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Al-Mg and U-Pb chronological records of Erg Chech 002 ungrouped achondrite meteorite --Manuscript Draft--

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Abstract:	<p>Achondrite meteorites are remnants of the earliest planetary differentiation processes in the Solar System. They have been used to anchor short-lived radiochronometers to absolute ages determined from long-lived radiochronometers. More specifically, when comparing the isotopic systematics of the short-lived ^{26}Al-^{26}Mg chronometer anchored to absolute U-corrected Pb-Pb ages, inferences about the distribution of ^{26}Al (half-life of $\sim 717\,000$ yr) in the protoplanetary disk can be evaluated. The ungrouped achondrite Erg Chech (EC) 002 has a distinct mineralogy and more evolved elemental composition compared to basaltic achondrites. In situ and solution ^{26}Al-^{26}Mg chronometry and ^{53}Mn-^{53}Cr chronometry suggest that EC 002 formed within ~ 0.7 to 2.2 Ma after the formation of Ca-Al-rich inclusions (CAIs), making it the oldest known sample of igneous crust in the Solar System (Barrat et al., 2021; Anand et al., 2022; Zhu et al., 2022; Fang et al., 2022). Here we present the U-corrected Pb-Pb age and ^{26}Al-^{26}Mg age obtained by MC-ICPMS solution analysis of the same mineral separate samples of EC 002. In addition, six merrillite grains were analyzed by in-situ SIMS to determine their Pb-Pb individual ages.</p> <p>The U isotope composition of EC 002 exhibits internal heterogeneities between leached pyroxene ($^{238}\text{U}/^{235}\text{U} = 137.766 \pm 0.027$) and the bulk rock ($^{238}\text{U}/^{235}\text{U} = 137.8190 \pm 0.0074$). The Pb isotope composition of progressively leached pyroxenes are characterized by radiogenic $^{206}\text{Pb}/^{204}\text{Pb}$ ratios (ranging from 41 to 23487). Using the U isotope composition of the leached pyroxenes, the resulting age of the $^{207}\text{Pb}/^{206}\text{Pb}$-$^{204}\text{Pb}/^{206}\text{Pb}$ isochron is 4565.87 ± 0.30 Ma (2σ). The weighted mean of the Pb-Pb ages of seven SIMS analyses of merrillites are 4564.3 ± 5.2 Ma (2σ). These similar ages (within uncertainty) indicate rapid cooling and the absence of significant thermal events after ~ 4559 Ma on the parent body of EC 002. The ^{26}Al-^{26}Mg isochron through a bulk, pyroxene, fine-grained and four plagioclase fractions defines an initial $^{26}\text{Al}/^{27}\text{Al}$ ratio of $[8.89 \pm 0.79] \times 10^{-6}$ corresponding to a formation age of 1.83 ± 0.12 Ma after CAIs ($[5.23 \pm 0.13] \times 10^{-5}$; Jacobsen et al., 2008). The initial ^{26}Al abundance is consistent with previous MC-ICP-MS ^{26}Al-^{26}Mg reported systematics for EC 002 (Fang et al., 2022), but 0.46 ± 0.13 Myr older than the in situ SIMS ^{26}Al-^{26}Mg age previously reported by Barrat et al. (2021). When anchored to the</p>

absolute Pb-Pb age of CV3 CAIs (4567.30 ± 0.16 Ma; Connelly et al., 2012), the Al-Mg model age of EC 002 is 4565.47 ± 0.20 Ma, slightly younger than its U-corrected Pb-Pb age.

The concordance of the Pb-Pb and ^{26}Al - ^{26}Mg ages of ungrouped CC achondrites when anchored to EC 002 suggest that ^{26}Al was homogeneously distributed between the NC and CC reservoirs at the time of their parent body accretion. Furthermore, the presence of internal U isotope heterogeneities found between mineral and whole-rock samples of EC 002 supports the need of U isotope analysis of meteoritic samples dated using the Pb-Pb chronometer.

1 **Al-Mg and U-Pb chronological records of Erg Chech 002 ungrouped**
2 **achondrite meteorite**

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32 **Abstract**

33 Achondrite meteorites are remnants of the earliest planetary differentiation processes in the Solar
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44 analysis of the same mineral separate samples of EC 002. In addition, six merrillite grains were
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62 to EC 002 suggest that ^{26}Al was homogeneously distributed between the NC and CC reservoirs at
63 the time of their parent body accretion. Furthermore, the presence of internal U isotope
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65 isotope analysis of meteoritic samples dated using the Pb-Pb chronometer.

66

67

68 **Keywords**

69 early Solar System, planetary differentiation, achondrite, Al-Mg chronology, U-Pb chronology

70 1. Introduction

71 The use of radiometric dating has been a cornerstone of unravelling the mysteries surrounding the
72 formation and early evolution of the Solar System. Of the various chronometers, the absolute Pb-
73 Pb chronometer and the short-lived, relative ^{26}Al - ^{26}Mg chronometer have been particularly useful
74 in dating extra-terrestrial materials. The ^{26}Al - ^{26}Mg chronometer can resolve events that took place
75 within $\sim 30,000$ years, while the U-corrected Pb-Pb chronometer has, among other milestones,
76 established the age of the Solar System between ~ 4567 and 4568 Ma by dating Ca-Al-rich
77 inclusions in CV3 chondrites (CAIs; Amelin et al., 2010; Bouvier and Wadhwa, 2010; Connelly
78 et al., 2012; Sanborn et al., 2019). Despite the merits of both these chronometers, questions remain
79 about their significance. For instance, meaningful applications of the ^{26}Al - ^{26}Mg chronometer rely
80 on a homogeneous initial distribution of the parent nuclide ^{26}Al ($T_{1/2} = 0.717$ Ma, NuDat 3.0,
81 National Nuclear Data Center) between samples and a chronological reference (e.g., CAIs,
82 achondrites). To determine absolute and accurate U-corrected Pb-Pb ages using the dual long-lived
83 decay chains of ^{235}U to ^{207}Pb and ^{238}U to ^{206}Pb , the $^{238}\text{U}/^{235}\text{U}$ ratio needs to be precisely known. It
84 was assumed that the $^{238}\text{U}/^{235}\text{U}$ ratio was invariant at 137.88 among early Solar System materials,
85 until significant variations of up to 3.5% in different objects were discovered (Stirling et al., 2005;
86 Amelin et al., 2010; Brennecka et al., 2010). The assumption of a homogeneous distribution of
87 ^{26}Al in the protoplanetary disk has repeatedly been challenged, without a consensus in the scientific
88 community emerging. Studies involving CAIs, amoeboid olivine aggregates (AOAs), bulk
89 meteorites (Larsen et al., 2011), chondrules (Bollard et al., 2019) and angrites (Schiller et al., 2015)
90 suggested an initial $^{26}\text{Al}/^{27}\text{Al}$ ratio in the inner Solar System around 1×10^{-5} , about 80% lower
91 than the canonical value of $\sim 5.2 \times 10^{-5}$ (Jacobsen et al., 2008). Other studies have disagreed with
92 these conclusions (e.g., Wasserburg et al., 2012; Kita et al., 2013; Gregory et al., 2020). In

93 particular, combined U-corrected Pb-Pb and ^{26}Al - ^{26}Mg dating of achondrites, e.g., angrites or
94 unique, ungrouped meteorites, have been frequently utilized to evaluate the distribution of ^{26}Al
95 within the protoplanetary disk (e.g., Bouvier et al., 2011; Schiller et al., 2015; Wimpenny et al.,
96 2019). The Pb-Pb and ^{26}Al - ^{26}Mg ages of some meteorites that formed in the non-carbonaceous
97 inner Solar System (NC reservoir; Warren, 2011; Kruijer et al., 2020), such as the angrite
98 D'Orbigny (Amelin, 2008b; Schiller et al., 2015; Tissot et al., 2017), used as an anchor meteorite
99 for short-lived radionuclide chronometers, and the unique achondrite Asuka 881394 (Wadhwa et
100 al., 2009; Hublet et al., 2017; Wimpenny et al., 2019) do not match in model ages when anchored
101 to CAIs from the CV3 chondrites Efremovka and Allende using their reported U-corrected Pb-Pb
102 ages (Connelly et al., 2012) and canonical $^{26}\text{Al}/^{27}\text{Al}$ ratio of $[5.23 \pm 0.13] \times 10^{-5}$ (Jacobsen et al.,
103 2008). Conversely, the Pb-Pb and ^{26}Al - ^{26}Mg systematics of achondrites that formed in the
104 carbonaceous chondrite (CC) reservoir of the outer Solar System, such as NWA 6704 (Amelin et
105 al., 2019; Sanborn et al., 2019) and NWA 2976 (Bouvier et al., 2011b), exhibit concordant U-
106 corrected Pb-Pb and ^{26}Al - ^{26}Mg ages when anchored to the absolute age of CAIs of 4567.30 ± 0.16
107 Ma (Connelly et al., 2012), suggesting a homogeneous distribution of ^{26}Al between CAIs and CC
108 meteorites. The discordance between U-corrected Pb-Pb and ^{26}Al - ^{26}Mg ages of some NC
109 meteorites has been interpreted by different studies to reflect either heterogeneity of ^{26}Al in the
110 Solar System (Schiller et al., 2015), or that the ^{26}Al - ^{26}Mg systematics were disturbed through
111 thermal metamorphism or shock heating from impact events (Sanborn et al., 2015; Koefoed et al.,
112 2016; Sanborn et al., 2019; Wimpenny et al., 2019).

113 In order to further evaluate the distribution of ^{26}Al in the early Solar System, we have used both
114 the Pb-Pb and ^{26}Al - ^{26}Mg chronometers to date the formation of the unique achondrite Erg Chech
115 002. Erg Chech 002 is a recent meteorite find with an unusual chemistry and mineralogy: (1) the

116 whole-rock composition is comparable to terrestrial andesites in silica and alkaline element
117 concentrations (Barrat et al., 2021) and (2) it contains large (up to 9 cm) orthopyroxene megacrysts
118 (Gattacceca et al., 2021) in a groundmass of sodic plagioclase, pigeonite and silica (Fig. 1).
119 Geochemically, EC 002 exhibits a distinct trace element pattern compared to other andesitic
120 meteorites, with no significant anomalies in alkali and high field-strength elements relative to CI
121 chondrites (Barrat et al., 2021; Nicklas et al., 2022). Highly siderophile elements show
122 fractionation between the compatible and incompatible elements that suggest metal-silicate
123 separation occurred on the parent body of EC 002 (Nicklas et al., 2022).

124 Mass-independent $\Delta^{17}\text{O}$ values of EC 002 suggest an affinity with four anomalous eucrites,
125 Bunburra Rockhole, EET 92023, Emmaville and Asuka 881394, but these meteorites are
126 mineralogically different from EC 002 (Gattacceca et al., 2021). The O isotope composition, the
127 negative Tm anomaly of the whole-rock (Barrat et al., 2021), as well as the Cr and Nd stable
128 isotope anomalies (Anand et al., 2022; Fang et al., 2022; Zhu et al., 2022) indicate accretion of the
129 EC 002 parent body in the NC reservoir. The elemental composition supports formation by partial
130 melting of a chondritic precursor, followed by rapid cooling after cooling to 900°C, likely by
131 excavation or ejection from an impact event (Barrat et al., 2021).

132 EC 002 was first dated by in situ ^{26}Al - ^{26}Mg analysis using secondary ionization mass spectrometry
133 (SIMS), as well as noble gas geochemistry (Barrat et al., 2021). The Al-Mg systematics result in
134 an initial $^{26}\text{Al}/^{27}\text{Al}$ ratio of $[5.72 \pm 0.07] \times 10^{-6}$ and an age of 2.29 ± 0.04 Ma after CAI formation,
135 making it the oldest igneous meteorite found to date. A subsequent study of the ^{26}Al - ^{26}Mg
136 chronology by MC-ICP-MS reported a different initial $^{26}\text{Al}/^{27}\text{Al}$ ratio of $[8.89 \pm 0.09] \times 10^{-6}$,
137 resulting in an age of 1.83 ± 0.04 Ma after CAI formation (Fang et al., 2022). Additionally, the
138 formation age of EC 002 was modelled at 4565.6 ± 0.6 Ma and 4566.6 ± 0.6 Ma by two separate

139 studies using the short-lived ^{53}Mn - ^{53}Cr chronometer, albeit anchored to the D'Orbigny angrite
140 (Anand et al., 2022; Zhu et al., 2022). To get a better constraint on the crystallisation history of
141 EC 002, assess its potential as a new anchor for short-lived radionuclides and to further evaluate
142 the distribution of ^{26}Al in the early Solar System, we analyzed the Mg, Pb and U isotope
143 compositions of whole-rock and mineral fractions from the same sample piece by multi collector
144 inductively coupled plasma mass spectrometry (MC-ICP-MS) to determine the ^{26}Al - ^{26}Mg and U-
145 corrected Pb-Pb ages. We also analysed merrillite Ca-phosphates by *in-situ* secondary ion mass
146 spectrometry (SIMS) Pb-Pb analysis to constrain the crystallization history and thermal records of
147 EC002.

148

149 2. Methods

150 *2.1 Sample preparation*

151 For U-Pb and Al-Mg studies, a ~ 600 mg section of Erg Chech 002 (from Ben Hoefnagels, Big
152 Bang Meteorites) was rinsed in acetone and water, prior to being crushed in an agate mortar.
153 Roughly 120 mg were taken as a bulk rock fraction. Pyroxene and feldspar minerals were separated
154 by hand-picking under binoculars, with repeated crushing and sieving. Eight feldspar fractions
155 were hand-picked, along with a pyroxene and a fine-grained (<60 μm) fraction for Mg isotope
156 analysis.

