A quest for the moon

Narendra Bhandari

There has been a renewed interest in exploration of the moon in the past few years and a number of space missions are planned by various countries during this decade. The Indian Space Research Organization has also been discussing the possibility of an orbiter mission to the moon for remote sensing and has debated the scientific goals for such a mission in various fora such as the annual meeting of the Indian Academy of Sciences held at Lucknow.

The main motivation for the future missions is to resolve some of the problems regarding the formation and early stages of chemical evolution of the moon, which can be clearly defined now in view of the vast database on chemical, geological and chronological aspects that has become available as a result of thirty years of serious study of the samples of the moon and several orbiter missions. An attempt is made here to summarize the salient features of our current understanding of the moon so that scientific objectives of the future missions can be clearly formulated. Specifically, the desirability of a low altitude (100 km) lunar polar orbiter for simultaneous chemical, mineralogic and photogeologic mapping is discussed and some regions on the far side of the moon and polar regions are identified for a detailed study.

त्वं सोम प्र चिकितो मनीषा ।
त्वं रजिष्ठ मनु नेषि पन्थाम ।

'O moon! We should be able to know you through our intellect
You enlighten us through the right path'
Rg̣veda Part 1/91
(about 2000 years BC)

THE intellectual quest for the moon began when man first learnt to comprehend nature and correlate celestial phenomena. This got a big boost when the Russian satellite *Sputnik* demonstrated in 1957 that man could enter the space around and beyond the earth. This was followed by a series of orbiting, landing and sample-return missions to the moon by the former Soviet Union and USA. The Surveyor landings (Figure 1) provided the first analysis of lunar samples, and Lunokhods and lunar rovers explored large areas of the moon. The *Apollo* manned missions by USA provided the opportunity for direct exploration of the moon by man. As a result of these missions, lunar samples from nine well-documented sites became available for study and the past three decades have seen an all-out effort to understand the origin and evolution of the moon based on their laboratory analyses. Much has been learnt about the astronomical, physical, chemical, isotopic, geological and chronological aspects of the moon, but new questions have emerged which demand fresh efforts to be resolved. In spite of the large database, the mystery of the origin of the moon remains unsolved. With the development of new technology, there is a renewed interest in lunar exploration in many countries. Here I attempt to trace the major advances in lunar science and present a case for new missions to the moon.

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Lunar exploration and its scientific context

It has been over 30 years since *Apollo 11* returned the first batch of lunar samples. A vast amount of data have since been obtained from the study of rocks and soils from *Apollo* and *Luna* sample collections and of nearly twenty lunar meteorites (collected from Antarctica) which presumably originated from the moon and fell on the earth after a brief journey through space. These studies provide many constraints on the original source material of the moon, its subsequent differentiation into crust and mantle and its cratering history. A calendar of events has been constructed based on isotopic dating of these rocks which represent different regions of the moon. Geophysical experiments conducted on the lunar surface and remote sensing from lunar orbiters and the earth have provided additional data sets. Based on these results, some internally consistent though complicated models of origin and evolution of the moon have been proposed.

In addition to the origin and evolution of the moon, there are several other aspects which are of scientific interest. The moon represents an important link in the series of bodies which formed as the dust and gas of the

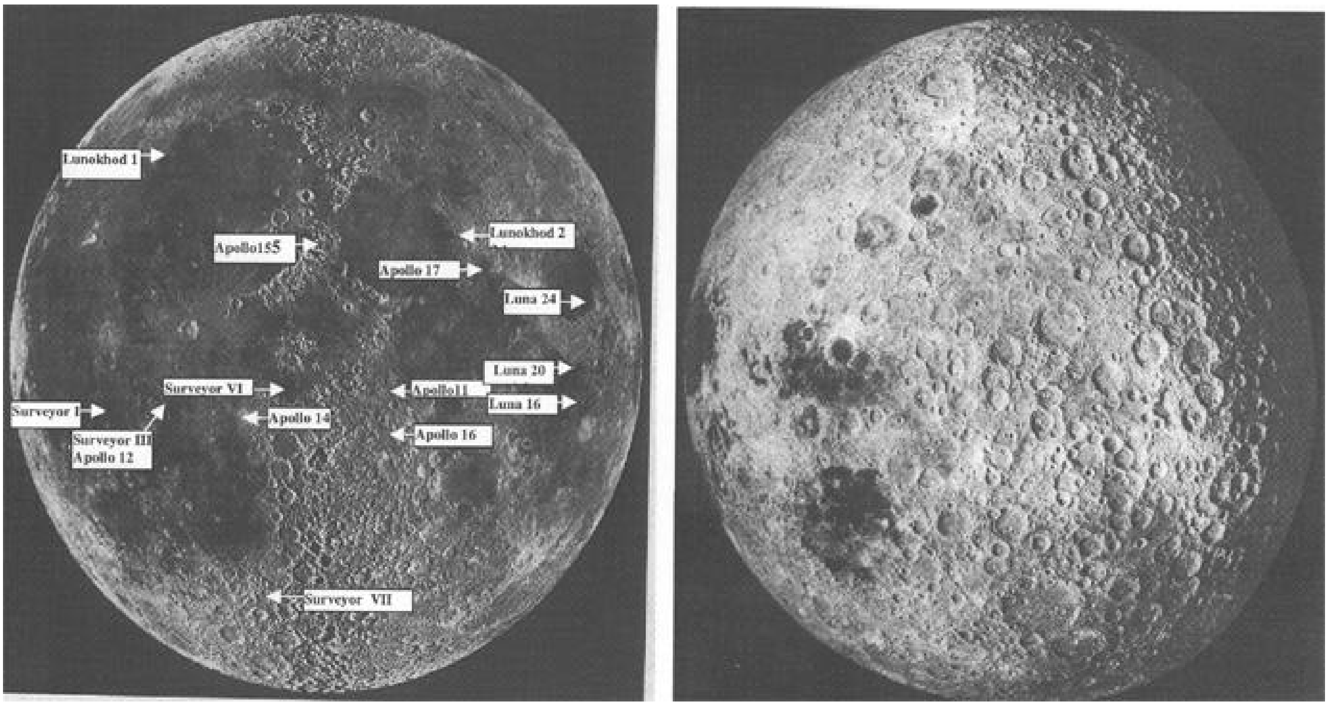

Figure 1. Earth-facing and far side pictures of the moon showing various lunar landing sites. These include the unmanned surveyor and Lunokhod sites for exploration of the moon and the *Luna* and *Apollo* (manned landing) sites from where the samples were brought back to the earth. The far side picture was taken by *Apollo 16* (Photo credit: NASA).

solar nebula condensed into grains, the grains aggregated into pebble-sized rocks and the rocks in turn grew into planetesimals, many of which finally accreted into planets. The moon may represent a preferred population of planetesimals that were formed in the infant solar system, since there are many planetary satellites (Ganymede, Callisto, Titan, etc.) in the solar system which are similar to the moon in size. The subsequent evolution of planets was size-dependent, the small asteroids and the moon stopped evolving soon after their formation (within $\sim 5 \times 10^6$–10^9 years), whereas large planets like the earth and Venus are still active and evolving. The moon and the earth, because of their proximity to each other, must have been subjected to the same interplanetary processes such as impacts and radiation. These records have been obliterated from the earth because of inherent geologic processes, but are preserved on the moon. Therefore a study of the moon can enable us to understand the physical environment of the ancient earth and its early evolutionary processes. In fact, the origin of the moon and the earth are intimately related as I shall discuss later in this article. In addition, being devoid of atmosphere and magnetic field, the moon has unmodulated record of solar particles in the form of solar wind, solar flares and energetic galactic particles from nearly the time the sun was formed. The moon can thus serve as a monitor of the solar and galactic radiations.

However, the main problem about the moon is the mystery of its origin. What is this mystery? Simply stated, how did the earth come about acquiring such a large satellite, which in many ways resembles the composition of the earth's mantle? Planetary satellites are either captured objects, as is the case with the two satellites Phobos and Deimos of Mars, or are formed by 'co-accretion' with their parent planet. Some of the satellites of the outer planets Jupiter, Saturn and Uranus, may be examples of co-accretionary processes. In case of the earth and the moon, co-accretion is ruled out as they do not have the same bulk composition (and density). The mass of the moon compared to the earth (M_M/M_E) is 0.012 whereas for other satellite–planet systems it is < 0.001, and its low density ($\boldsymbol{r}_M = 3.34\ \mathrm{g/cm^3}$ compared to $5.52\ \mathrm{g/cm^3}$ for the earth) makes the moon an enigmatic object. These observations led Harold Urey[1] and others in the pre-*Apollo* era to propose that the moon is a captured object, probably very primitive because its bulk density is similar to that of primitive chondrites. With myriads of objects going around in the solar system, capture is indeed a very probable process. Recent space missions have shown that even small asteroids have captured objects orbiting around them. The *Apollo* results indicating similarity of the composition of the moon with the earth's mantle, however, argued against such a simple capture. The moon thus, does not fit either mode of formation.

The five years following the *Apollo 11* landing on the moon in July 1969 witnessed tremendous advances in lunar science as a result of analyses of dust and rocks brought back by *Apollo* and *Luna* missions, and study of lunar surface processes based on direct measurements using instruments placed on the moon and on-board lunar orbiters. The landing sites of the six *Apollo* and the three *Luna* (16, 20 and 24) missions were close to the lunar equator on the earth-facing hemisphere (Figure 1) and therefore are not representative of the whole moon. However, subsequently more than 20 lunar meteorites were recognized from the Antarctic meteorite collection. Their analysis revealed that most of them originated from areas different from the *Apollo* and *Luna* sampling sites. Some of them may be from the far side of the moon, although it has not been possible to identify the place of their origin with certainty.

