

Theory of the Earth

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Chapter 1. The Terrestrial Planets

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Abstract:

Earth is part of the solar system. Although it is the most studied planet, it cannot be completely understood in isolation. The chemistry of meteorites and the Sun provide constraints on the composition of the bulk of the Earth. The properties of other planets provide ideas for and 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. In Chapter 1 I develop the external evidence that might be useful in understanding the Earth. In Chapter 2 I describe the Earth and Moon.

The Terrestrial Planets

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.

—EINSTEIN

Earth is part of the solar system. Although it is the most studied planet, it cannot be completely understood in isolation. The chemistry of meteorites and the Sun provide constraints on the composition of the bulk of the Earth. The properties of other planets provide ideas for and 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. In Chapter 1 I develop the external evidence that might be useful in understanding the Earth. In Chapter 2 I describe the Earth and Moon.

THEORIES OF PLANETARY FORMATION

The nature and evolution of the solar nebula and the formation of the planets are complex and difficult 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.

The planets are now generally thought to have 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 decrease both radially from its center and outward from its plane. The disk cools by radiation, mostly in the direction normal to the plane, and part of the incandescent gas condenses to solid "dust" particles. As the particles grow, they settle to the median plane by processes involving collisions with particles in other orbits, by viscous gas drag and gravitational attraction by the disk. The total pressure in the vicinity of Earth's orbit may have been of the order of 10^{-3} to 10^{-4} bar. The particles in the plane probably formed rings and gaps. The sedimentation time is fairly rapid, but the processes and time scales involved in the collection of small objects into planetary sized objects are not clear. The common thread of all cosmogonic theories is that the planets formed from dispersed material, that is, from a protoplanetary nebula. Comets, some meteorites and some small satellites may be left over from these early stages of accretion.

The following observations are the main constraints on theories of planetary origin:

1. Planetary orbits are nearly circular, lie virtually in a single plane, and orbit in the same sense as the Sun's rotation. The Sun's equatorial plane is close to the orbital plane. The planets exhibit a preferred sense of rotation.
2. The distribution of planetary distances is regular (Bode's Law).
3. The planets group into compositional classes related to distance from the Sun. The inner, or terrestrial planets (Mercury, Venus, Earth and Mars), are small, have high density, slow rotation rates and few satellites. The Moon is often classified as a terrestrial planet. The giant planets (Jupiter, Saturn, Uranus and Neptune) are large, have low density, rotate rapidly and have numerous satellites. Although the Sun contains more than 99 percent of the

mass of the solar system, the planets contain more than 98 percent of the angular momentum.

Apart from the mechanisms of accretion and separation of planetary from solar material, there are several important unresolved questions.

How dense was the protoplanetary nebula? A lower limit is found by taking the present mass of the planets and adding the amount of light elements necessary to achieve solar composition. This gives about 10^{-2} solar mass. Young stars in the initial stages of gravitational contraction expel large quantities of matter, possibly accounting for several tens of percent of the star's mass. Some theories therefore assume a massive early nebula that may equal twice the mass of the Sun including the Sun's mass. T-Tauri stars, for example, expel about 10^{-6} solar mass per year for 10^5 to 10^6 years.

What are the time scales of cooling, separation of dust from gas, growth of asteroidal size bodies, and growth of planets from meter- to kilometer-size objects? If cooling is slow compared to the other processes, then planets may grow during cooling and will form inhomogeneously. If cooling is fast, then the planets may form from cold material and grow from more homogeneous material.

The accretion-during-condensation, or inhomogeneous accretion, hypothesis would lead to radially zoned planets with refractory and iron-rich cores, and a compositional zoning away from the Sun; the outer planets would be more volatile-rich. 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.

The Safronov (1972) cosmogonical theory is currently the most popular. 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 goes unstable and 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 in the nebula.

The mechanism for bringing particles together and keeping them together to form large planets is obscure. A large body, with an appreciable gravity field, can attract and retain planetesimals. Small particles colliding at high speed disintegrate and have such small gravitational cross section that they can attract only nearby particles. Large collections of co-rotating particles, with minimum relative velocities, seems to be a prerequisite condition. Self-gravitation of the aggregate can then bring the particles together. Small bodies might also act as condensation nuclei and therefore add material directly from the gaseous phase. In the Safronov theory, accumulation of 97–98 percent of the Earth occurred in about 10^8 years. In other theories the accretion time is much shorter, 10^5 – 10^6 years.

If the relative velocity between planetesimals is too high, fragmentation rather than accumulation will dominate and planets will not grow. If relative velocities are too low, the planetesimals will be in nearly concentric orbits and the collisions required for growth will not take place. Safronov (1972) showed that 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 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 (such as Al_2O_3 , CaTiO_3 , $\text{Ca}_2\text{Al}_2\text{Si}_2\text{O}_7$) 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. 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 only 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 CO_2 and H_2O on the Earth has led some to propose that the Earth was made up entirely of cold carbonaceous chondritic material—the cold accretion hypothesis. Turekian and Clark (1969) assume that volatile-rich material came in as a late veneer—the inhomogeneous 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. The Earth and the Moon are deficient in not only the very volatile elements that make up the bulk of the Sun and the outer planets, but also the moderately volatile elements such as sodium, potassium, rubidium and lead.

A large amount of gravitational energy is released as the particles fall onto a growing 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. There are mechanisms for buffering the temperature rise and to retain material even if it vaporizes, but melting and vaporization are likely once the proto-Earth has achieved a given size, say lunar 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 is 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.

The initial temperature distribution in a planet can be estimated by using the equation of conservation of energy acquired during accretion:

$$\rho \frac{GM(r)}{r} dr = \varepsilon \sigma [T^4(r) - T_b^4] dt + \rho C_p [T(r) - T_b] dt$$

where t is the time, ρ is the density of the accreting particles, G is the gravitational constant, $M(r)$ is the mass of a growing planet of radius r , σ is the Stefan-Boltzmann constant, ε is the emissivity, C_p is the specific heat, T is the temperature at radius r , and T_b is the blackbody radiation temperature. The equation gives the 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. An impacting body cannot bury all of its heat since heat is transferred to the planetary material, and both the projectile and target fragments are thrown large distances through the atmosphere, cooling during transit and after spreading over the surface. Parts of the projectile and ejecta are buried and must, of course, conduct their heat to the surface before the heat can be radiated back to space. Gardening by later impacts can also bring buried hot material to the surface. Devolatilization and heating associated with impact can be expected to generate a hot, dense atmosphere that serves to keep the surface temperature hot and to trap solar radiation.

Intuitively, one would expect 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. A convenient expression for the rate of accretion that has these characteristics is (Hanks and Anderson, 1969)

$$\frac{dr}{dt} = k_1 t^2 \sin k_2 t$$

where k_1 and k_2 can be picked to give a specified final radius and accretion time. 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 up and melt the upper mantle.

One limiting case is that at every stage of accretion the surface temperature of the Earth is such that it radiates energy back into the dust cloud at precisely the rate at which gravitational energy is released by dust particles free-falling onto its surface. By assuming homogeneous accretion spread out over 10^6 years, the maximum temperature is 1000 K. For this type of model a short accretion time is required to generate high temperatures. The Earth, however, is unlikely to grow in radiative equilibrium. Higher internal temperatures can be achieved if the Earth accumulated partly by the continuing capture of planetesimal swarms of meteoritic bodies. These bodies hit the Earth at velocities considerably higher than free fall and, by shock waves, generate heat at depth in the impacted body.

Modern 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 1-1). 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 upper mantle can only occur if cold surface material is subducted into the mantle. This requires an unstable surface layer, that is, a very cold, thick thermal boundary layer that is denser than the underlying mantle. An extensive accumulation of basalt or olivine near the Earth's surface during accretion forms a buoyant layer that resists subduction. An extensively molten upper mantle is

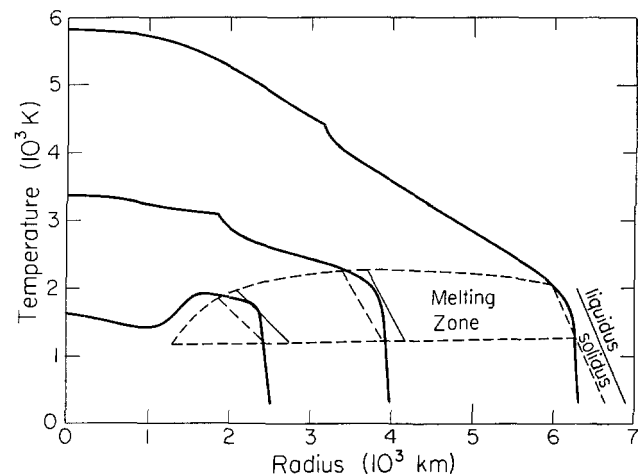


FIGURE 1-1

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 liquidus 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. Temperature estimates provided by D. Stevenson (personal communication).

therefore likely during accretion. As a thick basalt crust cools, the lower portions eventually convert to dense eclogite, and at this point portions of the upper mantle can be rapidly cooled. A dense primitive atmosphere is also an effective insulating agent and serves to keep the crust and upper mantle from cooling and crystallizing rapidly.

All things considered, it is likely that impact melting and gravitational separation combined with internal radioactive heating resulted in terrestrial planets that were being differentiated while they were accreting. Extensive upper-mantle melting gives an upward concentration of melts and the incompatible and volatile elements and burial of dense refractory crystals and melts. Dense melts include Fe, FeS and FeO-MgO-rich melts that form at high pressure. The enrichment of volatiles and incompatible elements in the Earth's atmosphere and crust and the depletion of siderophile elements in the upper mantle point toward a very effective chemical separation of this type, as does the presence of a high crust and a dense core. There is evidence from both the Earth and the Moon that these bodies were covered by deep magma oceans early in their history. The lifetime of such an ocean depends on the temperature of the atmosphere, the thickness of an insulating crust and the rate of energy delivery, from outside, by impacts and, from inside, through the thermal boundary layer at the base of the ocean. Removal of crystals from a crystallizing magma ocean and drainage of melt from a cooling crystal mush (also, technically a magma) are very much faster processes than cooling and crystallization times. Therefore, an expected result of early planetary differentiation is a stratified composition.

The following is a plausible variant of the *inhomogeneous* accretion hypothesis. Planets accrete as the nebula cools, and the accreting material has the composition of the solids that are in equilibrium with the nebula at that temperature plus the more refractory material that has condensed earlier and escaped accretion. The mean composition of a planet therefore becomes less refractory with time and with radius. After dissipation of the nebula, the terrestrial planets continue to slowly accrete material that has condensed in their vicinity and the more volatile material that condensed farther out in the solar system. It is not necessary that all of the refractories and iron be accreted before the silicates. There is always unaccreted material available for interaction with the gas. Iron, for example, is accreted as metal at the early stages but reacts with the silicates to form the ferromagnesian silicates that are accreted later. Likewise, calcium, aluminum, uranium and thorium are available for incorporation into the later condensates, but they are enriched in the early condensates. FeS will condense and accrete at low temperatures unless all the Fe metal has been removed by earlier processes.

The viability of the inhomogeneous accretion hypothesis depends on the relative time scales of nebula cooling and accretion in the early stages of condensation. If cooling is

slow relative to accretion rates, then the iron and refractories will form the initial nuclei of the planets. Alternatively, cooling can be rapid if temperatures in the vicinity of the Earth do not drop far into the olivine-pyroxene stability field before dissipation of the nebula. In this case, the majority of the mantle would be added by material perturbed into Earth orbit from cooler parts of the nebula. The earliest condensates also have more time for accretion and possibly experience more viscous drag in the dense, early nebula. The presence of water at the Earth's surface and the siderophile content of the upper mantle indicate that the later stages of accretion did not involve substantial amounts of metallic iron (Lange and Ahrens, 1984). Thus, there are several arguments supporting the view that the proto-Earth was refractory and became more volatile-rich with time. The atmospheres of the terrestrial planets are apparently secondary, formed by outgassing of the interior and devolatilization of late impacts. Primitive atmospheres are generally thought to have been blown away either by a strong solar wind or by giant impacts. Giant impacts in early Earth history may have blasted material into orbit to form the Moon (see Chapter 2) and, in later Earth history, been responsible for the various extinctions that punctuate the paleontological record. Some meteorites found on Earth are thought to have been derived from the surface of the Moon and Mars by large impacts on these bodies.

There is no particular reason to believe that there were originally only four or five terrestrial planets of Mars or Moon size or greater. The sweeping up of multiple small planets by the remaining objects is, in effect, a mechanism of rapid accretion. The early history of the surviving terrestrial planets is therefore violent and characterized by melting and remelting events.

METEORITES

Using terrestrial samples, we cannot see very far back in time or very deep into a planet's interior. Meteorites offer us the opportunity to extend both of these dimensions. Some meteorites, the chondrites, are chemically *primitive*, having compositions—volatile elements excluded—very similar to that of the sun. The volatile-rich carbonaceous chondrites are samples of slightly altered, ancient planetesimal material that condensed at moderate to low temperatures in the solar nebula. The nonchondritic meteorites are *differentiated* materials of nonsolar composition that have undergone chemical processing like that which has affected all known terrestrial and lunar rocks.

