Reports

Seismic Data from Man-Made Impacts on the Moon

Abstract. *Unusually long reverberations were recorded from two lunar impacts by a seismic station installed on the lunar surface by the Apollo 12 astronauts. Seismic data from these impacts suggest that the lunar mare in the region of the Apollo 12 landing site consists of material with very low seismic velocities near the surface, with velocity increasing with depth to 5 to 6 kilometers per second (for compressional waves) at a depth of 20 kilometers. Absorption of seismic waves in this structure is extremely low relative to typical continental crustal materials on earth. It is unlikely that a major boundary similar to the crust-mantle interface on earth exists in the outer 20 kilometers of the moon. A combination of dispersion and scattering of surface waves probably explains the lunar seismic reverberation. Scattering of these waves implies the presence of heterogeneity within the outer zone of the mare on a scale of from several hundred meters (or less) to several kilometers. Seismic signals from 160 events of natural origin have been recorded during the first 7 months of operation of the Apollo 12 seismic station. At least 26 of the natural events are small moonquakes. Many of the natural events are thought to be meteoroid impacts.*

Impacts on the lunar surface of the Apollo 12 Lunar Module (LM) ascent stage and the third stage of the Apollo 13 Saturn booster (S-IVB) generated seismic signals that were recorded by the seismometers installed on the moon by the Apollo 12 astronauts on 19 November 1969. The seismometers are part of the emplaced science station called ALSEP (Apollo Lunar Surface Experiments Package). [For details of the lunar seismic experiment, see (*1*).] Approximately 160 events of natural origin have been recorded by the seismometers during the first 7 months of operation. However, few criteria have emerged for estimating the time or location of these natural events from the study of their seismograms. Hence, the man-made impacts, whose time and location are well known from the National Aeronautics and Space Administration's tracking information, have emerged as extremely important tools for the seismic exploration of the moon. It is expected that NASA will include impacts of both the LM and the spent S-IVB stage of the Apollo booster on all future Apollo missions.

At least 26 of the natural lunar seismic events are believed to be shallow moonquakes. All of these events occur within 3 days of the time when the moon comes closest to the earth (perigee) during its monthly orbital cycle. At least one quake has been detected at each perigee crossing. Tidal strains, produced within the moon by the gravitational interaction between the earth and the moon, reach maximum values at perigee. Thus, tidal strains, which are much larger in the moon than in the earth, appear to be an important factor in the release of seismic energy in the outer shell of the moon.

Table 1. Expended LM ascent stage and S-IVB impact parameters. The Apollo 12 seismic station is located at 3°03′S, 23°25′W.

Parameter	LM	S-IVB
Time of impact (G.M.T.)	$22^h17^m16.4^s$	$01^h09^m39.0^s \pm 0.2^s$
Mass (kg)	2,383	13,925
Impact velocity (km/sec)	1.68	2.58
Kinetic energy of impact (ergs)	$3.36(10)^{16}$	$4.63(10)^{17}$
Equivalent energy of impact (lb of TNT)	$1.77(10)^3$	$2.44(10)^4$
Angle of impact from horizontal	3.7°	76.4°
Distance between point of impact and seismic station (km)	73	135
Azimuth from seismic station	112°	274°
Latitude	3°57′S	2°45′S
Longitude	21°12′W	27°52′W

All the moonquakes are small. The magnitude of the largest events is between 1 and 2 on the Richter scale. The magnitude of the lunar signals is obtained through Richter's empirical relationship between energy and magnitude. An earthquake of this magnitude would normally be barely perceptible even by persons in the immediate vicinity of the epicenter. The rate of lunar seismic energy release implied by these data is very low compared with the rate of seismic energy release in the earth. If the moon were as active as the earth and if the seismic sources were uniformly distributed throughout the outer shell of the moon, between 10 and 100 quakes with energy release equal to or greater than that of the S-IVB impact would have been recorded during the 7 months of operation considered in this report. No quakes of this magnitude have been recorded thus far. However, neither the recording period nor the area covered by lunar seismic stations is sufficient to permit generalizations on this point. Zones of high seismicity may be localized in regions far from the Apollo 12 station. Despite these qualifications, it is almost certain that the concept of plate tectonics as it applies to the earth, with large-scale deformation of the crust of the earth to form folded mountains and great trenches, does not apply to the moon at this point in its evolution. Nor do we see visual evidence of such tectonic activity in the past from study of the surface features of the moon. The outer shell of the moon appears to be very old and quite stable except for the disruptive influences of tidal stresses. Some of the natural events are believed to be produced by meteoroid impacts, which can usually be distinguished from the moonquake signals by their higher frequency content and their relatively peaked spectra. When Hawkins' (*2*) flux estimate is used, the rate of occurrence of the impact events is in approximate agreement with the number of meteoroids in the kilogram mass range that are expected to collide with the lunar surface within a radius of a few hundred kilometers from the seismic station. The study of these signals will eventually provide a quantitative estimate of the numbers and masses of meteoroids in space in the kilogram mass range and more detailed information on the internal structure of the moon.

The locations of the Apollo 12 seismic station and the impact points are shown in Fig. 1. Pertinent impact and seismometer parameters are listed in Table 1. The LM struck the lunar surface 73 km from ALSEP at a velocity of 1.68 km/sec and at an angle to the lunar surface of only 3.7°; it was moving toward ALSEP at the time of impact. The S-IVB struck the surface at nearly normal incidence, 135 km from ALSEP, with a velocity of 2.58 km/sec, headed toward the northeast.

