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Excerpt
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Part I

Planetary perspective

I want to know how God created this world. I am not interested in this or that phenomenon, in the spectrum of this or that element. I want to know his thoughts, the rest are details.

Albert Einstein

Overview

Earth is part of the solar system and it cannot be completely understood in isolation. The chemistry of meteorites and the Sun provide constraints on the composition of the planets. The properties of the planets provide ideas for and tests of theories of planetary formation and evolution. The Earth is often assumed to have been formed by the slow accumulation of planetesimals – small cold bodies present in early solar system history. In particular, types of stony meteorite called *chondrites* have been adopted as the probable primary material accreted by the Earth. This material, however, has to be extensively processed before it is suitable.

Study of the Moon, Mars and meterorites demonstrates that melting and basaltic volcanism is ubiquitous, even on very small bodies. Planets form hot, or become hot, and begin to differentiate at a very early stage in their evolution, probably during accretion. Although primitive objects have survived in space for the age of the solar system, there is no evidence for the survival of primitive material once it has been in a planet. One would hardly expect large portions of the Earth to have escaped this planetary differentiation, and to be 'primordial' and undegassed. The present internal structure of the Earth was mainly established 4.57 billion years ago. This is not a central dogma of current geochemical models but the use of high-precision short-lived isotope data promises to change this.

A large amount of gravitational energy is released as particles fall onto an accreting Earth, enough to raise the temperature by tens of thousands of degrees and to evaporate the Earth back into space as fast as it forms. Melting and vaporization are likely once the proto-Earth has achieved a given size. The mechanism of accretion and its time scale determine the fraction of the heat that is retained, and therefore the temperature and heat content of the growing Earth. The 'initial' temperature of the Earth was high. A rapidly growing planet retains more of the gravitational energy of accretion, particularly if there are large impacts.

The magma-ocean concept was developed to explain the petrology and geochemistry of the Moon. It proved fruitful to apply this to the Earth, taking into account the petrological differences required by the higher pressures on the Earth.

We now know that plate tectonics, at least the recycling kind, is unique to Earth, perhaps because of its size or water content. The thickness and average temperature of the lithosphere and the role of phase changes in basalt are important. *Any theory of plate tectonics must explain why the other terrestrial planets do not behave like Earth.* (Reminder: key words are embedded in the text, with the type face of the preceding sentence. These words and phrases can be entered into search engines to obtain background material, definitions and references.)

Chapter 1

Origin and early history

Earth is the namesake of the terrestrial planets, also known as the inner or rocky planets. The chemistry of meteorites and the Sun provide constraints on the composition of the bulk of these planets and they provide tests of theories of planetary formation and evolution. In trying to understand the origin and structure of the Earth, one can take the geocentric approach or the *ab initio* approach. In the former, one describes the Earth and attempts to work backward in time. For the latter, one attempts to track the evolution of the solar nebula through collapse, cooling, condensation and accretion, hoping that one ends up with something resembling the Earth and other planets. Planets started hot and had a pre-history that cannot be ignored. The large-scale chemical stratification of the Earth reflects accretionary processes.

Condensation of the nebula

The equilibrium assemblage of solid compounds that exists in a system of solar composition depends on temperature and pressure and, therefore, location and time. The condensation behavior of the elements is given in Figures 1.1 and 1.2.

At a nominal nebular pressure of 10^{-1} atm, the material would be a vapor at temperatures greater than about 1900 K. The first solids to condense at lower temperature or higher pressure are the refractory metals (such as W, Re, Ir and Os). Below about 1750 K refractory oxides of aluminum, calcium, magnesium and titanium condense, and metallic iron condenses near 1470 K (Table 1.1 and Figure 1.2). Below about 1000 K, sodium and potassium condense as feldspars, and a portion of the iron is stable as fayalite and ferrosilite with the proportion increasing with a further decrease in temperature. FeS condenses below about 750 K. Hydrated silicates condense below about 300 K.

