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Strength of the lithosphere: Constraints imposed by laboratory experiments

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Abstract. The concept of strength envelopes, developed in the 1970s, allowed quantitative predictions of the strength of the lithosphere based on experimentally determined constitutive equations. Initial strength envelopes used an empirical relation for frictional sliding to describe deformation along brittle faults in the upper portion of the lithosphere and power law creep equations to estimate the plastic flow strength of rocks in the deeper part of the lithosphere. In the intervening decades, substantial progress has been made both in understanding the physical mechanisms involved in lithospheric deformation and in refining constitutive equations that describe these processes. The importance of a regime of semibrittle behavior is now recognized. Based on data from rocks without added pore fluids, the transition from brittle deformation to semibrittle flow can be estimated as the point at which the brittle fracture strength equals the peak stress to cause sliding. The transition from semibrittle deformation to plastic flow can be approximated as the stress at which the pressure exceeds the plastic flow strength. Current estimates of these stresses are on the order of a few hundred megapascals for relatively dry rocks. Knowledge of the stability of sliding along faults and of the onset of localization during brittle fracture has improved considerably. If the depth to the bottom of the seismogenic zone is determined by the transition to the stable frictional sliding regime, then that depth will be considerably more shallow than the depth of the transition to the plastic flow regime. Major questions concerning the strength of rocks remain. In particular, the effect of water on strength is critical to accurate predictions. Constitutive equations which include the effects of water fugacity and pore fluid pressure as well as temperature and strain rate are needed for both the brittle sliding and semibrittle flow regimes. Although the constitutive equations for dislocation creep and diffusional creep in single-phase aggregates are more robust, few data exist for plastic deformation in two-phase aggregates. Despite the fact that localization is ubiquitous in rocks deforming both in brittle and plastic regimes, only a limited amount of accurate experimental data are available to constrain predictions of this behavior. Accordingly, flow strengths now predicted from laboratory data probably overestimate the actual rock strength, perhaps by a significant amount. Still, the predictions are robust enough that uncertainties in geometry, mineralogy, loading conditions and thermodynamic state are probably the limiting factors in our understanding. Thus, experimentally determined rheologies can be applied to understand a broad range of topical problems in regional and global tectonics both on the Earth and on other planetary bodies.

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Introduction

About 20 years ago, several researchers estimated the upper limit for the strength of the lithosphere as a function of depth based on the results of laboratory measurements on the mechanical properties of rocks, noting that the strength of Earth cannot be greater than the strength of the rocks of which it is composed [*Sibson*, 1977; *Bird*, 1978; *Goetze and Evans*, 1979; *Brace and Kohlstedt*, 1980; *Kirby*, 1980]. These initial analyses divided the lithosphere into either two or three rheological regions. In the uppermost part where the temperature and lithostatic pressure are relatively low, frictional sliding on preexisting fractures governs mechanical behavior. At greater depth due to increasing temperature and pressure, plastic deformation controls the strength. Between these two regions, brittle and plastic processes interact in a transitional zone. The resulting profiles or strength envelopes plot maximum rock strength versus depth, as illustrated schematically in Figure 1 for oceanic and continental lithosphere composed of a layer of crustal rock overlying mantle rock. The essential character of these profiles is that for each lithologic unit, the strength increases with increasing depth until the temperature is high enough that plastic flow occurs at a lower stress than does frictional sliding. For the subcontinental upper mantle, only rarely will stresses be high enough for brittle processes to occur [e.g., *Chen and Molnar*, 1983]. The strength envelopes reflect the experimental observations that frictional strength increases approximately linearly with increasing pressure and is relatively insensitive to temperature, while plastic strength decreases rapidly with increasing temperature but is relatively insensitive to pressure.

In the late 1970s and early 1980s, strength profiles were used to analyze the mechanical behavior of the lithosphere for a variety of geologic settings. For example, *Sibson* [1977, 1980] noted that higher stress releases are expected at depths at which brittle processes give way to plastic flow. *Bird* [1978] noted that a thrust sheet might readily form by delamination of the lithosphere at the Moho, as the strength of the upper mantle is far greater than that of the lower crust (Figure 1b). *Goetze and Evans* [1979] applied laboratory results on rock rheology to estimate the stress distribution in a bending lithospheric plate near a subduction zone and concluded that stresses in excess of 100 MPa must be supported. *Kirby* [1980] arrived at the same conclusion based on a calculation of the strength distribution for a specific steady state temperature gradient. *Brace and Kohlstedt* [1980] concluded, based on a comparison of stresses determined from overcoring measurements made in deep mines [*McGarr and Gay*, 1978] combined with a strength profile derived from laboratory experiments, that the pore pressure in these regions of the continental crust must be at or above the hydrostatic level. *Molnar and Tapponnier* [1981] posited that deformation associated with the collision between India and Eurasia occurred primarily in regions having recently undergone tectonic activity, as such regions have higher temperatures and thus weaker rocks. *Chen and Molnar* [1983] discussed the presence of an aseismic lower crust between a seismic shallow crust and a seismic uppermost mantle, with a relatively low plastic strength for the lower crust (Figure 1b). Since the publication of these initial papers applying laboratory results to determine the strength of the lithosphere, numerous studies have focused on aspects of lithospheric deformation in a variety of local and regional tectonic settings both on Earth [e.g., *Molnar*, 1992; *Furlong*, 1993, and references therein] and on Venus [e.g., *Bindschadler and Parmentier*, 1986; *Zuber*, 1987; *Grimm and Solomon*, 1988; *Phillips*, 1990; *Kaula*, 1990; *Williams et al.*, 1994].

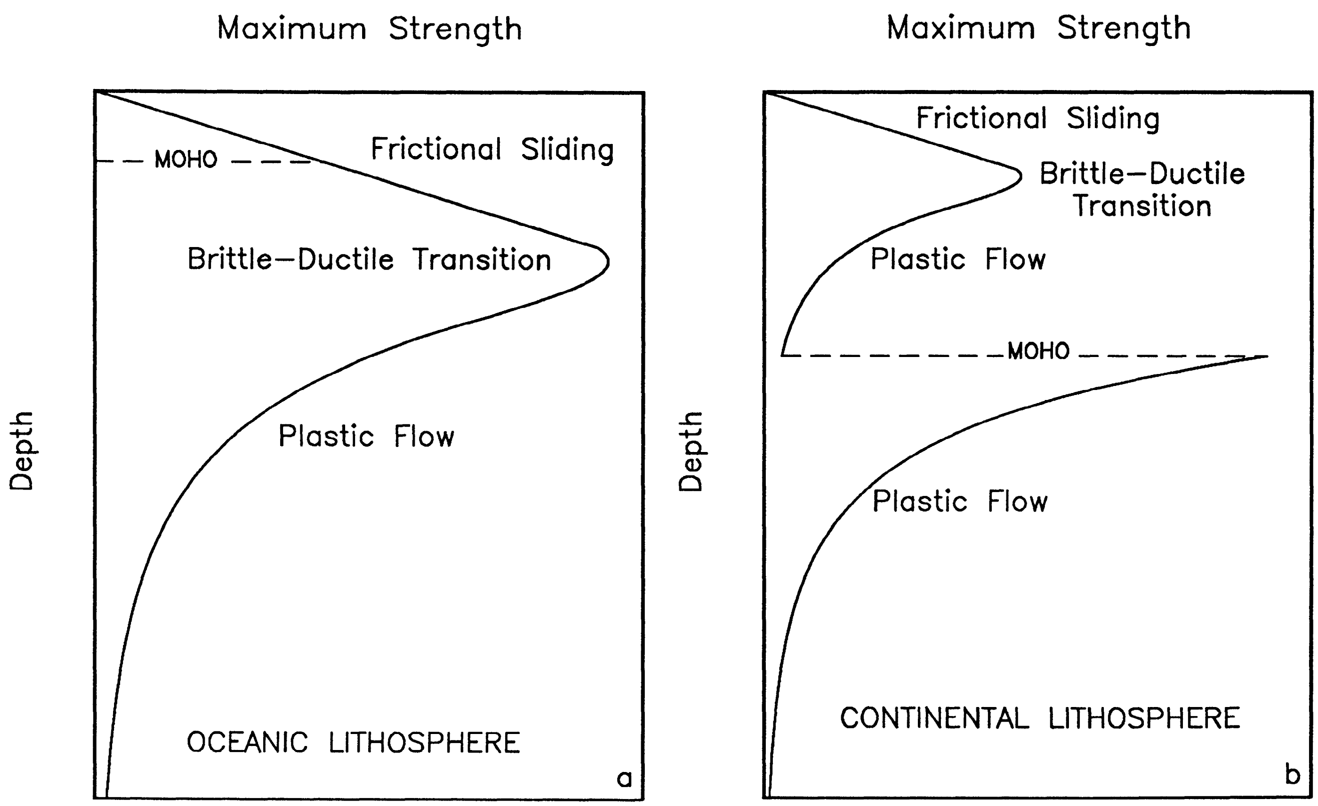

Figure 1. Schematic illustration of maximum rock strength as a function of depth for (a) the oceanic lithosphere and (b) the continental lithosphere.

In this paper, we first briefly review the principles learned from experimental rock mechanics that led to the development of strength profiles for the lithosphere. We then focus on the progress made since 1980 in constraining constitutive equations describing deformation of the lithosphere. Finally, we discuss some limitations and applications of strength profiles to understanding the dynamic behavior of the lithosphere. It is not our intent to provide a comprehensive list of rock deformation studies; indeed, several such surveys have recently been published [*Kirby*, 1983; *Carter and Tsenn*, 1987; *Tsenn and Carter*, 1987; *Kirby and Kronenberg*, 1987; *Evans and Dresen*, 1991; *Hickman*, 1991; *Evans and Kohlstedt*, 1995].

