Rb-Sr constraints on the age of Moon formation

Elsa Yobregat, Caroline Fitoussi, Bernard Bourdon

Rb-Sr constraints on the age of Moon formation

Elsa Yobregat[1], Caroline Fitoussi[1], Bernard Bourdon*[1]

[1] Laboratoire de Géologie de Lyon, ENS Lyon, CNRS and UCBL

*Correspondence to: bernard.bourdon@ens-lyon.fr

Keywords: Moon, Strontium isotopes, chronology, anorthosite

Abstract:

Determining the age of the Moon, which is commonly considered as the termination of Earth accretion has been a complex challenge for geochronology. A number of methods have been used to delineate the age of the Moon based either on absolute chronology of lunar rocks or have relied on more indirect methods using short-lived nuclides such as ^{182}Hf that was present in the early history of the Solar System. Model ages usually require some assumptions that are sometimes controversial or harder to verify.

In this study, new high precision Sr isotope data (2.4 ppm, 2SD) were obtained for a well-dated lunar anorthosite (60025) in order to better constrain the initial $^{87}Sr/^{86}Sr$ of the bulk silicate Moon. This new data is then used to model the Sr isotope evolution of the Earth-Moon starting from the beginning of the Solar System. To comply with the Hf-W and stable isotope constraints, we then assume that the Earth and Moon were equilibrated at the time of Moon formation. By investigating systematically all the sources of uncertainties in our model, we show that compared with previous work on anorthosite, one can tighten the constraints on the youngest age of Moon formation to no more than 79 Ma after the beginning of the Solar System, i.e. the Moon cannot be younger than 4488 Ma.

Highlights:

- New high precision Sr isotope data yields a constraint on Moon formation.
- The youngest age of the Moon is 79 Ma after the beginning of the Solar System.
- This model age includes all sources of uncertainties using MonteCarlo simulations.

1. Introduction

The Moon is thought to have formed after the collision of a large planetary object with the proto-Earth, yielding a disk of vapor and liquid that later accreted the Moon. Obtaining a reliable constraint on the age of this event has proven difficult, due to the absence of lunar rocks or minerals directly recording this event. An indirect approach based on the ^{182}Hf-^{182}W isotope system (Touboul et al., 2007) was used to argue that the Moon should have formed at least 45 Ma after the beginning of the Solar System based on the observation that the ^{182}W abundance in the Moon and the Earth were almost identical, despite having a difference between Hf/W ratios in lunar and terrestrial rocks. This approach was questioned by König et al. (2011) and (Münker, 2010) who argued that the Moon and Earth had almost identical Hf/W ratios, which implies that no ^{182}W difference would be expected between the Moon and the Earth, if they initially formed from the same material. Subsequently, the studies of Touboul et al. (2015) and (Kruijer et al. (2015) have shown that there was a small (≈15 ppm) but resolvable difference between Moon and Earth in ^{182}W abundances. However, this difference was ascribed to different proportions of late accreted material between Moon and Earth. More recently, Thiemens et al. (2019) have shown that reliable Hf/W ratios for the bulk Silicate Moon can only be obtained using low-Ti basalts, as high-Ti basalts have saturated metal and Ti-rich phases, which makes an estimate of their Hf/W source difficult. On this basis, the observed Hf/W ratios of low-Ti basalts ranging between 30-48 was used to infer the Hf/W ratio of the bulk silicate Moon (BSM), which proved significantly higher than that of the BSE (26); this difference between the Moon and the bulk silicate Earth was then used to determine an ^{182}Hf-^{182}W age for the Moon formation of 50 Ma based on the observed small 15 ppm difference in ^{182}W abundance between Moon and Earth (Kruijer et al., 2015; Touboul et al., 2015). In that study, however, it was considered that the excess ^{182}W was not related to late accretion. The small difference in ^{182}W abundance between present-day Earth and Moon was then interpreted as reflecting in-growth of ^{182}W in the Bulk Silicate Moon (BSM) relative to the BSE and a range of Moon formation ages between 45 and 60 Ma was inferred (Thiemens et al., 2019). However, this age determination was questioned by Kruijer et al. (2021) who showed that if one takes into account the possible difference between HSE budget between Earth and Moon, then there is no longer a difference in ^{182}W abundance between Earth and Moon, which means that the Hf-W constraints on the age of the Moon indicates that it must have formed later than ≈50 Ma (upper limit), as originally inferred in Touboul et al. (2007), even though the constraints on the Hf/W ratio in the BSM and BSE are admittedly not very tight.

Using a very different approach, Jacobson et al. (2014) demonstrated that the mass of material accreted to the Earth after the Moon-forming impact, the so-called 'late veneer' was a function of the time at which the Moon formed. This is simply because the mass of objects in the protoplanetary disk available to accrete to Earth is decaying with time. Thus, by measuring the abundance of highly siderophile elements in the Earth's mantle, which is a proxy for the mass of late accreted material, it is possible to constrain an age for the Moon formation to be 95±32 Ma at the earliest after the beginning of the Solar System. This approach has been reasserted by recent work showing the unfractionated and homogeneous HSE pattern in mantle derived rocks (e.g. Paquet et al. 2023) that is consistent with late accretion, rather than intra-mantle processes (see also Righter et al. 2023). Using a completely different method, Barboni et al. (2017) have argued that Hf isotope data for lunar zircons suggest that the Moon was already differentiated 58±20 Ma after the start of the Solar System and that the Moon must have formed before that time. However, this age determination has been questioned in the study of Borg and Carlson (2023) who argued that this data set was not consistent with earlier work on whole rocks (Gaffney and Borg 2014; Sprung et al. 2013) or with earlier Lu-Hf data on zircons (Taylor et al. 2009).

Based on internal isochrons obtained for ferroanorthosite rock 60025, Borg et al. (2011) obtained a consistent ^{146}Sm-^{142}Nd, ^{143}Nd-^{147}Sm and U-Pb age for ferroanorthosite rock 60025. If there had been a magma ocean in the history of the Moon, this age would indicate that the formation of anorthositic crust of the Moon lasted at least until 200 Myr after the beginning of the Solar System (see also Borg et al., 2020, 2019, 2011). Whether these anorthosite ages accurately represent the age of the Moon is a matter of debate. At the very least, the anorthosites that represent the oldest rocks found on the Moon give the youngest bound for the age of Moon formation. If one argues furthermore that the magma ocean was short-lived (at most 10 Ma, according to Elkins-Tanton et al., 2011), this age would indicate that the Moon would have formed only shortly before that. To reconcile this young age for the Moon with the relatively older sets of ages, several models for a long duration of the lunar magma ocean (LMO) have been proposed, with the latest version proposing a preferred model where the magma ocean crystallization stopped 170 Myr after accretion (Maurice et al., 2020). This study combining a thermal model with geochronological constraints concludes that the age of the Moon should be 143±25 Myr after the start of the Solar System. A more recent study has found that zircons from Apollo 17 impact melt breccia yielded U-Pb ages as old as 4460±31 Ma (Zhang et al., 2021), which yields an age of Moon formation at

107 Ma or earlier. Last, Connelly and Bizzarro (2016) have used the Pb isotope composition of the bulk silicate Earth to constrain the time of the Moon-forming impact indirectly and obtained a Pb model age as young as 4424-4417 Ma or 144-151 Ma after the beginning of the Solar System. Thus, there is a strong debate, that is still open, regarding the age of the Moon with some proponents for an 'Old Moon' based on Hf-W-HSE budgets, while the Sm-Nd/Lu-Hf U-Pb chronology of lunar rocks and thermal models would rather argue for a 'Young Moon'.

By assuming that the Sr initial isotope composition of the Solar System is known and that the Rb/Sr ratio of the Earth-Moon system is identical to that of the bulk Silicate Earth, Halliday (2008) calculated an age for the formation of the Moon that ranged between 50 and 100 Ma, if one considers the total uncertainties in Rb/Sr and $^{87}Sr/^{86}Sr$ ratios. Using a similar approach based on a two-component model for the Earth, Mezger et al. (2021) reported an age range between 45 and 75 Myr. More recently, Borg et al. (2022) used published Rb-Sr data on lunar rocks to derive constraints on both the age of the Moon and the volatile evolution of Earth and Moon building blocks. Their conclusions were that Theia (i.e. the Moon-forming impactor) and the proto-Earth must have formed from precursors that were already depleted in moderately volatile elements such as Rb. In their scenario, the Moon could have formed from 90% of a highly Rb-depleted precursor mixed with 10% of proto-Earth material. By assuming that proto-Earth and Theia did not equilibrate isotopically at the time of the impact, the age of the Moon can be as late as 137 Ma after the beginning of the Solar System (e.g. Table S1, Model 6). These conclusions relied on the hypothesis that Theia and proto-Earth were made of isotopically similar materials in terms of Ti, Cr and O isotopes (as in Dauphas et al., 2014), arguing that these bodies must have both come from the same reservoirs in the inner Solar System.

Here, we have measured precisely the initial $^{87}Sr/^{86}Sr$ of ferroanorthosite 60025 and revisited these interpretations, thereby providing a new view on Rb-Sr systematics of the Earth-Moon system. In this study, all the ages (unless specified) are reported relative to the beginning of the Solar System represented by the age of Calcium-Aluminium rich inclusions (CAI) dated at 4567 Ma (Amelin et al., 2010; Connelly et al. 2012) with the U-Pb method.

2. Methods

2.1 Sample preparation and Sr separation

The 60025 rock is a pristine coarse-grained ferroan anorthosite from the lunar highlands sampled during the Apollo 16 mission. This sample is a breccia that is thought to come from ejecta from surrounding large impact basins. Thus, FAN 60025 represents a plutonic rock cooled at depth that must have been brought to the surface from a depth possibly as large as 50 km (e.g. Spudis et al. 1989). The fragment analyzed in this study is nearly composed of pure plagioclase with only a few percent of mafic minerals. The sample was first coarsely crushed in an agate mortar, then mafic minerals, olivines, pyroxenes and oxides were handpicked under a binocular microscope. The plagioclase fraction (281.4 mg) was crushed into a fine powder and dissolved in a Savillex PFA beaker using an HF: HNO_3 mixture. The sample was then dried down and dissolved several times in concentrated HCl to remove fluorides.

The solution was loaded in 6 M HCl on a column filled with AG1 X8 (100-200 mesh size) anionic resin to remove Fe, which was retained in the column, while Sr was directly eluted. The sample was converted to nitrate form, then loaded in 3 M HNO_3 on 0.5 ml of Eichrom Sr-resin^{TM}. Matrix elements, including Rb, were eluted in 2.5 ml of 3 M HNO_3, then Ba was stripped from the resin with 6.5 ml of 7.5 M HNO_3 and finally Sr was collected with 10 ml of 0.05 M HNO_3. The Sr fraction was evaporated and redissolved in aqua regia several times to remove traces of organic material released from the Sr-resin. The whole procedure yielded Sr blanks less than 30 ± 5 pg and Rb blanks <45 pg (representing less than 10% of the Rb present in the aliquot used for measuring Rb). The Rb and Sr concentrations were measured on a ~20 mg aliquot using a Thermo Element high-resolution ICP-MS. This instrument has very low detection level (~20 ppt) required to measure the very low Rb content of ferroan anorthosites.

2.2 Mass spectrometric measurements of Sr isotope ratios

The $^{87}Sr/^{86}Sr$ ratios were measured on a Thermo Triton plus thermal ionization mass spectrometer following the methodology described in Yobregat et al. (2017). Approximately 1 µg of Sr was loaded onto single outgassed zone-refined Re filaments with a $TaCl_5$ activator. Intensities of 10×10^{-11} A for ^{88}Sr were maintained for at least 8 hours. Measurements were done in a multidynamic mode with the following Faraday cup configuration: ^{85}Rb (Axial), ^{86}Sr (H1), ^{87}Sr (H2), ^{88}Sr (H3) for line 1, ^{85}Rb (L1), ^{86}Sr (Axial), ^{87}Sr (H1), ^{88}Sr (H2) for line 2 and ^{85}Rb (L2), ^{86}Sr (L1), ^{87}Sr (Axial), ^{88}Sr (H1) for line 3. One run consisted of ~800 cycles with an 8 s integration

time for each line separated by a 5 s idle time. Amplifier rotation was used to improve uncertainties in final isotope ratios.

The NIST SRM 987 standard was measured over the course of this study and yielded a value of 0.7102487 ± 0.0000044 (0.6 ppm, 2 S.E., n = 15). It is common practice to use the 2SD reported for the standard values to determine the external reproducibility equal to 2.4 ppm (Table S1). The $^{87}Sr/^{86}Sr$ multidynamic ratios and $^{87}Rb/^{86}Sr$ ratios for 60025 are listed in Table S2 and the data of Carlson and Lugmair (1988) for the same sample is reported for reference. The external reproducibility of 2.4 ppm (2 S.D.) achieved for the $^{87}Sr/^{86}Sr$ ratio is about an order of magnitude better than the uncertainty of ~30 ppm previously reported for the same sample (Carlson and Lugmair, 1988) and a factor of 5 compared with the uncertainty reported in Borg et al. (2022) for the same sample. For the calculations, we have used the external reproducibility (2SD) obtained on the NIST 987 standard equal to 2.4 ppm. The ICP-MS measurement of Rb and Sr concentrations used in this study is less precise than the conventional isotope dilution method, but the values are identical within error (5 %) to the isotope dilution concentrations given by Carlson and Lugmair (1988).

3. Sr isotope evolution of ferroanorthosite 60025 and the initial $^{87}Sr/^{86}Sr$ ratio of the bulk silicate Moon

As shown in Table S2, we measured with high precision the $^{87}Sr/^{86}Sr$ ratio of ferroanorthosite (FAN) 60025 that has been well characterized in previous studies (Borg et al., 2011; Carlson and Lugmair, 1988) and obtained a mean value of 0.6990812 ± 0.0000009 (2 S.D., n = 3, Table S2) using the analytical technique developed in Yobregat et al. (2017). If one uses the external reproducibility derived by NIST 987 measurements, the relative uncertainty obtained for the plagioclase fraction was thus equal to 2.4 ppm (2 S.D.). In comparison the measured $^{87}Sr/^{86}Sr$ of the plagioclase fractions measured by Carlson and Lugmair (1988) was equal to 0.699096 and had a relative uncertainty of 28 ppm. If corrected for the difference in NIST 987 standard reported in Carlson and Lugmair (1988), this value is equal to 0.699085. The initial $^{87}Sr/^{86}Sr$ ratio of FAN 60025 at the time of its formation taken to be equal to 4.360 Ga (Borg et al., 2011) was calculated to be 0.699063±0.0000019, using error propagation and assuming a 5% uncertainty on the Rb concentration (the ^{87}Rb decay correction increases the uncertainty from 2.4 to 2.8 ppm). Note that the age of Borg et al. (2011) was chosen rather than that of Carlson and Lugmair (1988) as it was

consistent for three chronometers and also consistent with the age of other FANs (e.g. Borg and Carlson 2023). All these methods yielded consistent isochron ages with a preferred value of 4360±3 Ma, which can be used to pin down the $^{87}Sr/^{86}Sr$ of the FAN 60025 at the time this rock formed. It is thus possible to calculate the initial $^{87}Sr/^{86}Sr$ ratios of the mantle reservoir from which this rock was derived at the time of the Rb-Sr system closure (4360 Ma):

$$\left(\frac{^{87}Sr}{^{86}Sr}\right)_{60025}^{T_{FAN}} = \left(\frac{^{87}Sr}{^{86}Sr}\right)_{60025}^{today} - \left(\frac{^{87}Rb}{^{86}Sr}\right)_{FAN} \left(e^{\lambda T_{FAN}} - 1\right) \quad (1)$$

where T_{FAN} is the age of formation of FAN 60025, and λ is the decay constant of ^{87}Rb equal to 1.3971×10^{-11} yr^{-1} (Rothenberg et al. 2012). More complex scenarios that we consider unlikely are discussed in the Supplement.