Meteorites are assigned to three main categories. Irons (or siderites) consist primarily of metal; stones (or aerolites) consist of silicates with little metal; stony irons (or siderolites) contain abundant metal and silicates. These are further subdivided in various classification schemes, as listed in Table 1-1.

TABLE 1-1
 Classification and Characteristics of Stony Meteorites and Iron-rich Meteorites
 (Elements, Weight Percent; Ratios Based on Atomic Percent)

Meteorite class	Minerals	S	O	Fe ^T	Fe ⁰	Fe		Al/Si	Ca/Si	Mg/Si	Remarks
						Fe + Mg					
Stones (95 percent of meteorite population)											
Chondrites (86 percent)											
Carbonaceous (5 percent)	Layer-silicates ("clays")	5.9	46.0	18.4	0			0.086	0.072–0.088	1.05–1.06	No chondrules
C1 = CI e.g. Ivuna; Orgueil	18–22 pct. H ₂ O										
C2 = CM = CH e.g. Mighei; Murchison	6–16 pct. H ₂ O	3.4	41.5	21.9	0–1			0.084–0.094	0.072	1.02–1.04	
C3 = CIII e.g. CO = Ornans CV = Vigarano Allende	Olivine, refractory minerals <4 pct. H ₂ O	2.2	36.5	25.2	0–8	6–23		0.092, 0.12–0.14	0.074, 0.09	1.05 1.12	
Ordinary (81 percent)											
E = Enstatite e.g. Abee; Khairpur	Enstatite, Fe-Ni	3.3–5.9	29.3	25–33	19–25	.04–1.4	0.048		0.036	0.73–0.81	Fs ₀ , SiO ₂ , FeS, CaS
H = High iron = Bronzite	Olivine, bronzite, Fe-Ni	2.1	35.1	27.6	15–19	16–19	0.06–0.07	0.05	0.095–0.97	0.95–0.97	Fa _{15–19} , Fs _{15–17}
L = Low iron = "Hypersthene"	Olivine, bronzite, Fe-Ni	2.1	38.1	21.8	4–9	21–25	0.061	0.048	0.92–0.94	0.92–0.94	Fa _{22–25} , Fs _{19–22}
LL = Very low iron = Amphoterite = Soko-Banja	Olivine	2.3	38.8	20.0	0.3–3	27–32	0.062	0.046	0.92–0.93	0.92–0.93	Fa _{26–32} , Fs _{22–25}
Achondrites (9 percent)											
Calcium-poor	Olivine, pigeonite, Fe-Ni	0.5	45.4								
Ureilites = Olivine-pigeonite				14.5	0.3–6	10	0.016	0.023	1.37		
†Chassignites = Basaltic				20.6	1	25	0.042	0.015	1.28		
Aubrites = Enstatite	Enstatite			1.0	<1	<0.03	0.02	0.021	1.0		
*Diogenites = "Hypersthene," "Cumulates"	Orthopyroxene	0.4		13.5		25–27	0.034	0.029	0.74		Fs _{23–27} , An _{85–90}
Calcium-rich			42.4								
*Howardites and	Orthopyroxene, pigeonite, plagioclase	0.27		13.9	<0.3	25–40	0.191	0.137	0.47		
*Eucrites = Basaltic e.g. Juvinas, Pasarnonte, Stannern, Moore Co.		0.20		14.4	<0.1	50–67	0.290	0.234	0.22		
Angrites = Augite	Augite	0.45		7.5			0.233	0.595	0.34		
†Nakhlites = Basaltic 'Shergottites = Basaltic	cpx, ol cpx, plag	0.06		16.6			0.042	0.332	0.37		
Stony Irons (1 percent)											
*Mesosiderites	Orthopyroxene, plagioclase, Fe-Ni	1.1				44–52					
Pallasites	Olivine, Fe-Ni	0.19				69					
Irons (4 percent)											
Octahedrites	Fe-Ni	0.02–0.09				90.5–92					
Hexahedrites	Fe-Ni	0.06				93.6					
Nickel-rich Ataxites	Fe-Ni	0.08				79.6					

*The eucritic association.

†The SNC association.

TABLE 1-2
Major Minerals of Calcium-Aluminum-Rich Inclusions (CAI)

Mineral	Volume Percent	Condensation Temperature *
Spinel ($MgAl_2O_4$)	15–30	1513–1362 K
Melilite ($Ca_2Al_2SiO_7$)	0–85	1625–1450
Perovskite ($CaTiO_3$)	0–2	1647–1393
Anorthite ($CaAl_2Si_2O_8$)	0–50	1362
Pyroxenes	0–60	1450

Grossman (1972).

*Lower temperature is temperature at which phase reacts with nebular gas to form new phase.

Carbonaceous Chondrites

Carbonaceous chondrites contain unusually high abundances of volatile components such as water and organic compounds, have low densities, and contain the heavier elements in nearly solar proportions. They also contain carbon and magnetite. These characteristics show that they have not been strongly heated, compressed or altered since their formation; that is, they have not been buried deep inside planetary objects.

The C1 or CI meteorites are the most extreme in their primordial characteristics and are used to supplement solar values in the estimation of cosmic composition. The other categories of carbonaceous chondrites, CII (CM) and CIII (CO and CV), are less volatile-rich.

Some carbonaceous chondrites contain calcium-aluminum-rich inclusions (CAI), which appear to be high-temperature condensates from the solar nebula. The minerals (Table 1-2) include anorthite ($CaAl_2Si_2O_8$), spinel, diopside, melilite, perovskite ($CaTiO_3$), hibonite ($CaAl_{12}O_{19}$) and the Al-Ti pyroxene, fassaite. These inclusions are found in CV and CO chondrites, most notably (because of the total volume of recovered material) the Allende meteorite. Theoretical calculations show that compounds rich in Ca, Al and Ti, including the above minerals, are among the first to condense in a cooling solar nebula. Highly refractory elements are strongly enriched in the CAI compared to C1 meteorites, but they occur in C1, or cosmic, ratios.

Chondrites are named after the rounded fragments, or chondrules, that they contain. Some of these chondrules appear to be frozen drops of silicate liquid and others resemble hailstones in their internal structure. Whatever their origin, the presence of chondrules indicates the composite nature of meteorites and the melting or remelting episodes that characterized the history of at least some of their components. C1 "chondrites" are fine grained and do not contain chondrules. They are chemically similar, however, to the true chondrites (see Table 1-3).

Ordinary Chondrites

As the name suggests, ordinary chondrites are more abundant, at least in Earth-crossing orbits, than all other types of meteorites. They are chemically similar but differ in their contents of iron and other siderophiles, and in the ratio of

TABLE 1-3
Compositions of Chondrites (Weight Percent)

	Enstatite	Ordinary		Carbonaceous			
		H	L	CI	CM	CO	CV
Si	16.47–20.48	17.08	18.67	10.40	12.96	15.75	15.46
Ti	0.03–0.04	0.06	0.07	0.04	0.06	0.10	0.09
Al	0.77–1.06	1.22	1.27	0.84	1.17	1.41	1.44
Cr	0.24–0.23	0.29	0.31	0.23	0.29	0.36	0.35
Fe	33.15–22.17	27.81	21.64	18.67	21.56	25.82	24.28
Mn	0.19–0.12	0.26	0.27	0.17	0.16	0.16	0.16
Mg	10.40–13.84	14.10	15.01	9.60	11.72	14.52	14.13
Ca	1.19–0.96	1.26	1.36	1.01	1.32	1.57	1.57
Na	0.75–0.67	0.64	0.70	0.55	0.42	0.46	0.38
K	0.09–0.05	0.08	0.09	0.05	0.06	0.10	0.03
P	0.30–0.15	0.15	0.15	0.14	0.13	0.11	0.13
Ni	1.83–1.29	1.64	1.10	1.03	1.25	1.41	1.33
Co	0.08–0.09	0.09	0.06	0.05	0.06	0.08	0.08
S	5.78–3.19	1.91	2.19	5.92	3.38	2.01	2.14
H	0.13	—	—	2.08	1.42	0.09	0.38
C	0.43–0.84	—	—	3.61	2.30	0.31	1.08
Fe ⁰ /Fe _{tot}	0.70–0.75	0.60	0.29	0.00	0.00	0.09	0.11

Mason (1962).

TABLE 1-4
Normative Mineralogy of Ordinary Chondrites

Species	High Iron	Low Iron
Olivine	36.2	47.0
Hypersthene	24.5	22.7
Diopside	4.0	4.6
Feldspar	10.0	10.7
Apatite	0.6	0.6
Chromite	0.6	0.6
Ilmenite	0.2	0.2
Troilite	5.3	6.1
Ni-Fe	18.6	7.5

Mason (1962).

oxidized to metallic iron. As the amount of oxidized iron decreases, the amount of reduced iron increases. Olivine is the most abundant mineral in chondrites, followed by hypersthene, feldspar, nickel-iron, troilite and diopside with minor apatite, chromite and ilmenite (Table 1-4). The composition of the olivine varies widely, from 0 to 30 mole percent Fe_2SiO_4 (Fa). Enstatite chondrites are distinguished from ordinary chondrites by lower Mg/Si ratios (Table 1-5), giving rise to a mineralogy dominated by MgSiO_3 and having little or no olivine. They formed in a uniquely reducing environment and contain silicon-bearing metal and very low FeO silicates. They contain several minerals not found elsewhere (CaS, TiN, $\text{Si}_2\text{N}_2\text{O}$). In spite of these unusual properties, enstatite chondrites are within 20 percent of solar composition for most elements. They are extremely old and have not been involved in major planetary processing.

Achondrites

The achondrites are meteorites of igneous origin that are thought to have been dislodged by impact from small bodies in the solar system. Some of these may have come from the asteroid belt, others are almost certainly from the Moon, and one subclass (the SNC group) may have come from Mars. Many of the achondrites crystallized between 4.4 and 4.6 billion years ago. They are extremely diverse and are chemically dissimilar to chondrites. They range from al-

TABLE 1-5
Element Ratios (by Weight) in Four Subtypes
of Chondritic Meteorites

Ratio	C1	H	L	E6
Al/Si	0.080	0.063	0.063	0.044
Mg/Si	0.91	0.80	0.79	0.71
CalAl	1.10	1.11	1.08	1.06
Cr/Mg	0.025	0.025	0.026	0.024

most monomineralic olivine and pyroxene rocks to objects that resemble lunar and terrestrial basalts. Two important subgroups, classified as basaltic achondrites, are the *eucrites* and the *shergottites*. Two groups of meteoritic breccias, the *howardites* and the *mesosiderites*, also contain basaltic material. The eucrites, howardites, mesosiderites and diogenites appear to be related and may come from different depths of a common parent body. They comprise the *eucritic association*. The shergottites, nakhlites and chassignites form another association and are collectively called the SNC meteorites.

Eucrites are primarily composed of anorthite, $\text{CaAl}_2\text{Si}_2\text{O}_8$, and pigeonite, $\text{MgFeSi}_2\text{O}_6$, and are therefore plagioclase-pyroxene rocks similar to basalts. They also have textures similar to basalts. However, terrestrial basalts have higher abundances of sodium, potassium, rubidium and other volatile elements and have more calcium-rich pyroxenes. Eucrite plagioclase is richer in calcium and poorer in sodium than terrestrial basaltic feldspar. The presence of free iron in eucrites demonstrates that they are more reduced than terrestrial basalts.

Eucrites have reflectance spectra similar to Vesta, the second-largest asteroid, and this or a similar asteroid may be the parent body. Trace-element and petrological studies suggest that eucrites could form by about 5 to 15 percent melting, at low pressure, of material similar to chondritic meteorites, leaving behind olivine, pigeonite and plagioclase. Some eucrites exhibit high vesicularity, suggesting crystallization at low pressure. Others have characteristics of cumulates in layered igneous intrusions. Some are similar to lunar and ancient terrestrial calcic anorthosites.

Diogenites are achondrites that consist almost entirely of magnesian orthopyroxene (Fs_{23-27}), bronzite, and minor plagioclase ($\text{An}_{-,-}$). They appear to be metamorphosed cumulates related to eucrites, howardites, and mesosiderites, which have similar isotopic characteristics.

Howardites are mineralogically similar to eucrites, but are breccias that, from a variety of evidence (solar rare-gas content, particle tracks, micrometeorite craters), appear to be from the surface regolith of their parent body. They appear to be essentially mixtures of diogenites and eucrites.

Mesosiderites are stony irons (17–80 wt. percent metal) with silicate fractions similar to diogenites and eucrites. They appear to be a mixture of mantle and core-forming material from their parent body.

Studies of basalts from the Moon and the eucrite parent body *have several* important implications for the early history of the Earth and the other terrestrial planets. They show that even very small bodies can melt and differentiate. The energy source must be due to impact, rapid accretion, short-lived radioactive isotopes or formation in a hot nebula. The widespread occurrence of chondrules in chondritic meteorites also is evidence for high temperatures and melting in the early solar system. The depletion of volatiles in eucrites and lunar material suggests that small planets,

and the early planetesimal stage of planet formation, may be characterized by volatile loss. These extraterrestrial basalts also contain evidence that free iron was removed from their source region. The process of core formation must start very early and is probably contemporaneous with accretion.

