Vertical bars represent relative abundances of xenon isotopes.

Region of Nuclide Chart in near I-129. This unstable isotope has a half-life of 1.6E7 years. It's decay to Xe-129 produces an excess of that isotope. Stable isotopes in gray.
# Extinct Radionuclides

<table>
<thead>
<tr>
<th>Fractionation [^\text{b}]</th>
<th>Parent nuclide</th>
<th>Half-life (Myr)</th>
<th>Daughter nuclide</th>
<th>Estimated initial solar system abundance</th>
<th>Objects found in</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>Nebular</td>
<td>(^{26})Al</td>
<td>0.7</td>
<td>(^{26})Mg</td>
<td>(10^{-5} \times 10^{-5}) \times 26Al</td>
<td>CAIs, chondrules, achondrite</td>
<td>(2)</td>
</tr>
<tr>
<td></td>
<td>(^{10})Be</td>
<td>1.5</td>
<td>(^{10})B</td>
<td>(6 \times 10^{-4}) \times (^{9})Be</td>
<td>CAIs</td>
<td>(3)</td>
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<tr>
<td></td>
<td>(^{53})Mn</td>
<td>3.7</td>
<td>(^{53})Cr</td>
<td>(\sim 2 \times 10^{-5} \times (^{55})Mn</td>
<td>CAIs, chondrules, carbonates, achondrites</td>
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<td></td>
<td>(^{60})Fe</td>
<td>1.5</td>
<td>(^{60})Ni</td>
<td>(\sim 2 \times 10^{-7} \times (^{60})Fe</td>
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<td>(5)</td>
</tr>
<tr>
<td></td>
<td>(^{107})Pd</td>
<td>6.5</td>
<td>(^{107})Ag</td>
<td>(5 \times 10^{-7} \times (^{108})Pd</td>
<td>iron meteorites, pallasites</td>
<td>(6)</td>
</tr>
<tr>
<td></td>
<td>(^{182})Hf</td>
<td>9</td>
<td>(^{182})W</td>
<td>(10^{-4} \times (^{180})Hf</td>
<td>planetary differentiates</td>
<td>(7)</td>
</tr>
<tr>
<td></td>
<td>(^{129})I</td>
<td>15.7</td>
<td>(^{129})Xe</td>
<td>(10^{-4} \times (^{129})I</td>
<td>chondrules, secondary minerals</td>
<td>(8)</td>
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<tr>
<td></td>
<td>(^{92})Nb</td>
<td>36</td>
<td>(^{92})Zr</td>
<td>(10^{-4} \times (^{92})Nb</td>
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<tr>
<td></td>
<td>(^{244})Pu</td>
<td>82</td>
<td>Fission products</td>
<td>(7 \times 10^{-3} \times 238)U</td>
<td>CAIs, chondrites</td>
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<tr>
<td></td>
<td>(^{146})Sm</td>
<td>103</td>
<td>(^{142})Nd</td>
<td>(9 \times 10^{-4} \times (^{147})Sm</td>
<td>achondrites</td>
<td>(11)</td>
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</tbody>
</table>

References: (1) Srinivasan et al. (1994, 1996); (2) Lee et al. (1977); MacPherson et al. (1995); (3) McKeegan et al. (2000); (4) Birck and Allègre (1985); Lugmair and Shukolyukov (1998); (5) Shukolyukov and Lugmair (1993a); Tachibana and Huss (2003); (6) Chen and Wasserburg (1990); (7) Kleine et al. (2002a), Yin et al. (2002); (8) Jeffery and Reynolds (1961); (9) Schönberger et al. (2002); (10) Hudson et al. (1988); and (11) Lugmair et al. (1983).

\[^{a}\] Some experimental evidence exists suggesting the presence of the following additional isotopes, but confirming evidence is needed (half-lives are given after each isotope): \(^{9}\)Be—53 d (Chaussidon et al., 2002); \(^{99m}\)Te—0.2 Myr (Yin et al., 2000); \(^{30}\)Cl—0.3 Myr (Murty et al., 1997); \(^{205}\)Pb—15 Myr (Chen and Wasserburg, 1987).

\[^{b}\] Environment in which most significant parent–daughter fractionation processes occur.

