The Interior Structure of the Moon: What Does Geophysics Have to Say?

Mark A. Wieczorek*

Geophysical data obtained from orbit and surface stations show that the Moon is a differentiated body possessing a crust, mantle, and core. The crust is on average about 40 km thick, and impact events with asteroids and comets have excavated materials to great depths within the crust. Moonquakes that are correlated in time with Earth-raised tides occur about halfway to the center of the Moon and suggest that the deepest portion of the mantle might be partially molten. The lunar core is relatively small in comparison with the cores of the terrestrial planets, with a size less than one-quarter of the Moon's radius.

KEYWORDS: Moon, geophysics, internal structure

INTRODUCTION

Samples of the Moon obtained from the Apollo and Luna missions, as well as from lunar meteorites found haphazardly on Earth, are a testament to the violent origin and early evolution of the Earth–Moon system. The prevailing paradigm for the origin of the Moon derived from these samples posits that a Mars-sized object collided with the proto-Earth about 4.5 billion years ago, placing in Earth orbit a large quantity of material that eventually accreted to form the Moon (Canup 2004). As a result of the huge amount of energy that would have been liberated in such an event, a significant portion of the Moon is thought to have been initially molten (Pritchard and Stevenson 2000). Fractional crystallization of this "magma ocean" would most likely have led to the formation of an anorthositic flotation crust and to the partitioning of a large portion of the Moon's heat-producing elements into the last magmas that crystallized between the crust and mantle.

The subsequent 500 million years of lunar history were largely shaped by collisions with large comets and asteroids that formed impact basins hundreds of kilometers in diameter. During this time, the interior of the Moon slowly heated up from the decay of radioactive elements, and at some point before 4 billion years ago (Ga), the lunar mantle partially melted, giving rise to a thin veneer of flood basaltic lava flows that are visible today to the naked eye (Shearer et al. 2006). And then, around 3.5 Ga, the geologic activity of the Moon slowly came to end. While some lavas continued to erupt until about 1 billion years ago (Hiesinger et al. 2003) and "small" impact craters continued to form at a reduced rate, not much of great significance happened.

Most of the Moon's geologic evolution dates from a time that is poorly preserved on the planet we call home. As such, the Moon offers itself as a unique and invaluable laboratory to study processes that surely affected the Earth and other planets. In particular, the Moon is the only well-studied terrestrial object that currently possesses a primary crust formed during the initial stages of planetary differentiation. It possesses chemical and isotopic heterogeneities in its mantle that were frozen in place after only the briefest period of mantle convection. It also possesses some of the largest impact structures in the solar system, which, given the lack of weather on the Moon, are preserved in a relatively pristine state. However, in order to decipher the details of the Moon's origin and evolution, as well as to properly interpret the secrets that are locked inside the samples that we possess, it is necessary to have an understanding of what lies below the surface. Questions of fundamental importance to not only geophysicists but also geochemists include: What is the average thickness of the lunar crust? How does the crustal thickness vary across the surface? Is the mantle layered in composition? Has any portion of the Moon escaped primordial melting? What is the temperature of the lunar interior? And is any portion partially molten today? Does the Moon possess a metallic core? And if so, what is its size and composition? Could a lunar core have given rise to a dipolar magnetic field? All of these questions are ideally suited to investigations by geophysical means.

Although the data sets that lunar geophysicists make use of are vastly smaller than those available for Earth, a surprising number of high-quality experiments and measurements have, nevertheless, been performed both on the Moon's surface and from orbit. Indeed, a legacy of the Apollo program is the acquisition of data from long-lived geophysical stations (ALSEP, for Apollo Lunar Surface Experiment Package) that consisted of seismometers, heat-flow probes, magnetometers, and laser retroreflectors. Though we are still far from discovering all of the secrets of the Moon, the geophysical data have proven to be extremely useful. Unfortunately, more times than not, these measurements have posed more questions than they have resolved, and future missions (including a next-generation geophysical network) will certainly be required to obtain the same level of understanding of the lunar interior as we currently have for Earth.

