

MINERALOGICAL MAGAZINE

VOLUME 43 NUMBER 325 MARCH 1979

Mineralogy of the planets: a voyage in space and time

HALLIMOND LECTURE FOR CENTENARY OF MINERALOGICAL SOCIETY OF GREAT BRITAIN AND IRELAND, 1976

JOSEPH V. SMITH

Department of the Geophysical Sciences, University of Chicago, Chicago, Illinois 60637, USA

SUMMARY. Pertinent *general properties of the planets* are listed. The *condensation of the solar nebula* is set in the context of stellar evolution and meteorites with sections on astronomical observations, chemical composition of the solar nebula, physical properties of the solar nebula, and chemical and physical aspects of condensation, accretion, and planetary differentiation. A cool nebula is preferred to allow survival of pre-solar grains with isotopic anomalies. Equilibrium progressive condensation of the solar nebula is regarded as a useful theoretical boundary, but complex processes involving crystal-liquid differentiation in, and collisions between, planetesimals are used to interpret the properties of meteorites and terrestrial planets. *Chemical differentiation in the nebula* begins with condensation and aggregation of dust, which can yield oxidized and reduced products depending whether C/O is less or greater than unity. Simple models for direct accretion of condensed materials into planets are reviewed but not adopted. *Physical interactions involving small bodies* include collisional accretion of dust-covered bodies, and differentiation of silicate and metal from mechanical, magnetic, and electrostatic forces. *Physical and chemical differentiation involving large bodies* involves head-on and glancing collision of planetesimals, orbital deflection, and disintegration within the Roche limit, and collision with debris rings and moons. *Planetary accretion: dynamics, time scale, and heat sources* involves more rapid growth of a larger body than a smaller one with ultimate development of one planet in each feeding zone, which flares out and ultimately overlaps with adjacent zones. Mars is small, and a planet did not develop in the asteroid belt, because of perturbations from Jupiter. The giant planets deflected material into the inner solar system. Melting of early planetesimals is invoked to explain differentiated meteorites. *Chemical differentiation inside planet-*

esimals and planets describes the phase equilibria for metal, sulphide, and peridotite, either dry, wet, or containing CO₂. A wet body could begin crystal-liquid differentiation near 1250 K with sinking of Fe,S-rich liquid and rising of basaltic melt. The peridotitic residuum might undergo a subsequent differentiation at higher temperature under volatile-free conditions. Mineralogical storage of H₂O, CO₂, S, Cl, F, and alkalis is discussed. *Chemical differentiation in planetary atmospheres* briefly mentions escape of light species.

For the *Earth*, the *early history* is constrained by Archaean rocks dating from -3.8×10^9 yr whose properties indicate a non-reducing atmosphere, and a mantle that yielded volcanic rocks mostly similar to recent ones. The *upper mantle (above 200 km depth)* contains peridotitic rocks attributable to crystal-liquid differentiation and metamorphism. Volatile elements exist in mica and other minerals, but are sparse. Abundances of siderophile and chalcophile elements are high enough to require late accretion of material rich in these elements, the presence of a barrier between upper mantle and core, and some extraction by sinking sulphide. The *mantle (deeper than 200 km) and core* are inaccessible to direct study but interpretation of seismic data coupled with high-pressure laboratory studies requires inversion to dense phases in the mantle (especially perovskite?) and presence of light elements in the core (mainly S?). The *bulk composition* is modelled by cosmochemical analogy constrained by geophysical and geochemical parameters. The early condensate may be augmented by $1.5 \pm 0.5\%$ over (Mg + Si), and metal by $1.2 \pm 0.1\%$, while alkalis are probably depleted six-fold. Radial heterogeneity from a reduced interior to an oxidized exterior is suggested. For the *Moon*, sections cover *observations, petrologic interpretations, and bulk chemical composition*. For the *origin of the*

Earth and Moon, age constraints, chemical constraints, and dynamical and accretional constraints allow comparison of suggested origins, with the conclusion that the Moon formed by either fission or disintegrative capture during early growth of the Earth, followed by simultaneous accretion coupled with disintegrative capture of planetesimals.

Mercury must be Fe-rich, but the silicate may not be just early condensate. For Venus, reviews are given of the surface properties, atmosphere, speculations on bulk composition and speculations on surface mineralogy and atmospheric compositions. The CO₂ is in the atmosphere; most H may have been lost with concomitant oxidation of rocks; K/U ratios suggest basaltic and granitic rocks, and the high-surface temperature (c. 740 K) implies granulitic metamorphism. For Mars, reviews are given of surface morphology, atmosphere and volatiles, mineralogy and petrology, and geophysical and geochemical models. Prolonged emission of Fe-rich lavas is suggested. Volatiles were removed by mineralogical processes from the atmosphere to give ice caps and sediments affected by aeolian processes and oxidation from photochemically generated H₂O₂. Fe-rich layer silicates, maghemite, and Mg-sulphate may dominate the sediments.

The composition of comets and interplanetary dust may be inferable from micrometeorites whose complex properties are suggestive of carbonaceous meteorites. Asteroids should be supplying at least many of the meteorites to Earth. A general description and review of remote-sensing studies culminate in a review of spatial descriptions and implications. The main belt is dominated by dark C-type asteroids assumed to be the primordial inhabitants produced by primary condensation and local accretion. The inner zone contains some brighter S-type asteroids interpreted as having undergone separation of metal from silicate as in stony-iron meteorites, as well as some E- and M-types perhaps matching enstatite-bearing meteorites and irons. Perhaps these differentiated(?) asteroids, as well as basaltic Vesta, are strays perturbed outwards from the inner solar system.

A review of meteorites covers carbonaceous meteorites,

ordinary chondrites, enstatite chondrites and achondrites, reduced irons, forsterite-bearing meteorites and silicate inclusions in irons, irons, pallasites, chassignites and nakhlites, ureilites and lodranite, eucrites and shergottites, diogenites, howardites and mesosiderites, oxygen isotopes, and original location of meteorites. Emphasis is placed on mineralogical properties demonstrating crystal-liquid differentiation, brecciation, agglomeration, and even aqueous alteration and vapour transfer, though some evidence remains of direct condensation of gas to solid. The meteorites demonstrate the existence of many parent bodies (probably over 60), and most macrometeorites are ascribed to collision debris from the main-belt asteroids that has been deflected past Mars. The range of oxygen isotopes is explained by mixing of supernova debris rich in ¹⁶O with gas-solid differentiates of the nebula.

The culmination of the review is a suggested synthesis, which emphasizes a new model of heterogeneous accretion from planetesimals whose diverse compositions range from reduced material near Mercury to oxidized material from the asteroid zone outwards. Growth of a terrestrial planet begins from near-by slow planetesimals, and ends with distant fast planetesimals. Only the Earth and Venus are big enough to retain debris from late volatile-rich planetesimals deflected to high speed by the giant planets. Mercury, Moon, and Mars are volatile-poor. The Earth is zoned from a reduced interior, composed of crystal-liquid segregates of planetesimals captured early, to an oxidized exterior containing some material captured from outside the orbit of Mars. A mantle barrier hindered chemical equilibration. Hydrogen loss augmented oxidation. The mantle may be chemically zoned inwards from olivine-rich composition to pyroxene-rich composition, and the core should be reduced and contain substantial S and perhaps C and P. Mercury and Venus should have accreted reduced material, and only Venus should contain volatiles obtained from late planetesimals from outside the Martian orbit. Mars should have accreted mainly S-type asteroids, and have captured little volatile-rich material.

Index

	page		page
General properties of the planets	4	Moon:	
Condensation of the solar nebula	8	Observations	36
Astronomical observations	9	Petrologic interpretations	37
Chemical composition	9	Bulk chemical composition	39
Physical properties	10	Origin of earth and moon:	
Chemical and physical aspects of condensation	11	Age constraints	40
Chemical differentiation	11	Chemical constraints	40
Physical interaction of small bodies	14	Dynamical and accretional constraints	40
Physical and chemical differentiation, large bodies	15	Comparison of suggested origins	41
Planetary accretion: dynamics, time scale, heat sources	17	Mercury	42
Chemical differentiation inside bodies	20	Venus:	
Chemical differentiation in planetary atmospheres	25	Surface properties	
Earth:		Atmosphere	44
Early history	26	Speculations on bulk composition	45
Upper mantle	28	Speculations on surface mineralogy and atmospheric composition	45
Lower mantle and core	30	Mars:	
Bulk composition	31	Surface morphology	48
		Atmosphere and volatiles	50
		Mineralogy and petrology	51

Geophysical and geochemical models	52	Irons	67
Phobos and Deimos	54	Pallasites	70
Comets, asteroids, and meteorites	55	Chassignites and nakhlites	71
Comets and interplanetary dust	55	Angrite	71
Asteroids:		Ureilites and lodranite	72
General description	56	Eucrites and shergottites	73
Remote-sensing studies	57	Diogenites	74
Spatial descriptions and implications	59	Howardites and mesosiderites	74
Meteorites	60	Oxygen isotopes in meteorites	75
Carbonaceous chondrites	60	Original location of meteorites	76
Ordinary chondrites	62	Suggested synthesis:	
Enstatite chondrites and achondrites: reduced irons	64	Pre-planetary stage	77
Forsterite-bearing meteorites and silicate inclu- sions in irons	66	Earth	78
		Other terrestrial planets	80
		Conclusion	81

THE Mineralogical Society of Great Britain and Ireland has a distinguished record in promoting research in geochemistry, mineralogy, and petrology over the past century, and I was delighted but awed to accept the invitation to talk at this Centenary on some topic that looks forward to the next century. Predicting the future is uncertain whether it relates to science (recall the complacency of classical physicists at the end of the nineteenth century), to personal matters (as Ecclesiastes reminds us), or to public affairs.

Direct exploration of the mineralogy of the solar system began with the manned and unmanned missions of the last two decades, four centuries after a natural philosopher in Padua, Galileo Galilei, used his 32-power telescope to discover the roughness of the Moon, the Medicean satellites of Jupiter, the phases of Venus, and the spots on the Sun. Fearful of revolutionary ideas the Roman church placed Galileo under house arrest, but John Milton visited him and was undoubtedly stimulated in writing *Paradise Lost*: e.g. lines 563–75 in Book VII.

The Planets in thir station list'ning stood,
While the bright Pomp ascended jubilant.
'Open, ye everlasting Gates', they sung,
'Open, ye Heav'ns, your living doors, let in
The great Creator from his work return'd
Magnificent, his Six days' work, a World;
Open, and henceforth oft; for God will deign
To visit oft the dwellings of just Men
Delighted, and with frequent intercourse
Thither will send his winged Messengers
On errands of supernal Grace.'

In spite of numerous historical records of meteorite falls, a century of scientific scepticism about these messengers from the gods elapsed before the nineteenth-century exploration of the mineralogical mysteries of meteorites, typified by the microscopic studies of Henry Sorby in Edin-

burgh and Gustav Tschermak in Vienna (Wood, 1968). [Milton's winged messengers would be comets not meteorites.]

Voyages to the Moon were envisaged at least two millenia ago (Braun and Ordway, 1975), even though cultural memory of the Sun and planets as inviolable gods persisted strongly into Elizabethan times: indeed this mythical inheritance still remains in astrology, and may even have produced psychological resistance of scientists to melting of the poetical pale and cold Moon. A Parisian lawyer, Jules Verne, extrapolated current technology, including that of ballooning, in his fanciful but scientifically based adventure tales *La Voyage au Centre de la Terre* in 1864 and *De la Terre à la Lune* in 1865. Here in London, an unsuccessful teacher, H. G. Wells, made his fortune with scientific fantasies including *The War of the Worlds* in 1898 and *The First Men in the Moon* in 1901. Science-fiction writers continue to speculate on exploration of solar systems, but leadership is passing to engineers and physicists whose plans include establishment of human communities on artificial satellites (e.g. O'Neill, 1974, 1975) and planetary surfaces. A session on 'Utilization of lunar materials and expertise for large scale operations in space' on 16 March 1976, at NASA-Johnson Space Center, demonstrated the difficulties of construction of human dwellings in space. It is easy to scoff at suggestions for capturing comets and asteroids, and for wholesale modifications of the atmosphere of Venus to lower the temperature enough for human habitation but, when account is taken of the achievements of technology in relation to then current knowledge (e.g. the building of Neolithic temple-observatories, the Great Wall of China, nuclear reactors, and space rockets) and of the possibility of new scientific discoveries, it is perhaps safer to believe that some suggestions will ultimately bear fruit.

Although space exploration has been one of the

major forces underlying the revolution in the earth sciences, and although practical needs will require even more intense study of the Earth from both observational and theoretical viewpoints, it is perhaps optimistic to hope that scientific exploration of the solar system will receive high priority over the next century as unconfined populations exacerbate the mad passions of religious, political, social, and economic tensions. But what a magnificent dream would be the conversion of money spent on dreadful war machines and dangerous drugs into exploration of the near-by planets. Even if the Earth were spared human catastrophes, establishment of manned laboratories on planets or asteroids is unlikely in the next half-century (and perhaps century), and scientific study probably will be based largely on remote observations, except for meteorites and lunar samples. Small samples from Mars may be returned to Earth within 20 years. Establishment of a scientific laboratory on the Moon may result as an ancillary to engineering developments.

Study of lunar samples and meteorites has shown how difficult is the detective work even for samples dissected and probed by the most sophisticated instruments. Shock and thermal metamorphism have removed evidence in many samples, and we must accept that certain questions cannot be answered with certainty. Even adherence to Chamberlin's method of multiple working hypotheses may not lead to a definitive answer, and humility rather than cocksure assertion is preferable in facing such complex questions as the emplacement of the Moon around the Earth.

Looking backwards, the Earth begins to lose its pre-eminence because of the absence of rocks older than -3.8×10^9 yr. Early hopes that the Moon would be a cold body with undisturbed relics of the

early solar system were thwarted by the discovery of extensive melting and brecciation. Mineralogical and chemical studies of meteorites, and remote sensing of asteroids, are demonstrating the great complexity of products resulting from chemical differentiation of the solar nebula. Probably the asteroids and comets preserve samples of most if not all stages of development of the solar nebula. Astronomical observations of stars and interstellar clouds should provide boundary conditions for theories of the solar system. As orbiting telescopes are developed with facilities for observation without loss of spectral range in the Earth's atmosphere, many ideas of the mineralogy of the surfaces of solar-system bodies should be tested.

Therefrom, the mineralogy of the planets must be envisaged in a very wide framework embracing many aspects of astronomy, geochemistry, and geophysics. Perhaps no scientist can hope to understand all the relevant factors, and all of us must recognize the limitations of our competence. With great hesitation I try to assemble ideas from many disciplines but recognize that the present view of the mineralogy of the planets is highly imperfect. Let future scientists take joy in their success in reaching a higher level of understanding than that attained here.

The textbooks by Hartmann (1972), Kaula (1968), Haymes (1971), Short (1975), and King (1976) cover many pertinent properties of the planets, but are becoming obsolescent as new discoveries appear. The September 1975 issue of *Scientific American* contains useful surveys.

For brevity, referencing is less detailed for general concepts than for mineralogical ones. The lecture has been revised to include publications up to June 1978.

GENERAL PROPERTIES OF THE PLANETS

The heavens themselves, the planets, and this centre,
Observe degree, priority, and place.

WILLIAM SHAKESPEARE, *Troilus and Cressida*, I. iii. 85

Table I lists pertinent physical properties of the planets, which bear no simple relation to the distance from the sun. Properties of the outer planets and especially of their satellites are difficult to measure from telescopes (Newburn and Gulkis, 1973; Morrison and Cruikshank, 1974; *Scientific American*, Sept. 1975, pp. 26 and 152).

For massive planets the real density must be adjusted for the compression before being used to limit chemical composition. The estimated density at 10 kb and c.300 K depends on the variation of pressure, temperature, chemistry, and crystal struc-

ture with depth, but all reasonable distributions should give a density within c.0.1 g/cm³ of the listed value. [Note that 1 bar = 10^5 N/m² = 10^5 Pa.]

The pressure at the centre of a planet can be estimated accurately only when the density distribution is known. The gravitational pressure P is related to the density $\rho(r)$ by $dP/dr = -GM(r)\rho(r)/r^2$, where $M(r)$ is the mass from the centre to radius r , and G is the gravitational constant. The density distribution can be determined from inversion of seismic data, as typified by density models obtained from free oscillations of

TABLE I. Selected physical properties of planets

Planet	Mass 10^{27} g	Radius 10^3 km	Real density g/cm ³	Estimated density at 1 GPa g/cm ³	Estimated central pressure 10^{10} Pa	Escape velocity km/sec	T of solid surface K	Distance to Sun 10^8 km
Mercury	0.33	2.44	5.44	5.3	< 4.4	4	100–650	0.58
Venus	4.87	6.05	5.27	3.9	< 30	10	740 ± 10	1.08
Earth	5.98	6.38	5.52	4.0	37	11	230–320	1.50
(Moon)	0.07	1.74	3.34	3.4	0.47	2	100–400	1.50
Mars	0.65	3.39	3.96	3.9	< 3.6	5	150–300	2.28
Jupiter	1900	70	1.3	—	$c.10^4$	$c.60$	—	7.78
Saturn	570	60	0.7	—	$c.6 \times 10^3$	$c.40$	—	14.3
Uranus	87	26	1.2	—	$c.4 \times 10^3$	$c.20$	—	28.7
Neptune	100	25	1.7	—	$c.6 \times 10^3$	$c.25$	—	45.0
Pluto	0.7?	3??	3??	—	?	?	—	59

Data from various sources.

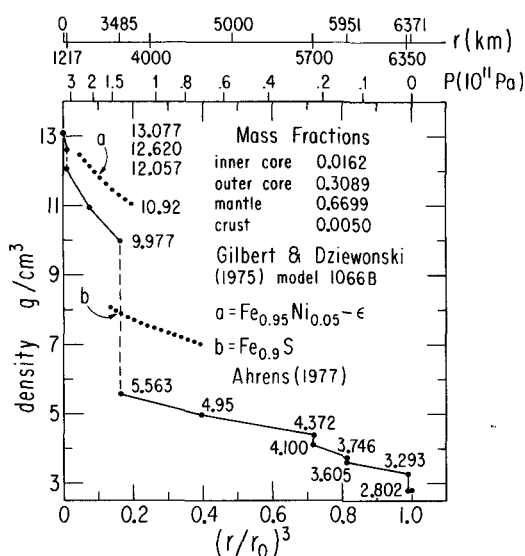


FIG. 1. Density-depth profile of Earth in model 1066B of Gilbert and Dziewonski (1975). Use of third power of radius r (normalized to $r_0 = 6371$ km) results in linear segments. See Ahrens (1978) for discussion of accuracy of shock-wave data in curves a and b, respectively from McQueen and Marsh (1966) and Ahrens (1976).

the Earth (e.g. fig. 1 based on Gilbert and Dziewonski, 1975). For the Earth a uniform density would give a central pressure of 17×10^{10} Pa (1.7×10^6 bar) compared to $c.37 \times 10^{10}$ Pa for models with an Fe-rich core (fig. 1). The density profiles of Mercury, Venus, and Mars are unknown, but the central pressures are unlikely to be greater and may be less than the values listed in Table I, which pertain to models with a heavy Fe-rich core. Accurate determination of the moment of inertia would provide a strong control on the degree of chemical differentiation of these planets.

The higher is the escape velocity $V_e = (2GM/R)^{1/2}$, the more difficult is loss of material by volatilization (M and R are the planetary mass and radius), and the easier is retention of impact debris during accretion. Modelling of impacts responsible for the large basins on the Moon, Mercury, and Mars is difficult (e.g. Baldwin, 1963; French and Short, 1968; Moore *et al.*, 1975; O'Keefe and Ahrens, 1977a, b) but probably most impact debris is retained on planets with $R > 2000$ km for impact velocities < 10 km/s.

The temperature at the solid surface depends on the heat flow from inside, the heat flow from the Sun, heat transfer by an atmosphere, and the heat loss to space. Atmosphere-free bodies must experience a strong range of surface temperature in response to variation of heat inflow from the Sun during rotation. Because of thermal resistivity of surface rocks the temperature wave at the surface is damped within a few centimetres for rotation periods up to 10^2 days, which covers all known stony bodies. For an atmosphere-free body rotating with its equator in the plane of the ecliptic, the smoothed surface temperature $T(\Psi)$ varies with latitude Ψ as $\Delta + T(0)\cos^n\Psi$ where n ranges from $\frac{1}{4}$ to $\frac{1}{2}$ depending on surface roughness (Bastin and Butt, 1975) and Δ , the equilibrium temperature for material unheated by the Sun, depends on the internal heat flow and the galactic temperature. On the Moon the temperature variations should be mainly between 400 and 100 K and on Mercury between 600 and 100 K, though somewhat lower values might occur in shaded spots near the poles. The heavy atmosphere and fairly smooth surface of Venus should lead to little temperature variation at the surface (< 30 K?). Mars has complex variations of surface temperature with latitude, elevation, and season.

The variation of temperature with depth below

the surface is poorly known for the terrestrial planets, and is model-dependent. Perturbation of the solar wind by the Moon leads to estimates of the electrical conductivity with depth, and the electrical conductivity can be transformed to temperature, assuming that the Moon is composed of either Mg-rich olivine or orthopyroxene (Dyal, Parkin, and Daily, 1975; Sonett and Duba, 1975). Particularly important is the assumption of complete divalency of the iron, which results in a lower electrical conductivity for a given temperature than for the assumption of some trivalency. Seismic data indicate lower Q near the Moon's centre than for the outer part, but presence of melt is disputable, as indeed would be the chemical composition of any inferred melt. Because of the evidence for early crystal-liquid differentiation of the Moon under dry conditions, the pressure-temperature curve (denoted *prete*) for at least the outer part of the early Moon probably became near to that for the beginning of melting of dry peridotite. The exact position of the curve depends on the detailed chemical composition of the peridotite, on the possible separation of an Fe,S-rich core, and on the time relation between accretion and melting (see many papers in *Proc. Lunar Sci. Conf.*). Whereas some workers have argued for complete early melting of the Moon (not necessarily at one time), others have argued for early melting only of the exterior (e.g. Solomon and Chaiken, 1976; Solomon, 1977*a*). Considerable cooling occurred in the exterior of the present Moon, and the present lunar-*prete* is subject to considerable uncertainty.

The two theoretical geotherms for the outer part of the present Earth (fig. 2, curves 1 and 2) are generalized for oceanic and shield areas, as reviewed by Wyllie (1971, pp. 30-2). Pyroxene thermobarometry of peridotite xenoliths from kimberlites (Boyd, 1973; Nixon and Boyd, 1977) apparently confirms the model geotherms for shield and oceanic areas, but details are controversial. Dynamic models associated with the theory of plate tectonics indicate complex geographic variation of the geotherm, as does the existence of 'hotspots'. The liquid nature of the outer core and the solidness of the inner core place constraints on the temperature, but these are weak because of uncertainty of chemical composition.

Unfortunately, there are severe problems in estimating planet-*prete*s by geophysical calculations principally because of uncertainty in the size and position of heat sources and sinks as a function of time, and in transfer mechanisms. Calculations for the Earth by Hanks and Anderson (1969), for Mars by Anderson (1972) and by Johnston, McGetchin, and Töksoz (1974), and for Mercury by Siegfried and Solomon (1974), illustrate the

problems with estimating the heat sources. New developments in the theory of heat transfer during convection (Daly and Richter, 1978) require modification of earlier calculations.

Solomon (1977*b*, 1978*a, b*) reaffirmed suggestions in earlier papers that the timing and extent of planetary differentiation is revealed by the presence or absence of compressional or extensional tectonic features at the surface. Specifically he argues for early core formation in Mercury, only partial early melting of the Moon, and the possibility of late or sluggish core formation in Mars.

Perhaps simple geochemical arguments will prove valuable in estimating planet-*prete*s. Plains on Mars and Mercury are commonly interpreted as lava flows, and Venus must have been chemically differentiated by crystal-liquid processes (see later). Terrestrial planets must be dominated by various proportions of O, Si, Fe, Mg, Ca, Al, and S. The beginning of melting involves complex phase relations, which can be crudely illustrated by reference to curves in fig. 2. When S and Fe metal are present, the Fe-FeS eutectic might form at $c.1250^{\circ}\text{K}$ from 0 to $c.60$ kb with a slow increase of temperature for higher pressures. The silicate portion of all the inner planets must be dominated by Mg-rich olivine and pyroxene compositions with minor Ca,Al-rich silicates. Such peridotitic materials show very complex melting relations in the presence of H_2O , CO_2 , and other volatiles (e.g. Wyllie, 1978). Some features are discussed later (p. 22), and it suffices here that 'wet peridotite' tends to melt between $c.1250$ and 1400°K for most pressures up to 60 kb, which temperature range is close to the Fe-FeS eutectic temperature (fig. 2). Anhydrous peridotite could yield basaltic and picritic liquids at the considerably higher temperatures on the curve labelled 'dry peridotite'. A planet composed of 'wet peridotite' could begin to melt near 1250°K , and most of the volatiles could be released to the surface taking the radioactive heat sources with them. Subsequent melting of the interior would require heating to the solidus for 'dry peridotite', which process might be difficult to achieve because of paucity of deep heat sources. The 'wet' differentiation could yield a 'basaltic' or 'andesitic' surface, which could undergo complex crystal-liquid differentiation, ultimately producing some 'granitic' material. Generalized melting curves for these materials (fig. 2) lie at considerably lower temperatures than for peridotite. Subduction and convection would lead to further complexities in the planet-*prete*.

Melting curves for Fe and enstatite (fig. 2) may pertain to the differentiation of Mercury, if Mercury accreted before condensation of feldspar from the solar nebula. Curve 5 from Siegfried and

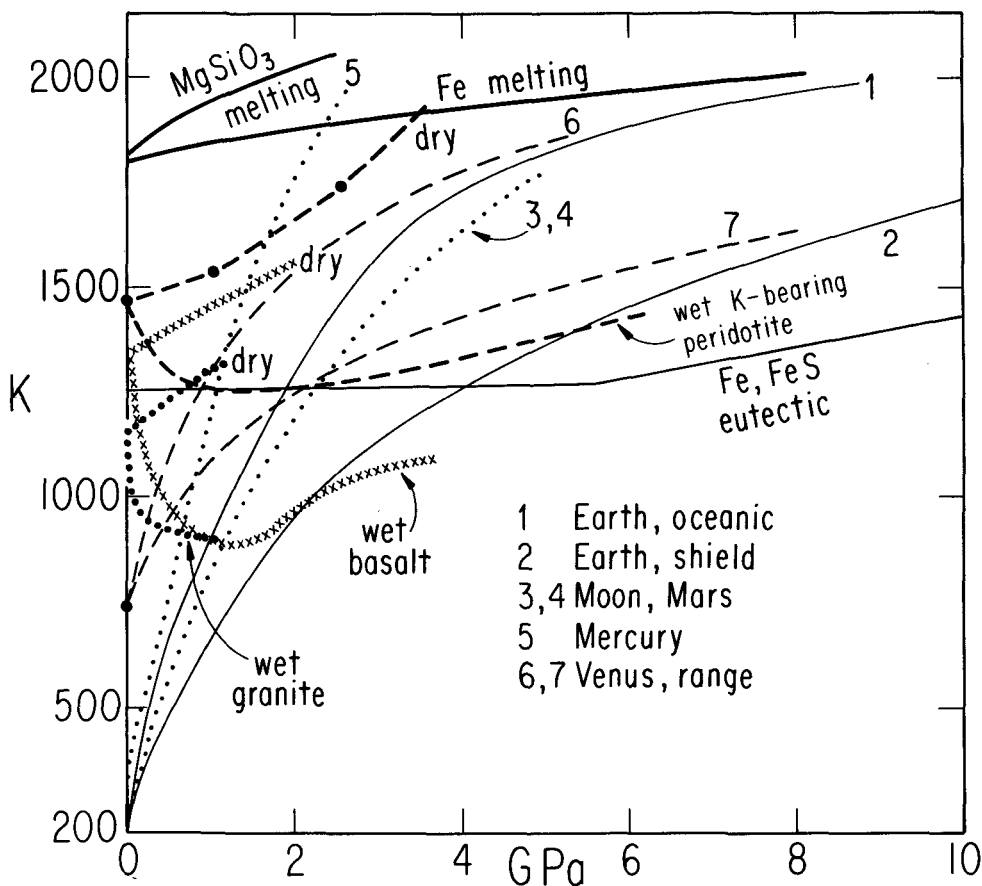


FIG. 2. Planetopretes and some phase boundaries. Curves 1 and 2: mathematical estimates of oceanic and shield geotherms (quoted in Wyllie, 1971). Curve 5: Mercury model (Siegfried and Solomon, 1974, Fig. 6). Curves 3 and 4 for Moon and Mars were so close that only one line is given: curve 3: Toksöz and Johnston (1974, Fig. 14, upper bound); curve 4, Johnston, McGetchin, and Toksöz (1974, Fig. 11, mean A and B). Curves 6 and 7: qualitative estimates for Venus. Beginning of melting for MgSiO_3 : Boyd, England, and Davis (1964). Melting of Fe and Fe-FeS eutectic: Usselman (1975). Beginning of melting for dry and H_2O -saturated basalt: Wyllie (1971). Beginning of melting for dry and wet peridotite—see fig. 8.

Solomon (1974, fig. 6) is approximately consistent with such an assumption, when account is taken of cooling of the exterior of Mercury during the past 4×10^9 yr.

The present suggested prete for Mars (fig. 2, curve 3) is consistent with early melting of 'wet peridotite' and incomplete heating to the curve for 'dry peridotite'.

The suggested range of pretes for Venus (fig. 2, curves 6 and 7) is constrained by the assumption that the Venusian interior is similar to that of the Earth, while the Venusian surface has been held at $c.740$ K for the past 4×10^9 yr. Perhaps plate-tectonic processes occur on Venus with complications caused by convection and subduction.

Chemical compositions of the planets are discussed later. Before looking at the processes involved in the accretion of the planets, it is perhaps desirable to point out the paucity of direct information on the chemistry and mineralogy of the planets, and the need to supplement the observations by implications from the chemistry of the planetary atmospheres. For the Earth direct information is available only for $c.200$ km, which is limited by the peridotite and eclogite xenoliths in kimberlite pipes. All chemical models for deeper-seated material ultimately rely on cosmochemical speculations, and on assumptions about the expulsion of volatile elements, some of which enter the atmosphere. That crystal-liquid differentiation has

occurred on Venus is necessitated by the strong concentrations of K, U, and Th at Venera landing sites, but equally important mineralogical implications ensue from the presence of CO_2 , H_2SO_4 , HCl , and HF in the Venusian atmosphere. The chemistry of the Martian atmosphere with its strong evidence for exospheric escape of light molecules has profound implications for the mineralogy and geology,

and provides a warning that the present oxidizing atmospheres of Earth and Venus may have been considerably different just after accretion.

The next section attempts to bring together cosmochemical and cosmophysical concepts useful in placing limits on planetary compositions and mineral assemblages.

CONDENSATION OF SOLAR NEBULA SET IN CONTEXT OF STELLAR EVOLUTION AND METEORITES

Who can number the clouds in wisdom? or who can stay the bottles of heaven,
When the dust groweth into hardness, and the clods cleave fast together?

Job xxxviii, 38-9

Although many plausible ideas and calculations now exist on the presumed origin of the solar system by condensation of a nebula, Jehovah's chiding of Job is still salutary.

Williams and Cremin (1968) reviewed ideas for the origin of the solar system, and claimed that derivation of the Sun and planets from the same gas cloud leads to the least difficulties in explaining the dynamical properties, particularly the distribution of angular momentum between the Sun and planets. Astronomical observations of nebulae, stars, and interstellar dust in the Galaxy provide a valuable guide to what might have happened in the solar nebula about 4.6×10^9 yr ago, as reviewed by

Hartmann (1972) and Strohmaier (1972). Spectrochemical analysis of the Sun's photosphere (Ross and Aller, 1976) yielded a bulk composition that fits well with that of C1 meteorites (except, of course, for volatile gases). This bulk composition is used along with estimates of pressure and temperature in thermochemical models for gas-to-solid condensation (Grossman and Larimer, 1974). The condensed products are envisaged to coalesce into 'clods', then into planetesimals, and ultimately into planets by complex dynamical processes (Safronov, 1972), which are complicated by chemical differentiation (fig. 3). The next sub-sections review features relevant to the mineralogy of the planets and

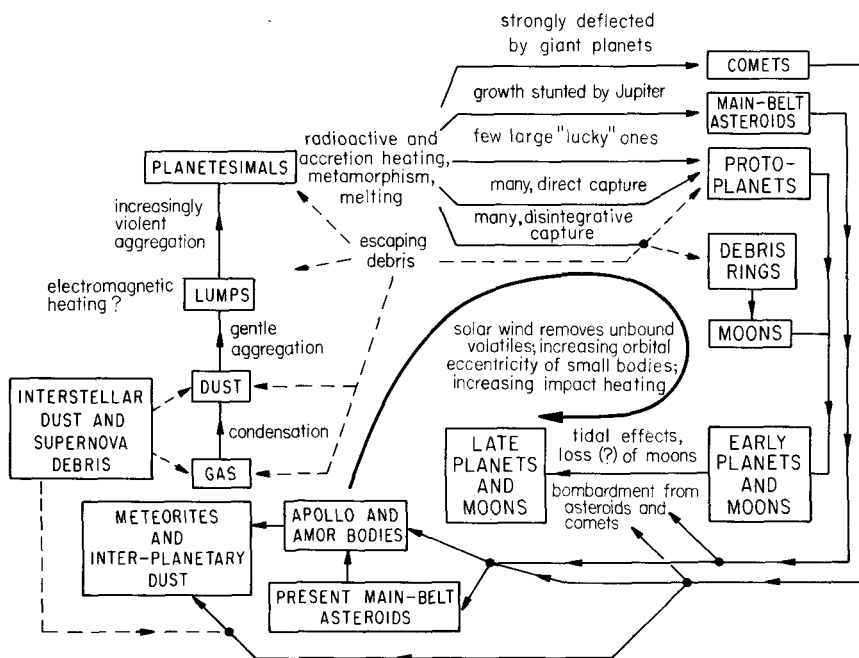


FIG. 3. Summary of possible development of planets and moons from the solar nebula.

meteorites. A general review (Kaula, 1975) on 'The Seven Ages of a Planet' was very useful, as was a review (Kaula, 1977b) on mechanical differentiation of protolunar material.

Astronomical observations

Cocoon nebulae around young stars are inferred to contain warm dust from the high ratio of infrared to visible radiation. Herbig-Haro objects also have excess infra-red radiation, and show irregular brightness. Stars passing through the T-Tauri phase show strong irregular changes of brightness and are surrounded by a variable nebula in which particles move up to 100 km/s. The nebula rotates rapidly and has a strong magnetic field, which is an important support for theories of electromagnetic transfer of angular momentum from the Sun to the planets. A big mass loss (c.0.1 total mass) from the star is inferred as the solar wind intermittently floods outwards.

The properties of dust in inter-stellar and intergalactic space were reviewed by Wesson (1974) and in papers collected by Field and Cameron (1973). Severe technical problems exist in the chemical interpretation of light scattered from small dust particles especially if very small and coated with condensed molecules (Aannestad, 1975; Allen and Robinson, 1975; Day, 1975; Wickramasinghe, 1974; Zaikowski, Knacke, and Porco, 1975), but the telescopic spectral data are not inconsistent with the dust consisting of minerals found in carbonaceous meteorites, especially phyllosilicates. Many molecules have been detected (Zuckerman, 1977), of which C-bearing ones are particularly relevant to chemical models for condensation of the solar nebula. Grain growth from inter-stellar gas involves complex processes that are poorly understood (e.g. Snow, 1975), and indeed grains can be destroyed by sputtering and by radioactive and collisional heating near a supernova (e.g. Falk and Scalo, 1975).

There are numerous speculations about the origin of star clusters and about the effects of supernovae on the interstellar clouds. Currently, the old idea of a supernova trigger for the origin of stars is popular, and indeed some scientists, including Cameron and Truran (1977), are exploring whether the gravitational collapse of the solar nebula was initiated by a shock wave from a nearby supernova. Such a collapsing solar nebula might contain substantial amounts of 'shrapnel-like' grains and gas from the supernova, and Lattimer *et al.* (1978) tried to predict the isotopic anomalies that might survive in refractory grains ejected from the various nucleosynthetic zones of the supernova. Alternatively, the solar nebula might have col-

lapsed without a supernova trigger, and might have inherited granular and gaseous debris from an earlier supernova (e.g. D. D. Clayton, 1977a, b).

Actually, grains of the appropriate mineral types occur in meteorites. White inclusions of the Allende meteorite carry the melilite, spinel, diopside, and perovskite predicted for oxidizing supernova zones with either $C/O < 1$ or $(S+Si)/O < 1$, while enstatite meteorites carry reduced phases including oldhamite (CaS) predicted for reducing zones with $C/O > 1$ or $(S+Si)/O > 1$. Observed correlations of excess ^{26}Mg with $^{27}\text{Al}/^{24}\text{Mg}$ in anorthite from the Allende meteorite (Lee *et al.*, 1977a, b); Hutcheon *et al.*, 1977; Lorin *et al.*, 1977) are consistent with crystallization of anorthite from aluminium enriched in nucleosynthetic ^{26}Al . Clayton *et al.* (1977) interpreted enrichments of ^{16}O (relative to terrestrial abundances) in C2, C3, and C4 meteorites in terms of admixture of 1 to 5% ^{16}O ascribed to pre-solar carriers, and Clayton and Mayeda (1977) found correlations between O and Mg isotopic anomalies. Manuel *et al.* (1972) had earlier attributed xenon anomalies to nucleosynthesis in a supernova. Recent data on isotopic anomalies in the early solar system are reviewed by R. N. Clayton (1978).

Chemical composition of the solar nebula

The simplest approach is to assume that the solar nebula was spatially homogeneous when condensation began, and indeed there is reasonable agreement between the spectral estimate of the chemical composition of the solar photosphere and the bulk composition of C1 meteorites (Table II) when account is taken of depletion of volatile elements in C1 meteorites. Furthermore, estimated depletion factors for the Earth, Moon, and presumed parent body of eucrites with respect to C1 composition are consistent, within the somewhat large uncertainties, with the assumptions of a single reservoir in the solar nebula, and variable depletion factors related to element volatility and metal-silicate fractionation (Anders, 1977; Smith, 1977).

However, different ratios of the three isotopes of oxygen among terrestrial, lunar, and meteoritic samples (R. N. Clayton and Mayeda, 1978) require at least nine separate reservoirs related by mass fractionation and by incorporation of various amounts of ^{16}O . The Earth and Moon have isotopic ratios consistent with derivation mainly from enstatite meteorites, but not from ordinary and carbonaceous chondrites; however, mixtures of several meteorite types are more plausible than just a single type. Mechanical admixture of ^{16}O might be explained by incorporation of supernova debris (preceding section). If ordinary and enstatite

TABLE II. *Chemical composition of solar photosphere and certain meteorites*

	a atomic	b atomic	c wt %	d wt %	e wt %	f wt %	g wt %	h wt %	i wt %
H	12	8.36	< 2	< 1	< 0.04	—	—	—	—
C	8.62(12)	7.43(20)	3.1	2.2	0.5	0.4	c.0.2	c.0.3	0.07
N	7.94(15)	6.28(40)	0.26	c.0.2	c.0.05	c.0.03	0.005	c.0.005	0.004
O	8.84(7)	8.47	46.0	41.5	36.5	27	35	39	—
Na	6.28(5)	6.36(4)	0.51	0.38	0.35	0.82	0.58	0.66	c.0.35
Mg	7.60(15)	7.61(1)	9.56	11.8	14.5	10.7	14.2	15.2	23
Al	6.52(12)	6.51(3)	0.85	1.08	1.37	0.79	1.01	1.10	c.0.5
Si	7.65(8)	7.59(1)	10.5	13.0	15.7	16.9	17.0	18.3	27
P	5.50(15)	5.57(20)	0.11	0.10	0.13	0.21	0.12	0.11	0.01?
S	7.20(15)	7.28(7)	5.90	3.4	2.2	5.8	2.0	2.2	c.0.4
Cl	5.5(4)	5.17(30)	0.05?	0.03?	0.03?	0.07?	0.01?	0.01?	c.4 ppm
K	5.16(10)	5.15(6)	0.054	c.0.05	c.0.04	0.09?	0.08?	0.09?	0.02?
Ca	6.35(10)	6.44(10)	1.06	1.34	1.70	0.84	1.19	1.28	c.1
Ti	5.05(12)	4.97(3)	0.043	0.054	0.087	0.057	0.062	0.066	c.0.03
Cr	5.71(14)	5.68(6)	0.24	0.31	0.35	0.32	0.34	0.37	c.0.05
Mn	5.42(16)	5.55(7)	0.19	0.16	0.15	0.32	0.22	0.24	0.14
Fe	7.50(8)	7.53(1)	18.4	21.9	25.1	33.0	27.6	21.8	1.02
Ni	6.28(9)	6.25(13)	1.0	1.3	1.4	1.8	1.8	1.2	0.02
Zn	4.45(15)	4.68(1)	0.030	0.02	0.01	0.004	0.005	0.006	0.001
U	< 0.6	—0.40(12)	9.1 ppm	12 ppm	16 ppm	11 ppm	11 ppm	13 ppm	5 ppm

a. Solar spectrum (Ross and Aller, 1976); uncertainty in brackets, e.g. C 8.62 ± 0.12 ; normalized to $\log H = 12$; $\log$ atomic. b. C1 meteorite; $\log$ atomic normalized to $\log Mg + \log Si = 15.2$; data sources in Smith (1977, Table 2, col. 1). c. C1 meteorites. d. C2 meteorites. e. C3 meteorites. f. enstatite chondrites, mainly E4. g. ordinary H chondrites, mainly unequilibrated. h. ordinary L chondrites, mainly unequilibrated. i. enstatite achondrites. Data taken from compilation in Mason (1971a). The present table is designed to show important features of meteorite compositions that may be a guide to planetary compositions, and readers are referred to original papers for many complications.

chondrites, along with differentiated meteorites, are assumed to have formed in the inner solar system, as is highly probable, then the solar nebula was not isotopically homogeneous. Only if most meteorites are assumed to be strangers to the solar system (e.g. the residua of captured comets) can the convenient assumption of a homogeneous primordial nebula be retained.

Perhaps the simplest temporary assumption is that the primordial nebula was sufficiently homogeneous in major elements for condensation calculations to be applied towards the origin of the planets, yet sufficiently inhomogeneous in some isotopes to yield the isotopic variations found in ordinary chondrites and differentiated meteorites. The carbonaceous meteorites might be ascribable to cometary debris. At the other extreme are possibilities involving major inhomogeneities of the solar nebula: e.g. Manuel and Sabu (1975) and Sabu and Manuel (1976) proposed that the solar nebula is composed largely of supernova debris, that the iron cores of the inner planets and the iron meteorites derive principally from the centre of the supernova, and that the outer planets are highly

enriched in elements from the rim of the supernova. The problem of the highly reduced enstatite meteorites is discussed later.

Two possible causes of chemical variation in the solar nebula are momentum differentiation, in which lighter gases tend to move to lower gravitational potential than heavier ones, unless turbulent mixing is strong, and effects from the solar wind, which has been jokingly described as a 'magic broom' allowing theorists to sweep away any unwanted volatiles!

Physical properties of the solar nebula

Unfortunately the physical properties cannot be calculated from first principles, and all estimates are controversial.

The original solar nebula must have had a mass greater than that of the present Sun and planets. Assuming that the Sun completed most of its growth before the planets accreted, the non-solar part of the nebula need not have been greater than c.0.01 solar mass if it had solar-photospheric composition (Hoyle and Wickramasinghe, 1968).

Safronov (1972) used 0.05 solar mass, but Cameron and Pine (1973) used a massive nebula with 1 solar mass. Safronov (1977) argued that such a large nebula could not have been dispersed by the solar wind after < 0.01 solar mass had entered the planets, and indeed Handbury and Williams (1976) have severely questioned the efficacy of the solar wind in removing even a small nebula. Cameron (pers. comm., 1976) has apparently abandoned the concept of a hot massive nebula.

Ringwood (1975, pp. 563–8) assumed that a small nebula was relatively cool, perhaps not above 300 K in the source region of the inner planets and c.200 K for the giant planets. Such a cool nebula would allow survival of interstellar grains, and is in complete contrast to the concept of a hot nebula (e.g. Cameron and Pine, 1973) in which initial temperatures of c.2000 K resulted in the vapourization of all interstellar grains. Hoyle and Wickramasinghe (1968) discussed the expansion of the solar nebula by transfer of angular momentum via a magnetic field of 10^2 gauss, and deduced a temperature $c. T_s(r/R)^{-1/2}$ where T_s is the temperature of the Sun's surface, R the Sun's radius, and r is distance from the Sun. A temperature of 1500 K was obtained between Venus and Earth and 150 K between Uranus and Neptune. Lewis's (1974a) suggestion that the chemistry of the planets enforces an adiabatic profile on the solar nebula was criticized by Kaula (1975), but some kind of pressure-temperature gradient must occur.

A cool nebula would rule out the appealing and highly developed concept of simple progressive condensation from c.2000 K of a chemically uniform gas, as used in many models for chemical differentiation of planets and meteorites. It can be argued that the high-temperature minerals of C3 meteorites formed outside the solar system. Anders (1971a) summarized evidence on the 'condensation temperature' of ordinary chondrites, the eucrites, the Earth, and the Moon, but the significance of the values of 460 to 600 K estimated from the contents of volatile elements is not clear when one considers all possible chemical processes in the accretion of planets via planetesimals. There is abundant meteoritic evidence for melting of even small planetesimals, but this does not necessarily require accretion from hot condensates. Radiation loss could allow grains to be cooler by some hundreds of degrees than the surrounding gas (Arrhenius and De, 1973). Against all these permissive arguments for cool grains in a solar nebula is the undoubted evidence for strong depletion of volatiles in the inner planets (especially for Mercury), and the weaker evidence for enrichment of the Earth in refractory material (Ganapathy and Anders, 1974; Smith, 1977). For the time being it is desirable to

keep an open mind on the temperature of the solar nebula, and especially not to accept uncritically the well-developed concepts of progressive condensation of a fully gaseous nebula.

Pressure estimates of the solar nebula are also uncertain, but fortunately are not critical to condensation calculations. The pressure depends on the degree of flattening of the nebula. Hoyle and Wickramasinghe suggested that the density is proportional to r^{-3} , and Cameron and Pine suggested a pressure range from 10^{-2} bar at Mercury to 10^{-6} bar at Neptune.

Chemical and physical aspects of condensation, accretion, and planetary differentiation

Condensation and aggregation of dust. Although simple progressive condensation of the solar nebula cannot explain all the properties of meteorites and planets, it provides a useful theoretical boundary. Grossman and Larimer (1974) summarized the results of many workers in which the early chemical history of the solar system is developed in terms of thermochemical calculations. Solids in thermal equilibrium with the nebula gas condense when vapour saturation occurs, and change composition or sometimes disappear upon reaction with the surviving gas at lower temperature. Table III(a) summarizes the essential features for thermal

TABLE III. *Effect of C/O on condensation of solar nebula at 10 Pa (0.1 millibar)*

(a) Cosmic 0.6		(b) Reduced c.1.0	
1800 K	refractory metals		
1600–1400	Ca, Al, Ti oxides	1500 K	SiC, CaS, TiN, Fe ₃ C
1400–1300	Fe, Ni, Co and Mg, Si oxides	1400	Fe, Ni, Co
700	some Fe → sulphide	1200–1100	Ca, Al, Ti oxides
500	some Fe → oxide	1100–1000	Mg, Si oxides
400	OH-silicate		
150	water ice		
100–50	NH ₃ , CH ₄ clathrates		
20	CH ₄ , Ar ices		

Abstracted from Grossman and Larimer (1974) and Larimer (1975).

condensation of a nebula with composition similar to that in Table II, and with the specific C/O ratio of 0.6, and fig. 4 shows some details. The pressure was assumed to be 10^{-4} bar, but only small differences in the condensation sequence occur for pressure changes up to 100-fold. Prediction of condensation of refractory metals near 1800 K and of Ca, Al, Ti-oxides at 1600–1400 K fits beautifully with the observations of tiny nuggets containing

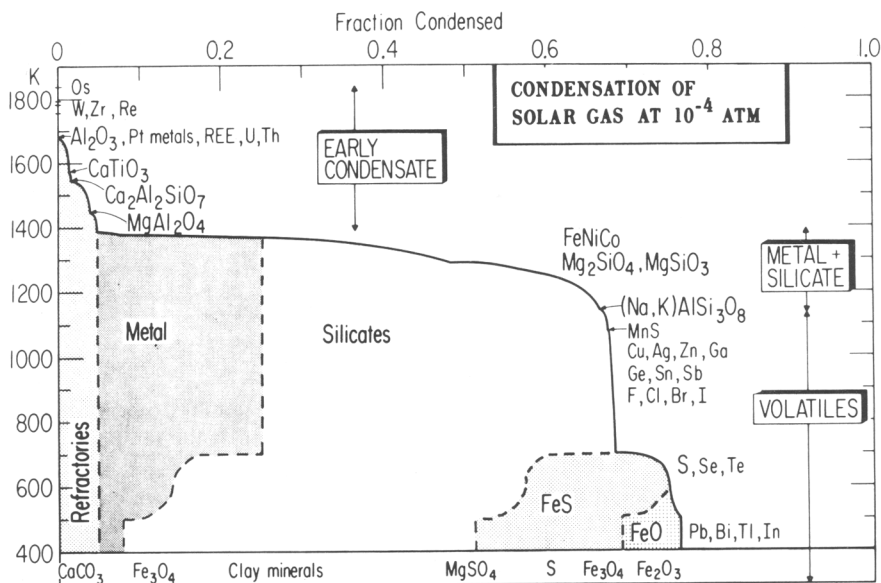


FIG. 4. Condensation of the elements from a gas of 'solar' composition at 10 Pa (10^{-4} bar) (Ganapathy and Anders, 1974). This composition has C/O ratio of 0.6. See also Grossman and Larimer (1974, Fig. 2).

Mo, W, and Ru, Os, Ir, Pt (Wark and Lovering, 1976) and of melilite-perovskite-spinel aggregates (many references) in the Allende C3 meteorite; however, Blander and Fuchs (1975) argued for a liquid rather than a gaseous origin of some Ca, Al-rich inclusions in the Allende meteorite. Fe, Ni, Co-metal, and Mg-silicates (Fe-free forsterite and then enstatite) should condense under equilibrium at 1400–1300 K. At higher pressure ($> 10^{-3}$ bar), Fe-rich metal begins to condense before forsterite, but most of the metal condenses simultaneously with Mg, Si-oxides over pressures from 10^{-5} to 10^{-1} bar. At about 700 K much of the Fe is removed from the metal and converted to troilite, FeS, and from 750 to 500 K some of the remaining Fe is oxidized to FeO, which should be incorporated into the Mg-silicates. Magnetite, Fe_3O_4 , becomes stable only below 400 K. Hydroxylated phyllosilicates become stable near 350 K, water ice at 150 K, various clathrates at 100–50 K, and methane and argon ices near 20 K (Lewis, 1972b; Sill and Wilkening, 1978). Particularly important are the predictions that U and Th should be incorporated into the perovskite, which is a high-temperature condensate, whereas Na and K condense as a solid-solution with anorthite only in the medium-temperature range 1200–1000 K. It must be emphasized that this condensation sequence applies only to a low-pressure nebula that retains its bulk composition during cooling, and in which solid phases maintain thermal equilibrium with

the remaining gas, and do not separate mechanically from the gas.

The simplest attempts (Lewis, 1972a; Grossman, 1972) to explain the mineralogy of the solar system implicitly assumed that the chemical composition of the solar nebula was spatially uniform, that the temperature and pressure fell off monotonically from the Sun, that equilibrium condensation occurred as the nebula cooled, that the solar wind terminated condensation by removing remaining gas, and that the planets accreted only from material in their immediate radial zone without significant overlap with the radial zones of the other planets. The detailed proposal by Lewis, which has been reiterated and copied by so many other investigators that it has almost become a dogma, is summarized in fig. 5. Mercury formed at 1440 K and $c.10^{-2}$ bar would contain a massive Fe, Ni core, a small mantle of Fe-free Mg-silicates and high-temperature Ca, Al, Ti-oxide condensates but only traces of alkali metals; sulphur and FeO should be absent unless 'due perhaps to "contamination" by infalling debris'. Venus formed at 940 K and $c.10^{-3}$ bar would have a relatively smaller core and larger mantle but each of similar chemical composition to those in Mercury; sulphur should be 'virtually absent'. Earth formed at 630 K would contain considerable sulphur as FeS melt, while Mars would contain only FeS in its core, and would contain considerable H_2O because it accreted tremolite. Unfortunately there may be

serious problems with this simple model, such as the presence of water on Earth, the relative density of Venus and Earth (Ringwood and Anderson, 1976), and especially the strong evidence for an oxidizing CO_2 -rich atmosphere on Venus. Furthermore, the model does not consider mechanical differentiation, overlap of accretion zones, and chemical effects produced by intense heating of an accreting or differentiating planet, as emphasized by Ringwood (1975, Chapter 16). Of course, the model can be patched, and indeed Lewis (1974b) proposed that late impacts from comets added volatile elements to Venus.

Related strongly to the above condensation model of the planets is the inhomogeneous accretion model proposed by Turekian and Clark (1969) and developed for the Earth by Clark *et al.* (1972). They adopted the concept of a hot nebula and assumed rapid cooling and direct accretion of the condensate into the Earth in not more than 10^5 yr. Effectively the Earth was supposed to accrete its iron-rich core and Mg,Si,O-rich mantle sequentially from material condensing at successively lower temperature from the solar nebula. Actually the first condensates should be rich in Ca,Al,Ti-

oxides, and the first Mg-silicates should be Fe-free. This simple model has been heavily criticized by scientists who favour accretion of planets via planetesimals, and it will not be considered further in its simple form; however, heterogeneity of accretion probably makes some significant contribution to the complex mineralogy of the planets, especially with respect to the outer zone.

Returning to the condensation sequence for C/O c.0.6 (Table IIIa), at a pressure of 10^{-4} bar hydrogen remains in the vapour until hydroxylated silicates are formed at about 400 K. The metal phases coexist with oxide phases, and there is no condition for condensation of reduced minerals such as Si-bearing metal, sinoite ($\text{Si}_2\text{N}_2\text{O}$), and osbornite (TiN). In order to explain such minerals in enstatite chondrites and achondrites, another mechanism is needed, probably involving a drastic composition change of the gas. Larimer (1975) invoked condensation at low pressure from a solar nebula with similar composition to that in Table II except that C/O is near 1.0 or higher (Table IIIb). If the enstatite chondrites and achondrites actually formed in the solar system, and were not brought in by comets, it will be necessary to find some mechanism for major change of the chemical composition of the nebula; indeed, if the E-asteroids really have surfaces that match enstatite achondrites, it will be necessary to modify a substantial region of the nebula. Perhaps some clues may arise from study of carbon-rich stars, for which calculations show that reduced species should occur in the surrounding nebula (Gilman, 1969). This is an extremely important problem that must be resolved: in the meantime, it will be assumed that at least the major part of the solar nebula had solar composition with C/O c.0.6.

Since the equilibrium condensation model reviewed by Grossman and Larimer (1974) relies on continuous adjustment between gas and crystalline condensates, diffusion barriers or mechanical separation may cause complications. How would grains of Fe,Ni become oxidized and the resultant FeO then enter Fe-free olivine or enstatite? Intervention of melting would ease problems, and indeed there is considerable evidence of melting in planetesimals and perhaps for dust (occurrence of chondrules). Blander and Abdel-Gawad (1969) attempted to explain the origin of meteorites in terms of nonequilibrium condensation constrained by occurrence of supersaturation and solid-state diffusion barriers, but Grossman and Larimer (1974) argued that meteorite chemistry is more consistent with equilibrium condensation.

An alternative approach (Arrhenius and Alfvén, 1971) emphasized ionization of the solar nebula, thereby ruling out simple condensation based on

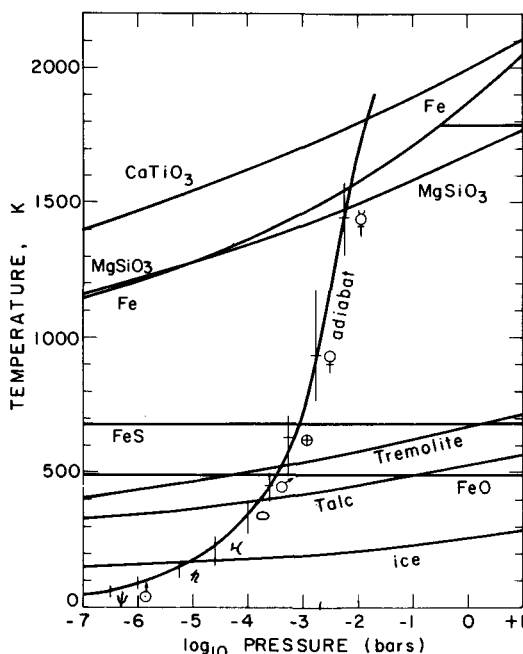


FIG. 5. Hypothetical condensation model of the planets based on direct equilibrium condensation of the solar nebula with C/O = 0.6 to a supposed adiabat (Lewis, 1972a, Fig. 1). See fig. 4 for details of condensation at 10^{-4} Pa ($= 10^{-4}$ bar). The symbols represent Mercury (upper), Venus, Earth, Mars, asteroids, Jupiter, Saturn, Uranus, and Neptune.

thermal equilibrium between gas and solid. Hydrogen, O, Cl, and Hg would be weakly ionized, whereas In, Tl, Bi, Pb, Pd, Ni, and Fe would be strongly ionized. Furthermore, solid particles could have a much different temperature of lattice vibrations than the kinetic temperatures of the electrons, ions, and atoms of the surrounding gas because of incomplete adjustment between energy transfers by radiative and collisional processes. Ion-sputtering experiments yielded crystalline and glassy aggregates on substrates at temperatures as low as 300 K. Several processes of chemical separation in a partially ionized nebula were also considered, especially with respect to a magnetic field associated with a solar wind. Arrhenius and Alfvén concluded that the meteorites might range between high-iron varieties formed from ion-rich plasma and carbonaceous varieties formed in a vapour dominated by neutral atoms and molecules. Meyer (1971) found that Ca was not fractionated from Al during sputtering, but that fractionation of Mg and Si with respect to Al and Ca occurred as a function of the substrate temperature. Electron diffraction study indicated that enstatite and two unknown crystalline materials were produced by sputtering at temperatures even as low as c. 400 K.

Physical interactions involving small bodies. Because small particles travelling at 1–50 m/s in near-vacuum ($c. 10^{-3}$ bar) or air would adhere spontaneously to targets with rough or dust-coated surfaces that ‘absorbed’ the impact energy, Hartmann (1978) deduced that no other mechanism was needed for accretion of condensed dust into meteoroid and asteroid-sized bodies. Growth would be faster for the larger bodies, and would accelerate when a regolith developed. When the larger bodies had reached 2–5 km radius, the impact velocities of the swarm would reach 2–50 m/s. Smaller bodies would become fragmented, and the fragments would either be swept up by the larger bodies or driven out of the solar system by radiation pressure. Finally the very largest bodies would ‘run away’ from the other large bodies as the collisional cross-section became augmented by the gravitational attraction.

The greater mechanical strength of iron than silicates allows the theoretical possibility that repeated collisions would result in large bodies of iron separating from a cloud of fine silicate debris (Whipple, 1964; Orowan, 1969). Shock metamorphism is a very complex process depending not only on the relative velocity of the colliding bodies and the mechanical strength of their minerals, but also on the relative porosity and initial temperature of the rocks. On average, however, for rocks starting at c. 300 K, feldspar and silica minerals begin to fracture upon release from

shock pressure achieved on the Hugoniot curve at 100 kb and to melt at 200 kb, whereas pyroxene and olivine are affected only at about twice the pressure (Stöffler, 1972, 1974). Kamacite transforms near 130 kb to ϵ -iron whose distinctive etch pattern reveals the solid-state recrystallization (Jain *et al.*, 1972). Shock loading causes hardening, and coherence is retained even at a peak pressure of 1 Mb, where kamacite merely undergoes partial recrystallization and shearing offset of inclusions of phosphide and taenite (Wasilewski, 1976). Perhaps iron fragments could accrete efficiently by welding together upon impact, so long as the temperature is high enough for ductile behaviour. The above pressures could be obtained only for high-impact velocities (c. 1–5 km/s?) when planetesimals had become large ($c. 10^2$ – 10^3 km radius). During early accretion the impact velocities should be too low for effective mechanical segregation of metal from silicate, especially as residual gas would cause viscous drag; however, at some stage of growth mechanical segregation can be envisaged. If stony-iron planetesimals had differentiated into an iron core and silicate mantle, impacts would tend to segregate silicate debris from surviving iron cores.

Magnetic attraction might cause mechanical separation of silicate from iron-rich metal (Curie point: Fe 1043 K, Co 1404, Ni 631) and perhaps magnetite. Thus Anders (1971b) emphasized that the Curie points of metal with 5–6% Ni in ordinary and enstatite chondrites would be somewhat above 900 K, which temperature is above estimates of final accretion temperatures of the Earth and of parent bodies of meteorites (c. 450–480 K; Grossman and Larimer, 1974). Particles smaller than 20 nm would have single magnetic domains, while those larger than 1 μ m would be multidomained. Only the former should have significant magnetostatic attraction.

Arrhenius and Asunmaa (1973) measured the electrostatic attraction of lunar soil particles and concluded that the range of adhesion force ($c. 10^{-4}$ to 2×10^{-3} N) and charge density (10^{-6} to 10^{-10} C/cm²) would cause particles 0.1 μ m across to adhere upon collision at a relative velocity up to 30 cm/s. Smaller grains could adhere at higher velocities, and pyroxene apparently adheres less easily than feldspar or glass.

