Bulk Composition of the Earth:

Purpose — find out what the $\oplus$ and other planets in the solar system made of?

We have learned that

talk about decompression here! if skip on Monday

$$\boxed{\begin{aligned} \bar{\rho}_\oplus &= 5500 \text{ kg/m}^3 \\ \bar{\rho}_\oplus \text{ uncompressed} &\approx 4300 \text{ kg/m}^3 \end{aligned}}$$

Show Table 3-3 from Broecker — vital statistics of other planets

Terrestrial planets — Mercury, Venus, $\oplus$, Mars, Pluto

♇ mass now known from

~~spacecraft and~~ ~~[illegible]~~

Hubble telescope observations of its moon Charon

Giant gaseous planets — Jupiter, Saturn, Uranus, Neptune

Also a number of very large moons

	radius (km)	$\bar{\rho}$ (kg/m^3)
Earth		
Moon	1740	3360
Jupiter		
Io	3630	3500
Europa	3140	2900
Ganymede	5260	1900
Callisto	4800	1800
Saturn		
Titan	5150	1900

So, what could the $\oplus$ and other planets be made of to have these densities?

insert page 2 1/2 here

Look at surface rocks to get an idea.

Three main types of rocks – chapter on each in book:

1. <u>igneous</u> – volcanic & plutonic

expelled from volcanoes

i.e. intrusive

[redacted]

Divided into a number of different types depending on chemistry

What controls the density of a substance anyway?

Two things:

(1) how heavy are the atoms out of which it is made?

(2) how closely packed are those atoms in the crystal lattice of the constituent minerals.

Simply what something is made of has the most important influence on ρ — hence knowledge of planetary ρ's has something to tell us about their chemistry

Example: Table 3-4 Broecker

H_2O : atomic weight $1+1+16=18$
6 nucleons/atom

$CaCO_3$: $40+12+3(16) = 100$
20 nucleons/atom

density ~ nucleons/atom

Table 3-3. Characteristics of the planets:*

Planet Name	Radius 10^8 cm	Volume 10^{26} cm^3	Mass 10^{27} gm	Density gm/cm^3	Corrected density† gm/cm^3
Mercury	2.44	0.61	0.33	5.42	5.4
Venus	6.05	9.3	4.9	5.25	4.3
Earth	6.38	10.9	6.0	5.52	4.3
Mars	3.40	1.6	0.64	3.94	3.7
Jupiter	71.90	15,560	1900	1.31	<1.3
Saturn	60.20	9130	570	0.69	<0.7
Uranus	25.40	690	88	1.31	<1.3
Neptune	24.75	635	103	1.67	<1.7
Pluto	1.6	0.17	?	?	?

*The mass of the Sun is 1.99×10^{33}gm, 1000 times the mass of Jupiter.

†Density a planet would have in the absence of gravitational squeezing.

The Solar System

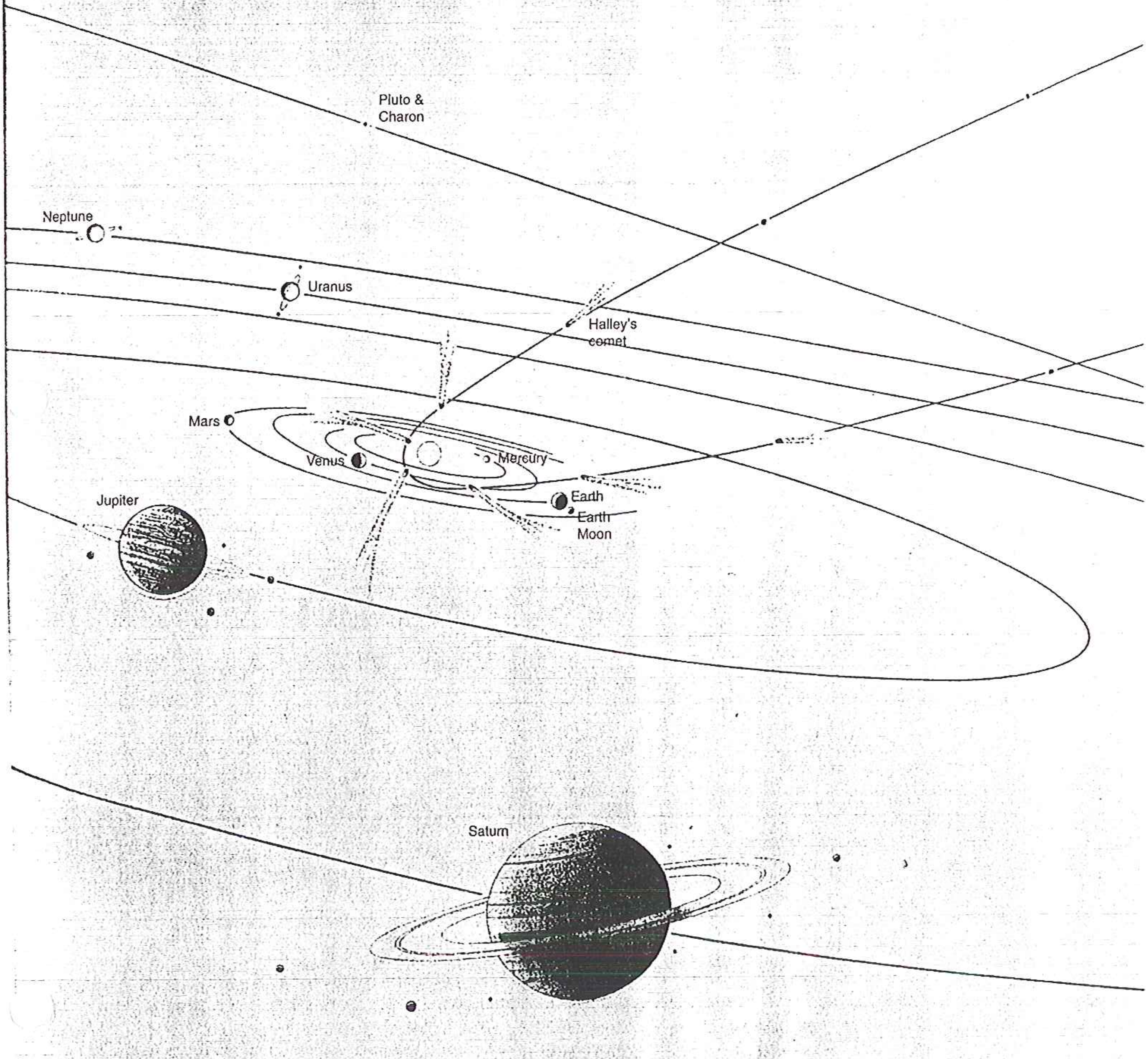


TABLE 1.1

Physical Characteristics of the Planets and Their Major Satellites

Planetary Body	Semi-Major Axis (AU for Planets, 10^3km for Satellites)	Orbital Period (Days or Years (y))	Rotation Period (Days)	Density (g/cm^3)	Diameter (km)	Surface Composition	Atmosphere Composition
Mercury	0.387	87.97	58.65	5.44	4,800	basaltic	Na (thin)
Venus	0.723	224.7	243.0 R	5.25	12,104	basaltic	CO_2
Earth	1.000	365.26	1.00	5.52	12,756	basaltic & H_2O	$N_2 + O_2$
Moon	384	27.3	27.3	3.34	3,476	basaltic	None
Mars	1.524	686.98	1.03	3.93	6,787	basaltic	CO_2
Largest Asteroids							
Vesta	2.362	3.63 y	0.22	2.9	520	basaltic	None
Ceres	2.768	4.61 y	0.38	?	932	DCS	None
Pallas	2.773	4.62 y	0.33	?	533	D S	None
Jupiter	5.203	11.86 y	0.41	1.3	143,800		H_2 and He
Io	422	1.77	1.77	3.50	3,640	S compounds	SO_2 (thin)
Europa	671	3.55	3.55	3.03	3,130	water ice	None
Ganymede	1071	7.15	7.15	1.93	5,280	water ice D	None
Callisto	1884	16.69	16.69	1.79	4,840	water ice D	None
Saturn	9.54	29.46 y	0.43	0.69	120,660		H_2 and He
Mimas	186	0.94	0.94	1.12	392	water ice	None
Enceladus	238	1.37	1.37	1.00	500	water ice	None
Tethys	295	1.89	1.89	1.00	1,060	water ice	None
Dione	377	2.74	2.74	1.49	1,120	water ice	None
Rhea	527	4.52	4.52	1.24	1,530	water ice	None
Titan	1222	15.94	15.9	1.88	5,150	water ice C	N_2
Hyperion	1484	21.3	?	?	250	water ice	None
Iapetus	3562	79.33	79.33	1.03	1,436	H_2O ice DCS	None
Phoebe	12930	550.4 R	0.4	?	220	H_2O ice DC?	None
Uranus	19.18	84.01 y	0.72	1.28	51,120		H_2 and He
Miranda	130	1.41	1.41	1.35	470	water ice	None
Ariel	191	2.52	2.52	1.66	1,150	water ice	None
Umbriel	266	4.14	4.14	1.51	1,170	water ice	None
Titania	438	8.70	8.70	1.68	1,580	water ice	None
Oberon	586	13.46	13.46	1.58	1,520	water ice	None
Neptune	30.07	164.79 y	0.73	1.64	49,560		H_2 and He
Triton	355	5.88 R	5.88	2.01	2,700	N_2 and CH_4 ice	N_2, CH_4
Proteus	118	1.12	1.12	?	400	D H_2O ice	None
Nereid	5562	359.9	?	?	340	D H_2O ice	None
Pluto	39.44	247.7	6.4	2.06	2,284	nitrogen ice	N_2
Charon	17	6.39	6.4	2.06	1,192	H_2O ice	None

D = dark materials; silicates, carbonaceous, or methane
C = carbonaceous materials
S = silicates
R = retrograde orbit

PERIODIC TABLE OF THE ELEMENTS

Mn — Chemical Symbol
25 — Atomic Number
Manganese — Element Name
54.94 — Atomic Weight

Strong tendency for outermost electrons to be lost to make full outer shell

Transition elements: valence electrons not in outer shell

Tendency to fill outer electron shell by electron sharing and gain or loss of electrons

Strong tendency to gain electrons to make full outer shell

Noble gases: outer shells filled; no tendency to gain or lose electrons

H 1 Hydrogen 1.01																	**He** 2 Helium 4.00
Li 3 Lithium 6.94	**Be** 4 Berylium 9.01											**B** 5 Boron 10.81	**C** 6 Carbon 12.01	**N** 7 Nitrogen 14.01	**O** 8 Oxygen 16.00	**F** 9 Fluorine 19.00	**Ne** 10 Neon 20.18
Na 11 Sodium 22.99	**Mg** 12 Magnesium 24.31											**Al** 13 Aluminum 26.98	**Si** 14 Silicon 28.09	**P** 15 Phosphorus 30.97	**S** 16 Sulfur 32.06	**Cl** 17 Chlorine 35.45	**Ar** 18 Argon 39.95
K 19 Potassium 39.10	**Ca** 20 Calcium 40.08	**Sc** 21 Scandium 44.96	**Ti** 22 Titanium 47.90	**V** 23 Vanadium 50.94	**Cr** 24 Chromium 52.00	**Mn** 25 Manganese 54.94	**Fe** 26 Iron 55.85	**Co** 27 Cobalt 58.93	**Ni** 28 Nickel 58.70	**Cu** 29 Copper 63.55	**Zn** 30 Zinc 65.38	**Ga** 31 Gallium 69.72	**Ge** 32 Germanium 72.59	**As** 33 Arsenic 74.92	**Se** 34 Selenium 78.96	**Br** 35 Bromine 79.90	**Kr** 36 Krypton 83.80
Rb 37 Rubidium 85.47	**Sr** 38 Strontium 87.62	**Y** 39 Yttrium 88.91	**Zr** 40 Zirconium 91.22	**Nb** 41 Niobium 92.91	**Mo** 42 Molybdenum 95.94	**Tc** 43 Technetium (98)	**Ru** 44 Ruthenium 101.07	**Rh** 45 Rhodium 102.91	**Pd** 46 Palladium 106.4	**Ag** 47 Silver 107.87	**Cd** 48 Cadmium 112.41	**In** 49 Indium 114.82	**Sn** 50 Tin 118.69	**Sb** 51 Antimony 121.75	**Te** 52 Tellurium 127.60	**I** 53 Iodine 126.90	**Xe** 54 Xenon 131.30
Cs 55 Cesium 132.91	**Ba** 56 Barium 137.33	* (see below)	**Hf** 72 Hafnium 178.49	**Ta** 73 Tantalum 180.95	**W** 74 Tungsten 183.85	**Re** 75 Rhenium 186.21	**Os** 76 Osmium 190.2	**Ir** 77 Iridium 192.22	**Pt** 78 Platinum 195.09	**Au** 79 Gold 196.97	**Hg** 80 Mercury 200.59	**Tl** 81 Thallium 204.37	**Pb** 82 Lead 207.2	**Bi** 83 Bismuth 208.98	**Po** 84 Polonium (209)	**At** 85 Astatine (210)	**Rn** 86 Radon (222)
Fr 87 Francium (223)	**Ra** 88 Radium 226.03	** (see below)	**Rf** 104 Rutherfordium 261.11	**Db** 105 Dubnium 262.11	**Sg** 106 Seaborgium (263)	**Bh** 107 Bohrium (262)	**Hs** 108 Hassium (264)	**Mt** 109 Meitnerium (266)									
IA	IIA	IIIB	IVB	VB	VIB	VIIB	VIIIB	VIIIB	VIIIB	IB	IIB	IIIA	IVA	VA	VIA	VIIA	VIIIA

Lanthanide (Rare Earth) Elements

La *57 Lanthanum 138.91	**Ce** 58 Cerium 140.12	**Pr** 59 Praseodymium 140.91	**Nd** 60 Neodymium 144.24	**Pm** 61 Promethium (145)	**Sm** 62 Samarium 150.4	**Eu** 63 Europium 151.96	**Gd** 64 Gadolinium 157.25	**Tb** 65 Terbium 158.93	**Dy** 66 Dysprosium 162.50	**Ho** 67 Holmium 164.93	**Er** 68 Erbium 167.26	**Tm** 69 Thulium 168.93	**Yb** 70 Ytterbium 173.04	**Lu** 71 Lutetium 174.97

Actinide Elements

Ac **89 Actinium 227.03	**Th** 90 Thorium 232.04	**Pa** 91 Protactinium 231.04	**U** 92 Uranium 238.03	**Np** 93 Neptunium 237.05	**Pu** 94 Plutonium (244)	**Am** 95 Americium (243)	**Cm** 96 Curium (247)	**Bk** 97 Berkelium (247)	**Cf** 98 Californium (251)	**Es** 99 Einsteinium (252)	**Fm** 100 Fermium (257)	**Md** 101 Mendelevium (258)	**No** 102 Nobelium (259)	**Lr** 103 Lawrencium (260)

Table 3-4. The relationship between the density of a substance and the average number of nuclear particles in its constituent atoms:

Substance	Formula	Nuclear particles Atom	Density gm/cm^3	Ratio*
Water	H_2O	6.0	1.00	6.0
Calcite	$CaCO_3$	20.0	2.72	7.4
Quartz	SiO_2	20.0	2.65	7.5
Gypsum	$CaSO_42H_2O$	14.3	2.32	6.2
Olivine	Mg_2SiO_4	18.3	3.20	5.7
Hematite	Fe_2O_3	32.0	5.26	6.1
Magnetite	Fe_3O_4	33.1	5.18	6.4
Diamond	C	12.0	3.50	3.4
Iron	Fe	56	7.50	7.5
Gold	Au	197	17.10	11.6

*Nuclear particles per atom divided by density. Were the relationship between the average number of nuclear particles per atom and density perfect, this ratio would be exactly the same for all ten compounds. For 8 of the 10 compounds the range is small (5.7 to 7.5). For gold and for diamond the deviation from the mean for the other 8 is sizable (a factor of about 2).

