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Rheology of the Lithosphere: Selected Topics

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We review recent results concerning the rheology of the lithosphere with special attention to the following topics: 1) the flexure of the oceanic lithosphere, 2) deformation of the continental lithosphere resulting from vertical surface loads and forces applied at plate margins, 3) the rheological stratification of the continents, 4) strain localization and shear zone development, and 5) strain-induced crystallographic preferred orientations and anisotropies in body-wave velocities. We conclude with a section citing the 1983–1986 rock mechanics literature by category.

INTRODUCTION

Improved geophysical observations, continuum mechanical modeling, and application of laboratory measurements of mechanical properties of rocks to problems associated with plate dynamics have led to advances during the period 1983-1986 in our understanding of the rheology of the earth's lithosphere. Rheological models for the oceanic lithosphere, applied to large-scale deformations at plate boundaries and within plate interiors, have been further developed using elastic, elastic-plastic, and viscoelastic formulations. These models have been further refined by incorporating nonlinear stress and temperature dependencies into the viscous response of Maxwell-type viscoelastic rheological models, consistent with experimental measurements of the mechanical properties of rocks at elevated temperatures. In addition to flexure at trench-rise systems, deformation of the oceanic lithosphere has been examined within plate interiors in response to large horizontal compressional forces and thermally-derived stresses, and constrained by measured ocean floor topographic profiles, marine geoid anomalies, and the distributions and focal mechanisms of earthquakes.

Rheological models for the continental lithosphere have likewise emerged in this quadrennial period, based upon a continuum approach to the large-scale structures developed in diverse tectonic settings, and upon experimentally-determined mechanical responses of crustal and mantle lithologies. The mechanical behavior of the continental lithosphere is complicated by its compositional heterogeneity and complex thermal history, and cannot, as yet, be as closely constrained as that of the oceanic lithosphere. Nevertheless, favorable comparisons of model results with observed structures have led to insights into the tectonics and mechanical response of the continents.

Rather than attempt a discussion of the entire literature pertinent to the rheologies of the oceanic and continental lithospheres, we select several current lines of research for discussion which we feel are particularly important and noteworthy. We review recent results concerning 1) the flexure of the oceanic lithosphere, 2) deformation of the continental lithosphere resulting from vertical surface loads and forces applied at plate margins, 3) the rheological stratification of the continents, 4) strain localization and shear zone development, and 5) strain-induced crystallographic preferred orientations and anisotropy of elastic wave velocities. We conclude with a section citing the 1983–1986 rock mechanics literature by category.

1. FLEXURE OF THE OCEANIC LITHOSPHERE

The concept of rigid plates constituting a lithosphere overlying a more fluid-like asthenosphere has been most successful in describing the tectonics of the ocean basins, owing in part to the relatively high strength of the oceanic lithosphere. As shown by the bulges in sea floor topography, geoid anomalies, and other flexural features which extend into the oceanic lithosphere from loads applied at deep ocean trenches and seamounts, the oceanic lithosphere is capable of supporting large differential stresses over extended geologic times. Consistent with these observations, yield envelopes for the oceanic lithosphere, based upon experimentally-determined mechanical properties of rocks which constitute the oceanic crust and upper mantle [*e.g., Goetze and Evans,* 1979; *Brace and Kohlstedt,* 1980; *Kirby,* 1983], require loads in excess of those generally available for significant inelastic deformations within plate interiors. Thus, with the exception of relatively gentle flexural features, displacements and deformation tend to localize at plate boundaries.

Analyses of flexure of the oceanic lithosphere have been particularly rewarding due to the relatively simple geometries involved and the wide range of geophysical constraints which can be placed upon

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model results. Models of plate flexure have included elastic, as well as elastic-plastic rheologies, based upon experimentally measured mechanical properties, and have been compared with observed vertical seafloor surface displacements, gravity anomalies, distributions of seismicity, and the inferred loading and environmental conditions. Flexural models over the period 1983–1986 have been extended to oceanic plates of widely differing ages and within differing tectonic settings based upon elastic, elastic-plastic, viscoelastic, and layered rheological models. Among the most important developments which have come from these studies has been the definition of the lithosphere-asthenosphere interface based upon time-dependent yield strength, which coincides with the seismically defined lithosphere and, for a given time of loading, corresponds approximately to an isothermal contour within the upper mantle.

Elastic plate models, although largely surpassed by more realistic rheological models, have provided a useful first order approximation to the behavior of the lithosphere and have recently been applied to evaluate the state of stress near ridge-transform intersections [*Morgan and Parmentier,* 1984] and in determining the thermoelastic bending stresses generated by lateral variations in heat flow [*Bills,* 1983]. Within the context of flexural features of the ocean floors, elastic plate models have been used to characterize plate bending resistance in terms of an effective flexural rigidity and elastic plate thickness. Comparisons of calculated flexural rigidities and effective plate thicknesses for oceanic plates of differing ages at the time of loading have revealed a particularly important trend of increasing flexural strength with age [*e.g., Watts,* 1978, 1982; *Bodine et al.,* 1981] which closely parallels models of plate cooling. Recent contributions have been made by matching gravitational anomalies calculated from an elastic flexure model with SEASAT altimeter profiles of globally distributed oceanic trench systems [*McAdoo and Martin,* 1984; *McAdoo et al.,* 1985] and determining effective elastic thicknesses for plates ranging in age between 22 and 160 m.y. Effective plate thicknesses determined over this interval ranged from 27 to 63 km, in agreement with the relationship between plate thickness and the square root of lithospheric age as suggested by *Bodine et al.* [1981]. The improved geographic coverage provided by the SEASAT altimeter data has indicated that regional compressional stresses normal to trench trends are not needed to account for the observed geoid profiles [*McAdoo et al.,* 1985].

While the central core of lithospheric plates may remain elastic during flexure, stresses within the upper lithosphere are likely to be limited by pressure-dependent brittle failure. Changes in mechanical properties of the oceanic lithosphere with age, comparable with trends of cooling, reveal the importance of temperature-dependent ductile processes at the base of the lithosphere. Recent modeling efforts of outer rise-trench systems have therefore been focused on the development of more realistic, composite layer models which are consistent with the results of experimental rock mechanics [*Goetze and Evans,* 1979; *Kirby,* 1980, 1983]. In addition to matching bathymetric and gravity profiles [*McAdoo et al.,* 1985], these layered rheological models predict lithospheric plate thicknesses more consistent with those derived from maximum depths of intraplate seismicity [*Wiens and Stein,* 1983, 1984, 1985]. Significantly, the base of this mechanically-based lithosphere appears to correspond to an isotherm of between 700 to 800° C, when compared with thermal cooling models of the oceanic plates [*Parsons and Sclater,* 1977], reflecting the exponential temperature dependence of creep.

In addition to models of flexure near plate margins, layered rheological models have been applied to a unique example of intraplate flexural buckling within the central Indian Ocean [*McAdoo and Sandwell,* 1985; *Zuber,* in press] apparently resulting from large horizontal compressional stresses associated with the Indian-Eurasian plate collision. *McAdoo and Sandwell* [1985] examined the thinning of the elastic core of an elastic-plastic plate as yield conditions associated with brittle fracture and plastic flow within the upper and lower regions of the plate, respectively, were reached. Using both plastic and lab-based nonlinear viscous models for the lithosphere overlying a viscous asthenosphere, *Zuber* [in press] examined both the flexural buckling of a plate of uniform thickness and the growth of instabilities in a hydrodynamic flow model of the lithosphere. While differing in approach, these models predict fold wavelengths of 200 km, consistent with those of the seafloor topographic undulations and geoid anomalies associated with buckling.

As the combined thicknesses of the basalts and gabbros of the oceanic crust do not generally exceed 6–7 km, the mechanical properties of the oceanic lithosphere are likely to be controlled by the materials of the upper mantle. Experimental determinations of the fracture, frictional, and flow properties of peridotites and of olivine, the predominant phase of the upper mantle, have been extensive, spanning an enormous range of environmental conditions and have had an important influence upon modeling efforts. Given the distribution of pressure and temperature with depth in the oceanic lithosphere, stresses within the upper regions of the lithosphere may be predicted by Coulomb laws for fracture and laws for frictional sliding on pre-existing fractures, whereas ductile flow within the lower regions of the lithosphere may be constrained by laboratory-based ductile creep relations. The principal uncertainties of these applications stem from conjectures regarding fluid pressure and chemistry, hydrothermal alteration, and olivine grain size within the lithosphere.