Table 1. $^{87}Sr/^{86}Sr$ ratios and elemental Rb and Sr concentrations for FAN 60025

		$^{87}Sr/^{86}Sr$	2 s.e.	Rb (ppm)	Sr (ppm)	$^{87}Rb/^{86}Sr$
This study	NIST 987 (± 2 S.D.)	0.7102487	0.0000004			
	60025 plagioclase*	0.6990809	0.0000011	0.020	196.2	2.95×10^{-4}
		0.6990809	0.0000010			
		0.6990817	0.0000013			
	Mean (± 2 S.D.)	0.6990812	0.0000009			
Carlson and Lugmair (1988)	60025 plagioclase	0.699102	0.000020	0.0192	196.5	2.83×10^{-4}
		0.699090	0.000022			
	60025 mafic minerals	0.699151	0.000040	0.0017	4.976	9.89×10^{-4}

Data are normalized to $^{86}Sr/^{88}Sr = 0.1194$. The data of Carlson and Lugmair (1988) was renormalized to the same value of the NIST987 standard reported in Table S1. The mass of plagioclase separates used for the three aliquots was 281.4 mg.

4. A model for the earliest age of the Moon based on Rb-Sr systematics

4. 1 Input parameters

4.1.1 The initial $^{87}Sr/^{86}Sr$ ratio of the Solar System

It has been shown that the nucleosynthetic composition of Sr in planetary materials is not identical throughout the Solar System (Hans et al., 2013; Moynier et al., 2012). Most notably, carbonaceous chondrites and calcium-aluminium rich inclusions (CAI) show significant anomalies expressed as $\mu^{84}Sr$ (Charlier et al., 2021; Kleine et al., 2012; Moynier et al., 2012). These anomalies were initially attributed to either a p-component excess in ^{84}Sr or and alternatively it

could be due to an r-process component excess in ^{86}Sr and ^{88}Sr that would translate to an ^{84}Sr anomaly after mass fractionation correction using the $^{86}Sr/^{88}Sr$ ratios as is commonly done. However, Charlier et al. (2021) have argued that it is unlikely to be an r-process component as this would yield large ^{87}Sr anomalies in the Allende leachates analyzed in that study. Thus, the preferred interpretation of Charlier et al. (2021) is that these anomalies are caused by an excess of p-process nuclide ^{84}Sr. The presence of these anomalies reveals that the early Solar System was not fully homogenized with respect to Sr isotopes, unlike what had been argued in earlier studies (Palme, 2000). This observation for Sr isotopes is fully consistent with other observations made on refractory elements (Cr, Ti, Mo) that show a clear dichotomy between planetary materials from the inner solar system (e.g. Trinquier, 2007; Warren 2011; Kleine et al. 2020) and a second reservoir that could be the outer solar system (i.e. beyond Jupiter). The observed heterogeneity at the scale of a few astronomical units, indicates that the planetary dust was not well mixed within the early Solar System.

Thus, it has become questionable to use $^{87}Sr/^{86}Sr$ measured in CAI as a reliable initial for the inner Solar System. One evidence for this comes from measurements in angrites and eucrites, both classified as non-carbonaceous materials, (e.g., Kleine et al., 2020) that give similar initial $^{87}Sr/^{86}Sr$ ratios (Hans et al. 2013). The initial $^{87}Sr/^{86}Sr$ ratio of angrites and eucrites is difficult to reconcile with the $^{87}Sr/^{86}Sr$ inferred from CAI, since the nucleosynthetic anomalies in ^{84}Sr are solely due to an excess of p-process nuclide. As shown in Hans et al. (2013), if one use the CAI D7 that has the lowest $^{87}Sr/^{86}Sr$ ratio ever found as the initial Sr isotope ratio of the Solar System (Papanastassiou and Wasserburg, 1978) to calculate the age of Rb depletion of angrites, it gives an age of 7.1±1.6 Ma, while it yields 4.6±1.5 Ma with CAI E38 (note that we used the Rb/Sr ratio from Barrat et al. (2012) for calculating the Rb depletion age). This means that Rb loss would postdate the age of accretion of angrites and would correspond to the time when the first crustal angrite formed. Altogether, these ages are not consistent with those inferred with Hf-W systematics (Kleine et al., 2012) and with U-Pb and ^{26}Al-^{26}Mg systematics (Schiller et al. 2015). It is thus more reasonable to consider that the initial $^{87}Sr/^{86}Sr$ inferred for angrites and eucrites better represents the initial ratio of the *inner* solar system, as assumed in recent studies (Mezger et al. 2021 or Borg et al., 2022) without considering that the CAI could represent the appropriate initial for Sr isotopes.

In brief, it is legitimate to use the angrite initial $^{87}Sr/^{86}Sr$ (Hans et al. 2013) as an anchor for the estimation of the age of the Moon, in what follows.

4.1.2 Rb/Sr ratio in the bulk silicate Earth and bulk silicate Moon.

The $^{87}Rb/^{86}Sr$ ratio of the bulk Silicate Earth (BSE) has been determined in previous studies (Hofmann and White, 1983; McDonough et al., 1992). For the sake of this study, we redetermined a new value of the $^{87}Rb/^{86}Sr$ ratio of the bulk silicate Earth using the more recent Ba/Sr ratio determined in the Orgueil CI chondrite (Barrat et al., 2012). This Ba/Sr value for chondrites (0.318) is also consistent with the Ba/Sr ratio found in the comprehensive study of Allende (Stracke et al., 2012; Ba/Sr=0.302) or the study of Barrat et al. (2014) for enstatite chondrites yielding an average Ba/Sr equal to 0.318 (excluding EL6 which are heavily metamorphosed, Stracke et al., 2012). We then assume that the Ba/Sr ratio of the BSE is chondritic as both elements are refractory and lithophile. Since Rb and Ba are both highly incompatible in magmatic processes, it is then possible to use the constant Rb/Ba ratios measured in terrestrial rocks to infer the $^{87}Rb/^{86}Sr$ ratio of the BSE:

$$\left(\frac{Rb}{Sr}\right)_{BSE} = \left(\frac{Rb}{Ba}\right)_{BSE} \left(\frac{Ba}{Sr}\right)_{chondrite}$$

The calculated $^{87}Rb/^{86}Sr$ ratio is equal to 0.079. The $^{87}Rb/^{86}Sr$ value obtained with EH chondrites is equal to 0.0792, which is marginally different. However, this ratio may not be representative of the Rb/Sr of the Earth at the time of the Moon formation since this event predates late accretion (also called late veneer), which is generally assumed to consist of volatile-rich materials. For this reason, one can also correct the $^{87}Rb/^{86}Sr$ ratio of the bulk Earth for the effect of the late veneer. The $^{87}Rb/^{86}Sr$ of the Earth prior to the addition of the late veneer is simply inferred by subtracting a mass fraction of 0.5 % of CI chondrites to the present-day Earth (Day et al., 2007; Walker, 2009). By subtracting the late veneer contribution, the $^{87}Rb/^{86}Sr$ ratio of the BSE shifts from 0.079 to 0.077. Since CI chondrites represent the most Rb-rich material, this calculation provides a *minimum* value for the Rb/Sr ratio for the Earth, as long as the percentage of Earth's mass added by the late veneer does not exceed 0.5%. If the late veneer materials did not consist of CI chondrites and were depleted in volatile elements compared to CI chondrites (as suggested by Ru isotope observations, Fisher-Godde et al. 2017), the Rb/Sr ratio of the Earth before the addition of the late veneer would be comprised between the value calculated above (0.077) and the ratio of the present-

day BSE. In the calculations that follow we have considered that the $^{87}Rb/^{86}Sr$ could range between this minimum value (0.077) and the maximum value taken by Borg et al. (2022) equal to 0.089. For the bulk Silicate Moon (BSM), a similar approach was used. The mean Rb/Ba ratio estimated by McDonough et al. (1992) is 0.017, this yields a value of $^{87}Rb/^{86}Sr$ equal to 0.015 for the bulk silicate Moon. A new estimate of this value was calculated using a compilation of Rb/Ba ratios in thirty-four lunar rocks using only isotope dilution data (Murthy et al., 1971; Philpotts et al., 1972; Philpotts and Schnetzler, 1970; Schnetzler and Nava, 1971) and we obtained a value of 0.0154±0.0028 (2 S.D.). This translates into a new estimate of the $^{87}Rb/^{86}Sr$ for the bulk silicate Moon equal to 0.0135±0.0024 (2 S.D.). Our calculated $^{87}Rb/^{86}Sr$ for the BSM is significantly lower than the $^{87}Rb/^{86}Sr$ suggested in previous publications for FAN 60025 (e.g., Carlson et al. 2014; Carlson and Lugmair, 1988). However, in order to be conservative, we have considered in our calculations a broader range from $2.7x10^{-4}$ (assuming it is equal to the Rb/Sr of FAN) to 0.022.

4.2 Determination of the age of the Moon

A striking observation is the similarity in the isotope composition of refractory elements in the Moon and Earth such as Si, Ca, Mg, Cr, Ti, O (Armytage et al., 2012; Fitoussi and Bourdon, 2012; Mougel et al., 2018; Schiller et al., 2018; Sedaghatpour et al., 2013; Young et al., 2016; Zhang et al., 2012). Understanding these observations has been a challenging task over the years. The quasi-coincidence in isotope signatures between Moon and Earth had initially been used to infer that the Moon and the bulk silicate Earth have equilibrated with each other at the time of the impact (Armytage et al., 2012; Fitoussi and Bourdon, 2012; Young et al., 2016). A mean of alleviating this constraint could be that Theia and the proto-Earth contributed to equal proportions in the Moon and the Earth resulting in homogeneous isotope composition (Canup 2012). As an alternative, Dauphas et al. (2014) have hypothesized that the Earth and Theia came from similar regions in the inner Solar System and had in fact similar O, Cr and Ti isotope compositions. This interpretation would no longer require an equilibration process as described in various studies (Lock et al., 2018; Lock and Stewart, 2017; Pahlevan and Stevenson, 2007). Borg et al. (2022) went further by stating that this hypothesis was also consistent with Rb/Sr ratios lower than that of chondrites, both in Theia and the proto-Earth, indicating a global volatile element depletion that pre-dated the Giant Impact. However, this interpretation conflicts with the Hf-W (Kruijer and Kleine 2017; Fischer et

al. 2021) and Si isotope evidence (Armytage et al., 2012; Fitoussi and Bourdon, 2012). Since Theia and the proto-Earth were unlikely to have had the same evolution in terms of core formation (Hf/W ratio and age of core formation), it is very unlikely that they had similar ^{182}W abundances in their silicate mantles. This was even shown quantitatively by Kruijer and Kleine (2017) who demonstrated that the probability that Moon and Earth initially shared the same $\varepsilon^{182}W$ is less than 5%. Thus, in other words, the similarity in $^{182}W/^{184}W$ ratios for Earth and Moon is not just a coincidence, it must have resulted from a homogeneization process following the giant impact.

A similar argument can be made based on Si, Ca, Mg and Fe stable isotope signatures (e.g., Armytage et al., 2012; Fitoussi and Bourdon, 2012; Fu et al. 2023). The Si isotope data show identical $\delta^{30}Si$ in the Moon and the BSE, while they are distinct from all other planetary materials. Si isotopes can be affected by the processes of planetary accretion (Hin et al., 2017; Kadlag et al. 2021) and core differentiation (Hin et al., 2014; Shahar et al. 2011). There is no particular reason to believe that Theia would have inherited the same $\delta^{30}Si$ as the proto-Earth, even though it might have carried the same nucleosynthetic anomalies. Thus, if the Moon consisted mainly of material derived from Theia as argued in Borg et al. (2022), it would have inherited a silicon isotope signature distinct from that of the Earth. Fu et al. (2023) further emphasized that in the context of the traditional impact models, one does not expect an equilibration between BSE and BSM simply because the Moon stems mostly from the Theia and the energy associated with the impact is insufficient to produce thorough mixing. This is another argument for an isotope equilibration between the Moon and the bulk silicate Earth following the giant impact. As an additional argument, we have calculated the condensation temperatures of Sr, Ca and W in the conditions of the Moon formation. The results (Figure 1) show that Sr was volatile in the early stages of the proto-lunar disk evolution and that its volatility is similar to that of Ca and W. Thus, this is a direct new argument for Sr isotope equilibration, since there is direct evidence for equilibration of Ca and W isotopes between Moon and Earth, this must be the case for Sr as well.

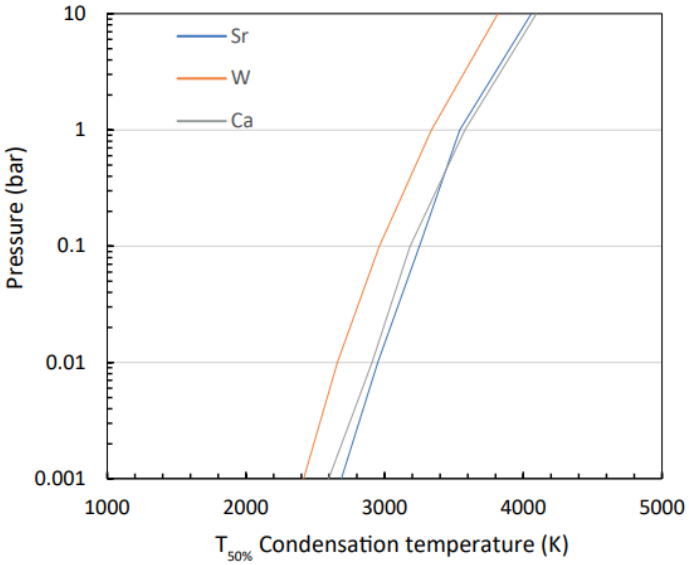

Figure 1

Condensation temperature (50%) for Ca, Sr and W calculated using the approach described in Ivanov et al. (2022) as a function of pressure as described in the Supplementary materials. The results show that Sr is of similar volatility to Ca and W, which are both elements that have equilibrated isotopically between the Moon and bulk silicate Earth, following the Moon forming impact.

In this study, similarly to Halliday (2008) and Mezger et al. (2021), we have considered that the proto-Earth and Theia have equilibrated and therefore considered that the $^{87}Sr/^{86}Sr$ of the Moon and the Earth were identical at the time of the Moon-forming impact due to efficient mixing, which is consistent with a high energy structure, as that described in Lock et al. (2018). Reaching such a high energy structure stems directly from the models of Cuk and Stewart (2012) or Canup (2012), where the final thermal structure of the proto-lunar disk or synestia directly depends on the parameters of the collision between proto-Earth and Theia (e.g. Nakajima et al. 2014).

With this hypothesis, it is possible to determine the $^{87}Sr/^{86}Sr$ ratio of the Bulk Silicate Moon as a function of the time of Moon formation (see details below). By further assuming that the proto-Earth started evolving at T_0 with an $^{87}Rb/^{86}Sr$ equal to that of the BSE+Moon system, it is then possible to calculate the age of the Moon by stating that at that time, the $^{87}Sr/^{86}Sr$ ratio in Moon

and Earth were equal (See Figure 2). As shown in Figure 2, if T_{Moon} is shifted to a younger value, this requires that the initial $^{87}Sr/^{86}Sr$ ratio of the inner Solar System has to be below the observed value. Thus, the calculated age corresponds to the youngest possible age for the Moon. In what follows, this approach is explained in more details.

