LUNAR VISCOSITY AS OBTAINED FROM THE SELENOTHERMS*

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Abstract. Based on the selenotherms $T(z)$ (= temperature-depth functions) and melting point-depth functions $T_m(z)$ viscosity values $\eta(z)$ are calculated. According to two different creep laws used, two sets of viscosity values are obtained. Viscosities in the outer part of the Moon are found to be larger than those anywhere on Earth. These high values of η explain the large elasticity Q found in lunar seismograms. Viscosities below about 500 km in depth are so small that, at present, some kind of convection or a flow of matter is possible. Tidegenerated moonquakes at depths of around 1000 km seem to be connected with some viscous process. From considerations of viscosities at the time period of mare filling, some selection of ancient selenotherms may be performed.

1. Introduction

Viscosity values for the outer part of celestial bodies may be obtained in 3 different ways. The first one, which may be called the 'relaxation time method', has been applied for calculating a 'regional' viscosity from the observation of the uplift of the Fennoscandian shield (Haskell, 1935; McConnel, 1965; Magnitzky, 1967; Artyushkov, 1967; Post and Griggs, 1973; and others) and the Canadian Shield (Crittenden, 1967; Berry and Fuchs, 1973; and others). For the Moon this method has been applied by Arkani-Hamed (1973a, b) in order to estimate the viscosity on the basis of calculated stress differences caused by the mascons. Munk and McDonald (1960) have interpreted the fact that the Earth is more oblated than it should be hydrostatically to mean that there is a lag in its adjustment to the gradually decreasing rate of rotation. The relaxation time of some 10^7 yr again is related to a global or average viscosity (of the lower mantle).

The second method for calculating viscosities is based on the relation between viscosity and temperature for a given material with a pressure-dependent melting point and makes use of geotherms (selenotherms) and melting point – pressure (depth) curves. This 'temperature method' has been suggested by Weertman (1968, 1970), based on empirical relationships of Sherby and Simnad (1961) and Shewmon (1963). It was applied to the mantles of the Earth and other planetary bodies by Weertman and to the Earth's crust by Meissner (1974).

The third method makes use of relations between seismic Q-values (elasticity) and viscosity assuming that both effects are caused by volume diffusion of vacancies. (McConnel, 1965; Anderson and O'Connel, 1967). This 'elasticity-method' seems to

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be the least reliable one, but may be used to check the values obtained by other viscosity estimations.

Although the mechanism of viscous flow and creep laws is hotly debated, there is a common agreement among Earth scientists, that viscosity distribution plays a major role for understanding the internal and external structure of planetary bodies. Gravity, topographical and global anomalies will decrease with time in a characteristic way depending on the viscosity-depth structure and dynamics. So, the Moon, Mercury, and parts of Mars with their prominent surface structures and gravity anomalies will be much more viscous, at least at shallow depths, than the Earth (and Venus).

After dealing shortly with methodical problems in obtaining viscosity values in general, a comparison will be made between the average lunar viscosity values of Arkani-Hamed and the values obtained by the 'temperature-method'.

2. Problems of Determining Viscosity Values Due to Different Creep Mechanisms

Weertman's 'effective viscosity' $\tilde{\eta}$ (Weertman, 1970) seems to be an efficient tool for obtaining viscosity values from temperature – depth curves $T(z)$ ($T(z)$ = geotherms or selenotherms) and melting point – depth curves $T_m(z)$. It is defined by

$$\tilde{\eta} = \sigma/\dot{\varepsilon}_{\mathrm{const}}, \tag{1}$$

σ = stress in cgs units,
$\dot{\varepsilon}$ = strain rate of $10^{-16}\ \mathrm{s}^{-1}$,
$\dot{\varepsilon} = f(\sigma, T, T_m)$,
T = temperature in K,
T_m = melting point in K.

In this formula $\dot{\varepsilon}$ is kept constant, for instance $\dot{\varepsilon} = 10^{-16}\ \mathrm{s}^{-1}$, by varying temperature and pressure according to the creep law applied (for instance an increasing T would cause a decreasing σ). Basically two creep laws have to be taken into account. One is the Nabarro-Herring Creep Law (Nabarro, 1948; Herring, 1950). Here, creep is produced by diffusional mass transport of lattice vacancies from one grain boundary to another as represented by a Newtonian body, in accordance with the equation

$$\dot{\varepsilon} = \frac{\alpha\Omega\sigma}{KL^2}\frac{D_0}{T}\exp(-gT_m/T) \approx C\sigma\exp(-gT_m/T), \tag{2a}$$

where

α = constant (≈ 5),
Ω = atomic volume,
K = Boltzmann's constant,
L = average grain diameter,
D_0 = constant ($10^{-1} < D_0 < 10^{+1}$),
T = temperature in K,
T_m = melting point in K,

g = constant ($13<g<25$; Weertman: $g=18$),
C = quasi constant.

