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Michael W. Broadley, David V. Bekaert, Bernard Marty, Akira Yamaguchi,
Jean-Alix J-A Barrat

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1 **Noble gas variations in ureilites and their implications for ureilite parent body**
2 **formation**

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5 Michael W. Broadley^{1*}; David V. Bekaert¹; Bernard Marty¹; Akira Yamaguchi²; Jean-Alix
6 Barrat³

7

8 ¹ *Centre de Recherches Pétrographiques et Géochimiques, UMR 7358 CNRS—Université*
9 *de Lorraine, 15 rue Notre Dame des Pauvres, BP 20, F-54501 Vandoeuvre-lès-Nancy,*
10 *France*

11 ² *National Institute of Polar Research, Tachikawa, Tokyo 190-8518, Japan*

12 ³ *Laboratoire Geoscience Océan, UMR 6538 CNRS—Université de Bretagne Occidentale et*
13 *Institut Universitaire Européen de la Mer, Place Nicolas Copernic, 29280 Plouzané, France*

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17

18 *Corresponding author - Email: broadley@crpg.cnrs-nancy.fr

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21

22 **Abstract**

23 Ureilites are equilibrated carbon-rich olivine-pyroxene rocks from the partially melted
24 mantle of a large (>500 km diameter) heterogeneous parent body. Recently the
25 ureilite parent body was interpreted as an incomplete mixture of material from two
26 carbon-rich chondritic reservoirs, one (Mg-rich) with reduced iron, low $\Delta^{17}\text{O}$ and low
27 $\delta^{13}\text{C}$, and the other with oxidised iron, high $\Delta^{17}\text{O}$ and high $\delta^{13}\text{C}$. Here we analyse
28 noble gases (Ar, Kr and Xe) in six equilibrated (unbrecciated) ureilites from
29 Northwest Africa (NWA 2236, NWA 7686, NWA 8049, NWA 8172, NWA 11032 and
30 NWA 11368). We observe weak positive and negative correlations of $\Delta^{17}\text{O}$ and Mg#
31 with the elemental ratios of Ar/Xe and Kr/Xe, respectively, as well as a weak positive
32 correlation of Mg# with the heavy isotopes of Xe. These correlations broadly support
33 the idea of the two-component mixing hypothesis. Our analyses further suggest that
34 the Mg-rich endmember was rich in Xe from presolar grains (HL-Xe) while the Mg-
35 poorer component may have contained solar-derived noble gases. The observed
36 correlations are less straightforward to reconcile with a recent model for the origin of
37 the ureilite parent body, involving oxidation of metal by H_2O from accreted ice with
38 'heavy' oxygen isotopes.

39

40 **1. Introduction**

41 Ureilites are a group of achondrites originating from an inner solar system body
42 (Downes et al., 2008). They are essentially the remnants of a mantle that has
43 undergone some degree of melt extraction (Goodrich et al., 2007). Ureilites are
44 predominantly composed of olivine and pyroxene, representing mantle restites after
45 the loss of trachyandesitic melts (Barrat et al., 2016; Bischoff et al, 2014). The ureilite
46 parent body (UPB) is considered to have accreted ~1.6 Myr after CAI formation and

47 therefore has an accretionary age intermediate between fully differentiated and non-
48 differentiated meteorite parent bodies (Budde et al., 2016). The early accretion of the
49 UPB during the existence of short-lived ^{26}Al resulted in significant heating and
50 segregation of a S-rich core (Barrat et al., 2015; Rankenburg et al., 2008; Warren et
51 al., 2006). The process of differentiation and core formation was brought to a halt,
52 possibly by a large impact event that resulted in the catastrophic breakup of the UPB
53 (Downes et al., 2008; Goodrich et al., 2015; Warren and Kallemeyn, 1989).

54

55 Ureilites are notable for their unusually high abundance of carbon, averaging ~3 wt%
56 (Warren and Huber, 2006). Olivines within ureilites exhibit rims that are depleted in
57 oxidised Fe as a result of being reduced. These rims are considered to have formed
58 during smelting reactions where carbon reacted with oxygen (Singletary and Grove,
59 2003). The rims of olivines (<100 μm ; Singletary and Grove, 2003) constitute
60 anywhere between 0 to 60% of the total volume of the olivines (Singletary and Grove,
61 2003). The localised reduction along the rims and the sharp boundaries between
62 rims and cores in olivines are considered to reflect a short lived process that
63 occurred during rapid cooling that followed the impact driven breakup of the UPB
64 (Miyamoto et al., 1985; 1993; Warren and Huber, 2006).

65

66 Olivines in ureilites are on average 1 mm in size, with curved intergranular
67 boundaries (Goodrich, 1992). Excluding the distinctive reduction rims, olivine core
68 compositions are uniform within a given sample, with limited intergrain or intragrain
69 variability in major and minor element concentrations (Goodrich et al., 1987). Whilst
70 the cores of olivines are uniform within a given sample, ureilites exhibit a large
71 compositional range from one meteorite to another, with the Mg# ($\text{Mg\#} = 100 \times \text{molar}$

72 Mg/(Mg+Fe)) of olivine cores ranging from 74 to 97 (Downes et al., 2008). Variations
73 in Mg# have been shown to positively correlate with FeO/MnO and $\Delta^{17}\text{O}$ (Clayton
74 and Mayeda, 1988; Goodrich and Delaney, 2000). The correlation between Mg# and
75 FeO/MnO is suggested to come about through redox processes, where FeO is
76 removed or added to an unknown meteoritic composition with a constant Mg/Mn via
77 reducing or oxidising reactions, respectively (Goodrich and Delaney, 2000;
78 Mittlefehldt, 1986). The redox-controlled variation in Mg# could have been
79 established within the UPB itself (Goodrich et al., 2007; 2013; Sanders et al., 2017),
80 or it could have been inherited either directly from nebular solids (Clayton and
81 Mayeda, 1988; Goodrich and Delaney, 2000; Warren et al., 2012), or from the
82 merging of compositionally distinct planetesimals (Barrat et al., 2017). The origin of
83 these correlated variations in Mg#, FeO/MnO and $\Delta^{17}\text{O}$ in ureilite olivines is therefore
84 still up for debate.

85

86 Noble gases are known to have resolvable elemental and isotopic signatures in
87 different cosmochemical reservoirs including the solar nebula (Meshik et al., 2014),
88 chondrites (Busemann et al., 2000) and comets (Marty et al., 2017), and can
89 therefore be used to help identify contributions of different solar system components
90 to growing planetesimals and planets (Busemann et al., 2006). Ureilites are
91 especially rich in trapped noble gases, with concentrations approaching, and
92 sometimes surpassing, those of carbonaceous chondrites (Rai et al., 2003a). Noble
93 gases in ureilites are predominantly hosted within diamonds and amorphous carbon
94 phases (Göbel et al., 1978; Rai et al., 2003a; Wacker, 1986), although graphite
95 (Okazaki et al., 2003) and an unknown acid soluble phase (Murty et al., 2010) have
96 also been shown to contain significant quantities of noble gases. Whether the noble

97 gases within ureilites retain evidence from multiple cosmochemical components that
98 have contributed to the formation of the UPB, remains to be seen (Barrat et al.,
99 2017). We therefore undertook high precision analyses of the concentrations and
100 isotopic compositions of Ar, Kr and Xe in a suite of ureilites covering the full range of
101 Mg# in an attempt to determine whether the UPB accreted from multiple
102 cosmochemical reservoirs, and whether noble gases can help in identifying the origin
103 of these distinct components.

