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The Early Earth

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Introduction- The several observations and scientific evidence suggests that the Earth at present has a layered structure. But the question arises how the layered structure of Earth has taken the present shape from the early Earth which was a product of condensation from the solar nebular. The Solar System is presumed to have begun after one or more local supernova explosions about 4.6 Ga ago. The planetesimals began to attain the proportions of planetary embryo as a result of collisions between them. The heat generated from the collisions must have melted substantial amount of the early planet resulting in the formation of a global magma ocean and the denser, refractory material which could not melt sank inwards. This discussion aims at illucidating the driving forces and the processes within the Earth which resulted in the present layered structure of Earth.

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The Early Earth

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I. Introduction

The several observations and scientific evidence suggests that the Earth at present has a layered structure. But the question arises how the layered structure of Earth has taken the present shape from the early Earth which was a product of condensation from the solar nebular. The Solar System is presumed to have begun after one or more local supernova explosions about 4.6 Ga ago. The planetesimals began to attain the proportions of planetary embryo as a result of collisions between them. The heat generated from the collisions must have melted substantial amount of the early planet resulting in the formation of a global magma ocean and the denser, refractory material which could not melt sank inwards. This discussion aims at illucidating the driving forces and the processes within the Earth which resulted in the present layered structure of Earth.

During the Hadean or the first 660 million years of the Earth's existence, the metallic core separated from the silicate mantle. Subsequently, the atmosphere and the hydrosphere were formed and melting of the silicate mantle produced the earliest crust. But there is no rock record of Hadean. Hence, theoretical modeling and geochemistry are the only tools to reveal the mechanisms of formation of the different layers in the Earth.

II. Heat Sources Necessary to Drive Planetary Differentiation

Planetary differentiation can be defined as a process by which planets develop concentric layering and each layer differ in chemical and mineralogical compositions. The generation of such layers results from a differential mobility of elements due to differences in their physical and chemical properties. When a rock is heated, different minerals within the rock will melt at different temperatures. This phenomenon is known as partial melting and is a key process in the formation of liquid rock or magma. Once the elements have been mobilized, they will begin to migrate under the influence of pressure or gravity.

If partial melting is the principal cause of differentiation, then the Earth needs to be heated before layering begins. The principal sources of heat are primordial heat source (accretional heat and heat generated due to core formation), tidal and radiogenic heating. They are discussed as follows:

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III. Accretional Heating

During the accretion, any planetesimal of mass 'm' falling towards the Earth will acquire a velocity because of the gravitational attraction towards the Earth and hence the body will acquire a kinetic energy $E=1/2mv^2$ due to its motion (where v is the velocity of the body immediately before impact). After collision, if all the kinetic energy of motion is converted into heat, then the increase in temperature ΔT can be calculated as follows:

$$\Delta T=mv^2/(2(m+M)C)$$

where m=mass of the body; M=mass of Earth; C= specific heat capacity of Earth material;

But all the impact material must not have arrived at the same time. Accretion took place over 10^7 years. Also, the entire kinetic energy would not be converted to heat as some of it was spent in excavation of craters and some radiated into space. Nevertheless, most estimates predict temperatures to have risen above the melting point of silicate minerals and Fe-Ni. This implies that Earth has gone through an early molten stage.

The metals and silicates got separated during the molten phase of the Earth. The 'falling inwards' of the Ni-Fe rich fraction to form the core would have released potential energy. The gravitational energy lost by the inward movement of Ni-Fe would have been first converted to kinetic energy and then into thermal energy. The core formation also contributed to the primordial heat source. However, if these primordial heat sources had remained the only way of heating the Earth, their intensity would have waned through time due to continual radioactive heat loss to space. The present day active volcanoes are indications of very large amount of Earth's internal heat. This requires additional processes of heat generation.

