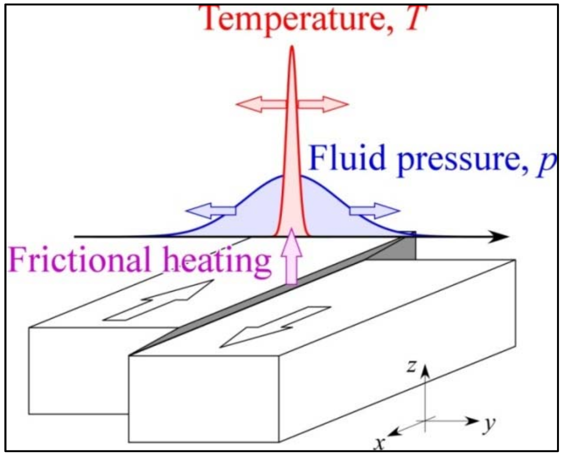
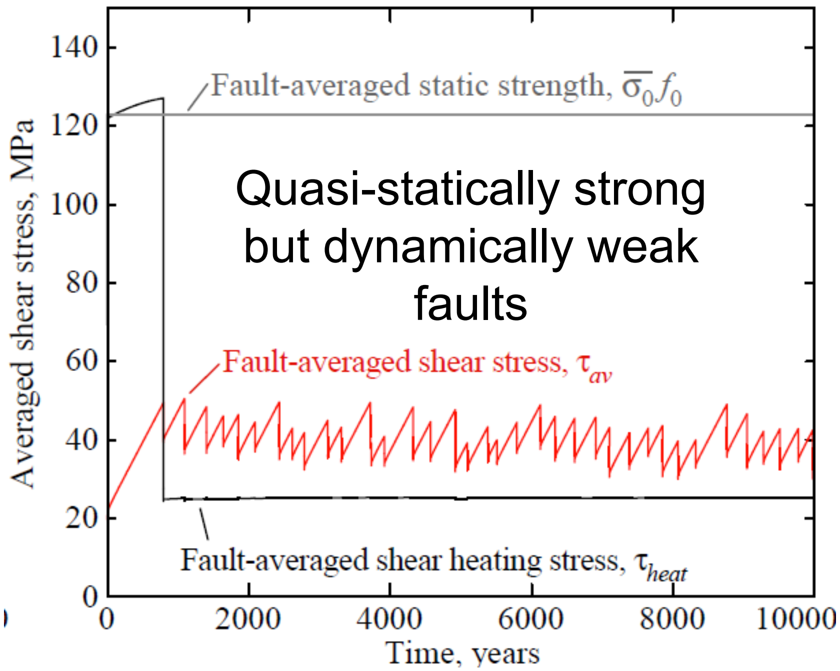
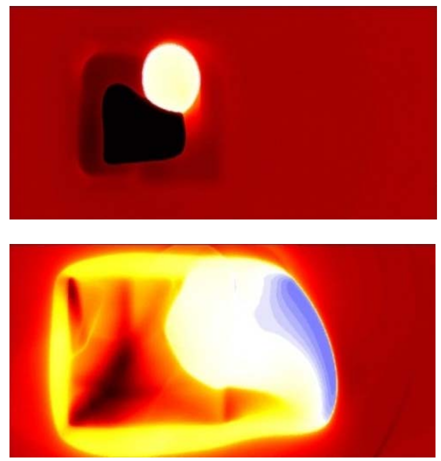
Long-term behavior of fault models with co-seismic weakening

Nadia Lapusta, Caltech

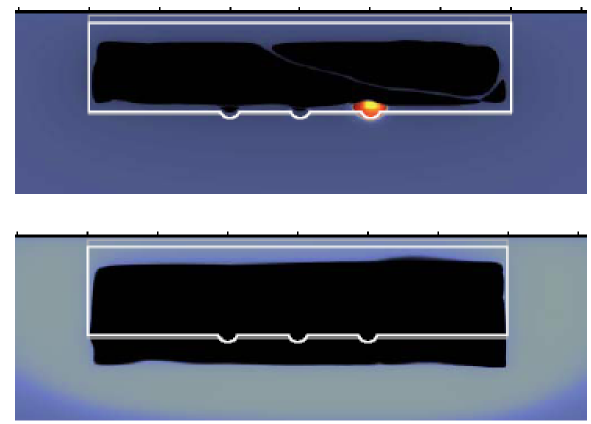
Work with Hiroyuki Noda, Jim Rice, Junle Jiang, Ting Chen







Can "decoupled" fault areas host large seismic slip?



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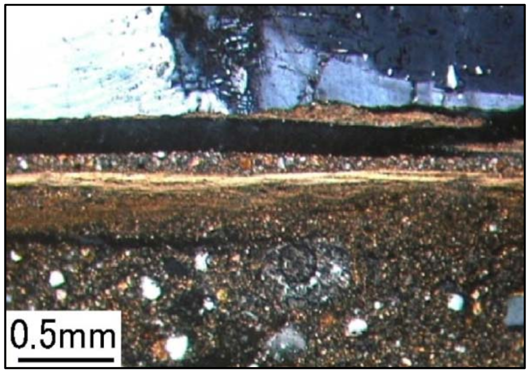
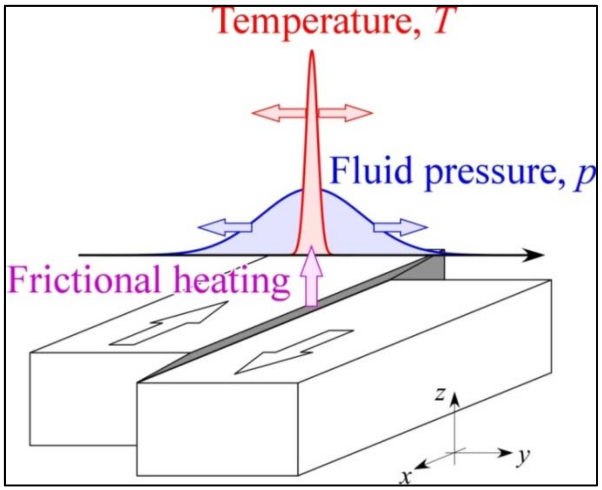
Fault constitutive behaviors



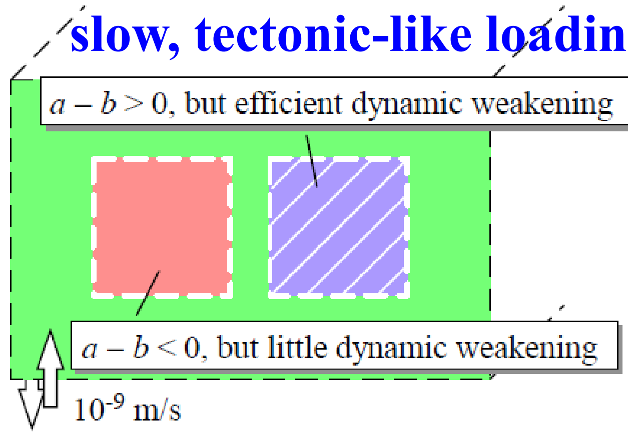


"Friction" laws

$$\tau = (\sigma - p)\left(f_o + a \ln \frac{V}{V_o} + b \ln \frac{V_o \theta}{L} \right)$$

$$\frac{d\theta}{dt} = 1 - \frac{V\theta}{L}$$

Simplified models with slow, tectonic-like loading



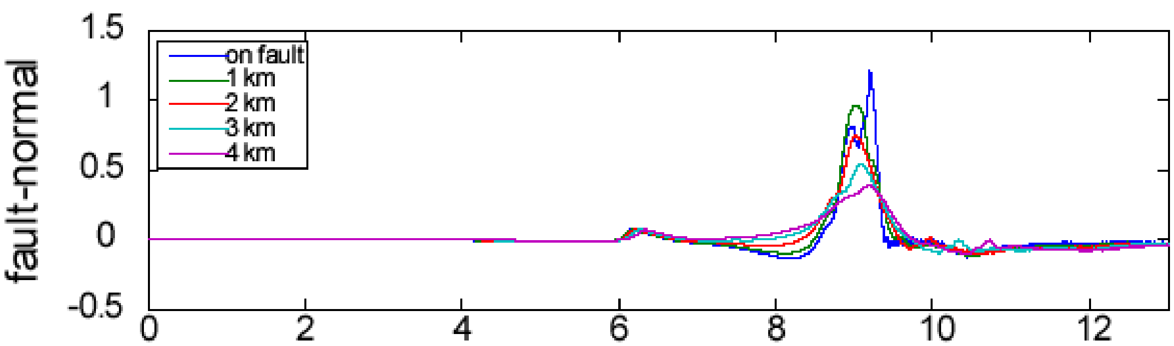
Comp. methodology

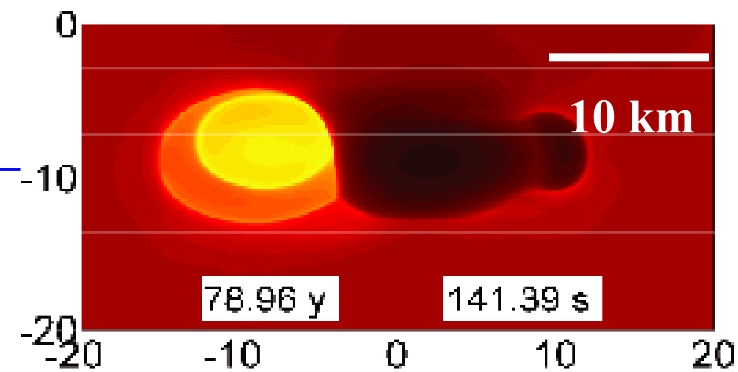
Boundary integral method

Spectral element method

Supercomputer

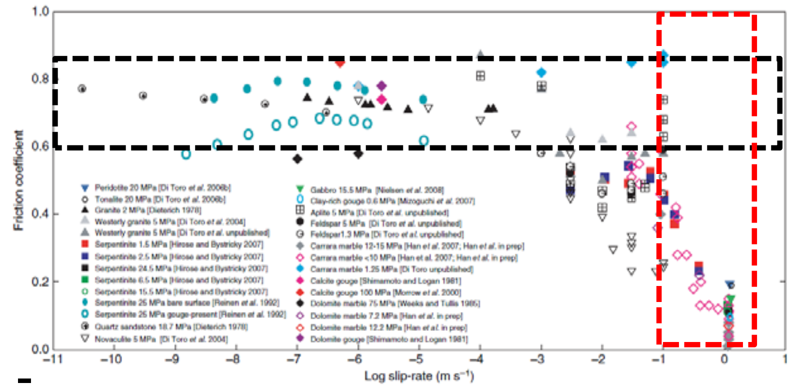
Spontaneous fault motions over 100-1000 years: sequences of earthquakes with all wave effects resolved, earthquake nucleation, post- and interseismic slip.

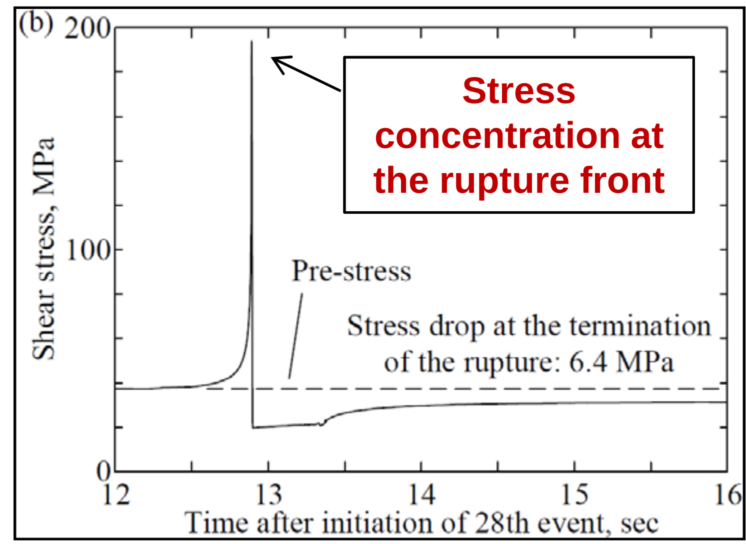


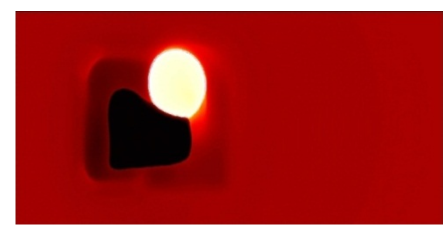


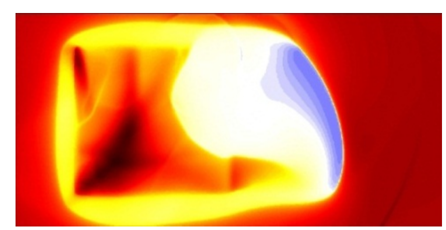
Conclusions

- Shear zones that host large earthquakes potentially experience ***extreme coseismic weakening***, due to shear heating and other physical mechanisms.
- Models that incorporate such weakening can reconcile
  -- the apparent weakness of mature faults,
  -- their high quasi-static strength,
  -- typical stress drops for large events,
  -- low heat generation on mature faults,
  -- compact short-duration pulse-like rupture mode.
- In such fault models, **availability and nature of spots favorable to earthquake nucleation** is important.
- Creeping ("decoupled") fault patches may be susceptible to coseismic weakening, joining earthquakes to produce large coseismic slip.
- A model with such patch qualitatively explains observations on a range of temporal scales for two well-studied earthquakes (1999 Chi-Chi & 2011 Tohoku-Oki).
- Earthquakes may penetrate below the traditionally defined seismogenic zone due to coseismic weakening.





