

Formation of the Moon and the Earth from a Common Supraplanetary Gas–Dust Cloud (Lecture Presented at the XIX All-Russia Symposium on Isotope Geochemistry on November 16, 2010)

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Abstract—A hypothesis is proposed on the formation of the Earth and the Moon from a large-scale gas–dust cloud, the size of which is limited by the Hill radius, i.e., approximately one million kilometers. The compression of the supraplanetary gas–dust cloud resulted in an adiabatic temperature increase in its interior parts and evaporation of volatiles, including iron, from the surface of particles. At a certain stage, within 50–70 Ma after solar system formation, the supraplanetary gas–dust disk was fragmented, the Moon was separated, and the Earth embryo was formed. The remaining part of the gas–dust material was accreted mainly to the Earth. During this process, the gas dominated by primordial hydrogen was squeezed out of the disk. Vapor was removed together with hydrogen from the interparticle space. The hydrodynamic lifting resulted in the loss of volatiles, including Rb, Xe, and Pb, which is reflected in the Rb–Sr, Xe–I–Pu, and U–Pb isotopic systems. The gas–dust accretion was accomplished within 110–130 Ma (most likely, ~120 Ma) after the beginning of solar system formation. Since then, the hydrodynamic lifting and volatile loss have ceased, and the history of the Earth as a condensed body has started.

Keywords: origin of the Moon, Earth accretion, planetary formation, ^{238}U – ^{206}Pb , ^{235}U – ^{207}Pb , ^{87}Rb – ^{87}Sr , ^{182}Hf – ^{182}W .

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INTRODUCTION

This contribution presents my ideas on the origin of the Earth–Moon system, which are strongly different from the currently most popular model.

They can be summarized into two suggestions.

The first is that the Moon was not produced by a giant impact (this hypothesis is widely accepted, especially in the western literature), but rather through the fragmentation of a common initial gas–dust body, which was also parental to the Earth.

The second is the generalization of the first suggestion and postulates that, in general, planets are not formed by the accretion of solid planetesimals, which is also considered as a commonly accepted truism, but rather through the development of gas–dust clouds, which grow and assemble into large (on the scale of the Hill radius) gas–dust bodies parental for particular planets and their satellites. Hereafter, they will be referred to as supraplanetary gas–dust bodies (clouds).

It is appropriate to present these concepts to experts in geochemistry and isotopic cosmochemistry, because the arguments that will be invoked are related

to the analysis of isotopic systems, including Hf–W, Rb–Sr, I–Xe, and U–Pb.

THE BEGINNING OF THE SOLAR SYSTEM

The development of isotopic studies has allowed very detailed reconstructions of the events that occurred 4.5 billion years ago (it is a pleasure to make this conclusion, because our studies contributed to this progress).

The solar system was formed owing to the collapse of an interstellar gas–dust cloud, which was evidently triggered by a supernova explosion. The new star (Sun) formed very rapidly, within one million years, which resulted in the preservation of short-lived isotopes produced during nucleogenesis (Podosek and Cassen, 1994). In particular, most of ^{26}Al with a half-life of only 730 thousand years was retained.

The history of the protoplanetary cloud surrounding the Sun dates from the appearance of the earliest solid objects. These are refractory globules enriched in Ca and Al, which were first studied in the Allende meteorite and called CAI (Calcium–Aluminum Inclusions). Their age is tightly constrained (Amelin

et al., 2002; Bouvier et al., 2008) as 4567.5 ± 0.5 Ma and is considered as the age of the solar system.

Chondrules were formed within 1.7–2.0 million years from the origin of the solar system. The accretion of chondrites completed 2–4 Ma after this. Chondrites of different petrologic classes were formed sequentially in time (Kurahashi et al., 2008; Kleine et al., 2008).

The most primitive matter of the solar system is represented by CI carbonaceous chondrites (Ivuna, Orgueil, and Murchison). Their compositions approach most closely solar element abundances. Important geochemical ratios, including Rb/Sr and Hf/W, are identical in carbonaceous chondrites and the solar nebula (Allegre et al., 2008). They are rich in carbon and contain organic matter, including amino acids and hydroxy acids, hydrous minerals (hydrosilicates), and carbonates. They were probably produced by the agglomeration of the dust component of the interstellar gas–dust nebula, the collapse of which resulted in the formation of the Sun. It did not undergo any significant temperature influence during the formation of the Sun, except for minor hydrothermal alteration. Carbonaceous chondrites of other classes (CM, CO, and CR) are mixtures of chondrules and a thermally metamorphosed matrix.

Meteorite parent bodies were 10–100 km and larger in size. The interior parts of large bodies underwent melting and differentiation. The application of Hf–W thermometry showed that the maximum heating was reached by approximately 6 Ma (Kleine et al., 2008). Then, the bodies cooled owing to the decay of short-lived isotopes, the heat production of which (in particular, ^{26}Al) sustained melting (Carlson and Lugmair, 2001). Fragments of such asteroidal materials were sources of achondrites. Within 5–10 Ma, large planetesimals transformed into differentiated bodies (Horan et al., 1998).

According to the theory, the subsequent collision of planetesimals resulted in the accumulation of increasingly larger bodies, until four planets (Mercury, Venus, Earth, and Mars) eventually formed in the inner part of the solar system. The theory of solid-phase accumulation was developed both in Russia (Safronov, 1969) and in western countries (Harris and Kaula, 1975; Stewart and Kaula, 1980; Wetherill, 1980; Wetherill and Cox, 1985).

The hypothesis of a giant impact, i.e., the formation of the Moon owing to the collision of two planet-sized bodies at the final stage of the solid-phase accumulation is in line with this accepted concept of planetary formation.

IMPACT HYPOTHESIS OF MOON FORMATION

The giant impact hypothesis was proposed in the mid-1970s by two groups of American researchers

(Hartmann and Davis, 1975; Cameron and Ward, 1976). According to this hypothesis, the Moon was formed by the ejection of the molten material of the Earth's silicate mantle owing to the impact of a Mars-sized body.

Any attempt to explain the formation of the Moon must answer the question why the Moon is strongly depleted in iron relative to the Earth. The abundances of iron in the Earth and Moon are 33.5 and 10–15%, respectively.

The giant impact hypothesis provided a simple explanation: the Moon was formed at the expense of the material of the Earth's mantle, when a considerable portion of iron was already concentrated in the metallic core.

In addition, the catastrophic impact explained the high angular momentum of the Earth–Moon system.

From a geochemical viewpoint, another convincing argument is that the Moon and the Earth are completely identical in such an important cosmogenic characteristic as the proportion of three oxygen isotopes ($^{16}\text{O}/^{17}\text{O}/^{18}\text{O}$). The trends of oxygen isotope fractionation are distinctive for different bodies of the solar system. Therefore, the coinciding trends of the oxygen isotope fractionation of the Moon and the Earth are compelling evidence for the genetic relations of their matter.

However, more comprehensive of the dynamics of investigations of the process showed that the molten material ejected into orbit by a giant impact includes at least 80% of the impactor and is not dominated by the Earth's mantle material (Melosh, 2000; Canap, 2004). Since the origin and chemical composition of the impactor are unknown, the geochemical arguments in favor of the impact hypothesis (including the iron abundances) were invalidated. As to the oxygen isotope characteristics of the Moon and the Earth, any accidental coincidence of fractionation trends is very improbable for genetically unrelated bodies. Various suggestions have been made to save the hypothesis; for instance, it was supposed that the protolunar material occurred in the atmosphere of silicate vapor ejected from the Earth by the giant impact, and the oxygen isotopic compositions of the Moon and the Earth were equilibrated (Stevenson, 2005). Experts in isotopic geochemistry can hardly concur with such an explanation. Another suggestion was that the Earth and the impactor were formed in the same zone of the solar system, practically on the same orbit, and inherited identical oxygen isotopic ratios from this zone. One variant of this hypothesis is the accumulation of the material of the future impactor in the Lagrange points (Belbruno and Gott, 2005). The stable accumulation and occurrence in a Lagrange point of a massive body comparable with the Earth is practically impossible.

Another pitfall with the giant impact hypothesis is the absence of evidence for the isotopic fractionation of elements lost as volatiles. It is known that the Moon is depleted in volatiles compared with the Earth. The dynamic modeling of the giant impact indicates that the material ejected into near-Earth orbit must consist of 10–20% melt and 80–90% vapor (Canap and Esposito, 1996; Canap, 2004). It was experimentally shown that the isotopic compositions of K, Mg, and Si can change by several per mil during melt evaporation (Wang et al., 1999; Humayun and Clayton, 1995). The scenario accepted for the giant impact implies that the evaporation and volatile loss had to be accompanied by a kinetic isotope effect (Humayun and Cassen, 2000). However, no isotopic shift was detected between the lunar and terrestrial materials. The explanation that this is due to vapor loss in a hydrodynamic regime (Pahlevan and Stevenson, 2008) is inadequate, because the kinetic isotope effect occurs during both gas dissipation and liquid–vapor phase transition.