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McKeegan & Davis (2003)
Dickin (2005)

Fig. 15.6. Xe–Xe plot for stepwise-degassed samples of the Richardton meteorite, showing the line of constant initial $^{129}$I/$^{127}$I ratio. Solid and open symbols indicate gas fractions released above and below 1100 °C, respectively. After Hohenberg et al. (1967).

Fig. 15.11. Xe/Xe ages for individual chondrules for several meteorites relative to whole-rock samples of the standard Bjurböle. The achondrite Shallowater is shown for reference. Modified after Whitby et al. (2002).
10. Correlation diagram to test the concordancy of I-Xe and Pb/Pb ages on apatite separates from Acanthlaetic meteorites. Ages are given relative to Acapulco.

Dickin (2005)
Evidence for Extinct Al-26

\[ \frac{^{26}\text{Mg}}{^{24}\text{Mg}} \text{ vs. } \frac{^{27}\text{Al}}{^{24}\text{Mg}} \]

- Anorthite-G
- Anorthite-B
- Low aluminum silicates
Figure 29 $^{26}$Al–$^{26}$Mg isochron diagram for a large number of CAIs. Data taken from numerous sources listed in MacPherson et al. (1995)
Figure 30  Histogram of calculated initial $^{26}$Al/$^{27}$Al values, comparing normal and FUN CAIs. FUN CAIs rarely show evidence of live $^{26}$Al, and when evidence is available (in the case of the hibonite-rich Allende CAI HAL), the initial value is very low. Data sources as in Figure 29.

Figure 31  Histogram of calculated initial $^{26}$Al/$^{27}$Al values in chondrules. Data from numerous sources as summarized in Table 2 of Huss et al. (2001)

MacPherson (2003)
Fig. 15.12. $^{25}\text{Mg}^{24}\text{Mg}$ versus $^{26}\text{Mg}^{24}\text{Mg}$ isotope diagram, showing deviations of Allende inclusions from the normal solar-system value, in parts per mil ($\delta$). These may be due to mass fractionation in the solar nebula (open symbols) or decay of extinct $^{26}\text{Al}$ (solid symbols). After Wasserburg and Papanastassiou (1982).
Fig. 15.13. Plot of $\delta$ Mg against Al/Mg ratio for the ilmenite EGG-3 inclusion, showing the best-fit line of instant initial $^{26}\text{Al}/^{27}\text{Al}$ ratio with a value of $4.9 \times 10^{-5}$. After Armstrong et al. (1984).

Fig. 15.14. Schematic illustration of a model in which solar-system collapse is promoted by a supernova, which also seeds it with short-lived nuclides. After Wasserburg (1985).

Dickin (2005)
Trinquier et al. (2008): Chondritic meteorites
C1 – C4 – carbonaceous, OC- ordinary
EC – enstatite;

EPB – eucrite parent body (Asteroid 4 Vesta: eucrites, diogenites, mesosiderites)

C3, C4 – CV3, CO3

Mean of chondrites: 0.20 ± 0.10

lower than previously reported

Fig. 1. Present-day ε_{OC} values in the inner Solar System. Chondrites and the EPB average excesses of 0.20 ± 0.10 (2SD) and 0.11 ± 0.06, respectively, are significantly lower than previously reported in Lugmair and Shukolyukov (1998). Error bars represent 2σ uncertainties (95% confidence level).
Fig. 6. $^{55}\text{Mn} / ^{53}\text{Cr}$ vs. $\varepsilon^{53}\text{Cr}$ excesses in the inner Solar System. The correlation among OC–EC–CI–CV–CO chondrites and terrestrial planets and planetesimals likely represents early Mn/Cr fractionations in the protoplanetary disk associated with a loss of the moderately volatile components (Palme et al., 1988; Palme and O’Neill, 2003). The excessive scatter of data points for individual OC, CC and EC samples (inset) stems from sample heterogeneity. However, the means of chondrite groups and the bulk of terrestrial planets (black circles and squares in the main graph) average out the sampling heterogeneity of the Mn/Cr ratios and generate a correlation with identical $^{53}\text{Mn} / ^{55}\text{Mn}$ to that obtained for Orgueil (Fig. 4). CR–CK–CM chondrites plot at elevated $\varepsilon^{53}\text{Cr}$ relative to the correlation line, most likely due to incomplete dissolution of the most refractory $^{53}\text{Cr}$ deficient minerals in bulk samples (star symbols in the inset). St-Aubin iron meteorite chromite is indistinguishable from the Solar System $\varepsilon^{53}\text{Cr}$ initial.

