

Lecture 31

Planetary Accretion – the raw materials and the final compositions

Reading this week: White Ch 11 (sections 11.1 -11.4)

Today

1. Boundary conditions for Planetary Accretion

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Growth and Differentiation of Planet Earth

It is difficult to separate the planetary processes of

- a. **accretion** and
- b. **differentiation** (into core, mantle, crust, atmosphere/hydrosphere)

because a number of features suggest they overlapped in time.

Thus, we will work at this problem from both ends over the next few lectures, using as many of the *Boundary Conditions* for earth's early history as we can to infer a likely sequence of events.

Today we examine constraints on planetary compositions at the “beginning” and “end” of the accretion sequence.

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Growth and Differentiation of Planet Earth

compositional:

- ♣ relative proportions of refractory, silicate, metal and volatile components on earth compared to the Sun and C1-3 chondrites
- ♣ how and when these components were segregated into our present core, mantle, crust and atmosphere.
- ♣ redox conditions during accretion.
- ♣ density of earth and its neighboring planets.

physical:

- ♣ rates of accretion and thermal history of planetesimals

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evidence from planetary density

As per last lecture, the solar nebula probably originally had a **fairly regular radial density gradient**. (Density anomalies at Mercury, Mars and the Asteroids have been related to peculiarities early in the accretion process.)

Compare that to modern planetary densities:

Planet	density (g/cm ³)	
Mercury	5.42	-----
Venus	5.25	stony
Earth	5.52	(terrestrial)
Mars	3.94	----- planets
Asteroids	2.2	(Ceres)
Jupiter	1.3	-----
Saturn	0.7	gaseous
Uranus	1.3	planets
Neptune	1.7	
Pluto	1.1	-----

Note: Silicate rocks at the **earth's surface** have mean density **~2.7 g/cm³**.

Thus, much **denser material** must exist inside the Earth for the planet to have a bulk density of 5.52. This material is in the metallic core, where the density is closer to 8 g/cm³. The **core is ~32% of Earth's mass**.

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The density variations between Earth and the other stony planets have been ascribed to a combination of two factors:

- ♣ Accretion of **different proportions of metal and silicate condensates** in different zones of the solar nebula. As discussed last lecture, this requires some sort of physical density (or magnetically controlled) segregation of materials rather early in the accretion sequence.

stony-Fe meteorites show separate blobs of Fe and silicate material, so some segregation is possible on a small scale in early solar system history.

1

Metal-Silicate Physical Fractionation Hypothesis

Table 6.3 Mean densities of terrestrial planets at 10 kbar, assuming that they are composed of varying proportions of metal phase ($\rho_{10} = 7.9 \text{ g/cm}^3$) and silicate phase ($\rho_{10} = 3.3 \text{ g/cm}^3$)

Planet	Mean density	Mean density at 10 kb	Percent nickel-iron phase
Mercury	5.44	5.3	65
Venus	5.24	3.96	28.8
Earth	5.52	4.07	32.5
Moon	3.34	3.40	5
Mars	3.94	3.73	20

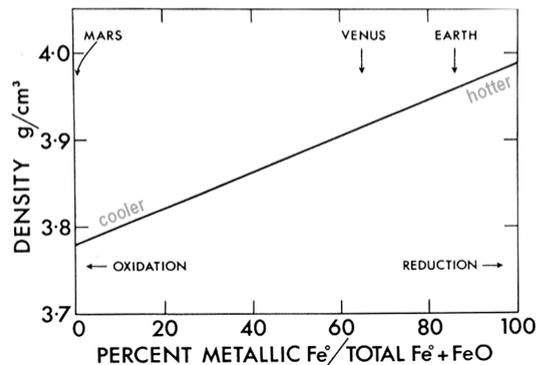
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- ♣ **Redox state variations** with location during accretion. Different redox conditions affect the density of Fe-rich condensates by changing the Fe/FeO ratio. Fe is denser than FeO.

