Table 6.7 Isotopic data for lunar norite 77215

<table>
<thead>
<tr>
<th>Sample</th>
<th>Whole rock</th>
<th>Plagioclase</th>
<th>Pyroxene</th>
</tr>
</thead>
<tbody>
<tr>
<td>Rb (ppm)</td>
<td>2.67</td>
<td>6.59</td>
<td>1.09</td>
</tr>
<tr>
<td>Sr (ppm)</td>
<td>92.6</td>
<td>216</td>
<td>7.80</td>
</tr>
<tr>
<td>87Rb/86Sr</td>
<td>0.0834</td>
<td>0.0884</td>
<td>0.4041</td>
</tr>
<tr>
<td>87Sr/86Sr</td>
<td>0.704124</td>
<td>0.704723</td>
<td>0.724811</td>
</tr>
<tr>
<td>Sm (ppm)</td>
<td>3.52</td>
<td>8.35</td>
<td>2.224</td>
</tr>
<tr>
<td>Nd (ppm)</td>
<td>12.13</td>
<td>47.78</td>
<td>6.082</td>
</tr>
<tr>
<td>147Sm/144Nd</td>
<td>0.1754</td>
<td>0.1566</td>
<td>0.221</td>
</tr>
<tr>
<td>143Nd/144Nd:</td>
<td>0.512016</td>
<td>0.511491</td>
<td>0.513320</td>
</tr>
<tr>
<td>error</td>
<td>0.000004</td>
<td>0.000004</td>
<td>0.000005</td>
</tr>
<tr>
<td>Lu (ppm)</td>
<td>0.642</td>
<td>0.444</td>
<td>0.897</td>
</tr>
<tr>
<td>Hf (ppm)</td>
<td>2.70</td>
<td>3.501</td>
<td>3.988</td>
</tr>
<tr>
<td>176Lu/177Hf</td>
<td>0.0338</td>
<td>0.01801</td>
<td>0.03193</td>
</tr>
<tr>
<td>176Hf/177Hf</td>
<td>0.282778</td>
<td>0.281421</td>
<td>0.282601</td>
</tr>
</tbody>
</table>

Source: Data from Carlson et al. [2014].
$^{176}\text{Lu} \rightarrow ^{176}\text{Hf} + \beta^- + \nu$

\[ \lambda^{176}\text{Lu} = 1.867 \times 10^{-11} \text{ a}^{-1} \]

\[ t_{1/2} = 37.13 \text{ Ga} \]

less than 0.5% of $^{176}\text{Lu}$ decay is e.c. to $^{176}\text{Yb}$

\[ \frac{^{176}\text{Hf}}{^{177}\text{Hf}} = \left( \frac{^{176}\text{Hf}}{^{177}\text{Hf}} \right)_{1} + ^{176}\text{Lu} \left( e^{\lambda t} - 1 \right) \]

Table 6.3  Lu–Hf isotopic composition in atom %

<table>
<thead>
<tr>
<th>Element</th>
<th>174</th>
<th>175</th>
<th>176</th>
<th>177</th>
<th>178</th>
<th>179</th>
<th>180</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lu</td>
<td></td>
<td></td>
<td>97.42</td>
<td>2.58</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Hf\textsuperscript{a}</td>
<td>0.161</td>
<td>5.258</td>
<td>18.595</td>
<td>27.282</td>
<td>13.621</td>
<td>35.083</td>
<td></td>
</tr>
</tbody>
</table>

\textsuperscript{a} Hf atomic abundances assume a $^{176}\text{Hf}/^{177}\text{Hf}$ ratio of 0.282785.
Fig. 9.12. Partition coefficients for REE and Hf between minerals (cpx, garnet) and kimberlite magma. After Fujimaki et al. (1984).
Garnet Sm-Nd and Lu-Hf dating
Smit et al., 2013

Fig. 2. Major- and trace-element (REE, Hf) concentration profiles of four representative garnet grains from large (a), (b) and small (c), (d) grain-size classes (U-chondrite normalizing factors).

Fig. 6.8. Two-point garnet-whole-rock tie line ages plotted against grain size for the data provided in Table 6.13. (Source: Smit et al. [2013]. Reproduced with permission of Elsevier.)
Garnet Sm-Nd and Lu-Hf dating

Potential complications
1. Numerous mineral inclusions that can screw up P/D ratios.
2. Post-metamorphic ingrowth must be large enough to overcome sometimes high initial ratios.
3. Garnet growth may occur at, below, or above Tc, complicating thermochronologic interpretations.
4. Daughter Hf may diffuse more slowly than parent Lu...this is not how a thermochronometer works.

