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1 **Ultra-depleted 2.05 Ga komatiites of Finnish Lapland: Products of grainy**
2 **late accretion or core-mantle interaction?**

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41 34

Abstract

Rhenium-Os, Pt-Os, Sm-Nd, Lu-Hf, and Hf-W isotope data, together with lithophile trace element and highly siderophile element (HSE: Re, Os, Ir, Ru, Pt, and Pd) abundances, are reported for 2.05 Ga Jeesiörova and Kevitsa komatiites from the Central Lapland Greenstone Belt, Fennoscandia, Finland. Both komatiites are closely genetically related, with the Kevitsa dikes having served as feeding magma conduits to the Jeesiörova pillowed and massive lavas. The parental komatiite magma is estimated to have contained ~25 wt.% MgO and was, thus, derived from a mantle source at least as hot as those of some of its late Archean counterparts. A suite of Jeesiörova and Kevitsa whole-rock komatiite samples and olivine and chromite separates define an internal Re-Os isochron with an age of 2049 ± 13 Ma and an initial $\gamma^{187}\text{Os} = -0.2 \pm 0.2$ (2SE), indicating long-term chondritic Re/Os in the mantle source. By contrast, Pt-Os data for a set of Jeesiörova chromite separates define an average initial $\mu^{186}\text{Os} = +29 \pm 2$ (2SE), indicating a long-term history of suprachondritic Pt/Os in the mantle source. The absolute HSE abundances in the mantle source of the Jeesiörova-Kevitsa komatiite system are estimated to have been $120 \pm 5\%$ of the present-day Bulk Silicate Earth (BSE). This is the first komatiite system for which excess HSE in the mantle source, relative to modern BSE, has been documented.

The ^{147}Sm - ^{143}Nd and ^{176}Lu - ^{176}Hf data yield isochron ages and initial ratios of, respectively, 2046 ± 22 Ma with $\epsilon^{143}\text{Nd} = +3.7 \pm 0.3$, and 2072 ± 20 Ma with $\epsilon^{176}\text{Hf} = +8.7 \pm 0.4$ (2SE), indicating a long-term history of depletions of Nd relative to Sm, and Hf relative to Lu. The measured $\mu^{182}\text{W} = +1.5 \pm 3.3$ is indistinguishable from the modern mantle value. Despite being strongly depleted in highly incompatible lithophile trace elements, the Th-Nb-La systematics of the komatiites indicate ~1% crustal contamination of the original komatiite magma, assuming the contaminant was similar in composition to the calculated Fennoscandian Tonalite Average (FTA). This level of contamination would have also significantly modified the Nd, Hf, and W isotope compositions of the original komatiitic magma, but not the Os isotope compositions or HSE abundances. The calculated original komatiite magma, corrected for the effects of crustal contamination, would have had initial $\epsilon^{143}\text{Nd} \sim +4.9$, $\epsilon^{176}\text{Hf} \sim +10.2$, and $\mu^{182}\text{W} \sim -10$.

Our modeling indicates that the initial $^{186,187}\text{Os}/^{188}\text{Os}$ isotopic compositions and suprachondritic HSE abundances, coupled with the projected negative $\mu^{182}\text{W}$, are best explained by either (1) derivation from a mantle domain characterized by an excess of late accreted, differentiated planetesimal core metal, i.e., “grainy” late accretion, or (2) addition of chemically fractionated terrestrial core metal to the mantle source domain of the komatiites. The presence of these characteristics in the Jeesiörova-Kevitsa komatiite mantle source provides further evidence for the early creation and long-term survival of chemically diverse domains within the mantle.

1. Introduction

The chemically and isotopically heterogeneous nature of the mantle has long been established, largely through the study of modern rocks (e.g., Gast et al., 1964; Hart and Brooks, 1977; Zindler et al., 1982; White and Hofmann, 1982; White and Patchett, 1984; Hofmann, 1984, 1997; Jacobsen, 1988; Galer and Goldstein, 1991). However, the origin of the heterogeneities for different elements and isotopic systems (e.g., $^{146,147}\text{Sm}$ - $^{142,143}\text{Nd}$, ^{176}Lu - ^{176}Hf , ^{182}Hf - ^{182}W), as well as their length scales and residence times, remain topics of debate. Some of the heterogeneities have been argued to be primordial, reflecting initial planetary accretion/differentiation, including magma ocean crystallization processes (e.g., Goldstein and Galer, 1992; Albarède et al., 2000; Boyet and Carlson, 2005; Touboul et al., 2012; Jacobsen and Yu, 2015; Puchtel et al., 2013; 2016a; Rizo et al., 2016b; Mundl et al., 2017; Mundl-Petermeier et al., 2019; Tusch et al., 2019), while others may have originated as a result of a protracted accretion history (Willbold et al., 2011; Kruijer et al., 2015; Touboul et al., 2015; Puchtel et al., 2018; Archer et al., 2019).

Osmium isotope and highly siderophile element (HSE, including Re, Os, Ir, Ru, Pt, and Pd) abundance systematics of the mantle through time provide information about certain aspects of early Earth processes. For example, the observation that the HSE occur in approximately chondritic relative proportions in the Bulk Silicate Earth (BSE), and that absolute abundances of at least some of the HSE are higher than would be expected from metal-silicate equilibration during core formation, have led to the concept of late accretion, which purports that 0.5-1% of Earth's mass was accreted to the mantle subsequent to cessation of core formation (Kimura et al., 1974; Chou et al., 1983; Morgan, 1985, 1986). Issues related to the nature of late accretion include the composition of the late accreted materials and the time frame within which they were delivered to Earth and homogenized within the mantle (e.g., Bennett et al., 2002; Maier et al., 2009; Coggon et al., 2013; Walker et al., 2015; van de Löcht et al., 2018).

The Os isotopic composition and absolute HSE abundances of portions of the Archean mantle have been the subject of considerable prior study (e.g., Brüggmann et al., 1987; Walker et al., 1988; Foster et al., 1996; Gangopadhyay and Walker, 2003; Puchtel et al., 2014, 2016a,b; 2018; Maier et al., 2009). These studies have mostly focused on komatiites because komatiites are valuable probes of absolute HSE abundances and Os isotopic compositions in some portions of the mantle. This stems from the fact that most komatiites were formed by sufficiently high degrees of partial melting that the dominant sulfide hosts of HSE in the mantle were dissolved into the melts. Under such conditions, the HSE abundances in

107 komatiitic melts are much more similar to the initial mantle source compositions than those in
108 basaltic melts. These high degrees of partial melting also led to comparatively high Os
109 abundances in komatiitic melts, which make their Os isotopic compositions insensitive to
110 crustal contamination. The high Os concentrations in komatiites, coupled with typically low
111 Re/Os ratios, mean that they can provide precise initial Os isotopic compositions of mantle
112 sources.

113 Compared to Archean komatiites, HSE and Os isotopes have been much less extensively
114 studied in their Proterozoic counterparts, primarily due to the paucity of Proterozoic
115 komatiites in the rock record. Here, we report combined ^{147}Sm - ^{143}Nd , ^{176}Lu - ^{176}Hf , ^{187}Re -
116 ^{187}Os , ^{190}Pt - ^{186}Os , ^{182}Hf - ^{182}W , and HSE and lithophile trace element abundance data for the
117 Proterozoic Kevitsa komatiitic dikes and associated Jeesiörova komatiites, all part of the
118 Central Lapland Greenstone Belt in Fennoscandia, Finland. The Kevitsa komatiite dikes in
119 particular display strong chemical differentiation trends, excellent preservation of igneous
120 minerals, including olivine, and no signs of interaction with their wall rocks (Mutanen, 1997;
121 Huhma et al., 2018). They hence provide ample opportunity for dating and obtaining
122 information on the composition of the original komatiitic magmas. We use these data to (i)
123 constrain the long-term evolution of the mantle domain from which the Fennoscandian Shield
124 was produced and which also gave rise to the komatiite parental magmas, and (ii) further
125 evaluate the extent of Archean-Paleoproterozoic mantle heterogeneity.

2. Geological background

126 Paleoproterozoic komatiites of Finnish Lapland are part of the Central Lapland
127 Greenstone Belt (**Fig. 1**), which consists of a thick supracrustal rock series spanning 570 Ma
128 (Hanski and Huhma, 2005). The komatiites are known to occur at two stratigraphic levels
129 and, as such, have been referred to as the lower and upper komatiites (Hanski et al., 2001).
130 The lower komatiites are found near the bottom of the succession and are assigned to the
131 Kuusamo Group. They were emplaced subaerially and, locally, lie directly on Archean
132 granitoid basement. A type occurrence has been described from the Möykkelmä area, where
133 komatiites form part of a 250-m-thick komatiite-tholeiite sequence with a strong upper crustal
134 signature in their chemical composition (Hanski and Huhma, 2005). Stratigraphically, the
135 lower komatiites can be correlated with the komatiitic lavas of the Vetreny Belt in the SW
136 part of the Fennoscandian Shield, which have an internal Re-Os isochron age of 2407 ± 6 Ma
137 (Puchtel et al., 2016b).

139 The upper komatiites are associated with phyllites and black schists of the Savukoski
140 Group whose sedimentary precursors were deposited on cratonic siliciclastic sediments of the
141 Sodankylä Group (Hanski and Huhma, 2005). The upper komatiitic volcanic rocks extend
142 from Finnish Lapland to northern Norway over a distance of ~400 km and are associated with
143 Ti-rich picrites and basalts (Hanski et al., 2001). Unlike typical Archean komatiites, the upper
144 komatiites are characterized by the abundance of volcanoclastic eruptions (Saverikko, 1985;
145 Barnes and Often, 1990). They also occur as massive lava flows, pillow lavas, tuffs, and rare
146 layered flows. The volcanic structures and types of the associated sedimentary rocks attest to
147 submarine conditions of eruption in a deepened sedimentary basin. Although belonging to the
148 Al-undepleted, or Munro-type of lavas of Nesbitt et al. (1979) typical of late Archean
149 komatiite sequences, the upper komatiites are further distinct from Archean komatiites in
150 having elevated TiO₂ contents and, therefore, they were classified as a Ti-enriched komatiite
151 type (Hanski et al., 2001).

152 The upper komatiites are best preserved in the Jeesiörova area studied by Hanski et al.
153 (2001). Magmatic olivine has not been reported to have been preserved in komatiitic lavas
154 anywhere in Finnish Lapland, but magmatic clinopyroxene is well preserved at Jeesiörova,
155 occurring as poikilitic intergrowths in olivine cumulates, as prismatic grains in gabbroic
156 rocks, and as needle-like crystals in pillowed and massive lavas. The presence of magmatic
157 clinopyroxene in the lavas allowed Hanski et al. (2001) to obtain a weighted average Sm-Nd
158 isochron age of 2056±25 Ma for clinopyroxene-whole-rock pairs for the Jeesiörova
159 komatiites. The initial $\epsilon^{143}\text{Nd}$ values range from +2 to +4 for individual samples, with the
160 most LREE-depleted samples having the highest initial $\epsilon^{143}\text{Nd}$ values. This inverse correlation
161 between the La/Sm and $\epsilon^{143}\text{Nd}$ values was interpreted to be the result of contamination by
162 material of the upper crustal rocks through which the Jeesiörova komatiite magma ascended
163 (Hanski et al., 2001).

164 Gangopadhyay et al. (2006) analyzed a set of whole-rock Jeesiörova komatiite samples
165 and chromite separates for Re-Os isotope systematics. Although the Re-Os system in the
166 whole-rock samples was found to be disturbed by post-magmatic processes, by analyzing
167 chromite separates, the authors constrained the initial $^{187}\text{Os}/^{188}\text{Os}$ ratio for the komatiite lavas
168 to be chondritic ($\gamma^{187}\text{Os}(\text{T}) = +0.1 \pm 0.6$).

169 The Kevitsa komatiitic dikes were discovered by Mutanen (1997) during an exploration
170 drilling campaign of the Kevitsa Ni-Cu sulfide deposit. The dikes were penetrated by at least
171 a dozen drill holes in an area of ~1 km², covering the ore-bearing part of the Kevitsa mafic-

172 ultramafic intrusion (**Fig. 1c**). This intrusion is located approximately 50 km east of the
173 Jeesiörova area within pelitic metasediments of the Savukoski Group (**Fig. 1a**) and has a U-
174 Pb zircon age of 2058 ± 4 Ma (Mutanen and Huhma, 2001), which is identical, within
175 uncertainty, to the Sm-Nd age of the Jeesiörova komatiites. Each of the drill cores contains
176 from one to four dike intersections at depths varying from 12 to 870 m beneath the surface.
177 Although the true thickness of the dikes is less than 7 m, the lengths of the intersections reach
178 up to 16 m because the drill holes cut through the dikes at sharp angles ($\leq 30^\circ$).

179 The dikes are generally subvertical, with the dip varying between 65° and 90° , and strike
180 approximately ENE (Mutanen, 2005). They are characterized by strong internal
181 differentiation related to olivine fractionation, having an olivine-enriched central part and
182 finer-grained chilled zones near contacts with the wall rocks, which are composed of olivine-
183 pyroxene cumulates of the Kevitsa intrusion. The differentiation is rather symmetrical across
184 the dikes, suggesting that originally they were also steeply dipping. The dikes are chemically
185 similar to the Jeesiörova komatiites and have been interpreted to represent their feeder magma
186 conduits (Huhma et al., 2018); the absolute age of the dikes, however, has so far remained
187 unknown.

3. Samples

189 Sampling of the Kevitsa komatiite dikes was done at the National Drill Core Depot of the
190 Geological Survey of Finland, Loppi, Finland. Samples for chemical and isotopic analyses
191 were selected from drill core DDH 814, which contains two dike intersections at depths of
192 152.10–159.55 and 175.00–188.50 m, with the corresponding estimated thicknesses of the
193 dikes being ~1.3 and ~6.8 m.

194 For this study, we also prepared powders from hand specimens of the best preserved and
195 most primitive komatiite lava samples 12D-PPR, 13-EJH, and 17.1-PPR from the Jeesiörova
196 area (Hanski et al., 2001). In addition, we powdered sample LP-10 that was recently collected
197 from an olivine-phyric massive komatiite lava flow from the same area.

4. Analytical techniques

199 Details of the majority of the analytical techniques used in this study have been reported in a number of
200 previous publications (e.g., Puchtel et al., 2016a,b; 2018); these, therefore, are only briefly summarized here. In
201 contrast, full details are provided for those techniques implemented here for the first time.

4.1. Analysis of mineral compositions

204 Mineral compositions were determined using a *JEOL JXA-8530F Plus Hyper Probe* instrument at the
1 205 Center of Material Analysis, University of Oulu. The analytical conditions were an accelerating voltage of 15 kV
2 206 and a beam current of 15 nA. Peak and background counting times were 10 and 5 s, respectively. The standard
3 207 built-in ZAF correction routine was used.
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6 208 **4.2. Re-Os isotopic compositions and HSE abundances**

7 209 The measurements of Ru, Pd, Re, Ir, and Pt were performed at the *Plasma Laboratory (PL)* on Faraday cups
8 210 of a *ThermoFisher Neptune Plus* ICP-MS in static mode using 10^{13} Ohm amplifiers. Isotopic mass fractionation
9 211 was monitored and corrected for by interspersing samples and standards. The external precision of the analyses
10 212 was estimated, on the basis of standard measurements performed during the period of the analytical campaign, to
11 213 be $^{185}\text{Re}/^{187}\text{Re} = 0.25\%$, $^{99}\text{Ru}/^{101}\text{Ru} = 0.26\%$, $^{191}\text{Ir}/^{193}\text{Ir} = 0.15\%$, $^{194}\text{Pt}/^{196}\text{Pt} = 0.10\%$, and $^{105}\text{Pd}/^{106}\text{Pd} = 0.08\%$
12 214 relative (2SD). The accuracy of the data was assessed by comparing the results for the reference materials IAG
13 215 MUH-1 (Austrian harzburgite), IAG OKUM (ultramafic komatiite), and NRC TDB-1 (Diabase PGE Rock
14 216 Material) obtained at the *Isotope Geochemistry Laboratory (IGL)* with the reference values. Concentrations of
15 217 all HSE and Os isotopic compositions obtained at the *IGL* are within the uncertainties of the certified reference
16 218 values (**Supplementary Table A1**).
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18 219 The average total analytical blank (TAB) measured during the present analytical campaign was (in pg): Ru
19 220 6.0, Pd 17, Re 0.53, Os 0.42, Ir 1.2, and Pt 174 ($N = 9$). For the whole-rock komatiite samples, the average TAB
20 221 constituted less than 0.1% for Os, Re, Ir, Ru, and Pd, and less than 1% for Pt of the total amount of element
21 222 analyzed. For the olivine and chromite separates, the TAB for Os and Ir constituted less than 0.1%, for Ru less
22 223 than 0.4%, for Re between 0.1 and 3%, for Pt between 14 and 19%, and for Pd less than 1% of the total amount
23 224 of element analyzed. We, therefore, cite $\pm 0.1\%$ as the uncertainty on the concentrations of Os, $\pm 0.2\%$ for Ir,
24 225 between 0.3 and 0.4% for Ru, between 0.1 and 1% for Pd, between 0.1 and 3% for Re, and between 1.0 and 19%
25 226 for Pt, of the total amount of element analyzed. The uncertainty on the Re/Os ratio for each sample was
26 227 calculated by multiplying the estimated uncertainties on the Re and Os abundances. These uncertainties vary
27 228 between 0.30 and 3.0% relative.
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30 229 **4.3. Pt-Os isotopic data.**

31 230 In order to obtain the amount of Os required for high-precision measurements of $^{186}\text{Os}/^{188}\text{Os}$ and $^{187}\text{Os}/^{188}\text{Os}$
32 231 (~ 100 ng), ~ 1.0 g of pure chromite separates from Jeesiörova komatiite samples 12D-PPR and 13-EJH,
33 232 representing different fractions in terms of magnetic susceptibility and density, were digested in Carius tubes.
34 233 For the initial unspiked digestions, six ~ 1 g fractions of chromite separate 12D-PPR and two fractions of 13-EJH
35 234 were digested. After the digestion was complete, the tubes were chilled, opened, and $\sim 2\%$ of the acid sample
36 235 solution from each CT were transferred into a 25 mL Pyrex™ CT for precise determination of Pt/Os, Re/Os, and
37 236 Ir/Os. The spiked aliquots were processed using the same procedure utilized for the Re-Os and HSE analyses.
38 237 From the remaining part of the unspiked acid sample solutions, Os was extracted and purified using the same
39 238 protocol utilized for the Re-Os work. The Os cuts from the batch of two CT containing sample 13-EJH were
40 239 combined into one cut and used for the precise measurements of $^{186}\text{Os}/^{188}\text{Os}$ and $^{187}\text{Os}/^{188}\text{Os}$ in this sample. For
41 240 sample 12D-PPR, every Os cut from six individual CT was used for the precise measurements of Os isotopic
42 241 compositions. For TAB-corrections of the data from the spiked aliquots, since the ID digestion introduced $\sim 98\%$
43 242 of the total blank, the TAB of the initial (IC) digestion was considered negligible.
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243 The high-precision measurements of the $^{186}\text{Os}/^{188}\text{Os}$ and $^{187}\text{Os}/^{188}\text{Os}$ ratios were performed by *N-TIMS* in
1 244 static mode on a *ThermoFisher Triton*[®] mass spectrometer at the *IGL*. During each run, between 1200 and 2400
2 245 ratios were collected for each sample load; the in-run uncertainties on the measured $^{186}\text{Os}/^{188}\text{Os}$ and $^{187}\text{Os}/^{188}\text{Os}$
3 246 ratios are quoted as 2SE. The possible isobaric interference of $^{186}\text{W}^{16}\text{O}_3^-$ on $^{186}\text{Os}^{16}\text{O}_3^-$ was monitored and
4 247 corrected for by measuring masses $^{184}\text{OsO}_3^-$ and $^{183}\text{W}^{16}\text{O}_3^-$ using the electron multiplier.

7 248 The mean of the Johnson-Matthey Os standard runs during the period of data collection was 0.001302 ± 2 for
8 249 $^{184}\text{Os}/^{188}\text{Os}$, 0.1198432 ± 18 (± 15 ppm) for $^{186}\text{Os}/^{188}\text{Os}$, and 0.1137950 ± 18 (16 ppm) for $^{187}\text{Os}/^{188}\text{Os}$ (2SD, $N =$
9 250 14); these long-term reproducibilities were used to assess the true uncertainty on the measured $^{186}\text{Os}/^{188}\text{Os}$ and
10 251 $^{187}\text{Os}/^{188}\text{Os}$ ratios for individual samples.

14 252 **4.4. Tungsten isotopic compositions and abundances**

16 253 The W isotope analyses were carried out at the *IGL*. For each sample, approximately 100 grams of powder
17 254 were processed to obtain the ~ 1 μg of W necessary for high-precision W isotope measurements. The sample
18 255 powders for each sample were digested in four 300 mL Savillex Teflon screw-cap vials using a 5:1 mixture of
19 256 double-distilled concentrated HF and HNO_3 on a hot plate at 150°C for one week. Tungsten was separated and
20 257 purified using the four-stage ion-exchange chromatography protocol described in Peters et al. (2019), with minor
21 258 modifications. The third stage involving a 1.5 mL anion-exchange column was repeated to improve the
22 259 separation of Ti from W, and this step significantly increased W ionization efficiency. Tungsten recovery using
23 260 this procedure was better than 90%.

27 261 Tungsten isotopic compositions were measured by *N-TIMS* on a *ThermoFisher Triton* mass spectrometer at
28 262 the *IGL* using a 2-line multi-static acquisition protocol and following a slightly modified technique described by
29 263 Archer et al. (2017). All data are reported as $\mu^{182}\text{W}$ and $\mu^{183}\text{W}$, which are the part per million (ppm) deviations
30 264 of $^{182}\text{W}/^{184}\text{W}$ and $^{183}\text{W}/^{184}\text{W}$, respectively, in a given sample from those of the in-house *Alfa Aesar* laboratory W
31 265 standard. Uncertainties on $\mu^{182}\text{W}$, based on the long-term 2SD of our *Alfa Aesar* laboratory standard, were ± 4.5
32 266 ppm. The measured $^{183}\text{W}/^{184}\text{W}$ ratios were identical within uncertainties (± 6 $\mu^{183}\text{W}$ units) to those in the average
33 267 *Alfa Aesar* standard data.

35 268 Tungsten abundances were determined by isotope dilution ICP-MS. Tungsten was purified using a
36 269 previously established anion-exchange chromatography technique (e.g., Kleine et al., 2004). Concentrations
37 270 were measured using the *Element 2* single-collector ICP-MS at the *PL*. Typical uncertainties were $\sim 5\%$ relative.

45 271 **4.5. Sm-Nd isotopic compositions and abundances**

47 272 The Sm-Nd isotopic data were collected at the *IGL*. Approximately 300 mg of whole-rock komatiite sample
48 273 powder, or between 300 and 400 mg of pure clinopyroxene separates, were digested. Measurements of the Nd
49 274 isotopic compositions were performed on Faraday cups of the *ThermoFisher Neptune Plus* ICP-MS at the *PL* in
50 275 static mode followed by off-line correction for spike contributions. For each sample, 200 to 300 ratios were
51 276 collected with 8 s integration times in blocks of 100 ratios each. For every block of data collected, the peaks
52 277 were centered, and a 30 s baseline measurement was performed for each Faraday cup/amplifier pair by beam
53 278 deflection. The effects of instrumental mass fractionation were corrected for relative to $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$
54 279 using an exponential law. A total of five to six 60 ppb AMES Nd standard solutions were run at the beginning
55 280 and end of the analytical session, with 200 to 300 ratios collected during each measurement. The in-run precision

281 of the measured $^{143}\text{Nd}/^{144}\text{Nd}$ ratio for both samples and standards was between 5 and 7 ppm (2SE). During the
1 282 course of the present analytical campaign, the external reproducibility of the 60 ppb AMES Nd standard solution
2 283 measurements for $^{143}\text{Nd}/^{144}\text{Nd}$ was ± 7.0 ppm (2SD). The average $^{143}\text{Nd}/^{144}\text{Nd}$ measured for the AMES Nd
3 284 standard during each analytical session was used to calculate the instrumental mass bias coefficient for
4 285 correction of the measured $^{143}\text{Nd}/^{144}\text{Nd}$ in the samples using the long-term average $^{143}\text{Nd}/^{144}\text{Nd}$ value measured
5 286 precisely for the AMES Nd standard on the *IGL ThermoFisher Triton* (0.512152 ± 2 ; Puchtel et al., 2018).
6 287

7 288 In order to assess the accuracy of the analyses, several separate powder aliquots of USGS GRM BIR-1 and
8 289 BCR-1 were processed and analyzed using the same analytical protocol. The measured $^{143}\text{Nd}/^{144}\text{Nd}$ ratios for
9 290 both GRM are identical, within their respective uncertainties, to the GeoRem preferred values (**Supplementary**
10 291 **Table A2**), as well as to the average high-precision value for BCR-1 obtained on the *IGL ThermoFisher Triton*
11 292 (0.512645 ± 1 ; Puchtel et al., 2018).
12 293

13 294 Measurements of the Sm isotopic compositions were also performed on Faraday cups of the *ThermoFisher*
14 295 *Neptune Plus* ICP-MS at the *PL* in static mode. For each sample, 20 ratios were collected with 8 s integration
15 296 times in blocks of 10 ratios each. The effects of instrumental mass fractionation were corrected for relative to
16 297 $^{147}\text{Sm}/^{152}\text{Sm} = 0.56081$ using an exponential law. The uncertainty on the Sm/Nd ratio in the samples analyzed
17 298 was estimated on the basis of replicate analyses of the USGS GRM BCR-1 ($^{147}\text{Sm}/^{144}\text{Nd} = 0.13941\pm 0.00014$) to
18 299 be 0.10% relative (2SD).
19 300

26 298 **4.6. Lu-Hf isotopic compositions and abundances**

27 299 The Hf isotopic compositions and Lu and Hf concentrations in the Jejsiörova komatiites were determined at
28 300 the *IGL* and the Department of Terrestrial Magnetism (*DTM*), Carnegie Institution for Science. At the *IGL*,
29 301 approximately 500 mg of the whole-rock sample powder and pure clinopyroxene separate were weighted out in
30 302 15 mL Teflon inserts of Parr bombs with 7 mL double-distilled concentrated HF, 1.5 mL HNO_3 , and an
31 303 appropriate amount of mixed ^{176}Lu - ^{178}Hf spike, sealed in stainless steel jackets and digested in an oven at 170°C
32 304 for one week. The solutions were dried down, 0.7 mL of concentrated *SeaStar* HClO_4 added to the residue, the
33 305 inserts re-sealed and kept on a hotplate at 180°C for 24 hours. The solutions were then dried down at $\sim 230^\circ\text{C}$,
34 306 and the residues converted into the chloride form using 6M HCl. This step was repeated several times until clear
35 307 solutions were obtained, and then dried down.
36 308

37 309 At *DTM*, the residue was re-dissolved in 5 ml of a 1M HCl – 0.1M HF mixture, and loaded onto a 0.6×20
38 310 cm column filled with AG50W-X8 200-400 mesh cation-exchange resin. High field strength elements, including
39 311 Hf, were eluted in the loading solution and an additional 5 mL of 1M HCl – 0.1M HF. After eluting 44 ml of
40 312 2.5M HCl, heavy REE, including Lu, were eluted in 12 mL of 4M HCl (Blichert-Toft, 2001). Hafnium was
41 313 purified on a 0.6×10 cm column of Eichrom LN resin (100-150 μm bead size) by loading the sample in 5 mL of
42 314 2.5N HCl, followed by washes of 10 mL 2.5M HCl, 10 mL 6M HCl, 4 mL H_2O , 40 mL of a mixture consisting
43 315 of 0.09M Citric acid – 0.45M HNO_3 – 1% H_2O_2 , 5 mL of the same solution without H_2O_2 , and, finally, 20 mL of
44 316 6M HCl – 0.06N HF. Hafnium was then eluted in 8 mL of 6M HCl – 0.4M HF. Lutetium was purified using
45 317 Eichrom LN resin (50-100 μm bead size) on a 0.4×7 cm column. The HREE fraction from the first column was
46 318 loaded in 2.5M HCl, followed by a wash of 30 mL 2.5M HCl to remove much of the other HREE, then Lu was
47 319 collected in 10 mL 6M HCl. The resultant Hf and Lu cuts were used for the measurements of the Hf and Lu
48 320 isotopic compositions.
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320 Measurements of the Hf isotopic compositions were performed on the Faraday cups of the *ThermoFisher*
321 *Neptune Plus* ICP-MS at the *PL* in static mode followed by off-line correction for spike contributions. For each
322 sample, 200 to 300 ratios were collected with 8 s integration time in blocks of 100 ratios each. Before each
323 sample measurement, a 60 s baseline measurement was performed for each Faraday cup/amplifier pair without
324 beam deflection, and this baseline was then automatically subtracted from the sample beam. Instrumental mass
325 fractionation was corrected for relative to $^{179}\text{Hf}/^{177}\text{Hf} = 0.7325$ using an exponential law. The Yb, Lu, Ta, and W
326 isobaric interferences were corrected for using $^{173}\text{Yb}/^{176}\text{Yb} = 1.256$, $^{175}\text{Lu}/^{176}\text{Lu} = 37.70$, $^{181}\text{Ta}/^{180}\text{Ta} = 8129$, and
327 $^{183}\text{W}/^{180}\text{W} = 109.0$. A total of five 30 ppb JMC-475 Hf standard solutions were run at the beginning and end of
328 the analytical session, with 200 ratios collected during each measurement. During the measurements, the signal
329 intensities for both the standard and the samples were kept at constant levels, between 1.8 and 2.6 V on the ^{178}Hf
330 mass; the in-run precision of the measured $^{176}\text{Hf}/^{177}\text{Hf}$ ratio for both samples and standards was between 11 and
331 13 ppm (2SE). During the course of the present analytical campaign, the external reproducibility of the 30 ppb
332 JMC-475 Hf standard solution measurements for $^{176}\text{Hf}/^{177}\text{Hf}$ was ± 14 ppm (2SD); this value was used to
333 estimate the true uncertainty on the Hf isotopic analyses. The average $^{176}\text{Hf}/^{177}\text{Hf}$ ratio measured for the JMC-
334 475 Hf standard solution during the analytical session was used to calculate the instrumental mass bias
335 coefficient for correction of the measured $^{176}\text{Hf}/^{177}\text{Hf}$ ratio in the samples using the true value of $^{176}\text{Hf}/^{177}\text{Hf} =$
336 0.282163 ± 0.000009 as determined by a multiple dynamic analysis protocol of Blichert-Toft et al. (1997).