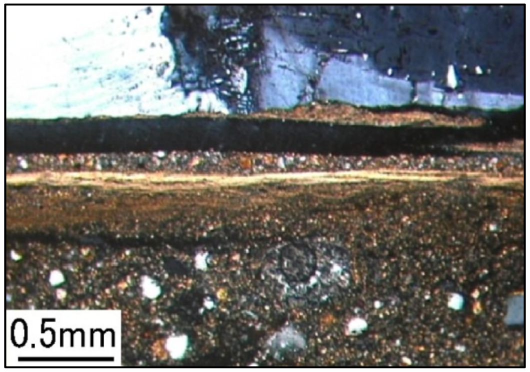




Fault deformation modeling is multiscale on several levels

Multiscale Aspect I

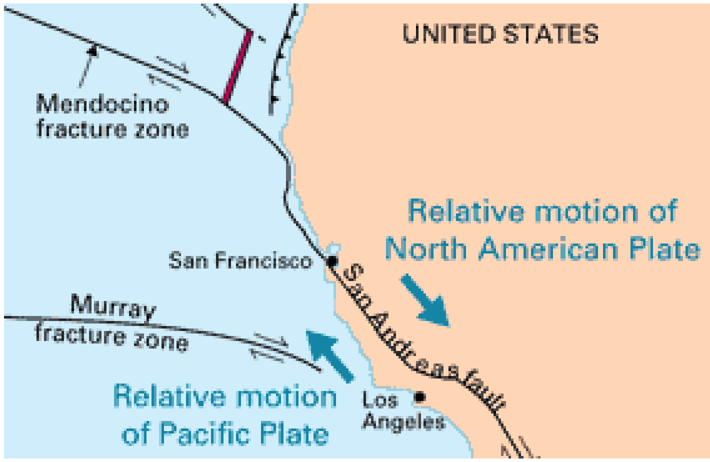
Constitutive response of the earthquake shearing zone



Multiscale Aspect II

Spontaneous slip accumulation on a planar interface under slow loading assuming simple (elastic) bulk

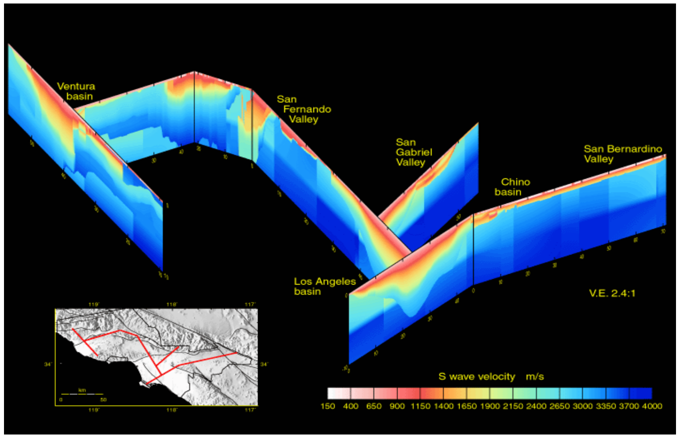
10^{9}-10^{10} s slow loading / aseismic slip / slow deformation
10^{5}-10^{6} s accelerating nucleation process
10 -100 s duration of a large inertially-controlled event
10^{-3}-10^{-1} s variation of stress and slip rate at rupture front



Multiscale Aspect III

Heterogeneous damaged temperature- and pressure-dependent visco- poro- elasto- plastic bulk material

Locally non-planar shear zone with varying thickness



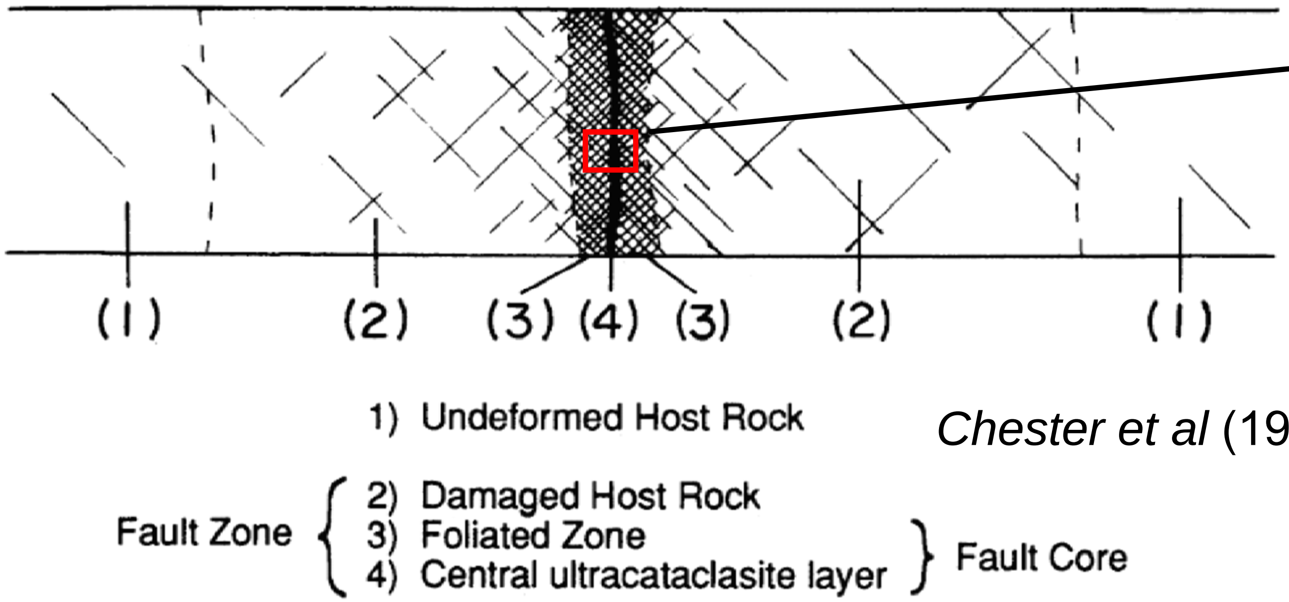
Multiscale Aspect IV

Hierarchy of shear zones, interaction between them; large-scale fault system structure

⇒ Need appropriately formulated laws, multiple physical inputs, and advanced numerical methods

Constitutive response of the earthquake shear zone

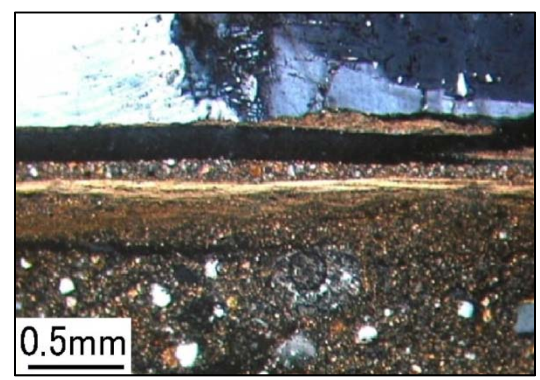
Localized layer (1-5 mm) of finely granulated material
(Particle size range: 10 nm to 100 μm; $d_{50} \sim 1$ μm)
embedded within a broader zone of damaged material



Chester et al (1993)

Fig. 2. Schematic section across the North Branch San Gabriel fault zone illustrating position of the structural zones of the fault. The diagram is not to scale.

Chester and Chester, 1998



Mizoguchi et al (2004)

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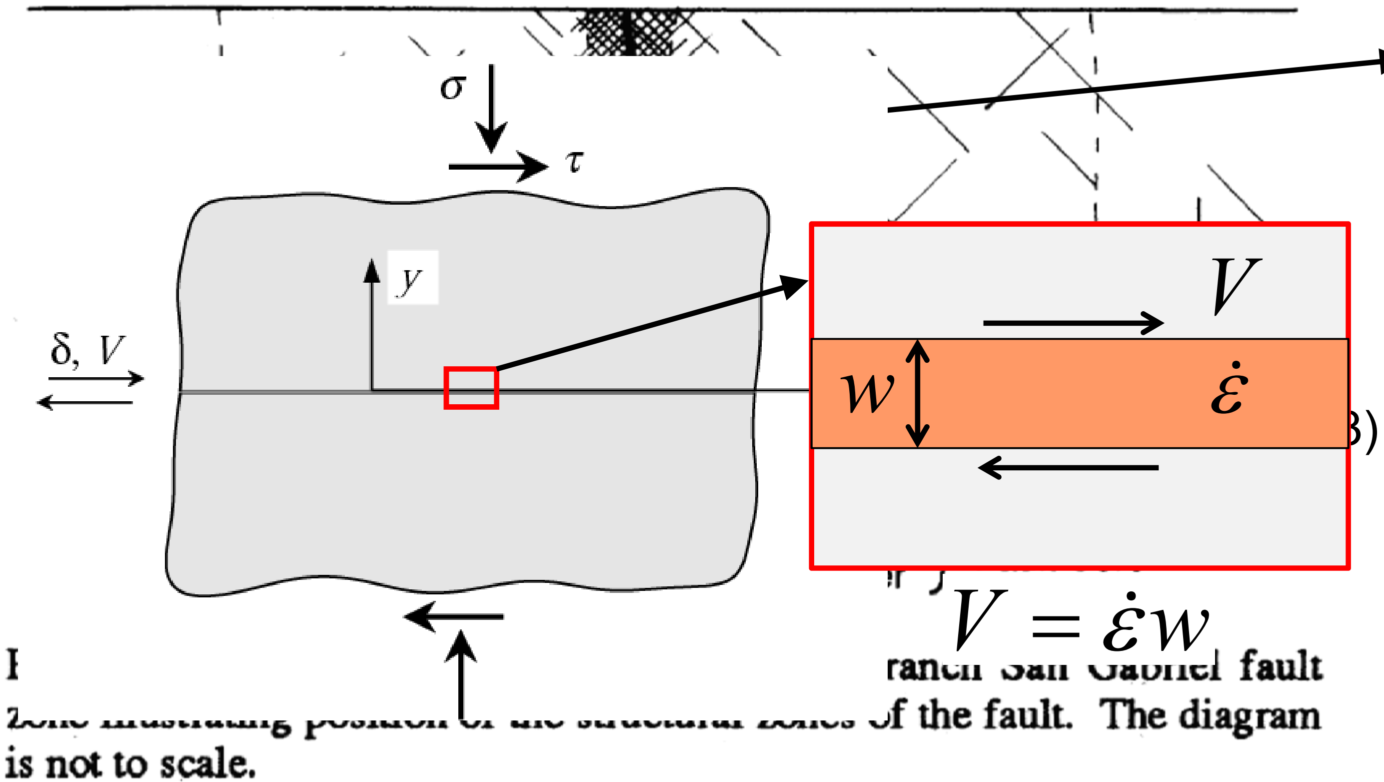
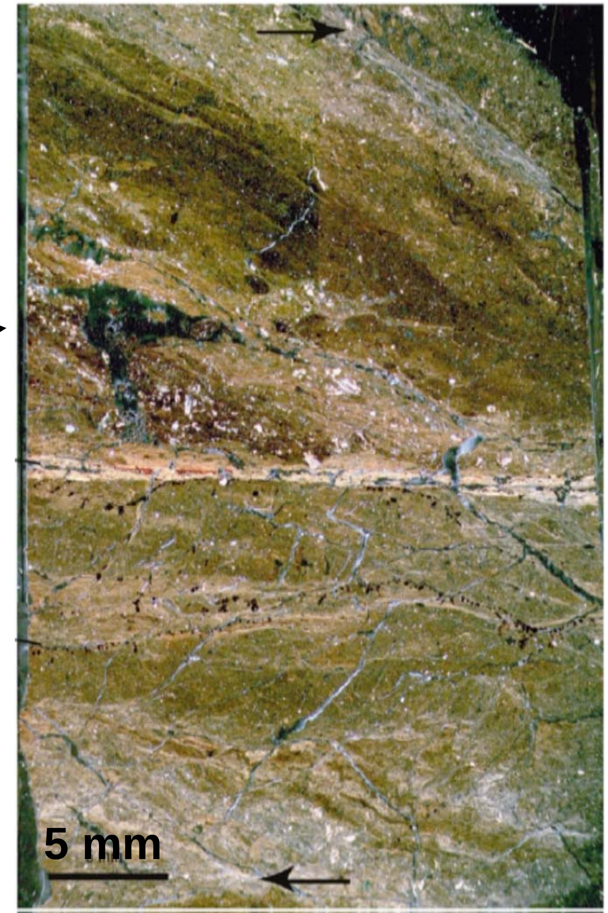
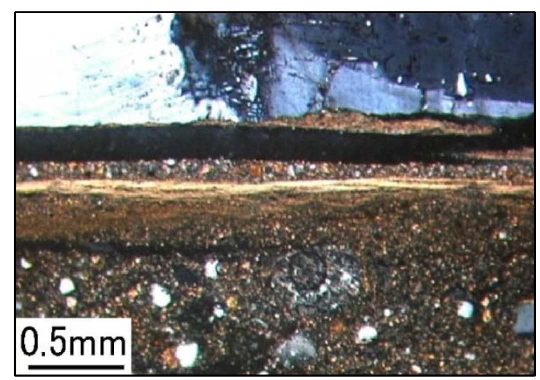




Chester and Chester, 1998

For numerical tractability,
we need a law prescribing fault strength:

$$\tau = f\,(\delta,\, V,\, \boldsymbol{\theta_i},\, T,\, \ldots,\, \sigma,\, p)$$



Mizoguchi et al (2004)

Important conceptual advance: Low-velocity rate and state friction

Laboratory-derived (Dieterich, Ruina, Tullis, Marone, and others) **for slip velocities** small **(~ 10^{-9} – 10^{-3} m/s) compared to the seismic range**.

Unique tool for simulating earthquakes and slow slip in their entirety,
from accelerating slip in **slowly expanding nucleation zones**
to rapid dynamic propagation of **earthquake rupture**
to **post-seismic slip** and interseismic creep
to **fault healing** between seismic events.

$$\tau = \bar{\sigma} f = (\sigma - p) f = \bar{\sigma}\left(f_o + a \ln \frac{V}{V_o} + b \ln \frac{V_o \theta}{L} \right); \quad \frac{d\theta}{dt} = 1 - \frac{V\theta}{L}$$

Base friction $f_o = 0.6$ at $V_o = 1$ μm/s

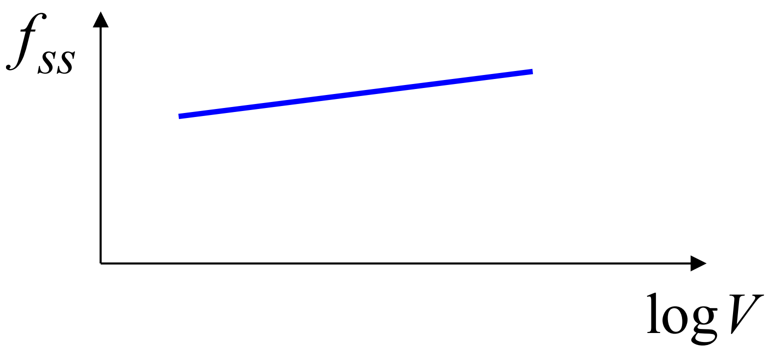
Variations $a = 0.015$, $b = 0.019$, $L = 1$-100 μm (lab values)

Numerous successful applications:
earthquake nucleation, earthquake sequences, postseismic slip, earthquake triggering, aftershock sequences, slow slip transients, scaling of repeating earthquakes

$$\tau/(\sigma - p) = f = f_o + a\ln\frac{V}{V_o} + b\ln\frac{V_o\theta}{L}; \quad \frac{d\theta}{dt} = 1 - \frac{V\theta}{L}$$

$$V \text{ constant}, \quad \theta_{ss} = L/V, \quad \tau_{ss}/(\sigma - p) = f_{ss} = f_o + (a-b)\ln(V/V_o)$$

$a - b > 0$, velocity strengthening



Aseismic slip under slow loading

Factors that favor VS in experiments:

High temperatures ($\gtrsim$ 300° C)

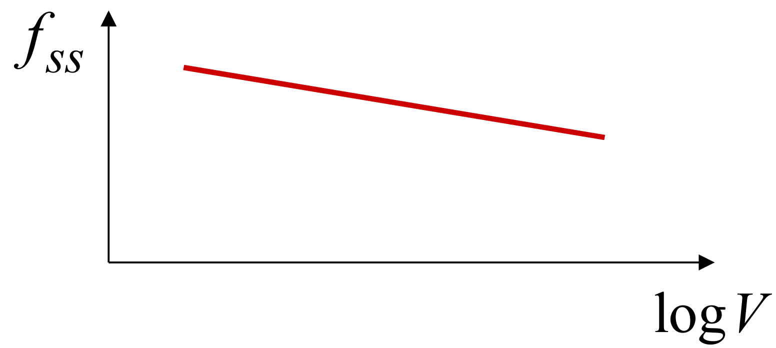
$\Rightarrow$ **Below certain depth**

Low effective normal stress

$\Rightarrow$ **Shallow VS layers**

Certain types of rocks and fault gouge

$a - b < 0$, velocity weakening



Seismic slip in *large enough* regions

Aseismic slip in smaller regions

Estimates of the critical size

(Rice and Ruina, 1983; Rice, Lapusta, Ranjith, 2001; Rubin and Ampuero, 2005):

$$h^* \propto \frac{\text{shear modulus} \times \text{char. slip}}{(\text{normal stress - pore pressure}) \times F(a,b)}$$

$$h^*_{RR} \propto \frac{\mu L}{(\sigma - p)(b-a)}; \quad h^*_{RA} \propto \frac{\mu L}{(\sigma - p)(b-a)^2/b}$$

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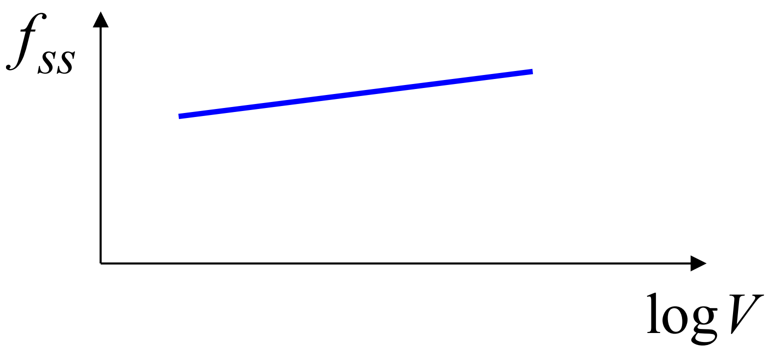

