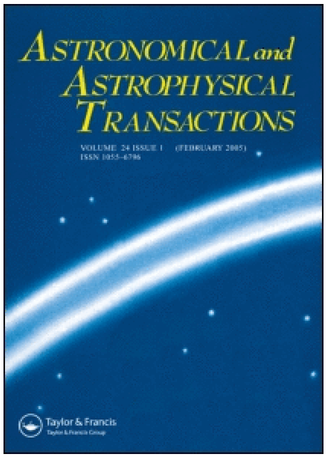

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EARLY STAGES IN THE DEVELOPMENT OF THE EARTH–MOON SYSTEM

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A possible mechanism of formation of the Moon and of its influence on the development of the Earth is considered. Some alternative hypotheses of formation of the Earth and of the terrestrial core are discussed. It is shown that the appearance of a massive satellite (the Moon) in a close near-Earth orbit span the Earth up and considerably accelerated its tectonic development. In addition to this, it is shown that all the hypotheses supposing an early melting of the Earth, which was accompanied by formation of the Earth's core and early degassing of the Earth, meet obvious and insuperable inconsistencies with numerous geological and geochemical data. The only model that satisfies all geological and geochemical data to the greatest extent is the Schmidt–Safronov model of the planets' formation. According to this model the young Earth had a homogeneous composition and was a rather cool and, therefore, tectonically passive planet. For the first time, melting of the Earth's interiors at the level of the upper mantle took place under the influence of lunar tides and decay of radioactive elements only 600 millions years after the formation of the Earth. This moment is precisely fixed by the beginning of basaltic magmatism on the Moon about 4 Ga ago and appearance of the most ancient rocks of the Earth's crust about 3.8 Ga ago. Since the same time, the density differentiation of the Earth's substance began with the release of iron and its oxides, which subsequently formed the Earth's core. However, the isolation of the Earth's core took place only at the very end of the Archaean, about 2.6 Ga ago, though the process of formation of the Earth's core is continuing.

KEY WORDS Origin of the Earth and Moon, tidal evolution of the Earth and Moon, Roche limit, origin of the Earth's core, early Precambrian, evolution of the lead ratio

1 INTRODUCTION

Models of formation of the Solar-system planets and, especially, of the Earth are important not only as world outlook, but also as starting points and basis in elaboration of the most general geological concepts of the Earth's development. On the other hand, geological and, at present, cosmogeological data allow us not only to check, but, if necessary, to improve models of origin of the Earth and planets of the terrestrial group.

From these points of view, elaboration and presentation of the ideas of O. Yu. Schmidt (1948) and his followers, especially V. S. Safronov (1969), were extremely fruitful. The theory of origin of the planets was presented most comprehensively and consistently in the monograph of V. S. Safronov 'Evolution of the Protoplanetary Nebula and Formation of the Earth and Planets' (1969). Note that the fundamental Schmidt–Safronov model of the origin of the planets fully satisfies modern geological and geochemical data on the Earth's development. Moreover, these data just confirm the adequacy of the model. At the same time, attempts of a further 'improvement' of the theory of the origin of the planets especially those aspiring to substantiate the early differentiation of the Earth (see, e.g., Vityazev *et al.*, 1990, Vityazev and Pechernikova 1996, and other works), resulted only in obvious inconsistencies with geological (and not only geological) data.

What are the main conclusions of the theory of the formation of the planets that we are considering? Firstly it follows that the young Earth was a rather cool planet and the temperature in its interiors did not rise anywhere above the melting temperature of the Earth's substance. Secondly, owing to homogeneous accretion of planets, the composition of the primary-Earth's substance was practically uniform throughout its volume and, hence, the Earth did not yet have a dense core.

These simple conclusions of the theory of origin of the planets have implications that are very important for geology; they can be basically tested with the geological annals of the Earth. In particular, it follows hence that the young Earth was a tectonically passive planet and was characterised by complete absence of magmatic activity. This conclusion of theory is easily checked by the ages of the most ancient magmatic rocks of the Earth. According to geological data, the first melts in the Earth's history appeared at the beginning of the early Archaean, i.e., only 600–800 Ma after the Earth's formation. The reliably established age of the most ancient igneous rocks of the Earth's crust is 3.8 Ga (Moorbath, 1980; Barton, 1981; Ray, 1983). It should be also taken into account that the crust rocks are products of chemical-density differentiation of the terrestrial substance, and, therefore, they are always lighter than the primary substance of the mantle. For this reason, these rocks, once risen to the Earth's surface, cannot be completely destroyed by subsequent tectonic processes or dive into the mantle (this is hindered by the Archimedean principle). Hence it appears that indeed no magmatic processes took place in the Earth's interiors during the long time interval after formation of the Earth, i.e., throughout the Catarchaean, as follows from the Schmidt–Safronov theory. According to the estimates of Safronov (1969) himself, such a latent phase could last for about 1 Ga.

The second major conclusion from the theory of planetary origin considered for geology suggests that the young Earth did not have a dense core. We will discuss the problem of isolation of the Earth's core in detail below. Here we note only that this is indicated, in particular, by paleomagnetic data. These data show that the dipole magnetic field of the present-day type, which is usually ascribed to convection currents in the Earth's core, appeared on the Earth only about 2.6 Ga ago (Hale, 1987), i.e., approximately 2 Ga after the formation of our planet (see Figure 1).

In this connection, it is obvious that, for a correct understanding of basic regularities of the Earth's development, it is of primary importance to establish the beginning of its tectonic activity. Analysing the reasons of the origin of the Earth's primary tectonomagmatic activation, V. E. Khain (1977) noted that the primary

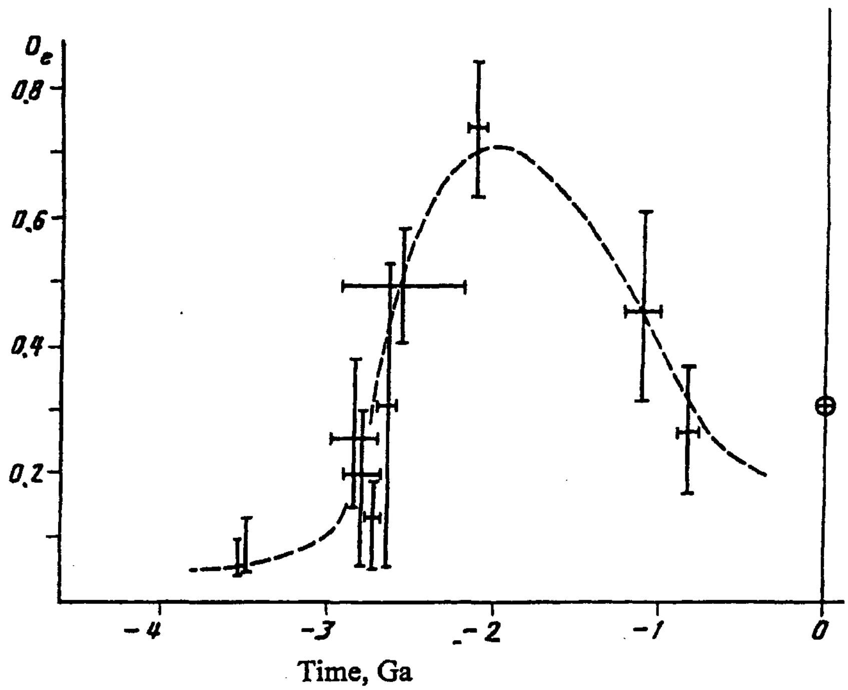

Figure 1 Variation of the magnetic-field intensity of the Earth according to paleomagnetic data (Hale, 1987). Horizontal and vertical lines show the confidence intervals of measurements; the crossed circle marks the intensity of the present-day geomagnetic field.

granitization on the Earth about 4 Ga ago was probably stimulated by causes exterior to it, because at the same time major tectonic events happened on the Moon, too. Seemingly, this was indeed the case. Therefore, the moment of the onset of the Earth's tectonomagmatic activation can be further refined based only on the age of the lunar rocks(!). However, for this purpose it is first necessary to at least briefly review a possible mechanism of the Moon's formation and evolution of tidal interactions in the Earth–Moon system, which were presented in more detail by Sorokhtin (1988), Sorokhtin and Ushakov (1989,1991).

2 FORMATION OF THE MOON AND ROTATION OF THE EARTH

The Earth and the Moon are connected by tidal interactions. At present, the Earth's proper rotation is appreciably braked by the tidal interaction with the Moon, and the Moon itself is moving away from our planet. Hence it follows that in the past geological epochs the Moon was much closer to the Earth, especially at the earliest stages of development of this system, and the tidal interaction that arose between them could reach an extreme intensity. Nowadays, the most popular models of the Moon's origin assume its formation in a giant grazing impact (megaimpact) on the Protoearth of a certain planet of a Martian size (Hartmann and Davis, 1975; Cameron and Ward, 1976). In subsequent works, it was supposed that a single megaimpact or a multitude of small 'macroimpacts' formed a disc of debris around the Earth which had arisen from these impacts, whereas the Moon formed at the expense of the further accretion of this disc (Cameron and Benz, 1991; Cameron, 1997; Ida *et al.*, 1997). A direct modelling of gravity interactions of many bodies in a heterogeneous protoplanetary disc around a growing planet was completed by Ohtsuki and Ida (1998). They showed that the gravity captures, especially of small-size cosmic bodies (which, however, always end in grazing impacts of planetesimals against the surface of the planet) are quite plausible. They usually happen in the same (prograde) direction and represent the principal cause of planets' spinup. Unfortunately, in the numerical experiment described there was no modelling of destruction of the captured massive bodies at the Roche limit. Therefore, the authors also connect the origin of the Moon with either a grazing impact of a larger protoplanetary body, or accretion of the Protoearth debris disc.