Differences in planetary composition may depend on the location of the planet, the location and width of its feeding zone and the effects of other planets in sweeping up material or perturbing the orbits of planetesimals. In general, one would expect planets closer to the Sun and the median plane of the nebula to be more refractory rich than the outer planets. On the other hand, if the final stages of accretion involve coalescence of large objects of different eccentricities, then there may be little correspondence between bulk chemistry and the present position of the terrestrial planets (Table 1.2).

There is evidence that the most refractory elements condensed from the solar nebula as a group, unfractionated from one another, at temperatures above the condensation temperature of the Mg-silicates. Hence, the *lithophile refractory elements* (Al, Ca, Ti, Be, Sc, V, Sr, Y, Zr, Nb, Ba, rare-earth elements, Hf, Ta, Th and U and, to some extent, W and Mo) can be treated together. From the observed abundance in samples from the Moon, Earth and achondrites, there is strong support for the idea that these elements are present in the same ratios as in Cl chondrites. The abundance of the refractory elements in a given planet can be weakly constrained from

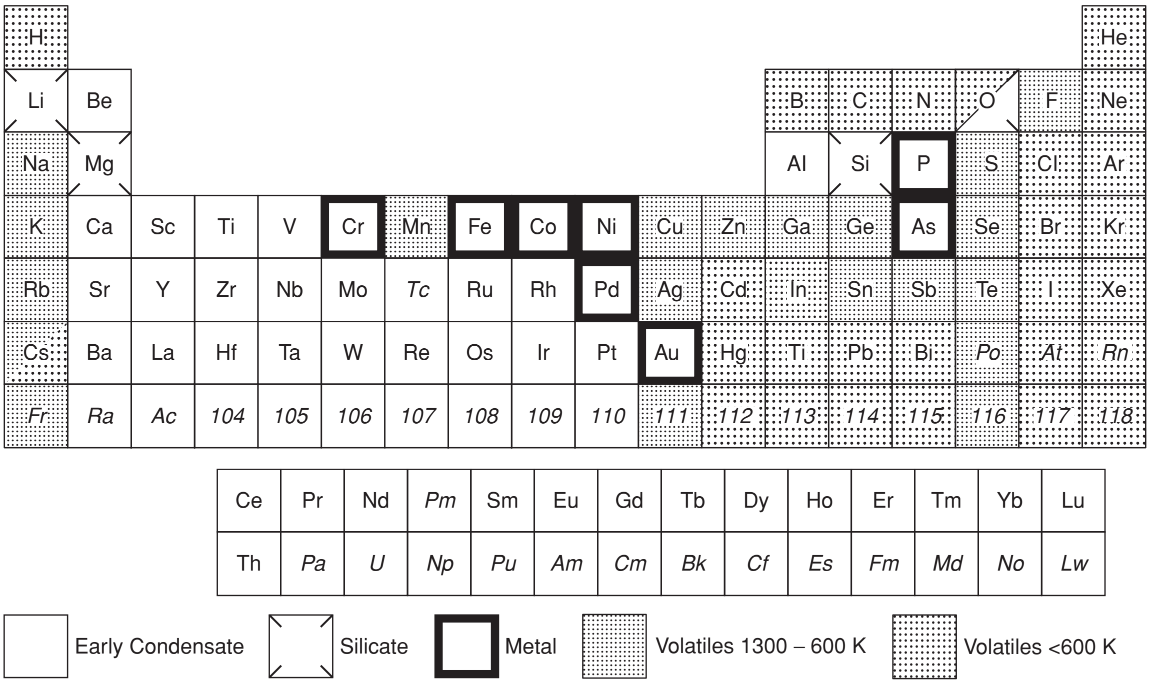

Fig. 1.1 Condensation behavior of the elements. Short-lived radioactive elements are shown in italics (after Morgan and Anders, 1980).

