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EARTHQUAKES AND ROCK DEFORMATION IN CRUSTAL FAULT ZONES

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INTRODUCTION

The physical origin of earthquakes lies ultimately in the geological structure of fault zones and the deformation processes that occur therein in response to tectonic stress. Although the possibility now exists that the nucleation regions for large earthquakes at depths of ~ 10 km may shortly become directly accessible by deep drilling, our present knowledge of fault structure and the shallow earthquake source is derived largely from a variety of *indirect* sources. These include seismological studies, surface studies of fault zones and earthquake ruptures, geodetic information on modes of fault slip, geophysical constraints on fault zone structure and rheology, and information garnered from materials science and experimental rock deformation.

Studies of fault zone structure and the rock products of faulting provide complementary information on deformation processes at depth in fault zones. Although descriptions of fault rocks are widespread in the geological literature (e.g. Spry 1969, Higgins 1971), it is only in the past decade that they have begun to be interpreted in the context of the physical conditions and processes prevalent in seismically active fault zones at different crustal depths (Sibson 1977, Watts & Williams 1979, Anderson et al 1983, Wise et al 1984). Such interpretations are still at an early stage, but the deformation textures and structural associations of fault rocks have already been shown to have the potential to yield information on such diverse topics as shear stress levels, power dissipation, and seismic efficiency during earthquake faulting; on fluid pressure levels and episodic fluid flow accompanying

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faulting; on physical mechanisms for earthquake rupture arrest; and on factors governing the depth of the seismogenic zone in continental crust.

This review seeks to demonstrate the extent to which field data from fault rocks and associated structures can be integrated with information from the other sources to build *conceptual fault zone models* accounting for a wide range of observed fault behavior. It is not intended to be a comprehensive review of such studies but, rather, a demonstration of our present level of understanding, focusing in particular on the areas of uncertainty and the problems in interpretation.

SAMPLING CONSIDERATIONS

Methodology

In developing conceptual fault zone models, the general aim is to establish for appropriate host rocks the changes in macroscopic faulting style (e.g. slip on discrete planes, homogeneous or heterogeneous shear across zones of finite width) and the associated mineral deformation mechanisms that occur with depth for both seismic and aseismic slip modes. Information of this kind may be gathered from ancient fault zones exposed at different erosion levels or from dip-slip fault zones, perhaps still active, which have associated with them rock products of faulting generated over a range of depths. Note that fault rocks derived from deeper crustal levels should generally occur on the hanging walls of reverse-slip faults (e.g. Sibson et al 1979) and on the footwalls of normal-slip faults (e.g. Davis 1983). The distribution, metamorphic state, and textural-microstructural characteristics of different fault rocks may then be correlated with the associated style of faulting to build up fault models (Figure 1).

Fault Zone Evolution

Conceptual models relating to long-established fault systems are of greatest interest because the largest earthquake ruptures and most intense subsidiary seismicity are localized within tabular fault zones, commonly 10–10^3 m in thickness, that have presumably undergone considerable strain

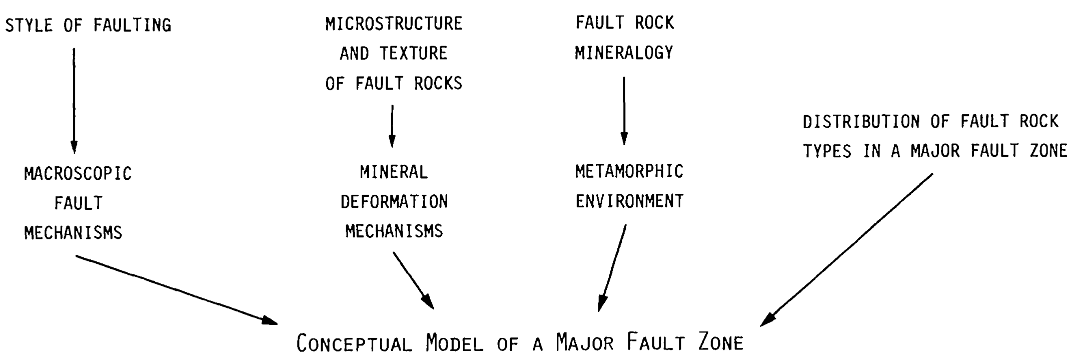

Figure 1 Methodology for evolving conceptual models of fault zones.

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softening with respect to the surrounding crust. Our prime concern, then, is with the *residual* infrastructure of mature fault zones, which have arisen from the progressive evolution and coalescence of faults, fractures, and ductile shear zones that originally nucleated in more-or-less intact crust (Segall & Pollard 1983). Major fault zones may also undergo substantial geochemical evolution with time through their role as conduits for fluid flow at all crustal levels (Beach 1980, Kerrich et al 1984). For example, progressive hydrothermal alteration of fault rocks to clay-rich assemblages within the seismogenic regime may drastically change the mechanical properties of fault zones with time (Wu 1978, Wang 1984). Unfortunately, the residual infrastructure of mature fault zones is inherently less easy to study than that of low-displacement faults and shear zones, where comparatively undamaged country rock may, for example, allow correlation of finite displacement with style and intensity of deformation (e.g. Ramsay & Graham 1970, Sibson 1975, Aydin & Johnson 1978, Gay & Ortlepp 1979, Mitra 1979, Muraoka & Kamata 1983, Simpson 1983).

Further complications arise in connection with the evolution of faults with a major component of dip-slip and with fault zones that have undergone progressive unroofing throughout their history. In such situations, fault rocks generated at one crustal level may be transported to become texturally overprinted in different deformation environments (Grocott 1977, Watts & Williams 1979). Perturbation of the thermal environment by sustained dip-slip, by shear heating, or by some combination of these effects (Brewer 1981) may cause additional problems in interpretation.

Range of Slip and Strain Rates

A major consideration in the interpretation of fault-related deformation is the enormous variety of slip and strain rates prevalent within active fault zones, which has the important implication that rates of energy dissipation within fault zones likewise vary over a broad range (Sibson 1977). Measured slip rates across active fault zones span ten orders of magnitude, from steady aseismic rates of 1–30 mm yr^{-1} to transitory seismic slip rates that may reach 0.1–2 m s^{-1} for periods of up to a few seconds (Brune 1976), repeating at intervals commonly between 10^2 and 10^4 yr. Slip at intermediate rates may also occur. Aseismic shear strain rates inferred for mylonite belts at depth range from perhaps 10^{-9} to 10^{-12} s^{-1}, but much higher rates may pertain locally where strain continuity is maintained during seismic slip.

Deformation and the Earthquake Stress Cycle

Within the seismogenic regime, which generally occupies at least the top third of continental crust (Chen & Molnar 1983, Sibson 1983), it is now

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generally accepted that the bulk of fault displacements occur by large-scale earthquake rupturing (Thatcher et al 1975, Hyndman & Weichert 1983), though aseismic slip may predominate over local fault segments, such as along the San Andreas fault in central California (Burford & Harsh 1980). Thus, as a general rule one cannot think in terms of steady progressive deformation at constant stress within the upper levels of fault zones, nor of a simple succession of deformational events. Rather, one must consider deformation in relation to multiple cycles of stress and elastic strain energy accumulation, with intermittent, abrupt stress and energy release accompanying each major rupture event.