2

Redox Hypothesis

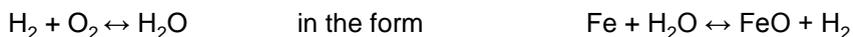
Figure 6.9 Relationship between density and oxidation state of primordial material (represented by Type 1 carbonaceous chondrites on a volatile-free and sulphur-free basis). Densities were calculated to correspond to mineral assemblages stable at 10 kbar. Positions of planets in relation to redox states according to hypothesis of Ringwood (1959) are indicated. (From Ringwood and Anderson 1977, with permission.)



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How can redox state vary during accretion?

Essentially, with our old friend



Reaction products are **avored at lower temperature** (such as is found at say, Mars relative to Earth)

The implication of both hypotheses (the redox state and metal-silicate segregation) is that we must be very careful in estimating both the oxidation state and the proportion of Fe metal to other elements in the material from which the Earth accreted.

Interestingly, $\text{Fe}/\text{Si}_{\text{earth}} \approx \text{Fe}/\text{Si}_{\text{sun}}$. So, if silicate-metal segregation was the main process, it *conveniently* produced **roughly equal proportions of metal and silicate in the Earth as in the sun (within 25%)**.

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Chemical Evidence

Goldschmidt's Classification and the Geochemical Periodic Chart

It is useful at this point to briefly discuss the affinities of elements for the four general types of material thought to exist during accretion:

Siderophile: iron-“liking” elements (liking zero-valent Fe; i.e., metal)

Chalcophile: sulfide-liking (S^{2-})

Lithophile: silicate-liking ($[\text{SiO}_4]_n$, also O-loving in practice)

Atmophile: gas-phase-liking

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Chemical Evidence

Goldschmidt's Classification and the Geochemical Periodic Chart

This classification was proposed in the 1920s by geochemist Victor Goldschmidt.

It is qualitatively useful for describing the origin of the Earth from materials present in the early solar system—particularly once liquids (iron and silicate) had become important.

Goldschmidt compared the distributions of elements in silicate, metal-rich and gas phases in

1. *metal-ore smelter materials*
2. *meteorites,*
3. *the modern Earth.*

He recognized a pattern in how elements were distributed with other elements.

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Goldschmidt Classification/Geochemical Periodic Chart

Elements can be assigned to more than one group depending on the situation, so this scheme provides only generalities.

Atmophile				Siderophile				Chalcophile		
H	(C)	N	(O)	He	Fe	Co	Ni	Cu	Zn	Ga
				Ne	Ru	Rh	Pd	Ag	Cd	In
				Ar	Os	Ir	Pt		Hg	Tl
				Kr		plus			plus	
				Rn	Mo		Au	(Ge)	As	S
					(W)	Re		(Sn)	Sb	Se
						and		Pb	Bi	Te
					C				and	
						P		(Cr)	(Fe)	
					Ge	(As)		(Mo)		
					Sn					
					(Pb)					

Lithophile									
Li	Be	B	(C)		O	F			
Na	Mg	Al	Si	P		Cl			
K	Ca	(Ga)	(Ge)			Br			
Rb	Sr					I			
Cs	Ba	(Tl)							
		plus							
	Sc	Ti	V	Cr	Mn	(Fe)			
	Y	Zr	Nb						
	La	Hf	Ta	W					
	REE	Th		U					
		and							
		(H)							

*The elements in parentheses occur primarily in another class.

- Siderophile** – iron liking (zero-valent Fe)
- Chalcophile** – sulfide liking (S^{2-})
- lithophile** – silicate liking ($[(SiO_4)_n]$, also O loving in practice)
- Atmophile** – gas phase liking

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Goldschmidt Classification/Geochemical Periodic Chart

The groups have a general relationship to the periodic chart, reflecting an underlying relationship to the electronic configurations of the elements in their common forms.

