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The Initiation of Rifting at Constant Tectonic Force: Role of Diffusion Creep

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We have investigated the effect of diffusion creep on lithospheric extension using a one-dimensional numerical model that assumes a constant force is available to drive extension. The model is motivated by the fact that continental areas with average heat flow should be too strong to rift using standard estimates of lithospheric strength. However, such areas do rift and diffusion creep is a mechanism by which lithosphere may deform at a lower stress level than is required for the usually assumed dislocation creep. We consider the evolution of strain rate and temperature at the center of an idealized pure shear rift. The strain rate will either increase or decrease with time depending on whether lithospheric weakening or strengthening dominates as extension progresses. Given an initial thermal condition and an assumed lithospheric rheology, the applied force must be such that the initial strain rate is greater than some critical value for weakening to dominate. The force at which this condition is met we will term the critical force. Diffusion creep is more efficient at smaller grain sizes and the model results indicate that mantle grain sizes would have to be less than 1 mm for diffusion creep to significantly reduce the critical force. We speculate that during the initial stages of continental rifting, grain size reduction and diffusion creep deformation mechanisms may have an important effect on lithospheric strength.

INTRODUCTION

Treating the lithosphere as having a few layers with simple rheological properties has been a successful approach for predicting a wide variety of phenomena in many tectonic settings [*Kirby and Kronenberg*, 1987]. Most models describe brittle deformation using Byerlee's law, which is independent of rock type [e.g., *Goetze and Evans*, 1979; *Brace and Kohlstedt*, 1980]. Plastic deformation is usually modeled as steady state flow controlled by dislocation creep, which is independent of mineral grain size. Because dislocation creep is strongly temperature dependent, the depth to the brittle/plastic transition should be controlled by the thermal structure of the lithosphere. *Meissner and Strehlau* [1982] demonstrated this by showing that the depth of seismogenic faulting correlates with the inverse of the heat flow in many tectonic settings. Another early success in using simple rheological models was explaining the flexural response of the lithosphere to applied loads, especially as a function of oceanic plate thermal structure [*Bodine et al.*, 1981; *McNutt*, 1987]. *Bird* [1989] included a simple description of rheology to model compression in western North America. In continental areas undergoing extension, models using simple rheologies have been used to estimate how the heat flow and crustal thickness affect the pattern of extensional strain [*Sonder*, 1986; *Houseman and England*, 1986; *Sonder et al.*, 1987; *Kusznir and Park*, 1987; *Buck*, 1991].

There are several regions where there is evidence that deformation occurs at a lower stress level than predicted by simple rheological models, however. The most well known example is the San Andreas fault, where stress measurements indicate that the maximum principle stress is almost orthogonal to the fault plane so that the resolved shear stress on the fault is much lower than what should be necessary to cause slip [*Mount and Suppe*, 1987; *Zoback et al.*, 1987]. In addition, the heat flow directly over the fault is much lower than expected, implying that little frictional heating occurs [*Lachenbruch and McGarr*, 1990]. This indicates that the fault is much weaker than one would expect based on standard rheological models. Similarly, in extensional settings, areas of low-average heat flow should have enough mechanical strength to resist extension and thinning unless the tectonic driving forces are very large (Figure 1). For example, if the surface heat flow is less than 60 mW/m^2, the integrated strength of the lithosphere is greater than 20 TN/m (TN $\equiv 10^{12}$ newtons). This is much greater than estimates of the forces available to drive plate tectonics. Forces that could drive extension, such as slab pull or trench suction, are 1-4 TN/m and ridge push forces are approximately 3-5 TN/m [*Bott*, 1991; *Bott et al.*, 1989; *Forsyth and Uyeda*, 1975]. If the surface heat flow away from the zone of active continental extension (that is, beyond the rift flanks) is indicative of the thermal condition at the onset of rifting, then it is clear that cold areas do rift. The background heat flow adjacent to the Kenya rift, East Africa is 35-45 mW/m^2, while the background heat flow adjacent to the Baikal Rift is 35-50 mW/m^2 [*Morgan*, 1982]. Thus in cold continental areas, either tectonic forces associated with rifting are much greater than is currently believed or cold lithosphere is much weaker than conventional models would imply.

In this paper we consider the temporal evolution of a simple one-dimensional model of lithospheric extension that assumes the tectonic driving force is constant. We show that if the applied force is less than some critical value, acceleration of thinning and continued extension does not occur. We examine the effect of rheology on the critical force and propose one mechanism which may allow cold areas to deform: a local reduction in grain size. If tectonic grain size reduction can occur, then it may be possible for diffusion creep to become important and deformation could occur at much lower stress levels than would be possible by dislocation creep alone. Many have suggested that diffusion creep could be an important mechanism in mantle deformation [*Ranalli*, 1984; *Ranalli and Fischer*, 1984; *Karato et al.*, 1986; *Kirby and Kronenberg*, 1987]. However, the effect of this mechanism on geodynamical models has received relatively little attention. We look at the grain sizes that would be required for cold

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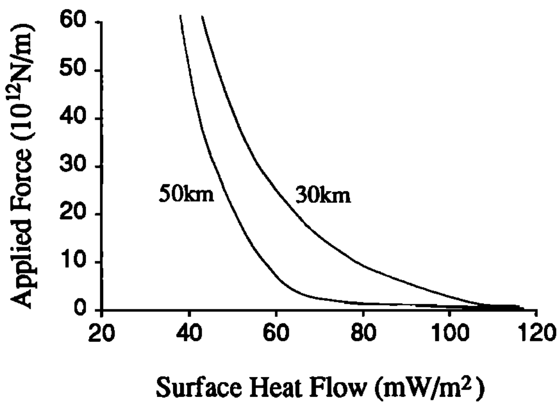

Fig. 1. The applied force required to get large-scale lithospheric extension for different crustal thicknesses as a function of surface heat flow. Under cold conditions (<50 mW/m^2) the integrated strength of the lithosphere is quite large (>20 TN/m) even for thick crust. From Buck [1991].

lithosphere to rift at reasonable levels of tectonic force and in geologically reasonable lengths of time. Observational evidence implying diffusion creep in the mantle has been summarized by *Karato and Wu* [1993]. We emphasize that the inclusion of diffusion creep does not exclude the importance of dislocation creep in mantle deformation. Rather, we wish to examine the conditions under which diffusion creep should be rate-controlling.