The region of the impacts is located in the southeastern edge of Oceanus Procellarum. The relatively smooth area, particularly between the LM impact point and ALSEP, is believed to be covered by igneous rock, which flooded the region early in the moon's history. Several separate episodes of flooding may have occurred. The rough terrain between the S-IVB impact point and ALSEP presumably consists of older lunar material, which extends up through the lava fill. The nature of the older material is unknown. It may be primitive lunar material that accreted during the final stage of lunar evolution; it may be material thrown out from large craters; or it may be igneous rock formed during an early stage of melting. On the basis of the apparent flooding of older craters, we estimate the overlying mare material to be approximately 1 to 2 km thick between the LM impact point and the seismic station and to be variable in thickness from 0 to 2 km between the S-IVB impact point and the station; however, these estimates are subject to considerable uncertainty (*3*).

The impact signals are shown on a compressed time scale in Fig. 2, along with the signals from the two largest events of natural origin recorded thus far. The signal from a missile impact recorded at White Sands, New Mexico, by Latham *et al.* (*4*) is also shown for comparison. All the lunar signals are clearly similar in character but, as a class, are quite different from the White Sands missile impact signals. The following remarks can be made concerning the general characteristics of the lunar signals.

1) The lunar signals have extremely long duration. The LM impact signal was detectable for about 1 hour, and the S-IVB impact signal was detectable for more than 4 hours. A missile impact signal on earth would last only a few minutes at an equivalent distance. The lunar signals build up gradually to a maximum and then decay very gradually. The LM impact signal reaches its maximum intensity approximately 7 minutes after its beginning, and the S-IVB impact signal reaches its maximum value after approximately 12 minutes. The smoothed envelopes of the signals reach maximum peak-to-peak ground displacement amplitudes of 9.5 and 75 nm for the LM and S-IVB, respectively.

2) The earth missile impact signal contains distinct phases corresponding to the arrival of various types of seismic waves. Such phases are less distinct in the lunar signals.

3) Spectra of the impact signals (shown in Fig. 3) are relatively broad for samples taken near the beginnings of the wave trains. For the S-IVB signals, the spectra of samples taken at later times in the wave trains show a gradual decrease of energy in the high-frequency end of the spectrum, owing, presumably, to the greater absorption of high-frequency seismic energy. The gradual loss of high-frequency energy tends to narrow the spectra, thus emphasizing a single dominant spectral

Table 2. Travel times and velocities for seismic waves recorded from LM and S-IVB impacts.

Impacts	Wave type	Travel time (sec)	Velocity (km/sec)
LM	*P*	23.5 $^{+2.1}_{-3.7}$	3.1 (2.9–3.7)
	S	40.4	1.8
S-IVB	*P*	29.1	4.64
	PP	34.8	3.88
	S	54.0	2.50