A range of stress-time and displacement-time regimes may be envisaged for seismically active fault zones (Figure 2). Crude sawtooth stress

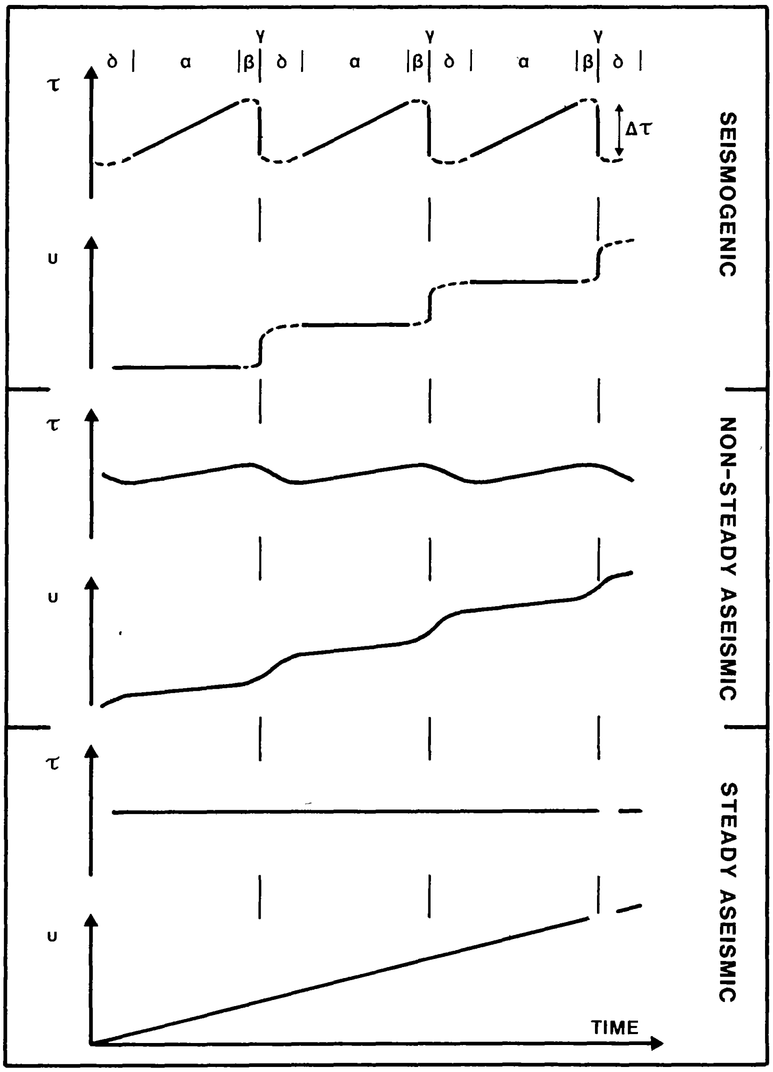

Figure 2 Hypothetical stress-time and displacement-time relationships at different levels in crustal fault zones.

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oscillations should prevail throughout the seismogenic regime and presumably to some depth beneath it, with the oscillations progressively becoming damped and smoothed. At still greater depths, aseismic shearing should proceed at near-constant stress. The average amplitude of the stress oscillation within the seismogenic regime, corresponding to the static stress drop of moderate to large earthquakes ($\Delta\tau$) is well constrained to 1–10 MPa (Kanamori & Anderson 1975, Hanks 1977). This may be equated with a characteristic release of elastic shear strain ($\Delta\gamma \sim 10^{-4}$–$10^{-5}$). However, it is important to note that at present there is no consensus on the average absolute level of tectonic shear stress (τ) within crustal fault zones; current estimates range from $\sim$10 MPa to $\sim$100 MPa (see discussion by Hanks & Raleigh 1980).

The stress cycle for a large shallow earthquake is conveniently partitioned into four phases: an α-phase of secular, mainly elastic strain accumulation; a β-phase of preseismic anelastic deformation perhaps involving dilatant cracking and terminating in foreshock activity and accelerating precursory slip; a coseismic γ-phase of mainshock slip, rupture propagation, and energy release; and a postseismic δ-phase of decelerating afterslip and aftershock activity decaying inversely with time in accordance with Omori's law (Figure 2). With a view to understanding details of the faulting process, it is clearly desirable that deformation arising from different phases of the stress cycle should be distinguished. For example, β-phase deformation might provide an explanation for observed precursory phenomena, while γ-phase deformation could potentially yield information on power dissipation during seismic slip and on earthquake efficiency (Sibson 1980a).

Unfortunately, the cyclical nature of fault deformation leads to major interpretative difficulties: Structural features developed in one particular phase of a stress cycle are liable to be overprinted or obliterated by deformation during a later phase of the same cycle or during subsequent cycles. The seriousness of this problem becomes apparent when one considers that a finite displacement of, say, 10 km might have developed by $\sim 10^4$ 1-m seismic slip increments. Clearly, features related to particular phases of the earthquake stress cycle may be easier to distinguish around faults with small total displacements, but such faults then may not have attained the mature residual infrastructure determining long-term behavior.

Foreshock-Mainshock-Aftershock Sequences

Within the seismogenic regime, the primary stress cycle for large earthquakes may be locally perturbed by subsidiary stress release accompanying foreshocks and, more particularly, by the extensive aftershock

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sequences that invariably follow large crustal earthquakes. While the cumulative slip produced by such sequences is generally subordinate to that of the main earthquake, subsidiary deformation may be distributed over a broad region surrounding the mainshock rupture. Figure 3 illustrates concentrations of aftershock activity following two moderate strike-slip ruptures that terminated in different structural configurations. It is apparent that substantial deformation may be induced well outside the causative fault zone by a large earthquake.

DEFORMATION PROCESSES IN FAULT ZONES

Crustal fault processes encompass diverse rock and mineral deformation mechanisms as a consequence of the wide range of slip and strain rates involved and of the different physical environments encountered. Much experimental rock deformation has been devoted to establishing conditions for uniform flow of rocks and constitutive flow laws (see reviews by Tullis 1979, Schmid 1982, Kirby 1983) and to studies of frictional sliding with the particular aim of understanding the instabilities leading to stick-slip behavior (Byerlee 1978, Dieterich 1981, Rice 1983, Okubo & Dieterich 1984, Shimamoto & Logan 1986). However, it is important to appreciate

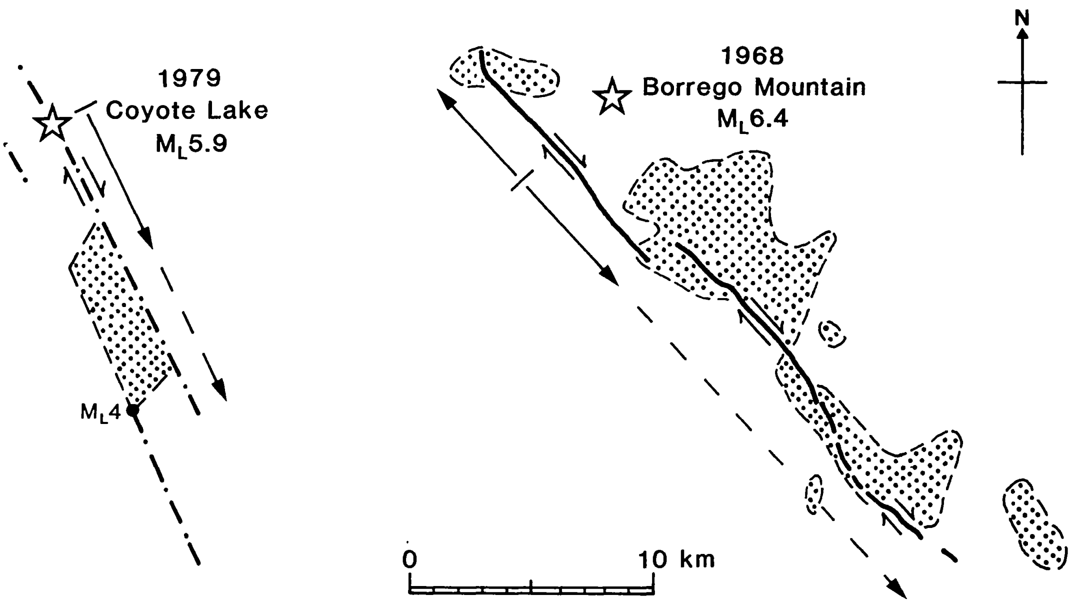

Figure 3 Seismotectonic maps illustrating aftershock concentrations associated with right-lateral strike-slip rupturing within the San Andreas fault system, California (after Sibson 1985). (*Left*) The 1979 $M_L = 5.9$ Coyote Lake earthquake rupture, which terminated in a dilational fault jog; (*Right*) The 1968 $M_L = 6.4$ Borrego Mountain rupture, which was at least partly arrested at an antidilational jog. (Epicenters represented by stars, propagation direction and extent of mainshock ruptures by arrows, surface breaks by broad lines, microearthquake lineaments by dash-dot lines, areas of intense aftershock activity by stippling.)