LITHOSPHERIC RHEOLOGY AND STRENGTH

Plastic Rheology

Steady state flow of materials by plastic creep is a thermally activated process and can be expressed as an Arrhenius relation of the form

$$\dot{e} = A\,\sigma^n d^{-m} \exp\left[\frac{-Q}{RT}\right] \quad , \qquad (1)$$

where $\dot{e}$ is the strain rate, σ is the stress, Q is an activation energy of the particular deformation mechanism, R is the universal gas constant, T is the temperature in Kelvins, d is the mineral grain size, and A is a constant. For determining the yield stress as a function of depth, this is rewritten as

$$\sigma = \left(\frac{\dot{e}}{A}\right)^{1/n} d^{\,m/n} \exp\left[\frac{Q}{nRT}\right] \quad . \qquad (2)$$

A, Q, n, and m are established experimentally leaving the strain rate, temperature, and grain size to be specified. If we assume a temperature profile for the lithosphere and a mineral grain size, the stress is only a function of strain rate.

The experimentally determined parameters (A, Q, n, and m) are dependent on the micromechanism of flow. It is generally assumed that under any given temperature and applied stress conditions, one mechanism will be more efficient than any other and will thus be rate-controlling. Because estimated lower crustal and upper mantle conditions usually fall in the dislocation creep field, nearly all rheological models of the lithosphere use dislocation creep flow laws [e.g., Brace and Kohlstedt, 1980; Sonder, 1986; Sonder et al., 1987; Kusznir and Park, 1987; Buck, 1991; Bird, 1989]. Others, however, have argued that there may be some conditions where diffusion creep could have an important effect on lithospheric strength [Karato, 1984; Karato et al., 1986; Rutter and Brodie, 1988]. Here, we will examine the conditions under which diffusion creep mechanisms could play an important role in lithospheric extension. There are several differences between dislocation creep and diffusion creep that are important from the geophysical modeling standpoint. In dislocation creep, the stress required for flow is less sensitive to strain rate than is diffusion creep. Also, dislocation creep is independent of the mineral grain size, whereas diffusion creep is not.

Theoretical treatments of dislocation creep show that the stress exponent should be large and is $n = 3$. A reduction in strain rate therefore has little effect on reducing the yield stress. Reducing the strain rate by three orders of magnitude will only reduce the yield stress by a single order of magnitude. Extensive laboratory experiments have verified that many rock-forming minerals have such high stress exponents ($n \geq 3$) when dislocation creep appears to be the dominant deformation mechanism [Kirby and Kronenberg, 1987]. In contrast, diffusion creep should have a small stress exponent of $n = 1$. It is therefore much more sensitive to changes in strain rate as shown in Figure 2*a*. Under slowly straining conditions, the yield stress should be quite small. This has been verified in the limited amount of laboratory work done on this deformation mechanism [Karato et al., 1986].

Dislocation creep depends primarily on the crystalline structure of the material being deformed. The deformation rate is independent of the mineral grain size so that $m = 0$ in equation (2). In contrast, diffusion creep is very sensitive to the mineral grain size (see Figure 2). Experimental evidence from Karato et al. [1986] shows that the grain size dependence varies with the amount of water present. For wet samples, they found that $m \approx 3$ and for dry samples that $m \approx 2$. In all cases, $m > 1$ so that diffusion creep becomes more efficient at smaller grain sizes.

If the strain rate is low and the grain size is small, then diffusion creep may be the rate-controlling flow mechanism. The effect of diffusion creep on a yield strength envelope is demonstrated in Figure 2*b*. For conditions of low to average heat flow, it is normally assumed that the mantle controls the strength evolution of the lithosphere [e.g., England, 1983]. In our model calculations, we also considered the effect of changing the crustal rheology. We found, however, that including pressure solution of quartz in the crust changed the overall results by only a few percent. Therefore, in the remainder of the paper we focus exclusively on varying the rheology of the olivine mantle. One of the central questions that will be addressed is the grain size that would be required for diffusion creep to be an important deformation mechanism in geophysical models. The parameters used for the different creep laws in our model calculations are listed in Table 1 and are taken from Chopra and Paterson [1981], Karato et al. [1986], and Koch [1983].

A wide range of rheologic parameters has been reported for dislocation creep of olivine [e.g., Kirby and Kronenberg, 1987]. We use the results for the relatively strong Aheim dunite given by Chopra and Paterson [1981]. Chopra and Paterson [1981] report that for the weaker Anita Bay dunite, accommodation processes at grain boundaries has occurred and that some of the weakness in those samples is probably due to enhanced diffusion along grain boundaries. Thus, some component of diffusion creep seems to have occurred in the Anita Bay dunite, but not in the Aheim dunite. In the experiments of Karato et al. [1986], a lower stress exponent (3-

