Earthquake nucleation

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Caltech Seismolab

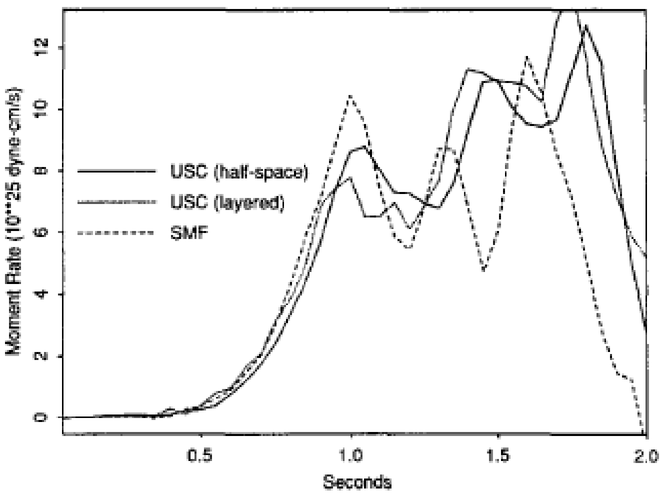
How do earthquakes start?

Do small and large earthquakes start differently?

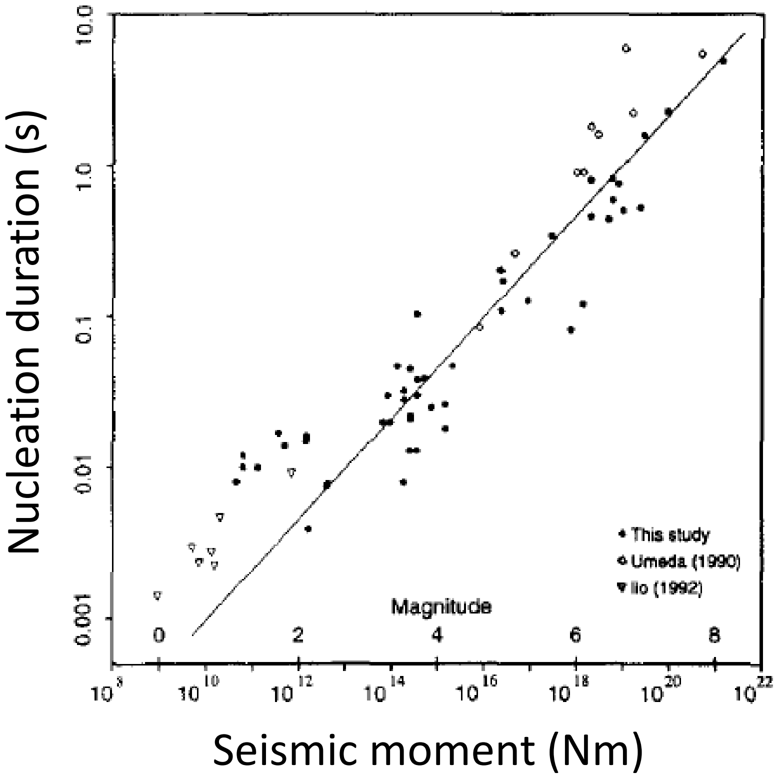
Predictive value of earthquake onset and foreshock sequences?

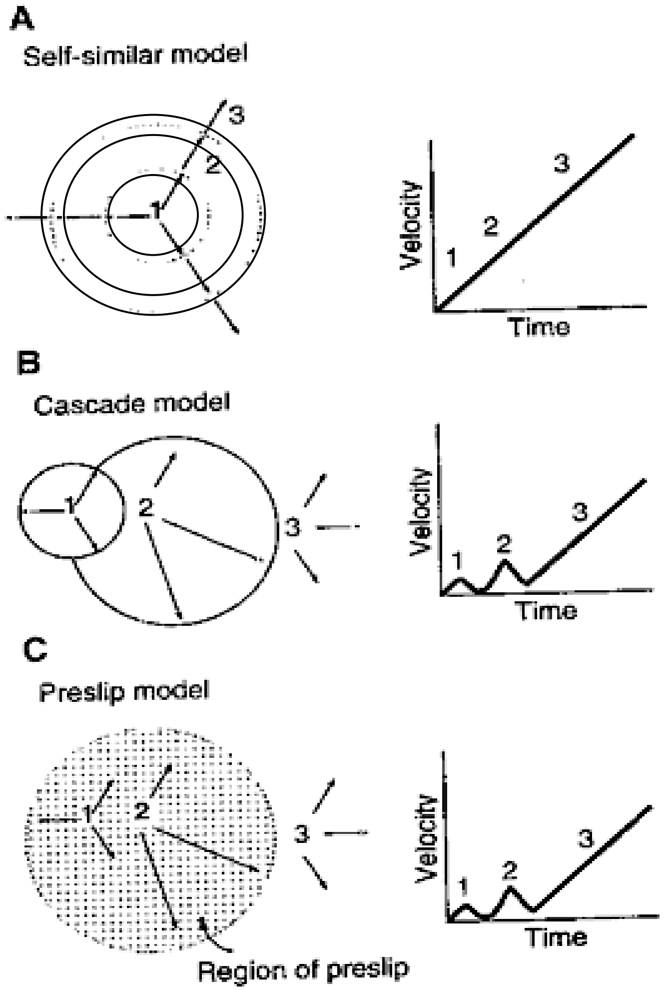
- Seismological observations: seismic nucleation phase, foreshock sequences
- Laboratory observations
- Friction laws and earthquake nucleation
- Further implications: stress drop, recurrence time, ~~seismicity rate~~, ~~tremors~~

Seismological observations



Ellsworth and Beroza (1995)
Beroza and Ellsworth (1996)