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that adequate laboratory simulation has yet to be achieved for many natural deformation processes for which there is abundant field evidence. Pressure solution (water-assisted diffusive mass transfer) is an important example of one such mechanism (Rutter 1983). Furthermore, laboratory friction experiments have yet to simulate the displacements and anticipated levels of frictional energy dissipation likely to accompany moderate or greater earthquake ruptures. Some physical appreciation of these fast dissipation processes can be gained by considering that slip at seismic rates of $\sim 1\ \mathrm{m\ s^{-1}}$ against an assumed constant shear resistance of 10 MPa (the lowest likely value) leads to power dissipation at $10\ \mathrm{MW\ m^{-2}}$ on the fault, albeit only for a few seconds at a time (see Sibson 1980a).

Thus, in considering fault-related deformation, we emphasize those processes for which there is abundant field evidence in the form of textures and microstructures. It is convenient to partition the deformation mechanisms into those capable of allowing steady aseismic shearing, those associated with the intermittent bursts of energy dissipation accompanying the δ slip phase, and those that may be associated with the β and γ phases of the earthquake stress cycle, with the recognition that varieties of intermediate behavior may also occur. Note also that individual mechanisms rarely operate in isolation.

Mechanisms Allowing Steady Shearing

In truly steady-state deformation, shearing would occur at a uniform rate under constant stress across a tabular zone of fixed thickness containing a statistically invariant microstructure. These requirements are unlikely to be met in most natural fault environments; we therefore relax the definition to include here all mechanisms that allow steady shearing under more-or-less constant stress. It is useful to distinguish further between mechanisms for stable sliding on more-or-less discrete surfaces and those involving volumetric shearing flow in tabular zones, although intermediate behavior is again possible and transitions may occur with progressive deformation.

PLANAR SLIDING Stable frictional sliding is well documented from experimental studies (Engelder et al 1975, Dieterich 1978) and leads to the progressive accumulation of gouge by processes of frictional wear involving ploughing of asperities into opposing walls, sidewall cracking and plucking, and shearing of asperities (Engelder 1978). Clearly, with time this may lead to shear across a cataclastically flowing gouge layer, though there are usually indications that deformation continues to be concentrated near the gouge-rock interface, with the layer growing laterally with increasing displacement. In natural, fluid-saturated fault zones it seems inevitable that subcritical crack growth involving stress corrosion (Atkinson 1982, 1984)

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becomes an integral part of frictional wear processes accompanying slow stable sliding.

Evidence in the form of fibrous growth "slickensides" coating some natural fault surfaces, and the presence of related features such as stylolites, suggests the existence of a nonfrictional mechanism for stable sliding involving dissolution of asperities, diffusive mass transfer in an aqueous phase, and precipitation into fibrous accretion steps (Durney & Ramsay 1973); this mechanism has been termed pressure solution slip by Elliot (1976). Rutter & Mainprice (1979) predict linear viscous behavior with an inverse dependence of sliding rate on grain size. They also point out that the process may lead to substantial local fluctuations in porosity, possibly affecting fluid pressures and fault stability. The frequent association of these fiber-coated fault surfaces with arrays of extension veins, interpretable as natural hydraulic fractures, suggests that this mechanism is commonly operative under conditions of very low effective confining pressure (Sibson 1980a).

SHEARING FLOW Schmid (1982) has reviewed deformation mechanisms allowing steady flow of crustal rocks. Cataclastic flow, intracrystalline dislocation creep, and different variants of grain boundary diffusion creep may contribute to aseismic shearing flow within crustal fault zones. Intracrystalline lattice diffusion (Nabarro-Herring creep) is unlikely to be significant. Deformation maps have been constructed for quartz and calcite illustrating the dominant mechanisms operating under different grain size, stress, temperature, and strain rate conditions (Rutter 1976, White 1976, McClay 1977, Etheridge & Wilkie 1979, Schmid 1982). However, there is at present no adequate means for representing fields of cataclastic flow on such maps. Moreover, the boundaries must still be regarded as preliminary, and their application to deformation of polymineralic rocks remains uncertain. In most crustal environments, flow of polymineralic rocks probably involves a mixture of mechanisms. On a further cautionary note, one should also consider inferences drawn by Etheridge et al (1984) to the effect that under metamorphic conditions, high-strain deformation invariably occurs at near-zero effective confining pressure in the presence of a mobile, high-pressure fluid. If true, this has profound implications for the rheology and strength of fault zones in the mid-to-deep crust.

Cataclastic flow This mechanism may involve both brittle fragmentation of rocks (brecciation) and comminution of mineral grains, so that grain size distribution and microstructure evolve with time (Engelder 1974). Micromechanisms involve intragranular and transgranular cracking, frictional grain boundary sliding, and grain rotation. No adequate constitutive law has yet been devised, but the mechanism is highly sensitive to effective

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confining pressure (high pressures inhibit cracking and frictional grain sliding) and is relatively unaffected by variations in temperature and strain rate. Subcritical crack growth by stress corrosion is again likely to play an integral role in natural cataclastic flow. Possible examples of cataclastic flow related to faulting have been described by Engelder (1974), Brock & Engelder (1977), House & Gray (1982), Mitra (1984), and Blenkinsop & Rutter (1986). Comparison of natural gouge infrastructure with that developed during experimental stable shearing of gouge layers has revealed similarities in the form of systematic arrays of Riedel shears and subsidiary fractures (Logan et al 1979).

Dislocation creep With the exception of calcite-rich rocks, where large strains may develop by twinning, the dominant intracrystalline deformation mechanism in crustal rocks is dislocation creep. Steady-state flow occurs by a combination of dislocation glide leading to work hardening and of processes of recovery involving dislocation climb and cross slip (see Nicolas & Poirier 1976). The recovery processes are thermally activated, so that in general steady flow is only achieved for absolute temperatures $T > 0.5\ T_m$, where T_m is the melting temperature. Flow to large strains by dislocation creep generally involves dynamic recrystallization and the development of a strong crystallographic preferred orientation, possibly with a statistically invariant microstructure (White 1976). Dislocation creep in silicates is facilitated by hydrolytic weakening through the incorporation of structural "water" into the lattice (Griggs 1967), though the details of the weakening process are not yet fully understood (Kirby 1984). Constitutive flow laws for dislocation creep take the form of a power law

$$\dot{e} = A \exp(-H/RT)\sigma^n, \tag{1}$$

where e and σ are the flow strain rate and the differential stress, respectively, A is a material constant, R is the gas constant, H is an activation energy, and $3 < n < 5$. Thus, there is no explicit dependence on grain size nor on confining pressure, though recent experimental work suggests that the diffusion rate of "water" into silicates to allow hydrolytic weakening is affected by pressure and may also lead to a grain size dependence (Blacic & Christie 1984, Kronenberg & Tullis 1984).