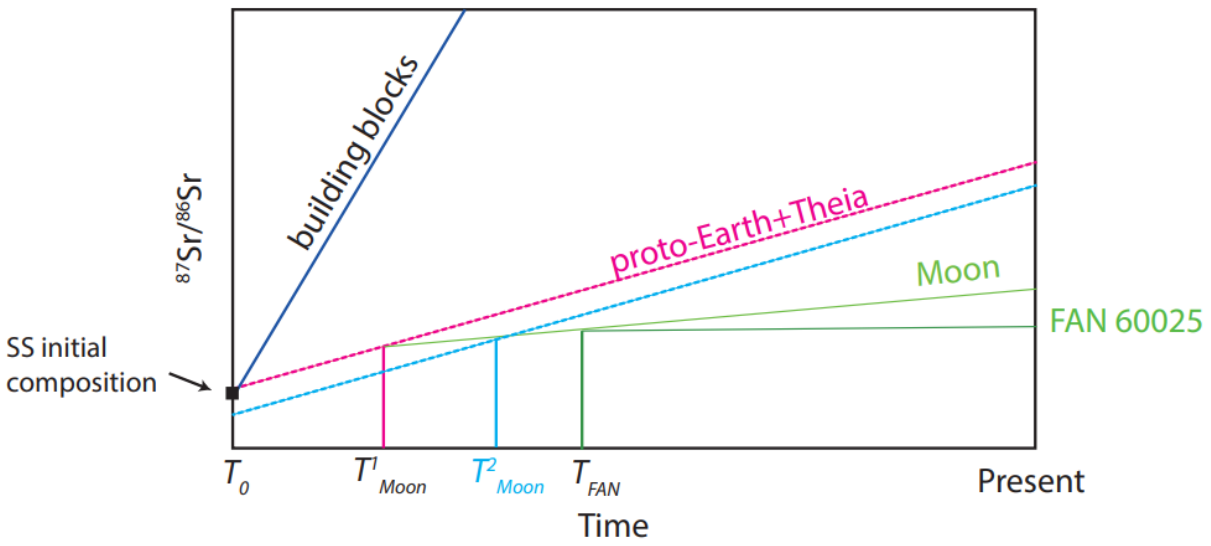

Figure 2. Schematic view of the $^{87}Sr/^{86}Sr$ temporal evolution for the bulk system (Proto-Earth+Theia) and the Moon illustrating the principle used for calculating the youngest age of the Moon. The initial $^{87}Sr/^{86}Sr$ of FAN 60025 is calculated at 4.360 Ga (dark green line). This ratio is then used to calculate the $^{87}Sr/^{86}Sr$ of the Moon at T^1_{Moon} the age of its formation (light green line). Assuming that the bulk system evolved from initial $^{87}Sr/^{86}Sr$ of the Solar System 'SS' at $T_0 = 4.567$ Ga to a common composition at T^1_{Moon}, it is possible to calculate the age that matches this constraint (intersect of red and light green lines). If the age of Moon formation is shifted to a younger age T^2_{Moon}, the initial $^{87}Sr/^{86}Sr$ of the Solar System has to be below the observed value. This defines a youngest age for the Moon.

If one assumes that the initial $^{87}Sr/^{86}Sr$ ratio in the inner Solar System is indeed that defined based on angrites and eucrites (Hans et al. 2013; Mezger et al. 2021; Borg et al. 2022), it is possible to estimate the $^{87}Sr/^{86}Sr$ of the bulk system (proto-Earth+Theia) by assuming that it evolved as a closed system until the Moon formed (Figure 2).

$$\left(\frac{^{87}Sr}{^{86}Sr}\right)^{T_{Moon}}_{E+M} = \left(\frac{^{87}Sr}{^{86}Sr}\right)^{T_0}_{E+M} + \left(\frac{^{87}Rb}{^{86}Sr}\right)_{E+M} \left(e^{\lambda T_0} - e^{\lambda T_{Moon}}\right) \quad (2)$$

where E+M designates the bulk system (Earth+Moon), T_0, the age of the beginning of the Solar System as defined by CAI and T_{Moon} is the age of Moon formation. The bulk system consists of the proto-Earth and Theia, both of which have an unknown composition. However, if we assume mass conservation, then the composition of the bulk system is equal to the composition of the Earth+Moon system. In recent models for the Moon formation, the volatile elements that may have been lost from the Moon are reaccreted onto the Earth (Charnoz and Michaut 2015; Lock et al. 2018; Charnoz et al. 2021). The $^{87}Rb/^{86}Sr$ of the bulk system can thus be calculated as:

$$\frac{^{87}Rb}{^{86}Sr}_{E+M} = \frac{^{87}Rb}{^{86}Sr}_{BSE} X^{Sr}_{BSE} + \frac{^{87}Rb}{^{86}Sr}_{BSM} (1 - X^{Sr}_{BSE})$$

where $X^{Sr}_{BSE}$is the mass fraction of ^{86}Sr from the bulk silicate Earth in the E+M system (=0.982) assuming a lunar core representing 1% of the lunar mass. If we then assume that the Moon and the Earth had identical $^{87}Sr/^{86}Sr$ ratios at the time of the giant impact, it is possible to estimate the age of the Moon-forming impact by calculating the time at which the compositions of the bulk system $^{87}Sr/^{86}Sr$ and of the Moon $^{87}Sr/^{86}Sr$ coincided, as shown in an $^{87}Sr/^{86}Sr$ (t) versus time diagram (Figure 2). The intersection of evolution line of the bulk system and that of the bulk silicate Moon composition defines the age of the Moon (Figure 2). The parameters used for this calculation are given in Table S3. The assumption that the Earth and the Moon had similar Sr isotope compositions at the time the Moon formed is written as:

$$\left(\frac{^{87}Sr}{^{86}Sr}\right)^{T_{Moon}}_{Moon} = \left(\frac{^{87}Sr}{^{86}Sr}\right)^{T_{Moon}}_{E+M} \quad (3)$$

Fortunately, it is also possible to determine independently the $^{87}Sr/^{86}Sr$ ratio of the bulk silicate Moon at the time of Moon formation, T_{Moon} based on lunar rock data. The revised crystallization age for FAN 60025 equal to 4.360 ± 0.003 Ga obtained by Borg et al. (2011) implies that the initial $^{87}Sr/^{86}Sr$ ratio of FAN 60025 cannot be taken as representative of the Moon initial $^{87}Sr/^{86}Sr$ ratio as assumed in Halliday (2008) and Halliday and Porcelli, (2001), unless the age of anorthosite is equal to the age of Moon. This is an important difference with these previous studies. Anorthosites such as FAN 60025 are thought to have formed in the last 70-80% of the lunar magma ocean crystallization by plagioclase flotation. Based on their history, it is possible to consider two possibilities. As a first possibility (Model A, Figure S1), one can assume that the mantle source of

anorthosites is similar to that of the bulk silicate Moon with an $^{87}Rb/^{86}Sr$=0.014-0.019 (Halliday and Porcelli, 2001; Ringwood and Kesson, 1977; Borg et al. 2022) (Model A, see Figure S1). This makes the implicit assumption that the preceding stages of the lunar magma ocean crystallization induced little variation in the Rb/Sr ratio. Given that phases crystallizing prior to anorthosite have been shown to be olivine and orthopyroxene (Elardo et al., 2011; Elkins-Tanton et al., 2011), this is a reasonable assumption because these phases should not greatly fractionate the Rb/Sr ratio. A second possibility (Model B) is to assume that the mantle reservoir producing the FAN was as depleted in Rb as the anorthosite itself (Model B, Rb/Sr_{FAN}=Rb/Sr_{BSM}, see Figure S1) since the time of Moon formation. This second option implicitly assumes that the anorthosite reservoir would have formed rapidly after lunar accretion, a hypothesis that would imply a rapid crystallization of the magma ocean (Elkins-Tanton et al., 2011) and that this reservoir evolved as a closed system until formation of the anorthosite itself. In this case, the late Sm-Nd and U-Pb age of FAN 60025 could be attributed to thermal resetting or remelting of existing anorthosite-rich material.

We first consider that the Sr isotope composition of 60025 was that of the bulk silicate Moon between the time of Moon formation until the time of the anorthosite formation T_{FAN}, (Model A) and using a standard equation for radioactive decay in a closed system, the $^{87}Sr/^{86}Sr$ ratio for the Moon when the Moon formed can be deduced from the following equation:

$$\left(\frac{^{87}Sr}{^{86}Sr}\right)_{60025}^{T_{FAN}} = \left(\frac{^{87}Sr}{^{86}Sr}\right)_{Moon}^{T_{Moon}} + \left(\frac{^{87}Rb}{^{86}Sr}\right)_{BSM}\left(e^{\lambda T_{Moon}} - e^{\lambda T_{FAN}}\right) \quad (4)$$

where T_{FAN} is the age of formation of the ferro-anorthosite. Alternatively, in the case of Model B, we can also assume that the $^{87}Rb/^{86}Sr$ of this sample is equal to that of FAN 60025 rather than that of the BSM from T_{Moon} to T_{FAN}. The Sr isotope composition of FAN 60025 at the time of its formation T_{FAN} can be deduced from the measured value (Table S2) using equation (4) shown above. Combining equations (1)-(4), it is then possible to calculate the age of the Moon:

$$T_{Moon} = \frac{1}{\lambda}\ln\left(\frac{\left(\frac{^{87}Sr}{^{86}Sr}\right)_{60025}^{today} - \left(\frac{^{87}Rb}{^{86}Sr}\right)_{FAN}\left(e^{\lambda T_{FAN}}-1\right) - \left(\frac{^{87}Sr}{^{86}Sr}\right)_{E+M}^{T_0} + \left(\frac{^{87}Rb}{^{86}Sr}\right)_{BSM}e^{\lambda T_{FAN}} - \left(\frac{^{87}Rb}{^{86}Sr}\right)_{E+M}e^{\lambda T_0}}{\left(\frac{^{87}Rb}{^{86}Sr}\right)_{BSM} - \left(\frac{^{87}Rb}{^{86}Sr}\right)_{E+M}}\right) \quad (5)$$

The equation for Model B is identical, except that we replace $\left(\frac{^{87}Rb}{^{86}Sr}\right)_{BSM}$ by $\left(\frac{^{87}Rb}{^{86}Sr}\right)_{FAN}$.

The age of the Moon was calculated using equation (5) and the results are plotted in Figures 3 and 4 for various values of the input parameters. In these calculations, the most sensitive parameter is the $^{87}Rb/^{86}Sr$ of the bulk Silicate Moon, which is used as the x-axis for plotting the results.

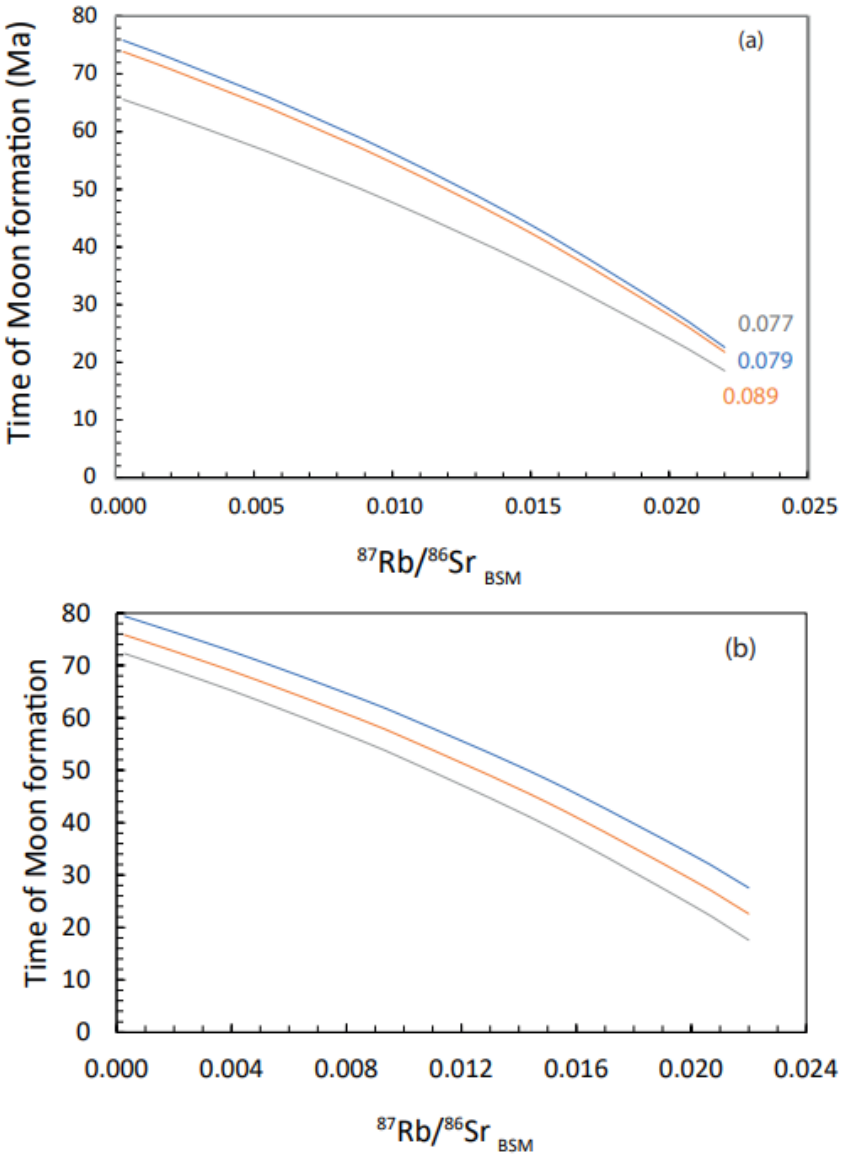

Figure 3 (a) Forward model for the calculation of the age of the Moon based on Rb-Sr isotope systematics showing the sensitivity of the age determination to $^{87}Rb/^{86}Sr$ ratio of the bulk Silicate Moon (BSM) and to the uncertainty to the $^{87}Rb/^{86}Sr$ ratio of the Bulk Silicate Earth with the values labeled on the curves. (b) Forward model for the calculation of the age of the Moon based on Rb-

Sr isotope systematics showing the sensitivity of the age determination to $^{87}Rb/^{86}Sr$ ratio of the bulk Silicate Moon (BSM) and to the uncertainty on the initial $^{87}Sr/^{86}Sr$ ratio of the inner Solar System (Hans et al. 2013). The blue curve corresponds to the mean value μ=0.698978, while the yellow and grey curves corresponds to $\mu+2\sigma$ and $\mu-2\sigma$, respectively with 2σ=0.000004.

If we assume that Model A is correct, the age of the Moon is equal to 45 Ma for an $^{87}Rb/^{86}Sr$ ratio in the BSE equal to 0.079. If we now assume that Model B is correct, then the age of the Moon is 73.9 ± 0.8 Ma for an $^{87}Rb/^{86}Sr$ ratio in the BSE equal to 0.079 (the quoted uncertainty takes only into account the uncertainty on the measurements of FAN 60025, corresponding to fixed $^{87}Rb/^{86}Sr$ values in the BSM and BSE. For the sake of comparison, if one uses the analytical uncertainties on the measured $^{87}Sr/^{86}Sr$ ratio for FAN 60025, reported by Carlson and Lugmair (1988), with the same values of $^{87}Rb/^{86}Sr$ ratios in the Moon and Earth, then the maximum age of the Moon would range between 24.7 and 68.2 Ma with Model A and between 58.0 and 93.7 Ma with Model B (Figure 4). Thus, the improvement in analytical uncertainty of FAN 60025 is critical for improving the age uncertainty (see also below for a detailed assessment).