What peculiarities of celestial mechanics, structure, composition and geochemistry should be taken into account by a modem hypothesis of the Moon formation?

First of all, it should take into account that the total angular momentum of the Earth–Moon system to a good accuracy fits the situation where the Moon once could be at the Roche limit, and the Earth was at the same time rotating around its own axis with the angular velocity of a satellite at the same Roche limit and committed one revolution per approximately 6 hours. We will show below that the majority of other planets of the Solar system (Jupiter, Saturn, Uranus and Neptune) also rotate around their axes with similar velocities, close to orbital velocities of satellites at their Roche limits. The Roche limit (in units of the distance between the centres of gravity of the planet and satellite) L_R is determined by the following simple expression (Alfvén and Arrhenius, 1979)

$$L_R \approx 2.44 \left(\frac{\bar{\rho}_0}{\bar{\rho}_s}\right)^{1/3} R_0, \tag{1}$$

where R_0 is the planet's radius; $\bar{\rho}_0$ and $\bar{\rho}_s$, are the average densities of the planet and satellite.

Secondly, the hypothesis should take into account that average density of the Moon, $\bar{\rho}_L = 3.34$ g cm^{-3} is anomalously low; furthermore, the Moon has no large and dense core. Certainly, this testifies to a loss of the most part of iron from the primary lunar substance. According to the estimates of geochemists, the Moon is depleted with iron by a factor of approximately four in comparison with the Earth (Henderson, 1982). If as in the case of the Earth, such loss of iron had not taken place, the average density of the Moon, in view of its smaller compression, would be approximately 4.1 g cm^{-3}. Together with iron, the Moon lost the most part of siderophile and chalcophile elements (Ringwood, 1979; Henderson, 1982) and nearly all of its primary lead (Mason and Melson, 1970). As a result, the ratios of the lead isotopes in lunar rocks are anomalously high. Such extreme values of the lead isotopic ratios cannot be accounted for by any 'megaimpacts' or debris of 'macroimpacts' thrown out to the near-Earth space. It is a very strong argument against the 'megaimpact' hypotheses of the Moon's origin, and it cannot be neglected.

Finally, it should also be noted that the lunar crust thickness reaches 80 km and consists of anorthosites, rocks that represent the extreme term of ultrabasic and basaltic magma differentiation. The age of anorthosites, about 4.6–4.2 Ga, corresponds to the epoch of the Solar-system planets' formation. Except for continental anorthosites, basalts only are widespread on the Moon, which have flown out in lunar maria. The lunar basalts are somewhat enriched with titanium, but in general their chemical composition is rather close to tholeiitic basalts of the rift zones of the Earth (Ringwood, 1979). It is worth mentioning that the age of the most ancient lunar basalts, approximately 4.0–3.8 Ga, closely coincides with the age of the most ancient rocks of the Earth's crust, 3.8 Ga (Khain, 1977, 1994).

An important conclusion follows from the above data: at the moment of its formation, the Moon was overheated and completely molten. As a result, already at the very beginning of its life, it has passed full differentiation. In addition to this, at the same time its parent body, which we have termed Protomoon, should be at the Roche limit and should be destroyed by tidal forces from the Earth. Otherwise, it is simply impossible to explain the simultaneous loss by the lunar substance of the greater part of iron, siderophile and chalcophile elements, with simultaneous full conservation of all the lithophile elements in the composition of its rocks.

Hence it is concluded that the Protomoon could still be captured by the gravitational field of the growing Earth from the neighbouring (nearest) orbit of the protoplanetary disc. The probability of such capture under the action of tidal forces is justified at the qualitative level by Singer (1970), Alfvén and Arrhenius (1972, 1979). Unfortunately, the exact solution of this problem has not been found until now. Around practically all exterior planets there exist systems with a multitude of satellites that are usually rotating in the equatorial planes of their planets (one of which, Neptune, is rotating 'lying on its side' with respect to the ecliptic), and, in our opinion, this rather convincingly tells us about gravity captures that happened in the past. Moreover, it is otherwise impossible to explain the retrograde rotation of the satellites Phoebe in the system of Saturn and Triton in the system of Neptune, and also recent impact on the Jupiter of the Comet Shoemaker–Levy. Most probably, at the stage of formation of the planets when their masses did not remain constant, but were increasing with time, the gravity capture of satellites from the neighbouring orbits was not only possible, but had a mass character.

Taking into account what was said, we have proposed a hypothesis, according to which, during the Earth growth, its gravity field captured from a neighbouring orbit a small planet, the Protomoon (Sorokhtin, 1988). In the course of this capture, under the influence of tidal deformations, which accompanied the process of extinguishing the eccentricity of the initial orbit of the Protomoon, it was overheated (by approximately 2000–3000°C), melted, and passed a complete density differentiation. Because of tidal interactions with the Earth, the liquid Protomoon approached the Roche limit, began to plunge into it and to gradually decay (to spill over to the

Earth), thus spinning the central planet up. The Moon itself formed from the remnants of the silicate substance of the exterior tidal hump of the Protomoon, whereas the substance of the interior tidal hump and its liquid ferrous core, which got into the Earth's Roche sphere, precipitated onto the Earth's surface. The process of the Protomoon capture and destruction, should finally spin the Earth up in the prograde direction (in the direction of the Moon's revolution) up to the limiting rate of the satellite rotation at the Roche limit (for the Earth, this rate is approximately one revolution per 6 hours). After this moment, the Moon, together with other satellites (then still abundant in the near-Earth space) under the influence of tidal perturbations began only to be repelled and to move away from the Earth (more detailed calculations of this hypothesis were presented by Sorokhtin and Ushakov, 1989, 1991).

As already noted, one of the main testimonies in favour of the above-mentioned hypothesis of the origin of the Moon is the quantitative calculation of the planetary-mass dependence of the specific momentum of axial rotation of the central planet, spun up to the angular velocity of the satellite rotation. This dependence is derived under the assumption that the planet was spun up at the expense of its tidal interactions with the satellites and because of precipitation of some part of the satellites' substance from the Roche limit onto the planet

$$A_R = JR_0^2\Omega = J\left(\frac{\gamma}{2.44^3(\rho_0/\rho_s)(4\pi\rho_0/3)^{1/3}}\right)^{1/2} M^{2/3}, \qquad (2)$$

where J is the planet's dimensionless moment of inertia (for the modern Earth $J \approx 0.33$, for the young and yet not differentiated Earth $J \approx 0.37$); R_0 is its radius; Ω is the angular velocity of its own axial rotation; γ is the gravity constant; ρ_0 and ρ_s are the average density of the central planet and of the satellite destroyed at the Roche limit, respectively; M is the mass of the planet. If we now mentally transfer the Moon from its modem orbit (384,400 km) to that of the Roche limit (about 17,000 km) with conservation of a constant momentum in the Earth–Moon system, the specific rotational momentum of the Earth will exactly fit expression (2). Such a coincidence can hardly be accidental.

It is interesting to note that our hypothesis about transfer of rotational momentum from satellites to central planets has correctly explained the known empirical dependence $A_R \sim M^{2/3}$ (Fish, 1967; Hartmann and Larson, 1967) and, moreover, has correctly estimated it quantitatively (see Figure 2). It should only be kept in mind that the derived expression for specific momentum $A_R = JR_0^2\Omega$ determines its greatest possible value. Under actual conditions for planets that were spun up in the prograde direction, condition $0 \le A(\Omega) \le A(\Omega)_R$ should be satisfied.

If we take into account that, according to our hypothesis, the substance of the Protomoon's core, essentially enriched with iron, has fallen onto the Earth surface, we can accept $\rho_s \approx 4.4$ g cm^{-3}. The average density of the Earth during its development experienced but minor variations; therefore, ρ_0 is ≈ 5.52 g cm^{-3}, as before. For the young Earth deprived of a core, $J \approx 0.37$; hence we find $W \approx 1.33 \times 10^{-5} c^{2/3}$. For other planets, the specific moment of inertia can be readily

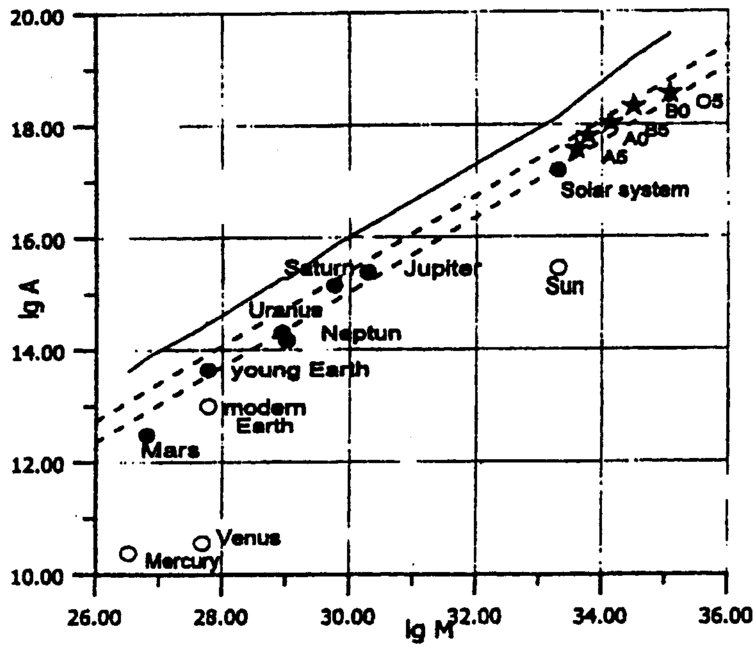

Figure 2 Specific momentum of axial rotation of the Solar system planets and some fast-rotating stars *versus* their masses. The dashed lines show the limiting values of momentum A in function of the density of the satellites that have fallen onto them from the Roche limit. The solid line shows the limit of rotational stability of the planets and stars.

found from empirical data, presented, for example, by Marov (1986). Note that our hypothesis has correctly determined also the constants of proportionality in the law considered. Depending on the average density and degree of differentiation of the planets, this factor varies from 1.33×10^{-5} cm^2 s^{-1} g$^{-2/3}$ for the young Earth to 1.75×10^{-5} cm^2 s^{-1} g$^{-2/3}$ for Saturn. At the same time, according to the empirical dependence of Fish–Hartmann–Larson, this factor is on the average 1.6×10^{-5} cm^2 s^{-1} g$^{-2/3}$.

