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Lunar basalt chronology, mantle differentiation and implications for determining the age of the Moon

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ABSTRACT

Despite more than 40 years of studying Apollo samples, the age and early evolution of the Moon remain contentious. Following the formation of the Moon in the aftermath of a giant impact, the resulting Lunar Magma Ocean (LMO) is predicted to have generated major geochemically distinct silicate reservoirs, including the sources of lunar basalts. Samples of these basalts, therefore, provide a unique opportunity to characterize these reservoirs. However, the precise timing and extent of geochemical fractionation is poorly constrained, not least due to the difficulty in determining accurate ages and initial Pb isotopic compositions of lunar basalts. Application of an in situ ion microprobe approach to Pb isotope analysis has allowed us to obtain precise crystallization ages from six lunar basalts, typically with an uncertainty of about ±10 Ma, as well as constrain their initial Pb-isotopic compositions. This has enabled construction of a two-stage model for the Pb-isopic evolution of lunar silicate reservoirs, which necessitates the prolonged existence of high-μ reservoirs in order to explain the very radiogenic compositions of these basalts. Further, once firm constraints on U and Pb partitioning behaviour are established, this model has the potential to help distinguish between conflicting estimates for the age of the Moon. Nonetheless, we are able to constrain the timing of a lunar mantle reservoir differentiation event at 4376 ± 18 Ma, which is consistent with that derived from the Sm–Nd and Lu–Hf isotopic systems, and is interpreted as an average estimate of the time at which the high-μ urKREEP reservoir was established and the Ferroan Anorthosite (FAN) suite was formed.

1. Introduction

The lunar magma ocean (LMO) model, proposed after the first analyses of the Apollo samples, remains the canonically accepted explanation for the magmatic differentiation of the Moon (Wood et al., 1970; Elkins-Tanton et al., 2011). In its current form, the model predicts the formation of a global magma ocean as a consequence of a Moon-forming “giant impact” between the Earth and a Mars-sized body (Hartmann and Davis, 1975) ~60 million years (Ma) after Solar System formation at 4567 Ma (Touboul et al., 2007; Connelly et al., 2012; Avice and Marty, 2014). Development of the major lunar rock suites then occurred as a result of cooling and differentiation of this magma ocean, followed by partial melting of the mantle reservoirs that formed during differentiation. However, the age of the LMO (and the Moon), as well as the time interval required for its crystallization, remain unclear. While W-isotope data, suggesting that the short-lived 182Hf was extinct by the time of lunar formation (Touboul et al., 2007), place the oldest limit for the age of the LMO at about 4500 Ma, the attempts to define the youngest limit are based on studies of the oldest identified lunar rocks, represented by the highland samples. These samples are in-
terpreted as remnants of either the primary anorthositic crust (Ferr-roan Anorthosites; FAN) that formed as part of the LMO differenti-
ation sequence, or early plutonic magmatic rocks (Magnesian- and
Alkali-Suite rocks; Elkins-Tanton et al., 2011; Carlson et al., 2014;
Borg et al., 2015; Pernet-Fisher and Joy, 2016). Crystallization age
estimates for FANs range from 4290–4570 Ma, while the
Magnesian-suite rocks have crystallization ages of 4110–4570 Ma,
but analytical uncertainties for a significant number of these dates
are often close to 100 million years or more (Borg et al., 2015).
Such large uncertainties, and the wide range of overlapping ages
for FAN and Magnesian-suite rocks, have led to the traditional LMO
model for lunar differentiation being challenged in recent stud-
ies (Borg et al., 2011; Carlson et al., 2014; McLeod et al., 2014;
Gaffney and Borg, 2014). In particular, some of these studies sug-
gested that the age of the Moon and the LMO may be more than
100 million years younger (i.e. ~4400 Ma) than indicated by other
isotopic constraints (e.g. Avice and Marty, 2014). The younger age
of the Moon, however, conflicts with dates as old as 4417 ± 6 Ma
determined in zircon grains identified in lunar breccias, which
have been interpreted as placing a lower time limit on the final
stages of LMO crystallization (Nemchin et al., 2009).

The Apollo mare basalts have younger crystallization ages, typi-
cally between 3800–3000 Ma (e.g. Albee et al., 1970; Turner, 1970;
Papanastassiou et al., 1970; Davis et al., 1971; Murthy et al.,
1971; Papanastassiou and Wasserburg, 1971; Compston et al.,
1971, 1972; Nyquist et al., 1975, 1979, 1981; Guggisberg et al.,
1979; Tartèse et al., 2013). Nevertheless, the melts from which these
rocks crystallized are thought to have been sourced from silicate
reservoirs in the lunar mantle generated during the LMO crys-
tallization, therefore providing a link to this initial phase of the
Moon’s evolution. As such, four mare basalts and two KREEP-rich
(material enriched in K, REE and Pb; Warren and Wasson, 1979)
basalt samples have been investigated using high spatial resolu-
tion Secondary Ion Mass Spectrometry (SIMS) to obtain a new set
of Pb isotope data and help constrain early lunar magmatic evolu-
tion.