157

158 *2.2 Characterisation and SIMS Pb-Pb dating of phosphates*

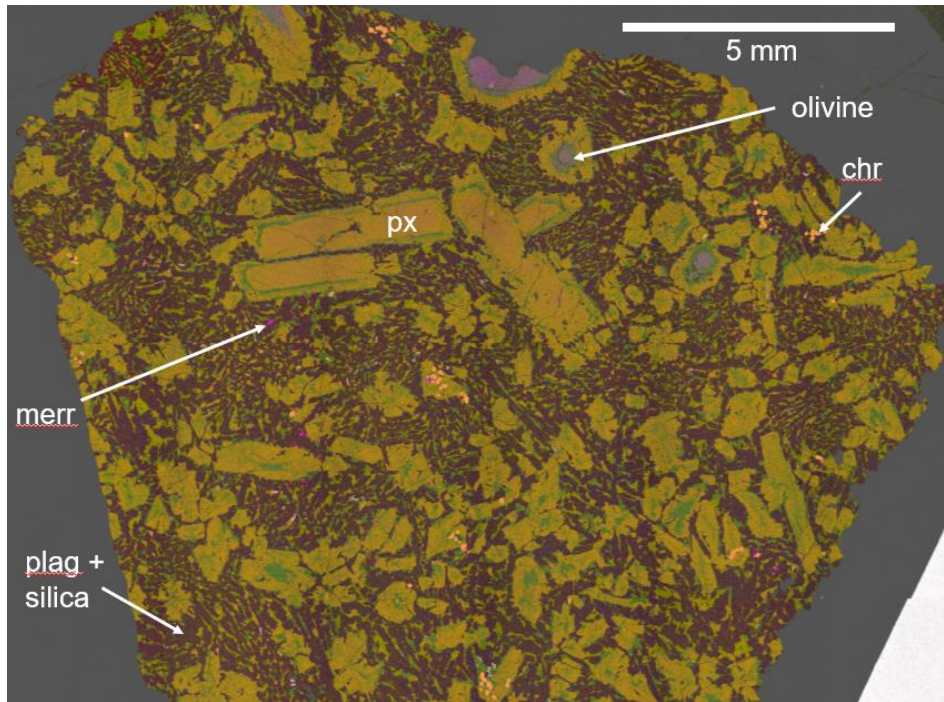
159 A 1-inch polished section mounted in epoxy was prepared from a 0.75g slice of EC 002 (Mirko
160 Graul Meteorite) at the Bayerisches Geoinstitut (BGI). The section was first carbon-coated and
161 analysed to obtain multi-field BSE-EDS element maps (Fig. 1; preselected to be Na, Mg, Al, Si,

162 P, O, Fe, Ca, Ti, Zr, Cr, Ni, S, Cl) using a focused ion beam scanning electron microscopy (FEI
163 Scios FIB-SEM) at BGI. For this purpose, the electron beam was tuned to 20 kV and 1.6 nA. We
164 selected six merrillite grains for SIMS analyses.

165 Further petrographic observations were carried out at BGI using a Zeiss Gemini 1530 scanning
166 electron microscope operated at 20 kV under BSE-EDS mode. Mineral compositions were
167 measured using a JEOL JXA 8200 electron microprobe equipped with five spectrometers operated
168 at 15 kV and 15 nA with $\sim 1 \mu\text{m}$ focused electron beam ($5 \mu\text{m}$ for merrillites) and counting time of
169 20s on peak and 10s on background (10s on peak for Na, Cl and F, and 5s on background to reduce
170 the beam induced damage and loss of volatile during analysis). The standards used were: Na and
171 Si (albite), Ca (wollastonite or apatite), K (orthoclase), Mg (enstatite), Fe (Fe metal), Mn and Ti
172 (MnTiO_3), Al (spinel), Ni (Ni metal), Cr (Cr metal), and P (apatite).

173 For SIMS, the carbon coating of the same polished mount of EC 002 was removed. The mount
174 was cleaned with alcohol, dried, and coated with a 50 nm gold film before SIMS analysis. High
175 precision ^{207}Pb - ^{206}Pb phosphate dating was performed using a Cameca IMS 1280-HR type
176 secondary ion mass spectrometer (SIMS) at the Heidelberg Ion Probe (HIP), Universität
177 Heidelberg. Duoplasmatron ion source was operated to generate a primary O^- ion beam with 23
178 kV impact energy, 20-40 nA current and beam size of 20-30 μm on sample surface. During each
179 analysis, a two-minute pre-sputtering was applied to remove potential surface contamination.
180 Positive secondary ions of $^{40}\text{Ca}_2^{31}\text{P}^{16}\text{O}_4^+$, $^{204}\text{Pb}^+$, $^{206}\text{Pb}^+$, $^{207}\text{Pb}^+$, $^{238}\text{U}^+$, $^{238}\text{U}^{16}\text{O}^+$ were extracted,
181 mass filtered at a mass resolution ($M/\Delta M$) of ~ 4000 before detected sequentially in a low-noise
182 ion-counting electron multipliers for 1s, 20s, 10s, 10s, 10s and 6s, respectively under single
183 collection mode. Thirty cycles per analysis were typically applied for better statistics.

184 $^{40}\text{Ca}_2^{31}\text{P}^{16}\text{O}_4^+$ was used for peak centering and $^{238}\text{U}^{16}\text{O}^+$ for calculating U and Pb secondary ion
185 yields.



186
187 **Figure 1:** Backscattered electron image overlaid with energy dispersive spectroscopy composed
188 as a multi-field phase colored map of Erg Chech 002 (Si in yellow, Na in dark blue, Al in grey,
189 Mg in red, Ca in pink) polished mount used for the SIMS analyses. Large pyroxene megacrysts
190 and rounded olivine crystals are surrounded by groundmass of sodic plagioclase, pigeonite and
191 accessory chromite, merrillite and silica polymorphs.

192

193

194 2.3 Determination of $^{27}\text{Al}/^{24}\text{Mg}$ ratios

195 A bulk rock fraction consisting of roughly 120 mg, one pyroxene and eight feldspar mineral
196 fractions prepared for Mg isotope analysis were dissolved in a 29 M HF- 15 M HNO₃ (5:1) mixture
197 at 120 °C, before being dried down and re-dissolved in 15M HNO₃. The samples were dried down

198 again and then converted to its chloride form with 6 M HCl. Determination of the $^{27}\text{Al}/^{24}\text{Mg}$ ratios
199 of the mineral separates was performed using a Thermo Fischer iCAP Qc quadrupole ICP-MS at
200 the University of Western Ontario (UWO). Of the eight feldspar fractions, after an initial analysis,
201 those with similar $^{27}\text{Al}/^{24}\text{Mg}$ ratios were combined in order to maximize Mg content, allowing for
202 analysis at higher concentrations, without sacrificing the spread of $^{27}\text{Al}/^{24}\text{Mg}$ ratios. This resulted
203 in four feldspar fractions, alongside the bulk rock, pyroxene and fine-grained fractions. During
204 analysis, gravimetrically prepared Al-Mg solutions, ranging from $^{27}\text{Al}/^{24}\text{Mg}$ ratios of 0.001 to 50,
205 were used to calibrate $^{27}\text{Al}/^{24}\text{Mg}$ ratios of the samples. Terrestrial standards BCR-2 and San Carlos
206 olivine, and bulk rock CV3 chondrite Allende were analyzed with the samples to verify the
207 accuracy of the analysis and are reported alongside their isotopic compositions in Table 1.
208 Uncertainties associated with $^{27}\text{Al}/^{24}\text{Mg}$ ratios is $\pm 5\%$, with exception of the bulk rock fraction,
209 which in two separate analyses resulted in distinct $^{27}\text{Al}/^{24}\text{Mg}$ ratios. To account for the increased
210 uncertainty, the error there is the 2SE of the average of both analyses.

211

212 *2.4 Magnesium isotope analysis*

213 An aliquot equivalent to 0.6 mg of the bulk rock fraction and corresponding to $\approx 20 \mu\text{g}$ Mg was
214 set aside for Mg separation, along with the pyroxene, fine-grained and feldspar fractions. We used
215 a two-stage ion-exchange chromatography protocol, following a modified version of the protocol
216 described by Wombacher et al. (2009).

217 The protocol uses about 1.2 ml of AG50W-X8 (200-400 mesh) resin deposited within HCl-cleaned
218 BioRad® plastic columns. For the first stage, the columns and resin are cleaned again together
219 with ca. 20 ml of 6 M HCl. The resin is then conditioned with 3.6 ml of 10 M HCl. The sample is
220 taken up in 0.3 ml 10 M HCl and then loaded onto the columns, before Mg and other matrix

221 elements (such as Al, Cr and Ni) are eluted by 3.2 ml of 10 M HCl. Subsequent Ca and Fe elution
222 are accomplished by adding 15 ml of 8 M HCl and 20 ml of 4 M HCl, respectively, to the columns.
223 To ensure sufficient recovery of Mg, the first 0.8 ml of the Ca elution step are collected separately
224 and dried down. The same procedure is repeated a second time with the separate cut which is then
225 collected and combined with the 3.2 ml from the previous run. Magnesium is then further separated
226 from the remaining matrix elements by a third stage, using the same columns and resin as the
227 previous stages. The resin is first conditioned with 3.6 ml 0.4 M HCl before the sample is loaded
228 onto the column in 0.3 ml of 0.4 M HCl. Remaining matrix elements are eluted by adding 33.6 ml
229 of 0.4 M HCl and then 7.2 ml of 0.15 M HF. For Mg elution, 1 ml of 1 M HCl is added before an
230 additional 20 ml are added (Table 2). This method results in a total yield of > 99.6 % of Mg. We
231 used dissolved homogenized powders of the San Carlos olivine and the Allende CV3 meteorite as
232 inter-laboratory secondary standards, that underwent the full protocol with each set of columns.
233 To evaluate that the column chemistry did not introduce any fractionation on Mg isotopes, a pure
234 Mg standard solution was analyzed without column processing, with column processing, and after
235 the solution being added to the Mg-free matrices of terrestrial standards BCR-2 and BIR-1.
236 Between these analyses, no resolvable differences in Mg isotope compositions were found (Table
237 1).

238 Magnesium isotope compositions of EC 002 fractions were analyzed by a Thermo Fisher Neptune
239 Plus MC-ICP-MS at the Friedrich-Alexander-Universität (FAU) in Erlangen. The ^{24}Mg , ^{25}Mg and
240 ^{26}Mg ion beams were measured simultaneously on the L4, Center and H4 Faraday cups,
241 respectively, with a 10^{10} amplifier on the L4 cup, and run in medium resolution mode with a Jet
242 sample cone and H skimmer cone interface. The sample and standard solutions were diluted to
243 approximately 300 ppb (≈ 90 V on ^{24}Mg), where Mg amounts allowed, in 2% HNO_3 and matched

244 to within 5% or less of their concentrations. Solutions were introduced using a Cetac Aridus II
 245 desolvating nebulizer system. Standard-sample bracketing with the DSM-3 standard was used to
 246 correct for mass fractionation and instrumental drift. Generally, analyses consisted of four sample
 247 measurements bracketed by five standard measurements, with each measurement consisting of a
 248 peak center, 20 cycles at 8.4 s of integration time followed by 150 s of wash time. The instrumental
 249 background was measured prior to each sample and standard analysis. Total procedural blanks
 250 were less than 5 ng, a negligible amount relative to the amount of Mg in the samples ($\approx 75 \mu\text{g}$).

251 We report our Mg isotope data in δ -notation (in per mil), the relative differences of $^x\text{Mg}/^{24}\text{Mg}$
 252 ratios of the sample to the bracketing standard, where $x = 25$ or 26 :

$$253 \delta^x\text{Mg} = ([^x\text{Mg}/^{24}\text{Mg}]_{\text{sample}}/[^x\text{Mg}/^{24}\text{Mg}]_{\text{standard}} - 1) \times 1000 \quad (\text{Eq. 1})$$

254 Mass-independent variations in ^{26}Mg due to ^{26}Al decay ($\delta^{26}\text{Mg}^*$) were determined by internally
 255 normalizing the measured $^{26}\text{Mg}/^{24}\text{Mg}$ ratio to $^{25}\text{Mg}/^{24}\text{Mg} = 0.12663$ (Catanzaro et al., 1966) using
 256 the exponential law with fractionation factor $\beta = 0.511$:

$$257 \delta^{26}\text{Mg}^* = (\delta^{26}\text{Mg} - (1 + \delta^{25}\text{Mg}/1000)^{(1/\beta)} - 1) \times 1000 \quad (\text{Eq. 2})$$

258

259

260 **Table 1:** Al-Mg data for Mg solution standard Cambridge-1, terrestrial rock standard San Carlos
 261 olivine, and bulk rock CV3 chondrite Allende, as well as unprocessed and processed standard
 262 solution isotope compositions to verify the absence of fractionation effects during column
 263 chemistry. n.r.: not reported.

Standards	$^{27}\text{Al}/^{24}\text{Mg}$	\pm	$\delta^{25}\text{Mg}$ [‰]	2SE	$\delta^{26}\text{Mg}^*$ [‰]	2SE	n
Cambridge-1			-1.374	0.020	-0.005	0.006	22
<i>literature</i> ^a			-1.343	0.036	n.r.	n.r.	49
<i>literature</i> ^b			-1.335	0.032	n.r.	n.r.	
San Carlos olivine	0.0021	0.0009	-0.162	0.009	-0.005	0.005	9
<i>literature</i> ^c	0.0031	0.0004	-0.16	0.07	0.00	0.05	36

Allende	0.1300	0.0039	-0.174	0.026	0.013	0.009	4
<i>literature</i> ^c	0.1294	0.0007	-0.16	0.03	0.03	0.05	
BCR-2	3.796	0.052					
<i>literature</i> ^c	3.76	0.04					
UWO ICP Std ^d			-0.329	0.027	0.018	0.022	12
UWO ICP Std ^e			-0.296	0.027	-0.013	0.031	8
UWO ICP Std ^f			-0.323	0.020	0.006	0.041	8
UWO ICP Std ^g			-0.309	0.023	-0.003	0.033	16
UWO ICP Std ^h			-0.318	0.018	0.006	0.021	28

264 ^a From An and Huang (2014). ^b From Vogl et al. (2020). ^c From Bouvier et al. (2013). ^d Unprocessed. ^e
265 Column processed. ^f Added to Mg-free matrices of terrestrial standards BCR-2 and BIR-1, column
266 processed. ^g All column-processed combined (b+c). ^h All standard solution measurements combined
267 (a+b+c).

268 **Table 2:** Three-stage Mg separation scheme modified from Wombacher et al. (2009)

Stage 1 + 2			Stage 3		
Resin: 1.2 ml of AG50W-X8 (200-400 mesh)					
<i>Step</i>	<i>Acid</i>	<i>Volume [ml]</i>	<i>Step</i>	<i>Acid</i>	<i>Volume [ml]</i>
Resin cleaning	4 M HCl	ca. 20	Resin conditioning	0.4 M HCl	3.6
Resin conditioning	10 M HCl	3.6	Load sample	0.4 M HCl	0.3
Load sample	10 M HCl	0.3	Matrix elution	0.4 M HCl	33.6
Mg + matrix elution	10 M HCl	3.2	Be, Ti, Al elution	0.15 M HF	7.2
Post-Mg step	8 M HCl	0.8	Pre-Mg step	1 M HCl	1
Ca elution	8 M HCl	14.2	Mg elution	1 M HCl	20
Fe elution	4 M HCl	10	Wash out	6 M HCl	5
Wash out	4 M HCl	10			

269

270

271 *2.5 Lead isotope analysis*

272 Pervasive terrestrial Pb contamination (Pb_c) of meteoritic samples is a major concern. Meaningful

273 ^{207}Pb - ^{206}Pb isochrons are linear arrays of a two-component system made up of radiogenic Pb (Pb_r),

274 produced by the decay of ^{235}U and ^{238}U , and, in most cases, initial Pb (Pb_i). Adding a third

275 component, Pb_c , to this system can pull data points off the $Pb_r + Pb_i$ line towards the Pb_c isotope

276 composition, affecting the linearity of the isochron. This requires effective removal of all Pb_c from
277 samples (e.g., Connelly et al., 2017). Alternatively, Pb-Pb ages can also be accurately calculated
278 from linear arrays made of Pb_r and Pb_c , if the sample contains no initial Pb (Pb_i). This is achieved
279 through a step-wise dissolution procedure where crushed samples were leached first by weak acids
280 to remove surface contamination before increasingly strong acids were used to dissolve minerals
281 (Connelly and Bizzarro, 2009). The pyroxene fraction, totalling ~ 61 mg, underwent such a
282 procedure as described in Bouvier et al. (2011). The sample was ultrasonicated in 0.5 M HBr, then
283 Milli-Q water and twice in 0.5 M HNO_3 for 15 minutes each. Subsequent steps involved a hot-
284 plate fluxing in 6 M HCl and 7 M HNO_3 (twice) at 120°C for an hour, ultrasonication in 1 M HF
285 for an hour and 12-hour-long hot-plate fluxing at 90°C in 1 M HF (Table 3). Since the first three
286 steps are designed to remove terrestrial Pb and are unlikely to have any chronological meaning
287 (e.g., Merle et al., 2020), these were collected all together in one beaker. Steps 5 and 6 (both 7 M
288 HNO_3) were also mixed, as the amount of Pb (0.13 ng total) would have been insufficient for a
289 precise determination of the Pb isotope ratios if these two leaching steps had been analyzed
290 separately.

