# Reports

## Earth's Viscosity

Abstract. *Seismic methods are now being used to determine not only Earth's elastic properties, but also by how much it departs from a perfectly elastic body. The seismic anelasticity* (Q) *varies by several orders of magnitude throughout the mantle, the main feature being an extremely dissipative zone in the upper mantle above 400 kilometers. Recent determinations of viscosity by McConnell show a similar trend. The two sets of data indicate that the ratio of viscosity to* Q *is roughly a constant, at least in the upper mantle of Earth. On the assumption that this relation is valid for the rest of Earth, viscosities are estimated in regions that are inaccessible for direct measurement. The implied presence of a low-viscosity zone in the upper mantle, overlying a more viscous, less deformable, lower mantle, reconciles viscosites calculated from the shape of Earth and from postglacial uplift. The mismatch of the deformational characteristics at various levels in Earth, coupled with the changing rate of rotation, may be pertinent to the rate of release of seismic energy as a function of depth.*

Nonelastic processes are responsible for some of the most interesting and basic problems regarding Earth and its evolution. These problems include the shape of Earth, isostasy, mountain building, polar wandering, formation of the core, earthquake sequences, change in length of day, tidal heating, tidal acceleration of the moon, postglacial uplift, and the possibility of continental drift and convection in the mantle. Seismology has provided a complete description of Earth's elastic properties and of their variation with depth. Elastic waves supply the only direct means of penetrating the deep interior of Earth; their velocities yield the basic data for calculating its internal elastic structure. The elastic properties, however, are essentially irrelevant to the above problems that involve rate processes, except that the variation in nonelastic properties in Earth is probably as complicated as the variation in elastic properties. Most attempts to compute the nonelastic response of Earth have assumed homogeneity—for example, a constant viscosity with depth.

Even in those processes that are amenable to mathematical formulation, the rheological properties of Earth are not well enough known for even a rough estimate of rates, or even for determination that a given process, such as convection, is possible. Direct measurement of rates, which would yield an estimate of a measure of nonelasticity such as a relaxation time or a viscosity, is difficult because of the extremely slow motion and long time-scales involved.

Estimates of viscosity have come from the uplift of previously loaded portions of Earth and from its nonequilibrium shape. The uplifts of Fennoscandia and Lake Bonneville gave viscosities of $10^{22}$ and $10^{21}$ poises, respectively (*1*), while the nonequilibrium shape of Earth yielded a value of $10^{26}$ poises (*2*). For these results pure Newtonian or Maxwellian flow was assumed in a homogeneous material with constant viscosity. The phase delay of the solid tides and the damping of the Chandler wobble have not yet yielded satisfactory values for viscosity. This uncertainty in viscosity—five orders of magnitude—precludes definitive conclusions regarding, among other questions, the controversial problem of convection in the mantle.

Seismic waves, our most powerful tool for probing the interior of Earth, can in principle yield information regarding anelasticity as well as elasticity, and methods are being developed to determine anelasticity as a function of depth at seismic frequencies (*3*).

The most direct manifestation of anelasticity is the damping of seismic waves with distance or with the decay with time of Earth's free oscillations. Such data have already supplied the best estimates of the departure of Earth from a perfectly elastic body in its various regions. The seismic measure of anelasticity is the dimensionless quality factor $Q$, which seems to be roughly independent of frequency for homogeneous materials (*3*). The observed frequency dependence of $Q$ for long-period surface waves and free oscillations is attributed to the variation of $Q$ with depth in Earth. These studies have demonstrated the existence of a low-$Q$ region in the upper mantle in the general vicinity of the Gutenberg low-velocity layer. Below a depth of about 400 km the attenuation of seismic waves decreases markedly—that is, $Q$ increases rapidly. This is also the region in which the elastic properties increase abruptly. $Q$ varies by several orders of magnitude between top and base of the mantle. In principle the method can give the anelasticity at seismic frequencies throughout Earth, although it gives only a measure of anelasticity and not the physical mechanism.

A convenient measure of anelasticity more appropriate to long-term processes is viscosity. This quantity can be determined in principle from the rates of geological processes, which in practice are so slow that direct measurement is usually impossible and that recourse must be made to the geologic record. McConnell (*4*) recently estimated the variation of viscosity with depth in the upper mantle from relaxation time versus wave length in Fennoscandia, finding a low-viscosity region in the upper mantle and a rapid increase in viscosity with depth below about 400 km.

Figure 1 shows the viscosity profile determined by McConnell and two $Q$ profiles determined from seismic shear waves. The major features of these two measures of anelasticity are remarkably similar; low $Q$ seems to imply low viscosity, and vice versa. We expect an intimate relation between these two measures of anelasticity since, as activated processes, they are similar functions of temperature and pressure and

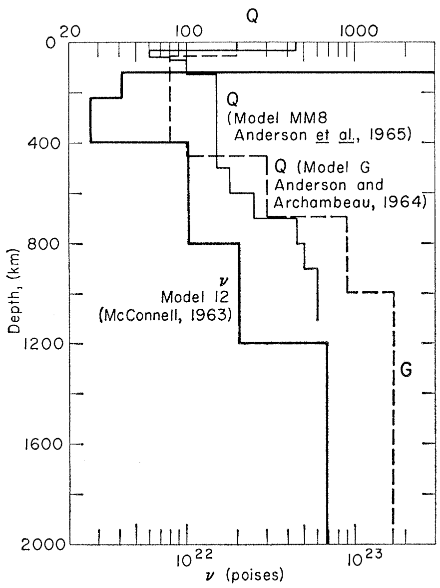


Fig. 1. Variation of seismic anelasticity in shear, $Q$, and viscosity ($\nu$) as a function of depth in Earth. The parameter $Q$ is a measure of the energy loss per cycle.

both are measures of the defect structure of the mantle.

