Heating, weakening and shear localization in earthquake rupture

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Heating, Weakening and Shear Localization in Earthquake Rupture


Abstract: Field and borehole observations of active earthquake fault zones show that shear is often localized to principal deforming zones of order 0.1 mm to 10 mm width. The presentation addresses how frictional heating in rapid slip weakens faults dramatically, relative to their static frictional strength, and promotes such intense localization. Pronounced weakening occurs even on dry rock-on-rock surfaces, due to flash heating effects, at slip rates above ~0.1 m/s (earthquake slip rates are typically of order 1 m/s). But weakening in rapid shear is also predicted theoretically in thick fault gouge in the presence of fluids (whether native ground fluids or volatiles such as H2O or CO2 released by thermal decomposition reactions), and the predicted localizations are compatible with such narrow shear zones as have been observed. The underlying concepts show how fault zone materials with high static friction coefficients, ~0.6 to 0.8, can undergo strongly localized shear at effective dynamic friction coefficients of order 0.1, thus fitting observational constraints, e.g., of earthquakes producing negligible surface heat out-flow and, for shallow events, only rarely creating extensive melt. The results to be summarized include those of collaborative research published with Nicolas Brantut (Univ. Col. Lond.), Eric Dunham (Stanford Univ.), Nadia Lapusta (Caltech), Hiroyuki Noda (JAMSTEC, Japan), John D. Platt (Carnegie Inst. Sci., now at *gram Labs), Alan Rempel (Oregon State Univ.), and John W. Rudnicki (Northwestern Univ.).

Introduction:

A long-standing quandary in seismology is that the friction coefficient \( f \) for typical fault rocks, as estimated from both lab (Byerlee [1978]) and field (Zoback and Townend [2001], Zoback et al. [2002]) studies, is relatively high, e.g., \( f \sim 0.6 \) to 0.8. Here \( f \) appears in a representation of fault shear strength \( \tau \) as \( \tau = f \times (\sigma_n - p) \), where \( \sigma_n \) is the normal stress clamping the fault shut, and \( p \) is the pore pressure in an infiltrating fluid phase along the contacting surfaces, that phase typically being groundwater in the crust.

However, zones of significant slip along mature faults are often remarkably thin. E.g., Chester and Chester [1998] reported that a zone of sub-cm thickness, which they called a “persistent slip zone”, along the Punchbowl Fault, had apparently accommodated km-scale slips, presumably in many individual earthquakes, with high localization of slip (that despite apparently wide fault-bordering damage zones of, say, 1 m breadth or larger). A subsequent thin-section study of a part of that fault zone by Chester and Goldsby [2003] suggested that the most intense straining was localized to a remarkably thin zone of \( \sim 0.1 \) - 0.3 mm thickness (see also the review by Chester et al. [2004] and Fig. 1 of Rice [2006]). Around the same time Heermance et al. [2003] reported an analysis of borehole core retrieved across the Chelungpu fault, which hosted the 1999
Mw 7.6 Chi-Chi, Taiwan, earthquake, and concluded that slip, at a 328 m depth traverse of the rupture by the borehole, was accommodated within an, again, remarkably thin zone, of ~ 0.05 - 0.3 mm width. Later core studies by Boullier et al. [2009, 2011], in differently located borehole traverses of that Chi-Chi rupture, at 1136 m and 1111 m depths, found thin, if not so hyper-thin, shear zones of ~3 mm and ~20 mm widths, respectively.

