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Chemical Differentiation of the Galilean Satellites of Jupiter: 4. Isochemical Models for the Compositions of Io, Europa, and Ganymede

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Abstract—Models for the composition and structure of the Galilean satellites of Jupiter (Io, Europa, and Ganymede) were constructed using geophysical data provided by the *Galileo* mission on the mass, average density, and moment of inertia, as well as thermodynamic data on the equation of the state of water, high-pressure ices, and meteoritic materials. The distribution of density, pressure, temperature, and gravity acceleration in the interiors of the satellites was determined. A simulation of the internal structure of the satellites showed the possibility of identical bulk compositions for water-free Io and the rock-iron cores of Europa and Ganymede (i.e., satellites without their outer ice–water shells). The sizes of the satellites' cores (Fe with 10 wt % S) and the thicknesses of the ice–water shells of Europa (120 km) and Ganymede (900 km) were also estimated. These satellites contain 7 and 47% H_2O, respectively. The radii of Fe–10% S cores are 737 km for Io, 695 km for Ganymede, and 576 km for Europa. The ratios of the radii and masses of the Fe–S cores and rock-iron cores of Io, Ganymede, and Europa are almost identical and equal R(Fe–10%S core)/R_{Cor} = 0.4 and M(Fe–10% S core)/M_{Cor} = 10.55 ± 0.3 wt %. It was shown that the geochemical parameters of the rock-iron constituent of the satellites are similar to the material of L/LL chondrites. The silicate fraction of the satellites contains about 16 wt % FeO and shows an Fe/Si mass ratio of 0.53. The total iron to silicon mass ratio is also identical in the three satellites: $(Fe_{tot}/Si)_{Cor} = 0.99 \pm 0.02$. This value is different from that in the bulk compositions of the most oxidized carbonaceous chondrites and the most reduced H chondrites. Io, Europa, and Ganymede could be formed in the accretion disk of Jupiter from a material similar to L/LL chondrites under relatively low temperatures, not higher than the evaporation temperature of Fe and Fe–Mg silicates.

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INTRODUCTION

The investigation of the outer regions of the solar system started by the *Pioneer* and *Voyager* missions in 1973–1979 stimulated the exploration of the internal structure, thermal history, and geologic evolution of the regular (Galilean) satellites of Jupiter [1–11]. This work provided a basis for the modern models of the Galilean satellites and predicted many unique features of their structure, including the possibility of the existence of a liquid layer at the base of Europa's icy crust, the degree of differentiation of Ganymede and Callisto, and vigorous volcanic activity on Io.

The *Galileo* spacecraft was launched in 1989 and reached Jupiter's orbit in December 1995. Numerous *Galileo* probe flybys of the Jovian satellites, Io, Europa, Ganymede, and Callisto, provided detailed information on their gravitational and magnetic fields. An analysis of the data on gravitational fields allowed estimating the moment of inertia and average density [12–15] of water-free Io and ice-bearing or ice–water-bearing Europa, Ganymede, and Callisto (Table 1). This information served as a basis for the construction of modern models for the internal structure of the Galilean satellites and the estimation of the distribution of density and chemical composition [16–29]. It was shown that Io, Europa, and Ganymede have experienced extensive differentiation to a metallic Fe–FeS core, a silicate mantle, and an ice or water–ice shell (Ganymede and Europa). In contrast to these three satellites, Callisto was not differentiated to an ice shell, a mantle, and a Fe–FeS core [15, 23, 25, 29]; its heat sources sufficed only for the partial differentiation of Callisto. The outer water–ice shell of Callisto is underlain by the mantle consisting of a mixture of H_2O ices, silicate rocks, and metal alloy of the Fe–FeS composition [25].

Hereafter, the rock-iron core of a satellite (or the Fe–Si core for brevity) refers to its water-free inner shell differentiated into the silicate crust, mantle, and central iron–sulfide (Fe–FeS) core. Io having no ice shell consists of a rock-iron core only. In order to keep the common terminology, the rock-iron core of Io is equivalent to the whole satellite.

A comparison of estimates for the chemical composition of Io and geochemical data on ordinary and car-

Table 1. Physical parameters and internal structures of the Moon and Jupiter's satellites

Parameter	Moon	Io	Europa	Ganymede
M_{Sat}, km	0.31×10^{23}	0.89×10^{23}	0.48×10^{23}	1.48×10^{23}
I/MR^2	0.3931 ± 0.0002	0.37685 ± 0.00035	0.346 ± 0.005	0.3105 ± 0.0028
ρ, g/cm^3	3.3437 ± 0.0016	3.5278 ± 0.0029	2.989 ± 0.046	1.936 ± 0.022
R_{Sat}, km	1738	1821.3	1565	2634
R_{Cor}, km	1738	1821.3	1445	1734
$R_{Fe-10\%\,S}$, km	445	737	576	695
$M_{Fe-10\%\,S}/M_{Sat}$, %	2.9	10.91	9.52	5.52
$M_{Fe-10\%\,S}/M_{Cor}$, %	2.9	10.91	10.25	10.48
M_{Cor}/M_{Sat}	1	1	0.929	0.533
M_{Ice}/M_{Sat}	0	0	0.071	0.47
H_{Ice}, km	0	0	120	900
$(Fe_{tot}/Si)_{Cor}$	0.47	1.00	0.97	0.99
$P_{m\text{-}cor}$, kbar		58.5	36	73
P_0, kbar	50	79.7	49	93.4

Note: R_{Sat}, ρ, and I/MR^2 are the radius, average density, and dimensionless moment of inertia of the satellite; R_{Cor} is the radius of the rock-iron core; $R_{Fe-10\%\,S}$ is the radius of the central core; $M_{Fe-10\%\,S}/M_{Sat}$ is the ratio of the mass of the central Fe–10% S core to the total mass of the satellite; $M_{Fe-10\%\,S}/M_{Cor}$ is the ratio of the mass of the central Fe–10% S core to the mass of the rock-iron core; H_{Ice} is the thickness of the outer shell (water–ice for Europa and ice for Ganymede); M_{Ice}/M_{Sat} is the mass fraction of ice; $(Fe_{tot}/Si)_{Cor}$ is the mass ratio of total iron to silicon in the rock-iron core; $P_{m\text{-}cor}$ is the pressure at the boundary between the mantle and Fe–10% S core; and P_0 is the pressure in the center of the satellite.

bonaceous chondrites suggested [20–22] that the bulk chemistry of Io is not consistent with the material of C chondrites, because of the violation of constraints on the mass and moment of inertia and it is more similar to the compositions of L and LL chondrites. It was also demonstrated that CM, CV, and L/LL chondritic compositions can be regarded as models for either the primary material of Europa (carbonaceous chondrites) or its anhydrous rock-iron material (ordinary chondrites) [24]. It was also supposed that the composition of the rock-iron core of Europa is closer to L/LL chondrites than to carbonaceous chondrites with respect to the degree of oxidation. Similarly, the bulk composition of Ganymede's rock-iron core similar to L/LL chondrites is in agreement with the known values of its mass and moment of inertia [22]. Thus, it is supposed that the rock-iron cores of Io, Europa, and Ganymede are probably composed of a material chemically similar to L/LL chondrites.

The average density of the Galilean satellites decreases with increasing distance from Jupiter, which indicates a higher H_2O abundance in the composition of the outer satellites: from 0 in water-free Io, 6–9% in Europa, to 45–55% in Ganymede and Callisto [22–25]. At great heliocentric distances, water ice becomes the major component of planetary bodies, such as the satellites of giant planets, Uranus and Neptune, and cores of comets. However, this fact has no bearing on the bulk composition and average density of the rock-iron cores of the Galilean satellites. It was supposed [5, 30, 31] that the high luminosity of proto-Jupiter prevented ice condensation in the innermost orbits (Io), but not in the more distant orbits of icy satellites, where ice condensed from the disk at $T < 250$ K [32]. The condensation of ice did not occur in that part of the nebula where Io accumulated at $T > 500$ K.

An increase in the H_2O fraction with increasing distance from the central body supports the model of heating of the inner parts of the satellite swarm by a hot giant planet and explains the different densities of the inner and outer satellites. This is related to the fact that the concentrations of condensing major elements could either be almost constant or decrease, while the concentrations of volatile components had to increase with increasing orbital distance during disk cooling. If this is the case, Io and the rock-iron cores of the icy satellites must be either similar to each other and have almost identical compositions and Fe_{tot}/Si ratios or fall on a common compositional trend with the distance from Jupiter. This trend must be identical to that observed for terrestrial planets with respect to their distance from the Sun (a decrease in the average density and Fe_{tot}/Si ratio from Mercury to Mars). In the former case, the Jovian satellites formed during disk cooling must differ in water content but not in bulk composition, i.e., the element ratios, such as Fe_{tot}/Si, must be approximately constant. In the latter case, the density and Fe_{tot}/Si ratio of the rock-iron cores of the satellites must decrease with an increasing distance from Jupiter.

The occurrence of a particular scenario is controlled by the physicochemical parameters of the protosatellite disk of Jupiter (temperature, pressure, H_2O/H_2 ratio, thermal state, viscosity, and mass) and the conditions of accretion (collision and crushing of bodies, their partial or total evaporation, and recondensation). Early models were based on the hypothesis of Jupiter's formation by gravitational instability in the near-solar disk. According to modern concepts [30–38], the Galilean satellites were generated from the material of the accretionary gas–dust disk that existed around Jupiter during its formation and was much smaller in size and mass than the solar protoplanetary nebula. This approach is based on the assumption of the accretionary character of the protosatellite disk of Jupiter (similar to the solar disk), i.e., the main mass of gas and dust material that entered Jupiter's disk moved in the radial direction and fell then on the central body, Jupiter. Studies on the simulation of satellite formation processes [34–36, 38] considered both massive (hot) and low-mass (relatively cold) models for the protodisk. These studies showed that it is difficult to construct a model that was simultaneously compatible with constraints on the distribution of temperature in the accretionary disk (in agreement with an increase in water content in the Galilean satellites) and on the minimum mass of the disk estimated from the mass of the Galilean satellites. The disk compatible with the temperature constraints appeared to be too light, and that compatible with the mass constraints was too hot [34, 36].