337 In order to assess the accuracy of the analyses, an aliquot of USGS GRM BCR-1 was processed and
338 analyzed using the same protocol as that used for the samples. The measured $^{176}\text{Hf}/^{177}\text{Hf}$ ratio of BCR-1 is
339 identical, within the uncertainty, to the GeoRem preferred value.

340 Measurements of the Lu isotopic compositions were also performed on the Faraday cups of the
341 *ThermoFisher Neptune Plus* ICP-MS at the *PL* in static mode. For each sample, 40 ratios were collected with 4 s
342 integration time in blocks of 20 ratios each. Before each sample measurement, a 60 s baseline measurement was
343 performed for each Faraday cup/amplifier pair without beam deflection, and this baseline was then automatically
344 subtracted from the sample beam. Instrumental mass fractionation was corrected for by analysis of a 10 ppb Lu
345 standard solution before and after every 4 samples. The Yb and Hf isobaric interferences were corrected for on-
346 line using $^{176}\text{Yb}/^{173}\text{Yb} = 0.7962$ and $^{176}\text{Hf}/^{177}\text{Hf} = 0.2822$. The Lu concentration obtained for the SRM BCR-1 in
347 this study was identical, within the uncertainty, to the GeoRem recommended value. The propagated uncertainty
348 on the Lu/Hf ratio in the samples analyzed is estimated to be 0.5%. The total analytical blanks were 70 pg for Hf
349 and 2 pg for Lu.

350 5. Results

351 5.1. Petrography of the komatiites

352 The studied samples of the Jeesiörova komatiites are massive rocks with up to 1 mm-sized
353 euhedral, completely serpentinized olivine phenocrysts in a groundmass of prismatic or
354 needle-like clinopyroxene crystals and devitrified glass (Fig. 2a in Hanski and Kamenetsky,
355 2003). The rocks also contain euhedral chromite grains up to 0.3 mm in size, often containing
356 melt inclusions (Figs. 2b and 2d in Hanski and Kamenetsky, 2003).

358 Compared to the Jeesiörova rocks, the Kevitsa komatiites analyzed in this study are
1 359 characterized by a much better preserved magmatic mineralogy. The rocks are olivine-
2 360 porphyritic, with the most coarse-grained varieties found in dike centers, due to the presence
3 361 of olivine phenocrysts (**Fig. 2A, B, E**). The size of the phenocrysts is commonly less than 6
4 362 mm in the center of the dikes, decreasing gradually to 0.1-0.2 mm near the dike margins.
5 363 Occasionally, large amoeboid and embayed megacrysts are observed, with a maximum grain
6 364 size of ~2 cm (**Fig. 3**). Olivine phenocrysts show various morphologies; they are usually
7 365 euhedral to subhedral, but sometimes have irregular shapes with ragged outlines (**Fig. 2D**).
8 366 The latter are interpreted to be broken pieces of larger phenocrysts. Also observed are
9 367 elongated olivine bars up to 1 cm in length, with an aspect ratio of up to 20 (**Fig. 2B**).
10 368 Although olivine phenocrysts contain ubiquitous chrome spinel inclusions 0.1-0.3 mm in size,
11 369 some of the largest phenocrysts are nearly devoid of them; in these cases, chrome spinel
12 370 grains are concentrated along the olivine grain boundaries.

13 371 The holocrystalline groundmass is composed of clinopyroxene, orthopyroxene, olivine,
14 372 plagioclase, brown amphibole, and opaque minerals, including two kinds of spinels, chrome
15 373 spinel, and lamellar ilmenomagnetite (**Fig. 2C**). These oxides occur as individual or
16 374 composite grains, in which chrome spinel is partly surrounded by ilmenomagnetite.
17 375 Plagioclase forms lath-like grains up to 1 cm in length, but other silicates are anhedral or
18 376 equant, with their grain size ranging commonly between 0.1 and 0.5 mm.

19 377 The chilled margins of the dikes contain olivine microphenocrysts (<0.5 mm in size) and
20 378 some chrome spinel grains in a fine-grained to cryptocrystalline groundmass composed
21 379 mostly of clinopyroxene and plagioclase. It is noteworthy that no vesicles are found in the
22 380 contact zones, indicating that either the pressure of emplacement was too high for volatiles to
23 381 form a gas phase, or the volatile content was very low.

24 382

25 383 **5.2. Mineral chemistry**

26 384 Representative electron probe analyses of igneous minerals from the Kevitsa komatiites
27 385 are listed in the **Supplementary Table A3**.

28 386 **5.2.1. Olivine.** Olivine exhibits a continuous compositional range from Fo_{94.1} to Fo_{73.8},
29 387 with the highest forsterite content reported for large phenocrysts and the lowest for small
30 388 grains in the groundmass. Microphenocrysts in the chilled margin samples vary from Fo_{83.5} to
31 389 Fo_{93.2}. The CaO content of all analyzed olivine grains is 0.20–0.32 wt.%, typical of magmatic
32 390 olivine. This, together with a negative correlation between MgO and MnO, indicates the

391 absence of xenocrystic mantle olivine among the grains analyzed. Instead, all olivine grains
392 are considered to be cognate phases crystallized from the same komatiitic parental magma.

393 **5.2.2. Pyroxenes and plagioclase.** Groundmass clinopyroxene has an augitic average
394 composition of $\text{En}_{49}\text{Fs}_{10}\text{Wo}_{41}$ with Al_2O_3 ranging from 2.4 to 5.8 wt.% and TiO_2 from 0.12 to
395 0.71 wt.%. Orthopyroxene grains average $\text{En}_{71}\text{Fs}_{25}\text{Wo}_4$ and have 3.4–4.5 wt.% Al_2O_3 and
396 0.40–0.57 wt.% TiO_2 . The anorthite content of the analyzed plagioclase grains in the
397 groundmass ranges from 57.7 to 72.1 mol.%.

398 **5.2.3. Chrome spinel.** The most primitive chrome spinel grains have MgO contents of
399 around 15–16 wt.% and corresponding Mg# of 0.71–0.76, whereas in the most evolved
400 grains, MgO and Mg# drop to 2.7 wt.% and 0.14, respectively. The MgO-rich grains show an
401 average Al_2O_3 content of 14.7 wt.% and an average TiO_2 concentration of 0.38 wt.%, yielding
402 $\text{Al}_2\text{O}_3/\text{TiO}_2$ typical of well-preserved chrome spinel grains in komatiitic lavas in Central
403 Lapland, but distinct from the more Ti-rich spinels that occur in the associated picrites
404 (Hanski and Kamenetsky, 2013).

405 **5.2.4. Amphibole.** The compositions of the brown groundmass amphibole (**Fig. 2B, C**)
406 straddle the boundary between Ti-rich pargasite and Ti-rich magnesio-hastingsite. The TiO_2
407 contents fall between 2.9 and 4.0 wt.%, consistent with the magmatic nature of the mineral.
408 Amphibole shows moderate Na_2O contents of 2.3–3.0 wt.%, whereas K_2O is low, 0.10–0.26
409 wt.%, consistent with the low level of K_2O in the parental komatiite magma.

410 **5.3. Major and lithophile trace element abundances**

411 The major and trace element concentration data for the Jeesiörova and Kevitsa komatiites
412 are listed in **Tables 1 and 2**, and selected elements are plotted against MgO contents on
413 variation diagrams in **Fig. 4** and as BSE-normalized values in **Fig. 5**.

414 The MgO abundances range between 25.3 and 28.6 wt.% in the Jeesiörova komatiites and
415 between 16.2 and 28.3 wt.% in the Kevitsa komatiites. In the Kevitsa komatiite dikes, the
416 lowest abundances are observed in the chilled margin samples and the highest abundances in
417 the samples from the central parts. The abundances of TiO_2 , CaO, Al_2O_3 and incompatible
418 lithophile trace elements in the dikes plot on well-constrained trends with negative slopes that
419 intersect the MgO axes at ~51 wt.%; these intersections correspond to the composition of
420 liquidus olivine obtained in this study and also reported for the dikes by Nicklas et al. (2019).
421 The abundances of the compatible elements Ni and Cr show positive correlations with MgO

422 contents (**Fig. 4**). These strong correlations reflect olivine and chromite control during
423 magmatic differentiation of the dikes, and also indicate immobile behavior of the elements
424 analyzed during secondary alteration processes.

425 The four samples of the Jeesiörova komatiites analyzed in this study have slightly higher
426 Ti, Cr, La, Ce, Th, and U concentrations compared to the Kevitsa komatiites (**Fig. 4**), but for
427 the rest of major and trace elements, plot on the same trends. The Jeesiörova and Kevitsa
428 komatiites have $\text{Al}_2\text{O}_3/\text{TiO}_2$ (11.7 ± 0.3 and 13.9 ± 0.1 , respectively, 2SE) that are lower, and
429 $(\text{Gd}/\text{Yb})_N$ (1.53 ± 0.08 and 1.49 ± 0.01 , respectively, 2SE) that are higher, than the BSE values
430 of 22.4 and 1.00, respectively. This is due to enrichment of the parental komatiite magma in
431 Ti and the middle REE, rather than to depletion in Al and the heavy REE, as is evident from
432 comparison with the typical Al-undepleted komatiites from Pyke Hill, Ontario, Canada (**Fig.**
433 **4A**). As such, they are classified as the Al-undepleted komatiite type of Nesbitt et al. (1979).

434 The BSE-normalized lithophile trace element abundances (**Fig. 5**) are characterized by
435 strong depletions in light REE ($(\text{La}/\text{Sm})_N = 0.347\pm 0.022$ and 0.264 ± 0.002 in the Jeesiörova
436 and Kevitsa komatiites, respectively; 2SE), sloping HREE ($(\text{Gd}/\text{Yb})_N = 1.53\pm 0.08$ and
437 1.49 ± 0.01 , respectively; 2SE), and enrichments in Th and U relative to the neighboring
438 element with similar incompatibility, i.e., Nb ($(\text{Nb}/\text{Th})_N = 0.595\pm 0.019$ and 0.583 ± 0.011 ,
439 respectively).

5.4. Re-Os isotopic compositions and HSE abundances

442 The Re-Os isotopic and HSE abundance data for the Jeesiörova and Kevitsa komatiites
443 and olivine and chromite separates are listed in **Table 3** and plotted on the $^{187}\text{Re}/^{188}\text{Os}$ versus
444 $^{187}\text{Os}/^{188}\text{Os}$ diagram in **Fig. 6**, on a CI chondrite-normalized plot in **Fig. 7**, and on MgO
445 versus HSE variation diagrams in **Fig. 8**.

446 Thirty-three whole-rock samples and olivine and chromite separates define a regression
447 line with a slope corresponding to an ISOPLOT Model 3 age of 2049 ± 13 Ma and a chondritic
448 initial $^{187}\text{Os}/^{188}\text{Os} = 0.11284\pm 28$ ($\gamma^{187}\text{Os} = -0.2\pm 0.2$). This age is in good agreement with the
449 internal Sm-Nd isochron age of 2056 ± 25 Ma obtained by Hanski et al. (2001) for the
450 Jeesiörova komatiites. The initial $\gamma^{187}\text{Os} = -0.2\pm 0.2$ is identical, within uncertainty, to the less
451 precise average initial $\gamma^{187}\text{Os} = +0.1\pm 0.6$ obtained for chromite separates from the Jeesiörova
452 komatiites by Gangopadhyay et al. (2006).

453 In CI chondrite-normalized HSE diagrams (**Fig. 7**), the Jeesiörova and Kevitsa komatiite
454 samples display typical komatiitic patterns with moderate enrichments in incompatible Pt, Pd,

455 and Re relative to compatible Os, Ir, and Ru ($(\text{Pd}/\text{Ir})_{\text{N}} = 2.84\text{-}10.1$). By contrast, the two
1 456 olivine separates analyzed show an inverse relationship, being enriched in compatible relative
2 457 to incompatible HSE, with the exception of the elevated Re abundance in one of the separates.
3 458 This elevated Re concentration is, however, supported by the correspondingly more
4 459 radiogenic Os isotopic composition of this separate, indicating that the olivine likely
5 460 contained inclusions of interstitial melt enriched in incompatible elements.

10 461 In the MgO versus HSE diagrams (**Fig. 8**), variations in the Os and Ir contents show broad
11 462 positive correlations with the MgO abundances, with the large degree of scatter most likely
12 463 due to the presence of Os-Ir alloy on the liquidus, along with olivine. The Ru abundances
13 464 display a more regular behavior and were likely mostly controlled by chromite fractionation
14 465 during differentiation of the komatiites.

19 466 The Pt, Pd, and Re abundances show strong negative correlations with MgO contents and,
20 467 with the exception of some scatter in the Re content, especially among the Jeesiörova
21 468 komatiites, plot on the trends that pass through the olivine compositions; these variations are
22 469 consistent with olivine control, indicating both incompatible behavior of these HSE during
23 470 komatiite differentiation and their immobile behavior during post-magmatic processes. Of
24 471 note, the Jeesiörova and Kevitsa rocks show a subchondritic $(\text{Pt}/\text{Pd})_{\text{N}} = 0.78 \pm 0.01$, which is
25 472 typical of komatiites (e.g., Puchtel et al., 2009b), but is ~20% lower than the near-chondritic
26 473 ratios in the Ni-Cu-PGE sulfide deposits of Finnish Lapland (Mutanen 1997; Törmänen et al.,
27 474 2016).

38 476 *5.5. Pt-Os isotope systematics*

39 477 The high-precision Os isotopic data for the chromite separates from the Jeesiörova
40 478 komatiite samples 12D-PPR and 13 EJH are presented in **Table 4** and plotted on the
41 479 $^{190}\text{Pt}/^{188}\text{Os}$ versus $^{186}\text{Os}/^{188}\text{Os}$ diagram in **Fig. 6B**. The data for the seven chromite separates
42 480 define a regression line with a slope corresponding to an age of ~2.0 Ga and an average initial
43 481 $^{186}\text{Os}/^{188}\text{Os} = 0.1198369 \pm 3$ ($\mu^{186}\text{Os} = +29 \pm 2$, 2SE) calculated for the time of the komatiite
44 482 emplacement derived from the Re-Os isochron (2049 Ma).

53 484 *5.6. W isotopic compositions and abundances*

54 485 The W abundances and isotopic compositions of the Jeesiörova and Kevitsa komatiites are
55 486 listed in **Tables 2 and 5** and plotted in **Fig. 4B** and as BSE-normalized values in **Fig. 5**. The
56 487 W abundances vary between 8.8 and 13 ppb in the Kevitsa komatiites; the only analyzed
57 488 sample of the Jeesiörova komatiite contains 27 ppb W. In the MgO versus W variation

489 diagram (**Fig. 4B**), the data for the Kevitsa komatiites plot with some scatter around a nearly
490 horizontal trend line, whereas the only data point for the Jeesiörova komatiite plots above the
491 trend line.

492 The BSE-normalized trace element abundances (**Fig. 5**) for the Kevitsa komatiites are
493 characterized by small negative to positive W abundance offsets relative to neighboring
494 elements (i.e., Th and U) with similar incompatibility during mantle melting ($W/W^* =$
495 0.95 ± 0.13 , where $W/W^* = W_N / (\sqrt{[Th_N \times U_N]})$ and N are BSE-normalized values from Arevalo
496 and McDonough (2008) and Hofmann (1988)). The W/W^* ratio in the only Jeesiörova
497 komatiite sample analyzed (LP10) is 2.1.

498 The two Kevitsa komatiite samples and two replicates analyzed in this study for W
499 isotopic composition, are characterized by $^{182}\text{W}/^{184}\text{W}$ ratios that are indistinguishable from
500 $^{182}\text{W}/^{184}\text{W}$ measured in the laboratory reference material, with an average $\mu^{182}\text{W} = +1.5 \pm 3.3$
501 (2SE, $n = 4$), where $\mu^{182}\text{W}$ is the parts per million deviation of $^{182}\text{W}/^{184}\text{W}$ of a given sample
502 from that of the reference material, which, by definition, is equal to zero (**Table 5**).

504 ***5.7. Sm-Nd and Lu-Hf isotopic compositions and abundances***

505 The ^{147}Sm - ^{143}Nd isotopic and concentration data for the whole-rock samples and
506 clinopyroxene separates of the Jeesiörova and Kevitsa komatiites are listed in **Table 6** and
507 plotted in **Fig. 9**. Regression of the combined Sm-Nd isotopic data (**Fig. 9a**) for 13 samples
508 yields a Model 1 ISOPLOT internal isochron age of 2046 ± 22 Ma (MSWD = 2.4), which is
509 identical, within the uncertainty, to the Re-Os isochron age of 2049 ± 13 Ma. The initial $\epsilon^{143}\text{Nd}$
510 derived from the isochron is $+3.8 \pm 0.9$. The individual samples have very uniform initial Nd
511 isotopic compositions, with initial $\epsilon^{143}\text{Nd}$ ranging between $+3.5$ and $+3.9$. A more precise
512 initial $\epsilon^{143}\text{Nd} = +3.7 \pm 0.3$ (2SD, $n = 13$) is obtained by averaging the initial $^{143}\text{Nd}/^{144}\text{Nd}$ ratios
513 calculated for each sample using the measured $^{147}\text{Sm}/^{144}\text{Nd}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios. Our results
514 are consistent with those obtained earlier for the most LREE-depleted Jeesiörova komatiites
515 (Hanski et al., 2001).

516 The ^{176}Lu - ^{176}Hf data (**Fig. 9b**) for the Jeesiörova whole-rock komatiite sample 13-EJH
517 and clinopyroxene separate define a regression line in $^{176}\text{Lu}/^{177}\text{Hf}$ - $^{176}\text{Hf}/^{177}\text{Hf}$ space with a
518 slope corresponding to an ISOPLOT Model 1 age of 2072 ± 37 Ma, which also overlaps,
519 within uncertainties, the Re-Os emplacement age of the Jeesiörova and Kevitsa komatiites.
520 The initial $\epsilon^{176}\text{Hf}$ value derived from the regression is $+8.7 \pm 0.7$. The average initial $\epsilon^{176}\text{Hf}$

521 value calculated for the two samples analyzed ($+8.7\pm 0.3$, 2SD, $n = 2$) is identical to the one
522 derived from the isochron.

523 Plotted in **Fig. 9c** are the Nd and Hf isotopic data for the Jeesiörova komatiites, together
524 with the Nd-Hf terrestrial array from Blichert-Toft and Puchtel (2010). In contrast to other
525 late Archean and post-Archean komatiite systems (e.g., Kostomuksha: Blichert-Toft and
526 Puchtel, 2010; Abitibi: Puchtel et al., 2018; Vetryny: Puchtel et al., 2016a; Gorgona:
527 Thompson et al., 2003), the calculated initial $\varepsilon^{143}\text{Nd}$ and $\varepsilon^{176}\text{Hf}$ values of the Jeesiörova
528 komatiites (**Fig. 9c**) do not plot on the terrestrial array of Blichert-Toft and Puchtel (2010).

530 **6. Discussion**

531 **6.1. Jeesiörova-Kevitsa emplaced komatiite magma composition**

532 Nicklas et al. (2019) reported major and trace element abundance data for olivine
533 phenocrysts and whole-rock samples of the Kevitsa komatiites. Based on the Mg-Fe olivine-
534 komatiite melt equilibria, these authors estimated that the parental magma to the Kevitsa
535 komatiites contained 27.4 ± 2.4 wt.% MgO. By comparison for this study, in order to estimate
536 the parental magma composition, we used an approach different from that of Nicklas et al.
537 (2019). We performed mathematical incremental step-wise addition of olivine to a melt
538 composition represented by sample KD-14 (MgO = 16.3 wt.%) until the melt was in
539 equilibrium with olivine $\text{Fo}_{94.1}$, the most MgO-rich olivine population found in the Kevitsa
540 komatiites (Nicklas et al., 2019, and this study). To select appropriate olivine-melt Fe-Mg
541 distribution coefficients (K_D), we utilized the results of the isothermal crystallization
542 experiments of komatiitic liquids carried out by Sossi and O'Neill (2016). Using the empirical
543 K_D defined as $(\text{Fe}^{2+}/\text{Mg})_{\text{ol}}/(\Sigma\text{Fe}/\text{Mg})_{\text{melt}}$ and the T-MgO- K_D - $\log f\text{O}_2$ relationships from their
544 study, we adjusted the K_D values at each step of olivine addition. Because K_D defined in this
545 way using bulk Fe in the melt is dependent on the oxygen fugacity, two different redox
546 conditions were considered, $\Delta\text{FMQ} = +0.9$ and $\Delta\text{FMQ} = 0.0$. The former choice is based on
547 the observed V partitioning between olivine and melt in the Kevitsa komatiites (Nicklas et al.,
548 2019) and the latter on the composition of chrome spinel ($\text{Fe}^{3+}/(\text{Fe}^{3+}+\text{Al}+\text{Cr}) = 0.06$). The K_D
549 values were varied between 0.28–0.26 and 0.29–0.27 in these two cases, respectively, and the
550 required additions of olivine were 34 and 40%.

551 The two resulting major element compositions of the emplaced komatiite melt are similar,
552 with the most important difference being the MgO content. The more oxidized conditions
553 yielded an MgO content of 25.2 wt.%, while the FMQ buffer conditions gave an MgO content

554 of 26.3 wt.%. These estimates are identical, within the uncertainties, to those obtained by
1 555 Nicklas et al. (2019) using an independent technique, and corroborate the komatiitic
2 556 composition of the emplaced Kevitsa komatiite magma. The results indicate that the parental
3 557 magmas to the Kevitsa, and, therefore, also Jeesiörova, komatiites approached those of their
4 558 late Archean counterparts from the Abitibi and Belingwe greenstone belts (Puchtel et al.,
5 559 2009b) in terms of MgO content, and, therefore, their liquidus temperatures.

6.2. *Lithophile trace element and isotope systematics of the komatiites*

560
561
562 The Jeesiörova and Kevitsa komatiites have low $(\text{La}/\text{Sm})_{\text{N}}$ (0.347 ± 0.022 and 0.264 ± 0.002 ,
563 respectively), and high positive initial ϵNd and ϵHf ($+3.7\pm 0.3$ and $+8.7\pm 0.3$, respectively),
564 indicating derivation from a mantle source that was strongly depleted in highly incompatible,
565 lithophile trace elements. These komatiites are ~ 2.5 times more strongly depleted in the light
566 REE than, e.g., modern N-MORB (average $(\text{La}/\text{Sm})_{\text{N}} = 0.654$: Hofmann, 1988). Yet, the
567 lavas are also characterized by enrichments in Th, U, and W relative to Nb, an element with
568 similar incompatibility during melting of spinel peridotite ($(\text{Nb}/\text{Th})_{\text{N}} = 0.583\pm 0.011$), and a
569 negative Nb anomaly ($\text{Nb}/\text{Nb}^* = 0.582\pm 0.010$, where $\text{Nb}/\text{Nb}^* = \text{Nb}_{\text{N}}/\sqrt{(\text{Th}_{\text{N}}*\text{La}_{\text{N}})}$). This is
570 most likely indicative of minor, yet significant crustal contamination.

571 Upper crustal rocks are generally characterized by low $(\text{Nb}/\text{Th})_{\text{N}}$ and $(\text{Nb}/\text{La})_{\text{N}}$ ratios and,
572 as a result, display pronounced negative Nb anomalies relative to the elements with similar
573 incompatibility during mantle melting (Th and La) on BSE-normalized trace element
574 diagrams, i.e., their $\text{Nb}/\text{Nb}^* \ll 1.0$ (Rudnick and Fountain, 1995; Rudnick and Gao, 2014).
575 By contrast, primary uncontaminated komatiitic magmas commonly do not show a Nb
576 anomaly, i.e., their $\text{Nb}/\text{Nb}^* = 1.00$ (Jochum et al., 1991; Puchtel et al., 1998). Jochum et al.
577 (1991) noted that the elements most sensitive to crustal contamination in komatiite magmas
578 include Th, U, and light REE and, to a lesser extent, Nb. As a result, the Nb/Nb^* ratio can be
579 utilized as a measure of the degree of crustal contamination and, when applied in combination
580 with the lithophile element isotopic systems, to discriminate between contaminated and
581 uncontaminated lavas. Based on these observations, it is likely that the original Jeesiörova-
582 Kevitsa komatiite magma was contaminated with upper crustal rocks of the Fennoscandian
583 Shield *en route* to the surface.

584 In order to test this hypothesis and calculate the major and trace element and Nd, Hf, and
585 W isotopic compositions of the original Jeesiörova-Kevitsa komatiite magma, we performed
586 model mixing calculations, the results of which are presented in **Table 8**. For the crustal
587 contaminant endmember, we assumed an average composition of the Vodla Block tonalitic

588 gneisses that are the dominant upper crustal rock type within the adjacent Karelian granite-
589 greenstone terrane (Puchtel et al., 2016b). These Karelian tonalites are similar in age and
590 major- and trace element composition to their counterparts from Finnish Lapland (e.g.,
591 Huhma et al., 2012), but, unlike the latter, have been better characterized in terms of the
592 isotopic systematics of interest (**Table 8**). This average tonalite composition from Puchtel et
593 al. (2016b) is henceforth referred to as the Fennoscandian Tonalite Average (FTA). In order
594 to calculate the degree of contamination and the trace element abundances and Nd and Hf
595 isotopic compositions of the original, uncontaminated Jeesiörova-Kevitsa komatiite magma,
596 we assumed that it had a $Nb/Nb^* = 1.00$. Calculations show that to produce the emplaced
597 Jeesiörova-Kevitsa komatiite magma with $Nb/Nb^* = 0.582$, the original komatiite magma had
598 to be mixed with ~1% of the average FTA ($Nb/Nb^* = 0.118$).