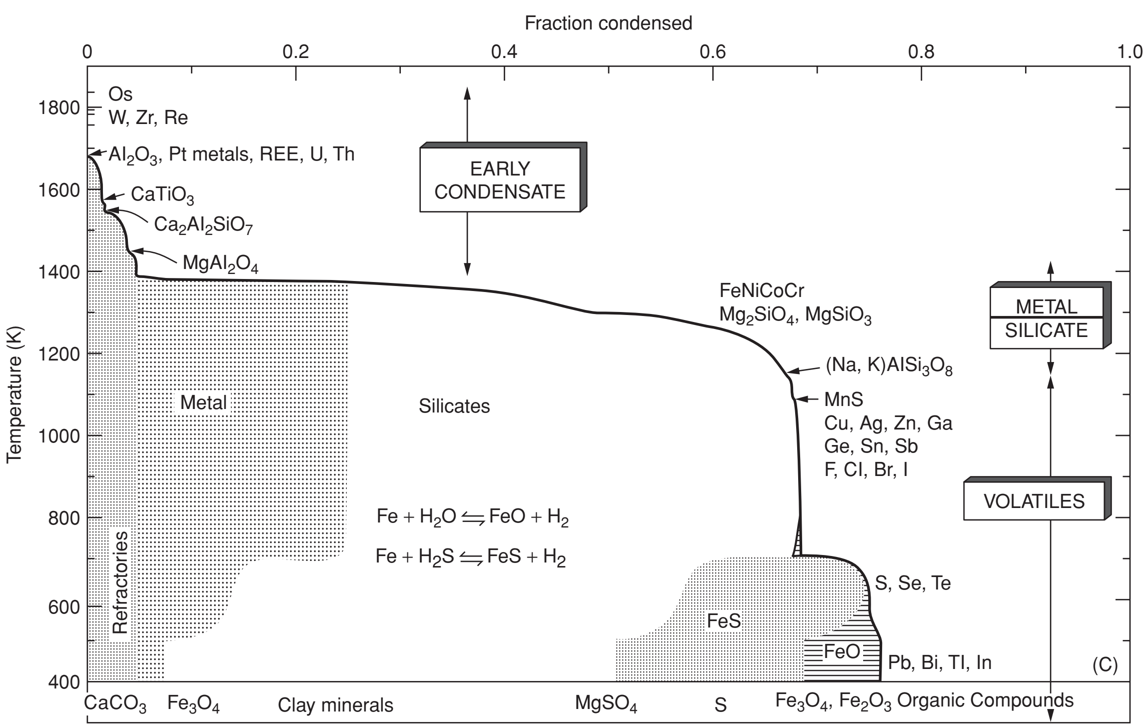

Fig. 1.2 Condensation of a solar gas at 10^{-4} atm (after Morgan and Anders, 1980).

Table 1.1 Approximate sequence of condensation of phases and elements from a gas of solar composition at 10^{-3} atm total pressure

Phase	Formula	Temperature
Hibonite	$CaAl_{12}O_{19}$	1770 K
Corundum	Al_2O_3	1758 K
Platinum metals	Pt, W, Mo, Ta Zr, REE, U, Th Sc, Ir	
Perovskite	$CaTiO_3$	1647 K
Melilite	$Ca_2Al_2SiO_7$– $Ca_2Mg_2Si_2O_7$	1625 K
	Co	
Spinel	$MgAl_2O_4$	1513 K
	Al_2SiO_5	
Metallic iron	Fe, Ni	1473 K
Diopside	$CaMgSi_2O_6$	1450 K
Forsterite	Mg_2SiO_4	1444 K
Anorthite	$CaAl_2Si_2O_8$	1362 K
	Ca_2SiO_4	
	$CaSiO_3$	
Enstatite	$MgSiO_3$	1349 K
	Cr_2O_3	
	P, Au, Li	
	$MnSiO_3$	
	MnS, Ag	
	As, Cu, Ge	
Feldspar	$(Na,K)AlSi_3O_8$	
	Ag, Sb, F, Ge	
	Sn, Zn, Se, Te, Cd	
Reaction products	$(Mg,Fe)_2SiO_4$ $(Mg,Fe)SiO_3$	1000 K
Troilite, pentlandite	FeS, (Fe, Ni)S Pb, Bi, In, Tl	700 K
Magnetite	Fe_3O_4	405 K
Hydrous minerals	$Mg_3Si_2O_7 2H_2O$, etc.	
Calcite	$CaCO_3$	<400 K
Ices	H_2O, NH_3, CH_4	<200 K

Anders (1968), Grossman (1972), Fuchs and others (1973), Grossman and Larimer (1974).