IV. Tidal Heating

One heat source known to be generated within the planetary bodies is tidal heating, which is created by the distortion of shape resulting from mutual gravitational attraction. This effect of tidal force is manifested as the ebb and flow of tides seen around the coast. The solid Earth is also distorted by these forces and produces tides that reach a maximum amplitude of about 1m on the rocky surface. This deformation causes heating within the planet, although precisely this kind of heating is dependent on Earth's internal property. Earth shows this kind of heating within the crust and mantle.

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V. Radiogenic Heating

The experiments conducted on primordial lead in meteorites demonstrated that the formation of the Earth occurred about 4.6 Ga ago. John Joly, an Irish physicist, was one of the first to suggest that radioactive decay, leading to radiogenic heating, was an important independent source of heat within the Earth that supplements that remaining from the primordial sources.

Most elements have different isotopes (that is atoms having same number of protons but different number of neutrons).Some of these isotopes are unstable and decay to stable forms. For example, isotopes deficient in protons decay by transformation of a neutron into a proton and an electron, which expelled from the nucleus. During this process known as beta decay, the mass of the nuclide does not change significantly. In alpha decay, heavy atoms decay through the emission of an α-particle (He^{2+}) which consists of two protons and two neutrons. This process reduces the mass of the nuclide. α- and β-particle collision with adjacent nuclei during decay causes heating through the loss of kinetic energy.

The rate of decay of a radioactive parent nuclide to form a stable daughter product is proportional to the number of atoms, n, present at any time, t:

$dn/dt = -\lambda n$ where λ=decay constant characteristic of the radionuclide which is decaying.

After integration from time t=0 to t=t, we get $n=n_0e^{-\lambda t}$ where n_0= the initial number of atoms present at time t=0.

An alternative way of referring to the rate of decay of a radionuclide is by its half-life (t1/2), which is the time required for half of the parent atoms to decay.

On substituting $n=n_0/2$ and t=t1/2 into the equation $n=n_0e^{-\lambda t}$, we get,

$t1/2=\ln 2/\lambda=0.693/\lambda$, where ln 2 is the natural log of 2.

The number of radiogenic daughter atoms formed (D*) is equal to the number of parent atoms consumed (Figure to be shown on transparency). So:

$D=D_0+D^*$

$\rightarrow D=D_0+n_0-n$ where $D^*=n_0-n$

$\rightarrow D=D_0-n+ne^{\lambda t}$

where $n_0=ne^{\lambda t}$

$\rightarrow D=D_0+ n(e^{\lambda t}-1)$

If radio nuclides have short half-lives and are not replenished by the decay of other isotopes, then they may be lost altogether. One such short-lived extinct nuclide is ^{26}Al, which has a half-life of 0.73 Ma.

All those with half-lives significantly less than the age of the Earth, that is 4.6 Ga, are extinct, namely: ^{26}Al, ^{129}I, ^{146}Sm, ^{182}Hf and ^{244}Pu. The others, principally isotopes of ^{40}K, ^{87}Rb, ^{147}Sm, ^{232}Th, ^{235}U, ^{238}U are still active today.

^{26}Al dacays to ^{26}Mg and hence anomalously high abundances of ^{26}Mg relative to other isotopes of Mg is observed in materials from the early Solar System. For example the carbonaceous chondrites show high $^{26}Mg/^{24}Mg$ ratios and this suggests that a significant proportion of Al present at the time of condensation of the solar nebula was the unstable isotope ^{26}Al. This was originally created during a supernova explosion pre-dating the birth of our Solar System, and cannot be replenished by the spontaneous decay of other radiogenic elements.

The half-life of ^{26}Al is only 0.73 Ma, so the time between the supernova explosion that generated the ^{26}Al and the accretion of the meteorite parent body must have happened on a similar timescale of a few million years.

While such short lived isotopes may have been important heat sources during early stages of terrestrial planet evolution, study of Earth material indicates that it is the isotopes of the elements U, Th and K that are responsible for the most of the radiogenic heating that has occurred throughout the history of the planet. These isotopes, which have particularly long half-lives are termed as long-lived radiogenic nuclides and were present in sufficient quantities after condensation and accretion to ensure that they have remained abundant within present-day Earth.

