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Seismic Fault Rheology and Earthquake Dynamics

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ABSTRACT

As preparation for this Dahlem Workshop on *The Dynamics of Fault Zones*, specifically on the subtopic "Rheology of Fault Rocks and Their Surroundings," we addressed critical research issues for understanding the seismic response of fault zones in terms of the constitutive response of fault materials. This requires new concepts and a host of new observations and experiments to document material response, to understand the shear localization process and the inception of earthquake instability, and especially to understand the mechanisms of fault weakening and dynamics of rupture tip propagation and arrest during rapid, possibly large, slip in natural events. We examine in turn the geological structure of fault zones and its relation to earthquake dynamics, the description of rate and state friction at slow rates appropriate to the interseismic period and earthquake nucleation, and the dynamics of fault weakening during rapid slip. The last topic gets special attention in view of the important recent advances in theoretical concepts and experiments to probe the range of slip rates prevailing during earthquakes. We then address the assembly of the constitutive framework into viable, but necessarily simplified, conceptual and computational models for description of the dynamics of crustal earthquake rupture. This is done principally in the slip-weakening framework, and we examine some of the uncertainties in doing so, and issues of how new understanding of the rapid large slip range will be integrated to model the traction evolution and the weakening process during large slip episodes.

INTRODUCTION

The mechanics of earthquakes and faulting is currently investigated by collecting observations and performing theoretical analyses and laboratory experiments on three main physical processes: tectonic loading, fault interaction through stress transfer, and rheological response of fault zones. Earthquakes are certainly one of the most important manifestations of faulting, and the understanding of dynamic fault weakening during the nucleation and propagation of a seismic rupture is a major task for seismologists and other Earth scientists. In the literature, the study of the initiation, propagation, and arrest of an earthquake rupture is associated with the understanding of coseismic processes. However, the three main physical processes cited above also control fault behavior during the interseismic period (i.e., the time interval between two subsequent earthquakes) and the postseismic period (i.e., the time interval immediately following a seismic event). In this chapter we focus primarily on the most recent advances and progresses of research in the understanding of rheological and constitutive properties of active fault zones. Fault rheology and constitutive properties help to understand the depth extent of the seismogenic zone. They are associated with the characterization of thickness of the seismogenic zone (see Sibson 2003, and references therein), the material properties of fault gouge, and the properties of contact surfaces within the slipping zone (Dieterich 1979; Ohnaka 2003). The latter are our primary focus here.

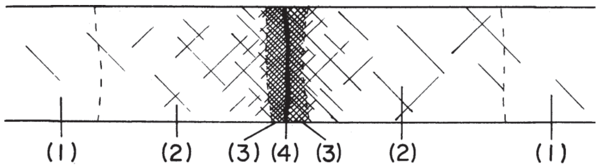
We present geological observations of fault zones to constrain a model of a seismogenic structure capable of localizing strain and generating earthquakes. We review results of laboratory experiments aimed at understanding fault-constitutive properties. In the context of this study, earthquakes are considered as instabilities of a complex dynamic system governed by assigned frictional laws and other constitutive laws (e.g., for the damage regions bordering the zone of concentrated slip). Therefore, we briefly present the physical origin and analytical expressions of these friction laws and discuss the different competing physical mechanisms that contribute to dynamic fault weakening during earthquakes. In particular, there is now the awareness that, although the properties of the contact surface play a relevant role in controlling dynamic slip episodes, frictional heat, thermal pressurization of pore fluids, and mechanical lubrication can contribute to explain dynamic fault weakening and to control fault friction at high slip rates (Sibson 1973; Lachenbruch 1980; Mase and Smith 1987; Tsutsumi and Shimamoto 1997; Kanamori and Brodsky 2001; Andrews 2002; Goldsby and Tullis 2002, 2003; Tullis and Goldsby 2003; Di Toro et al. 2004; Fialko 2004; Hirose and Shimamoto 2005; Rice 2004, 2006). The definition of a fault zone model (i.e., to determine the size of the slipping zone with respect to the surroundings) and its characterization in terms of the dominant physical processes are extremely important tasks which will be a focus for future scientific research.

To describe a fault zone, we must address two main problems: (a) the definition of the geometrical and mechanical properties of a fault zone (see reviews by Ben-Zion and Sammis 2003; Biegel and Sammis 2004) and (b) the understanding of the spatial and temporal scale dependence of relevant physical processes. Only by clarifying these issues can we integrate the sometimes conflicting evidence on fault zone properties. The effect of fault geometry is important because it concerns the three-dimensional structure of a fault zone (i.e., thickness of the slipping region, of the fault core, or the damage zone), the fragmentation of the rupture surface (i.e., bending, branching, step over, etc.), as well as the complexity of the fracture network (i.e., fractal distribution of fractures). Scale dependence is extremely important because it allows us to establish a hierarchy to characterize where (i.e., the spatial extension) and when (i.e., the temporal evolution) the different physical processes govern crustal faulting. The contribution of each physical process controlling dynamic fault weakening and fault evolution depends on the size of the slipping zone as well as on the values of the main physical parameters (such as thermal and hydraulic diffusivity, permeability, and porosity) that appear within the mathematical representation of fault zones. In this chapter, we discuss these topics in an effort to describe a fault zone model, to discuss the nature and the dominant physical processes of fault zones, and to determine their scale of relevance. Our primary task is to stimulate a discussion and an exchange of views and ideas to clarify perspectives and new horizons in fault mechanics.

GEOLOGICAL OBSERVATIONS OF FAULT ZONES AND THEIR RELATION TO RUPTURE DYNAMICS

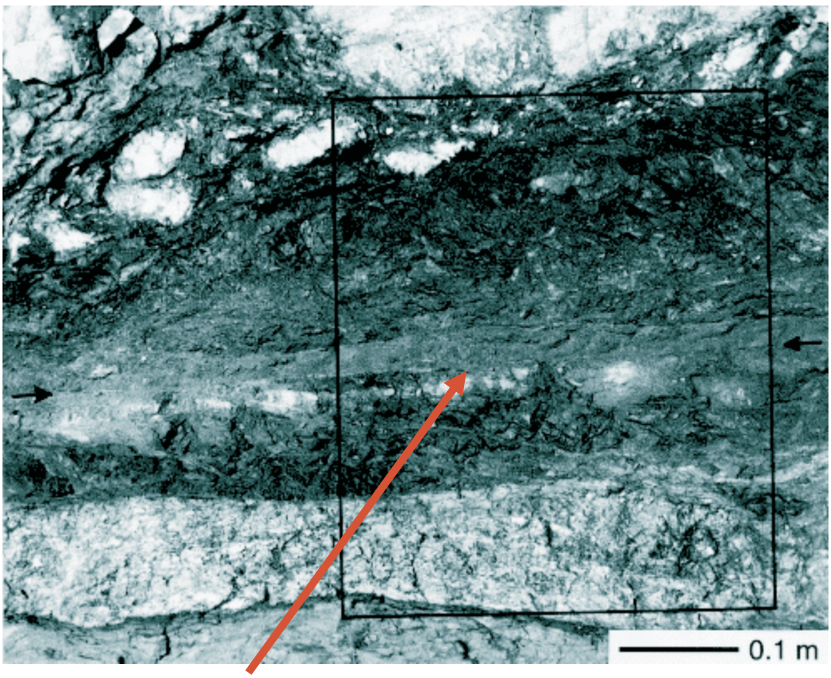
Many recent investigations have focused on the internal structure of fault zones to improve knowledge of microscale processes, fault zone rheology, and dynamic weakening processes. These studies pointed out that coseismic slips on mature, highly slipped fault zones often occur within ultracataclastic, possibly clayey, zones of order tens to hundreds of millimeters thick, but that the zone of principal seismic shearing may be localized to a thickness less than 1–5 mm width within that ultracataclasite core. A broad damage zone, on the order of one to hundreds of meters thick, surrounds the fault core and it is characterized by highly fractured and possibly granulated materials which, because of their porosity, must usually be assumed to be fluid-saturated. Figure 5.1a displays a sketch of the fault zone model discussed here. Evidence for such a model has been collected from the Punchbowl fault (southern California; see Chester and Chester 1998, and Figure 5.1b), the nearby North Branch San Gabriel fault (Chester et al. 1993), the Median Tectonic Line (Japan; see Wibberley and Shimamoto 2003), the Nojima fault (which ruptured in the 1995 Kobe earthquake in Japan and has been penetrated by drill holes; Lockner et al. 2000), as

(a)



(1) Undamaged host rock
(2) Damage zone, highly cracked; 10s m to 100 m wide, minor faults may reach 1 km
(3) Gouge or foliated gouge; 1 m to 10s m wide
(4) Central ultracataclasite shear zone, may be clay rich; 10s mm to 100s mm wide
(5) [within (4), not marked above] Prominent slip surface; may be < 1 to 5 mm wide

(b)



Prominent slip surface (pss) is located in the center of the layer (black arrows)

Figure 5.1 Internal structure of a major fault zone: (a) schematic representation of the inner structure of a fault zone within the brittle upper crust based on Chester et al. (1993); (b) picture of the ultracataclastic layer (with prominent slip surface indicated by the arrows) at the Punchbowl fault, taken from Chester and Chester (1998); see also Sibson (2003).

well as from other observations summarized in Sibson (2003), Ben-Zion and Sammis (2003), and Biegel and Sammis (2004). According to these observations, slip is generally accommodated along a single, nearly planar surface.

The damage zone is characterized by a higher fracture density than the surroundings. At a much larger scale of a plate boundary, the fault zone model described above can be part of a broader shear zone, whose thickness can be on the order of several hundreds of meters, which is composed by many different fracture zones and is responsible for stress reorientation with respect to the direction of remote tectonic loading. However, this "large-scale" fault zone model may describe only mature faults such as San Andreas or plate margin transform faults. Although we point out that the spatial scale at which we attempt to characterize the fault zone structure does matter, here we focus our attention on the fault zone at a local scale such as that depicted in Figure 5.1a. It is important to point out that the width and the complexity of fault zones inferred from the analysis of surface ruptures depend on the faulting mechanism (i.e., whether reverse, normal, strike-slip, or oblique) and can be affected by the presence or absence of sedimentary cover as well as by other free surface effects.

The geological observations of fault zones presented above raise several important issues that must be addressed to understand the mechanical properties of faults as well as the dynamic weakening processes occurring during earthquakes. The first concerns *frictional heating* caused by *large slip* (≥ 1 m) during an earthquake within such an *extremely thin slipping zone*. The temperature increase caused by a meter of slip within a few millimeters-thick slipping zone would be larger than 1000°C over most of a seismogenic zone under the assumption of uniform adiabatic shearing, at least if it occurred (as recent evidence suggests it may not; see below) in the absence of any mechanism to reduce strength rapidly once slip begins. Such a sudden frictional-driven temperature change should lead to melting and formation of pseudotachylites. The understanding of this issue may contribute to solving the well-known heat flow paradox along the San Andreas fault; that is, to explain the absence of measurably enhanced heat outflow along the fault. Moreover, if melts produced during frictional heating have a low viscosity, they may lubricate faults and thus reduce dynamic friction (Sibson 1975; Spray 1993; Brodsky and Kanamori 2001; Kanamori and Brodsky 2001). Such rapid changes in friction at high slip rates, not all due to melting, have also been observed in laboratory experiments (Tsutsumi and Shimamoto 1997; Goldsby and Tullis 2003; Tullis and Goldsby 2003; Di Toro et al. 2004). However, according to Sibson (2003) the evidence of localized slip and the apparent scarcity of pseudotachylites suggest that pseudotachylite is rarely preserved within mature fault zones, or that melting is rare because other phenomena play a dominant role in controlling dynamic fault weakening. We discuss some of these phenomena and, in particular, we address frictional heating and presently known thermal weakening mechanisms (flash heating at micro-contacts, thermal pressurization of pore fluid) in more detail in the next section.

A second issue concerns the relatively *simple structure of the slipping zone*. If the slipping zone of real faults is *extremely thin*, comparable to the thickness of gouge layers in some laboratory experiments, we may ask what is the contribution of rupture surface topography (and contact properties) in controlling fault friction during sliding? The resolution of this issue is related to the interpretation of results of numerous laboratory experiments on fault friction (distributed vs. localized strain within the gouge layer; grain rolling/sliding vs. grain fracture), and it is also related to the previous issue concerning flash heating on a localized zone versus broadly distributed shear of gouge layers. In general, the principal slipping zone contains wear materials or gouge, which can be cohesive or incohesive. These products of faulting have formed by macroscopic fracturing, frictional wear, and cataclastic comminution, in combination with alterations by reactions with fluids and mineral deposition (Chester et al. 1993; Sibson 2003). Thus, it is important to understand the gouge layer evolution both during single earthquakes (coseismically) and during the interseismic periods. It is likely that repeated slip episodes continuously modify the grain size and the properties of gouge materials. Power and Tullis (1991) found that roughness of natural fault surfaces in the slip direction has considerably smaller amplitude than the roughness in the direction normal to the slip. Therefore, we may expect that cumulative slip tends to smooth the fault surface, an expectation that is well documented at larger scales (Stirling et al. 1996). This view of the gouge layer evolution is commonly associated with other precepts that gouge texture is fractal (Steacy and Sammis 1991) and that gouge surface energy yields a negligible contribution to the earthquake energy balance. However, Wilson et al. (2005) have recently investigated the gouge texture of two seismic faults having very different cumulative slip and extension and report that both faults display similar gouge characteristics and the grain-size distribution is not fractal. These authors also proposed that fracture surface energy is a nonnegligible part of the energy budget, a view questioned by Chester et al. (2005), and suggested that gouge evolution is not related to quasi-static cumulative slip, but rather formed by dynamic rock pulverization during the propagation of a single earthquake. Thus, we have two competing interpretations of the gouge and damage zone evolution: the former can be defined as a large-scale geological view in which both coseismic and interseismic processes are involved; the latter solely invokes coseismic processes during individual events.

