

Several considerations on the early history of the Earth

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1. Introduction

Since the classical works by Urey, Ruby and Vinogradov in the early fifties it is commonly accepted that formation of the Earth proceeded by accretion of solid bodies, that neither the planet as a whole nor its upper mantle had ever been completely melted, that the rise of the ocean and the atmosphere was due to degassing of the mantle, and that the Earth crust was made up of derivatives of partial melting of the mantle rocks and therefore enriched in incompatible elements and large ion lithophiles.

These principles are well established physico-chemically and appear to be in agreement with geological observations connected with the relatively late geological history of the Earth. However, it is difficult to reconstruct the events of the early Earth from the accepted theory. This paper devoted to Professor D Lal on his 60th birthday is an attempt to look at this problem from another angle.

2. Problem of the ancient Earth crust

One would think that the early formation of the hydrosphere and the atmosphere and hence the development of the sedimentation process must have given rise to early stabilization of lithospheric blocks. However, as is well known, there is no geological formation on the Earth whose age exceeds 3.8–3.9 billion years. One might suggest that absence of any traces of the ancient sialic lithosphere is due to its subsequent reworking. However, when the Sm/Nd technique was applied to the study of Precambrian rocks, it had been discovered that ϵ_{Nd}^t -values for Archean rocks fell on the line of the Nd-isotopic evolution of chondritic reservoir. This meant that granites, gneisses and gabbros which composed the oldest nuclei of the Earth crust, derived directly from the mantle, and not from any differentiated precursor. Since petrological considerations argue against significant generation of acid magmas from a substance of mantle composition, a model was proposed that preliminary differentiation in fact existed. But the lifetime of the precursors was too short to show any appreciable change in their Nd-isotopic composition, so that the whole cycle including magmatism, weathering, sedimentation and remelting would not have exceeded 50 million years (e.g. De Paolo 1981). In terms of Sm/Nd systematics, this is equivalent to

immediate origin from the mantle. This idea was related to the concept of intensive mantle convection in the early Earth.

However as more data appeared, it became evident that ϵ_{Nd}^i -values for a majority of the Archean rocks deviated slightly from the line of chondritic evolution. Compilation of data shows that ϵ_{Nd}^i -values range between +0.5 and +3. Positive ϵ_{Nd}^i -values characterize rocks which are derived from a source previously depleted in light rare earths, in particular Nd relative to Sm ($f_{\text{Sm}/\text{Nd}} > 0$). It is difficult to reconcile this fact with the accepted model, as the depletion of the mantle has been suggested to be a consequence of the chemical differentiation which occurred due to the extraction of the crust. The positive ϵ_{Nd}^i values indicate that mantle was depleted even before formation of the earth's crust.

Geochemistry of the Early Archean in many aspects is unique. The Archean granites (tonalites and trondhjemites) are different in composition from the later ones. They are impoverished in K and enriched in Na. Sharp increase of K content in the crust occurred before 2.6–2.9 billion years. This boundary is also marked by a strong change of $^{87}\text{Sr}/^{86}\text{Sr}$ in the oceans. This suggests that before this time different mechanisms and sources were involved in the formation of the earth's crust.

3. Suggested model of formation of planets and meteorites

As mentioned earlier the classical view relates formation of the earth's crust to partial melting processes leading to the enrichment of incompatible elements in the liquid phase and eventually in the crust. However does a process exist which would produce crustal material depleted rather than enriched with incompatible elements? Yes, it does. Such a process apparently occurred on the moon. Many investigators agree that the upper mantle of the early moon was completely melted. As it cooled and crystallization began, heavy oxides containing Fe, Ti etc. and minerals of relatively high density like olivine formed the lower peridotitic layer while lighter minerals like plagioclase floated to the surface leading to the formation of the gabbro-anorthositic crust.

As this plagioclase crust is a cumulate and not a product of partial melting, it has to be depleted in incompatible elements. The same process could have occurred during formation of the Earth. However the belief exists that the Earth and other inner planets of the solar system have grown by a relatively low temperature accretion of solid bodies. And the "hot" model is not consistent with the presence of hydrosphere and other volatiles in the Earth.