Table 1.2 Properties of the terrestrial planets

	GM 10^{18} cm^3/s^2	R km	ρ g/cm^3	I/MR^2	D* km
Earth	398.60	6371	5.514	0.3308	14
Moon	4.903	1737	3.344	0.393	75
Mars	42.83	3390	3.934	0.365	>28
Venus	324.86	6051	5.24	?	?
Mercury	22.0	2440	5.435	?	?

*Estimated crustal thickness.

the inferred abundance of their heat-producing members, uranium and thorium, and the global heat flux. But the present surface heat flow does not accurately represent the current rate of heat production. A large fraction of the present heat flow is due to cooling of the Earth, which means that only an upper bound can be placed on the uranium and thorium content. Nevertheless, this is a useful constraint particularly when combined with the lower bound on potassium provided by argon-40 and estimates of K/U and Th/U provided by magmas and the crust. There is little justification for assuming that the volatile elements joined the planets in constant proportions. In this context the volatiles include the alkali metals, sulfur and so forth in addition to the gaseous species.

Theories of planetary formation

The nature and evolution of the solar nebula and the formation of the planets are complex subjects. The fact that terrestrial planets did in fact form is a sufficient motivation to keep a few widely dispersed scientists working on these problems. There are several possible mechanisms of planetary growth. Either the planets were assembled from smaller bodies (planetesimals), a piece at a time, or diffuse collections of these bodies, clouds, became gravitationally unstable and collapsed to form planetary-sized objects. The planets, or protoplanetary nuclei, could have formed in a gas-free environment or in the presence of a large amount of gas that was subsequently dissipated. Some hypotheses speculate that large amounts of primordial helium dissolved in an early molten Earth. Others assume that the bulk of the Earth assembled gas-free and volatiles were brought in later. The intermediate stages of planetary assembly involved impacts of large objects. The final stages involved sweeping up the debris and collecting an outer veneer of exotic materials from the Sun and the outer solar system.

The planets originated in a slowly rotating disk-shaped 'solar nebula' of gas and dust with solar composition. The temperature and pressure in the hydrogen-rich disk decreased radially from its center and outward from its plane. The disk cooled by radiation, mostly in the direction normal to the plane, and part of the incandescent gas condensed to solid 'dust' particles. As the particles grew, they settled to the median plane by collisions with particles in other orbits, by viscous gas drag and gravitational attraction by the disk. The total gas pressure in the vicinity of Earth's orbit may have been of the order of 10^{-1} to 10^{-4} of the present atmospheric pressure. The particles in the plane formed rings and gaps. The sedimentation time is rapid, but the processes and time scales involved in the collection of small objects into planetary-sized objects are not clear. Comets, some meteorites and some small satellites may be left over from the early stages of accretion.

The accretion-during-condensation, or inhomogeneous-accretion, hypothesis leads to radially zoned planets with refractory and iron-rich cores, and a compositional zoning away from the Sun; the outer planets are more volatile-rich because they form in a colder part of the nebula. Superimposed on this effect is a size effect: the larger planets, having a larger gravitational cross section, collect more of the later condensing (volatile) material but they also involve more gravitational heating.

In the widely used `Safronov cosmogonical theory` (1972) it is assumed that the Sun initially possessed a uniform gas–dust nebula. The nebula evolves into a torus and then into a disk. Particles with different eccentricities and inclinations collide and settle to the median plane within a few orbits. As the disk gets denser, it breaks up into many dense accumulations where the self-gravitation exceeds the disrupting tidal force of the Sun. As dust is removed from the bulk of the nebula, the transparency of the nebula increases, and a large temperature gradient is established.

If the relative velocity between planetesimals is high, fragmentation rather than accumulation will dominate and planets will not grow. If relative velocities are low, the planetesimals will be in nearly concentric orbits and the collisions required for growth will not take place. For plausible assumptions regarding dissipation of energy in collisions and size distribution of the bodies, mutual gravitation causes the mean relative velocities to be only somewhat less than the escape velocities of the larger bodies. Thus, throughout the entire course of planetary growth, the system regenerates itself such that the larger bodies would always grow. The formation of the giant planets, however, may have disrupted planetary accretion in the inner solar system and the asteroid belt.