Creep is proportional to the applied stress σ.

A second creep law which seems to be gaining more support recently, (Turcotte and Oxburgh, 1969; Weertman, 1967, 1970; Post and Griggs, 1973; Kohlstedt and Goetze, 1974) is based on theoretical and experimental evidence (Post, 1973). Here, creep is related to stress by a power-law like

$$\dot{\varepsilon} = \frac{\alpha^* \Omega \sigma^3}{K\mu^2} \frac{D_0}{T} \exp(-gT_m/T) \approx C^* \sigma^3 \exp(-gT_m/T), \tag{2b}$$

where

α^* = constant ($\approx 2.26 \times 10^{12}$ cm^{-2}; Weertman),
μ = shear modul ($\approx 10^{12}$ dyn cm^{-2}),
C^* = quasi constant,
other symbols: see (2a).

It seems probable (Weertman, 1970) that (2a) applies to small stresses, while (2b) may be valid for higher stresses, so both laws must be taken into account when determining $\dot{\varepsilon}$ and $\tilde{\eta}$ from temperature values. Although the coefficients in (2b) differ by about 200 for Post (1973) and Weertman (1970) it seems well justified from many investigations of glacial rebound to assume $\eta \approx 10^{21}$ P for the low velocity layer of the Earth's mantle, a value which would apply to a stress of about 0.1 bar and $\dot{\varepsilon} = 10^{-16}$ s^{-1} (Weertman, 1970) and to a stress of about 20 bar and $\dot{\varepsilon} = 5 \times 10^{-13}$ s^{-1} (Post, 1973). This 'fixpoint' for η determines for practical use the 'quasi constant' C in (2a) as well as C^* in (2b), so that by means of (1), (2a) and (2b) one gets

$$\ln \tilde{\eta}_{\mathrm{NH}} = -\ln C + gT_m/T, \qquad \text{(Nabarro-Herring creep)} \tag{3a}$$

$$\ln \tilde{\eta}_{\mathrm{W}} = -\ln K^* + \tfrac{1}{3} g T_m/T, \qquad \text{with} \quad \ln K^* = \ln C^* + \tfrac{2}{3} \ln(\dot{\varepsilon}/C^*) \tag{3b}$$

(subgrain $\approx$ dislocation glide creep)
(Weertman).

By use of (3a) and (3b) respectively, two extreme sets of $\tilde{\eta}$-values may be obtained from selenotherms (geotherms) $T(z)$ and melting points $T_m(z)$. The first set (3a) applying for small stress (possibly $\sigma < 10^{-2} \ldots 10$ bar) leads to higher viscosities than using the second set (3b). The gradient $(\ln \eta)/(T_m/T)$ is for the second set only one third of that of the first one. As the region where the transition from one creep law to the other takes place seems to be related to some unknown stress and crystal size, the range between $\tilde{\eta}_{\max} = \tilde{\eta}_{\mathrm{NH}}$ and $\tilde{\eta}_{\min} = \tilde{\eta}_{\mathrm{W}}$ may be considered as one of possible viscosity values.

3. Temperatures and Related Viscosity Distribution

Geotherms and selenotherms may be calculated on the basis of heat flow values, thermal conductivity, heat production and initial temperature. For the Moon many calculations of present and ancient selenotherms have been performed (Fricker *et al.*, 1967; Sonett *et al.*, 1971; Duba *et al.*, 1972; Dyal *et al.*, 1972; Wood, 1972; Toksőz