291 The collected washes from each step were then dried down before being dissolved in a conc. HF-
292 HNO_3 mixture for ~48 hrs. After ultrasonication for an hour and subsequent drying down, the
293 samples are re-fluxed in conc. HNO_3 , before being ultrasonicated once again and taken up in 6 M
294 HCl to convert the samples to their chloride form. Prior to Pb separation, the fractions were
295 analyzed for their U and Pb concentrations using Thermo Fischer iCAP Qc quadrupole ICP-MS at
296 the University of Western Ontario (UWO).

297

298 **Table 3:** Leaching protocol for samples prepared for Pb isotopic analysis.

Step	Acid	Processing	Duration	Repeat	Collected as...	
1	0.5 M HBr	Ultrasonication washing	15 min	once	Wash-1 (W1)	Combined as W1-3
2	Water	Ultrasonication washing	15 min	once	Wash-2 (W2)	
3	0.5 M HNO ₃	Ultrasonication washing	15 min	twice	Wash-3 (W3)	
4	6 M HCl	Hot plate fluxing at 120°C	1 hr	once	Wash-4 (W4)	Combined as W5+6
5	7 M HNO ₃	Hot plate fluxing at 120°C	1 hr	once	Wash-5 (W5)	
6	7 M HNO ₃	Hot plate fluxing at 120°C	1 hr	once	Wash-6 (W6)	
7	1 M HF	Ultrasonication washing	1 hr	once	Wash-7 (W7)	
8	1 M HF	Hot plate fluxing at 90°C	12 hrs	once	Wash-8 (W8)	

299

300 For Pb separation, PTFE columns with 100 µl of AG1-X8 (200-400 mesh) anion-exchange resin
301 are used following the procedure described in Bouvier et al. (2007), which was based on Marsh et
302 al. (1978). Before loading, the samples are dried down and taken up in 1.5 M HBr and centrifuged
303 to separate any remaining undissolved materials. Matrix elements are first eluted using 1.5 M HBr
304 after the sample load. Lead is finally eluted using 0.5 M HNO₃. This procedure is repeated a second
305 time to guarantee sufficient purification of Pb from matrix elements. The eluted matrix elements
306 were collected together for U isotope analysis.

307 Lead isotope compositions were determined using a Thermo Fisher *Neptune Plus* MC-ICP-MS
308 equipped with a Cetac Aridus II desolvating nebulizer at Université Clermont-Auvergne (UCA)
309 and followed the protocols of Ancellin et al. (2019). The total Pb masses in the various leachates
310 range from 0.13 ng to 3 ng and were taken up in 0.5 to 1.5 ml of 0.05M HNO₃, depending on their
311 Pb concentration. Sample and standard solutions were both doped with the NBS 997 Tl standard
312 to a concentration of 1 ppb and bracketed with the NBS 981 Pb standard (Abouchami et al., 2000)
313 to correct internally and externally for instrumental mass fractionation. In addition, the radiogenic
314 NBS 983 Pb standard was included in the analytical sessions as a secondary standard. Before every
315 standard and sample measurement, a background measurement with baseline analysis was taken,
316 which was subtracted from the standards or samples during data reduction. Peak center and
317 baseline analysis were performed on all standard measurements but only on samples diluted in 1.5
318 ml of 0.05M HNO₃, to ensure sufficient material for was available for the isotopic analysis. Two
319 samples, where enough ²⁰⁴Pb was present, were analyzed in a Faraday cup (FC) setup (EC W1-3
320 and ECW4). The other samples were analyzed with a secondary electron multiplier (SEM) on
321 ²⁰⁴Pb, as the intensity in a FC setup would be below 0.3 mV. This setup has the disadvantage that
322 ²⁰⁸Pb cannot be measured on H4 due to the fixed spacing imposed by the pin attaching H3 and H4
323 (Table 4). Sample analyses consisted of 60 cycles with 8.4 s integration time in the FC setup,
324 whereas the analyses in the SEM setup consisted of only 25 cycles, with no peak center or baseline
325 analysis.

326 The measured Pb isotopic data were reduced offline in Excel and R, and “inverse” isochrons were
327 calculated using IsoplotR (Vermeesch, 2018) and Isoplot 4.15 in Excel (Ludwig, 2008). Total
328 procedural blanks ranged from 0.3 to 1.8 pg, resulting in sample/blank ratios from 131 to 3808,
329 and compositions were corrected accordingly, following the equations provided in the Appendix

330 of Bouvier et al. (2007). Errors on $^{204}\text{Pb}/^{206}\text{Pb}$ and $^{207}\text{Pb}/^{206}\text{Pb}$ ratios for isochron regression were
 331 either the correlated errors or the daily external reproducibility of the NBS 981 or NBS 983
 332 standards (depending on which standard is more representative of the sample ratios), whichever is
 333 larger.

334 The full session Pb isotope data averages are reported in Table 5.

335

336 **Table 4:** Overview of cup configurations used for Pb isotope analysis with all Faraday cups (FC)
 337 or with central Secondary Electron Multiplier (SEM).

Cup	L3	L2	L1	C (or SEM)	H1	H2	H3
Pb (FC)	^{202}Hg	^{203}Tl	^{204}Pb	^{205}Tl	^{206}Pb	^{207}Pb	^{208}Pb
Pb (SEM)		^{202}Hg	^{203}Tl	^{204}Pb	^{205}Tl	^{206}Pb	^{207}Pb

338

339

340 *2.6 Uranium isotope analysis*

341 Analysis of the U isotope composition of a bulk rock sample and leached pyroxenes (included
 342 W5+6, W7, W8 and R) of EC 002 were performed with a Neptune MC-ICP-MS at the Leibniz
 343 Universität Hannover, following the protocols of Roebbert et al. (2021). The collected matrix
 344 elements of two aliquots of the terrestrial standard BCR-2, equal to ~ 5 ng and ~ 20 ng of U that
 345 underwent prior Pb separation (as for EC 002), were analyzed along with the meteorite samples to
 346 ensure that no U isotope fractionation occurred during the separation of Pb (Table 6). The ^{236}U -
 347 ^{233}U double-spike solution IRMM 3636 was used to correct for instrumental mass-bias (Richter et
 348 al., 2010). Each sample was analyzed twice, bracketed by the CRM-112a standard, over 60 cycles
 349 at 4 s integration time. Concentrations between standards and samples were individually matched
 350 and intensities on ^{238}U were 2.5 V and 12 V for the leached pyroxene and the bulk rock fraction,

351 respectively. Because of the relatively low ion beams, all U isotopes, including ^{238}U , were
352 measured using amplifiers with 10^{-11} Ohm resistors. The standard IRMM 184 was analyzed along
353 with the samples and its results are reported in Table 6.

354

355 3. Results

356 *3.1 Al-Mg systematics*

357 Magnesium isotope compositions and $^{27}\text{Al}/^{24}\text{Mg}$ ratios of the analyzed mineral fractions of EC
358 002 are summarized in Table 7. The $^{27}\text{Al}/^{24}\text{Mg}$ ratios span a range from 0.98 for the Px fraction to
359 34.4 in feldspars. Excesses in ^{26}Mg are not resolved from the terrestrial fractionation line for the
360 bulk rock and Px fractions at $\delta^{26}\text{Mg}^* = 0.013 \pm 0.020\text{‰}$ and $0.010 \pm 0.010\text{‰}$, respectively. In the
361 feldspar fractions and fines fraction, radiogenic ^{26}Mg excesses are resolved and range from 0.132
362 $\pm 0.010\text{‰}$ to $2.058 \pm 0.015 \text{‰}$. A regression line through the data is obtained with Isoplot 4.15
363 (Ludwig, 2008), with a slope corresponding to the $^{26}\text{Al}/^{27}\text{Al}$ at the time of crystallization of $[8.89$
364 $\pm 0.79] \times 10^{-6}$ and an y-intercept $\delta^{26}\text{Mg}^*_0$ of $-0.058 \pm 0.026\text{‰}$ (both 2σ), representing the initial
365 Mg isotope composition at time of crystallization (MSWD = 4.6). This $^{26}\text{Al}/^{27}\text{Al}_0$ corresponds to
366 a crystallization age of 1.83 ± 0.12 Ma after CAI formation (Fig. 2a). This corresponds to an
367 absolute age of 4565.47 ± 0.20 Ma when anchored to the canonical $^{26}\text{Al}/^{27}\text{Al}$ value of $[5.23 \pm 0.13]$
368 $\times 10^{-5}$ (Jacobsen et al., 2008) and the absolute age of CAIs at 4567.30 ± 0.16 Ma (Connelly et al.,
369 2012). When anchored to the D'Orbigny angrite (Pb-Pb age: 4563.51 ± 0.29 (Amelin, 2008b;
370 Tissot et al., 2017), initial $^{26}\text{Al}/^{27}\text{Al}$: $[3.98 \pm 0.15] \times 10^{-7}$ (Schiller et al, 2015), $[5.06 \pm 0.15] \times 10^{-7}$
371 (Spivak-Birndorf et al., 2009)), the ^{26}Al - ^{26}Mg formation age of EC 002 is 4566.72 ± 0.32 Ma and
372 4566.47 ± 0.40 Ma, calculated from the initial $^{26}\text{Al}/^{27}\text{Al}$ ratios of Schiller et al. (2015) and Spivak-

373 Birndorf et al. (2009), respectively. These ages are 1.26 ± 0.38 Myr and 1.01 ± 0.45 Myr,
374 respectively, older than when anchored to CAIs.

375 Regressing the subset of feldspar fractions results in an initial $^{26}\text{Al}/^{27}\text{Al}$ of $[8.36 \pm 0.49] \times 10^{-6}$ and
376 a $\delta^{26}\text{Mg}^*_0$ of $-0.010 \pm 0.032\text{‰}$ (MSWD = 0.0017), equal within the stated uncertainties of the
377 regression through the complete dataset (Fig. 2b).

378 The $\delta^{25}\text{Mg}$ values of EC 002 fractions lie between $-0.729 \pm 0.012\text{‰}$ and $-1.104 \pm 0.013\text{‰}$, which
379 are distinctly lighter than the compositions of the terrestrial San Carlos olivine and the CV3
380 chondrite Allende.

381 382 *3.2 Uranium isotope composition*

383 The results for the $^{238}\text{U}/^{235}\text{U}$ ratios in both the bulk rock and the re-combined W5-R leachates and
384 residue of EC 002 are summarized in Table 6. Both samples were measured twice, with their
385 weighted average $^{238}\text{U}/^{235}\text{U}$ ratios being 137.819 ± 0.010 for the bulk rock, and 137.766 ± 0.027
386 for the leachates (both 2σ). The uncertainty associated with the analyses, particularly on the
387 leachates, is mainly due to the low U concentration of this sample (2.5 V on ^{238}U from ≈ 5 ng U).

388

389 **Table 5:** Mass, Pb and U elemental data, and Pb isotope data for EC 002 and full session Pb isotope data for standards NBS 981 and
390 NBS 983. Nantan model ages are calculated from the primordial Pb isotope composition, represented by the Nantan troilite (Blichert-
391 Toft et al., 2010). SK model ages are calculated from the modern terrestrial Pb composition as given by Stacey and Kramers (1975).

Sample	mass [mg]	Pb [ng]	U [ng]	Sample/ Blank	$^{206}\text{Pb}/$ ^{204}Pb raw	$^{204}\text{Pb}/^{206}\text{Pb}$	2SE [%]	$^{207}\text{Pb}/^{206}\text{Pb}$	2SE [%]	Nantan model age [Ma]	±	SK model age [Ma]	±
EC-W1-3 *	1.6	3.2	2.2	3808	41.5	0.024092	0.015	0.72121	0.0024	4529.73	0.10	4571.38	1.54
EC-W4 *	1.0	1.3	6.5	1549	42.5	0.023515	0.035	0.71755	0.0054	4526.72	0.16	4565.57	1.49
EC-W5+6 *	0.3	0.1	0.2	131	343.9	0.00268	3.42	0.6370	0.045	4568.19	1.07	4571.84	1.00
EC-W7 &	4.1	0.6	0.3	302	10902.7	—	—	0.6250	0.021	4566.15	0.30	4566.14	0.30
EC-W8 &	44.2	0.8	0.8	1331	6953.1	0.000122	7.25	0.62538	0.0085	4565.63	0.14	4565.78	0.14
EC-R &	42.9	1.2	1.0	1366	16604.9	0.000039	21.3	0.62513	0.0085	4565.91	0.14	4565.96	0.14
<i>Standards</i>						$^{206}\text{Pb}/^{204}\text{Pb}$ <i>corr</i>	<i>2SD</i> [%]	$^{207}\text{Pb}/^{206}\text{Pb}$ <i>corr</i>	<i>2SD</i> [%]		<i>n</i>		
NBS 983													
- SEM setup 0.25 ppb						2762.5	1.29	0.071186	0.25		5		
- FC setup 1 ppb						2721.75	7.85	0.07124	0.60		10		
- literature ^a						2695.4	5.4	0.07120	0.06				
NBS 981													
- SEM setup 0.5 ppb						16.9437	0.26	0.91466	0.06		10		
- FC setup 1 ppb						16.9412	0.26	0.91475	0.06		14		
- literature ^b						16.9405	0.01	0.91475	-				

^a Catanzaro et al. (1968). ^b Abouchami et al. (2000). * measured using Faraday cup setup. & measured using SEM setup.

392

393

394 **Table 6:** Weighted average U isotope compositions of EC 002 bulk rock and pyroxene samples and of BCR-2 terrestrial basalt standard,
 395 as well as repeat analyses of the IRMM 184 U isotope standard (Richter et al., 2010).

Sample	Mass [mg]	U [ng]	$^{238}\text{U}/^{235}\text{U}$	2σ	n	^{238}U intensity [V]
EC 002 Bulk rock	~120	15	137.819	0.007	2	12
EC 002 Px leachates	~160	5	137.766	0.027	2	2.5
BCR-2 a	11.8	20	137.797	0.017	2	15
BCR-2 b	2.9	5	137.815	0.046	2	4
IRMM 184			137.668	0.008	8	2.5 - 15

396

397 **Table 7:** Al-Mg data for EC 002 fractions.

