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Relationship Between the Seismic Quality Factor Q and the Effective Viscosity η*

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Abstract. Two very similar relationships are developed for the effective viscosity η and the seismic quality factor Q with regard to the ratio of the solidus temperature T_m to temperature T. This relationship is linear both for $\log Q$ and $\log \eta$ and seems to be valid for $0.95 > T/T_m > 0.5$. Within this temperature interval two different creep laws dominate; also two different attenuation mechanisms are observed from a compilation of seismic body wave data and T_m/T values. The transition from the linear Nabarro-Herring creep to power law creep and that from one attenuation mechanism to the other both take place around 0.7 to 0.8 T/T_m. For the above mentioned temperature regime, which covers the lower lithosphere and the asthenosphere, and for body wave frequencies around 1 to 10 Hz the following linear relationship between $\ln \eta$ and $\ln Q$ is tentatively established:

$$\ln \eta = 4.4 \ln Q + 22 \quad \text{(for } \eta \text{ given in poise).}$$

Activation energies for attenuation are only about 23% of those for creep.

Key words: Seismic quality factor – Attenuation – Effective viscosity – Activation energies.

1. Introduction

The seismic quality factor Q and the effective viscosity η are governed by different physical processes. At first glance there seems to be little hope of finding a relationship between them. Scattering effects dominate in the near surface area, especially in sediments, and hence Q is strongly frequency dependent, increasing for decreasing frequencies (McDonal et al., 1958; Lütjen, 1978). Although geometrical scattering is not related to the anelastic properties of the medium but rather to certain boundary conditions, its effect cannot be removed from seismic attenuation measurements. The effective viscosity in near surface areas down to the middle or lower crust is also influenced by boundary

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conditions: An especially low viscosity associated with fault zones determines the viscosity values in all areas where faults are present (Meissner, 1978; Vetter and Meissner, 1979). Effective fault zone viscosities as low as 10^{14} poise have been calculated for the upper few hundred meters of the crust whereas Q-values between 5 and 25 are found in young sediments at shallow depths (Meissner, 1965). It seems, however, that Q-values in the upper crust are connected with microcracks and are more a bulk property of the whole rock assemblage than are the η-values, but information on this subject is scarce. Estimates of η are derived from temperature models (Meissner and Vetter, 1976) and from the uplift of salt domes and diapirs (Hunsche, 1977). Calculations of Q are based on measurements of the attenuation of body waves (Sutton et al., 1967; Barazangi and Isacks, 1971) and surface waves (Hart et al., 1967; Solomon et al., 1972).

Calculation of η for the whole lithosphere and the asthenosphere can be obtained using the temperature method, postglacial uplift data (Walcott, 1973; Crittenden, 1967; Brennen, 1974), and the movement of plates (Meissner and Vetter, 1976), while amplitude decay spectra of free oscillations in addition to the surface wave and body wave data are used for an estimation of Q in the whole earth (Sailor and Dziewonski, 1978). Figure 1 shows a summary of η- and Q-values versus depth with a smooth curve between data of the authors mentioned above. This figure provides information only on the general behavior of η and Q and certainly can not be used for detailed considerations. It is important to note, however, that Q and η both increase in the upper crust and decrease down to the asthenosphere.

The present study is mainly concerned with the relation between η and Q in the lower lithosphere and asthenosphere with some extrapolations to the middle and lower mantle. The area under consideration covers mainly the temperature range between about $0.95\,T_m > T > 0.5\,T_m$ (with T_m = solidus temperature), though some extrapolations to lower and higher T-values will be made. Also the considered frequency range Δf is small; most body wave data cover the range between 1 and 10 Hz. From theoretical studies (Jackson and Anderson, 1970; Anderson and Hart, 1978; Goetze, 1977) it seems that Q in this $\Delta T - \Delta f$ field is dominated by a grain-boundary relaxation mechanism with an exponential relation to T_m/T. A similar exponential relation has been established for η (Weertman, 1970; Kohlstedt and Goetze, 1974; Kohlstedt et al., 1976; Mercier et al., 1977; among others). Physically, steady state creep may consist of grain boundary diffusion (Coble, 1963) or of a dislocation glide mechanism (Weertman, 1970) – the so called power law creep. The possible relevance of both creep laws for conditions in the lithosphere and asthenosphere has been investigated by Weertman and Weertman (1975), Meissner and Vetter (1976), and Vetter and Meissner (1977). It seems that the likelihood that a close relation between Q and η might exist is greatest in the above mentioned high temperature field where grain boundary processes dominate.