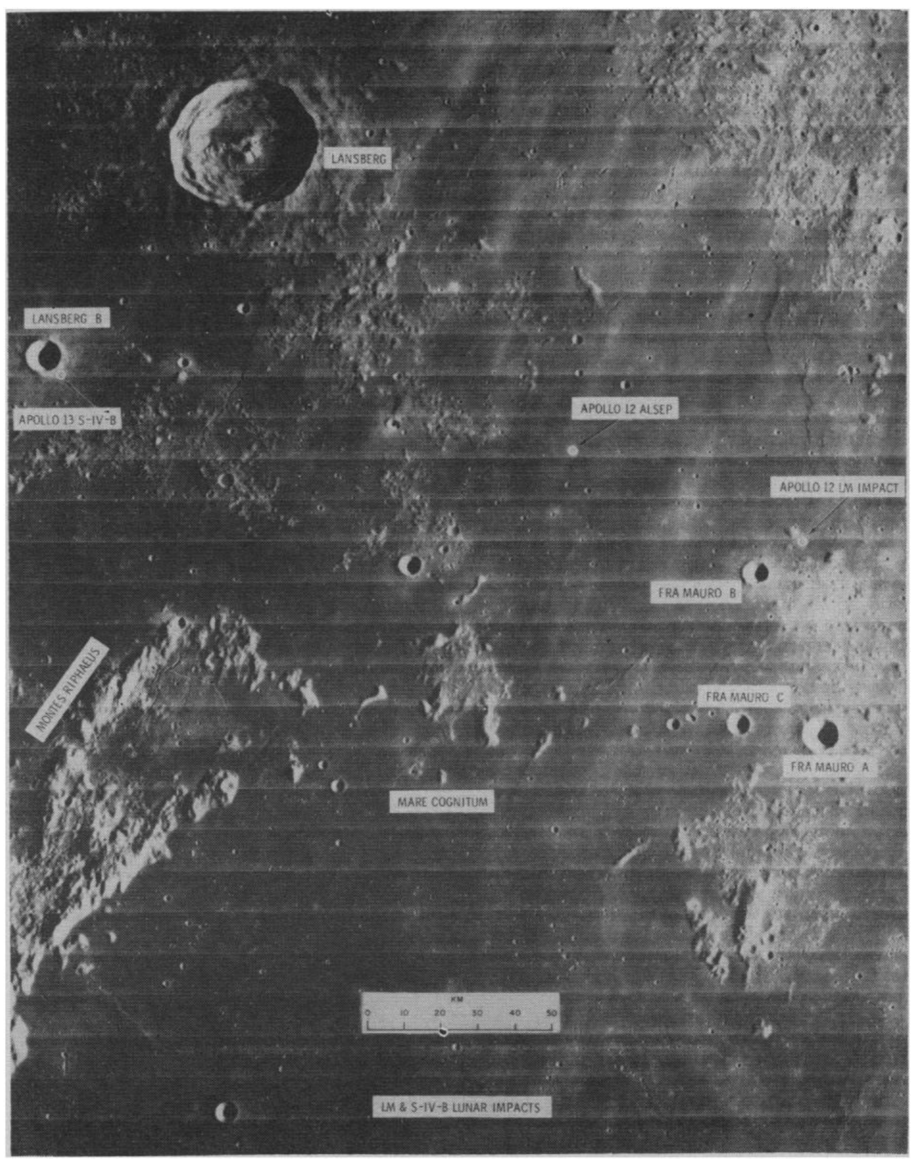


Fig. 1. Locations of the LM and S-IVB impacts and the Apollo 12 seismic station. The region shown is in the southeastern edge of Oceanus Procellarum. The coordinates of these sites are listed in Table 1.

Table 3. Summary of seismic energies and conversion efficiencies for impact signals.

Impacts	Calculated seismic energy coupled into the lunar structure (ergs)		Conversion efficiency (ratio of seismic energy to impact kinetic energy)	
	Scattering hypothesis	Dispersion hypothesis	Scattering hypothesis	Dispersion hypothesis
LM	$1.5 \cdot 10^{10}$	$2.8 \cdot 10^{10}$	$4.5 \cdot 10^{-7}$	$8.3 \cdot 10^{-7}$
S-IVB	$8.5 \cdot 10^{11}$	$4.7 \cdot 10^{12}$	$1.8 \cdot 10^{-6}$	$1.0 \cdot 10^{-5}$

peak for the later portions of the signals. The spectra of LM impact signals remain relatively broad. Although the peak frequency is relatively stable with time for each signal, it does vary from signal to signal. The spectral maxima are at 1.1 and 0.40 hz for the LM and S-IVB signals, respectively. Low-frequency energy (below 0.25 hz) is notably absent in the signals.

4) The impact signals are complex with little or no phase correlation between any two components of motion, except for the very early parts of the wave trains at times of body wave arrivals.

The beginnings of the impact signals are shown on an expanded time scale in Fig. 4. Signals corresponding to the arrival of compressional waves (P) and shear waves (S), waves that travel through the body of the moon, can be identified on the seismic records. The identification of these arrivals is based primarily on their associated particle motions and on their relative times of arrival. The first arrival from the LM impact is very small in amplitude, and it is difficult to specify the exact arrival time. This arrival may correspond to a wave that has reflected once from the lunar surface (PP), as will be discussed below. The prominent signal that nearly obscures the P wave onset from the LM impact was produced by a sudden tilt of the instrument. Such tilts are produced at random by thermal expansion and contraction of the Mylar insulating blanket that surrounds the instrument. The remaining phases are more easily distinguished. The travel times and wave speeds are listed in Table 2.

Our objective in the interpretation of the lunar impact seismograms is to derive a model for the outer shell of the moon that is consistent with the observed travel times of seismic body waves and that will explain the unusual characteristics of the remainder of the wave train. With data from only two artificial impacts recorded at a single station, detailed answers are difficult to achieve.