Sample	$^{27}\text{Al}/^{24}\text{Mg}$	\pm	$\delta^{25}\text{Mg}$ [‰]	2SE	$\delta^{26}\text{Mg}^*$ [‰]	2SE	n
EC Px	0.98	0.05	-0.825	0.014	0.010	0.010	4
EC Bulk rock	1.05	0.18	-0.736	0.037	0.013	0.020	7
EC fines	3.45	0.17	-0.840	0.054	0.139	0.006	8
EC fsp2-4	6.33	0.32	-0.729	0.012	0.370	0.004	4
EC fsp1+8	9.03	0.45	-1.104	0.013	0.532	0.005	4
EC fsp7	10.7	0.54	-0.957	0.010	0.636	0.016	4
EC fsp6	34.4	1.72	-0.956	0.015	2.058	0.015	4

398

399 **Table 8:** Pb isotope data and corresponding Pb-Pb ages of merrillites analyzed by SIMS

Merrillites	$^{207}\text{Pb}/^{204}\text{Pb}$	$\pm 1\text{s.e.}$	$^{206}\text{Pb}/^{204}\text{Pb}$	$\pm 1\text{s.e.}$	$^{207}\text{Pb}/^{206}\text{Pb}$	$\pm 1\text{s.e.}$	$^{207}\text{Pb}/^{206}\text{Pb}$ corr [Ma] ^a	$\pm 1\text{s.e.}$	$^{207}\text{Pb}/^{206}\text{Pb}$ age [Ma] ^b	$\pm 1\text{s.e.}$
M1	2368	352	3799	565	0.6245	0.0036	0.6226	0.0037	4561.0	8.6
M4	1738	191	2773	306	0.6243	0.0033	0.6226	0.0032	4560.9	7.5
M5	24360	8636	38710	13750	0.6271	0.0023	0.6268	0.0023	4570.7	5.2
M8-1	5252	889	8429	1426	0.6233	0.0027	0.6228	0.0025	4561.4	5.8
M8-2	8178	1597	13080	2554	0.6254	0.0032	0.6249	0.0032	4566.5	7.4
M12	5315	1474	8442	2342	0.6244	0.0048	0.6223	0.0044	4560.4	10.2
M14	2620	429	4164	682	0.6243	0.0038	0.6225	0.0036	4560.8	8.4
Weighed mean age \pm 95% confidence interval (Ma)									4564.3 \pm 5.2	

400 ^a ^{207}Pb - ^{206}Pb corr. compositions are corrected from primordial Pb contribution. ^b ^{207}Pb - ^{206}Pb ages are calculated using the $^{238}\text{U}/^{235}\text{U}=137.82 \pm 0.01$ of the measured

401 whole-rock composition of EC 002 and correlated errors between Pb and U isotopic measurements.

402 3.3 MC-ICP-MS Pb isotopic analysis

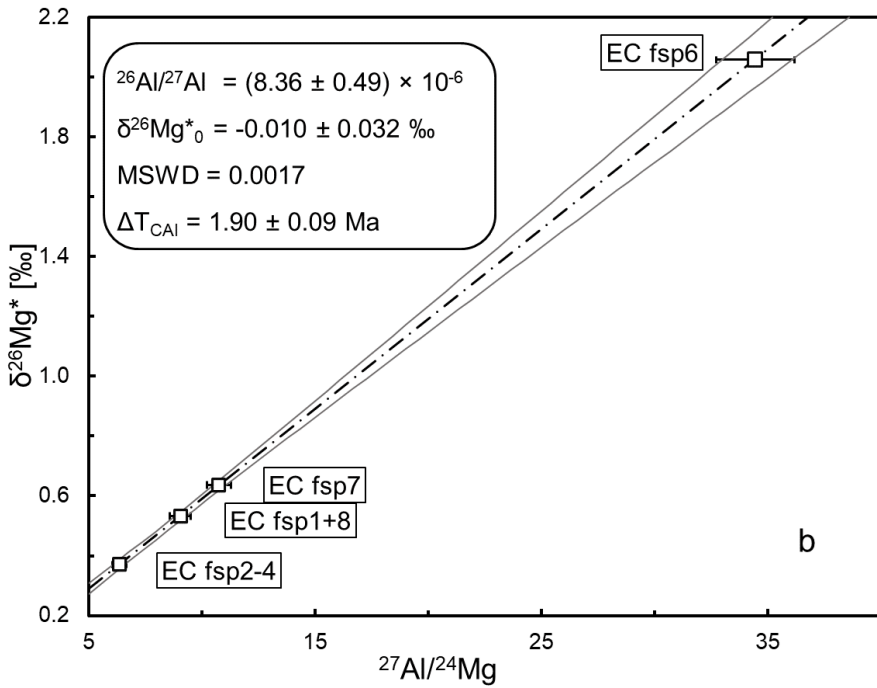
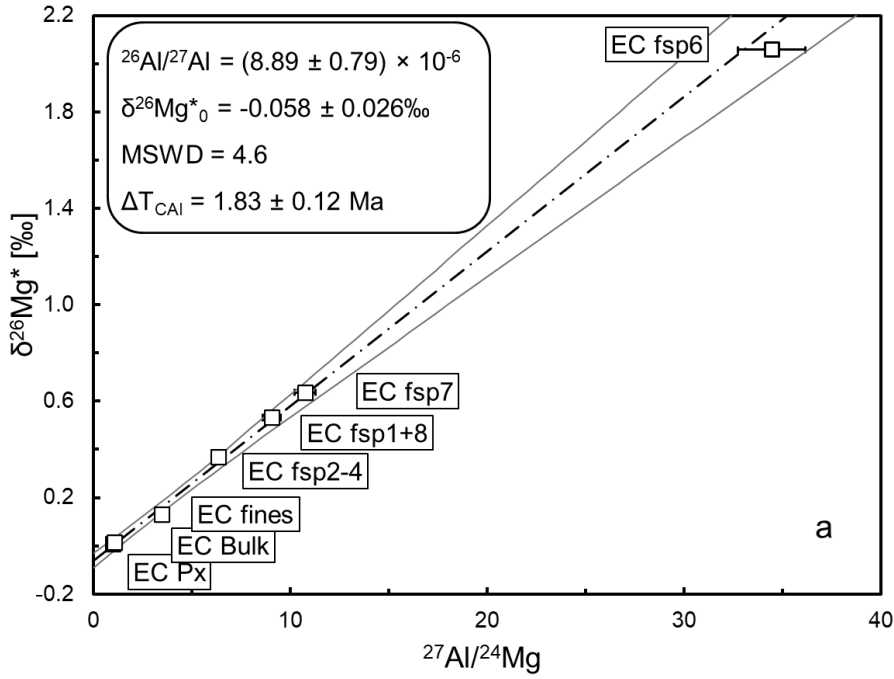
403 Lead isotope data for leachates and residue of a pyroxene fraction are presented in Table 5.
404 Leachates W7 and W8, as well as the residue, are highly radiogenic, with raw $^{206}\text{Pb}/^{204}\text{Pb}$ ratios
405 ranging from 6953 to 16604. The first leachates (W1-3 to W5+6) have $^{206}\text{Pb}/^{204}\text{Pb}$ ratios in the
406 range of 41 to 344, with a low sample-to-blank ratio of 131 for W5+6. The sample-to-blank ratios
407 of the remaining leachates range from and 1549 to 3808, with the exception of W7 at 302. In an
408 “inverse” $^{204}\text{Pb}/^{206}\text{Pb}$ - $^{207}\text{Pb}/^{206}\text{Pb}$ space, the leachates and residues are significantly overdispersed
409 (MSWD = 391). Nevertheless, they plot along a line projecting towards the point of terrestrial Pb
410 as defined by the model of Stacey and Kramers (1975), indicating that the pyroxene in the sample
411 is free of initial, primordial Pb (Fig. 3) and that the Pb isotope compositions of the leachates and
412 residues are a mixture of only radiogenic Pb ($^{204}\text{Pb} = 0$) and terrestrial Pb. This has been previously
413 observed in the angrites Sahara 99555 (Amelin, 2008a; Connelly et al., 2008) and NWA 1670
414 (Schiller et al., 2015), as well as some chondrules (Bollard et al., 2017).

415 A statistically meaningful isochron (probability of fit > 0.05) can be calculated using leachates
416 W4, W7, W8 and the residue R, however, W7 is omitted due to its negative $^{204}\text{Pb}/^{206}\text{Pb}$ ratio after
417 background- and blank-correction, despite it still plotting along the isochron. This leaves three
418 fractions defining the isochron (MSWD = 3.3), resulting in an age of 4565.87 ± 0.30 Ma (2σ) (Fig.
419 4), calculated with the $^{238}\text{U}/^{235}\text{U}$ ratio 137.766 ± 0.027 , determined from the pyroxene fraction. All
420 ages given include error propagation on the U isotope composition, except for individual
421 $^{207}\text{Pb}^*/^{206}\text{Pb}^*$ model ages of the leachates and residue.

422 The $^{207}\text{Pb}^*/^{206}\text{Pb}^*$ model ages (Table 5), calculated from both initial Pb as represented by the
423 Nantan troilite (Blichert-Toft et al., 2010) and the Stacey & Kramers (SK) point representing
424 terrestrial Pb, show no age difference for a given highly radiogenic leachate or the residue, as they

425 plot close enough to the y-axis to not make any significant difference in y-intercept
426 ($= {}^{207}\text{Pb}^*/{}^{206}\text{Pb}^*$). The agreement between those leachates and the residue also demonstrates that
427 no fractionation occurred due to the use of HF in the leaching procedure. For the less radiogenic
428 leachate W4, the Nantan and SK model ages differ significantly, the SK model ages being in line
429 with the other leachates and residue of its respective fraction, indicating once again that there likely
430 was no primordial Pb present in the sample. The weighted average of the SK model ages for W4,
431 W8 and R results in an Pb-Pb age of 4565.87 ± 0.40 Ma (MSWD = 2.1). Adding the SK model
432 age for W7, where only its ${}^{207}\text{Pb}/{}^{206}\text{Pb}$ ratio is considered, to the weighted average gives an age of
433 4565.90 ± 0.36 Ma (MSWD = 1.7).

434

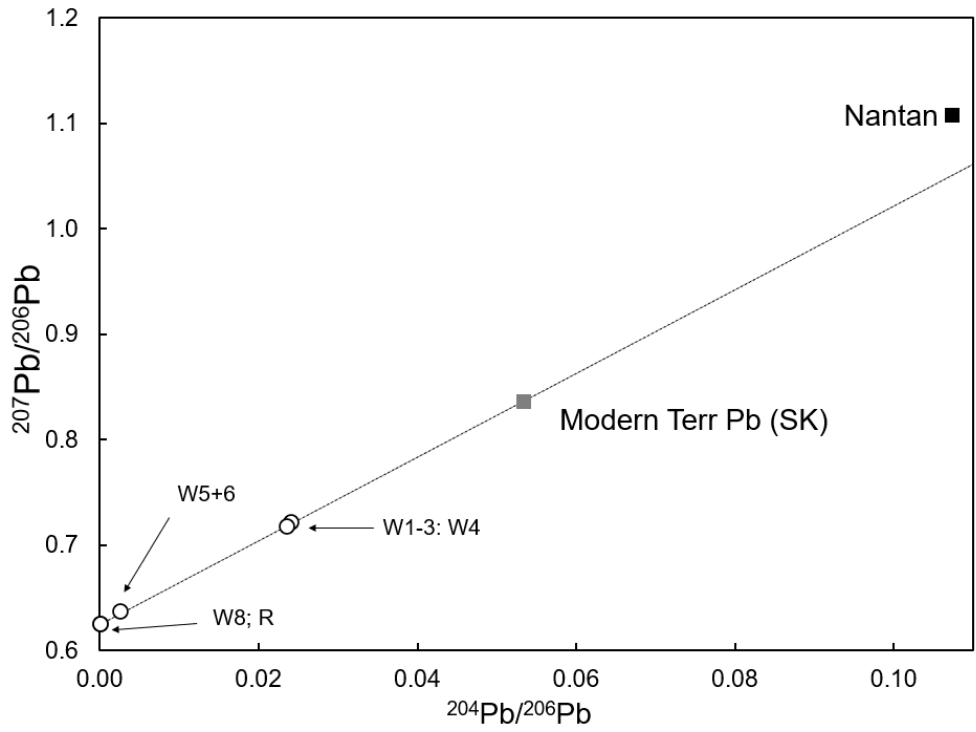


435

436 **Figure 2:** Internal ^{26}Al - ^{26}Mg isochrons defined by a) pyroxene, bulk rock, fine-grained and
 437 plagioclase fractions, corresponding to a $^{26}\text{Al}/^{27}\text{Al}$ ratio of $(8.89 \pm 0.79) \times 10^{-6}$ and a formation age
 438 of 1.83 ± 0.12 Ma after CAI formation, based on the canonical $^{26}\text{Al}/^{27}\text{Al}$ abundance (Jacobsen et

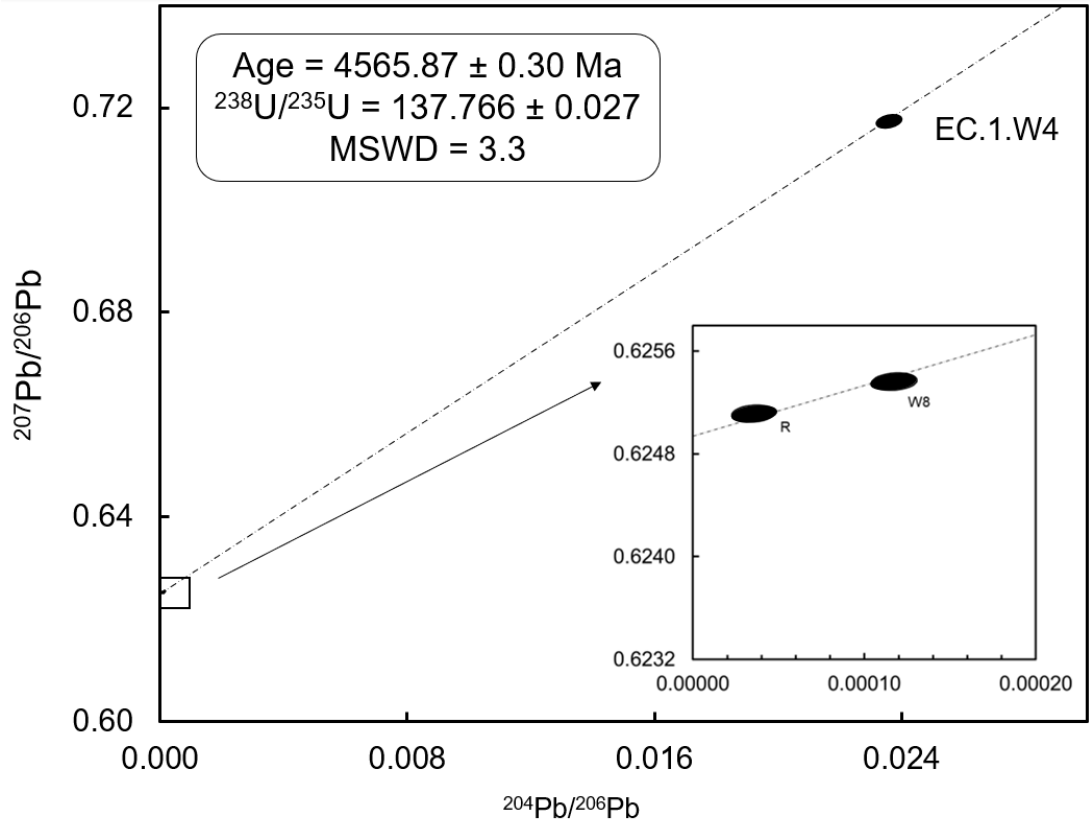
439 al., 2008); and **b**) only plagioclase fractions, corresponding $(8.36 \pm 0.49) \times 10^{-6}$ and a formation
440 age of 1.90 ± 0.09 Ma after CAI formation. Uncertainties are 2SE errors on $\delta^{26}\text{Mg}^*$ measurements
441 and $\pm 5\%$ on $^{27}\text{Al}/^{24}\text{Mg}$ ratios.