A third issue related to the previous two concerns the *presence of fluids* within a fault zone and the way in which they control fault strength evolution during and after a dynamic slip episode. One of the most important parameters to model the evolution of shear zone fluid pressure is the *hydraulic diffusivity*, which depends on the permeability, fluid viscosity, and fluid-pore compressibility. Several investigations have attempted to constrain the values of hydraulic diffusivity within the fault zone as well as to estimate the value of permeability (see Lockner et al. 2000; Wibberley and Shimamoto 2003; Sulem et al. 2004, and references therein). These studies have shown that the gouge zones forming the fault core have a much lower permeability than that measured in the surrounding damage zone, which can be highly variable. Permeability within the fault core ($\sim 10^{-19}$ m^2) can be three orders of magnitude smaller that that in the damage zone ($\sim 10^{-16}$ m^2). In both regions, permeability is reduced as the effective normal stress is increased. For example, in the case of the ultracataclastic gouge core material containing the slip zone in the Median Tectonic Line, permeability is 10^{-19} m^2 at 10 MPa effective confining stress, 10^{-20} m^2 at 70 MPa, 4×10^{-21} m^2 at 120 MPa, and 3×10^{-21} m^2 at 180 MPa (Wibberley and Shimamoto 2003); assuming hydrostatic pore pressure, 126 MPa corresponds to the effective overburden stress at 7 km, a representative centroidal depth for the slip zone of crustal earthquakes. It is important to assess the hydrodynamic behavior of fault zone fluids during dynamic slip episodes, which requires a detailed examination of permeability and poroelastic properties of fault core as well as their variations with effective pressure. In fact, hydraulic diffusivity can change during a dynamic slip episode because of effective normal stress and temperature changes: an increase in temperature can decrease fluid viscosity, therefore increasing hydraulic diffusivity during sliding. Moreover, as discussed above, porosity can evolve not just during an earthquake but also during the interseismic period. The latter may allow a creep-compaction mechanism which isolates and pressurizes fluids, as proposed to explain weakness of mature faults (Sleep and Blanpied 1992).

A fourth issue with mature fault zones is that they present *variations in elastic and seismic properties* in the *direction perpendicular to the slip surface*. At large scale, a major fault may have brought distinct lithologies into contact with one another, for example, seafloor crust and accreted sediments, along a subduction fault. At the scale of the damage zone, the highly variable materials, as just discussed, will cause property variations. That brings a new ingredient into rupture dynamics, because spatially inhomogeneous slip (like during earthquake rupture) along a fault, which is not a plane of mirror symmetry, alters not only the shear stress but also the normal stress along the fault. The effects have been shown to allow extremely unstable behavior even along faults that have been idealized to have a constant friction coefficient f (Weertman 1980; Andrews and Ben-Zion 1997; Cochard and Rice 2000) and hence would be stable if in a configuration with mirror symmetry. This means that it is critical to understand fault zone rheological response when there is *rapid change in normal stress*. In fact, it has been established (Cochard and Rice 2000; Ranjith and Rice 2001) that deviations from the classical formulation of Coulomb friction, of a type seen in shock wave experiments which deliver an abrupt change in normal stress to a slipping surface (Prakash and Clifton 1992; Prakash 1998), allow models of rupture along dissimilar material interfaces to be well posed mathematically. The classical formulation is not well posed in that circumstance. Nevertheless, the critical feature seen in the shock wave experiments, namely, that an abrupt change in normal stress does not cause a corresponding abrupt change in shear strength, has obviously not been duplicated in much lower speed friction experiments (Linker and Dieterich 1992; Boettcher and Marone 2004) with less abrupt changes of normal stress. In reality, rupture of a fault separating dissimilar materials elicits both the effect of coupling slip to alteration of normal stress discussed above and the effect of weakening of friction with slip or slip rate that would exist for a fault between identical materials. Both effects seem important to a complete computational and experimental description of rupture along dissimilar material interfaces (Harris and Day 1997; Xia et al. 2005).

Fifth, calculations (Poliakov et al. 2002; Andrews 2005; Rice et al. 2005) suggest that even if primary shear is confined to a thin zone, the adjoining *damage zone* is likely to *deform inelastically* as the rupture tip passes by because of the localized high stressing near the tip. The effect becomes particularly marked as the rupture propagation speed approaches the Rayleigh speed. The resulting inelastic straining is likely to interact with stressing and energy flow to the slipping process, in a way analogous to what has been studied over many years for tensile crack growth in elastic–plastic solids (typically, structural metals). Thus it is crucial to understand the high strain-rate constitutive response of the highly cracked and granulated material of the damage zone and then to address the mechanics of interaction between inelastic response there and on the main fault. A subtle interaction, yet to be quantified, is that nonlinear constitutive response off the fault plane (which would, in any event, inevitably be asymmetric relative to that plane because of the dependence of shear strength on normal stresses) has the generic effect of altering the normal stress on the fault plane itself. It is not yet known if this has a negligible or perhaps major effect on the shear rupture dynamics.

In this section we have presented a model of a fault zone based on geological observations that are sometimes corroborated by results of laboratory experiments. Such a fault zone model is also consistent with seismological observations based on the analysis of fault zone trapped waves (see Li et al. 1994). These studies have a resolution of meters and show that the damage zone is characterized by lower body wave velocities than the surrounding host rocks, a difference that can reach 50%. The thickness of the damage zone at depth inferred from these studies is consistent with geological observations (10–100 m wide).

The observations presented in this section allow the proposition of a fault zone model characterized by the presence of localized slip in a thin zone, the presence of frictional wear or gouge, a fault core composed of cataclasite and ultracataclasite, and a broader damage zone (highly fractured, anisotropic, and poroelastic). Other observational evidence comes from laboratory experiments on fault friction, which is discussed in the next section. Thereafter we review observational and experimental evidence to shed light on the conflicting and supporting interpretations of data and theoretical modeling.

FAULT MATERIAL RESPONSE TO INTERSEISMIC AND EARTHQUAKE STRESSING

Aseismic to Seismic Transition, Interseismic Stressing, and Earthquake Nucleation

Two concepts concerning what underlies the *transition from ductile to brittle fault response* and the *depth extent of seismogenesis* are to be found in the current literature. The concepts are not obviously identical, but are surely interrelated: an issue is to understand how they are interrelated, and to incorporate both into a proper understanding of seismic phenomena.

The older concept, of a transition from *localized friction* to *broadly distributed creep*, is rooted in ideas of Brace, Evans, Goetz, Kohlstedt, Meissner, Strehlau, and Sibson; for recent assessments see Handy and Brun (2004) and Chapter 6. It leads to the famous pine tree-like plots of crustal strength versus depth. One asks what stress distribution would allow the shallow lithosphere to adjust to remotely imposed plate motions by temperature-dependent creep processes (dislocation and/or grain boundary creep, or fluid-assisted dissolution and transport). Where that stress is less than the friction strength, which is generally assumed to increase linearly with depth, the material is declared ductile; where creep strength is greater than friction strength, it is assumed that the latter dominates and that the response of that part of the lithosphere is brittle, occurring in earthquakes. A problem with this interpretation is that widths over which plate motions would be accommodated by pure creep processes, and hence the strain rates, are not readily estimated; the creep laws are somewhat forgiving on that since stress depends on strain rate raised to a low exponent. Greater problems are that not all localized frictional sliding is unstable and that the distribution of creep strains cannot really be analyzed independently of the episodic stressing pulses (and hence rapid transient creep) delivered by earthquakes to the lithosphere below.

The somewhat more recent concept is that of a transition from *potentially unstable* to *inherently stable* but still localized friction (Tse and Rice 1986; see also Scholz 1990, 1998). This built on earlier, unpublished, modeling concepts by Mavko, on the Brace and Byerlee (1970) results of an absence of stick slip at higher temperatures, and on high-temperature friction data from Stesky et al. (1974) and Stesky (1978) (see Tse and Rice 1986). This approach assumes that over some depth range extending below that where earthquakes can nucleate, the deformation is strongly localized, so that we can discuss response in terms of a constitutive relation between slip rate, temperature, appropriate state variables, and stress. Laboratory studies then show that the constitutive law may exhibit either steady-state rate-weakening ($a - b < 0$, in the formulation given below), or rate-strengthening ($a - b > 0$). If the latter is the case then, at least for the simple one-state-variable class of constitutive laws, earthquakes cannot nucleate. If $a - b < 0$, then they can nucleate provided that a large enough patch of fault—greater than the "nucleation size" (see below)—is made to slip. Thus the depth of seismogenesis is, from this view, not limited by a transition from *localized friction to distributed creep*, but rather by a transition from *potentially unstable* ($a - b < 0$) to *inherently stable* ($a - b > 0$) localized friction. Based on data from Blanpied et al. (1991, 1995) that transition is expected to take place, for a wet granitic gouge composition, at temperatures around 350°C. The difficulty here is that the model assumes localized deformation at all depths considered. We must therefore ask if there is a region slightly downdip of the seismogenic zone in which deformation, while possibly being indeed localized when large earthquakes rupture downward into it, does nevertheless exhibit broadly distributed deformation throughout the interseismic period (see related discussion in Chapter 6).

Both approaches result in a temperature limit for depth of seismogenesis, which can be made to agree within the uncertainty in choosing constitutive parameters and might reflect the same micromechanisms (e.g., onset plastic flow in wet quartz). Also, both mechanisms contribute to limiting seismic ruptures to shallow depths, because they allow for continuing interseismic creep deformation below the locked seismogenic zone. That means the stress prevailing beneath the seismogenic zone at the time of an earthquake is smaller than it must be to allow rapid frictional slip to begin. Thus as the tip of a propagating rupture begins its downward penetration into the hotter material, a negative stress drop (stress becomes higher during rupture) develops that weakens the stress concentration at the tip and soon stops the downward penetration.

Rate and State Frictional Constitutive Laws

These laws begin with the empirical observations and formulations by Dieterich (1978, 1979) and Ruina (1983) and focused on the slow slip range appropriate for nucleation of slip instability under slowly increasing load. They were soon applied broadly to interpretation of laboratory experiments (Tullis 1986), crustal earthquake sequences (Tse and Rice 1986), and descriptions of aftershocks and induced seismicity rate changes (Dieterich 1994). The basic form for these laws, as understood more recently (see Rice et al. 2001, and references therein) is that there is a thermally activated slip process at the stressed microscopic asperity contacts and that contact properties evolve with the maturity of a contact, in a way that strengthens the contact with increase in its age θ. Frictional strength is regarded as being due to atomic bonding (e.g., like at a defect-rich, high-angle grain boundary) at those contacts. The coherence of the contact increases with its age θ, whether due to creep within the contact region which drives the contact towards a less misfitting, lower energy, boundary configuration at the atomic scale, or to desorption of impurities (e.g., water molecules) which were trapped at the contact at levels beyond their equilibrium concentration when it was first formed, and whose desorption thus also lowers energy of the boundary. Those processes would cause the activation energy E for shear processes within the contact to increase with age θ. Also, local creep flow in the vicinity of the contact, at the scale of the contact diameter, can allow the contact area to grow in time (e.g., Dieterich and Kilgore 1996) so that, for a given macroscopic normal stress σ, the average normal stress σ_c at the contacts decreases with age θ.

These contacts may be between roughness asperities on nominally bare surfaces or may be the places where particles of a gouge contact one another. Shear strength τ_c is very high at the contacts and is estimated to be of order $0.1 \times$ shear modulus μ in minerals in which dislocation motion is difficult. (Such estimates can be obtained by using $\tau/\tau_c = \sigma/\sigma_c$, where τ and σ are the macroscopic shear and normal stresses and where both sides of the equation correspond to the ratio of contact area to nominal area, and by measurement of friction coefficient $f = \tau/\sigma = \tau_c/\sigma_c$ and estimate of σ_c from microhardness measurements or, consistently in transparent materials, from optical inference of true contact area; Dieterich and Kilgore 1994, 1996). For the constitutive law, we consider atomic scale thermally activated jumps over energy barriers within the contact zone, with activation energy written as $E - \tau_c\Omega$. Here E is the barrier in absence of bias by the contact shear stress, and Ω is an activation volume.

Thus, if V_1 is the pre-factor in an Arrhenius description (V_1 is estimated to be of order shear wave speed times the fraction of the contact area that slipped by a lattice spacing in an elementary activated event; Rice et al. 2001), the slip rate is

$$V = V_1\left(\exp[-(E - \tau_c\Omega)/k_BT] - \exp[-(E + \tau_c\Omega)/k_BT]\right) \qquad (5.1)$$

The second term, representing backward jumps (in direction opposite to the driving shear traction) is generally negligible except at small positive, or at negative, applied stress. We can usually neglect it. Using $f = \tau/\sigma = \tau_c/\sigma_c$ as above, the Arrhenius law, with backward jumps neglected, is thus equivalent to writing $f = a\ln(V/V_1) + E/\sigma_c\Omega$, which is of a familiar structure in the rate and state formulation. The direct-response parameter a, characterizing the response to a sudden change in V at fixed state of the contacts, is thereby identified as $a = k_BT/\sigma_c\Omega$. This relation together with experimental constraints on a, f and σ_c for quartzite and granite at room temperature led to estimates of Ω equal to a few atomic volumes and $E \sim 1.7$–1.8 e.v. in Rice et al. (2001). Those results had to rely on an estimate of the pre-factor V_1 and could be improved by appropriate experiments to determine response to a stress jump as a function of temperature.