Lower bounds to inelastic yielding within the upper oceanic lithosphere have been based upon the frictional response of rocks with pre-existing fractures. Frictional behavior of rocks and minerals are relatively insensitive to rate of deformation at room temperature and exploratory experiments suggest that temperature also has a small effect up to 400°C. With the exception of hydrous minerals, these data may be described by two relatively simple, linear friction laws, depending on the range of normal stresses [*Byerlee,* 1968; *Brace and Kohlstedt,* 1980; *Kirby,* 1983]. Written in terms of principal stresses and assuming fracture surfaces of all orientations

$$(\sigma_1 - \sigma_3) = 3.9\ \sigma_3 \qquad \text{for } \sigma_3 < 120 \text{ MPa} \quad (1a)$$

$$(\sigma_1 - \sigma_3) = 210 + 2.1\ \sigma_3 \qquad \text{for } \sigma_3 > 120 \text{ MPa} \quad (1b)$$

Although σ_3 in these relations may be well constrained by the lithostatic load under dry conditions, pore fluid pressures P_f may reduce the effective minimum principal stress $\sigma_3' = \sigma_3 - P_f$, leading to reductions in differential stresses. Recent results for dunite [*Pinkston and Kirby,* 1982; *Pinkston et al.,* 1986] have roughly confirmed these relations for olivine under anhydrous conditions at pressures of up to 700 MPa and temperatures to 600°C but with somewhat lower σ_3 coefficients. Large departures from this relation were observed, however, for samples with only trace quantities of water and hydrous alteration products on grain boundaries. Whether a result of pore pressures decreasing the effective pressure, the markedly lower frictional strengths of the hydrous minerals, or the result of alteration reactions, the presence of fluids at shallow depths within the oceanic lithosphere may lead to significant reductions in its strength within the brittle field.

The base of the mechanical lithosphere in ocean basins may be defined physically by the exponential temperature dependence of creep for olivine. While experimental work is still needed to characterize its transient creep properties, steady state flow laws for olivine are now well established under anhydrous conditions and in the presence of water for both oriented single crystals and coarse-grained polycrystalline aggregates (Table 1) and work is well on its way to determining the effects of grain size, water, oxidation states within the mantle, defect chemistry, and presence of partial melts. Creep laws for olivine under conditions which favor dislocation processes may be represented by a thermally activated power law

$$\dot{\epsilon}_s = A\sigma^n \exp(-H^*/RT) \qquad (2)$$

where $\dot{\epsilon}_s$ is the steady-state creep rate, $\sigma = (\sigma_1 - \sigma_3)$ is the differential stress, T is absolute temperature, A is a material constant, $\underline{n}$ a dimensionless constant of the order 3.4 to 4.5, and $H^* = E^* + PV^*$ is the activation enthalpy (E^* is the activation energy, V^* is the activation volume, and P is the hydrostatic pressure or mean normal stress). Of greatest impact to flexural models of the lithosphere are the combined effects of geothermal gradients and the strong dependence of creep upon temperature, the non-Newtonian power-law dependence upon stress, and the time-dependent nature of strength within the lowermost regions of the lithosphere. While creep rates may be presumed to vary smoothly with increasing temperature at depth, a critical temperature can be defined operationally within a given time frame, corresponding to an effective mechanical discontinuity between the lithosphere and asthenosphere. The non-linear dependence of strain rate, characteristic of dislocation creep, also has an important geophysical impact, affecting the distribution and pattern of strains resulting from various loading sources. Although eqn. (2) is non-Newtonian by definition, it can be expressed in terms of a simple viscous relation

$$\dot{\epsilon}_s = \frac{1}{2\eta}\sigma \qquad (3)$$

by defining an effective viscosity

$$\eta \text{ (effective)} = \frac{\sigma^{1-n}\exp(H^*/RT)}{2A} \qquad (4)$$

which may vary locally as a function of stress. The final feature we emphasize in the relation (2) is its time dependence. If we invert eqn. (2), the steady-state stress at fixed strain rate is

$$\sigma_s = \left(\frac{\dot{\epsilon}}{A}\right)^{1/n} \exp(H^*/nRT) \qquad (5)$$

is time dependent for a given strain increment. σ_s is the strain-independent counterpart of $\dot{\epsilon}_s$ in constant stress tests. In addition, this form of the flow law exhibits a powerful exponential effect of inverse temperature on the steady state strength, an effect that leads to plate-like behavior. Over the time of flexural loading, viscous relaxation within the lower lithosphere may lead to reductions in lower lithosphere stresses and the amplification of stresses within the upper lithosphere [*Kusznir,* 1982; *Bott and Kusznir,* 1984].

The experimental data for the flow of olivine are among the most extensive of earth materials; however, applications to the rheology of the lower lithosphere involves uncertainties associated with the important effects of water [*Chopra and Paterson,* 1981; *Mackwell et al.,* 1985] and possible contributions of grain-size sensitive diffusional creep at fine grain sizes [*Karato,* 1984; *Karato et al.,* 1986; *Chopra,* 1986]. As fluid inclusions within mantle xenoliths are composed primarily of CO_2 [*e.g., Green and Gueguen,* 1983; *Bergman and Dubessy,* 1984; *Rovetta et al.,* 1986; *Tingle et al.,* 1986], H_2O is not expected to be the dominant fluid within the upper mantle. Nevertheless, only

TABLE 1. Steady-State Flow Law Parameters for Olivine: Dislocation Creep

$\dot{\epsilon}_s = A\sigma^n \exp(-H^*/RT) \qquad H^* = E^* + PV^*$

	$\log_{10} A$ ($MPa^{-n}\ s^{-1}$)	n	H^* ($kJ\ mol^{-1}$)	V^* ($m^3\ mol^{-1}$)
Dry				
Karato *et al.* [1982] GS = 0.02 to 0.2 mm P = 0.1 MPa	3.9	3.5±0.6	528±63	—
Kirby [1983] interpretation of single-crystal rheology and diffusion data P = 0.1 MPa	4.8±1.2	3.5±0.6	533±60	$(17 \pm 4) \times 10^{-6}$
Chopra and Paterson [1984] GS = 0.1 and 0.9 mm P = 300 MPa	4.46±0.18	3.6±0.2	535±33	—
Green and Hobbs [1984], Green and Borch [1986] P = 1000–3000 MPa	—	—	—	$(28 \text{ to } 36) \times 10^{-6}$
Zeuch [1984] GS = 0.3 mm P = 1000–1500 MPa (exponential stress dependence at differential stresses of 370–1290 MPa)	—	—	594	—
Karato *et al.* [1986] GS = 0.03 to 0.06 mm P = 300 MPa	—	3–3.5	—	—
Wet				
Chopra and Paterson [1981] Anita Bay dunite GS = 0.1 mm P = 300 MPa	4.0±0.2	3.4±0.2	444±24	—
Aheim dunite GS = 0.9 mm P = 300 MPa	2.6±0.2	4.5±0.2	498±38	—
Chopra [1986] GS = 0.01 mm P = 300 MPa (at T = 1100°C)	—	3.3	—	—
Karato *et al.* [1986] GS = 0.03 to 0.06 mm P = 300 MPa	—	3–3.5	—	—

GS = Grain Size.

trace quantities of intracrystalline H_2O are required [*Mackwell et al.*, 1985] for hydrolytic weakening, well within the range of concentrations measured in mantle-derived olivines [*Miller and Rossman,* 1985] and garnets [*Aines and Rossman,* 1984]. Recent experimental efforts have also been aimed at the mechanical behavior of polyphase peridotites, olivine-basalt partial melts, and very fine-grained olivine aggregates (Table 2). Of particular importance has been the discovery of nearly linear rheologies for partial melts and fine-grained olivine aggregates associated with diffusional transfer creep mechanisms [*Cooper and Kohlstedt,* 1984; *Chopra,* 1986; *Karato et al.,* 1986]. These rheologies similarly are expressed in terms of eqn. (2), but incorporating a grain size dependence in the pre-exponential term A, and with values of n ranging between 0.9 to 1.5. Application of these recent results to the mantle will require models of olivine grain size [*Ross,* 1983; *Karato,* 1984] and extrapolations of the competing dislocation and diffusional creep rheologies, in addition to those of its thermal structure. While linear viscous flow in the mantle requires serious consideration [*Ranalli,* 1984; *Ranalli and Fischer,* 1984; *Karato et al.,* 1986], the predominance of microstructural and textural evidence from naturally deformed ultramafic xenoliths and massifs [*e.g., Gueguen and Nicolas,* 1980; *Ross,* 1983] suggest that dislocation creep processes are important in the upper mantle. In addition, while