Constitutive Equations

Observations of the microstructures of rocks deformed naturally at low temperatures suggest that a number of deformation mechanisms are possible: granular processes including grain sliding and rotation, cataclastic processes including fracture and frictional sliding, as well as crystal plastic processes including twinning, dislocation creep, and pressure solution (for a review, see *Groshong* [1988]). Field observations, microstructural studies of naturally deformed rocks, laboratory experiments, and theories of material strength all suggest that as temperature and pressure increase, deformation involves increasing amounts of crystal plasticity, diffusional flow, recrystallization, and flow of partially molten material. Fracture and frictional sliding tend to be suppressed, although both could occur even at very high temperatures, provided that the effective pressure (the difference between the confining pressure and the pore pressure) is reduced by increasing fluid pressure [e.g., *Brace*, 1972; *Dell'Angelo and Tullis*, 1988; *Davidson et al.*, 1994].

In addition to changes in the partitioning of strain among different deformation mechanisms, the macroscopic appearance of failed rocks changes profoundly with increasing temperature and pressure. Fault rocks may be characterized in three broad groups, incoherent breccia, coherent cataclasites, and mylonites, based on the amount of cohesion and the fabric [*Sibson*, 1977]. Laboratory samples show a similar change in failure mode, tending to form localized shear zones at low temperatures and pressures but exhibiting distributed permanent strain at higher pressures and temperatures (for reviews, see *Heard* [1960]; *Ross and Lewis* [1989]; *Evans et al.* [1990]; *Wong* [1990]; *Tullis and Yund* [1992]).

Frictional Strength

In experiments performed on nominally dry rocks at low temperatures and low effective pressures, the stress required to initiate frictional sliding is substantially smaller than that required for fracture. Thus the strength of pervasively jointed or fractured rock in the upper few kilometers of the lithosphere is probably limited by frictional resistance along faults. To first order, the stress necessary to cause sliding along a fault in nominally dry rocks at low temperature is most sensitive to normal load and relatively insensitive to surface conditions, sliding rate, mineralogy, and temperature. Except for a few minerals, such as swelling clays, the shear stress τ at which frictional sliding begins is related to the normal stress σ_n by Amonton's law

$$\tau = \mu_f \sigma_n + C_f \,, \tag{1}$$

where μ_f is the coefficient of friction and C_f is the frictional cohesive strength. More explicitly [*Byerlee*, 1978]

$$\tau = 0.85\,\sigma_n \qquad 3 < \sigma_n < 200 \text{ MPa} \tag{2a}$$

$$\tau = 60 + 0.6\sigma_n \qquad 200 < \sigma_n < 1700 \text{ MPa} \tag{2b}$$

The remarkable generality of this rule has led to its being called Byerlee's law. Of course, being an empirical generalization, it does not have the status of a physical law and may itself need to be modified to include, for example, the chemical effects of aqueous pore fluids. Hence, in the remainder of this paper, equation (2) will be referred to as Byerlee's rule. When pore fluids are present, both chemical and physical processes can influence strength. The physical effect of pore pressure P_p can be incorporated into these equations by using effective stresses $\sigma'_{ij} = \sigma_{ij} - \alpha P_p \delta_{ij}$, where $\delta_{ij} = 1$ if $i = j$ and $\delta_{ij} = 0$ if $i \neq j$. The coefficient α is of order unity but may vary depending on the specific situation [*Paterson*, 1978, pp. 71-73]; for simplicity, we take $\alpha = 1$. If equation (2) is now rewritten for a fault whose normal is 60° from the maximum principal stress σ_1, then [*Sibson*, 1974; *Brace and Kohlstedt*, 1980; *McGarr et al.*, 1982; *McGarr*, 1984]

$$\sigma_1 - P_p \approx 4.9(\sigma_3 - P_p) \qquad \sigma_3 - P_p < 100 \text{ MPa} \tag{3a}$$

$$\sigma_1 - P_p \approx 3.1(\sigma_3 - P_p) + 210 \qquad \sigma_3 - P_p > 100 \text{ MPa} \tag{3b}$$

where σ_3 is the minimum principal stress. In this analysis, one of the principal stresses σ_V is oriented vertically and is equal to the lithostatic pressure

$$\sigma_V = \rho_r g z \,, \tag{4}$$

where ρ_r is the density of the overlying rock, g is the acceleration due to gravity, and z is the depth. In continental crust the vertical stress gradient is ~27 MPa km^{-1}, while in oceanic lithosphere it is ~34 MPa km^{-1}. In the case of a thrust fault, where the horizontal principal stress σ_H is greater than σ_V, $\sigma_H = \sigma_1$ and $\sigma_V = \sigma_3$ [e.g., *McGarr and Gay*, 1978; *Nicolas*, 1987, p. 60]. The pore pressure is commonly expressed in terms of the lithostatic pressure P_l via the parameter λ

$$\lambda = \frac{P_p}{P_l} \,. \tag{5}$$

Hence, for example, if the pore pressure is hydrostatic such that $P_p = \rho_f g z$, then $\lambda \equiv \lambda_H \approx 0.38$, where ρ_f is the density of the pore fluid. For a more detailed discussion of the application and appropriateness of Byerlee's rule (equation (2)), to natural fault systems, see *Hickman* [1991].

Frictional sliding may be either stable or unstable, depending on the material, the effective pressure, the conditions along the frictional surface or gouge layer, the temperature and the rate of sliding [*Tullis*, 1986]. Unstable or stick-slip sliding has been suggested as a process analogous to the earthquake mechanism [*Brace and Byerlee*, 1966], and even small changes in the coefficient of friction are important in determining stability. Analysis indicates that slip weakening is a critical condition for unstable or seismogenic slip [*Stesky et al.*, 1974; *Tse and Rice*, 1986]. Thus many laboratory experiments have focused on understanding the evolution of the coefficient of friction with either slip displacement or slip rate and with the state of the surface [e.g., *Tullis*, 1986; *Beeler et al.*, 1994]. One important conclusion is that the onset of stable sliding does not necessarily coincide with the onset of plastic flow mechanisms [*Tse and Rice*, 1986].

Fracture Strength

If the rocks under consideration are not pervasively jointed or if the length scale between the joints is large, then the strength of the rock mass may be limited by the compressive fracture strength. As with frictional slip, compressive fracture strength of relatively intact, nominally dry rock is most sensitive to effective pressure and mineralogy; it depends to a much lesser extent on temperature and strain rate [*Paterson*, 1978]. Fracture strength is also affected by sample size [*Pratt et al.*, 1972; *Paterson*, 1978; *Pinto de Cuhna*, 1990; *Lockner*, 1995]; the scale dependence of strength changes for different rock types, amounts of weathering, and loading conditions. *Pratt et al.* [1972] observed a decrease in strength of 70% per decade increase in sample size up to sizes of a few meters; however, the rate of weakening may decrease for yet larger samples. Presumably at pressures below that for the brittle-ductile transition, the lower limit of fracture strength is given by the stress to cause frictional sliding.

Rock fracture under low to intermediate confining pressure occurs by formation of dilatant cracks, followed by localization along a shear zone. Failure in this region can be approximated as a linear relation between maximum compressive strength and confining pressure, expressed here in terms of shear stress at failure τ_f and normal stress

$$\tau_f = \mu_i \sigma_n + C_i, \tag{6}$$

where μ_i is the coefficient of internal friction and C_i is the internal cohesive strength. This relationship is often called the Mohr-Coulomb failure criterion. At higher pressures, the pressure sensitivity of strength decreases and the linear relation is less accurate [*Paterson*, 1978]. At low pressures, the compressive strength is greater than the stress to cause sliding on a fault, but as pressure increases, the fracture strength intersects the friction curve at the brittle-ductile transition, as illustrated in Figure 2.

At low temperatures for pressures less than about 600 MPa, the initial stage of failure of silicate rocks involves dilatant behavior, during which preexisting flaws extend in a direction parallel to the greatest principal stress [e.g., *Wong*, 1982]. At or near the peak load, the small axial cracks coalesce to form a shear plane inclined by about 30-40° to the greatest compressive stress.

A framework for understanding localization has been built from continuum mechanics models [*Rudnicki and Rice*, 1975], micromechanical models utilizing fracture mechanics [*Costin*, 1983; *Horii and Nemat-Nasser*, 1986; *Sammis and Ashby*, 1986; *Kemeny and Cook*, 1987a,b], and numerical studies [*Lockner and Madden*, 1991a,b]. The initial dilatancy that occurs before the peak stress is reached can be modeled by considering the growth of ancillary wing cracks from flaws oriented at the correct angle for maximum shear stress. The theories relate the cohesive strength and the internal coefficient of friction to physical parameters including the number of flaws per unit volume, the average length of the initial flaws, and the length of the wing cracks. Although the theories qualitatively describe many aspects of brittle failure, further work is necessary for a complete, quantitative rationalization of the failure criterion.

At pressures above about 600 MPa, silicates deforming in the brittle regime undergo a second transition in failure mode [*Shimada and Cho*, 1990; *Hirth and Tullis*, 1989, 1994; *Tullis and Yund*, 1992]. Failure still occurs by cataclastic processes but without the coalescence of axial dilatant cracks. Instead, fracturing takes place dominantly by shear cracking (mode II fracture in the parlance of fracture mechanics). Observations using optical and transmission electron microscopy suggest that the fault along which most strain is localized is quite narrow in extent and is characterized fully as a shear offset. Axial dilatant cracks are absent or much less apparent than in samples deformed at lower pressures. For quartzite, granite, gabbro, eclogite and dunite, the transition from one failure mode to another occurs at about 600 MPa and does not appear to depend on temperature. For nominally dry silicate rocks, the dependence of the ultimate strength on strain rate and temperature is relatively small. If water is added, creep effects become important, even at room temperature [*Lockner*, 1995]. Similarly, if even a trace amount of fluid is present, the fracture strength can be influenced significantly by temperature [*Wong*, 1982]. The addition of water can increase the kinetics for several deformation mechanisms, including stress corrosion (subcritical crack growth) [*Atkinson*, 1984], pressure solution [*Paterson*, 1995], and plastic flow [*Paterson*, 1989] (also see below).

