

1 **Determining the impactor of the Ordovician Lockne crater: Oxygen and**
2 **neon isotopes in chromite versus sedimentary PGE signatures**

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19 **ABSTRACT**

20 Abundant chromite grains with L-chondritic composition in the resurge deposits of the
21 Lockne impact crater (458 Myr old; dia. ~10 km) in Sweden have been inferred to represent
22 relict fragments of an impactor from the break-up of the L-chondrite parent body at 470
23 Ma. This view has been challenged based on Ir/Cr and platinum group element (PGE)
24 patterns of the same resurge deposits, and a reinterpretation of the origin of the chromite
25 grains. An impactor of the non-magmatic iron meteorite type was proposed instead. Here
26 we show that single-grain oxygen and noble-gas isotope analyses of the chromite grains
27 from the resurge deposits further support an origin from an L-chondritic asteroid. We also
28 present PGE analyses and Ir/Cr ratios for fossil L-chondritic meteorites found in mid-
29 Ordovician marine limestone in Sweden. The L-chondritic origin has been confirmed by
30 several independent methods, including major element and oxygen isotopic analyses of
31 chromite. Although the meteorites show the same order-of-magnitude PGE and Cr
32 concentrations as recent L chondrites, the elements have been redistributed to the extent
33 that it is problematic to establish the original meteorite type from these proxies. Different

34 PGE data processing approaches can lead to highly variable results, as also shown here for
35 the Lockne resurge deposits. We conclude that the Lockne crater was formed by an L-
36 chondritic impactor, and that considerable care must be taken when inferring projectile type
37 from PGEs in sedimentary ejecta deposits.

38

39 **1. Introduction**

40 Abundant fossil, L-chondritic meteorites in marine limestone from a mid-Ordovician (470
41 Ma) quarry in Sweden, and a two orders-of-magnitude increase in L-chondritic
42 micrometeorites in sedimentary strata of the same age worldwide, provide strong evidence
43 for a breakup of the L-chondrite parent body in the asteroid belt at that time (Schmitz et al.,
44 1996, 2001, 2008; Cronholm and Schmitz, 2010; Heck et al., 2010). Already in the 1960s
45 the young K-Ar gas retention ages of many recently fallen L chondrites were inferred to
46 reflect a major parent-body breakup event at around 500 Ma (Anders, 1964). Recently,
47 refined $^{40}\text{Ar}/^{39}\text{Ar}$ measurements of L chondrites indicate an age of 470 ± 6 Ma for the event,
48 which is the same age as for the sediments rich in L-chondritic material (Korochantseva et
49 al., 2007). Although there is robust evidence for a dramatic increase in the flux of
50 micrometeorites and meteorites to Earth for a few million years after the breakup event,
51 model predictions of a coeval increase in the flux of larger L-chondritic asteroids to Earth
52 are more difficult to test (Zappalà et al., 1998). Prominent changes around 470 Ma in
53 Earth's fauna, such as the onset of the Great Ordovician Biodiversification Event (Schmitz
54 et al., 2008), as well as an order-of-magnitude overrepresentation of mid-Ordovician impact
55 craters among Earth's well-dated craters (Schmitz et al., 2001) give some support for an
56 asteroid shower. More robust evidence, however, must come from studies of impact ejecta
57 layers in the geological record, and from identifying the type of impacting projectiles. The
58 problem is that large projectiles tend to become completely vaporized upon impact, leaving
59 behind only a chemical fingerprint that may be fractionated by various processes and thus
60 difficult to interpret. In rare cases pieces of the impactor are preserved, like in the Eltanin or
61 Morokweng impact events (Kyte, 2002; Maier et al., 2006), allowing safe identification of
62 projectile type. For one mid-Ordovician impact crater, the well-preserved, ca. 10 km
63 diameter Lockne crater (458 Ma) in central Sweden, Alwmark and Schmitz (2007) reported
64 an extreme enrichment of extraterrestrial chromite grains in the resurge deposits, the so
65 called Loftarstone. More than 75 extraterrestrial chromite grains per kg of Loftarstone were
66 found, which is three to four orders of magnitude more grains than in normal marine
67 limestone. The chromite grains have an L-chondritic element composition and hence were