104

105 **2. Samples and Analytical Methods**

106 Six North West African ureilites were chosen for noble gas analysis. The details of
107 the samples used in this study have been presented previously, by Barrat et al.,
108 (2015; 2017), and in the Meteoritical Database (Table 1). All samples analysed are
109 unbrecciated main group ureilites, which are known to have extremely uniform and
110 well-equilibrated olivine core compositions with limited to no intergrain or intragrain
111 variability (see Barrat et al., 2015 for an example from sample NWA 8049). Olivine
112 cores within the samples analysed as part of this study have Mg# ranging from 76.9
113 to 96.9 (Barrat et al., 2015, 2017). These samples were specifically chosen to span
114 the range of compositions previously measured within ureilites (Mg# = 74 - 97;
115 Downes et al., 2008).

116

117 Samples weighing ~100 mg each were first crushed in to sub millimetre sized chips.
118 Between 6 and 32 mg of the crushed material was wrapped in Ta foil packets, which
119 were loaded two at a time into a filament furnace consisting of 3 alumina-coated
120 tungsten evaporation baskets (Bekaert et al., 2018). One basket was loaded with an
121 empty Ta foil packet to determine the blank contribution. Prior to analysis, the

122 furnace was baked at 150°C for 48 hours to remove any adsorbed atmospheric noble
123 gas from the samples and furnace. Noble gases were then released from the
124 samples by step heating at 600°C, 1100°C and 1700°C. The samples and the blank
125 basket were step heated sequentially, i.e. all were analysed at 600°C before moving
126 on to the next temperature step in order to avoid passive heating of other samples
127 during higher temperature extractions. Blanks analysed prior to samples had
128 consistently atmospheric noble gas isotopic ratios, indicating that no noble gas was
129 released from adjacent samples by passive heating during analysis.

130

131 Samples were heated for a total of 10 minutes at the required temperature. Extracted
132 gases were purified by being passed through an inline Ti-sponge getter at 650°C.
133 Krypton and xenon were then trapped onto a quartz cold finger held at liquid nitrogen
134 temperature for 20 minutes. The Ar fraction, which remains in the gas phase under
135 these conditions, was then exposed to two pairs of hot (550°C) and cold (60°C) Ti-
136 sponge getters, respectively, before being admitted to the mass spectrometer (Helix
137 MC plus, Thermo Fisher Scientific) for analysis. Argon was analysed using multi-
138 collection mode, with ^{40}Ar measured on the faraday collector and $^{36,38}\text{Ar}$ measured on
139 compact dynode electron multipliers. After Ar was analysed, Kr and Xe were
140 released from the quartz cold finger at room temperature and purified using the same
141 procedure as described for Ar. Typically, 5% of the Kr and <1% of the Xe was not
142 trapped in the quartz cold finger and therefore was lost in the Ar fraction. This effect
143 has been corrected for in the presented dataset. Krypton and xenon were then
144 simultaneously admitted to the mass spectrometer, with Xe isotopes being analysed
145 first, followed by Kr. Krypton and xenon isotopes were analysed using peak jumping
146 mode over 20 repeat analytical cycles. Krypton and xenon isotopes were measured

147 on an axial compact dynode electron multiplier. Sufficient gas release at the 1100°C
148 and 1700°C temperature steps allowed multiple aliquots of the gas to be analysed
149 (n=3), thereby improving the accuracy of the determination of isotopic compositions.

150

151 Sensitivity and mass discrimination of the mass spectrometer were monitored by the
152 daily analysis of standards with an atmospheric isotopic composition (Bekaert et al.,
153 2018). Blank levels for the 1700°C temperature step were 5.1×10^{-15} , 7.7×10^{-17} and
154 9.7×10^{-18} mol for ^{36}Ar , ^{84}Kr and ^{132}Xe , respectively. Typical blank contributions at
155 1700°C were ~5% for ^{36}Ar and ^{84}Kr , and ~1% for ^{132}Xe , with the exception of the gas
156 poor sample NWA 11032, which had a blank contribution roughly 4 times higher than
157 other samples. The blank contribution to ^{40}Ar was considerably higher given the low
158 ^{40}Ar concentration in ureilite samples (Rai et al., 2003a), reaching 90% of the total
159 amount released from certain samples. Uncertainties in isotopic ratios (2σ) include
160 blank correction, external (reproducibility of the standards) and internal (statistical
161 and internal mass fractionation) error propagation.

162

163 **3. Results**

164 Noble gas (Ar, Kr and Xe) abundances and isotopic data are presented in tables 1, 2
165 and 3. Bulk concentrations (1100°C + 1700°C) of ^{36}Ar , ^{84}Kr and ^{132}Xe vary by an
166 order of magnitude across samples (Table 1; $^{36}\text{Ar} = 4.1 - 40.3 \times 10^{-12}$, mol/g, $^{84}\text{Kr} =$
167 $4.6 - 25.1 \times 10^{-14}$ mol/g, $^{132}\text{Xe} = 3.7 - 47.6 \times 10^{-14}$ mol/g). Concentrations of Ar, Kr and
168 Xe are within the range previously published for ureilites, with the exception of
169 sample NWA 8172, which has lower ^{36}Ar and ^{132}Xe concentrations than previously
170 measured (Figure 1). Amounts of Ar, Kr and Xe released during the 1100°C and
171 1700°C temperature steps are generally of the same order of magnitude, with the

172 exception of sample NWA 11368, which released significantly (10 to 100 times) more
173 Ar, Kr and Xe during the 1700°C temperature step than during the 1100°C release
174 (Table 1). The 600°C temperature steps released significantly less gas. This is
175 consistent with the majority of the noble gases being contained in diamonds, which
176 release noble gases at higher temperature (>800°C) than does amorphous carbon
177 (≤ 500 °C; Rai et al., 2002). The 600°C temperature releases consistently contained
178 atmospheric-derived noble gases and, as such, we do not consider these data further
179 in the discussion.

180

181 With regard to elemental ratios, the $^{36}\text{Ar}/^{132}\text{Xe}$ and $^{84}\text{Kr}/^{132}\text{Xe}$ range between 11.5-
182 160.7 and 0.4-1.7, respectively falling within the same range as previously reported
183 for ureilite samples (Figure 2), but do not attain the maximum values of 1090.9 and
184 3.0, respectively, measured in the carbon-rich residue of Goalpara (Göbel et al
185 1978). In general the $^{36}\text{Ar}/^{132}\text{Xe}$ and $^{84}\text{Kr}/^{132}\text{Xe}$ are higher for the 1700°C release
186 compared to the 1100°C one (Table 1). Both the $^{36}\text{Ar}/^{132}\text{Xe}$ and $^{84}\text{Kr}/^{132}\text{Xe}$ are
187 negatively correlated with Mg# (with the exception of the gas-poor sample NWA
188 8172), which also appears to be the case for previously analysed ureilites (Figure 3).

189

190 In general, the noble gas isotopic signatures of ureilites are similar to that of phase Q
191 (Busemann et al., 2000). The $^{38}\text{Ar}/^{36}\text{Ar}$ ranges from 0.1842 (± 0.0006) to 0.2053 (\pm
192 0.0024), similar to values previously reported for ureilites (table 1). Deviations in
193 $^{38}\text{Ar}/^{36}\text{Ar}$ from phase Q (0.187; Busemann et al., 2000) indicate that cosmogenic Ar
194 ($^{38}\text{Ar}/^{36}\text{Ar} = 1.54$; Hohenburg et al., 1978) has contributed a small, less than 2%,
195 proportion to the total Ar measured. Krypton and xenon isotopes (Tables 2 and 3) are
196 dominated by a Q-like signature, with negligible cosmogenic contribution. Isotopic

197 variations exist for both Kr and Xe, with values both isotopically heavier and lighter
198 than phase Q (Figure 3, 4). The Kr and Xe isotopic signatures of all samples lie
199 intermediate between solar and the presolar HL component (Huss and Lewis, 1994a;
200 Figure 4). The maximum measured $^{134}\text{Xe}/^{132}\text{Xe}$ (0.382 ± 0.001) and $^{136}\text{Xe}/^{132}\text{Xe}$
201 (0.319 ± 0.001) are similar than has been previously reported maximum values for
202 ureilite samples ($^{134}\text{Xe}/^{132}\text{Xe} = 0.381 \pm 0.004$ and $^{136}\text{Xe}/^{132}\text{Xe} = 0.318 \pm 0.001$; Nagao
203 et al., 2014) corresponding to up to a 1.1 ± 0.3 % excess (NWA 2236) relative to
204 phase Q (Table 3, Figure 3, 4). Interestingly, $^{136}\text{Xe}/^{132}\text{Xe}$ ratios are positively
205 correlated with Mg# (Figure 3). However, we find no clear correlation between Mg#
206 and Kr isotopic ratios (Figure 3).