Fig. 6.8. Two-point garnet–whole-rock tie line ages plotted against grain size for the data provided in Table 6.13. (Source: Smits et al. [2013]. Reproduced with permission of Elsevier.)
Garnet Sm-Nd and Lu-Hf dating
Smit et al., 2013

Fig. 7. $T_{c}^{Hi}$ (a) and $T_{c}^{Nd}$ (b) as a function of diffusion domain radius $r$. Both were solved for different cooling rates and activation energies of 250 and 300 kJ mol$^{-1}$, which likely bracket $E_a$ for cation diffusion in garnet.

Fig. 8. $T_{c}^{Nd}$ as a function of diffusion domain radius $r$ obtained by comparing apparent Sm–Nd ages to the externally calibrated cooling curve for the PGD at Cauchon Lake. Data for garnet from the same research area by Mezger et al. (1992) are also plotted (M92). $T_{c}^{Nd}$ calculated for the PGB ($dT/dt = 1.5 ^\circ$C Myr$^{-1}$; Mezger et al., 1989b; Kooijman et al., 2010) using Nd diffusion data of (I) Ganguly et al. (1998) and (II) van Orman et al. (2002) are shown for comparison.
Fig. 9.12. Partition coefficients for REE and Hf between minerals (cpx, garnet) and kimberlite magma. After Fujimaki et al. (1984).
Isotopic evolution of Hf in a chondritic reservoir. The primordial $^{176}\text{Hf}/^{177}\text{Hf}$ ratio is 0.27978, its present $^{176}\text{Lu}/^{177}\text{Hf}$ ratio is 0.0334, the present $^{176}\text{Hf}/^{177}\text{Hf}$ ratios is 0.28286, and the age of the reservoir is $4.55 \times 10^9$ years (Patchett et al. 1981). Partial melting in the reservoir at $t_1$ produces liquids having lower Lu/Hf ratios, whereas the residual solids have higher Lu/Hf ratios than the chondritic reservoirs. Subsequent melting of the residual solids (depleted mantle) at $t_2$ produces magmas whose $^{176}\text{Hf}/^{177}\text{Hf}$ ratios are greater than those of the chondrite reservoir and their $\varepsilon^t$ (Hf) values are therefore positive. Renewed melting of crustal rocks at $t_2$ results in magmas that are depleted in $^{176}\text{Hf}$ and have negative $\varepsilon^t$ (Hf) values. The blending of melts derived from the depleted mantle and the continental crust results in intermediate $\varepsilon$-values for igneous rocks that form from such contaminated magmas.
Fig. 6. Initial $\varepsilon_{Hf}$ vs. $\varepsilon_{Nd}$ plot of whole-rock data from Archean to recent sediments from a range of depositional environments; modern OIBs, ORBs and IABs; granitoids; and continental basalts. Also shown are data for juvenile samples (unfilled circles) and early Archean gneisses (squares) from this paper. All data shown here (excluding early Archean data) plot along an array (Vervoort et al., 1999; $\varepsilon_{Hf} = 1.36 \varepsilon_{Nd} + 3.0$; upper black line) indistinguishable from that of the juvenile samples ($\varepsilon_{Hf} = 1.40 \varepsilon_{Nd} + 2.1$; lower grey line) reported in this paper. As is apparent from this plot, which contains samples of wide-ranging ages and representing different crust and mantle reservoirs, Hf and Nd isotopes behave remarkably coherently. Although there is decoupling of Hf and Nd isotopes in some unique environments it is clear that these effects are not translated into larger-scale features in the major crust and mantle reservoirs. In addition to the data presented in this paper, sources of data include the following. Hf isotopes (some with Nd isotopes): Beard and Johnson (1993); Johnson and Beard (1993); Patchett (1983a; 1983b); Patchett and Tatsumoto, (1980c); Salters (1996); Salters and Hart (1991); Stille et al. (1983; 1986); Vervoort and Patchett, (1996); Vervoort et al. (1999); White and Patchett (1984). Nd isotope data only: Abouchami et al. (1990); Ben Othman et al. (1989); Boher et al. (1992); Gleason et al. (1995); McLennan et al. (1990); O’Nions et al. (1977); Patchett et al. (1984); Premo (1991); Samson et al. (1989); Stern and Abdelsalam (1998); Stern and Kröner (1993); Stevenson et al. (1990); White and Hofmann (1982); White et al. (1985); Wirth and Vervoort (1995).
Fig. 7. Depleted-mantle evolution plots of initial $\varepsilon_{\text{Nd}}$ (a) and $\varepsilon_{\text{Hf}}$ (b) vs. age of juvenile mantle-derived samples. Nd isotope data compiled from Shirey (1991) and augmented with data from Jacobsen and Wasserburg (1988), Bennett et al. (1993), and Bowring and Harland (1995). All Hf isotope data have been determined by us with the exception of the zero-age island arc data (White and Patchett, 1984). Grey lines represent approximate upper limit of the ranges. Although there is still insufficient coverage of crucial time periods for Hf isotope data (most notably the period from 3.6 to 2.7 Ga) both Hf and Nd isotope records are similar, with the exception of the early Archean samples.
$^{146}\text{Sm} \rightarrow ^{142}\text{Nd}$