Calculations of P–T conditions for the massive model showed that iron and magnesium silicates are partially evaporated in the inner zone of the disk at $r < 17\ R_J$ (i.e., in the orbits of Io, Europa, and Ganymede). This could result in the fractionation of the dust fraction and the Fe_{tot}/Si ratio depending on the distance from Jupiter [35, 36]. In contrast, under the P–T conditions of the low-mass (cold) model, metals and silicates are not affected by evaporation and selective fractionation, and the chemical features of Jupiter's satellites must be related to the compositions of solid ice and ice–rock planetesimals captured by the disk and affected in it by high-velocity collisions [34]. Such models of the protodisk of Jupiter are compatible with the identical chemical compositions of the rock-iron cores of the satellites. Kuskov and Kronrod [19] hypothesized that Io, Europa, and Ganymede have structurally similar and chemically identical rock-iron cores and differ from one another only in the thickness and structure of the outer ice (water–ice) shell. In such a case, water-free Io (i.e., a satellite with an icy shell of zero thickness) can represent the material of the nonvolatile fraction of Jupiter's disk.

The estimates of the moment of inertia of Europa calculated on the basis of the hypothesis of the similarity of the rock-iron cores of the satellites coincided with experimental measurements [18, 19], which provided indirect support for this hypothesis. It was shown [20–22] that the rock-iron cores of the three satellites (Io, Europa, and Ganymede) could be formed from a composition similar to L/LL chondrites, which suggests that the bulk satellites without icy shells are probably isochemical. However, the models of Europa and Ganymede are also compatible with bulk compositions different from the composition of L/LL chondrites [22, 24]. An important argument in favor of the isochemical character of the rock-iron components of the Galilean satellites can be obtained from the construction of a model of these satellites with identical bulk compositions of rock-iron cores. This problem is addressed in detail in this paper.

INTERNAL STRUCTURE OF THE ROCK-IRON CORES OF THE SATELLITES

Moment of Inertia and Average Density of Rock-Iron Cores

Let us consider a model satellite consisting of a water–ice shell, the thickness of which varies from zero for Io to H_{Ice} for Europa and Ganymede, and a rock-iron core. The conservation equations for the mass and momentum of the satellite can be written as

$$\begin{aligned} M_{Ice} + M_{Cor} &= M_{Sat}, \\ I^0_{Ice} + I_{Cor}M_{Cor}R^2_{Cor} &= I_{Sat}M_{Sat}R^2_{Sat}, \end{aligned} \tag{1}$$

where M is the mass, R is the radius, and I is the reduced moment of inertia of the satellite ($I = I^0/MR^2$, where I^0 is the total moment of inertia) according to [12–14] (Table 1); the subscripts Sat, Ice, and Cor refer to the whole satellite, the outer ice or water–ice shell, and the rock-iron core, respectively. The mass and moment of inertia of a spherically symmetrical body can be calculated from the equations

$$\begin{aligned} M &= 4\pi \int_0^{R_{Sat}} \rho(r) r^2 dr, \\ I^0 &= \frac{8}{3}\pi \int_0^{R_{Sat}} \rho(r) r^4 dr, \end{aligned} \tag{2}$$

where ρ is the density, and r is the radial coordinate. The reduced moment is hereafter referred to merely as moment.

Using Eqs. (1) and (2) with the equations of state for water and high-pressure ices [23], the density (ρ_{Cor}) and moment of inertia (I_{Cor}) of the rock-iron core can be determined as functions of the thickness of the outer shell (H_{Ice}) (Figs. 1a, 1b). It can be seen that, at certain H_{Ice} values, the curves $\rho_{Cor} = \rho_{Cor}(H_{Ice})$ and $I_{Cor} = I_{Cor}(H_{Ice})$ for Europa ($H_{Ice} \approx 105$–120 km) and Ganymede ($H_{Ice} \approx 870$ km) pass close to the parameters of Io ($I_{Io} = 0.37685 \pm 0.00035$ and $\rho_{Io} = 3.5278 \pm 0.0029$ g/cm^3 [14]). This means that the average densities and moments of inertia of the Fe–Si cores of Europa and Ganymede can be identical or similar to the

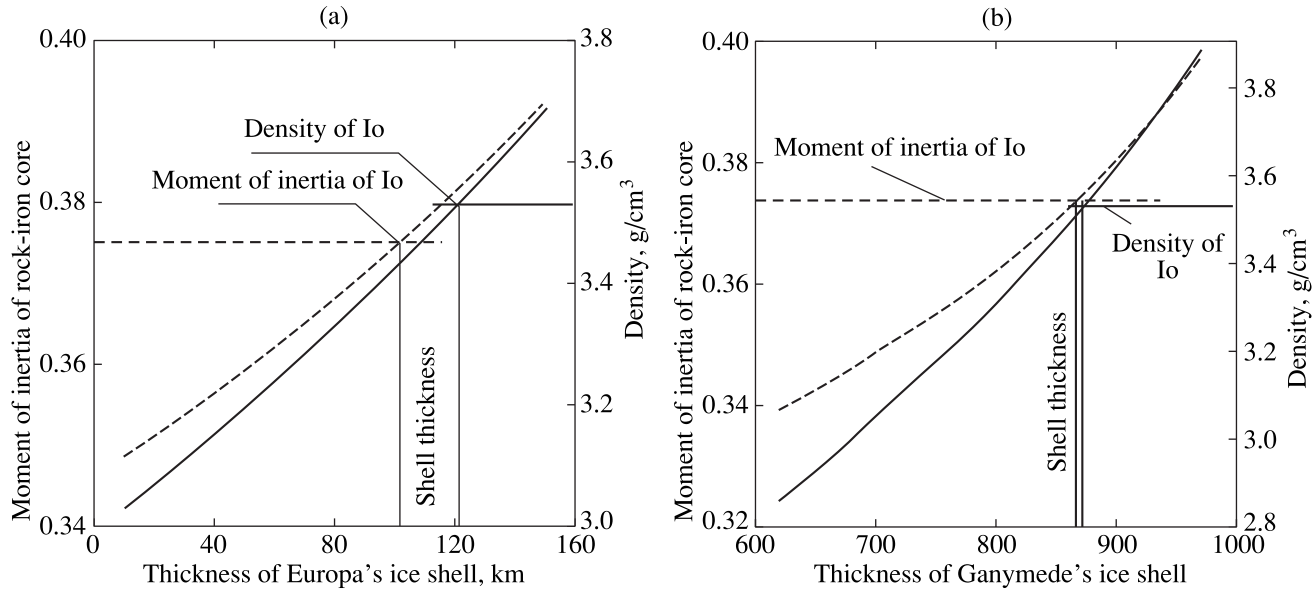

Fig. 1. Moment of inertia (dashed line) and average density (solid line) of the rock-iron cores of (a) Europa and (b) Ganymede as functions of the thickness of the ice shell, H_{Ice}. Under certain H_{Ice} values, the curves $\rho_{Cor} = \rho_{Cor}(H_{Ice})$ and $I_{Cor} = I_{Cor}(H_{Ice})$ for Europa (H_{Ice} = 105–120 km) and Ganymede ($H_{Ice} \approx 870$ km) pass near the point of Io, $I_{Io} = 0.37685 \pm 0.00035$ and $\rho_{Io} = 3.5278 \pm 0.0029$ g/cm^3 [14].

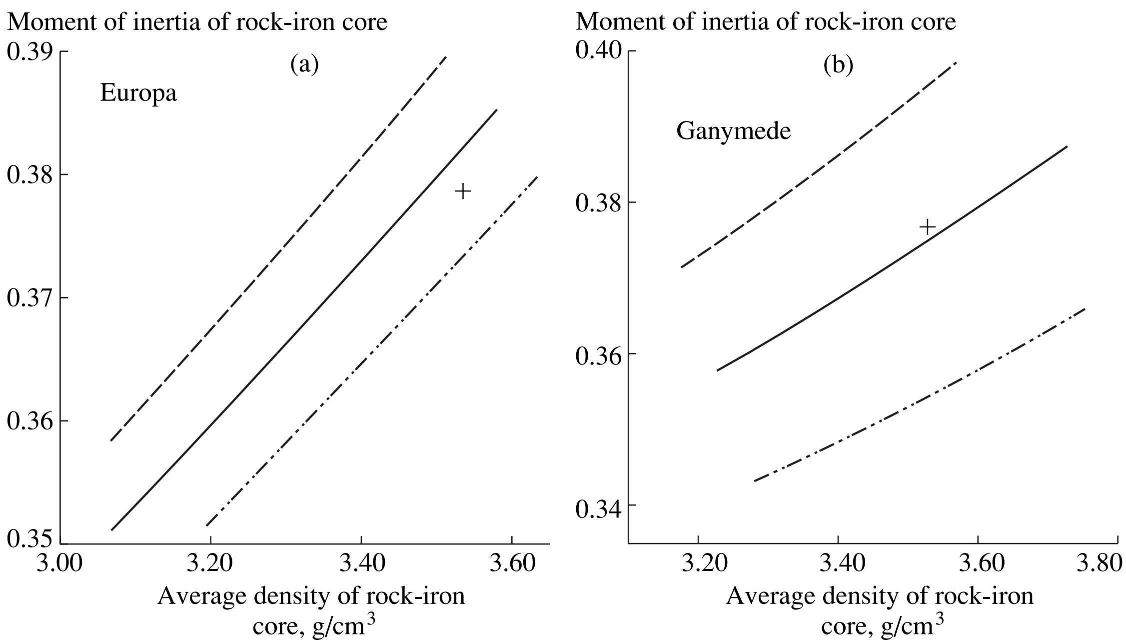

Fig. 2. Moments of inertia and average densities of the rock-iron cores of Europa and Ganymede at various values of the total moment of inertia of the satellites. The cross indicates the parameters of Io [14]. (a) Europa. The solid line corresponds to the experimental value of the moment of inertia of Europa, $I = 0.346$ [13]; the dashed line, $I = 0.353$; and the dash–dot line, $I = 0.339$. (b) Ganymede. The solid line corresponds to the experimental value of the moment of inertia of Ganymede, $I = 0.3105$ [12]; the dashed line, $I = 0.3155$; and the dash–dot line, $I = 0.3055$. It can be seen that the solid lines pass near the point of Io.

parameters of water-free Io. The question is whether this coincidence is accidental or related to the compositions and properties of materials composing the satellites.