TABLE 2. Flow Law Parameters of Mantle Materials: Effects of Grain Size, Melt, and Mineralogy

$$\dot\epsilon_s = A\sigma^n \exp(-H^*/RT)$$

Material	$\log_{10} A$ (MPa^{-n} s^{-1})	n	H^* (kJ mol^{-1})	Comments	Ref.†
Synthetic olivine aggregate (Mg, Fe)$_2$SiO$_4$ GS = 0.01 mm	—	3.3	—	at 1100°C σ =350–672 MPa	1
P = 300 MPa	—	1.5	—	at 1200°–1300°C σ =28–150 MPa	1
Synthetic dunite aggregate (Mg, Fe)$_2$SiO$_4$ GS = 0.007 to 0.06 mm	—	3–3.5	—	at 1300°C GS = 0.03–0.06 mm	2
P = 300 MPa	—	1.4	—	at 1300°C GS = 0.007–0.03 mm	2
Olivine + basalt liquid partial melt GS = 0.003–0.013 mm P = 0.1 MPa	—	0.9±0.2	385	at 1300°–1400°C	3
Spinel lherzolite partial melt P = 1000 MPa	-11.6	2.95	36–117	at 900°–1100°C	4, 5
Fe$_2$SiO$_4$ single crystals	22.2	6±1	932±220		6

†References: 1. Chopra [1986], 2. Karato *et al.* [1986], 3. Cooper and Kohlstedt [1984], 4. Bussod and Christie [1983], 5. Bussod and Christie [1984], 6. Ricoult and Kohlstedt [1984].

the success of flexural models incorporating nonlinear olivine rheologies in predicting seafloor bathymetric and gravitational profiles are not necessarily diagnostic of the underlying creep laws, plate thicknesses derived from dislocation creep laws are in excellent agreement with the depth distribution of intraplate earthquakes [*Wiens and Stein,* 1983, 1984, 1985].

Due to the relatively similar times of loading involved in flexure of the oceanic lithosphere within various outer rise-trench systems, modeling efforts have not required the incorporation of the time dependence of creep explicitly. However the strength of the lower lithosphere may differ under other loading conditions and time duration. In order to capture the time dependence of the mechanical lithosphere, *DeRito et al.* [1986] have developed a viscoelastic plate model for flexure in which elastic stresses within the lower lithosphere decay by time-dependent nonlinear creep. Using a Maxwell-type model, a characteristic time τ_M

$$\tau_M = \frac{\epsilon_{\text{elast}}}{\dot\epsilon_{\text{creep}}} = \frac{2\eta}{E} \qquad (6)$$

can be defined [*Melosh,* 1980] as the time required for the inelastic creep strain under a constant load to equal the elastic strain due to the same load (and can be expressed equivalently in terms of viscosity η and Young's modulus E). Using the effective viscosity (eqn. 4) for power law creep

$$\tau_M \text{ (effective)} = \frac{\exp(H^*/RT)}{EA\sigma^{n-1}} \qquad (7)$$

depends upon temperature as well as stress.

Turcotte and Schubert [1982] similarly have defined a characteristic relaxation time τ_r as the time required for an elastic strain to relax by non-linear viscous flow to half its initial value in a spring and non-linear dashpot model subject to a *constant total strain constraint* (as opposed to constant stress). This rheological parameter τ_r exhibits identical temperature and stress dependencies as the Maxwell time.

During flexure, three subhorizontal rheological layers develop as functions of plate loading and geothermal gradients. Within the cold upper regions of the plate, characteristic times τ_M are much greater than the time of loading $t(\tau_M > t)$ corresponding to elastic behavior, whereas at intermediate levels, $\tau_M \cong t$ corresponding to transitional viscoelastic behavior, and at deeper levels, $\tau_M < t$ corresponding to a predominantly viscous rheology. Under flexural loads, stresses and τ_M within the plate vary with depth. A *strength parameter* Σ is defined as

$$\Sigma = \frac{\sigma}{\sigma_{\text{elast}}} = \frac{\tau_M}{\tau_M + t} \quad \text{(generally bounded by } 0 < \Sigma \leq 1) \qquad (8)$$

(differential stress is normalized with respect to the elastic stress that would be present at the same strain at $t = 0$). Σ varies with depth just as τ_M varies with temperature and flexural stress. Using this model, *DeRito et al.* [1986] determined contours of equal Σ and showed an approximate correspondence of the base of the lithosphere, as defined by the $\Sigma \cong 0.3$ contour, to the 700°C isotherm based on a plate cooling model. Thus the lithosphere and asthenosphere are defined by the Maxwell time in relation to the time scale of loading.

2. DEFORMATION OF THE CONTINENTAL LITHOSPHERE

Despite the far more complex thermal and mechanical structure of the continents, significant developments in our understanding of the continental lithosphere have resulted from simple, yet elegant, continuum mechanical plate models employing elastic, viscous, viscoelastic, and plastic rheologies. Among these, flexural models have been developed for the subsidence of continental plate margins during rifting and crustal thinning associated with thermal heating and extension [*Park and Westbrook,* 1983; *Alvarez et al.,* 1984], the development of large-scale intracontinental basins [*Bills,* 1983; *Lambeck,* 1983; *Garner and Turcotte,* 1984; *Nunn and Sleep,* 1984], and the response of the continental crust to vertical loads associated with surface topography, erosion [*Stephenson,* 1984] and plate scale faulting [*Owens,* 1983]. Constrained by sedimentation and erosional histories, these models have produced estimates of elastic flexural rigidities, time-dependent viscoelastic responses, and effective mechanical plate thicknesses. In addition, models of rifting and graben formation [*Bott and Mithen,* 1983; *Keen,* 1985] have been developed using an upper brittle layer to represent the shallow crust overlying temperature-dependent viscous and viscoelastic layers representing the lower crust and mantle. On an entirely different time scale, models have been developed for the elastic strain accumulation, coseismic, and postseismic viscoelastic response associated with great earthquakes using models of a plate-scale crack within an elastic lithosphere overlying a viscoelastic asthenosphere [*Li and Rice,* 1983*a,b*; *Melosh and Raefsky,* 1983; *Bonafede et al.,* 1984, 1985; *Cohen and Kramer,* 1984; *Thatcher and Rundle,* 1984; *Li and Kisslinger,* 1985; *Reilinger,* 1986]. Models of stress diffusion and associated surface displacements, combined with active monitoring of seismically active faults should provide close constraints on the rheological properties of the continental lithosphere on this time scale.

Perhaps the most provocative results concerning the large-scale structures and tectonics of the continents have come from model studies of continental deformations associated with convergent, divergent, and transcurrent plate motions. Compared with deformations within the oceanic lithosphere, deformation of continental plates is far more penetrative and complex. However, neglecting heterogeneities in crustal lithologies and in environmental conditions, continental deformations have been modeled to first order by examining the mechanical response of continental plates with relatively simple, uniform rheologies to applied displacements at their boundaries. Applying displacement boundary conditions to the continental Eurasian plate associated with its collision with India, *Tapponnier and Molnar* [1976] were able to match many of the tectonic features of the Himalayan arc and, on the basis of plastic slip-line analysis, predicted the patterns of strike-slip faulting and seismic activity [*Khattri and Tyagi,* 1983].

More recently, the continental lithosphere has been modeled as a thin viscous plate, with either Newtonian or power-law stress dependencies, overlying an inviscid asthenosphere [*Bird and Piper,* 1980; *England and McKenzie,* 1982; *England et al.,* 1985; *England and Houseman,* 1986; *Houseman and England,* 1986*b*]. Assuming that vertical gradients of horizontal velocities within the thin plate are small, vertical averages of strain rates and stresses were related by a depth-averaged rheology, integrating the plate's temperature-dependent viscosity over its vertical temperature gradient. Comparisons of lateral deformation fields within this thickness-averaged plate resulting from boundary conditions associated with continental collisions, extension, and strike-slip plate motions [*England et al.,* 1985] have yielded striking relationships between the length scales of penetrative continental deformation, the directions of relative plate motions, and lithosphere rheology. Deformation fields associated with compressional and extensional plate interactions may be four times wider than those associated with transcurrent plate motions. Widths of intraplate deformation were also affected by the stress exponent, decreasing approximately as $n^{-1/2}$.