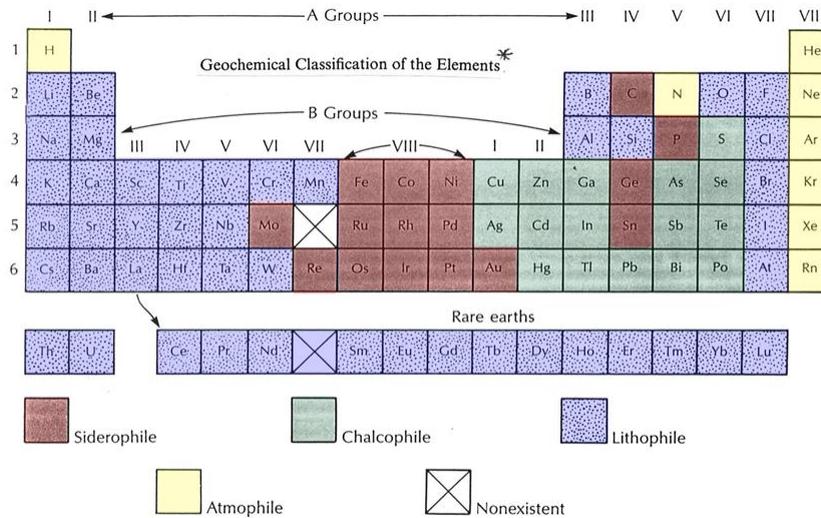


Figure 7.3 Geochemical classification of the elements in the periodic table. The classification is based on the way the elements distribute themselves between an iron liquid (siderophile), a sulfide liquid (chalcophile), a silicate liquid (lithophile), and a gas phase (atmosphile).

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Element Relationships: Earth and C1 Chondrites

Among three of the most important siderophile and lithophile elements, compositional data for the sun, chondrites and bulk Earth indicate that the **Earth has higher Fe/Si and Mg/Si than the chondrites.**

Table 3-6. Comparison between the chemical compositions of various meteorite classes and that of the bulk Earth: Despite the great change in the fraction of iron in oxide form from class to class (see last column), the relative abundances of the three major constituent metals (i.e., Si, Mg, and Fe) remain nearly unchanged in the high-iron chondrites. As can be seen, the Earth is richer in magnesium and iron relative to silicon than are meteorites.

	Fraction of mass as SiO ₂ , MgO, FeO, and iron metal*	Relative atom abundance				Relative mass abundance				Fraction of iron in oxide form
		Si	Mg	Fe	O†	Si	Mg	Fe	O†	
LOW-IRON CHONDRITES										
Enstatite chondrites	.92	100	92	60	325	100	80	119	185	.55
HIGH-IRON CHONDRITES										
Carbonaceous chondrites	.78	100	104	84	380	100	90	167	216	.90
Olivine-Pigeonite chondrites	.92	100	101	78	357	100	87	155	203	.72
Olivine-bronzite chondrites	.91	100	96	79	340	100	83	157	194	.56
Olivine-Hypersthene chondrites	.92	100	84	82	285	100	73	163	162	.01
WHOLE EARTH										
Whole Earth	.94	100	131	126	359	100	114	250	199	.11

*FeS present in meteorites included with iron metal.

†Excludes oxygen associated with metals other than silicon, magnesium, and iron.

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...And the **Earth is variably depleted in volatile elements** (e.g., K, Rb, Cs, etc.) relative to chondrites.

TABLE 4.3. *Element abundances in the Earth and selected meteorites*

	Ordinary chondrites	Earth	Eucrites
Na %	0.68	0.07	0.3
K ppm	850	130	360
Rb ppm	2.8	0.42	0.24
Cs ppm	4 to 619	9	12
Ca %	1.21	0.95	7.49
Sr ppm	11	12	85
Ba ppm	4.1	5.3	35
Sc ppm	8.0	16.5	27
Y ppm	2.0	4.2	23
La ppm	0.24	0.52	3.7
Th ppb	43	35	440
U ppb	12	12	130

Based on Larimer (1971).