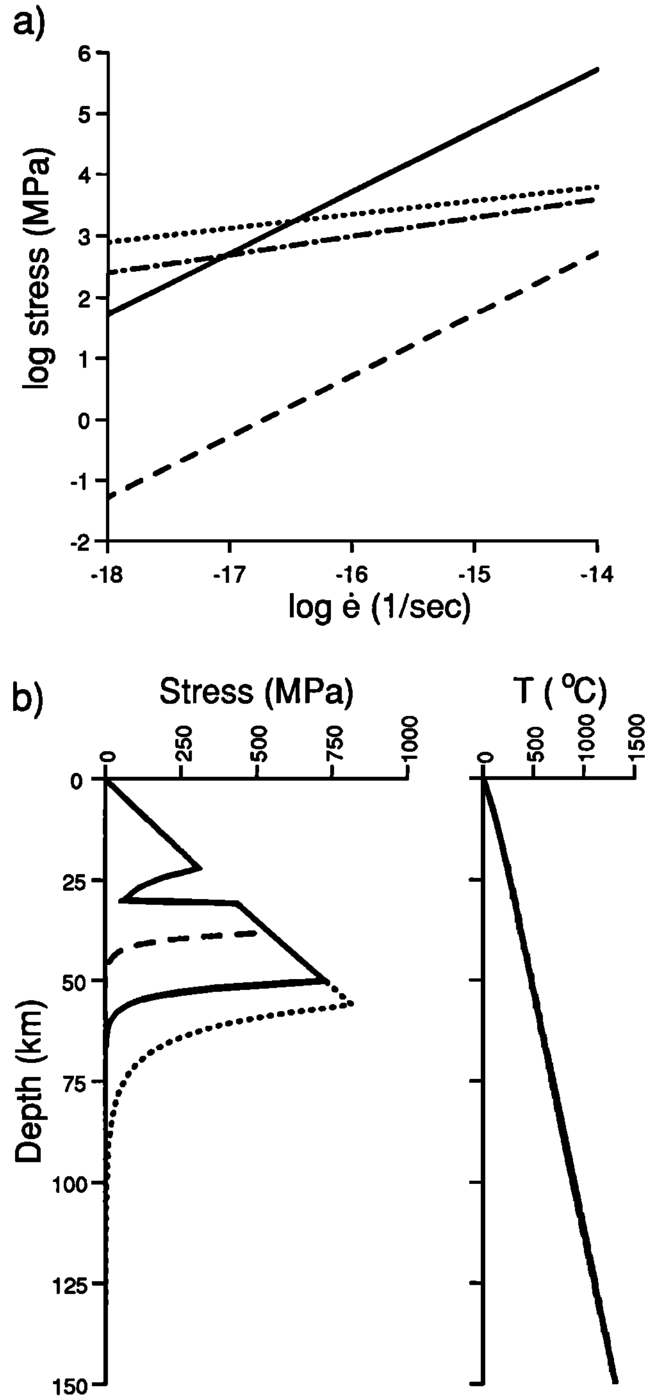

Fig. 2. (*a*) Plot of stress and strain rate for olivine rheologies at 500°C. A solid line denotes diffusion creep using a grain size of 1 mm; a dashed line denotes diffusion creep at a grain size of 0.1 mm.; a dotted line denotes dislocation creep of Aheim dunite; and a dot-dashed line denotes dislocation creep of Anita Bay dunite. All rheologic parameters are given in Table 1, except for the Anita Bay dunite which are A = 9550 $s^{-1}Mpa^{-n}mm^{m}$, n = 3.5, m = 0, and Q = 444 kJ/mol. (*b*) The effect of different olivine rheologies in the mantle on a yield strength envelope. The thermal structure is shown to the right (surface heat flow is 45 mW/m^2), and the strain rate is assumed to be 1×10^{-17} s^{-1}. The solid line here denotes diffusion creep using a grain size of 1 mm; the dashed line denotes diffusion creep at a grain size of 0.1 mm; and the dotted line denotes dislocation creep of Aheim dunite.

3.5) was determined for the dislocation creep regime than what we use here. They too report significant grain boundary migration and could not rule out grain boundary accommodation processes, although they dispute that the weakening is due to intergranular phenomena. Since one of the primary objectives of this study is to treat the deformation mechanisms separately, we use the Aheim dunite results. This is not too important a point, however, because the difference in rheological parameters between the two dunites is small compared to the differences between dislocation creep and diffusion creep (as seen in Figure 2*a*). Because of the inherent uncertainties in extrapolating experiments to mantle conditions and the difficulty in clearly distinguishing between different creep mechanisms in the constituitive relations determined in the laboratory, appropriate cautions should be taken in interpreting the model results.

Brittle Rheology and Total Strength

We assume that brittle rheology of the lithosphere is controlled by Byerlee's friction law [Byerlee, 1968]. If fractures exist throughout the lithosphere in all orientations, then as soon as enough stress is applied to overcome friction on optimally oriented fractures, slip will occur. In the calculations we use the same formulation for Byerlee's law as that used by Brace and Kohlstedt [1980]. Brace and Kohlstedt [1980] assume a constant lithostatic pressure gradient of 25 MPa/km, whereas we calculate the vertical stress explicitly as $\sigma_I = g \int \rho(z)\, dz$ for an assumed density structure with depth. We further assume hydrostatic pore pressure in calculating the stress differences. The model parameters used are listed in Table 2.

The total strength of the lithosphere can be defined as the minimum applied force necessary to cause large-scale deformation. Assuming that at any depth a rock will deform at a stress level corresponding to the weakest possible mechanism (including both brittle and plastic rheologies), it is possible to define a yield stress envelope with depth, $\sigma_y(z)$ [Brace and Kohlstedt, 1980]. The lithospheric strength B is simply the integral over depth of the yield stress:

$$B = \int_0^{z_L} \sigma_y(z)\, dz \ , \tag{3}$$

where z_L is the depth to the base of the lithosphere.

CONSTANT FORCE MODEL FORMULATION

Consider pure shear thinning of the lithosphere in response to some applied tectonic force (Figure 3). We treat this as a one-dimensional problem and model the evolution of the center of the rift. Because plastic rheologies are strongly temperature dependent, the yield stress is dependent on the temperature profile of the lithosphere. Thus we consider the one-dimensional heat flow equation:

$$\frac{\partial T(z,t)}{\partial t} = \kappa \frac{\partial^2 T(z,t)}{\partial z^2} - v(z)\frac{\partial T(z,t)}{\partial z} + H(z) \ , \tag{4}$$

where $T(z,t)$ is the thermal profile of the lithosphere, $v(z)$ is the vertical velocity of material that upwells in response to tensional forces, κ is the thermal diffusivity and $H(z)$ represents heat sources in the lithosphere. For the initial condition, an analytic, steady state ($\partial T/\partial t = 0$) solution is found by specifying an upper boundary condition of 25°C and a heat flow across the base of the crust (the mantle heat flow), q_m. The depth to the base of the lithosphere is then defined as the

TABLE 1. Plastic Rheology Parameters

Mineral	A, s^{-1} MPa^{-n} mm^{m}	n	m	Q, kJ mol^{-1}
quartz*	5×10^{-6}	3	0	149
olivine - dislcation creep†	3.98×10^{2}	4.5	0	498
olivine - diffusion creep‡	1.5×10^{-3}	1	3	250 §

* Koch [1983].