207

208 **4. Discussion**

209 **4.1 Evaluating the origin of trapped heavy noble gases in ureilites**

210 The primordial heavy noble gas inventory of chondrites is primarily dominated by
211 phase Q, which is thought to be associated with refractory organic compounds
212 (Busemann et al., 2000). Relative to those of solar noble gases, the elemental
213 abundances of phase Q are highly fractionated in favour of Ar, Kr, and Xe, while the
214 isotopic compositions of each element are fractionated in favour of the heavier
215 isotopes (Busemann et al., 2000; Rai et al., 2003a). In addition, up to 10% of the
216 heavy noble gas budget in chondrites can be derived from presolar grains (Ott,
217 2014). These grains contain a variety of noble gas components (HL, P3 and P6),
218 which have their own distinct noble gas isotopic signatures (Huss and Lewis, 1994a).

219

220 Primitive achondrites such as ureilites often retain high concentrations of primordial
221 noble gases (Wieler et al., 2006). Heavy noble gas isotopes in primitive achondrites

222 are dominated by a Q-like component (Rai et al., 2003), indicating that achondrites
223 may have formed directly from the processing of chondritic material (Wieler et al.,
224 2006). In addition to chondritic phase Q, achondrites have also been shown to
225 contain solar-derived noble gases introduced either during their formation
226 (Busemann et al., 2006), or through later implantation of solar wind (Busemann and
227 Eugster, 2002).

228

229 Processes previously advocated for the trapping of noble gases in ureilite
230 carbonaceous phases include: low temperature equilibrium solubility or physical
231 adsorption (Wacker, 1989), ion implantation within the nebular plasma (Göbel et al.,
232 1978; Kuga et al., 2015; Rai et al., 2003a) and multicomponent mixing (Wilkening
233 and Marti, 1976). Among the aforementioned processes, the introduction of noble
234 gases during ion implantation has been favoured among the community (Göbel et al.,
235 1978; Matsuda et al., 1991; Rai et al., 2003a). Laboratory experiments have shown
236 that, under ionising conditions, gases with a nebular-like composition can produce
237 organic solids, structurally similar to chondritic organics whilst displaying the same
238 elemental and isotopic fractionation for noble gases as that exhibited by phase Q
239 (Kuga et al., 2015, 2017). A similar ionisation regime has been proposed for the
240 introduction of noble gases into ureilite diamonds and amorphous carbon (Matsuda
241 et al., 1991; Matsuda and Yoshida, 2001; Rai et al., 2003a). The similarity between
242 the isotopic heavy noble gas signature in ureilites and phase Q indeed suggests that
243 ureilites may have accreted chondritic material, or acquired their noble gases
244 independently in the same way as chondrites (Rai et al., 2003a).

245

246 Mass-independent variations in oxygen isotopes within ureilites have also been
247 previously ascribed to nebular processes (Clayton and Mayeda, 1988). Taken
248 together with the fact that ureilites fall along the ~1 CCAM line in the oxygen three
249 isotope plot, the oxygen isotopes in ureilites are considered to be the result of mixing
250 between isotopically distinct nebular reservoirs (Clayton and Mayeda, 1999). When
251 the $^{36}\text{Ar}/^{132}\text{Xe}$ and $^{84}\text{Kr}/^{132}\text{Xe}$ ratios of ureilites found in the literature are plotted
252 against $\Delta^{17}\text{O}$ (Figure 5), it is apparent that noble gas elemental ratios are positively
253 correlated with oxygen isotopes. The correlation between noble gas elemental ratios
254 and oxygen isotopes indicates that the variation in noble gas elemental ratios and
255 $\Delta^{17}\text{O}$ in ureilites were produced by similar, perhaps nebular, processes.

256

257 The hypothesis of a common nebular origin for noble gases and oxygen in ureilites is
258 however not without problems. Firstly, signatures inherited from the nebula would
259 have to have survived planetary melting and recrystallisation processes within the
260 UPB mantle whilst it underwent differentiation and a cataclysmic impact (Goodrich et
261 al., 2015). Furthermore, no evidence of presolar diamonds has been found in
262 ureilites, therefore requiring that planetary processing destroyed any presolar
263 diamonds accreted by the UPB, whilst co-existing nebular diamonds were somehow
264 able to survive (Wieler et al., 2006). Whereas ureilites and phase Q show a similar
265 extent of fractionation in Ar, Kr and Xe isotopes from solar, ureilites exhibit a much
266 larger range of elemental fractionation i.e. $^{36}\text{Ar}/^{132}\text{Xe}$ and $^{84}\text{Kr}/^{132}\text{Xe}$ ratios, with
267 values ranging from Q/HL towards values more similar to solar. These two
268 observations could be attributed the nebular plasma, from which the ions were
269 implanted, being at a higher and more variable temperatures than was the case for
270 phase Q (Rai et al., 2003a). However, how the UPB preserved evidence for high

271 temperature nebular processes whilst no other planetesimal or chondritic body,
272 including those that presumably formed closer to the Sun, was able to do so remains
273 unknown. Finally, and perhaps most crucially, if the noble gas-bearing carbon phases
274 and the oxygen-isotope-bearing silicates had a similar, nebular origin (Figure 5), then
275 one would expect silicates to also contain implanted noble gases, which is not the
276 case (Göbel et al., 1978; Weber et al., 1976).

277

278 Figure 3 suggests that variation in $^{36}\text{Ar}/^{132}\text{Xe}$ and $^{84}\text{Kr}/^{132}\text{Xe}$ within ureilites is also
279 related to the processes controlling the variation in Mg#. Variations in Mg# are most
280 likely controlled by differences in redox conditions within the UPB (Goodrich et al.,
281 2007; Sanders et al., 2017), or within the accretionary materials that formed the UPB
282 (Barrat et al., 2017). We therefore suggest that the range of noble gas signatures
283 within ureilites is not the result of implantation within a nebular plasma, of varying
284 temperatures, but most likely represents the range of compositions that were present
285 within the carbon-rich mantle of the UPB, either as the result of processing within the
286 mantle of the UPB, or inherited from different accretionary material. Noble gases
287 within ureilite diamonds would therefore have been either incorporated during their
288 formation at high pressure and temperature within the UPB mantle, or during shock
289 events, rather than being implanted into diamonds in the solar nebula
290 (Lipschutz, 1964; Miyahara et al., 2015; Nabeie et al., 2018; Takeda et al., 2001).

291

292 The *in situ* origin of ureilite diamonds has been previously contested due to the fact
293 that nitrogen isotopes released at high temperature from the diamonds ($\delta^{15}\text{N} < -$
294 100‰) are distinct from those measured within graphitic and amorphous carbon
295 phases ($\delta^{15}\text{N} > +20\text{‰}$) (Rai et al., 2003b). It was therefore concluded that diamonds

296 could not have formed directly from the conversion of graphite or amorphous carbon
297 (Rai et al., 2003b). However, nitrogen may be carried in multiple phases, and so may
298 not be associated with the same carbonaceous phases as those carrying noble
299 gases within ureilites (Yamamoto et al., 1998). The original source of nitrogen now
300 found in ureilite diamonds, and its potential relationship to noble gases, therefore
301 remains to be resolved.