Table 3-7. *Relative abundance of the first 28 elements and their fates during the formation of the terrestrial planets:*

Element Number	Element Name	Compound Solid	Gas	Rel. Abundance In Sun*	Fate†	Rel. Abundance In Chondrites*
1	HYDROGEN	H ₂		40,000,000	(1)	—
2	HELIUM	He		3,000,000,000	(1)	trace
3	LITHIUM	Li ₂ O		60	(3)	50
4	BERYLLIUM	BeO		1	(3)	1
5	BORON	B ₂ O ₃		43	(2)	6
6	CARBON	CH ₄		15,000,000	(1)	2,000
7	NITROGEN	NH ₃		4,900,000	(1)	50,000
8	OXYGEN	H ₂ O**		18,000,000	(2)	3,700,000
9	FLUORINE	HF		2,800	(1)	700
10	NEON	Ne		7,600,000	(1)	trace
11	SODIUM	Na ₂ O		67,000	(2)	46,000
12	MAGNESIUM	MgO		1,200,000	(3)	940,000
13	ALUMINUM	Al ₂ O ₃		100,000	(3)	60,000
14	SILICON	SiO ₂		1,000,000	(3)	1,000,000
15	PHOSPHORUS	P ₂ O ₅		15,000	(3)	13,000
16	SULFUR	FeS	H ₂ S	580,000	(2)	110,000
17	CHLORINE	HCl		8,900	(1)	700
18	ARGON	Ar		150,000	(1)	trace
19	POTASSIUM	K ₂ O		4,400	(2)	3,500
20	CALCIUM	CaO		73,000	(3)	49,000
21	SCANDIUM	Sc ₂ O ₃		41	(3)	30
22	TITANIUM	TiO ₂		3,200	(3)	2,600
23	VANADIUM	VO ₂		310	(3)	200
24	CHROMIUM	CrO ₂		15,000	(3)	13,000
25	MANGANESE	MnO		11,000	(3)	9,300
26	IRON	FeO, FeS, Fe		1,000,000	(3)	690,000
27	COBAL	CoO		2,700	(3)	2,200
28	NICKEL	NiO		58,000	(3)	49,000

*Relative to 1,000,000 silicon atoms.

(1) Highly volatile; mainly lost.

(2) Moderately volatile; partly captured.

(3) Very low volatility; largely captured.

**Plus metal oxides.

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More broadly, relative to C1 chondrite abundances, the Earth's **estimated primitive mantle** (i.e., after removal of Fe to the core but assuming no extraction of crust from the mantle) has (see also Lecture 31):

- refractory elements (Al, Ca, U, Th, Si, Ba, rare earths,...) $\approx 3 \times$ C1 values.
- depletions in other elements relative to C1 values (including Mg and Si).

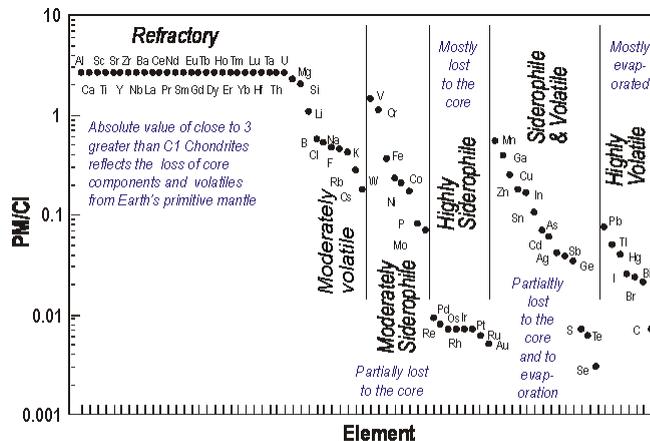


Figure 11.10. Abundances of the elements in the Primitive Mantle compared to C1 chondrites. The two primary processes that resulted in relative elemental loss were incorporation into the core and volatilization modified from White, *Geochemistry*

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As we discussed last lecture, we use a “CHUR” model (**chondritic uniform reservoir**) as the bulk starting composition for Earth and the other planetary bodies on our solar system.

The refractory elements (~40 of them) match this model quite well.