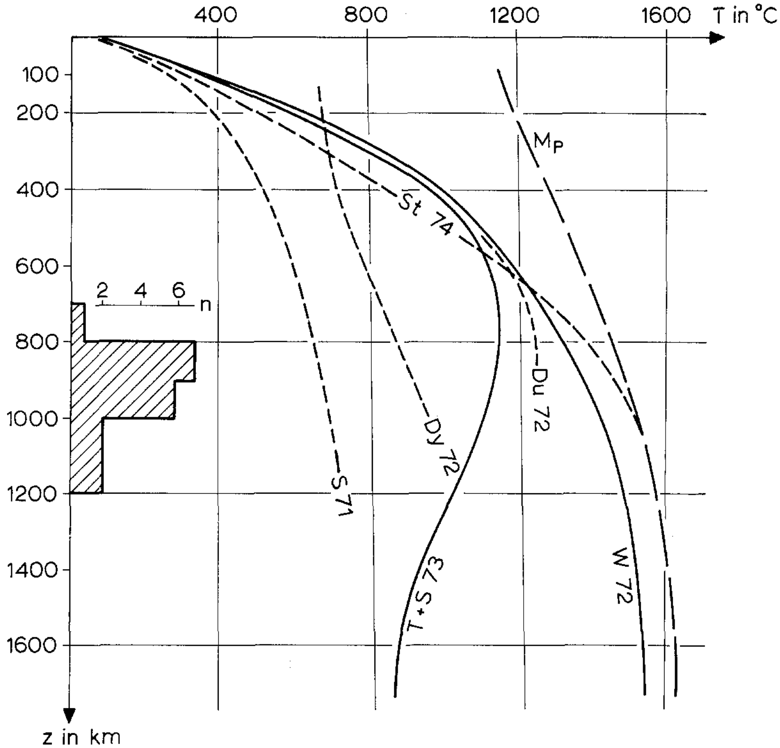

Fig. 1. Selenotherms with melting point – depth function.

S 71	= Sonett *et al.* (1971),
Du 72	= Duba *et al.* (1972),
Dy 72	= Dyal *et al.* (1972),
T + S 73	= Toksőz and Solomon (1973),
St 74	= Strangway and Sharpe (1974),
W 72	= Wood (1972),
M_P	= melting point curve,
n	= number of moonquakes (Latham *et al.*, 1973b).

and Solomon, 1973; Duba and Ringwood, 1973; Strangway and Sharpe, 1974). Some of the recently presented selenotherms, based mainly on magnetotelluric methods and heat flow data of Apollo 15 and 17, are shown in Figure 1. From these widely differing curves, those which are consistent with the heat flow values have been selected as the more reliable ones. In fact, on the basis of the heat flow data no large deviation of the selenotherms is possible for the outer part of the Moon. Two of the selenotherms were plotted in a temperature-pressure diagram together with some terrestrial curves. See Figure 2. As the pressure increases much more rapidly in the Earth compared with the Moon, very different terrestrial and lunar temperature-depth curves may show about the same temperature-pressure relation. It seems interesting, too, that the moonquakes (Latham *et al.*, 1973a) prefer about the same temperature-pressure regime than do small earthquakes near plume and rift zones.

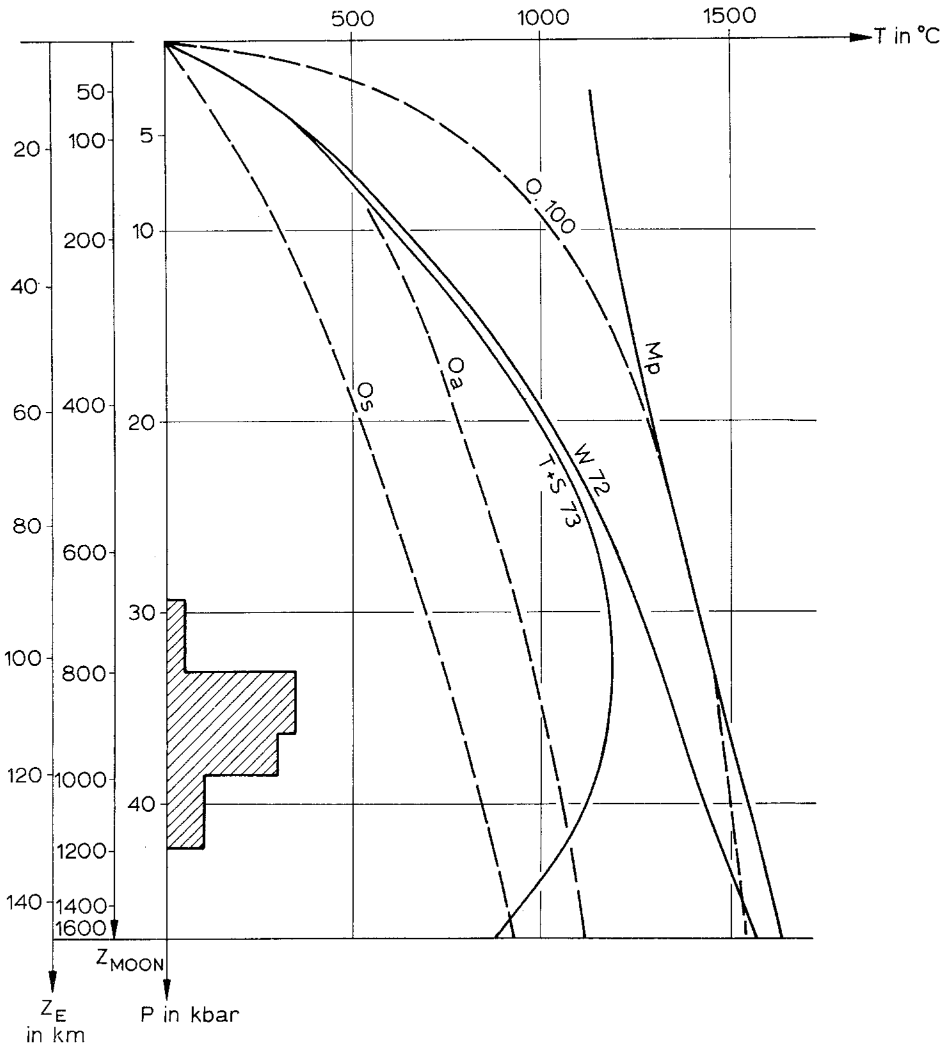

Fig. 2. Geotherms, selenotherms, and melting point-curve vs pressure.