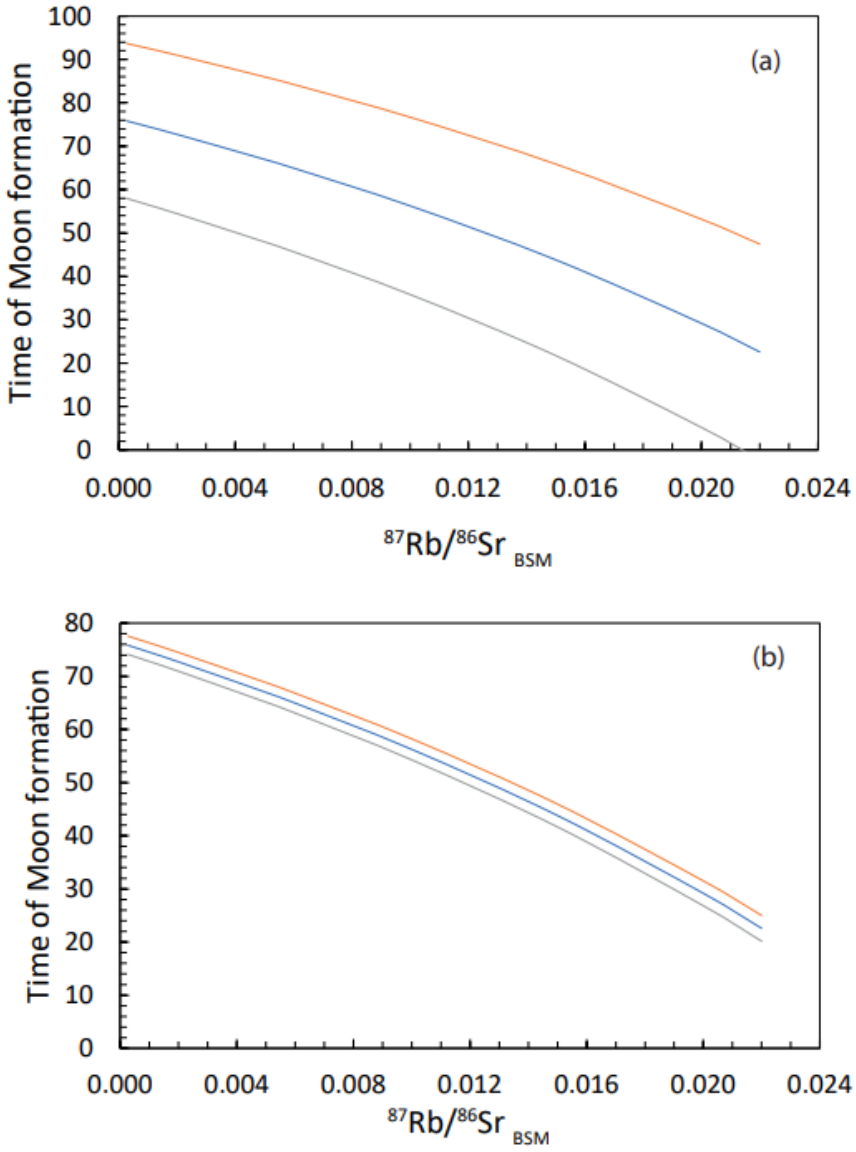

Figure 4 Forward model for the calculation of the age of the Moon based on Rb-Sr isotope systematics showing the sensitivity of the age determination to the initial $^{87}Sr/^{86}Sr$ ratio of the ferro-anorthosite 60025 (a) uncertainties are those reported in Carlson and Lugmair (1988) which is approximately 28 ppm. (b) uncertainties are those reported in Table S2, this study. There is a clear reduction of uncertainty on the age determination based on the new data.

A further refinement is that one can consider the existence of late accretion that could have modified the Rb/Sr ratio of the BSE and this late accretion was not identical for Moon and Earth

(Kruijer et al., 2015; Touboul et al., 2015). If we substract the contribution of late accreted material to the BSE assuming that this contribution consisted of chondrites, then the estimated age of the Moon becomes 46.4 Ma instead of 45.0 Ma (Model A). This indicates that if we combine the results obtained from Model A and B, the youngest possible age of the Moon is 79.4 Ma after the beginning of the Solar System. In Figures 3 and 4, we have investigated the sensitivity of the Moon age on possible input parameters, namely $^{87}Rb/^{86}Sr$ of BSE, $^{87}Rb/^{86}Sr$ of BSM, initial $^{87}Sr/^{86}Sr$ of the Solar System, initial $^{87}Sr/^{86}Sr$ of FAN 60025. One of the most sensitive parameters is the $^{87}Rb/^{86}Sr_{BSM;}$ as shown in Figure 3-4, the youngest age (79.4 Ma) is obtained for a low $^{87}Rb/^{86}Sr$ in the BSM while an older age (17.6 Ma) corresponds to the highest $^{87}Rb/^{86}Sr_{BSM}$ value (0.022). As mentioned above, the calculated age is also very sensitive to the uncertainty on the initial $^{87}Sr/^{86}Sr$ of FAN 60025. If one takes the uncertainty of 28 ppm on $^{87}Sr/^{86}Sr$ reported in Carlson and Lugmair (1988) then the youngest age becomes 94 Ma with the lowest bound being 0 Ma (Figure 4a). In contrast, if one uses the newly determined $^{87}Sr/^{86}Sr$ with an uncertainty of 2.4 ppm, then this range becomes 75.8-79.4 Ma (Figure 4b). The least sensitive parameter is $^{87}Rb/^{86}Sr_{BSE}$, given a reasonable range of input values from 0.077 to 0.089 as shown in Figure 3. Similarly, the uncertainty on the initial $^{87}Sr/^{86}Sr$ of inner Solar System also has a limited impact on the calculated Moon age (Figure 3b).

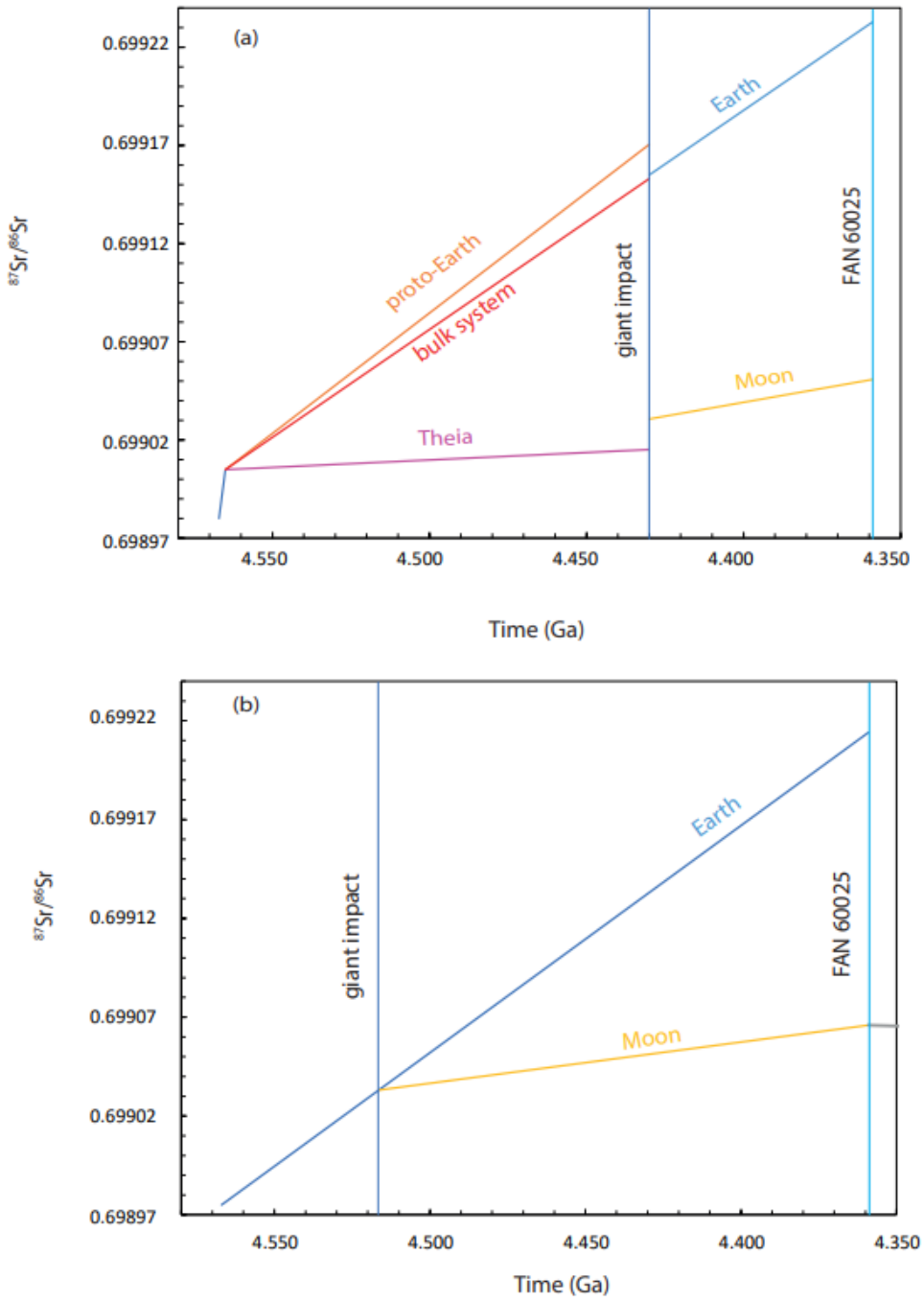

Figure 5

$^{87}Sr/^{86}Sr$ versus time diagram to illustrate the differences between (a) the model of Borg et al. (2022) and (b) the model presented in this study. The model of Borg et al. (2022) shown here was redrawn based on the parameters given in their Table S1 (Model 6). During the first stage, the system evolves as a closed system with $^{87}Rb/^{86}Sr = 0.832$ (CI chondrites). After 2 Ma, proto-Earth

and Theia form and evolve as closed systems until the time of the impact. The time evolution of the bulk system (Theia + proto-Earth) was recalculated and is shown as a red line. Note that it is very similar to the evolution of the BSE after the impact. This model clearly shows that the composition of the bulk silicate Earth is more radiogenic than that of the Moon, which consists mainly of Theia. The time of the impact is fixed arbitrarily to match the composition of the Moon using plausible mixing proportions. (b) in this study, the $^{87}Sr/^{86}Sr$ of the bulk system (recalculated based on the Moon + BSE mass balance) evolves as a closed system, until the time of the impact when Earth and Moon equilibrated isotopically. The time of the impact is fixed such that the $^{87}Sr/^{86}Sr$ ratio of the Moon is equal to that of the bulk system.

Our results clear differ from those of Borg et al. (2022) and for the sake of clarity, we have reproduced their calculations using their model 6 (Table S1 in Borg et al. 2022). The results are plotted in Figure 5a and directly compared with one of our models (Figure 5b) using the same scale in X and Y. The obvious difference is that they do not assume that Earth and Moon had similar $^{87}Sr/^{86}Sr$ due to isotope equilibration after the giant impact, hence the $^{87}Sr/^{86}Sr$ of the post impact Earth is significantly more radiogenic than that of the Moon. We have calculated in addition the $^{87}Rb/^{86}Sr$ ratio of the bulk system and it is in fact very close to that of the bulk silicate Earth after the impact. This diagram also illustrates the closed system evolution of Theia and proto-Earth in the stage preceding the impact with a Rb loss taking place 2 Ma after the beginning of the Solar System. In what follows, we propose a model where the Rb loss is not necessarily linked with a unique event of Rb loss, which is strongly restrictive.

5. MonteCarlo simulations of the age of Moon formation.

The model presented in the previous sections illustrates conceptually the method for calculating the age of the Moon but it is not fully satisfactory from a conceptual viewpoint. In the previous scenario (similarly to all published models), we have made some simple assumptions about the $^{87}Rb/^{86}Sr$ evolution of the system by stating that proto-Earth+Theia system had the $^{87}Rb/^{86}Sr$ ratio of the BSE (+Moon) since the beginning of the Solar System and that it remained a closed system since then. However, this scenario is rather restrictive. By essence, the accretion process does not fulfil the conditions of a closed system and it is known that planetary bodies of the inner Solar System do not share a common $^{87}Rb/^{86}Sr$ value, with values ranging between that

of angrites (5×10^{-3}) up to the maximum $^{87}Rb/^{86}Sr$ found in CI chondrites (0.8-0.9). Thus, it is more realistic to consider a model where the $^{87}Rb/^{86}Sr$ of the accreting Earth constantly evolves due to contribution of bodies with variable Rb-Sr history.

In what follows, we describe a MonteCarlo approach that includes all sources of uncertainties and considers a more realistic Rb-Sr evolution prior to the Moon formation. The evolution starting from the beginning of the Solar System is described by N steps of equal duration (Δt) until the time of Moon formation (T_{Moon}). At each step, a fraction of Sr, f_{Sr}^{i} is added to the growing Earth whose initial mass of Sr is assumed to be M_0. The new $^{87}Sr/^{86}Sr$ at step i is thus:

$$\frac{^{87}Sr}{^{86}Sr}_{init} = f_{Sr}^{i}\frac{^{87}Sr}{^{86}Sr}_{i-1} + \left(1 - f_{Sr}^{i}\right)\frac{^{87}Sr}{^{86}Sr}_{in}$$

A similar equation can be written for $^{87}Rb/^{86}Sr$. At each time step, an object with a random mass was accreted and the ^{87}Rb-^{87}Sr characteristics of this object were determined according to the following rules: the $^{87}Sr/^{86}Sr$ ratio of accreted objects was assumed to evolve starting from the initial of the Solar System defined by angrites, until it is accreted with an $^{87}Rb/^{86}Sr$ ratio that is allowed to take a random value between 0 (≈angrites) and 0.832 (CI chondrites). In addition, the accretionary process could randomly lower this ratio upon accretion, mimicking the effect of volatile element depletion after a giant impact or any other thermal process. Although this process is not well constrained, the volatile loss could be due to the impact itself or to the generation of a magma ocean that would lose its volatile (e.g. Hin et al., 2017). A decay equation is then used to calculate the in-grown ^{87}Sr:

$$\frac{^{87}Sr}{^{86}Sr}_{i} = f_{Sr}^{i}\frac{^{87}Sr}{^{86}Sr}_{i-1} + \left(1 - f_{Sr}^{i}\right)\frac{^{87}Sr}{^{86}Sr}_{in} + k_{Rb}\frac{^{87}Rb}{^{86}Sr}\left(e^{\lambda T_{i-1}} - e^{\lambda T_i}\right)$$

The proto-Earth, is then assumed to grow until it reaches a mass fraction of Sr equal to 1 corresponding to a full accretion. The parameter f_{Sr}^{i} represents the mass fraction of Sr added at each time step and k_{Rb} is a coefficient ranging between 0 and 1 representing a random Rb loss linked with accretion. Since Sr is a refractory element, we assume that its concentration is constant in the accreted material. In this case, the differential equation describing the evolution of $^{87}Sr/^{86}Sr$ in the silicate mantle can be written as:

$$\frac{d\,^{87}Sr/^{86}Sr}{dt} = \frac{1}{M}\frac{dM}{dt}\left[\frac{C_{me}^{87}}{C_{me}^{86}}\frac{C_{me}^{86}}{C_{m}^{86}} - \frac{C_{m}^{87}}{C_{m}^{86}}\right] - \frac{C_{m}^{87}}{C_{m}^{86}}\left[\frac{C_{me}^{86}}{C_{m}^{86}} - 1\right]$$

where the subscripts m and me represent the mantle of Earth and the mantle of accreted material, respectively and C_k^i denotes the concentration of isotope i in reservoir k (i.e. me or m). Since $C_{me}^{86} = C_m^{86}$, the above equation can be simplified to:

$$\frac{d\,{}^{87}Sr/{}^{86}Sr}{dt} = \frac{1}{M}\frac{dM}{dt}\left[\frac{C_{me}^{87}}{C_{me}^{86}} - \frac{C_m^{87}}{C_m^{86}}\right]$$

If this equation is discretized over a time step Δt; it becomes:

$$\frac{{}^{87}Sr}{{}^{86}Sr}_i - \frac{{}^{87}Sr}{{}^{86}Sr}_{i-1} = \frac{1}{M}\frac{dM}{dt}\Delta t\left[\frac{{}^{87}Sr}{{}^{86}Sr}_{me} - \frac{{}^{87}Sr}{{}^{86}Sr}_{i-1}\right]$$

In this expression, one can recognize that:

$$f_{Sr} = \frac{1}{M}\frac{dM}{dt}\Delta t$$

If this equation is discretized, the accreted mass at time iΔt can be calculated as a function of f_{Sr}^i defined at step i:

$$f_{Sr}^i = \frac{1}{M_i}(M_i - M_{i-1}) = \frac{\Delta M_i}{M_i}$$

In other words, f_{Sr}^i represent a relative incremental mass accreted at time step i. The accretion of the proto-Earth was assumed to take place over N time steps (N=120) until the Earth reached its present mass. If we assume that the size of M_0 can vary between a 1 km to 1000 km object, then a large range of growth curves can be constructed (Figure 6). In order to generate growth curves that are similar to the growth curves of planetary embryos (e.g., Thommes et al., 2003), we used the following function for the accreted mass:

$$\Delta M_i = rand^i$$

where rand is a random number between 0 and 1 and i represents the time step. For each trial, a new random value of $f_{Sr}^i$is calculated at each time step.