Interest in lunar science was recently renewed when the imaging system on-board *Galileo* sent pictures of some of the previously unexplored regions of the moon. *Galileo* identified a large impact basin, about 2500 km in diameter and 10 to 12 km deep in the South Pole Aitken Region (SPAR) on the far side of the moon, which was not recognized by the earlier missions. In 1994, the joint European–American Clementine mission[2], equipped with a Laser Image Detection and Ranging (LIDAR) System and High Resolution Cameras (HIRES) photographed the lunar surface in ultraviolet, visible, near IR and long-wave IR bands and provided the first global data sets for lunar gravity, topography and multispectral imaging. These results enabled mineralogical mapping of the lunar surface. The data allowed identification of several features in the SPAR, mineralogically and chemically similar to the expected lower crust or upper mantle of the moon which may have been excavated during the formation of this large impact basin. The Lunar Prospector[2] was a technological marvel with a number of sophisticated instruments on-board which included gamma-ray spectrometer (GRS), neutron spectrometer, α-spectrometer, magnetometer, electron reflectometer and Doppler gravity experiment. Apart from preparing chemical maps of the moon for elements like Th, K, Fe, Al, etc., it also identified the presence of hydrogen-bearing compounds (probably water?) in the permanently shadowed north and south pole regions of the moon.

Several hypotheses of the origin of the moon have been proposed before the *Apollo* era: (a) fission from the earth; (b) co-accretion with the earth and (c) capture by the earth. In view of the results obtained from the Soviet lunar exploration in 1966, which indicated basic (40 to 50% silica) and ultrabasic (< 40% silica) rocks as the major constituents of the lunar surface, and from the laboratory analysis of returned lunar samples, it became clear that none of these hypotheses are consistent with the new experimental evidence. New models, within as well as outside the framework of the general planet-forming processes, have therefore been proposed. At present, the most acceptable hypothesis includes the disintegrative capture of a giant planetesimal and collisional ejection of terrestrial material (out of which the moon formed) known as the giant impact hypothesis[3].

Most of the experimental results and models have been the subject of several recent conferences[3,4]. However, many of these studies have considered the earth–moon system in isolation and not within the general framework of formation of planet–satellite systems. Absence of satellites in case of the three inner planets Mercury, Venus and Mars (the two small Martian satellites Phobos and Deimos are recently captured objects and would not last long enough to be considered as permanent satellites) indicates that satellites do not form as a natural consequence of planetary formation process. The co-accretion, fission, capture or collisional ejection models should work as well on Venus (since Venus is in many ways similar to the earth), if not on the much smaller Mercury or Mars. Therefore, the moon must have become the earth's satellite in a very special way.

Synthesis of lunar data

The important physical, astronomical, chemical and geological features bearing upon the origin and evolution of the moon are summarized below and discussed within the framework of various models proposed for the formation of the moon. The basic orbital parameters of the moon and the earth are listed in Table 1.

Physical features

(i) The bulk density of the moon (3.344 ± 0.003 g/cm^3) is much smaller than that of the earth (5.52 g/cm^3) and the other inner planets (Venus, 5.23 g/cm^3; Mercury, 5.46 g/cm^3; Mars, 3.92 g/cm^3). The moon's density is much less than the earth's density even when corrected for gravitational compression (4.45 g cm^{-3}), but is similar to the density of the earth's mantle. This suggests that the moon may be depleted in heavy elements (like iron), relative to the inner planets. If the density difference is entirely due to compositional difference, then the moon should have only 25% of the cosmic proportion of iron. Ringwood[5] suggested that the low density can partly be explained if the oxidation state of iron (FeO or Fe_2O_3) in the moon differs from the inner planets.

Table 1. Basic parameters of the moon and the earth

	Moon	Earth
Semi-major axis	384.4×10^3 km	1.000 AU
Revolution period	27.32 days	365.26 days
Orbit inclination	18.3° to 29° w.r.t. earth's equator	0.00 w.r.t. ecliptic plane
Eccentricity	0.055	0.017
Obliquity	6.7°	23.4°
Rotation period	27.32 days	23 h 56 min
Radius	1738 km	6378 km
Mass	7.35×10^{22} kg	5.98×10^{24} kg
Mean density	3340 kg/m^3	5520 kg/m^3
Escape velocity at surface	2.38 km/s	11.2 km/s

(ii) The density of the moon is similar to that of some stony meteorites, but significant differences in bulk chemical composition preclude formation of the moon directly by accumulation of stony meteorites.

(iii) The centre of mass of the moon is not at its geometric centre (Figure 2) but about 0.9 km to the north and about 1.1 km nearer to the earth. This indicates that the moon is not homogeneous.

(iv) The normalized moment of inertia of the moon is 0.391 ± 0.002, which is slightly less than that of a homogeneous sphere (0.400). While the measured moment of inertia is consistent with a slightly stratified mantle, it implies either a very small (< 450 km) iron core or no core at all[6]. The moments of inertia towards the earth and along the polar and equatorial axes of the moon indicate departures from isostatic equilibrium, implying that the moon was probably never totally molten.

(v) The earth–moon mass ratio is 81.3, much less than in other planet–satellite systems ($\geq 10^3$).

Astronomical features

(i) The moon is at present in a synchronous orbit around the earth. Observations of its orbital parameters show that it is retreating from the earth and its orbital period is increasing. The current rate of increase of the semi-major axis a is + 3.2 cm/y, inclination $i = -1''.9 \times 10^{-6}$/y and eccentricity e is $+1.2 \times 10^{-10}$/y. At the same time the earth is slowing down at the rate of 1.8×10^{-3} s/century. Extrapolating the moon's drift back to the beginning, i.e. 4.5×10^9 years ago, it could have been within ten earth radii (even within its Roche limit of $\sim 2.89\, R_E$ or 18400 km).

(ii) The earth–moon system is in the ecliptic, which is inclined by about 6° to the equatorial plane of the sun.

(iii) The lunar orbital plane is inclined to both the earth's equatorial and orbital planes and by 5.15° to the ecliptic.

(iv) The orbit of the moon is such that its polar regions are not exposed to sunlight. The temperature is about +130°C on the sunlit side, whereas it is about –170°C on its dark hemisphere.

(v) The angular momentum of the earth–moon system (i.e. sum of rotational angular momentum of the earth and the orbital angular momentum of the moon) is unusually high, $\sim 3.45 \times 10^{41}$ rad.g.cm^2/s compared to other bodies of the solar system.

(vi) Bodies of size comparable to the moon abound in the solar system (Ganymede 5280 km; Callisto 4840 km, Io 3440 km, Europa 3130 km, Titan 5140 km, Triton 4400 km and Pluto 3100 km) and this may indicate a preferred size range of planetesimals formed in the early stages of the solar system.

Chemical and isotopic features

Conventionally, the moon's surface is divided into two major rock types, the bright crustal highlands and the dark basaltic mares. Analyses of *Apollo* and *Luna* samples showed that, in addition, there are rocks rich in K, REE, P, etc., called KREEP. Based on detailed chemical and geologic nature, three major zones are distinguishable on the moon: The Feldspathic Highland Terrane (FHT), Procellarum KREEP Terrane (PKT) and the South Pole Aitken Terrane (SPAT). The FHT mainly consists of calcic plagioclase feldspar, rich in calcium (11.3%) and aluminum (13%) and poor in iron (5.1%) and magnesium (4.1%). The large impact basins, mostly on the earth-facing side (Figure 2), are filled with basalt. Mare basalts cover about 17% of the moon's surface. The basalts have high iron (13–15%) and low calcium (~8%) and aluminum (4.6–7.2%). The composition of mare basalts is similar to a group of meteorites called basaltic achondrites (eucrites). In fact, based on their chemical similarity, the eucrites were once proposed to originate from the moon. However, significant difference in isotopic composition of eucrites and lunar basalts indicates that they were formed from different source materials, though in similar magmatic processes on different bodies of the solar system. The KREEP is low in Ca and Al and enriched in incompatible elements like K, REE, P and U. It forms a minor constituent (2–3%) of the lunar surface. The SPAT consists of a primitive impact basin where the deep crustal or upper mantle material may be exposed. Various groups of rocks found on moon, based on their chemical composition, are listed in Table 2. Based on some realistic models, the bulk moon composition has been calculated, which is also given in Table 2. The composition of the moon compares well with the primitive mantle of the earth, particularly with major and refractory elements although many volatile elements are relatively depleted in the moon[5,7], indicative of a heating episode. Similarity of refractory elements in the moon and the earth's mantle argue in favour of their common source material. However, the chemical composition is process-specific and not uniquely diagnostic of the source material, but the definite proof for identical source material for the earth and the moon comes from the isotopic systematics of oxygen[8] which is source-specific and does not depend on the processes of chemical differentiation.

