THE VELOCITY STRUCTURE OF THE LUNAR CRUST*

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Abstract. Seismic refraction data, obtained at the Apollo 14 and 16 sites, when combined with other lunar seismic data, allow a compressional wave velocity profile of the lunar near-surface and crust to be derived. The regolith, although variable in thickness over the lunar surface, possesses surprisingly similar seismic properties. Underlying the regolith at both the Apollo 14 Fra Mauro site and the Apollo 16 Descartes site is low-velocity brecciated material or impact derived debris. Key features of the lunar seismic velocity profile are: (i) velocity increases from 100–300 m s^{-1} in the upper 100 m to $\sim$4 km s^{-1} at 5 km depth, (ii) a more gradual increase from $\sim$4 km s^{-1} to $\sim$6 km s^{-1} at 25 km depth, (iii) a discontinuity at a depth of 25 km and (iv) a constant value of $\sim$7 km s^{-1} at depths from 25 km to about 60 km. The exact details of the velocity variation in the upper 5 to 10 km of the Moon cannot yet be resolved but self-compression of rock powders cannot duplicate the observed magnitude of the velocity change and the steep velocity-depth gradient. Other textural or compositional changes must be important in the upper 5 km of the Moon. The only serious candidates for the lower lunar crust are anorthositic or gabbroic rocks.

1. Introduction

A seismic refraction experiment was used to study the characteristics of the lunar near-surface at the Apollo 14 and 16 landing sites. The results of these experiments, together with the recording of Lunar Module (LM) ascent stage and upper stage of the Saturn rocket (S-IVB) impacts allow an interpretation of the seismic velocity structure in the lunar interior to be made. In this paper we want to examine some of the following questions: What are the acoustic or seismic properties of the lunar near-surface material? How thick is the lunar regolith at the Apollo 14 and 16 sites? Are there distinct seismic horizons and do they correlate with any geological horizons? Is permafrost present near the lunar surface? Are there characteristic differences in the shallow seismic velocities between the maria and the highlands? How do the shallow seismic results interface with the larger-scale seismic results? What are the implications, compositional or otherwise, of the seismic results for the outer lunar interior?

In the study of the Earth's crust seismic refraction profiling has often been used. The technique involves the recording of explosive charges at various distances from an array of seismometers (geophones). On the Moon a 91 m linear array of three geophones was deployed by the Apollo astronauts and several seismic energy sources were used: an astronaut-activated thumper device with small explosive initiators, a

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mortar package containing rocket launched grenades and the impulse produced by the thrust of LM-ascent stage as it departed from the lunar surface. Technical details of the experiment can be found in Kovach *et al.* (1971) and Kovach *et al.* (1972).

2. Apollo 14 Results

Apollo 14 landed at latitude 3°40′24″S, longitude 17°27′55″W in the Fra Mauro region. Here the lunar surface is covered by a stratigraphic unit, called the Fra Mauro formation, which is presumed to be the ejecta deposit produced by the large impact which formed the Imbrium basin. At the landing site the Apollo 14 crew deployed the string of three geophones in a southerly direction across the lunar surface and used the thumper device to generate seismic signals. The thumper was fired at 4.57 m intervals along the geophone line.