O.s. = oceanic plate, subducing,
O.a. = oceanic area, average,
O. 100 = oceanic area, 100 km from rift axis,
W 72 = selenotherm after Wood (1972),
T + S 73 = selenotherm after Toksőz and Solomon (1973),
Mp = melting point curve of dry peridotite-pyroxenite,
//// = zone of moonquakes.

Temperature curves of Figures 1 and 2 were further used for the calculation of T_m/T where $T_m(z)$ has been taken from pressure dependent melting point curves of dry pyroxenite-peridotite. Figure 3 shows T_m/T and possible viscosity values versus depth for the selenotherms by Wood (1972), Toksőz and Solomon (1973), and Strangway and Sharpe (1974). It may be seen that due to the relatively high heat flow

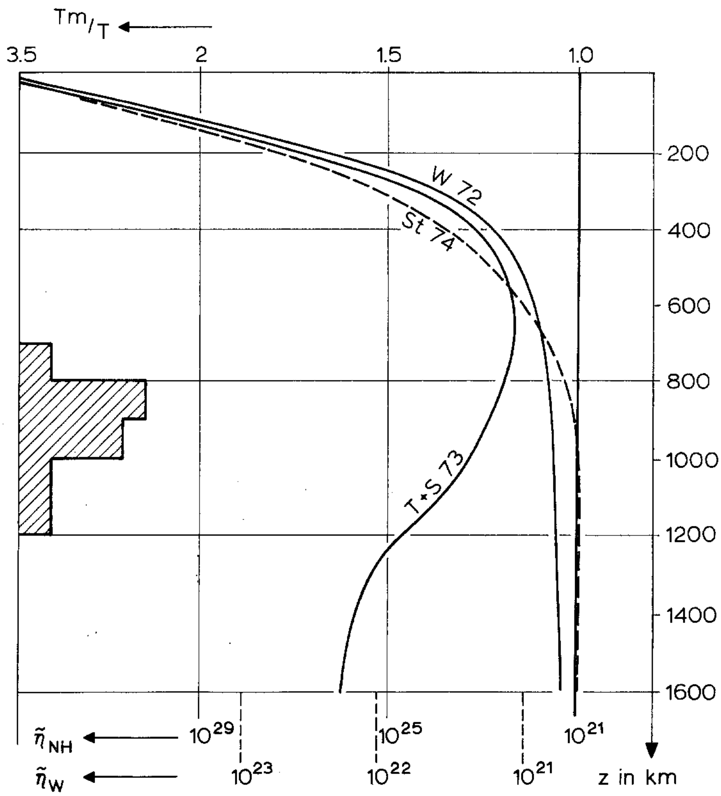

Fig. 3. T_m/T vs depth, estimation of viscosity-depth relation.

$\tilde{\eta}_{NH}$ = viscosity values calculated assuming Nabarro-Herring creep,
$\tilde{\eta}_W$ = viscosity values calculated assuming Weertman's dislocation glide,
W 72 = viscosities as derived after selenotherms from Wood (1972),
T + S 73 = viscosities as derived from selenotherms by Toksőz and Solomon (1973),
St 74 = viscosities as derived from selenotherms by Strangway and Sharpe (1974).

values on the surface, around 30 erg cm^{-2} s^{-1}, and the K and U content of basalts and anorthosites the gradient of temperature and viscosity is rather steep for the upper 400 km of the Moon. T_m/T-values for the upper 400 km are between 1.8 and 1.9. The two viscosity values are 10^{27} P–10^{28} P for Nabarro-Herring creep and 0.8–1 $\times$ 10^{23} *P* for Weertman's dislocation glide. It seems at first glance problematic to decide which of the two sets of viscosity values should be preferred. On Earth, isostatic rebound processes like the uplift of the Fennoscandian and Canadian shields are combined with small stresses, so that even Weertman (1970) assumes a Nabarro-Herring creep, while Post and Griggs (1973) favour Weertman's dislocation glide. Relating to stress differences as calculated by Arkani-Hamed (1973c), maximum stress values up to 170 bar may be caused by the mascons with their density anomalies in the upper part of the Moon, so that dislocation glide (with low viscosities) would apply at least to the mascon boundaries. On the other hand, the bulk of the isostatic sinking of the mascons by the load effect of the mare basalts will certainly be accompanied by small stresses and Nabarro-Herring creep with $\eta \approx 10^{27}$–10^{28} P as an average for the uppermost 400 km depth. Calculations of realistic stresses of such a rebound process using the finite element method are underway.

