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1 **Caleta el Cobre 022 martian meteorite: increasing nakhlite diversity**

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20

21 **Abstract**

22 Caleta el Cobre (CeC) 022 is a martian meteorite of the nakhlite group, showing an unbrecciated
23 cumulate texture, composed mainly of clinopyroxene and olivine. Augite shows irregular core
24 zoning, euhedral rims and thin overgrowths enriched in Fe relative to the core. Low-Ca pyroxene
25 is found adjacent to olivine. Phenocrysts of Fe-Ti oxides are titanomagnetite with exsolutions of
26 ilmenite/ulvöspinel. Intercumulus material consists of both coarse plagioclase and fine-grained
27 mesostasis, comprising K-feldspars, pyroxene, apatite, ilmenite, Fe-Ti oxides and silica. CeC 022
28 shows a high proportion of martian aqueous alteration products (iddingsite) in olivine (45.1 vol%
29 of olivine) and mesostasis. This meteorite is the youngest nakhlite with a distinct Sm/Nd
30 crystallization age of 1.215 ± 0.067 Ga. Its ejection age of 11.8 ± 1.8 Ma is similar to other nakhlites.
31 CeC 022 reveals contrasted cooling rates with similarities with faster-cooled nakhlites, such as
32 Northwest Africa (NWA) 817, NWA 5790 or Miller Range (MIL) 03346 nakhlites: augite irregular
33 cores, Fe-rich overgrowths, fine-grained K-feldspars, quenched oxides and high rare earth element
34 content. CeC 022 also shares similarities with slower-cooled nakhlites, including Nakhla and NWA
35 10153: pyroxene modal abundance, pyroxenes crystal size distribution, average pyroxene size,
36 phenocryst mineral compositions, unzoned olivine and abundant coarse plagioclase. Moreover,
37 CeC 022 is the most magnetic nakhlite and represents an analogue source lithology for the strong
38 magnetization of the martian crust. With its particular features, CeC 022 must originate from a
39 previously unsampled sill or flow in the same volcanic system as the other nakhlites, increasing
40 martian sample diversity and our knowledge of nakhlites.

41 **1. Introduction**

42 Nakhlites constitute a meteorite group originating from Mars (McSween 1985; Treiman et
43 al. 2000). To date, the group comprises 21 meteorites: 10 from Northwest Africa (NWA) including
44 the recently approved NWA 12542, 7 from Antarctica, 1 fall (Nakhla) and 2 other finds from
45 various provenances, with the addition of Caleta el Cobre (CeC) 022. After pairing this number is
46 reduced to 12 meteorites (Udry et al., 2012; Udry and Day, 2018; Meteorite Bulletin Database,
47 <https://www.lpi.usra.edu/meteor/>). Nakhlites all share broadly similar textures, whole-rock
48 compositions and mineralogy. In addition, they all present a crystallization age within the same
49 range from 1.26 to 1.42 Ga, consistent over four different dating techniques $^{87}\text{Rb}/^{86}\text{Sr}$, $^{147}\text{Sm}/^{144}\text{Nd}$,
50 $^{39}\text{Ar}/^{40}\text{Ar}$ and $^{176}\text{Lu}/^{176}\text{Hf}$ (from data compilation of Nyquist et al., 2001, Shih et al., 2010,
51 Korochantseva et al., 2011, Cohen et al., 2017, and Udry and Day 2018). Lastly, cosmogenic
52 isotopes yield an exposure age of 9 to 13 Ma, which indicates a single ejection event, strengthening
53 their genetic relationship (Herzog and Caffee, 2013, Nyquist et al. 2001, Wieler et al., 2016, Cohen
54 et al., 2017).

55 Nakhlites are clinopyroxene-rich cumulates, composed mainly of euhedral clinopyroxene
56 and less abundant olivine. The intercumulus materials consist either of coarse plagioclase, or a fine-
57 grained mesostasis composed of various amounts of fayalite, feldspar, low-Ca pyroxene,
58 titanomagnetite, sulfide, silica and apatite (e.g., Treiman, 2005). Post-eruption alteration phases
59 have also been reported in those meteorites, such as iddingsite which is described as a hydrous
60 iron-magnesium silicate (Bunch and Reid 1975; Bridges and Grady 2000; Treiman and Irving,
61 2008; Hallis and Taylor, 2011; Lee et al., 2018), as clay minerals referred to as serpentine (Irving
62 et al., 2002) or smectite (Treiman et al., 1993; Gillet et al., 2002), or as jarosite (Noguchi et al.,
63 2009). These secondary phases formed through hydrothermal alteration on Mars (Changela and

64 Bridges 2010) and have been tentatively dated around 633 ± 23 Ma with K-Ar system (Borg and
65 Drake, 2005 and references therein).

66 The formation of the nakhlites started with low-degree partial melting of a long-lived, trace
67 element depleted source that occurred at around 1.34 Ga (McCubbin et al., 2013). The cumulus
68 part of the nakhlites with coarse subhedral crystals suggests slow growth over a long period of time
69 (Lentz et al., 1999) at low pressure in a shallow magma chamber (Day et al. 2006). In contrast, the
70 intercumulus material is composed either of fine-grained mesostasis attesting a relatively fast
71 cooling, or coarse plagioclase crystallized under lower cooling rates. Indeed, nakhlites final cooling
72 would have taken place at the martian surface over short periods of time from 5 to 184 days (Day
73 et al., 2006; Udry and Day, 2018), with cooling rates from 1 to 6°C/hr based on zoning in olivine
74 (Sautter et al., 2002) and intercumulus crystals surface area to volume (Hammer and Rutherford
75 2005). After emplacement as sill or flow, crystal gravitational settling and compaction, as well as
76 dynamical sorting of a phenocryst-rich magmatic mush (Corrigan et al., 2015; Day et al., 2006)
77 and flow dynamics (Daly et al., 2019) have shaped the final texture of nakhlites, based on
78 petrographic textural evidence.

79 The comparable crystallization ages and ejection ages of nakhlites led to the hypothesis of
80 a suite of igneous rocks originating from a single martian volcanic system. The most widely
81 accepted model for the petrogenesis of nakhlites is that the cumulus phases crystallized at depth
82 followed by emplacement at the surface as a flow or shallow subsurface sill (Bunch and Reid 1975;
83 Treiman 1986; Lentz et al., 1999; Mikouchi et al., 2003; Day et al., 2006; Mikouchi et al., 2012,
84 2017; Richter et al., 2016; Udry and Day, 2018). All previously studied nakhlites present slight
85 differences in their modal abundances, intercumulus material and minor minerals, mineral
86 chemistry, crystallinity, rare earth element (REE) enrichment and Ar-Ar ages, which led to various
87 models of emplacement. Based on differences in mineralogy and textures, a stratigraphy of

88 nakhrites within a single cumulate pile was first proposed (Treiman, 1986; Mikouchi et al., 2003,
89 2012; Treiman, 2005; Bridges and Warren, 2006; Day et al., 2006; McCubbin et al., 2013; Corrigan
90 et al., 2015). Based on textural and chemical compositional differences among the various
91 nakhrites, Jambon et al. (2016) proposed nakhrite emplacement in at least three different flows
92 piled on top of one another from Miller range (MIL) 03346 at the uppermost part of the pile, down
93 to Northwest Africa (NWA) 5790. Based on geochemistry and petrography, Balta et al. (2017)
94 proposed another model with one body chamber and two separated lobes. Cohen et al. (2017)
95 proposed a model of multiple flows using geochronological evidences supporting a layered
96 volcanic sequence with at least four discrete eruptive events. Recently, Udry and Day (2018)
97 proposed in a comprehensive study that nakhrites originated from at least six different lava flows
98 or sills, distributed around one or several volcano vents.

99 Caleta el Cobre 022 was approved by the Meteoritical Society Nomenclature Committee in
100 April 2018. It is the 11th unpaired nakhrite (the only one from Atacama Desert) and brings new
101 information to help constrain the diversity, stratigraphy and emplacement conditions of the nakhrite
102 group. This paper is focused on the description of this new nakhrite (petrology, mineralogy and
103 geochemistry) and its comparison with previously studied nakhrites. We observe that CeC 022
104 combines evidence for both high and low cooling rates, with features never observed in other
105 nakhrites. This will open discussion on the crystallization conditions of cumulus minerals, the
106 crystallization sequence and the relation of CeC 022 with the other nakhrites.

107 **2. Analytical Methods**

108 Petrographic observation and elemental micro-analyses were performed on three thick polished
109 sections (referred as A, B and D in the following) and one thin section (C).

110 **2.1. Modal abundances**

111 Modal abundances of major crystals were measured on all sections using hand-drawn grain
112 boundaries in *Adobe Illustrator* (see section 2.1.2.) and pixel counting using the *ImageJ* software.
113 The modal abundances of minor phases were measured using composite elemental maps obtained
114 by Scanning Electron Microscopy (SEM) and Energy Dispersive Spectrometry (EDS) and using
115 pixel counting with the *ImageJ* software. The average modal abundance of CeC 022 is calculated
116 from the 4 sections, all showing similar surface sizes of about 1.5 cm². The total surface analyzed
117 in the 4 sections is 7.1 cm². All phases were characterized using a Leica DM2500P petrographic
118 microscope. An Olympus BH₂ microscope operating in reflected light was used to first characterize
119 Fe-Ni-Cu sulfides. The abundance of sulfides was determined on section B by point counting (2000
120 points, 50 μm step).

