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Some Aspects of the Formation of the Solar System

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Some Aspects of the Formation of the Solar System

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Vsevolod N. Anfilogov
Institute of Mineralogy
Russian Academy of Sciences
Miass
Russia

Yurij V. Khachay
Institute of Geophysics
Russian Academy of Sciences
Ekaterinburg
Russia

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Preface

This book is devoted to the problems that occur when attempting to understand and construct a concise representation of the original conditions, composition and dynamics of the evolution of the Earth-Moon system in the form in which it is seen today. This volume will perhaps contribute to a better understanding of what is necessary to research the dynamics of the Solar system.

The complexity of the problem demands great effort from many generations of outstanding scientists. In the introduction, which by no means claims that this work is comprehensive, we wish to illustrate the idea that any weaknesses in the conclusions and the obtained consequences are mainly due to the limitations of the observed experimental data, the analytical base of the geochemical and isotopic research, and the possibilities of astronomy and geophysical methods. That problem is addressed in the first chapter.

In other chapters, attention is mainly directed to the presentation of researching results on the dynamics of the formation of the Earth and Moon, which are based on Safronov's accumulation model and the results of numerical mathematical modelling. These results show that the account of heat release from the decay of short-living radioactive elements leads to the early heating of pre-planetary bodies and makes the collisions inelastic. The consequence of that is the realization of the process of matter differentiation, which results in the combining of fragments enriched by iron with a low melting temperature, while the more infusible silicate parts remain in the proto-planetary cloud. A significant role in the thermal balance of the growing planet was played by the existence of a poorly transparent atmosphere with a large amount of silicate dust. Finally, we present the first results obtained by numerical solution of the evolutionary problem for a 3-D medium model, which describes the development of thermal heterogeneities, which are stipulated by the random distribution of the bodies and particles falling on the surfaces of the growing Earth and Moon.

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Miass, December 2014 Vsevolod N. Anfilogov
Ekaterinburg Yurij V. Khachay

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Chapter 1
Introduction

Abstract At present, there is no generally recognized model that can describe the whole diversity of the conditions which led to the formation of the Solar system. The numerous hypotheses in which authors try to solve this problem can be divided into two groups.

Keywords Solar system · Molecular clouds · Supernova · Meteorite · The Earth · The Moon · Formation

The first group includes hypotheses that proceed from the assumption that the Sun and planets of the Solar system were formed during the evolution of the uniform gas-dust cloud. The second group combines hypotheses that assume that the planetary material was captured by the Sun from some other cosmic object or was ejected from the Sun as a result of the interaction of gravity with such an object, which came near to the Sun at some distance at which the interaction was significantly strong.

One of the first attempts to answer the question how the Solar system occurred was made by Immanuel Kant. He assumed that the planets had been accumulated from the dust clouds that rotated around the Sun at distances which were near to the present day planetary orbits, and that each planet had been formed from its own cloud [1]. After half a century, Laplace gave that idea a strong mathematical formulation. Laplace assumed that there had been a proto-solar gas cloud which could rotate, and that this rotation had been a necessary condition for the balancing of gravitational force during the process of the Solar system's formation. As a result of the action of the gravitational and centrifugal forces, the gas cloud contracted to a thick proto-planetary disc and, by further compression, that disc was divided into the central part, from which the Sun was formed, and into concentric rings, which were transformed into planets. Laplace's hypothesis, which was very popular in the XIXth century, was rejected by Maxwell, who showed theoretically that the gas rings could not condense to planets, because the masses of their matter in the rings were insufficient to form under the action of their own gravitational forces and to gather into clusters of planetary sizes [2].

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Hypotheses about the separate origin of the Sun and planets can also be divided into two groups [3]: 1—the hypotheses of Buffon, Moulton, Chamberlain, Jeans et al., which assume that the proto-planetary material was ejected from the already formed Sun [4]; and 2—hypotheses that the material had been captured by the Sun from the interstellar medium, assumed by Alfven, Schmidt and Littleton [4, 5]. At present, most researchers prefer the variant of the Sun's formation as a result of the collapse of the interstellar gas–dust cloud during the process when the Sun was formed together with the planets.

One possible variant of the universe's evolution, which is connected with the origin of the planetary systems, was proposed by Weizsäcker in 1944 [4]. Weizsäcker assumed that the universe was a gas cloud which, during the evolution process, had been divided into numerous turbulent vortices, and that the numerous nebulae were formed from these objects. Inside each nebula, there occurred turbulent currents that formed the maternal bodies of the stellar clusters. The Sun is one of the constant stars formed in this way. It is assumed that it was surrounded by a rapidly rotating gas nebula, in which, due to the rotation, there also occurred turbulent vortices, in the centre of which clusters of the proto-planetary material formed, from which the planets were then in turn formed.

The ideas of Weizsäcker were modified and developed by Kuiper [4]. He considered that the cloud that had surrounded the Sun and had been composed of gas and thin dust later became a flat disk, which was divided into turbulent vortices, in the centre of which the planets were formed. At the present time, the idea of the planets dividing from the Sun has been abandoned, and all models are now constructed using the assumption that the Solar system was formed from a single initial gas–dust cloud. The characteristics of the interstellar clouds, from which, as it was assumed, the Solar system was formed, are presented in Chapter 1.

At the present time, there is active discussion in the literature of the role of the supernova in the formation of the Solar system. Herbst and Azusa [6] and Shramm [7] consider that the presence of an anomalous concentration of ^{26}Mg, formed by decay of the short-living isotope ^{26}Al, is the main evidence for the role of the supernova in the Solar system's formation. The sources from which the short-living isotopes were introduced into the Solar system are discussed in [8, 9]. In the work of Ouellette, two models are considered: the trigger model and the air gel model. The trigger model presumes that the explosion of a supernova acted as a trigger mechanism for collapse of the molecular cloud, into the core of which the short-living isotopes were introduced. This raises the question: what matter would be injected into the proto-Solar cloud from outside for the short-living isotopes, if we assume that the whole composition of the proto-Solar cloud material could be delivered by the supernova? We think that the delivery of those elements from different sources, as assumed in different models, will not solve that problem, because besides the short-living isotopes of aluminum, manganese and elements of the iron group in the Solar system there exist heavy elements that could only be formed by the explosion of the supernova.

The physical characteristics of the Solar system are given in Sect. 2.3.

The analysis of the existing models of the Earth's formation is given in Chap. 2. It is shown that the model of homogeneous accumulation from the cold gas–dust cloud, which enjoys the greatest recognition, cannot solve the main problem: to find or to define the mechanism of the separation of the initial homogeneous proto-Earth into its metallic core and silicate mantle. The existence of the meteorites CAI-inclusions, enfolded by minerals enriched by calcium and aluminum, and also iron meteorites allows us to conclude that the initial proto-planetary cloud was gaseous, heated to a temperature of about 2000 K and that the mineral condensation from the gaseous phase coincides with the sequence calculated by Grossman [10]. This conclusion is in good agreement with the relative age of CAIs and iron meteorites [11]. That conclusion then allows us to assume that the initial Earth's pre-planetary bodies were composed of material of CAI, which were located in their centers, and of Fe–Ni alloy, which was located in the middle part of these bodies. The outer envelope of the pre-planetary bodies may have been composed of material similar to pallasite. During the decay of ^{26}Al, the protoplanetary bodies were heated to temperatures higher than the melting temperature of iron.

The formation of the Earth's core began from the collision of the initial melted pre-planetary bodies, as a result of which the melted material of CAI was displaced from the pre-planetary body's centre and replaced by melted iron. A significant question concerns the existence in the Earth's core of components of low density. In our book we prove that one such component is the iron oxide FeO, which occurs from the melt or by melting of olivine, which contains the fayalite component. The growing temperature of the Earth's core is supported by the decay in the initial stage of ^{26}Al and by the kinetic energy that is exuded at inelastic collisions of partially melted pre-planetary bodies and transformed into thermal energy. The results of mathematical modeling have shown that the temperature of the core surface can reach 3000 K in the final stage [12–16].

The composition of the Earth's silicate mantle is determined by the composition of the meteoritic material condensed from the gas phase after iron condensation and formed from fragments of planetesimals caused by their collisions with the growing Earth. The age and the evolution of meteoritic material are discussed in Sect. 4.1. These data have allowed us to determine the modeling composition of the silicate mantle. This material was sediment on the core surface, which was heated to a temperature higher than the melting temperature of silicate. Therefore the silicate melt layer was formed on the core–mantle boundary, its thickness reaching 800–900 km. The silicate melt began to crystallize into the bottom of the melt layer at this thickness. The crystallized layer was composed of Mg-perovskite and magnesio-wüstite minerals on the core–mantle boundary and the melt layer moved up. The thickness of the melt layer decreased as the Earth grew and it became around 400 km at the final stage. Olivine was crystallized into the layer bottom at this thickness. The Earth's mantle transition layer at the depth of 400 km is composed of this mineral. The melt was enriched by FeO and Al_2O_3 as the olivine crystallized, and anorthite was crystallized in the upper part of the layer as the melt cooled [17].

Carbonaceous chondrite material, which contains about 7 % H_2O, is sediment on the Earth's surface at the final stage of heterogeneous accumulation. The outer solid envelope, with a thickness of 40–50 km, was formed from this material. The main masses of water H_2O and CO_2 were exuded from this envelope as a result of its heating by the melt layer and the aggressive hot ocean was formed during a very short period [18].

This model of the Earth's heterogeneous accumulation allows us to propose a new principle of the mechanism of the Moon's formation. The Moon formed simultaneously with the Earth. Fragments formed from the initial pre-planetary collisions passed into the Earth's satellite orbits and formed the material from which the Moon was formed in the initial stage. These fragments were composed of melted material of CAI, drops of melted iron and pallasite material. The major mass of iron passed to the Earth's core at the initial pre-planetary collisions. Therefore, the central part of the Moon is depleted of iron and enriched by calcium and aluminum. The melted state of the fragments allowed them to maintain the high temperature of the Moon in the process of its growth [19].

When the main part of the book had been written, we obtained the first results of a numerical solution of the Earth's and Moon's thermal evolution for a 3-D model. The results allowed us to quantitatively research the thermal after-effects of the random bodies and particles falling during the Earth's and Moon's formation [20, 21]. These results were obtained thanks to the active participation of A. Antipin, and are presented in the Appendix.

References

1. Urey H (1951) The origin and development of the Earth and other terrestrial planets. Geochim et Cosmochim Acta 2:209–277
2. Fisher D (1987) The birth of the Earth. Columbia University Press, New York
3. Vitjazev A, Pechernikova G, Safronov V (1990) The Earth's group planets. Nauka, Moscow (in Russian)
4. Miyake Y (1965) Elements of geochemistry. Muruzen, Tokyo
5. Schmidt O (1962) The origin of the Earth and planets. Academy of Science USSR Press (in Russian)
6. Herbst W, Assouza G (1982) The role of supernova on the star formation and spiral structure. In: Gehrels T (ed) Protostars and planets. The studies of star formation and of the origin of the Solar system. University of Arizona Press. Russian version. Mir, Moscow, pp 421–429
7. Shramm D (1982) Supernova and the solar system formation. In. Gehrels T (ed) The studies of star formation and of the origin of the Solar system. University of Arizona Press. Russian version. Mir, Moscow, pp 440–457
8. Ouellette N, Desh S, Bizzaro M et al (2008) The injection mechanisms of short-living radio nuclides and their homogenization. Geochim et Cosmochim Acta 73:4946–4962
9. Huss G, Meyer B, Srinivasan B et al (2008) Stellar sources of short-living radionuclides in the early Solar system. Geochim et Cosmochim Acta 73:4922–4945
10. Grossman L (1972) Condensation in the primitive solar nebula. Geochim et Cosmochim Acta 36:597–619
11. Amelin Y, Krot A (2007) Pb isotopic age of the Allende chondrules. Meteorit Planet Sci 42:1321–1337
12. Anfilogov V, Khachay Y (2005) A possible scenario of material differentiation at initial stage of the Earth formation. Doklady Earth Sci 403A:954–947 (in Russian)

13. Anfilogov V, Khachay Y (2013) Origin of the Kimberlite diamond bearing lithosphere of cratons. Dokl Akad Nauk SSSR 451:537–540
14. Anfilogov V, Khachay Y (2013) Evolution of core and silicate envelopes at heterogeneous accumulation of the Earth. Lithosphera 4:146–153 (in Russian)
15. Khachay Y, Anfilogov V (2009) The temperature distribution numerical models for the Earth envelopes at its accumulation stages. Geodynamics. Deep structure. The Earth's thermal field. Geophysical fields interpretation. In: Proceedings on 5th conference on behalf of Bulashevich. Ekaterinburg, pp 520–522 (in Russian)
16. Khachay Y, Anfilogov V (2009) The variant of the temperature distributions in the Earth on its accumulation. The study of the Earth as a planet by methods of geophysics, geodesy and astronomy. In: Proceedings on 6th Orlov conference. Kiev, pp 197–202
17. Anfilogov V, Khachay Y (2012) Differentiation of mantle material at the Earth's accumulation and the initial crust formation. Lithosphera 6:3–13 (in Russian)
18. Khachay Y, Anfilogov V (2013) Numerical results of Earth's core accumulation 3-D modeling. EGU Geophys Res Abs 15:51
19. Anfilogov V, Khachay Y, Antipin A (2014) Numerical 3D simulation of temperature distribution at the stage of the Earth's accumulation and peculiarity of its initial minerageny. Ural Geophys Bull 1:81–85 (in Russian)
20. Khachay Y, Anfilogov V, Antipin A (2014) Numerical results of 3-D simulation of Moon accumulation. EGU Geophys Res Abs 16:1011
21. Khachay Y, Antipin A (2014) The results of numerical 3D simulation of temperature distribution in the lunar envelopes at its accumulation. Monitoring. Nauka i Technologii 1:28–32 (in Russian)

Chapter 2
The Solar System

Abstract A very brief overview is presented of current conceptions about the state of the interstellar medium, interstellar clouds, and the role of the supernova in their evolution. The main information available on the physical parameters of the Solar system is presented and there is a discussion of the modern conceptions about the processes that brought about the formation and evolution of the Solar system.

Keywords Interstellar medium · Supernovas · Formation of the solar system · Role of short-living radioactive isotopes

2.1 Interstellar Clouds

Any model which attempts to describe the process of formation of the planetary systems must proceed from the definite initial condition of the material state from which the stars and the planetary systems were formed. At the present time, it is assumed that such state coincides with the state of the interstellar gas–dust clouds [1]. The following data, according to [2], are the basis of that assumption:

1—Young stars exist in surroundings of gaseous and gas-dust clouds. Many of them have massive optically dense disks with diameters from 10 to 1000 AU and a life time from 1 to 10 million years; 2—at least 4 % of the main part of the stars has planetary systems. The masses in these systems have values from 1 to 10 Jupiter masses and orbital radii from 0.05 to 5 AU. Below, we give a brief data overview of the sizes, composition and particles concentration in the interstellar clouds and the possible transformation mechanisms of scattered cloud matter into stars. The main characteristics of molecular clouds and the interstellar medium surrounding them are presented in Table 2.1.

The density of the matter in the core of molecular clouds is 10^3–10^5 molecules/cm^3 [4]. 50 different molecules are detected in the gaseous phase of the molecular clouds. These are mainly hydrogen, H_2O, CO and relatively small molecules of organic compounds up to ethanol: C_2H_5OH. About 10 % of the Galaxy mass is concentrated

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Table 2.1 The main dynamical characteristics of the interstellar media [3]

Phase	Indicator	Temperature (K)	Concentration/cm^3	mass %	Volume %
Molecular clouds	CO, OH	10–60	10^{-2}–10^{-7}	40	1/2
Clouds HI	λ 21 cm	50–100	1–50	40	5
Intercloud media	λ 21 cm	7000	0.2	20	40
Clouds H II	Hα, OIII	10^4	10^{-2}–10^{-3}	Low	Low
Coronal gas	O VI	10^{-5}–10^{-6}	10^{-3}–10^{-4}	0.1	50

in the molecular interstellar clouds [5]. Besides the gaseous phase, the interstellar clouds contain dust particles with sizes around 0.1 μ, the composition of which is not defined, and their masses part is equal to 1 %.

According to Evans, the molecular clouds can be divided into two groups. In group A, Evans classifies clouds in which the temperature is lower than 20 K, and in group B are clouds, in some areas of which the temperature is higher than 20 K [1]. The physical conditions in clouds A and B are presented in Table 2.2.

It is presumed that stars are formed in the cores of massive clouds or in the whole volume of small clouds. Evans distinguished three stages of the evolution of massive cloud cores [1].

1. At low temperature and high density the cloud begins to compress. The cloud density increases, but the temperature remains low at this stage. With growing density, the number of collisions with dust particles also grows and the kinetic temperature becomes linked with the dust temperature.
2. If at the first stage a massive proto-star occurs, it will heat the remaining matter in the core dust.
3. The third stage passes with high temperature and low densities. At this stage, this can occur only through the formation of a very massive star, when an area with low density but high temperature forms [1]. From this it is inferred that if the initial material from which stars and planets form is molecular clouds, the material must necessarily evolve through a high temperature gas condition.

Table 2.2 Physical characteristics of molecular clouds [1]

Characteristic	Group A		Group B	
	Envelope	Core	Envelope	Core
T K	10	10	10–20	20–70
ncm^{-3}	10^2–10^3	10^4	10^3?	10^5–10^6
M/Mo	20–?	?	10^3–10^5	10^2–10^3
MJ/MO	8–24	2	?	2–4
Strong IR radiation	No	No	No	Yes
n(HCO+)/n	?	2×10^{-9}	?	2×10^{-11}

2.2 The Role of the Supernova in the Formation of the Solar System

The role of the supernova in the formation of planet systems of the solar type is determined by two factors. The supernova explosion, which is accompanied by the ejection of a tremendous mass of material, can be a trigger mechanism for collapse of the molecular cloud. Furthermore, the molecular cloud, from which a system like the Solar system could be formed as a result, must contain short-living isotopes and heavy elements, which are absent in the molecular clouds and which could be formed only by the supernova explosion [6]. Therefore, the question about the source of these elements and the process of their formation is very significant for any model of the formation of the Solar system.

Herbst and Assusa [5] and Shramm [6], as the main proof of supernova participation in the formation of a solar type system, indicated the presence in the Solar system of an anomalous concentration of ^{26}Mg, which was formed by the decay of the short-living isotope ^{26}Al. In [6], the following scenario was suggested. According to Shramm [6], the universe existed and evolved for 7–15 billion years before the processes began that lead to the formation of the Solar system. It is assumed that the interstellar material contained gas and dust and heavy elements entered into the dust composition, which had been formed earlier by supernova explosions. The Galactic spiral wave passed through the cloud every 10^9 years, stimulating the formation of stars. One or more massive stars were formed when this wave passed through the gas and dust, then, in the vicinity of the present Solar system, one or some stars were formed in the cloud, and quickly evolved and exploded. The ejected material had a composition that was specific to the r-process. It is possible that this material was ejected in the form of jets [6]. The next spiral wave, which passed through the gas-dust cloud 4.6 billion years ago, again initiated the formation of stars, which were introduced during the evolution process into the proto-solar cloud ^{26}Al together with ^{16}O and other anomalous infusible elements. It is important to note that ^{26}Al could not be present in any clouds whose age was more than 10 million years.

The physical characteristics of the matter which had been ejected by the explosion of the supernova will differ significantly from the initial material of the molecular cloud. Firstly, the concentration of particles from that release could be higher than in the molecular cloud, and secondly the ejected material will have an initial temperature such that it will be in the gaseous state and, during its cooling, equilibrium condensation of the solid phases could be possible.

The works of [7] are devoted to the problems of the sources from which the short-living isotopes were introduced into the Solar system. In the works of Ouellette et al. two models are considered: the trigger model and aero gel model. The trigger model assumes that the explosion of the supernova was the mechanism that triggered the collapse of the molecular cloud, into the core of which the short-living isotopes were introduced. This model provides the development of the process, according to the second scenario, which is discussed in the work of Herbst [5]. The trigger model has been considered in more detail by Boss [4]. It is evident that this model makes sense only if, after the enrichment of the molecular cloud by short-living isotopes and heavy elements, the sun, as a star of small dimensions, and around the Sun a proto-planetary disk, could be formed. In that case, the interval between the supernova explosion and the solid fragments of CAI occurring, enriched by ^{26}Al and Ca, must be no more than 1 million years [4].

The aero gel model is based on the assumption that the Sun is a star that has a small mass, formed in the direct vicinity of the massive stars cluster [4]. It is suggested that the population of stars existed at a distance of less than 2000 AU from the Sun and that 70–90 % of the proto-stars were formed in that cluster [8]. The short-living isotopes were synthesized into the supernova from that cluster and injected into the proto-sun cloud as dust particles. Williams and Gaidos [9] and Gounelle and Meibom [10] showed that the possibility of the injection of these elements in the 1 million years after a supernova explosion is less than 1 %. Nevertheless, Ouellette et al. consider that this possibility is enough to explain the origin of short-living isotopes and especially ^{60}Fe in the Solar system [7], with the use of the aero gel model. From that, the question arises: for what reason was it necessary to inject short-living isotopes into the proto-Sun cloud from the outside, if we can assume that the proto-Sun cloud itself represents material that could be ejected by the supernova? It seems that the delivery of these elements from the different sources assumed in different models does not solve this problem, because besides the short-living isotopes of aluminum, manganese and elements of the iron group in the Solar system, heavy elements also exist, which could occur only after a supernova explosion.