Such thinness of earthquake slip zones implies that if those typical $f \sim 0.6$ to 0.8 values prevail during seismic slip, and if $p$ is ~hydrostatic, then $\tau$ is large enough that we should find measurable heat outflow near major faults, and/or extensive melt signatures along exhumed faults. But neither effect is generally found, although thin pseudotachylite layers along faults are indeed sometimes reported. E.g., to quote from a helpful anonymous review of the original version of this paper, “… moderate to large in size crustal earthquakes nucleate at 7-25 km in depth and seismic inversion studies suggest that the maximum slip is often localized in fault patches located at several km depth (e.g., 6-8 km depth for the L’Aquila Mw 6.3 2009 earthquake, Cirella et al. [2016]). Several field studies and their comparison with experimental studies (Fondrist et al. [2013]; Siman-Tov et al. [2013]; De Paola et al. [2015]; Demurtas et al. [2016], etc.), present evidence of extreme strain localization during seismic slip in carbonate-bearing faults exhumed from 2-4 km depth. This would extend the evidence of the presence of extremely thin slipping zones to [a] deeper level in the continental crust. The occurrence of such extremely thin slipping zones could be further extended to 60 km depth by considering the thin pseudotachylyte-bearing faults veins hosted in crustal and mantle rocks [Swanson, 1988; Austrheim and Boundy, 1994; Di Toro et al., 2005; Ueda et al., 2008].”

That (mostly) negative result concerning extensive melts has motivated a sequence of studies to understand the underlying physics by many in the earthquake science community, including studies by the author and his co-workers, as summarized here. Those began with Rice and Cocco [2007] (presented 2005), and continued with Rice [2006], Rempel and Rice [2006], and in later studies with J. W. Rudnicki, J. Platt, and N. Brantut (Rice et al. [2014], Platt et al. [2014, 2015]).

It seems at first puzzling that such extreme localization of seismic slip evaded observation for so long. However, a plausible reason, consistent with a brief comment by Sibson [2003], was suggested by Rockwell and Ben-Zion [2007]. Based on rupture surfaces and strands observed from trenching across the Johnson Valley segment of the 1992 Landers earthquake, they noted that what is observed at Earth’s surface to be an ~2 m wide zone of significant distributed damage had gradually localized with increasing depth below surface into a single thin zone of highly concentrated shear, like those discussed above, accommodating the ~2.6 m earthquake slip at depths of ~3m (and presumably greater). Not being aware of the localization of significant shear with increasing depth, it would be easy to assume, as many (at least this writer) did, that the damage zone width as gauged at Earth’s surface marked the width that sheared and damaged significantly in the upper crust during individual earthquakes.
That noted, studies by Mitchell and Faulkner [2012] of crack density vs. distance from strike slip faults, particularly, those exposed by erosion in the Atacama fault system in northern Chile, show that there are significant microcrack densities, and associated permeability, developed in the fault-bordering zones; see their Fig. 2. Those densities and permeability diminish towards background level over ~20 m to ~150 m distance scales perpendicular to the fault (the larger the accumulated slip on the fault, 35 m to 5,000 m in the data set for their Fig. 2, the larger the distance over which that significant microcracking extends). So there are extensive fault-bordering damage zones. They need not contain long connected fault features, analogous to the fresh surface-breaking features mentioned above. However, over the time scales of evolution of major faults, the microcracks of those zones have, nevertheless, enough connectivity to notably enhance the near-fault permeability, at least outside the highly granulated, low-permeability, core zone along the rupture surface. Such is, e.g., consistent with permeability measured by Lockner at al. [2000] (their Fig. 4) as a function of distance from the Nojima Fault, which ruptured in the 1995 Kobe, Japan, earthquake.

While such is not the main focus in this brief report, it may be recalled that the stresses generated in the sidewalls of a rupturing fault, over distances up to a few 10s of meters from the rupture surface, may often exceed a Mohr-Coulomb threshold for fault rupture in that border region, e.g., Rice et al. [2005], Templeton and Rice [2008]. Such may possibly nucleate a subsidiary rupture (which may then propagate so as to “branch off” from the main fault). Further, when a propagating rupture along the fault on which it has been nucleated reaches an intersection with a branching fault structure, the rupture may stay on the main fault path, or take the branch, or continue on both, or perhaps arrest. These features have been discussed extensively, and enabling conditions for activation of rupture along an intersecting fault branch have been identified (Poliakov et al. [2002]; Kame et al. [2003]) and applied and further expanded upon for a variety of natural settings, e.g., by Bhat et al. [2004] to slip transfer from the Denali to Totschunda faults in the 2002 Denali, Alaska, earthquake; by Fliss et al. [2005] to branching and jumping between fault segments in the 1994 Landers, Mohave Desert area earthquake; and by Bhat et al. [2007] to off-fault damage patterns developed during the supershear rupture phase of the 2001 M_w 8.1 Kokoxili (Kunlun) Tibet earthquake.