Crustal thickening in compressional regimes associated with continental collisions have been modeled using this same thickness-averaged model [*Houseman and England,* 1986*b*; *England and Houseman,* 1986], as well as a thin visco-plastic plate model [*Vilotte et al.,* 1984, 1986]. Although the rheological relations of continental crust lithologies are not known with great confidence (see next section), application of these models to the continental Indian-Eurasian collision have provided calculated distributions of crustal thickness which closely resemble the topographic patterns of the Himalayan arc and Tibetan plateau. These models have furthermore shown that once a thickened crustal plateau has formed, further increases in crustal thickness are inhibited by buoyancy forces and strain rates within the plateau are significantly reduced.

Crustal thinning and necking associated with extensional deformations of continental plates have been examined in simple layer models [*Fletcher and Hallet,* 1983; *Ricard and Froidevaux,* 1986] with plastic and power-law flow behavior. Modeling the brittle, upper lithosphere as a plastic plate and the underlying lithosphere as a power law material whose effective viscosity decreases with depth, *Fletcher and Hallet* [1983] evaluated the development and spacing of extensional flow instabilities. Choosing a plastic plate thickness of 10–15 km, consistent with the seismically-determined brittle/ductile transition within the Basin-and-Range Province of the western United States, they predicted necking instabilities with a spacing of 35–

60 km, in excellent agreement with the observed horst and graben spacings of 20–50 km.

Although simple, thickness-averaged plate models have provided extremely valuable insights into the large-scale structures and dynamics of continental deformations, more elaborate rheological models will be required to evaluate these complex tectonic regimes.

3. RHEOLOGICAL STRATIFICATION OF THE CONTINENTAL LITHOSPHERE

Even the oceanic lithosphere with its thin crust and its simple mineralogy dominated by olivine and pyroxenes is not likely to be rheologically monolithic. Systematic variations in environmental parameters such as lithostatic pressure (vertical normal stress σ_{zz}), the state of stress σ_{ij}, fluid pressure P_f, temperature T and the chemical effects of reactive fluids can give rise to spatial variations in the relative activities of inelastic processes, processes that place limits on the stresses that can be supported by the lithosphere. These processes include jointing, hydraulic fracturing, brittle shear faulting, "ductile faulting," semi-brittle deformation (distributed microfracturing and intracrystalline plasticity), low temperature intracrystalline plasticity, high-temperature recovery creep, high-temperature transient creep and grain-size-sensitive high-temperature creep. In view of the fact that many of the rheological laws that characterize these processes are not known with confidence for olivine-bearing rocks and that even the distribution of environmental parameters that influence rock strength are not firmly established, it would not be surprising that current rheological models for the oceanic lithosphere are oversimplified [see *Chapple and Forsyth,* 1979; *Goetze and Evans,* 1979; *Ashby and Verall,* 1978; *Kirby,* 1977, 1980, 1983, 1985; *Brace and Kohlstedt,* 1980; *McNutt and Menard,* 1982; *Carter and Tsenn,* 1987; *Tsenn and Carter,* 1987].

Consider now the added complexities of the continental lithosphere. First, the thicker continental crust is mineralogically more complex, with at least ten minerals needed to describe it at the 2% level in abundance. Second, the crust has segregated radiogenic elements that are important heat sources, the distribution of which is crucial in predicting the spatial variation of temperature and hence, ductile strength. Third, the crust also tends to segregate fluids such as melts, hydrothermal fluids and CO_2 because partial melting in the mantle very effectively segregates the volatile species into melts and because lower density mafic and more acidic melts are gravitationally unstable in the mantle and rise in the crust to the point of neutral buoyancy. Also the movement of hot fluids affects the thermal structure. Fourth, a whole host of petrological and geochemical processes attend the presence and movement of hot, chemically aggressive fluids, processes that include melt wetting of grain boundaries, hydrothermal alteration, metasomatism, hydrothermal dissolution and crystal growth and intracrystalline diffusion of hydrogen, water and related species. These petrological and geochemical processes give rise to a spectrum of weakening processes, such as hydrolytic weakening, chemically assisted crack growth, solution transfer creep, melt transfer creep, and solute effects on creep processes [see reviews by *Sibson,* 1984 and *Kirby,* 1983, 1984, 1985].

A simplified view of the rheology of the continental lithosphere is to consider only the effects of gross crustal mineralogy, neglecting the physical and chemical effects of fluids. Olivine retains high strength to temperatures as high as 1000–1200°C at typical laboratory strain rates and high confining pressures. This is in contrast with the thermal weakening of crustal rocks and minerals (Table 3) at temperatures as much as 500°C below the corresponding weakening temperature T_c of olivine [*Bird,* 1978; *Brace and Kohlstedt,* 1980; *Chen and Molnar,* 1983; *Kirby,* 1985; *Carter and Tsenn,* 1987]. These interpretations of the rock-mechanics literature have brought rock-mechanics support to the concepts of a *crustal asthenosphere* and the interpretation of the crust-mantle boundary as a possible *rheological discontinuity.* If the temperature at the crust mantle boundary is below T_c for olivine but above T_c for the rocks appropriate to the lower crust, then the lower crust will be weak and the mantle below the crust-mantle boundary will be comparatively strong.

The above interpretations of the rock-mechanics data have been motivated by independent geological and geophysical observations that suggest locally weak lower continental crust and, at the same time, strong uppermost mantle. These observations include:

1) Evidence for decoupling of upper crust from the upper mantle during large-scale thrusting connected with continental collisions and evidence for large-scale intraplate thrust faults soling into the lower crust [*Bird,* 1978].
2) Stress relaxation in the middle to lower crust implied by the vertical deflections in response to rapid changes in small surface loads on continental interiors, loads such as glacial lakes [*McAdoo,* 1985, 1987].
3) On a larger scale and over longer load duration, the evidence for isostatic compensation in the lower crust suggested by the relatively uniformly high topography in the northern Himalayas and Tibetan plateau [*Bird,* 1978].
4) Small wavelength scales of basin-and-range topography in the extensional tectonic regime, implying flow within the lower continental crust in connection with "pinch and swell" extensional deformation [*Zuber et al.,* 1986], with large wavelength features corresponding to flow in the mantle asthenosphere below.
5) The general lack of seismicity in the lower crust [*Sibson,* 1982, 1983, 1984*a,b*; *Meissner and*

TABLE 3. Steady-State Flow Law Parameters for Crustal Rocks and Minerals

$$\dot{\epsilon}_s = A\sigma^n \exp(-H^*/RT)$$

Material[a]	$\log_{10} A$ ($\mathrm{MPa}^{-n}\ \mathrm{s}^{-1}$)	n	H^* ($\mathrm{kJ\ mol}^{-1}$)	Comments	Ref.†
Albite rock	18	3.9	234		1
Anorthosite	16	3.2	238		1
Quartzite	9.0	2.0	167	α-quartz field	1
	11	2.9	149	α-quartz field	2
	6.9	1.9	123	α-quartz field	3
	10.4	2.8	184	α-quartz field	4
	—	4	300	β-quartz, vacuum dried at 800°C	5
	—	—	195	α-quartz field, transient strains to 0.8%	6
	—	—	51	β-quartz, transient strains to 0.8%	6
Quartzite (wet[b])	10.4	2.4	160	α-quartz, water from talc	2
	10.8	2.6	134	α-quartz, 0.4 wt. % water added	5
	9.1	1.8	167	α-quartz, 0.4 wt. % water added	3
Aplite	12	3.1	163		1
Westerly granite	8.5	2.9	106	α-quartz field	7
	6.4	3.4	139	α-quartz field	3
	—	—	165	α-quartz field, transient strains to 0.8%	6
	—	—	44	β-quartz, transient strains to 0.8%	6
Westerly granite (wet[b])	7.7	1.9	137	α-quartz field	3, 8
Quartz diorite	11.5	2.4	219	α-quartz field	3
Biotite single crystals	-19	10	30	compression direction at 45°to (001)	9
Clinopyroxenite	17	2.6	335		1
	-260	83	220	at 230°-900°C	10
	-5	5.3	380	at 800°-1100°C	10
Clinopyroxenite (wet[b])	5.17	3.3	490		11
Diabase	17	3.4	260		1
Carrara marble	48.6	7.6	418	drying procedure not described	12
	33.2	4.2	427	drying procedure not described	12
Natural rocksalt	-7.24	4.10	33.6	Avery Island	13, 14, 15
	-6.82	1.39	28.8	Paradox Formation	13, 14, 15
	-2.33	4.50	72.0	Permian Basin	13, 14, 15
	-1.59	5.01	82.3	Richton Dome	13, 14, 15
	-5.41	4.90	50.2	Salado Formation	13, 14, 15
	-2.06	2.22	62.9	Vacherie Dome	13, 14, 15
Synthetic rocksalt	—	—	37, 74	for n = 6, 3, respectively	16
	-0.7	5.8	96	pure NaCl	17
	-1.4	6.5	126	NaCl (+0.3% K^+)	17
	0.8	4.6	115	NaCl (+0.3% Mg^{2+})	17
	-3.9	5.7	72	NaCl (+0.3% Ca^{2+})	17
Anhydrite	—	1.5-2.0	114-152		18
Bischofite	—	4.4	59	at $\sigma < 3.0$ MPa	19
	—	1.5	67	at $\sigma > 1.5$ MPa	20
Carnallite	—	4.5	—		20
Ice I_h	-2.8	4.7	36	at $T \leq 195$ K	21
	5.10	4.0	61	at 195-240 K	21
	11.8	4.0	91	at 240-258 K	21