The theory of collisional evolution of solid bodies developed by Safronov (1969) is still most popular. There exists an alternative model suggested by Gurevich and Lebedinskij (1950), the computer simulation of which was later developed by Kozlov and Eneyev (1977). This model envisages accumulation of a diffuse body by amalgamation of wisps of gas and dust. Proceeding from this model one can suggest that planets grew by compaction of diffuse disk-like protoplanetary bodies consisting of small particles.

What was the chemical composition of those particles? The protoplanetary material was earlier considered a product of condensation of high temperature gas of solar composition. However the discovery of the isotope anomalies in meteorites shows that grains of interstellar matter in the initial protosolar cloud were not

completely evaporated during the formation of the Sun. They survived and might serve as initial material during creation of the planets. Therefore it is possible that the chemical composition of the initial particles was close to that of interstellar dust or roughly to the chemical composition of the most primitive carbonaceous chondrites.

Heat radiated by the condensed central body, due to adiabatic increase of temperature during compaction, must have led to melting of the particles in the surroundings. Volatiles evolved from the melted particles were swept away by the solar wind and degassed droplets (chondrules) fell to the surface of the growing planet. A gradient of chemical and physical properties was established along the profile of the compacting diffuse body from the melted and partly evaporated silicates in the central body through the shell consisting of melted and exhaustively degassed droplets in the inner part and less altered material on the periphery. I think that the different types of meteorites represent different parts of the former diffuse body which on compaction was catastrophically disrupted at a relatively early stage. I believe that it was an unrealized planet which presumably grew between Mars and Jupiter and was destroyed by the tidal forces from the side of Jupiter. The abbreviation UP (unrealized planet) is used hereafter to refer to this object.

After destruction of the UP, its material was scattered over the solar system. Part of it remained in the orbit and gave rise to a variety of asteroids and meteorites. The former droplets provided the main source of chondrites whose eventual composition (from H to L type of ordinary chondrites or from C₃ to C₁ type of carbonaceous chondrites) was determined by the position of the starting material within the original diffuse body. Achondrites represent silicate material of the central condensed body and iron meteorites—its metallic core.

The suggested model implies that the metallic core of planets formed immediately after the accretion process. It has been inferred from geophysical data that the core of the Earth contains, apart from Fe and Ni, some light elements like H, C, O, Si and S. But our model rules out H, C and S in the core as these elements were lost during accretion. The same applies to Si as its incorporation in the core requires the presence of reducing agents like H or C. The model is consistent with the presence of FeO as an additional component in the core. This idea was suggested by Dubrovskij and Pan'kov (1972). It is essential that solubility of FeO in molten iron depends on pressure. It amounts to 4 mol% under 100 kbar, 11 mol% under 300 kbar and 52 mol% under 600 kbar (Ringwood 1979). In this connection it is noteworthy that the Moon, where even in the central part the pressure does not exceed 50 kbar, contains only a small core (~2% of the whole mass of the Moon). Apparently it consists only of metallic Fe and Ni. This explains the exceptionally low concentration of siderophiles and at the same time relatively high concentration of FeO (15–25%) in the lunar mantle. One can suggest that in general at the early stage of accretion only relatively small metallic Fe-Ni core can form. As the growing planet exceeds a certain size further increase of core becomes possible by dissolution of FeO in metallic iron. It is of interest to note that the inner part of the Earth core, which is believed to consist of metallic Fe + Ni, is also about 2% of the mass of the planet. Massive cores could be expected only in relatively large bodies. Indeed the Earth and the Venus have cores which are approximately 30% of their total mass. Mars has a much smaller core (~20% of the mass of the planet). The exception is Mercury, in which the core constitutes 70% of its mass. However because of its proximity to the Sun it must have lost not only the volatiles but also a significant part of silicates.

Probably the Moon acquired only inner part of its initial nebula whereas the external part was captured by the Earth with which the Moon formed a binary system. Apparently the external part of the proto-Martian diffuse shell was also disturbed and partly lost during the catastrophic event which occurred when a protoplanetary body developed between Mars and Jupiter.

The fact that iron meteorites do not contain FeO shows that during the catastrophic destruction the inner condensed part of UP was approximately of lunar size.