68 interpreted as relict fragments of an impactor related to the 470 Ma L-chondrite breakup
69 event. Tagle et al. (2008) challenged this view based on Ir/Cr and platinum group element
70 (PGE) ratios of the Loftarstone, as well as a reinterpretation of the compositional signature
71 of the chromite grains. They suggested that the impactor was a non-magmatic iron
72 meteorite (NMI). This is a viable suggestion considering that PGE signatures of
73 sedimentary ejecta have been shown to be usable under some conditions to identify
74 projectile types (e.g., Evans et al., 1993).

75 In order to further constrain the origin of the Lockne crater we present here single-
76 grain oxygen and noble gas isotopic data for the purported L-chondritic chromite grains
77 from the Loftarstone. High precision oxygen isotopic analyses of fossil extraterrestrial
78 chromite has proven to be a reliable method to identify precursor meteorite types
79 (Greenwood et al., 2007; Heck et al., 2010). Neon isotopes can discern if an extraterrestrial
80 chromite grain originates from a micrometeorite (solar-wind implanted Ne), meteorite
81 (cosmic-ray Ne) or the interior of an extraterrestrial body with a size greater than the
82 penetration depth (> ca. 2 m) of galactic cosmic rays (Heck et al., 2008; Meier et al., 2010).
83 We also present PGE and Ir/Cr data for mid-Ordovician fossil meteorites for which the L-
84 chondritic origin has been confirmed based on independent proxies, e.g. oxygen isotopes,
85 element composition and silicate inclusions in chromite, and petrography including
86 chondrule appearances. In light of these data we discuss the utility of PGE patterns in
87 sediments for determining impactor types.

88

89 **2. Materials and methods**

90 The chromite grains (ca. 63-100 μm in diameter) studied here originate from the two
91 Loftarstone samples FF2 and FF4 collected at the edge of the inner Lockne crater (see
92 figure 1 in Sturkell, 1998, for sample locations). The samples contain 2.5 and 2.0 ppb Ir,
93 respectively (Sturkell, 1998). Alwmark and Schmitz (2007) recovered abundant
94 extraterrestrial chromite grains from sample FF2, however, because of shortage of this
95 material 50 g of sample FF2 and 300 g of sample FF4 were used for the present study. The
96 procedure of recovering grains was the same as in Alwmark and Schmitz (2007) except that
97 the HCl- and HF-leached residue fractions were not heated to remove coal, because this
98 could have an effect on the noble gases.

99 Seven chromite grains with L-chondritic element composition according to Alwmark
100 and Schmitz (2007) were selected from the Loftarstone for O isotope analysis, using a
101 CAMECA IMS-1280 ion microprobe at the WiscSIMS Laboratory, University of

102 Wisconsin-Madison (Kita et al., 2009). The grains were mounted in the center of 25 mm
103 epoxy plugs with chromite standards UWCr-2 and UWCr-3 and polished to a flat low-relief
104 surface (Heck et al., 2010). We performed oxygen-three-isotope analyses on the grains
105 using the same instrument setup and analytical conditions as in the approach optimized for
106 chromite analyses by Heck et al. (2010). The primary ion beam at 5 nA intensity was
107 focused to a 15 μm spot. In total 10 analyses, bracketed by standard analyses, were
108 obtained on the chromite grains. We are able to achieve precisions $\leq 0.3\%$ (2SD) for
109 $\delta^{18}\text{O}$ and $\delta^{17}\text{O}$ and $\sim 0.2\%$ for $\Delta^{17}\text{O} = (\delta^{17}\text{O} - 0.52 \times \delta^{18}\text{O})$ from single spot analysis. The
110 contribution of OH⁻ interference to $^{17}\text{O}^-$ was typically less than 0.1‰ and the uncertainties
111 of the correction on $^{17}\text{O}/^{16}\text{O}$ ratios were insignificant ($< 0.05\%$). This precision is sufficient
112 to distinguish H versus L or LL chondrites. Values of $\Delta^{17}\text{O}$ from chromite grains from
113 recently fallen H, L and LL chondrites, analyzed during the same session and reported by Heck
114 et al. (2010), fell onto $\Delta^{17}\text{O}$ group averages obtained from bulk meteorite fluorination analysis
115 (Clayton et al., 1991), and demonstrate the reliability of our analytical method (Fig. 1). The
116 seven polished grains were also analysed for major and minor elements with a CAMECA
117 SX-51 electron probe microanalyzer (EPMA) at UW-Madison. For further details on the
118 SIMS and EPMA procedures, see Heck et al. (2010).