Elastic collisions result in less massive bodies gaining velocity while the more massive bodies lose velocity. The latter should tend to move towards gravitational highs at the centre of the Sun and planets while the latter should tend to drift away. Since metal bodies are stronger than silicate bodies, disintegrative collisions might ultimately have resulted in Fe-rich metal drifting towards the Sun, and locally towards the planets, while silicate

tended to move away. Perhaps this factor may be important in yielding a high Fe-content for Mercury, and a higher Fe content of the Earth than the Moon (Kaula, 1977a, b).

Physical and chemical differentiation involving large bodies. Whatever the details, nebular condensates must have aggregated by several processes, almost certainly at low relative velocity ($c.1.5 \times$ escape velocity of the larger bodies; Hartmann, 1978). The resulting 'clouds' would sink by gravitational attraction to the medial plane of the nebula, and would accrete into planetesimals (Goldreich and Ward, 1973; Safronov, 1972). Many features of meteorites are explainable by chemical differentiation in planetesimals at least tens of km in diameter, and perhaps hundreds of km; furthermore, the main-belt asteroids range up to 1000 km radius. Consequently this paper follows the lead of Urey (1952) in building planets from large planetesimals, rather than assuming continuous direct accretion of condensed material on to planets (Clark *et al.*, 1972).

The dynamics of two-body interaction are fundamental to discussion of the physical and chemical differentiation produced by collision and near-collision. Interaction of two bodies in a single event leads to either a direct hit, a glancing hit, a miss within the Roche stability limit, or a miss outside the Roche limit (fig. 6).

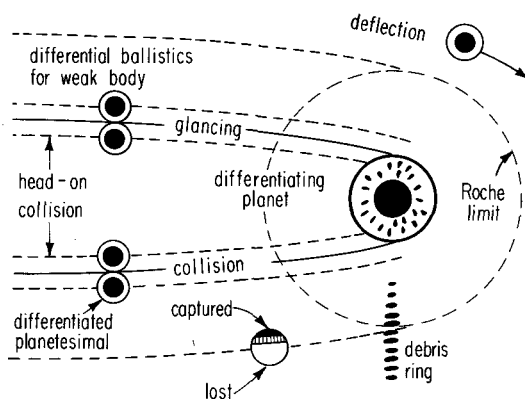


FIG. 6. Cartoon of possible two-body interactions. The inner zone for head-on collision is enclosed by the two zones for glancing collision and disintegration of a *weak* body in the Roche stability zone, and by the region just outside the Roche zone, for which substantial deflection ('sling-shotting') occurs.

The collisional radius R_c of a body with geometrical radius R_g and escape velocity V_e for smaller incoming bodies of non-gravitational relative velocity V_i is augmented by gravitational focusing:

$$R_c = R_g(1 + V_e^2/V_i^2)^{1/2}$$

This effect is shown schematically in fig. 6 by the tapering area marked 'head-on collision'. Reducing V_i/V_e increases R_c/R_g especially as V_i falls below V_e ; thus for an olivine-rich planetesimal of radius $c.2000$ km with escape velocity $c.3$ km/sec, the collisional radius is 2100 km for V_i at 10 km/sec, 2800 km for 3 km/sec, 6000 km for 1 km/sec, and extremely large as V_i falls below 100 m/sec. This is a major cause of the 'tadpole' effect in which the larger members of a population of planetesimals 'eat' the smaller ones (but see later for complications).

Head-on collision could lead to disruption of both bodies with dispersal of the debris, but a large enough body would survive the impact and capture the smaller body. The products of the impact are governed principally by the strength of the bodies, the impact velocity, and the starting temperature. Increasing the escape velocity increases the fraction of captured debris; increasing the relative velocity increases the maximum shock temperature and pressure until ultimately the impacting projectile explodes into a cloud of vapour, which tends to escape; increasing the ratio of the strength of the small body to the large body (which might have either an uncompacted regolith or a surface weakened by volcanic activity) results in deeper penetration and greater chance of accretion. A thick atmosphere, and especially a ring of debris, could have major effects on the incoming body and on the impact products. Modelling major impact events is extremely difficult even when data for natural and man-made explosions are used as a control. Many pertinent references are listed in Head *et al.* (1975), O'Keefe and Ahrens (1975), and Wetherill (1976). Recent calculations by O'Keefe and Ahrens (1977b) suggest that for the present Moon, Mercury, and Mars (assumed to have a feldspar-rich surface), meteorite impacts result in a mass gain only for impact velocities less than 20, 35, 45 km/s, respectively, for feldspar-rich meteorites and less than 25, 35, and 40 km/s for iron meteorites. Hence these small planets would retain much less of the material deflected into the inner solar system by the giant planets than would Earth and Venus.

Glancing collision formally occurs in an annular region of width equal to the diameter of the incoming body (fig. 6). The fate of the debris from glancing or near-glancing collisions was predicted by Öpik (1972), Hartmann and Davis (1975), and especially by Cameron and Ward (1976) with reference to possible formation of the Earth's Moon from a Mars-sized planetesimal hitting at $c.10$ km/s. The outward-facing part of the planetesimal might shear off and fragment whereas the inward part might undergo severe shock including vapourization of the silicate. The combination

of differential ballistics with shock-induced dispersal followed by coagulation of only part of the debris might lead to rapid formation of a refractory, metal-poor Moon near the Roche limit, but it will not be easy to quantify the qualitative suggestions (see Kaula, 1977*a*, for some sceptical remarks).

The next annular zone involves passage through the Roche stability limit for self-gravitation. The classical mathematical treatment by Darwin (1906) is for a tiny satellite of no mechanical strength rotation-locked in a circular orbit about the central body. Inside the Roche radius R_r , the self-gravitation is insufficient to compensate for the disruptive tidal forces. A simple three-body treatment (e.g. Haymes, 1971, pp. 207–8) gives

$$R_r = (6M_c/\pi\rho_s)^{\frac{1}{3}} \quad (3)$$

where M_c is the mass of the central body and ρ_s is the satellite density. The two-body tidal treatment gives

$$R_r/R_c = 2.455(\rho_c/\rho_s)^{\frac{1}{3}} \quad (4)$$

where R_c and ρ_c are the geometrical radius and density of the central body (Darwin, 1906). For the present Earth (Smith, 1974), R_r/R_c ranges from 2.24 for liquid iron ($c.7 \text{ g/cm}^3$) to 2.88 for the bulk density of the Moon (3.34) to near 3 for feldspar-rich liquid ($c.2.6$). Because iron should tend to accrete earlier than silicate into planetesimals because of mechanical stability, ρ_c should tend to be greater than ρ_s ; as a rough approximation, the Roche limit is placed at $R_r \approx 3R_c$ in fig. 6.

Several complications arise. Mechanical cohesion can prevent gravitational disruption inside the Roche limit (Jeffreys, 1947; Mitler, 1975); this effect decreases when the incoming body is hot, and especially if it is partly molten. Relative rotation of the incoming and central bodies, and especially a non-circular orbit of the incoming body, increase the instability region. The last factor is part of the very complex problem of tidal influence on satellite orbits (Kaula and Harris, 1975, pp. 366–8). Obviously the Roche limit is merely a theoretical concept, and many processes can actually occur within several radii of the central body.

The simplest model for disintegration inside the Roche limit involves passive fragmentation of the incoming body into pieces that move independently (Wood and Mitler, 1974) with those nearest to the central body going into elliptical orbits about it, those farthest from the body being lost into hyperbolic orbits, and those in an intermediate zone going into such elongated orbits that escape is likely. For capture of a significant fraction, V_i must be less than 2 km/s, which for the Earth requires that the semi-major axis of an

incoming body must lie between 0.8 and 1.2 AU (Kaula and Harris, 1975). Unfortunately, the simple calculation is only a mathematical boundary condition because fragmentation would not be passive, the fragments would not simply assume the velocity of the parent planetesimal, and much of the material that flew off into heliocentric orbit would ultimately impact the Earth and any surrounding cloud or moon(s) on succeeding orbits.

The fourth zone in fig. 6, which must overlap the Roche zone for strong bodies, involves merely orbital deflection and trivial tidal heating. In a simple two-body encounter the effect is apparently trivial but taken in the context of the entire system of bodies orbiting the Sun this ‘slingshot’ effect is extremely important as it changes the eccentricity of the orbit thereby causing overlap of accretion zones of the planets (see later).

Finally the simple two-body encounter must be extended to include collision with any debris ring (including possible moons) in orbit around a growing planet or planetesimal. Kaula and Harris (1973) showed that collision of a Moon-sized object with 2×10^{-5} of its own mass in Earth orbit dissipated as much energy as by the tidal effect during a non-collisional near-encounter. Ruskol (1972) extended earlier papers on the origin of the Moon by discussing how an incoming body could either destroy bodies in geocentric orbit or be captured to augment the orbiting material. In general the larger the body or bodies in Earth orbit, the greater their chance of survival and the less the likelihood of impact debris ending up on Earth. Of course, a very thick, deep atmosphere such as the present atmosphere of Venus, and particularly the early planetary atmosphere postulated in the single-stage model of Ringwood (1966*b*), would contribute to the effect of a debris ring.

Attention became focused on disintegrative capture because of the need to explain the coupling of the low iron and volatile contents with the somewhat high content of anorthite inferred for the bulk Moon with respect to the Earth. Numerous authors listed the possible chemical effects that could accompany the dynamic effects of disintegrative capture (e.g. Kaula, 1971, 1977*a*; Ruskol, 1977), but quantification is difficult. An incoming body might have chemically differentiated into a metal core and a silicate-rich mantle and crust, thereby permitting several types of mechanical separation such as stripping off the outer part by collision with debris in Earth orbit (Kaula and Harris, 1973), ballistic differentiation (Wood and Mitler, 1974), and momentum differentiation of massive Fe-rich fragments with less massive silicate-rich ones (Smith, 1974; Kaula, 1977*a*). Major loss of volatiles could

occur in any collision involving relative velocities greater than a few km/s: of course, the greater the temperature rise the greater the range of volatile elements that would be significantly depleted. Although these processes cannot be quantified it will be assumed that they typically occurred during the interaction of planetesimals with growing planets and that they substantially modified the primary condensates from the solar nebula.

Before facing up to multi-body problems it is convenient to mention briefly the problems of tidal interactions. Once a moon has formed it causes tidal perturbations on its planet, and undergoes perturbations itself. Even ignoring impacts by late accreting bodies the tidal evolution can be complicated (Kaula, 1971; Kaula and Harris, 1975). A satellite orbiting sufficiently slowly with respect to the rotation of the central body will recede if the angular momentum vectors are parallel (i.e. prograde rotation of the satellite), and will spiral inwards for retrograde rotation. If the central body is rotating very slowly with respect to the orbital period of the satellite, the latter can be captured whatever its direction of rotation (Burns, 1973; Counselman, 1973; Ward and Reid, 1973), especially when the central body is close to the Sun. Indeed Singer (1970) postulated that the slow retrograde rotation of Venus results from capture of a retrograde moon. Tidal recession of a satellite results in rotation-locking such that the satellite always shows the same face to the central body, except for small librations. Many complications arise if the two bodies change mass and velocity during simultaneous accretion (Harris and Kaula, 1975).

Because of the existence of mechanisms for destruction of satellites, the absence of satellites around Venus and Mercury, and the small size of the ones around Mars, will not be taken to imply that substantial debris clouds and perhaps large moons never existed around these terrestrial planets. Furthermore, the presence of only one Moon around the Earth does not preclude the earlier existence of several moons and a debris cloud. Indeed Ruskol (1960, 1963, 1977) and Öpik (1972) developed detailed mechanisms for generation of the Earth's moon by aggregation of several moonlets produced from a thick debris cloud, and Harris (1978) presented further considerations.

Planetary accretion: dynamics, time scale, and heat sources. The dynamical processes of planetary accretion from planetesimals are too complex to be reviewed satisfactorily here, and indeed the calculations are highly dependent on the uncertain physics of inelastic collisions. Readers are referred initially to the book by Safronov (1972), which covers the ideas developed in the 1960s by Soviet mathemati-

cians, and to the book by Öpik (1976). The present account briefly abstracts the summary in Safronov (1977), and refers to some recent amendments.

At the beginning of accretion of the planetesimals the collision probabilities can be calculated by statistical models. As the protoplanets grow the gravitational potential becomes non-uniform, and the two-body collision outcomes become more complex, thereby resulting in the mathematical models becoming less accurate. Ultimately most of the mass has entered the planets, and the motions of the surviving planetesimals can then be modelled (again statistically) by Monte Carlo calculations, whose accuracy depends on the validity of the assumed mass-space-velocity distribution of the planetesimals and on the rate of accumulation of small errors caused by approximations in the successive orbits.

A basic equation for capture of small planetesimals by head-on collision with a protoplanet is:

$$dm/dt = 4\pi R_g^2(1 + V_e^2/\bar{V}^2) \cdot \sigma/P$$

where dm/dt is the increase of protoplanetary mass with time, R_g is the geometrical radius, V_e is the escape velocity, $\bar{V}$ is the mean relative velocity of small planetesimals with respect to the planet, and σ is the surface density of planetesimals in the zone swept out by the protoplanet as it orbits around the Sun with period P . Of key importance is the increase of capture cross-section with escape velocity as described in the preceding section, leading to a few large bodies capturing the smaller ones. If the planetesimals have exactly circular orbits in the ecliptic plane each protoplanet could sweep up material only from a torus of area $\pi R_g^2(1 + V_e^2/\bar{V}^2)$, and from a larger volume for bodies precessing outside the ecliptic plane. Thus continued growth depends on capture of material with non-circular orbits, and here is the crucial and probably insoluble problem. Elastic collisions between numerous small bodies cause 'jostling' and a tendency for the bodies to adopt circular orbits with very small relative velocity (as in Saturn's rings). Accretion of planetesimals may begin by gentle aggregation into an enormous number of small bodies (< 1 km radius) composed of lumps that have fallen into the plane of the ecliptic after condensation. By chance a few of these small bodies will coalesce thereby reducing the population, but planets can form only as the planetesimal orbits become non-circular, and indeed the final stage of accretion must involve capture of planetesimals with orbits at least as eccentric as half the distance to the next planet.

Unfortunately it is not possible to accurately calculate the changes of $\bar{V}$ and orbital eccentricity as the planet grows. The key factor is the relative

$c.10^8$ at -4.0 , and $c.10^9$ at -1.0 . Calculations of the resulting mass distribution yielded a mean lifetime of $c.3 \times 10^7$ yr for 10^{24} g bodies and $c.2 \times 10^6$ yr for 10^{22} g. Using Wetherill's procedure, Hartmann (1976) estimated that about half of the late accretion of Mercury, Venus, Earth, and Mars came from planetesimals that began in the feeding zones of other planets. Furthermore, he suggested that the Moon received 60% of its late accreted mass from planetesimals that grew in the Venus zone. These calculations refer only to an assumed population of late planetesimals, and only further study can test whether the feeding zones of the planets would overlap significantly at an earlier stage of growth. In this lecture it will be assumed that at least 90% and probably at least 95% of the mass of an inner planet is derived from material accreted near its present orbit, and that the bulk composition is indicative of the condensed material of the local zone of the nebula: however, it would be wise to investigate the consequences of greater mixing.

Safronov (1977) deduced that the Earth accreted with a half-life of $c.10^7$ yr and that 98% accretion occurred in $c.10^8$ yr. Interpretation of radiometric chronology of the early solar system is not easy, and may need revision if the solar nebula was incorporating supernova debris. Wetherill (1975) interpreted the evidence for meteorites in terms of rapid accretion of small bodies into planetesimals ($c.100$ km radius) in $c.10^4$ yr with subsequent heating and cooling over 10^7 – 10^8 yr. Roughly 10^8 yr is needed from the nebular period to establishment of the reservoirs of lead in the Earth.

A serious problem for simple accretion models is the relative time scale for accretion of the inner and outer planets. The existing accretion models of Safronov-type lead to much longer accretion times for the outer than for the inner planets, as can be deduced simply but not rigorously from the σ/P term in the earlier equation for dm/dt . If the Earth accreted in 10^8 yr, Mars and the outer giant planets would have required at least 10^9 yr, according to the simple Safronov model (Safronov, 1977; Weidenschilling, 1976). Since accretion of the outer planets should have resulted in deflection of many planetesimals into the inner solar system, and the simple interpretation of the morphology of Mars assumes essential completion of accretion in time for a differentiated crust to retain a record of the same population of bodies as formed the lunar basins, it is desirable to investigate models for more rapid accretion of the outer planets, and perhaps for the inner planets as well. Weidenschilling pointed out that the Safronov model actually gives a low estimate of accretion time because it neglects the difficulty of capturing planetesimals at the edge

of a feeding zone, thereby accentuating the problem.

The Safronov model predicts that the mass distribution of bodies in each feeding zone is very strongly biased. The protoplanet is by far the largest body, and the simple model suggests that the next largest body has not more than $c.1\%$ of the mass of the protoplanet. If true, this would rule out the model of disintegrative capture of a Mars-size body from whose chemically differentiated debris the Moon was formed. However, Hartmann and Davis (1975) made further calculations, which suggest that the second-largest body in the Earth zone might have a radius of 500 to 3000 km, which upper limit corresponds to a volume five times that of the Moon. Wetherill (1976) questioned the accuracy of the Safronov-type model and suggested change in the parameter θ . I suggest that large bodies might develop at the interfaces of planetary feeding zones because of lack of competition with protoplanets when $\bar{V}$ is low, and recommend investigation of the possible capture by the Earth of a large body growing between Earth and Mars (fig. 7).

The present masses of Mars and the asteroids are much lower than the estimated mass of non-volatile material in the relevant zones of the solar nebula (cf. Wetherill, 1976; Weidenschilling, 1976) and probably an enormous mass of material was ejected from these zones by Jupiter (Ip, 1978). Most of the mass would be ejected outwards with some ending up as comets and some colliding with planetesimals in the outer solar system to end up in the giant planets, but a significant fraction should have ended up in the terrestrial planets. Together with the material deflected by Saturn, Neptune, and Uranus, there is scope for developing theories of bombardment of the terrestrial planets by a considerable number of volatile-rich bodies even though most of the material would arrive at such high velocity (especially at Mercury) that perhaps only the larger planets Earth and Venus would be able to retain the collision debris while the smaller planets Mercury, Moon, and, to a lesser degree, Mars, would lose most of the debris because of their lower ratio of escape velocity to impact velocity. Consequently it is not unreasonable to invoke the qualitative possibility of late heterogeneous accretion of volatile-rich material on the larger terrestrial planets. Furthermore, it is possible to envisage that the planetesimals of the inner part of the solar system (Mercury–Mars) were almost free of volatiles (as indicated by many meteorites), and that substantial amounts of volatiles condensed only from the asteroid zone (or perhaps Mars zone) outwards [this refers to volatiles condensing below 600 K, and perhaps as high as 1100 K]. Indeed Sill

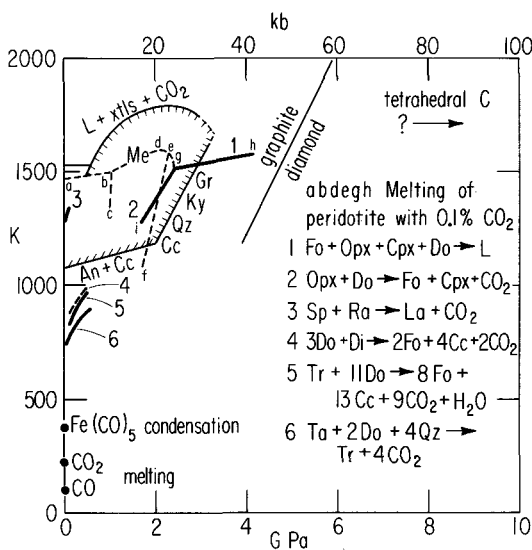
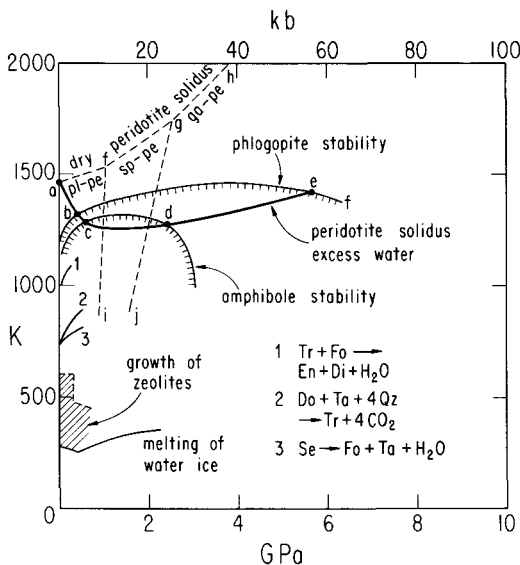
and Wilkening (1978) concluded that the atmospheres of the terrestrial planets cannot be explained by condensation models in which the Earth accreted at 500 K, and suggested that the noble gases of the Earth were provided by about 1 ppm of clathrates, which originally condensed in the outer solar system.

Finally, heat sources need discussion (Safronov, 1977; Weidenschilling, 1976). Conversion of gravitational energy of infalling bodies into heat is trivial for small planets and planetesimals, and becomes important only when planets have grown large. Estimates of $\bar{V}$ from Safronov-type models are too low for significant heating from the collision of planetesimals whose velocities have not been 'pumped up' by mature planets or the Sun, but can provide substantial heating when $\bar{V}$ reaches several km/s. The higher values of $\bar{V}$ envisaged by Wetherill (1976) may be significant. Heat loss from a planet's surface becomes more important when accretion slows down. Models of radioactive heating do not lead to early melting (at $\sim 4.5 \times 10^9$ yr) of a terrestrial planet unless the accreting planetesimals are hot (> 1000 K), whereas it is easy to find conditions for late melting of at least the outer zone. Actually there is so much mineralogical evidence for high temperatures in differentiated meteorites that even small planetesimals ($c. 10^2$ to 10^3 km radius) of the inner solar system must have

reached high-metamorphic temperatures or even melted. Perhaps short-lived radioactive elements (e.g. ^{26}Al) were responsible, as proposed by many scientists (see Herndon and Herndon, 1977), or perhaps some other mechanism or mechanisms (including electromagnetic induction heating; Sonett *et al.*, 1970) were responsible. For the rest of this lecture, it will be assumed that most planetesimals in the region of the terrestrial planets reached or nearly reached melting temperatures, but not necessarily those in the region of the giant planets, and that terrestrial planets accreted from hot material and underwent an early melting with out-sweeping of the elements that entered H_2O - CO_2 -bearing melts before accretion had ended.

Chemical differentiation inside planetesimals and planets. This section begins with a review of phase equilibria particularly pertinent to crystal-liquid differentiation in terrestrial planets and meteorite parent-bodies.

Figs. 2, 8, and 9 show the important phase boundaries for the major minerals and rocks expected in terrestrial planets. Crystal-liquid differentiation of a bulk composition composed only of enstatite (MgSiO_3) and iron metal would be governed by the melting curves (fig. 2) for iron (Sterrett *et al.*, 1965) and enstatite (Boyd *et al.*, 1974). Substitution of 10% Ni and 0.5 wt% Co would reduce the melting point of metallic iron by



FIGS. 8 and 9: FIG. 8 (left). Possible phase relations involving H_2O . Di diopside, Do dolomite, En enstatite, Fo forsterite, Qz quartz, Se serpentine, Ta talc, Tr tremolite. FIG. 9 (right). Possible phase relations involving CO_2 . An anorthite, Cc calcite, Cpx clinopyroxene, Di diopside, Do dolomite, Fo forsterite, Gr grossular, Ky kyanite, La larnite, Me meionite-sapolite, Opx orthopyroxene, Qz quartz, Ra rankinite, Sp spurrite, Ta talc, Tr tremolite.

only 10 K. Forsterite (Mg_2SiO_4) has an even higher melting point than enstatite. Thermochemical models of the solar nebula predict condensation of Ca,Al,Ti,O-rich mineral assemblages prior to condensation of Mg-silicates and Fe,Ni metal. Detailed melting studies have not been made on the specific bulk composition implied for Mercury by the model of Lewis (1972a), but the phase equilibria for 'peridotitic' compositions have been studied extensively. Peridotite compositions (Wyllie, 1971) free of H_2O and CO_2 begin to melt along the curve *afgh* (fig. 8) producing increasing amounts of melt as the temperature rises. The melt is 'basaltic' in composition near the solidus and becomes increasingly richer in pyroxene and olivine components as melting proceeds. Below the solidus the peridotitic compositions crystallize into assemblages of olivine and pyroxenes with either plagioclase (low pressure field: *pl-pe* outlined by *afi*), or spinel (medium pressure field: *sp-pe* outlined by *ifgj*) or garnet (high pressure field: *ga-pe* outlined by *jgh*). At even higher pressures peridotitic compositions yield oxide and perovskite minerals. Whatever the details of the condensation of the solar nebula and accretion of planetesimals, some approximation to 'peridotitic' composition can be expected for all the inner planets, and the 'dry peridotite solvus' marks an upper limit to the beginning of melting. See the Proceedings of the Lunar Science Conference for experimental details of many specific compositions.

The phase relations change drastically and become very complex indeed as soon as Na, K, H_2O , and CO_2 enter the planetary composition. Sodium can be accommodated in plagioclase and pyroxene minerals, but will tend to enter magma produced by melting of peridotitic composition. Potassium does not enter the dense minerals composing dry peridotite (except at pressure over 40 kb) and is strongly concentrated into magma or slightly into plagioclase. Addition of H_2O and CO_2 causes drastic effects whose details have been explored only in the last decade. The present account is based on a forthcoming review of the peridotite- CO_2 - H_2O system by Wyllie (1978) and utilizes a series of studies including Eggler (1976, 1977), Wyllie and Huang (1976a, b; 1977), and Mysen (1977).

In the presence of H_2O , Na, and K (as for peridotites on Earth, probably on Mars, and possibly on Venus), amphibole and mica become stable, as delineated by Boettcher and co-workers. The stability limits of these two minerals vary with many factors such as $P_{\text{H}_2\text{O}}$, Mg/Fe, Ca/Na, etc., but the two generalized curves in fig. 8 are within 100 K and 5 kb of most conditions to be expected in Earth and Venus. Mg-rich mica (phlogopite) is stable to higher pressures and temperatures than amphibole,

and a K-bearing peridotite will begin to melt along the curve *abcde* (fig. 8). Roughly speaking, H_2O can be retained by mica and amphibole in a wet planet up to c. 1200–1400 K. Titanoclinohumite and titanian chondrodite are very rare minerals in the Buell Park kimberlite, and might be important in subduction zones of the upper mantle (Akimoto *et al.*, 1977).

Presence of CO_2 leads to phase relations that are even more complex, but the main features are illustrated in fig. 9. Melting in the presence of 0.1 % CO_2 begins along the curve *abdegh*, whose temperature varies only about 100 K for the pressure range 0 to 40 kb. The maximum at *d* was not properly recognized in early work, and is important for models of explosive loss of CO_2 from rising peridotitic melt. Fields *abc*, *cbdef*, and *fegh* outline the stability of plagioclase-, spinel-, and garnet-peridotites. Dolomite-rich carbonate (abbreviated *Do*) is stable in the region bounded by curve *gh* (no. 1) for the melting of forsterite, two pyroxenes, and *Do*, and by curve *gi* (no. 2) for the transformation of orthopyroxene + *Do* to forsterite, clinopyroxene, and CO_2 . Curve 2 extrapolates into the cluster of curves 4, 5, and 6 for the breakdown of dolomite in the presence of various silicates at low pressure (see later). Presence of both H_2O and CO_2 leads to subtle complexities reviewed by Wyllie (1978).

Sulphur is a common constituent of meteorites as troilite, pyrrhotine, or sulphate minerals. If it occurs as troilite in accreting planets, the Fe,Ni metal could react with the troilite at the Fe,FeS eutectic (Usselman, 1975), which changes composition from 27 wt % S at 30 kb to 24 % at 100 kb. The Fe,FeS eutectic stays at c. 1250 K for pressures up to 55 kb, and increases slowly for higher pressures.

Limits can now be placed on the initial melting of a planetesimal or planet containing volatile elements. If S is present and K, Na, H_2O , and CO_2 are absent, melting could begin at c. 1250 K leading to sinking of a heavy Fe,S-rich liquid (together with associated Fe,Ni metal) to form a core surrounded by a mantle of unmelted peridotite. Further heating would lead to 'peridotite' melting near the 'dry peridotite solvus'. Addition of Na and K to the peridotite would lead to somewhat lower melting temperatures, but not as low as the Fe,FeS eutectic, while addition of CO_2 and especially of water would lead to considerably lower temperatures. Indeed hydrous K-bearing peridotite would melt near the Fe,FeS eutectic for pressures of 10 to 30 kb, and 'basaltic' melt could rise as Fe,S melt descended. Actually there are minor complexities involving the mutual solubility between basaltic and sulphidic melts. Perhaps such a double differentiation occurred as the proto-Earth was growing, but perhaps water was absent in the early stages

with differentiation involving mainly the sulphidic component. For the double-differentiation, sulphur- and metal-loving elements would concentrate into the core while magmaphile elements would enter the 'basaltic' fraction. Particularly important would be the fate of radioactive elements: for double-differentiation, the U, Th, and K would probably concentrate almost entirely into the 'basaltic' fraction, but for H_2O -free differentiation, these elements might be fractionated substantially into the sulphidic liquid, especially if a highly reduced state leads to an increased chalcophilic behaviour of these elements (Smith, 1978).

The pressure-temperature curve for a proto-planet depends critically on the temperature of accreting material, the collisional heating of accretion, and the heat loss from the surface. All three factors are highly uncertain, but most simple models yield a temperature maximum closer to the surface than the centre, often about one-tenth to one-quarter down. Sinking Fe,S-rich diapirs and rising 'basaltic' diapirs can be envisaged as partial melting begins. Gravitational energy is released by diapiric separation, and catastrophic heating and differentiation might occur for an Earth-sized planet, as suggested by Urey (1952) and emphasized by Ringwood (1966*b*, 1975). However, such catastrophic differentiation need not occur for a proto-planet that begins to differentiate early in its growth.

Accretion of large planetesimals ($c.10^2$ to 10^3 km radius) should cause catastrophic crystal-liquid differentiation in large regions of a growing planet, especially if the planet is at near-melting temperatures. Large overturns might occur, and the possibility that the outer part of the growing Earth was essentially 'stirred' by impacting bodies should be considered.

A 'basaltic' veneer on the Earth could undergo many complex differentiations as secondary processes (igneous, metamorphic, and ultimately sedimentary) occurred. Two important considerations are the relative melting relations of wet vs. dry 'basalt' and of wet vs. dry 'granite', for which generalized curves (fig. 2) are abstracted from Wyllie (1971, 1978), Stern and Wyllie (1973), and Winkler (1974). These curves are not pertinent to the primary differentiation of planets because the starting compositions are dominated by olivine and pyroxene, which components must be segregated into a mantle before the feldspar minerals and quartz can dominate the phase relations of upward segregates.

After cessation of bombardment by planetesimals the Earth apparently developed a 'granitic' crust and water-borne sediments by -3.8×10^9 yr. The development of plate-tectonic processes is

controversial, but present evidence indicates subduction to at least 500 km and perhaps deeper (Davies, 1977). Theoretically, one can imagine that a cooling planet could resorb its hydrosphere (and some of the atmosphere) when the radioactive heating diminished enough to allow hydrous minerals and carbonates to be stable at considerable depth. For this to happen subduction must carry volatile-rich sediments deep into the planet. Much of the lower crust and upper mantle of the present Earth would adsorb H_2O and CO_2 if transported downwards from the hydrosphere and upper crust. Such readsorption could occur even if parts of the upper mantle were still outgassing volatiles by explosive volcanism. If magmas from the upper mantle could react with a granulitic lower crust during transport to the surface, they could lose their volatiles: indeed the concept of volatile traps corresponding to the stability limits of mica, amphibole, and other minerals would become viable.

Much further work is needed to explore the possibilities of primary and secondary crystal-liquid differentiation of a growing terrestrial planet, and the even more complex possibilities of return of volatile elements into a dying planet. Further phase equilibria are now explored to place limits on mineral reservoirs for H_2O and CO_2 .

Fig. 8 pertains to H_2O . The melting temperature of H_2O -ice varies slightly with pressure (Clark, 1966) and may provide an effective lower boundary for the crystallization of zeolites. On Earth large beds of zeolites formed from alteration of volcanic ash (Munson and Sheppard, 1974). The stippled area in fig. 8 for growth of zeolites under metamorphic conditions is limited by the reactions of breakdown of heulandite, laumontite, and wairakite (Winkler, 1974, Fig. 12—11). Analcime is stable only to about 480 K in the presence of quartz, but is stable to magmatic temperatures in some alkaline compositions (Kim and Burley, 1971). The breakdown of serpentine to forsterite + talc + water (curve 3) is one of many possible reactions involving hydroxylated layer silicates (Winkler, 1974, Fig. 11—5). For amphiboles the stability depends strongly on the chemical composition. Tremolite and forsterite react to enstatite + diopside + H_2O some 300 K higher than the breakdown of serpentine, and anthophyllite breaks down to enstatite + quartz + H_2O at a similar temperature to that of tremolite (Winkler, 1974, Fig. 11—7). Aluminium-rich hornblendes from peridotitic rocks are stable to even higher temperatures, and the dashed curve marks the approximate limit based on many experimental syntheses (Mysen and Boettcher, 1975). For micas the maximum temperature stability is for Fe-free phlogopite, which breaks down to forsterite

+ liquid in the presence of enstatite (Modreski and Boettcher, 1972). Fe-bearing phlogopite found in xenoliths from the Earth's mantle is stable to a slightly lower temperature. If the Martian rocks are very Fe-rich the stability fields of both mica and amphibole would be lowered several hundreds of degrees with respect to curves in fig. 2. Phase relations for muscovite (and talc) are given by Wyllie (1973) and Huang and Wyllie (1974).

All the above phase equilibria are for $P_{\text{H}_2\text{O}} = P_{\text{total}}$. Phase relations for systems containing both CO_2 and H_2O are very complex, as reviewed in Winkler (1974), but the reaction dolomite + talc + quartz going to tremolite + CO_2 (fig. 2, curve 2) is fairly straightforward because it applies only to a high ratio of CO_2 to H_2O .

Many other minerals should be added to fig. 8 including the epidote and chlorite groups, montmorillonite, and staurolite. Systematic perusal of Winkler (1974) will reveal details which cannot be given here. Comparison of figs. 2 and 3 shows that theoretically many minerals could store water in all the terrestrial planets except for Venus in which only the amphibole and mica groups are serious candidates. Of course, water-bearing minerals would not occur if the planet were bone-dry, as apparently is the Moon.

Because of the presence of CO_2 in Mars, Earth, and Venus, and the high abundance of C in the Sun, a skeletal review of the phase relations of C-bearing species is useful (fig. 9). Carbon monoxide and CO_2 melt near 74 and 216 K at low pressure; the former boils at 82 K and the latter sublimates at 195 K (Weast, 1969). The melting point of CO_2 increases to 400 K at 15 kb (Clark, 1966, p. 373). Methane melts at 91 K and boils at 109 K (Weast, 1969), which values are important for the giant planets. The $\text{CO}_2 \cdot 6\text{H}_2\text{O}$ clathrate, isostructural with *craigite*, breaks down to H_2O -ice and CO_2 at 150 K and 0.004 bar and at 190 K and 2 bars (Miller and Smythe, 1977).

Iron occurs as a vapour deposit in lunar rocks, and transportation as pentacarbonyl (melting point 262 K, boiling point 374 K; Weast, 1969) may be responsible.

The surface temperature of Venus (c. 740 K) is close to the range of temperatures at 10^7 Pa for a group of decarbonation reactions. Three reactions involving dolomite and various silicates are shown in fig. 9 (curves 4, 5, and 6), and many other reactions, together with a thorough description of the complexities of the phase relations are given in Winkler (1974). Reaction of spurrite + rankinite to larnite + CO_2 (curve 3) marks the upper temperature stability of calcium carbonate silicates.

The meionite variety of scapolite has a wide stability range at high temperature and pressure in

dry bulk compositions rich in the anorthite component (Goldsmith, 1976; Newton and Goldsmith, 1975). Its stability field is bounded by reaction to either anorthite + calcite, or grossular + kyanite + quartz + calcite, or liquid + crystalline phases + CO_2 . Meionite is only one of the components of natural scapolites on Earth, but exemplifies the high pressure and temperature to which this mineral group is stable. Scapolite occurs in deep-seated granulite rocks and could trap CO_2 escaping from the Earth's mantle. Nitrate-scapolite (Goldsmith, Newton, and Moore, 1974) and nitrate-cancrinite are also stable at high P and T , but probably do not occur on Earth and on other planets because the oxidation state is too low. However, sulphate-scapolite (Goldsmith, 1976) is stable on Earth (e.g. in the Sittampundi Complex and in Australian *granulites*, Lovering and White, 1964), and must be considered as a possible storage site for sulphur in other planets.

The eruption of carbonatite and kimberlite rocks on Earth testifies to the storage of CO_2 in the upper mantle, and comparison of figs. 2 and 9 shows that similar storage is theoretically possible on other planets if their bulk compositions are suitable. Under reduced conditions (Brey and Green, 1977; Rosenhauer *et al.*, 1977), carbon can exist as either graphite or diamond, and occurrence of these polymorphs in mantle-derived xenoliths is consistent with the experimental transition curve (Boyd, 1973; Dawson and Smith, 1975; McCallum and Eggler, 1976). In the deep mantle carbon might be envisaged as substituting for silicon in tetrahedral coordination, but this has not been demonstrated in natural minerals. Cohenite, $(\text{Fe}, \text{Ni})_3\text{C}$, occurs in iron meteorites, but the conditions for its stability are uncertain (Anders, 1964), and do not require high pressure as was once believed.

The occurrence of sulphur in the planets is governed by many complex phase equilibria. Progressive condensation of the solar nebula should give large amounts of FeS (troilite), which mineral is indeed present in iron meteorites and as an accessory in other meteorites and lunar rocks. Sulphur is an obvious possible constituent of metal-rich cores for the terrestrial planets, as proposed for Earth by Murthy and Hall (1970). Small amounts of Fe, Cu, Ni-sulphides occur in mantle-derived xenoliths (Bishop *et al.*, 1975), and sulphur dissolves in small amounts (0.1 wt%) in mafic and ultramafic melts (Haughton *et al.*, 1974; Shima and Naldrett, 1975). From electron microprobe analyses, Anderson (1974) found 0.005–0.28 wt% S in glasses trapped inside megacrysts of volcanic rocks from Hawaiian and circum-Pacific areas, whereas vesicular glasses contained only 0.005–0.04 wt%. Apparently the magmas were

saturated before eruption with respect to a sulphide melt, and lost considerable SO_2 during eruption. The geochemical distribution for sulphur is very complex, involving emission of volcanic gases into the atmosphere, and transport into aqueous and solid reservoirs, and probably some recycling back into volcanic environments. A review of the phase relations in the system $(\text{Na}, \text{K}, \text{Mg}, \text{Ca}) \cdot (\text{Cl}, \text{SO}_4)$ by Rowe, Morey, and Zen (1972) points up the many potential ways in which sulphate can exist near cool planetary surfaces (e.g. of Mars). Numerous papers cover the complexities of S-bearing ore-forming fluids of which Holland (1965) and Naldrett (1969) are representative. Anhydrite is stable to high pressures and temperatures in systems of the appropriate composition (e.g. occurrence in S.W. African gneisses, Nash, 1972), and sulphate-melinite is stable to even higher pressures and temperatures when the anorthite component is available and the system is dry (Goldsmith, 1976, Fig. 9). Formation of sulphate component in scapolite was ascribed to oxidation of sulphide by concomitant reduction of Fe_2O_3 in iron oxide to FeO component in silicate. Sodalite and cancrinite minerals are stable to high temperatures: thus sulphate-nosean dissociates at 1528 K (Tomisaka and Eugster, 1968).

Chlorine and fluorine are further elements whose occurrence at planetary surfaces may be indicative of the internal composition. Anderson (1974) observed 0.01–0.22 wt% Cl in the aforementioned glassy inclusions, and concluded that as water boils away from an erupting lava it carries Cl with it. Hards (1976) showed experimentally that Cl tends to concentrate in an aqueous vapour during differentiation of acid and alkaline magmas whereas F tends to remain in the silicate melt and reduce the solidus and liquidus temperatures. Wellman (1970) described phase relations involving Cl in synthetic syenites, and Tomisaka and Eugster (1968) showed that chloride-sodalite dissociates at 1503 K. Concentrated brines can deposit halite and sylvine as part of the group of evaporite minerals, and when isolated these minerals melt at 1073 and 1043 K, respectively. However, evaporite minerals are very mobile and reactive in metamorphic terrains, and would react with silicates to form feldspathoids at lower temperatures. Small amounts of Cl occur in amphibole and mica. Fluorine is strongly stabilized in apatite, mica, and amphibole minerals (Ekström, 1972; Paul, Buckley, and Nixon, 1976; Prins, 1973; Stormer and Carmichael, 1971).

Space forbids similar surveys for other elements, but a few comments are perhaps necessary for large-ion lithophile elements, which concentrate in the late liquids differentiated by melting of basaltic and peridotitic bulk compositions. Potassium is

found in diopside from mantle-derived peridotites (probably from c. 150–200 km depth) up to 0.2 wt% (Erlank, 1973; Sobolev, 1972; Bishop *et al.*, 1978) and in pyroxene inclusions from diamond up to 0.9 wt% (Prinz *et al.*, 1975) but not so far in garnet (Bishop *et al.*, 1978). Whether substantial potassium is stored in the deep mantle, and if so how, is not known, but perovskite should be tested as a possible host. Phlogopite is a major host on Earth (Carswell, 1975; Smith and Dawson, 1975), probably as deep as c. 150 km and perhaps even to 300 km. In and near the crust, feldspar, mica and amphibole (Best, 1974) are major hosts, while some evaporite minerals store potassium close to the surface. Occurrence of K in metal cores of planets would provide an attractive heat source for convection, but the chemical relations are controversial (summarized in Ringwood, 1975, p. 553). Sodium occurs in pyroxene from mantle-derived xenoliths up to 8 wt% and in garnet up to 0.2 wt% (Sobolev and Lavrent'ev, 1971; Bishop *et al.*, 1978). In the crust feldspar is the dominant host but many other minerals carry some sodium. If Na and K are depleted in the Earth to the same degree, much of the Earth's Na must occur below 200 km depth (Smith, 1977). In meteorites Na-rich plagioclase is the dominant alkali species but several other minerals occur of which the large K-feldspar crystals in the Colomera iron and the Na-bearing phosphate minerals are particularly interesting. Isotopic data reviewed by Moorbath (1975) and Sun and Hanson (1975) show that Sr and Pb have been stored in the Earth's mantle all through geologic time with intermittent eruption of primary alkali basalt and nephelinite. These and other data show that on Earth the large-ion-lithophile elements did not move entirely into the surface and that mineral reservoirs exist perhaps down to at least 600 km. Whether the reservoirs are independent minerals, or whether they are merely microscopic inclusions trapped in stable minerals produced during primary differentiation is not known.

The simplest assumption for the terrestrial planets is that the temperature did not reach vapourization temperature and that the only loss of material was escape of light species from the atmosphere.

Ringwood (1966a), in a review of the genesis of chondritic meteorites, interpreted ordinary and enstatite chondrites as the possible products of volatile loss and autoreduction at high temperature of primitive material like that of CI meteorites. He emphasized the effects of selective volatility and of selective solubility in supercritical fluids of H_2O , CO_2 , H_2S , and H_2 . Numerous studies of lunar basalts and breccias demonstrated the mobility of

C and S; thus Gibson and Moore (1973) found that 12 to 30% of the C and S in lunar breccias could be volatilized under vacuum in 1 day at 1023 K and almost all at 1373 K. Smith and Steele (1976) reviewed the pertinent lunar mineralogy with respect to chemical processes involving C, P, and S. Particularly relevant to the chemistry of the early solar system are the studies of implantation of the solar wind into fine-grained regolith and the subsequent effect on metamorphism. In general these processes lead to reduction with removal of Fe from silicate, phosphate, and sulphide, and formation of iron metal, schreibersite, and probably cohenite.

Carrying further his preference for carbon-reduction in a condensed environment (e.g. a planetesimal) rather than hydrogen-reduction in a low-pressure nebula, Ringwood in a series of papers (see Ringwood, 1975) developed the single-stage chemical model for accretion and differentiation of the terrestrial planets, particularly the Earth and its Moon. High temperatures result from rapid accretion in which accretion energy overwhelms radiation loss at the surface, and from formation of an iron-rich core, which releases gravitational energy. The inner region of the solar nebula would condense to oxidized assemblages of minerals like those in CI carbonaceous meteorites (including magnetite, hydroxylated silicates, and carbon-rich compounds) while the outer region would be dominated ultimately by ices of various compositions including methane and ammonia. Ringwood developed the terrestrial planets by rapid accretion ($< 10^6$ years) before the T Tauri wind began. Such rapid accretion produced temperatures exceeding 1300 K during most of the growth of the Earth, and from 1800 to over 2300 K at the end. A cool, oxidized, volatile-rich nucleus of olivine, magnetite, hydrated Fe,Mg,Ni-silicates, and (Fe,Ni)S, with surface temperature near 1000 K, became reduced by carbonaceous material and degassed as further accretion to about one-fifth of final mass brought the surface temperature up to 1500 K. A primitive reducing atmosphere developed from the reaction $\text{CH}_4 + \text{H}_2\text{O} \rightarrow \text{CO} + 3\text{H}_2$. Further accretion with rise of surface temperature to 1800 K caused reduction and volatilization of some relatively volatile elements including alkali metals, Zn, Hg, Pb, S, and Cl. Accreting material reacted with the atmosphere to give metallic iron and Fe-free silicates. Further temperature rise to over 1800 K caused volatilization of SiO and accretion of forsterite. At even higher temperature Mg vapour would be formed and metallic Si would enter iron metal. Calcium and aluminium silicates would accrete. Gravitational instability with sinking of metal to form a core would yield 2.5 J/kg, and the

entire outer part of the Earth would volatilize into a disc. Condensation into a sediment ring that accreted into a Moon, and removal of the volatiles by an intense T Tauri wind complete the single-stage model. This impressive model has many attractive features but has been criticized. How can enormous amounts of CO be lost from the Earth? How can substantial amounts of SiO be volatilized from the Earth? Some modifications to the single-stage model are described in Ringwood (1975, pp. 591–7). I prefer to consider models that do not require escape of substantial amounts of heavy molecules from planets with escape velocities greater than 5 km/s.

Chemical differentiation in a planetary atmosphere and consequent effect on the solid part. Escape of an atom, molecule, or ion can occur from a planetary atmosphere when the escape velocity is exceeded. Simple kinetic theory of gases leads to a classical distribution of velocities whose mean value for an ideal gas is $\bar{V} = (8kT/\pi m)^{1/2}$, where k is Boltzmann's constant, T is temperature, and m is mass. Escape occurs from the high-energy tail of the distribution. Unfortunately, the loss from planetary atmospheres is complicated by the variation of temperature and chemical composition with height, the extent of ionization from solar radiation (principally ultraviolet), and the effect of the solar wind. At one extreme the T Tauri wind has been used to sweep away the entire atmosphere, irrespective of mass (Ringwood, 1975), while at the other extreme, the upper atmosphere might act as a cold trap inhibiting or preventing the loss of all but the lightest species. Volatiles can be ejected to high altitude by volcanic emission, and removed by adsorption in soil or by direct freezing. It is difficult to place fundamental constraints on the escape processes, and readers are referred to Hunten (1973), Hunten and Donahue (1976), Curtis and Hartle (1977), Visconti (1977), and articles listed in Mars and Venus sections.

It is convenient here to mention Turekian's latest model for the Earth, which depends greatly on loss of volatile species. A brief summary (Turekian, 1976) mentions the following features: hot condensates accreted sequentially and rapidly on to the growing Earth, followed by a rubble veneer, which became progressively enriched in low-temperature products; the rubble veneer did not equilibrate with the iron-nickel core; most of the iron from the nebula was sequestered in large bodies during condensation and did not react with the remaining gas; low-temperature condensates in the rubble veneer lost water to the atmosphere during the first 15×10^6 yr, and most of the protons in the water were lost from Earth, leaving oxygen which combined with metallic iron to give the Fe^{2+} and Fe^{3+}

in the present crust and upper mantle; after dissipation of the early atmosphere the atmosphere would be primarily composed of H_2O and CO_2 with the minimum oxygen pressure determined by the oxidation state of the crust and upper mantle; oxidation of Fe metal to FeO resulted in conver-

sion of some pyroxene ($c.\text{MgSiO}_3$) to olivine $(\text{Mg,Fe})_2\text{SiO}_4$.

Instead of following Turekian's model, the alternative of losing most of the H from planetesimals should be considered.

EARTH

Descend into that crater of Sneffels Yocul . . . audacious traveller, and you will reach the centre of the earth. I have done it.

Jules Verne, *Journey to the Centre of the Earth*

We can only envy Arne Suknusse in our puny attempts to explore the interior of the Earth!

This section is restricted to mineralogical, petrological, and geochemical properties particularly relevant to the general mineralogy of the planets. Speculations on the earliest history of the Earth range from complete melting and partial volatilization, even of Si, to slow accretion with segregation of a core and proto-crust by localized partial melting and diapirism. The latter approach permits many variants including chemically homogeneous vs. inhomogeneous accretion, and rapid vs. slow differentiation into a core, mantle, and crust. Emphasis is placed here on the relics of the oldest crust, and on the upper-mantle xenoliths carried up by alkali basalts and kimberlites. General reviews of the properties of and processes inside the Earth were given by Stacey (1977), Ringwood (1975), and Wyllie (1971).

Early history

The accretion age of the Earth is hard to define, and all estimates depend on geochemical models. An estimate of 67 ± 20 million years for the formation interval of the Earth (Pepin and Phinney, 1976) is consistent with interpretation of initial Pb-isotope data for the Amitsoq gneiss, W. Greenland (Gancarz and Wasserburg, 1977).

Advances in the interpretation of structural deformation in polymetamorphic rocks are leading to decipherment of early Archaean terrains, as summarized by contributions to a NATO Advanced Study Institute (Windley, 1975). Comparison with data for Proterozoic terrains (-2.5 to -0.6×10^9 yr) and later assemblages allowed Windley (1977) to summarize the evolution of the continents. As far as possible references are made to articles by listed authors in Windley (1975).

That the oldest known crystallization age for a terrestrial rock is about -3.8×10^9 yr (Moorbath) is not surprising in view of the lunar evidence for bombardment by major projectiles (20–100 km

diameter?) as late as -3.9×10^9 yr. Coupled with the trebled production of radioactive heat in the pre-Archaean era, the disruptive effect of the projectiles should have produced major volcanic activity in the outer part of the Earth. There is no need to invoke substantial accretion of material after -3.8×10^9 yr, though minor amounts have arrived as meteorites, comets, and dust.

Petrologically the Archaean rocks can be separated into two metamorphic suites. A high-grade one is dominated by tonalitic-granodioritic gneisses of calc-alkaline affinity, intercalated with layered igneous complexes and metamorphosed volcanic and sedimentary rocks of which marbles, graphite-bearing mica schists, and banded magnetite-quartzites may be particularly significant. A low-grade suite of greenstone belts was originally composed of volcanic rocks of ultramafic, mafic, and calc-alkaline affinity, and sedimentary rocks including cherts and banded iron formations.

Attempts to compare the chemistry of these rocks with modern ones are complicated by the probability of extensive metasomatism (Gunn), as suggested by the depletion of mobile elements (e.g. K, Rb) in granulites compared to amphibolites, and by the certainty that higher heat flow would result in a different tectonic style (Burke, Lambert) and in different processes of partial melting in Archaean than in modern eras. Both Archaean suites contain relics of ultramafic volcanic rocks, including komatiites characterized by spinifex-textured Mg,Cr-rich olivine. Such olivine-rich rocks are relatively less abundant in later eras, which supports ideas that the higher heat production in the Archaean era allowed a greater degree of fractional melting (Green, 1975).

Since blue-schists are virtually absent in Precambrian rocks, while green-schists are abundant back to the Archaean, plate tectonic processes must have changed considerably with time, if indeed such processes can be applied to the Archaean era. Bearing this in mind, Windley and Smith noted the resemblance between the rocks in the Archaean

high-grade terrains and those in active continental margins along Cordilleran fold belts, while Tarney *et al.* used the 'Rocas Verdes' marginal basin as a model for greenstone belts. Windley (1977, pp. 60–3) incorporated these features into a plate-tectonic model for the Archaean suites. Even more speculative are the models for a pre-Archaean 'scum' dominated by basic rocks (Lambert, Murthy, Shaw), which turned into proto-continents via a 'permobile' tectonic regime (Burke). Perhaps Ca-rich amphibole and Na-rich plagioclase dominated the proto-crust in contrast to Ca-poor pyroxene and Na-poor plagioclase in the lunar crust.

Interpretation of Rb–Sr and U–Pb isotopic data depends considerably on models in which the rate of, and extent of, separation into separate reservoirs by partial melting and metasomatism provide difficult problems. Preliminary investigations (Moorbath) imply that Archaean and Proterozoic gneiss complexes, greenstone belts, and 'granitic' plutons are mostly new additions to the continental crust by episodic partial melting of the mantle, while regeneration of the crust, except for rocks rich in K-feldspar, is a minor process. This implies that feldspar-rich rocks are irreversibly differentiated from the mantle with little returned by subduction. Models for the origin of the Earth–Moon system must allow for storage of really substantial amounts of large-ion-lithophile elements in the Earth's mantle for at least 10^9 yr after accretion. These ideas are compatible with the conclusion by A. T. Anderson (1974) that there is a gradual and permanent transfer of Cl, S, and H_2O from the interior of the Earth to surface reservoirs on a time scale of the Earth's life.

Clues to the development of the atmosphere and hydrosphere have been inferred from reconnaissance studies of the sediments. From the relation between sedimentary and volcanic rocks in Archaean terranes, Sutton suggested deposition of sediments on stable shelves. The stratigraphic succession for the Barberton greenstone belt has ultramafic–mafic volcanic rocks overlain by calc-alkaline volcanic rocks, and finally by sediments ranging from a lower argillaceous group of shales, sandstones, and greywackes to an upper arenaceous group of conglomerates, quartzites, limestones, and banded iron formations. These sediments can be assigned to a range of deep- to shallow-water environments (see also Lowe and Knauth, 1977).

The extreme depletion of noble gases on Earth rules out direct capture of gases from the solar nebula, and all recent models assume that the present atmosphere was derived by out-gassing of the interior, modified by an unknown loss of light gases through the atmosphere and ionosphere.

Present volcanic gases are composed of roughly equal amounts of H_2O , CO_2 , and SO_2 , possibly in original equilibrium with magnetite-bearing rocks, and certainly diluted during ascent by erratic amounts of contaminants (Nordlie, 1977).

Current theories (Schidlowski, Walker) assume that early volcanic gases were dominated by H_2O and CO_2 , and indeed the sedimentary and volcanic rocks dating back to -3.8×10^9 yr have been interpreted in terms of an early H_2O – CO_2 -rich atmosphere, essentially free of O_2 . Loss of H through the ionosphere was coupled with capture of the released O_2 by Fe^{2+} , C, and S. Free oxygen developed in the atmosphere by fermentation produced by organisms, which became abundant in the Proterozoic era. However, it is not clear what was the temporal variation of oxygen in the atmosphere since -3.8×10^9 yr, and what was the nature of the atmosphere before -3.8×10^9 yr, especially during the accretion of the Earth. Whether methane and ammonia were present in the early atmosphere is of considerable biological significance because of the speculations on the origin of amino acids and other organic molecules.

Some quartzites contain abundant magnetite in bands, and these banded iron formations, dating back to -3.8×10^9 yr, typically occur with volcanic rocks, perhaps consistent with derivation of the iron and silica during fumarolitic activity, as suggested by Goodwin (Windley, 1977, p. 42). Alternatively, the iron and silica may have derived from weathering of continental rocks in a reducing atmosphere such that iron was washed away as Fe^{2+} rather than staying as Fe^{3+} as in modern sediments. Schidlowski discussed Cloud's suggestion that autotrophic organisms were protected from ultraviolet radiation by an aqueous shield, and released O_2 , which combined with hydrated Fe^{2+} ions to ultimately yield magnetite in sediments. Arguing against the weathering model is the absence of Al_2O_3 -rich residues in Archaean continental relics, which would be expected after leaching of Fe^{2+} (Windley, 1977). Perhaps a combination of fumarolitic activity with biological fixation provides a viable model. Whatever the source of the banded iron formations they could act as a sink for oxygen produced by the exospheric loss of H from atmospheric water. Banded iron formations became sporadic after -2×10^9 yr, and continental 'red beds', closely associated with evaporites, became abundant thereafter. Schidlowski suggested that these hematite-bearing rocks began to form when oxygen was released to the atmosphere resulting in conversion of weathered Fe^{2+} into immobile Fe^{3+} . Perhaps a slow-down of fumarolitic activity might be involved. Ultimately the concentration of oxygen in the atmosphere reached

the 'Pasteur' level of one-hundredth the present level, at which stage breathing organisms developed. Fryer (1977) observed anomalously large concentrations of Eu in chemical sediments earlier than -1.9×10^9 yr and anomalously high Ce for those later than -2.3×10^9 yr. This was interpreted as the result of increasing oxidation such that Eu^{2+} and Ce^{3+} were stable during formation of the earlier sediments and Eu^{3+} and Ce^{4+} for the later sediments.

Graphite is widespread in metasedimentary Archaean rocks including mica schists, limestones, and banded iron formations such as the -3.8×10^9 yr Isua occurrence (Allaert). Carbon isotopic ratios indicate that a sedimentary reservoir of organic carbon had built up by -3.7×10^9 yr, indicating early existence of photosynthesis (Schidlowski). Morphological evidence of early biological activity is controversial (cf. Schopf vs. Muir and Grant). The ratio of C/CO₂ in sedimentary rocks has been roughly constant over the past 3×10^9 yr. Holland suggested that decreasing heat production slowed down geochemical cycles, and in particular lowered the injection of CO₂ and reducing gases into the atmosphere. Carbonate rocks date back to -3.8×10^9 yr, with a temporal trend from dolomite-rich varieties in the Archaean to calcite-rich ones in modern sediments (Holland).

Opposing the concept of an early oxygen-poor atmosphere, Kimberley and Dimroth interpreted uraninite and pyrite in Archaean sediments as products of diagenesis rather than as indicators of a low partial pressure of O₂ in the atmosphere ($< 4 \times 10^{-3}$ bar; Holland). They ruled out radiative synthesis of hydrocarbons in the atmosphere because of the absence of bitumen in sediments, and noted that carbonaceous mudrocks of all ages are rich in pyrite. Finally they explained the absence of Archaean red beds and sulphate deposits by lack of preservation of appropriate shelf facies. Simpson and Bowles (1977) also suggested that the early atmosphere was non-reducing because of evidence for SO_4^{2-} and UO_2^{2+} ions in solution; furthermore they argued that uraninite grains of the Witwatersrand and Dominion Reef Systems were detrital, and had survived abrasion because of hardening from Th substitution.

Obviously there is great scope for study of all Archaean rocks, perhaps especially of sediments. For comparative planetology it seems safe to conclude that by -3.7×10^9 yr the Earth had a non-reducing atmosphere, perhaps even containing some molecular oxygen, and that proto-continent shelves were becoming sufficiently stable to survive later tectonism (cf. Hargraves, 1976). Furthermore, the upper mantle was spewing out volcanic rocks mostly similar in chemistry to those

of recent times, except perhaps for absence of kimberlitic and carbonatitic types thought to be characteristic of large stable cratons, and for presence of the ultrabasic types indicative of a greater degree of partial melting (e.g. Villaume and Rose, 1977). In detail, there is evidence of mantle heterogeneity and chemical evolution with time (Sun and Nesbitt, 1977), but for this general review it is probably better to point to resemblances rather than differences between Archaean and modern crustal rocks. Speculations that the Earth had a massive primitive atmosphere (e.g. 300 bars H₂O and 50 bars CO₂; Walker) are not ruled out by direct evidence from Archaean rocks. However, there is no evidence in support of proposals for a massive reducing atmosphere, dominated by methane, from which organic molecules could be generated. It is necessary to turn to evidence in the upper mantle in order to proceed further with a mineralogical study of the Earth.

Upper mantle (above 200 km depth)

Kimberlite magmas have formed in stable continental regions at depths up to c.150–200 km, and during explosive emplacement have torn off and transported fragments of the wall rocks—hence the name 'Poor Man's Mohole' for a kimberlite pipe. Basalt magmas provide information on the site of partial melting, and alkali basalts have transported pieces of the crust and mantle down to c.60 km. The details are very complex (e.g. Ringwood, 1975; Meyer, 1977; Sobolev, 1977), but an over-all picture has emerged on the basis of the following eight concepts: the dominant rock types of the upper mantle are Ca,Al-rich peridotites, Ca,Al-poor harzburgites, and subordinate eclogites and dunites. The peridotites are dominated by olivine and orthopyroxene and are subdivided into spinel- and garnet-varieties, which can be attributed respectively to approximate depth ranges of c.25–c.60 km, and $> c.60$ km when the oceanic and continental-shield geotherms are compared with a generalized phase diagram for peridotites (fig. 2). Partial melting of Ca,Al-rich peridotites would result in extraction of various basaltic magmas as the Ca,Al-rich minerals were dissolved, leaving behind refractory residues dominated by olivine and orthopyroxene, thereby explaining most harzburgites. Although considerable amounts of Ca,Al-rich material have entered the crust as basaltic regions near the surface and metamorphosed amphibolitic to granulitic regions at depth, large amounts remain in the upper mantle. There is a very large range of compositions for the pyroxenes and garnets in eclogites and garnet peridotites, and diverse origins are needed including metamorphism of arrested partial melts

and of subducted rocks. There may be a tendency for density differentiation with heavier fertile peridotites underlying lighter barren peridotites (Boyd and McCallister, 1976). [Note: 'fertile' refers to the chemical capacity to yield basalt upon partial melting, as proposed by M. J. O'Hara.] There is strong evidence for substantial heterogeneity of the source regions of basalts, though the details are based on controversial interpretations of geochemical data (e.g. Hofmann and Hart, 1978; Kay and Hubbard, 1978; Tatsumoto, 1978). Finally, there is no need to assume major changes in the chemistry of the source regions of basalts since Archaean times (O'Nions and Pankhurst, 1978).