The moments of inertia and densities of satellites depend on the physicochemical characteristics of their Fe–Si cores and water–ice shells. Let us derive the dependency of the moment of inertia of a rock-iron core on its density, $I_{Cor} = I_{Cor}(\rho_{Cor})$. To this end, for the given thickness of the water–ice shell, H_{Ice}, its moment of inertia (I^0_{Ice}) and mass (M_{Ice}) are calculated. Then, I_{Cor} and ρ_{Cor} are determined from Eq. (1). According to Eq. (1), any given values of H_{Ice}, I_{Sat}, and ρ_{Sat} correspond to a single pair of values for the moment of inertia (I_{Cor}) and density (ρ_{Cor}) of the rock-iron core, i.e., a single point in Fig. 2. Varying the thickness of the water–ice shell, we obtain a single curve, $I_{Cor} = I_{Cor}(\rho_{Cor})$, for the experimental values of I_{Sat} and ρ_{Sat} (Fig. 2). Therefore, the existence of a point in which the moment of inertia and density of Io coincide with those of the rock-iron cores of Europa and Ganymede is possible only under certain values of the moment of inertia and density of Europa and Ganymede. If the moments of inertia of Europa and Ganymede were different from the measured values, the functions $I_{Cor} = I_{Cor}(\rho_{Cor})$ for these satellites would not pass near the point corresponding to the properties of Io. In order to illustrate this statement, the moments of inertia of the satellites are changed in Fig. 2 by two times the uncertainties given in Table 1. Indeed, an increase in the moment of Europa by 0.007 results in that the point closest to Io has coordinates of I_{Cor} = 0.382 and ρ_{Cor} = 3.43 g/cm^3, which are significantly different from the parameters of Io (Fig. 2a). Even greater deviations from the coordinates of Io were observed when the moment of inertia of Ganymede was increased (decreased) by 0.005 (Fig. 2b). Thus, the relationships shown in Fig. 2 demonstrate that the coincidence of I_{Cor} and ρ_{Cor} values for the three bodies is not accidental and must be related to the compositions and properties of the Fe–Si cores of the satellites.

In such a case, the following similarity conditions must hold:

$$I^{E}_{Cor} = I^{G}_{Cor} = I_{Io},\qquad \rho^{E}_{Cor} = \rho^{G}_{Cor} = \rho_{Io}, \tag{3}$$

which, as will be shown below, could result from the equality of the chemical compositions of the Fe–Si cores.

Distribution of Density in the Satellites

Similarity conditions (3) allow us to make important conclusions on the distribution of density in the interiors of the satellites. The analysis of Eqs. (2) and (3) provides the following possible law of density distribution in the interiors of the satellites:

$$\rho^{E}_{Cor}(r^0) = \rho^{G}_{Cor}(r^0) = \rho_{Io}(r^0),\quad r^0 = (r/R_{Cor}). \tag{4}$$

The moments of inertia and densities of the rock-iron cores of the three satellites are equal, if the values of density (ρ) in the Fe–Si cores are identical at a given value of the reduced radius r^0. An example of satisfying condition (4) is presented below.

Let us consider a two-layer model for a satellite consisting of an Fe–FeS core and silicate mantle with a constant density in each layer. Under a given core density, only two variables must be determined from the system of conservation equations for the mass and moment of inertia of the satellite Eq. (1): the density of the mantle and the radius of the Fe–FeS core. If the moments of inertia and densities of the rock-iron cores of the three satellites are equal, Eq. (4) implies the identity of mantle densities and dimensionless radii r^0 at the core–mantle boundary:

$$\rho^{E}_{Man} = \rho^{G}_{Man} = \rho^{Io}_{Man},\qquad (r^0)^{E}_{Cor} = (r^0)^{G}_{Cor} = (r^0)^{Io}, \tag{5}$$

where $r^0 = R_{Fe–FeS}/R_{Cor}$, and $R_{Fe–FeS}$ is the radius of the central Fe–FeS core.

For the models of satellites with identical densities of Fe–FeS cores, Eq. (5) implies the identity of the mass ratios of the Fe–FeS core to the whole rock-iron core (i.e., to the mass of silicate crust + mantle + Fe–FeS core):

$$M^{E}_{Fe–FeS}/M^{E}_{Cor} = M^{G}_{Fe–FeS}/M^{G}_{Cor} = M^{Io}_{Fe–FeS}/M^{Io}. \tag{6}$$

It was previously shown that the mantle density of Moon-sized satellites weakly changes with depth, because the effects of pressure and temperature on the current values of density largely cancel out [39]. Variations in the density of the core from the troilite composition to pure iron from 4.7 to 8.0 g/cm^3 also exert a negligible influence on the distribution of density in the mantle of the satellites [22]. Therefore, as a first approximation, the density of the mantle of the satellites can be estimated from two-layer models. In such a case, the conditions of similarity Eq. (3) are equivalent for satisfying conditions (5) and (6).

Hypothesis of the Isochemical Material of the Satellites

Kuskov and Kronrod [21, 22] showed that the composition of the mantle can be estimated from density distribution. Therefore, given the equality of the average densities of the mantles of Io, Europa, and Ganymede, it can be supposed that the compositions of the silicate mantles of the three satellites are identical. The isochemical character of mantle materials and Eq. (6) yield the following geochemical similarity conditions:

$$\begin{gathered}(Fe_{tot}/Si)^{G}_{Cor} = (Fe_{tot}/Si)^{E}_{Cor} = (Fe_{tot}/Si)_{Io},\\ FeO^{G}_{Sil} = FeO^{E}_{Sil} = FeO^{Io}_{Sil},\\ (Fe_m/Fe_{tot})^{G}_{Cor} = (Fe_m/Fe_{tot})^{E}_{Cor} = (Fe_m/Fe_{tot})_{Io},\end{gathered} \tag{7}$$

where $(Fe_{tot}/Si)_{Cor}$ is the mass ratio of total iron to silicon in the rock-iron core of the satellite; FeO_{Sil} is the mass concentration of FeO in the silicate crust and mantle, and $(Fe_m/Fe_{tot})_{Cor}$ is the mass ratio of the abundance of metallic iron (Fe_m) in the central Fe–FeS core to the bulk content of iron in the whole Fe–Si core of the satellite.

Thus, we obtained geophysical (Eq. (3)) and geochemical (Eq. (7)) similar conditions for the internal structures of Io and the Fe–Si cores of Europa and Ganymede. In fact, there is no exact coincidence of the moments of inertia and densities of the rock-iron cores of Europa and Ganymede and the parameters of Io (Figs. 1, 2). The observed discrepancies can be attributed to differences in the distribution of temperature and pressure in the interiors of the satellites. It is reasonable to expect that accounting for the effects of compressibility and thermal expansion will provide more accurate parameters for the Fe–Si cores and the thicknesses of the water–ice shells of Europa and Ganymede.

PROBLEM FORMULATION FOR THE NUMERICAL SIMULATION OF THE INTERNAL STRUCTURE OF IO, EUROPA, AND GANYMEDE

Similar conditions (3) and (7) for the internal structures of Io and the Fe–Si cores of Europa and Ganymede imply identical chemical compositions of the rock-iron constituents of the three satellites. In order to test the isochemical hypothesis, we constructed models for the satellites compatible with the main geophysical (mass and moment of inertia) and geochemical constraints (composition of the metallic Fe–FeS core and the silicate fraction of the Fe–Si component) and minimizing the deviation of the solutions from the similar conditions, i.e., the function $\delta f_{S,Z} = |f_S - f_Z|$, where

$$f = \{\rho_{Sil}, (Fe_{tot}/Si)_{Cor}, FeO_{Sil}, (Fe_m/Fe_{tot})_{Cor}\},$$
$$S = E, G, Io; \quad Z = E, G, Io; \text{ and } S \neq Z, \tag{8}$$

where δf is the objective function (residual of f); ρ_{Sil} is the density of the silicate fraction of the rock-iron core, i.e., the average density of the material of core and mantle recalculated to the reference values of pressure (P_1) and temperature (T_1), and Fe_m is the abundance of metallic iron in the central Fe–FeS core.

It was not required that conditions (3) and (7) were exactly satisfied, because the temperature profile and density in the mantle and core were determined mainly from approximate models. The conditions of the minimization of residuals (8) were replaced by the inequalities

$$\delta f \leq \Delta f, \tag{9}$$

where Δf is the maximum permissible residual for the function f. The values of Δf are prescribed. In our case, taking into account the assumption in the problem formulation, misfits of 2% were assigned for Fe_{tot}/Si and 0.5% for the average density of the silicate fraction of the Fe–Si core (ρ_{Sil}). The fulfillment of conditions (8) and (9) under such uncertainties implies that the chemical compositions of the rock-iron cores of Europa, Ganymede, and Io can in principle be identical.

The chemical compositions and physical properties of the satellites can be further constrained using the compositions of the silicate fraction of ordinary (H, L, and LL) and carbonaceous (CI, CM, and CV) chondrites. That is, the composition of the Fe–Si cores is constrained to lie between those of reduced H chondrites and oxidized C chondrites. The chemical composition of the silicate fraction of the Fe–Si cores was determined in the course of the solution.

The composition of chondrites was taken from [40] and recalculated to the volatile-free Na_2O–TiO_2–CaO–FeO–MgO–Al_2O_3–SiO_2–Fe–FeS system (NaTiCFMAS–Fe–FeS). Equilibrium phase assemblages were calculated by Gibbs free energy minimization using the THERMOSEISM program package and database [22]. The NaTiCFMAS system included the following phases: binary solutions of olivine, plagioclase, ilmenite, and spinel; pyrope–almandine–grossular garnet; five-component orthopyroxene solution; and six-component clinopyroxene solution. The mixing functions of solid solutions were described by regular and subregular models. The database includes the mutually consistent thermodynamic properties of minerals, and their equations of state were calculated using the Mie–Grüneisen–Debye approximation [41]. The errors in the density of phase associations were usually no higher than 1%. This approach is described in detail elsewhere [21–24].

Sulfur is one of the most important minor elements in the compositions of the metallic Fe–Ni–S cores of planetary bodies. Since iron is present in meteorites both as metal and as FeS, Kuskov and Kronrod [20–22] considered various models of the composition of satellite cores, from pure iron to troilite. They showed that the suggestion that Io's core consists of troilite (FeS) and has a mass of 18–20% is not consistent with any meteorite composition [20, 21]. The eutectic Fe–FeS composition also contains excess sulfur compared with chondrites. In this study, we accepted a model with a central Fe–10% S core containing 10 wt % S ($Fe_{0.84}S_{0.16}$) with $\rho = 5.7$ g/cm^3 at 50 kbar and 1500°C [23, 42–44]. It was assumed that the core is homogeneous in composition and density.