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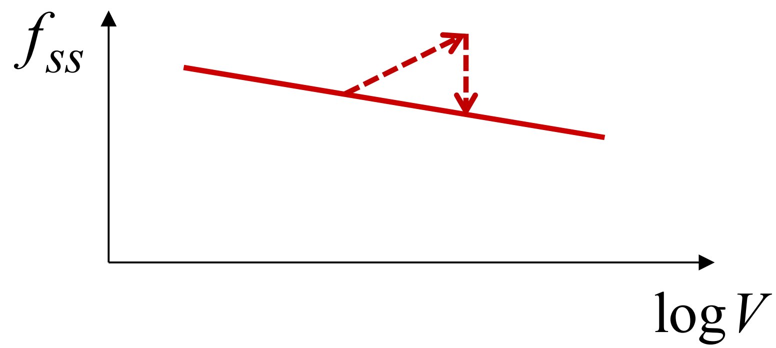
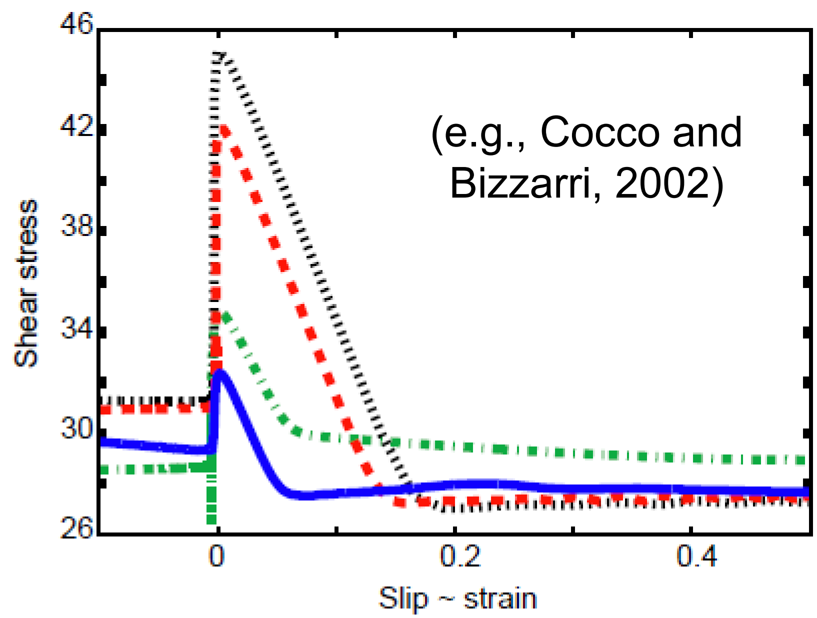
Relation to creep laws:

$$\dot{\varepsilon} = \frac{\dot{\varepsilon}_o}{\exp[(a - b)(\sigma - p)]} \exp(\Delta \tau_{ss})$$

$a - b < 0$, velocity weakening



Behavior at the dynamic rupture tip:



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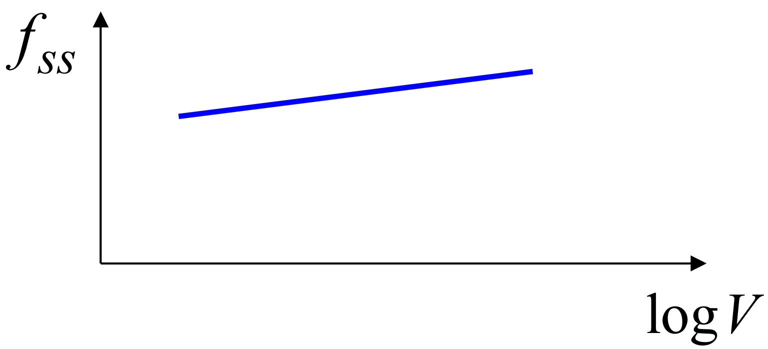

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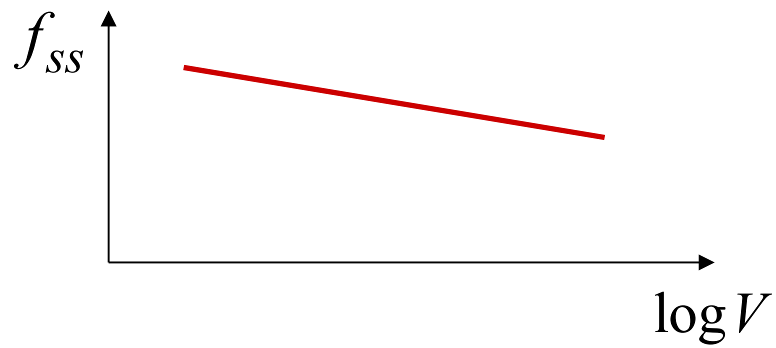

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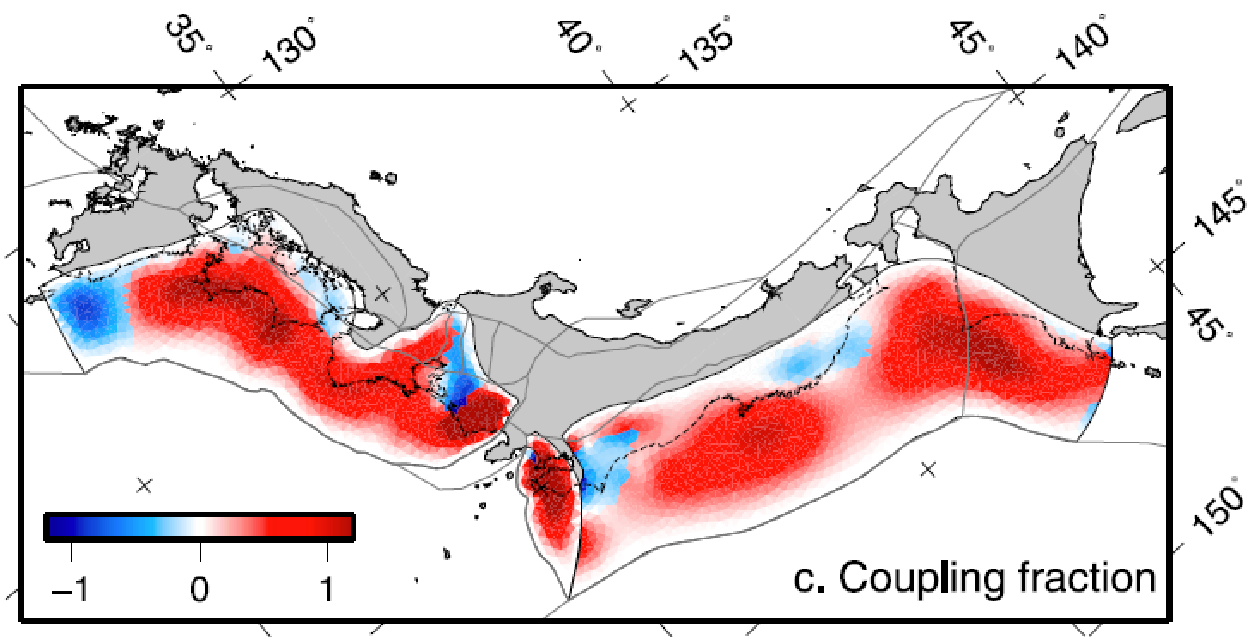
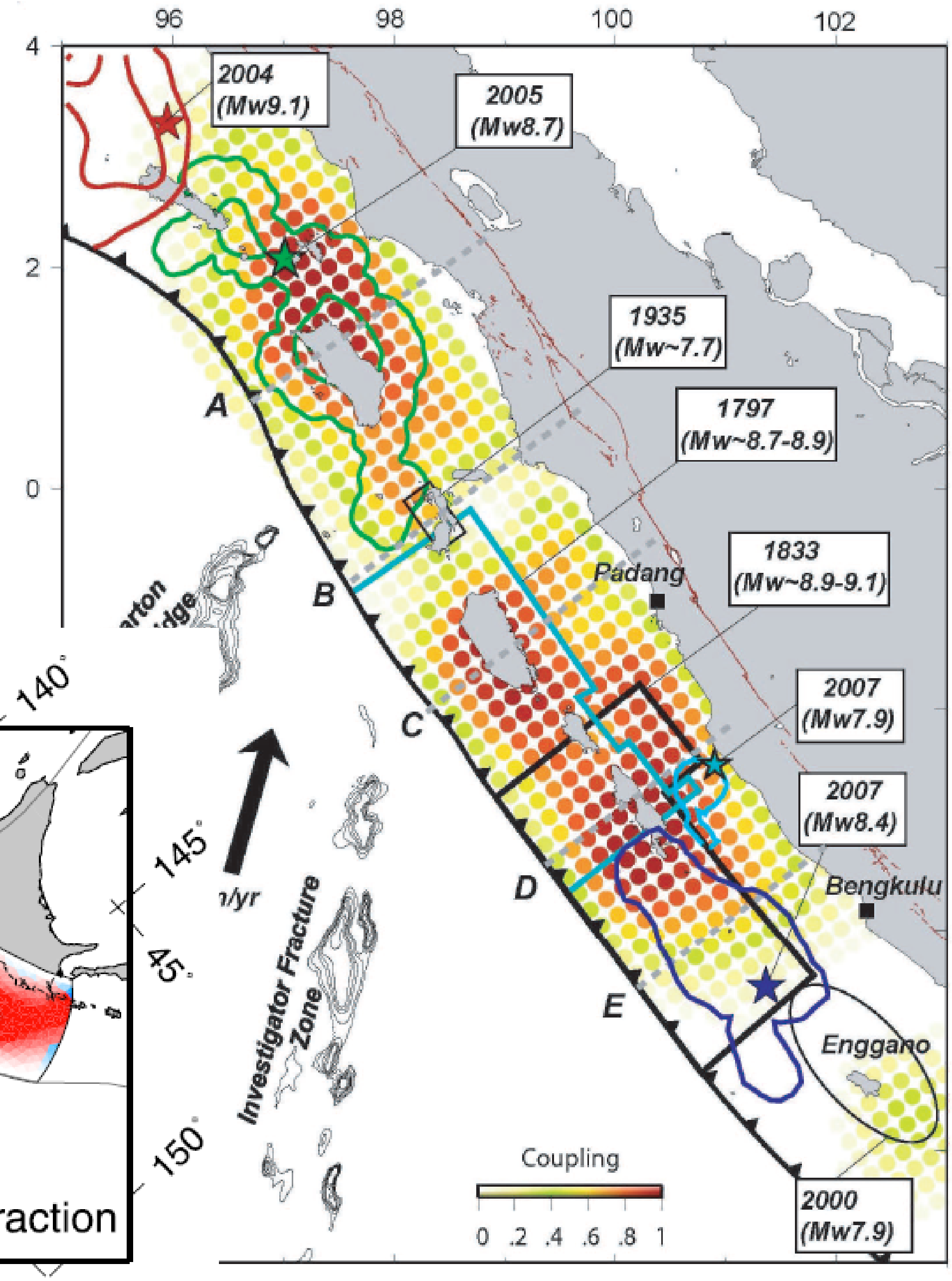
Is fault separation into stable/unstable areas persistent?
(Convenient picture but potentially too simplified)

Slowly moving (creeping) areas
⇔ rate-strengthening friction
⇔ "barriers" to earthquake rupture

Locked segments
⇔ rate-weakening friction
⇔ "seismic asperities"

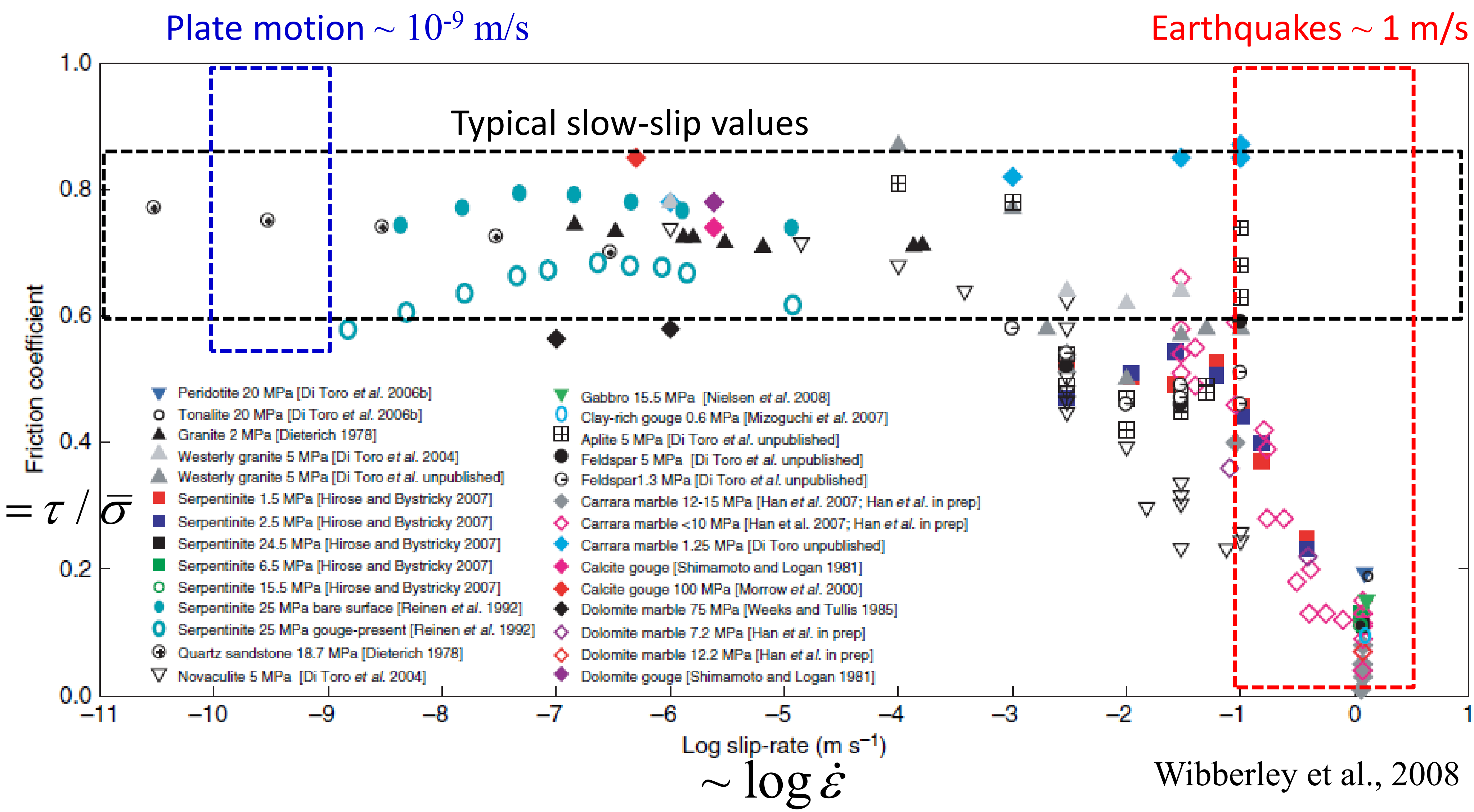


Japan (Loveless and Meade, 2010)



Sumatra (Chlieh et al., 2008)

Potential complication: Substantial add. weakening at high slip rates



Wibberley et al., 2008

Theories, experimental evidence for the much smaller shear zone resistance at fast slip rates

Shear heating mechanisms

Flash heating of contacts at small slips (e.g., Bowden and Thomas, 1954, Lim and Ashby, 1987, Molinari et al., 1999, Rice, 1999; 2006; Beeler and Tullis, 2003)

$$\tau_{ss} = (\sigma - p)\left[f_w + \frac{f_{ss}^{\text{rate/state}} - f_w}{1 + V / V_w} \right]$$