The details of upper-mantle mineralogy are extremely complex, as is apparent from the Proceedings of the First and Second International Conferences on Kimberlites (*Phys. Chem. Earth*, 9; to be published), and many papers scattered among journals. Although most olivines and orthopyroxenes from upper-mantle rocks have Mg/Fe ≈ 10 , the contents of trace and minor elements are highly variable; e.g. harzburgites can be divided into 'barren' and 'fertile' varieties on the basis of Ca and Al substitutions in the orthopyroxene (Hervig *et al.*, 1977). Major and minor elements in garnets and pyroxenes vary considerably in the upper mantle, and cluster analysis is yielding classification schemes (Dawson and Stephens, 1975; Stephens and Dawson, 1977) that should lead to petrogenetic interpretations. Controversy remains over the experimental and theoretical calibration and the practical application of the pyroxene and garnet barometer-thermometers for garnet peridotites (many papers), but the general concepts and the application to estimation of a shield geotherm (Boyd, 1973) mark a high point in upper-mantle petrology. In future years trace and minor elements should become increasingly used in thermo-barometry, and two possible candidates are the Na distribution between garnet and pyroxene (Sobolev, 1977; Bishop *et al.*, 1978), and the Cr content of silicate and oxide minerals (Malinowsky and Doroshev, 1977).

The roles of volatile elements are particularly intriguing. Depression of the melting temperature of peridotites by H_2O and CO_2 has been explored in detail (see earlier). Under terrestrial shields, it appears that the maximum depth for stability of amphibole is ≈ 100 km and for phlogopite is ≈ 200 km, though this latter value might possibly be too low for an unusually cool region. These minerals should be major reservoirs for H_2O (actually as OH) in the upper mantle. Whether substantial H_2O can be stored at greater depths is totally speculative, but synthesis of hydroxylated Mg-silicates (Akimoto *et al.*, 1977) is indicative. Of course, it is

easy to postulate hypothetical hosts, such as inclusions trapped in refractory minerals during accretion, and indeed the emission of primordial noble gases from the mantle testifies to deep-seated reservoirs that have persisted for 4.5×10^9 yr (Craig and Lupton, 1976; Dymond and Hogan, 1978). However, it seems easiest to assume that an early differentiation flushed out most of the water from the mantle below ≈ 200 km. A similar conclusion is probably true for CO_2 , but again there is experimental evidence indicative of potential incorporation of CO_2 into crystalline material at lower-mantle pressures (Jeanloz and Ahrens, 1977b).

Absolutely crucial to the controversies over the accretion of the Earth are the contents of chalcophile and siderophile elements in the upper mantle (Smith, 1977), which indicate core-mantle disequilibrium, as emphasized by Ringwood (1966b). The Fe/Ni ratio of upper-mantle olivines is close to the cosmic ratio, thereby indicating that Ni was not stripped away significantly from the upper mantle by reaction with either metallic iron or sulphide at low pressure. Silicate-sulphide partitioning (Binns and Grove, 1976; MacLean and Shimazaki, 1976; Rajamani and Naldrett, 1978) depends on complex factors, but the partition coefficients for Ni favour sulphide so strongly that it must be concluded that sulphide is only a minor component of the upper mantle, even though it is a ubiquitous trace constituent of peridotite and basalt. The noble metals in peridotite xenoliths (Smith, 1977) are strongly depleted over cosmic composition, but the depletion factors are not as high as would be expected for equilibration of peridotite with metal or sulphide (Kimura *et al.*, 1974; Rajamani *et al.*, 1977). These data are strong evidence for late accretion of material either undepleted or only slightly depleted in chalcophile and siderophile elements, and it may be possible to go even further and argue that the Earth then had a stable mantle barrier to prevent equilibration with a metal-rich core.

Diamonds have a triple significance: pressure indication (fig. 9), redox indication (Brey and Green, 1977; Rosenhauer *et al.*, 1977), and preservation of inclusions. Existing data indicate growth in both peridotitic and eclogitic hosts, and the possibility of survival of diamond during partial-melting events should be considered. Of great potential importance are the observations of high values of K_2O in some pyroxene inclusions (Sobolev, 1977) and of the very high content of Cr in spinel inclusions (Meyer, 1977). Although diamonds have been found *in situ* in eclogite and peridotite xenoliths, the silicate and spinel inclusions of random diamond grains mostly do not match in detail the chemistry of silicate and spinel minerals of the xenoliths, and it may turn out that

the inclusions 'remember' an earlier igneous event whereas the independent minerals are mostly controlled by metamorphism, perhaps following an earlier igneous event.

The trace-element composition of the upper mantle is usually inferred from partial-melt models applied to observed chemistry of basalts. Since partial-melt models involve complex assumptions about the melting process and the extent of crystal-liquid differentiation during magma ascent, they are not pursued here. For general reviews readers are referred to Wyllie (1971), Carmichael *et al.* (1974), and Yoder (1976), and for more recent papers to references listed by Sun and Nesbitt (1977) and those in the Memorial Volume to P. W. Gast (*Earth Planet. Sci. Lett.* **38**, no. 1). Ideally, this approach should be tested by analysis of peridotite and eclogite xenoliths. Unfortunately, metasomatism during transport to the surface causes systematic chemical bias of bulk samples, and it is desirable to proceed to chemical analysis of unaltered regions. Most existing bulk analyses of mineral separates are affected by mechanical impurities, as demonstrated by high concentrations of incompatible elements in olivine, but the data of Shimizu (1975) for K, Rb, Cs, Sr, and Ba in clinopyroxenes from ultramafic xenoliths appear especially good, probably because of careful hand-picking of clean grains. Thorough attention to the background is allowing detection levels of c.20 ppm for many elements measured with an electron microprobe, and Smith *et al.* (1978) were able to compare K/Rb/Ba ratios for micas of several textural types in peridotite xenoliths and kimberlite host with ratios for various types of basalts. Micas with primary texture in peridotites have K/Rb close to that for CI meteorite (c.300), while K/Ba is about six times lower. Many basalts could have obtained their K, Rb, and Ba contents by total melting of mica from peridotite, whereas those with high K/Rb (c.10³) could not. The next decade should see a plethora of data on trace elements in mantle-derived minerals using ion microprobes, as indicated by measurements of transition elements (Shimizu and Allègre, 1977), thereby placing limits on models for partial melting and metamorphism when matched with experimental data for element partitioning (see especially the recent *Year Books* of the Carnegie Institution of Washington).

Because the upper mantle is heterogeneous as a result of partial melting and mechanical transport, it is not surprising that estimates of the bulk composition vary considerably. Analyses of garnet peridotites range several-fold in Ca and Al, probably as a result of partial melting, and attempts have been made to select compositions supposedly

undepleted or unaugmented in basaltic component. Other attempts involve addition of enough basaltic component to peridotite to give CI-meteorite ratios of rare earths. Some analyses of garnet peridotites are listed in Smith (1977, Table 4). Ringwood (1975) obtained a model 'pyrolite' by adding one part of 'basalt' to three parts of 'peridotite', and this is a useful but not definitive composition for speculation about the mineralogy of the Earth's mantle.

Mantle (deeper than 200 km) and core

All mineralogical speculation for these regions depends on interpretation of seismic evidence in terms of a density-depth profile (fig. 1) and velocity-depth profiles (not shown), cosmochemically based speculation on the bulk composition, and experimental measurement of the physical properties of likely minerals. A major uncertainty is the temperature profile (McKenzie and Weiss, 1975), which is constrained only weakly by the liquid-solid transition in the core. Thomsen (1977) questioned the theoretical foundations of equations of state for terrestrial planets, and the present review will emphasize the uncertainty of present interpretations.

The properties of the core (reviewed by Brett, 1976, and Jacobs, 1975) are important for speculation on the bulk composition of the Earth (next section). Although the solid inner core must be rich in iron and siderophile elements (particularly Ni), the chemical details are unknown. The liquid outer core is less dense than an Fe,Ni liquid (fig. 1), but the light constituents cannot be uniquely defined. Troilite or pyrrhotine occur in most meteorites, and partial melting of chondritic compositions would give heavy Fe,FeS-rich liquids which should sink in a growing planet. Thus the suggestion that S is the major light element in the Earth's core (Murthy and Hall, 1972; Murthy, 1976) is very popular, and Ahrens (1978) concluded that 9-12 wt% S is sufficient to explain the lower density. Ringwood originally argued for Si, by analogy with Si-bearing metal in highly reduced enstatite meteorites, but has now (1977) switched to oxygen because of two arguments: first, placing c.10% S in the core would lead to greater depletion of K (six-fold) than S (three-fold) with respect to solar composition, which is contrary to the Grossman-Larimer type of model for simple progressive condensation of the solar nebula, and second, thermochemical calculations indicate complete miscibility between Fe and FeO above 3800 K at high pressure. Smith (1977) argued that S might be protected from mechanical disaggregation in Fe-metal cores of differentiated planetesimals while K was lost

preferentially from the bombarded surface. Currently, it is desirable to consider the two extremes of S-rich and O-rich outer cores. Simple analogy with iron meteorites suggests that C, N, and P may also substitute in the Earth's core. Smith (1978) tentatively suggested that U and Th may partition into troilite under reducing conditions, thereby providing a possible heat source for the Earth's dynamo, but the available evidence from troilite nodules in meteorites is poor, and probably unfavourable. Whether substantial K enters a S-rich core is highly controversial (Goettel, 1974; Ringwood, 1975) and the latest experimental partition coefficients between silicate and sulphide (Ganguly and Kennedy, 1977) apparently place an upper limit of 100 ppm K in the Earth's core.

The mineralogy of the lower and middle mantle is quite speculative. Several transitions are indicated by seismic studies, as exemplified by a density-depth profile (fig. 1). Refinement of seismic data may lead to minor revision of the profiles, especially if subducted slabs actually penetrate to great depth (c. 700 km). To a first approximation the mantle can be regarded as an assemblage of close-packed oxygen ions with interstices occupied by cations, and the physical properties can be modelled by equations which depend principally on the mean atomic number $\bar{Z}$ and not on the detailed crystal structure. This approximation is inadequate, and must be replaced by equations that utilize data measured for specific crystal structures. The early structural data obtained by Ringwood and collaborators (collected in Ringwood, 1975) are being extended and essentially confirmed by new X-ray data obtained in diamond-anvil cells. All the silicates found in peridotite (viz. olivine, pyroxenes, and garnet) transform into denser structures as the depth increases. Particularly important is the perovskite structure, which is probably the major carrier of $(\text{Mg,Fe})\text{SiO}_3$ in the lower mantle, and quite possibly of other components including Ca- and Al-bearing chemical units (Liu, 1974, 1975, 1976; Liu and Ringwood, 1975; Liebermann *et al.*, 1977; Sawamoto, 1977; Ito and Matsui, 1978; Yagi *et al.*, 1977, 1978). Other important structure types are garnet and $\beta\text{-Mg}_2\text{SiO}_4$ (upper mantle), periclase and ilmenite (middle mantle), and perhaps a structure with 8-coordinated Si (similar to fluorite?) in the very deep mantle (Jeanloz and Ahrens, 1977a).

There are four major difficulties in disentangling the various chemical factors. First, the Mg/Si ratio may vary with depth. Ringwood favoured volatilization of SiO from a very hot earth resulting in a mantle dominated by Mg_2SiO_4 , whereas D. L. Anderson favoured a mantle dominated by MgSiO_3 . Jeanloz and Ahrens (1977a) suggested

that olivine and pyroxene compositions have similar physical properties at high pressure, whereas Sawamoto (1977) interpreted the lower mantle in terms of pyroxene composition and perovskite structure. Second, there is uncertainty in the depth variation of Fe/Mg, ranging from a model with substantial increase of FeO with depth (e.g. reaching 21 wt % in the deep mantle; Anderson, 1977) to the possibility of a smaller change of FeO (e.g. Davies, 1974), and to the interpretation by Sawamoto (1977) that the ratio actually increases with depth from 0.05–0.18 in olivine composition at 200–350 km depth to 0.11–0.25 in olivine composition at 400–600 km and then decreases to 0–0.18 in pyroxene composition at 700–1500 km depth. Third, there is essentially complete uncertainty of the spatial distribution of Ca, Al, Ti, and all other elements of lower concentration, and fourth, disproportionation of 3FeO into Fe and Fe_2O_3 may occur above 100 kb (Mao and Bell, 1977a). Undoubtedly these uncertainties will be reduced by detailed crystallographic studies with diamond-anvil devices (now operating up to 1.5 Mb, Mao and Bell, 1977b), phase-equilibrium studies to 200 kb or higher (e.g. Akimoto and Akaogi, 1977), and crystalchemical models (e.g. Hazen and Prewitt, 1977; Burns, 1974) based on detailed crystal structure analysis (e.g. spinel at high pressure, Finger *et al.*, 1977). To illustrate technical improvement, note how revision of the cell dimensions of quenched MgSiO_3 -perovskite changed the density from 4.00 g/cm^3 (Sawamoto, 1977) to 4.10 (Ito and Matsui, 1978; Yagi *et al.*, 1978), thereby allowing a lower Fe content for any part of the mantle containing perovskite.

In the meantime Anderson (1977) suggested transformation of pyroxene to garnet at 250–370 km depth, olivine to β -phase(?) at c. 400 km, and pyroxene to ilmenite(?) near 500 km, whereas Liu (1977) suggested that the 400 km discontinuity marks the transformation of olivine to β -phase and gradual absorption of pyroxene into garnet, the 650 km discontinuity involves formation of perovskite, ilmenite, and periclase structures, and the lower mantle has perovskite replacing ilmenite.

Bulk composition

Speculation on the bulk chemical composition, especially its radial distribution, is important for constraining models for the chemical composition and mineralogy of the inner planets, and especially of the Earth-Moon system. As emphasized in preceding sections, it is impossible to model the Earth merely from direct observations, and it is necessary to use cosmochemical information. Several attempts by geochemists to model the

Earth in the 1960s were followed by Larimer (1971) who compared twenty-two elements plus H_2O in the Earth's crust with those in ordinary chondrites and eucrites, by Shaw (1972, 1976) who developed a detailed model for chemical differentiation, and by Clark *et al.* (1972) and Hutchison (1974, 1976), whose models invoke two different kinds of heterogeneous accretion. This section is a heavily truncated version of Smith (1977), with some new ideas.

Rather than model the Earth with mixtures of meteorites (Anderson *et al.*, 1971; Murthy and Hall, 1972), Ganapathy and Anders (1974) assumed that the inner planets were made by 'exactly the same processes as the ordinary chondrites'. Equilibrium progressive cooling of the solar nebula at 10^{-4} bar with C/O 0.6 yielded the theoretical condensates shown in fig. 4. Ganapathy and Anders split the elements up into five groups belonging to early condensate, silicate, metal, volatiles condensing at 1300–600 K [lithophile and siderophile sub-groups], and volatiles condensing at < 600 K [two sub-groups]. They assumed that each group was unfractionated after condensation except for the melting event that formed chondrules. Relative fractionations between groups were calculated by assuming CI-meteorite composition for the element ratios in the nebula and the following Earth data: 0.018 ppm U to model the early condensate; 36 wt % Fe in the bulk Earth to give the metal/silicate ratio; observed Fe/Mn ratio to give oxidation state of iron; assumed S content to give FeS; K/U ratio of 9440 to model loss of volatiles (1300–600 K) during assumed chondrule formation; and Ti/U ratio of 0.27 to give volatiles (< 600 K). Brevity forbids further details, and the final composition is given in Table IV. Ignoring details the resulting model has the Earth enriched in 'early condensate' and 'metal' about 1.5-fold with respect to Mg-silicate, and depleted about four-fold in 'volatiles 1300–600 K' and about fifty-fold for 'volatiles < 600 K'. Special treatment for H, C, and N resulted in depletions of 5×10^{-7} (H), 5×10^{-4} (C), and 3×10^{-5} (N).

Smith (1977) made further estimates of the bulk Earth trying to utilize observed compositions as far as possible. There was no problem for the atmosphere and hydrosphere. Important elements in the crust were modelled after Larimer (1971), and others were taken from compilations by Ronov and Yaroshevskiy and by Taylor. Probably the major uncertainty is in the lower crust, where high-grade metamorphism may have caused substantial redistribution of mobile elements; however, there is substantial agreement between a new estimate by Ronov and Yaroshevskiy (1976) for the whole crust and the estimates by Leyreloup *et al.* (1977) for the lower continental crust under a French volcanic

TABLE IV. *Suggested chemical compositions of the planets*

	wt	a	b	c	d	e	f	g
H	ppm	78	66	66	0.8	—	—	—
C	ppm	350	220	220	11	—	—	—
N	ppm	9.1	19	19	0.1	—	—	—
O	%	28.5	31.1	31.3	42.0	35.3	15.5	41.6
Na	ppm	1580	850	850	710	300	—	—
Mg	%	13.21	14.65	13.7	17.8	15.0	5.34	20.5
Al	%	1.77	1.30	1.83	3.92	1.2	3.17	1.70
Si	%	14.34	16.09	15.1	19.2	16.4	6.26	19.2
P	ppm	2150	1700	1830	690	(1000)	3700	—
S	%	1.84	4.34	2.91	570	?	—	—
Cl	ppm	25	45	45	0.47	—	—	—
K	ppm	170	130	130	76	30	—	830
Ca	%	1.93	1.62	2.28	4.28	1.3	3.93	2.00
Ti	ppm	1030	660	928	2300	700	1600	—
Cr	ppm	4780	3700	4160	1700	(3800)	8000	—
Mn	ppm	590	470	470	1140	100	—	—
Fe	%	35.87	28.19	31.7	11.5	29.2	61.1	13.0
Ni	%	2.04	1.53	1.72	0.65	1.7	3.4	—
Zn	ppm	93	220	147	3.1	—	—	—
U	ppb	18	14	20	40	(12)	34	—

a. Earth. Ganapathy and Anders (1974). b. Earth. S1 model (Smith, 1977) based on total condensation of metal, silicate and high-temperature fractions, and various depletions of volatile elements. c. Earth. New S2 model in which the S1 model is revised to have 10% S in core, 'metal' elements augmented by 1.2, 'early condensate' elements by 1.5 over (Mg + Si). d. Moon. Morgan *et al.* (1978b); model 4b based on lowered U content and assumed core mass of 2% (mostly Fe, Ni, S). See listed references for other estimates, including ones with lower Ca, Al, and U. e. Eucrite parent body. Morgan *et al.* (1978a); bracketed elements estimated from expected cosmic ratios; 11.2 wt % Fe is in the core. See listed references for other estimates. f. Mercury. See text for simple model based on oxidized nebula. Minor S, C, and N would be expected for condensation from reduced nebula. g. Mars. Fe-rich mantle proposed by Smyth and McGetchin (1978); note that this model would have less Fe if the moment-of-inertia is reduced. Minor and trace elements not considered.

Note. All the above compositions are based on the theory of equilibrium condensation with 'stepped' depletion factors for an oxidized nebula with C/O = 0.6. Somewhat different compositions would result for a reduced nebula, especially for C, N, S, Cr and elements which become chalcophilic under reduced conditions.

pipe and by Sibley and Wilband (1977) for average sediment. For the upper mantle Smith used the mean of twenty garnet peridotites calculated by Carswell (which is very close to a new estimate by Maaløe and Aoki, 1977), together with a rag-bag of trace-element data culled from many papers. The Carswell composition was also used for the remainder of the mantle, except that those elements that should enter either basaltic liquid or FeS-rich liquid during partial melting of peridotite were assumed arbitrarily to occur only in 10% of the mantle. Ringwood's (1975) model for a lower mantle that had not lost basaltic component would

require a higher content of Ca, Al, etc. than the Smith model; furthermore the Anderson model would require substantial increase of Fe with depth. Sulphur was assumed to be the major light element of the core, but the original concentration of 14 wt% should now be reduced to c.10 wt% (Ahrens, 1978), and even lower if other light elements are present (Si, C, O?). Minor element concentrations of meteoritic troilite vary considerably with redox state, and Smith assumed that substantial Cr (1 wt%) and other chalcophile elements are present in the FeS component of the Earth. No single type of iron meteorite was suitable for modelling the core, and Type III may be slightly better than Types I and II. Large variations of trace refractory metals occur in Types II and III irons, probably because of crystal-liquid fractionation, and log-median values were taken.

Ignoring details, the important features are shown in fig. 10. The ordinate is the ratio of concentrations in Earth to $1.5 \times C_1$ meteorite, with the factor 1.5 accounting for the excess of volatile elements (mainly H_2O and carbon compounds) in the meteorite. The abscissa is the column number of the Periodic Table as arranged by Ganapathy and Anders. Elements that accreted on to the Earth with the same efficiency from an assumed C_1 -type nebula should plot at the same horizontal level in fig. 10a. At first sight there is a lot of scatter, but much of it can be interpreted in terms of uncertainty in the estimates of the Earth's composition. There is no need to abandon for now the convenient assumptions behind the Ganapathy-Anders type of model, though amendments can be expected in the future. Nine specific comments follow: Figs. 10(b) and (c) show major fractionations of the crust and upper mantle with respect to the bulk Earth (fig. 10a), which correspond to crystal-liquid partitioning during the segregation of the crust and upper mantle from lower zones. The pattern for the bulk Earth (fig. 10a) has most elements within a factor of 2 of the requirements for a Ganapathy-Anders type of model, and there are likely explanations for the deviant elements; e.g. old inaccurate analysis for Cd; uncertainty of locations of Hg and Bi; substantial atmospheric loss of H, as shown by arrow. Magnesium is slightly higher than Si because of the high olivine content of peridotite, but could be reduced by putting substantial $MgSiO_3$ (perovskite structure) in the lower mantle, or by decreasing the Fe/Mg ratio there. The 'metal' elements occur at roughly the same level as Mg and Si, and will tend to become slightly higher if extra Fe, Cr, and P are placed in the lower mantle and core. The elements of the 'early condensate' scatter considerably, but plausible adjustments are available: e.g. increase of the Ca and Al com-

ponents of the lower mantle on the assumption that it has not differentiated upwards all its 'basaltic' component (Ringwood, 1975; Tatsumoto, 1978); reduction of Ba by assigning it only to the upper 2–5% of the mantle in conformity with probable occurrence only in amphibole and mica; adjustment of the rare earths by garnet-pyroxene partitioning (many papers listed in Smith, 1977); increase of V in the core. The 'lithophile volatiles 1300–600 K' scatter widely except for Rb, K, and Mn, but F should probably be reduced in order to concentrate it into the upper mantle, and Na probably should be increased greatly, as shown by the arrow, because it enters garnet at high pressure and might couple with high-valence cations (Si, P, Al?) in deep-seated minerals; only Cs is likely to be aberrant, and indeed, it is expected to be more volatile than its companions in the solar nebula. The 'siderophile volatiles 1300–600 K' match quite well within the large uncertainties. Particularly important is the contribution of the troilite component to Cu and Zn, but not to Ga and Ge, which are siderophilic; removal of S and chalcophilic elements from the Earth's core, as in the Ringwood (1977) oxygen-based model, would drastically worsen the fit thereby providing a powerful argument in favour of substantial S in the Earth's core. The 'volatiles < 600 K', shown as open squares, scatter widely, but there are plausible reasons for ignoring Cd, B, and Hg, and lowering Pb; probably the most reliable elements are Th, Cl, Br, and I, which concentrate in the crust and which match quite well. Most of the C and N is assigned to the core by analogy with iron meteorites, whereas H can be doubled by assuming atmospheric escape. Because there is a large tolerance in these elements there is no need to abandon a simple model at this stage, though special processes may be involved for the very volatile elements as suggested by many data for carbonaceous meteorites.

The key role of U as the control element for the early condensate necessitates detailed discussion. If all the radioactive heat were instantly transported to the surface and there were no other heat sources, the U content of the bulk Earth would be near 0.030 ppm (there is some uncertainty involved with processes at the ocean ridges) assuming plausible U/Th and U/K ratios. For C_1 -proportions of 'early condensate', 'silicate' and 'metal', the U content should be c.0.014 ppm (Table IV, column b). Smith (1977) questioned the validity of estimates of U content from heat flow, and Daly and Richter (1978) showed that convection processes could retard heat flow up to about two-fold. By assuming that Ba has been differentiated entirely into the outer 200 km it may be possible to estimate its

abundance accurately from crustal and upper-mantle rocks. In the meantime it seems certain that the elements of the 'early condensate' are not depleted and might be augmented by $1.5(\pm 0.5?)$ over (Mg + Si), as in Table IV, column c.

The bulk Fe content of the Earth has a strong effect on the mean density, which is accurately known. When the equation-of-state has been determined accurately for possible high-pressure phases, it may prove possible to place accurate bounds on the ratio of Si to the sum of the other major elements. In the meantime it is certain that Fe is not depleted and might be augmented by $1.2(\pm 0.1?)$ over (Mg + Si) in C1 composition. Actually the Fe content of C1 meteorites is poorly defined (Kerridge, 1977a), and substantial separation of Fe and silicate has occurred in chondrites (especially enstatite chondrites); hence it is not really certain what was the Fe/(Si + Mg) ratio of the nebula, and the above value of $1.2(\pm 0.1?)$ used in Table IV, column c, must be treated cautiously.

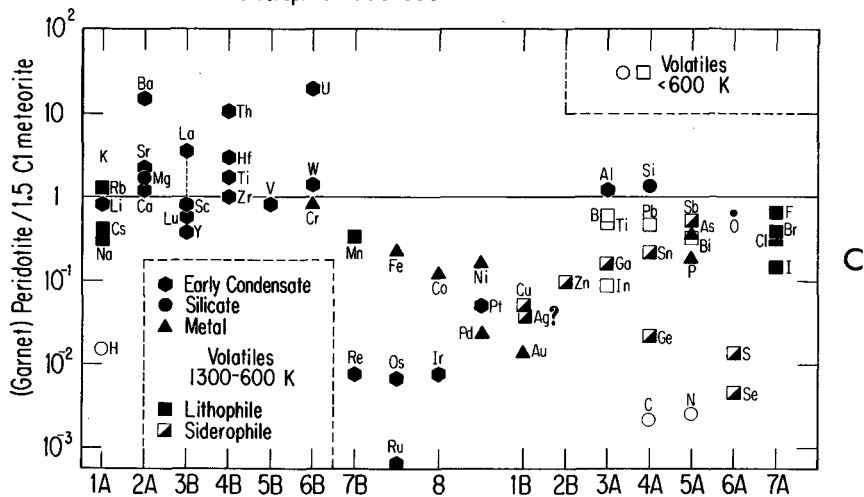
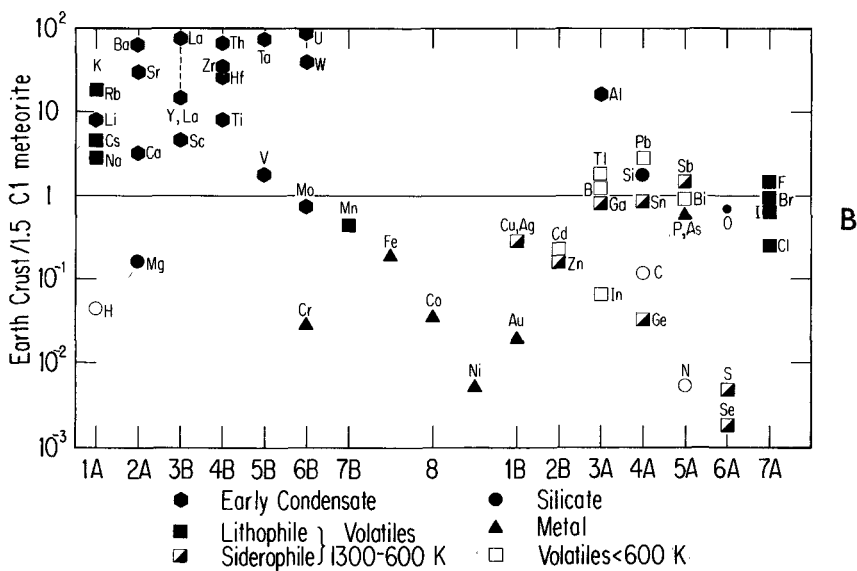
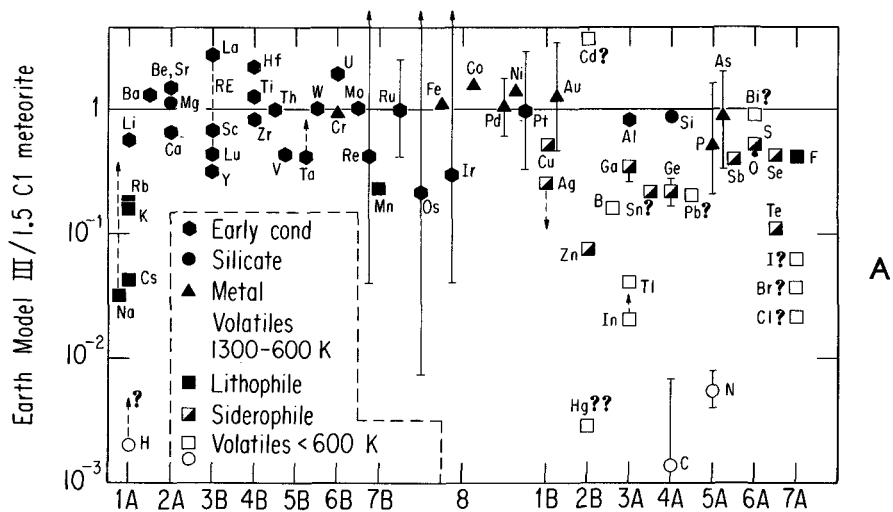
The bulk K and Rb contents of the Earth are highly controversial, though there is no need to invoke substantial difference of the K/Rb ratio of the Earth from the C1 reference. An absolute lower limit of 90 ppm K results from the assumption that all ^{40}Ar is in the atmosphere, and a model-dependent upper limit of 170 ppm is based on the assumption that all Ar was released to the atmosphere as easily as H_2O . Unfortunately there are three permissive factors: present-day emission of noble gases from the mantle argues for incomplete differentiation, and hence for retention of substantial K in deep-seated minerals; pyroxene may be stable down to c.400–500 km, and may carry substantial K as mentioned earlier; and substantial K may enter a S-rich core, especially at high pressure (Goettel, 1974). I would not like to increase the abundance of K more than two-fold in fig. 10(a) because there is no plausible way to increase the abundance of K over that of Na in nebular condensates and indeed it seems more plausible to reduce the Na content of the mantle with depth rather than increase it, and secondly, because K

could be flushed outwards with 'basaltic' liquid during early differentiation of the Earth, and not react at high pressure with sunken FeS-rich material (Smith, 1977).

Obviously there are many plausible and non-definitive arguments which could be discussed indefinitely, but the following five chemical features of the Earth must be explained in any model for its origin: First, its bulk composition corresponds generally to accretion models involving only slight fractionation of the 'early condensate', 'silicate', and 'metal' components, and major depletion of elements expected to condense below 1300 K and especially below 600 K. Second, major crystal-liquid differentiation of the Earth flushed out elements which enter 'basaltic' liquid. Third, substantial volatilization of the bulk Earth is ruled out, except perhaps at an early stage of growth, because of the similarity of depletion factors for elements of widely different thermal volatility [note that 'thermal volatility' is quite different from volatility in the solar nebula]. Fourth, siderophile and chalcophile elements were retained substantially in the upper mantle and crust (fig. 10b, c) at a late stage of accretion, and fifth, storage of major amounts of Ca, Al, Na, Ti, etc. occurs deeper than 200 km. Furthermore, there is the probability of substantial loss of H through the atmosphere, and major storage of C and N in the core and mantle.

Detailed discussion is needed on the variation of redox state with depth. Certainly the present crust contains substantial ferric iron, while upper-mantle peridotites have a high ratio of $\text{Fe}^{2+}/\text{Fe}^{3+}$. The occurrence of diamond in peridotites, and the expulsion of CO_2 in carbonatitic and kimberlitic magmas, together with the indication that volcanic gases are buffered by magnetite-silicate (Nordlie, 1977), testify to absence of extreme redox states in the upper mantle. However, it is not clear what is the redox state of the lower mantle and core, especially in view of possible effects of very high pressure on transition metals and even alkali metals. It is probably wise not to rule out the possibilities that the Earth accreted substantial

FIG. 10. Relation between Earth composition and C1 composition. The base is labelled with column numbers from the Periodic Table, and the elements are classified according to the sequence of equilibrium condensation of a solar nebula with C/O c.0.6 and 10 Pa pressure. The 1.5 factor allows for removal of very volatile elements from C1 composition. (a, top) Model III for the bulk Earth is based on 38 wt % troilite of meteoritic composition and 62 wt % type IIIAB iron meteorite in the core and garnet peridotite composition for the entire mantle except for occurrence of large-ion-lithophile elements only in the outer 10 % (putting Na in the whole mantle would raise Na above Rb and K as shown by arrow), 'andesitic' crust, and the hydrosphere and atmosphere. The primordial content of H might have been about three times higher as shown by the arrow. Particularly uncertain elements are shown by question marks. Ranges for elements result from element partitioning in type III meteorites, and the symbol is placed at the log median value. (b, middle) Earth's crust. (c, bottom) Peridotite (mainly garnet) as model for uppermost mantle. The significance of this type of plot is explained in Ganapathy and Anders (1974), and the many sources of data, along with detailed tables, are given in Smith (1977).



amounts of highly reduced material, and that the outer 200 km became oxidized either from late accretion of oxidized material or from major atmospheric loss of H and subsequent subduction of oxidized rocks.

It is not clear how much of the radial heterogeneity of the Earth is the result of heterogeneous accretion, and how much is the result of post-accretion crystal-liquid differentiation. At one extreme, accretion might have been near-homogeneous with crystal-liquid differentiation proceeding by partial melting, whereas at the other extreme, accretion might have been strongly heterogeneous with 'early condensate' and 'metal' fractions accreting first. Whatever the outcome of

discussion, cosmochemical modelling implies that the middle and lower mantle contain large amounts of Na, Al, and Ca, and that the early differentiation swept out only some of the 'basaltic' component to the surface. Further discussion is needed on the implications of this important conclusion, but it may mean that the Earth was never fully molten, or that it was not molten after it had accreted to (say) half its present mass. Perhaps the Earth is now moving increasingly from an early era dominated by crystal-liquid differentiation of 'wet' K-bearing peridotite to a new era dominated by second-stage differentiation of deep-seated 'dry' K-free peridotite. Will plate-tectonic processes involve subduction to increasingly greater depths?

MOON

'Then . . . the moon must be older than the earth?'. 'No!' said Barbicane decidedly, 'but a world which has grown old quicker . . .'

Jules Verne, *From the Earth to the Moon*

This section is deliberately brief because of the existence of numerous overlapping reviews (e.g. over-all view, Taylor, 1975; classical mineralogy, Frondel, 1975; petrologically oriented mineralogy, Smith and Steele, 1975; survey of lunar rock types, Wood, 1977; mare basalts, Papike *et al.*, 1976; Papike and Vaniman, 1978; ANT-suite rocks, Prinz and Keil, 1977; KREEP-rich basalts, Meyer, 1977). For brevity, detailed references are given here only to controversial subjects. Readers can find appropriate up-to-date primary references to most subjects in the *Proceedings of the Lunar Science Conferences*, and secondary reviews in many places, including Massey *et al.* (1977), *Reviews Geophysics and Space Physics*, and *Phys. Chem. Earth*.

Observations

Although the Moon's surface has been reworked by impact of bodies ranging from tens of kilometres diameter down to micrometeorite size, the absence of plate-tectonic processes and the lack of mineral alteration has allowed reconstruction of the major events in the Moon's petrologic history merely from detailed study of surface rubble returned from nine sites. Most returned homogeneous rocks are mare basalts attributed to late volcanic flows in impact basins, and the early lunar crust is represented mainly by breccias, though a few homogeneous rocks survived (Phinney *et al.*, 1977). Painstaking study of tiny fragments in breccias and soils was one of the keys to reconstruction of the petrologic history of the Moon, while photo-geologic mapping coupled with geochemical mapping from Earth-based telescopes and orbiting satellites was essential for placing the petrologic

information in a geochemical and geophysical context. Great emphasis was placed on the chemistry of minerals and rocks in the development of petrogenetic models, and the lack of stratigraphic control was thereby mitigated.

The lunar rocks and minerals can be crudely classified into five groups: Very rare 'ultrabasic' material (mostly small fragments and mineral grains) contains Mg-rich olivine and pyroxene, which can be attributed to early products of crystal-liquid differentiation. A second group of rocks, minerals, and glasses has chemical, and sometimes petrographic, properties similar to anorthositic, noritic, or troctolitic rocks: this ANT suite is attributed to an early heterogeneous crust rich in plagioclase, low-Ca pyroxene, and olivine. A highly diverse group of fragments (mostly from metamorphosed breccias), whose acronym KREEP refers to enrichment in K, REE(Y, Ln), P, Ba, Th, and U has a considerable range of enrichment and bulk composition; although separation of 'basaltic' liquids from olivine, plagioclase, and pyroxene crystal residues is needed to give the enrichment of this third group, it is not clear what are the relative contributions of primary crystal-liquid differentiation, of secondary partial melting of crystalline cumulates with trapped liquids, and of multi-step processes and hybridism; particularly confusing is the controversy over glasses and heterogeneous rocks with Low-K-Fra-Mauro basaltic composition, which contains substantial KREEP component, but which may have a variety of origins (Reid *et al.*, 1977). A fourth group of mare basalts and mafic glasses with wide variation of Fe, Ti, and Al has generally high ratios of Fe/Mg, which require

advanced stages of crystal-liquid differentiation, as do the trace-element signatures; partial melting of various heterogeneous Fe-rich cumulates (some also Ti-rich or Al-rich) is the most popular mechanism for origin of mare basalts (e.g. Shih and Schonfeld, 1976), but the original sources of the cumulates, the extent of crystal-liquid fractionation during emplacement, and the possibility of hybridism of magmas are the subject of intense, non-definitive debate. Finally the debris from meteoritic bombardment is represented by rare fragments of meteorites, metals, and alloys, and by ghostly signatures of non-lunar trace elements attributed to dispersed meteoritic debris.

Numerous experimental syntheses at controlled temperature, pressure, and oxidation state provide controls on the possible crystal-liquid fractionations in the Moon, but there is considerable controversy over the depth of origin of mare basalts because of disputes over the crystallization sequence near the solidus, and of uncertainties about the effect of crystal separation on the bulk composition of supposed primary magmas. Certain elemental ratios (e.g. Mg/Fe and Cr/Al/Ti) provide good guides to the direction of differentiation. Many lunar rocks can be represented closely by compositions in the olivine-plagioclase-silica ternary system, and there seems to be a growing consensus that olivine dominates over pyroxene in the bulk Moon; furthermore, the pyroxenes tend to be Ca-poor except in mare basalts. Numerous trace-element ratios provide fairly good controls over the efficiency of crystal-liquid differentiation on the Moon, and are the basis of several estimates of the bulk composition.

Geophysical data provide eight constraints: First, the density of 3.34 g/cm^3 enforces an upper limit of $c.400 \text{ km}$ for an Fe-rich core and $c.700 \text{ km}$ for an FeS-rich core if the mantle is dominated by Mg-rich olivine (atomic Mg/Fe $c.10$). Second, estimates of the moment-of-inertia, which have fallen by amounts greater than stated errors to recent values near 0.392 , require a centre-heavy Moon. Third, seismic profiles (Goins *et al.*, 1977), whose interpretation is complicated by closely similar transmission properties of Ca-rich plagioclase and Mg-rich pyroxene are consistent with a crust rich in plagioclase and pyroxene overlying a mantle rich in olivine; furthermore, poor transmission of S-waves through the Moon's centre requires a hot interior. Fourth, the offset of 2 km between the centres of mass and figure, when taken together with other data, requires a light plagioclase-rich crust, ranging in thickness from perhaps $50 \pm 25 \text{ km}$ on the near-side to $100 \pm 50 \text{ km}$ on the far side, lying on a denser mafic mantle. Fifth, remanent magnetism of early lunar rocks, especially breccias, is controver-

sially attributed either to a dynamo in an early liquid core or to special processes involving external magnetic fields that were especially effective during impact brecciation. Sixth, the electrical-conductivity profile, derived from the perturbation of the solar wind, yielded model-dependent temperature profiles, which allow, but do not require, crystal-liquid differentiation of a lunar interior of Mg-rich silicates. Seventh, estimates of the radioactive heating from the surface heat-flow indicate, but perhaps do not definitively require, two-fold enrichment of U in the bulk Moon with respect to devolatilized CI-meteorite composition, and, eighth, the occurrence of mass anomalies only for late basins shows that such mare-basalt fill arrived too late for isostatic adjustment with adjacent non-mare regions.

Orbital missions have provided detailed photo-geologic maps, which were compared with chemical maps of some of the equatorial regions for Mg/Al/Si (X-ray technique) and Th, K, U, and Fe (gamma-ray technique). Important results are: a first-order distinction between mare and non-mare (highland) regions; second-order distinctions correlatable with details of mare flows and impact ejecta; a general correlation between crustal thickness (estimated from isostatic compensation of elevation) and chemistry (Fe negative; Al positive; Th complex); irregular concentrations in the Imbrium-Procellarum region of KREEP-rich material; and a tendency for Al/Si/Mg to change with longitude for both inter-mare and highland areas so that the feldspar content appears to increase from west to east on the near side. Spectral reflectance data from earth-based telescopes were correlated with direct observations of returned samples to show that several types of mare basalt occur in the same basin, that several mare units are not represented in returned samples, and that numerous dark deposits of possible pyroclastic origin need to be sampled.

Petrologic interpretations

Taken together, these data provide a good framework for a general petrologic model of the Moon, but the details are highly tentative and difficult to test definitively because of the considerable tolerance in estimates of relative proportions of rock types. The following model incorporates chronological information obtained from isotope studies.

First, the Moon underwent major crystal-liquid differentiation at $c. -4.5 \times 10^9 \text{ yr}$, probably involving complex diapiric processes triggered by impact of large bodies, and perhaps involving the entire Moon. The differentiation produced four products:

a heterogeneous 'crust' composed of feldspar-rich rocks (including anorthositic, noritic, and troctolitic varieties ranging from plutonic to volcanic types); a heterogeneous 'mantle', perhaps statistically stratified from a lower mantle rich in olivine and Mg-rich complex oxides to an upper mantle rich in low-Ca pyroxene and complex oxides less rich in Mg (note that simple crystal-liquid separation of a completely molten Moon could have yielded a gravitationally unstable mantle with heavier Fe-rich minerals lying above lighter Mg-rich minerals, whereas convective overturn or local diapiric processes could have yielded different spatial distributions); residual liquids, rich in KREEPBaThU components, which tended statistically to concentrate between the 'crust' and 'upper mantle', but which might have become distributed erratically throughout the 'crust' and 'upper mantle' by volcanic and impact-induced processes; and perhaps a small core of Fe,S-rich material. The time-temperature-depth history of the Moon is poorly constrained in spite of numerous model calculations, and whether a single deep 'magma ocean' actually existed is highly speculative. The extent of gravity-separation or elutriation is debatable, and plagioclase flotation in an asymmetric gravity field, controlled by a near-by Earth, is just one potential way to develop the Moon's asymmetric crust, while differential bombardment is another.

Second, the primary differentiation tailed off into a period of complex secondary differentiation whose details are quite uncertain. Perhaps little happened in the 'lower mantle' and 'core', while the major action occurred in the 'crust' and 'upper mantle' which were subjected to impacts from late planetesimals, and to radioactive heating, especially from KREEPBaThU-rich regions. Partial melting in the 'crust' may have resulted in cotectic melts, while huge impacts may have homogenized portions of the crust giving a range of rock types and Low-K-Fra-Mauro compositions. Whether there was a major basin-forming 'cataclysm' at -3.9×10^9 yr or whether the recognizable basins were formed randomly from -4.3 to -3.9×10^9 yr is highly controversial, but from the petrologic viewpoint there can be no doubt that much of the lunar surface was affected by complex metamorphic processes associated with major impacts (Phinney *et al.*, 1977). An alternative to the suggestion of a quiescent 'lower mantle' is that trapped liquid from the primary differentiation seeped upwards and hybridized with the material in the 'upper mantle' and 'crust'. Whether the lunar crust is stratified into upper 'anorthositic gabbro' composition overlying middle 'Low-K-Fra-Mauro' composition and lower 'mafic ANT'

composition (Ryder and Wood, 1977) is debatable, as are the relative effects of primary and secondary differentiation in achieving such a presumed stratification.

Third, cessation of the basin-forming impacts, and reduction of the radioactive heating, allowed the lunar surface to largely stabilize into the present stratigraphic provinces of ancient highlands, impact debris, old irregular maria, and young circular maria. Partial melts from the 'upper mantle' yielded mare basalts. These may have been forming before cessation of the basin-forming impacts, and the returned mare basalts may be providing information mainly on the final capping of the maria after a presumed long history of igneous and metamorphic activity in the early impact debris and the uprising mantle (and crust?) under the impact basin. The late mare basalts must have arisen from considerable depth in the Moon ($c. 50-500$ km?) in order to allow a sufficiently rigid crust above the weak source region, but the estimates of depth from presumed cotectic birth of the magmas is controversially dependent on subtle details of experimental phase equilibria. Crystallization ages of mare basalts range from -3.9 to -3.1×10^9 yr, but earlier ones may have been lost in the metamorphosed debris of impact basins, while later ones indicated by photogeologic mapping did not turn up in samples returned to Earth. Large compositional variations of Fe, Al, and Ti in the mare-basalts and mafic glasses indicate derivation from heterogeneous source regions, or crystal-liquid differentiation during ascent and emplacement, or both. Paucity of mare basalts on the far side of the Moon might be explained by the greater hydrostatic head needed to penetrate the thicker far-side crust, or by lower production of mare basalt in the far-side mantle, or both.

This generalized model stitches together and modifies the original general ideas of early major differentiation to produce a feldspar-rich crust and an olivine-rich mantle (Smith *et al.*, 1970a, b; Wood *et al.*, 1970) and partial melting of igneous cumulates to produce mare basalts (Philpotts and Schnetzler, 1970). Probably most lunar petrologists are currently accepting these two general features, but many of them prefer to differentiate only some hundreds of kilometres rather than the entire Moon. Detailed discussions of crystal-liquid differentiation in a magma ocean, together with controls imposed by mineral and rock chemistry (e.g. Mg/Fe ratio, (Y,Ln) signature, Ni and Cr contents) can be found in recent *Proceedings of the Lunar Sciences Conference*.

A second type of model was developed from the original proposition by Ringwood and Essene (1970) that Apollo 11 basalts originated by partial

melting of a (primordial) pyroxene-rich Moon generated from a hot fission cloud thrown off the Earth. The latest version (Ringwood and Kesson, 1977) has adopted the Smith-Wood concept of an early major differentiation of the Moon (but only to 400 km depth), which produced a gabbroic crust (c.50 km thick) overlying an intermediate layer (c.50 km) and a refractory residue of dunite (c.300 km), but invoked generation of mare basalts by partial melting of a diapir rising out of deep-seated (c.500 km) primordial olivine pyroxenite into the overlying dunite zone. Presumably the central zone of the Moon still remains as undifferentiated primordial pyroxenite.

Unfortunately it will prove difficult to distinguish the Philpotts-Smith-Wood class of models from the Ringwood-Kesson model because of four reasons. The RK model has adopted an early differentiation of half the Moon's volume according to the Smith-Wood model; the rising melt from the RK diapir can interact with the cumulates produced in the early differentiation, which are the source of mare basalts in the Philpotts model; the experimental partition coefficients for element distribution between liquid and crystals have considerable tolerance; and the initial chemical parameters of model Moons (e.g. Mg/Si ratio) have some tolerance. Geophysical tests are probably inconclusive because of poor control over the properties of the Moon's interior. If the lunar magnetism could be uniquely attributed to a dynamo (Runcorn, 1976), complete differentiation of the Moon would be indicated. Ringwood and Kesson (1977) pointed to the problem of heat sources for melting the entire Moon, but Toksöz and Johnston (1977) were able to develop a thermal model for complete differentiation of the Moon. Brett (1977) also questioned whether the bulk of the Moon melted. Solomon and Chaiken (1976) suggested that the absence of major fractures in the lunar surface limited an early 'magma ocean' to only 200 ± 100 km depth (now revised to 300 ± 100 km; Solomon, pers. comm.), but detailed study is needed to check whether such a thin ocean can provide a plagioclase-rich crust thick enough (c.75 km?, Kaula, 1977a) to explain the offset between the centres of figure and mass and allow sufficient trace-element enrichment of mare basalts.

Probably it will be difficult for geochemists, mineralogists, and petrologists to come up with convincing controls on the deep interior of the Moon, but there is great scope for further study of near-surface differentiation.

Bulk chemical composition

Fortunately from the viewpoint of planetary mineralogy there is a substantial convergence of

opinion about the bulk chemical composition of the Moon. This arises mainly from the tendency to assume that at least half of the Moon's mass participated in an early major differentiation into a plagioclase-rich crust and an olivine-rich complement, together with likelihood that the undifferentiated interior (if any exists) is dominated by Mg-rich silicate rather than non-peridotitic compositions. Early extreme views, such as a Moon dominated by Ca,Al-rich nebular condensate (Anderson, 1973), are quite unacceptable when the surface rocks are dominated by 'basaltic' compositions, and recognition of the uncertainty of the estimate of U content from heat-flow measurements (Langseth *et al.*, 1976) has allowed reduction of early estimates of high U and cosmochemically associated Ca, Al, and Ti (Ganapathy and Anders, 1974). All recent estimates of lunar composition (Taylor, 1976; Binder and Voss, 1978; Taylor and Jakeš, 1977; Dreibus *et al.*, 1977; Kaula, 1977a; Kesson and Ringwood, 1977; Longhi, 1977; Ringwood and Kesson, 1977) are dominated modally by olivine, pyroxene, and plagioclase, as in early models by Smith *et al.* (1970a, b) and Wood *et al.* (1970), and can be described as peridotitic.

The eight key features of all recent model Moons (Table IV) are: the low content of FeO (10–14 wt%), whose maximum value is strongly constrained by the mean density for all models dominated by Mg-rich olivine; the very low content of volatile elements, as typified by the calcic nature of almost all lunar plagioclase (thereby requiring low Na and K), and the almost complete absence of H₂O in lunar rocks; the very low content of metal-seeking elements (e.g. Ir and Au) in lunar surface rocks, thereby allowing the extent of meteoritic contamination to be estimated by measuring these elements in soils and breccias (Hertogen *et al.*, 1977); the dominance of Mg-rich olivine in the differentiated lunar interior in order to allow primary crystal-liquid differentiation to proceed to ANT-like compositions (e.g. Drake, 1976; Longhi, 1977); the limitation of Mg/Fe ratio in the olivines by observed mineral compositions and crystal-liquid fractionation factors; the limitation of Ca and Al from constraints in the amount of Ca-rich pyroxene and plagioclase produced during differentiation (recent estimates ranging from 4 to 8 wt% of Al₂O₃ are lower than earlier values of 11 wt% linked to high estimates of U from the heat flow); the limitation of many trace elements from model calculations of crystal-liquid differentiation (e.g. Taylor, 1975); and absence of precise control over the amount of S in a hypothetical lunar core, except that an upper limit of 1 wt% seems likely if the Moon is dominated by Mg-rich olivine.

ORIGIN OF THE EARTH AND MOON

‘... and who can say that the moon has always been a satellite of the earth?’

Jules Verne, *From the Earth to the Moon*

Because the origin of the Earth and Moon is so highly controversial, the existing geochemical-mineralogic-petrologic data must be utilized as much as possible in placing boundaries on models. If one accepts the concept of planetesimals as intermediaries between nebular condensates and planets, one must expect the Earth-Moon system to have developed in a complex way. Perhaps most lunar scientists would now agree that the three simple models of fission of the Moon from the Earth, non-disintegrative capture of the Moon by the Earth, and simultaneous accretion of the Moon and the Earth, are insufficient, and that it is necessary to move to more complex models. Readers should note that I have discussed disintegrative capture of a large proto-Moon (Smith, 1974), but in this review I shall try to stand aloof in the tradition of Chamberlin's *Method of Multiple Working Hypotheses*, and indeed now suggest that no single process can explain the origin of the Earth-Moon system.

Age constraints

The crystallization ages of the mare basalts require that the Moon was orbiting the Earth at least by -4.0×10^9 yr, and it is easiest to interpret the model ages of radioactive species in terms of an early major differentiation at $c. -4.5 \times 10^9$ yr. Capture of the Moon by the Earth would be so catastrophic that major geologic provinces would have been obliterated. It is easiest to assume that the lunar chronology began with the Moon in Earth orbit by -4.5×10^9 yr.

The Moon was certainly bombarded by large projectiles up to $c. -3.9 \times 10^9$ yr, and the Earth must have been affected by the same population of bodies. Furthermore the effects on the Earth should have been more catastrophic than on the Moon, and it is necessary to consider impact-triggered volcanic processes in the outer part of the Earth up to $c. -3.9 \times 10^9$ yr. This is consistent with the earliest crystallization ages of $c. -3.8 \times 10^9$ yr for terrestrial rocks.

Chemical constraints

This section is based largely on Smith (1977, 1978).

Incomplete equilibration of the Earth is indicated by the well-known occurrence of substantial amounts of metal-seeking elements in the Earth's upper mantle (e.g. 0.3 wt% Ni in upper-mantle

olivine). No matter whether the Earth accreted homogeneously or near-homogeneously, it is necessary to assume that a mantle barrier blocked equilibration between the upper mantle and core. Unfortunately it is extremely difficult to calculate the rate at which equilibration would occur even in a completely molten Earth, but it is qualitatively easier to envisage lack of equilibration in an Earth cool enough to have a solid mantle when (say) 80–90% of the Earth had accreted.

The alkali metals in the bulk Earth are depleted not more than six-fold with respect to water-free, CI-meteorite composition, and it is difficult or impossible to retain these alkalis in Ringwood's (1975) volatilization model in which refractory elements such as Mg, Si, Ca, and Al are supposed to have volatilized from a very hot Earth to form the Moon.

Estimates of the bulk composition of the Earth are subject to considerable uncertainty from lack of direct knowledge of the lower mantle and core. It is possible to satisfy geophysical constraints with models in which the major elements Si, Mg, Fe, Ca, and Al have nearly the same ratios as in CI meteorites, though perhaps there is enrichment of Ca and Al (1.5-fold?) and Fe (1.2-fold?). Volatile elements are considerably depleted, but there is no need to attribute the depletions to any special factors in the origin of the Earth-Moon system; rather, it is initially easier to attribute them to processes involved in nebular condensation (Ganapathy and Anders, 1974) and catastrophic accretion of planetesimals (many authors).

Perhaps the chemical composition of the Moon is more closely constrained than that of the Earth. It has several important features (Kaula, 1977a): low abundance of Fe (three-fold lower than Earth?), and almost certainly of all siderophile elements; not more than two-fold enrichment of early-condensate lithophile elements with respect to CI composition (Ca, Al, U); very low abundance of all volatile elements with respect to CI composition, and even with respect to Earth; and dominance by Mg-rich olivine composition, as is also likely in the bulk Earth.

Dynamical and accretional constraints

Origin of the Moon by simple rotational fission faces momentum and energy problems (e.g. Kaula, 1971, 1977b), but these can be obviated by combining rotational fission with other mechanisms.

It may be possible to develop useful constraints on accretion. If Safronov's models for the size distribution of accreting planetesimals can be adequately developed to cover the complex processes involved in collisions, it may prove possible to place an upper limit on the size of planetesimals; current theory is incompletely developed (Wetherill, 1976). Although the ratio of escape velocity to impact velocity is extremely important in determining the fate of colliding bodies, it is too early to rely on model calculations; present calculations suggest that the Moon is and was very much less efficient in retaining collision debris than the Earth (e.g. O'Keefe and Ahrens, 1977*a, b*), especially for material arriving at 10 km/s or greater.

Comparison of suggested origins

Fission. Shoemaker (1977*b*) suggested that accretion of the Earth from large planetesimals would inevitably result in major changes of angular momentum as the result of random off-centre collisions, and indeed he proposed that off-centre collision by one or more particularly large planetesimals would lead to one or more episodes of fission. Actually one can expect a whole gamut of events ranging from head-on collision via glancing collision (Hartmann and Davis, 1975; Cameron and Ward, 1976) to disintegrative capture of mechanically weak bodies in the Roche instability zone (Öpik, 1972; Smith, 1974; Mitler, 1975), and it may prove impossible to estimate the relative proportion of the types of events. Three types of chemical differentiation can be envisaged: on Earth, such that a piece of the mantle was flung or torn off, leaving the lower mantle and core in place (Binder, 1976; Ringwood and Kesson, 1977; Ram-mensee and Wänke, 1977); in orbit, such that volatiles were lost, and mechanical differentiation occurred for fragments of different chemical composition (Smith, 1974; Kaula, 1977*a, b*); and thirdly, during partial capture of the colliding body, which initiated fission (Smith, 1974; Mitler, 1975), especially if it had already differentiated chemically in heliocentric orbit.

There is no great difficulty in accounting for a Moon low in volatiles by a fission-dominated event. There is no problem in obtaining an olivine-rich Moon since both the outer mantle of the proto-Earth and colliding planetesimals might be olivine-rich. Deficiency of Fe and metal-seeking elements in the Moon could be explained by obtaining the Moon largely from the outer mantle of the proto-Earth; furthermore, the deficiency might be augmented by more efficient capture of silicate debris from planetesimal(s) that initiated fission. Smith (1977) emphasized that it is difficult to retain

substantial amounts of volatile elements on Earth if simple rotational fission had occurred, but in 1978 pointed out that late accretion of volatile-rich material by the Earth after fission had occurred could account for the volatile elements so long as the accreting material arrived at too high a velocity (c. 10–20 km/s?) to be retained significantly on the Moon.

Disintegrative capture. One extreme version of this mechanism involves a single large planetesimal (c. $4 \times$ present lunar size?) which had differentiated chemically into a metal-rich core and silicate-rich mantle, while another consists of fortuitous head-on collision of two planetesimals near the Earth (Levin and Mayeva, 1977). The Moon would be formed from debris produced during the capture, and numerous possibilities exist for separation of silicate from metal components as in the preceding paragraph. Less extreme versions might involve disintegrative capture of many planetesimals, which produced debris that accumulated over a period of time to form the Moon, and indeed such versions might be classified under simultaneous accretion. Unfortunately it is impossible to pin down the many possible chemical fractionations that are available, especially because of the unknown factors involved in the relative mechanical stability of an iron-rich core and a silicate-rich mantle. Many complex possibilities are also involved in the accretion of Earth-orbiting debris (Öpik, 1972; Safronov, 1972; Kaula, 1977*b*).

From the general similarity of the chemical composition of the Moon to that of the supposed eucrite planetesimal, Anders (1977) suggested that a special process is not needed to explain the origin of the Moon. Indeed it might be argued that a supposed proto-Moon already had a Moon-like composition prior to disintegrative capture. However, the eucrites are particularly depleted in volatile and in siderophile elements, and it might be unwise to press analogies with the Moon.

Since planetesimals growing at the edges of planetary feeding zones should grow longer than ones in the middle of the zones, might a proto-Moon have formed approximately mid-way between Earth and Mars, and been captured near the end of Earth's accretion?

Simultaneous accretion. This process could not have begun unless a proto-Earth and orbiting proto-Moon had formed by either capture (disintegrative?) or fission. Furthermore, simultaneous accretion must have occurred to a greater or lesser degree for models dominated by fission or disintegrative capture. Whether the difference of impact velocity for the Earth and Moon could account for most of the chemical differences between the Earth and Moon is debatable.

Ganapathy and Anders (1974, p. 1201) argued that the different chemical compositions of the Earth and Moon result merely from selective accretion of aerodynamically sorted dust and that the escape velocity is unimportant. However, this ignores accretion of large bodies for which O'Keefe and Ahrens (1977*a, b*) imply poorer collection efficiency for bodies impacting the Moon than those hitting the Earth.

Particularly important are the ideas of Ruskol (1977) for simultaneous accretion of the Earth and Moon, in which great importance is attached to the formation of a swarm of small bodies around the Earth during its growth. The capture efficiency of the swarm is greatly reduced if the swarm accretes into a proto-Moon or several moonlets. Incoming bodies might disrupt the proto-Moon or moonlets. Ruskol suggested that the Moon accreted most of its material only at the end of the accretion period of the Earth, and that its distinctive chemistry arises because the late heliocentric bodies had undergone more collisions than the ones captured early by the growing Earth; these late bodies would be volatile-poor and richer in brittle silicate than

coherent iron. Kaula (1977*b*) adopted Ruskol's general model, but augmented it by further discussion of chemical differentiation of impact debris in Earth orbit. Complex dynamical factors are also important, and one can envisage complex possibilities for accretion near the Roche limit, and sequential capture of moonlets (Öpik, 1972; Smith, 1974).