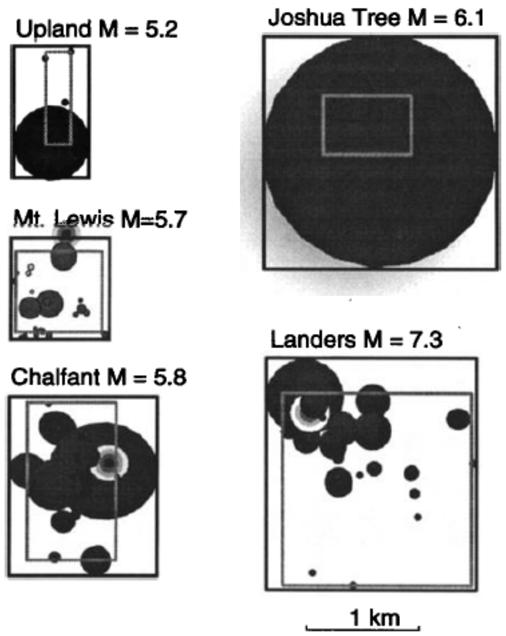


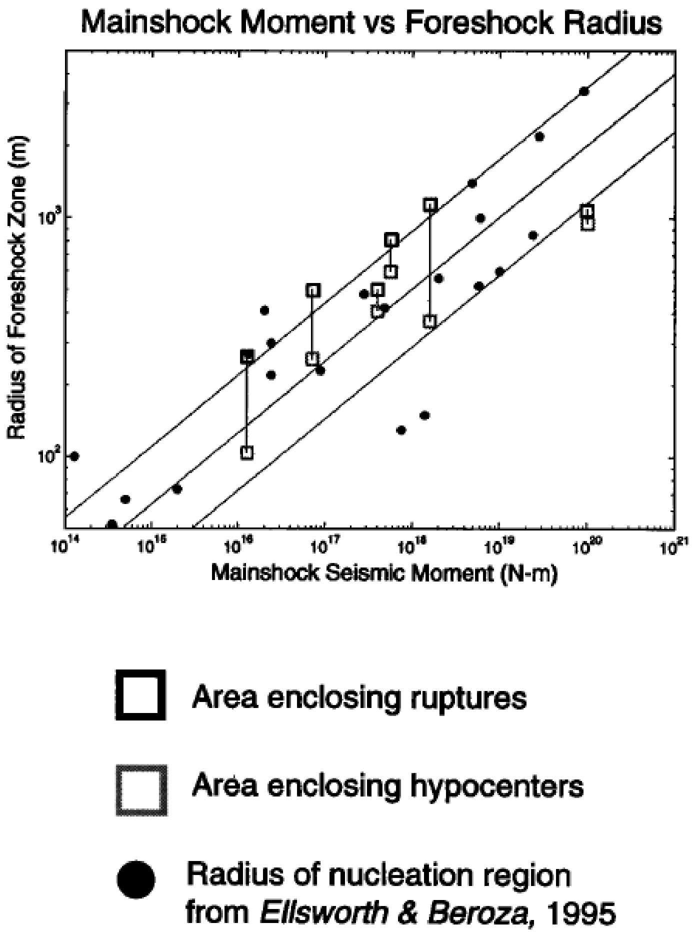


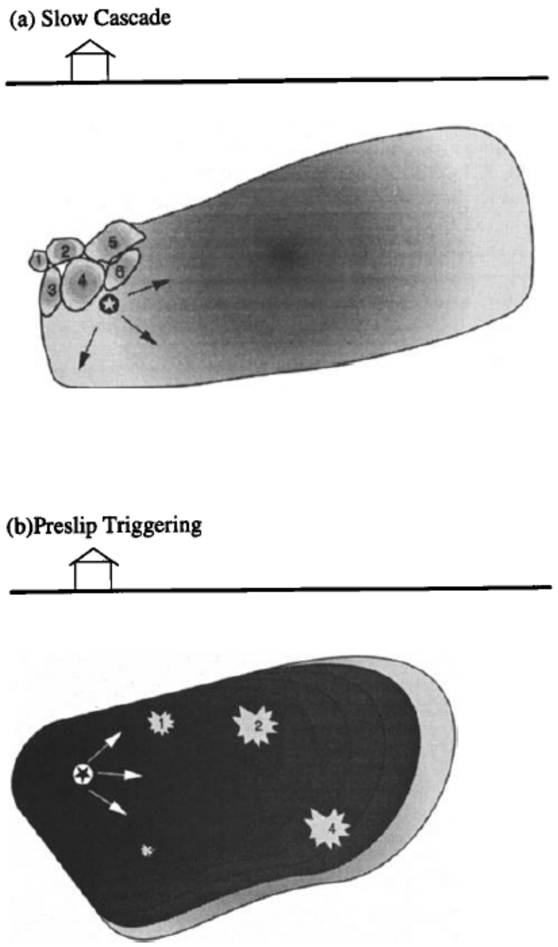
Seismological observations

Foreshock sequences

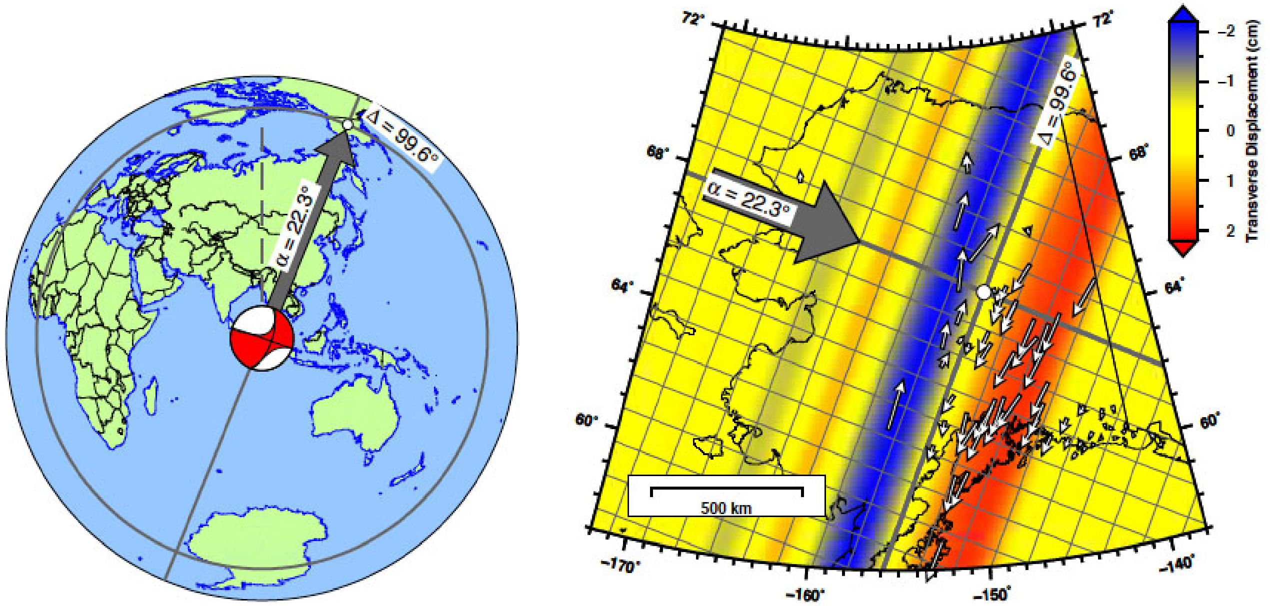
Dodge et al (1996)

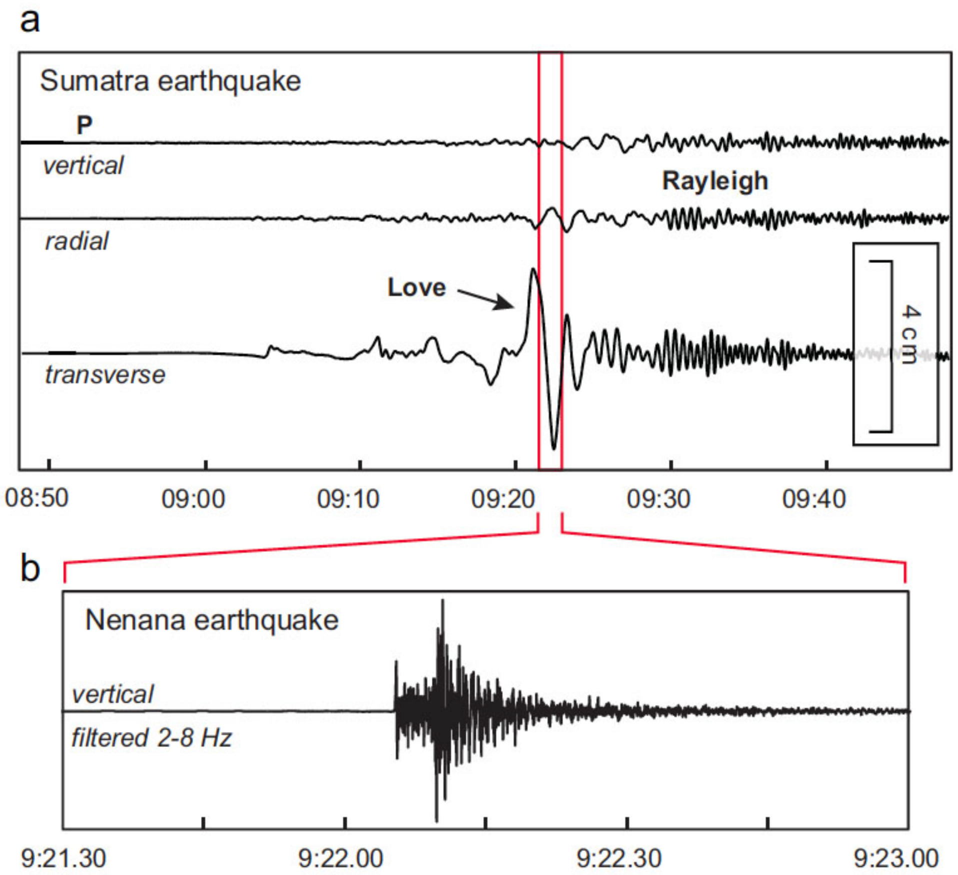






Seismological observations

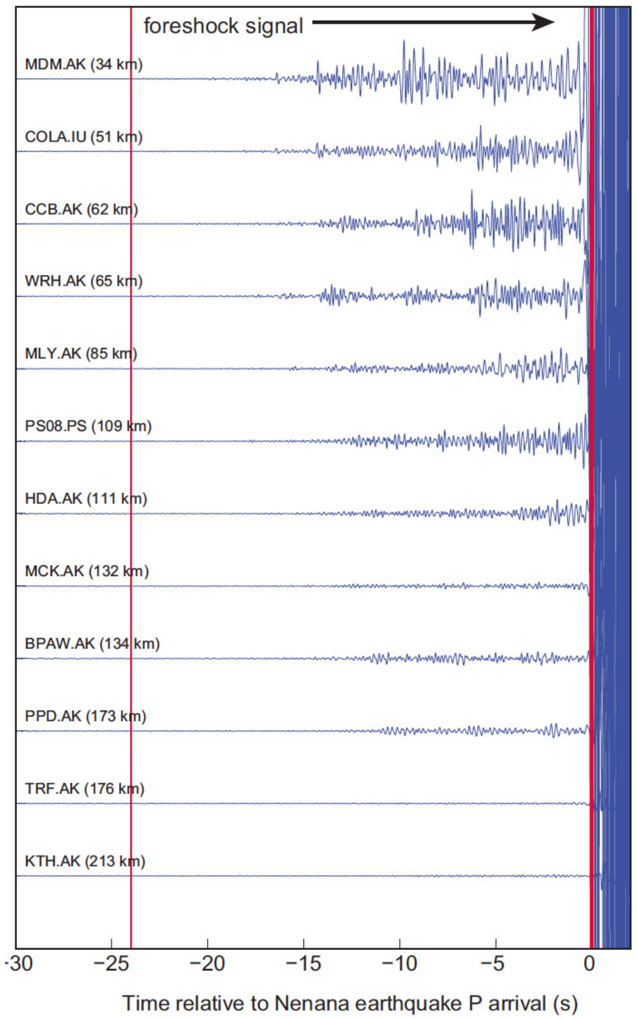


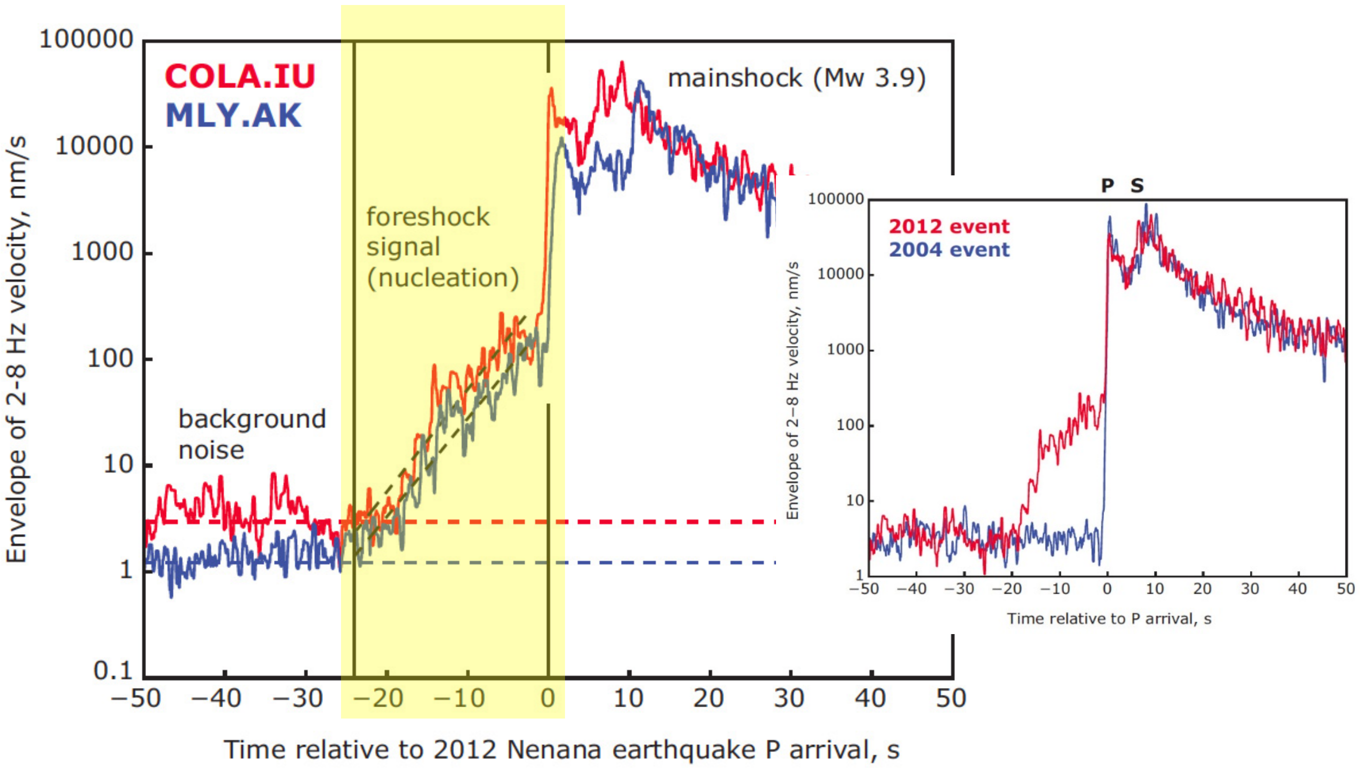


A Mw3.9 earthquake in Alaska triggered by Love waves from the April 11, 2012 Mw 8.6 Sumatra earthquake

Tape et al (2013)