The results of calculation of the specific moment of axial rotation of the Earth, fast rotating of planets, stars and Solar system are presented in Figure 2. A comparison shows that for all these celestial bodies, whose masses cover seven orders of magnitude, theoretical dependence (2), derived from the hypothesis accepted, indeed approximates fairly well the angular velocities of their rotation. Such a coincidence, too, can hardly be accidental. Most likely, it testifies to a frequent occurrence of the gravity capture of satellites by more massive planets and consequent spinup of such planets by tidal interactions with their satellites and shedding of the satellites' substance onto their central planets. Anyway, at the stages of planetary systems' formation, the probability of such events could be rather high.

However, this mechanism is not universal, because there are planets that obviously do not fit into this regularity, for example Venus possessing retrograde rotation. Probably this was connected with Venus' formation close to the Sun; this hindered the acquisition of satellites, its rotation is therefore due only to accretion of protoplanetary substance and turned out to be retrograde. We cannot tell anything definite about the past rotation of Mercury. Most likely, it came to the resonance rotation from the very moment of its formation.

Mars, most probably because of its small mass in the epoch of the planets' formation, did not get a sufficient number of satellites and consequently was not spun up to the limiting velocity. For the same reason, the closest satellite of Mars, Phobos, is now not moving away, on the contrary, it is approaching Mars and even has already plunged into the Roche sphere of Mars. In future Phobos will inevitably fall onto the Martian surface (judging from the system of furrows on the satellite's surface, recalling tracks of moving snow avalanches in the mountains, the process of destruction of this satellite has already begun).

3 THE BEGINNING OF THE EARTH'S TECTONIC ACTIVITY IS MARKED ON THE MOON'S SURFACE

After the destruction of the Protomoon at the Roche limit, all further evolution of the Earth–Moon system was connected with the tidal braking of the Earth and with the Moon's recession from it. However, it is known (Ruskol, 1975) that velocity of this recession $\dot{L}$, other things being equal, is inversely proportional to distance between planets L to power 5.5 and is directly proportional to the value of dissipation function of the central planet Q_μ^{-1}, where Q_μ is the factor of its tidal anelasticity.

$$\dot{L} = 3\gamma^{1/2} k_2 \frac{m(M+m)^{1/2}}{MQL^{5.5}} R_0^5. \tag{3}$$

Here $k_2 = 0.301$ is the Love number, determined by the ratio of the potential originating from the planet's tidal strains to the tide-forming potential on the planet's surface; m is the satellite's mass; M is the planet's mass.

The initial location of the Moon was very close to the Earth (about 17,000 km between the centres of the planet and satellite). The factor of anelasticity (Q-factor) of the cool primary Earth was extremely high: $Q_\mu \approx 5000$ and because of the generation of tidal and radiogenic energy, by the end of the Catarchaean the Q-factor had gradually decreased to approximately $Q_\mu \approx 1000$. Exactly at the beginning of the Archaean, about 4 Ga ago, the tectonic activity of the Earth began, the first melts appeared, and degassing of the mantle started. After the formation of the first shallow sea basins at this time, the Q-factor of the Earth sharply decreased (according to our calculations, to 17 in the early Archaean). Then, as the water mass in the World Ocean was accumulating, the Q-factor again began to proportionally increase, approximately to 100 at the end of the Proterozoic. In the Phanerozoic, after the beginning of the sea transgressions and mass formation

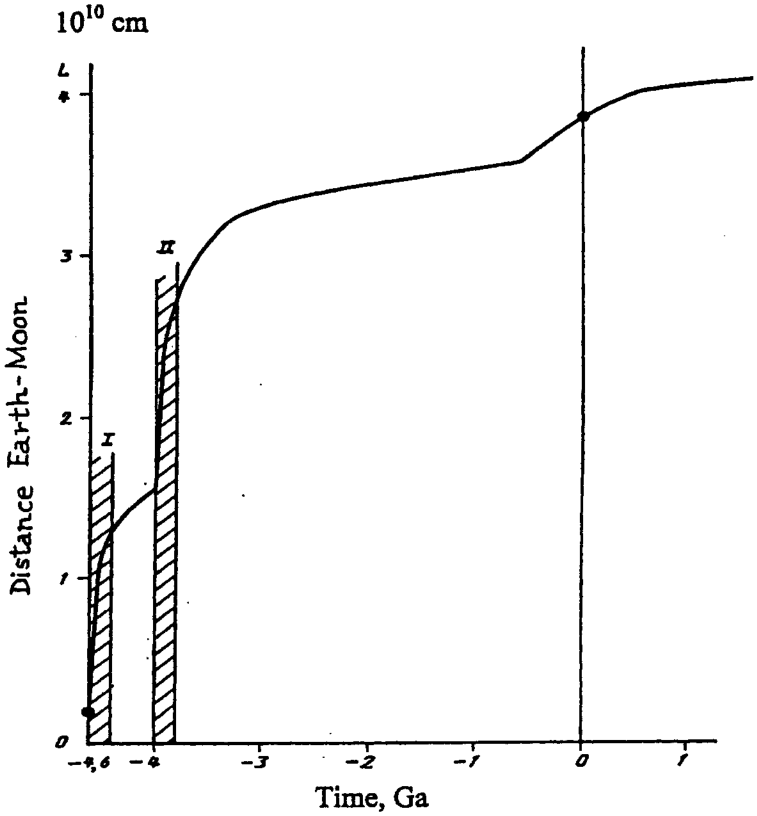

Figure 3 Evolution of the Earth–Moon distance, caused by variations of the anelasticity factor Q_μ of the Earth. Epochs of the lunar magmatism: (I) anorthositic; (II) basaltic.

of shallow epicontinental seas, the Q-factor again decreased to the approximately present-day level of 13. In this connection, it infers that in the history of tidal interaction of our planets there were three episodes of an accelerated recession of the Moon from the Earth (Figure 3). For the first time, it was right after the formation of the Moon about 4 Ga ago (this pulse of the accelerated recession of the Moon was connected with its very small distance from the Earth). For the second and third time, this took place after a sharp decrease of the Q-factor owing to formation of shallow seas on the Earth's surface. However, the rate of recession of the Moon from the Earth in the Phanerozoic was already insignificant, because by this time the tidal interaction of planets had essentially decreased (owing to a considerable increase of the distance between them).

The velocities of the tidal repulsion of the satellites from the central planet are always proportional to masses of the satellites (3). This circumstance resulted in 'sweeping-out' of smaller cosmic bodies from the near-Earth space by the large satellites; the small bodies inevitably fell on the surfaces of their more massive neighbours (during approaches and due to intersections of their orbits). It is natural that the Moon, being the most massive satellite, 'swept out' all the contents of the near-Earth satellite swarm more efficiently than the objects of the swarm. Such 'sweeping' was especially active during the phases of the fastest recession of the Moon from the Earth at the beginning of the system development in the Catarchaean and in the early Archaean.

Simultaneously with the Moon, the orbits of the rest bodies of the near-Earth satellite swarm were extending, too, at rates proportional to their masses; thus, by the time of collision of such bodies with the Moon, the more massive satellites had time to be removed to larger distances from the Earth than the small satellites. Therefore, initially (i.e., in the Catarchaean) collisions of the Moon with small- or medium-mass bodies should happen. At the second stage of the Moon's recession from the Earth (i.e., in the early Archaean), such impacts already happened to the largest and most massive of the remaining satellites (which also had had time to grow by absorption of small bodies of the near-Earth satellite swarm). Let us see now, the resulting consequences for the Moon of such 'bombardments' of its surface.

In the epoch considered, the Moon had not yet had time to cool down completely and under its lithosphere there was a 'magmatic ocean' with density-stratified melts: light anorthositic magmas at the top and basaltic melts at the bottom. Therefore, at the beginning, when the thickness of the lithosphere did not yet exceed 10–20 km, the 'bombardment' of the lunar surface resulted in effusions of only anorthositic melts.

Later, 600–800 Ma after that, i.e., at the very beginning of the Archaean, the thickness of the lunar lithosphere already increased to 100–120 km due to the cooling of the planet. Only large satellites with diameters from several tens to hundreds of kilometres could break through such a lithosphere. But at this particular time, larger satellites should also fall onto the Moon, piercing its lithosphere and thus opening paths to effusions of basaltic magmas.