Thermal pressurization of pore fluids/decomposition products in the fault zone (e.g., Sibson, 1973; Lachenbruch, 1980; Mase & Smith, 1985, 1987; Andrews, 2002; Garagash & Rudnicki, 2003; Rice, 2006; Noda et al., 2009; Brantut et al., 2008 and others)

Partial or full melting of the shearing layer (e.g., Jeffreys, 1942; McKenzie and Brune, 1972; Tsutsumi and Shimamoto, 1997; Hirose and Shimamoto, 2005; Di Toro et al., 2006)

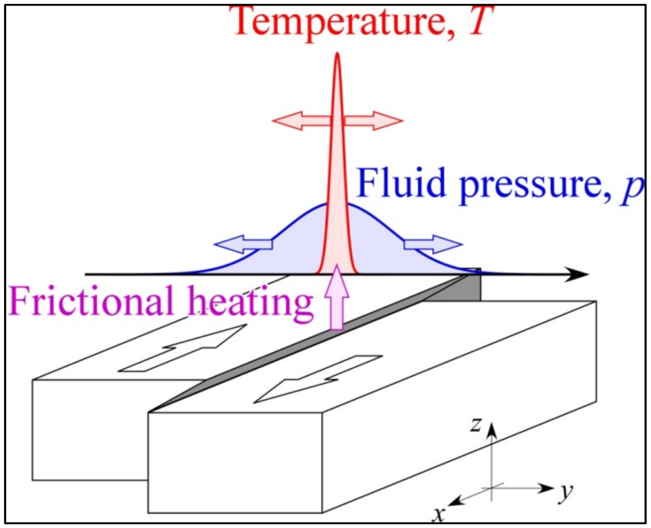
Other possibilities

Lubrication by silica gel layer (Goldsby and Tullis, 2003; Di Toro et al., 2004)

Normal stress reduction from elastic mismatch (Weertman, 1963, 1980 and others)

…

Rapid shear heating ⇒ Temperature and pore pressure evolution



Effective stress law:

$$\tau = f\bar{\sigma} = f\left(\sigma - p\right);\ f \equiv \text{rate-and-state law}$$

Temperature evolution (with diffusion normal to the fault):

$$\frac{\partial T(x,y,z,t)}{\partial t} = -\alpha_{th}\frac{\partial^2 T}{\partial y^2} + \frac{\omega}{\rho c}$$

Heat source:

$$\omega = \frac{\tau V}{w\sqrt{2\pi}}\exp\left(-\frac{y^2}{2w^2}\right)$$

T : Temperature
α_{th} : Thermal diffusivity
ω : Heat generation per unit volume
ρ : Density
c : Heat capacity per unit mass
w : Half width of the shear zone

Pore fluid pressure evolution (with diffusion normal to the fault):

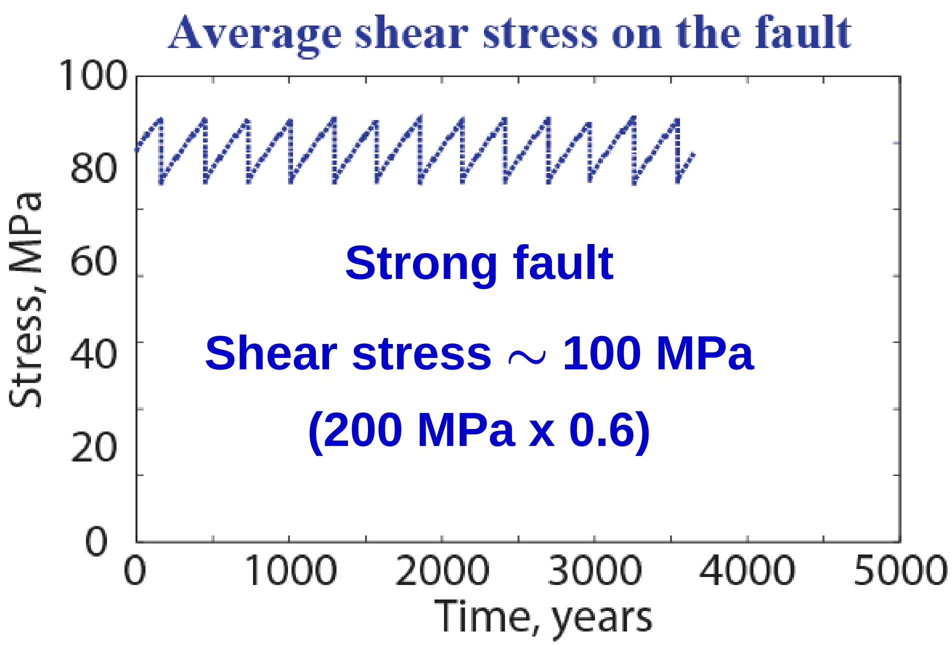
$$\frac{\partial p(x,y,z,t)}{\partial t} = -\alpha_{hy}\frac{\partial^2 p}{\partial y^2} + \Lambda\frac{\partial T}{\partial t}$$

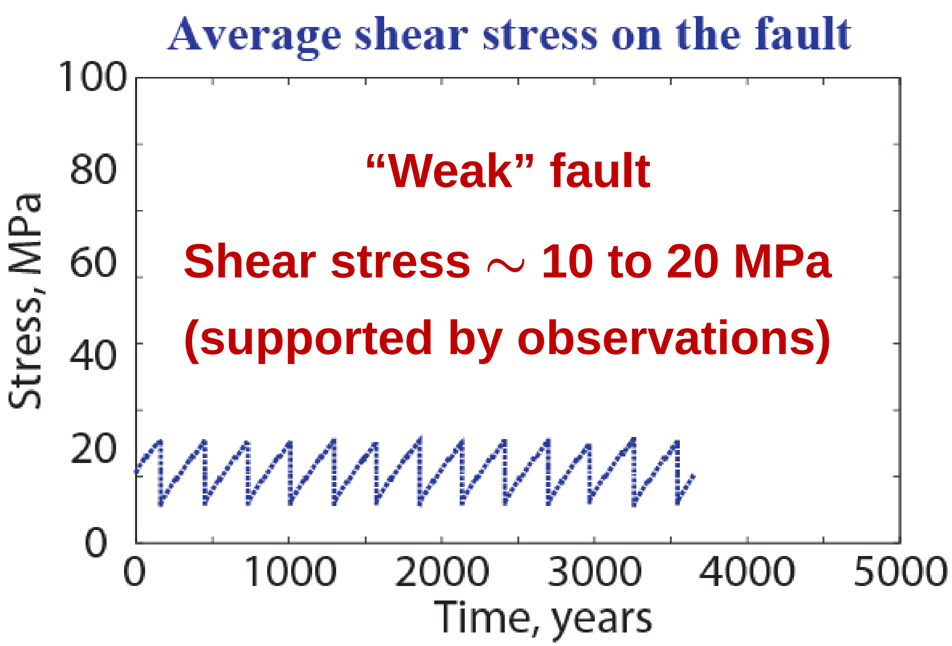
α_{hy}: **Hydraulic diffusivity (depends on permeability)**
Λ : Fluid pressure change / temperature change

Noda and Lapusta (2010)

Major mature faults appear to be "weak"

Definition based on *average* shear stress level





Evidence for "weakness" of mature faults:

Steep angles between the max principal stress and fault trace (e.g., Townend and Zoback, 2004; Hickman and Zoback, 2005)

Long-term heat outflow (Brune et al., 1969; Henyey & Wassenburg, 1971; Lachenbruch & Sass, 1973, 1980).

Fault temperature measurements after earthquakes in drill sites (e.g., Kano et al., 2007)

Significant rotations of principal stresses due to stress drops (e.g., Wesson and Boyd, 2007).

Geometry of thrust-belt wedges (Suppe, 2007).

Why would faults be "weak" on average?

$$\tau = \bar{\sigma} f = (\sigma - p) f = (\text{normal stress} - \text{pore pressure}) \times \text{friction coefficient}$$

Common explanations

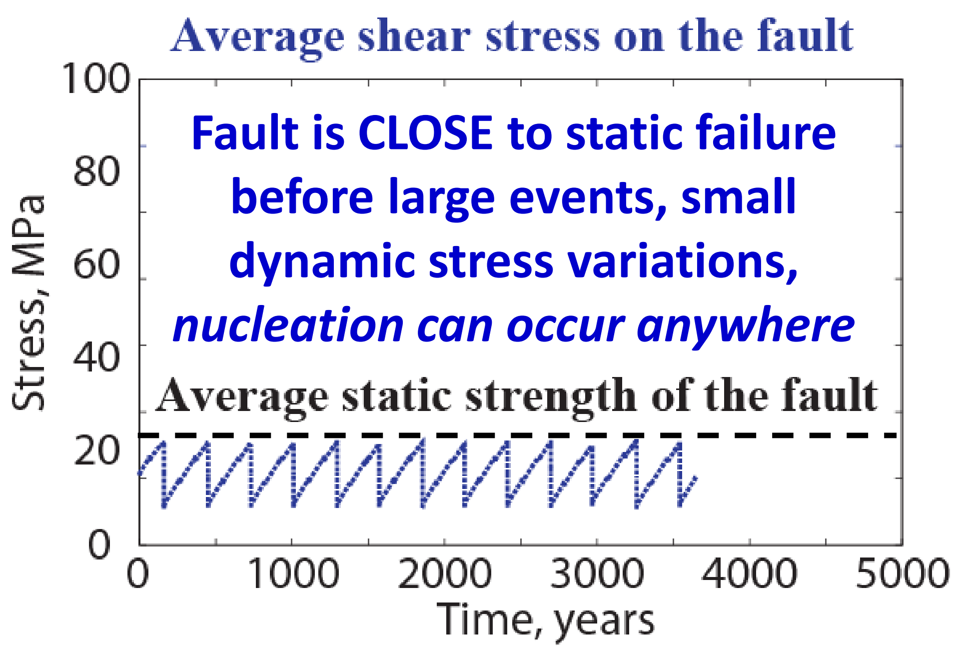
Low effective normal stress (< 40 MPa) (high pore pressure) OR low static friction coefficient (< 0.1) OR both

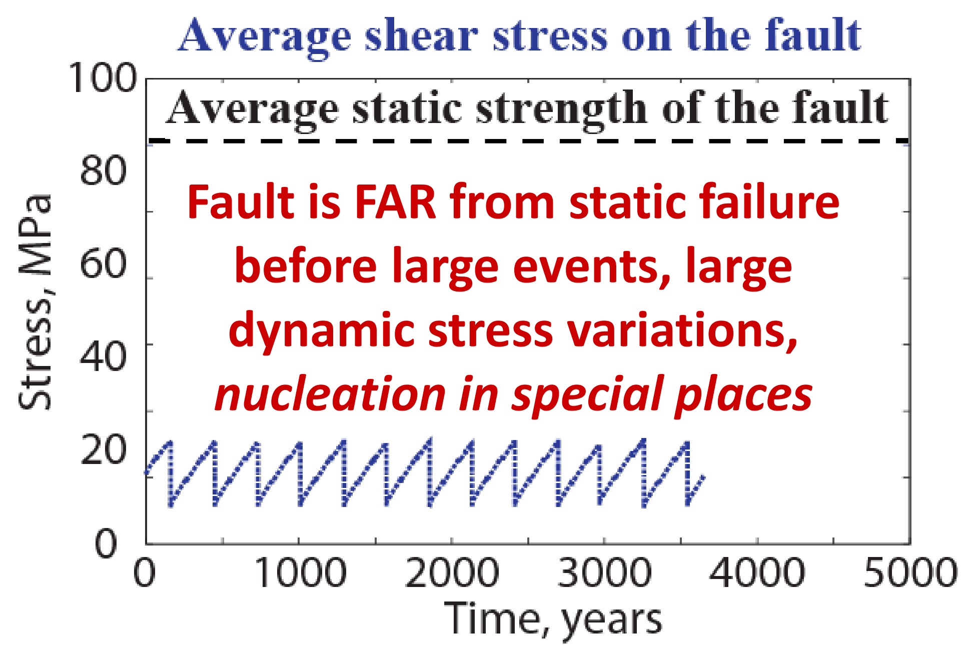
One more possibility

Faults are strong (~100 MPa) at low slip rates, but weak (~10–20 MPa) at high slip rates, with favorable spots to nucleate quakes

(Name? "Strong but very brittle"?)

(Lapusta, Noda, Rice, 2012)



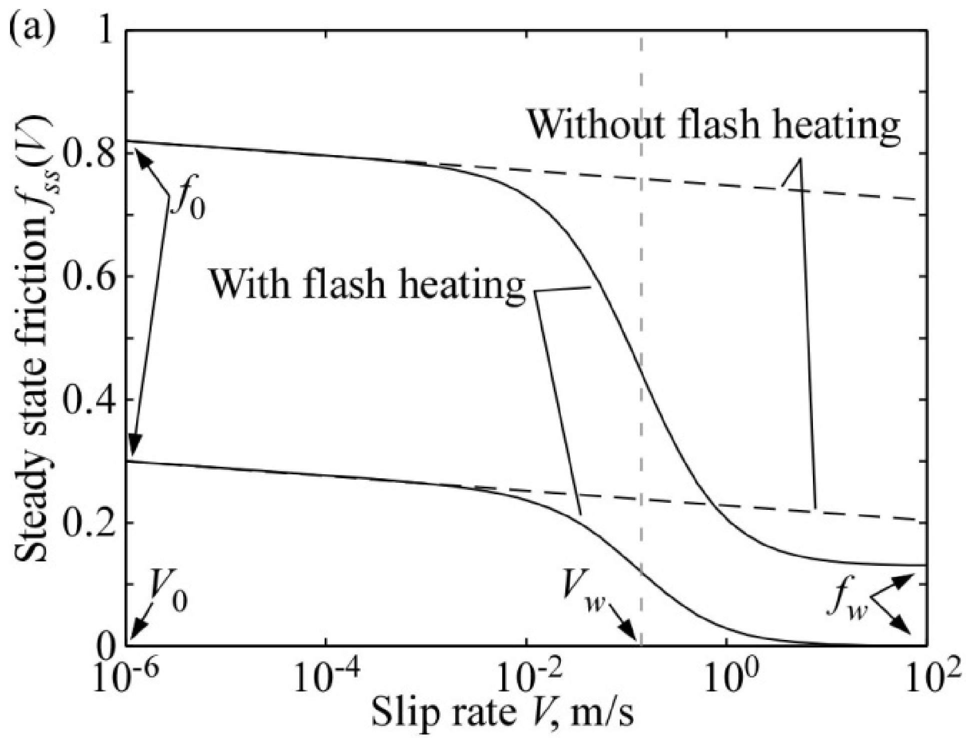


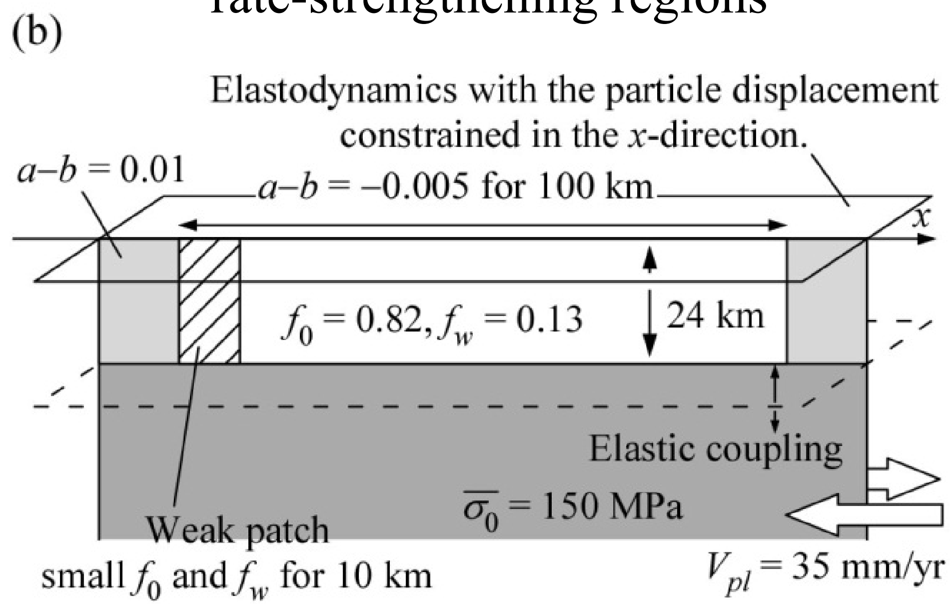
Fault model with enhanced co-seismic weakening