In order of decreasing silica

$\rho \sim 2700$ kg/m^3

- granitic ~ 70% SiO_2
  e.g. Sierra Nevada batholith
- andesitic ~ 60% SiO_2
  island arcs

$\rho \sim 2900$ kg/m^3

- mafic or basaltic ~ 50% SiO_2
  by far most common
  sea floor — oceanic crust - 5 km thick
  Hawaii
  cover surface of other terrestrial planets
  Moon, Venus, Mars

Mineralogy shown in J&R Fig. 3.10

- granite 30% quartz SiO_2
  35% K feldspar $KAlSi_3O_8$
  25% Na feldspar (plagioclase) $NaAlSi_3O_8$
  10% mica & amphibole
- basalt 35% olivine $(Mg, Fe)_2SiO_4$
  35% Ca plagioclase $CaAl_2Si_2O$
  30% pyroxene $Ca(Mg, Fe, Al)[(Si, Al)O_3]_2$

$MgSiO_3$
enstatite

Finally, a fourth type, rare but important for our purposes

- ultramafic — samples of rock from mantle

  $\rho = 3300$ kg/m^3

  up to 95% olivine $(Mg,Fe)_2SiO_4$ with some pyroxene

  pure olivine rock — dunite
  with some pyroxene — peridotite

2. Sedimentary rocks

clastic sediments

- shale (mudrock) — by far most common — continental margins, river deltas, basins

- sandstone
- conglomerate

$\rho = 2500 - 2700$ kg/m^3

Also chemical sediments

- limestone — biogenic — $CaCO_3$ shells
- chert — biogenic SiO_2 — diatoms & radiolaria

Key minerals — already studied in lab

quartz SiO_2

K feldspar $KAlSi_3O_8$ orthoclase

Na feldspar $NaAlSi_3O_8$ } plagioclase

Ca feldspar $CaAl_2Si_2O_8$ }

olivine $(Mg, Fe)_2SiO_4$

pyroxene $Ca(Mg, Fe, Al)[(Si, Al)O_3]_2$

e.g. enstatite $MgSiO_3$

amphibole e.g. anthophyllite

$Mg_7Si_8O_{22}(OH)_2$

↗ hydrated

blackboard

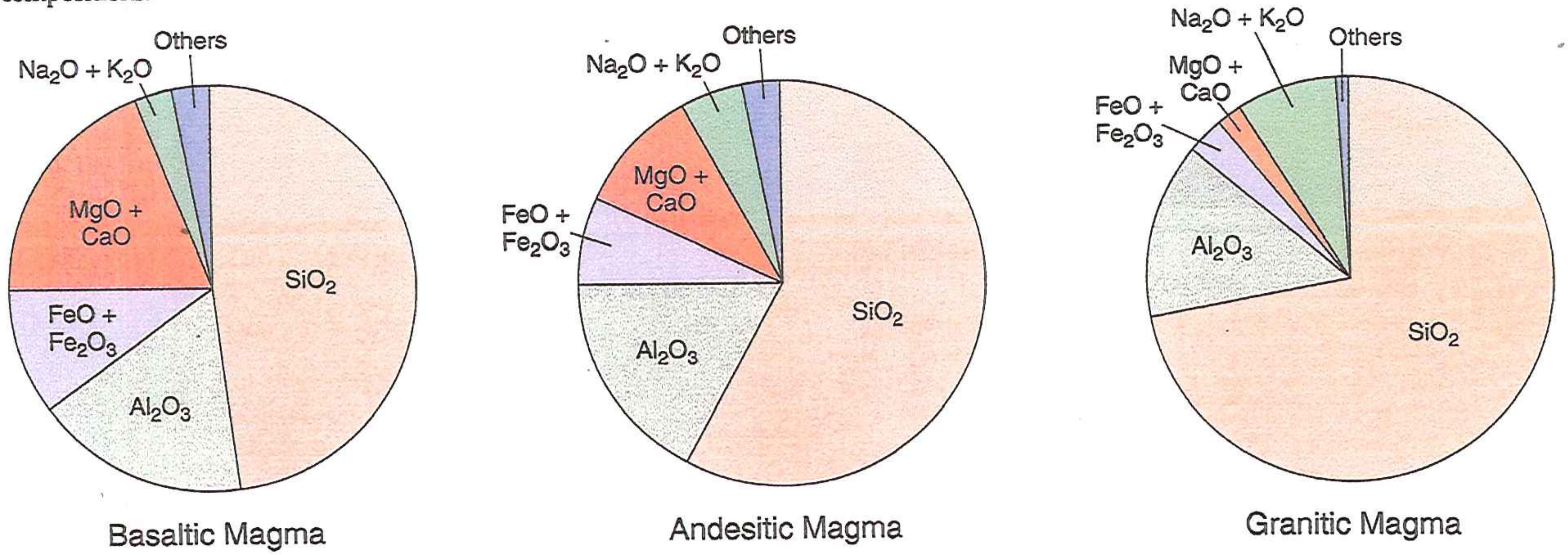
Write as

olivine Mg_2SiO_4 ↙ Fe substitutes

pyroxene $MgSiO_3$ ↖ Al substitutes

↑ Ca, Fe, Al substitutes

FIGURE 3.1 As shown in these pie diagrams, basaltic magmas contain much less silica and more iron, magnesium, and calcium than do granitic magmas. Andesitic magmas have intermediate compositions.



Basaltic Magma

Andesitic Magma

Granitic Magma

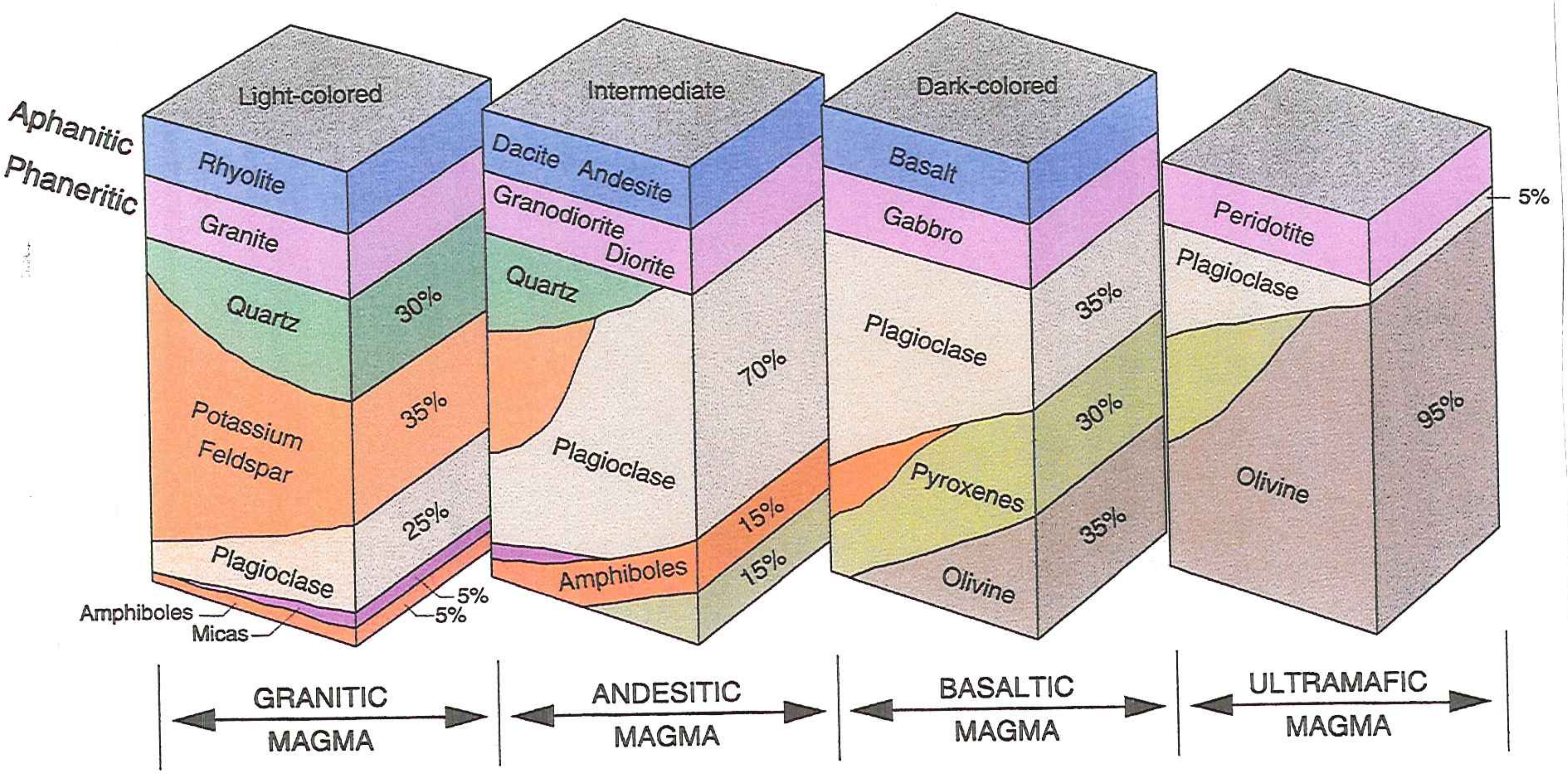


FIGURE 3.10 A simple classification of igneous rocks, based on rough proportions of key minerals. Magma compositions, from granitic to ultramafic, are shown across the bottom of this diagram; textures (aphanitic or phaneritic) are indicated at the top. The broad divisions shown here are meant to be suggestive, not precise. In nature, the boundaries between granitic and andesitic rocks, or between andesitic and basaltic ones, are not sharp.

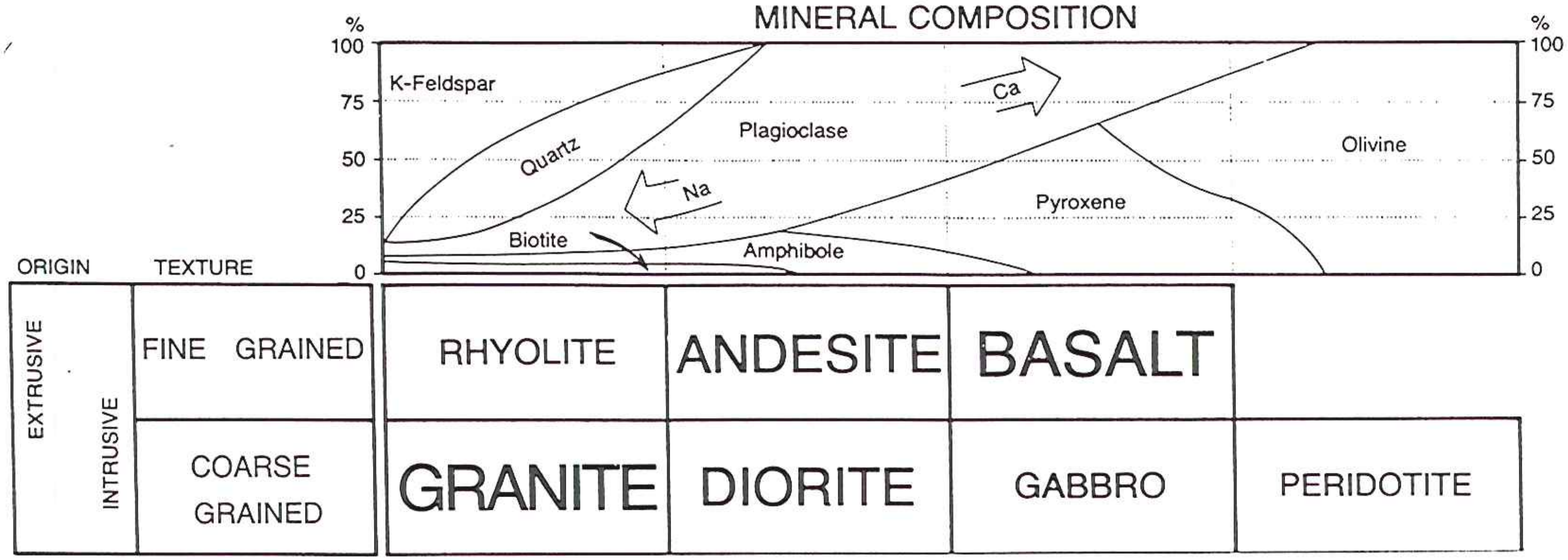


Figure 2.16

The classification of igneous rocks is based on composition and texture. Granite is the most abundant intrusive rock, whereas basalt is the most abundant extrusive rock on Earth.

- evaporites (salt NaCl
  gypsum $CaSO_4 \cdot 2H_2O$)

only found on ⊕ and Mars

3. <u>metamorphic rocks</u>

rocks that have recrystallized upon being subjected to high PT conditions

- metasediments
- metaigneous
  (metabasalt , etc.)

~~insert page 5½ here~~

None of these surface rocks — not even the ultramafic rocks — represent the primordial material from which the ⊕ formed by accretion.

We must turn to other sources to find such primordial samples — meteorites.

Classic meteorite is essentially pure iron — e.g. Meteor Crater, near Winslow, Arizona

FIG. 3-5 The Arizona Meteor Crater, a well-preserved and dramatically exposed impact structure on Earth. Diameter is 1.2 km; depth to surface of alluvial fill inside it, ~100 m. Photograph courtesy of D. J. Roddy and the U.S. Geological Survey.

