Creep Laws for the Mantle of the Earth [and Discussion]

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Creep laws for the mantle of the Earth

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The analyses of glacial rebound data by Cathles and by Peltier and Andrews have led them to the conclusion that the flow law of the mantle of the Earth is Newtonian and that the viscosity is essentially a constant (10^{22} P) throughout the mantle. In this paper it is concluded that no large strain, steady-state creep process in mantle rock can account for a Newtonian, constant viscosity mantle. It is suggested that small strain, transient creep and not steady-state creep is involved in the isostatic rebound phenomenon. Since convective motion in the mantle involves large creep strains, conclusions about the effective viscosity of mantle rock undergoing such flow that is based on isostatic rebound data are likely to be wrong. If Post's and Carter & Mercier's laboratory results of the stress dependence of the grain size in a mantle type rock are representative of the actual grain sizes in the mantle, power law creep is almost certainly the creep law that governs convective flow in the mantle.

INTRODUCTION

The temperature of the rock in the Earth's crust and mantle ranges from a small fraction of the melting (solidus) temperature, at shallow depths near the Earth's surface, up to the melting (solidus) temperature at depths within the low-velocity, low strength zone from which magmas of many volcanoes originate. Rocks in the Earth's crust and mantle also are subjected to a wide range of stresses. At teleseismic distances from an earthquake source the stresses are much smaller than those required to activate the Frank–Read sources necessary for plastic deformation by dislocation movement. At the other end of the spectrum explosive events, natural and man-made, may cause the theoretical shear strength of the rock to be exceeded.

The type of response of rock to the stress that deforms it depends both on the magnitude of that stress and the temperature of the rock. The stress and temperature ranges of the different types of responses are conveniently indicated on a diagram in which one axis is a normalized stress σ/μ, where σ is the stress and μ is the shear modulus of rock, and the other axis is the normalized temperature T/T_m, where T is the temperature and T_m is the melting (solidus) temperature.

Figure 1 shows such a diagram, which we have called a creep diagram when applied to creep phenomena (Weertman & Weertman 1965, 1975), for crystalline material like rock minerals, whose dislocations cannot move easily at the lower temperatures because of the existence of a high Peierls stress. The effective Peierls stress at the higher temperatures is reduced because of thermally produced stress fluctuations in the thermal sound waves within the solid. It should be pointed out that most mechanisms that give rise to anelastic deformation, and thus produce the damping of seismic waves, also can operate at stresses above the Peierls stress-dislocation source stress boundary in figure 1. Moreover deformation mechanisms, such as the Nabarro–Herring mass transport of point defects between grains boundaries by diffusion, can operate below the Peierls stress-dislocation source stress boundary. In this region Nabarro–Herring creep can produce large plastic deformation in fine-grained material although the effect would be negligible in large crystals. Rock within the Earth's mantle also is subjected to high hydrostatic pressure. Since hydrostatic pressure raises the melting temperature of most materials the effect of pressure can be taken into account in a diagram such as figure 1 through its influence on the melting temperature.

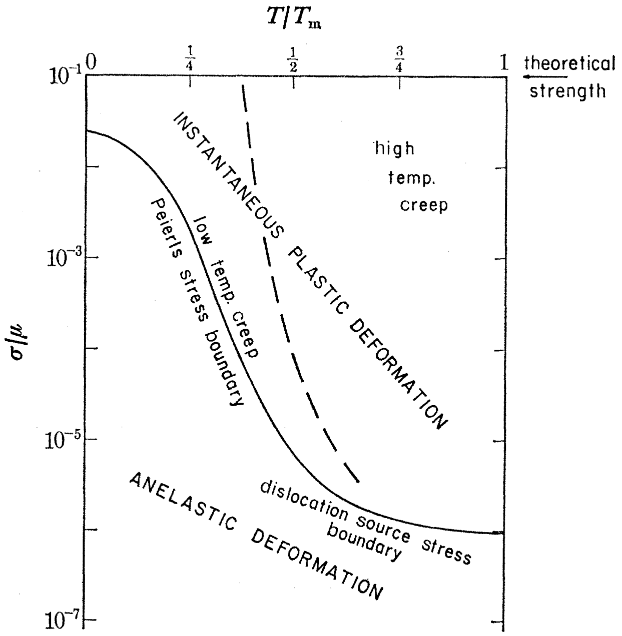

FIGURE 1. Schematic diagram of normalized temperature and normalized stress showing fields where different types of deformation and creep occur for crystalline material with a large Peierls stress.

In the case of steady-state creep processes a qualitative diagram like figure 1 can be made quantitative by plotting on it lines of constant creep rate. This is what Ashby and his co-workers (Ashby 1972; Stocker & Ashby 1973) have done and the resultant plots have been called by him deformation maps. The constant creep rate curves can be obtained over part of the diagram from experimental data, if it is available. In regions where no experimental data exist, theoretical curves can be calculated from theories that give steady-state creep rates based on the creep mechanism judged most likely to be dominant in the particular region. The deformation maps have not included transient creep data up to now but transient creep maps could be made by plotting, say, fixed times to reach some standard creep strain.

A geophysicist who works on a problem such as the process of continental drift or isostatic rebound from glacial loads wants to know the creep behaviour of mantle rock at the stresses, temperatures and pressures that exist within the Earth's mantle. He needs a deformation map which contains plots of constant creep rate in the area that corresponds to mantle conditions, but this area cannot be reached in laboratory experiments. So he must try to get there, by hook or crook, from extrapolation of experimental data, from geophysical field data, and from theories. Review articles on creep of rock and creep of the mantle (Gordon 1965, 1967; McKenzie 1968; Weertman 1970; Stocker & Ashby 1975; Weertman & Weertman 1975; Kirby & Raleigh 1973; Kohlstedt, Goetze & Durham 1976; Tsukahara 1974, 1976; Meissner & Vetter 1976; Heard 1976; Murrell 1976; Lorimer 1976; Nicolas 1976; Nicolas & Poirier 1976; Carter 1976), of which Carter's and Nicolas & Poirier's are the most complete and up to date, have summarized the experimental data and theories used in attempts to fill in the mantle area of the deformation map.

In recent years there has been controversy over what is the most likely creep law and creep mechanism for rock which is deforming under the stress, temperature, and pressure conditions found within the mantle. This paper is addressed to this controversy. The main point in dispute is the following: Is the appropriate creep law a linear law (Newtonian flow) or a power law with a power significantly larger than one?