† Chopra and Paterson [1981].

‡ Karato et al. [1986].

§ The activation energy for diffusion creep is not well constrained experimentally. It is generally assumed that the mechanism is rate limited by the slowest diffusing species. For olivine, the diffusion activation energies of Si and O are used. See Karato et al. [1986] and Karato and Wu [1993] for a more complete discussion and explanation.

TABLE 2. Model Parameters

Symbol	Name	Value
ρ_m	mantle density	3300 kg m^{-2}
ρ_c	crustal density	2800 kg m^{-2}
κ	thermal diffusivity	1×10^{-6} m^2 s^{-1}
H_s	surface heat production	2×10^{-6} W m^{-3}
h_r	heat production length scale	1×10^4 m
R	univeral gas constant	8.3143 J mol^{-1} K^{-1}
h_c	Initial crustal thickness	30 km

depth to the 1330°C isotherm and will depend on the mantle heat flow. During the time-dependent calculation, we use constant temperature boundary conditions of 25°C at the surface and 1330°C at the base of the lithosphere (i.e., constant T at a fixed depth determined by the initial condition). We solve the thermal equation by center differencing the diffusive terms and upwind differencing the advective terms.

In pure shear thinning, the vertical velocity is a linear function of depth and strain rate, $v(z) = z\dot{e}$. It is widely accepted that the decay of radioactive elements is the only important heat source to consider. Most of these elements are concentrated in the crust and their abundance decreases exponentially with depth [Birch et al., 1968; Lachenbruch, 1970; Sclater et al., 1980]. The heat source term is thus

$$
\begin{aligned}
H(z) &= H_s \exp\left[\frac{-z}{h_r}\right] & z < z_c \\
H(z) &= 0 & z > z_c
\end{aligned}
, \qquad (5)
$$

where H_s is the surface heat production, h_r is a length scale, and z_c is the depth to the base of the crust. In the formulation here, crustal heat production contributes 20 mW/m^2 to the total surface heat flow. In the discussion of the results, we generally refer to the mantle heat flow that was used as an initial condition. The total surface heat flow q_s is simply the sum of the mantle heat flow and the crustal heat production assuming that the thermal conductivity is constant throughout the lithosphere. Although in pure shear deformation the length scale h_r should change in proportion to the amount of strain that has accumulated, we ignore this in our model calculations. In general, the effect will be minor. In addition, we stop the calculation soon after the strain rates become geologically fast so that the overall amount of accumulated strain in the calculation is small (strain <25%). In models with a larger amount of strain, the effect should be included. The various parameters and thermal constants we use in the model calculation can be found in Table 2.

Previous one-dimensional extension models have examined how the strength of the lithosphere evolves with time, assuming that the extensional strain rate is constant [England, 1983; Kusznir and Park, 1987]. Here we will assume that the force available to drive extension is constant and determine how the strain rate evolves through time. We do not address the problem of where the driving force for extension comes from, as have some previous workers [e.g., Houseman and England, 1986], nor do we consider the effect of changes in

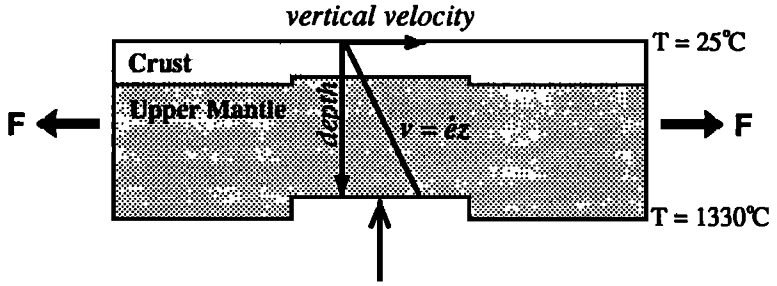

Fig. 3. Pure shear deformation assumed in the model formulation. The velocity is a simple function of strain rate and depth. The applied force F is assumed to be constant in the calculation. We consider the temporal evolution of temperature and strain rate at the center of the rift.

force (particularly gravitational potential energy changes) due to the deformation. We simply assume that force remains constant and consider the evolution of strain rate. We define a function,

$$F_a - B\left(t, \dot{e}\right) = 0\,, \qquad (6)$$

where F_a is the applied tectonic force and B is the integrated strength of the lithosphere at time t. Given the temperature structure at each time step, we calculate the integrated strength of the lithosphere at different strain rates until the above equation is satisfied. We can iterate to the correct solution of equation (6) using any numerical root-finding method. Here we used the secant method for two reasons. First, no information about the derivatives of the function is required. Second, a good initial guess to the correct strain rate can usually be provided so that convergence to the solution is rapid. The model is similar to that of Houseman and England [1986], who find a strain rate consistent with the magnitude of an assumed driving force for extension. Constant force calculations are also described by Christensen [1992], except that he uses Newton's method to find the root of equation (6) and by Takeshita and Yamaji [1990].

RESULTS AND DISCUSSION

By assuming that the regional tectonic driving force is constant, the strain rate of the deforming region will either increase or decrease with time depending on whether or not the area is getting weaker or stronger. As in all thermally dependent models, there is a trade-off between advective and conductive processes. When the strain rate is very small, then conduction dominates over advection in the heat flow equation (equation 4), and the temperature profile is little changed from the initial state. However, the crust eventually is thinned and is replaced with stronger mantle. The overall lithospheric strength increases, and the strain rate decreases in order to compensate. When the strain rate is large, then conduction cannot keep up with advection. If the rate of weakening due to advective thinning is greater than the strengthening due to crustal thinning and conductive cooling of the mantle, then the overall lithospheric strength decreases, and the strain rate will increase. The model has intuitive appeal because it is unlikely that an area experiencing extension will instantaneously begin deforming at high strain rates. Rather, it is more likely that an area will thin slowly at first and then accelerate to more rapid rates if advective weakening dominates. For example, in the Gulf of Suez there is strong evidence that extensional rates are very slow during the initial phase of rifting and is followed by a second phase of rapid extension [Steckler et al., 1988]. As another example, kinematic studies of rift propagation indicate that the area in front of a rift tip should have negligible deformation rates but must then increase as the rift propagates through the area [e.g., Hey et al., 1980; Courtillot, 1982].