A significant factor that defines the possibility of the Solar system forming from the material ejected by the supernova explosion is the distribution of short-living isotopes in the volume of the proto-sun cloud. It is evident that, if that assumption is right, the distribution of ^{26}Al in the proto-sun cloud must be uniform and the value of the ratio of ^{26}Al/^{27}Al must be near to the value 5×10^{-5} [11].

The basis of the conclusion about the heterogeneous ^{26}Al distribution in the proto-sun cloud are the values of the ratio of ^{26}Al/^{27}Al, obtained for the fragments of CAI enriched by Ca, and Al inclusions in carbonaceous chondrites, as presented in [12].

But these data [11] do not provide a unique conclusion about the heterogenic distribution of ^{26}Al in the proto-solar cloud. The value of the ^{26}Al/^{27}Al ratio can differ from the canonic one, which is 5×10^{-5}, for two reasons. Firstly, condensation of the corundum and other elements enriched by Al that enter the composition of CAI occurs over 1–3 million years and the later the condensed mineral was formed, the less would be the ratio in it of ^{26}Al/^{27}Al. Secondly, corundum and gibbonite, for which ratios are measured, are often associated with spinel, and part of the ^{26}Mg can move to spinel from these minerals by diffusion. This is in agreement with the finding that the ratio of ^{26}Al/^{27}Al with the minimum values can be obtained by analyzing specimens with the minimum values of the ratio of ^{27}Al/^{24}Mg [12].

2.3 Physical Parameters of the Solar System

The Solar system has the following most important physical characteristics.

1. The planetary rotation around the Sun occurs in the same direction as the Sun's rotation.
2. All planets rotate round their axis in the same direction as the Sun rotates, except for Venus and Uranus.
3. The planets are distributed around the Sun non-randomly. The planetary distances from the Sun can be described by empirical equations, which have been obtained by different authors.
4. The planets are divided into two groups: 1—the outer planets, which have large masses and whose compositions are close to the solar composition; 2—the inner planets, with masses of 0.45 % of the Solar system mass and whose compositions significantly differ from the composition of the rest of the Solar system.
5. 98.7 % of the Solar system mass is concentrated in the Sun and only 1.3 % of the mass is concentrated in the planets, whereas 98 % of the angular momentum belongs to the outer planets.

The first mathematical approximation of the relation for the distances of the planets from the Sun is known as the Titsius-Bode law, which is as follows:

$$R_{n+1}/R = b,$$

where *n*—planetary number, *b*—empirical parameter. The values of *b* for different planets are presented in Table 2.3 [13].

From the table it is seen that the value of this parameter is not constant, but varies: $1.4 < b < 2$. Furthermore, many researchers have attempted to relate the planetary distances from the Sun with the process of planetary formation [13]:

Schmidt attempted to refine the empirical relation of Titsius-Bode. His equation was as follows:

$$\sqrt{R_n} = a' + b'.$$

This equation satisfactorily describes the planetary distances from the Sun, if the parameters a′ and b′ can be determined separately for the outer planets and planets of the Earth's group (Tables 2.4 and 2.5) [14].

Table 2.3 The values of the parameter b in the Titsius-Bode equation [13]

Planet pairs	b	Planet pairs	b	Planet pairs	b
Venus–Mercury	1.87	Asteroids (Cercera)–Mars	1.82	Uranus–Saturn	2.00
Earth–Venus	1.38	Jupiter–Asteroids	1.88	Neptune–Uranus	1.58
Mars–Earth	1.52	Saturn–Jupiter	1.84		

Table 2.4 The distance between the outer planets and the Sun, calculated using the Schmidt equation [14]

Planet	Rtheor. AU	Rfact. AU	Discrepancy (%)	R Titsius-Bode AU
Jupiter	5.20	5.20	0	5.20
Saturn	10.76	9.54	13	10.0
Uranus	18.32	19.19	−5	19.6
Neptune	27.88	30.07	−7	38.8
Pluto	39.44	39.52	0.2	77.2

Table 2.5 The distance between the inner planets and the Sun, calculated using the Schmidt equation [14]

Planet	Rtheor. AU	Rfact. AU	Discrepancy (%)	R Titsius-Bode AU
Mercury	0.39	0.39	0	0.4
Venus	0.67	0.72	−7	0.7
Earth	1.04	1.00	4	1.0
Mars	1.49	1.52	−2	1.6

Besides the relations presented above, the planetary distances from the Sun can be accurately described by the logarithmic spiral equation, suggested by us:

$$R = 0.225(\exp(0.172\varphi)), \tag{2.1}$$

where φ—angle of the ray turning round the polar spiral; R—distance of the planet from the Sun, AU.

It should be noted that Eq. (2.1) does not describe the planetary space location, only the values of their orbits. Equation (2.1) represents the empirical dependence of the planetary distances from the Sun, identical to Titsius-Bode, but it is possible that the spiral distribution of the planets has a genetic sense. More on the assumption of the vortex structure of the proto-planet cloud is suggested in the models of Weizsäcker and Kouper [15].

2.4 The Mass Material Distribution in the Solar System

An important characteristic of the Solar system is the relationship between the masses of the planets and their distances from the centre of the system [14].

Moreover, the composition of the terrestrial group planets differs in principle from the composition of the outer planets and the Solar system as a whole. That could be understandable if these differences had been presented for the outer planets, but the assumption about the intrusion of alien material into the inner part of the Solar system is not real. Also unreal is the assumption about the initial lack of the mass in the proto-planetary cloud in the supply area of growing inner planets, and about the difference in composition of the other part of the cloud.

The alternative of the initially heterogenic composition of the proto-planetary cloud is a variant that contains an assumption of the loss by the inner planets of the major part of their material during the process of the formation of the Sun and outer planets. This includes the fact that the material from the supply area was redistributed in opposite directions: between the Sun and Jupiter.

The next step consists in determining the composition of the material that was lost by the inner planets, and it's mass. We assume that the composition of the cloud from which the Solar system was formed corresponds to the abundance of the chemical elements in the cosmic space. The main element in this distribution is hydrogen, which amounts to 2.8×10^8 atoms per 10,000 atoms of Si [16]. Let us further suppose that hydrogen and helium were the main elements ejected from the proto-planet cloud in the area of the inner planets during the formation of the Solar system. For comparison, we can take the iron element Fe. In the cosmic material, to every 1 g-atom of Fe is related 0.67×10^5 g-atom of H and 0.5×10^4 g-atom of He. The Earth contains 34.0 mass% of Fe, which corresponds to 6.07×10^{-3} m_o g-atoms (m_o—Earth's mass). The amount of g-atoms H and He, for which the initial material of the Earth must contain that amount of Fe, is equal to 1.1×10^3 m_o. That value is comparable with the mass of the material, which must be at a distance, on which the Earth is located. It is impossible to explain the loss of such an amount of material by the planets of the terrestrial group through the process of the dissipation of volatile components from the atmospheres of the formed planets.

2.5 The Nature of the Asteroids Belt

The active role of Jupiter in the formation of the Solar system is evidenced by the fact that, during the formation process, there occurred a significantly sharp boundary, which was fixed in space by the asteroids belt. The nature of that boundary in all models of planetary formation is not discussed. We can assume that it in fact presents a zone of gravitational equilibrium between the Sun and Jupiter, which existed during the process of simultaneous growing. The location of this zone was not fixed. The Sun's gravitational pull grew more quickly than that of Jupiter, and therefore the gravitational equilibrium zone moved in the direction from the Earth to the belt of asteroids. We can assume that Mars, during its formation, was in the area of Jupiter's attraction. This makes the nature of asteroids belts more understandable. They can be considered as pre-planetary bodies for the planets, which were formed at the initial stage of the planets' formation [17]; they occurred in the zone of gravitational equilibrium, and they could not form together as a planet, whose location would be between Jupiter and Mars.

2.6 The Possible Mechanism of the Formation of the Sun and Planets

As mentioned in the chapter above, in discussing the role of the supernova explosion in the formation of the Solar system, we concluded that the presence of short-living isotopes of aluminum, iron and manganese was evidently the basis for the formation of the Solar system from material thrown out by the supernova explosion. That assumption radically changed our ideas about the mechanism of the formation of the Sun and the planets of the Solar system. Firstly, the duration of the period when the main mass of the initially dissipated proto-solar material was concentrated into the growing Sun decreases sharply. That period is no more than 10^6 years, which is the time after the supernova explosion up to the occurrence in the near-solar space of the material of CAI. That velocity of the Sun's material accumulation could be achieved, probably, only in the case that the material ejected by the supernova explosion was in the form of a jet, which was transformed into a spiral vortex by the action of gravitational and electromagnetic forces, as a result of which the material moving along the spiral became concentrated in its centre. Secondly, because the proto-planetary material was in a gaseous state, the initial and boundary conditions that were in the models of the planetary formation from the gas–dust cloud [13, 18, 19] were changed. It can be assumed that the proto-planetary clots of the material could appear in the form of turbulent perturbations in the spiral vortex, from which the Sun formed. This is proved by the fact that the largest satellites of Jupiter are located on the branches of the logarithmic spiral. Thirdly, the formation of these proto-planetary clots is also supported by the distances distribution of Jupiter's satellites, which is described by the logarithmic spiral equation. Thirdly, the proto-planetary clots were also formed over a period of 10^6 years, and in that time interval the main mass of the material that had been in a gaseous state in the area of the formation of the terrestrial group planets, was absorbed by the growing Sun. The suggested variant of the formation of the Solar system does not disagree with the data that Chambers presented as the reason for the formation of the Solar system from the dust molecular clouds [2]. It must be noted that only about 4 % of the main part of young stars have planetary systems, which is evidence that the conditions in which planetary systems can be formed are very rare. Therefore the existence of young stars having gas–dust disks, but having no planets, can be considered as a situation in which the necessary conditions were not realized. An additional argument, which satisfies the possibility of the Solar system's formation from the supernova material, is the fact that the planetary systems could have stars no older than $\sim 10^6$ years, from which it flows that the average velocity of the material accumulation by the central star forming in such systems would achieve 10^{-7} times the Sun's mass per year [2]. Taking into account that the relation of the accumulation velocity to time has a maximum, and that the maximum value of the velocity at that point can be significantly larger, than its average value, therefore the most Sun's mass could be absorbed during the first million years from the moment of the supernova explosion.

The most complicated question for models of planetary formation from the initial homogeneous near-solar disk is to explain the characteristics of the Solar system's distribution of angular moment. Because the Sun together with the proto-planetary disk is an isolated system, it has to fulfill internally the condition of moving moment conservation, or a mechanism for transformation into another form of energy must exist. One possible variant to explain the distribution of the angle moment can be suggested based on the mass distribution of the Solar system. From that distribution, it follows that the large material mass that was initially located in the interval between the Sun and the Earth was absorbed by the Sun. Together with this mass, also transferred into the Sun was kinetic energy, which was transformed into thermal energy and was then used for the heating of solar material. Therefore it can be assumed that the values of 98 % of the moving moment, which are presently linked with the moving of the outer planets, are really only equal to 2 % from the initial moment of the proto-planetary cloud. The remaining 98 % was transformed into thermal energy by the formation of the Sun.

References

1. Evans NI (1982) The formation of stars in molecular clouds. In: Moroz VI (ed) Protostars and planets. The studies of star formation and of the origin of the Solar system. Mir, Moscow, pp 171–185 (in Russian)
2. Chambers J (2003) Planet formation. In: Davis A, Holland HD (eds) Treatise on geochemistry. V.1. Meteorites, comets and planets, pp 461–473 (2003)
3. Clayton D (1982) Interstellar clouds. In: Moroz VI (ed) Protostars and planets. The studies of star formation and of the origin of the Solar system. Mir, Moscow, pp 18–52 (in Russian)
4. Boss AP (2007) Evolution of the solar nebula. VII. Spatial and temporal heterogeneity of short-lived radioisotopes and stable oxygen isotopes. Astrophys J 660:1707–1714
5. Herbst W, Assouza G (1982) The role of supernova on the star formation and spiral structure. In: Gehrels T (ed) Protostars and planets. The studies of star formation and of the origin of the Solar system. University of Arizona Press. Russian version. Mir, Moscow, pp 421–429
6. Schramm D (1982) Supernova and the solar system formation. In: Gehrels T (ed) The studies of star formation and of the origin of the Solar system. University of Arizona Press. Russian version. Mir, Moscow, pp 440–457
7. Ouellette N, Desh S, Bizzaro M et al (2008) Injection mechanisms of short-lived radionuclides and their homogenization. Geochim et Cosmochim Acta 73:4946–4962
8. Lada C, Lada E (2003) Embedded clusters in molecular clouds. Ann Rev Astron Astrophys 41:51–115
9. Williams J, Gaidos E (2007) The similarity of supernova enrichment of proto-planetary disks. Astrophys J 663:L33–L36
10. Gounelle M, Meibom A (2008) The origin of short-living radionuclides and astrophysical environment of solar system formation. Astrophys J 548:781–792
11. MacPherson G, Huss R, Davis A (2003) Extinct ^{10}Be in the type A CAI from CV chondrites. Geochim et Cosmochim Acta 67:3165–3179
12. Makide K, Nagoshima K, Krot A et al (2013) Evidence for corundum-bearing refractory in carbonaceous chondrites. Geochim et Cosmochim Acta 110:190–215
13. Vitjazev A, Pechernikova G, Safronov V (1990) The Earth's group of planets. Nauka, Moscow (in Russian)
14. Zharkov V, Trubitsin V (1980) Physics of the Earth and planets. Nauka, Moscow (in Russian)

15. Urey H (1956) The origin of the Earth. In: Faul H (ed) Nuclear geology. Moscow, pp 437–489 (in Russian)
16. Voitkevich G, Miroshnikov A, Povarennikh A et al (1977) Handbook on geochemistry. Nedra Press, Moscow (in Russian)
17. Anfilogov V, Khachay Y (2005) A possible scenario of material differentiation at initial stage of the Earth's formation. Doklady Earth Sci 403A:944–947
18. Schmidt O (1957) Four lectures on the Earth's origin theory. Academy of Sciences of USSR Press, Moscow (in Russian)
19. Safronov V (1969) Evolution of the proto-planetary cloud and formation of the Earth and the planets. Nauka, Moscow (in Russian)

Chapter 3
The Model of the Earth's Heterogeneous Accumulation

Abstract This chapter presents an overview of the main conditions demanded by the experimental data that the models of the Earth's formation have to satisfy. We suggest a new model of the Earth's heterogeneous accumulation. The results of the numerical modeling, presented in this chapter, of the temperature distribution in the pre-planetary body allow us to determine that the heat released due to the decay of short-living radioactive elements can provide the melted state in the inner parts of bodies with sizes greater than 50 km.

Keywords Heterogeneous accumulation · Short-living isotopes · Experimental data · Numerical modeling · Pre-planetary body

3.1 Models of the Earth's Formation

The principal moment for all hypotheses on the formation of the Solar system as a whole and of Earth in particular is the question: did the Sun and planets form from the very high temperature gaseous clouds or from cold clouds composed of gas and solid dust particles. On this basis, two variants of the hypothesis have been researched in detail: homogeneous and heterogenic accumulation (accretion) of the planets. In the homogeneous accumulation variants, which were suggested in the 1950s–1970s, it is supposed that the Earth and other planets were formed from a homogeneous mixture of cold dust particles containing iron and silicate minerals, where the average composition of these minerals corresponds to the composition of ordinary chondrite [1]. In these variants, it is considered to belong to meteorite matter, which, it was assumed, has a composition close to the composition of the material from which the planets were formed in the solar nebula, before planetary formation. Note that the homogeneous accumulations were created during a period when data about the age of the meteorite material and about the presence in the Solar system of short-living isotopes, particularly ^{26}Al, were absent.

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Based on the fact that, in the homogeneous accumulation models, the initial matter was cold and the gravitational energy released during the planetary accumulation process had dissipated, the Earth, in these models, formed as a homogeneous cold body (Fig. 3.1).

It is assumed that, after the end of the accumulation process over 10^9 years, there then took place the heating of the Earth and the division of the matter into the iron core and silicate mantle [1]. The most serious negative statements countering these hypotheses of homogeneous accumulation are given in [3].

It is not evident where so much iron in the planets, and of which their cores consist, originated. For example, the mass of the core of Mercury core is 60–70 % relative to the planet's total mass.

1. The composition of Ni in the silicate part of the mantle is much greater than it should be for these rocks to be in equilibrium with the Fe-Ni material of the Earth's core.
2. It is unknown what the energy source was that could melt the iron during the creation of drip congestion, which could then sink to the Earth's centre.
3. There is no adequate mechanism that allows the transfer of iron, which was initially uniformly distributed, into the Earth's volume and towards its centre.

In [4], one of the suggested variants of the differentiation of the Earth's matter is presented, where all the problems linked with the Earth's division into the iron core and silicate mantle become evident. It is very difficult to find a mechanism that allows the gathering of uniformly distributed particles of iron into lenses and then emits them towards the Earth's centre.

The alternative to the homogeneous hypothesis is the hypothesis of the Earth's heterogenic accumulation. The first such hypothesis was proposed by Eucken, who

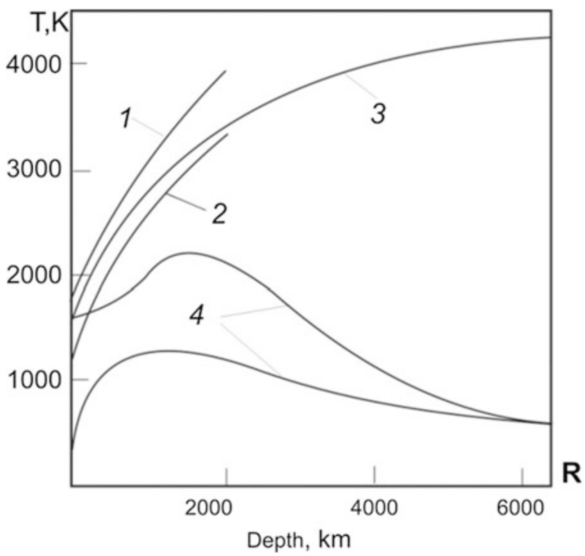


Fig. 3.1 The Earth's temperature distribution after the end of accumulation. *1*—the temperature of the full mantle material melting; *2*—the temperature of the solid mantle; *3*—the melting iron temperature; *4*—the temperature distribution in the Earth according to Safronov [2] (*lower curve*), and according to Ringwood (*upper curve*) [1]

supposed that the iron had first been condensed in the proto-planetary cloud, after which the silicates were condensed. Eucken considered that these sequences allow us to explain the zonal structure of the Earth [5]. The heterogenic accumulation is possible under the following condition: it occurs simultaneously with the condensation and the composition of the condensate changes according to a defined sequence. That assumption was made by Turekian and Clark in [6]. They suggested the following sequence of condensation, which led to the stratified formation of the Earth's structure.

1. First, the iron, which the Earth's core contains, condenses.
2. The outer layers of the Earth are more oxidized than the inner layers, because during the process of cooling the composition of the proto-planetary gas changed, possibly due to the loss of hydrogen. According to the sequence of condensation, the relatively oxidized material never mixed with the rest of the core's material.
3. H_2O, CO_2 and other volatile components are concentrated near the Earth's surface (in the crust and upper mantle), which did not require the degassing of the whole Earth's volume for their extraction.

Turekian and Clark considered that the initial temperature in the proto-planetary cloud was more than 2000 K and the pressure was from 10^2 to 10^5 PA [6].

From the discrepancies in the heterogenic accumulation hypothesis, which were considered in [6], we should note the main points. According to the data of Grossman [7], iron does not condense first into the proto-planetary cloud as assumed by Turekian and Clark, and instead minerals with a high concentration of CaO and Al_2O_3 condensed first, and thus the central part of the growing Earth had to consist of these phases. Neither of the heterogeneous accumulation models can explain how this material was replaced by iron or how the iron core formed. Furthermore, the ideas of heterogeneous accumulation of the Earth were considered by many researchers [8–11], but for various reasons were rejected [1].

3.2 The Sequence of Mineral Condensation from the Gaseous Phase and Accumulation of the Solid Phase

In this section, we shall consider two processes: the condensation of solid phases from the gaseous phase and accumulation of the solid particles formed during the condensation process, which led to the formation of pre-planetary bodies. By cooling of the proto-planetary cloud formed from the resulting material after the supernova explosion, in principle, two alternative scenarios are possible: (1) dust particles are formed during the condensation process, and contain the whole spectrum of solid phases; only after mixing with the dust material and forming a cloud with a homogeneous composition does the process of forming planetesimals begin; (2) the process of accumulation of dust particles and the formation of planetesimals begins at the moment when the highest temperature solid phase occurs in the cloud. The possibility of simultaneous condensation and accumulation of the solid particles and the formation of the pre-planetary body decisively defines the mechanism of the planetary formation of the Solar system.

The thermodynamic calculation of the equilibrium condensation sequence of the gaseous proto-planetary material was performed by Grossman, and is reported in Table 3.1 [12] and Fig. 3.2 [4].

The calculations made later did not make any significant corrections to that sequence and only gave more accurate values of the condensation temperatures of the minerals contained in the CAI, enriched by intrusions of calcium and aluminum in the carbonaceous chondrite [13].