Stress range – creep rate range

The order of magnitude of the creep rates and stresses of interest to flow processes within the mantle can be estimated with little difficulty. Plate motion is known to occur with velocities of the order of 1–10 cm a^{-1}. The convection within the mantle that gives rise to the motion of plates must take place over distances which are not much smaller than the thickness of a plate (about 100 km), and no larger than the thickness of the mantle (3000 km). Hence the creep rates for convective motion are within the range of $10^{-16}\,s^{-1}$ to $3 \times 10^{-14}\,s^{-1}$. The isostatic rebound that follows the removal of a glacial load from the Earth's surface produces a maximum vertical displacement of the crust of about one third the thickness of the ice (*ca.* 3 km) that is removed. Since the horizontal scale of a large ice sheet is several thousand kilometres, rebound probably involves the mantle down to depths of comparable distances. Thus the total strain is of the order of 10^{-3}. Walcott (1972) has estimated uplift velocities over the last 6000 a of the order of 2 cm a^{-1} near the centres of the former ice age ice sheets. This velocity would correspond to creep rates of the order of $10^{-15}\,s^{-1}$, a value similar to that estimated from the convective motion.

A creep rate of the order of $10^{-15}\,s^{-1}$ is well beyond a meaningful measurement in a laboratory experiment. At that creep rate in a test of one year's duration the total creep strain is only 3×10^{-8}. To produce such a strain, each dislocation in a sample with a dislocation density of $10^{12}\,m^{-2}$ ($10^{8}\,cm^{-2}$) would have to move, on the average, about one half a lattice spacing. In the case of creep produced by dislocation motion, to reach steady-state creep each dislocation source must produce many dislocation loops and each loop must move through distances orders of magnitude larger than the interatomic distance.

An estimate can be made of the order of magnitude of the stress needed to produce a creep rate of about $10^{-15}\,s^{-1}$ in the mantle. An ice sheet with an average thickness of 2–3 km pushes the Earth's crust downwards with a pressure of the order of 20–30 MPa (200–300 bar). The non-hydrostatic stresses that this applied load sets up within the mantle before isostatic adjustments, take place may be $\frac{1}{2}$–$\frac{1}{4}$ this amount; that is, about 5–10 MPa. Removing the ice load after equilibrium has been reached would produce stresses of the same magnitude but of the opposite sign. Walcott (1972) estimates that about 300 m of uplift still will take place in Canada. This present depression of the Earth's crust corresponds to a non-hydrostatic stress component of about 2–5 MPa (20–50 bars).

For convective flow in the mantle a rough estimate of the order of magnitude of the stress can be made. The stresses should be some fraction of the quantity $\rho g \alpha_t \Delta T L$, where ρ is the density of rock, g is the gravitational acceleration, α_t is the coefficient of thermal expansion of the rock, ΔT is the average temperature difference that drives the convection, and L is a characteristic distance over which the convection is taking place. The amount of heat transported per unit time by convection to the Earth's surface is equal to $\mathscr{A}Q$, where $\mathscr{A}$ is the area of the Earth and Q is the average heat flux at the Earth's surface. The heat transport also is of the order of magnitude of $\mathscr{A}vc\Delta T$, where v is the average velocity of rock within the mantle and c is the specific heat of rock. Thus the magnitude of the stresses is some fraction of the expression $\rho g \alpha_t LQ/vc$. For $\rho = 5 \times 10^3\,\mathrm{kg\,m^{-3}}$, $g = 9.8\,\mathrm{m\,s^{-2}}$, $\alpha_t = 2 \times 10^{-5}\,\mathrm{K^{-1}}$, $L = 1000\,\mathrm{km}$, $Q = 0.052\,\mathrm{W\,m^{-2}}$, $v = 1\text{–}10\,\mathrm{cm\,a^{-1}}$, and $c = 5\,\mathrm{MJ\,m^{-3}\,K^{-1}}$, the stresses are some fraction of 3–30 MPa (30–300 bar).

It seems not unreasonable to expect that the stresses that produce flow in the mantle are of the order of magnitude of 5 MPa (50 bar) plus or minus half an order of magnitude.

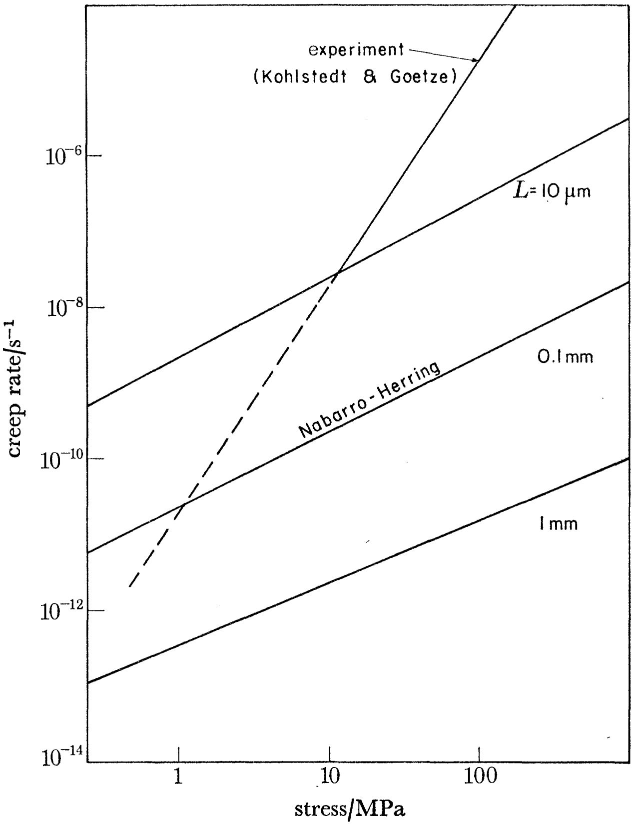

Figure 2. Plot of creep rate against stress for olivine, normalized to a temperature of 1400 °C. Solid part of line is in region where data were obtained by Kohlstedt & Geotze (1974). Also shown are creep rates predicted from the Nabarro–Herring creep equation for various grain sizes. Data used to obtain the plots are given in the text.