According to the isotopic age determinations of lunar rocks, brought to the Earth during the *Apollo* program, the first pulse of the lunar magmatism was indeed characterised only by an anorthositic composition and was taking place between 4.6 and 4.4 Ga ago. The second pulse of the Moon magmatism, this time having a basaltic composition, began about 4.0 Ga ago and was proceeding to 3.6–3.2 Ga ago (Jessberger *et al.*, 1974a, b; Tera and Wasserburg, 1974, 1975; Tera *et al.*, 1974). Traces of such large holes, filled with basaltic effusions, are still visible today on the Moon's surface as so-called 'lunar maria'.

Thus, the beginning of the Moon's basaltic magmatism about 4 Ga ago marks very precisely the moment of a sharp decrease of the Earth's anelasticity factor, i.e., the occurrence of the first melts in the Earth's interiors, beginning of the Earth tectonomagmatic activity and formation of the first shallow seas. Therefore, we can state with a high degree of probability that in the first 600 Ma of its existence the Earth really remained a tectonically passive planet. However, the Moon had not only fixed the beginning of the tectonomagmatic activity of the Earth, but, as it will be shown below, considerably hastened its onset.

4 INFLUENCE OF THE MOON ON THE DEVELOPMENT OF THE EARTH

For the clear reasons, the temperature distribution in the primary Earth can be estimated only theoretically, proceeding from the ideas available about the formation of the Solar system planets and the above-mentioned data when the Earth was in a tectonically passive state. For the considered model of the Earth's formation (due to accretion of a cool dusty protoplanetary nebula), but disregarding its latent phase of development, such calculations were done by V. S. Safronov (1969). Let us recall that, according to this model, a major part of the thermal energy of the growing Earth was generated in its interiors because of transformation of the kinetic energy of planetesimals falling on the Earth's surface to heat. In spite of the huge energy of the accretion, it was released mainly in the near-surface parts of the forming planet; therefore, the heat generated in its upper layers was readily lost with the thermal radiation of the growing planet. Owing to the fact that, simultaneously with the Earth's growth, the sizes of the planetesimals falling on it also increased, the temperature in the interiors of the young Earth was increasing from the centre to the periphery. However, near the Earth's surface it again was decreasing because of the faster rate of cooling of parts near the surface. Figure 4 shows the theoretical temperature distribution in the young Earth, compared with the distributions for other geological epochs. According to Safronov's calculations, at that time the Earth's interiors were everywhere cooler than the melting temperature of the Earth's substance. The temperature of the young Earth peaked to 1500–1600 K at depths of about 800–1000 km, and toward the planet's centre it again fell to approximately 800 K.

The temperature dependence presented in Figure 4 may correctly reflect the general character of the temperature distribution in the primary Earth. However, when using it, it should be kept in mind that the temperature estimates presented are rather approximate, because the solution of this problem essentially depends on several model parameters, which are difficult to determine. The same figure shows the curve of the Earth's substance melting taken from Zharkov (1983), extrapolation of melting temperatures of the $Fe \cdot FeO$ eutectic alloy ('core substance') and the temperature of the present-day Earth (Sorokhtin and Ushakov, 1991).

A comparison of these curves implies, in particular, that for melting and differentiation of the young Earth (as postulated in some hypotheses of the planets' origin) its overheating by several thousand degrees is required, even in comparison with its present-day heat regime. However, this is completely unrealistic, because such overheating of the Earth would leave indelible tracks in its geological annals, but these tracks are simply absent.

According to our estimates (Monin *et al.*, 1987; Sorokhtin and Ushakov, 1991; and this work), during the Catarchaean latent stage of the Earth development (i.e., from 4.6 to 4.0 Ga ago), in its interiors about 2.08×10^{37} erg of tidal energy and 1.24×10^{37} erg of radiogenic energy, in total 3.32×10^{37} erg, were generated. All this energy was spent to heat the Earth. Owing to this, the Earth on the average was additionally heated by approximately 450°C. However, the greatest heating took

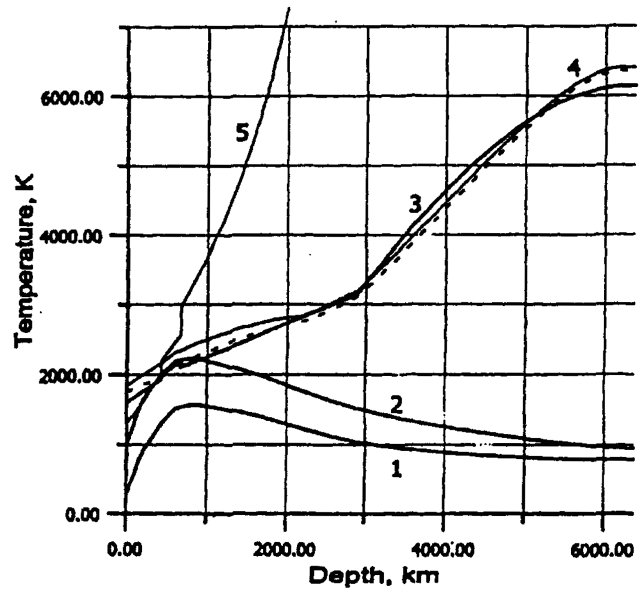

Figure 4 Temperature distributions in the Earth: (1) at the moment of its formation 4.6 Ga ago (Safronov, 1969); (2) at the Catarchaean–Archaean boundary about 4 Ga ago (Sorokhtin and Ushakov, 1989); (3) at the present epoch (Sorokhtin and Ushakov, 1991); (4) the melting temperature of Fe at high pressures (Boehler, 1993, 1996); (5) melting temperature of the mantle substance. Dotted line: extrapolation of the experimental data on melting of the Fe·FeO eutectic alloys for the conditions of the lower mantle and Earth's core at $p \gg 200$ Kbar (Ohtani and Ringwood, 1984; Ohtani *et al.*, 1984; Boehler, 1993).

place at the bottom of the upper mantle of the equatorial belt of the Earth, where the tidal strains reached the greatest amplitude. Here the temperature of the Earth rose by approximately 1000°C and about 4 Ga ago reached the temperature of the Earth's substance melting (see Figure 4); in fact, after this the tectonomagmatic activity of our planet began.

The theoretical temperature distribution in the young Earth immediately after its formation is presented by Safronov (1969) (see Figure 4). Hence we can determine the initial heat reserve of the young Earth: it is approximately (7–8) $\times 10^{37}$ erg (including the tidal energy, released during the gravity capture of the Protomoon, $\sim 1.3 \times 10^{37}$ erg). For a comparison we note that the total heat content of the present-day Earth is 14.94×10^{37} erg. Assuming that at the very beginning of the Archaean the temperature distribution still remained similar to the primary one, we can now determine (using the criterion of contact of the geotherm with the melting curve of the Earth's substance, see Figure 4), to determine the total heat content of the Earth at the beginning of the Archaean. We found it to be 10.58×10^{37} erg. With this, we can readily update the heat reserve of the primary Earth at the moment of its formation, it is $10.58 \times 10^{37} - 3.18 \times 10^{37} = 7.4 \times 10^{37}$ erg, close to the theoretical value (Safronov, 1969). This once again proves the theory of the 'cool' planets' origin considered here.

Thus, the origin of a massive satellite, the Moon, in a close Earth's orbit not only span the Earth up in the prograde direction, but also essentially accelerated its tectonic development. If our planet did not have a massive satellite, the Earth, like Venus, would be delayed in its tectonic development by 2.5–3 Ga. In this case, the conditions of the Archaean with a dense carbon-dioxide atmosphere (with a pressure of up to 9–10 atm) and high temperatures (up to 90–100°C) would dominate now on the Earth, and, instead of the modem highly-organised life, the Earth would be populated only by primitive single-cell prokaryotes of the archiabacteria type.

5 TIME OF THE EARTH'S CORE FORMATION

The second important consequence following from the considered theory of the planets' origin: the young Earth did not have a dense core. Now it exists. Therefore, it is very important to find out when the Earth's dense core appeared and how it formed, at one time or gradually. This problem is closely connected to the heat regime of the Earth's formation. In many, even serious, works, the early isolation of the Earth's core is simply postulated or is justified by the solution of ill-posed problems of heating and cooling of the growing Earth during fall-outs of planetesimals of different sizes onto it (Vityazev *et al.*, 1990; Vityazev and Pechernikova, 1996). In view of the importance of this problem, we will consider it in more detail.

In all the hypotheses with a short-term accretion of planets (within about ten or several tens of millions of years) and with an overestimated depth of intermixing of the Earth's substance during impacts of planetesimals, it was obtained that the Earth should be melted even during its formation. But if this happened, the Earth, as well as the Protomoon, would undergo a fast and full differentiation of its substance, which would contribute a considerable amount of energy to melting of the Earth. As a result, right at the beginning of the Earth's life, about 4.6 Ga ago, a dense iron core would appear, a molten layer of a powerful anorthositic crust would form, and an early degassing of the Earth's substance with formation of a very powerful fluidic water-carbon-dioxide atmosphere would also take place.

If, as a result of such differentiation, a powerful (up to 80 km) and rather light (with a density of 2.7 g cm^{-3}) anorthositic Earth's crust with an age of 4.6 Ga would indeed separate, it should already remain forever on the Earth (as required by the Archimedean law). The same can be said about a powerful initially-basaltic crust.

However, in spite of all efforts of geologists, neither tracks of such primitive ancient crust nor other indications of early catastrophic differentiation of the Earth were found. According to geological data, as noted above, the Earth's crust was forming gradually, only since approximately 3.8 Ga ago. Moreover, if in the young Earth, about 4.6 Ga ago, a metal core separated and a powerful anorthositic crust formed the most part of radioactive elements would go to this crust, like the Moon.