Previous attempts to apply Pb isotope systematics to fundamen-
tal questions related to the early history of the Moon have utilized
Thermal Ionization Mass Spectrometry (TIMS) analyses of chemi-
sically separated Pb fractions. While this approach has been succes-
sfully applied to the investigation of differentiation processes on the
Earth (e.g. Zartman and Doe, 1981; Kramers and Tolstikhin, 1997),
the inherently low Pb concentrations in lunar samples (relative to
those from the Earth) renders the data particularly susceptible to
the influence of laboratory contamination. The low Pb content in
lunar samples is thought to be a consequence of extensive loss of
volatile Pb early in the history of the Moon (e.g. Tatsumoto,
1970), most likely resulting from the giant Moon-forming im-
pact. This concept is consistent with previous attempts to mea-
sure Pb isotopes in lunar samples (Tatsumoto, 1970; Tera and
Wasserburg, 1972; Tatsumoto et al., 1987; Gaffney et al., 2007a;
Nemchin et al., 2011), which indicate that the ratio of $^{238}$U/$^{204}$Pb
($\mu$) in many lunar rocks is significantly higher (~100–600) than
$\mu$-values inferred for the Earth’s mantle (typically ~8–10; Zartman
and Doe, 1981; Kramers and Tolstikhin, 1997). These high $\mu$-values
of lunar rocks also result in very radiogenic Pb isotope composi-
tions, making it difficult to distinguish between analyses represent-
ing Pb accumulated in the samples by in situ radioactive decay of U
(radiogenic Pb) and those reflecting the initial Pb composition ini-
tiated during crystallization, or various mixtures of the two. The
high spatial resolution SIMS approach undertaken here makes it
possible to limit contamination, focus on the individual phases in
which Pb is concentrated, and overcome the issues associated with
the multi component Pb mixtures present in lunar samples, help-
ing to determine both the crystallization ages of these samples and
their initial Pb compositions.

2. Analytical methods

2.1. Data acquisition

Four of the samples (10044,645; 12039,44; 14072,61 and
15386,46) were thin sections prepared at NASA Johnson Space Cen-
ter and cleaned with isopropyl alcohol before being carbon coated
for Scanning Electron Microscope (SEM) analysis. The 12038,263
and 12063,330 samples were chips mounted in EPOTEK epoxy
resin blocks at Birkbeck College, University of London. Initial Back
Scattered Electron (BSE) and elemental mapping of the four thin
sections was performed at the Open University with a Quanta 3D
Focused Ion Beam (FIB) Scanning Electron Microscope (SEM), while
12038,263 and 12063,330 were mapped at Birkbeck College using a
JEOL JXA-8100 electron microprobe (Fig. A.1). Both instruments
were fitted with Oxford Instruments INCA energy dispersive X-ray
(EDS) detectors. Acquisition of additional BSE and elemental maps
was performed with a Quanta 650 Field Emission Gun (FEG) SEM
and accompanying Oxford Instruments EDS detector at Stockholm
University. For all three SEM setups, the mapping and analysis of
the samples was conducted with an electron beam accelerating
voltage of 15–20 kV at a working distance of 10–15 mm. The BSE
and element maps were then used to identify phases for SIMS
analyses; these included plagioclase and pyroxene grains, as well
as accessory phases (such as K-Feldspar and K-rich glass; Fig. A.2),
which were identified based on regions of high-K and high-Si.