Cross-over stress for dislocation creep and Nabarro–Herring creep

There has been a persistent controversy whether the dominant type of creep in the mantle is that produced by dislocation motion or is that produced by the diffusional mass transport of atoms between grain boundaries: the Nabarro–Herring creep mechanism. The relative importance of these two mechanisms in the mantle depends upon the stress level and the grain size. This point can be understood most easily through figure 2. This figure shows for olivine

$$((\mathrm{Mg_{0.92}Fe_{0.08}})_2\mathrm{SiO_4})$$

at 1400 °C and atmospheric pressure logarithmic plots of creep rate $\dot{\epsilon}$ against stress calculated for various grain sizes from the Nabarro–Herring creep equation

$$\dot{\epsilon} = \alpha(D/L^2)\,(\sigma\Omega/kT), \tag{1}$$

where α is a dimensionless constant ($\alpha \approx 5$), L is the grain size, k is Boltzmann's constant, T is the temperature, Ω is the atomic volume, and D is the diffusion coefficient. The diffusion coefficient is given by the equation

$$D = D_0 \exp(-Q/kT) \exp(-P\Delta V/kT), \tag{2}$$

where D_0 is a constant, Q is the activation energy of diffusion, ΔV is the activation volume of diffusion, and P is the hydrostatic pressure. (In our earlier papers (Weertman 1970; Weertman & Weertman 1975) we have assumed that the diffusion coefficient is also given by

$$D = D_0 \exp(-gT_{\mathrm{m}}/T), \tag{3}$$

where g is a dimensionless constant ($g \approx 29$) and T_{m} is the melting temperature at the hydrostatic pressure P.) The Nabarro–Herring plots in figure 2 were calculated using

$$\Omega = 1.71 \times 10^{-28}\,\mathrm{m}^3$$

and the diffusion data estimated by Goetze & Kohlsteds (1973) from climb of dislocation loops ($D_0 = 3\,\mathrm{m^2\,s^{-1}}$ and $Q = 565\,\mathrm{kJ\,mol^{-1}}$ ($135\,\mathrm{kcal\,mol^{-1}}$).)

Figure 2 also shows a plot of the steady-state creep rate of olivine single crystals measured by Kohlstedt & Goetze (1974) and normalized to a temperature of 1400 °C using the measured activation energy of creep of $524\,\mathrm{kJ\,mol^{-1}}$ ($125\,\mathrm{kcal\,mol^{-1}}$). The magnitude of the creep rate of these data seems to be representative of that reported in the literature for dunite under dry conditions. Only data of Griggs & Post (1973) and Post (1973) have substantially smaller creep rates. The dashed part of the line shown in figure 2 is an extrapolation from the stress range where the experimental data was obtained. The experimental data obey the following power law creep equation with a stress exponent $n = 3$:

$$\dot{\epsilon} = C\sigma^3 \exp(-Q/kT), \tag{4}$$

where C is a constant ($C = 4.2 \times 10^{-13}\,\mathrm{Pa^{-3}\,s^{-1}}$) and Q is the activation energy of creep. (The two activation energies used in figure 2 are within experimental error of each other.) Equation (4) is of the form of theoretical creep equations that are based on a dislocation mechanism. A third power dislocation creep equation (stress exponent $n = 3$) can be expressed as

$$\dot{\epsilon} = \gamma(D/b^2)\,(\mu\Omega/kT)\,(\sigma/\mu)^3, \tag{5}$$

where γ is a dimensionless constant whose value depends upon the particular mechanism that controls the dislocation motion, b is the length of the Burgers vector of the dislocation, and μ is the shear modulus. (If $C = 4.2 \times 10^{-13}\,\mathrm{Pa^{-3}\,s^{-1}}$, $\mu = 79.1\,\mathrm{GPa}$, and $b = 0.698\,\mathrm{nm}$, then $\gamma \approx 0.058$.) Combining equations (1) and (5) gives the following equation for the stress at which the Nabarro–Herring creep equation and the dislocation creep equation predict the same creep rate:

$$\sigma = \mu(\alpha/\gamma)^{\frac{1}{2}}\,(b/L). \tag{6}$$

The cross-over stress that is given by equation (6) is independent of temperature and hydrostatic pressure except as μ varies with pressure. Equation (6) can be rearranged to give the grain size at which equations (1) and (5) predict the same creep rate for a constant stress. The equation is

$$L = b(\alpha/\gamma)^{\frac{1}{2}}(\mu/\sigma). \tag{7}$$

From figure 2 and equation (6) it is seen that the cross-over stress for olivine is 10 MPa (100 bar) for a grain size of 10 μm, 0.1 MPa (1 bar) for 1 mm, and 10 kbar (0.1 bar) for 1 cm. These grain sizes, particularly the larger ones, are all within the range of possible ones for mantle rock. The cross-over stresses they give bracket the rough estimates for stresses existing in the mantle that is given in the previous section.

Post (1973) has found in high-temperature creep experiments carried out on dunite that contains 98 % olivine in the deviatoric stress range of 100 MPa–1 GPa (1–10 kbar) that recrystallization occurred during the creep runs and that the recrystallized grain size is given by the equation

$$L/L_0 \approx C'^{\frac{3}{2}}(\sigma_0/\sigma)^{\frac{3}{2}}, \tag{8}$$

where the constants $L_0 = 1\,\mu\text{m}$, $\sigma_0 = 100\,\text{MPa}$, and $C' = 19$. Equation (8) applies whether the creep runs are made under 'wet' or 'dry' conditions. Carter and Mercier (reported in Carter 1976) also measured the grain size as a function of stress on a dunite from the same source as used by Post and on a synthetic dunite. They also found that equation (8) gives the grain size, and values of the constants L_0, σ_0, and C' that they determined are essentially the same as those found by Post.

If the grain size L that is given by equation (8) is larger than that given by equation (7) power law creep is dominant. Because equation (7) gives $L \sim 1/\sigma^{\frac{3}{2}}$ there is a critical stress *below* which the dislocation creep mechanism (power law creep) is dominant and *above* which the Nabarro–Herring creep mechanism is dominant (assuming that equation (8) can be extrapolated outside the stress range for which it is experimentally established). This critical stress is found by combining equations (7) and (8) to give

$$\sigma = C'^3(\gamma/\alpha)\,(\sigma_0^3/\mu^2)\,(L_0/b)^2. \tag{9}$$

For $\alpha/\gamma = 86$ and the values of C', L_0 and σ_0 quoted above the critical stress given by equation (9) is equal to $\sigma = 26\,\text{GPa}$ (260 kbar). Thus if equation (8) offers a reasonable estimate of the grain size of mantle rock it would be very unlikely that Nabarro–Herring creep is the important creep mechanism in the mantle for steady-state creep. (It should be noted that for a stress of $\sigma = 5\,\text{MPa}$ the grain size predicted by equation (8) is $L = 7.4\,\text{mm}$, a not unreasonable value at such a low stress level. From figure 2 it is clear that Nabarro–Herring will not be the rate controlling creep mechanism if the dislocation creep mechanism does indeed operate at 5 MPa.)

At lower temperatures and for finer grain sizes grain boundary diffusion can increase significantly the Nabarro–Herring creep rate (Coble 1963). The Nabarro–Herring–Coble creep equation is the following modification of equation (1):

$$\dot{\epsilon} = \alpha(D/L^2)\,(\sigma\Omega/kT)\,[1 + (\pi\delta/l)\,(D_\text{B}/D)], \tag{10}$$

where D is again the bulk or lattice diffusion coefficient, D_B is the grain boundary diffusion coefficient, and δ is the effective thickness of a grain boundary as a fast diffusion path ($\delta \approx b$).