Such results on creation and activation of off-fault damage, may provide a useful perspective on the long-term evolution of major fault systems. E.g., they are possibly relevant to such recent studies as by Perrin et al. [2016a,b] who note the development of “off-fault tip splay networks” which are preferentially placed relative to the main-fault long-term growth direction, such that the damaged/faulted border region is widest towards the oldest end of the main fault), and which may possibly be features which control the slip distribution in subsequent earthquakes (with slip generally inferred to be largest on the oldest part of the fault).

Mechanisms of heating and weakening

There are two major streams of the more recent understanding of heating and weakening of fault surfaces or fault zones. The first involves an adaptation to the seismological
context (Rice [1999, 2006]; Beeler et al. [2008]; Rice et al. [2009] of a known "flash heating" mechanism of weakening at high slip rates in metal on metal friction [Bowden and Thomas [1954]; Archard [1958/59], Ettles [1996]; Lim and Ashby [1997]. That is summarized in Fig. 1, and involves the rapid heating and weakening of frictional asperity contacts between sliding surfaces of, essentially, coherent rock as discussed in the next section.

The other major stream relies on the fact that shear zones are often zones of finely granulated material and, being (for the situations considered here) zones which lie within Earth's crust, it is frequently the case that such zones are fluid infiltrated. Thus we must be concerned with the thermal pressurization of native ground fluids by frictional dissipation [e.g., Sibson, 1973; Lachenbruch, 1981; Mase and Smith, 1985, 1987; Andrews, 2002; Vardoulakis, 2002; Sulem et al., 2005; Veveakis et al., 2007, 2012] raising $p$ towards $\sigma$, and hence weakening the shear strength $\tau$. The earlier of those studies represented the fault as a saturated granulated layer of uniform thickness, say, $h$, separating blocks of coherent rock that were either non-deforming and unable to conduct heat or pore fluid.

A more extreme version of that pressurization mechanism is caused when the heating raises temperature sufficiently to trigger thermal decomposition reactions, releasing highly pressurized volatiles such as $\text{H}_2\text{O}$ from clay mineral or serpentine fragments within the gouge, or $\text{CO}_2$ from carbonate fragments, e.g., [Sulem and Famin, 2009; Veveakis et al., 2012]. Those processes likewise promote extreme localization of shear [Platt et al., 2015].

### Flash heating and weakening of frictional asperity contacts

To address flash heating in rock-on-rock friction by a simplified model (see Rice [1999, 2006], Beeler et al. [2008], Rice et al. [2009]), consider Fig. 1. There $D$ is a representative asperity contact diameter, and asperity shear ($\tau_c$) and normal ($\sigma_c$) stresses are comparable to the maximum sustainable shear stress and to the normal stress at contact indentation of a punch of a yet-stronger material. Force equilibrium demands that $\tau_c / \sigma_c$ (if the terms are understood as a ratio of the averages of $\tau_c$ and $\sigma_c$ over all contacts) must equal the same ratio $\tau / \sigma$ based on the macroscopic average stresses, and hence be equal to the friction coefficient $f \sim 0.6-0.8$ during slip. Such reasoning, together with studies with hard indentors, led to the conclusion that the local shear stress $\tau_c$ resisting sliding at a typical rock-on-rock asperity contact is of order $\tau_c \sim 0.1 \mu$ where $\mu$ is the elastic shear rigidity. The asperities are assumed to remain strong, at $\tau_c \sim 0.1 \mu$, until the asperity contact, initially at ambient fault temperature $T_f$, reaches a "weakening temperature" $T_w$ (which would generally correspond to forming a melt layer at the contact, but might in some cases involve enhanced plasticity, e.g., by increased mobility of crystal dislocations or local phase transformations at elevated temperature). A simple heat transfer model [Rice, 1999, 2006] approximately estimates, by the expression shown
in Fig. 1, the sliding velocity $V_w$ such that the weakening temperature is just reached at the moment when the contact has been slid out of existence.