If the "CHUR" model for bulk earth composition is correct, then the depletion for the other elements in Earth's mantle (in the previous slide) implies that these other elements were:

- ♣ were “rearranged” within the Earth during differentiation (i.e., partially put in other sub-reservoirs, like the core and atmosphere),

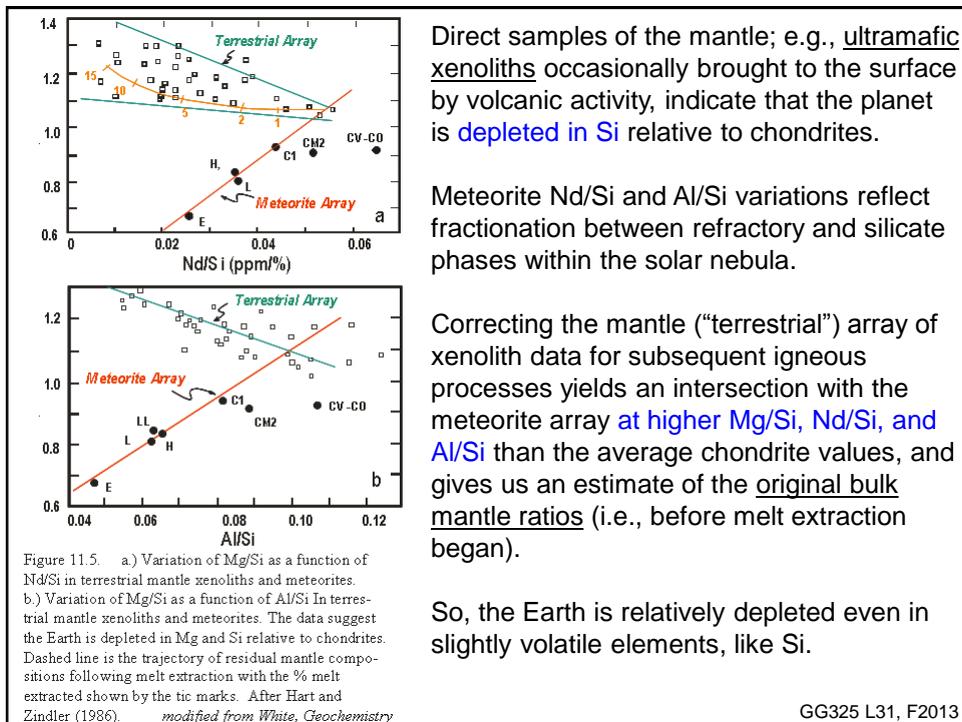
Or

- ♣ were partly lost (volatilized) to space during accretion.

We know that some “depleted” elements in this category, they are now partially in the core (**Siderophile** elements), crust (**lithophile**) and atmosphere (**Atmophile**).

So how can we get at the bulk Earth composition?

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In fact, elemental abundances in **ordinary chondrites** (the slightly metamorphosed cousins of the carbonaceous chondrites) relative to C1 values imply some amount of **volatility-based element fractionation** occurred **before** large-scale accretion.

♣ So, some volatile elements were lost from chondrites well before planetary differentiation occurred. Thus, we **must be very careful in estimating the volatile content of the starting material for the Earth** from a specific type of chondrite, even for the moderately volatile elements.

no single type of chondrites serves as a good model for the Earth for all elements.

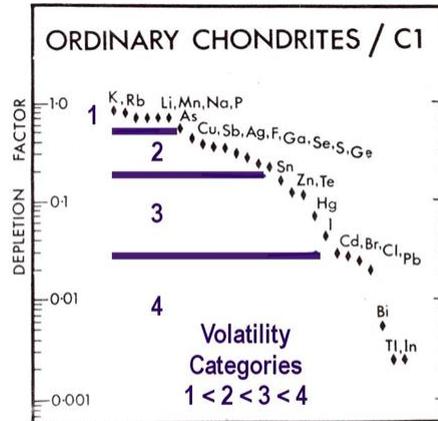
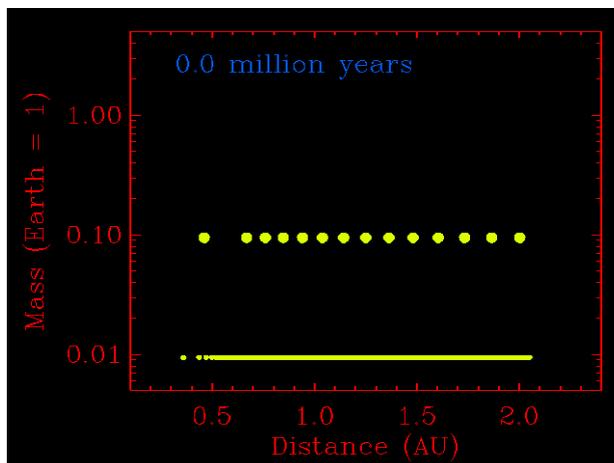


Figure 9.2 Comparison of abundances of volatile elements in ordinary chondrites with corresponding abundances in C1 chondrites. (From Ringwood and Kesson, 1977; with permission.)