2. Derivation of Formulas for η and Q in the High Temperature Field

As shown by Weertman (1970), Kohlstedt et al. (1976) and others, creep properties of igneous rocks are governed by the general creep equation

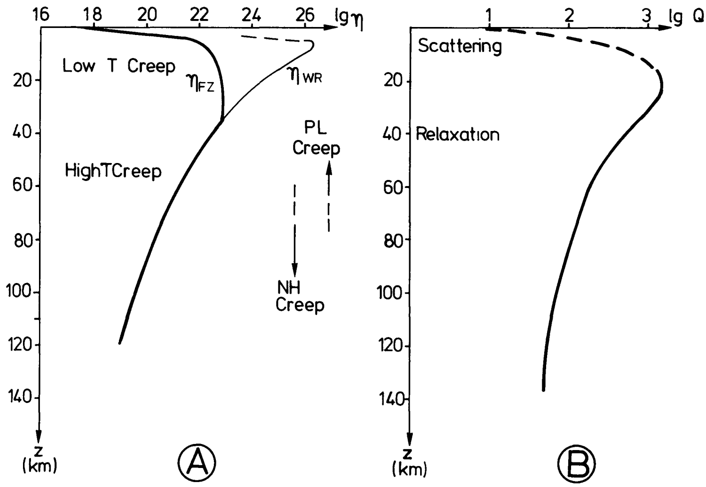

Fig. 1. A General behavior of η versus depth; **B** General behavior of Q versus depth in the lithosphere/asthenosphere after various authors (see text). η_{FZ}: effective fracture zone viscosity. η_{WR}: effective whole rock viscosity. PL: Power creep law and NH: Nabarro-Herring creep law = high T creep. Q-data from body wave investigations

$$\dot{\varepsilon} = C_n \sigma^n \exp - [(U + pV)/RT] \tag{1}$$

with

C_n = creep constant $\dot{\varepsilon}$ = creep rate
σ = stress p = pressure
n = exponent of the creep law (1...3)
U = activation energy V = activation volume
R = gas constant T = absolute temperature.

Weertman (1970) has shown that a good approximation of (1) is

$$\dot{\varepsilon} = C_n \sigma^n \exp(-g^* T_m/T) \tag{2}$$

with g^* = constant [about 29 for olivine after Kohlstedt and Goetze (1974)]

T_m = melting temperature (solidus).

Introducing the effective viscosity

$$\eta \approx \sigma/\dot{\varepsilon} \tag{3}$$

and inserting (2) into (3) one obtains

$$\ln \eta = g^* T_m/T - \ln C_n - \ln \sigma^{n-1}. \tag{4}$$

(η always in poise).

If mass transport takes place by general diffusional processes (Nabarro-Herring creep) or along grain boundaries by means of grain-boundary diffusion

(Coble creep) the exponent n will be equal to 1 or at least near 1 (Langdon, 1970). Hence, (4) may be written as,

$$\ln \eta_{NH} = g^* T_m/T - \ln C_1 \quad \text{for } n=1 \tag{5}$$

i.e., η_{NH} does not depend on stress or creep rate. Dislocation glide mechanisms, as thoroughly investigated by Weertman (1970), Kohlstedt and Goetze (1974), and Carter (1976), may also be derived from (4) using $n=3$. For $\dot{\varepsilon}=$const. one obtains after inserting $\sigma \approx \eta \dot{\varepsilon}$

$$\ln \eta_{PL} = 1/3\, g^* T_m/T - 1/3 \ln C_3 - 2/3 \ln \dot{\varepsilon}. \tag{6}$$