Several features stand out when the chemical composition of the moon, grouped as a function of volatility is compared with the earth, particularly its mantle and meteorites. The most important observations (Table 2) are:

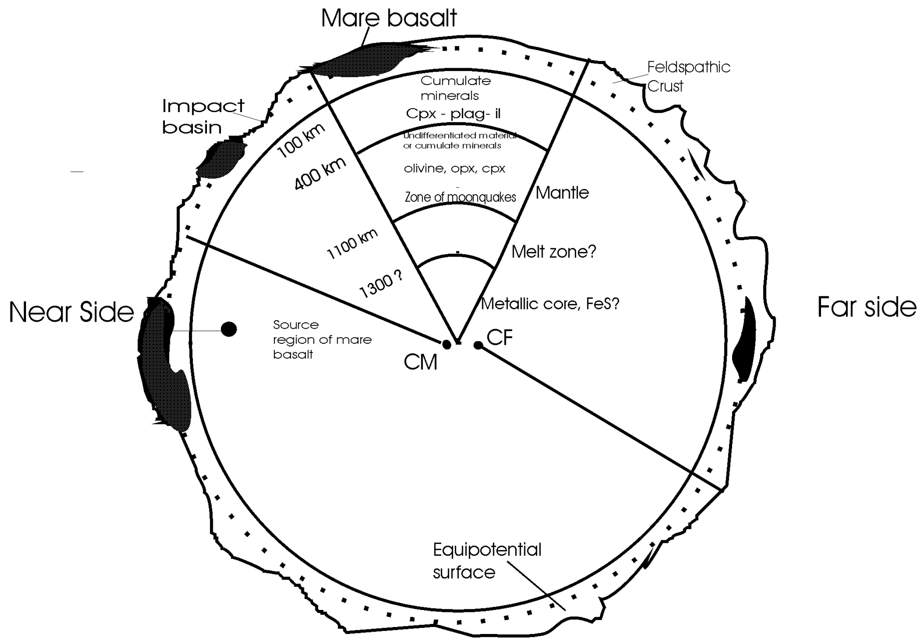

Figure 2. External shape and internal structure of the moon (not to scale). The centre of mass (CM) and geometrical centre (CF) are not coincident. The major minerals expected in various zones are shown.

Table 2. Major composition of model[7] bulk moon and various typical lunar rocks

		Highland rock						Mare basalt		
Element	Bulk moon	Highland crust	KREEP basalt	Ferro anorthosite	Norite	Gabbronorite	Troctolite	High Ti	Low Ti	High Al
Al (wt %)	3.17	13.0	8.95	18.85	7.93	6.98	10.53	4.66	5.66	7.19
Mg (wt%)	19.3	4.1	4.70	0.16	7.5	7.68	12	5.04	3.9	5.1
Ca (wt%)	3.22	11.3	6.73	14.56	6.50	8.28	7.71	7.64	8.43	8.00
Si (wt%)	20.3	21.0	24.47	20.74	23.81	23.63	20.04	17.6	21.57	21.62
Fe (wt%)	10.6	5.1	7.02	0.163	7.69	7.70	3.89	15.3	15.46	13.05
Na (wt%)	0.06	0.33	0.65	0.27	0.30	0.67	0.17	0.27	0.21	–
O (wt%)	40.4	43.1	44.53	46.28	44.15	44.06	44.35	40.3	41.5	42.33
Ti (wt%)	0.18	0.336	1.14	0.012	0.20	0.24	0.03	7.8	2.4	1.56
U (ppm)	0.033	0.24	3.37	0.002	0.60	<0.6	0.054	0.13	0.22	0.59
K (wt%)	0.0083	0.06	0.44	0.01	0.14	0.014	0.02	0.041	0.06	0.08
Th (ppm)	0.125	0.90	10.3	0.004	1.84	0.19	0.19	0.34	0.84	2.1

(i) The highland samples of the moon are enriched in calcic plagioclase feldspar, some being close to 95% anorthosite. The average highland may be ~75% plagioclase.

(ii) Depletion of volatile elements (K, Rb, Cs, etc.) by about an order of magnitude compared to the most primitive meteorites (C1 chondrites) and by a factor of about 1.6 to 2 relative to the earth's mantle and enrichment (factor of ~ 2) of refractory elements (Sr, U, La, Eu, V).

(iii) The rare earth element distribution shows no evidence for the moon's material to have been heated above 1100–1200 K, precluding vaporization as the main cause of depletion.

(iv) Concentrations of siderophile elements Fe, Ni, Co, W, P, Os, Ir, S, Te and Se are similar (within a factor of 2) in mare basalts and terrestrial oceanic basalts, suggesting analogous source materials. Some siderophile elements like P, Mo, Ge and Re are more depleted in the moon than in the earth's mantle suggesting that, in addition to the primary fractionation, a secondary depletion occurred on the moon, probably during the core-forming process.

(v) Isotopic composition of K (^{39}K, ^{41}K) in the moon is the same as in C1 chondrites and the earth, arguing against any isotopic fractionation of potassium between the earth, moon and the solar nebula. The giant impact hypothesis envisages high-temperature processes which should have resulted in significant isotopic fractionation in volatile elements like potassium. Therefore, one has to postulate that the material was transferred from the earth to the moon in bulk without any fractionation.

(vi) The moon has about 13% FeO compared to only 8% in the earth's mantle.

(vii) The remarkable similarity among V, Cr and Mn in the earth's mantle and the moon points to their common origin. The comparison of concentrations of various elements suggests that the moon must, in large parts, consist of the earth's mantle like material which was reheated and whereby some volatiles were depleted.

(viii) Normative mineral composition for the moon is 34% olivine and 66% orthopyroxene, which will match the observed density.

(ix) The earth and the moon have the same relative proportion of $^{18}O:^{17}O:^{16}O$, which is also similar to the enstatite chondrites, but different from other meteorites[8] (carbonaceous chondrites, H and L type chondrites, eucrites and the Martian (SNC) meteorites) representing their source composition (Figure 3). Different proportions of oxygen isotopes in various bodies of the solar system show that the solar nebula was isotopically inhomogeneous. Identical $^{18}O:^{17}O:^{16}O$ for the earth and the moon suggest a common source material and therefore, they must have formed in the close vicinity of each other, probably out of enstatite chondrites.

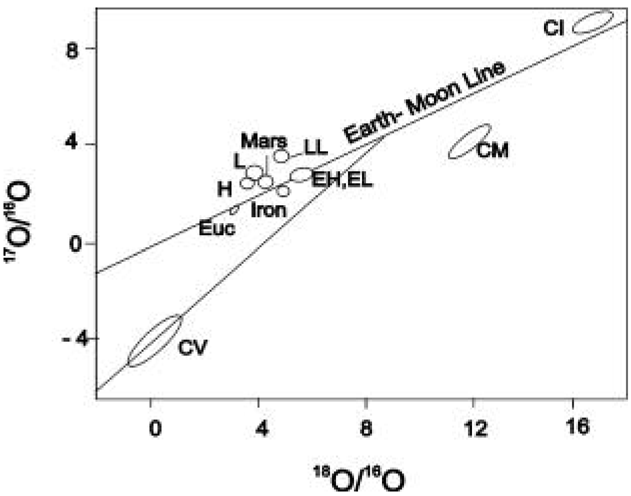

Figure 3. Oxygen isotope systematics of lunar samples, earth and various types of stony meteorites. Data for the earth–moon system and enstatite meteorites fall on the same fractionation line which is different from other meteorites (after Clayton[8]).

The timing of volatile depletion from the moon is important from the point of view of early heating events. A clue comes from the initial $^{87}Sr/^{86}Sr$ ratio in the moon which is similar to basaltic achondrites[9]. Since ^{87}Sr is derived by the decay of the volatile ^{87}Rb, the volatile (i.e. Rb) depletion should have taken place early, i.e. before $\sim 10^8$ y required for planetary accretion[3,7].

Water on moon

Water can be expected on the moon in spite of its weak gravity, because comets and meteorites which contain water have been hitting the moon all through its history. In addition, some juvenile water existing since its formation may still be preserved. Solar-wind protons impinging on the moon can reduce the oxides present on the lunar surface and produce some water molecules. However, even trace amounts of water were not found in lunar rocks and soils. Since the lunar surface essentially acts as a still because of high temperature (~130°C) on the sunlit face and low temperature on the dark hemisphere (–170°C), water and other volatiles are expected to be distilled-off to the cooler hemisphere, eventually depositing in the permanently shadowed polar regions of the moon. The Lunar Prospector carrying a neutron spectrometer found a reduction of epithermal neutron fluxes around the north and south poles of the moon, indicating that a thermalizing element was present in these regions[10]. Whether this thermalization of neutrons is due to the presence of water or some hydrogen-bearing organic compounds has been debated, but it is difficult to arrive at any conclusion. Assuming the signal to be entirely due to water, it is estimated that $\sim 2 \times 10^9$ tons of water is spread over 2.2×10^3 km^2 and 10^3 km^2 of the south and north poles respectively[10].

Geologic features

Study of surface features, ages and chemical composition of lunar samples, and the seismic experiments conducted on the moon to determine its internal structure have revealed that several geologic processes have been operative on the moon. Some of the important observations are summarized below.

(i) The primitive crust of the moon is still preserved in large parts. The average thickness of the lunar crust is 61 km (near-earth side, 55 km; far-side, 67 km), occasionally exceeding 100 km over large areas, comprising 12.3% of the lunar volume. The crust is feldspathic (density 2.76 g/cm^3) and hence it can be surmised that the moon, very early in its history, was covered with molten magma (~3.35 g/cm^3) in which this low-density anorthositic fraction floated and subsequently cooled to form the crust.