The intrinsic resistance of the contact to shear, represented by E, will increase with the maturity of the contact, and those maturing processes will depend on temperature of the fault zone, on the fluid environment outside the contact, and, of course, on the lifetime θ of the contact. Likewise, growth of the contact area due to creep processes taking place on the scale of the contact diameter corresponds to a reduction in σ_c with increase of θ. The more precise elucidation of those variations with θ is an important goal for tribological research. Assuming that the temperature and external fluid environment remain unchanged during the lifetime of a given contact, we may then assume that $E/\sigma_c\Omega$ is some increasing function of contact lifetime θ; such a function must presently be represented empirically. That naturally introduces a state variable θ into the formulation and, to simplify, it is usual to associate θ with the average lifetime of the contact population. (It may be more fundamental to have the parameters in $E/\sigma_c\Omega$ depend not directly on θ but rather on a measure of lifetime that is stretched or contracted according to temperature, e.g., like $\theta\exp(-Q/k_B T)$ (see Blanpied et al. 1995, 1998, for a related concept). Also, it is well known (Ruina 1983; Tullis 1986) that more than one state variable does better than one at fitting experiments; that may reflect either diversity in the contact populations or presence of more than one maturing process, or both.) Making the assumption that $E/\sigma_c\Omega$ grows logarithmically with contact age θ then leads to a law which can be put into the form of the classic Dieterich-Ruina "ageing," or "slowness," law $f = f_o + a\ln(V/V_o) + b\ln(\theta/\theta_o)$, where b is a new dimensionless parameter (actually, a function of T). Here V_o is an arbitrarily chosen reference value, θ_o is the contact lifetime during sustained steady sliding at rate V_o, and f_o is the associated steady-state friction coefficient. The evolution law for θ is now postulated as a simple law which meets the requirements that (a) $\mathrm{d}\theta/\mathrm{d}t = 1$ when $V = 0$, and (b) θ scales inversely with V in sustained steady sliding (so that in steady state $V\theta = L$, a constant—often denoted by d_c—equal to the sliding distance to renew the contact population; then $f = f_o + (a - b)\ln(V/V_o)$ in steady state). A simple law which accomplishes these features is the one normally used in the ageing formulation (Ruina 1983), namely, $\mathrm{d}\theta/\mathrm{d}t = 1 - V\theta/L$. The above logarithmic form in the expression for f is not sensible for V near zero or negative, but we can handle all such cases by not then neglecting the backward jumps in the original Arrhenius law, leading to the more general expression $f = a\,\mathrm{arcsin}h((V/2V_1)\exp[E/a\sigma_c\Omega])$; that is the version, with logarithmic dependence of $E/\sigma_c\Omega$ on θ like assumed above, sometimes used in simulations of earthquake sequences (e.g., Rice and Ben-Zion 1996; Lapusta et al. 2000).

It will be important to determine how to generalize those concepts to the regime of rapid slips, at rates much greater than those examined in developing the rate and state formulation discussed above, a subject on which there has been some attention already (Prakash 1998). Some rapid-slip models, below, have so far advanced to giving the steady state f as a function of slip rate V, but a description of friction in the form $f = f(V)$ with rate-weakening ($\mathrm{d}f(V)/\mathrm{d}V < 0$) has been shown to either provide an ill-posed model, for which no mathematical solutions generally exist to problems of sliding between elastically deformable continua, or in a limited parameter range when solutions do exist, to predict nonobserved phenomena like rupture fronts that propagate faster than the fastest elastic wave speed (Rice et al. 2001). Thus it is critical to bring any viable constitutive description with steady state rate-weakening to the form $f = f(V, \text{state})$ where "state" represents some set of evolving state parameters like θ and perhaps also contact temperature, and where $a > 0$ (where we recall that $a = V\partial f(V, \text{state})/\partial V$).

Existing treatments of fault response under variable normal stress, following Linker and Dietrich (1992), attempt to map changes in normal stress into changes in θ, because those changes alter the contact population, but more fundamental work is needed too on that. Further, in existing formulations of rate and state friction, it is considered that the deformation is always localized to some thin zone of dimension set by smaller-scale physics, so that overall slip, and not the strain distribution through the thickness of the fault zone, is the only needed descriptor of deformation. However, an issue still to be resolved arises when considering the shear of granular layers that show steady-state rate-weakening. Then because $a > 0$, it is possible that response to a rapid increase in overall slip rate would create broad shear strain throughout the granular layer, whereas the steady-state rate-weakening would take over in sustained slip at that faster rate, promoting highly localized shear. Such effects, to the extent that they matter, have not yet been incorporated into existing models.

Dynamic Weakening Processes (Thermal, Fluid) during Seismic Slip

The materials physics of dynamic weakening had been largely ignored in theoretical modeling of earthquake processes up to relatively recent times. It is now an area of vigorous research. Given that earthquake slips are often accommodated within thin zones, but that evidence of melting is not pervasive, especially at the shallow depths of activity represented by surface exposures, it is reasonable to suspect that strong weakening mechanisms must exist during rapid, large slip. A combination of observation and theory has now identified some important candidates, summarized here. There may be others. The first of these requires rapid but not necessarily large slip. All are discussed in primarily theoretical terms here, citing experimental evidence, although specific challenges for experimental resolution are outlined in the next section. Some of these mechanisms imply a rate-weakening of (steady-state) friction that is much stronger than can be inferred by extrapolating the above rate and state laws to the seismic regime. Such a stronger rate dependence is likely to have a significant role (Cochard and Madariaga 1996; Beeler and Tullis 1996; Zheng and Rice 1998; Nielsen and Carlson 2000) in inducing the *self-healing* rupture mode, which has been advocated on the basis of seismic slip inversions for large events by Heaton (1990). In that mode, slip at a point effectively ceases at a time after passage of the rupture front which is much smaller than the overall event duration.

Flash Heating and Weakening of Micro-asperity Contacts

As noted, shear strength τ_c is very high (estimated to be of order 0.1 × shear modulus μ) at contacts in typical rock systems, and thus when forced to shear, they generate intense but highly localized heating during their limited lifetime, which is of order L/V (here again L is the slip needed to renew the asperity contact population, and V is slip rate). If slip is fast enough, the significantly heated zone is just a thin (relative to contact diameter) region adjoining the contact. The contact's shear strength is diminished by temperature increase, but because the affected zone is thin, the capacity of the contact to support normal stress, and also the net area of contact, are not much affected. Thus the friction coefficient reduces with slip rate V. An elementary first model (Rice 1999, 2004) considers contacts of uniform size L, hence lifetime L/V, and assumes that their shear strength remains at the low-temperature value τ_c until temperature has reached a weakening value T_w, above which shear strength is taken to be zero. (Their temperature rise is estimated from a simple one-dimensional heat conduction analysis, with heating rate $\tau_c V$ per unit area at the sliding contact interface.) The modeling thus identifies a critical slip rate V_w, such that there is no weakening if $V < V_w$, but strong weakening if $V > V_w$. That is, the friction coefficient f (precisely, a *steady state* friction coefficient at slip rate V), which has the value f_o at low slip rates, is given in this simple model by

$$f = f_o \quad \text{if} \quad V < V_w \ , \quad f = f_o \frac{V_w}{V} \quad \text{if} \quad V > V_w \ ;$$
$$\text{here} \quad V_w = \left(\frac{\pi\alpha}{L}\right)\left[\frac{\rho c(T_w - T_f)}{\tau_c}\right]^2 \tag{5.2}$$

where α is thermal diffusivity, ρc is heat capacity per unit volume and T_f is the average temperature of the fault surface; T_f increases gradually due to the heat streaming in at the sliding contacts. [Beeler and Tullis (2003) and Beeler et al. (2006) assume that some low contact strength is retained at $T > T_w$ and modify the latter to $f = f_w + f_o V_w / V$ where $f_w < f_o$.] Goldsby and Tullis (2003) and Rice (1999, 2006) have estimated V_w to range from 0.1 to 0.5 m s^{-1}, when it is recognized that $\tau_c \sim 0.1\mu$. Since the average slip rate in an earthquake is thought from seismic slip inversions (Heaton 1990) to be of order 1 m s^{-1}, the theoretical expectation is that f is reduced significantly from its low speed value f_o during seismic slip. Laboratory experiments imposing rapid slip (Tsutsumi and Shimamoto 1997; Hirose and Shimamoto 2005; Goldsby and Tullis 2003; Tullis and Goldsby 2003; Prakash 2004) are indeed consistent which such an anticipated friction reduction at higher V. The results suggest the possibility that f of order 0.2 to 0.3 may prevail at average seismic slip rates for rocks whose low-speed f is of order 0.6 to 0.7. The weakened f implies a slower heating rate (i.e., that more slip is needed to achieve a given temperature change). When combined with the next thermal weakening mechanism, this leads to effective values of the slip-weakening parameter D_c (see later) which may reach a size ~0.2 m (Rice 2004, 2006).

Thermal Pressurization of Pore Fluid

This mechanism (Sibson 1973; Lachenbruch 1980; Mase and Smith 1985, 1987) assumes that fluids (water, typically) are present within the fault gouge which shears, and that the shear strength τ during seismic slip can still be represented by the classical effective stress law $\tau = f(\sigma - p)$, where σ is normal stress and p is pore pressure (see Cocco and Rice 2002). Frictional heating then would cause the fluid, if it was unconstrained, rather than caged by the densely packed solid particles, to expand in volume much more than would the solid cage. Thus, unless shear-induced dilatancy of the gouge cage overwhelms the thermal expansion effect, or unless the gouge is highly permeable, a pressure increase must be induced in the pore fluid. Since σ typically remains constant during slip, strength τ is reduced, ultimately towards zero, as shear heating continues to raise temperature so that p approaches σ. Calculations by Rice (2006) evaluated this mechanism using permeability and poroelastic properties and shear zone thicknesses based on properties of the Median Tectonic Line fault (Wibberley 2002; Wibberley and Shimamoto 2003), Nojima fault (Lockner et al. 2000), and Punchbowl fault (Chester and Chester 1998; Chester and Goldsby 2003; Chester et al. 2003). Rice (2004) as well as Cocco and Bizzarri (2004) found that predictions based on thermal pressurization enabled plausible estimates of the fracture energy of earthquakes, as have been established independently in seismological studies (for recent summaries, see Abercrombie and Rice 2005, Rice et al. 2005; Tinti et al. 2005; Rice 2006; Bizzarri and Cocco 2006a, b), and could explain why strength loss over all but deeper portions of crustal seismogenic zones is too rapid for melting to take place. That seems consistent with general conclusions that fault zone pseudotachylites of tectonic earthquake origin have generally formed deep in the seismogenic zone.

An illustration is provided by the analysis (following Rice 2004, 2006) for slip at speed V on the fault plane $y = 0$ (zero thickness shear zone) in a poroelastic solid under constant normal stress σ_n. With standard simplifications, the pore pressure p and temperature T then satisfy:

$$\text{In } |y| > 0 \ , \quad \frac{\partial T}{\partial t} = \alpha_{th} \frac{\partial^2 T}{\partial y^2} \quad \text{and} \quad \frac{\partial p}{\partial t} - \Lambda \frac{\partial T}{\partial t} = \alpha_{hy} \frac{\partial^2 p}{\partial y^2} \ , \tag{5.3}$$

with conditions on

$$y = 0^{\pm} \ , \quad -\rho c \alpha_{th} \frac{\partial T}{\partial y} = \pm \frac{1}{2} f(\sigma_n - p) V \quad \text{and} \quad \frac{\partial p}{\partial y} = 0 \ .$$

Here α_{th} and α_{hy} are the respective thermal and hydraulic diffusivities, Λ is the value of dp/dT under undrained conditions, and ρc is the specific heat per unit mass. The two partial differential equations express energy conservation, assuming conductive heat transfer but neglecting advective transfer, and conservation of fluid mass during the increase of T and p and Darcy transport. The boundary conditions express that the frictional work rate provides the heat input at the fault, and that there is no fluid outflow from a vanishingly thin zone. These can be solved for the case of constant V and f, and when the solution is written in terms of slip δ (= Vt), the p and T on the fault plane $y = 0$ are

$$\begin{aligned} p - p_o &= (\sigma_n - p_o)\left[1 - \exp\left(\frac{\delta}{L^*}\right)\operatorname{erfc}\left(\sqrt{\frac{\delta}{L^*}}\right)\right] \\ T - T_o &= \left(1 + \sqrt{\frac{\alpha_{hy}}{\alpha_{th}}}\right)\frac{p - p_o}{\Lambda} \end{aligned} \tag{5.4}$$

where $L^* = \dfrac{4}{f^2}\left(\dfrac{\rho c}{\Lambda}\right)^2 \dfrac{\left(\sqrt{\alpha_{hy}} + \sqrt{\alpha_{th}}\right)^2}{V}$,

and where p_o and T_o are the initial values; T_o would be the ambient temperature but, if we sensibly assume that the onset of shear should be associated with some inelastic dilatancy of material near the fault zone, inducing a sudden suction in it (Segall and Rice 1995), then p_o should be assumed to be reduced from the ambient value by that suction. The shear stress transmitted across the fault plane is thus predicted to be

$$\tau = f(\sigma_n - p) = f(\sigma_n - p_o)\exp\left(\frac{\delta}{L^*}\right)\operatorname{erfc}\left(\sqrt{\frac{\delta}{L^*}}\right) . \tag{5.5}$$