After this, the Earth, having lost all the sources of endogenous energy, like the Moon, would become a tectonically dead planet. Taking these reasons into account, we supposed that the process of the Earth's core isolation began only 600 Ma after the Earth's formation, i.e., about 4.0 Ga ago (Monin and Sorokhtin, 1982a, b).

Sometimes, to prove an early heating and differentiation of the Earth, the data on the noble gas isotope distribution in the atmosphere and mantle are invoked. The excess abundance of the xenon ^{129}Xe radiogenic isotope in the Earth's atmosphere (its concentration is approximately 7% higher than it is supposed to be for the composition of the primary xenon) is especially indicative in this respect. But the ^{129}Xe isotope arises from the decay of a short-lived radioactive isotope of iodine ^{129}I with decay constant $\lambda_{129} = 4.41 \times 10^{-8}$ year^{-1}. Hence it is concluded that the excess of ^{129}Xe in the Earth's atmosphere testifies to an early differentiation and degassing of the Earth, which took place still before disappearance of the ^{129}I isotope from the Earth's substance (Tolstikhin, 1986; Ozima and Podosek, 1983; Azbel and Tolstikhin, 1988). However, it should be noted that in samples of the Earth's rocks excesses of ^{129}Xe are also sometimes found, even exceeding those in the atmosphere; this rather indicates a late degassing of the Earth. Marking an inconsistency of the interpretation of the xenon isotopic ratios, M. Ozima and F. Podosek, who are indisputable authorities in geochemistry of noble gases, note: 'Increase of the contents of radiogenic isotopes of xenon in the atmosphere, as well as the excess of ^{129}Xe, require that the degassing should be extremely fast; this not only contradicts the models for argon and helium, but also is internally inconsistent. On the other hand, the proximity of the isotopic composition of the major part of the mantle xenon to the atmospheric xenon and absence of significant excess of ^{136}Xe, connected with ^{129}Xe, indicates a slow degassing' (Ozima and Podosek, 1983, p. 310). As an alternative, Ozima and Podosek propose a two-stage model: in the beginning, at a very early stage of development of the Earth, a violent and fast degassing happened, during which the major part of noble gases was released to the atmosphere, and then, during the subsequent life of the Earth, its gradual degassing was developing.

The model of Ozima–Podosek can hardly be rejected, except for a 'minor' detail: not the Earth itself, but only the planetesimals that fell onto the Earth were subject to early degassing. Of course, this process was rather violent, because during the impacts against the Earth's surface and heat explosions the planetesimals could even evaporate. But all chemically active gases (CO_2, H_2O and other volatile species) quickly reacted with porous regolith of ultrabasic composition, which covering the growing Earth at that time, and were soon removed from the primitive Earth's atmosphere (Sorokhtin and Ushakov, 1989, 1991). At the same time, in the primary atmosphere only the noble gases and partially nitrogen were predominantly preserved and accumulated. It is obvious that such a violent degassing of the planetesimals could in no way characterise the heat regime of the Earth and, particularly, could not be an indicator of its early differentiation.

However, there is also direct evidence that the young Earth had never melted and that it had no dense metal core. For example, many differences between the geochemistry of lunar rocks and those of the Earth can be explained only if the parental body of the Moon (i.e., the Protomoon), in contrast to the Earth, was completely molten soon after its formation. Meanwhile, the Protomoon passed a full differentiation with isolation of a metal core and anorthositic crust. This is indicated, for example, by a powerful anorthositic crust on the Moon, and a strong depletion of all siderophile and chalcophile elements in the lunar rocks (as compared to the rocks found on Earth) (Ringwood, 1979; Henderson, 1982). However, the most vivid and virtually indisputable testimony for this are the isotopic ratios of lead on the Moon and the Earth. In lunar rocks that evident had separated after a complete melting of the Protomoon, the ratios of the radiogenic isotopes of lead with atomic weights 206, 207 and 208 (formed from the decay of uranium 238 and 235 and thorium 232) to the stable isotope 204 are extremely high. In lunar rocks these ratios reach on the average 207, 100, 226 respectively and greater, whereas for the rocks found on Earth, the ratios averaged in the oceanic reservoir of pelagic sediments, are 19.04, 15.68 and 39.07. For the primary leads (judging from the isotopic composition of the Canyon Diablo iron meteorite, Arizona, USA) the ratios are even lower, only 9.50, 10.36 and 29.45 (Voytkevich *et al.*, 1990).

The above ratios imply that during melting of the Protomoon the lunar substance indeed lost 96 to 98% of the primary (nonradiogenic) lead (passed to the Protomoon's core), and only the radiogenic lead was accumulating in the lunar crust and basalts. Such a loss of primary lead from the lunar substance can only be explained only by full melting of the Protomoon's substance and transfer of lead and its sulphides to the Protomoon's core. The Canyon Diablo iron meteorite, in which the lead isotopes indeed closely fit their primary ratios, must be considered as a fragment of the core of a satellite, which, like the Protomoon, underwent tidal melting, differentiation and destruction as early as at the stages of the planets' formation.

The presented ratios of isotopes of lead almost unambiguously fix the fact of full melting and differentiation of the lunar substance, and also convincingly show that the Earth had never melted completely and was not subject to such a radical differentiation.

For the same reason we cannot agree with numerous hypotheses of the formation of the Moon at the expense of so-called 'megaimpacts' or 'macroimpacts', according to the hypothesis of Vityazev and Pechernikova (1996). If the Moon was really formed of splinters of the Earth's mantle, which had been thrown out to near-Earth space by a grazing impact (or impacts) of planet-like bodies, then the same ratios of lead isotopes would be observed on the Moon as in the rocks of the Earth's mantle. In this case, these ratios in the lunar rocks would be: $^{206}Pb/^{204}Pb \approx 18$–$19$; $^{207}Pb/^{204}Pb \approx 15$–$16$ and $^{208}Pb/^{204}Pb \approx 37$–$38$, instead of the above-mentioned violent values from 100 to 220 (Voytkevich *et al.*, 1990).

As opposed to the lunar substance, the Earth's substance has never been exposed to a fast and radical differentiation. This is explained by the fact that the Earth's core was forming gradually and without melting of silicates, owing to the barodiffusion mechanism of differentiation of the mantle substance; the rate of action of this mechanism was always restrained by extremely low coefficients of diffusion in the mantle silicates (Monin and Sorokhtin, 1981). In addition to this, both the primary lead and its radiogenic isotopes, which could have been accumulated by the time of the Earth's substance differentiation, passed to the Earth's core simultaneously. Hence the intermediate value is rather close to the primary ratio of the lead isotopes in the Earth's rocks (in comparison with the same ratios in the lunar substance).

Nevertheless, when the lead-isotope age determinations of lead deposits on the Earth were checked by the standard method (based on the assumption of a uniform and constant mantle-crust reservoir for lead, uranium and thorium, according to the model of Holmes–Houtermans), it turned out that all such determinations underestimated the age of the Earth's rocks by approximately 400–500 Ma (Faure, 1986). Systematic discrepancies between the single-stage Pb ages and the ages of the same rocks determined from other isotopic or geological data, have required a refinement of the model of evolution of the Earth's lead. Taking this into account, Stacey and Kramers (1975) proposed a two-stage model of variations of the isotopic ratios in leads of different ages. The parameters of this model were chosen such that it best approximated the empirical data on the ages of the majority of the world's lead deposits, whose ages could be reliably determined by other methods (for example, from the rubidium–strontium, potassium–argon or samarium-neodymium ratios as well as from geological data based on the uranium–lead ratios in zircons). According to the Stacey–Kramers model, the evolution of the isotopic ratios of lead began 4.57 Ga ago in a closed reservoir, but later, approximately 3.7 Ga ago, the U/Pb and Th/Pb isotopic ratios sharply changed as a result of geochemical differentiation of the Earth's substance. It is remarkable that in this model the beginning of the accumulation of excess radiogenic lead approximately coincides with the beginning of formation of the Earth's crust, and, hence, of the Earth's core.

In the concept of the Earth's global evolution which was developed earlier (Monin and Sorokhtin, 1981, 1982a, b; Sorokhtin and Ushakov, 1991), we quantitatively determined the basic regularities of the Earth's core isolation. Using this concept, we tried to transform the known equations describing the dependence of the lead isotopic ratios on the age. We took into account that some part of lead passed to the Earth's core, whereas radioactive elements mainly moved to the Earth's crust. Transfer of iron and its oxides (i.e., core substance) to the Earth's core is determined by simple relationship (Monin and Sorokhtin, 1982a)

$$C = C_0 \frac{1 - x}{1 - C_0 x}, \tag{4}$$

where C is the mantle concentration of iron and its oxides recalculated to the eutectic alloy $Fe \cdot FeO$ (i.e., to the core substance Fe_2O); $C_0 = 0.37$ is the total content of the core substance in the Earth; $x = M_c/M$ is the evolution parameter of the Earth, introduced by Monin and determining the relative mass of the Earth's core; M_c is the mass of the Earth's core; M is the Earth's mass. The present-day value of the Earth's evolution parameter x_0 is 0.863. The dependence of C on time t is set in terms of parameter x, defined by Monin and Sorokhtin (1981), Sorokhtin and Ushakov (1991). Figure 5 shows parameter x *versus* time t.