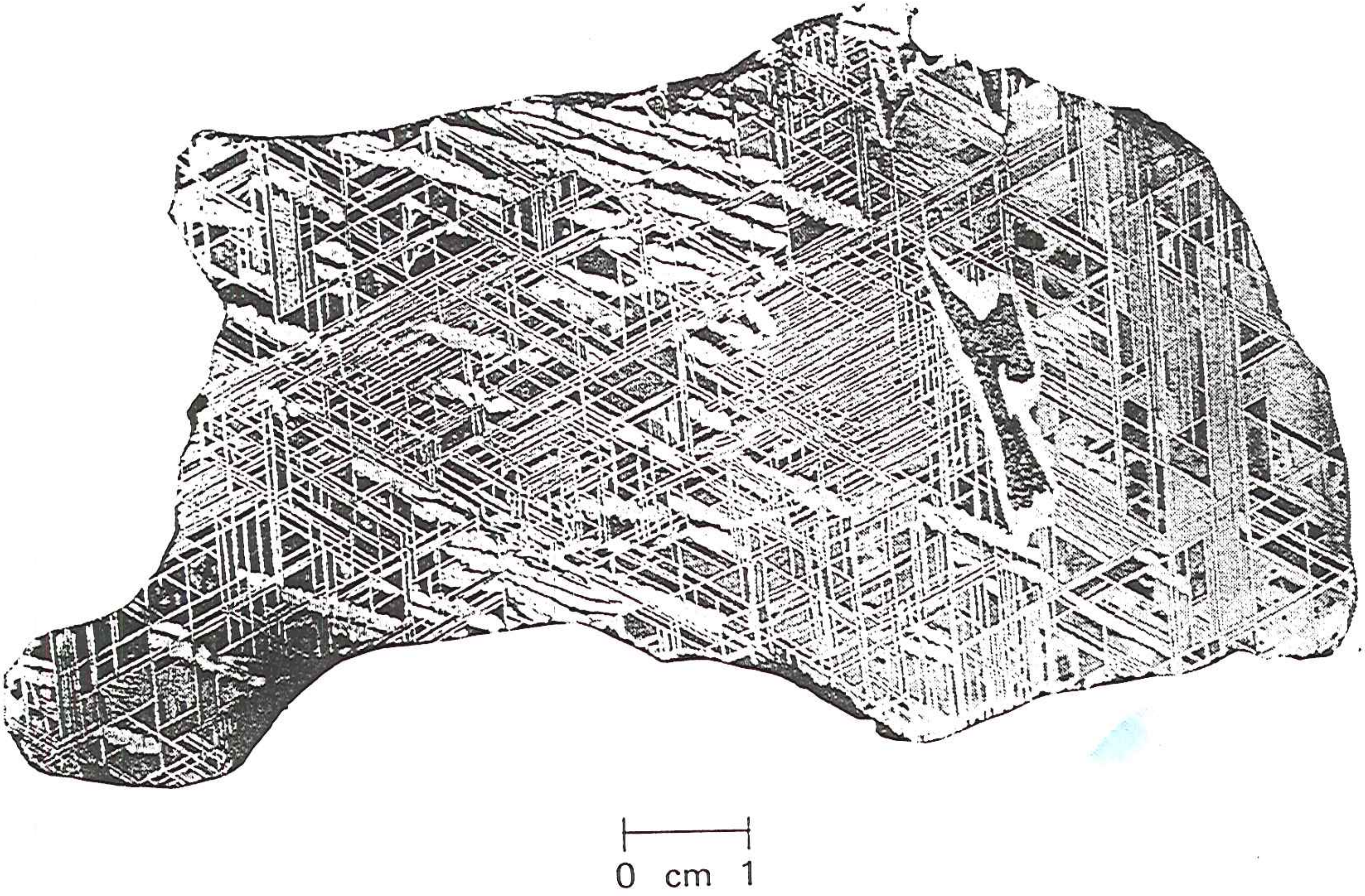


FIG. 5-10 Polished and etched slab of the Edmonton (Kentucky) octahedrite, illustrating the Widmanstätten structure. Narrow parallel bands are cross sections of plates of the low-nickel alloy kamacite; the spaces between them are filled with higher-nickel alloys. Smithsonian Astrophysical Observatory photograph.

Most meteorites are pieces of rock from the asteroid belt between Mars & Jupiter — knocked into ⊕-crossing orbit by asteroidal collisions.

A few very rare meteorites are from the Moon or Mars

Comets, in contrast to meteorites, have very elliptical orbits — compositionally they are dirty snowballs.

Meteorites classified into observed falls (with trackable orbits) and finds

About 150 with mass > ½ kg fall every year — most are not recovered.

About 1000 recovered falls

About 10–100 times as many documented finds — most from Antarctica — easy to see against ice

Special meteorite-collecting field parties every Antarctic summer — find ~1000 in a season.

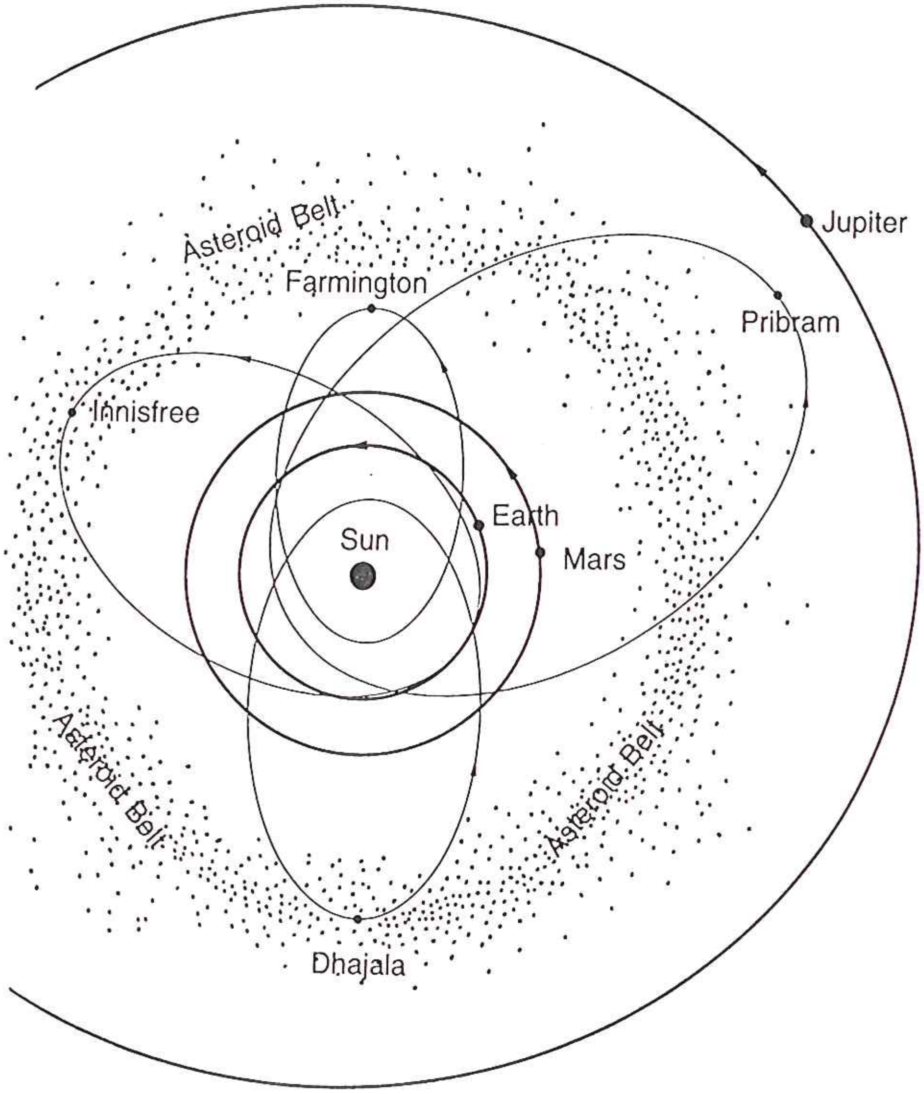


Figure 3.6

Most asteroids have circular orbits that take them between the orbits of Mars and Jupiter in an area known as the asteroid belt. Many other asteroids orbit outside of the main belt—some across Earth's orbit. Asteroids were probably the source of the bodies that impacted Earth and other planets during their early histories.

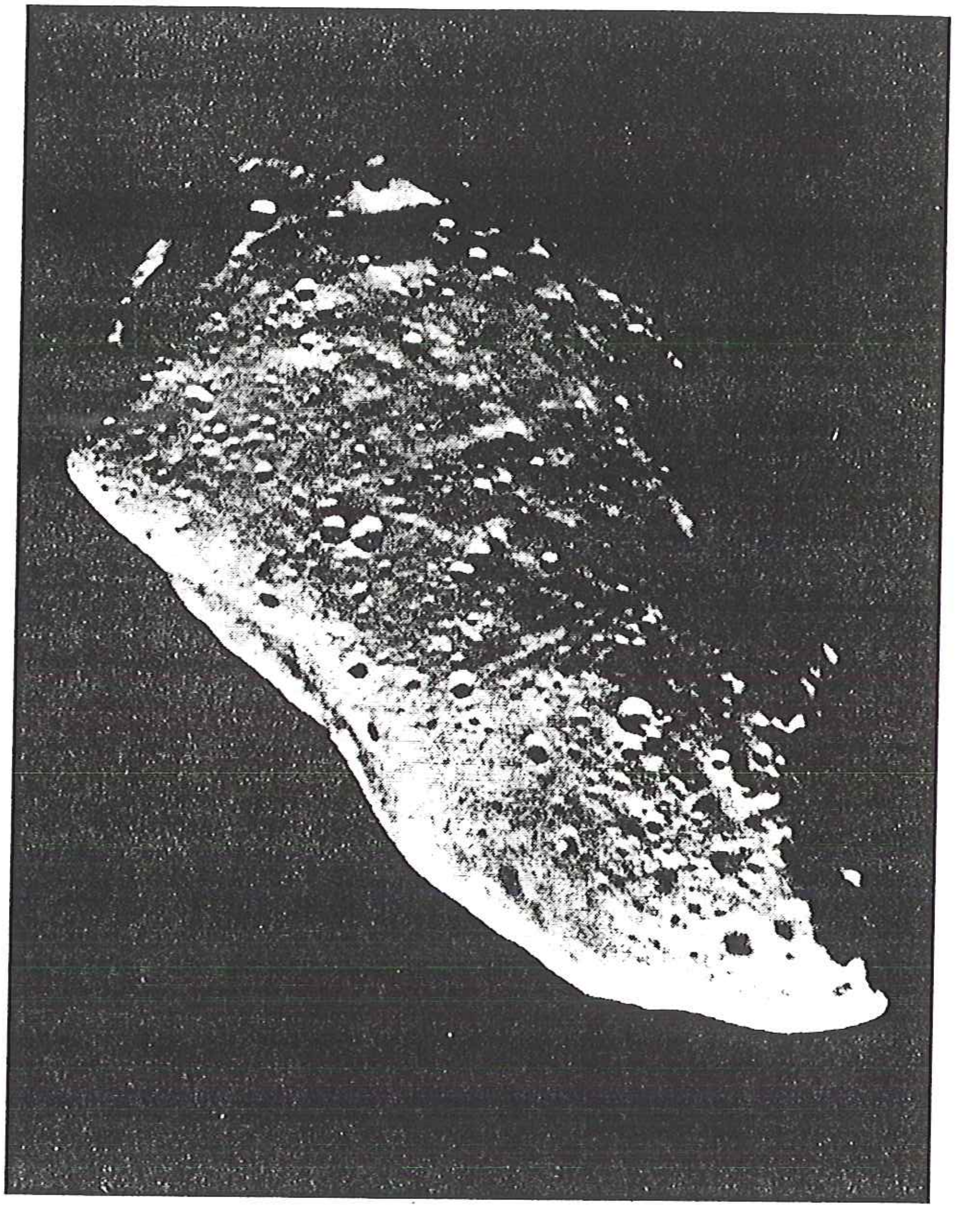


Figure 3.8

Gaspra, the first asteroid to be photographed by a passing spacecraft, shows many of the characteristics we expect for many other asteroids.

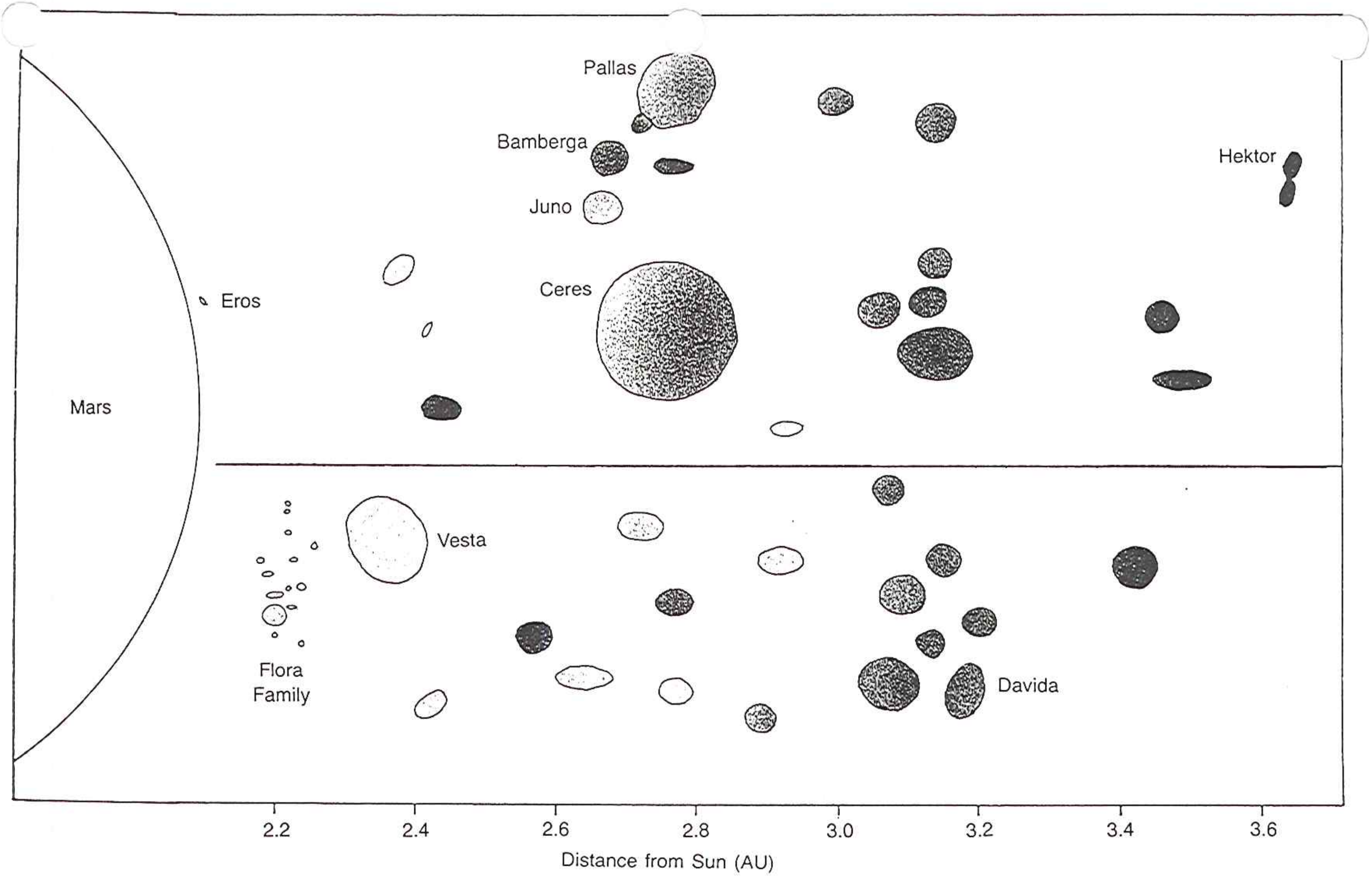


Figure 3.7

Asteroids come in a variety of sizes, shapes, and colors. This diagram illustrates these properties along with relative distances from the Sun and orbital inclinations for all 33 asteroids larger than 200 km in diameter. To give an indication of the size of smaller asteroids, all members of the Flora family are also shown. Thousands of others have comparable sizes. The limb of Mars is shown for comparison. Asteroids near the center of the diagram revolve in near circular, noninclined orbits; those near the top or bottom of the diagram have elliptical and/or inclined orbits. Asteroids become detectably darker with distance from the Sun. This is thought to correlate with the proportion of dark carbonaceous materials at the asteroids' surfaces.