599 Using 1% contamination and the Nd and Hf abundances and initial Nd and Hf isotopic
600 compositions of the FTA calculated at the time of the Kevitsa komatiite magma emplacement,
601 the original Kevitsa komatiite magma and, thus, its source is calculated to have had initial
602 ϵNd and ϵHf values of +4.9 and +10.2, respectively.

603 On the ϵNd versus ϵHf evolution diagram in **Fig. 9c**, the calculated datum for the
604 Jeesiörova-Kevitsa komatiite source plots within uncertainty of the terrestrial evolution curve
605 of Blichert-Toft and Puchtel (2010). This indicates coupled, or congruent, behavior of the two
606 isotope systems in the source of the original komatiite magma. It appears, therefore, that the
607 Nd-Hf isotope decoupling observed in the Jeesiörova-Kevitsa emplaced komatiite magma was
608 caused by contamination with the older tonalitic crust that had isotopically evolved away
609 from the terrestrial evolution curve by the time it was mixed with the original komatiite
610 magma (**Table 8**). The datum for the Jeesiörova-Kevitsa komatiite source also plots near the
611 Depleted MORB Mantle (DMM) value of Goldstein and Jacobsen (1988) at 2049 Ma (**Fig.**
612 **9c**), indicating derivation from a mantle source that was already as long-term depleted in
613 incompatible lithophile trace elements as DMM at 2.05 Ga.

615 **6.3. Osmium isotope - HSE systematics of the Jeesiörova-Kevitsa komatiite source**

616 Although the Os isotopic compositions and absolute and relative HSE abundances in the
617 mantle sources of early and late Archean komatiites have been the subject of numerous
618 studies over nearly four decades (e.g., Brüggemann et al., 1987; Walker et al., 1988; Foster et
619 al., 1996; Gangopadhyay and Walker, 2003; Puchtel et al., 2009a; 2014, 2016b; Maier et al.,
620 2009), studies of post-Archean komatiites have been rare, in part because the number of
621 Proterozoic and younger komatiites is far more limited.

622 In order to characterize the HSE present in the Paleoproterozoic mantle source of the
623 Jeesiörova-Kevitsa komatiites, the effects of crustal contamination must first be evaluated.
624 We again use the average FTA composition from Puchtel et al. (2016b) to assess the effects
625 of 1% contamination of the original Jeesiörova-Kevitsa komatiite magma on its Os isotopic
626 composition and HSE abundances. Because of the very low concentrations of these elements
627 in the FTA, the effects are negligible. Contamination of this magnitude would have increased
628 $\mu^{186}\text{Os}$ and $\gamma^{187}\text{Os}$ values by only 0.004 (based on the calculated $^{186}\text{Os}/^{188}\text{Os} = 0.1198767$ at
629 2.05 Ga for the FTA assuming its derivation from a BSE-type mafic source at 3.2 Ga, per
630 Puchtel et al., 2016b) and 0.01 (based on the calculated $^{187}\text{Os}/^{188}\text{Os} = 1.2091$ at 2.05 Ga for
631 the FTA assuming its derivation from a BSE-type mafic source at 3.2 Ga, per Puchtel et al.,
632 2016b) units, respectively, which are well within the uncertainties of the current estimates for
633 these values. Similarly, effects on absolute HSE abundances would have been $\ll 1\%$.

634 Estimation of the HSE abundances in the mantle source of the Jeesiörova-Kevitsa
635 komatiites is achieved by the bootstrap method of Puchtel et al. (2004b) that sequentially
636 combines Os isotopic and HSE concentration data. First, we used the initial $^{187}\text{Os}/^{188}\text{Os}$ and
637 $^{186}\text{Os}/^{188}\text{Os}$ ratios calculated for the Jeesiörova-Kevitsa komatiites to estimate the time-
638 integrated evolution of Re/Os and Pt/Os in their mantle source. In order to place minimum
639 constraints on these long-term source ratios, we calculated the parent/daughter elemental
640 ratios necessary to arrive at the Os isotopic composition at 2049 Ma by assuming generation
641 of this mantle domain soon after Solar System formation. It is estimated that this source
642 evolved from an early Solar System $^{187}\text{Os}/^{188}\text{Os} = 0.09517$ at 4567 Ma (Archer et al., 2014) to
643 the initial $^{187}\text{Os}/^{188}\text{Os} = 0.11285 \pm 23$ at 2049 Ma with $^{187}\text{Re}/^{188}\text{Os} = 0.397 \pm 5$. This ratio is well
644 within the range of bulk chondritic meteorites. Using the early Solar System $^{186}\text{Os}/^{188}\text{Os} =$
645 0.1198269 at 4567 Ma (Brandon et al., 2006) requires a source with a $^{190}\text{Pt}/^{188}\text{Os} = 0.00269 \pm 8$
646 to have evolved to its $^{186}\text{Os}/^{188}\text{Os} = 0.1198369 \pm 3$ at 2049 Ma. In contrast to Re/Os, the
647 required minimum $^{190}\text{Pt}/^{188}\text{Os}$ ratio is $\sim 50\%$ higher than the average $^{190}\text{Pt}/^{188}\text{Os} = 0.00180 \pm 17$
648 (2SD) in bulk chondritic meteorites (as compiled from the data of Horan et al. (2003),
649 Brandon et al. (2005, 2006), Fischer-Gödde et al. (2010), and van Acken et al. (2011)).

650 The calculated initial $\gamma^{187}\text{Os}$ of the Jeesiörova-Kevitsa komatiite source is also within the
651 range of those for the majority of Archean and Paleoproterozoic komatiite sources, which
652 were characterized by initial $\gamma^{187}\text{Os}$ values between -0.1 and $+1.3$, also well within the range
653 of 95% of chondritic meteorites (**Fig. 10a**). The $^{186}\text{Os}/^{188}\text{Os}$ isotope data available for
654 Archean and Paleoproterozoic komatiite sources indicate that mantle sources of the 2.7 Ga

655 Abitibi (Canada), 2.7 Ga Belingwe (South Africa), and 2.4 Ga Vetreny (Fennoscandia)
1 656 komatiite systems evolved with time-integrated Pt/Os within the chondritic range (**Fig. 10b**),
2 657 whereas the coupled high initial $^{186,187}\text{Os}/^{188}\text{Os}$ of the 2.8 Ga Kostomuksha komatiite system
3 658 (Fennoscandia) require long-term suprachondritic Pt/Os and Re/Os (Puchtel et al., 2005). By
4 659 contrast, the 3.5 Ga Komati and 3.3 Ga Weltevreden komatiite systems (South Africa)
5 660 evolved with non-chondritic Pt/Os, but chondritic Re/Os ratios (Puchtel et al., 2014),
6 661 displaying decoupling of the Re-Os and Pt-Os isotopic systems. The calculated
7 662 suprachondritic initial $\mu^{186}\text{Os} = +29 \pm 2$ in the Jeesiörova-Kevitsa komatiite source obtained in
8 663 this study is similar to $\mu^{186}\text{Os} = +22 \pm 7$ in the source of the Weltevreden komatiite system
9 664 (Puchtel et al., 2014).

10 665 Because of the availability of high-precision Os isotopic data and immobile behavior of
11 666 HSE in the Kevitsa komatiites during post-magmatic alteration, absolute abundances of
12 667 incompatible Re, Pt, and Pd and compatible Os in the Jeesiörova-Kevitsa komatiite source
13 668 can be estimated using the bootstrap protocol of Puchtel et al. (2004b).

14 669 As was shown in a number of previous studies (e.g., Alard et al., 2000; Lorand and Alard,
15 670 2001; Luguet et al., 2007), $\geq 90\%$ of the HSE budget of the mantle resides in two major types
16 671 of sulfides. The high-temperature Os-Ir-Ru-rich Fe-Ni monosulfide solid solution (*mss*)
17 672 usually occurs as inclusions in liquidus phases, primarily olivine, whereas lower-temperature
18 673 Cu-Ni sulfides occupy interstitial space between the liquidus phases. During partial melting of
19 674 mantle peridotite, Cu-Ni sulfides enter the melt, whereas *mss* remains trapped in the melting
20 675 residue. It takes $\sim 18\text{-}25\%$ partial melting (Barnes et al., 1985; Keays, 1995; Luguet et al.,
21 676 2007; Mungall and Naldrett, 2008; Naldrett, 2010; Fonseca et al., 2011; 2012; Kiseeva et al.,
22 677 2017) for all the low-temperature Cu-Ni sulfides to get exhausted in the mantle. As the degree
23 678 of melting continues to increase, the magma becomes sulfide-undersaturated. It has also
24 679 recently been shown that decrease in $f\text{S}_2$ with increase in the degree of melting triggers
25 680 exsolution of Os-Ir alloys from the refractory *mss* in the residue (Fonseca et al., 2011; 2012).
26 681 All low-degree (basalts) and the majority of higher-degree (picrites and komatiites) partial
27 682 melts are characterized by compatible behavior of Os and Ir during magmatic differentiation,
28 683 indicating that their parental magmas remained saturated in Os-Ir alloys (Puchtel et al.,
29 684 2004b; Barnes and Fiorentini, 2008). However, some lavas, such as the 2.8 Ga Kostomuksha
30 685 and the 3.55 Ga Schapenburg komatiites, exhibited an incompatible behavior of Os and Ir
31 686 during magma differentiation, likely indicating near-complete exhaustion of Os-Ir alloys in
32 687 the mantle sources of these komatiites (Puchtel and Humayun, 2005; Puchtel et al., 2009a).

688 Apparently, the sulfide saturation status of komatiite magmas is a crucial parameter that
1 689 plays a key role in determining whether their HSE abundances accurately reflect those in their
2 690 mantle sources. In the following discussion, we evaluate the sulfide saturation status of the
3 691 Jeesiörova-Kevitsa komatiite magma during both mantle melting and after emplacement.
4

5 692 Sulfide saturation of a mafic magma is determined by several key parameters, the most
6 693 important of which are pressure (P), temperature (T), and FeO content. In their recent study,
7 694 Steenstra et al. (2020) combined the results of their own work with previously published
8 695 experimental data of Mavrogenes and O'Neill (1999), Li and Agee (2001), Holzheid and
9 696 Grove (2002), O'Neill and Mavrogenes (2002), Jugo et al. (2005), Liu et al. (2007), Brennan
10 697 (2008), Kiseeva and Wood (2013, 2015), Ding et al. (2014, 2018), Fortin et al., (2015), Wood
11 698 and Kiseeva (2015), Smythe et al. (2017), and Steenstra et al. (2018) to produce a refined
12 699 parameterization model for the sulfur concentration at sulfide saturation (SCSS) as a function
13 700 of P, T, and major and trace element abundances in silicate liquids (Eq. 4 in Steenstra et al.,
14 701 2020).

15 702 In order to calculate the P-T parameters of the Jeesiörova-Kevitsa komatiite magma
16 703 generation, we used the mantle melting parameterization model of McKenzie and Bickle
17 704 (1988). These calculations yield the following parameters: potential mantle temperature of the
18 705 Jeesiörova-Kevitsa komatiite mantle source $T_{pot} = 1763^{\circ}\text{C}$, liquidus temperature upon
19 706 emplacement $T_{liq} = 1544^{\circ}\text{C}$, depth of melting initiation 294 km, which translates into a
20 707 pressure of 9.5 GPa. Based on these parameters, the temperatures and pressures at which
21 708 melting terminated for the Jeesiörova-Kevitsa komatiite magma was calculated using the
22 709 model of Herzberg et al. (2010) to be 1730°C and 5.0 GPa. Using these parameters, the
23 710 chemical composition of the original Jeesiörova-Kevitsa komatiite magma from Table 8, and
24 711 Eq. 4 of Steenstra et al. (2020), the SCSS of the Jeesiörova-Kevitsa komatiite magma upon
25 712 separation from the melting residue and ensuing emplacement are calculated to be 2270 and
26 713 5200 ppm, respectively.

27 714 In order to estimate the S content in the Jeesiörova-Kevitsa komatiite magma and, thus, its
28 715 sulfide saturation status, knowledge of the sulfur content of its mantle source, the S partition
29 716 coefficient, and the degree of melting is required. The average S content of modern DMM is
30 717 estimated to be 206 ± 25 ppm (Sun et al., 2020). Since the Jeesiörova-Kevitsa komatiite source
31 718 was as strongly melt-depleted as contemporary DMM, we assume it also had a similar S
32 719 content.

33 720 To estimate the degree of partial melting that produced the original Jeesiörova-Kevitsa
34 721 komatiite magma, we used a batch melting model and the calculated abundances of

722 moderately incompatible lithophile elements (e.g., Al, Ti, Gd) in this magma and its garnet-
723 free mantle source of spinel lherzolite composition, as indicated by their Al-undepleted nature
724 and the depth of melting initiation of ~290 km. These calculations yield 49% partial melting.
725 Assuming that sulfur was highly incompatible during komatiite melting, with the degree of
726 incompatibility similar to that of La, the original Jeesiörova-Kevitsa komatiite magma is
727 calculated to have contained 420 ppm S. This S content is, thus, ~5 times lower than the
728 SCSS of the Jeesiörova-Kevitsa komatiite magma upon separation from the melting residue
729 and ~12 times lower than the SCSS of the Jeesiörova-Kevitsa komatiite magma upon
730 emplacement, indicating that it was strongly sulfur-undersaturated upon separating from the
731 source and remained as such all the way to the surface.

732 The sulfur-undersaturated nature of the Jeesiörova-Kevitsa komatiite magma upon
733 emplacement is also attested to by the incompatible behavior of the highly chalcophile
734 element Cu during magma differentiation after emplacement. The Cu abundances in the
735 Kevitsa komatiites strongly correlate with the MgO content and plot on an olivine control line
736 in the MgO versus Cu diagram (**Fig. 4a**), indicating that sulfide liquid was not a fractionating
737 phase within the compositional range represented by the samples of this study.

738 To estimate the abundances of incompatible Pt and Pd in the Jeesiörova-Kevitsa komatiite
739 source, we extrapolated the differentiation trends for each of these elements to a mantle MgO
740 = 38 wt.% using ISOPLOT. The concentrations of Pt and Pd in the source were, thus,
741 calculated to be 9.54 ± 0.22 and 8.13 ± 0.19 ppb, respectively, and the total Pt and Pd
742 abundances to be 17.7 ± 0.8 ppb, or $120 \pm 5\%$ of those in the average estimates for modern BSE
743 of 14.7 ± 2.0 ppb (Becker et al., 2006).

744 The concentration of compatible Os in the source was estimated to be 3.4 ± 0.1 ppb using
745 the calculated Pt content in the source and the long-term $^{190}\text{Pt}/^{188}\text{Os}$ derived from the initial
746 $^{186}\text{Os}/^{188}\text{Os}$. Further, using the calculated Os concentration and the estimated long-term
747 $^{187}\text{Re}/^{188}\text{Os}$ derived from the initial $^{187}\text{Os}/^{188}\text{Os}$, the Re concentration in the source is
748 calculated to be 0.285 ± 0.008 ppb. Unlike the Pt and Pd concentrations, therefore, the
749 calculated Os and Re concentrations in the source of the Jeesiörova-Kevitsa komatiites are,
750 respectively, 13% and 19% lower than the average estimates for BSE.

751 The presence of residual Os/Ir alloys in the melting residue, as evidenced by the
752 compatible behavior of Os and Ir during the Jeesiörova-Kevitsa magma differentiation, makes
753 it possible for the Os/Ir ratio to fractionate between the source and the komatiitic magma.
754 Furthermore, as discussed earlier, chromite fractionation likely had a strong effect on Ru
755 concentrations in the Jeesiörova-Kevitsa komatiite magma. Based on these observations, we

756 were unable to constrain the abundances of Ir and Ru in the source with any meaningful
1 757 degree of accuracy.

3 758 The calculated total Pt and Pd abundances in the Jeesiörova-Kevitsa komatiite source are
4
5 759 also plotted as a function of age in **Fig. 11** and compared with the data for other Archean and
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7 760 Paleoproterozoic komatiite systems studied to date and the BSE estimate. The calculated total
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9 761 Pt and Pd abundances in the source of the Jeesiörova-Kevitsa komatiites are significantly
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11 762 higher than in any other komatiite system examined so far. The significance of this
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13 763 observation will be discussed in the following sections.

15 764 ***6.4. The siderophile and lithophile element isotope conundrum***

17 765 There is an apparent decoupling between the radiogenic lithophile trace element isotope
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19 766 systematics (e.g., Sm-Nd and Lu-Hf) and Re-Os isotope systematics in komatiite mantle
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21 767 sources worldwide. Many komatiite systems, such as those from the Abitibi (Canada) and
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23 768 Belingwe (South Africa) greenstone belts (Puchtel et al., 2009b; Puchtel et al., 2018), the
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25 769 Weltevreden and Komati (South Africa) Formations (Puchtel et al., 2014), and the Sumozero-
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27 770 Kenozero (Fennoscandia) greenstone belt (Puchtel et al., 2007), are characterized by strong,
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29 771 long-term depletions in incompatible lithophile trace elements, e.g., Nd versus Sm, but
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31 772 chondritic Re/Os. By contrast, certain other komatiite systems, such as the Kostomuksha in
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33 773 Fennoscandia (Puchtel et al., 2005) and the Schapenburg in South Africa (Puchtel et al.,
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35 774 2016a), while being long-term depleted in light REE, evolved with long-term suprachondritic
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37 775 Re/Os ratios. In this study, we address this apparent conundrum using the Sm-Nd, Lu-Hf, and
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39 776 Re-Os isotope and lithophile trace element abundance data for the Jeesiörova-Kevitsa
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41 777 komatiite system.

42 778 In Section 6.2, we calculated that the mantle source of the Jeesiörova-Kevitsa komatiites
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44 779 had the initial $\epsilon^{143}\text{Nd} = +4.9$ and $\epsilon^{176}\text{Hf} = +10.2$. This source was similar to DMM at 2.05 Ga
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46 780 (Goldstein and Jacobsen, 1988) and had long-term coupled Sm/Nd and Lu/Hf isotope
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48 781 systematics (Blichert-Toft and Puchtel, 2010). If the source had originally chondritic Sm/Nd
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50 782 and Lu/Hf, then it experienced a melt-extraction event a minimum of several hundred million
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52 783 years prior to komatiite emplacement. The same melt extraction event might also have been
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54 784 expected to result in depletion of incompatible Re relative to compatible Os, as well as more
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56 785 modest depletion in Pt relative to Os. Over time, these chemical effects would have led to the
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58 786 development of negative $\gamma^{187}\text{Os}$ and $\mu^{186}\text{Os}$ values. Instead, as shown in Section 6.3, the
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787 Jeesiörova-Kevitsa komatiite source evolved within the range of chondritic Re/Os and
788 suprachondritic Pt/Os.

6.5. Tungsten isotopic composition of the Jeesiörova-Kevitsa komatiite source

790 The Paleoproterozoic mantle was recently shown to have been characterized by an
791 oxidation state similar to that of the modern mantle (Nicklas et al., 2019). As a result, during
792 melting of the Paleoproterozoic mantle, W likely was characterized by a degree of
793 incompatibility similar to that of Th and U (Newsom et al., 1996; Arevalo and McDonough,
794 2008; König *et al.*, 2011). Therefore, the BSE-normalized W abundance in the emplaced
795 Jeesiörova-Kevitsa komatiite magma is expected to be similar to those of Th and U. This is
796 the case, with an average calculated $W/W^* = 1.0$. This also implies that W, similarly to Th
797 and U, was characterized by immobile behavior during the limited postmagmatic alteration of
798 the Jeesiörova-Kevitsa komatiites. Further, similarly to the other lithophile trace elements,
799 such as Nd and Hf, upper crustal rocks have significantly higher W abundances than
800 komatiites; they also likely had a different W isotopic composition compared to komatiites,
801 and, therefore, addition of even a small amount of upper crustal material can significantly
802 change both the W isotopic composition and W abundances of a komatiite magma.

803 In this study, we used the measured W isotopic composition and W abundance of the
804 emplaced Jeesiörova-Kevitsa komatiite magma to calculate those of the original, non-
805 contaminated komatiite magma. In order to remove the effect of the 1% crustal contamination
806 estimated on the basis of the lithophile trace element systematics, we used the W isotopic
807 composition and W abundances of the average FTA from Puchtel et al. (2016b). The
808 endmember parameters and the results of the mixing calculations are given in **Table 8**. The
809 calculations suggest that the original Jeesiörova-Kevitsa komatiite magma contained ~5.8 ppb
810 W, implying that ~50% of the total W budget in the emplaced Jeesiörova-Kevitsa komatiite
811 magma was derived from the crustal contaminant.

812 The average $\mu^{182}\text{W}$ value of the FTA reported by Puchtel et al. (2016b) was $+12.6 \pm 4.5$, so
813 contamination of the original komatiite magma with W-rich crust would have led to an
814 increase in the $\mu^{182}\text{W}$ value of the original magma. If the assumptions regarding the original
815 uncontaminated komatiite magma concentration of W, and concentration and ^{182}W isotopic
816 composition of the crustal contaminant are accurate, the W isotopic composition of the
817 original Jeesiörova-Kevitsa komatiite magma and, by extension, of its mantle source, after
818 correction for crustal contamination, is calculated to be $\mu^{182}\text{W} = -10 \pm 5$. Here, the uncertainty

819 solely reflects propagated measurement uncertainties for the komatiite and the FTA. Because
820 it is a projected original melt composition, this $\mu^{182}\text{W}$ must be considered tentative.
821 Nevertheless, negative $\mu^{182}\text{W}$ values of this magnitude are common in modern ocean island
822 basalt systems (Rizo et al., 2019; Mundl-Petermeier et al., 2019; 2020) and have also been
823 observed in the 3.6 Ga Schapenburg komatiites (Puchtel et al., 2016a), South Africa, and
824 some spatially associated, Archean glacial diamictites (Mundl et al., 2018). By contrast, the
825 majority of evolved Archean crustal rocks similar to those envisioned for the FTA
826 contaminant, are characterized by positive ^{182}W anomalies (e.g., Touboul et al., 2014;
827 Willbold et al., 2015; Rizo et al., 2016b).

6.6. Reconciling isotopic characteristics of the Jeesiörova-Kevitsa komatiites

829 The combined chemical and isotopic data for the Jeesiörova-Kevitsa komatiites are
830 complex. The Nd and Hf isotopic data are consistent with derivation from a mantle domain
831 that experienced at least one melting event prior to the one that produced the komatiitic
832 magma. The Nd-Hf isotope characteristics of this mantle domain are similar to projections of
833 the modern DMM to 2049 Ma. The ^{187}Os isotopic data indicate derivation from a broadly
834 chondritic mantle source that was not strongly long-term melt depleted. Because the long-
835 term effects of low degrees of partial melting on Re/Os, compared to Sm/Nd and Lu/Hf, are
836 relatively small (Puchtel et al., 2004b; 2009b), the chondritic $^{187}\text{Os}/^{188}\text{Os}$ of the mantle source
837 can be considered at least qualitatively consistent with the Nd and Hf isotopic data. By
838 contrast, the fractionated Pt/Os of the mantle source, as revealed by the suprachondritic initial
839 $^{186}\text{Os}/^{188}\text{Os}$, and the correspondingly fractionated HSE pattern for the source, requires an
840 event that could have strongly modified HSE relative abundances early in Earth history.
841 Finally, if the negative $\mu^{182}\text{W}$ value for the Jeesiörova-Kevitsa komatiite system obtained
842 after correction for crustal contamination is accurate, it requires invoking one of a limited
843 number of possible processes to account for the projected anomalous composition for the Hf-
844 W short-lived radiogenic isotope system.

845 To address these complexities, we consider two melting models. The parameters used in
846 both models are as follows. (1) The mantle domain that ultimately became the source of the
847 Jeesiörova-Kevitsa komatiites initially had a composition consistent with that of the BSE. The
848 chemical and isotopic parameters of this domain were compiled from Jacobsen and
849 Wasserburg (1980), Hamilton et al. (1983), Bouvier et al. (2008), Hofmann (1988), Shirey
850 and Walker (1998), Brandon et al. (2006), and Becker et al. (2006). (2) This domain had a

851 modal mineral composition typical of spinel lherzolite, with 65% olivine, 24% orthopyroxene,
1 852 10% clinopyroxene, and 1% spinel. (3) The melting of the domain was modeled using the
2 853 batch melting equation of Shaw (1970). The bulk D^{Sm} , D^{Nd} , D^{Lu} , and D^{Hf} during melting were
3 854 calculated to be 0.040, 0.024, 0.090, and 0.037, respectively, based on the above source
4 855 modal composition and crystal-liquid D values for the respective mineral phases from Green
5 856 (1994). The bulk D^{Re} and D^{Pt} under the redox conditions of $\Delta\text{FMQ} +0.90\pm 0.33$ determined by
6 857 Nicklas et al. (2019) were estimated to be 0.20 and 2.0, respectively (Rehkämper et al., 1999;
7 858 Mallmann and O'Neill, 2007). The bulk D^{Os} for the low-degree partial melting, consistent
8 859 with formation of a MORB-type magma, was adopted to be 50 (e.g., Gannoun et al., 2016 and
9 860 references therein).
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19 861 In Model 1, we consider an early, global differentiation event. This model requires a
20 862 minimum degree of melting and a melt extraction that would lead to fractionation of Sm/Nd
21 863 and Lu/Hf in the BSE domain from the chondritic value to that required to bring the $\varepsilon^{143}\text{Nd}$
22 864 and $\varepsilon^{176}\text{Hf}$ of the domain to the corrected values of +4.9 and +10.2, respectively, by the time
23 865 of komatiite formation at 2049 Ma. The effect of this melt extraction event on the Re-Os and
24 866 Pt-Os isotope systems is then considered. Model 2 considers the youngest-possible, maximum
25 867 degree of melting event that would fractionate the Sm/Nd and Lu/Hf ratios in the BSE domain
26 868 from the chondritic values to those calculated for the Jeesiörova-Kevitsa komatiite *source*,
27 869 and again bring the $\varepsilon^{143}\text{Nd}$ and $\varepsilon^{176}\text{Hf}$ values of this domain to +4.9 and +10.2, respectively,
28 870 by 2049 Ma. The effect of this melt extraction event on the Re-Os and Pt-Os isotope systems
29 871 is then also considered.
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41 872 For Model 1, we assume that the global differentiation event occurred 100 Ma after Solar
42 873 System formation. This timing is consistent with the recent data on the early formation of the
43 874 Moon at ~ 4.51 Ga (Barboni et al., 2017). The Jeesiörova-Kevitsa komatiite source would
44 875 have evolved from chondritic ratios of $^{143}\text{Nd}/^{144}\text{Nd} = 0.506807$ and $^{176}\text{Hf}/^{177}\text{Hf} = 0.279863$ at
45 876 4.467 Ga to the initial ratios of $^{143}\text{Nd}/^{144}\text{Nd} = 0.510236\pm 8$ and $^{176}\text{Hf}/^{177}\text{Hf} = 0.281762\pm 18$ at
46 877 2049 Ma with long-term ratios of $^{147}\text{Sm}/^{144}\text{Nd} = 0.2123\pm 5$ and $^{176}\text{Lu}/^{177}\text{Hf} = 0.0394\pm 4$. Such
47 878 modest change in $^{147}\text{Sm}/^{144}\text{Nd}$ and $^{176}\text{Lu}/^{177}\text{Hf}$ from the chondritic ratios of 0.1967 and
48 879 0.0336, respectively, would require only $\sim 0.6\%$ melting, followed by removal of the melt. At
49 880 the same time, this melt extraction event would have decreased the $^{187}\text{Re}/^{188}\text{Os}$ in the melting
50 881 residue from 0.4019 to 0.3912; the latter ratio is essentially identical to the calculated long-
51 882 term $^{187}\text{Re}/^{188}\text{Os} = 0.397\pm 8$ determined for the Jeesiörova-Kevitsa komatiite source. However,
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883 such low-degree melting event would not have had any measurable effect on the Pt/Os ratio
1 884 of the source. Therefore, Model 1 cannot explain the observed $^{186}\text{Os}/^{188}\text{Os}$ composition of the
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3 885 Jeesiörova-Kevitsa komatiite system.
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5 886 For Model 2, using the lithophile trace element data, we first calculate the $^{147}\text{Sm}/^{144}\text{Nd}$
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7 887 and $^{176}\text{Lu}/^{177}\text{Hf}$ ratios in the Jeesiörova-Kevitsa komatiite source to be 0.2912 and 0.0704,
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9 888 respectively. Modeling indicates that 9% melting would have generated a residual mantle
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11 889 domain with these Sm/Nd and Lu/Hf ratios. The latest time of the melting event required for
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13 890 the mantle domain with such Sm/Nd and Lu/Hf ratios to have evolved to the initial $\epsilon^{143}\text{Nd} =$
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15 891 $+4.9$ and $\epsilon^{176}\text{Hf} = +10.2$ by 2049 Ma is 2.45 Ga. Any event that occurred later than that would
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17 892 not have afforded sufficient time to develop such radiogenic $\epsilon^{143}\text{Nd}$ and $\epsilon^{176}\text{Hf}$ in the mantle
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19 893 domain. It is noteworthy that 2.45 Ga coincides with the ages of abundant mafic layered
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21 894 intrusions (Alapieti et al., 1990; Balashov et al., 1993; Amelin et al., 1995; Yang et al., 2016)
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23 895 and komatiitic lavas (Puchtel et al., 1997; 2016b) throughout the Fennoscandian Shield,
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25 896 indicating that large volumes of mafic magma were indeed produced in this part of the planet
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27 897 at that time.