The initial stage in the formation of a planet is the condensation in the cooling nebula. The first solids appear in the range 1750–1600 K and are oxides, silicates and titanates of calcium and aluminum and refractory metals such as the platinum group. These minerals (such as corundum, perovskite, melilite) and elements are found in white inclusions (chondrules) of certain meteorites, most notably in Type III carbonaceous chondrites. These are probably the oldest surviving objects in the solar system. Metallic iron condenses at relatively high temperature followed shortly by the bulk of the silicate material as forsterite and enstatite. FeS and hydrous minerals appear at very low temperature, less than 700 K. Volatile-rich carbonaceous chondrites have formation temperatures in the range 300–400 K, and at least part of the Earth must have accreted from material that condensed at these low temperatures. The presence of He, CO_2 and H_2O in the Earth has led some to propose that the Earth is made up almost entirely of cold carbonaceous chondritic material – the *cold-accretion hypothesis*. Even in some current geochemical models, the lower mantle is assumed to be gas-rich, and is speculated to contain as much helium as the carbonaceous chondrites. This is unlikely. The volatile-rich material may have come in as a late veneer – the inhomogenous accretion hypothesis. Even if the Earth accreted slowly, compared to cooling and condensation times, the later stages of accretion could involve material that condensed further out in the nebula and was later perturbed into the inner solar system. A drawn-out accretion time does not imply a cold initial condition. Large impacts reset the thermometer.

The early history of planets was a very violent one; collisions, radioactive heat and core formation provided enough energy to melt the planet. Cooling and crystallization of the planet over timescales of millions of years resulted in its chemical differentiation – segregation of material according to density. This differentiation left most of the Earth's mantle different in composition from that part of the mantle from which volcanic rocks are derived. There must be material that is complementary in composition to the materials sampled by volcanoes.

The Earth and the Moon are deficient in the very volatile elements that make up the bulk of the Sun and the outer planets, and also the moderately volatile elements such as sodium, potassium, rubidium and lead. Mantle rocks contain some `primordial noble gas isotopes`. (*Reminder*: `primordial noble gas isotopes` *is a Googlet. If it is typed into a search engine it will return useful information on the topic, including definitions and references. These Googlets will be sprinkled throughout the text to provide supplementary information.*) The noble gases and other very volatile elements were most likely brought in after the bulk of the Earth accreted and cooled. The ^{40}Ar content of the atmosphere demonstrates that the Earth is an extensively degassed body; the atmosphere contains about 70% of the ^{40}Ar produced by the decay of ^{40}K over the whole age of the Earth. This may imply that most of the K and other incompatible elements are in the crust and shallow mantle.

Magma ocean

A large amount of gravitational energy is released as particles fall onto an accreting Earth, enough to evaporate the Earth back into space as fast as it forms. Even small objects can melt if they collide at high velocity. The mechanism of accretion and its time scale determine the fraction of the heat that is retained, and therefore the temperature and heat content of the growing Earth. The 'initial' temperature of the Earth was likely to have been high even if it formed from cold planetesimals. A rapidly growing Earth retains more of the gravitational energy of accretion, particularly if there are large impacts that can bury a large fraction of their gravitational energy. Evidence for early and widespread melting on such small objects as the Moon and various meteorite parent bodies attests to the importance of high initial temperatures, and the energy of accretion of the Earth is more than 15 times greater than that for the Moon. The intensely cratered surfaces of the solid planets provide abundant testimony of the importance of high-energy impacts in the later stages of accretion.