2004 U.S. LIFE SCIENCES SALARY SURVEY, PAGE 1829

18 June 2004

Science

Vol. 304 No. 5678
Pages 1701–1856 $10

Stardust
AT COMET WILD 2

AAAS

TABLE 3.2
Major Meteorite Types

Type	Abundance (percent)	Composition
Stony Meteorites	94	
Chondrites	86	
Ordinary	82	Metamorphosed chondrites
Carbonaceous	4	Carbon- and volatile-rich, undifferentiated
Achondrites	8	Igneous textures, differentiated
Stony-Iron Meteorites	1	Silicate-metal mixtures, differentiated
Iron Meteorites	5	Iron metal, differentiated

Abundances are percentages of each type of meteorite among all meteorites seen to fall on Earth.

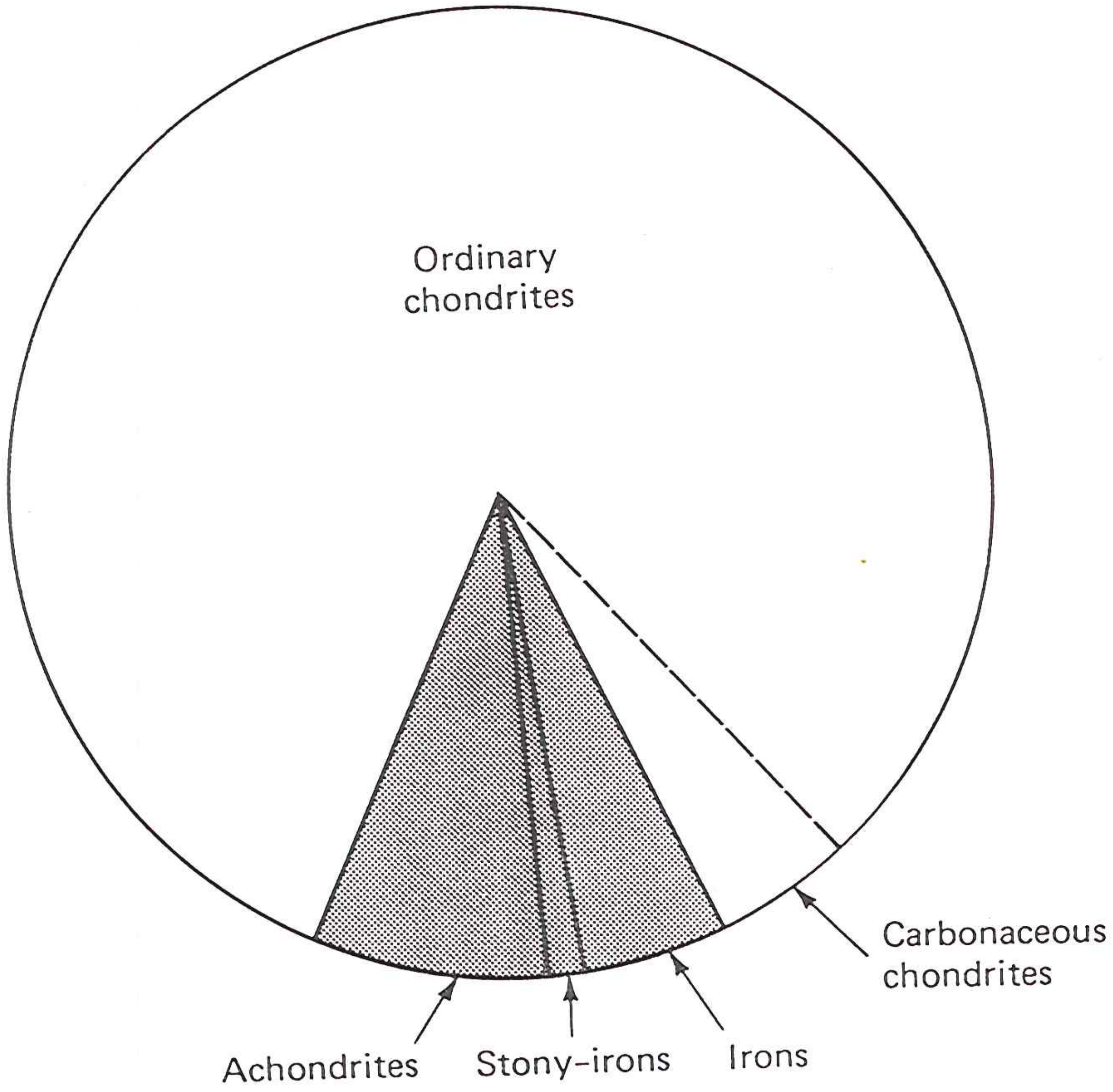
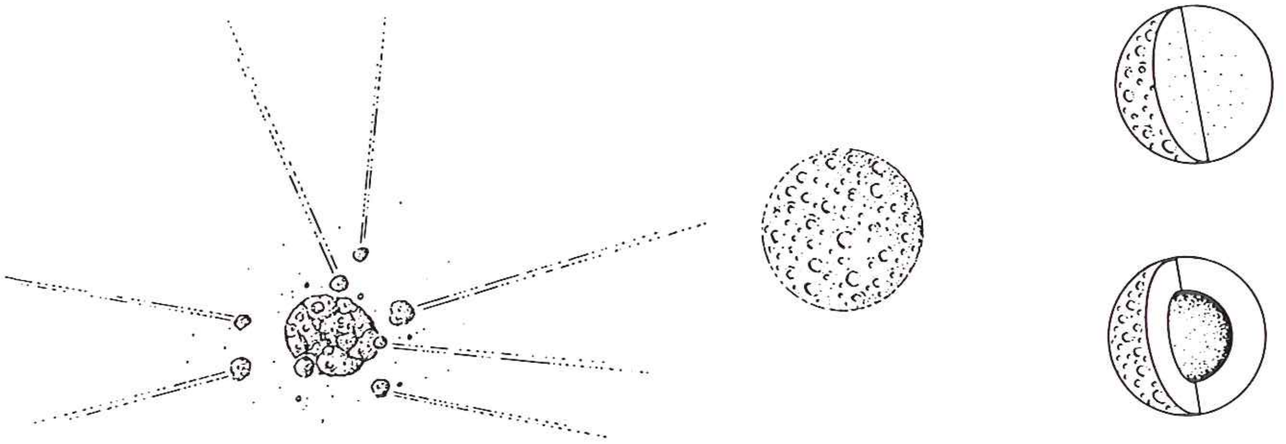
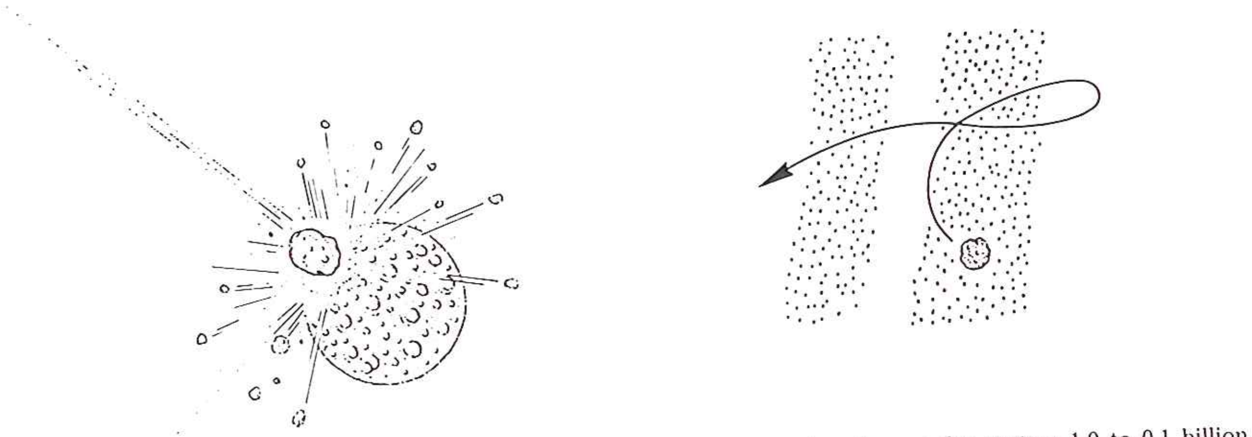


FIG. 5-2 Proportions of major types among meteorites observed to fall to Earth (falls). The enstatite chondrite subtype is included with "ordinary chondrites."



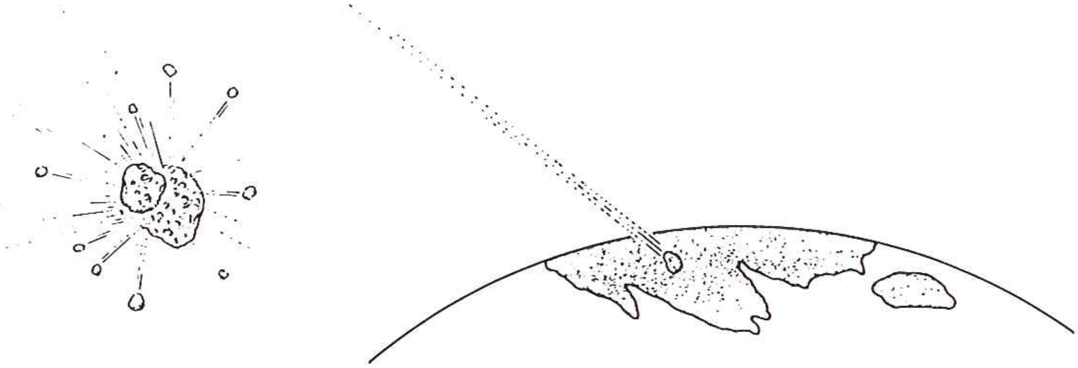
(A) Accretion of a planetesimal 4.6 billion years ago.

(B) Differentiation of interior to form iron core or metamorphosed rock 4.4 to 4.6 billion years ago.



(C) Collision and fragmentation 1 to 0.1 billion years ago.

(D) Deflection into inner solar system 1.0 to 0.1 billion years ago.



(E) Further collisions less that 0.1 billion years ago.

(F) Impact on Earth today.

Figure 3.15

The evolution of asteroids as meteorite parent bodies is summarized in this diagram. The events that led to the delivery of fragments of the asteroids to Earth and other inner planets are emphasized.

Gives a representative sampling of meteorite compositions

~ 94 % stony
~ 6 % iron or stony-iron

Stony meteorites further subdivided

chondrites 86 %
  ordinary (metamorphosed) 82 %
  carbonaceous 4 %
achondrites 8 %
  ↑ look like terrestrial basalts

Why are meteorite compositions so variable? Because of their complicated histories — see Fig. 3.15 from Christiansen & Hamblin

Most interesting are the <u>carbonaceous chondrites</u>:

contain mm-sized <u>chondrules</u> that are believed to be samples of <u>undifferentiated solar nebula</u>: the material out of which the ⊕ and other planets condensed.

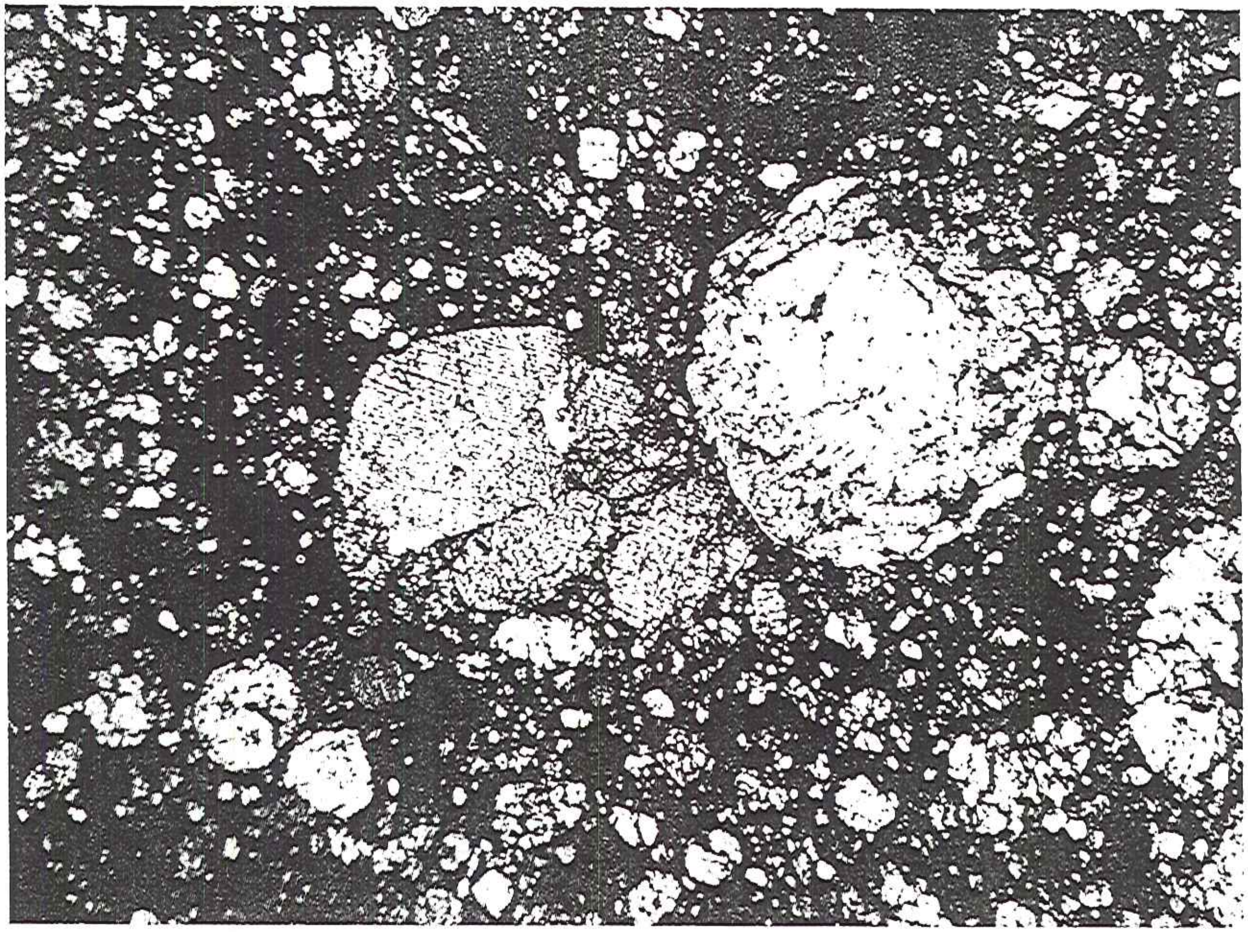
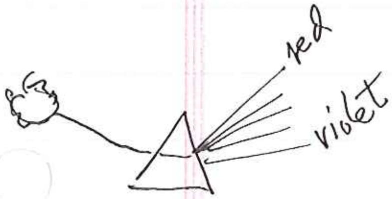


Figure 3-1. **Photomicrograph of chondrules from a stony meteorite:** The chondrules are about 1 millimeter in diameter.