This shows continued weakening at an ever-decreasing rate over a very broad range of size scales, Figure 5.2. How large is L^*, the single length scale which enters into the description of the weakening process under the conditions considered? Lachenbruch (1980) gives $\rho c \approx 2.7$ MPa °C^{-1} and $\alpha_{th} \approx 1$ mm^2 s^{-1}. Supplementing his data set by results from Wibberley (2002) and Wibberley and Shimamoto (2003) for gouge from the central slip zone of the Median Tectonic Line at 130 MPa effective confining stress ($\approx$ effective overburden at 7 km depth, for hydrostatic pore pressure), with porosity 0.04 (Wibberley, pers. comm.), gives $\Lambda \approx 0.8$ MPa °C^{-1} and $\alpha_{hy} \approx 1.8$ mm^2 s^{-1}, using a permeability of 10^{-20} m^2. That gives, with $V = 1$ m s^{-1} and $f = 0.25$ to allow for flash heating, $L^* = 4$ mm. That estimate directly uses lab data on undisturbed gouge samples, whereas in the natural situation there may be initial dilatant deformation and

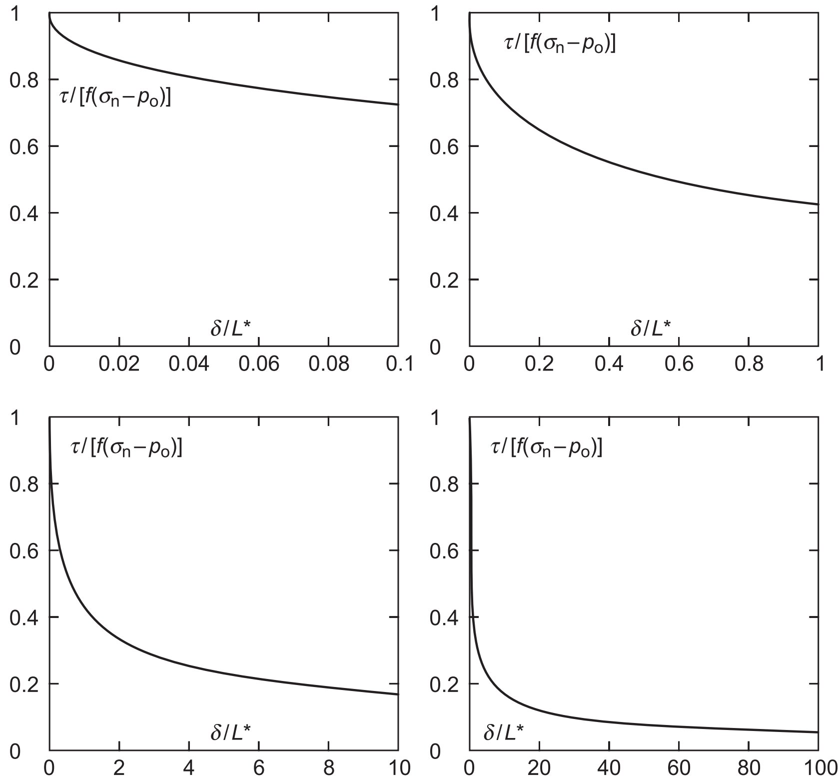


Figure 5.2 Prediction of shear strength τ versus slip δ, due to *thermal pressurization of pore fluid* during slip on a plane, at constant rate V and with constant friction coefficient f, in a fluid-saturated solid. The four panels show the solution in Eq. 5.5 for different δ/L^* ranges, extending from 0 to, respectively, 0.1 (upper left), 1.0 (upper right), 10 (lower left), and 100 (lower right). After Rice (2004, 2006). Note the multiscale nature of the weakening. Parameter L^* (see text) is estimated to be in range 4–30 mm at representative centroidal depths, ~ 7 km, of the slipping region during crustal earthquakes. Here σ_n is fault-normal stress and p_o is the pore pressure just after its reduction from ambient pressure by any dilatancy at onset of shear.

damage near the rupture front (e.g., Poliakov et al. 2002; Rice et al. 2005) at the start of shear. As guesses as to how such effects might change parameters, a tenfold greater permeability, 10^{-19} m^2, and doubled pressure-expansivity of the pore space (making $\Lambda \approx 0.5$ MPa °C^{-1} and $\alpha_{hy} \approx 12$ mm^2 s^{-1}), would instead give $L^* = 33$ mm. (These estimates were adopted from a preliminary version, Rice (2004), of those published in Rice [2006]. The final 2006 version has a much fuller analysis of the underlying data, including its T and p dependence. The resulting poro-thermo-elastic parameter values in the final version are different from those just quoted but, nevertheless, $L^* = 4$ mm and 30 mm remain plausible "low end" and "high end" values, as representative of properties of intact gouge and of gouge with the guessed effects of damage as above, respectively.) Thus the upper right panel in Figure 5.2, showing slip up to $\delta = L^*$, could correspond to a maximum slip between 4 and perhaps 30 mm slip, and the lower right panel, showing slip up to $\delta = 100L^*$, to a maximum slip between 0.4 and 3 m.

Bizzarri and Cocco (2006a,b) have performed three-dimensional simulations of the spontaneous nucleation and propagation of a dynamic rupture governed either by slip-weakening or rate- and state-dependent laws (defined in more detail below), thus including a cohesive zone where dynamic weakening occurs and accounting for frictional heating (by solving the heat flow equation; see Fialko 2004) and thermal pressurization of pore fluids (by solving the above fluid diffusion equation, like in Andrews 2002). These authors analyzed the traction evolution as a function of time or slip for different values of fault zone thickness and hydraulic diffusivity. In these simulations, slip velocity evolves spontaneously and is not constant. Bizzarri and Cocco (2006a,b) have shown that fault zone thickness and hydraulic diffusivity modify the shape of the traction versus slip curves and affect the stress drop and the critical slip-weakening distance (D_c, defined below; see Figure 5.4). For particular configurations they found that traction evolution shows a gradual and continuous weakening such as those shown in Figure 5.2 and predicted by Eq. 5.5.

The fracture energy G associated with an event with slip δ may be calculated from a slip-weakening function $\tau = \tau(\delta)$ by

$$G = G(\delta) = \int_0^\delta [\tau(\delta') - \tau(\delta)] d\delta' \ ,$$

that definition (Abercrombie and Rice 2005) generalizes Ida (1972) and Palmer and Rice (1973) to forms of $\tau(\delta)$ for which weakening continues at an ever-decreasing rate to slips greater than that in the event. Then, for the above $\tau(\delta)$, there results (Rice 2006)

$$G(\delta) = f(\sigma_n - p_o)L^* \left[\exp\left(\frac{\delta}{L^*}\right) \operatorname{erfc}\left(\sqrt{\frac{\delta}{L^*}}\right)\left(1 - \frac{\delta}{L^*}\right) - 1 + 2\sqrt{\frac{\delta}{\pi L^*}} \right] . \quad (5.6)$$

This is plotted as the solid lines in Figure 5.3, for the low and high poromechanical estimates of L^* above, using $V = 1$ m s^{-1}, $f = 0.25$ as an approximate representation of flash heating (Hirose and Shimamoto 2005; Prakash 2004; Goldsby and Tullis 2003), and $\sigma_n - p_o = 126$ MPa (initial effective overburden at 7 km depth, a representative centroidal depth for slip during crustal events). Seismic data assembled recently by Abercrombie and Rice (2005) and, for the seven large events of Heaton (1990), by Rice et al. (2005) is also shown there. The seismic data and theoretical modeling have many uncertainties.

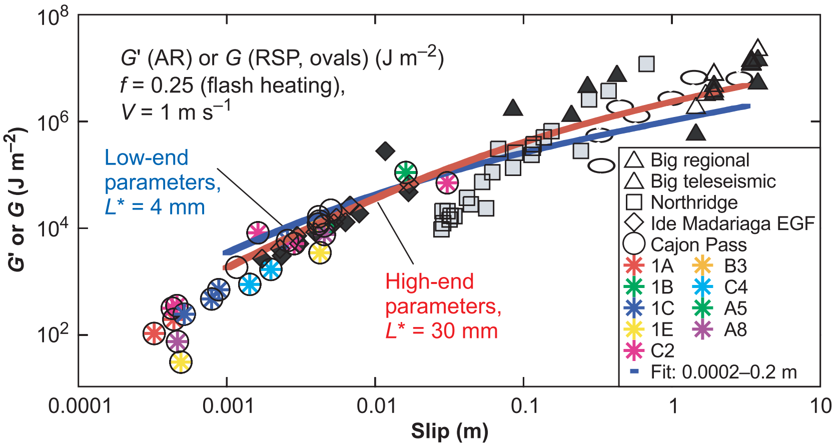


Figure 5.3 Lines show theoretical predictions of earthquake fracture energy G versus slip δ in the event, based on combined effects of *thermal pressurization* of pore fluid and *flash heating*, with simplified represention at constant friction coefficient f and slip rate V. Symbols represent estimates of G from seismic data. The basic plot, to which the curves and oval symbols have been added, is from Abercrombie and Rice (2005); it shows their parameter G', thought to be of the same order as G ($G' = G$ when final dynamic sliding strength and final static stress coincide). Oval symbols at large slip from Rice et al. (2005) for the seven large earthquakes with slip inversions reported by Heaton (1990). (The figure is a preliminary version, taken from Rice (2004), of a corresponding figure in Rice (2006); the 2006 version compares theoretical predictions to an enlarged seismic data set for large earthquakes, including results from Tinti et al. (2005) and other sources.)

Nevertheless, the rough coincidence does lend credibility to the concept that flash heating together with thermal pressurization of pore fluid may be dominant processes in fault weakening during earthquakes. (Figure 5.3 is from a preliminary version, taken from Rice (2004), of the corresponding figure in Rice (2006); the final 2006 version is similar but compares to a larger seismic data set including results from Tinti et al. (2005).)

There is a critical need for laboratory tests of this mechanism. Also, it is important to understand if and how it might operate in conditions that might exist at midcrustal depths, possibly involving an extensively mineralized pore space with isolated, unconnected pockets of liquid water, for which the classical dependence of strength on $\sigma - p$ might then not hold. For example, if shear of such a zone has been initiated by a propagating rupture tip, so that a possibility of fluid connectivity is reestablished, can it then be assumed that the effective stress characterization of strength, $\tau = f(\sigma - p)$, holds again? Further, we need to understand how to represent the shear strength of a dense, rapidly shearing, gouge with $\sigma = p$. Is it sufficiently small to be negligible compared *to* $f(\sigma - p_{\text{initial}})$? Or does it represent a substantial fraction of that value? Can the condition of p

approaching σ actually be achieved on a real fault? Or will any such highly pressurized pore fluids hydraulically crack their way into the already damaged walls of the fault, (since p will then be at least as great, and generally greater, than the least principal compressive stress)? Will such hydraulic cracking effectively increase permeability and bring the fault weakening to a halt (Sibson 1973), perhaps abruptly halting slip?

Finally, it is essential to constrain better the dilatancy of a gouge under shear (Marone et al. 1990; Segall and Rice 1995), and also the effect of dilatancy and shear on the instantaneous permeability and poroelastic moduli. Those values are essential inputs for quantitative estimates of fault weakening by thermal pressurization, but at present they are known only for nondeforming gouge. It is also essential to understand how the evolution of porosity is related to that of surface contact properties, if we aim to constrain a constitutive behavior for the principal slipping zone, as well as to determine where porosity evolution is of relevance (within the damage zone, the fault core, or the slip zone).

Silica Gel Formation

This mechanism, reported by DiToro et al. (2004) (see also Goldsby and Tullis 2002; Roig Silva et al. 2004), and supported by observations of fracture surface morphologies, applies most directly for large slip (>0.5–1.0 m) and moderately rapid slip (>1 mm s^{-1}), in the presence of water. It assumes that there is a silica component (quartz) within the shearing zone. Present understanding is linked to the observation in friction experiments on a quartzite, Arkansas novaculite, of "now solidified … flow-like textures that make it … evident that at the time the deformation was going on, a thin layer coating the sliding surface was able to flow with a relatively low viscosity" (Tullis, pers. comm.). The concept is that granulation within the shear zone produces fine silica particles which adsorb water to their surfaces and form a gel. It is weak but would gradually consolidate into a strong, amorphous solid if shear was stopped. However, the presence of shear continuously disrupts particle bonding (thixotropic response) so that the fluidized gel mass deforms at low strength. Fuller limits to their range in which the mechanism is active have not yet been identified, and it is not known if it could contribute during seismic slip. Nevertheless, for a given shear and within a given velocity range for which the mechanism was plausibly established for pure quartzite rocks (Arkansas novaculite), weakening for other rock types seems to be ordered by their silica content (Roig Silva et al. 2004):

quartzite (*novaculite*) > *granite* > *gabbro.*

Granite and a pure albite-feldspar rock with nearly identical silica content show essentially identical weakening, although the "solidified flow structures have so far only been seen for novaculite" (Tullis, pers. comm.). A gel can be thought of for some purposes as a water-infiltrated porous medium. That raises the question of whether there could be any connection between this weakening mechanism and the Sibson (1973) mechanism of weakening by thermal elevation of pore fluid pressure. Nevertheless, even if so, a new ingredient here is the strong thixotropic response, causing a fractionally much greater regain of strength after cessation of shear than what might be expected that a granular gouge would show on a comparable timescale.