121 **2.2. Quantitative textural analyses**

122 Crystal Size Distribution (CSD) is a quantitative tool defining kinetic textures of a specific
123 mineral crystal population and provides a plot representing the relationship between the natural
124 logarithm of the population density and crystal size (Cashman and Marsh, 1988; Marsh 1988,
125 1998). Spatial distribution pattern (SDP) analyses investigate ordering and framework of a crystal
126 population and constrain processes of rock formation, such as compaction, flow settling and grain
127 orientation. Spatial distribution pattern analyses are based on both R-value and the porosity or melt
128 content of the thin section, corresponding to the inverse of pyroxene content in the thin section
129 (Jerram et al. 1996, 2003). The R-value is defined- as: $R = 2\sqrt{\rho} \sum r/N$, where ρ is the observed
130 population density, r is the distance between the center of a crystal and the center of its nearest
131 neighboring crystal and N is the total number of crystals measured (Jerram et al. 1996, 2003). The
132 R-value represents the ratio of the observed versus predicted mean distribution of nearest neighbor

133 distances using the density of the observed distribution. The CSD and SDP analyses were
134 conducted for the pyroxene population of the thin section (C). Pyroxene grains were hand-drawn
135 in *Adobe Illustrator* using mosaic pictures in transmitted and reflected light, X-ray elemental maps
136 and an optical microscope to distinguish grain boundaries. We used the *ImageJ* software to obtain
137 the area of each crystal, the total area of the section, the major and minor axes of ellipses, as well
138 as the centroid location. We followed the same quantitative textural methods as the one described
139 in Udry and Day (2018).

140 **2.3. Physical properties**

141 Grain density was measured with a Quantachrome helium steropycnometer at Centre Européen
142 de Recherche et d'Enseignement en Géosciences de l'Environnement (CEREGE). Magnetic
143 susceptibility was quantified with MFK1 or KLY2 instruments depending on sample size. Curie
144 temperatures were estimated by measuring the variation of susceptibility with temperature with a
145 CS3 furnace on the MFK1 susceptibility meter. Saturation magnetization was determined with a
146 Princeton model 2900 Micromag vibrating sample magnetometer. All magnetic measurements
147 were performed at CEREGE.

148 **2.4. Mineral major and minor elements**

149 Backscattered electron (BSE) images, microanalyses by energy dispersive spectroscopy (EDS)
150 and chemical maps were acquired with a Hitachi S3000-N SEM operated at 15 kV and equipped
151 with a Bruker X-ray X-Flash detector and a Spirit analyzer at CEREGE, France. Semi-quantitative
152 analyses used a variety of natural and synthetic standards. Additional field emission gun (FEG)
153 SEM images were obtained on a ZEISS Gemini 500 SEM operated at accelerating voltages from
154 5 to 15 kV at Centre Pluridisciplinaire de Microscopie Electronique et de Microanalyse (CP2M),
155 Marseille. Compositions of sulfides were assessed at the National Museum of Natural History

156 MNHN (Paris) using a Tescan VEGA II LSU SEM operated at 15kV and equipped with an SD3
157 (Bruker) EDS detector.

158 Quantitative major element analyses were conducted using an electron microprobe (EMP)
159 Cameca SX100 at UPMC CAMPARIS facility. The operating conditions were 15 kV accelerating
160 voltage with a current of 10 nA and a counting time of 30 s, with a focused beam (1 μm) for
161 pyroxene, olivine, plagioclase and apatite, and defocused beam (5 μm) for fine-grained mesostasis
162 and alteration phases. Both natural and synthetic standards were used for calibration: albite for Na;
163 anorthite for Al; apatite for P; diopside for Mg, Si, Ca; orthoclase for K; pyrite for S; MnTiO_3 for
164 Mn and Ti; Cr_2O_3 for Cr; Fe_2O_3 for Fe; and NiO for Ni. The detection limits are (in ppm): 445 for
165 Na, 466 for Al, 714 for Mg, 669 for Si, 605 for Ca, 602 for K, 392 for S, 1044 for Mn, 419 for Ti,
166 496 for Cr, 1271 for Fe, and 1331 for Ni.

167 Sulfide microphases were analyzed at 15 kV accelerating voltage by a PhiRhoZ EDS
168 standardless procedure. Accuracy of our SEM analytical procedure analyses was checked by
169 repeated analysis of a Del Norte County-DNC (Oregon, USA) natural troilite (stoichiometric FeS),
170 a well-known terrestrial troilite occurrence previously studied by X-ray diffraction and repeatedly
171 analyzed by EMPA and SEM (Chevrier et al., 2011, Lorand et al. 2012, Gattaceca et al., 2013).
172 Gattaceca et al. (2013) shown that the SEM analyses of DNC troilite perfectly match the
173 composition of stoichiometric troilite (FeS). Our own analyses yielded a mean atomic Fe+Ni/S
174 ratios (metal-to-sulfur ratios) of 1.0 ± 0.01 (Figure S2). DNC troilite was used as external
175 calibration standard and analyzed every 10 to 15 analyses to evaluate potential drifts on metal-to-
176 sulfur ratios of CeC 022 pyrrhotites.

177 **2.5. Major and trace elements bulk analyses**

178 From 200 mg of powdered sample of CeC 022 using an agate mortar, ~50 mg was digested
179 with a 4:1 mixture of lithium metaborate and tetraborate. After heating in a graphite crucible for
180 10 minutes at 1000°C, the bead was dissolved in 5% HNO₃. After dilution, to control possible
181 instrumental drift, the solution was doped with In for trace elements measurements using an Agilent
182 7700 inductively coupled plasma mass spectrometry (ICP-MS) and Y for major elements
183 measurements using a Thermoscientific iCA inductively coupled plasma optical emission
184 spectroscopy (ICP-OES) at Laboratoire G-Time (Université Libre de Bruxelles-ULB). In both
185 cases, the calibration curves were obtained on synthetic standards and reproducibility was checked
186 using USGS standard (BHVO-2). For major elements, the relative standard deviation (RSD) is
187 systematically better than 2% (except Na, 5 %), while for trace elements, it is better than 10%,
188 except for Lu (46%) and U (33%).

189 **2.6. Noble Gases**

190 All noble gas isotopes of He to Xe were extracted in one step and measured at ETH Zurich
191 in a 50.9 mg sample by fusion at ~1700 °C in a Mo crucible according to the established procedures
192 detailed by Riebe et al. (2017). Blank corrections are for ³He, ⁴He, ²¹Ne, ²²Ne < 0.4 %, for ²⁰Ne,
193 ³⁶Ar, ³⁸Ar, ⁴⁰Ar, ⁸⁴Kr and ¹³²Xe 3.3 %, 13 %, 2.1 %, 2.5 %, 8 % and 1.8 %, respectively.

194 **2.7. Isotopic measurements**

195 Mineral separation was carried out at Laboratoire G-Time at Université Libre de Bruxelles.
196 The sample has been crushed and sieved. A first separation using the 150-250 µm fraction was
197 made using dense liquid (methyl iodine) and a second one using a Frantz magnetic separator. The
198 different fractions were then crushed and dissolved in a 1/3 mixture of sub-boiled concentrated HF
199 and HNO₃ and left 2 days on hot plate at 120°C using high-pressure vessels. After drying, the

200 fractions were re-dissolved using 12N HCl until the solutions were clear. A 5% aliquot was
201 removed from the initial dissolution and spiked using ^{150}Sm - ^{148}Nd spike. REE were first purified
202 using cationic resin, being collected in HCl 6N. For both spiked and unspiked cuts, the REE were
203 further separated using HDEHP resin by increasing the HCl molarity for collecting Nd and spiked
204 Sm. The whole rock fraction was further purified for ^{142}Nd measurements. First, Ce was removed
205 by using HDEHP resin after oxidation of Ce using NaBrO_3 . The added Na was then cleaned out
206 using cationic resin, similarly to the first purification step. All fractions but the whole rock one,
207 spiked and unspiked, were measured on the HR-MC-ICP-MS Nu-Plasma 2 using an Aridus
208 desolvator, using 0.05N HNO_3 . Spike deconvolution was performed off-line by iteration as
209 described in Debaille et al. (2007). The unspiked fraction were corrected for a Rennes standard
210 value of $^{143}\text{Nd}/^{144}\text{Nd}$ of 0.511961 (Chauvel and Blichert-Toft, 2001). Total reproducibility
211 including standards ran at 20 and 5 ppb (n=24) is better than 50 ppm (2σ). The whole rock fraction
212 was measured on the Thermoscientific Triton Plus at ULB in multidynamic mode as in Debaille et
213 al. (2007). The Rennes standard (n=3) gave a value of 0.511957 ± 0.000002 and no correction was
214 required for comparison with the data set measured on the MC-ICP-MS.

215 Oxygen isotopes were measured at the Stable Isotopes Laboratory of CEREGE. Molecular
216 oxygen was extracted using laser fluorination. Aliquots of 1.5 mg obtained from a bulk 12 mg
217 powdered samples were heated with a 30 W CO_2 IR laser in the presence of 100 hPa of BrF_5 . The
218 released gas was purified through two cryogenic nitrogen traps and one heated KBr trap. Molecular
219 oxygen was trapped for 10 mn in a molecular sieve cooled at -196°C . The gas was then expanded
220 at 100°C and passed through a molecular sieve cooled at -114°C slush for 5 mn to refreeze possible
221 interfering gases. The gas was then trapped again for 5 mn in the molecular sieve cooled at -196°C ,

222 then expanded again at 100°C directly into the bellow of a dual-inlet Thermo-Finnigan Delta Plus
223 mass spectrometer.