It is well known that planets and meteorites are depleted in volatiles relative to solar abundances of the corresponding elements. The degree of depletion increases with increase of volatility of the elements: Si, Cr, Mn, Na, K, As, Ga, Rb, Tl, Cs, In, Ag, Zn etc (Ringwood 1979). This has been interpreted as a consequence of condensation of protoplanetary material from an initial high temperature gas and served as a cornerstone of the theory of heterogeneous accretion. However it may also be due to evaporation and loss of volatiles by melted droplets during accretion. The logic of the suggested model implies that the Moon should be depleted in volatiles to a greater extent than the Earth. In turn the Earth should be depleted relative to the ordinary chondrites and the latter relative to the carbonaceous chondrites.

The degree of depletion can be estimated by comparing elements which behave similarly in geochemical processes but show different volatility. For instance the K/U ratio is known to change only slightly in different terrestrial rocks. However, as one can see from table 1, this ratio is completely different for the planetary bodies with different accretion history. Depletion of K in the Earth relative to the ordinary chondrites is characterized by factor 0.15–0.18. For the Moon, this factor is 0.02–0.03.

It is noteworthy that the Moon and the achondrites have almost similar K/U ratios. Since achondrites in our model represent the silicate phase of the inner condensed body of UP, it is clear that this body was approximately of lunar size before destruction.

The Moon is depleted in Rb even to a greater extent than in K. Rb is heavier but a more volatile element. Consequently the difference in depletion is actually controlled by the difference in volatility. This is consistent with the idea that loss of volatiles occurred during compaction from the surface of small melted particles but not from the surface of a planetary-sized body.

The Moon is depleted not only in K and Rb but also in Na. Consequently, coming back to the problem of the origin of the earth crust, we can conclude that the uppermost layer of the primitive Earth, even if it was formed in the same way as the lunar gabbro-anorthositic crust, could not be chemically identical to the latter. The earth's mantle contains 8–10 times more K and 5–6 times Na than the lunar mantle. The protocrust of the Earth has to be enriched in K_2O (~ 0.8%) and Na_2O (~ 3.0%).

Table 1. K/U and K/La ratios for the Earth, Moon and meteorites.

Ratio	Earth	Moon	Ordinary chondrites	Achondrites (eucrites)
$K/U \times 10^{-3}$	11.2	1.6	70.8	2.7
$K/La \times 10^{-5}$	4.5	0.7	24.2	1.2

In order to evaluate the chemical composition of the protocrust of the Earth one can modify the composition of the anorthositic lunar crust by a corresponding addition of potash feldspar and albite components and subtraction of an equivalent portion of anorthite component. Obviously it will be more acidic than the lunar crust. It would correspond to a pyroxene-feldspathic composition, with the normative plagioclase being close to 50, i.e. to andesine-labradorite.

The protocrust of this composition fits well with the Archaean tonalites and pyroxene granulites in a certain proportion. It could be a source of tonalitic magma as pyroxene and plagioclase are on the liquidus of tonalitic composition below 10 kbar. A high degree of melting along with the initial relatively low content of K in the protocrust explains the specific low K and high Na composition of the Archaean granites and gneisses. Besides, enrichment of the Archaean rocks in Sr and Ba, excess of Eu, positive ϵ_{Nd}^t -values and some other features can be explained.

4. Beginning of the ocean formation

As mentioned earlier ordinary chondrites are depleted in volatiles to a lesser extent than the Earth. Therefore even if ordinary chondrites contain $< 0.02\%$ of carbon and negligible amount of water, the Earth must have been depleted in carbon and water almost completely. It follows from the model that the surface of the Earth by the end of its accretion was low in water and other volatiles and could not gain them through degassing of the mantle. The question then arises when and how water comes to the surface of the Earth? More than 20 years ago, I suggested that the earth's crust received its volatiles from carbonaceous-chondrite-like-material (Galimov 1968). I based my theory upon the unique similarity in isotope composition between the volatile elements of the earth's crust and carbonaceous chondrites. Different versions of this idea have been proposed in the literature from time to time.

The origin of the hydrosphere of Earth can be investigated by oxygen isotope studies. Water plays an important role in the oxygen isotope balance and the oxygen isotope distribution between geospheres. In table 2, the oxygen isotope inventory is presented. It illustrates the well-known fact that the earth's crust-ocean system is enriched in ^{18}O relative to the upper mantle. The present calculation gives $\delta^{18}\text{O} = +8.8\%$.