(ii) Evidence of impact by large planetary bodies (≥ 100 km) is preserved in the form of brecciated rocks around large ringed mares, but these impacts occurred quite late in the history of the moon (4.26 to 3.85 Ga) as indicated by their ages (Table 3).
(iii) There are about 50 large impact basins on the moon. Those on the earth-facing side, because of thinner crust, got filled with basaltic magma from the lunar interior, whereas many on the far side are still preserved as impact basins.
(iv) The presence of mass concentrations (MASCONS) gives a positive gravitational anomaly at the large mare basins and is due to higher density of basalt compared to the surrounding crust. The observed range of extreme deviations is about 900 mgal, about 0.5% of the lunar gravity. In a basin, the anomaly is anti-correlated with height.
(v) Seismic measurements are sparse and their interpretation is somewhat uncertain. However, the data indicate that the upper mantle consists of predominantly olivine/pyroxene composition with MgO/(MgO + FeO) in the range of 0.70 to 0.85 (Figure 2). Seismic velocities at > 500 km depth are consistent with a large abundance of garnet or an increase in MgO/(MgO + FeO), characteristic of mafic silicates.
(vi) There is no direct evidence for a lunar core. However, models indicate a small core of about 330–460 km radius, less than 1–3% of the total moon by weight. The central density may be too low for iron and is consistent with FeS or silicates.
(vii) Average heat flow (29 $\mathrm{mWm^{-2}}$) is less than half that of the earth, indicating uranium higher than chondritic but lower than terrestrial values.

Chronological constraints

Ages of various components present in lunar rocks and dust give information on the timing and processes responsible for their formation. For example, dating of the highland rocks allows us to determine the formation age of the lunar crust (FHT). Ages of impact-generated breccias and melts define the period of large impacts on the moon. Ages of basaltic fragments in lunar soils and volcanic glasses give the span of basaltic volcanism (mare-filling) on the moon.

There are several methods available for dating rocks based on radioactive nuclide-daughter product systematics. These include Pb–Pb, Sm–Nd, Rb–Sr and ^{40}Ar/^{39}Ar methods. Extinct radionuclide systems, e.g. ^{244}Pu–Xe, ^{192}W–^{192}Hf, etc. enable us to determine the time elapsed between the initiation of solar system formation, when the radionuclides were injected into the solar nebula and their incorporation in rocks. *Apollo 16* landed on the highland region and collected many crustal rocks. Based on detailed work, the oldest crystallization age based on Pb–Pb method is found to be 4.5 Ga for one of the *Apollo 16* rocks. One may therefore infer that the moon was formed by 4.5 Ga. The Sm–Nd closure age of this rock is 4.440 ± 20 Ma, indicating that the crust was already formed by this time. The ages of breccias, defining the period of major impacts, range between 4.26 and 3.85 Ga. The oldest basalts are dated at 4.3 Ga and the age distribution (3.08 to 3.85) indicates that volcanism continued, off and on, for about 800 m.y. or so. The ages of mare basalts can also be determined by crater counts on the freshly created mare surfaces[11]. The radiometric ages of mare basalts and the crater count ages of mare surfaces are given in Table 3. Based on the distribution of ages and chemical composition, a sequence of major events on the moon has been constructed. These are listed in Table 4.

Any model for the formation and evolution of the moon must be consistent with the various features summarized above. The emerging scenario is that the moon was formed close (~10 R_E) to the earth, about 4.5 Ga and was at least partially molten resulting in a global magma ocean. This led to the separation of the crust and possibly the core and was followed by a period of large impacts on the moon and subsequent eruption of basaltic lava in the basins as summarized below.

Table 3. Radiometric ages of the moon rocks and crater count ages of mare surfaces[11]

Mare	Breccia age (10^9 y)	Mare basalts age (10^9 y)	Mare surface crater count age (10^9 y)
Serenitatis	4.26	3.79 ± 0.2	3.89 ± 0.01
Nectaris	4.25 ± 0.50		3.92 ± 0.03
Fecunditatis	4.20 ± 0.04	3.50	
Tranquillitatis	4.20 ± 0.05	3.70 ± 0.2	
Humorum	4.11 ± 0.04		
Crisium	4.05 ± 0.05		3.89 ± 0.02
Imbrium	3.95 ± 0.05	3.30 ± 0.1	3.85 ± 0.03
Orientale	3.85 ± 0.05		

Early events on the moon

Magma ocean and crystallization of the lunar crust

The formation of the lunar crust early in its history indicates that, soon after its formation, the moon had a magma ocean in which the low-density feldspathic crust floated and crystallized. How much of the moon was molten is still not certain. The igneous rocks of the crust (Table 2) can be grouped according to their mineral composition into ferroan-anorthositic, magnesian and alkali suite of rocks. The ages of ferroan suite range from 4.55 to 4.36 Ga, the magnesian suite from 4.55 to 4.1 Ga and the alkali suite is younger, 4.34 to 3.85 Ga. Mineral compositions and assemblages show that most of the

Table 4. Chronology of the major lunar events with respect to formation of solar system

Formation of solar system (oldest meteorite)	t_o = 4566 Ma
Giant impact on the earth and formation of the moon?	
Formation of magma ocean and solidification of lunar crust	4530 ± 20 Ma ~4400 Ma
Lunar cataclysm (late heavy bombardment) by large planetesimals	4250–3850 Ma
Mare volcanism	3800–3200 Ma

rocks formed at shallow depths and only a few of them formed at 40 to 50 km. Rocks having the composition of dunite, granite or felsite are extremely rare. Therefore the depth of the magma ocean should be small, between half and a quarter of the moon. It is difficult to imagine separation and formation of a metallic core without the whole moon being molten. The composition and size of the core can settle this question. Radioactive heating is proposed for melting of the moon, but olivine and orthopyroxene cumulates which sink down in the magma ocean hardly contain any radioactive heat sources, making it difficult to produce even partial melts at 1150–1450 km.

Late heavy bombardment

Models of planet formation show that significant amount of debris is left in heliocentric orbits after their formation. Wetherill[12] used Monte Carlo technique to simulate the dynamical behaviour of these bodies, particularly in relation to the inner planets. These calculations indicate that the rate of impact due to this debris decayed with a half-life of 15–20 Ma as the debris is transferred to the Mars crossing orbit by planetary perturbations. Later it returned to the earth–Mercury crossing orbit. Due to collisions, the size distribution of this debris changed drastically; whereas small bodies ($< 10^{22}$ g) got destroyed in collision, the big ones ($> 10^{23}$ g) were accreted. Wetherill[12] has suggested that these bodies are responsible for the late heavy bombardment (LHB or lunar cataclysm) which gave rise to the impact basins on the moon after a delay of about 600 Ma (Table 4) since the formation of the moon and planets. Some additional contribution to this late heavy bombardment probably came from bodies in the outer solar system, particularly if Uranus and Neptune formed late and deflected icy bodies to the inner solar system.

Observationally, evidence for LHB having occurred throughout the inner solar system comes from Mercury. The earth, Venus and Mars may have gone through a similar bombardment history, but surface processes have subsequently erased the records. The crater density on both Mercury and the moon's pre-mare crust for craters (diameter > 10 km) is ~ $200/10^6$ km^2. The moon's highland region is in saturation (steady-state crater density), whereas there is a high density of > 50 km craters on Mercury. Both Moon and Mercury have several large (≥ 1000 km) ringed impact basins which are comparable in size, e.g. Imbrium (moon) and Carolis (Mercury). The post-basin formation crater density is estimated to be $120/10^6$ km^2 which further reduced to $40/10^6$ km^2 in basin-filled regions. Thus the first decline in crater production rate occurred between 4500 and 4200 Ma and a slow decline of the peak due to LHB between 3900 and 3000 Ma.

Mare volcanism

The basins excavated by planetesimal impacts were subsequently (3.8–3.2 Ga) filled by basaltic lava erupting out of the lunar interior. What the heat source was remains a big question. Many more mares were filled on the earth-facing side because of thinner lunar crust compared to the far side (Figure 1). The global asymmetry also occurs in the composition of major terranes. For example, while Procellarum KREEP terrane (PKT) on the near side is enriched in Th and other heat-producing elements, SPAT on the far side has low Th, an indication that lower crust may be exposed in this basin.

In view of the various events, the geologic history on the moon is stratigraphically divided into five epochs: Pre-Nectarian (> 3.92 Ga), Nectarian (3.85–3.92 Ga), Imbrian (3.1–3.85 Ga), Eratosthenian (1–3.1 Ga) and Copernican (1 Ga to Present).

Models of formation of the solar system and the origin of the moon

There are several ways in which a sun-like star with orbiting planetary bodies can form from a molecular cloud. However, experimental evidence related to the early formative stages of our solar system is quite sketchy and is sufficient to define only a few basic constraints. The picture that is emerging indicates that the solar system formed from a rotating, cold, dense molecular cloud which contracted and collapsed about 4.6 Ga. There is some evidence that the collapse of the nebula was triggered by shock waves from nearby supernovae explosions. The proto-solar nebula ($M = 1.2\,M_{\odot}$) itself was neither chemically nor isotopically homogeneous. The contracting nebula became hot and the central mass ($1\,M_{\odot}$) evolved to become the sun. The peripheral gas and dust ($0.2\,M_{\odot}$), as a result of early cooling processes, led to grain formation which collapsed into a protoplanetary disc. Grain aggregation resulted in the formation of larger objects which eventually accumulated into 100–1000 km size planetesimals[13]. The planetary material went through these stages quickly, within about a million years. Melting and internal differentiation of large planetesimals into crust, mantle or core occurred within a few million years due to radioactive heating[14] from the decay of ^{26}Al. Runaway gravitational accretion formed Mars-sized bodies within time scales of 10^5 years, but formation of planets like the earth took a longer time, a few tens of million years. There is much debate on homogeneous and hot accretion of planets versus heterogeneous and relatively cold accretion. While the planets formed in this way, the protosun went through the FU Orionis stage and became a T-Tauri star. In the meanwhile, due to formation of the planetesimals, the surrounding disc of dust got cleared and the sun, going through the weak lined T-Tauri stage arrived on the Main sequence. Figure 4 schematically shows the various developmental stages in the formation of the sun and planets after the solar nebula fragmented from the molecular cloud.