224 The isotopic compositions are expressed in standard δ -notation, relative to Vienna standard
225 mean ocean water (VSMOW): $\delta^{18}\text{O} = (^{18}\text{O}/^{16}\text{O})_{\text{sample}} / (^{18}\text{O}/^{16}\text{O})_{\text{VSMOW}} - 1$ and $\delta^{17}\text{O} = (^{17}\text{O}/^{16}\text{O})_{\text{sample}}$
226 $/ (^{17}\text{O}/^{16}\text{O})_{\text{VSMOW}} - 1$ (expressed in ‰), $\Delta^{17}\text{O}$, also expressed in ‰, is defined as $\delta^{17}\text{O} - \lambda \times 0.5247$
227 $\delta^{18}\text{O}$. For comparison with previous data obtained on Martian meteorites (Franchi et al. 1999), we
228 use $\lambda = 0.52$. Measured $\delta^{18}\text{O}$ and $\delta^{17}\text{O}$ values of the samples were corrected daily using the
229 laboratory quartz standard itself calibrated against the international standard NBS28 ($\delta^{18}\text{O} = 9.60$
230 ‰, $\delta^{17}\text{O} = 0.52 \times \delta^{18}\text{O} = 4.992$ ‰). Reproducibility (1 sigma) of the quartz laboratory standard are
231 0.119 ‰ for $\delta^{18}\text{O}$, 0.061 ‰ for $\delta^{17}\text{O}$, and 0.024 ‰ for $\Delta^{17}\text{O}$ ($n = 63$) with a $\Delta^{17}\text{O}$ of -0.015 ± 0.024 ‰.

232 **2.8. Secondary alteration phases analyses**

233 Infrared (IR) spectroscopy was performed at the Institut de Planétologie et d'Astrophysique
234 de Grenoble (IPAG) using the SHADOWS instrument (Potin et al., 2018). The instrument was
235 used in normal standard mode as well as in a μ -beam mode enabling measurements of IR
236 reflectance spectra of areas of the order of 3 mm^2 . Spectra were measured under nadir incidence
237 and with an observation angle of 30° . Spectra were normalized to infragoldTM and spectralonTM.
238 The Raman experiments were performed at Laboratoire de Géologie de Lyon (ENS Lyon – France)
239 using a Labram HR800 spectrometer equipped with a Nd ion laser providing 532 nm excitation.
240 The laser spot is $<2\mu\text{m}$ with the used x100 microscope objective. Individual spectra were acquired
241 at various locations in veins of iddingsite, optically localized throughout the sections A and B of
242 CeC 022.

243 **3. Results**

244 **3.1. General description and physical properties**

245 Caleta el Cobre 022 was found casually by tourists traveling in the Atacama Desert, Chile,
246 in November 2016, lying at the surface in the CeC dense collection area, which is part of the densest
247 meteorite collection area in hot deserts (Hutzler et al. 2016). It is a single stone, almost fully crusted.
248 The fusion crust has a few cracks and the typical glassy gloss of achondrites and was identified as
249 such by R. Martinez from the Museo del Meteorito in San Pedro de Atacama (Chile), when he
250 observed the stone a few days after the find. The type specimen of 21g was cut from the main mass
251 of 450 g at the Museo del Meteorito and later given to CEREGE, where the classification was done.
252 Cut face reveals a dark interior with a few fractures penetrating the whole stone. Pinkish-orange
253 mm-sized minerals (altered olivine) are visible over a dark green background (Fig. 1).

254 Grain density, measured on a 15.42 g sample, is $3.342 \pm 0.002 \text{ g/cm}^3$. Magnetic
255 susceptibility, measured on 8 samples ranging from 19.7 mg to 340 g, is homogeneous over this
256 size range with a coefficient of variation of 3.4% and an average mass weighed value of $8.41 \cdot 10^{-6}$
257 m^3/kg . Saturation magnetization is $1.51 \text{ Am}^2/\text{kg}$. Saturation remanent magnetization is $4.59 \cdot 10^{-1}$
258 Am^2/kg . Two Curie temperatures are observed at 525°C and 585°C (Fig. S1).

259 **3.2. Petrography**

260 The meteorite shows an unbrecciated cumulate texture with euhedral elongated augite
261 crystals (56.5 vol%) and a lower amount of olivine (4.2 vol%), set in an intercumulus made of
262 coarse plagioclase (19.4 vol%) and fine-grained mesostasis (15.9 vol%).

263 **3.2.1. Phenocrysts**

264 ***Pyroxene***

265 Clinopyroxene is the major mineral with 52.3-60.3 vol% (average 56.5 vol% among four
266 sections) (Table 1) and measure on average 0.27 mm (up to 2 mm). On BSE images, they present
267 an irregular zoning with rounded cores and a Fe-rich euhedral rim (Fig. 2), surrounded by a thin,
268 sharply delimited Fe-enriched overgrowth, when in contact with the mesostasis. The zoning can be
269 discontinuous or gradual, depending on the orientation of the grains. The sharp overgrowth, usually
270 10-15 μm wide, gets thinner or absent when in contact with other phenocrysts (pyroxene,
271 plagioclase, titanomagnetite, or olivine). Some rounded clinopyroxene ($\sim 100 \mu\text{m}$) can be found
272 included inside olivine crystals. Also, smaller clinopyroxene ($\sim 10 \mu\text{m}$) with Fe-oxides are found
273 in linear planes of inclusions, or exsolutions, in the olivine (Fig. 3). Low-Ca pyroxenes (up to 100
274 μm , Fig. 2) were frequently observed adjacent to olivine crystals, sometimes presenting augite
275 irregular cores. Based on grain boundary geometries, low-Ca pyroxene appears to have crystallized
276 after augite rim and olivine.

277 ***Olivine***

278 Olivine crystals represent 2.0-8.9 vol% (average 4.2 vol% among four sections) with a
279 maximum measured grain size of 2.5 mm and an average length of 0.66 mm (Table 1). They do
280 not display a euhedral shape, rather a resorbed border. They do not show any zoning on BSE images
281 (Fig. 4). They display large melt inclusions up to 100 μm , rich in Na-K feldspars and apatite.
282 Olivine crystals present high abundance (45.1 vol%) of aqueous alteration products (iddingsite)
283 found in crosscutting veins and sometimes extending to the edge of the olivine and the mesostasis
284 around the olivine. These secondary alteration products are described in more details in part 3.4.

285 ***Plagioclase***

286 Caleta el Cobre 022 contains coarse plagioclase with a maximum measured grain size of 1
287 mm, representing in average 19.4 vol% (17.1-24.2 vol% among three sections) of the meteorite
288 (Table 1). The association of augite cumulate crystals with coarse plagioclase forms a doleritic
289 texture.

290 ***Opaque minerals***

291 Intergranular Fe-Ti oxides are abundant (2.6 vol% among three sections) (Table 1) and up
292 to 1 mm size. On BSE images, they show fine-scale (0.5 μm wide) Fe-Ti oxides (ilmenite or
293 ulvöspinel) exsolutions in what appears to be titano-magnetite (Fig. 5, right). At higher resolution
294 (FEG-SEM), finer nanometric (\sim 1-3 nm) exsolutions can be observed in between these micron
295 sized Fe-Ti oxides exsolutions (Fig. 5, left). These oxides phenocrysts show inclusions of augite
296 as well as melt inclusions.

297 Caleta el Cobre 022 is sulfide-rich (0.15 ± 0.05 vol.%), with sulfides mostly under the form
298 of pyrrhotite. Those intercumulus pyrrhotites are frequently associated with magnetite without
299 ilmenite exsolution. A continuum of textural associations between pyrrhotite and magnetite was
300 observed such as discontinuous or continuous magnetite rims on pyrrhotite (Fig. 6A-B), residual
301 pyrrhotite inclusions in coarse granular magnetite (Fig. 6B), and rounded inclusions of magnetite
302 inside the pyrrhotite coexisting with coarser magnetite grains (Fig. 6C). Plagioclase- or
303 clinopyroxene-hosted pyrrhotite are free from magnetite. Chalcopyrite was observed in only three
304 grains as marginal blebs (Fig. 6D). Pentlandite flames inside pyrrhotite are very scarce.

305 **3.2.2. Mesostasis**

306 Mesostasis in CeC 022 represents on average 15.9 vol% (12.7-16.0 vol% among three
307 sections) of the meteorite (Table 1). Mesostasis pockets contain mainly K-feldspar, silica, dendritic

308 Ti-magnetite, low-Ca pyroxene, apatite and ilmenite. Most oxides in the mesostasis are skeletal Ti-
309 magnetite. Apatite is mainly found as small elongated euhedral crystals embedded in the feldspar.
310 Adjacent to the olivine grains, apatite displays larger sizes (up to 700 μm long) in association with
311 pigeonite rims (Fig. 7). Fine-grained pyrrhotite is hosted in the microcrystalline mesostasis, along
312 with skeletal Ti-magnetite.

313 **3.2.3. Terrestrial weathering**

314 Terrestrial weathering is minimal on CeC 022. The meteorite shows a fresh black fusion
315 crust with few alteration products (Fig. 1). Caleta el Cobre 022 is a highly fractured rock (Fig. 1,
316 Fig. 3, Fig. 4). On BSE images, we observed that those fractures are devoid of terrestrial alteration
317 products (e.g., Fig. 1, Fig. 3, Fig. 4), such as evaporite minerals that are classically encountered in
318 meteorites weathered in hot deserts (Bland et al. 2006). Terrestrial weathering is observed only
319 close to the fusion crust and is minimal. It is found as partly altered sulfides and pyrrhotite cracks
320 invaded by iron oxyhydroxides.

321 **3.3. Quantitative textural analyses**

322 We measured a total of 2103 pyroxene grains and obtained a R^2 of 0.88 (Table S1.). Caleta
323 el Cobre 022 pyroxene population displays a best habit ratio of 1.00:1.25:1.80 and shows a CSD
324 slope of $-5.31 \pm 0.09 \text{ mm}^{-1}$, with an intercept of 6.04 ± 0.05 (Fig. 8). A downturn for the smaller
325 sizes ($<0.03 \text{ mm}$) is observed. Spatial distribution analyses show that the pyroxene population of
326 CeC 022 has an R-value of 1.277 and falls within the cluster touching framework, similar to the
327 other nakhlites (Fig. S3, Table S1).