Shergottites are remarkably similar to terrestrial basalts. They are unusual, among meteorites, for having very low crystallization ages, about 10^9 years, and, among basalts, for having abundant maskelynite or glassy shocked plagioclase. Shergottites also contain augite, pigeonite and magnetite. In contrast to eucrites, the plagioclase-maskelynite contains terrestrial-type abundances of sodium and calcium. The shergottites are so similar to terrestrial basalts that their source regions must be similar to the upper mantle of the Earth. The similarities extend to the trace elements, be they refractory, volatile or siderophile, suggesting a similar evolution for both bodies. The young crystallization ages imply that the shergottites are from a large body, one that could maintain igneous processes for 3 billion years. Cosmic-ray exposure ages show that they were in space for several million years after ejection from their parent body. Shergottites are slightly richer in iron and manganese than terrestrial basalts, and, in this respect, they are similar to the eucrites. They contain no water and have different oxygen isotopic compositions than terrestrial basalts. The major-element chemistry is similar to that inferred for the martian soil. The rare-gas contents of shergottites are also similar to the martian atmosphere, giving strong circumstantial support to the idea that these meteorites may have come from the surface of Mars. In any case, these meteorites provide evidence that other objects in the solar system have similar chemistries and undergo similar processes as the Earth's upper mantle. The growing Earth probably always had basalt at the surface and, consequently, was continuously zone-refining the incompatible elements toward the surface. The corollary is that the deep interior of a planet is refractory and depleted in volatile and incompatible elements. The main difference between the Earth and the other terrestrial planets, including any meteorite parent body, is that the Earth can recycle material back into the interior. Present-day basalts on Earth may be recycled basaltic material that formed during accretion and in early Earth history rather than initial melts from a previously unprocessed peridotitic parent. Indeed, no terrestrial basalt shows evidence, if all the isotopic and geochemical properties are taken into account, of being from a primitive, undifferentiated reservoir.

PLANETARY ATMOSPHERES

The very volatile elements are concentrated near the surface of a planet and provide important clues as to the average composition of the interior, the mode of formation and the

outgassing history. In this section I use the term "volatile" to refer to the very volatile (atmospheric) elements. It is not yet known whether the present planetary atmospheres formed while the planets were accreting, or in early catastrophic events, or in a continuous fashion over geological time. It also is not known if the volatiles were uniformly distributed in the accreting material or if most of the volatiles were brought in as a thin veneer in the terminal stages of accretion. In most theories of atmospheric origin, the early atmospheres are thought to have been dissipated, either by solar activity or by violent impacts, and the present atmospheres are secondary, having been formed by outgassing of the interior. This outgassing must be relatively efficient for the Earth since an appreciable fraction of the argon-40 produced by the decay of potassium-40 in the interior resides in the atmosphere. On the other hand, primordial gases such as helium-3 are still being expelled from the Earth's mantle. Estimates of the extent of outgassing of the Earth and the efficiency of crustal formation are in the 30–70 percent range. This refers to secondary processes, not the accretional outgassing and melting. Accretional devolatilization and melting concentrate the volatiles into the atmosphere and such LIL elements as uranium and potassium into the near-surface layers.

The present atmospheres contain only part of the volatile inventory. Earth has a large amount of CO, tied up in organic limestones, and this must be counted as part of *the* prebiotic atmosphere. Most of the water, of course, is presently in the oceans, ice caps and porous near-surface rocks. Mars has appreciable CO₂ and some H₂O in its polar caps. Venus apparently had an appreciable water content as evidenced from its high deuterium-hydrogen abundance ratio (Donahue and others, 1982). When these factors are taken into account, the early atmospheres of Earth and Venus, similar-sized planets, may have been more similar than they are now.

Differences in the atmospheric abundances and compositions between the terrestrial planets may be due to (1) chemistry of the accreting planetesimals, (2) incomplete outgassing of the interior, (3) trapping in surface regions, (4) catastrophic loss of an early atmosphere and (5) gradual escape of the lighter constituents from the top of the atmosphere.

Among the terrestrial planets there is little correlation of volatile content, as inferred from atmospheric composition, and distance from the sun, but there is a tendency for the larger bodies to be better endowed with volatiles than the smaller planets. Gravitational escape may, in part, be responsible for this trend but cannot explain the persistence of the trend to higher molecular weight species. The larger planets may be more efficiently outgassed or may have been able to accrete more of the volatile-rich late condensates. The study of isotopes, particularly of the noble gases, provides important clues.

In the currently available data on the absolute abun-

dances and ratios of constituents of the volatile inventory of Earth, Venus, Mars and the Sun (Table 1-6) are three important trends: (1) The absolute abundances of N_2 and CO_2 are essentially the same for Venus and Earth and much lower for Mars. (2) The absolute abundance of argon-36 and the $^{36}Ar/^{14}N$ ratio decrease by several orders of magnitude from Venus to Earth to Mars, and a similar increase occurs for $^{40}Ar/^{36}Ar$. (3) Ratios such as $^{20}Ne/^{36}Ar$, $^{38}Ar/^{36}Ar$, $^{18}O/^{16}O$ and $^{13}C/^{12}C$ are similar for the three planets. The ratios of primordial rare-gas species are similar to those of chondritic meteorites but differ from solar values.

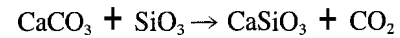
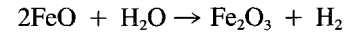
The similarity of the planetary rare-gas ratios and their differences from solar values, and the argon, carbon and nitrogen trends dictate against the trapping of solar nebula or solar wind by the accreting planetesimals, the "primary atmosphere model." The volatile content of the terrestrial planets is best explained if the volatiles are brought in as a late veneer by carbonaceous chondritic material (Anders and Owen, 1977). This implies that some carbon and nitrogen has entered the cores of the terrestrial planets. These elements are important in carbonaceous chondrites but are less so in the atmosphere-crust-mantle system.

The high $^{40}Ar/^{36}Ar$ of the Earth suggests that late outgassing may have been more efficient than on Venus. This is probably a consequence of plate tectonics and continuous overturning of the mantle. The high surface temperature on Venus (about 740 K) may preclude deep subduction and the consequent displacement of a large part of the upper mantle to the near surface where it can be outgassed.

Mars has an $^{40}Ar/^{36}Ar$ ratio 10 times greater than the terrestrial values but on the basis of absolute abundances appears to be less outgassed than the Earth or endowed with much lower primitive abundances of the volatiles. The most likely explanations are that Mars is depleted in the very

volatile elements such as argon-36 or that it lost its early atmosphere. In addition, it is also likely that a small body such as Mars, with little evidence of plate tectonics, is less outgassed than the Earth.

The high $^{15}N/^{14}N$ ratio for Mars indicates that material has been lost from the atmosphere, presumably by thermal escape. Thermal escape is most likely for low atomic weight elements such as hydrogen, helium and neon. There are also various nonthermal mechanisms, such as impact erosion, for the escape of atmospheric material. Reactions such as



between the atmosphere and the surface can also alter the composition and amount of the atmosphere.

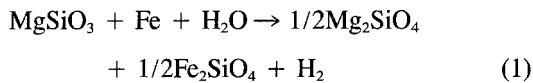
The similarity in nitrogen, argon and CO , abundances between Venus and Earth suggest that both planets experienced a similar degree of outgassing, but the higher $^{40}Ar/^{36}Ar$ of Earth suggests that later outgassing was more efficient for this body. The "missing water" on Venus apparently escaped (Donahue and others, 1982). The lifetime for helium-4 in Venus's atmosphere is about 10^9 years, about 300 times longer than for the Earth. The longer lifetime reflects the higher abundance and the lower escape efficiency for Venus. The present 4He abundance in the atmosphere of Venus is about that expected to have been produced over the past 10^9 years if the uranium and thorium content of Venus is similar to the Earth. The relative average abundances of the argon-40 in the terrestrial and venusian atmospheres suggest that the rate of argon-40 outgassing from Venus is about a factor of four less than from the Earth (Donahue and others, 1982).

TABLE 1-6
Volatile Abundances

	Earth	Venus	Mars	Chondrites	Sun
N_2	3×10^{-6}	4×10^{-6}	10^{-7}	—	1.3×10^{-3}
Ar	10^{-8}	7×10^{-9}	5×10^{-10}	—	10^{-4}
^{40}Ar	10^{-8}	3×10^{-9}	5×10^{-10}	—	—
Ne	10^{-11}	2×10^{-10}	4×10^{-14}	—	1.7×10^{-3}
Kr	3×10^{-12}	10^{-10}	2×10^{-14}	—	9.7×10^{-8}
CO_2	2×10^{-4}	0.95×10^{-4}	$>3.5 \times 10^{-8}$	—	10^{-2}
H_2O	2.8×10^{-4}	$>10^{-6}$	$>5 \times 10^{-6}$	—	10^{-2}
$^{40}Ar/^{36}Ar$	292	1.2	3000	—	—
$^{38}Ar/^{36}Ar$	0.2	0.2	0.2	0.2	—
$^{20}Ne/^{36}Ar$	0.5	0.5	0.5	0.2	31
$^{84}Kr/^{36}Ar$	0.036	0.01	0.03	0.022	0.00027
$^{18}O/^{16}O$	2×10^{-3}	2×10^{-3}	2×10^{-3}	—	—
$^{13}C/^{12}C$.011	0.012	0.011	—	—
$^{14}N/^{15}N$	277	277	165	—	—
DIH	1.6×10^{-4}	1.6×10^{-2}	—	—	—

Donahue and others (1982), Anders and Owen (1977), Pollack and Black (1979).

Nonreactive components of the atmosphere such as the rare gases are presumably incorporated into a planet by being dissolved in crystal interiors and absorbed in grain boundaries and surfaces, or by rapid accretion of a gas-dust mixture. Reactive components such as H₂O and CO₂ are contained in silicates, such as Mg(OH)₂, Mg₃Si₂O₅(OH)₄, and in carbonate minerals. H₂O and CO₂ are released from these minerals by high-velocity impacts. The high temperatures associated with impact, however, can cause such reactions as



that, for chondritic abundances, uses up the available water and allows FeO to be incorporated into the mantle. Free water will only be available at the end of accretion if there is a deficiency of free iron in the late-stage accreted materials (Lange and Ahrens, 1984). Reaction 1 leads to a high FeO content for the mantle unless the core contains appreciable oxygen. Free H₂O in the atmosphere and hydrosphere therefore requires slightly inhomogeneous accretion (more free iron in the early stages of accretion) and a core containing iron and FeO.

Estimates of the accretion time of the Earth range from about 10⁷ to 2.5 × 10⁸ years, but the bulk of the Earth was accreted on a much shorter time scale. For most of the accretion time, the gravitational energy of accretion was sufficient to melt and vaporize infalling material. Much out-

gassing was therefore contemporaneous with accretion. Likewise, the process of core formation was probably synchronous with accretion rather than a later event occurring after the Earth had assembled. Rapid infall of material may have allowed some volatiles to be trapped and buried by subsequent debris-forming events. Rapid accretion and accretion of large, deeply penetrating bodies also causes the early Earth to be a hot body. Since accretional energy increases as the planet grows, it is possible that the early temperature gradient of the mantle was negative.

There are, therefore, several processes that are responsible for the incorporation of volatiles in the interior of a planet: (1) the initial low-energy accretion that allows infalling particles to retain their volatiles, (2) subsequent high-energy accretion that allows a fraction of the incoming volatile inventory to be deeply buried or covered, and (3) late-stage accretion of volatile-rich bodies. Correspondingly, there are several time scales of degassing: (1) the devolatilization of incoming material that is contemporaneous with accretion, (2) the slow and inefficient process of outgassing that involves convection and the cycling of material to the near surface.