Since Io and Europa are located at close orbital distances from the central body (6.0–9.5 R_J), Jupiter influences the satellites forming considerable tidal waves, the dissipation of which into heat has a pronounced effect on the composition and structure of the satellites. This heat source is probably responsible for the volcanism of Io, as well as the cryovolcanic activity, water–ice diapirism, formation of surface faults, and expansion of the interiors of Europa. The presence of metallic cores suggests that the interiors were heated to high temperatures sufficient for the dehydration of water-bearing minerals. With this in mind, we assumed that the mantles of the three satellites (Io, Europa, and Ganymede) consist of dehydrated silicates. The simulation of the chemical and mineral compositions of the mantles and cores of these satellites was performed for the "dry" NaTiCFMAS-Fe–FeS system free of water and other volatiles.

Models of Europa and Ganymede

Numerical solutions resulted in the models of Europa and Ganymede differentiated into an outer H_2O shell, a mantle, and a Fe–FeS core [22, 24]. By analogy with other planetary bodies (e.g., the Earth and Moon), the appearance of a light crust is expected during the differentiation of the satellites. The calculations showed that the (Fe_{tot}/Si) ratio is much more sensitive to variations in the composition of model chondritic mantle than to the thickness and density of the crust [24]. Therefore, the density of the crust was accepted by analogy with the lunar anorthositic crust with a density of ~3.0 g/cm^3 and a thickness of 60 km [45]. The density of Europa's crust was taken to be 2.7 g/cm^3 on the surface and 3.0 g/cm^3 at the crust–mantle boundary with a linear dependency on the depth. Similar density distribution was accepted for Ganymede's crust with a small correction for pressure due to the overlying ice shell. The thickness of the crust was a fitted parameter calculated from relationships (8) and (9).

The physical properties and phase state of the outer shells of Europa and Ganymede are controlled by phase equilibria in the H_2O system, including water, ice I, and high-pressure ice modifications. The latter are characteristic of Ganymede and are lacking in Europa. The densities of these phases are given in [23] as functions of P and T. According to the phase diagram of H_2O [23, 46], the melting temperature (T_m) of ice I decreases with increasing P, and the minimum T_m value (251 K at 2.07 kbar) is attained in the ternary point (water + ice I + ice III). Owing to such peculiar behavior of water, the liquid phase (ocean) can exist beneath the ice cover of satellites under favorable T conditions. However, this possibility is still a matter of debate. The existence of a liquid layer has received some support from the data of the *Galileo* mission on the morphology of the ice surface of Europa and the magnetic field of the satellites [16, 47, 48].

A layer of liquid water can exist in the outer shell of Europa beneath the ice cover, which has a thickness in the range 3–30 km [49, 50]. In our model, the thickness of the ice shell of Europa was taken to be 10 km, and it is underlain by the ocean extending to the boundary with the crust. Two models were considered for the composition of Ganymede's outer shell [22, 51]: (1) polymorphous modifications of ice, and (2) a layer of ice I, 100–140 km thick, underlain by a liquid water layer. The density was calculated from the equations of state of water and high-pressure ices [23]. Note that the thicknesses of both the outer shells of Europa and Ganymede and their silicate crusts were determined from conditions (8) and (9).

Model of Io

Io is a satellite differentiated into an outer shell (silicate crust + asthenosphere), mantle, and an iron–sulfide core [20–22]. According to modern concepts, the energy of tidal heating exerts a significant influence on the volcanic activity and structure of Io. The calculations of the thermal history of Io [52] showed that the ordinary abundances of radioactive elements (similar to those of the Earth and meteorites) cannot explain the volcanic activity of Io. The spectral observations of the *Galileo* spacecraft documented brightness temperatures of up to 1500 K for some hot spots, which provides an estimate of about 1700 K for the erupting lavas [53]. Such high temperatures correspond to MgO-rich ultrabasic melts. These observations are insufficient for the determination of the composition and thickness of the crust and the composition of the erupting lavas. Most authors explain the properties of Io by heating owing to the dissipation of tidal deformations [54, 55]. It is universally accepted that tidal heating has played a key role in the thermal and geochemical evolution of Io. The existing models differ in assuming dissipation either in the whole mantle or in a relatively thin asthenospheric layer.

The mass and radius of Io are similar to those of the Moon, but in contrast to the latter, the extensive internal heating due to the dissipation of tidal energy controls the present-day thermal regime of Io. Our model for Io consists of a thin solid crust, a partially molten asthenosphere, a solid mantle, and a Fe–FeS core [6, 9, 21]. The distribution of hot spots on the surface of Io is more consistent with the existence of an asthenospheric layer than with the extensive melting of the mantle [56]. We admit that several limited regions of partial melting can be present in the mantle [57], but they have no considerable influence on the average density characteristics of the mantle. According to various authors, the thickness of Io's outer shell (solid crust + asthenosphere) ranges from 30 to 90 km [11, 21]. In our case, the thickness of the outer shell is determined by calculations, with the density of the solid crust, 1.5 km thick, taken as 2.15 g/cm^3. The density of the asthenosphere varies from 2.2 g/cm^3 at the boundary with the solid crust to 3.25 g/cm^3 at the boundary with the mantle [26] with a linear dependency on depth.

Distribution of Temperature in the Satellites

The distribution of temperature is assigned using the available thermal models of Jupiter's satellites. It is assumed that uncertainties in the model of the temperature field have a negligible influence on the main calculated parameter of the model, the distribution of density in the mantle [29].

Europa and Ganymede. A surface temperature of T_0 = 130 K was taken for the satellites [2, 49]. Conductive heat transfer and linear temperature variations were assumed for the ice cover from the surface to a depth of

10 km. Convective heat transfer [58] and adiabatic T distribution were taken for the deeper zones of the ocean and ice shell of the Ganymede. The silicate crust of Europa and Ganymede is characterized by linear temperature variations. The temperature at the crust–mantle boundary ($T_{Cr–M}$) was determined from conditions (8) and (9).

The mechanism of heat transfer in the mantle of Europa and Ganymede is unknown. Theoretically, both conductive and convective mechanisms are possible. Assuming an enrichment of radioactive elements in the material of the Galilean satellites, similar to meteoritic abundances, it is reasonable to suppose that heat transfer in the mantle of the rock-iron cores of the satellites is similar to that operating in the Moon, i.e., via the conductive mechanism [59]. An additional constraint on the distribution of temperature is the absence of density inversions in the mantle. Within the model of a solid mantle, the maximum T values must be below the solidus. It was previously shown [39] that, within each zone of the lunar mantle, the density profile is almost invariant with depth, i.e., the temperature profile in the lunar mantle is such that the temperature- and pressure-related variations in density tend to cancel each other. Based on such an analogy, a temperature profile is constructed for the mantle of Europa and Ganymede providing the minimum density gradient with depth under the condition $d\rho/dH > 0$. The temperature profile in the Fe–10% S cores is taken to be adiabatic, i.e., the temperature is almost constant near the center of the core. Based on the above assumptions, the temperature variations in the rock-iron cores of Europa and Ganymede were approximated by following expressions:

$$\begin{aligned}
T_{\mathrm{Cor}}^{\mathrm{E}} &= T_{\mathrm{Cr–M}}^{\mathrm{E}} + 1966H_0 - 1416.1H_0^2 + 211.7H_0^3 \\
&\qquad (H_0 < 0.9), \\
T_{\mathrm{Cor}}^{\mathrm{G}} &= T_{\mathrm{Cr–M}}^{\mathrm{G}} + 2606H_0 - 2662.9H_0^2 + 906.7H_0^3 \\
&\qquad (H_0 < 0.9), \\
dT_{\mathrm{Cor}}^{\mathrm{E}}/dH &= 0, \quad dT_{\mathrm{Cor}}^{\mathrm{G}}/dH = 0 \quad (H_0 \geq 0.9), \\
H_0 &= (H - H_{\mathrm{Cr}})/(R_{\mathrm{Cor}} - H_{\mathrm{Cor}}),
\end{aligned} \tag{10}$$

where H is the distance from the surface of the Fe–Si core ($H \geq H_{Cr}$, where H_{Cr} is the thickness of the crust), and $T_{Cr–M}^{E}$ and $T_{Cr–M}^{G}$ are the temperatures at the crust–mantle boundaries determined from the solution.

Io. The temperature of Io depends in a complex manner on thermal history, energy sources, and heat transfer mechanisms. There is no universally accepted opinion on the temperature conditions in the interiors of Io, but it is reasonable to suggest that the mantle temperature is close to the solidus. A linear law of temperature variations was accepted for the solid crust of Io, from the surface (130 K) to the asthenosphere boundary. Heat transfer in the asthenosphere is most likely convective [55], and the temperature profile was taken to be adiabatic. Based on the model of convective solid-state heat transfer [6], adiabatic temperature profiles were also assumed for the mantle and Fe–FeS core. The unknown parameter is the temperature at the mantle–asthenosphere boundary, $T_{A\text{-}M}^{Io}$, which is calculated within the interval 900–1250°C from conditions (8) and (9). The temperature profile in the rock-iron core of Io, except for the crust + asthenosphere layer, is approximated by the following polynomial:

$$\begin{aligned}
T_{\mathrm{Io}} &= T_{\mathrm{A–M}}^{\mathrm{Io}} + 220.8H_0 - 106.3H_0^2 + 36.0H_0^3, \\
H_0 &= (H - H_{\mathrm{A–M}})/(R_{\mathrm{Io}} - H_{\mathrm{A–M}}),
\end{aligned} \tag{11}$$

where H is the distance from the surface of the Fe–Si core, and $H \geq H_{A–M}$, where $H_{A–M}$ is the thickness of the solid crust and asthenosphere, i.e., the depth to the mantle–asthenosphere boundary.

Mathematical Model

Models for the internal structure of the satellites are described by the system of equations, including equations for the momentum of inertia and mass (2); the dependence of pressure on radius R and gravity acceleration g in a hydrostatic equilibrium approximation,

$$dP/dR = -\rho(R)g(R); \tag{12}$$

equation for the determination of gravity acceleration,

$$dg/dR = -4\pi G\rho(R) - 2g(R)/R, \tag{13}$$

where G is the gravitational constant; and equation of state for the determination of the density of material in the mantle (ρ_M) and rock-iron core (ρ_{Cor}),

$$\rho_{\mathrm{M}} = \rho_{\mathrm{M}}(P, T), \quad \rho_{\mathrm{Cor}} = \rho_{\mathrm{Cor}}(P, T). \tag{14}$$

The distribution of density is constrained to be free of inversions with depth, i.e., $d\rho/pH > 0$. Temperature is calculated from Eqs. (10) and (11).

The system of Eqs. (2) and (10)–(14) is solved under conditions (8) and (9) to determine the distribution of physical parameters in the Fe–Si cores, their radii, the radii of Fe–10% S cores, the thicknesses of water–ice shells and silicate crust for Europa and Ganymede, and the thickness of the upper silicate shell (solid crust + asthenosphere) of Io.