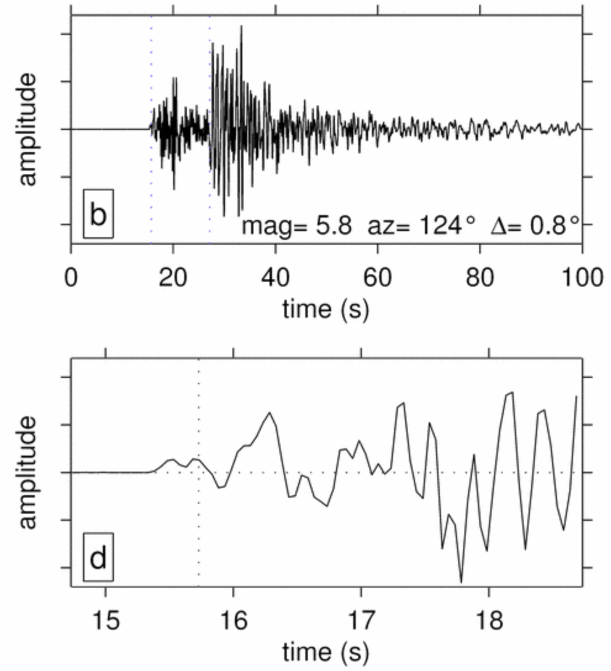
Seismological observations





Nucleation phase of the Mw3.9 Alaska triggered earthquake
Tape et al (2013)

Seismological observations

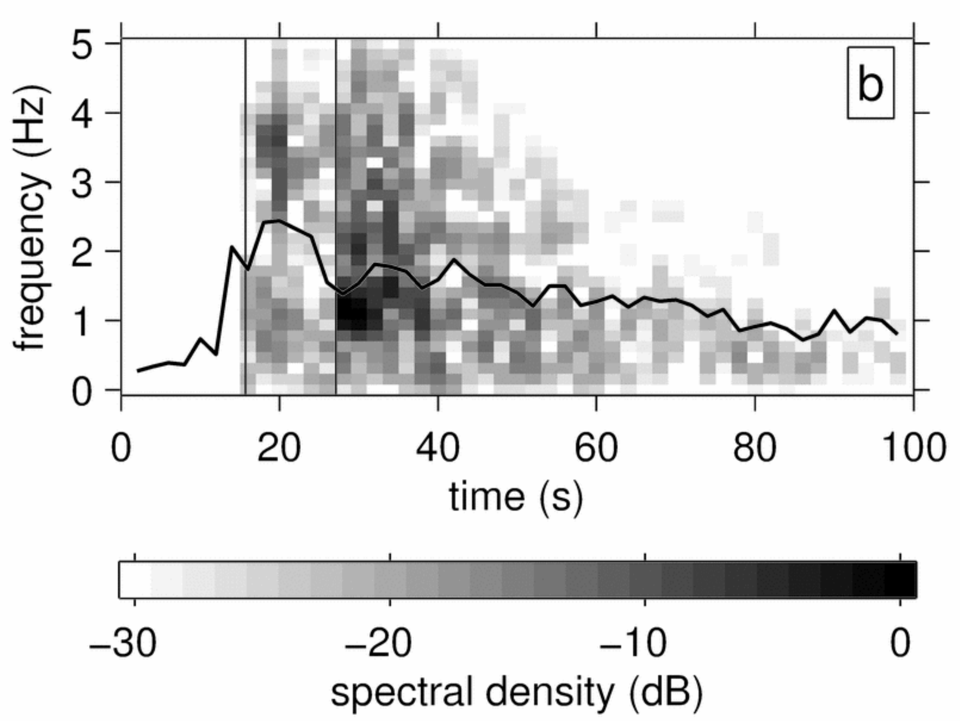


Simons et al (2006)

$$\tau_c^2 = 4\pi^2 \frac{\int_0^{\tau_0} u^2(t)\,\mathrm{d}t}{\int_0^{\tau_0} \dot{u}^2(t)\,\mathrm{d}t}$$

$1/\tau_c \sim$ instantaneous frequency

Nakamura (1988)

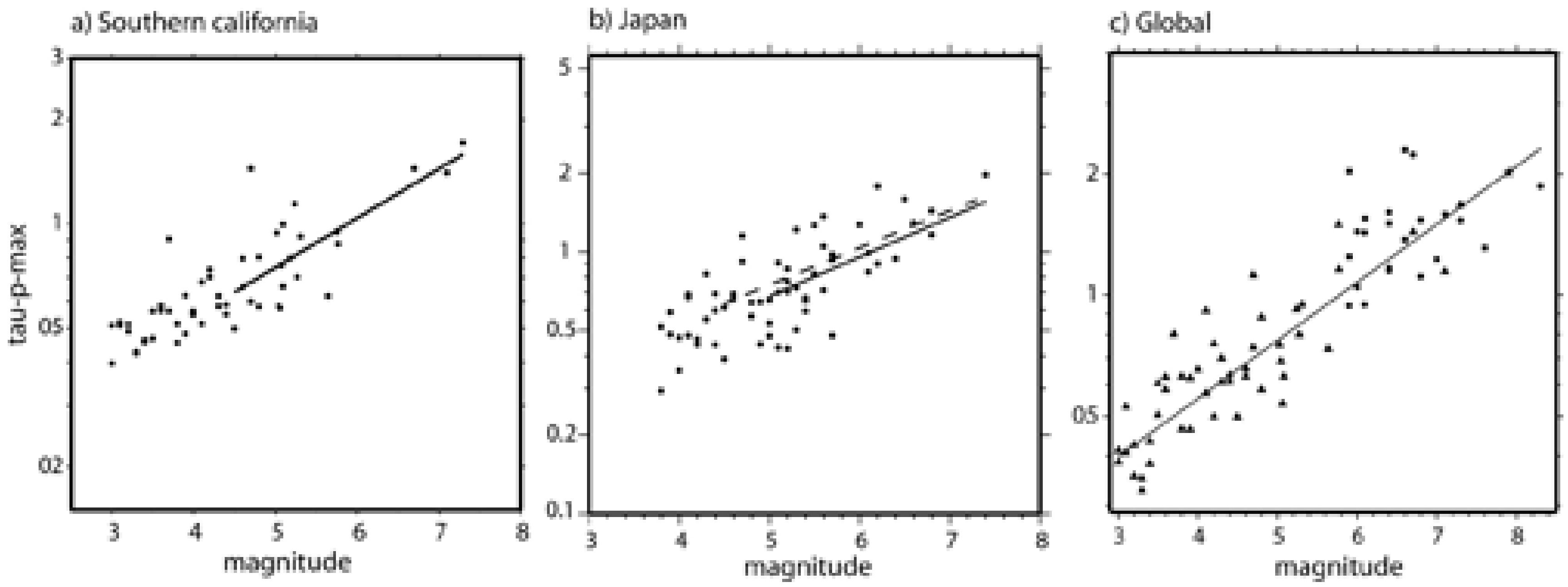


Seismological observations

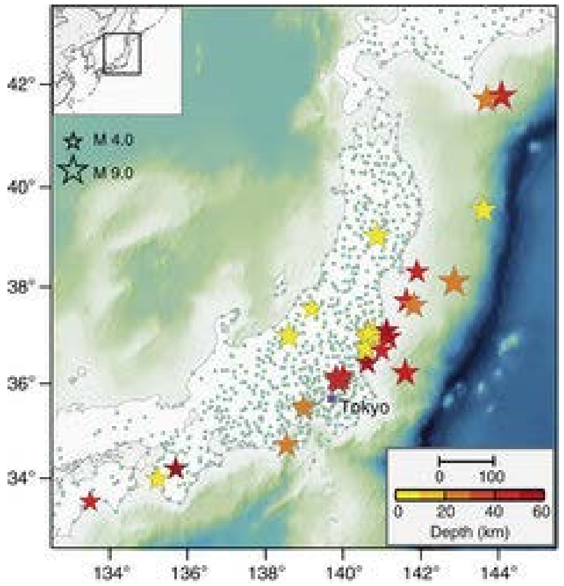
$$\tau_c^2 = 4\pi^2 \frac{\int_0^{\tau_0} u^2(t)\mathrm{d}t}{\int_0^{\tau_0} \dot{u}^2(t)\mathrm{d}t}$$

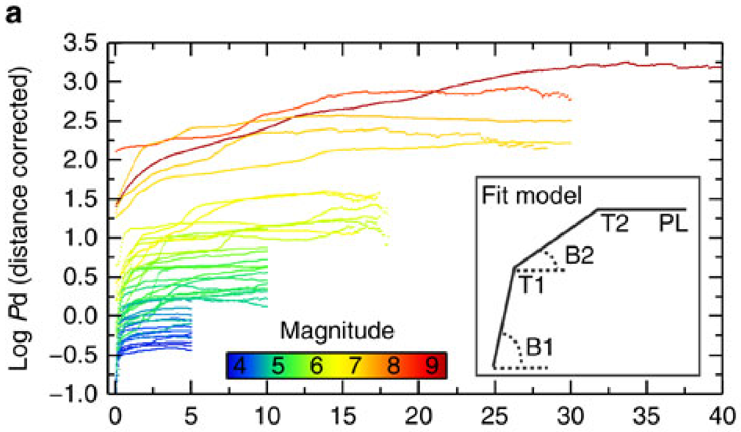
Magnitude dependence of early dominant period

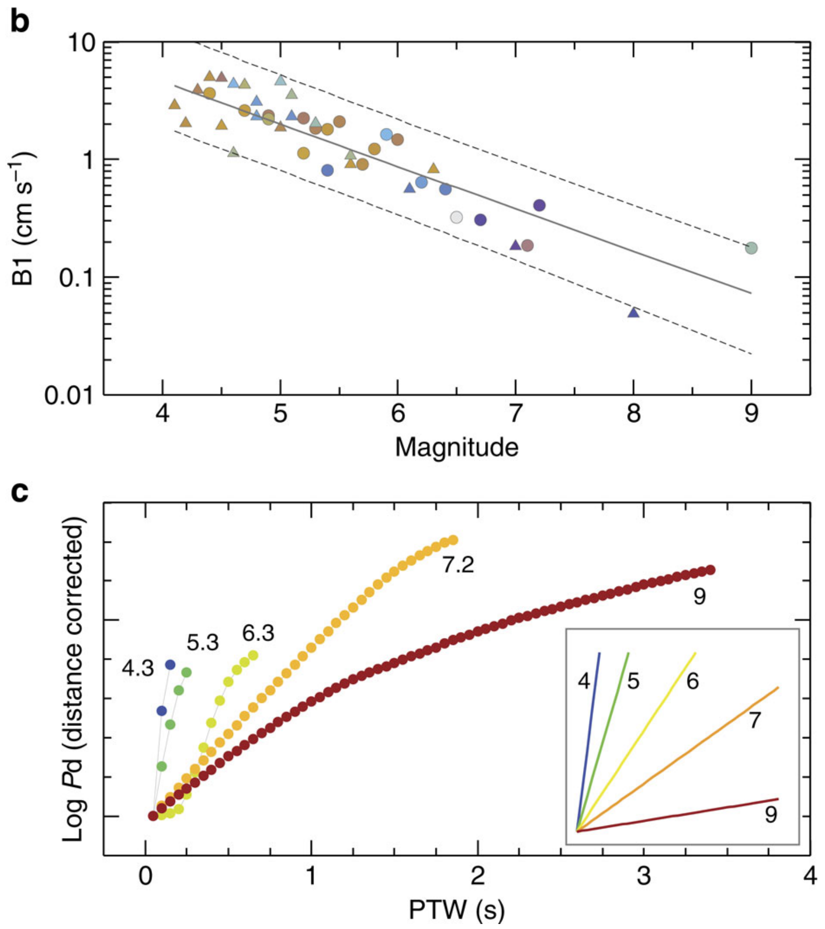
Allen and Kanamori (2003)