The steady decrease in the atmospheric argon-36 abundance, per gram of planet, from Venus to Earth to Mars may represent a chemical gradient in the solar nebula, different degrees of outgassing or atmospheric erosion. The abundance of argon-40, which represents late outgassing from the decay of potassium-40 in the interior, decreases from Earth to Venus to Mars. The factor of 16 difference between

TABLE 1-7
Cosmic Abundances of the Elements (Atoms/10⁶Si)

1	H	2.72 × 10 ¹⁰	24	Cr	1.34 × 10 ⁴	48	Cd	1.69	72	Hf	0.176
2	He	2.18 × 10 ⁹	25	Mn	9510	49	In	0.184	73	Ta	0.0226
3	Li	59.7	26	Fe	9.00 × 10 ⁵	50	Sn	3.82	74	W	0.137
4	Be	0.78	27	Co	2250	51	Sb	0.352	75	Re	0.0507
5	B	24	28	Ni	4.93 × 10 ⁴	52	Te	4.91	76	Os	0.717
6	C	1.21 × 10 ⁷	29	Cu	514	53	I	0.90	77	Ir	0.660
7	N	2.48 × 10 ⁶	30	Zn	1260	54	Xe	4.35	78	Pt	1.37
8	O	2.01 × 10 ⁷	31	Ga	37.8	55	Cs	0.372	79	Au	0.186
9	F	843	32	Ge	118	56	Ba	4.36	80	Hg	0.52
10	Ne	3.76 × 10 ⁶	33	As	6.79	57	La	0.448	81	Tl	0.184
11	Na	5.70 × 10 ⁴	34	Se	62.1	58	Ce	1.16	82	Pb	3.15
12	Mg	1.075 × 10 ⁶	35	Br	11.8	59	Pr	0.174	83	Bi	0.144
13	Al	8.49 × 10 ⁴	36	Kr	45.3	60	Nd	0.836	90	Th	0.0335
14	Si	1.00 × 10 ⁶	37	Rb	7.09	62	Sm	0.261	92	U	0.0090
15	P	1.04 × 10 ⁴	38	Sr	23.8	63	Eu	0.0972			
16	S	5.15 × 10 ⁵	39	Y	4.64	64	Gd	0.331			
17	Cl	5240	40	Zr	10.7	65	Tb	0.0589			
18	Ar	1.04 × 10 ⁵	41	Nb	0.71	66	Dy	0.398			
19	K	3770	42	Mo	2.52	67	Ho	0.0875			
20	Ca	6.11 × 10 ⁴	44	Ru	1.86	68	Er	0.253			
21	Sc	33.8	45	Rh	0.344	69	Tm	0.0386			
22	Ti	2400	46	Pd	1.39	70	Yb	0.243			
23	V	295	47	Ag	0.529	71	Lu	0.0369			

Anders and Ebihara (1982).

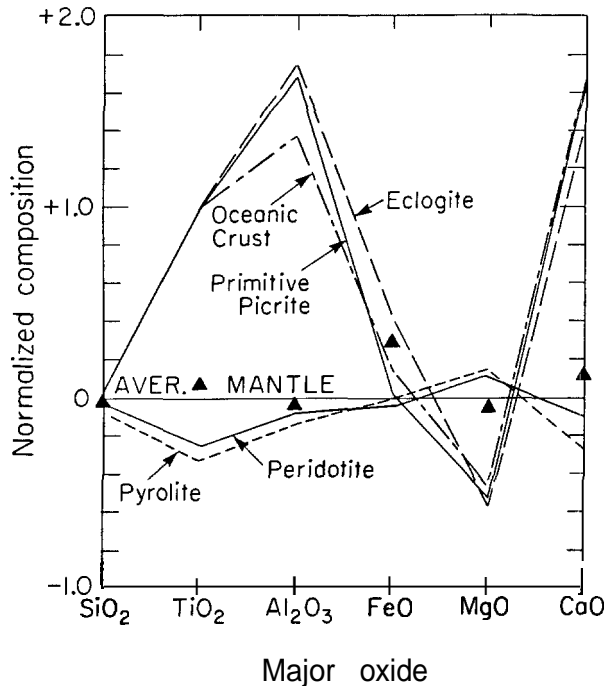


FIGURE 1-2
Bulk chemistry of ultramafic rocks (peridotite) and basic, or basaltic, rocks (oceanic crust, picrite, eclogite) normalized to average mantle composition based on cosmochemical considerations and an assumption about the FeO content of the mantle. Pyrolite is a hypothetical rock proposed by some to be representative of the whole mantle. A composition equivalent to 80 percent peridotite and 20 percent eclogite (or basalt), shown by triangles, is a mix that reconciles petrological and cosmochemical major-element data. Allowance for trace-element data and a possible MgSiO_3 -rich lower mantle reduces the allowable basaltic component to 15 weight percent or less.

Earth and Mars probably represents a lower volatile content, including potassium, for Mars and the lesser outgassing expected for a small, relatively cold planet.

COMPOSITION OF THE TERRESTRIAL PLANETS

It is now generally believed that with the exception of a few elements such as Li, Be and B, the composition of the solar atmosphere is essentially equal to the composition of the material out of which the solar system formed. (This ignores the possibility that the Sun is a chemically zoned object.) The planets are assumed to accrete from material that condensed from a cooling primitive solar nebula. Various attempts have been made to compile tables of "cosmic" abundances. The Sun contains most of the mass of the solar system; therefore, when we speak of the elemental abundances in the solar system, we really refer to those in the Sun, assuming that the abundances in the Sun, its surface,

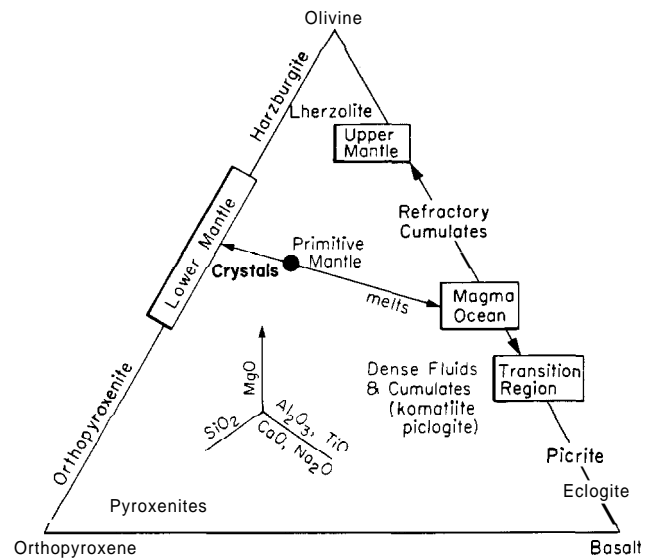


FIGURE 1-3
Representation of mantle components in terms of olivine and orthopyroxene (the high melting point minerals) and basalt (the most easily fusible component). Primitive mantle is based on cosmic abundances. Melting of chondritic material at high temperature gives an MgO-rich melt (basalt + olivine) and a dense refractory residual (olivine + orthopyroxene). Crystallization of a magma ocean separates clinopyroxene and garnet from olivine. Melting during accretion tends to separate components according to density and melting temperature, giving a chemically zoned planet. If melting and melt-crystal separation occur primarily at low pressure, the upper mantle will be enriched in basalt, olivine and the incompatible elements relative to the lower mantle and relative to the chondritic starting material.

and in the primitive solar nebula are the same. The spectroscopic analyses of elemental abundances in the solar photosphere do not have as great an accuracy as chemical analyses of solid materials. CI carbonaceous chondrite meteorites, which appear to be the most representative samples of the relatively nonvolatile constituents of the solar system, are used for compilations of the abundances of most of the elements (Table 1-7). For the very abundant volatile elements, solar abundance values are used.

The very light and volatile elements (such as H, He, C and N) are extremely depleted in the Earth relative to the Sun or carbonaceous chondrites. Moderately volatile elements (such as K, Na, Rb, Cs and S) are moderately depleted in the Earth. Refractory elements (such as Ca, Al, Sr, Ti, Ba, U and Th) are generally assumed to be retained by the planets in their cosmic ratios. It is also likely that magnesium and silicon occur in a planet in chondritic or cosmic ratios with the more refractory elements. The Mg/Si ratio, however, varies somewhat among meteorite classes. Sometimes it is assumed that magnesium, iron and silicon may be fractionated by accretional or pre-accretional

processes, but these effects, if they exist, are slight. The upper mantle is olivine-rich and has a high Mg/Si ratio compared to the cosmic ratio (Figures 1-2 and 1-3). If the Earth is chondritic in major-element chemistry, then the deeper mantle must be rich in pyroxene.

It is not clear that the planets accreted homogeneously from material of uniform composition. The mean uncompressed densities of the terrestrial planets decreases in the order Mercury, Earth, Venus, Mars, Moon (Figure 1-4 and Table 1-8). The high density of Mercury probably means a high iron content, and the reverse is implied for the Moon. The other planets may have a variable iron content or may differ in the oxidation state of the iron, that is, the FeO/Fe ratio. Venus and Earth are so close in mass and mean density that they may have nearly identical major-element chemistries. The high surface temperature of Venus decreases the density of near-surface materials, and this along with the low pressures depress the depths of phase changes such as basalt-eclogite, making Venus overall a less dense body. The uncompressed density of Mars is less than that of Earth and Venus, and it therefore differs in composition. The outer planets and satellites are much more volatile-rich than the inner planets. Meteorites also vary substantially in composition. The above considerations suggest that there may be an element of inhomogeneity in the accretion of the planets, perhaps caused by temperature and pressure gradients in the early solar nebula. Early forming planetesimals would have been refractory- and iron-rich and the later forming planetesimals more volatile-rich. If planetary accretion was occurring simultaneously with cooling and condensation, then the planets would have formed inhomogeneously. As a planet grows, the gravitational energy of accretion increases, and impact vaporization becomes more

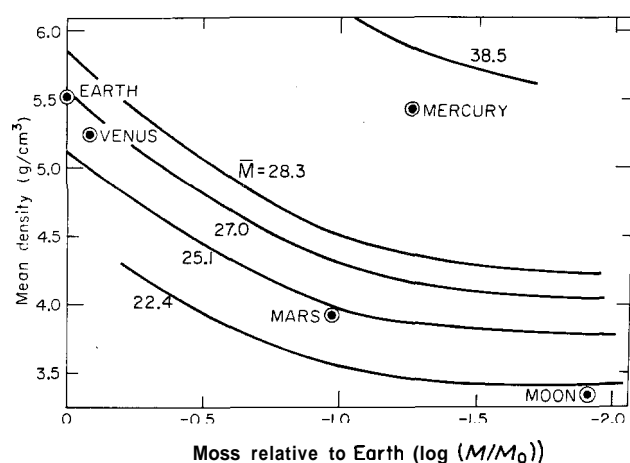


FIGURE 1-4
Mean density versus mass, relative to Earth, of planets having the same structure as the Earth and various metal/silicate ratios, expressed as \bar{M} , mean atomic weight. Earth and Venus have similar bulk chemistries while Mars and Moon are clearly deficient in iron. Mercury is enriched in iron.

important for the larger planets and for the later stages of accretion. The assumption that Earth has cosmic abundances of the elements is therefore only a first approximation but is likely to be fairly accurate for the involatile elements. There is little dispersion of the refractory elements among the various stony meteorite classes, suggesting that these elements are not appreciably fractionated by pre-accretional processes. Fortunately, the bulk of a terrestrial planet is iron, magnesium, silicon, calcium, and aluminum and their oxides. The bulk composition of a terrestrial planet can therefore be discussed with some confidence.

During large solar flares a sample of the corona can be accelerated to high energies. Solar energetic particles can be used to infer the composition of the corona and photosphere (Breneman and Stone, 1985). Some recent results are given in Table 1-9. The iron abundance is about 40 percent higher than previous estimates of cosmic or solar abundances. This agrees with recent spectroscopic values for the photosphere (Grevasse, 1984). The refractory elements calcium and titanium are also more abundant than chondritic values. A terrestrial planet formed from the solar nebula can therefore have much more free iron, FeO and diopside ($\text{CaMgSi}_2\text{O}_6$) than previously supposed. With these new values Mercury is still enriched in iron, and Mars and Moon are depleted in iron compared to the Sun.

It is usually assumed that the Sun and planets formed more or less contemporaneously from a common mass of interstellar dust and gas. There is a close similarity in the relative abundances of the condensable elements in the atmosphere of the Sun, in chondritic meteorites and in the Earth. To a first approximation one can assume that the planets incorporated the condensable elements in the proportions observed in the Sun and the chondrites. On the other hand, the differences in the mean densities of the planets, corrected for differences in pressure, show that they cannot all be composed of materials having exactly the same composition. Variations in iron content and oxidation state of iron can cause large density variations among the terrestrial planets. The giant, or Jovian planets, must contain much larger proportions of low-atomic-weight elements than Mercury, Venus, Earth, Moon and Mars. The condensation behavior of the elements is given in Figure 1-5.

The equilibrium assemblage of solid compounds that exists in a system of solar composition depends on temperature and pressure and, therefore, with location and time. At a nominal nebular pressure of 10^{-3} atm, the material would be a vapor at temperatures greater than about 1900 K (Larimer, 1967; Grossman, 1972). The first solids to condense at Power 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 1-10 and Figure 1-6). 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

TABLE 1-8
Properties of the Terrestrial Planets

	GM $10^{18} \text{cm}^3/\text{s}^2$	R km	P 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.

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.