442



443

444 **Figure 3:** Regression line through the analyzed Pb fractions. Leachate W7 is omitted due to its
445 negative blank-corrected $^{204}\text{Pb}/^{206}\text{Pb}$ ratio. Also plotted are data points for the Nantan troilite,
446 representing the primordial Pb composition of the Solar System (Blichert-Toft et al., 2010), and
447 the Stacey and Kramers (1975) data point representing the terrestrial Pb composition. Symbols of
448 data points have been enlarged for clarity, larger than their uncertainties.



449

450 **Figure 4:** Inverse Pb-Pb isochron diagram for Erg Chech 002. The uncertainty on W4 is smaller
 451 than the ellipse, as it is enlarged for clarity. The inset shows the highly radiogenic leachate W8
 452 and residue. Leachate W7 would plot along the regression line in negative $^{204}\text{Pb}/^{206}\text{Pb}$ vs.
 453 $^{207}\text{Pb}/^{206}\text{Pb}$ space.

454

455 *3.4 Pb-Pb merrillite ages by SIMS*

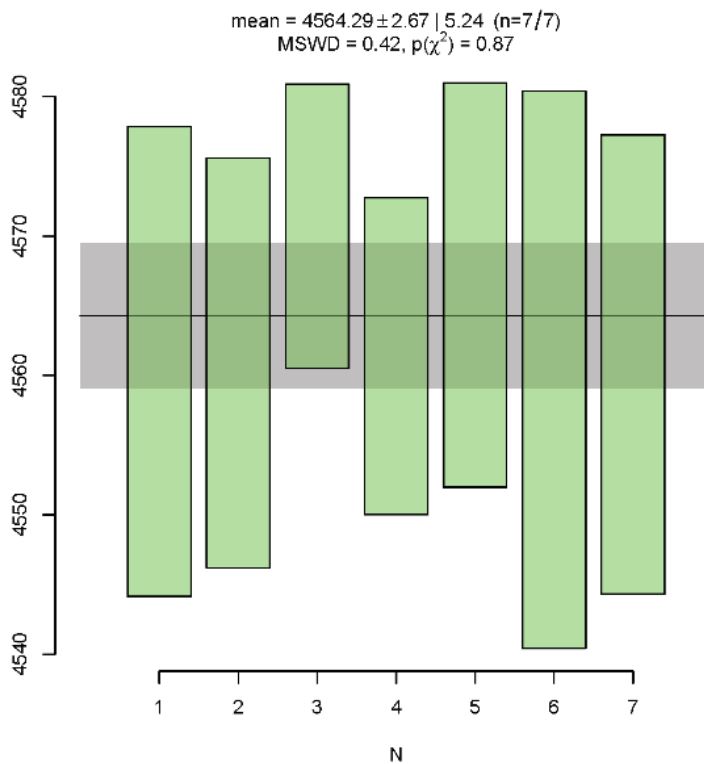
456 Merrillites have a homogeneous and H-free composition of $\text{Ca}_{8.5}\text{Na}_{0.8}\text{Mg}_{0.7}\text{Fe}_{0.3}(\text{PO}_4)_7$ (n=4). Raw
 457 Pb isotopic data were reduced and calibrated against measurements of standard Madagascar apatite
 458 (485 Myr, Thomson et al., 2012) using in-house-developed software (ZIPS version 3.1.1, by
 459 Christopher Coath). Common lead correction was made using the composition and model of
 460 Stacey and Kramers (1975) and primordial Pb isotopic composition reported for Canyon diablo

461 troilite by Tatsumoto et al. (1973). In the absence of suitable merrillite standard, U and Pb
462 secondary ion yields were not known precisely, as such, U/Pb systematics cannot be measured
463 with satisfactory precision. But if the analysed points were used as hypothetical standard, all
464 results plotted tightly into a concordant cluster, strongly suggesting that the observed discrepancy
465 derived chiefly from matrix effect (i.e., using apatite as standards for merrillite samples). Due to
466 the extremely high radiogenic Pb isotopic ratios measured for merrillites in Erg Chech 002
467 (common lead $^{204}\text{Pb} < 0.1$ cps, and $^{204}\text{Pb}/^{206}\text{Pb}$ ratio on the order of 10^{-4} - 10^{-5}), the choice of the
468 apatite standard (comparably aged apatite or merrillite standards are not available) as well as
469 primordial Pb isotopic ratio did not affect the calculated Pb-Pb merrillite ages.

470 The seven ^{207}Pb - ^{206}Pb ages of individual analysis (Table 8) were calculated from radiogenic $^{207}\text{Pb}^*$ -
471 $^{206}\text{Pb}^*$ ratio using the U decay constants from Steiger and Jäger (1977), with uncertainties of 0.2-
472 0.4% ($\pm 2\sigma$), and a $^{238}\text{U}/^{235}\text{U}$ ratio of 137.82 ± 0.01 (from our whole-rock measurement). In the
473 end, the ^{207}Pb - ^{206}Pb merrillite age for EC 002 was reported with a weighted mean of 4564.3 ± 5.2
474 Ma (2σ), as seen in Fig. 5.

475

476



477

478 **Figure 5:** Weighted mean age in Ma with 1σ and 95% confidence interval for 7 spot analyses of
479 6 individual merrillite grains of EC 002 (Table 8).

480

481 4.Discussion

482 *4.1 Lead isotope systematics*

483 Even though only three fractions make up our ^{207}Pb - ^{206}Pb isochron, with a rather non-radiogenic
484 leachate and a highly radiogenic leachate and residue, it is statistically meaningful (Pb-Pb age =
485 4565.87 ± 0.09 Ma, MSWD = 3.3, probability of fit = 0.068). Three-point isochrons are generally
486 considered to be unreliable, as the regression may be subject to increased bias by outlying data
487 points and data filtering. These concerns are valid, however, in this case, the highly radiogenic
488 leachates used in the regression argue for the reliability of our age determined from the isochron:

489 Pb-Pb ages are determined by the intercept of the regression line with the y-axis in an “inverse”
490 Pb-Pb diagram ($^{207}\text{Pb}^*/^{206}\text{Pb}^*$). In the case of EC 002, the later leachates and residue are highly
491 radiogenic, defining a precise intercept, making any scatter of the less radiogenic leachates less
492 influential on the Pb-Pb age. This is visible when including leachates that were not used in the
493 final isochron and in the $^{207}\text{Pb}^*/^{206}\text{Pb}^*$ model ages. For example, including W5+6 instead of, or in
494 addition to, W4 results in overdispersed errorchrons with an absolute age of 4565.78 ± 0.31 Ma
495 (MSWD = 8.3) or 4565.91 ± 0.50 (MSWD = 27), respectively. In addition to this example, the
496 weighted average of the $^{207}\text{Pb}^*/^{206}\text{Pb}^*$ SK model ages of the W4, W8 and R fractions ($4565.87 \pm$
497 0.36 Ma, MSWD = 2.1) is consistent with the isochron age, as is the weighted average of the model
498 ages when including the W7 leachate (4565.90 ± 0.40 Ma, MSWD = 1.7). Hence, we interpret the
499 absolute age of 4565.87 ± 0.30 Ma as defined by the isochron as the crystallization age of EC 002
500 (Fig. 4).

501 Seven SIMS measurements of merrillites are consistent within errors and provide a weighted mean
502 age of 4564.3 ± 5.2 Ma (2σ). While relatively much less precise than the age determination by
503 solution analyses, this Ca-phosphate formation age is similar within uncertainty to the pyroxene
504 Pb-Pb age. This reinforces the conclusions made by Barrat et al. (2021) suggesting a rapid cooling
505 of the parent body. The diffusion rate of Pb in Ca-phosphate is faster than in pyroxene and therefore
506 the closure temperature of the U-Pb system in phosphate is protracted toward lower temperatures
507 (Cherniak, 2001) and more prone to resetting temperatures such as those sustained during impact
508 events. This additional thermochronological information supports that EC 002 cooled rapidly after
509 crystallization and was neither subjected to substantial impact events nor extensively thermally
510 metamorphosed after ~ 4559 Ma. Detailed thermochronological records can be used to constrain
511 EC 002’s parent body accretion history and size, and burial depth.

512

513 *4.2 Uranium-corrected Pb-Pb age of Erg Chech 002*

514 The $^{238}\text{U}/^{235}\text{U}$ ratio of bulk rock EC 002 is 137.819 ± 0.010 , distinct from the combined leachates
515 (W5 to R) of the Px fractions at 137.766 ± 0.027 by 0.38 ± 0.21 ‰. The resulting U-corrected Pb-
516 Pb ages are 4566.43 ± 0.14 Ma and 4565.87 ± 0.30 Ma, respectively, a difference of 0.56 ± 0.28
517 Myr. With that in mind, the choice of $^{238}\text{U}/^{235}\text{U}$ ratio for calculating the Pb-Pb age is an important
518 consideration, as the interpretation of the result may change depending on which value is used.

519 Since the Pb isotope systematics of EC 002 were determined on a pyroxene fraction of EC 002,
520 we use the $^{238}\text{U}/^{235}\text{U}$ ratio of the same mineral fraction to calculate the U-corrected Pb-Pb age. The
521 Pb-Pb ages in achondrites are usually determined from leachates and residues of pyroxene, which
522 incorporate more U (e.g., Amelin, 2008b; Bouvier et al., 2011). The U isotope compositions of the
523 same meteorites are however determined on unleached whole-rocks, sometimes from different
524 laboratories (Brennecka and Wadhwa, 2012; Tissot et al., 2017). If U contamination in secondary
525 minerals formed during desert weathering (e.g. carbonates, sulfates) or U isotope heterogeneity
526 between pyroxenes and bulk rock samples exist in dated samples, U-corrected Pb-Pb ages would
527 require further correction. Ideally, whenever Pb-Pb ages are determined from mineral fractions, so
528 should the $^{238}\text{U}/^{235}\text{U}$ ratio. In practice, however, this is most often hampered by the large sample
529 amounts necessary (> 100 mg) to ensure sufficient U to obtain a precise analysis. Analyzing the
530 same mineral fractions for U and Pb isotopic compositions ensures consistency and avoids the
531 assumption that there are no U isotope variations between pyroxene and bulk rock fractions of the
532 sample or contribution from terrestrial U contamination or secondary minerals (e.g., carbonates)
533 in whole-rock composition, which is a great concern for desert find meteorites.

534 The difference in $^{238}\text{U}/^{235}\text{U}$ ratio between the bulk rock and pyroxene fractions requires an
535 explanation and could reasonably have originated through i) internal heterogeneities between
536 minerals and the bulk rock sample; ii) fractionation during the acid-leaching procedure; or iii)
537 fractionation during column chemistry and/or isotopic analysis.

538 The third possibility can be ruled out, as two BCR-2 aliquots underwent the same chromatographic
539 and analytical procedures with comparable amounts of U as the two EC 002 samples without
540 producing any evidence of fractionation within their statistical uncertainties (Table 6). These
541 results also rule out any significant interference on mass 235 during analysis of small U amounts,
542 as suggested by Connelly et al. (2019) to explain an outlier $^{238}\text{U}/^{235}\text{U}$ ratio in a troilite nodule of
543 IVA meteorite Muonionalusta (Blichert-Toft et al., 2010).

544 Previously, Brennecka and Wadhwa (2012) reported a slightly lighter U isotope composition (by
545 0.30 ± 0.25 ‰) of the leachate fraction compared to the residue of the angrite NWA 6291, for
546 which they invoked the removal of terrestrial components during leaching. Tissot et al. (2017),
547 however, pointed out that the leachate contained ~ 64% of total U of the sample and that such an
548 amount was unlikely to be exclusively of terrestrial origin, and instead argued for U isotope
549 variability between pyroxene and phosphate, caused by magmatic differentiation, to explain
550 different $^{238}\text{U}/^{235}\text{U}$ ratios between plutonic and volcanic angrites.

551 In the case of EC 002, the early leachates (W1 to W4) contain almost 80% of total U of the sample.
552 Following the argumentation of Tissot et al. (2017), this suggests that the origin of the distinct
553 $^{238}\text{U}/^{235}\text{U}$ ratios is isotopic heterogeneity between pyroxene and any remaining minerals. The acid-
554 leaching procedure should not have affected the U isotope composition of the pyroxene, as the
555 acids used in the early leaching steps (0.5 M HBr, water, 0.5 M HNO₃ and 6 M HCl; Table 3),
556 whose leachates were not included in the analyzed fraction, do not dissolve pyroxene. Rather, they

557 dissolve feldspar, phosphates, troilite, as well as any potential secondary terrestrial components,
558 that may have remained attached to pyroxene grains after hand-picking (Bouvier et al., 2005; Ito
559 et al., 2019). Therefore, the U isotope composition of the “re-combined residue” consisting of
560 leachates W5-8 and the residue from the leaching procedure should represent the true $^{238}\text{U}/^{235}\text{U}$
561 ratio of the EC 002 pyroxene megacrysts, free of other mineral or terrestrial contamination.

562 Based on this, a mass balance calculation estimates a $^{238}\text{U}/^{235}\text{U}$ ratio of ~ 137.83 for the removed
563 components (re-combined W1 to W4 leachates), a difference of up to $\sim 0.45\%$. This is larger than
564 observed in NWA 6291 (Brennecka and Wadhwa, 2012), but comparable to results reported in a
565 conference abstract by Huyskens et al. (2020) between merrillite and the bulk rock fraction in the
566 plutonic angrite Angra dos Reis.

567 These results show that the bulk rock U isotope composition may not be wholly representative for
568 age correcting pyroxene fractions using Pb-Pb chronometry. Age corrections may be necessary to
569 account for potential isotopic heterogeneities between pyroxene and bulk rock fractions of certain
570 achondrites. This discussion re-emphasizes the need for more detailed and systematic
571 investigations into the U isotope compositions of achondrites and their minerals, but also the effect
572 that acid leaching potentially has on them.