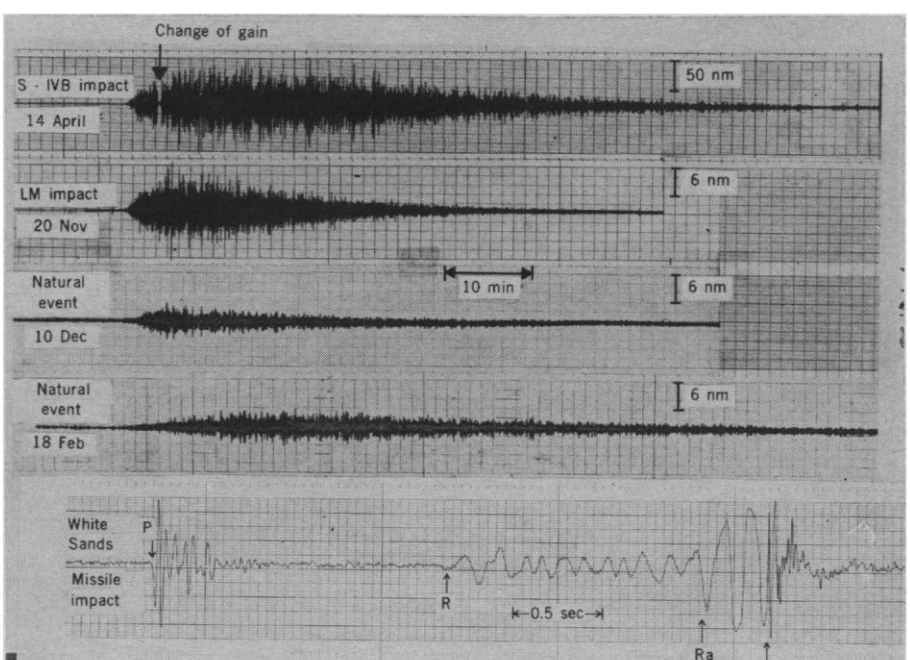


Fig. 2. Signals from the LM and S-IVB impacts and from two of the largest natural events recorded to date. All signals were recorded on the long-period vertical component seismometer. A record of the seismic signal from a missile impact recorded at the White Sands Missile Range is also shown for comparison. For the White Sands record: $P = P$ wave; $R =$ Rayleigh wave; $Ra =$ air-coupled Rayleigh wave; $A =$ atmospheric acoustic arrival; distance, 1.5 km; kinetic energy, $1.5 \cdot 10^{15}$ ergs. Note that the time scale of the White Sands impact signal is greatly expanded relative to that of the lunar signals.

An independent source of information is provided by measurement of seismic velocities on returned lunar samples. The laboratory measurements are obtained by placing a rock sample under pressure and measuring the speeds of waves passing through it by ultrasonic techniques. Since increasing pressure is equivalent to increasing depth within the moon, estimates of seismic velocities as a function of depth within the moon are thus obtained. Experimental results of this type have been reported for the Apollo 11 samples by Kanamori *et al.* (*5*) and Schreiber *et al.* (*6*).

The sample analyzed by Kanamori *et al.* (sample 10057) is a basalt, which contains numerous voids and microfractures. It appears to be moderately shocked, presumably by the meteoroid impact that blasted it out of the lunar material and deposited the rock on the lunar surface. The sample has a bulk density of 2.88 g/cm^3 and an intrinsic density of 3.38 g/cm^3. The data given by Schreiber *et al.* were obtained from a fine-grained vesicular rock (sample 10017) which has a bulk density of 3.1 g/cm^3. Both sets of experimental data show very low surface velocities and a rapid increase in velocity with depth in the upper 20 km of the moon to between 4.8 and 5.6 km/sec for compressional waves. The low surface velocities result from the presence of open pores and microfractures in the samples. The rapid increase in velocity with depth is produced by the closing of cracks and voids under pressure. Complete consolidation of rock material may not occur under the low lunar gravity until depths of at least 20 km have been reached.

Low seismic velocities for the upper few meters of lunar material have been confirmed by measurement of seismic signals generated by the LM and from Surveyor results. From LM signals, a compressional velocity of 108 m/sec for the top several meters of material has been reported (*7*). From Surveyor data, a compressional velocity of 45 m/sec and a shear velocity of 23 m/sec for the top few centimeters of material have been derived (*8*).

The time-distance curves for seismic compressional and shear waves con-

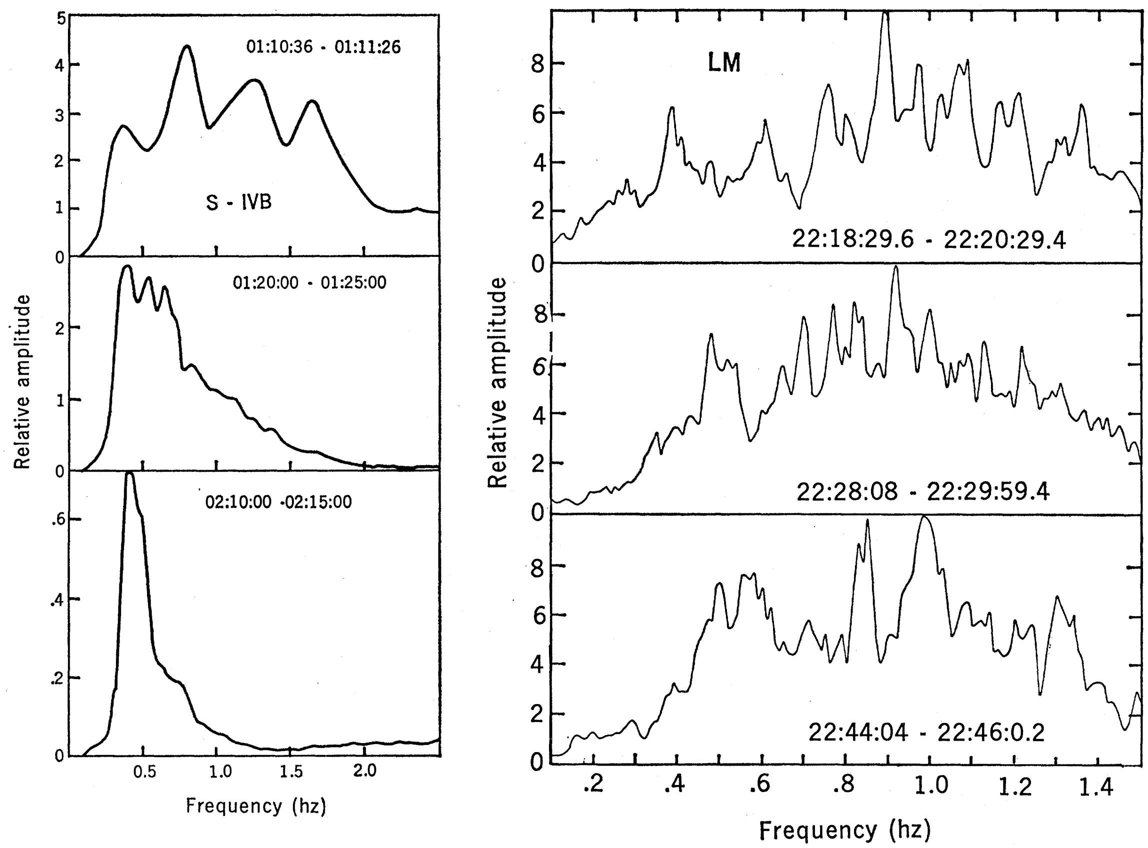


Fig. 3. Frequency spectra of impact signals. The S-IVB signal spectra are corrected for instrument response; the LM spectra are not corrected. Times (G.M.T.) for the beginning and ending of each sample are given on the spectral plots.