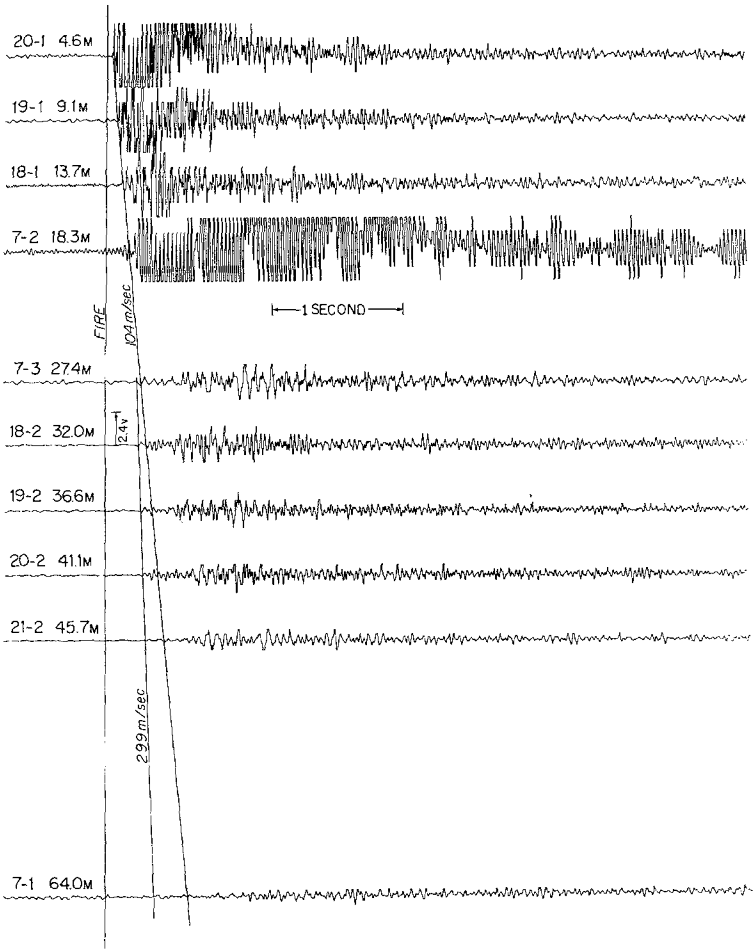

Fig. 1. Composite seismic record section aligned in time to same instant of thumper firings at the Apollo 14 site. First number refers to shot number. Second number to geophone on which recorded.

A composite seismic record section aligned in time to the same instant of thumper firings at the Apollo 14 site is shown in Figure 1. The first number refers to the thumper explosive initiator used and the second number to the geophone on which the data were recorded. The northernmost geophone which was closest to the central telemetry station was designated geophone 1.

Two compressional wave (P-wave) velocities are recognized on the composite record section. A first arrival with a P-wave velocity of 104 m s^{-1} is observed out to a distance of 22.7 m where a faster arrival with a velocity of 299 m s^{-1} is identified. No variation in P-wave velocities was observed across the horizontal section sampled, as was evidenced by the conformance of the seismic velocities measured along the geophone line. The lunar near-surface at the Apollo 14 site is thus characterized by a surface layer of 104 m s^{-1} velocity overlying material possessing a seismic velocity of 299 m s^{-1}.

Seismic signals were also generated from the rocket thrust produced by the ascent of the landing module from the Moon and recorded by the Apollo 14 passive seismometer at a distance of 178 m (584 ft). These travel-time data are shown in Figure 2 compared to the extrapolated travel-time distance curves determined from the astronaut thumper firings. The observed travel-times are in close agreement with the extrapolated travel-times. However, a first arrival was observed with a travel-time somewhat faster than that predicted by a refraction from the top of the 299 m s^{-1} horizon indicating that a material with a faster intrinsic compressional-wave velocity lies beneath the 299 m s^{-1} material.

If the assumption is made that the material underlying the 299 m s^{-1} material possesses an infinite compressional wave velocity, a maximum estimate of the 299 m s^{-1} material can be derived. Furthermore, as shown in Figure 2, if the critical distance for

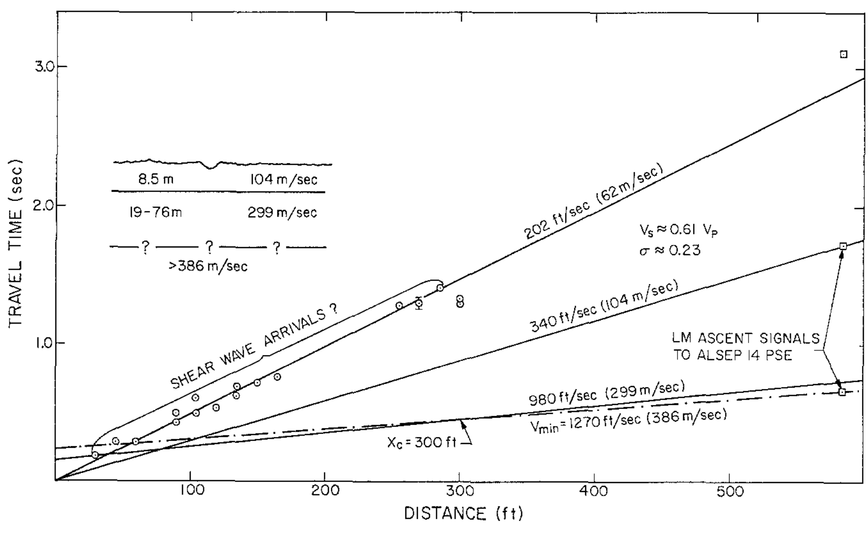

Fig. 2. Travel-times measured at the Apollo 14 site.

the arrival of the head wave traveling through the underlying material is taken as 100 m (the length of the geophone line) a minimum velocity of 386 m s^{-1} is inferred for the underlying material. The range of possible thickness, therefore, for the 299 m s^{-1} material is 17 to 88 m.

The depth to the 299 m s^{-1} refracting horizon is 8.5 m. The low seismic velocity of 104 m s^{-1} is suggestive of a porous and highly brecciated rock material and thus believed to be representative of the fragmental veneer of unconsolidated particulate debris – the lunar regolith that covers the surface of the Fra Mauro site. This measured thickness of 8.5 m for the regolith agrees well with that estimated solely on geological evidence. By photographic studies of the depth at which blocky floors appear in fresh craters, it can be inferred that the fragmental, surficial layer that overlies the more consolidated or semiconsolidated substrate at the Fra Mauro site ranges in thickness from 5 to 12 m (Offield, 1970).