Trinquier et al. (2008): Pb/Pb age of LEW as reference: 4567.1 Ma
Fig. 7. Solar System $\varepsilon_{\text{Fe}}$ isotopic evolution diagram vs. time. The isotopic evolution of individual samples through time was calculated using the slope and initial ratios obtained from isochron diagrams and present-day $\varepsilon_{\text{Fe}}$ values (Table 1). Present-day $\varepsilon_{\text{Fe}}$ in chondrites and terrestrial planets are consistent with primordial nebular Mn/Cr fractionation. Estimates of Solar System $^{53}\text{Mn} / ^{55}\text{Mn}$ and $\varepsilon_{\text{Fe}}$ initials inferred from C1 Orgueil as well as back calculated to the time of CAI formation using SM systematics are indistinguishable. The best estimates are $^{53}\text{Mn} / ^{55}\text{Mn}_{\text{Initial}} = (6.28 \pm 0.66) \times 10^{-6}$ and $\varepsilon_{\text{Fe}}_{\text{Initial}} = -0.23 \pm 0.09$.

Trinquier et al. (2008): Pb/Pb age of LEW as reference: 4567.1 Ma
Minarik B.: The core of planet formation

Figure 1 Timing of core formation. The Earth formed through accretion, absorbing planetesimals (lumps of rock and ice) through collisions. Did the Earth accrete undifferentiated material that then separated into shell and core — in which case, did the planet reach its present mass before differentiating, or was it a more gradual process? Alternatively, core formation might have happened rapidly inside growing planetesimals, so that the Earth’s core is a combination of these previously formed cores. Isotopic evidence supports the latter model, and now Yoshino et al.1 demonstrate a mechanism for the physical process.

Figure 2 Degrees of separation. Metal alloys with a lower melting point than silicate minerals form a melt between the grain boundaries of silicate crystals. The dihedral angle (left) is a measure of how likely it is that pockets of melt will connect and separate from the silicate matrix. The photomicrograph (right) shows sulphide melt trapped in a matrix of Fe,Mg oxide and olivine (the silicate that makes up most of the Earth’s upper mantle). The dihedral angle is near 90° and, if the melt makes up a sufficiently large fraction of the sample volume, the melt pockets start to interconnect. The field of view of the photomicrograph is about 500 μm.

Haack & McCoy (2003)

Figure 3 Illustration of the dihedral angle θ and a depiction of two end-member microstructures for static systems. If θ < 60°, an interconnected network will form and melt migration can occur. If θ > 60°, melt forms isolated pockets. In experimental systems, inter-connectedness only occurs in anion-rich static systems.

Figure 1 $\varepsilon_W$ values of carbonaceous chondrites compared with those of the Toluca iron meteorite and terrestrial samples analysed in this study. The values for Toluca, Allende, G1-RF and IGDL-GD are the weighted averages of four or more independent analyses.

Figure 2 $\varepsilon_W$ versus $^{180}\text{Hf}/^{184}\text{W}$ for different fractions of the H chondrites Ste Marguerite (a) and Forest Vale (b). NM-1, NM-2 and NM-3 refer to different nonmagnetic fractions, $M$ is the magnetic fraction. We interpret the positive correlation of $\varepsilon_W$ with $^{180}\text{Hf}/^{184}\text{W}$ as an internal Hf–W isochron whose slope corresponds to the initial $^{180}\text{Hf}/^{184}\text{Hf}$ ratio at the time of closure of the Hf–W system.
**Figure 3** Time of core formation in Myr after CAI condensation for Vesta, Mars, Earth and Moon versus planet radius as deduced from Hf–W systematics. For the Moon, the two data points refer to the endmember model ages. The Moon plots distinctly to the left of the correlation line defined by Vesta, Mars and Earth, suggesting a different formation process.

Kleine et al. (2002)

**Figure 5** Well-defined deficiency in $^{182}$W in early metals and carbonaceous chondrites relative to the silicate Earth (source Lee and Halliday, 1996; Horan et al., 1998; Kleine et al., 2002).