302

303 **4.2 Evolution of noble gases within the ureilite parent body**

304 In the following section, we outline and discuss the processes that could have
305 resulted in the observed correlation between noble gases, Mg# and $\Delta^{17}\text{O}$ on the
306 UPB.

307

308 **4.2.1 Mass dependent fractionation during gas loss**

309 Variations in $^{36}\text{Ar}/^{132}\text{Xe}$ and $^{84}\text{Kr}/^{132}\text{Xe}$ measured in ureilites, from values similar to
310 phase Q and presolar HL towards more elevated values more similar to the solar
311 composition, have previously been attributed to diffusive mass-dependent
312 fractionation during gas loss (Okazaki et al., 2003). If ureilites originated with a high
313 $^{36}\text{Ar}/^{132}\text{Xe}$ and $^{84}\text{Kr}/^{132}\text{Xe}$, similar to the maximum values found in the diamond
314 bearing Goalpara ureilite (Göbel et al., 1978), then preferential loss of the light noble
315 gases during episodes of heating could have lowered the $^{36}\text{Ar}/^{132}\text{Xe}$ and $^{84}\text{Kr}/^{132}\text{Xe}$
316 towards (and even beyond) Q-like values (Figure 2). To explain the full range of
317 $^{36}\text{Ar}/^{132}\text{Xe}$ and $^{84}\text{Kr}/^{132}\text{Xe}$ within ureilites would require that different ureilites have
318 undergone variable degrees of heating and degassing, within the UPB itself or during
319 its break up (Rai et al., 2003a).

320

321 It remains unclear how exactly mass-dependent fractionation of noble gases during
322 gas loss could result in the correlation seen between noble gas elemental ratios and
323 Mg#. One possibility is that the loss of noble gases occurred with the release of CO
324 during smelting (Goodrich et al., 1987, 2007). Given their incompatible nature, noble
325 gases would likely be expelled along with CO during smelting reactions, and release
326 favoured the lighter elements (Ar), then the residue in the mantle would be
327 preferentially enriched in the heavy noble gases (Kr and Xe). Smelting however fails
328 to account for both the range of $\Delta^{17}\text{O}$ and the correlation seen between oxygen and
329 Mg# in ureilites (Figure 3, 5; Sanders et al., 2017), and cannot therefore be
330 responsible for the correlation between noble gases, Mg# and $\Delta^{17}\text{O}$.

331
332 The other possibility is that the UPB was heterogeneous with regard to Mg#, and Mg-
333 rich ureilites simply underwent larger degrees of gas loss. Mg-rich ureilites record
334 significantly higher temperatures than the ferroan ureilites (Warren, 2012), which
335 could reflect greater relative loss and fractionation of the noble gases during the
336 break up of the UPB. This would however imply that low $^{36}\text{Ar}/^{132}\text{Xe}$ and $^{84}\text{Kr}/^{132}\text{Xe}$,
337 Mg-rich ureilites lost proportionally more gas than the more ferroan samples,
338 whereas no correlation is found between noble gas concentrations and the Mg#, or
339 the $^{36}\text{Ar}/^{132}\text{Xe}$ and $^{84}\text{Kr}/^{132}\text{Xe}$, of the samples (Figure 1). In fact, some of the Mg-rich
340 ureilites with low, Q-like $^{36}\text{Ar}/^{132}\text{Xe}$ and $^{84}\text{Kr}/^{132}\text{Xe}$ are some of the most gas-rich
341 samples (Yamamoto et al., 1998). Therefore, to attain Q-like $^{36}\text{Ar}/^{132}\text{Xe}$ and
342 $^{84}\text{Kr}/^{132}\text{Xe}$ values in the Mg-rich ureilites from a starting composition similar to that of
343 Goalpara (Göbel et al., 1978) would require the loss of > 99% of the noble gases via
344 diffusive loss (Figure 2). High Mg# ureilites such as ALH 77257 (Yamamoto et al.,
345 1998) should then have originated with ^{132}Xe concentration more than 2 orders of

346 magnitude greater than that of phase Q (Busemann et al., 2000). Given that phase Q
347 is the main carrier of heavy noble gases in chondritic material, it would be unlikely for
348 ureilites to have been enriched in noble gases to such a degree. We therefore
349 conclude that diffusive gas loss controls neither the range of noble gas elemental
350 ratios in ureilite, nor their correlation with Mg# and $\Delta^{17}\text{O}$. We do however note that
351 diffusive loss may still play a role in explaining why some ureilite samples such as
352 ALH 82130 (Rai et al., 2003a) have $^{36}\text{Ar}/^{132}\text{Xe}$ and $^{84}\text{Kr}/^{132}\text{Xe}$ lower than that of
353 primordial phase Q or HL (Figure 2; Rai et al., 2003a).

354

355 **4.2.2 Two component mixing**

356 Two-component mixing has been advocated previously to explain the correlation
357 between $\Delta^{17}\text{O}$, FeO/MnO and Mg# within ureilites (Barrat et al., 2017; Sanders et al.,
358 2017), as well as the range of $^{36}\text{Ar}/^{132}\text{Xe}$ and $^{84}\text{Kr}/^{132}\text{Xe}$ in ureilites (Murty et al.,
359 2010). Ideally, the same two-component mixing models should also be able to
360 account for the noble gas signatures, given that Mg# correlates with the elemental
361 ratios of Ar/Xe and Kr/Xe, and also weakly correlates with the isotopic ratio of
362 $^{136}\text{Xe}/^{132}\text{Xe}$. The first model, proposed by Sanders et al., (2017), suggests that the
363 UPB accreted low $\Delta^{17}\text{O}$ silicates together with variable amounts of iron metal and
364 high $\Delta^{17}\text{O}$ nebular ice (Sanders et al., 2017). The second model, proposed by Barrat
365 et al., (2007), envisages that the UPB formed from two distinct C-rich planetesimals
366 with different $\Delta^{17}\text{O}$ and Mg#. In following sections, we evaluate how well each of
367 these two models can reproduce the trends defined by the noble gas data.

368

369 **4.2.2.1 Chondritic phase Q and nebular ice**

370 In the model proposed by Sanders et al., (2007), nebular ice with high $\Delta^{17}\text{O}$ accreted
371 together with low $\Delta^{17}\text{O}$ silicates to the UPB. Water derived from the melting of the
372 nebular ice component resulted in the uneven aqueous oxidation of metal within the
373 UPB, yielding secondary FeO-bearing minerals with high $\Delta^{17}\text{O}$. The incorporation of
374 these FeO-bearing phases into silicates during igneous and/or metamorphic
375 processes could then have resulted in the observed correlation between oxygen
376 isotopes and redox-controlled Mg# and FeO/MnO (Sanders et al., 2017).