Finally, the giant impact hypothesis is inconsistent (this will be discussed below) with evidence from the Hf–W, I–Xe, and U–Pb isotopic systems or requires artificial or untenable assumptions.

FORMATION OF THE EARTH–MOON SYSTEM THROUGH THE FRAGMENTATION OF A GAS–DUST CLOUD

The above difficulties, some of which were detected recently and some have long been recognized, instigate speculations on possible alternative models.

An alternative hypothesis was proposed by Galimov (1990, 1995, 2008) and Galimov et al. (2005). Its essence is that the Moon was formed simultaneously with the Earth as a fragment of a double system rather than owing to a giant impact.

ON THE EVOLUTION OF THE SOLAR NEBULA

In order to avoid misunderstanding, it should be noted that the early formation and growth of solid bodies in the solar system are beyond doubt and can be considered as observed facts. The meteorite materials available for study were formed during the first millions of years of the existence of the solar system. The ages of respective events are accurately constrained. The meteorites themselves are products of disintegration of larger bodies of asteroidal sizes. Vestiges of this process can be observed as numerous impact craters on the surface of planets and asteroids.

My suggestion is that this was probably not the only way of the evolution of the protoplanetary nebula. In addition, large-scale gas–dust clouds appeared and grew.

The precipitation of dust particles in the central plane of the solar nebula accompanying its radial con-

traction resulted in the formation of a thin dense sub-disk, the dusty material of which probably gave birth to solid-state evolution (Goldreich and Ward, 1973; Vityazev et al., 1990).

Simultaneously, there existed still the rather thick low-density part of the disk. The lifetime of the solar nebula is on the order of 10^7 yr (Podosek and Cassen, 1994).

Observations of extrasolar planetary systems indicate that stars older than several millions of years are usually devoid of visible gas–dust disks (Hairish et al., 2001; Chambers, 2004).

It is possible that the disk decomposed to local clumps, which could assemble to rather large (up to the Hill radius) gas–dust bodies.

The possible formation of gas–dust clumps has been discussed by many authors starting from the pioneering study of Gurevich and Lebedinskii (1950). As early as in the 1970s, Eneev (1979) and Eneev and Kozlov (1977) reported numerical simulation of the formation of planets and satellite systems from a gas–dust state. The appearance and evolution of clumps could be related to the development of turbulence in the gas–dust disk (Kolesnichenko and Marov, 2006). The investigations of these processes are part of our biospheric program (Marov et al., 2008; Adushkin et al., 2008; Zabrodin et al., 2008).

It should be emphasized that, rather than consider the role of clumps as precursors for planetesimals, i.e., an intermediate step in solid-phase accumulation, it is fundamentally important for the model proposed here to estimate the possibility of such an evolution which would eventually result in the formation of a large-scale protoplanetary gas–dust cloud, a supraplanetary body.

The hypothesis of the formation of the Earth–Moon system was based on such a scenario. The presence of gas in the cloud prevents its gravitational contraction. However, as the gas is blown off by the solar wind, conditions are reached when gravitational interaction became sufficient for the compression of the dust cloud. The compression of the particle cloud is accompanied by an adiabatic temperature increase. The solid particles are heated and partly evaporated. If the angular momentum is sufficiently high, the contracting dust cloud is separated into two or, sometimes, more fragments. This could result in the formation of the proto-Moon and proto-Earth from the supraplanetary body. In such a case, the Earth and the Moon (more precisely, their embryos) were formed from a single source (Galimov et al., 2005).

IRON LOSS FROM THE PROTO-MOON

A new interesting aspect of this hypothesis is that the heating of particles during the collapse provides a possible explanation for the iron loss from the Moon.

Component	CI wt %	Residue after 40 wt % evaporation	Moon
SiO ₂	35.0	42.9	43.4
MgO	23.4	31.9	32.0
FeO	36.9	15.8	13 + Fe in core
CaO + Al ₂ O ₃	4.6	9.4	10.8

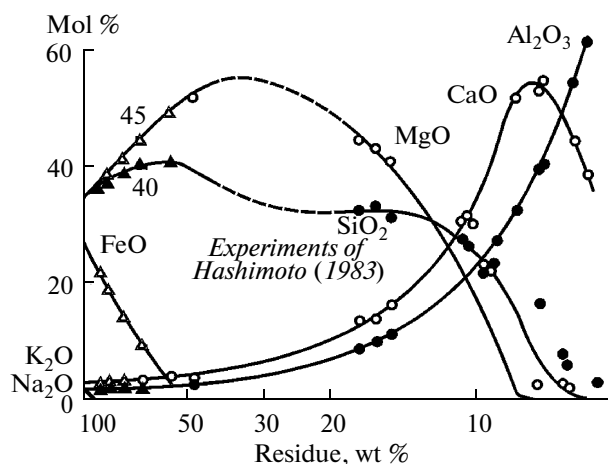


Fig. 1. Comparison of the lunar composition (Taylor, 1986) with the residue produced by 40% evaporation of the CI composition calculated using the experimental data of Hashimoto (1983).

Since the publication of the classical condensation sequence (Grossman and Larimer, 1974; Larimer, 1979, 1986), iron has been considered as a refractory element.

Indeed, the first condensed substances are Ca, Al, and Ti oxides followed by Fe and Ni. Silicate and oxides of Mg and Si are formed later. Among the major elements, relatively volatile are K and Na, and even more volatile are Rb and Pb.

However, if the evaporation of carbonaceous chondritic material enriched in iron oxide is considered instead of high-temperature gas-to-solid condensation (CAI and chondrules), it appears that iron can be lost as FeO. Ferrous iron is not volatile by itself. It is decomposed to elemental iron and oxygen, and iron vaporizes from a silicate or oxide mineral more efficiently than from a pure metallic phase (Hashimoto, 1983).

With respect to volatility, ferrous iron is next to K₂O and Na₂O (De Maria et al., 1971; Hashimoto, 1983; Markova et al., 1986). The depletion of the Moon in FeO relative to CI is similar to that of Mn and Cr, which show identical volatilities. Calculations on the basis of the experimental data of Hashimoto showed that the evaporation of 40 wt % of initial chondritic material is accompanied by iron depletion to typical lunar values (Galimov et al., 2005). This can be seen from the table in Fig. 1 showing the compositions of carbonaceous chondrites, the Moon, and the residue after the 40% evaporation of the mineral composition of carbonaceous chondrites.

It should be noted that this is true for the lunar composition by Taylor (1986) (Table 1), who postulated that the Moon is enriched in refractory elements (Al, Ca, and Ti). But this opinion is not accepted by all geochemists. For instance, Taylor estimated the Al₂O₃ abundance of the Moon as 6%, whereas Ringwood

(1986) advocated a value of 3.7%, which implies that the Moon is not enriched in Al relative to the Earth. O'Neill (1991) reported similar estimates.

The upper shell of the Moon is enriched in Al, Ca, and Ti minerals, which makes Taylor's estimates reasonable. However, the internal chemical composition of the Moon is unknown. It is possible that the lunar interior is depleted in refractory elements, and the bulk abundances of Al and other refractory elements are not higher than the terrestrial values.

The elastic properties of rocks are correlated with their mineral and chemical compositions. The chemical composition of the mantle can be estimated from seismic data. Such estimates were made by Kuskov and Kronrod (2008). They indicated elevated Al contents in the lunar mantle and, consequently, high abundances of refractory elements in the Moon.

Unfortunately, the Apollo seismic data set (Nakamura, 1983) characterizes only part of the lunar section.

This is why we proposed in 1995 the Luna-Globe cosmic program aimed at investigating the internal structure of the Moon (Galimov, 1995; Galimov et al., 1999). This project was postponed for 15 years. In 2010, the problem of the investigation of the internal structure of the Moon was omitted from the Luna-Globe program, because it was claimed that the remaining time before the scheduled launch in 2012 was insufficient for manufacturing penetrators; i.e., the main challenge of the initial project appeared to be discarded.

Currently, we accept Taylor's estimates as most plausible (although they require further substantiation), which implies that the Moon is enriched in Al, Ca, and Ti.

During the whole period of the compression of the supraplanetary disk, particles of chondritic composition were evaporated.