The proof that the planetary formation had been according to the second scenario lies within the data, that there are iron meteorites that do not contain other phases apart from melt of Fe-Ni, and the existence in the carbonaceous chondrite minerals of CAI, enriched by intrusions of calcium and aluminum, consisting of the high temperature condensation products: corundum, gibbonite, perovskite, melilite, spinel and anorthite [14]. Condensation of these phases at a pressure of 10^6–3×10^5 bar occurred at the temperature range of 1760–1360 K.

The components of CAI in the carbonaceous chondrite present themselves as separate inclusions in the low temperature matrix; therefore, they could not be formed in the parental body. At the same time, they could be formed only in the case that the agglomeration of the dust particles, of which these minerals consist, occurred during the moment of condensation, while other phases that did not belong to the CAI composition in the proto-planetary cloud were also absent. The mineral composition of coarse-grained CAI is presented in Table 3.2.

Table 3.1 Temperature of mineral condensation from the gas phase [12]

Mineral	Mineral formula	Temperature of condensation beginning (K)	Temperature of the condensation end (K)
Corundum	Al_2O_3	1758	1513
Perovskite	$CaTiO_3$	1647	1393
Melilite	$Ca_2Al_2SiO_7$	1625	1450
Spinel	$MgAl_2O_4$	1513	1362
Iron	(Fe, Ni) + Co + Cr	1471	–
Diopside	$CaMgSi_2O_6$	1450	–
Forsterite	Mg_2SiO_4	1444	–
Anorthite	$CaAl_2Si_2O_8$	1362	–
Enstatite	$MgSiO_3$	1349	–
Rutile	TiO_2	1125	–
Alkaline feldspar	$NaAlSi_3O_8$ $KAlSi_3O_8$	~1000	–
Troilite	FeS	700	–
Magnetite	Fe_3O_4	405	–

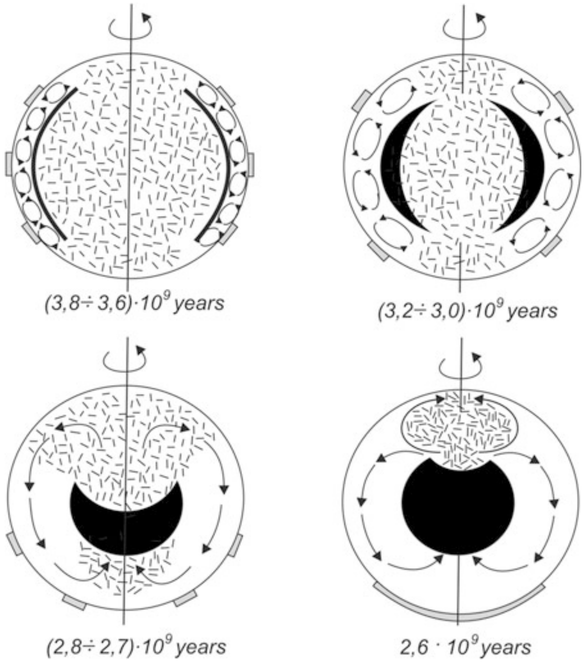


Fig. 3.2 Mechanism of core formation in the initial homogeneous Earth according to Sorokhtin and Ushakov [5]

Table 3.2 Dependence of the temperature in the centre of the pre-planetary body with the content of Al_2O_3 [15]

Radius of the pre-planetary body (km)	Concentration of Al_2O_3 mass%	3.0	4.6	9.0
	1.0			
	Temperature in the pre-planetary body centre (K)			
5	1240	1701	1734	1825
100	1676	1752	1812	1978
150	1690	1793	1876	2104
200	1701	1828	1928	2206
250	1711	1856	1972	2290
300	1718	1878	2006	2359
400	1730	1912	2059	2461

The structure of the inclusions is evidence of the fact that these minerals are products formed by crystallization of the melt [16].

Presently it is determined that the geochemical variations of the elements in the iron meteors IIIAB and IVAB are also the result of crystallization of the melted iron cores into single parental bodies [17]. That is also only possible if there was a differential condensation of the iron, which led to the formation of parental bodies, in which the greatest volume was that of iron, without any inclusions of other solid phase.

Currently, the view on the processes of planetary formation and the meteorite matter of the Solar system has changed significantly. It was established that the most ancient crystalline material in the Solar system are CAI, inclusions that were discovered in the carbonaceous chondrite. The age of CAI, defined by different methods, is estimated at 4567–4568 million years [18–20]. The use of the ^{182}Hf-^{182}W isotopic method with the refined, as compared with earlier data [21], value of ε^{182}W has allowed a significantly more precise definition of the age of the iron meteorites matter [17, 22, 23]. A significant step has been made in defining the possible energy source that was necessary for heating the CAI material and iron meteorites to above their melting temperature. The modeling calculations fulfilled in [24] showed that for the ratio, equal to $^{26}Al/^{27}Al = 5 \times 10^{-5}$ [25] as the result of decay of ^{26}Al, bodies of asteroid size can be heated to higher temperatures than 2000 K. It is noted [26] that the most significant role of that thermal source was during the first million years of the accumulation of proto-planetary bodies. The data presented above has allowed the formulation of two important conclusions: 1—at least in the initial stage of Earth's formation, the planetesimals occurred according to the regime of heterogeneous accumulation; 2—the iron meteorite material, which, as it was proposed, also passed through the stage of melting, was formed 1–3 million years after the formation of CAI [17, 22, 23]. This leads to a set of models with two stages of planetary formation, according to which the first stage during the heterogenic accumulation saw the formation of significantly large planetesimals and the separation in them of a significant part of the core matter. These planetesimals, during the first 3 million years, were the building blocks for the proto-Earth. The final stage of the accretion of the inner planets, according to the authors of these models, was the result of collisions with large bodies with iron cores, and these were comparable in size with Mercury and Mars [27]. It is assumed that, as a result of these collisions, the Earth was partly or wholly melted, which created the conditions for establishing equilibrium between the material of the crust and the mantle. The variants of the two-stage models of the Earth's formation are represented in [23, 27, 28, 30]. Below, we shall consider a principally different model for the two-stage formation of the Earth, which takes into account that, due to the collisions of the initial pre-planetary bodies, heated up to the melted state, the amount of the potential energy sharply increases, and is transformed into the thermal energy of the growing Earth, as a result of which the process of heterogenic accumulation took place, caused by temperatures higher than the melting temperature of the matter of the core and the silicate mantle.

3.3 The Initial Earth's Pre-planetary Bodies

The abovementioned peculiarities of the composition and structure of CAI allow us to assume that, in the area of the formation of the terrestrial group of planets at the initial moment, planetesimals were formed, consisting of the minerals of CAI, and these planetesimals appear as the initial pre-planetary bodies for the planets [15]. The possible structure of the initial pre-planetary bodies is discussed in [13].

The absence of metallic Fe in CAI composition (Table 3.1) allows us to suggest that the central parts of the initial pre-planetary bodies contained minerals that had been condensed at temperatures between 1760 and 1500 K. The middle part of the pre-planetary bodies may be composed of metallic iron Fe and silicate material, which had condensed at temperatures between 1500 and 1350 K. Thus, the amount of iron decreases from the centre to the outer parts. At the same time, the existence of iron meteorites is evidence that at a definite temperature range the condensation of Fe-Ni material occurred without the silicate mineral admixture, while in the middle part of the initial pre-planetary bodies a Fe-Ni layer could be formed.

The high concentration of ^{26}Al and possible ^{60}Fe in the initial pre-planetary bodies [30] creates the condition for their heating up to melting temperature, which can be reached in their central parts, which are composed of material of CAI. Merk et al. were the first to draw attention to this fact and carried out numerical modeling of the heating of an asteroid size body caused by the energy of ^{26}Al decay [24]. The authors adopted the following initial conditions. Heating of the growing pre-planetary bodies up to a temperature higher than the temperature of the surrounding medium began after the time value t_0, when the radius of the pre-planetary body became $R_{p0} \sim 1$ km. The initial $^{26}Al/^{27}Al$ ratio at $t = t_0$ was taken as equal to 5×10^{-5} [25] and the corresponding energetic contribution was calculated using the following relation:

$$Q\ (t) = Q_0 \exp[-\lambda(t + t_0)].$$

It is assumed that the velocity of the growth of the pre-planetary bodies is described by the linear function of time and that for the pre-planetary bodies their radius R_{pmax} became 100 km over about 1 million years. Calculations were performed for the values $5 < R_{pmax} < 100$ km and for the time values $0 \le t_0 \le 4t_{1/2}$, where $t_{1/2}$ is the time of ^{26}Al half-decay, equal to 0.720 million years. The temperature distribution in a body with radius R = 100 km, without taking into account the growth velocity (at the top) but taking into account the linear relation of growth velocity from the time (at the bottom), is shown in Fig. 3.3.

The authors of the work [24] solved a particular problem: revealing the possible source of energy necessary for heating and metamorphism of meteorite material in small asteroid bodies (Fig. 3.4).

We obtained similar results for the initial pre-planetary bodies of the planets, whose central parts were composed of material resembling CAI, the mass of which grows over time, according to the Safronov model [2]. The temperature distribution in the growing pre-planetary bodies is shown in Fig. 3.4.

For the calculations, the following values of the thermal parameters were used: heat capacity, $c = 1 \times 10^3$ J/(kg grad); heat conductivity at standard conditions, $\lambda_0 = 2$ W/(m grad).; heat of melting, $L = 4 \times 10^5$ J/kg; energy of heat allocation ^{26}Al, $q = 1.5 \times 10^{-7}$ W/kg; the period of half-decay was 7.2×10^5 years [31]. The temperature reached in the centre of a pre-planetary body of a given radius depends decisively on the concentration of Al_2O_3 (Table 3.2).

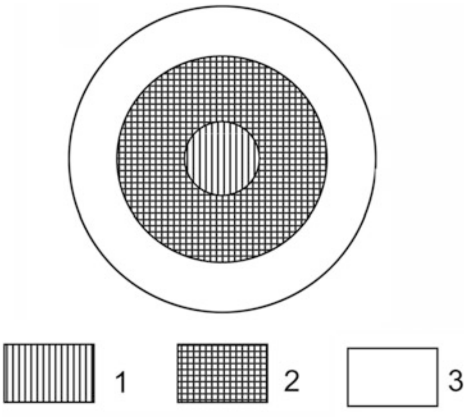
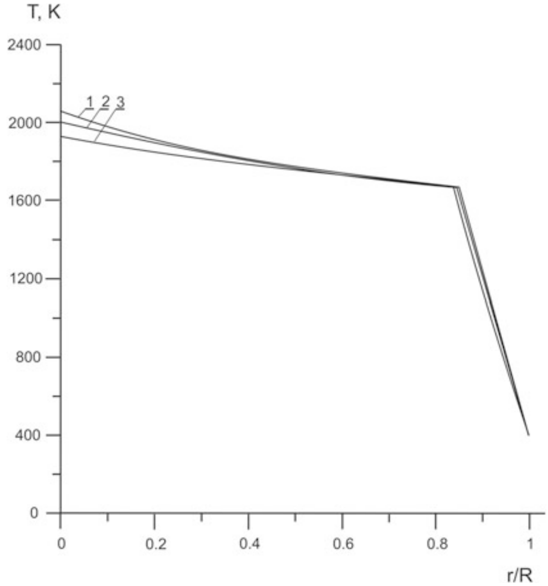


Fig. 3.3 Possible scheme of the initial Earth's pre-planetary body structure according to the model of temperature distribution. *1*—the core of the pre-planetary body is composed of CAI minerals; *2*—the middle part of the envelope is composed of iron; *3*—the outer part of the solid envelope

Fig. 3.4 Distribution of the temperature in the growing pre-planetary body. Its radius is: 400 km—*1*; 300 km—*2*; 250 km—*3*



References

1. Ringwood A (1982) Origin of the Earth and Moon. Nedra, Moscow (in Russian)
2. Safronov V (1969) Evolution of the protoplanetary cloud and formation of the Earth and the planets. Nauka, Moscow (in Russian)
3. Kuskov O, Khitarov N (1982) Thermodynamics and geochemistry of the Earth's core and mantle. Nauka, Moscow (in Russian)
4. Sorokhtin O, Ushakov S (1991) The global evolution of the Earth. Moscow University Press, Moscow (in Russian)
5. Miyake Y (1965) Elements of geochemistry. Muruzen, Tokyo
6. Turekian K, Clark S (1969) Inhomogeneous accumulation of the Earth from the primitive solar nebula. Earth Planet Sci Lett 6:346–348
7. Grossman L (1972) Condensation in the primitive solar nebula. Geochim et Cosmochim Acta 36:597–619
8. Anders E (1968) Chemical processes in the early solar system, as inferred from meteorites. Account Chem Res 1:289–298
9. Anderson D, Sammis C, Jordan T (1972) Composition of the mantle and core. In: Robertson E (ed) The nature of the solid Earth. McGraw-Hill, New York, pp 41–66
10. Cameron A (1973) Accumulation processes in the primitive solar nebula. Icarus 18:407–450
11. Voitkevich G, Miroshnikov A, Povarennikh A et al (1977) Handbook on geochemistry. Nedra Press, Moscow (in Russian)
12. Dodd R (1986) Meteorites. A petrologic-chemical synthesis. Mir, Moscow (in Russian)
13. Davis A, Richter F (2003) Condensation and evaporation of Solar system material. In: Davis A, Holland HD, Turekian KK (eds) Treatise on geochemistry. Meteorites, comets and planets, vol 1, pp 407–427
14. Grossman L, Larimer J (1974) Early chemical history of the solar system. Rev Geophys Space Phys 12:71–101
15. Anfilogov V, Khachay Y (2005) A possible scenario of material differentiation at initial stage of the Earth's formation. Dokl Earth Sci 403A:954–947
16. Stolper E (1982) Crystallization sequences of Ca-Al inclusions from Allende: An experimental study. Geochim et Cosmochim Acta 46(11):2159–2180
17. Shersten A, Elliot T, Hawskesworth C et al (2006) Hf-W evidence for rapid differentiation of iron meteorite parental bodies. Earth Planet Sci Lett 241:530–542
18. Amelin Y, Krot A (2007) Pb isotopic age of the Allende chondrules. Meteorit Planet Sci 42:1321–1337
19. Connelly R, Amelin Y, Krot A et al (2008) Chronology of the solar system's oldest solids. Astrophys J 675:L121–L124
20. Krot A, Amelin Y, Bland P et al (2009) Origin and chronology of chondritic components. A review. Geochim et Cosmochim Acta 73:4963–4997
21. Harper C, Jacobsen S (1996) Evidence for ^{182}Hf in the early solar system and constraints in the timescale of terrestrial accretion and core formation. Geochim et Cosmochim Acta 60:1131–1153
22. Kleine N, Mezger K, Palme H et al (2005) Early core formation and late accretion of chondrite parent bodies: Evidence from ^{182}Hf-^{182}W in CAIs, metal rich chondrites and iron meteorites. Geochim et Cosmochim Acta 69:5805–5818
23. Kleine T, Touboul M, Bourdon B et al (2009) Hf-W chronology of accretion and early evolution of asteroids and terrestrial planets. Geochim et Cosmochim Acta 73:5150–5188
24. Merk R, Breuer D, Spohn T (2002) Numerical modeling of ^{26}Al-induced radioactive melting of asteroids concerning accretion. Icarus 159:183–191
25. Lee T, Papanastassiou T, Wasserburg G (1976) Demonstration of ^{26}Mg excess in Allende and evidence for ^{26}Al. Geophys Res Lett 3:109–112
26. Pechernikova G, Vitjasev A (2005) Impacts and the early Earth evolution in catastrophic action of cosmic bodies. In: Adushkin VV (ed) Academkniga Press, Moscow (in Russian)

27. Walter M, Tronnes R (2004) Early Earth differentiation. Earth Planet Sci Lett 225:253–269
28. Chambers J, Wetherill G (1998) Modeling of the terrestrial planets: N-body integration of preplanetary bodies in three dimensions. Icarus 136:304–327
29. Agnor C, Capur R, Levison H (1999) On the character and consequences of large impacts in the late stage of terrestrial planets formation. Icarus 142:219–237
30. Nichols R (2000) Short-living radionuclides in meteorites: constraints on nebular time scales to the production of solids. Space Sci Rev 1–2:113–122
31. Ghosh A, Sween H (1998) The thermal model for the differentiation of asteroid 4 Vesta, based on radiogenic heating. Icarus 143:187–206

Chapter 4
Formation of the Earth's Core

Abstract This chapter describes the physical model that is used for numerical solution of the problem of temperature distribution in the forming core of the growing Earth. It defines the composition of the Earth's core, on which depends, in particular, the changes during the process of accumulation and the melting temperature distribution from the pressure. There is discussion of the conditions that describe the energy balance on the surface of the growing planet. The obtained variants of the numerical temperature distribution in dependence on the values of the model parameters are presented. All variants of the solution are described for the moment when the core ceases to grow, the formation of the inner solid core and melted outer core. It is shown that different conditions ensure either the solid state or the melted state of the layer that forms at the bottom of the forming mantle.

Keywords Content of the initial core matter · Numerical modelling · Variants of the temperature distribution · Temperature in the core

4.1 Mechanism of the Earth's Core Formation and Its Possible Composition

The possibility of the initial Earth's pre-planetary bodies heating up to the melting temperature of CAI allowed us to propose a principally new mechanism of the Earth's core formation [1]. The sense of it is as follows. In the initial stage of the Earth's formation in the proto-planetary cloud, the initial pre-planetary bodies of the Earth were formed with sizes of few hundred kilometers, the central parts of which had been heated and melted. Together with the phases enriched with Al and Ca, iron was also melted, and was mainly located in the middle part of the pre-planetary bodies (Chap. 3, Fig. 3.4). According to our supposed model, the iron–nickel of the Earth's core was formed by the collisions of the initial pre-planetary bodies. Due to the fact that the main volume of the initial pre-planetary bodies was in the melted state, their collisions were inelastic. From those inelastic collisions, the

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fusion of the pre-planetary bodies occurred, the fragments of iron resulting from the collisions were combined, and they then displaced from the central parts of the initial pre-planetary bodies the dense melted aluminum-silicate material, which together with the fragments of the solid cold envelope was thrown into the planetary supply zone. As a result of these collisions, new pre-planetary bodies were formed, in which the central parts were composed of the melted iron–nickel material (Fig. 4.1).

According to geophysical data, the modern distribution of the density of the Earth's core requires that, besides iron, it must also include a considerable amount of light components. In our variant, in the initial stage of the process of heterogenic accumulation, core growth occurred from the materials which arrived on the surface of the growing Earth from the supply zone at low pressure. Therefore in the core composition there can be only those components that were by that time present in the supply zone and that were dissolved in substantial amounts in the melted iron by

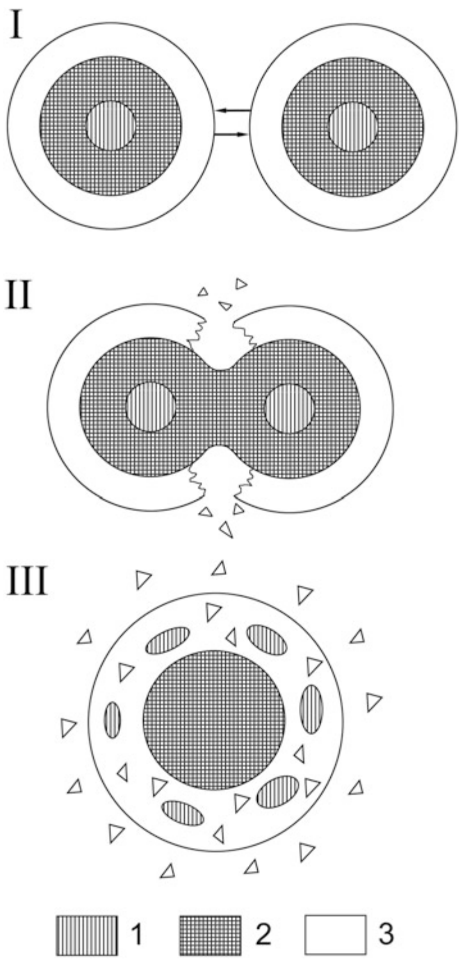

Fig. 4.1 Scheme of a secondary pre-planetary body with melted core formation by collision of two initial pre-planetary bodies. *1*—the core of the pre-planetary body is composed of CAI minerals; *2*—the middle part of the envelope is composed of iron; *3*—the outer part of the solid envelope

low pressure. The main light component in the core is FeO, according to the assumption of [2]. In the initial stage, when the initial pre-planetary Earth's bodies are formed, the composition of the iron–nickel material corresponds to the composition of iron meteorites, in which the composition of FeO is negligibly low. Troilite, which is considered as a possible light component of the core [2] and which exists in iron meteorites, formed in them later than iron, and it is not in an equilibrium state with the iron [3]. Therefore, the entry of considerable amounts of sulphur into the Earth's core during the heterogeneous accumulation is problematical.