573

574

575

576 *4.3 ^{26}Al - ^{26}Mg systematics*

577 Regressing the ^{26}Al - ^{26}Mg data from seven fractions of EC 002 results in an isochron with excess
578 scatter (MSWD = 4.6) and a crystallization age of 1.83 ± 0.12 Ma after CAI formation (Fig. 2a).

579 The four plagioclase fractions of our dataset define an isochron with an age of 1.90 ± 0.09 Ma after

580 CAI formation, identical within the stated uncertainties but with no excess scatter (MSWD =
581 0.002; Fig. 2b). When anchored to the absolute age of CAIs (4567.30 ± 0.16 ; Connelly et al.,
582 2012), they result in absolute ages of 4565.47 ± 0.20 Ma and 4565.40 ± 0.18 Ma, respectively. The
583 excess scatter of the regression through all fractions is likely the result of the presence of accessory
584 phases in the fines fraction, and/or xenolithic material in the bulk rock and pyroxene fractions (e.g.,
585 Barrat et al., 2021, Nicklas et al, 2022). The plagioclase-only isochron is less likely to be affected
586 by xenolithic material, which could explain the low MSWD. The difference is not resolvable at
587 this level of precision, therefore for chronological purposes we conservatively use the isochron
588 that includes all fractions (1.83 ± 0.12 Ma after CAI formation) going forward.

589 The crystallization age of 1.83 ± 0.12 Ma after CAI formation is consistent with, albeit less precise
590 than the age of 1.83 ± 0.04 Ma determined by Fang et al. (2022). Both these ages are respectively
591 0.46 ± 0.13 and 0.46 ± 0.05 Myr older than the ^{26}Al - ^{26}Mg age of 2.29 ± 0.04 Ma after CAI
592 formation (both re-calculated with a ^{26}Al half-life of 0.717 Myr) reported by Barrat et al. (2021),
593 which was determined by SIMS (Fig. 6). Our age is strengthened by the analysis of different
594 plagioclase fractions with a good spread in $^{27}\text{Al}/^{24}\text{Mg}$ ratios, and with that subset of our data in
595 agreement with our overall isochron. The differences of these results to those of Barrat et al. (2021)
596 have also been discussed by both Fang et al. (2022) and Zhu et al. (2022). Potentially, when
597 analyzing sample spots with very low amounts of Mg by SIMS, such as those with $^{27}\text{Al}/^{24}\text{Mg}$ ratios
598 between 2000 and 5000, matrix effects have to be strongly considered when using in-situ analytical
599 methods and counting statistics errors could become a factor, where background corrections could
600 hypothetically overcorrect the amount of Mg isotopes, leading to spuriously low ^{26}Mg excesses.
601 Furthermore, the calibrating standard used, Miyake-Jima plagioclase, has a $^{27}\text{Al}/^{24}\text{Mg}$ ratio of

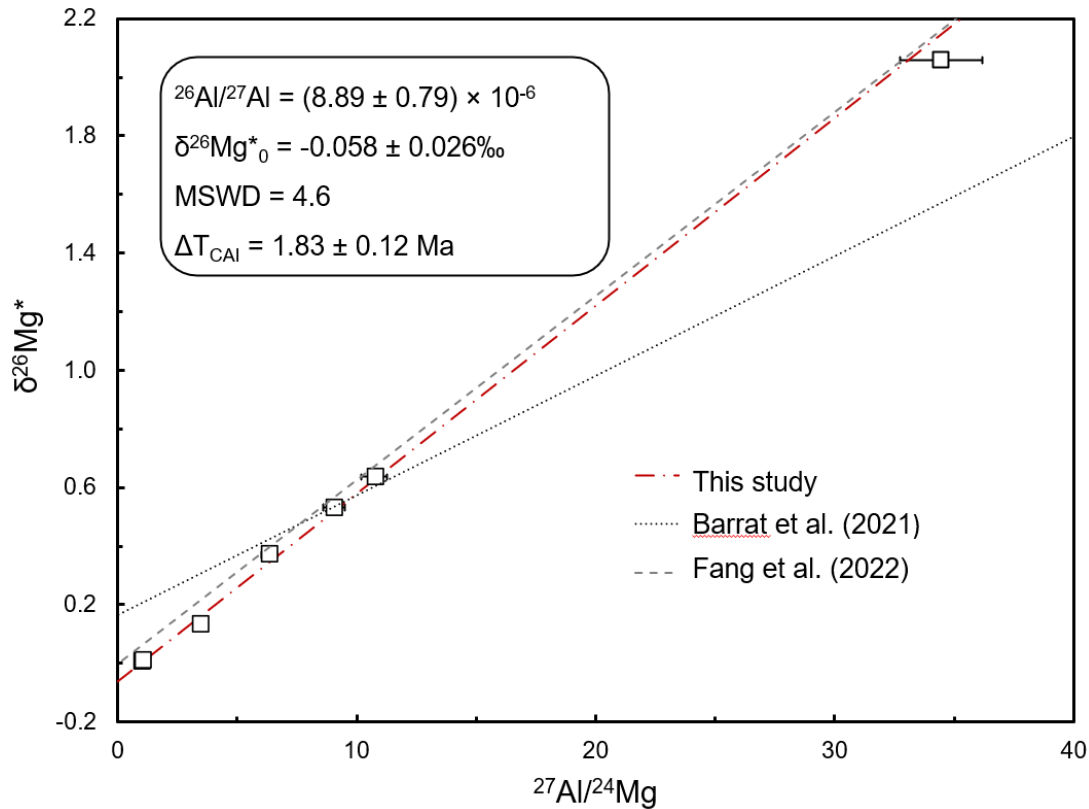
602 396.3, an order of magnitude lower than the plagioclase analyzed in their sample of EC 002, which
603 could lead to miscalibration for the measurements with higher $^{27}\text{Al}/^{24}\text{Mg}$ ratios (Zhu et al., 2022).
604 An alternative interpretation is that the isochron determined by Barrat et al. (2021) represents the
605 time of closure of pure plagioclase ($^{27}\text{Al}/^{24}\text{Mg} > 1500$), whereas our isochron with fractions that
606 contain at least some proportion of other minerals represents the time of closure of a mineral
607 mixture with a higher closure temperature for Mg diffusion. Closure temperatures of pyroxene and
608 anorthite have been estimated at 900°C to 1000°C and 700°C to 800°C, respectively (Wadhwa et
609 al., 2009). With the age difference of 0.46 ± 0.13 Myr between isochrons, based on those closure
610 temperatures as an upper limit, the cooling rate would be 465°C/Myr, several orders of magnitude
611 slower than what Barrat et al. (2021) estimated from the petrology of EC 002 (5°C/yr from 1200
612 °C to 1000°C, and >0.1 to 1°C/day below 900°C). Such a large difference cannot be explained by
613 the isochrons representing different times of closure of minerals and still satisfy the petrological
614 constraints established on the cooling rate of EC 002. Fang et al. (2022) argued similarly that the
615 plagioclase cores analyzed by the SIMS ion probe were not a closed system for Mg diffusion. The
616 younger SIMS ^{26}Al - ^{26}Mg age (Barrat et al., 2021) may be the result of the resumption of Mg
617 diffusion due to a re-heating event above the closure temperature in plagioclase.

618 In addition, the $\delta^{26}\text{Mg}^*_0$ values (initial $^{26}\text{Mg}/^{24}\text{Mg}$ ratio) between our study, Barrat et al. (2021)
619 and Fang et al. (2022) are distinct (Fig. 7). Barrat et al. (2021) first reported the $\delta^{26}\text{Mg}^*_0$ of 0.065
620 ± 0.085 ‰, followed by the -0.009 ± 0.005 ‰ of Fang et al. (2022). The $\delta^{26}\text{Mg}^*_0$ for our regression
621 through all fractions is -0.058 ± 0.026 ‰, distinctly lower than both previous studies. The $\delta^{26}\text{Mg}^*_0$
622 of the regression through the plagioclase fractions only is -0.010 ± 0.032 ‰, which is similar to
623 the initial value of Fang et al. (2022). Our two initial values are less precise than the one reported
624 by Fang et al. (2022) due to the absence of data points with $^{27}\text{Al}/^{24}\text{Mg}$ ratios < 0.9 , but more precise

625 from the higher value reported by Barrat et al. (2021). The difference between the all-fraction and
626 plagioclase-only isochron may be attributed to earlier-formed xenolithic material affecting the bulk
627 rock and pyroxene fractions. The $\delta^{26}\text{Mg}^*_0$ of -0.058 ± 0.026 ‰, while within uncertainty of the
628 “canonical” initial Solar System value (-0.034 ‰; Jacobsen et al., 2008; Schiller et al., 2010),
629 would require crystallization coeval with the formation of CAIs, which is inconsistent with all
630 reported ^{26}Al - ^{26}Mg ages. Some studies have reported “sub-canonical” initial $\delta^{26}\text{Mg}^*_0$ values up to
631 -0.128 ‰ in otherwise “canonical” CAIs (Wasserburg et al., 2012; MacPherson et al., 2017),
632 arguing for $\delta^{26}\text{Mg}^*_0$ heterogeneity in the early Solar System. While this could explain the low
633 $\delta^{26}\text{Mg}^*_0$ of -0.058 ± 0.026 ‰, there is no evidence for this in other Al-Mg studies of chondrites
634 and achondrites. Part of this discrepancy in $\delta^{26}\text{Mg}^*_0$ values reported so far is that both Barrat et al.
635 (2021) and Fang et al. (2022) calculated the mass-independent $\delta^{26}\text{Mg}^*$ values using the
636 equilibrium fractionation factor ($\beta = 0.521$). For EC 002, which is a rapidly cooled sample, we
637 argue that kinetic mass fractionation ($\beta = 0.511$) more appropriately models Mg isotope
638 fractionation, as equilibrium mass fractionation is negligible under these conditions (Teng et al.,
639 2007). This is further supported by the mass-dependent fractionation of our rock standard data of
640 the San Carlos olivine and CV3 chondrite Allende (Figs. S1 and S2). As Fang et al. (2022) did not
641 publish the stable Mg isotope compositions of their mineral fractions, their $\delta^{26}\text{Mg}^*$ values cannot
642 be re-calculated with the kinetic fractionation factor to estimate the effect of fractionation factor.
643 Re-calculating the data from Barrat et al. (2021) with the kinetic fractionation factor results in a
644 $\delta^{26}\text{Mg}^*_0$ value of 0.024 ± 0.055 ‰. This is closer to the value of Fang et al. (2022) and our
645 plagioclase-only isochron, indicating a likely better approximation of the true $\delta^{26}\text{Mg}^*_0$ value of
646 EC 002. For completeness, using the equilibrium fractionation factor instead shifts the $\delta^{26}\text{Mg}^*_0$

647 values of our isochrons even lower to -0.086 ± 0.035 ‰ and -0.037 ± 0.032 ‰ for the full and
648 plagioclase-only isochron, respectively.

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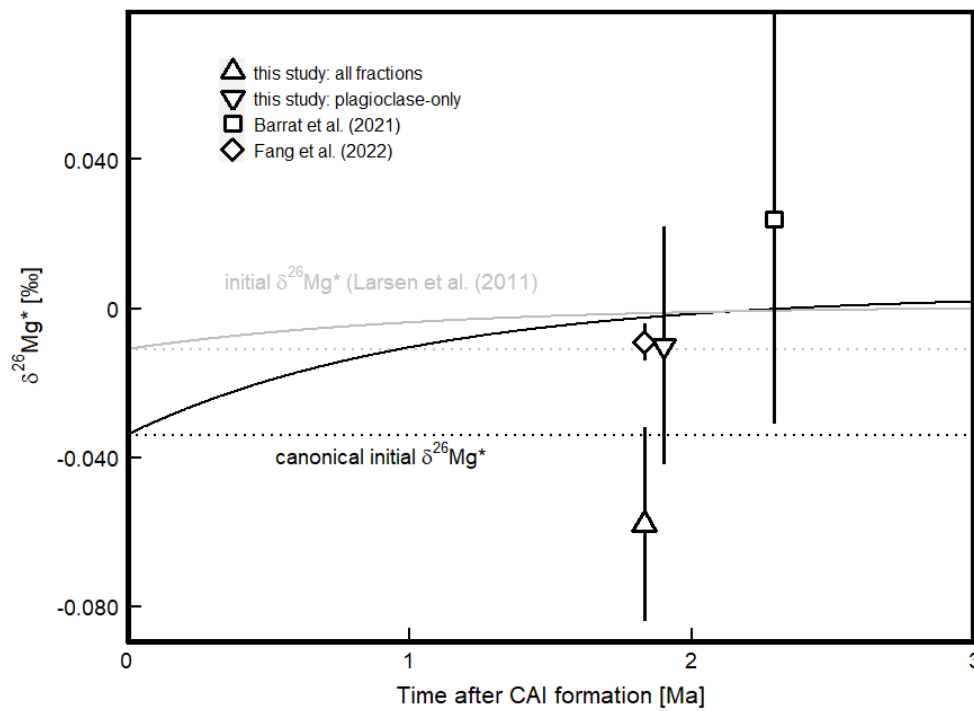


650

651 **Figure 6:** Comparison of initial $^{26}\text{Al}/^{27}\text{Al}$ ratios between this study, Fang et al. (2022) and Barrat
652 et al. (2021). Error envelopes of isochrons have been omitted for clarity.

653 The $\delta^{25}\text{Mg}$ mass-dependent values of the analyzed mineral fractions are isotopically light
654 compared to terrestrial rocks, bulk chondrites and other achondrites (e.g., Sedaghatpour and Teng,
655 2016). Due to the absence of reported mass-dependent Mg isotope data by Fang et al. (2022), we
656 cannot compare our data to this other study of EC 002. Standards that underwent analysis and the
657 same Mg separation procedure as the one employed for samples are consistent with their reported
658 compositions (Table 1). Any process to fractionate Mg isotopes would therefore have had to occur

659 between crystallization and dissolution of the samples. Terrestrial alteration may be a possibility.
 660 EC 002 shows signs of oxidation and secondary weathering products (Gattaccecca et al., 2021),
 661 but terrestrial weathering should cause an enrichment in heavier Mg isotopes, not a depletion (Teng
 662 et al., 2010). While the exact provenance of these light $\delta^{25}\text{Mg}$ values is unknown, the good
 663 agreement of the initial $^{26}\text{Al}/^{27}\text{Al}$ ratio with Fang et al. (2022) suggests that the responsible process,
 664 be it natural or analytical, did not affect the mass-independent values nor the ^{26}Al - ^{26}Mg
 665 chronometry.



666
 667 **Figure 7:** Presentation of the different $\delta^{26}\text{Mg}^*_0$ values of Barrat et al. (2021), Fang et al. (2022),
 668 as well as our isochron containing all fractions and the isochron containing only plagioclase
 669 fractions. Plotted alongside are the $\delta^{26}\text{Mg}^*$ evolution lines assuming canonical and sub-canonical
 670 $^{26}\text{Al}/^{27}\text{Al}$ ratios and $\delta^{26}\text{Mg}^*_0$ values (5.23×10^{-5} , -0.034 ‰: Jacobsen et al., 2008; 1.6×10^{-5} ,
 671 -0.016 ‰: Larsen et al., 2011).