According to the planetary formation scenario, several planetesimals as large as Mars (~0.1 M_E) existed in the early solar system. Deviations of planetary orbits from solar equatorial plane and of rotational axes from orbital planes suggest that the planets did not only form from a myriad of small grains randomly or orderly oriented but major addition of rotational and orbital angular momentum occurred, probably due to large-body collisions during the terminal stages of planetary formation. Wetherill[3,12] and Hartmann and Davis[15] have shown that

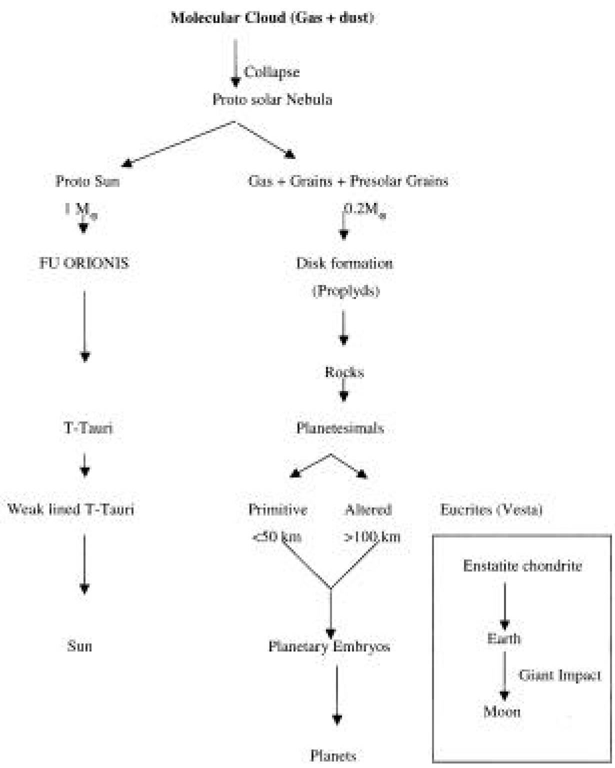

Figure 4. Schematic diagram showing evolutionary stages of the solar nebula resulting in the sun and planets.

collisions of planetesimals with protoplanets were highly probable during this period.

One of the mechanisms of formation of satellites, while the planets are accumulating by collision of planetesimals with protoplanets, requires that grazing incidence or tidal forces disrupt them and generate debris in heliocentric orbits. Subsequently, the planetesimals and/or their debris can be transferred to planetocentric orbits, which can eventually accumulate into satellites. The mechanism has been studied in detail, particularly with reference to the earth–moon system and plausible scenarios, described below, have been constructed. Although satellites can be formed in the manner envisaged above, it appears that formation of a large moon orbiting around the earth should be a consequence of rare coincidence of events.

Implications to the origin of the moon

Several models on the origin of the moon were proposed during the pre-*Apollo* era. These include fission of the earth, co-accretion with the earth and capture (intact as well as disintegrative) by the earth. These are described below.

Brush[16] has reviewed the historical development of ideas regarding formation and evolution of the moon. Based on the observation of the secular acceleration of the moon, Laplace[17] and George Darwin[18] proposed the co-accretion and fission hypotheses. The nebular hypothesis of Laplace envisaged that the moon condensed from a ring spun-off from the rotating gaseous proto-earth. According to Darwin's fission hypothesis, the moon fissioned out of the rapidly rotating earth by tidal effects of the sun. Urey[1] advanced the capture theory based mainly on physico-chemical arguments, i.e. (i) density of the moon is similar to that of stony meteorites, and (ii) dynamic calculations of Arnold[19] had shown that the probability of the earth intercepting meteorites originating from the moon is high. The analyses of lunar samples took most scientists by surprise as they did not fit in any of these hypotheses of fission, co-accretion or capture.

Fission and precipitation hypotheses

The fission hypothesis has its origin in the fact that extrapolating the moon's orbit in the past, due to conservation of angular momentum, leads to a rapidly rotating earth. Even so, the initial rotation rate is insufficient for fission and therefore Darwin[18] invoked resonance between period of tides raised by the sun on the earth and the slowest free oscillation of a fluid of earth's size, resulting in tidal bulge, growing higher and higher and ultimately breaking off from the earth. However, frictional damping in the earth cannot allow so much distortion to result in fission, although a much faster rotating earth with a period of ~2.6 h could bring about rotational instability and cause fission. Based mainly on chemical pattern of elements, Ringwood[5] modified the fission hypothesis to the 'precipitation hypothesis' which considered temperature rising to 2000 K due to rapid accretion of the earth. A hot vapour cloud rich in silica, spun-off the rapidly rotating earth. This hot cloud, already deficient in siderophiles, lost volatiles into space before condensing into an object matching with the observed lunar composition. The similarity in the composition of the moon and the mantle of the earth requires that the formation of the earth's core preceded the fission of the silica-rich clouds.

Co-accretion

The nebular hypothesis of Laplace[17] envisages that the moon condensed from a ring spun-off from the rotating gaseous proto-earth, just as the earth itself condensed from a ring spun-off from the rotating solar nebula. Recent variants of this model assume that the accreting earth accumulated a swarm of geocentrically orbiting particles, including collisional debris of the approaching heliocentric planetesimals, to grow eventually into the moon. In the binary accretion version of this model, the earth and the moon formed independently but close to each other.

Capture hypothesis

Capture was invoked mainly to account for the high angular momentum of the earth–moon system. There are several versions of the capture hypothesis. The intact capture, the disintegrative capture, the collisional ejection model and the giant impact hypothesis.

Disintegrative capture model

According to the disintegrative capture model of Opik[20], a planetesimal passing close-by the earth will be tidally decelerated, disrupted within Roche limit and captured. In the 'hetegonic' model of Alfven and Arrhenius[21], initial accretion of the moon occurred in a jet stream in the vicinity of the earth's jet stream. The accretion is mainly heterogeneous and layered. Subsequently, the moon was captured in a retrograde orbit during the early history of the earth, when it was still accreting. The earth itself had its own original satellite system which was destroyed during the tidal evolution of the captured moon, resulting in the formation of mare basins.

Giant impact hypothesis

The giant impact[3,22] or collisional ejection hypothesis for the formation of the moon takes into consideration all the post-*Apollo* data discussed in the previous section. In both these scenarios[3,22], a terminal giant impactor of the size of Mars ($M = 0.1\,M_E$), or larger, travelling at a velocity of ~5 km/s had a grazing impact with the early earth before it was fully formed ($0.7\,M_E$). According to Hartmann[3], the earth and the impactor were already differentiated into a metallic core and silicate mantle at the time of impact. As the impactor accelerated away from the earth, its metallic core separated from the mantle, decelerated and accreted onto the earth. The mantle of the impactor and ejected earth material formed a disc orbiting around the earth, partly outside and partly inside the Roche limit, broke up and accreted to form a partly molten moon. The giant impact hypothesis, in a way, can be considered as an impact-induced fission of the earth.

Cameron and Ward[22] postulated an off-axis collision that would impart much of the earth's prograde angular momentum to the debris created from such an impact. Thus, after collisions and viscous interaction between the ejecta placed in orbit have cancelled most of its random angular momentum, a disc would be left in prograde orbit around the earth. A huge amount of energy was released due to collision, so that the interface attained a temperature of about 5500 K and pressure of 600–700 bars, and a dense hot plume of vapour was formed. This model explains in a natural way the obliquity of the earth, lack of volatiles on the moon, similarity of the moon's composition to the earth's mantle, and its low density, and is consistent with the presence of many large bodies during the formative stage of the early solar system. However, the timing of the impact after core formation, makes it a very special event. Recent computer simulations[23] indicate that the moon could have formed by impact on the fully formed earth. After the impact, it took only about 24 h to form the proto-moon and the associated disc in the circum-terrestrial orbit and the moon quickly formed thereafter in a period of days. If this model is correct, then chance rather than destiny has been responsible for the formation of the moon, and it was a catastrophic and the rarest of the rare events that gave the earth its large satellite.

Some unresolved problems in lunar science

It is clear from the foregoing discussion that although a framework for the formation of the moon, consistent with the available astronomical, physical, chemical, geological and chronological data has been developed, much remains to be understood about its origin and early evolution. It is desirable to identify problems with the current models and gaps in our understanding, so that goals for future missions to the moon can be defined. Such an exercise cannot be exhaustive and, at the same time, would be subjective. Bearing these limitations, we enumerate some interesting points here.