Melting

This is the ultimate mechanism of thermal weakening, but it is not a simple mechanism. Comparison of strength during rapid shear in the range prior to, versus shortly after, the transition to macroscale melting (i.e., when a coherent melt layer has formed along the whole sliding surface) shows that at or near the transition, there is an abrupt increase in frictional strength (Tsutsumi and Shimamoto 1997; Hirose and Shimamoto 2005). At least, that increase is observed for the small normal stresses σ in experiments reported thus far; Tsutsumi and Shimamoto (1997) find that, at $\sigma = 1.5$ MPa, the friction coefficient $f = \tau/\sigma$ shoots up to ~0.9. Similar behavior is known in other parts of the tribological literature. There is complex response on the way to the transition: Sizeable blobs of melt form near the larger frictional contact asperities, as an extreme form of the flash heating process. These get smeared out along the sliding surfaces and rapidly solidify at least when the average temperature of those surfaces is low enough, so that the surfaces are spot-welded together. During macroscopic melting, the fault strength is $\tau = \eta V/h$ where h is the thickness of the shearing layer and viscosity $\eta = \eta(T$, melt composition). In the early phases of macroscale melting, h is small and T is low, compared to what it will become with continued shear, and thus τ is relatively high. It is only with increasing h, so that shear rate decreases, or with increasing T, so that η decreases, that the melted fault zone weakens to a friction level comparable to what it showed (due to flash heating) at a comparable slip rate shortly before the transition to macroscale melting.

Important problems for the future are to understand and quantify shear resistance in the transition range (Hirose and Shimamoto 2005; Di Toro et al. 2006) and the macroscale melting range. With the present understanding, as outlined here, if water or other fluids are completely absent, we would have only frictional heating (at a reduced f due to flash heating) and then a transition to melting. Melting itself is complex and takes place under strongly non-equilibrium conditions. For example, whereas a hydrous granitic composition could be expected to equilibrium melt over an ~50°C interval, pseudotachylite studies (Otsuki et al. 2003; Di Toro et al. 2005) suggest a 500–600°C melting interval (starting at 750°C) for the melting of a granitic composition on the earthquake timescale (~1 s for 1 m of slip, since the average V is inferred to be ~1 m s^{-1}; that is based on seismic slip inversions for large events, Heaton 1990, and is obtained by dividing his inferred slip by inferred slip duration at a point). For the macroscale melting range, a primary issue is to describe the evolution of h, T, and melt composition (including reactions with initial pore water of the fault zone). Presumably h is set by the balance between melt generation and either freezing or loss by melt injection into cracks in the fault wall. Concerning injection, we must recognize that the melt layer is under pressure $p = \sigma$, the fault-normal stress, and that in general the propensity to inject should increase with the increasing difference between σ and the least principal compressive stress at the fault wall.

New Advances Needed in Laboratory Experiments on Fault Friction

Rapid Slip

New advances have been reported quite recently in achieving slip in rocks at high rates, approaching or exceeding what is thought to be the average slip rate of order 1 m s^{-1} during seismic instability. The work achieves rates up to 0.3 m s^{-1} (Goldsby and Tullis 2003) using a hydraulically-driven mechanical testing machines capable of torsional loading, and rates of 3 to 30 m s^{-1} (Prakash 2004) based on sudden unloading of a pre-torqued Kolsky bar or, for the higher rates, adaptation of the oblique shock impact arrangement of Prakash and Clifton (1992) and Prakash (1998). These new approaches have the promise in opening a new chapter in making laboratory rock friction studies relevant not just to nucleation of unstable slip, but to characterizing the fault constitutive response during the dynamic rupture process itself. They, or such other new experimental techniques as may be devised, need to be more widely adopted to develop this important phase of the subject. They will be needed to confidently characterize such mechanisms of weakening as flash heating at microscopic asperity contacts and thermal pressurization of pore fluids discussed above.

Fault Strengthening

A severe limitation of current frictional constitutive modeling in the rate and state framework is the almost completely empirical manner by which time dependent fault strengthening is brought into the description. For example, in the commonly adopted ageing, or slowness, version of the Dieterich-Ruina friction law discussed earlier, f is assumed to increase in proportion to $\log(\theta)$ where θ is interpreted as contact lifetime (or at least as its average value). The physical basis of actual healing, and its dependence on θ or other appropriate variables to be identified, needs to be better quantified by appropriate experiments. The physical chemistry of water is relevant here; water-assisted processes have been shown, in variable humidity studies, to play a major role in time-dependent strengthening at least at room temperature (Dieterich and Conrad 1984; Frye and Marone 2002). Exclusion of water seems to remove most of that time-dependent strengthening. In the

Earth, very long times of (nearly) stationary contact are of interest for the interseismic period, and their effect on strengthening needs fuller experimental clarification, and basic physical understanding, at temperatures up to those slightly higher than for the seismogenic depth range. Significant contact strengthening by mineral deposition, alterations by reactions with fluids (the kinetics of hydration is critical for that; see Chapter 12), and creep flow may be important; those processes are also of interest for overall energy balances in faulting. Creep compaction of fine fault gouge and depositional processes may also seal off fluids into noncommunicating pore spaces and drive pore pressure above hydrostatic values, allowing low-strength shear (Sleep and Blanpied 1992). Constitutive response after such long hold times at high temperature is directly relevant to nucleation of large events (e.g., Tse and Rice 1986; Lapusta and Rice 2003), which usually occurs towards the lower depth range of the effectively locked seismogenic zone. It is also relevant to evaluating aftershock production (Dieterich 1994).

Localized versus Distributed Shear

In many attempts to represent fault gouge effects, by shear of granular layers in the lab (e.g., Marone 1998; Scruggs and Tullis 1998), there is a variation in deformation response between stable shearing that may span the full width of the granular layer versus localized shearing that may take place on a family of imperfectly aligned shear structures, or may take place on a through-going surface. This remains an important process to characterize and understand, especially at higher shear rates typical of seismic stressing, for which thermal weakening processes may contribute to localization in an otherwise stable granular shear flow. Beeler et al. (1996) proposed that considering the combined dependence of strength on slip and slip rate might be of relevance and developed a relationship between the change of slip zone thickness and the strength change in such a formulation.

A more fundamental starting point might be to begin with a formulation in terms of shear rate and evolving state parameters, including porosity, of the fault gouge. Nevertheless, in any such study, we run up against a pervasive difficulty, recognized for many years in the macroscopic mechanics of ductile and granular materials, which is that of representing the full transition from distributed deformation to highly localized deformation like in a shear band or fault. This must generally be addressed by appeal to length parameters of the smaller-scale physics that limits localization zone thickness, for example, by adding positive-definite strain gradient terms to the strength expression (by dimensional considerations, their coefficients will necessarily introduce a length scale), or using a spatially nonlocal relation between strength and deformation, or by setting the cell size of a finite element or finite difference computational grid to the length parameter, or in cases for which it is appropriate like adiabatic shear localization in temperature-weakening materials, by including transport phenomena (e.g., α_{th}/V provides a length scale). In all but the latter approach, the present level of understanding is such that these must generally carried out as ad hoc procedures, often without clear identification of the relevant microscopic physics and length scales, or convincing demonstration that it leads to the localization-limiter procedure adopted. It is generally accepted that localized shear zones span ~5–30 $d_{50\%}$ in aggregates of relatively equiaxed granular particles like for sands (Desrues and Viggiani 2004), and possibly as much as 200 $d_{50\%}$ in fine-grained clays with platelet particles (Morgenstern and Tschalenko 1967; Vardoulakis 2003), where $d_{50\%}$ is the particle diameter dividing the aggregate into two equal masses. However, we do not presently know how such results generalize to real fault gouges, with a wide particle size distribution (Chester et al. 1993; Ben-Zion and Sammis 2003) from, say, 0.01–100 µm (the diameter to be used is unknown), with particles that are often angular and are susceptible to cracking.

Variable Normal Stress

This arises in slip propagation along dissimilar material interfaces (Weertman 1980; Andrews and Ben-Zion 1997; Harris and Day 1997; Cochard and Rice 2000; Xia et al. 2005), because gradients in slip then induce changes in normal stress, and also when the propagating front of a rupture encounters a bend or branch in the fault path (Poliakov et al. 2002; Kame et al. 2003), rapidly increasing both normal and shear stress there. It is also relevant to rapid alterations of pore pressure due to shear heating and thermal pressurization of pore fluid. There remains conflicting evidence on its effect of frictional strength: Precise laser diagnosis of oblique shock wave experiments at slip rates of order 10 m s^{-1} involving abrupt normal stress changes by a reflected sharp shock front (Prakash and Clifton 1992; Prakash 1998) suggest that there is no correspondingly abrupt change in shear strength, but only an evolution with continuing slip (over a few µm scale). In contrast, conventional slow friction studies with much less abrupt change of normal stress (Linker and Dieterich 1992; Boettcher and Marone 2004) suggest that shear strength changes on the same timescale as the normal stress change, but changes only partly towards what it will ultimately evolve to after a few µm of further slip.

Scaling from Lab to Natural Faults

It is commonly asserted that results of laboratory experiments must be "scaled," in some manner yet to be determined, to the geometrically much larger natural fault scale. That would surely be a valid point of view if the lab experiments were done on other than natural fault materials (or hopefully similar lithologies), and were done to produce laboratory analogs of crustal earthquakes. However, the laboratory studies that we have discussed are aimed at determining local constitutive response of fault materials. A viable proposition to be discussed is that they require no scaling whatsoever and are directly applicable to the Earth, as descriptors of local response on a fault. All that may be called "scaling" is then just a description of those predictions which result when appropriate boundary and initial value problems are set for mathematical models, incorporating those constitutive relations, to predict earthquake behavior. The tacit assumption, then, is that the experimental fault rock or gouge zone responds to any given history of slip rate and normal stress with the same history of shear traction as that which would occur along the actual fault zone. That is a reasonable assumption when there is a clear separation between the scales of microphysical processes determining the local response and the (presumably) more macroscopic scales over which the locally averaged stress and slip vary. An area of some uncertainty here, however, involves the fractal-like roughness of faults (Power and Tullis 1991). Friction response as we understand it seems to be controlled at the multi-micron size scale by properties of contacts. Is there then any effect, on what we use for constitutive laws, of the nonplanarity of faults at larger size scales? Or is it properly accommodated (not that such is yet done in practice) by regarding the normal stress supported by the fault as a variable which has an effectively random component at larger size scales?

That concept—that there may be no scaling to be imposed on laboratory (constitutive equation) results—should not be confused with a scaling-like compromise that fault modelers must generally make to fit their problems on current-day computer systems. Computational tractability of crustal scale earthquake models requires use of far larger constitutive length scales L in description of frictional weakening than can be justified experimentally. That is because the required size of numerical discretization cells, for a numerical solution to represent the solution of the underlying continuous system of equations on a rate-weakening fault, must be small compared to the size of a slipping patch at the transition from aseismic to seismic slip (which is called the nucleation size). That nucleation size scales roughly as (Lapusta et al. 2000) L times a large number, of order $5\mu/[(b-a)(\sigma-p)]$ where μ is the elastic shear modulus and other parameters are as above. At 10 km depth, identifying σ as overburden and taking p as hydrostatic, and taking $b-a=0.004$ (Blanpied et al. 1991), ends up meaning that the grid size should be small compared to 2×10^5 L, which would mean small compared to a 1–5 m nucleation size if $L = 5$–50 μm, representative lab values. Such resolution is unattainable in simulations which try to resolve the entire crustal scale. The conventional approach is then to choose an artificially large L so as to make the required discretization size large enough, and then to try to get some sense of what happens as L is reduced as close as one can get to the laboratory range. However, it has recently been realized that this is a nontrivial extrapolation, in that new populations of smaller events emerge near the downdip end of the seismogenic zone as L is reduced in size (Lapusta and Rice 2003), a complication which does nevertheless have compatibility with natural observations of earthquake locations.

REPRESENTATION OF FAULT CONSTITUTIVE BEHAVIOR FOR PREDICTIVE EARTHQUAKE MODELS

Slip Weakening

To use experimental and theoretical results on constitutive response, as in the last section, for predicting large scale fault rupture behavior, it is necessary to contend with the art of simplification, that is, of identifying simple, tractable modeling procedures, but not so simplified as to lose essential features. To that aim, a main result that is relevant to characterizing dynamic fault weakening during an earthquake is the traction evolution. Weakening during rupture propagation is now conventionally represented by the traction drop associated with slip increase (see Figure 5.4), resulting in the well known slip-weakening model. Different physical processes can yield a traction evolution consistent with that behavior (Cocco and Bizzarri 2002). This shear stress degradation during dynamic propagation occurs in a finite extended zone at the crack front called the cohesive zone (Barenblatt 1959; Ida 1972; Palmer and Rice 1973; see Figure 5.5). Although the shape of the slip-weakening curve can differ among different constitutive formulations, such a traction variation with slip must be common to any constitutive relation proposed to model rupture propagation. Because of its simplicity and in order to prescribe the traction evolution within the cohesive zone (see Figure 5.5), the slip-weakening model has been widely used as a

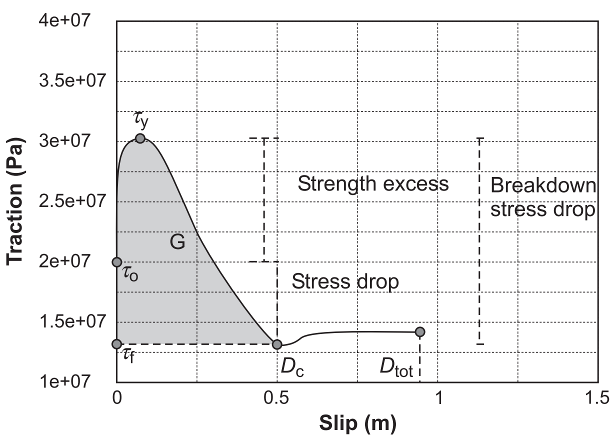


Figure 5.4 Traction evolution as a function of slip obtained by a numerical experiment of spontaneous dynamic propagation on a fault. This kind of evolution is common to different constitutive formulations (see Bizzarri and Cocco 2003, and references therein). τ_o is the initial stress, τ_y the upper yield stress, and τ_f the kinetic friction level; the characteristic slip-weakening distance is D_c, and D_{tot} is the final slip value. The difference ($\tau_y - \tau_o$) is usually named the strength excess, while ($\tau_o - \tau_f$) is the stress drop. The shaded area indicated by G yields an estimate of the fracture energy.

constitutive relation to model dynamic rupture with theoretical and numerical approaches (Andrews 1976a, b). The main parameters of this model are the initial stress (τ_o), the upper yield stress (τ_y) or peak strength, the kinetic friction level (τ_f) or residual strength, and the characteristic slip-weakening distance D_c (see Figures 5.4 and 5.5). Dependence of traction on slip has been observed in dynamic laboratory experiments (e.g., Ohnaka and Yamashita 1989).