Textural evidence from naturally deformed rocks suggests that dynamic recrystallization accompanying dislocation creep becomes significant in quartz under greenschist facies metamorphic conditions ($T > \sim 300$°C) and in feldspars under high greenschist to amphibolite facies conditions ($T > \sim 450$°C) (Voll 1976, Tullis et al 1982, Hanmer 1982, Simpson 1985). Grain refinement accompanying dynamic recrystallization in quartz-bearing rocks appears to play a major role in promoting strain localization

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within mylonitic shear belts in the mid-to-deep crust (Bell & Etheridge 1973, White 1976, Etheridge & Wilkie 1979, White et al 1980, Tullis et al 1982).

Grain boundary diffusion Coble creep and pressure solution are two deformation mechanisms that both involve diffusive mass transfer along grain boundaries; in the latter, more geologically significant mechanism, diffusion is believed to be greatly enhanced by the presence of an aqueous intergranular film (Rutter 1976, 1983, McClay 1977). In the broad sense, pressure solution encompasses all processes of stress-controlled solution, diffusional mass transfer, and redeposition. Solute transport may occur on the scale of individual grains, or over distances of millimeters to meters with redeposition in extension veins (the crack-seal mechanism of Ramsay 1980), or the dissolved material may be flushed from the deforming system, which leads to volume losses as high as 50% (Etheridge et al 1984). For diffusive mass transfer on the grain scale, strain rate depends inversely on the cube of grain size. Situations may thus arise where grain boundary diffusion takes over as the dominant deformation mechanism once initial grain size has been reduced below some critical value, perhaps by cataclastic processes or by dynamic recrystallization. Field studies of regional (low strain rate) metamorphic terrains suggest that pressure solution is an important flow mechanism for fine-grained quartz and calcite-bearing rocks over the temperature interval 150–450°C (McClay 1977, Kerrich et al 1977). However, the role of these processes in comparatively fast strain rate shear zones is less clear. Textures characteristic of pressure solution have been recognized in the fine groundmass of some cataclastic shear zones (Brock & Engelder 1977, Mitra 1978, 1984, Rutter 1983, Blenkinsop & Rutter 1986) and also in some mylonites (White & White 1983).

Superplastic flow It has been postulated that intense grain refinement accompanying dynamic recrystallization in mylonitic shear zones may allow a transition to low-stress superplastic flow (Boullier & Gueguen 1975). In this mechanism, large strains develop dominantly by nonfrictional grain boundary sliding (involving diffusive mass transfer and/or dislocation processes) in a constant microstructure of fine equiaxed grains. Grain switching and rotation occur readily; superplastic tectonites should not therefore possess a strong crystallographic preferred orientation. A superplastic flow regime has been attained experimentally for Solnhofen limestone (Schmid et al 1977), and on textural grounds the mechanism has been identified in the fine-grained (<10 μm) Lochseiten calc-mylonite along the Glarus overthrust (Schmid et al 1981). Kerrich et al (1980) have inferred superplastic behavior within a quartzo-feldspathic ductile shear zone. The possibility of low-temperature superplastic flow has also been

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raised for ultrafine-grained cataclastic shear zones in quartzo-feldspathic host rocks (Phillips 1982, Anderson et al 1983). However, in general the extent to which quartz may develop superplastic behavior, even under mylonitic conditions, remains controversial (Etheridge & Wilkie 1979, White et al 1980).

Preseismic (β-Phase) Deformation

Despite its potential importance in accounting for observed precursory behavior to large earthquakes, little has been done to distinguish permanent deformation around ancient fault zones that might be related to the preseismic phase of the earthquake stress cycle. Possibilities include high-stress microcrack dilatancy, intensified through repetitive stress cycling (Scholz & Kranz 1974); cyclic dilatancy involving the reactivation of old or the development of new macrofracture systems, perhaps by subcritical growth of extension cracks (Crampin et al 1984); and possibly some form of sand-grain dilatancy within the fault zone itself (Nur 1975). Arrays of macroscopic extension veins in the vicinity of faults with textures recording incremental development in a shared stress regime have been interpreted as resulting from cyclical preseismic hydrofracture dilatancy (Sibson 1981). This form of dilatancy is a low-stress phenomenon that can only develop at shallow depths or under conditions of abnormally high fluid pressure, but it seems plausible that much of the repetitive crack-seal deformation described by Ramsay (1980) may also be caused by similar fault-related stress cycling. Mawer & Williams (1985) have attributed periodic microfracturing and crack healing in crystalline rocks to seismic activity.

Coseismic (γ-Phase) Deformation

Through consideration of likely power dissipation effects, it appears that the character of deformation accompanying seismic slip is critically dependent on the thickness of the slip zone (Sibson 1980a). Field evidence from deeply exhumed fault zones (Flinn 1977, Aydin & Johnson 1978, Davis et al 1980, Segall & Pollard 1983, Grocott 1981) suggests the existence, at least locally, of concentrated slip zones commonly centimeters or less in thickness (and even discrete sliding surfaces) throughout the seismogenic zone to depths of 10–15 km. Often, these principal slip surfaces (PSS) are located at the margins of broader cataclastic shear zones.

Given the high strain rates and levels of power dissipation anticipated for seismic slip, one may expect related deformation to be primarily cataclastic, though minor crystal plastic strains may develop by dislocation glide and through twinning and kinking of appropriate minerals. Within the broader slip zones, frictional wear processes may include attrition brecciation of the wall rock and cataclastic grain comminution. Transitory fluid pressure

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differentials caused by seismic slip transfer across dilational fault jogs may also induce brecciation of wall rock by hydraulic implosion (Sibson 1985). High-dilation breccias resulting from this process generally possess a matrix of hydrothermal minerals with textures recording multiple episodes of brecciation and cementation.

Depending on the degree of slip localization, significant increases in temperature can be expected to develop locally as a consequence of the rapid energy dissipation (Cardwell et al 1978). Overpressuring of fluid inclusions from sudden temperature rises may induce crystal fragmentation, with thermally induced stresses also contributing to fracturing adjacent to slip surfaces (Moore & Sibson 1978). Provided that seismic slip is localized to within a few millimeters, sufficient heat may be generated under *dry* conditions for friction-melting to occur (McKenzie & Brune 1972, Sibson 1975, Allen 1979). This is borne out by the restricted occurrence of pseudotachylyte friction-melt to minor faults in crystalline rock (Sibson et al 1979, Grocott 1981, Maddock 1983, Macaudiere et al 1985). In mature fluid-saturated fault zones, high transient fluid pressures are likely to develop on faults as a consequence of initial power dissipation at the onset of slip, leading to dramatic reductions in shear resistance and preventing large temperature increases (Sibson 1980a, Lachenbruch 1980, Raleigh & Evernden 1981). Possible field evidence for these high transient fluid pressures is sometimes found in the form of gouge-laden clastic dikes leading off slip surfaces (Gretener 1977).

These alternative feedback processes for lowering shear resistance during seismic slip (friction melting and the development of high transient fluid pressures) should both lead to near-total stress relief and high seismic efficiency, provided that the fluids remain contained within the slip zones. Acoustic fluidization has been suggested as a further possible mechanism for drastically lowering shear resistance during earthquake slip (Melosh 1979).