* Institut de Physique du Globe de Paris
  4 avenue de Neptune
  94100 Saint Maur des Fossés, France
  E-mail: wieczor@ipgp.fr

THE CRUST

Two key parameters used in deciphering both lunar evolution and the lunar samples are the average thickness of the crust and how the crustal thickness varies from place to place. As an example, the average crustal thickness is believed to be a direct by-product of the Moon's initial differentiation; it therefore depends upon several factors, including the depth of the magma ocean and the efficiency with which crystallizing plagioclase was able to float (e.g. Solomatov 2000). Data obtained from seismometers placed on the lunar surface at the Apollo 12, 14, 15, and 16 stations offer the most direct means of assessing this quantity (see FIGURE 1 for the locations of the Apollo and Luna sampling stations and the outline of the Apollo seismic network). Analyses during the Apollo era suggested initially that the crust beneath the "Apollo zone" was about 60 km thick. However, recent independent analyses by two different research groups now suggest that the crustal thickness is much less in this area, probably somewhere between about 30 and 38 km (Khan and Mosegaard 2002; Lognonné et al. 2003). While one might think that this dramatic revision would have large consequences on estimates of the bulk composition of the Moon, it is now realized that the globally averaged thickness of the crust is somewhat greater than that measured beneath the Apollo stations (the globally averaged thickness is probably between 40 and 45 km; see Chenet et al. 2006 and Hikida and Wieczorek 2007). Nevertheless, as a complicating factor, several measurements show that there are also large lateral and vertical variations in the composition of the crust, which were not fully appreciated until after the Clementine and Lunar Prospector missions (Jolliff et al. 2000). In particular, orbital gamma-ray data and in situ heat-flow measurements indicate that heat-producing and incompatible elements are concentrated within a single geologic province that encompasses Mare Imbrium and Oceanus Procellarum (i.e. the Procellarum KREEP Terrane; see Wieczorek and Phillip 2000), and remote sensing data show that the surrounding highlands crust becomes increasingly mafic with depth.