Table 1. Comparison of the compositions of carbonaceous chondrites (without H₂O, C, N, and other volatile components), the Earth's mantle, and the Moon estimated by various authors

Major oxide	CI carbonaceous chondrites	Earth (mantle and crust)	Moon		
			Taylor (1986)	Ringwood (1986)	O'Neill (1990)
SiO ₂	34.2	49.9	43.5	43.2	44.6
TiO ₂	0.11	0.10	0.3	0.3	0.17
Al ₂ O ₃	2.44	3.64	6.0	3.7	3.9
FeO	35.8	8.0	13.0	12.2	12.4
MgO	23.7	35.1	32.0	36.9	35.1
CaO	1.89	2.9	4.5	3.0	3.3
Na ₂ O	0.98	0.34	0.09	0.06	0.05
K ₂ O	0.10	0.02	0.01	—	—

Al-normalized elemental compositions of CI, the Earth's mantle, and the Moon (Taylor, 1986)

Element	CI	Earth	Moon
Si	12.3	12.0	6.4
Ti	0.05	0.05	0.05
Al	1.0	1.0	1.0
Fe	21.5	24.7	3.2
Mg	11.0	10.9	6.1
Ca	1.04	1.05	1.02
Na	0.57	0.13	0.02
K	0.06	0.01	0.0025

Note that, similar to the impact hypothesis, the loss of iron and, consequently, other volatile components from the Moon seems to be at odds with the absence of evidence for isotope fractionation on the Moon. However, there is a fundamental difference. The kinetic isotope effect occurs during evaporation from the surface of a condensed phase and vapor dissipation into open space. It is inevitable at any giant impact scenario. Evaporation in a closed system occurs differently. In a gas–dust cloud, materials evaporate from the surface of particles into the internal medium. The particles attain isotope exchange equilibrium with the enclosing vapor. Elements enter the gas phase in accordance with their volatility, and isotope fractionation is controlled by the thermodynamic isotope effect. The latter, in contrast to the kinetic isotope effect, is negligible at elevated temperatures. Moreover, the gas phase does not dissipate freely but is squeezed and completely removed from the surface of the gas–dust cloud by the solar wind. There is no isotope fractionation under such conditions. This is why isotopic shifts were not observed on the Moon for most of the elements which are evaporated under the condition of isotopic exchange, including K, Mg, and Si. However, isotope fractionation may occur, if this condition is violated owing to some particular features of

the process. This is exemplified by Fe and will be discussed below.

DYNAMIC MODEL OF FRAGMENTATION

The process of contraction was explored by particle dynamics simulation (Hockney and Eastwood, 1988; Krivtsov and Krivtsova, 2002; Le-Zakharov and Krivtsov, 2008). A particle ensemble is described by the classical equation of Newton dynamics. The dynamics of dust cloud compression was considered in the variant of a two-dimensional disk, whose angular momentum corresponded to that of the Earth–Moon system (Galimov et al., 2005).

In a standard problem, the interaction force between particles is controlled by three components: gravitational attraction, collision-related elastic interactions, and a component characterizing dissipative losses, i.e., the conversion of mechanical energy to heat. In addition to the angular momentum of the whole dust body, the initial conditions included a random vector of particle velocity simulating their chaotic movement and a certain radial distribution of the density of particles.

The computer simulation of disk compression showed that there is no fragmentation in the standard

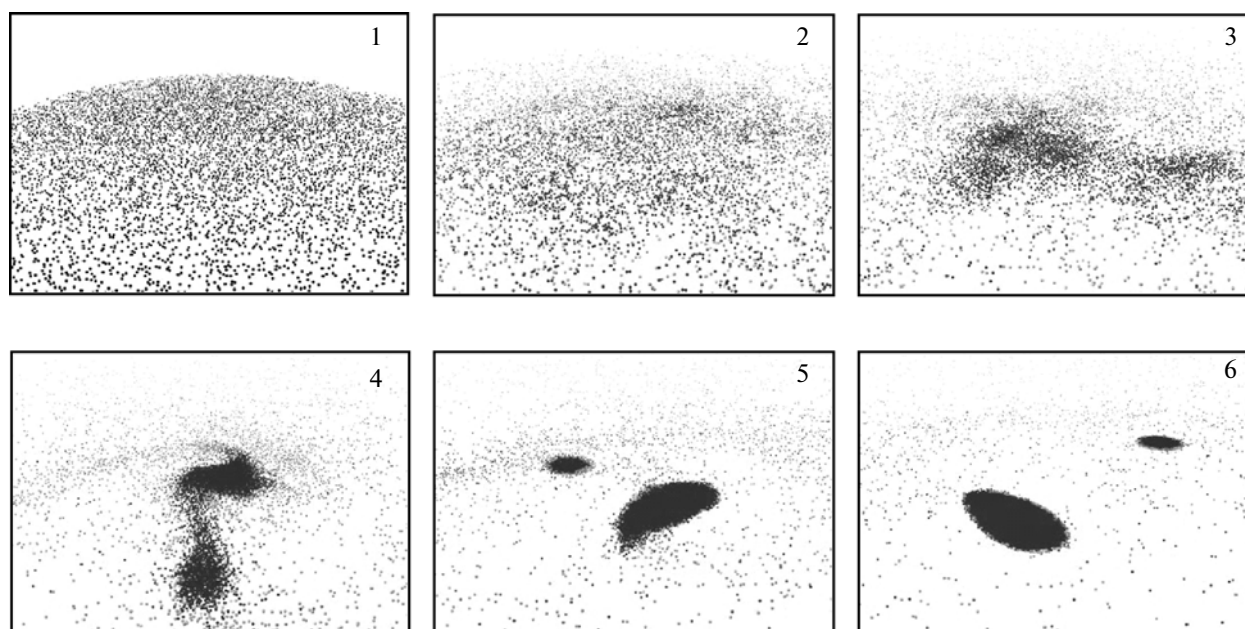


Fig. 2. Computer modeling of sequential phases of the fragmentation of the gas–dust cloud from which the Earth and Moon were formed (Galimov et al., 2005).

variant. The angular momentum in the Earth–Moon system is insufficient for fragmentation. This result is not surprising. It was previously hypothesized that the Moon was separated from the Earth. However, it was shown that, under any reasonable assumptions, the angular momentum of the Earth is insufficient for the separation of a Moon-sized fragment.

Nonetheless, the problem can be circumvented by including expulsion due to gas emission from the surface of evaporating particles in the balance of forces controlling the dynamics of dust cloud compression. This expulsion is subtracted from the gravitational attraction. In such a case, fragmentation can be reached at the angular momentum values that are characteristic of the Earth–Moon system.

Figure 2 shows phases of the development of fragmentation obtained by the computer simulation of contraction accounting for particle evaporation. These results were described in detail by Galimov et al. (2005) and Le-Zakharov and Krivtsov (2008).

Thus, the factor of evaporation provides a new interpretation for the Fe-depleted Moon composition and has a profound influence on the dynamic behavior of a contracting gas–dust cloud.

Both the fragments that gave rise to the Moon and the Earth were initially depleted in volatiles and iron almost to the same extent. However, according to computer simulation (Vasil'yev et al., 2004), if one of the fragments appeared (accidentally) somewhat more massive than the other, the larger embryo grows much faster. Therefore, the composition of the smaller embryo changes only slightly and retains its depletion

in volatiles and enrichment in refractory elements (future Moon), whereas the larger embryo (future Earth) scavenges almost all primary matter of the cloud and eventually approaches the composition of primary chondritic matter, except for the most volatile components.

GEOCHEMICAL ASPECTS OF THE MODEL

The supposed process of the formation of the Earth and its satellite from a contracting gas–dust body is beyond the traditional concepts on planet accumulation. Hence, the occurrence of some unusual and intriguing geochemical mechanisms should be expected (Fig. 3).

It is assumed that the parental gas–dust cloud of the Earth and Moon was chemically similar to the solar nebula. The dust component approached the composition of CI carbonaceous chondrites (in particular, iron occurred as FeO), whereas the gas medium was dominated by hydrogen, similar to the solar nebula. The resources of hydrogen were tremendous. Its mass in the composition of the supraplanetary body should be initially more than an order of magnitude greater than the total mass of particles.

Hydrogen executes two functions. First, it is a carrier gas providing the hydrodynamic removal of compounds and elements migrating into the gas phase during the compression of the gas–dust cloud. Second, it is an efficient reducer.