672

673 *4.4 Combined short-lived chronology and U-corrected Pb-Pb ages*

674 Time anchors are used to calculate absolute model ages of samples that were dated by short-lived
675 chronometers. To anchor relative ^{26}Al - ^{26}Mg to absolute ages, analyzed samples are most-
676 commonly anchored to the absolute age of CAIs (4567.30 ± 0.16 Ma; Connelly et al., 2012).
677 Some achondrites have also been proposed and used as anchors, such as the angrite D'Orbigny,
678 particularly for the ^{53}Mn - ^{53}Cr chronometer due to the lack of Mn in CAIs (e.g., Glavin et al.,
679 2004; Wadhwa et al., 2009; Sanborn et al., 2019).

680 The distribution and abundance of the parent nuclide at time of formation between individual
681 meteorites as well as meteorite formation reservoirs can be then made. For example, previous
682 studies of angrites has suggested that the initial $^{26}\text{Al}/^{27}\text{Al}$ of the inner Solar System (NC
683 reservoir) was $\sim 1 \times 10^{-5}$, significantly lower than the accepted “canonical” value ($\sim 5 \times 10^{-5}$)
684 determined from CAIs (Schiller et al., 2015). This was based on a discrepancy of ~ 1 to 2 Myr
685 between Pb-Pb ages and ^{26}Al - ^{26}Mg ages of the angrites when anchored to CAIs. They posit that
686 the inner Solar System (NC reservoir) was depleted in ^{26}Al through thermal processing due to it
687 being closer to the proto-Sun (Schiller et al., 2018), before CI-like, ^{26}Al -enriched dust was
688 admixed to that formation reservoir (Bollard et al., 2019). To further investigate this hypothesis,
689 we anchor ^{26}Al - ^{26}Mg ages of different NC and CC achondrites to the absolute age of EC 002 as
690 determined by our U-corrected Pb-Pb chronology. The early formation, rapid cooling history and
691 undisturbed history of EC 002 (Barrat et al., 2021) make it a suitable candidate as an anchor
692 meteorite.

693 The recent reporting of the Cr isotope systematics of EC 002 (negative bulk $\epsilon^{54}\text{Cr}$ values; Anand
694 et al., 2022; Zhu et al., 2022) combined with O isotope data (Anand et al., 2022; Gattacceca et
695 al., 2021) and negative Tm* anomaly (Barrat et al., 2021) indicate that EC 002 formed in the NC

696 reservoir, near the angrites, brachinites and HEDs. Therefore, its parent body should have formed
697 with the same ^{26}Al abundance as the angrite parent body. This can be easily examined by anchoring
698 the ^{26}Al - ^{26}Mg ages of three angrites (D'Orbigny, Sahara 99555 and NWA 1670) to EC 002. These
699 three angrites also have published Pb-Pb ages, though the $^{238}\text{U}/^{235}\text{U}$ ratio of NWA 1670 has not
700 been analyzed (Schiller et al., 2015), so that any potential discrepancies between the ^{26}Al - ^{26}Mg
701 and U-corrected Pb-Pb chronometers can be examined. The resulting ^{26}Al - ^{26}Mg ages are visualized
702 in Fig. 8 and listed in Table S3. To summarize, the U-corrected Pb-Pb age of D'Orbigny, Sahara
703 99555 and NWA 1670 are systematically older by ~ 0.5 to 1.5 Myr than their ^{26}Al - ^{26}Mg ages
704 anchored to EC 002. This suggests that ^{26}Al heterogeneity is not the underlying cause of these age
705 discrepancies, but rather that there is some chronological disturbance in the angrites. This has been
706 previously discussed based on apparent isotopic disturbances of Sm-Nd systematics in D'Orbigny
707 (Sanborn et al., 2015). Furthermore, Sahara 99555 exhibits evidence of terrestrial alteration (e.g.,
708 Floss et al., 2003) and potential re-distribution of Pb on the mineral grain scale (Amelin, 2008a),
709 and NWA 1670 may have been shocked (Keil, 2012 and references therein), all of which could
710 have affected the accuracy of the radiometric systems. Evidence for post-crystallization Mg
711 isotope re-equilibration in plagioclase has also been observed in the NC ungrouped achondrite
712 Asuka 881394 (Nyquist et al., 2003; Wadhwa et al., 2009; Hublet et al., 2017; Wimpenny et al.,
713 2019), which shares a similar age discrepancy when anchored to EC 002 (Fig. 8, Table S3).
714 In contrast to the angrites and Asuka 881394, the ^{26}Al - ^{26}Mg ages (anchored to EC 002) of the
715 two ungrouped carbonaceous achondrites NWA 2976 and NWA 6704 are concordant to their U-
716 corrected Pb-Pb ages (Fig. 8, Table S3: Schiller et al., 2010a; Bouvier et al., 2011b). This implies
717 that, within the resolution of the chronometers, there were no significant variations of the ^{26}Al
718 abundance between the NC (EC 002) and CC reservoir (NWA 2976 and NWA 6704).

719 In addition, it is important to note that using the U-corrected Pb-Pb age with the bulk rock
720 $^{238}\text{U}/^{235}\text{U}$ ratio of 137.819 ± 0.010 and the corresponding older age of 4566.43 ± 0.14 Ma does
721 not significantly alter the interpretation of these findings. Of the three angrites, the ^{26}Al - ^{26}Mg age
722 of D'Orbigny (when anchored to EC 002) becomes concordant with its U-corrected Pb-Pb age.
723 Those of NWA 1670 and also Sahara 99555, when using the more precise ^{26}Al - ^{26}Mg data of
724 Schiller et al. (2015), still are over 0.5 Myr younger than their U-corrected Pb-Pb ages. Crucially,
725 both the ^{26}Al - ^{26}Mg ages of the two ungrouped carbonaceous achondrites NWA 2976 and NWA
726 6704 remain concordant with their U-corrected Pb-Pb ages (Table S3). All these pieces of
727 evidence do not suggest that ^{26}Al was distinctly different between the NC and CC reservoirs, and
728 that ^{26}Al heterogeneities are not the cause of discordant ages between the ^{26}Al - ^{26}Mg and U-
729 corrected Pb-Pb chronometers.

730 One remaining issue between the ^{26}Al - ^{26}Mg and U-corrected Pb-Pb ages of EC 002 is their
731 own discordance when anchored to the absolute age of CAIs. The age difference between the U-
732 corrected Pb-Pb age and the ^{26}Al - ^{26}Mg age of EC 002 is 0.40 ± 0.36 Myr (our data) or 0.40 ± 0.34
733 Myr (Fang et al., 2022). This results in $^{26}\text{Al}/^{27}\text{Al}$ ratios at the time of CAI formation of $[3.54^{+1.82}/$
734 $_{1.22}] \times 10^{-5}$ and $[3.54^{+1.43}/_{-1.03}] \times 10^{-5}$, respectively. These values appear lower than the “canonical”
735 $^{26}\text{Al}/^{27}\text{Al}$ ratio, though both are within uncertainty of at least some internal isochrons of CAIs that
736 also have old Pb-Pb ages (e.g., Jacobsen et al., 2008; Bouvier and Wadhwa, 2010). Both values,
737 however, are clearly higher than the suggested $^{26}\text{Al}/^{27}\text{Al}$ ratio of the angrite parent body (APB) at
738 time of CAI formation ($[1.33^{+0.21}/_{-0.18}] \times 10^{-5}$; Schiller et al., 2015). These age differences could
739 be resolved by invoking earlier formation of CAIs at ~ 4567.7 Ma. The timing and duration of CAI
740 formation has been discussed largely by previous workers (Bouvier and Wadhwa, 2010; Bouvier
741 et al., 2011a; Sanborn et al., 2019; Wimpenny et al., 2019) and is beyond the scope of this study.

742 The most recent U-corrected absolute Pb-Pb age of CAIs (4567.30 ± 0.16 Ma; Connelly et al.,
743 2012) is based on a weighted mean of four CAIs from the CV3 chondrites Efremovka and Allende
744 (Amelin 2010, Connelly et al., 2012), both of which underwent aqueous alteration that may have
745 re-distributed isotopes and disturbed the Pb-Pb systematics (e.g., Bouvier and Wadhwa, 2010; Krot
746 et al., 2021). The initial $^{26}\text{Al}/^{27}\text{Al}$ ratios of those four CAIs have also not been determined *in situ*
747 by SIMS, thereby not confirming if they actually formed with a “canonical” $^{26}\text{Al}/^{27}\text{Al}$ ratio.

748 Furthermore, there have been reports of older U/Th or U-corrected Pb-Pb ages of two CAIs,
749 2364-B1 from the CV3 chondrite NWA 2364 (Bouvier and Wadhwa, 2010) and 6991-B4 from the
750 CV3 chondrite NWA 6991 (although not peer-reviewed, Bouvier et al., 2011a), with absolute Pb-
751 Pb ages of $4568.2 (^{+0.2}/_{-0.4})$ Ma and 4567.94 ± 0.31 Ma, respectively. Both D’Orbigny and NWA
752 6704 are concordant when anchored to the 2364-B1 CAI age, in contrast to the Efremovka +
753 Allende CAIs (Sanborn et al., 2019). The U isotope composition of CAI 2364-B1 was not
754 measured but estimated using the low measured Th/U ratio of the sample. Bouvier and Wadhwa
755 (2010) used the correlation found between U/Th with $^{238}\text{U}/^{235}\text{U}$ (Brennecka et al., 2010) to propose
756 an adjustment of up to -0.3 Myr to the internal mineral separate Pb-Pb age of the 2364-B1 CAI.
757 The 2364-B1 U/Th-corrected Pb-Pb age is therefore consistent with the U-corrected 6991-B4 Pb-
758 Pb age. The correlation between U/Th ratios and U isotopic compositions in CAIs was confirmed
759 by Tissot et al. (2016) as produced by the decay of short-lived ^{247}Cm . The CAIs measured by two
760 other studies did not fall on this correlation (Amelin et al., 2010; Connelly et al., 2012). Therefore,
761 those samples did not support this proxy for correcting Pb-Pb ages of CAIs. The process
762 responsible for such U isotopic variations in CAIs (beside from ^{247}Cm decay) is unknown.

763 The U and Al-Mg isotope compositions of NWA 6991 compact type A CAI (called B4)
764 were measured (Bouvier et al., 2011a), but these data have not been published in a peer-reviewed

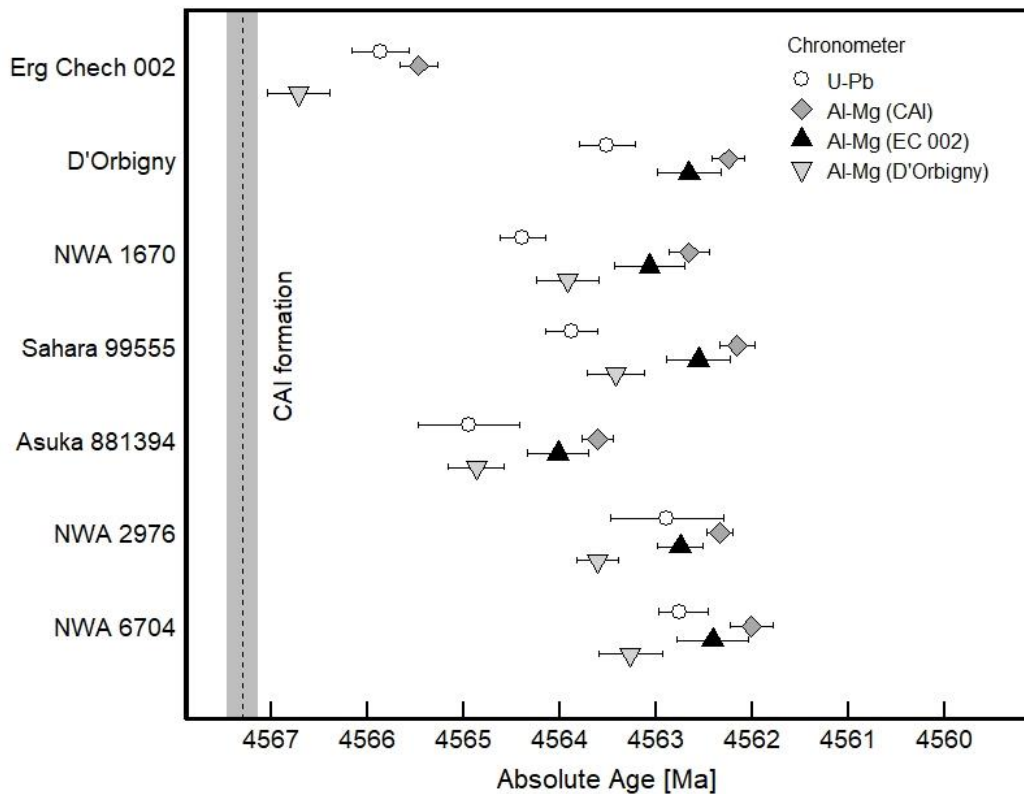
765 journal and remain available only in an abstract form. A protracted formation interval of CAIs of
766 up to 0.4 Myr, as suggested by Kawasaki et al. (2019, 2020), based on *in situ* ^{26}Al - ^{26}Mg dating,
767 could potentially resolve the differences in absolute ages. This would, however, still require
768 investigation into the absolute timing of such CAIs, and not necessarily impact the age of the oldest
769 CAIs that define the initial $^{26}\text{Al}/^{27}\text{Al}$ ratio in the Solar System. As it currently stands, the age of
770 4567.30 ± 0.16 Ma (Connelly et al., 2012) remains the most robust U-corrected Pb-Pb age for a
771 CAI. Nevertheless, renewed attention to U-corrected Pb-Pb ages combined with Al-Mg
772 systematics of CAIs from more pristine chondrites may be warranted.

773 Two recent studies aimed to determine the initial $^{53}\text{Mn}/^{55}\text{Mn}$ ratio of EC 002 to calculate its
774 formation age using the ^{53}Mn - ^{53}Cr chronometer (Zhu et al., 2022; Anand et al., 2022). Notably,
775 the results of the two studies do not agree with each other (Fig. 9, Table S3). Zhu et al. (2022)
776 determined an initial $^{53}\text{Mn}/^{55}\text{Mn}$ ratio of $[5.85 \pm 0.45] \times 10^{-6}$ from an isochron based on various
777 mineral fractions, resulting in an absolute age of 4566.66 ± 0.56 Ma when anchored to D'Orbigny.
778 Conversely, the initial $^{53}\text{Mn}/^{55}\text{Mn}$ ratio of $[4.76 \pm 0.39] \times 10^{-6}$, determined by Anand et al. (2022)
779 from leachates as well as a whole-rock, silicate and chromite fractions, results in an absolute age
780 of 4565.56 ± 0.59 Ma, therefore younger by 1.10 ± 0.95 Myr than the study by Zhu et al. (2022).
781 Anand et al. (2022) argue that the older age of Zhu et al. (2022) stems from higher fractions of
782 xenolithic material within their sample set, as evidenced by larger $\epsilon^{54}\text{Cr}$ variations in them. While
783 a sample like EC 002 is susceptible to heterogeneities between samples due to the presence of
784 large megacrysts, it is questionable that this would result in an age difference as large as ~ 1 Myr.
785 We note that Zhu et al. (2022) excluded xenolithic fractions (based on their $\epsilon^{54}\text{Cr}$ values) from
786 their isochron. Certainly, these incongruent results warrant further measurements.