A velocity of 299 m s^{-1} is comparable to that observed in the upper part of Meteor Crater ejecta material (Watkins and Kovach, 1971) and it is thus quite reasonable to assume that this velocity is representative of ejecta from Mare Imbrium. A thickness estimate of 19 to 76 m is comparable to geological estimates of 100 m or so for the Fra Mauro formation (Offield, 1970). The returned lunar samples have also definitely revealed that the Fra Mauro formation is primarily comprised of breccias (Wilshire and Jackson, 1972).

The relatively low compressional wave velocities that were measured in the lunar near-surface argue against the presence of substantial amounts of any shallow permafrost at this particular site. Measured velocities in permafrost vary greatly – depending on such factors as lithology, porosity, and degree of interstitial freezing – but typically range from 2438 to 4572 m s^{-1} (8000 to 15000 ft s^{-1}) (Barnes, 1966).

3. Apollo 16 Results

The landing site for Apollo 16 was at latitude 8°59′29″S, longitude 15°30′52″E. This location, known as the Descartes site, was chosen to sample and study the constructional units, called the Cayley formation and the associated Descartes material which are a part of the lunar highlands. It had been presumed by some that these formations were primarily volcanic in origin and that lava flows might be found at this location (Milton and Hodges, 1972).

At this site the three geophones were deployed on the Cayley formation on a highly cratered uneven area at a bearing of 287° (clockwise from north) from the central telemetry station. Geophone 3 was deployed 90 m northwest of geophone 1 (Figure 3).

Figure 4 is a record section aligned in time to the same instant of firing for thumper shots 2 to 10 as recorded at geophone 2. There is no difficulty in recognizing the onset of the seismic signals out to a distance of 40 m; but at greater distances the onset of the seismic wave arrivals are more uncertain because of the emergent beginnings. Figure 5 is an expanded time scale record section for thumper shots 2 to 9 as recorded at geophone 3.

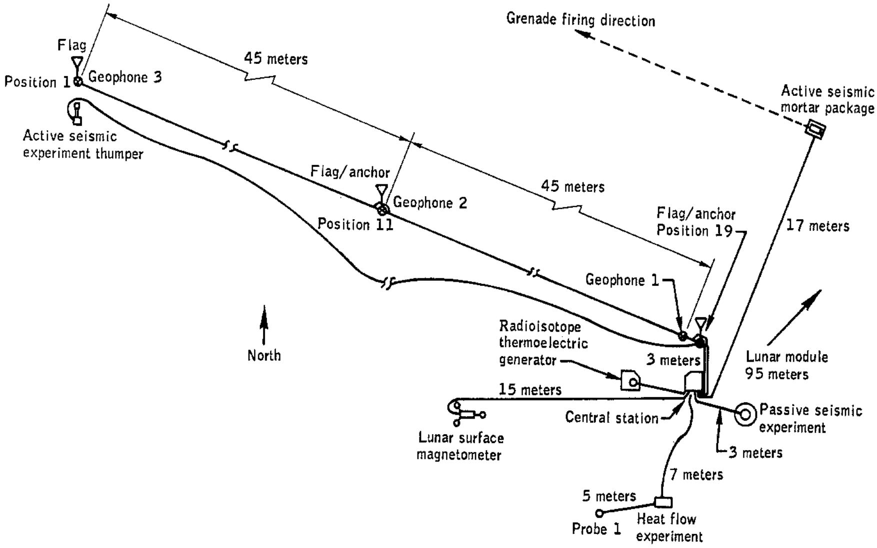

Fig. 3. Schematic diagram showing deployment of geophysical experiments at the Apollo 16 site.

The travel-time-distance data from the Apollo 16 thumper firings are shown in Figure 6. Only one compressional wave velocity is evident in these data – a direct arrival possessing a velocity of 114 m s^{-1}. Again no variation in the measured surface velocity was noted across the geophone line.

Seismic signals of particular interest at the Apollo 16 site were generated by the ascent of the lunar landing module and the detonations of rocket propelled grenades. The seismic signals from the ascent of the landing module are shown in Figure 7.

The mortar package assembly was located 14 m from geophone 1 pointed to fire parallel to the geophone line and downrange toward geophone 3 (Figure 4).

The seismic signals produced by the detonation of the nearest grenade (0.1 lb of high explosive) are shown in Figure 8. The records are noisy prior to the onset of the detonation signal because the grenade launch itself produced a disturbance which had not completely decayed to low level prefiring conditions. However, the desired signals can be recognized on the basis of a change in frequency inasmuch as the detonation signals possess a predominant signal frequency of 10 Hz compared to 15–20 Hz for the grenade launch itself. Figure 9 shows the seismic signals produced by the detonation of 0.3 lb of explosive at a distance of about 400 m. 0.6 lb of high explosives were also detonated at a distance of 1000 m from the geophone line. However, because of the reverberation produced from the grenade launch only weak surface waves could be recognized.