From the present topography and the large gravity anomalies of the mascons, it is clear that the Moon's viscosity must be much greater in its outer part than that of the Earth. Figure 4 shows a comparison of the upper 150 km depth of the Moon and the Earth. Terrestrial values are from Meissner (1974), and show some 'jumps' in viscosity at chemical boundaries because of the smooth geotherms and material-dependent melting points. From Figure 4 it becomes clear that the outer part of the Moon is definitely much more viscous than that of the Earth, even that of the 'stable' shield areas. The upper 100 to 150 km of the Moon with viscosity values $\eta > 10^{28}$ P will certainly be able to sustain large stresses, although specific calculations have still to be carried out. Please note that viscosity values of $\eta < 10^{21}$ should be considered with caution because of possible strong convection effects.

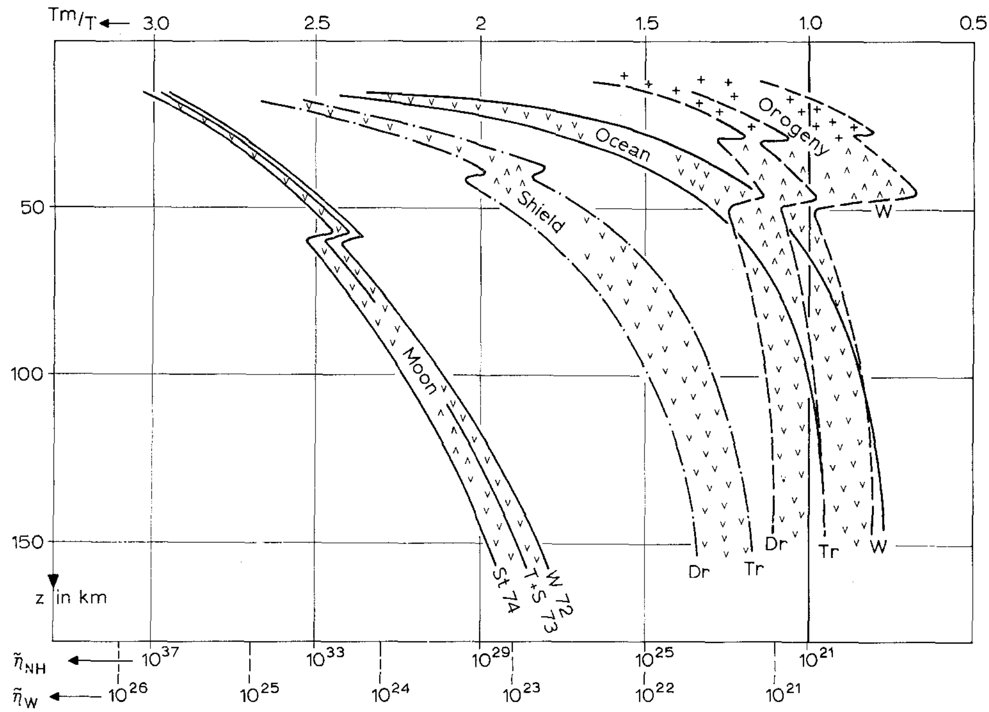

Fig. 4. Viscosity-depth relation for uppermost part of Moon and Earth.

W 72 = viscosities as derived from selenotherms by Wood (1972),
T + S 73 = viscosities as derived from selenotherms by Toksöz and Solomon (1973)
St 74 = viscosities as derived from selenotherms by Strangway and Sharpe (1974),
Dr = values based on dry solidus melting curve,
Tr = values based on melting curve with traces of water,
W = values based on wet solidus curve.

Viscosity values larger than 10^{27} P in the upper 200 km of the Moon do not preclude convective currents below a depth of about 500 km, where viscosities definitely are lower than 10^{23}–10^{24} P. In fact, a limited convection or at least a flow of matter should certainly be expected from the activity of small moonquakes (Meissner *et al.*, 1970, 1973; Latham *et al.*, 1973a, b; Lammlein *et al.*, 1974). Therefore, this part of the conclusions derived from the viscosity considerations are in contrast to those of Arkani-Hamed (1973c). His calculations of viscosity on the basis of the possible sinking of the heavy mascons give (Arkani-Hamed, 1973a, b) a present-day average figure of about 10^{27} P for the whole Moon, which may be compared to the same values derived above as an average of the upper 400 km.