377

378 Aqueous alteration processes akin to those that occurred on the parent body of
379 carbonaceous chondrites from the melting of nebular ice can also affect noble gas
380 elemental ratios. The ratios of $^{36}\text{Ar}/^{132}\text{Xe}$ and $^{84}\text{Kr}/^{132}\text{Xe}$ in carbonaceous chondrites
381 are correlated with the petrographic type of the meteorites, with the most aqueously
382 altered samples having lower $^{36}\text{Ar}/^{132}\text{Xe}$ and $^{84}\text{Kr}/^{132}\text{Xe}$ (Bekaert et al., 2019;
383 Busemann et al., 2000; Yamamoto et al., 2006). Laboratory experiments showed that
384 this preferential loss of Ar and Kr occurs irrespective of the composition of the
385 alteration fluid (in this case air saturated laboratory water with $^{36}\text{Ar}/^{132}\text{Xe}$ and
386 $^{84}\text{Kr}/^{132}\text{Xe}$ ratio of 434 and 15, respectively; Yamamoto et al., 2006). Therefore, even
387 if the nebular ice component was depleted in Xe relative to lighter noble gases, as
388 suggested by laboratory-based experiments (Bar-Nun and Owen, 1998; Notesco et
389 al., 2003) and measurements of cometary ice (Marty et al., 2017; Rubin et al., 2018),
390 it would only serve to lower the $^{36}\text{Ar}/^{132}\text{Xe}$ and $^{84}\text{Kr}/^{132}\text{Xe}$ during periods of aqueous
391 interaction with the original phase Q component. These observations lead us to
392 conclude that aqueous alteration is unlikely to be the main process driving the
393 evolution of originally chondritic Q-like $^{36}\text{Ar}/^{132}\text{Xe}$ and $^{84}\text{Kr}/^{132}\text{Xe}$ towards higher
394 values.

395

396 **4.2.2.2 Chondritic phase Q and a solar component**

397 The model proposed by Barrat et al., (2017) calls for the incomplete mixing between
398 two planetesimals with different oxygen and carbon isotopic signatures, as well as
399 different Mg#, in order to explain the range of compositions found in ureilites. If these
400 two planetary reservoirs had different noble gas compositions then this model could
401 also account for the correlation between Mg#, $\Delta^{17}\text{O}$ and noble gas elemental ratios in
402 ureilites.

403

404 Given the general similarity in Ar, Kr and Xe isotopes between phase Q and ureilites,
405 regardless of their Mg#, it is fair to assume that phase Q constituted the major source
406 of noble gases in the UPB (Figure 4). However, the range of $^{36}\text{Ar}/^{132}\text{Xe}$ and
407 $^{84}\text{Kr}/^{132}\text{Xe}$ in ureilites requires another noble gas component with higher $^{36}\text{Ar}/^{132}\text{Xe}$
408 and $^{84}\text{Kr}/^{132}\text{Xe}$ than phase Q was also accreted to the UPB (Figure 2). One potential
409 source for the high $^{36}\text{Ar}/^{132}\text{Xe}$ and $^{84}\text{Kr}/^{132}\text{Xe}$ within ureilites is solar-derived noble
410 gases. Solar noble gases are thought to have been trapped within the interior of
411 planetary bodies, including the interiors of Mars, the Earth and the angrite parent
412 body, during their accretion (Busemann et al., 2006; Matthew and Marti, 2001;
413 Williams and Mukhopadhyay, 2019). Solar noble gases may be added to planetary
414 interiors through the gravitational capture of solar nebula gas (Williams and
415 Mukhopadhyay, 2019), or by the irradiation of pre-accretionary material by the solar
416 wind (Busemann et al., 2006).

417

418 Noble gas elemental ratios within ureilites, with the exception of ALH 82130 (Rai et
419 al., 2003a) and the carbon-rich residue of Goalpara (Göbel et al., 1978) fall along a

420 mixing line between a chondritic Q/HL-bearing component and subsolar (Figure 2;
421 Patzer and Schultzi, 2002). The reason why ALH 82130 and the residue from
422 Goalpara do not fall on this trend is not known. It has been suggested that ALH
423 82130 may have undergone a peculiar heating event leading to significant gas loss
424 and fractionation (Nagao et al., 2014), and it is noted that the bulk value for Goalpara
425 does fit within the chondritic Q/HL-implanted solar mixing trend (Mazor et al., 1970;
426 Göbel et al., 1978).

427

428 The trend shown in figure 3 indicates that solar-derived noble gases may have been
429 present within the low Mg# component in ureilites. The solar component within the
430 glass of the D'Orbigny angrite has a trapped $^{20}\text{Ne}/^{22}\text{Ne}$ ratio of 11.9 ± 0.3 ,
431 intermediate between the values for solar wind (13.8; Benkert et al., 1993), and
432 fractionated solar wind ($^{20}\text{Ne}/^{22}\text{Ne} = 11.2$; Busemann et al., 2006). This value is
433 higher than the Q-like trapped Ne component ($^{20}\text{Ne}/^{22}\text{Ne} = 10.7$, regularly found in
434 ureilites; Busemann et al., 2000), potentially arguing against the contribution to the
435 UPB of noble gases that consist only of solar or fractionated solar wind. However,
436 recent analyses of Ne isotopes in the main group ureilite JAH 422 suggest it contains
437 a trapped solar wind component (Leya and Stephenson, 2019).

438

439 Implanted solar wind is also postulated to be at the origin of the subsolar component,
440 which is primarily found in enstatite chondrites (Patzer and Schultz, 2002). The
441 subsolar component in enstatite chondrites is often found in conjunction with small
442 deviations in Xe (and presumably Kr) isotopes from phase Q towards solar-like
443 values (Patzer and Schultz, 2002). This is however not always the case, as subsolar
444 elemental ratios are often found without evidence for isotopically solar noble gases

445 (Okazaki et al., 2010). The addition of subsolar noble gases to the UPB could
446 therefore introduce high $^{36}\text{Ar}/^{132}\text{Xe}$ and $^{84}\text{Kr}/^{132}\text{Xe}$, without significantly changing the
447 Q-like isotopic ratios. Interestingly, the NWA 8172 (this study) and JAH 422 (Leya
448 and Stephenson, 2019) samples, which potentially contain solar-derived noble
449 gases, have low Mg# of 76.9 and 78-80 (Janots et al., 2011), respectively, further
450 suggesting that the solar-derived noble gas component could be associated with a
451 low Mg# endmember in ureilites (Figure 3). We therefore propose that a two
452 component mixing within the UPB is capable of producing the range of noble gas
453 elemental and isotopic ratios measured within ureilites, and can also account for the
454 correlation between noble gas elemental ratios, Mg# and $\Delta^{17}\text{O}$ (Figure 3, 5). In the
455 next section we outline in detail the composition and potential origin of these two
456 components, as well as discuss their significance with regard to volatile distribution in
457 the early Solar System.

458

459 **4.3 Heterogeneous volatile element distribution in the early Solar System**

460 Incorporating the noble gases into the mixing models proposed by Barrat et al.,
461 (2017) results in one component having high Mg# olivines, $\Delta^{17}\text{O} \sim -2.5\text{‰}$, $\delta^{13}\text{C} \sim -$
462 8‰ , and low, Q/HL-like noble gas elemental ratios. The other component is then
463 inferred to have low Mg#, $\Delta^{17}\text{O} \sim -0.5\text{‰}$, $\delta^{13}\text{C} \sim 0\text{‰}$, and higher (potentially solar-
464 derived) noble gas elemental ratios. Noble gas elemental ratios do not appear
465 correlated with the $\delta^{13}\text{C}$ values given by Barrat et al., 2017 (Figure 5c,d). Similarly, a
466 weak correlation is exhibited between $\delta^{13}\text{C}$ and Mg# with Mg# between 80 and 95
467 due to a hyperbolic mixing relationship (Barrat et al., 2017).