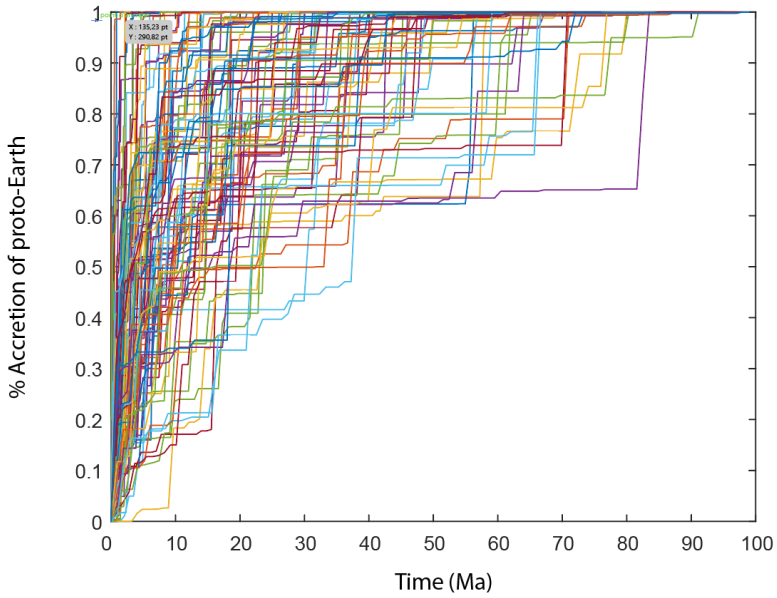

Figure 6 Sample of random growth curves for the Earth generated in the MonteCarlo simulations mimicking typical accretion curves produced in numerical simulations (e.g., Thommes et al. 2003) A value of 1 indicates 100% accretion.

Table S2 shows the range of a priori input parameters used in the MonteCarlo simulation. Figure 7 shows examples of $^{87}Rb/^{86}Sr$ evolution in the proto-Earth as a function of time for various simulations.

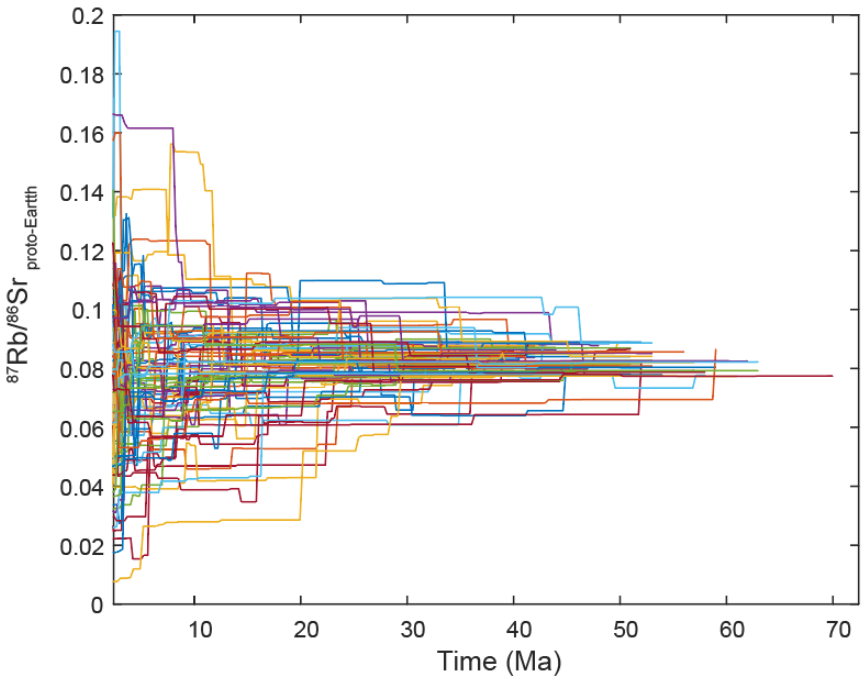

Figure 7 $^{87}Rb/^{86}Sr$ in the accreting proto-Earth as a function of time produced in MonteCarlo simulation for solutions that match the criteria described in the main text. Only the final values of $^{87}Rb/^{86}Sr$ in the bulk silicate Earth range between 0.077 and 0.089.

This model does not assume that the average $^{87}Rb/^{86}Sr$ of the growing Earth is equal to the $^{87}Rb/^{86}Sr$ of the BSE. Rather, the $^{87}Rb/^{86}Sr$ can evolve partly independently of the $^{87}Sr/^{86}Sr$ ratio in the proto-Earth due to the lack of a closed-system assumption. The only constraint being that at the time of the Moon formation, Sr isotopes in the Moon and bulk Silicate Earth equilibrate and the $^{87}Rb/^{86}Sr$ ratio of the final Earth is equal to that of the BSE, as shown in Figure 7 that represents the $^{87}Rb/^{86}Sr$ ratio of the accreting Earth as a function of time. In Figure 8, we also show the mean $^{87}Rb/^{86}Sr$ ratio of the final Earth obtained for a series of successful models.

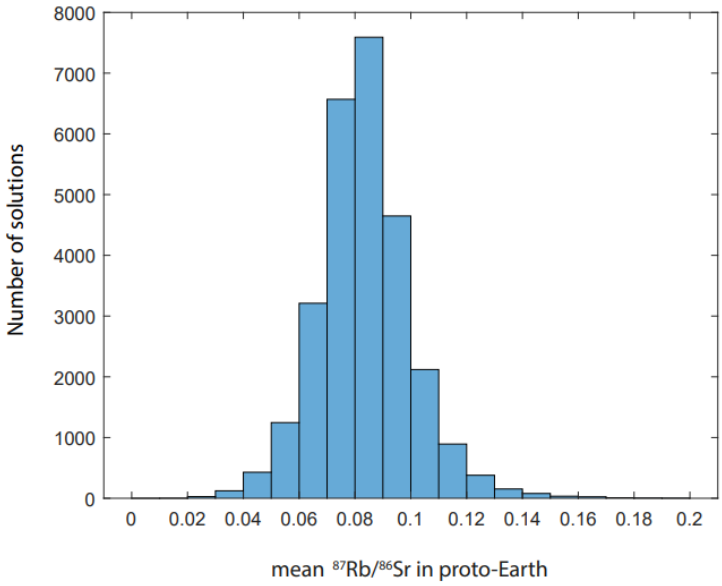

Figure 8 Histogram of mean $^{87}Rb/^{86}Sr$ ratios (over time) for the Earth that are solutions of the MonteCarlo model (section 4.). These ratios range between 0.0137 and 0.260 are both below and above the value $^{87}Rb/^{86}Sr_{BSE}$ (0.077-0.089). This illustrates the open-system behavior of the accretion model.

This model clearly illustrates that our open system model no longer has the constraint that the mean $^{87}Rb/^{86}Sr$ is equal to $^{87}Rb/^{86}Sr_{BSE}$. Thus, this model simulates more reliably the Earth accretion than previously existing model that make assumptions about the building blocks (Mezger et al. 2021), or use two component models where the two components evolve as closed system (no Rb loss is allowed during accretion; Borg et al. 2022) or assume an $^{87}Rb/^{86}Sr$ ratio equal to that of the BSE (Halliday 2008).

As there are significant uncertainties in the $^{87}Rb/^{86}Sr$ ratios of the bulk silicate Moon and bulk silicate Earth, we assumed a range for these parameters (see discussion above). An evolutionary path was considered a successful solution if the $^{87}Sr/^{86}Sr$ ratios of the proto-Earth at the time of Moon formation was equal to that of the Moon within error. The most significant source of uncertainty in this calculation is the assumed value of the Rb/Sr ratio of the BSM. To investigate the influence of this ratio, the age determination was directly calculated for various $^{87}Rb/^{86}Sr$ values for the Bulk Silicate Moon ranging between that of FAN 60025 (similar to the scenario of

Model B) and 0.022 (Model A). This model was also run for the whole range of $^{87}Rb/^{86}Sr$ in the bulk silicate Earth (0.077-0.089).

The results of the MonteCarlo calculation are plotted in Figure 9 and show that the age of Moon formation ranges between 18 and 79 Ma after CAI for the large range in Rb/Sr assumed for the BSM and BSE. For $^{87}Rb/^{86}Sr_{BSM}$ equal to 0.0135±0.0024, as estimated in the Supplementary materials based on isotope dilution data, the minimum age of lunar formation is 50.4±6.0 Ma.

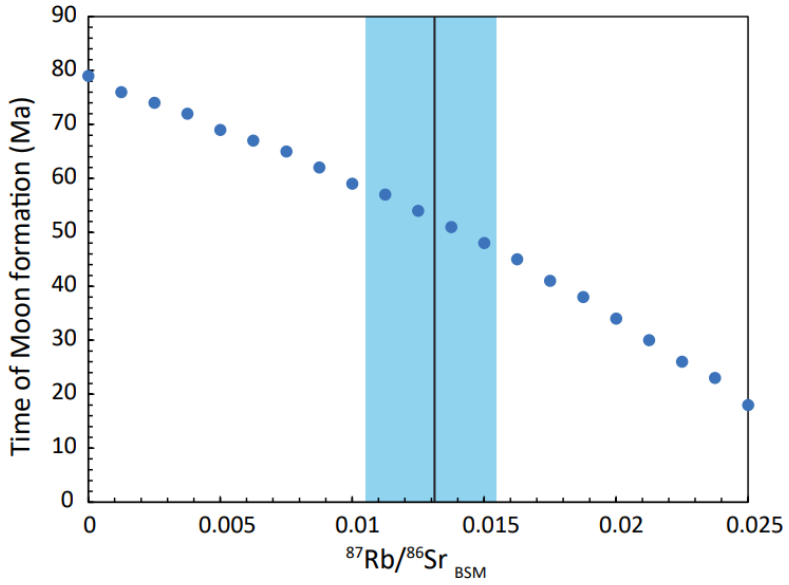

Figure 9 Determination of the maximum age of Moon formation as a function of the $^{87}Rb/^{86}Sr$ ratio of the bulk Silicate Moon using a MonteCarlo simulation for the $^{87}Sr/^{86}Sr$ ratio of the proto-Earth as a function of time (see description in section 5.). This simulation takes into account the joint variability of the parameters shown in Figures 3-4 and the possibly complex open-system Rb-Sr evolution of the proto-Earth until the Moon forming impact. The blue band marks the preferred $^{87}Rb/^{86}Sr_{BSM}$ value calculated in this study (0.0135±24).

The results of MonteCarlo simulations in Figure 9 shows that considering all sources of uncertainties (including $^{87}Rb/^{86}Sr_{BSM}$, $^{87}Rb/^{86}Sr_{BSE}$, $^{87}Sr/^{86}Sr_{FAN}$ and $^{87}Sr/^{86}Sr_{SS}$), the youngest age of the Moon should range between 18 and 79 Ma after CAI, depending on the value of $^{87}Rb/^{86}Sr_{BSM}$. To illustrate how the improvement in uncertainties affects the calculated results, A 3-D diagram (Figures S2 and S3; Supplementary materials) showing the solutions obtained with our new data for FAN 60025 can be compared with the solutions obtained with the uncertainties

reported in Carlson and Lugmair (1988). The range of solutions is clearly broader with the uncertainties from Carlson and Lugmair (1988).

This approach provides a robust bound on the youngest age of the Moon, independently of the exact Sr isotope evolution of the proto-Earth, which is currently unknown. Note that our simulation shows that the Rb/Sr ratio of the BSE has rather small leverage on the calculated age (see Figure 3) as also demonstrated in section 4.

Incidentally, the 60025 FAN anorthosite used for this study with a chronometric age of 4.360 Ga (or 200 Ma after CAI) was used for determining a new Rb-Sr age for the Moon. If there had been a resetting the Sm-Nd or U-Pb ages, this had probably little effect on the $^{87}Sr/^{86}Sr$ value of anorthite of FAN 60025, since this mineral phase contains more than 99.95% of Sr in the whole rock, as this mineral represents 98% of the rock with 40 times more Sr than in the mafic minerals (Table S2). Thus, our approach that considers the FAN evolution starting from the time of the giant impact until 4.36 Ga (Model B) does in fact consider implicitly the effect of resetting and our estimate of $^{87}Sr/^{86}Sr$ at the time of Moon formation depends only on the assumed $^{87}Rb/^{86}Sr$ ratios.

6. Implications

Our high precision Sr data on lunar rocks implies that the Moon has to have formed earlier than 79 Ma after the beginning of the Solar System. If we use the preferred value for the $^{87}Rb/^{86}Sr$ (=0.0135) of the Bulk Silicate Moon, this value falls to 51 Ma. This age is remarkably close to the most recent estimate for the earliest age of Moon formation given by Hf-W constraints (Kruijer and Kleine 2017), assuming again that Earth and Moon were isotopically equilibrated (as evidence by their similar ^{182}W abundances, after accounting for differences in the late accreted material). Using the most recent data for Hf/W ratios in the BSM and BSE, this maximum age is 48 Ma to obtain a difference between BSM and BSE that does not exceed 15 ppm (Kruijer and Kleine 2017). These conclusions are clearly at odds with those of Borg et al. (2022) based on the Rb-Sr system. The difference in our conclusions can be understood as Borg et al. (2022) have not made the hypothesis that $^{87}Sr/^{86}Sr_{Moon}=^{87}Sr/^{86}Sr_{BSE}$ at the time of Moon formation. With this hypothesis, it is possible for the bulk silicate Earth to evolve to a higher $^{87}Sr/^{86}Sr$ than that of the Moon (see Figure 5a), and this allows the possibility of delaying the time of Moon formation. Owing to relaxing the hypothesis of isotope equilibration between Earth and Moon, it is then possible to obtain younger ages for the Moon. For example, in the Models 6 of (Borg et al., 2022) (see their

Table S1 for details) the $^{87}Sr/^{86}Sr_{BSE}$ at the time of Moon formation (=0.699168) can be higher than that of the BSM (=0.699005), which means that the $^{87}Sr/^{86}Sr$ can evolve longer than in the case of our model. This salient feature of the Borg et al. (2022) model is illustrated in Figure 5a to better explicit the implications of their hypothesis regarding the absence of Moon-Earth equilibration. However, as mentioned above this hypothesis is not consistent with the Hf-W and stable isotope observations (e.g. Fu et al. 2023) and was shown to be very unlikely (Kruijer et al. 2021). This question was also investigated by Fischer et al. (2021) who also concluded that Earth and Moon should have equilibrated isotopically after the giant impact.

An advantage of our model that differs from that of Borg et al. (2022) is that we consider a realistic Rb-Sr evolution of the proto-Earth, as it is being accreted. In contrast, Borg et al. (2022) have assumed a single-stage 'closed system' from 1-2 Ma until the time of Moon formation for both Theia and the proto-Earth. To be more specific, the two components in their model evolve separately as closed system (after an initial loss of Rb at the beginning of the Solar System). This means that the accretion process implies no further loss of Rb. This is a strong hypothesis that does not fit with the idea that the accretion process of the Earth is by essence an open system where numerous accreting objects of variable Rb/Sr ratios are being added. Such a complex process is not well described by a single stage evolution.