The ratio of viscosity to $Q$ is roughly a constant:

$$\nu/Q \simeq 0.4 \times 10^{20}. \tag{1}$$

Although it has the limitations of any empirical relation, particularly in regard to extrapolation, the good correlation between $Q$ and viscosity in regions where they have both been measured encourages us to estimate viscosities in regions of Earth in which direct determination has not been possible. Even an estimate within an order of magnitude will be useful in view of our previous complete lack of information regarding the mechanical behavior of the deep mantle.

Kovach and Anderson (*3*) determined average values of $Q$ in shear for the whole mantle (600), the upper 600 km of the mantle (200), and the lower 2300 km of the mantle (2200); these values lead to estimates of viscosity (Eq. 1) of $2.4 \times 10^{22}$ poises for the whole mantle, $8 \times 10^{21}$ for the upper mantle, and $10^{23}$ for the lower mantle. Their estimate of 5000 for the $Q$ at the base of the mantle gives a viscosity of $2 \times 10^{23}$ poises. These viscosities are all much lower than the $10^{26}$ estimated for the mantle by MacDonald (*2*) from the nonequilibrium shape of Earth, on the assumption that viscosity is constant throughout the mantle. All this evidence suggests, however, that large-scale deformations of Earth involve the upper mantle more than the lower mantle; a viscous-layer model is clearly more appropriate.

If we assume that flow is restricted to a relatively thin layer in the upper mantle, then the relaxation time of Earth's bulge can be estimated from Jeffreys's formula (*5*) for flat-lying layers:

$$\tau = 6L^2\nu/\rho g H^3\pi^2 \simeq (1.8 \times 10^{-4})(L^2/H^3) \tag{2}$$

where $\tau$ is the relaxation time, $g$ is the gravitational acceleration, $H$ is the thickness of the viscous layer, $L$ is the wavelength of the deformation, and $\rho$ is the density. If $H$ is 400 km, $\rho$ is 3.5 g $\text{cm}^{-3}$, $L$ is $2\pi a$ ($n + 1/2$), and $n$ is 2,

$$\tau \approx (7.2 \times 10^{-9})\nu. \tag{3}$$

The average $Q$ for the upper 400 km, as determined from surface-wave and free-oscillation data, lies between 88 and 135. The corresponding viscosities from Eq. 1 are between $3 \times 10^{21}$ and $5 \times 10^{21}$ poises. The relaxation time for the shape of Earth is then about 2 to $4 \times 10^{13}$ seconds. Munk and MacDonald (*6*) obtained $3 \times 10^{14}$ seconds from the observed shape and the change in angular acceleration. This is satisfactory agreement in view of the oversimplified one-layer model, the neglect of sphericity, and the assumed absence of flow in the lower mantle; agreement is better if $H$ is 200 km. Thus the nonequilibrium shape of Earth seems not inconsistent with values of viscosity determined from Fennoscandian uplift (*7*).

If the whole mantle participates equally in the deformation, as assumed by MacDonald, then $H \simeq 3 \times 10^8$ cm, $\nu \simeq 2.5 \times 10^{22}$ poises, and $\tau \simeq 4 \times 10^{11}$ seconds—three orders of magnitude less than the "observed" relaxation time for the shape of Earth. This is very approximate since $H$ is now a large fraction of the radius of Earth and the neglect of sphericity becomes critical.

Alternatively, from the "observed" relaxation time, $\tau = 3 \times 10^{14}$ seconds, we obtain $\nu \simeq 2 \times 10^{25}$ poises, which is much higher than the estimates from $Q$. MacDonald obtained an even higher value by assuming that rigidity rather than buoyancy was the restoring force, precluding direct comparison with the Fennoscandia data. His approach is, however, probably useful in providing an absolute upper bound on the viscosity of the mantle.

The lowest $Q$ measured seismically is about 60, in a thin layer at the base of the crust, yielding a value of about $2 \times 10^{21}$ poises. This layer may be responsible for the low viscosity of $10^{21}$ poises measured from the uplift of the relatively small Lake Bonneville region. It is also likely, however, that temperatures in the upper mantle under this region are anomolously high, and viscosities are therefore correspondingly low.

Although crude, my estimates of viscosity are determined directly from the most relevant data available for the interior of Earth, namely, the seismic anelasticity. Previous estimates of viscosity assumed a mechanism and required estimates of such unknowns as temperature, pressure, activation energy, activation volume, and grain size; my estimates are based on the experimental evidence that the ratio $\nu : Q$ remains constant with depth.

The ability of Earth to deform in response to internal or external forces differs markedly in different regions of the mantle, especially between the upper and lower mantle and, possibly, between the top of the low-velocity, low-$Q$ layer and the overlying crust. This mismatch, coupled with the changing angular velocity of Earth, may be a mechanism for concentrating earthquakes at these depths.

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### References and Notes

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