In our model, as well as in the Stacey–Kramers model, in an age interval of 4.6–4.0 Ga the U/Pb and Th/Pb ratios developed in the closed reservoir of the young

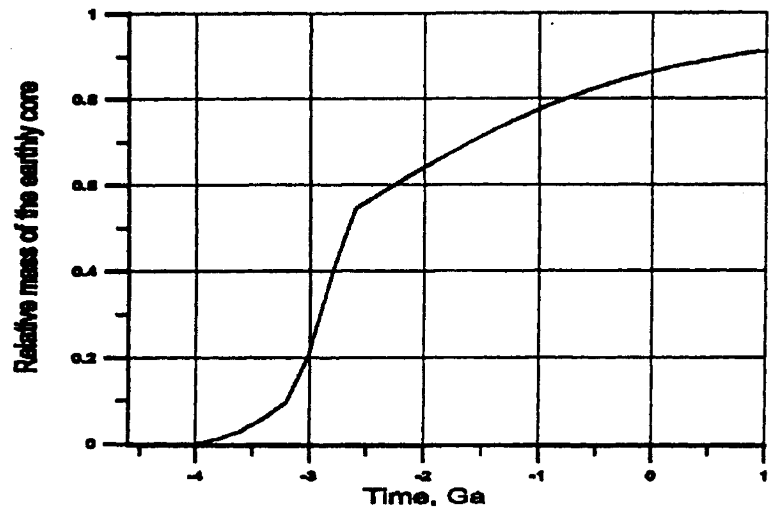

Figure 5 Time dependence of the Earth's evolution parameter x.

Earth without loss of lead. After the moment 4.0 Ga ago, the ratios of the radiogenic leads to primary ^{204}Pb in the mantle are already affected by a gradual transfer of lead to the Earth's core. In this, the main relocation of radioactive elements to the continental crust followed the laws close to expression (4). To simplify the equations that follow, we introduce the following notation:

$$\frac{^{206}\mathrm{Pb}}{^{204}\mathrm{Pb}} = a; \quad \frac{^{207}\mathrm{Pb}}{^{204}\mathrm{Pb}} = b; \quad \frac{^{208}\mathrm{Pb}}{^{204}\mathrm{Pb}} = c. \tag{5}$$

We will also designate, with a subscript, the age in billions of years of the moment to which the parameter considered refers. In this case, for the initial (Catarchaean) stage of the Earth's development, we can write the usual relationships for the dependence of the lead isotopic ratios on its age (Faure, 1986)

$$\begin{aligned} a_i &= a_{4.6} + \mu(e^{\lambda_1 t_{4.6}} - e^{\lambda_1 t_i}), \\ b_i &= b_{4.6} + \mu\frac{(e^{\lambda_2 t_{4.6}} - e^{\lambda_2 t_i})}{137.88}, \\ c_i &= c_{4.6} + \omega(e^{\lambda_3 t_{4.6}} - e^{\lambda_3 t_i}), \end{aligned} \tag{6}$$

were $\lambda_1 = 0.15513 \times 10^{-9}$; $\lambda_2 = 0.98485 \times 10^{-9}$ and $\lambda_3 = 0.049475 \times 10^{-9}$ year^{-1} are the constants of decay of ^{238}U, ^{235}U and ^{232}Th; $\mu = {}^{238}U/{}^{204}Pb$ and $\omega = {}^{232}Th/{}^{204}Pb$. In the Stacey–Kramers two-stage model, at the initial stage, prior to the geochemical differentiation, $\mu_1 = 7.192$ and $\omega_1 = 32.208$. At the second stage, after the beginning of the Earth's substance differentiation, $\mu_2 = 9.735$ and $\omega_2 = 36.837$.

In our model, we have changed slightly the initial values of factors $\mu = 7.767$ and $\omega = 33.593$, but included into them the corrections for transfer of lead to the

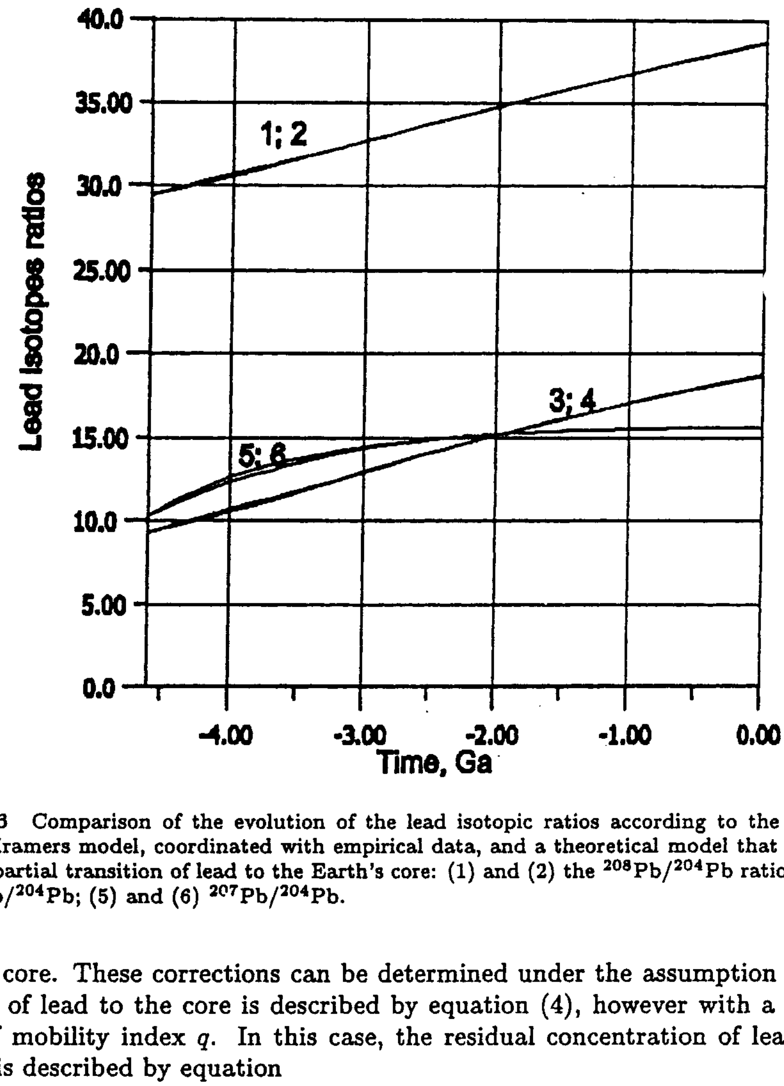

Figure 6 Comparison of the evolution of the lead isotopic ratios according to the two-stage Stacey–Kramers model, coordinated with empirical data, and a theoretical model that takes into account partial transition of lead to the Earth's core: (1) and (2) the $^{208}Pb/^{204}Pb$ ratios; (3) and (4) $^{206}Pb/^{204}Pb$; (5) and (6) $^{207}Pb/^{204}Pb$.

Earth's core. These corrections can be determined under the assumption that the transfer of lead to the core is described by equation (4), however with a different value of mobility index q. In this case, the residual concentration of lead in the mantle is described by equation

$$X_i = 1 - q\left[1 - \frac{(1 - x_i)}{1 - C_0 x_i}\right]. \tag{7}$$

Let us now set in our model the primary and present-day ratios of the lead isotopes (i.e., the boundary conditions of the problem) the same as in the Stacey–Kramers model: $a_{4.6} = 9.307$; $b_{4.6} = 10.294$; $c_{4.6} = 29.476$; $a_{0.0} = 18.700$; $b_{0.0} = 15.628$; $c_{0.0} = 38.630$. In this case, for the Archaean and subsequent epoch we find a new, updated set of equations

$$a_i = a_{4.6} + \mu \frac{(e^{\lambda_1 t_{4.6}} - e^{\lambda_1 t_i})}{X_i},$$

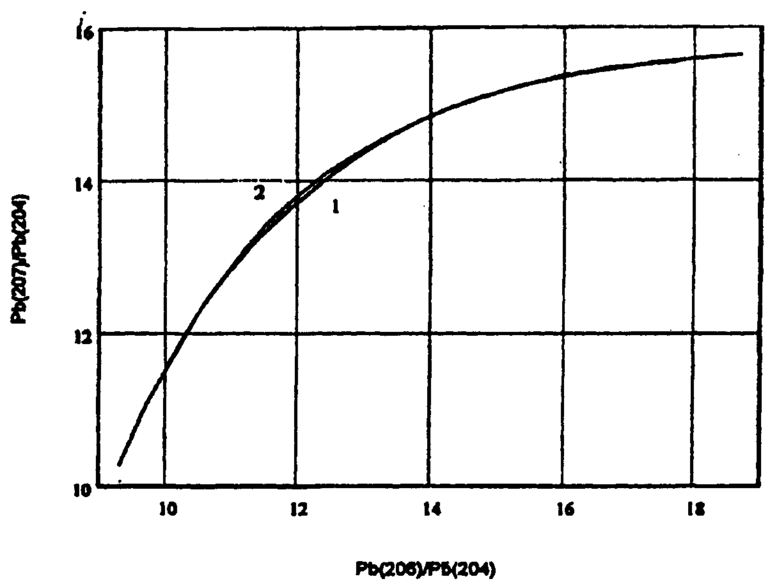

Figure 7 Evolution of the lead isotopes: (1) two-stage Stacey–Kramers model; (2) two-stage model taking into account the transition of some part of lead to the Earth's core.

$$b_i = b_{4.6} + \mu \frac{(e^{\lambda_2 t_{4.6}} - e^{\lambda_2 t_i})}{137.88 X_i},$$
$$c_i = c_{4.6} + \omega \frac{(e^{\lambda_3 t_{4.6}} - e^{\lambda_3 t_i})}{X_i}. \quad (8)$$

It should be noted that mobility factor q in relationship (8) already describes not the transition of lead to the core, but only the changes (caused by such transitions) of the isotopic ratios' formed during the Earth's core formation, i.e., during the last 4 Ga. But the formation rate of the radiogenic lead isotopes is different, too. For example, during the geological development of the Earth, i.e., the same last 4 Ga, the ratio of isotopes $^{206}Pb/^{204}Pb$ has increased by 42%, whereas the ratios of isotopes $^{207}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$ did so only by 19% and 21%. In accordance with this, for different isotopes of lead their effective mobility factors q are also different, varying from 0.184 for ^{206}Pb to 0.074 and 0.087 for ^{207}Pb and ^{208}Pb.