Temperatures sufficient for FeO evaporation will be reached in the central part of the fragmenting

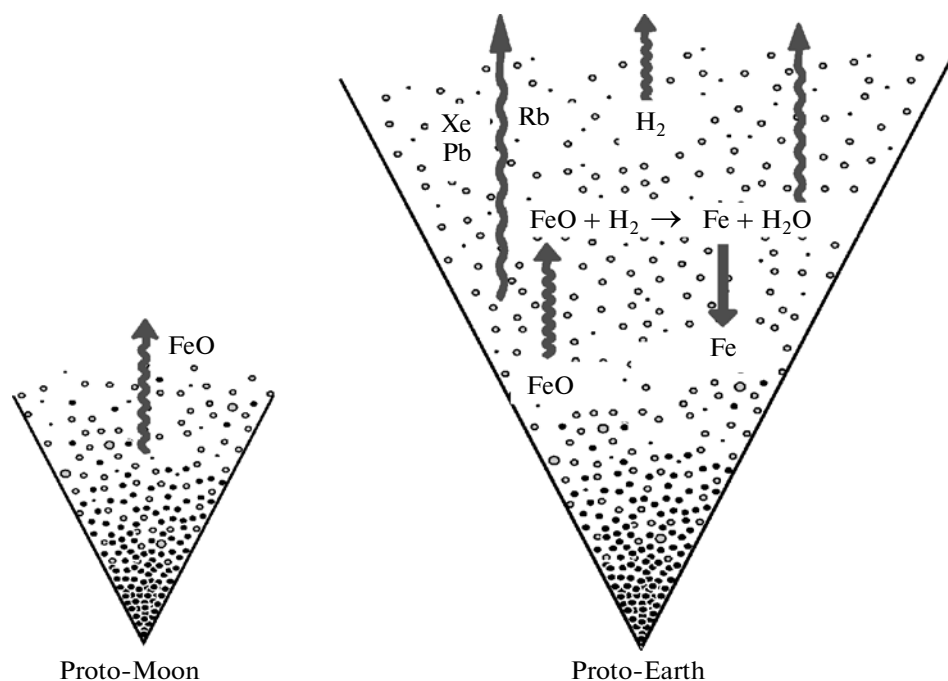


Fig. 3. Segregation of metal and the hydrodynamic escape of volatiles in the model of the formation of the Earth and Moon from a supraplanetary gas–dust body.

cloud. This is our starting assumption. The evaporation rate should be small at the time scale of the fragmentation (10^5 – 10^6 yr).

Experiments showed that the rate of FeO_s evaporation is $1.14 \cdot 10^{-5} \text{ g cm}^{-2} \text{ s}^{-1}$ at 1900 K (Kazemas and Tsvetkov, 2007). At such a temperature, half of FeO will evaporate from a particle of 1 mm radius within $0.2 \cdot 10^4$ s. Obviously, temperatures had to be no higher than 1000 K. There are no experimental data for such temperatures.

The evaporation of FeO is incongruent. In fact, Fe and O_2 occur in the vapor. The oxygen is fixed by hydrogen, and the metallic iron is separated in a dispersed state.

The embryos of the Moon and the Earth are depleted in Fe owing to FeO loss. However, during the continuing accretion of the Earth occurring in the hydrogen medium, the dispersed iron will be condensed into metallic particles.

The solid metallic particles can precipitate against the gas flow toward the center of mass.

Thus, **the formation of the Earth's future core begins owing to the precipitation of metallic iron through the vapor–gas phase.** This process imitates heterogeneous accretion.

The Moon retained residual iron mainly as FeO , whereas the Earth gained metallic iron during further accretion.

Such a scenario has a number of important consequences.

It has always been a difficult problem to explain why the Earth and other planets contain massive metallic cores, whereas oxidized iron occurs in the primary matter. The primary material of carbonaceous chondritic composition contains iron only in the FeO form. Ordinary chondrites contain very minor amounts of metallic iron.

The formation of a massive metallic iron core requires the removal of the equivalent mass of oxygen. No adequate mechanism has been proposed for this process.

Iron reduction by primary hydrogen during the compression of the particle cloud provides a natural solution to this problem. The water produced by this process is squeezed together with hydrogen in a vapor–gas phase from the compressed cloud of particles and expelled from the proto–Earth cloud. This is the process of hydrodynamic escape.

It is important to note in this connection that the general problem of Earth degassing is interpreted in a new manner. The Earth not only lost the volatile compounds of light elements, such as water, nitrogen, methane, and CO_2 , but also became depleted in heavy volatiles, which should have been retained. The hydrodynamic conditions in a hydrogen flow result in the removal of heavy volatile elements, including Xe and other noble gases, as well as nonatmophile heavy elements, such as Rb and Pb, occurring in a vapor phase under the P – T conditions of the gas–dust medium. Such removal is impossible from the surface of the

planet and without the tremendous resources of the carrier gas.

The loss of volatiles, including heavy elements, is a natural component of the scenario in which the forming planet had a large envelope consisting of particles mixed with a hydrogen-dominated gas phase.

The giant impact hypothesis invoking extensive, if not complete, melting of the Earth inevitably implies close to equilibrium partitioning of siderophile elements between silicate and metal melts, in particular, corresponding to equilibrium in the mantle–core system. However, the observed partitioning is far from equilibrium (Jones and Drake, 1986).

Siderophile properties depend on temperature, pressure, and oxygen fugacity. Therefore, at a certain combination of conditions, quantitative relations corresponding to equilibrium partition coefficients can be found. For instance, at certain temperatures and pressures in the mantle, the observed depletion of Mo and P corresponds to equilibrium (Righter et al., 2010). Changing the degree of mantle oxidation and the depth of the terrestrial magma ocean, the contents of Nb, V, and Cr can be made consistent with equilibrium partitioning in the system silicate melt–metallic melt (Wood et al., 2008). However, each element or element group requires its own combination of conditions.

Given the hydrodynamic mechanism of iron segregation, the metallic particles descending to the center of mass and silicate particles suspended in the gas do not participate in chemical exchange. Therefore, the partitioning of siderophile elements must not, in principle, correspond to equilibrium partition coefficients in the silicate melt–metal system, which is consistent with observations.

The volatility of NiO approaches that of FeO, and the behavior of Ni is similar to that of Fe (Kazenas and Tsvetkov, 2008). During reduction, nickel converts to the metallic phase and is transported into the core together with iron. Thus, an iron–nickel core is formed. On the other hand, such elements as Rh, Pt, Ru, Ir, and Os, which are much more siderophile than Ni, remain in silicate particles owing to their refractory nature and low volatility. They do not partition into the precipitating metallic phase. This is why their content in the silicate component is several orders of magnitude higher than could be expected on the basis of their equilibrium partitioning between silicate and metallic melts.

Although in the subsequent history of the condensed Earth there were obviously conditions for iron segregation in the silicate melt–metal system, the above feature of the formation of the Earth's core (mechanism heterogeneous to some extent) was responsible for the observed nonequilibrium distribution of siderophile elements in the Earth.

The giant impact hypothesis gives no grounds for expecting differences in the distribution of siderophile elements in the Earth and Moon. If there are such differences, the Moon must be less depleted in siderophile elements, because the relative size of its core (up to 5%) is much smaller than that of the Earth's core (~30%). In fact, the opposite is observed: the Moon is more strongly depleted in siderophile elements (Newsom, 1986). It is obvious from our model (Fig. 3) that hydrodynamic escape dominates during the accretion period after fragmentation; i.e., it is relevant mainly to the Earth. As to the Moon, after its formation as a consolidated body, the segregation of the lunar core occurred mainly in the silicate melt–metallic melt system. Correspondingly, the distribution of siderophile elements in the Moon is described by equilibrium segregation in the silicate–melt system accounting for partial volatility and **partial** melting (Newsom, 1986; Galimov, 2004).

Let us compare now to what extent these two alternative models agree with isotopic data and satisfy isotope geochemical constraints.

ISOTOPE FRACTIONATION, $^{57}\text{Fe}/^{54}\text{Fe}$

We begin from noting that the acceptance of a common source for terrestrial and lunar materials revives the geochemical arguments that were proposed in connection with the initial giant impact model, which postulated that the Moon was formed from the material of the Earth's mantle (Wanke and Dreibus, 1986; Ringwood, 1986).

This concerns primarily the coincidence of the ^{16}O – ^{17}O – ^{18}O fractionation trends, which naturally follows from the model proposed but can hardly be explained within the giant impact hypothesis.

It was noted above that the Moon shows no isotopic shifts for those elements that are depleted relative to the Earth, for instance, K, Si, and Mg. Significant isotopic shifts were observed for these elements in experiments on silicate melt evaporation (Humayun and Cassen, 2000; Wang et al., 1999), but they are not manifested in the composition of the Moon. This fact is difficult to reconcile with the giant impact hypothesis.

The absence of isotope fractionation is explained in our model by the fact that evaporation occurs in such a case into the space between particles. This leads to isotope exchange and negligible isotopic differences (thermodynamic isotope effect) between the particles and vapor (Fig. 4).