In the following it will be shown that formulas like (4), (5), and (6) can also be obtained by assuming a similar grainboundary relaxation process as postulated by Anderson and Hart (1978) and O'Connell and Budiansky (1978). The relaxation time for the inelastic relaxation after a small displacement is

$$\tau = \tau_0 \exp[(\tilde{U} + p\tilde{V})/RT] \tag{7}$$

with $\tilde{U}$ and $\tilde{V}$=activation energy and volume respectively for small scale displacements – as, for instance, caused by the passing of a seismic wave.

τ_0=characteristic time which according to Anderson and Hart (1978) should be related to the atomic jump frequency. $\tilde{U}$ and $\tilde{V}$ are certainly not identical with U and V of Eq. (1) and may be considerably smaller. From (7), a similar approximation as that of (2) gives

$$\tau = \tau_0 \exp(g_Q T_m/T). \quad \text{For } g_Q \text{ see (14).} \tag{8}$$

The introduction of the general Debye equation for Q gives

$$Q = (c_0/\Delta c) \cdot (1 + \omega^2 \tau^2)/\omega\tau \tag{9}$$

with

c_0=low frequency elastic wave velocity
Δc=difference between low and high frequency wave velocity
ω=angular frequency.

We apply a high frequency approximation and obtain

$$Q = (c_0/\Delta c)\,\omega\tau \quad \text{for } \omega^2\tau^2 \gg 1. \tag{10}$$

As τ appears to be in the range of 10 to 100 s (Anderson and Hart, 1978) the approximation of (10) seems to be valid for frequencies larger than 0.1 Hz, i.e., certainly for body wave data from explosions and small earthquakes.

Inserting (8) into (10) one obtains

$$\ln Q = g_Q(T_m/T) + \ln(c_0/\Delta c) + \ln(\omega\tau_0). \tag{11}$$

This equation is of the same form as (4), (5), and (6). It describes the strong dependence of Q on the ratio of T_m/T which is similar to that of η. Equation (11) also contains a direct relationship between Q and ω which generally appears in theoretical derivations (Knopoff, 1964) as well as in some experimental studies (Goetze, 1977). This relationship is hard to understand in view of the large amount of seismic field observations which are consistent with a frequency independent Q (Anderson and Hart, 1978). Recent experimental work of Berckhemer and his coworkers (personal communication) shows (as the earlier experiments of Gordon and Davis (1968)) that Q at high temperature may be independent of frequency over at least 2 decades. We suggest an inverse relationship between ω and τ_0 and will mention this problem later again.

In order to compare both η and Q with values of T_m/T, as needed for a later calculation of activation energies, we define the gradient g_v of the logarithm of η as a partial derivative from (5), (6), and (11):

$$g_1 = \frac{\partial(\ln \eta)}{\partial(T_m/T)} = g^* \tag{12}$$

for NH creep and PL creep with constant σ

$$g_2 = \frac{\partial(\ln \eta)}{\partial(T_m/T)} = 1/3\, g^* \tag{13}$$

for PL creep with constant $\dot{\varepsilon}$ and with $n=3$

$$g_3 = \frac{\partial(\ln Q)}{\partial(T_m/T)} = g_Q \tag{14}$$

with g_Q between 0.07 and 0.23 g^* as will be shown in the next section.