structed from the laboratory data are plotted in Fig. 5. The observed travel times of waves from the impacts are also indicated. Seismic waves traveling downward in a medium in which the velocity increases with depth will be refracted according to Snell's law, along curved ray paths that are concave upward, and they will eventually return to the surface at a time that increases with distance as indicated by the time-distance curve. Rays may reflect from the surface one or more times before reaching the detector. The *PP* phase, for example, corresponds to a compressional wave that reflects once from the surface at a point midway between the source and the receiver (for a source at the surface). Thus the travel time for the *PP* phase is equal to twice that of the direct *P* wave, which would be recorded at half the distance. The travel time of the *PP* phase from the S-IVB can, therefore, be plotted on the *P* wave curve at a distance of 67.5 km. In this connection, we note that the ground motion amplitude for the *PP* phase is approximately twice the amplitude of the *P* wave for the S-IVB impact. Thus, it is possible that the first arrival detected from the LM impact is *PP* and that the earlier *P* wave was too small to be detected. In Fig. 5 we have also plotted the LM first arrival on the alternate assumption that it is the *PP* phase.

The velocity-depth curves derived from laboratory measurements are applicable only if the outer 20 to 40 km of Oceanus Procellarum, which corresponds to the greatest depth of penetration of seismic waves from the impacts, consists of rock material similar to the crystalline rocks used in the measurements. The approximate agreement between the travel-time curves observed from the impacts and those computed from the laboratory measurements indicates that this similarity may exist. If so, no major boundary similar to the crust-mantle interface on earth, where an abrupt increase in velocity to about 8.1 km/sec occurs, is present in the outer 20 km of the mare.

The assumption of a model consisting of homogeneous rock under self-compaction leads to a travel-time curve with no sharp changes of slope. This simple model is consistent with the few travel-time data obtained so far, but future impact observations may necessitate its modification. It is clear that the material found at the surface of the mare is too dense to constitute the entire moon unless internal temperatures are high enough so that thermal expansion offsets the effects of compression (*9*). Alternately, the density of the material sampled in the mare may be anomalously high (*10*).

Quality factor Q *of lunar material.* Attenuation of energy in a vibrating system is frequently specified by the quantity Q (quality factor) for the system; or $1/Q$ is the dissipation function, where $2\pi/Q$ is the fractional loss of energy per cycle of vibration of the system. Thus, a high Q implies low attenuation. The Q of a system can be determined by measuring the rate of decay of system energy with time. The value of Q for the lunar material in the region of the Apollo 12 site, as measured by the rate of decay of the LM and S-IVB signal amplitudes, is approximately 3000. This value is in contrast with values of Q between 10 and 300 for most materials of the earth's continental crust.

The relatively high Q inferred for the lunar material may be a consequence of the nearly complete absence of fluids within the outer shell of the moon. Some experimental evidence that supports this possibility has been presented by Pandit and Tozer (*11*). These authors state that evacuation of porous terrestrial rocks to pressures of 10^{-2} torr typically increases the Q of the sample by a factor of 5 over the value measured at 1 atmosphere. Low temperatures in the mare material might also increase the Q. However, no substantive information on subsurface lunar temperatures are available at present.

Lunar seismic reverberation. To explain the unexpectedly long duration of the impact signals, we must either assume that the effective source mechanism was prolonged in some manner or that the long duration is a propagation effect. However, the fact that signals from events of internal origin (moonquakes) and events of external origin (impacts) both show the reverberation is strong evidence for a propagation effect.

The possible propagation mechanisms fall into two general categories: true dispersion effects, where coherent waves propagate at differing group velocities dependent upon wavelength; or scattering effects, where the effective path lengths are increased owing to numerous reflections from acoustic discontinuities. An explanation must take into account that a surface impact source is expected to generate most of its seismic energy in the form of surface waves. Normally, surface waves appear on earthquake or shallow explosion records as trains of sinusoidal waves dispersed by variation of velocity with wavelength due to a vertical gradient of elastic properties in the wave guide. There are usually clear and consistent phase relations between the horizontal and vertical components of motion. Consistent phase relations are lacking in the extended wave trains of the lunar impact signals, as was previously described.