One of the problems that arise from discussion of the question about FeO entering into the core composition is the problem of the possibility of achieving in the proto-planetary cloud the value of partial oxygen pressure necessary for forming FeO. A priori that value is not known. In the literature there are many variants for obtaining the required value P_{O2}, including the participation of water [2]. Actually, that problem is solved by nature itself. The existence in olivine pallasites up to 30 % of the fayalite molecule [4] is evidence that the iron oxide condensed from the gaseous phase in the form of olivine together with the metallic iron. From melting of the olivine in the melt, free iron oxide is formed, which is not linked with SiO_2. It occurs as a result of partial decay of the fayalite component [5]:

$$Fe_2SiO_4 = FeSiO_3 + FeO$$

The relation between the content of the free iron oxide FeO and the melt composition in the MgO-FeO-SiO_2 system is shown in Fig. 4.2 [5]. If the olivine melt is in an equilibrium state with the melted iron, the part of free FeO will dissolve in the iron and the equilibrium of that reaction will shift to the right. The conditions of FeO dissolving in the melted iron were investigated in detail by Othani et al. [6]. According to their estimation, for the core density to correspond with the geophysical data, the content of the iron oxide in the core has to reach a value of 40 mol% FeO [6]. By the melting temperature of olivine and by low pressure, the solubility of FeO in the melt of Fe does not exceed 4.0 mol%, but that value sharply increases as the temperature increases [6]. The high solubility of iron oxide is achieved by the increase of pressure [6]. This is because the molecular volume of the FeO dissolved in the melted Fe is 3.8 cm^3 less than in the melt of FeO [6]. Nevertheless, the possibility of dissolving 40 mol% FeO in the melted core is far from obvious. The problem lies in the fact that the main olivine mass, from which the iron oxide FeO enters into the core, is concentrated in the surface layer of initial pre-planetary bodies and as a result of these conditions it is impossible to provide the uniform distribution of FeO in the middle iron–nickel envelopes. It can be distinguished only in the formation of secondary pre-planetary bodies, when, as a result of collisions, mixing of the material of the initial pre-planetary bodies occurs. With the radius of the secondary pre-planetary bodies reaching more than 200 km, the necessary solubility of FeO in the core is achieved due to high internal pressure.

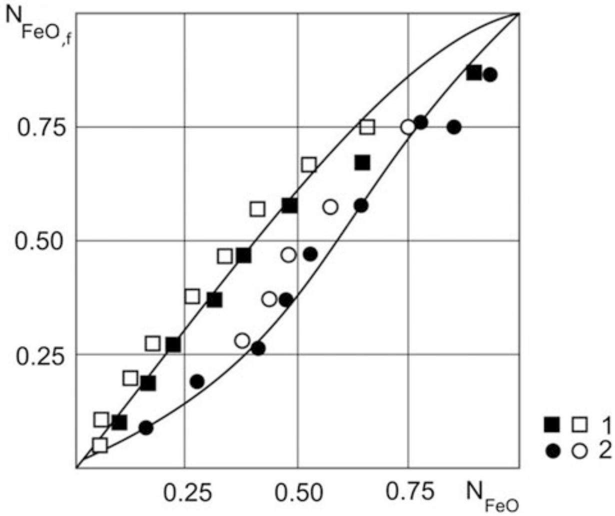

Fig. 4.2 The dependence of free iron oxide content in MgO-FeO-SiO_2 system melt on the mole fraction of FeO. Melt composition: *1*—(0.5MgO-0.5SiO_2) – 1.0FeO; *2*—(0.2MgO-0.8SiO_2) − 1.0FeO; 2173K—*black*, 1873K—open symbols

Additional information about the possible composition of the Earth's core can be obtained from the density distribution in the present core. From the curve of the core's radial distribution density (Fig. 4.3), three intervals can be distinguished.

For the interval from 6370 to 5600 km, the density gradient is equal to 0.0002 g/cm^3 per km. For the interval from 5600 to 4800 km, it increases to 0.43 g/cm^3 per km and finally, for the interval from 4800 to 3000 km, it increases to 1.04 g/cm^3 per km. The subsequent density distribution allows us to propose that the core for the interval from 6370 to 5600 km has a constant composition and contains the iron–nickel melt without light components. The density change for that interval is stipulated by the dependence of the specific volume of Fe-Ni alloy on pressure. In the interval from 5600 to 4800 km, iron oxide can appear in the composition of the inner core. The compressibility of this material increases with increasing FeO content. The strong increase of the density gradient in the third interval is brought about because in that interval the material is in the melted state and its compressibility is significantly greater than that of solid matter. This interpretation of the radial distribution of density agrees well with the given mechanism of the formation of the Earth's core.

Thus, the process of the formation of the Earth's core can be divided into three stages: 1—formation of the initial pre-planetary bodies, in which the middle envelope is mainly composed of melted iron; 2—formation of the secondary pre-planetary bodies, in which the Fe–Ni melt is located in the central part of the new pre-planetary bodies; 3—combination of the secondary pre-planetary bodies into

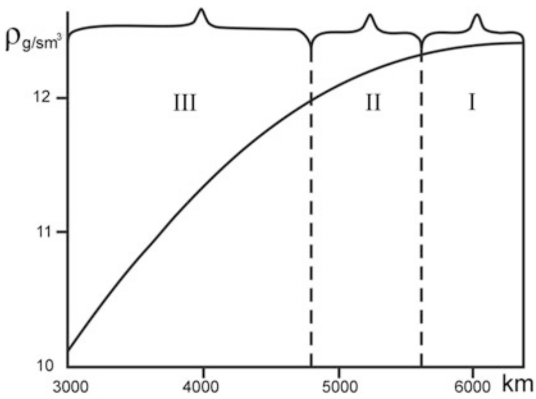

Fig. 4.3 Radial density distribution in the modern earth's core. *I*—the central part of the inner core, composed of Fe-Ni alloy without FeO; *II*—outer part of the inner core, which consists of a different quantity of FeO; *III*—outer melted core area

one growing planet. At the third stage, the core continues to increase due to the iron melting, which reaches the Earth's surface as meteorite content and sinks to the core's surface as melted iron. Based on the average iron composition in meteorites H, L and LL, at that stage about 30 % of the present core mass passed into the core. The role of the meteorite material will be considered in more detail in the Chap. 5.

4.2 The Temperature Regime During the Process of the Earth's Core Growing

Further evolution of the growing Earth depends on the temperature that was reached in the core. We can estimate it by the following. The temperature distribution in the body of the increasing radius is obtained from the numerical solution of the boundary value problem for the system of heat equation with a convective term, the balance equation of impulse, mass, gravitation potential, and the equation of the Stefan problem for phases of boundary shifting [1]. In this stage the solution can be derived using the 1-D model taking account of the possibility of melting appearing without explicit location of the crystallization of the front boundary, and taking parametrical account of convective heat transfer in the melting zone [7, 8]. On the surface of the growing body we assign the conditions that provide the balance between the incoming part of potential energy of gravitational bodies interaction, heat expenditure for heating of the falling matter, and thermal energy radiating to space, taking account of the transparency of outer space. The separation of the proto-planetary material into the metallic and silicate components that occurred in the stage of core growth must be taken into account in the mathematical modelling of the core's thermal regime. The concentration of short-living isotopes decreases at the final stage of core formation and the energy contribution from their decay becomes minor.

Mathematical modelling of the thermal evolution of the growing planet is based on the above scheme of the process. For the velocity of the growth of the planet's pre-planetary body, from the model of Safronov in variant [9], the following equation is used:

$$\frac{\partial m}{\partial t} = 2(1 + 2\theta)r^2\omega(1 - \frac{m}{M})\sigma \tag{4.1}$$

where: ω—angular velocity of the orbital motion, σ—surface density of the matter in the "supply" area of the planet, M—modern mass of the planet, r—the radius of the growing pre-planetary body, q—statistical parameter which takes into account the distribution of particle masses and velocities in the supply zone. The temperature distribution in the body, whose radius is increasing, was discovered from numerical solution of the boundary problem for the heat equation, with consideration of the possibility of melt occurring, but without taking into account the crystallization front position and parametric counting of the convective heat transport in the melt [10, 11]:

$$c_{ef}\rho\frac{\partial T}{\partial t} = \nabla(\lambda_{ef}\nabla T) + Q \tag{4.2}$$

where: c_{ef}, λ_{ef}—the effective values of the heat capacity and thermal conductivity, T—temperature at the desired point in time t, Q—volumetric capacity of internal heat sources. On the surface of a growing body of given conditions, to ensure the balance between the incoming part of the potential energy of the gravitational interaction of bodies, loss of heat to heating of the sediment material, and radiated heat into space with the transparency of the environment:

$$k\rho\frac{GM}{r}\frac{dr}{dt} = \varepsilon\sigma[T^4 - T_1^4] + \rho\, c_P[T - T_1]\frac{dr}{dt} \tag{4.3}$$

where: ρ—density of the matter, G—gravitational constant, M—mass of the growing planet, r—its radius. T and T_1—correspondingly, the temperature of the body boundary and outer space, ε—coefficient of medium transparency, c_p—specific heat capacity, k—the part of the potential energy transformed into heat. The velocity of change of the radius according to the chosen density model is defined from (4.1). At present, the more perfect expressions are suggested for the Earth's accumulation time [12]. They provide a qualitative similarity with (4.1) and give the best correspondence with the given data, but the calculation of Δr leads to additional technical complexity, which is why we use (4.1) at each step.

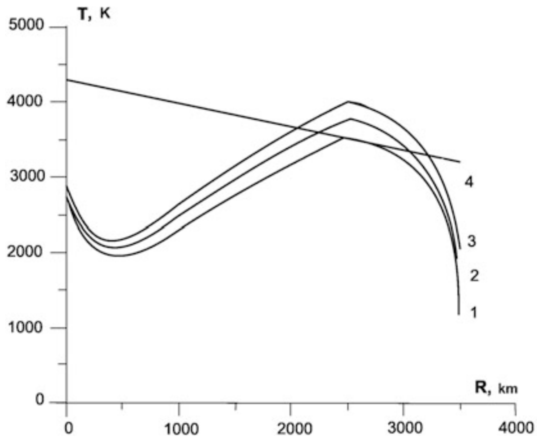

Fig. 4.4 Variants of the temperature distributions in the growing earth's core [7]. *1*—k = 0.3 for iron proto-core; k = 0.01 in silicate mantle. k is varied linear for the interval 2500–3500 km. *2*—k = 0.4 for iron proto-core; k = 0.02 in silicate mantle. *3*—k = 0.4 for iron proto-core; k = 0.03 in silicate mantle. *4*—Dependence of the melting temperature on pressure in the modern core

The problem is solved by use of the constant step for the space grid $\Delta r = 100$ m, and not a uniform step for the time grid, which is defined from (4.1) for given Δr. We used the implicit, conservative, stable scheme, with the second approximation order for the space coordinate and with the first order for the time coordinate. For the successive time moments, at which the planet radius grows in the step Δr, we define the distributions of temperature, lithostatic pressure, adiabatic temperature, melting temperature, viscosity, and effective thermal conductivity, linked with the solid state value through the Nusselt number. It is assumed that in the area with a developed convection there are realized not very critical flows, for which Nu $\leq$ 2. That corresponds to the variant in which the temperature distributions in convective areas give their upper estimation.

Some possible variants of the temperature distribution in the core, when the Earth's size is equal to 3500 km, are shown in Fig. 4.4.

The main difference between these variants and the variants given before consists in the existence of the minimum values of T corresponding to 400–500 km. By that time, the value of energy being released decreases significantly with ^{26}Al decay. At the same time, as the mass of the proto-planet increases, the amount of the kinetic energy of falling accumulated bodies and particles also increases. In the final stage of the core accumulation, account is taken of the decrease of the part of the energy transformed into thermal energy, which is due to the part of the solid silicate component of the collision bodies, which leads to the significant decrease in temperature of the forming layers.

The possible variants of the temperature distribution up to the moment when the planet size reaches 6300 km are shown in Fig. 4.5.

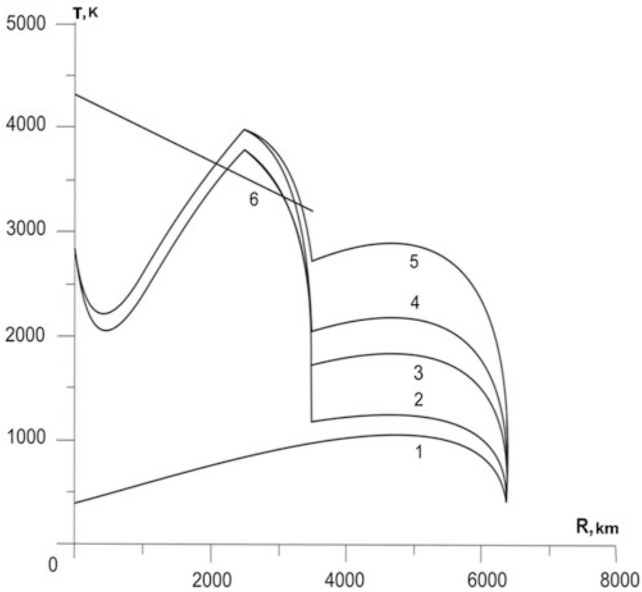

Fig. 4.5 Possible variants of the temperature distributions up to the moment when the earth is 6300 km in size [8]. *1*—The accumulation from small particles. The generation of heat by short-living radioactive elements is not taken into account. k = 0.001 for growing core; k = 0.001 for mantle; *2* and *3*—the generation of heat by short-living radioactive elements is taken into account; *2*—k = 0.3 for growing core; k = 0.02 for mantle; *3*—k = 0.4 for growing core; k = 0.01 for mantle; *4*—k = 04 for growing core; k = 0.02 for mantle; *5*—k = 0.5 for growing core; k = 0.05; *6*—The distribution of the core's melting temperature with depth by lithostatic pressure [13]

As seen from the results presented in Fig. 4.1, the temperature distribution is determined by the heat output caused by the decay of short-living radioactive isotopes only in the initial stage, and up to the moment when the radius reaches 300 km. The further energy balance depends on the part of the potential gravitation energy that results from collisions of the accumulated bodies, on the heat, and on the other part, which is lost by radiation. According to the mechanism described above in the presented variants, the differentiation is explicitly on account of the fact that, in the growing stage of the greater part of the iron core, the collisions that occur are practically inelastic and most of the potential energy is transformed into thermal energy. In the final stage of the core's growth, the pre-planetary body can already retain the outer brittle envelope of the collision bodies. The impact becomes more elastic, which is explained by the decrease in the part of the potential energy used for heating. The presented results show that, by the end of core formation, the temperature distribution in the variants obtained in accordance with the experimental relation of the iron's melting temperature, the Fe–FeO mixture, and the pressure [14], support the melting state of the outer core and the solid state of the inner core.

References

1. Anfilogov V, Khachay Y (2005) A possible scenario of material differentiation at initial stage of the Earth formation. Dokl Earth Sci 403A:954–947 (in Russian)
2. Ringwood A (1982) Origin of the earth and moon. Nedra, Moscow (in Russian)
3. Park F, Bunch T, Masalski T (1966) A study of the silicate and other phases in Campo del Cielo meteorite. Geochim Cosmochim Acta 30:399–414
4. Mason B (1962) Meteorites. Wiley, New York (in Russian)
5. Bobylev I, Anfilogov V (1986) The calculation of the free form metal oxide concentration and the structural parameters of poly component silicate melts. Izv USSR Ser Met 2:23–27 (in Russian)
6. Othani E, Ringwood A, Hibberson W (1984) Composition of core 3, II. Effect of high pressure on solubility of FeO in melted iron. Earth Planet Sci Lett 71:94–103
7. Khachay Y, Anfilogov V (2009) The temperature distribution numerical models for the Earth envelopes at its accumulation stages. In: Geodynamics. Deep structure. The earth's thermal field. Geophysical fields interpretation. Proceedings 5th conference on behalf of Bulashevich, Ekaterinburg, pp 520–522 (in Russian)
8. Khachay Y, Anfilogov V (2009) Variant of temperature distributions in the Earth on its accumulation. In: The study of the earth as planet by methods of geophysics, geodesy and astronomy. Proceedings 6th Orlov conference, Kiev, pp 197–202 (in Russian)
9. Safronov V (1969) Evolution of the protoplanetary cloud and formation of the earth and the planets. Nauka, Moscow (in Russian)
10. Samarsky A, Moiseenko B (1965) The economic scheme of transparent account Stefan problem. J Calculating Math 5:816–827 (in Russian)
11. Tikhonov A, Liubimova E, Vlasov V (1969) About the evolution of fusion zones in the thermal history of the earth. Dokl Acad Sci USSR 188:338–342 (in Russian)
12. Pechernikova G, Vitjasev A (2005) The impacts and the early earth evolution by catastrophic action of cosmic bodies. In: Adushkin V (ed) Academkniga Press, Moscow, pp 251–265 (in Russian)
13. Stacey F (1969) Physics of the Earth. Wiley, New York
14. Boehler R, Chopelas A, Zerr A (1995) Temperature and chemistry of the core-mantle boundary. Chem Geol 120:199–205

Chapter 5
Formation of the Earth's Silicate Mantle

Abstract This chapter discusses the energy source and the matter content from which the initial Earth's mantle was formed. The PT conditions by which the mantle accumulation is provided are defined. Attention is drawn to the role of chondrites of different composition in the frame of the Earth's heterogeneous accumulation model. The conditions are formulated by which a melted layer at the bottom of the mantle is formed. At the bottom boundary of the layer fraction crystallization occurs. The crystallization of Mg-pyroxene and magnesio-wüstite will lead to the formation at the bottom of the mantle of a layer comprising a mixture of these minerals. We note that, based on seismic data, it can be concluded that, namely by that mineral association, there exists a transition layer "D" on the modern core–mantle boundary. The boundaries between layers shift following the body's growing surface.

Keywords Matter composition · Initial mantle · Mineral association of the transition layer "D" · Shifting of the layer's boundary

5.1 The Sequence of Formation and the Age of Meteoritic Material

Before proceeding to consideration of the forming process of the Earth's silicate envelope, it is necessary to determine the source and composition of the material from which that envelope is formed. The unique information source about the composition of the matter from which planets were formed is the composition of meteorites. So far, the science of meteorites has developed as an independent area of knowledge and the possibility of a direct relation with the process of planetary formation and the sequence of meteorite matter formation has not been considered. This situation arose because, until recently, precise data about the age of meteorite matter did not exist and, without that information, we could not develop such a sequence.

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This led to the situation that meteorites are considered as an initial matter, which provides an independent history and evolution of the formation process.

The most notable model illustrating that approach was proposed by Ringwood [1]. He considered that initially, from the cold dust cloud, a certain parental body formed, which had a carbonaceous chondrite composition [1]. As a result of the secondary heating, that body was melted and destroyed. It is assumed that all kinds of meteorites were formed from the destruction of that body. As a possible sequence, Ringwood considered the following sequence of meteorite formation: initial carbonaceous chondrite (Orguel type) → enstatite chondrite → iron meteorites, pallasites, mesosiderites, and howardites.

Urey suggested that two consistent generations of parental bodies existed [2]. The initial bodies were accumulated by the low temperature and they were of lunar or larger size. The bodies were heated with local sources of melting formation. Iron, iron-stone meteorites and chondrites were formed by the destruction of the initial bodies, and the secondary bodies and chondrites were formed by the accumulation of crushed down initial bodies. The hypothesis of meteorite formation in bodies of asteroid size was also suggested by Fisher [3]. They assumed that the asteroid material, due to the action of accidental sources, was heated up to the melting temperature and differentiated. Unfortunately, neither of these hypotheses suggests the energy source that was necessary for the heating and metamorphism of the meteoritic material. We assume that these hypotheses not only do not correspond to the role that the meteoritic material played in planetary formation, but also do not explain the numerous peculiarities of the mineral composition, structure and thermal history of that material. They could exist, until there would appear certain definitions of the absolute age of the meteorite matter.

The situation radically changed when, by using new isotopic methods, we obtained reliable data about the age of meteorite matter. These data showed that the formation of the meteoritic material had been during the whole period of planetary accumulation and thus not only the age but also the composition of the material from which the meteorite matter had been formed had changed. We must add that numerous meteorites present themselves as poly mix breccias, whose composition contains fragments of meteoritic and planetary material, which formed at different stages of planetary accumulation. The age of these fragments can differ by tens of millions of years [4–6].

A large number of absolute and relative age determinations of CAI, magmatic iron meteorites and chondrites have been made in recent years. An overview of these investigations is given in [7–9]. The most ancient material of the Solar system and the initial accounting point for beginning the process of planetary and meteorite formation is the matter of CAI—enriched by aluminum and calcium inclusions discovered in meteorites from Allende and Efremovka. Their age, defined by the U-Pb method, varies within the limits of 4568.6–4567.2 million years [5, 9, 10]. According to the mechanism of the formation of the Earth's core, that material was part of the composition of the central parts of the initial pre-planetary bodies. As a result of their destruction and the formation of the secondary pre-planetary bodies, that material was thrown into the planetary supply zone. The produced age interval between the age of CAI and iron meteorites corresponds to the time interval during which the initial pre-planetary bodies were formed, in the centers of which was the CAI material, and in which its heating and melting were provided.

The difference between the age of iron meteorites and CAI inclusions according to data of ^{187}Re-^{187}Os isotopes varies from 1.0 to 3.0 million years [8, 11]. For the meteorites of group IIIAB the difference is 2.4 million years [8]. This is evidence that the material of the iron meteorites was condensed from the gaseous phase directly after CAI and was heated and melted simultaneously. According to our model, this happened in initial planetesimals and pre-planetary bodies. The Rb–Sr age of ordinary chondrites, determined from the gross samples, varies from 4590 to 4520 million years for H-chondrites to 4460 million years for the LL group [12]. The detailed investigations of the chondrites age performed by Bouvier and co-authors produced a more detailed picture. They established that the time of the exit of H-chondrites fragments from parental bodies (the authors considered the process as a process of chondrite material cooling to the given temperature) for Santa Marguerite chondrite varies from 4565 to 4556 million years [6]. For the Nadiabondi chondrite, this interval is 4559–4556 million years, while for Forest City chondrite it is 4556–4530 million years [6]. Below we shall show that it corresponds to the model of heterogenic accumulation's accepted composition sequence of chondrite material sediment during the process of the formation of silicate envelopes in the inner planets.