119 In addition six L-chondritic chromite grains from the Loftarstone, each grain
120 weighing between 1 and 4 μg , were analyzed for cosmogenic ^3He and ^{21}Ne at ETH-Zürich.
121 As the amount of cosmogenic noble gases was expected to be very small, due to the small
122 size of the grains, an ultra-high-sensitivity mass spectrometer and a low-blank extraction
123 line were used for the measurements. The mass spectrometer concentrates gases into the ion
124 source by a molecular drag pump (compressor), which gives a ca. two orders of magnitude
125 higher sensitivity than the same instrument without a compressor ion source (Baur, 1999).
126 Detection limits were $\sim 4 \times 10^{-16} \text{ cm}^3 \text{ STP}$ for ^{21}Ne and $\sim 2 \times 10^{-16} \text{ cm}^3$ for ^3He , and are defined
127 as the 2σ scatters of the blank. For further details on the instrument, analytical procedures
128 and calculations, see Heck et al. (2004, 2008) and Meier et al. (2010).

129 The selection of the 13 chromite grains discussed above was based on semi-
130 quantitative element analyses of unpolished, whole grains, using an energy-dispersive
131 spectrometer (Inca X-sight from Oxford instruments) with a Si detector, mounted on a
132 Hitachi S-3400 scanning electron microscope at Lund University. Cobalt was used for
133 standard, see Alwmark and Schmitz (2009a) for further details. The analyses are mostly of

134 sufficient quality to determine if a grain has an ordinary chondritic composition. Grains
135 were discarded if there was any doubt of a chondritic origin.

136 Whole-rock samples (ca. 40-70 mg) of fossil L chondrites from mid-Ordovician
137 marine limestone in Sweden (Schmitz et al., 2001) were analysed for PGEs at Woods Hole
138 Oceanographic Institution by isotope dilution with ICP-MS after NiS fire assay
139 preconcentration according to methods described in Hassler et al. (2000) and Peucker-
140 Ehrenbrink et al. (2003). The following meteorites were analysed: Österplana 003, 008,
141 009, 019, 027, 032, 035 and Gullhögen 001 (Connolly et al., 2007). Concentrations were
142 calculated using one (Ir), two (Ru, Pd, Pt), or three (Os) isotope ratios, and concentrations
143 based on multiple ratios typically agree to better than 10% (Ru) and 5% (Pd, Pt). External
144 reproducibility of PGE data was investigated by multiple analyses of standard reference
145 materials (SRM) with certified PGE concentrations (Pd, Pt). The most homogenous of the
146 SRM we tested with moderate low PGE concentrations is the Trembley Lake Diabase
147 (TDB-1). The average values and reproducibility (95% confidence interval) of this
148 standard is 122 ± 5 pg Os/g, 78 ± 5 pg Ir/g, 4.4 ± 0.2 ng Pt/g, and 24.8 ± 1.0 ng Pd/g (see
149 Peucker-Ehrenbrink et al., 2003 for details). Certified or provisional values for this SRM
150 are 5.8 ng Pt/g, 22.4 ng Pd/g (both certified), 150 pg Ir/g (provisional). Total procedural
151 blanks contribute less than 1% of the total analyte.