468

469 Although the two potential planetesimals that merged to form the UPB are required to
470 have been dominated by phase Q-like noble gas isotopic signatures (Figure 4), as is
471 the case for most planetary bodies in the solar system (Busemann and Eugster,
472 2002), they may have had slight differences in their respective amounts of HL-Xe
473 (Figure 6), as evidenced by $^{134}\text{Xe}/^{132}\text{Xe}$ and $^{136}\text{Xe}/^{132}\text{Xe}$ of the Mg-rich samples being
474 in excess of phase Q. Variable contributions of HL-Xe within phase Q from different
475 chondrites were originally interpreted to represent different degrees of processing on
476 planetary bodies leading to loss of the presolar components (Gilmour, 2010; Huss
477 and Lewis, 1994b). It has also been proposed that the HL-Xe contribution within
478 phase Q increased with increasing heliocentric distances in the solar system, with
479 carbonaceous chondrites containing a higher proportion of presolar HL-Xe than
480 enstatite chondrite, which formed closer to the sun (Bekaert et al., 2019). Although
481 the Xe isotopic composition group around those of phase Q, the Mg-rich samples
482 exhibit a resolvable excess in $^{136}\text{Xe}/^{132}\text{Xe}$ ($1.1 \pm 0.3\%$ for the highest Mg# sample
483 NWA 2236), consistent with a greater contribution from Xe-HL compared to low Mg#
484 ureilites (Figure 3, 6). However, if the low Mg# ureilites contained some isotopically
485 solar Xe, it could have lowered their bulk $^{136}\text{Xe}/^{132}\text{Xe}$ ratios, so both planetesimals
486 may have originated with a similar excess in Xe-HL. As ureilites do not contain any
487 presolar diamonds, the primary carriers of presolar HL-Xe (Huss and Lewis, 1994a),
488 the Xe-HL component was most likely accreted to the UPB as part of the high Mg#
489 planetesimal. The excess in Xe-HL from Q measured in the high Mg# ureilites
490 approaches that found within the insoluble organic matter of carbonaceous
491 chondrites (Figure 6; Bekaert et al., 2019), suggesting that the high Mg# component
492 experienced limited thermal processing prior to its accretion to the UPB (Gilmour,
493 2010; Huss and Lewis, 1994b).

494

495 The nature of the secondary solar-derived noble gas component within the UPB is
496 less well constrained. The implantation of solar wind within silicate material is
497 considered to be most effective within close proximity to the Sun, hence explaining
498 why the subsolar component is most commonly found in enstatite chondrites
499 (Okazaki et al., 2001). Although primarily found in enstatite chondrites, this
500 "component" has also been discovered in ordinary and carbonaceous type
501 chondrites (Ott, 2014; Schelhaas et al., 1990) and the implantation of solar wind in
502 pre-accretionary materials has been proposed as the origin for the solar component
503 in the angrite parent body (Busemann et al., 2006). Solar wind irradiated material
504 therefore appears to have been efficiently transported throughout the protoplanetary
505 disk.

506

507 **Conclusion**

508 The noble gas elemental ratios of ureilites are shown to be heterogeneous from one
509 sample to the next, with the majority of $^{36}\text{Ar}/^{132}\text{Xe}$ and $^{84}\text{Kr}/^{132}\text{Xe}$ values being
510 intermediate between those of chondritic phase Q and presolar Xe-HL, and an
511 implanted solar wind component. The $^{36}\text{Ar}/^{132}\text{Xe}$ and $^{84}\text{Kr}/^{132}\text{Xe}$ ratios exhibit a
512 negative correlation with Mg# and positive correlation with $\Delta^{17}\text{O}$ indicating that noble
513 gases share a genetic link with the silicate fraction. The range of noble gas elemental
514 ratios measured within the ureilites can be explained the mixing of two distinct noble
515 gas components, one inherited directly from chondritic-like precursors (phase Q), and
516 the other a solar-derived component from implanted solar wind. The variations in
517 noble gas elemental ratios within ureilites indicate that the noble gases have resisted
518 homogenisation during the formation and later break-up of the UPB.

519

520 The Kr and Xe isotopes measured as part of this study are dominated by trapped
521 noble gases that show an isotopic signature similar to that of phase Q, consistent
522 with previous studies. Our new data, in combination with those already published,
523 however show that the Xe isotopic signatures of ureilites range from values similar to
524 fractionated solar wind towards isotopically heavier values corresponding to material
525 enriched in presolar HL. The range of Xe isotope compositions measured within
526 ureilites is consistent with the idea of mixing between two distinct reservoirs,
527 indicating that planetesimals that formed in the disk accreted material with a variety
528 of noble gas signatures, potentially reflecting the heterogeneous nature of the early
529 inner solar system.

530

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538

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813

814 **Figure Captions**

815 **Figure 1.** Magnitude of concentration expressed in mol/g (on a logarithmic scale) of
816 ^{36}Ar (a), ^{84}Kr (b) and ^{132}Xe (c) within ureilites plotted against Mg# (Mg# = $100 \times$ molar
817 $\text{Mg}/(\text{Mg}+\text{Fe})$) measured within olivine cores. Noble gas concentrations of the
818 samples measured in this study are the bulk values released over the 1100°C and
819 1700°C temperature steps. Literature data are from Göbel et al., 1978; Mazor et al.,
820 1970; Müller and Zähringer, 1969; Ott et al., 1985; Weber et al., 1971; Rai et al.,
821 2003a; Wacker, 1986; Wilkening and Marti, 1976; Yamamoto et al., 1998. 2σ
822 uncertainties on the measurements from this study are smaller than symbol sizes.

823

824 **Figure 2.** Compilation of new and previously published (round symbols) elemental
825 ratios of trapped noble gases measured within ureilites. The $^{36}\text{Ar}/^{132}\text{Xe}$ and $^{84}\text{Kr}/^{132}\text{Xe}$
826 ratios form a trend, which ranges from values similar to phase Q towards higher
827 values. The trend also appears to correlate with the Mg# of the samples (identified by
828 the colour spectrum, see online version), with high Mg# samples being associated
829 with low $^{36}\text{Ar}/^{132}\text{Xe}$ and $^{84}\text{Kr}/^{132}\text{Xe}$. We show the predicted trends for two possible
830 mechanisms that may have given rise to the observed pattern; (i) diffusive gas loss
831 (curved dashed black line) from the highest $^{36}\text{Ar}/^{132}\text{Xe}$ and $^{84}\text{Kr}/^{132}\text{Xe}$ samples
832 (carbon-rich residue of Goalpara; Göbel et al., 1978), and (ii) two component mixing
833 (solid black line) between primitive Q/HL component and subsolar (Patzner and
834 Schultz, 2002). The HL-subsolar mixing line also intercepts the trapped solar
835 component found in D'Orbigny (dotted line; Busemann et al., 2006). The combined
836 1100°C and 1700°C data are shown for samples from this study, with the exception
837 of samples NWA 8049 and NWA 8172 which represent solely the 1700°C extraction.
838 Ureilite literature data are the same as referenced in Figure 1. The results for
839 cometary ice from Comet 67P/C-G are also shown (Rubin et al., 2018).

840

841 **Figure 3.** Noble gas elemental and isotopic ratios relative to Mg#. The $^{36}\text{Ar}/^{132}\text{Xe}$ (a),
842 $^{36}\text{Kr}/^{132}\text{Xe}$ (b) and $^{136}\text{Xe}/^{132}\text{Xe}$ (c) show a weak correlation with Mg#. The highest Mg#
843 samples appear to include a higher contribution from presolar Xe-HL. Krypton
844 isotopes (d) do not appear to be correlated with Mg# for the samples measured in
845 this study. Literature data are from the same sources referenced in Figure 1. Linear
846 trends are fitted through all the data to highlight the correlation. It should however be

847 noted that there is no *a priori* reason for the correlations to be linear. Uncertainties
848 from this study are shown to 2σ . Symbols for samples analysed in this study are the
849 same as in figure 1, although colour now represents the extraction temperature (see
850 online version).

851

852 **Figure 4.** Xenon and krypton isotopic ratios measured in the six ureilites. In general,
853 the Xe (a) and Kr (b) isotopes in ureilites are similar to phase Q. Some of the
854 samples show an excess in heavy Xe isotopes and plot along a mixing line between
855 phase Q and the presolar HL component. There does not appear to be a significant
856 contribution of solar wind Xe to the ureilite samples, although there is potentially
857 some evidence for a solar wind contribution in the Kr isotopes. Mixing line between
858 solar and Kr-HL is shown for reference. All uncertainties from this study are shown to
859 2σ .