The elements U, Th and K (and their radiogenic isotopes) are particularly concentrated in the silicate-dominatedouter layers of the Earth as they are the incompatible elements. They are particularly concentrated in the continental crust and virtually absent from the core. As a result of the incompatibility of the heat producing elements, the radiogenic heat produced per unit mass of the continental crust is, on average, over one hundred times greater than that of the underlying mantle. But because the mantle is so much more massive than the crust, in effect this means the overall radiogenic heat budget is roughly split equally between the mantle and the crust despite much greater mass of the mantle material. Ti is the decay of the long-lived radiogenic isotopes that provides sufficient heat energy to keep the Earth geologically active. Therefore, the surface heat flux is not simply the slow cooling of a once molten body, as originally envisaged by Kelvin.

VI. Heat Transfer within the Earth

Accretion, core formation and the radioactive decay heated the Earth. This internal heat is transferred to the surface by three main mechanisms and these are; conduction, convection and advection.

VII. Conduction

This is the process of heat transfer experienced when the handle of a pan becomes hot. Heat is conducted from the stove to the pan and then to its handle. Different materials, such as rocks of various compositions, conduct heat at different rates, and the efficiency of heat transfer in this manner is known as conductivity. This form of heat transfer is the most important in the outermost layer of the Earth (that is the lithosphere).

VIII. Convection

This process involves the movement of hot material from regions that are hotter to those that are cooler and return the cool material to warmer regions. During this transfer the material gives up its heat. It is particularly a efficient method of heat transfer, the medium through which transfer takes place must be fluid. The mantle can also flow when subject to temperature differences in a process known as the solid state convection and the rates are no more than few centimeters per year. This is the most efficient form of heat transfer within all but the outermost part of the mantle.

IX. Advection

Advection is the final process of transferring heat when molten material (magma) moves up through fractures in the lithosphere and remains there. Advection operates when magma spreads out at the surface as a lava flow or if it is injected, cools and crystallize within the lithosphere itself.

X. The Age of the Earth and its Layers

Radioactivity allows absolute ages to be determined from measurements of long-lived radioactive isotopes and their daughters. Several isotope systems are used to date events and processes from throughout the Earth history, but three most commonly used are the K-Ar, U-Th-Pb and Rb-Srsystems. The isotope data is illustrated on an isochron plot (or isochron diagram or isotope evolution diagram) examples of which are to be shown in figure.

Rb-Sr isotope data from a series of ordinary chondrites that define an isochron age of 4.5 Ga. This age relates to the last time the Rb and Sr were fractionated from each other by a particular process. In case of Rb and Sr, both elements are lithophile, so it is unlikely that they are fractionated by the separation of a metallic phase from a silicate fraction. However, Rb being a Group 1 alkali metal, is significantly more volatile than Sr, a Group 2 element similar to Ca, which is one of the early condensing elements. Hence Rb/Sr fractionation may relate to the loss of a volatile phase; the age indicates when the Rb/Sr ratio in ordinary chondrites was last disturbed.

U-Th system has two parent isotopes, ^{235}U and ^{238}U decaying to ^{207}Pb and ^{206}Pb respectively. By combining these two, it is possible to eliminate the U/Pb ratio and determine an age from the plot of ^{207}Pb/^{204}Pb against ^{206}Pb/^{204}Pb. In this case, the ages represent the time at which U was fractionated from Pb and, as Pb is moderately volatile element and can be lithophile, siderophile or chalcophile in different environments, it is less easy to define the process that led to U/Pb fractionation. However, iron meteorites are rich in Pb and poor in lithophile U, so the age probably represents the timing of the separation of a metallic phase. Given that the chondrite isochron passes through the Pb isotope ratio of most iron meteorites, it adds further support to this idea.