28 898 Using this extent and time of melting, and applying a bulk $D^{\text{Re}} = 0.2$, model calculations
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30 899 indicate that the melting event that produced this light REE depletion would have reduced the
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32 900 $^{187}\text{Re}/^{188}\text{Os}$ ratio in the chondritic reservoir from 0.4019 to 0.2694 at 2.45 Ga. The resultant
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34 901 melt-depleted mantle domain, if allowed to evolve from 2.45 Ga until the time of the
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36 902 komatiite formation at 2049 Ma with this Re/Os ratio, would have developed a negative initial
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38 903 $\gamma^{187}\text{Os} = -0.8$, which is close to the observed value and well within the range of chondritic
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40 904 meteorites at 2049 Ma (**Fig. 10**). At the same time, this melting event would only have
41
42 905 reduced the $^{190}\text{Pt}/^{188}\text{Os}$ ratio in the mantle domain by 0.5%. The 400 Ma of growth from 2.45
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44 906 to 2.05 Ga with this slightly reduced Pt/Os ratio will decrease the $\mu^{186}\text{Os}$ value of the system
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46 907 by 0.4 ppm, not increase it by 29 ppm, as observed for the Jeesiörova-Kevitsa komatiite
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48 908 source. Hence, similarly to Model 1, Model 2 can account for the Nd, Hf, and ^{187}Os isotopic
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50 909 compositions of the komatiites, but it also fails for ^{186}Os .

51 910 One significant modification to Models 1 and 2, which would make it possible to account
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53 911 for all the observed isotopic and elemental data for the komatiites, is assuming early
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55 912 perturbation of the Pt/Os ratio in the komatiite source, without affecting Sm-Nd and Lu-Hf
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57 913 evolution, as well as not modifying Re-Os evolution outside of the chondritic range. The most
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59 914 parsimonious process that can achieve this perturbation is crystal-liquid fractionation of
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61 915 metal, given that fractional crystallization of metal can strongly fractionate Pt/Os, but only

16 modestly fractionates Re/Os, with solid metal-liquid metal bulk distribution coefficients in
17 this order: $D_{Os} > D_{Re} \gg D_{Pt} > 1$ (e.g., Cook et al., 2004). Thus, addition to a mantle domain of
18 evolved metal either through late accretion of a fragment of a differentiated planetesimal core,
19 similar to the mechanism proposed by Fischer-Gödde et al. (2011), or addition of metal
20 derived from the Earth's core, but partially crystallized at or near the core-mantle boundary,
21 similar to the mechanism of core-mantle interaction proposed by Humayun (2011), can best
22 achieve the Os isotopic requirements for the komatiites. Either process would have the added
23 benefit of decreasing the $\mu^{182}W$ value of the mixed domain to a negative value, as suggested
24 by the contamination-corrected $\mu^{182}W = -10$ for the komatiites.

25 Compilation of time-integrated $^{190}Pt/^{188}Os$ and $^{187}Re/^{188}Os$ ratios for IIAB, IIIAB, and
26 IVA magmatic iron meteorites from Cook et al. (2004) and McCoy et al. (2011) reveals large,
27 up to three orders of magnitude, variations in the $^{190}Pt/^{188}Os$ ratios that resulted from
28 magmatic solid metal-liquid metal fractionation, accompanied by relatively small, about a
29 factor of 2.5, variations in the $^{187}Re/^{188}Os$ ratios in the same magmatic systems. The $^{186,187}Os$
30 isotope systematics of the Jeesiörova-Kevitsa komatiite source can be accounted for by
31 mixing with a number of fractionated iron compositions. We performed mixing calculations
32 for Os, Re, and Pt concentrations, as well as for W isotopic composition and W abundances,
33 results of which are illustrated in **Fig. 12**. For the meteoritic endmember, we chose a low-Os
34 group IIIAB iron meteorite, Tieraco Creek; similar results are also obtained for other low-Os
35 irons from this group, e.g., Grant and Chupaderos. The mantle source endmember used in the
36 mixing calculations was the BSE. The HSE and W abundance data and W isotopic
37 composition for the endmembers were adopted from Cook et al. (2004), Becker et al. (2006),
38 Arevalo and McDonough (2008), Wasson (1999), and Kruijer et al. (2017); these are
39 presented in **Table 9**. Based on the results of calculations, addition of 0.2% of material with
40 the composition of Tieraco Creek to a mantle domain of BSE composition can explain the
41 HSE and W abundances and Os and W isotopic compositions of the mantle domain that gave
42 rise to the Jeesiörova-Kevitsa komatiite system. One caveat to this model is that the mixing
43 event must have occurred very early, within the first 100 Ma of Solar System history, while
44 late accretion of substantial mass was still possible.

45 Addition of fractionated metal to the Moon (and presumably Earth, too) in the form of a
46 late accretionary component was advocated by Fischer-Gödde and Becker (2012) in their
47 study of lunar impact melt breccias to explain suprachondritic Pt/Ir, Ru/Ir, and Pd/Ir ratios in
48 some of the impact melts. It has also been argued that the impactors that created the largest
49 impact basins on the Moon and Mars, and presumably Earth as well, were hundreds of km in

950 diameter (Ryder, 2002; Strom et al., 2005; Bottke et al., 2010) and represented by
1 951 differentiated planetesimals and, thus, would be capable of generating large HSE and negative
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3 952 ^{182}W heterogeneities in portions of the mantle (Marchi et al., 2018, 2020). Cores of
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5 953 differentiated planetesimals have up to two orders of magnitude higher W abundances and up
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7 954 to 350 ppm less radiogenic $^{182}\text{W}/^{184}\text{W}$ than the modern BSE (Kleine et al., 2002, 2004;
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9 955 Schoenberg et al., 2002; Yin et al., 2002). Thus, any mantle domain to which an excess of this
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11 956 material was added would be ^{182}W -depleted. Such a mantle domain would also be expected to
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13 957 be enriched in HSE, compared to BSE, particularly in Pt and Pd.

14 958 In order to further evaluate the effect of addition of such material on $^{182}\text{W}/^{184}\text{W}$ and HSE
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16 959 abundances, we plot the $\mu^{182}\text{W}$ in the Jeesiörova-Kevitsa komatiite system versus the total
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18 960 calculated HSE abundances in its mantle source relative to those in the estimates for modern
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20 961 BSE (**Fig. 13**). This proportion corresponds to the fraction of the total HSE budget of the BSE
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22 962 added during late accretion assuming an HSE-free mantle following core formation. The
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24 963 $^{182}\text{W}/^{184}\text{W}$ of the BSE prior to late accretion is constrained by mass-balance calculations (e.g.,
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26 964 Kleine and Walker, 2017) and is supported by the $^{182}\text{W}/^{184}\text{W}$ data for the lunar mantle of
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28 965 $+25\pm 5$ ppm (Kruijer et al., 2015; Touboul et al., 2015; Kruijer and Kleine, 2017). Calculations
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30 966 indicate that the estimated 10 ppm deficit in $\mu^{182}\text{W}$ is consistent with the calculated total HSE
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32 967 abundances in the source of the Jeesiörova-Kevitsa komatiite system and the results of the
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34 968 mixing calculations presented above.

35 969 A second type of processes to consider is interaction between Earth's core and mantle.
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37 970 This type of process was originally proposed by Walker et al. (1995, 1997) and Brandon et al.
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39 971 (1998, 1999) as a means of generating coupled $^{186,187}\text{Os}$ enrichments observed in some ocean
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41 972 island basalts. Later, Humayun (2011) proposed a mechanism of core-mantle interaction that
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43 973 involved outer core liquid becoming trapped in a cumulate pile of Fe-rich non-metallic
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45 974 precipitates (FeO, FeS, Fe₃Si) at the top of the core. This metal is presumed to undergo partial
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47 975 fractional crystallization. The resulting, chemically-evolved liquid is then incorporated into
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49 976 the base of the mantle, where mantle plumes subsequently entrain it. If this metal had a
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51 977 composition similar to that of evolved iron meteorites, such as Tieraco Creek, its entrainment
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53 978 into the plume that gave rise to the Jeesiörova-Kevitsa komatiites could explain the $^{186,187}\text{Os}$
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55 979 isotope systematics of these komatiites. As with the grainy late accretion model, this
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57 980 mechanism has the added benefit of explaining the W isotope systematics of the komatiites.
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59 981 Based on the discovery of negative $\mu^{182}\text{W}$ values in modern ocean island basalts (Mundl et
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61 982 al., 2017), Rizo et al. (2019) and Mundl-Petermeier et al. (2020) have argued for some form

983 of core-mantle exchange/equilibration in order to transfer the inferred negative $\mu^{182}\text{W}$ value
984 of the core ($\mu^{182}\text{W} = -220$) to plumes rising from the core-mantle boundary. Although ^{186}Os -
985 ^{182}W data for the same rocks are still limited, all Hawaiian lavas with negative $\mu^{182}\text{W}$ (ranging
986 between -7.8 and -20.2) are also characterized by positive $\mu^{186}\text{Os}$ (ranging between +18.6 and
987 +62.8) and $\gamma^{187}\text{Os}$ (ranging between +2.0 and +7.2) values (**Fig. 14**). The projected negative
988 $\mu^{182}\text{W}$ value for the Jeesiörova-Kevitsa komatiites is within the range of $\mu^{182}\text{W}$ values for
989 Hawaii, implying that these komatiites could be a Proterozoic equivalent of Hawaiian picrites.
990 This process would be expected to have decreased $^{182}\text{W}/^{184}\text{W}$ in the komatiite source relative
991 to the ambient mantle, and also potentially have increased the HSE abundances over the
992 ambient mantle levels, which is consistent with the observations.

993 We conclude that at present, both grainy late accretion and core-mantle interaction remain
994 viable mechanisms for explaining the observed combined lithophile trace- and siderophile
995 element isotope and abundance systematics of the Jeesiörova-Kevitsa komatiite system.

7. Conclusions

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998 The Jeesiörova-Kevitsa komatiites of the Fennoscandian Shield are unique among the
999 komatiite systems studied so far. They have a relatively young age of 2.05 Ga that has not
1000 been reported for the existing terrestrial komatiite record elsewhere, yet are as MgO-rich as
1001 some of their late Archean counterparts. These komatiites were derived from a contemporary
1002 DMM-like mantle source in terms of Nd-Hf isotope systematics; this source was
1003 characterized by a negative ^{182}W anomaly, total absolute HSE abundances ~20% higher than
1004 of the modern BSE, and chondritic initial $^{187}\text{Os}/^{188}\text{Os}$, but suprachondritic initial $^{186}\text{Os}/^{188}\text{Os}$
1005 isotopic compositions. When considered together, these constraints are reconciled within a
1006 model that requires derivation of the parental Jeesiörova-Kevitsa komatiite magma from a
1007 mantle domain that experienced a melting and melt depletion event within the first 100 Ma of
1008 Solar System history. This mantle domain is argued here to have acquired its HSE and Os
1009 isotope systematics as a result of either (1) grainy late accretion of differentiated metallic
1010 impactors or (2) core-mantle interaction via addition of chemically fractionated outer core
1011 metal to the mantle source domain of the komatiites. The survival of the early-formed
1012 Jeesiörova-Kevitsa komatiite mantle source for ≥ 2.5 billion years implies that portions of the
1013 mantle remained poorly mixed with respect to HSE and W until at least the Paleoproterozoic.

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21025 Alapieti, T.T., Filen, B.A., Lahtinen, J.J., Lavrov, M.M., Smolkin, V.F., and Voitsekhovskiy,
31026 S.N., 1990. Early Proterozoic layered intrusions in the northeastern part of the
41027 Fennoscandian Shield. *Mineralogy and Petrology* **42**: 1-22.
- 51028 Alard, O., Griffin, W. L., Lorand, J.-P., Jackson, S. E., O'Reilly, S. Y., 2000. Non-chondritic
71029 distribution of the highly siderophile elements in mantle sulfides. *Nature* **407** (6806): 891-
81030 894.
- 91031 Albarède, F., Blichert-Toft, J., Vervoort, J. D., Gleason, J. D., Rosing, M., 2000. Hf-Nd
101032 isotope evidence for a transient dynamic regime in the early terrestrial mantle. *Nature* **404**
111033 (6777): 488-490.
- 121034 Amelin, Y.V., Heaman, L.M., and Semenov, V.S., 1995. U-Pb geochronology of layered
131035 mafic intrusions in the eastern Baltic Shield: implications for the timing and duration of
141036 Paleoproterozoic continental rifting. *Precambrian Research* **75**(1-2): 31-46.
- 151037 Archer, G. J., Mundl, A., Walker, R. J., Worsham, E. A., and Bermingham, K. R., 2017.
161038 High-precision analysis of $^{182}\text{W}/^{184}\text{W}$ and $^{183}\text{W}/^{184}\text{W}$ by negative thermal ionization mass
171039 spectrometry: Per-integration oxide corrections using measured $^{18}\text{O}/^{16}\text{O}$. *International*
181040 *Journal of mass-spectrometry* **414**: 80-86.
- 191041 Archer, G. J., Ash, R. D., Bullock, E. S., and Walker, R. J., 2014. Highly siderophile elements
201042 and ^{187}Re - ^{187}Os isotopic systematics of the Allende meteorite: Evidence for primary
211043 nebular processes and late-stage alteration. *Geochimica et Cosmochimica Acta* **131**: 402-
221044 414.
- 231045 Archer, G. J., Brennecka, G. A., Gleißner, P., Stracke, A., Becker, H., and Kleine, T., 2019.
241046 Lack of late-accreted material as the origin of ^{182}W excesses in the Archean mantle:
251047 Evidence from the Pilbara Craton, Western Australia. *Earth and Planetary Science Letters*
261048 **528**: 115841.
- 271049 Arevalo, R. and McDonough, W. F., 2008. Tungsten geochemistry and implications for
281050 understanding the Earth's interior. *Earth and Planetary Science Letters* **272** (3-4): 656-665.
- 291051 Balashov, Y. A., Bayanova, T. B., and Mitrofanov, F. P., 1993. Isotope data on the age and
301052 genesis of layered basic-ultrabasic intrusions in the Kola Peninsula and northern Karelia,
311053 northeastern Baltic Shield. *Precambrian Research* **64**(1-4): 197-205.
- 321054 Barboni, M., Boehnke, P., Keller, B., Kohl, I. E., Schoene, B., Young, E. D., and McKeegan,
331055 K. D., 2017. Early formation of the Moon 4.51 billion years ago. *Science Advances* **3** (1):
341056 e1602365.
- 351057 Barnes, S.-J., Naldrett, A. J., Gorton, M. P., 1985. The origin of the fractionation of platinum-
361058 group elements in terrestrial magmas. *Chemical Geology* **53** (3-4): 303-323.
- 371059 Barnes, S.-J., Often, M., 1990. Ti-rich komatiites from Northern Norway. *Contributions to*
381060 *Mineralogy and Petrology* **105** (1): 42-54.
- 391061 Barnes, S. J., Fiorentini, M. L., 2008. Iridium, ruthenium and rhodium in komatiites:
401062 Evidence for iridium alloy saturation. *Chemical Geology* **257** (1-2): 44-58.
- 411063 Becker, H., Horan, M. F., Walker, R. J., Gao, S., Lorand, J.-P., Rudnick, R. L., 2006. Highly
421064 siderophile element composition of the Earth's primitive upper mantle: Constraints from
431065 new data on peridotite massifs and xenoliths. *Geochimica et Cosmochimica Acta* **70** (17):
441066 4528-4550.
- 451067 Begemann, F., Ludwig, K.R., Lugmair, G.W., Min., K., Nyquist, L.E., Patchett, P.J., Renne,
461068 P.R., Shih, C.-Y., Villa, I.M., and Walker, R.J., 2001. Call for an improved set of decay
471069 constants for geochronological use. *Geochimica et Cosmochimica Acta* **65**(1): 111-121.
- 481070 Blichert-Toft, J., Chauvel, C., and Albarède, F., 1997. Separation of Hf and Lu for high-
491071 precision isotope analysis of rock samples by magnetic sector multiple collector ICP-MS.
501072 *Contributions to Mineralogy and Petrology* **127**(3): 248-260.

- 1073 Blichert-Toft, J., 2001. On the Lu-Hf isotope geochemistry of silicate rocks. *Geostandards*
11074 *Newsletter* **25**, 41-56.
- 21075 Blichert-Toft, J., Puchtel, I. S., 2010. Depleted mantle sources through time: Evidence from
31076 Lu-Hf and Sm-Nd isotope systematics of Archean komatiites. *Earth and Planetary Science*
41077 *Letters* **297** (3-4): 598-606.
- 61078 Bottke, W. F., Walker, R. J., Day, J. M. D., Nesvorny, D., Elkins-Tanton, L., 2010. Stochastic
71079 Late Accretion to Earth, the Moon, and Mars. *Science* **330** (6010): 1527-1530.
- 81080 Boyet, M., Carlson, R. W., 2005. ¹⁴²Nd evidence for early (> 4.53 Ga) global differentiation
91081 of the silicate Earth. *Science* **309** (5734): 576-581.
- 101082 Brandon, A. D., Walker, R. J., Morgan, J. W., Norman, M. D., and Prichard, H. M., 1998.
121083 Coupled ¹⁸⁶Os and ¹⁸⁷Os evidence for core-mantle interaction. *Science* **208** (5369): 1570-
131084 1573.
- 141085 Brandon, A. D., Norman, M. D., Walker, R. J., and Morgan, J. W., 1999. ¹⁸⁶Os-¹⁸⁷Os
151086 systematics of Hawaiian picrites. *Earth and Planetary Science Letters* **174** (1-2): 25-42.
- 161087 Brandon, A. D., Humayun, M., Puchtel, I. S., Zolensky, M., 2005. Re-Os isotopic systematics
181088 and platinum group element composition of the Tagish Lake carbonaceous chondrite.
191089 *Geochimica et Cosmochimica Acta* **69** (6): 1619-1631.
- 201090 Brandon, A.D., Walker, R.J., and Puchtel, I.S., 2006. Platinum-osmium isotope evolution of
221091 the Earth's mantle: Constraints from chondrites and Os-rich alloys. *Geochimica et*
231092 *Cosmochimica Acta* **70**(8): 2093-2103.
- 241093 Brenan, J. M., 2008. Re-Os fractionation by sulfide melt-silicate melt partitioning: A new
251094 spin. *Chemical Geology* **248** (3-4): 140-165.
- 261095 Brüggemann G. E., Arndt N. T., Hofmann A. W., and Tobschall H. J., 1987. Noble metal
281096 abundances in komatiite suites from Alexo, Ontario, and Gorgona Island, Colombia.
291097 *Geochimica et Cosmochimica Acta* **51** (8): 2159-2169.
- 301098 Chou, C.-L., Shaw, D. M., Crocket, J. H., 1983. Siderophile trace elements in the Earth's
311099 oceanic crust and upper mantle. *Journal of Geophysical Research* **88** (S2): A507-A518.
- 321100 Cook D. L., Walker R. J., Horan M. F., Wasson J. T., and Morgan J. W., 2004. Pt-Re-Os
341101 systematics of Group IIAB and IIIAB iron meteorites. *Geochimica et Cosmochimica Acta*
351102 **68** (6): 1413-1431.
- 361103 Creaser, R. A., Papanastassiou, D. A., Wasserburg, G. J., 1991. Negative thermal ion mass-
371104 spectrometry of osmium, rhenium, and iridium. *Geochimica et Cosmochimica Acta* **55**
381105 (1): 397-401.
- 401106 Ding, S., Dasgupta, R., and Tsuno, K., 2014. Sulfur concentration of martian basalts at sulfide
411107 saturation at high pressures and temperatures - Implications for deep sulfur cycle on Mars.
421108 *Geochimica et Cosmochimica Acta* **131**: 227-246.
- 431109 Ding, S., Hough, T., Dasgupta, R., 2018. New high pressure experiments on sulfide saturation
451110 of high-FeO*— basalts with variable TiO₂ contents - Implications for the sulfur inventory
461111 of the lunar interior. *Geochimica et Cosmochimica Acta* **222**: 319-339.
- 471112 Fischer-Gödde, M., Becker, H., Wombacher, F., 2010. Rhodium, gold and other highly
481113 siderophile element abundances in chondritic meteorites. *Geochimica et Cosmochimica*
501114 *Acta* **74** (1): 356-379.
- 511115 Fischer-Gödde M., Becker H., and Wombacher F., 2011. Rhodium, gold and other highly
521116 siderophile elements in orogenic peridotites and peridotite xenoliths. *Chemical Geology*
531117 **280** (3-4): 365-383.
- 541118 Fischer-Gödde M. and Becker H., 2012. Osmium isotope and highly siderophile element
551119 constraints on ages and nature of meteoritic components in ancient lunar impact rocks.
571120 *Geochimica et Cosmochimica Acta* **77**: 135-156.

- 1121 Fonseca, R. O. C., Laurenz, V., Mallmann, G., Luguët, A., Hoehne, N., Jochum, K. P., 2012.
11122 New constraints on the genesis and long-term stability of Os-rich alloys in the Earth's
21123 mantle. *Geochimica et Cosmochimica Acta* **87**: 227-242.
- 31124 Fonseca, R. O. C., Mallmann, G., O'Neill, H. S. C., Campbell, I. H., Laurenz, V., 2011.
41125 Solubility of Os and Ir in sulfide melt: Implications for Re/Os fractionation during mantle
51126 melting. *Earth and Planetary Science Letters* **311** (3-4): 339-350.
- 71127 Fortin, M.-A., Riddle, J., Desjardins-Langlais, Y., Baker, D. R., 2015. The effect of water on
81128 the sulfur concentration at sulfide saturation (SCSS) in natural melts. *Geochimica et*
91129 *Cosmochimica Acta* **160**: 100-116.
- 101130 Foster, J. G., Lambert, D. D., Frick, L. R., Maas, R., 1996. Re-Os isotopic evidence for
111131 genesis of Archaean nickel ores from uncontaminated komatiites. *Nature* **382** (6593): 703-
131132 706.
- 141133 Galer, S. J. G., Goldstein, S. L., 1991. Early mantle differentiation and its thermal
151134 consequences. *Geochimica et Cosmochimica Acta* **55** (1-2): 227-239.
- 161135 Gangopadhyay, A. and Walker, R. J., 2003. Re-Os systematics of the ca. 2.7 Ga komatiites
171136 from Alexo, Ontario, Canada. *Chemical Geology* **196** (1-4): 147-162.
- 191137 Gangopadhyay, A., Walker, R. J., Hanski, E., Solheid, P.A., 2006. Origin of Paleoproterozoic
201138 Ti-enriched komatiitic rocks from Jeesiörova, Kittilä Greenstone Complex, Finnish
211139 Lapland. *Journal of Petrology* **47**, 773–789.
- 231140 Gannoun A., Burton K. W., Day J. M. D., Harvey J., Schiano P., and Parkinson I. J., 2016.
241141 Highly Siderophile Element and Os Isotope Systematics of Volcanic Rocks at Divergent
251142 and Convergent Plate Boundaries and in Intraplate Settings. *Reviews in Mineralogy and*
261143 *Geochemistry* **81**: 651-724.
- 271144 Gast, P.W., Triton, G.R., Hedge, C.E., 1964. Isotopic composition of lead and strontium from
291145 Ascension and Gough islands. *Science* **145**, 1181-1185.
- 301146 Goldstein, S. L., Galer, S. J. G., 1992. On the trail of early mantle differentiation: $^{142}\text{Nd}/^{144}\text{Nd}$
311147 ratios of early Archean rocks. *Eos* **73** (30): 323.
- 321148 Goldstein, S. J. and Jacobsen, S. B., 1988. Nd and Sr isotopic systematics of river water
341149 suspended material: implications for crustal evolution. *Earth and Planetary Science*
351150 *Letters* **87** (3): 249-265.
- 361151 Green T. H., 1994. Experimental studies of trace-element partitioning applicable to igneous
371152 petrogenesis - Sedona 16 years later. *Chemical Geology* **117** (1-4): 1-36.
- 381153 Hamilton, P. J., O'Nions, R. K., Bridgwater, D., Nutman, A. P., 1983. Sm-Nd studies of
401154 Archaean metasediments and metavolcanics from West Greenland and their implications
411155 for the Earth's early history. *Earth and Planetary Science Letters* **62** (2): 263-272.
- 421156 Hanski, E., Huhma, H. 2005. Central Lapland Greenstone Belt. In: M. Lehtinen, P. Nurmi,
431157 O.T. Rämö (Eds.), *Precambrian Bedrock of Finland – Key to the Evolution of the*
441158 *Fennoscandian Shield*, Elsevier, Amsterdam, p. 139–194.
- 461159 Hanski, E.J., Kamenetsky, V.S., 2013. Chrome spinel-hosted melt inclusions in primitive
471160 Paleoproterozoic volcanic rocks, northern Finland: evidence for coexistence and mixing
481161 of komatiitic and picritic magmas. *Chemical Geology* **343**, 25–37. DOI:
491162 10.1016/j.chemgeo.2013.02.009.
- 501163 Hanski, E., Huhma, H., Rastas, P., Kamenetsky, V. S., 2001. The Paleoproterozoic komatiite-
521164 picrite association of Finnish Lapland. *Journal of Petrology* **42** (5): 855-876.
- 531165 Hart, S. R., Brooks, S., 1977. The geochemistry and evolution of Early Precambrian mantle.
541166 *Contributions to Mineralogy and Petrology* **61** (2): 109-128.
- 551167 Herzberg, C., Condie, K., and Korenaga, J., 2010. Thermal history of the Earth and its
571168 petrological expression. *Earth and Planetary Science Letters* **292** (1-2): 79-88.
- 581169 Hofmann, A. W., 1984. Geochemical mantle models. *Terra Cognita* **4**: 157-165.