If the model of the Earth suggested above had a primitive plagioclase crust, the difference between the isotope composition of the crust and the mantle could be partly due to isotope fractionation during crystallization. Plagioclases are enriched in ^{18}O relative to olivine. For andesine the isotope effect is about 1.5% below 1000°C . However it could give utmost $\delta^{18}\text{O} \approx +7.5\%$ for the protocrust, which is not sufficient to explain the observed enrichment of the crust in ^{18}O . The earth crust is composed of primary magmatic rocks and secondary rocks through the sedimentary cycle. It is obvious that there are basic differences between the oxygen isotope composition of the secondary and the primary rocks. The sedimentary and metamorphic rocks have an average oxygen isotopic composition of $+13.8\%$. Part of the granites is also secondary. Even if we take half of all granites to be secondary, the $\delta^{18}\text{O}$ -value for secondary rocks as a whole would be no less than $+12.8\%$. On the other hand, for primary rocks in total, $\delta^{18}\text{O}$ cannot be higher than $+7.0\%$. The oceanic water has $\delta^{18}\text{O} = 0\%$, and that of the juvenile water is about $+6\%$, since

Table 2. ^{18}O -budget in the Earth crust.

	Mass $\times 10^{24}$ g*	O_2 -content $\times 10^{22}$ mol	$\delta^{18}\text{O}\%$
Sedimentary rocks			
Sand and sandstones	0.43	0.69	+ 15
Clays and shales	1.14	0.81	+ 15
Carbonates	0.71	1.10	+ 25
Evaporites	0.02	0.03	—
Metamorphic rocks			
Metasandstones	0.47	0.75	+ 12
Paragneisses and shales	4.74	7.55	+ 12
Metamorph. carbonates	0.18	0.28	+ 15
Iron formation rocks	0.06	0.09	+ 6
Magmatic rocks			
Granites	5.68	8.69	+ 9
Alkalines	0.01	0.01	+ 7
Basalts	15.0	22.0	+ 6
Ultrabasic	0.02	0.03	+ 5
Earth crust:	28.46	43.06	+ 8.8
Ocean:	1.4	3.89	0
Total:		46.95	+ 8.1
Upper mantle rocks			+ 6.0

* Taken from Ronov (1980)

water in contact with molten silicates acquires isotope composition very close to the oxygen isotope composition of minerals.

Hence one can obtain an estimate of the minimum water flow through the mantle-crust boundary, ΔM , by considering the isotope balance as

$$M_{\text{SR}}\delta^{18}\text{O}_{\text{SR}} + (M_{\text{OW}} + \Delta M)\delta^{18}\text{O}_{\text{OW}} = M_{\text{PR}}\delta^{18}\text{O}_{\text{PR}} + (M_{\text{JW}} + \Delta M)\delta^{18}\text{O}_{\text{JW}}$$

where M is the mass of the corresponding oxygen reservoir (SR, secondary rocks; OW, oceanic water; PR, primary rocks; JW, juvenile or primary water). The calculation gives ΔM equal to 11.9×10^{22} moles. This means that the amount of water passing through the mantle-crust boundary during geological history exceeds three times the present volume of the world oceans. This value gives an idea of the scale of lithospheric circulation of water and in particular about the scale of hydrothermal deposits related to this cycle.

Thus owing to the global circulation of water ^{18}O is transferred from the mantle to the ocean. On the contrary, as isotope fractionation under low temperature results in enrichment of ^{18}O in sedimentary minerals (e.g. + 28‰ in calcite and + 34‰ in silica relative to water under 25°C), sedimentary process takes out ^{18}O from the ocean. The present isotope composition of the ocean is under the control of these two competing processes. The oxygen isotopic composition of sedimentary rocks reflects the isotopic composition of contemporary oceanic water. Perry (1967) was the first to draw attention to the enigmatic trend of oxygen isotopic composition of cherts and carbonates with geological time. Figure 1 illustrates such a trend for carbonates. The data are taken from different literature sources and supplemented by measurements

made recently in our laboratory. The latter, labelled with asterisk cover a 600 million year interval of the Late Proterozoic which was poorly covered by $\delta^{18}\text{O}$ data before. One can recognize some details of the background of the general trend. The most pronounced shifts towards positive $\delta^{18}\text{O}$ values occur during periods of tectonic activities, Caledonian, Hercynian and Alpinian. Apparently during these periods the global circulation of water increased. On the contrary during periods of steady sedimentation the trend weakened or was even reversed. The ^{18}O -trend indicates unsteady state of oxygen isotope distribution between the ocean and the earth's crust which might be caused by initial depletion of the Archaean ocean in ^{18}O as suggested by Perry (1967). Some investigators, however, interpret the observed change in oxygen isotope composition of carbonates and cherts as a secondary effect which is not related to change of the isotope composition of the ocean. The matter needs further study. However the suggested depletion of the Archaean ocean in ^{18}O is most likely. I believe that this phenomenon is due to change with geological time of the relation