Noda, Lapusta, and Rice, in prep

$$f_{ss} = \frac{\tau_{ss}}{(\sigma - p)} = f_w + \frac{f_{ss}^{\text{rate/state}} - f_w}{1 + V / V_w}$$

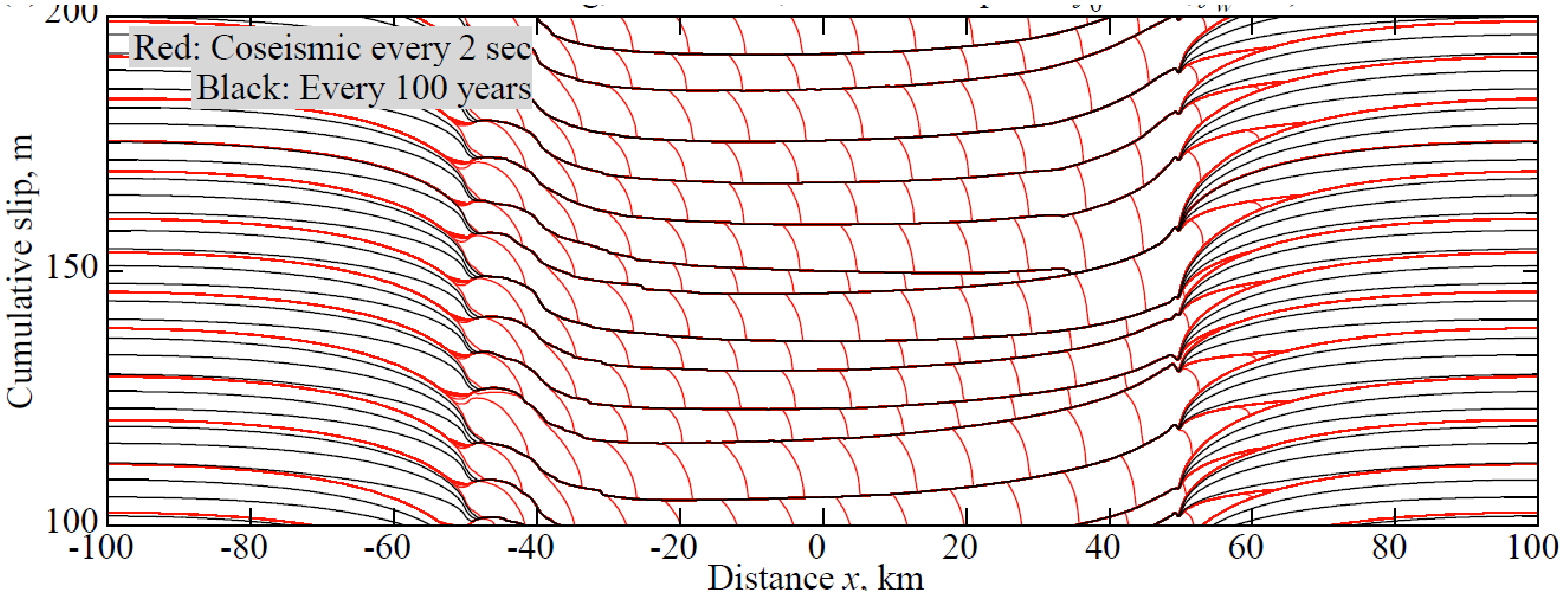
- 2D model with 1D strike-slip fault that is averaged through the depth;
- Low-slip-rate behavior: rate-weakening segment surrounded by rate-strengthening regions





Numerical simulation methodology for long-term fault slip punctuated by earthquakes with all wave effects: Lapusta et al. (2000); Lapusta and Liu (2009); Noda and Lapusta (2010)

Behavior of such “strong but very brittle” faults



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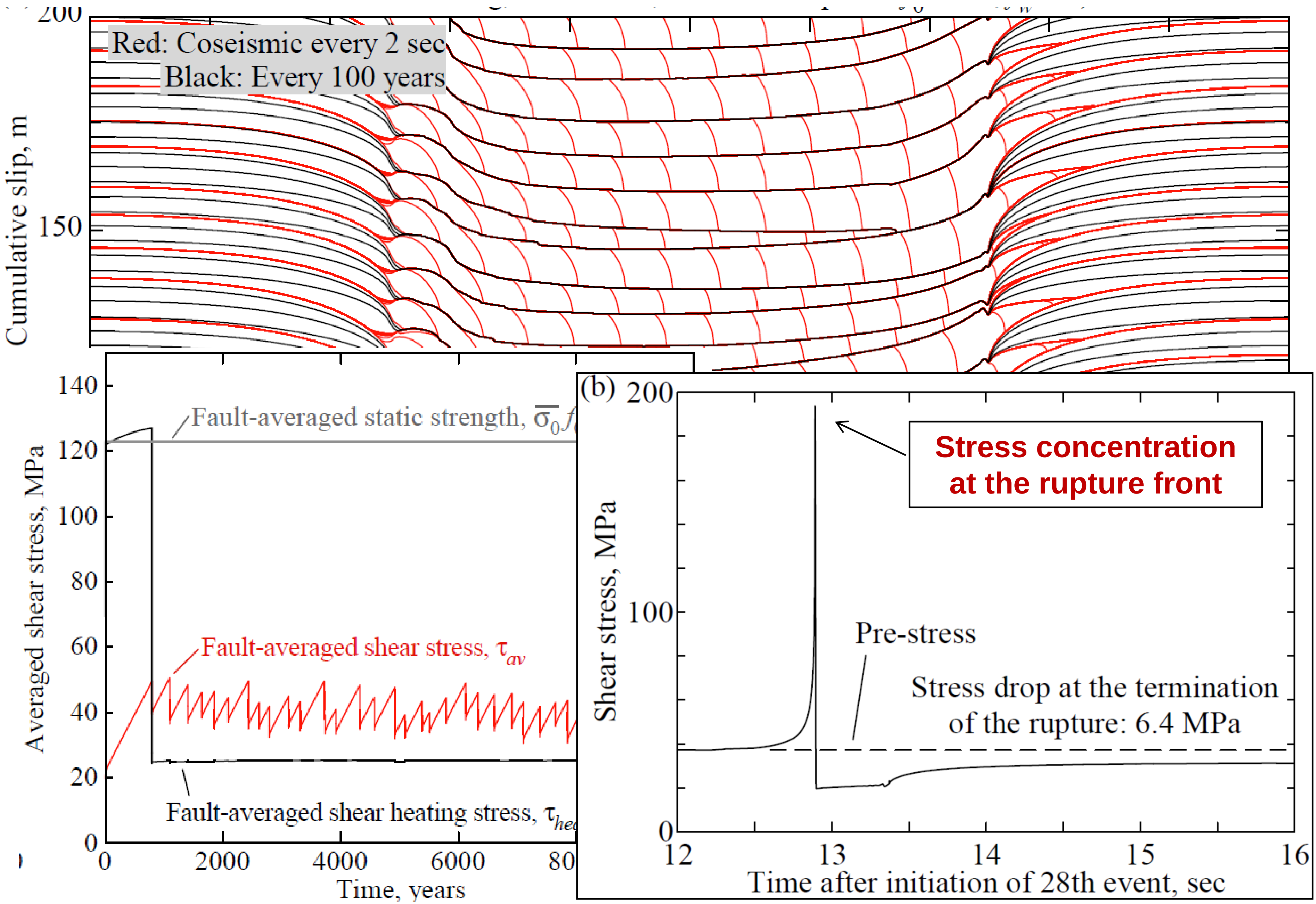
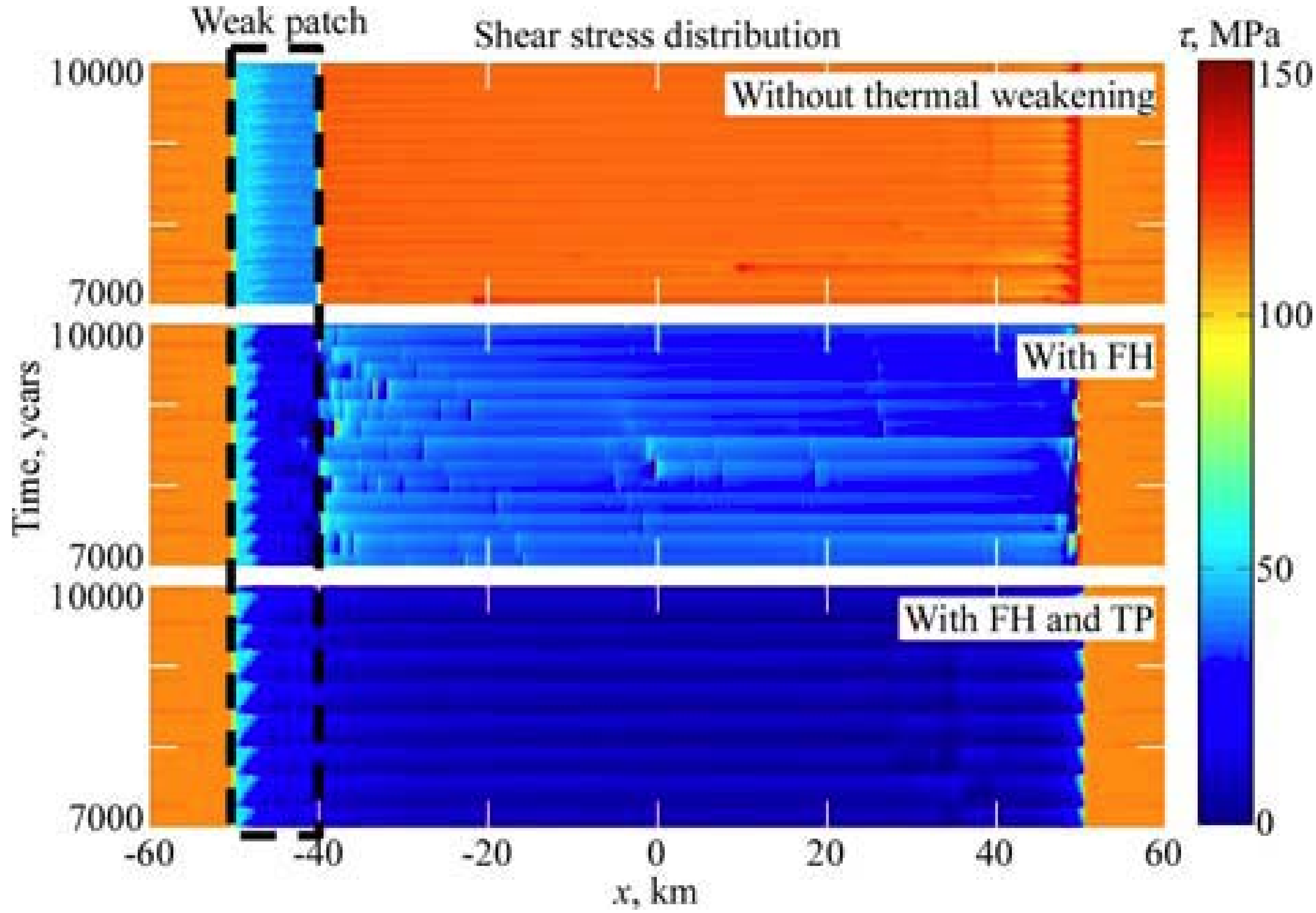


Low shear stress on the fault: determined by the co-seismic (dynamic) fault strength