860

861 **Figure 5.** $^{36}\text{Ar}/^{132}\text{Xe}$ and $^{84}\text{Kr}/^{132}\text{Xe}$ plotted against $\Delta^{17}\text{O}$ (a and b) and $\delta^{13}\text{C}$ (c and d).
862 Noble gas elemental ratios within ureilites appear to be positively correlated with
863 $\Delta^{17}\text{O}$. When multiple $^{36}\text{Ar}/^{132}\text{Xe}$ and $^{84}\text{Kr}/^{132}\text{Xe}$ values exist in the literature we have
864 highlighted the maximum value (green squares). The linear trends in figures a and b
865 is fitted through the samples with the maximum measured $^{36}\text{Ar}/^{132}\text{Xe}$ and $^{84}\text{Kr}/^{132}\text{Xe}$,
866 as these are assumed to be more representative of the original values since any gas
867 loss during metamorphism, atmospheric entry etc. is assumed to lower these ratios.
868 The $\delta^{13}\text{C}$ data from Barrat et al., 2017 and does not appear to correlate with the
869 noble gas elemental ratios (c and d). The linear trend in figures c and d is fitted
870 through all the data. These linear fits are only shown to highlight the correlation as
871 the relationship between the noble gases, oxygen and carbon may not be linear, as
872 is the case with carbon and Mg#; Barrat et al., 2017. Noble gas elemental ratios for
873 ureilites are provided by the same references given in the caption of Figure 1. The O
874 isotope data are from Clayton and Mayeda, 1996 and Greenwood et al., 2017.

875

876 **Figure 6.** Xenon three-isotope diagram ($^{134}\text{Xe}/^{132}\text{Xe}$ vs. $^{136}\text{Xe}/^{132}\text{Xe}$). The range of Xe
877 isotopic compositions previously measured in ureilites is shown together with
878 published data from Nagao et al., 2014 (grey filled circles). The $^{134}\text{Xe}/^{132}\text{Xe}$ and
879 $^{136}\text{Xe}/^{132}\text{Xe}$ ratios of ureilites range from values similar to that of fractionated solar

880 wind (Gilmour, 2010) and the average value of inner solar system enstatite
881 chondrites (EC; Patzer and Schultz, 2002), towards values measured within the
882 insoluble organic matter of carbonaceous chondrites (CC IOM; Bekaert et al., 2019)
883 and the bulk average carbonaceous chondrites composition (AVCC; Pepin, 2003).
884 Ureilites therefore retain a wide range of Xe isotopic signatures, perhaps indicating
885 that the ureilite parent body accreted material from different parts of the
886 protoplanetary disk. The dotted line represents the mass dependent fractionation line
887 (MDF).

Samples	mass (g)	T °C	Mg#	Shock Grade	³⁶ Ar (mol/g)	±	⁸⁴ Kr (mol/g)	±	¹³² Xe (mol/g)	±	³⁶ Ar/ ¹³² Xe	±	⁸⁴ Kr/ ¹³² Xe	±	³⁸ Ar/ ³⁶ Ar	±	
					x 10 ⁻¹²	x 10 ⁻¹²	x 10 ⁻¹⁴	x 10 ⁻¹⁴	x 10 ⁻¹⁴	x 10 ⁻¹⁴							
NWA 2236	0.01	600.00	97	Low	1.002	0.002	1.243	0.034	1.159	0.020	86.42	1.49	1.07	0.03	0.1847	0.0014	
					1100.00	10.721	0.013	5.761	0.221	14.466	0.118	74.08	0.61	0.40	0.02	0.1866	0.0011
					1700.00	17.602	0.316	19.303	1.418	15.434	0.117	114.05	2.23	1.25	0.09	0.1850	0.0016
					Total	28.319	0.317	25.063	1.435	29.900	0.166	94.71	1.18	0.84	0.05		
NWA 8172	0.03	600.00	76.9	Moderate	n.d.	n.d.	2.077	0.259	0.122	0.006	n.d.	n.d.	17.07	2.27	0.1986	0.0065	
					1100.00	n.d.	n.d.	1.984	0.007	2.644	0.016	n.d.	n.d.	0.75	0.01	0.2053	0.0024
					1700.00	2.046	0.047	4.074	0.007	2.384	0.013	85.82	1.93	1.71	0.01	0.1982	0.0024
					Total	2.046	0.047	6.057	0.010	5.028	0.021	85.82	1.93	1.20	0.01		
NWA 11368	0.01	600.00	89.9 - 90.7		0.147	0.001	1.004	0.202	0.635	0.011	23.18	0.41	1.58	0.32	0.1899	0.0022	
					1100.00	0.504	0.008	1.985	0.254	4.392	0.032	11.49	0.20	0.45	0.06	0.2024	0.0021
					1700.00	14.803	0.411	11.337	0.007	20.951	0.013	70.66	1.96	0.54	0.00	0.1843	0.0007
					Total	15.307	0.411	13.322	0.254	25.342	0.034	60.40	1.62	0.53	0.01		
NWA 7686	0.01	600.00	91		0.730	0.045	2.658	0.246	0.738	0.015	98.85	6.40	3.60	0.34	0.1995	0.0039	
					1100.00	10.961	0.433	8.658	0.468	15.608	0.130	70.22	2.90	0.55	0.03	0.1842	0.0006
					1700.00	29.321	0.991	16.328	0.163	32.023	0.193	91.56	3.14	0.51	0.01	0.1842	0.0014
					Total	40.283	1.085	24.986	0.496	47.631	0.233	84.57	2.32	0.52	0.01		
NWA 8049	0.01	600.00	84.3	Moderate	0.141	0.008	2.077	0.259	0.659	0.015	21.41	1.27	1.23	0.16	0.2056	0.0260	
					1100.00	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.		
					1700.00	18.263	1.280	14.006	1.447	11.367	0.066	160.66	11.30	1.23	0.13	0.1993	0.0061
NWA 11032	0.01	600.00	82.6 - 82.9	High	0.053	0.001	0.249	0.223	0.122	0.006	43.34	2.29	2.04	1.84	0.1958	0.0367	
					1100.00	0.868	0.007	1.265	0.309	1.282	0.018	67.71	1.10	0.99	0.24	0.2027	0.0008
					1700.00	3.264	0.006	3.305	0.390	2.398	0.035	136.13	2.01	1.38	0.16	0.1929	0.0011
					Total	4.133	0.009	4.569	0.498	3.680	0.040	112.29	1.23	1.24	0.14		
Phase Q ^a										76	7	0.81	0.05	0.1873	0.0007		
HL ^b										50	20	0.48	0.04	0.2268	0.0031		
Solar Wind ^c										22705	2783	9.50	1.0	0.1998	0.0002		
Subsolar ^d										582	270	2.40	1.42	0.1835	0.0013		

888 Table 1. Noble gas concentrations and elemental ratios of six Northwest African ureilites. Mg# have been previously presented in
889 Barrat et al., 2015^a; 2017^b and in the Meteoritical Bulletin Database^c. Total signifies gas released during the 1100°C and 1700°C
890 temperature steps. For sample NWA 8049 the total gas is not is presented as the gas from 1100°C temperature step was lost. n.d.
891 = not determined. ^a Busemann et al., 2000, ^b Huss and Lewis 1994a, ^c Heber et al., 2009, ^d Patzer and Schultz, 2002.
892