One of the problems with the Apollo seismic data is that the network spans only a small portion of the Moon's central nearside hemisphere. Fortunately, the thickness of the crust can be estimated outside of the Apollo zone by using a combination of the Moon's surface relief, gravitational field, and reasonable assumptions about the density of the crust and mantle. FIGURE 2 shows the topography of the Moon (upper left) obtained by the Clementine mission (Smith et al. 1997; USGS 2002), as well as the best estimate of the radial gravitational acceleration at the surface (upper right), derived primarily from the Lunar Prospector mission (Konopliv et al. 2001). Since the gravitational field is obtained by measuring small Doppler shifts in radio signals emitted by orbiting spacecraft, the resolution over the farside hemisphere of the Moon, where spacecraft are not visible from Earth, is considerably poorer than for the nearside hemisphere. Immediately visible in the topographic image are the depressions associated with numerous giant impact basins—the giant basin on the farside hemisphere is the South Pole–Aitken basin, which is currently the largest recognizable impact structure in the solar system. Furthermore, some of the impact basins are seen to possess large central positive gravitational anomalies, which are colloquially referred to as "mascons," short for mass concentrations. While the origin of mascons is not completely resolved, it is likely that they result from a combination of dense mare basaltic lava flows (which are perhaps several kilometers thick in places) and structural uplift of relatively dense mantle materials (for a review, see Wieczorek et al. 2006).
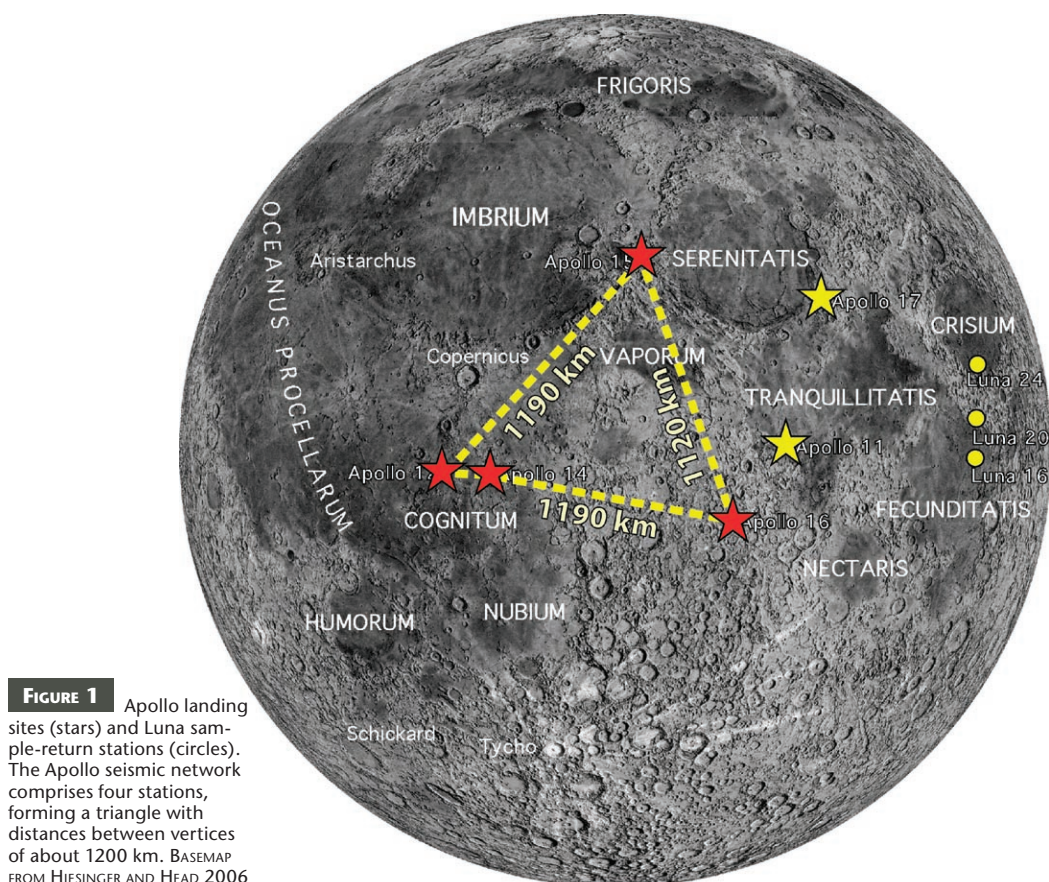
FIGURE 1 Apollo landing sites (stars) and Luna sample-return stations (circles). The Apollo seismic network comprises four stations, forming a triangle with distances between vertices of about 1200 km. BASEMAP FROM HIESINGER AND HEAD 2006
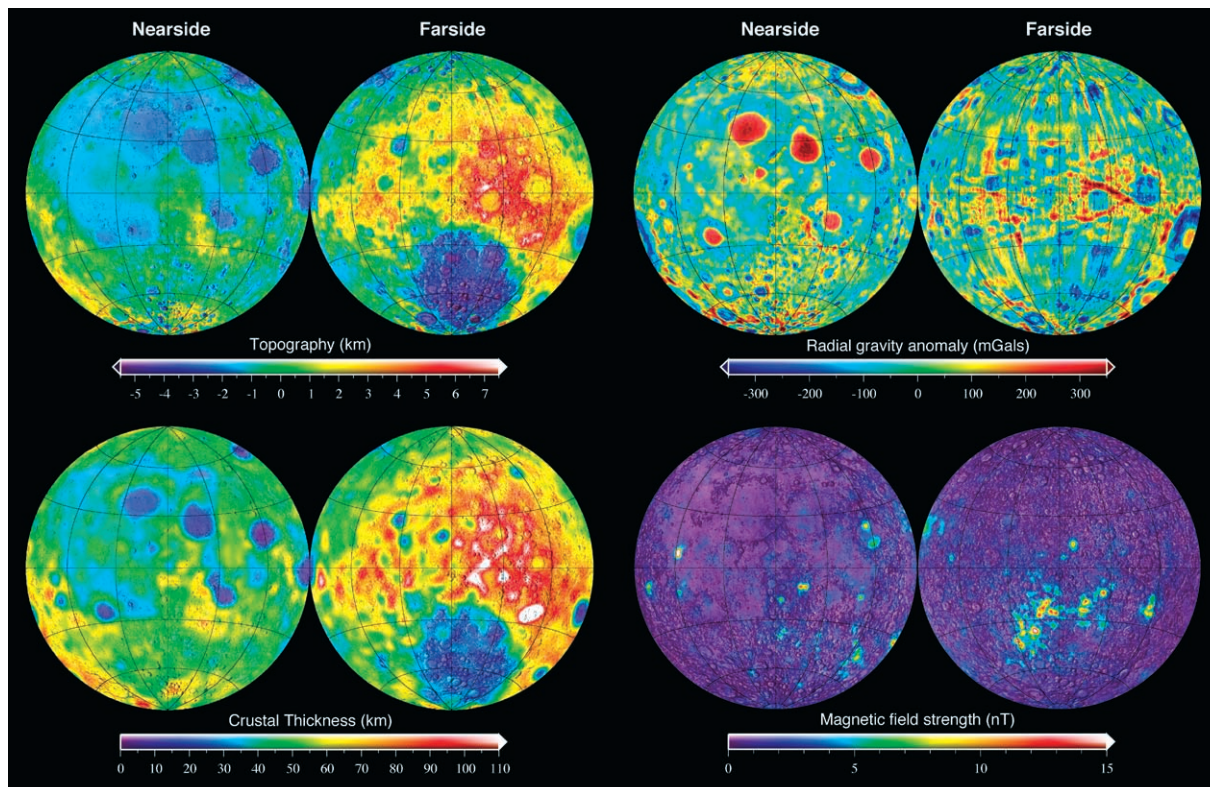
By subtracting the gravitational contribution associated with the surface relief and mare basalts from the observed gravity field, the remaining signal can be modeled in terms of variations in relief along the crust–mantle interface, providing us with a global map of the Moon's crustal thickness. One such model is shown in FIGURE 2 (lower left), which demonstrates that the crust is considerably thinned, and in some cases nearly absent, beneath the largest impact basins. This thinned crust is a natural consequence of the impact-cratering process, which is thought to excavate materials from depths of up to one-tenth of the crater's diameter. Thus, even though all of the samples that we have from the Moon were collected on the surface, many of them could have originated tens of kilometers below the surface, perhaps even in the mantle.