- 1170 Hofmann, A. W., 1988. Chemical differentiation of the Earth: The relationship between
11171 mantle, continental crust and oceanic crust. *Earth and Planetary Science Letters* **90** (3):
21172 297-314.
- 31173 Hofmann, A.W., 1997. Mantle geochemistry: the message from oceanic volcanism. *Nature*
41174 385 (6613): 219-229.
- 61175 Holzheid, A. and Grove, T. L., 2002. Sulfur saturation limits in silicate melts and their
71176 implications for core formation scenarios for terrestrial planets. *American Mineralogist* **87**
81177 (2-3): 227-237.
- 91178 Horan, M. F., Walker, R. J., Morgan, J. W., Grossman, J. N., Rubin, A. E., 2003. Highly
101179 siderophile elements in chondrites. *Chemical Geology* **196** (1-4): 5-20.
- 121180 Huhma H., Mänttari I., Peltonen P., Kontinen A., Halkoaho T., Hanski E., Hokkanen T.,
131181 Hölttä P., Juopperi H., Konnunaho J., Layahe Y., Luukkonen E., Pietikäinen K.,
141182 Pulkkinen A., Sorjonen-Ward P., Vaasjoki M., and Whitehouse M., 2012. The age of the
151183 Archean greenstone belts in Finland. In: *The Archaean of the Karelia Province in Finland*.
161184 Hölttä P. (Ed). Geological Survey of Finland Special Paper 54: 74-105.
- 181185 Huhma, H., Hanski, E., Kontinen, A., Vuollo, J., Mänttari, I., Lahaye, Y. 2018. Sm-Nd and
191186 U-Pb isotope geochemistry of the Palaeoproterozoic mafic magmatism in eastern and
201187 northern Finland. *Geological Survey of Finland, Bulletin* **405**, 150 p.
- 221188 Humayun M., 2011. A model for osmium isotopic evolution of metallic solids at the core-
231189 mantle boundary. *Geochemistry, Geophysics, Geosystems* **12** (3): 1-23.
- 241190 Ireland T. J., Walker R. J., and Brandon A. D., 2011. ¹⁸⁶Os-¹⁸⁷Os systematics of Hawaiian
251191 picrites revisited: New insights into Os isotopic variations in ocean island basalts.
261192 *Geochimica et Cosmochimica Acta* **75** (16): 4456-4475.
- 281193 Jacobsen, S. B., 1988. Isotopic and chemical constraints on mantle-crust evolution.
291194 *Geochimica et Cosmochimica Acta* **52** (6): 1341-1350.
- 301195 Jacobsen, S. B., Wasserburg, G. J., 1980. Sm-Nd isotopic evolution of chondrites. *Earth and*
311196 *Planetary Science Letters* **50** (1): 139-155.
- 321197 Jacobsen, S. B., Yu, G., 2015. Extinct isotope heterogeneities in the mantles of Earth and
331198 Mars: Implications for mantle stirring rates. *Meteoritics & Planetary Science* **50** (4): 555-
341198 567.
- 351199 Jochum, K. P., Arndt, N. T., and Hofmann, A. W., 1991. Nb-Th-La in komatiites and basalts:
361200 constraints on komatiite petrogenesis and mantle evolution. *Earth and Planetary Science*
371201 *Letters* **107** (2): 272-289.
- 401203 Jugo, P. J., Luth, R. W., and Richards, J. P., 2005. An Experimental Study of the Sulfur
411204 Content in Basaltic Melts Saturated with Immiscible Sulfide or Sulfate Liquids at 1300°C
421205 and 1.0 GPa. *Journal of Petrology* **46** (4): 783-798.
- 431206 Keays, R. R., 1995. The role of komatiitic and picritic magmatism and S-saturation in the
441207 formation of ore deposits. *Lithos* **34** (1-3): 1-18.
- 461208 Kimura, K., Lewis, R. S., Anders, S., 1974. Distribution of gold and rhenium between nickel-
471209 iron and silicate melts; implications for abundance of siderophile elements on the Earth
481210 and Moon. *Geochimica et Cosmochimica Acta* **38** (5): 683-701.
- 491211 Kiseeva, E. S. and Wood, B. J., 2013. A simple model for chalcophile element partitioning
501212 between sulphide and silicate liquids with geochemical applications. *Earth and Planetary*
511212 *Science Letters* **383**: 68-81.
- 521213 Kiseeva, E. S. and Wood, B. J., 2015. The effects of composition and temperature on
531214 chalcophile and lithophile element partitioning into magmatic sulphides. *Earth and*
541215 *Planetary Science Letters* **424**: 280-294.
- 561216 Kiseeva, E. S., Fonseca, R. O. C., and Smythe, D. J., 2017. Chalcophile Elements and
571217 Sulfides in the Upper Mantle. *Elements* **113**: 111-116.

- 1219 Kleine, T., Münker, C., Mezger, K., Palme, H., 2002. Rapid accretion and early core
11220 formation on asteroids and the terrestrial planets from Hf-W chronometry. *Nature* **418**
21221 (6901): 952-955.
- 31222 Kleine, T., Mezger, K., Münker, C., Palme, H., Bischoff, A., 2004. ¹⁸²Hf-¹⁸²W isotope
41223 systematics of chondrites, eucrites, and martian meteorites: Chronology of core formation
51224 and early mantle differentiation in Vesta and Mars. *Geochimica et Cosmochimica Acta* **68**
71225 (13): 2935-2946.
- 81226 Kleine, T., Walker, R. J., 2017. Tungsten isotopes in planets. *Annual Review of Earth and*
91227 *Planetary Sciences* **45** (1): 389-417.
- 101228 König, S., Münker, C., Hohl, S., Paulick, H., Barth, A. R., Lagos, M., Pfander, J., Büchl, A.,
121229 2011. The Earth's tungsten budget during mantle melting and crust formation. *Geochimica*
131230 *et Cosmochimica Acta* **75** (8): 2119-2136.
- 141231 Kruijjer, T. S., Kleine, T., 2017. Tungsten isotopes and the origin of the Moon. *Earth and*
151232 *Planetary Science Letters* **475** (Supplement C): 15-24.
- 161233 Kruijjer, T. S., Kleine, T., Fischer-Gödde, M., Sprung, P., 2015. Lunar tungsten isotopic
181234 evidence for the late veneer. *Nature* **520** (7548): 534-537.
- 191235 Kruijjer, T. S., Kleine, T., Borg, L. E., Brennecka, G. A., Irving, A. J., Bischoff, A., Agee, C.
201236 B., 2017. The early differentiation of Mars inferred from Hf-W chronometry. *Earth and*
211237 *Planetary Science Letters* **474** (Supplement C): 345-354.
- 231238 Li, J. and Agee, C. B., 2001. The effect of pressure, temperature, oxygen fugacity and
241239 composition on partitioning of nickel and cobalt between liquid Fe-Ni-S alloy and liquid
251240 silicate: implications for the earth's core formation. *Geochimica et Cosmochimica Acta* **65**
261241 (11): 1821-1832.
- 271242 Liu, Y., Samaha, N.-T., and Baker, D. R., 2007. Sulfur concentration at sulfide saturation
291243 (SCSS) in magmatic silicate melts. *Geochimica et Cosmochimica Acta* **71** (7): 1783-1799.
- 301244 Lorand, J.-P., Alard, O., 2001. Platinum-group element abundances in the upper mantle: new
311245 constraints from in situ and whole-rock analyses of Massif Central xenoliths (France).
321246 *Geochimica et Cosmochimica Acta* **65** (16): 2789-2806.
- 331247 Luguët, A., Shirey, S. B., Lorand, J.-P., Horan, M. F., Carlson, R. W., 2007. Residual
351248 platinum-group minerals from highly depleted harzburgites of the Lherz massif (France)
361249 and their role in HSE fractionation of the mantle. *Geochimica et Cosmochimica Acta* **71**
371250 (12): 3082-3097.
- 381251 Maier, W. D., Barnes, S. J., Campbell, I. H., Fiorentini, M. L., Peltonen, P., Barnes, S. J.,
401252 Smithies, R. H., 2009. Progressive mixing of meteoritic veneer into the early Earth's deep
411253 mantle. *Nature* **460** (7255): 620-623.
- 421254 Mallmann, G. and O'Neill, H. St C., 2007. The effect of oxygen fugacity on the partitioning of
431255 Re between crystals and silicate melt during mantle melting. *Geochimica et*
441256 *Cosmochimica Acta* **71** (11): 2837-2857.
- 451257 Marchi, S., Canup, R. M., and Walker, R. J., 2018. Heterogeneous delivery of silicate and
471258 metal to the Earth by large planetesimals. *Nature Geoscience* **11**: 77-81.
- 481259 Marchi S., Walker, R. J., and Canup, R. M., 2020. A compositionally heterogeneous martian
491260 mantle due to late accretion. *Science Advances* **6** (7): eaay2338.
- 501261 Mavrogenes, J. A. and O'Neill, H. S. C., 1999. The relative effects of pressure, temperature
521262 and oxygen fugacity on the solubility of sulfide in mafic magmas. *Geochimica et*
531263 *Cosmochimica Acta* **63** (7-8): 1173-1180.
- 541264 McCoy, T. J., Walker, R. J., Goldstein, J. I., Yang, J., McDonough, W. F., Rumble, D.,
551265 Chabot, N. L., Ash, R. D., Corrigan, C. M., Michael, J. R., and Kotula, P. G., 2011. Group
571266 IVA irons: New constraints on the crystallization and cooling history of an asteroidal core
581267 with a complex history. *Geochimica et Cosmochimica Acta* **75** (22): 6821-6843.

- 1268 McDonough, W. F. and Sun, S. S., 1995. The composition of the Earth. *Chemical Geology*
11269 **120** (3-4): 223-253.
- 21270 McKenzie, D. and Bickle, M. J., 1988. The volume and composition of melt generated by
31271 extension of the lithosphere. *Journal of Petrology* **29** (3): 625-679.
- 41272 Morgan, J. W., 1985. Osmium isotope constraints on Earth's late accretionary history. *Nature*
61273 **317** (6039): 703-705.
- 71274 Morgan, J. W., 1986. Ultramafic xenoliths: Clues to Earth's late accretionary history. *Journal*
81275 *of Geophysical Research* **91** (B12): 12375-12387.
- 91276 Mundl, A., Touboul, M., Jackson, M. G., Day, J. M. D., Kurz, M. D., Lekic, V., Helz, R. T.,
101277 Walker, R. J., 2017. Tungsten-182 heterogeneity in modern ocean island basalts. *Science*
121278 **356** (6333): 66-69.
- 131279 Mundl, A., Walker, R. J., Reimink, J. R., Rudnick, R. L., Gaschnig, R. M., 2018. Tungsten-
141280 182 in the upper continental crust: Evidence from glacial diamictites. *Chemical Geology*
151281 **494**: 144-152.
- 171282 Mundl-Petermeier, A., Walker, R.J., Jackson, M.G., Blichert-Toft, J., Kurz, M.D., and
181283 Haldórsson, S.A. (2019) Temporal evolution of primordial tungsten-182 and ³He/⁴He
191284 signatures in the Iceland mantle plume. *Chemical Geology* **525**, 245-259.
- 201285 Mundl-Petermeier, A., Walker, R.J., Fischer, R.A., Lekic, V., Jackson, M.G., and Kurz, M.D.,
211286 2020. Anomalous μ¹⁸²W signatures in high ³He/⁴He ocean island basalts – fingerprints of
221287 Earth's core? *Geochimica et Cosmochimica Acta* **271**: 194-211.
- 241288 Mungall, J. E. and Naldrett, A. J., 2008. Ore deposits of the platinum-group elements.
251289 *Elements* **4** (4): 253-258.
- 261290 Mutanen, T. 1997. Geology and ore petrology of the Akanvaara and Koitelainen mafic
271291 layered intrusions and the Keivitsa-Satovaara layered complex, northern Finland.
291292 Geological Survey of Finland, Bulletin 395, 233 p.
- 301293 Mutanen, T. 2005. The Akanvaara intrusion and the Keivitsa–Satovaara Complex, with stops
311294 at Kaikkivaara and Särkivaara intrusions. Geological Survey of Finland, Guide
321295 51b, 124 p.
- 341296 Mutanen, T., Huhma, H. 2001. U-Pb geochronology of the Koitelainen, Akanvaara and
351297 Keivitsa layered intrusions and related rocks. Geological Survey of Finland, Special Paper
361298 **33**: 229–246.
- 371299 Naldrett, A. J., 2010. Secular Variation of Magmatic Sulfide Deposits and Their Source
381300 Magmas. *Economic Geology* **105** (3): 669-688.
- 401301 Nesbitt, R. W., Sun, S. S., Purvis, A. C., 1979. Komatiites: geochemistry and genesis.
411302 *Canadian Mineralogist* **17** (2): 165-186.
- 421303 Newsom, H. E., Sims, K. W. W., Noll, P. D., Jaeger, W. L., Maehr, S. A., Beserra, T. B.,
431304 1996. The depletion of tungsten in the bulk silicate Earth: Constraints on core formation.
441305 *Earth and Planetary Science Letters* **60** (5): 1155-1169.
- 461306 Nicklas, R. W., Puchtel, I. S., and Ash, R. D., 2016. High-precision determination of the
471307 oxidation state of komatiite lavas using vanadium liquid-mineral partitioning. *Chemical*
481308 *Geology* **433**: 36-45.
- 491309 Nicklas R.W., Puchtel I.S., Ash R.D., Piccoli P., Hanski E., Nisbet E.G., Waterton P., Pearson
501310 D.G., and Anbar A.D., 2019. Secular Mantle Oxidation across the Archean-Proterozoic
511311 Boundary: Evidence from V partitioning in Komatiites and Picrites. *Geochimica et*
521312 *Cosmochimica Acta* **250**: 49-75.
- 541313 O'Neill, H. St C. and Mavrogenes, J. A., 2002. The Sulfide Capacity and the Sulfur Content at
551314 Sulfide Saturation of Silicate Melts at 1400°C and 1 bar. *Journal of Petrology* **43** (6):
561315 1049-1087.
- 581316 Peters B. J., Mundl-Petermeier A., Horan M. F., Carlson R. W. and Walker R. J., 2019.
591317 Chemical Separation of Tungsten and Other Trace Elements for TIMS Isotope Ratio

- 1318 Measurements Using Organic Acids. *Geostandards and Geoanalytical Research* 43 (2):
11319 245-259.
- 21320 Puchtel, I.S., Haase, K.M., Hofmann, A.W., Chauvel, C., Kulikov, V.S., Garbe-Schönberg,
31321 C.D., and Nemchin, A.A., 1997. Petrology and geochemistry of crustally contaminated
41322 komatiitic basalts from the Vetreny Belt, southeastern Baltic Shield: Evidence for an early
51323 Proterozoic mantle plume beneath rifted Archean continental lithosphere. *Geochimica et*
61324 *Cosmochimica Acta* **61** (6): 1205-1222.
- 81325 Puchtel I. S., Hofmann A. W., Mezger K., Jochum K. P., Shchipansky A. A., and Samsonov
91326 A. V., 1998. Oceanic plateau model for continental crustal growth in the Archaean: A case
101327 study from the Kostomuksha greenstone belt, NW Baltic Shield. *Earth and Planetary*
111328 *Science Letters* 155 (1-2): 57-74.
- 131329 Puchtel, I. S. and Humayun, M., 2000. Platinum group elements in Kostomuksha komatiites
141330 and basalts: Implications for oceanic crust recycling and core-mantle interaction.
151331 *Geochimica et Cosmochimica Acta* 64 (24): 4227-4242.
- 171332 Puchtel, I. S., Humayun, M., 2005. Highly siderophile element geochemistry of ¹⁸⁷Os-
181333 enriched 2.8-Ga Kostomuksha komatiites, Baltic Shield. *Geochimica et Cosmochimica*
191334 *Acta* **69** (6): 1607-1618.
- 201335 Puchtel, I. S., Brandon, A. D., Humayun, M., 2004a. Precise Pt-Re-Os isotope systematics of
211336 the mantle from 2.7-Ga komatiites. *Earth and Planetary Science Letters* **224** (1-2): 157-
221337 174.
- 241338 Puchtel, I. S., Humayun, M., Campbell, A., Sproule, R., Leshner, C. M., 2004b. Platinum
251339 group element geochemistry of komatiites from the Alexo and Pyke Hill areas, Ontario,
261340 Canada. *Geochimica et Cosmochimica Acta* **68** (6): 1361-1383.
- 271341 Puchtel, I. S., Brandon, A. D., Humayun, M., Walker, R. J., 2005. Evidence for the early
281342 differentiation of the core from Pt-Re-Os isotope systematics of 2.8-Ga komatiites. *Earth*
291343 *and Planetary Science Letters* **237** (1-2): 118-134.
- 301344 Puchtel, I. S., Humayun, M., Walker, R. J., 2007. Os-Pb-Nd isotope and highly siderophile
311345 and lithophile trace element systematics of komatiitic rocks from the Volotsk suite, SE
321346 Baltic Shield. *Precambrian Research* **158** (1-2): 119-137.
- 351347 Puchtel, I. S., Walker, R. J., Anhaeusser, C. R., Gruau, G., 2009a. Re-Os isotope systematics
361348 and HSE abundances of the 3.5 Ga Schapenburg komatiites, South Africa: Hydrous
371349 melting or prolonged survival of primordial heterogeneities in the mantle? *Chemical*
381350 *Geology* **262** (3-4): 355-369.
- 401351 Puchtel, I. S., Walker, R. J., Brandon, A. D., Nisbet, E. G., 2009b. Pt-Re-Os and Sm-Nd
411352 isotope and HSE and REE systematics of the 2.7 Ga Belingwe and Abitibi komatiites.
421353 *Geochimica et Cosmochimica Acta* **73** (20): 6367-6389.
- 431354 Puchtel, I. S., Blichert-Toft, J., Touboul, M., Walker, R. J., Byerly, G., Nisbet, E. G.,
441355 Anhaeusser, C. R., 2013. Insights into early Earth from Barberton komatiites: Evidence
451356 from lithophile isotope and trace element systematics. *Geochimica et Cosmochimica Acta*
461357 **108**: 63-90.
- 481358 Puchtel, I. S., Walker, R. J., Touboul, M., Nisbet, E. G., Byerly, G. R., 2014. Insights into
491359 Early Earth from the Pt-Re-Os isotope and Highly Siderophile Element abundance
501360 systematics of Barberton komatiites. *Geochimica et Cosmochimica Acta* **125**: 394-413.
- 521361 Puchtel, I. S., Blichert-Toft, J., Touboul, M., Horan, M. F., Walker, R. J., 2016a. The coupled
531362 ¹⁸²W-¹⁴²Nd record of early terrestrial mantle differentiation. *Geochemistry, Geophysics,*
541363 *Geosystems* **17** (6): 2168-2193.
- 551364 Puchtel, I. S., Touboul, M., Blichert-Toft, J., Walker, R. J., Brandon, A. D., Nicklas, R. W.,
561365 Kulikov, V. S., Samsonov, A. V., 2016b. Lithophile and siderophile element systematics
571366 of the mantle at the Archean-Proterozoic boundary: Evidence from 2.4 Ga komatiites.
581367 *Geochimica et Cosmochimica Acta* **180**: 227-255.

- 1368 Puchtel, I. S., Blichert-Toft, J., Touboul, M., and Walker, R. J., 2018. ^{182}W and HSE
11369 constraints from 2.7 Ga komatiites on the heterogeneous nature of the Archean mantle.
21370 *Geochimica et Cosmochimica Acta* **228**: 1-26.
- 31371 Rehkämper M., Halliday A. N., Fitton J. G., Lee D.-C., Wieneke M., and Arndt N. T., 1999.
41372 Ir, Ru, Pt and Pd in basalts and komatiites: New constraints for the geochemical behavior
51373 of the platinum group elements in the mantle. *Geochimica et Cosmochimica Acta* 63 (22):
71374 3915-3934.
- 81375 Rizo H., Andraut D., Bennett N. R., Humayun M., Brandon A. D., Vlastelic I., Moine B.,
91376 Poirier A., Bouhifd M. A., and Murphy D. T., 2019. ^{182}W evidence for core-mantle
101377 interaction in the source of mantle plumes. *Geochemical Perspectives Letters* 11: 6-11.
- 121378 Rizo, H., Walker, R. J., Carlson, R. W., Horan, M. F., Mukhopadhyay, S., Manthos, V.,
131379 Francis, D., Jackson, M. G., 2016a. Preservation of Earth-forming events in the tungsten
141380 isotopic composition of modern flood basalts. *Science* **352** (6287): 809-812.
- 151381 Rizo, H., Walker, R. J., Carlson, R. W., Touboul, M., Horan, M. F., Puchtel, I. S., Boyet, M.,
171382 Rosing, M. T., 2016b. Early Earth differentiation investigated through ^{142}Nd , ^{182}W , and
181383 highly siderophile element abundances in samples from Isua, Greenland. *Geochimica et*
191384 *Cosmochimica Acta* **175**: 319-336.
- 201385 Rudnick, R.L., and Fountain, D.M., 1995. Nature and composition of the continental crust: a
211386 lower crustal perspective. *Reviews of Geophysics* **33** (3): 267-309.
- 231387 Rudnick, R.L., and Gao, S., 2014. Composition of the Continental Crust. Treatise on
241388 Geochemistry, Macmillan Publishers Limited: 1-51.
- 251389 Ryder, G., 2002. Mass flux in the ancient Earth-Moon system and benign implications for the
261390 origin of life on Earth. *Journal of Geophysical Research* **107** (E4): 6.1-6.13.
- 271391 Saverikko, M. 1985. The pyroclastic komatiite complex at Sattasvaara in northern Finland.
291392 *Bulletin of the Geological Society of Finland* 57, 55–87.
- 301393 Schoenberg, R., Kamber, B. S., Collerson, K. D., Eugster, O., 2002. New W-isotope evidence
311394 for rapid terrestrial accretion and very early core formation. *Geochimica et Cosmochimica*
321395 *Acta* **66** (17): 3151-3160.
- 341396 Shirey, S. B., Walker, R. J., 1998. The Re-Os isotope system in cosmochemistry and high-
351397 temperature geochemistry. *Annual Reviews of Earth and Planetary Sciences* **26**: 423-500.
- 361398 Smoliar, M. I., Walker, R. J., Morgan, J. W., 1996. Re-Os ages of Group IIA, IIIA, IVA, and
371399 IVB iron meteorites. *Science* **271** (5762): 1099-1102.
- 381400 Smythe, D. J., Wood, B. J., and Kiseeva, E. S., 2017. The S content of silicate melts at sulfide
401401 saturation: New experiments and a model incorporating the effects of sulfide composition.
411402 *American Mineralogist* **102** (4): 795-803.
- 421403 Sossi, P.A., O'Neill, H.St.C., 2016. Liquidus temperatures of komatiites and the effect of
431404 cooling rate on element partitioning between olivine and komatiitic melt. *Contributions to*
451405 *Mineralogy and Petrology* **171**: 49.
- 461406 Steenstra, E.S., Berndt, J., Klemme, S., Rohrbach, A., Bullock, E.S., van Westrenen, W.,
471407 2020. An experimental assessment of the potential of sulfide saturation of the source
481408 regions of eucrites and angrites: Implications for asteroidal models of core formation, late
491409 accretion and volatile element depletions. *Geochimica et Cosmochimica Acta* **269**: 39-62.
- 511410 Steenstra, E. S., Seegers, A. X., Eising, J., Tomassen, B. G. J., Webers, F. P. F., Berndt, J.,
521411 Klemme, S., Matveev, S., van Westrenen, W., 2018. Evidence for a sulfur-undersaturated
531412 lunar interior from the solubility of sulfur in lunar melts and sulfide-silicate partitioning of
541413 siderophile elements. *Geochimica et Cosmochimica Acta* **231**: 130-156.
- 561414 Strom, R. G., Malhotra, R., Ito, T., Yoshida, F., and Kring, D. A., 2005. The origin of
571415 planetary impactors in the inner Solar system. *Science* **309** (5742): 1847-1850.
- 581416 Sun, Z., Xiong, X., Wang, J., Liu, X., Li, L., Ruan, M., Zhang, L., and Takahashi, E., 2020.
591417 Sulfur abundance and heterogeneity in the MORB mantle estimated by copper

- 1418 partitioning and sulfur solubility modelling. *Earth and Planetary Science Letters* **538**:
11419 116169.
- 21420 Thompson, P. M. E., Kempton, P. D., White, R. V., Kerr, A. C., Tarney, J., Saunders, A. D.,
31421 Fitton, J. G., and McBirney, A., 2003. Hf-Nd isotope constraints on the origin of the
41422 Cretaceous Caribbean plateau and its relationship to the Galapagos plume. *Earth and*
61423 *Planetary Science Letters* **217** (1-2): 59-75.
- 71424 Törmänen, T., Konnunaho, J., Hanski, E., Moilanen, M., Heikura, P., 2016. The
81425 Paleoproterozoic komatiite-hosted PGE mineralization at Lomalampi, Central Lapland
91426 Greenstone Belt, northern Finland. *Mineralium Deposita* **51**: 411-430.
- 101427 Touboul, M., Puchtel, I. S., Walker, R. J., 2012. ^{182}W Evidence for Long-Term Preservation
121428 of Early Mantle Differentiation Products. *Science* **335**: 1065-1069.
- 131429 Touboul, M., Liu, J., O'Neil, J., Puchtel, I. S., Walker, R. J., 2014. New Insights into the
141430 Hadean Mantle Revealed by ^{182}W and Highly Siderophile Element Abundances of
151431 Supracrustal Rocks from the Nuvvuagittuq Greenstone Belt, Quebec, Canada. *Chemical*
171432 *Geology* **383**: 63-75.
- 181433 Touboul, M., Puchtel, I. S., Walker, R. J., 2015. Tungsten isotopic evidence for
191434 disproportional late accretion to the Earth and Moon. *Nature* **520** (7548): 530-533.
- 201435 Tusch, J., Sprung, P., van de Löcht, J., Hoffmann, J. E., Boyd, A. J., Rosing, M. T., and
221436 Münker, C., 2019. Uniform ^{182}W isotope compositions in Eoarchean rocks from the Isua
231437 region, SW Greenland: The role of early silicate differentiation and missing late veneer.
241438 *Geochimica et Cosmochimica Acta* **257**: 284-310.
- 251439 van Acken, D., Brandon, A. D., and Humayun, M., 2011. High-precision osmium isotopes in
261440 enstatite and Rumuruti chondrites. *Geochimica et Cosmochimica Acta* **75** (14): 4020-
281441 4036.
- 291442 Walker, R. J., 2016. Siderophile Elements in Tracing Planetary Formation and Evolution.
301443 *Geochemical Perspectives* **5** (1): 1-145.
- 311444 Walker, R. J., Morgan, J. W., and Horan, M. F., 1995. ^{187}Os enrichment in some plumes:
321445 evidence for core-mantle interaction? *Science* **269** (5225): 819-822.
- 341446 Walker, R. J., Morgan, J. W., Beary, E. S., Smoliar, M. I., Czamanske, G. K., and Horan, M.
351447 F., 1997. Applications of the ^{190}Pt - ^{186}Os isotope system to geochemistry and
361448 cosmochemistry. *Geochimica et Cosmochimica Acta* **61** (22): 4799-4807.
- 371449 Walker, R. J., Shirey, S. B., and Stecher, O., 1988. Comparative Re-Os, Sm-Nd and Rb-Sr
391450 isotope and trace element systematics for Archean komatiite flows from Munro Township,
401451 Abitibi belt, Ontario. *Earth and Planetary Science Letters* **87** (1-2): 1-12.
- 411452 Walker, R. J., Horan, M. F., Morgan, J. W., Becker, H., Grossman, J. N., Rubin, A. E., 2002.
421453 Comparative ^{187}Re - ^{187}Os systematics of chondrites: Implications regarding early solar
431454 system processes. *Geochimica et Cosmochimica Acta* **66** (23): 4187-4201.
- 451455 Walker, R. J., Bermingham, K., Liu, J., Puchtel, I. S., Touboul, M., and Worsham, E. A.,
461456 2015. In search of late-stage planetary building blocks. *Chemical Geology* **411**: 125-142.
- 471457 Wasson, J. T., 1999. Trapped melt in IIIAB irons; solid/liquid elemental partitioning during
481458 the fractionation of the IIIAB magma. *Geochimica et Cosmochimica Acta* **63** (18): 2875-
501459 2889.
- 511460 White, W. M. and Hofmann, A. W., 1982. Sr and Nd isotope geochemistry of oceanic basalts
521461 and mantle evolution. *Nature* **296** (5860): 821-825.
- 531462 White, W. M. and Patchett, J., 1984. Hf-Nd-Sr isotopes and incompatible element abundances
541463 in island arcs: implications for magma origins and crust-mantle evolution. *Earth and*
561464 *Planetary Science Letters* **67** (2): 167-185.
- 571465 Willbold, M., Elliott, T., Moor bath, S., 2011. The tungsten isotopic composition of the Earth's
581466 mantle before the terminal bombardment. *Nature* **477** (7363): 195-198.