Why do we think this?

- presence of minerals that are unstable above 100°C — have never been heated above this temperature by radiocative heating or impact heating

- presence of volatiles, including carbon (reason for name) also attests to lack of differentiation



sketch solar spectrum

- agreement with composition of solar atmosphere — determined by presence of Fraunhofer diffraction lines in the solar spectrum (rainbow)

red green blue indigo violet

| | |
blackness of line ⇒ amount

This striking correlation seen in Broecker Fig. 3-2 : for elements of low to moderate volatility

$Si \equiv 1$

Most abundant : Si, Fe, Mg, O (not shown)
Least abundant : Thulium, Thorium, ...
seven orders of magnitude less

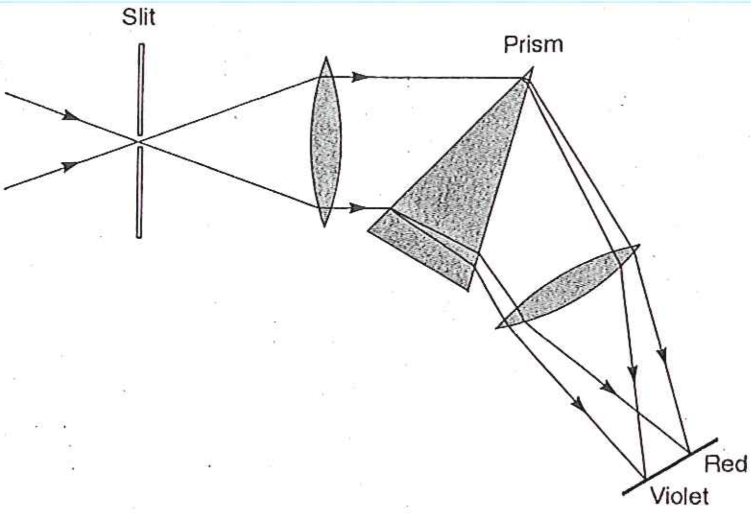


Figure 3.8. How a spectrograph operates. The light enters from the left through a small slit and is concentrated, as in Newton's experiment (shown in Figure 3.1), by a collecting lens and projected onto a prism, which disperses the rays of light into different directions according to their colour. A second lens collects the light dispersed by the prism and produces a band of colour, running from violet to red. If this is observed from the rear with an eyepiece, the instrument is a spectroscope, but if the spectrum is allowed to fall on a photographic plate, it becomes a spectrograph

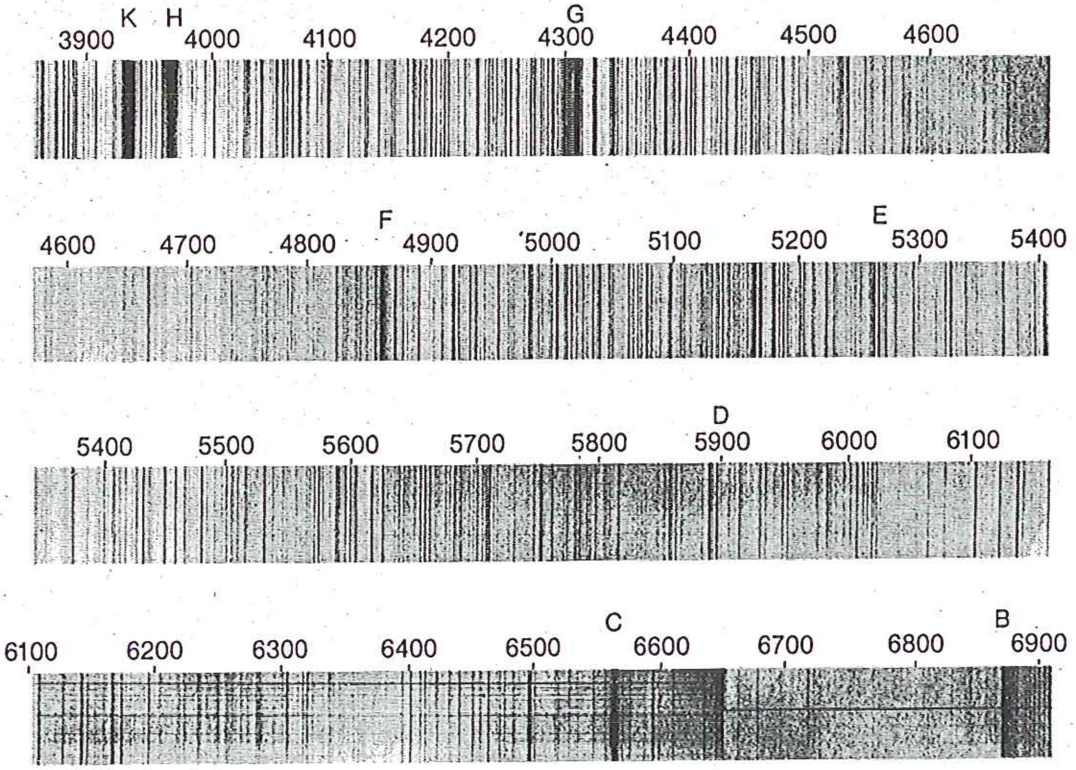


Figure 3.9. The spectrum of the disk of the Sun, divided into four sections, stretching from the violet (top left) to red (bottom right). The continuous spectrum is crossed by thousands of dark absorption lines. The figures are the wavelengths in ten-millionths of a millimetre. The letters are Fraunhofer's designation of spectral lines. The K line in the violet is a calcium line, which is discussed on p. 94. The C line in the red is the hydrogen-alpha line

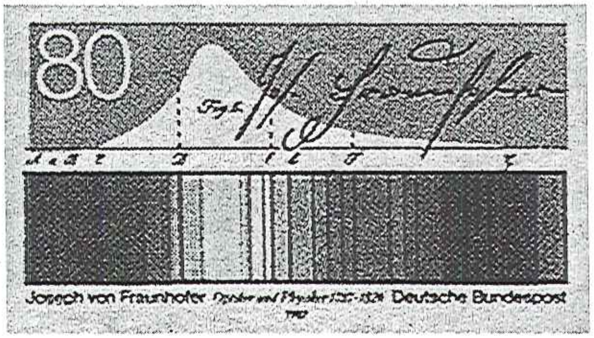


FIGURE 3.3 Fraunhofer's original spectrum, shown on a German postage stamp. The spectrum is crossed by the dark absorption lines that Fraunhofer accurately mapped in the Sun's spectrum.

PLATE 41. Violet region of the Sun's spectrum at the centre (above) and at the limb (below) from 4220Å to 4325Å (Arcetri).

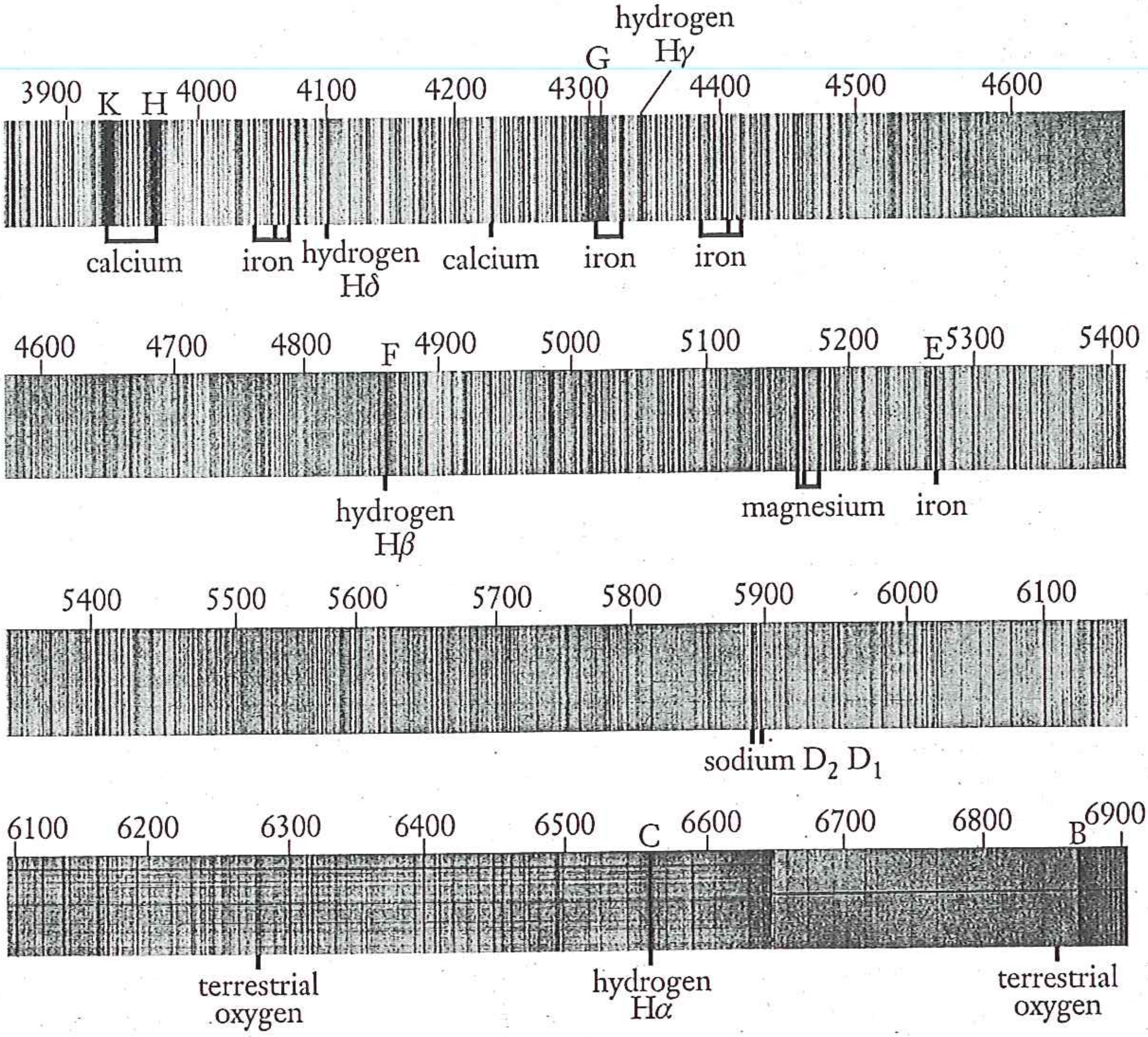
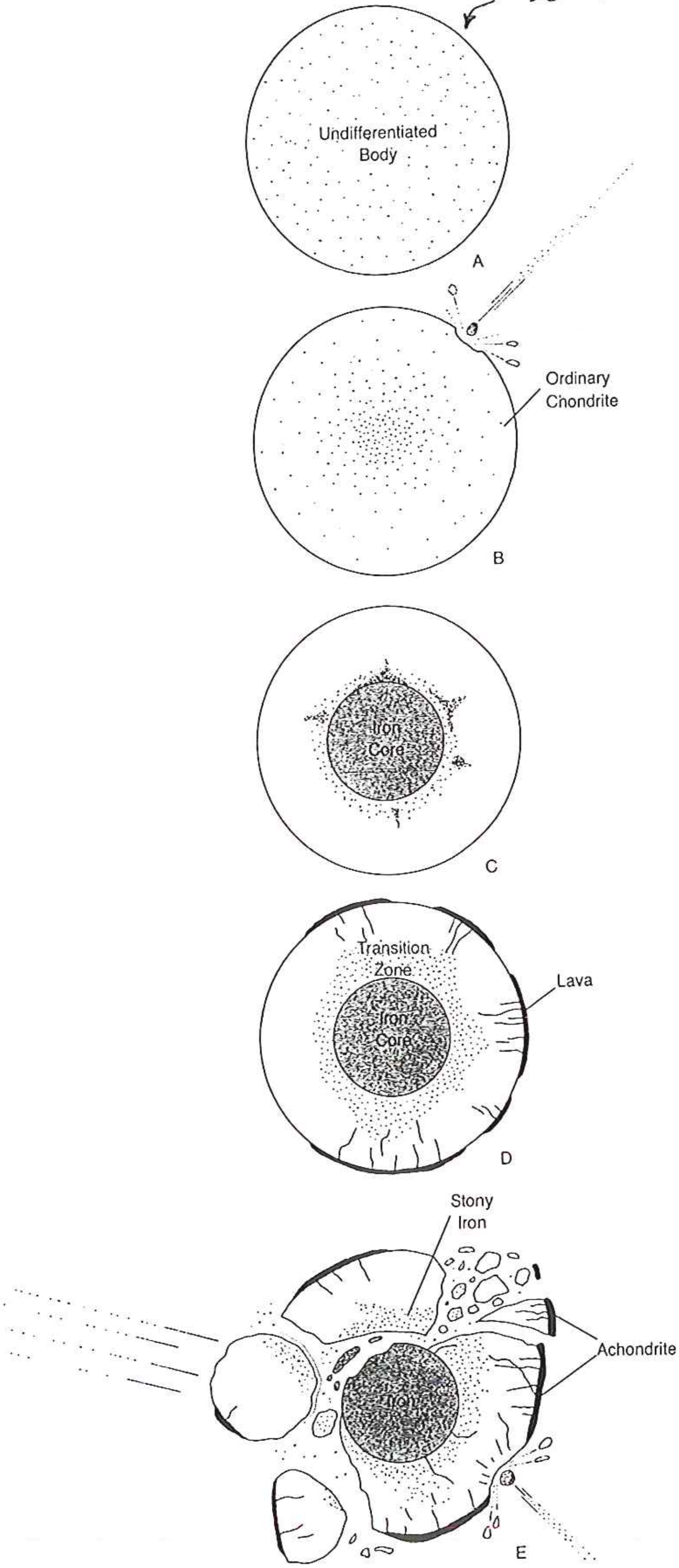


FIGURE 3.4 A solar spectrum with some prominent spectral lines identified. The wavelength is in angstroms; 3,900 angstroms = 400 nanometers.