Seismological observations



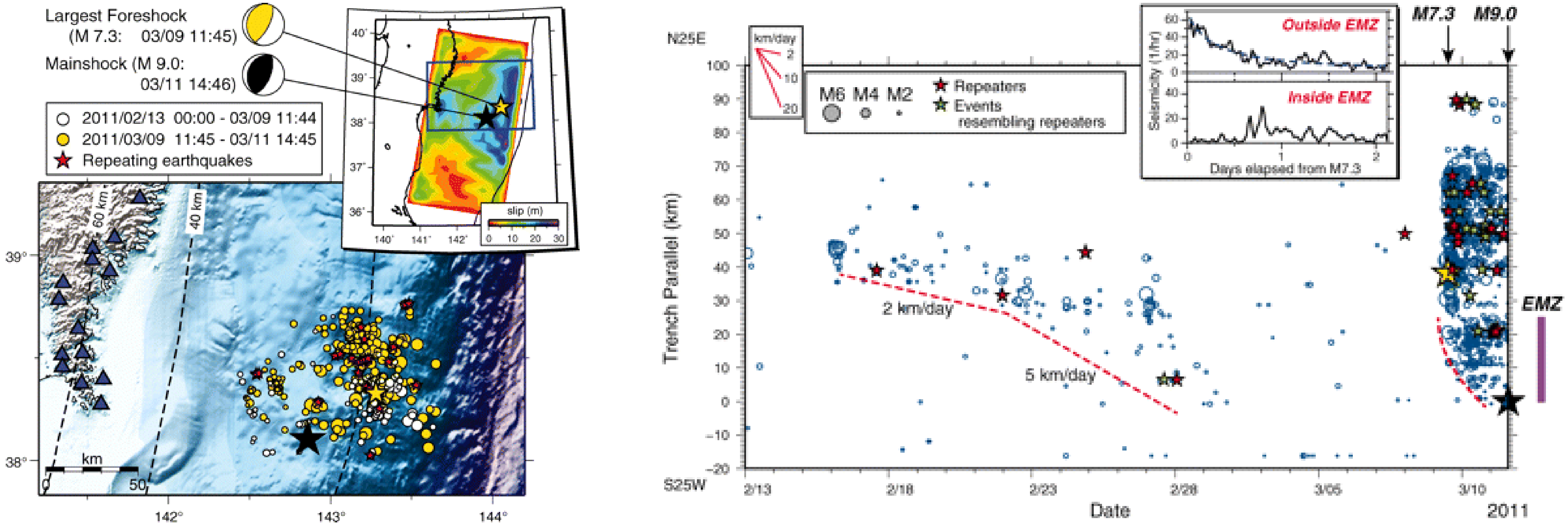




Peak ground displacement (Pd) grows exponentially.
Growth rate depends on magnitude

Colombelli et al (2014)

Seismological observations

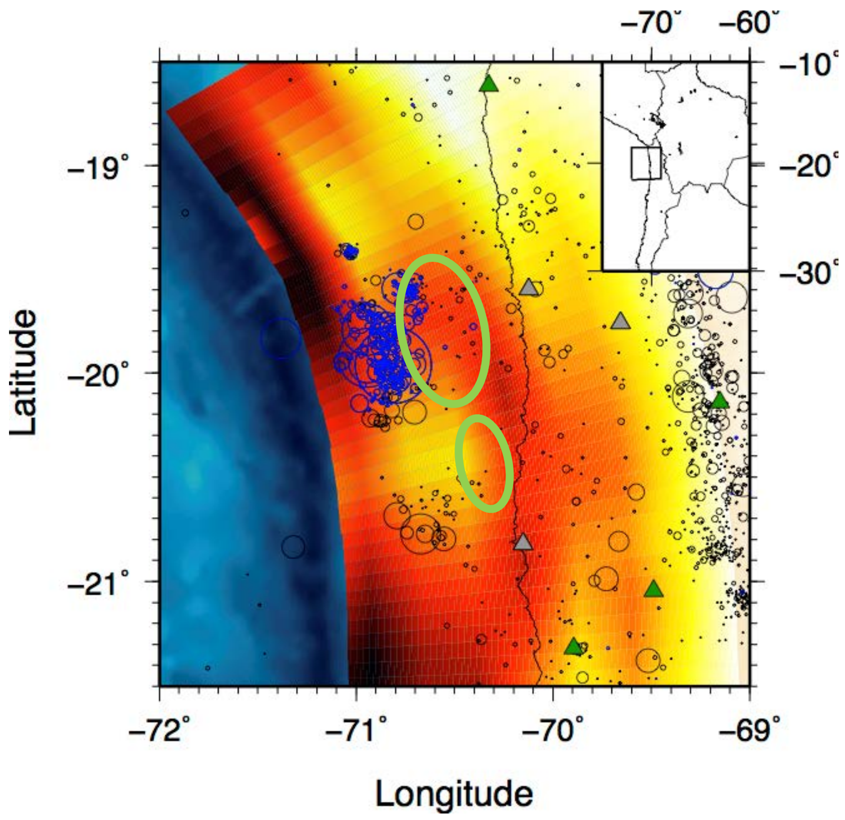


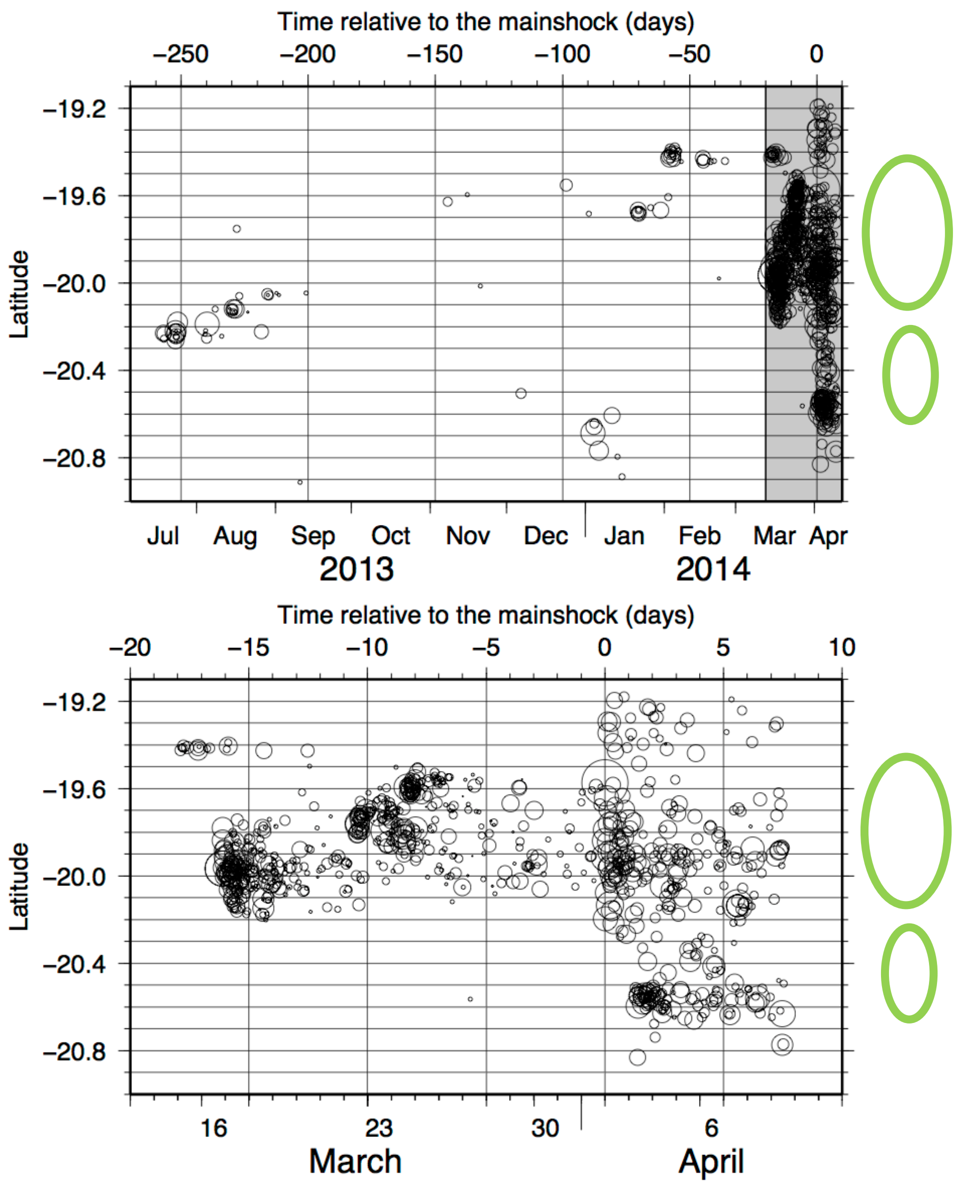
Foreshock sequence of the 2011 Tohoku earthquake

Kato et al (2012)

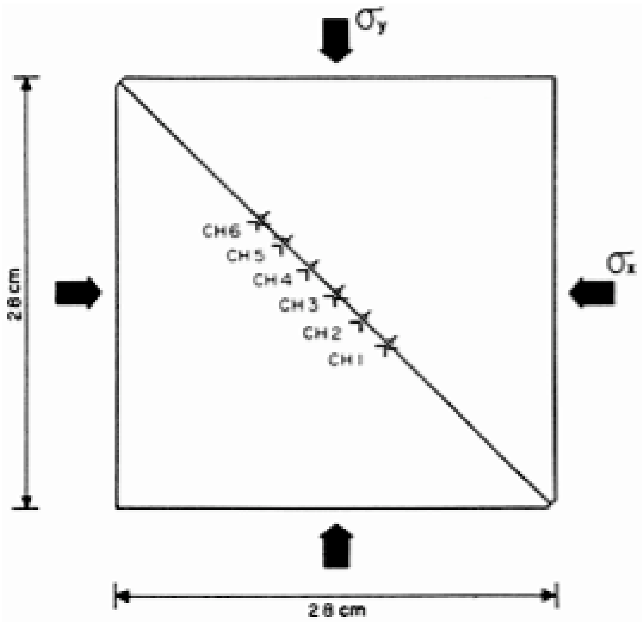
Seismological observations

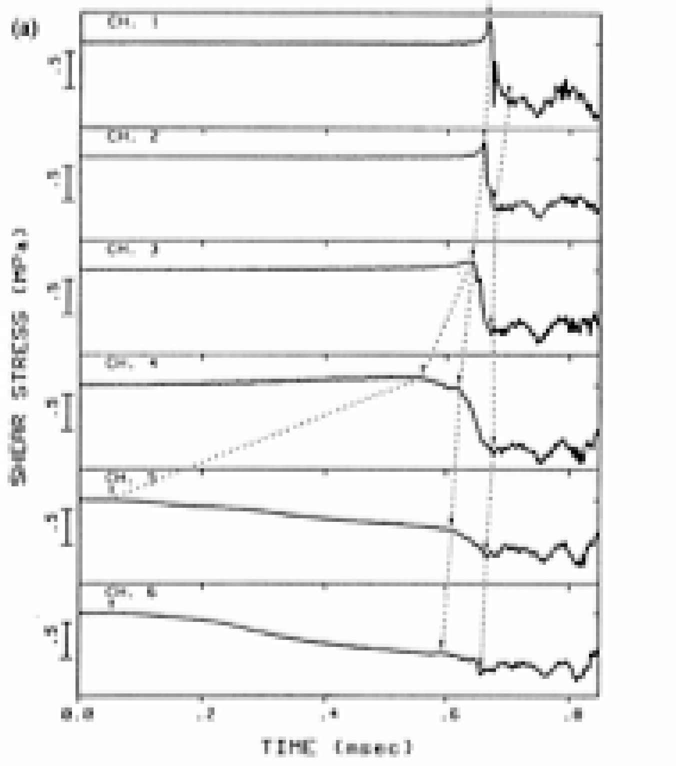
Foreshocks of the 2014 Iquique (Chile) earthquake