However, the elements that are irreversibly separated from the vapor into an independent phase must be an exception. This concerns iron, which is condensed into an independent phase after evaporation. A kinetic isotope effect must operate in such a case. The metallic particles migrating to the core must be

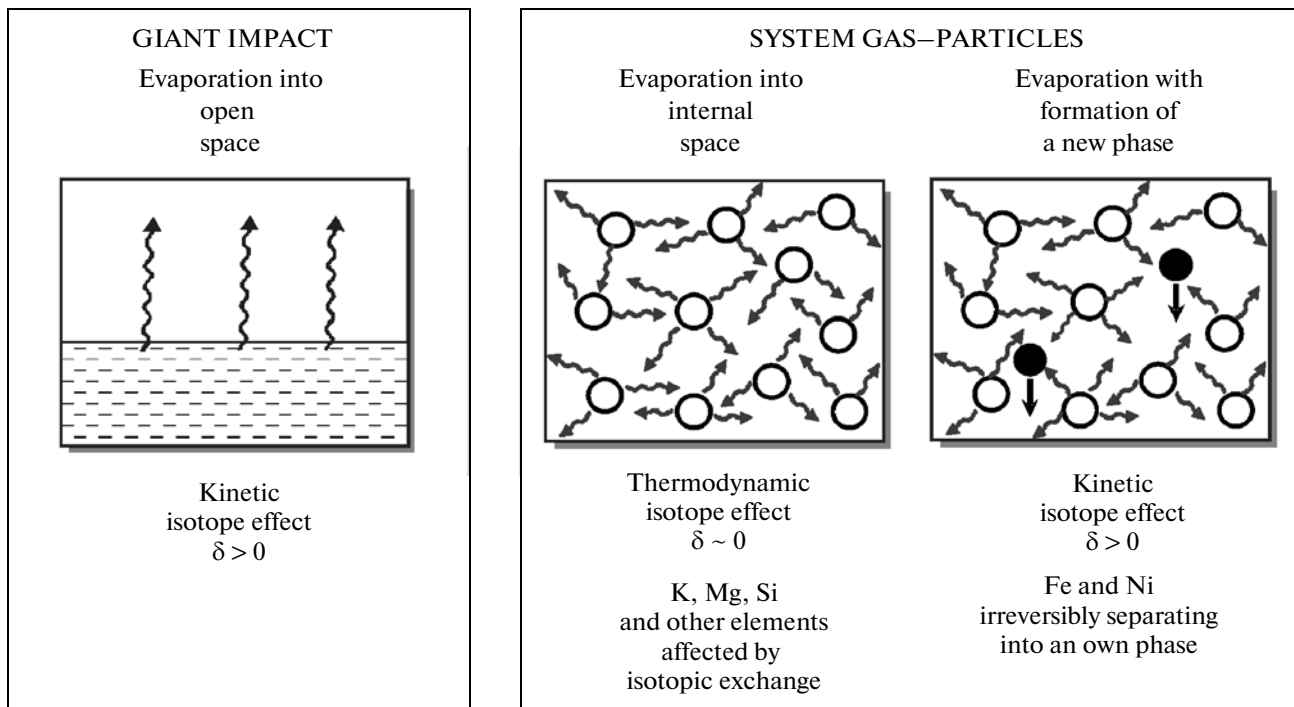


Fig. 4. Illustration of the types of isotope fractionation discussed in the text.

enriched in the light isotope (^{54}Fe), and the iron retained in silicate particles must be enriched in ^{57}Fe . This is consistent with observations.

Iron meteorites, which are considered as fragments of the metallic cores of asteroidal bodies, are enriched to some extent in the heavy isotope ^{57}Fe relative to chondrites (Poitrasson et al., 2004; Williams et al., 2005). Had it been related to differentiation in the metal–silicate system, the iron of the silicate component would be complementarily depleted in ^{57}Fe (Weyer et al., 2005).

However, it appears that both the Earth's mantle and the silicate phase of the Moon are enriched in ^{57}Fe (by 0.1 and 0.2‰, respectively) compared with chondrites. This is in agreement with the proposed scenario for the formation of the Earth's core.

Nickel must show a similar behavior, because it evaporates from particles in a similar way as NiO, converts into Ni in the vapor phase, and migrates, then, to the core together with Fe. It should be expected that the silicate Ni will be enriched in the heavy isotope.

The enrichment of the Moon in ^{57}Fe could have been attributed to the giant impact. Poitrasson (2007) suggested that the enrichment of the Moon in the heavy Fe isotope is “best explained by loss of light iron isotopes during the high temperature event accompanying the interplanetary impact that led to the formation of the Moon”. However, this author avoided the question why the Moon is not enriched in the heavy

isotopes of other elements that were lost from the Moon as volatiles during the giant impact.

The observed distribution of stable isotopes in the Earth–Moon system is consistent in detail with the hypothesis proposed here and is at odds with the giant impact hypothesis.

ISOTOPIC SYSTEMS

Hf–W

The Hf–W systematics is most appropriate for the analysis of the problem of the formation of the Earth and the Moon.

It is known that ^{182}Hf is a short-lived isotope transforming into ^{182}W with a half-life of 8.9 Ma. Hafnium and tungsten are fractionated during iron separation from silicates. Hafnium is a lithophile element and is retained in silicates, whereas W shows a siderophile affinity and partitions into the metallic phase. Since the precursor of ^{182}W , the ^{182}Hf radioisotope, remains in the silicate part, the $^{182}\text{W}/^{184}\text{W}$ ratio of the silicate increases compared with undifferentiated chondritic material. The earlier the differentiation, the higher the isotopic shift (ϵ_w). The differentiation must occur within the earliest tens of million years, before the complete decay of short-lived ^{182}Hf . Early studies (Lee and Halliday, 1995; Lee et al., 1997; Halliday and Lee, 1999) reported rather high positive ϵ_w values for the Moon and negligible isotopic shifts for the Earth.

However, the analytical method used in these studies was subsequently challenged (Schoenberg et al., 2002; Yin et al., 2002). The revision of isotopic data resulted in the re-estimation of the time of the formation of the Earth's core and the Moon (Yin et al., 2002). Additional uncertainty was associated with the fact that the interaction of Ta with cosmogenic neutrons, $^{181}\text{Ta} (n) \rightarrow \text{Ta}^{182}(\beta) \rightarrow ^{182}\text{W}$, can significantly contribute to the production of excess ^{182}W (Leya et al., 2000). Recent investigations of W isotopic systematics in Ta-free lunar metals showed that there is no excess ^{182}W in such a case (Kleine et al., 2005; Touboul et al., 2007).

Eventually, a consensus was reached after the revision of the previous results, and the following conclusions were formulated (Kleine et al., 2009).

The Hf/W ratios of the present-day Earth and Moon are different: 17 ± 5 and 26 ± 2 , respectively.

The Earth and the Moon have almost identical W isotopic compositions,

$$\Delta\epsilon^{182}\text{W} = 0.09 \pm 0.1.$$

The W isotopic compositions of the Earth and the Moon are slightly different from the chondritic composition,

$$\epsilon^{182}\text{W} = +1.9.$$

Consider the consequences of these results in the context of the hypothesis of the giant impact, on the one hand, and the hypothesis of fragmentation, on the other hand.

Since the isotopic compositions of the Moon and the Earth are identical but their Hf/W ratios are different, the observed difference had to appear after the almost complete decay of the precursor of radiogenic ^{182}W , i.e., ^{182}Hf .

Kleine et al. (2009) reported the following estimate. In order to obtain a $\Delta\epsilon^{182}\text{W}$ value within 0.09 ± 0.1 (if the value $+0.09$ is significant!), the age of the event must be ~ 60 Ma after solar system formation. In any case, the Moon could not be formed earlier than 50 Ma.

This value is an important time constraint for events in both hypotheses. In the fragmentation hypothesis, it corresponds to the time of fragmentation, when the geochemical histories of the Moon and the Earth diverged. Until that moment, they consisted of the same material, and their W isotopic compositions were obviously identical. After ~ 60 Ma, the Hf/W ratios became different, but the W isotopic compositions of the Earth and the Moon could not already change significantly.

In the giant impact hypothesis, this value corresponds to the time of the impact. But the situation is more dubious in that case. The bodies colliding during the giant impact had experienced different geochemi-

cal histories before this event. Therefore, either they accidentally had identical W isotopic characteristics, or the giant impact had to lead to the isotopic homogenization of the whole mass formed during the event. This problem is similar to that arising in the giant impact hypothesis in connection with the observed coincidence of the oxygen isotope fractionation trends of the Moon and the Earth.

In any case, this is in conflict with the dynamic model of a giant impact, which indicates that the impactor material accounts for more than 80% of the ejected material and, correspondingly, the hypothetical Moon.