4. Estimations About the Viscous History

Conclusions about the thermal and viscous history of the Moon are necessarily very vague. Using temperature-depth curves of Wood (1972), Toksőz and Solomon (1973),

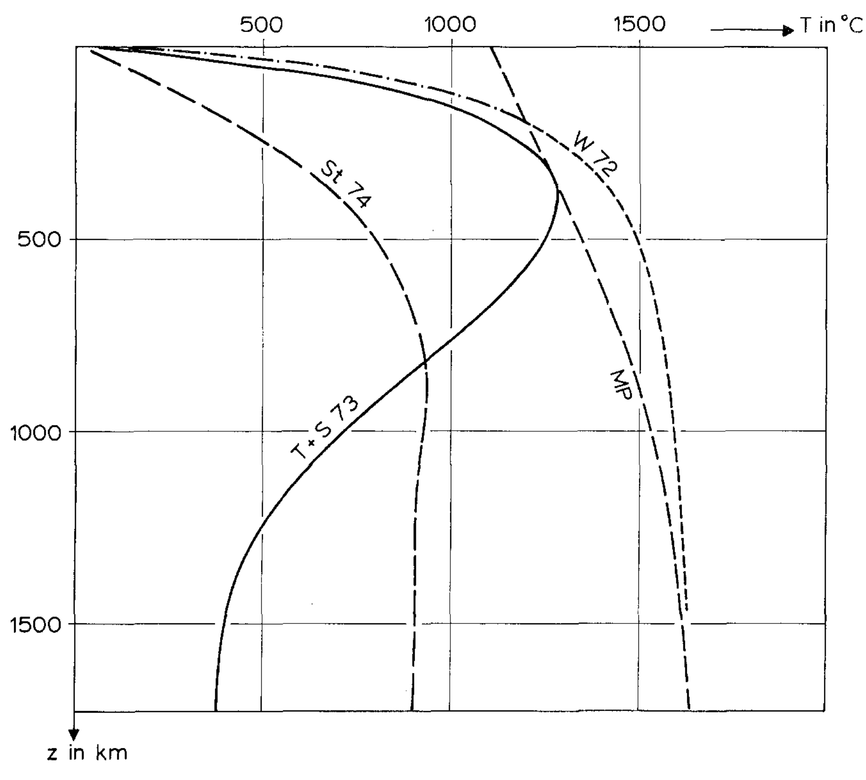

Fig. 5. Selenotherms at 3.1×10^9 yr.

W 72 = curves from Wood (1972),
T + S 73 = curves from Toksőz and Solomon (1973),
St 74 = curves from Strangway and Sharpe (1974),
*M*P = melting point curve.

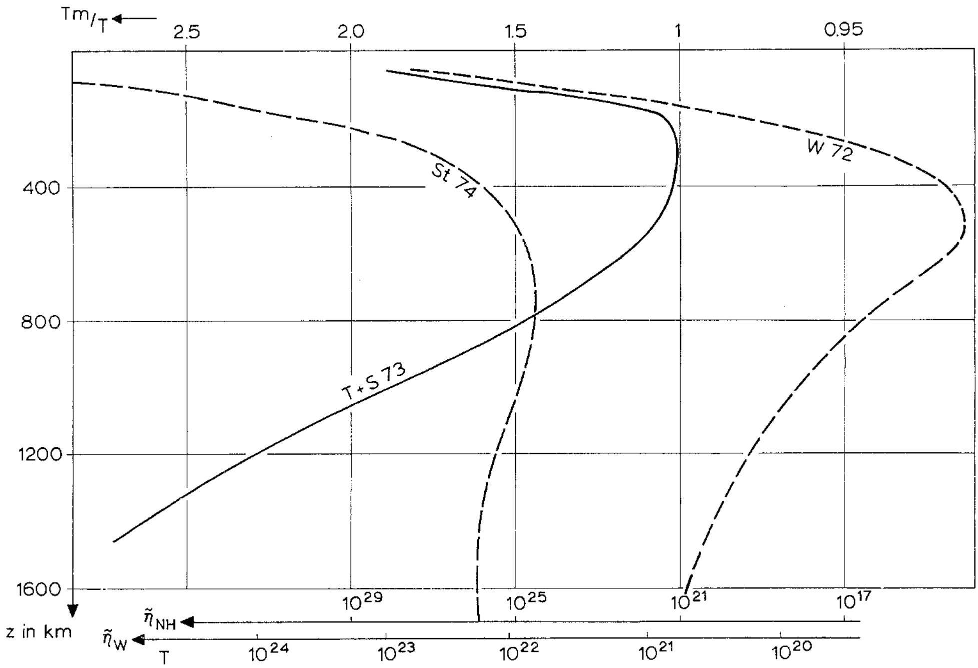

Fig. 6. T_m/T vs depth; estimation of viscosity-depth relations 3.1×10^9 yr ago.