The travel-time data for the seismic signals recorded at the Apollo 16 site are shown in Figure 10. Seismic arrivals with velocities of 114 m s^{-1} and 250 m s^{-1} are observed. The depth to the top of the 250 m s^{-1} refracting horizon is 12.2 m. Again it is inferred that the material with a seismic velocity of 114 m s^{-1} is representative of the regolith

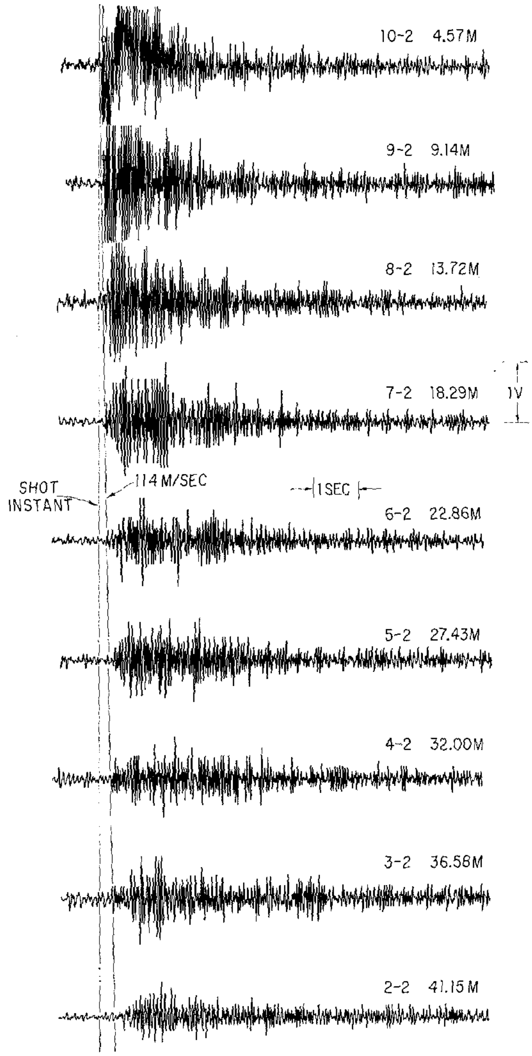

Fig. 4. Seismic record section recorded at Apollo 16 site. Line marked 114 m s^{-1} shows first arrival of P-wave.

at the Apollo 16 site. Estimates of the regolith thickness at the Apollo 16 site based solely on geological considerations can be ambiguous. Regolith thicknesses are commonly estimated from the total crater population with a higher density of craters implying a greater regolith thickness. If one assumes that all craters observed on the Cayley formation at the Apollo 16 site are impact craters, the indicated mean regolith thickness is 22 m (Oberbeck, 1971). On the other hand, restricting the analysis to only concentric craters suggested a regolith thickness of about 7 m. The refraction from the 250 m s^{-1} horizon could not be recognized out to a distance of 445 m, but if it is indeed present the maximum thickness for the 250 m s^{-1} material is 220 m.