Following SEM documentation of the samples and prior to the
SIMS analyses, the samples were cleaned with isopropyl alcohol
before applying a 30 nm thick gold coat. The Pb isotopic com-
positions of various phases (Table B.1) were determined over three
analytical sessions using a Cameca IMS 1280 ion microprobe at
the NordSIMS facility in the Swedish Museum of Natural His-
tory, Stockholm, using a methodology similar to that outlined in
previous studies (Whitehouse et al., 2005; Nemchin et al., 2011;
Bellucci et al., 2015). Apertures in the primary column were used
to generate a slightly elliptical $\Omega$ sample probe with dimensions
appropriate to the target. The smaller accessory phases (including
K-feldspar and K-rich glass) were analysed using a ~10 $\mu$m spot
beam current ca. 2–3 nA), while larger accessory phases and pla-
 gioclase grains were analysed with either a ~20 $\mu$m spot (beam
current ca. 10–12 nA), or a ~30 $\mu$m spot (ca. 19–20 nA) (Fig. A.2).
Prior to each measurement, an area of 20–35 $\mu$m around the spot
location was rastered for 60 s in order to remove the gold coating
and minimise possible surface contamination. The instrument
was operated in high-transmission mode, corresponding to a transfer
magnification of 160×. In this mode, the field aperture size was
chosen to limit the field of view on the sample surface (i.e. the
area from which ions will be admitted to the mass spectrome-
tor) to be bigger than the un rastered spot but smaller than the
rastered area, further minimising the possibility of surface contami-
nation. The mass spectrometer was operated at a nominal mass
resolution of 4860 (M/ΔM), sufficient to resolve Pb from known
molecular interferences. An NMR field sensor regulated the stabil-
ity of the magnetic field. Pb-isotopes were measured simulta-
neously in multi-collector mode using four low-noise (~0.01 counts
per second) ion counting electron multipliers (Hamamatsu 416) with
electronically gated deadtimes of 60 ns. Background counts for
each channel were measured at regular intervals during each
session. The average background counts for each session were used
to correct the sample analyses and are presented in Table B.2.

Analyses of the USGS basaltic glass reference material, BCR-2G,
were used to correct for mass fractionation and detector gain cal-
boration. The BCR-2G analyses were all within ±5% of the values
presented by Woodhead and Hergt (2000), and these deviations
were used to generate correction factors for the data. The re-
producibility of the BCR-2G analyses for all three sessions were
will lead to points falling to the right of the isochron and towards the composition of modern terrestrial Pb (Stacey and Kramers, 1975). As such, the data for each sample were filtered so as to define a statistically significant isochron (MSWD < 2; probability of fit >0.1) on a $^{207}\text{Pb}^{206}\text{Pb}$ vs. $^{204}\text{Pb}^{206}\text{Pb}$ plot with no analytical points lying to the left of it (Fig. 1). The analyses removed during this filtering process include those with the lowest count rates and, consequently, the largest analytical uncertainties (Fig. A.4; Table B.1). An exception to this was the Apollo 15 KREEP basalt sample (15386,46), for which it was not possible to obtain an isochron with an MSWD <5.3 based on the current dataset. In addition to the three main components discussed, some contribution to the measured Pb isotope compositions may in theory also come from meteoritic material and contamination from an old (>3.9 Ga) lunar Pb component identified in previous studies (Tera and Wasserburg, 1974; Borg et al., 2011), which has been interpreted as being distributed across the lunar surface due to volatilization of Pb by impacts during the period of heavy bombardment that characterized early lunar history. A conceivable mechanism for introducing one or both of these components into unaltered basaltic samples that postdate this period is through interaction of the lower part of the basaltic flow with the lunar surface during the basalt extrusion. However, the bulk of the basaltic flow is unlikely to have been affected. Furthermore, mixing of these components during eruption of the basalts will have resulted in an effective mixing and homogenization of different initial Pb components on the sample scale and will, therefore, have no effect on the way the Pb/Pb isochrons are constrained. The possible presence of such contamination may be confirmed or disproved with further analyses of samples interpreted to represent different parts of individual basalt flows.

The data were also filtered in order to remove any analyses where the count rates for any of the Pb isotopes were very low and effectively below detection limits (<3× the measured background count rates; Tables B.1; B.2). In the dataset presented here, it is also notable that the analyses with the largest analytical errors and lowest Pb counts are also typically those with the highest $^{204}\text{Pb}^{206}\text{Pb}$ ratios (Fig. A.3; Table B.1). This is interpreted as evidence that the points with the lowest counts of Pb are also those that are most affected by the presence of terrestrial contamination.

3. Results

The Apollo 11 basalt (10044) yields a crystallization age of 3688 ± 5 Ma, while the three Apollo 12 basalts (12038, 12039 and 12063) have crystallization ages of 3242 ± 13 Ma, 3129 ± 10 Ma and 3193 ± 11 Ma, respectively (Fig. 2). A crystallization age of 3905 ± 8 Ma is determined for the high-Al (and KREEP-rich) Apollo 14 basalt (14072), and 3884 ± 76 Ma for the Apollo 15 KREEP basalt (15386; Fig. 2). The uncertainties for each of these ages are quoted at the 2σ level.