The results are illustrated in Figure 4*a* where the strain rate as a function of time is plotted for a calculation assuming dislocation creep flow laws for olivine and an initial mantle heat flow of q_m = 24 mW/m^2 (equivalent to a surface heat flow of 44 mW/m^2). Curves are shown for different applied force levels. At small forces, the strain rate never increases, indicating that overall, the lithosphere is strengthening due to crustal thinning. Once enough force is applied, however, the strain rate always increases. In these cases, the increase is

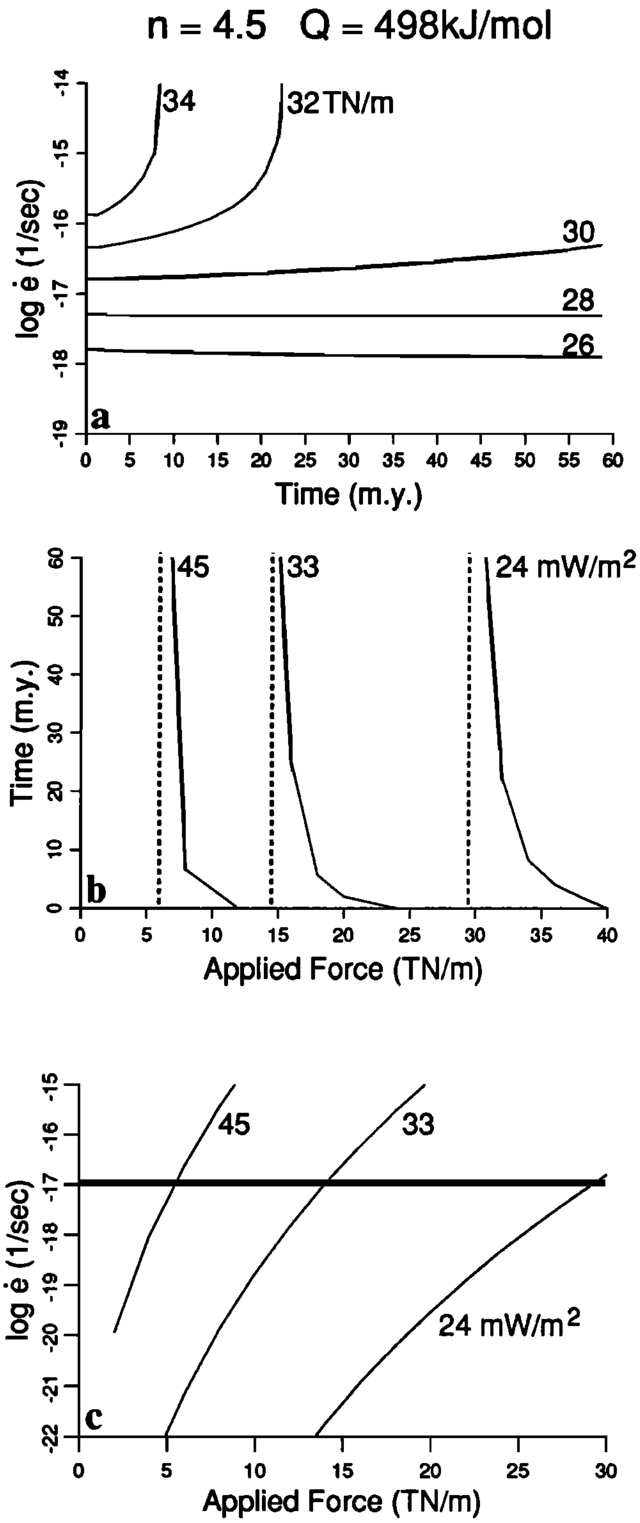

Fig. 4. (*a*) Strain rate as a function of time for different applied forces in TN/m. The assumed rheologic parameters are experimentally determined values for dislocation creep (Table 1). Below 29 TN/m, the strain rate does not accelerate. Inital mantle heat flow is 24 mW/m^2, which is equivalent to a surface heat flow of 44 mW/m^2. (*b*) Plot of the time required to accelerate to 1×10^{-14} s^{-1} for a given applied tectonic force at different initial mantle heat flows. The dashed lines represent the critical force F_c below which acceleration of strain rate does not occur. (*c*) The initial strain rate for a given applied force at different initial mantle heat flows. The critical force is where the curves intersect the critical strain rate, here approximated as a constant strain rate of 1×10^{-17} s^{-1}.

slow at first and then accelerates rapidly. As expected, there are two distinct modes of behavior depending on whether advective processes or conductive processes dominate. For a given set of initial conditions, if the applied force is such that the initial strain rate is greater than some critical value, weakening always occurs and the strain rate increases. If the initial strain rate is less than the critical value, strengthening dominates. The calculation of Takeshita and Yamaji [1990] shows a similar behavior to that just described. However, they did not consider the effect of varying the rheologic parameters and only considered two different initial conditions, both of which were very hot (>60 mW/m^2 total surface heat flow).

A simple scaling argument predicts that to first order, the critical strain rate depends on the inverse square of the depth to the brittle-ductile transition in the mantle, z_{bd}. Assume that the isotherm initally at the depth z_{bd} is advected upward at a velocity of $\dot{e}z_{bd}$. On the basis of analogy with half-space cooling, conduction should make the isotherm move down at a velocity porportional to κ/z_{bd}. An expression for the critical strain rate can be found by setting these velocities equal: $\dot{e}_{crit} = C\kappa/z_{bd}^2$, where C is a constant that is a function of the initial thermal condition. The depth z_{bd}, and hence the critical strain rate, depends on the initial thermal condition, the applied force, and the assumed rheology. For the range of conditions that we consider, the critical strain rate varies from $\approx 5 \times 10^{-18}$ to 8×10^{-17} s^{-1}. For simplicity, we ignore the variation of the critical strain rate due to the applied force and assumed rheology, and consider a constant value of $\dot{e}_{crit} = 1 \times 10^{-17}$ s^{-1}. Thus the magnitudes of the results presented below are not precise, but the relative relationships and ultimate conclusions remain the same. The applied force that yields an initial strain rate equal to the critical value we term the critical force F_c. At all forces greater than this, the initial strain rate is above the critical value and acceleration always occurs.