In general, the Eqs. (5), (6), and (11) can be used to derive a relation between η and Q. Solving (11) for T_m/T and inserting T_m/T into (5) and (6), gives

$$\ln \eta = (g^*/g_Q) \ln Q - (g^*/g_Q) \ln(c_0/\Delta c) - (g^*/g_Q) \ln(\omega \tau_0) - \ln C_1 \tag{15}$$

for NH creep and

$$\ln \eta = 1/3 (g^*/g_Q) \ln Q - 1/3 (g^*/g_Q) \ln(c_0/\Delta c) - 1/3 (g^*/g_Q) \ln(\omega \tau_0) - 1/3 \ln C_3 - 2/3 \ln \dot{\varepsilon} \quad (\text{for } \eta \text{ in poise}) \tag{16}$$

for Pl creep with $n=3$. While g^* is rather well known from experiments of steady state creep processes and seems to be in agreement with in situ observations, g_Q has not yet been determined.

3. Comparisons Between Q and η From Field Observations

Figure 2A gives some well known relations between η and T_m/T resp. T/T_m. Effective viscosities were calculated on the basis of the rheological constants of olivine (Kohlstedt and Goetze, 1974), which is supposed to be the main constituent of mantle material. T- and T_m-values of an oceanic mantle have been used for this figure as in Vetter (1978). These values are in good agreement with

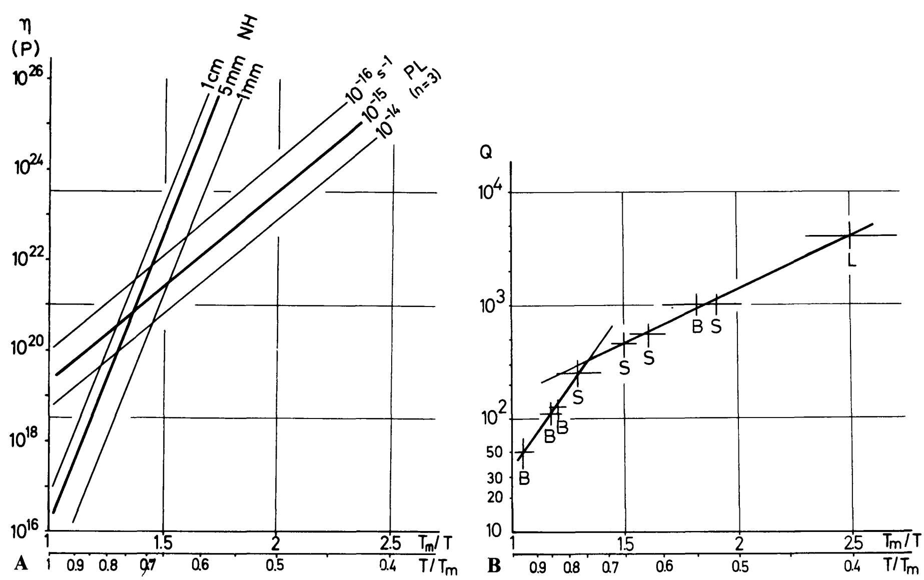

Fig. 2. A Effective viscosities according to Nabarro-Herring (*NH*) and power law (*PL*) creep for different grain sizes and creep rates as a function of the ratio of solidus (T_m) to temperature (T). For global relations a grain size of 5 mm and a mean creep rate of $\dot{\varepsilon}=10^{-15}\,s^{-1}$ is used. [For details see Vetter and Meissner (1978)]. **B** Q-values as a function of T_m/T resp. T/T_m. *B*: Barazangi and Isacks (1971); *S*: Sutton et al. (1967); *L*: Latham et al. (1970) (size of crosses-measure for the variance). T_m = dry pyrolite solidus (after Stocker and Ashby, 1975)