For 10044, the analysis with the highest $^{207}\text{Pb}^{206}\text{Pb}$ ratio (and therefore the most likely to represent an initial Pb component) is within error of several analyses obtained in a separate study for a similarly aged, low-K, ilmenite basalt 10047 (Fig. 2a; Rasmussen et al., 2008). As such, we have taken an average of the analyses in our dataset with the highest $^{207}\text{Pb}^{206}\text{Pb}$ ratios and those in the data presented by Rasmussen et al. (2008), as being the best indication of the initial Pb component in the Apollo 11 ilmenite basalt suite. In 12038 and 12063, the highest $^{207}\text{Pb}^{206}\text{Pb}$ compositions are represented by a single measurement in each sample. Therefore, we have no option but to use these compositions as the closest estimate for the initial Pb component in each sample. Although these initial Pb values may slightly underestimate the $^{207}\text{Pb}^{206}\text{Pb}$ and $^{204}\text{Pb}^{206}\text{Pb}$ ratios of the true initial composition, they nonetheless provide the lowest possible limit. In 12039, we interpret a clustering of K-feldspar compositions as represent-
Fig. 2. $^{207}\text{Pb}/^{206}\text{Pb}$ vs. $^{204}\text{Pb}/^{206}\text{Pb}$ plots of filtered datasets from: (a) 10044,645; (b) 12038,263; (c) 12039,44; (d) 12063,330; (e) 14072,61; and (f) 15386,46. The isochrons determined for each sample have been indicated with dashed lines. Also shown are the growth lines for our two-stage model of lunar Pb isotope growth for each sample, from the model Pb isotopic composition at $t_1$ (4376 ± 18 Ma) to the time of crystallization. Error bars are at 2σ uncertainties.
ing the initial Pb component of the sample (Fig. 2c). A weighted average of all analyses from this cluster was taken as the best estimate for this initial Pb composition. For 14072, three plagioclase analyses were interpreted as providing the closest approximation to initial Pb compositions (Fig. 2e). In the case of 15386, the analyses with the highest $^{207}$Pb/$^{206}$Pb ratios fall significantly to the right of the sample isochron in the $^{207}$Pb/$^{206}$Pb vs. $^{204}$Pb/$^{206}$Pb coordinate space (Fig. 3). This is interpreted to be a result of minor terrestrial contamination. Consequently, an initial Pb composition for this sample has been estimated by calculating the intercept between the sample isochron and the mixing trend between the initial Pb component and the terrestrial contaminant (Fig. 3).

The measured $^{238}$U/$^{208}$Pb ratios determined in several of the samples provide a first order check that the obtained Pb isotopic compositions do indeed represent a mixture between initial and radiogenic Pb. In the locations interpreted as having Pb isotopic compositions closer to initial values, the U content is confirmed as being lower than in the locations interpreted as containing radiogenic Pb supported by the in situ decay of U (Fig. 4).

4. Discussion

4.1. Comparison of basalt ages with previous studies

The crystallization ages determined here are in broad agreement with previous studies (Fig. 5; Table B3; cited in the following discussion with 2σ uncertainties). Two separate studies of 10044 determined Rb–Sr ages of 3700 ± 70 Ma and 3710 ± 110 Ma (Albee et al., 1970; Papanastassiou and Wasserburg, 1971). Early analyses of this sample also determined a range of $^{39}$Ar–$^{40}$Ar plateau ages; 3740 ± 50 Ma, 4000 ± 70 Ma and 3710 ± 40 Ma (Turner, 1970; Davis et al., 1971; Guggisberg et al., 1979). Note, these $^{39}$Ar–$^{40}$Ar and Rb–Sr ages reflect the original reported values and have not been recalculated for updated monitor ages and decay constants. More recently, Tartèse et al. (2013) determined a $^{207}$Pb/$^{206}$Pb age of 3722 ± 11 Ma for the sample with in situ analyses of tranquillityte. The tranquillityte $^{207}$Pb/$^{206}$Pb age is significantly older than the age determined in this study for 10044 (3688 ± 5 Ma), but was determined from the weighted average of $^{208}$Pb/$^{206}$Pb ages from very radiogenic Pb isotope compositions, which were not corrected for the presence of initial Pb. These compositions all lie at the extreme lower end of the isochron in Fig. 2a. If the tranquillityte...
The initial Pb compositions estimated for all the investigated samples are extremely radiogenic, indicating prolonged growth of radiogenic Pb in high-μ reservoirs, which is consistent with the hypothesis that the Moon experienced profound Pb-loss during its early evolution. Although single stage model ages are clearly an oversimplification of the true Pb isotope evolution of lunar reservoirs, they can provide an estimate of the time when such high-μ reservoirs were established, providing that there was no significant change in the μ-values of the different reservoirs since the initial loss of Pb from the Moon. These model ages were calculated for the source reservoirs of each of the samples based on the obtained Pb/Pb isochron ages and initial Pb isotope compositions, giving: 4525 ± 32 Ma for 10044, 4423 ± 20 Ma for 12038, 4466 ± 10 Ma for 12039, 4417 ± 20 Ma for 12063, 4423 ± 77 Ma for 14072 and 4384 ± 63 Ma for 15386 (all uncertainties are 2σ). The scatter of these model ages indicates that the evolution of Pb recorded by the analysed samples cannot be explained by a single simple differentiation event on the Moon that coincided with its formation and simultaneously resulted in the overall Pb loss from the Moon and differentiation of the lunar mantle into several reservoirs with distinct μ-values.