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Accretion of Planetary Embryos



Here's an animation of a recent numerical model of late-stage accretion in the inner solar system. It assumes:

- 14 planetary embryos
- hundreds of asteroid-size objects
- Jupiter and Saturn are present.

Taylor (2008) after Chambers (2001)

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What does the differentiated Earth indicate about its composition during accretion?

Seismological studies of the interior demonstrate that today the Earth is radially zoned, with layers of increasing density toward the center.

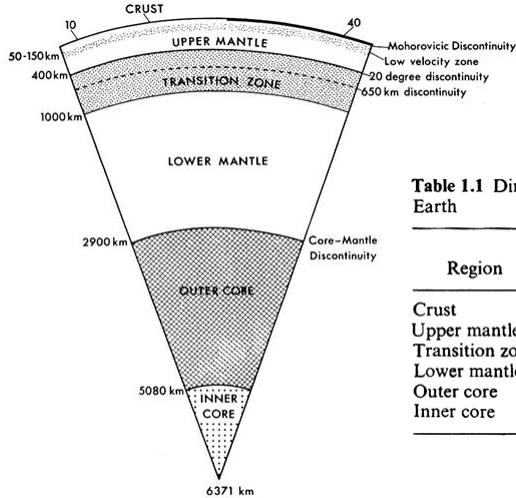


Table 1.1 Dimensions and masses of the internal layers of the Earth

Region	Depth to boundaries (km)	Mass (10 ²⁵ g)	Fraction of total mass
Crust	0-Moho	2.4	0.004
Upper mantle	Moho-400	62	0.10
Transition zone	400-1000	100	0.17
Lower mantle	1000-2900	245	0.41
Outer core	2900-5154	177	0.30
Inner core	5154-6371	12	0.02

Figure 1.2 Principal internal subdivisions of the Earth. (From Ringwood 1975a, with permission.)

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Density of the solid substrate can be inferred from the seismological data

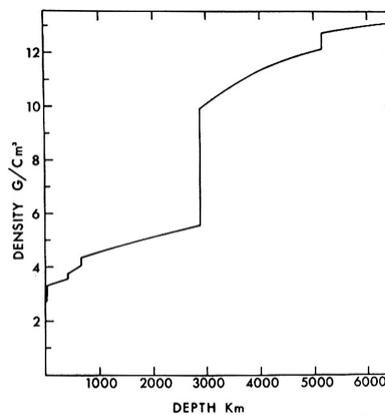
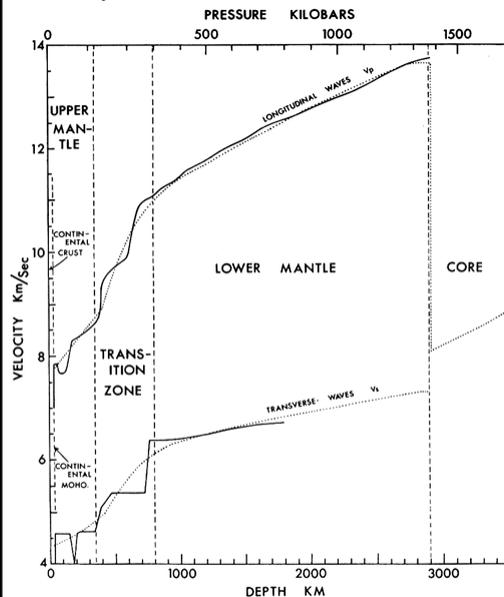


Figure 1.8 Distribution of density with depth throughout the Earth. The model is from Dziewonski et al. (1975) and is very similar to the classical Model A derived by Bullen (1940).