temperature models as well as viscosity values from other methods such as uplift data and plate movements. The creep process which operates with the lowest effective viscosity will always dominate about the other one. An average grain size of about 5 mm and a viscosity of 10^{21} poise were found to prevail in the continental shield asthenosphere [for details see Meissner and Vetter (1976), and Vetter and Meissner (1977)]. The transition from *NH* to *PL* creep occurs at about 1 bar and at T/T_m between 0.7 and 0.8, depending on grain size and creep rate $\dot{\varepsilon}$. We presume, based on our former work, that the power law creep in the upper part of a fast moving oceanic asthenosphere may follow the curve of $\dot{\varepsilon}=10^{-14}\,s^{-1}$ which grades into the lithosphere with $\dot{\varepsilon}\approx 10^{-15}$ and $10^{-16}\,s^{-1}$. Accordingly, the viscosity-values for lower T/T_m have to be taken from those $\dot{\varepsilon}$-curves if a special area is considered. On the other hand, an average of $\dot{\varepsilon}=10^{-15}\,s^{-1}$ will be used for obtaining an average η for *PL* creep on a global scale. A grain size of 5 mm will be used for *NH* creep in accordance with our former work.

Figure 2B shows the observed relations between Q and T_m/T. Q-values from body wave data (Sutton et al., 1967; Barazangi and Isacks, 1971; Anderson and Kovach, 1964) were considered more reliable than those derived from surface wave inversion. They can more easily be located with respect to depth and position and can therefore better be related to the temperature regime. The lunar value is from Latham et al. (1970). The relation of Q to the T_m/T regime was made for a depth of 100 ± 50 km, mostly relying to the P_n-wave or direct P-wave data at these depths. The T_m/T-values in various heat flow provinces were estimated from those of Vetter and Meissner (1976) for terrestrial data and Meissner (1975) for the lunar data point. We are aware of the fact that the lunar data point may not be representative for terrestrial material because of the high vacuum and the total absence of any traces of water. We have added it, though, to the terrestrial data because of the petrological similarity between the terrestrial and the lunar mantle and our belief that possibly Q and η might be equally influenced by the strange lunar environment.

As seen from Fig. 2B, it is not possible to combine the data points by a straight line in the semilog diagram. Two straight lines, considered as extreme values in the higher and lower temperature field, have been plotted with gradients of $\partial(\ln Q)/\partial(T_m/T) = 6.6$ and 2.2, respectively. The transition between the two gradients, like that between the two creep laws, is between 0.7 and 0.8 T_m. The two straight line approximations can be expressed as

$$\lg Q = 0.94\, T_m/T + 1.25 \qquad \text{for} \quad T/T_m < 0.7 \tag{17}$$

and

$$\lg Q = 2.87\, T_m/T - 1.17 \qquad \text{for} \quad T/T_m > 0.8. \tag{18}$$

A detailed comparison between η and T_m/T as well as between Q and T_m/T may be obtained by a cross section through a well known subduction zone. We combine recent microearthquake observations of Hasegawa et al. (1978), which clearly show a double planed structure in the descending Pacific plate near Japan, with a temperature model of Toksöz et al. (1971). Figure 3A shows Hasegawa's results which were obtained by taking into account the velocity structure of the descending plate and its vicinity. Figure 3B gives the temperature model of Toksöz et al. (1971) with some modifications in the back arc area where an observed heat flow of more than 2 HFU was used as a basis to modify the temperature model. Figure 4 is obtained by transforming the temperature curves into T/T_m-curves and superimposing them onto Fig. 3A. The coincidence between the double plane earthquake structure and the isoline $T/T_m = 0.6$ ($T_m/T = 1.67$) is really striking. Earthquakes only occur within an area with $T/T_m \lesssim 0.6$. This value had previously been considered to be the lower limit of T/T_m where enough stress for a subsequent stress release by quakes can accumulate (Meissner, 1978; Vetter and Meissner, in preparation). The temperature relation of $T/T_m = 0.6$ corresponds to whole rock viscosities of $\eta \approx 10^{23}$ poise. Lower viscosities do not allow the build-up of significant stresses because the whole material creeps. In fact, no quakes occur outside the double plane structure where whole rock viscosities are below 10^{23} poise with the exception of an anomalous area in the upper right side of the figure where apparently a new slab of cold material is formed which has not been incorporated into the thermal model.