†References: 1. Shelton and Tullis [1981], 2. Koch [1983, manuscript], 3. Hansen and Carter [1982], 4. Jaoul *et al.* [1984], 5. Kronenberg and Tullis [1984], 6. Ross *et al.* [1983], 7. Carter *et al.* [1981], 8. Hansen and Carter [1982], 9. Kronenberg *et al.* [1985], 10. Kirby and Kronenberg [1984], 11. Boland and Tullis [1986], 12. Schmid *et al.* [1980], 13. Pfeifle and Senseny [1982], 14. Handin *et al.* [1986], 15. Wawersik and Zeuch [1986], 16. Gangi [1983], 17. Heard and Ryerson [1986], 18. Müller *et al.* [1981], 19. Urai [1983], 20. Urai [1985], 21. Kirby *et al.* [1987].

[a] All samples oven dried at 100°-200°C before testing unless noted otherwise.

[b] "Wet" samples: water added in sealed capsule, unless noted otherwise.

Strehlau, 1982; *Chen and Molnar,* 1983] and the occurrence of mantle earthquakes in the Tibetan plateau and other localities around the world [*Chen and Molnar,* 1983], suggesting locally a weak lower crust and strong uppermost mantle.

6) The large theoretical effect of yielding in the lower continental crust on reducing the resistance to bending of the continental lithosphere [*DeRito et al.,* 1986] indicates that internal yielding must be considered in flexural models of the continental lithosphere with relatively thick crust.
7) Reconciliation of the average deviatoric stress levels due to geodynamic forces and topographic loads and the yield stresses of crustal materials based on experimental rock mechanics suggests that stress relaxation can take place in the lower crust, amplifying the deviatoric stress by reducing the thickness of the load-bearing section of the continental lithosphere [*Kusznir and Park,* 1984].
8) The relatively narrow zone of accumulation of strain and its release along the San Andreas fault suggests a viscoelastic response of the middle to lower crust to plate-scale loading [*Turcotte et al.,* 1984].
9) The preferential rifting of continental crust and lithosphere compared to the oceanic lithosphere, leading to ridge jumps, the formation of new ocean basins, and the development of micro-continents and displaced terranes [*Vink et al.,* 1984]. Despite the steeper average geothermal gradients of the oceanic regions, the extensional loads required to rift continents, consisting of crustal lithologies of substantial thicknesses overlying mantle lithologies, appear to be lower than those required to rift oceanic plates made up primarily of olivine and pyroxenes.

4. STRAIN SOFTENING AND STRAIN LOCALIZATION IN SHEAR-ZONES

Geological and geophysical observations in the last decade have provided compelling evidence that large deformations are accommodated by the continental crust through faulting involving strain localization in shear zones both in the shallow crust, involving "brittle" faulting and crustal seismicity and in the mid-crust, producing primarily aseismic macroscopically ductile deformation in shear zones [see reviews in *Carreras et al.,* 1980; *Sibson,* 1977, 1982, 1986; *Kirby,* 1985]. These observations include the study of "ductile" shear zones in deep continental crust exhumed by uplift and erosion, comparison of crustal deformation rates (based on geologic and geodetic observations) with seismicity and the developing concept that the loading of the seismogenic zone involves deeper aseismic strain localization. It follows, then, that understanding the earthquake source and the overall non-hydrostatic stresses supported by the crust depends on improvements in our knowledge of how shear zones are created and what is their specialized rheology.

Extreme strain localization in shear zones is demonstrated by offsets and "drag" in pre-existing strain markers that cross these zones as well as the exclusive presence of shear-zone deformation features that are known only to develop in the laboratory at very high strain. What are the characteristics of the shear-zone materials compared to the rock matrix, and what do these tell us about the causes of the "soft" shear zone rheology? These characteristics are: 1) *Extreme grain size reduction.* In brittle faulting, this is caused by microfracturing and associated grain comminution. In deeper shear zones, recrystallization and the creation of new grains of new minerals cause grain size reduction. 2) *Other aspects of rock texture,* such as more extreme foliation development as defined by grain shape or mineral distribution, are also distinctive. 3) *Preferred orientations of ductile minerals* such as quartz are usually more strongly developed than in the host rock and bear clear orientation relationships to the plane of shear and displacement direction of the shear zone [for recently-published examples, see *Evans and White,* 1984; *Law et al.,* 1984, 1986; *Burg,* 1986; *Schmid and Casey,* 1986; *Platt and Behrmann,* 1986]. 4) *The mineralogy and mineral chemistry* of shear zones is typically different than the host rock from which it was derived [*Brodie,* 1980; *Beach,* 1980; *White et al.,* 1980; *Rubie,* 1983; *Knipe and Wintsch,* 1985; *Watts and Williams,* 1983]. This reflects evidently greater access of aqueous solutions to the zone and/or enhanced kinetics of metamorphic reactions.

How do shear zones nucleate? Three factors appear to be involved here. First, the generally non-linear stress-strain rate relations of rocks (as outlined in earlier sections) would tend to promote localization if a shear zone is only moderately softer than the host rock [see *Kirby,* 1985, p. 16]. Moreover, non-linear materials exhibit more localized deformation even in the absence of shear zone softening [*Melosh,* 1980]. Second, pre-existing zones of weakness can facilitate strain localization by providing stress concentrations as these flaws are exploited and grow as shear faults. For example, higher-than-regional non-hydrostatic stress (and related higher strain rates) can promote finer recrystallized grain size or aid the kinetics of metamorphic reactions, both of which can produce a softer shear zone rheology. Pre-existing fractures, in addition to their role as stress concentrators, can localize later shear deformation by providing access of hydrothermal solutions and making possible a variety of water-weakening processes in the zone adjacent to the fracture [*Segall and Pollard,* 1983; *Segall and Simpson,* 1986]. Distributed microcracking, fluid infiltration, and localized ductile deformation connected with hydrothermal alteration may be processes that occur simultaneously or cyclically in shear zones [*White and White,* 1983; *Etheridge et al.,* 1984].