Combined processes. Currently it seems plausible to attempt to explain the origin of the Earth and Moon in terms of combined processes. In general, it seems best to consider fission (coupled with disintegrative capture) only for the first half of the Earth-Moon accretional history, while simultaneous accretion (again coupled with disintegrative capture) became the dominant process late in the history. Probably the evidence on the details of accretion of the Earth and Moon has mostly disappeared leaving only ghostly hints such as the mare basins, and the chemical disequilibrium of the Earth. The iron depletion of the Moon could be the sum of effects occurring in helio- and geo-orbiting bodies, and the low abundance of volatiles could be the sum of several processes.

MERCURY

Look here upon . . . the herald Mercury
New-lighted on a heaven-kissing hill.

William Shakespeare, *Hamlet*, III. iv. 53.

Because of the paucity of direct evidence on mineralogy, petrology, and chemistry of Mercury, all ideas ultimately depend on cosmochemical models. The following account is based mainly on papers in vol. 28, no. 4 of *Icarus*.

Mercury has an eccentric orbit (the result of a late large impact?), and a rotation period (58.65 days) precisely in two-thirds resonance with its orbital period (almost certainly from tidal interaction with the Sun). Its uncompressed density ($c.5.3 \text{ g/cm}^3$) can be explained only by a high content of Fe, as correctly adduced by Urey (1952). For an Fe,Ni core and a diopside-rich mantle, approximately two-thirds of the mass is contained in the core. Siegfried and Solomon (1974) estimated 0.66 assuming $\text{Fe}_{0.9}\text{Ni}_{0.1}$ in the core, and silicates with uncompressed density 3.2 g/cm^3 in the mantle. Substitution of a light element in the core (e.g. S or Si) would require compensating reduction in the proportion of silicate, and indeed the uncompressed density of troilite (4.8 g/cm^3) is well below that of Mercury. The Mercurian atmosphere can be explained by temporary capture of light species from the solar wind, and there is no evidence for substantial volcanic outgassing (Kumar, 1976) or significant weathering of craters.

Earth-based telescopic study shows a continuous increase of reflectance from 0.32 to $1.05 \mu\text{m}$ similar to that for spectra of the lunar surface except for the lack of a significant absorption band at $c.0.95 \mu\text{m}$ in the Mercurian spectrum (Vilas and McCord, 1976; Hapke, 1977). This absence indicates $< 6 \text{ wt } \% \text{ FeO}$ at the Mercurian surface, and the general trend between reflectance and wavelength indicates a mature regolith, probably dominated by breccias and glasses. Adams and McCord (1976) suggested that the surface composition is anorthositic, but I suggest that other minerals with a featureless spectrum should be considered, especially as Mercury need not have accreted substantial amounts of feldspar according to simple condensation models of the solar nebula: likely candidates are diopside, spinel, forsterite, enstatite, and melilite. The Mariner photographs of the Mercurian surface (Murray *et al.*, 1975) revealed smaller variations of albedo and colour than on the Moon, indicative of lower range and absolute level of Fe and Ti.

The surface of Mercury is heavily cratered, and displays conspicuous basins (especially Caloris) and fairly smooth areas subdivided into pre-basin 'inter-crater plains' and post-basin 'smooth plains'. Long, low cliffs ('lobate scarps') were attributed to

compression of the crust as a result of continued cooling of the interior after consolidation of the crust. Although there is no unequivocal evidence of volcanism, such as flow fronts, the plains are commonly ascribed to volcanic activity (Trask and Strom, 1976) though Wilhelms (1976) attributed them to impact melt or other basin ejecta by analogy with light plains on the Moon. Lowman (1976) attempted to match the inter-crater plains with the pre-Imbrian lunar highlands and the smooth plains with lunar maria, thereby requiring two episodes of igneous differentiation on Mercury. Solomon (1977b) argued that compressional tectonic features require early melting. Whatever the outcome of the speculation on the surface history of Mercury it seems safe to conclude that any major crystal-liquid fractionation of the planet must have taken place before the basin-forming events (? before -3.9 Gyr).

Particularly important is the firm evidence for a dipole magnetic field, with its implications on the thermal evolution. Ness *et al.* (1976) demonstrated severe difficulties in attributing the dipole to fossil remanent magnetization, principally because the expected heat sources should melt Mercury, and because subsequent cooling would allow only a thin outer shell (probably very low in Fe-metal) to go below the Curie point. Ironically, the alternative assumption of a dynamo raises the problem of finding an energy source to drive it at the present time (latent heat of crystallization of an inner core?). Existing calculations (Siegfried and Solomon, 1974; Cassen *et al.*, 1976; Solomon, 1976, 1978a) suggest that an iron core would solidify during 1 to 3×10^9 yr unless substantial heat sources remain in the lower mantle or core, or perhaps if the viscosity and thermal conductivity were incorrectly estimated. The cosmochemical expectation of no volatile elements and no Fe substitution in Mg-silicates actually leads to correct qualitative changes in the viscosity and thermal conductivity, but quantitative values have not been estimated so far. Cassen *et al.* discussed the possibility of an iron-sulphide shell, while recognizing that this is inconsistent with the simplest cosmochemical models. Perhaps the assumption that U and Th are completely or nearly partitioned into the silicate fraction may be incorrect, especially as the silicates have higher melting-points than iron. Experiments are needed to test whether U and Th may not partition significantly (e.g. 20%) into Fe-rich liquid with respect to Fe-free Mg-silicates.

As mentioned earlier, the simplest cosmochemical model (fig. 5) for Mercury assumed termination of accretion in the Mercury 'zone' of the solar nebula after most of the iron had condensed and just after Mg began to condense as forsterite. This apparently implies a nebular pressure greater than $c.10^{-4}$ bar (Grossman and Larimer, 1974, Fig. 7), but uncertainty in the thermochemical data must be considered. For 1×10^{-4} bar the final temperature might be $c.1360$ K, which is close to the beginning of condensation of anorthite. If Mercury grew from planetesimals, and if there were significant overlap of the planetary feeding zones, it should contain small amounts of materials condensing at a lower temperature and pressure. The estimated bulk composition of Mercury (Table IV) assumes 90% condensation of Fe alloyed with Ni, Co, and Cr in cosmic proportion; 100% condensation of Al as $MgAl_2O_4$, Ti as $CaTiO_3$, and remaining Ca as $CaMgSi_2O_6$; and sufficient Mg as Mg_2SiO_4 to bring the mass fraction of the core down to 0.66. This simplest model has 66% iron alloy, 20% diopside, 8% $MgAl_2O_4$, 5% forsterite, and 0.5% perovskite. For 100% condensation of iron alloy, there would be 18% diopside, 7.5% $MgAl_2O_4$, 8% forsterite and 0.5% perovskite. Many other variants are possible, including incorporation of some Ca as $CaAl_2Si_2O_8$ and some Mg as $MgSiO_3$ if overlap occurs of planetary feeding zones. For 100% condensation of U and Th, and 90% condensation of Fe, the weight fractions in Mercury of U and Th are 34 and 0.150 ppm. Melting experiments are needed to provide mineralogical models for the Mercurian mantle, but the simplest chemical model cannot produce a surface resembling a terrestrial basalt or anorthosite.

Because Mercury is close to the Sun the impact velocity of late-scattered bodies is high (> 20 km/s), and impacts lead to violent explosions with significant loss of debris. Consequently there may be no or very little gain of C1-type material from comets and Jupiter-perturbed bodies: indeed, impacts may have removed a significant amount of the surface material as Mercury was growing.

In conclusion, the arguments of Kaula (1976) provide a warning about the application of a simple condensation model to Mercury, and emphasize the importance of planetesimal scattering. It is fascinating to see how the early ideas of Urey (1952) and Ringwood (1966b) still can be seen underlying the more complex ideas of recent years.

VENUS

And the planet of Love is on high
Beginning to faint in the light that she loves
On a bed of daffodil sky.

Alfred, Lord Tennyson, *Maud*, I. XXII. i

There are no detailed mineralogical data on the 'Queen of Heav'n, with crescent horns', but γ -ray spectra from Venera landers indicated granitic and basaltic compositions, thereby implying planetary differentiation. Although the bulk composition of Venus is only weakly constrained the similarity of K/U ratio for terrestrial and Venusian surface rocks, and the approximately equal abundance of CO_2 , suggest that Venus and Earth have similar bulk compositions. However, there is considerable tolerance in the ratios of Fe, Si, Mg, S, and O. The mineralogy at the Venusian surface must be quite different from on Earth because of the higher surface temperature ($c.740 \pm c.10$ K) and thick atmosphere ($c.90$ bars = 9×10^6 Pa) dominated by CO_2 , and indeed the crust should be dominated by high-grade metamorphic rocks. The present account relies heavily on a review by Hunten *et al.* (1977).

Surface properties

Radar interferometry allows penetration of the thick clouds, but even the most detailed maps produced up to 1976 (Campbell *et al.*, 1976; Goldstein *et al.*, 1976) have a formal resolution of 10–20 km and a geologically meaningful resolution of perhaps $c.100$ km. Variations in reflectivity may result from surface roughness. Deviations from sphericity are mostly less than 6 km, perhaps supporting Mueller's (1969) speculation (qualified by some chemical caveats) that the higher temperature of the Venusian surface results in lower viscosity of Venusian than terrestrial rocks. Some near-circular features ranging from 100 to 1000 km diameter might be tentatively identified as craters, some ten times shallower than lunar craters. At least two bright areas of elevated terrain might be volcanic shields. Saunders and Malin (1977) reported that the radar-bright feature 'Beta' is 700 km diameter with elevation up to 10 km above the surrounding terrain. A 60×90 km depression at the summit is part of the evidence that 'Beta' is a huge volcanic construct analogous to ones on Mars. Linear features may be canyons or rifts. Venus is brighter at radar wavelengths than Mercury or the Moon, perhaps indicating a thinner regolith overlying solid rock. This is not surprising since the thick atmosphere 'blankets' the surface from small meteoroids. There is insufficient evidence to test whether plate tectonic processes operate on Venus.

The Venera 9 camera revealed boulders littered on a fine-grained surface. The boulders appear to be angular to discoidal with horizontal dimension up to 70 cm and depth up to 20 cm. At least two are layered, and one looks like a hamburger. The Venera 10 photograph shows a curious mottled pattern apparently of low relief suggesting fine-grained material partly covering rock outcrops. There is no obvious morphological clue to the degradation processes necessary to explain the photographs (Florensky *et al.*, 1977). Observed wind velocities of 0.4–1.3 m/sec should be sufficient to erode loose material because of the denseness of the atmosphere, and chemical and aeolian erosion may act in conjunction.

Gamma-ray measurements yielded the following analyses:

	Venera 8	Venera 9	Venera 10
K	4.0 (1.2) wt %	0.47 (8)	0.30 (16)
Th	6.5 (0.2) ppm	3.65 (42)	0.70 (34)
U	2.2 (0.7) ppm	0.60 (16)	0.46 (26)

(Surkov *et al.*, 1974; Florensky *et al.*, 1977). Within the experimental errors these analyses lie in the ranges for terrestrial granite (Venera 8) and basalt (Veneras 9 and 10). The Venera 10 rock densitometer gave a density of 2.8 ± 0.1 g/cm³.

Atmosphere

Thick clouds of spherical droplets, 1 μm across, extend from 70 to 50 km above the surface. The refractive index and infra-red absorption spectrum are consistent with H_2SO_4 , but other acids cannot be ruled out as minor constituents, and might be expected as the ultimate products of volcanic exhalations. The yellow colour of the clouds probably results from native sulphur.

The atmospheric pressure at the surface is 90 bar. Carbon dioxide accounts for over 90%, and the balance (0–7%?) can be attributed largely to Ar and N_2 , whose concentrations are still unmeasured but have been estimated by Turekian and Clark (1975). Water amounts to $c.0.1\%$ below the clouds, but some hundred times lower above the clouds. Minor observed constituents are CO 50 ppm, He 10 ppm, H_2SO_4 $c.10$ ppm, HCl 0.4 ppm, HF 0.01 ppm, and $\text{O}_2 < 1$ ppm.

Complex chemical reactions, including catalytic and photo-dissociation processes, are invoked to

explain the composition of the atmosphere. The cloud tops are near 250 K, and the Venus atmosphere acts as a greenhouse with respect to absorbed solar radiation, in spite of the higher albedo of the Venus clouds than of planetary regolith. Radioactive heating from the Venusian interior has only a trivial effect on the heat balance of the atmosphere. Current atmospheric models do not utilize the concept of continuous emission of volcanic gases, though this must have been important for initial development of the atmosphere.

Speculations on bulk composition

Before considering the interaction of the atmosphere with the surface minerals it is necessary to review speculations on the bulk composition of Venus.

The simplest model has identical chemical compositions of Venus and Earth, except for presumed major loss of H from Venus. However, Lewis (1972a) and particularly Ringwood and Anderson (1977) argued that the uncompressed density of Venus is too low with respect to that of Earth. [Note that the conversion factor from the real density depends somewhat on the assumed chemical composition, and the depth distribution of density, pressure, and temperature.] In particular, Ringwood and Anderson estimated that the uncompressed density of Venus is 1.9% less than that of Earth, and that uncertainties in the pretes have little effect, especially as Earth and Venus apparently have similar K/U ratios. Revision of the Venus density from 5.24 g/cm³ to 5.27 (Hunten *et al.*, 1977) reduces the above difference to 1.2%.

Lewis (1972a) suggested that Venus accreted at a higher temperature than Earth, resulting in a mantle of Fe²⁺-free magnesium silicates, a core of Fe, Ni metal, and a silica-rich crust similar to the Earth's crust. Sulphur was supposed to be 'virtually absent', and Lewis (1974) appealed to cometary impacts for supply of S to the atmosphere. Ringwood and Anderson (1977) strongly criticized this model with four propositions: its density is c.1% too high; the K/U ratio of surface rocks is like that of Earth rocks; sulphur is a major component of the clouds, and should also be abundant in surface rocks; and, finally, the high CO₂ content of the atmosphere is not explainable by release from a reduced mantle. Lewis has not had time to reply, but the cumulative effect of these criticisms is rather compelling, and difficult to refute quantitatively though perhaps capable of being met by qualitative arguments based on late addition of CI material and selective loss of atmospheric components.

Ringwood and Anderson preferred the Ringwood model in which Venus has the same relative

abundances of Fe, Si, Mg, Al, and Ca as Earth, but in which Venus is more oxidized than Earth. The greater proportion of oxygen lowers the density, and transfers Fe from the metallic core to FeO in the mantle. The core (16 wt% S) amounts to 23% of the Venus mass, and the mantle contains 13.4 wt% Fe giving mol Mg/(Mg + Fe) 0.76. This assumption of 16 wt% S in the Venusian core is inconsistent with Ringwood's (1977) assumption that the light element in the Earth's core is oxygen and not sulphur, unless nebular condensates have an unexpected variation with a planet's distance from the Sun.

Ringwood and Anderson concluded that Venus accreted at lower mean temperature than Earth because of its smaller size and gravitational accretion energy. They dismissed silicate-metal segregation in the accretion process as implausible because it is based on undemonstrated *ad hoc* processes. Nevertheless I think that metal-silicate segregation must have taken place during the planetesimal stage, as evidenced by many properties of meteorites, and I would not rule out a lower content of Fe metal in Venus than in Earth. I find the similarity of K/U/Th ratios on Earth and Venus compelling evidence that Venus and Earth accreted material with similar volatile content. This allows a simple explanation of the similar content of CO₂ on the two planets (next section) and the substantial S content of the Venusian clouds. In the next section it will be assumed that Venusian surface rocks have similar chemical composition to those on Earth except for differences caused by reaction with the atmosphere and by biological processes. If the Ringwood-Anderson model is correct the Venusian silicate-bearing rocks should have higher Fe/Mg than on Earth, but other models with smaller Fe/Mg ratios are possible.

Speculations on surface mineralogy and atmospheric compositions

The following account largely follows the ideas of R. F. Mueller, who established the key features of the interaction between the atmosphere and surface minerals. Only minor modifications were needed as new data became available.

Urey (1952) suggested that the high content of CO₂ in the Venusian atmosphere resulted from failure of silicates (e.g. enstatite and wollastonite) to react with atmospheric CO₂ to produce quartz and carbonate (magnesite or calcite), perhaps because water was not available as a catalyst. Mueller (1963) and Mueller and Kridelbaugh (1973) considered the rate of reactions, and in particular the latter authors concluded that equilibration should take place in geologically short periods even for dry

rocks. Mueller's conclusion that all the CO_2 of Venus is in the atmosphere, and that there is no buffering by surface minerals was reaffirmed by Orville (1975). Since basaltic rocks should dominate the surface, Orville re-examined possible reactions involving CO_2 for bulk compositions in the forsterite–enstatite–diopside ternary, and found that such compositions would occur as silicates at 741 K and pressures up to several hundred bar of CO_2 . Acid igneous rocks (e.g. diorites and granites) would not react with CO_2 at the Venusian surface. Several reactions listed by Lewis (1970) involve minerals that could not be stable at the surface (e.g. jadeite and åkermanite).

I now suggest that some CO_2 might be stored inside Venus. The suggested range (fig. 2) of pretes for Venus not only intersects the field of stability of meionite, a variety of scapolite, but allows the assemblage orthopyroxene + Ca,Mg-carbonate to be stable with respect to forsterite + clinopyroxene + CO_2 , and allows calcium carbonate-silicate minerals to be stable, as typified by the assemblage spurrite + rankinite with respect to larnite + CO_2 . Furthermore, carbon might be stored inside Venus as graphite or diamond in the mantle, or perhaps as a solid solution with iron in the core.

Evaluation of these possibilities will require very detailed speculation on the availability of suitable rock compositions. Especially difficult will be evaluation of the pressure–temperature–time history of Venus and the extent of geochemical cycling between atmosphere, crust, and mantle. At this stage of thinking about Venus it is simplest to assume that essentially all the CO_2 is in the atmosphere where it arrived by volcanic action, which stripped the mantle clean. Since the amount of CO_2 in the Venusian atmosphere is similar to that in the outer zones of the Earth, it would be surprising from cosmochemical arguments if large amounts of CO_2 were hidden below the surface of Venus. If the surface temperature of Venus were to drop below c.680 K, Mg-silicates should begin to react with CO_2 with the ultimate possibility of a thin atmosphere with no greenhouse heating.

The height distribution of CO in the Venusian upper atmosphere was modelled by photochemical reactions and vertical mixing (Sze and McElroy, 1975). If the CO/CO_2 ratio of the lower atmosphere were shown to depend on reaction with surface minerals, or on volcanic action, there might be important implications for the oxidation state either at the surface or inside Venus.

The partial pressure of c.0.1 bar H_2O in the lower Venus atmosphere might be explained either by input from continuing volcanism and cometary and meteoroid debris, or by buffering from surface minerals, or by some hybrid non-equilibrium state.

Although there is a psychological tendency to think that H_2O has left the Venus interior along with the CO_2 and that it must have continued to move outwards rather than returning, possible storage in the interior should be considered. The present content of H_2O in Venus is hard to estimate though an upper limit can be calculated fairly accurately. Even more uncertain is the primordial content because of the difficulty of quantifying the potential loss through the atmosphere and ionosphere.

All known minerals containing H_2O molecules are unstable on Venus, and many hydroxylated ones are unstable. Of the major hydroxylated minerals, only amphibole and mica could be important stores (Mueller, 1964). For peridotitic rocks, amphibole (of appropriate composition) could be stable to 15–20 kb for low Fe/Mg (cf. figs. 2 and 8), but to lower pressures as Fe/Mg increases. Phlogopite-rich mica could be stable to a higher pressure whose limit is difficult to ascertain because of the near-asymptotic approach of a prete to the curve for breakdown (fig. 8). On Earth, amphibole may occur down to c.100 km and phlogopite to c.150 km, and probably these depths are not exceeded on Venus. Of course, even mica and amphibole may not occur on Venus because volcanic action may have driven the water into the atmosphere, where it was lost through the ionosphere before it could be reacted at the surface and subducted downwards.

It turns out that mica can be a significant store even if Venus has no more K than Earth, as is required by cosmochemical models. Paragonite and margarite cannot be significant contributors, and all micas must be K-rich with $\text{K}/\text{H}_2\text{O}$ c.2. If Venus contains 130 ppm K, mica could store c.60 ppm H_2O , which is about one-tenth the amount of H_2O estimated for Earth (Smith, 1977). Of course, much or all of the K on Venus might be tied up in feldspar, thereby reducing the above estimate. The maximum amount of amphibole can be estimated by placing a stability limit of c.60 km, corresponding to c.400 ppm H_2O in Venus. Thus the outside limit of c.500 ppm H_2O is similar to the estimated c.600 ppm for Earth, but many arguments can be invented to suggest that Venus has much less than 500 ppm H_2O .

Loss of water from Venus depends on volcanic emission and transport upwards in the atmosphere. Models for generation and loss of water are discussed in detail by Mueller (1964, 1970), Fricker and Reynolds (1968), Palm (1969), Rasool and DeBergh (1970), Turekian and Clark (1975), Prinn (1975), Walker (1975), and Sill (1976). Loss of water from Venus cannot be estimated *a priori* because it depends on complex diffusion-controlled processes. Uncertainty in the transport mechanism in

the atmosphere, especially in relation to a sulphuric acid barrier, is particularly severe.

Volatiles can be modelled by analogy with meteorites and C1 material has been used. The simplest reaction relates organic material, represented by CH_2 , and magnetite, Fe_3O_4 : $\text{CH}_2 + 3\text{Fe}_3\text{O}_4 \rightarrow \text{CO}_2 + \text{H}_2\text{O} + 9\text{FeO}$. This requires equal abundances of CO_2 and H_2O on Venus, except for H_2O lost through the atmosphere. If H_2O were lost *in toto* there is no irreversible change in the oxidation state, but the more likely prospect that H is lost more efficiently than O requires that the remaining O must be taken up by the crustal rocks. Loss only of H corresponds to oxidation of Fe^{2+} to Fe^{3+} producing some 10 km of magnetite or c.100 km of oxidized basalt. Sill (1976) suggested that volcanic gases on Venus were like those on Earth in containing roughly equal molar proportions of H_2O , CO_2 , and SO_2 , and he proposed the reaction $\text{SO}_2 + \text{H}_2\text{O} \rightarrow \text{SO}_3 + \text{H}_2\uparrow$. The SO_3 was incorporated into sulphate by reactions such as $\text{CaMgSi}_2\text{O}_6 + 2\text{SO}_3 \rightarrow \text{CaSO}_4 + \text{MgSO}_4 + 2\text{SiO}_2$. Complex models for the atmosphere have been developed with production of large amounts of O_2 in the early atmosphere (Walker, 1975) and use of the 'runaway greenhouse' mechanism (Rasool and DeBergh, 1970).

Mueller's (1965) investigation of the stability of sulphur compounds on Venus is partly obsolete because of revisions in the atmospheric composition, but the basic features remain intact (Sill, 1976). At the surface, anhydrite and the feldspathoids sulphate-scapolite (Goldsmith, 1976) and haüyne are potential sinks for S, and iron sulphides are unstable if the oxidation state is controlled by magnetite and hematite. Sill (1976) also considered the possibility of a reduced state for the Venusian surface in which troilite and perhaps pyrite are stable, thereby allowing for a significant partial pressure of S vapour. This S vapour might explain the yellow colour of the clouds. However, I prefer the concept of an oxidized surface, which could be tested experimentally by magnetic measurements (cf. Mariner experiments on Mars).

The small but non-zero mixing ratios of HCl and HF in the atmosphere require detailed explanation (Mueller, 1968; Sill, 1976). Several minerals including fluorite, mica, and amphibole offer potential sinks or buffers for F, while scapolite and cancrinite

could serve for Cl. Perhaps continuing volcanic activity is supplying HCl and HF to the atmosphere, and perhaps there are substantial amounts dissolved in H_2SO_4 in the clouds.

Concluding remarks

Major uncertainties remain about the mineralogy, petrology, and geochemistry of Venus, and it will be very difficult to obtain detailed mineralogical information about the surface because of the strain placed upon instrumental design by the high temperature. Perhaps magnetic and spectral reflectance studies will provide useful data, especially on the oxidation state. Very useful would be determination of the extent of volcanic activity. If radar studies of high spatial resolution indicate volcanic morphologies, it might ultimately be possible to fly a mass spectrometer over possible cones to test for emission of gases. In the meantime, mineralogists must rely mainly on further study of the composition and height distribution of the atmosphere, preferably by high-resolution mass spectrometry.

The thermal evolution of Venus may be complicated by capture of one or more moons (as invoked to explain the slow retrograde rotation: Singer, 1970; French, 1971; Ward and Reid, 1973). Because the planet is so large and because its K/U ratio is like that of Earth it is tempting to invoke the presence of plate tectonic processes continuing to the present time. If so, the volcanic gases should be providing information on the interior. Can it be argued that the paucity of water in the atmosphere results from a dry interior, or is water removed by surface reactions as fast as it is supplied? It is certain that low-grade metamorphic rocks cannot exist on Venus, and it is likely that the prete corresponds roughly to the boundary between medium and high grades of metamorphism as defined for terrestrial rocks (Winkler, 1974, Fig. 7—2). Does anatexis occur? Can most crustal rocks be loosely described as granulites? Whatever the answers, Venus must have much simpler mineralogy and petrology than Earth and many abundant minerals on Earth (including muscovite and gypsum) cannot occur on Venus, while others (including microcline) can be stable only at shallow depths.

MARS

And yonder all before us lie
Deserts of vast eternity.

Andrew Marvell, *To his coy Mistress*

The spectacular Mariner and Viking missions revealed the major features of the atmosphere and surface of the 'red' planet of war and death, but provided only tantalizing hints of the mineralogy, petrology, and chemistry of the sub-surface. At the time of first writing, debate was just beginning on the complex results from the Viking missions (V1, Chryse Planitia; V2, Utopia Planitia), and the present account is based largely on articles in *Science* (27 Aug., 1 Oct., 17 Dec. 1976). Reviews of the geology were given by Mutch and Saunders (1976) and Mutch *et al.* (1976); of the general properties of Mars by J. B. Pollack in *Scientific American*, Sept. 1975; and of the atmosphere by Moroz (1976). Implications for the evolution of the Earth from Martian observations were discussed by Rasool *et al.* (1977). The September issue of *J. Geophys. Res.* appeared during revision, as did an article on the abundance of volatiles on Earth and Mars (Anders and Owen, 1977).

Surface morphology

Bearing in mind the difficulties of interpretation of physiography and the lack of an absolute time scale for Martian stratigraphy, the following summary provides a temporary basis for mineralogical speculation.

The oldest units are the undivided cratered terrain, which covers rather more than one hemisphere centred on $c.55^{\circ}$ S, 325° W, and the undivided plains, which occupy the remaining antipodal fraction. Almost certainly the heavy cratering of the undivided cratered terrain results, by analogy with the lunar highlands, from the period earlier than $c. -4.0 \times 10^9$ yr. Lower abundance of craters in the undivided plains can be explained by obliteration of craters by extensive volcanic activity. Whereas most craters of the northern plains have sharp rims, those in the southern terrain have a range of sharpness suggesting considerably greater erosion for older craters. There is no obvious explanation for the asymmetry of the early Mars surface and its angular displacement from the present polar axis.

Huge near-circular structures, including the Argyre, Isidis, and Hellas basins (the largest with diameter 2000 km and depth 4 km) have a size and frequency that would allow matching with lunar and Mercurian impact basins (Wood and Head,

1976), probably providing a time marker near -4×10^9 yr.

Several types of plains have been distinguished, but it is uncertain whether they are surfaced by volcanic flows or impact debris.

Huge masses of volcanic rocks definitely poured out of Mars (Carr *et al.*, 1977b). The saucer-shaped Alba Patera, 1600 km across, may be the largest central-vent volcano in the solar system. Its very low profile is not matched by volcanic structures on Earth, Moon, or Mercury. The Tharsis and Elysium plateaux are capped by enormous shield volcanoes, typified by Olympus Mons whose cone (with a summit caldera 80 km across) rises 25 km above a scalloped escarpment 600 km across and 6 km high. Surrounding the cone is a ring of plains 100–200 km wide and an outer incomplete girdle of grooved terrain 200–700 km across. Perhaps the complex data on volcanic landforms will be fitted into an over-all sequence beginning with planet-wide volcanism in an era of heavy impacts and isostatic adjustment, and terminating with extrusion of lavas in uncompensated shield volcanoes that grew over stationary hot-spots in the mantle for long periods of unknown duration (10^8 to 10^9 yr?).

Although the circular structures are particularly eye-catching, the major volume of volcanic rocks probably occurs in inconspicuous plains. The morphology of most lava flows that can be recognized as individual units indicates a flow volume much greater than on Earth and a low viscosity. Flows near the summit of the Martian shield volcanoes are shorter than those at lower elevation, perhaps indicating higher viscosity of late lavas. Three types of lava channels (Masursky *et al.*, 1977) were distinguished from three types of channels attributed to flow of water.

These latter channels were classified into broad channels abutting areas of chaotic terrain (Carr *et al.*, 1977a, Fig. 7), possibly caused by catastrophic melting of frozen ground by subsurface heating; sinuous channels of widespread distribution attributed to underground melt-water produced from permafrost by climatic warming; and dendritic channels, which terminate abruptly on plains or crater walls, and which may result from rainfall. Carr and Schaber (1977) attributed the following features to extensive permafrost in the ancient cratered uplands: fretted terrain, chaotic terrain, alases, canyons, patterned ground, and unusual

morphology of impact craters (see also Carr *et al.*, 1977c). Although water, derived either from permafrost or rainfall, is a probable scouring agent for many Martian channels, rifting coupled with landslides and aeolian erosion may be responsible for other linear features.

The enormous region of Valles Marineris, which spans one-quarter of the equator, was attributed by Blasius *et al.* (1977) to extensional tectonic activity followed by a variety of erosional and depositional processes. Large blocks have shifted vertically and tilted with respect to each other. Massive landslides, perhaps lubricated by fluid, modified the valley shapes, and the whole set of processes probably lasted over 10^8 – 10^9 yr into recent times.

Huge dust clouds periodically obscure Mars as spring ends in the southern hemisphere. The layered terrain near 75° N and S consists of numerous uncratered terraced layers with an unconformity, all overlain by deposits of permanent ice (Cutts *et al.*, 1977). Its origin is not understood, but it may result from episodic deposition of dust at the edge of a mid-latitude storm belt that was affected by long-term climatic changes. Major dune fields adjacent to the layered terrain, and minor dunes at the Viking sites, testify to aeolian processes.

Thermal and albedo mapping revealed two major types of Martian surface material and a covering of fine particles on bright surfaces (Kieffer *et al.*, 1977b). Frey (1974) suggested that there were two distinct types of crustal material on Mars by simple analogy with Earth and Moon, but there are no specific chemical indications. The causes of albedo changes on Mars are not understood, and dust deposition may be only one factor.

Three types of aerosols were detected over the Viking landers (Pollack *et al.*, 1977): ground fog during late summer nights, equivalent to $2\text{ }\mu\text{m}$ of precipitable H_2O ; high-level ice clouds in the autumn; and mineral particles ($0.4\text{ }\mu\text{m}$, non-spherical, rough), whose imaginary refractive index fits well with $10 \pm 5\%$ of magnetite mixed with an unidentifiable material. Pollack *et al.* suggested that a coating of CO_2 would result in more rapid fall-out of mineral particles and associated H_2O -ice over the winter poles.

The permanent portions of the polar caps consist of water ice and dirt (Kieffer *et al.*, 1977a), and are augmented by small seasonal ice caps containing H_2O and CO_2 ices, which migrate between the poles by vapourization, atmospheric transport, and condensation. The thickness of the ice caps is unknown. Dzurism and Blasius (1975) discussed the effects of higher surface elevation and lower pressure of the southern cap, and suggested that the

caps are 1–2 km thick in the south and 4–6 km in the north (see also Farmer *et al.*, 1977).

Dating of surfaces from crater counts is highly controversial because of lack of calibration. Whereas Neukum and Wise (1976) argued for completion of the major volcanic episodes in the first 1.5×10^9 yr with Olympus Mons completed by 2.5×10^9 yr, Hartmann (1977) suggested a much longer volcanic history with Olympus Mons only 0.06 – 0.6×10^9 yr old, and the channel system formed at 3 to 0.3×10^9 yr. Whatever the absolute time scale the growth period of the volcanic shields on Mars is much longer than on Earth (Blasius, 1976). See also Chapman and Jones (1977).

Small craters have sharp rims suggesting low erosion in recent times (10^8 – 10^9 yr?) in spite of the dust storms. The great thickness of layered deposits near the poles and the high rate of obliteration of early craters were attributed to an early episode of strong erosion, perhaps associated with a thick atmosphere (Jones, 1974).

Topographically, the northern plains are lower by about 2 km than the southern cratered terrain. Mars is mostly at isostatic equilibrium, except for the Tharsis plateau whose positive anomaly may be related to the negative anomaly of the adjacent Chryse region by underlying mantle convection. There is no evidence of plate-tectonic processes, but major fractures probably mark incipient rifting.

Panoramic views at the Chryse and Utopia landing sites (Mutch *et al.*, 1977a, b) revealed boulder-strewn deserts. Greeley *et al.* (1977) interpreted the geology of Chryse Planitia in terms of a basement of heavily cratered ancient terrain covered first by deposits of aeolian, fluvial, and volcanic types, and more recently by deposits, probably of volcanic rocks and impact debris. Binder *et al.* (1977) attributed the view from the V1 lander to volcanic terrain exposed as bed-rock on some crests, upon which reside numerous blocks of several rock types, whose pitting indicates a grain size in the mm to cm range, and fine-grained sediment and duricrust. The view from the V2 lander is also dominated by blocks lying on a fine-grained surface. Mutch *et al.* (1977c) suggested that Utopia Planitia has volcanic domes, and that its sediments and dunes are part of the mantle extending northward from 35° latitude. The blocks have similar surfaces whose pits of mm–cm size were thought to represent vesicles rather than the weathering of soft crystals. A polygonal network was attributed to ice wedging of contraction cracks. Pebbles sampled by the X-ray fluorescence experiment have similar composition to soil except for extra S, and probably represent fragments of duricrust 1–2 cm thick. No chemical data were

obtained for the boulders and the surface homogeneity and even distribution argue against attribution to the ejecta blanket of nearby Mies crater. Fragmentation of lava flows followed by sedimentary and aeolian processes may explain the boulders.

Early attempts to understand the surface of Mars relied on its red colour, but Huck *et al.* (1977) discovered that the Martian surface is actually yellowish brown with local change to grey. Some VI rocks appear dark grey, but most soils and rocks range from dark-brown to strong yellowish-brown and even moderate olive-brown. The reflectance has a strong band centred at $c.0.93 \mu\text{m}$, which Huck *et al.* prefer to match with nontronite rather than limonite. Binder *et al.* (1977) prefer to retain the suggestion by Binder and Cruickshank that limonite is responsible, whereas Hargraves *et al.* (1977b) prefer maghemite ($\gamma\text{-Fe}_2\text{O}_3$). Whether the yellow-brown colour merely results from a surface patina on grey rocks (basaltic?) and clay minerals is not certain, and neither is it clear whether air-borne dust and aeolian sediments differ in colour.

Atmosphere and volatiles

The pressure at the surface ranges from about 5 to 10 mbar, decreasing with elevation and changing slightly with season and time of day. Mixing ratios from various techniques are CO_2 95%, N_2 2.7%, Ar 1.6%, O_2 0.13%, CO 0.07%, H_2O variable but $c.0.03\%$, Ne $c.2.5$ ppm, Kr 0.3 ppm, Xe 0.08 ppm, and O_3 0.03 ppm (Owen *et al.*, 1977). The water content of the atmosphere varies greatly, but, although sometimes high enough to produce local mists, amounts only to 1 km^3 for the whole planet.

Enrichment of ^{15}N over ^{14}N in the Martian atmosphere at $c.130 \text{ km}$ height ($^{15}\text{N}/^{14}\text{N}$ 0.0064) over Earth (0.0037) can be explained by differential loss of the lighter isotope from the upper atmosphere thereby suggesting a substantially higher concentration of nitrogen in earlier times (McElroy *et al.*, 1977b). Ratios for $^{18}\text{O}/^{16}\text{O}$ (0.0020) and $^{13}\text{C}/^{12}\text{C}$ (0.0115) match those for the Earth within experimental error, and imply large accessible reservoirs (at least equivalent to several bars of atmospheric pressure) of O and C for exchange with the atmosphere if the differential escape mechanism used for N_2 is not to produce detectable changes for O and C (Nier *et al.*, 1977).

Mixing ratios and isotopic ratios of the rare gases place important restraints on models for outgassing and atmospheric escape. Owen and Biemann (1977) found that $^{40}\text{Ar}/^{36}\text{Ar}$ is 2750 ± 500 , which is ten times greater than on Earth. Owen *et al.* (1977) detected Xe and Kr, and found that ^{129}Xe

is much more abundant than ^{132}Xe and ^{131}Xe . They found that non-radiogenic noble gases in the Martian atmosphere have the same relative abundance as in the terrestrial atmosphere, but with 100-fold lower abundance when scaled to planetary mass. The deficiency of Xe in the Earth's atmosphere with respect to chondritic abundance has been attributed to storage in shales and other sediments, and the probable deficiency in the Martian atmosphere may similarly be attributed to sedimentary reservoirs.

Some suggestions published before the Viking mission are now obsolete because of the erroneous assumption of twenty-fold higher abundance of Ar in the Martian atmosphere, but their general ideas are still useful (e.g. Fanale, 1976; Levine, 1976; Owen, 1976). Current speculations involve five features: a denser early atmosphere, perhaps dense enough (0.1 to 1 bar?) to allow heavy rainstorms; considerable exospheric loss of light gases to explain the $^{15}\text{N}/^{14}\text{N}$ ratio; reaction of C- and O-bearing species with crustal reservoirs; migration of much of the early atmosphere into the regolith, either by freezing (permafrost), adsorption, or chemical reaction; and only partial outgassing, especially of ^{40}Ar produced from the decay of ^{40}K . The climate and atmosphere may have changed considerably because of either change of the energy output of the Sun (Hartmann, 1974) or the obliquity and eccentricity of the Martian orbit (Ward, 1974), and these changes even may have been complicated by greenhouse-type mechanisms (Sagan *et al.*, 1973).

The photochemical reactions in the atmosphere have profound implications for the surface mineralogy. The following section is based on a lecture by D. M. Hunten in November 1977 in which he followed up ideas in Hunten (1974), McElroy and Kong (1976), Kong and McElroy (1977), and McElroy *et al.* (1977a).

The key is photolysis of H_2O in the atmosphere by solar ultraviolet radiation producing H and OH radicals. Various photochemical reactions, whose details are irrelevant here, then produce H_2O_2 and other species including HO_2 . The hydrogen peroxide freezes out under almost the same conditions as H_2O , while HO_2 condenses at a lower temperature. The estimated production of approximately 10^{17} molecules of H_2O_2 per cm^2 per year would suffice to explain the absence of organic molecules in the soil and the evolution of gases in the biologically oriented experiments on the Viking landers (Biemann *et al.*, 1977; Horowitz *et al.*, 1977; Oyama and Berdahl, 1977; Ponnampetuma *et al.*, 1977). Ironically this photochemical scheme results in water destroying the possibility of life on Mars rather than fostering life as on Earth!

Mineralogy and petrology

Weight restrictions limited Viking instrumentation to an energy-dispersive X-ray fluorescence spectrometer that could not detect elements lower than Mg. Furthermore, the poor energy resolution of the proportional counter (resulting from the impracticability of using a high-resolution crystal cooled to liquid N₂ temperature) caused strong overlap of contributions from elements adjacent in the Periodic Table. The early analyses of fines from both Viking sites (Clark *et al.*, 1976) were remarkably similar in having high Fe (13 ± 2 wt %), moderate Mg (5 ± 2.5), Ca (4 ± 0.8), and S (3 ± 0.5), low Al (3 ± 0.9) and K (< 0.25), and very low trace elements (e.g. Sr c.80 ppm; Y c.60 ppm; Rb < 30 ppm). Other elements are: Si 20.9 ± 2.5 ; Cl 0.7 ± 0.3 ; Ti 0.5 ± 0.2 . Computer modelling led to a chemical mixture of 81 % Fe-rich clay (59 % nontronite; 22 % montmorillonite), 12 % Mg-sulphate (kieserite), 6 % calcite, 1 % halite, and 0.5 % rutile. Some 1 to 7 % of the fines became attached to magnets (Hargraves *et al.*, 1977b), thereby indicating some type of magnetic iron oxide. Although details are uncertain the analyses were attributed to the weathering products of mafic igneous rocks (Baird *et al.*, 1977). Unpublished analyses of 2–12 mm fragments at V2 yielded 25–50 % more S than for fines, but no sympathetic increase of Ca, Al, and Fe. A sample of fines, protected from aeolian modification by burial under a rock, gave 10–15 % less Fe than the earlier fines. The correction procedure for the analyses assumes that the material is spatially uniform. A thick skin of iron oxide coating silicate material would yield falsely high Fe and *over-all* sum of calculated oxides, and Clark *et al.* (1977) placed an upper limit of 0.24 μ m on such a skin. Baird *et al.* (1977) concluded that the Martian surface has an abundance of mafic parents for subsequent clay formation, a lack of large exposed alkali-rich granitic materials, and a sufficiency of water and ice to form clays. Schonfeld (1977, oral presentation, Lunar Science Conference) argued that the analyses can be well represented by a mixture of C1 meteorite and basalt, but details have not appeared in print.

Using a gas-chromatograph-mass-spectrometer, Biemann *et al.* (1977) did not detect organic molecules at the 10^{-3} ppm level, and observed release only of H₂O and CO₂ upon heating fines to 50°, 200°, 350°, and 500°C for 30 seconds. From 0.1 to 1 wt % H₂O and 50–700 ppm CO₂ were released, but identification of the mineral hosts is unclear because of uncertainty in the kinetics. Perhaps the water was released from hydroxylated silicate or hydrated sulphates or both, while the CO₂ might have derived from carbonates less stable than

calcite (e.g. magnesite or siderite). Absence of S-bearing gases rules out presence of weakly bonded sulphides, and the most likely hosts for the sulphur are strongly bonded sulphates not containing substantial Al, Ca, or Fe.

Revision of the colour of the Martian surface from reddish to yellow-brown changed the mineralogical implications. Instead of favouring a thin film ($< 0.25 \mu$ m) of hematite on surfaces, the colour favours maghemite (γ -Fe₂O₃). Hargraves *et al.* (1977a) had preferred to interpret the magnetic material at the Viking sites as magnetite (possibly titanian to explain the 0.5 ± 0.2 % Ti in the XRF analyses), but later (1977b) switched their preference to maghemite.

Toulmin *et al.* (1977) pointed out that nontronite breaks down to maghemite and siliceous material under oxidizing or inert atmospheres, and that maghemite does not transform to hematite below 800 °C. Huguenin (1976 and earlier references) showed experimentally that ultraviolet radiation produced photo-stimulated oxidation at the surface of magnetite and Fe-bearing silicates under dry conditions in the presence of O₂ gas. In the preceding section the photochemical conversion of H₂O into H₂O₂ was described, and reaction with hydrogen peroxide (and perhaps other 'super-oxides') might be more important than direct photo-stimulated oxidation at a mineral surface. Furthermore, burial and later exhumation of sediments might result in complex low-grade reactions dominated by surface reactions, perhaps of a catalytic nature.

Spectral data for dust clouds need re-examination. Aronson and Emslie (1975) interpreted infrared data for Martian dust in terms of a strongly polymerized silicate, preferably feldspar and possibly a layered mineral, and ruled out goethite from absence of 800 and 900 cm⁻¹ bands. Perhaps the suggestion by Pollack *et al.* (1977) that magnetite is involved may need revision: maghemite might also be a suitable candidate because of its similar crystal-chemical properties.

The S content of XRF analyses of Viking fines can be interpreted in terms of hydrated Mg-sulphate(s) and the chlorine content (0.7 ± 0.3) may result from salt(s). Malin (1974) suggested that salt weathering, which is believed to occur in Antarctic dry valleys, may occur on Mars. The salts may derive from volcanic gases, and the pressure and temperature, especially for equatorial latitudes, are consistent with temporary formation of brine (Ingersoll, 1970). Malin also pointed out that some quartz diorites from Antarctica had weathered to brown red.

Whether melting of the clathrate CO₂.6H₂O is responsible for floods is disputed (Milton, 1974;

Peale *et al.*, 1975) but the clathrate was predicted to occur in association with water ice in polar regions with temperature below *c.*153 K (Miller and Smythe, 1970; Dobrovolskis and Ingersoll, 1975). The conditions which govern the melting or evaporation of water on Mars are complex and apparently strongly dependent on a dust cover (Farmer, 1976).

Temporarily, the simplest synthesis of the clues to the mineralogy of Mars is that the dominant extrusive rocks were Fe-rich 'basalts', thereby explaining the low Al in the XRF analyses, and the apparently low viscosity indicated by flow morphology; that the weathering of the 'basalts' occurred in a hydrous, oxidizing environment producing Fe-rich clay minerals (perhaps mainly nontronite) coated with maghemite and perhaps other iron oxides or hydroxides; that S- and C-bearing gases were incorporated into sulphates and carbonates (probably mainly of Fe and Mg); and that various species, especially H₂O and CO₂, did not react to form minerals, but formed ices, which temporarily melted into brines or even evaporated. Perhaps most of the material analysed by XRF ultimately derived from dust clouds of weathered volcanic rocks.

Geophysical and geochemical models

Unfortunately the seismic experiment on the Viking missions has not recorded any event strong enough to provide constraints on the Martian interior, and geophysical modelling is based merely on the density and the moment-of-inertia. Revision of the latter from $C/MR^2 = 0.377 \pm 0.001$ to 0.365 ± 0.001 (Reasenbergs, 1977) resulted in major transfer of mass towards the planetary centre, but the history of the lunar moment-of-inertia does not inspire confidence in the latest Martian value. Cosmochemical models for the planets are also highly uncertain, and present data on the volatiles on Mars are perhaps more easily interpreted in terms of a volatile-poor than a volatile-rich planet (Anders and Owen, 1977). Nevertheless the early models for Mars are important for construction of better models.

Simple models for condensation of the solar nebula lead to proposals that Mars should have a full complement of 'refractory' and 'silicate' groups of elements, a higher content of S and oxidized iron than Earth, and a greater content of volatile elements. Specifically, Lewis (1972*a*) proposed that Mars has a core dominated by FeS, perhaps with a little Fe metal, and a mantle with Fe/Mg *c.*1. He also proposed that Mars accreted well within the condensation field of tremolite, and estimated one H₂O for every 2Ca (i.e. H₂O *c.*0.3 wt% of bulk

Mars). Anders and Owen (1977) concluded that Mars has only 3% of the Earth's complement of volatiles (proportioned to planetary mass), and obtained them from the same source. This apparently rules out the Lewis model for Mars. Perhaps the volatiles are derived from late impacting planetesimals that condensed in the outer part of the solar system, and perhaps the lower escape velocity of Mars resulted in retention of only a few per cent of such material. If so, Mars may have accreted mostly from bodies formed from the solar nebula at temperatures higher than that for condensation of tremolite. Perhaps the non-volatile composition of Mars is quite similar to that of Earth as befits the similar values for uncompressed density, but perhaps the degree of oxidation and sulphidization is higher by analogy with the simple Lewis model. It is also possible to modify the ratio of refractory condensate to Mg-silicate and metal fractions between the two planets, as in the Ganapathy-Anders model for the Earth. With these factors in mind the existing models can be reviewed.

Using the old moment-of-inertia, Anderson (1972) modelled Mars with several mixtures of meteorites thereby ending up with substantial S in the core. Varying the core composition from pure Fe to pure FeS resulted in the core radius changing from 1220 to 2040 km. The zero-pressure density of the mantle was limited to 3.54–3.49 g/cm³, implying 24–21 wt% FeO unless some Fe metal remains in the mantle. For all chondritic models Mars has a smaller and less dense core than the Earth and a denser mantle. The bulk Fe content is 25–8 wt%.

Binder and Davis (1973), following Binder (1969), used new physical data for Mars, together with the possibility of S in the core, to place revised limits on the core and mantle. Particularly important is the use of an isostatically compensated equatorial bulge of 8 km to yield a crust (assumed 2.8 g/cm³) of thickness ranging from 24 to 65 km depending on the mantle density and the crustal thickness at the poles. Many complex possibilities are allowed, but for a core density near 7.1 g/cm³ (corresponding to about 30% FeS in the core), the core is about 1250 km in radius and contains *c.*10% of the mass while the mantle has olivine composition (Fo_{65–80}). If the crust is not fully compensated isostatically there are greater tolerances.

Earlier, Ringwood and Clark (1971) investigated the consequences of Ringwood's model for derivation of Mars from C1-meteoritic material, but more oxidized than Earth and ordinary chondrites. Using a simplified C1 composition that ignored all volatiles and S, they considered three mantle-core models: (i) all Fe in magnetite and all silicates Fe-free; (ii) half Fe as FeO incorporated in olivine and pyroxene, and half as magnetite; and (iii) most

Fe as FeO and little as Fe₂O₃. All models could be made to yield the observed density, but all had too high a moment-of-inertia even for the old value. In order to decrease the moment-of-inertia a crust of density 3.0 g/cm³ was added. Finally the probable presence of FeS was considered. Ultimately the following complex model IIB was proposed: depth 0–c.50 km crust; 50–1100 upper mantle of olivine and pyroxene; 1100–1650 lower mantle of spinel and garnet phases of olivine and pyroxene composition; 1650–1850 outer core of molten Fe–S–O; 1850–3388 inner core of Ca-ferrite, a high-pressure form of magnetite composition.

Johnston *et al.* (1974) set the internal structure of Mars in the context of the possible thermal state. Geophysical parameters were chosen in order to obtain an early differentiation producing a core, mantle, and crust, a prolonged partial melting of the mantle producing extensive volcanism, and a surface rigidification to 200 km depth in order to support the present non-hydrostatic shape. They did not rule out present melting of the core, as had Binder and Davis, because of new evidence for a weak dipole magnetic field (but see Dolginov, 1978, for controversy with C. T. Russell). In order to obtain early melting of Mars (before -4×10^9 yr), it was necessary to assume, even for the low-melting Fe–FeS eutectic, either a high accretion temperature, or a high content of radioactive elements, or both. Johnson *et al.* assumed 0.031 ppm U, which is over twice the value expected for the Lewis model of Mars. For rapid accretion (10^5 yr) of material at 273 K, core formation occupied the period c. -4.2 to -3.1×10^9 yr, and melting of dry silicate began at -1.3×10^9 yr at 600 km depth. For material accreted at 773 K, core formation began almost immediately and was completed at -3.6×10^9 yr while dry silicate began to melt at -2.6×10^9 yr. As discussed earlier, there is a general problem of finding heat sources for the early melting of planets, and there are severe problems in attempting to model early core formation on Mars if the accreted material is low in U (c.0.013 ppm) and if it accreted slowly at a low temperature. It seems very difficult or perhaps impossible to explain the surface morphology of Mars and its present geophysical properties without invoking early differentiation, and Solomon (1978a) argues that the surface morphology of Mars requires an originally cold interior coupled with restriction of early differentiation to the outer part. If this is accepted the analysis by Johnston *et al.* must be a good guide to the thermal development of Mars, except that there may be a different distribution of heat sources at the beginning, and a smaller production of radioactive heat. Whatever the details, one can imagine an early differentiation involving separation of a

heavy Fe–FeS eutectic or near-eutectic liquid and a light hydrous basaltic liquid, and a later differentiation involving partial melting at higher temperature of a mantle of dry peridotite.

Johnston *et al.* modelled Mars with a 50 km crust of density 3.0 g/cm³, a mantle of pyrolite with added FeO to bring the density up to 3.74 g/cm³, and a core with 85 wt % Fe and 15 % S. This results in a mantle with 73 % olivine (65 mol % Fo), 25 % garnet (grossular₅₃ almandine₁₉ pyrope₂₈; wt %) and 2 % Fe,Mg-oxide. The mantle composition is SiO₂ 34.6 Al₂O₃ 2.7 Fe₂O₃ 0.3 FeO 29.5 MgO 28.7 CaO 2.4 K₂O 0.1 wt %. Use of a revised moment-of-inertia by Johnston and Toksöz (1977) resulted in movement of Fe from the olivine (now 75 mol % Fo) into the core (1500–2000 km radius; more Fe-rich than the Fe–FeS eutectic) giving an estimated bulk composition close to that in Table IV.

It is obvious that there is considerable tolerance in models for Mars, and that the key factors are the assumed concentrations of S and O in the core, and the degree of differentiation of Fe between the mantle and core. Uncertainties in the pressure–temperature distribution are not as important. Because of the high S content of the fines at the Viking site, I prefer to assume that the volcanic gases and volcanic rocks derived from sources containing substantial S, and hence I prefer models of Mars in which S is a major element, perhaps largely concentrated in a core but with some remaining in a silicate-rich mantle and crust. Recent preprints by Okal and Anderson (1978), McGetchin and Smyth (1978), and Solomon (1978a, b) extend the above discussion.

Siever (1974) compared the differentiation of Mars and Earth with particular emphasis on geochemical transport. He pointed out that weathering does not occur simply because a mineral is brought to low temperature, but requires reaction with water and dissolved acids. Reactions in feldspar are governed by diffusion through an alteration rim, and it is difficult to estimate *a priori* the rate of alteration on Mars. He suggested that Huguenin's (1976) mechanism of photo-stimulated oxidation of magnetite might also be ultimately controlled by diffusion. Huguenin suggested that surface scaling would greatly enhance the rate of weathering. Booth and Kieffer (1978) concluded that carbonate minerals can form even under arid conditions. There are many complex problems, and it seems impossible to make detailed predictions on the mechanisms and extent of weathering on Mars, especially in view of the relation to photochemical processes in the atmosphere. Siever emphasized that there is no evidence for vertical motions on Mars large enough to allow recycling of sediments followed by volcanic activity. Unlike Earth

sediments, all the Martian sediments should be still at the surface or covered only by other sediments or lava flows. Volcanic emissions from Mars should enter the atmosphere, react with surface minerals, and remain mostly trapped there forever. Wind transport should redistribute the altered minerals from one place to another, and volcanic and climatic heating should cause local melting of frost, and perhaps some devolatilization of weakly bonded minerals, but there should be no major geochemical cycles as on Earth.

For the atmosphere Siever mentioned thermal escape of hydrogen and helium, non-thermal escape of oxygen (Liu and Donahue, 1976), and absence of S- and Cl-species. He suggested that most S occurs on Mars as solid sulphide, and pointed to the stability of pyrite in terrestrial Precambrian sandstones before bacterial oxidation became important. However, the dissociation of H_2O in the Martian atmosphere (McElroy and Kong, 1976) and at mineral surfaces (Huguenin, 1976) should lead to more effective inorganic oxidation of sulphide than on Earth, and indeed the Viking experiments are more easily interpreted in terms of sulphate than sulphide. Siever suggested that Cl is removed from the atmosphere by freezing out with CO_2 and H_2O ice, and doubted the existence of evaporite sediments because of the ephemeral nature of liquid water on Mars.

Because of the absence of subduction on Mars, Siever suggested that extrusive rocks would become drier with time, beginning with early wet basalts of low viscosity and ending with dry basalts of higher viscosity.

Perhaps partial melting on Mars might be complex. For a hydrated silicate composition with high Mg/Fe, partial melting (fig. 2) might begin below 1000 K at a pressure of 10–20 kb. Uprise of the magma might remove the radioactive elements making it difficult for the remainder to become heated high enough for further partial melting. The key to the outcome is how small a degree of partial melting is required before (diapiric?) separation begins. Presence of FeS and Fe metal would cause further complications, as would a high Fe/Mg ratio in the silicate fraction. There is a rich field here for study. Whatever the outcome Fe-rich lavas are to be expected on Mars, as is consistent with the high Fe/Al ratio in the Viking XRF analyses.

Water- and CO_2 -rich basaltic liquids might retain most of their volatiles in crystallized minerals (e.g. amphibole and calcite) rather than releasing them to the atmosphere. Any ash deposits with substantial glass might alter to clay minerals or zeolites. There is no morphological evidence on Mars for the extrusion or intrusion of viscous

rhyolitic lavas like those on Earth, but ferro-rhyolites might have low enough viscosity to match high-level flows on Martian volcanoes.

With this geochemical and geophysical background it is not surprising that the inorganic analyses of fines at the Viking sites appear to be dominated by weathered Fe-rich silicate and sulphate. Plausible mechanisms can be invented for termination of volcanism on Mars some 10^8 – 10^9 yr ago, especially if Mars melted almost completely before -4×10^9 yr, and it may be possible for Mars to have incompletely differentiated if H_2O - and CO_2 -rich partial melts removed the radioactive elements from the interior. Finally, it is quite possible to envisage mineralogical traps for volatiles, thereby allowing a thin atmosphere even for bulk Martian compositions rich in K, H_2O , CO_2 , N_2 , and inert gases (Rasool and Le Sergeant, 1977; Fanale and Cannon, 1978; Fanale *et al.*, 1978). Unfortunately it will prove very difficult to place mineralogical limits on the storage capacity for the volatiles, and hence to decide what types of material contributed to the accretion of Mars. Currently it is probably easier to model Mars as a volatile-poor planet (Anders and Owen, 1977), and to explain the mineralogy in terms of a much lower content of H_2O than the one H_2O for every two Ca proposed by Lewis (1972).

Future mineralogical explanation of Mars should bear in mind the complex weathering processes, recently considered from the thermodynamic viewpoint by Gooding (1978). Determination of the chemical composition of unweathered volcanic blocks, preferably of a range of ages, would be more valuable than study of the fine-grained material. If possible, the surface layer should be spalled off to yield access to unaltered interior. Return of a soil sample to Earth would result in much valuable information, but every effort should be made to design a mission to recover unaltered volcanic rock, perhaps from the rim of a young impact crater, or perhaps from a drill hole.

Phobos and Deimos

The mineralogical nature of these two tiny irregular moons is unknown. The low albedo and heavy cratering may indicate a regolith (Zellner and Capen, 1974; Veverka, 1974; Veverka and Duxbury, 1977). If spectral data and the density ($c.2 \text{ g/cm}^3$) correctly indicate a composition like a carbonaceous chondrite, these moons are unlikely to be the last survivors of a debris cloud formed by fission or disintegrative capture, and it is perhaps more profitable to speculate that they are captured C-type asteroids.

COMETS, ASTEROIDS, AND METEORITES

There's not the smallest orb which thou behold'st
 But in his motion like an angel sings,
 Still quiring to the young-eyed cherubins.

William Shakespeare, *The Merchant of Venice*, V. i. 10.

Planets underwent so much differentiation that it is difficult to 'see' back to the solar nebula. Meteorites yield crucial information on the early processes of differentiation, but uncertainty about their original locations hinders speculation about chemical differentiation of the condensing nebula. Spectral studies on asteroids (often called minor planets) are opening up new vistas, but interpretations and speculations are controversial. Little is known about the mineralogy of comets. Extremely important for understanding the mineralogy of the planets will be landing and fly-by missions to asteroids and comets, though technical problems will be severe. Until then, most speculations about the inter-related origin of asteroids, comets, and meteorites will be difficult to evaluate. A book *Comets, Asteroids, Meteorites, Interrelations, Evolutions and Origins* (Delsemme, 1977), was in press when this section was written.

Comets and interplanetary dust

Differentiation of the solar nebula probably involves ejection of much of the mass as either gas and dust propelled by the solar wind, or as planetesimals perturbed by the giant planets. [Note. Small particles could spiral towards the Sun by the Poynting-Robertson momentum process.] Other stars would be ejecting material, while condensation would be taking place in the inter-stellar regions. Ejected planetesimals would undergo several types of orbital perturbations, and it is qualitatively easy to envisage the Sun being surrounded by a huge cloud of heterogeneous bodies with orbits ranging from elliptical to parabolic, or even hyperbolic. This Oort cloud has been estimated to consist of 10^6 to 10^{11} bodies, but only those bodies are visible that approach the Sun closely, and have enough volatiles to form a visible tail. Comets with small elliptical orbits, produced by planetary perturbations, lose volatiles during each close approach to the Sun, and should ultimately burn out and become asteroids if they have a coherent residue. From the mineralogic viewpoint the only 'direct' evidence is that comets have several types of tails. Those with linear streamers give spectra of ionized molecules, including CO^+ , N_2^+ , CO_2^+ , OH^+ , and ionized atoms. Diffuse tails are dominated by the reflection of sunlight, and must consist of dust

grains. Whipple's model of a 'dirty snowball' is generally accepted. The proto-comets would accrete dust particles and ices in the outer solar system, and some might scavenge gas and condensates in inter-stellar space. Upon approaching the Sun the ices would become unstable under sunlight.

Dispersed particles are the cause of meteor showers, and interplanetary dust particles may result dominantly from short-period comets and only slightly from collisional debris from asteroids. Lunar surface rocks are pitted by micrometeorite craters, but the high impact velocity results in major loss of material. Readers should consult the *Proceedings of the Lunar Sciences Conference* for numerous papers, and should note that lunar soil contains about 1% of C1 composition attributable to meteorite accretion.

Perhaps the best indication of cometary material is in the fine particles settling into the Earth's stratosphere (Brownlee *et al.*, 1976, 1977). Theoretical calculations for spherical particles arriving at 15 km/s indicate that ones greater than $10\text{ }\mu\text{m}$ are heated to $1000\text{ }^\circ\text{C}$ or higher for a few seconds whereas ones below $2\text{ }\mu\text{m}$ are hardly affected. Particles collected at 20 km height consist of five basic types, namely 'chondritic', Ni-bearing Fe-sulphide, Fe-poor olivine and pyroxene, Ni-Fe metal, and others; 60% of the 5– $10\text{ }\mu\text{m}$ particles are 'chondritic' and 30% are sulphide, whereas particles $>30\text{ }\mu\text{m}$ are dominated by olivine and pyroxene. The particles are more porous and finer-grained than those in C1 and C2 meteorites, and commonly are aggregates of grains in the 10–500 nm range. Electron microprobe analyses of thirteen micrometeorites (3– $30\text{ }\mu\text{m}$) have a two- to ten-fold range of major elements, with a general similarity to those of C2 meteorites but with slightly lower Fe, Mg, Ca, Al, Ni, Cr, and Ti and higher S. Some chondritic aggregates have C and S $>4\text{ wt}\%$, and some contain magnetite and Ni-pyrrhotite as well as Fe-poor olivine and pyroxene, and Ca, Al, Ti-rich material. Together with tentatively identified hydrated silicates in some particles the assemblage is strongly reminiscent of carbonaceous meteorites.