The authors are aware of this difficulty. It is pertinent to cite Kleine et al. (2009): “that two such different objects as the proto-Earth and impactor would evolve to identical W isotope compositions in their mantles seems highly unlikely...,” and “the identical W isotope compositions of the lunar and terrestrial mantles could indicate that the Moon is largely derived from terrestrial mantle material, but this is inconsistent with results from numerical simulations, all of which indicate that the Moon predominantly consists of impactor material”.

The data of Hf–W isotope analysis pose difficult and, in my opinion, insurmountable problems for the giant impact hypothesis. In contrast, the identical W isotopic characteristics of the Moon and Earth appear natural for the fragmentation hypothesis. Simultaneously, a numerical estimate for the time of fragmentation is obtained: approximately 60 million years after the birth of the solar system.

Previous studies attributed the difference in ϵ_w between the chondritic (primordial) and terrestrial materials to the formation of the core. Indeed, this is the most obvious possibility related to the tendency of Hf–W fractionation in the silicate–metal system. On the other hand, an ϵ_w shift can merely result from a difference between the assumed chondritic and primary proto-terrestrial (lunar) Hf/W ratios. Calculations show that the observed $\epsilon_w = +1.9$ could be formed if the Hf/W ratio of the proto-terrestrial gas–dust body was on average 2.2 times higher than the chondritic value for 50–60 Ma. Note that the Hf/W ratio of carbonaceous chondrites, $^{180}\text{Hf}/^{184}\text{W} = 1.21 \pm 0.06$, is twice as high as that of H chondrites, $^{180}\text{Hf}/^{184}\text{W} = 0.63 \pm 0.20$ (Kleine et al., 2008).

The experimentally measured age of the oldest lunar rocks constrains any hypothesis. While the lower boundary for Moon formation is 50 Ma, the upper boundary is defined by the age of the oldest rocks.

It can be seen from Table 2 that the age of the oldest known lunar rocks is 4.44–4.50 Ga; i.e., the Moon was formed between 50 and 70 Ma after solar system formation.

Table 2. Age estimates for the oldest lunar rocks

Object	Age		Isotope system	Reference
	Before present (Ga)	After solar system formation (Ma)		
Ferroanorthosite 60025	4.5 ± 0.01	67 ± 10	²⁰⁶ Pb/ ²⁰⁷ Pb Recalculated data of Hanan and Tilton (1987)	<i>Hanan, Tilton, 1987</i> <i>Halliday, 2008</i>
	4.45 ± 0.1	117 ± 100		
Lunar crust Descartes breccia 67215	4.46 ± 0.04	107 ± 40	U-Pb ¹⁴⁷ Sm- ¹⁴³ Nd	<i>Norman et al. 2003</i> <i>Tera et al., 1973</i>
	4.47 ± 0.02	97 ± 20		
Ferroanorthosite 67016	> 4.5 (*)	< 70	¹⁴⁷ Sm- ¹⁴³ Nd Sm - Nd	<i>Alibert et al., 1994</i> <i>Borg et al., 1999</i>
Ferroanorthosite 62236	4.29 ± 0.03	277 ± 30		
KREEP	4.42 ± 0.07	137 ± 70	Sm - Nd	<i>Nyquist, Shih, 1992</i> <i>Shih et al., 1993</i>
Norite from breccia 15445	4.46 ± 0.07	107 ± 60		
Ferroanorthosite 60025	4.44 ± 0.02	127 ± 20		

* 4.562 ± 0.068

Rb/Sr

Rubidium-87 is a long-lived isotope with a half-life of $T_{1/2} = 48.8 \cdot 10^9$ yr and a decay constant of $\lambda_{87} = 1.42 \cdot 10^{-11}$ yr⁻¹.

One characteristic feature of the Rb-Sr system relevant to the model discussed is that Rb is a rather volatile element, which implies that the Moon must be significantly depleted in Rb.

The starting point of the evolution of the Rb-Sr system is defined by the solar $^{87}\text{Rb}/^{86}\text{Sr} = 0.92$ (Grevese et al., 1998) and the initial Sr isotope ratio of the earliest CAI: $(^{87}\text{Sr}/^{86}\text{Sr}) = 0.69892$ (Carlson and Lugmair, 1988). The Moon has $^{87}\text{Rb}/^{86}\text{Sr} = 0.018$ and the initial Sr isotope ratio $(^{87}\text{Sr}/^{86}\text{Sr})_{i\text{-Moon}} = 0.69906$ (Carlson and Lugmair, 1988).

Given the observed $^{87}\text{Rb}/^{86}\text{Sr}$ value of the Moon (0.018), the Sr isotope ratio would only 0.69893 after 60 Ma, i.e., lower than the lunar ratio (0.69906).

In our model, this is the primary gas-dust body. Variations in Sr isotopic composition in a gas-dust body gradually losing Rb can be described in the following way.

In a given moment (*t*), the Sr isotopic composition is specified as

$$\left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}}\right)_t = \left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}}\right)_0 + \frac{^{87}\text{Rb}}{^{86}\text{Sr}} [e^{\lambda_{87}T_0} - e^{\lambda_{87}(T_0-t)}]. \quad (1)$$

If the Rb/Sr ratio changes owing to Rb loss, the structure of Eq. (1) can be retained by introducing the effective ratio $(\text{Rb}/\text{Sr})_{\text{eff}}$:

$$\left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}}\right)_t = \left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}}\right)_0 + \left(\frac{^{87}\text{Rb}}{^{86}\text{Sr}}\right)_{\text{eff}} e^{\lambda_{87}(T_0-t)} (e^{\lambda_{87}t} - 1). \quad (2)$$

The exponent term $e^{\lambda_{87}(T_0-t)}$ is denoted as a_{87} .

At $t = 0.60$ Gyr and $T_0 = 4.568$ Ga, $a = 1.07$,

$$[\varphi_{\text{Rb}}(t)]_{\text{av}} = a_{87} \left(\frac{^{87}\text{Rb}}{^{86}\text{Sr}}\right)_{\text{eff}} = \frac{1}{t} \int_0^t \varphi_{\text{Rb}}(t) (dt). \quad (3)$$

where $\varphi_{\text{Rb}}(t)$ is the law of Rb/Sr variations in the initial gas-dust supraplanetary cloud.

If $(^{87}\text{Rb}/^{86}\text{Sr})_{0.60}$ is taken to be identical to $(^{87}\text{Sr}/^{86}\text{Sr})_{i\text{-Moon}}$, i.e., 0.69906, we obtain a value of 0.17. This parameter characterizes the integrated mean Rb/Sr ratio from the formation of the solar system ($^{87}\text{Rb}/^{86}\text{Sr} = 0.92$) to the moment of fragmentation (Moon separation). The $^{87}\text{Rb}/^{86}\text{Sr}$ value at the moment of fragmentation is unknown and can be calculated if the function $\varphi_{\text{Rb}}(t)$ is specified.

The exact form of this function is unknown. However, its approximate shape can be deduced from the rules of construction of averaging functions. There exists a set of curves compatible with the same mean value of 0.17. Figure 5 shows a curve quantitatively corresponding to the character of $\varphi_{\text{Rb}}(t)$ between 0 and 60 Ma. It corresponds to the flux of the volatile component (Rb) from the contracting cloud.

The $^{87}\text{Rb}/^{86}\text{Sr}$ value characteristic of the Moon itself (0.018) was produced by the catastrophic loss of Rb in a relatively short and high-temperature fragmentation event, which did not significantly affect $(^{87}\text{Sr}/^{86}\text{Sr})_{i\text{-Moon}}$ attained by that time.

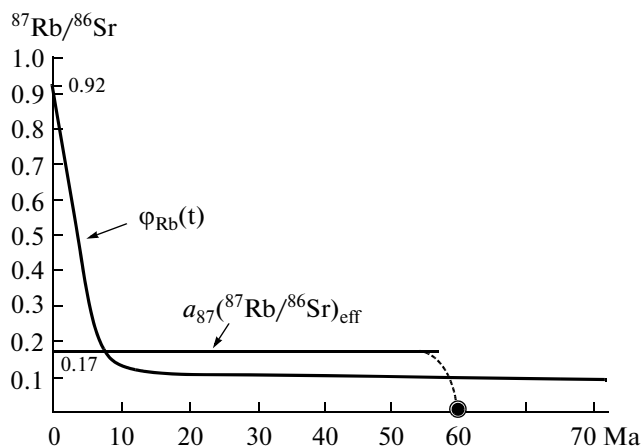


Fig. 5. Curve approximating the function $\varphi_{\text{Rb}}(t)$ at an initial value of 0.92 and a mean value of 0.17 between 0 and 60 Ma.

The application of the analysis of the Rb–Sr system to the case of a giant impact is strictly speaking impracticable, because the geochemical characteristics of the impactor are unknown.