During accretion there is a balance between the gravitational energy of accretion, the energy radiated into space and the thermal energy produced by heating of the body. Latent heats associated with melting and vaporization are also involved when the surface temperature gets high enough. The ability of the growing body to radiate away part of the heat of accretion depends on how much of the incoming material remains near the surface and how rapidly it is covered or buried. Devolatization and heating associated with impact generate a hot, dense atmosphere that serves to keep the surface temperature hot and to trap solar radiation. One expects the early stages of accretion to be slow, because of the small gravitational cross section and absence of atmosphere, and the terminal stages to be slow, because the particles are being used up. The temperature profile resulting from this growth law gives a planet with a cold interior, a temperature peak at intermediate depth, and a cold outer layer. Superimposed on this is the temperature increase with depth due to self-compression and possibly higher temperatures of the early accreting particles. However, large late impacts, even though infrequent, can heat and melt the upper mantle. Formation of 99% of the mass of Earth probably took place in a few tens of millions of

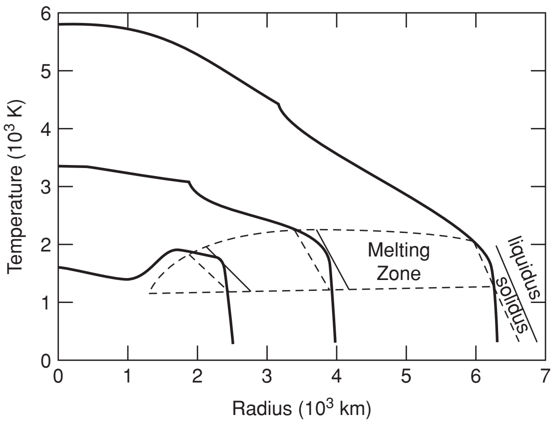

Fig. 1.3 Schematic temperatures as a function of radius at three stages in the accretion of a planet (heavy lines). Temperatures in the interior are initially low because of the low energy of accretion. The solidi and liquidi and the melting zone in the upper mantle are also shown. Upper-mantle melting and melt-solid separation is likely during most of the accretion process. Silicate melts, enriched in incompatible elements, will be concentrated toward the surface throughout accretion. The Earth, and perhaps the mantle, will be stratified by intrinsic density, during and after accretion. The Melting Zone in the upper mantle or a near-surface magma ocean processes accreting material. Temperature estimates provided by D. Stevenson.

years, around 4.55 billion years ago. The core was forming during accretion and was already in place by its end. There was likely not a `core-forming event`.

Accretional calculations, taking into account the energy partitioning during impact, have upper-mantle temperatures in excess of the melting temperature during most of the accretion time (Figure 1.3). If melting gets too extensive, the melt moves toward the surface, and some fraction reaches the surface and radiates away its heat. A hot atmosphere, a thermal boundary layer and the presence of chemically buoyant material at the Earth's surface, however, insulates most of the interior, and cooling is slow. Extensive cooling of the interior can only occur if cold surface material is subducted into the mantle. This requires a very cold, thick thermal boundary layer that is denser than the underlying mantle. This *plate tectonic mode* of mantle convection – with subduction and recycling – may only extend back into Earth history about 1 Ga (10^9 years ago). An extensive accumulation of basalt or olivine near the Earth's surface during accretion forms a buoyant layer that resists subduction. An extensively molten, slowly cooling, upper mantle, and a very slowly cooling deeper mantle are predicted.

A magma ocean freezes from the bottom but a thin chill layer may form at the surface. As various crystals freeze out of the ocean they will float or sink, depending on their density. On the Moon, plagioclase crystals float when they freeze and this is one explanation of the anorthositic highlands. On the much larger Earth, the aluminum enters dense garnet crystals and a deep eclogite-rich layer is the result. Although a magma ocean may be convecting violently when it is hot, or being stirred by impacts, at some point it cools through the crystallization temperatures of its components and the subsequent gravitational layering depends on the relative cooling rate and sinking rates of the crystals. Meanwhile, new material is being added from space and is processed in the magma ocean. A chemically stratified Earth is the end result. Accretional and convective stirring is unlikely to dominate over gravitational settling.