Deep-sea sediments (Blanchard *et al.*, 1978) contain numerous spheres of three types. Iron-rich spheres, which contain magnetite accompanied by varying amounts of wüstite, hematite, taenite, and trevorite, were identified a century ago, and are

attributed to ablation of iron bodies because of identical mineralogy with fusion crusts of iron meteorites. Silicate spheres have etched surfaces revealing euhedral olivine and magnetite crystals, but the interiors have surviving glass of various compositions. Pentlandite, Ni-rich troilite, and chromite also occur. Although most of the silicate spheres melted completely and then recrystallized into various textures, including the barred-chondrule type, some contain surviving grains of forsterite, enstatite, Fe-spinel, chromite, pentlandite, and Ni-Fe. Bulk electron microprobe analyses of the least-etched spheres showed a range of compositions (SiO_2 31–52; Al_2O_3 1–3; FeO 18–45; MgO 18–27; CaO 1–3; NiO 1–3) generally describable as chondritic except for low Ni. This second group of spheres is attributed to ablation of a wide range of stony bodies. The third group of glassy spheres with skeletal magnetite may derive from ablation of metal-rich silicate bodies. Rarity of sulphide may result from preferential loss by vapourization as meteoroids hit the Earth's atmosphere.

Perhaps sophisticated mineralogical and microscopic studies will reveal details of crystallization and irradiation histories, and allow definitive conclusions about the ultimate source of the interplanetary grains. There is an enormous literature on the composition of interstellar grains (to be located in astrophysics journals: see obsolete book edited by Field and Cameron, 1973). Bearing in mind the observational problems coupled with theoretical uncertainties (e.g. on light scattering from silicate grains coated with ices; competition between spluttering and condensation), it seems reasonable to interpret the current observations on interstellar grains in terms of material similar to or identical with the interplanetary grains, and to relate both to fine-grained material found in meteorites.

Asteroids

General description. The asteroids are difficult to describe succinctly because of the diversity of orbital elements and surface properties. They can be separated into three main groups: the Trojan asteroids, which lie within 20° of the Lagrangian stability points in the orbit of Jupiter; the Earth-crossing and Mars-crossing asteroids; and the main-belt asteroids whose orbits lie between Mars and Jupiter.

Trojan asteroids are too remote for easy spectral study from Earth. Their low albedo would not be inconsistent with theoretical expectations that they are primitive planetesimals containing dark minerals such as magnetite, troilite, and carbon. Published spectra are unlike any from the main-belt asteroids, and Chapman (1976) stated that they

do not match with any known meteorites, including carbonaceous ones.

Some thousands of main-belt asteroids are known, of which the largest are Ceres (diameter 1000 km), Pallas (610), and Vesta (540). Their orbits have a mean eccentricity of 0.15 and a mean inclination of 10° , which are greater than for the planets and less than for comets. Their mean distances from the Sun range from 2.0 to 3.3 AU (compared with Mars 1.5 and Jupiter 5.2). Asteroids are erratically distributed over the main belt, but are absent or rare at or near the Kirkwood gaps, which correspond to orbital resonance with Jupiter. The present root-mean-square collision velocity of 5 km/s in the asteroid belt is some ten times the escape velocity of several hundred m/s for Ceres and Vesta, and much greater than the escape velocity of the smaller asteroids. Many known asteroids with similar orbits can be grouped into Hirayama families.

Dynamical interpretation is difficult, but the ideas of several workers were brought together and developed by Chapman (1976), Shoemaker (1977a), and Wetherill (1977). Much more mass was originally present in the main belt (perhaps $c.10^{26}$ – 10^{27} g compared to a present estimate of $c.10^{24}$ g), but perturbations from Jupiter increased the encounter velocity and orbital eccentricity so much that planetesimals smashed each other to pieces instead of accreting into a planet. Most debris was removed into the inner and outer parts of the solar system by various mechanisms. The present population consists of a few fortunate survivors, which are being degraded by collisions, some of which are sufficiently catastrophic to yield large fragments belonging to Hirayama families. Material near the Kirkwood gaps (especially the 2:1 gap) is being perturbed into eccentric orbits, which ultimately approach Mars. Further perturbations can lead to capture by Mars, or even deflection into orbits that approach the inner planets or the Sun. In early days the asteroid belt would have been bombarded by planetesimals and debris deflected from the outer (and inner) solar system, but calculations of the collisional frequency and consequences are very difficult.

All Mars- and Earth-crossing asteroids are small, probably with maximum diameter up to some tens of km across (e.g. Eros $13 \times 15 \times 36$ km). The Apollo group, with orbits intersecting the Earth's orbit, is of particular interest. Shoemaker (1977b) estimated that there are 10^3 Apollo asteroids with absolute visual magnitude of at least 18, and that there are 10^5 of size 100 m or greater. Only two dozen have been detected so far. Estimated impact velocities on Earth are 15 to 40 km/s with root-mean-square velocity 25 km/s, and the present production rate of

impact craters 10 km across should be about one per million years, in good agreement with observed values.

Some Earth- and Mars-crossing asteroids have elongated orbits similar to those of short-period comets, and might actually be cores of burnt-out comets. Are Earth-crossing asteroids derived primarily from comets or from perturbed fragments of main-belt asteroids, and are they the main sources of meteorites?

Earth-crossing debris less than c.200 m across is extremely difficult to detect, and the numbered main-belt population is incomplete for diameters smaller than 50 km. The recent determination of the orbit of a tiny asteroid between Saturn and Neptune suggests that there may be undiscovered belts between the giant planets.

Remote-sensing studies. Over the past decade reflectance spectrophotometry, optical polarimetry, and thermal radiometry have been developed for remote sensing of planetary surfaces, and the results have been checked by direct sampling of the lunar surface. Most studies of spectral reflectance of asteroids were made from c.0.3–1.1 μm , which wavelength range encompasses the visible region 0.4–0.7 μm . Unfortunately, the mineralogical interpretations are uncertain, but recently the available range was extended deeper into the infra-red by use of bigger telescopes fitted with new detectors. The narrow-band study (0.9–2.7 μm ; 28 cm^{-1} resolution) of Larson *et al.* (1976) on the Apollo asteroid 433 Eros points the way to utilization of the detailed spectral differences of rock-forming minerals in the infra-red. The present account is based mainly on the reviews by Chapman, Morrison, and Zellner (1975), Chapman (1976), Morrison (1977), and Gaffey and McCord (1977, 1978), but is too brief to do justice to the controversial details.

Asteroids rotate rapidly with a period of several hours, and the constancy of relative spectral reflectance and polarization except for Vesta (Degewij and Zellner, 1978) indicates that there is no major chemical variation from one side to the other, while the large variations of total flux for some asteroids indicate a highly irregular shape. Because of their low escape velocity, asteroids are being 'sand-blasted' by collisions, but some debris must remain on the surface (at least several mm) to explain optical and radar reflectances. Optical reflectances are interpreted in terms of light penetrating deeply into an unconsolidated coating of dust particles whose chemical composition is representative of the outer part, but not necessarily of the deep interior of the asteroid. The particles are 'fresh' and have not undergone the age-darkening that occurred for a mature lunar regolith.

The simplest classification of asteroids uses the geometric albedo. Variation of the polarization of reflected light with phase angle demonstrates that asteroidal surfaces are dusty, and yields an estimate of the albedo. The absolute brightness then yields an estimate of the scattering cross section, and hence the mean diameter, which can be checked from thermal infra-red measurements. Zellner *et al.* (1977a) obtained calibrations by reflection polarimetry of lunar soils and meteorites, and were able to match asteroidal polarization data only with samples in which large particles were completely covered by fine dust. Perhaps electrostatic forces are important, as for lunar fines. Asteroidal albedos are strongly bimodal, and were classified into C and S types with albedos near 0.035 and 0.15, respectively. The albedos of the very dark C type were lower than those of lunar fines and meteorites, though carbonaceous meteorites were only slightly brighter (albedo c.0.05–0.07). Artificial mixtures of Mg-silicates with 10% carbon were similar in darkness to C types, and Zellner *et al.* (1977b) suggested that the interplanetary material studied by Brownlee *et al.* (1977) might be related to C asteroids, thereby implying a 'supercarbonaceous chondrite' composition.

Spectrophotometry in the visible and near infra-red regions offers the best clues to the mineralogy. Laboratory experiments with mineral mixtures showed that the spectral signature is relatively unaffected by grain size so long as the grains have similar reflectances, and the individual grains have low absorbance. Gross dissimilarity of reflectance, as between silicate and metal, causes serious problems for interpretation of spectra from mixtures, as does large grain size (c.1 cm) of silicates. Telescopic studies of reflectance of the Moon have shown good correlation with mineralogical identification of returned fines, and for the time being it is easiest to assume that asteroidal surfaces are covered by well-mixed fine material.

Two approaches have been used for interpretation of spectral signatures from asteroids: direct comparison with laboratory spectra for powdered meteorites measured by Gaffey (1976), and attempted 'reconstruction' of the observed spectrum from those for individual minerals that are abundant in meteorites (Adams, 1975).

The first approach (numerous studies reviewed by Gaffey and McCord, 1977) was particularly successful for the asteroid Vesta, whose spectrum matches those for basaltic achondrites, especially with respect to absorption bands at 0.92 μm and 1.9 μm indicative of pyroxene with 10–12 mole% Ca. Quite good matches were obtained between other meteorites and asteroids (e.g. Chapman, 1976, Fig. 1), but many asteroidal spectra could

not be matched. Indeed Gaffey and McCord (1978) questioned the validity of some of the earlier apparent matches, and emphasized the desirability of the reconstruction procedure.

Briefly, the matching approach led Chapman (1976) to conclude that the darker asteroids (C type) may be subdivided into ones like C2 meteorites and ones (called C*) perhaps similar to the metamorphosed C4 Karoonda meteorite; that the redder and brighter asteroids (S type) mostly cannot be matched with meteorites, but may consist of a mixture of silicates and iron in the ratio 2:1 to 1:4, probably closer to stony-iron meteorites than ordinary chondrites; that Vesta matches with basaltic achondrites, perhaps most closely with eucrites; that a few asteroids (M type) match enstatite chondrites, or just possibly iron meteorites or enstatite-iron mixtures; that approximately forty spectral sub-types can be tentatively recognized (e.g. the fourteen listed in his Fig. 4); and that most meteorite types are not matched by the presently studied asteroids.

Gaffey and McCord (1978) were unhappy with Chapman's use of the words Carbonaceous and Siliceous or Stony-iron as the sources of the designations C and S, because the important spectral features of a C type might match with a black chondrite or carbon-poor C4 meteorite and not necessarily with a carbon-rich meteorite. Nevertheless they agreed that the C-S classification was useful for reconnaissance, in which a simple measurement of colour or polarization is made.

The important spectral features for common minerals in the solar system are: Ices have high albedos like the Martian ice caps, and are not possible for asteroids. Feldspar has no absorption feature in the 0.3–1.1 μm range, but has a weak absorption near 1.25 μm if Fe^{2+} is a minor substituent, as in lunar volcanic rocks. Iron-rich metal has no absorption feature, but the general trend of reflectance with frequency is more linear than for silicate. The spectral signature may vary with Ni content. All silicates containing abundant transition-metals have a strong increase of absorption in the blue and ultra-violet range caused by charge-transfer. Pyroxenes give a range of spectra: Fe-bearing varieties have a symmetrical absorption near 0.95 μm , whose position depends on the Ca/Mg/Fe ratios; Fe-free varieties have a flat and featureless spectrum; and Ti-rich specimens have a strong absorption near 0.6 μm . Olivines also give a range: Fe-rich varieties have a broad, non-symmetrical absorption near 1 μm , whereas Fe-free forsterite has a flat and featureless spectrum. Hydroxylated and hydrated silicates (clay minerals) have a weak absorption near 0.6 μm and a strong absorption near 3 μm . Hydroxylated iron oxides

have strong absorption near 0.7 μm . Graphite and carbon-rich compounds have a flat, featureless spectrum with very low albedo. Spectra from mixtures are determined principally by the darker components, e.g. relatively transparent olivine is difficult to detect in the presence of relatively opaque pyroxene, while graphite and other very dark materials such as magnetite strongly reduce spectral features from intermixed silicates. The finer is the grain size of a darker phase, the greater is its dominance in spectra from mixtures with bright phases. So far, the important phase troilite has not been considered in the interpretation of asteroid spectra, but it may be a significant opaque phase on asteroids, especially if it is fine-grained, or if it is coating other minerals.

Originally McCord and Chapman (1975) emphasized the following parameters for asteroid spectra: R/B, reflectance ratio at red (0.7 μm) and blue (0.4 μm), which tends to separate C from S types; BEND, curvature, which tends to measure silicates/(opaques and metal); DEPTH, depth of pyroxene band at 0.95 μm ; and IR, reflectance at 1.05 μm minus that at 0.73 μm , negative values of which indicate olivine. The difficulty of detailed interpretation is pointed up by the controversy over the different types of data for Eros (Larson *et al.*, 1976; Pieters *et al.*, 1976).

Gaffey and McCord (1978) classified asteroid spectra as follows: The RA type has near-linear decrease of reflectance vs. frequency for visible light (hence Fe,Ni-metal component) and a silicate absorption feature near 1 μm (hence olivine or pyroxene or both). Sub-groups RA-1 and RA-2 were chosen depending whether olivine or pyroxene is indicated as the dominant silicate. Many subtle differences were discussed including the Ca content of the pyroxene, the Ni content of the iron, the presence of plagioclase, and possible presence of a wüstite coating. The RR type has near-linear decrease of reflectance vs. frequency, but no detectable silicate absorption (hence Fe,Ni-metal component and possible Fe-free silicate). Low albedo of some RR asteroids might result from a fine-grained opaque phase. There is a possible weak absorption near 0.6–0.7 μm . The RF type has decreasing reflectance with frequency in the visible range and near-constant reflectance in the infra-red (hence dominated by metal plus Fe-free silicate?, but Ni-content of metal and possible presence of opaque phases such as carbon or magnetite is uncertain). Perhaps troilite will prove important in RR and RF spectra. The A type for Vesta is indicative of basaltic achondrite (eucrite?), and for Dembowska is indicative of olivine-rich achondrite with possible minor pyroxene and metal. The F type has a flat spectrum from 0.4–1 μm

(and low albedo 0.02–0.05); interpretation is very difficult, but Gaffey and McCord preferred a mixture of Fe-bearing silicate (olivine?) and magnetite, while not ruling out a mixture of opaque phases and Fe-poor clay minerals; a definitive interpretation is needed since the large asteroids Ceres and Pallas belong to this group. The *T* type (for transitional between *F* and *R*) has a fairly flat spectrum with weak absorption at 0.6–0.7 μm and a UV absorption feature; most C-type asteroids belong to this group, and some undetermined mixture of opaque phase(s) and Fe^{2+} – Fe^{3+} –silicate(s) is indicated. Sub-types *TA*, *TB*, and *TC* were proposed with approximate correspondence with C1 and C2 meteorites. Sub-types *TD* and *TE* with intermediate albedo (9–15%) and a significant UV feature indicative of mafic silicate (but no infra-red features) pose serious identification problems, though magnetite and/or carbon-rich material (or troilite?) are indicated.

The Gaffey–McCord approach provides the following possible matches between main-belt asteroids and meteorites (1977, Table 1): *A*, eucrite and olivine achondrite; *F*, C4 (Karoonda); *RA*, various stony-irons, including mesosiderite, pallasite, and olivine- or pyroxene-rich stony-irons; *RF*, enstatite chondrite or iron; and *T*, C1, C2, and C3.

Zellner *et al.* (1977) classified 44 Nysa, 64 Angelina, and 434 Hungaria as *E*-type asteroids, whose spectra indicate Fe-free silicate and a plausible match with enstatite achondrite. Nysa is highly aspherical, and its dynamic association with 135 Hertha and several small objects in a Hiryama family suggested fragmentation of a precursor with Hertha as core and Nysa as a mantle fragment.

Data for Earth- and Mars-crossing asteroids are sparse. Zellner and Bowell (1977) found only one C-type object in a dozen members of the Apollo and Amor groups. Particularly important is the possible matching of Eros (Zellner, 1976) and Toro with high-grade chondrites (H or L; grade 5–6?) and Alinda with C3.

Spatial distributions and implications. For the simple S–C–M–E–U classification [U—unclassifiable], there is a strong tendency for increasing abundance ratio (S/C) with decreasing distance from the Sun of main-belt asteroids (Chapman, 1976; Zellner and Bowell, 1977), and this trend apparently carries inwards to the Apollo and Amor asteroids (though the statistics are poor). An early suggestion that S objects predominated over C objects for asteroids with c.150 km diameter was apparently the result merely of statistical bias. Gaffey and McCord (1977, Fig. 10) concluded that the C2 and C3 types (assumed to be primitive condensates, not heated significantly) occur throughout the main belt, whereas C4 and differ-

entiated metal–silicate types tend to occur mainly in the inner half of the main belt. Details are disputable (especially with regard to the C4 type), but the over-all spatial trend of C/S or primitive/differentiated appears plausible.

The implications of the observations are unclear in view of uncertainty in the evolution of asteroidal orbits and the extent of catastrophic fragmentation (refs. in Wetherill, 1977; Shoemaker, 1977*a, b*). The main-belt asteroids could have formed in place or have been perturbed into the main-belt either from the inner or outer solar system. Of course, the simplest idea is that C- or primitive asteroids were the 'primordial inhabitants' produced by primary condensation, whereas S- or differentiated asteroids were perturbed into the main belt from the inner solar system. Attempts to reconstruct the precursor asteroids from presumed collisional fragments in Hiryama families (e.g. Gradie and Zellner, 1977; Zellner *et al.*, 1977) must be regarded as tentative, though fascinating. The sources of heat for differentiated asteroids are unknown, though plausible possibilities exist (Sonett *et al.*, 1970), and it is not known when the heating occurred, except by implication from meteorites. Gaffey and McCord (1977) proposed that all the larger asteroids apparently have surfaces of differentiated material (either silicate, metal, or C4 type), but whether this results merely from greater mechanical strength and resistance to disintegration of heated asteroids or some correlation between heat sources and size is not clear. The recent discovery of water of hydration on Ceres, the largest asteroid, by Lebofsky (1978) is inconsistent with this general proposition, and may require some revision of the spectral interpretation of C-type asteroids.

Although there are possible spectral matches between asteroidal surfaces and most types of meteorites, it is far from clear how meteorites derive from asteroids. Current dynamical theories (refs. in Wetherill, 1977) utilize the concept of perturbation of collision fragments from asteroids near the Kirkwood gaps thereby ultimately providing Earth-crossing material. Actually the spectral characteristics of Apollo asteroids appear to be dominated by S-type material in contrast to C-type dominance in the main belt. Perhaps the primitive main-belt asteroids yield finer debris than differentiated asteroids, and perhaps further study of Apollo asteroids will yield a correlation between size and spectral type. Particularly problematical from the dynamic viewpoint is attempted assignment of the supposed eucrite parent planet to the asteroid Vesta, which is far from an orbital resonance with Jupiter, but see Consolmagno and Drake (1977) and Hostetler and Drake (1978). Readers are

referred to Chapman (1976) and Gaffey and McCord (1977, 1978) for provocative speculations about the origin and development of asteroids, and to O'Leary (1977) for a proposal to mine Apollo and Amor asteroids, which just might enhance funding for scientific exploration.

Meteorites

All planetary models are explicitly or implicitly based on analogies with meteorites. The 'Heroic Age' of meteoritics (i.e. before the Apollo 11 mission) produced valuable but over-simplified ideas, and recent petrographic data interpreted in the context of solid-state theories are leading to complex interpretations involving condensation, accretion, crystal-liquid differentiation, collisional destruction, and metamorphism. For brevity, the following sections are rather selective and didactic. Recent general reviews are: Buchwald (1975), Nagy (1975), Wasson (1974), and Mason (1971*b*). Older but still valuable are: Keil (1969*a*), Anders (1964), and Mason (1962).

Carbonaceous meteorites. Three dozen silicate-rich meteorites carry substantial C and were characterized from bulk chemistry into Wüik types I, II, and III (data from Mason, 1971*b*):

	wt %								
	Si	Al	Fe	Mg	Ca	Na	C	S	H ₂ O
CI	10	0.8	18	10	1.1	0.5	3	6	20
CII	13	1.1	22	12	1.3	0.4	2	3	13
CIH	16	1.4	25	15	1.7	0.4	0.5	2	1

All carbonaceous meteorites are mechanical mixtures of materials that have reacted with gases or liquids at diverse temperatures. They coalesced at low temperature, and are grossly out of mineralogical equilibrium. A petrologic classification into C1, C2, C3, and C4 types depends on ill-defined features indicative of increasing consolidation, and detailed study is leading to sub-division of the crude chemical classification.

Five CI meteorites (typified by Orgueil) have bulk compositions whose non-volatile elements fit closely with the composition of the solar photosphere (Table II). They are regarded as the best indicators of the chemical composition of the solar nebula when appropriate amounts of H, rare gases, N, and C are added. However, they could not have formed as direct equilibrium condensates from the solar nebula, and the bulk chemical composition results from addition of many components. The CI meteorites are friable, and hydration of sulphates may have occurred since landing on Earth. Nagy (1975) describes the many mineralogical studies and lists twenty-five minerals. Phyllosilicates, too fine-grained for easy study, dominate the matrix. Bass (1971) identified montmorillonite and serpen-

tine, and Kerridge (1976, 1977*b*) and McSween and Richardson (1977) interpreted bulk electron-microprobe analyses of fine-grained matrix in terms of Mg-rich phyllosilicate(s) and a NiS-bearing component, perhaps the complex product of aqueous alteration. The carbon occurs mainly as a wide variety of organic compounds (Nagy, 1975), which have been interpreted by Anders and co-workers as the products of abiogenic synthesis. Carbon also occurs as several carbonates (Fe-bearing magnesite, calcite, dolomite). Magnetite occurs in several morphologies. Rare olivine and pyroxene grains have wide composition ranges (Reid *et al.*, 1970), and Kerridge and McDougall (1977) discovered particle tracks in only 15% of the olivine grains. Sulphur occurs as very rare troilite, pyrrhotite, and pentlandite, as native sulphur, and mainly as sulphates in complex veins: hydrated Mg- and Ca-sulphates occur, but the extent of terrestrial hydration is uncertain. There are no chondrules in CI meteorites. Mechanical aggregation of diverse material on a disturbed planetesimal regolith, together with aqueous alteration (Du Fresne and Anders, 1962), provides a plausible origin.

The remaining members of the carbonaceous chondrites actually do contain chondrules, though fine-grained matrix is dominant. Nomenclature depends on features selected by the classifier (Mason, 1971*a*; Van Schmus and Hayes, 1974; Wasson, 1974, 1977; Fitzgerald and Jones, 1977). Type CIII (c.C3) was subdivided into Ornans (O) and Vigarano (V) varieties (Van Schmus, 1969), the latter including the large Allende shower. Type CII (c.C2) includes the Murchison, Murray, and Mighei (M) meteorites. All these meteorites are mechanical mixtures of several components, though the bulk chemical compositions (Table II) can be approximated by simple mixing models related to equilibrium condensation theory. The Karoonda (C4) meteorite is interpreted as the product of open-system metamorphism of a C3 meteorite on the basis of volatile-element depletion (Matza and Lipschutz, 1977). Chemical and mineralogical properties are too complex for detailed description here. When normalized to Si, the abundances of Ca, Al, and Ti are lower and of Fe and Mn are higher in C1 and most C2 and C3(O) meteorites than in C3(V) and a few C2, C3, and C4 ones. Volatile elements tend to correlate inversely with the abundance of high-temperature minerals, and the bulk composition can be matched with equilibrium-condensation theory (e.g. the 'step' pattern in C3(O) and C3(V) meteorites), but with the possible complication of dust-gas fractionation in the Ornans sub-type (Anders *et al.*, 1976). But in detail the history of these meteorites cannot be interpreted in terms of equilibrium condensation, and in

particular the oxygen isotopes require mixing of unrelated components.

The complex features of CM meteorites are yielding to modern techniques, but the observations and interpretations are incomplete and somewhat controversial. Detailed studies of Murchison (Fuchs *et al.*, 1973) and of Murchison, Murray, and clasts in the Jodzie howardite (Bunch and Chang, 1978*a, b*; Bunch *et al.*, 1978) provided most of the following description. Textural relations, which include schlieren, foliation, and boudinage structures, indicate that the CM meteorites once resided in a regolith breccia. Pseudomorphing of clasts by phyllosilicates and poorly characterized phases indicates subsequent alteration. Murray contains xenoliths mainly of CM type, but 10% of C3 type in which phyllosilicates replaced olivine and pyroxene. Chemical and structural identification of the phyllosilicates in the CM meteorites is difficult and Bunch and Chang (1978*b*) found at least four types in Murray, probably corresponding to either septeclorites or serpentines. Electron microprobe analyses showed considerable C, S, and Ni, which probably result from intergrown submicroscopic phases. Opaque materials for several textural varieties, collectively described as poorly characterized phases, consist of a mixture of an Fe-rich phyllosilicate and a phase enriched in Fe, Ni, C, and S (Bunch and Chang, 1978*b*). These authors attributed these opaque materials to alteration of breccia matrix, rather than to direct nebular condensation (McSween and Richardson, 1977). Of the sulphides, pyrrhotine and pentlandite are texturally early, as is a mysterious phase that probably resembles the carrier of a strange xenon component in Allende (Lewis *et al.*, 1975). Olivine grains comprise 10–30% of CM meteorites (Mason, 1963*a*; Wood, 1967), mostly with 0–5% Fa, but up to 80% Fa in Murchison with Fe and Ni positively correlated (Fuchs *et al.*, 1973). Direct condensation of the olivines from the solar nebula was favoured by Grossman and Olsen (1974) and Olsen *et al.* (1977) because of presence of Ni, Cr, P-rich metal inclusions. Attribution of the olivine to release from chondrules upon conversion of glassy mesostasis to friable phyllosilicate (Richardson and McSween, 1978) is apparently negated by the observation of C-rich mounds on the surface of the olivine grains (Bunch and Chang, 1978*b*). The occurrence of rare hibonite and perovskite (high-temperature minerals) with gypsum and whewellite (low-temperature minerals) reinforces the other evidence for mineralogical disequilibrium. Compaction of olivine and phyllosilicate at -4.42×10^9 yr for Murray and -4.22×10^9 yr for Nogoya is indicated by fission-track densities (MacDougall and Kothari, 1976).

For the C3 meteorites, McSween (1977*b, c*) emphasized the petrographic diversity. The Vigarano subgroup has larger chondrules and greater abundance of dark matrix than the Ornans subgroup, as well as higher Ca, Al, and Ti and lower Fe. Of the fourteen Vigarano meteorites, two (Renazzo and Al Rais) have phyllosilicate matrices and two (Coolidge, Mulga West) have been metamorphosed to Van-Schmus–Wood grades 4, 5, or 6. The main group of ten meteorites (including Allende) has 32–51% by volume of Type I chondrules (granular forsterite, Fa_{0-10} , clinoenstatite, Fs_{0-7} , Ca-Al-rich glass; some with no opaque minerals, but most riddled with metal, magnetite, and/or sulphides). There are traces of Type II chondrules (olivine, Fa_{20-50} , and Ca, Al-rich glass) and Type IV chondrules (melilite, anorthite, spinel, diopside, glass), and four types of melilite-rich chondrules and aggregates were distinguished by Martin and Mason (1974). Amoeboid olivine (Fa_{1-36}) aggregates with minor pyroxene (four types), spinel, anorthite, perovskite, nepheline, and sodalite (Grossman and Steele, 1976) amount to 1–9%. Two types of Ca, Al-rich inclusions [Type A 80–85% melilite, 15–20% spinel, 1–2% perovskite, rare anorthite, hibonite, wollastonite, grossular, pyroxene ($< 9\% \text{Al}_2\text{O}_3$, $< 0.7\% \text{Ti}$); Type B 35–60% pyroxene ($> 15\% \text{Al}_2\text{O}_3$, $> 1.8\% \text{Ti}$), 15–30% spinel, 5–25% anorthite, 5–20% melilite] amount to 3–9%, but see Blander and Fuchs (1975) for a four-fold classification and arguments for crystallization from a liquid rather than direct condensation at high temperature from a gas; of course, melting could have occurred after condensation, as also suggested by Grossman (1975). Rock and mineral fragments (1–3%) include isolated olivine grains, which were attributed to release from chondrules (McSween, 1977*a*), but see preceding paragraph. The matrix (35–51%) consists mainly of olivine grains, free from radiation damage (Green *et al.*, 1971) and coated and interspersed with carbonaceous material (graphitic?) of uncertain nature (Bauman *et al.*, 1973), and a few per cent of opaque minerals including metal, sulphide, and magnetite.

McSween subdivided the Vigarano group into three groups: five reduced meteorites with 0.5–5% kamacite plus taenite, 1.5–2.6% troilite, and 0.1–0.6% magnetite; five oxidized meteorites, including Allende, with 0–0.2% metal, 1.6–4% Ni-rich pentlandite plus minor troilite, and 1.2–3.2% magnetite; and a reduced pair (Renazzo, Al Rais) classifiable as petrographic type C2. Perhaps magnetite resulted from oxidation of sulphide (Herndon *et al.*, 1976). High Co contents of kamacite fall outside the main range of meteoritic metal (Fuchs and Olsen, 1973) and perhaps require crystallization in

an olivine-differentiated environment (Smith and Steele, 1975).

Several Vigarano-type meteorites, including Allende, are agglomerates, and many types of clasts await detailed study. Fe-rich rims on fine-grained inclusions in Allende indicate metamorphic transfer of Fe from the matrix, and coarse-grained inclusions have Na-rich rims and complex textural features (Hutcheon *et al.*, 1978). It is not clear whether chemical changes occurred prior to or after final consolidation, or both. Certainly the Vigarano-type meteorites have some chemical and mineralogical features attributable to high-temperature condensates (e.g. Grossman and Clark, 1973; Grossman and Ganapathy, 1977; Grossman *et al.*, 1977) but petrographic data require complex sequential processes, as advocated by Fuchs, Grossman, and others (e.g. Kurat *et al.*, 1975; McSween, 1977*d*). Oxygen isotope data (see later) require survival of pre-solar grains enriched in ^{16}O (Clayton *et al.*, 1977), and a mass of data on isotope anomalies in Allende inclusions (Clayton, 1978) is pointing to a pre-solar history probably associated with a supernova. Trace-element data require complex interpretations (Chou *et al.*, 1976; Takahashi *et al.*, 1978), whose correctness will require testing by sophisticated chemical studies of individual fractions selected with petrographic control.

The Ornans-type meteorites contain olivine-rich chondrules, inclusions (some Ca,Al-rich), rock fragments, matrix, metal, and sulphide. McSween (1977*b*) placed them in a metamorphic sequence from Kainsaz to Warrenton and perhaps Karoonda, in which the olivine and pyroxene become richer in Fe while the metal loses Cr and gains Co. The ratio of matrix to chondrules changes.

A tentative new group of carbonaceous chondrites (denoted CK) consists of the Kakangari and Adelaide chondrites, together with the lunar fragment Bench Crater (references in Fitzgerald and Jones, 1977). All three have chondrules in a magnetite-bearing unequilibrated matrix, and the principal bulk chemical feature is low Ca/Al (c.0.5 atomic) perhaps resulting from condensation of a gas that had lost Ca-rich materials. Some xenoliths in meteorites may belong to this group. Data for Kakangari are given later.

Many features of the carbonaceous meteorites would have been obliterated by metamorphism at temperatures exceeding 500–1000 K. There is abundant evidence of metasomatism in CI and CII meteorites probably associated with water or steam. Coexistence of isotopically anomalous Ca,Al-rich inclusions with olivine-rich matrix in CIII meteorites requires survival of presolar grains.

Ordinary chondrites. The ordinary chondrites have similar chemical compositions (Table II), but can be separated into three groups on the basis of total Fe: high (H) c.28 wt%, low (L) c.22 and very low (LL) c.20. They contain rounded bodies (chondrules), and display petrographic properties ranging from the Van-Schmus-Wood type 3 (sharply defined chondrules; much opaque matrix and glass; compositionally diverse olivines and pyroxenes; no feldspar and taenite) to type 6 (indistinct chondrules; recrystallized matrix and no glass; uniform olivines and pyroxenes; distinct feldspar and taenite). In the Heroic Age it was possible to envisage accretion of chondrules and fine debris from the solar nebula into three planetesimals each of near-uniform composition, followed by metamorphism varying in intensity with depth. Unfortunately this attractive story (e.g. Wood, 1962) is insufficient, and it is necessary to consider six processes for each chemical group: formation of a debris cloud of diverse components during a major collision of planetesimals; mechanical sorting of the cloud under the influence of a central gravitational field; sequential accretion with change of composition, especially of volatile elements; thermal metamorphism varying with depth; formation of a regolith, yielding mainly monomict breccias with varying content of adsorbed solar gases, but producing some polymict breccias incorporating late-accreting material; and disruption of the regolith and underlying metamorphosed material during the collision that started fragments on their paths towards the Earth.

Many petrographic features can be interpreted by analogy with lunar breccias (Prinz *et al.*, 1977*a*). Chondrites composed of light- and dark-textured components with the same bulk composition (e.g. Frederiksson and Keil, 1963; Bunch and Stöffler, 1974) but different contents of solar-implanted gases can be explained by impact mixing of a regolith. Brecciated chondrites are mostly monomict, indicating a uniform regolith on the parent planetesimal, but rare specimens contain inclusions of a different type. Wilkening and Clayton (1974) summarized existing data on inclusions of carbonaceous type in brecciated ordinary chondrites, and further examples were given by Fodor and Keil (1976), Fodor *et al.* (1976), and Wilkening (1976). Complex regolith annealing and disruption are implied by occurrence of unequilibrated inclusions in a more equilibrated host in the Weston brecciated chondrite (Noonan and Nelen, 1976). Strong shock caused melting in irregular veins of some chondrites, with formation of disseminated sulphide droplets and veinlets (e.g. Wahl, 1963; Bergemann and Wlotzka, 1969), and severe shock caused conversion of plagioclase, olivine, and

pyroxene to maskelynite, ringwoodite, and majorite in some chondrites (Tenham, Coorara, etc.), and emplacement of melt in the St. Mesmin chondrite (Dodd and Jarosewich, 1976). Weaker shock resulted in lithification of some gas-rich chondrites (Ashworth and Barber, 1976).

Most ordinary chondrites have textures essentially unaffected by impact, and they can be classified into the Van-Schmus-Wood types 3 to 6. There can be no doubt that types 4 to 6 have undergone prolonged thermal metamorphism, and attention is directed first to the unequilibrated ordinary chondrites of type 3, which retain textures from the original aggregation. Such type 3 chondrites consist of two types of chondrules set in a fine-grained complex matrix. 'Droplet' chondrules have textures (excentroradial; barred) definitely resulting from supercooling of spheres of silicate-rich liquid (e.g. Kurat, 1967*a*; Blander *et al.*, 1976), while 'lithic' chondrules have textures matching those of igneous rocks (e.g. Kurat, 1967*a*; Dodd, 1978). The early literature did not always distinguish between the two types, but the later literature recorded features indicative of their origin: e.g. Dodd (1971, 1974) noted that droplet chondrules were unshocked whereas lithic ones were shocked in the Sharps and Hallingeborg chondrites. Olivines and pyroxenes have a range of Mg/Fe ratios and diverse CaO contents indicative of low to high crystallization temperatures (e.g. Dodd, 1973). These features suggest derivation of lithic chondrules by mechanical abrasion of a variety of igneous and metamorphic rocks. The fine-grained matrix contains volatile-rich material not in chemical equilibrium, such as the nepheline-rich and silica-oversaturated components of the Mezö-Madaras brecciated chondrite (Kurat, 1967*b*; see also Binns, 1968; Van Schmus, 1967). The Bovedy L3 chondrite contains An_{85} glass (Graham *et al.*, 1976) whereas most chondritic glasses have high Na/Ca. Detailed microscopy of the Tieschitz (H3) chondrite (Christophe Michel-Levy, 1976) revealed a fine coating around the chondrules, an interstitial deposit of Si,Al-rich material, primary cavities in chondrules and secondary ones produced by leaching of glass, and, finally, debris of broken chondrules. Clastic matrices of high porosity were observed in L3 and LL3 chondrites (Ashworth, 1977) as well as non-clastic matrices attributed to solid-state recrystallization.

The chondrules have roughly the same size distribution in all ordinary chondrites (Dodd, 1969). Controversy over the details probably results from selective disintegration of some chondrules, and the data of Dodd (1976) and Hughes (1977, 1978) obey a log-normal distribution. Hughes (1977) noted that the mass distribution

matches that in active meteor streams, and Dodd (1976) found that metal-rich particles were smaller than silica-rich ones. Dodd (1969) described evidence for mechanical sorting during accretion, and in 1976 proposed that this resulted during laminar flow of a particle cloud around a growing planetesimal, using aerodynamic ideas of F. L. Whipple. Such a cloud might result from disintegrative collision of two planetesimals, or disintegrative capture of one planetesimal by another.

Evidence from lunar rocks does not favour generation of ordinary chondrites by minor impacts into a regolith, because of the preponderance of agglutinates over chondrules in such a process (Kerridge and Kieffer, 1977). Volcanic fire fountains (Ringwood, 1961*a*, 1966*a*) could not produce bodies large enough to give chondrites with a large variation of metamorphic grade. Impact of dust grains (Lange and Larimer, 1973) and direct condensation of liquid droplets from the solar nebula (Wood, 1963; McSween, 1977*d*) do not provide a plausible explanation of the association of droplet and lithic chondrules in ordinary chondrites if the lithic chondrules form by mechanical abrasion of rocks. However, these other processes may well have occurred in the pre-planetesimal stage, and direct condensation followed by aggregation may be responsible for porous aggregates in carbonaceous meteorites (e.g. Grossman and Steele, 1976; Wood and McSween, 1978).

Another process for formation of chondrules is melting upon high-speed impact into a planetary atmosphere, but this also fails to give the association of droplet chondrules with a range of lithic chondrules.

Early mineralogic studies of metamorphism (reviewed in Dodd, 1969) have been extended by many detailed studies with the electron microprobe. Temperature estimates based on Mg/Fe/Ca contents of coexisting pyroxenes and olivines are somewhat uncertain but suggest *c.* 900 K for type 4 to *c.* 1300 K for type 6. Oligoclase develops as distinct grains in type 6, and the olivines and pyroxenes have narrow composition ranges in types 5 and 6. The Shaw chondrite was placed in type 7 because of evidence for anatexis and high metamorphic temperatures near 1550 K (Dodd *et al.*, 1975). Partial melting of metal and sulphide from the Shaw chondrite should have yielded a composition similar to but not the same as the San Cristobal IAB iron (Rambaldi and Larimer, 1976). Chromite increases five-fold from type 3 to type 6, and Cr decreases while Al and Ti increase (Bunch *et al.*, 1967). Urey and Mayeda (1959) interpreted the metal particles of chondrites as fragments of a disrupted planetesimal that had cooled slowly

before accumulation into a secondary object and reheating to 770 K. From zoning profiles of clear taenite in twelve chondrites, Taylor and Heymann (1971) deduced cooling rates of 1 to 10 K/10⁶ yr and crystallization temperatures of 610–660 K while Smith and Goldstein (1977) deduced very rapid cooling rates of 1 and 100 K/day for two severely reheated chondrites. An unresolved problem is the deduction by Pellas *et al.* (1978) that the Shaw chondrite must have cooled very slowly to retain fission tracks whereas metallographic data by Berkley *et al.* indicate rapid cooling. Gallium, Ge, and W increase in metal with petrologic type (Chou and Cohen, 1973; Rambaldi, 1976), perhaps because of extraction from silicate.

Although the major elements (especially Fe) of the ordinary chondrites allow classification into the H, L, and LL groups, irrespective of the details of brecciation or petrologic type, the minor and trace elements vary with petrologic type: specifically, Binz *et al.* (1976) found that Bi, In, and Tl were tenfold lower in equilibrated than in unequilibrated ordinary chondrites, and attributed this to pre-accretion processes in the nebula rather than to thermal metamorphism. Much more study is needed to sort out the effects of pre-accretion chemical variation and post-accretion metamorphic changes for which the heating studies of Ikramuddin *et al.* (1977 *a, b*) are providing valuable controls. In the Heroic Age the bulk chemistry of the chondrites was interpreted as the mixing of a high-temperature condensate with a low-temperature condensate, as detailed by Keays *et al.* (1971). Actually the success of this model must rely on statistical averaging of several materials into each type of supposed condensate, and planetesimal formation must intervene between nebular condensation and chondrule formation if the Dodd model is sustained. The chemical analyses of trace elements in individual chondrules by Osborn *et al.* (1973), and in separated minerals by Rambaldi,

point the way to further understanding of the genesis of ordinary chondrites.

It will be assumed now that the ordinary chondrites are the result of accretion of dust clouds produced by three separate collisions. Perhaps the ordinary chondrites represent only the outer parts of accreted bodies and the inner parts contain coarse breccias of blocks too heavy to go into the dust clouds: of course, melting might have caused major crystal-liquid differentiation of such material. Such a brecciated interior might be represented by the Netschaëvo iron (Bild and Wasson, 1977), which contains 25 % angular blocks of silicate with uniformly distributed metal. Perhaps this silicate-rich brecciated iron meteorite might be the product of collision of a chondritic planetesimal with an iron planetesimal. Anyway the silicate material is less Fe-rich than the H chondrites, and may point the way to discovery of other types of ordinary chondrites.

Enstatite chondrites and achondrites: reduced irons. Whereas the CI and CM meteorites have essentially no Fe in metal and sulphide, and the ordinary chondrites have Fe distributed between metal-sulphide and silicate (H c.3/4 of Fe in metal-sulphide; L c.1/2; LL c.1/3), the enstatite chondrites and achondrites (Table II) have no or very little Fe in the silicates. Indeed the enstatite meteorites are so reduced that metal contains Si (Ringwood, 1961*b*) and troilite contains Ti (Keil, 1969*b*), but with variable partitioning with silicate (Easton and Hey, 1968). Also the minerals oldhamite (CaS), sinoite (Si₂N₂O), osbornite (TiN) occur. The mineral assemblages and compositions match those predicted by Larimer (1975) for a nebula of cosmic composition with C/O ≥ 1 rather than the 0.6 used to explain the mineralogy of ordinary chondrites and carbonaceous meteorites (Table III).

Most enstatite chondrites (Mason, 1966) can be roughly classified into two main chemical and mineralogical groups (Keil, 1968):

	chondrules	volatiles	Fe	Si/Mg S	Plagioclase*			kamacite	troilite	oldhamite		enstatite				
					Ab	An	Or	Si	Ti	Mg	Mn	CaO	MnO	FeO		
I	+	to +	+ solar abundance	30–33	1.4	6	93	2–3	1–4	3.5	0.3–0.6	1.4	0.3	0.2	0.1	0.6
II	—	—	depleted	21–29	1.2	3–4	80–83	13–17	3–4	1.3	0.6–0.7	0.5	1.1	0.7	0	0.2

* Mol%; the rest wt%.

Type I belongs to petrologic grade 4, and type II to grades 5 and 6, but they cannot be related by isochemical metamorphism (Keil, 1968; Shaw, 1974). Detailed textural studies are needed to complement the careful mineralogical data. Type I chondrites have several types of sharply defined chondrules plus angular fragments (especially of metal) set in a diverse matrix. Type II chondrites are highly recrystallized with rarity or absence of

chondrules. Presence of silica polymorphs (mostly tridymite but some quartz in II) demonstrates a highly undersaturated composition, while djerfisherite (K-bearing Fe sulphide) shows that K is chalcophilic under highly reduced conditions. The sulphides (Ca, Mg, Mn, Fe)S have equilibrated to low temperatures (Skinner and Luce, 1971), and Larimer and Buseck (1974) deduced equilibration at 910–1110 K (possible systematic error of 150 K)

for the Types I and II chondrites at P_{S_2} $c.10^{-8}$ – 10^{-12} bar and P_{O_2} $c.10^{-28}$ – 10^{-37} bar.

Although there are general similarities among the enstatite achondrites, detailed mineralogical data indicate that sub-types will be needed.

Olsen *et al.* (1977) distinguished the Happy Canyon specimen as a separate type with a crystal cumulate texture attributed to melting of EII chondrite at a slightly higher oxidation state (to explain absence of Si in kamacite; 1% Ti and 5% Cr in troilite; enstatite $Fs_{0.4}$) in a body big enough to give congruent melting of enstatite. Perhaps the diopside ($En_{52.3}Fs_{0.9}Wo_{46.8}$) exsolved out of a solid solution of an enstatite polymorph and the pyroxenes equilibrated to $c.1150$ K based on the Lindsley and Dixon (1976) calibration. Congruent melting of enstatite requires a pressure greater than $c.2$ kb (Boyd *et al.*, 1964). Perhaps the pressure was high enough to yield ortho-enstatite instead of proto-enstatite. The plagioclase (An_{26}) is more calcic than that in EII chondrites (An_{13-17}), and is sporadically abundant (5–10%).

The Cumberland Falls meteorite has angular blocks of black chondrite set in an enstatite-rich breccia indistinguishable from other brecciated enstatite achondrites (Mason, 1962).

Unlike the silica-bearing enstatite chondrites the principal group of enstatite achondrites contains several per cent of forsterite in association with dominant enstatite and minor plagioclase, kamacite, and essentially the same group of accessory minerals as in the enstatite chondrites. The unbreciated Shallowater meteorite may have a cumulate texture (Foshag, 1940) with equant to spherical grains of forsterite (5%) and irregular grains of oligoclase (2%) poikilitically enclosed in enstatite laths (81%). Irregular masses of Ni,Fe (9%) up to 3 mm across show Neumann lines and pearly texture, and are sometimes associated with troilite (1.5%).

All the other E-achondrites are monomict breccias with coarse fragments of enstatite set in a groundmass dominated by enstatite. The Norton County achondrite (Beck and LaPaz, 1951; Keil and Fredriksson, 1963) has metal inclusions 5 cm across composed of Ni-poor kamacite (Ni 3.6 wt%), and disseminated troilite (Ti 0.4–4 wt%), along with copper, alabandine, daubréelite ($FeCr_2S_4$), and graphite. The Mayo Belwa achondrite (Graham *et al.*, 1977a) has highly shocked enstatite ($Wo_{0.6}$; FeO 0.02 wt%), forsterite, diopside ($En_{55}Fs_1Wo_{44}$; hence $c.1350$ K from Lindsley-Dixon calibration), albite ($Ab_{88}An_9Or_3$), Si-bearing metals (with Ni 2.5, 7.6, and 21 wt%), Cr,Ti-troilite, daubréelite, oldhamite, and fluor-amphibole needles projecting into 1 cm vugs. The Khor Temiki achondrite (Hey and Easton, 1967)

has albite ($Ab_{92}An_4Or_4$) and metal (inclusions with 50% Ni set in matrix with 4.5%). Enstatites are ordered orthoenstatite in unbreciated Shallowater, but are sometimes disordered in brecciated ones, probably because of impact deformation (Pollack, 1966; Reid and Cohen, 1967).

Systematic study showed a negative correlation between Si in metal and Fs in enstatite for some enstatite chondrites and achondrites (Wasson and Wai, 1970; Wai and Knowles, 1972). Metal particles occur inside enstatite crystals, and the grain size of chondrites increases as the chondrule content decreases. Wasson and Wai (1970) proposed that a planetesimal accreted enstatite chondrules from a cold solar nebula with mechanical separation of metal and silicate, and that external heating yielded achondrites on the outside, but not by igneous activity.

Based on the textural similarities of enstatite chondrites and ordinary chondrites, I tentatively prefer to generate the former from a debris cloud formed by planetesimal collision, but await detailed study to test this idea. The pyroxene relationships may favour derivation of the achondrites at considerable depth, and I prefer to consider them very tentatively as brecciated cumulates. Probably the present data are most easily interpreted in terms of several planetesimals, but detailed mineralogical and chemical studies may allow coordination of some sub-types into a single chemically zoned planetesimal.

Bulk chemical data were interpreted in terms of nebular condensation, metal-silicate-sulphide mechanical fractionation, and migration of mobile elements (Baedeker and Wasson, 1975; Binz *et al.*, 1974; Ikramuddin *et al.*, 1976) but analyses of separated mineral fractions are needed to test the models.

Melting of a hypothetical planetesimal might have produced an E-achondrite mantle and an iron-rich core, thereby explaining the low Fe and S content of E-achondrites. Only three iron meteorites have Si in the metal (Tucson, 0.8 wt%; Horse Creek, 2.5; Nedagolla, 0.14 wt%; Wai and Wasson, 1969, 1970). Troilite is probably absent in Horse Creek, Tucson, and Nedagolla (Buchwald, 1975), but Tucson contains brezinaite (Cr_3S_4) and abundant silicate inclusions (Bunch and Fuchs, 1969). Although these inclusions contain abundant forsterite (FeO 0.23; CaO 0.11 wt%) and minor orthopyroxene ($En_{99}Fs_{0.4}Wo_{0.6}$; $Al_2O_3 < 0.03$ wt%), which might be matched with enstatite meteorites, the rare diopside ($En_{54.8}Fs_{0.5}Wo_{44.7}$) contains 7.3 wt% Al_2O_3 and the feldspar is pure anorthite. The Tucson iron has a texture consistent with rapid cooling of taenite (Buchwald, 1975). The Nedagolla iron was remelted (Miyake and

Goldstein, 1974), and both it and the Tucson iron were cooled very rapidly (Nedagolla 1 K/min; Tucson 1 K/1000 yr).

The Bencubbin metal-silicate breccia contains clasts related to carbonaceous (c.CM-CO) and LL chondrites (Kallemeyn and Wasson, 1977) plus clinoenstatite and rare olivine crystals set in a matrix of opaque silicates (Newsom and Drake, 1977). Abundant metal clasts are rich in Cr and S, and preliminary analyses indicated Si substitution in some.

Obviously much study is needed to explain the chemistry, origin, and development of the enstatite meteorites and reduced irons.

Forsterite-bearing meteorites and silicate inclusions in irons. The iron contents of olivine and pyroxene are a guide to the extent of oxidation, and, in the preceding sections, the olivines have ranged as follows: pure forsterite crystals in CII meteorites (perhaps early condensate from the nebula) to unequilibrated Fe-bearing olivines in other C-meteorites; unequilibrated Fe-bearing olivines in type 3 ordinary chondrites to equilibrated ones in the higher grades (H,Fa₁₈; L,Fa₂₃; LL,Fa₂₈), and almost pure forsterite in E-chondrites.

Graham *et al.* (1977b) described four meteorites with substantial Mg-rich olivine and orthopyroxene, and similar compositions of troilite and metal. The Kakangari chondrite was briefly mentioned earlier as a member of the tentative CK group of carbonaceous chondrites. The remaining three meteorites lack chondrules, and have chemical and mineralogical analogies with silicate inclusions in Type I iron meteorites (Bild, 1977; Davis *et al.*, 1977). The Pontlyfni meteorite (Graham *et al.*) has abundant sulphides (21 % troilite, daubréelite) and metal (25 % kamacite, taenite) associated with silicates (forsterite, enstatite, diopside, oligoclase) in a fine-grained mosaic. The Mount Morris (Wis) and Winona stones are badly altered, but contain minerals of similar composition, as well as graphite and chromite (refs. and data in Bild, 1977).

Silicate inclusions occur in half of the group-IAB irons (Buchwald, 1975), and probably result from trapping in pools of liquid metal. Most other groups of irons do not contain silicates, but a few anomalous and IIE and IIICD irons do. Bunch *et al.* (1970) described three groups and three single irons, and further data are summarized in Bild (1977). The Odessa (IA) type of silicate inclusions has > 5 % graphite and troilite, rounded to irregular shape, granoblastic texture, olivine Fa₃₋₄, enstatite En₉₁₋₉₂Wo_{1.2-1.9}, diopside En₅₁₋₅₄Fs₃Wo₄₃₋₄₆ (hence equilibrated to sub-solidus temperatures c. 1300 K), plagioclase An₉₋₁₄Or₃₋₄, zincian chromite of distinctive composition (Bild, 1977), troilite

with 0.2–0.6 % Cr, alabandine, sphalerite with 22–31 wt % Fe, Cl-apatite, merrillite, kamacite, taenite, and copper. [Whitlockite was the name used in most papers up to 1978 for the calcium phosphate mineral in meteorites. Because of small structural differences from terrestrial whitlockite, the name merrillite has been resuscitated by the I.M.A. Commission on New Minerals and Mineral Names for the lunar and meteoritic mineral. Merrillite has been used here for all meteoritic occurrences, but definite proof is needed especially for C1 meteorites.] The Copiapo (IA) type has < 5 % graphite and troilite, angular shape, diopside En₅₂₋₅₅Fs₂₋₄Wo₄₂₋₄₄, enstatite En₉₀₋₉₅Fs₄₋₉Wo₁₋₂, forsterite Fa₁₋₇, plagioclase An₁₁₋₂₂Or₂₋₄, Cr-spinel, and Cr-troilite. Scott and Wasson (1975) did not accept the distinction between Odessa and Copiapo types. The alkali-rich Weekeroo Station (IIE) type has no graphite, variable texture, drop-like inclusions, augite En₄₆Fs₁₈Wo₃₆, bronzite En₇₅Fs₂₂Wo₃, no olivine, Cr-spinel, Cr-free troilite, phosphate, and K-rich feldspars (Bunch and Olsen, 1968). The shocked Kodaikanal member (Bence and Burnett, 1969) contains high-albite solid solution (Ab₆₁₋₈₆Or₁₁₋₃₉ with 0–10 % excess SiO₂), glass Ab₄₂Or₄₄Qz₁₄, augite En₅₂Fs₁₀Wo₃₈, orthopyroxene En₈₁Fs₁₇Wo₂, chromite, Ca-phosphate, and kamacite-taenite Widmanstätten intergrowth indicating rapid cooling of 10²–10³ K/10⁶ yr. Enon (Anom) with Fe-rich olivine and orthopyroxene, and oligoclase An₁₅Or₆, resembles mesosiderites. The Steinbach stony-iron has metal of IVA type, and the silicate portion (c. 50 %) contains tridymite, orthopyroxene En₈₅(CaO 0.23 wt %) and clinopyroxene inverted from protopyroxene (En₈₅; CaO 0.17), chromite, troilite, and schreibersite. The original assemblage should have equilibrated at 1470 K, *f*_O, 10⁻¹² bar, and total pressure < 2 kbar; rapid cooling at high temperature is indicated by the clinopyroxene texture, and slow cooling (5–20 K/10⁶ yr) at low temperature by the diffusion profile of the Widmanstätten pattern and the Mg,Fe ordering in the orthopyroxene. Kendall County (Anom) has graphite with minor Fe-free olivine and pyroxene, and variable plagioclase Ab₈₇–Ab₅₁. Netschaëvo (IIE-Anom) has recrystallized relict chondrules, diopside En₄₉Fs₆Wo₄₅, enstatite En₈₄Fs₁₄Wo_{1.04}, oligoclase An₁₄Or₄, chromite, merrillite, chlorapatite, metal, troilite and schreibersite (Olsen and Jarosewich, 1971).

Bunch *et al.* questioned why the pyroxene thermometer gave temperatures (c. 1300–1500 K) much higher than the ones required to give the kamacite-taenite assemblage (< 700 K?), and suggested trapping of metamorphosed silicate in younger Ni-Fe melt. However, this problem also

occurs in other meteorite groups and may result from easier solid-state diffusion in metal than in silicate.

The Winona meteorite contains zincian chromite

	Kakangari	Mt. Morrison	Winona	Pontlyfni
olivine (Fa at %)	3-9	3	5	0.8-1.3
orthopyroxene (at %)	Fs ₃₋₁₄ Wo ₂	Fs ₄ Wo ₂	Fs ₇ Wo ₂	Fs ₁₋₃ Wo ₂
diopside (at %)	not found	Fs ₁ Wo ₄₆	—	Fs ₁ Wo ₄₆
feldspar (at %)	—	An ₂₃ Or ₃	An ₁₆ Or ₄	An ₁₂ Or ₃
troilite (Cr wt %)	0.3	0.4	—	0.8
kamacite (Ni wt %)	6.0	6.0	6 to	6.1
taenite (Ni wt %)	26-32	35-44	14	23-28
total Fe (wt %)	23	20	—	34

The Mundrabilla iron (IA-Anom) is remarkable for the high S content resulting from c.35 vol% troilite nodules (Ramdohr, 1976; Bild, 1977), some containing angular silicate inclusions (olivine Fa₃, orthopyroxene Fs_{6.5}Wo₂, clinopyroxene Fs₃Wo₄₅, plagioclase An₁₁Or₄). The troilite contains 0.5 wt% Cr and 0.1% Cu (Weinke, 1977) and other sulphides are daubreelite, alabandine, and sphalerite. Graphite occurs mainly at grain boundaries, often associated with schreibersite. The Waterville meteorite (Scott and Wasson, 1976) is mineralogically similar, but finer-grained and lower in Ir (0.30 ppm vs. 0.87).

The Campo del Cielo (IA) iron has mineralogical and chemical properties indicative of heating to a higher temperature than other IA irons with silicate inclusions (Wlotzka and Jarosewich, 1977).

Unquestionably the discovery of silicate inclusions in irons depends somewhat on chance. Temporarily it seems best to assume that the IAB irons are characteristically associated with silicates whose Mg/Fe ratios indicate aggregation into a planetesimal from material of oxidation state intermediate between those for ordinary chondrites and enstatite chondrites. Perhaps the IAB irons, the Mt. Morrison, Winona and Pontlyfni stones, and the Mundrabilla troilite-iron, are samples of a modally heterogeneous planetesimal composed of collision debris that became hot enough for partial melting but not core separation. Other bodies are needed to accommodate the Weekeroo Station (IIE) irons and anomalous irons not in the IAB group.

Irons. Iron meteorites are valuable indicators of the accretion processes and metal-sulphide-silicate differentiation of at least a dozen and perhaps sixty planetesimals, or regions of planetesimals, for most of which there are no obvious silicate-rich meteorites. This results from the augmentation of thirty-one fallen irons by c.530 finds, and the lesser fraction of finds of stony meteorites. The Heroic Age produced a textural classification, which corresponds roughly with bulk Ni, and a burgeoning chemical classification based principally on Ga and

and silicates of similar compositions to those in the Odessa-type irons. The following table shows other comparisons between forsterite-bearing meteorites:

Ge (Scott and Wasson, 1975). All members of a Ga-Ge group are believed to arise from the same planetesimal, while the range of textures (if any) gives some clue to the physical locations of the irons in the parent body. Some irons were severely reheated. That many have been shocked (Jain *et al.*, 1972; Wasilewski, 1976) is not surprising since irons should tend to be deep-seated in differentiated planetesimals.

Principal textural classes are hexahedrites (H) [large single crystals of kamacite ($\equiv \alpha \equiv$ body-centred iron); 5.5 wt% Ni; Ga-Ge group IIA; some recrystallized to fine grain size upon reheating], octahedrites [kamacite-taenite intergrowths attached on the octahedral planes of the original high-temperature single-phase taenite ($\equiv \gamma \equiv$ face-centred iron); the Widmanstätten structure varies in coarseness from coarsest octahedrite (Ogg) via Og(coarse), Om(medium), Of(fine), and Off(finest) to plessitic octahedrite (Opl)], ataxites (D) [heterogeneous, fine-grained; Ni-rich ones are kamacite-taenite intergrowths] and anomalous irons [often fine-grained, or with abundant sulphide or silicate, or both]. The over-all features can be explained by the binary relations between low-temperature kamacite and high-temperature taenite in the Fe-Ni phase diagram (reviewed by Goldstein and Short, 1967), but many complexities result from nucleation features related to minor substituents (especially, P, C, and S; e.g. Clarke and Goldstein, 1978). Diffusion profiles lead to estimates of cooling rate, which are based on the key assumption of monotonic cooling.

Important chemical features of the twelve major Ga-Ge groups are listed in Table V taken mainly from Buchwald (1975; Table 27), but extended with data from Scott and Wasson (1976, Table 2).

It is quite impossible to do justice here to the magnificent compilation of mineralogical and textural data in Buchwald (1975) together with the detailed metallographic and geochemical studies of many workers. From the viewpoint of planetary mineralogy emphasis is placed here on features

TABLE V. *Summary of iron meteorites*

Ga-Ge group	Approx. number	Textural class	Cooling* rate	Ni wt %	Ga ppm	Ge ppm	Ir ppm
IA	82	Om-Ogg	2-3	6-9	55-100	190-520	0.6-5.5
IB†	8	D-Om	—	9-25	11-55	25-190	0.3-2.0
IC‡	10	diverse	1-10 000	6-7	42-54	85-250	0.07-9
IIA	43	H	2; 10	5.3-5.7	57-62	170-185	2-60
IIB	14	Ogg	—	5-7	45-60	105-185	0.01-0.5
IIC	7	Opl	100-500	9-12	35-40	85-115	4-11
IID	11	Of-Om	—	10-11	70-85	80-100	3-20
IIE	12	variable	—	7-10	21-28	62-75	0.5-8
IIIA	117	Om	2	7-9	17-22	32-46	0.15-20
IIIB	40	Om	2-10	9-11	16-22	28-46	0.01-0.15
IIIC	6	Of	—	11-13	11-26	8-35	0.07-0.55
IIID	5	Off-D	—	16-23	1-5	1-4	0.01-0.07
IIIE	7	Og	—	8-9	17-19	34-36	0.05-0.6
IIIF	5	—	5-200	7-8	6.3-7.2	0.7-1.1	1.3-7.9
IVA	39	Of	7-200§	7-9	1.6-2.4	0.09-0.14	0.1-4
IVB	11	D	5-200?	16-18	0.17-0.27	0.03-0.08	4-36
Anom	92	All	—	5-60	0.05-87	0.005-2000	0.007-54

Data mainly from Buchwald (1975, Table 27); see also Wasson (1974, Table II—7), Scott and Wasson (1975, Table 2; 1976) and Scott (1978c).