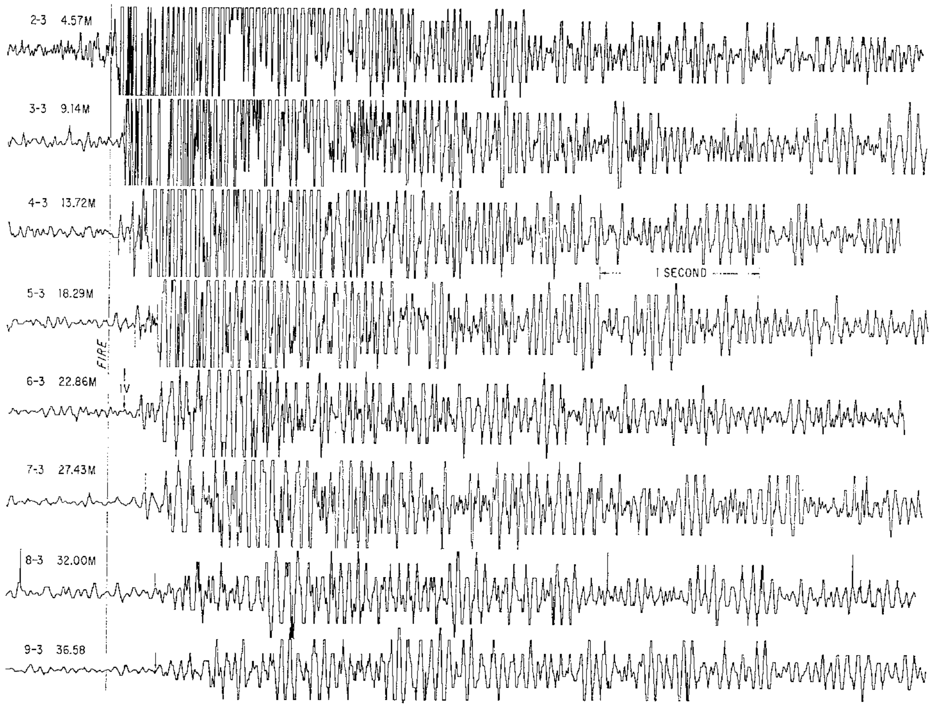

Fig. 5. Expanded time playout of thumper firings 2–9 recorded at geophone 3 at the Apollo 16 Descartes site. Arrows indicate the first arrival. Note the more emergent beginning with an increase in distance.

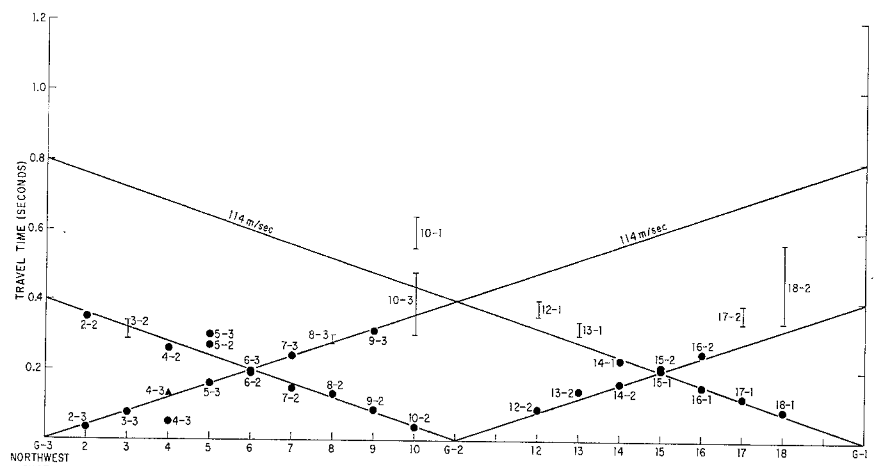

Fig. 6. Travel-time data for thumper firings at the Apollo 16 site. The data points are shown as black circles; the first number refers to the thumper firing, the second number to the geophone on which the data were recorded. Distance between firing locations (except skipped positions 11 and 19) is 4.57 m (15 ft). Note uniformity of measured velocity across section sampled.

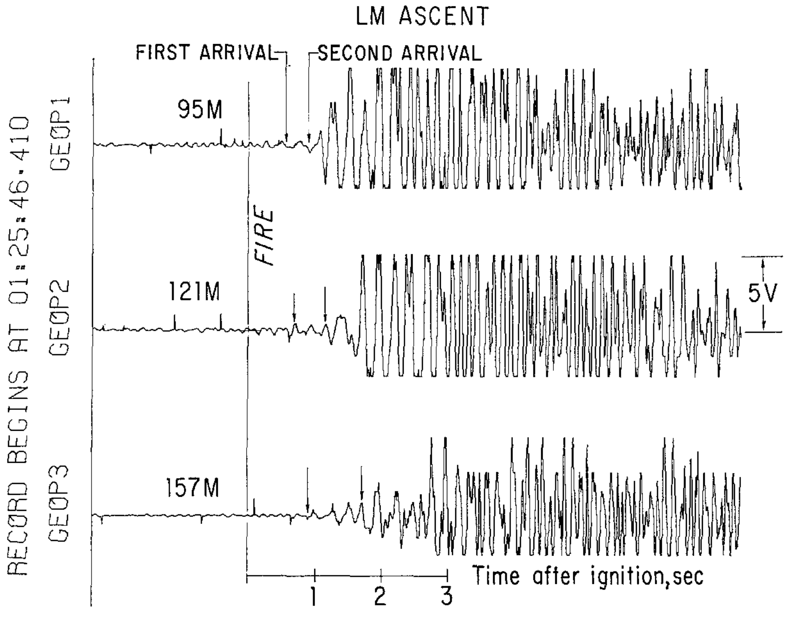

Fig. 7. Seismograms recorded on Apollo 16 geophone array from ascent of Lunar Module (LM) from lunar surface. Arrows indicate first and second arrivals.