If it is assumed that the grain boundary activation energy is approximately two thirds of that of the bulk diffusion activation energy then for olivine with a value of $Q = 565\,\text{kJ/mol}$ (135 kcal/mol), the ratio $D_\text{B}/D = 7.9 \times 10^5$ at 1400 °C and $D_\text{B}/D = 3.7 \times 10^4$ at 1890 °C (the melting temperature of an olivine that contains almost no iron (Yoder 1976)). Thus the Coble creep rate is faster than the Nabarro–Herring creep rate in olivine for grain sizes smaller than 1.7 mm at 1400 °C (and 8 μm at the melting temperature of 1890 °C). In figure 2 the creep rate for a grain size of $L = 1$ mm is increased by a factor 2.7, the curve with $L = 0.1$ mm by a factor 17, and the curve of $L = 10$ μm by a factor of 170 if grain boundary diffusion is taken into account.

If equations (5) and (10) are combined the cross-over stress is found to be equal to

$$\sigma = \mu(\alpha/\gamma)^{\frac{1}{2}}\,(b/L)\,[1+(\pi\delta/L)\,(D_{\mathrm{B}}/D)]^{\frac{1}{2}}. \tag{11}$$

If equations (8) and (11) are combined, the following equation is found for the stress *above* which Nabarro–Herring–Coble creep is dominant and *below* which dislocation creep is the important creep mechanism

$$\sigma = C'^{3}\frac{\gamma}{\alpha}\frac{\sigma_0^3}{\mu^2}\left(\frac{L_0}{b}\right)^2\Bigg/\left[1+\frac{\pi\delta}{L_0}\frac{D_{\mathrm{B}}}{D}\left(\frac{\sigma}{\sigma_0}\right)^{\frac{3}{2}}C'^{-\frac{3}{2}}\right]. \tag{12}$$

At temperatures so low that the ratio D_{B}/D is a very large number, equation (12) reduces to

$$\sigma = \sigma_0^{\frac{3}{5}}\,\sigma^{*\frac{2}{5}}C'^{\frac{3}{5}}(L_0/\pi\delta)^{\frac{2}{5}}\,(D/D_{\mathrm{B}})^{\frac{2}{5}}, \tag{13}$$

where $\sigma^* = C'^3(\gamma/\alpha)\ (\sigma_0^3/\mu^2)\ (L_0/b)^2$. With the values of the terms α, γ, etc., used previously at one half of the melting temperature of magnesium rich olivine the stress σ given by equation (13) is $\sigma = 14$ MPa. Thus even at this rather low temperature it is doubtful that a Nabarro–Herring–Coble creep mechanism can control the creep rate in the mantle.

It is known that the power law creep rate of rock is increased substantially when water is present (the hydrolytic weakening effect discovered by Blacic & Griggs: see the discussion in Nicolas & Poirier 1976). If hydrolytic weakening only increased the power law creep rate and not the Nabarro–Herring–Coble creep rate the presence of water in mantle rock would make it even more probable that the Nabarro–Herring–Coble creep mechanism does not control the creep rate of the mantle. Of course, it is possible that the hydrolytic weakening effect occurs because it reduces the activation energy of diffusion. If such is the case the presence of water will increase the creep rates of power law creep and Nabarro–Herring–Coble creep by equal amounts.

Superplastic creep

Gueguen & Boullier (1976) have seen evidence of superplastic flow in the deformation textures of some mantle peridotite nodules they have examined. Twiss (1976) has proposed, in fact, that at least in the upper mantle superplastic creep may be the dominant creep mechanism. Superplastic creep occurs at a faster rate than Nabarro–Herring–Coble creep and it has a stress exponent n that is smaller than that found in dislocation produced creep. Generally, the exponent is of the order of $n = 1.5$–2 but values as low as $n = 1$ are reported. (A review of superplastic creep is given in the article of Edington, Melton & Cutler (1976). It also is discussed by Nicolas & Poirier (1976) as applied to rocks.) Thus if superplastic creep is the dominant creep mechanism in the mantle, the mantle would be Newtonian or near Newtonian in its flow properties.

Twiss based his quantitative calculations on the probability of superplastic flow on the following superplastic creep equation that was derived by Ashby & Verrall (1973):

$$\dot{\epsilon} = \alpha^*(D/L^2)\,(\sigma\Omega/kT)\,[1+(\pi\delta/L)\,(D_{\mathrm{B}}/D)], \tag{14}$$

where the dimensionless constant $\alpha^* \approx 100$. Equation (14) is essentially the Nabarro–Herring–Coble creep equation but with a creep rate that is one order of magnitude larger because the constant α^* is an order of magnitude larger than the constant α of equations (1) and (10).

Recently Nix (1976, private communication) has objected (and I believe his objection is well founded) to the model used by Ashby & Verrall (1973) in the derivation of their equation (14). The reason that the Ashby–Verrall creep equation is an order of magnitude faster than the Nabarro- Herring–Coble creep equation is that the main diffusional flow paths in their model do not start at one grain boundary and end at a different grain boundary. Instead, as shown in figure 3*a*, the diffusion paths generally start and terminate at the same grain boundary. The paths are shorter and the amount of material transported needed to attain the same creep strain is smaller than in the Nabarro–Herring–Coble model. Hence, the predicted creep rate is faster. However, Nix points out that the grain boundary sliding that is part of the Ashby–Verrall model produces an effective dislocation distribution at the grain boundaries that is shown on the left hand side of figure 3*b*. The stress fields of these dislocations are approximately the same as those of the 'superdislocations' shown on the right hand side of figure 3*b*. These latter dislocations have stress fields that will cause diffusional mass transport along paths also indicated on the right hand side of figure 3*b*. This mass transport is symmetrical and is not the same as that used to obtain equation (14) and will not produce an increase in the creep rate.