Fortunately, Pb isotopes also offer an opportunity to evaluate two-stage models, with the advantage of gaining additional information related to the Pb isotope evolution of the Moon, with an assumption that the formation of the Moon and its differentiation are separated by some time exceeding the errors defined by the analytical uncertainties. These two-stage source reservoir Pb evolution models were calculated, assuming the sources of mare basalts and the primordial KREEP reservoir (i.e. urKREEP; Warren and Wasson, 1979) differentiated at approximately the same time (t1) from a single primordial lunar reservoir with a common μ-value (μ1), which acquired its Pb isotope composition as a result of radioactive decay before differentiation (between t0 and t1; Fig. 7a). Starting with the Pb isotope composition of Canyon Diablo Troilite (CDT values; Göpel et al., 1985), a two-stage model was calculated for a range of t0 times (between 4567 and 4400 Ma) in order to obtain corresponding values for μ1 (primordial lunar reservoir) and μ2 (six values for each basalt investigated), as well as the age of differentiation and the Pb isotope composition at this time (see Appendix C for discussion of model calculations). The model was calculated using nonlinear regression fitting function in Matlab in order to optimize the fit with the ages and initial Pb compositions of all investigated samples. These calculations were initially performed assuming that the μ-values of the Earth–Moon precursor materials were low enough that there was no significant evolution of the Pb isotope compositions away from CDT values between 4567 Ma (Connelly et al., 2012) and t0. Additional model runs were then performed assuming a range of precursor μ-values, allowing for Pb isotope compositions more radiogenic than CDT prior to the formation of the Moon. The results (Figs. 7b and 8a) indicate that changing t0 does not produce any significant variation in time of differentiation, with t1 being centred at 4376 ± 18 Ma (2σ) for t0 of 4500 Ma and estimates for all
other \( t_0 \) values ranging within this 18 Ma uncertainty (Table B.4). A similarly limited variability is observed for the Pb isotope composition at the time of differentiation (\(^{204}\text{Pb}/^{206}\text{Pb} = 0.036 \pm 0.004\) and \(^{207}\text{Pb}/^{206}\text{Pb} = 1.59 \pm 0.02\); Fig. 7b) and, as a consequence, the \( \mu_2 \) values (372 ± 14 for 10044, 534 ± 4 for 12038, 643 ± 4 for 12039, 419 ± 7 for 12063, 2604 ± 185 for 14072 and 3675 ± 271 for 15386; Table B.4). The \( \mu_1 \) value is the only parameter that changes significantly with changing the model start time (274 ± 50 when \( t_0 = 4567 \text{ Ma} \) and 1083 ± 184 when \( t_0 = 4425 \text{ Ma} \); Fig. 8a). The uncertainties on these model values are quoted at the 2σ level. The model calculation becomes less stable when \( t_0 \) moves closer to \( t_1 \), indicating that it is becoming increasingly difficult to fit all the analytical data into a two-stage model. This loss of stability is reflected in the increased uncertainty of \( \mu_1 \) estimates corresponding to the younger \( t_0 \) values (e.g. \( \mu_1 = 1557 \pm 727 \) when \( t_0 = 4400 \text{ Ma} \); Fig. 8a; Table B.4).

Initial \(^{208}\text{Pb}/^{204}\text{Pb} \) compositions were determined for all of the mare basalts and 14072 (it was not possible to determine the initial \(^{208}\text{Pb}/^{204}\text{Pb} \) composition of 15386 with the current data set). Within the uncertainties of these initial compositions, the samples can all be modelled as having evolved from \( t_1 \) (4376 Ma) until their respective crystallization ages with a \( k \)-value (\(^{232}\text{Th}/^{238}\text{U} \)) of 3.90 ± 0.64 (Fig. 9), which is similar to estimates for the terrestrial and martian systems (Zartman and Doe, 1981; Kramer and Tolstikhin, 1997; Bellucci et al., 2015).