Although the geophysical crustal-thickness models are invaluable for deciphering both the impact-cratering process and the provenance of the lunar samples, several vexing questions remain. For instance, the model in FIGURE 2 predicts that the crustal thickness is on average greater on the farside hemisphere than on the nearside, but this could conceivably be an artifact of the geophysical model that assumes the crustal and mantle densities are everywhere the same. In addition, even though the South Pole–Aitken basin is the largest impact structure on the Moon, this model predicts that there is still a thin "crust" in the basin's interior. These materials could represent potentially deep crustal materials that were exposed by the impact (Pieters et al. 2001), or perhaps a differentiated impact melt sheet. Alternatively, the assumptions of the geophysical model might be inappropriate for this region of the Moon, and the mantle might be closer to the surface than is predicted. Given that this basin is the oldest recognizable impact structure on the Moon, it is plausible that this impact might have occurred while the crust was still forming during the terminal stages of magma-ocean crystallization.

THE MANTLE

The Apollo seismic data are the primary source of information bearing on the structure and composition of the lunar mantle. Events detected over the approximately eight-year operation of the Apollo seismic network included about 1800 meteoroid impacts, 28 energetic shallow moonquakes (with body wave magnitudes up to 5 and hypocenters about 100 km below the surface), and about 7000 extremely weak deep moonquakes that were located about halfway to the center of the Moon (FIGURE 3 shows the seismic ray paths as a function of epicentral distance and hypocenter depth for all stations). The deep moonquakes are very enigmatic in that their occurrences are correlated in some way with the tides raised by the Earth, they involve very low stress drops (less than 1 bar), they appear to originate from about 300 "nests" that are repeatedly activated, and almost all of these nests are located on the Moon's nearside hemisphere (Nakamura 2003; Bulow et al. 2007). The nearside distribution of the deep moonquakes could perhaps indicate that the farside hemisphere is seismically inactive. Such a hypothesis is possible, especially if one considers that almost all of the Moon's mare basalts erupted on its nearside hemisphere. Alternatively, it is possible that the deepest interior of the Moon attenuates seismic signals originating from farside events. In support of this hypothesis, shear-wave arrivals appear to be absent for those ray paths that probe the deepest portions of the mantle, perhaps indicating the presence of a partial melt in this zone (Nakamura 2005). The presence of a highly attenuating zone deep in the lunar mantle is also consistent with analyses of the lunar laser ranging data (these indicate a bulk solid-body "quality factor" Q of about 30, which is comparable to that of the Earth and Mars; Williams et al. 2001). One scenario that might be consistent with the
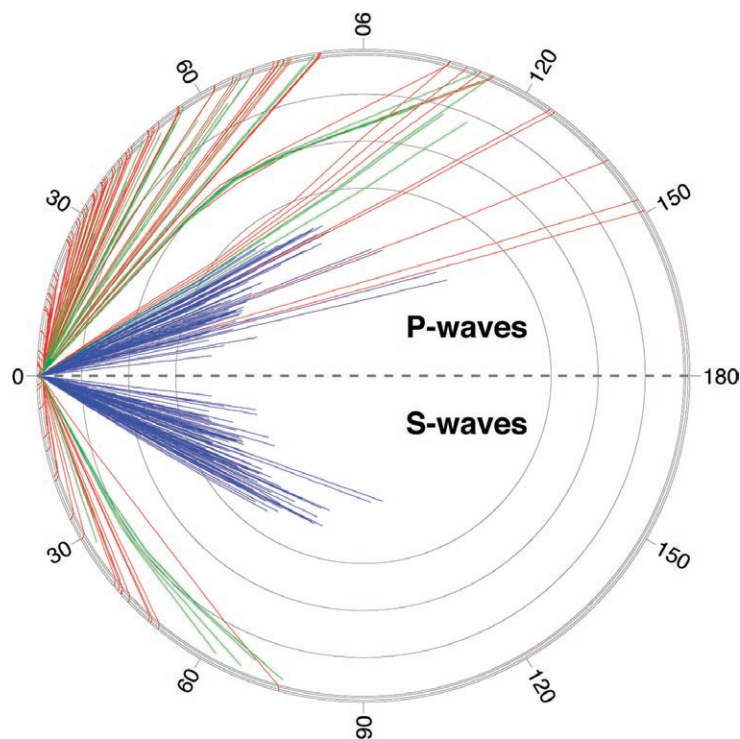
FIGURE 3 P-wave (upper) and S-wave (lower) ray paths in the Moon as viewed from each seismic station as a function of epicentral distance and hypocenter depth (after Lognonné et al. 2003). Deep moonquakes are in blue, shallow moonquakes are in green, and the artificial and meteoroid impacts are in red. The light concentric circles correspond to artificial seismic discontinuities that were used in creating the seismic velocity model.

above observations is that numerous magma-filled fractures currently located in the deep mantle are capable of relieving the small stresses induced by Earth-raised tides.