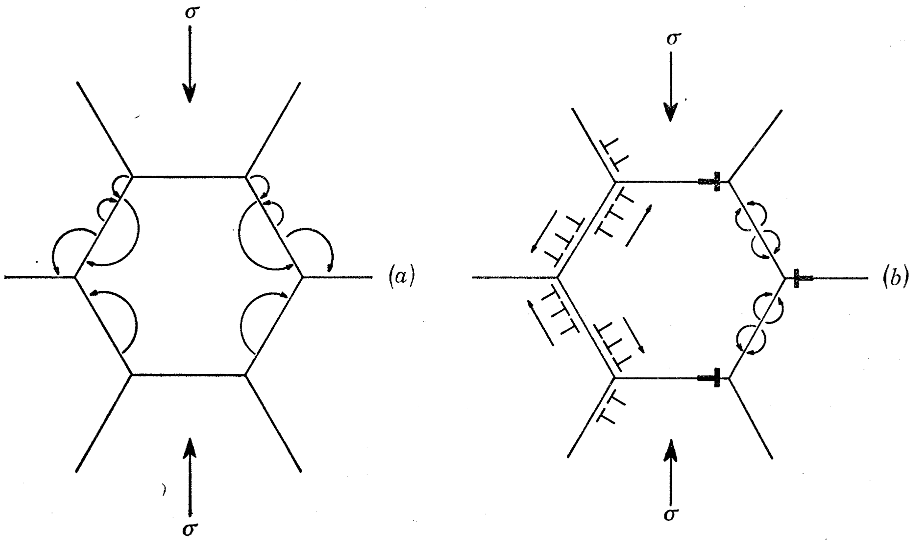

Figure 3. (*a*) Mass transport from grain boundaries as shown in fig. 7 of Ashby & Verrall (1973). (*b*) On the left hand side of the figure is shown the equivalent dislocations at grain boundaries that dislocation sliding would produce. These dislocations are roughly equivalent to the 'super' dislocations shown on the right hand side of the figure. The stress fields of the dislocations would cause mass transport as indicated by the arrows.

Whether the derivation of equation (14) is correct or not Ashby and Verrall have shown that it does predict the right creep rate for at least one superplastic alloy (Zn–0.2 mass % Al) of 3.5 μm grain size. Moreover, for a number of superplastic alloys the grain size dependence of the creep rate does agree with equation (14). That is $\dot{\epsilon} \propto 1/L^m$ where $m = 2$–3 (Edington *et al.* 1976). It seems not unreasonable to use their equation, as Twiss has done, in considering the possibility of superplastic flow in the mantle. The equations for cross-over stress, etc., discussed in the previous section are unaltered if the Ashby–Verrall equation is used. It is only necessary to increase the value of the constant α in those equations by one order of magnitude.

Cathles–Pelter–Andrews constant viscosity Newtonian mantle

Cathles (1975) has analysed the post glacial rebound data from N. America in a very ambitious attempt to determine the viscosity of the Earth's mantle from shallow to the deepest depths. He assumed that the flow law of the mantle was that of a Newtonian solid and found that the field data was consistent with this assumption and that, moreover, the coefficient of viscosity was essentially a constant and independent of depth within the mantle. The viscosity coefficient has the value of 10^{22} P (10^{21} Pa s). Cathles's results rule out the existence of a thick (*ca.* 350 km) low viscosity channel. However, he does conclude that a thin (*ca.* 75 km) low viscosity channel does exist in the upper mantle which has a viscosity (4×10^{20} P) smaller by a factor of 25 than for the rest of the Earth. Peltier & Andrews (Peltier & Andrews 1976; Peltier 1974, 1976) have reached essentially the same conclusions as did Cathles using a different mathematical analysis of the rebound data.

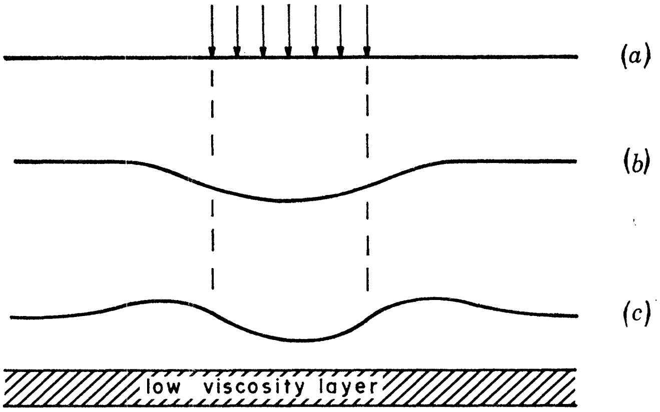

Figure 4. (*a*) Stress applied to a limited area on a half space. (*b*) Deformation produced by stress of (*a*) of surface of half space made of material that obeys a linear flow law with constant viscosity (or constant elastic constants). (*c*) Deformation of surface when material obeys a linear flow law but a layer of significant thickness exists below the surface within which the viscosity is very much lower than elsewhere in the half space. The deformation of the half space surface shown in (*c*) also occurs, whether the low viscosity layer exists or not, if the flow law of the half space material obeys a power law equation with an exponent significantly larger than unity.

Cathles's and Peltier & Andrews's conclusions depend not only on numerical results, but also on the qualitative character of the results. The general type of qualitative results used can be explained with the aid of figure 4. In figure 4*a* is shown a stress, that might correspond to a glacial load, pushing down over a limited area on the surface of a half space. If the material of the half space obeys the linear flow law with a constant viscosity everywhere, the deformation produced by the stress in (*a*) will be that shown schematically in figure 4*b*. The result to note is that the surface is lowered adjacent to the area subjected to the stress. If a low-viscosity channel of appreciable thickness exists below the free surface at a depth that is comparable with the horizontal extent of the surface area that is stressed, as is shown in figure 4*c*, the surface deformation will be that shown schematically in figure 4*c*. The surface adjacent to the stress area is raised rather than lowered. If the direction of the stress is reversed, as occurs during unloading by the melting away of an ice sheet, these deformations are also reversed. (The actual surface displacement histories actually are more complex than indicated by the schematic figure 4.)

The areas covered by Lake Bonneville and the Fennoscandia Ice Sheet were considerably smaller than that covered by the Laurentide Ice Sheet. Cathles found that the Lake Bonneville and the Fennoscandia rebound data corresponded essentially to the model of figure 4*b* and the Laurentide rebound data to figure 4*c*. Since the Laurentide rebound data sample the mantle deformation behaviour to deeper depths than do the rebound data of Lake Bonneville and Fennoscandia, Cathles and Peltier & Andrews were able to reach the conclusion that the Earth's mantle has constant viscosity except for a 75 km thick channel where the viscosity is reduced by a factor of 25.

Cathles ruled out a mantle governed by power law creep with an exponent n significantly larger than 1. A power law creep equation will produce a surface deformation that is shown in figure 4*c*. The qualitative behaviour is the same as found in channel flow. Thus the qualitative nature of Cathles's and Peltier & Andrews's results appear to be evidence against a power law creep being important for the mantle *if it is assumed that steady-state creep equations apply to the isostatic rebound phenomenon.*

The Fennoscandia rebound data have been used as evidence for a mantle that obeys a power law creep equation (Griggs & Post 1973; Post 1973). Because both channel flow and power law flow give the same qualitative rebound response the surface deformation history cannot be used as evidence for power law creep against Newtonian creep. Griggs and Post based their conclusion about power law creep on the curve of uplift at the centre of the rebounding area versus time. They found that the observed rebound curve fitted the one predicted using a power law creep equation (with $n = 3.21$) and not the one using a Newtonian creep equation ($n = 1$). However, as Cathles has pointed out, the shape of the uplift versus time curve depends upon the value estimated of the final uplift after infinite time. Cathles disputes the final uplift estimate used by Griggs and Post.