Very rigid and highly viscous zones should not show quakes either if they are surrounded by a less viscous material. In fact, only very few quakes are observed in the very interior of the subduced plate inside the two earthquake belts, which may be considered as two "fault zones" with a slightly reduced effective fault

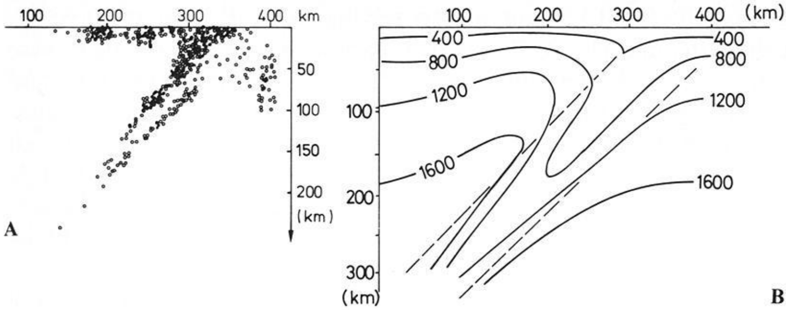

Fig. 3. A Microearthquakes as determined by recent array studies under the Japanese Islands (Hasegawa et al., 1978). **B** Thermal model for a descending plate in a subduction area with $v_{sub} = 8$ cm/y; adiabatic compression, conduction, phase changes, strain heating, and back arc spreading are taken into account. Temperatures in °C. [After Toksöz et al. (1971), modified in the back arc area]

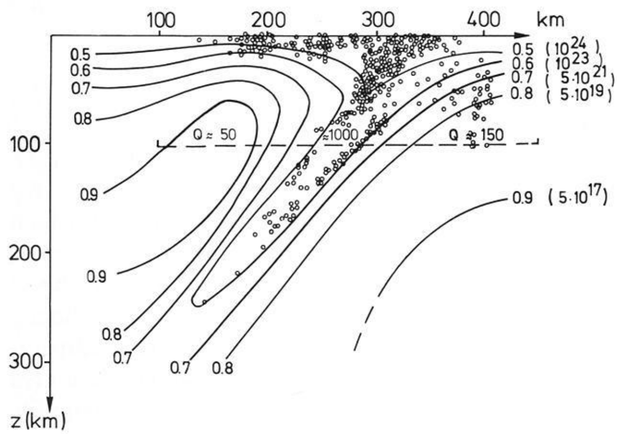

Fig. 4. T/T_m-values as calculated from Toksöz's et al. (1971) model related to Hasegawa's et al. (1978) microearthquake distribution. *Numbers in brackets* are viscosity values in poise calculated for *PL*-creep ($\dot{\varepsilon} = 10^{-14} \ldots 10^{-16}\,\mathrm{s}^{-1}$) and for *NH*-creep (grain size 5 mm). Quakes in the *upper right* hand side of the figure possibly show a new onset of subduction not contained in the thermal model

zone viscosity. Quakes in the high viscosity area between the "fault zones" are scarce because high fracture stresses tend to translate all stresses to the next "weaker" zones, zones which apparently are be double planed "master faults".

The good agreement between viscosity and seismicity in this example seems to be a solid basis to relate the η- to the Q-values. Adopting the Q-model of Barazangi and Isacks (1971), Fig. 5 shows the smooth η-curve together with the average Q-values from P body waves for a depth of 100 km. There is certainly a close relationship between both parameters.