152

153 **3. Chemical and isotopic signatures**

154 *3.1 Loftarstone chromite grains*

155 The oxygen isotope results for chromite grains from the Loftarstone are summarized in
156 Table 1, and compared with oxygen isotope results for recently fallen H, L and LL
157 meteorites in Fig. 1. It is clear that the Loftarstone grains show typical L or possibly LL
158 composition. The grains also show the same oxygen isotopic composition as chromite
159 grains from fossil Österplana meteorites (Heck et al., 2010). The $\Delta^{17}\text{O}$ SIMS data for the
160 Loftarstone chromite show little variability (0.11‰, 2SD; n=10) and average at 1.15
161 ± 0.08 ‰ (weighted average based on individual 2SD errors). This is consistent with $\Delta^{17}\text{O}$ of
162 chromite from modern Ergheo L5 fall (1.09 ± 0.07 ‰, weighted average as above, Heck et
163 al., 2010), and with the $\Delta^{17}\text{O}$ group average of modern L or LL chondrites (1.07 ± 0.18 ‰,
164 2SD; n=26; 1.26 ± 0.24 ‰, 2SD, respectively) (Clayton et al., 1991). Oxygen isotope ratios
165 of the Loftarstone chromite grains lie on a mass-dependent fractionation line within the
166 analytical uncertainty, while the weighted average is similar to that of Ergheo L5 (Fig. 1).

167 The $\Delta^{17}\text{O}$ value of the Loftarstone grains ($1.15 \pm 0.08\text{‰}$) is clearly distinct from $\Delta^{17}\text{O}$ of
168 NMI meteorites ($-0.48 \pm 0.20\text{‰}$; 2SD; n=23), including IAB and III CD irons, analyzed by
169 Clayton and Mayeda (1996). Oxygen and nitrogen isotopes of silicate inclusions of some
170 magmatic iron meteorites of type IIE are consistent with H chondrite data (Clayton et al.,
171 1983; Mathew et al., 2000). However, we can clearly distinguish H chondrites from L and
172 LL chondrites, and can exclude from that observation alone a IIE origin of the Loftarstone
173 chromites. There is not even a remote possibility that the abundant chromite in the
174 Loftarstone could have an NMI origin.

175 The elemental concentrations determined with EPMA for the grains analysed for O
176 isotopes are presented in Table 2. The results are very similar to results for the 73
177 extraterrestrial chromite grains recovered from the Loftarstone by Alwmark and Schmitz
178 (2007), and chromite grains from 26 mid-Ordovician fossil meteorites analysed by Schmitz
179 et al. (2001). The data in tables 1 and 2 allow a detailed grain-by-grain comparison of
180 oxygen isotopic composition and elemental composition. No correlation of $\delta^{18}\text{O}$, $\delta^{17}\text{O}$, or
181 $\Delta^{17}\text{O}$ with any measured element concentration is observed.

182 The noble gas results for six chromite grains from the Loftarstone are shown in Table
183 3, and compared with similar data for extraterrestrial and terrestrial chrome spinels from
184 meteorite-rich mid-Ordovician limestone. The Loftarstone grains show no or very low
185 concentrations of ^3He and $^{20,21,22}\text{Ne}$, i.e. the grains have acquired no or insignificant
186 amounts of cosmogenic, solar or nucleogenic noble gases. This is in clear contrast to other
187 extraterrestrial chromite grains recovered from mid-Ordovician sediments. Sediment
188 dispersed chromite grains representing parts of micrometeorites typically contain 3-4 orders
189 of magnitude higher concentrations of solar noble gases (Heck et al., 2004; 2008; Meier et
190 al., 2010). Chromite from fossil meteorites instead contain significant concentrations of
191 cosmic-ray induced neon. Terrestrial chrome spinel grains from mid-Ordovician sediments,
192 on the other hand, show a similar absence of ^3He or $^{20,21,22}\text{Ne}$ as the Loftarstone grains
193 (Meier et al., 2010). The semi-quantitative analyses by SEM-EDS of the six chromite
194 grains used for the destructive noble gas analyses are presented in Table 4. Despite the
195 lower quality of the analyses it is obvious, based on previous experience, that the grains
196 originate from ordinary chondrites. Over the years our group has screened in excess of a
197 thousand unpolished chromite grains in order to identify chondritic grains by means of
198 semi-quantitative SEM-EDS analyses. Subsequent high-quality analyses of the grains after

199 polishing have always confirmed the preliminary assessment (e.g., Schmitz et al., 2001;
200 Cronholm and Schmitz, 2010).