Calculation of Density in the Mantle of the Satellites

The following approximate method was used to calculate density as a function of composition, temperature, and pressure. Let the density at the parameters P_1 and T_1 be ρ_1. It is necessary to calculate density ρ_2 in the point P_2, T_2. The total differential of density as a function of depth (H), temperature (T), pressure (P), and phase composition (C_i) is expressed as

$$\frac{d\rho}{dH} = \left(\frac{\partial\rho}{\partial T}\right)_{C,P} dT + \left(\frac{\partial\rho}{\partial P}\right)_{C,T} dP + \sum_i \rho_i\left(\frac{\partial C_i}{\partial T}dT + \frac{\partial C_i}{\partial P}dP\right), \tag{15}$$

where $\rho_i = \rho(P, T, C_i)$. Taking into account that $\partial\rho/\partial T = \rho\alpha$ and $\partial\rho/\partial P = \rho(1/K_T)$, where α is the thermal expansion coefficient, and K_T is the isothermal bulk modulus, Eq. (15) can be rewritten as

$$\frac{d\rho}{\rho} = F(H)dH,$$
$$F(H) = \left(\alpha dT + \frac{1}{K_T}dP + \sum_i \frac{\rho_i}{\rho}\left(\frac{\partial C_i}{\partial T}dT + \frac{\partial C_i}{\partial P}dP\right)\right). \tag{16}$$

The integration of Eq. (16) over the depth interval from H_1 to H_2 gives the density ratio of phase associations in points *1* and *2*:

$$\ln\frac{\rho_1}{\rho_2} = \int_{H_1}^{H_2} F(H)dH = \phi(P, T). \tag{17}$$

The density ratio in Eq. (17) depends on the functions $P = P(H)$, $T = T(H)$, $\alpha = \alpha(P, T)$, $K_T = K_T(P, T)$, $C_i = C_i(P, T)$, and $\rho_i/\rho = \rho^0(P, T)$, where $i = 1, 2, \ldots, n$. The functions α and K_T are conservative and usually taken to be constant in the mantle, and a linear pressure dependence is assumed for K_T.

In this study, we proposed a method of density correction for pressure and temperature variations on the basis of the assumption that the function $\phi(P, T)$ is identical for compositions differing by 2–3 wt % in MgO, FeO, and SiO_2 concentrations, with all the other concentrations being constant. That is, the functions α, K_T, C_i, and ρ^0 are considered identical within a relatively small range of chemical composition. Using these assumptions, Eq. (17) can be simplified to

$$\rho_2 = \rho_1(\rho_2^*/\rho_1^*), \tag{18}$$

which allows the calculation of density ρ_2 in point (P_2, T_2) from the density determined in the reference point at $P_1 = 20$ kbar and $T_1 = 1000°C$. Subscripts 1 and 2 in Eq. (18) refer to points *1* and *2* or depths H_1 and H_2; the asterisk denotes the density calculated for a certain reference composition [composition (*)]. The density of composition (*) was calculated using approximations obtained from the THERMOSEISM program [22]. The density value ρ_1 is the desired parameter, which is determined using an iterative procedure. The densities in all other points of the mantle are calculated by Eq. (18).

The chemical composition (*) was determined by the method of successive approximations in the NaTiCFMAS system. A composition lying between the silicate fractions of ordinary and carbonaceous chondrites was taken for the first iteration, and this composition was subsequently refined using the calculated average density of the silicate fraction (crust ± asthenosphere + mantle) of the Fe–Si cores (ρ_{Sil}) reduced to P_1 and T_1. The density ρ_{Sil} was calculated from the average density of the crust for the icy satellites, the density of the crust + asthenosphere system for Io, and the average density of mantle material recalculated to P_1 and T_1. The computations showed that at least three iteration steps were required. In our case, the following composition of (*) was obtained at the last iteration step (wt %): $C_{SiO_2} = 47.2$, $C_{FeO} = 17.2$, $C_{MgO} = 29.4$, $C_{Al_2O_3} = 2.7$, $C_{CaO} = 2.2$, $C_{Na_2O} = 1.15$, and $C_{TiO_2} = 0.15$. The density of this composition (ρ*) was calculated by the polynomial

$$\rho_{Sil}^* \ (\text{g/cm}^3) = 3.521 + 0.004P - 0.00014T - 0.1766 \times 10^{-4}P^2 - 0.138 \times 10^{-7}T^2 + 0.58 \times 10^{-6}PT, \tag{19}$$

where T is in K, and P is in kbar.

A comparison with the initial data set, calculated by the THERMOSEISM program, showed that the maximum error associated with approximation by Eq. (19) is no higher than 1% and in most cases much smaller than 1%. According to Eq. (19), the density of the composition (*) at the last iteration at $P_1 = 20$ kbar and $T_1 = 1000°C$ is $\rho_{Sil}^* = 3.407$ g/cm^3.

Density variations in the Fe–10% S core were described by the equation

$$\rho(\text{Fe–10\%S core}) = 5.7[1 + (P - 50)/1250 - 0.00005(T - 1773)], \text{ g/cm}^3. \tag{20}$$

Equations (18)–(20) allow for the calculation of current values of density in the mantle and metallic core. Since the parameters of reference point *1* are taken to be equal for the three satellites, the average density of the rock-iron core can be readily calculated for the (P_1, T_1) conditions and compared with the average densities of the Fe–Si cores of other satellites.

Estimation of FeO and SiO_2 Concentrations in the Silicate Fraction of Satellite Material

Density is the only parameter that provides insight into the chemistry of the silicate fraction of the rock-iron cores. The dependence of FeO concentration on the mantle density was previously described by a linear function [21, 22]. Similarly, the dependence of the mass concentration of iron oxide (C_{FeO}) on the density of silicates (ρ_1) in the point (P_1, T_1) was approximated at each iteration step for the compositions (*) using the THERMOSEISM program. Together with the refinement of the composition (*), new linear dependencies were obtained for density. At the final iteration step, C_{FeO} was calculated as

$$C_{FeO} = 178.54(\rho_{Sil})/\rho_1^* - 161.34. \quad (21)$$

The concentration of SiO_2 was refined using the calculated C_{FeO} values:

$$C_{SiO_2} = 50.38 + 0.146C_{FeO} - 0.019C_{FeO}^2. \quad (22)$$

Equation (22) was obtained using the data of [40] for carbonaceous and ordinary chondrites.

Principles of the Calculation Procedure

The goal of the numerical simulation was to determine using conditions (8) and (9) the thickness of the water–ice shells, H_{Ice}; the crust thickness, H_{Cr}; the temperature at the upper boundary of the mantle, T_{Cr-M} and T_{A-M} (crust–mantle boundary for Europa and Ganymede and the asthenosphere–mantle boundary for Io); the density of mantle material; and the sizes of the metallic Fe–10% S cores.

The calculation procedure of the desired parameters can be divided into two stages. The first stage included the solution of the system of Eqs. (2) and (10)–(14) for the three satellites. During this stage, the distribution of density in the mantle and Fe–10% S core, the radius of the Fe–10% S core, and all the parameters from similarity conditions (8) and (9) were determined from the thickness of the water–ice shell (H_{Ice}^{E} and H_{Ice}^{G}), the thickness of the silicate crust (H_{Cr}) for Europa and Ganymede, the thickness of the crust + asthenosphere layer (H_{A-M}) for Io, and the temperature at the upper boundary of the mantle (T_{Cr-M}^{E}, T_{Cr-M}^{G}, and T_{A-M}^{Io}). Then, using the obtained values, the input parameters (H_{Ice}, H_{Cr}, T_{Cr-M}^{E}, T_{Cr-M}^{G}, and T_{A-M}^{Io}) were refined, and the procedure was repeated until the similar conditions were satisfied to the desired accuracy. In order to solve the system of Eqs. (2) and (10)–(14), we developed a rapidly converging iteration procedure accounting for the specific features of the simulated object.

The initial approximation for the thickness of ice shells was obtained from the following qualitative considerations. Since pressure and temperature in the Fe–Si core of Europa are significantly lower than in Io (Figs. 3a, 3b), it can be supposed that the lower temperature compensates to a great extent the effects of compressibility, i.e., the average density of the Fe–Si core of Europa is similar to that of Io. Then, the initial approximation for the thickness of the water–ice cover can be readily obtained from Fig. 1a, which allows an estimation of the moment of inertia of the rock-iron core of Europa.

The pressure profile of the rock-iron core of Ganymede lies mainly above that of Io, whereas its temperature is mainly significantly lower (Figs. 3a, 3b). Therefore, the density of the rock-iron core of Ganymede must be higher than the average density of Io. The thickness of the ice shell is determined from Fig. 1b. The initial approximations for the thicknesses of the crust of Europa and Ganymede and the crust + asthenosphere system of Io were taken to be 60 km.

RESULTS OF CALCULATIONS: THE INTERNAL STRUCTURE OF THE SATELLITES UNDER ISOCHEMICAL CONDITIONS

The main result of our study is the possibility of constructing satellite models satisfying the conditions of geochemical similarity of Io and the rock-iron cores of Europa and Ganymede [Eqs. (3) and (7)]. The distribution of P, T, ρ, and g, thicknesses of the water–ice shells and crust, and the geochemical constraints on the bulk compositions of the satellites were also determined (Figs. 3–5; Tables 1, 2).

Io. The thickness of the crust + asthenosphere system is 70 km. It is underlain by the silicate mantle and the Fe–10% S core with a radius of 737 km. The temperature varies from 1200°C at the asthenosphere–mantle boundary, 1304°C at the mantle–core boundary, and 1304–1350°C in the core (Fig. 3a). The pressure in the center is 80 kbar (Fig. 3b). The density of the isochemical mantle increases monotonously from 3.392 g/cm^3 at the upper boundary to 3.570 g/cm^3 at the core–mantle boundary (Fig. 3c). The density of the silicate fraction (ρ_{Sil}) is 3.386 g/cm^3 at P_1 = 20 kbar and T_1 = 1000°C; and the average density of the material of Io, $\rho_{Cor}(P_1, T_1)$ = 3.534 g/cm^3, is similar to the average astronomical density.