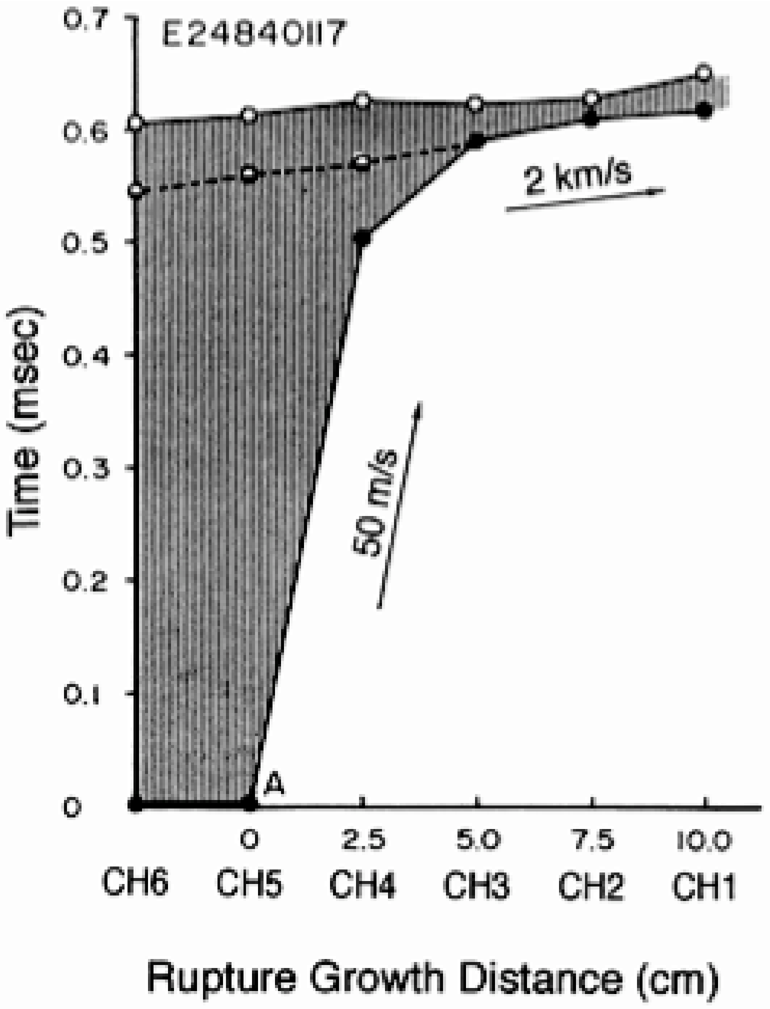




Laboratory experiments

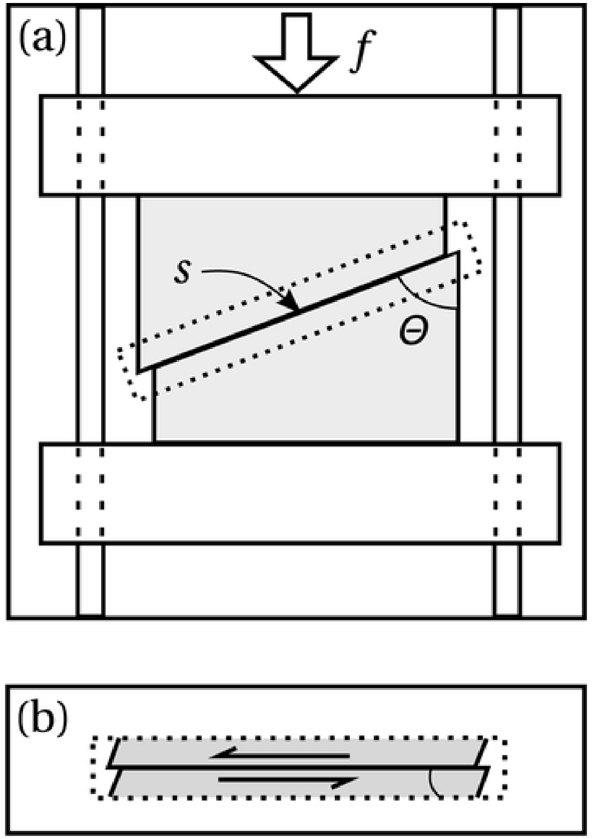


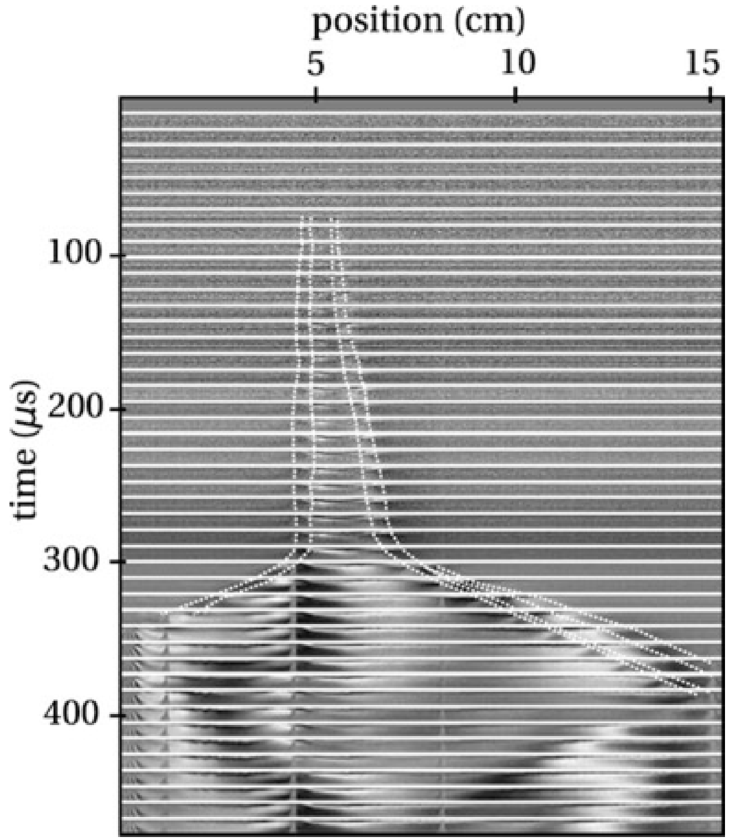




Ohnaka (1990)

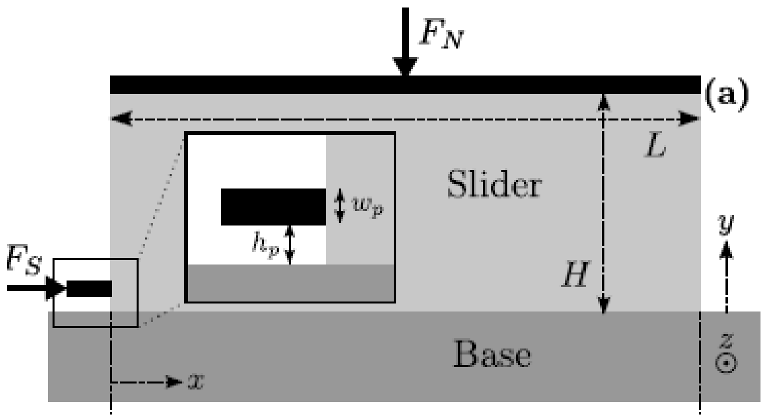
Laboratory experiments

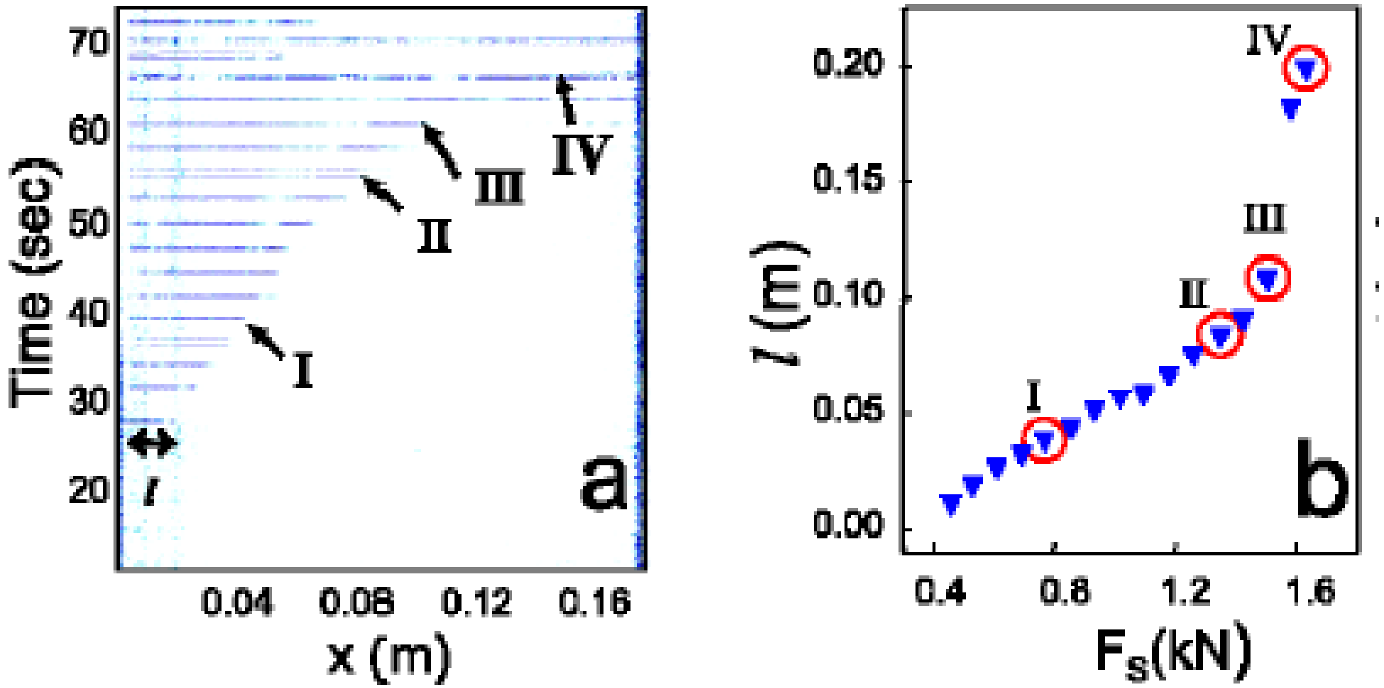




Nielsen et al (2010)