1467 Willbold, M., Mojzsis, S. J., Chen, H. W., Elliott, T., 2015. Tungsten isotope composition of
11468 the Acasta Gneiss Complex. *Earth and Planetary Science Letters* **419**: 168-177.
21469 Wood, B. J. and Kiseeva, E. S., 2015. Trace element partitioning into sulfide: How lithophile
31470 elements become chalcophile and vice versa. *American Mineralogist* **100** (11-12): 2371-
41471 2379.
61472 Yang, S.-H., Hanski, E., Li, C., Maier, W. D., Huhma, H., Mokrushin, A. V., Latypov, R.,
71473 Lahaye, Y., O'Brien, H., and Qu, W.-J., 2016. Mantle source of the 2.44-2.50-Ga mantle
81474 plume-related magmatism in the Fennoscandian Shield: evidence from Os, Nd, and Sr
91475 isotope compositions of the Monchepluton and Kemi intrusions. *Mineralium Deposita* **51**:
101476 1055-1073.
121477 Yin, Q., Jacobsen, S. B., Yamashita, K., Blichert-Toft, J., Telouk, P., Albarede, F., 2002. A
131478 short timescale for terrestrial planet formation from Hf-W chronometry of meteorites.
141479 *Nature* **418** (6901): 949-952.
151480 Zindler, A., Jagoutz, E., Goldstein, S. L., 1982. Nd, Sr and Pb isotopic systematics in a three-
161481 component mantle: a new perspective. *Nature* **298** (5874): 519-523.

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Fig. 1. (a) Geological map showing the distribution of the 2.05 Ga komatiites and related picrites in the Central Lapland Greenstone Belt, Finland (modified after Hanski et al., 2001), and the locations of the Jeesiörova komatiites and the Kevitsa mafic-ultramafic intrusion. (b) Geological map of the Kevitsa intrusion (modified after Mutanen and Huhma, 2001). (c) Location of drill hole DDH 814 and other drill holes penetrating Kevitsa komatiite dikes (red dots).

Fig. 2. Photomicrographs of olivine-phyric Kevitsa komatiite dikes. (a) Sample showing various olivine morphologies. (b) Olivine phenocrysts in a groundmass composed of clinopyroxene, plagioclase, olivine, brown amphibole, and oxide. (c) Close-up view of the groundmass. (d) Large olivine phenocryst with ragged grain boundaries. (e) Thin olivine-phyric vein injected into olivine-pyroxene cumulate of the Kevitsa intrusion. (f) Chilled margin with small olivine and clinopyroxene phenocrysts set in a fine-grained groundmass. Am = brown amphibole, Pl = plagioclase, Ol = olivine, Cp = clinopyroxene. Plane-polarized light in all panels except for (c).

Fig. 3. Thin section scanner micrographs in transmitted light showing olivine megacrysts and phenocrysts in the Kevitsa komatiite. Note the large chromite grains attached to the megacryst in (a).

Fig. 4. Abundances of selected major (wt.%) and lithophile trace (ppm) elements and W (ppb) plotted against MgO (wt.%) in the Jeesiörova and Kevitsa komatiites.

Fig. 5. BSE-normalized abundances of lithophile trace elements in the Jeesiörova (a) and Kevitsa (b) komatiites arranged in order of decreasing incompatibility during mantle melting. The normalizing value for W is from Arevalo and McDonough (2008) and, for the other elements, from Hofmann (1988).

Fig. 6. (a) Re-Os isochron diagram for whole-rock samples and olivine and chromite separates from the Jeesiörova and Kevitsa komatiites analyzed in this study. (b) Pt-Os isotopic data for chromite separates from the Jeesiörova komatiites.

Fig. 7. CI chondrite-normalized HSE abundances in whole-rock samples of the (a) Jeesiörova komatiites, (b) Kevitsa komatiites, and (c) olivine separates from the Kevitsa komatiites.

Fig. 8. Abundances of HSE in the Jeesiörova and Kevitsa komatiites and olivine separates (in ppb) plotted against MgO contents (wt.%).

Fig. 9. (a) ^{147}Sm - ^{143}Nd and (b) ^{176}Lu - ^{176}Hf isochron diagrams for the Jeesiörova-Kevitsa komatiites and clinopyroxene separates. See text for additional details. (c) Diagram illustrating average initial $\epsilon^{143}\text{Nd}$ and $\epsilon^{176}\text{Hf}$ values for the emplaced Jeesiörova-Kevitsa komatiites (light-green symbol) and their mantle source (dark-green symbol), as well as for the Chondritic Uniform Reservoir (CHUR) and Depleted MORB Mantle (DMM) calculated at the time of the emplacement (2049 Ma). The CHUR and DMM parameters

are from Jacobsen and Wasserburg (1980), Bouvier *et al.* (2008), and Blichert-Toft and Puchtel (2010). Note the coupled, or congruent, behavior between the two isotope systems in the source of the Jeesiörova-Kevitsa komatiites, after correction of the komatiite data for 1.0% crustal contamination.

Fig. 10. (a) Initial $^{187}\text{Os}/^{188}\text{Os}$ isotopic compositions, expressed in terms of $\gamma^{187}\text{Os}$, of Archean and Paleoproterozoic komatiite systems plotted as a function of age. The data are from Foster *et al.* (1996); Puchtel *et al.* (2004a; 2005; 2007; 2009a; 2009b; 2014; 2016a; 2016b; 2018), and this study. The data for chondritic meteorites are compiled from Walker *et al.* (2002), Brandon *et al.* (2005), and Fischer-Gödde *et al.* (2010). **(b)** Initial $^{186}\text{Os}/^{188}\text{Os}$ isotopic compositions, expressed in terms of $\mu^{186}\text{Os}$, of Archean and Paleoproterozoic komatiite systems studied to date, plotted as a function of age. The data are from Puchtel *et al.* (2004a, 2005, 2009b, 2014, 2016b; 2018), and this study. The data for chondritic meteorites are compiled from Horan *et al.* (2003), Brandon *et al.* (2005, 2006), and Fischer-Gödde *et al.* (2010). See text for more details.

Fig. 11. Calculated total Pt and Pd abundances in the sources of Archean and Paleoproterozoic komatiite systems plotted as per cent of the total Pt and Pd abundances in the estimates for the modern BSE of Becker *et al.* (2006). Data sources are as follows: 2.05 Ga Jeesiörova-Kevitsa - this study; 2.41 Ga Vetryny Belt - Puchtel *et al.* (2016b); 2.69 Ga Belingwe - Puchtel *et al.* (2009b); 2.72 Ga Pyke Hill and Alexo - Puchtel *et al.* (2004b); 2.72 Ga Boston Creek - Puchtel *et al.* (2018); 2.82 Ga Kostomuksha - Puchtel and Humayun (2005); 2.88 Ga Volotsk-Kamennoozero - Puchtel *et al.* (2007); 3.26 Ga Weltevreden and 3.48 Ga Komati - Puchtel *et al.* (2014); 3.55 Ga Schapenburg - Puchtel *et al.* (2016a). Uncertainties are 2SD. See text for additional details.

Fig. 12. Diagram illustrating potential genetic relationships between long-term $^{187}\text{Re}/^{188}\text{Os}$ and $^{190}\text{Pt}/^{188}\text{Os}$ in IIIAB fractionated iron meteorites and in the Jeesiörova-Kevitsa (J-K) komatiite source. J-K source 1 and 2 denote the calculated Re/Os and Pt/Os in the source after and before the global melt depletion event, respectively, that is assumed to have occurred within the first 100 Ma of Solar System history and is necessary to account for the Nd-Hf isotope systematics of the Jeesiörova-Kevitsa komatiites. The data for the IIIAB iron meteorite Tieraco Creek used in the modeling are from Cook *et al.* (2004). Addition of 0.2% of Tieraco Creek metal to the mantle domain with BSE Re/Os and Pt/Os *prior* to the global melt depletion event can explain the $^{186,187}\text{Os}/^{188}\text{Os}$ isotope systematics of the Jeesiörova-Kevitsa komatiite source. See text for more details.

Fig. 13. $\mu^{182}\text{W}$ (ppm) versus total calculated HSE abundances in the sources of Archean and Paleoproterozoic komatiite systems studied to date relative to those in the estimates for the present-day BSE of Becker *et al.* (2006). This proportion corresponds to the fraction of the total HSE budget of the BSE added during late accretion assuming an HSE-free terrestrial mantle prior to late accretion. The $\mu^{182}\text{W}$ value of the BSE prior to late accretion is constrained via mass-balance calculations and is supported by the W isotopic data for the lunar mantle ($\mu^{182}\text{W} = +25 \pm 5$ ppm: Kruijer *et al.*, 2015; Touboul *et al.*, 2015).

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The W isotopic data and estimates of the HSE contents for the komatiite systems are from Puchtel and Humayun (2005), Touboul et al. (2012), Puchtel et al. (2014; 2016a; 2016b; 2018), and this study.

Fig. 14. $\mu^{182}\text{W}$ versus $\mu^{186}\text{Os}$ data (ppm) for the Jeesiörova-Kevitsa komatiites and Hawaiian picrites. The Os data for Hawaii are compiled from Brandon et al. (1998) and Ireland et al. (2011), and the W data - from Mundl et al. (2017) and Mundl-Petermeier et al. (2020).

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Table 1. Major and minor element data for Jeesiörova and Kevitsa komatiites.

Sample	12D-PPR	13-EJH	17.1-PPR	LP10	KD03	KD05	KD06	KD07	KD08	KD09	KD10	KD11	KD12	KD13	KD14
SiO ₂	44.8	43.8	43.0	43.6	46.6	45.0	44.9	44.6	44.1	43.4	43.5	44.1	45.1	45.0	46.9
TiO ₂	0.691	0.632	0.681	0.583	0.730	0.612	0.628	0.533	0.554	0.488	0.487	0.582	0.662	0.679	0.778
Al ₂ O ₃	7.70	7.47	8.01	6.94	10.1	8.55	8.74	7.34	7.94	6.50	6.35	8.20	9.09	9.57	10.6
Fe ₂ O _{3tot}	12.0	12.4	13.7	12.1	12.4	12.6	12.6	12.0	12.3	12.2	12.3	12.4	12.7	12.9	13.0
MnO	0.169	0.221	0.227	0.301	0.190	0.179	0.177	0.180	0.176	0.178	0.183	0.178	0.181	0.184	0.190
MgO	25.3	27.1	25.7	28.6	18.1	22.6	22.3	26.4	25.2	28.3	28.3	24.2	21.2	20.3	16.2
CaO	8.21	7.33	7.51	6.86	10.2	8.74	8.88	7.42	7.88	7.18	7.16	8.41	9.14	9.41	10.6
Na ₂ O	0.077	0.074	0.092	0.090	1.21	1.13	1.15	0.92	1.11	0.94	0.88	1.15	1.25	1.28	1.29
K ₂ O	0.004	0.018	0.056	0.019	0.068	0.040	0.047	0.135	0.086	0.043	0.119	0.062	0.054	0.069	0.089
P ₂ O ₅	0.053	0.033	0.034	0.027	0.063	0.039	0.040	0.039	0.036	0.031	0.028	0.039	0.045	0.047	0.056
LOI	11.3	7.20	6.39	7.49	0.13	0.63	0.55	1.72	1.40	1.47	2.15	1.25	1.01	1.22	0.09
Total	99.85	99.81	100.00	99.55	100.00	99.90	100.22	100.05	99.68	100.21	99.88	99.88	99.92	99.84	99.85
Cr	3834	3674	3267	3806	1847	2309	2241	2895	2843	3333	3280	2719	2317	2127	1557
Al ₂ O ₃ /TiO ₂	11.1	11.8	11.8	11.9	13.9	14.0	13.9	13.8	14.3	13.3	13.0	14.1	13.7	14.1	13.7

31 1570 Data (in wt.% except for Cr in ppm) were recalculated on an anhydrous basis, but not re-normalized to 100% in order to preserve information on
32 1571 the quality of the analyses. Here and in Tables 2 and 3, samples from 12D-PPR to LP10 are Jeesiörova komatiites and those from KD03 to
33 1572 KD14 - Kevitsa komatiite dikes.
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Table 2. Trace element data for Jeesiörova and Kevitsa komatiites.

Sample	12D-PPR	13-EJH	17.1-PPR	LP-10	KD-03	KD-05	KD-06	KD-07	KD-08	KD-09	KD-10	KD-11	KD-12	KD-13	KD-14	BSE
Sc	28.8	26.4	28.4	24.8	33.6	30.4	29.3	26.8	26.6	25.0	25.3	28.4	31.2	31.7	35.0	14.9
V	217	196	207	184	277	238	238	204	210	182	190	223	245	251	294	77.0
Ni	1547	1616	1535	1734	856	1173	1138	1479	1389	1611	1624	1322	1062	1014	748	2080
Cu	141	5.22	9.45	6.01	129	114	114	96	104	89.0	89.9	108	119	123	139	28.0
Ga	10.6	9.51	10.6	8.90	13.9	11.9	11.5	10.0	10.5	9.21	9.23	11.1	12.4	12.6	14.7	4.00
Th	0.0768	0.0722	0.0769	0.0677	0.0843	0.0752	0.0746	0.0643	0.0674	0.0584	0.0583	0.0696	0.0763	0.0794	0.0888	0.0813
W	n.d.	n.d.	n.d.	26.7	n.d.	9.93	8.76	n.d.	12.7	12.1	11.6	11.4	11.9	11.8	n.d.	13.0
U	0.0281	0.0269	0.0275	0.0240	0.0247	0.0208	0.0210	0.0190	0.0199	0.0173	0.0168	0.0212	0.0221	0.0225	0.0260	0.0203
Nb	0.359	0.331	0.345	0.294	0.379	0.335	0.320	0.267	0.296	0.263	0.266	0.310	0.355	0.347	0.393	0.618
La	0.687	0.658	0.589	0.561	0.652	0.569	0.557	0.489	0.501	0.442	0.439	0.522	0.596	0.610	0.687	0.614
Ce	2.23	2.13	1.98	1.79	2.26	1.93	1.90	1.70	1.75	1.51	1.49	1.83	2.03	2.11	2.40	1.60
Pr	0.414	0.410	0.391	0.345	0.502	0.429	0.416	0.361	0.387	0.328	0.333	0.399	0.448	0.460	0.526	0.242
Nd	2.71	2.65	2.61	2.25	3.43	2.91	2.87	2.50	2.62	2.25	2.30	2.77	3.06	3.13	3.62	1.19
Sm	1.19	1.15	1.18	1.01	1.57	1.34	1.35	1.15	1.19	1.04	1.05	1.25	1.41	1.45	1.66	0.387
Hf	0.803	0.775	0.782	0.706	1.036	0.924	0.913	0.797	0.822	0.720	0.723	0.873	0.944	0.967	1.093	0.268
Zr	28.4	25.2	28.2	24.3	36.9	31.9	32.0	26.5	28.3	25.3	25.0	29.3	33.2	34.5	39.1	9.71
Eu	0.566	0.268	0.311	0.343	0.646	0.553	0.525	0.462	0.468	0.381	0.372	0.501	0.593	0.593	0.677	0.146
Gd	1.81	1.69	1.63	1.54	2.20	1.91	1.90	1.60	1.70	1.49	1.48	1.79	1.99	2.04	2.32	0.513
Tb	0.311	0.289	0.307	0.264	0.381	0.326	0.325	0.282	0.293	0.257	0.259	0.312	0.348	0.355	0.404	0.0940
Dv	1.98	1.86	1.96	1.74	2.48	2.14	2.14	1.81	1.93	1.71	1.70	2.01	2.27	2.34	2.63	0.638
Y	9.79	9.20	9.51	8.34	12.3	10.5	10.7	9.07	9.76	8.50	8.45	10.0	11.0	11.4	12.9	3.94
Ho	0.398	0.373	0.387	0.353	0.504	0.433	0.432	0.362	0.386	0.340	0.339	0.399	0.445	0.463	0.533	0.142
Er	1.09	1.01	1.07	0.955	1.37	1.17	1.18	1.00	1.05	0.919	0.915	1.09	1.23	1.27	1.44	0.417
Tm	0.148	0.140	0.142	0.132	0.186	0.157	0.161	0.138	0.143	0.126	0.126	0.152	0.167	0.172	0.196	0.0643
Yb	0.911	0.880	0.929	0.817	1.21	1.02	1.02	0.87	0.923	0.815	0.816	0.964	1.08	1.11	1.26	0.414
Lu	0.129	0.127	0.129	0.117	0.170	0.145	0.147	0.125	0.132	0.115	0.116	0.137	0.154	0.158	0.179	0.0637
Ti/Zr	146	151	145	144	119	115	118	120	118	115	116	119	120	118	119	112
Nb/Nb*	0.566	0.550	0.586	0.546	0.585	0.587	0.568	0.545	0.582	0.593	0.601	0.588	0.602	0.570	0.576	1.00
W/W*	n.d.	n.d.	n.d.	2.07	n.d.	0.785	0.691	n.d.	1.09	1.19	1.16	0.931	0.902	0.874	n.d.	1.00
(La/Sm) _N	0.362	0.359	0.315	0.350	0.261	0.267	0.261	0.268	0.264	0.268	0.264	0.263	0.267	0.265	0.260	1.00
(Gd/Yb) _N	1.60	1.56	1.41	1.53	1.47	1.51	1.49	1.48	1.49	1.48	1.47	1.50	1.49	1.48	1.48	1.00

38 1575 Abundances recalculated on an anhydrous basis in ppm, except for W (ppb). *N*- normalized to the BSE values of Hofmann (1988) for all
 39 1576 elements except Ga (McDonough and Sun, 1995) and W (Arevalo and McDonough, 2008), as reported in the table.
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Table 3. Highly siderophile element abundances (in ppb) and Re-Os isotopic data for Jeesiörova and Kevitsa komatiites and olivine and chromite separates.

Sample No.	Re	Os	Ir	Ru	Pt	Pd	¹⁸⁷ Re/ ¹⁸⁸ Os	¹⁸⁷ Os/ ¹⁸⁸ Os	$\gamma^{187}\text{Os}(2049)$
12D-PPR WR	0.9867	5.434	4.226	7.436	17.89	15.35	0.8766±0.0035	0.14236±7	-1.0
13-EJH WR	0.6650	4.269	3.279	6.235	18.26	14.87	0.7517±0.0030	0.13836±6	-0.7
17.1-PPR WR	0.8497	3.213	2.637	5.869	19.17	15.22	1.280±0.005	0.15817±7	+0.6
LP-10 WR	0.3756	4.435	3.631	6.657	16.75	13.93	0.4082±0.0016	0.12871±6	+1.3
KD-03 WR	1.188	1.528	1.472	2.573	22.31	19.01	3.805±0.015	0.24505±2	-0.1
KD-05 WR	1.007	2.149	1.845	3.549	19.77	16.92	2.277±0.009	0.19117±8	-0.9
KD-06 WR	1.010	5.528	3.006	3.824	19.57	16.49	0.8817±0.0035	0.14353±7	-0.1
KD-07 WR	0.8141	4.161	2.571	4.082	16.82	14.72	0.9452±0.0038	0.14662±1	+0.7
KD-08 WR	0.8693	3.684	2.824	4.300	17.77	15.32	1.141±0.005	0.15200±8	-0.6
KD-09 WR	0.7062	3.956	2.870	5.001	16.02	13.74	0.8620±0.0035	0.14307±6	+0.1
KD-10 WR	0.6813	3.190	2.434	5.027	16.22	13.60	1.032±0.004	0.14820±7	-0.6
KD-11 WR	0.9146	3.701	2.708	4.262	19.20	15.87	1.195±0.005	0.15471±7	+0.1
KD-12 WR	1.051	2.297	1.753	3.545	20.43	17.04	2.223±0.009	0.19087±10	+0.5
KD-13 WR	1.127	3.456	2.166	3.436	21.10	17.97	1.579±0.006	0.16626±10	-1.4
<i>Replicate</i>	1.109	2.063	1.933	3.384	21.04	17.76	2.617±0.010	0.20361±9	-0.3
KD-14 WR	1.284	2.035	1.593	2.168	23.68	20.01	3.076±0.012	0.21969±2	-0.2
12D-PPR Chr	1.195	122.7					0.0469±0.0004	0.11378±5	-0.8
17.1-PPR Chr	1.233	11.13					0.5341±0.0044	0.13226±8	+0.6
12D-PPR_1 Chr	1.688	156.7	102.2		27.85		0.0518±0.0010	0.11462±7	-0.2
12D-PPR_2 Chr	0.8898	141.0	94.4		33.03		0.0304±0.0010	0.11357±6	-0.5
12D-PPR_3 Chr	0.6325	149.8	102.5		36.05		0.0203±0.0009	0.11336±6	-0.4
12D-PPR_4 Chr	0.5442	153.3	103.6		38.92		0.0171±0.0009	0.11339±6	-0.2
12D-PPR_5 Chr	0.6303	150.2	101.7		47.32		0.0202±0.0009	0.11329±7	-0.4
12D-PPR_6 Chr	0.5624	157.5	108.0		52.97		0.0172±0.0009	0.11337±5	-0.3
13-EJH_1 Chr	0.2597	53.32	38.33		6.164		0.0234±0.0027	0.11394±6	+0.1
13-EJH_2 Chr	0.3449	53.59	38.45		7.314		0.0310±0.0026	0.11393±7	-0.2
KD-05 Ol	0.04468	3.952	2.262	2.296	1.525	1.426	0.0544±0.0009	0.11455±4	-0.4
KD-10 Ol	0.6178	3.126	2.285	4.158	1.936	1.698	0.9547±0.0050	0.14627±4	0.0
KD-05 Chr	0.4528	35.29					0.0617±0.0016	0.11497±6	-0.2
KD-08 Chr	0.5359	15.54					0.1660±0.0023	0.11837±8	-0.4
KD-10 Chr	1.197	78.14					0.0737±0.0011	0.11531±7	-0.3
KD-12 Chr	0.7288	7.693					0.4568±0.0062	0.12945±13	0.0
LP-10 Chr	1.014	55.28					0.0883±0.0019	0.11655±7	-0.2

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HSE abundances recalculated on an anhydrous basis. Initial $\gamma^{187}\text{Os}$ values were calculated at $T = 2049$ Ma derived from the isochron using the parameters specified in the text. The HSE normalizing values are from Horan et al. (2003). *Replicate* – replicate digestion of the sample. Chromite separates 12D-PPR_1 through 12D-PPR_6 and 13-EJH_1 and 13-EJH_2 denote spiked subsamples of unspiked digestions for the Pt-Os study.

Table 4. High-precision Os isotopic data and Re/Os and Pt/Os ratios for chromite separates from the Jeesiörova komatiites.

Sample No.	$^{187}\text{Re}/^{188}\text{Os}$	$^{187}\text{Os}/^{188}\text{Os}$	$\gamma^{187}\text{Os}(T)$	$^{190}\text{Pt}/^{188}\text{Os}$	$^{184}\text{Os}/^{188}\text{Os}$	$^{186}\text{Os}/^{188}\text{Os}$	$\mu^{186}\text{Os}(T)$
12D-PPR_1 Chr	0.0518±0.0010	0.1146036±06	-0.23	0.0001697±34	0.0013031±2	0.1198369±6	+25±5
12D-PPR_2 Chr	0.0304±0.0010	0.1135064±05	-0.54	0.0002238±45	0.0013028±2	0.1198381±5	+34±4
12D-PPR_3 Chr	0.0203±0.0009	0.1134565±07	-0.27	0.0002299±46	0.0013023±3	0.1198377±7	+30±6
12D-PPR_4 Chr	0.0171±0.0009	0.1134060±05	-0.22	0.0002425±48	0.0013033±2	0.1198381±5	+33±4
12D-PPR_5 Chr	0.0202±0.0009	0.1134294±06	-0.29	0.0003009±60	0.0013033±2	0.1198378±6	+29±5
12D-PPR_6 Chr	0.0172±0.0009	0.1134048±06	-0.22	0.0003213±64	0.0013033±2	0.1198376±6	+27±5
13-EJH Chr	0.0272±0.0019	0.1138779±11	-0.11	0.0001204±24	0.0013024±4	0.1198370±10	+27±8

Initial $\gamma^{187}\text{Os}$ and $\mu^{186}\text{Os}$ values were calculated at $T = 2049$ Ma using the parameters specified in the text.

1584 **Table 5.** Tungsten isotopic compositions of Kevitsa komatiites.

Sample	$\mu^{182}\text{W}$ ($\pm 2\text{SE}$)	$\mu^{183}\text{W}$ ($\pm 2\text{SE}$)
KD-08	-1.7 \pm 3.6	-0.8 \pm 4.3
<i>Replicate</i>	+2.2 \pm 4.3	-2.1 \pm 4.8
KD-12	+5.9 \pm 3.0	+2.1 \pm 3.6
<i>Replicate</i>	-0.2 \pm 3.3	-2.6 \pm 4.1
Average ($\pm 2\text{SE}$)	+1.5\pm3.3	-0.8\pm2.1

131585 Uncertainties on W isotopic compositions for individual samples are the 2SE in-run
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 151586 uncertainties of the individual analyses.

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1588 **Table 6.** Sm-Nd isotope and concentration data for Jeesiörova and Kevitsa komatiites.

Sample	Sm, ppm	Nd, ppm	¹⁴⁷ Sm/ ¹⁴⁴ Nd	±2SD	¹⁴³ Nd/ ¹⁴⁴ Nd	±2SE	ε ¹⁴³ Nd(T)
13-EJH WR	1.081	2.446	0.2673	0.0005	0.513768	0.000004	+3.5
17.1-PPR WR	1.123	2.445	0.2779	0.0006	0.513917	0.000003	+3.6
KD-05 WR	1.379	2.966	0.2811	0.0006	0.513959	0.000004	+3.6
KD-08 WR	1.176	2.546	0.2794	0.0006	0.513950	0.000003	+3.8
KD-10 WR	0.9099	1.950	0.2821	0.0006	0.513981	0.000003	+3.7
KD-12 WR	1.501	3.245	0.2798	0.0006	0.513946	0.000003	+3.7
13-EJH Cpx	0.9271	1.542	0.3636	0.0007	0.515069	0.000003	+3.5
17.1-PPR Cpx	0.9256	1.496	0.3743	0.0008	0.515217	0.000003	+3.6
KD-05 Cpx	0.6775	1.309	0.3131	0.0006	0.514395	0.000003	+3.7
KD-08 Cpx	0.6564	1.335	0.2975	0.0006	0.514177	0.000004	+3.5
KD-10 Cpx_1	0.6936	1.368	0.3066	0.0006	0.514317	0.000004	+3.9
KD-10 Cpx_2	0.7036	1.388	0.3066	0.0006	0.514316	0.000003	+3.8
KD-12 Cpx	0.8599	1.676	0.3104	0.0006	0.514366	0.000003	+3.8

Initial ε¹⁴³Nd values were calculated at $T = 2049$ Ma using the parameters specified in the text. WR - whole-rock samples; Cpx - clinopyroxene separates.

1592 **Table 7.** Lu-Hf isotope and concentration data for the Jeesiörova komatiites and USGS SRM
 1593 BCR-1.