328 **3.4. Geochemistry**

329 **3.4.1. Phenocrysts**

330 *Pyroxenes*

331 Electron microprobe analyses reveal augite cores range in composition from $\text{En}_{32}\text{Fs}_{30}\text{Wo}_{38}$
332 to $\text{En}_{36}\text{Fs}_{24}\text{Wo}_{40}$ (average $\text{En}_{33}\text{Fs}_{27}\text{Wo}_{40}$) and $\text{FeO}/\text{MnO} = 35\text{-}45$ with $\text{Mg\#} [=100 \times (\text{Mg}/(\text{Fe}^{2+} +$
333 $\text{Mg}))] = 55$ (Table 2, Fig. 9). They are surrounded by augite richer in Fe of average composition
334 $\text{En}_{31}\text{Fs}_{29}\text{Wo}_{40}$ ($\text{Mg\#} \sim 52$). Microprobe traverse analyses across augite crystal reveal that Al, Ti and
335 Mn anticorrelate with Mg#. In contrast, Ca and Cr show a correlation with Mg#, increasing from
336 the rim to the core of the augite, and Na shows no variation throughout the crystal. Augite crystals
337 have relative low Al, Ti and Na concentration ($\text{Al}_2\text{O}_3 = 1.1 \pm 0.1 \text{ wt\%}$; $\text{TiO}_2 = 0.4 \pm 0.1 \text{ wt\%}$; Na_2O
338 $= 0.3 \pm 0.1 \text{ wt\%}$), comparable to terrestrial pyroxene crystallized at low pressure (Harvey and
339 McSween, 1992). The thin overgrowth around the augite rim shows Fe-rich compositions going
340 up to $\text{En}_{30}\text{Fs}_{31}\text{Wo}_{39}$ (Mg\# of ~ 49). This overgrowth is also more enriched in MnO (up to 0.6 wt%),
341 Al_2O_3 (up to 1.5 wt%) and TiO_2 (up to 0.6 wt%), but poorer in Na and Ca. In the olivine, planes of
342 inclusions contain Fe-oxides and augites of composition $\text{En}_{31}\text{Fs}_{26}\text{Wo}_{43}$, similar to cumulus augite
343 irregular cores (Fig. 3).

344 Low-Ca pyroxenes differ from augite by their texture as well as their composition. They
345 are found adjacent to olivine crystals (Fig. 2, Fig. 7). Their augite core compositions are similar to
346 those previously described, but the low-Ca pyroxene overgrowth around the cores has a
347 composition of $\text{En}_{36}\text{Fs}_{60}\text{Wo}_4$ ($\text{Fe}/\text{Mn} = 28$; $\text{Mg\#} 38$) (Table 2, Fig. 9). The low-Ca overgrowths
348 show an enrichment in Fe, Mn and Mg and lower Ti, Al and Na, compared to the augite pyroxene.

349 *Olivine*

350 Olivine compositions show no zoning, with an average composition of $\text{Fo}_{30.3}$ ($\text{FeO}/\text{MnO} =$
351 49 ± 3) and minor intragrain variations of $\text{Fo}_{27.3}$ to $\text{Fo}_{32.5}$ (Table 3, Fig. 10).

352 ***Plagioclase***

353 The coarse plagioclase grains found in the meteorite have a homogenous composition of
354 $An_{29.7\pm 4.3}Ab_{67.2\pm 4.0}Or_{3.1\pm 0.5}$ (Fig. 11).

355 **3.4.2. Mesostasis**

356 Mesostasis in CeC 022 is dominated by K-feldspar (Fig. 11) with compositions in the range
357 $An_{6.7}Ab_{84.6}Or_{8.7}$ to $An_{4.5}Ab_{60.4}Or_{35.1}$. Also, some low-Ca pyroxene can be found in the mesostasis
358 with compositions of $En_{31}Fs_{65}Wo_4$ (Fe/Mn = 30) (Table 2). These compositions are similar to the
359 ones of low-Ca pyroxenes rims found next to olivine crystals. Apatite is also observed in the
360 mesostasis with Cl-rich compositions (Table S2).

361 **3.4.3. Bulk rock geochemistry**

362 Major elements in CeC 022 are reported in Table 4, together with other nakhlites. It has a
363 high bulk Mg# of 53.8, with a low FeO/MnO of 38.0 (Table 4). Compared to the other nakhlites,
364 the bulk composition of CeC 022 has low CaO contents (12.8 wt%) but high MgO and Al₂O₃ (12.2
365 wt% and 3.32 wt% respectively). Sodium and potassium content are low (1.13 wt% and 0.25 wt%).

366 Rare earth element compositions are reported in Table 4 and Fig. 13 and similar to other
367 nakhlites. Conversely an enrichment in Ce was measured in CeC 022 that is not described in other
368 nakhlites. Bulk $(La/Lu)_{CI}$ is 4.4, within the range of nakhlites (Udry and Day, 2018).

369 **3.4.4. Oxygen isotopes**

370 Analyses of three 1.5 mg aliquots gave $\delta^{18}O = 4.683\text{‰} \pm 0.428\text{‰}$, $\delta^{17}O = 2.683\text{‰} \pm$
371 0.210‰ , and $\Delta^{17}O = 0.248 \pm 0.013\text{‰}$ (mean $\pm 1\sigma$).

372 **3.5. Alteration products**

373 **3.5.1. Petrography**

374 The main martian alteration products in CeC 022 occur in olivine phenocrysts as a reddish-
375 orange phase in optical microscopy in transmitted light. This material is found mainly as veins
376 along cracks in the olivine with Fe-oxides and sulfides situated at the center of the veins, but it can
377 also be located as patches in the mesostasis. This alteration product has already been described in
378 nakhlites under various names such as iddingsite, smectite, illite, saponite, siderite, sulfates (Bunch
379 et al., 1975; Reid and Bunch, 1975; Boctor et al., 1976; Gooding et al. 1991; Gooding, 1992;
380 Treiman et al., 1993; Bridges et al., 2000; Gillet et al. 2002). Caleta el Cobre 022 contains a high
381 abundance of iddingsite, replacing 45.1 vol% of the crystals, found mainly in crosscutting veins
382 inside olivine. A noteworthy feature of CeC 022 is the occurrence of pyrrhotite and Fe-oxides
383 clusters or stringers inside iddingsite veins across olivine (Fig. 4, Fig. 6, Fig. 14). Also, a few
384 deeply altered former sulfides were observed in the vicinity of olivine grains, coexisting with
385 iddingsite aggregates. Such ovoid grains are made from highly porous iron oxyhydroxides-like
386 phases showing detectable amount of sulfur (EDS spectra) (Fig. 6F).

387 Backscattered electron images reveal two types of iddingsite found in association with each
388 other (Fig. 14). Energy dispersive spectrometry (EDS) and EMP analyses show that a brighter type
389 of iddingsite in contact with olivine in the BSE image is enriched in Fe and Mn (~52 wt% FeO)
390 and is poorer in Si (~27 wt% SiO₂), Mg and K, compared to the darker type (~30 wt% FeO and
391 ~50 wt% SiO₂) (Table 5). Low totals (<90 wt%) of the EMP analyses suggest that these phases are
392 hydrated. The Fe-rich type of iddingsite has a Raman spectrum exhibiting well defined and narrow
393 bands in contrast to the spectrum related to the Si-rich type. The Fe-rich iddingsite is also
394 associated, at the micron scale, with a few oxides, possibly hematite or magnetite and possibly
395 carbonates or pyroxene (Fig. S4, right). The wide and poorly defined Raman bands exhibited by
396 the Si-rich iddingsite reflect a degree of crystallization lower than in Fe-rich type (Fig. S4, right).
397 Raman spectroscopy reveals that the Si-rich type of iddingsite is a poorly crystallized hydrated

398 assemblage of minerals (Fig. S4, right). Both Raman and IR (Fig. S4) spectroscopy show weak
399 bands of carbonate and/or phyllosilicate and H₂O bending and H₂O + OH stretching modes,
400 evidencing the hydration of these alteration products. Raman spectra show that each type of
401 iddingsite is homogenous throughout the sample, independent of the location in the meteorite.

402 **3.6. Geochronology**

403 **3.6.1. Sm-Nd systematics and crystallization age**

404 The crystallization age of CeC 022 has been measured at 1215 ± 67 Ma using a Sm-Nd
405 isochron using the Isoplot software (MSWD of 1.2). The initial $^{143}\text{Nd}/^{144}\text{Nd}$ using Isoplot is
406 0.511600 ± 0.000069 (Table S3). Expressed in 10^4 deviation to the chondritic average, this
407 corresponds to an $\epsilon^{143}\text{Nd(i)}$ of $+10.4 \pm 1.4$, similar to the $\epsilon^{143}\text{Nd(i)}$ measured on the bulk rock
408 fraction, of $+10.5 \pm 0.1$. Concerning the ^{142}Nd systematics, expressed in 10^6 deviation to the
409 average modern Earth, the $\mu^{142}\text{Nd}$ of CeC 022 is equal to $+56 \pm 3$.

410 **3.6.2. Ejection age**

411 Results of the noble gas measurements are given in Table S4. The measured He and Ne
412 isotope ratios indicate that there is no trapped component in He and Ne. Hence, the measured ^3He ,
413 ^{21}Ne and $^{21}\text{Ne}/^{22}\text{Ne}$ are entirely cosmogenic. The (inverse) shielding indicator $(^{21}\text{Ne}/^{22}\text{Ne})_{\text{cosm}}$ of
414 0.831 ± 0.06 is low. This value suggests low shielding in a small meteoroid, which is often observed
415 for rare meteorites, particularly martian ones (Wieler et al. 2016). Also, the presence of some solar
416 cosmic ray (SCR)-derived Ne cannot be excluded (Wieler et al. 2016). However, using the Mg /
417 (Al + Si) vs. $^{21}\text{Ne}/^{22}\text{Ne}$ diagram (Wieler et al., 2016), SCR-Ne is most likely not predominant.
418 Nevertheless SCR-Ne could still cause some overestimation of the Cosmic Ray Exposure (CRE)
419 ages and underestimation of shielding. The $^{36}\text{Ar}/^{38}\text{Ar}$ ratio in CeC 022 is 0.773 and hence close to

420 the expected pure cosmogenic composition of ~ 0.65 (e.g., Wieler et al., 2016) and a correction for
421 trapped, e.g. atmospheric, ^{38}Ar would be in the order of 2-3%. Trapped ^{38}Ar is thus negligible
422 compared to the other uncertainties related to the determination of CRE ages.