^{129}I – ^{129}Xe and ^{244}Pu – ^{136}Xe

There is a problem in the isotopic geochemistry of xenon, which is sometimes referred to as the xenon paradox (Podosek and Ozima, 2000; Swindle and Podosek, 1988; Allegre et al., 1995). Paradoxically, the isotopic composition of terrestrial Xe indicates that the Earth had lost the Xe that was formed during the earliest 110–130 Ma, although it had to be retained by gravity.

The fractions of radiogenic ^{129}Xe and ^{136}Xe , which are produced by the decay of ^{129}I and ^{244}Pu , respectively, in atmospheric Xe are much lower than could be expected. The relatively short-lived ^{129}I and ^{244}Pu isotopes have half-lives of 17 and 81 Ma, respectively. It was shown (Podosek and Ozima, 2000) that, both for the ^{129}I – ^{129}Xe and ^{244}Pu – ^{136}Xe systems, the observed proportions of Xe isotopes correspond to the Xe that has been retained by the Earth since 110–130 Ma (Podosek and Ozima, 2000; Allegre et al., 2008).

Leaving aside some exotic hypotheses, the explanation of this fact is reduced to the suggestion that the Earth's primordial atmosphere that had accumulated owing to its degassing during accretion (this concerns collisional accumulation) was lost in a catastrophic event at ~110–130 Ma (Porcelli et al., 2001).

According to Halliday (2008), this event was the giant impact that produced the Moon. But this is contradicted at least by the existence of lunar rocks that formed earlier than 110 Ma after solar system formation (Table 2). If the Moon-forming giant impact occurred earlier, this excludes subsequent large impact

that could result in the loss of the atmosphere (Canap and Asphaug, 2001).

Within this concept, the loss of early Xe is a consequence of the removal of released volatiles in a hydrodynamic flow during the compression of the gas–dust body, which was discussed above. Then, the time ~120 Ma corresponds to the completion of hydrodynamic lifting. In other words, the boundary ~120 Ma is the closure age of the Xe system. This is the moment when the process of accretion terminated and the history of the Earth as a consolidated body started.

Note that it is known that terrestrial Xe is fractionated relative to meteoritic Xe (Podosek and Ozima, 2000). Such fractionation is possible owing to gravitational separation during hydrodynamic escape (Hunten et al., 1987). This is well consistent with the model discussed here. But such an effect is impossible at the instantaneous loss of the atmosphere.

U – Pb

Similar to Xe and Rb, approximately 97% of initial terrestrials Pb were lost. Primary (solar) Pb shows a $^{238}\text{U}/^{204}\text{Pb}$ ratio (designated as μ) of 0.27 (Anders and Grevesse, 1989), whereas terrestrial Pb has $^{238}\mu$ of 8–10 (Allegre et al., 1995).

Allegre et al. (2008) suggested that the loss of early Xe and the loss of early Pb were related and reflected a common major event during Earth differentiation at ~4.45 Ga (~117 Ma after solar system formation). In the case of Pb, its removal into the core in a sulfide form can be considered. It is supposed that the core segregation, which occurred then, was accompanied by tremendous energy release resulting in the formation of the magma ocean and the loss of the atmosphere.

The suggestion that Pb was removed to the core does not explain why the Moon is more depleted in Pb than the Earth, although its core is much smaller. Lead is in fact a lithophile rather than siderophile element, although it can show siderophile properties in a sulfide form at low redox potential. But its affinity to the metal phase is hardly sufficient to explain the 30-fold depletion in the Earth's mantle, higher than that of some typical siderophile elements, for instance, W and Co.

It is a reasonable suggestion that Pb has escaped, similar to Xe and Rb, in a hydrodynamic flow. Lead is one of the most volatile elements. Its depletion in the Earth and, to a greater extent, in the Moon is logically included in our model.

Let us assume that hydrodynamic lifting ceased and the loss of Pb terminated at t_* . Then, the following relation holds for the $^{238}\text{U}/^{206}\text{Pb}$ system:

$$\left(\frac{^{206}\text{Pb}}{^{204}\text{Pb}}\right)_{\text{present}} = \left(\frac{^{206}\text{Pb}}{^{204}\text{Pb}}\right)_{*} + \mu [e^{\lambda_{238}(4.568 - t_*)} - 1], \quad (4)$$

where $(^{206}\text{Pb}/^{204}\text{Pb})_*$ is the isotopic composition of Pb by the moment of the termination of Pb loss.

If the system were closed, the following relation would have held by t_* :

$$\left(\frac{^{206}\text{Pb}}{^{204}\text{Pb}}\right)_* = \left(\frac{^{206}\text{Pb}}{^{204}\text{Pb}}\right)_0 + {}^{238}\mu e^{\lambda_{238}(4.568-t_*)} (e^{\lambda_{238}t_*} - 1). \quad (5)$$

But, since hydrodynamic escape of volatile Pb occurs during the contraction stage, ${}^{238}\mu$ should be changed by some effective value, μ_{eff} . Let $\exp(\lambda_{238}(4.568 - t_*))$ be denoted as a_{238} . Then,

$$\left(\frac{^{206}\text{Pb}}{^{204}\text{Pb}}\right)_* = \left(\frac{^{206}\text{Pb}}{^{204}\text{Pb}}\right)_0 + \mu_{\text{eff}} a_{238} (e^{\lambda_{238}t_*} - 1). \quad (6)$$

By analogy with the Rb–Sr system, $\mu_{\text{eff}} \cdot a_{238}$ is defined as an integral mean value varying from $(^{238}\text{U}/^{204}\text{Pb})_0 = 0.27$ to $\mu \cdot a_{238}$:

$$\mu_{\text{eff}} \cdot a_{238} = \frac{1}{t_*} \int_0^{t_*} \varphi_{238}(t) dt. \quad (7)$$

$$18.279 = 9.307 + \mu_{\text{eff}} e^{0.155125(4.568-t_*)} (e^{0.155125t_*} - 1) + {}^{238}\mu [e^{0.155125(4.568-t_*)} - 1], \quad (11)$$

and

$$15.491 = 10.294 + \frac{\mu_{\text{eff}}}{137.88} e^{0.98485(4.568-t_*)} (e^{0.98485t_*} - 1) + \frac{{}^{238}\mu}{137.88} [e^{0.98485(4.568-t_*)} - 1]. \quad (12)$$

If the ${}^{238}\mu$ value were precisely known, μ_{eff} and t_* could have been determined from Eqs. (11) and (12). Given the estimated range of ${}^{238}\mu$ from 8 to 10, the time of accretion completion from the gas–particle medium is constrained between approximately 110 to 130 Ma.

At ${}^{238}\mu = 9$, we obtain $t_* = 120$ Ma, $\mu_{\text{eff}} = 0.7$, $a_{238} = 1.99$, and $a_{235} = 0.41$; correspondingly, the mean values of functions are $(\varphi_{238})_{\text{mean}} = 1.4$ and $(\varphi_{235})_{\text{mean}} = 0.29$.

Since, in contrast to the Rb–Sr system, the escaping component of the U–Pb system appears in the denominator, $\varphi_{\text{Pb}}(t)$ should be compared with the function $\varphi_{238}^{-1}(t)$.

Correspondingly, the scale of values for $^{204}\text{Pb}/^{238}\text{U}$ is the following: $(^{238}\text{U}/^{204}\text{Pb})_0^{-1} = 0.27^{-1} = 3.7$, $(\varphi_{238})_{\text{mean}}^{-1} = 0.71$, $({}^{238}\mu a_{238})^{-1} = 0.06$; and for

Similarly, for the ^{235}U – ^{207}Pb system:

$$\left(\frac{^{207}\text{Pb}}{^{204}\text{Pb}}\right)_{\text{present}} = \left(\frac{^{207}\text{Pb}}{^{204}\text{Pb}}\right)_* + \frac{{}^{238}\mu}{137.88} [e^{\lambda_{235}(4.568-t_*)} - 1] \quad (8)$$

and

$$\left(\frac{^{207}\text{Pb}}{^{204}\text{Pb}}\right)_* = \left(\frac{^{207}\text{Pb}}{^{204}\text{Pb}}\right)_0 + \frac{\mu_{\text{eff}}}{137.88} e^{\lambda_{235}(4.568-t_*)} (e^{\lambda_{235}t_*} - 1). \quad (9)$$

Let us denote $a_{235} = \frac{1}{137.88} e^{\lambda_{235}(4.568-t_*)}$, then

$$\mu_{\text{eff}} \cdot a_{235} = \frac{1}{t_*} \int_0^{t_*} \varphi_{238}(t) dt. \quad (10)$$

Expressing $(^{206}\text{Pb}/^{204}\text{Pb})_*$ in Eq. (4) through Eq. (6) and $(^{207}\text{Pb}/^{204}\text{Pb})_*$ in Eq. (8) through Eq. (9) and using the known values of $\lambda_{238} = 0.155125 \text{ Gyr}^{-1}$, $\lambda_{235} = 0.98485 \text{ Gyr}^{-1}$, $(^{206}\text{Pb}/^{204}\text{Pb})_{\text{present}} = 18.279$, $(^{207}\text{Pb}/^{204}\text{Pb})_{\text{present}} = 15.491$, and initial solar ratios (Canion Diablo meteorite) $(^{206}\text{Pb}/^{204}\text{Pb})_0 = 9.307$ and $(^{207}\text{Pb}/^{204}\text{Pb})_0 = 10.294$, we obtain

$^{204}\text{Pb}/^{235}\text{U}$: $(^{235}\text{U}/^{204}\text{Pb})^{-1} = 0.085^{-1} = 11.72$; $(\varphi_{238})_{\text{mean}}^{-1} = 2.43$, and $({}^{235}\mu a_{235})^{-1} = 0.19$. On the same scale, the shapes of the $(\varphi_{238})_{\text{mean}}^{-1}$ and $(\varphi_{235})_{\text{mean}}^{-1}$ curves are identical.

