THE EARLY HISTORY OF THE EARTH

Edited by

Brian F. Windley

The Early History of the Earth

BASED ON THE PROCEEDINGS OF A NATO ADVANCED STUDY INSTITUTE HELD AT THE UNIVERSITY OF LEICESTER 5–11 APRIL, 1975

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Preface

The early history of the Earth was so long and complicated that current students usually work only on one or two aspects of the subject. But the research subjects include field geology, geochemistry, structural geology and tectonics, the evolution of the atmosphere and the oceans, palaeontology, geochronology and metallogeny, so it is not surprising if specialists in one branch commonly have little knowledge of another; and yet many of these fields overlap and knowledge of one should influence and contribute to that of another. The oldest rocks known at present have an age of about 3.8 b.y., but most geologists know little of that pre-geological period before 3.8 b.y. when the core, mantle and protocrust formed. The aim of the NATO Advanced Study Institute, held from 5–11th April, 1975, at Leicester, was to bring together specialists from many fields to produce their latest findings and ideas and to discuss the present state of knowledge on early Earth history in the period 4.5–2.5 b.y. ago. Much of the success of the meeting was due to the fact that 2 hours of every day was devoted to discussion of specific problems. This volume contains the text of papers presented at the meeting with the object of providing an inter-disciplinary approach for the student, teacher and researcher to the problem of how the Earth evolved in its early stages. With such a large subject it is obviously impossible to be comprehensive, but it is hoped that this volume goes some way in providing an integrated compilation. The papers vary from long reviews of major subjects to short reports of recent research. Some are quantitative, some speculative and some highly controversial, but this variation reflects the state of current research in the subject. No one pretends that research into the early history of the Earth has reached an advanced stage of development—far from it; there are not enough constraints to enable one to choose between various alternatives and models in almost every field. I hope that this volume imparts to the student the controversial nature of our knowledge of the Archaean and pre-Archaean.

The idea of having a NATO Advanced Study Institute on the early history of the Earth came from Professor J. V. Smith in 1971 when I was working with him on Archaean rocks in Chicago and when he was involved in the early organization of the 1972 Feldspar NATO ASI at Manchester. I am grateful to him and to Professor J. Sutton (Imperial College, London) for many discussions since then on the organization of the meeting, and to Professor P. C. Sylvester-Bradley of the Department of Geology, Leicester University, who made many useful suggestions and gave much encouragement.

The Institute was attended by 141 scientists from 11 NATO and 67 other countries. The organizers are grateful for a generous grant from the Scientific Affairs Division of NATO which covered the accommodation and partial travel costs of participants from NATO countries, partial support for the field excursions in Scotland, and the organizational expenses of the meeting. The National Science Foundation of the U.S.A. kindly provided travel grants for three research students, and the International Union of Geological Sciences financed two visitors from non-NATO countries. UNESCO gave a grant to enable five delegates from non-NATO countries to attend the first full committee meeting of the International Geological Correlation Project on Archaean Geochemistry, convened by Dr A. Glikson and held during the Conference.

Thanks are due to many people who helped to make the meeting a success, especially: Drs J. Peal and K. Davies, Wardens of Villiers and Gilbert Murray Halls of Residence, for providing excellent facilities; the technical and secretarial staff of the Geology Department of Leicester University for their services; Miss J. Baker and Mr M. Clarke for general assistance throughout the meeting; and Professor J. Watson who led two field excursions, to the Scourian of NW Scotland and of the Outer Hebrides, before and after the meeting. In particular, I am grateful to Judith, my wife, for indispensable administrative and secretarial assistance pre-, syn- and post- the meeting; she also organized the field excursions.

June, 1975 BRIAN F. WINDLEY

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The Early Earth–Moon System

Development of the Earth–Moon System with Implications for the Geology of the Early Earth

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Introduction

The literature on the Earth–Moon system is bedevilled by naïveté and special pleading. This brief review cannot encompass all ideas, and I deliberately select recent accessible papers. For the Moon, I do not reference well-known data, and refer readers to the Proceedings of the Lunar Science Conferences and the journal *The Moon*: controversial ideas and references in other journals are documented more thoroughly. Let me emphasize my predilection for catastrophic processes which lead to molten planets, while recognizing that cooler, less catastrophic processes also occur: also note my psychological bias in favour of the petrologic model for the Moon by Smith et al. (1970a,b) with its crust of plagioclase-rich rocks and basalts, its olivine- and pyroxene-rich mantle, and Fe-rich core (Fig. 1): see also Wood et al. (1970). [The manuscript was prepared before the Sixth Lunar Science Conference, but brief references are made to the preprints. The book *Lunar Science: A Post-Apollo View* by Taylor (1975) appeared during final revision of the manuscript, and quick perusal reveals an excellent survey of the properties of the Moon, together with the Taylor–Jakeš geochemical model of partial melting.]