Plastic Strength

Steady state plastic flow of rocks can take place by a variety of mechanisms including dislocation motion, solid-state diffusion, and solution-diffusion-precipitation processes [e.g., *Poirier*, 1985, Chapters 4 and 7; *Guéguen and*

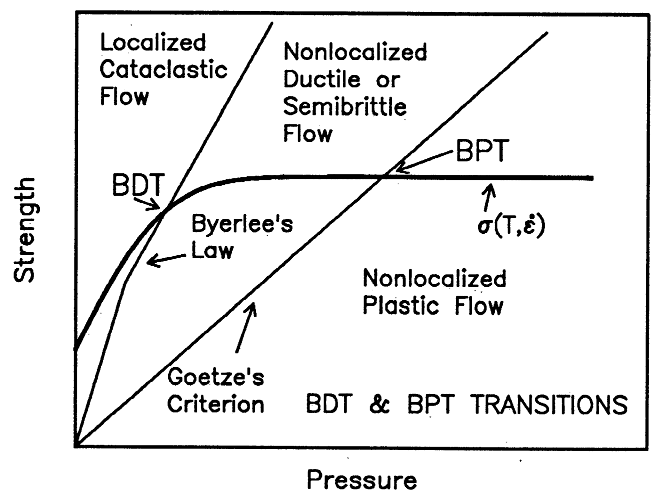

Figure 2. Schematic illustration for the transition from localized brittle failure to semibrittle flow (BDT) and from semibrittle flow to fully plastic flow (BPT) for a fixed temperature and strain rate based on laboratory studies of nominally dry intact rocks. Some rocks (e.g., calcite and halite) exhibit plasticity at room temperature, while most silicates require elevated temperature and pressures to initiate plastic flow.

Palciauskas, 1994, Chapter 6]. Direct observations of strong preferred orientations in crustal rocks collected from ductile shear zones and indirect observations of seismic anisotropy in mantle rocks demonstrate that dislocation creep dominates at the relatively high stresses (≥10 MPa) generally encountered in the lithosphere [*Wenk*, 1985; *Karato*, 1989a]. The flow law commonly used to describe dislocation flow in crystalline materials is a power law relation between strain rate $\dot{\epsilon}$ and differential stress $\sigma_1 - \sigma_3$ [e.g., *Weertman*, 1978]:

$$\dot{\epsilon} = A_d (\sigma_1 - \sigma_3)^n \exp\left(- \frac{Q_d + PV_d}{RT} \right), \tag{7}$$

where A_d is a material parameter, n is a constant characteristic of the creep process, Q_d is the activation energy, V_d is the activation volume for dislocation creep, P is pressure, R is the gas constant, and T is absolute temperature. The reader is referred to a review by *Tsenn and Carter* [1987] for a summary of other forms of the steady state creep equation. For differential stresses less than approximately 1% of the shear modulus, most data on plastic deformation of silicate rocks can be fit to equation (7) within measurement uncertainties using single values for A_d, n, Q_d, and V_d [e.g., *Tsenn and Carter*, 1987]. However, it should be noted that high-resolution creep experiments on olivine single crystals demonstrate that a single creep mechanism does not dominate over the full range of lithospheric pressure-temperature conditions [*Bai and Kohlstedt*, 1992a]. Based on the results of these experiments, the constitutive equation describing the plastic strength of the lithosphere can be expressed in terms of a combination of two or three power law relations of the form given in equation (7), each representing a specific creep mechanism characterized by a unique value for the activation energy (and for A_d) but with a single value for the stress exponent of $n \approx 3.5$ for all of the mechanisms [*Mackwell et al.*, 1990; *Bai et al.*,1991; *Jin et al.*, 1994]. The PV_d term in equation (5) has generally been omitted in modeling the rheology of the lithosphere; for $V_d = 17\times10^{-6}$ m^3 mol^{-1} [e.g., *Kirby*, 1983], stress increases by a factor of about 2 over a depth range of 50 km.

At the higher stresses in the colder portions of the lithosphere, the strain rate for plastic deformation increases approximately exponentially with increasing differential stress. A number of constitutive equations have been proposed to describe dislocation flow in this high-stress regime, often referred to as the power law breakdown regime (for a summary, see *Tsenn and Carter* [1987]). For example, *Goetze* [1978] used the following equation to describe the low-temperature, high-stress rheology of olivine:

$$\dot{\epsilon} = \dot{\epsilon}_0 \exp\left[- \frac{Q_d'}{RT} \left(1 - \frac{(\sigma_1 - \sigma_3)}{\sigma_p} \right)^2 \right] \tag{8}$$

where σ_p, the Peierl's stress, is the lattice resistance to dislocation glide and Q_d' is the activation energy for dislocation glide.

Within ductile shear zones, localization of deformation and presence of fluids may result in a significant reduction in grain size d due to dynamic recrystallization. Under these conditions, grain boundary diffusional creep processes, accommodated by a substantial component of grain boundary sliding, may predominate over dislocation mechanisms. The flow law is then given by [e.g., *Coble*, 1963]

$$\dot{\epsilon} = A_{gb} \frac{(\sigma_1 - \sigma_3)}{d^3} \exp\left(- \frac{Q_{gb} + PV_{gb}}{RT} \right) \tag{9}$$

where the subscript gb indicates grain boundary. The occurrence of equiaxed grains and lack of preferred orientation in some fault zones have been used as evidence for such a deformation mechanism [e.g., *Boullier and Guéguen*, 1975].

Brittle-Ductile and Brittle-Plastic Transitions

Field observations of deformed rocks within fault zones indicate that the transition from purely brittle to purely plastic deformation processes may occur over a relatively broad range of temperature and pressure [*Sibson*, 1977]. A change in failure mode from localized faulting to distributed shear accompanies the change in mechanism. The distinction between the transition in deformation mechanism and the transition in failure mode has sometimes been blurred, but we follow the terminology of *Rutter* [1986], denoting the change in mode as the brittle-ductile transition (BDT) and the change in dominant mechanisms as the brittle-plastic transition (BPT). If deformation involves both plastic and brittle mechanisms, it is called semibrittle flow. If deformation involves brittle failure only but strain is not macroscopically localized, it is called cataclastic flow. To illustrate these points, a schematic diagram is shown in Figure 2.

For some minerals, including halite and calcite, plastic slip or twinning occurs even at room temperature; in these cases, the BDT and the BPT can be induced in a deformation experiment simply by increasing the confining pressure [*Heard*, 1960; *Fredrich et al.*, 1989; 1990]. For quartzites, which do not show plasticity at room temperature, localized failure takes place at all pressures accessible in the laboratory; distributed failure in laboratory experiments begins only once some plasticity starts at temperatures >500°C and elevated pressure [*Hirth and Tullis*, 1994]. Feldspar rocks can exhibit cataclastic flow but only at temperatures above 300°C and pressures greater than 500 MPa [*Tullis and Yund*, 1987; 1992]; apparently, the presence of several cleavage planes allows significant strain to be accommodated by these rocks without concurrent localization.

The tendency for deformation in a brittle material to localize during compressive loading is increased if the material strongly dilates, if it strain softens, or if pressure sensitivity decreases. Using a bifurcation analysis, *Rudnicki and Rice* [1975] predicted the critical hardening modulus necessary for localization, h_{cr}, to be

$$\frac{h_{cr}}{G} = \frac{1+\nu}{9(1-\nu)}(\beta - \mu)^2 - \frac{1+\nu}{2}\left(N + \frac{\beta + \mu}{3} \right)^2, \tag{10}$$

where ν is Poisson's ratio, β is the dilatancy factor, μ is a coefficient of internal friction defined by the incremental slope of the yield surface expressed as a function of pressure, and N is $1/\sqrt{3}$ for axisymmetric compression. This analysis indicates that localization should be suppressed by any process that decreases dilatancy or promotes work hardening. For intact rocks without throughgoing faults, many of the conditions that suppress localization have been delineated by laboratory experiments on nominally dry samples of silicates, carbonates, and ceramics for pressures up to 600 MPa at a variety of temperatures below the brittle-plastic transition.

For example, factors which suppress dilatancy (including increased mode II cracking and cracking associated with easy cleavage [*Hirth and Tullis*, 1989, 1994; *Shimada and Cho*, 1990; *Lockner and Madden*, 1991a,b; *Shimada*, 1993], increased brittle grain crushing as well as collapse of preexisting pores [*Wong*, 1990], cracking associated with dispersed second-phase particles [*Dresen and Evans*, 1993], and crack-tip plasticity and distributed plastic flow [*Heard*, 1960; *Tullis and Yund*, 1977; 1992; *Evans et al.*, 1990]) tend to enhance ductility or delocalization.

Empirically, under nominally dry conditions, the BDT occurs when the Mohr-Coulomb criterion for brittle failure intersects the frictional sliding curve [*Byerlee*, 1968]. The BPT (Figure 2) occurs when the stress necessary for plastic flow is about equal to the effective confining pressure, that is, at the intersection of the plastic yield strength curve with the line

$$\sigma_1 - \sigma_3 = \sigma_3' . \tag{11}$$

This criterion is referred to here as Goetze's criterion (C. Goetze, private communication, 1975). *Shimada and Cho* [1990] have suggested that the fracture mode of rocks deformed at room temperature changes at a pressure of 0.8, 1.0, and 2.0 GPa for granite, dunite, and eclogite, respectively. In a study of quartzites deformed at temperatures between 300 and 1123 K, *Hirth and Tullis* [1994] observed a similar transition in failure mode at a pressure of 0.6 GPa, which they associate with an increase in the amount of mode II (shear) cracking.