Now we can calculate the dependences of b_i on a_i, and of all parameters a_i, b_i and c_i on time t_i. The results of such calculations in comparison with the Stacey–Kramers model are presented in Figures 6 and 7.

It is visible that both two-stage models coincide with each other very well. But the Stacey–Kramers model was coordinated with other independent age determinations of actual lead deposits; therefore our model also should fit the same empirical data fairly well. Hence, we can assert that our model of lead transition in the Earth's core is indeed consistent with the available empirical data on the Earth's lead isotopes.

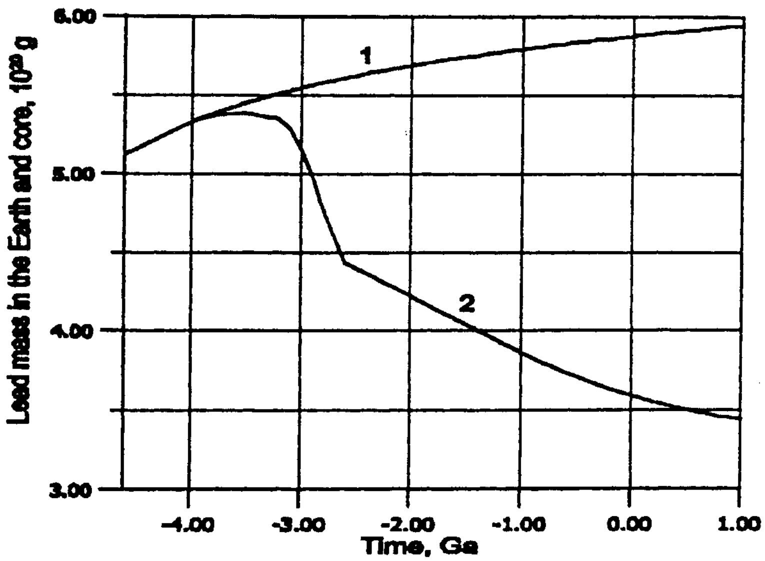

Figure 8 Evolution of the lead mass in the Earth: (1) the total lead mass in the Earth; (2) the total lead mass in the mantle and continental crust. The distance between the two curves defines the lead mass that has passed to the Earth's core.

In connection with the gradual transition of lead to the growing core of the Earth, a part of its radiogenic isotopes in the total mass of the mantle (and crust) lead, naturally, should gradually rise. It is easy to calculate that if lead did not pass to the Earth's core, then, according to the Stacey–Kramers model, the present-day isotopic ratios would be: $(a_0)' = 16.8$, $(b_0)' = 15.04$ and $(c_0)' = 37.71$, and in our model, 17.32; 13.32 and 38.0 respectively, instead of the commonly accepted values for the modem crust isotopic ratios: 18.7, 15.63 and 38.63. Taking into account the above values of the lead isotopic ratios, we can estimate that approximately 30% of this metal has sunk to the Earth's core. Supplementing these data with determinations of the radioactive elements' distribution in the Earth (Sorokhtin and Ushakov, 1991), we can also determine the mass of lead that has passed to its core during the geological development of the Earth: in total about 2.29×10^{20} g (see Figure 8).

With reference to the lead ratios in the lunar substance $a_L \approx 200$; $b_L \approx 100$ and $c_L \approx 230$, we find that during melting and differentiation of the Protomoon up to 96 to 97% of the primary lead was removed from the lunar reservoir in the due time.

There is a natural question: why, during the differentiation of planets with identical initial compositions, did the lead conversion factors in the planetary cores differ from each other so widely? The answer is simple: the Protomoon was overheated by tidal deformations, underwent full melting and also full gravity differentiation by means of liquation of melts. The Earth has never melted completely (we can speak only about partial melting of the substance of the upper mantle), and the isolation of the Earth's core was taking place gradually and without melting of silicates. At the same time, lead belongs to disseminated elements, does not form a free phase in the mantle substance and is contained in the crystalline lattices of silicates or sulphides. Therefore, the release of lead from the mantle substance and its transition to the Earth's core is, most likely, due to the same barodiffusion mechanism (but with different diffusion coefficients) that releases the oxides of iron forming the Earth's core, i.e., without melting of the mantle substance (Monin and Sorokhtin, 1981).

In addition to the above-mentioned geochemical and geological data, which indicate that the young Earth was not subject to melting and radical differentiation, we will present purely ecological evidence for this. In the case of full melting of the young Earth, accompanied by isolation of the Earth's core, its interiors would have been completely degassed. In this case, about 5×10^{23} g of carbonic gas, nowadays connected in carbonate rocks, and more than 2.5×10^{24} g of water would be supplied to the Earth's atmosphere in a rather short time. A single-moment formation of such a dense carbon–dioxide atmosphere with a pressure of about 100 atm would produce an extremely strong greenhouse effect with a rise in the surface temperature above the critical temperature of water (+374°C). After that the oceans would boil, and the pressure of the Earth's atmosphere would rise still higher, approximately by 500 atm. As a result on the Earth, like Venus, an irreversible (we emphasise, *irreversible!*) greenhouse effect would be established, with average temperatures stably exceeding 550–600°C. In this case, on the Earth there would be no liquid phase of water and no hints to even the most primitive life. Fortunately for us and all creatures living on the Earth, this has not taken place.

6 A POSSIBLE SCENARIO OF THE EARTH'S CORE FORMATION

Thus, how did the Earth's core form? We believe the following scenario to be the most probable one and this is considered in more detail in our monograph (Sorokhtin and Ushakov, 1991). In the Catarchaean the tidal energy was released mostly in the equatorial belt of the Earth, because at that remote time the Moon was revolving around our planet in the plane of the equator (Sorokhtin and Ushakov, 1989). In the same belt, there formed the first ring zone of the Earth's substance differentiation with a gradually sinking layer of molten iron and its oxides (see Figure 9). For the same reason the tectonic activity of the Earth originally should also show itself only in its equatorial belt.

It is interesting to note that, together with iron and its oxides, the majority of siderophile and chalcophile elements also passed to the ring layer of melts. Therefore, the convection mantle above the sinking ring layer of zonal differentiation of the Earth's substance in the Archaean was depleted with iron and with these elements. Just for this reason, probably, the Archaean continental shields and greenstones belts in them do not display an enhanced metallogenetic potential. As the front

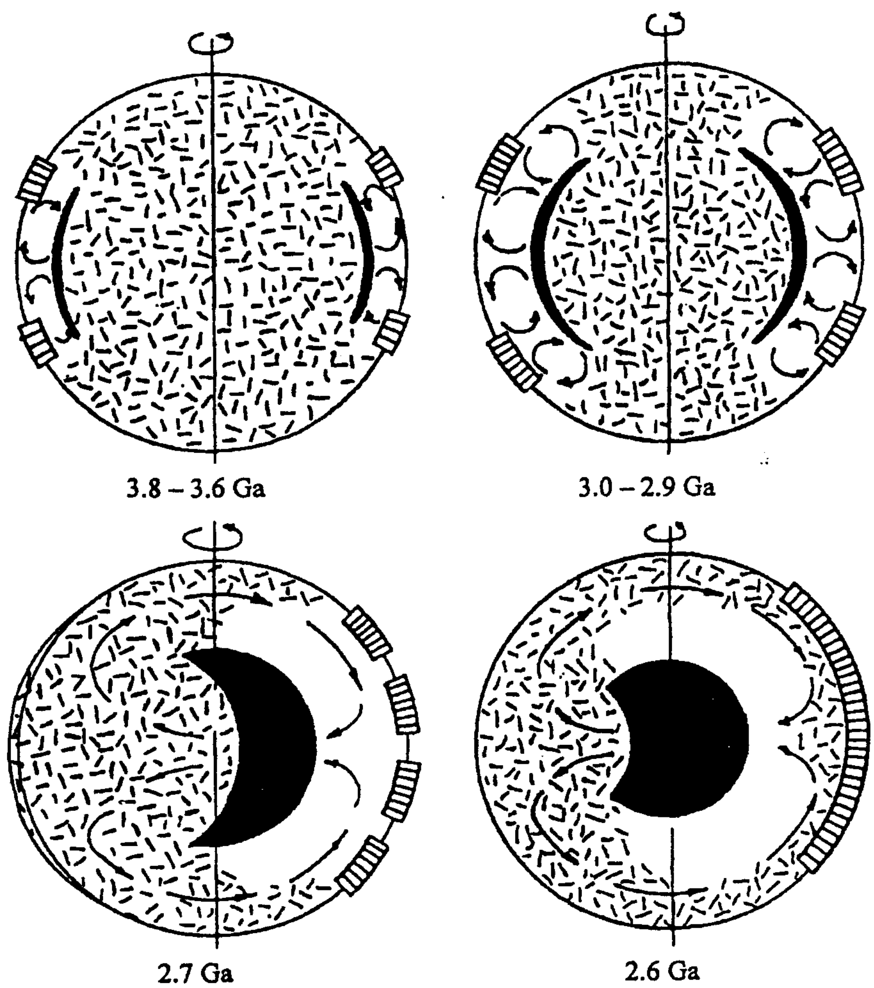

Figure 9 Consecutive stages of development of the zonal differentiation process of the Earth's substance and the Earth's core formation. Black: the melts of iron and its oxides; dashes: primary Earth's substance; radial shading: continental massifs (the formation of the Monogea supercontinent about 2.6 Ga ago is shown).

of differentiation was penetrating deeper into the Earth, the ring zone of differentiation of the Earth's substance was gradually extending. The molten iron and its oxides could not flow down to the centre of the Earth because of the very large viscosity of the cool Earth's substance in its central parts ($\eta \gg 10^{30}$ poise). As a result, a strong gravitational instability of the planet arose, when heavy ferrous melts were located above the less dense, but very rigid 'pith' of the Earth. Such an instability should finally terminate in a catastrophic event: floating-up of the rigid

Earth's 'pith' in the equatorial zone of one of the Earth's hemispheres and sinking of heavy melts to the centre of the Earth in the opposite hemisphere, as shown in Figure 9.