Analyses of the Apollo seismic data have given rise to a data set of P- and S-wave first-arrival times from which it is possible to invert for a seismic velocity profile of the crust and mantle. Several investigations have attempted to do this, and a debate has arisen as to whether or not any seismic discontinuities exist in the lunar mantle. The final Apollo-era analysis indicated that a discontinuity might potentially exist about 500 km below the surface, and many researchers subsequently assumed that this depth most plausibly represented the base of the lunar magma ocean. A reanalysis of these data by one group confirmed that a discontinuity at this depth was probable (Khan and Mosegaard 2002); however, an independent analysis by a second group, using a somewhat different data set, obtained a contradictory result (Lognonné et al. 2003). A more recent inversion by the first group using a different methodology showed that they could in fact fit the seismic data using a mantle that was homogeneous in elemental composition and that did not possess any discontinuities at all (Khan et al. 2007).

Given that seismology is one of the most powerful techniques for investigating the interior structure and composition of a planetary body, it is somewhat disappointing that no consensus has arisen in regard to the lunar mantle. The range of possibilities presented by these studies is in some sense related to the data themselves. In particular, given the weak nature of the deep moonquakes, the uncertainty in the first-arrival times can sometimes exceed 10 seconds. Furthermore, there are currently two independent arrival-time data sets, and each of these has been subjected to a different set of analysis techniques. Regardless of the quality of the existing data, one needs to keep in mind that the Apollo seismic network only covered a very limited portion of the Moon's surface and that its eight years of operation represent only a fraction of the largest tidal periodicity of 18.6 years. Given that mare volcanism occurred primarily on the nearside hemisphere of the Moon, it is

quite possible that the seismic properties of the farside hemisphere will turn out to be much different from those beneath the Apollo network.

The origin of the deep moonquakes, as well as the nature by which energy is dissipated in the deep mantle, could be further addressed if the temperature profile of the Moon was known. Two techniques used to constrain this have exploited the known dependence of a mineral's seismic velocity and electrical conductivity on temperature. The electrical-conductivity profile of the Moon has been estimated by using electromagnetic-sounding techniques, whereby the relationship between simultaneous time variations in the ambient magnetic field as measured from orbit and the resulting field measured on the surface was analyzed (for a review, see Hood 1986). By calculating hypothetical mineral assemblages that would be stable at a given temperature, it was possible to invert for not only the mantle temperature profile, but also its composition. Unfortunately, the current electromagnetic and seismic data sets only place weak constraints on the interior temperature profile. Nonetheless, these analyses indicate that the lunar mantle is composed primarily of olivine and orthopyroxene, with lesser abundances of clinopyroxene and garnet (Khan et al. 2006, 2007).

THE CORE

All of the terrestrial planets and many of the icy satellites in the outer solar system possess iron-rich cores whose size is about half the radius of the parent body (the core of Mercury is probably somewhat larger). While the Moon too probably possesses some kind of core, its relative size is thought to be considerably smaller. In particular, analyses of the Moon's mass, radius, and moment of inertia (a measure that is sensitive to how density varies with depth) imply that if the Moon does possess an iron-rich core, it must be smaller than about 460 km, which is only about one-quarter of the Moon's radius (for reviews, see Hood 1986 and Wieczorek et al. 2006). Other geophysical observations are generally consistent with this picture. Namely, electromagnetic-sounding data place an upper limit of about 500 km on the core radius (Hood 1986), and the

measurement of a weak, induced dipolar magnetic field as the Moon passes through the Earth's geomagnetic tail implies the existence of a high-electrical-conductivity core with a radius of about 360 km (Hood et al. 1999). Furthermore, analyses of minute rotational signatures obtained from the lunar laser ranging experiment indicate that energy is currently being dissipated at the boundary between a molten core and solid mantle of approximately the same radius (Williams et al. 2001).