Europa. Europa has a water–ice shell with a thickness of 120 km (7 wt % H_2O), which is in agreement with our previous calculations [22, 24]. The thickness of the silicate crust is 50 km, and the radius of the Fe–10% S core is 576 km. The temperature of the upper shells of Europa's mantle is significantly lower than that of Io (Fig. 3a). The temperature at the crust–mantle boundary is T ~ 570°C, but it rises in deeper levels faster than in Io, and at the core–mantle boundary is almost as high as that in Io, 1293°C. Such a behavior of temperature is related to the different mechanisms of heat transfer. The density of the mantle is practically constant and ranges within 3.453–3.463 g/cm^3. The densities of the silicate fraction (crust + mantle) and the rock-iron core reduced to P_1 and T_1 are almost identical to the corresponding parameters of Io (Table 2).

Ganymede. Two models were considered for Ganymede. For the model without an ocean, the thickness of the ice shell is 900 km (47 wt % H_2O); in accor-

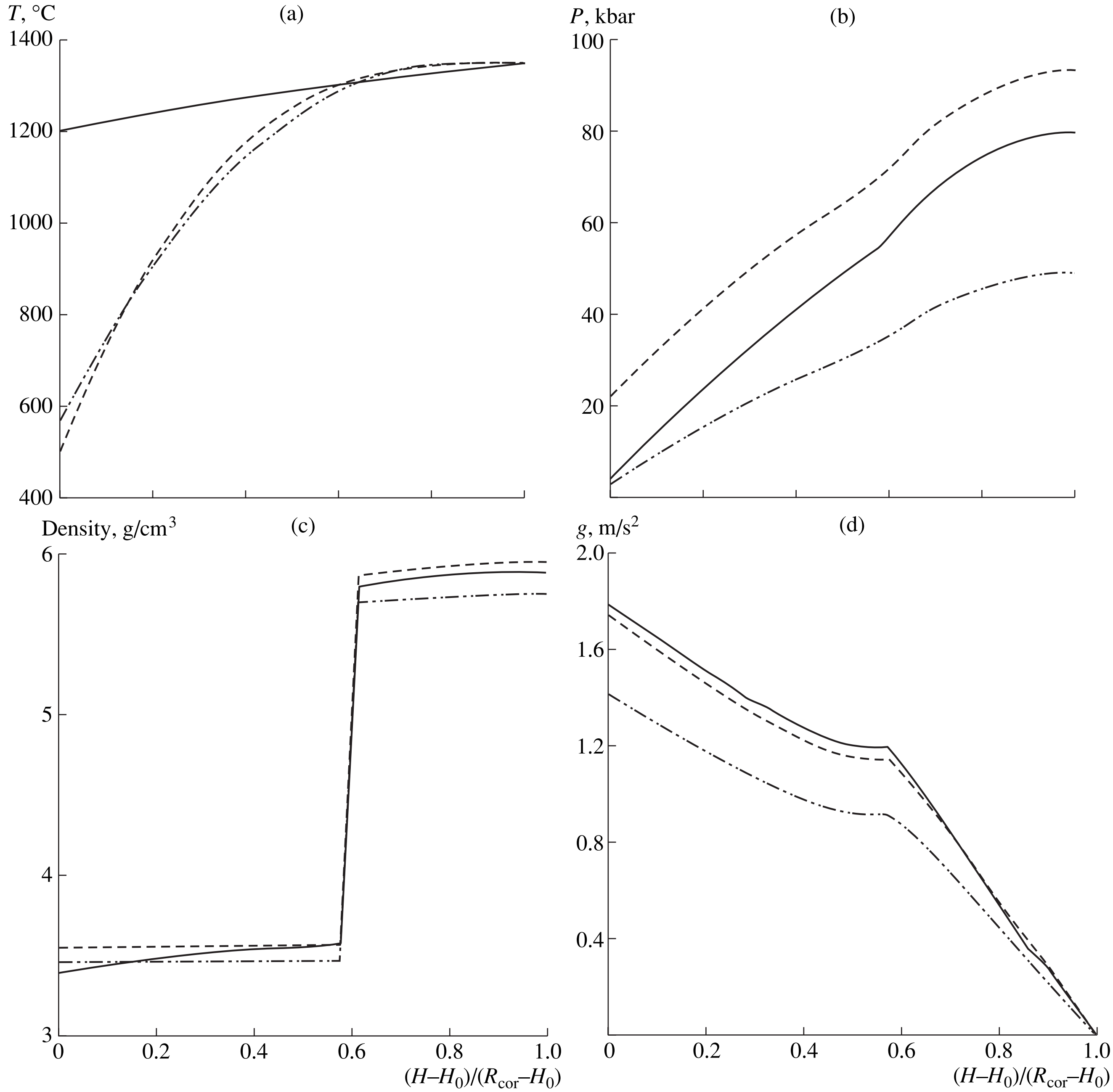

Fig. 3. Distribution of (a) temperature, (b) pressure, (c) density, and (d) gravity acceleration in the mantle and metallic Fe–10% S cores of Io (solid line), Europa (dash–dot line), and Ganymede (dashed line). H is the distance from the surface of the rock-iron core; H_0 is the thickness of the silicate crust of Europa and Ganymede and the thickness of the crust + asthenosphere layer for Io, and R_{Cor} is the radius of the rock-iron core.

dance with the phase diagram, it consists of ice I and layers of high-pressure ices. For the model with an ocean, the conditions of similarity are satisfied only if its thickness is no more than 40–50 km, and the thickness of the ice cover is 130–140 km, which is in agreement with the data of [51]. The total thickness of the water–ice shell of the satellite approaches 900 km, which is consistent with the results of [22].

The thickness of the silicate crust is 55 km. The radius of the Fe–10% S core is 695 km, which is 42 km smaller than the radius of Io's core. The radius of the rock-iron core (1734 km) is greater than that of Europa but is 87 km smaller than that of Io (Table 1). Thus, the Fe–Si core of Ganymede is almost a twin of Io and differs from the latter mainly in temperature distribution within the crust and mantle. The temperature profile of the rock-iron core of Ganymede is very similar to that of Europa but shows somewhat higher gradients in the upper mantle levels (Fig. 3a). Similar to Europa, the density is almost independent of depth (Fig. 3c). The

Table 2. Geochemical parameters of Jupiter's satellites calculated from the conditions of the similarity of the internal structures of their rock-iron cores compared with the materials of ordinary chondrites

Parameter	Io	Europa	Ganymede	Average	LL	L
H_{crust}, km	70	50	55	0.035 R_{Cor}	–	–
$(Fe_{tot}/Si)_{Cor}$	1.00	0.97	0.99	0.986	1.03 ± 0.04	1.18 ± 0.06
$(Fe/Si)_{Sil}$	0.534	0.531	0.539	0.534	0.714	0.607
$(FeO)_{Sil}$, wt %	16.1	16.04	16.28	16.14	19.66	17.20
Fe_m, wt %	9.82	9.22	9.43	9.49	6.33 ± 2.27	11.04 ± 1.46
Fe_m/Fe_{tot}	0.47	0.45	0.45	0.46	0.31+0.1	0.49+0.05
ρ_{Sil}, g/cm^3	3.386	3.389	3.385	3.387	3.431	3.396
ρ_{Cor}, g/cm^3	3.543	3.532	3.540	3.538	–	–

Note: H_{crust} is the thickness of the silicate crust for Europa and Ganymede and the thickness of the crust + asthenosphere layer for Io; $(Fe_{tot}/Si)_{Cor}$ is the total iron to silicon mass ratio in chondrites and the rock-iron cores of the satellites; $(FeO)_{Sil}$ and $(Fe/Si)_{Sil}$ are the concentrations of FeO and Fe/Si mass ratio in the silicate fractions of the Fe-Si cores of the satellites and chondrites; Fe_m is the mass percentage of metallic iron in the central Fe–10% S core relative to the total mass of the rock-iron core of the satellites and the abundance of metallic iron in chondrites calculated as $Fe_m = Fe_m^\circ + Fe_m$ of FeS [24]; (Fe_m/Fe_{tot}) is the mass fraction of metallic iron in chondrites [24] and the central Fe–10% S cores relative to the total amount of iron in the satellites; ρ_{Sil} and ρ_{Cor} are the densities of the silicate fraction of the rock-iron cores of the satellites and chondrites and the average density of the rock-iron cores recalculated to P = 20 kbar and T = 1000°C.

density is 3.541 g/cm^3 in the upper 340 km and increases monotonously at greater depths to 3.569 g/cm^3.

Structure of the Rock-Iron Cores of the Galilean Satellites

All the rock-iron cores of the satellites satisfy the similarity conditions of Eqs. (8) and (9) with an error not higher than 0.1% for the reduced average density of rock-iron cores, $\rho_{Cor}(P_1, T_1)$ = 3.538 g/cm^3, and 0.3% for the reduced density of the silicate fraction, $\rho_{Sil}(P_1$ = 20 kbar, T_1 = 1000°C) = 3.387 g/cm^3 (Table 2). The ratio of the radius of the Fe–10% S core to the radius of the rock-iron core is practically identical for Io, Ganymede, and Europa and averages R(Fe–10% S core)/R_{Cor} = 0.4. In contrast, the mass ratios of the Fe–10% S core to the rock-iron core, $M($Fe–10% S core)/M_{Cor}, of Io and Europa differ by 6% owing to differences in temperature and pressure (Table 1). The lowest M(Fe–10% S core)/M_{Cor} value was obtained for Europe (10.25%); the highest value, for Io (10.9%); and an intermediate value, for Ganymede (10.5%). The average M(Fe–10% S core)/M_{Cor} value of the three satellites is ~10.5%. The average thickness of the silicate crust of the icy satellites and the crust + asthenosphere system of Io is 0.032–0.038 of the radius of the Fe–Si core (R_{Cor}). The proportions of the sizes of the crust, mantle, and core show that the similarity conditions of Eqs. (3) and (7) are satisfied at almost exact geometric similarity of the internal structures of the rock-iron cores [Eq. (4)], which is the reason for the similarity of their moments of inertia.