The giant impact hypothesis, though able to explain most of the observations, appears to be ad hoc. It also considers only two bodies in isolation, a large impactor and the earth, whereas some debris from previous collisions on the earth and several moonlets may be already existing at the time of the terminal giant impact. Their role in the formation of the moon remains to be ascertained. There are also questions related to the magma ocean. The extent of magma ocean is not known, although there are indications that less than half of the moon was involved. The question whether the moon has a core or not, and its size and composition, has been extensively debated, but there are no seismological observations which can provide a direct and conclusive answer. The formation of the moon's core is also related to the size of the magma ocean. The extremely high concentration of incompatible elements (KREEP, U, Th, etc.) in the Procellarum–Imbrium–Frigoris regions on the near-side of the moon and their relatively low concentrations in the SPAR on the far side (Figures 5 and 6) have a bearing on the extent and the homogeneity of the magma ocean, chemical inhomogeneities of the crust and chemical processes responsible for formation of the large differentiated regions. Opinions are widely divided on the LHB episode. Does it really represent a peak in lunar cratering history or is it simply the tailing of the frequency of accretionary impacts? Studies of large basins on Mercury, Mars and asteroids have raised the question of this peak in cratering frequency: whether it was solar system-wide, confined to the inner planets or only to the moon[24]? The LHB was caused by planetesimals moving in heliocentric orbits or by moonlets in geocentric orbits. The composition of impactors[25] would be useful in understanding the earth's accretionary history and the nature of the source material as well. Period and mechanism of the formation of some large mare basins, e.g. South Pole Aitken basin, is not understood. It requires about a 1000-km impactor with low relative velocity to form this basin. It was probably formed about 4.2 Ga and may still have some surviving original impact melt material. A sample return mission to SPAR should be informative. The lunar cataclysm has catastrophic consequences for living organisms on the earth. The fossil records and biogenic isotopes found in terrestrial rocks show that life may have existed on the earth as early as 3.8 Ga and water as early as 4.2 Ga, well within the time span covered by large impacts on the earth–moon system. While large impacts occurred on the moon (4.2–3.8 Ga), how could life evolve or survive on the earth, which should have been subjected to at least ten times more impacts, so energetic that the oceans and environment on the earth would be sterilized? Since the time bracket of LHB is based on only a few breccias (Table 3), the cataclysmic era is not well constrained. It is important to define the cut-off time of large impacts on the moon as well as on the earth from the point of view of the biological evolution. For this reason, samples of the far side of the moon are very crucial.

Some of the problems stated above can be resolved by long cores (several km) from selected sites on the moon. Long cores through the bedrock can provide information about stratigraphic relations, composition of the lunar interior and heat flow (which depends on the radioactive content). A soil core going to the bedrock can provide insight into the nature of the solar activity (solar wind, heavy nuclei, solar energetic particles), way back in time

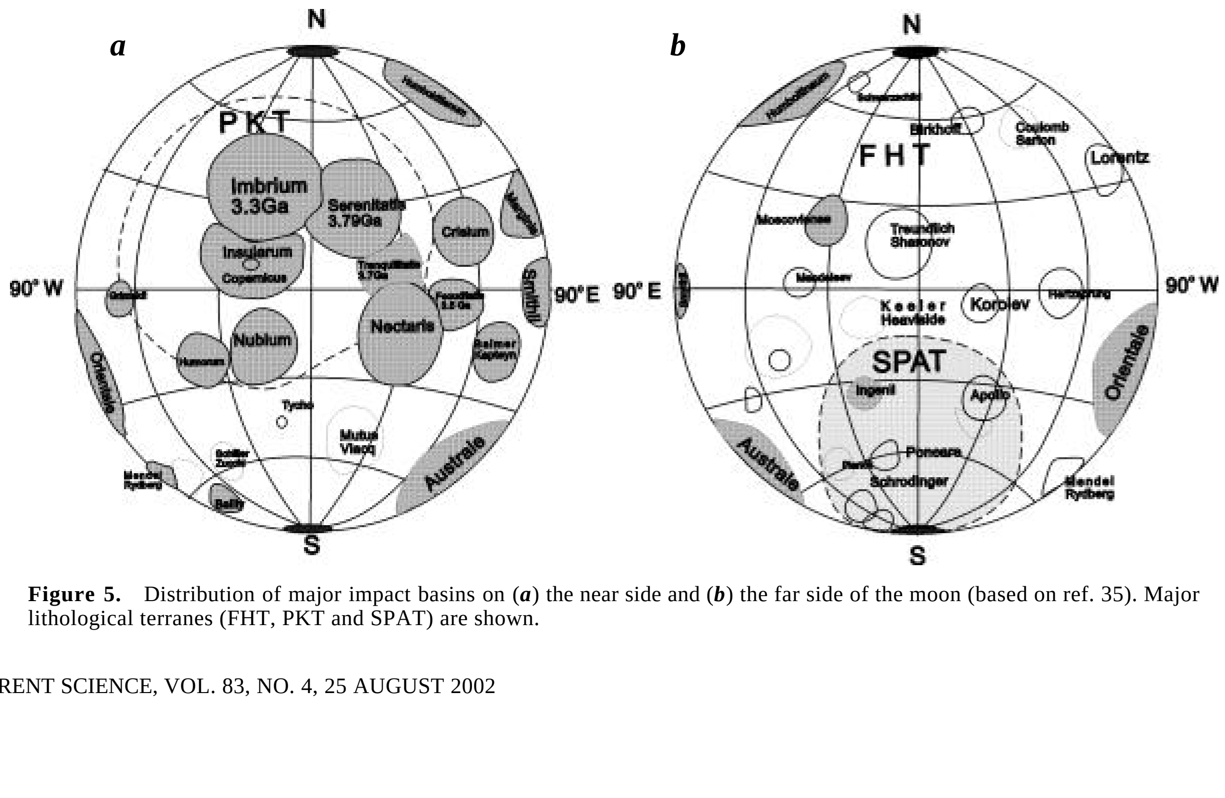

Figure 5. Distribution of major impact basins on (***a***) the near side and (***b***) the far side of the moon (based on ref. 35). Major lithological terranes (FHT, PKT and SPAT) are shown.

Figure 6. South Pole Aitken Region (Photo credit: NASA).

when the lunar crust had just formed. Their flux and isotopic composition may be related to the nuclear reactions in the sun. Bhandari *et al.*[26] have shown that the soil column broadly represents a time sequence. The 2-m long *Apollo 15* core represented roughly 1 Ga of records and therefore extending it to 4.5 Ga may not be difficult. However, taking long cores through soil or bedrock may pose challenging technical problems. It is, therefore, necessary to explore scientific programmes that are realizable and inexpensive.

Recent studies have raised questions regarding the various hypotheses related to early lunar chemical processes. For example, the thickness of individual flows within a basin is not well determined. Volcanism on the far side of the moon and its chronology still remain to be quantified. It is believed that filling of mare basins was a prolonged process which took several hundred million years. The onset and termination of lava-filling should be precisely determined. We ought to know where the youngest and oldest basalts on the moon are exposed and what are their absolute age, composition and stratigraphy. An important objective, therefore, could be to get a time-stratigraphic sequence from the regolith as well as from mare basalts. The mineralogical stratigraphy of lunar mare basalts and crust is crucial for understanding the evolution of the moon. The best current imagery suitable for crater counts is that of Lunar Orbiter and *Apollo* missions, but is not available for the entire moon. The data are particularly lacking for the far side. Stratigraphic investigations require high-resolution imaging under relatively low sun angle, so that crater rims and flow fronts can be clearly identified. On the other hand, mineralogical studies need high sun angle so that shadows can be avoided. Clementine and Lunar Prospector have provided global mineralogical and chemical data[27,28], but superior data with high spatial resolution can be obtained with new sensors with high spectral resolution available now.

Remote sensing and lunar surface chemistry

As discussed above, two types of basalts (high and low Ti types), KREEP and anorthositic highland rocks have been identified on the lunar surface (Table 2). There may be other suites of rocks on the moon, which are yet to be characterized. Deep material thrown out during large impacts can give information about the lunar interior. Horizontal stratigraphy around large impacts, formed by the material thrown out laterally by the cratering event, may reflect vertical stratigraphy of the basin.

The chemical composition and provenance of various minerals on the surface can be easily measured by gamma ray spectrometry. Reedy[29] has discussed the mechanisms of production of gamma ray lines from the lunar surface. There are basically four sources (Table 5): natural radioactivity (U, Th, K, Rb, etc.), spallation radionuclides (^{26}Al, ^{22}Na, ^{54}Mn, etc.), de-excitation gamma rays from major elements like Fe, Ca, Al, Si, Mg, etc. which go to the excited states due to interaction of solar protons, and the neutron capture gamma rays produced mainly below the surface due to Co(n, g) and Gd(n, g) type of reactions in rare earths, etc. These mechanisms are quantitatively well understood and therefore the observed gamma-ray spectra can be deconvoluted to obtain elemental composition. Such a study was made earlier by Metzger *et al.*[30] and by Lunar Prospector Mission[2]. In the *Apollo 15* and *Apollo 16* missions a NaI (Tl) scintillator was used and the Lunar Prospector used a bismuth germanate (BGO) detector with a spatial resolution of ~ 150 km. Poor spectral resolution of these detectors did not allow correct identification of characteristic peaks for some elements, because of interference by other overlapping lines. Currently available, large solid-state detectors (HPGe, CdZnTe, etc.) can be used with much advantage for high resolution chemical mapping of the lunar surface.