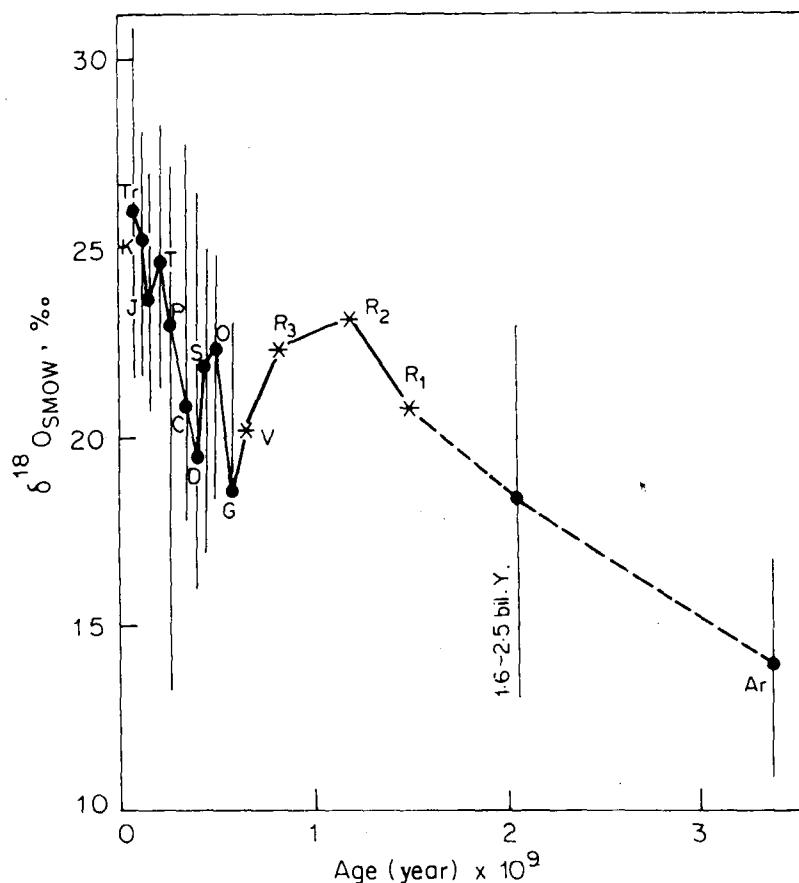


Figure 1. Variations of oxygen isotope composition of carbonates during geological time. The horizontal lines are ranges of variation of $\delta^{18}\text{O}$ (relative to SMOW) of carbonates of a given age. The dots are the averaged values. The points marked by an asterisk are the data obtained recently in our laboratory for representative samples of the Late Proterozoic (V, R₁-R₃) carbonates of the Russian platform prepared by G A Kazakov. The data for the Phanerozoic rocks are compilation from different literature sources. Archaean and the Early Proterozoic data have been taken from Veizer and Hoefs (1976).