Magma is one of the most buoyant products of mantle differentiation and will tend to stay near the surface. A hot, differentiated planet cools by the `heat-pipe cooling mechanism of mantle convection`; pipes or sheets of magma remove material from the base of the proto-crust and place it on top of the basaltic pile, which gets pushed back into the mantle, cooling the interior. As a thick basalt crust cools, the lower portions eventually convert to dense eclogite – instead of melting – and delaminate. This also cools off the interior. As the surface layer cools further, the olivine-rich and eclogitic parts of the outer layer become denser than the interior and subduction initiates. At this point portions of the upper mantle are rapidly cooled and the thermal evolution of the Earth switches over to the plate-tectonic era. Plate tectonics is a late-stage method for cooling off the interior, but it is restricted to those parts of the interior that are less dense than slabs. A dense primitive atmosphere and buoyant outer layers are effective insulators and serve to keep the crust and upper mantle from cooling and crystallizing as rapidly as a homogenous fluid, radiating to outer space.

On a large body, such as the Earth, the dense mineral garnet forms and sinks into the interior; on a small body plagioclase forms and rises to the surface. In both cases an aluminum-poor, residual mantle forms, composed of olivine and pyroxene. The chemical stratification that forms during accretion and magma ocean crystallization may be permanent features of the planet. The importance of these processes during the earliest history of the planet cannot be over-emphasized. No part of the interior is likely to have escaped extensive heating, melting and degassing. What happened at high temperature and relatively low pressure is unlikely to be reversed.

The 'initial' state of a planet

Partial differential equations require boundary conditions and initial conditions; so do geodynamic and evolutionary models. The present surface boundary condition of the Earth is a continuously evolving system of oceanic and continental plates. The initial condition usually adopted employs one edge of Occam's razor; *the mantle started out cold and homogenous and remains homogenous today*. The more probable initial condition is based on the other edge of Occam's razor. Although a homogenous mantle with constant properties is the simplest imaginable assumption about the *outcome*, it is not consistent with a simple *process*. No one has simply explained how the mantle may have arrived at such a state, except by slow, cold, homogenous accretion. This is an unstated assumption in the standard models of mantle geochemistry. The accretion of Earth was more likely to have been a violent high temperature process that involved repeated melting and vaporization and the probable end result was a hot, gravitationally differentiated body.

That the Earth itself is efficiently differentiated there can be no doubt. Most crustal elements are in the crust, possibly all the ^{40}Ar – depending on the uncertain potassium content – is in the atmosphere and most of the siderophile elements – such as Os, Ir – are in the core. Given these circumstances, it is probable that the mantle is also zoned by chemistry and density. Large-degree melts from primitive mantle can have relatively unfractionated ratios of such elements as Sm, Nd, Lu and Hf, giving 'chondritic' isotope ratios. This has confused the issue regarding the possible presence of *primordial unfractionated reservoirs*.

The assumed starting composition for the Earth is usually based on *cosmic* or *meteoritic abundances*. The refractory parts of carbonaceous, ordinary or enstatite chondrites are the usual choices. These compositions predict that the lower mantle has more silicon than the olivine-rich buoyant shallow mantle and that only a small fraction of the mantle, or even the upper mantle, can be basaltic. The volatile components that are still in the Earth were most likely added to Earth as a late veneer after most of the mass had already been added and the planet had cooled to the point where it could retain volatiles.

A process of RAdial ZOne Refining (RAZOR) during accretion may remove incompatible and volatile elements and cause purified dense materials to sink. Crystallizing magma oceans at the surface are part of this process. The formation of a deep reservoir by perovskite fractionation in a magma ocean is not necessary. The magma ocean may always have been shallower than the perovskite-phase boundary – roughly 650 km depth – but as the Earth accretes, the deeper layers will convert to high-pressure phases. There is no need for material in the upper mantle to have been in equilibrium with the dense phases that now exist at depth.

Prior to the era of plate tectonics, the Earth was probably surfaced with thick crustal layers, which only later became dense enough to sink into the mantle. But because of the large stability field of garnet, there is a subduction barrier, currently near 600 km. The great buoyancy of young and thick oceanic crust, particularly oceanic plateaus, dehydration of recycled material, the low melting temperature of eclogite, and the subduction barrier to eclogite (and harzburgite) probably prevents formation of deep fertile and radioactive layers, even after the onset of plate tectonics.