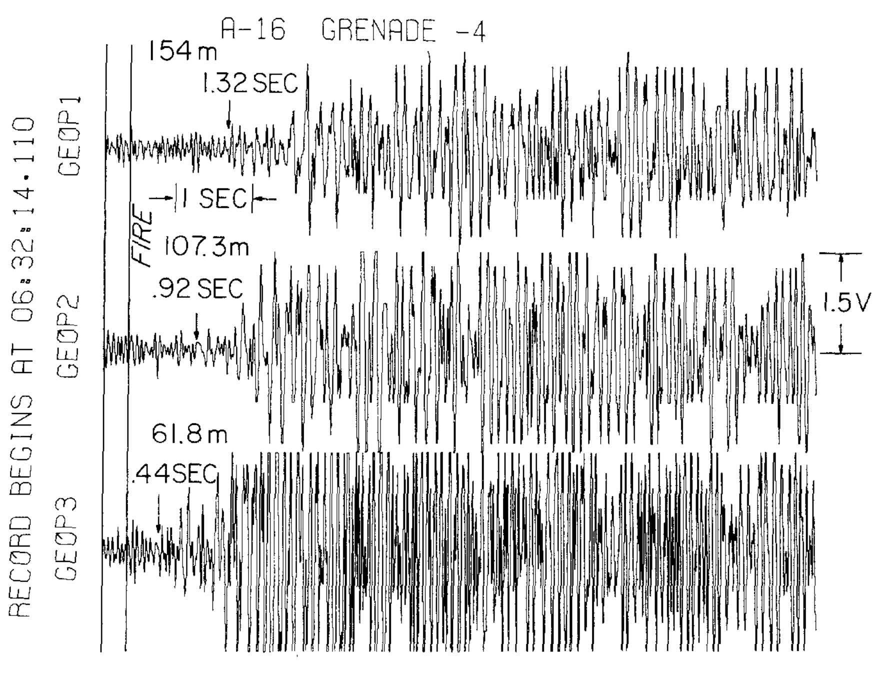

Fig. 8. Seismograms produced by detonation of 0.1 lb of high explosive at the Apollo 16 site.

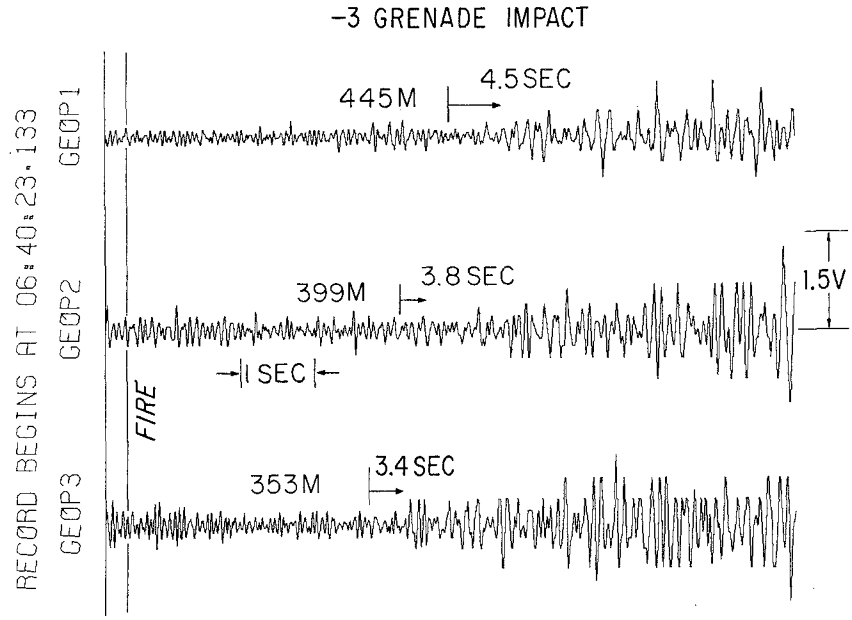

Fig. 9. Seismogram produced by detonation of 0.3 lb of high explosive at the Apollo 16 site recorded on geophone array at distances shown.

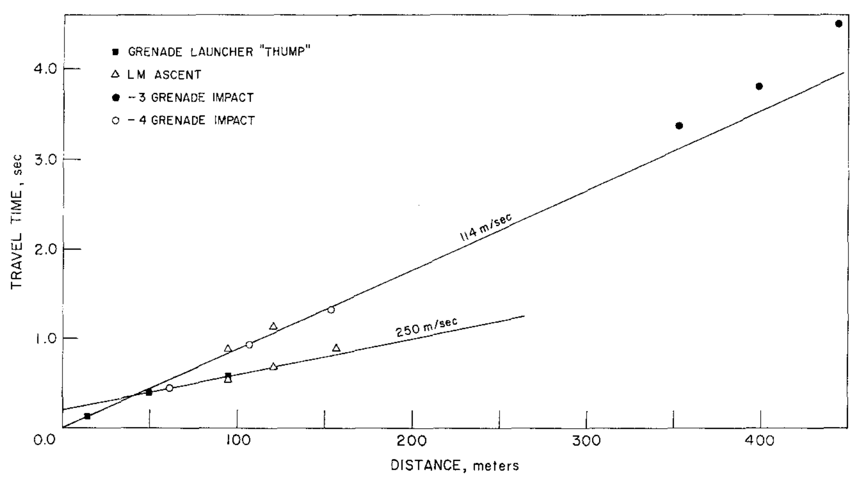

Fig. 10. Travel-times measured at the Apollo 16 site.