Figure 1.1 Variation of seismic P and S wave velocities throughout the mantle and outermost core. P waves—solid line: Johnson (1967, 1969); S waves—solid line: Nuttli (1969); broken lines: Jeffreys (1939). (From Ringwood, 1975a, with permission).

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From a combination of high-pressure mineral experiments, modeling, and seismic studies of high-pressure minerals, we know the main phases associated with each of the major transitions in density in the mantle.

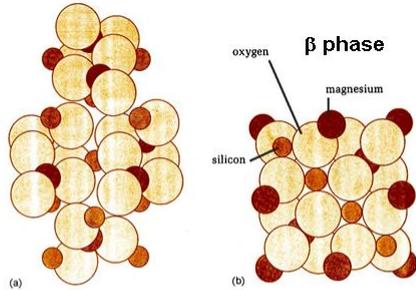


Figure 17-36
Olivine (Mg_2SiO_4 with Fe_2SiO_4) is a major mineral in the Earth's mantle. Part (a) shows olivine in its low-pressure form. The large pale-brown atoms are oxygen, the medium-brown atoms are silicon, and the dark-brown ones are magnesium (or iron). When the pressure reaches a critical value, corresponding to a depth of about 400 km in the Earth, the molecule collapses into a denser form (b). Note the decrease in void space in the high-pressure form, in which the oxygen atoms are more closely packed. Seismologists have found where this transition occurs in the Earth, and petrologists have observed the transition in high-pressure laboratory experiments.

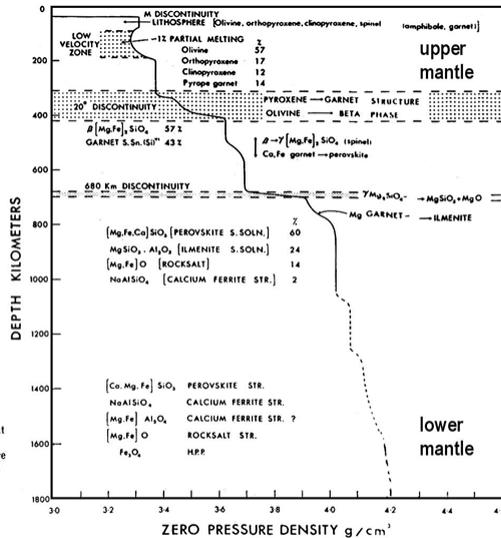


Figure 1.9 Mineral assemblages and corresponding zero-pressure densities for a model mantle of pyrolite composition based on the results of static high-pressure, high-temperature experimentation.

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For the core, the proportions are:

Fe ($\approx 85\%$)

Ni ($\approx 5\%$)

light elements ($\approx 10\%$)

The light elements are believed to mostly be O, S, Si and/or C.

The light elements can not be directly determined; instead, they are inferred from the presumed conditions of core formation and remain a topic of debate.

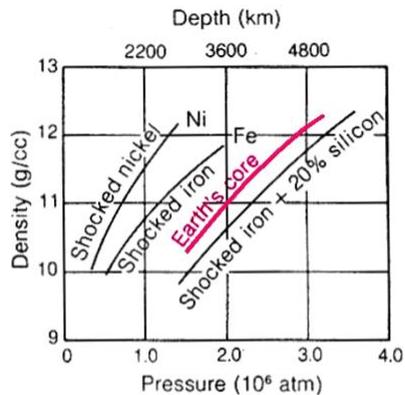


Figure 17-37
Density in the Earth's fluid core plotted against depth below the surface and against pressure (black curve). Comparison with the densities for iron, nickel, and iron-silicon mixtures measured in laboratory studies enables seismologists to conclude that the core is mostly iron but slightly less dense than pure iron, as if a small amount of a "lightening" element like silicon were present.

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We can approximate the bulk major-element composition of the primitive upper mantle pretty closely by taking a C1 chondrite, removing most of its volatile elements, and removing most of the Fe and Ni (i.e., segregating them to the core).

pyrolite is a slight modification of this major element composition. It is commonly used as a model for the starting composition of the mantle after core formation.