Sample	Lu (ppm)	Hf (ppm)	$^{176}\text{Lu}/^{177}\text{Hf}$	$\pm 2\text{SE}$	$^{176}\text{Hf}/^{177}\text{Hf}$	$\pm 2\text{SE}$	$\epsilon^{176}\text{Hf}(\text{T})$
13-EJH WR	0.1742	1.054	0.02343	0.00001	0.282630	0.000003	+8.6
13-EJH Cpx	0.1326	0.4958	0.03791	0.00001	0.283201	0.000004	+8.8
BCR-1	0.4980	4.949	0.01426	0.00001	0.282872	0.000004	
BCR-1*	0.4988 \pm 0.0048	4.92 \pm 0.06			0.282867	0.000015	

1594 Initial $\epsilon^{176}\text{Hf}$ values were calculated at $T = 2049$ Ma, using the parameters specified in the
 1595 text. *Denotes the *GeoRem* recommended values.

Table 8. Compositions of the emplaced Jeesiörova-Kevitsa komatiite magma, crustal contaminant endmember, and the calculated original komatiite magma.

	1. Emplaced komatiite magma	2. Crustal contaminant	3. Original komatiite magma
SiO₂	44.1	68.2	43.9
TiO₂	0.554	0.335	0.556
Al₂O₃	7.94	17.1	7.85
Fe₂O_{3_{tot}}	12.3	3.28	12.4
MnO	0.176	0.048	0.177
MgO	25.2	1.35	25.4
CaO	7.88	4.01	7.92
Th	0.0674	3.51	0.0327
W	11.7	595	5.81
U	0.0199	0.980	0.0102
Nb	0.296	3.06	0.268
La	0.501	25.0	0.253
Ce	1.75	46.2	1.30
Pr	0.387	4.99	0.340
Nd	2.62	17.2	2.48
Sm	1.19	2.57	1.18
Hf	0.822	3.81	0.792
Gd	1.70	2.01	1.70
Tb	0.293	0.243	0.293
Dy	1.93	1.33	1.94
Ho	0.386	0.252	0.387
Er	1.05	0.704	1.06
Tm	0.143	0.0987	0.144
Yb	0.923	0.652	0.926
Lu	0.132	0.0970	0.133
Nb/Nb*	0.582	0.118	1.07
W/W*	1.00	1.00	1.00
(Nb/Th)_N	0.577	0.115	1.08
(La/Sm)_N	0.264	6.14	0.135
(Gd/Yb)_N	1.49	2.50	1.48
ε¹⁴³Nd(<i>T</i>)	+3.7±0.3	-14.5±1.0	+4.9±0.5
ε¹⁷⁶Hf(<i>T</i>)	+8.7±0.3	-21.8±1.9	+10.2±0.6
μ¹⁸²W	+1.5±3.3	+12.6±4.5	-10±5

1- emplaced magma composition for the Jeesiörova-Kevitsa komatiites; 2 - FTA composition compiled from the average Vodla Block tonalite composition of Puchtel et al. (2016b). 3 - original parental komatiite magma for the Jeesiörova-Kevitsa komatiites. The initial εNd and εHf values were calculated at $T = 2049$ Ma.

Table 9. Results of mixing calculations between an IIIAB iron meteorite endmember and the BSE mantle domain.

	Tieraco Creek	BSE	BSE+0.2% TC	Mantle source
Os, ppb	31.7	3.40±0.50	3.50±0.50	3.39±0.10
Re, ppb	3.10	0.289±0.007	0.295±0.010	0.285±0.010
Pt, ppb	2089	6.1±1.6	10.3±1.6	9.54±0.22
W, ppb	200	13.0±10	13.4±10	5.8±5
¹⁸⁷ Re/ ¹⁸⁸ Os	0.464	0.402	0.403	0.397±0.008
¹⁹⁰ Pt/ ¹⁸⁸ Os	0.0630	0.00171	0.00284	0.00269±0.008
$\mu^{182}\text{W}$	-335	0	-10	-10±5

The composition of the Jeesiörova-Kevitsa komatiite mantle source is calculated to have resulted from mixing of 0.2% of material of the IIIAB iron meteorite Tieraco Creek into a mantle domain with the BSE composition. Uncertainties on the HSE abundances in the average estimate for the BSE are 1SE (Becker et al., 2006).

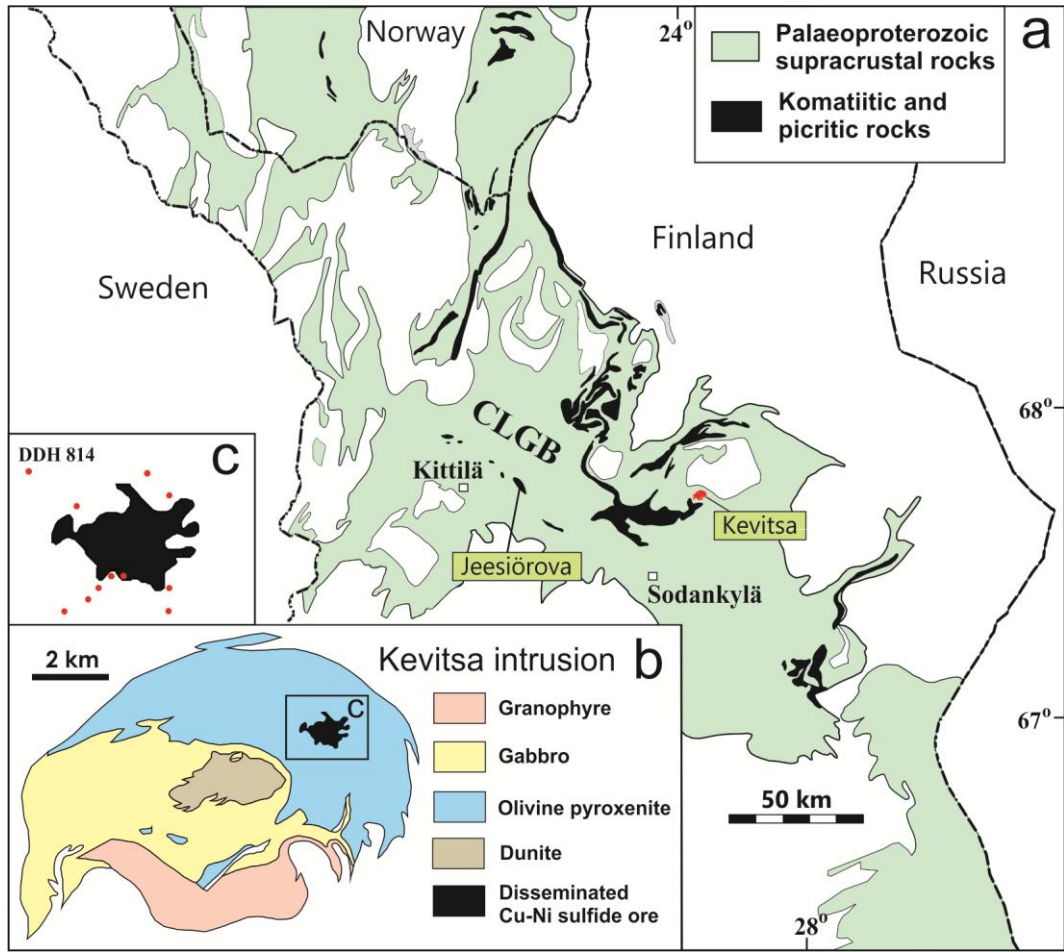


Fig. 1.

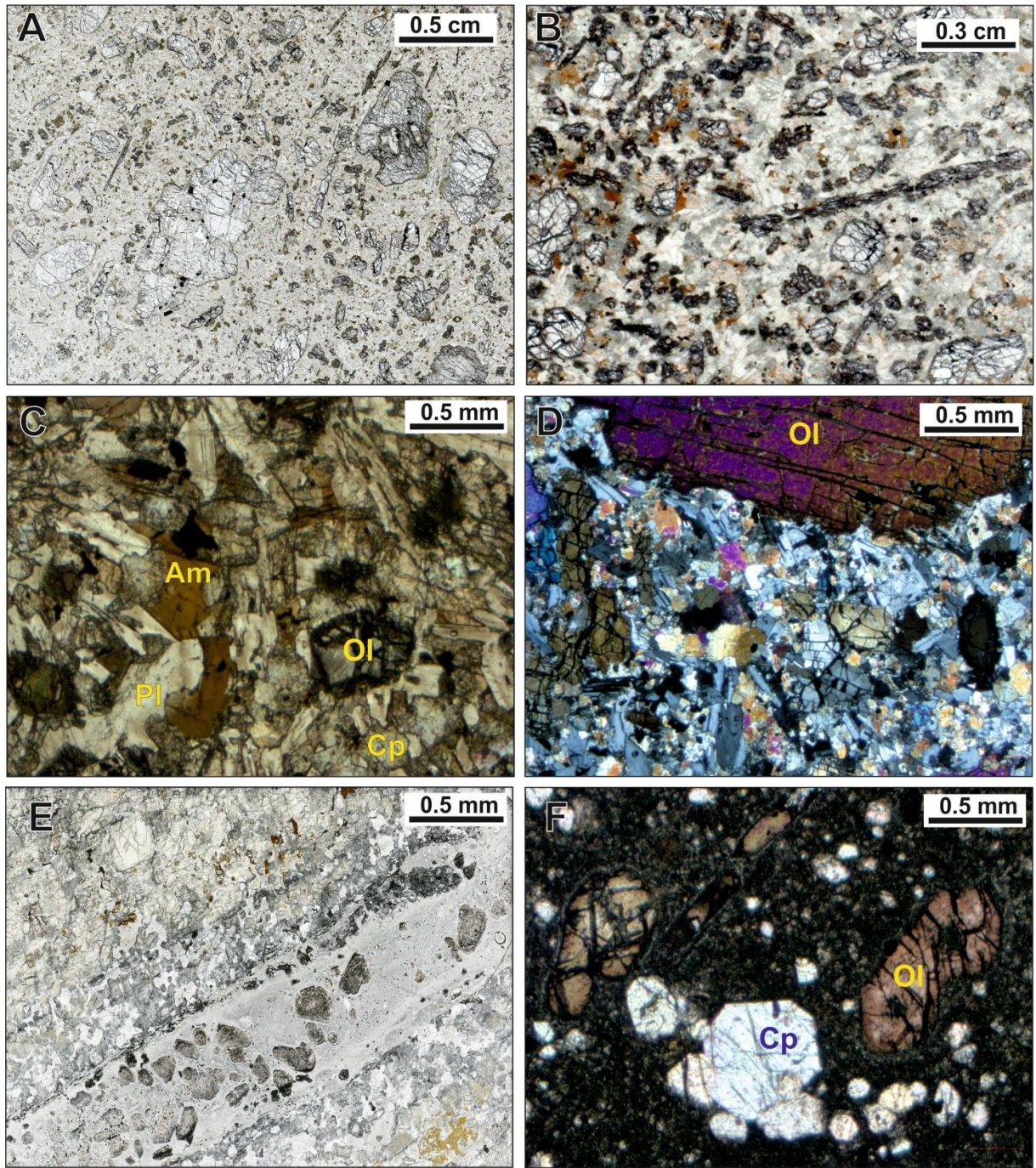


Fig. 2.

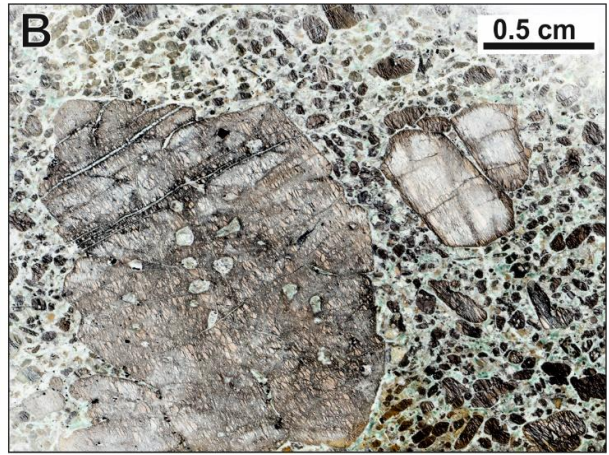
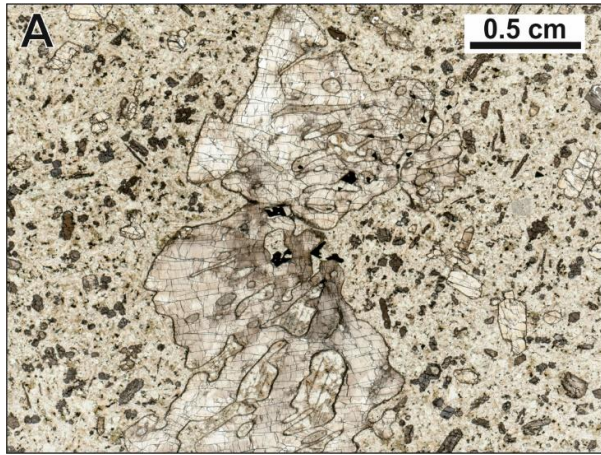


Fig. 3.

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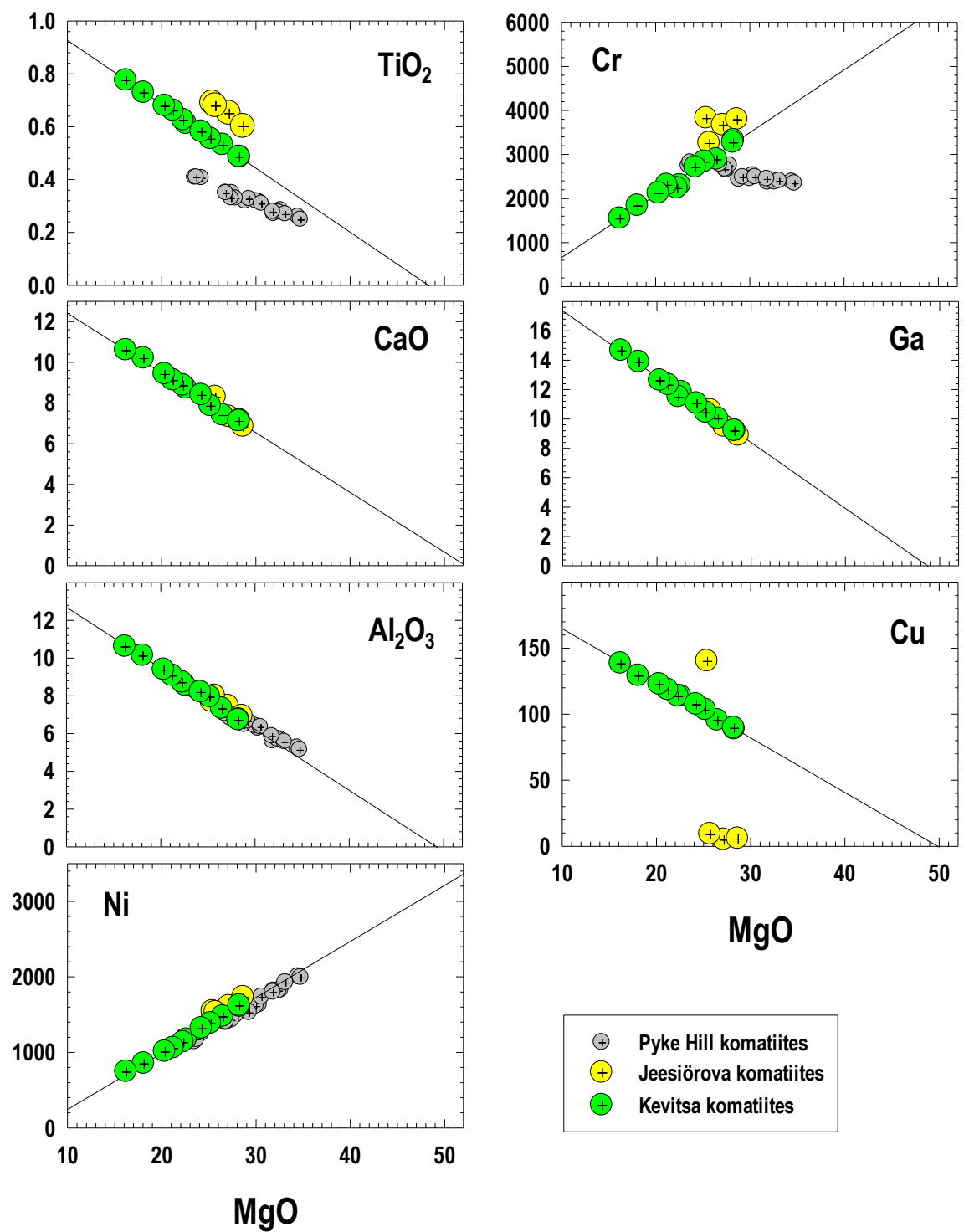


Fig. 4A.

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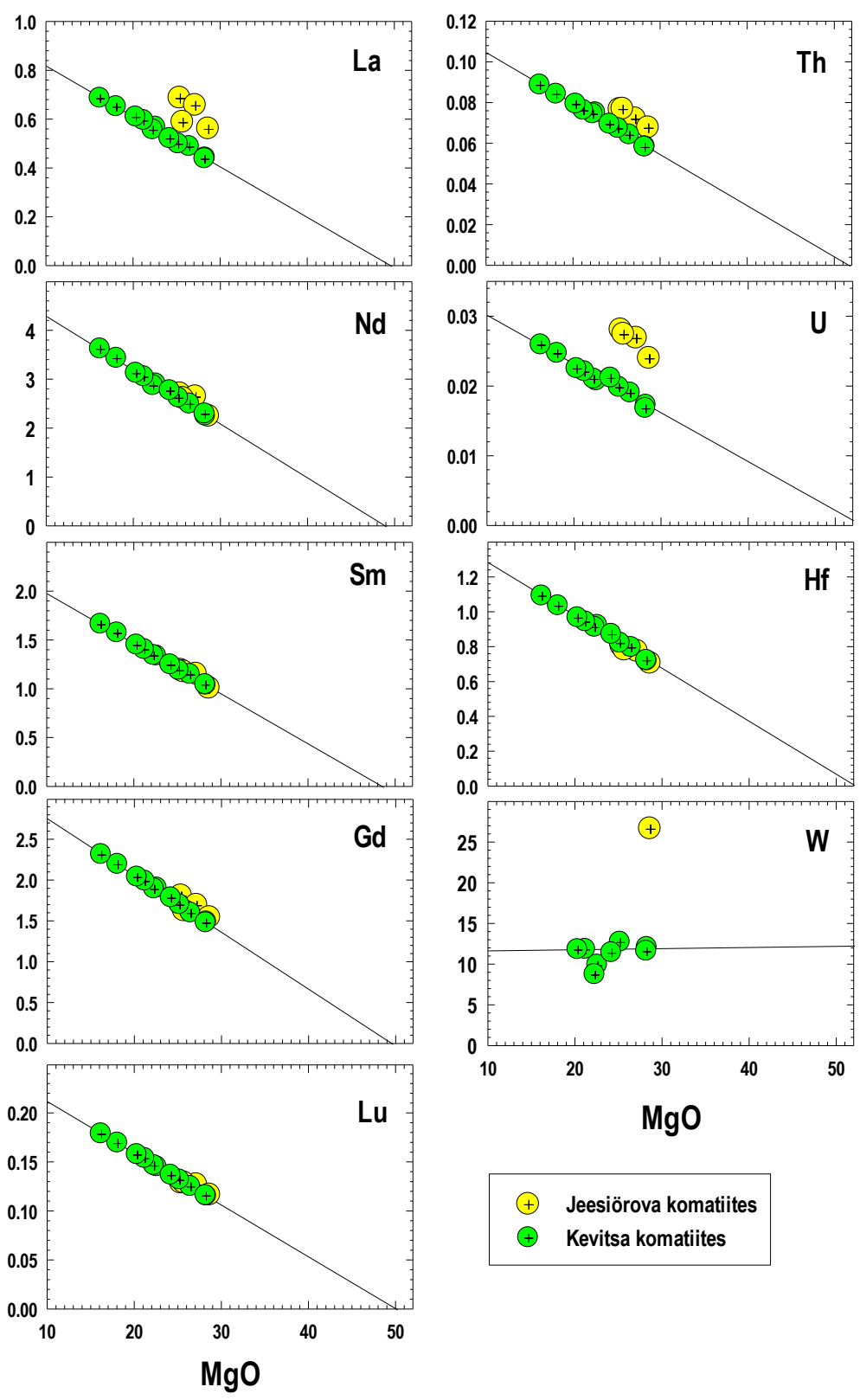


Fig. 4B.

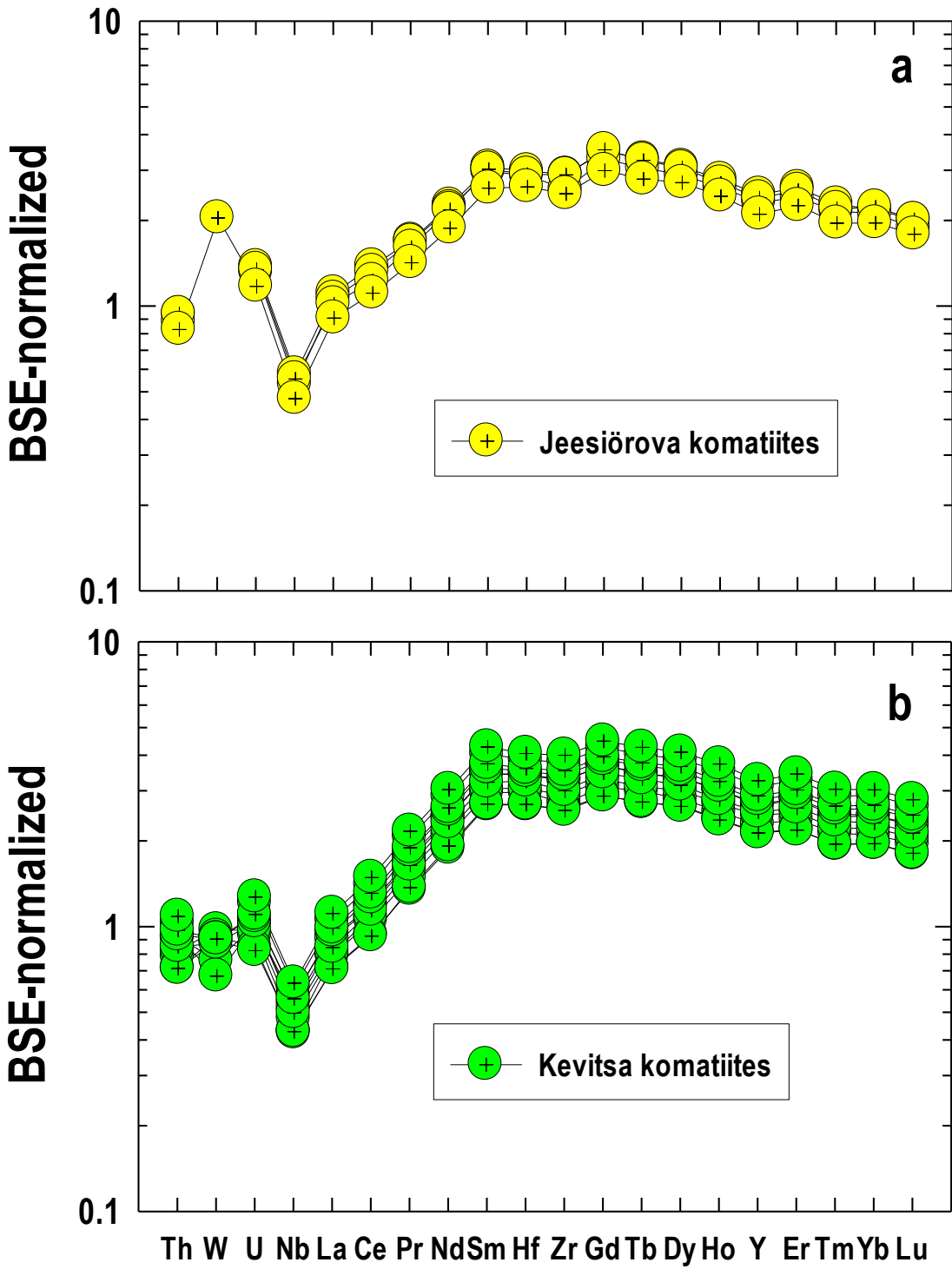


Fig. 5.

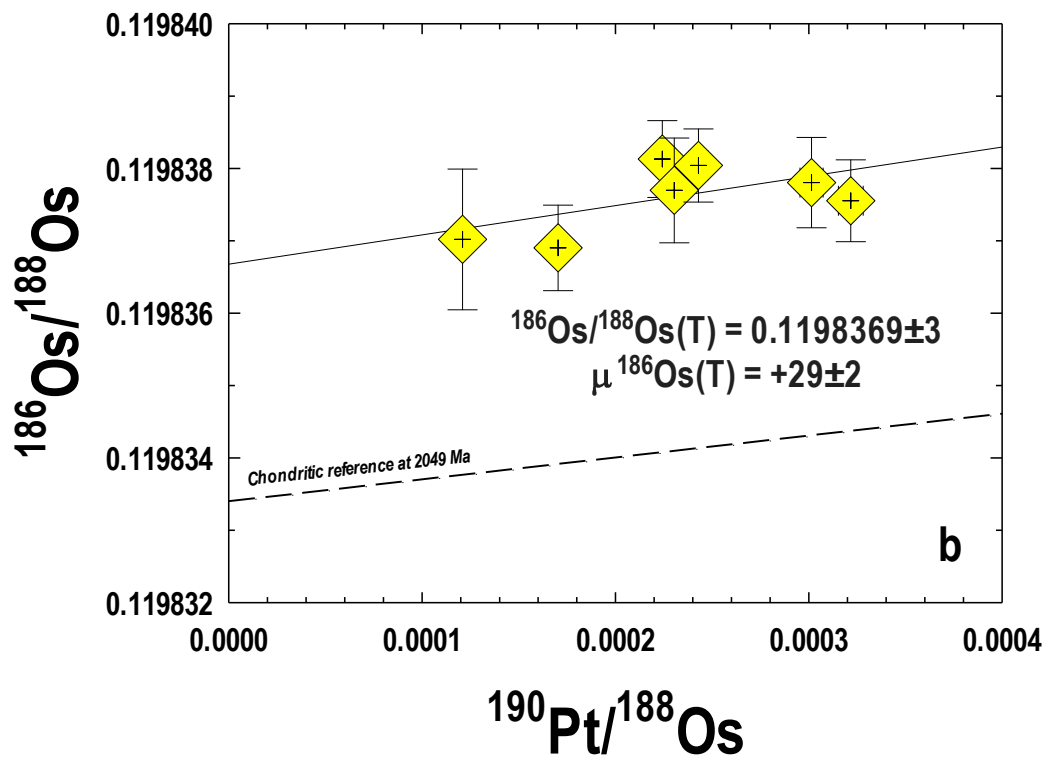
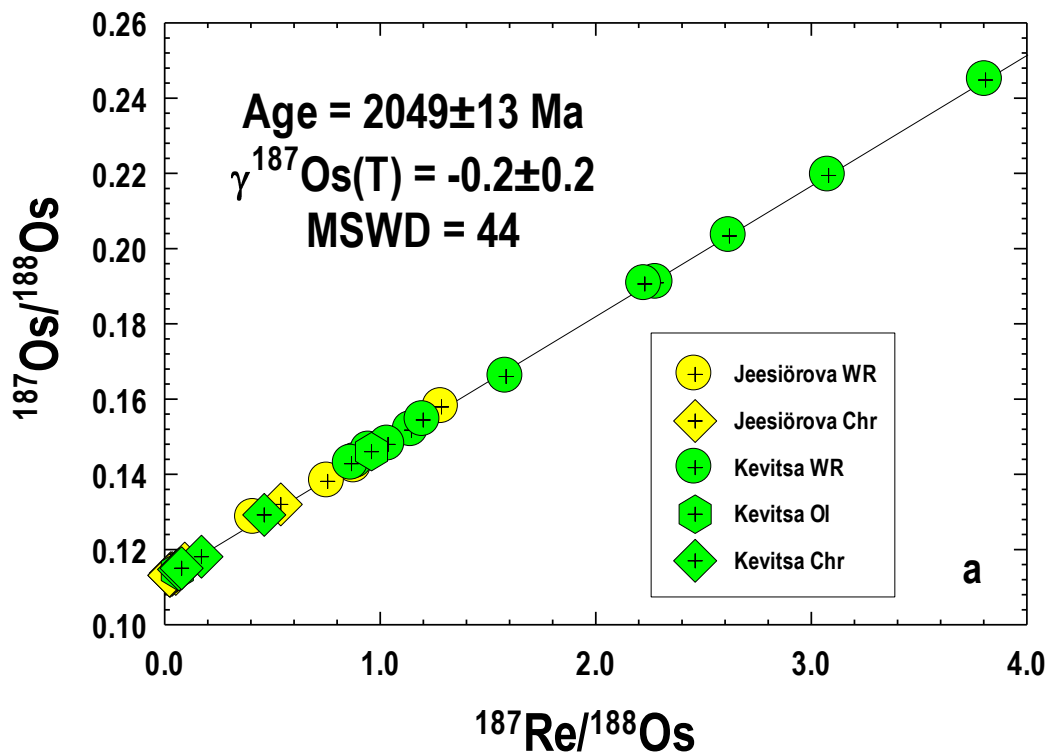


Fig. 6.