Experimental Constraints on Strength Profiles

Frictional Sliding

The original strength curves formulated by *Sibson* [1977], *Goetze and Evans* [1979], *Brace and Kohlstedt* [1980] and others assumed that the strength of the upper, brittle regions of the lithosphere is governed by frictional strength along faults. For dry rocks, equation (2) conveniently expresses the relationship between the normal stress on a fault and the stress necessary to cause sliding. Remarkably, frictional strengths obey this relation over a broad range of temperatures and sliding rates for a wide variety of rock types with different roughness, provided that water is not present. If the pore pressure is given by a hydrostat, strengths predicted by Byerlee's rule provide upper limits to in situ stresses measured in a number of wells to depths of a few kilometers (for a review, see *Hickman* [1991]).

The coefficient of friction changes, however, if aqueous pore fluids are present. Substantial weakening occurs for slip along faulted rocks with saturated quartz gouge [*Chester and Higgs*, 1992; *Chester*, 1994] or with saturated granite gouge [*Blanpied et al.*, 1991, 1995] when temperature is increased and strain rate is decreased. Weakening in fine-grained quartz gouge is associated with changes in the microstructure, suggesting that deformation occurs via solution transfer in the pore fluid [*Chester and Higgs*, 1992]. Above a certain temperature that depends on material properties and slip rate, chemical interactions with the aqueous pore fluid may reduce the coefficient of friction. Thus frictional behavior in the presence of a fluid might be considerably more complex than suggested by Byerlee's rule, reflecting contributions from several different frictional slip mechanisms that vary with temperature, pressure, slip rate, and/or mineralogy [*Chester*, 1995]. The net effect would be to introduce a set of frictional slip mechanisms intermediate to those described by Byerlee's rule and fully plastic deformation mechanisms.

The rate effects that are important in understanding the stability of frictional sliding are often described by the rate-state equation [*Dieterich*, 1979; *Ruina*, 1985]

$$\tau = \mu^* \sigma_n + A_\mu \sigma \ln\left(\frac{V}{V^*}\right) + B_\mu \sigma \theta \tag{12a}$$

$$\frac{d\theta}{dt} = -\frac{V}{\delta_c}\left[\theta + \ln\left(\frac{V}{V^*}\right)\right] \tag{12b}$$

where V is the slip velocity, θ is a poorly defined variable that describes the state of the surface, δ_c is a decay distance, the starred quantities are reference values, and A_μ and B_μ are constants describing the direct effect and the indirect effect of velocity on the coefficient of friction. The influence of temperature on the coefficient of friction can be described by assuming that the shear strain rate $\dot{\gamma}$ within the gouge layer follows an Arrhenius relationship with temperature, $\dot{\gamma} = \dot{\gamma}_o \exp(-Q_\mu / RT)$, where Q_μ is the activation energy governing frictional sliding. If it is further assumed that the effect of temperature on friction can be included by adding terms $\sigma_n f(1/T)$ and $g(1/T)$ to equations (12a) and (12b), respectively, then for a gouge layer of thickness ζ, the substitution $\dot{\gamma} = V/\zeta$ yields [*Chester and Higgs*, 1992]

$$\tau = \mu^* \sigma_n + A_\mu \sigma_n \ln\left(\frac{V}{V^*}\right) + \frac{Q_A}{R}\left(\frac{1}{T} - \frac{1}{T^*}\right) + B_\mu \sigma_n \theta \tag{13a}$$

$$\frac{d\theta}{dt} = -\frac{V}{\delta_c}\left[\theta + \ln\left(\frac{V}{V^*}\right) + \frac{Q_B}{R}\left(\frac{1}{T} - \frac{1}{T^*}\right)\right]. \tag{13b}$$

Experiments suggest that the friction mechanism controlling the sliding process changes as thermodynamic conditions change. Thus the coefficient of friction will depend on temperature, sliding rate, pore fluid pressure, water fugacity, and possibly total pressure. Further, the activation energies Q_A and Q_B are expected to depend on mineralogy and mechanism of sliding.

Estimates of the frictional strength for quartz and granite gouge are plotted in Figure 3a and 3b for sliding velocities of 1.0 and 0.001 μm s^{-1}, respectively, for wet quartz gouge [*Chester and Higgs*, 1992; *Chester*, 1994, 1995] and for a rate of 0.01 μm s^{-1} for granite gouge [*Blanpied et al.*, 1991, 1995]. Deviations from Byerlee's rule are particularly striking at higher temperatures and lower sliding rates.

Brittle Fracture

For formations with few preexisting faults or for faults with extensive induration, the strength of the rock mass may be greater than the stress necessary to cause sliding on an optimally oriented fault. Thus fracture experiments provide an upper bound to the strength of the rock mass. Recent evidence indicates that the failure mode changes as pressure increases [*Hirth and Tullis*, 1989, 1994; *Shimada and Cho*, 1990; *Shimada*, 1993]. Above about 600 MPa in quartzite under nominally dry conditions, the angle between the normal to the final fault surface and the shortening direction

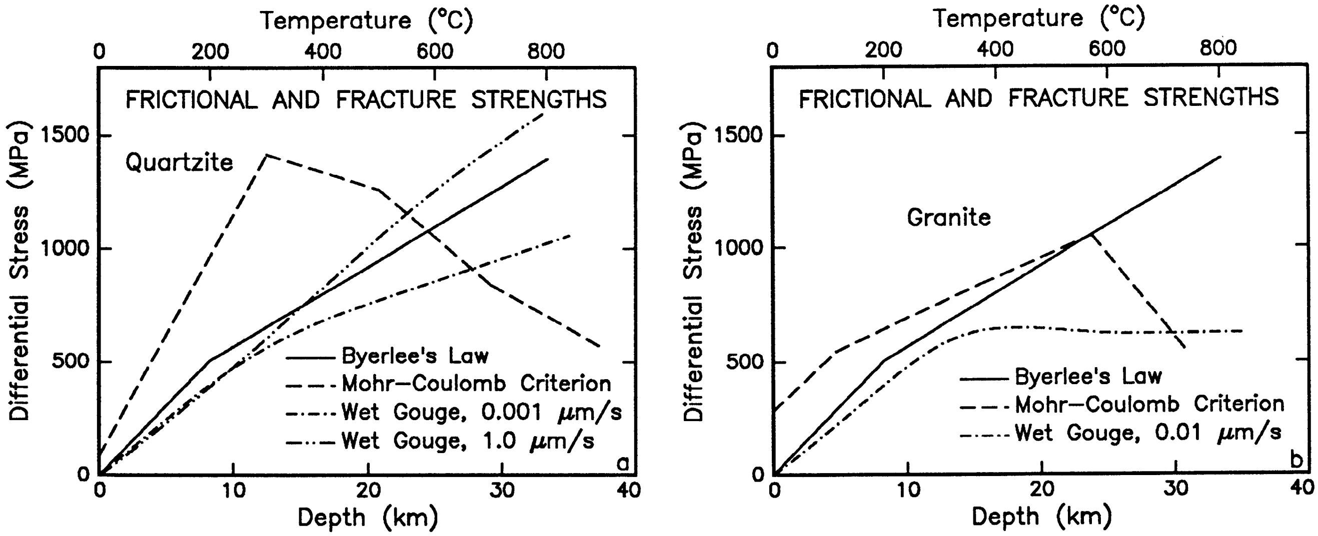

Figure 3. (a) Comparison of fracture strength and frictional strengths for quartzite rocks. The Mohr-Coulomb criterion for fracture strength is obtained from experiments on nominally dry quartzite at the pressure and temperature appropriate for each depth [*Hirth and Tullis*, 1994]. The curves were reduced by 30% at all temperatures and pressures to estimate the effect of reducing the strain rate from 10^{-5} s^{-1} to 10^{-15} s^{-1}. For the stress to cause frictional sliding, Byerlee's rule as determined for nominally dry rocks is compared with data from experiments on fault gouge under hydrothermal conditions [*Chester and Higgs*, 1991; *Chester*, 1994]. (b) Similar plot for fracture strength of granite from *Wong* [1982] and *Tullis and Yund* [1977] reduced by 30%. Friction curves are for granite samples with fine-grained granite gouge saturated with distilled water from *Blanpied et al.* [1994].

decreases from 60° to 45°. Apparently, dilatant mode I cracking is suppressed as pressure is increased, while mode II shear cracks become more abundant. The appearance of the fault surface changes from a diffuse region of cracked material formed by coalescence of mode I cracks, to a rather sharp fault with substantial shear offset.

If fluids are present, stress corrosion and creep effects are significant. For example, compressive fracture strength is reduced by 10% for a decrease in pressure of 25 MPa, for an increase in temperature of 200°C, or for a decrease in strain rate by a factor of 1000 (for a comprehensive review, see *Lockner* [1995]). The exact constitutive equation for stress corrosion creep is still being debated [*Kranz*, 1980; *Atkinson*, 1984; *Lockner*, 1993], but assuming that low-temperature creep is activated by stress as well as temperature, *Lockner* [1993] calculated an activation energy of about 150 kJ mol^{-1}. To estimate the weakening due to a decrease in strain rate from 10^{-8} to 10^{-14} s^{-1}, fracture strengths for granite and quartzite are plotted in Figures 3a and 3b, respectively, after reduction by 30% over the entire temperature and pressure range. The actual weakening is probably greater especially at greater depths, because the stress corrosion creep rate probably depends on the fugacity of water.