Finally, Fig. 6 summarizes all available Q-values and shows their relation to η. This figure was obtained by converting the data of Figs. 2A and B into a

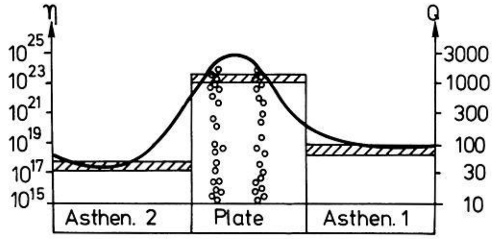

Fig. 5. η and Q in a cross section through the subduction area at a depth of approximately 100 km. *open dots:* zone of quakes inside the descending plate; *Asthen. 1:* "normal" oceanic asthenosphere; *Asthen. 2:* asthenosphere in the back arc area hatched bars = average Q-values of Barazangi and Isacks (1971); η-curve from values of Fig. 4

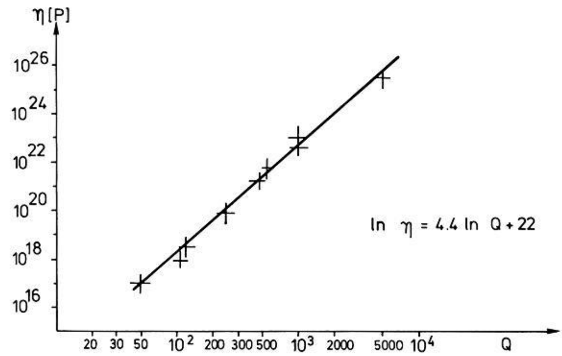

Fig. 6. General relation between η and Q as obtained from the data of Fig. 2A and B

combined log-log diagram. As the two gradients in Fig. 2A and those of Fig. 2B both differ by a factor of about 3 and change at about the same value of T/T_m it was not surprising to us to observe a rather well established linear relationship between $\log\eta$ and $\log Q$:

$$\lg\eta = 4.4\log Q + 9.6 \tag{19}$$

and

$$\ln\eta = 4.4\ln Q + 22 \qquad \text{(for } \eta \text{ in poise).} \tag{20}$$

This means that apparently different values of g_Q are related to the different creep laws, i.e., NH-creep with $g_1 = g^* = 29$ is connected with $g_Q^{(1)} = 6.6$ and *PL*-creep ($n=3$) with $g_2 = 1/3\, g^* = 29/3$ is connected with $g_Q^{(2)} = 2.2$ $(=6.6/3)$. As mentioned before, an average value of $\dot{\varepsilon} = 10^{-15}\,\mathrm{s}^{-1}$ was used for the calculation of η.

4. Discussion

As seen from Figs. 2B and 6, the error bars are rather large. As mentioned before, the values of the 2 different gradients $\partial(\ln Q)/\partial(T_m/T)$ should be considered as extremes or asymptotic values. Also, the gradient of $\partial(\ln \eta)/\partial(T_m/T)$ should be understood as asymptotic; further, as mentioned before, the dependence of η on $\dot{\varepsilon}$ causes uncertainties in η because of the uncertain values of $\dot{\varepsilon}$. In another approach we have tried to calculate η for different depths by assuming a systematic change of $\dot{\varepsilon}$ from the asthenosphere into the lithosphere. In doing so we arrive at smaller gradients for $\partial(\ln \eta)/\partial(\ln Q)$. However, as $\dot{\varepsilon}$ on a global scale is hard to define and because all creep experiments as the basis for the calculation of creep constants are performed at constant $\dot{\varepsilon}$-values, we preferred to present the general relation between Q and η for a constant $\dot{\varepsilon}$. We have also compared our results with those of Anderson and O'Connell (1967) who gave a constant ratio of η/Q of about $4 \cdot 10^{19}$ poise. As seen from Fig. 6, our ratio η/Q is about 10^{16} poise for $Q=50$–100 and 10^{20} for a Q around 1000.

The linear relation between $\ln Q$ and $\ln \eta$ is certainly hard to understand in view of the small stresses associated with wave transmission and the stress dependent different creep processes which are related to grain boundary diffusion or power law dislocation glide mechanisms. This means that not only does the creep mechanism change between 0.7 and 0.8 T_m, but also the relaxation mechanism responsible for Q changes its character at these temperatures. Apparently only small stresses can build up in the range of $T/T_m > 0.8$, where creep is not dependent on creep rate or stress. In contrast, at $T/T_m < 0.7$ the lithosphere generally exhibits some stress, $\sigma > 1$ bar, which may be released in earthquakes if the temperature drops a little more and more stress is built-up.