Cathles's and Peltier & Andrews's conclusions about the value of the viscosity of the upper and lower mantle are not in major conflict with those of others (see the review article of Walcott 1973; O'Connell 1971) except for McConnell's (1968) conclusion that the rebound data indicates that the viscosity of the lower mantle increases significantly with depth. However, it is very difficult to account for the findings that the viscosity of the lower mantle is low in magnitude and is a constant and that the flow of the lower mantle is Newtonian.

The most popular mechanism that has been invoked to give a Newtonian law for the mantle is the Nabarro–Herring one. Nabarro himself proposed this mechanism for creep in the Earth in his original paper (Nabarro 1948). The first quantitative calculations that used this mechanism to determine the viscosity of the mantle were made by Gordon (1965), who showed that the viscosity should vary over many orders of magnitude. The reason that the variation is so large is that the ratio T/T_m, where T is the temperature and T_m is the melting (solidus) temperature, at a given depth in the mantle presumably varies from a value near 1 in the low viscosity channel to a value of the order of $\frac{1}{2}$ at the core–mantle boundary. The takeover of Coble creep at the lower values T/T_m reduces the variation of the predicted viscosity with depth but the variation is still in violent disagreement with the conclusion of a constant viscosity mantle. (From data in fig. 7 of Weertman & Weertman 1975, the variation of viscosity over a temperature range of

$\frac{1}{2} \leqslant T/T_m \leqslant 1$ for Nabarro–Herring–Coble creep for a grain size of 1 mm is a millionfold.) If superplastic creep exists in the mantle the predicted viscosity variation is just as large as that for Nabarro-Herring–Coble creep.

If a power law creep equation governs mantle flow the variation of 'effective' viscosity with depth is greatly reduced (Weertman 1970; Weertman & Weertman 1975). But this reduction is no help in explaining Cathles' and Peltier & Andrews' results because they conflict qualitatively with the surface displacements predicted with a power law creep equation. In fact, any diffusion controlled steady-state creep mechanism has great difficulty in giving an account of a Newtonian Earth with essentially a constant viscosity. Only in the very unlikely event that the ratio T/T_m is almost constant through the mantle is it possible to do so.

It would appear that we are at an impasse. Cathles's and Peltier & Andrews's conclusions about the Earth's viscosity and predictions made using steady-state creep equations just do not agree. A way out of the dilemma is to abandon, at least for the lower mantle, the steady-state creep equations.

The creep that occurs during glacial rebound is not necessarily a steady-state type of creep. It has already been mentioned that the total strain that occurs during rebound is of the order of 10^{-3}. The rebound data that Cathles used to obtain information about the viscosity of the lower mantle involve strains that actually are considerably smaller than 10^{-3}. For example, the measured uplift curves from regions far south of the edge of the Laurentide Ice Sheet (see Cathles 1975, pp. 226–228) have total displacements no larger than 10 m. For these data to be a sample of deformation of the lower mantle at depths of the order of 2000–3000 km implies that the total strain involved is only of the order of $10\,\text{m}/2500\,\text{km} = 4 \times 10^{-6}$. Strains of such small magnitude can hardly be considered to be produced in ordinary steady-state creep.

It seems reasonable to conclude that the viscosity of the Earth (at least of the lower mantle) as measured from glacial rebound is a measure of the transient creep properties of mantle rock, not a measure of the large strain, steady-state creep properties of mantle rock that determine the convective flow in the mantle and the motion of the plates.

It is possible for transient creep, over a limited range of strain, to be Newtonian. In Chalmers's original experiments on microcreep (Chalmers 1936) it was observed that at very small creep strains that the creep rate (of tin) is proportional to the stress. Thus microcreep is Newtonian. Moreover, the creep rate is much larger than the true steady-state creep rate measured at large creep strains. I have argued that other cases of Newtonian behaviour seen at small creep strain, including observations in ice, is Chalmers's microcreep phenomenon (Weertman 1967, 1969). (The explanation of microcreep is that, because the creep strains are so small, the dislocation density remains essentially constant during creep. If the dislocation velocity is proportional to the stress and the dislocation density remains constant the creep rate produced by the dislocation motion is proportional to the stress.)

Transient, Newtonian creep behaviour also may occur over a rather large range in strain. Harper & Dorn (1957) discovered in high-purity polycrystalline (3 mm grain size) aluminum that at low stresses ($\sigma \approx 2 \times 10^6 \mu$) the creep strain was proportional to time and the creep rate proportional to the stress for creep runs carried out to strains as large as 0.01. The creep rates were 3 orders of magnitude larger than that predicted from the Nabarro–Herring creep equation. Subsequent work (Harper, Shepard & Dorn 1958; Barrett, Muehleisen & Nix 1972; Muehleisen 1969; Mohamed, Murty & Morris 1975) has shown that Harper–Dorn creep occurs in single crystals of aluminium and in other metals and that it is not likely to be produced by

Nabarro–Herring subgrain creep. The status of our understanding of Harper–Dorn creep was reviewed recently by Mohamed *et al.* (1975).

It seems reasonably clear (Mohamed *et al.* 1975) that in Harper–Dorn creep the dislocation density is independent of stress. Application of a low stress will not immediately cause the dislocation density to change through dislocation multiplication. (In fact, the attainment of equilibrium dislocation density appropriate to the applied stress level might even require the annihilation of dislocations. For example, in unstressed crystal the equilibrium dislocation density is no dislocations at all. Yet dislocations exist in almost all crystals, regardless of how long they have been annealed at a high temperature.) A creep strain as large as 0.01, however, cannot be reached before appreciable dislocation multiplication is needed. The creep strain is equal to the product of the Burgers vector, the average distance a dislocation moves, and the total length of dislocation line per unit volume that moves during creep. For a dislocation density of $10^8\,\mathrm{m/m^3}$ ($10^4\,\mathrm{cm/cm^3}$) that have been reported for Harper–Dorn creep specimens and for an average dislocation displacement equal to the average spacing between dislocations (*ca.* 170 μm) the total creep strain would be only of the order of 1×10^{-5}. Thus dislocation multiplication is needed or much larger dislocation displacements are required to obtain a strain of 10^{-2}. Therefore, Harper–Dorn creep and Chalmers microcreep must be different. It has yet to be made clear in any paper so far published how the Newtonian behaviour of Harper–Dorn creep comes about.

Can Harper–Dorn creep – if it exists in rock – be important in the mantle? The activation energy of Harper–Dorn creep is that of self-diffusion. Therefore, the same objection can be raised against it as against Nabarro–Herring–Coble creep: if Harper–Dorn creep controls glacial rebound then the viscosity of the mantle should vary over many orders of magnitude. Since Chalmers's microcreep mechanism also involves diffusion mechanisms the objection raised against Harper–Dorn creep and Nabarro–Herring–Coble creep also can be raised against it.