We demonstrate the critical force using two different approaches. In the first approach, we note that when a strain rate of 1×10^{-16} s^{-1} is reached, the strain rate-time curves begin to steepen and acceleration of thinning becomes rapid. A strain rate of 1×10^{-14} s^{-1} is equivalent to the upper range of strain rates estimated for rift formation [Martinez and Cochran, 1988] and we stop the calculation at this point. In Figure 4*b*, the applied tectonic force is plotted as a function of the time required to reach a strain rate of 1×10^{-14} s^{-1}. The curve flattens out and defines the critical force. In the second approach, we simply plot the initial strain rate as a function of applied force (Figure 4*c*). The critical force is where the initial strain rate exceeds 1×10^{-17} s^{-1} ($\dot{e}_{crit}$). The values of F_c obtained in this way agree to within a few TN/m to those obtained above. The discrepancy is primarily due to the fact that we have taken the critical strain rate to be independent of the applied force.

Figures 4*b* and 4*c* also show calculations at different initial mantle heat flow conditions. As one would expect, the colder the initial condition, the greater the critical force that is needed to get acceleration. For the calculation in Figure 4*a*, the initial mantle heat flow is 24 mW/m^2 and the critical force is large, 28 TN/m < F_c < 30 TN/m. Using a higher initial mantle heat flow of 45 mW/m^2 yields a smaller critical force $F_c \approx 7$ TN/m. A lower stress exponent (3 rather than 4.5) has the effect of reducing the critical force by a few TN/m.

To test the effect of varying the rheological parameters, we repeated the calculations shown in Figure 4 using the olivine diffusion creep flow laws in Table 1. Figure 5 shows the results for the same sets of conditions as those in Figure 4 using an olivine grain size of 1 mm. Comparison of Figures 4 and 5 shows that the effect of diffusion creep is to reduce the critical force required to get weakening at a given initial mantle heat flow. Using the assumed grain size of 1 mm, F_c for diffusion creep is two-thirds to three-fourths the value of F_c for

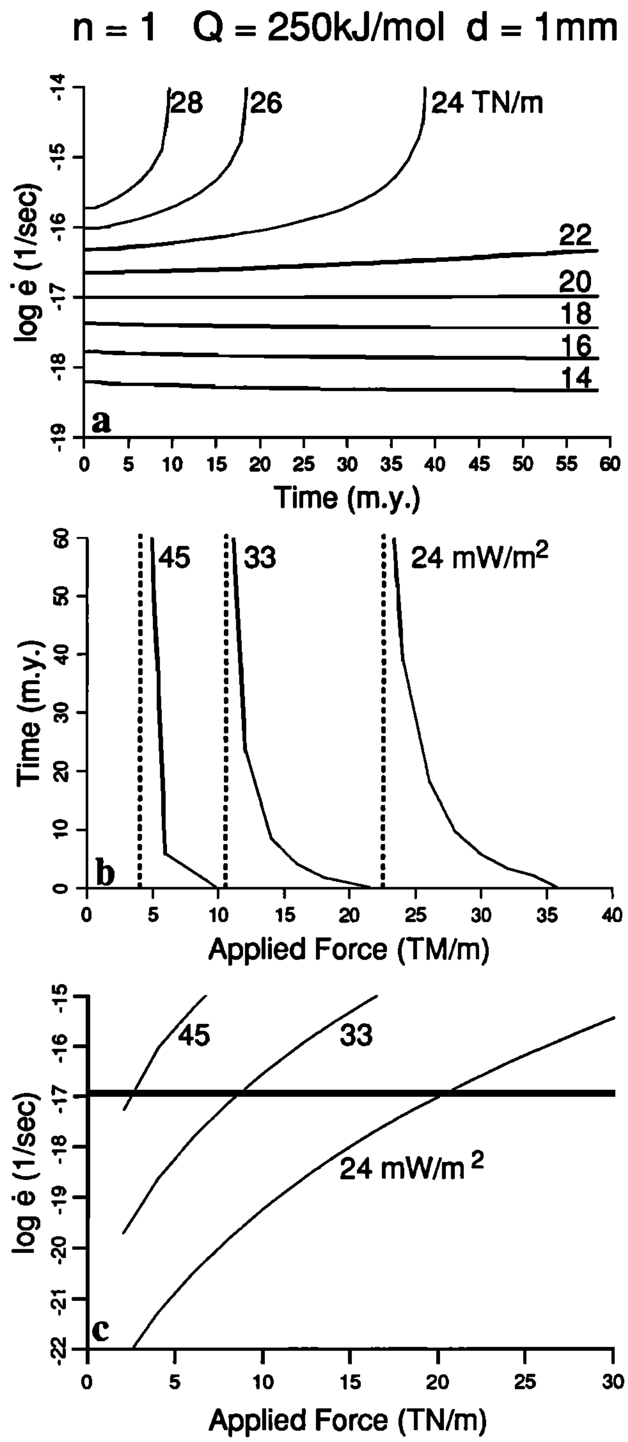

Fig. 5. The same calculations shown in Figure 4 using diffusion creep rheologic parameters instead of dislocation creep. The assumed olivine grain size is 1.0 mm. (*a*) Strain rate as a function of time for different force levels and an initial mantle heat flow of 24 mW/m^2. (*b*) Time required to accelerate to 1×10^{-14} s^{-1} as a function of applied force. (*c*) Initial strain rate as a function of applied force.

dislocation creep. As in all models of lithospheric deformation that use rheological parameters derived from laboratory experiments, some care must be taken in making such comparisons. The experimental data must be extrapolated over several orders of magnitude when using them in modeling exercises. Such extrapolations contain great uncertainties, but the important point is that under the low stress and low temperature conditions that are appropriate for some continental rifts, both mechanisms are likely to be of the same order of magnitude. Deformation models that consider either one or the other model are therefore incomplete. The full implications of multiple creep mechanisms on dynamic models has yet to be considered, and here we have shown that it could be very important.