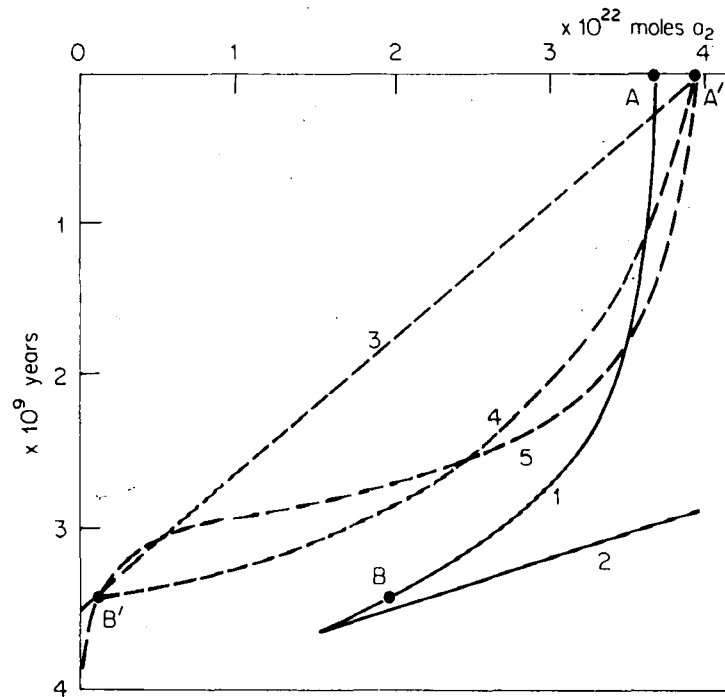


Figure 2. Suggested curves of evolution of mass of oxygen bound in the sedimentary shell (solid lines) and the ocean (dashed lines); 1. Accumulated mass of the sediments; 2. Deposited mass of the sediments; 3. Linear type of evolution of the ocean; 4. Exhaustive type of growth; 5. Exponential type of growth followed by exhaustion of the source. A and A' are masses of oxygen reservoirs related to the present sedimentary shell and the ocean correspondingly, B and B' are the same at 3,5 b.y. ago as suggested by the model.

between mass of the ocean (in terms of oxygen content) and mass of the sedimentary shell. Because of significant isotope fractionation between water and sediments (isotope separation coefficient $\alpha_{\text{CaCO}_3 - \text{H}_2\text{O}} = 1.028$ under 25°C), $\delta^{18}\text{O}$ of water depends strongly on water-mineral balance. For example, in a water-carbonate system with total $\delta^{18}\text{O} = +10\text{‰}$ if water-to-mineral ratio changes from 1:10 to 10:1 (in terms of oxygen mass), $\delta^{18}\text{O}$ of the water changes from -15.5‰ to $+7.5\text{‰}$ (at 25°C).

In order to evaluate oxygen isotope distribution between sedimentary shell and ocean one needs to know how their masses evolved during geological history. This is an unsolved problem and only a very approximate evaluation is attempted in the following. The growth of the sedimentary shell is determined by the rate of accumulation of the sedimentary rocks and the rate of their erosion and metamorphism. Analysis of data (Ronov 1980) on quantitative distribution of sediments in different intervals of geological time leads me to the conclusion that the decrease in amount of sediments of a given age due to erosion and metamorphism obeys the following relationship: $M_t = M_0 \exp(-1.35 \times 10^{-3}t)$, where M_0 is the initial mass, t is time (in 10^6 years) elapsed since sedimentation and M_t is the preserved mass at t . The rate of accumulation of sediments, averaged over large intervals of time, appears to be more or less constant, given by 3×10^{23} g/100 million years. The

integration then gives a generalized curve of growth of sedimentary shell as shown in figure 2. The present mass of sedimentary shell is 2.3×10^{24} g (3.63×10^{22} mol of oxygen) although over 4 billion years the sedimentation cycle produced about 12×10^{24} g of sediments. More than half of this amount was subjected to metamorphism and granitization and the rest recycled.

Since the ocean-crust system is an open system in terms of the oxygen balance, the rate of accumulation of the ocean cannot be inferred directly from the curve of growth of sedimentary shell and the known values of oxygen isotope distribution. However at the beginning of the process, a closed system model is an acceptable assumption. Calculation on this basis gives a value of about 0.05–0.08 for the ocean-sedimentary shell ratio 3.5 billion years ago. This means that the mass of the Early Archaean ocean was two orders smaller than its present value. As seen from figure 2 any conceivable curve of evolution of the mass of the ocean must begin from 3.9–4.1 billion years ago.

Thus one comes to the conclusion from the above considerations that during the first 500 million years, hydrosphere on the Earth was not significant, even if it existed at all.

5. Beginning of geological processes

After the accretion and formation of the core accompanied by evolution of great amounts of heat, the Earth became cool owing to convection and thermal radiation from the surface. Apparently during the growth of the Earth only the upper hundred kilometres of its mantle could be in the molten state. The melted layer concentrated incompatible elements. Therefore the upper mantle of the earth is enriched in such elements relative to ordinary chondrites. Solidification of the upper mantle proceeded simultaneously from below and above. The surficial plagioclases led to the formation of a primitive sialic protocrust—analogue to the lunar anorthositic crust but much more enriched in albitic component.