423 Apart from $^{129}\text{Xe}/^{132}\text{Xe}$, the Kr and Xe isotopic compositions are consistent with a mixing
424 of air with — in the light isotopes — cosmogenic Kr and Xe. The presence of a small martian
425 trapped component is visible only in the $^{129}\text{Xe}/^{132}\text{Xe}$ ratio = 1.02 ± 0.01 . Similarly, the trapped
426 element ratio $^{36}\text{Ar}_{\text{tr}}/^{132}\text{Xe}$ (4.2 ± 1.3) is very low suggesting significant Xe from air, in agreement
427 with the Xe isotopes. The ratios $^{84}\text{Kr}/^{132}\text{Xe}$ (1.25 ± 0.05) and $^{129}\text{Xe}/^{132}\text{Xe}$ are consistent with
428 previous results for nakhlites and suggest essentially no trapped martian component but merely
429 “Earth fractionated atmosphere” (cf. Fig 3A, Cartwright et al. 2014). The ^3He , ^{21}Ne and ^{38}Ar CRE
430 ages of CeC 022 were determined with physical model predictions for cosmogenic noble gas
431 production (Leya and Masarik, 2009) based on the chemistry (Table 4) and the shielding conditions
432 constrained by the shielding parameter given above. As a somewhat arbitrary but safely
433 conservative assumption, we used radii ≤ 85 cm. Most “allowed” shielding depths, based on
434 cosmogenic $^{22}\text{Ne}/^{21}\text{Ne}$, were in the range 0 to 2 cm. We obtain ^3He , ^{21}Ne and ^{38}Ar CRE ages of
435 11.4 ± 0.6 Ma, 9.9 ± 1.6 Ma and 11.8 ± 2.7 Ma, respectively, where the uncertainties include the
436 range of allowed shielding conditions and experimental errors.

437 If using a model specifically created for very small shergottites (see Wieler et al., 2016, based
438 on the original model by Leya and Masarik, 2009), only one shielding condition would be allowed:
439 a perhaps more realistic 7 cm pre-atmospheric radius and 6.4-6.6 cm depth (which excludes any
440 contribution from of SCR). Using that model, the age would slightly increase, to 12.3 ± 0.2 Ma,
441 10.9 ± 0.1 Ma and 14.2 ± 3.3 Ma, respectively. We consider the later data set more reliable and our
442 preferred CRE combined age for CeC 022 is 11.8 ± 1.8 Ma.

443 **4. Discussion**

444 **4.1. Martian origin for CeC 022**

445 All petrographic, geochemical and mineral composition data confirm that CeC 022 is a
446 martian meteorite. Grain density of CeC 022 is $3.342 \pm 0.002 \text{ g/cm}^3$, similar to the average of 3.41
447 ± 0.07 on four nakhlites (Macke et al. 2011), suggesting on overall similar mineralogy. The mineral
448 abundances are in the range observed in nakhlites (Table 1) (Udry and Day, 2018), although image
449 analyses using pixel counting methodology on *ImageJ* gives errors in calculations of modal
450 abundance that are up to 5 vol% (Maloy and Treiman, 2007), due to the heterogeneous nature of
451 the nakhlites sample (Corrigan et al., 2015). Major element bulk compositions are in the range of
452 the nakhlites, with a high FeO (19.0 wt%), Al₂O₃ (3.32 wt%), MgO (12.2 wt%) and CaO (12.8
453 wt%) (Table 4), reflecting the abundance of cumulus minerals olivine and pyroxene. Sodium and
454 potassium content are low (1.13 wt% and 0.25 wt%). Likewise, bulk key ratios also confirm the
455 martian origin (Mg#= 53.8, FeO/MnO= 37.9; Na₂O/Al₂O₃ = 0.34; K/La = 450) (Table 4). Bulk
456 REE compositions show a pattern parallel to other nakhlites (Table 4, Fig. 13). Bulk (La/Lu)_{CI} is
457 4.4, within the range of nakhlites (Udry and Day, 2018).

458 Oxygen isotopic analyses, with $\Delta^{17}\text{O} = 0.248 \pm 0.013 \text{ ‰}$ (n=3) are close to the Mars
459 fractionation line (MFL), confirming a martian origin. In detail, although this value is compatible
460 with the MFL value $\Delta^{17}\text{O} = 0.28 \pm 0.07 \text{ ‰}$ (31 analyses on 11 meteorites) defined by Clayton and
461 Mayeda (1996), it is significantly lower than the more recent MFL value $\Delta^{17}\text{O} = 0.321 \pm 0.013 \text{ ‰}$
462 (34 analyses on 11 meteorites) defined by Franchi et al. (1999). This small discrepancy may
463 originate from the relatively small mass of the initial bulk sample used in our analyses (12 mg),
464 compared to the 50-100 mg used by Franchi et al. (1999).

465 Caleta el Cobre 022's crystallization age of $1.215 \pm 0.067 \text{ Ga}$ is close to other nakhlites,
466 even though it is slightly younger than Nakhla (Treiman, 2005, Cohen et al., 2017, Udry and Day,

467 2018). The ejection age of CeC 022 is measured around 11 Ma is similar to other nakhlites (Herzog
468 and Caffee, 2013). Therefore, CeC 022 was sampled by the same ejection event as the other
469 nakhlites.

470 **4.2. Petrogenesis**

471 **4.2.1. Cumulus crystallization**

472 In a first stage of crystallization, pyroxene cores (Mg# 55) grew in a cumulate layer (Jambon
473 et al., 2016) before the later stage of euhedral rim and overgrowth. These augite cores come from
474 early crystallization of relatively evolved Fe-rich mafic melt, probably in a magma chamber (Imae
475 et al., 2005; Sautter et al., 2002; Jambon et al., 2016). A new injection of magma, with a different
476 composition than the augite cores, could have created a disequilibrium, shown by the resorbed
477 rounded irregular boundaries of the augite cores (Fig. 2). This injection could have enabled the
478 crystallization of the euhedral augite rims, with a composition slightly enriched in Fe and relatively
479 depleted in Mg (Mg# 52) (Table 2). Those euhedral augite with enriched Fe composition could
480 also be explained by a slight drop of the temperature that creates an enrichment in Fe in the minerals
481 (Fig. 15) but would not explain the resorbed rounded cores. Thus, those irregular rounded cores
482 could be rather interpreted as sector zoning or patchy zoning in augite. In both cases, the
483 conservation of zoning indicates that augites cooling was fast enough to preclude diffusive
484 homogenization during augite rim crystallization.

485 Olivine in CeC 022 could have crystallized at the same time as euhedral augite. Olivines in
486 CeC 022 are not zoned and their compositions are homogenous throughout the sample with a Fe-
487 rich composition (Table 3). This agrees with Treiman (2005) who noticed that homogenous
488 olivines are the most ferroan in nakhlites. The sharp petrographic contact between olivine and
489 augite rims show equilibrium and no diffusion (Fig. 4). Using Toplis and Carroll (1995)

490 experimental data, Jambon et al. (2016) showed the correlation between Mg# of coexisting augite
491 and olivine with decreasing temperature (Fig. 15). As observed in CeC 022, olivine with Mg# ~30
492 should be in equilibrium with augite euhedral rims of Mg#~50 at ~1075°C (Fig. 15) as seen in
493 other nakhlites (Treiman 1986; Longhi and Pan, 1989; Harvey and McSween, 1992b; Sautter et
494 al., 2002; Jambon et al., 2016). The crystallization of augite alone is possible at higher pressure
495 than the crystallization of both olivine and augite together (Jambon et al., 2016), thus a drop of
496 pressure would initiate the crystallization of olivine and augite rims (Fig. 15). Thence, after the
497 crystallization of pyroxene cores, olivine and augite rims crystallize from a common melt at a
498 shallower depth. Nonetheless, as Fe and Mg diffusion are must faster in olivine than in pyroxene,
499 the homogeneous olivine could be the result of diffusion of initially zoned olivine. The original
500 olivine would have had more magnesian core, with enriched Fe-rim, resembling zoned olivine in
501 other nakhlite, such as NWA 817 or MIL 03346 (core Mg# 40-45) (Sautter et al., 2002; Day et al.,
502 2006; Udry and Day, 2018). Looking at the equilibrium between olivine and pyroxene (Fig. 15),
503 olivine cores richer in Mg would have crystallized at higher temperature, so before augite rim, right
504 after augite cores. The model of crystallization of cumulus minerals in nakhlite has been previously
505 largely discussed in literature. Caleta el Cobre 022 resemble other nakhlites in the crystallization
506 of augite and olivine cumulus, with its differences to other nakhlites described in this paper. For
507 further details about the model of crystallization, we advise the reader to see Day et al. 2006.

508 Olivine contains planes of inclusions composed of augite and magnetite, which are different
509 in structures and composition from the lamellar symplectite exsolution observed in Nakhla and
510 Governador Valadares (Greshake et al., 2000; Mikouchi et al., 2000). Those augites included in
511 olivine are close in composition to the augite rim, hence likely formed at the same time as the
512 crystallization of augite euhedral rim and olivine, or a slightly after, before final crystallization of
513 intercumulus materials.