201

202 *3.2 PGEs in fossil meteorites and Loftarstone*

203 The results of PGE analyses of the nine small whole-rock pieces from eight fossil L
204 chondrites are shown in Table 5 and compared with PGE data from the literature for
205 recently fallen, fresh L chondrites. The L-chondrite classification of the fossil meteorites is
206 based on studies of the element and oxygen isotopic composition of chromite (Schmitz et
207 al., 2001; Heck et al., 2010), the composition of olivine and pyroxene inclusions in the
208 chromite (Alwmark and Schmitz, 2009b), and the average size distribution of relict
209 chondrule textures (Bridges et al., 2007). Most of the fossil meteorite samples show similar
210 high, or even higher, PGE concentrations as recent meteorites, e.g. 450-940 ng/g Ir in the
211 fossil meteorites compared to 385-490 ng/g Ir in recent L chondrites. The fossil meteorites
212 are almost completely pseudomorphed by secondary minerals, and chromite is the only
213 relict component known (Nyström and Wickman, 1991; Schmitz et al., 2001). Some of the
214 meteorite pieces have Ir concentrations as low as 60-200 ng/g, i.e. they have lost 50 to 90
215 percent of their original inventory. Other pieces show up to a factor two higher PGE
216 concentrations than recent meteorites. These data suggest considerable redistribution of
217 PGEs.

218 In Fig. 2 we show that the fossil L-chondrite samples give highly variable PGE
219 patterns, with no or only little resemblance to the patterns of recent unweathered L
220 chondrites. As a central argument for an NMI impactor of the Lockne crater, Tagle et al.
221 (2008) use the Ru/Ir ratio measured in the Loftarstone. They show that recent L chondrites
222 have ratios in the range between 1.42 and 1.62, and that the corresponding ratio, based on
223 regression slopes, in the Loftarstone lies at 2.00 ± 0.11 , outside the L-chondritic range but in
224 accord with an NMI impactor. The Ru/Ir ratios of the fossil L chondrites in Fig. 2 and
225 Table 5 vary between 1.00 and 3.83, however, none of the nine ratios fall in the range for
226 recent L chondrites. It would not have been possible to establish the L-chondritic origin of
227 the fossil meteorites from the individual PGE patterns, despite the fact that much of the
228 original PGEs are preserved. We attempted also regression analyses of the data (e.g., Ru/Ir,
229 Pt/Ir, Pd/Ir, Pt/Pd, Pt/Ru, Pd/Ru), but the scatter in the data is too large to give any
230 significant information about precursor meteorite type from the regression slopes.
231 Apparently diagenesis has dramatically altered the original ratios for each individual
232 sample. However, despite the large spread in platinum group inter-elemental ratios, the

233 average PGE content of the nine samples is close to identical to that from recent, fresh L
234 chondrites (Table 5; Fig. 2). The average Ru/Ir value is 1.57, i.e. perfectly L-chondritic.