The residual liquid, enriched in volatiles, incompatible elements and large ion lithophiles (LIL) like K, REE, U, Th, P, Rb etc, were concentrated at a certain depth, possibly 100–200 km. This sort of matter has been recognized as a source of KREEP-material on the Moon. With the Earth enriched in volatiles relative to the Moon, the reservoir of KREEP-like material must have played a much more significant role.

The surface of the Earth was almost free of water and other volatiles with the exception of some noble gases. The landscape must have been similar to that of the Moon at present. Appearance of water completely changed the situation. As is well known, water drastically decreases the melting point of silicates. For example, dry andesite is stable up to 1100°C. However in the presence of water andesite is melted at 650–700°C and its solidus intersects the Archaean geotherm at a depth 15–20 km. Magmatic differentiation of the primitive pyroxene-plagioclase protocrust must have given rise to formation of tonalites, grey gneisses, gabbro and granulites. Increased temperature which was characteristic of the Early Archaean upper mantle led to the high degree of partial melting and formation of komatiitic lavas.

Weathering of igneous rocks by water and chemically active gases in the atmosphere supplied sedimentary material to the newly formed water basins. Such basins were obviously limited in volume and isolated from each other. The mass of sediments accumulated in such basins could exceed many-fold the mass of water

contained in them. Accumulation of sediments alternating with effusive rocks resulted in formation of thick sedimentary-effusive complexes which now are known greenstone belts. At this early stage of geological history thermal conditions in the Earth were determined by heat evolved during accretion. The processes described occurred predominantly within the lithosphere possibly down to a depth of 80–150 km. Underlain layers including that enriched in KREEP-material were not touched at this stage. This layer was involved in active magmatism later when new increase of temperature occurred probably due to decay of radioactive elements. I believe that dramatic increase of the K content in rocks of the earth's crust from 2.5–2.9 billion years ago was related to volcanism which was supplied from this source.

6. Origin of the biosphere

Carbonaceous chondrites contain a great variety of organic molecules including amino acids, porphyrines, nucleic acids etc. These molecules being highly concentrated in incipient water basins must have given rise to fast prebiological evolution.

In an attempt to trace the earliest signs of life our attention was attracted by banded iron formations (BIF), firstly because rocks of this formation occur among the oldest rocks (3.8 billion years in Isua, West Greenland) and secondly because sedimentation of iron deposits calls for a source of molecular oxygen. Cloud (1976) was the first to put forward the idea that formation of the giant masses of iron ores in the Precambrian was due to release of oxygen by photosynthetic organisms. This idea has been disputed by others since molecular oxygen could be produced not necessarily by organisms but simply through photolysis of water. However, carbon isotope data support the idea that BIF is actually a product of the biosphere. Graphite is widely present in BIF. Its carbon isotope composition varies from $\delta^{13}\text{C} = -20\text{‰}$ to -28‰ . These values are typical of photosynthetically-produced organic carbon and indicate the biological source of graphites. BIF terranes often include layers of siderites. Figure 3 presents the reactions resulting in the deposition of oxidized iron. According to this scheme the carbon of siderite is derived both from bicarbonate and biogenic source. Carbon isotope composition of siderites varies mainly from $\delta^{13}\text{C} = -6\text{‰}$ to -13‰ . These are intermediate between $\delta^{13}\text{C}$ of biogenic and bicarbonate carbon.

Withdrawal of significant amounts of biogenic carbon from biospheric cycle during formation of iron deposits affects the carbon isotope balance and is eventually reflected in the isotope composition of contemporary marine carbonates. Depletion of the Precambrian carbonates in ^{12}C has been actually recorded. A pronounced depletion of carbonates in ^{12}C ($\delta^{13}\text{C}$ from $+8$ to $+13\text{‰}$) appears to have taken place during the Early Proterozoic (2.0–2.6 billion years ago) when 90% of the Precambrian iron ores was deposited.