The age of basaltic and cumulative eucrites varies from 4600 to 4400 million years [12, 13]. The material could be formed by the impact of planetesimals with the planetary surface and the ejection of melted material from the planetary surface layer. Interesting data have been obtained for the age of chondrules and matrices of carbonaceous chondrites. It is proposed that the matrix with an age of 50–60 million years is younger than the chondrules and the aggregates [14, 15]. This corresponds with the fact that in the carbonaceous chondrites the matrix composition contains the latest low-temperature condensates, which cemented the fragments of the material formed in the beginning and subsequent stages of the growth of matter and the planets.

Thus, it is possible to construct the following age sequence of meteorite matter formation: material of CAI and iron meteorites → material of ordinary chondrites (H → L → LL) → matrix of carbonaceous chondrites CI. This sequence is in good agreement with the assumption that the condensate composition changed during the process of the formation of the Earth's core and silicate envelopes in accordance with the sequence in [16].

The difference between the ages of chondrules and matrix of carbonaceous chondrites can be considered as the time during which the condensation of the proto-planetary cloud material occurred.

The main unsolved problem for all the early hypotheses is the problem of the energy source that was necessary for heating of the meteorite material in the parental bodies up to a temperature close to melting point.

The assumption of the secondary heating of parental bodies, which is made in all the early hypotheses on meteorite matter formation, is in explicit conflict with the currently accepted hypothesis of initial homogeneous planet formation from cold material. According to the results of calculations, the temperature in as large a body as the Earth, as it formed from a process of homogeneous accumulation, did not exceed 1000 °C [17], which is significantly less than the values at which the melting and metamorphism of the meteorite matter occurred. After that, when as the main energy source in the meteorites, the short-living radioactive isotopes and, above all ^{26}Al began to be considered, some definite progress was achieved [18, 19]. At present, most researchers consider that at least the material of CAI and iron magmatic meteorites passed through a stage of full or partial melting [8]. The problem of the energy source for melting of the parental bodies of ordinary chondrites remains unsolved.

In the frame of the model of heterogeneous accumulation, this becomes the simplest explanation of the formation process of iron meteorites and pallasites. The fragments, which have the composition of iron meteorites, were formed as a result of collisions of initial pre-planetary bodies. Most likely, after these collisions they combined in small bodies, which cooled down to a temperature lower than the melting temperature. Investigations conducted by Yang et al. [20, 21] showed that the real velocities of the formation of the Widmanstätten structure in iron meteorites are significantly higher than the value calculated earlier by the metallographic method [12], and it can be formed in bodies with a radius of 150 km at the initial temperature of 1750 K. According to the sequence of the mineral condensation from the gaseous phase, as given in Table 3.1 (Chap. 3), the iron condensation could occur simultaneously with olivine and, in the initial pre-planetary bodies on the boundary with the solid silicate envelope a layer with the composition near to pallasites could be formed. Pallasites could be formed from this material as a result of the collisions of initial pre-planetary bodies.

The presence of large roundish olivine crystals, which are uniformly distributed in the Fe–Ni matrix, is evidence that this material passed through a full melting state. It is impossible for this structure to form in any other way. The separation of material into iron and olivine compositions occurred in the melting state as a result of the immiscibility of the iron and silicate melts. So it can be concluded that the temperature of the iron–olivine melting mixture was higher than 1900 °C—the melting temperature of olivine. The uniform distribution of olivine crystals in the iron matrix of pallasites and their roundish form suggests that the structure in the bodies was formed when their sizes did not exceed the few hundreds of kilometers. The gravitational acceleration in these bodies was small and therefore with the formation of pallasites the separation of liquid phases by density did not occur.

Thus, as a result of the collisions and destruction of the initial pre-planetary bodies and passing to the planetary supply zone, the main mass of highly aluminous material was thrown out from the central parts of the initial pre-planetary bodies, from which the inclusions of CAI were formed, which were fixed later in the carbonaceous chondrites, and also the fragments of future iron meteorites and pallasites were formed. The further evolution process of the material of which CAI consists will be considered in a later chapter.

One of the problems discussed in the literature devoted to meteorites is the problem of chondrules. Brearley and Jones, the authors of a review devoted to chondrite meteorites, wrote that at present the most popular model is that of chondrules forming directly in the solar nebula [22]. At the same time, they noted that the mechanism of the heating of particles of millimetric size up to the melting temperature in the solar nebula and with gaseous pressure 10^2–10^1 PA is not clear. In reality, it is difficult to suggest that, in vacuum conditions, when the heat transfer and its dissipation are realized only by radiation from a remote source, it can have been possible to heat particles of millimetric sizes to above 1500 °C.

In the frame of the presented model of heterogeneous planetary accumulation, the formation of chondrules differing in composition and structure can be explained as follows. After the main iron mass was concentrated in the core of the growing planet, the lower temperature condensates, diopside, anorthite and enstatite, were deposited on its surface. Together with these phases, the fragments that had been formed by the destruction of the initial pre-planetary bodies and thrown out of central parts of these bodies, iron drops, fragments of pallasite composition and highly aluminous material, fell on the planetary surface. All this was then sediment on the core's surface, which was heated to a temperature significantly higher than the temperature of the silicate liquidus. As a result, between the core and the surface of the growing planet, the layer of the melt occurred, covered by a relatively thin layer of solid loose material. Due to the heat transferred by the convective flow to the bottom of the solid layer, at the surface of the growing planet four thermal zones occurred: 1—full melting zone; 2—partial melting zone; 3—zone of high temperature metamorphism; 4—zone of loose fine-grained material. As a result of the collisions of the planetesimals with the planet in the stage of the silicate envelope formation, the planetary surface was broken and from it was thrown out into the supplying zone material from all four zones: loose products of proto-planetary material condensation from the surface, metamorphic, non-equilibrium material from the deeper level, partially melted material from the area where the fallen material on the planet had melted. At the planetary growing stage, when its mass and the gravitational acceleration were small, the initial velocities at which the fragments thrown out by the collisions passed into the satellites orbits, where new planetesimals had formed, were also slow.

Let us consider the possible mechanism of chondrules formation. The kinetic processes of the melting and crystallization into planetesimals with sizes of a few hundreds of kilometers differ in principle from similar processes which occurred in the upper Earth's mantle and in the Earth's crust. The extreme value of gravitational acceleration in such bodies leads to processes that exist practically in conditions of weightlessness, where the convection melt mixing and the gravitational crystal silt are absent. Therefore, in the agglomerate which was formed during the process of planetary accumulation, the melting process will proceed in the diffusion regime and, first of all, it will proceed at the contacts of the mineral grains, the components of which constitute the most easily melted cotectic mixtures. In those conditions in the agglomerate, local melt zones occur, in which the melt composition is not in equilibrium with the average composition of the agglomerate. Due to the low velocity of diffusion processes of change with the components, drops of non-equilibrium melts can be formed, which are not mixed with each other [23]. A similar picture can be seen from experiments on granite melting by the action on it of a spherical converging shock wave [24]. In the centre of the spherical specimen, where the stress was maximal, a homogeneous glass of granite composition formed. In the intervals 2.25–6.75 cm from the centre, only the grains of feldspar melted and from that the glass formed by hardening preserved the configuration of the angular grains of the feldspar. The melting and cooling velocities in these experiments were so large that the gravitational and diffusion leveling of concentration was absolutely absent.

At the crystallization of cotectic mixtures, for example a mixture of olivine + plagioclase, in small bodies in conditions of weightlessness, a melt zone occurs around the growing crystal of olivine, and is enriched by plagioclase composition, which is typical for the olivine chondrules [22] in the chondrites, As a result of the collisions of the planetesimals with the surface of the growing planet, drops of the melt and these non-equilibrium fragments are thrown out to space and from their rapid cooling chondrules are formed, which then combine, forming the meteorite agglomerates.

The best demonstration of this conception is the composition of carbonaceous chondrites. The CO and CV chondrites are characterized by a large concentration of high-temperature phases and by a great variation of their composition. These types of chondrites can be considered as the transition composition differences between the ordinary and carbonaceous chondrites. In the CM chondrites the high-temperature phases account for 33–50 % and are represented by olivine, pyroxene, and Ca–Al silicate glass and also by gibbonite, perovskite and spinel [22]. The composition of the matrix of CM chondrite is similar to the composition of the CL chondrites. In the chondrites the matrix accounts for 99 % of its volume and it is composed of the lowest-temperature condensates: aqueous silicates, magnetite and carbon in organic compounds.

Interesting results that throw light on the origin of carbonaceous chondrites are presented in the work of Tomeoka and Ohnishi [25]. Researching the hydration of the chondrules and the matrices of CV chondrite of Mokoja, they established that the chondrules, which are composed of olivine and enstatite, and the rings around the chondrules are hydrated significantly more than the matrix, which is composed of olivine, enstatite, diopside and hedenbergite–andradite aggregate. Tomeoka and Ohnishi came to the conclusion that the hydration of chondrules occurred not in the meteorite that contains these chondrules, but in another parental body and in conditions that were different from these, which had existed in the carbonaceous chondrite Tomeoka. To that it must be added that the main mineral that forms into chondrites by the hydration of olivine and pyroxene is saponite—hydro mica, which forms in the Earth's conditions during the processes of rock weathering of the main composition in the presence of liquid water at temperatures higher than 0 °C. Hence, the parental body, in which hydration occurred, had to be significantly large to be able to retain the water in the liquid state, and heated so that the water would be able to stay in that state. Bodies in which these conditions could exist need to have a size similar to or larger than the moon, when their surface contained the material layer, close to the composition of carbonaceous chondrites, and it seems that the material that was returned to the Earth as meteorites had been thrown out from the surfaces of these bodies.

5.2 Possible Composition of the Earth's Mantle

Differentiation of the material formed during the process of the Earth's mantle formation depends on the P-T conditions in the growing planet and on the material of which the mantle is composed. The possible temperature distribution in the core and on its surface up to the moment when the mantle matter sediment began is shown in Figs. 4.5 (Chap. 4) and 5.1. Definition of the proto-planetary material composition from which the silicate Earth's envelope was formed is the next problem. Meteorites are the unique source of information about the composition of this material and for us it is important to recover from them the sequence of the sediment process on the growing Earth's surface. The definitions provided in recent years of the absolute age of the meteorite matter show that the processes of planetary and meteorite formations occurred synchronously.

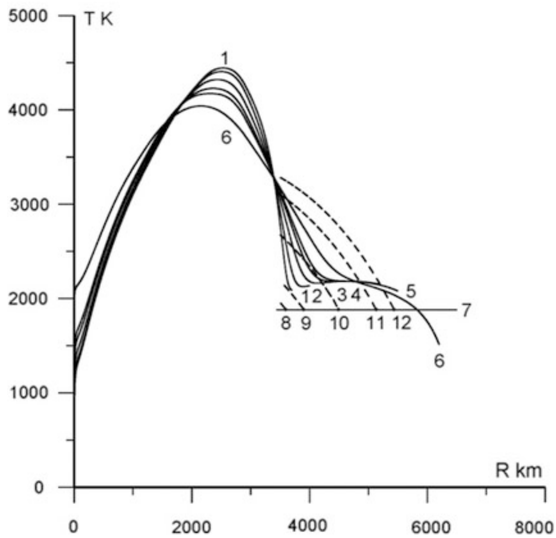

Fig. 5.1 Temperature distribution in the growing Earth at the moment when its size reached the determined radius (1–6): *1*—3500 km, *2*—3900 km, *3*—4300 km, *4*—4600 km, *5*—5600 km, *6*—6200 km; *7*—Liquidus temperature of KLB-1 at normal condition; *8–12*—the dependences of KLB-1 liquidus temperatures on the thickness of the silicate envelope for radiuses indicated in points 1–6

Considering the process of the heterogeneous Earth's accumulation and defining the role of the meteorite matter in that process, we have to keep in mind that the accumulation was not only a sedimentary process of the condensed material of given composition on the surface of the growing planet. It was accompanied by the destruction of the planetesimals by their collisions with the growing Earth and with each other, and by the throwing out of matter from the planetary surface. From this material in the planetary supply zone new planetesimals and parental meteorite bodies were formed. Besides material thrown out from the surface of the growing planets, the composition of parent meteorite bodies also included condensation products, which occurred in the proto-planetary cloud with the lowering of the temperature. Thus, during the process of planetary heterogeneous accumulation, a dynamic equilibrium was constantly fulfilled between the composition of matter, which had been sediment on their surfaces, and the sequence of meteorite matter formation.

This allows us to reinterpret the age intervals that had been determined for the iron meteorites and ordinary chondrites. Most researchers consider the time interval between the age of CAI and iron meteorites to be 1–3 million years as the time during which planetesimals cores of asteroid size were formed [8, 26, 27]. In our model that interval corresponds to the time during which the initial Earth's pre-planetary bodies were formed and destroyed [28]. At this time, as a result of the destruction of initial pre-planetary bodies, into the Earth's supply zone was thrown the main mass of mostly silicate material, including solid envelopes of initial pre-planetary bodies, and the matter from which the parental bodies of iron meteorites, mesosiderites and H-chondrites were formed. The main part of the core was formed simultaneously as the result of the combining of melted iron envelopes of the initial pre-planetary bodies and the formation from them of the secondary one. As the growing Earth's mass increased, on its surface the silicate material began to sediment and form the silicate envelope that covers the core. As a result, the throwing out of melted iron from the growing Earth and the formation of iron meteorites of that generation practically ceased, while the process of the core's formation continued. The time during which this process occurred can be defined by the relative age of the ordinary chondrites. As expected, the H-chondrites, which could have been formed from the material of silicate envelopes of the initial and secondary pre-planetary bodies, appeared 3–5 million years after CAI occurred and continued to form for up to 15 million years. L- and LL-chondrites appeared 13–15 million years after CAI and continued to form for up to 25–40 million years [7]. This data allows us to assume the following sequence of chondrite formation: H-chondrites → L-chondrites → LL-chondrites.

The material of H-chondrites was thrown out by the initial pre-planetary bodies, heated by the decay of short-living isotope ^{26}Al up to the iron melting temperature. By the time that the L- and LL-chondrites appeared, the isotope ^{26}Al had been practically destroyed. The small initial pre-planetary bodies had to cool down and, therefore, a new thermal source was needed to heat them up to a temperature higher than 1100 K [29]. In our model, L- and LL-chondrites were formed from the material thrown out from the surface of the growing Earth, heated due to kinetic

Table 5.1 Average compositions of ordinary chondrites, peridotite KLB-1 and meteorite Allende

Component	Type H [31, 32]	Type L [31, 32]	Type LL [31]	Peridotite KLB-1 [16]	Meteorite Allende CV-3 [33]
SiO_2	47.0	45.2	42.8	44.5	38.55
Al_2O_3	3.0	2.9	2.7	3.6	4.01
FeO	12.9	17.7	21.5	8.1	23.90
MgO	30.1	28.6	27.8	39.2	29.76
CaO	2.5	2.2	2.0	3.4	2.42
Na_2O	1.1	1.0	0.9	0.3	0.15
FeS	5.6	5.8	5.3	–	–
Fe	17.3	6.7	1.3	–	–

energy transformed into thermal energy by the inelastic collisions of bodies falling on the surface [30]. This allows us to define the composition of the matter from which the Earth's mantle formed.

The average chondrite compositions, without the composition of metallic iron and troilite, are presented in Table 5.1 [18, 31].

Also presented are the composition of pomegranate peridotite KLB-1 [16] and the gross composition of meteorite Allende [33]. From Table 2.1 (Chap. 2) it is seen that all the compositions are very similar, and for modelling of the heterogeneous accumulation formation process of the silicate mantle we can accept the average composition between the peridotite KLB-1 and meteorite Allende.

The diagrams are geometrically similar, in spite of the different compositions of FeO and MgO. This is due to Mg and Fe isomorphism in solid phases. The main difference between these diagrams is the low melting temperatures of the liquidus of the Allende meteorite, which is due to the high content of FeO. By the difference pressure of 30 Gpa, the temperature of the meteorite and peridotite differ by 300 °K. Table 5.1 shows that the meteorite Allende contains no mixed sulphide and silicate liquids, which do not considerably affect the phase equilibrium of the silicate part of the system, because the FeS solubility in silicate melt and in the solid phases is negligible.

5.3 Melting and Crystallization of Material During the Process of the Earth's Mantle Formation

Let us consider the temperature regime of the Earth's silicate envelope formation. It follows from the results shown in Fig. 4.5 (Chap. 4) that for most of the variants for the beginning of mantle formation the temperature on the core surface is higher than the temperature of the liquidus of the model compositions, which are presented by peridotite KLB-1 and meteorite Allende, due to the fact that at the boundary between the core and the forming silicate envelope a layer of silicate melt appears.

The further evolution of the melting and differentiation processes of the mantle's silicate material is determined by the thermal regime of the growing Earth and, above all, by the contribution of the potential energy transformation of planetesimals falling on the Earth's surface into thermal energy. The amount of this contribution depends on the degree of inelastic collisions of the accumulated bodies with the growing Earth's surface.

Figure 5.1 shows the temperature distributions calculated for the times when the radius of the planet achieved given values, together with the temperature dependence of the liquidus melting temperatures of peridotite KLB-1 on the lithostatic pressure in the silicate mantle.

The temperature in the layer of the silicate material on the boundary with the melted core (curves 8 and 9) is greater than the melting temperature of the peridotite (curve 7) and the silicate mantle material begins to melt on contact with the core. The temperature necessary for this melting will be supported by the high thermal conductivity of the core material, as well as the thermal energy generated by collisions of inelastic accumulated bodies. Because of the non-homogeneous heating of the silicate melt, there arises thermal convection, which ensures the effective heat conductivity in the layer. As a result, the upper melting boundary will move to the surface of the growing Earth along the melting curve, similarly to the variant described in [34]. This continues until the heat losses no longer compensate the heat income to its bottom boundary. As a result, a layer of the silicate melt is formed on the core's surface, and is covered by a sufficiently thin solid envelope.

With the increase of the melt layer's thickness and the lithostatic pressure, the melting temperature of the liquidus of the silicate material increases at the core–-mantle boundary. The melting temperature of the liquidus reaches more than 3000 K, when its thickness reaches 800 km. Simultaneously, due to the heat transfer from the inner part of the core to its surface, at that boundary the temperature of the melted core will increase, but since the liquidus temperature of the model peridotite, with increasing lithostatic pressure, will increase faster than the core's temperature, there will come a moment on the core–mantle boundary when the layer of the silicate melt at the boundary with the core will begin to crystallize. The thickness of the silicate melt layer, when the crystallization begins, can be considered as the depth of the magmatic "ocean", which forms during the process of the Earth's mantle formation. For the obtained variants of numerical solutions represented in Fig. 5.1, the thickness of this magmatic "ocean" may be 800–900 km.

The first crystalline phases, which crystallize on the bottom of the magmatic "ocean", according to the accepted model composition and diagrams of phase equilibrium [16, 33], are Mg-pyroxene with the structure of perovskite and magnesio-wüstite. The crystallization of the Mg-pyroxene and magnesio-wüstite will lead to the formation at the bottom of the mantle of a layer comprising a mixture of these minerals. Note that, based on seismic data, it can be concluded that it is just this mineral association that makes up the transition layer "D", located on the core–mantle boundary, as proposed by Saxena and other authors [35–37]. In our variant the composition of this layer is determined by the crystallization sequence of the accepted mantle model composition and the calculated variant of the temperature distribution during the process of accumulation.

A significant moment for the initial stage of crystallization is the distribution of iron between the melt and solid phases. The metallic iron that sediments on the Earth's surface with the chondrite material composition will melt, forming drops of melt liquid with no mixed silicate. By the way it will increase the concentration of iron oxide into the melt that will sediment on the core's surface. According to data in [38], at the pressure of 16 Gpa the solubility of FeO in the melted iron is about 25 mass%. These concentrations are significantly more than the mutual solubility of Fe and FeO at the low pressures at which the iron core formed. Having lower density than the density of the melted core, the iron oxide and the metallic iron, enriched by the iron, can form a melt layer on the core–mantle boundary, which will have a diffusion boundary with the melted core. The possibility of that layer's formation is considered in [37].