Interestingly, the results of our model are similar to those reported in Mezger et al. (2021) who reported a formation age for the Moon of 60±15 Myr. A notable difference is that our age for Moon formation represents a minimum age for the Moon, i.e. the Moon did not form later than 79 Ma. It should be mentioned also that the study of Mezger et al. (2021) made strong assumptions about the proportions and composition of components, one component being volatile-rich (15%) while the second one was volatile-depleted (85%). However, by considering the trace element data used to defined these components, one notes that the proportions are not better constrained than 5% (as opposed to the quoted uncertainty of 1%). If one takes these uncertainties into account the range of ages becomes 60±30 Myr. Furthermore, if one uses the mixing proportions (85-15%), to calculate the $^{87}Rb/^{86}Sr$ of the BSE, one obtains a value of 0.16, which is well above the observed value (≈0.08), which makes the reliability of the results problematic. Another issue is that the history of the FAN 60025 used to derive an age was considered as a fixed value of the initial $^{87}Sr/^{86}Sr$ was considered. As explained in our study, one must also account for the poorly constrained value of the $^{87}Rb/^{86}Sr$ ratio of the bulk Silicate Moon. If this constraint had been

included, the Moon age derived by Mezger et al. (2021) would have been a maximum as in this study.

Our new results are also consistent with the conclusions of Jacobson et al. (2014) who inferred that the Moon must have formed before at 95 ± 32 Ma after the beginning of the Solar System.

The new Rb-Sr age constraint we propose also puts tighter constraints on the youngest age of Moon formation and these results may appear inconsistent with a very young age for the Moon based on recent work on lunar anorthosites or with the model age for Moon accretion given in Maurice et al. (2020). For example, Borg et al. (2011), measured internal isochrons for the FAN 60025 using the Pb-Pb and Sm-Nd isotope systems and obtained a mean age for this rock of 4.360 ± 0.003 Ga (or 208 Ma after CAI). As mentioned above, these authors further argued that if the lifetime of the magma ocean was short (as calculated in Elkins-Tanton et al., 2011), then the age of anorthosites would approximately represent the age of Moon formation. This conclusion was reemphasized based on more age determinations of anorthosites (Boyet et al., 2015). However, this interpretation has been questioned because it must be ascertained first that the ages of FAN really record magma ocean crystallization and there is the possibility of resetting the anorthosite age by a later event (Asphaug, 2014; McLeod et al., 2014). There is also petrological evidence for a complex history for the anorthosite lunar crust, indicating multiple events associated with anorthosite formation (Gross et al., 2014).

By considering a larger data set, Borg and Carlson (2023) have argued that the concordance of ages recorded for FAN, urKREEP, Mg-suite rocks as well as the Sm-Nd model age of mare basalts (Gaffney and Borg 2014) fall within a relatively narrow range between 4.33-4.36 Ga. Given that the cooling of the magma ocean must be relatively fast, this range of ages for samples from far distant sites (Borg et al. 2019) cannot be coincidental and must be closely related to the age of the formation of the Moon. However, given the uncertainties in the ages, the possible range of ages is actually 4.29-4.38 Ga, which span approximately 100 Ma. This first suggest that the duration of magmatism presumably associated with the formation of a magma ocean is considerably longer than assumed by the proponents of a young age for the Moon. The cooling of the lunar magma is probably extremely fast in its first stages because it is highly convective, so it is unlikely to have started 100 Ma after the accretion of the Moon. Nevertheless, the formation of a plagioclase lid, which is unavoidable, will take place after 70% crystallization (Schmidt and Kraettli 2022) and

this will considerably slow down the cooling of the lunar mantle. Thermal models that includes this process show that the formation of the KREEP and FAN reservoirs will take place late (Maurice et al. 2020). As a result, it is expected that the age of ferro-anorthosites (that start forming after 70% crystallization) will yield a range of ages reflecting the accumulation of anorthite at the base of the crust. As shown in Maurice et al. (2020) the duration of the last stages of crystallization could last up to ≈200 Ma, due in part to the low thermal conductivity of plagioclase. The recent ages of KREEP, FAN and the Mg-suite could just reflect this late stage of crystallization. Furthermore, although the 60025 rock was found at the lunar surface, its breccia texture indicates that it was excavated from depth by a crater forming impact. The depth of excavation could be tens of km (e.g. Spudis et al. 1989) which is consistent with its plutonic texture. This means that anorthosites such as 60025 could have recorded delayed cooling at depth in the lunar crust.

Alternatively, it has been shown that the survival of a slushy magma ocean (Michaut and Neufeld 2022) could lead to a delay in anorthosite crust formation. This model considers first the formation of a stagnant lid, while the sedimentation of crystals does not take place due to the low gravity of the Moon and an on-going convection of the slushy and viscous magma ocean. Within the framework of this scenario, the cooling of the LMO is slowed down and the formation of an anorthositic crust is delayed by several hundreds of Ma. In brief, both thermal models have the potential to reconcile the young age of FAN (i.e. ≈200 Ma after CAI; Borg et al. 2011; Boyet et al. 2015) with an older age for the Moon, in agreement with our inference based on Rb-Sr.

Our Rb-Sr age for the Moon can also be compared with estimates of the Moon accretion based on U-Pb systematics (Connelly and Bizzarro, 2016; Halliday, 2004). For example, Connelly and Bizzarro (2016) interpreted the U-Pb fractionation observed between the Earth and its precursor materials (i.e., chondrites) as reflecting a large-scale event of Pb volatilization during the Moon-forming impact, thereby dating the age of the Moon with ages ranging between 4.426 and 4.417 Ga, or 141-151 Ma after the beginning of the Solar System, which would seem to contradict our interpretation of the Rb-Sr record. However, as shown in Rudge et al. (2010), depending on the conditions of metal-silicate equilibration, on the Pb partitioning in the core (Wood and Halliday, 2010), or on the choice of U/Pb ratio for representing the bulk silicate Earth (Halliday, 2004), the constraints provided by U-Pb do not necessarily disagree with our Rb-Sr age. The U-Pb age constraints were recently reevaluated by Maltese and Mezger (2020). Similar to

Ballhaus et al. (2013), these authors showed that the Pb isotope systematics were mostly controlled by late accretion of volatile rich material that evolved with a high U/Pb ratio due to Pb incorporation into the Earth core as sulphide (Wood and Halliday, 2010) and this yielded a preferred age of 69±10 Ma, which seems in agreement with our Rb-Sr constraints. Overall, the interpretation of U-Pb for the Earth-Moon system are possibly consistent with the Rb/Sr constraints and is obviously a matter of debate.

In summary, our more precise Rb-Sr data for lunar anorthosite 60025 combined with a model that explores widely various sources of uncertainties and a more realistic accretion model provides new constraints on the age of the Moon. The main source of uncertainty is now the $^{87}Rb/^{86}Sr$ of the bulk Silicate Moon but based on existing data (compiled in this study), the youngest age of the Moon would be ~51 Ma after the beginning of the Solar System.

References

Amelin, Y., Kaltenbach, A., Iizuka, T., Stirling, C.H., Ireland, T.R., Petaev, M., Jacobsen, S.B., 2010. U–Pb chronology of the Solar System's oldest solids with variable $^{238}U/^{235}U$. Earth Planet Sci Lett 300, 343–350. https://doi.org/10.1016/J.EPSL.2010.10.015

Armytage, R.M.G., Georg, R.B., Williams, H.M., Halliday, A.N., 2012. Silicon isotopes in lunar rocks: Implications for the Moon's formation and the early history of the Earth. Geochim Cosmochim Acta 77, 504–514. https://doi.org/10.1016/j.gca.2011.10.032

Asphaug, E., 2014. Impact Origin of the Moon? Annu Rev Earth Planet Sci 42, 551–578. https://doi.org/10.1146/annurev-earth-050212-124057

Ballhaus, C., Laurenz, V., Münker, C., Fonseca, R.O.C., Albarède, F., Rohrbach, A., Lagos, M., Schmidt, M.W., Jochum, K.P., Stoll, B., Weis, U., Helmy, H.M., 2013. The U/Pb ratio of the Earth's mantle—A signature of late volatile addition. Earth Planet Sci Lett 362, 237–245. https://doi.org/10.1016/J.EPSL.2012.11.049

Barboni, M., Boehnke, P., Keller, B., Kohl, I.E., Schoene, B., Young, E.D., McKeegan, K.D., 2017. Early formation of the Moon 4.51 billion years ago. Sci Adv 3. https://doi.org/10.1126/sciadv.1602365

Barrat, J.A., Zanda, B., Moynier, F., Bollinger, C., Liorzou, C., Bayon, G., 2012. Geochemistry of CI chondrites: Major and trace elements, and Cu and Zn Isotopes. Geochim Cosmochim Acta 83, 79–92. https://doi.org/10.1016/j.gca.2011.12.011

Barrat, J.A., Zanda, B., Jambon, A., Bollinger, C., 2014. The lithophile trace elements in enstatite chondrites, Geochim. Cosmochim. Acta, 128, 71-94.

Borg, L.E., Brennecka, G.A., Kruijer, T.S., 2022. The origin of volatile elements in the Earth-Moon system. Proc Natl Acad Sci USA 119, e2115726119. https://doi.org/10.1073/PNAS.2115726119/SUPPL_FILE/PNAS.2115726119.SAPP.PDF

Borg, L.E., Cassata, W.S., Wimpenny, J., Gaffney, A.M., Shearer, C.K., 2020. The formation and evolution of the Moon's crust inferred from the Sm-Nd isotopic systematics of highlands rocks. Geochim Cosmochim Acta 290, 312–332. https://doi.org/10.1016/J.GCA.2020.09.013

Borg, L.E., Connelly, J.N., Boyet, M., Carlson, R.W., 2011. Chronological evidence that the Moon is either young or did not have a global magma ocean. Nature 477, 70–73. https://doi.org/10.1038/nature10328

Borg, L.E., Gaffney, A.M., Kruijer, T.S., Marks, N.A., Sio, C.K., Wimpenny, J., 2019. Isotopic evidence for a young lunar magma ocean. Earth Planet Sci Lett 523, 115706. https://doi.org/10.1016/J.EPSL.2019.07.008

Borg, L.E., Carlson, R.W., 2023. The evolving chronology of Moon formation, Annu. Rev. Earth Planet. Sci. 51, 25-52. https://doi.org/10.1146/annurev-earth-031621-060538

Boyet, M., Carlson, R.W., Borg, L.E., Horan, M., 2015. Sm-Nd systematics of lunar ferroan anorthositic suite rocks: Constraints on lunar crust formation. Geochim Cosmochim Acta 148, 203–218. https://doi.org/10.1016/j.gca.2014.09.021

Brenan, J.M., Mungall, J.E., Bennett, N.R., 2019. Abundance of highly siderophile elements in lunar basalts controlled by iron sulfide melt. Nature Geoscience 2019 12:9 12, 701–706. https://doi.org/10.1038/s41561-019-0426-3

Canup, R.M., 2012. Forming a moon with an Earth-like composition via a giant impact. Science (1979) 338, 1052–1055. https://doi.org/10.1126/science.1226073

Carlson, R.W., Borg, L.E., Gaffney, A.M., Boyet, M., 2014. Rb-Sr, Sm-Nd and Lu-Hf isotope systematics of the lunar Mg-suite: The age of the lunar crust and its relation to the time of Moon formation. Philosophical Transactions of the Royal Society A: Mathematical, Physical and Engineering Sciences 372. https://doi.org/10.1098/rsta.2013.0246

Carlson, R.W., Lugmair, G.W., 1988. The age of ferroan anorthosite 60025: oldest crust on a young Moon? Earth Planet Sci Lett 90, 119–130. https://doi.org/10.1016/0012-821X(88)90095-7

Charlier, B.L.A., Tissot, F.L.H., Vollstaedt, H., Dauphas, N., Wilson, C.J.N., Marquez, R.T., 2021. Survival of presolar p-nuclide carriers in the nebula revealed by stepwise leaching of Allende refractory inclusions. Sci. Adv 7.

Charnoz, S., Michaut, C., 2015. Evolution of the protolunar disk: Dynamics, cooling timescale and implantation of volatiles onto the Earth. Icarus 260, 440–463. https://doi.org/10.1016/j.icarus.2015.07.018

Charnoz, S., Sossi, P.A., Lee, Y.N., Siebert, J., Hyodo, R., Allibert, L., Pignatale, F.C., Landeau, M., Oza, A. v., Moynier, F., 2021. Tidal pull of the Earth strips the proto-Moon of its volatiles. Icarus 364, 114451. https://doi.org/10.1016/J.ICARUS.2021.114451

Connelly, J.N., Bizzarro, M., Krot, A.N., Nordlund, Å., Wielandt, D., Ivanova, M.A., 2012. The absolute chronology and thermal processing of solids in the solar protoplanetary disk. Science, 338, 651-655. 10.1126/science.1226919

Connelly, J.N., Bizzarro, M., 2016. Lead isotope evidence for a young formation age of the Earth–Moon system. Earth Planet Sci Lett 452, 36–43. https://doi.org/10.1016/j.epsl.2016.07.010

Ćuk, M., Stewart, S.T., 2012. Making the moon from a fast-spinning earth: A giant impact followed by resonant despinning. Science, 338, 1047–1052. https://doi.org/10.1126/science.1225542

Dauphas, N., Burkhardt, C., Warren, P.H., Teng, F.Z., 2014. Geochemical arguments for an Earth-like Moon-forming impactor. Philosophical Transactions of the Royal Society A: Mathematical, Physical and Engineering Sciences 372. https://doi.org/10.1098/rsta.2013.0244

Day, J.M.D., Pearson, D.G., Taylor, L.A., 2007. Highly siderophile element constraints on accretion and differentiation of the earth-moon system. Science (1979) 315, 217–219. https://doi.org/10.1126/science.1133355

Elardo, S.M., Draper, D.S., Shearer, C.K., 2011. Lunar Magma Ocean crystallization revisited: Bulk composition, early cumulate mineralogy, and the source regions of the highlands Mg-suite. Geochim Cosmochim Acta 75, 3024–3045. https://doi.org/10.1016/j.gca.2011.02.033

Elkins-Tanton, L.T., Burgess, S., Yin, Q.Z., 2011. The lunar magma ocean: Reconciling the solidification process with lunar petrology and geochronology. Earth Planet Sci Lett 304, 326–336. https://doi.org/10.1016/j.epsl.2011.02.004

Fischer-Gödde, M., Kleine, T., 2017. Ruthenium isotopic evidence for an inner solar system origin of the late veneer. Nature, 541:525–527.

Fischer, R.A., Zube, N.G., Nimmo, F., 2021. The origin of the Moon's Earth-like tungsten isotopic composition from dynamical and geochemical modeling, Nature Comm. 12, 35- https://doi.org/10.1038/s41467-020-2026

Fitoussi, C., Bourdon, B., 2012. Silicon isotope evidence against an enstatite chondrite earth. Science (1979) 335. https://doi.org/10.1126/science.1219509

Fu, H., Jacobsen, S.B., Sedaghatpour, F., 2023. Moon's high-energy giant-impact origin and differentiation timeline inferred from Ca and Mg stable isotopes, Commun. Earth Environ. 4, 307, https://doi.org/10.1038/s43247-023-00974-4

Gaffney, A.M., Borg, L.E., 2014. A young solidification age for the lunar magma ocean. Geochim. Cosmochim. Acta, 140:227–240.