514 **4.1.2. Intercumulus phase crystallization**

515 The crystallization of low-Ca pyroxene can be explained by the continuous crystallization
516 of augite and olivine with a decreasing temperature, which would lead to a Ca depleted and Fe
517 enriched melt, enabling the crystallization of low-Ca pyroxene. Also, a drop in the pressure during
518 crystallization could explain the low-Ca pyroxene rim. In basaltic igneous rocks, the peritectic
519 liquidus relationship of olivine and pyroxene is olivine + melt \rightarrow low-Ca pyroxene (e.g., Kushiro
520 1969; Morse 1980). Thus, a magma saturated in augite at higher pressure, would become
521 undersaturated in augite and saturated in low-Ca pyroxene at lower pressure (Treiman and Irving,
522 2008). Hence, low-Ca pyroxene can be found surrounding olivine in nakhlites. This demonstrates
523 that olivine and augite minerals finished growing in a closed system and the last melt crystallized
524 the low-Ca pyroxene rim and the mesostasis pockets with the fine Fe-rich overgrowth.

525 The thin and sharply delimited Fe-rich overgrowth is found on cumulus augite on their
526 contact with the mesostasis but is absent on low-Ca pyroxene rim. This overgrowth is out of
527 equilibrium with the olivine and augite rim (Fig. 15) and has no gradual transition between the rim
528 and the overgrowth, thus no diffusive exchange occurred in the crystal. Moreover, the euhedral
529 augite rims show no stage of resorption. Therefore, the crystallization of the Fe-rich pyroxene
530 overgrowth is due to rapid cooling and not to a new melt which would have disequibrated the
531 augite rim and olivine. In the mesostasis, the association of large apatites with low-Ca pyroxene
532 rim around olivines suggests that apatite growth is linked to the crystallization of low-Ca pyroxene,
533 so that the Ca content of apatite might come from the dissolution of augite (Treiman and Irving,
534 2008) (Fig.7). Intercumulus olivine crystals were not found in the mesostasis, where they have
535 possibly been altered in iddingsite. Indeed, iddingsite is found in the mesostasis as patches that
536 have sometimes shapes reminiscent of olivine crystals.

537 Sulfides in CeC 022 occur in unusually high amount (0.15 ± 0.05 vol.%) compared to other
538 nakhlites and show textural evidence for sulfide mobility and reactivity. The numerous occurrences
539 of sulfide blebs around and inside olivines may reflect a population of sulfide perhaps entrained
540 from deeper levels. In some cases, pyrrhotite-magnetite clusters were brecciated, which may be a
541 shock-related feature. Previous work on MIL's nakhlites has suggested that such large, sub-
542 rounded pyrrhotites, here almost systematically surrounded by magnetite, locally altered, may
543 represent a sulfide population of crustal origin assimilated during magmatic emplacement of the
544 nakhlites and are thus unstable in the matrix (Day et al., 2006; Franz et al., 2014)

545 **4.2.2. Secondary alteration**

546 Caleta el Cobre 022 exhibits one of the most extensive amounts of aqueous alteration
547 products reported in nakhlites (45.1 vol% of olivine crystal replaced by veins of alteration
548 products), resembling NWA 10153 in that regard (around 45.2 vol%). These alteration products,
549 called iddingsite, are assumed to be pre-terrestrial in this study on the basis of the close similarity
550 to iddingsite in other nakhlites, such as Nakhla and Lafayette that contain iddingsite veins truncated
551 by the fusion crust, which prove their extra-terrestrial origin (Gooding et al., 1991; Treiman et al.,
552 1993, Hicks et al., 2014).

553 In CeC 022, at least two types of iddingsite are detected: alteration veins in olivine and
554 patches in mesostasis pockets. The latter are interpreted as relicts of olivine grains in the mesostasis.
555 The Raman analyses reveal a poorly crystalline type and a more crystalline type (Fig. S4). EDS
556 analyses show that in comparison, the less crystalline phase is silicon-rich (~ 50 wt% SiO_2) and the
557 more crystalline is iron-rich (~ 52 wt% FeO). Both Raman and IR measurements reflect hydrated
558 alteration phases (Fig. S4). Also, the mesostasis shows alteration into iddingsite at the micrometer
559 scale (Fig. 16), which has been also observed in other nakhlites, but rarely studied compared to

560 iddingsite veins in olivine (Treiman et al., 2005; Daly et al., 2019). The sulfide and oxide stringers,
561 found inside iddingsite veins cross-cutting olivine, indicate that the fluid responsible for this
562 alteration event on Mars could also have transported reduced sulfur inside the nakhlites, after
563 completion of igneous crystallization of CeC 022. Our observations are consistent with previous
564 reports of magmatic sulfides oxidation to sulfate during weathering on Mars during episodic
565 periods of water activity in the vicinity of the nakhlite parent rock that precipitated
566 iddingsite/smectite as recently as 600-650 Ma (Franz et al., 2019 and reference therein). We may
567 surmise that hydrothermalism could have been able to transport sulfur inside the nakhlite while
568 bringing the water that was involved in local replacement of intercumulus sulfides of igneous origin
569 by iron oxyhydroxide-like phases.

570 **4.3. Geochronology and isotope systematics**

571 Caleta el Cobre 022 crystallization age was determined at 1.215 ± 0.067 Ga (Sm-Nd).
572 Considering only Sm-Nd ages, CeC 022 appears to be slightly younger than the three other
573 nakhlites dated by Sm-Nd (Nakhla, Lafayette and Gobernador Valadares) with an age range of 1.26
574 ± 0.07 – 1.37 ± 0.02 Ga (Nyquist et al., 2001), close to Nakhla within errors. With nakhlite
575 crystallization ages ranging from 1.26 to 1.42 Ga (from compilations of Sm-Nd, K-Ar, ^{40}Ar - ^{39}Ar ,
576 Rb-Sr, U-Th-Pb data, from Udry and Day, 2018), CeC 022 could be one of the youngest nakhlites
577 found yet. More recently, Cohen et al., 2017 presented an exhaustive study of ^{40}Ar - ^{39}Ar dating of
578 nakhlites, proposing four geochronologically distinct eruptions, the youngest age being Lafayette
579 with 1.321 ± 0.009 Ga. While the origin of the systematic offset between older ^{40}Ar - ^{39}Ar ages and
580 younger ages obtained with other dating systems is not clear, CeC 022 could be the youngest
581 nakhlite found yet and may extend the nakhlite eruptive sequence by about 50 Myr. As only one
582 radiochronologic system has been measured here and because different isotopic systems could lead

583 to different ages for a given nakhlite (Nyquist et al., 2001), this statement should be taken with
584 precaution. Thus, the comparison of CeC 022's Sm-Nd age with ^{40}Ar - ^{39}Ar ages is tentative and
585 ^{40}Ar - ^{39}Ar dating of CeC 022 is highly desirable.

586 The initial value of $\epsilon^{143}\text{Nd} = +10.5 \pm 0.1$, is clearly lower than the other nakhlites, with an
587 average of $\epsilon^{143}\text{Nd} = +15.7 \pm 1.4$ (2SD), from 14.6 ± 0.1 to 16.9 ± 0.1 (Caro et al., 2008; Debaille
588 et al., 2009). As demonstrated by the positive anomaly in Ce, this lower $\epsilon^{143}\text{Nd}$ could be related to
589 terrestrial contamination in the Atacama Desert. However, the $\mu^{142}\text{Nd}$ of CeC 022, of $+56 \pm 3$, is
590 very similar to other nakhlites, with $\mu^{142}\text{Nd}$ from $+51$ to $+67$ (Caro et al., 2008; Debaille et al.,
591 2009). Therefore, the decoupling between ^{143}Nd and ^{142}Nd systematics rules out terrestrial
592 disturbance of the Nd systematics as alteration would shift them concomitantly and the ^{142}Nd
593 systematics are identical to other nakhlites. Instead, this decoupling favors a slight change in the
594 source of nakhlite, possibly with time and/or metasomatic agent (McCubbin et al., 2013). However,
595 in any case, the contaminant agent must somehow be related to nakhlites. Indeed, the ^{142}Nd
596 systematics were established during the first 500 Myr of Mars geological history (while ^{146}Sm was
597 extinct) and could not have varied since then, except by mixing with another source. In this case,
598 it would mean that the second source would have had exactly the same ^{142}Nd value, which is highly
599 unlikely. One possibility to reconcile the nakhlite-like ^{142}Nd but slightly lower ^{143}Nd of CeC 022
600 would be that it is actually contaminated by older nakhlites that evolved towards lower $\epsilon^{143}\text{Nd}$
601 during the time between their emplacement and the emplacement of CeC 022. Considering the
602 $^{147}\text{Sm}/^{144}\text{Nd}$ ratios of nakhlites, a nakhlite-like sill or flow emplaced at least 700 Myr before CeC
603 022 would be able to generate a $\epsilon^{143}\text{Nd}$ low enough to decrease the ^{143}Nd systematics of CeC 022,
604 without changing its REE and ^{142}Nd systematics. This hypothesis is however impossible to test
605 with the current sampling of the nakhlite magmatic system.

606 The shielding indicator $(^{21}\text{Ne}/^{22}\text{Ne})_{\text{cosm}}$ of ~ 0.831 is low, as often observed in martian
607 meteorites. With the model specifically created for very small shergottites discussed in paragraph
608 3.6.2., and using a 7 cm pre-atmospheric radius, the recovered mass is estimated to be about 10%
609 of the pre-atmospheric mass of about 4 kg, which corresponds to the mass of the meteorite that was
610 found. The nominal CRE ages for ^3He , ^{21}Ne and ^{38}Ar are all consistent with ages found for other
611 nakhlites: around 11 Ma (Herzog and Caffee, 2013).

