Reports

New Seismic Data on the State of the Deep Lunar Interior

Abstract. *Direct shear-wave arrivals from seismic events originating on the far side of the moon are not observed at some of the stations of the Apollo seismic network. These data suggest that the material in the lunar interior at a depth of 1000 to 1100 kilometers is more dissipative for seismic shear waves than the lithosphere above, and possibly exists in a partially molten state akin to the earth's asthenosphere.*

The Descartes geophysical station installed during Apollo mission 16 in April 1972 completed a lunar seismic network of four stations (*1*). The network spans the near face of the moon in an approximate equilateral triangle with a spacing of 1100 km between stations (*2*), and greatly increases our ability to determine the origin time, epicenter, and focal depth for natural seismic events.

Previous seismic data, mainly from man-made impacts, have revealed a major discontinuity at a depth of about 60 km in the eastern part of Oceanus Procellarum (*1, 3*). By analogy with the earth, we refer to the zone above the discontinuity as the crust and the zone below as the mantle. Whether the crust is a regional or moon-wide feature cannot be determined from the data of the present seismic network. The velocity of compressional waves reaches a value between 6.3 and 7.0 km/sec in the lower half of the crust—a velocity range appropriate for the aluminous basalts and gabbroic anorthosites that predominate in samples from the lunar highlands. The compressional-wave velocity increases to about 8 km/sec at the top of the mantle, and the shear-wave velocity there is about 4.6 km/sec.

Seismic data relevant to the structure and state of the deep lunar interior derive from deep moonquakes and distant meteoroid impacts. The presence or absence of shear waves and the travel times and relative amplitudes of both compressional (P) and shear (S) waves from these events are the most direct sources of information.

High-frequency S waves (0.5 hertz and higher) have been detected from all the moonquake focuses that have been located on the near side of the moon, at depths approaching 1000 km (*1*). This shows that wide-spread melting cannot occur in the outer 1000 km of the nearside. The locations of moonquake focuses are based on a seismic velocity model in which velocities in the upper mantle, determined primarily from artificial impacts, are extended to great depth. For two moonquake focuses that yield data sufficient to test this extrapolation, the observed travel times agree within a few seconds with those predicted from the model. Thus, we assume for the present discussion that the P-wave velocity remains within a few percent of 8 km/sec to a depth of nearly 1000 km. A slight increase in Poisson's ratio, or a slight decrease in both P-wave and S-wave velocities, is suggested by the available data.

In contrast, new data show that S waves cannot be identified in some of the signals from quakes originating on the far side of the moon. Twelve far-side moonquakes observed to date have a common source location, designated the A_{33} focus, tentatively placed at 3°N, 119°E, and a depth of 830 km. As shown in Fig. 1, S waves from these moonquakes are detected at stations 15 and 16, but not at station 14. Shear waves are normally observed as impulsive arrivals with amplitudes several times that of the P wave at the same station (*4*). These data indicate that S waves do not propagate, or are highly attenuated, below the depth of moonquakes.

Data from a newly identified large meteoroid impact that occurred on the far side of the moon on 17 July 1972 permit a more quantitative estimate of the depth and properties of the anomalous zone. The times of the P-wave arrivals listed in Table 1 are believed to be uncertain by several seconds because of the emergent character of the signals. Using the P-wave arrival times at stations 12, 15, and 16, and a seismic velocity model derived by extrapolating the velocity in the upper mantle to the center of the moon, we have tentatively determined the time and location of the impact to be 21h:50m:50s and 30°N, 147°E near the crater Moscoviense (*5*).

The characteristics of lunar impact seismograms differ markedly from those of typical earth recordings (*1, 6*). The most striking features of lunar seismic signals are their emergent beginnings, long duration, and slow decay. Following the first one or two cycles of initial P-wave motion, the ground motion becomes so complex that there is little or no correlation between any two components. The onsets of direct S waves are indistinct where they can be identi-

Table 1. Observed and derived parameters of category A_{33} moonquakes and the meteoroid impact of 17 July 1972. The moonquake arrival times are relative to the P-wave arrival time at station 15.