Laboratory experiments

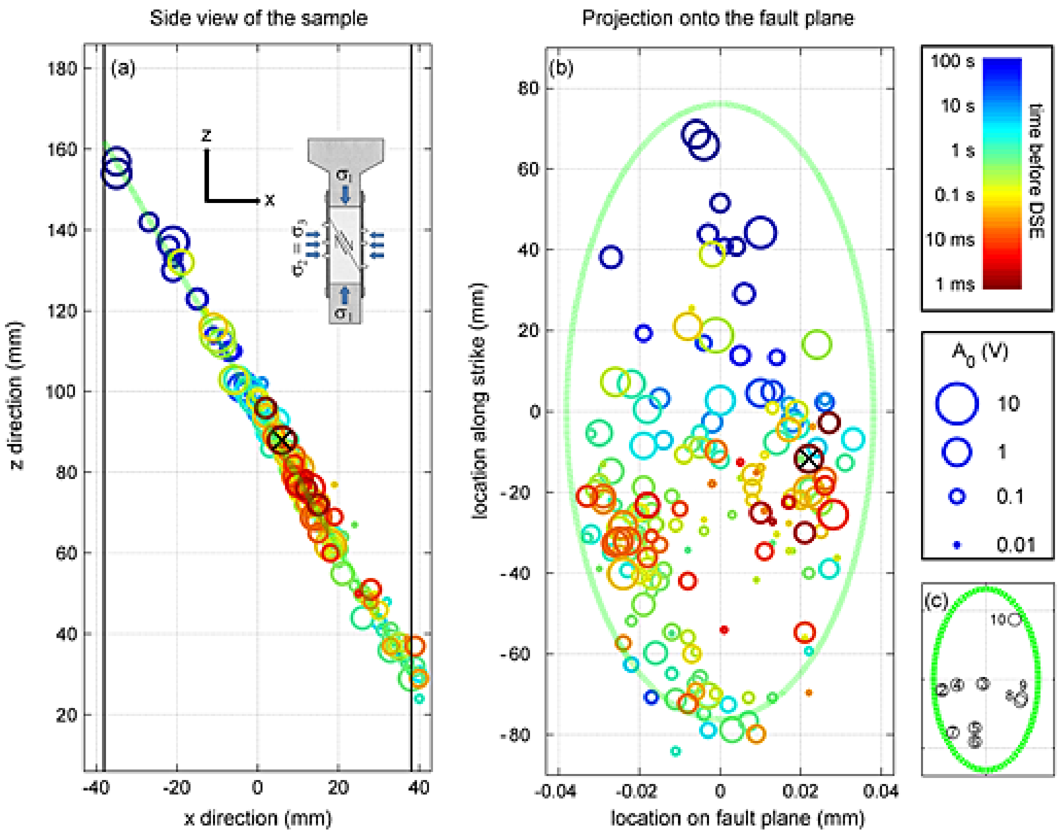




Laboratory foreshocks

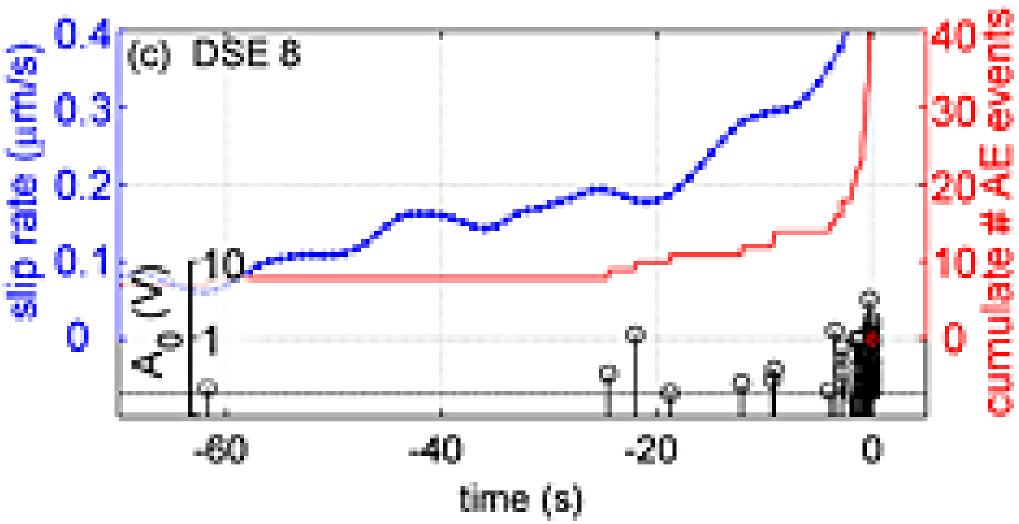
Rubinstein et al (2007)

Laboratory experiments



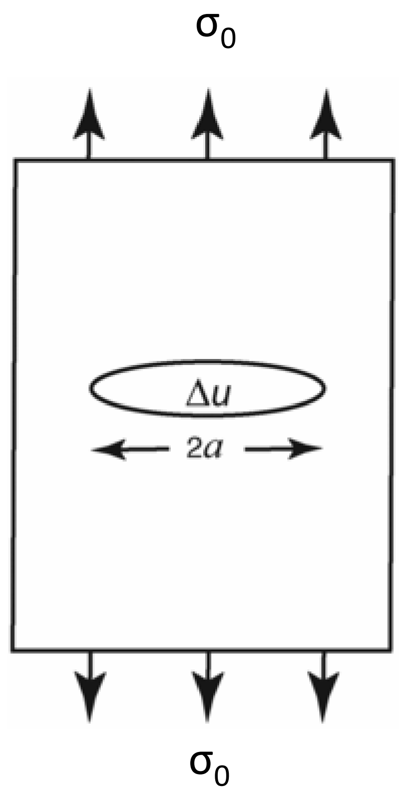
Foreshocks promoted by aseismic slip

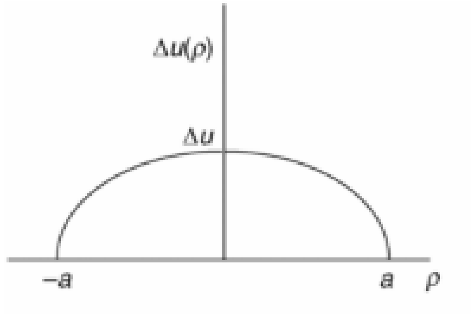
McLaskey and Kilgore (2014)

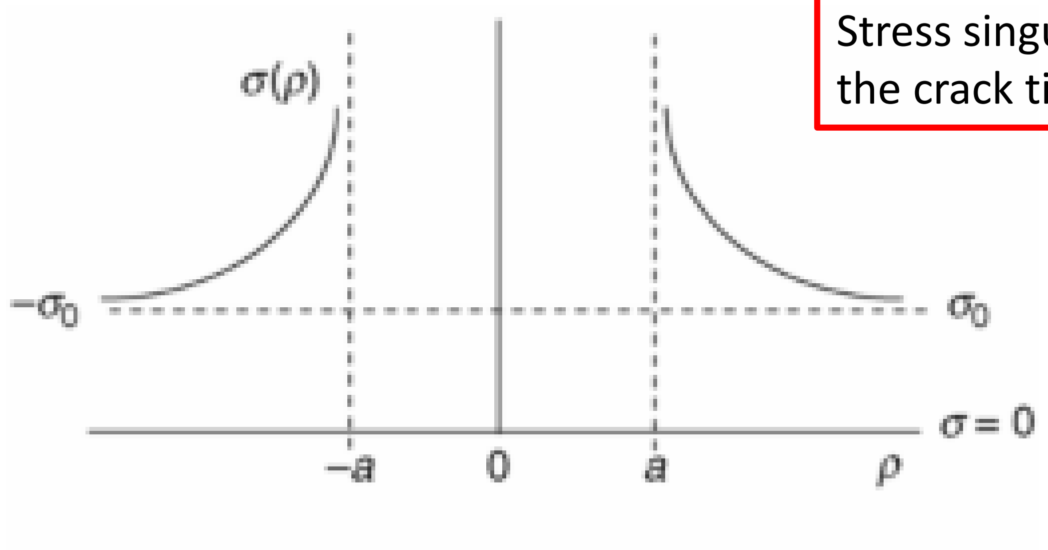


Fracture mechanics

Static equilibrium in a linear elastic solid with a crack and boundary conditions:
$\sigma(x) = \sigma_0$ for $|x| > a$ and
$\sigma(x) = 0$ for $|x| < a$.

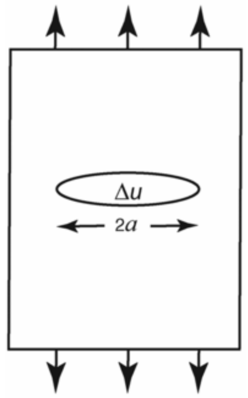


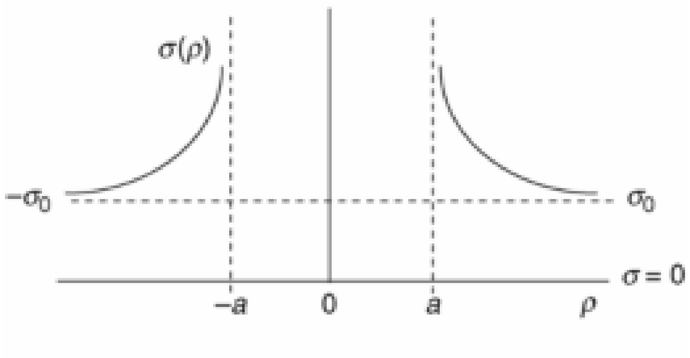




Stress singularity at the crack tips

Fracture mechanics





Stress singularity at the crack tips.