Occurrences of phosphorites and apatites related to BIF are often reported. Phosphorus plays a crucial role in the biosphere. Availability of phosphorus determines the mass of biosphere. Hence the relationship between development of biosphere and volcanism. It is of interest to note that explosive proliferation of life manifesting in exceptional development of BIF occurred just after activation of the layer enriched in KREEP-material. The correlation between occurrence of BIF and

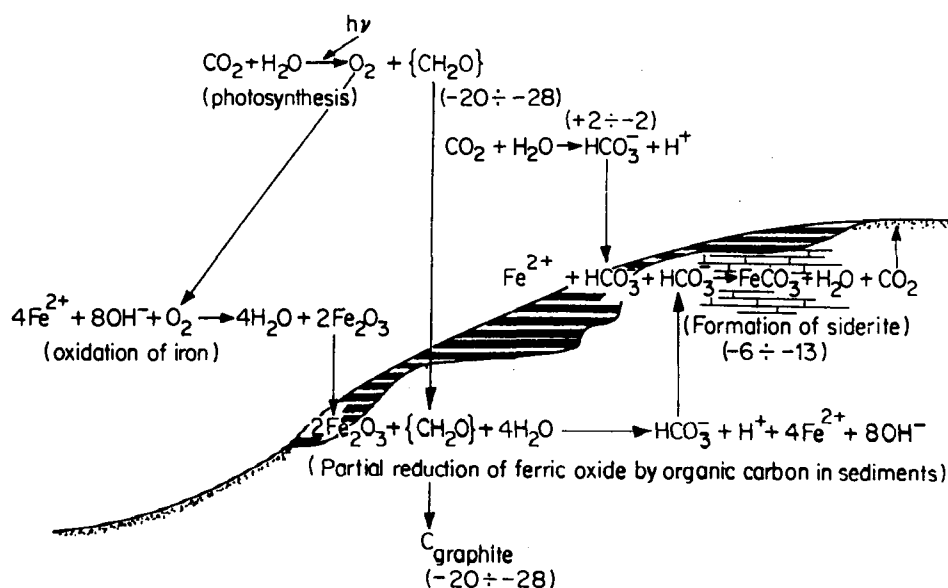


Figure 3. Isotopic evidences of involvement of biological carbon $\{CH_2O\}$ in the process of deposition of banded iron formation. Approximate carbon isotope composition ($\delta^{13}C$ in ‰, relative to PDB) of a compound is pointed out in brackets.

concentration of phosphorus is characteristic of deposits as old as 3.4–3.5 billion years.

Thus it appears that BIF is actually a phenomenon of the biosphere. Is it possible to extrapolate this conclusion to the Isua case? If it is, it would mean that photosynthetic organisms already existed 3.8–3.9 billion years ago. On the other hand as was just suggested, water and carbonaceous matter appeared on the surface of the Earth about 4.2–3.9 billion years ago. Then the time interval between the beginnings of the prebiological and biological evolution right up to photosynthesis is too short.

In this connection I suggested that photosynthesis, more precisely photochemical release of oxygen on the basis of organic compounds, appeared before life. In other words photosynthesis was suggested to be a prerequisite for the origin of life rather than its consequence. It seems logical that development of energy consuming and low entropy biological systems was preceded by the appearance of an effective and universal mechanism of conversion of external energy (light) into chemical energy. Besides experiments are known which demonstrate the release of oxygen by a simple system containing chlorophyll molecules and lipids. Porphyrins and bipolar compounds of lipid character have been found in carbonaceous chondrites and could be easily synthesized abiotically. Chlorophyll molecules on a lipid membrane are able to produce a photoeffect. The next step was the oxygen-release by the splitting of water molecules in the presence of an acceptor of hydrogen. Further development of this energy cell could result in realization of photoinduced chemical potential in the form of phosphorylation.

This could be one of the ways of prebiological evolution. In general prebiological evolution is conceivable as a separate development of functions inherent in different types of carbon compounds. In fact several combinations of amino acids may reveal

catalytic properties, and even short polymers of nuclear acids are capable of self-replication and bipolar lipids can form membranes. I believe that transition to the living matter consists in putting together these functions. Apparently the advantage achieved by interconnection of different components and coordination of their functions is as fundamental a property of open system as disintegration and increase of entropy is a property of isolated systems.

Such evolution may appear on the level of molecules, cells, organisms, or societies. In this sense life can be regarded as a type of evolution of matter. It cannot be adequately defined through the properties of an object. Therefore it is difficult to indicate a definite time of appearance of life. As a type of evolution of material it existed from the very beginning.

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