	Orgueil Wiik (1956)	High-temperature ¹ component		Primordial model mantle wt.-%	Pyrolite wt.-%
Metallic nickel-iron (~5% Ni)	—	34.1 *			
SiO ₂	21.7	32.8	SiO ₂	48.2	45.1
TiO ₂	0.1	0.2	TiO ₂	0.15	0.2
Al ₂ O ₃	1.6	2.8	Al ₂ O ₃	3.5	3.3
Cr ₂ O ₃	0.35	0.2	Cr ₂ O ₃	0.7	0.4
MnO	0.2	0.1	MgO	34.0	38.1
FeO	22.9	—	FeO	8.1	8.0
NiO	1.2	—	MnO	0.5	0.15
MgO	15.2	27.7	CaO	3.3	3.1
CaO	1.2	2.3	Na ₂ O	1.6	0.4
Na ₂ O	0.7	—	K ₂ O	0.15	0.03
K ₂ O	0.07	—			
P ₂ O ₅	0.3	—	CaO (MgO + SiO ₂)	0.040	0.037
Water	19.2	—	Al ₂ O ₃ (MgO + SiO ₂)	0.043	0.040
Organic compounds	9.7	—			
Sulphur	5.7	—			

¹ Based on devolatilized (including some Si, Cr, and Mn) CI composition and with all oxidized iron and nickel reduced to the metallic state (after Ringwood, 1977b).
* all Fe to core except enough to have $Mg^{2+}/(Mg^{2+} + Fe^{2+}) = 0.88$

* Based on average of chemical analyses of Orgueil and Ivuna (Wiik, 1956) on a volatile and sulphur-free basis. Siderophile components (NiO and most of the FeO) have been removed and assigned to the model "core" following the procedure of Ringwood (1966a), leaving sufficient FeO in mantle to produce an MgO/(MgO + FeO) ratio of 0.88.

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Continued differentiation of the Earth has caused the **upper mantle** to segregate into **compositional domains** (we'll talk more about this soon).

Differences in **seismic anisotropy** between two commonly inferred rock compositions (peridotite and pyroxenite or eclogite, both occasionally brought to the surface as xenoliths in volcanic rocks) are useful in modeling their distribution in the upper mantle.

Peridotite must be the most abundant rock type in the upper mantle.

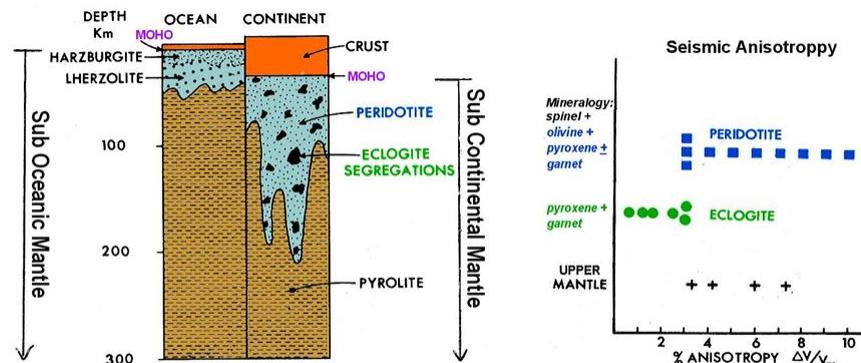


Figure 1.5 Chemically zoned model for the upper mantle. Harzburgite (olivine + orthopyroxene + chromite) and lherzolite (olivine + orthopyroxene + clinopyroxene + spinel) are varieties of peridotite. (From Ringwood, 1975a, with permission.)

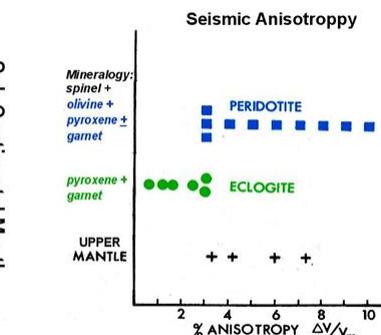


Figure 1.4 Percent anisotropies (expressed as difference between maximum and minimum observed velocities, ΔV , divided by the mean velocity V_m , for peridotites, eclogites and the upper mantle. (From Ringwood, 1975a, with permission.)

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