There are several ways in which interactions between the gaseous nebula and solid condensate particles might have controlled the composition of the planets. In one extreme, all or most of the material joining a planet may have equilibrated in a relatively narrow range of temperatures peculiar to that planet. In another extreme, one mineral af-

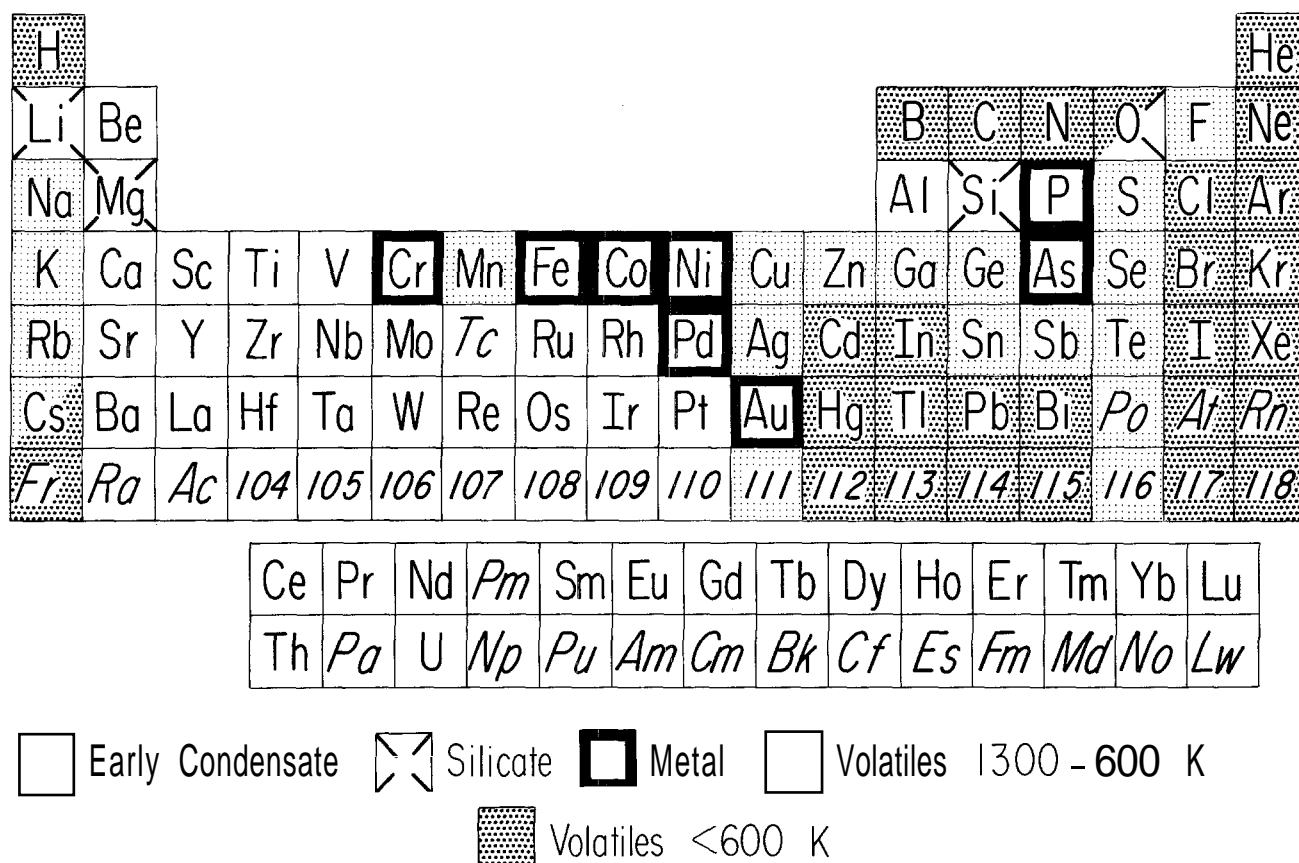


FIGURE 1-5
Condensation behavior of the elements. Short-lived radioactive elements are shown in italics (after Morgan and Anders, 1980).

TABLE 1-9
Solar and "Cosmic" Abundances in Atoms/1000 Si Atoms

Z	Element	Corona (1)	Photosphere (1)	"Cosmic" (2)
6	C	2350	6490	12,100
7	N	700	2775	2480
8	O	5680	22,900	20,100
9	F	0.28	1.1	0.843
10	Ne	783	3140	3760
11	Na	67.0	67.0	57.0
12	Mg	1089	1089	1075
13	Al	83.7	83.7	84.9
14	Si	1000	1000	1000
15	P	4.89	9.24	10.4
16	S	242	460	515
17	Cl	2.38	9.6	5.24
18	Ar	24.1	102	104
19	K	3.9	3.9	3.77
20	Ca	82	82	61.1
21	Sc	0.31	0.31	0.034
22	Ti	4.9	4.9	2.4
23	V	0.48	0.48	0.295
24	Cr	18.3	18.3	13.4
25	Mn	6.8	6.8	9.51
26	Fe	1270	1270	900
27	Co	<18.1	<18.1	2.25
28	Ni	46.5	46.5	49.3
29	Cu	0.57	0.57	0.514
30	Zn	1.61	1.61	1.26

(1) Breneman and Stone (1985).

(2) Anders and Ebihara (1982).

ter another condenses, as required to maintain thermodynamic equilibrium in the nebula, and immediately accretes into a planet. At some point the process is interrupted by dissipation of the nebula. Differences in the mean compositions of the planets would result when the nebula, with any remaining uncondensed elements, was removed. If temperatures declined outward in the nebula at the time when condensation ended, it could account qualitatively for the density differences in the planets. Mercury would have accreted mainly the calcium and aluminum silicates and metallic iron; planets farther out in the solar system would have condensed and accreted increasingly larger proportions of low-density silicates and volatiles. This process, described earlier in this chapter, is termed heterogeneous accretion (Turekian and Clark, 1969; Anderson, 1972a,b; Clark and others, 1972). The planets formed would be layered from the outset, having the highest temperature condensates at their centers and successively lower temperature condensates closer to their surfaces. Thus, planets would contain substantial metallic cores mainly as a result of accretion rather than subsequent interior melting and differentiation. At the early stages of accretion, it might be easier to accrete and retain material on a ductile iron nucleus than on a brittle silicate nucleus.

The chondritic meteorites may be samples of relatively primitive planetary material. Many of them consist of several components that have experienced very different temperature histories (Wood, 1962), which are mingled on the scale of a millimeter. Observed differences between the bulk compositions of the several chondrite subclasses can be explained as resulting from (1) processes that physically fractionated the various components from one another before they accreted into the parent chondrite planets and (2) differing nebula temperatures of accretion for the chondrite subclasses, which had the effect of excluding elements that required temperatures lower than the accretion temperature to condense. For example, siderophiles, which behave coherently during magmatic processes in a planet, might be fractionated in the nebula because they span a large range of condensation temperatures. Nebular fractionation processes include (1) gain or loss of early condensate by settling to the median plane of the nebula and (2) gain or loss of metal relative to silicate, by ferromagnetic or gravitational effects, and remelting of the primary condensate. Some typical compositions of all these components are listed in Table 1-11.

Various authors have suggested that the same physical components and fractionating processes that occurred in the chondrite subclasses are responsible for differences in the compositions and mean densities of the terrestrial planets (Mason, 1962; Morgan and Anders, 1980). Some of the differences among meteorite classes may be due to magmatic processes that occurred inside the parent body.

There are some constraints on the amounts or ratios of a number of key elements in a planet. For example, the mean density of a planet, or the size of the core, constrains the iron content. Using cosmic ratios of elements of similar geochemical properties (say Co, Ni, refractory siderophiles), a whole group of elements can be constrained. The uranium and thorium content are constrained by the heat flow and thermal history calculations. The K/U ratio, roughly constant in terrestrial magmas, is a common constraint in this kind of modeling. The Pb/U ratio can be estimated from lead isotope data. The amount of argon-40 in the atmosphere provides a lower bound on the amount of potassium in the crust and mantle. (This and some other rare-gas species are listed in Table 1-12.) Most of these are very weak constraints, but they do allow rough estimates to be made of the refractory, siderophile, volatile and other contents of the Earth and terrestrial planets. The elements that are correlated in magmatic processes have very similar patterns of geochemical behavior, even though they may be strongly fractionated during nebular condensation. Thus, some abundance patterns established during condensation tend not to be disturbed by subsequent planetary melting and igneous fractionation. On the other hand, some elements are so strongly fractionated from one another by magmatic and core formation processes that discovering a "cosmic" or "chondritic" pattern can constrain the nature of these processes.

TABLE 1-10
Approximate Sequence of Condensation of Phases and Elements
from a Gas of Solar Composition at 10^{-3} atm Total Pressure

Phase	Formula	Temperature
Hibonite	$\text{CaAl}_{12}\text{O}_{19}$	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	$\text{Ca}_2\text{Al}_2\text{SiO}_7$ $\text{Ca}_2\text{Mg}_2\text{Si}_2\text{O}_7$ Co	1625 K
Spinel	MgAl_2O_4 Al_2SiO_5	1513 K
Metallic iron	Fe, Ni	1473 K
Diopside	$\text{CaMgSi}_2\text{O}_6$	1450 K
Forsterite	Mg_2SiO_4	1444 K
Anorthite	$\text{CaAl}_2\text{Si}_2\text{O}_8$ Ca_2SiO_4 CaSiO_3	1362 K
Enstatite	MgSiO_3 Cr_7O_3 P, Au, Li MnSiO_3 MnS, Ag As, Cu, Ge	1349 K
Feldspar	$(\text{Na},\text{K})\text{AlSi}_3\text{O}_8$ Ag, Sb, F, Ge Sn, Zn, Se, Te, Cd	
Reaction products	$(\text{Mg},\text{Fe})_2\text{SiO}_4$ $(\text{Mg},\text{Fe})\text{SiO}_3$	1000 K
Troilite, pentlandite	FeS, (Fe, Ni)S Pb, Bi, In, Tl	700 K
Magnetite	Fe_3O_4	405 K
Hydrous minerals	$\text{Mg}_3\text{Si}_2\text{O}_7 \cdot 2\text{H}_2\text{O}$, 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).

There is strong 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 (Morgan and Anders, 1980). 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 as one component. From the observed abundance ratios of correlated elements in samples from the Moon, Earth and achondrites, there is excellent proof that these elements are present in the same ratios as in CI chondrites. The abundance of the refractory elements in a given planet can be calculated from the inferred abundance of their heat-producing members uranium and thorium, provided reliable data and interpretations of the global heat flux are available (Ganapathy and Anders, 1974). This assumes

that surface heat flow represents the current rate of heat production. A large fraction of the present heat flow, however, may be 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.

All of these complexities make planetary models somewhat variable. Table 1-13 lists typical model compositions for the terrestrial planets as well as one for the eucrite parent body. These and other models will be considered below in more detail.

Venus

Venus is 320 km smaller in radius than the Earth and is about 4.9 percent less dense. Most of the difference in density is due to the lower pressure, giving a smaller amount of self-compression and deeper phase changes. Venus is a much smoother planet than the Earth but has a measurable triaxiality of figure and a 0.34 km offset of the center of the figure from the center of mass. This offset is much smaller than those of the Moon (2 km), Mars (2.5 km) and Earth (2.1 km). The moment of inertia of Venus is not known.

In contrast to the bimodal distribution of Earth's topography, representing continent-ocean differences, Venus has a narrow unimodal height distribution with 60 percent of the surface lying within 500 m of the mean elevation. This difference is probably related to erosion and isostatic differences caused by the presence of an ocean on Earth. For both Earth and Venus the topography is dominated by long-wavelength features. Most of the surface of Venus is gently rolling terrain. The gravity and topography are positively correlated at all wavelengths. On Earth most of the long-wavelength geoid is uncorrelated with surface topography and is due to deep mantle dynamics.

The other respects in which Venus differs markedly from the Earth are its slow rotation rate, the absence of a satellite, the virtual absence of a magnetic field, the low abundance of water in the atmosphere and at the surface, the abundance of primordial argon, the high surface temperature and the lack of obvious signs of subduction (Phillips and others, 1980; Anderson, 1980, 1981; Head and Solomon, 1981). There is some evidence for spreading, ridge-like features and compressional features.

The moderate differences of the mass, mean density and solar distance of Venus from those of the Earth make it appropriate to discuss its bulk constitution (Table 1-14) in terms of differences from the Earth.