Once a shear zone is established, what deformation processes and structural features cause the continued strain localization? A host of localization factors are now recognized [see reviews by *White et al.*, 1980; *Kirby*, 1985; *Sibson*, 1986]. Most of these softening mechanisms come into play above some critical strain and this strain softening is an important part of the mechanics of strain localization [*Poirier*, 1980]. These strain mechanisms include:

1) *Softening caused by the direct effect of grain boundary migration* associated with recrystallization or the growth of new phases. Grain boundary migration can soften a crystalline aggregate by sweeping out dislocations created by earlier deformation, reducing the hardening effects of dislocation interactions in a manner analogous to the softening effects of annealing recovery. This softening mechanism is, in a sense, an extended primary creep. Examples in metals have been cited by *White et al.* [1980] and *Urai et al.* [1986], in ice by *Duval* [1979, 1981] and *Kirby et al.* [1987], and in silicates by *Zeuch* [1982, 1983] and *Tullis and Yund* [1985].
2) *Softening stemming from grain size reduction.* The formation of a gouge zone due to microfracturing and grain comminution is a familiar feature of brittle faulting and it is apparently the micromechanics of fine granular material under shear that governs the softer "rheology" of gouge zones compared to that connected with distributed microfracturing in the host rock. Mylonitic zones formed by recrystallization processes may also be softer because deformation mechanisms that are favored by fine grain size may operate, such as grain boundary sliding or those involving stress-directed diffusion to and from grain boundaries. To date, no firm evidence has been put forward proving that these deformation mechanisms operate in mylonitic rocks and, to the contrary, the strong preferred orientations often developed in quartz-bearing mylonites favor intracrystalline slip as the dominant deformation process [see references cited earlier]. *Kronenberg and Tullis* [1984] have studied grain-size effects on the steady-state strength of quartz aggregates under hydrothermal conditions and advanced the hypothesis that diffusion from wetted grain boundaries into grain interiors is a factor controlling strengths in their samples. Obviously fine grain size should facilitate such a process and lead to shear-zone softening. It is unclear what processes maintain fine grain sizes that are acquired at peak stress or recrystallization and further deformation occurs at lower stress via these grain-size sensitive mechanisms [see *White et al.*, 1985 for discussion of this issue].
3) *Softening caused by mineral preferred orientation,* often termed geometrical softening [*White et al.*, 1980; *Poirier*, 1980]. Grain orientations in a simple-shear setting progressively become more favorable for further intracrystalline slip as total shear strain increases because grains rotate to place the operating slip systems in orientations with higher resolved shear stress. This apparently is the major source of softening connected with reorientations associated with recrystallization and intracrystalline slip in shear experiments on ice [*Duval*, 1981], calcite [*Wenk and Takeshita*, 1984] and metals [see reviews by *White et al.*, 1980 and *Poirier*, 1980].
4) *Reaction softening.* Metamorphic reactions and polymorphic phase changes can aid strain softening and lead to shear-zone localization *via* a number of processes that attend phase changes. These include: (A) *Changes in texture*, especially reduced grain size promoted by transformation under stress and consequent weakening by grain-size sensitive deformation mechanisms [*White and Knipe*, 1978; *Rubie*, 1983, 1984]. (B) *Migration of grain boundaries* driven by the growth of the more favored minerals and the elimination of defects that may have work hardened the pre-existing mineral assemblage. (C) *Softening caused by latent heat released* by a transformation. (D) *Grain-scale and megascopic stresses* connected with the transformation volume changes can promote reaction rates and diffusional transport and lead to softening [*Poirier*, 1982; *Kirby*, 1985, 1987]. (E) *The transformation products may be softer* than the reactants, especially in retrograde metamorphic reactions producing phyllosilicates [*White and Knipe*, 1978; *Kirby*, 1985]. (F) The difference in free energy of hydrous transformation products and their anhydrous reactants can help drive dissolution, solute transport and growth of the hydrous assemblage along faults filled with hydrothermal fluid. This can *facilitate the accommodation of irregularities along fault surfaces* during shear displacement.

In summary, we emphasize that several factors are important in determining whether shear-zones develop: The nature of the far-field loading conditions (the tractions and displacements and their variations with time), the thermal and elastic properties of the medium, the inelastic properties of the medium including strain-softening behavior, and the existence of pre-existing flaws and heterogeneities in properties. Only a thorough continuum-mechanics approach, incorporating all of these factors, can realistically predict whether shear zones will develop in a given geological context.

5. SEISMIC ANISOTROPY AND FLOW IN THE LITHOSPHERE

During the quadrennial period, interest has been renewed in the anisotropy of seismic waves, particularly in the mantle. Progress has been spurred by the development of improved techniques for separating elastic anisotropy from regional velocity heterogeneities, by the study of fossil oceanic crust and mantle in ophiolite complexes and by developments in rock mechanics that have refined our knowledge of how preferred crystallographic mineral orientations and resulting elastic anisotropy are acquired by rocks during inelastic deformation. Observations of seismic anisotropy in the earth are important because they are revealing of the internal deformation and because preferred orientations developed during flow can greatly influence the rheological behavior of rocks.

Preferred Orientation Development and Deformation

There are many mechanisms and processes by which physical-property anisotropy can be acquired by rocks [see review by *Crampin,* 1984], but the two most important are preferred orientation development of mineral grains and preferred orientations of flaws such as cracks, both connected with inelastic deformation.

Vertical fluid-filled cracks with azimuths related to ridge orientations have been used to explain local azimuthal variations ($\lesssim \pm 4\%$) in V_p and particle-motion anomalies in the oceanic crust [*Stephen,* 1981; *White and Whitmarsh,* 1984; *Shearer and Orcutt,* 1985, 1986]. This is in spite of the fact that regional azimuthal variations in V_p have not been detected in the oceanic crust where P_n anisotropy is apparent [*Bibee and Shor,* 1976]. Such crustal anisotropy in V_p caused by crack preferred orientations probably exists in the continental crust but is masked by larger heterogeneities in lithology and V_p than occur in the oceanic crust [see papers in *Crampin et al.,* 1984]. Opening-mode (tensile) cracks nucleate and grow with preferred orientations normal to the least principal stress direction in isotropic rocks [see review by *Paterson,* 1976] and the velocity anisotropy produced by aligned cracks can be predicted from theory [see review by *Shearer and Orcutt,* 1986].

Mineral preferred orientations and hence property anisotropy generally develop under non-hydrostatic stress as a consequence of plastic deformation. The nature of the preferred orientation depends upon the plastic deformation mechanisms that operate [*Schmid,* 1982].

Intracrystalline slip leads to preferred grain orientations in mineral aggregates because slip is crystallographically oriented and because grain-grain continuity at grain boundaries requires progressive grain rotation when grains deform by shear on the slip plane. Much progress has been made in the last decade in our understanding of the relations between stress, strain and preferred orientation based on the Taylor-Bishop-Hill model for intracrystalline slip which assumes homogeneous grain deformation and minimum internal plastic work. This theory has been applied successfully to quartzite [*Lister et al.,* 1978; *Lister and Hobbs,* 1980; *Lister and Paterson,* 1979], to calcite marble [*Van Houttel et al.,* 1984; *Wagner et al.,* 1984; *Wenk et al.,* 1985, 1986] and to olivine [*Takeshita,* 1986]. What is particularly powerful about this approach is that it permits predictions of preferred orientation development for various states of stress and strain that are not easily achieved in the laboratory, and that anisotropies in the plastic *rheology* connected with preferred orientations can also be predicted [see, for example, *Wenk et al.,* 1986]. Some minerals, however, do not have sufficient slip systems to accommodate a general homogeneous strain on the grain scale and some degree of heterogeneity in grain strain is required. This has been successfully modeled by relaxing the homogeneous strain constraint in marble [*Wenk et al.,* 1986]. In any event, the ultimate preferred orientations expected from these models depend upon the operating slip systems, the type and magnitude of finite strain (uniaxial compression, extension, simple shear, etc.) and the strain path through which that finite strain was accomplished [*Schmid,* 1982].

Recrystallization under non-hydrostatic stress can lead to crystallographic preferred orientations. Early work suggested that new grains were independently nucleated and had orientations that depended on the state of stress. Research in the last decade suggests, however, that preferred orientations developed under conditions that favor recrystallization are not fundamentally different than those produced by intracrystalline slip and that grains nucleate by grain-boundary migration and/or subgrain rotation of pre-existing grains [*Urai et al.,* 1986; *Wilson,* 1986; *Schmid et al.,* 1987; *Burg et al.,* 1987]. Definitive experiments have not been done to explore the comparative roles of stress and strain in preferred orientation development during recrystallization, but the foregoing observations suggest that finite strain is the primary determinant of preferred orientations produced during recrystallization.

Grain boundary sliding, GBS, is a deformation process that depends upon accommodation mechanisms that allow necessary grain shape changes and is favored by small grain sizes. Experience in metals and in fine-grained rocks that are thought to deform by GBS shows that the process does not, of itself, lead to preferred orientations; on the contrary, GBS can randomize a pre-existing fabric [*Boullier and Nicolas,* 1975; *Gueguen and Boullier,* 1976; *Schmid et al.,* 1977, 1981, 1987; *Schmid,* 1982]. Weak preferred orientations can develop if GBS preferentially promotes another deformation process, such as slip [*Schmid et al.,* 1987]. Other grain-size-sensitive deformation processes, ones involving stress-directed diffusion to and from grain boundaries, are also not expected to develop preferred orientations.

