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Scattering attenuation profile of the Moon: Implications for shallow moonquakes and the structure of the megaregolith

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ABSTRACT

We report measurements of the attenuation of short period seismic waves in the Moon based on the quantitative analysis of envelope records of lunar quakes. Our dataset consists of waveforms corresponding to 62 events, including artificial and natural impacts, shallow moonquakes and deep moonquakes, recorded by the four seismometers deployed during Apollo missions 12, 14, 15 and 16. To quantify attenuation and distinguish between elastic (scattering) and inelastic (absorption) mechanisms we measure the time of arrival of the maximum of energy t_{max} and the coda quality factor Q_c. The former is controlled by both scattering and absorption, while the latter is an excellent proxy for absorption. Consistent with the strong broadening of seismogram envelopes in the Moon, we employ diffusion theory in spherical geometry to model the propagation of seismic energy in depth-dependent scattering and absorbing media. To minimize the misfit between predicted and observed t_{max} for deep moonquakes and impacts, we employ a genetic algorithm and explore a large number of depth-dependent attenuation models quantified by the scattering quality factor Q_{sc} or equivalently the wave diffusivity D, and the absorption quality factor Q_i. The scattering and absorption profiles that best fit the data display very strong scattering attenuation ($Q_{sc} \leq 10$) or equivalently very low wave diffusivity ($D \approx 2$ km^2/s) in the first 10 km of the Moon. These values correspond to the most heterogeneous regions on Earth, namely volcanic areas. Below this surficial layer, the diffusivity rises very slowly up to a depth of approximately 80 km where Q_{sc} and D exhibit an abrupt increase of about one order of magnitude. Below 100 km depth, Q_{sc} increases rapidly up to approximately 2000 at a depth of about 150 km, a value similar to the one found in the Earth's mantle. By contrast, the absorption quality factor on the Moon $Q_i \approx 2400$ is about one order or magnitude larger than on Earth. Our results suggest the existence of an approximately 100 km thick megaregolith, which is much larger than what was previously thought. The rapid decrease of scattering attenuation below this depth is compatible with crack healing through viscoelastic mechanisms. Using our best attenuation model, we invert for the depth of shallow moonquakes based on the observed variation of t_{max} with epicentral distance. On average, they are found to originate from a depth of about 50 km ± 20 km, which suggests that these earthquakes are caused by the failure of deep faults in the brittle part of the Moon.

1. Introduction

The Apollo Passive Seismic Experiment was conducted from 1969 to 1977 and allowed the recording of lunar seismic signals thanks to four seismometers deployed during Apollo missions 12, 14, 15 and 16. Lunar seismic signals are notably different from usual terrestrial seismic records: the energy rise at the onset of the signal is gradual, the S-wave arrival is difficult (or impossible) to detect, the maximum of energy is broad, and the energy decay in the coda is very slow. The signal can remain noticeably above the noise level for up to two hours, compared to a few minutes on Earth for an event of the same magnitude. Soon after the beginning of the experiment, several studies have established a link between the characteristics of lunar signals and strong scattering in the heterogeneous upper lunar crust (Latham et al., 1972b; Nakamura, 1977).

The surface layer is known as the regolith. The regolith is thought to be a few meters thick and consists of materials ranging from fine dust to boulders of a few meters. P-wave velocities in the regolith are about 100 m/s (Nakamura et al., 1975). This layer formed as a result of intensive meteoritic bombardment that caused cratering, cracking, distribution of ejecta and partial melting over the entire surface of the Moon. This layer extends at greater depths into a megaregolith, as cracks close gradually with

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confining pressure until the crust becomes competent. The thickness of the megaregolith has been revised several times since the beginning of the analysis of Apollo data. Latham et al. (1972a) suggested that the characteristics of lunar seismic signals could be explained by the presence of a scattering layer a few kilometers thick at the surface of the Moon, with a maximum thickness of 20 km due to the confining pressure. Simmons et al. (1973) increased the lower boundary for the depth extent of the megaregolith. These authors argue that microcracks are present in lunar rocks down to 25 km but not below, based on the velocity profile of Toksöz et al. (1972), which shows a gradual velocity increase up to a first discontinuous change in seismic properties at 25 km depth, and laboratory measurements of *P*- and *S*-wave velocities and isothermal compressibility by Todd et al. (1973). The authors add that confining pressure alone cannot explain a complete closure of cracks at only 25 km depth. The velocity model of Lognonné et al. (2003) suggests that impact-induced cracks might still be present below 30 km. In a review by Wieczorek et al. (2006), the 20-km seismic discontinuity described by Simmons et al. (1973) is interpreted as either a fracture discontinuity caused by annealing at depth thanks to thermal activation, or a compositional discontinuity. The latter hypothesis is supported by various geological studies, revealing a more mafic composition for the lower crust than for the upper crust. A combination of fracture and compositional discontinuity is also possible. A review by Jaumann et al. (2012) reports the following conservative estimates: ejecta blankets of 2–3 km thickness, a structural disturbance of at least 10 km thickness and in situ crust fracturing down to about 25 km. This brief summary indicates that the case of the megaregolith is far from clear and could benefit from independent geophysical observations.