4.3. Model interpretation and implications

Two-stage Pb evolution models, such as the one proposed by Stacey and Kramers (1975), have been an important initial step in the understanding of terrestrial evolution and chemical differentiation. Although these models cannot describe the full complexity of the Earth’s history, when combined with information obtained from other isotope systems, they have led to the development of the modern field of mantle chemical geodynamics over the past 40 years. The Stacey and Kramers (1975) model is still widely used in all chronological studies to determine composition of initial Pb when applying corrections to the U–Pb data obtained for the principal geochronometers, such as zircon, monazite, titanite etc. Similarly, the Pb evolution model presented here for the Moon is the first step in a better understanding of the dynamics and differentiation of lunar mantle. It is important, therefore, to determine the significance and limitations of the ages and \( \mu \)-values defined by this model.

4.3.1. \( \mu \)-Value of the Bulk Silicate Moon and implications for the age of the Moon

The two-stage model calculations based on the currently available data cannot provide a unique and precise set of values for the age of the Moon and the \( \mu \)-value of the Bulk Silicate Moon.
Moon. Sequestration of Pb into the lunar core as it formed could provide one alternative way to increase the μ-value of the primordial lunar reservoir. In the case of the Earth, formation of a volumetrically large core only generated, at most, a \( \sim 10^x \) increase in terrestrial mantle μ-values (Kramers and Tolstikhin, 1997). Additionally, the segregation of the Martian core did not increase the μ-value of the Martian mantle by a substantial amount (Bellucci et al., 2015). By comparison, the Moon has a relatively small core (\( \sim 20\% \) of its mean radius; Weber et al., 2011), making it impossible by mass balance to explain the high-μ (exceeding 300–400) by core segregation, which instead would have more likely resulted in a comparatively minor, and currently unresolved, μ-value increase during the first stage of the model (between \( \mu_1-\mu_2 \)).

An alternative to profound Pb loss during the formation of the Moon, which can possibly explain the Pb isotope composition observed in lunar samples, is to consider high-μ precursor material, for example, a high-μ impactor colliding with the Earth in the Moon forming giant impact. This scenario would allow for more radiogenic Pb isotope compositions to have evolved prior to \( t_0 \). In a simplistic case where the precursor material is considered as a single reservoir (i.e. neglecting the effects of mixing between the early Earth and the giant impactor), the Pb isotope compositions can be explained with a μ-1-value of \( 791 \pm 301 \) when \( t_0 \) is 4400 Ma, if the precursor has a μ-value of 200 and formed at about the same time as the Solar System (Fig. 8a). However, giant impact models commonly predict that the primordial Moon would contain tens of percent (by mass) of material from the Earth (Canup and Asphaug, 2001). Since a high-μ impactor would be expected to have a low concentration of Pb due to the volatile nature of Pb and refractory nature of U, even small amounts (e.g. 10%) of terrestrial material (assuming a μ-value of \( \sim 8 \)) will dominate the Pb budget of any mix with such an impactor. For example, even an impactor μ-value of 2000 (which seems highly unlikely given the comparatively low μ-values observed in other planetary bodies; Zartman and Doe, 1981; Kramers and Tolstikhin, 1997; Gaffney et al., 2007b; Amelin, 2008; Bellucci et al., 2015), would be little different to CDT starting composition (Fig. 8b), unless a high-μ early Earth is considered (Albarede, 2009).

4.3.2. Reservoir differentiation and crystallization of the Lunar Magma Ocean

The two isotopically distinct groups of reservoirs predicted in the second stage of this model have about a five-fold difference in μ-value. One is the source of the mare basalts and the other is a KREEP-rich reservoir (urKREEP; Warren and Wasson, 1979). Both are anticipated to form in a classic LMO model crystallization sequence (e.g. Wood et al., 1970; Elkins-Tanton et al., 2011). While the urKREEP represents a very small (a few percent maximum) portion of the LMO residual melt, the source of the mare basalts is considered to be an olivine-pyroxene cumulate formed during LMO crystallization (e.g. Wood et al., 1970; Snyder et al., 1992). Having estimates of the μ-values for these reservoirs allows for some simple first order constraints to be made in relation to the differentiation of the LMO based on the known behaviour of U and Pb in magmatic systems, even though exact partition coefficients for these elements are not very well constrained. Both are considered to be very incompatible with the available estimates of partition coefficients between most rock forming minerals and basaltic melt falling below 0.01 (e.g. Green, 1994; Fonseca et al., 2014). If that is the case, more than 90% of the original U and Pb will have remained in the residual melt, even after 99.99% of the LMO crystallized. Therefore, in the absence of phases concentrating significant amounts of Pb, fractional crystallization of the LMO appears not to be capable of changing the residual melt μ-value by more than about 10%, and certainly not the 2–10\( \times \) increase predicted by the model for urKREEP, unless the initial μ-value that the

\[ \text{Fig. 9. Plot of the best estimates for the initial } ^{206}\text{Pb}^{204}\text{Pb} \text{ compositions plotted against } \frac{^238\text{U}}{^206\text{Pb}} = k \frac{^235\text{U}}{^206\text{Pb}} (t_0 - \frac{^6}{^206} \mu_1), \text{ such that the slope of the trend is equal to the } \kappa \text{-value of the system, given that: } \frac{^206\text{Pb}}{^204\text{Pb}} = k \frac{^238\text{U}}{^206\text{Pb}} (t_0 - \frac{^6}{^206} \mu_1) + \frac{^206\text{Pb}}{^204\text{Pb}}. \]