Geochemical Characteristics and Bulk Composition

The geochemical characteristics of Io and the rock-iron cores of Europa and Ganymede are given in Table 2. The maximum difference from the average values (arithmetic mean) of geochemical parameters [Eq. (7)] was obtained for $(Fe_{tot}/Si)_{Cor}$ (about 1.5–2.0%). The discrepancies for C(FeO) and $(Fe_m/Fe_{tot})_{Cor}$ were not higher than 1%. It can be concluded that the misfit in the similarity conditions given by Eq. (7) is 1–2%. The following average geochemical parameters were obtained for the rock-iron cores of the satellites (Table 2):

$$\begin{aligned}(Fe_{tot}/Si)_{Cor} &= 0.99 \pm 0.02,\\ C(FeO)_{Sil} &= 16.15 \pm 0.15 \text{ wt \%},\\ (Fe/Si)_{Sil} &= 0.53 \pm 0.04,\\ (Fe_m/Fe_{tot})_{Cor} &= 0.46 \pm 0.01,\\ M_{Fe}/M_{Cor} &= 9.5 \pm 0.3 \text{ wt \%},\\ M_{Fe-10\%S}/M_{Cor} &= 10.55 \pm 0.3 \text{ wt \%}.\end{aligned} \tag{23}$$

$(Fe_{tot}/Si)_{Cor}$ is the total iron to silicon mass ratio in the rock-iron core; $C(FeO)_{Sil}$ and $(Fe/Si)_{Sil}$ are the concentration of FeO and the iron to silicon mass ratio in the silicate fraction of the Fe–Si core; $(Fe_m/Fe_{tot})_{Cor}$ is the mass ratio of the content of metallic iron in the central Fe–10% S core to the total content of iron; M_{Fe}/M_{Cor} is the mass ratio of Fe in the central Fe–10% S core to the whole Fe–Si core; and $M_{Fe-10\%S}/M_{Cor}$ is the mass ratio of the central core to the whole Fe–Si core.

DISCUSSION

The physical and chemical parameters of the models of the composition and internal structure of the Jovian satellites are compared with those of the Moon [39, 60] in Figs. 3–5 and Tables 1 and 2. Based on the isochemical models of Io, Europa, and Ganymede, the results of numerical simulation provided important geochemical constraints on the compositions of the rock-iron cores of the satellites consistent with geophysical similarity conditions [Eq. (3)]. It was shown that Io and the rock-iron cores of Europa and Ganymede are very similar, and show almost identical compositions and Fe_{tot}/Si ratios. Figure 4 illustrates the general trend of a decrease in the Fe_{tot}/Si ratio with increasing heliocentric distance from Mercury to the Jovian satellites. Figure 5 shows an increase in the abundance of ice with increasing distance from Jupiter: from 0 in water-free Io, 7% in Europe, 47% in Ganymede, to 49–55% in Callisto [22–25]. The higher content of H_2O in the composition of the outer satellites correlates with a decrease in the average density of the Galilean satellites.

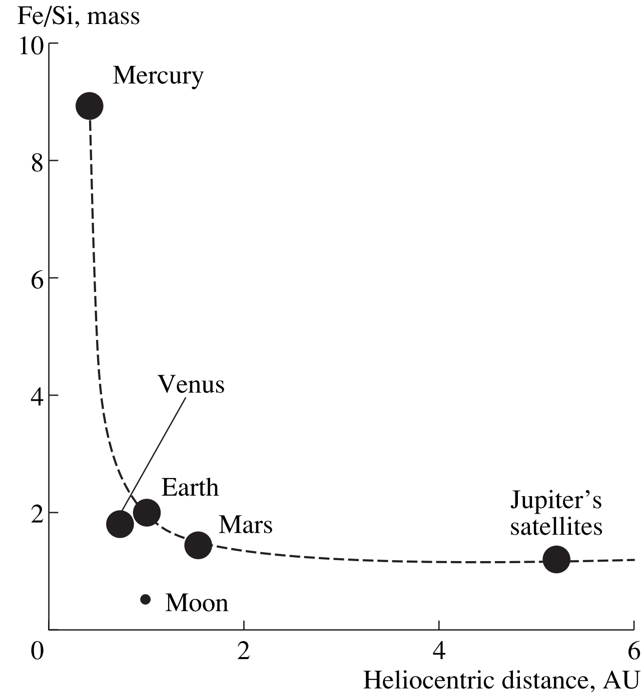

Fig. 4. Simplified dependency of the Fe_{tot}/Si ratio in planets and satellites on their heliocentric distance.

Based on a comparison of geochemical data on the compositions of terrestrial and lunar rocks and meteorites and the available geophysical constraints, it was concluded that the bulk composition of the Moon shows no genetic similarity to the material of either the Earth or chondrites [20, 39, 60]. As can be seen from Table 1, the composition of the Moon is also strongly different from the composition of the Fe–Si material of the Galilean satellites: the Fe_{tot}/Si ratio of the Moon is two times lower and the mass of the lunar Fe–10% S core is three times smaller than the corresponding parameters of the rock-iron cores of the Galilean satellites. The considerable depletion of iron in the composition of lunar rocks suggests that, in contrast to the Jovian satellites, the Moon was formed not from the accretion disk.

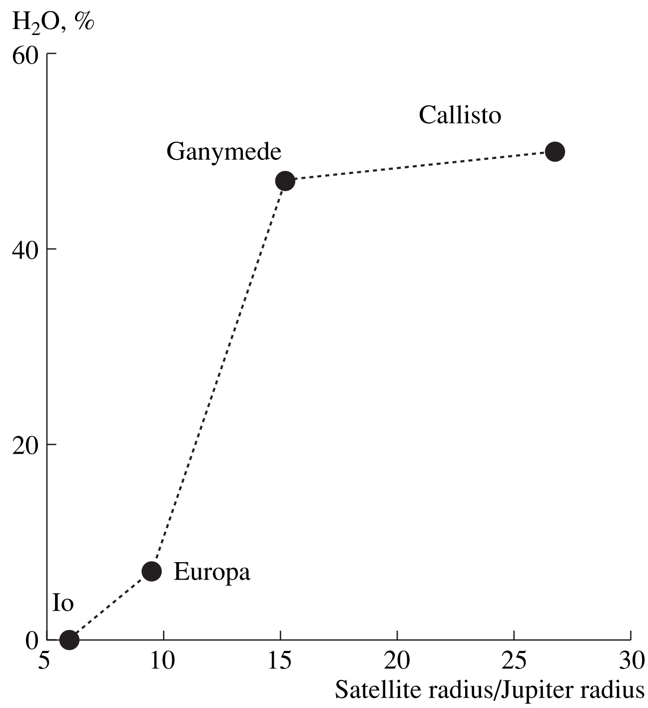

Fig. 5. Concentration of H_2O in the Galilean satellites versus the distance from Jupiter according to the calculations reported in this paper and [22–25].

The geochemical parameters of the bulk composition of the rock-iron constituents of Io, Europa, and Ganymede [Eq. (23)] estimated on the basis of condition (7) are similar to those of L and LL chondrites [40] (Table 2). The estimated uncertainties are 2% for $(Fe_{tot}/Si)_{Cor} = 0.99 \pm 0.02$ and 0.1% for ρ_{Cor}. The calculations of phase compositions for ordinary and carbonaceous chondrites are reported elsewhere [22, 24]. At 20 kbar and 1000°C, the phase assemblage of L/LL chondrites consists mainly of olivine (37–42 mol % *Ol*, $Fo_{72–75}$) and pyroxene (58–63 mol % *Opx* + *Cpx*), and has a density of 3.396 g/cm^3 (L chondrites) or 3.431 g/cm^3 (LL chondrites) (Table 2). The phase assemblage is obviously different from the pure olivine composition accepted a priori by Sohl et al. [29].

The satellites are closer to LL chondrites than to L chondrites with respect to only one parameter, Fe_{tot}/Si. This ratio is 0.97–1.0 for the satellites, 0.99–1.07 for LL chondrites, and 1.12–1.24 for L chondrites. It is interesting that the results of spectral and magnetic measurements of the surface layers of the asteroids 433 Eros and 243 Ida showed that the element ratios of Fe, Mg, Ca, and Al to Si and the ferromagnetic properties of the rocks are similar to those of ordinary chondrite samples [61–63]. On the other hand, the $(Fe_{tot}/Si)_{Cor}$ ratio of the satellites is considerably lower than that of H ordinary chondrites (1.6) and CI (1.73), CM (1.6), and CV (1.48) carbonaceous chondrites.

Sohl et al. [29] recently estimated the Fe_{tot}/Si ratio of the Galilean satellites. These authors obtained an unexpected increase in Fe_{tot}/Si with increasing distance from Jupiter: from ~0.7–1.6 for Io and Europa to ~2.0–5.0 for Ganymede. Leaving aside comments about these estimates and the method of their derivation, note that it is very difficult to find a reasonable explanation for the relatively high values of Fe_{tot}/Si and, consequently, high density of the rock-iron cores of the outer satellites formed far from Jupiter. Such a relation is not consistent with modern concepts on the processes of Fe/Si fractionation and formation of planetary bodies in the solar system [21, 22, 24, 35] (Fig. 4).

The element ratios and the masses of the Fe–10% S cores of the satellites (10.5 wt %) are in agreement with the bulk compositions and abundance of iron-sulfide phases in L chondrites (7.03 ± 0.95 wt % Fe and 5.76 ± 0.8 wt % FeS [40]). The abundance of metal in LL chondrites is significantly lower (2.44 ± 1.6% Fe [40]), although the content of FeS (5.79 ± 0.1%) is similar to those of L chondrites and the Fe–S cores of the satellites. In contrast, CI, CM, and CV carbonaceous chondrites are essentially free of metallic iron; they are more oxidized than ordinary chondrites and contain 4–7 wt % FeS and negligible amounts of Fe_m.

In the NaTiCFMAS system, the concentration of iron oxide in the silicate fraction of the satellites is $(FeO)_{Sil}$ = 16.1 wt %, which is somewhat lower than in L/LL chondrites (17–20 wt %). The density of the silicate fraction of the satellites (3.387 g/cm^3 at P = 20 kbar and T = 1000°C) differs by 0.4% from the density of the silicate fraction of L chondrites calculated for the same parameters (3.4 g/cm^3), which is related to the lower concentration of FeO in the mantle of the satellites. Taking into account the uncertainties in the problem formulation and numerical experiments, it can be concluded that the condition of identical bulk chemical compositions of the rock-iron cores is satisfied, if the material of the satellites is similar in composition to L/LL chondrites. These results supplement our previous estimates of the L/LL chondritic composition of Io, Europa, and Ganymede [20–22, 24]. In contrast to the previous studies, this paper reports the determination of the bulk composition of the satellites satisfying the condition of their isochemical character; moreover, we attempted to demonstrate that the equality of the chemical compositions of Io, Europa, and Ganymede must result from the moment of inertia and density estimates for the satellites.