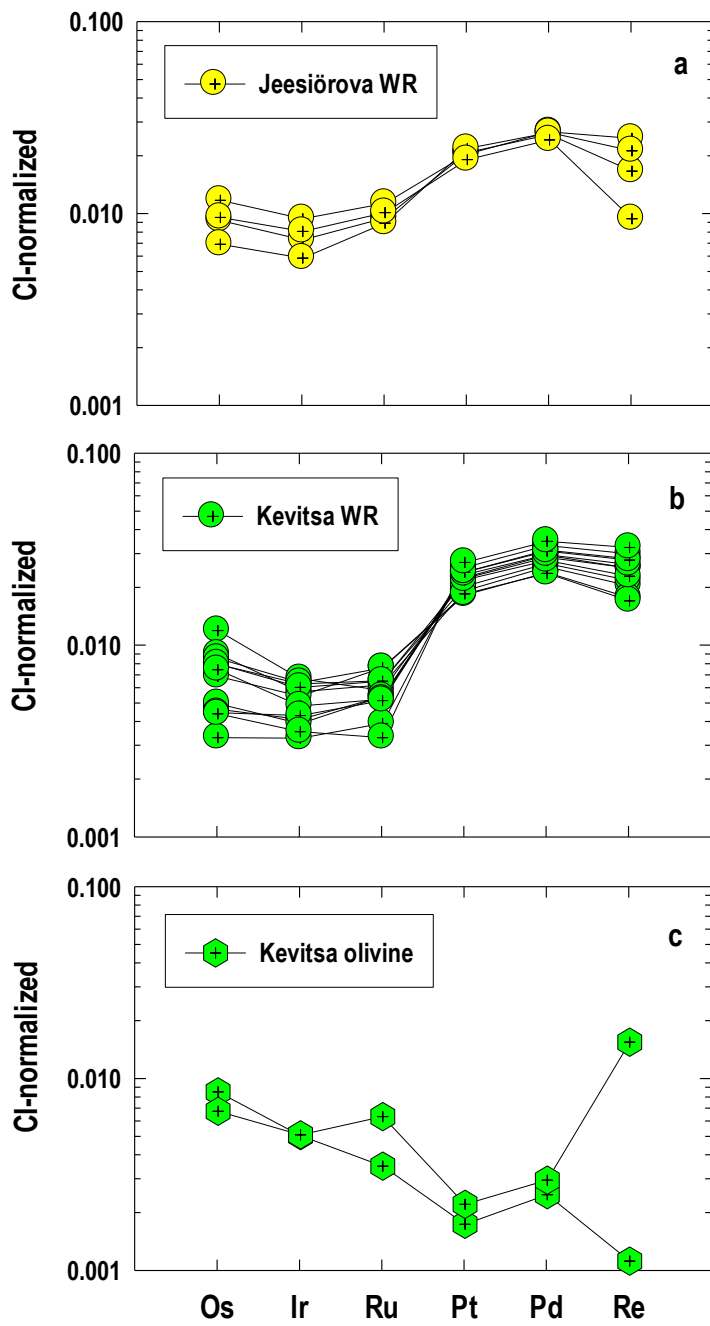


Fig. 7.

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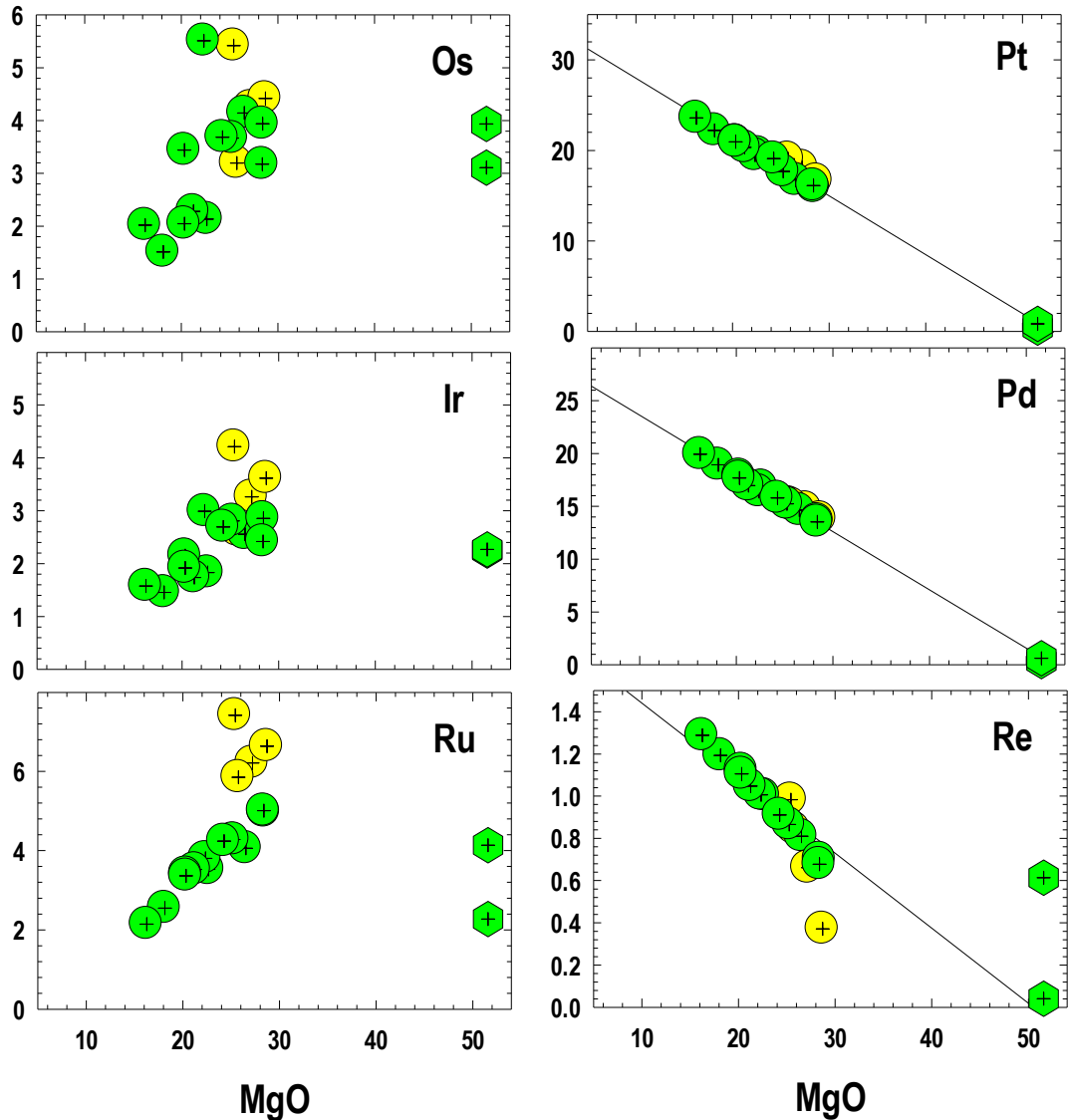


Fig. 8.

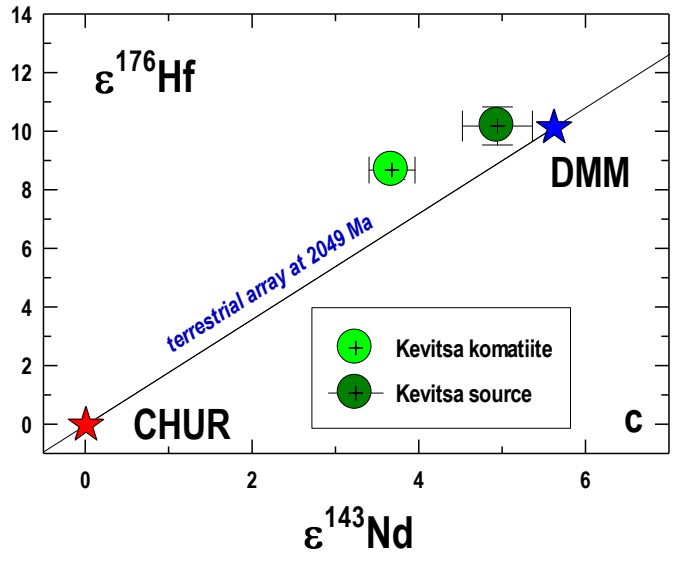
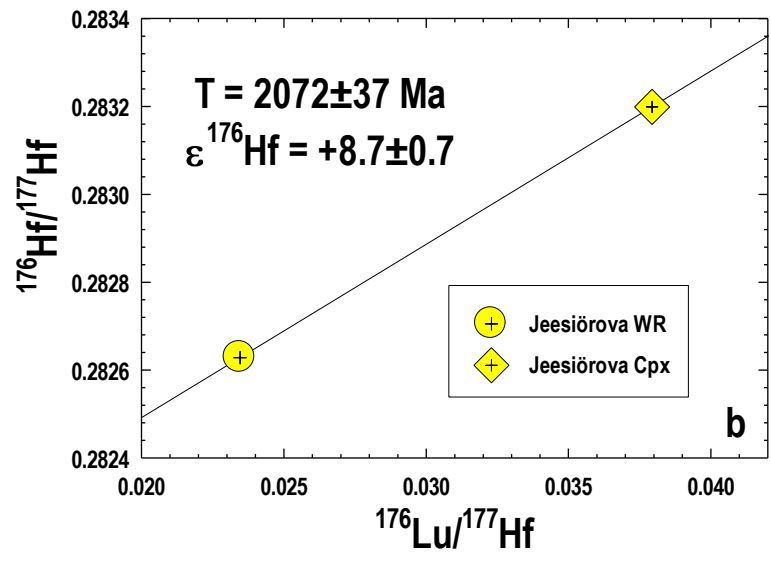
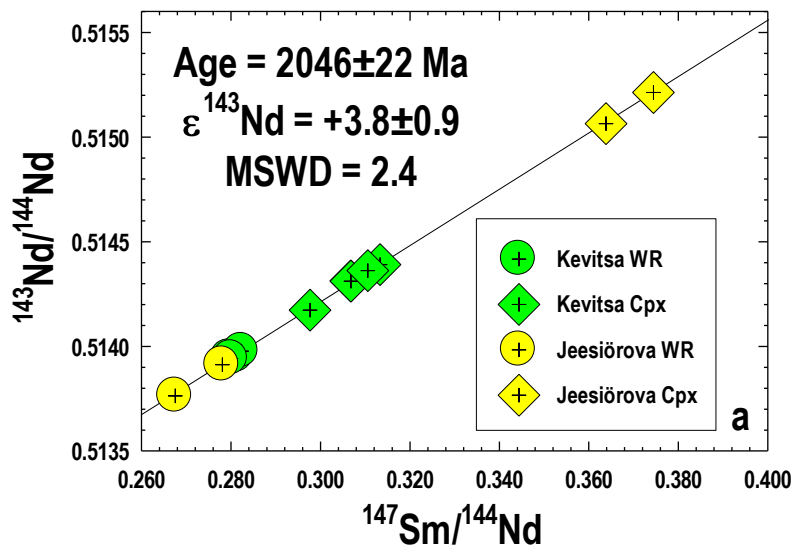


Fig. 9.

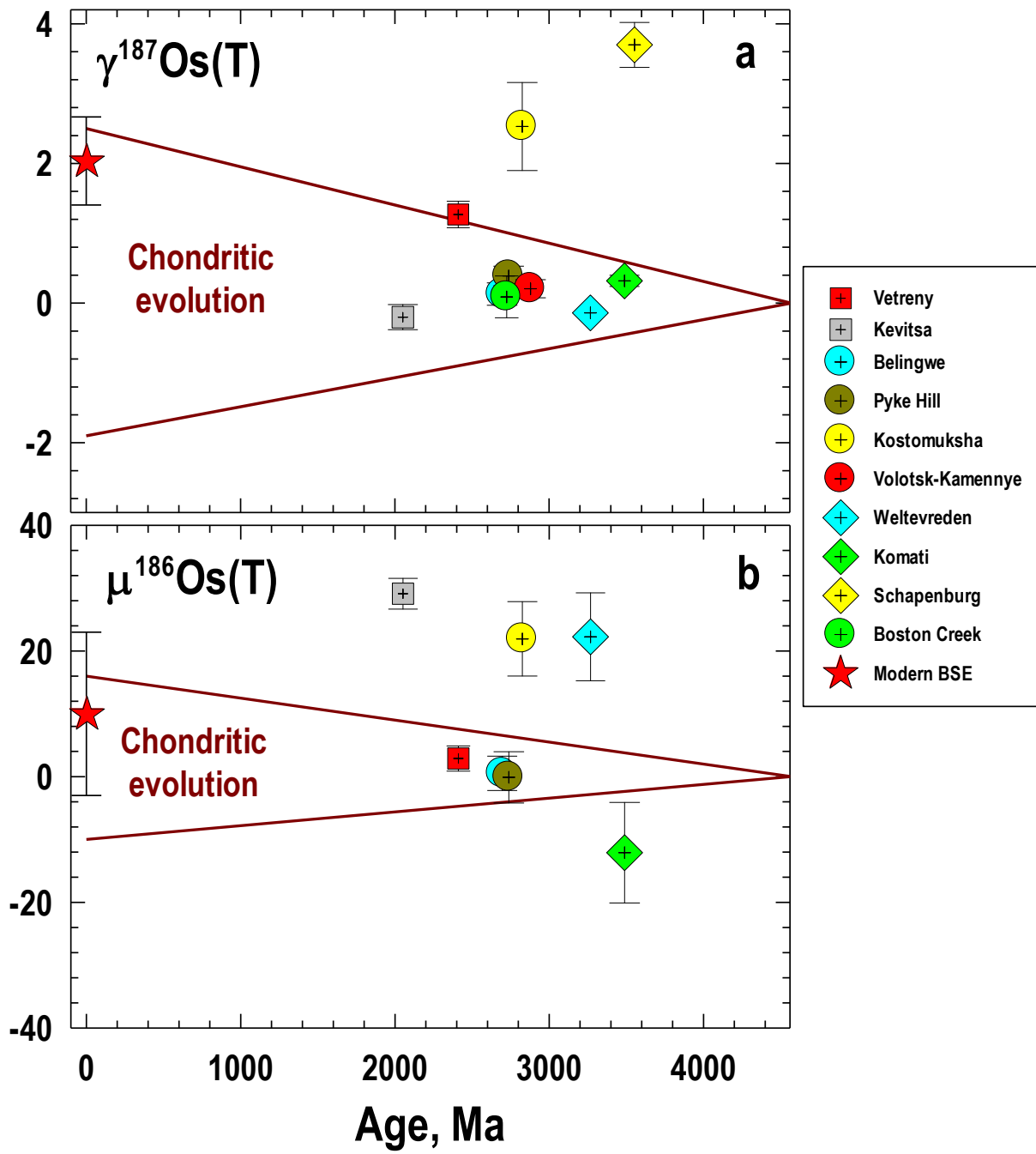


Fig. 10.

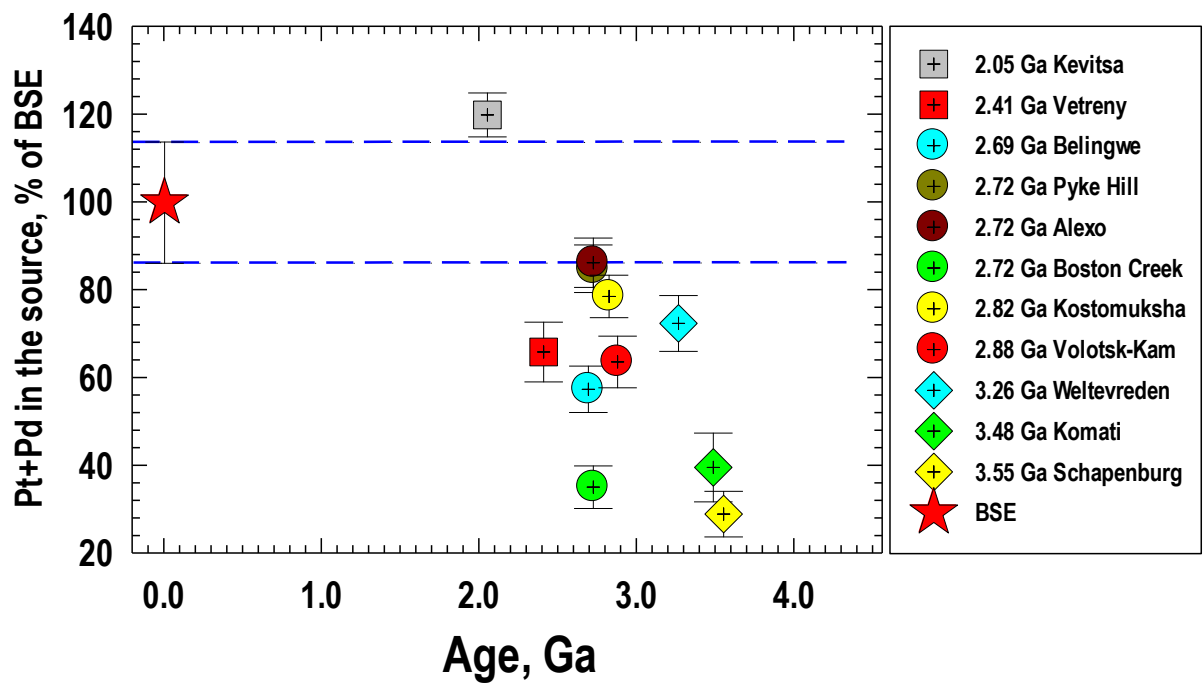


Fig. 11.

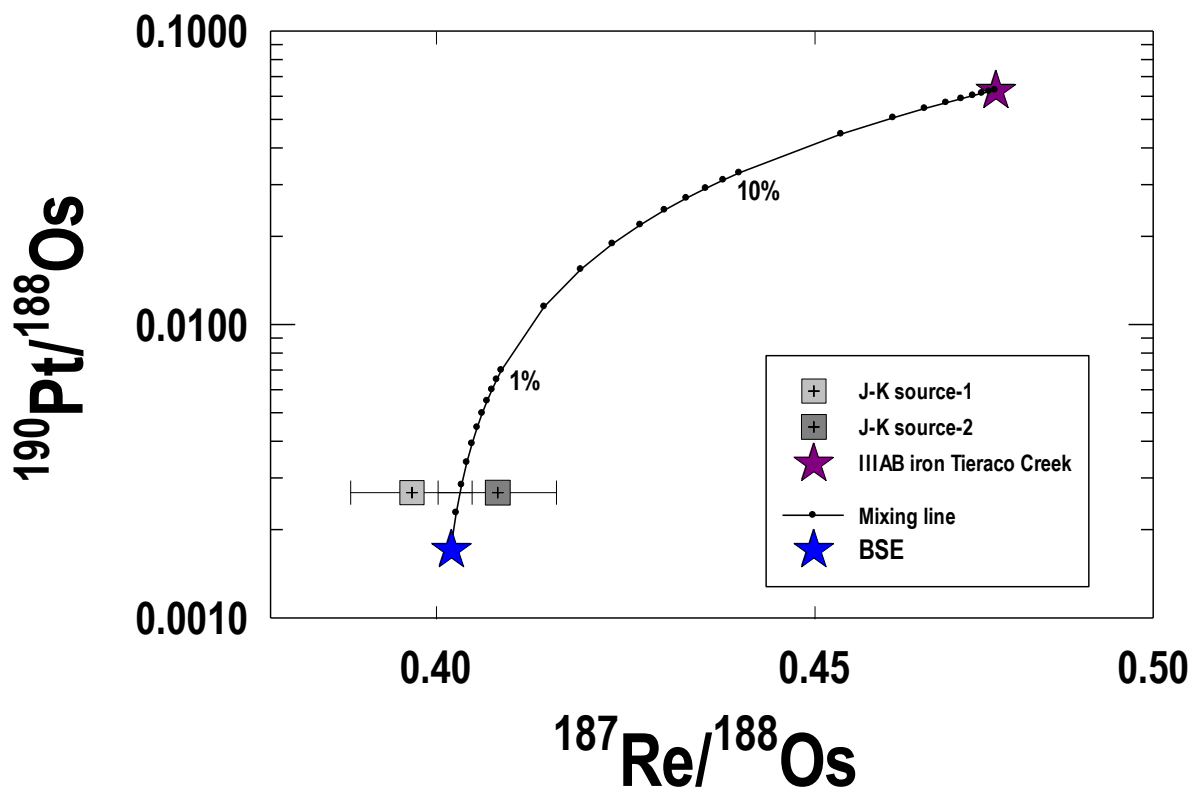


Fig. 12.

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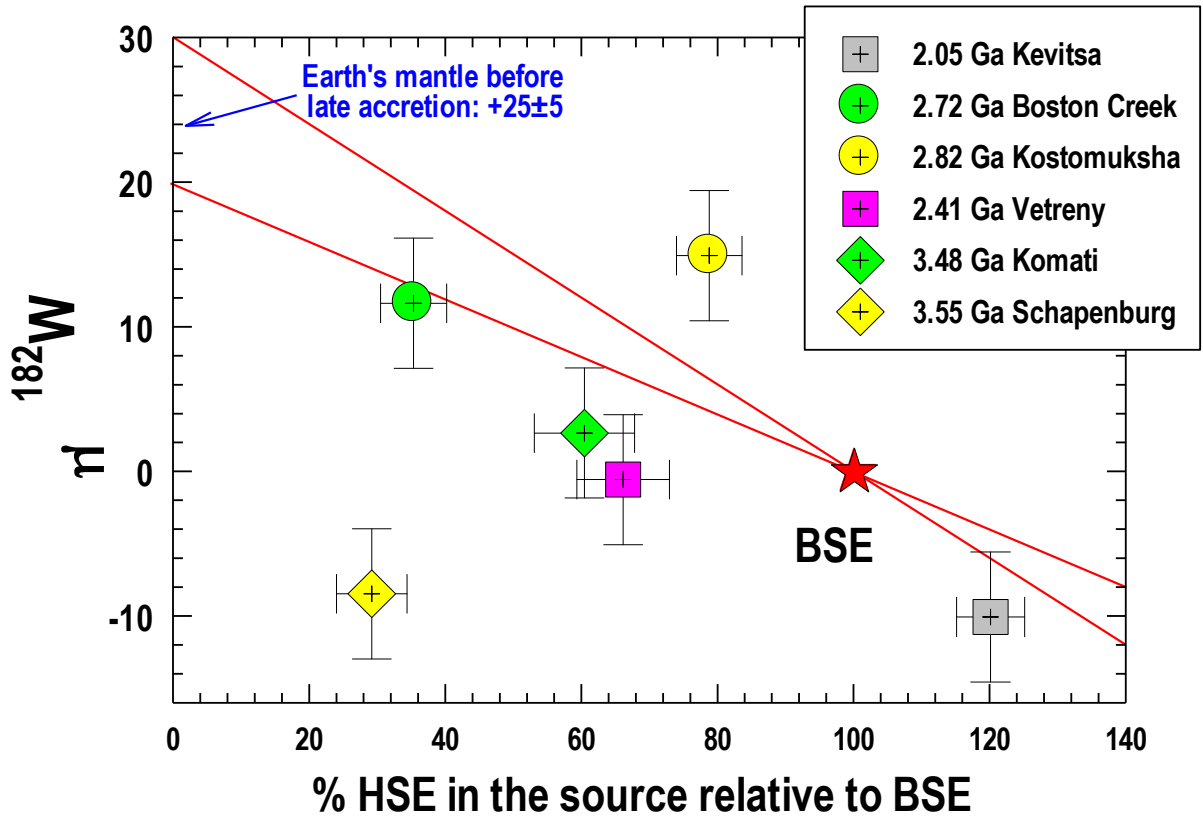


Fig. 13.

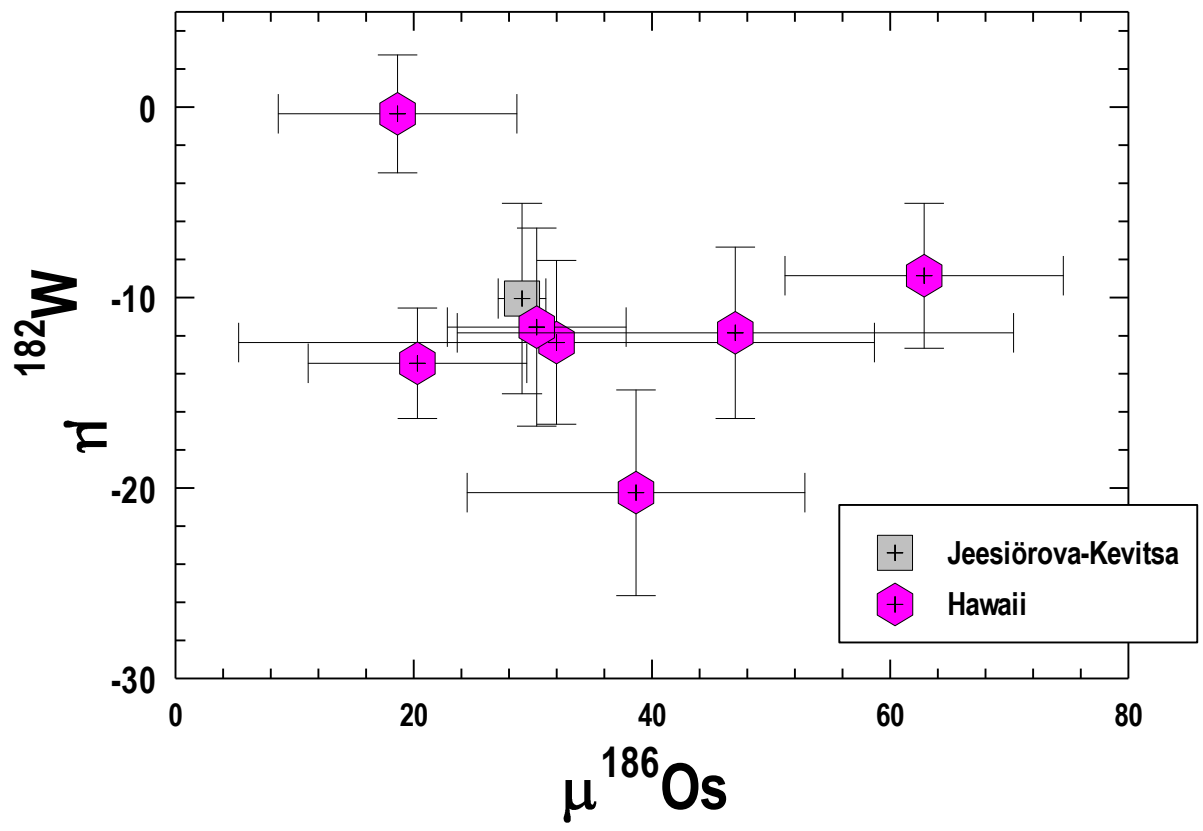


Fig. 14.

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Supplementary Tables A1 and A2

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Supplementary Table A3

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Declaration of interests

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.