If the Moon does indeed possess an iron-rich core, then it is possible that it could have generated a dipolar magnetic field at some point in its history. Some of the lunar samples have strong magnetizations (see Fuller and Cisowski 1987), and orbital magnetic measurements from the Lunar Prospector mission show that portions of the crust are magnetized (see Fig. 2, lower right, for one model of the magnetic-field strength; Purucker 2008). Some of these magnetizations could potentially have been acquired as magmatic rocks cooled in the presence of a stable, internal dipolar field. If true, the limited magnetization age constraints from the lunar samples appear to suggest that such a dynamo might have turned on relatively late (~3.9 billion years ago) and lasted for only a short duration (see Stegman et al. 2003). A potential problem with the dynamo hypothesis, however, is that the implied field strengths are difficult to generate from a core as small as the Moon's. As an alternative hypothesis, large impact events could have generated strong transient fields that magnetized portions of the lunar crust. In support of this scenario, high crustal fields appear to be correlated with basin ejecta deposits (Halekas et al. 2001). Furthermore, a curious correlation exists between some strong crustal fields and the antipodes of some of the youngest and largest impact basins, such as Imbrium, Serenitatis, Orientale, and Crisium. One explanation for this observation is that these impact events might have generated a partially ionized plasma cloud that expanded around the Moon and amplified ambient fields (of either internal or external origin) at the point diametrically opposite the crater (Hood and Artemieva 2008). A currently unresolved issue of great importance is whether the magnetic fields as observed from orbit are a consequence of magnetized sources that are located at the surface (such as impact ejecta) or deep in the crust (such as from magmatic intrusions).

While the available evidence indicates that the Moon possesses a small molten core, the geophysical data cannot unambiguously constrain its composition. Nearly pure molten iron or molten iron-sulfide compositions are both compatible with the available constraints. Indeed, one could even explain the available data with a core composed of a dense silicate magma that was rich in iron and titanium (Wieczorek et al. 2006). Perhaps the best way to discriminate among these possibilities would be to determine the seismic velocity of the core. Unfortunately, as is evident from Figure 3, none of the well-determined ray paths pass through the deepest portion of the lunar interior.

SYNTHESIS

Figure 4 shows one interpretation of the Moon's interior structure based on geophysical measurements. Seismic, gravity, and topography data suggest that the crust is on average about 40–45 km thick and that it is considerably thinned beneath the largest impact basins. Furthermore, seismic data show that a few energetic moonquakes occur per year within the crust and upper mantle, and a probable explanation for these events is that they are caused by the slow accumulation of stresses as the Moon cools and contracts over time. The seismic data are at present some-