Parameters	Station			
	12	14	15	16
Category A_{33} moonquakes				
Relative arrival time (sec)				
P-wave		30	0	
S-wave			208	188
Epicentral distance (deg)	142	137	111	104
Meteoroid impact, 17 July 1972				
Arrival time (hr:min:sec)				
P-wave	21:57:56	21:58:01	21:57:00	21:57:29
Beginning of second wave train		22:06:30	22:02:00	22:04:36
S-wave (expected)	22:03:06	22:03:02	22:01:29	22:02:19
SS-phase (expected)	22:06:39	22:06:28	22:03:18	22:04:44
Epicentral distance (deg)	151	150	114	130
Depth of penetration (km)				
Direct wave	1315	1290	800	1010
SS-phase	390	380	235	300

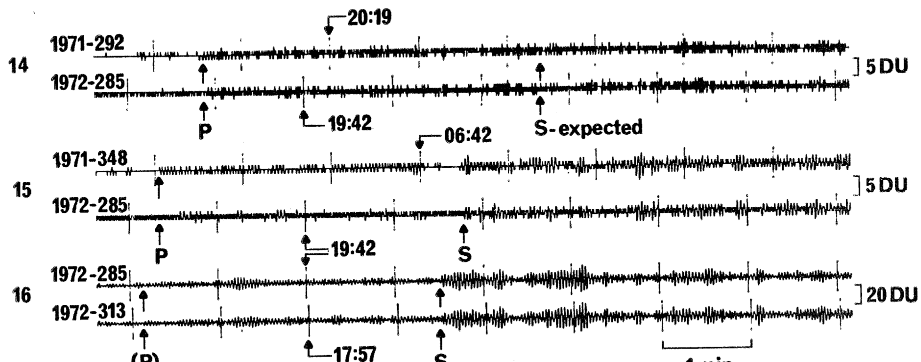

Fig. 1. Initial seismogram sections of a category A_{33} moonquake detected at stations 14, 15, and 16, and of another moonquake for each station. The signal was too weak to be detected at station 12. Only the seismogram for the long-period horizontal (LPY) component is shown for each event at each station. These events are believed to have originated within the far half of the moon. Note that the S wave is not observable at station 14 when expected (marked *S-expected*). Digital units, *DU*, represent amplitude; times are given as hour : minute.

fied at all. These and other unusual characteristics of the lunar seismic signals have been interpreted as resulting from intensive, but nearly loss-free, scattering of seismic waves in a heterogeneous surface layer presumably produced by meteoroid bombardment of the moon. Shear waves from impacts are most readily identified in the filtered playouts of lunar signals in a narrow band of frequencies between 2 and 4 hertz. At these frequencies, the rise and decay times of the seismic wave trains are sufficiently short that the S-wave arrival appears as a characteristic bulge in the seismograms displayed on a compressed time scale. Signals in this frequency range are best recorded by the short-period, vertical-component seismometer (SPZ), which is one of four seismometers placed at each station. This seismometer is operational at all stations except station 12.

Filtered SPZ seismograms for the impact of 17 July (Fig. 2) clearly show the S-wave arrival at station 15. The arrow marked *S* indicates its expected beginning, which is estimated to occur at about 1.7 minutes before the peak of the wave train at 4 hertz. The corresponding signal, however, cannot be identified in the records from stations 14 and 16. For these stations, small increases in signal intensity late in the wave train are tentatively interpreted as shear waves reflected from the surface (SS), or perhaps from the bottom of the crust-mantle interface, at a point approximately midway between the source and the station (*7*). The arrival times of these late phases are also listed in Table 1 together with the theoretical times of arrival for S and SS phases. In the model used in this calculation it is assumed that the P- and S-wave velocities observed for the upper mantle extend throughout the lunar interior. A comparison of the observed and expected arrival times reveals that (i) direct S-wave arrivals are not observed at stations 14 and 16 near the expected times of arrival; (ii) the arrival time of the direct S wave at station 15 is approximately 30 seconds later than expected; and (iii) late arrivals at stations 14 and 16 are at about the right time for the SS phase. The first observation is a strong indication that high-frequency S waves are not effectively transmitted in the deep interior of the moon. The second and third observations suggest that the average S-wave velocity in the lower mantle is less than that in the upper mantle. A decrease of the S-wave velocity averaged over the upper 800 km of the moon by 5 percent from the model velocity is sufficient to account for the 30-second discrepancy between the expected and observed S-wave arrival times at station 15.