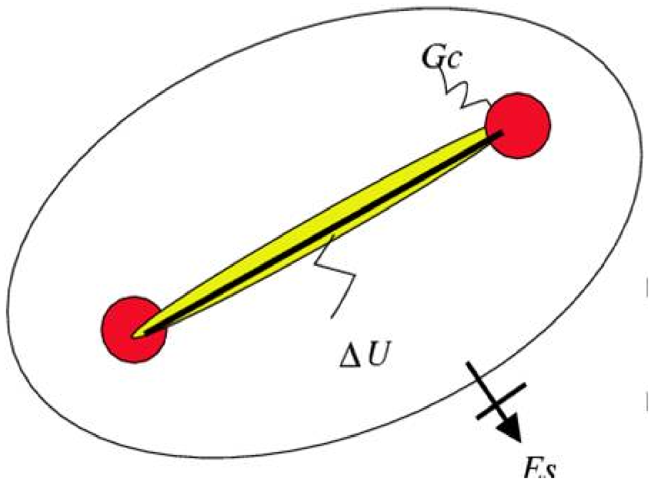
Asymptotic form: $\boxed{\sigma = \frac{K_I}{\sqrt{2\pi r}}} + O(\sqrt{r})$

where r is the distance to a crack tip,
K is the ***stress intensity factor***
and Δσ the stress drop (here, σ_0 - 0)

$$\boxed{K_I = \Delta\sigma\sqrt{a/2}}$$

In reality, stresses are finite: singularity accommodated by inelastic deformation.

Fracture mechanics



Energy release rate = energy flux to the crack tip per unit of crack advance:

$$G(a) = \frac{K(a)^2}{2\mu} = \pi \frac{\Delta\tau^2 a}{2\mu}$$

During quasi-static crack growth, this energy is dissipated at the crack tip into fracture energy

$$G(a) = G_c$$

During dynamic growth, $G(a) > G_c$
At rest, $G(a) \leq G_c$

Nucleation size

At the onset of rupture (critical equilibrium): $G_c = \pi \frac{\Delta\tau^2 a}{2\mu}$

→ earthquake initiation requires a minimum crack size (*nucleation size*)

$$\boldsymbol{a_c = \frac{2\mu G_c}{\pi \Delta\tau^2}}$$

($\mu \sim 30$ GPa, $\Delta\tau \sim 5$ MPa)

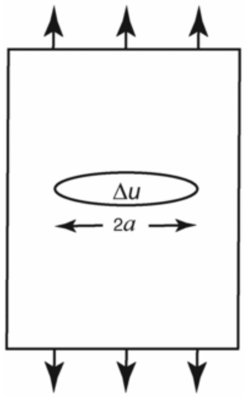
Estimates for large earthquakes: $G_c \sim 10^6$ J/m² → $a_c \sim 1$ km

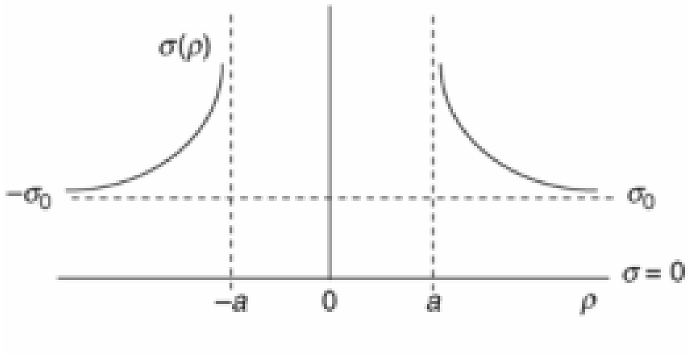
… how can M<4 earthquakes nucleate ?!

Laboratory estimates: $G_c \sim 10^3$ J/m² → $a_c \sim 1$ m (M -2)

→ **G_c scaling problem**

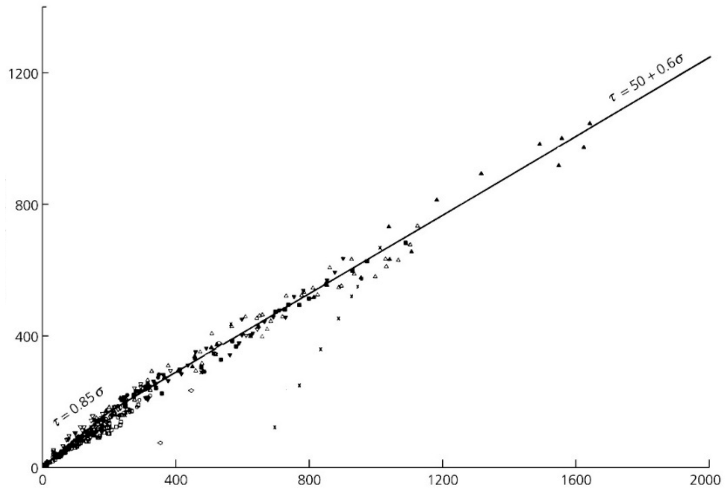
Limitations of fracture mechanics: Rock strength is finite





$$\sigma = \frac{K_{\mathrm{I}}}{\sqrt{2\pi r}}$$

Figure 5.7-10: Relation between shear stress and normal stress for frictional sliding.

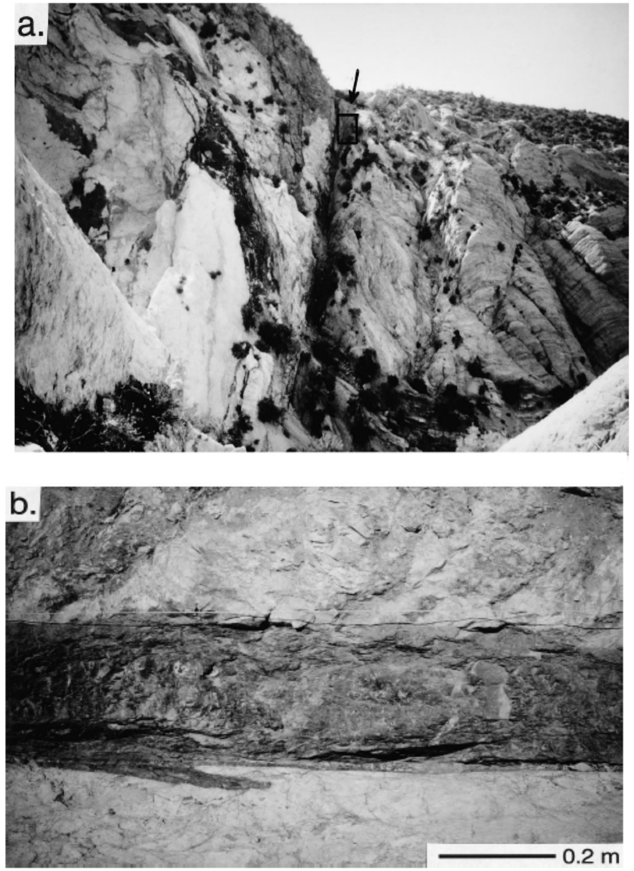


Shear stress (Mpa)

Normal stress (Mpa)

Byerlee's law
$\tau \sim 0.6\sigma$

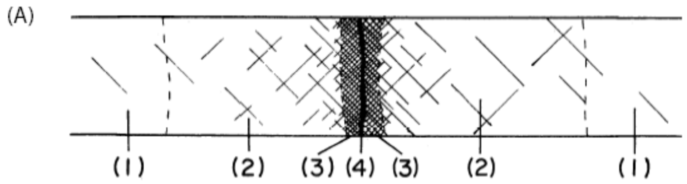
Fault zone damage



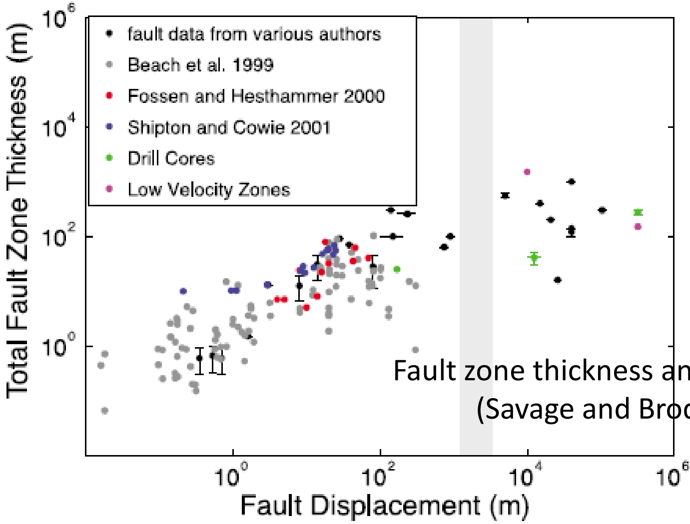
[Chester and Chester, 1998]

Internal Structure of a Major Fault Zone

(after Chester et al., 1993; Chester & Chester, 1998; Sibson, 2003)



(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



Fault zone thickness and maturity
(Savage and Brodsky, 2011)

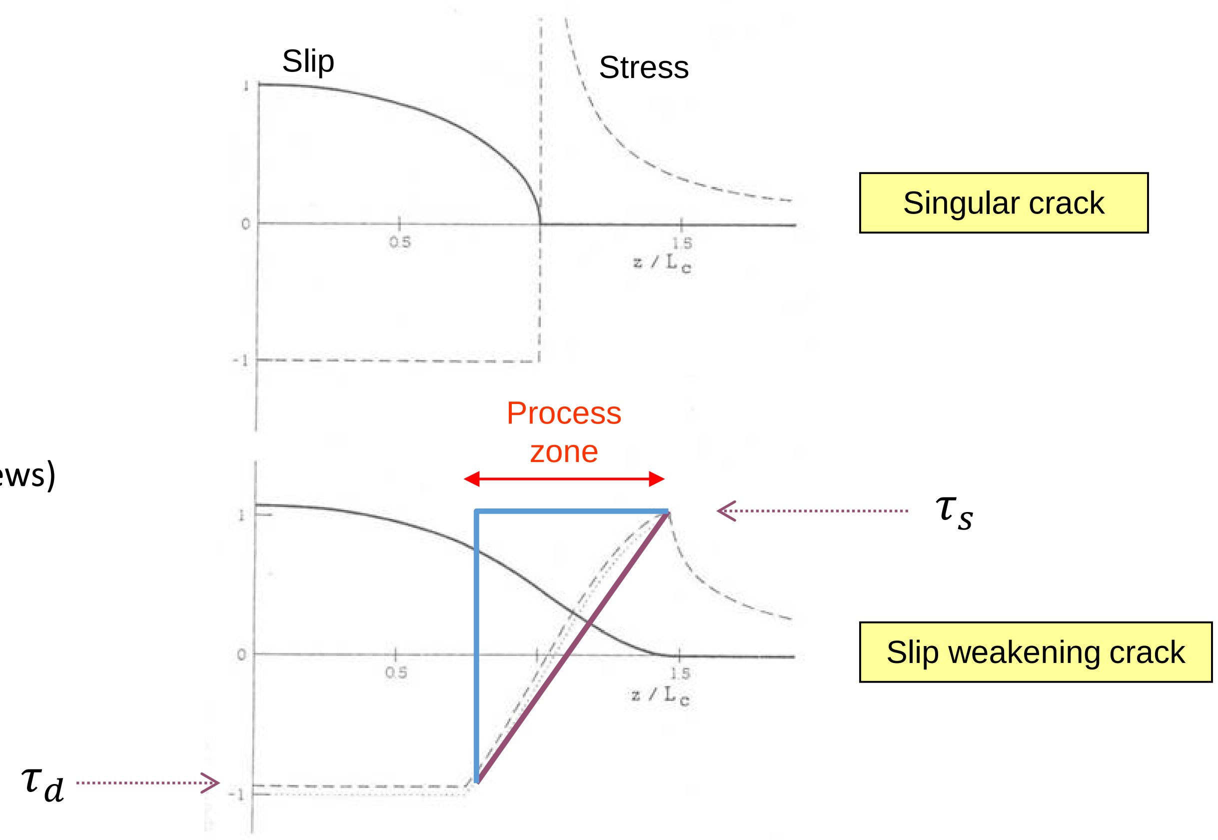
Cohesive zone models

Assumption: dissipative processes are mapped onto the fault plane, represented by a distribution of **cohesive stresses** near the crack tip