X-ray fluorescence spectrometers consisting of proportional counters were carried on *Apollo 15* and *Apollo 16* missions. The spectral resolution of such counters is ~ 25% and the spatial resolution of the collimator was about 80 km. The detector technology has undergone a tremendous improvement in the past decade. The X-ray CCDs, swept charge devices, Si pin detectors and CdTe detectors can provide more than an order of magnitude improvement in spectral resolution and better collimators

Table 5. Sources of X-ray and g-ray lines from the moon

Natural radioactive series	U, Th, K, La, Lu, ^{210}Pb …
Solar and galactic cosmic ray produced radionuclides	^{26}Al, ^{22}Na, ^{54}Mn …, etc.
Deexcitation of spallation nuclides	Fe, Ti, Al, O, Si …, etc.
Neutron elastic and inelastic scattering	Mg, Al, Si, Ca, Fe …, etc.
Neutron capture reactions	Mg, Al, Si, Co, REE …

can provide good spatial resolution. Lunar Prospector has identified some chemically anomalous regions on the moon. Their origin and processes responsible for the anomalies need to be understood. High-resolution chemical mapping with high spatial resolution can thus provide crucial observations to resolve some of the problems stated above.

Lunar interior

The internal structure of the moon, particularly the presence or absence of a metallic core, remains an important puzzle. Geophysical data, e.g. heat flow and seismic data should provide valuable clues to our understanding of the structure and bulk composition of the moon. The crustal thickness models of the moon based on seismic, gravity and topographic data currently assume a single or dual layered crust. Multispectral studies of the central peaks of craters and uplift structures of basins may enable us to constrain or improve upon these simple models. These models may also allow us to infer the structure beneath the large basins. Some indications of the composition of the moon's interior are provided by the composition of pyroclastic material erupted and brought up to the surface from different depths within the upper mantle.

The model of the magma ocean depends on the composition and structure of the moon's crust. Distribution of plagioclase and corresponding cogenetic mafic minerals and incompatible elements such as KREEP on the lunar surface can be used to construct such a model. Their variability and relationship with the global differentiation layers, intrusive rocks or differentiates of basin impact melts have to be understood for this purpose. Although major highland and mare rock types are known from the *Apollo* analysis, little is known about depth of their origin. As mentioned above, the surface expression of lunar materials is related to the moon's internal structure and evolution. X-ray and γ-ray mapping can provide us with information about the location where large bodies of anorthosite crop-out on the moon. Global, high-resolution multispectral mineralogical data coupled with global chemical mapping using γ-ray spectrometry can constrain the way these materials vary laterally and vertically in the crust. Uplifted structures associated with large impact craters and basins may expose crustal igneous rocks and mineralogical remote sensing of these structures can provide important clues to the lateral distribution of different types of igneous rocks and their pre-impact depths of formation. The composition of the mega regolith can provide information about the types of rocks exhumed from basin impacts. Considering that anorthosite is the original product of a magma ocean, the magnesian suit of plutonic rocks, including norite, troctolite and gabbro should enable us to determine the structure of the crust at various places on the moon.

As discussed above, the central upland regions and ejecta of large-impact craters may contain material from the lunar interior. Large impact basins, particularly those on the far side which are not filled with basalts may also have interior material present within and around them. If the spatial resolution of X-ray and gamma-ray spectrometers is better than the size of the uplands or ejecta units, their composition can be determined and chemical stratigraphy can be constructed. It is expected that lunar composition becomes more mafic with depth and some diagnostic parameters like mg# (MgO/(MgO + FeO)) should be useful for determining the depth of their origin.

The offset between the lunar centre of mass and its geometric centre (Figure 2) implies either an increase in crustal density for the lunar near-side or a thicker far-side crust. The presence or absence of ejected mantle material from large basins and its composition may help constrain the thickness of the crust and may be useful to determine, if there is a global compositional asymmetry. Compositional asymmetry, if it predates the formation of the major near-side basins, may imply crustal heterogeneity on a global scale. On the other hand, if it was acquired during the formation of basins, then we may be able to better understand the way KREEP residua were formed.

The mare basalts, although volumetrically minor, were formed by partial melting of the lunar mantle and thus record composition, mineralogy and processes from as deep as 200–400 km. Although major volcanism on the moon occurred between 3.8 and ~3.1 Ga, significantly younger (up to 800 Ma) volcanism is indicated by stratigraphic relations and crater densities. On the other hand, volcanism as early as 4.2 Ga has been inferred from basalt clasts present in impact breccias. The sampled near-side basalts and related pyroclastic glasses cover a broad range in composition, e.g. TiO_2 varies between < 1 and 16 wt%. Little is known about the far-side mare and volatile-rich pyroclastic eruptions, which were an important part of mare volcanism and need to be mapped.

Transport of volatiles on the lunar surface

Volatiles like mercury, water and gases like radon emanating from the decay of U, Th chain nuclides continuously degas from the hot sunlit regions and get deposited on the colder dark side[31], ultimately depositing on the north and south poles of the moon which are in the permanent shadow[32]. They may also degas from faults and fractures. It is important, particularly from the point of view of habitation of the moon, to understand the transportation of volatiles on the lunar surface and their eventual deposition in the polar regions. Attempts to determine radon distribution on the moon by α-spectrometry on-board Lunar Prospector met with some difficulties due to interference from intense alpha particle fluxes from the sun, and therefore gamma-ray spectrometry may be better-suited for this purpose.

^{210}Pb, the 22.4 year daughter of radon can serve as a tracer for understanding the volatile transport on the moon. ^{222}Rn (half life 3.8 days), degassing by diffusion or through faults and fractures from the lunar interior, quickly decays into a series of daughter nuclides, resulting in a thin coat of ^{210}Pb on the lunar surface. A part of it eventually accumulates in the polar regions. The distribution of ^{210}Pb on the moon can be mapped by gamma-ray spectrometry of its 46.5 keV line.

Scientific objectives for future missions

Orbiting missions

Remote sensing of the moon is a powerful technique for geochemical mapping. Natural radiation (α, X- and γ-rays) can be used to determine the distribution of many elements. The radiation environment of the moon is shown in Figure 7. A useful approach for orbiting missions would be simultaneous chemical and mineralogical mapping and photogeology with high spatial resolution. The objectives for such missions could be: (1) Multispectral imaging to determine distribution of various minerals having different spectral response so as to map the central peak regions of large craters, to understand the composition of the lunar interior and the composition of ejecta as a function of distance from large impact basins. (2) X-ray mapping in 0–10 keV region for elements like Mg, Al, Si, Ca and Fe. The spectral resolution of the detector should be such that it can distinguish the various rock types given in Table 2. Concentration ratios of some of these elements are characteristic of lunar rock types. Some rocks can be easily distinguished by their magnesium number. (3) Mapping in the hard X-ray–gamma ray region (10–200 keV). This region has not been studied so far and contains several γ-rays of interest like the 46.5 keV line of ^{210}Pb and X-rays from radon, uranium and thorium (Table 6). Besides, distribution of some of the rare earth elements which give neutron capture gamma lines in this window can be mapped. (4) High-energy gamma-ray (0.2 to 10 MeV) mapping. This region has been scanned earlier by *Apollo* and Lunar Prospector gamma-ray spectrometers, but better solid-state detectors with higher spectral resolution now available, make it an attractive proposition for determining concentration of some of the elements with better accuracy and reliability. Of particular interest is the 2.2 MeV line of H, which may indicate presence of water, if it can be distinguished from the nearby lines due to Al, Th, etc. Also Fe, Si, Mg, Al, Ca and Ti maps will give the distribution of various minerals like ilmenite ($FeTiO_3$), feldspars, pyroxenes and olivines.

Table 6. Gamma-ray lines expected in 15–200 keV band

Element	Energy (keV)
^{222}Rn	16.77
^{210}Pb	46.5
^{238}U	48
^{234}U	53, 120
Th K_α	89.9
U K_α	94.7
Gd	103
Sm	103

The objective of an orbiter mission should be to prepare a high-resolution 3D atlas in the visible/UV, near IR regions, and to superimpose X- and γ-ray maps and study some areas of specific interest related to crustal inhomogeneity, presence of water, and the bulk composition of the moon.

Landing missions

Landing missions with geophysical sounding (conductivity, composition, seismic studies, etc.) can resolve many questions related to the existence and dimensions of the core, presence of water and chemical composition, particularly of some unexplored areas on the far side. If water is found, then measurement of D/H will provide crucial answers to the source of water, whether it is juvenile, cometary, asteroidal or was formed by interaction of solar wind hydrogen. Three interesting sites for lunar landing can be the two poles and the SPAR.

Sample return missions

Much can be learnt if samples from some critical areas of the moon can be brought to the earth for laboratory analysis, particularly chemical and isotopic composition and chronology. Interesting areas from this point of view are on the far side of the moon, the SPA basin and the north and south pole regions. These will also provide ground truth for interpretation of multispectral images from the Clementine and other future remote sensing missions.

Future missions to moon

Important missions currently being planned are SMART-1 (ESA), SELENE (Japan) and *Luna A* (Japan) missions. The mission configurations and their scientific objectives are briefly described below.