η_{NH} = viscosity values calculated assuming Nabarro-Herring creep,
η_W = viscosity values calculated assuming Weertman's dislocation glide,
W 72 = viscosities as derived after selenotherms from Wood (1972),
T + S 73 = viscosities as derived after selenotherms from Toksőz and Solomon (1973),
St 74 = viscosities as derived after selenotherms from Strangway and Sharpe (1974).

and Strangway and Sharpe (1974) calculated for 3.6 and 2.6×10^9 yr, and interpolating them for the time of the end of the basin filling with lava (3.1×10^9 yr, Figure 5) possible viscosity-depth values for this time period may be obtained (see Figure 6). If we consider the fact that the Moon's surface at that time had been rigid and viscous already since about 1.4×10^9 yr, it seems that the giant impacts excavating the mare basins had several effects:

(i) providing additional heat,

(ii) causing additional cracks at least down to depths comparable to the crater diameter (some 100 km), and

(iii) causing strong stresses and so favouring a dislocation glide mechanism with reduced viscosities at least in the time period of isostatic uplift of the excavated basin and the subsequent filling and sinking of this area.

So, the areas of the circular maria may well have been regions of strongly reduced viscosities, and also adjacent areas like the Oceanus Procellarum or the Mare Frigoris may have been effected. Considering these areas as thermal and viscous anomalies, the viscosities may here have been near to those obtained from the selenotherms of Toksőz and Solomon. In the remaining old terra regions and the bulk of the Moon with its rigid and viscous behaviour, Strangway and Sharpe's selenotherm seems to be more appropriate.

On the other hand, following Strangway and Sharpe's reasoning, the present-day temperature curve would meet the melting curve at about 1000 km depth, so that the appearance of moonquakes down to 1200 km in a nearly completely melted inner Moon seems problematic. Several aspects of moonquakes, for instance the perigee appearance of A_1-type quakes (Meissner *et al.*, 1970, 1973) indicate that stress in the hypocenter regions can be stored for about a month. This may be possible in a visco-elastic, but not in a nearly completely melted region. It must be concluded, therefore, that the present day temperature and the thermal development will be somewhere in between the presented curves of Strangway and Sharpe (1974), and those of Toksőz and Solomon (1973), with a slight preference for the former ones. It seems important that both curves were calculated on the basis of an initially cold Moon.

5. Relation Between η and Q Values

Anderson (1966) and Anderson and O'Connel (1967) suspect that viscosity η (or creep rate $\dot{\varepsilon}$) and seismic inelasticity Q^{-1} may be related to each other, as both are defect controlled and, hence, should be expected to be similar functions of depth. Both η and Q^{-1} should really be most probably exponential functions of temperature and also be pressure dependent to a minor degree by their connection to grain- or subgrain size effects. Anderson and O'Connel (1967) suspect a relation such as

$$\eta/Q = \text{const} \approx 4 \times 10^{19}\ \text{P}. \tag{4}$$

On the Moon Q-values obtained from the Apollo seismograms are definitely very high (Latham *et al.*, 1970, 1973a, b). Values of 2000 and 5000 have been observed from the amplitude behaviour of moonquakes as well as from artificial impacts. It seems interesting to compare these high Q-values, which are representative of the upper mantle of the Moon between about 50 and 100 km depth with the high viscosities of the Moon in this depth zone. Figure 7 shows the two values of $\tilde{\eta}$ compared to Q-values for the uppermost mantle of the Moon and some ones of two regions of the Earth as calculated from Sutton *et al.* (1967). Q-values obtained from shear waves and surface waves for the deeper part of the Earth's mantle have not been taken for this comparison, as below about 400 km depth, by the phase transition phenomena, possibly different mechanisms for Q and η might apply. From Figure 7 may be seen that Anderson and O'Connel's curve doesn't fit the observed data from the upper mantles. Instead, a much steeper curve, such as $\Delta\eta/\Delta Q \approx 3 \times 10^{28}$ should be used, supposing Nabarro-Herring creep. The low velocity or Gutenberg zone in the Earth's upper mantle with $\eta \approx 10^{21}$ P then should have a Q-value of 150–200, regardless what set of viscosity values is taken. Similar values should dominate in the lower part of crusts in orogenic regions and in the inner part of the Moon if the temperature approaches the melting point. So, the high Q-values including the long 'ring' of the