A set of important questions arise as we try to unite this established rupture dynamics methodology with new physical understanding and laboratory documentation of friction behavior. This has been partly addressed, as we explain later, for interpreting dynamic predictions based on the now classical, logarithmic, rate and state friction laws in terms of slip-weakening concepts. However, the recent theories and experiments for the range of rapid, large slip involve, in some cases, much stronger steady-state rate-weakening at seismic slip rates than predicted by extrapolation of the logarithmic laws. Could that bring on new types of dynamic response, like inducement of self-healing of the rupture (Cochard and Madariaga 1996; Beeler and Tullis 1996; Zheng and Rice 1998; Nielsen and Carlson 2000), which might be obscured by mapping laws for the range of rapid, large slips into the slip-weakening framework? Also, some of the theoretical constitutive modeling of weakening in large slips by thermal

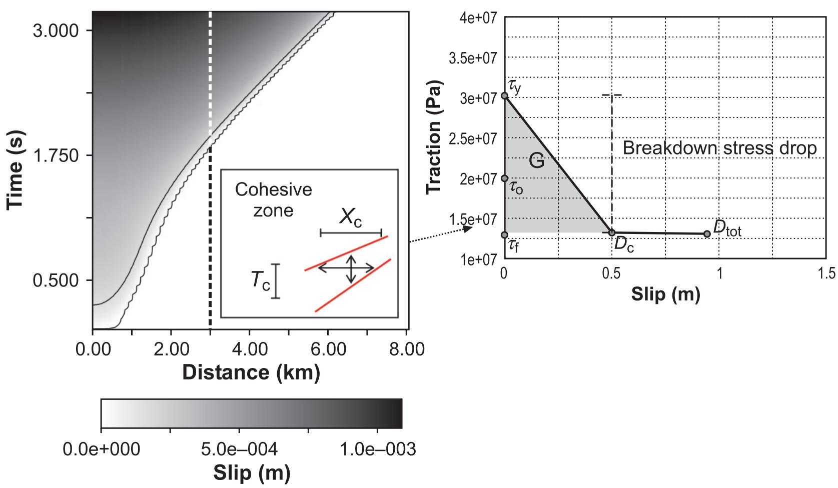


Figure 5.5 Spatiotemporal evolution of slip obtained by a numerical experiment of spontaneous dynamic propagation on a two-dimensional fault obeying to a slip-weakening law stated in Eq. 5.8 and shown in the right panel. This sketch allows the identification of the cohesive or breakdown zone, which is defined as the region shear stress degradation from the upper yield stress to the kinetic friction level. The spatial dimension of the cohesive or breakdown zone (X_c) is different from the critical slip-weakening distance D_c. The difference (τ_y–τ_f) is the breakdown strength drop in the terminology of Ohnaka and Yamashita (1989).

pressurization of fluid (Figures 5.2 and 5.4), as well as seismic attempts to look at the scaling of fracture energy with slip in an earthquake (Abercrombie and Rice 2005; Rice et al. 2005; Tinti et al. 2005), suggest that the effective slip-weakening law might have a multiscale character not envisioned in the classical formulations of slip weakening. That multiscale character means that discernible weakening continues, at an ever-diminishing rate with slip, out to large slips, say, of order of a meter. Such response, when fitted to classical linear slip-weakening models (Figure 5.5), has led instead to the interpretation (Ohnaka 2003) that D_c depends on the slip in an earthquake.

Different Constitutive Formulations

The choice of a fault constitutive law is necessary to solve numerically the elastodynamic equations and to model spontaneous nucleation and propagation of an earthquake rupture. A constitutive law relates the total dynamic traction to fault friction and allows the absorption of a finite fracture energy (named G in Figures 5.3 and 5.5) at the crack tip. Different constitutive relations have been proposed in the literature. They can be grouped in two main classes: slip-dependent (Andrews 1976a, b; Ohnaka and Yamashita 1989) and rate- and state-dependent (R&S) laws (Dieterich 1979; Ruina 1983). The former assumes that friction is a function of the fault slip only, whereas the latter implies that the friction is a function of slip velocity and state variables. (The former can in fact be considered as the time- and rate-insensitive limit of a law formulated in a general R&S framework.) The analytical expression of the classical slip-weakening law (Andrews 1976a, b) is:

$$\tau = \begin{cases} \tau_y - \left(\tau_y - \tau_f\right)\dfrac{\Delta u}{D_c} & \Delta u < D_c \\ \tau_f & \Delta u \geq D_c \end{cases} \tag{5.7}$$

where Δu is the slip. The traction evolution associated to this law is shown in Figure 5.5 and it is characterized by a constant and linear traction decay and by a constant kinetic friction level (τ_f). The analytical expression of R&S friction laws is, as explained earlier, composed by two equations (written here generally, and then in their most commonly used forms): the strength and the evolution laws. These are, respectively,

$$\begin{aligned} \tau &= \Im(V, \Phi, \sigma_n^{eff}) = f(V, \Phi) \cdot \sigma_n^{eff} \\ &= \left[f_* + a \ln\left(\frac{V}{V_*}\right) + b \ln\left(\frac{V_* \Phi}{L}\right)\right] \cdot \sigma_n^{eff} \\ \frac{d\Phi}{dt} &= g(V, \Phi) = 1 - \frac{\Phi V}{L} \end{aligned} \tag{5.8}$$

where f is the coefficient of friction, σ_n^{eff} is the effective normal stress, V is the slip velocity, a and b are constitutive parameters, together with length L, and the reference parameters f_* and V_* (of which one can be chosen arbitrarily). Φ is the state variable (like denoted by θ earlier), which provides a memory of previous slip episodes and its evolution equation guarantees a time dependence of friction. In particular, the analytical expression stated in Eq. 5.8 is the ageing (slowness) law proposed by Ruina (1983) and Dieterich (1986), and motivated by Dieterich (1979). It represents one possible formulation among different ones in the literature (see Beeler et al. 1994).

The characteristic length scale parameters of these two constitutive formulations are the slip-weakening distance D_c and the parameter L: the former represents the slip required for traction to drop; the latter is the characteristic length for the renewal of a population of contacts along the sliding surface and controls the evolution of the state variable. These two length scale parameters are different (Cocco and Bizzarri 2002): Bizzarri and Cocco (2003) proposed analytical relations to associate slip weakening and R&S constitutive parameters. Such an association is possible because the commonly used versions of R&S laws lead to predictions which mimic those of the slip-weakening description when applied to a situation of rapid increase in slip rate, like at a propagating rupture front. In fact, the most general form of R&S law is broad enough that by choice of $f(V,\Phi)$ and $g(V,\Phi)$, we can duplicate an arbitrarily chosen slip-weakening law $f = f(\text{slip})$; e.g., set $g(V,\Phi) = V$ so that Φ is the slip, and choose $f(V,\Phi) = f(\Phi)$.

R&S Friction Laws Applied to Earthquake Modeling

The R&S constitutive formulation allows the modeling of rupture nucleation (Lapusta et al. 2000, and references therein), dynamic rupture propagation (Bizzarri et al. 2001, and references therein) as well as fault restrengthening during the interseismic period; therefore, it has been used to simulate repeated seismic cycles. R&S constitutive laws allow the definition of different frictional regimes: an earthquake is associated with an instability occurring in a velocity weakening field, meaning that there is negative change of fault strength with slip rate, $d\tau_{ss}/d\ln(V) < 0$ (where τ_{ss} is the steady-state traction), which corresponds to $(a - b) < 0$. Theoretical and observational evidence shows that, while spontaneous nucleation and dynamic propagation of an earthquake rupture are governed by the constitutive behavior (and the analytical form of Eq. 5.8 does matter), rupture arrest is more likely associated with geometrical complexities as well as frictional or rheological heterogeneities, which can stop the propagating dynamic rupture front. In the framework of the R&S laws, a velocity-strengthening regime (defined by $d\tau_{ss}/d\ln(V) > 0$, or $(a - b) > 0$, as occurs at higher temperatures [Blanpied et al. 1991, 1995]) can arrest the rupture, although it can also participate in a dynamic rupture if loaded enough and $(a - b)$ is relatively small. Rice (1993) and Boatwright and Cocco (1996) discussed the frictional control of crustal faulting and presented a classification of fault heterogeneity in terms of nonuniform distribution of the R&S constitutive parameters on the fault plane. This represents understanding of rate-weakening effects that it is important to extend to stronger weakening processes, for example, based on the flash heating mechanism discussed earlier which seems likely to be important at seismic slip rates.

Short Slip Duration

Another feature of the constitutive relation and overall fault model concerns their prediction of the duration of dynamic slip. Seismological observations demonstrate that slip duration is relatively short compared to the duration of the whole rupture propagation, and a mechanism has to be identified to explain the healing of slip. Two different interpretations have been recently proposed: one associates the healing of slip with strong heterogeneity of stress or strength on the fault plane (Beroza and Mikumo 1996; Day et al. 1998), whereas the other associates healing with strong rate-weakening in the constitutive relation (Cochard and Madariaga 1996; Beeler and Tullis 1996; Zheng and Rice 1998; Nielsen and Carlson 2000). These two mechanisms yield different traction evolutions, as shown in Figure 5.6. If the healing of slip is caused by the constitutive relation a fast restrengthening occurs immediately after the dynamic weakening. On the contrary, if strength or stress heterogeneity controls slip duration, the stress

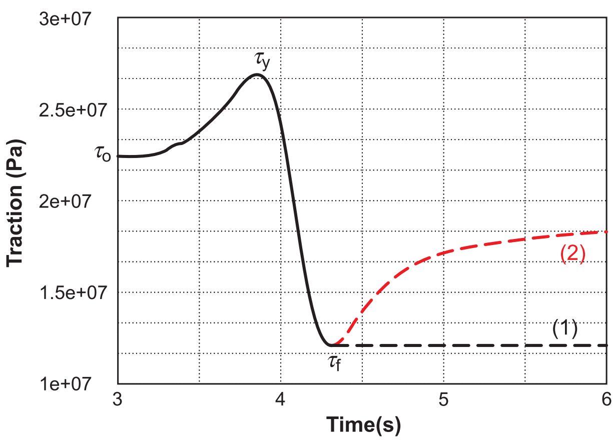


Figure 5.6 Sketch showing the traction evolution as a function of time in a generic point on the fault: dynamic traction increases from its initial value (τ_o) to the upper yield stress (τ_y) and therefore drops to the kinetic friction level (τ_f). The two subsequent evolutions refer to case (1), when slip occurs at a constant kinetic friction level, or to case (2) which is characterized by a fast restrengthening causing the healing of slip. In these two configurations, the final traction when slip is healed is different.

seems to remain near to or somewhat below (dynamic overshoot) the kinetic friction level. Effects of different elastic properties on the two sides of the fault plane can contribute to healing (Weertman 1980; Andrews and Ben-Zion 1997). Yet another possibility is that abrupt cessation of a weakening mechanism, for example, by hydraulic cracking of thermally pressurized fluid into the fault walls (Sibson 1973) would induce healing. There is evidence in lab studies (Tsutsumi and Shimamoto 1997; Hirose and Shimamoto 2005) that frictional resistance might increase abruptly in association with the earliest phases of melting and formation of a continuous melt layer; that too is a possible basis for healing, provided that it is effective before a broader, hotter and weaker melt layer can develop (Fialko 2004). Understanding the mechanisms controlling slip duration and estimating its value during real earthquakes is relevant to estimating frictional heating and radiated seismic energy, and therefore to the earthquake energy balance.

Slip Inversions and Rupture Parameters

Seismological observations and the modeling of ground motion waveforms recorded during large magnitude earthquakes allow the imaging of the slip-time history and its distribution on the fault plane. Numerous papers show that slip and rupture time distribution on the fault plane are heterogeneous, thus supporting the complexity of the processes controlling the mechanics of faulting and earthquakes. The availability of kinematic slip models (see Mai 2004) obtained by inverting geophysical data and by fitting observations makes feasible the estimate of the dynamic parameters strength excess ($\tau_y - \tau_o$), dynamic stress drop ($\tau_o - \tau_f$) or strength breakdown ($\tau_y - \tau_f$), and critical slip-weakening distance D_c (Ide and Takeo 1997; Guatteri and Spudich 2000; Piatanesi et al. 2004). Seismological estimates of D_c yield large values of this parameter (of the order of 0.1–1 m) and suggest that it is a large fraction (up to 80%) of the total slip during the earthquake (Dalguer et al. 2002, among others). That might reflect a multiscale weakening process in which weakening continues at an ever decreasing rate with respect to slip, out to large slip (Abercrombie and Rice 2005), a feature also found in some thermal pressurization models (Rice 2004, 2006; Figures 5.2 and 5.4). These findings raise important issues which affect the estimate of fracture energy and the earthquake energy balance. The first concerns the problem of bridging laboratory and seismological estimates of length scale parameters. This is equivalent to the problem of interrelating microscale and macroscale processes controlling dynamic fault weakening. The second issue concerns the definition of fracture energy or the understanding of the way in which the work done during sliding is spent, including the much neglected possibility of dissipation in inelastic deformation of highly stressed material in the damage zone. Slip inversions for large earthquakes often lead to the inference of supershear rupture propagation (e.g., 1999 Izmit, Turkey; 2001 Kunlun, Tibet; 2002 Denali, Alaska), a phenomenon also seen in laboratory studies

(Rosakis 2002; Xia et al. 2004), and according to existing theoretical understanding of the conditions for transition to that regime (Andrews 1976b; Dunham et al. 2003), those observations should place some at least loose constraints on fault stressing and strength at the times of the events.