(i.e. \( \mu_1 \)) acquired during its formation. Nevertheless, the model highlights the relationship between the time of lunar formation and the \( \mu_1 \)-value (Fig. 8), such that if one of these parameters can be determined independently, the other one would be defined by the Pb isotope systematics described in this study. For example, accepting the recently proposed young age for the Moon (Borg et al., 2011; Carlson et al., 2014; McLeod et al., 2014), requires a comparatively high first stage \( \mu_1 \)-value (1063 ± 184 when \( t_0 = 4425 \) Ma or 1557 ± 727 when \( t_0 = 4400 \) Ma), as well as a 2–5\( \times \) decrease of the \( \mu_2 \)-values in the mare basalts relative to that of the undifferentiated LMO and a 2–3\( \times \) increase of the \( \mu_2 \)-values in urKREEP. Conversely, older lunar formation requires more moderate \( \mu_1 \)-values (e.g. 462 ± 46 when \( t_0 = 4500 \) Ma), a negligible change in μ-values between the undifferentiated LMO and mare basalts, and a 5–10\( \times \) increase to reach the urKREEP \( \mu_2 \)-values. Regardless of the precise age of the Moon, the model places constraints on LMO crystallization and the formation of major lunar silicate reservoirs (Fig. 8a; Table 8.4), as the Pb isotope modelling restricts the range of \( \mu_2 \)-values for both the mare basalt sources (\( \sim 300–700 \)) and urKREEP (\( \sim 2500–3500 \)).

So long as U and Pb partitioning behaviour between melt and the main rock forming minerals is well constrained, any LMO differentiation model that defines a specific set of \( \mu_1-\mu_2 \) relationships will uniquely determine the age of the Moon. While estimates of U and Pb partition coefficients are available (e.g. Green, 1994; Bindeman et al., 1998; Bindeman and Davis, 2000; Fonseca et al., 2014), to date, no study has determined coefficients for both elements from a single set of experiments. Thus, prediction of the relative behaviour of these elements in magmatic fractionation processes will be imprecise and prevent full application of the Pb-isotope model presented here.

While the two-stage Pb evolution model appears to have the capacity to describe the early evolution of the Moon within the framework of the LMO crystallization sequence, it is necessary to consider several complications, which may violate the model. The estimated first stage μ-value is likely to be an integrated result of several discrete or continuous changes in the U–Pb ratio of the primordial lunar reservoir, rather than a single increase at the starting time of the model. However, achieving a high-μ early in the differentiation history is necessary to attain the observed radiogenic Pb compositions. This supports the conclusion of earlier work that Pb volatility is the major factor in its depletion in the
Moon acquired during its formation was similar to that estimated for urKREEP. Alternatively, the observed increase in the urKREEP $\mu$-value could be unrelated to the magmatic differentiation process and explained instead by a significant loss of Pb from the residual LMO liquid. For example, degassing of the residual LMO liquid has been proposed following comparative studies of Cl isotope signatures in mare basalts and KREEP-rich rocks (Boyce et al., 2015; Barnes et al., 2016).

Two groups of minerals provide likely candidates as Pb sinks in the assemblages crystallizing from the LMO; sulfides and feldspars. If all U stays in the residual melt, fractionation of an assemblage with a bulk Pb partition coefficient of around 0.25 is required to achieve a $2 \times$ increase in the $\mu$-value of the remaining 5% of melt. Taking into account that it is likely only part of the entire LMO cumulate pile is characterized by increased Pb concentrations, a partition coefficient of $0.4-0.5$ would be more realistic to explain the urKREEP $\mu$-value. Fractionation of sulfides during LMO crystallization is feasible, but the lack of samples representing this process in the existing collection of lunar rocks means that this possibility cannot be tested in great detail. However, the FAN samples thought to represent the primary feldspathic crust are an important product of LMO crystallization. Results of Pb isotope studies of the FAN samples (e.g. Premo et al., 1999) indicate that plagioclase in these samples has a range of $\mu$-values between $\sim1$ and 30, supporting the assumption that this mineral can extract two orders of magnitude more Pb than U out of its parent melt. This is further supported by studies of U and Pb partitioning in plagioclase Bindeman et al., 1998; Bindeman and Davis, 2000.