Postseismic (δ-Phase) Deformation

As previously discussed, distributed aftershock activity must lead to substantial deformation, much of which may be well away from the mainshock rupture (Figure 3). However, significant postseismic deformation may also develop within the primary slip zone as a consequence of afterslip. Following some large earthquakes, afterslip has increased surface displacement significantly over periods of weeks or months (e.g. Bucknam et al 1978). Such behavior can be expected to impose substantial strains and fabrics on the rock products of the preceding coseismic slip phase. A possible example of this behavior could be the local imposition of strong

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fabrics within some pseudotachylyte fault veins (Sibson 1980b, Passchier 1982).

In the postseismic phase, displacements may also extend with time over a much greater area than the coseismic rupture. Thatcher (1975) has shown from geodetic data that in the decades following the great 1906 San Francisco earthquake, displacements matching the coseismic strike-slip of ~4 m in the top 10 km of the crust extended downward to depths of perhaps 30 km. Short-term aseismic deformation at unusually high but decaying strain rates may thus be induced in the surrounds of the mainshock rupture.

Interseismic Healing

It seems probable that earthquake rupture surfaces undergo some form of healing and strengthening between successive major events. Apart from the time-dependent frictional "healing" observed in laboratory experiments (Dieterich 1978), processes of recrystallization, pressure solution, and hydrothermal cementation from circulating fluids may contribute to this effect (Angevine et al 1982). The concept of interseismic healing accords with observations of a correlation between recurrence interval and magnitude of stress drop (Kanamori & Allen 1986).

CONTINENTAL FAULT ZONE MODEL

A general model for a mature active fault zone in quartzo-feldspathic crust has been developed on the basis of the deformation textures and the distribution of fault rocks in ancient fault zones (Sibson 1977, 1983, Anderson et al 1983), and has been used to account for the depth distribution of earthquakes in continental crust. In this model, a seismogenic frictional regime dominated by pressure-sensitive deformation involving cataclasis and frictional sliding gives way with increasing depth and temperature to a quasi-plastic regime where largely aseismic and continuous shearing is localized within mylonite belts (Figure 4). The passage from frictional to quasi-plastic behavior is believed to be determined most commonly by the changing response of quartz to deviatoric stress with increasing temperature; dislocation creep becomes an important quartz deformation mechanism at $T > \sim 300°C$, corresponding to the onset of a greenschist facies metamorphic environment. Complex transitional behavior incorporating mixed continuous and discontinuous deformation over a large range of strain rates is inferred in the vicinity of the frictional/quasi-plastic transition.

These fault models have been crudely quantified using available

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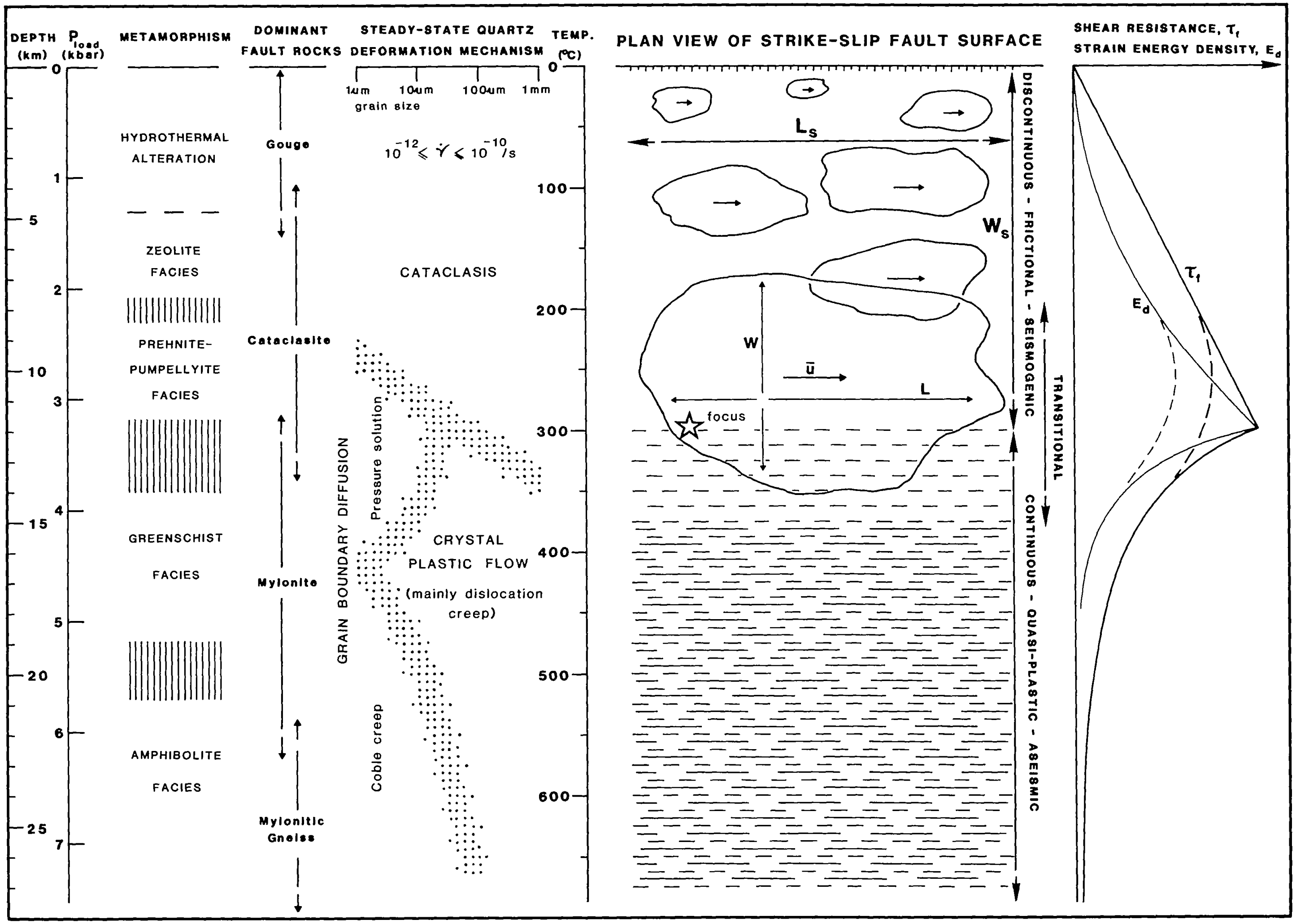

Figure 4 Conceptual model for a major strike-slip fault zone in continental crust (uniform density of 2.80 g cm^{-3}, and geothermal gradient of 25°C km^{-1}), schematically relating different fault regimes to likely metamorphic environment, dominant quartz steady-state deformation mechanism, and associated fault rocks (from Sibson 1983). Rupture parameters (length L, width W, mean slip $\bar{u}$) shown for events within a seismogenic regime of total dimensions $L_s \times W_s$. Profiles of shear resistance and distortional strain energy density are in arbitrary units.