Because diffusion creep is strongly dependent on mineral grain size, it is necessary to consider the effect of varying the grain size on the calculations. At smaller grain sizes, diffusion creep becomes very efficient and the critical force decreases accordingly. This is shown in Figure 6*a* by plotting the initial strain rates for grain sizes of 0.1 mm, 1 mm, and 5 mm using an initial mantle heat flow of 24 mW/m^2. For a grain size of 0.1 mm, the critical force is ≈10 TN/m (or one-third that required for dislocation creep under the same conditions). Figure 6*b* shows the same calculation using an initial mantle heat flow of 33 mW/m^2 and is included for comparison. The full numerical results are summarized in Figure 7, where the force required to accelerate to 1×10^{-14} s^{-1} in 20 m.y. and the critical force are plotted as a function of grain size.

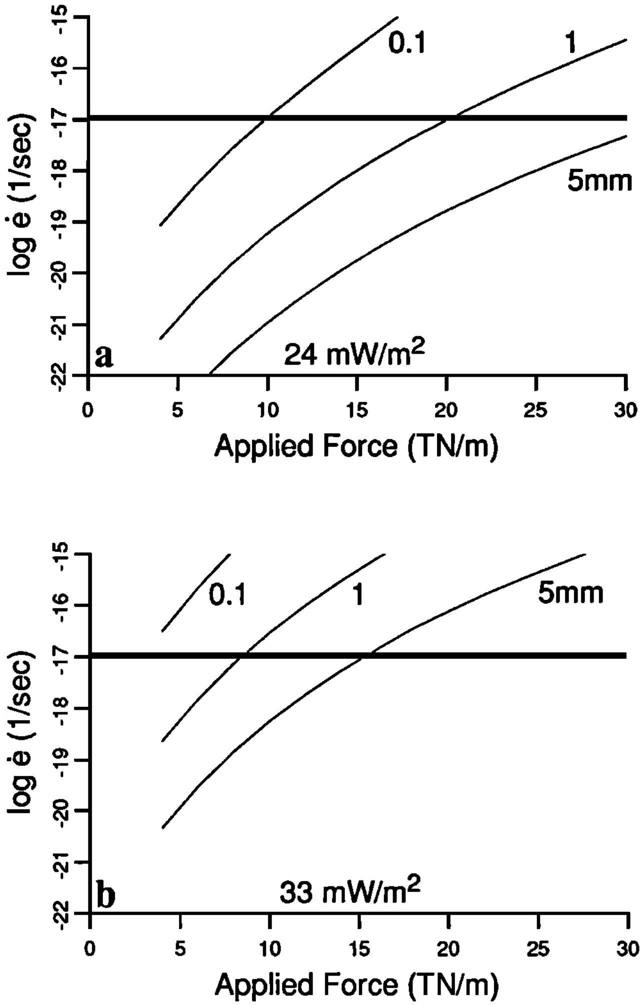

Fig. 6. Effect of grain size on the critical force level determined for diffusion creep at initial mantle heat flow values of (*a*) 24 mW/m^2 and (*b*) 33 mW/m^2. Plots are of the initial strain rate as a function of applied force. The critical force is approximately where the curves interesect 1×10^{-17} s^{-1}.

Thus it is clear that for diffusion creep to have a significant effect in controlling deformation rates during continental extension, grain sizes in the mantle would have to be very small. At a grain size of 1 to 2 mm, the critical force for dislocation creep and diffusion creep are roughly the same so that for diffusion creep to be the rate controlling mechanism, grain sizes less than 1 mm are required. Experimental evidence has shown that the recrystallized grain size of deformed olivine samples is inversely proportional to the applied deviatoric stress [Karato et al., 1980; Ross et al., 1980; Mercier et al., 1977]. The experiments of Karato et al. [1980] show that at stresses believed to be appropriate for the upper mantle under extension (10-500 MPa), grain sizes ranging from 0.5 mm to less than 0.01 mm would be expected. Ave-Lallement et al. [1980] further show that at depths of 50-60 km, grain sizes of mantle xenoliths taken from continental extension zones are on the order of 1 mm. Thus it may not be unreasonable to expect small grain sizes in areas experiencing tensional forces.

Another possibility is that some form of localized grain size reduction can occur that greatly weakens the integrated strength of the upper mantle. Etheridge and Wilkie [1979] and Rutter and Brodie [1988] discuss the implications of grain size reduction and the localization of strain into shear zones, where diffusion creep and grain boundary sliding mechanisms could be important. The scenario we envision is one in which rift propagation would play an important role. The stress concentrations in vicinity of the tip of a propagating rift could be sufficient that the area falls into the dislocation creep regime leading to recrystallization and smaller grain sizes. The smaller grain size allows for an enhanced weakening effect as diffusion creep can then operate. In turn, this permits rapid acceleration of rifting at much lower applied force levels. This is similar to the situation described by Karato and Wu [1993] and Karato [1992]. The most serious problem with this idea is that dynamic recrystallization usually requires relatively large amounts of strain (>40%)[Karato et al., 1980]. These strains are much larger than what is assumed in our simple one-dimensional model. Rather than having high strain over the entire lithosphere, however, it may be possible to get locally high-strain regions while the lithosphere as a whole strains only a little. The locally high-strain regions would become the shear zones as proposed by the above previous workers. These scenarios cannot be treated in the greatly simplified one-dimensional model presented and so the effect of grain size reduction and onset of diffusion creep necessarily remains speculative in the current context.