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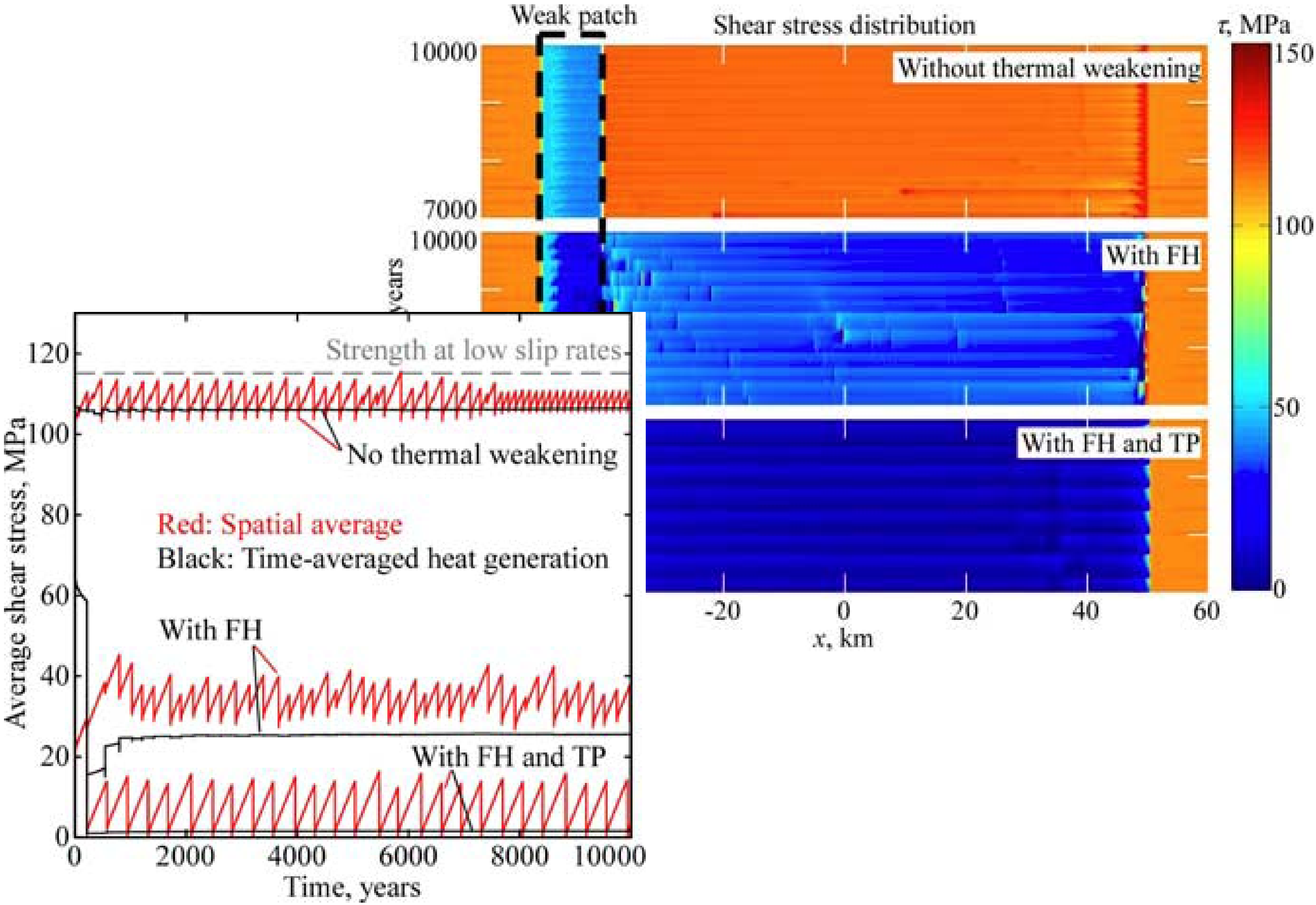
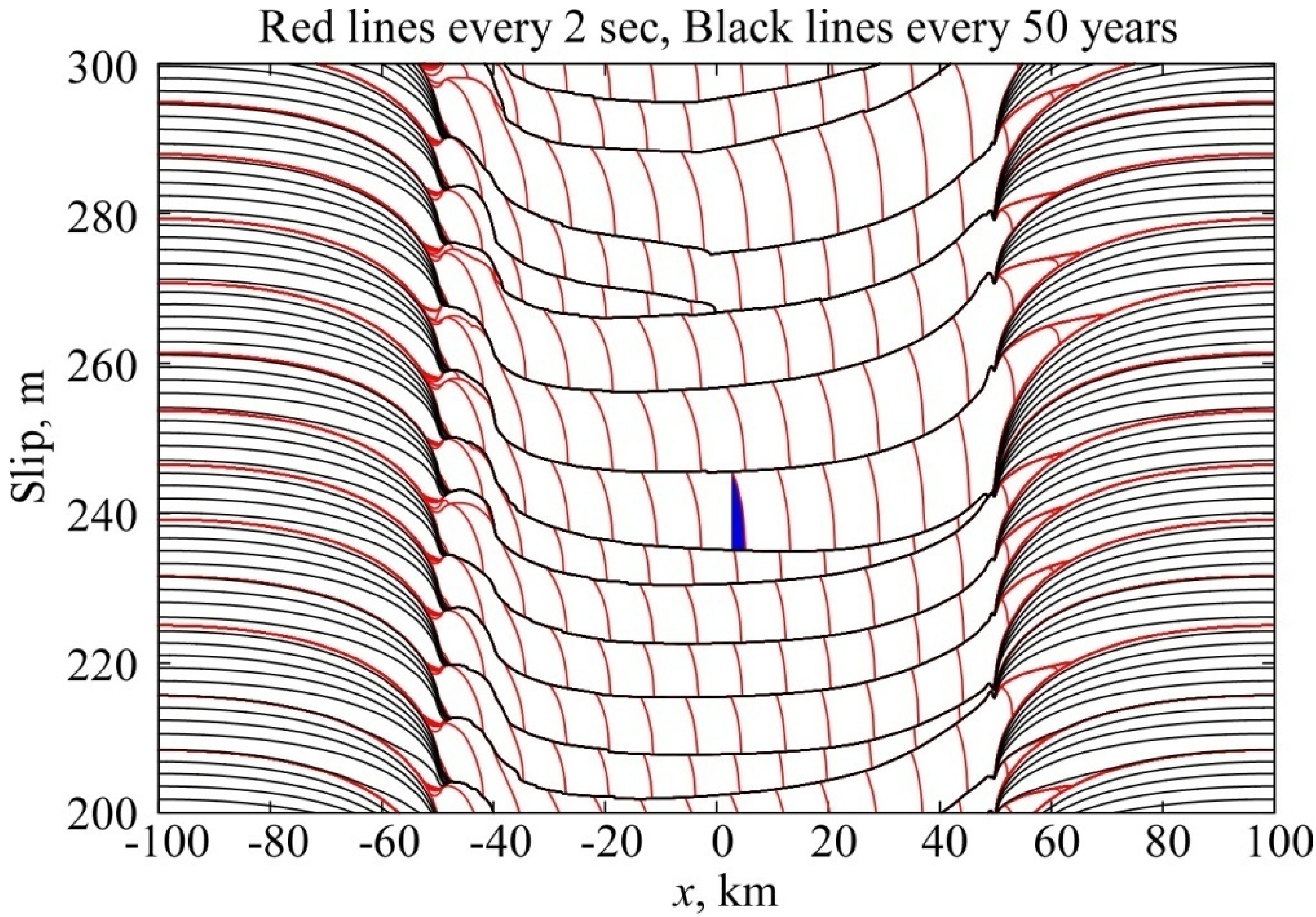
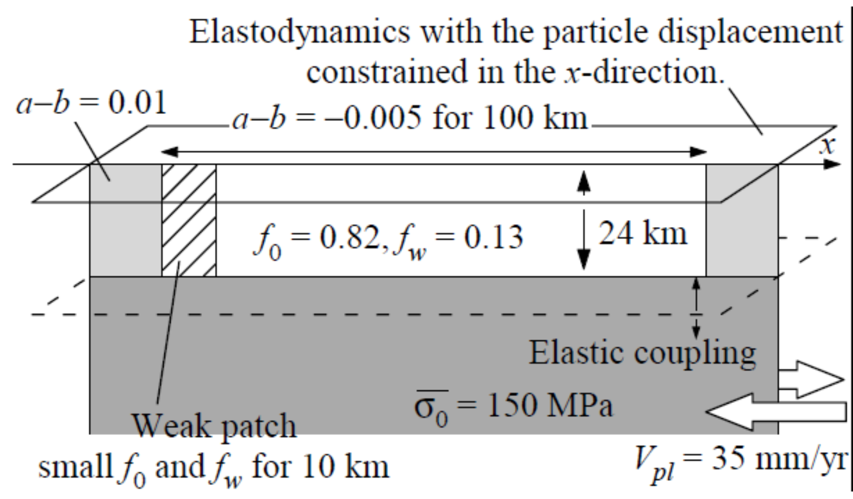


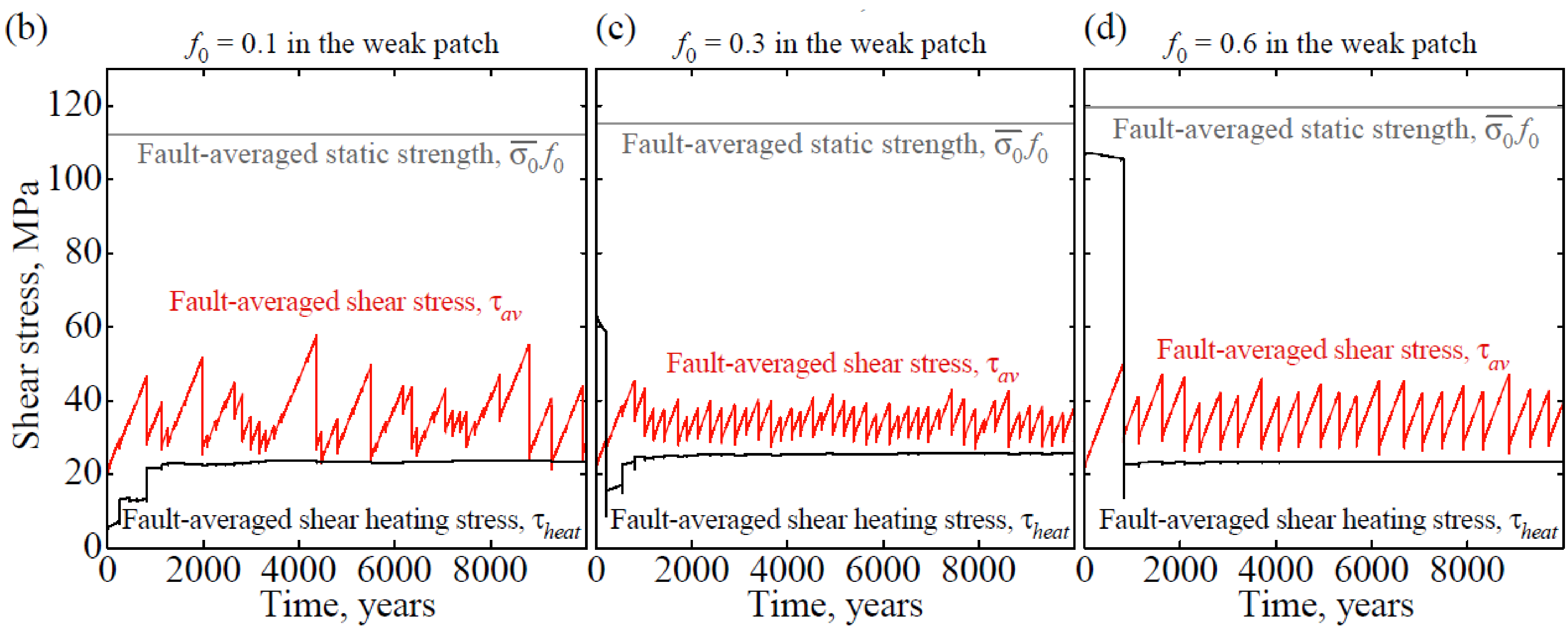
Ruptures propagate as short-duration narrow slip pulses

(as observed, e.g. Heaton, 1990)

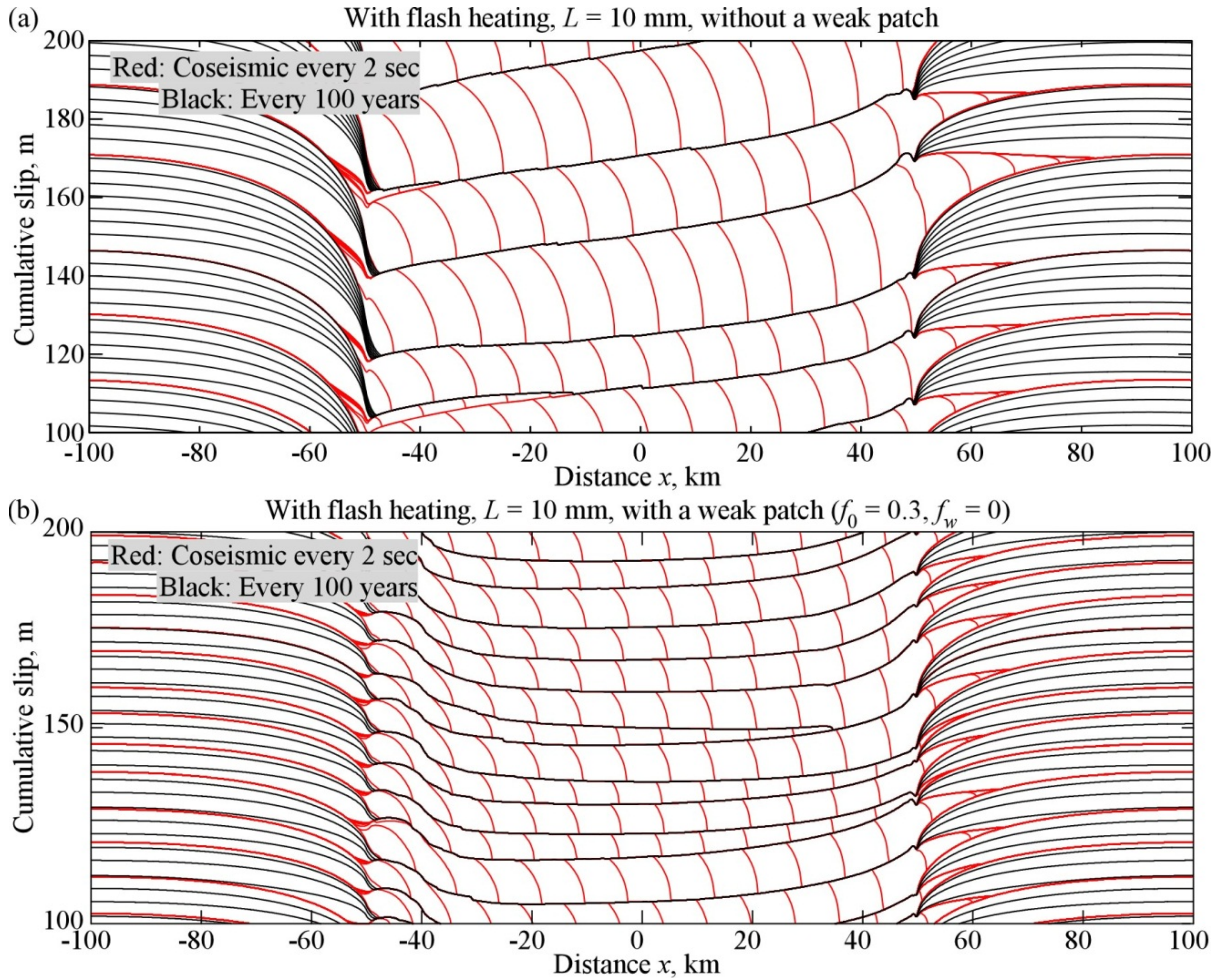


Characteristics of nucleation spots affect the overall fault dynamics

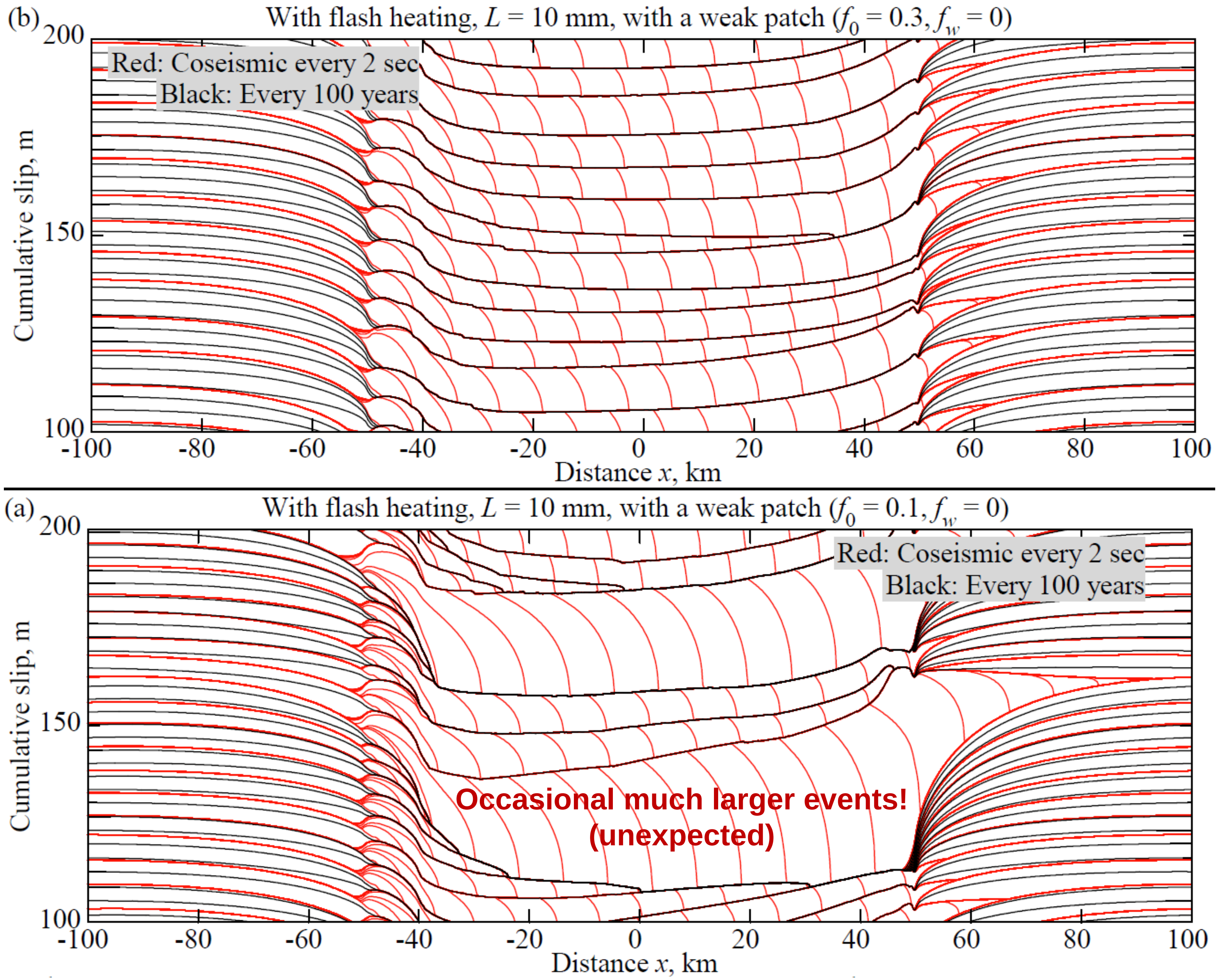




No weak spot (f_0 = 0.82 everywhere) vs. with weak spot (f_0 = 0.3)