In normal transient creep the creep rate decreases as the creep strain increases. Thus the effective viscosity in transient creep increases with increasing strain. Since the creep strain involved in the isostatic rebound phenomenon must decrease with depth in the mantle, this strain effect would cause a decrease in the effective viscosity with increasing depth in the lower mantle. This decrease might, therefore, compensate for the increase in effective viscosity at a fixed strain that occurs because the temperature ratio T/T_m decreases with depth in the lower mantle. Therefore, it is possible to see the possibility of an explanation of why the mantle has a constant viscosity if normal transient creep occurs during isostatic rebound.

Discussion

'The resistance arising from want of lubricity in parts of a fluid, is, other things being equal, proportional to the velocity with which the paths of the fluid are separated from one another' (Newton 1947).

Up to now the discussions about whether the mantle of the Earth lacks lubricity and is or is not Newtonian have not distinguished between large strain creep deformation (which occurs in convective motion and which requires at least one of many possible steady-state creep mechanisms to produce it) and small strain creep deformation such as occur during glacial rebound. It appears to be very important that this distinction be made. Cathles and Peltier & Andrews have shown that glacial rebound data can only be explained with a Newtonian mantle of essentially constant viscosity. It is impossible for steady-state creep mechanisms to account for this result. Since glacial rebound requires only relative small creep strains the viscosity values determined from rebound data undoubtedly are a measure of transient creep that occurs in the mantle rather than steady-state creep. Unfortunately transient creep has not been studied, either experimentally or theoretically, as extensively as has steady-state creep and we are not as well equipped to make comparisons between rebound data and experimental data and theories; but perhaps it will prove possible in the future to decide if the viscosity of the mantle as measured with rebound data can be explained with transient creep phenomenon.

The value of the viscosity determined in rebound measurements most probably sets a lower limit to the value of the effective viscosity of large strain flow. The larger the transient creep strain, and hence the shallower the depth in the mantle, the closer this limit will be to the true effective viscosity of large strain flow.

Whether the large strain, steady-state creep flow that occurs during convection is or is not Newtonian depends upon the value of the grain size of mantle rock. If the experimental results of Post (1973) and Carter and Mercier (Carter 1976) on the stress dependence of the grain size in dunite are at all representative of the stress dependence of the grain size of rock in the mantle it is extremely unlikely that the mantle is Newtonian at stress levels that are likely to occur during convective motion. When power law creep operates the effective viscosity of the mantle, as calculated under conditions of either constant stress or constant strain rate, varies over about 4–6 orders of magnitude (Weertman 1970; Weertman & Weertman 1975; Stocker & Ashby 1973). The effective viscosity in the lower mantle is likely to be several orders of magnitude larger than the glacial rebound value of the viscosity, but it still may be low enough to permit convection there.

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Discussion

J. WEERTMAN. D. C. Tozer pointed out in the discussion after the presentation of my paper (see also Tozer 1977) that convection in the mantle self-regulates the viscosity value of rock of the mantle. (That is, if the mantle were cold its viscosity would be very high, but radioactive heating in time would raise the temperature of the mantle and reduce the value of the viscosity. On the other hand, if the temperature were very high, the viscosity of the mantle would be very low, but the resultant fast convective motion within the mantle would reduce the mantle temperature and increase the value of the viscosity of the mantle rock.) He proposed that the viscosity of the mantle is a constant throughout because of this self-regulation.

I agree with him that the mantle self-regulates to a large extent the value of its viscosity, and point out that if power law creep is the important creep law for the mantle that the mantle can even self-regulate the value of its effective viscosity by changing the stress level without even changing the mantle temperature. In my paper (Weertman 1970), in which the importance for the mantle of power law creep was stressed, it was pointed out that the variation with depth of effective viscosity is smaller for power law creep than for Newtonian creep. However, self-regulation is no guarantee that the effective viscosity of the mantle will have a constant value everywhere. It is only necessary to note the obvious fact that self-regulation has not kept the viscosity of the outer core of the Earth from being numerous orders of magnitude smaller and the effective viscosity of the Earth's crust from being orders of magnitude larger than that of the mantle.

I am not convinced that Tozer's self-regulation mechanism is the explanation of the constant viscosity values that come out of the glacial rebound analysis. Convective motion within the mantle drives the temperature profile of the mantle towards an adiabatic temperature profile. The pressure gradient of an adiabat in the mantle presumably is smaller than the pressure gradient of a melting temperature curve (Jacobs, Russell & Wilson 1974). (However, in the liquid outer core of the earth Kennedy & Higgins (1973) believe that the adiabat gradient is larger than the melting point gradient.) Thus the temperature profile within the mantle should lie between the melting temperature profile and an adiabat. If there is no zone of appreciable thickness within the mantle in which the actual temperature is above the solidus temperature then the ratio T/T_m (where T_m is the solidus temperature and T is the actual temperature) will be smaller the greater the depth within the mantle. Thus the effective viscosity at a fixed strain rate or stress should increase with depth in the lower mantle.