If Venus had an identical bulk composition and structure to the Earth (see Figure 1-4), then its mean density

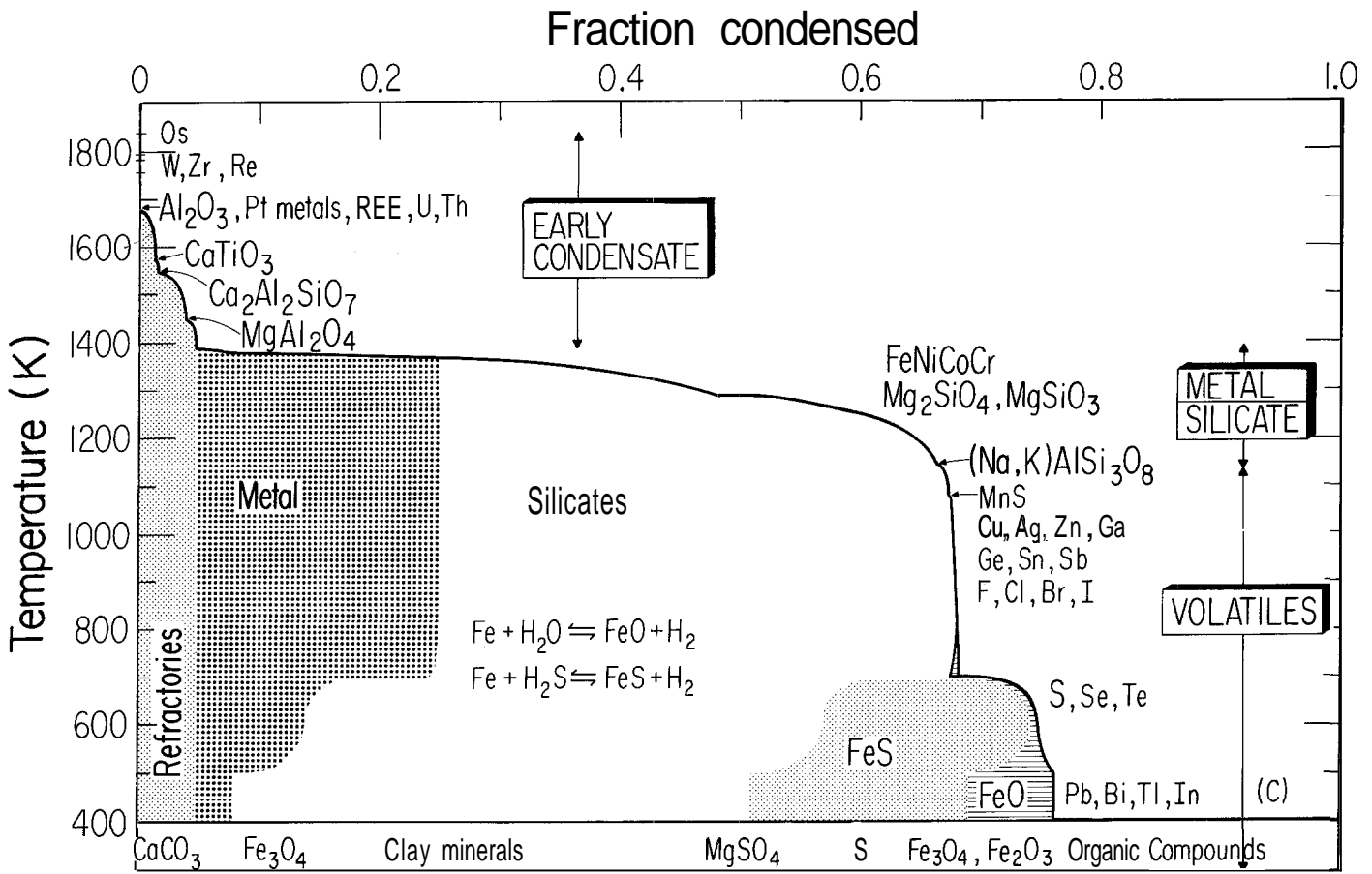


FIGURE 1-6 Condensation of a solar gas at 10^{-4} atm (after Morgan and Anders, 1980).

would be about 5.34 g/cm^3 . By "identical structure" I mean that (1) most of the iron is in the core, (2) the crust is about 0.4 percent of the total mass and (3) the deep temperature gradient is maintained by convection. The high surface temperature of Venus, about 740 K, would have several effects; it would reduce the depth at which the convectively controlled gradient is attained, it would deepen temperature-sensitive phase changes and it may prevent mantle cooling by subduction.

The density of Venus is 1.2 to 1.9 percent less than that of the Earth after correcting for the difference in pressure. This has been attributed to differences in iron content, sulfur content, oxidation state of the mantle and deepening of the basalt-eclogite phase change. Most of the original basaltic crust of the Earth subducted when the upper mantle temperatures cooled into the eclogite stability field. The density difference between basalt and eclogite is about 15 percent. Because of the high surface temperature on Venus, the upper-mantle temperatures are likely to be 200–400 K hotter in the outer 300 km or so than at equivalent depths on Earth, or melting is more extensive.

Since the deep interior temperatures of the two planets are likely to be similar, the near-surface thermal gradient is

controlled by the thickness, δ , of the conductive thermal boundary layer. This is related to the Rayleigh number, Ra , and the thickness, D , of the underlying convecting layer by

$$\delta \approx Ra^{-1/3} D$$

where

$$Ra = \frac{\alpha g \Delta T D^3}{\kappa \nu},$$

which is written in terms of the gravitational acceleration g , temperature rise across the layer ΔT , the thermal expansion and diffusivity α and κ , and the kinematic viscosity ν . The largest difference between Earth and Venus is likely to be the viscosity since this depends exponentially on temperature and pressure. The higher temperatures in the outer layer of Venus and the lower pressure throughout serve to decrease the average viscosity of the Venus mantle by at least several orders of magnitude. The other parameters in the Rayleigh number are only weakly dependent on temperature or on the size of the planet. Because of the thinner boundary layer, the near-surface thermal gradient is steeper in Venus than the Earth. The net result is that the deep mantle adiabat is brought closer to the surface. This has inter-

TABLE 1-11
Compositions of Possible Components of the Terrestrial Planets
(Percent or ppm)

Species	C1	EC	HTC
SiO ₂	30.9	39.1	20.2
TiO ₂	0.11	0.06	1.9
Al ₂ O ₃	2.4	1.9	36.5
Cr ₂ O ₃	0.38	0.35	—
MgO	20.8	21.3	7.1
FeO	32.5	1.7	—
MnO	0.25	0.14	—
CaO	2.0	1.6	34.1
Na ₂ O	1.0	1.0	—
K (ppm)	800	920	—
U (ppm)	0.013	0.009	0.19
Th (ppm)	0.059	0.034	0.90
Fe	0	26.7	—
Ni	1.3	1.7	—
S	8.3	4.5	—

C1: Average C1 carbonaceous chondrite, on a C-, H₂O-free basis (Wood, 1962).

EC: Average enstatite chondrite (Wood, 1962).

HTC: High-temperature condensate (Grossman, 1972).

esting implications for the phase relations in the upper mantle and the evolution of the planet. In particular, partial melting in the upper mantle of Venus is probably much more extensive than is the case for the Earth. Crust can be much thicker because of the deepening of the basalt-eclogite phase boundary.

Schematic geotherms are shown in Figure 1-7 for surface temperatures appropriate for Earth and Venus. With the phase diagram shown, the high-temperature geotherm crosses the solidus at about 85 km. With other phase relations the eclogite field is entered at a depth of about 138 km. For Venus, the lower gravity and outer layer densities increase these depths by about 20 percent; thus, we expect a surface layer of 100 to 170 km thickness on Venus composed of basalt and partial melt. On the present Earth, the eclogite stability field is entered at a depth of 40–60 km.

A large amount of basalt has been produced by the Earth's mantle, but only a thin veneer is at the surface at any given time. There must therefore be a substantial amount of eclogite in the mantle, the equivalent of about 200 km in thickness. If this were still at the surface as basalt, the Earth would be several percent less dense. Correcting for the difference in temperature, surface gravity and mass and assuming that Venus is as well differentiated as Earth, only a fraction of the basalt in Venus would have converted to eclogite. This would make the uncompressed density of Venus about 1.5 percent less than Earth's without invoking any differences in composition or oxidation state. Thus, Venus may be close to Earth in composition. It is possible that the present tectonic style on Venus is similar to that of Earth in the Archean, when temperatures and temperature gradients were higher.

The high degree of correlation between gravity and elevation on Venus might suggest that surface loads are supported by a thick, strong lithosphere. Because of the high surface temperature, this is unlikely. In the other extreme, in a purely viscous, convecting planet, such a correlation is expected if hot upwellings deform the surface upward to an extent that more than compensates the lower density. This, in general, is the expected situation. A thick buoyant crust would also give this effect if isostasy prevails. The relationship between elevation and geoid, or gravity, gives the depth of compensation. High topography may also be maintained dynamically by horizontal compression, but this is expected to be a transient situation.

The depth of compensation on Venus has been estimated to be 115 ± 30 km (Phillips and others, 1980). A thick buoyant crust on Venus is therefore a distinct possibility. If the depth of compensation corresponds to the crustal thickness, then it would be six times the average crustal thickness on Earth and comparable in thickness to the seismic lithosphere under continental shields, which has been interpreted as olivine-rich residual material that is lighter than "normal" fertile (high garnet content) mantle. The amount of implied crust for Venus is not unreasonable, considering the amount of CaO, Al₂O₃, Na₂O and so on that is

TABLE 1-12
Rare-gas Isotopes in Planetary Atmospheres and Chondrites

	³⁶ Ar (10 ⁻¹⁰ cm ³ /g)	⁴⁰ Ar (10 ⁻⁸ cm ³ /g)	⁴⁰ Ar/ ³⁶ Ar	¹²⁹ Xe (10 ⁻¹² cm ³ /g)	¹²⁹ Xe/ ¹³² Xe
Mars	1.6	48	3000	1.12	1.49
Earth	210	612	291	1.04	0.067
Venus	21,000	225	1.07	—	—
Chondrites					
C1	7700	476	6.2	380	0.05
C3V	3440	1500	44	780	0.26
E4	3310	8606	260	3890	4.05

Mazor and others (1970), Anders and Owen (1977), Morgan and Anders (1980), Wacker and Marti (1983), von Zahn and others (1983).

TABLE 1-13
 Representative Model Compositions of Terrestrial Planets and Other Bodies
 Based Mainly on Cosmochemical Considerations (Weight Percent and ppm)

Species	Mercury	Venus	Earth	Mars	Moon	EPB
		Mantle + Crust				
SiO ₂	43.8	53.9	48.0	40.0	45.6	46.0
TiO ₂	—	0.20	0.27	0.1	0.2	0.15
Al ₂ O ₃	4.4	3.9	5.2	3.1	4.6	3.1
Cr ₂ O ₃	—	—	1.1	0.6	0.4	0.8
MgO	47.7	38.3	34.3	27.4	32.4	32.5
FeO	0	2.1	7.9	24.3	13.0	14.4
MnO	—	—	0.12	0.2	0.18	0.4
CaO	4.1	3.6	4.2	2.5	3.8	2.5
Na ₂ O	—	1.5	0.33	0.8	0.06	0.07
H ₂ O	—	—	0.11	0.9	—	—
K (ppm)	—	1318	262	573	60	62
U (ppm)	0.024	0.021	0.028	0.017	0.027	0.018
Th (ppm)	0.11	0.096	0.100	0.077	—	0.07
		Core				
Fe	94.5	84.7	89.2	72.0	—	—
Ni	5.5	5.2	5.7	9.3	—	—
S	0	10.0"	5.1 *	18.6*	—	—
		Relative Masses				
Mantle + Crust	33.0	69.1	63.9	88.1	—	—
Core	67.0	30.9	36.1	11.9	—	—

Mercury: Cosmic Al:Mg ratio, sufficient SiO₂ for anorthite, diopside, and forsterite phases in mantle, *core/mantle* mass ratio satisfying mean density (BVP, 1980).

Venus: Equilibrium condensation modified by use of feeding zones (Weidenschilling, 1976; BVP, 1980). Venus can be identical in composition to the Earth (Anderson, 1980, 1981).

Earth: Ganapathy and Anders (1974).

Mars: Anderson (1972a).

Moon: Wänke and others (1977).

Eucrite Parent Body: Dreibus and others (1977), Dreibus and Wänke (1979).

*A fraction of the cosmic complement of sulfur is assumed to be in the core.

likely to be incorporated into planetary interiors and considering the relative thickness of the martian and lunar crusts. It is Earth that is anomalous in total crustal thickness, and this can be explained by crustal recycling and the shallowness of the basalt-eclogite boundary in the Earth. Most of the Earth's "crust" probably resides in the transition region of the mantle. Estimates of bulk Earth chemistry can yield a basaltic layer of about 10 percent of the mass of the mantle.

There is little evidence for Earth-style plate tectonics on Venus, but this does not mean that Venus is tectonically dead. If the surface is choked with thick buoyant crust, the manifestation of mantle convection would be quite different than on Earth, where oceanic lithosphere can subduct to make room for the new lithosphere formed at midocean ridges. The rifts and highlands on Venus may be recent transient features, and the surface may be constantly reorganizing itself more on the style of pack ice in the polar oceans. If the crust and lithosphere on Venus is buoyant, the cooling effect of subduction is precluded, and the venerian mantle may be hotter than the Earth's mantle.

The very slow retrograde spin of Venus and its great abundance of primordial argon might be explained by the impact of a major body (Cameron, 1982). The slow rotation rate may also be at least partly responsible for the small magnetic field, which is smaller than either Earth's or Mars's. The magnetic dipole moment of Venus is at least four orders of magnitude less than Earth's. Other factors that might be involved in the small magnetic field of Venus include the roughness of the core-mantle boundary, the temperature of the core and the absence of chemical sedimentation in the core. It is also remotely possible that the field is temporarily low, as occurs on Earth when the field is reversing polarity.

The Soviet gamma-ray data from the surface of Venus (Table 1-15) are consistent with those expected for basaltic rocks.