612 **4.4. Relation with other nakhlites**

613 **4.4.1. Nakhlites typology**

614 Independently of the different models of emplacement into a single pile (Treiman, 1986;
615 Mikouchi et al., 2003, 2012; Treiman, 2005; Bridges and Warren, 2006; Day et al., 2006;
616 McCubbin et al., 2013; Corrigan et al., 2015) or different units (Jambon et al., 2016; Balta et al.,
617 2016; Cohen et al., 2017; Udry and Day., 2018), nakhlites can be divided into two subgroups based
618 on mineralogy, mineral compositions, REE bulk composition, rock texture and CSD: the slower-
619 cooled nakhlites and the faster-cooled nakhlites.

620 The faster-cooled subgroup comprises NWA 817, MIL's nakhlites, NWA 10645 and NWA
621 5790. They show zoned olivine with a large range of composition $\text{Fo}_{55}\text{-Fo}_{95}$ (Fig. 10). Pyroxenes
622 show patchy zoning in NWA 10645 (Udry and Day, 2018) and in NWA 5790 (Jambon et al., 2016),
623 with a similar Fe-rich composition (Table 6). They also display Fe-rich overgrowth on augite
624 phenocryst in contact with the mesostasis, produced by rapid cooling of the last residual melt. The
625 zoning observed in pyroxene and olivine is retained due to rapid cooling preventing diffusive
626 homogenization. The mesostasis of faster-cooled nakhlites show quenched oxides with skeletal
627 morphology and fine-grained feldspar, both characteristics of high cooling rates.

628 The slower-cooled subgroup is composed of Nakhla, Lafayette, Governador Valadares,
629 NWA 998, Yamato (Y) 000593 and the pairing group of NWA 10153 (NWA 11013, NWA 10153,
630 NWA 10659 and NWA 10720; Udry and Day, 2018). They present unzoned olivine due to
631 diffusion homogenization enabled by slow crystallization. They also show oxide phenocrysts
632 formed through slow crystallization, in contrast to quenched oxides in faster-cooled nakhlites. The
633 intercumulus materials is mainly composed of coarse plagioclase with very little microcrystalline
634 mesostasis, or none.

635 In addition, REE composition of nakhlites reflects these two different subgroups. The
636 nakhlites with higher cooling rates have a higher content of REE, except for NWA 10113,
637 compared to the nakhlites showing lower cooling rates features, which have lower REE content
638 (Fig. 13), as mentioned by Udry and Day (2018). This grouping is also comparable to the
639 mesostasis grouping by Corrigan et al. (2015), who used modal abundance of major phases and
640 McCubbin et al. (2013) who worked on volatile-bearing minerals morphology and composition to
641 present a model of crystallization sequence of the nakhlites.

642 Caleta el Cobre 022 shows evidence of rapid cooling, as seen in the faster-cooled subgroup
643 and at the same time, shows features consistent with a lower cooling rate, similar to the slower-
644 cooled subgroup. Caleta el Cobre 022 uniqueness is the coexistence of those characteristics of both
645 lower and higher cooling rates.

646 **4.4.2. Slower-cooled nakhlite characteristics**

647 *Olivine*

648 Caleta el Cobre 022 presents a low content of olivine (4.2 vol%), being the second olivine-
649 poorest nakhlite after NWA 10645 (1.2 vol%). Olivines in CeC 022 are the second most fayalitic
650 analyzed (Mg# 30), after the exception of NWA 10645 (Mg# 22), also close to the composition of

651 Lafayette (Mg# of 34) and NWA 998 (Mg# of 36), both slower-cooled nakhlites (Table 7, Fig. 10).
652 Also, olivines do not show any zoning, as seen only in the olivines of NWA 998 and Lafayette, a
653 characteristic explained by a stage of diffusion at low cooling rates.

654 *Pyroxene rims*

655 The large rims of augite richer in Fe around the irregular cores of augite show similar
656 mineral compositions ($\text{En}_{31}\text{Fs}_{29}\text{Wo}_{40}$, Mg# ~ 52) as in the slower-cooled nakhlites, such as the
657 Nakhla subgroup (Udry and Day, 2018) (Fig. 9). Also, as seen in both slower and faster-cooled
658 nakhlites, like Nakhla and NWA 817 (Treiman, 2005), pyroxenes are also observed as rounded
659 inclusions (~100 μm) in olivine crystals (Fig. 4).

660 *Plagioclase*

661 Caleta el Cobre 022 contains a high proportion of millimeter-sized plagioclase grains (19
662 vol%), also seen in NWA 10153 (32 vol%) and NWA 998 (10.5 vol%) (Table 1). The compositions
663 of those plagioclase are also comparable within errors, for example $\text{An}_{35}\text{Ab}_{61}\text{Or}_4$ for NWA 10153
664 and $\text{An}_{25}\text{Ab}_{72}\text{Or}_3$ for NWA 998 and $\text{An}_{29.7\pm 4.3}\text{Ab}_{67.2\pm 4.0}\text{Or}_{3.1\pm 0.5}$ for plagioclase in CeC 022 (Fig. 11).
665 Those coarse plagioclases are only seen in slower-cooled nakhlites as intercumulus material, and
666 are characteristic of slower crystallization of the last melt, in contrary of fine-grained mesostasis,
667 created by rapid cooling of the last melt.

668 *Pyroxene overgrowth*

669 The fine overgrowth of Fe-rich augite is found on the euhedral augite rim and only when in
670 contact with the mesostasis. This stage of final fast cooling is observed on all nakhlites, but the
671 compositions of this pyroxene sharp overgrowths and intercumulus pyroxenes found in the
672 mesostasis are similar to the subgroup of slower-cooled nakhlite Gobernador Valadares and Nakhla
673 but richer in iron (up to $\text{En}_{35}\text{Fs}_{61}\text{Wo}_4$) (Fig. 9).

674 *Sulfides*

675 The population of large-sized intercumulus sulfides closely associated with magnetite looks
676 similar to the slower-cooled nakhlite subgroup, such as NWA 998 (Chevrier et al., 2011).

677 *Quantitative textural analysis*

678 The habit ratio and grain sizes of the pyroxene population of slower-cooled nakhlites are
679 similar to all nakhlites, except NWA 10645, which contains pyroxene double in length than the
680 nakhlite average (Udry and Day, 2018). The similar CSD slope, CSD intercept and average
681 pyroxene size of slower-cooled Nakhla and Lafayette with CeC 022 (Fig. 8) demonstrate similar
682 emplacement histories for the pyroxene population. Based on CSD analysis, CeC 022 might
683 originate from the same unit as these nakhlites, but closer to the surface as shown by the quenched
684 oxides and lower content of pyroxene (settling). The SDP analyses show that CeC 022 falls close
685 to the other nakhlites, especially NWA 11013 (Fig. S3).

686 **4.4.3. Faster-cooled nakhlite characteristics**

687 *Pyroxenes cores*

688 Caleta el Cobre 022 contains a low abundance of pyroxene phenocrysts (58 vol%) similar
689 to NWA 5790 (57.8 vol%) and NWA 11013 (55.0 vol%), which have the lowest pyroxene
690 abundance in nakhlites (Table 1). The irregular rounded cores with the euhedral rim richer in Fe
691 are described so far only in the faster-cooled nakhlites NWA 5790 (Jambon et al., 2016) and NWA
692 10645 (Udry and Day, 2018). The augite cores in CeC 022 are the most Fe-rich measured in
693 nakhlites yet (16.2 wt%), richer than NWA 5790 (15.7 wt%), NWA 10645 (14.5 wt%), the
694 nakhlites showing those irregular cores (Table 6; Udry and Day, 2018). The conservation of those
695 cores, with a different composition than the surrounding rim, is an indicator of rapid cooling, fast
696 enough to prevent diffusive homogenization (Jambon et al., 2016).

697 ***Mesostasis***

698 Caleta el Cobre 022 presents a high abundance (16 vol%) of mesostasis, resembling NWA
699 5790 (36.3 vol%) and NWA 817 (24.5 vol%), from the faster-cooled subgroup. This abundance of
700 mesostasis can reflect the packing efficiency, thus the burial depth of the rocks. The high abundance
701 of mesostasis in Caleta el Cobre 022 suggests a weak packing efficiency, thus a shallow burial
702 depth, which could induce a higher cooling rate. The abundance of mesostasis could also reflect
703 the crystallization of the nakhlite from dynamically sorted phenocryst-rich magma (Corrigan et al.,
704 2015). The mesostasis pockets contain dendritic cross-shaped oxides, low-Ca pyroxene and
705 apatites, embedded in fine-grained alkali feldspar and silica groundmass, resembling faster-cooled
706 nakhrites NWA 817, MIL 03346 and NWA 5790 (Sautter et al., 2002; Day et al., 2006; Jambon et
707 al., 2016). The fine-grained K-feldspars in the mesostasis has similar composition to the faster-
708 cooled nakhrites NWA 817 and NWA 5790 (Fig. 11). Caleta el Cobre 022's main peculiarity is in
709 those pockets of mesostasis caused by rapid cooling, associated with coarse plagioclase (Fig. 12).

710 This association of contrasted intercumulus materials has never been observed before in
711 other nakhrites. It could result from a change in cooling rate after crystallization of phenocrysts and
712 plagioclase, during eruption of the magma or emplacement as shallow sill with a drastic
713 temperature drop, leading to fast crystallization of mesostasis pockets and pyroxene overgrowth.
714 Alternatively, this association could be due to a new injection of magma after crystallization of
715 phenocrysts and plagioclase, inducing eruption and faster crystallization.

716 ***Sulfides***

717 Caleta el Cobre 022 is sulfide-rich (0.15 ± 0.05 vol.%) compared to other nakhrites so far
718 studied for sulfides (Chevrier et al., 2011; Franz et al., 2014; Dottin et al., 2018; Mari et al., 2019).