The underlying material with a seismic velocity of 250 m s^{-1} is certainly not indicative of competent lava flows which typically, on Earth, have seismic velocities greater than about 800 m s^{-1} (Watkins *et al.*, 1972). The returned lunar samples have also revealed that the Cayley formation does not consist of lava flows as had been postulated (Milton and Hodges, 1972), but rather consists of perhaps interstratified breccias. No bedrock appears to have been sampled by the Apollo 16 crew. The measured value of 250 m s^{-1} is close to the value of 299 m s^{-1} measured for the underlying Fra Mauro breccias at the Apollo 14 site. Thus it seems clear that at both the Apollo 14 and 16 sites the lunar surface is underlain by low-velocity brecciated material or impact derived debris.

4. Discussion and Conclusions

The seismic velocity of the 114 m s^{-1} measured at the Descartes site can be compared to the velocities for the regolith of 104, 108 and 92 m s^{-1} measured at the Apollo 12, 14 and 15 sites respectively (Latham *et al.*, 1972; Kovach *et al.*, 1971). Even though there is some variability in the velocity from site to site the process of fragmentation and communition by meteoroid impacts has produced a layer of surprisingly similar seismic properties, arguing against any major regional difference in the near-surface acoustical properties of the Moon.

The velocity model for the upper 25 km of the Moon, compatible with the impact data recorded by the network of Apollo seismometers (Toksőz *et al.*, 1972) and the lunar seismic refraction experiments is shown in Figure 11. Several salient features can be pointed out.

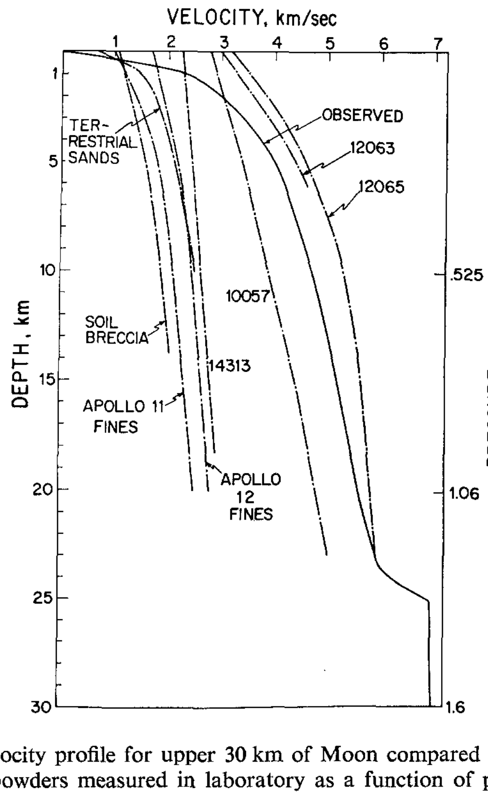

Fig. 11. Observed velocity profile for upper 30 km of Moon compared to velocities of lunar and terrestrial rocks and powders measured in laboratory as a function of pressure. Lunar rocks are identified by sample number.

(i) The seismic velocity increases very rapidly from values of 100–300 m s^{-1} in the upper 100 m or so of the Moon to a value of ~4 km s^{-1} at a depth of 5 km.

(ii) The velocity increases more gradually from ~4 km s^{-1} at 5 km depth to ~6 km s^{-1} at a depth of 25 km.

(iii) A discontinuity in seismic velocity is present at a depth of 25 km.

(iv) A nearly constant value of ~7 km s^{-1} from a depth of 25 km to about 60 km.

The exact details of the velocity variation in the upper 5–10 km of the Moon cannot yet be resolved (i.e. whether it is smooth as depicted or a stepwise increase) but one simple observation can be made. Self-compression of any rock powder such as the Apollo 11 or 12 soils or terrestrial sands cannot duplicate the observed magnitude of the lunar velocity change and the steep velocity-depth gradient (~2 km s^{-1} km^{-1}).

Experiments on returned lunar soils (Anderson *et al.*, 1970; Kanamori *et al.*, 1970, 1971; Mizutani *et al.*, 1972; Warren *et al.*, 1971) and truly hydrostatic measurements on terrestrial sands and basaltic ash (Talwani *et al.*, 1972) suggest velocity-depth gradients of 0.4 to 0.8 km s^{-1} km^{-1} but such gradients only persist to a pressure of ~50 bars (depth ≈1 km in Moon). At higher pressures the velocity gradient decreases to values 10 to 20 times less than the gradient at low pressure.