Even if the actual temperature of the mantle is close to the melting point everywhere, as suggested by Kennedy & Higgins (1972) and by Stacey (1975), there is another reason why the mantle viscosity may increase with depth in the lower mantle. The viscosity is inversely proportional to the diffusion coefficient D. The diffusion coefficient D depends upon the melting temperature of the material by the empirical equation (3), $D = D_0 \exp(-gT_m/T)$, of the text. (For magnesium-rich olivine–forsterite of a melting temperature 2163 K and a diffusion activation energy of 565 kJ/mol the value of g is 31.5.) In solid solutions the solidus temperature is a reasonable temperature to use for T_m in equation (3) because it is the temperature at which a solid of a given composition begins to melt as the temperature is raised. The liquidus temperature, on the other hand, gives the temperature at which a solid of a *different* composition begins to freeze out of a melt as the temperature is lowered. The only available test of equation (3) for metal alloys which form single-phase solid-solutions below the solidus temperature (Birchenall 1951; Shewmon 1963) shows that the diffusion coefficient varies with change in composition in the same way that the solidus temperature does. (However, in the alloy systems studied the liquidus temperatures were close to the solidus temperatures and hence it is not possible to decide if the solidus or the liquidus temperature is the better temperature to use for T_m.) In the lower mantle it has been suggested that olivine, which has transformed to a spinel structure at intermediate depths, transforms into periclase (MgO) and stishovite (SiO_2) (Kennedy & Higgins 1972). More recently it has been proposed that the transformation is into periclase and a perovskite phase ($MgSiO_3$) (Liu 1976*a*, *b*; Ito 1977). Suppose that the creep properties of the lower mantle are determined by the creep properties of the periclase mineral. Suppose further that the actual temperature in the lower mantle is near an eutectic temperature of a two or more phase mixture, as suggested by Kennedy & Higgins (1972). What is the melting temperature T_m to be used in equation (3)? It is not the eutectic temperature. Consider figure 5 which shows a schematic phase diagram for an eutectic system. Suppose periclase is the α phase indicated in the figure. If the actual temperature T is somewhat below the eutectic temperature, as indicated in figure 5, the α phase in a two-phase mixture has a composition that is different from what it would have at the eutectic temperature. The solidus temperature of the α phase of a composition at temperature T in a two-phase mixture is that indicated by the temperature T_m in figure 5. The solidus temperature T_m can be considerably higher than the eutectic temperature. It is the most reasonable temperature to use in equation (3) for the melting temperature of the α phase of a two-phase mixture. It is the melting temperature of the α phase of a composition that is different from the composition the α phase has at the eutectic temperature. In other words, the melting temperature to use in equation (3) for a two-phase mixture will itself depend on the actual temperature because the composition of the phases varies with temperature. Therefore, even if the actual temperature is close to the eutectic temperature it may not be close to the appropriate melting temperature T_m. Only if the actual temperature is exactly equal to the eutectic temperature is T_m equal to the eutectic temperature. Newton, Jayaraman & Newton (1962) have pointed out that an eutectic trough deepens with increasing pressure. That is, the melting temperature of the end members of an eutectic system increases much more rapidly with increasing pressure than does the eutectic temperature. Hence even if the mantle were everywhere slightly below an eutectic temperature the ratio T/T_m could decrease with increasing depth in the mantle.

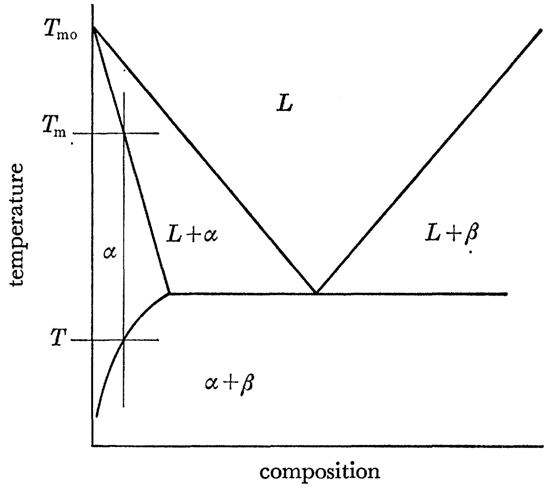

FIGURE 5. Schematic phase diagram for an eutectic system.

It should also have been pointed out in the text that an increase of an order of magnitude in the viscosity will be produced in the lower mantle (for power law creep) because the elastic constants there increase by a factor of 3. The transformation into denser phases also can produce a decrease of one to two orders of magnitude in the diffusion coefficient and thus a comparable increase in the viscosity (Shewmon 1963).

In an earlier paper (Weertman 1970) I had suggested that if the glacial rebound stresses are superimposed on the stresses that are produced during convection in the mantle and if the rebound stresses are small compared with the later one, then the rebound phenomenon will be pseudo-Newtonian although the creep rate is determined by the power law creep equations. I still believe this possibility. However, if the effective viscosity for convection in the lower mantle is, say, 10^{24} P, it would not be possible for the pseudo-Newtonian viscosity for the rebound phenomenon to be lower than this value (in steady-state creep).

It also should have been pointed out in the paper that data on the recrystallized grain size in olivine that was considered by Kohlstedt, Geotze & Durham (1976) also is described by equation (8). Sellars and co-workers (references given in Nicolas & Poirier 1976) have found in metals that the recrystallized grain size also is proportional to $\sigma^{-\frac{3}{2}}$.

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S. White (*Department of Geology, Royal School of Mines, Imperial College, London SW7 2BP*). Phase transformations are often overlooked when mantle deformation is considered, although they are among the least speculative of mantle phenomena. There is a growing volume of literature associating deep seated earthquakes with phase transformations (see Sung & Burns 1976). However, these earthquakes are directly related to the volume changes associated with the phase transformation. There is another process associated with transformations, namely transformation enhanced ductility which is termed 'transformation superplasticity' in the metallurgical literature. Greenwood & Johnston (1965) showed that volume changes associated with phase changes in metals can lead to the generation of appreciable stresses across the phase boundary. These stresses lower the applied stress required to induce flow in the weaker phase. In some instances the stresses may exceed the yield stress and consequently may induce flow in the absence of an applied stress. Transformation enhanced ductility has also been observed in a variety of ceramics and minerals (Edington, Melton & Cutler 1976).

Pressure induced phase transformations appear to be common in the Earth's mantle (Akimoto, Matsui & Syona 1975). A major zone of transformation occurs at a depth of 400 km where olivine transforms to a spinel structure. It is reasonable to assume that appreciable stresses, especially when compared to the 10 MPa differential stress thought to be responsible for most mantle deformations, develop in the above zone. Consequently transformation enhanced ductility should be associated with transformation zones and rocks moving through these should exhibit an abnormal weakness. This can lead to homogeneous deformation in the deep mantle.

Normally, only small strains are induced by a transformation in metals, and cycling through the transformation is necessary to obtain large strains. However, the strains associated with mantle phase transformations may be greater because of the very slow, continuous passing of rock through the transformation. Consequently transformation enhanced ductility may make a significant contribution to mantle deformation processes.

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J. Weertman. Transformational superplasticity certainly is a possible deformation mode in those parts of the mantle in which phase changes occur. I would doubt that large plastic strains are produced because of the need to cycle back and forth through the phase transformation to obtain this type of superplasticity. However, for the rebound phenomenon, which involves only small plastic strains, it might be an important deformation mechanism.

Alan H. Cook, F.R.S. (*Department of Physics, Cavendish Laboratory, Madingley Road, Cambridge CB3 0HE*). It is generally agreed that a series of polymorphic changes occurs in minerals in the upper mantle. Olivine, in particular, is known to change to a so-called β-phase, then to a spinel structure and finally, probably, to a mixture of high pressure forms of the constituent oxides. Could Professor Weertman say how such changes will affect creep rates within the upper mantle?

J. Weertman. A phase change to a more close packed phase is known to decrease the diffusion coefficient and the creep rate of metals by one to two orders of magnitude (Shewmon 1963; Sherby & Burke 1967). I would expect that a similar decrease in the magnitude of the creep rate would occur for polymorphic changes into denser phases in the mantle. The effect of the phase change could also be accounted for by a change in the melting temperature T_m in equation (3).

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