Table 3.1 The most abundant elements in the solar photosphere.

For each: 1,000,000	atoms of hydrogen
There are: 98,000	atoms of helium
850	atoms of oxygen
400	atoms of carbon
120	atoms of neon
100	atoms of nitrogen
47	atoms of iron
38	atoms of magnesium
35	atoms of silicon
16	atoms of sulfur
4	atoms of argon
3	atoms of aluminum
2	atoms of calcium
2	atoms of sodium
2	atoms of nickel

source of carbonaceous chondrites



(A) An original body of primitive composition (carbonaceous, outer belt; ordinary, inner belt) forms by accretion.

(B) This small planetesimal is heated by short-lived radioactivity to the point of mild metamorphism, driving off some volatiles.

(C) Or perhaps it is heated to the melting point of metallic iron. Dense segregations of iron drains toward the center of the body to form a core or several smaller accumulations of metal.

(D) If heating is intense enough, the silicates in the chondritic interior may melt to produce magma, which erupts at the surface to produce a thin veneer of lava and associated intrusive rocks. Eventually, the asteroid cools as heat is radiated away into space; the core and mantle become solid. Depending on (among other factors) size, composition, and distance from the Sun, for a specific asteroid this differentiation process may have ended at any point of the evolutionary scheme.

(E) Fragmentation of such differentiated or undifferentiated bodies could then produce the spectrum of observed asteroid and meteorite types. In fact, a variety of types could come from one body, as illustrated in (E). Most asteroids were not heated beyond the first or second step, and only a very few have exposed metallic cores stripped of their silicate cloaks.

Figure 3.14
Stages in the evolution of a meteorite parent body.

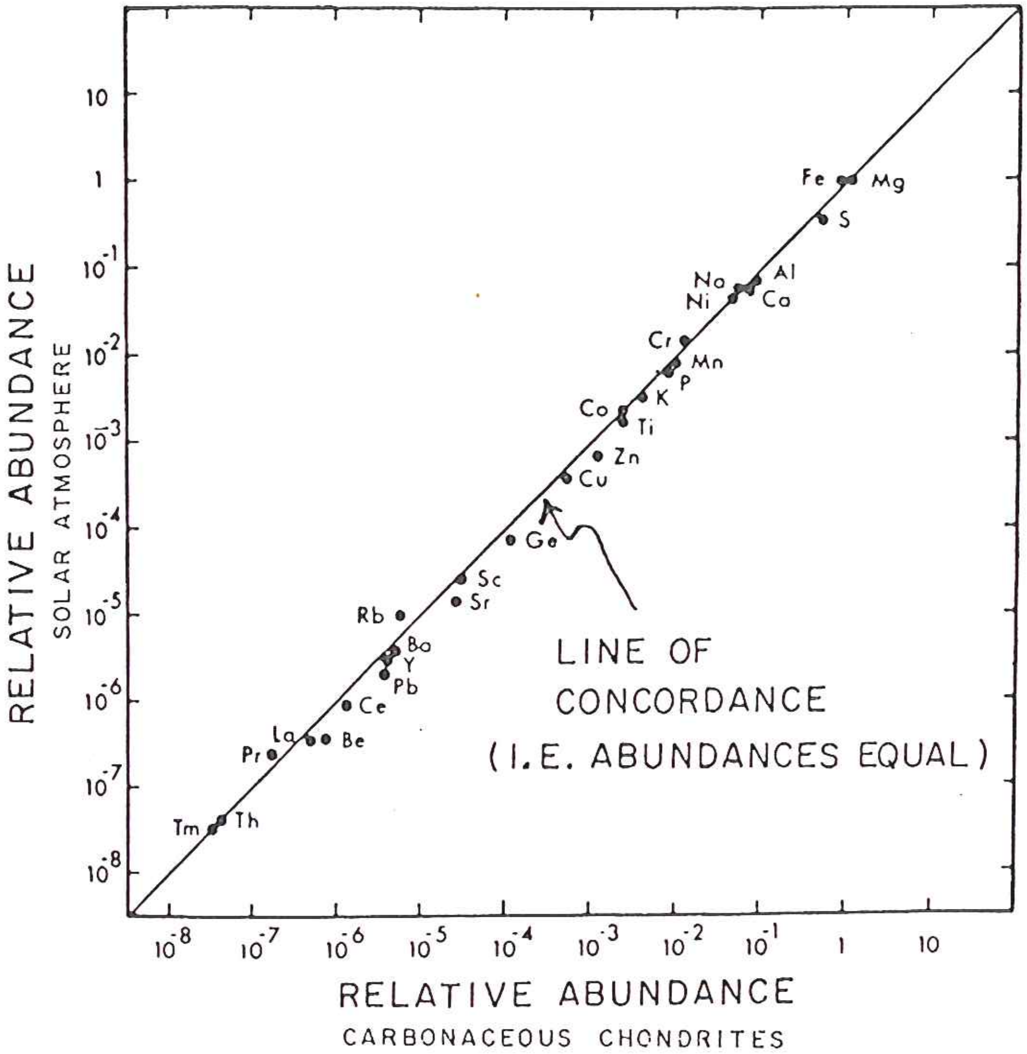


Figure 3-2. Comparison of the relative abundances of elements of low and moderate volatility in the Sun's atmosphere with those in carbonaceous chondrites: Clearly for these elements, carbonaceous chondrites provide a chemically unbiased sample of bulk solar system matter. Because the element silicon is the reference for comparison, it does not appear in the diagram.

Gives us confidence in <u>both</u> determinations

Carbonaceous chondrites are samples of the solar nebula that we can can actually hold in our hand and analyze.

Table 3-5 shows abundances of <u>metals</u> in chondrites:

Mg
Si } 91% of ~~[illegible]~~ total
Fe

Al } next most abundant
Ca

Most important anion is oxygen O

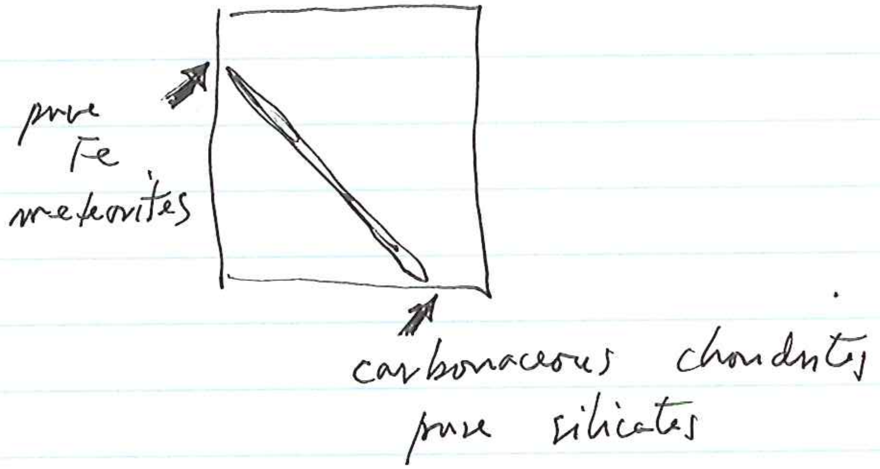
~~Roughly speaking:~~

~~$(Mg, Fe) : Si : O = 2 : 1 : 4$~~

~~Stoichiometry of olivine $(Mg, Fe)_2 SiO_4$~~

Iron Fe in meteorites is not always bound into silicates, i.e. mated with oxygen

Fig 3-3 Broecker plots Fe in metallic Fe + FeS form and Fe bound to oxygen in silicates



Linearity ⇒ total Fe is the same in all meteorites ≈ chondritic composition

But the parent bodies have differentiated.

~~Recipe for construction of the E⊕~~

~~a. form by accretion of planitesimals having the chondritic composition for major elements Fe : Mg : Si = 1 : 1 : 1~~

Table 3-5. Abundances of metallic elements in chondritic meteorites:

	Percent of total metal atoms	
Magnesium (Mg)	32	91%
Silicon (Si)	33	
Iron (Fe)	26	
Aluminum (Al)	2.2	7%
Calcium (Ca)	2.2	
Nickel (Ni)	1.6	
Sodium (Na)	1.3	
Chromium (Cr)	0.40	minor
Potassium (K)	0.25	
Manganese (Mn)	0.20	
Phosphorus (P)	0.19	
Titanium (Ti)	0.12	
Cobalt (Co)	0.10	

[redacted]

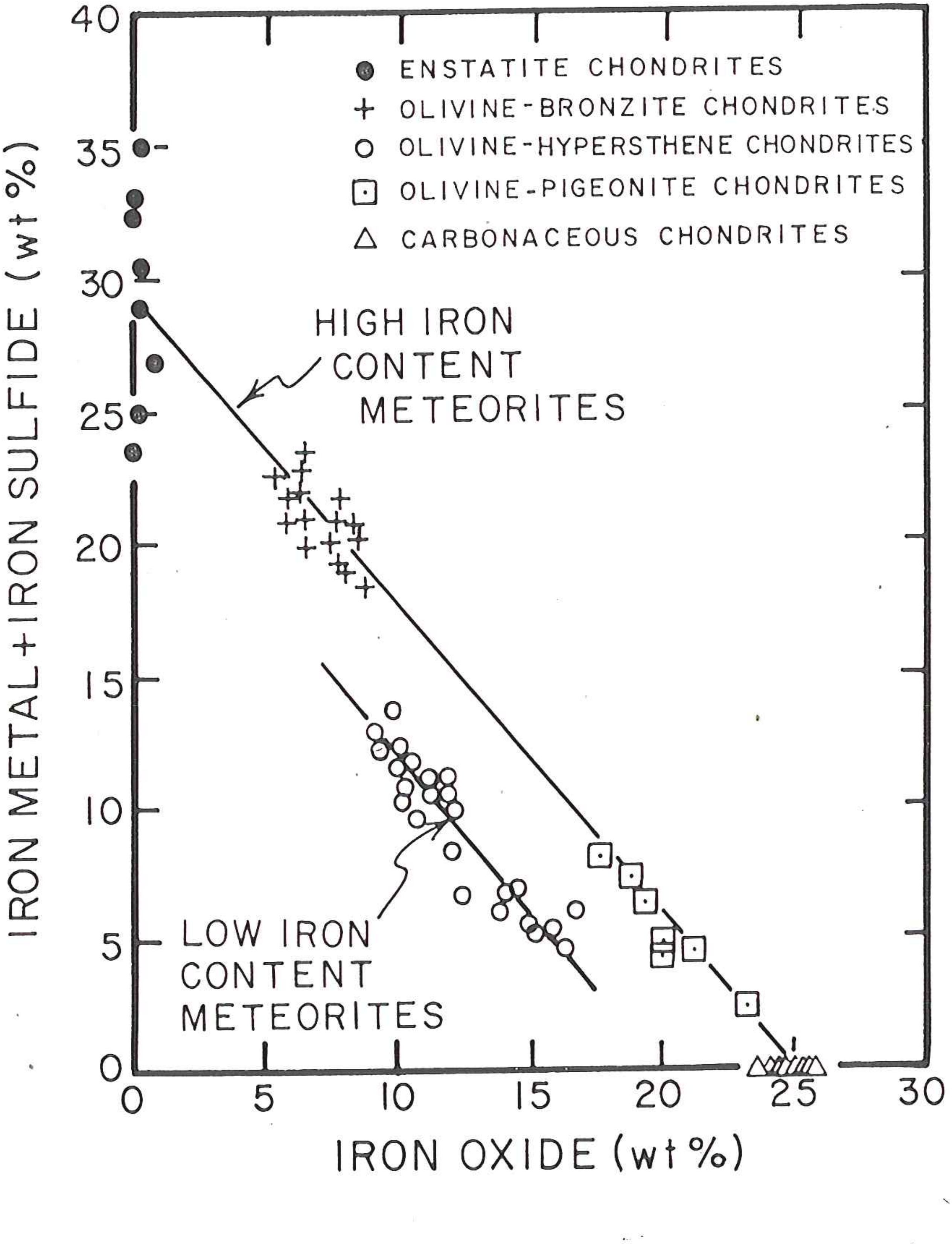


Figure 3-3. Gradations in chemical form of iron in chondrites: The range extends from all oxide in carbonaceous chondrites to all metal (or sulfide) in olivine-hypersthene chondrites. Except for enstatite chondrites, which are depleted in iron, the total iron content remains nearly the same.