A number of recent chronological results are easier to explain if the observed increase in the $\mu$-value characterizing KREEP-rich materials is linked to FAN formation, in particular younger ages of some FAN samples similar to the Nd and Hf model ages of KREEP-rich samples (Borg et al., 2011; Carlson et al., 2014; McLeod et al., 2014). In fact, all of the related analytical data recently obtained, the most difficult to explain is the range of $^{142}$Nd model ages of several FAN samples presented by Boyet et al. (2015), which indirectly supports protracted crystallization of the anorthositic lunar crust, and the large $^{142}$Nd deficit in one sample (62255), which is impossible to explain if it formed after $\sim4440$ Ma. The two-stage Pb model likely provides a good approximation for the time ($4376 \pm 18$ Ma) for the crystallization of the mineral assemblage that ultimately resulted in FAN formation as well as producing the enrichment of incompatible elements in the residual melt, leading to the high-$\mu$ isotopic signature of urKREEP. This Pb model age is further supported by a majority of the Nd and Hf model ages for similar rocks (Borg et al., 2011; Carlson et al., 2014; McLeod et al., 2014), and may not date the precise timing of urKREEP (and FAN) formation, but rather an average time of a potentially complex process that could have extended over a few million to a few tens of millions of years.

As discussed in the previous section (4.3.1), since U and Pb partitioning behaviour is not well understood, the two-stage Pb model presented here is currently not in conflict with either an “old” or a “young” Moon. Resolution of the conflict regarding the formation of the Moon has implications for defining a unique time for the formation of the mare basalt sources. If the Moon is “old”, and the $\mu$-values of the mare basalt sources are similar to that of the primordial lunar reservoir, then the time of mare basalt source formation is difficult to determine as it will effectively be invisible within the two-stage model, where formation of urKREEP and FAN dominates the fractionation of U and Pb in the LMO. More incompatible behaviour of U (relative to Pb), will imply late formation of both the LMO and the Moon, in which case the $4376 \pm 18$ Ma age will also encompass the time of mare basalt source formation, satisfying the main assumption of the two-stage model, that the differentiation took place at a single point in time from a previously undifferentiated reservoir.

Similar uncertainty with low-Ti basalts exists in the Nd isotopic dataset. The olivine and pigeonite basalts investigated by McLeod et al. (2014) are not fractionated far enough from the primitive reservoir to provide any definitive differentiation time constraints on their own in the $^{142}$Nd-$^{144}$Nd system. This behaviour is remarkably similar to that of the Pb isotope systematics described above, which indicates that the mare basalt sources do not show a significant change in either their $\mu$-value or Sm–Nd ratio, relative to that of the undifferentiated Moon and, consequently, the model age of differentiation defined by both systems at about 4370–4380 Ma defines the average time of urKREEP differentiation and FAN formation, but not necessarily the separation of the mare basalt sources. As such, neither the Pb nor Nd isotope systems can be used to determine exactly when these mare basalt mantle sources formed. Strictly speaking the “young” ($\sim4340–4390$ Ma) ages in the McLeod et al. (2014) data are defined by an Apollo 15 KREEP basalt and Apollo 17 high-Ti basalts, which are assumed to represent some of the final differentiates of the LMO, i.e. ilmenite-rich cumulates and urKREEP (e.g. Warren and Wasson, 1979; Snyder et al., 1992).

5. Conclusions

By using SIMS analyses to determine the Pb isotope compositions of multiple phases in lunar basalts, crystallization ages have been determined that are both consistent with previous studies and, in most cases, significantly more precise. In addition to these ages, the data have been used to constrain the isotopic compositions of Pb incorporated into the basalts when they crystallized, from which a two-stage model has been constructed for the Pb isotopic evolution of major lunar silicate reservoirs. This model is necessarily a simplification of the early magmatic differentiation of the Moon, omitting the effects of processes such as core segregation. Nonetheless, it provides a framework upon which subsequent studies can build, and several important constraints for the timing of key stages in this evolution and Pb isotope compositions at these stages. In principle, the model is capable of resolving uncertainty over the age of the Moon, but this requires better knowledge of U and Pb partitioning behaviour in magmatic systems. The Pb isotope data support a major magmatic event on the Moon at $4376 \pm 18$ Ma, possibly representing the average age of urKREEP and FAN formation. Following this event the mare basalt sources evolved with relatively low $\mu$-values (370–640), while the KREEP reservoir acquired a significantly larger $\mu$-value (2600–3675).

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Appendix A. Supplementary material

Supplementary material related to this article can be found online at http://dx.doi.org/10.1016/j.epsl.2016.07.026.