References

1. Ringwood A (1982) Origin of the earth and moon. Springer-Verlag, New York (1979). Russian version: Nedra, Moscow (in Russian)
2. Urey H (1956) Abundances of the elements. Rev Mod Phys 28:53–57
3. Fisher D (1987) The birth of the earth. Columbia University Press, New York
4. Gopel C, Manhes G, Allegre C (1994) U-Pb systematic of phosphates from unequilibrated ordinary chondrites. Earth Planet Sci Lett 121:153–171
5. Amelin Y, Connelly J, Zartman R et al (2009) Modern U-Pb chronometry of meteorites: advancing to higher time resolution reveals new problems. Geochim Cosmochim Acta 73:5212–5223
6. Bouvier A, Blichert-Toft V, Moynier F et al (2007) Pb-Pb dating constraints on accretion and cooling history of chondrites. Geochim Cosmochim Acta 71:1583–1604
7. Krot A, Amelin Y, Bland P et al (2009) Origin and chronology of chondritic components: a review. Geochim Cosmochim Acta 73:4963–4997
8. Shersten A, Elliot T, Hawskesworth C et al (2006) Hf-W evidence for rapid differentiation of iron meteorite parent bodies. Earth Planet Sci Lett 241:530–542
9. Kleine N, Mezger K, Palme H et al (2005) Early core formation and late accretion of chondrite parent bodies: evidence from ^{182}Hf-^{182}W in CAIs, metal rich chondrites and iron meteorites. Geochim Cosmochim Acta 69:5805–5818
10. Amelin Y, Krot A (2007) Pb isotopic age of the Allende chondrules. Meteorit Planet Sci 42:1321–1337
11. Kleine T, Touboul M, Bourdon B et al (2009) Hf-W chronology of accretion and early evolution of asteroids and terrestrial planets. Geochim Cosmochim Acta 73:5150–5188
12. Dodd R (1986) Meteorites: a petrologic-chemical synthesis. Mir, Moscow (in Russian)
13. Mittelefehldt D, McCoy T, Goodrich C et al (1998) Non-chondritic meteorites from asteroid bodies. Rev Min 36:1–195
14. Chen J, Tilton G (1976) Isotopic lead investigations of the Allende carbonaceous chondrite. Geochim Cosmochim Acta 40:635–643
15. Tatsumoto M, Unruh D, Desborough G (1976) U-Th-Pb and Rb-Sr systematics of Allende and U-Th-Pb systematics of Orgueil. Geochim Cosmochim Acta 40:617–634
16. Herzberg C, Zang J (1996) Melting experiments on anhydrous peridotite KLB-1: compositions of magmas in the upper mantle and transition zone. J Geophys Res 101(B4):8271–8295

17. Safronov V (1969) Evolution of the protoplanetary cloud and formation of the earth and the planets. Nauka, Moscow (in Russian)
18. Merk R, Breuer D, Spohn T (2002) Numerical modeling of ^{26}Al-induced radioactive melting of asteroids concerning accretion. Icarus 159:183–191
19. Anfilogov V, Khachay Y (2005) A possible scenario of material differentiation at initial stage of the Earth's formation. Dokl Earth Sci 403A:954–947 (in Russian)
20. Yang J, Goldstein J, Scott E (2007) Iron meteorite evidence for early formation and catastrophic disruption of proto-planets. Nature 446(7138):888–891
21. Yang J, Goldstein J, Scott E (2008) Metallographic cooling rates and origin of IVA iron meteorites. Geochim Cosmochim Acta 12:3043–3061
22. Brearley A, Jones R (1998) Chondrite meteorites. Rev Min 36:83–190
23. Anfilogov V, Bykov V, Osipov A (2005) Silicate melts. Nauka, Moscow (in Russian)
24. Belogub E, Kozlov E, Yu Z et al (1999) Transformation of rocks of spherical stress waves. Uralskiy Mineralogicheskiy Sbornic № 9. Ural Branch of RUS Press. MIASS, pp 206–223 (in Russian)
25. Tomeoka K, Ohnishi I (2011) A hydrated class in the Mokoia CV3 carbonaceous chondrite: evidence for intensive aqueous alteration in the CV parent body. Geochim Cosmochim Acta 75:6064–6079
26. Burkhardt C, Kleine T, Bourdon B et al (2008) Hf-W mineral isochronal for Ca, Al-rich inclusions: Age of the solar system and timing of core formation in planetesimals. Geochim Cosmochim Acta 72:6177–6197
27. Babechuk M, Kamber B, Greig A et al (2010) The behavior of tungsten during mantle melting revised with implications for planetary differentiation time scale. Geochim Cosmochim Acta 74:1448–1470
28. Anfilogov V, Khachay Y (2013) Evolution of core and silicate envelopes at heterogeneous accumulation of the Earth. Lithosphera 4:146–153 (in Russian)
29. Khachay Y, Anfilogov V (2009) Variant of temperature distributions in the Earth on its accumulation. In: The study of the earth as planet by methods of geophysics, geodesy and astronomy. Proceedings 6th Orlov Conference, Kiev, pp 197–202
30. Khachay Y, Anfilogov V (2009) The temperature distribution numerical models for the Earth envelopes at its accumulation stages. In: Geodynamics. Deep structure. The Earth's thermal field. Geophysical fields interpretation. Proceedings 5th Conference on behalf of Bulashevich. Ekaterinburg, pp 520–522 (in Russian)
31. Marakushev A (1991) The early earth's crust on meteorite investigation data. In: Early Crust: The Composition and Age. Nauka, Moscow, pp 27–38 (in Russian)
32. Mason B (1962) Meteorites. Wiley, New York
33. Agee C, Li J, Shannon M et al (1995) Pressure-temperature phase diagram for the Allende meteorite. J Geophys Res 100(B9):17725–17740
34. Tikhonov A, Liubimova E, Vlasov V (1969) About the evolution of fusion zones in the thermal history of the earth. Dokl Acad Sci USSR 188:338–342 (in Russian)
35. Saxena S, Lasor P, Dubrovinsky L (2000) A model of earth's deep interior based on mineralogical data. Mineral Petrol 69(1):1–10
36. Herlund J, Thomas C, Trackley P (2005) A doubling of the post-perovskite phase boundary and structure of the earth's lower mantle. Nature 434:882–886
37. Ozava H, Hirise K, Mitome M et al (2009) Experimental study of reaction between perovskite band molten iron to 146 GPa and implication for chemically distinct buoyant layer at the top of the core. Phys Chem Miner 36:365–363
38. Ringwood A, Hibberson W (1990) The system Fe-FeO revised. Phys Chem Miner 17:313–319

Chapter 6
Formation of the Earth's Proto-Crust

Abstract In this chapter, we analyze the model composition of the Earth's upper mantle, which may have formed as a result of the evolution of the PT conditions obtained. Consideration is given to the variant in which, as the melt layer decreases, it is depleted by the oxide of aluminum and enriched by alumina. With the further cooling of the melt layer, the conditions for the formation of anorthosite in the surface layers occur.

Keywords Heterogeneous melt layer · Fractioning · Pyrolyte · Anorthosite · Greenstone

6.1 The Possible Composition and Structure of the Earth's Upper Mantle

The crystallization of the melt, which occurs at the bottom of the melt layer, proceeds into the conditions of the open system, because new material is constantly entering the melt, after being sedimented on the Earth's surface during the accumulation process. For those conditions the evolution of the silicate envelope formation depends on the velocity of the Earth's growth and on the amount of kinetic energy that is transformed into thermal energy during the process of accumulation. Independently from the accepted model of the Earth's accumulation (homogeneous or heterogenic), the velocity of the Earth's growth up to ~ 60 % of the current radius is constant and then it begins to gradually decrease [1].

When the velocity of the Earth's growth decreases, the thickness of the melt layer also decreases and the phases composition, which is derived from the melt at the bottom of the layer, will change according to the curve of the liquidus (Chap. 5, Figs. 5.2 and 5.3). For peridotite KLB-1 the sequence of phase composition is as follows: 1—liquidus phases: L + MgPv + Mw in the interval of the melt layer thickness of 800–670 km. 2—liquidus phases: L + Gt + Mw in the interval of the melt layer thickness of 670–440 km. We attempt to connect this sequence to the

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radius of the growing Earth. For that we can use the distribution of mantle material density determined by geophysical methods [2].

Above, we assumed that the association of the magnesium-perovskite and magnesium-wüstite MgPv + Mw crystallized on the core–mantle boundary. For the current structure of the Earth, that association is located in the composition of the transition layer "D". At that time, the layer of the melt had a thickness of 800 km and it was located in the radius interval of the growing Earth from 3000 km (the core–mantle boundary) up to 4000 km. The value 4000 km is determined by summing the following values: the core radius (3000 km) + the thickness of the transition layer ($\sim$200 km) + the thickness of the melt ($\sim$800 km). At the depth of 680 km the mantle is composed of the association presenting magnesio-wüstite and garnet. These phases are crystallized at the bottom of the melt, when its thickness decreases to 680 km (Chap. 4, Fig. 4.5).

That thickness of the layer has to exist when the Earth's radius reaches 5700 km. Although the presented calculations are approximate, they allow us to connect the change of the mineral associations by the crystallization in the melt layer with the possible mineral composition and physical characteristics of the Earth's mantle. Thus, it is necessary to take into account that the change of the mineral associations with the decrease of the melt layer thickness occurs gradually.

The melt crystallization in the interval of the growing Earth's dimensions from 3000 to 5570 km occurred only in the bottom of the layer in the open system conditions. At that time the composition of the meteorite matter gradually changed from H-chondrite to LL-chondrite. Accordingly, the composition of the model matter also changes from peridotite KLB-1 to Allende meteorite (Chap. 5, Table 5.1). For these conditions at the bottom of the mantle, the crystalline phases had sedimented, and are located on the liquidus curve of KLB-1. It can be expected that the liquidus KLB-1 phases change to the phases that are allocated on the liquidus curve of the meteorite Allende, when the thickness of the melt layer becomes less than 500 km. That transition decisively defines the peculiarities of the crystalline differentiation process of the melt into the melt layer in the final stage of the Earth's accumulation, when its radius achieves its current value, and the crystallization of the melt will proceed in the closed system.

When the melt layer thickness reaches the value of 420 km, the garnet, which extracts the alumina Al_2O_3 from the melt, will vanish from the liquidus phases and the alumina Al_2O_3 will itself accumulate in the melt. Olivine crystallization on the liquidus curve [3, 4] will be accompanied by depletion of the melt MgO and enrichment of the melt FeO. This creates favourable conditions for the next plagioclase crystals flotation and their accumulation in the upper part of the melt layer. According to data of [5], An_{89} is floated by the content of 14.1–14.5 mass% Al_2O_3 and 11–12 mass% FeO of the iron oxides sum.

According to the suggested model [6], at least on the Earth's surface will sediment matter that has a composition near to the carbonaceous CI chondrites, which contain the lowest temperature products of the proto-planetary matter condensation (Table 6.1) [7].

Table 6.1 Compositions of carbonaceous chondrites [7]

Component	Migei	Staroe boriskino	Grosnaya
SiO_2	27.8	26.7	33.8
TiO_2	0.05	0.15	–
Al_2O_3	2.15	3.11	3.45
FeO	19.1	29.8	28.8
MgO	19.5	19.4	23.6
CaO	1.66	2.2	3.20
Na_2O	0.63	1.07	0.06
K_2O	0.05	0.20	0.30
H_2O+	12.9	8.72	–
H_2O-	–	2.96	–
H	–	–	0.17
FeS	10.0	1.02	5.37

From this material, the initial solid Earth's envelope formed. Its thickness can be estimated from the following considerations. The mass of water in the Earth's hydrosphere constitutes 0.024 % of the Earth's mass. If we assume that the average content of H_2O in the carbonaceous chondrites is 7 % and the whole of the water initially belongs to the composition of the outer envelope, its thickness will be 20 km. It is possible that the carbonaceous chondrite material sometimes sediments simultaneously with the material of ordinary chondrites. In that case, the thickness of the upper envelope, which contains water, can increase to 40–50 km. The time of the sedimentation on the Earth's surface of the carbonaceous chondrites material can be considered as being near to the end of the accumulation process. We can consider it to be equal to the age of carbonaceous chondrites, which consist of the lowest temperature condensates. According to data of [8], this is about 60 million years. Significant results, which are evidence of the possibility of the sedimentation of carbonaceous chondrites during the finishing stage of the Earth's accumulation, as described in Chap. 4, are presented in [9]. The possible structure of the Earth after the sedimentation of the carbonaceous chondrite material on its surface is shown in Fig. 6.1.

During the cooling of the melt layer, which was covered by the material of the carbonaceous chondrites, crystallization will occur not only at the bottom of the layer, but also at contact with the covering solid envelope. Plagioclase begins to crystallize in this contact region, when the temperature decreases to 1200–1250 °C. These crystals will float and on the contact of the melt layer with the solid envelope, a magmatic "mesh" which consists of plagioclase crystals and basaltic melt will form. This magmatic "mesh" will be ejected as diapirs through the covering solid layer of the carbonaceous chondrites to the surface, forming on the Earth's surface large clusters of anorthosite, similar to the anorthosite of which the ancient Moon crust consists. The mechanism of the Moon's anorthosite formation is considered in detail by Warren in [10, 11]. As the base, as well as in our variant, is the assumption that, in the final stage of the Moon's formation there existed on its subsurface area a

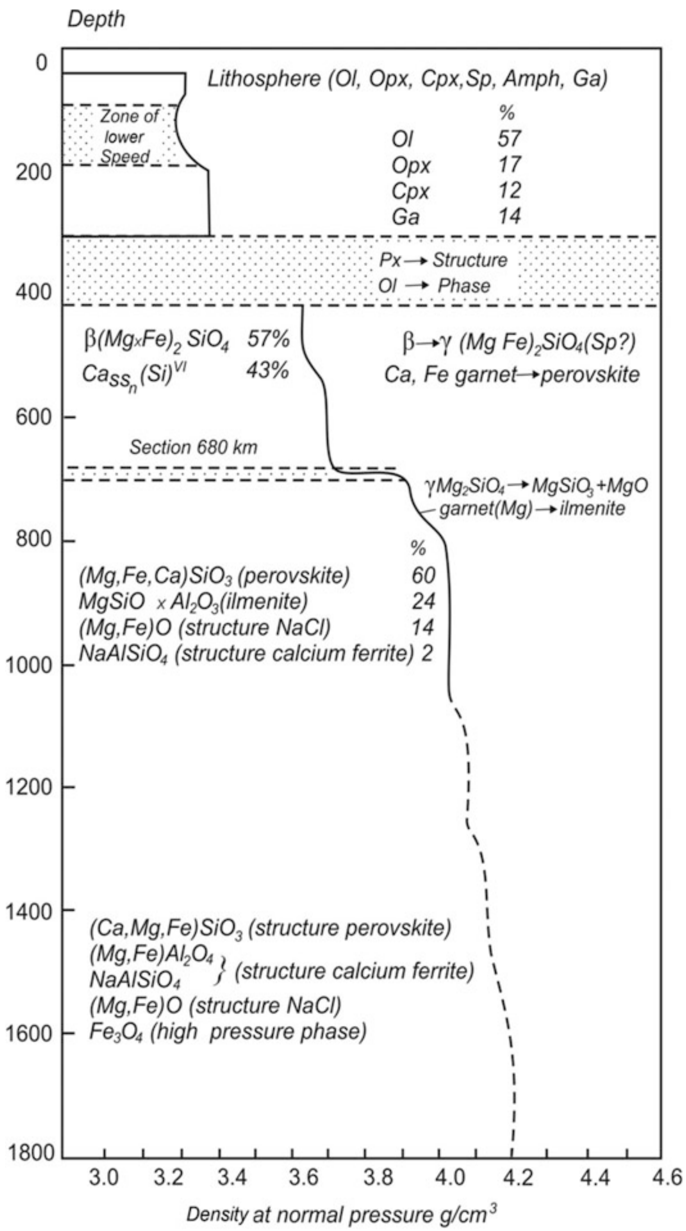

Fig. 6.1 The mineral consist and density of the Earth's mantle [2]

"magma ocean" with a thickness up to 400 km. The possibility of plagioclase flotation in the melt of the main composition was experimentally demonstrated by [5]. They determined that the real density of plagioclase is somewhat lower than the theoretical one, and the lifting force that occurs in the melt of the main composition is respectively higher. From the melt of the outer solid Earth's layer, which consists of the material of carbonaceous chondrites, the main mass of water and carbon dioxide seceded, then, as a result, the density increased and the blocks containing the dehydrated material of the carbonaceous chondrites could sink into the melt layer as a whole down to its bottom. By means of that interaction with the melt and the upper hybrid, the heterogeneous mantle layer was formed. The composition of this layer may be close to Ringwood's pyrolyte [2]. As the result of that process, the mantle's heterogeneity was fixed to a depth of 300–350 km in the transversal wave vertical sections for different crust types [12].

The peculiarities described here of the Earth's upper mantle formation at the finishing stage of the active stage of planetary accumulation allow us to analyse in a new way the existing data on the formation of the proto-crust.

6.2 Condition of Archaean Proto-Crust Formation

One of the main peculiarities of the Precambrian Earth's crust, which has rarely attracted the attention of researchers, is the large volume in its composition of sialid rocks, which are present as quartzite, grey gneisses of diorite composition and garnets. The thickness of the sialid crust reaches 40 km [12]. This peculiarity may be explained by assuming that, due to water allocation from the initial mantle, at that stage a special matter differentiation mechanism acted, which led to the formation of large volumes of the garnet melt. Really, by means of water from the mantle material, small volumes of the melt of diorite or garnet composition can be melted, but these volumes are insufficient to provide the sialid composition of the Precambrian lithosphere. We suggest that the necessary material for the formation of the Archaean "granitic" lithosphere was produced by specific matter conditions that occurred at the final stage of the Earth's accumulation.

As we noted above, by means of the anorthosite mesh entering the outer solid Earth's envelope, which consisted of the carbonaceous chondrite material, it was heated. Due to the fact that the main mass of water was contained in that material, carbon dioxide and other gases evolved from it, and there occurred the instantaneous—on the geological time scale—formation of the ocean and atmosphere. Initially, these components, which formed a dense atmosphere around the Earth, similarly formed the atmosphere of Venus. With the increasing of the atmosphere's thickness and its cooling, the gases condensed and formed a hydrosphere, as a shallow ocean.

Due to the high temperature and subsequently the small viscosity of the subsurface areas, the Earth's surface at that stage may have been near to equipotential and would not have had any significant relief anomalies. Therefore the depths of the initial ocean could not be great and the ocean covered the whole Earth's surface. That created the conditions of a "greenhouse" effect that heated the Earth and the ocean surface. The same conclusion concerning the time and the ocean formation mechanism was reached by Salop and many other researchers [13, 14]. Due to the high temperature of the Earth's surface, the temperature of the initial ocean may have exceeded 150 °C and the pressure, in equilibrium with the atmosphere, may have exceeded 10^6 PA [13]. It is assumed that the water of the initial ocean could contain up to 5000 cm^3 of CO_2 per liter. Moreover, it must have contained significantly many strong acids, notably HCl and H_2SO_4 [13]. Investigations of the inclusions in the quartz of ancient Aldanian quartzite showed that they contain up to 60 % of CO_2, about 35 % of H_2S, SO_2, NH3, HCl, HF, and 1–8 % N_2 and rare gases [15]. Therefore the ancient lithosphere was aggressive and able to extract from the rocks strong bases. Salop considered that the ancient quartzite, which had been located at the bottom of the Katarchaean section, had been formed as a result of erosion and chemical weathering of granitites. The chemogenic nature of the quartzite is testified by the absence in it of relicts of fragmentary structure and also by interstratifications of quartzite with the high alumina rocks, sillimanite and corundum, containing gneisses. However, to explain the existence of the large volumes of magmatic acidic rocks with ages of more than 4400 million years is hard. The problem can be solved if we assume that it was not granitites, but the initial anorthosite crust that underwent chemical weathering.

Due to the high acidity of the initial ocean, calcium extracted from the decay of the anorthosite was retained in the ocean water in a dissolved state and, with decreasing of the water temperature, entered into the composition of carbonates located above the Fedorovsky Katarchaean formation (Fig. 6.2), while alumina, stable in the acidic medium, entered the composition of quartzite as an interlayer of high alumina rocks after the decomposition of anorthosite, owing to the high concentration of powerful acids.

The Al_2O_3, which is stable in acid medium, entered the quartzite composition as a high alumina interlayer. In recent years the idea of the early ocean's formation has received serious verification due to the investigation of oxygen isotope composition in zircon from West Australian quartzite, and the identification of conglomerates, aged up to 4400 million years, defined by the U-Pb method [16, 17].

Interlayers of aluminum-silicate rocks in these quartzites are similar to Aldan interlayers [17]. Isotope composition investigations in zircon allow us to conclude that the ocean and land fragments had been formed earlier than 4400 million years ago [18, 19].

Intrusions of anorthosite and the formation of ocean and ancient quartzite were accompanied by submarine basalt eruptions, arising from the melt layer.

Basalts are fixed in the section of the Iengersky stratigraphic complex, by the pyroxene and amphibole crystalline orthoshales [13].

The global character of the formation process of the foundation of the Katarchaean proto-crust is testified by the section correlation of the Katarchaean complexes presented in [13].

The analysis outlined above allows us to propose the following scheme of the geological evolution of the Katarchaean Earth's crust in the period from 4.4 to 3.5 billion years.

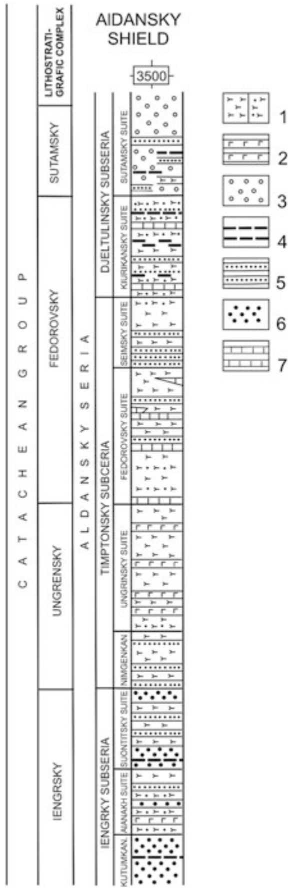

Fig. 6.2 Lithostratigraphic Katarchaean complexes of Aldansky shield. *1*—Basic crystalline ortho schist, gneisses, and other metabasites; *2*—ultra basic crystalline schist and meta pyroxenite; *3*—granitized gneisses; *4*—leucocratic granulites; *5*—high-aluminum gneisses; *6*—garnetiferous biotite gneisses; *7*—quartzites [13]

- 4400 million years—formation of the initial anorthosite crust.
- less than 4400 million years—formation of the initial hot ocean, with a significant chloride content; deep chemical decomposition of the initial crust matter; sedimentation of the chemogenic siliceous and high alumina rocks; eruption of lava of the main content.