235

236 **4. Discussion**

237 *4.1 Chromite for projectile identification*

238 Only for very few craters larger than 1.5 km has the impactor type been established, mainly
239 because physical pieces of the impactors generally are missing (see, Alwmark and Schmitz,
240 2007; Tagle et al., 2008; Tagle and Claeys, 2005). The extremely abundant chromite grains
241 with L-chondritic composition in the Loftarstone were interpreted by Alwmark and
242 Schmitz (2007) as representing relict physical fragments of an impactor in relatively deep
243 water (ca. 500 m). Using a grossly exaggerated scale when presenting the chromite major
244 element data of Alwmark and Schmitz (2007), Tagle et al. (2008) highlighted some minor
245 deviations for some grains relative to the "ideal" L-chondritic composition. Based on this
246 they questioned the L-chondritic origin of the grains, but suggested no other likely origin.
247 Alwmark and Schmitz (2007) argue that the chromite in the Loftarstone sometimes shows
248 minor deviations from a typical L-chondritic composition, related to hydrothermal
249 alteration of the grains after the impact, but on the whole the majority of the grains still
250 show a clear L-chondritic elemental signature, as now also confirmed by O-isotopic
251 analyses. The NMI meteorites do not contain common chromites of the type found in the
252 Loftarstone (Bunch et al., 1970). It is noteworthy that the Loftarstone is the sediment with
253 the highest content of extraterrestrial chromite ever observed. It contains on the order of
254 one extraterrestrial chromite grain per 0.01 kg rock, in contrast to one grain per 100 kg for
255 slowly formed (ca. 2 mm kyr⁻¹) limestone from periods in Earth's history not influenced by
256 the excess flux of meteoritic matter immediately following the break-up of the L-chondrite
257 parent body (Schmitz and Häggström, 2006; Cronholm and Schmitz, 2007, 2010). In
258 condensed sediments formed within the first few million years after the break-up event, the
259 concentrations are 1-3 grains per kg (Schmitz and Häggström, 2006). The absence of
260 cosmogenic ³He and ^{20,21,22}Ne in the chromite grains from the Loftarstone indicates that
261 they have not been exposed to galactic cosmic rays. This is to be expected for
262 extraterrestrial material that has been transported to Earth in a large body (asteroid), as the
263 penetration depth of galactic cosmic rays is on the order of 1-2 m. For the Lockne impactor,
264 which had an approximate size of ~600 m in diameter (Ormö and Lindström, 2000), this
265 means that only ca. 2% of the chromite grains could have been exposed to galactic cosmic
266 rays as they would have been situated in the outermost two meters of the body. The results

267 also indicate that the chromite grains are not reworked grains derived from the enhanced
268 rain of micrometeorites in the mid-Ordovician. Such grains almost always contain solar
269 wind Ne (Heck et al., 2008; Meier et al., 2010). In this alternative explanation,
270 micrometeoritic chromite grains dispersed in the target rock would have lost their solar
271 wind Ne either during impact, because of elevated temperature and pressure, or due to
272 hydrothermal activity in the crater after the impact. However, while the solar wind Ne,
273 which is only implanted a few nm into the surface of a grain, might be easily lost that way,
274 complete loss of noble gases, including cosmogenic Ne, requires more or less total melting
275 of the grains. Most grains show a perfect, unaltered L-chondritic major element and oxygen
276 isotopic composition, disproving melting and pointing towards an initial deficiency in such
277 gases. Degassing of the Loftarstone chromite grains on million year time scales after the
278 impact event is neither a likely scenario, since the long-term burial heating history in the
279 Lockne region was not very different from that at other sites in Sweden, e.g. Kinnekulle,
280 where extraterrestrial chromite grains in sediments have retained all or most of their solar-
281 wind and cosmogenic gases until today (Heck et al., 2008; Meier et al., 2010).

282 The extraterrestrial chromite grains in the Loftarstone most likely represent the relict
283 residues of weathered, small pieces of the impactor. When an asteroid impacts in deep
284 water such pieces may escape vaporization as shown by the abundant unmelted meteorite
285 fragments found in sediments from the late Pliocene Eltanin impact site in the Southern
286 Ocean (Kyte, 2002). Based on robust biostratigraphy we know that the Lockne crater
287 formed ca. 10 million years after the first micrometeorites from the break-up of the L-
288 chondrite parent body showered the Earth (Alwmark and Schmitz, 2009a). This time lag is
289 in agreement with modeling simulations of large break-up events showing that the larger,
290 km-sized bodies typically tend to reach Earth on the order of 1-30 million years later than
291 the dust particles (e.g. Zappalà et al., 1998; Dermott et al., 2002). Poynting-Robertson light
292 drag is important in transferring particles <500 μm directly from the asteroid belt to the
293 inner solar system, whereas kilometer-sized objects are ejected to the inner solar system
294 first after having drifted into orbital resonance positions.