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laboratory data on the frictional and rheological properties of crustal rocks to construct composite profiles illustrating the variation of shear resistance with depth for different faulting modes and heat flow provinces (Meissner & Strehlau 1982, Sibson 1982, 1983, 1984, Smith & Bruhn 1984). The peak shear resistance occurs at the frictional/quasi-plastic transition, though in reality its value is likely to be considerably reduced and smoothed out below the theoretical sharp intercept (Figures 4, 6). Observed depth distributions of microearthquakes defining the seismogenic zone can be correlated fairly satisfactorily with the modeled depths of frictional interaction in different heat flow provinces (Figure 5). Larger ($M_L > 5.5$) earthquakes tend to nucleate toward the base of this seismogenic zone in the region of inferred peak shear resistance, rupturing upward and laterally

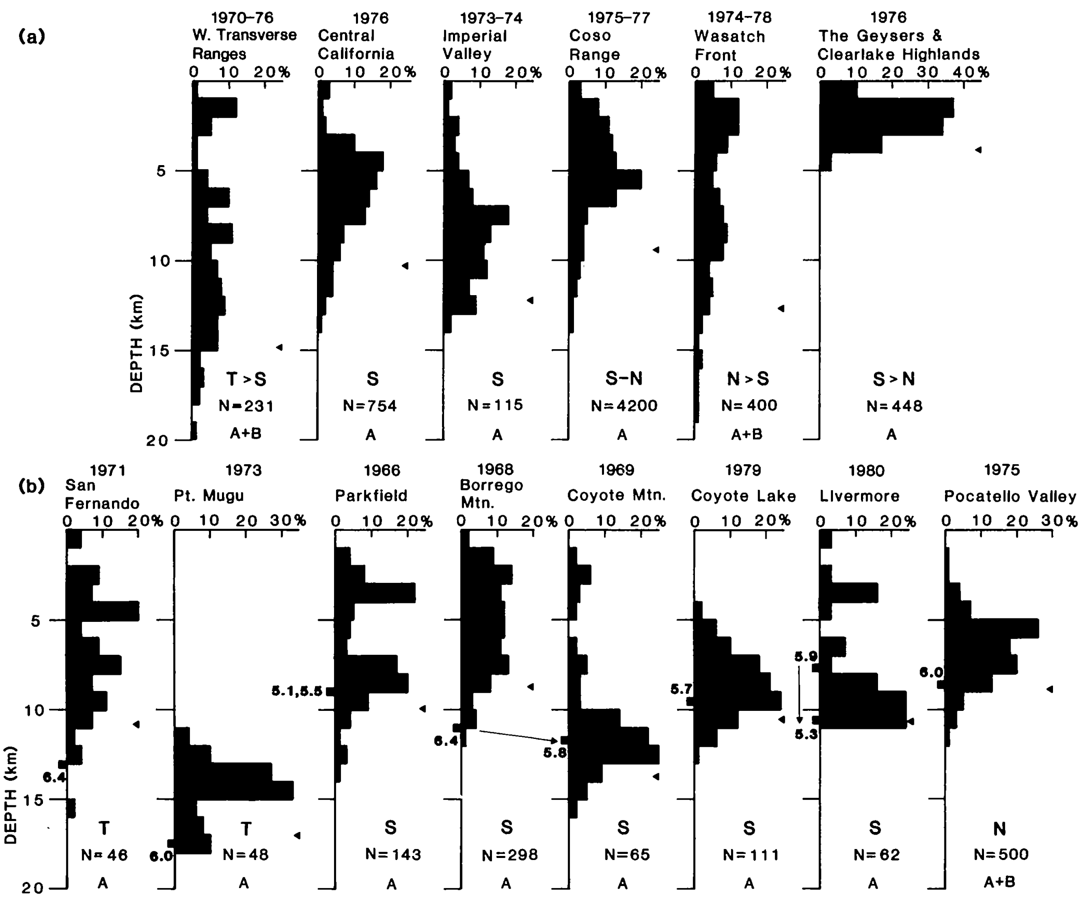

Figure 5 Earthquake depth distributions in the western United States (from Sibson 1984). Solid triangles indicate depths above which 90% of the activity occurs; the dominant faulting mode (T, thrust; S, strike-slip; N, normal), the sample size, and the data quality are listed at the base of each histogram: (*a*) background microearthquake activity (note that the Geysers–Clearlake Highlands region is an area of intense geothermal activity); (*b*) aftershock sequences following moderate earthquakes. Depth and local magnitude (M_L) of mainshock or mainshocks are indicated by solid squares. Related larger events are linked by arrows.

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so that their aftershock sequences are also largely restricted to the same zone (Sibson 1982, Chen & Molnar 1983). Variations in conductive heat flow and crustal composition (especially the quartz-feldspar ratio) appear the most important of a range of factors affecting the depth of frictional interaction and seismic activity (Figure 6; Sibson 1984).

Given the heterogeneous nature of continental crust and the variety of deformation mechanisms involved in faulting, it is clear that these simple fault models can only represent bulk fault behavior to a first approximation. Many uncertainties remain in our understanding of the fine structure of fault zones and the changing physical conditions and deformation processes operative at different crustal levels. To illustrate these problems, we consider each of the main fault regimes in turn.

Frictional Regime

BRITTLE INFRASTRUCTURE Detailed maps of the surface traces produced by individual and repeated earthquake ruptures (Vedder & Wallace 1970, Clark 1972, Tchalenko & Berberian 1975) and of the brittle infrastructure of exhumed fault zones (Flinn 1977, Segall & Pollard 1983), when coupled with high-precision aftershock studies (Reasenberg & Ellsworth 1982), suggest that throughout the seismogenic regime much of the displacement within mature fault zones is localized on principal slip surfaces (PSS), which are commonly discontinuous or curved. Segmentation may be systematically en echelon, or the PSS may step abruptly from one margin to another of major fault zones that range up to a kilometer or so in width. Geomorphic evidence indicates that the broad infrastructure of PSS may persist in a particular configuration through many episodes of rupturing over lengthy time periods, in some instances exceeding 10^4 yr. Detailed seismological analyses of rupture propagation, precision microearthquake studies, and theoretical consideration of the effects of fault segmentation (Bakun et al 1980, Segall & Pollard 1980, King & Nabelek 1985, Sibson 1985) suggest that the broad PSS infrastructure exerts major controls on rupture nucleation and arrest and on patterns of aftershock activity (see Figure 3).

FAULT ROCKS AND DEFORMATION The dominant rock products of faulting in the frictional regime are incohesive gouge, breccias, and cohesive fault rocks of the cataclasite series (Sibson 1977). Cataclastic deformation may be localized to thin tabular zones associated with PSS or may be distributed through large volumes. In general, there is abundant evidence of hydration in mature fault zones; it seems probable that most seismogenic faulting takes place in the presence of an aqueous fluid that is at least at hydrostatic pressure. Arrays of hydrothermal extension veins in the vicinity of some

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faults indicate the local presence of suprahydrostatic fluid pressures exceeding the least principal compressive stress (Sibson 1981).

One of the main problems in interpreting fault rocks within the seismogenic regime arises from the lack of good depth control on most samples. A major controversy exists concerning the relative dominance of cataclasite series material or clay gouge, rich in montmorillonite. This has important implications for the strength of fault zones. Microbreccias and cataclasites may crudely be expected to follow Byerlee's (1978) general rock friction relationship, with a static shear resistance approximated by

$$\tau_f = \mu_s \sigma'_n = \mu_s(\sigma_n - P), \quad (2)$$

where σ_n is the normal stress across the sliding surface, P is the fluid pressure, and $\mu_s \sim 0.75$ is the static friction coefficient (Sibson 1983). In contrast, montmorillonite-rich gouge may have an effective friction coefficient ≤ 0.35 and a complex mechanical response under shear (Bombolakis et al 1978, Bird 1984). Moreover, the low permeability of clay gouges may contribute to the development and maintenance of suprahydrostatic fluid pressures in fault zones (Morrow et al 1984). It can be argued on thermodynamic and other grounds that given an appropriate geochemical environment, the montmorillonite-rich gouge frequently found at high levels in fault zones remains the dominant intrafault material to depths in excess of 10 km (Wu 1978, Bird 1984, Wang 1984). However, studies of ancient exhumed fault zones generally reveal a preponderance of microbreccias and cataclasites (Brock & Engelder 1977, Flinn 1977, Davis et al 1980, Anderson et al 1980, 1983), with the implication that clay-rich gouges are largely restricted to the top few kilometers. This view is supported by studies of clastic shales undergoing diagenesis during progressive burial; these studies demonstrate that the smectite-illite transition generally begins below temperatures of $\sim 90°C$ and is largely complete by 150°C (Ramseyer & Boles 1986). It is tempting to speculate that the comparatively low level of microseismic activity generally observed in the top few kilometers of active fault zones (e.g. Bakun et al 1980, Reasenberg & Ellsworth 1982) may correlate with the depth extent of weak montmorillonite-rich gouge.