Mars

Mars is about one-tenth of the mass of Earth. The uncompressed density is substantially lower than that of Earth or

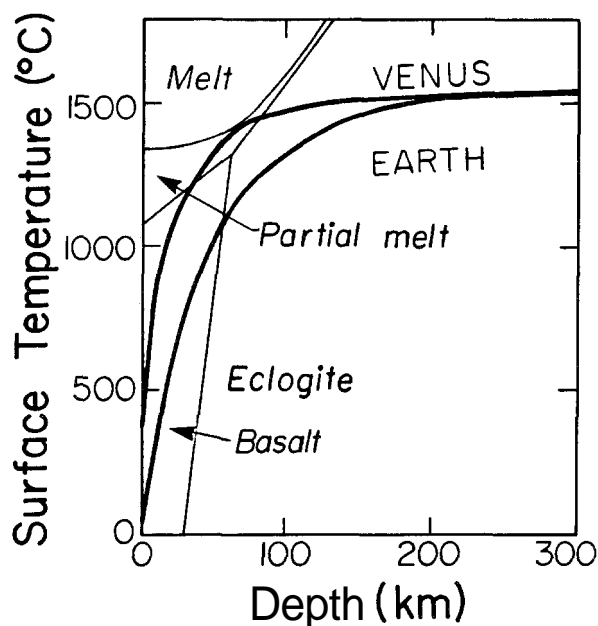


FIGURE 1-7
Schematic geotherms for the Earth with different surface temperatures. Note that the eclogite stability field is deeper for the higher geotherms and that a partial melt field intervenes between the basaltic crust and the rest of the upper mantle. Basaltic material in the eclogite field will probably sink through the upper mantle and be replaced by peridotite. Shallow subduction of basaltic crust leads to remelting in the case of Venus and the early Earth but conversion to eclogite and deep subduction for the present Earth. The depth scale is for an Earth-size planet with the colder geotherm and present crust and upper-mantle densities. For Venus, with smaller g , higher temperatures and low-density crust replacing part of the upper mantle, the depths are increased by about 20 percent.

Venus and is very similar to the inferred density of a fully oxidized (less C and H_2O) chondritic meteorite. The moment of inertia, however, requires an increase in density with depth over and above that due to self-compression and phase changes, indicating the presence of a small core. This in turn indicates that Mars is a differentiated planet.

Mars has long been known as the red planet. Its soils are apparently rich in iron oxides, possibly formed by weathering of an FeO-rich basalt. Models of the mantle of Mars are also rich in FeO.

The tenuous atmosphere of Mars suggests that it either is more depleted in volatiles or has experienced less outgassing than Earth or Venus. It could also have lost much of its early atmosphere by large impacts. Geological evidence for running water on the surface of Mars suggests that a large amount of water is tied up in permafrost and ground water as well as in the polar caps. The northern polar cap of Mars is mainly solid CO_2 , while water ice is an important component of the "permanent" southern polar cap.

The high $^{40}Ar/^{36}Ar$ ratio on Mars, 10 times the terrestrial value, suggests either a high potassium-40 content plus

efficient outgassing, or a net depletion of argon-36 and, possibly, other volatiles. Early outgassed argon-36 could also have been removed from the planet.

The SNC meteorites, described earlier, have trapped rare-gas and nitrogen contents that differ from other meteorites but closely match those in the martian atmosphere (Bogard and Johnson, 1983; Becker and Pepin, 1984). The discovery of meteorites in Antarctica that appear to have come from the Moon increases the possibility that impacts on other planets can launch fragments that eventually land on the Earth. If SNC meteorites do come from Mars (Dreibus and Wänke, 1985), then a relatively volatile-rich planet is implied, and the atmospheric evidence for a low volatile content for Mars would have to be rationalized by the loss of the early accretional atmosphere. Mars, of course, is more susceptible to atmospheric escape than Venus or Earth owing to its low gravity. A model for Mars based on SNC composition is compared with other models in Table 1-16.

The surface of Mars appears to be basaltic, and the large volcanoes on Mars are similar in form to shield volcanoes on Earth. The composition of the soil is consistent with weathering from basaltic parent materials (Table 1-17).

The topography and gravity field of Mars indicate that parts of Mars are grossly out of hydrostatic equilibrium and that the crust is highly variable in thickness. If variations in the gravity field are attributed to variations in crustal thickness, with a constant density ratio between crust and mantle, then reasonable values of the density contrast imply that

TABLE 1-14
Cosmochemical Model Compositions of Venus

Species	V1	v 3
Mantle Plus Crust		
SiO_2	52.9	49.8
TiO_2	0.20	0.21
Al_2O_3	3.8	4.1
Cr_2O_3	—	0.87
MgO	37.6	35.5
FeO	0.24	5.4
MnO	—	—
CaO	3.6	3.3
Na_2O	1.6	0.28
H_2O	0	0.22
K (ppm)	1442	221
U (ppm)	0.020	0.022
Th (ppm)	0.094	0.079
Core		
Fe	94.4	88.6
Ni	5.6	5.5
S	0	5.1
Relative Masses		
Mantle + Crust	69.8	68.0
Core	30.2	32.0

V1: Equilibrium condensation (BVP, 1980)

V3: Morgan and Anders (1980).

the average crustal thickness is at least 30 km (Bills and Ferrari, 1978). This minimal bound is based on the assumption of zero crustal thickness in the Hellas basin. An impact large enough to excavate the Hellas basin would easily remove a 30-km-thick crustal layer. This minimal average crustal thickness on Mars gives a crust/planet mass ratio that is more than five times the terrestrial value (0.4 percent), indicating a well-differentiated planet.

The crust of the Earth is enriched in CaO, Al₂O₃, K₂O, and Na₂O in comparison to the mantle, and ionic radii considerations and experimental petrological results (Chapter 16) suggest that the crust of any planet will be enriched in these constituents. A minimal average crustal thickness for a fully differentiated chondritic planet can be obtained by removing all of the CaO possible, with the available Al₂O₃, as anorthite to the surface. This operation gives a crustal thickness of about 100 km for Mars. Incomplete differentia-

tion and retention of CaO and Al₂O₃ in the mantle will reduce this value, which is likely to be the absolute upper bound. (Earth's crust is much thinner due to crustal recycling and the basalt-eclogite phase change.)

The average thickness of the crust of the Earth is 15 km, which amounts to 0.4 percent of the mass of the Earth. The crustal thickness is 5–10 km under oceans and 30–50 km under older continental shields. The situation on the Earth is complicated, since new crust is constantly being created at the midoceanic ridges and consumed at island arcs. It is probable that some of the crust is being recycled. If the present rate of crustal genesis was constant over the age of the Earth and none of the crust was recycled, then 17 percent of the Earth would be crustal material.

The Moon apparently has a mean crustal thickness greater than that of the Earth. If the average composition of the Moon is similar to chondrites minus the Fe-Ni-S, then the crust could be as thick as an average of 62 km. This is about the thickness of the crust determined by seismic experiments on the nearside of the Moon but less than the average thickness inferred from the gravity field.

Atmospheric analyses suggest that Mars is a less outgassed planet than Earth, but the high ratio of argon-40 relative to the other inert gases suggests that the martian crust is enriched in potassium-40 in relation to the Earth's crust. The only direct evidence concerning the internal structure of Mars is the mean density, moment of inertia, topography, and gravity field. It is possible to calculate models of the martian interior with plausible chemical models and temperature profiles that satisfy these few constraints; however, the process is highly nonunique (Anderson and others, 1971; Anderson, 1972a, 1977; BVP, 1980).

The mean density of Mars, corrected for pressure, is less than that of Earth, Venus and Mercury but greater than that of the Moon. This implies either that Mars has a small total Fe-Ni content, in keeping with its density, or that the FeO/Fe ratio varies. Plausible models for Mars can be constructed that have solar or chondritic values for iron, if most or all of it is taken to be oxidized (Anderson, 1972a).

With such broad chemical constraints, mean density and moment of inertia and under the assumption of a differentiated planet, it is possible to trade off the size and density of the core and density of the mantle. Most models favor an FeO-enrichment of the martian mantle relative to the mantle of the Earth. Anderson (1972a) concluded that Mars has a total iron content of about 25 weight percent, which is significantly less than the iron content of Earth, Mercury or Venus but is close to the total iron content of ordinary and carbonaceous chondrites. The high zero-pressure density of the mantle suggests a relatively high FeO content in the silicates of the martian mantle. The radius of the core can range from as small as one-third the martian radius for an iron core, or a core similar in composition to the Earth's core, to more than half the radius of the planet if it is pure FeS. With chondritic abundances of Fe-FeS, the size of the

TABLE 1-15
Surface Composition of Venus

	Venera 8	Venera 9	Venera 10
K, percent	4.0 ± 1.2	0.47 ± 0.08	0.30 ± 0.16
U, ppm	2.2 ± 0.7	0.60 ± 0.16	0.46 ± 0.26
Th, ppm	6.5 ± 0.2	3.65 ± 0.42	0.70 ± 0.34
K/U	20,000	8000	7000
Th/U	3.0	6.1	1.5

Surkov (1977)

TABLE 1-16
Model Compositions of Silicate Portions of Mars and Earth

Species	Mars		Earth
	(1)	(2)	(3)
SiO ₂ (percent)	44.4	40.0	47.9
TiO ₂	0.14	0.1	0.2
Al ₂ O ₃	3.02	3.1	3.9
Cr ₂ O ₃	0.76	0.6	0.9
MgO	30.2	27.4	34.1
FeO	17.9	24.3	8.9
MnO	0.46	0.2	0.14
CaO	2.45	2.5	3.2
Na ₂ O	0.50	0.8	0.20
P ₂ O ₅	0.16	0.34	0.01
K (ppm)	315	573	200
Rb	112	—	0.4
Zn	74	—	46
Ga	6.6	—	4.6
In (ppb)	14	—	10
Cl (ppm)	44	—	8
Th	0.56	0.077	0.060
U	0.016	0.017	0.020

(1) SNC model (Dreibus and Wanke, 1985).

(2) Meteorite mix model plus geophysical constraints (Anderson, 1972a).

(3) Morgan and Anders (1980).

TABLE 1-17
Mars Surface Chemistry

Species	Soil	Igneous Component
SiO ₂ (percent)	43	53.3
TiO ₂	0.63	0.78
Al ₂ O ₃	7.2	8.9
Fe ₂ O ₃	18.0	2.23
MgO	6	7.4
CaO	5.7	7.1
SO ₃	7.6	—
Cl	0.75	—
Rb (ppm)	<30	—
Sr	60 ± 30	—
Y	70 ± 30	—
Zr	30 ± 20	—
K	2500	—
U	1.1 ± 0.8	—

Taylor (1982).

core would be about 45 percent of the planet's radius, or about 12 percent by mass. A small dense core would imply a high-temperature origin or early history because of the high melting temperature of nickel-iron, while a larger light core, presumably rich in sulfur, would allow a cooler early history, since sulfur substantially reduces the melting temperature. A satisfactory model for the interior of Mars can be obtained by exposing ordinary chondrites to moderate temperature, allowing the iron to form a core. The mechanisms for placing oxygen in the core and for reducing FeO to metallic iron would be less effective in Mars than Earth.

We know the equations of state (Chapter 5) and the locations of phase changes (Chapter 16) for most of the materials that might be expected to be important in the interiors of the terrestrial planets. With this information, we can completely define the structure of a two-zone planet (for example, one containing a mantle and a core) in terms of the zero-pressure densities of the mantle and the core and the radius of the core. For Mars two of these parameters can be found as a function of the third. If sulfur is the light alloying element in the core, the density of the core will increase as it grows and become richer in iron because of the nature of the Fe-FeS phase diagram.

The mantle of Mars is presumably composed mainly of silicates, which can be expected to undergo one or two major phase changes, each involving a 10 percent increase in density. To a good approximation, these phase changes will occur at one-third and two-thirds of the radius of Mars. The deeper phase change will not occur if the radius of the core exceeds one-third of the radius of the planet. With these parameters we can solve for the radius and density of the core, given the density of the mantle and the observable mass, radius and moment of inertia for Mars. The results are given in Table 1-18.

The curve in Figure 1-8, giving these results in terms of the density of the core and the radius of the core, is the locus of possible Mars models. Clearly, the data can accommodate a small dense core or a large light core. The upper limit to the density of the core is probably close to the density of iron, in which case the core would be 0.36 of Mars's radius, or about 8 percent of its mass. To determine a lower limit to the density, one must consider possible major components of the core. Of the potential core-forming materials, iron, sulfur, oxygen and nickel are by far the most abundant elements.

The assumption of a chondritic composition for Mars leads to values of the relative radius and mass of the core: $R_c/R = 0.50$ and $M_c/M = 0.21$. The density of the mantle is less than the density of the silicate phase of most ordinary chondrites.

Three kinds of chondrites, HL (high iron, low metal or ornasites), LL (low iron, low metal or amphoterites) and L (low iron or hypersthene-olivine) chondrites, all have lower amounts of potentially core-forming material than is required for Mars, although HL and LL have about the right silicate density (3.38 g/cm³). If completely differentiated, H (high iron) chondrites have too much core and too low a silicate density (3.26–3.29 g/cm³). We can match the properties of H chondrites with Mars if we assume that the planet is incompletely differentiated. If the composition of the core-forming material is on the Fe side of the Fe-FeS eutectic, and temperatures in the mantle are above the eutectic composition, but below the liquidus, then the core will be more sulfur-rich and therefore less dense than the potential core-forming material.