SMART-1

The SMART-1 is an acronym for Small Mission for Advance Research and Technology and is planned by

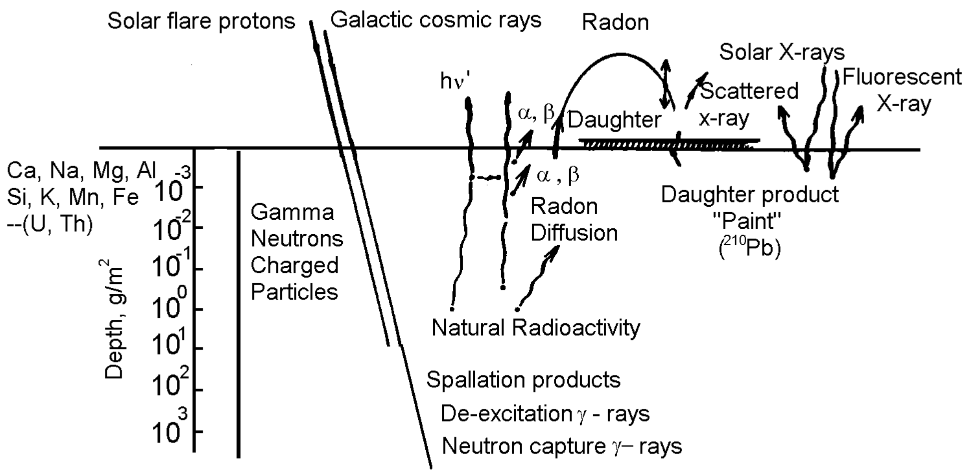

Figure 7. Radiation environment on the moon (modified from Adler *et al.*[36]).

ESA. Its primary objective is to test solar-electric propulsion; it has several other technology goals connected with deep space telemetry and communications in X and Ka bands, radioscience investigation, a laser link experiment which will provide optical link to a ground station in Tenerife, and a system of autonomous navigation based on image processing.

It will be launched by Ariane-5 rocket in early 2003. After cruising with the ion propulsion system employing xenon, the mission will orbit the moon for a nominal period of ~6 months with a perilune of 300–1800 km and apolune of 1000 km. The scientific payload of SMART-1 includes a high-resolution camera, a near infrared spectrometer and a compact X-ray spectrometer with a new type of detector (D-CIXS) and micro collimator. The D-CIXS for X-ray imaging in the energy range 0.5 to 10 keV and a solar X-ray monitor for absolute calibration will provide distribution of the main rock-forming elements, i.e. silicon, magnesium, iron, sodium, aluminum and oxygen, with a 30 km spatial resolution at perilune. During the cruise period, the instrument will monitor X-ray variability of some bright cosmic sources.

The infrared spectrometer is a monolithic grating spectrometer sensitive in the 0.9 to 2.4 micron range. It will perform spot reflectance IR spectrometry of minerals providing discrimination among the major minerals like pyroxene, olivine and feldspar. Weighing about 1.5 kg, the main purpose of the spectrometer is to provide mineralogy of the boundary zones between terranes, lunar crust stratigraphy through the spectra of crater ejecta and possible resource potential. The CCD-based micro imager weighs 45 g. It includes a teleobjective mount allowing a field of 5.3° and a resolution of 30 m per pixel from 300 km altitude and weighs ~ 1.8 kg. It will monitor illumination in permanently shadowed polar areas for subsequent polar exploration, besides contributing to surface mineralogy and geology.

The spacecraft also contains Langmuir probes mounted on short booms for monitoring the spatio-temporal variations of the plasma, electron and dust environment for measurement in the energy range of a few tens of eV and plasma density in the range of 0.1 to 1000 particles per cm^3.

SELENE

SELENE (Selenological and Engineering Explorer) is a ISAS/NASDA joint mission to be launched by H-IIA rocket in 2004. The mission will consist of a main orbiting satellite at about 100 km altitude in polar circular orbit and two sub-satellites in elliptical orbit, with apolune at 2400 and 800 km. Its scientific objectives include mapping of lunar topography and surface composition, measurement of magnetic field and study of lunar and solar terrestrial environments. The life time of the mission is expected to be at least one year. Elemental abundances are measured by X-ray and γ-ray spectrometers. Alpha-spectrometer is mainly devoted to measurement of radon and its daughter polonium. The mineralogical characterization is performed by a multi band imager with nine spectral bands ranging from 0.4 to 1.6 μm with a spatial resolution typically of 20 m. The mineralogical composition will be mapped by continuous spectrum analysis in the range 0.5–2.6 μm with 500 m resolution. The mission also includes a high-resolution terrain camera and a laser altimeter. The terrain camera has a field of view of 35 km with a spatial resolution of 10 m to provide stereo images. The laser altimeter will measure the altitude every 1.6 km, with a vertical resolution of 5 m and a spot size of 30 m.

The sounder experiment using a 5 MHz radio wave will reveal the lunar internal structure up to 5 km below the surface, with a vertical resolution of 100 m. The magnetometer provides data on the lunar surface magnetic field, which will be used to understand the origin of lunar palaeomagnetism and impact-induced paloeomagnetism. Doppler tracking of the orbiter via the relay satellite when the orbiter is in the far side will be used to determine the gravity field of the far side. Radio sources on the two subsatellites will be used to conduct the differential VLBI observation from ground stations. Studies of the lunar environment of high-energy particles, electromagnetic fields and plasma will also be made. Radioscience experiments using coherent x and s band carriers from the orbiter will be conducted to detect the tenuous lunar ionosphere. Some observations related to solar–terrestrial phenomena from the lunar orbiter, i.e. imaging of the earth to understand the macroscopic dynamics of the terrestrial plasma environment and aurora activities, are also planned. Observations of planetary radiation from Jupiter and Saturn will also be made by the orbiter.

Lunar A

The Lunar A mission of Japan, proposed for launch in 2003, is mainly for seismic and heat-flow studies. It will use two penetrators, one on the near side and the other on the far-side of the moon. Each penetrator has an accelerometer, several temperature sensors, thermal conductivity probes and seismometer systems. The life time of the penetrator is expected to be one year. The seismometer system consists of two short-period electromagnetic seismometers with a resonant period of about 1 s. These seismometers will be aligned orthogonally and will record deep moonquakes. The foci of moonquakes, the amplitude and travel time of seismic waves will provide information about the internal structure of the moon. The objective of the mission is to determine whether the moon has a core and if so, its size and physical properties.

Planetary mission capabilities of Indian Space Research Organization

The Polar Satellite Launch Vehicle (PSLV) of Indian Space Research Organisation (ISRO) is capable of undertaking planetary missions for direct exploration of the moon, Mars and Venus. It can, for example, put a 480 kg spacecraft in circum-lunar orbit at an altitude of 100 km and can carry about 280 kg and 300 kg respectively, on Mars and Venus fly-by missions. For a 1000 km circular orbiter mission to Mars and Venus, the carrying capacity reduces to about 140 and 100 kg. The Geosynchronous Satellite Launch Vehicle (GSLV), which will shortly be commissioned, can significantly enhance this capacity and can be used for missions to inner planets (Mercury, Venus and Mars) as well as to asteroids and comets.

A mission to the moon is under active study by ISRO[33,34]. The best choice for the first mission appears to be a lunar polar orbiter with a circular orbit, having an altitude of ~100 km. Considering the weight of the lunar craft and the fuel required for a nominal life of about two years, about 60 kg should be available for science payloads. Discussion of scientific objectives and choice of payloads within the weight constraints, scientific priorities and expertise available indicate that simultaneous photogeological and chemical mapping should provide a good approach to resolve some of the problems discussed in the foregoing sections. The following payloads are being considered[34]: terrain mapping camera (TMC); hyperspectral spectral imager (HySI); laser ranging instrument (LLRI); low energy X-ray spectrometer covering 0–10 keV region (LEX) and hard X–γ-ray spectrometer covering 15–200 keV region (HEX).

The TMC uses three-strip imaging configuration (fore and aft looking at $\pm 19.4^{\circ}$ and the third at nadir) and is designed to have 5 m spatial resolution and 40 km swath. It can achieve a height resolution of about 10 m. The HySI employing a wedge filter operating in 400–900 nm band, will have a ground resolution of 80 m and 32 channels will select the spectral bands of interest with a spectral resolution of 15 nm for mineralogic mapping. It can easily distinguish anorthosite (highland), basalt (mare) and some minerals expected to be present at some depth within the moon. The LEX, using well-collimated X-ray CCD or swept charge device is designed to measure the concentration of Mg, Al, Si, Ca, Fe and Ti, whereas the HEX, using a CdZnTe solid state detector and CsI anticoincidence system, would measure the distribution of ^{222}Rn, Th, U and some rare earths on the moon. The superior energy resolution and lower background of these detectors should enable us to map the abundances of Fe, Ti, U, Th, K, etc. with reasonable spatial resolution (< 100 km) using their de-excitation or decay radiations. An infra red detector covering wavelengths up to 2.5 μm (for mineral identification) and gamma-ray spectrometer based on hyperpure germanium detector may be suitable for detection of K, H and other elements of interest. Though desirable, it may be difficult to include these instruments in the first lunar mission because of cooling requirements, power and weight constraints.

This study should enable us to determine the stratigraphic correlation of various surface units in selected areas of the moon, which include some features in the SPAR and areas of the north and south poles.

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ACKNOWLEDGEMENTS. I am grateful to Dr K. Kasturirangan for encouragement and to Dr J. R. Arnold and Dr M. B. Duke for useful suggestions. Details of the Indian mission are based on the study reports prepared by various members of the moon mission task force constituted by ISRO and I would like to thank them, specially Dr G. Joseph, P. C. Agrawal, V. Adimurthy and A. S. Kiran Kumar. I also thank Dr K. Gopalan, P. N. Shukla, K. L. Pruseth, S. V. S. Murty and J. N. Goswami for several useful discussions and A. D. Shukla and K. R. Nambiar for their help in the preparation of this manuscript. This article has been written as a part the National Planetary Science and Exploration Programme (PLANEX) supported by ISRO. Constraint of space did not allow citing all the relevant references, some of which can be found elsewhere[3,4,37].

Received 23 January 2002; revised accepted 18 June 2002