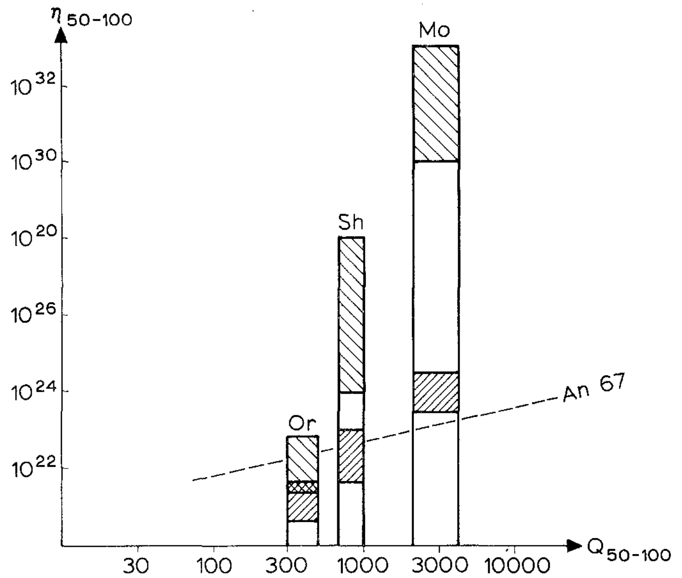

Fig. 7. Possible relations between viscosity and anelasticity Q^{-1} in the uppermost mantle.

Or = orogeny,
Sh = shield area,
Mo = Moon,
An = Anderson's Q-relation,
Q = values from Sutton *et al.* (1967),
η = values from Meissner (1974),
\\\\ = Nabarro-Herring $\sigma - \eta$ relation (3a),
//// = Weertman $\sigma - \eta$ relation (3b).

Moon after an impact or a quake may be easily understood when considering the high η-values in the upper mantle. On the other hand, a relation between η and Q, even for similar material and temperature, has to be investigated in much more detail in order to use one of the variables to determine the other.

6. Conclusions

(1) Two sets of viscosity values according to two creep laws seem to exist on the Earth and on the Moon, one set for low stress and small grain size, leading to high viscosity values (Nabarro-Herring creep law with $\dot{\varepsilon} \propto \sigma$) and another one for higher stresses resulting in lower viscosities (diffusion glide or subgrain creep laws with $\dot{\varepsilon} \propto \sigma^3$; Weertman (1970)). The region where the transition of one creep mechanism to the other takes place is probably between 10^{-2} and 10 bar. Some indication for Nabarro-Herring creep mechanism comes from high viscosity values used by a number of scientists in order to explain the excess equatorial bulge of the Earth. Diffusion glide would possibly give too low viscosity values for the lower Earth's mantle. For the Moon, Nabarro-Herring creep seems to explain far better the necessary high viscosity values in the upper 200–400 km. If we assume lower viscosity

values, it is difficult to explain the present day existence of the large mascons which have survived over 3×10^9 yr. On the other hand, considering the small confining pressure in the upper 200–400 km (see Figure 2), here some kind of stickslip mechanism may also be predominant, generating at times a shallow moonquake releasing the accumulated stress.

Indications for diffusion glide creep mechanism, based on good experimental evidence, is found at terrestrial plate boundaries. This kind of creep may play a major role at all places where topography and/or density differences and dynamics lead to large stresses. So it is suggested that mascon boundaries may well reduce viscosity values at some depth by causing a transition to diffusion glide creep processes. This would explain why the latest lava flows seem to have prefered the boundaries of the circular maria. Also the appearance of transient events preferentially in these areas (Middlehurst, 1967) may be better understood.

(2) Viscosities in the Moon's outer part are larger than in any comparable region on Earth. Previously calculated average viscosity values (Arkani-Hamed, 1973a, b) obtained on the basis of relaxation studies of mascons could be verified for the outer 400 km of the Moon. However, the viscosity-depth curve deduced here by the temperature-method results in much lower viscosity values below 400 km. Thus convection and a limited flow of matter may be possible inside the Moon below a depth of about 500 km. The mechanism of moonquakes at depths of around 1000 km seems to be similar to processes near terrestrial plume and rift zones as viscosities, temperatures, and pressures are similar.

(3) The high Q-values for the Moon's uppermost mantle may easily be understood by considering the high viscosities in this region. For similar material and for a pressure smaller than that necessary for a phase transition there seems to be a relation between Q and η.

(4) Introducing viscosity as an additional physical parameter some selection of selenotherms can be performed and indications of the thermal and viscous history of the Moon may be gained. Viscosity constraints must apply for the present time period as well as for the period of lava flows and the origin of mascons in areas of reduced visscoity.

Note added in proof. Recalculation of hypocenters of moonquakes (Nakamura *et al.*, 1974) have revealed maximum depths of about 800 km for tide-generated quakes. These reduced depths further emphasize the conclusions drawn from temperature curves of Figures 1 and 2 and the viscosity curves of Figure 3 that quakes seem to prefer the region where temperatures are slightly below melting point and viscosities must not yet have reached their minimum value. Correlations between the pressure-temperature regime of these quakes and those occurring below terrestrial oceanic ridges are improved.

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