Prolonged weathering and metamorphism in a water-rich environment together with continuing igneous activity and continental drift have destroyed most of the Earth's early

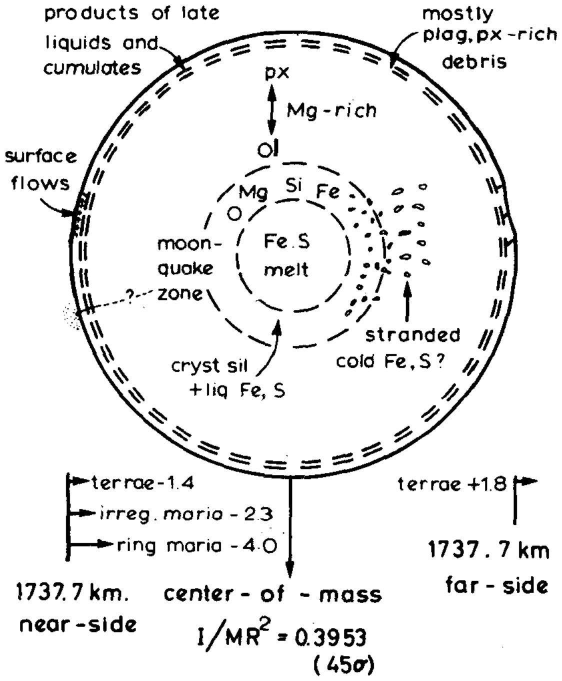

Fig. 1. Model for recent Moon (Smith, 1974) based on model of Smith et al. (1970a,b). Near-side basins are filled with mare basalts whereas far-side ones are unfilled. The moon-quake zone at ~700–1000 km depth is ascribed here to Fe,Ni,S liquid enclosed in solid silicate mantle, but probably would derive from melting of silicate if recent temperature estimates are correct. The postulated core might be as large as 700 km radius if the Fe,Ni,S liquid is rich in sulphide. The displacement of surface features with respect to a mass-centred sphere of mean radius 1737·7 km is shown at the bottom (Kaula et al., 1974)

crust. Initial hopes that an inert Moon would supply the missing evidence have been partly justified, but differences in chemical composition and dynamic environment of the two bodies require cautious interpretation. In 1975, chemical and morphological data for the Moon allow serious testing of petrologic and geochemical models but incomplete stratigraphic and geophysical controls lead to severe frustration. Exploration of Mercury, Mars and Venus is mostly confined to morphology and bulk properties, but these clearly favour major differentiation of the silicates, perhaps with separation of a metal-rich core. Theoretical calculations of chemical differentiation in a cooling solar nebula together with dynamic models for accretion and fragmentation can be combined with direct observations on meteorites and telescopic observations of asteroids and comets to provide a basis for the evolving system of planets. All these evidences combine to yield a plausible model in which high temperatures and intense bombardment prevent establishment of a stable crust on Earth before ~ −4 Giga years. In addition, the morphological and age relations on the Moon allow plausible speculations on the transition period in which early crust survived bombardment and remelting.