Gross, J., Treiman, A.H., Mercer, C.N., 2014. Lunar feldspathic meteorites: Constraints on the geology of the lunar highlands, and the origin of the lunar crust. Earth Planet Sci Lett 388, 318–328. https://doi.org/10.1016/j.epsl.2013.12.006

Halliday, A.N., 2008. A young Moon-forming giant impact at 70-110 million years accompanied by late-stage mixing, core formation and degassing of the Earth. Phil. Trans. Roy. Soc. A, 366, 4163–4181. https://doi.org/10.1098/rsta.2008.0209

Halliday, A.N., 2004. Mixing, volatile loss and compositional change during impact-driven accretion of the Earth. Nature 427, 505–509. https://doi.org/10.1038/nature02275

Halliday, A.N., Porcelli, D., 2001. In search of lost planets - The paleocosmochemistry of the inner solar system. Earth Planet Sci Lett 192, 545–559. https://doi.org/10.1016/S0012-821X(01)00479-4

Hans, U., Kleine, T., Bourdon, B., 2013. Rb-Sr chronology of volatile depletion in differentiated protoplanets: BABI, ADOR and ALL revisited. Earth Planet Sci Lett 374, 204–214. https://doi.org/10.1016/j.epsl.2013.05.029

Hin, R.C., Fitoussi, C., Schmidt, M.W., Bourdon, B., 2014. Experimental determination of the Si isotope fractionation factor between liquid metal and liquid silicate. Earth Planet Sci Lett 387. https://doi.org/10.1016/j.epsl.2013.11.016

Hin, R., Coath, C., Carter, P., Nimmo, F., Lai, Y.-J., Pogge von Strandmann, P.A.E., Willbold, M., Leinhardt, Z.M., Walter, M.J., Elliott, T., n.d. Magnesium isotope evidence that accretional vapour loss shapes planetary compositions. Nature 549, 511–515.

Hofmann, A.W., White, W.M., 1983. Ba, Rb and Cs in the Earth's Mantle. Zeitschrift fur Naturforschung - Section A Journal of Physical Sciences 38, 256–266. https://doi.org/10.1515/zna-1983-0225

Jacobson, S.A., Morbidelli, A., Raymond, S.N., O'Brien, D.P., Walsh, K.J., Rubie, D.C., 2014. Highly siderophile elements in Earth's mantle as a clock for the Moon-forming impact. Nature 508, 84–87. https://doi.org/10.1038/nature13172

Kadlag, Y., Tatzel, M., Frick, D.A., Becker, H., Kühne, P., 2021. In situ Si isotope and chemical constraints on formation and processing of chondrules in the Allende meteorite, Geochim. Cosmochim. Acta, 304, 234-257.

Kleine, T., Budde, G., Burkhardt, C., Kruijer, T.S., Worsham, E.A., Morbidelli, A., Nimmo, F., 2020. The Non-carbonaceous–Carbonaceous Meteorite Dichotomy. Space Sci Rev 216, 1–27. https://doi.org/10.1007/S11214-020-00675-W/FIGURES/7

Kleine, T., Hans, U., Irving, A.J.A.J., Bourdon, B., 2012. Chronology of the angrite parent body and implications for core formation in protoplanets. Geochim Cosmochim Acta 84, 186–203. https://doi.org/10.1016/j.gca.2012.01.032

König, S., Münker, C., Hohl, S., Paulick, H., Barth, A.R., Lagos, M., Pfänder, J., Büchl, A., 2011. The Earth's tungsten budget during mantle melting and crust formation. Geochim Cosmochim Acta 75, 2119–2136. https://doi.org/10.1016/j.gca.2011.01.031

Kruijer, T.S., Archer, G.J., Kleine, T., 2021. No ^{182}W evidence for early Moon formation. Nature Geoscience 2021 14:10 14, 714–715. https://doi.org/10.1038/s41561-021-00820-2

Kruijer, T.S., Kleine, T., 2017. Tungsten isotopes and the origin of the Moon. Earth Planet Sci Lett 475, 15–24. https://doi.org/10.1016/j.epsl.2017.07.021
Kruijer, T.S., Kleine, T., Fischer-Gödde, M., Sprung, P., 2015. Lunar tungsten isotopic evidence for the late veneer. Nature 520, 534–537. https://doi.org/10.1038/nature14360
Lock, S.J., Stewart, S.T., 2017. The structure of terrestrial bodies: Impact heating, corotation limits, and synestias. J Geophys Res Planets 122, 950–982. https://doi.org/10.1002/2016JE005239
Lock, S.J., Stewart, S.T., Petaev, M.I., Leinhardt, Z., Mace, M.T., Jacobsen, S.B., Cuk, M., 2018. The Origin of the Moon Within a Terrestrial Synestia. J Geophys Res Planets 123, 910–951. https://doi.org/10.1002/2017JE005333
Maltese, A., Mezger, K., 2020. The Pb isotope evolution of Bulk Silicate Earth: Constraints from its accretion and early differentiation history. Geochim Cosmochim Acta 271, 179–193. https://doi.org/10.1016/J.GCA.2019.12.021
Maurice, M., Tosi, N., Schwinger, S., Breuer, D., Kleine, T., 2020. A long-lived magma ocean on a young Moon. Sci Adv 6, eaba8949. https://doi.org/10.1126/SCIADV.ABA8949
Mezger, K., Maltese, A., Vollstaedt, H., 2021. Accretion and differentiation of early planetary bodies as recorded in the composition of the silicate Earth, Icarus, 365, 114497.
McDonough, W.F., Sun, S.S., Ringwood, A.E., Jagoutz, E., Hofmann, A.W., 1992. Potassium, rubidium, and cesium in the Earth and Moon and the evolution of the mantle of the Earth. Geochim Cosmochim Acta 56, 1001–1012. https://doi.org/10.1016/0016-7037(92)90043-I
McLeod, C.L., Brandon, A.D., Armytage, R.M.G., 2014. Constraints on the formation age and evolution of the Moon from 142Nd-143Nd systematics of Apollo 12 basalts. Earth Planet Sci Lett 396, 179–189. https://doi.org/10.1016/j.epsl.2014.04.007
Mougel, B., Moynier, F., Göpel, C., 2018. Chromium isotopic homogeneity between the Moon, the Earth, and enstatite chondrites. Earth Planet Sci Lett 481, 1–8. https://doi.org/10.1016/j.epsl.2017.10.018
Moynier, F., Day, J., Wataru, O., Yokoyama, T., Day, J.M.D., Okui, W., Bouvier, A., Walker, R.J., Podosek, F.A., 2012. Planetary-scale strontium isotopic heterogeneity and the age of volatile depletion of early Solar System materials. Article in The Astrophysical Journal 758, 45. https://doi.org/10.1088/0004-637X/758/1/45
Münker, C., 2010. A high field strength element perspective on early lunar differentiation. Geochim. Cosmochim. Acta 74, 7340–7361. https://doi.org/10.1016/j.gca.2010.09.021
Murthy, V.R., Evensen, N.M., Jahn, B.M., Coscio, M.R., 1971. Rb-Sr ages and elemental abundances of K, Rb, Sr, and Ba in samples from the Ocean of Storms. Geochim. Cosmochim. Acta 35, 1139–1153. https://doi.org/10.1016/0016-7037(71)90030-5
Nakajima, M., Stevenson, D.J., 2014. Investigation of the initial state of the Moon-forming disk: Bridging SPH simulations and hydrostatic models. Icarus 233, 259–267. https://doi.org/10.1016/j.icarus.2014.01.008
O'Brien, D.P., Walsh, K.J., Morbidelli, A., Raymond, S.N., Mandell, A.M., 2014. Water delivery and giant impacts in the 'Grand Tack' scenario. Icarus 239, 74–84. https://doi.org/10.1016/J.ICARUS.2014.05.009
Pahlevan, K., Stevenson, D.J., 2007. Equilibration in the aftermath of the lunar-forming giant impact. Earth Planet Sci Lett 262, 438–449. https://doi.org/10.1016/j.epsl.2007.07.055
Papanastassiou, D.A., Wasserburg, G.J., 1978. Strontium isotopic anomalies in the Allende Meteorite. Geophys Res Lett 5, 595–598. https://doi.org/10.1029/GL005I007P00595
Paquet, M., Day, J.M.D., Brown, D.B., Waters, C.L., 2022. Effective global mixing of the highly siderophile elements into Earth's mantle inferred from oceanic abyssal peridotites, Geochim. Cosmochim. Acta, 316, 347-362.
Philpotts, J.A., Schnetzler, C.C., 1970. Potassium, rubidium, strontium, barium, and rare-earth concentrations in lunar rocks and separated phases. Science, 167, 493–495. https://doi.org/10.1126/science.167.3918.493

946 Philpotts, J.A., Schnetzler, C.C., Bottino, M.L., Schuhmann, S., Thomas, H.H., 1972. Luna 16: Some
947 Li, K, Rb, Sr, Ba, rare-earth, Zr, and Hf concentrations. Earth Planet Sci Lett 13, 429–435.
948 https://doi.org/10.1016/0012-821X(72)90120-3
949 Righter, K., Boujibar, A., Humayun M., Yang, S., Rowland II, R., Pando, K., 2023. Activity model
950 for 36 elements in Fe-Ni-Si-SC liquids with application to terrestrial planet accretion and mantle
951 geochemistry: New data for Ru, Re, Pt, Os, Ti, Nb, and Ta, Geochim. Cosmochim. Acta, 354,
952 211-228.
953 Ringwood, A.E., Kesson, S.E., 1977. Basaltic magmatism and the bulk composition of the moon - II.
954 Siderophile and Volatile Elements in Moon, Earth and Chondrites: Implications for Lunar
955 Origin. The Moon 16, 425–464. https://doi.org/10.1007/BF00577902
956 Rotenberg, E., Davis, D.W., Amelin, Y., Ghosh, S., & Bergquist, B.A., 2012. Determination of the
957 decay-constant of ^{87}Rb by laboratory accumulation of ^{87}Sr. Geochim. Cosmochim. Acta, 85, 41–
958 57. https://doi.org/10.1016/J.GCA.2012.01.016
959 Rudge, J.F., Kleine, T., Bourdon, B., 2010. Broad bounds on Earths accretion and core formation
960 constrained by geochemical models. Nat Geosci 3. https://doi.org/10.1038/ngeo872
961 Schiller, M., Bizzarro, M., Fernandes, V.A., 2018. Isotopic evolution of the protoplanetary disk and
962 the building blocks of Earth and the Moon. Nature 555, 501–510.
963 https://doi.org/10.1038/nature25990
964 Schmidt, M.W., Kraettli, G., 2022. Experimental crystallization of the lunar magma ocean, initial
965 selenotherm and density stratification, and implications for crust formation, overturn and the
966 bulk silicate Moon composition. J. Geophys. Res., 127, e2022JE007187. https://doi.
967 org/10.1029/2022JE007187
968 Schnetzler, C.C., Nava, D.F., 1971. Chemical composition of Apollo 14 soils 14163 and 14259. Earth
969 Planet Sci Lett 11, 345–350. https://doi.org/10.1016/0012-821X(71)90191-9
970 Sedaghatpour, F., Teng, F.Z., Liu, Y., Sears, D.W.G., Taylor, L.A., 2013. Magnesium isotopic
971 composition of the Moon. Geochim Cosmochim Acta 120, 1–16.
972 https://doi.org/10.1016/j.gca.2013.06.026
973 Shahar, A., Hillgren, V.J., Young, E.D., Fei, Y., Macris, C.A., Deng, L., 2011. High-temperature Si
974 isotope fractionation between iron metal and silicate, Geochim. Cosmochim. Acta, 75, 7688-
975 7697
976 Sprung, P., Kleine, T., Scherer, E.E., 2013. Isotopic evidence for chondritic Lu/Hf and Sm/Nd of the
977 Moon. Earth Planet. Sci. Lett. 380:77-87.
978 Stracke, A., Palme, H., Gellissen, M., Münker, C., Kleine, T., Birbaum, K., Günther, D., Bourdon, B.,
979 Zipfel, J., 2012. Refractory element fractionation in the Allende meteorite: Implications for solar
980 nebula condensation and the chondritic composition of planetary bodies. Geochim Cosmochim
981 Acta 85. https://doi.org/10.1016/j.gca.2012.02.006
982 Taylor, D.J., McKeegan, K.D., Harrison, T.M., 2009. Lu–Hf zircon evidence for rapid lunar
983 differentiation. Earth Planet. Sci. Lett. 279:157–64.
984 Thiemens, M.M., Sprung, P., Fonseca, R.O.C., Leitzke, F.P., Münker, C., 2019. Early Moon
985 formation inferred from hafnium–tungsten systematics. Nat Geosci 12, 696–700.
986 https://doi.org/10.1038/s41561-019-0398-3
987 Thommes, E.W., Duncan, M.J., Levison, H.F., 2003. Oligarchic growth of giant planets. Icarus.
988 https://doi.org/10.1016/S0019-1035(02)00043-X
989 Touboul, M., Kleine, T., Bourdon, B., Palme, H., Wieler, R., 2007. Late formation and prolonged
990 differentiation of the Moon inferred from W isotopes in lunar metals. Nature 450.
991 https://doi.org/10.1038/nature06428
992 Touboul, M., Puchtel, I.S., Walker, R.J., 2015. Tungsten isotopic evidence for disproportional late
993 accretion to the Earth and Moon. Nature 520, 530–533. https://doi.org/10.1038/nature14355
994 Trinquier, A., Birck, J.L., Allègre, C.J., 2007. Widespread 54Cr heterogeneity in the inner solar system
995 Astrophys. J. 655, 1179—1185.

Trinquier, A., Elliott, T., Ulfbeck, D., Coath, C., Krot, A.N., Bizzarro, M., 2009. Origin of nucleosynthetic isotope heterogeneity in the solar protoplanetary disk, Science, 324, 374—376.
Walker, R.J., 2009. Highly siderophile elements in the Earth, Moon and Mars: Update and implications for planetary accretion and differentiation. Chemie der Erde 69, 101–125. https://doi.org/10.1016/j.chemer.2008.10.001
Wood, B.J., Halliday, A.N., 2010. The lead isotopic age of the Earth can be explained by core formation alone. Nature 465, 767–770. https://doi.org/10.1038/nature09072
Yobregat, E., Fitoussi, C., Bourdon, B., 2017. A new method for TIMS high precision analysis of Ba and Sr isotopes for cosmochemical studies. J Anal At Spectrom 32. https://doi.org/10.1039/c7ja00012j
Young, E.D., Kohl, I.E., Warren, P.H., Rubie, D.C., Jacobson, S.A., Morbidelli, A., 2016. Oxygen isotopic evidence for vigorous mixing during the Moon-forming giant impact. Science (1979) 351, 493–496. https://doi.org/10.1126/science.aad0525
Zhang, B., Lin, Y., Moser, D.E., Hao, J., Liu, Y., Zhang, J., Barker, I.R., Li, Q., Shieh, S.R., Bouvier, A., 2021. Radiogenic Pb mobilization induced by shock metamorphism of zircons in the Apollo 72255 Civet Cat norite clast. Geochim Cosmochim Acta 302, 175–192. https://doi.org/10.1016/J.GCA.2021.03.012
Zhang, J., Dauphas, N., Davis, A.M., Leya, I., Fedkin, A., 2012. The proto-Earth as a significant source of lunar material. Nat Geosci 5, 251–255. https://doi.org/10.1038/ngeo1429

Acknowledgements: We thank the NASA curatorial staff for giving us access to the Apollo lunar samples, Guillaume Caro, Chloé Michaut, Stéphane Labrosse and Herbert Palme for discussions, and Philippe Telouk for maintaining the TRITON in good operation. We thank Richard Carlson and an anonymous reviewer for their insightful comments that helped improve the manuscript. **Funding**: This study was funded both by the ANR project ISOVOL (ANR-12-BS06-0002) and by the Labex LIO (ANR-10-LABX-0066) and received partial support from the ERC project COSMOKEMS (#694819).

Author contributions: EY: Investigation, conceptualization, writing. CF: Funding acquisition, conceptualization, writing. BB: Funding acquisition, conceptualization, writing, formal analysis.