* Quoted as $K/10^6$ yr assuming monotonic cooling through 773 K; data from Kelly and Larimer (1977) and Randich and Goldstein (1978).

† See Buchwald (1975) for details of I-Anomalous irons listed here in group IB.

‡ Scott (1977d).

§ Disputed by J. I. Goldstein and J. T. Wasson groups.

resulting from accretion and from metal-sulphide-silicate differentiation.

Scott (1972) interpreted the trend between Ni and various trace and minor elements of a Ga-Ge group in terms of igneous fractionation within a single parent body while the distinctions between Ga-Ge groups were attributed to primary fractionation in the solar nebula. Of course, this binary distinction can be blurred if some planetesimals accreted inhomogeneously. Furthermore the model of progressive equilibrium condensation of the solar nebula may be far too simple. For the latest compilation of trace elements, Scott (1978a) concluded that the non-members of the twelve major Ga-Ge groups are sparse samples of some fifty additional planetesimals.

Kelly and Larimer (1977) set up a detailed chemical model. At 10^{-5} bar the first condensate contains Fe-rich metal with 19 ± 3 wt% Ni. It is enriched in all metals more refractory than Ni (viz. Os Re Mo Ir Ru Pt Rh), and strongly depleted in elements (Cu Zn Au Ga Ge) more volatile than Ni (Wasson and Wai, 1976; Scott, 1978b). Equilibrium cooling results in dilution of the Ni and refractory elements with condensing iron to give c.6 wt% Ni. The refractory elements are also diluted, while the volatile elements are augmented up to the cosmic

ratio. Cobalt and Pd condense simultaneously with Ni, and the Co/Ni and Pd/Ni ratios remain constant. All the above elements, except Cu and Mo at high temperature, are more noble than Fe, and oxidation of Fe (with incorporation into silicate) results in an increase in Ni content, while ratios of noble elements stay constant. Sulphidization of the Fe will also increase the Ni content and not change the ratios of non-chalcophilic elements; however, the sulphide could remove chalcophilic elements (e.g. Cu, Zn). Melting and gravitational segregation in a planetesimal is very complex, but partial melting can yield an Fe,FeS eutectic melt, which is gravitationally segregated from immiscible silicate liquid or solid silicate residue to yield a core. Alternatively, segregation is incomplete leaving 'raisins' of Fe,FeS in a silicate host, or melting does not occur, and there is merely solid-state partitioning of elements. Crystallization of the Fe,FeS melt is controlled by the partition of elements between solid and liquid phases. Chalcophilic elements should concentrate into the sulphide melt. Elements that melt at high temperature should favour solid Fe-rich metal over Fe-rich liquid. Controlled experiments on these phenomena are only just beginning by M. J. Drake, J. I. Goldstein, and co-workers, and present interpretations must

be based largely on the element correlations observed by Scott for the various Ga-Ge groups. All the above processes are complicated by the redox state, and by the nucleation phenomena associated with minor amounts of phosphides and carbides.

The four elements in Table V are sufficient to give clues to the origin of the major groups of iron meteorites, though readers are referred to many details in published papers.

Group IVB is enriched in refractories (Ir 4–36 ppm) and depleted in volatiles (Ge 0.03–0.08 ppm), and has high Ni (16–18 wt %). Kelly and Larimer suggested accretion at 1270 K shortly after metal began to condense from the nebula. Minor metal-silicate fractionation may have occurred. The fine texture precludes accurate measurement of the cooling rate, but rapid cooling is indicated. Group IIID also has high Ni (16–23) but the Ir content (0.01–0.07) is much too low for a simple refractory condensate, and strong solid-liquid fractionation in a planetesimal is indicated.

Group IVA is strongly depleted in volatile elements but has low Ni (7–9) and Ir (0.1–4). Trace elements correlate with Ni as expected for fractional crystallization.

The abundant IIIAB irons have element correlations indicative of fractional crystallization in a single core, but the range of cooling rates 2 to 10 K/10⁶ yr may indicate more than one metal reservoir. These irons resemble ordinary chondrites in having five-fold depletion of volatile elements (e.g. Ga, Ge), a similar redox state, and exposure ages indicating a major collision at -6.5×10^8 yr. Detailed study of the huge Cape York iron revealed many trace minerals (Kracher *et al.*, 1977): four Na-bearing phosphates (including buchwaldite NaCaPO₄), six sulphides (daubréelite, chalcopyrrhotine, djerfisherite, another K-rich sulphide, sphalerite, Fe-alabandine), chromite, copper, carlsbergite (CrN), and a silica mineral. The main group of pallasites (next section) is chemically related to the high-Ni IIIAB irons (Scott, 1977a, c).

The abundant IIAB irons have low Ni and high Ga and Ge indicative of low-temperature accretion, and strong Ir variation indicative of strong crystal-liquid differentiation in a single core. Six IIA hexahedrites have cooling rates of 2 K/10⁶ yr whereas one has 10 K/10⁶ yr (Randich and Goldstein, 1978). The low Ni content of 5.3–5.7% for hexahedrites indicates that all the Fe and Ni were in the metal phase, and none was in a silicate phase. However, hexahedrites are not as reduced as enstatite meteorites because they do not contain Si in the metal.

Group IAB irons have trace element trends interpreted by Kelly and Larimer as the result of

seepage of metal during partial melting but without aggregation. These irons are characterized by silicate inclusions described in the preceding section. The angularity of the silicate inclusions suggests derivation from a metal-silicate agglomerate. Gravitational segregation of metal and silicate occurred only locally. Solid-state metamorphism yielded a silicate-equilibrated mosaic with pyroxene grains established at c.1300 K. The kamacite and taenite intergrowth results from solid-state annealing at c.600 K of a taenite solid solution established at >1000 K.

Group IC irons (Scott, 1977d) have low Ni and high Ge and Ga, together with an Ir-Ni correlation indicative of metal-liquid fractionation. Cohenite and chromite are abundant and the cylindrical sulphide inclusions are free of silicate and graphite. Schreibersite and minor carlsbergite occur. The wide range of cooling rate (1–10⁴ K/10⁶ yr) was attributed to burial at different depths of impact debris from a body undergoing metal-liquid fractionation.

It is convenient here to mention briefly the evidence on the pressure at which meteorites crystallized. The early conclusion by Anders (1964) that the maximum static pressure was not above c.3 kb still stands for at least the great majority of meteorites. This conclusion rests on the lack of inversion of tridymite to quartz, the evidence for protopyroxene, the attribution of diamond to shock rather than static pressure, and the reinterpretation of early speculations that high pressure was needed for cohenite (Brett, 1967) and kamacite-taenite intergrowths (Goldstein and Doan, 1972). For meteoritic silicates there is no need to invoke static pressures greater than a few kilobars, except just possibly for pallasites (next section). For irons, the Fe content of sphalerite, when in equilibrium with troilite, provides a barometer if the temperature can be estimated. Schwarcz *et al.* (1975) estimated pressures of 1.7–3.1 kb for the Bougou, Gladstone, Odessa, and Sardis irons of Type I. The pressure increases as Ge decreases, suggesting a chemically zoned planetesimal of radius up to c.180 km if made only of iron and c.350 km if silicate-rich. Further studies of sphalerite-bearing irons have not yet appeared, except for an estimate of 10–15 kb for the Cape York IIIA iron (Kracher *et al.*, 1977), and c.1 kb for the sphalerite with 28 wt % Fe in Mundrabilla (Ramdohr, 1976).

The importance of estimating the cooling rate from kamacite-taenite intergrowths was recognized by Wood (1964), and later work has demonstrated the complications caused by supercooling and heterogeneous nucleation on inclusions (reviews by Goldstein and Short, 1967; Goldstein and Axon, 1973). Controversy continues over details,

but Widmanstätten intergrowths must have exsolved from a supercooled parent taenite solid solution at a temperature below c.900–1000 K that depends on the bulk Ni content. Furthermore, the cooling histories of iron meteorites must vary considerably, and the possibility of complex cooling histories, such as annealing in a breccia after disruption of a planetesimal must be considered. Utilization of phosphide inclusions in hexahedrites (Randich and Goldstein, 1978) points the way to further estimates of cooling rate using exsolution of minor elements, while new analytical techniques may permit study of fine-grained plessites (Lin *et al.*, 1977).

The occurrence of S, C, P, and N in iron meteorites is particularly relevant to planetary differentiation. Troilite is the most abundant sulphide mineral. Complete separation of an Fe,FeS eutectic might be expected in differentiated planetesimals with subsequent crystallization into metal and troilite, which might become mechanically segregated in a gravitational field. Meteorites composed entirely of troilite are not known, perhaps because fragmentation of a planetesimal core would shock-melt or vapourize a troilite region. Most iron meteorites carry little troilite (<2%), but a few contain a lot, of which the Mundrabilla (preceding section) and Soroti irons are important examples (Buchwald, 1975). It would be very useful to attempt to estimate the metal-sulphide ratio from the relative content of chalcophilic to siderophilic elements in iron meteorites, but different condensation properties from the solar nebula cause problems. Taken at face value the planimetric data of Buchwald (1975, Table 30) indicate that the S content of most iron meteorites is much less than expected for the cosmic Fe/S ratio, but it is not certain whether this results from lower condensation of volatile S than less-volatile iron, or sampling effects. There is no obvious correlation between S content and assignment of iron meteorites to either a core, raisins, or incipient melt regions, except for the high S content of I-Anom irons attributed to non-segregated melts. Troilite is absent in the reduced irons Tucson and Nedagolla, but brezinaite occurs in the former and daubréelite in the latter.

Carbon is sparse in irons except for group I with 0.2–2 wt% graphite, cohenite, and haxonite, and groups IC, IIICD, and IIIE with somewhat lower abundances of carbides (Scott and Wasson, 1975, Table 3). Graphite is commonly associated with troilite, while cohenite occurs in metal, especially in IC irons. In spite of this low abundance in irons simple analogy suggests that an iron-rich core may be the major store of C in terrestrial planets, rather than the atmosphere and crust (Smith, 1977).

Nitrogen has a very low abundance in irons (2–130 ppm; Gibson and Moore, 1971), but simple analogy again suggests that a planetary core could be the major reservoir rather than the atmosphere of terrestrial planets. Carlsbergite, CrN, occurs particularly in low-Ni irons of groups IIIA, IIA, and I (Buchwald, 1975).

Phosphorus is important in planetary differentiation as an indicator of oxidation state. Schreibersite (Fe,Ni,Co)₃P, and its textural variety rhabdite, are almost ubiquitous in iron meteorites (Doan and Goldstein, 1969; Scott and Wasson, 1975, Table 3) and arise either from P-rich liquid or exsolution in the metal. This proves that the irons developed in an environment sufficiently reducing that the P was not dominantly in the oxidized form (as in merrillite). Rare phosphates, some associated with phosphide, occur in IIIAB irons (Buchwald, 1975) and in the Verkhne Dnieprovsk and Baranca Blanca IIE (Olsen and Fredriksson, 1966; Bild, 1974) and Dayton IIID irons (Fuchs *et al.*, 1967); four of the phosphates contain sodium (Kracher *et al.*, 1977). The implications for the oxidation state await experimental synthesis, and calculation of possible oxygen diffusion.

The highly reduced state of the irons containing Si-bearing metal and Ti-bearing troilite was described earlier. Chromium is probably a common minor constituent of troilite from iron meteorites, whereas Ti probably occurs only in the very reduced ones. Systematic study of these and other elements in troilite and other sulphides from irons might provide a guide to the redox state, but the complex sulphide mineralogy must be taken into account (e.g. El Goresy, 1965, 1967) when interpreting distribution coefficients.

From this brief and highly selective survey it is abundantly clear that iron meteorites derive from a variety of nebular condensates that have been modified in many ways, including differentiation in many planetesimals. Further study, especially of anomalous irons, should increase the evidence for diverse origins.

Pallasites. These stony-iron meteorites consist essentially of olivine and metal, whose complex textures require prolonged solid-state annealing. Those with c.65 vol% olivine were interpreted by Buseck (1977) as cumulates of olivine crystals whose interstices were filled gently by liquid metal, but Scott (1977b) argued for violent mixing of molten metal with brecciated olivine. Those with > 65 vol% olivine were interpreted as adcumulates by Buseck (1977). A 'mean' pallasite has 65 vol% olivine, 31% metal, 2.3% troilite, 1.2% schreibersite, 0.4% chromite, and 0.2% phosphate giving a bulk composition: P₂O₅ < 0.1, SiO₂ 19.7, Cr₂O₃ 0.2, FeO 6.1, MgO 21.0, Fe 44.4, Ni 5.7, Co 0.2, S 0.2,

P 0.3 wt%. Most pallasites contain metal with similar Ni, Ga, Ge, Cr, W, As, and Au to high-Ni IIIAB irons (Scott, 1977a, c), and the chemical and textural properties of the latter would be consistent with derivation from the core of a planetesimal of which the pallasites represented the core-mantle interface. However, the apparent cooling rates inferred from the Widmanstätten intergrowth are 2 to 10 K/10⁶ yr for the irons and about 1 K/10⁶ yr for the pallasites (Buseck and Goldstein, 1969). Perhaps this can be resolved by appeal to different annealing conditions after disruption of the original parent planetesimal, but mineralogical evidence might be difficult to interpret. Three pallasites (including Eagle Station) have high Ge/Ga, and may be akin to some anomalous irons. Six pallasites are unique.

Buseck (1977) emphasized cumulate texture, angular shape, kink-banding of olivine, troilite eutectic with kamacite, and deformed plessite. Scott (1977b) emphasized disintegration of olivine masses into smaller angular fragments as the result of intrusion by molten metal. Adjacent olivine grains inside the fragments have triple contacts typical of solid-state annealing while olivines in contact with metal may have many small faces defining a nearly rounded interface. Scott suggested that the gravitational separation of olivine and metal was probably prevented by incipient solidification. Wood (1978) concluded that because olivine deforms easily olivine-metal cumulates would squeeze to pure dunite and metal in bodies of radius > 10 km. In order to obtain the slow cooling indicated by the Widmanstätten texture it would be necessary to aggregate small solid bodies into > 100 km bodies before final cooling of pallasite through 800 K. Probably it will be necessary to consider at least the following processes: melting, cumulation, deformation and solid-state annealing, violent disruption, further solid-state annealing, and perhaps final fragmentation. Cooling was probably not monotonic, and estimation of radii of parent bodies may be difficult or impossible.

The olivine and metal have Fa_{10.5-13} (atomic) and Ni8-12 wt% (main group), and Fa₁₉₋₂₀ and Ni 14-15 (Eagle Station trio). Olivine grains contain tubes 1-5 μm wide, and in tiny regions contain 4-5 wt% P₂O₅ (Buseck, 1977). Rare symplectites occur with pyroxene (Fs₁₂ or Fs₁₇). Cobalt is depleted 700 times in the olivine compared to metal, whereas Ni is depleted c. 7000 times (Turekian *et al.*, 1977). This is the prime basis of the argument that the Earth's upper mantle (not necessarily the lower mantle) did not equilibrate with liquid metal. Kolomeitseva (1975) interpreted Fe, Mn, Ni, Cr, and Co of pallasitic olivine as the result of metal-silicate

equilibration at 1400-1500 K and 20-70 kb, but this cannot be accepted until partition coefficients have been measured on synthesized phases.

Phosphorus occurs as widespread schreibersite and trace phosphates (merrillite with minor Na and K, stanfieldite, and farringtonite; Buseck and Holdsworth, 1977). Trace minerals include low-Ti chromite, sphalerite, pentlandite, copper, and graphite.

Chassignites and nakhlites. In view of all the evidence for differentiation, it is surprising that there are so few olivine- and pyroxene-rich materials. Perhaps olivine and pyroxene tend to fragment upon impact unless bound together by other minerals.

The Chassigny meteorite (Mason *et al.*, 1976; Floran *et al.*, 1977a) is texturally a shocked cumulate of high-Fe olivine (Fa₃₂; high CaO 0.16 wt%) with minor low- and high-Ca pyroxenes, plagioclase (Ca 4.63 Na₂O 8.15 K₂O 1.47 wt%), sanidine and chromite, and traces of chlorapatite, troilite, hydroxy-fluor-kaersutite (in melt inclusions in olivine), pentlandite, ilmenite, rutile, and baddeleyite. Bulk chemical data (Boynton *et al.*, 1976) fit with an olivine-rich cumulate from a chondritic magma.

A second chassignite, the recently discovered Brachina meteorite (Johnson *et al.*, 1977), contains dominant olivine (Fa₃₃; CaO 0.25 wt%) in a subhedral granular texture along with minor augite (En₄₆Fs₁₄Wo₄₀) and plagioclase (CaO 4.71 Na₂O 8.92 K₂O 0.26 wt%) and accessory chromite, troilite, pentlandite, and inferred phosphate. Olivine and chromite grains project into voids, and the meteorite is porous and unshocked.

The Nakhla and Lafayette meteorites are identical pyroxenites (Bunch and Reid, 1975; Bockor *et al.*, 1976) with cumulus augite En₃₈Fs₂₃Wo₃₉, minor olivine (Fa₆₅₋₆₈; CaO 0.45), interstitial oligoclase-sanidine (An₃₄Or₄-Or₇₅An₃), Cr-Ti-magnetite with exsolved ilmenite, pyrite with marcasite lamellae, troilite, chalcopyrite, and F-Cl-apatite.

The Fe-rich nature of the olivines is notable in all four meteorites, and it may prove possible to interpret the nakhlites as later differentiates than the chassignite in a planetesimal of Fe-rich peridotitic composition, perhaps chemically related to low-Fe chondrites. Laul *et al.* (1972) found the volatile elements of nakhlite to resemble those of ocean-ridge basalts, and Gale *et al.* (1975) reaffirm a Rb-Sr isochron that indicates igneous crystallization at -1.24×10^9 yr.

Angrite. The unique Angra dos Reis achondrite (Hutchison, 1972; consortium of papers, see Prinz *et al.*, 1977b) has 93% fassaite (Wo₅₄En₃₄Fs₁₂; TiO₂ 2.2 Al₂O₃ 10 wt%) occurring in a texture

interpreted as oikocrysts set in a lineated ground-mass. Triple junctions were interpreted as the result of metamorphism following igneous cumulation. Rare olivine ($\text{Fe}_{53}, \text{CaO } 1.3 \text{ wt } \%$) occurs with Mg-kirschsteinite, hercynitic spinel (requiring Fe^{3+}), merrillite, Ni-iron, troilite, plagioclase ($\text{An}_{13}\text{Or}_{0.7}$), celsian ($\text{Ca}_{90}\text{Al}_8\text{Ab}_2$), baddeleyite, and Ti-magnetite. High pressure is not required to explain the Al content of the pyroxene. A parent body of unusual composition is required for this meteorite, perhaps very rich in high-temperature condensate (i.e. like Ca-Al-rich inclusions in CIII meteorites).

Ureilites and lodranite. The Lodran meteorite (Bild and Wasson, 1976; Prinz *et al.*, 1978) contains three-eighths olivine (inversely zoned to grain boundaries and veins from Fa_{12} to Fa_6) and three-eighths unzoned orthopyroxene ($\text{En}_{84}\text{Fs}_{13}\text{Wo}_3$ with exsolved augite $\text{En}_{50}\text{Fs}_7\text{Wo}_{43}$), interpreted as a cumulate harzburgite surrounded by one-quarter metal (mainly kamacite; minor taenite) and troilite. Minor phases are schreibersite (which yields a cooling rate of $10 \text{ K}/10^6 \text{ yr}$), chromite and 'granitic' inclusions. The olivine zoning indicates reduction after crystallization, and various chemical features (including oxygen isotopes) indicate a different parent body than ureilites in spite of similar mineralogy. Fukuoka *et al.* (1978) interpreted bulk chemistry in terms of partial melting of a hybrid between H and E chondrites.

The seven ureilites have remarkable properties, which involve a complex history and controversial interpretations. Berkley *et al.* (1976, 1978) describe them as a pile of igneous cumulates with graphite (probably primary), which caused reduction of olivine (as for the Lodran meteorite). Strong shock produced diamond and lonsdaleite. The ureilites were classified as:

		wt % FeO/ (FeO + MgO) wt % C	
group 1	Dyalpur, Dingo		
(top)	Pup Donga,		
	North Haig	0.18–0.22	3.6
'group' 2	Haverö	0.27	2.1
group 3	Novo Urei,		
(bottom)	Kenna, Goalpara	0.25–0.32	1.5–2.2

Wasson *et al.* (1976) also interpreted the ureilites as a section of a planetesimal but from measurements of Ni Zn Ga Ge Cd In Ir and Au placed five ureilites in a different order from top to bottom: Goalpara (heavily shocked, low Ir and Ir/Ni), Haverö, Dyalpur, Novo-Urei, and Kenna (moderately shocked, high Ir, high Ir/Ni). External heating was supposed to have caused partial melting with greater extraction of metal in Goalpara than in Kenna. A major impact then injected C-rich and metallic material, which was not primary as in the model of Berkley *et al.*, into hot silicates. Olivine zoning requires reduc-

tion and rapid cooling. Diamond formed either during injection or later. The primary metal was a high-Ir nebular condensate, partly lost by partial melting, and the injected metal was low in Ir. Boynton *et al.* (1976) from bulk chemistry interpreted the ureilites as a mixture of ultramafic silicates with higher Co/Ni than metal, an indigenous metal phase remaining after partial melting, and carbon-rich material added after partial melting. The C-rich material is remarkable not only for the diamonds and lonsdaleite, but also for being rich in volatiles and primordial gases (Weber *et al.*, 1976). CI-type material was ruled out by Binz *et al.* (1975), who suggested a late 'distillate' of mobile elements. Higuchi *et al.* (1976) observed ten-fold enrichment of refractory Ir and Re over moderately volatile Ni and Au, and paucity of volatiles except for Ge, C, and noble gases. They interpreted ureilites as the residue from partial melting of a C3V-like body with incomplete separation of liquid. Vein material was injected later. Carbon was derived by surface catalytic reactions at c. 500 K. Noble gases but few other volatiles were trapped. Clayton *et al.* (1976) explained the oxygen isotopes by mixing of c. 0.2 % ^{16}O with anhydrous phases in C2 and C3 meteorites, or similar material.

Returning to mineralogy and petrography Berkley *et al.* (1976) provided detailed evidence in the Kenna ureilite for cumulation of olivine (core $\text{Fa}_{21}, \text{CaO } 0.42, \text{Cr}_2\text{O}_3 \text{ } 0.63 \text{ wt } \%$; rim to Fa_1) and pigeonite ($\text{En}_{73}\text{Fs}_{18}\text{Wo}_9$) in a lineated texture with triple junctions. Secondary melt inclusions and veins in olivine and pigeonite contain clinopyroxene ($\text{En}_{52-60}\text{Fs}_{6-25}\text{Wo}_{22-42}$), K-feldspar, andesine, chromite, and glass. Thin metal veins and blebs are rich in Co (mean 1.2 wt %) and have Ni 4–9 wt %. The carbonaceous interstitial material contains graphite, diamond, and lonsdaleite, and often occurs as veins tapering into metalliferous extensions, which contain some troilite and taenite. Brecciation and extreme mosaicism of olivine occur in Goalpara, Dingo Pup Donga, North Haig, and Haverö ureilites, but not in Kenna. Textural features in the Kenna silicates indicate shock in the 50–250 kb range, which should be high enough to form the diamond and lonsdaleite.

There can be no doubt that melting has occurred, and that rapid cooling and shock deformation followed, but the other events are highly controversial. Probably the graphite is not a cumulate phase because it would not be in equilibrium with Fe-rich olivine. Introduction as a vapour phase with grain-boundary reduction of olivine, perhaps at low temperature over a long period, seems more likely, especially as this could coincide with entrance of rare gases. Perhaps a shock event tended to fracture the igneous cumulate along

grain boundaries, thereby providing channels for the vapour phase (note that eclogite nodules from kimberlite typically are altered in the grain boundaries). The lineated texture may be difficult to explain in a residue from partial melting unless some directed stress were available; igneous cumulation seems more plausible. Perhaps the ureilites result from six events: igneous processes in a parent body containing some ^{16}O -rich material; disruption and incorporation as hot debris in another parent body; vapour deposition into fractures, mainly along grain boundaries; further disruption with formation of diamond; rapid cooling of the shock debris; and transport to Earth.

Eucrites and shergottites. These two groups of basaltic achondrites are mostly monomict breccias dominated by plagioclase and high-Fe pyroxenes, but a few unbrecciated specimens occur. The early survey by Duke and Silver (1967) outlined the evidence for magmatic crystallization followed by various degrees of impact and thermal metamorphism, and recent detailed studies are profiting by experience on lunar rocks.

The highly shocked Shergotty and Zagami meteorites (Duke, 1968; McSween and Stolper, 1978; Stolper and McSween, 1978; Smith and Hervig, 1978) are characterized by about 70% pyroxene (pigeonite $\text{En}_{70}\text{Wo}_{11}$ – $\text{En}_{20}\text{Wo}_{20}$ and augite $\text{En}_{48}\text{Wo}_{32}$ – $\text{En}_{33}\text{Wo}_{32}$), c.20% maskelynite ($\text{An}_{56}\text{Or}_3$ – $\text{An}_{41}\text{Or}_3$), c.2% Ti-magnetite, c.0.5% ferrian ilmenite, and c.4% mesostasis of fayalite, silica, and K,Na,Ca-feldspars. Melting experiments indicate approximately one-third cumulus pyroxene and two-thirds intercumulus liquid formed at c.1410 K. The shergottites are notable for the intermediate composition of the plagioclase glass, which contrasts with the oligoclase and albite in ordinary and enstatite chondrites and the calcic plagioclase in the eucrites, howardites, and C3 meteorites. Other volatile elements besides Na have a similar intermediate relationship between chondrites and eucrites. The shergottites are also distinguished from the eucrites by a higher oxidation state, and the magnetite–ilmenite assemblage equilibrated near the fayalite–silica–magnetite buffer. Minor pyrrhotine (Smith and Hervig, 1978) and merrillite occur in and near the mesostasis.

The eucrites also contain ferropigeonite and ferroaugite but the plagioclase is very calcic (An_{85-100}). The unbrecciated eucrites have igneous textures, and the Ibitira specimen even has vesicles (Wilkens and Anders, 1975), which survived later metamorphic events (Steele and Smith, 1976). The Moore County (Hess and Henderson, 1949), Moama (Lovering, 1975), and Serra de Mãe (Prinz *et al.*, 1977c) eucrites have cumulate texture, and the Moore County eucrite has pyroxene assem-

blages indicative of slow cooling (Hostettler and Drake, 1978). The brecciated eucrites have complex textures, but clasts display igneous textures. They contain much more U and Th than the unbrecciated eucrites (Morgan and Lovering, 1973), which are very depleted in large-ion-lithophile elements. The Ibitira unbrecciated eucrite contains no mesostasis and is extremely depleted in alkalis.

From melting experiments Stolper (1977) explained most eucrites by low-pressure fractionation of pigeonite and plagioclase from liquids similar in composition to the Sioux Co. and Juvinas eucrites. The compositions of the Stannern and Ibitira eucrites cannot be produced by fractionation of liquids with compositions like those of the other eucrites. Stolper suggested that a primitive undifferentiated body yielded batches of magma by partial melting of olivine Fa_{35} , pigeonite $\text{En}_{65}\text{Wo}_{5}$, plagioclase An_{94} , Cr-spinel, and metal. Initial melts yielded liquid compositions like those of Stannern, Ibitira, and Sioux Co., while later melts after exhaustion of plagioclase yielded Mg-rich liquids responsible for plutonic crystallization of Mg-rich pyroxenes and olivine found in howardites, diogenites, and mesosiderites. The cumulate eucrites may have accumulated from liquids resulting from extensive fractional crystallization of advanced partial melts.

Minor amounts of olivine, silica minerals, low-Ni metal, troilite, chromite, apatite, and merrillite occur in eucrites, but systematic petrographic and mineralogic studies are needed.

The pyroxenes show complex exsolution textures from which Takeda *et al.* (1976) and Miyamoto and Takeda (1977) deduced that eucrites crystallized from 0.5 to 12 km depth. This range is plausible for a large planetesimal, but the accuracy of the deduction may be affected by a multi-stage metamorphic history, as proposed for the Ibitira eucrite (Steele and Smith, 1976).

In general the eucrites resemble lunar basalts (McCarthy *et al.*, 1973) but the details rule out derivation from the Moon. The simplest assumption that the eucrites come from the same parent body has allowed estimates of bulk composition from several lines of geochemical reasoning. Details vary, but most features are similar (e.g. Anders, 1977; Consolmagno and Drake, 1977; Morgan *et al.*, 1978a). The bulk composition in Table IV (Morgan *et al.*, 1978a) resembles that for H-chondrites except for the lower content of moderately volatile elements Na, K, and Rb in eucrites, and the lower Fe,Ni metal in eucrites, though total Fe is similar. Perhaps the parent body of the shergottites is closer to H-chondrite than eucrites because of the greater abundance of alkalis.

Diogenites. The nine hypersthene achondrites (Mason, 1963*b*; one Antarctic meteorite) are composed essentially of orthopyroxene Fs_{23-28} with accessory bytownite, olivine, troilite, chromite, tridymite, and metal. Six are monomict breccias, Tatahouine is unbrecciated, and Garland and Aïoun El Trouss are polymict. Their chemical properties are quite similar, and cosmic-ray exposure ages (Herzog and Cressy, 1977) suggest that five originated in a collision at -14×10^6 and perhaps three at $c. -24 \times 10^6$ yr.

The Johnstown diogenite contains coarse clasts in a fine matrix. Orthopyroxene $\text{En}_{72-74}\text{Fs}_{23-25}\text{Wo}_{2-3}$ occurs in fractured grains up to 2 cm across, some with triple junctions. Minor phases are bytownite $\text{An}_{82-90}\text{Or}_{0.1}$, clinopyroxene $\text{En}_{46-48}\text{Fs}_{9-10}\text{Wo}_{44-45}$, olivine Fa_{28-29} , silica mineral, troilite, metal, and chromite. Floran *et al.* (1977*b*) explained these data by impact deformation and recrystallization of a primary cumulate. Gooley and Moore (1976) found that metal in eight diogenites occurs with chromite and troilite as arrays of inclusions predating fractures in orthopyroxene, and as larger grains in matrix. Most metal has relatively low Ni (usually 0–5 wt %) and relatively high Co (mostly 0–7 wt %), but taenite (Ni 15–53 wt %) occurs in three diogenites. A complex origin is indicated: reduction of pyroxene with migration of siderophile elements and silica to fractures and grain boundaries; annealing and recrystallization of pyroxene with formation of new grain boundaries; and possible addition of metal from an external source. Miyamoto *et al.* (1975) deduced slow cooling from the Mg/Fe ordering of Johnstown orthopyroxene, and noted that the Ca content was too low to give high-temperature exsolution.

McCarthy *et al.* (1973) emphasized the association of eucrite and diogenite clasts in the polymict brecciated howardites (next section), and Stolper's model for eucrites involves partial melting of olivine Fa_{35} , pigeonite $\text{En}_{65}\text{Wo}_5$, and plagioclase An_{94} , Cr-spinel, and metal, all of which minerals are represented in diogenites. However, detailed studies are needed to test whether the plagioclase in diogenites (bytownite) is too sodic to yield eucritic liquid by partial melting. Perhaps one planetesimal supplied both the eucrites and diogenites, but there were separate reservoirs on it. Radial zoning might have been produced during accretion.

Howardites and mesosiderites. Both groups of meteorites have complex unequilibrated mixtures of mineral and rock clasts, which have undergone varying degrees of metamorphism both before and after final agglomeration. The twenty mesosiderites are approximately equal mixtures of silicates and metal plus troilite, whereas the nineteen howardites

have only a little metal. Texturally and mineralogically the howardites match regolith breccias from the Moon (Bunch, 1975; Prinz *et al.*, 1977*a*) but chemical compositions rule out derivation from the Moon. Although eucrite and hypersthene achondrite (diogenite) clasts occur in both howardites and mesosiderites, the difference in other clast populations suggests derivation from different regoliths (Floran and Prinz, 1978), perhaps on different planetesimals formed after separate collisions with a single eucrite–diogenite planetesimal, or perhaps from more than one eucrite–diogenite planetesimal.

Duke and Silver (1967) described the basic mineralogy and petrography of howardites, and Bunch (1975) gave a survey of the clasts, which are dominated by pyroxenites and basaltic to gabbroic rock fragments resembling diogenites and eucrites, but which also consist of types not hitherto recognized in meteorites.

The range of rock types is so similar to that in lunar breccias that details need not be given here; let it suffice (Bunch, 1975), that the howardites must have originated in a body that differentiated the same suite of rocks as the Moon. Clasts of CH meteorites were observed by Bunch, Wilkening, and Dymek *et al.* The Kapoeta howardite (Dymek *et al.*, 1976) is rich in rock clasts set in a glassy matrix comprised of comminuted mineral grains. The petrographic features require plutonic-to-volcanic differentiation of basaltic material dominated by pyroxene prior to thermal and impact metamorphism in a regolith. Dymek *et al.*, concluded that this howardite is not a simple mixture of eucrite and diogenite, and noted that the Fe/Mn ratio rules out derivation from the Moon. Crystallization ages demonstrated differentiation at $c. -4.55$ to -4.6×10^9 yr. The Malvern howardite (Desnoyers and Jerome, 1977) has an unusually large amount of glass (20%). Hewins and Klein (1978) found that the Ni/Co pattern of the metal has too little Co to match metals equilibrated with silicates, and suggested that it results from 'meteoritic' metal impacting the regolith.

Powell (1969, 1971) interpreted the mesosiderites as unequilibrated polymict breccias because of the complex textures and the occurrence of Mg-rich olivine with tridymite. The non-metallic portion is dominated by orthopyroxene (Fs_{20-40}), pigeonite ($\text{Fs}_{40-50}\text{Wo}_{10-13}$), plagioclase (An_{80-98}), and chromite. Rare metal fragments have Widmanstätten texture, but most kamacite nucleated at irregular grain boundaries. The inferred cooling rate of $0.1 \text{ K}/10^6 \text{ yr}$ is even slower than for pallasites. Mason and Jarosewich (1973) recorded a range of pyroxenes from $\text{Fs}_{18-26}\text{Wo}_2$ (Veramin) to $\text{Fs}_{35-52}\text{Wo}_5$ (Dyarrl Island) and noted an inverse

relation between bulk wt% and vol% of metal. Nehru *et al.* (1978) recorded the following components in the Emery mesosiderite: mineral clasts of pyroxene, olivine, and metal; rock clasts of diogenite, eucrite, and cumulate eucrite; recrystallized metal-silicate clasts; recrystallized clasts of all the above; and a norite clast. Floran *et al.* (1978) concluded that an impact-melt was the source of at least the silicate portion of the Simondium, Hainholz, and Pinnaroo mesosiderites, and suggested that the impact target was pyroxene-rich from the preponderance of low-Ca pyroxene clasts. The abundance of anorthite clasts was ascribed to cumulate eucrite and basaltic eucrite. Finally, Floran and Prinz (1978) classified the mesosiderites, and concluded that silicate clasts strongly dominate over rock and metal clasts, that silicate clasts are dominated by low-Ca pyroxene (orthopyroxene and pigeonite), plagioclase and olivine, that diogenite is the dominant rock clast while cumulate eucrite is abundant, and that there is a range of recrystallization textures.

Although the mesosiderites must have developed in a shocked regolith, they were released from the regolith by an event that did not shock them greatly, judging from the preservation of inter-clast textures.

The mesosiderites have a common origin (Mason and Jarosewich, 1973) because of sequential composition range of pyroxene, near-constant Ca,Al ratio, and near-uniformity of trace elements and Ni in the metal. Wasson *et al.* (1974) found that the metal in seventeen mesosiderites (Ni 7–9 wt%; Ga 13–16 Ge 47–58 Ir 2.4–4.4 ppm) is in the same general range as IIIAB and IIE irons, pallasites and H-chondrites, but does not fit exactly with any group. Weckeroo-type irons have higher Ni, Ga, Ge, and Ir on an extension of the mesosiderite trend, but the feldspar is K-rich rather than Ca-rich as in the mesosiderites.

The complex polymict nature of the howardites and mesosiderites makes a fitting close to this mineralogic and petrographic survey of the meteorites. There can be no doubt that eucrite and diogenite fragments were reworked in one or more planetary regoliths, and were mixed with other material to give the howardites and mesosiderites, but many uncertainties remain, including the possibility that the eucrite and diogenite fragments differ in detail from the eucrite and diogenite meteorites.

Oxygen isotopes. The ratios of the three oxygen isotopes constrain the genetic relations between the Earth, Moon, and meteorites (Clayton and Mayeda, 1978). The relations in fig. 11 can be explained only by invoking both mass fractionation between different phases reacting with the

same reservoir of oxygen, and mechanical mixing of material from at least two reservoirs.

Simple mass fractionation produces phases joined by a line of slope near 0.5, such as the heavy line representing materials from the Earth and Moon, which apparently derived their oxygen from the same reservoir or the same mixture of reservoirs. A solid phase condensing from the solar nebula below c.700 K is isotopically heavier than a coexisting gas phase (inset diagram), and the isotopic separation is greater at lower temperature (LT) than at higher temperature (HT). Mathematically the C2, Bencubbin, and Kakangari meteorites could be related by mass fractionation, but not with any other specimens in fig. 11.

Rocks from the upper mantle of the Earth fall at M, the position for the Moon. The oxygen isotopic ratio of the Earth's lower mantle is unknown, and will be assumed to be identical with that of the upper mantle. Equilibrium crystal-liquid differentiation of the whole Earth might give a lower mantle that is isotopically heavier or lighter than point M on the heavy line.

Since the meteorites do not lie on a single mass-fractionation line, mixing from more than one reservoir is needed. The range for C3 meteorites defines a line of slope 1 which is explainable by addition of a minor component rich in ^{16}O , and the simplest assumption is that the solar nebula was inhomogeneous, probably because of addition of variable amounts of supernova debris rich in ^{16}O . The meteorites can then be related by means of mass-fractionation from gas-liquid-solid processes coupled with addition of ^{16}O .

For this model the order of increasing addition of ^{16}O is:

- (i) L and LL chondrites;
- (ii) H and C1 chondrites; silicate inclusions in IIE irons;
- (iii) nakhlites;
- (iv) E chondrites and achondrites; Moon and outer part of Earth;
- (v) eucrites, howardites, diogenites, most mesosiderites, two main groups of pallasites, silicate inclusions in IAB irons, and Winona;
- (vi) ureilites;
- (vii) C2 and CK chondrites; Bencubbin and Weatherford stony-iron breccias;
- (viii) C3 and C4 chondrites;
- (ix) Eagle Station and Itzawisis pallasites.

Varying degrees of gas-solid, liquid-solid, or solid-solid mass fractionation can be deduced. The carbonaceous meteorites are isotopically heavier than other meteorites because of condensation at lower temperature of solid from the residual gaseous nebula. Solid-liquid or solid-solid fractiona-

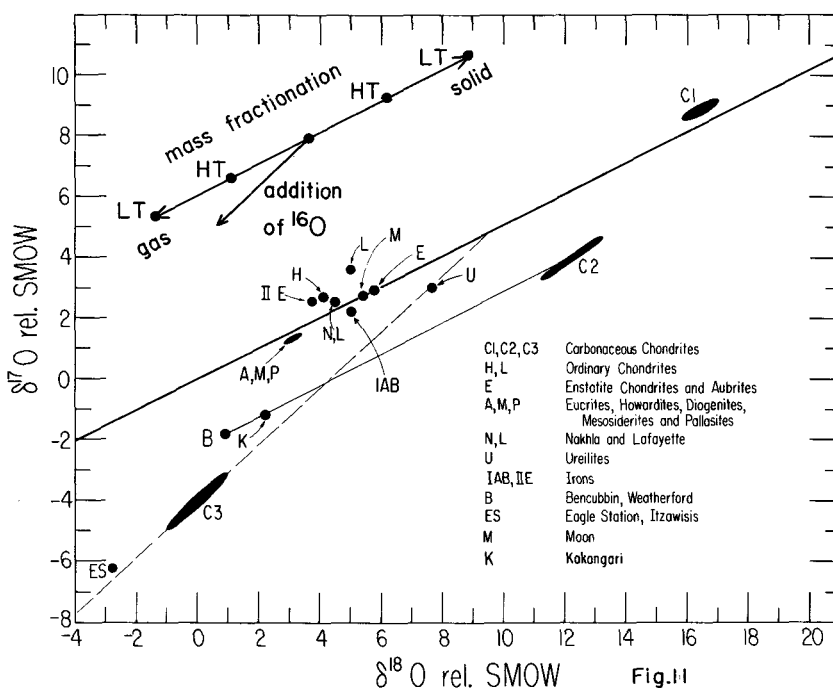


FIG. 11. Oxygen isotopic ratios of meteorites referred to SMOW standard. The heavy line is the mass-fractionation line for earth materials (Clayton, 1978).

tion at high temperature could explain the small spread between the achondrites in group (v), but other evidence suggests that more than one planetesimal is involved. The silicate inclusions in IAB irons are too heavy isotopically to be related to any of the group (v) achondrites, and a different planetesimal is needed. Clayton and Mayeda (1978) suggested that the IAB silicates might be related to the E-meteorites by addition of ^{16}O , as might H chondrites to L and LL chondrites. The silicate inclusions in IIE irons are close enough to H chondrites to be explainable by mass fractionation.

These isotope data reinforce the mineralogic and petrographic evidence for a complex history of meteorite formation, and conclusively demonstrate the need for many parent bodies condensing from a chemically inhomogeneous nebula. Although E-meteorites provide the closest match of oxygen isotopes to the Earth and Moon, so many mixing models are possible (e.g. of ordinary chondrites + C2 chondrite + basaltic achondrites) that there is little isotopic control over the relation of meteorites to the Earth, except to rule out more than small amounts of meteorites with extreme compositions (e.g. all the carbonaceous meteorites).

Original location of the meteorites. That most meteorites underwent prolonged metamorphism and crystal-liquid differentiation, together with

long exposure to cosmic rays, and that many passed through a regolith, argues strongly for derivation from substantial planetesimals whose relics are now represented by asteroids (Anders, 1975). Only the Apollo and Amor asteroids are currently reaching into Earth-crossing orbit, but these must contain material migrating from the main-belt asteroids, mostly too small to be seen by telescopes. Although incomplete matching of meteorite and asteroid spectra can be explained in many ways, including the two possibilities that meteorites represent only a tiny fraction of the diverse range of asteroidal materials, and that asteroidal surfaces undergo mineral sorting or selective degradation (e.g. Anders, 1978), there is considerable concern that the spectra of main-belt asteroids do not provide a match for the most abundant meteorites—the ordinary chondrites. A host of differentiated silica-rich materials is required to match the iron meteorites, which themselves may represent only a small fraction of the M asteroids.

Some micrometeorites must come from the tails of comets, but there is no concrete evidence that any macrometeorites come from comets. It is possible to envisage ejection by Jupiter of planetesimal fragments from the inner solar system and subsequent transport by comets back to Earth-intersecting orbit after a complex history in the

outer solar system. Comets may transport material that originally condensed in the outer solar system. It may be easier to invent ways of deriving carbonaceous meteorites from comets than the crystal-liquid differentiated meteorites, but one must not

overlook the strong evidence for polymict textures and aqueous alteration. The next section will assume that all differentiated meteorites result from condensation of the solar nebula (gas plus solid particles) inside the orbit of Jupiter.

SUGGESTED SYNTHESIS

So many worlds, so much to do
So little done, such things to be.

Alfred, Lord Tennyson, *In Memoriam*

Whereas the preceding sections have tended to recite and criticize the available data and ideas on individual planets, the present section attempts to provide a comprehensive framework for further studies.

Pre-planetary stage

To explain the oxygen isotopic ratios let us assume that the nebula was a mixture of dust and gas, with some dust carrying supernova debris rich in ^{16}O . The nebula did not undergo simple progressive condensation with equilibrium reaction of solids with residual gas. Some early condensates were incorporated into substantial bodies, without addition of more volatile components, as evidenced by the irons rich in Ni and refractory siderophile elements. Many chemical types of planetesimals were formed. There is abundant evidence of crystal-liquid differentiation. Probably early formed bodies were melted by short-lived radioactivity whereas late-formed ones stayed cool (time scale of 10^6 yr for ^{26}Al). Collisions of planetesimals produced complex products including breccias and chondrule-agglomerates. The planets were assembled from many materials, which passed through complex planetesimal stages. Chemical models based on simple progressive condensation are inaccurate in detail, but essentially correct because of the statistical averaging of the components entering the planets.

A crucial question is the redox state of the nebula, which is controlled by the C/O ratio when H is highly abundant. Meteorites demonstrate a wide redox state from the super-reduced enstatite meteorites and Si-bearing irons, via a wide range of reduced irons and stones with troilite and low-Fe silicates, to the oxidized CI meteorites with magnetite. Let us assume that the oxidized meteorites come from a cool part of the nebula in which hydroxylated and hydrated silicates could form; to be more precise, let us assign them mainly to the outer part of the main-belt asteroid region, and perhaps partly to the regions between the giant planets. Then, the reduced and super-reduced meteorites must be assigned to the inner solar

system, probably reaching to the inner realm of the main-belt asteroids. If so, how can Earth and Venus have oxidized surfaces? Let us assume that the inner planets began their accretion almost entirely from reduced material in near-circular orbits (fig. 7), and ended their accretion with oxidized material projected into the inner solar system by the late-growing outer planets. The larger bodies (Earth and Venus) could have captured essentially all the high-speed oxidized material whereas the smaller bodies (Mars, Mercury, and Moon) would have captured very little. Finally, the planets could have lost volatiles through the atmosphere and ionosphere, and even the Earth might have lost most of its primordial H with concomitant oxidation of C, S, and Fe. These assumptions of heterogeneous accretion and planetary gas loss allow the Earth and Venus to have begun accretion from reduced, and perhaps super-reduced material. Almost the entire population of planetesimals inside the orbit of Mars would have been captured by the inner planets, but a few survivors might have reached the main belt of asteroids by a series of lucky encounters. This interpretation of the redox state implies a radial gradient of C/O from ≥ 1 in the inner solar system (near Mercury and Venus?) to < 1 (near 0.6? in the middle to outer solar system; outside the Martian orbit?). But how was this accomplished? Did it involve grains of graphite in the primordial nebula? Could radial migration of gas and solids have been important (propulsion by solar wind or radiation pressure?)? The present synthesis will fail if answers do not appear to these questions.

There must have been some outward gradient of pressure and temperature in the solar nebula, but it seems desirable for the temperature to be low enough for survival of isotopically unusual grains, probably derived from a super-nova. If this is correct the early lumps (fig. 3) could be a mixture of surviving grains and new solar condensates. Chemical equilibration would occur by melting in early planetesimals heated by short-lived radioactivity rather than by solid-gas reactions in the nebula.

Earth

The following framework is suggested for a new model of heterogeneous accretion (fig. 12). Growth began with accretion of hot reduced material from near-by planetesimals, perhaps dominated by super-reduced enstatite-rich materials, as suggested by naïve interpretations of oxygen isotopic ratios (fig. 11). Continuous crystal-liquid differentiation resulted in separation of an Fe,Ni,S,C-rich core (including some K, U, and Si??) and a 'basaltic' rim (*sensu lato*) from a pyroxene-rich mantle. The growing Earth would be surrounded by a thick cloud of collision debris and gas.

Accretion continued with expansion of the feed-

ing zone as the growing Earth 'pumped up' the relative velocity of uncaptured planetesimals. Because of the greater competition from Venus than Mars, the captured planetesimals would tend with time to come increasingly from the Mars side with a chemical tendency for higher redox state. Volatile loss from colliding planetesimals would tend to convert Fe-poor pyroxene and Fe-rich metal into Fe-bearing olivine. Perhaps most of the last quarter of accreting material consisted of differentiated stony-iron planetesimals, represented in meteorites by the A, M, P group in fig. 11, while a small amount would consist of material like ordinary chondrites, and an even smaller amount would be of carbonaceous type, perhaps closer in mean composition to C2 than C1 meteorites. The propor-

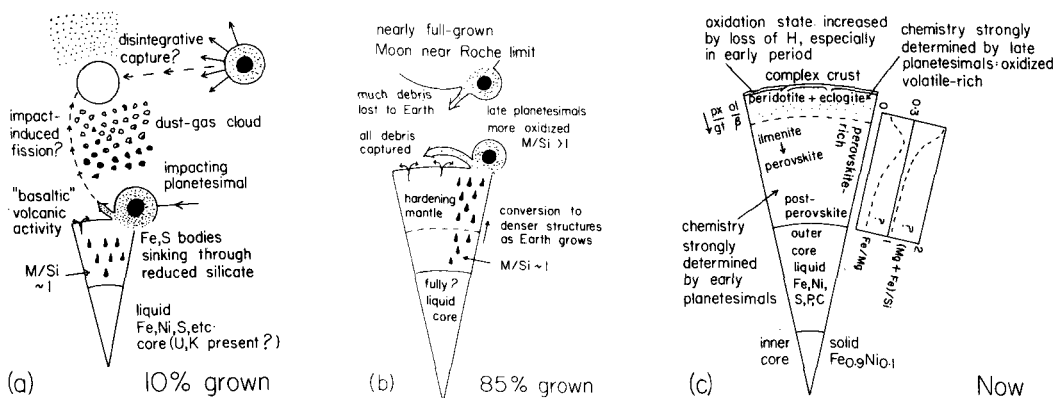


FIG. 12. Cartoon of growth of Earth and Moon. (a) Early stage in which the baby Earth has grown from hot bodies in the middle of the Earth growth zone (fig. 7). These bodies are assumed to resemble enstatite meteorites, and have differentiated into a liquid core rich in Fe, Ni, S, etc. (perhaps with U, Th, and K?; Smith, 1978) and an enstatite-rich exterior through which negative diapirs of Fe,Ni,S-rich material are sinking. Strong volcanic activity is present at the surface. A baby Moon may have grown by disintegration of an incoming planetesimal either with or without collision with the Earth. A dust/gas cloud surrounds the Earth, and aerodynamic sorting tends to grade the impact debris from fine particles at high altitude to heavy bodies near the Earth. (b) Late stage in which the Earth is now accreting planetesimals that originally formed at considerable distance from the Earth, with a tendency to come from the Mars rather than the Venus side. These planetesimals tend to be more oxidized and have a greater ratio of octahedral cations (M) to Si (i.e. they contain substantial olivine). The core is fully liquid, and the inner solid core will not develop for another $c.10^9$ yr when sufficient cooling has occurred. The lower mantle does not equilibrate completely with the incoming material and is beginning to rigidify. Debris from Earth-colliding planetesimals is captured by the Earth and rotational fission does not occur. Much of the debris from collision with the Moon ends up on Earth. Bodies perturbed by Jupiter arrive at such a high velocity that the Moon retains very little of the collision debris. (c) Present Earth segregated into a solid inner core especially rich in Ni and siderophile elements, a liquid outer core rich in S, P, C, and both chalcophile and siderophile elements, a solid mantle zoned from pyroxene-rich chemistry in the lower mantle to olivine-rich chemistry in the upper mantle, and a complex crust containing most of the large-ion-lithophile elements. The redox state of the mantle inherits the trend of the accreting planetesimals, and is modified by high-pressure effects in Fe-bearing minerals. An extreme model has $(\text{Mg} + \text{Fe})/\text{Si}$ ranging from $c.1$ to $c.2$ as Fe/Mg increases from $c.0.1$ to 0.25 and then falls to $c.0.1$ in the uppermost mantle. Convective overturning would modify this simple trend (which assumes merely gravitational separation of Fe-rich liquid from heavier Mg-rich crystals), and the Fe/Mg variation may be considerably different. The uppermost mantle has 'pyrolite' bulk composition and consists of garnet and spinel peridotites with minor eclogite, harzburgite, and dunite. Plate-tectonic processes would cause major local variations. In the depth range 300–650 km, several mineralogical transitions occur including the incorporation of pyroxene into garnet solid solution and olivine into β -phase. Below 650 km perovskite is the dominant structure type (perhaps in more than one chemical variety), but other structure types including ilmenite should be present. In the lowermost mantle a structure (fluorite type?) denser than perovskite may occur.

tions of the different types of accreting material must satisfy the oxygen isotopic ratios.

As the pressure of the lower mantle passed 100–200 kb (c.2000 to 3000 km radius), low-pressure silicates typical of peridotitic compositions would begin to transform into high-pressure silicates or oxides, unless the growing Earth became totally molten from a catastrophic event. A growing mantle in a relatively cool Earth would increasingly hinder the sinking of Fe,Ni,S-differentiate from accreting material. Then the outer 500 km of the present Earth would be able to retain some chemical features of the late-accreting material. Volcanic activity and crust-mantle recycling would not directly involve the lower mantle which would be below the melting point because of the pressure increase caused by subsequent accretion. Negative diapirs of Fe,S-rich material would sink through the mantle without triggering off complete overturning. Loss of hydrogen would cause oxidation in the outer part of the Earth. Some combination of these factors would allow the siderophile and chalcophile elements of the outer 500 km to be 'controlled by' oxidized iron in silicates coupled with some depletion caused by sinking of sulphide-rich liquid.

This model of cool differentiation of the later stage of the Earth's growth does not preclude an earlier history of hot differentiation. Indeed, the possibility of collision-induced fission (fig. 12a) to produce a proto-Moon must be considered during growth of the first half of the Earth. Alternatively, the Moon may have begun its growth near the Roche stability limit from a massive debris cloud that did not contain material torn off from the Earth in a single catastrophic event. Whatever the origin of the Moon it would accrete together with the Earth after it became established in orbit. Many complex processes would cause differences in the chemistry of material captured by the two bodies. The Moon would be capturing only the low-velocity fraction of the material, and would retain very little of the volatile-rich material from the outer solar system.

The sequence of a hot early Earth and a cool late Earth is plausible qualitatively from the energy viewpoint only if three factors are valid: The early planetesimals captured by the Earth must be hot and the late planetesimals should be relatively cool. Crystal-liquid differentiation should occur continuously during growth so that there is not a single catastrophic overturning during the late stages of accretion. Finally, accretion must be slow enough so that the gravitational energy of accretion can be dissipated sufficiently by radiation from the surface. Detailed calculations are needed to quantify these factors, and it must be emphasized

that the present model relies on chemical evidence to claim that a mantle barrier existed during the final stage of accretion.

After completion of the major phase of accretion, volcanic activity at the surface would still be intense, and would be triggered locally by the decaying flux of the projectiles that formed basins on the Moon. Major impacts ceased by -3.9×10^9 yr thereby allowing survival of some crustal relics dating back to -3.8×10^9 yr.

Many speculations are possible about the Earth's atmosphere, but the key features for the present model are the major loss of H with concomitant oxidation of C, S, and Fe in the outer part of the Earth, and the establishment of a non-reducing and probably slightly oxidizing atmosphere by -3.8×10^9 yr. Perhaps an intense solar wind caused total dissipation of an early atmosphere, but pronounced loss of volatiles from hot planetesimals must also be considered in models for the distribution of the noble gases.

During the Archaean and Proterozoic eras the volcanic processes would simmer down, and at some stage plate tectonic events would develop. However, these events did not cause multiple cycling through the lower mantle. The only losses from the lower mantle were gases and some other volatile species, perhaps including alkali elements.

For the future there is the possibility that plate-tectonic processes might evolve into whole-mantle convection in which a secondary differentiation sweeps out material not involved in the primary differentiation. Such material might be dominated by Na-Al-rich components, for which cosmochemically based models of the Earth's composition indicate substantial amounts in the mantle.

Finally, in the distant future, a cooling crust might absorb most of the oceans and atmosphere by conversion of anhydrous minerals into hydroxylated, hydrated, and carbonated ones. Details are difficult to predict because of the effects of diffusion and mechanical instability on the transport phenomena.

From the chemical viewpoint the present model of heterogeneous accretion allows the possibility that the lower mantle has a lower Fe content than the upper mantle because of the lower content of Fe in silicates from reduced planetesimals than from oxidized planetesimals and because of the concentration of Fe into rising liquid overlying Fe-poor crystal cumulates; however, this simple conclusion would be modified by gravitational overturning before the lower mantle became too rigid. The lower mantle would not have the same composition as the upper mantle, and the garnet-peridotite and pyrolite models would apply only to the latter. A pyroxene-rich composition for the

lower mantle allows models for the bulk Earth in which the Mg/Si ratio is close to that for C1 meteorite. The Earth's core would be reduced rather than oxidized, and the major light elements would be S and perhaps C and P rather than O.

Other terrestrial planets

There is little to add for the Moon, except to emphasize that it may have started as a fission product from the Earth (induced by disintegrative glancing capture of a large planetesimal from the Mars side of the Earth?), and proceeded to accrete simultaneously with the Earth. The lower abundance of iron, siderophile elements, and volatile elements can be explained by a combination of processes. Because the Earth and Moon have different chemical compositions, it is necessary to explain the near-identity of oxygen isotopic ratios by fortuitous averaging of different mixtures of accreting material; perhaps the higher content of C2-like material in the Earth is balanced by a concomitant amount of material resembling ordinary chondrites.

If this proposed synthesis is correct, Mercury and Venus must have accreted reduced material from the inner solar system, rather than the oxidized material assumed by earlier workers such as Lewis.

Under reduced conditions Fe,Ni-rich alloy can condense before Ca,Al,Ti-oxides and Mg-silicates (Table III), thus providing a simple explanation for the high content of Fe in Mercury. However, this simple idea is almost certainly complicated by mechanical fractionation of metal and silicate at the planetesimal stage, especially because of the high collision velocities near the Sun. Furthermore, the Mercurian crust might have been partly spalled off by late high-speed impacts. The bulk composition in Table IV is based on oxidized condensates, and should be replaced by one based on reduced condensates. There is ample scope for speculation on the detailed composition of the core, mantle, and crust of Mercury. Melting studies are needed for bulk compositions in which the proportions of Ca,Al,Ti-oxides and Mg-silicates are changed. The core should be low in S, and might contain substantial amounts of C, P, N, and perhaps even Si; perhaps the Ni content is higher than 5% because of incomplete reaction of early condensate (c.20% Ni) with remaining gas. Low-Fe pyroxene, spinel, and perovskite should be major minerals in the mantle, and garnet may also be important. Volcanic rocks should have unusual low-Fe compositions undersaturated in silica. Unfortunately the Fe-free minerals have featureless reflectance spectra, which will inhibit chemical

characterization by remote sensing. Furthermore, the Mercurian surface may contain some Fe-rich material projected into the innermost solar system.

Venus may have had a similar history to the Earth except for capture of all the orbiting debris (and probably a retrograde moon), and a higher surface temperature. Perhaps Venus lost nearly all its hydrogen with concomitant oxidation of the atmosphere, crust, and perhaps outer mantle. However, the deep interior may be reduced. Although the bulk composition of Venus may be generally similar to that of the Earth, many speculations about details are possible, including the amount of S in the core, and the extent of metal-silicate mechanical fractionation prior to accretion. Venus should have received rather less C2-like material than the Earth, and measurements of greater accuracy are needed to determine if it has a lower K/U ratio. Because Venus is smaller than the Earth a greater fraction of the mantle occurs as low-pressure silicates rather than as high-pressure phases. Sulphides should be absent in the crust and probably in the upper mantle, and the sulphur should occur at the surface as sulphates and perhaps as feldspathoids. Granitic and basaltic rocks should be less abundant on Venus than on Earth. The mantle should contain less Fe than the terrestrial one if the planetesimals in the Venusian zone were more reduced than in the terrestrial zone.

Predictions for Mars are very uncertain because it may lie near the transition zone from reduced to oxidized planetesimals, and because it is strongly affected by gravitational effects from Jupiter. The present main-belt asteroids are merely the remnants of a larger population, and at least some asteroids have moved either towards or away from the Sun since initial aggregation. I assume that the over-all distinction between an inner region dominated by Siliceous or Stony-Iron asteroids and an outer region dominated by Carbonaceous(?) ones is a relic of primordial accretion, while the unusual asteroids (e.g. E asteroids and Vesta) have changed position. Perhaps the E asteroids originally formed inside the orbit of Mars and were perturbed by Earth and Venus. If the above assumption is correct, Mars accreted mainly from S-asteroids, whose assumed stony-iron nature should indicate early melting, destruction, agglomeration, and metamorphism based on the evidence from stony-iron meteorites. If so, Mars should have a fairly reduced interior since the stony-iron meteorites contain no ferric iron, and contain sulphide rather than sulphate (the sulphide is mainly troilite). Mars should have accreted a small amount of material from the outer part of the main belt of asteroids, and detailed mineralogical

identification of the C-asteroids is badly needed. For the present I assume that the C-asteroids tend to be closer to C₂ and C₃ meteorites than to the highly oxidized C₁ types, which are tentatively assigned to the outer solar system. The oxidation state of the Martian crust may be governed by a combination of loss of hydrogen and of inheritance of the oxidation state of C-asteroids. This assumes that Mars did not become completely molten, or did not melt sufficiently for the whole body to equilibrate. Perhaps the lavas forming the huge volcanoes are reduced, except for the surface, which has reacted with the oxidizing atmosphere. This framework is opposed to the concept of a highly oxidized Mars with a magnetite core. Rather, I would prefer to explore the idea of a core rich in Fe, Ni, and S, and a mantle with a lower average Fe content than envisaged by other scientists. For a partly reduced Mars the moment of inertia would be lower (*c.o.* 37) than the present suggested value (0.377).