719 The sulfide droplets inside skeletal mesostasis minerals coupled with Ni-free pyrrhotite
720 were described in faster-cooled nakhlites as NWA 817 (Chevrier et al., 2011) and MIL's nakhlites
721 (Day et al., 2006; Dottin et al., 2018). Similarly, the high sulfide content, coupled with Ni-free
722 pyrrhotite compositions in CeC 022, also resemble MIL's nakhlites (Day et al., 2006; Franz et al.,
723 2014; Dottin et al., 2018). Assimilation of crustal sulfur during the igneous crystallization stage
724 could account for CeC 022 pyrrhotite modal abundance and Ni-poor composition, as repeatedly
725 documented in nakhlites, especially in the fast-cooled subgroup (Farquhar et al., 2007; Franz et al.,
726 2014; Dottin et al., 2018; Mari et al., 2019; Franz et al., 2019 and reference therein). The mean
727 Fe/S ratio of CeC 022 pyrrhotite ($\text{Fe}_{0.89}\text{S}$; hexagonal pyrrhotite) (Fig. S2), coupled with textural
728 evidence of magnetite, indicates similar oxidizing crystallization conditions near the FMQ
729 synthetic buffer (Chevrier et al., 2011), as usually reported for nakhlite pyrrhotites that inherited
730 abundant crustal sulfur in their S isotopic composition (Farquhar et al., 2007b; Franz et al., 2014;
731 Dottin et al., 2018).

732 ***Rare earth element compositions***

733 Caleta el Cobre 022 shows a high bulk content in REE, resembling faster-cooled nakhlites
734 such as NWA 10645, NWA 817 and MIL 03346 except for a clear positive anomaly in Ce (Fig.
735 13). As mentioned in Pourkhorsandi et al. (2017), a positive Ce anomaly is observed in REE
736 patterns of meteorites from the Atacama Desert, as is CeC 022. This Ce anomaly is likely due to
737 terrestrial alteration specific to Atacama, as Ce in its oxidized form (Ce^{4+}) is a quite mobile element.

738 **4.4.4. Magnetic properties**

739 Magnetic susceptibility average mass weighed value of $8.41 \times 10^{-6} \text{ m}^3/\text{kg}$ is to be compared
740 to the average value of $4.10 \pm 1.73 \times 10^{-6} \text{ m}^3/\text{kg}$ for the other nine nakhlites (Rochette et al. 2005,
741 Jambon et al. 2016 and our measurement of NWA 10153 at $5.25 \times 10^{-6} \text{ m}^3/\text{kg}$). Saturation remanent

742 magnetization (M_{RS}) is $0.459 \text{ Am}^2/\text{kg}$, compared to $0.160 \pm 0.108 \text{ Am}^2/\text{kg}$ for the other nine
743 nakhlites (Rochette et al. 2005, Jambon et al. 2016, and our measurement of NWA 10153 at 0.135
744 Am^2/kg .) Saturation magnetization is $1.51 \text{ Am}^2/\text{kg}$, to be compared with $0.66 \pm 0.37 \text{ Am}^2/\text{kg}$ for
745 the other nine nakhlites. Caleta el Cobre 022 is the most magnetic nakhlite (Fig. 17).

746 The presence of two Curie temperatures of 525°C and 585°C evidences two families of
747 ferromagnetic iron oxides with compositions $\text{Fe}_{2.9}\text{Ti}_{0.1}\text{O}_4$ (more abundant) and Fe_3O_4 , respectively
748 (Lattard et al., 2006). The microprobe analyses on magnetite phenocrysts give a composition richer
749 in Ti of $\text{Fe}_{2.6}\text{Ti}_{0.4}\text{O}_4$. This difference is explained by the exsolutions observed at nanometric scales
750 in these iron oxides grains (Fig. 5). Thus, the microprobe analyses represent a combination of the
751 nanoscale exsolution with the oxide surrounding them. Using the composition $\text{Fe}_{2.9}\text{Ti}_{0.1}\text{O}_4$, the
752 saturation magnetization indicates a magnetite content of 2.90 vol%. This estimate, that does not
753 include the ilmenite/ulvöspinel exsolution since these two minerals are paramagnetic at room
754 temperature, is in good agreement with the modal abundance of oxide phenocrysts estimated from
755 image analyses (2.6 vol%).

756 Besides the petrological implications, the study of magnetic properties of martian
757 meteorites is also of interest for the interpretation of the magnetic anomalies that have been
758 documented over large parts of the Noachian crust at satellite altitude (Acuña et al., 1999). The
759 strong intensity of these anomalies (over 1000 nT at 100 km altitude) imply the existence of a past
760 dynamo operating on Mars, but also the existence of rocks capable of acquiring strong remanent
761 magnetization. It has been estimated that such rocks should satisfy the condition $M_{RS} > 0.29 \text{ Am}^2/\text{kg}$,
762 regardless of their present-day natural remanent magnetization that may have been later modified
763 by a number of processes including shock demagnetization at the surface of Mars, or contamination
764 by magnets by meteorite collectors (Gattacceca et al., 2014). It is noteworthy that this condition is
765 met by only a handful of martian meteorites: MIL 03346, NWA 5790, NWA 7034 and paired

766 stones (Rochette et al., 2005; Gattacceca et al., 2014; Jambon et al., 2016; Herd et al., 2017). With
767 $M_{RS}=0.459 \text{ Am}^2/\text{kg}$, CeC 022 is only the fourth martian meteorite and the third nakhlite whose
768 magnetic properties may account for the strong intensity of the martian magnetic anomalies (Fig.
769 17). Although nakhlites are obviously younger than the age of the end of the martian dynamo (e.g.,
770 Lillis et al., 2008), CeC 022 still represent an analogue source lithology for the strong
771 magnetization of the martian crust.

772 **4.4.5. Caleta el Cobre 022: a new subgroup**

773 Caleta el Cobre 022 shows similarities with faster-cooled nakhlites, as well with slower-
774 cooled nakhlites. A new characteristic, never observed before in nakhlites, is the presence of both
775 fine-grained mesostasis and coarse millimeter-size plagioclase in CeC 022, showing two different
776 cooling histories in intercumulus materials. In addition, compared to other nakhlites, CeC 022 has
777 the lowest CaO content amongst nakhlites (12.9 wt%), similar to faster-cooled nakhlites NWA
778 5790 (12.9 wt%) or NWA 817 (13.1 wt%). Its low content in FeO (19 wt%) is also close to NWA
779 817, NWA 5790, or MIL nakhlites (Jambon et al., 2016). Caleta el Cobre 022 has high MgO
780 content, similar to Lafayette, NWA 998 and Nakhla (Udry and Day, 2018), with a Mg# of 53.8,
781 closer to the slower-cooled nakhlites (Table 4). This bulk composition and the abundance and
782 composition of mineral phases are a reflection of the presence of both faster and slower-cooled
783 nakhlite subgroup features in CeC 022. In addition, martian aqueous alteration phases are found in
784 high abundance compared to most other nakhlites, and the occurrence of sulfides and Fe-oxides at
785 the core of iddingsite veins is a unique and puzzling feature, giving new insights into the complex
786 thermal and alteration histories of nakhlites.

787 With all its different characteristics, CeC 022 cannot be part of any previously proposed
788 nakhlite subgroup, and therefore increases the diversity of the martian sample suite. In the Udry

789 and Day (2018) model presenting six different units in the nakhlite group based on petrography,
790 crystallinity and composition, CeC 022 may thus represent a 7th subgroup. Also, CeC 022 has a
791 slightly younger Sm-Nd age than other nakhlites. Caleta el Cobre 022 must originate from the same
792 volcanic system as the other nakhlites, according to their similar crystallization age and
793 petrography, compositions and physical properties, but it probably samples a different flow or sill
794 than the other nakhlites.

795 **5. Conclusion**

796 All petrographic, chemical and isotopic analyses confirm that CeC 022 is a nakhlite,
797 coming from the same volcanic system as other nakhlites, although with characteristics never
798 observed before in other nakhlites. On one hand, it shows similarities with faster-cooled nakhlites,
799 such as NWA 817, NWA 5790 or MIL's nakhlites: conservation of augite irregular cores, Fe-rich
800 overgrowth, high proportion of mesostasis with fine-grained K-feldspars, quenched oxides and
801 high REE content. On the other hand, CeC 022 also shares similarities with slower-cooled
802 nakhlites, like Nakhla and NWA 10153 (Udry and Day, 2018), including: pyroxene modal
803 abundance, CSD slope and intercept, average pyroxene size, phenocrysts minerals composition,
804 unzoned olivine and abundant plagioclase phenocrysts. One unique feature of CeC 022 is the
805 presence of both fine-grained mesostasis and millimeter-size coarse plagioclase, showing different
806 cooling rates of the intercumulus material through time, a feature never observed before in nakhlites.
807 With these characteristics, CeC 022 may thus represent a 7th nakhlite subgroup in Udry and Day
808 (2018) model of six different subgroups. Its slightly younger Sm-Nd age also supports CeC 022
809 was emplaced in a different flow or sill. Moreover, higher abundance of martian aqueous alteration
810 phases compared to other nakhlites, showing sulfides and Fe-oxides at the core of iddingsite veins
811 is a new characteristic. Magnetic susceptibility and saturation remanent magnetization are higher
812 than for other nakhlites, indicating an overall higher content of magnetite. This elevated magnetite

813 content makes CeC 022 a potential analogue lithology to account for the strong magnetic anomalies
814 observed at the surface of Mars. With its multistage thermal history, its abundant and complex
815 aqueous alteration, its younger age and its unusual magnetic properties, CeC 022 increases the
816 diversity of the martian sample suite and gives new insights into the complex thermal and alteration
817 histories of nakhlites.

818

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