Plastic Deformation

Numerous experimental studies have been published on the creep strength of a broad range of crustal and mantle minerals and rocks [e.g., *Kirby and Kronenberg*, 1987; *Tsenn and Carter*, 1987]. However, problems with apparatus design and control of chemical environment in many of the studies performed prior to 1980 limit the application of experimentally determined rheologies to geologic situations. In this review, emphasis is placed on flow laws that were obtained in well-controlled chemical environments using apparatuses with sufficient stress resolution for accurate extrapolation. Of particular note are recent advances in solid-medium machines using low-strength salt cells, which have vastly improved the quality of flow laws at higher pressures [*Green and Borch*, 1990; *Gleason and Tullis*, 1993]. At the same time, improved jacketing procedures and furnace design have extended the range of chemical environment and temperature attainable in the gas-medium apparatus [*Paterson et al.*, 1982; *Paterson*, 1990].

For crustal rocks, reproducible rheologies are now available for single-phase aggregates of quartz, clinopyroxene, and calcite. Only limited data are currently available for feldspars and polyphase aggregates. For mantle rocks, well-constrained flow laws have been determined for rocks of peridotitic composition, as well as for single-phase olivine aggregates.

Crustal rheologies. For quartz, the major issue dominating laboratory studies of rheology is the effect of water on high-temperature creep, as quartzites without added water are generally too strong to be deformed appreciably by plastic flow mechanisms at realistic crustal stresses (for a review, see *Paterson* [1989]). In the present paper, we focus on results from experiments in which quartz grains were saturated with water, so that the water fugacity is well-defined. As diffusion of water in quartz is relatively slow at experimental temperatures [*Kronenberg et al.*, 1986], we use the results of *Kronenberg and Tullis* [1984] on novaculites (grain size <5 μm) with added water, of *Luan and Paterson* [1992] on synthetic polycrystalline quartz fabricated under hydrous conditions from silicic acid, and of *Gleason and Tullis* [1995] on wet Black Hills quartzite. In the experiments on novaculite, 0.5 wt % water was added to the sample capsule and, at least in some cases, water was present at the end of the deformation run. For the experiments on synthetic quartzite, the water pressure was estimated from the total water content and porosity. In both cases, the water pressure was near or at the total confining pressure. The Black Hills quartzite contains several thousand parts per million hydroxyl; and the molten-salt assembly used in these experiments results in a high hydrogen activity, inhibiting diffusion of hydrogen out the sample capsule.

To estimate the dependence of strain rate on water fugacity, we used the stress-strain curves for novaculite at pressures between 350 and 1590 MPa from *Kronenberg and Tullis* [1984] Figure 3a, and assumed that water pressure was approximately equal to confining pressure. Water fugacity was calculated from the results of *Pizner and Sterner* [1994]. As illustrated in Figure 4, at constant stress and temperature the strain rate increases systematically with increasing water fugacity f_{H_2O} according to a relation of the form

$$\dot{\epsilon} \propto f_{H_2O}^{p}. \tag{14}$$

If the steady state stress levels estimated from the results of *Kronenberg and Tullis* [1984] are converted to strain rate at a common stress using equation (7) with their value of $n = 2.6$ for the stress exponent, then $p \approx 0.8$; if $n = 4.0$ is used [e.g., *Luan and Paterson*, 1992; *Gleason and Tullis*, 1995], then $p \approx 1.2$.

Based on this relationship between strain rate and water fugacity, the experimental data of *Kronenberg and Tullis* [1984] obtained in a solid-medium apparatus can be compared to those of *Luan and Paterson* [1992] measured with a gas-medium apparatus, as shown in the plot of stress versus strain rate in Figure 5. The recent results of *Gleason and Tullis* [1995], using a molten-salt cell, for the deformation of Black Hills quartzite containing ~0.15 wt % water are included for comparison. Results from all three studies were reduced to a common temperature using values reported by the authors for activation energy and to a common pressure using a water fugacity exponent in equation (14) of $p = 1.0$; no correction was made for the effect of pressure through the activation volume term in equation (7). The agreement among the three sets of data is striking. In particular, the results obtained in the gas-medium apparatus [*Luan and Paterson*, 1992] and those acquired with the molten-salt assembly [*Gleason and Tullis*, 1995] are nearly identical. The strength measured in the solid-medium apparatus [*Kronenberg and Tullis*, 1984] is somewhat higher than the other two values and the stress exponent is somewhat lower (2.6 versus 4.0), consistent with a recent comparison of results from solid-medium and molten-salt cells [*Gleason and Tullis*, 1993].

For calcite, experiments have been performed on both coarse-grained (Carrara and Yule marbles) and fine-grained (Solenhofen limestone) rocks [*Heard and Raleigh*, 1972; *Schmid et al.*, 1977, 1980; *Rutter*, 1986] as well as fine-grained synthetic aggregates [*Walker et al.*, 1990]. Thus rheologies have been determined in the dislocation creep [*Heard and Raleigh*, 1972; *Schmid et al.*, 1980] and diffusional creep [*Schmid et al.*, 1977] regimes, as well as the power law breakdown regime [*Rutter*, 1986]. These results have been used to construct the strength profiles present in Figure 6. Plastic deformation in the fine-grained Solenhofen limestone [*Schmid et al.*, 1977, 1980] and fine-grained synthetic calcite aggregate [*Walker et al.*, 1990] occurs predominantly by twinning and power law breakdown at higher stresses (>400 MPa at <800°C) and by grain-size dependent diffusional creep at lower stresses and higher temperatures. The coarse-grained Carrara and Yule marbles also deform by twinning and power law breakdown at higher stresses but by power law creep at lower stresses and higher

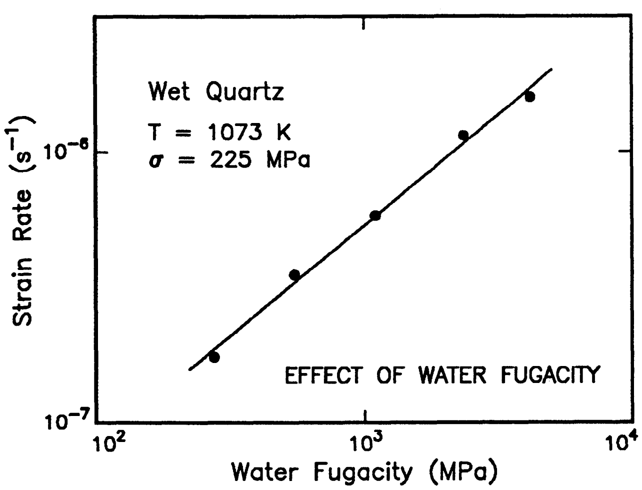

Figure 4. Strain rate versus water fugacity for fine-grained quartzite deformed in the dislocation creep regime using the stress-strain data in Figure 3a of *Kronenberg and Tullis* [1984]. For experiments in which steady state creep was not attained by the end of the run, the steady state flow stress was estimate from an extrapolation of the stress-strain data to a strain of ~25%. Flow stresses were converted to strain rates at a common stress of 225 MPa using a stress exponent of 2.6, the value reported for water-added experiments at a confining pressure of 1500 MPa. For the sample deformed at a confining pressure of 350 MPa, the flow stress exceeded the confining pressure; hence, as noted by Kronenberg and Tullis, optically visible microcracks may have dominated the deformation behavior of this sample.

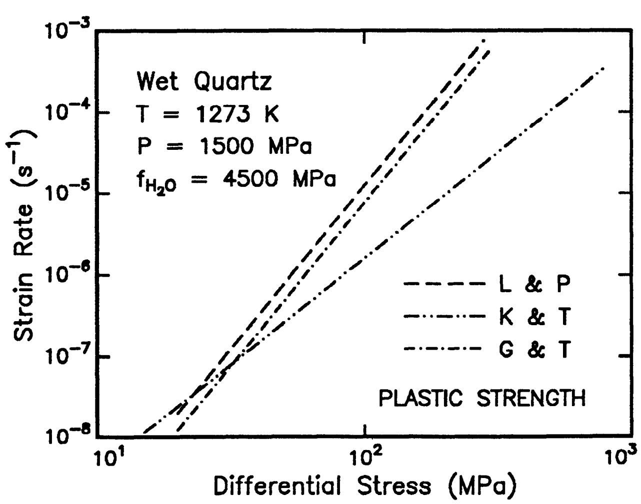

Figure 5. Strain rate versus stress for wet quartzite including rheologies from *Luan and Paterson* [1992], *Kronenberg and Tullis* [1984], and *Gleason and Tullis* [1994]. The data of Luan and Paterson were extrapolated from a fluid pressure of 300 MPa to 1500 MPa using a water fugacity exponent of unity.