This process should be accompanied by generation of a huge energy, of the order of 10^{37} erg, by intense convection currents in the mantle, completely and radically rebuilding all the previous regime of tectonic development of our planet, and by formation of Monogea, the first supercontinent in the history of the Earth. This event, most probably, took place at the very end of the Archaean. The formation of the Earth's core about 2.6 Ga ago can probably be explained in such a way. In particular, this scenario is supported by paleomagnetic data, which also testify that the dipole magnetic field of the modern type appeared on the Earth only about 2.6 Ga ago (Hale, 1987), i.e., just at the boundary between the Archaean and Proterozoic (see Figure 1).

7 GEOLOGICAL EVIDENCE FOR THE EARTH'S CORE FORMATION AT THE ARCHAEAN-PROTEROZOIC BOUNDARY

We have already noted that in the Archaean, when the zonal differentiation of metal iron took place, the convection mantle was depleted with iron and siderophile elements (see Figure 10). At the very end of the Archaean – beginning of the Proterozoic, during the isolation of the Earth's core, the composition of the convection mantle changed radically. Indeed, at this particular time, the substance of the former pith of the Earth, with its primary concentration of iron (about 13%), iron oxides (about 24%), siderophile elements, sulphides of chalcophile metals and other ore elements, including platiniferous minerals, was added to the mantle.

The direct evidence for this phenomenon are unique differentiated intrusions of basic and ultrabasic rocks of the early Proterozoic age on many ancient shields. The most conspicuous and classical complexes of this type are the famous Bushveld stratified intrusion massif in South Africa (iron, titanium, platinum, chromium), the Great Dike intrusion in Zimbabwe (chromium, platinum), intrusions of norites of Sudbury in Canada (cobalt, nickel, copper, platinum).

In Russia, there is the Burakovsky intrusion in the southeast part of the Baltic shield with chromium, nickel, vanadium, platinum and, probably, gold metallogeny. The Pechenga gabbro-norit complex (cobalt, nickel, copper), the Panskaya intrusion on the Kola Peninsula and magmatic formations of the Olongskaya group in Karelia (platinum) should probably also be classified as structures of a similar type.

Let us emphasise here that the intrusion formations of this type with such a high concentration of ore elements were not formed again, either before the early Proterozoic or after it. This indicates once more the uniqueness of the early Proterozoic metallogenic epoch and supports our model of enrichment of the mantle in this epoch by the primary Earth's substance, which had risen from the central parts of the Earth during the formation of the Earth's core at the very end of the Archaean, as shown in Figure 9.

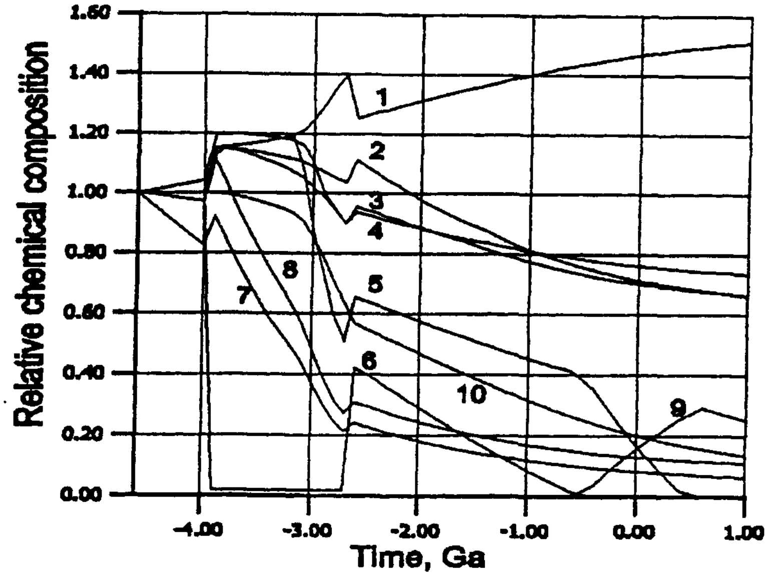

Figure 10 The evolution of the chemical composition of the convection mantle in relative concentrations (unit is the concentration of the given element in the primary substance of the Earth): (1) SiO_2; MgO; CaO; Al_2O_3; Na_2O; (2) Pb (total, the mantle plus the crust); (3) H_2O; (4) K_2O; (5) FeO; (6) Fe; (7) U; (8) Th; (9) Fe_3O_4; (10) core substance: iron and oxides of iron recalculated to eutectic $Fe \cdot FeO$ (or Fe_2O). The variation of the siderophile ore elements concentration in the mantle is similar to curve 6.

For this reason, right at the end of the Archaean – early Proterozoic the concentration of iron in the mantle should also rise. According to our estimates, the average concentration of metal iron in the mantle then reached 5.5%, and that of bivalent iron, 15% (see Figure 10). In oceanic rift zones the hot iron together with its sulphides and sulphides of other metals (copper, zinc, lead) rose to the surface of the Earth and contacted the ocean waters. In contact with water in an oxygen-free environment, characteristic of the early Proterozoic, the hot iron was oxidised to soluble bivalent hydroxide and was spread by currents over all the oceans. In the near-surface conditions the bivalent iron, because of the vital activities of microbes and microalgae, was oxidised to the trivalent condition (or up to magnetite) and precipitated, gradually forming unique iron-ore formations of the Precambrian. About 90% of all world iron resources on the Earth are concentrated in the deposits formed during this period. This once again confirms the uniqueness of the Archaean – early Proterozoic boundary and convincingly testifies to the mechanism of the Earth's core isolation we have described here.

An additional, however weighty, argument for the model accepted with reference to the formation of the Earth's core and transfer of the primary substance of the Earth's interiors to the convective mantle at the Archaean – early Proterozoic boundary can be the osmium $^{187}Os/^{186}Os$ ratios in the rocks of the Bushveld complex. Indeed, the ^{187}Os radiogenic isotope forms by β-decay of rhenium ^{187}Re, but osmium is a strongly pronounced siderophile element, it belongs to the iron group and is transferred to the Earth's core together with iron, whereas rhenium belongs to the manganese group and is mainly concentrated in the Earth's crust. For this reason and also because of a comparatively long half-life of rhenium (4.56×10^{10} years), in the convective mantle the $^{187}Os/^{186}Os$ ratios changed relatively little: from the value typical for iron meteorites (0.8) to their values in the secondary peridotites (1.05), while for the Bushveld massif this ratio is 1.45 (Allegre and Luck, 1980; Luck and Allegre, 1983). Such an enhanced osmium isotopic ratio in an obvious mantle intrusion can be explained only by the possibility that, shortly before its intrusion in the continental crust, some primary substance with an appreciable concentration of ^{187}Re, which had not yet undergone differentiation, was added to the convective mantle. The decay of ^{187}Re in this substance resulted in excessive amounts of the ^{187}Os radiogenic isotope. It would be interesting to check the osmium isotopic ratios in this respect in other mantle intrusions of the early Proterozoic.

8 CONCLUSIONS

Thus, an analysis of the isotopic ratios of lead in the Earth's and lunar rocks, as well as numerous geological data, which were mentioned above, almost unambiguously testify that the young Earth was originally a rather cool and homogeneous planet, without a dense core and Earth's crust, as follows from the fundamental theory of the planets' origin of Schmidt–Safronov. During the first 600–800 Ma after its formation (in the Catarchaean) the Earth remained a tectonically passive planet, it never completely melted, and its core was separated gradually, beginning from approximately 4–3.8 Ga ago and throughout all the subsequent history of its geological development. The process of the Earth's core growth is now continuing, being thus the main energetic process that is powering the tectonic activity of the Earth.

The Moon, most likely, was formed by the destruction, at the Roche limit, of a larger cosmic body (Protomoon) captured by the gravity field of the growing Earth from a neighbouring orbit. As a consequence of the tidal interactions with the Earth, the Protomoon was overheated, melted and underwent a full gravity differentiation during the formation of the Earth–Moon system. During the destruction of the Protomoon, all the substance of its interior tidal hump and ferrous core fell onto the Earth, whereas the Moon basically was formed only from the differentiated silicate substance of the exterior tidal hump. This can explain all the specificity of its geochemistry and chemical composition.

In conclusion, we should note that, when improving the theory of planets' origin, it is necessary to take into account modern geological, geochemical and cosmochemical data, some part of which was presented and interpreted in this paper. Therefore, the creation of a comprehensive theory of formation of the Earth, Moon and other planets of the Solar system is possible only in a full unification of theoretical constructions with actual geological, geochemical and cosmogeological data.

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