893

Samples	T °C	$^{78}\text{Kr}/^{84}\text{Kr}$	\pm	$^{80}\text{Kr}/^{84}\text{Kr}$	\pm	$^{82}\text{Kr}/^{84}\text{Kr}$	\pm	$^{83}\text{Kr}/^{84}\text{Kr}$	\pm	$^{86}\text{Kr}/^{84}\text{Kr}$	\pm
NWA 2236	600	0.0061	0.0014	0.0407	0.0038	0.2040	0.0719	0.2014	0.0136	0.3019	0.0374
	1100	0.0061	0.0001	0.0398	0.0010	0.2015	0.0039	0.2022	0.0033	0.3060	0.0038
	1700	0.0061	0.0001	0.0397	0.0016	0.2027	0.0025	0.2014	0.0027	0.3065	0.0023
NWA 8172	600	0.0061	0.0019	0.0409	0.0053	0.2050	0.0640	0.2014	0.0140	0.3025	0.0313
	1100	0.0060	0.0001	0.0395	0.0016	0.2019	0.0050	0.2016	0.0057	0.3102	0.0093
	1700	0.0059	0.0002	0.0388	0.0022	0.2011	0.0029	0.1999	0.0084	0.3101	0.0025
NWA 11368	600	0.0064	0.0013	0.0405	0.0041	0.2054	0.0551	0.2032	0.0150	0.3047	0.0254
	1100	0.0060	0.0010	0.0395	0.0025	0.2019	0.0037	0.2040	0.0093	0.3082	0.0055
	1700	0.0060	0.0001	0.0395	0.0017	0.2018	0.0026	0.2022	0.0062	0.3085	0.0033
NWA 7686	600	0.0062	0.0005	0.0403	0.0026	0.2030	0.0502	0.2018	0.0091	0.3029	0.0094
	1100	0.0061	0.0002	0.0396	0.0014	0.2019	0.0056	0.2024	0.0029	0.3073	0.0030
	1700	0.0060	0.0001	0.0396	0.0016	0.2020	0.0021	0.2024	0.0057	0.3074	0.0021
NWA 8049	600	0.0064	0.0009	0.0401	0.0022	0.2021	0.0331	0.2011	0.0110	0.3014	0.0150
	1100	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
	1700	0.0063	0.0002	0.0396	0.0010	0.20318	0.0015	0.2021	0.0038	0.3088	0.0063
NWA 11032	600	0.0070	0.0009	0.0407	0.0019	0.2030	0.0057	0.1977	0.0084	0.3048	0.0102
	1100	0.0062	0.0003	0.0398	0.0015	0.2016	0.0042	0.2008	0.0084	0.3047	0.0060
	1700	0.0056	0.0002	0.0378	0.0008	0.1979	0.0014	0.1998	0.0084	0.3136	0.0024
Phase Q ^a		0.00603	0.00003	0.0394	0.0001	0.2018	0.0002	0.2018	0.0002	0.3095	0.0005
HL ^b		0.0042	0.0020	0.0305	0.0010	0.1590	-	0.1989	0.0010	0.3623	0.0018
Solar Wind ^c		0.00642	0.00005	0.0413	0.0002	0.2055	0.0002	0.2034	0.0002	0.3012	0.0004

894

895 Table 2. Krypton isotopic composition of six Northwest African ureilites. Uncertainties are to 2σ . n.d. = not determined. ^a Busemann896 et al., 2000, ^b Huss and Lewis 1994a, ^c Meshik et al., 2014

Sample	T °C	$^{124}\text{Xe}/^{132}\text{Xe}$	±	$^{126}\text{Xe}/^{132}\text{Xe}$	±	$^{128}\text{Xe}/^{132}\text{Xe}$	±	$^{129}\text{Xe}/^{132}\text{Xe}$	±	$^{130}\text{Xe}/^{132}\text{Xe}$	±	$^{131}\text{Xe}/^{132}\text{Xe}$	±	$^{134}\text{Xe}/^{132}\text{Xe}$	±	$^{136}\text{Xe}/^{132}\text{Xe}$	±
NWA 2236	600	0.0042	0.0003	0.0037	0.0003	0.0784	0.0020	1.018	0.015	0.159	0.002	0.811	0.012	0.384	0.007	0.324	0.006
	1100	0.0046	0.0002	0.0040	0.0002	0.0829	0.0008	1.042	0.012	0.163	0.001	0.822	0.001	0.382	0.001	0.319	0.001
	1700	0.0045	0.0001	0.0040	0.0001	0.0822	0.0006	1.035	0.016	0.162	0.001	0.817	0.002	0.382	0.001	0.319	0.002
NWA 8172	600	0.0056	0.0004	0.0046	0.0004	0.0743	0.0025	0.932	0.026	0.143	0.004	0.763	0.022	0.403	0.012	0.321	0.010
	1100	0.0046	0.0002	0.0041	0.0003	0.0824	0.0018	1.032	0.026	0.162	0.003	0.819	0.003	0.378	0.002	0.314	0.001
	1700	0.0046	0.0002	0.0041	0.0003	0.0821	0.0015	1.031	0.012	0.162	0.002	0.818	0.002	0.379	0.001	0.316	0.002
NWA 11368	600	0.0043	0.0002	0.0038	0.0003	0.0786	0.0016	1.017	0.016	0.159	0.003	0.809	0.061	0.384	0.003	0.327	0.004
	1100	0.0045	0.0001	0.0040	0.0001	0.0814	0.0012	1.035	0.027	0.161	0.001	0.819	0.003	0.379	0.001	0.319	0.002
	1700	0.0044	0.0001	0.0040	0.0002	0.0815	0.0005	1.032	0.016	0.161	0.001	0.818	0.002	0.380	0.002	0.318	0.002
NWA 7686	600	0.0040	0.0002	0.0036	0.0002	0.0755	0.0022	1.004	0.013	0.156	0.003	0.798	0.068	0.382	0.002	0.323	0.006
	1100	0.0046	0.0001	0.0041	0.0001	0.0813	0.0006	1.037	0.020	0.163	0.001	0.822	0.004	0.379	0.003	0.316	0.002
	1700	0.0046	0.0001	0.0041	0.0001	0.0816	0.0006	1.032	0.017	0.162	0.001	0.816	0.004	0.379	0.001	0.317	0.001
NWA 8049	600	0.0044	0.0003	0.0039	0.0002	0.0792	0.0016	1.014	0.018	0.159	0.004	0.806	0.066	0.379	0.003	0.320	0.005
	1100	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
	1700	0.0046	0.0001	0.0041	0.0001	0.0823	0.0006	1.032	0.020	0.161	0.001	0.818	0.002	0.379	0.001	0.316	0.001
NWA 11032	600	0.0047	0.0003	0.0040	0.0002	0.0765	0.0017	0.966	0.015	0.153	0.003	0.757	0.014	0.392	0.008	0.317	0.007
	1100	0.0044	0.0002	0.0037	0.0002	0.0793	0.0012	1.032	0.012	0.161	0.002	0.815	0.002	0.378	0.003	0.316	0.004
	1700	0.0046	0.0002	0.0041	0.0001	0.0823	0.0006	1.035	0.011	0.162	0.001	0.821	0.003	0.377	0.001	0.314	0.001
Phase Q ^a		0.00455	0.00002	0.00406	0.00002	0.0822	0.0002	1.042	0.002	0.1619	0.0003	0.8185	0.0009	0.3780	0.0011	0.3164	0.0008
HL ^b		0.00839	0.00009	0.00564	0.00008	0.0905	0.0006	1.056	0.002	0.1542	0.0003	0.8457	0.0013	0.6356	0.0013	0.6991	-
Solar Wind ^c		0.00491	0.00007	0.00416	0.00009	0.8420	0.0003	1.041	0.001	0.1650	0.0004	0.8256	0.0012	0.3691	0.0007	0.3001	0.0006

897

898 Table 3. Xenon isotopic composition of six Northwest African ureilites. Uncertainties are to 2σ . n.d. = not determined. ^a Busemann

899 et al., 2000, ^b Huss and Lewis 1994a, ^c Meshik et al., 2014