General Features of Earth and Moon

Established facts for the Earth include its density, seismic properties (which require an irregular crust, mantle and core); magnetism; general chemical properties of the upper crust with probable inferences for the lower crust and upper mantle (ranging from sediments via granitic, basaltic and granulitic to peridoditic and dunitic rocks); time scale running from −3·7 Gy for surviving remnants of the early crust; and evidence of bombardment. The inferred properties of the outer 100–200 km, plus those of the hydrosphere and atmosphere, show that temperatures did not rise high enough and long enough for loss of all the volatile components H_2O, CO_2 and alkalies although some loss can be expected on the basis of the noble gas data (e.g. Ringwood volatilization model: see Ringwood (1975) for many references). Quite uncertain is how these residual volatiles were originally stored and how they subsequently migrated (e.g. mica was probably stable down to ~ 200 km, and pyroxene could store Na and K at greater depths). Qualitatively, the observed properties of the outer 100–200 km are most easily explained by temperatures reaching the melting of silicates while permitting retention of volatile elements in amphibole, mica, scapolite, liquid, glass, etc. The inferred properties of the middle mantle are consistent with predominance of the $(Mg,Fe)_2SiO_4$ composition, perhaps occurring mainly as a mixture of dense oxides (e.g. stishovite and periclase). Inferences for the lower mantle and core are even more uncertain, but S-bearing Fe-rich material is likely for the latter. The above model provides sufficient framework for present purposes.

Established facts for the Moon include its density (3·34 gm/cm^3); its moment of inertia (~0·395 implying near-uniformity but slightly centre-heavy); its asymmetry (centre-of-mass about 2 km closer to Earth than the centre-of-volume); its surface morphology (highlands, irregular and ringed mare basins, craters of all sizes, lineaments and wrinkles); the refractory and reduced nature of, low content of metal-seeking elements in, and ubiquity of feldspar-rich breccias and basalts in the Apollo samples; the crystallization ages ranging from −3·9 to −3·1 Gy for mare basalts, and some less certain values earlier than −4 Gy for possible fragments of the early crust (e.g. Schaeffer and Husain, 1973; Jessberger et al., 1974); the model ages which indicate that lunar rocks originated mostly from a single differentiated reservoir about −4½ Gy; its weak seismicity; remanent magnetism of the crustal rocks; mass anomalies, especially positive ones over the centres of basalt-filled ringed basins and negative ones of late craters.

The seismic data are sparse, and mostly confined to travel paths near part of the Earth side. The latest interpretation (Nakamura et al., 1974) distinguishes four and perhaps five zones whose seismic properties are consistent with: I plagioclase-rich crust 50–60 km thick ranging from rubble at the surface to consolidated rock at depth (note that the seismic velocity of a consolidated rock is independent of grain size); II olivine-pyroxene upper mantle 250 km thick (note that olivine has only slightly higher seismic velocity than pyroxene); III 500 km middle mantle with high Poisson's ratio (0·33–0·36); IV lower mantle with high attenuation of shear waves, perhaps resulting from partial melting; V core 170–360 km radius with low P velocity, perhaps resulting from an iron-rich melt. The details are model-dependent, but the evidence for moon quakes at ~ 600-1000 km depth, plus poor transmission of shear waves, definitely favours a hot interior with partial melting. The crust is probably thicker on the far side, perhaps about 100 km thick. Although quite tentative at this time, the easiest interpretation of the seismic data utilizes an olivine-rich mantle and an Fe-rich core. Perturbation of the solar wind yields an estimated profile of the electrical conductivity with depth. The inferred temperature profile requires a mineralogical model and an estimate of the oxidation state of iron. Whatever the details, an olivine or pyroxene mantle with iron in the reduced state allows high temperatures up to and including the melting range of basaltic and Fe,S-rich liquids (Duba and Ringwood, 1973; Schwerer et al., 1974). The simplest explanation of the magnetic properties invokes a dynamo in the early lunar core, but there are many unsolved problems involving dynamics and energy sources (e.g. Sonett and Runcorn, 1973). Many aspects of the internal composition and evolution of the Moon are reviewed by Solomon and Toksöz (1973) and Dainty et al. (1974).

Taking all the data together, a possible model (e.g. Smith and Steele, 1975) utilizes (a) accretion near −4·5 Gy with only minor later accretion at the surface from impacting bodies: the original bulk composition was dominated by Mg-rich olivine and pyroxene, Ca-rich plagioclase, ~5% Fe,S,Ni, minor double-oxides, apatite, etc., thereby requiring the refractory and low-Fe nature to be inherited from the accretion process, (b) total crystal-liquid differentiation near −4·5 Gy producing plagioclase-rich asymmetric crust, olivine-rich mantle and Fe,Ni,S liquid core, (c) intense early bombardment down to ~ −4 Gy at which stage the crust had thickened sufficiently to sustain distinct impact basins, (d) intense brecciation and crystal-liquid fractionation of the crust from −4·5 to −4·0 Gy, (e) uprising of the mantle under basins, differentiation of the debris and underlying rock, remelting of early cumulates at the crust–mantle interface to produce basalts which flooded the mare basins from −4 to −3 Gy, (f) consolidation and shrinkage of the crust and upper mantle plus minor volcanism from −3 Gy to present (e.g. Muehlberger, 1974), (g) retention of liquid in the lower mantle and core, (h) declining flux of projectiles. [Note that complete simultaneous melting of the Moon is not necessary, and progressive partial melting with complex crystal-liquid fractionation gives an easier interpretation of the temperature profile.] Fig. 2 is a possible model of the *present* Moon.