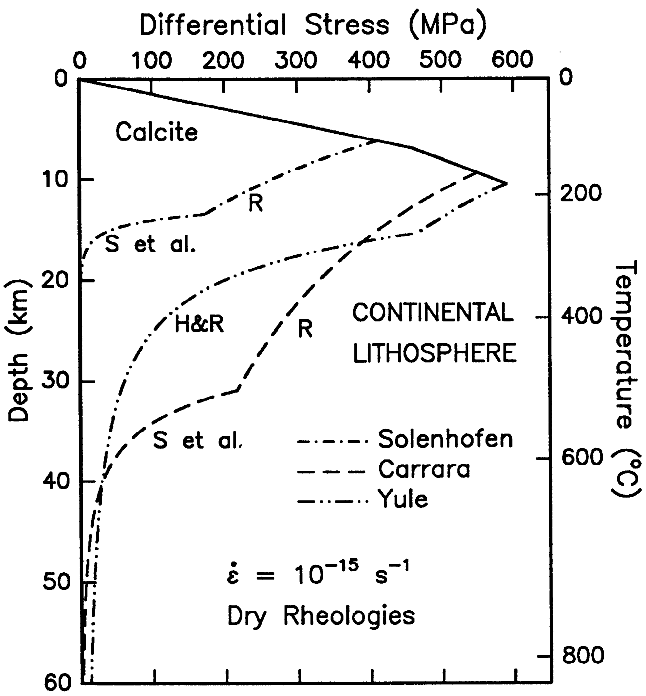

Figure 6. Strength envelope based on rheologies for dry calcite. Rheologies are plotted based on the following data: R indicates data from *Rutter* [1986] for power law breakdown creep in Solenhofen limestone, Carrara marble, and Yule marble; S et al. indicates data from *Schmid et al.* [1977] for diffusion creep of Solenhofen limestone and from *Schmid et al.* [1980] for power law creep of Carrara marble; and H & R indicates data from *Heard and Raleigh* [1972] for power law creep of Yule marble. The geotherm used in this strength profile is that of *Chapman* [1986] for continental crust with a surface heat flow of 60 mW m^{-1}.

temperatures. Although no grain-size dependent flow has been observed experimentally in coarse-grained marbles, diffusional creep may become important at geologic strain rates.

For clinopyroxene, several studies have reported on the deformation behavior of diopside single crystals [e.g., *Kolle and Blacic*, 1982, 1983; *Raterron and Jaoul*, 1991] and polycrystalline aggregates [e.g., *Kirby and Kronenberg*, 1984; *Boland and Tullis*, 1986]. The rheologies of materials with compositions dominated by diopside are sometimes used as analogues for lower crustal rocks. The results of the recent studies on diopsidite were used to construct the strength profiles shown in Figure 7. While the study of *Boland and Tullis* [1986] was carried out under hydrous conditions, that of *Kirby and Kronenberg* [1984] was performed under nominally anhydrous conditions. Curiously, their results, when extrapolated to the lower temperatures and slower strain rates of the middle and lower crust, indicate higher strengths for the wet samples; this result is more likely an artifact of the extrapolation of data with large uncertainties in the activation energy for creep than a real effect (see later discussion). Other than some preliminary data by *Mackwell* [1992], no results have been reported for deformation in the diffusional creep regime, although it may be important for creep under the conditions expected in Earth's crust. The diopsidite rheologies indicate strengths that are intermediate between those determined for shallow crustal rocks (wet quartzites and dry carbonates) and mantle rocks (wet and dry dunites and peridotites).

Mantle rheologies. In the past decade, the rheologies for olivine-rich rocks have been investigated at high pressures and temperatures under both dry and wet conditions. Under dry conditions at a confining pressure of 300 MPa, the flow law for dislocation creep determined by *Chopra and Paterson* [1981, 1984] for two natural dunites and that reported by *Karato et al.* [1986] for a synthetic olivine aggregate agree within a factor of 2 in stress, the finer-grained synthetic aggregates being somewhat weaker than the coarser-grained rocks. This difference in strength between coarse-grained and fine-grained samples in the dislocation creep regime has recently been connected with a contribution to the strain associated with grain boundary sliding in samples with smaller grain sizes [*Hirth and Kohlstedt*, 1995].

Under wet conditions, experiments have been carried out at confining pressures from 300 MPa [*Chopra and Paterson*, 1984; *Karato et al.*, 1986] to 2000 MPa [*Borch and Green*, 1989]. (Note that although the samples of Borch and Green were initially thought to be dry, subsequent infrared analyses have revealed the presence of a significant amount of hydroxyl (H.W. Green II, private communication, 1994).) As illustrated in Figure 8, the results of the experiments at 300 MPa on two natural dunites and a synthetic dunite agree within a factor of 2 in stress at laboratory strain rates of 10^{-4} to 10^{-5} s^{-1}. However, the synthetic harzburgite (olivine plus 3% orthopyroxene) deformed at higher pressures is substantially weaker than any of the samples deformed at lower pressures. This discrepancy apparently arises for two

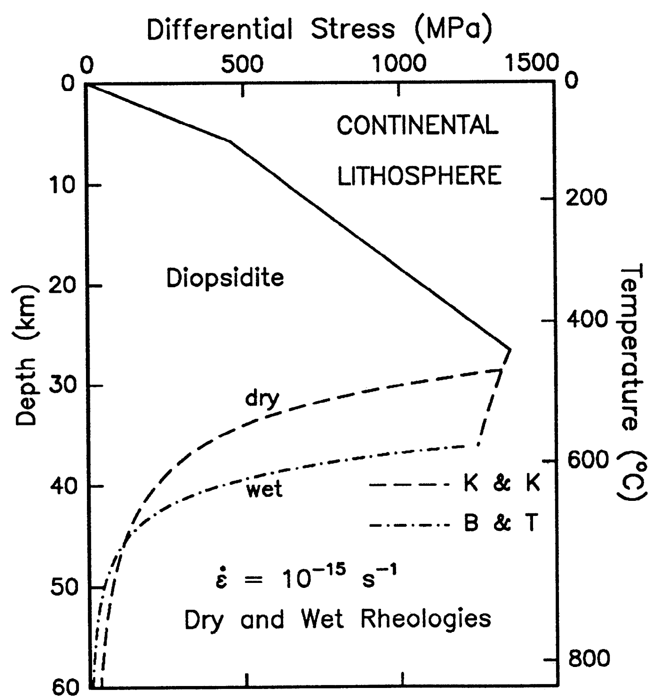

Figure 7. Strength envelope based on rheologies for diopsidite. Dry rheologies (K & K) are plotted using the data of *Kirby and Kronenberg* [1984] for power law breakdown creep and power law creep in Sleaford Bay diopsidite. A wet rheology (B & T) is also plotted based on the data of *Boland and Tullis* [1986]. The geotherm used in this strength profile is that of *Chapman* [1986] for continental crust with a surface heat flow of 60 mW m^{-1}.

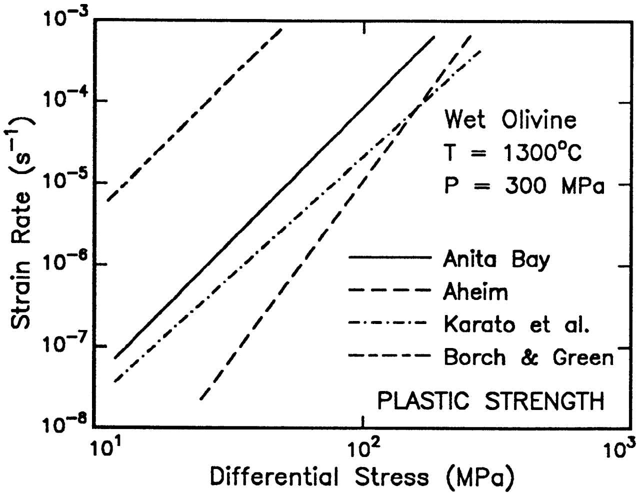

Figure 8. Strain rate versus differential stress for four olivine samples plotted for a temperature of 1300°C and a confining pressure of 300 MPa. The results for Anita Bay and Aheim dunite are from *Chopra and Paterson* [1981, 1984]. The results of *Borch and Green* [1989] were reduced to a confining pressure of 300 MPa using the pressure correction reported in their power law creep equation. No correction has been made for differences in water fugacity or oxygen fugacity between gas-medium and molten-salt assemblies.

reasons: First, the samples used in the experiments performed at 300 MPa were jacketed in iron, so that the oxygen fugacity was buffered by the Fe-FeO reaction. In the higher pressure experiments, platinum jackets were employed, so that the oxygen fugacity was fixed near the oxidizing boundary of the olivine stability field, as indicated by the formation of small amounts of hematite (H.W. Green II, private communication, 1992). It is well established from deformation experiments on olivine single crystals at 1 atm and at 300 MPa that creep rate increases with increasing oxygen fugacity f_{O_2} according to the relation

$$\dot{\epsilon} \propto f_{O_2}^{q} \tag{15}$$

with $0 < q < 0.4$ [*Bai et al.*, 1991]. A limited series of experiments on synthetic olivine aggregates yields an oxygen fugacity exponent of $q \approx 1/3$ [*Beeman and Kohlstedt*, 1993]. If the data of *Borch and Green* [1989] are adjusted to an oxygen fugacity corresponding to that fixed by an Fe-FeO buffer, the strain rate would decrease by a factor of 10 to 100. Second, the effect of water fugacity must be taken into account. Because the dependence of creep rate on water fugacity for olivine-rich rocks has not yet been determined, it might be assumed as a zeroth-order approximation that the dependence of creep rate on water fugacity mimics the dependence of water solubility on water fugacity; in this case, creep rate would vary as water fugacity to ~0.8 power [*Bai and Kohlstedt*, 1992b]; preliminary data are consistent with this assertion (G. Hirth and S.J. Mackwell, private communication, 1995). If the samples of *Borch and Green* [1989] were water-saturated, then strain rates measured at a confining pressure of 1500 MPa should be reduced by almost a factor of 20. Although a complete comparison between experiments carried out in the molten-salt cell and those performed in the gas apparatus is not possible without better constraints on the thermodynamic conditions in the former, agreement between the two groups of experiments is substantially better than indicated in Figure 8, if the effects of oxygen and water fugacity are taken into account. (It should be noted that the dependence of creep rate on water fugacity reflects the dependence of water or hydroxyl solubility on water fugacity; at high pressures, the relationship between hydroxyl solubility and water fugacity deviates substantially from linearity [*Kohlstedt et al.*, 1995].)