DISCUSSION

To face the problems of comprehensively describing earthquake phenomena, we have emphasized here the critical need for new types of laboratory and natural fault observations, together with theory, for moving the conceptual background beyond what is now available. This is needed to address, for example, at short timescales, the weakening during rapid, large slips of significant seismic events as well as, at long timescales, the interseismic restrengthening of fault zone materials, including the effects of temperature and fluid reactions, and consequences for subsequent earthquake nucleation and (to go back to short timescales) reinitiation of failure at a propagating rupture tip. The present common formulations of R&S laws have represented a milestone advance for the field, but they have been proposed to interpret the results of laboratory experiments at low slip velocity (e.g., velocity stepping experiments usually with $V < 1$ mm s^{-1}), and usually under constant normal stress. Moreover, different interpretations exist in the literature concerning the state variable and its evolution with slip and time, all of which have only the barest basic physical, versus empirical, foundations. The two most common interpretations rely on considering the state variable as representative of the evolution of properties of the micro-asperity contacts during sliding (Dieterich 1986), or of the granular packing density within the shear zone (Sleep and Blanpied 1992; Sleep 1997; Sleep et al. 2000). Both of these are expected to change during dynamic slip.

An important implication for future research is the understanding of the different temporal evolutions of those and other processes (like flash heating, thermal pressurization, gouge gelation, and local and macroscale melting), and of where, in the complex fault zone model here discussed (Figure 5.1), and under what conditions, those processes occur. Along with the necessity for a new generation of laboratory experiments—some performed at high slip rates and under widely variable normal stress conditions, some under long hold times at temperature—it will be most convincing to use, in the laboratory, rock samples taken from fault cores from depth, through drilling programs, as well as from exhumed faults. The properties of natural fault zones, while in need of further elucidation, seem from recent studies (Chester et al. 1993; Chester and Chester 1998; Wibberley 2002; Ben-Zion and Sammis 2003; Chester and Goldsby 2003; Chester et al. 2003; Otsuki et al. 2003; Sibson 2003; Wibberley and Shimamoto 2003; Sulem et al. 2004) to be more complex than usually believed. We should account for that not just in choice of laboratory materials, but certainly also in our theoretical and numerical interpretations.

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REFERENCES

Abercrombie, R.E., and J.R. Rice. 2005. Can observations of earthquake scaling constrain slip weakening? *Geophys. J. Intl.* **162**:406–424, doi: 10.1111/j.1365-246X.2005.02579.x.

Andrews, D.J. 1976a. Rupture propagation with finite stress in antiplane strain. *J. Geophys. Res.* **81**:3575–3582.

Andrews, D.J. 1976b. Rupture velocity of plane strain shear cracks. *J. Geophys. Res.* **81**:5679–5687.

Andrews, D.J. 2002. A fault constitutive relation accounting for thermal pressurization of pore fluid. *J. Geophys. Res.* **107**:B12, 2363, doi: 10.1029/2002JB001942.

Andrews, D.J. 2005. Rupture dynamics with energy loss outside the slip zone. *J. Geophys. Res.* **110**:B1, B01307, doi: 10.1029/2004JB003191.

Andrews, D.J., and Y. Ben–Zion. 1997. Wrinkle-like slip pulse on a fault between different materials. *J. Geophys. Res.* **102**:B1, 553–571.

Barenblatt, G.I. 1959. The formation of brittle cracks during brittle fracture. General ideas and hypotheses. Axially-symmetric cracks. *Appl. Math. Mech.* **23**:1273–1282.

Beeler, N.M., and T.E. Tullis. 1996. Self-healing slip pulses in dynamic rupture models due to velocity-dependent strength. *Bull. Seismol. Soc. Am.* **86**:1130–1148.

Beeler, N.M., and T.E. Tullis. 2003. Constitutive relationships for fault strength due to flash-heating. In: 2003 SCEC Annual Meeting Proceedings and Abstracts, vol. 13, p. 66. Los Angeles: Southern California Earthquake Center, Univ. of Southern California.

Beeler, N.M., T.E. Tullis, M.L. Blanpied, and J.D. Weeks. 1996. Frictional behavior of large displacement experimental faults. *J. Geophys. Res.* **101**:B4, 8697–8715.

Beeler, N.M., T.E. Tullis, and D.L. Goldsby. 2006. Constitutive relationships and physical basis of fault strength due to flash-heating. *J. Geophys. Res.*, in press.

Beeler, N.M., T.E. Tullis, and J.D. Weeks. 1994. The roles of time and displacement in the evolution effect in rock friction. *Geophys. Res. Lett.* **21**:1987–1990.

Ben-Zion, Y., and C.G. Sammis. 2003. Characterization of fault zones. *Pure & Appl. Geophys.* **160**:677–715.

Beroza, G., and T. Mikumo. 1996. Short slip duration in dynamic rupture in the presence of heterogeneous fault properties. *J. Geophys. Res.* **101**:22,449–22,460.

Biegel, R.L., and C.G. Sammis. 2004. Relating fault mechanics to fault zone structure. *Adv. Geophys.* **47**:65–111.

Bizzarri, A., and M. Cocco. 2003. Slip–weakening behavior during the propagation of dynamic ruptures obeying to rate- and state-dependent friction laws. *J. Geophys. Res.* **108**:B8, 2373, doi: 10.1029/2002JB002198.

Bizzarri, A., and M. Cocco. 2006a. A thermal pressurization model for the spontaneous dynamic rupture propagation on a 3D fault. Part I: Methodological approach. *J. Geophys. Res.* ***111:*** B05303, doi:10.1029/2005JB003862.

Bizzarri, A., and M. Cocco. 2006b. A thermal pressurization model for the spontaneous dynamic rupture propagation on a 3D fault. Part II: Traction evolution and dynamic parameters. *J. Geophys. Res.* **111:** B05304, doi:10.1029/2005JB003864.

Bizzarri, A., M. Cocco, D.J. Andrews, and E. Boschi. 2001. Solving the dynamic rupture problem with different numerical approaches and constitutive laws. *Geophys. J. Intl.* **144**:656–678.

Blanpied, M.L., D.A. Lockner, and J.D. Byerlee. 1991. Fault stability inferred from granite sliding experiments at hydrothermal conditions. *Geophys. Res. Lett.* **18**:609–612.

Blanpied, M.L., D.A. Lockner, and J.D. Byerlee. 1995. Frictional slip of granite at hydrothermal conditions. *J. Geophys. Res.* **100**:B7, 13,045–13,064.

Blanpied, M.L., C.J. Marone, D.A. Lockner, J.D. Byerlee, and D.P. King. 1998. Quantitative measure of the variation in fault rheology due to fluid-rock interactions. *J. Geophys. Res.* **103**:B5, 9691–9712.

Boatwright, J., and M. Cocco. 1996. Frictional constraints on crustal faulting. *J. Geophys. Res.* **101**:B6, 13,895–13,909

Boettcher, M.S., and C. Marone. 2004. Effects of normal stress variation on the strength and stability of creeping faults. *J. Geophys. Res.* **109**:B3, B03406.

Brace, W.F., and J.D. Byerlee. 1970. California earthquakes: Why only shallow focus? *Science* **168**:1573–1575.

Brodsky, E.E., and H. Kanamori. 2001. Elastohydrodynamic lubrication of faults. *J. Geophys. Res.* **106**:B8, 16,357–16,374.

Chester, F.M., and J.S. Chester. 1998. Ultracataclasite structure and friction processes of the Punchbowl fault, San Andreas system, California. *Tectonophysics* **295**:199–221.

Chester, F.M., J.P. Evans, and R.L. Biegel. 1993. Internal structure and weakening mechanisms of the San Andreas fault. *J. Geophys. Res.* **98**:771–786.

Chester, J.S., F.M. Chester, and A.K. Kronenberg. 2005. Fracture surface energy of the Punchbowl fault, San Andreas system. *Nature* **437**:133–136, doi: 10.1038/nature03942.

Chester, J.S., and D.L. Goldsby. 2003. Microscale characterization of natural and experimental slip surfaces relevant to earthquake mechanics. In: 2003 Annual Progress Report to the Southern California Earthquake Center. College Station: Texas A & M University.

Chester, J.S., A.K. Kronenberg, F.M. Chester, and R.N. Guillemette. 2003. Characterization of natural slip surfaces relevant to earthquake mechanics. *EOS Trans. AGU* **84**: Fall Meeting Suppl., abstract S42C-0185.

Cochard, A., and R. Madariaga. 1996. Complexity of seismicity due to highly rate-dependent friction. *J. Geophys. Res.* **101**:B11, 25,321–25,336.

Cochard, A., and J.R. Rice. 2000. Fault rupture between dissimilar materials: Ill-posedness, regularization, and slip pulse response. *J. Geophys. Res.* **105**:891–907.

Cocco, M., and A. Bizzarri. 2002. On the slip-weakening behavior of rate- and state-dependent constitutive laws. *Geophys. Res. Lett.* **29**:1–4.

Cocco, M. and A. Bizzarri. 2004. Dynamic fault weakening caused by thermal pressurization in an earthquake model governed by rate- and state-dependent friction. *EOS Trans. AGU* **85**: Fall Meeting Suppl., Abstract T22A-06.

Cocco, M., and J.R. Rice. 2002. Pore pressure and poroelasticity effects in Coulomb stress analysis of earthquake interactions. *J. Geophys. Res.* **107**:B2, 2030, doi:10.1029/2000JB000138.

Dalguer, L.A., K. Irikura, W. Zhang, and J.D. Riera. 2002. Distribution of dynamic and static stress changes during 2000 Tottori (Japan) earthquake: Brief interpretation of the earthquake sequences; foreshocks, mainshock and aftershocks. *Geophys. Res. Lett.* **29**: doi: 10.1029/2001GL014333.

Day, S.M., G. Yu, and D.J. Wald. 1998. Dynamic stress changes during earthquake rupture. *Bull. Seismol. Soc. Am.* **88**:512–522.

Desrues, J., and G. Viggiani. 2004. Strain localization in sand: An overview of the experimental results obtained in Grenoble using stereophotogrammetry. *Intl. J. Numer. Anal. Meth. Geomech.* **28**:279–321.

Dieterich, J.H. 1978. Time-dependent friction and the mechanics of stick slip. *Pure & Appl. Geophys.* **11**:790–806.

Dieterich, J.H. 1979. Modeling of rock friction. 1. Experimental results and constitutive equations. *J. Geophys. Res.* **84**:2161–2168.

Dieterich, J.H. 1986. A model for the nucleation of earthquake slip. In: Earthquake Source Mechanics, ed. S. Das, J. Boatwright, and C.H. Scholz, Geophys. Monograph vol. 37, Maurice Ewing Ser. no. 6, pp. 37–47. Washington D.C.: Am. Geophys. Union.

Dieterich, J.H. 1994. A constitutive law for rate of earthquake production and its application to earthquake clustering. *J. Geophys. Res.* **99**:2601–2618.

Dieterich, J.H., and G. Conrad. 1984. Effect of humidity on time-dependent and velocity-dependent friction in rocks. *J. Geophys. Res.* **89**:B6, 4196–4202.

Dieterich, J.H., and B.D. Kilgore. 1994. Direct observation of frictional contacts: New insights for state-dependent properties. *Pure & Appl. Geophys*. **143**:283–302.

Dieterich, J.H., and B.D. Kilgore. 1996. Imaging surface contacts: Power law contact distributions and contact stresses in quartz, calcite, glass and acrylic plastic. *Tectonophysics* **256**:219–239.

Di Toro, G., D.L. Golbsby, and T.E. Tullis. 2004. Friction falls toward zero in quartz rock as slip velocity approaches seismic rates. *Nature* **427**:436–439.

Di Toro, G., T. Hirose, S. Nielsen, G. Pennacchioni, and T. Shimamoto. 2006. Natural and experimental evidence of melt lubrication of faults during earthquakes. *Science* **311**:647–649.

Di Toro, G., S. Nielsen, and G. Pennacchioni. 2005. Earthquake rupture dynamics frozen in exhumed ancient faults. *Nature* **436**:1009–1012.

Dunham, E.M., P. Favreau, and J.M. Carlson. 2003. A supershear transition mechanism for cracks. *Science* **299**:1557–1559.

Fialko, Y.A. 2004. Temperature fields generated by the elastodynamic propagation of shear cracks in the Earth. *J. Geophys. Res.* **109**:1–14, art. no. B01303, doi: 10.1029/2003JB002497.

Frye, K.M., and C. Marone. 2002. Effect of humidity on granular friction at room temperature. *J. Geophys. Res.* **107**:B11, 2309, doi: 10.1029/2001JB000654.

Goldsby, D.L., and T.E. Tullis. 2002. Low frictional strength of quartz rocks at subseismic slip rates. *Geophys. Res. Lett.* **29**:1844.

Goldsby, D.L., and T.E. Tullis. 2003. Flash heating/melting phenomena for crustal rocks at (nearly) seismic slip rates. In: 2003 SCEC Annual Meeting Proceedings and Abstracts, vol. 13, pp. 88–90. Los Angeles: Southern California Earthquake Center, Univ. of Southern California.