The main source of error in the solution is the uncertainty in core composition and the real distribution of temperature in the interiors of the satellites. An increase in the abundance of iron in the central Fe–S core results in a smaller and less massive core and a lower (Fe_{tot}/Si) value of the satellite [22]. However, the maximum difference between the (Fe_{tot}/Si) values of Io's core for pure Fe and Fe–FeS eutectic compositions is less than 10%. The variant used in this study is intermediate in density and iron content. Therefore, the error in the $(Fe_{tot}/Si)_{Sat}$ estimate must be lower than 5%. The main uncertainty in the distribution of temperature is related to the choice of the mass transfer mechanism in the mantle, either conductive or convective. In this study, the choice of a heat transfer mechanism was based not only on a priori information, but also on the results of numerical experiments. For instance, the conductive mechanism in the mantles of the three satellites, including Io, did not satisfy similarity conditions (3) and (7). After a series of preliminary calculations, we selected the conductive mechanism for the mantle of Europa and Ganymede and the convective mechanism for the mantle of Io.

Perturbations in the magnetic field of the Galilean satellites were observed during the flybys of *Galileo* [47, 48, 64]. If a planetary body possesses a liquid conducting layer (aqueous electrolyte solution or a liquid core), electric currents induced in it by the strong electromagnetic field of Jupiter generate an imposed magnetic field. Europa contains about 7% H_2O. Ganymede has a 900-km-thick ice shell, and its ice–rock mass ratio is 47/53, which is in agreement with our previous calculations [22], but differs from the cosmic ratio (40 wt % of ice components) calculated from the solar composition assuming complete chemical equilibrium in the C–O–H system. This means that, despite the abundance of H_2O, Ganymede is enriched in the rock-iron component compared with the solar composition, which was previously noted by McKinnon [49]. Both Europa and Ganymede have probably thick water layers and relatively small cores.

It was previously concluded that the liquid phase is stable (does not freeze) below the ice crust of Callisto, and the thickness of the water layer was estimated as 120–180 km [23]. Albeit the surface temperature of the satellites is low (100–130 K), the ice crust could serve as a heat insulator providing the stability of the ocean in the past or in the present day. This problem is of fundamental importance for the geology of the icy satellites and is still a subject of heated debate [16, 23, 24, 27, 50, 51].

The source of the magnetic field is still unknown and can be related to convective motions either in the partially molten core or in seawater. Such a dichotomy can be preliminarily accepted both for the induced field of Europa and for the intrinsic field of Ganymede, although the dynamo mechanism can be supposed for the latter [65]. Water-free Io has no appreciable magnetic field [64], but it can be disguised by electromagnetic phenomena in the surrounding plasma torus. It is not clear if the absence of the field implies that Io's core is completely solid or completely liquid. According to calculations, the temperature at the mantle–core boundary is about 1300°C in all the satellites (Fig. 3a). The melting points of iron–sulfide alloys are much lower than those of silicates. The high-pressure phase diagram of the Fe–S system [43, 44] suggests that at least the outer part of the Fe–10% S core can be partially molten at temperatures of ≥1300°C.

The isochemical model of the three Galilean satellites suggests the isochemical character of the primary rock-iron material of the satellites. If this is the case, the *P–T* conditions of the accretion protodisk in the zones of satellite formation did not cause chemical differentiation of the iron–rock components along the disk radius. This means that the temperatures in the disk at the orbits of Io, Europa, and Ganymede were lower than the evaporation temperatures of metallic Fe and Fe–Mg silicates. In such a case, the Fe–Si material from which the satellites were formed must reflect the chemical composition of the solar disk in the Jupiter orbit.

The models of satellite formation from the accretion disk surrounding Jupiter assume that the mass of the disk is equal to the total mass of the Galilean satellites (39.2×10^{25} g). The minimum mass of the disk must provide the existing total mass of the rock component of the satellite material (without water) [34]. Without the ice component, the mass of the rock-iron material of the satellites ranges between 26.5×10^{25} and 27.1×10^{25} g depending on the amount of ice in Callisto (49–55% [25]). Hydrogen, helium, and ice are added to the masses of the satellites to achieve their correspondence to the solar chemical composition. Taking into account their relative content ($\sim 5 \times 10^{-3}$), the minimum mass of material required for the formation of the Galilean satellites is $\sim 6 \times 10^{28}$, i.e., ten Earth masses or 0.03 the present-day mass of Jupiter [34, 66].

As was noted in the introduction, the existing models of the Jovian protodisk can be classified into two groups, massive and low-mass [34–38]. The low-mass model [34] suggests that the total mass of material flowing through the disk during the stage of satellite formation was 6×10^{28} g. According to the massive model, the same value of 6×10^{28} g corresponds to the maximum instantaneous mass of the disk attained by the end of the stage of accretion of material on Jupiter and the disk from the solar nebula [66]. That is the two groups of models, being alternative in dynamic aspects, are strongly different in the surface density of the disk (see [34, 66] for more details).

The models of the massive disk are characterized by high density (pressure in the central plane of the disk ranges from 10^{-2} to several bars) and high temperature. Mosquera and Estrada [37] estimated the duration of Callisto and Ganymede formation as $\sim 10^6$ and 10^3–10^4 yr, respectively. Canup and Ward [38] argued that the formation of the Galilean satellites from the gas-rich Jovian subnebula with a mass of 0.02 M_J (hot massive model) poses considerable difficulties. This is related to the fact that rapid accretion must occur in a massive disk, and satellites must be formed within $\leq 10^3$ yr, which cannot be reconciled with the time required for the formation of undifferentiated Callisto (more than 10^5 yr).

In contrast, the formation of satellites in a low-mass gas-poor disk (gas-starved accretion disk after [38]) lasts $\geq 10^5$ yr, which is compatible with the structure of undifferentiated Callisto [25]. The low-mass models [34, 38] imply a density close to the background density of the solar disk (pressure in the central plane of the disk is lower than 10^{-4} bar) and low temperature (below 200–300 K). This means that the low-mass Jovian disk is relatively cold despite viscous heating, because its surface density and optical thickness are low. The low temperature provides the stability of ice in the orbit of Ganymede and hydrous minerals in the orbit of Europa. The material of carbonaceous chondrites is too poor in water to be the only source of Ganymede and Callisto consisting of ice and nonvolatile components mixed in the mass proportion ~50 : 50 [23–25]. Moreover, according to modern concepts, the material of carbonaceous chondrites is a product of secondary chemical interactions in the parent bodies between the anhydrous nebular material and liquid water fluid [67].

Canup and Ward [38] showed that the models of the low-mass disk are favorable. The same conclusion was reached by Makalkin and Ruskol [66] on the basis of an analysis of the time of gas dissipation from the protosatellite disk. The time of gas dissipation for the massive disk models is greater than the lifetime of the solar system, whereas the low-mass models give about 10^7 yr for gas dissipation in the orbits of the Galilean satellites. Thus, the recent theoretical studies support the possibility of the existence of the low-mass protodisk of Jupiter and, consequently, the possibility of the isochemical distribution of rock-iron components in the Galilean satellites. The reconstruction of the chemical compositions of Io, Europa, and Ganymede using geochemical and geophysical data allowed us to conclude that the satellites were formed from a material similar in composition to L/LL ordinary chondrites at relatively low temperatures, and the composition of the satellites corresponds in general to the composition of the solar system in the Jovian orbit during the formation of the Galilean satellites. A comparison of Fe_{tot}/Si ratios in the terrestrial planets and Jovian satellites (Fig. 4) suggests an occurrence of metal–silicate fractionation during early stages of solar system evolution [21, 23]. However, the mechanism of the fractionation is still unknown.

CONCLUSIONS

In this study, the models of the composition and structure of the Jovian satellites were constructed using the geophysical constraints obtained by the *Galileo* mission on the mass, average density, and moment of inertia of the satellites; geochemical data on the composition of meteorites; and thermodynamic data and equations of the state of water, high-pressure ices, and chondritic materials. The phase compositions and properties of satellite materials were modeled in the Na_2O–TiO_2–CaO–FeO–MgO–Al_2O_3–SiO_2–Fe–S system. The results of this analysis allowed us to compare the bulk chemical compositions of the rock-iron cores of the icy satellites (Europa and Ganymede) with water-free Io and chondritic materials; using these results, many important aspects of the compositions and structures of the satellites were elucidated. The main conclusions are the following.

(1) Using geophysical data on the mass, average density, and moment of inertia of Io, Europa, and Ganymede, it was shown that the internal structures of the satellites may be similar, and the bulk compositions of their rock-iron cores may be identical. The results of our study suggest that the three satellites are differentiated. Io consists of a 70-km-thick crust + asthenosphere shell, a solid silicate mantle, and a metallic Fe–10% S core. Europa and Ganymede are differentiated into water–ice shells, silicate crusts, (50–55 km thick), mantles, and iron–sulfide cores. The thicknesses of the water–ice shell of Europa and the ice shell of Ganymede were estimated as 120 and 900 km, respectively. The contents of H_2O in these satellites are 7 and 47%, respectively.

(2) The internal structures of Io, Europa, and Ganymede were numerically simulated under the condition of their isochemical nature. The most important geochemical and geophysical parameters of the composition and internal structure of the satellites were estimated. The distribution of temperature, pressure, density, and gravity acceleration was determined for the interiors of the satellites. The isochemical conditions for the rock-iron cores are satisfied at $(Fe_{tot}/Si)_{Cor} = 0.99 \pm 0.02$; $M_{Fe}/M_{Cor} = 9.5 \pm 0.3\%$; $M_{Fe-10\%\ S}/M_{Cor} = 10.5 \pm 0.3\%$, and $(Fe_m/Fe_{tot})_{Cor} = 0.46 \pm 0.01$. The silicate fraction of the satellites shows $(FeO)_{Sil} = 16.15 \pm 0.15$ wt % and $(Fe/Si)_{Sil} = 0.53 \pm 0.04$. The radii of the Fe–10% S cores are 737 km for Io, 695 km for Ganymede, and 576 km for Europa.

(3) A comparison of element ratios and geochemical parameters of the satellites with the corresponding characteristics of chondrites led to the conclusion that the bulk composition of Io and the compositions of the rock-iron cores of Europa and Ganymede are similar to the compositions of L/LL chondrites but very different from the geochemical characteristics of H ordinary chondrites and carbonaceous chondrites. The identical chemical compositions of three Galilean satellites implies that the primary rock-iron material of the satellites was isochemical and there was no radial Fe/Si fractionation under the P–T conditions of the accretion disk. Consequently, Io, Europa, and Ganymede were formed from a material chemically similar to L/LL ordinary chondrites at relatively low temperatures, below the temperature of evaporation of iron and silicates. In such a case, the rock-iron material from which the satellites were formed must correspond to the chemical composition of the solar disk in the orbit of Jupiter.

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