295

296 *4.2 PGEs and Cr for projectile identification*

297 Several studies have previously pointed out the general difficulties of using PGE
298 patterns to determine the type of impactor. Farley (2009) evaluated in detail the claims by
299 Tagle and Claeys (2005) of an L-chondritic impactor for the Popigai crater based on PGE
300 patterns in impact glasses, and showed that the impactor signature is highly sensitive to the

301 assumptions and methods used in the regression. By recomputing the Ru/Ir and Ru/Rh, for
302 example, simply by switching x and y axes, or by including single data points omitted for
303 unspecified reasons by Tagle and Claeys (2005), Farley (2009) showed that no robust
304 conclusion about projectile type for the Popigai crater can be obtained from the PGEs.
305 Regression slopes may be highly dependent on single outlier data points whereas an
306 average value, like for the PGEs of our fossil meteorite, weights each data point to the same
307 extent. As shown here, the average PGE ratios of the fossil meteorite samples give a clear
308 chondritic pattern, whereas regression slopes following the approach of Tagle et al. (2008)
309 are not usable for determining the origin. This insight led us to calculate the average Ru/Ir
310 value for the 17 samples from the Loftarstone analysed by Tagle et al. (2008; their figure
311 3). This gives a Ru/Ir ratio of 1.76, rather than the 2.00 ± 0.11 calculated by Tagle et al.
312 (2008) from regression slopes. The average Ru/Ir value of the Loftarstone thus lies just
313 slightly outside the L-chondritic range, 1.42-1.62, whereas the regression slope for the same
314 data set instead indicates an NMI impactor. We argue that the different results from the two
315 approaches shows, analogous to the Popigai case, that the conclusion of Tagle et al. (2008)
316 about an NMI impactor based on the Loftarstone Ru/Ir regression slopes can not be
317 considered robust. It should be noted that the method of averaging PGEs rather than using
318 slopes will not work in samples that have a large fraction of target PGEs.

319 The Cr/Ir ratios of Loftarstone bulk sediment samples were plotted by Tagle et al.
320 (2008), and mixing lines assuming two-components mixing between target rock and
321 impactor were compared to chondrite-target rock mixing lines (Fig. 3). It was argued that
322 the Loftarstone Cr/Ir regression slope indicates a ratio of 13.1 compared with 7.8 for L-
323 chondrites, a fact used to argue for an NMI impactor. In Table 6 and Fig. 3 we present a
324 compilation from the literature of the Cr/Ir ratios of nine samples from six fossil L
325 chondrites (Schmitz et al., 1996, 1997). It is clear from Fig. 3 that because of element
326 redistribution it is not possible to establish the L-chondritic origin of the fossil meteorites
327 from the Cr/Ir ratios, despite the fact that a major fraction of the Ir and Cr of the meteorites
328 is preserved. The Cr/Ir ratios ($\times 10^3$) of the fossil meteorite samples are highly variable, in
329 the range 5.50 to 37.7 compared to 7.92 for recent L chondrites (Table 6). In a transect of
330 three samples from the central to the outer part of fossil meteorite Österplana 009 there is a
331 gradual shift in Cr/Ir from 5.50 to 11.3 to 32.3, clearly indicating a higher mobility of Ir
332 than Cr during diagenesis. This is further indicated by the decrease in Cr/Ir ratio to 15.4 in
333 the limestone just outside the meteorite, where Ir mobilized from the fossil meteorite was
334 redeposited. We argue that in an open sediment such as the initially porous Loftarstone the

335 observation by Tagle et al. (2008) of slightly higher Cr/Ir ratios than in L chondrites most
336 likely reflects the higher mobility of Ir during early diagenesis, rather than a non-chondritic
337 impactor.