4000–3750 million years—denudation of ancient rocks, local sedimentations of carbonaceous and iron-siliceous rocks, eruption of basaltic and acid lavas, occurrence of fragmental sediments.

3750–3500 million years—Saamian diastrophism, intense folding, early tectonic intrusions of the base and ultra base content, big masses of granitite composition for syntectonic intrusions (granodiorite, quartz diorite), metamorphism of granulites and amphibolites associations.

Beginning from 3700 million years, the volume of the granite material in the Katarchaean lithosphere sharply increases. The unique source from which the acidic melts in the Precambrian crust could be formed is the Katarchaean proto-crust, consisting of quartzite enriched by alumina. The acidic melts could be produced from this material as the result of interaction with the deepest melts of the basic content. Therefore, as a rule, they have a hybrid nature [20]. The process of the hybrid acidic melt formation could be realized only by conditions of high temperature and pressure achieved at a depth of not less than 15 km. Therefore, its development was preceded by the accumulation of the sedimentary material and formation of the sedimentary cover with a thickness of more than 15 km. The formation of this thick cover took about 900 million years and large volumes of the hybrid rocks, which were present as the so-called grey gneisses of the tonality content, which occurred only during the time interval of 3000–3700 million years [20]. In many cases, the grey gneisses are the foundation on which the greenstone belts were formed, in which the granites and acidic vulcanites play the dominant role [20].

References

1. Vitjazev A, Pechernikova G, Safronov V (1990) The Earth's group planets. Nauka, Moscow (in Russian)
2. Ringwood A (1982) Origin of the Earth and Moon. Nedra, Moscow (in Russian)
3. Herzberg C, Zang J (1996) Melting experiments on anhydrous peridotite KLB-1: compositions of magmas in the upper mantle and transition zone. J Geophys Res 101(B4):8271–8295
4. Agee C, Li J, Shannon M et al (1995) Pressure-temperature phase diagram for the Allende meteorite. J Geophys Res 100(B9):17725–17740
5. Campbell I, Roeder P, Dixon J (1978) Plagioclase buoyancy in basaltic liquids, as determined with centrifuge furnace. Contrib Miner Petrol 67(4):369–378
6. Anfilogov V, Khachay Y (2012) Differentiation of mantle material at the Earth's accumulation and the initial crust formation. Lithosphere 6:3–13 (in Russian)
7. Vdovikin G (1974) Meteorites. Nauka, Moscow (in Russian)
8. Chen J, Tilton G (1976) Isotopic lead investigations of the Allende carbonaceous chondrite. Geochim Cosmochim Acta 40:635–643
9. Tomeoka K, Ohnishi I (2011) A hydrated class in the Mokoia CV3 carbonaceous chondrite: evidence for intensive aqueous alteration in the CV parent body. Geochim Cosmochim Acta 75:6064–6079
10. Warren P (1990) Lunar anorthosites and the magma-ocean plagioclase-flotation hypothesis: importance of FeO enrichment in parent magma. Amer Miner 75:46–58
11. Warren P (1993) A concise compilation of petrologic information on possible pristine nonmare Moon rocks. Amer Miner 78:360–376
12. Dorman J, Ewing M, Oliver J (1960) Study of shear—velocity distribution in the upper mantle by mantle Rayleigh waves. Bull Seismol Soc Am 50:87–115
13. Salop L (1982) Geological evolution of the Earth in Precambrian. Nedra, Leningrad (in Russian)
14. Shuldiner V (1991) The initial crust and its evolution. In: Early crust: the composition and age. Nauka, Moscow, pp 87–94 (in Russian)
15. Kasanski Y, Kataeva V, Shugurova N (1973) About the composition of ancient gas inclusions of quartz rocks. In: Geochemistry of precambrian and paleozoic siberian sediments. Novosibirsk, pp 5–12 (in Russian)

16. Cavoise A, Valley J, Wilde S (2005) Magmatic δ ^{18}O in 4400–3900 Ma. Detrtial zircons: A record of the alteration and recycling of crust in early Archaean. Earth Planet Sci Lett 235:663–681
17. Spaggiaria C, Pidgeon R, Wilde S (2007) The Jack hills greenstone belt. Western Australia: Part 2: Lithological relationships and implications for the deposition of ≥4.0 Ga detrital zircons. Precambr Res 155(3–4):261–296
18. Valley J, Peck W, King E et al (2002) A cool early earth geology 30:351–354
19. Walter M, Tronnes R (2004) Early Earth differentiation. Earth Planet Sci Lett 225:253–269
20. Chadwick B (1980) New data on Archaean geology of South-West Greenland. In: Windly B (ed) The early history of the Earth. Mir, Moscow, 212–223 (in Russian)

Chapter 7
About the Origin of the Moon Matter

Abstract This chapter presents a brief overview of the research state of the Moon at the present time. It is shown that the existence of the energy that follows the decay of short-living isotopes stipulates a principally new mechanism of matter differentiation into the proto-planetary cloud and allows the construction of a dynamic model of the Moon's formation as a development of the results of Schmidt, Safronov, Maeva, Ruskol, and Kaula without the model of mega impact. The results of numerical modelling of the temperature distribution dynamics in the inner areas of the Moon during the stage of accumulation in the frame of a spherical symmetrical model are presented.

Keywords Short-living radioactive elements · Matter differentiation of the proto-planetary cloud · Earth-Moon system · Numerical modelling · Initial moon's temperature formation

7.1 The Hypothesis of the Moon's Formation

The problem of the origin and composition of the Moon's material is of fundamental importance to understanding the mechanism of the planetary Solar system's formation. As far as the hypothesis of the mega impact is concerned, in the literature there is discussion of three main mechanisms of the Moon's formation: 1—the hypothesis of the Moon's separation from the Earth; 2—the hypothesis of capture; 3—the hypothesis of joint formation or co-accretion of the Earth and Moon. The defects of these hypotheses have been considered by [1–3].

The idea of the Moon's separation from the Earth was suggested in 1880 by Darwin [4]. Its discrepancy with the laws of celestial mechanics was considered in [1]. According to these authors, in the case of rotational instability occurring, which is the reason for part of the material separating from the rotating body, a soft separation of the satellite from the basic body was not possible. The material that was thrown out as a result of the rotational instability either flew away or fell back.

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Later Ringwood attempted to modify Darwin's hypothesis. He supposed that the material from the Earth's mantle was thrown out into the lunar orbit as the result of impacts of large meteorites [5].

The hypothesis of the mutual formation of the Earth and Moon was proposed by Schmidt and Ruskol [6–8]. Schmidt assumed that the Moon had accumulated in the growing Earth's vicinity from the pre-Earth's cluster of bodies, continuously replenished from the proto-planetary cloud. According to [1], "the hypothesis of O.Y. Schmidt is based on the processes that necessarily had to proceed during the Earth's accumulation and from the mechanical point of view it could be more perspective". However, in the frame of that hypothesis, it would be expected that the Moon and the Earth have the same composition and the attempt to prove that this process can lead to the different composition of these bodies, according to the authors of [1], is dubious.

7.2 Current Data on Composition and Inner Structure of the Moon

Significant progress in the study of the Moon has been achieved from research on the material delivered by the automatic stations Luna-16, Luna-20 and Apollo expeditions from the Moon to the Earth. The data obtained from the study of these materials allowed us to determine some rigid limitations on the possible material composition of the Moon and the research on its gravitational field. The first seismic experiments yielded information about the density distribution and structure of the inner areas of the Moon [5, 9, 10]. Detailed analysis of these results and the problems that occurred, in relation to obtaining new data, is provided in the work of Kaula and Levin and Maeva [1, 11]. Let us note the most important of their results. 1—It was determined that the Moon had undergone magmatic differentiation, which occurred simultaneously with its growth, and that is why currently its interior is in a partially melted state. As a result of this discovery, it was necessary to have a high initial temperature for the Moon. Nobody could suggest the energy source that had been necessary for the realization of that state. Kaula assumed that the asthenosphere of the Moon, located at a depth of 700 km, is evidence that the area located above the asthenosphere was too cool for early differentiation arrangement [11]. Attempts to explain the high temperature in the lunar interior by a higher concentration of radio-active elements are not quite decisive and create a new problem: it then becomes necessary to find the source of extended concentrations of these elements. 2—A marked lack of iron is determined, compared with the Earth. According to [1], if we take into account the non-dimensional inertial moment of the Moon equal to: 0.005 C/MR* = 0.395, the iron core mass, if it exists in general, cannot exceed 1–2 % of the Moon's mass [1], whereas the Earth's iron core mass amounts to about 30 mass%. The Moon's third peculiarity is the enrichment by minerals with high concentrations of Ca, Al and apparently Ti.

The presence on the Moon of a thick anorthosite crust and partially melted central area suggests that the whole of its volume passed through the melted state and in the final stage of its formation there was a layer of melt with a thickness of 400 km near to its surface [12]. Kaula assumed that the only suitable model of the Moon's formation has to explain the three main differences in its chemical composition from that of chondrite meteorites: 1—the loss of volatile components; 2—the loss of iron; 3—enrichment by plagioclase [11].

The inadequacy of the three main hypotheses for the Moon's formation and the above listed problems, the main one of which is the heating of lunar material up to the melting temperature, have led to the emergence of the mega impact hypothesis [13–17]. Tailor considered that the high content of infusion elements in the rocks of the lunar surface and the sharp re-enrichment of volatile components are so large that it is difficult to consider the Moon and the Earth as objects that were formed from a unified physicochemical system [17].

There was the opinion that if the Earth-Moon system is unique, then it is possible that its genesis is also unusual, and one such unusual variant, which is now widely discussed, is the hypothesis of the Moon's origin as a result of Earth's collision with a cosmic body of planetary size (with the mass of Mars or more). The mega impact is considered also as a source of the energy needed for initiation of the separation process of initial homogeneous Earth into an iron core and silicate mantle. Note that Mercury, Venus and Mars also have a metallic core, but do not have satellites and, moreover, it is hard to believe that a unique method of core formation resulting from collision with other cosmic bodies existed for all the inner planets of the Solar system. The hypothesis of the mega impact does not solve the problem, and merely replaces the solution with a speculative scheme. Arbitrarily varying in the composition and dimensions of the impact body, this hypothesis allows us to explain any given differences in the composition of the Earth and the Moon, as well as the Moon's deviations from the ecliptic plane and the specific angular moment of the momentum for the Earth + Moon pair, but there are no algorithms to verify this. A set of geochemical contradictions, which are incompatible with the hypothesis of mega impact, is considered by Galimov [2, 3]. Many papers are devoted to the hypothesis of the mega impact, but because of their speculative nature it is not necessary to discuss them further.

7.3 The Possible Source of Material in Its Early Stage of Growth

Before explaining the differences in the nature of the chemical composition of the Earth and Moon and the mechanism of the Moon's heating at the initial stage of its formation, we have to choose a definite dynamic model of the Moon's formation. We believe that the model suggested by Schmidt and further developed by Ruskol and Kaula is the most correct [8, 11, 18, 19]. Ruskol and Kaula proposed that the dynamically possible method is one where the Moon is a (subtended) by-product of the Earth's formation, but this thesis could have a double interpretation. In the variant accepted by Ruskol and Kaula, the Moon was formed from the cluster of satellites, which were particles before becoming satellites of the Earth, rotating on heliocentric orbits. This variant assumes that the Earth and the Moon were formed from the same material. This then creates insuperable difficulties in explaining the differences in the chemical composition of the Earth and the Moon.

We think that in the initial stage the Moon was formed from fragments that were formed at the destruction of initial Earth's pre-planetary bodies. This variant is a logical consequence of our suggested model of heterogeneous accumulation [20–22], for which the foundation is the two-stage mechanism of the formation of the Earth's pre-planetary bodies.

7.4 The Composition of the Moon's Matter

Our suggested two-stage mechanism of the Earth's formation allows us to propose, coordinated with this, the variant of the Moon's formation, which explains the existing lack of iron and the high initial temperature in its interior. The fundamental difference between our model and the Schmidt-Ruskol model consists in the fact that the material from which the Moon was formed in the initial stage is represented not by particles located on the heliocentric orbits, but mainly by initial fragments of the Earth's pre-planetary bodies, which were thrown out by their destruction on to the satellite orbits. That material consists of melted material with composition near to that of CAI and partially melted iron and fragments of the outer solid envelopes.

The destruction of the initial Earth's pre-planetary bodies occurs at the stage when their masses are small and the velocities necessary for the ejection of fragments from the supply zone of the growing Earth, caused by the collisions and destruction of relatively small bodies, would be quite accessible. Therefore, the amount of fragments of the initial pre-planetary bodies, which after their collisions passed into the Earth's satellite orbits, was sufficient for the formation of the Moon's central part. After this, the Moon's growth was provided by material that was not on the heliocentric orbits, together with fragments that had been formed from the collisions of planetesimals with the growing Earth. Therefore, from the beginning of a certain moment, the material composition of which the Earth and its satellite were formed is practically identical.

Just in the initial stage, the heterogeneous iron distribution between the Earth and the Moon arises. In the secondary Earth's pre-planetary bodies, the main mass of iron passes into their central parts, from which the Earth's core is then formed. On the Moon, the fragments of the initial pre-planetary bodies, in which the main part consists of the material of CAI enriched by Al_2O_3, fall. On the surface of the main part of the Earth's core and on the surface of the growing planet, silicate chondrite material with relatively low iron concentration sediments after the formation of the main part of the Earth's core is finished.

The second problem that must be solved by any adequate model of the Moon's formation is the problem of the lack of Na, K, Cs, Rb, Mn and practically the total absence of H_2O and carbon dioxide in the composition of the lunar rocks. According to the suggested model of the heterogeneous Earth's accumulation [21], during the last stage, material with a composition close to carbonaceous CI chondrites must be sedimented, which consists of the lowest temperature products of the condensation of proto-planetary matter, enriched by volatile components. There may be two reasons for the absence of traces of these elements on the Moon. As the radius of the Earth's gravitational field increases, the flow of fragments being ejected from its supply zone will decrease. Therefore, the later low temperature condensates could not pass beyond the limits of the Earth's gravity area and the Moon became depleted of these components. More likely is another variant, which supposes that the material of carbonaceous chondrites sedimented not only on the Earth but also on the surface of the Moon, and the part that sedimented on the Moon was proportional to the ratio of the masses of the Moon and the Earth, that is, 0.012. During the Moon's magmatic activity and the bombing of its surface by meteorites, that material may have been completely transformed, with the volatile components dissipated into space. The possibility of the realization of this second variant is borne out by the existence on the Moon of rocks of the alkaline group, including granites [23], which could have formed only by water existing in the initial Moon's matter.

7.5 Thermal State and Evolution of the Moon's Material

One of the fundamental achievements of the Apollo expeditions was to prove the high density of the heat flow on the Moon's surface. Accordingly, it was concluded that the Moon's interior is in a partly melted state and that the Moon passed through a stage of intense magmatic differentiation [1]. The very decisive proof is the existence of the Moon's ancient anorthosite crust, which could have formed only as a result of crystalline differentiation of the melt layer with a thickness of about 400 km, which must have existed near the Moon's surface during the finishing stage of its formation [24–27]. This depth of the magmatic ocean near the Moon's surface corresponds with experimental data about the crystallization of lunar basalts and picrite glasses and with the location of the bottom boundary of the middle mantle, which is at a depth of 500 km.

All these results demand special features for the thermal Moon models; the main one is the existence of the thermal energy for its early heating. The available results showed [1] that for such a short time interval the known internal sources of energy would not have been able to heat the Moon up to the temperature necessary for its melting and for magmatic differentiation.

Our model proposes that the Moon began to form from the fragments of the initial planetesimals, the average temperature of which was higher than 2000 K [22]. The cooling of these fragments, with sizes of several tens of kilometers, which could lead, as pre-planetary bodies, to the Moon's formation, lasted for n × 10^4 years. This is quite enough for the formation of a new body, with a size of a few hundred km, in which the heat could be retained for n × 10^6 years.

The decay of ^{26}Al is an additional heat source that could maintain the high temperature in the central part of the growing Moon. As we noted above, the destruction of the initial Earth's pre-planetary bodies began 1 million years after CAI formation. That time interval is determined by the relative age of iron meteorites, which is 1–3 million years [28]. The time during which the bodies with a radius up to 400 km were heated by the energy of ^{26}Al decay depends on the period of the half-decay, which is equal to 7.38 × 10^5 years and on the concentration of Al_2O_3 (Chap. 3, Table 3.2) [20]. In the material, which is near to the CAI composition, from which the Moon's central part had to be formed, the concentration of Al_2O_3 is equal to about 30 mass% [30]. For these concentrations, that source of energy will last for at least 1 million years after the material of CAI enters into the composition of the Moon, and during that time the main part of the growing Moon will be in the melted state. Simultaneously, during the process of the Moon's growth another source of energy will be active. The collisions of planetesimals with the melted Moon will be inelastic and a significant part of the energy precipitated by the collisions will be transformed into the thermal energy of the Moon. The thermal model of the Moon and the temperature distribution during the process of its formation will be discussed in the Appendix.

As significant parameters, which define the thermal and chemical evolution of the Moon, we can consider the possible composition and dimensions of its core. The numerical modelling of these parameters done by Kuskov and Konrad [26] led the authors to two alternative variants: to an iron core with a radius of 330–390 km and density of 8.0 g/cm^3, and to a pyrrothine core with a radius of 490–590 km and density of 4.7 g/cm^3. Note that the main parameter that defines the core's dimensions in this model is not the composition, but the core's assumed density. It is understood that the density of 8.0 g/cm^3 is only possible with an iron core, whereas for the density of 4.7 g/cm^3 there are possible alternative variants. One such variant is where the Moon's core consists of iron and minerals that enter into the composition of CAI, and from which, according to our model, the cores of the initial Earth's pre-planetary bodies were constructed. Accepting the value of the core density calculated in [26] as 4.7 g/cm^3, the average density of CAI as 3.0 g/sm^3 and the iron density of 8.0 g/cm^3, we can calculate the parts of these components in the composition of the lunar core for the radius of 490–590 km. These parts are equal to 0.7 for iron and 0.3 for aluminum silicate composition.

The gravitational acceleration into bodies of sizes up to 500 km is small and therefore in the formation of the central part of the Moon there was no density separation of iron and silicate phases. It can be assumed that the structure of the lunar core at the moment of its formation was similar to the structure of pallasites. The material of CAI uniformly distributed in the volume of the lunar core played the role of olivine in pallasites. Later the silicate material and iron were able to divide, resulting in an iron core being formed with a radius of 330–390 km.

The problem of the existence of excessive angular momentum in the Earth-Moon system will be discussed as a conclusion. Supporters of the mega impact model consider that fact as one of the main evidence of the Moon's formation as a result of the Earth's impact with a cosmic body [17]. Mercury and Venus have cores like that of the Earth, which probably consist of iron–nickel material. This gives the basis to suggest that their formation, like that of the Earth, occurred in two stages. The differences in the formation processes of Mercury and Venus on the one hand and the Earth on the other arise because the Earth, located significantly far from the Sun, was able to keep the fragments of the silicate envelopes in the area of its gravitation and the Moon was then formed from these fragments. However, in the case of Mercury and Venus this material occurred in the gravitational field of the Sun and was absorbed by the Sun. The same situation took place during the formation of Mars. Part of the fragmentary material was absorbed by the growing Jupiter and then it passed into the asteroid belt.