Recent experimental results on unconsolidated sands and rock powders show that no unique relation exists between seismic velocity and porosity in granular material. Secondly, velocities in unconsolidated materials do not exhibit excess pressure memory (Talwani *et al.*, 1972). Rather there is an irreversible change in porosity accompanying a completely reversible variation in velocity as the material is repeatedly cycled to pressures of 2.4 kbar (~50 km lunar depth).

In view of the lack of verifying experimental data it would seem unlikely to expect that a deep rock powder layer (of several km) with a velocity gradient of 1.35 km s^{-1} km^{-1}, as proposed by Gold and Soter (1970) and further discussed by Gold (1971a, b), can be invoked to explain the shallow lunar velocity variation. It has also been argued that plastic deformation, sintering and excess pressure memory (presumably from large meteorite impacts) make the lunar *in situ* velocities much higher than is observed in the laboratory (Jones, 1972). The important observation is that the increase of seismic velocity in the shallow lunar interior is much too large to be due simply to self-compaction of rock powders. *The implication is that other effects such as composition or textural changes must be important in the upper 5 km of the Moon.* It, of course, might be possible to invoke *ad hoc* hypotheses to explain this velocity variation but the seismic experiment planned for Apollo 17 should shed light on this important paradox. The primary purpose of the Apollo 17 seismic profiling experiment is to gather needed travel-time data in the distance range of 0.1 km to 10 km. Explosive charges will be detonated out to distances of 2.5 km and the LM ascent stage is targeted to impact within 10 km.

From about 4 km to 25 km depth the physical properties of the lunar rocks are probably dominated by cracks, pores and intergranular effects. Rocks have intercrystalline cracks and pores and have very low seismic velocities at atmospheric pressure.

Seismic velocities increase rapidly over the pressure range from zero to 1.5 kbar ($\sim$25 km depth in Moon) where velocity values change only minor amounts with a further increase in hydrostatic pressure (Nur and Simmons, 1971).

Because of this large pressure effect on seismic velocities over this depth range in the Moon, probably not much can unambiguously be said about composition. However, the observed velocity variation does lie between the measured velocity values for returned lunar basaltic rocks 10057, 12065 and 12063 suggesting compatibility with a basaltic composition.

At a depth of 25 km the seismic velocity increases only very slightly to a value of 7.0 km s^{-1} at a depth of 65 km. Because at these pressures (>1.5 kbar) much of the crack and porosity effects in rocks have been minimized it is possible to use laboratory measurements on rocks to make some compositional inferences. Temperature effects can also be excluded because at the anticipated lunar temperatures their effect is, at worst, to produce a decrease in seismic velocity by 0.1–0.2 km s^{-1}.

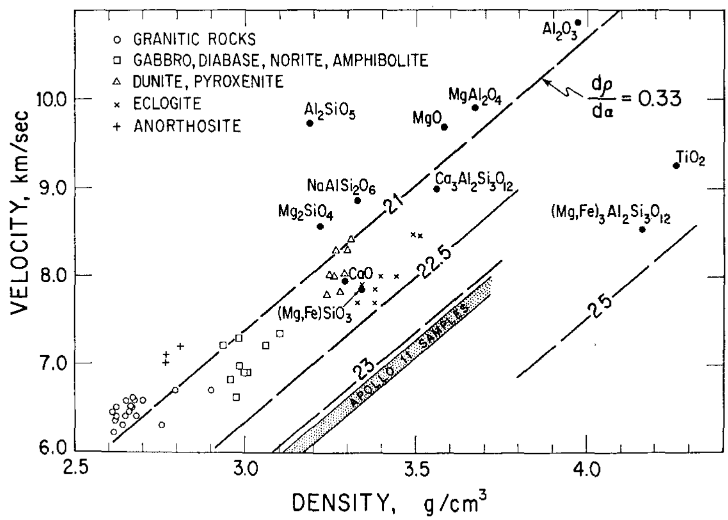

Fig. 12. Velocity and density of rocks and minerals as a function of mean atomic weight.

Figure 12 is a plot of the seismic velocity in rocks and minerals as a function of mean atomic weight. For a given mean atomic weight seismic velocity increases as the density increases. The effect of an increase in iron content is to increase the density and mean atomic weight but decrease the seismic velocity. It can be seen that for the various rocks shown only anorthosite or gabbro (including diabase, norite or amphibolite) have a seismic velocity of $\sim$7 km s^{-1} and can be considered as serious candidates for the lower lunar crust. Eclogites, dunites and pyroxenites all have seismic velocities higher than that observed and should not be considered further. It is also obvious that seismic data alone cannot distinguish between gabbro and anorthosite and petrologic and geochemical arguments must be introduced.

Acknowledgements

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