787 Regardless of the cause of these different ages, the chronological repercussions of the 1 Myr gap
788 are substantial: the younger age obtained by Anand et al. (2022) is consistent with the U-corrected
789 Pb-Pb age of EC 002, and with the ^{26}Al - ^{26}Mg ages anchored to CAIs (this study; Fang et al., 2022).
790 Furthermore, the ^{53}Mn - ^{53}Cr ages of the three angrites D'Orbigny, NWA 1670 and Sahara 99555
791 (Glavin et al., 2004; Sugiura et al., 2005), NC achondrite Asuka 881394 (Wimpenny et al., 2019)
792 and CC achondrite NWA 6704 (Sanborn et al., 2019), when anchored to the U-corrected Pb-Pb
793 age of EC 002 are all concordant within the resolution of the chronometers. This result contrasts
794 with the older ^{53}Mn - ^{53}Cr age by Zhu et al. (2022), to which none of the above achondrites are
795 concordant (Fig. 10, Table S3). Given this context, it would seem that the initial $^{53}\text{Mn}/^{55}\text{Mn}$ ratio
796 of $[4.76 \pm 0.39] \times 10^{-6}$ (Anand et al., 2022) may be more representative of EC 002. The
797 concordance between ^{53}Mn - ^{53}Cr ages and U-corrected Pb-Pb ages when anchored to EC 002 with
798 that particular value furthermore suggests that the ^{53}Mn - ^{53}Cr chronometer was not disturbed in
799 Asuka 881394 and the angrites, in comparison with the ^{26}Al - ^{26}Mg chronometer. This would imply
800 that the cause of chronological disturbance in those samples is due to Mg diffusion in plagioclase
801 through later thermal events, rather than a more general re-setting of the chronometers. For ^{26}Al -
802 ^{26}Mg chronometry, D'Orbigny should be avoided as an anchor based on these findings.

803 In summary, the age gap between U-corrected Pb-Pb and ^{26}Al - ^{26}Mg ages (when anchored to CAIs)
804 in some angrites and Asuka 881394 is most likely to be attributed to later re-setting of the ^{26}Al -
805 ^{26}Mg chronometer. Alternatively, U-corrected Pb-Pb ages of CAIs from the Efremovka and
806 Allende meteorites that may have been affected by secondary alteration, resulting in spuriously
807 young Pb-Pb ages. Heterogeneous $^{238}\text{U}/^{235}\text{U}$ ratios in bulk meteorites, as evidenced in pyroxene
808 separates of EC 002, may contribute to age shifts of up to ~ 0.5 Myr, but are unlikely the main
809 cause of age discordances of up to 2 Myr. Further advances in the analytical precision of both U

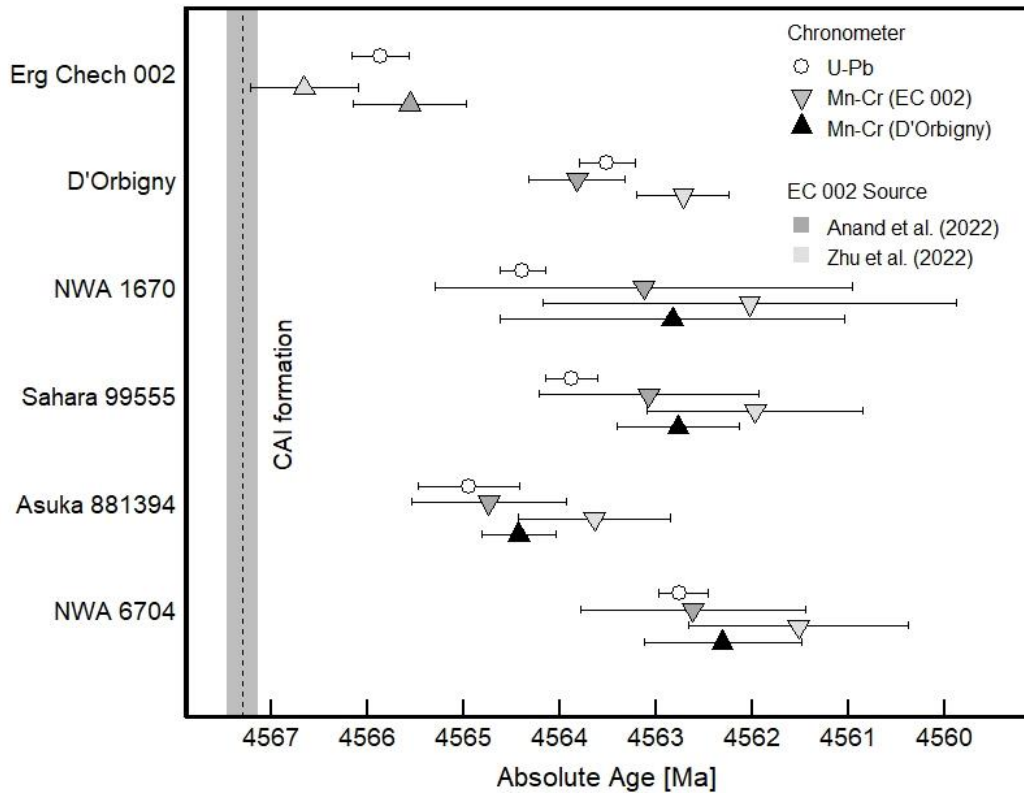
810 and Pb isotope measurements could potentially reveal more and/or larger differences in ages
 811 between the ^{26}Al - ^{26}Mg and Pb-Pb chronometers, indicating smaller scale chronological
 812 disturbances or secondary processes in meteorite parent bodies.
 813



814
 815 **Figure 8:** Age comparison between U-corrected Pb-Pb ages and ^{26}Al - ^{26}Mg ages anchored to EC
 816 002 (4565.87 ± 0.30 Ma: this study), D'Orbigny (4563.51 ± 0.29 Ma: Amelin, 2008a; Tissot et
 817 al., 2017) and CAIs (4567.30 ± 0.30 : Connelly et al., 2012) of various achondrites. Data and
 818 their sources are presented in Table S3. For clarity, the plotted U-corrected Pb-Pb age of Sahara
 819 99555 is the weighted mean of Amelin (2008a) and Connelly et al. (2008), corrected by Tissot et
 820 al. (2017). For Asuka 881394 and NWA 2976, the ^{26}Al - ^{26}Mg ages plotted are those reported by
 821 the same studies that reported their U-corrected Pb-Pb ages (Bouvier et al., 2011b; Wimpenny et
 822 al., 2019).

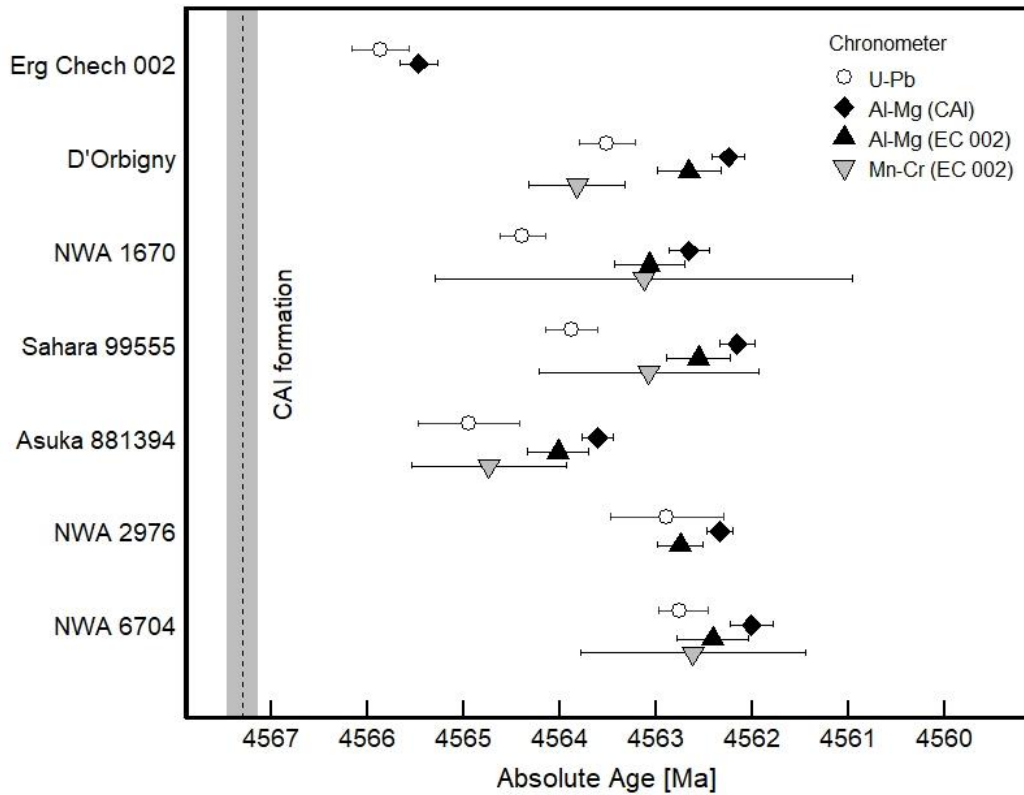
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825

826 **Figure 9:** Age comparison between U-corrected Pb-Pb ages and ^{53}Mn - ^{53}Cr ages anchored to EC
827 002 (4565.87 ± 0.30 Ma: this study) and D'Orbigny (4563.51 ± 0.29 Ma: Amelin, 2008a; Tissot
828 et al., 2017) of different achondrites, highlighting the ^{53}Mn - ^{53}Cr age differences between of
829 Anand et al. (2022) (dark grey colouring) and Zhu et al. (2022) (light grey colouring). Data and
830 their sources are presented in Table S3. For clarity, the U-corrected Pb-Pb age of Sahara 99555
831 plotted is the weighted mean of Amelin (2008a) and Connelly et al. (2008), corrected by Tissot
832 et al. (2017).



833

834 **Figure 10:** Age comparison between U-corrected Pb-Pb ages, and ^{26}Al - ^{26}Mg ages anchored to
 835 CAIs (4567.30 ± 0.30 : Connelly et al., 2012), and both ^{26}Al - ^{26}Mg and ^{53}Mn - ^{53}Cr ages anchored
 836 to EC 002 (4565.87 ± 0.30 Ma: this study) of various achondrites. Data and their sources are
 837 presented in Table S3. The initial $^{53}\text{Mn}/^{55}\text{Mn}$ ratio reported by Anand et al. (2022) is used to
 838 calculate ^{53}Mn - ^{53}Cr ages in this diagram. For clarity, the U-corrected Pb-Pb age of Sahara 99555
 839 plotted is the weighted mean of Amelin (2008a) and Connelly et al. (2008), corrected by Tissot
 840 et al. (2017), and for Asuka 881394 and NWA 2976, the ^{26}Al - ^{26}Mg ages plotted are those
 841 reported by the same studies that reported their U-corrected Pb-Pb ages (Bouvier et al., 2011b;
 842 Wimpenny et al., 2019).

843

844 5. Conclusions

845 The ungrouped achondrite Erg Chech 002 was dated by U-corrected Pb-Pb and ^{26}Al - ^{26}Mg
846 chronology in mineral and bulk rock fractions, as well as *in situ* U-Pb chronology of Ca-phosphates
847 by SIMS. Compared to previous studies that dated EC 002 by *in situ* ^{26}Al - ^{26}Mg analysis (Barrat et
848 al., 2021), the ^{26}Al - ^{26}Mg age was determined by MC-ICP-MS analysis of seven mineral fraction
849 and results in an age that is 0.43 ± 0.11 Myr older, at 1.83 ± 0.12 Ma after CAI formation and an
850 absolute age of 4565.47 ± 0.20 Ma when anchored to the absolute age of CAIs. This age is
851 consistent with the reported ^{26}Al - ^{26}Mg age by Fang et al. (2022). The U-corrected Pb-Pb age of
852 EC 002 is 4565.87 ± 0.30 Ma, determined by pyroxene residue and leachates, and makes EC 002
853 the oldest dated achondrite so far. The U isotope composition was determined for both a bulk rock
854 and a leached pyroxene fraction, resulting in a notable difference between them that is likely
855 caused by internal heterogeneities between mineral phases. For the calculation of the Pb-Pb age,
856 the $^{238}\text{U}/^{235}\text{U}$ ratio of the (re-combined) leached pyroxenes was used.

857 The U-Pb age determined by *in situ* analysis of merrillites by SIMS is 4564.3 ± 5.2 Ma, equal
858 within uncertainty to the U-corrected Pb-Pb age determined on leached pyroxenes. This indicates
859 that the parent body of EC 002 was not exposed to thermal metamorphism or any substantial
860 impact events after ~ 4559 Ma and underwent rapid cooling.

861 The good agreement between ^{26}Al - ^{26}Mg and U-corrected Pb-Pb ages of CC achondrites when
862 anchored to EC 002, indicates that ^{26}Al was distributed homogeneously between the NC and CC
863 reservoirs, in contrast to previous interpretations suggesting the opposite (e.g., Schiller et al.,
864 2015). The age differences between ^{26}Al - ^{26}Mg and U-corrected Pb-Pb ages of angrites and NC
865 achondrite Asuka 881394 when anchored to EC 002 and CAIs are most likely to have been caused
866 by diffusion of Mg isotopes in plagioclase during re-heating episodes or later thermal events. The

867 concordance between ^{53}Mn - ^{53}Cr ages and U-corrected Pb-Pb ages of those same achondrites
868 suggest that the ^{53}Mn - ^{53}Cr chronometer was not disturbed. Given its old age, precise ^{26}Al - ^{26}Mg
869 and U-corrected Pb-Pb ages and straightforward thermal history, EC 002 shows promise to be used
870 as an anchor to date other objects with short-lived chronometers.

871

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880

881 **Appendix A. Supplementary Material**

882 Table S1: EPMA elemental compositions of merrillites in EC 002.

883 Table S2: ^{26}Mg isotopic anomalies ($\delta^{26}\text{Mg}^*$) of EC 002 mineral fractions calculated with different
884 fractionation (kinetic vs. equilibrium) factors.

885 Table S3: compilation of published U-corrected Pb-Pb ages, ^{26}Al - ^{26}Mg ages and ^{53}Mn - ^{53}Cr ages
886 of EC 002 and other achondrites.

887 Figures S1 and S2: linearized three-isotope plots of Mg of individual analyses of rock standards
888 San Carlos olivine and CV3 chondrite Allende, respectively.

889

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