Halliday (2003)
Fig. 1. Hf-W isochrons for CAIs. Error bars represent 2σ uncertainties, m = initial $^{182}\text{Hf}/^{180}\text{Hf}$, i = initial $\epsilon_W$. Regressions (model 1) were calculated using ISOPLOT (Ludwig, 1991). The absolute ages are calculated from their initial $^{182}\text{Hf}/^{180}\text{Hf}$ ratios relative to the well-dated H chondrite Ste. Marguerite (see Appendix for details). (a) Hf-W data for separates from All-MS-1. M = magnetic fraction, NM = nonmagnetic fraction, WM = weakly magnetic fraction, WR = whole-rock. Fine and NM-fractions from All-MS-1 consist mostly of plagioclase and melilite, the WM-fraction mostly of pyroxene and metal. (b) Combined CAI isochron including Hf-W data for All-MS-1, bulk CAIs A37 and A44a (Yin et al., 2002) and carbonaceous chondrites (CC).

Fig. 2. Hf-W isochron for the CH chondrite Acfer 182. Error bars represent 2σ uncertainties, m = initial $^{182}\text{Hf}/^{180}\text{Hf}$, i = initial $\epsilon_W$. Regressions (model 1) were calculated using ISOPLOT (Ludwig, 1991). The absolute age is calculated from initial $^{182}\text{Hf}/^{180}\text{Hf}$ ratios relative to the well-dated H chondrite Ste. Marguerite (see Appendix for details). NM-fractions from Acfer 182 consist mostly of olivine and pyroxene.

Kleine et al. (2005)
Fig. 3. εW values for iron meteorites compared to the initial εW values of CAIs and metal-rich chondrites. The latter are taken from the intercepts of the isochrons shown in Fig. 1 and Fig. 2. Tungsten isotope data for Toluca and Yanhuitlan are from the literature: (1) Kleine et al. (2002), (2) Markowski et al. (2004). Error bars represent 2σ uncertainties. Dashed vertical lines indicate the 2σ uncertainties in initial εW values for individual groups or samples. The left vertical hatched bar indicates the weighted average initial εW of the IIIAB, IVA, IVB, and IC iron meteorites of −3.79 ± 0.06 (2σ); the right vertical hatched bar indicates the initial εW of CAIs of −3.47 ± 0.20 (2σ).

Fig. 6. Hf-W ages for iron meteorites compared to Al-Mg and Pb-Pb ages for chondrules. These ages are linked to each other using the Hf-W data for CAIs from this study and Al-Mg data for CAIs from Bizzarro et al. (2004) and U-Pb data from Amelin et al. (2002). The vertical dashed line represents the CAI reference and is set to a time of 0 Myr. The horizontal dotted lines indicate the uncertainty on some Hf-W ages caused by possible cosmogenic effects. All chondrule ages and the Hf-W age for Yanhuitlan are from the literature: (1) Amelin and Krot (2005), (2) Amelin et al. (2002), (3) Russell et al. (1996), (4) Huss et al. (2001), (5) (Mostefaoui et al., 1999), (6) (McKeegan et al., 2000), (7) Kita et al. (2000), (8) Kunihiro et al. (2004), (9) Bizzarro et al. (2004), (10) Markowski et al. (2004).

Kleine et al. (2005)
Fig. 9. Measured \(^{26}\text{Al}/^{27}\text{Al}\) and \(^{182}\text{Hf}/^{180}\text{Hf}\) ratios for CAIs (E60, WA) and angrites D’Orbigny and Sahara 99555. The solid line is the best-fit line given by the CAI and angrite data and has a slope that is consistent with that expected from the \(^{26}\text{Al}\) and \(^{182}\text{Hf}\) half-lives. The dashed line is the decay line calculated using the Hf–W and Al–Mg data for D’Orbigny and Sahara 99555 and the \(^{182}\text{Hf}\) and \(^{26}\text{Al}\) half-lives. Data are from the following references: Al–Mg (Lee et al., 1977; Amelin et al., 2002; Spivak-Birndorf et al., 2005); Hf–W (Kleine et al., 2008a).