Although the available data are not sufficient to derive a detailed seismic velocity model for the deep interior,

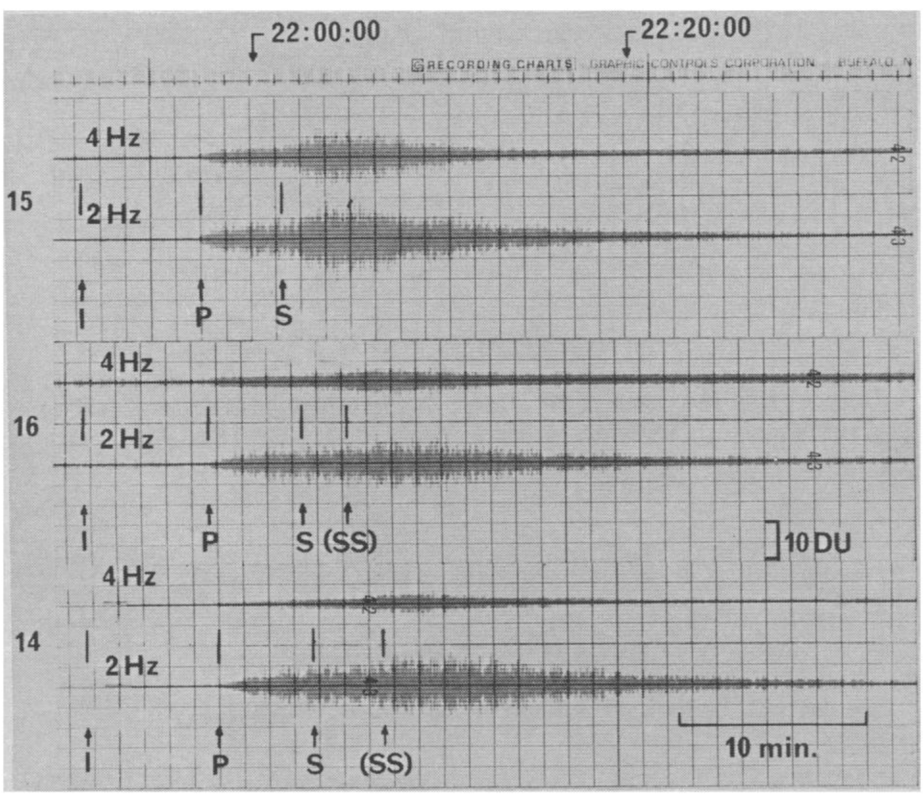

Fig. 2. Filtered short-period seismograms of the seismic event detected on 17 July 1972. The signal is believed to be from a meteoroid impact on the far side of the moon. Traces for two different filter settings are shown for each station. The frequencies, 4 and 2 hertz (*Hz*), are the center frequencies of the narrow band-pass filters used in the data playback. *I*, the estimated time of impact; *P*, observed arrival time of the direct P wave; *S*, arrival time of the direct S wave expected from extrapolation of upper mantle velocities; *SS*, arrival time of S wave reflected from the surface. The characteristic S-wave bulge, visible at station 15, is missing at stations 16 and 14 when expected. Times are given at the top as hour : minutes : second; digital units, *DU*, represent amplitude.

these observations can be explained by introducing a "core" with a radius between 600 and 800 km, in which S waves are highly attenuated [Q of about 100 or less (8)]. The P-wave velocity within this zone may be slightly lower than that in the mantle. The maximum allowable velocity decrease for P waves is about 0.3 km/sec. These limiting values are found to be relatively insensitive to assumed velocities in the mantle.

Seismic wave attenuation is strongly dependent on temperature, showing a rapid increase with increasing temperatures, and a sharp increase with the onset of melting (9). Thus, a high temperature approaching the solidus temperature in the lunar interior may account for the lack of S-wave transmission indicated by the seismic data. A silicate interior would have a solidus temperature between 1450° and 1650°C at a depth of 1000 to 1100 km. This model is in substantial agreement with several thermal models recently proposed by Toksöz *et al.* (10). Partial melting of silicate material is considered to be a possible explanation of the zone of low Q values and low seismic velocities in the upper mantle of the earth (9, 11). Thus, partial melting of a few percent of the material may explain the present data. A completely molten core of the size indicated is not likely, however, because it would require a decrease of the P-wave velocity by more than the value of 0.3 km/sec obtained in our preliminary analysis. Other possibilities, such as increased volatiles in the deep interior of the moon, cannot be ruled out at present.

The possibility of a very dense, molten, metallic core similar to that of the earth is ruled out by considerations of both moments of inertia and seismic wave velocities. The radius of about 700 km is too large for such a core.