For most of the rock products of faulting within the frictional regime, there are major problems in distinguishing deformation that results from seismic slip (γ-phase) from that induced during other phases of the earthquake stress cycle or during stable aseismic sliding and cataclastic flow. For example, while the development of shape fabrics in some cataclasites (House & Gray 1982, Chester et al 1985) seems likely to be a consequence of slow aseismic flow, it is unclear whether the flow was stable or was instead induced by δ-phase afterslip. High-dilation, hydrothermally recemented breccias associated with dilational fault jogs are inferred to

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have developed by hydraulic implosion induced by seismic slip (Sibson 1985). Gouge-filled clastic dikes sometimes observed leading off sliding surfaces (Gretener 1977) may be diagnostic of transient high fluid pressures developed as a consequence of temperature increases during earthquake rupturing (Sibson 1980a). Anomalous vitrinite reflectances recognized on some thrust faults may likewise result from seismic power dissipation (Bustin 1983). Pseudotachylyte friction-melt, the product of localized seismic slip under dry conditions, can be recognized only rarely in crystalline rocks that have preserved their integrity. Fibrous slickensides diagnostic of aseismic pressure solution slip (Elliot 1976) are often associated with extension vein arrays characterizing areas of anomalously high fluid pressure. Microstructural evidence for pressure solution has likewise been recognized in the groundmass of a range of gouge and cataclasite material (Brock & Engelder 1977, Mitra 1978, 1984, Rutter 1983, Blenkinsop & Rutter 1986).

Thus, given the heterogeneity of continental crust, the picture that emerges of fault deformation within the frictional regime is of extreme variability. In mature fault zones, fluid pressures may be expected to range upward from hydrostatic values; portions of faults probably obey the Byerlee friction relationship, but these may give way both laterally and vertically to clay-rich patches of lower strength or to sliding surfaces and zones where pressure solution mechanisms are operative.

Quasi-Plastic Regime

Deformation within this regime is largely aseismic and continuous. Strain rates range upward from $\sim 10^{-14}$ s^{-1} to the higher values of 10^{-12}–10^{-10} s^{-1} inferred for mylonitic shear zones. However, significant strain rate fluctuations, dying out with depth, must still occur as a consequence of the intermittent stress relief accompanying earthquake rupturing in the frictional regime. Shearing is mostly localized in mylonite belts, with widths that in the mid-crust typically range from tens to hundreds of meters but in the deep crust may broaden to as much as 10 km (Jegouzo 1980, Sorensen 1983). Deformation is heterogeneous over a wide range of scales; typically, an anastomosing mesh of high-strain shear belts encloses lozenges of comparatively undeformed material (Mitra 1979, Simpson 1983, Sorensen 1983). Characteristic rock products are well-ordered L-S tectonites of the mylonite series and, in the lower crust, mylonitic gneisses (Sibson 1977). High-shear strain gradients may give rise to considerable mesoscopic structural complexity (Bell & Hammond 1984).

Mylonitization of quartzo-feldspathic rocks under mid-crustal greenschist facies conditions at $T > \sim 300°C$ involves flow of quartz by dislocation creep around resistant feldspar grains; this gives rise to the

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characteristic fluxion texture of mylonites. This contrasting mechanical response becomes less marked as dynamic recrystallization of feldspars and other minerals sets in under amphibolite facies conditions in the deep crust (Voll 1976, Tullis et al 1982, Hanmer 1982, Simpson 1985). Microstructural characteristics of mylonitic rocks are reviewed by Simpson & Schmid (1983) and Lister & Snoke (1984). Although there is no question that dislocation creep plays a major role in mylonitic shear zones, especially during their initial development, the possibility of a transition to mechanisms involving grain boundary diffusion, and perhaps even to superplastic flow, has to be considered once grain size has been sufficiently reduced by dynamic recrystallization (Boullier & Gueguen 1975). If such a transition occurs and, more particularly, if it involves shearing flow under near-zero effective confining pressure, as Etheridge et al (1984) surmise, the shear resistance will drop well below values inferred from dislocation creep flow laws.

Transitional Regime

Deformation processes associated with the frictional/quasi-plastic transition are of particular interest because of the tendency for larger earthquake ruptures to nucleate in the inferred transition region, where shear resistance should be at a maximum (Sibson 1982). Problems of strain compatability in the vicinity of the transition, arising during time-dependent loading of the frictional regime from below by aseismic quasi-plastic shearing, are likely to play a key role in the nucleation process. Moreover, it has been suggested that behavior in the transitional regime may control whether or not moderate earthquakes develop into very large ruptures with lengths greatly exceeding their width (Das 1982). However, as pointed out by Carter & Kirby (1978), deformation within the transition region is likely to be highly varied, involving mixed continuous and discontinuous behavior over an enormous range of slip and strain rates. For a vertical extent of perhaps several kilometers, aseismic flow at varying strain rates may involve a mixture of cataclasis, pressure solution, and crystal plastic processes, intermittently punctuated by the sudden energy dissipation accompanying brittle earthquake rupturing. Much experimental work is now being directed at understanding this complex semibrittle behavior (Shimamoto & Logan 1986). Possible examples of the mixed deformation styles derived from this regime have been described by Sibson (1980b), Passchier (1982), Mitra (1978, 1984), and White & White (1983).

Shear Resistance Profiles

Composite profiles illustrating the variation of fault shear resistance with depth (e.g. Figure 6) are constructed on a number of simplifying assump-

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tions. In the frictional regime, it is usually assumed that frictional resistance is independent of rock type, obeying Byerlee's (1978) simple friction laws with $\mu_s \sim 0.75$, and that fluid pressures are hydrostatic. Within the quasi-plastic regime, flow shear resistance is estimated for an appropriate conductive geotherm and a constant strain rate using a laboratory-determined flow law for an assumed representative rock type (Meissner & Strehlau 1982, Sibson 1982, 1984). More complex multilayer rheologies may also be modeled (Smith & Bruhn 1984). The power laws employed depend largely on the flow of a dominant constituent such as quartz by dislocation creep. Choice of the quasi-plastic strain rate depends on the assumption of a uniformly deforming lower crust ($\sim 10^{-14}$ s^{-1}) or on the more reasonable inference from geologic evidence that