$\lambda \sim 6.73 \times 10^{-9} \text{ a}^{-1}$

$t_{1/2} \sim 103 \text{ Ma}$
NWA 4590 U/Pb age = 4557.8 ± 0.4 Ma

$\left(\frac{^{146}\text{Sm}}{^{144}\text{Sm}}\right)_0 = 0.00814$

other estimates: 0.00828 ± 0.00044, 0.0085 ± 0.0007.

$t \sim 9.2$ Ma...
Fig. 6. $\varepsilon^{142}$Nd results for Archean rocks from West Greenland, Barberton, and Acasta, and modern oceanic basalts. Rock sample measurements were systematically bracketed with analyses of the Nd Ames standard. All rock samples from the West Greenland craton display well-defined $^{142}$Nd excesses compared to modern oceanic basalts and Ames Nd standard. Note that all data from Table 3 are represented (including duplicates). Error bars represent the internal precision of individual measurements ($2\sigma_m$).
instantaneous depleted mantle formation

continuous depleted mantle formation

continuously interacting depleted mantle and crust

Caro et al., 2006
Boyet and Carlson, 2005

Fig. 1. $^{142}\text{Nd}/^{144}\text{Nd}$ ratios measured for chondrites and eucrites compared to the La Jolla terrestrial Nd standard ($\varepsilon^{142}\text{Nd}$). All chondrites and basaltic eucrites have negative $\varepsilon^{142}\text{Nd}$ values outside the external analytical error of $\pm 0.07$ $\varepsilon$ units ($2\sigma$) (shaded area). Cumulate eucrites have positive $\varepsilon^{142}\text{Nd}$ values in agreement with their high Sm/Nd, resulting from igneous processes on their parent body. The error bars correspond to the internal precision ($2\sigma_{\text{mean}}$). Terrestrial samples (MORBs, kimberlites, and carbonatites of different ages and collected in diverse locations) measured using the same procedure ($2\sigma$) have been added to demonstrate the significant excess of 0.2 $\varepsilon$ units in all the terrestrial material (samples and standard) relative to the mean chondritic value. All terrestrial samples were measured several times using the same procedures as were used for the chondrites. The uncertainties reported on the mean are $2\sigma$.

Boyet and Carlson, 2006

Fig. 1. $^{142}\text{Nd}/^{144}\text{Nd}$ in the samples analyzed here, expressed ($\varepsilon_{142\text{Nd}}$) as parts in 10,000 deviation from the average measured for the La Jolla Nd standard, which is approximately 0.2 $\varepsilon_{142\text{Nd}}$ higher than the average measured for chondrites and eucrites [9]. The grey bar shows the average external precision obtained on repeat analyses of the same sample. Error bars on the individual points show the $2\sigma$ deviation obtained for multiple measurements of each sample. Only 5 samples from Isua show deviation outside of analytical uncertainty from the $^{142}\text{Nd}/^{144}\text{Nd}$ measured for La Jolla Nd.
Figure 18.3. Sr and Nd isotope ratios of the suboceanic mantle as sampled by oceanic basalts. Three fields are shown for MORB from the 3 ocean basins, other symbols are various oceanic islands and island chains.
Fig. 3. Sr–Nd isotope compositions of the SLCP and PK cpx compared with the Pacific MORB, Pacific seamounts and Hawaiian lavas. Symbols as in Fig. 1 and data from Table 1. Other peridotite data: open triangles PK peridotites from [34], small circles SLCP from [11,16]. Fields for Pacific MORB and other Pacific seamount data are compiled from the GEOROC [73] and PETDB [74] databases. The Pacific MORB samples with the most radiogenic $^{143}$Nd/$^{144}$Nd ratios from the Garret fracture zone [37] are shown separately for comparison. The data fields for the HV and Koʻolau lavas from Oʻahu are shown separately for comparison. Most of the SLCP overlap the HV lavas. The PK peridotites extend to more radiogenic $^{143}$Nd/$^{144}$Nd ratios than the SLCP or the Hawaiian lavas, similar to the most radiogenic Pacific MORB (Garret FZ). The three SLCP with low $^{143}$Nd/$^{144}$Nd for a given $^{87}$Sr/$^{86}$Sr are pointed out separately. The SLCP field labeled “SLC V, O” encloses three SLCP data from [11,16] with relatively radiogenic $^{87}$Sr/$^{86}$Sr ratios. A plagioclase separate from a plagioclase peridotite [36] is the only reported Oʻahu peridotite with Nd–Sr isotope compositions that overlap the Koʻolau lavas.
We will return to this later, with Re/Os system.
Gehrels zircon Lu/Hf approaches
Fig. 5. BSE (37) normalized trace element abundances for early-differentiated reservoirs of different sizes. These curves are calculated by assuming that the $^{147}\text{Sm}/^{144}\text{Nd}$ of the EDR is 0.21, needed to evolve the 20-ppm terrestrial excess in $^{142}\text{Nd}/^{144}\text{Nd}$ for a reservoir formation age of 30 My after Earth formation. The remaining elemental compositions of the EDR are then calculated by adding average continental crust (29) to average MORB source mantle (30) until this Sm/Nd is achieved. The composition of the EER is then calculated on the assumption that the composition of the EER and EDR sum to BSE (37) for the various assumed mass ratios of EER and EDR shown in the figure.
Lavas with primordial $^3\text{He}/^4\text{He}$ don’t have primitive chondritic $^{143}\text{Nd}/^{144}\text{Nd}$