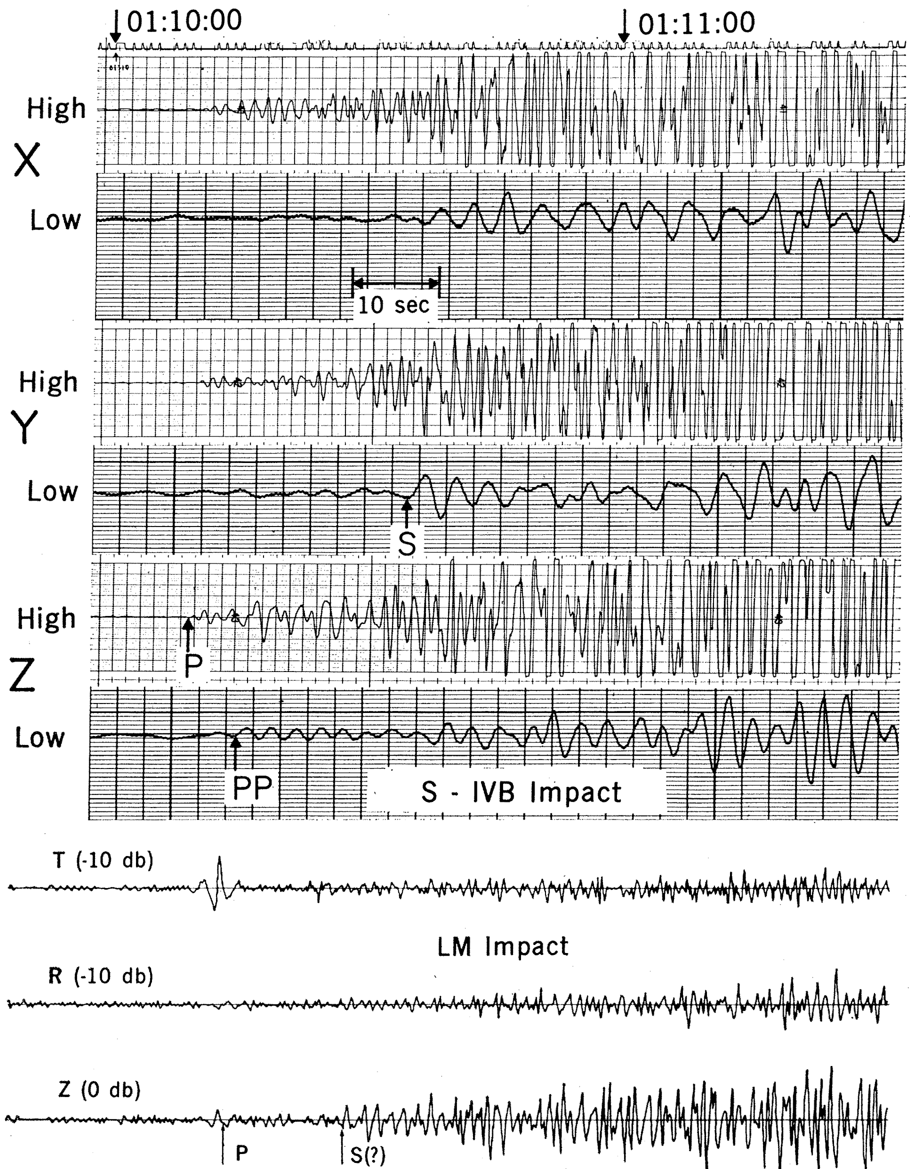


Fig. 4. Initial portions of impact signals on an expanded time scale. (Top) S-IVB signal with high-pass filtering (*High*) to emphasize the *P* wave, and low-pass filtering (*Low*) to emphasize the *PP* and *S* waves (filter corners are at 5 seconds). (Bottom) LM impact signal showing *P* (or *PP*) and *S* phases. For the S-IVB signal, *X* and *Y* are horizontal component seismometers, and *Z* is the vertical component. The *X* and *Y* components are approximately transverse and radial, respectively, for the S-IVB impact. For the LM signal, *R* and *T* are the radial and transverse (horizontal) components and *Z* is the vertical component. The S-IVB *PP* phase is also plotted on the *P* travel-time curve at half the distance and half the travel time. The LM *P* phase is also plotted in this manner under the assumption that it may actually be a *PP* arrival (see text). Time on the records is the time at which the data are recorded at the receiving station on earth.

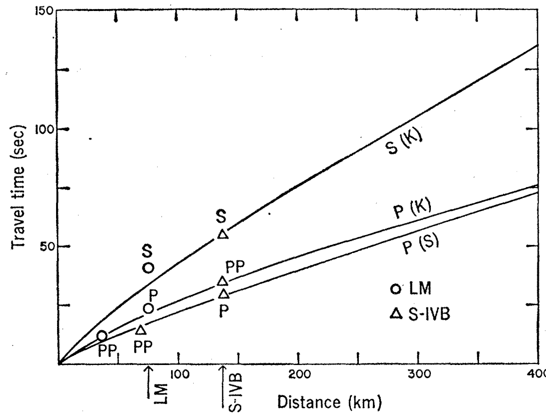
Scattering hypothesis. The scattering hypothesis explains the character of the impact signals by postulating that the moon not only has a high Q but is also very heterogeneous, at least in the near-surface region of the mare. If the lunar medium contains inhomogeneities comparable in scale to the wavelength of the propagating seismic waves, significant scattering will occur. Scattering would tend to increase the duration of the observed seismic waves, would suppress the appearance of distinct phases within the wave train, and would reduce coherence between the horizontal and vertical components of surface motion.