Guatteri, M., and P. Spudich. 2000. What can strong-motion data tell us about slip-weakening fault-friction laws? *Bull. Seismol. Soc. Am.* **90**:98–116.

Handy, M.R., and J.-P. Brun. 2004. Seismicity, structure, and strength of the continental lithosphere. *Earth Planet. Sci. Lett.* **223**:427–441.

Harris, R., and S. Day. 1997. Effects of a low velocity zone on a dynamic rupture. *Bull. Seismol. Soc. Am.* **87**:1267–1280.

Heaton, T.H. 1990. Evidence for and implications of self-healing pulses of slip in earthquake rupture. *Phys. Earth & Planet. Inter.* **64**:1–20.

Hirose, T., and T. Shimamoto. 2005. Growth of a molten zone as a mechanism of slip weakening of simulated faults in gabbro during frictional melting. *J. Geophys. Res.* **110**:B05202, doi:10.1029/2004JB003207.

Ida, Y. 1972. Cohesive force across the tip of a longitudinal-shear crack and Griffith's specific surface energy. *J. Geophys. Res.* **77**:3796–3805.

Ide, S., and M. Takeo. 1997. Determination of constitutive relations of fault slip based on seismic wave analysis. *J. Geophys. Res.* **102**:B12, 27,379–27,391.

Kame, N., J.R. Rice, and R. Dmowska. 2003. Effects of pre-stress state and rupture velocity on dynamic fault branching. *J. Geophys. Res.* **108**:B5, 2265, doi:10.1029/2002JB002189.

Kanamori, H., and E.E. Brodsky. 2001. The physics of earthquakes. *Phys. Today* **54**:34–39.

Lachenbruch, A.H. 1980. Frictional heating, fluid pressure, and the resistance to fault motion. *J. Geophys. Res.* **85**:B11, 6097–6122.

Lapusta, N., and J.R. Rice. 2003. Nucleation and early seismic propagation of small and large events in a crustal earthquake model. *J. Geophys. Res.* **108**:B4, 2205, doi:10.1029/2001JB000793.

Lapusta, N., J.R. Rice, Y. Ben-Zion, and G. Zheng. 2000. Elastodynamic analysis for slow tectonic loading with spontaneous rupture episodes on faults with rate- and state-dependent friction. *J. Geophys. Res.* **105**:23,765–23,789.

Li, Y.G., J. Vidale, K. Aki, C. Marone, and W.K. Lee. 1994. Fine structure of the Landers fault zone: Segmentation and the rupture process. *Science* **265**:367–380.

Linker, M.F., and J.H. Dieterich. 1992. Effects of variable normal stress on rock friction: Observations and constitutive equations. *J. Geophys. Res.* **97**:B4, 4923–4940.

Lockner, D., H. Naka, H. Tanaka, H. Ito, and R. Ikeda. 2000. Permeability and strength of core samples from the Nojima fault of the 1995 Kobe earthquake. In: Proc. Intl. Workshop on the Nojima Fault Core and Borehole Data Analysis, Tsukuba, Japan, Nov. 22–23, 1999, ed. H. Ito, H. Fujimoto, H. Tanaka, and D. Lockner, Open File Report 00-129, pp. 147–152. Menlo Park, CA: U.S. Geol. Survey.

Mai, M. 2004. Database of finite-source rupture models. http://www.seismo.ethz.ch/srcmod/, ETH, Zurich.

Marone, C. 1998. Laboratory-derived friction laws and their application to seismic faulting. *Ann. Rev. Earth & Planet. Sci.* **26**:643–696.

Marone, C., C.B. Raleigh, and C.H. Scholz. 1990. Frictional behavior and constitutive modeling of simulated fault gouge. *J. Geophys. Res.* **95**:B5, 7007–7025.

Mase, C.W., and L. Smith. 1985. Pore-fluid pressures and frictional heating on a fault surface. *Pure & Appl. Geophys.* **92**:6249–6272.

Mase, C.W., and L. Smith. 1987. Effects of frictional heating on the thermal, hydrologic, and mechanical response of a fault. *J. Geophys. Res.* **92**:B7, 6249–6272.

Morgenstern, N.R., and J.S. Tschalenko. 1967. Microscopic structures in kaolin subjected to direct shear. *Géotechnique* **17**:309–328.

Nielsen, S.B., and J.M. Carlson. 2000. Rupture pulse characterization; self-healing, self-similar, expanding solutions in a continuum model of fault dynamics. *Bull. Seismol. Soc. Am.* **90**:1480–1497.

Ohnaka, M., and T. Yamashita. 1989. A cohesive zone model for dynamic shear faulting based on experimentally inferred constitutive relation and strong motion source parameters. *J. Geophys. Res.* **94**:4089–4104.

Ohnaka, M.A. 2003. A constitutive scaling law and a unified comprehension for frictional slip failure, shear fracture of intact rock, and earthquake rupture. *J. Geophys. Res.* **108**:B2, 2080, doi: 10.1029/2000JB000123.

Otsuki, K., N. Monzawa, and T. Nagase. 2003. Fluidization and melting of fault gouge during seismic slip: Identification in the Nojima fault zone and implications for focal earthquake mechanisms. *J. Geophys. Res.* **108**:B4, 2192, doi: 10.1029/2001JB001711.

Palmer, A.C., and J.R. Rice. 1973. The growth of slip surfaces in the progressive failure of over-consolidated clay. *Proc. R. Soc. Lond. A* **332**:527–548.

Piatanesi, A., E. Tinti, M. Cocco, and E. Fukuyama. 2004. The dependence of traction evolution on the earthquake source time function adopted in kinematic rupture models. *Geophys. Res. Lett.* 31:L04609, doi: 10.1029/2003GL019225.

Poliakov, A.N.B., R. Dmowska, and J.R. Rice. 2002. Dynamic shear rupture interactions with fault bends and off-axis secondary faulting. *J. Geophys. Res.* **107**:B11, 2295, doi: 10.1029/2001JB000572.

Power, W.L., and T.E. Tullis. 1991. Euclidean and fractal models for the description of rock surface roughness. *J. Geophys. Res.* **96**:415–424.

Prakash, V. 1998. Frictional response of sliding interfaces subjected to time varying normal pressure. *J. Tribology* **120**:97–102.

Prakash, V. 2004. Pilot studies to determine the feasibility of using new experimental techniques to measure sliding resistance at seismic slip rates. In: 2004 Annual Progress Report to the Southern California Earthquake Center. Cleveland, OH: Case-Western Reserve Univ.

Prakash, V., and R.J. Clifton. 1992. Pressure-shear plate impact measurement of dynamic friction for high speed machining applications. In: Proc. VII Intl. Congress on Experimental Mechanics, pp. 556–564. Bethel, CT: Soc. Experimental Mechanics.

Ranjith, K., and J.R. Rice. 2001. Slip dynamics at an interface between dissimilar materials. *J. Mech. Phys. Solids* **49**:341–361.

Rice, J.R. 1993. Spatio-temporal complexity of slip on a fault. *J. Geophys. Res.* **98**: 9885–9907.

Rice, J.R. 1999. Flash heating at asperity contacts and rate-dependent friction. *EOS Trans. Amer. Geophys. Union* **80**: Fall Meeting Suppl., F681l.

Rice, J.R. 2004. Thermal weakening in large seismic slips and effects on rupture dynamics. In: 2004 Annual Progress Report to the Southern California Earthquake Center. Cambridge, MA: Harvard University.

Rice, J.R. 2006. Heating and weakening of faults during earthquake slip. *J. Geophys. Res.* **111**: B05311, doi:10.1029/2005JB004006.

Rice, J.R., and Y. Ben-Zion. 1996. Slip complexity in earthquake fault models. *PNAS* **93**:3811–3818.

Rice, J.R., N. Lapusta, and K. Ranjith. 2001. Rate and state dependent friction and the stability of sliding between elastically deformable solids. *J. Mech. Phys. Solids* **49**:1865–1898.

Rice, J.R., C.G. Sammis, and R. Parsons. 2005. Off-fault secondary failure induced by a dynamic slip-pulse. *Bull. Seismol. Soc. Am.* **95**:109–134, doi: 10.1785/0120030166.
Roig Silva, C., D.L. Goldsby, G. Di Toro and T.E. Tullis 2004.The role of silica content in dynamic fault weakening due to gel lubrication. *EOS Trans. AGU* **85**: Fall Meeting Suppl., Abstract T21D-06.
Rosakis, A.J. 2002. Intersonic shear cracks and fault ruptures. *Adv. Phys.* **51**:1189–1257.
Ruina, A.L. 1983. Slip instability and state variable friction laws. *J. Geophys. Res.* **88**:B12, 10,359–10,370.
Scholz, C.H. 1990. The Mechanics of Earthquake and Faulting. New York: Cambridge Univ. Press.
Scholz, C.H. 1998. Earthquakes and friction laws. *Nature* **391**:37–41.
Scruggs, V.J., and T.E. Tullis. 1998. Correlation between velocity dependence of friction and strain localization in large displacement experiments on feldspar, muscovite, and biotite gouge. *Tectonophysics* **295**:15–40.
Segall, P., and J.R. Rice. 1995. Dilatancy, compaction, and slip instability of a fluid-infiltrated fault. *J. Geophys. Res.* **100**:22,155–22,171.
Sibson, R.H. 1973. Interaction between temperature and pore-fluid pressure during earthquake faulting: A mechanism for partial or total stress relief. *Nature* **243**:66–68.
Sibson, R.H. 1975. Generation of pseudotachylite by ancient seismic faulting. *Geophys. J. Roy. Astr. Soc.* **43**:775–794.
Sibson, R.H. 2003. Thickness of the seismic slip zone. *Bull. Seismol. Soc. Am.* **93**: 1169–1178.
Sleep, N.H. 1997. Application of a unified rate and state friction theory to the mechanics of fault zones with strain localization. *J. Geophys. Res.* **102**:B2, 2875–2895.
Sleep, N.H., and M.L. Blanpied. 1992. Creep, compaction, and the weak rheology of major faults. *Nature* **359**:687–692.
Sleep, N.H., E. Richardson, and C. Marone. 2000. Physics of friction and strain rate localization in synthetic gouge. *J. Geophys. Res.* **105**:25,875–25,890.
Spray, J.G. 1993. Viscosity determinations of some frictionally generated silicate melts: Implications for fault zone rheology at high strain rates. *J. Geophys. Res.* **98**: 8053–8068.
Steacy, S.J. and C.G. Sammis. 1991. An automaton for fractal patterns of fragmentation. *Nature* **353**:250–252.
Stesky, R.M. 1978. Rock friction effect of confining pressure, temperature, and pore pressure. *Pure & Appl. Geophys.* **116**:690–704.
Stesky R.M., W.F. Brace, D.K. Riley, and P.Y.F. Robin. 1974. Friction in faulted rock at high-temperature and pressure. *Tectonophysics* **23**:177–203.
Stirling, M.W., S.G. Wesnousky, and K. Shimazaki. 1996. Fault trace complexity, cumulative slip, and the shape of the magnitude-frequency distribution for strike-slip faults: A global survey. *Geophys. J. Intl.* **124**:833–868.
Sulem, J., I. Vardoulakis, H. Ouffroukh, M. Boulon, and J. Hans. 2004. Experimental characterization of the thermo-poro-mechanical properties of the Aegion fault gouge. *C. R. Geoscience* **336**:455–466.
Tinti, E., P. Spudich, and M. Cocco. 2005. Earthquake fracture energy inferred from kinematic rupture models on extended faults. *J. Geophys. Res.* **110**:B12, B12303, doi: 10.1029/2005JB003644.
Tse, S.T., and J.R. Rice. 1986. Crustal earthquake instability in relation to the depth variation of frictional slip properties. *J. Geophys. Res.* **91**:9452–9472.

Tsutsumi, A., and T. Shimamoto. 1997. High velocity frictional properties of gabbro. *Geophys. Res. Lett.* **24**:699–702.

Tullis, T.E., ed. 1986. Friction and faulting, topical issue. *Pure & Appl. Geophys.* **124**.

Tullis, T.E., and D.L. Goldsby. 2003. Laboratory experiments on fault shear resistance relevant to coseismic earthquake slip. In: 2003 Annual Progress Report to the Southern California Earthquake Center. Providence, RI: Brown Univ.

Vardoulakis, I. 2003. Dynamic thermo-poro-mechanical analysis of catastrophic landslides. *Géotechnique* **53**:523–524.

Weertman, J. 1980. Unstable slippage across a fault that separates elastic media of different elastic constants. *J. Geophys. Res*. **85**:1455–1461.

Wibberley, C.A.J. 2002. Hydraulic diffusivity of fault gouge zones and implications for thermal pressurization during seismic slip. *Earth Planets Space* **54**:1153–1171.

Wibberley, C.A.J., and T. Shimamoto. 2003. Internal structure and permeability of major strike-slip fault zones: The Median Tectonic Line in Mid Prefecture, Southwest Japan. *J. Struct. Geol.* **25**:59–78.

Wilson, B., T. Dewers, Z. Reches, and J. Brune. 2005. Particle size and energetics of gouge from earthquake rupture zones. *Nature* **434**:749–752.

Xia, K., A.J. Rosakis, and H. Kanamori. 2004. Laboratory earthquakes: The sub-Rayleigh-to-supershear transition. *Science* **303**:1859–1861.

Xia, K., A.J. Rosakis, H. Kanamori, and J.R. Rice. 2005. Inhomogeneous faults hosting earthquakes in the laboratory: Directionality and supershear. *Science* **308**:681–684.

Zheng, G., and J.R. Rice. 1998. Conditions under which velocity-weakening friction allows a self-healing versus a cracklike mode of rupture. *Bull. Seismol. Soc. Am.* **88**:1466–1483.