Jackson et al. *(EPSL, 2007)*
Nd- and Pb-isotopic composition of Baffin Island lavas from Padloping Island.

MG Jackson et al.  
*Nature* **466**, 853-856 (2010)  
doi:10.1038/nature09287
Table 6.2 Sm–Nd isotopic composition in atom%

<table>
<thead>
<tr>
<th>Element</th>
<th>Mass</th>
<th>142</th>
<th>143</th>
<th>144</th>
<th>145</th>
<th>146</th>
<th>147</th>
<th>148</th>
<th>149</th>
<th>150</th>
<th>152</th>
<th>154</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sm</td>
<td>27.168</td>
<td>3.075</td>
<td>14.99</td>
<td>11.24</td>
<td>13.82</td>
<td>7.38</td>
<td>26.74</td>
<td>22.75</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Nd\textsuperscript{a}</td>
<td>12.197</td>
<td>23.794</td>
<td>8.290</td>
<td>17.177</td>
<td>5.748</td>
<td>5.626</td>
<td></td>
<td></td>
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<td></td>
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</tr>
</tbody>
</table>

\textsuperscript{a}Nd atomic abundances assume a \textsuperscript{143}Nd/\textsuperscript{144}Nd ratio of 0.51263.
Our results demonstrate that chondrites are the most appropriate proxy for the elemental composition of the Earth. However, they also highlight that chondrites cannot be the actual building blocks of the Earth because they are deficient in a presolar component that contains s-process matter. The s-process deficit increases from enstatite via ordinary to carbonaceous chondrites, indicating that the distribution of presolar matter in the solar protoplanetary disk either varied as a function of heliocentric distance or over time. For instance, the nucleosynthetic isotope heterogeneity within the disk may reflect differences in the thermal processing of stellar-derived dust, imparting isotopic heterogeneity on an initially homogeneous disk, but it could also reflect distinct compositions of material added to the disk from the molecular cloud at different times\textsuperscript{18,28–30}. Either way, the increasing deficit in s-process matter with increasing heliocentric distance provides a new means for identifying genetic relationships among planetary bodies. For instance, Mars formed at a greater heliocentric distance than Earth and should, therefore, be characterized by an s-process deficit that may be similar to those observed for enstatite and ordinary chondrites. Thus, high-precision Nd isotopic data for Martian

Bouvier & Boyet, 2016
Figure 3 | Comparison of absolute internal Pb–Pb ages of CAIs with the Al–Mg, Hf–W and Mn–Cr model ages for CAIs anchored on the isotope systematics of the D’Orbigny angrite. The Pb–Pb age and Al–Mg model age for CAI 2364-B1 (this study), and Mn–Cr and Hf–W model ages for CAIs (based on previous studies), calculated relative to the D’Orbigny angrite (vertical grey band; see Supplementary Table S3 for parameters used for calculations of the model ages, with their associated errors, and corresponding references). For comparison and consistency, we show the range of Pb–Pb internal isochron ages for CAIs from Allende and Efremovka CV chondrites\textsuperscript{1,4-6} and the range of Pb–Pb ages for chondrules from the CR and CV chondrites\textsuperscript{1,26} after recalculating the ages assuming $\frac{^{238}\text{U}}{^{235}\text{U}} = 137.84$ (ref. 15). Also shown here is the range and the relative frequency of Al–Mg ages of chondrules from carbonaceous and ordinary chondrites\textsuperscript{28}. 

Bouvier and Wadhwa, 2010