That the outer shell of the moon might be highly heterogeneous is not surprising in view of the extreme age of the surface. Meteoroid bombardment would certainly have shattered the original lunar material to depths approaching 50 km, and the terrestrial geologic processes that remove these scars would be absent. Also, the surface lava is reported to have a very low viscosity and a high thermal coefficient of expansion (*12*). A lava with these properties might form extensive networks of lava tubes within the flow and would fracture extensively after solidifying. Alternately, the heterogeneity might simply be characteristic of the structure produced by the influx of blocks of material during the final stages of accumulation of the moon.

If scattering is sufficiently intense, the propagation of seismic waves might properly be described by diffusion theory, in which seismic energy "flow" is proportional to the gradient of energy density. With proper selection of parameters, the envelopes of the impact reverberations can be fit quite accurately by diffusion theory except for the first few minutes, when scattering would not be expected to be the predominant mechanism (*13*). If scattering is an important factor in lunar seismic wave transmission, heterogeneity must exist within the lunar material on a scale of from several hundred meters, or less, to several kilometers. An alternate approach to the scattering problem has been given by Steg and Klemens (*14*).

Dispersion hypothesis. At least two seismic wave guides are likely to occur in the outer shell of the moon. One is the unconsolidated surface debris layer (the lunar regolith) which, together with a shattered crystalline layer, forms a low-velocity zone overlying the more competent, higher-velocity basement rocks. A deeper wave guide is formed by the velocity gradient in the basement rock, which is inferred from the pressure coefficient of velocity measured in the laboratory on lunar crystalline rocks.

Both of these wave guides would "trap" surface waves (Love waves and Rayleigh waves), but important distinctions between their effects are to be expected. The very high modes of the basement wave guide can be thought of as multiply reflected body waves analogous to the Sofar channel in the oceans or as a "whispering gallery" phenomenon. Neither low nor high modes of surface waves in the basement channel can account for the prolonged reverberation without invoking scattering, because the difference between the maximum and minimum group velocities is small for the frequencies of interest. The situation is different for the surface low-velocity zone, where the dispersion curves (Fig. 6) for Rayleigh waves of the fundamental mode show a maximum group velocity of 1660 m/sec and a minimum group velocity of 30 m/sec for the period range of from 1.2 to 1.5 seconds. For this model, surface waves with predominant periods of from 1.2 to 1.5 seconds and a duration of 74 minutes for the S-IVB source distance of 135 km would be expected. The predominant period varies directly with the thickness of the low-velocity surface layer, and inversely with the shear velocity. The observed signal for the S-IVB impact shows a lower frequency than does the LM impact, which implies a thicker surface layer (or lower shear velocity) for the S-IVB wave path, or excitation of a higher mode by the LM impact. The signal duration predicted by this model is satisfactory for the LM but requires some multipathing for the S-IVB impact. Multipath propagation or a mixture of modes is also required to explain the lack of correlation between the three components of ground motion.



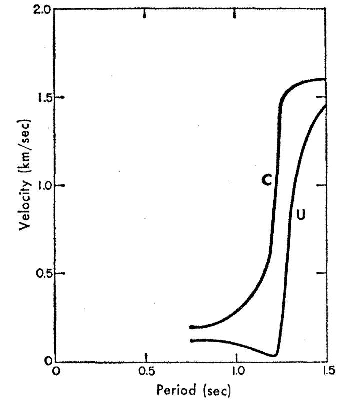


Fig. 5 (left). Travel times of seismic waves from the lunar impact signals. Solid curves are derived from laboratory measurements of seismic velocities on returned lunar samples. $P(S)$ = compressional wave velocities measured for a lunar rock sample by Schreiber *et al.* (*6*); $P(K)$, $S(K)$ = compressional and shear wave velocities measured for a lunar rock sample by Kanamori *et al.* (*5*). Fig. 6 (right). Phase (C) and group velocity (U) curves for Rayleigh waves of the fundamental mode. The model consists of a 30-m low-velocity layer with shear velocity of 0.1 km/sec, which overlies basement material with shear velocity of 1.8 km/sec.

The closest terrestrial analog to the lunar impact signals discovered thus far are surface waves (first-shear-mode Rayleigh waves) recorded on the ocean bottom from shallow focus earthquakes originating in the Mid-Atlantic Ridge (*15*). These signals show a gradual buildup and decay, very long duration with nearly constant period, and inferred Q values from 1500 to 2000 for the oceanic wave guide. The signals are observed only when significant amounts of unconsolidated sediments are present along the path between the source and the receiver. If we ignore the water layer, the oceanic structure may be quite similar to the lunar structure, with the sediment layer being equivalent to the low-velocity outer zone of the moon.

The presence of a thick permafrost zone overlain by 100 to 200 m of low-velocity material, as suggested by Shubert *et al.* (*16*), might also provide a model consistent with this interpretation.

The terrestrial oceanic signals clearly demonstrate that the major features of the lunar seismic reverberation can be explained by dispersion in a high Q wave guide that contains material of low shear velocity. If the lunar signals are higher-mode Rayleigh waves, this would also explain the absence of low-frequency energy in these signals, since higher modes cannot propagate at frequencies below a fixed cutoff frequency. The observed differences in the reverberation frequencies of the natural events would be explained by the dispersion hypothesis as resulting from lateral variations in the shear velocity or thickness of the surficial layer.