Figure 6 shows a curve constructed for the case of $\varphi_{238}^{-1}(t)$ in the same manner as the $\varphi_{\text{Rb}}(t)$ curve in Fig. 5. It illustrates the general evolution of the Pb–U system under the above parameters.

Given the time of the completion of Earth accretion (~120 Ma), the $^{87}\text{Rb}/^{86}\text{Sr}$ evolution line can be continued up to the moment when $\varphi_{\text{Rb}}(t)$ will correspond to the $^{87}\text{Rb}/^{86}\text{Sr}$ value characteristic of the Earth, i.e., 0.09.

The loss of Rb, Pb, and Xe, if it is related to hydrodynamic escape, must occur in a similar manner; i.e., the $\varphi_{\text{Rb/Sr}}(t)$ and $\varphi_{\text{Pb/U}}^{-1}(t)$ functions must be more or

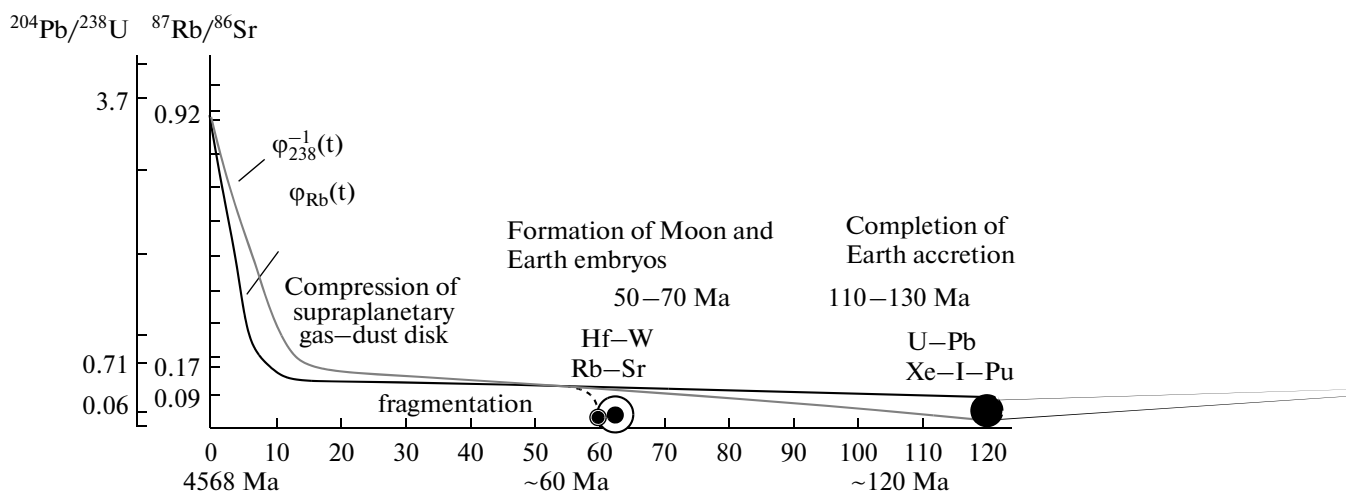


Fig. 6. Evolution of the prototerrestrial supraplanetary gas-dust body estimated from variations in the Rb/Sr and Pb/U ratios.

less conformable. To a first approximation, such a similarity is observed.

Thus, the time of the completion of the evolution of the gas-dust body from which the Earth and Moon were formed is estimated as 110–130 Ma, the most probable value being ~120 Ma. According to the Hf–W and Rb–Sr isotopic systematics, the fragmentation and the separation of the Moon occurred 50–70 Ma after solar system formation with the most probable estimate of ~60 Ma.

CONCLUSIONS

A hypothesis was proposed for the formation of the Earth and the Moon from a large gas-dust cloud, the size of which is limited by the Hill radius, i.e., approximately 1 million kilometers. The compression of this supraplanetary gas-dust body resulted in an adiabatic temperature increase in its interior parts and evaporation of volatile components, including iron, from the surface of particles. At a certain stage, within 50–70 Ma after the beginning of solar system formation, the supraplanetary gas-dust disk underwent fragmentation, the Moon was separated, and the embryo Earth was formed. The remaining part of the gas-dust material was accreted mainly to the Earth. During this process, gas dominated by primary hydrogen was squeezed out of the disk. Vapor was expelled from the interparticle space together with hydrogen. Hydrodynamic lifting resulted in the loss of volatiles, including Rb, Xe, and Pb, which is reflected in the Rb–Sr, Xe–I–Pu, and U–Pb isotopic systems. Gas-dust accretion was completed at 110–130 Ma (most likely, ~120 Ma) after the beginning of solar system formation. This moment was marked by the cessation of hydrodynamic lifting and volatile loss and the beginning of the history of the Earth as a condensed body.

What problems does the hypothesis solve, and what facts can it explain?

The model proposed here is alternative to the giant impact hypothesis and overcomes most of its major difficulties. The existence of a common source for the Moon and the Earth solves the problem of the coincidence of the ^{16}O – ^{17}O – ^{18}O fractionation trends for the Earth and the Moon. The identical tungsten isotopic compositions ($^{182}\text{W}/^{184}\text{W}$) of the Moon and Earth are also natural, whereas this fact dramatically complicates the interpretation of Hf–W isotopic systematics in the giant impact hypothesis. The proposed hypothesis explains the depletion of the Moon in Fe. It was shown that the loss of Fe is correlated with the enrichment of the Moon in refractory elements (Al, Ca, and Ti). Isotopic exchange in the vapor–particle system explains why volatile loss from the Moon was not accompanied by isotope fractionation. This fact is difficult to account for in the giant impact hypothesis. The dependence of the probability of fragmentation on the angular momentum of the gas-dust disk explains why the Earth has a satellite, whereas Venus does not.

The origin of a planet from a gas-dust cloud implies long-term existence of an extended envelope consisting of particles and gas. The hydrodynamic regime imposed by an ascending hydrogen flow (hydrodynamic lifting) solves the problem of the loss of a range of volatile elements, including Rb, Xe, and Pb, from the Earth during the earliest 100–130 Ma. This phenomenon was attributed to various processes: catastrophic impact-related removal of the primordial atmosphere (in the case of Xe), Pb migration into the core, etc. However, each of these explanations contradicted some observations.

The proposed hypothesis gives rise to the concept of core formation by segregation from a vapor–gas

state, i.e., by a mechanism imitating heterogeneous accretion. This provides a solution for the long-standing problem of the formation of a metallic core from the initial FeO state, which requires the removal of the tremendous equivalent amount of oxygen from the Earth.

The scenario of core formation considered above is consistent with the fundamentally nonequilibrium distribution of siderophile elements.

The Hf–W, Rb–Sr, Xe–I–Pu, and U–Pb isotopic systems provide constraints for the main events of the evolution of the supraplanetary prototerrestrial gas–dust disk: fragmentation and Moon formation occurred ~60 Ma, and gas–dust accretion completed ~120 Ma after the beginning of solar system formation, after which the history of the condensed Earth started.

What is yet not fully understood?

The real mechanism of the evolution of the protosolar nebula resulting in the formation and prolonged (~10⁸ yr) development of large-scale supraplanetary gas–dust bodies. There are only some tentative approaches, and 3D computer modeling is needed.

A more sophisticated model should be constructed for processes in a compressed and fragmenting supraplanetary disk, including the estimation of spatiotemporal variations in temperature and pressure. This will allow us to quantify the models that are as yet only qualitative.

The main conclusion is that, if the proposed hypothesis is correct in principle, the mechanism of planetary formation and many important geochemical relationships should be fundamentally revised.

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