Recently, the GRAIL mission collected gravity data that were used by Wieczorek et al. (2013) to infer a 34–43 km deep crust with an average porosity of 12%, which supports the idea of a highly-fractured environment up to great depths. This inference and the spindle-like shape of short period lunar seismic signals support the idea that seismic waves propagate in the multiple scattering regime, and that energy transport can be modeled using diffusion theory. In seismology, Wesley (1965) introduced for the first time a diffusion equation to model the propagation of seismic energy emitted by underground nuclear explosions and earthquakes in the near range. Dainty et al. (1974) made the first attempt at modeling the scattering processes in the Moon using diffusion theory, and inferred a thickness of 25 km for the scattering layer. However, these modeling efforts were limited to the planar case. Considering the size of the Moon and the duration of lunar seismic signals, it is essential to take sphericity into account. As will be illustrated in this study, stratification of scattering properties and spherical geometry are key to the modeling of lunar seismic signals.

The paper is organized as follows. We first introduce the dataset and discuss the observables which will help us quantify the heterogeneity of the interior of the Moon. We then introduce our modeling approach, based on diffusion theory, and the inversion scheme. After presenting the profiles of diffusivity, scattering quality factor and absorption quality factor for the Moon, we explain how our results can be used to improve the depth location of shallow moonquakes and discuss the implications for the structure and depth extent of the megaregolith.

2. Data processing and error analysis

2.1. The seismic network

The lunar seismometers consisted of one 3-component long-period sensor unit, and one short-period sensor unit sensitive to vertical motion. The long-period sensor was able to operate in two modes: a flat mode, with an instrument frequency response ranging from 0.004 Hz to 2 Hz, and a peaked mode, around 0.45 Hz. The short-period sensor has a flat frequency response in the 0.05–20 Hz range. As illustrated in Fig. 1, the stations are located at the vertices of an approximately equilateral triangle, with stations 12, 15 and 16 about 1100 km apart. Stations 12 and 14 are located at the same vertex, about 180 km apart from each other. The four stations operated simultaneously from April 1972 to September 1977. A total of 12,000 events were recorded (Nakamura et al., 1981).

2.2. Event types

The events used in this study can be sorted into three categories: impacts, shallow moonquakes and deep moonquakes. Impacts are mostly due to meteoroids, whose mass varies from 100 g to 100 kg (Oberst and Nakamura, 1991). There are also a few artificial impacts, caused by the crash of rocket boosters and parts of lunar modules from the successive missions at take-off. These crashes were programmed to provide seismic events of known spatial and temporal origin. Shallow moonquakes are thought to originate from depths between 30 and 200 km, with an important uncertainty on the location of their sources. Their mechanism is also poorly understood. Various explanations have been proposed, from the deformation of a rigid crust over a less rigid asthenosphere by tidal forces (Lammlein, 1977), to nuggets of strange quark matter (Frohlich and Nakamura, 2006; Banerdt et al., 2006). They are the most energetic events recorded on the Moon, with magnitudes as large as 5 (Goins et al., 1981). Deep moonquakes originate from a population of 316 nests around 1000 km deep (Nakamura, 2005). These nests are regularly reactivated, and produce very similar signals with observed periodicities of 14 days, 27 days, 206 days and 6 years (Lammlein, 1977; Nakamura, 2005). This periodic activity is correlated with tides, extreme librations, and Earth-Moon separation periods (Lammlein, 1977). The variety of source depth among the different events will be of particular interest to our study.

2.3. Dataset

The available dataset is composed of the following records: 39 artificial impacts, 269 natural impacts, 384 shallow moonquakes and 152 deep moonquakes from both long- and short-period sensors, across all available components. A high number of those signals were discarded on initial visual inspection because they showed either a high number of digital spikes, or a low signal-to-noise ratio. In the case of broadband records, analysis of the power spectral density of the signals showed that most of the energy is found at frequencies between 0.2 and 1.5 Hz, with a sharp peak at 0.45 Hz consistent with the instrumental response. For short period records, most of the energy is found at frequencies below 5 Hz with a sharp peak around 2 Hz for impacts. Shallow moonquakes release energy between 0.05 and 15 Hz with a peak around 7 Hz which is due to the instrument response. The latter are the only events to exhibit significant energy beyond 3 Hz.

Prior to the measurements, preliminary processing of the signals is necessary and summarized hereafter. For impacts and shallow moonquakes, signals are first band-pass filtered between 0.25 Hz and the Nyquist frequency (3.15 Hz for low frequency signals and 22.87 Hz for high frequency signals). To eliminate the remaining spikes in the signal, we use a robust median despiking algorithm. In a 700-period overlapping moving window, any amplitude larger than 5 times the median of the window is recalculated using a simple interpolation procedure (see Bulow et al., 2005 for details). Deep moonquake signals were kindly provided