CONCLUSION

Looking forward to the twentieth century there are plenty of opportunities for planetary exploration by the members of the Mineralogical Society of Great Britain and Ireland. In spite of the great advances of the past two decades there are many uncertainties in our understanding even of the Earth and Moon. Exploration techniques are now developed sufficiently to allow determination of the chemistry, mineralogy, and petrology of the surfaces of all the terrestrial planets, but budget restrictions will limit the rate of advance. Badly needed to control theoretical speculation are seismic velocity and density profiles as well as accurate values of the moments-of-inertia. Perhaps the greatest mineralogical advances in the next fifty years will come from coupled study of meteorites and asteroids, while a new era will open up from close-range studies of the satellites of the giant planets, for which long-range studies indicate remarkable properties (e.g. alkali emission from Io).

But no matter how many observations are made most of the detailed information about the early solar system has disappeared. Furthermore, there seems little hope of obtaining definitive answers to the depth profiles of minor and trace elements in planets. We must be content with plausible answers

to many questions, though we need not be as pessimistic as Ecclesiastes: 'That which is far off and exceeding deep, who can find it out?' We can feel confident that we are slowly working towards an over-all understanding of the development of the solar system, and that the great Architect has not placed barriers to our scientific exploration of the principal features.

Returning to *Paradise Lost*, Book VIII, 66-84, let us ponder how John Milton's words pertain to our studies:

'To ask or search I blame thee not, for Heav'n
Is as the Book of God before thee set,
Wherein to read his wond'rous Works and learn
His Seasons, Hours, or Days, or Months, or
Years:

This to attain, whether Heav'n move or Earth,
Imports not, if you reck'n right; the rest
From Man or Angel the great Architect
Did wisely to conceal, and not divulge
His secrets to be scann'd by them who ought
Rather admire; or if they list to try
Conjecture, he his Fabric of the Heav'n's
Hath left to thir disputes, perhaps to move
His laughter at thir quaint Opinions wide
Hereafter, when they come to model Heav'n
And calculate the Stars, how they will wield
The mighty frame, how build, unbuild, contrive
To save appearances, how gird the Sphere
With Centric and Eccentric scribb'l'd o'er,
Cycle and Epicycle, Orb in Orb.'

Acknowledgements. It is impossible to list all those scientists who provided the basic knowledge used in this lecture, ranging from my mentors in crystallography and mineralogy at Cambridge University to my colleagues at the Geophysical Laboratory, Pennsylvania State University, and the University of Chicago who taught me the rudiments of geochemistry and meteoritics. A first draft of this lecture was criticized constructively by the following, who must not be blamed for shortcomings in this revised version: T. J. Ahrens, R. Brett, T. E. Bunch, R. N. Clayton, M. J. Drake, D. M. Hunten, W. M. Kaula, B. Mason, E. R. D. Scott, D. M. Shaw, S. C. Solomon, G. W. Wetherill, and J. A. Wood. A particular debt is owed to M. H. Hey for editorial and scientific advice. Irene Baltuska typed the manuscripts with her usual skill, and Brenda Smith improved the style. During the past ten years in which the present ideas simmered, financial support was provided by the National Aeronautics and Space Administration and the National Science Foundation for which current grants are NASA 14-001-171 and NSF 76-03604.

References

- Aannestad (P.A.), 1975. Absorptive properties of silicate core-mantle grains. *Astrophys. J.* **200**, 30-41.
- Adams (J.B.), 1975. Interpretation of visible and near-infrared diffuse reflectance spectra of pyroxenes and other rock-forming minerals. In *Infrared and Raman Spectroscopy of Lunar and Terrestrial Minerals*, 91-116; ed. Karr (C.), New York (Academic Press).
- , and McDowell (T.B.), 1976. Mercury: evidence for an anorthositic crust from reflectance spectra. *Lunar Science Institute Contrib.* **262**, 1.
- Ahrens (T.J.), 1976. Shock wave data for pyrrhotite and constraints on composition of the outer core. *U.S.-Japan Seminar on High-Pressure Research Applications in Geochemistry*, Honolulu, Abstracts, p.31.
- , 1978. Equations of state of iron sulfide and constraints on the sulfur content of the Earth. Ms.
- Akimoto (S.) and Akaogi (M.), 1977. Pyroxene-garnet solid solution equilibria in the Earth's mantle. *Second Int. Kimberlite Conf.*, Ext. Abstr., unpag.
- , Yanamoto (K.) and Aoki (K.), 1977. Hydroxyl-clinochumite and hydroxyl-chondrodite: possible H₂O-bearing minerals in the upper mantle, 163-72. In *High-pressure Research: Applications in Geophysics*, eds. Mangham (M.) and Akimoto (S.), New York (Academic Press).
- Allen (M.) and Robinson (G.W.), 1975. Formation of molecules on small interstellar grains. *Astrophys. J.* **195**, 81-90.
- Anders (E.), 1964. Origin, age and composition of meteorites. *Space Sci. Rev.* **3**, 583-714.
- , 1971a. How well do we know "Cosmic" abundances? *Geochim. Cosmochim. Acta* **35**, 16-22.
- , 1971b. Meteorites and the early solar system. *Ann. Rev. Astron. Astrophys.* **9**, 1-34.
- , 1975. Do stony meteorites come from comets? *Ioanus* **24**, 363-71.
- , 1977. Chemical compositions of the Moon, Earth, and eucrite parent body. *Phil. Trans. R. Soc. London A* **285**, 23-40.
- , 1978. Most stony meteorites come from the asteroid belt. In *Asteroids: An Exploration Assessment*, NASA Special Paper, to be published.
- , and Owen (T.), 1977. Mars and Earth: origin and abundance of volatiles. *Science* **198**, 453-65.
- , Higuchi (H.), Ganapathy (R.) and Morgan (J.W.), 1976. Chemical fractionations in meteorites - IX.33 chondrites. *Geochim. Cosmochim. Acta* **40**, 1131-9.
- Anderson (A.T.), 1974. Chlorine, sulfur, and water in magmas and oceans. *Geol. Soc. Amer. Bull.* **85**, 1485-92.
- Anderson (D.L.), 1972. Internal constitution of Mars. *J. Geophys. Res.* **77**, 789-95.
- , 1973. The composition and origin of the Moon. *Earth Planet. Sci. Lett.* **18**, 301-16.
- , 1977. Composition of the mantle and core. *Ann. Rev. Earth Planet. Sci.* **5**, 179-202.
- , Sammis (C.) and Jordan (T.), 1971. Composition and evolution of the mantle and core. *Science* **171**, 1103-12.
- Aronson (J.R.) and Emslie (A.G.), 1975. Composition of the Martian dust as derived by infrared spectroscopy from Mariner 9. *J. Geophys. Res.* **80**, 4925-31.
- Arrhenius (G.) and Alfven (H.), 1971. Fractionation and condensation in space. *Earth Planet. Sci. Lett.* **10**, 253-67.
- , and Asunuma (S.K.), 1973. Aggregation of grains in space. *The Moon* **8**, 368-91.
- , and De (B.R.), 1973. Equilibrium condensation in a solar nebula. *Meteoritics* **8**, 297-313.
- Ashworth (J.R.), 1977. Matrix textures in unequilibrated ordinary chondrites. *Earth Planet. Sci. Lett.* **35**, 25-34.
- , and Barber (D.J.), 1976. Lithification of gas-rich meteorites. *Earth Planet. Sci. Lett.* **30**, 222-33.
- Baedecker (P.A.) and Wasson (J.T.), 1975. Elemental fractionations among enstatite chondrites. *Geochim. Cosmochim. Acta* **39**, 735-65.
- Baird (A.K.), Toulmin (P., III), Clark (B.C.), Rose (H.J., Jr.), Keil (K.), Christian (R.P.) and Gooding (J.L.), 1977. Mineralogic and petrologic implications of Viking geochemical results from Mars: interim report. *Science* **194**, 1289-93.
- Baldwin (R.B.), 1963. *The Measure of the Moon*. Chicago (University of Chicago Press).
- Bass (M.N.), 1971. Montmorillonite and serpentine in Orgueil meteorite. *Geochim. Cosmochim. Acta* **35**, 139-47.
- Bastin (J.A.) and Butt (R.V.), 1975. Latitude effects in lunar thermal evolution. *Science* **191**, 25-27.
- Bauman (A.J.), Devaney (J.R.) and Bollin (E.M.), 1973. Allende meteorite carbonaceous phase: intractable nature and scanning electron morphology. *Nature* **241**, 264-7.
- Beck (C.W.) and LaPaz (L.), 1951. The nortonite fall and its mineralogy. *Am. Mineral.* **36**, 45-59.
- Bence (A.E.) and Burnett (D.S.), 1969. Chemistry and mineralogy of the silicates and metal of the Kodakana meteorite. *Geochim. Cosmochim. Acta* **33**, 387-407.
- Bergemann (F.) and Wlotzka (F.), 1969. Shock induced thermal metamorphism and mechanical deformations in the Rumdort chondrite. *Geochim. Cosmochim. Acta* **33**, 1351-70.
- Berkley (J.L.), Brown (H.G., IV), Keil (K.), Carter (N.L.), Mercier (J.-C.C.) and Huss (G.), 1976. The Kenna ureilite: an ultramafic rock with evidence for igneous, metamorphic, and shock origin. *Geochim. Cosmochim. Acta* **40**, 1429-37.
- , Taylor (G.J.) and Keil (K.), 1978. Ureilites: origin as related magmatic cumulates. *Lunar and Planetary Sci.* **19**, 73-5.
- Best (M.J.), 1974. Mantle-derived amphibole within inclusions in alkalic-basaltic lavas. *J. Geophys. Res.* **79**, 2107-13.
- Bienemann (K.) and others, 1977. The search for organic substances and inorganic volatile compounds in the surface of Mars. *J. Geophys. Res.* **82**, 4641-58.
- Bild (R.W.), 1974. New occurrences of phosphates in iron meteorites. *Contrib. Mineral. Petrol.* **45**, 91-8.
- , 1977. Silicate inclusions in group IAB irons and a relation to the anomalous stones Winona and Mt. Morris (Wis.). *Geochim. Cosmochim. Acta* **41**, 1439-56.
- , and Wasson (J.T.), 1976. The Lodran meteorite and its relationship to the ureilites. *Mineral. Mag.* **40**, 721-35.
- , and Wasson (J.T.), 1977. Metchaëvo, a new class of chondritic meteorite. *Science* **197**, 58-62.
- Binder (A.B.), 1969. Internal structure of Mars. *J. Geophys. Res.* **74**, 3110-18.
- , 1976. On the petrology and early development of the crust of a moon of fission origin. *The Moon* **15**, 275-314.
- , and Davis (D.R.), 1973. Internal structure of Mars. *Phys. Earth Planet. Int.* **7**, 477-85.
- , and Voss (J.), 1978. The moon's internal structure and its compositional and seismic implications. *Lunar and Planetary Sci.* **19**, 97-9.
- , and 7 others, 1977. The geology of the Viking Lander 1 sites. *J. Geophys. Res.* **82**, 4439-51.
- Bins (R.A.), 1968. Cognate xenoliths in chondritic meteorites: examples in Mezo-Madras and Ghabara. *Geochim. Cosmochim. Acta* **32**, 299-317.
- , and Groves (D.I.), 1976. Iron-nickel partition in metamorphosed olivine-sulfide assemblages from Perseverance, Western Australia. *Am. Mineral.* **61**, 189-7.
- Binz (C.M.), Kurimoto (R.K.) and Lipschutz (M.E.), 1974. Trace elements in primitive meteorites - V. Abundance patterns of thirteen trace elements and interelement relationships in enstatite chondrites. *Geochim. Cosmochim. Acta* **38**, 1579-1606.
- , Ikramuddin (M.) and Lipschutz (M.E.), 1975. Contents of eleven trace elements in ureilite achondrites. *Geochim. Cosmochim. Acta* **39**, 1576-9.
- , Ikramuddin (M.), Rey (P.) and Lipschutz (M.E.), 1976. Trace elements in primitive meteorites - VI. Abundance patterns of thirteen trace elements and interelement relationships in unequilibrated ordinary chondrites. *Geochim. Cosmochim. Acta* **40**, 59-71.
- Bishop (F.C.), Smith (J.V.) and Dawson (J.B.), 1975. Pentlandite-magnetite intergrowth in DeBeers spinel therzolit: review of sulphides in nodules. *Phys. Chem. Earth* **9**, 323-37.
- , and others, 1978. Na, K, P and Ti in garnet, pyroxene and olivine from peridotite and eclogite xenoliths from African kimberlites. *Lithos* **11**, 155-73.
- Blanchard (M.B.), Brownlee (D.E.), Bunch (T.E.), Hodge (P.W.) and Kyte (F.T.), 1978. Spheres from deep sea sediments: a way to study meteoroid compositions with time. *Barek Planet. Sci. Lett.* To be submitted.
- Blander (M.) and Abdel-Gawad (M.), 1969. The origin of meteorites and the constrained equilibrium condensation theory. *Geochim. Cosmochim. Acta* **33**, 701-16.
- , and Fuchs (L.H.), 1975. Calcium-aluminum-rich inclusions in the Allende meteorite: evidence for a liquid origin. *Geochim. Cosmochim. Acta* **39**, 1605-19.
- , and Planner (H.N.), Keil (K.), Nelson (L.S.) and Richardson (N.L.), 1976. The origin of chondrules: experimental investigation of metastable liquids in the system MgSiO₃-SiO₂. *Geochim. Cosmochim. Acta* **40**, 889-96.
- Blasius (K.R.), 1976. The record of impact cratering on the great volcanic shields of the Tharsis region of Mars. *Ioanus* **25**, 343-61.
- , Cutts (J.A.), Guest (J.E.) and Masursky (J.), 1977. Geology of the Valles Marineris: first analysis of imaging from the Viking 1 orbiter imaging mission. *J. Geophys. Res.* **82**, 4067-91.
- Bocor (N.Z.), Meyer (H.O.A.) and Kullerød (G.), 1976. Lafayette meteorite: petrology and opaque mineralogy. *Earth Planet. Sci. Lett.* **32**, 69-76.
- Booth (M.C.) and Kieffer (H.H.), 1978. Carbonate formation in Marslike environments. *J. Geophys. Res.* **83**, 1809-15.
- Boyd (F.R.), 1973. A pyroxene geotherm. *Geochim. Cosmochim. Acta* **37**, 2533-46.
- , and McCallister (R.H.), 1976. Densities of fertile and sterile garnet peridotites. *Geophys. Res. Lett.* **3**, 509-12.
- , England (J.L.) and Davis (B.T.C.), 1964. Effects of pressure on the melting and polymorphism of enstatite, MgSiO₃. *J. Geophys. Res.* **69**, 2101-09.
- Boynton (W.V.), Starzyk (P.M.) and Schmitt (R.A.), 1976. Chemical evidence for the genesis of the ureilites, the achondrite Chassigny and the nakhlites. *Geochim. Cosmochim. Acta* **40**, 1439-47.
- Braun (W.von) and Ordey (F.J., III), 1975. *History of Rocketry and Space Travel*. 3rd Ed., New York (Thomas Crowell).
- Brett (R.), 1967. Coherent: its occurrence and a proposed origin. *Geochim. Cosmochim. Acta* **31**, 143-59.
- , 1976. The current status of speculations on the composition of the core of the earth. *Rev. Geophys. Space Phys.* **14**, 375-83.
- , 1977. The case against early melting of the bulk of the Moon. *Geochim. Cosmochim. Acta* **41**, 443-5.
- Brey (G.) and Green (D.H.), 1977. Petrogenesis of olivine melilitite and kimberlite and melting of peridotite-C-O-H. *Second Int. Kimberlite Conf.*, Ext. Abstr., unpag.
- Brownlee (D.E.), Tomand (D.), Blanchard (M.B.), Ferry (G.V.) and Kyte (F.T.), 1976. *An Atlas of Extraterrestrial Particles collected with NASA U-2 Aircraft*. NASA TMX 73, 152.
- , and Olszewski (E.), 1977. Interplanetary dust: A new source of extraterrestrial material for laboratory study. *Proc. Lunar Sci. Conf.* **8th**, 149-60.
- Buchwald (V.F.), 1975. *Iron Meteorites*. Berkeley (University of California Press).
- Burch (T.E.), 1975. Petrography and petrology of basaltic achondrite polymict breccias (howardites). *Proc. Lunar Sci. Conf.* **8th**, 469-92.
- , and Chang (S.), 1978a. Carbonaceous chondrite (CM) phyllosilicates: condensation or alteration origin? *Lunar and Planetary Sci.* **19**, 134-6.
- , and others, 1978b. Carbonaceous chondrite phyllosilicates and their bearing on parent body surface conditions. *Geochim. Cosmochim. Acta*. To be submitted.
- , and Fuchs (L.H.), 1969. A new mineral, breznite Cr₃S₄, and the Tucson meteorite. *Am. Mineral.* **54**, 1509-18.
- , and Olsen (E.), 1968. Potassium feldspar in Wekeroo Station, Kodakana, and Colomera iron meteorites. *Science* **160**, 1223-5.
- , and Reid (A.M.), 1975. The nakhlites Part I: Petrography and mineral chemistry. *Meteoritics* **10**, 303-15.
- , and Stöffler (D.), 1974. The Kelly chondrite: a parent body surface metabreccia. *Contr. Mineral. Petrol.* **44**, 157-71.
- , Chang (S.) and Frick (U.), 1978. Characterization and significance of carbonaceous chondrite (CM) xenoliths in the Jodzie howardite. *Geochim. Cosmochim. Acta*, submitted.
- , Keil (K.) and Olsen (E.), 1970. Mineralogy and petrology of silicate inclusions in iron meteorites. *Contr. Mineral. Petrol.* **25**, 297-340.
- , and Snetinger (K.G.), 1967. Chromite composition in relation to chemistry and texture of ordinary chondrites. *Geochim. Cosmochim. Acta* **31**, 1569-82.
- Burns (J.A.), 1973. Where are the satellites of the inner planets? *Nature* **242**, 23-5.
- Burns (R.G.), 1974. Partitioning of transition metals in mineral structures of the mantle, 555-72. In *Physics and Chemistry of Minerals and Rocks*, ed. Strens (R.G.), Chichester, England (John Wiley).
- Buseck (P.R.), 1977. Pallisite meteorites-mineralogy, petrology and geochemistry. *Geochim. Cosmochim. Acta* **41**, 711-40.
- , and Goldstein (J.I.), 1969. Olivine compositions and cooling rates of palisitic meteorites. *Geol. Soc. Amer. Bull.* **80**, 2141-55.
- , and Holdsworth (E.), 1977. Phosphate minerals in palisite meteorites. *Mineral. Mag.* **41**, 91-102.
- Cameron (A.G.W.), 1975. Clumping of interstellar grains during formation of the primitive solar nebula. *Ioanus* **24**, 128-33.
- , and Pine (R.R.), 1973. Numerical models of the primitive solar nebula. *Ioanus* **18**, 377-406.
- , and Truran (J.W.), 1977. The supernova trigger for formation of the solar system. *Ioanus* **30**, 447-61.

- , and Ward (W.R.), 1976. The origin of the moon. *Lunar Sci.* 7, 120-2.
- Campbell (D.B.), Byer (R.B.), and Pettengill (G.H.), 1976. New radar image of Venus. *Science* 193, 1123-4.
- Carmichael (I.S.E.), Turner (F.J.), and Verhoogen (J.), 1974. *Igneous Petrology*. New York (McGraw-Hill).
- Carr (M.H.) and Schaber (G.G.), 1977. Martian permafrost features. *J. Geophys. Res.* 82, 4339-54.
- , and 12 others, 1977a. Preliminary results from the Viking Orbiter imaging experiment. *Science* 193, 766-76.
- , Greeley (R.), Blasius (K.R.), Guest (J.E.), and Murray (J.B.), 1977b. Some Martian volcanic features as viewed from the Viking orbiters. *J. Geophys. Res.* 82, 3985-4015.
- , Masursky (H.), 1977c. Martian impact craters and emplacement of ejecta by surface flow. *J. Geophys. Res.* 82, 4055-65.
- Carswell (D.A.), 1975. Primary and secondary phlogopites and clinopyroxenes in garnet hercynite xenoliths. *Phys. Chem. Earth* 9, 417-29.
- Cassen (P.), Young (R.E.), Schubert (G.), and Reynolds (R.T.), 1976. Implications of an internal dynamo for the thermal history of Mercury. *Icarus* 28, 501-8.
- Chapman (C.R.), Jones (K.L.), 1977. Cratering and obliteration history of Mars. *Ann. Rev. Earth Planet. Sci.* 5, 515-40.
- , 1976. Asteroids as meteorite parent bodies: the astronomical perspective. *Geochim. Cosmochim. Acta* 40, 701-19.
- , Morrison (D.), and Zellner (B.), 1975. Surface properties of asteroids: a synthesis of polarimetry, radiometry and spectrophotometry. *Icarus* 25, 104-30.
- Chou (C.-L.) and (A.J.), 1973. Gallium and germanium in the metal and silicates of L- and LL-chondrites. *Geochim. Cosmochim. Acta* 37, 315-27.
- , Baedeker (P.A.), and Nasson (J.T.), 1976. Allende inclusions: volatile-element distribution and evidence for incomplete volatilization of presolar solids. *Geochim. Cosmochim. Acta* 40, 85-94.
- Christophe Miché (M.), 1976. Le météorite "E. blanche de la chondrite de Tieschitz (H3)". *Earth Planet. Sci. Lett.* 30, 143-50.
- Clark (B.C.) and 7 others, 1977. Inorganic analyses of Martian surface samples at the Viking landing sites. *Science* 194, 1283-8.
- Clark (S.P., Jr.), 1966. Ed. *Handbook of Physical Constants*. Geol. Soc. America, Mem.
- , Turkian (K.K.), and Grossman (L.), 1972. Model for the early history of the Earth, 3-18, in *The Nature of the Solid Earth*, ed. E.C. Robertson, New York (McGraw-Hill).
- Clarke (R.S., Jr.) and Goldstein (J.I.), 1978. Schreibersite growth and its influence on the metallography of coarse-structured iron meteorites. *Smithsonian Contrib. Earth Sci.* no. 21, 80pp.
- Clayton (D.D.), 1977a. Solar system isotopic anomalies: supernova neighbor or presolar carriers. *Icarus* 32, 255-69.
- , 1977b. Cosmordogenic ghosts and the origin of Ca-Al-rich inclusions. *Earth Planet. Sci. Lett.* 33, 398-410.
- Clayton (R.N.), 1978. Isotope anomalies in the early solar system. *Am. Rev. Nuclear Sci.*, submitted.
- , and Mayeda (T.K.), 1977. Correlated oxygen and magnesium isotope anomalies in Allende inclusions. I. Oxygen. *Geophys. Res. Letters* 4, 295-6.
- , and Mayeda (T.K.), 1978. Genetic relations between iron and stony meteorites. *Earth Planet. Sci. Lett.* 40, 168-174.
- , Onuma (N.), and Mayeda (T.K.), 1976. A classification of meteorites based on oxygen isotopes. *Earth Planet. Sci. Lett.* 30, 10-18.
- , Grossman (L.), and Mayeda (T.K.), 1977. Distribution of the pre-solar component in Allende and other carbonaceous chondrites. *Earth Planet. Sci. Lett.* 33, 209-24.
- Consolmagno (G.), and Drake (M.J.), 1977. Composition and evolution of the eucrite parent body: evidence from rare earth elements. *Geochim. Cosmochim. Acta* 41, 1271-82.
- Counselman (C.C.), 1973. Outcomes of tidal evolution. *Astrophys. J.* 180, 307-14.
- Craig (H.) and Lupton (J.E.), 1976. Primordial neon, helium and hydrogen in oceanic basalts. *Earth Planet. Sci. Lett.* 31, 369-85.
- Curtis (S.A.) and Hartle (R.E.), 1977. Wind enhanced planetary escape: collisional modifications. *J. Geophys. Res.* 82, 1206-10.
- Cutts (J.A.), Blasius (K.R.), Briggs (G.A.), Carr (M.H.), Greeley (R.), and Masursky (H.), 1977. North polar region of Mars: imaging results from Viking 2. *Science* 194, 1329-37.
- Daly (S.F.) and Richter (F.M.), 1978. Convection with decaying heat sources: a simple thermal evolution model. *Lunar and Planetary Science* 19, 213-4.
- Darwin (G.H.), 1906. On the figure and stability of a liquid satellite. *Phil. Trans. R. Soc. London* 206, 161-248.
- Davies (G.F.), 1974. Limits on the constitution of the lower mantle. *Geophys. J. R. Astron. Soc.* 38, 479-503.
- , 1977. Whole-mantle convection and plate tectonics. *Geophys. J. Astron. Soc.* 49, 459-86.
- Davis (A.M.), Ganapathy (R.), and Grossman (L.), 1977. Pontilyni: a differentiated meteorite related to the group IAB irons. *Earth Planet. Sci. Lett.* 35, 19-24.
- Dawson (J.B.) and Smith (J.V.), 1975. Occurrence of diamond in a mica-garnet hercynite xenolith from Kimberlite. *Austral. J. Geol.* 25, 580-1.
- , and Stephens (W.E.), 1975. Statistical classification of garnets from kimberlite and associated xenoliths. *J. Geol.* 83, 589-607.
- Day (K.L.), 1975. Measured extinction of small olivine spheres. *Astrophys. J.* 199, 660-2.
- Degewij (C.), and Zellner (B.), 1978. Asteroid surface variegation. *Lunar and Planetary Sci.* 19, 235-7.
- Delsemme (A.H.), 1977. Ed. of *Comets-Asteroids-Meteorites*, Toledo, Ohio (University of Toledo Press).
- Desnoyers (C.) and Jerome (D.Y.), 1977. The Malvern howardite: a petrological and chemical discussion. *Geochim. Cosmochim. Acta* 41, 81-6.
- Doan (A.S.) and Goldstein (J.I.), 1969. The formation of phosphides in iron meteorites, 763-79. In *Meteorite Research*, ed. P.M. Millman, Dordrecht, Holland (D. Reidel).
- Dobrovolskis (A.) and Ingersoll (A.P.), 1975. Carbon dioxide-water clathrate as a reservoir of CO₂ on Mars. *Icarus* 28, 363-7.
- Odd (R.T.), 1969. Metamorphism of the ordinary chondrites: A review. *Geochim. Cosmochim. Acta* 33, 161-203.
- , 1971. The petrology of chondrules in the Sharps meteorite. *Contrib. Mineral. Petrol.* 31, 201-27.
- , 1973. Minor element abundances in olivines of the Sharps (H-3) chondrite. *Contrib. Mineral. Petrol.* 42, 159-67.
- , 1974. The petrology of chondrules in the Hallingberg meteorite. *Contrib. Mineral. Petrol.* 47, 97-112.
- , 1976. Accretion of the ordinary chondrites. *Earth Planet. Sci. Lett.* 30, 281-91.
- , 1978. The composition and origin of large microporphyrific chondrules in the Manych (L-3) chondrite. *Earth Planet. Sci. Lett.* 39, 52-66.
- , and Jaroszewicz (E.), 1976. Olivine microporphyry in the St. Mesmin chondrite. *Meteoritics* 11, 1-20.
- , Gröner (J.T.), and Brown (G.E.), 1975. Pyroxenes in the Shaw (L-7) chondrite. *Geochim. Cosmochim. Acta* 39, 1585-94.
- Dolgintov (Sh.Sh.), 1978. On the magnetic field of Mars: Mars 5 evidence. *Geophys. Res. Lett.* 5, 93-5.
- Drake (M.J.), 1976. Evolution of major mineral compositions and trace element abundances during fractional crystallization of a model lunar composition. *Geochim. Cosmochim. Acta* 40, 401-11.
- Dreibus (G.), Kruse (H.), Spettel (B.), and Wänke (H.), 1977. The bulk composition of the moon and the eucrite parent body. *Proc. Lunar Sci. Conf.* 8th, 211-27.
- Dufresne (E.R.) and Anders (E.), 1962. On the chemical evolution of the carbonaceous chondrites. *Geochim. Cosmochim. Acta* 26, 1085-114.
- Duke (M.B.), 1968. The Shergotty meteorite: magmatic and shock metamorphic features. In *Shock Metamorphism of Natural Materials*, ed. B.M. French and N.M. Short, 613-21, Baltimore, Md. (Mono Books Corp.).
- , and Silver (L.T.), 1967. Petrology of eucrites, howardites and mesosiderites. *Geochim. Cosmochim. Acta* 31, 1637-65.
- Dyal (P.), Parkin (C.W.), and Daily (W.D.), 1975. Lunar electrical conductivity and magnetic permeability. *Proc. 6th Lunar Sci. Conf.*, 2909-26.
- Dymek (R.F.), Albee (A.L.), Chodos (A.A.), and Wasserburg (G.J.), 1976. Petrography of isotopically dated clasts in the Kapotea howardite and petrologic constraints on the evolution of its parent body. *Geochim. Cosmochim. Acta* 40, 1115-30.
- Dymond (J.), and Hogan (L.), 1978. Factors controlling the noble gas abundance patterns of deep-sea basalts. *Earth Planet. Sci. Lett.* 38, 117-28.
- Dzurim (D.) and Blasius (K.R.), 1975. Topography of the polar layered deposits of Mars. *J. Geophys. Res.* 80, 3285-3306.
- Easton (A.J.) and Hey (M.H.), 1968. Minor elements present in the silicate phase of enstatite chondrites. *Mineral. Mag.* 36, 740-2.
- Eggle (D.H.), 1976. Does CO₂ cause partial melting in the low-velocity layer of the mantle? *Geology* 4, 69-72.
- , 1977. The principle of the zone of invariant vapor composition: an example in the system CaO-MgO-SiO₂-CO₂-H₂O and implications for the mantle solidus. *Carnegie Inst. Washington Yearb.* 75, 426-35.
- Ekström (T.K.), 1972. The distribution of fluorine among some coexisting minerals. *Contrib. Mineral. Petrol.* 34, 192-200.
- El Goresy (A.), 1965. Mineralbestand und Strukturen der Graphit- und Sulfideinschlüsse in Eisenmeteoriten. *Geochim. Cosmochim. Acta* 29, 1131-51.
- , 1967. Quantitative electron microprobe analyses of co-existing sphalerite, daubereite and troilite in the Odessa iron meteorite and their genetic implications. *Geochim. Cosmochim. Acta* 31, 1667-76.
- Erlank (A.J.), 1973. Kimberlite potassic richterite and the distribution of potassium in the upper mantle. *Carnegie Inst. Washington Yearb.* 68, 433-9.
- Falk (S.W.) and Scalo (J.M.), 1975. Behavior of grains in early supernova environments. *Astrophys. J.* 202, 690-5.
- Fanale (F.P.), 1976. Martian volatiles: their degassing history and geochemical fate. *Icarus* 28, 179-202.
- , and Cannon (W.A.), 1978. Mars: the role of the regolith in determining atmospheric pressure and the atmosphere's response to insolation changes. *J. Geophys. Res.* 83, 2321-5.
- , and Owen (T.), 1978. Mars: regolith adsorption and the relative concentrations of atmospheric rare gases. *Geophys. Res. Lett.* 5, 77-80.
- Farmer (C.B.), 1976. Liquid water on Mars. *Icarus* 28, 279-89.
- , Davies (D.W.), and LaPorte (D.D.), 1977. Mars: northern summer ice cap-water vapor observations from Viking 2. *Science* 194, 1339-41.
- Field (G.B.) and Cameron (A.G.W.), eds., 1973. *The Dusty Universe*. New York (Neville-Tamming Academic Publications).
- Finger (L.W.), Hazen (R.M.), and Yagi (T.), 1977. High-pressure crystal structures of the spinel polymorphs of Fe₂SiO₄ and Ni₂SiO₄. *Carnegie Institution of Washington Yearb.* 76, 504-8.
- Fitzgerald (M.J.) and Jones (J.B.), 1977. Adelaide and Bench Crater - members of a new subgroup of the carbonaceous chondrites. *Meteoritics* 12, 443-58.
- Florán (R.J.) and Prinz (M.), 1978. Silicate petrography and classification of the mesosiderites. *Lunar and Planetary Sci.* 19, 329-31.
- , Hlava (P.F.), Keil (K.), Mehr (C.E.) and Gault (J.D.), 1977a. Chassigny revisited: a cumulate dunite with hydrous amphibole-bearing melt inclusions. *Meteoritics* 12, 225-6. (Abstr.)
- , Spettel (B.) and Wänke (H.), 1977b. The Johnston orthopyroxene (diogenite) and its relationship to meteoritic cumulates. *Meteoritics* 12, 226-7. (Abstr.)
- , Impact-melt origin for the Sironium, Hainholz, and Pinnaro mesosiderites. *Lunar and Planetary Sci.* 19, 326-8.
- Florensky (C.P.) and 8 others, 1977. The surface of Venus as revealed by Soviet Venera 9 and 10. *Geol. Soc. Am. Bull.* 88, 1937-45.
- Fodor (R.V.) and Keil (K.), 1976. A kamufite-like lithic fragment with spinifex texture in the Eua meteorite: origin from a supercooled impact melt of chondritic parentage. *Earth Planet. Sci. Lett.* 29, 1-6.
- , Wilkening (L.L.), Bogard (D.D.), and Gibson (E.K.), 1976. Origin and history of a meteorite parent-body regolith breccia: carbonaceous and noncarbonaceous lithic fragments in the Abbot, New Mexico, chondrite. *New Mexico Geol. Soc. Spec. Publ.* no. 6, 206-18.
- Foshag (W.F.), 1940. The Shallower meteorite: a new aubrite. *Am. Mineral.* 25, 779-86.
- Frederiksson (K.) and Keil (K.), 1963. The light-dark structure in the Pantar and Kapotea stone meteorites. *Geochim. Cosmochim. Acta* 27, 1717-39.
- French (B.M.), 1971. How did Venus lose its angular momentum? *Science* 173, 169-70. Reply by S. F. Singer, 170.
- , and Short (N.M.), 1968. Eds. of *Shock Metamorphism of Natural Materials*. New York (Neville-Tamming Academic Publications).
- Frey (B.J.), 1974. Surface features on Mars: ground-based albedo and radar compared with Mariner 9 topography. *J. Geophys. Res.* 79, 3927-16.
- Fricker (P.E.) and Reynolds (R.T.), 1968. Development of the atmosphere of Venus. *Icarus* 9, 221-30.
- Fronzel (J.M.), 1975. *Lunar Mineralogy*. New York (John Wiley).
- Fryer (B.J.), 1977. Rare earth evidence in iron-formations for changing Precambrian oxidation states. *Geochim. Cosmochim. Acta* 41, 361-7.
- Fuchs (L.H.) and Olsen (E.), 1973. Composition of metal in Type III carbonaceous chondrites and its relevance to the source-assignment of lunar metal. *Earth Planet. Sci. Lett.* 18, 379-84.
- , and Henderson (E.P.), 1967. On the occurrence of brinites and panethites, two new phosphate minerals from the Dayton meteorite. *Geochim. Cosmochim. Acta* 31, 1711-9.

- and Jensen (K.J.), 1973. Mineralogy, mineral-chemistry, and composition of the Murchison (C2) meteorite. *Smitthsonian Contr. Earth Sciences*, no. 10.
- Fukunaga (T.), Ma (M.-S.), Wakita (H.) and Schmitt (R.A.), 1978. Lodran: the residue of limited partial melting of matter like a hybrid between H and E chondrites. *Lunar and Planetary Sci. IX*, 356-8.
- Gaffey (M.J.), 1976. Spectral reflectance characteristics of the meteorite classes. *J. Geophys. Res.* **81**, 905-20.
- and McCord (T.B.), 1977. Asteroid surface materials: mineralogical characterizations and cosmological implications. *Proc. Lunar Sci. Conf. 8th*, 113-43.
- and , 1978. Asteroid surface materials: mineralogical characterizations from reflectance spectra. *Space Science Reviews*, submitted.
- Gale (N.H.), Arden (J.W.) and Hutchison (R.), 1975. The chronology of the Nakhla achondritic meteorite. *Earth Planet. Sci. Lett.* **26**, 195-206.
- Ganapathy (R.) and Anders (E.), 1974. Bulk compositions of the moon and earth, estimated from meteorites. *Proc. Lunar Sci. Conf. 5th*, 1181-1206.
- Gancarz (A.J.) and Wasserburg (G.J.), 1977. Initial Pb of the Amitsoq gneiss, West Greenland, and implications for the age of the Earth. *Geochim. Cosmochim. Acta* **41**, 1283-1301.
- Ganguly (J.) and Kennedy (G.C.T.), 1977. Solubility of K in Fe-S liquid, silicate-K-(Fe-S)₁₉ equilibria, and their planetary implications. *Earth Planet. Sci. Lett.* **35**, 411-20.
- Gibson (E.K., Jr.) and Moore (C.B.), 1971. The distribution of total nitrogen in iron meteorites. *Geochim. Cosmochim. Acta* **35**, 877-90.
- and Moore (G.W.), 1973. Carbon and sulfur distributions and abundances in lunar fines. *Proc. Fourth Lunar Sci. Conf.* **2**, 1577-86.
- Gilbert (F.) and Dziewonski (A.M.), 1975. An application of normal mode theory to the study of structural and source mechanisms from seismic spectra. *Phil. Trans. R. Soc. London* **27B**, 187-269.
- Gilman (R.C.), 1969. On the composition of interstellar grains. *Astrophys. J.* **155**, L185-7.
- Goettel (K.A.), 1974. Potassium in the Earth's core: evidence and implications. In *Physics and Chemistry of Minerals and Rocks*, ed. Strens (R.G.J.), London (John Wiley).
- Goins (N.R.), Dainty (A.M.) and Toksöz (M.N.), 1977. The deep seismic structure of the Moon. *Lunar Sci. Conf. 8th*, 471-86.
- Goldreich (P.) and Ward (W.R.), 1973. The formation of planetesimals. *Astrophys. J.* **181**, 1051-61.
- Goldsmith (J.R.), 1978. Scapolites, granulites, and volatiles in the lower crust. *Geol. Soc. Am. Bull.* **87**, 161-8.
- and Newton (R.C.) and Moore (P.B.), 1974. Silicate-nitrate compounds: high-pressure synthesis and stability of a nitrate scapolite. *Am. Mineral.* **59**, 768-74.
- Goldstein (J.I.) and Axon (H.J.), 1973. The Widmanstätten figure in iron meteorites. *Nature* **240**, 313-21.
- and Doan (A.S., Jr.), 1972. The effect of phosphorus on the formation of the Widmanstätten pattern in iron meteorites. *Geochim. Cosmochim. Acta* **36**, 51-69.
- and Short (J.W.), 1967. The iron meteorites, their thermal history and parent bodies. *Geochim. Cosmochim. Acta* **31**, 1733-70.
- Goldstein (R.M.), Green (R.R.) and Runney (H.C.), 1976. Venus radar images. *J. Geophys. Res.* **81**, 4807-17.
- Gooding (J.L.), 1976. Chemical weathering on Mars. Thermodynamic stability of primary minerals (and their alteration products) from mafic igneous rocks. *Icarus* **33**, 483-513.
- Gooley (R.) and Moore (C.B.), 1976. Native metal in diogenite meteorites. *Am. Mineral.* **61**, 373-8.
- Gradie (J.) and Zellner (B.), 1977. Asteroid families: observational evidence for common origins. *Science* **197**, 254-5.
- Graham (A.L.), Hutchison (R.J.) and Jérôme (D.V.), 1976. The Bovey meteorite: mineral chemistry and origin of its Ca-rich glass inclusions. *Geochim. Cosmochim. Acta* **40**, 529-35.
- and Hutchison (R.J.), 1977a. The Mayo Belwa achondrite: a new enstatite achondrite fall. *Mineral. Mag.* **41**, 487-92.
- and , 1977b. Forsterite in chondrites: the meteorites Kakangari, Mount Morris (Wisconsin), Pontlyfni and Winoona. *Mineral. Mag.* **41**, 201-10.
- Greeley (R.), Theilig (E.), Guest (J.E.), Carr (M.H.), Masursky (H.) and Cutts (J.A.), 1977. Geology of Chryse Planitia. *J. Geophys. Res.* **82**, 4093-109.
- Green (D.H.), 1975. Genesis of Archaean peridotitic magmas and constraints on Archaean geothermal gradients and tectonics. *Geology* **3**, 15-8.
- Green (H.W., II), Radcliffe (S.V.) and Heuer (A.H.), 1971. Allende meteorite: a high-voltage electron photomicrographic study. *Science* **172**, 936-9.
- Grossman (L.), 1972. Condensation in the primitive solar nebula. *Geochim. Cosmochim. Acta* **36**, 597-619.
- and , 1975. Petrography and mineral chemistry of Ca-rich inclusions in the Allende meteorite. *Geochim. Cosmochim. Acta* **39**, 433-54.
- and Clark (S.P., Jr.), 1973. High-temperature condensates in chondrites and the environment in which they formed. *Geochim. Cosmochim. Acta* **37**, 635-49.
- and Ganapathy (R.), 1976. Trace elements in the Allende meteorite - I. Coarse-grained, Ca-rich inclusions. *Geochim. Cosmochim. Acta* **40**, 331-44.
- and Larimer (J.W.), 1974. Early chemical history of the solar system. *Rev. Geophys. Space Physics* **12**, 71-101.
- and Olsen (E.), 1974. Origin of the high temperature fraction of C2 chondrites. *Geochim. Cosmochim. Acta* **38**, 173-87.
- and Steele (I.M.), 1976. Amorphous olivine aggregates in the Allende meteorite. *Geochim. Cosmochim. Acta* **40**, 149-55.
- and Ganapathy (R.) and Davis (A.M.), 1977. Trace elements in the Allende meteorite: III. Coarse-grained inclusions revisited. *Geochim. Cosmochim. Acta* **41**, 1647-64.
- Handbury (M.J.) and Williams (I.P.), 1976. The dispersal of the solar nebula by the solar wind. *The Observatory* **36**, 140-2.
- Hanks (T.C.) and Anderson (D.L.), 1969. The early thermal history of the Earth. *Phys. Earth Planet. Int.* **2**, 19-29.
- Hapke (B.), 1977. Interpretations of optical observations of Mercury and the Moon. *Phys. Earth Planet. Int.* **15**, 264-74.
- Hards (N.J.), 1976. The role of fluorine and chlorine in late-stage magmatic processes: an experimental study. *Mineralogical Society Bulletin* no. 30, 4 (Abstr.).
- Hargraves (R.B.), 1976. Precambrian geologic history. *Science* **193**, 363-71.
- and Collinson (D.W.), Arvidson (R.E.) and Spitzer (C.R.), 1977a. Viking magnetic properties investigation: further results. *Science* **194**, 1303-9.
- and , 1977b. The Viking magnetic properties experiment: primary mission results. *J. Geophys. Res.* **82**, 4547-58.
- Harris (A.W.), 1978. Satellite formation. II. *Icarus* **34**, 128-45.
- and Kaula (W.M.), 1975. A co-accretional model of satellite formation. *Icarus* **24**, 516-24.
- Hartmann (W.K.), 1972. *Moons and Planets*. Tarrytown, New York (Gordon and Quigley).
- and , 1974. Martian and terrestrial paleoclimatology: relevance of solar variability. *Icarus* **22**, 301-11.
- and , 1976. Planet formation: compositional mixing and lunar compositional anomalies. *Icarus* **27**, 553-9.
- and , 1977. Relative crater production rates on planets. *Icarus* **31**, 260-76.
- and , 1978. Planet formation: mechanism of early growth. *Icarus* **33**, 50-61.
- and Davis (D.R.), 1975. Satellite-sized planetesimals and lunar origin. *Icarus* **24**, 504-15.
- Haughton (D.R.), Roeder (P.L.) and Skinner (B.G.), 1974. Solubility of sulfur in mafic magmas. *Econ. Geol.* **69**, 451-67.
- Haynes (R.C.), 1971. *Introduction to Space Sciences*. New York (John Wiley).
- Hazen (R.M.) and Prewitt (C.T.), 1977. Effects of temperature and pressure on interatomic distances in oxygen-based minerals. *Am. Mineral.* **62**, 309-15.
- Head (J.W.), Settle (M.) and Stein (R.S.), 1975. Volume of material ejected from major lunar basins and implications for the depth of excavation of lunar samples. *Proc. Lunar Sci. Conf. 8th*, 2805-29.
- Herndon (J.M.) and Herndon (M.A.), 1977. Aluminum-26 as a planetoid heat source in the early solar system. *Meteoritics* **12**, 459-65.
- and Rowe (M.W.), Larson (E.E.) and Watson (D.E.), 1976. Thermomagnetic analysis of meteorites. 3. C3 and C4 chondrites. *Earth Planet. Sci. Lett.* **33**, 283-90.
- Hertogen (J.), Janssens (M.J.), Takahashi (H.), Palme (H.) and Anders (E.), 1977. Lunar basins and craters: evidence for systematic compositional changes of bombarding population. *Proc. Lunar Sci. Conf. 8th*, 17-45.
- Hervig (R.L.), Smith (J.V.) and Dawson (J.B.), 1977. Minor element content of olivine and orthopyroxene in upper-mantle xenoliths. *Second Int. Kilauea Conf.*, Ext. Abstr., unpag.
- Herzog (G.F.) and Chesley (P.J., Jr.), 1977. Diogenite exposure ages. *Geochim. Cosmochim. Acta* **41**, 127-34.
- Hess (H.H.) and Henderson (E.P.), 1949. The Moore County meteorite: A further study with comment on its primordial environment. *Am. Mineral.* **34**, 494-507.
- Hewins (R.H.) and Klein (L.C.), 1978. Provenance of metal and melt rock textures in the Malvern howardite. *Lunar and Planetary Sci. IX*, 503-5.
- Hey (M.H.) and Easton (A.J.), 1967. The Khor Temiki meteorite. *Geochim. Cosmochim. Acta* **31**, 1789-92.
- Higuchi (H.), Morgan (J.W.), Ganapathy (R.) and Anders (E.), 1976. Chemical fractionations in meteorites - X. Ureilites. *Geochim. Cosmochim. Acta* **40**, 1563-71.
- Hofmann (A.W.) and Hart (S.R.), 1978. An assessment of local and regional isotopic equilibrium in the mantle. *Earth Planet. Sci. Lett.* **38**, 44-62.
- Holland (H.D.), 1965. Some applications of thermochemical data to problems of ore deposits. II. Mineral assemblages and the composition of ore-forming fluids. *Econ. Geol.* **60**, 1101-66.
- Horowitz (N.H.), Hobby (G.L.) and Hubbard (J.S.), 1977. Viking on Mars: the carbon assimilation experiments. *J. Geophys. Res.* **82**, 4659-62.
- Hostetler (C.J.) and Drake (M.D.), 1978. Quench temperatures of Moore County and other eucrites: residence time on eucrite parent body. *Geochim. Cosmochim. Acta* **42**, 517-22.
- Hoyle (F.) and Wickramasinghe (N.C.), 1968. Condensation of the planets. *Nature* **217**, 415-8.
- Huang (W.L.) and Myllie (P.J.), 1974. Melting relations of muscovite with quartz and sanidine in the K₂O-Al₂O₃-SiO₂-H₂O system to 30 kilobars and an outline of paragonite melting relations. *Amer. J. Sci.* **274**, 376-95.
- Huck (F.O.) and 6 others, 1977. Spectrophotometric and color estimates of the Viking lander sites. *J. Geophys. Res.* **82**, 4401-11.
- Hughes (D.W.), 1977. A comparison between the mass distribution indices of chondrules and cometary meteoroids. *Earth Planet. Sci. Lett.* **42**, 38-56.
- and , 1978. A disaggregation and thin section analysis of the size and mass distribution of the chondrules in the Bjurböle and Chajupur meteorites. *Earth Planet. Sci. Lett.* **38**, 391-400.
- Huguénin (R.L.), 1976. Chemical weathering as a massive volatile sink. *Icarus* **28**, 203-12.
- Hunten (D.M.), 1973. The escape of light gases from planetary atmospheres. *J. Atmos. Sci.* **30**, 1481-94.
- and , 1974. Aeronomy of the lower atmosphere of Mars. *Rev. Geophys. Space Phys.* **12**, 529-35.
- and Donahue (T.M.), 1976. Hydrogen loss from the terrestrial planets. *Ann. Rev. Earth Planet. Sci.* **4**, 265-92.
- McMill (G.E.) and Nagy (A.F.), 1977. Current knowledge of Venus. *Space Sci. Rev.* **20**, 265-82.
- Hutcheon (I.D.), Steele (I.M.), Smith (J.V.) and Clayton (R.N.), 1978. Ion microprobe, electron microprobe and cathodoluminescence data for Allende inclusions with emphasis on plagioclase chemistry. *Proc. Ninth Lunar Planetary Sci. Conf.*, accepted.
- and Solberg (T.N.), Clayton (R.N.) and Smith (J.V.), 1977. Ion microprobe measurements of excess 26Mg in Allende inclusions. *Meteoritics* **12**, 262. (Abstr.).
- Hutchinson (R.), 1972. The Angra dos Reis (stone) mineral assemblage and the genesis of stony meteorites. *Nature (Phys. Sci.)* **240**, 58-9.
- and , 1974. The formation of the Earth. *Nature* **250**, 556-8.
- and , 1976. Strontium and lead isotopic ratios, heterogeneous accretion of the Earth, and mantle plumes. *Geochim. Cosmochim. Acta* **40**, 482-5.
- Ikramuddin (M.), Binz (C.M.) and Lipschutz (M.E.), 1976. Thermal metamorphism of primitive meteorites - II. Ten trace elements in Abee enstatite chondrite heated at 400-1000°C. *Geochim. Cosmochim. Acta* **40**, 133-42.
- and , 1977a. Thermal metamorphism of primitive meteorites - III. Ten trace elements in Krymka L3 chondrite heated to 400-1000°C. *Geochim. Cosmochim. Acta* **41**, 393-401.
- and Matza (S.) and , 1977b. Thermal metamorphism of primitive meteorites - V. Ten trace elements in Tieschitz H3 chondrite heated at 400-1000°C. *Geochim. Cosmochim. Acta* **41**, 1247-56.
- Ingersoll (A.P.), 1970. Mars: occurrence of liquid water. *Science* **168**, 972-3.
- Ip (W.-H.), 1978. Model consideration of the bombardment event of the asteroidal belt by the planetesimals scattered from the Jupiter zone. *Icarus* **34**, 117-27.
- Ito (E.) and Matsui (Y.), 1978. Synthesis and crystal-chemical characterization of MgSiO₃ perovskite. *Earth Planet. Sci. Lett.* **38**, 443-50.
- Jacobs (J.A.), 1975. *The Earth's Core*. London (Academic Press).
- Jain (A.V.), Gordon (R.B.) and Lipschutz (M.E.), 1972. Hardness of kamacite and shock histories of 119 meteorites. *J. Geophys. Res.* **77**, 6940-54.
- Jeanloz (R.) and Ahrens (T.J.), 1977a. Pyroxenes and olivines: structural implications of shock-wave data for high pressure phases, 439-61. In *High-pressure Research*, eds. Mangham (M.H.) and Akimoto (S.), New York (Academic Press).
- and , 1977b. Equation of state of CaO at high pressures. *Eos* **58**, 1236. (Abstr.).

- Jeffreys (H.), 1947. The relation of cohesion to Roche's limit. *Mon. Not. R. Astron. Soc.* 107, 260-2.
- Johnson (J.E.), Scrymgeour (J.M.), Jarosewich (E.) and Mason (B.), 1977. Brachina meteorite - a chondritic from South Australia. *Records South Australian Museum* 12, 309-17.
- Johnston (D.H.) and Toksöz (M.N.), 1977. Internal structure and properties of Mars. *Icarus* 32, 73-84.
- , McGetchin (T.R.) and Toksöz (M.N.), 1974. *J. Geophys. Res.* 79, 3959-71.
- Jones (K.L.), 1974. Evidence for an episode of crater obliteration intermediate in Martian history. *J. Geophys. Res.* 79, 3917-31.
- Kallemeyn (G.W.) and Wesson (J.T.), 1977. The Eucumbin meteorite. *Meteoritics* 12, 270-1. (Abstr.).
- Kaula (W.M.), 1968. *An Introduction to Planetary Physics: The Terrestrial Planets*. New York (John Wiley).
- , 1971. Dynamical aspects of lunar origin. *Rev. Geophys. Space Phys.* 9, 217-36.
- , 1975. The seven ages of a planet. *Icarus* 26, 1-15.
- , 1976. Comments on the origin of Mercury. *Icarus* 28, 429-33.
- , 1977a. On the origin of the moon, with emphasis on bulk composition. *Proc. Lunar Sci. Conf.* 8th, 321-31.
- , 1977b. Mechanical processes affecting differentiation of protolunar material. *NASA SP-370*, 805-13.
- , and Harris (A.W.), 1973. Dynamically plausible hypotheses of lunar origin. *Nature* 245, 367-9.
- , and —, 1975. Dynamics of lunar origin and orbital evolution. *Rev. Geophys. Space Phys.* 13, 949-71.
- Kay (R.W.) and Hubbard (K.J.), 1978. Trace elements in ocean ridge basalts. *Earth Planet. Sci. Lett.* 38, 95-116.
- Keays (R.R.), Ganapathy (R.) and Anders (A.), 1971. Chemical fractionations in meteorites - IV Abundances of fourteen trace elements in L-chondrites; implications for cosmochemistry. *Geochim. Cosmochim. Acta* 35, 337-63.
- Keil (K.), 1968. Mineralogical and chemical relationships among enstatite chondrites. *J. Geophys. Res.* 73, 6945-76.
- , 1969a. Meteorite composition, 78-115. In *Handbook of Geochemistry*, Vol. 1, ed. Wedepohl (K.H.). Berlin (Springer).
- , 1969b. Titanium distribution in enstatite chondrites and achondrites and its bearing on their origin. *Earth Planet. Sci. Lett.* 2, 243-8.
- , and Fredriksson (K.), 1963. Electron microprobe analysis of some rare minerals in the Norton County achondrite. *Geochim. Cosmochim. Acta* 27, 939-47.
- Kelly (W.R.) and Larimer (J.W.), 1977. Chemical fractionations in meteorites - VIII. Iron meteorites and the cosmochimical history of the metal phase. *Geochim. Cosmochim. Acta* 41, 93-111.
- Kerridge (J.F.), 1976. Major element composition of pyroclasticites in the Orgueil carbonaceous meteorite. *Earth Planet. Sci. Lett.* 29, 194-200.
- , 1977a. Iron: whence it came, where it went. *Space Sci. Rev.* 20, 3-68.
- , 1977b. Correlation between nickel and sulfur abundances in Orgueil pyroclasticites. *Geochim. Cosmochim. Acta* 41, 1163-4.
- , and MacDougall (J.D.), 1976. Mafic silicates in the Orgueil carbonaceous meteorite. *Earth Planet. Sci. Lett.* 29, 341-8.
- , and Kieffer (S.W.), 1977. A constraint on impact theories of chondrule formation. *Earth Planet. Sci. Lett.* 35, 35-42.
- Kesson (S.E.) and Ringwood (A.E.), 1977. Further limits on the bulk composition of the moon. *Proc. Lunar Sci. Conf.* 8th, 411-31.
- Kieffer (H.W.), Chase (S.C., Jr.), Martin (T.Z.), Miner (E.D.) and Palluconi (F.D.), 1977a. Martian northern pole summer temperatures: dirty water ice. *Science* 194, 1341-4.
- , Martin (T.Z.), Peterfreund (A.R.), Jakosky (B.M.), Miner (E.D.) and Palluconi (F.D.), 1977b. Thermal and albedo mapping of Mars during the Viking primary mission. *J. Geophys. Res.* 82, 4249-92.
- Kim (K.-T.) and Burley (B.J.), 1971. Phase equilibria in the system $\text{NaAlSi}_3\text{O}_8\text{-NaAlSi}_2\text{O}_6\text{-H}_2\text{O}$ up to 15kbar: a theoretical discussion. *Can. J. Earth Sci.* 8, 549-57.
- Kimura (K.), Lewis (R.S.) and Anders (A.), 1974. Distribution of gold and rhodium between nickel-iron and silicate melts: Implications for the abundance of siderophile elements on the Earth and Moon. *Geochim. Cosmochim. Acta* 38, 683-701.
- King (E.A.), 1976. *Space Geology*. New York (John Wiley).
- Kolomoitseva (L.N.), 1975. On the conditions for equilibrium in pallastites. *Meteoritika* 24, 52-6.
- Kong (T.Y.) and McElroy (M.B.), 1977. The global distribution of ozone on Mars. *Planet. Space Sci.* 25, 839-57.
- Kracher (A.), Kurat (G.) and Buchwald (V.F.), 1977. Cape York: The extraordinary mineralogy of an ordinary iron meteorite and its implication for the genesis of IIAB irons. *Geochim. J.* 11, 207-17.
- Kumar (S.), 1976. Mercury's atmosphere: a perspective after Mariner 10. *Icarus* 28, 579-91.
- Kurat (G.), 1967a. Zur Entstehung der Chondren. *Geochim. Cosmochim. Acta* 31, 491-502.
- , 1967b. Einige Chondren aus dem Meteoriten von Mezö-Madaras. *Geochim. Cosmochim. Acta* 31, 1843-57.
- , Hoinkes (G.) and Fredriksson (K.), 1975. Zoned Ca-Al-rich chondrule in Bali: new evidence against the primordial condensation model. *Earth Planet. Sci. Lett.* 26, 140-4.
- Lang (D.E.) and Larimer (J.W.), 1973. Chondrules: an origin by impacts between dust grains. *Science* 182, 920-2.
- Langseth (M.G.), Rehm (S.J.) and Peters (K.), 1976. Revised lunar heat-flow values. *Proc. Lunar Sci. Conf.* 7th, 3143-71.
- Larimer (J.W.), 1971. Composition of the earth: chondritic or achondritic? *Geochim. Cosmochim. Acta* 35, 769-86.
- , 1975. The effect of C/O ratio on the condensation of planetary material. *Geochim. Cosmochim. Acta* 39, 389-93.
- , and Buseck (P.R.), 1974. Equilibration temperatures in enstatite chondrites. *Geochim. Cosmochim. Acta* 38, 471-77.
- Larson (H.P.), Fink (U.), Treffers (R.R.) and Gautier (T.N., III), 1976. The infrared spectrum of asteroid 433 Eros. *Icarus* 28, 95-103.
- Lattimer (J.M.), Johnson (D.H.) and Grossman (L.), 1978. Condensation in supernova ejecta and isotopic anomalies in meteorites. *Astrophys. J.* 219, 230-49.
- Laul (J.C.), Keays (R.R.), Ganapathy (R.), Anders (E.) and Morgan (J.W.), 1972. Chemical fractionations in meteorites - V. Volatile and siderophile elements in achondrites and ocean ridge basalts. *Geochim. Cosmochim. Acta* 36, 329-45.
- Lebofsky (L.A.), 1978. Asteroid 1 Ceres: evidence for water of hydration. *Mon. Not. R. Astron. Soc.* 182, 17P-21P.
- Lee (T.), Papanastassiou (D.A.) and Wasserburg (G.J.), 1977a. Aluminum-26 in the early solar system: fossil or fuel? *Astrophys. J.* 211, L107-10.
- , and —, 1977b. Mg and Ca isotopic study of individual microscopic crystals from the Allende meteorite by the direct loading technique. *Geochim. Cosmochim. Acta* 41, 1473-85.
- Levin (O.Yu.) and Mayeva (S.V.), 1977. Riddles about the origin and thermal history of the Moon. *NASA SP-370*.
- Levine (J.S.), 1976. A new estimate of volatile outgassing on Mars. *Icarus* 28, 165-9.
- Lewis (J.S.), 1970. Venus: Atmospheric and lithospheric composition. *Earth Planet. Sci. Lett.* 10, 73-80.
- , 1972a. Metal/silicate fractionation in the solar system. *Earth Planet. Sci. Lett.* 15, 286-90.
- , 1972b. Low temperature condensation from the solar nebula. *Icarus* 16, 241-52.
- , 1974a. The temperature gradient in the solar nebula. *Science* 186, 440-3.
- , 1974b. Volatile element influx on Venus from cometary impacts. *Earth Planet. Sci. Lett.* 22, 239-44.
- Lewis (R.S.), Srinivasan (B.) and Anders (E.), 1975. Host phase of a strange xenon component in Allende. *Science* 190, 1251-62.
- Leyroloup (A.), Dupuy (C.) and Andriambololona (R.), 1977. Catanzonal xenoliths in French neogene volcanic rocks: constitution of the lower crust. *Contr. Mineral. Petro.* 62, 283-300.
- Liebermann (R.C.), Jones (L.E.A.) and Ringwood (A.E.), 1977. Elasticity of aluminates, titanates, stannates and germanates compounds with the perovskite structure. *Phys. Earth Planet. Int.* 14, 165-78.
- Lin (L.S.), Goldstein (J.I.) and Williams (D.B.), 1977. Analytical electron microscopy study of the pliesite structure in the Carlton iron meteorite. *Geochim. Cosmochim. Acta* 41, 1861-74.
- Lindsley (D.H.) and Dixon (S.A.), 1976. Diopside-enstatite equilibria at 850° to 1400°C, 5 to 10 kbar. *Am. J. Sci.* 275, 1285-1301.
- Liu (L.), 1974. Silicate perovskite from phase transformations of pyrope-garnet at high pressure and temperature. *Geophys. Res. Lett.* 1, 277-80.
- , 1975. Post-oxide phases of forsterite and enstatite. *Geophys. Res. Lett.* 2, 417.
- , 1976. Orthorhombic perovskite phases observed in olivine, pyroxene, and garnet at high pressures and temperatures. *Phys. Earth Planet. Int.* 11, 289-98.
- , 1977. The system enstatite-pyroxene at high pressures and temperatures and the mineralogy of the earth's mantle. *Earth Planet. Sci. Lett.* 36, 237-45.
- , and Ringwood (A.E.), 1975. Synthesis of perovskite-type polymorph of CaSiO_3 . *Earth Planet. Sci. Lett.* 28, 209-11.
- Liu (S.C.) and Bonahue (T.M.), 1976. The regulation of hydrogen and oxygen escape from Mars. *Icarus* 28, 231-46.
- Longhi (J.), 1977. Magma oceanography 2: chemical evolution and crustal formation. *Proc. Lunar Sci. Conf.* 8th, 601-21.
- Lorin (J.C.), Shimizu (N.), Lévy (M.C.) and Allègre (C.J.), 1977. The Mg isotope anomaly in carbonaceous chondrites: an ion-pore study. *Meteoritics* 12, 299-300 (Abstr.).
- Lovering (J.F.), 1975. The Moama eucrite - a pyroxene-plagioclase cumulate. *Meteoritics* 10, 101-14.
- , and White (A.J.R.), 1964. The significance of primary scapolite in granulitic inclusions from deep-seated pipes. *J. Petro.* 5, 195-218.
- Low (D.R.) and Knauth (L.P.), 1977. Sedimentology of the Overwacht group (3.4 billion years), Transvaal, South Africa, and its bearing on the characteristics and evolution of the early earth. *J. Geol.* 85, 699-723.
- Lownan (P.D., Jr.), 1976. Crustal-evolution in terrestrial planets: implications for the origin of continents. *J. Geol.* 84, 1-26.
- Maalé (S.) and Aoki (K.), 1977. The major element composition of the upper mantle estimated from the composition of hercynites. *Contr. Mineral. Petro.* 63, 161-73.
- MacDougall (J.D.) and Kothari (B.K.), 1976. Formation chronology for C2 meteorites. *Earth Planet. Sci. Lett.* 33, 36-44.
- Maclean (W.H.) and Shimazaki (H.), 1976. The partition of Co, Ni, Cu, and Zn between sulfide and silicate liquids. *Econ. Geol.* 71, 1049-57.
- Malin (M.C.), 1974. Salt weathering on Mars. *J. Geophys. Res.* 79, 3888-94.
- Malinowsky (I.Yu.) and Doroshov (A.M.), 1977. Evaluation of P-T conditions of diamond formation with reference to chrome-bearing garnet stability. *Second Int. Kimberlites Conf. Rep. Abstr.*, unpag.
- Manuel (O.K.) and Sabu (D.D.), 1975. Elemental and isotopic inhomogeneities in noble gases: the case for local synthesis of the chemical elements. *Trans. Missouri Acad. Sci.* 9, 104-22.
- , Hennecke (E.W.) and Sabu (D.D.), 1972. Xenon in carbonaceous chondrites. *Nature (Phys. Sci.)* 240, 99-101.
- Mao (H.K.) and Bell (P.M.), 1977a. Generation of static pressure to 1.5 Mbar. *Carnegie Inst. Washington Yearb.* 76, 644-50.
- , and —, 1977b. Disproportionation equilibrium in iron-bearing systems at pressures above 100kbar with applications to the Earth's mantle. In *Energy and Geology of Geological Processes*, ed. Saxena (S.) and Bhattacharji (S.), Heidelberg (Springer).
- Martin (P.M.) and Mason (B.), 1974. Major and trace elements in the Allende meteorite. *Nature* 249, 333-4.
- Mason (B.), 1962. *Meteorites*. New York (John Wiley).
- , 1963a. Olivine composition in chondrites. *Geochim. Cosmochim. Acta* 27, 1011-23.
- , 1963b. The hypersthene achondrites. *Amer. Mus. Novitates* 2156, 1-13.
- , 1966. The enstatite chondrites. *Geochim. Cosmochim. Acta* 30, 23-40.
- , 1971a. The carbonaceous chondrites - a selective review. *Meteoritics* 6, 59-70.
- , 1971b. ed. of *Handbook of Elemental Abundances in Meteorites*. New York (Gordon and Breach).
- , and Jarosewich (E.), 1973. The Barea, Dyrrol Island, and Emery meteorites, and a review of the mesosiderites. *Mineral. Mag.* 39, 204-15.
- , Nelen (J.A.), Muir (P.) and Taylor (S.R.), 1976. The composition of the Chassigny meteorite. *Meteoritics* 11, 21-7.
- Massey (H.), Brown (G.H.), Eglington (G.), Runcom (S.K.) and Urey (H.C.), 1977. eds. of *The Moon: A New Appraisal from Space Missions and Laboratory Analyses*. London (Royal Society).
- Masursky (H.), Boyce (D.M.), Dial (A.L.), Schaber (G.G.) and Strohbell (M.E.), 1977. Classification and time of formation of Martian channels based on Viking data. *J. Geophys. Res.* 82, 4016-38.
- Matz (S.L.) and Lipschutz (W.E.), 1977. Volatile/mobile trace elements in Karoonda (C4) chondrite. *Geochim. Cosmochim. Acta* 41, 1398-1401.
- McCallum (M.E.) and Eggler (D.H.), 1976. Diamonds in an upper mantle peridotite nodule from kimberlite in southern Wyoming. *Science* 192, 253-6.
- McCarthy (T.S.), Erland (A.J.) and Willis (J.P.), 1973. On the origin of eucrites and diogenites. *Earth Planet. Sci. Lett.* 18, 433-42.
- McCord (T.B.) and Chapman (C.R.), 1975. Asteroids: spectral reflectance and color characteristics. *Astrophys. J.* 192, 553-62 and 1977, 781-90.
- McElroy (M.B.) and Kong (T.Y.), 1976. Oxidation of the Martian surface: constraints due to chemical processes in the atmosphere. *Geophys. Res. Lett.* 3, 569-72.
- , and Yung (Y.L.), 1977a. Photochemistry and evolution of Mars' atmosphere: a Viking perspective. *J. Geophys. Res.* 82, 4379-88.