Brittle-Ductile Transition

Empirically, for dry rocks, the intersection of the brittle-fracture criterion with the Byerlee's friction curve marks the highest stress at which macroscopic stress drops occur in intact materials. The transition from velocity weakening to velocity strengthening along a frictional fault is a distinct and separate point, with important implications for the stability of sliding along the faulting surface [*Tse and Rice*, 1986]. Because of the paucity of data for mechanical behavior under hydrothermal conditions, it is not clear if a similar transition in failure mode will occur for the intersection of the failure criterion for stable crack extension and the stress to cause sliding when friction occurs by creep mechanisms such as solution transfer.

Constitutive equations based on detailed micromechanical observations do not exist for semibrittle deformation behavior. *Kirby* [1980] estimated the pressure at the onset of semibrittle behavior by assuming that the transition would occur when the effective pressure is about 0.4 times the frictional strength. *Chester* [1988] proposed that the total strength in the semibrittle regime can be written in terms of the creep strength σ_c and the fracture strength σ_f as

$$\sigma_1 - \sigma_3 = \phi\sigma_c + (1-\phi)\sigma_f \,, \tag{16}$$

where ϕ is a mixing parameter given by $\tanh(\beta\sigma_3)$ and β is a material constant. Although this equation provides an estimate of strain partitioning between brittle and ductile processes, it supposes that the processes are independent and makes no allowance for the influence of temperature. In lieu of more detailed information, we estimated the onset of semibrittle flow by assuming that limited plastic flow can occur when the yield or flow stress is less than 5 times the frictional stress. This criterion assumes that stress concentrations along the fault elevate the local stress to values high enough to cause yielding. Although this estimate is clearly qualitative, at the least it provides rate and temperature dependencies of the transition pressure. All of the current estimates are lacking in that they are not based on detailed physical models and they do not account for the effects of strain rate, grain size, porosity, second phases, water fugacity, and temperature.

Application of Strength Envelopes to Earth

Extrapolation From Laboratory to Geologic Conditions

Based on the information from the previous section, strength profiles for oceanic and continental lithosphere are presented in Figures 9a and 9b. The following approach was used in their construction: First, for simplicity we use Byerlee's frictional sliding rule as determined on nominally dry rocks (equation (2)) to define the strength profile at shallow depths; pore pressure was taken to be hydrostatic. The actual stress to cause sliding is probably lower because of the chemical effect of water. Second, the rheologies for quartz and/or olivine aggregates delineate the strength of the plastic portions of the lithosphere; as discussed in the preceding and following sections, robust rheologies for polymineralic rocks are not yet available. Third, the brittle-ductile transition (BDT) truncates the frictional sliding curve at the depth at which the flow stress is five times the frictional strength, as discussed in the previous section. Fourth, the flow strength curve is terminated at the point at which it intersects Goetze's criterion given in equation (11) for hydrostatic pore pressure (i.e., at the BPT). The BDT is connected to the BPT using a dotted line, as no constitutive equation is available for this region.

The strength profile for 60-m.y.-old oceanic lithosphere [*Turcotte and Schubert*, 1982, pp. 163-167] in Figure 9a based on rheologies of dry rocks is characterized by a relatively broad brittle-to-plastic transition region. The calculated strength reaches a maximum value of ~800 MPa at a depth of ~35-40 km. A rheology for dry mantle rocks was used because water is effectively removed from the rock during decompression melting in upwelling mantle. Thus the pore fluid pressure is assumed to be very low. The peak strength might be reduced by as much as a factor of two, if the rheology for dry olivine were replaced by one for wet olivine. However, a high-stress (e.g., equation (8)) flow law for wet olivine is not available to constrain further the peak stress or the depth at which it occurs. Similarly, the frictional strength might also be reduced by the presence of aqueous fluids, but, again, no data exist.

The strength profile in Figure 9b for continental lithosphere with a geotherm corresponding to a surface heat flow of 60 mW m^{-1} [*Chapman*, 1986] reveals a relatively weak lower crust. While the top of the mantle reaches a strength of ~300 MPa, the base of the crust has a strength of <10 MPa. Flow laws for wet quartzite are included for the crust, based on evidence for fluid pressures at or above the hydrostatic level [*Brace and Kohlstedt*, 1980]. Although the quartzite strengths determined by *Gleason and Tullis* [1995] and *Luan and Paterson* [1992] are in reasonable agreement at laboratory conditions (Figure 5), extrapolation to crustal temperatures leads to substantial differences in predicted rock strength, as illustrated in Figure 9b. Laboratory experiments are generally performed at strain rates that are 5 to 10 orders of magnitude faster than those found under normal geologic conditions; in the ductile regime, experiments are also frequently carried out at temperatures significantly higher than those appropriate for the lithosphere. Extrapolation of strength curves determined in the laboratory to natural conditions requires that the activation energy for creep be determined accurately [e.g., *Chen and Molnar*, 1983] and that no changes in deformation mechanism occur [*Paterson*, 1976, 1987]. In the present comparison, the activation energy of 223 kJ mol^{-1} reported by *Gleason and Tullis* [1995] is ~50% larger than the values of 130 and 150 kJ mol^{-1} reported by *Kronenberg and Tullis* [1984] and *Luan and Paterson* [1992], respectively, leading to marked differences in strengths predicted for crustal conditions (e.g., over a factor of 10 at a depth of 17.5 km and a temperature of 300°C). The uncertainty in strength is significantly smaller for a dry dunite rheology, however; for an activation energy of 535±35 kJ mol^{-1}, it is less than a factor of 2 at 45 km and 700°C. (It should be noted that this uncertainty in strength is much smaller than that quoted by *Molnar* [1992], in some part because we used a smaller uncertainty

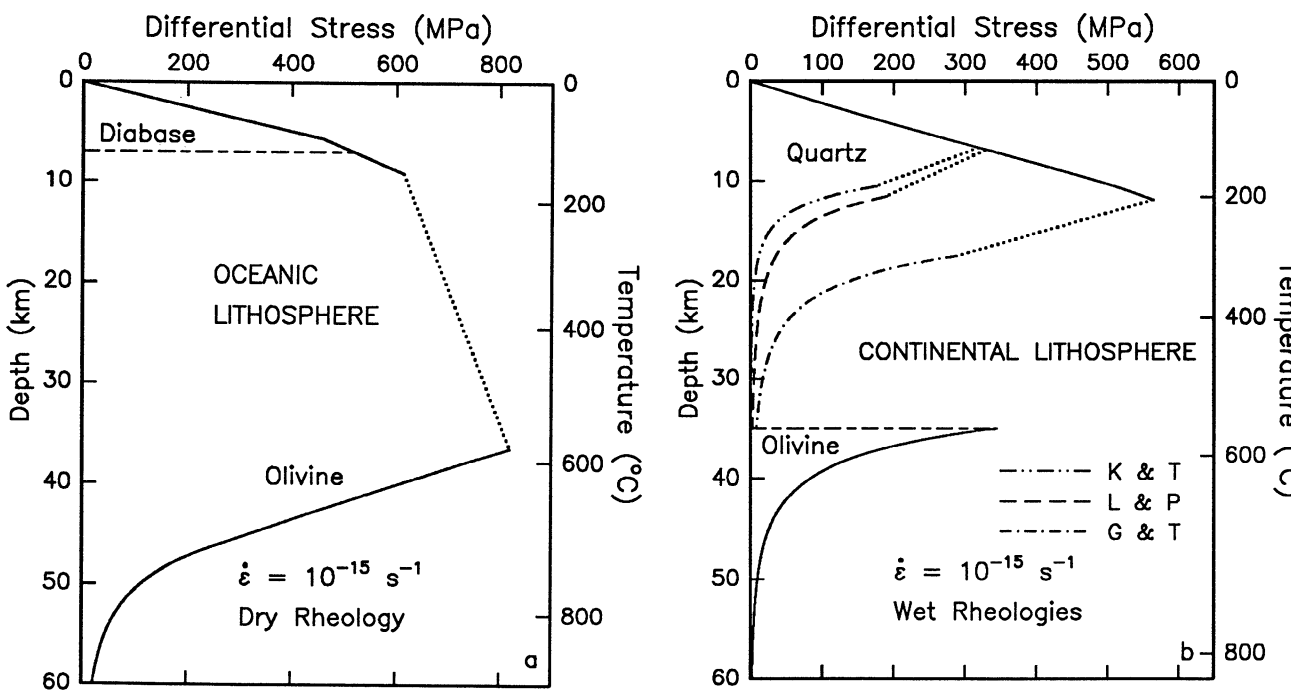

Figure 9. Strength envelopes for oceanic and continental lithosphere. (a) For the oceanic lithosphere, a geotherm for 60-m.y.-old lithosphere was used [e.g., *Turcotte and Schubert*, 1982 pp. 163-167]. A rheology for dry olivine [*Chopra and Paterson*, 1984] was used because water strongly partitions into the melt during partial melting. (b) For the continental lithosphere, a geotherm for a surface heat flow of 60 mW m^{-1} was employed [*Chapman*, 1986]. The rheologies for wet quartzite are those used in Figure 5; the olivine rheology is for wet Anita Bay dunite from *Chopra and Paterson* [1984]. Wet rheologies were used, consistent with high fluid pressures in fault zones. Plastic flow strength was corrected for water fugacity using a water fugacity exponent of unity and assuming lithostatic pore pressure. The BDT and BPT, determined as described in the text, have been connected by a dotted line.

in activation energy but in large part because he neglected to adjust the value of the material parameter A_d in equation (7) for different values of Q_d.)

Role of Fluids

Frictional slip on mature faults can apparently occur at stresses considerably below those predicted from Byerlee's rule with hydrostatic pore pressure [*Hickman*, 1991]. Recent measurements of in situ stresses, heat flow, and focal plane mechanisms indicate that the San Andreas fault is probably weak compared with equation (2) and weak relative to the surrounding material [*Rice*, 1992]. At least two general explanations for the weakening exist: First, effective pressures are much less than those predicted by the difference between the lithostatic and hydrostatic pressures [e.g., *Hubbert and Rubey*, 1959; *Byerlee*, 1990; *Rice*, 1992] or, second, the coefficient of friction is substantially less than that predicted by equation (2) [*Blanpied et al.*, 1995; *Chester and Higgs*, 1992].