Fig. 11. Tungsten model ages for magmatic iron meteorites, calculated using the range of \(\varepsilon^{182}\text{W}\) values for several groups of iron meteorites after correction for cosmogenic effects as reported in Qin et al. (2008b) and using an initial \(\varepsilon^{182}\text{W}\) of CAIs of \(-3.28 \pm 0.12\), as determined in the present study. Numbers in parentheses correspond to the range of corrected \(\varepsilon^{182}\text{W}\) for groups of magmatic iron meteorites. Note that the uncertainties in the W model ages include the uncertainty on the initial \(\varepsilon^{182}\text{W}\) of CAIs. The IVB irons have corrected \(\varepsilon^{182}\text{W}\) values that are lower than the initial \(\varepsilon^{182}\text{W}\) of CAIs, suggesting that the correction procedure employed by Qin et al. (2008b) did not fully correct the cosmic-ray effect on the \(^{182}\text{W}/^{184}\text{W}\) ratio.

Angrites: rapid cooling meteorites;
Ages with different systems should be the same

Core formation of iron meteorites within less than \(-1\) Ma after CAI formation
Pb/Pb ages of Allende CAI does not correspond to angrite ages: Pb loss during shock processes?

Fig. 10. Measured $^{182}$Hf/$^{180}$Hf ratios and $^{207}$Pb–$^{206}$Pb ages in angrites, CAIs and the H5 chondrite Richardton. Data are from the following references: Hf–W data for angrites (Markowski et al., 2007; Kleine et al., 2008a); $^{207}$Pb–$^{206}$Pb ages for angrites (Amelin, 2008b; Connelly et al., 2008b); $^{207}$Pb–$^{206}$Pb age for CAI E60 (Amelin et al., 2002; Amelin et al., 2006); Hf–W data for Richardton (Kleine et al., 2008b); $^{207}$Pb–$^{206}$Pb age for Richardton (Amelin et al., 2005). The $^{182}$Hf/$^{180}$Hf for Northwest Africa 2999 is calculated from a pyroxene-metal isochron. Note that angrites plot on a straight line whose slope is consistent with that expected from the $^{182}$Hf half-life. Note further that the Hf–W ages of angrites and the H5 chondrite Richardton would be inconsistent with their $^{207}$Pb–$^{206}$Pb ages if the $^{207}$Pb–$^{206}$Pb CAI age of 4567.11 ± 0.16 Ma were used as an age anchor.
Figure 1  Pb–Pb isochrons for acid-washed fractions of two CAIs from CV3 Efremovka and for the six most radiogenic fractions of acid-washed chondrules from the CR chondrite Acfer 059. The $^{207}\text{Pb}/^{206}\text{Pb}$ data are not corrected for any assumed common lead composition; 2σ error ellipses are shown. Isochron ages for the two CAIs overlap with a weighted mean age of $4,567.2 \pm 0.6$ Ma, which is $-2.5$ Myr older than the chondrules. Data and figure from Amelin et al. (2002) (reproduced with permission of the American Association for the Advancement of Science from Science 2002, 297, 1678–1683).
Figure 8. Timeline for early solar system events integrating the $^{26}Al-^{26}Mg$ and $^{53}Mn-^{51}Cr$ short-lived chronometers with the absolute timescale provided by the Pb-Pb chronometer. The anchor points (vertical dashed lines) are (i) the Pb-Pb age of CAIs (Amelin et al., 2002) with "canonical" $^{26}Al$ and (ii) the Pb-Pb age of Angrites (Lugmair and Galtier, 1997) with the $^{53}Mn/Mn$ ratio in LEW (Lugmair and Shukolyukov, 1998). Pb-Pb ages are indicated for the filled symbols read against the absolute timescale (central axis); the top axis shows the initial $^{26}Al$/$^{27}Al$ values measured in various phases. Vertical axes intersect at the open symbols in the bottom panel. Squares represent CAIs, diamonds—chondrules, triangles—chondrites, right-triangles—carbonates, inverted triangles—angrites, crossed circles—angrites, diamonds—fayalite. The diamond labeled "HED" represents the $^{26}Al$/$^{27}Al$ ratio of the HED meteorite. The gray arrow represents the $^{26}Al$/$^{27}Al$ ratio of the $^{26}Al$/$^{27}Al$ ratio in the Earth's mantle. The gray square represents the $^{26}Al$/$^{27}Al$ ratio in the Earth's mantle, with the error bar indicating the range of values observed in meteorites.
Exposure ages of meteorites

Figure 1  $^{21}$Ne exposure ages (Myr) of carbonaceous chondrites. Large symbols show group averages.