Efficiencies of impacts as seismic sources. The calculation of seismic energy coupled into the lunar structure by the impacts is dependent on whether we assume scattering or dispersion to be the predominant mechanism that explains the seismic reverberations. Calculated seismic energies based on both hypotheses are given in Table 3. For these calculations, we have assumed a density of lunar material of 3.3 g/cm^3; Q of 3000; constant signal frequencies of 1 and 0.5 hz for the LM and S-IVB, respectively; and cylindrical spreading of energy with thicknesses of layers through which the waves are transmitted of 2 and 4 km for the LM and S-IVB signals, respectively.

The efficiency of the S-IVB impact is 4 to 12 times greater than that of LM impact (see Table 3). This difference was anticipated, since the LM struck the surface at a very shallow angle and would thus have transferred a small fraction of its momentum to the surface at the primary impact point, whereas the S-IVB struck the surface at a much steeper angle and would have transferred essentially all of its momentum at the primary impact point.

Conversion efficiencies of between $1 \cdot 10^{-5}$ and $5 \cdot 10^{-5}$ were obtained for missile impacts at White Sands, New Mexico (*3*). McGarr *et al.* (*17*) obtained efficiencies of between $8 \cdot 10^{-7}$ and $6 \cdot 10^{-5}$ for various target materials for hypervelocity impacts. Thus, the efficiency of the S-IVB impact for production of seismic energy is in approximate agreement with efficiencies obtained for impact experiments on earth.

Conclusions. We summarize below in four major conclusions:

1) Body wave velocities measured from the LM and S-IVB impacts are in reasonably close agreement with the velocities predicted from laboratory measurements on lunar rock samples. This result implies that the rock material collected at the surface of the mare near the Apollo 12 landing site is similar to the material that forms the mare to depths of at least 20 km.

Present data for the mare region near the Apollo 12 landing site suggest that the outer shell of the moon consists of low-velocity material near the surface, with velocity increasing rapidly with depth in the upper 20 km. It is unlikely that a major discontinuity similar to the Mohorovicic discontinuity, which defines the lower limit of the crust of the earth, exists in the outer 20 km of the moon.

The fact that NASA was able to achieve such high targeting accuracy for the S-IVB and that the resulting seismic signals were so large suggests that planned future impacts can be extended to ranges up to at least 500 km and that the data returned will provide information on lunar structure to depths of several hundred kilometers.

2) The lunar seismic reverberation can be explained equally well as resulting from dispersion of surface waves or from scattering or, perhaps more likely, from a combination of these mechanisms. Scattering of surface waves implies the presence of heterogeneity in the outer shell of the moon on a scale from several hundred meters, or less, to several kilometers. Surface irregularities may contribute significantly to the scattering. The dispersion hypothesis requires the presence of a low-velocity outer zone. The presence of this zone to depths of several meters has been confirmed by measurements of seismic waves from sources associated with the LM and from Surveyor measurements. Very low absorption of seismic wave energy in the lunar material is inferred, independent of the assumed mode of propagation. This may be a consequence of the absence of fluids in the near-surface materials or low temperature or a combination of these factors.

3) Estimates of the fraction of impact kinetic energy that is converted to seismic wave energy are reasonably close to results obtained from impact experiments on earth.

4) Seismic signals of natural origin are produced by moonquakes and by meteoroid impacts. The low level of detectable lunar seismic activity relative to the earth suggests that the outer shell of the moon is tectonically stable as compared with the lithosphere of the earth. Tidal stresses appear to be an important factor in the occurrence of moonquakes. Meteoroid flux in the kilogram mass range, as it has been inferred from seismic data, is in approximate agreement with the Hawkins flux estimate (*2*).

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18. We thank L. Sykes and C. Scholz for reviewing the manuscript and offering helpful suggestions. Supported by the National Aeronautics and Space Administration under contracts NAS9-5957 and NAS9-5632. Computer facilities were provided by the NASA Goddard Space Flight Center, Institute for Space Studies, New York. Lamont-Doherty Geological Observatory Contribution 1586. Hawaii Institute of Geophysics Contribution 345.

9 July 1970; revised 25 August 1970

Mechanism for the Water-to-Air Transfer and Concentration of Bacteria

Abstract. *Air bubbles breaking at the air-water interface can remove bacteria that concentrate in the surface microlayer and eject the bacteria into the atmosphere. The bacterial concentrations (numbers per milliliter) in the drops ejected from the bubbles may, depending on drop size, be from 10 to 1000 times that of the water in which the bubbles burst.*

Woodcock (*1*) found that small droplets in the air were most likely the causative factor in human respiratory irritation associated with high concentrations of plankton in the sea. He showed that droplets carrying the irritant were easily generated by bubbling through the water rich in plankton. He later showed (*2*) that organic surface films could be removed from the water by the drops and suggested that bacteria in the surface layers of either sea water or fresh water might be removed in a similar way. Higgins (*3*), following up on this suggestion, collected the aerosol produced by bubbling through water that contained several species of bacteria. The ratios of the recoveries of some of the bacteria in the aerosol were higher than expected. He deduced from this that *Serratia marcescens* in particular must be concentrated at the surface of the liquid. The biological implications of these surface phenomena have been reviewed (*4*).

Material in the upper few microns of the surface is easily removed by drops from bursting bubbles. It has been investigated in the laboratory (*5, 6*) and it has been shown that the nat-