Table 2-1 Anderson shows that

$\text{Fe} : \text{Mg} : \text{Si} : \text{O} \approx 1:1:1:20$
cosmic – not counting H & He

↳ in atoms – <u>not</u> weight %

But the oxygen is volatile and easily lost

Ratios in bulk ⊕ are more like

$\text{Fe} : \text{Mg} : \text{Si} : \text{O} = 1:1:1:3\tfrac{1}{2}$

<u>Recipe</u> for <u>making</u> <u>the</u> ⊕ :

1. form by accretion of planitesimals having composition

   $\text{Fe} : \text{Mg} : \text{Si} : \text{O} \approx 1:1:1:20$

2. heat due to k.e. of impact + radioactive heating

   - drive off most oxygen
     $\text{Fe} : \text{Mg} : \text{Si} : \text{O} \approx 1:1:1:3\tfrac{1}{2}$
   - form core of molten Fe

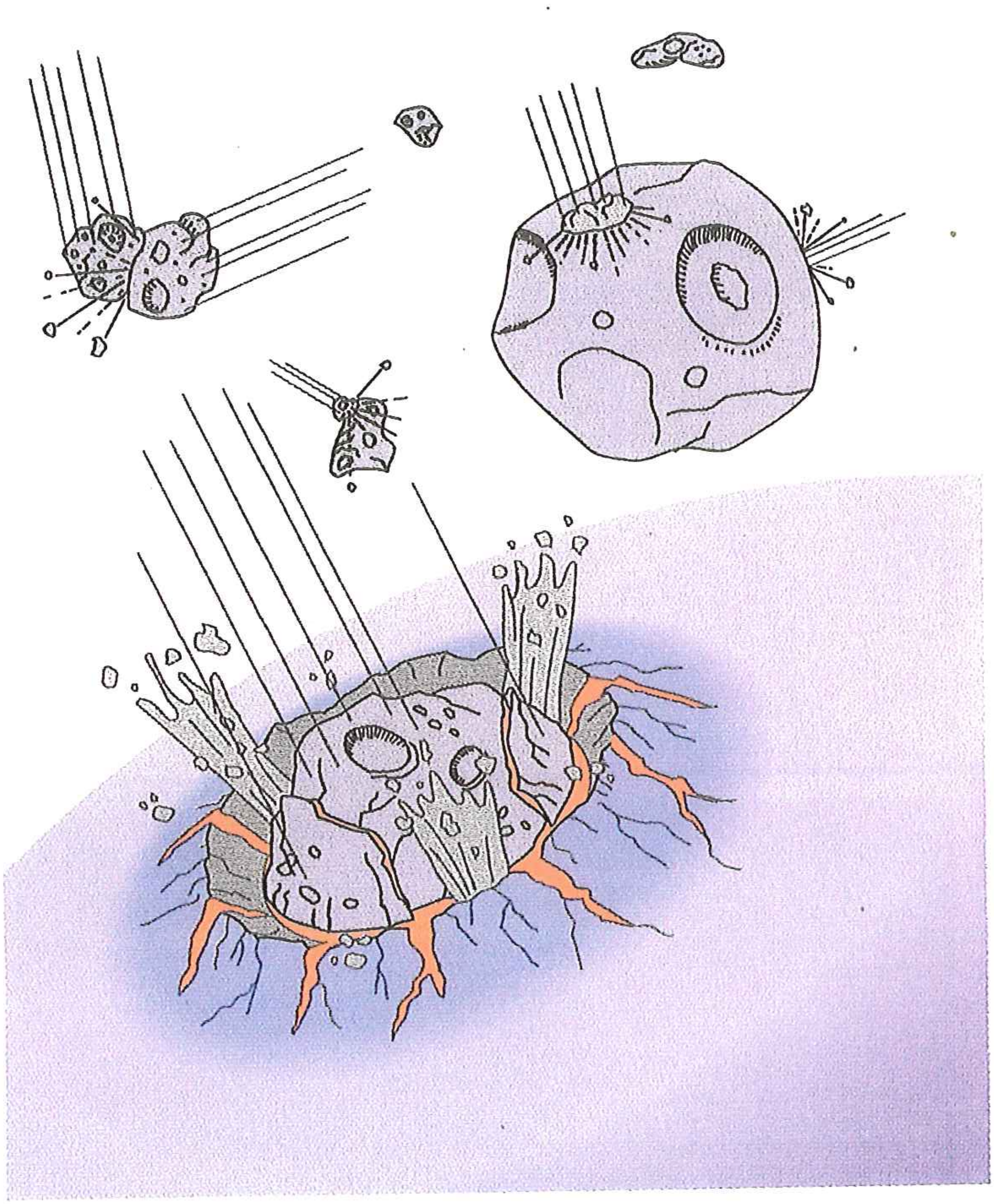


FIGURE 8.1 Four and a half billion years ago a lot of debris, ranging in size from dust to asteroids, still orbited the young Sun. Collisions between these materials, called *planetesimals*, were frequent. As planetesimals gradually accumulated to form larger bodies, these developed stronger gravitational fields and began to attract more particles from nearby space. Each impact added heat to the planet and buried earlier hot rocks under a new blanket of debris. According to some estimates, the Earth could have grown to full size by this process of accretion in a few tens of millions of years. Long before it did, the planet had become so hot that its outer region melted completely to form a global-scale ocean of magma

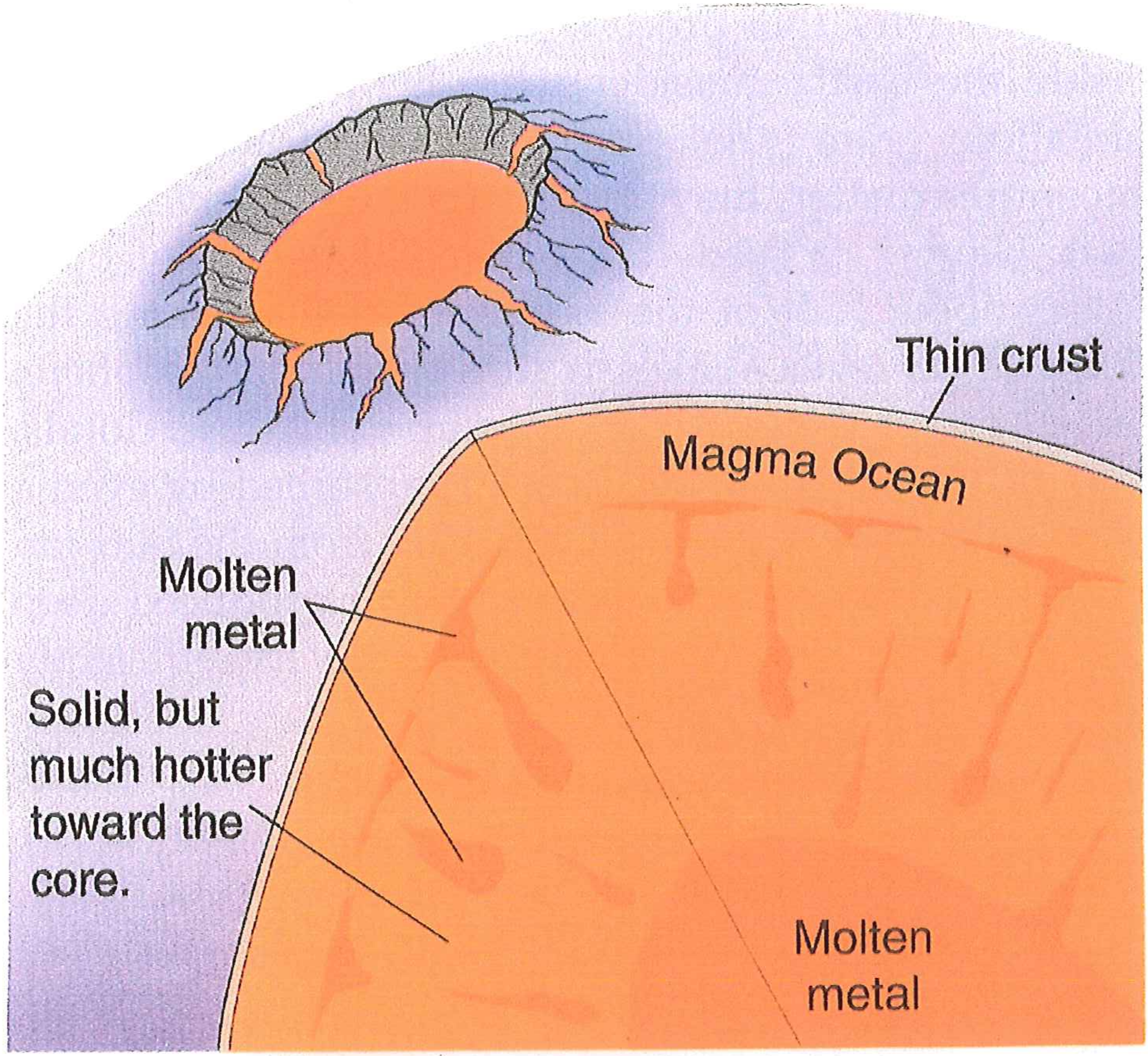


FIGURE 8.2 Liquid iron/nickel, separated from the outer portion of the young Earth, gradually settled toward the center of the planet because of its high density. As it "fell" through the mantle and had to overcome its viscosity, some of the metal's energy of motion was converted into heat. This "iron catastrophe" was a second source of gravitational heating in the young Earth.

TABLE 2-1

Short Table of Cosmic Abundances (Atoms/Si)

Element	**Cameron (1982)**	**Anders and Ebihara (1982)**
O	18.4	20.1
Na	0.06	0.057
Mg	1.06	1.07
Al	0.085	0.0849
Si	1.00	1.00
K	0.0035	0.00377
Ca	0.0625	0.0611
Ti	0.0024	0.0024
Fe	0.90	0.90
Ni	0.0478	0.0493

Most of Fe in $\oplus$ (~90% migrates to core)

Mg : Si ratio $\approx 1 \Rightarrow$ average composition of mantle silicate is

$MgSiO_3$ — enstatite (pyroxene)

not Mg_2SiO_4 — fayalite (olivine)

~~Suppose [illegible] lower mantle~~

In fact upper mantle — above 660 km — is largely

90% Mg since 90% Fe → core

$$\boxed{(Mg_{0.9}\ Fe_{0.1})_2\ SiO_4 \text{ — olivine}}$$

Whereas lower mantle is

$$\boxed{(Mg_{0.9}\ Fe_{0.1})\ SiO_3 \text{ — pyroxene, but in perovskite structure}}$$

Suppose we construct a two-layer $\oplus$ made of these two ingredients

~~[illegible] iron meteorite~~

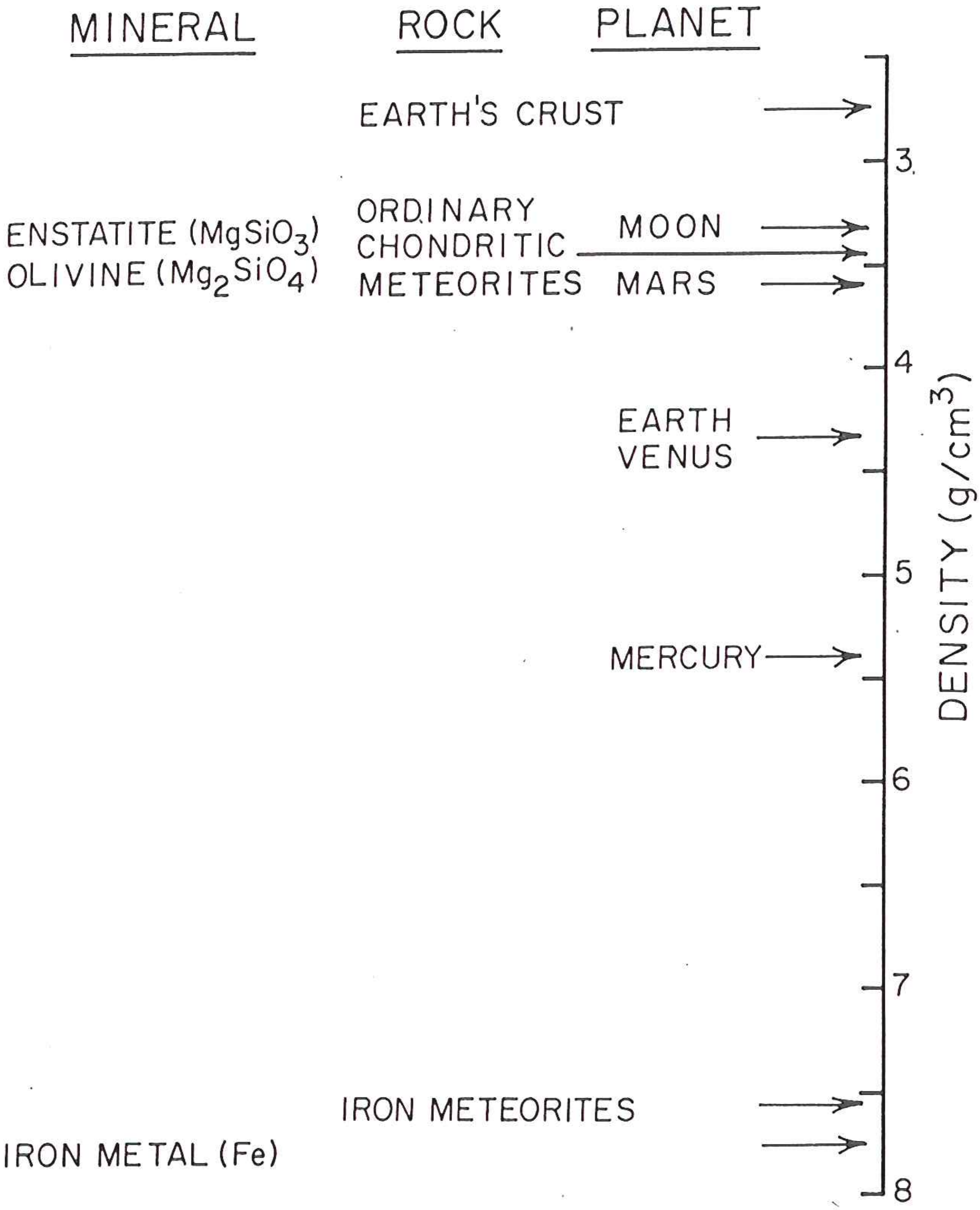


Figure 3-5. Bulk densities of various minerals, rocks, and planets: In the case of the planets, the densities shown have been corrected for gravitational compaction. The planet-to-planet density differences are in part the result of differences in the Fe/Mg+Si ratio and in part the result of differences in the iron/iron oxide ratio.