Supplementary materials

1. Radioactive correction in FAN 60025 for deriving the initial $^{87}Sr/^{86}Sr$ at 4.36 Ga

2. Calculation of the condensation temperature of Sr and W in the conditions of Moon formation

3. Additional figures

1.Radioactive correction in FAN 60025 for deriving the initial $^{87}Sr/^{86}Sr$ at 4.360 Ga

One question that arises is whether this sample has remained a closed system since its formation dated at 4.360±3 Ga using the Sm-Nd and U-Pb systems (Borg et al. 2011). As shown in Borg et al. (2011) and Carlson and Lugmair (1988), the mineral separates analyzed in this sample plot on an isochron that are indicative of a closed-system behavior for the Sm/Nd and U/Pb systems. Nevertheless, it is worth investigating whether a more complex history for Rb-Sr systematics in this rock is possible. To explore the maximum possible uncertainties, we consider two extreme scenarios.

First, one could consider that the Rb contained in the sample was added late and to obtain the maximum effect, it would have to be added at the last minute (no in-growth of ^{87}Sr). In this case, the measured $^{87}Sr/^{86}Sr$ ratio would have evolved with Rb/Sr = 0. The age of the Moon was recalculated and the results are plotted in Figure S4.

Second, in this case, we consider that the FAN 60025 has evolved with a higher Rb that was recently lost. The maximum Rb/Sr ratio that one could consider is that it was equal to $^{87}Rb/^{86}Sr_{BSM}$=0.014-0.022. In this case the calculated initial $^{87}Sr/^{86}Sr$ at 4.36Ga ranges between 0.69820217 and 0.69769987, which is below the value of the initial $^{87}Sr/^{86}Sr$ for the inner Solar System. Thus, this second scenario is considered even more unlikely.

2. Calculation of the condensation temperature of Sr and W in the conditions of Moon formation

In order to provide further arguments for the isotope equilibration between the Moon and the bulk Silicate Earth after the giant impact, the 50% condensation temperatures of Ca, Sr and W were calculated using the formalism described in Ivanov et al. (2022) using the FactSage™ software. The composition of the bulk Silicate Earth given in Ivanov et al. (2022) was used. Fegley et al.

(2023) have assumed that the activity coefficient of Sr was equal to 1. However, the software used here (FactSage™) includes in its data base a comprehensive data set especially developed for Sr, based on the PhD work of Shukla (2012) who has studied several binary and ternary compositions including Sr yielding an activity coefficient different from 1.

The 50% condensation temperature was calculated for various pressures ranging between 10 bar and 0.001 bar, corresponding to the conditions in the proto-lunar (or synestia) disk as described in Lock et al. (2018). The results are depicted in Figure 1 and clearly show that the condensation temperatures for the three elements are relatively close to each other. Thus, if there has been Ca and W isotope equilibration between the Moon and the bulk Silicate Earth, this must have been the case for Sr as well.

Table S1. $^{87}Sr/^{86}Sr$ ratios in NIST SRM 987

Standards	$^{87}Sr/^{86}Sr$	2 SE	2SE (ppm)	N blocks
std 1	0.7102485	0.0000014	2.0	986
std 2	0.7102487	0.0000015	2.2	957
std 3	0.7102501	0.0000012	1.7	616
std 4	0.7102477	0.0000013	1.8	564
std 5	0.7102478	0.0000013	1.8	417
std 6	0.7102504	0.0000010	1.5	971
std 7	0.7102483	0.0000011	1.6	798
std 8	0.7102482	0.0000012	1.7	1021
std 9	0.7102491	0.0000013	1.8	594
std 10	0.7102491	0.0000012	1.7	558
std 11	0.7102486	0.0000011	1.6	697
std 12	0.7102498	0.0000011	1.5	991
std 13	0.7102476	0.0000014	1.9	767
std 14	0.7102490	0.0000012	1.7	983
std 15	0.7102482	0.0000012	1.6	552
		2SD	2SD (ppm)	
Mean	0.7102487	0.0000017	2.4	15

Table S2 Values of input parameters in MonteCarlo simulations

Input parameter	Description	Range
M_0	Initial mass of protoEarth	1-1000 km radius
f_{Sr}^{i}	Relative incremental mass accreted at stage i	0-1
$^{87}Rb/^{86}Sr_{BSM}$	Parent daugther ratio bulk silicate Moon	0.00027-0.022
$^{87}Rb/^{86}Sr_{BSE}$	Parent daugther ratio bulk silicate Earth	0.077-0.089
$^{87}Sr/^{86}Sr_{Solar\ System}$	Initial ratio of Solar System measured in CAI	0.698978
T_{Moon}	Age of Moon formation	0-200 Ma

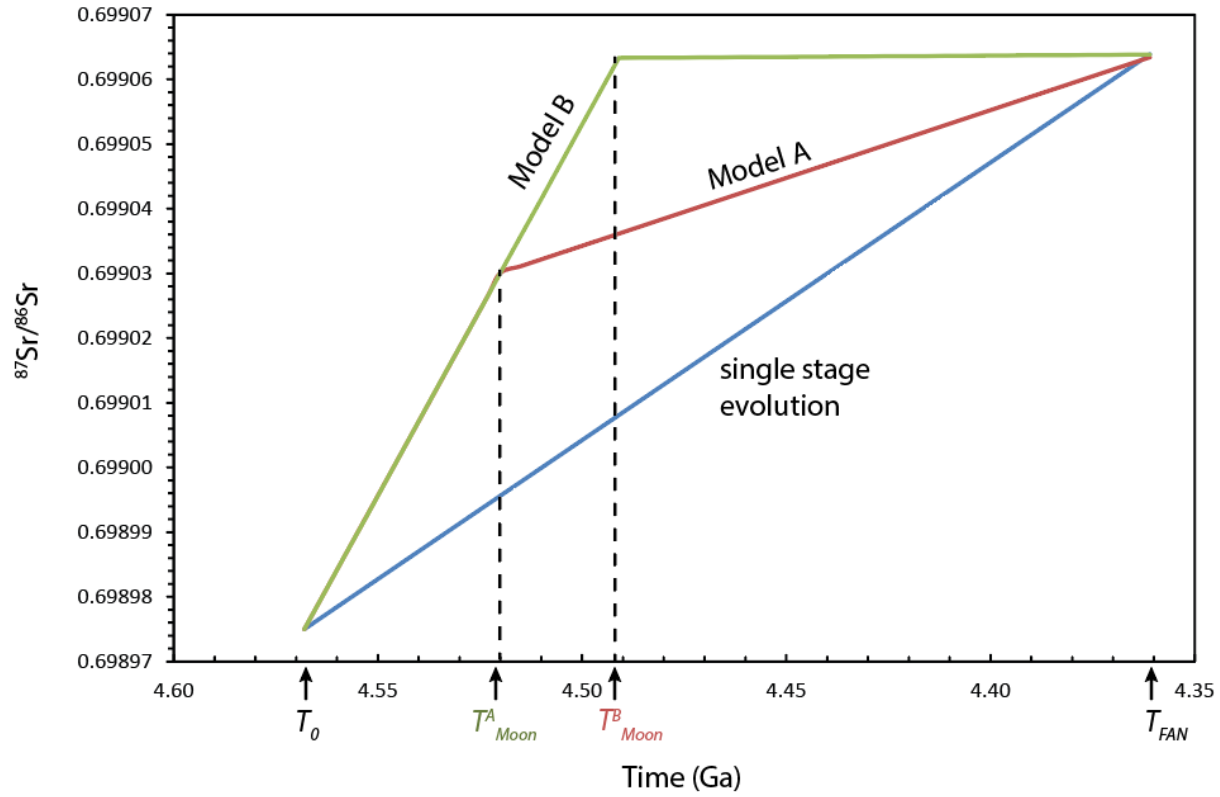

Figure S1

Evolution of the $^{87}Sr/^{86}Sr$ ratios for ferroanorthosite 60025 as a function of time for Model A and Model B compared with a single stage evolution starting from the beginning of the Solar System at T_0 as described in Carlson et al., (2014) and Carlson and Lugmair, (1988). Model A and Model B both assume that the Moon and bulk Silicate Earth had similar $^{87}Sr/^{86}Sr$ at the time of Moon formation (T^A_{Moon} and T^B_{Moon}) which is not the case of the single stage model. The $^{87}Rb/^{86}Sr$ of the bulk silicate Moon or the FAN 60025 (given by the slope of the evolution lines) are much lower than in the single stage model. The first stage of $^{87}Sr/^{86}Sr$ evolution had a chondritic $^{87}Rb/^{86}Sr$ (green line), the second stage had a $^{87}Rb/^{86}Sr=0.014$ for model A (red line) and $^{87}Rb/^{86}Sr=2.8\times10^{-4}$ for model B (flat green line).

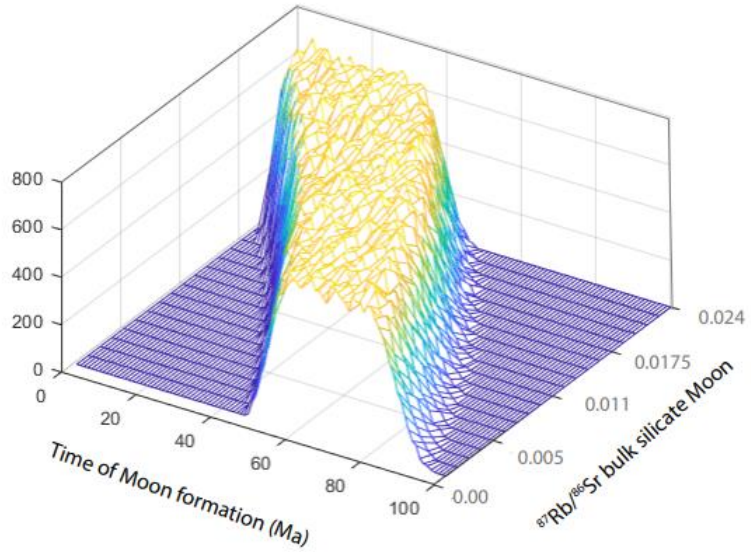

Figure S2 MonteCarlo solutions obtained for the age of the Moon versus $^{87}Rb/^{86}Sr$ in the bulk silicate Moon. The vertical axis represents the number of solutions. In this calculation, the uncertainty in the $^{87}Sr/^{86}Sr$ ratio of FAN 60025 is that measured obtained by Carlson and Lugmair (1988) i.e. 28 ppm.

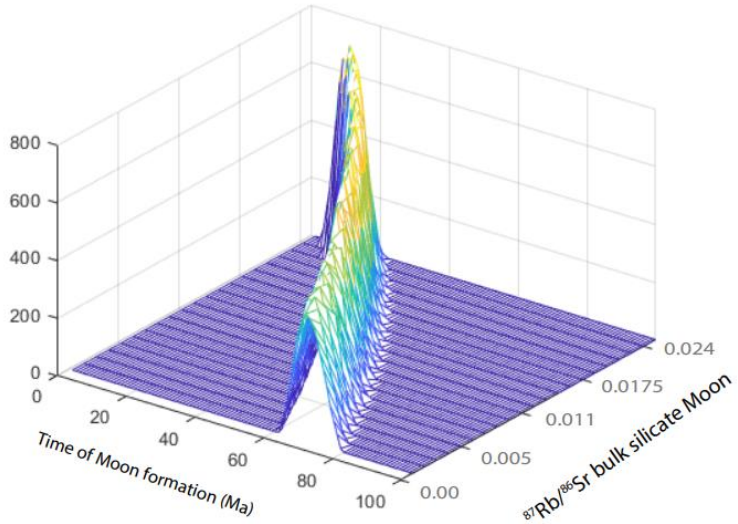

Figure S3 MonteCarlo solutions obtained for the age of the Moon versus $^{87}Rb/^{86}Sr$ in the bulk silicate Moon. The vertical axis represents the number of solutions. In this calculation, the uncertainty in the $^{87}Sr/^{86}Sr$ ratio of FAN 60025 is that measured in this study (2.4 ppm). The age range of solutions is far more restricted in this second case, which illustrates the benefit of the more precise measurement in constraining the age of the Moon.

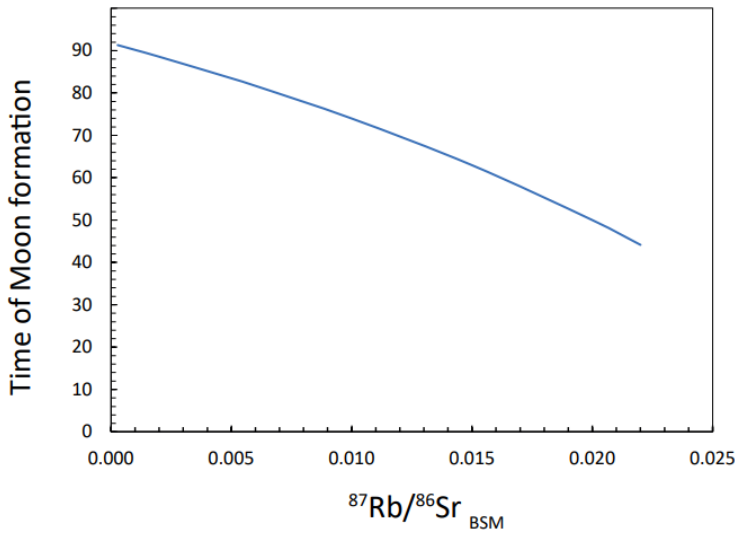

Figure S4 Forward model for the calculation of the age of the Moon based on Rb-Sr isotope systematics assuming that the Rb in FAN 60025 was not indigenous (i.e. added at the last minute). We consider this hypothesis unlikely as FAN 60025 shows isochrons for three isotope systems and the 60025 plagioclase falls on a Rb-Sr isochron defined with other FANs (Shih et al. 2005).

References

Borg, L.E., Connelly, J.N., Boyet, M., Carlson, R.W., 2011. Chronological evidence that the Moon is either young or did not have a global magma ocean. Nature 477, 70–73. https://doi.org/10.1038/nature10328

Carlson, R.W., Borg, L.E., Gaffney, A.M., Boyet, M., 2014. Rb-Sr, Sm-Nd and Lu-Hf isotope systematics of the lunar Mg-suite: The age of the lunar crust and its relation to the time of Moon formation. Phil. Trans. Roy. Soc. A 372. https://doi.org/10.1098/rsta.2013.0246

Carlson, R.W., Lugmair, G.W., 1988. The age of ferroan anorthosite 60025: oldest crust on a young Moon? Earth Planet Sci Lett 90, 119–130. https://doi.org/10.1016/0012-821X(88)90095-7

Fegley, Jr, B., Lodders, K., Jacobson, N.S., 2023. Chemical equilibrium calculations for bulk silicate earth material at high temperatures, Geochemistry 83, 125961. https://doi.org/10.1016/j.chemer.2023.125961

Ivanov, D., Fitoussi, C., Bourdon, B., 2022. Trace element volatility and the conditions of liquid-vapor separation in the proto-lunar disk, Icarus 386, 115143. https://doi.org/10.1016/j.icarus.2022.115143

Lock, S.J., Stewart, S.T., Petaev, M.I., Leinhardt, Z., Mace, M.T., Jacobsen, S.B., Cuk, M., 2018. The Origin of the Moon Within a Terrestrial Synestia. J Geophys Res Planets 123, 910–951. https://doi.org/10.1002/2017JE005333

Shih, C.Y., Nyquist, L.E., Reese, Y., Yamaguchi, A., Takeda, H., 2005. Rb-Sr and Sm-Nd Isotopic Studies of Lunar Highland Meteorite Y86032 and Lunar Ferroan Anorthosites 60025 and 67075. 36th Annual Lunar Planet. Sci. Conf., 1433.

Shukla (2012) Development of a Critically Evaluated Thermodynamic Database for the Systems Containing Alkaline-Earth Oxides, PhD Thesis, Ecole Polytechnique, Montreal, 350 pp. https://central.bac-lac.gc.ca/.item?id=NR91235&op=pdf&app=Library&is_thesis=1&oclc_number=1019464323