338 Element mobility during early diagenesis is probably the greatest problem when
339 attempting to use PGE patterns from sedimentary ejecta deposits, however, fractionation in
340 the hot impact plume and during subsequent condensation may also obscure original
341 patterns (e.g., Evans et al., 1993). Post-depositional mobility of PGEs in sediments has
342 been described in several papers, e.g. Colodner et al. (1992) showed that Pt, Re and Ir in
343 abyssal sediments are redistributed by changes in sedimentary redox conditions. Wallace et
344 al. (1990) showed that the PGEs of the Late Proterozoic Acraman impact ejecta similarly
345 were highly mobile and affected by redox chemistry. Evans et al. (1993) compiled PGE
346 data for a large number of continental and marine Cretaceous-Tertiary boundary clays, and
347 concluded that the boundary has a chondritic PGE signature, but only when the integrated
348 values on a global scale are considered. For each individual site non-chondritic values were
349 the rule rather than an exception. This conclusion is similar to that obtained here for the
350 PGEs of the fossil meteorites, i.e. only the integrated value for all samples gives a clear
351 chondritic signature. This also lends support to the significance of the near-chondritic
352 average Ru/Ir ratio of the Loftarstone.

353

354 **5. Conclusions**

355 High-precision oxygen isotope SIMS analyses confirm that the abundant extraterrestrial
356 chromite grains in the Loftarstone are L (or LL) chondritic (see, Alwmark and Schmitz,
357 2007). The isotopic results are clearly incompatible with meteorites of the NMI type.
358 Analyses of Cr and PGEs of fossil L-chondritic meteorites show that meteoritic elemental
359 ratios in ancient sedimentary environments are significantly affected by element mobility,
360 and only integration of a large data set can give clues about original ratios. The approach is
361 very sensitive to how the integration is made. We find no robust support for the claim by
362 Tagle et al. (2008) that the Lockne crater was caused by an NMI impactor. The Lockne
363 crater is likely related to the L-chondrite parent body break-up at 470 Ma.

364

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484

485 **Figure captions**

486 **Fig. 1:** Oxygen-three-isotope diagram. Loftarstone chromite individual analyses (open
487 circles) are shown with 2 SD error bars. The weighted Loftarstone average (solid circle) is
488 shown with its weighted average error based on individual 2 SD errors. Gol 001 is a bulk
489 analysis of ca. 100 chromite grains shown with 2 SD error bars from the fossil meteorite
490 Österplana 029 (Gol 001) reported by (Greenwood et al., 2007). Chromite data from recent
491 ordinary chondrites (triangle and box symbols) are weighted averages of SIMS data from
492 Heck et al. (2010). Mass-dependent fractionation lines are shown for terrestrial samples
493 (TFL; dashed line), for average compositions of group H, L and LL bulk ordinary
494 chondrites (Clayton et al., 1991) and NMI meteorites (Clayton and Mayeda, 1996) (solid
495 lines) and their standard deviations (shaded boxes).

496

497 **Fig. 2:** L-chondrite normalized PGE patterns for nine samples from fossil L-chondritic
498 meteorites Öst 003, 008, 009, 019, 027, 032, 035 and Gul 001. Recent L-chondrite data
499 from Wasson and Kallemeyn (1988) and Tagle and Claeys (2005).

500

501 **Fig. 3:** Ir and Cr content of nine samples from fossil L-chondritic meteorites Öpl 001, 007,
502 009, 011, 030, and 036, compared with mixing lines by Tagle et al. (2008) for assumed two
503 component mix between target rock and impactor for the Loftarstone based on two data

504 sets. Included in the figure are also expected mixing lines with different chondritic
505 impactors, following Tagle et al. (2008) and Wasson and Kallemeyn (1988). Slopes for the
506 Loftarstone mixing lines were based on Ir contents <4.5 ppb, whereas fossil meteorites
507 contain 100-880 ppb Ir.

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