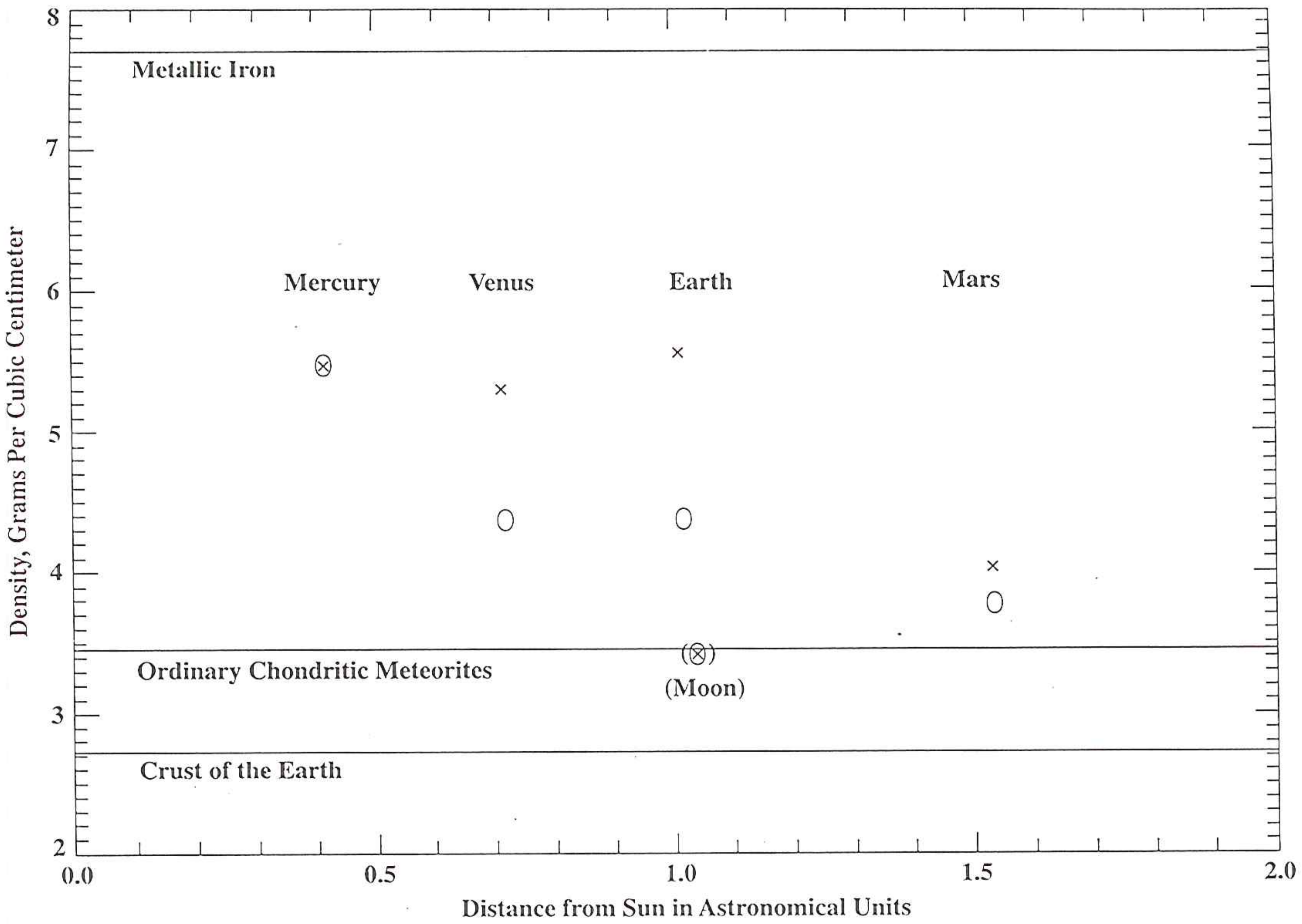


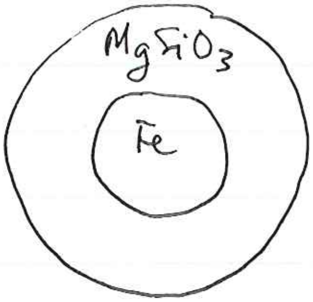
Figure 11.1. Densities of terrestrial planets and candidate mineral components. The planets are plotted as a function of their distance from the Sun. The symbol X indicates the measured density; the symbol O refers to the uncompressed density of the planet when the effect of gravitational compression of the material is removed. The compressed and uncompressed densities for the Moon are shown in parentheses, to distinguish them from Earth's. The density of Earth's crust also is shown; for comparison, the density of liquid water under low pressure is 1 gram per cubic centimeter.

Fe $\rho_{Fe} \approx 8000 \ kg/m^3$

silicate

$MgSiO_3$ perovskite

$\rho_{Si} \approx 3500 \ kg/m^3$



$$V_\oplus = V_{Fe} + V_{Si}$$

$$M_\oplus = M_{Fe} + M_{Si}$$

$$X_{Fe} = \frac{V_{Fe}}{V_\oplus} \ , \quad X_{Si} = \frac{V_{Si}}{V_\oplus}$$ volume fractions

$$\boxed{X_{Fe} + X_{Si} = 1}$$

$$\bar{\rho}_\oplus V_\oplus = \rho_{Fe} V_{Fe} + \rho_{Si} V_{Si}$$

$$\bar{\rho}_\oplus = \rho_{Fe} X_{Fe} + \rho_{Si} X_{Si}$$ ← all densities uncompressed

$$= \rho_{Fe} X_{Fe} + \rho_{Si} (1 - X_{Fe})$$

$$\boxed{X_{Fe} = \frac{\bar{\rho}_\oplus - \rho_{Si}}{\rho_{Fe} - \rho_{Si}} = \frac{4300 - 3500}{8000 - 3500} = 17\%}$$

Volume of a sphere $V = \frac{4}{3}\pi R^3$

$$\left(\frac{R_{core}}{R_\oplus}\right)^3 = 0.17$$

$$\boxed{R_{core} = \sqrt[3]{0.17} \times 6371 = 3500 \ km}$$

In fact from seismology we know the radius of $\oplus$ core is

$$\boxed{R_{core} = 3480 \text{ km}}$$

Compare with 5.19 (a) showing ρ and decompressed ρ_0 in PREM

- inner core, radius 1200 km solid — slightly higher density than liquid Fe core

- upper mantle phase transitions

410 km : olivine – spinel.
660 km : spinel – perovskite

Bulk chemistry Fe : Mg : Si : O thought to be constant throughout mantle

$$\boxed{\begin{array}{l} Fe : Mg : Si : O \approx 1 : 1 : 1 : 3\tfrac{1}{2} \\ \uparrow \\ \text{but 85\% of} \\ \text{this in core} \end{array}}$$

Magnesiowüstite MgO also present in lower mantle

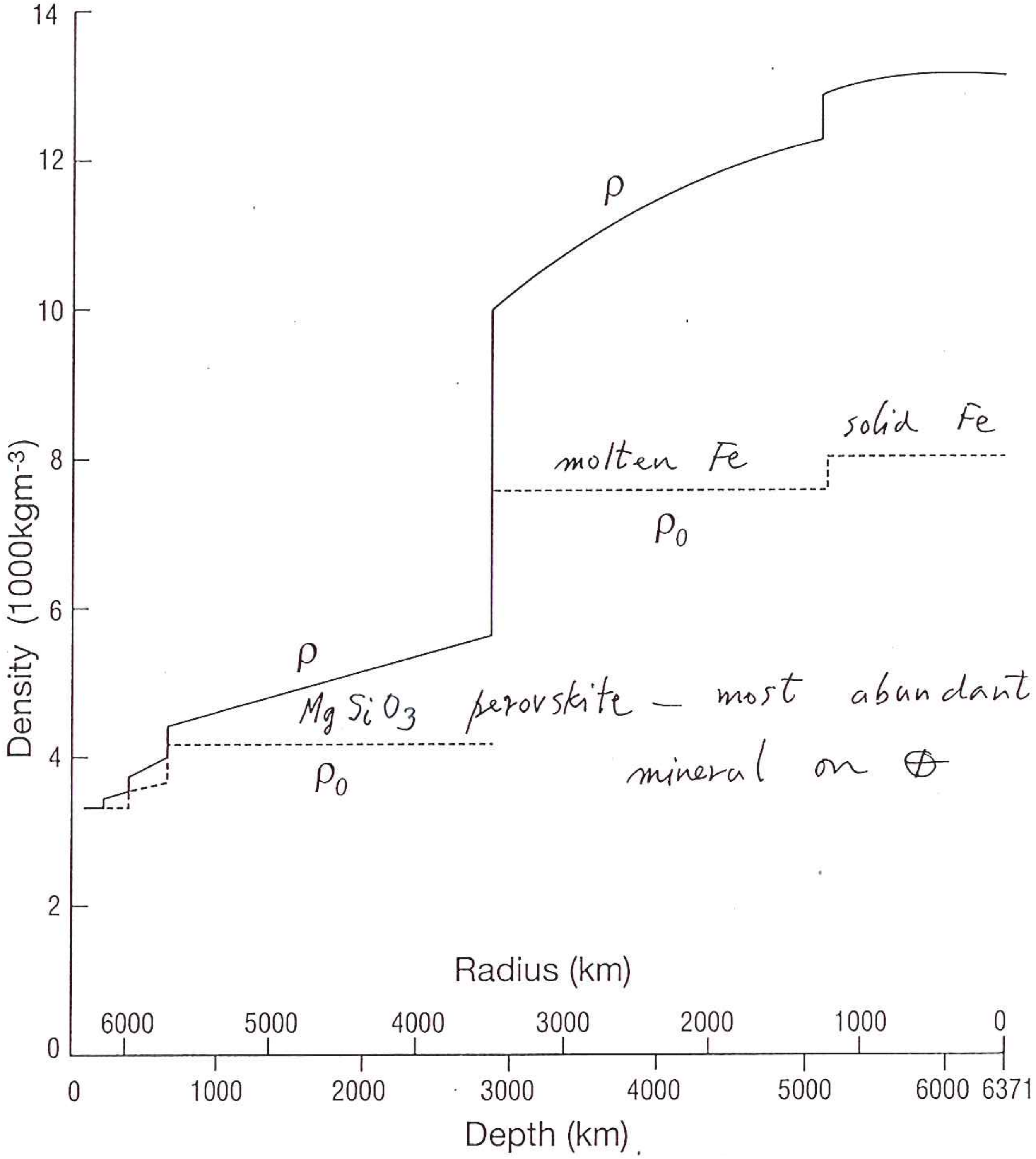


Figure 5.19(a). Profile of density, ρ, through the earth model PREM with corresponding zero pressure, low temperature density, estimated by finite strain theory.

pressure at center of ⊕ : 360 GPa
responsible for the compression

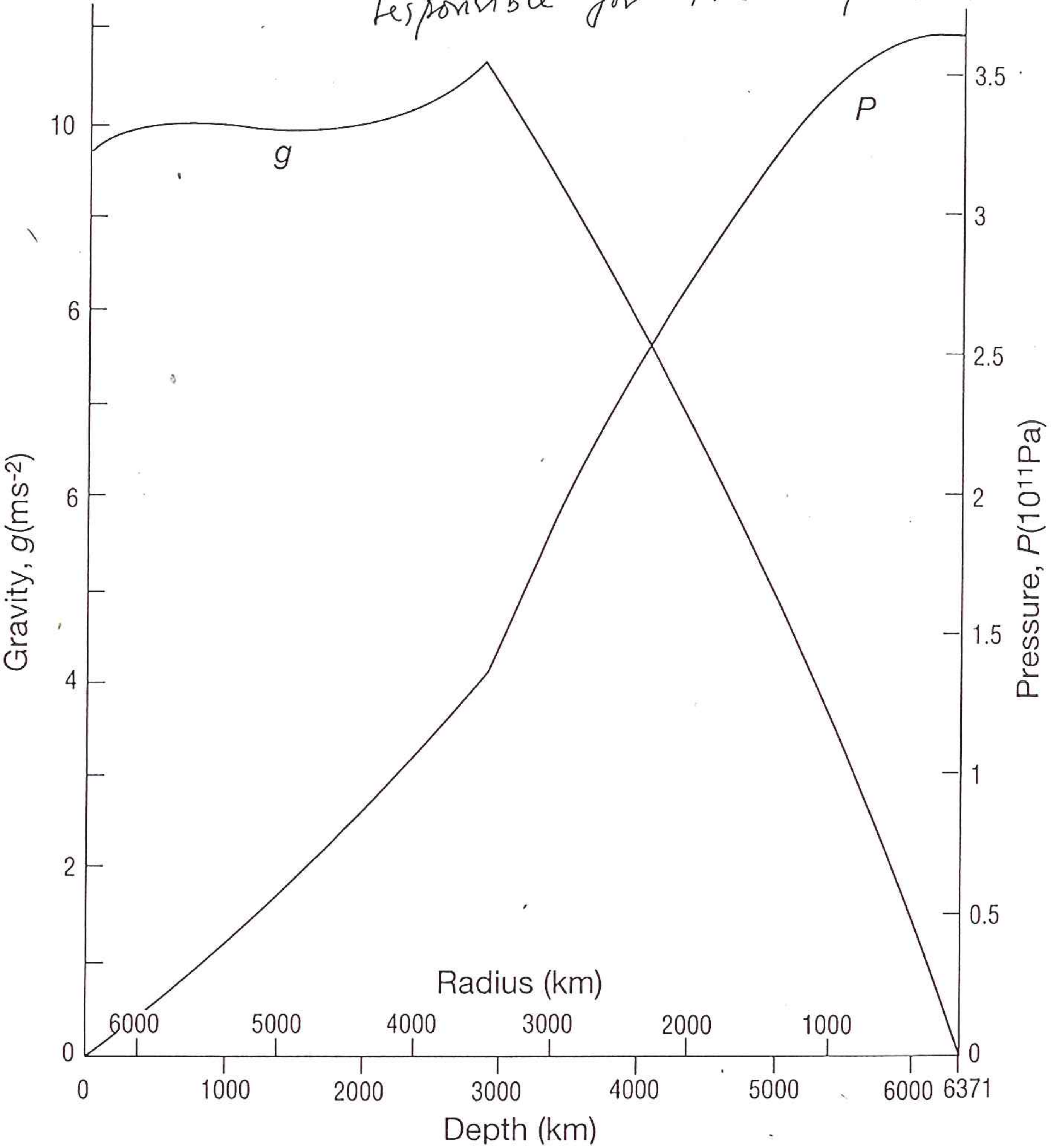


Figure 5.19(b). Profiles of gravity, g, and pressure, P, corresponding to the density profile in Fig. 5.19(a).

Table 5-3 gives an estimate of bulk composition $\oplus$ mantle

To convert to atoms :

Set Si = 1

Atomic weight M = moles/gm

For O in chondrites

$$\frac{32.3 \text{ gms} / 16 \text{ moles/gm}}{16.3 \text{ gms} / 28 \text{ moles/gm}} = 3.5$$

$\oplus$ mantle :

Mg:Fe ratio

$Mg_{90}Fe_{10}$

$\boxed{Fe : Mg : Si : O \approx 0 : [illegible] : 4\tfrac{1}{2}}$

$0.2 : 1\tfrac{1}{2} : 1$

~~Bulk composition of lower mantle more like~~

~~$Fe : Mg : Si : O \approx 0 : 1 : 1 : 3$~~

~~lower mantle ≈ $MgSiO_3$~~

Table 5-3. **Chemical composition** (in percent by weight) of the two most important rock types in the Earth's crust compared to the composition of Earth's mantle and to the composition of chondritic meteorites.

Atomic Weight		Chondritic meteorites		Earth's mantle		Basalt	Granite
16	O	32.3	3.5	43.5	4.7	44.5	46.9
56	Fe	28.8	0.9	6.5	0.2	9.6	2.9
28	Si	16.3	1	21.1	1	23.6	32.2
24	Mg	12.3	0.9	22.5	1.6	2.5	0.7
27	Al	1.4		1.9		7.9	7.7
40	Ca	1.3		2.2		7.2	1.9
23	Na	0.6		0.5		1.9	2.9
39	K	0.1		0.02		0.1	3.2
	Other	5.9		1.7		2.7	1.6

⇧

ratios in atoms