Important Problems Involving the Moon

Extent of Melting

Many early lunar models utilized a cold Moon in which a molten zone developed near the surface and moved inwards only a short distance. Most present models have accepted the concept of early (~ −4·5 Gy) melting sufficient to produce an ~ 50 km plagioclase-rich crust, but many models involve melting of only the outer part of the Moon with retention of primary accreted material at the centre (e.g. Taylor and Jakeš, 1974, who consider both possibilities).

Time of Crustal Differentiation and Nature of Bombardment

Taylor (1975) reviews the general features of the early basins. These basins seem to occur randomly over the entire Moon. Faint features indicate partial destruction of earlier basins by later ones. Basins on the near-side contain basalt, while most on the far-side are free of

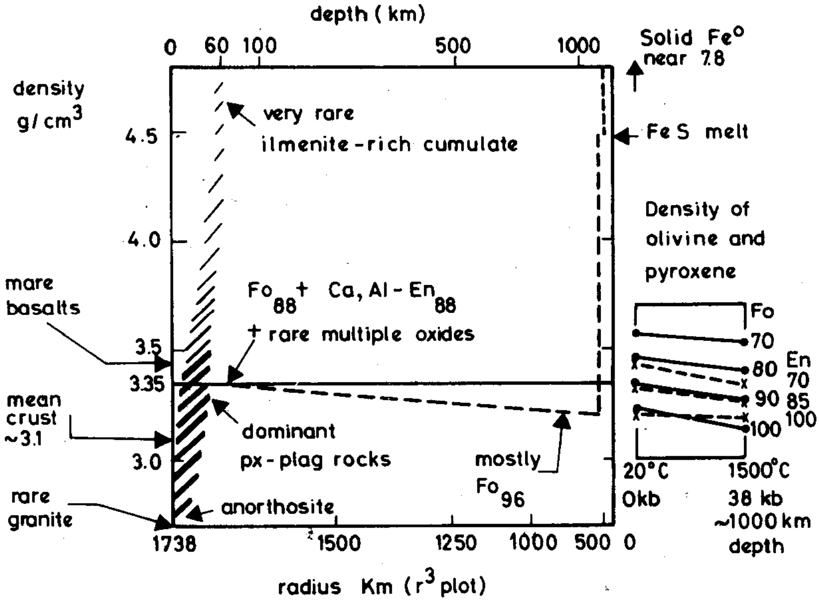

Fig. 2. Possible cross-section of present Moon showing density profile and location of rock types for simplest interpretation of seismic data (Smith and Steele, 1975). The radius is plotted as r^3 to show graphically how the zones contribute to the bulk density. An Fe-rich core could range in density from ~4·5 for FeS liquid to ~7·8 for solid Fe. The densities of olivine and low-Ca pyroxene are shown at the small right-hand diagram. The mantle varies from olivine-rich at the base to olivine plus pyroxene at the top. The lunar crust is dominated by rocks containing plagioclase and pyroxenes but contains some Ba-rich granite and ilmenite-rich cumulate

basalt except for minor quantities. See Stuart-Alexander and Howard (1970), Wilhelms and McCauley (1971) and Wilshire and Jackson (1972) for basin stratigraphy based on surface morphology: note also suggestion of Gargantuan basin (Cadogan, 1974) which covered most of NW quadrant of Moon prior to Imbrium impact. These morphological features require early development of a global crust which was disrupted by impact of large projectiles prior to out-pouring of mare basalts.