Not only are rock fabrics dependent on the operating deformation mechanisms but they are also dependent on the relation between the state of stress and the finite strain state. Of particular interest is whether the stress and strain states are *coaxial* or *non-coaxial* (*i.e.,* whether the principal stress and principal finite strain directions are parallel to each other). For example, intracrystalline slip under uniaxial compression or extension (coaxial) results in the progressive rotation of the operating slip plane(s) normal toward the compression direction and, for minerals deforming primarily by one slip system, creep rates should decrease with strain and should never reach steady state in the absence of grain boundary migration and recrystallization. In contrast, the progressive rotation of slip planes toward the shear plane by intracrystalline slip in a simple shear environment can lead to a steady-state preferred orientation and creep rate. Simple shear (biaxial or torsion) experiments and measurement of resultant fabrics have been done in a number of non-metallic materials [ICE: *Kamb,* 1972; *Byers,* 1973; *Lile,* 1978; *Duval,* 1981; *Bouchez and Duval,* 1982; *Burg et al.,* 1987; CALCITE: *Kern and Wenk,* 1983; *Schmid et al.,* 1987; QUARTZITE: *Dell'Angelo and Tullis,* 1987] and the results are generally consistent with the above predictions. The development of preferred orientations causes materials to exhibit a transient rheological response to changes in the stress state [*Griggs and Miller,* 1951; *Handin and Griggs,* 1951; *Heard and Raleigh,* 1972; *Byers,* 1973; *Duval,* 1981; *Duval and Le Gac,* 1982; *Gao and Jacka,* 1987]. This may be important in the deformation of the oceanic lithosphere where preferred orientations caused by basal shear deformation connected with plate motion or by deformation along transform faults could influence the rheological response of the lithosphere to changes in the stress state such as the bending deformation at trench-rise systems or at island loads.

Seismic Observations of Velocity Anisotropy

The basic seismological observations of velocity anisotropy in the uppermost mantle are reviewed by *Crampin et al.* [1984], *Kawasaki and Kon'no* [1984], *Christensen* [1984], *Nicolas* [1986] and *Kawasaki* [1986]. Foremost among them is the azimuthal variation in P_n velocity in the oceanic mantle, first interpreted in the eastern Pacific by *Hess* [1964], and confirmed by subsequent refraction surveys in the same region [*Raitt et al.,* 1969; *Morris et al.,* 1969; *Raitt et al.,* 1971; *Keen and Barrett,* 1971; *Bibee and Shor,* 1976; *Clowes and Au,* 1982]. More recently, a prominent P_n anisotropy was shown to apply to the western Pacific and Sea of Japan as well [*Shimamura et al.,* 1983; *Shimamura,* 1984; *Hirahara and Ishikawa,* 1984; *Okada et al.,* 1978; *Shearer and Orcutt,* 1985, 1986]. The direction of maximum P_n with rare exception [*Whitmarsh,* 1971; *Talandier and Bauchon,* 1979] is approximately perpendicular to the magnetic lineations between the source and receiver and peak-to-trough variations of 3–10% with azimuth are typically observed. Similar observations of Pn anisotropy have been made in the continental lithosphere in southern Germany [*Bamford,* 1977; *Fuchs,* 1983], the western U.S. [*Bamford et al.,* 1979] including southern California [*Vetter and Minster,* 1981; *Hearn,* 1984] and indirect evidence for P-wave anisotropy in northern Australia [*Leven et al.,* 1981]. Analysis of P-wave travel-time data worldwide by *Dziewonski and Anderson* [1983] suggest that P-wave velocity anisotropy may be deep-seated in the upper mantle and vary smoothly in relation to tectonic provinces in the continental lithosphere. In the examples of southern California and Germany the direction of maximum P_n approximately parallels the traces of plate-scale faults.

Although high values (4.9 ± 0.1 km s^{-1}) of the shear-wave phase S_n have been measured in the western Pacific [*Shimamura et al.,* 1977; *Shimamura and Asada,* 1983], suggesting S_n velocity anisotropy, the three other studies of S_n elsewhere show typical values of 4.6 ± 0.1 km s^{-1}, independent of direction, even though P_n varies significantly in the same regions [*Clowes and Au,* 1982; *Talandier and Bouchon,* 1979; *Shearer and Orcutt,* 1986]. Other effects of elastic anisotropy on body waves include split shear waves with different polarizations and velocities and particle motions of P-waves that are out of the vertical plane connecting the source and receiver [*Shearer and Orcutt,* 1985]. Shear-wave polarization anisotropy has been observed for steeply inclined S and ScS phases from deep earthquakes beneath Japan [*Ando et al.,* 1980, 1983; *Fukao,* 1984] and other areas worldwide [*Ando,* 1984]. The depth over which the polarized S-wave splitting of ScS phases are acquired is not known but is likely to be in the upper mantle because lower mantle minerals are not known to be particularly anisotropic [*Jeanloz and Thompson,* 1983] and preferred orientations of some mantle analogue materials are not especially strong [*Toriumi,* 1984]. Also, since the splittings of direct S-waves are similar to those for ScS waves this suggests that the delays between the polarized phases occurs within the upper mantle [*Ando,* 1984]. Also, the polarization direction of the fastest ScS wave is approximately parallel to the direction of maximum P_n velocity offshore east of Japan [*Shimamura et al.,* 1983], again suggesting that time separations between the split ScS phases occur in the upper mantle.

Forsyth [1975] and *Kawasaki and Kon'no* [1984] have detected a significant azimuthal variation in Rayleigh wave group velocities in overlapping areas of the eastern Pacific, with those surface waves traveling approximately parallel to the prominent fracture zones (and approximately normal to the magnetic lineations) being about 2–3% faster than those traveling perpendicular to those directions. Rayleigh wave studies over oceanic paths of greatly variable spreading directions not surprisingly have failed to detect azimuthal anisotropy in Rayleigh wave velocity [*Schlue and Knopoff,* 1976, 1977; *Mitchell and Yu,* 1980; *Anderson and Regan,* 1983]. Love wave anisotropy is always small (<1%) [*Forsyth,* 1975; *Kawasaki and Kon'no,* 1984; *Tanimoto and Anderson,* 1984, 1985] apparently reflecting the effective isotropy of SH elastic wave motion in oceanic paths.

The shear-wave velocities inferred from mantle Rayleigh wave and Love wave dispersion data are different, with SH values consistently higher than SV in oceanic paths (see review by *Anderson and Dziewonski* [1982]). The spatial (especially depth) distribution of this anisotropy inferred from surface wave data is dependent on the specific inversion model; some suggest an isotropic seismic lithosphere and anisotropic upper asthenosphere [*Schlue and Knopoff,* 1977, 1978; *Anderson and Regan,* 1983; *Regan and Anderson,* 1984] and others infer lithosphere anisotropy [*Yu and Mitchell,* 1979; *Mitchell and Yu,* 1980; *Forsyth,* 1975; *Kawasaki,* 1986]. Many of these surface-wave studies have assumed that the oceanic mantle is transversely isotropic with velocities for propagation in the horizontal plane averaged and deemed isotropic and having a unique vertical direction with different velocities than in the horizontal. This probably captures the differences in the vertical and averaged horizontal velocities but de-emphasizes the important azimuthal variations in velocity, a point raised by *Tanimoto and Anderson* [1984], *Kawasaki and Kon'no* [1984], *Kawasaki* [1986] and *Estey and Douglas* [1986], as noted below.