References

1. Levin B, Maeva C (1975) The problems of the Moon's origin and thermal history. In: Cosmochemistry of the Moon and planets. Nauka, Moscow, pp 238–298 (in Russian)
2. Galimov E (2004) About the Moon's material origin. Geokhimia 7:691–706 (in Russian)
3. Galimov E (2005) Dynamic model of the Moon's origin. Geokhimia 11:1139–1150 (in Russian)
4. Darwin G (1880) On the secular changes in the orbit of a satellite revolving around a tidally distributed planet. Phil Trans Roy Soc London 171:713–891
5. Ringwood A (1986) Composition and origin of the moon. In: Hartman W, Phillips R, Taylor G (eds) Origin of the Moon. Lunar and Planetary Institute, Houston, pp 673–698
6. Schmidt O (1950) The origin of the planets and their satellites. Izvestiya AN USSR Phys Series 14(1):29–45 (in Russian)
7. Schmidt O (1957) Four lectures on the theory of the Earth's origin. Academy of Sciences of USSR Press, Moscow (in Russian)
8. Ruskol E (1972) About the Moon's formation from the near-Earth swarm. Izvestiya AN USSR Earth Phys 7:99–108 (in Russian)
9. Nakamura Y (1983) Seismic velocity structure of the lunar mantle. J Geophys Res 88:677–686
10. Hays J, Walker J (1975) Igneous lunar rocks and the nature of the Moon's interior nature. Cosmochemistry of the Moon and planets. Nauka, Moscow, pp 275–282 (in Russian)
11. Kaula W (1971) Dynamical aspects of lunar origin. Geophys Rev Sp Phys 9:217–238
12. Warren P (1985) The magma ocean concept and lunar evolution. Ann Rev Earth Planet Sci 13:201–240
13. Cameron A, Ward W (1976) The origin of the Moon. In: Proceedings 7th Lunar Science Conference, Houston, pp 120–122
14. Cameron A, Benz W (1991) The origin of the Moon and single-impact hypothesis IV. Icarus 92:204–216
15. Benz V, Slattery W, Cameron A (1986) The origin of the Moon and single-impact hypothesis I. Icarus 66:51–535
16. Benz V, Slattery W, Cameron A (1987) The origin of the Moon and single-impact hypothesis II. Icarus 71:30–45
17. Taylor S (1987) The unique lunar composition and its bearing on the origin of the Moon. Geochim Cosmochim Acta 51(5):1297–1310

18. Ruskol E (1960) The origin of the Moon. 1. Formation of a swarm of bodies around the Earth. Sov Astron J 4:657–688 (in Russian)
19. Ruskol E (1963) On the origin of the Moon. 2. The growth of the Moon from the protoplanetary cloud. Sov Astron J 7:221–227 (in Russian)
20. Anfilogov V, Khachay Y (2005) A possible scenario of material differentiation at initial stage of the Earth's formation. Dokl Earth Sci 403A:954–947 (in Russian)
21. Anfilogov V, Khachay Y (2012) Differentiation of mantle material at the Earth's accumulation and the initial crust formation. Lithosphere 6:3–13 (in Russian)
22. Anfilogov V, Khachay Y (2013) Origin of the Kimberlite diamond-bearing lithosphere of cratons. Dokl Akad Nauk SSSR 451(5):537–540
23. Sharkov E, Bogatikov O (2010) The evolution of the tectonic-magmatic processes in the Earth and Moon history. Geotectonics 2:3–32 (in Russian)
24. Binder A, Manfred A (1980) The thermal history, thermal state and related tectonics of a Moon of fission origin. J Geophys Res 85(B6):3194–3208
25. Binder A (1986) The initial thermal state of the Moon. In: Hartman W, Phillips R, Tailor G (eds) Origin of the Moon. Lunar and Planetary Institute, Houston: 425–433 (1986)
26. Kuskov O, Konrod V (1998) The model of the Moon's chemical evolution. Petrologia 6:615–633 (in Russian)
27. Solomon S (1986) On the early thermal state of the Moon. In: Hartman W, Phillips R, Tailor G (eds) Lunar and Planetary Institute, Houston, pp 435–452
28. Kleine N, Mezger K, Palme H et al (2005) Early core formation and late accretion of chondrite parent bodies: evidence from ^{182}Hf-^{182}W in CAIs, metal rich chondrites and iron meteorites. Geochim Cosmochim Acta 69:5805–5818
29. Nichols R (2000) Short lived radionuclides in meteorites: constraints on nebular time scales to the production of solids. Space Sci Rev 1–2:113–122
30. Brearley A, Jones R (1998) Chondrite meteorites Rev Min 36:83–190

Chapter 8
Conclusions

Abstract The content of short-living radioactive elements in the silicate part of the proto-planetary cloud matter is sufficient, in the inner part of even small proto-planetary bodies with a radius of about 50 km, to provide a temperature higher than the melting temperature of iron.

Keyword Dynamics · Bodies collision · Proto-planetary body · Differentiation · 3-D model · P-T distribution · Accumulation process

The results presented in this book showed the following:

The fact of short-living radioactive elements gives rise to a principally new dynamics of the collision of bodies. During these collisions melted bodies with a thin upper envelope were involved, rather than solid bodies. The masses of small bodies provided not very high relative collision velocities. Thus the collision of the melted inner parts was realized as an inelastic impact and these parts were then combined. The fragments of the solid upper envelope again entered the proto-planetary cloud. But the melted part had mainly iron content, while the solid fragments of the envelope contained silicate matter. So a new mechanism of matter differentiation is realized in the proto-planetary cloud.

Using the mechanism of differentiation and the available data about the matter content of the proto-planetary cloud, it was possible to obtain the variants of the hydrostatic pressure and the temperature in the inner parts of the Earth and the Moon at the stage of their heterogenic accumulation. Variants were obtained of a solution that describes the existence of the Earth's core, the solid state of the inner core and the melted state of the outer core towards the end of the planetary accumulation. In this was considered the matter differentiation of the initial mantle and of the Earth's core.

In the Appendix the numerical results of the problem solution in the frame of a 3-D model are presented. For the first time, the evolution is traced of the local thermal heterogeneities during the planetary accumulation process, which were the result of the random distribution by masses and velocities of the accumulated bodies and particles.

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All these results were obtained, assuming the suggested mechanism of the heterogeneous accumulation of the planets of the Earth's group and they are supported by the available geochemical and geophysical experimental data. The further development of our understanding of the processes of the Solar system's evolution will be determined as well by the detailization of the experimental data and by opportunities presented by numerical experiments.

Appendix

Results of Numerical 3D Simulation of the Thermal Evolution of the Earth and the Moon at the Process of Their Accumulation

To analyze the temperature distribution in the growing proto-planet we used Safronov's [1] model, the most valid model available today, of the accumulation of the planets of the Earth's group and his equation (Chap. 4, (4.1)), which describes the changing of the proto-planet's mass with the growing velocity. The process of body collision depends on their mass, composition and state and research of that process is a great problem (see, for instance, [2, 3]). In that statement the quantitative description of planetary accumulation from the proto-planet cloud is beyond the realm of current computer engineering. In the Safronov equation the whole of the statistics of the collision process is approximated by a parameter θ—a statistical parameter that takes into account the distribution of the particles by masses and velocities into the planets' "power" zone. Beginning from the papers [4, 5], we take into account the 3-D distribution of falling bodies by defining boundary conditions as modified boundary conditions (4.3) (Chap. 4).

On the outer surface of the spherical layer with a thickness Δr during time dt the whole increase of the energy in it is described by a term in the left part of equation (Chap. 4, (4.3)), which depends on the coordinates of the surface points. The solution is discovered in the 1/8 part of the spherical body. Since for further solution of the problem we use the finite-difference method, for each cell of the surface we define the part of energy that is obtained from the whole amount of obtained energy, using a random number generator. In Fig. A.1 we can see the variants of such temperature distribution.

As seen from the results obtained, the temperature distribution, which is responsible for collisions of accumulated bodies and particles, is very heterogeneous and during the time interval required for forming the next layer it does not become smooth. This is clearly seen in Fig. A.2 from [6], which presents a set of variants of temperature sections of the growing Moon with a subsequently growing radius.

The main peculiarities of the temperature distribution in the Moon's model composition are as follows: after reaching the end of the accumulation process,

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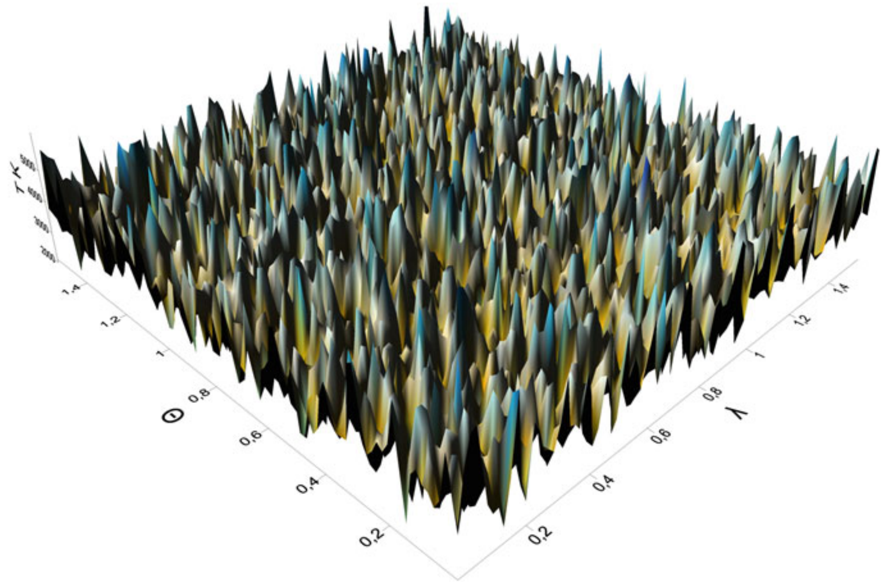

Fig. A.1 The variant of the temperature distribution, stipulated by the random distribution of the accumulated bodies by their values and kinetic energy on the planet's surface at the radius R = 1000 km [5]

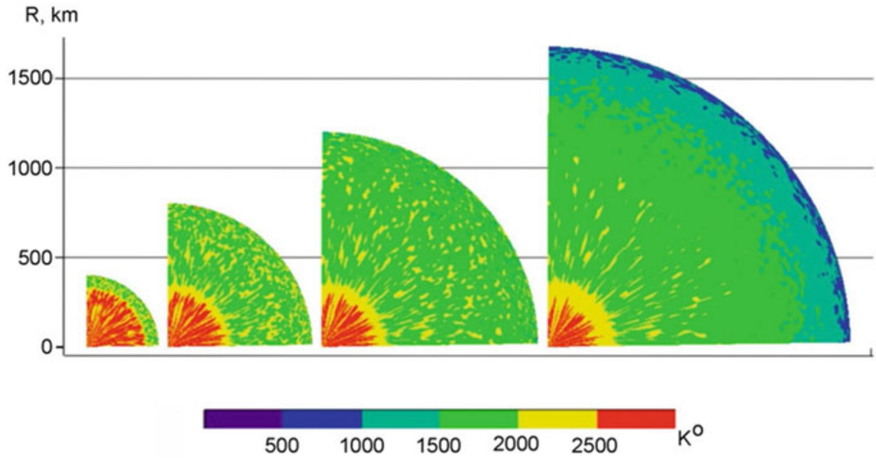

Fig. A.2 The variant of the temperature distribution for the radial Moon's sections of the successfully increasing radius. The thermal heterogeneities are stipulated by their random location, the variant of which is presented in Fig. A.1 [6] a small size melted inner core is formed, and in the mantle a thick melted and a partially melted layer of matter is formed.

The greatest difficulties are linked with the numerical solution of the Earth's thermal evolution, because we need to take into account the adiabatic compressibility of the matter, which leads to the increase of the matter density from 2310 kg/m^3 on the crust–mantle boundary to the density of 7200 kg/m^3 on the mantle–core surface.

The numerical solution of the problem of the temperature distribution in the inner parts of the Earth at the stage of accumulation in the model, when taking account of adiabatic compression, includes an additional numerical block. The initial conditions are defined in it. For the presented variant, the initial value of the radius is R_0 = 1000 m. The temperature inside and on the surface of the pre-planetary body at the initial time moment is T = 320 K, the density and the module of volume compression of iron in the pre-planetary body are $\rho_0 = 7.6 \times 10^3$ kg/m^3, $K = 16 \times 10^{10}$ PA, and the mass of the pre-planetary body is:

$$m_0 = \frac{4}{3} \cdot \pi \cdot R_0^3 \cdot \rho_0$$

The step of the planetary radius growing by passing on to the next time layer is established as const ΔR, while the step for time is variable and it is calculated at each stage of planetary growth from Safronov's equation (Chap. 4, (4.1)) [1]. It is assumed that in each layer with a thickness ΔR the values of the density, module of compression and pressure remain constant.

In each time layer the planetary radius increases such that: $R_{j+1} = R_j + \Delta R$

Using a 1-D spherical-symmetric model, we obtain a new value for the bodies' mass and the pressure distribution. The lithostatic pressure is defined, using the pressure of the upper layers:

$$p_i = \sum_{i=0}^{i=j} \rho_i \cdot g_i \cdot \Delta R_i \tag{A.1}$$

where g_i—acceleration of the gravity force of the sphere with radius R_i is equal to:

$$g_i = \frac{G \cdot m}{R_i^2} \tag{A.2}$$

where G—gravitational constant, m—mass of the sphere with the radius R_i.

The module of compression K according to the accepted definition is equal to:

$$K = \rho \cdot \frac{\Delta p}{\Delta \rho} \tag{A.3}$$

Thus, for the density distribution we obtain:

$$\rho_i = \rho_{i+1} + \rho_{i+1} \cdot \frac{p_i - p_{i+1}}{K_i} \tag{A.4}$$

Using Eqs. (A.1)–(A.3), we can obtain a new value of the bodies' mass and pressure in the layers. After that, we can derive the value of the compression module from (A.3):

$$K = \rho_i \cdot \frac{p_i - p_{i+1}}{\rho_i - \rho_{i+1}} \tag{A.5}$$

After calculating the physical parameters with a newly formed layer, we can calculate the time step from equation (Chap. 3, (3.1)) and the whole time of planetary growth for further numerical solution of the boundary problem equations (Chap. 4, (4.2)–(4.3)). The temperature distribution in the body with increasing radius is obtained from the numerical solution of the boundary problem for the equation of heat conductivity, taking account of the possibility of a melting area appearing without explicit release of the boundary location of the crystallization front and parametrical account of convective heat transfer in the melted area after [7]:

$$c_{ef}\,\rho \frac{\partial T}{\partial t} = \nabla(\lambda_{ef}\,\nabla T) + Q \tag{A.6}$$

where: c_{ef}, λ_{ef}—effective values of heat capacity, and heat conductivity, which take account of the melting heat in Stefan's problem after [8] and the existence of convective heat transfer; T—is the temperature at a point and a moment t, Q—is the volume capacity of the inner sources of heat. The problem is solved using the method of finite differences together with a whole implicit monotonic, conservative scheme. The dimensional step for the space grid is constant. The step for the time grid is variable and for the given distribution of the density, as a function of the depth, is calculated from equation (Chap. 4, (4.1)). Using that equation for each time step, we calculate the mass of the growing planet and the distribution of lithostatic pressure in the inner areas. For each value of the dimension reached by the growing planet, we calculate the distribution of the melting temperature. In the core the function of the melting temperature, mainly of the iron composition, is calculated after [9]:

$$\begin{gathered}
\rho\,[\frac{\partial \vec{V}}{\partial t} + (\vec{V}\,\nabla)\vec{V}] = -\nabla P + \eta \Delta\,\vec{V} + (\frac{\eta}{3} + \xi)\nabla(\nabla\vec{V}) - \rho\,\nabla W \\
\rho\,T\,[\frac{\partial S}{\partial t} + (\vec{V}\,\nabla)\,S] = \lambda\,\Delta\,T + Q \\
\Delta\,W_1 = -4\pi\gamma\,\rho \quad W = W_1 + W_2 \\
\frac{\partial \rho}{\partial t} + \nabla(\rho\,\vec{V}) = 0 \\
L\,\frac{\partial \vec{\psi}}{\partial t} = \vec{q}|_{\xi+0} - \vec{q}|_{\xi-0}
\end{gathered} \tag{A.7}$$

where: $\vec{V}$—fluid velocity, P—pressure, S—entropy, W_1—gravitational potential, W_2—centrifugal potential, ρ—density, η and ξ—coefficients of the first and second viscosity, λ—coefficient of thermal conductivity, γ—gravitational constant, Q—summarized capacity of the inner sources in the volume unit, L—heat of the phase transfer, $\frac{\partial \vec{\psi}}{\partial t}$—the velocity of the boundary location of the phase division, $\vec{q}|_{\xi+0}$ *and* $\vec{q}|_{\xi-0}$—density of the heat flow, correspondingly, in front of and behind the phase boundary, ∇ and Δ–"nabla" and Laplace operators.

In the area of the forming mainly silicate mantle, the relation of the temperature with pressure is used after [10]. The zone of complete or partial melting is defined in each time layer by comparison of the calculated temperature distribution with the melting temperature at the given depth. Nevertheless, the large gradients of the density and the temperature in the large areas, and the necessity of taking into account the dependence of the matter viscosity on temperature and pressure, lead to a need to derive the solution in a more comprehensive statement.

The solution of Eqs. (A.7, 1) and (A.7, 5) is obtained using the natural parameters: the vector of velocity and pressure [11, 12].

The solution of the boundary problems of the first equation (A.7, 2) of that system, which is known as the Navier-Stokes equation, presents difficulties. And

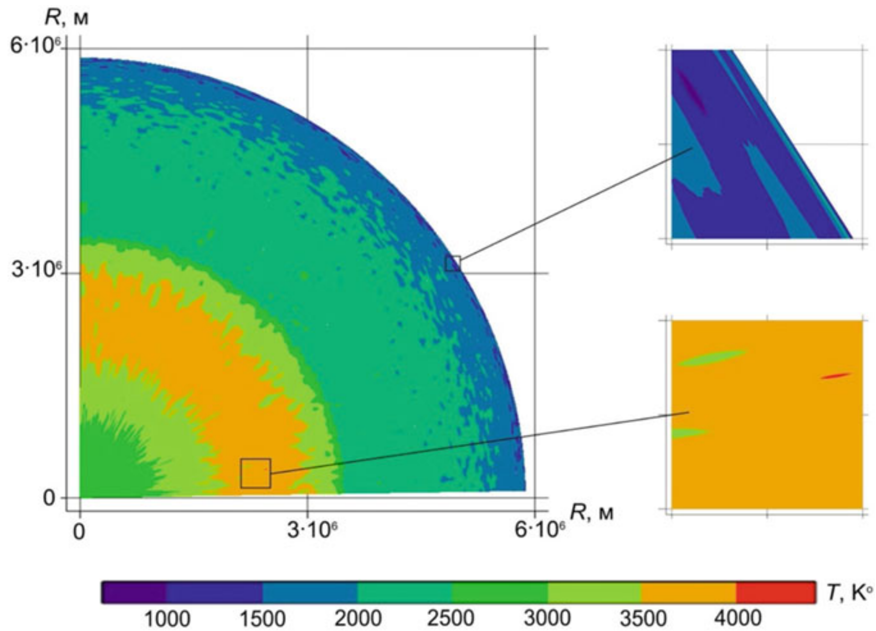

Fig. A.3 The variant of the temperature distribution for one section of the Earth's model up to the end of accumulation. One of the radial sections is shown. The layer of the silicate melt is formed on the core–mantle boundary. The structure of the thermal heterogeneities is shown in the small pictures [14] for the approximation with constant coefficients of viscosity, as used in (A.7, 3), for a 3-D spherical layer, obtaining a numerical solution represents an essential problem. Besides that, in the frame of equation (A.7, 2) it is hard to describe the forced mixing of convective matter near the surface of the growing body by the falling of some other bodies. Obtaining the solution in the frame of a 3-D model is a complicated and cumbersome problem [13], therefore the mass numerical modelling was made in the approximation (Chap. 4, 4.1–4.3) for the 3-D model.

For the geometrical model of the forming Earth we used the same model for the 1/8 part of the spherical volume as for the Moon model. The problem (4.1)–(4.3) was solved taking account of the adiabatic compression and the heat dissipation by that adiabatic compression. It is interesting to compare the obtained variants of the temperature sections of the Earth and the Moon with the finishing of their accumulation. The quantitative estimations of the temperature of these bodies in their central parts coincide; there they had been controlled by heat dissipation through the decay of short-living elements. But the larger mass of the Earth compared to the Moon explains the melted state of the Earth's outer iron core, as against the solid state of similar areas of the Moon. This may be the reason for the development of the magneto-hydro-dynamic mechanism of the Earth's magnetic field generation and its absence on the Moon (Fig. A.3).

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References

1. Safronov V (1969) Evolution of the protoplanetary cloud and formation of the Earth and the Planets. Nauka, Moscow (in Russian)
2. Artemieva N, Shuvalov V (2008) Numerical simulation of high-velocity impact ejecta following falls of comets and asteroids onto the Moon. Solar Syst Res 42:329–334 (in Russian)
3. Grae J, Appel P, Hatsukawa Y et al (2009) The Earth-Moon system during the late heavy bom-bardment period—geochemical support for impacts dominated by comets. Icarus 204:368–380
4. Khachay Y, Anfilogov V (2013) Numerical results of Earth's core accumulation 3-D modelling. EGU Geophys Res Abs 15:51
5. Khachay Y, Anfilogov V, Antipin A (2009) 3-D model of the thermal heterogeneity evolution on the Earth's core on its accumulation. In: Proceedings 5th Bulashevich conference geodynamics. Deep structure. The Earth's thermal field. Geophysical fields interpretation, Ekaterinburg, pp 321–323 (in Russian)
6. Khachay Y, Antipin A (2014) The results of numerical 3D simulation of temperature distribution in the lunar envelopes at its accumulation monitoring. Nauka i Technol 1:28–32 (in Russian)
7. Tikhonov A, Liubimova E, Vlasov V (1969) About the evolution of fusion zones in the thermal history of the Earth. Doklady academy of sciences of USSR 188:338–342 (in Russian)
8. Samarsky A, Moiseenko B (1965) The economic scheme of transparent account Stefan problem. J Calc Math 5:816–827 (in Russian)
9. Stacey F (1969) Physics of the Earth. Wiley, New York
10. Kaula W (1971) Dynamical aspects of lunar origin. Geophys Rev Space Phys 9:217–238

11. Belocerkovskiy O (1984) Numerical modelling in the mechanics of a continuous medium. Nauka, Moscow (in Russian)
12. Berkovskiy V, Nogotov G (1976) Numerical methods of investigation of heat exchanging problems. Nauka i Technika, Minsk (in Russian)
13. Khachay Y (2013) MHD process in the layer of gravitating sphere growing radius. Magnetohy-drodynamics 49(1–3):81–86
14. Khachay Y, Antipin A (2014) Numerical modelling of temperature distribution at the stage of Earth's accumulation in the frame of 3-D model and peculiarities of its initial minerageny. Ural Geophys Bull 1:81–85