Regarding the activation energies, a value of 125 kcal/mol (for olivine) corresponding to $g^* = 29$ is found for NH creep, while an "effective activation energy" of only $125/3 = 41.7$ kcal/mol is found for PL creep (because of $g = g^*/3$). The activation energy at higher temperatures is 28.3 for the attenuation process; at slightly lower temperatures it is 9.44 kcal/mol. Certainly, the transmission of an elastic wave does not activate as many atoms or grain-boundaries as does a steady state creep process; however, at higher temperatures a larger number of particles is apparently activated.

So far, a dependence of Q on the frequency of waves has not yet been proved by field observations. Many authors present convincing evidence that these data are consistent with Q being independent of frequency at least between 1 and 60 s (Anderson and Hart, 1978). Equation (11) however contains a frequency term. Several authors, among them Lomnitz (1957) and Kanamori and Anderson (1977), try to avoid this dilemma by assuming a whole spectrum of relaxation times, the superposition of them then giving a quasi frequency independent Q. While this may be an interesting approach for deriving Q-values for different depths, compositions, and densities, it fails to explain the Q-values of the lower lithosphere and asthenosphere where olivine certainly dominates. The only way to remove the dependence of frequency from Eq. (11) is to assume that the term $\omega \tau_0$ is constant. This means that τ_0 must be proportional to $1/\omega$. Its value may be obtained by comparing (11) with (17) and (18). It is about $2.3 \cdot 10^{-4}$ s for a 10 Hz frequency and $T/T_m < 0.7$ and about $1.4 \cdot 10^{-4}$ s for 10 Hz and $T/T_m > 0.8$. This means that τ_0 certainly is too large to be related directly to the atomic jump frequency and therefore seems to describe a more general relaxation upon very small stresses. The higher the frequency, the lower is τ_0.

5. Conclusion

It has been shown by a simple theoretical approach that the parameters Q and η can be expressed by a very similar exponential relationship with regard to the ratio of the solidus temperature T_m to temperature T, at least for body wave frequencies larger than about 1 Hz and for $0.95 > T/T_m > 0.5$. The attenuation seems to obey two different laws, one at higher and one at lower temperatures; as such there is a strong similarity in this behavior to that of creep in that there are also two dominating processes, the Nabarro-Herring or Newtonian linear creep law in a small stress-high temperature regime and the power law dislocation glide mechanism at slightly lower temperatures. Effective activation energies for seismic attenuation seem to be only 23% of that of steady state creep. Both creep and attenuation exhibit lower effective activation energies for lower temperatures. The change of effective activation energies for both creep and attenuation to about a third of that for high temperature values takes place between 0.8 and 0.7 T_m. From this, a linear relationship between $\lg Q$ and $\lg \eta$ has tentatively been established. It agrees with field observations of high frequency body waves and known creep processes in the earth's asthenosphere. The relation is substantiated by more evidence than were those of Anderson and O'Connell (1967) or Meissner (1975) but will certainly be subject to revision when more data become available. Moreover, other relations definitely exist in the upper lithosphere especially for sediments, and possibly also for the middle and deeper part of the mantle, where phase transitions change the structure of minerals. It should, however, be mentioned that data from surface wave and free oscillation inversions and those from viscosity estimations may well agree with the empirical relation presented in this study.

Acknowledgement. When presenting our paper at the EGS-ESC meeting in Strasbourg in September 1978 we noticed that Berckhemer and his coworkers had found quite a similar relationship between Q and η, based on experimental high temperature studies in Frankfurt. Subsequent discussions helped us considerably in developing our arguments. We thank Dr. A. Binder for critically reading the manuscript.

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