Field Measurements of Preferred Orientations and Anisotropy in Mantle Materials

Paralleling remote measurements of velocity anisotropy in the oceanic lithosphere have been direct studies of structures and textures in ophiolite complexes, for which persuasive arguments have been put forward that they represent oceanic lithosphere emplaced in the crust by large displacement thrust faulting (see reviews by *Christensen* [1984] and *Nicolas* [1986]). The basal peridotites representing oceanic mantle generally show well-developed deformation textures and marked regional preferred orientations of olivine and less well-developed pyroxene fabrics [*Christensen,* 1984]. The olivine fabrics generally show an $\underset{\sim}{a}$-axis maximum approximately parallel to the crust-mantle boundary and perpendicular to the sheeted dikes in the crustal section (and presumably parallel to the paleo-spreading direction). The $\underset{\sim}{b}$- and $\underset{\sim}{c}$-axes range from point maxima to partial girdles around the $\underset{\sim}{a}$-axis maxima, indicating orthorhombic to uniaxial symmetry of the crystallographic orientations. Similar olivine preferred orientations are also observed in mantle peridotite xenoliths from the continental and oceanic lithosphere [*Mercier and Nicolas,* 1975; *Peselnick et al.,* 1977]. Olivine is extremely anisotropic in its elastic properties and the resulting anisotropy in the velocity of elastic wave propagation shows similar symmetries. Shear-wave velocities depend on the polarization direction and, in general, two mutually-polarized shear waves travel at different velocities. Except for propagation directions parallel to the crystallographic axes, particle motion is not purely compressional or purely shear but mixed and the structure imposes the polarization directions on the two shear waves. V_p varies from 9.9 km s^{-1} parallel to $\underset{\sim}{a}$ to 7.7 km s^{-1} parallel to $\underset{\sim}{b}$, while shear waves vary from 4.9 to 4.6 km s^{-1} in the same propagation directions, averaged over all polarization directions for those propagation directions. Quasi-shear wave velocities as high as 5.5 km s^{-1} can occur for off-axis wave normals [*Leven et al.,* 1981].

The relation between crystallographic preferred orientation in polycrystalline olivine and the resultant anisotropy in elastic wave velocity is now well established in theory [*Kumazawa,* 1964; *Crossin and Lin,* 1971; *Baker and Carter,* 1972; *Carter et al.,* 1972; *Peselnick and Nicolas,* 1978; *Crampin,* 1981; *Johnson and Wenk,* 1985, 1986; *Bunge,* 1985; *Kern and Wenk,* 1985] and experiment [*Christensen,* 1966, 1971; *Christensen and Ramananantoandro,* 1971; *Peselnick et al.,* 1974; *Peselnick et al.,* 1977; *Meissner and Fakhimi,* 1977; *Peselnick and Nicolas,* 1978]. The effects of pyroxenes, spinel and garnet have been investigated and are known to dilute the anisotropy due to olivine preferred orientation; the degree of dilution depends on the mineral proportions and pyroxene preferred orientation [*Leven et al.,* 1981; *Christensen and Lundquist,* 1982; *Fuchs,* 1983; *Christensen,* 1984; *Estey and Douglas,* 1986].

The orthorhombic to uniaxial symmetries of the preferred orientations of olivine in ophiolite peridotites correspond to the same symmetries in seismic velocities with direction. Based on the hypothesis that the preferred orientations and velocity anisotropies in ophiolite peridotites represent those of the oceanic lithosphere ($\underset{\sim}{a}$-axis maximum parallel to the spreading direction at the time of lithosphere formation) and that the mafic-ultramafic contacts in ophiolites were originally horizontal, the anisotropic seismic-velocity behavior of the oceanic mantle lithosphere can be predicted with surprising fidelity. In particular the P_n anisotropy of 3–10% observed in refraction experiments in the oceans is consistent with an olivine $\underset{\sim}{a}$-axis maximum typical of ophiolite peridotites parallel to spreading direction and an isotropic distribution of $\underset{\sim}{b}$ and $\underset{\sim}{c}$ axes normal to the spreading direction, diluted by 0 to 40% pyroxene [*Christensen,* 1966; *Christensen and Crossen,* 1968; *Crossen and Christensen,* 1969; *Christensen and Salisbury,* 1979; *Christensen and Lundquist,* 1982; *Kasahara and Kon'no,* 1984; *Kasahara,* 1986; *Estey and Douglas,* 1986; *Shearer and Orcutt,* 1986]. The predicted azimuthal variation of S_n body waves and Love surface waves is smaller than the resolution in measuring the velocities of those phases [*Kawasaki and Kon'no,* 1984; *Kawasaki,* 1986; *Shearer and Orcutt,* 1986], whereas the time delays of split *ScS* would be detectable. Lastly, the polarization anisotropy of shear waves predicted for the uniaxial model is within a range consistent with observation (0–0.2 km s^{-1}) [*Kawasaki and Kon'no,* 1984; *Kawasaki,* 1986]. *Estey and Douglas* [1986] have proposed an anisotropy model in which olivine and pyroxene have preferred orientations of orthorhombic symmetry with olivine $\underset{\sim}{a}$ and pyroxene $\underset{\sim}{c}$ axes parallel to spreading direction and olivine $\underset{\sim}{b}$ and pyroxene $\underset{\sim}{a}$ axes vertical and normal to the Moho, based on the expected easy slip systems in these minerals. This model is, however, at variance with the experience in ophiolite complexes that olivine $\underset{\sim}{b}$ and $\underset{\sim}{c}$ axes show partial girdles about the $\underset{\sim}{a}$-axis maximum or point maxima with no particular relation of olivine

$\underset{\sim}{b}$ axes with respect to vertical [*Christensen*, 1984; *Nicolas*, 1986]. Moreover, the quasi S-wave velocities for an orthorhombic model vary from $QSH = 4.86$–5.51 km s^{-1} for horizontally polarized waves traveling in the (010) plane and $QSV = 4.42$–4.89 km s^{-1} for vertially polarized waves [*Leven et al.*, 1981], a much wider range than actually observed. In particular, the QSH anisotropy is inconsistent with the lack of evidence for Love wave anisotropy.

Tectonic Models for Velocity Anisotropy of the Oceanic Lithosphere

Given the success of the uniaxial preferred orientation model for olivine in predicting the primary features of elastic wave anisotropy of the oceanic lithosphere, what are its implications for the state of stress and strain in the oceanic mantle? Various models have been put forward to account for azimuthal P_n anisotropy of the oceanic lithosphere:

1) *Hess* [1964] suggested that plastic flow associated with *simple shear along oceanic fracture zones* (with shear direction parallel to the fracture zone) causes P_n anisotropy, pointing out that fabrics of foliated olivine-bearing rocks often show preferred orientations consistent with fast V_p parallel to the fracture zone.
2) *Francis* [1969] noted that Hess' mechanism is unlikely to pervade the entire oceanic lithosphere and that *basal shear strain connected with plate motion* could produce $\underset{\sim}{a}$-axis maxima parallel to the direction of plate motion, consistent with the easy slip direction in olivine (and ophiolite studies), thereby producing fast V_p in the direction of plate motion. *Ishikawa* [1984] has followed up on this idea by including thickening of the lithosphere and freezing in of basal-shear deformation connected with plate motion as the lithosphere cools and the zone of active shear deformation deepens with age. Analysis of long-period surface wave dispersion data by *Regan and Anderson* [1984] and *Tanimoto and Anderson* [1984] suggests that the fast $\underset{\sim}{a}$-axis direction aligned parallel to the flow direction may also be deep seated in the upper mantle and consistent with modern numerical models for convection in the asthenosphere.
3) *Avé Lallemant and Carter* [1970] and *Carter et al.* [1972] considered the expected *preferred orientation of olivine due to recrystallization in relation to the stress state presumed to occur in the lithosphere connected with basal shear.* As noted earlier, it is more likely that preferred orientations develop with reference to the finite strain (flow field) and strain path, and hence the preferred orientations predicted by the above authors are probably incorrect.
4) *Ida* [1984] suggests that P_n anisotropy is caused by *plate stretching parallel to the direction of plate motion.* This is unlikely because large stretching strains would be required to develop significant preferred orientations and there is no evidence for such stretching deformations.

To summarize, the basal-shear model of *Francis* [1969], as refined by *Ishikawa* [1984] and *Anderson* and his colleagues, is consistent with the seismic constraints and the preferred orientation model of *Kawasaki* [1986]. The latter appears to account for the first-order observations of body-wave and long-period seismology.

6. ROCK MECHANICS: GUIDE TO THE LITERATURE

Laboratory studies of the mechanical properties of rocks over the quadrennial period have been extensive, encompassing the fracture, frictional behavior, and flow of rocks and minerals. The emphasis of much of this work has been towards understanding deformation mechanisms and establishing physically-based constitutive relations. Major advances along these lines have been made in our understanding of hydrolytic weakening and the effects of chemical environment upon surface states and internal defects which affect the deformation processes. Steady-state rheologies relevant to the oceanic lithosphere and upper mantle are summarized in Tables 1 and 2, and rheologies of crustal rocks and minerals are summarized in Table 3.

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6.1 ROCK FRACTURE

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6.2 ROCK FRICTION

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