With weak spot (f_0 = 0.3) vs. with even weaker spot (f_0 = 0.1)

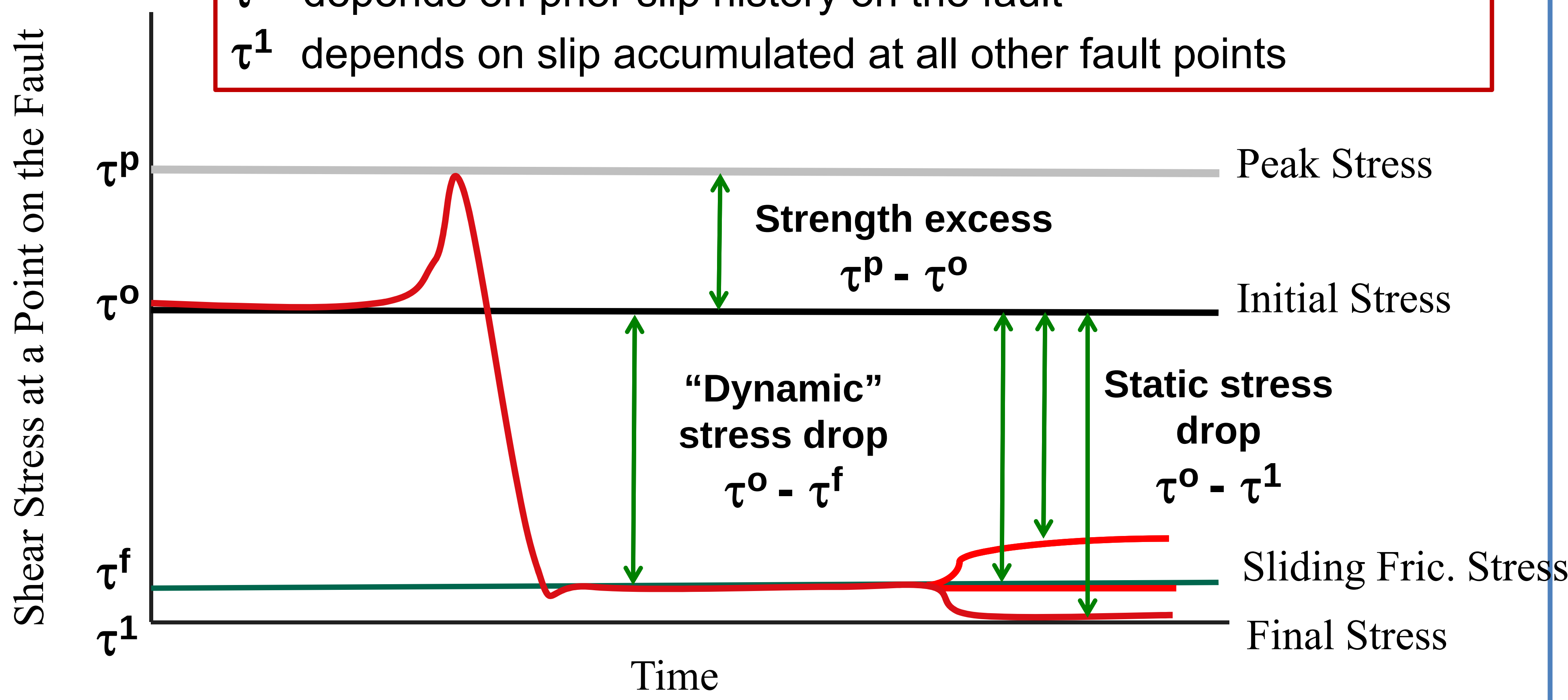


What controls local strength/stress changes, and how are they related to fault behavior (quakes, slow slip)?

τ^p, τ^f may depend both on variations in fault material properties AND details of the rupture itself (slip rate, its history, etc)

τ^o depends on prior slip history on the fault

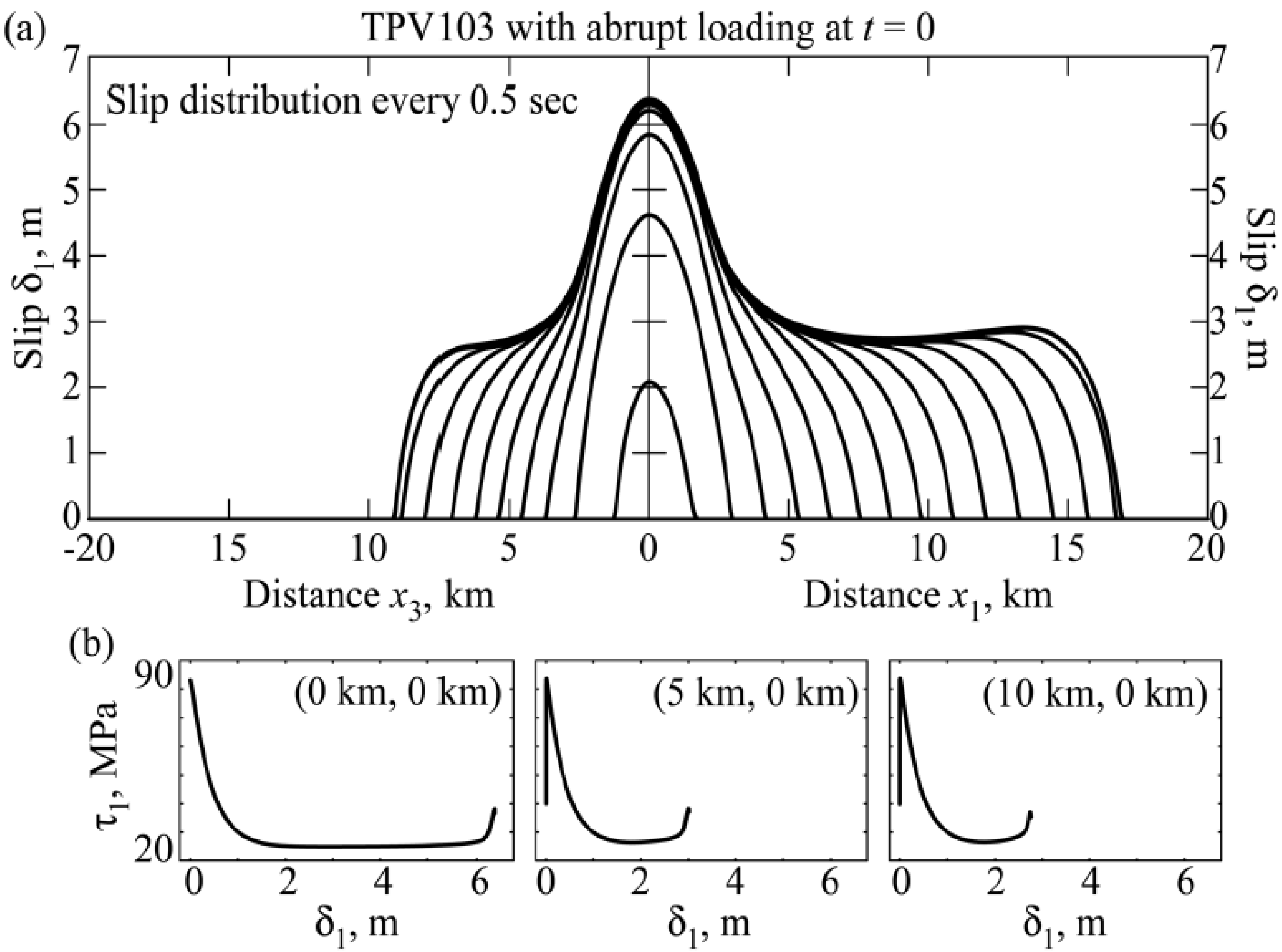
τ^1 depends on slip accumulated at all other fault points



Modified from Ralph Archuleta

Comment on relation between local stress changes and energy budget

Noda and Lapusta (JAM, 2012), Noda, Lapusta, and Kanamori, in revision



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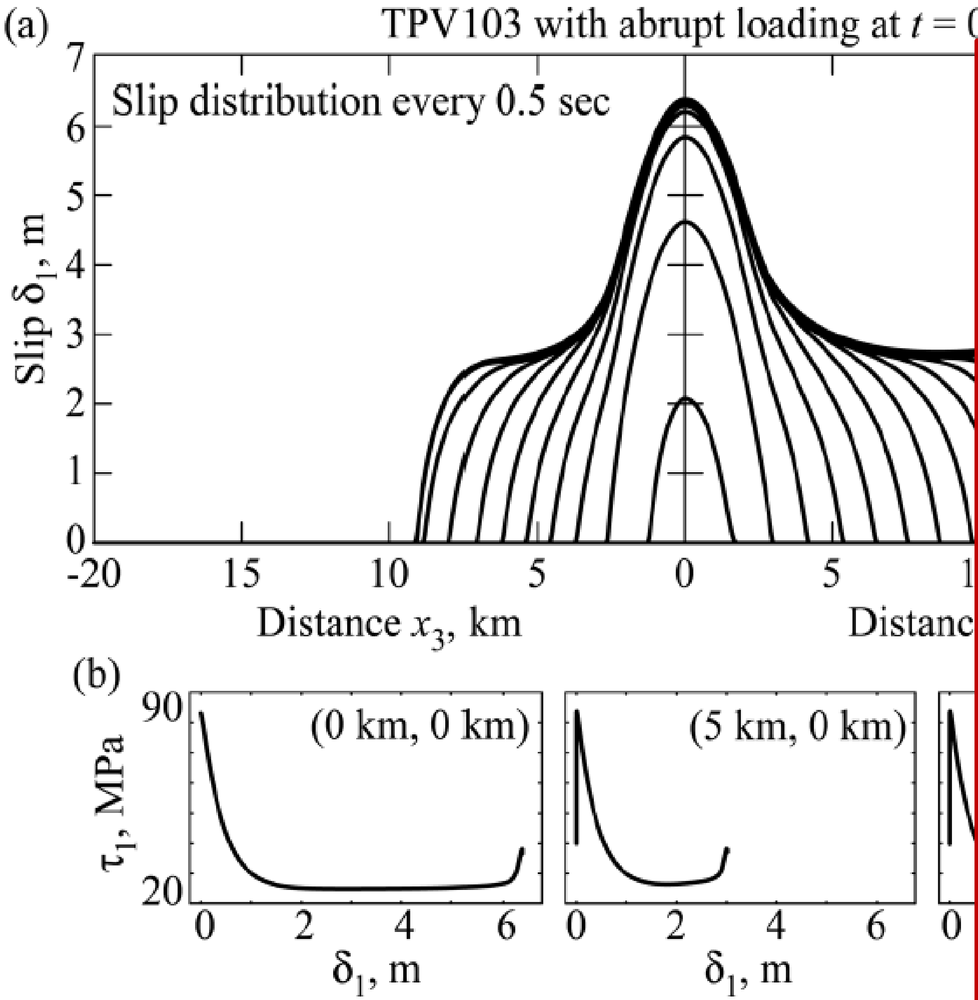
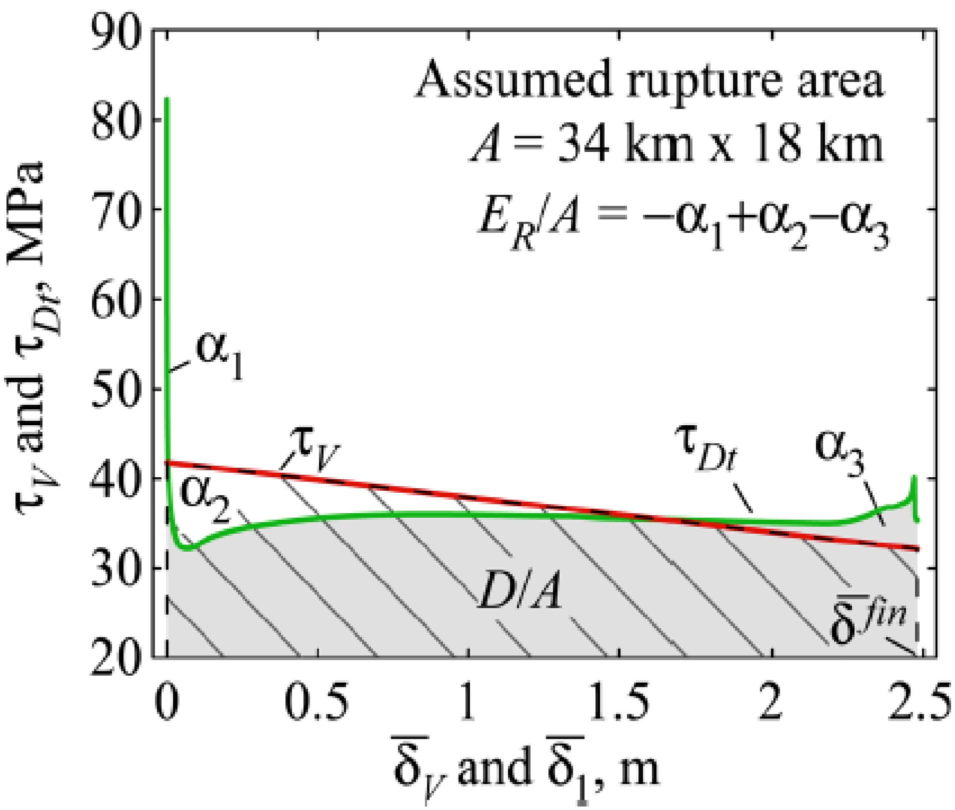
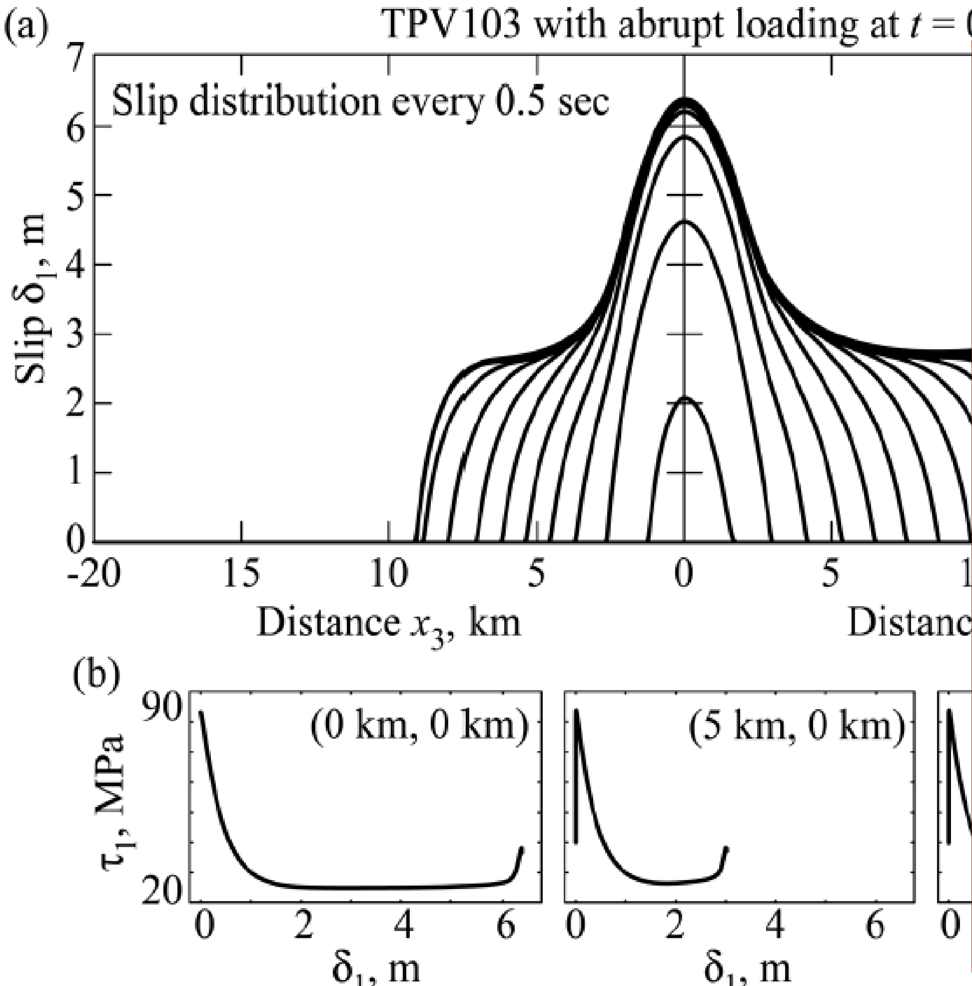