The actual time of crustal differentiation and nature of the bombardment are controversial. Tera et al. (1974) and Tera and Wasserburg (1974) interpreted U–Pb data in terms of formation of the lunar crust near −4·42 Gy with widespread metamorphism in a 'terminal lunar cataclysm' at −4·0 to −3·8 Gy which involved blanketing of the Moon's nearside by the debris from basin-forming impacts. Jessberger et al. (1974) used high-resolution Ar data for breccias and mineral separates to suggest that the basins formed over $\sim 10^8$ y between −4·0 and −3·9 Gy. Schaeffer and Husain (1974), also using ^{40}Ar–^{39}Ar ages but of small fragments, proposed that the terminal lunar cataclysm results merely from the predominance of ejecta from the Imbrium basin, and proposed that basins developed over some hundreds of millions of years: viz., Nectaris −4·25 ± 0·05, Crisium/Humorum −4·20 to −4·05, Imbrium −3·95 ± 0·05, Orientale −3·85 ± 0·05 Gy. Nunes et al. (1974) claimed that two-stage models are insufficient to explain the U–Pb evolution of many lunar samples; argued for a complex bombardment history from ~ −4·5 to −3·9 Gy with basin chronology similar to that of Schaeffer and Husain;

and suggested that mare basalts retained no Pb from earlier rocks. Further data and polemics are given by Kirsten and Horn (1974), Nunes and Tatsumoto (1975) and Tera and Wasserburg (1975).

Crucial to settling this controversy will be reliable assignment of ejecta to the various basins. McGetchin et al. (1973) developed a semi-empirical formula for ejecta thickness t in terms of the crater radius R and distance from the crater centre r (all in metres) where $t = 0{\cdot}14R^{0{\cdot}74}(r/R)^{-3{\cdot}0}$.

This model predicts the following thicknesses of Imbrium ejecta: Apollo 15 800 m, Apollo 14 130 m, Apollo 17 100 m and Apollo 16 50 m. Pike (1974) estimated even greater thicknesses. Moore et al. (1974), using the McGetchin et al. model and a spherical Moon, predicted average thicknesses of 0·2 km at 1000 km from the basin centre, falling to 10 metres at 4000 km and rising again to 0·2 km at the antipode.

Morgan et al. (1974) utilized the very low content of siderophile elements in the indigeneous lunar rocks to detect the amount of meteoritic contamination in surface breccias. They calculated the expected thicknesses at the Apollo and Luna sites for 12 basins, and utilized the proportions of less to more volatile siderophile elements to identify six distinct meteoritic components, five of which were tentatively assigned to the bodies responsible for the Imbrium, Serenitatis, Crisium, Nectaris and Humorum or Nubium basins. Using ingenious but model-dependent ideas, they concluded that the bodies consisted of an extinct population of planetesimals or moonlets of roughly chondritic composition, undifferentiated, with 15–40% Fe, and striking at velocities generally less than 8 km/sec. None of the bodies resembled the Earth and Moon in bulk composition, while several appeared to be less refractory and richer in volatiles. Utilizing an estimate that the lunar crust contains no more than 2% of meteoritic material, and that none was lost during impact, Morgan et al. concluded that between 40 and 160 basin-forming objects hit the Moon from the time of formation of the crust (~ −4·5 Gy) until the Imbrium impacts at −3·9 Gy. This is about twice the number identified by Hartmann and Wood (1971), but early basins could have been obliterated, of course.

Perhaps at this time the most plausible model involves (a) development of the crust, mantle and ? core at −4·5 ± 0·1 Gy in one major episode, (b) intense bombardment by a variety of projectiles down to −3·9 Gy, only the late arrivals leaving distinct basins and ejecta blankets. Undoubtedly considerable tolerance exists for varying the timing and nature of the bombardment, and clustering of projectiles is not unlikely especially if a planetisimal disintegrates.

Bulk Chemical Composition

Apart from late meteoritic accretion which is trivial volumetrically, the remarkably uniform partitition coefficients for many geochemical indicators (e.g. Ba, Nb, Hf, Th, vs. REE, Taylor and Jakeš, 1974) appear to rule out significant heterogeneous accretion of the crust, and major loss of volatiles from the surface.