When $V < V_w$, there is no thermal weakening of the contact in its brief lifetime, whereas when $V > V_w$, the contact is weak for part of its lifetime. The resulting prediction of the form of the $f$ vs. $V$ relation (i.e., the form $f = \text{Const}_1 + \text{Const}_2 / V$, as shown in Fig. 1) has been found to give a reasonable description of laboratory friction measurements for a wide variety of rocks (Tullis and Goldsby [2003], Beeler et al. [2008]; Goldsby and Tullis [2011], Kohli et al. [2011], Proctor et al. [2014]). Proctor et al. [2014] also emphasize that to fit the model as outlined here to data involving substantial slip it is important to recognize that the appropriate ambient fault temperature $T_f$ to use in the model will not remain constant but, rather, will increase with accumulated slip from frictional warming of the fault zone and its environs. By correcting for that effect of a slip-history dependence of $T_f$ (note that the slip rate $V_w$ at which weakening initiates depends on $T_f$ and diminishes as $T_f$ rises), they obtained plausible agreement with the model of Fig. 1 not only during a phase of increasing imposed slip rate, but also during a subsequent phase of decreasing slip rate, a phase for which previous comparisons of the model to data had been less satisfactory.

**Thermal pressurization by frictional heating of native fluids in fault zone gouge**

Chester et al. [2005] and subsequently Kitajima et al. [2010] reported rotary shear friction experiments on the ultracataclasite gouge from the Punchbowl fault and concluded "That significant weakening is only observed at high rates and that the critical slip distance for weakening decreases with an increase in normal stress, imply that weakening is a thermally activated process. Moreover, slide-hold-slide tests show rapid strength recovery consistent with transient thermal effects." (The author understands that their phrase "weakening is a thermally activated process" in this context is meant to convey that rapid temperature rise and weakening co-occur, and not that the weakening processes are necessarily Arrhenius thermally activated rate processes -- although such activated rate processes are part of modern understanding of rate and state frictional processes (e.g., Rice et. Al. [2001]) operative at asperity contacts, at least in a low slip rate range prior to nucleation of unstable seismic slip.)

In a recent refinement of understanding of the thermal pressurization process, Rice [2006] devised a solution for slip on a mathematical plane, generalizing an analysis of Mase and Smith [1985, 1987] and considering the combined effects of thermal and fluid diffusion (whereas Mase and Smith considered only a single transport process in this particular aspect of their pioneering study, which did elsewhere address the combined effects). Other studies of thermal stability of deformation in fluid-infiltrated granular media have been contributed by Andrews [2002] and, while not specifically focused on seismic rupture, by Benallal and Comi [2003] and Benallal [2005].

Subsequently, Rice et al. [2014] showed that relatively homogeneous shear within a layer of fluid-saturated gouge, like considered in Lachenbruch [1981] and other earlier studies mentioned, is actually an unstable deformation mode, unless the gouge layer is thinner
than a certain computable value. That drives extreme localization of shear in such configurations, in a manner which Platt et al. [2014] showed to agree closely, at increasing slip, with the Rice [2006] solution for slip on a mathematical plane. That is illustrated here in Figure 2.