Primitive carbonaceous chondrites are thought to be among the least differentiated material in the Solar System. Among other things, they contain chondrules and Ca- and Al-rich inclusions (CAIs). Chondrules are millimeter-sized spherical droplets believed to have been produced when mineral grain assemblages were flash heated and cooled quickly. CAIs are typically cm-sized and consist of the first minerals to condense at equilibrium from a gas of solar composition. A detailed study of CAIs and chondrules yielded a ^{206}Pb/^{207}Pb isotopic age for CAIs of 4567.2±0.6 Ma, whereas that of chondrules(more primitive) is 4564.0±1.2 Ma.

The data gives an interval of 3.2±1.8 Ma between formation of the CAIs and chondrules-carbonaceous chondrites must have formed at or after the time of formation of the chondrules, that is 4564 Ma.

These data show that the oldest components of meteorites, and hence the Solar System, must be close to 4.57 Ga old, but how do we know that this age also applies to the Earth.

In the figure of Pb isotopes, the average Pb isotope ratios of Pacific sediments are compared with the data from the chondrules. The sediment data fall on or close to the meteorite isochron, implying ultimate derivation from a similar source or common parent.

XI. Radioactivity Applied to Dating

The number of ^{87}Sr daughter atoms produced by the decay of ^{87}Rb in a rock or mineral since its formation t years ago is given by substitution into the radioactive decay equation:

$$^{87}Sr = {}^{87}Sr_i + {}^{87}Rb\,(e^{\lambda t}-1)$$

where $^{87}Sr_i$ is the number of ^{87}Sratoms initially present.

But mass spectrometers can measure isotope ratios to very high presion and accuracy and so it is more convenient to work with isotope ratios.

$$^{87}Sr/^{86}Sr = (^{87}Sr/^{86}Sr)_i + (^{87}Rb/^{86}Sr)\,(e^{\lambda t}-1)$$

where ^{86}Sr=the stable isotope of the Sr and hence remains constant with time.

The above equation is in the form of y=c+mx which is a straight line equation and the equation will be valid under the assumption the system has been closed to Rb and Sr mobility from the time t to the present.

But it is difficult to measure the initial ratio. The $^{87}Sr/^{86}Sr$ at present is plotted along the y axis and the (^{87}Rb /^{86}Sr) is plotted along the x axis. Hence the intercept of the staright line gives the initial ratio. On such a diagram, a suite of cogenetic rocks or minerals having the same age define a line termed an isochron. The slope of the line gives the age of the rock or mineral.

XII. Core Formation and Magma Oceans

One potential mechanism for Fe-Ni metal separation or segregation is that the metal melts and forms an interconnected network. Whether or not this happens depends on a property known as the dihedral angle, θ. The dihedral angle is that formed by the liquid in contact with the two solid grains, which in the case of the mantle will be silicate or oxide grains. If $\theta \leq 60$, the melt will fill channels between the solid grains and form an interconnected network. If θ is greater than 60 degrees, the melt is confined to pockets at grain corners and cannot easily move unless the melt fraction is more than 10 percent.

If melt is able to connect, its rate of migration is quite rapid, and can be calculated using Darcy's law:

$$v=k/\eta\Delta\rho g$$

where v=velocity of melt relative to the solid matrix; k=permeability, η=viscosity of the melt measured in Pa s, $\Delta\rho$=density difference between the melt and the solid, and g=acceleration due to gravity.

Permeability, $k=a^2\phi/24\pi$ where a=mean radius and ϕ=melt fraction.

There was more than 40 percent of silicate melting in which the dense metal droplets sank inward. But such high percentage of melt requires tremendous amount of heat for its generation.(possible source; the collisions during accretion+radioactive heat which is much greater than today).

XIII. Core-Mantle Equilibration

The core-mantle separation resulted in the separation of the metal-loving siderophile elements and their partitioning into the core. However, trace amounts of the siderophile elements are retained in the mantle and if metal segregation were an equilibrium process then these elements would provide information about the conditions of core formation.

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