The bulk composition of the Moon is very difficult to estimate since surface rocks are strongly biased by crustal material. Mare basalts have often been interpreted as partial melts of the lunar mantle, but some tricky geochemical problems remain (e.g. Ringwood, 1974). Smith et al. (1970a) argued for a hybrid origin of mare basalts, and Smith and Steele (1975) presented a model for origin of mare basalts at the crust–mantle transition zone (see later). They also argued that minor ultrabasic fragments could be interpreted in terms of an Mg-rich mantle composed principally of olivine. Table 1, column 3, gives their estimate of the bulk composition of the Moon based on a crust 80 km thick of 50% plagioclase An_{93} and 50% pyroxene $(En_{63}Fs_{27}Wo_{10})_{98}$ $(Al_2O_3)_2$, a mantle composed of 80% olivine $Fo_{91}Fa_9$ and 20% pyroxene $(En_{86}Fs_7Wo_7)_{99}$ $(Al_2O_3)_1$, and a hypothetical core with 5·5 wt.% Fe and 1·0 wt.% FeS (Fig. 2). If mare basalts actually derive from the mantle, the olivines and pyroxenes therein must contain more Fe (e.g. $\sim Fa_{15-20}$): this

Table 1. Estimates of chemical composition (wt.%) of Moon and Earth

	1	2	3	4	5	6	7	8
SiO_2	44·9	44·0	42·5	39·8	33·7	45·2	44·6	30·7
TiO_2	0·56	0·3	ne	0·56		0·7	0·2	0·17
Al_2O_3	24·6	8·2	2·6	11·0	26·6	3·5	3·9	3·3
Cr_2O_3	0·10	0·19	ne	0·17		0·4	0·5	0·7
FeO	6·6	10·5	5·6	3·5	2·3	8·0	8·4	4·7
MnO	0·1	0·1	ne	0·04		0·14	0·2	0·08
MgO	8·6	31·0	41·0	28·8	13·1	37·5	37·4	21·9
CaO	14·2	6·0	1·7	8·9	21·6	3·1	3·7	2·7
Na_2O	0·45	0·11	0·05	0·12	1·1	0·57	0·35	0·21
K_2O	0·075	0·012	ne	0·12		0·13	0·12	0·021
Fe			(5·5)	5·6				29·0
FeS			(1·0)	1·1				5·0

See original references for other elements, and for qualifications concerning some listed elements.

1 Lunar crust, 60 km thick, Taylor and Jakeš (1974), Taylor (1975).
2 Whole Moon, geochemical model, Taylor and Jakeš (1974).
3 Whole Moon, simple mineralogic model, Smith and Steele (1975): ne not estimated but values near TiO_2 0·2 Cr_2O_3 0·1 MnO 0·1 K_2O 0·01 would be acceptable: Fe and FeS arbitrary guesses.
4 Whole Moon, solar condensation model, Ganapathy and Anders (1974), similar composition used for phase-equilibria experiments by Hodges and Kushiro (1974): also Ni ~0·5.
5 Allende Ca–Al aggregate, Clarke et al. (1970).
6 Pyrolite III, Ringwood (1966a), used as Moon model by Binder (1974). Also Fe_2O_3 0·5 NiO 0·2.
7 Archean pyrolite, Green (1975). Also Fe_2O_3 0·7 NiO 0·3.
8 Whole Earth, solar condensation model, Ganapathy and Anders (1974): also Ni ~2.

results in the higher FeO content of the Taylor–Jakeš model (column 2).

Taylor and Jakeš (1974) developed a geochemical model of the Moon using (a) a crust based on orbital remote-sensing data, observed data for various rock types, and selection of a model of 80% anorthositic gabbro (often called highland basalt: note the common use of rock names in a *chemical* rather than a *petrographic* sense) and 20% low-K basalt, (b) an interior whose negative REE anomaly matches the positive one of the crust and whose mineralogic composition is dominated by olivine and pyroxene with properties obeying bulk physical properties and derivation of mare basalt by partial melting. Fig. 3 shows the model developed for −4·4 Gy. Table 1, columns 1 and 2, shows predicted amounts of major and minor elements in the differentiated part of the bulk Moon and the 60 km crust. Note that this model is basically similar to the one in Figs. 1, 2 and 4, except for the greater depth of origin of the mare basalts, and the concept of a frozen primitive crust to provide high Ni and Mg in crustal rocks.