The governing equations assumed within the shearing layer, stated with reference to the panel at the upper right of Fig. 2, treat the layer as a saturated granular material offering frictional resistance to shear. Referring the reader to Rice et al. [2014] and Platt et al. [2014] for further details, we assume that two poroelastic half-spaces, assumed non-yielding, are forced to move relative to each other at a speed $V$ (estimated, from seismically inferred ratios of slip to slip duration, to be $V \sim 1$ m/s). All inelastic deformation is accommodated in the gouge layer, of thickness $h$, leading to an average (i.e., through the layer thickness) layer-parallel shear strain rate $\dot{\gamma}_o = V / h$ for the layer (the local strain rate $\dot{\gamma}$ in the layer, averaging to $\dot{\gamma}_o$, will vary with $y$ and with time $t$ from the start of the imposed shear motion). Then mechanical equilibrium (accelerations are negligible in this context), conservation of energy, and conservation of fluid mass within the layer lead, respectively, to

\[
\frac{\partial \tau}{\partial y} = 0 \quad \text{and} \quad \frac{\partial \sigma_n}{\partial y} = 0 \quad , \quad \frac{\partial T}{\partial t} - \alpha_h \frac{\partial^2 T}{\partial y^2} = \frac{\tau \dot{\gamma}}{\rho c} \quad , \quad \text{and} \quad \frac{\partial p}{\partial t} - \alpha_{hy} \frac{\partial^2 p}{\partial y^2} = \Lambda \frac{\partial T}{\partial t} , \tag{1}
\]

where

\[
\tau = f(\dot{\gamma})(\sigma_n - p) \quad \text{and} \quad f(\dot{\gamma}) = f_0 + (a - b) \log(\dot{\gamma} / \dot{\gamma}_o) \quad , \quad \text{with} \quad a - b \equiv \left( \frac{d f(\dot{\gamma})}{d \dot{\gamma}} \right)_{\dot{\gamma}=\dot{\gamma}_0} > 0 , \tag{2}
\]

thus describing a rate strengthening friction (often appropriate to higher temperature situations, like for rapidly sheared gouge) which, given the large strains to be experienced, is treated as being always at steady state corresponding to the momentary strain rate. Also, $\tau$ and $\sigma_n$ are the local shear and layer-perpendicular normal stress within the layer, $T$ is local temperature, $p$ is local pore pressure, $\alpha_{hy}$ and $\alpha_h$ are the respective hydraulic and thermal diffusivities in the layer, $\rho c$ is specific heat, and $\Lambda$ is a poro-thermo-mechanical parameter characterizing how temperature rise causes pore pressure rise under undrained conditions.

Figure 2 (see its caption for further details) shows an important set of solutions of these equations, verifying that they predict strong shear localization in gouge, so much so that at sufficiently large slip their solution agrees with the solution for slip on a mathematical plane between poro-thermo-elastic half-spaces.

### Thermal decomposition reactions releasing high-pressure volatile phases

Platt et al. [2015] have shown that this situation, mentioned earlier, can be modeled to an acceptable approximation, following J. Sulem and co-workers (e.g., Sulem and Famin [2009]; Sulem et al.[2009], Veveakis et al. [2012]), by introducing two thermo-chemical parameters, labeled $E_r$ and $P_r$, into the above set of equations, so that the energy and fluid mass conservation equations generalize to
\[
\frac{\partial T}{\partial t} - \alpha_{th} \frac{\partial^2 T}{\partial y^2} - E_r \frac{\partial \xi}{\partial t} = \frac{\tau \gamma}{\rho c} \quad \text{and} \quad \frac{\partial p}{\partial t} - \alpha_{hy} \frac{\partial^2 p}{\partial y^2} + P_r \frac{\partial \xi}{\partial t} = \Lambda \frac{\partial T}{\partial t}
\]