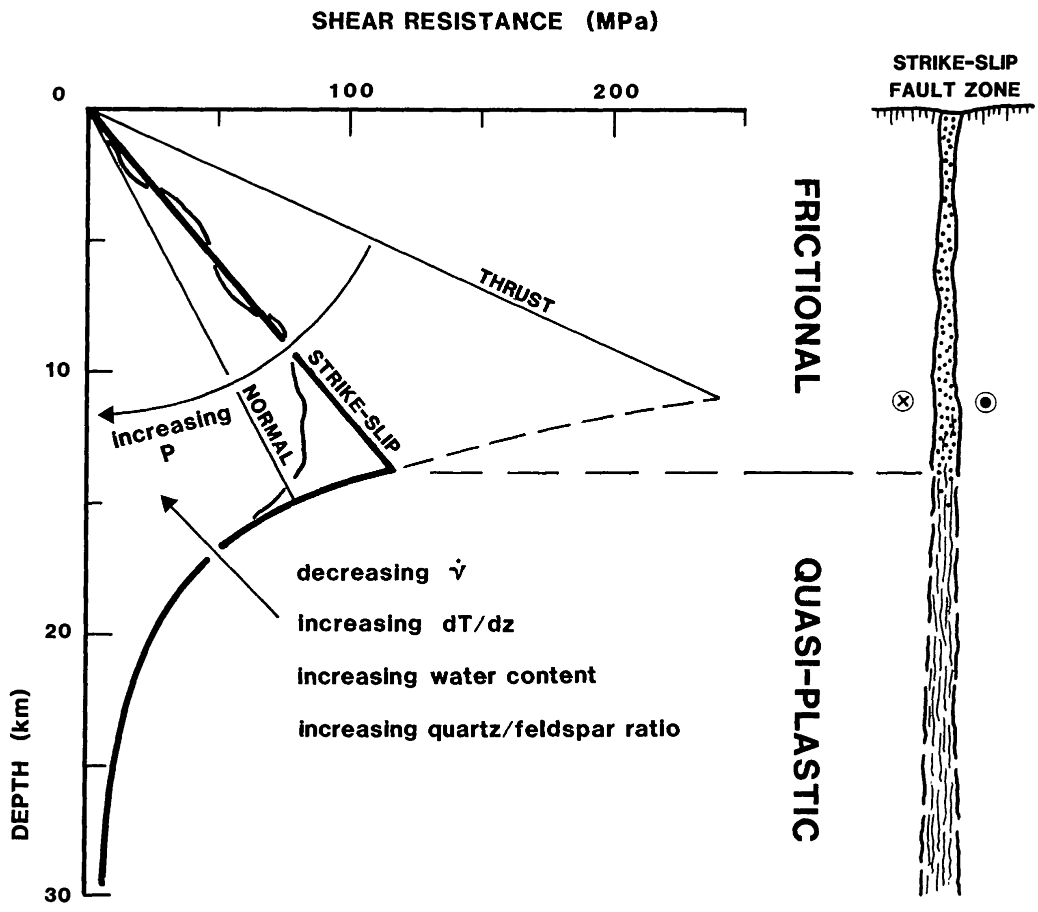

Figure 6 Composite shear resistance profile for a simple fault model (constructed for $\mu_s = 0.75$, hydrostatic fluid pressures P, a geothermal gradient $dT/dz = 25°\mathrm{C\ km^{-1}}$, and a Westerly Granite flow law at a strain rate of 10^{-11} s^{-1}). Expected subsidiary roughness in the frictional regime and the smoothing out of the peak shear resistance are illustrated schematically for the strike-slip profile. Effects of perturbing factors on the general profiles are indicated (after Sibson 1984).

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quasi-plastic shearing is concentrated in narrow mylonitic shear belts ($\sim 10^{-12}$–10^{-10} s^{-1}).

It is instructive to consider the implications of the profiles with regard to the controversy over shear stress levels in crustal fault zones (see Hanks & Raleigh 1980). The strike-slip profile given in Figure 6 yields a peak shear resistance of ~120 MPa at the frictional/quasi-plastic transition and an average shear resistance of ~60 MPa over a 14-km-deep seismogenic zone. Within the quasi-plastic regime, values of flow shear resistance are broadly consistent with the shear stresses of 10–100 MPa derived from paleopiezometric studies of mylonites (Kohlstedt & Weathers 1980, Ord & Christie 1984). However, the average frictional shear stress is three times the 20-MPa maximum inferred from heat flow measurements around the San Andreas fault (Lachenbruch & Sass 1980).

If we take account of the simplifying input assumptions, there are, in fact, good grounds for believing that the modeled profiles generally represent upper bounds to possible stress values in quartz-bearing crust. Shear resistance in the frictional regime is likely to be lowered by concentrations of impermeable montmorillonite-rich gouge, by suprahydrostatic fluid pressures, and (especially in the vicinity of the frictional/quasi-plastic transition) by pressure solution processes. Within the quasi-plastic regime, the effect of shear zone broadening with depth will be to decrease strain rates and the flow shear resistance. Note, though, that diminishing quartz content and a change to a feldspar-dominated rheology would increase the flow resistance. A transition to diffusional mechanisms and superplastic flow consequent on grain size reduction would lead to linear viscous behavior and extreme softening (White et al 1980). This weakening would be carried to extremes if, as postulated by Etheridge et al (1984), quasi-plastic mylonite belts generally act as conduits for high-pressure aqueous fluids, so that shearing is occurring under near-zero effective confining pressure.

DISCUSSION: OUTSTANDING PROBLEMS

The conceptual models developed to date account for a wide variety of observed fault behavior and earthquake source characteristics, but they can only be regarded as preliminary. Refinement of existing models will come through the combination of new field, theoretical, and experimental studies. The following interrelated topics appear to be particularly fruitful avenues for future research.

Fault zone geometry Further investigations of the geometrical characteristics of fault systems at all crustal levels and on all scales are needed.

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Particular attention should be paid to large-scale brittle infrastructure and to the continuation of fault zones beneath the seismogenic regime for different faulting modes; both may play critical roles in determining the character of earthquake rupturing.

Evolution of fault zones Our understanding of fault maturation is still limited, and a number of questions remain unanswered. What factors control how a fault zone widens with time and with increasing displacement? What degree of strain softening is involved at different crustal levels, and how does it occur? How does a residual brittle infrastructure develop, and what controls its longevity in one particular configuration? What processes of interseismic healing occur within the frictional regime?

Pressure-temperature controls Much more and improved quantitative information is needed on the *P-T* conditions and, in particular, the depths of formation of the different fault rock types, especially within the seismogenic regime.

Physical conditions and stress levels The level of tectonic shear stress remains the outstanding problem in crustal fault mechanics. Fault rock studies on power dissipation in relation to seismic slip bear directly on this important question. It is intimately related to other poorly constrained physical controls on deformation, such as fluid pressure levels, and to whether heat flow in the vicinity of major faults is dominantly a conductive or a convective process. The relative dominance of impermeable clay gouge or cataclasite within the frictional regime and the effects of pressure solution and hydrothermal self-sealing on fault permeability all have implications for the development of suprahydrostatic fluid pressures requiring further assessment. Systematic evaluation of the commonness of hydrothermal extension veining, diagnostic of near-lithostatic fluid pressures, is needed within both the frictional and quasi-plastic regimes of ancient fault zones, particularly in view of the hypothesis that mylonitic deformation generally occurs in the presence of a high-pressure fluid phase.

Rheological modeling As new laboratory-determined constitutive flow laws become available, it should be possible to develop more elaborate rheological models for fault zones that reflect continental heterogeneity. Prime requirements are a better appreciation of the clay gouge to cataclasite ratio throughout the frictional regime, its controls and mechanical effects, quantitative understanding of the role of pressure solution processes at all crustal levels in realistically "dirty" fault zone environments, and improved knowledge of the factors affecting hydrolytic weakening of silicates. Detailed modeling of the time-dependent loading of the frictional regime to failure and of the likely effects throughout the quasi-plastic

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regime of sudden stress drops in the seismogenic zone would be instructive. A rheological model for fault zones in oceanic crust should also be developed.

Rupture controls Research should be directed toward improving our nascent understanding of the role of brittle infrastructure in controlling earthquake rupture nucleation and arrest within the frictional regime. An improved ability to distinguish the rock products of steady aseismic shearing from those associated with different phases of the earthquake stress cycle is desirable in this regard. The complex mixture of continuous and discontinuous deformation inferred to be associated with the nucleation of larger ruptures in the frictional/quasi-plastic transitional regime warrants special attention.

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