Ganapathy and Anders (1974) predicted bulk compositions for the Earth and Moon by use of model compositions developed from theoretical studies of condensation of a hypothetical solar nebula (constrained by estimates of pressure, temperature and of the bulk composition of the sun) together with inferences from meteorite texture and chemistry. They used three principal condensates: refractory, early condensate; metallic Ni,Fe; and Mg-rich silicate. During cooling, metal reacts with H_2S to give FeS and with H_2O to give FeO which enters the silicate. Remelting, probably during collisions, leads to loss of volatiles, and reversion of FeS to metal. Finally, a volatile-rich, Mg–Fe silicate component is incorporated. The resulting seven components were adjusted for the Earth and Moon using observed values for certain elements, including U, Th, Mn and K, to give the following:

	Moon (mean of models 3 and 3a)	Earth
Early condensate	0·30	0·09 mass
Metal, remelted	0·05	0·24 fraction
Metal, unremelted	—	0·07
Troilite	0·009	0·05
Silicate, remelted	0·57	0·42
Silicate, unremelted	0·07	0·11
Volatile-rich material	0·0004	0·01
Mg/(Mg + Fe) atomic	0·90	0·89
Fe/Si atomic	0·24	1·26

The resulting bulk compositions are given in Table 1, columns 4 and 8.

Anderson (1973) proposed that the entire Moon is composed of differentiation products of a high-temperature condensate, such as the Ca,Al-rich aggregate from the Allende meteorite (column 5). However, Seitz and Kushiro (1974) and Hodges and Kushiro (1974) found from melting experiments that neither the Allende aggregate nor a 2 : 3 mixture of aggregate and bulk Allende meteorite was suitable since the interior of the Moon would contain considerable clinopyroxene and minor melilite, partial melting of which would yield melts too low in Si and too high in Ca to fit with known lunar rocks. Of course,

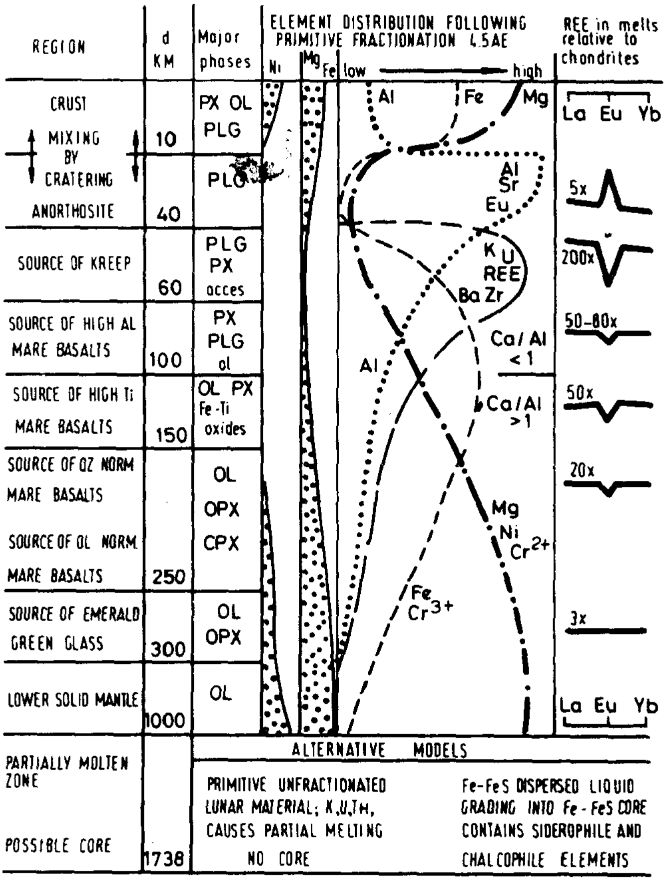

Fig. 3. Geochemical model of Moon proposed by Taylor and Jakeš (1974) for ~ −4·4 Gy. A frozen crust rich in olivine provides the high Ni and Mg and low Al of surface rocks. Melting of the underlying layer down to 1000 km leads to mineralogic zoning similar to that of Figs. 1 and 2. The frozen crust is mixed by impacts with the plagioclase-rich floated cumulate. The upper part of the lithosphere is zoned to provide successively deeper sources for the extrusive rocks ranging from basalts rich in K, REE and P to breccias rich in emerald green glass.