Field structural indicators for transient periods of high pore fluid pressure are abundant and include evidence from fluid inclusions, vein formation, oxygen isotope analysis, metamorphic assemblages, and hydrothermal ore deposits (for reviews, see *Fyfe et al.* [1978], *Etheridge et al.* [1984], *Walder and Nur* [1984], *and Nur and Walder* [1992]). Permeable zones with hydrostatic fluid pressure gradients exist at moderate depths in many natural situations [*Brace*, 1980], but regions of overpressured fluids are also observed [e.g., *Fertl*, 1976; *Neuzil*, 1986]. If high pore pressures along faults are to be maintained over geologic time, either rock permeability must be rather low [*Bredehoeft and Hanshaw*, 1968; *Nur and Walder*, 1992] or an active source of overpressured fluids must exist at depth [*Rice*, 1992]. Observations of extensive fluid flow may be reconciled with low permeability, if permeability and pore structure vary spatially and temporally [*Sibson*, 1982; *Walder and Nur*, 1984; *Guéguen et al.*, 1986; *Byerlee*, 1990; *Nur and Walder*, 1992; *Rice*, 1992]. Recent friction experiments provide an example of the effect of sealing on the production of elevated pore fluid pressure and the associated reduction of permeability and rock strength [*Blanpied et al.*, 1992].

Under laboratory conditions, at least, the presence of water during sliding experiments can decrease the coefficient of friction by more than 30%, as illustrated in Figure 3. Additionally, high pore fluid pressure can substantially reduce the maximum strength of the lithosphere. To illustrate this point, pore pressures corresponding to $\lambda = P_p/\sigma_3$ of 0, 0.38, 0.76, and 0.94 or equivalently, relative to a hydrostatic pore pressure, $\lambda_H = \rho_f/\rho_r$, $\lambda/\lambda_H = 0$, 1, 2, and 2.5 were used in constructing Figure 10. As fluid pressure builds up, rock strength decreases and the depth at which the maximum stress is reached increases. When faulting occurs, permeability increases and pore pressure drops; λ then increases with increasing time as solution-precipitation processes, for example, seal fractures in the fault zone. As pore pressure again approaches the lithostatic value, the fault becomes weak. In the example presented in Figure 10, crustal strength decreases from ~400 MPa for hydrostatic pore pressure to ~50 MPa for 2.5 times that value.

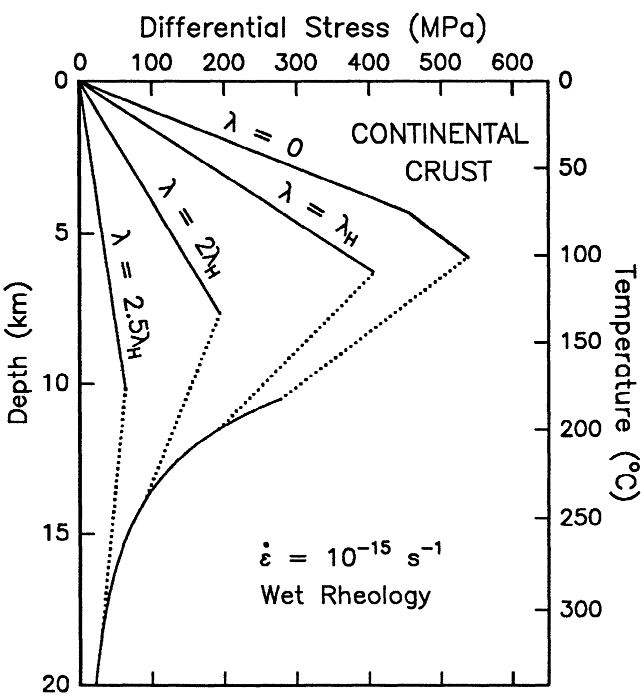

Figure 10. Maximum differential stress supported by rocks as a function of depth for several values of pore pressure. The dotted line connects the BDT to the BPT, as described in the text. The rheology for plastic flow is based on the data of *Luan and Paterson* [1992] for wet quartzite; strength is corrected for the effect of water fugacity as a function of depth by taking the fluid pressure to be lithostatic.

Rheologies for Polymineralic Rocks

All of the ductile rheologies considered above are for single-phase aggregates. Since laboratory measurements are usually carried out at temperatures higher than expected under geologic conditions in order to obtain measurable strain rates (e.g., equations (7)-(9)), experiments on polymineralic rocks are very often complicated by melting reactions. For example, dunites and even harzburgites can be deformed at laboratory strain rates at a temperature of 1300°C with stresses of less than 100 MPa. If clinopyroxene and spinel are added to form lherzolite, the solidus temperature is reduced to ~1150°C, below which the rock has a strength greater than 300 MPa at a strain rate of 10^{-6} s^{-1} (M. Zimmerman, personal communication, 1994). Hence several authors have recently examined techniques for calculating rheologies for polymineralic rocks from measured rheologies of the constituent phases.

Analytical solutions for the rheology of two-phase mixtures derived from flow laws for the constituent phases exist only for highly idealized phase distributions [*Chen and Argon*, 1979] and hence are applicable at most to only a limited range of rocks. Several averaging schemes have been proposed for calculating the composite rheology from the single-phase flow laws and their volume fractions [*Tullis et al.*, 1991; *Ji and Zhao*, 1993; *Handy*, 1994]. While these schemes provide a good approximation for the rheology of aggregates composed of phases with similar strengths and equant shapes, they do not accurately predict the flow behavior of rocks composed of two phases with markedly different strengths. For example, the addition of only 20% quartz to a calcite matrix results in a disproportionate increase in strength of a factor of ~10 (G. Dresen, M. Eckhardt, B. Evans, and D.L. Olgaard, private communication, 1995). For many multiphase rocks, therefore, it will be necessary to use finite element techniques to calculate the aggregate strength from monomineralic flow laws [*Tullis et al.*, 1991].

Experimental Rheologies and Strength of the Lithosphere

From the preceding discussion, it is clear that first-order descriptions of strength distributions in the oceanic and continental lithosphere can be illustrated using strength envelopes, such as those given in Figure 9. These models present the maximum strengths of the rocks, assuming simple mineralogical sequences, for the lithosphere deforming at a constant strain rate. While rather simplistic, such models are very useful in understanding large-scale effects, such as seamount loading and postglacial rebound. Used judiciously, they may also provide insight into more localized lithospheric deformation, such as that associated with the San Andreas fault system [e.g., *Verdonck and Furlong*, 1992; *Furlong*, 1993]. In their study, *Verdonck and Furlong* [1992] used the rheologies of *Karato et al.* [1986] and *Zeuch* [1983] to characterize the strain distribution in the San Francisco Bay area region and explain the existence of a parallel set of major strike-slip faults.

Although this article has focused on plastic deformation in the steady state regime, transient or time-dependent creep may dominate in certain tectonic settings. Transient creep is expected to be important when total strains are small, as in the case of postglacial rebound [*Yuen et al.*, 1986; *Sabadini et al.*, 1987; *Karato*, 1989b]. *Smith and Carpenter* [1987] reanalyzed published creep data to determine the nature of the transient response for mantle minerals and aggregates, and *Hanson and Spetzler* [1994] measured transient creep behavior in natural and synthetic olivine single crystals. However, as noted by Hanson and Spetzler, scaling of these data to deformation in the mantle remains problematic, due to the limited number of experimental studies and the incomplete development of an appropriate flow law.

Although it is comparatively straightforward to apply experimental rheologies for mantle rocks to the behavior of the lithospheric mantle, application of experimental rheologies under conditions favoring plastic deformation to the behavior of the continental crust is rather more difficult. The oceanic crust is comparatively simple mineralogically, with rocks of predominantly basaltic composition. In contrast, the continental crust is quite complex, with both lateral and vertical variations in mineralogy and texture [*Kay and Kay*, 1981]. Thus the concept of "typical" continental crustal rheology is likely to be a gross oversimplification. In addition, abundant evidence from seismology and structural geology demonstrates that a large component of the deformation occurring in the lower continental crust may be concentrated in ductile shear zones, in which mineralogies and textures may be significantly different from those of the surrounding country rock. For example, recent research on the deep structure of the San Andreas fault system by *Brocher et al.* [1994] indicates the presence of a subhorizontal ductile shear zone at a depth of about 15 km in the crust, as predicted by the rheological study by *Verdonck and Furlong* [1992] and *Furlong* [1993].

In regions in which strain is predominantly concentrated in faults and ductile shear zones, localized strain rates will be higher than those expected if deformation occurred homogeneously throughout the crust. Thus, processes such as dynamic recrystallization and grain-size dependent creep may be favored in such regions. In addition, as such zones are believed to be major conduits for fluid transport through the crust, the ready availability of water might promote hydrolytic weakening or favor deformation mechanisms different from those in the neighboring country rock [*Karato et al.*, 1986; *Rutter and Brodie*, 1988a,b]. Syntectonic metamorphic reactions may also have important effects on the nature of deformation [*Rutter and Brodie*, 1988c], due to processes such as the release of fluids through dehydration or decarbonation, changes in the stress state associated with solid-state volume changes, enhanced deformation resulting from fining of the grain size during phase transitions, and changes in chemical potential gradients due to solid-state reactions. Consequently, strength envelopes based on a constant rate of strain and a constant lithology throughout the lithosphere must be applied with caution to real continental crust, which may have a highly heterogeneous and time-dependent strain distribution.

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