As a way to succinctly summarize the calculations, we repeated the calculations of Figures 4 and 5 over a wide range of applied force and initial mantle heat flow conditions and mapped out the conditions under which different mechanisms are important. In Figure 8, the critical force for different rheologies are shown. At forces less than critical, acceleration of rifting does not occur for a specified initial heat flow. Clearly, using well-established flow law parameters for dislocation creep, it is difficult to get the model lithosphere to

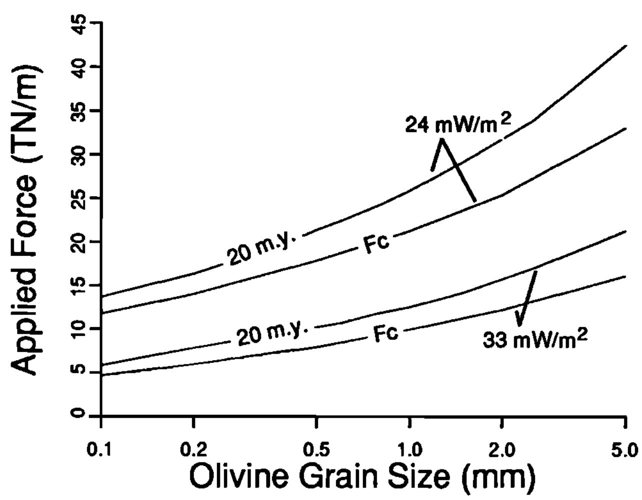

Fig. 7. The critical force as a function of olivine grain size assuming diffusion creep in the mantle and initial mantle heat flows of 24 mw/m^2 and 33 mW/m^2. The applied force required to accelerate to 1×10^{-14} s^{-1} in 20 m.y. as a function of olivine grain size is also shown.

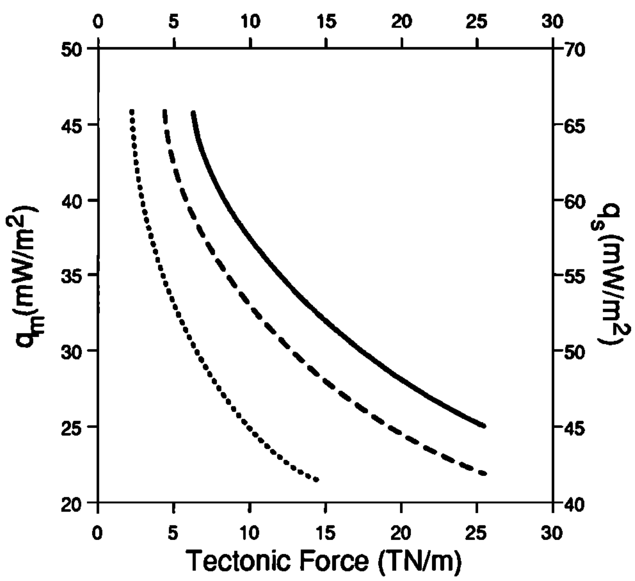

Fig. 8. Summary plot of a suite of calculations at different applied forces and mantle heat flows. The critical force is shown as a function of initial thermal condition. If the applied force is less than critical for a given initial heat flow, acceleration of thinning does not occur. The solid line denotes dislocation creep; the dashed line denotes diffusion creep using an olivine grain size of 1.0 mm; and the dotted line denotes diffusion creep using an olivine grain size of 0.1 mm.

rift at low force under cold conditions. Surface heat flows much greater than 70 mW/m^2 are required for an applied force of 5 TN/m if rifting is to accelerate. This would be an extremely hot initial condition for many areas of continental rifting. If the mineral grain size is small enough, however, it is possible to get acceleration of extension in short periods of time by including diffusion creep mechanisms.

We have only reported results for cases with an initial crustal thickness of 30 km. The crustal thickness in areas adjacent to several rifts is about this thick (e.g., the Gulf of Suez and the northern Red Sea) [Makris et al., 1981]. Increasing the initial crustal thickness will decrease the critical force, since the strong mantle is replaced by weak crust. For the initial thermal conditions we consider, the temperatures close to the crust-mantle boundary are such that the mantle yield strength is controlled by Byerlee's law. By assuming that the crust has negligible strength, we can estimate the upper limit to the reduction in critical force at different crustal thicknesses. Increasing the initial crustal thickness from 30 to 40 km reduces the critical force by about 5 TN/m.

We emphasize that including diffusion creep is not meant to imply that dislocation creep is unimportant. There is a continually growing body of evidence that demonstrates the importance of dislocation creep in mantle deformation [e.g., Kirby and Kronenberg, 1987]. Creep mechanisms may operate in parallel, and here we are primarily concerned with examining the possibility that during the early stages of extensional deformation, diffusion creep could be the rate-controlling mechanism. The constant force calculation shows that the initial deformation rates of an area under extension can be very small but still be sufficient to get rapid acceleration of rifting. The effect of rheology is to change the level of force required to achieve these strain rates and thus initiate the rifting process. We speculate that diffusion creep can operate at these small strain rates and low-stress conditions but that other mechanisms would become rate-controlling once thinning and heating accelerate to geologically reasonable rates.

CONCLUSIONS

Previous models of lithospheric extension suggest that continental areas should be quite strong under cold conditions. Assuming the rheological models used in these previous studies, the tectonic force that is required to rift cold areas is much greater than is thought to be available for rifting. Nonetheless, cold areas do rift. We have investigated a mechanism that may allow for such areas to rift at low force.

We have formulated a simple one-dimensional model for lithospheric extension assuming that the driving force is constant. The strain rate will increase or decrease with time depending on whether weakening due to advective thinning or strengthening due to conductive cooling is dominant. This is strongly dependent on the initial condition. If the applied force is such that the initial strain rate exceeds some critical value, acceleration of thinning always occurs. We have examined the effect of rheology on the critical force necessary to cause extension under a wide range of conditions.

By including the effects of diffusion creep for olivine in the upper mantle, we have considered the possibility that continental lithosphere could be much weaker than previous models have indicated. This would allow for the initiation of continental rifting at lower forces and lower temperatures than conventional rheological models, which only consider the effects of dislocation creep. The model calculations show that the grain size of olivine would have to be small for diffusion creep to be the rate-controlling deformation mechanism. Grain sizes in the mantle are poorly constrained, but there is some evidence that the small grain sizes required by the model may not be unreasonable in areas experiencing extension. In addition, it has been suggested that mineral grain size reduction occurs in localized lithospheric shear zones. However, this presents complications that cannot be treated easily in simple one-dimensional models. We speculate that during the initial stages of continental rifting, grain size reduction and diffusion creep deformation mechanisms may have an important effect on lithospheric strength.

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