Here \( \xi \) is the extent of the decomposition reaction, which can be modeled in the simpler situations by, e.g., a first-order reaction of form \( \frac{\partial \xi}{\partial t} = (1 - \xi) A \exp\left(-\frac{Q}{RT}\right) \), including the possibility of reactant depletion as \( \xi \to 1 \). The physical meaning of \( E_r \) and \( P_r \) can be understood by considering the idealized situation of a sample of material that is constrained to deform in a macroscopically homogeneous manner, such that the \( \frac{\partial^2}{\partial y^2} \) terms above vanish (that signaling zero fluid seepage and zero heat conduction within the layer), as would be appropriate to having impermeable and adiabatic sample boundaries which prevent fluid or heat outflows. In that situation it is clear that \( dp = \Lambda dT + P_r d\xi \) and that \( dT = \left(\tau / \rho c\right) d\gamma - E_r d\xi \), where \( P_r \) is the pore pressure increase and \( E_r \) is the temperature decrease (both relative to the situation without chemical reaction) per unit advance of that reaction. The properties \( E_r \) and \( P_r \) are obtained from complex calculations using thermo-chemical material properties; see Platt et al. [2015] who present the resulting estimates of \( E_r \) and \( P_r \) for decarbonation reactions in calcite, and dehydration reactions in lizardite, illite/muscovite, and talc.

Those analyses by Platt et al. [2015] predict extreme localization of shear, with localization zone thicknesses on the order of 10 \( \mu \)m, and also show how reactant depletion can lead to layered strain distributions in relic fault zones, i.e., with stripes of fully depleted material next to stripes of unreacted (or less reacted) material.

Some further thoughts on the thinness of shear zones

This section was added in response to anonymous review comments on the original version of the paper, which comments remarked that “… moderate to large in size crustal earthquakes nucleate at 7-25 km in depth and seismic inversion studies suggest that the maximum slip is often localized in fault patches located at several km depth (e.g., 6-8 km depth for the L’Aquila Mw 6.3 2009 earthquake, Cirella et al. [2016]). Several field studies and their comparison with experimental studies (Fondriest et al. [2012, 2013]; Siman-Tov et al. [2013]; De Paola et al. [2015]; Demurtas et al. [2016], etc.), present evidence of extreme strain localization during seismic slip in carbonate-bearing faults exhumed from 2-4 km depth. This would extend the evidence of the presence of extremely thin slipping zones to [a] deeper level in the continental crust. The occurrence of such extremely thin slipping zones could be further extended to 60 km depth by considering the thin pseudotachylyte-bearing faults veins hosted in crustal and mantle rocks [Swanson, 1988; Austrheim and Boundy, 1994; Di Toro et al., 2005; Ueda et al., 2008].”

Conclusions and Perspectives

• Faults are subject to thermal weakening and consequent shear localization processes, which may involve:
  - thermal pressurization
• of native ground fluid (e.g., H₂O)
• of volatiles (e.g., H₂O or CO₂) as fault core decomposition products;
- flash heating at asperity contacts;
- melting, the ultimate weakening process, not very fully discussed here but evidenced
  by thin pseudotachylite layers along some exhumed faults.
• Static fault strength $\tau_{\text{static}} = f_{\text{static}} \sigma_n$ may be an unreliable predictor of
  pre-earthquake stress on major, well-slipped, smooth faults.
• Such faults may operate near or slightly above $\tau_{\text{pulse}}$ (a $\tau$ level at which a small event,
  once nucleated, propagates for indefinitely large distance).
• Roughness may keep less mature faults from the full consequences of
dynamic weakening; to slip them, the fault side-walls must be deformed
• The fact that a segment is creeping may not preclude it from having large
coseismic slip through stress pulses that are sufficient to activate a
dynamic weakening mechanism (Chi-Chi, 1999, Tohoku-Oki, 2011); see Noda and
Lapusta [2013].