Averaging dissipation with time

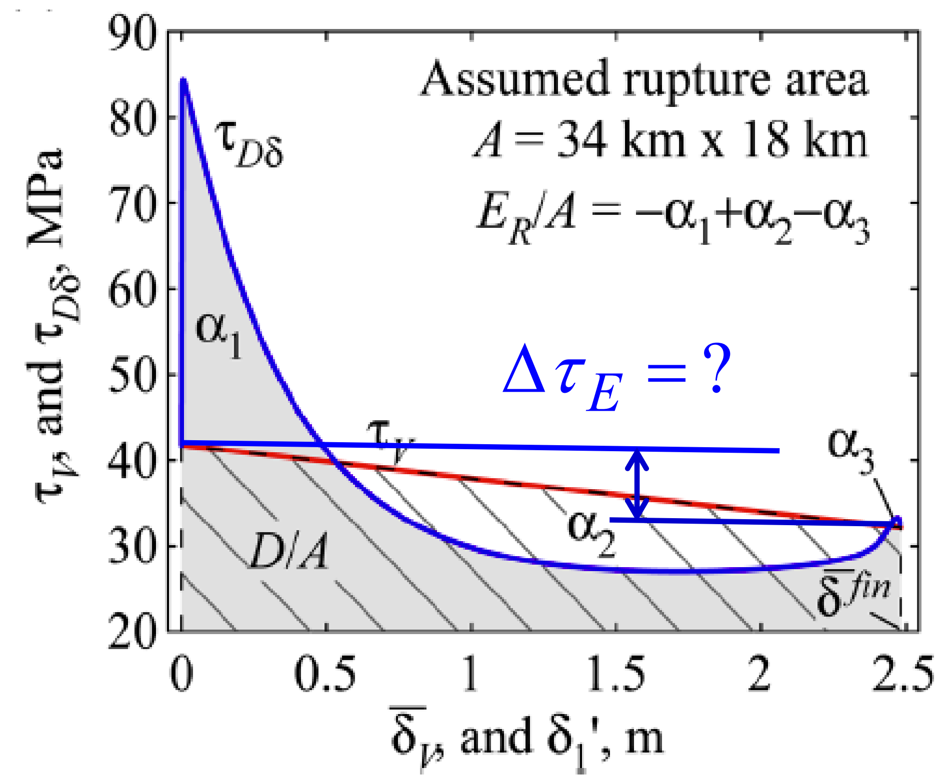


Comment on relation between local stress changes and energy budget

Noda and Lapusta (JAM, 2012), Noda, Lapusta, and Kanamori, in revision



Averaging dissipation with slip



$$\tau_{D\delta}(\delta') = \frac{\int_S \tau(\delta', \boldsymbol{x})\delta^{\text{fin}}(\boldsymbol{x})dS}{\int_S \delta^{\text{fin}}(\boldsymbol{x})dS} \Rightarrow \Delta\tau_E = \frac{\int_S [\tau^{\text{o}}(\boldsymbol{x}) - \tau^{1}(\boldsymbol{x})]\delta^{\text{fin}}(\boldsymbol{x})dS}{\int_S \delta^{\text{fin}}(\boldsymbol{x})dS} \geq \Delta\tau_M = C\frac{M_0}{r^3}$$

$$0 \leq \delta' \leq \bar{\delta}^{\text{fin}}$$

Stress drop from energy diagram , weighted by final slip distribution

Stress drop from seismic moment, weighted by elliptic slip distribution (Madariaga, 1979)

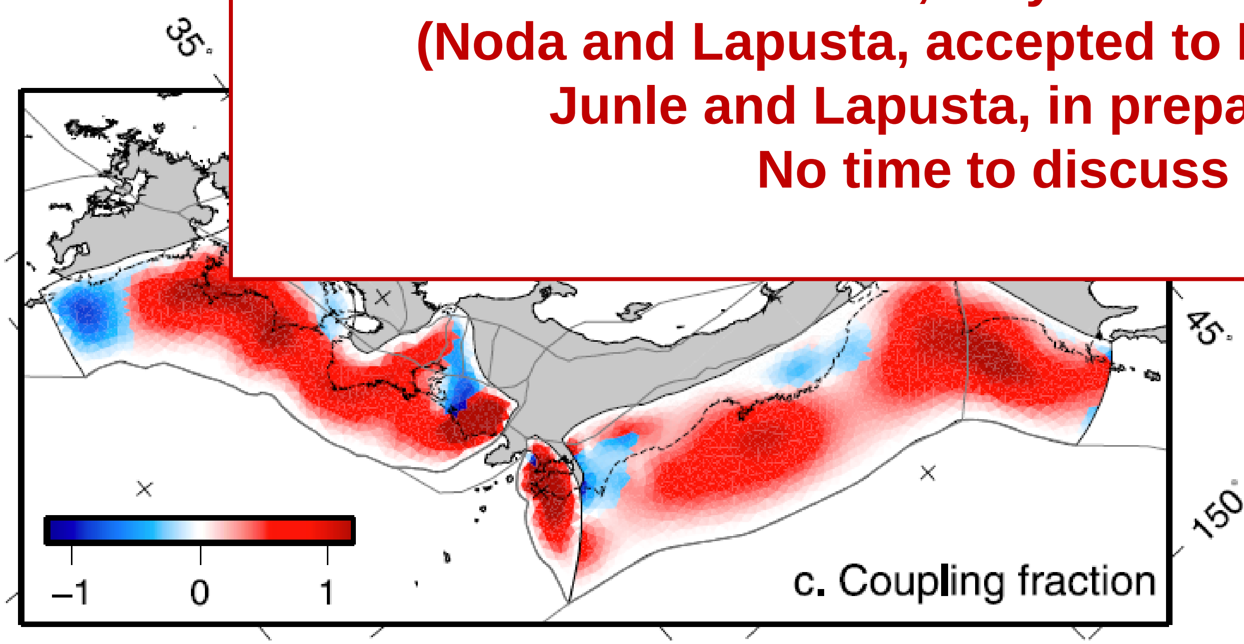
Is fault separation into stable/unstable areas persistent? (Convenient picture but potentially too simplified)

Slowly moving (creeping) areas
⇔ rate-strengthening friction
⇔ "barriers" to earthquake rupture

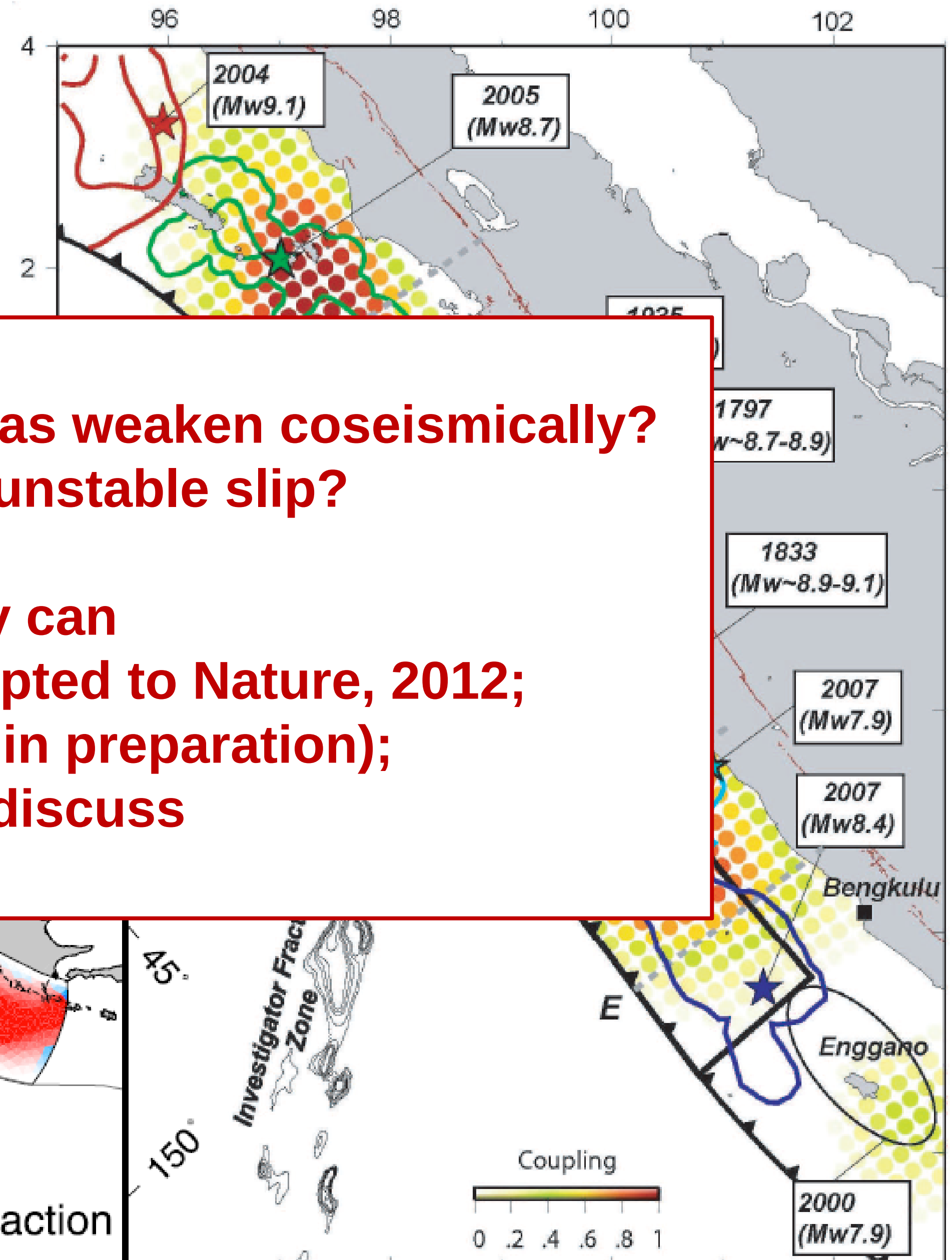
Locked
⇔ rate
⇔ "sei

What if rate-strengthening areas weaken coseismically?
Can they sustain unstable slip?

Yes, they can
(Noda and Lapusta, accepted to Nature, 2012;
Junle and Lapusta, in preparation);
No time to discuss

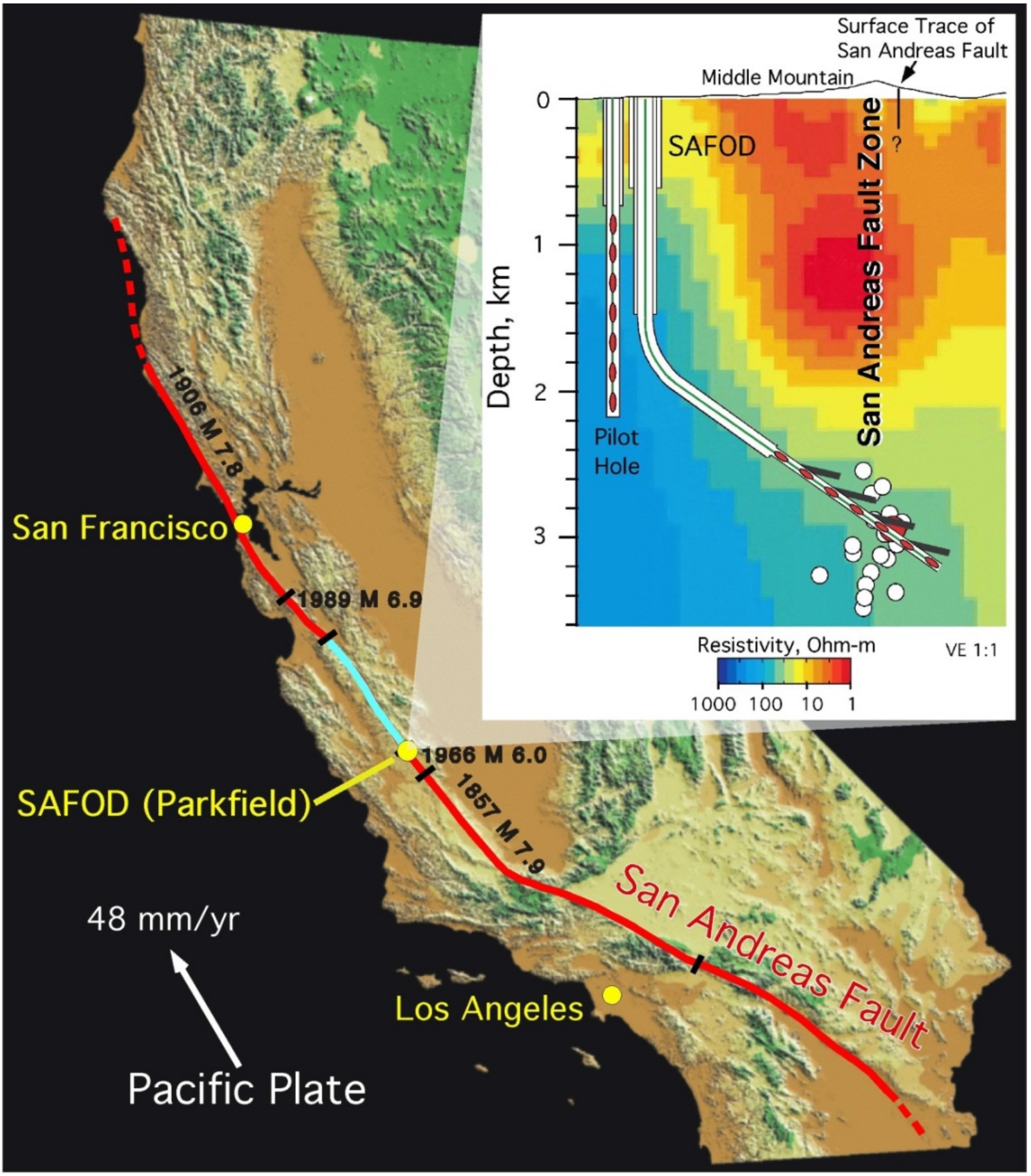


Japan (Loveless and Meade, 2010)



Sumatra (Chlieh et al., 2008)

Can a large earthquake propagate through the creeping section of San Andreas fault?



Hickman, Zoback, Ellsworth, 2004

Yes if the creeping segment has:

- Velocity-strengthening friction at interseismic slip rates
- Co-seismic weakening at seismic slip rates (e.g., due to shear heating)

Need more field, laboratory, and theoretical studies.

Potential indirect evidence of enhanced co-seismic weakening in small repeating earthquakes in the creeping section

(work with Ting Chen)

Conclusions

- Shear zones that host large earthquakes potentially experience ***extreme coseismic weakening***, due to shear heating and other physical mechanisms.
- Models that incorporate such weakening can reconcile
  -- the apparent weakness of mature faults,
  -- their high quasi-static strength,
  -- typical stress drops for large events,
  -- low heat generation on mature faults,
  -- compact short-duration pulse-like rupture mode.
- In such fault models, **availability and nature of spots favorable to earthquake nucleation** is important.
- Creeping (“decoupled”) fault patches may be susceptible to coseismic weakening, joining earthquakes to produce large coseismic slip.
- A model with such patch qualitatively explains observations on a range of temporal scales for two well-studied earthquakes (1999 Chi-Chi & 2011 Tohoku-Oki).
- Earthquakes may penetrate below the traditionally defined seismogenic zone due to coseismic weakening.