Figure 2  Exposure ages (Myr) of H chondrites (source Graf and Marti, 1995).
Figure 3  Exposure ages of L-chondrites. Different fills indicate varying degrees of precision (source Marti and Graf, 1992).

Figure 20  CRE ages of various meteorites versus aphelion of parent body or meteoroid.

Herzog (2003)

Figure 18  $^{40}$K/$^{40}$Ar CRE ages of iron meteorites as closed squares (Voshage, 1967; Voshage and Goldstein, 1976; Voshage et al., 1983). $^{34}$Ar/$^{38}$Ar/He ages of iron meteorites as open circles after Lavielle et al. (1985) with noble gas data taken from the compilation of Schultz and Franke (2002). Possible clusters of ages are marked with open circles (see also Eugster, 2003). The angular brackets denote group averages of the two types of ages for groups I-IV. For the ungrouped irons $<T_{40}> = 732 \pm 292$ Myr and $T_{38} = 469 \pm 567$ Myr.
Accretion ages

Figure 8 Initial strontium isotope composition of early lunar highland rocks relative to other early solar system objects. APB: Angrite Parent Body; CEPB: Cumulat Eucrite Parent Body; BSSI: Bulk Solar System Initi (source Halliday and Porcelli, 2001).

Figure 11 Lead isotopic modeling of the composition of the silicate Earth using continuous core formation and a sudden giant impact when the Earth is 90% formed. The impactor adds a further 9% to the mass of the Earth. The principles behind the modeling are as in Halliday (2000). See text for explanation. The field for the BSE encompasses all of the estimates in Galer and Goldstein (1996). The values suggested by Kramers and Tolstikhin (1997) and Murphy et al. (2002) are also shown. The figure is calibrated with the time of the giant impact (Myr). The $\mu$ values are the $^{238}\text{U}/^{204}\text{Pb}$ of the BSE. It is assumed that the $\mu$ of the total Earth is 0.7 (Allègre et al., 1995a). It can be seen that the lead isotopic composition of the BSE is hard to reconcile with formation of the Moon before $\sim45$ Myr after the start of the solar system.

Halliday (2003)
Early Earth, early solar system

Table 3  Recent estimates of the ages of early solar system objects and the age of the Moon.

<table>
<thead>
<tr>
<th>Object</th>
<th>Sample(s)</th>
<th>Method</th>
<th>References</th>
<th>Age (Ga)</th>
</tr>
</thead>
<tbody>
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<td>Earliest solar system</td>
<td>Allende CAIs</td>
<td>U–Pb</td>
<td>Göpel et al. (1991)</td>
<td>4.566 ± 0.002</td>
</tr>
<tr>
<td>Earliest solar system</td>
<td>Efremovka CAIs</td>
<td>U–Pb</td>
<td>Amelin et al. (2002)</td>
<td>4.5672 ± 0.0006</td>
</tr>
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<td>Chondrule formation</td>
<td>Acfer chondrules</td>
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<td>Amelin et al. (2002)</td>
<td>4.5647 ± 0.0006</td>
</tr>
<tr>
<td>Angrites</td>
<td>Angra dos Reis and LEW 86010</td>
<td>U–Pb</td>
<td>Lugmair and Galer (1992)</td>
<td>4.5578 ± 0.0005</td>
</tr>
<tr>
<td>Early eucrites</td>
<td>Chervony Kut</td>
<td>Mn–Cr</td>
<td>Lugmair and Shukolyukov (1998)</td>
<td>4.563 ± 0.001</td>
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<tr>
<td>Earth accretion</td>
<td>Mean age</td>
<td>U–Pb</td>
<td>Halliday (2000)</td>
<td>≥4.49</td>
</tr>
<tr>
<td>Earth accretion</td>
<td>Mean age</td>
<td>Hf–W</td>
<td>Halliday (2000)</td>
<td>≥4.55</td>
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<td>Sm–Nd</td>
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<td>Shih et al. (1993)</td>
<td>4.46 ± 0.07</td>
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<td>U–Pb</td>
<td>Tera et al. (1973)</td>
<td>4.47 ± 0.02</td>
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<td>Kleine et al. (2002)</td>
<td>4.54 ± 0.01</td>
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Figure 13  Schematic showing the timescales for various events through the "Dark Ages" of the Hadean.

Halliday (2003)
Quellen:


**Dickin, A.P. (2005)**: Radiogenic Isotope Geology.