- Yung (Y.L.) and Nier (A.O.), 1977b. Isotopic composition of nitrogen: Implications for the past history of Mars' atmosphere. *Science* **194**, 70-2.
- McGetchin (T.R.) and Smyth (J.R.), 1978. The mantle of Mars: some possible geological implications of its high density. *Icarus* **34**, 512-36.
- McKenzie (D.P.) and Weiss (N.), 1975. Speculations on the thermal and tectonic history of the Earth. *Geophys. J. R. Astron. Soc.* **42**, 131-74.
- McQueen (R.G.) and Marsh (S.P.), 1966. Shock-wave compression of iron-nickel alloys and the Earth's core. *J. Geophys. Res.* **71**, 1751-6.
- Discussion by T. Takahashi and W.A. Bassett **71**, 1757.
- McSween (H.Y., Jr.), 1977a. On the nature and origin of isolated olivine grains in carbonaceous chondrites. *Geochim. Cosmochim. Acta* **41**, 411-8.
- , 1977b. Carbonaceous chondrites of the Orans type: a metamorphic sequence. *Geochim. Cosmochim. Acta* **41**, 477-91.
- , 1977c. Petrographic variations among carbonaceous chondrites of the Vigarano type. *Geochim. Cosmochim. Acta* **41**, 1777-90.
- , 1977d. Chemical and petrographic constraints on the origin of chondrules and inclusions in carbonaceous chondrites. *Geochim. Cosmochim. Acta* **41**, 1843-60.
- and Richardson (S.M.), 1977. The composition of carbonaceous chondrite matrix. *Geochim. Cosmochim. Acta* **41**, 1145-61.
- and Stolper (E.), 1978. Shergottite meteorites. I: Mineralogy and petrography. *Lunar and Planetary Sci. IX*, 732-4.
- Meyer (C., Jr.), 1971. An experimental approach to circumstellar condensation. *Geochim. Cosmochim. Acta* **35**, 551-65.
- , 1977. Petrology, mineralogy and chemistry of KREEP basalt. *Phys. Chem. Earth* **10**, 239-60.
- Meyer (H.O.A.), 1977. Mineralogy of the upper mantle: a review of the minerals in mantle xenoliths from Kimberlite. *Earth-Science Reviews* **13**, 251-81.
- Miller (S.L.) and Smythe (W.D.), 1970. Carbon dioxide clathrate in the Martian ice cap. *Science* **170**, 531-3.
- Milton (D.J.), 1974. Carbon dioxide hydrate and floods on Mars. *Science* **183**, 654-6.
- Mittler (H.E.), 1975. Formation of an iron-poor moon by partial capture, or: Yet another exotic theory of lunar origin. *Icarus* **25**, 256-68.
- Miyake (G.T.) and Goldstein (D.I.), 1974. Nedapilla, a remelted iron meteorite. *Geochim. Cosmochim. Acta* **38**, 747-55.
- Miyamoto (M.) and Takeda (H.), 1977. Evaluation of a crust model of eucrites from the width of exsolved pyroxene. *Geochemical J.* **11**, 161-9.
- and Takano (Y.), 1975. Crystallographic studies of a bronzite in the Johnston achondrite. *Fortschr. Mineral.* **52**, 389-97.
- Modreski (P.J.) and Boettcher (A.L.), 1972. The stability of phlogopite + enstatite at high pressures: a model for micas in the interior of the earth. *Am. J. Sci.* **272**, 852-69.
- Moorthath (S.), 1975. Evolution of Precambrian crust from strontium isotopic evidence. *Nature* **254**, 395-9.
- Moore (H.J.), Hodges (C.A.) and Scott (D.H.), 1974. Multiringed basins—illustrated by Orientale and associated features. *Proc. Fifth Lunar Sci. Conf.*, p. 71-100.
- Morgan (J.W.) and Lovering (J.F.), 1973. Uranium and thorium in achondrites. *Geochim. Cosmochim. Acta* **37**, 1697-1707.
- , Higuchi (H.), Takahashi (H.) and Hertogen (J.) 1978a. A "chondritic" eucrite parent body: inference from trace elements. *Geochim. Cosmochim. Acta* **42**, 27-38.
- , Hertogen (J.) and Anders (E.), 1978b. The Moon: composition determined by nebular processes. *The Moon*, submitted.
- Moroz (V.I.), 1976. The atmosphere of Mars. *Space Sci. Rev.* **19**, 763-843.
- Morrison (D.), 1977. Asteroid sizes and albedos. *Icarus* **31**, 185-220.
- and Cruikshank (D.P.), 1974. Physical properties of the natural satellites. *Space Sci. Rev.* **15**, 641-739.
- Mueller (R.F.), 1963. Chemistry and petrology of Venus: preliminary deductions. *Science* **141**, 1046-7.
- , 1964. A chemical model for the lower atmosphere of Venus. *Icarus* **3**, 285-98.
- , 1965. Stability of sulfur compounds on Venus. *Icarus* **4**, 506-12.
- , 1968. Sources of HCl and HF in the atmosphere of Venus. *Nature* **220**, 55-7.
- , 1969. Effect of temperature on the strength and composition of the upper lithosphere of Venus. *Nature* **224**, 354-6.
- , 1970. Dehydrogenation of Venus. *Nature* **227**, 363-5.
- and Kridelbaugh (S.J.), 1973. Kinetics of CO₂ production on Venus. *Icarus* **19**, 531-41.
- Munson (R.A.) and Sheppard (R.A.), 1974. Natural zeolites: their properties, occurrences and uses. *Mineralog. Soc. Trans.* **6**, 19-34.
- Murray (B.C.), Strom (R.C.), Trask (N.J.) and Gault (D.E.), 1975. Surface history of Mercury: implications for terrestrial planets. *J. Geophys. Res.* **80**, 2508-14.
- Murthy (V.R.), 1976. Composition of the core and the early chemical history of the Earth, 21-31. In Windley (B.F.), ed., *The Early History of the Earth*, London (John Wiley).
- and Hall (H.T.), 1970. The chemical composition of the Earth's core: possibility of sulphur in the core. *Phys. Earth Planet. Int.* **2**, 276-82.
- and —, 1972. The origin and chemical composition of the Earth's core. *Phys. Earth Planet. Int.* **6**, 125-30.
- Mutch (T.A.) and Saunders (R.S.), 1976. The geologic development of Mars: a review. *Space Sci. Rev.* **19**, 3-57.
- , Arvidson (R.E.), Head (J.W., III), Jones (K.L.) and Saunders (R.S.), 1976. *The Geology of Mars*. Princeton, N.J. (Princeton Univ. Press).
- and 9 others, 1977a. The surface of Mars: the view from the Viking 1 Lander. *Science* **193**, 791-801.
- and 24 others, 1977b. The surface of Mars: the view from the Viking 2 Lander. *Science* **194**, 1277-83.
- , Arvidson (R.E.), Binder (A.B.), Guinness (E.A.) and Morris (E.C.), 1977c. The geology of the Viking Lander 2 site. *J. Geophys. Res.* **82**, 4452-67.
- Mysen (B.O.), 1977. The solubility of H₂O and CO₂ under predicted magma genesis conditions and some petrological and geophysical implications. *Rev. Geophys. Space Phys.* **15**, 351-61.
- and Boettcher (A.L.), 1975. Melting of a hydrous mantle: I. Phase relations of natural peridotite at high pressures and temperatures with controlled activities of water, carbon dioxide and hydrogen. *J. Petrol.* **16**, 520-48.
- Nagy (B.), 1975. *Carbonaceous Meteorites*. Amsterdam (Elsevier).
- Naldrett (A.J.), 1969. A portion of the system Fe-S-O between 900 and 1080°C and its application to sulfide ore magmas. *J. Petrol.* **10**, 171-201.
- Nash (C.R.), 1972. Primary anhydrite in Precambrian gneisses from the Swakopmund district, South West Africa. *Contr. Mineral. Petrol.* **36**, 27-32.
- Nehru (C.E.), Hewins (R.H.), Garcia (D.J.), Harlow (G.E.), Prinz (M.), 1978. Mineralogy and petrology of the Emery mesosiderite. *Lunar and Planetary Sci. IX*, 799-801.
- Ness (N.F.), Behannon (K.W.) and Lepping (R.P.), 1976. Observations of Mercury's magnetic field. *Icarus* **28**, 479-88.
- Neukum (G.) and Wise (D.U.), 1976. Mars: a standard crater curve and possible new time scale. *Science* **194**, 1381-7.
- Newburn (R.L., Jr.) and Gulkis (S.), 1973. A survey of the outer planets Jupiter, Saturn, Uranus, Neptune, Pluto and their satellites. *Space Sci. Rev.* **14**, 179-271.
- Newson (H.E.) and Drake (M.J.), 1977. Metal fractionation patterns in the BENCUBBIN meteorite. *Meteoritics* **12**, 326 (Abstr.).
- Newton (R.C.) and Goldsmith (J.R.), 1975. Stability of the scapolite melonite (3CaAl₂Si₂O₈·CaCO₃) at high pressures, and storage of CO₂ in the deep crust. *Contr. Mineral. Petrol.* **49**, 49-62.
- Nier (A.O.), McElroy (M.B.) and Yung (Y.L.), 1977. Isotopic composition of the Martian atmosphere. *Science* **194**, 66-70.
- Noonan (A.F.) and Melen (J.A.), 1976. A petrographic and mineral chemistry study of the Weston, Connecticut, chondrite. *Meteoritics* **11**, 111-30.
- Nordlie (B.E.), 1977. The composition of the magmatic gas of Kilauea and its behavior in the near surface environment. *Amer. J. Sci.* **277**, 417-63.
- Okal (E.A.) and Anderson (D.L.), 1978. Theoretical models for Mars and their seismic properties. *Icarus* **33**, 514-28.
- O'Keefe (J.D.) and Ahrens (T.J.), 1975. Shock effects from a large impact on the moon. *Proc. Lunar Sci. Conf.* **8th**, 2831-40.
- and —, 1977a. Impact-induced energy partitioning, melting, and vaporization on terrestrial planets. *Proc. Lunar Sci. Conf.* **8th**, 3357-74.
- and —, 1977b. Meteorite impact ejecta: dependence of mass and energy lost on planetary escape velocity. *Science* **198**, 1249-51.
- O'Leary (B.), 1977. Mining the Apollo and Amor asteroids. *Science* **197**, 363-6.
- Olson (E.) and Fredriksson (K.), 1966. Phosphates in iron and pallasite meteorites. *Geochim. Cosmochim. Acta* **30**, 459-70.
- and Jaroszewicz (E.), 1971. Chondrules: first occurrence in an iron meteorite. *Science* **174**, 583-5.
- , Grossman (L.) and Davis (A.), 1977. Origin of isolated olivine grains in the Murchison C2 meteorite. *Meteoritics* **12**, 336 (Abstr.).
- Paper accepted by *Earth Planet. Sci. Lett.* (1978) from Olson and Grossman.
- , Bunch (T.E.), Jaroszewicz (E.), Noonan (A.F.) and Huss (G.I.), 1977. Happy Canyon: a new type of enstatite achondrite. *Meteoritics* **12**, 109-23.
- O'Neill (G.K.), 1974. The colonization of space. *Physica Today* **27**, 32-40.
- , 1975. Space colonies and energy supply to the Earth. *Science* **190**, 943-7.
- O'Nions (R.K.) and Pankhurst (R.J.), 1978. Early Archaean rocks and geochemical evolution of the earth's crust. *Earth Planet. Sci. Lett.* **38**, 211-36.
- Öpik (E.V.), 1972. Comments on lunar origin. *Irish Astron. J.* **10**, 190-238.
- , 1976. *Interplanetary Encounters*. Amsterdam (Elsevier).
- Örögan (E.), 1969. Density of the Moon and nucleation of the planets. *Nature* **222**, 867.
- Orville (P.), 1975. Crust-atmosphere interactions. In *The Atmosphere of Venus*, ed. Hansen (J.E.). NASA SP-3482.
- Osborn (T.W.), Smith (R.H.) and Schmitt (R.A.), 1973. Elemental composition of individual chondrules from ordinary chondrites. *Geochim. Cosmochim. Acta* **37**, 1909-42.
- Owen (T.), 1976. Volatile inventories on Mars. *Icarus* **28**, 171-7.
- and Biemann (K.), 1977. Composition of the atmosphere at the surface of Mars: detection of argon-36 and preliminary analysis. *Science* **193**, 801-3.
- and 5 others, 1977. The composition of the atmosphere at the surface of Mars. *J. Geophys. Res.* **82**, 4635-9.
- Oyama (Y.I.) and Berdahl (B.J.), 1977. The Viking gas exchange experiment results from Chryse and Utopia surface samples. *J. Geophys. Res.* **82**, 4669-76.
- Palm (A.), 1969. The evolution of Venus' atmosphere. *Planet. Space Sci.* **17**, 1021-8.
- Papike (J.J.) and Vaniman (D.T.), 1978. The state of the mare basalt suite after Luna 24. Special Soviet Volume on Luna 24, in press.
- , Hodges (C.A.), Bence (A.E.), Cameron (M.) and Rhodes (J.M.), 1976. Mare basalts: crystal chemistry, mineralogy, and petrology. *Rev. Geophys. Space Phys.* **14**, 475-540.
- Paul (D.K.), Buckley (F.) and Nixon (P.H.), 1976. Fluorine and chlorine geochemistry of kimberlites. *Chem. Geol.* **17**, 125-33.
- Peale (S.S.), Schubert (G.) and Lingenfelter (R.E.), 1975. Origin of Martian channels: clathrates and water. *Science* **187**, 273-4.
- Pellas (P.), Bourot-Denise (M.) and Storzer (D.), 1978. Shaw revisited: cooling history and U-Pb distribution in phosphates. *Lunar and Planetary Science IX*, 879-81.
- Pepin (R.O.) and Phinney (D.), 1976. The formation interval of the Earth. *Lunar Sci.* **VII**, 682-4.
- Philpotts (J.A.) and Schnetzler (C.C.), 1970. Apollo 11 lunar samples: K, Rb, Sr, Ba and rare-earth concentrations in some rocks and separated phases. *Proc. Apollo 11 Lunar Sci. Conf.* **2**, 1471-86.
- Phinney (W.C.), Warner (D.L.) and Simonds (C.H.), 1971. Lunar highland rock types: their implications for impact-induced fractionation. NASA SP-370.
- Pieters (C.), Gaffey (M.J.), Chapman (C.R.) and McCord (T.B.), 1976. Spectrophotometry (0.33 to 1.07 μm) of 433 Eros and compositional implications. *Icarus* **28**, 105-15.
- Pollack (J.B.) and 6 others, 1977. Properties of aerosols in the Martian atmosphere as inferred from Viking Lander imaging data. *J. Geophys. Res.* **82**, 4479-96.
- Pollack (S.S.), 1966. Disordered orthopyroxene in meteorites. *Am. Mineral.* **51**, 1722-6.
- Pomameruma (C.), Shinyama (A.), Yanada (M.), Hobo (T.) and Pal (R.), 1977. Possible surface reactions on Mars: implications for Viking biology results. *Science* **197**, 455-7.
- Powell (B.N.), 1969. Petrology and chemistry of mesosiderites - I. Textures and composition of nickel-iron. *Geochim. Cosmochim. Acta* **33**, 789-810.
- , 1971. Petrology and chemistry of mesosiderites - II. Silicate textures and metal-silicate relationships. *Geochim. Cosmochim. Acta* **35**, 5-34.
- Prinn (R.G.), 1975. Venus: chemical and dynamical processes in the stratosphere and mesosphere. *J. Atmos. Sci.* **32**, 1237-47.
- Prins (P.), 1973. Apatite from African carbonates. *Earth* **6**, 133-44.
- Prinz (M.) and Keil (K.), 1977. Mineralogy, petrology and chemistry of ANT-suite rocks from the lunar highlands. *Phys. Chem. Earth* **10**, 215-37.
- , Manson (D.V.), Hlava (P.F.) and Keil (K.), 1975. Inclusions in diamonds: garnet therszollite and eclogite assemblages. *Phys. Chem. Earth* **9**, 797-815.

- , Fodor (R.V.) and Keil (K.), 1977a. Comparison of lunar rocks and meteorites: Implications to histories of the Moon and parent meteorite bodies. *NASA SP-370*, 183-99.
- , Keil (K.), Hlava (P.F.), Berkley (J.L.) and Gomes (C.B.), 1977b. Studies of Brazilian meteorites. III. Origin and history of the Angra dos Reis achondrite. *Earth Planet. Sci. Lett.* 35, 317-30.
- , Nehru (C.E.), Berkley (J.L.), Keil (K.), Jarosiewicz (E.) and Gomes (C.B.), 1977c. Petrogenesis of the Serra de Magé cumulate eucrite. *Meteoritics* 12, 341 (Abstr.).
- , Klimentidis (R.), Harlow (G.E.) and Hewins (R.J.), 1978. Petrologic studies bearing on the origin of the Lodran meteorite. *Lunar and Planetary Sci. IX*, 919-21.
- Rajamani (V.) and Naldrett (A.J.), 1978. Partitioning of Fe, Co, Ni, and Cu between sulfide liquid and basaltic melts and the composition of Ni-Cu sulfide deposits. *Econ. Geol.* 73, 82-93.
- , Chou (C.-L.) and Naldrett (A.J.), 1977. Partitioning of Pt-group elements between sulfide and basaltic melts. *Geol. Soc. Amer., Abstr. with Programs* 9, 1135-6.
- Rambaldi (E.), 1976. Trace element content of metals from L-group chondrites. *Earth Planet. Sci. Lett.* 31, 224-38.
- , and Larimer (J.W.), 1976. The Shaw chondrite. I. The case of the missing metal. *Earth Planet. Sci. Lett.* 33, 61-6.
- Ramdohr, P., 1976. The Mundrabilla Meteorite. *Fortschr. Mineral.* 53, 165-66.
- Rammensee (W.) and Wänke (H.), 1977. On the partition coefficient of tungsten between metal and silicate and its bearing on the origin of the moon. *Proc. Lunar Sci. Conf.* 8th, 399-409.
- Randich (E.) and Goldstein (J.I.), 1978. Cooling rates of seven hexahedrites. *Geochim. Cosmochim. Acta* 42, 221-33.
- Rasool (S.I.) and Debergh (C.), 1970. The runaway greenhouse and the accumulation of CO₂ in the Venus atmosphere. *Nature* 226, 1037-9.
- , and Le Sergeant (L.), 1977. Implications of the Viking results for volatile outgassing from Earth and Mars. *Nature* 266, 822.
- , Huxten (D.M.) and Kaula (W.M.), 1977. What the exploration of Mars tells us about the Earth. *Physica Today* 32-32.
- Reisenberg (R.), 1977. The moment of inertia and isostasy of Mars. *J. Geophys. Res.* 82, 369-75.
- Roid (A.M.) and Cohen (A.J.), 1967. Some characteristics of enstatite from enstatite achondrites. *Geochim. Cosmochim. Acta* 31, 661-72.
- , Duncan (A.R.) and Richardson (S.H.), 1977. In search of LKFM. *Proc. Lunar Sci. Conf.* 8th, 2321-36.
- , Bass (M.N.), Fujita (H.), Kerridge (J.F.) and Fredriksson (K.), 1970. Olivine and pyroxene in the Orgueil meteorite. *Geochim. Cosmochim. Acta* 34, 1253-5.
- Richardson (S.M.) and McSweeney (N.V., Jr.), 1978. Textural evidence bearing on the origin of isolated olivine crystals in C2 carbonaceous chondrites. *Earth Planet. Sci. Lett.* 37, 485-91.
- Ringwood (A.E.), 1961a. Chemical and genetic relationships among meteorites. *Geochim. Cosmochim. Acta* 24, 159-97.
- , 1961b. Silicon in the metal phase of enstatite chondrites and some geochemical implications. *Geochim. Cosmochim. Acta* 25, 1-13.
- , 1966a. Genesis of chondritic meteorites. *Rev. Geophys.* 4, 113-75.
- , 1966b. Chemical evolution of the terrestrial planets. *Geochim. Cosmochim. Acta* 30, 41-104.
- , 1975. *Composition and Petrology of the Earth's Mantle*. New York (McGraw-Hill).
- , 1977. Composition of the core and implications for the origin of the earth. *Geochemical J.* 11, 111-35.
- , and Essene (E.), 1970. Petrogenesis of Apollo 11 basalts, internal constitution and origin of the moon. *Proc. Apollo 11 Lunar Sci. Conf.* 1, 769-99.
- , and Clark (S.P.), 1971. Internal constitution of Mars. *Nature* 234, 89-92.
- , and Kesson (S.E.), 1977. Composition and origin of the moon. *Proc. Lunar Sci. Conf.* 8th, 371-98.
- Ronov (A.B.) and Yaroshevskiy (A.A.), 1976. A new model for the chemical structure of the Earth's crust. *Geochim. Int.* 13, 761-95.
- Rosenhauer (M.), Woermann (E.), Knecht (B.) and Ulmer (C.G.), 1977. The stability of graphite and diamond as a function of the oxygen fugacity in the mantle. *Second Int. Kimberlite Conf., Ext. Abstracts*, unpag.
- Ross (J.E.) and Aller (L.H.), 1976. The chemical composition of the Sun. *Science* 191, 1223-9.
- Rowe (J.J.), Morey (G.W.) and Zen (C.S.), 1972. The quinary reciprocal salt system. Na,K,Mg,Ca/Cl,SiO₄ - A review of the literature with new data. *Geol. Soc. Amer., Prof. Pap.* 741.
- Runcorn (S.K.), 1976. Inferences concerning the early thermal history of the moon. *Proc. Lunar Sci. Conf.* 7th, 3221-8.
- Ruskoj (E.L.), 1960. The origin of the moon, 1. Formation of a swarm of bodies around the earth. *Sov. Astron. A.J.* 4, 657-68.
- , 1963. On the origin of the moon, 2. The growth of the moon in the circumterrestrial swarm of satellites. *Sov. Astron. A.J.* 7, 221-7.
- , 1972. The origin of the moon, 3. Some aspects of the dynamics of the circumterrestrial swarm. *Sov. Astron. A.J.* 15, 646-54.
- Ruskoj (Ye.L.), 1977. The origin of the Moon. *NASA SP-370*, 815-22.
- Ryder (G.) and Wood (J.A.), 1977. Serenitites and Imbrium impact melts: implications for large-scale layering in the lunar crust. *Proc. Lunar Sci. Conf.* 8th, 655-68.
- Sabu (D.D.) and Manuel (O.K.), 1976. Xenon record of the early Solar System. *Nature* 262, 28-32.
- Safonov (V.S.), 1972. *Evolution of the Protoplanetary Cloud and Formation of the Earth and the Planets*. Jerusalem [Israel Program for Scientific Translations].
- , 1977. Time scale for the formation of the Earth and planets and its role in their geochemical evolution. *NASA SP-370*, 797-803.
- Sagan (C.), Toon (O.B.) and Gierasch (P.J.), 1973. Climatic change on Mars. *Science* 181, 1045-9.
- Saunders (R.S.) and Malin (M.C.), 1977. Geologic interpretation of new observations of the surface of Venus. *Geophys. Res. Lett.* 4, 547-50.
- Sawamoto (H.), 1977. Orthorhombic perovskite (Mg,Fe) SiO₃ and constitution of the lower mantle, 219-44. In *High-pressure Research*, eds. Mangham (M.H.) and Akimoto (S.), New York (Academic Press).
- Schwarzc (H.P.), Scott (S.D.) and Kissin (S.A.), 1975. Pressures of formation of iron meteorites from sphalerite compositions. *Geochim. Cosmochim. Acta* 39, 1457-66.
- Scott (E.R.D.), 1972. Chemical Fractionation in iron meteorites and its interpretation. *Geochim. Cosmochim. Acta* 36, 1205-36.
- , 1977a. Pallastites - metal composition, classification and relationships with iron meteorites. *Geochim. Cosmochim. Acta* 41, 349-60.
- , 1977b. Formation of olivine-metal textures in pallastite meteorites. *Geochim. Cosmochim. Acta* 41, 693-710.
- , 1977c. Geochemical relationships between some pallastites and iron meteorites. *Mineral. Mag.* 41, 265-72.
- , 1977d. Composition, mineralogy and origin of group IC iron meteorites. *Earth Planet. Sci. Lett.* 32, 273-84.
- , 1978a. Chemical fractionation of Ga and Ge in iron meteorites and the origin of anomalous irons. *Nature*, submitted.
- , 1978b. Primary fractionation of elements among iron meteorites. *Geochim. Cosmochim. Acta*, in press.
- , 1978c. Iron meteorites with low Ga and Ge concentrations - composition, structure and genetic relationships. *Geochim. Cosmochim. Acta*, in press.
- , and Wasson (J.T.), 1975. Classification and properties of iron meteorites. *Rev. Geophys. Space Physics* 13, 527-46.
- , and —, 1976. Chemical classification of iron meteorites. - VIII. Groups IC, 11E, 11IF and 97 other irons. *Geochim. Cosmochim. Acta* 40, 103-5.
- Shaw (D.M.), 1972. Development of the early continental crust. Part I. Use of trace element distribution coefficient models for the protoarchean crust. *Canadian J. Earth Sci.* 9, 1577-95.
- , 1974. R-mode factor analysis on enstatite chondrite analyses. *Geochim. Cosmochim. Acta* 38, 1607-13.
- , 1976. Development of the early continental crust. Part 2: Precambrian, protoarchean and later eras, 33-53. In Windley (B.F.), ed. *The Early History of the Earth*, London (John Wiley).
- Shih (C.) and Schonfeld (E.), 1976. Mare basalt genesis: A cumulate-remelting model. *Proc. Lunar Sci. Conf.* 7th, 151-92.
- Shima (H.) and Naldrett (A.J.), 1975. Solubility of sulfur in an ultramafic melt and the relevance of the system Fe-S-O. *Econ. Geol.* 70, 960-7.
- Shimizu (N.), 1975. Geochemistry of ultramafic inclusions from Salt Lake Crater, Hawaii, and from southern African kimberlites. *Phys. Chem. Earth* 2, 655-68.
- , and Allgrove (C.J.), 1977. Geochemistry of transition elements in garnet kherzolit nodules in kimberlites. *Second Int. Conf. Kimberlites*, Ext. Abstr., unpag.
- Shoemaker (E.M.), 1977a. Why study impact craters? In *Impact and Explosion Cratering*. New York (Pergamon).
- , 1977b. Astronomically observable crater-forming projectiles. Preprint.
- Short (N.M.), 1975. *Planetary Geology*. Englewood Cliffs, New Jersey (Prentice-Hall).
- Sibley (D.F.) and Wilband (J.T.), 1977. Chemical balance of the Earth's crust. *Geochim. Cosmochim. Acta* 41, 545-54.
- Stieglitz (R.W., II) and Solomon (S.C.), 1974. Mercury: internal structure and thermal evolution. *Icarus* 23, 192-205.
- Siever (R.), 1974. Comparison of Earth and Mars as differentiated planets. *Icarus* 22, 312-24.
- Sill (G.T.), 1976. Chemical reactions at the surface and in the atmosphere of Venus. Ph.D. thesis, Univ. of Arizona.
- , and Wilkening (L.L.), 1978. Ice clathrate as a possible source of the atmospheres of the terrestrial planets. *Icarus* 33, 13-22.
- Simpson (P.R.) and Bowles (J.F.W.), 1977. Uranium mineralization of the Witwatersrand and Dominion Reef system. *Phil. Trans. R. Soc. London* A286, 527-48.
- Singer (S.F.), 1970. How did Venus lose its angular momentum? *Science* 170, 1196-8.
- Skinner (B.J.) and Luce (F.D.), 1971. Solid solutions of the type (Ca,Mg,Mn,Fe) and their use as geothermometers for the enstatite chondrites. *Am. Mineral.* 56, 1269-96.
- Smith (B.A.) and Goldstein (J.I.), 1977. The metallic microstructures and thermal histories of several reheated chondrites. *Geochim. Cosmochim. Acta* 41, 1061-72.
- Smith (J.V.), 1974. Origin of moon by disintegrative capture with chemical differentiation followed by sequential accretion. *Lunar Science* 7, 718-20.
- , 1977. Possible controls on the bulk composition of the earth: implications for the origin of the earth and moon. *Proc. Lunar Sci. Conf.* 8th, 333-69.
- , 1978. Possible controls on the bulk composition and origin of the earth: further discussion. *Lunar Science* 11, 1071-3.
- , and Dawson (D.B.), 1975. Chemistry of Mg-rich micras from kimberlites and xenoliths, with implications for volatiles in upper mantle. *Geol. Soc. Amer., Abstr. with Programs* 7, 1275-6.
- , and Steele (I.M.), 1976. Lunar mineralogy: a heavenly detective story. Part II. *Am. Mineral.* 61, 1059-1116.
- , and Herwig (R.L.), 1978. Shergotty meteorite: mineralogy, petrography and minor elements. *Meteoritics*, submitted.
- , Herwig (R.L.), Ackermann (D.) and Dawson (J.B.), 1978. K,Rb and Ba in micras from kimberlite and peridotitic xenoliths, and implications for origin of basaltic rocks. *Proc. Second Int. Kimberlite Conf.*, unpag.
- , and others, 1970a. Petrologic history of the moon inferred from petrography, mineralogy and petrogenesis of Apollo 11 rocks. *Proc. Apollo 11 Lunar Sci. Conf.* 1, 897-925.
- , Anderson (A.T.), Newton (R.C.), Olsen (E.J.) and Willie (P.J.), 1970b. A petrologic model for the Moon based on petrogenesis, experimental petrology, and physical properties. *J. Geol.* 78, 381-405.
- Smyth (J.R.) and McGetchin (T.R.), 1978. The mantle of Mars: some possible geological implications of its high density. *Icarus*, submitted.
- Snow (T.P., Jr.), 1975. Depletion of interstellar elements and the interaction between gas and dust in space. *Astrophys. J.* 202, L87-90.
- Sobolev (N.V.), 1972. Petrology of xenoliths in kimberlitic pipes and indications of their abyssal origin. *24th Int. Geol. Congr., sect. 2*, 297-302.
- , 1977. Deep-seated inclusions in Kimberlites, trans. D. A. Brown and F. R. Boyd, Washington, D.C., American Geophysical Union.
- , and Lavrent'ev (Yu.G.), 1971. Isomorphous sodium admixture in garnets formed at high pressure. *Contr. Mineral. Petrol.* 31, 1-12.
- Solomon (S.C.), 1976. Some aspects of core formation in Mercury. *Icarus* 28, 509-21.
- , 1977a. Magma oceanography: 1. Thermal evolution. *Proc. Lunar Sci. Conf.* 8th, 583-99.
- , 1977b. The relationship between crustal tectonics and internal evolution in the Moon and Mercury. *Phys. Earth Planet. Int.* 15, 135-45.
- , 1978a. Formation, history, and energetics of cores in the terrestrial planets. *Phys. Earth Planet. Int.*, submitted.
- , 1978b. On volcanism and thermal tectonics on one-plate planets. *Geophys. Res. Lett.* 5, 461-4.
- , and Chaiken (J.), 1976. Thermal expansion and thermal stress in the moon and terrestrial planets: clues to early thermal history. *Proc. Lunar Sci. Conf.* 7th, 3229-44.
- Sonett (C.P.) and Dubs (A.), 1975. Lunar temperature and global heat flux from laboratory electrical conductivity and lunar magnetometer data. *Nature* 258, 118-21.
- , Golubov (G.S.), Schwartz (K.) and Keil (K.), 1970. The melting of asterothal parent bodies by uniaxial dynamo induction from a primordial T Tauri sun. *Astrophys. Space Sci.* 2, 446-88.
- Stacey (F.D.), 1977. *The Physics of the Earth*. 2nd ed. New York (John Wiley).
- Steele (I.M.) and Smith (J.V.), 1976. Mineralogy of the Ibitira eucrite and comparison with other eucrites and lunar samples. *Earth Planet. Sci. Lett.* 32, 273-84.
- Stephens (W.E.) and Dawson (J.B.), 1977. Statistical comparison between pyroxenes from kimberlites and their associated xenoliths. *J. Geol.* 85, 433-49.

- Stern (C.R.) and Wyllie (P.J.), 1973. Water-saturated and unsaturated melting relations of a granite to 35 kilobars. *Earth Planet. Sci. Lett.* 18, 163-7.
- Sterrett (K.T.), Klement (W., Jr.) and Kennedy (G.C.), 1965. Effect of pressure on the melting of iron. *J. Geophys. Res.* 70, 1979-84.
- Stöffler (D.), 1972. Deformation and transformation of rock-forming minerals by natural and experimental shock processes. I. Behavior of minerals under shock compression. *Fortschr. Mineral.* 49, 50-113.
- , 1974. Deformation and transformation of rock-forming minerals by natural and experimental shock processes. II. Physical properties of shocked minerals. *Fortschr. Mineral.* 51, 256-89.
- Stolper (E.), 1977. Experimental petrology of eucritic meteorites. *Geochim. Cosmochim. Acta* 41, 587-611.
- and McSwen (H.V., Jr.), 1978. Shergottite meteorites. II. Experimental petrology. *Lunar and Planetary Sci. IX*, 1119-21.
- Stormer (J.C.) and Carmichael (I.S.E.), 1971. Fluorine-hydroxyl exchange in apatite and biotite: a potential igneous geothermometer. *Contrib. Mineral. Petrogr.* 31, 121-31.
- Strohmeyer (W.), 1972. *Variable Stars*. Oxford (Pergamon).
- Sun (S.S.) and Hanson (G.N.), 1975. Evolution of the mantle: geochemical evidence from alkali basalts. *Geology* 3, 297-302.
- Sun (S.) and Nesbitt (R.W.), 1977. Chemical heterogeneity of the Archaean mantle, composition of the earth and mantle evolution. *Earth Planet. Sci. Lett.* 35, 429-48.
- Surkov (Yu. A.), Kirmozov (F.F.) and Vinogradov (A.P.), 1974. Uranium, thorium and potassium content of Venus rock measured by Venera 8. *Space Res.* 12, 679-87.
- Sze (H.G.) and McClell (H.B.), 1975. Some problems in Venus' aeronomy. *Planet. Space Sci.* 25, 763-86.
- Takahashi (H.), Janssens (M.-J.), Morgan (J.W.) and Anders (E.), 1978. Further studies of trace elements in C3 chondrites. *Geochim. Cosmochim. Acta* 42, 97-106.
- Takeda (H.), Miyamoto (H.), Ishii (T.) and Reid (A.M.), 1976. Characterization of crust formation on a parent body of achondrites and the moon by pyroxene crystallography and chemistry. *Proc. Lunar Sci. Conf. 7th*, 3535-48.
- Tatsumoto (M.), 1978. Isotopic composition of lead in oceanic basalt and its implication to mantle evolution. *Earth Planet. Sci. Lett.* 38, 63-67.
- Taylor (G.J.) and Heymann (D.), 1971. The formation of clear taenite in ordinary chondrites. *Geochim. Cosmochim. Acta* 35, 175-88.
- Taylor (S.R.), 1975. *Lunar Sciences: A Post-Apollo View*. New York (Pergamon).
- , 1976. Geochemical constraints on the composition of the moon. *Proc. Lunar Sci. Conf. 7th*, 3461-71.
- and Jakes (P.), 1977. Geochemical evolution of the moon revisited. *Proc. Lunar Sci. Conf. 8th*, 433-46.
- Thomsen (L.), 1977. Theoretical foundations of equations of state for the terrestrial planets. *Ann. Rev. Earth Planet. Sci.* 8, 491-513.
- Toksöz (M.N.) and Johnston (D.H.), 1974. The evolution of the moon. *Isaarus* 21, 389-414.
- and Johnston (D.H.), 1977. The evolution of the moon and the terrestrial planets. *MASA SP-370*, 295-327.
- Tomisaka (T.) and Eugster (H.P.), 1968. Synthesis of the sodalite group and subsolidus equilibria in the sodalite-noselite system. *Mineral. J. (Japan)* 5, 249-275.
- Toulin (P., III) and 7 others, 1977. Geochemical and mineralogical interpretation of the Viking inorganic chemical results. *J. Geophys. Res.* 82, 4625-34.
- Trask (N.J.) and Strom (R.G.), 1976. Additional evidence of Mercurian volcanism. *Isaarus* 23, 559-63.
- Turekian (K.K.), 1976. The evolution of the Earth's outer spheres. unpublished summary of paper presented at meeting of Amer. Assoc. Adv. Science.
- and Clark (S.P., Jr.), 1969. Inhomogeneous accretion of the Earth from the primitive solar nebula. *Earth Planet. Sci. Lett.* 6, 346-8.
- and —, 1975. The non-homogeneous accumulation model for terrestrial planet formation and the consequences for the atmosphere of Venus. *J. Atmos. Sci.* 32, 1257-61.
- , Davis (A.M.) and Clark (S.P., Jr.), 1977. Co, Ni and Fe partitioning between pallisitic phases. *Meteoritics* 12, 371-2 (Abstr.).
- Urey (H.C.), 1952. *The Planets: Their Origin and Development*. New Haven, Conn. (Yale University Press).
- and Mayeda (T.), 1959. The metallic particles of some chondrites. *Geochim. Cosmochim. Acta* 23, 113-24.
- Usselman (T.M.), 1975. Experimental approach to the state of the core: Part I. The liquidus relations of the Fe-rich portion of the Fe-Ni-S system from 30 to 100kb. II. Part II. Composition and thermal regime. *Amer. J. Sci.* 275, 278-90 and 291-303.
- Van Schmus (W.R.), 1967. Polymict structure of the Mezid-Madars chondrite. *Geochim. Cosmochim. Acta* 31, 2027-47.
- , 1969. The mineralogy and petrology of chondritic meteorites. *Earth-Science Reviews* 5, 145-84.
- and Hayes (J.M.), 1974. Chemical and petrographic correlations among carbonaceous chondrites. *Geochim. Cosmochim. Acta* 38, 47-64.
- Veverka (J.) and 8 others, 1974. A Mariner 9 atlas of the moons of Mars. *Isaarus* 23, 206-89.
- and Duxbury (T.C.), 1977. Viking observations of Phobos and Deimos: preliminary results. *J. Geophys. Res.* 82, 4213-23.
- Vilas (F.) and Leonard (T.B.), 1976. Mercury spectral reflectance measurements (0.33-1.06 μ m) 1974/75. *Isaarus* 28, 593-9.
- Villeneuve (J.F.) and Rose (A.W.), 1977. The geochemistry of some Archaean ultramafic lavas. *Chem. Geol.* 19, 43-60.
- Visconti (G.), 1977. Hydrogen escape in the terrestrial atmosphere at low oxygen levels: a photochemical model. *J. Atmos. Sci.* 34, 193-204.
- Wahl (W.), 1963. Cosmical changes and metamorphism of stony meteorites due to heating effects. *Geochim. Cosmochim. Acta* 27, 1025-31.
- Wai (C.M.) and Wasson (J.T.), 1969. Silicon concentrations in the metal of iron meteorites. *Geochim. Cosmochim. Acta* 33, 1465-71.
- and —, 1970. Silicon in the Mesodilla ataxite and the relationship between Si and Cr in reduced iron meteorites. *Geochim. Cosmochim. Acta* 34, 408-10.
- and Knowles (C.R.), 1972. The metal phase of the Bustee enstatite achondrite. *Mineral. Mag.* 38, 627-9.
- Walker (J.C.G.), 1975. Evolution of the atmosphere of Venus. *J. Atmos. Sci.* 32, 1248-56.
- Ward (W.R.), 1974. Climatic variations on Mars. *J. Geophys. Res.* 79, 3375-86.
- and Reid (M.J.), 1973. Solar tidal friction and satellite loss. *Moon. Res. Rep. Astron. Soc.* 164, 21-32.
- Wark (D.A.) and Lovering (J.F.), 1976. Refractory/platinum metal grains in Allende calcium-aluminum-rich clasts (CARC's): possible exotic presolar material? *Lunar Sci. VII*, 912-4.
- Wasilewski (P.), 1976. Shock-loading meteoritic b.c.c. metal above the pressure of transition: remanent-magnetization stability and microstructure. *Phys. Earth Planet. Inter.* 11, P5-11.
- Wasson (J.T.), 1974. *Meteorites: Classification and Properties*. New York (Springer).
- , 1977. Chondrite classification and origin. *Meteoritics* 12, 381-3 (Abstr.).
- and Wai (C.M.), 1970. Composition of the metal, schreibersite and perrite of enstatite achondrites and the origin of enstatite chondrites and achondrites. *Geochim. Cosmochim. Acta* 34, 169-84.
- and —, 1975. Explanation for the very low gallium and germanium concentrations in some iron meteorite groups. *Nature* 261, 114-6.
- , Schaudy (R.), Bild (R.W.) and Chou (C.), 1974. Mesosiderites - I. Compositions of their metallic portions and possible relationship to other metal-rich meteorite groups. *Geochim. Cosmochim. Acta* 38, 135-49.
- , Chou (C.-L.), Bild (R.W.) and Baedeker (P.A.), 1976. Classification of and elemental fractionation among urelites. *Geochim. Cosmochim. Acta* 40, 1449-58.
- Weast (R.C.), 1969. ed. *Handbook of Chemistry and Physics*. 50th ed. Cleveland (Chemical Rubber Co.).
- Weber (H.W.), Begemann (F.) and Hintsberger (H.), 1976. Primordial gases in graphite-diamond-kanakite inclusions from the Haverö urellite. *Earth Planet. Sci. Lett.* 29, 81-90.
- Weidenschilling (S.J.), 1976. Accretion of the terrestrial planets. II. *Isaarus* 22, 161-70.
- Weinke (H.M.), 1977. Chemical and mineralogical investigation of a Mundrabilla specimen. *Meteoritics* 12, 384-7. (Abstr.).
- Wesson (P.S.), 1974. A synthesis of our present knowledge of interstellar dust. *Space Sci. Rev.* 15, 469-82.
- Wetherill (G.W.), 1975. Radiometric chronology of the early solar system. *Ann. Rev. Astr. Soc.* 25, 283-328.
- , 1976. The role of large bodies in the formation of the earth and moon. *Proc. Lunar Sci. Conf. 7th*, 3245-57.
- , 1977. Evolution of the earth's planetesimal swarm subsequent to the formation of the earth and moon. *Proc. Lunar Sci. Conf. 8th*, 1-16.
- Whipple (F.L.), 1964. The history of the solar system. *Proc. Nat. Acad. Sci. Washington* 52, 565-94.
- Wickramasinghe (N.C.), 1974. Formaldehyde polymers in interstellar space. *Nature* 252, 462-3.
- Wilhelms (D.L.), 1976. Mercurian volcanism questioned. *Isaarus* 28, 551-8.
- Wilkening (L.), 1976. Carbonaceous chondritic xenoliths and planetary-type noble gases in gas-rich meteorites. *Proc. Lunar Sci. Conf. 7th*, 3549-59.
- and Clayton (R.N.), 1974. Foreign inclusions in stony meteorites II. Rare gases and oxygen isotopes in a carbonaceous chondritic xenolith in the Plainville gas-rich chondrite. *Geochim. Cosmochim. Acta* 38, 937-45.
- and Anders (E.), 1975. Some studies of an unusual eucrite. *Geochim. Cosmochim. Acta* 39, 1205-10.
- Williams (I.P.) and Crenin (A.W.), 1968. A survey of theories relating to the origin of the solar system. *Quart. J. R. Astron. Soc.* 9, 40-62.
- Windley (B.F.), 1975. ed. *The Early History of the Earth*. London (John Wiley).
- , 1977. *The Evolving Continents*. London (John Wiley).
- Winkler (H.G.F.), 1974. *Petrogenesis of Metamorphic Rocks*. 3rd ed. New York (Springer).
- Wlotzka (F.) and Jarosvich (E.), 1977. Mineralogical and chemical compositions of silicate inclusions in the El Tacio, Campo del Cielo iron meteorite. *Smithsonian Contrib. Earth Sci.* 19, 104-25.
- Wood (C.A.) and Head (J.W.), 1976. Comparison of impact basins on Mercury, Mars and the Moon. *Proc. Lunar Sci. Conf. 7th*, 3629-51.
- Wood (J.A.), 1962. Metamorphism in chondrites. *Geochim. Cosmochim. Acta* 26, 739-51.
- , 1963. On the origin of chondrules and chondrites. *Isaarus* 2, 152-80.
- , 1964. The cooling rates and parent planets of several iron meteorites. *Isaarus* 3, 429-59.
- , 1967. Olivine and pyroxene compositions in Type II carbonaceous chondrites. *Geochim. Cosmochim. Acta* 31, 2095-2108.
- , 1968. *Meteorites and the Origin of Planets*. New York (McGraw-Hill).
- , 1977. A survey of lunar rock types and comparison of the crusts of Earth and Moon. *MASA SP-370*, 313-24.
- , 1978. Pallisites and the growth of parent meteorite bodies. *Lunar and Planetary Science IX*, 1273-5.
- and Mitler (H.E.), 1974. Origin of the moon by a modified capture mechanism, or half of a loaf is better than a whole one. *Lunar Science V*, 857-3.
- and McSwen (H.V., Jr.), 1977. Chondrules as condensation products. 365-73. In *Comets-Asteroids-Meteorites*, ed. Delsemme (A.H.), Toledo, Ohio (University of Toledo).
- , Dickey (J.S., Jr.), Marvin (U.B.) and Powell (B.N.), 1977. Lunar anorthosites and a geophysical model of the moon. *Proc. Apollo 11 Lunar Sci. Conf. 1*, 965-88.
- Wyllie (P.J.), 1971. *The Dynamic Earth*. New York (John Wiley).
- , 1973. Experimental petrology and global tectonics - a preview. *Tectonophysics* 17, 189-209.
- , 1977. Effects of H_2O and CO_2 on magma generation in the crust and mantle. *J. Geol. Soc. London* 134, 215-34.
- , 1978. Mantle fluid compositions buffered in peridotite- CO_2 - H_2O by carbonates, amphibole and phlogopite. *J. Geol.* 86, in press.
- and Huang (W.L.), 1976a. Carbonation and melting reactions in the system $Ca-MgO-SiO_2-CO_2$ at mantle pressures with geophysical and petrological applications. *Contrib. Mineral. Petrogr.* 54, 79-107.
- and —, 1976b. High CO_2 solubilities in mantle magmas. *Geology* 4, 21-24.
- Yagi (T.), Mao (H.K.) and Bell (P.M.), 1977. Crystal structure of $MgSiO_3$ perovskite. *Carnegie Inst. of Washington Yearb.* 76, 516-9.
- , 1978. Structure and crystal chemistry of perovskite-type $MgSiO_3$. *Phys. Chem. Minerals*, submitted.
- Yoder (H.S., Jr.), 1976. *Generation of Basaltic Magma*. Washington, D.C. (National Academy of Sciences).
- Zaikowski (A.), Knacke (R.F.) and Porco (C.C.), 1975. On the presence of phyllosilicate minerals in the interstellar grains. *Astrophys. Space Sci.* 35, 97-115.
- Zellner (B.) and Bowell (E.), 1977. Asteroid compositional types and their distributions, 185-198. In *Comets-Asteroids-Meteorites*, ed. Delsemme (A.H.), Toledo, Ohio (University of Toledo).
- and Capen (R.C.), 1974. Photometric properties of the Martian satellites. *Isaarus* 23, 437-44.
- , Leake (M.), Morrison (D.) and Williams (J.G.), 1977. The E asteroids and the origin of the enstatite achondrites. *Geochim. Cosmochim. Acta* 41, 1759-67.

- Lebertre (T.), Duseaux (M.) and Dollfus (A.), 1977a. The asteroid albedo scale. I. Laboratory polarimetry of meteorites. *Proc. Lunar Sci. Conf. 8th*, 1091-110.
- Lebertre (T.) and Day (K.), 1977b. The asteroid albedo scale - II. Laboratory polarimetry of dark carbon-bearing silicates. *Proc. Lunar Sci. Conf. 8th*, 1111-17.
- Zuckerman (B.), 1977. Interstellar molecules. *Nature* **268**, 491-5.
- ### References
- The following papers and books, which are relevant to the major topics discussed in the Halliand Lecture, were noticed between June and November 1978.
- Bills (B.G.) and Ferrari (A.J.), 1978. Mars topography harmonics and geophysical implications. *J. Geophys. Res.* **83**, 3497-3508.
- Boudier (F.), 1978. Structure and petrology of the Lanzo peridotite massif (Piedmont Alps). *Geol. Soc. Amer. Bull.* **89**, 1574-91.
- Bowell (E.), Chapman (C.R.), Gradie (J.C.), Morrison (D.) and Zellner (B.), 1978. Taxonomy of asteroids. *Ioanus* **35**, 313-35.
- Burns, (J.A.), ed., 1978. *Planetary Satellites*. Tucson (University of Arizona Press).
- Butler (R.) and Anderson (D.L.), 1978. Equation of state fits to the lower mantle and outer core. *Phys. Earth Planet. Int.* **12**, 147-62.
- Clark (B.C.), 1978. Implications of abundant hygroscopic minerals in the Martian regolith. *Ioanus* **35**, 645-65.
- Collerson (K.D.) and Fryer (B.J.), 1978. The role of fluids in the formation and development of early continental crust. *Contr. Mineral. Petrol.* **62**, 151-67.
- Consolmagno (G.J.) and Lewis (J.S.), 1978. The evolution of icy satellite interiors and surfaces. *Ioanus* **34**, 280-93.
- Dermott (S.F.), ed., 1978. *Origin of the Solar System*. Based on a NATO Advanced Study Institute, 1976. Chichester, England (John Wiley).
- Dzurisin (D.), 1978. The tectonic and volcanic history of Mercury as inferred from studies of scarps, ridges, troughs, and other lineaments. *J. Geophys. Res.* **83**, 4883-906.
- Eggler (D.H.), 1978. The effect of CO₂ upon partial melting of peridotite in the system Na₂O-CaO-Al₂O₃-MgO-SiO₂-CO₂ to 35kb, with an analysis of melting in a peridotite-H₂O-CO₂ system. *Amer. J. Sci.* **278**, 305-43.
- Feldman (P.O.), 1977. The composition of comets. *Amer. Scientist* **65**, 299-309.
- Floran (R.J.), Prinz (M.), Hlava (P.F.), Keil (K.), Nehru (C.E.) and Hinthorne (J.R.), 1978. The Chassigny meteorite: a cumulate dunite with hydrous amphibole-bearing melt inclusions. *Geochim. Cosmochim. Acta* **42**, 1213-29.
- Fricker (P.E.), Reynolds (R.T.), Summers (A.L.) and Cassen (P.M.), 1976. Does Mercury have a molten core? *Nature* **259**, 293-4.
- Fuller (A.O.) and Hargraves (R.B.), 1978. Some consequences of a liquid water saturated regolith in early Martian history. *Ioanus* **34**, 614-21.
- Gooding (J.L.) and Keil (K.), 1978. Alteration of glass as a possible source of clay minerals on Mars. *Geophys. Res. Lett.* **5**, 727-30.
- Greenberg (R.), Wacker (J.F.), Hartmann (W.K.) and Chapman (C.R.), 1978. Planetesimals to planets: numerical simulation of collisional evolution. *Ioanus* **35**, 1-26.
- Head (J.W.), Wood (C.A.) and Mutch (T.A.), 1977. Geologic evolution of the terrestrial planets. *Amer. Scientist* **65**, 21-9.
- Heymann (D.), 1978. Solar gases in meteorites: the origin of chondrites and CI carbonaceous chondrites. *Meteoritics* **13**, 291-303.
- Johnson, (T.V.), 1978. The Galilean satellites of Jupiter: four worlds. *Ann. Rev. Earth Planet. Sci.* **6**, 93-125.
- Kulpecz (A.A., Jr.) and Hewins (R.H.), 1978. Cooling rate based on schreibersite growth for the Emery mesosiderite. *Geochim. Cosmochim. Acta* **42**, 1495-1500.
- Lambert (R.St.J.) and Chamberlain (V.E.), 1978. CO₂ permafrost and Martian topography. *Ioanus* **34**, 568-80.
- Lebofsky (L.A.), 1977. Identification of water frost on Callisto. *Nature* **269**, 785-7.
- Veeder (G.J.), Lebofsky (M.J.) and Matson (D.L.), 1978. Visual and radiometric photometry of 1580 Betulia. *Ioanus* **35**, 336-43.
- Loper (D.E.), 1978. The gravitationally powered dynamo. *Geophys. J. Roy. Astron. Soc.* **54**, 389-404.
- Matsui (T.) and Mizutani (H.), 1978. Gravitational N-body problem on the accretion process of terrestrial planets. *Ioanus* **34**, 146-72.
- McGeechin (T.R.) and Smyth (J.R.), 1978. The mantle of Mars: some possible geological implications of its high density. *Ioanus* **34**, 512-36.
- Merrill (R.B.) and Papike (J.J.), eds., 1978. *Mare Crisium. The View from Luna 24*. New York (Pergamon).
- Morrison (D.) and Wells (W.C.), eds., 1978. *Asteroids: an exploration assessment*. NASA Conference Publication 2053, 300 pp. Contains 15 papers and discussion.
- Olsen (E.) and Grossman (L.), 1978. On the origin of isolated olivine grains in type 2 carbonaceous chondrites. *Earth Planet. Sci. Lett.* **41**, 111-27.
- Rambaldi (E.R.), Cendales (M.) and Thacker (R.), 1978. Trace element distribution between magnetic and non-magnetic portions of ordinary chondrites. *Earth Planet. Sci. Lett.* **40**, 175-86.
- Sagan (C.), 1976. Erosion and the rocks of Venus. *Nature* **261**, 31.
- Scott (E.R.D.), 1978. Iron meteorites with low Ga and Ge concentrations - composition, structure and genetic relationships. *Geochim. Cosmochim. Acta* **42**, 1243-51.
- Sears (D.W.), 1978. Condensation and the composition of iron meteorites. *Earth Planet. Sci. Lett.* **41**, 129-38.
- Smith (P.H.), 1978. Diameters of the Galilean satellites from Pioneer data. *Ioanus* **35**, 167-76.
- Soderblom (L.A.) and Wanner (D.B.), 1978. Possible fossil H₂O liquid-ice interfaces in the Martian crust. *Ioanus* **34**, 622-37.
- Edwards (K.), Eliason (E.M.), Sanchez (E.M.) and Charette (M.P.), 1978. Global color variations on the Martian surface. *Ioanus* **34**, 446-64.
- Solomon (S.C.), 1978. On volcanism and thermal tectonics on one-plate planets. *Geophys. Res. Lett.* **5**, 461-4.
- Tedesco (E.), Drummond (J.), Candy (M.), Birch (P.), Nikoloff (I.) and Zellner (B.), 1978. 1580 Betulia, an unusual asteroid with an extraordinary light curve. *Ioanus* **35**, 344-59.
- Toksöz (M.N.) and Hsu (A.T.), 1978. Thermal history and evolution of Mars. *Ioanus* **34**, 537-47.
- Warner (J.L.) and Bickel (C.E.), 1978. Lunar plutonic rocks: a suite of minerals depleted in trace siderophile elements. *Amer. Mineral.* **63**, 1010-5.
- Weidenschilling (S.J.), 1978. Iron/silicate fractionation and the origin of Mercury. *Ioanus* **35**, 99-111.
- Willis (J.) and Wasson (J.T.), 1978. Cooling rates of group IVA iron meteorites. *Earth Planet. Sci. Lett.* **40**, 141-50.
- Moren (A.E.) and Goldstein (J.I.), 1978. Cooling rate variations of group IVA iron meteorites. 151-61. Willis (J.) and Wasson (J.T.), 1978. A core origin for group IVA iron meteorites: A reply to Moren and Goldstein, 162-67.
- Wyllie (P.J.), 1978. The effect of H₂O and CO₂ on planetary mantles. *Geophys. Res. Lett.* **5**, 440-2.
- Yagi (T.), Mao (H.-K.) and Bell (P.M.), 1978. Structure and crystal chemistry of perovskite-type MgSiO₃. *Phys. Chem. Minerals* **3**, 97-110.