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(Figures 1 and 2, and their captions, follow.)
Fig. 1. Model for frictional weakening by the flash-heating process (Rice [1999, 2006], Beeler et al. [2008], Rice et al., [2009]). Here, $D$ is a representative asperity contact size, and asperity shear ($\tau_c$) and normal ($\sigma_c$) stresses are comparable to the maximum sustainable shear stress and to the normal stress at contact indentation of a punch of a yet-stronger material. Force equilibrium demands that $(\text{avg. } \tau_c) / (\text{avg. } \sigma_c)$ be equal to the same ratio $\tau / \sigma$ based on the macroscopic average stresses, hence equal to the friction coefficient $f \sim 0.6$-0.8. Such reasoning, together with standard concepts involving hard indenters, leads to the conclusion that the local shear stress $\tau_c$ resisting sliding at a typical rock-on-rock asperity contact is of order $\tau_c \sim 0.1 \mu$ ($\mu$ is elastic shear rigidity).

$V_w$ is defined as the slip rate at which the asperity contact, initially at ambient fault temperature $T_f$, just reaches a "weakening temperature" $T_w$ (which may correspond to forming a melt layer at the contact) as it is slid out of existence. $V_w$ is estimated approximately as shown from a simplified one-dimensional heat conduction analysis ($\alpha_{th}$ is the thermal diffusivity, $\rho c$ the specific heat per unit volume). Thus, when the slip rate $V < V_w$, the asperities are assumed to remain strong, at $\tau_c \sim 0.1 \mu$, and the friction coefficient is the normal one in slow slip, called $f_{\text{slow}}$ here, and corresponding typically to the 0.6 to 0.8 range noted earlier. The 2nd to last expression for $f$ shown has the interpretation that when $V > V_w$ the contact is strong ($f = f_{\text{slow}}$) only for a fraction $V_w / V$ of its lifetime, but is weak ($f = f_{\text{weak}}$, which may be notably smaller than $f_{\text{slow}}$) for the remaining fraction $(1 - V_w / V)$. The approximate validity of that representation is well demonstrated: Beeler et al. [2008], Goldsby and Tullis [2011], [Proctor et al., 2014].
Fig. 2. **At top:** Extreme localization of earthquake shear due to thermal weakening of a groundwater-saturated granular fault gouge layer, of thickness $h = 1$ mm in this illustration based on Platt, Rudnicki, and Rice [2014]. The 1 mm thick gouge layer is forced to shear by imposed motion of two water-saturated, linear thermo-poroelastic blocks, forced to move relative to one another at a typical earthquake slip rate $V = 1$ m/s.
Significant straining ultimately localized to a zone of thickness \( W \sim 43 \, \mu m \) (<< \( h = 1,000 \, \mu m \)), as marked. That \( W \), also called \( W_{\text{nonlin. calc.}} \), corresponds to the width based on the full nonlinear numerical calculations of localization by Platt et al. [2014]. It is comparable to an approximate estimate of localization width, called \( W_{\text{lin. pert.}} \), derived by Rice et al. [2014] as half the shortest unstable Fourier wavelength in a linear perturbation analysis of an otherwise spatially uniform shear flow state with a spatially uniform strain rate. **Fig. 2. At bottom:** Weakening with increasing slip, also based on the full nonlinear calculations by Platt et al. [2014]. When the gouge is constrained to deform uniformly the solution agrees well with the Lachenbruch [1980] solution (not perfectly because Lachenbruch prohibited fluid or heat transfer into the bordering regions, which is not the case here -- the bordering regions are thermo-poroelastic media). When the gouge is not constrained to shear uniformly, the solution agrees closely, as slip increases, with the Rice [2006] solution for slip on a mathematical plane, which solution includes the simultaneous effects of both heat conduction (with thermal diffusivity \( \alpha_{th} \)) and poroelastic seepage (with hydraulic diffusivity \( \alpha_{hy} \)). An earlier Mase and Smith [1987] solution considered only one of those two transport processes in analysis of this particular problem (although they developed, in other contexts, analyses with both \( \alpha_{th} \) and \( \alpha_{hy} \) being non-zero). The Rice [2006] solution illustrated here reduces to theirs when either \( \alpha_{th} = 0 \) or \( \alpha_{hy} = 0 \).
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