Several observations suggest that there is a gradual transition from the solid material of the lower mantle to the material of different physical properties below: (i) the spectrum of S waves from moonquakes appears to shift toward lower frequencies with increasing depth for sources of approximately equal strength; (ii) the direct S waves recorded at station 15 from the large impact are abnormally small compared with the P waves; and (iii) no signals corresponding to reflections from a well-defined boundary at a depth of 1000 km have yet been observed.

The suggestion of a partially molten zone at great depth has an important bearing on the question of the focal mechanisms of moonquakes. The recognized categories of moonquake signals now number 41, corresponding to 41 different active focuses. Moonquake signals from numerous other active zones are too weak to be analyzed in detail. Moonquakes occur with a monthly periodicity, strongly suggesting that they are induced by lunar tides (4). The moonquake focuses for which depths have been determined (18 cases) are concentrated in the depth range from 800 to 1000 km. The localization of moonquake activity in a zone 200 km thick, falling immediately above the zone of high attenuation, may be explained as resulting from one or more factors: (i) reduced rigidity in the zone of high temperature or partial melting would tend to concentrate the dissipation of tidal energy at great depth; (ii) weak convective motions beneath a thick, rigid mantle might generate deep moonquakes without the surface manifestations associated with terrestrial plate tectonics; and (iii) fluids injected into the moonquake zone from below, under the influence of tides, would reduce the effective friction along zones of weakness, or weaken the silicate bond, leading to dislocation.

Making an analogy with the earth, we may consider the lunar lithosphere—the relatively rigid outer shell of the moon—to be about 1000 km thick. The central core of the moon would be equivalent to the asthenosphere (the low-velocity zone) of the earth.

Y. NAKAMURA, D. LAMMLEIN
G. LATHAM, M. EWING, J. DORMAN
Earth & Planetary Sciences Division, University of Texas Medical Branch, Galveston 77550
F. PRESS, N. TOKSÖZ
Department of Earth and Planetary Sciences, Massachusetts Institute of Technology, Cambridge 02139

References and Notes

1. G. Latham *et al.*, *Moon* **5**, 272 (1973).
2. The Apollo 12 and Apollo 14 stations together occupy a corner of the triangle.
3. M. N. Toksöz *et al.*, *Moon* **4**, 490 (1972).
4. G. Latham *et al.*, *Science* **174**, 687 (1971).
5. Station 14 data were not used because the long-period vertical component (LPZ) seismometer of this station was inoperative, adding uncertainty to the initial P-wave measurement.
6. G. V. Latham *et al.*, *Science* **167**, 455 (1970); *ibid.* **170**, 620 (1970).
7. This interpretation is based only on the arrival time of the wave train. No well-defined surface-reflected phases have previously been identified, probably because they are completely obscured by the presence of the direct-phase wave train. Thus, surface-reflected compressional waves are not identifiable on the same record.
8. The quality factor Q is used to specify the attenuation of energy in a vibrating system; $2\pi/Q$ is the fractional loss of energy per cycle of vibration of the system.
9. D. D. Jackson and D. L. Anderson, *Rev. Geophys.* **8**, 1 (1970).
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12. Supported by NASA contract NAS 9-13143. Earth & Planetary Sciences Division, Marine Biomedical Institute, Contribution No. 17.

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Methane in Lake Kivu: New Data Bearing on Its Origin

Abstract. *Lake Kivu, an African rift lake, contains about 50 cubic kilometers of methane (at standard temperature and pressure) in its deep water. Data resulting from two recent expeditions to the lake and a reevaluation of earlier data suggest that most of the methane was formed by bacteria from abiogenetic carbon dioxide and hydrogen, rather than being of volcanic origin or having formed from decomposing organic matter.*

Lake Kivu is the westernmost and most elevated of the string of rift lakes which curves through east-central Africa. It has a surface area of about 2400 km^2 and a maximum depth just under 500 m. The lake's peculiar hydrochemistry was first recognized by Damas (1), who discovered large quantities of dissolved gas in the deep water. The gas was later identified as consisting principally of carbon dioxide [up to 1.4 liters at standard temperature and pressure (STP) per liter of water] and methane [up to 0.37 liter (STP) per liter of water] (2). The lake is anoxic below 50 m and both temperature and salinity increase stepwise with depth, which suggests the existence of convecting layers (3). The total temperature increase below 50 m is about 3.2°C, and the content of Na, Mg, K, and Ca increases from 311 to 1355 mg/liter. The consequent density stratification of the water column severely restricts vertical mixing and the deep water acts as a trap for solutes. The large accumula-