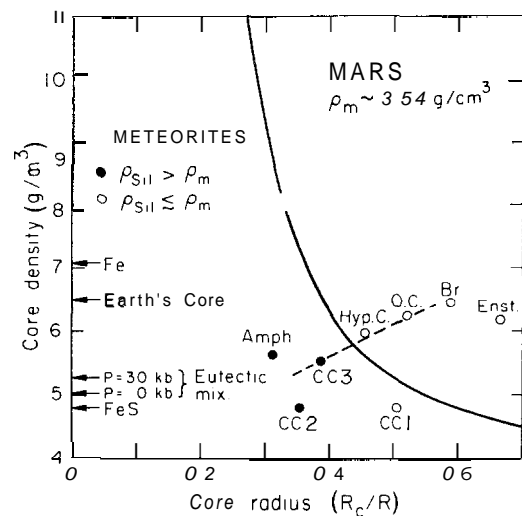


FIGURE 1-8
Radius of the core versus density of core for Mars models. The points are for meteorites with all of the FeS and free iron and nickel differentiated into the core. The dashed line shows how core density is related to core size in the Fe-FeS system (after Anderson, 1972a).

TABLE 1-18
Model Compositions of Mars

Species	Ma1	Ma2	Ma3	Ma4	Ma5
Mantle Plus Crust					
SiO ₂	43.6	43.9	41.6	40.0	41.0
TiO ₂	0.16	0.16	0.3	0.1	—
Al ₂ O ₃	3.1	3.2	6.4	3.1	3.2
Cr ₂ O ₃	—	—	0.6	0.6	—
MgO	31.0	31.2	29.8	27.4	34.1
FeO	17.2	16.7	15.8	24.3	16.8
MnO			0.15	0.2	—
CaO	2.9	3.0	5.2	2.5	2.8
Na ₂ O	1.4	1.4	0.1	0.8	—
H ₂ O	0.47	0.44	0.001	0.9	—
K (ppm)	1190	1199	69	573	1200
U (ppm)	0.017	0.017	0.033	0.017	0.017
Th (ppm)	0.078	0.078	0.113	0.077	0.079
Core					
Fe	59.8	60.4	88.1	72.0	—
Ni	5.9	5.8	8.0	9.3	—
S	34.3	33.8	3.5	18.6	—
Relative Masses					
Mantle + Crust	74.7	74.3	81.0	88.1	85.0
Core	25.3	25.7	19.0	11.9	15.0

Ma1: Equilibrium condensation (BVP, 1980).

Ma2: Equilibrium condensation modified by use of feeding zones (BVP, 1980).

Ma3: Morgan and Anders (1980).

Ma4: Anderson (1972a).

Ma5: McGetchin and Smyth, mantle only (in BVP, 1980).

Carbonaceous chondrites are extremely rich in such low-temperature condensates as H₂O and FeS, as well as carbon and "organic matter." If we ignore the water and carbon components, a fully differentiated planet of this composition would have a core of 15 percent by mass, composed mainly of FeS (13.6 percent FeS, 1.4 percent Ni), and a mantle with a density of about 3.5 g/cm³. However, these meteorites also contain about 19 percent H₂O, most of which must have escaped if Mars is to be made up primarily of this material. Otherwise, the mantle would not be dense enough.

In ordinary high-iron chondrites, the free iron content averages 17.2 percent by weight. The FeS content is approximately 5.4 percent (3.4 percent Fe, 2.0 percent S) and the nickel content is 1.6 percent. A planet assembled from such material, if completely differentiated, would yield a core of 24 percent of the mass of the planet, with Fe:S:Ni in the approximate proportions of 21:2:2 by weight. Low-iron chondrites would yield a core of 15 percent of the mass of the planet, with proportions of 12:1:2.

Carbonaceous chondrites have little or no free iron but contain 7–25 percent by weight FeS and about 1.5 percent nickel. The average core size for a planet made of carbonaceous chondrites would be 15 percent by mass, Fe:S:Ni being in the proportions 18:10:3. An absolute minimum core density can probably be taken as 4.8 g/cm³, corre-

sponding to a pure FeS core with a fractional core radius of 0.6 and a fractional mass of 26 percent. On these grounds, the mass of the martian core can be considered to lie between 8 and 24 percent of the mass of the planet.

This range can be narrowed considerably by further consideration of the compositions of meteorites. The points in Figure 1-8 represent most of the major categories of stony meteorites. The size and density of the "core" are computed from the amounts of iron, sulfur, and nickel in the meteorite. No single class of meteorites, fully differentiated into core and mantle, would satisfy the data for Mars, although carbonaceous chondrites and hypersthene (low-iron) chondrites come close. The open circles indicate silicate (mantle) densities less than the inferred density for the mantle of Mars; the closed circles indicate silicate densities that are too high. The meteorites above the curve can be migrated downward and to the left by placing some of the iron of the core in the mantle and thereby increasing the density of the mantle and decreasing the density and radius of the core. Physically, the result would correspond to a meteorite model that has been incompletely differentiated. If chondrites are an appropriate guide to the composition of Mars, possible core sizes would be further restricted to 12–15 percent by mass. Alternatively, the closed circles could be migrated to the locus of possible Mars models by reducing some of the FeO in the silicate phase and allowing

the iron to enter the core. This procedure is much more drastic and requires high temperature and, probably, the presence of carbon to effect the reduction. It is interesting that the meteorites in question do contain substantial amounts of carbon. The resulting cores would be about the same size as was previously inferred.

A third possibility would be to assemble Mars from a mixture of meteorites that fall above and below the curve. The meteorites below the curve, however, are relatively rare, although Earth may not be collecting a representative sample.

The size of the core and its density can be traded off. By using the density of pure iron and the density of pure troilite (FeS) as reasonable upper and lower bounds for the density of the core, its radius can be considered to lie between 0.36 and 0.60 of the radius of the planet.

The zero-pressure density of the mantle implies an FeO content of 21–24 weight percent unless some free iron has been retained by the mantle. The presence of CO, and H₂O, rather than CO and H₂, in the martian atmosphere suggests that free iron is not present in the mantle.

If chondrites are an appropriate guide to the major-element composition of Mars, the core of Mars is smaller and less dense than the core of Earth, and the mantle of Mars is denser than that of Earth. Mars contains 25–28 percent iron, independent of assumptions about the overall composition or distribution of the iron. Earth is clearly enriched in iron or less oxidized when compared with Mars or most classes of chondritic meteorites.

The lithosphere on Mars is apparently much thicker than on Earth and is capable of supporting large surface loads. The evidence includes the roughness of the gravity field, the heights of the shield volcanoes, the lack of appreciable seismicity and thermal history modeling. There is some evidence that the lithosphere has thickened with time. Olympus Mons is a volcanic construct with a diameter of 700 km and at least 20 km of relief, making it the largest known volcano in the solar system. It is nearly completely encircled by a prominent scarp several kilometers in height and it coincides with the largest gravity anomaly on Mars. The load is apparently primarily supported by thick lithosphere, perhaps greater than 150 km in thickness (Thurber and Toksoz, 1978).

The surface of Mars is much more complex than those of the Moon and Mercury. There is abundant evidence for volcanic modification of large areas after the period of heavy bombardment, subsequent to 3.8 Ga. Mars has a number of gigantic shield volcanoes and major fault structures. In contrast to Mercury there are no large thrust or reverse faults indicative of global contraction; all of the large tectonic features are extensional. The absence of terrestrial-style plate tectonics is probably the result of a thick cold lithosphere. The youngest large basins on Mars have gravity anomalies suggesting incomplete isostatic compensation and therefore a lithosphere of finite strength (Head and Solomon, 1981).

The data regarding an intrinsic magnetic field for Mars are inconclusive but allow, at most, only a small permanent field in spite of the rapid rotation rate and the probable presence of a dense core. There is no information yet on the physical state of the core.

Mercury

The planet Mercury has a mass of 3.30×10^{26} g and a mean radius of 2444 km, giving a density of 5.43 g/cm³. Although Mercury is only 5.5 percent of the mass of the Earth, it has a very similar density. Any plausible bulk composition satisfying this density (Table 1-19) is about 60 percent iron. This iron is largely differentiated into a core, because (1) Mercury has a perceptible magnetic field, appreciably more than either Venus or Mars; and (2) Mercury's surface has the appearance of being predominantly silicate. A further inference is that the iron core existed early in its history; a late core-formation event would have resulted in a significant expansion of Mercury. The presence of an internally generated magnetic field implies that the iron core is at least partially fluid.

Mercury's shape may have significantly changed over the history of the planet. Tidal despinning results in a less oblate planet and compressional tectonics in the equatorial regions. Cooling and formation of a core both cause a change in the mean density and radius. A widespread system of arcuate scarps on Mercury, which appear to be thrust

TABLE 1-19
Cosmochemical Model Compositions of Mercury
(Weight Percent or ppm)

Species	Me4	Me3
Mantle Plus Crust		
SiO ₂	43.5	47.1
TiO ₂	—	0.33
Al ₂ O ₃	4.7	6.4
Cr ₂ O ₃	—	3.3
MgO	47.7	33.7
FeO	0	3.7
MnO	—	0.06
CaO	4.1	5.2
Na ₂ O	0	0.08
H ₂ O	—	0.016
K (ppm)	0	69
U (ppm)	0.026	0.034
Th (ppm)	0.12	0.122
Core		
Fe	94.5	93.5
Ni	5.5	5.4
S	0	0.35
Relative Masses		
Mantle + Crust	35.2	32.0
Core	64.8	68.1

Me3: Morgan and Anders (1980).

Me4: Refractory condensate mixing model (Wood, 1962).

faults, provides evidence for compressional stresses in the crust (Solomon, 1977). The absence of normal faults, the result of extension, suggest that Mercury has contracted, perhaps by as much as 2 km in radius. This perhaps is evidence for cooling of the interior.

The primary factor affecting the bulk composition of Mercury is the probably high temperature in its zone of the solar nebula, so that it is formed of predominantly high-temperature condensates. If the temperature was held around 1300 K until most of the uncondensed material was blown away, then a composition satisfying Mercury's mean density can be obtained, since most of the iron will be condensed, but only a minor part of the magnesian silicates. Since the band of temperatures at which this condition prevails is quite narrow, other factors must be considered. Two of these are (1) dynamical interaction among the material in the terrestrial planet zones, leading to compositional mixing, and (2) collisional differentiation.

If the composition of Mercury is controlled by high-temperature condensation, it is unlikely to have significant SiO_2 beyond that necessary to combine with the MgO , Al_2O_3 , and CaO to make forsterite, diopside, and anorthite. Any such model of Mercury must have a ratio of $\text{Fe} + \text{Ni}$ to $\text{Al}_2\text{O}_3 + \text{CaO} + \text{MgO} + \text{SiO}_2$ higher than calculated from cosmic abundances to satisfy its mean density (BVP).

Mercury probably has an $\text{Al}:\text{Mg}$ ratio appreciably higher than cosmic but an $\text{Al}:\text{Fe}$ ratio somewhat lower, leading to an anorthite content on the order of 5–10 percent of its mass. If Mercury had a differentiation efficiency similar to that of the Moon (the nearest terrestrial planet in terms of size), it would have a crust approximately 75 to 150 km thick.

The areal density and size distribution of craters on the surface of Mercury are similar to lunar highland values. Both Moon and Mercury, and by inference the other terrestrial bodies, were subjected to a high flux of impacting objects in early planetary history. The end of the high-flux period can be dated from lunar studies at about 3.8 billion years ago. The large basins on the surface of Mercury probably formed during the period of high bombardment. Later cooling and contraction apparently were responsible for global compression of the outer surface and may have shut off volcanism. Tidal despinning may have affected the distribution of fault scarps over the surface.

COMPARATIVE PLANETOLOGY

Before the advent of space exploration, Earth scientists had a handicap almost unique in science: They had only one object to study. Compare this with the number of objects available to astronomers, particle physicists, biologists and sociologists. Earth theories had to be based almost entirely on evidence from Earth itself. Although each object in the

solar system is unique, we have learned some lessons that can now be applied to Earth.

Study of the Moon, Mars and the basaltic achondrites demonstrated that basaltic volcanism is ubiquitous, even on very small bodies. Planets apparently form, or become, hot and begin to differentiate at a very early stage in their evolution, probably during accretion. Although primitive objects, such as the carbonaceous chondrites, have survived 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.

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

We now know that the total crustal volume on the Earth is anomalously small, but it nevertheless contains a large fraction of the terrestrial inventory of incompatible elements.

The difference in composition of the atmospheres of the terrestrial planets shows that the original volatile compositions, the extent of outgassing or the subsequent processes of atmospheric escape have been quite different. The importance of great impacts in the early history of the planets is now clear; it has implications ranging from early, dense, insulating atmospheres, atmospheric escape and siderophile and volatile composition to formation of the Moon and angular momentum of the planets.

We now know that plate tectonics, at least the recycling kind, is unique to Earth. The thickness and average temperature of the lithosphere and the role of phase changes in basalt seem to be important. Any theory of plate tectonics must explain why the other terrestrial planets do not behave like Earth.

The compositions of lunar KREEP and terrestrial kimberlite are so similar that a similar explanation is almost demanded.

These lessons, some of which are just hints, should be kept in mind as we turn our attention to Earth.

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