The RAZOR process sets the initial stage for mantle evolution, including the distribution of radioactive elements. This step is often overlooked in geochemical and geodynamic models; it is usually assumed that most of the radioactive elements are still in the deep mantle. The initial temperatures may have been forgotten but the stratification of major and radioactive elements may be permanent.

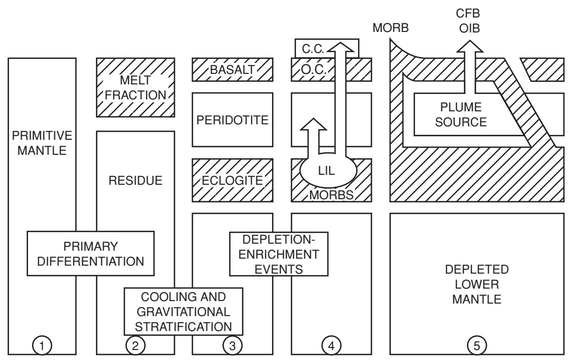

Fig. 1.4 A model for the early evolution of the mantle. Primitive mantle (1) is partially molten either during accretion or by subsequent whole-mantle convection, which brings the entire mantle across the solidus at shallow depths. Large-ion lithophile (LIL) elements are concentrated in the melt. The deep magma ocean (2) fractionates into a thin plagioclase-rich surface layer and deeper olivine-rich and garnet-rich cumulate layers (3). Late-stage melts in the eclogite-rich cumulate are removed (4) to form the continental crust (C.C.), enrich the shallow peridotite layer and deplete MORBs, the source region of oceanic crust (O.C.) and lower oceanic lithosphere. Partial melting of PLUME – or Primary Layer of Upper Mantle Enrichment – in the upper mantle (5) generates continental flood basalts (CFB), ocean-island basalts (IOB) and other enriched magmas, leaving a depleted residue (harzburgite) layer – perisphere – that stays in the upper mantle because of its buoyancy. Enriched or hot-spot magmas (EMORB, OIB, CFB) may be from a shallow part of the mantle and may represent delaminated C.C. Most of the mantle has been processed through the melting zone and is depleted in the heat-producing elements such as U and Th, which are now in the crust and upper mantle.

Evolution of a planet

Isotopic studies indicate that distinct geochemical components formed in the mantle early in its history. *Zone refining during accretion* and crystallization of a deep magma ocean are possible ways of establishing a chemically zoned planet (Figure 1.4). At low pressures basaltic melts are less dense than the residual refractory crystals, and they rise to the surface, taking with them many of the trace elements. The refractory crystals themselves are also less dense than undifferentiated mantle and tend to concentrate in the shallow mantle.

As the Earth accretes and grows, the crustal elements are continuously concentrated into the melts and rise to the surface. When these melts freeze, they form the crustal minerals that are rich in silicon, calcium, aluminum, potassium and the large-ion lithophile (LIL) elements. Melts generally are also rich in FeO compared to primitive material. This plus the high compressibility of melts means that the densities of melts and residual crystals converge, or even cross, as the pressure increases. They cross again as phase changes increase the density of the solids. Melt separation is therefore difficult at depth, and melts may even drain downward at very high pressure, until the silicate matrix undergoes a phase change. During accretion the majority of the melt-crystal separation occurs at low pressure. All of the material in the deep interior has passed through this low-pressure melting stage in a sort of continuous zone refining. The magnesium-rich minerals, Mg_2SiO_4 and $MgSiO_3$, have high melting temperatures and are fed through the melting zone into the interior. Even if the accreting material is completely melted during assembly of the Earth, these minerals will be the first to freeze, and they will still separate from the remaining melt. The downward separation of iron-rich melts, along with nickel, cobalt, sulfur and the trace siderophile elements, strips these elements out of the crust and mantle.

The aluminum, calcium, titanium and sodium contents in chondritic and solar material, restrict the amount of basalt that can be formed, but are adequate to form a crust some 200 km thick. The absence of such a massive crust on the Earth might suggest that the Earth has not experienced a very efficient differentiation. On the other hand, the size of the core