what ambiguous as to whether or not there are any seismic discontinuities in the mantle, though some studies are consistent with one being present about 500 km below the surface. Even if this discontinuity exists, the limited spatial distribution of the detected seismic events does not allow one to say if this is a global feature or if it is present only beneath the Apollo network, which was emplaced on the volcanically active nearside of the Moon. Deep in the Moon, about 1000 km below the surface, about 300 moonquake "nests" are repeatedly activated by the tides raised by Earth. While the locations of these nests are almost all found on the nearside hemisphere, perhaps consistent with an aseismic farside, the seismic and lunar laser ranging data indicate that the deepest portion of the mantle is highly attenuating and might perhaps be partially molten. If the deepest mantle were sufficiently attenuating, the nearside Apollo network might not be able to detect deep farside moonquakes. Several lines of evidence are consistent with the Moon possessing a dense molten core (<460 km radius), but its composition is not well constrained by the available data. Nevertheless, regardless of the core's composition, thermal-evolution models suggest that its temperature is probably low enough to have allowed some portion of it to crystallize, forming an even smaller solid inner core.
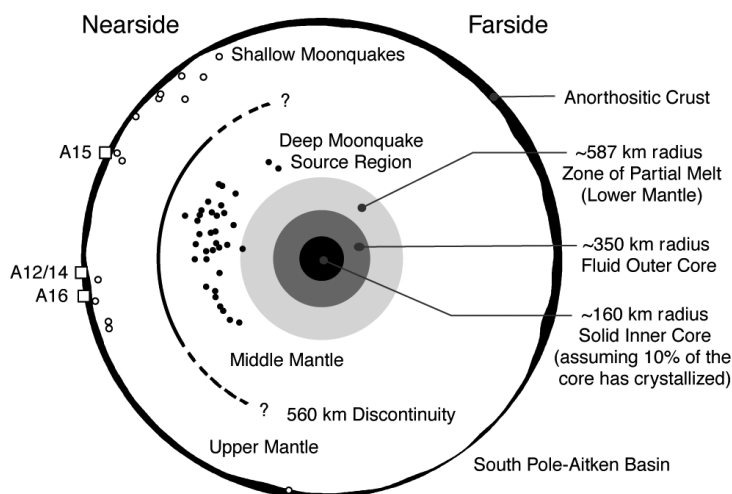
FIGURE 4 Schematic diagram of the Moon's interior structure as determined by geophysical means. Apollo landing sites are indicated by squares. From Wieczorek et al. 2006

FUTURE MISSIONS

The fact that a schematic diagram of the lunar interior can be drawn at all is a testament to the geophysical data that have been collected both from orbit and from the network of surface stations emplaced during the Apollo missions. Nevertheless, many unresolved questions about the interior of the Moon remain, particularly in regard to the composition of the mantle and core. Fortunately, geophysical measurements will be obtained from several spacecraft that are currently in orbit at the time of this writing or will be placed into orbit over the next several years. As examples, the Japanese Kaguya, Chinese Chang'e-1, and Indian Chandrayaan-1 missions are currently making precise topographic maps of the Moon, and this map will be further improved with the launch of the American Lunar Reconnaissance Orbiter in 2009. (Prior to these missions, the topography of both Venus and Mars was known more precisely than that of the Moon.) The Kaguya mission contains an orbital radar sounder that will attempt to determine the thickness of the mare basalts, map the thickness of the regolith, and detect buried regolith horizons. Kaguya

is also making some of the first direct farside gravitational measurements, and with the launch of the American mission GRAIL (Gravity Recovery And Interior Laboratory) in September 2011, we will finally obtain a global, high-resolution gravitational map of the Moon. In addition to allowing for more detailed studies of the crust and upper mantle, measurements of time-variable gravitational signatures from the GRAIL mission will help elucidate the size and properties of the lunar core and perhaps detect a solid inner core.

These future data sets will almost certainly lead to unexpected discoveries and change the way we view the interior and geologic evolution of the Moon. Nevertheless, a surface geophysical network consisting of seismometers, heat-flow probes, and magnetometers will ultimately be required to address the long list of questions that cannot be addressed from orbit. Such a network would be challenging to implement, as it would require a handful of stations on both the near- and farside hemispheres operating concurrently over a decade-long period. A simultaneously orbiting relay satellite would be required for the operation of the farside stations, and such an orbiter might also perform measurements of the ambient magnetic environment, as well as search for impact-produced flashes on the lunar surface that could be used as constraints in the seismic analyses. Several national space agencies are currently investigating the possibility of sending either landers or instrumented penetrators to the Moon. While these mission designs are still in preliminary phases, most call for payloads consisting of seismometers and heat-flow probes. NASA is currently promoting the concept of an International Lunar Network, where geophysical stations would be incrementally installed over time by various national agencies. Given the renewed interest in lunar exploration in the US and the optimistic lunar programs of the Chinese, Indian, Japanese, and European space agencies, such a strategy could turn out to be the best means for achieving this goal.

ACKNOWLEDGMENTS

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