Usual cohesive models:

- constant (Dugdale, Barenblatt)
- linearly dependent on distance to crack tip (Palmer and Rice, Ida)
- linearly dependent on slip (Ida, Andrews)

Slip and stress along a shear crack
(only half crack shown, Andrews 1976)



Cohesive zone size

- Cohesive stresses generate a negative stress intensity factor

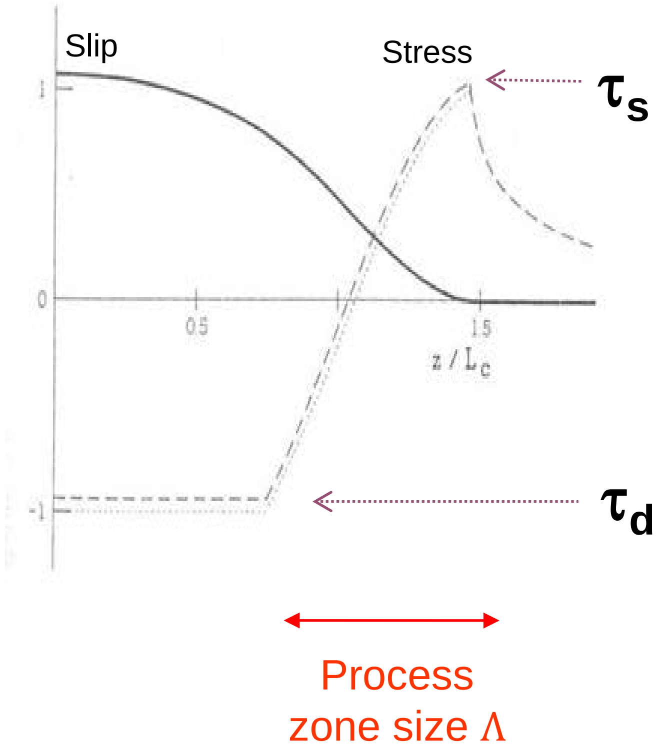
$$K_c \sim -(\tau_s - \tau_d)\sqrt{\Lambda}$$

that cancels the singularity: $K + K_c = 0$

- → **size of the cohesive zone** $\Lambda \sim \frac{K^2}{(\tau_s - \tau_d)^2}$

- $G_c = \frac{K^2}{2\mu}$

$$\boxed{\Lambda \sim 2\mu G_c/(\tau_s - \tau_d)^2}$$



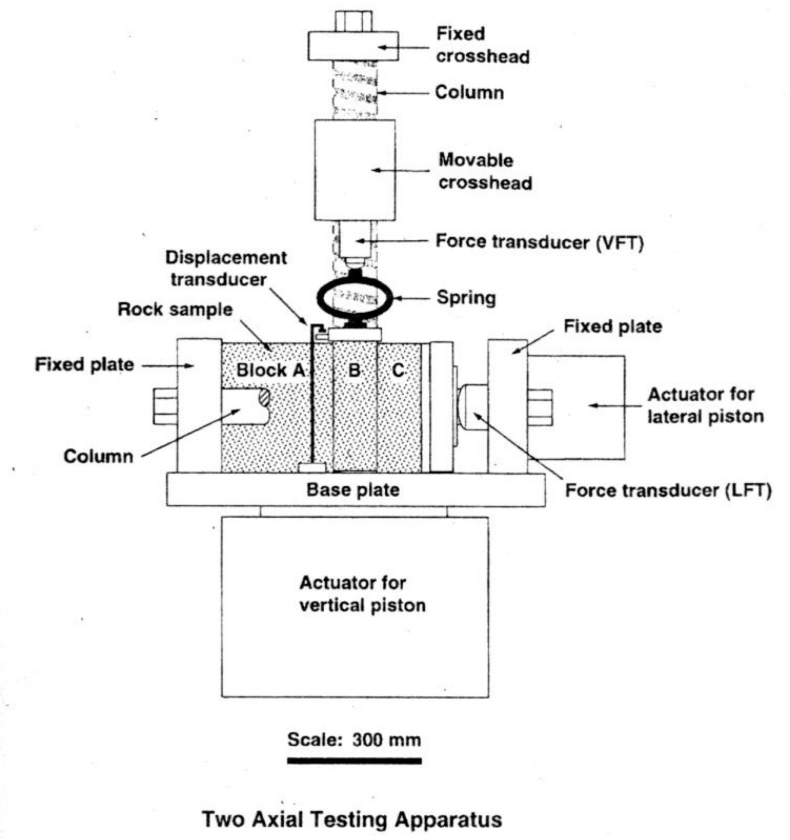
Laboratory-derived friction laws

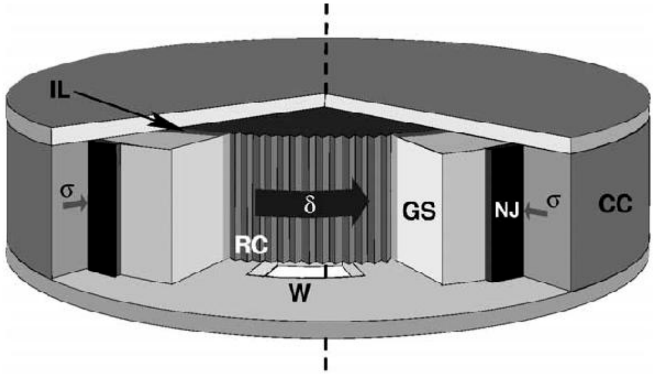
Requirements :

- High normal stress (100 MPa)
- High slip rate (1 m/s)
- Large displacements (>1 m)
- Large sample (>L_c) and high resolution
- Gouge + fluids

Only partially met by current experiments

Sandwich configuration (Ohnaka and Shen 1999)



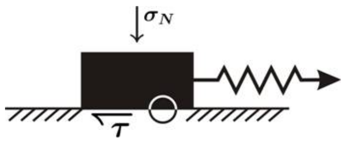


Rotary configuration (Chambon et al 2002)

Laboratory-derived friction laws

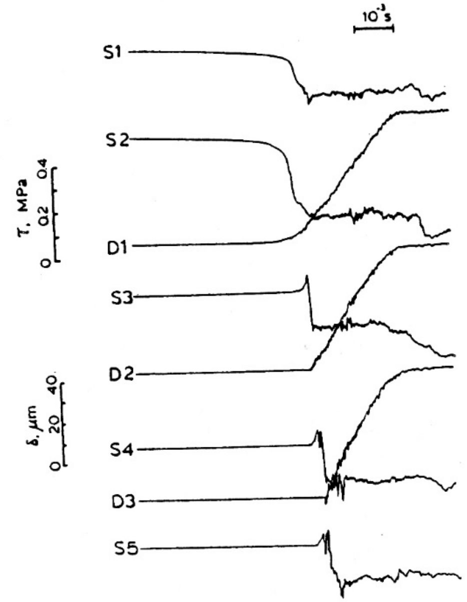
Low resolution experiments (≈ spring+block)
record the average stress and slip

→ macroscopic friction



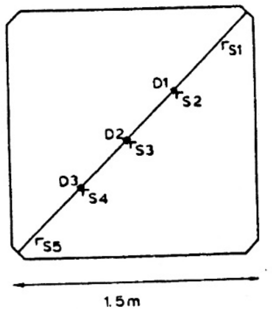
High resolution experiments are densely instrumented

→ local friction + rupture nucleation and propagation

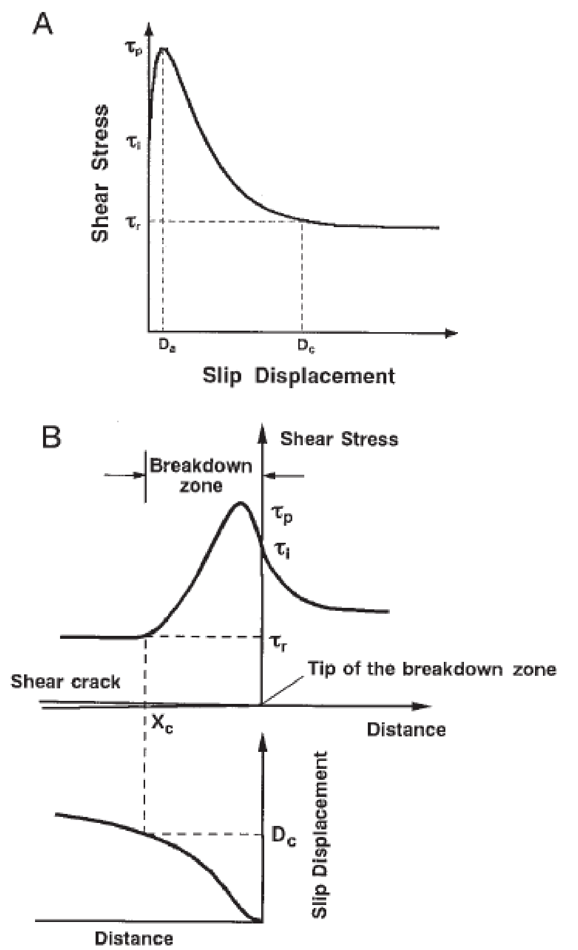


S = stress

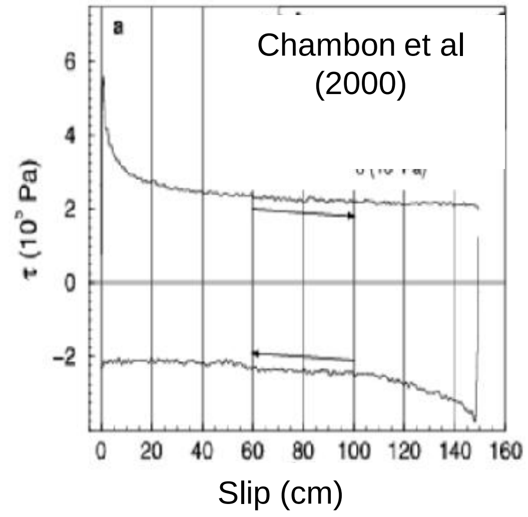
D = slip

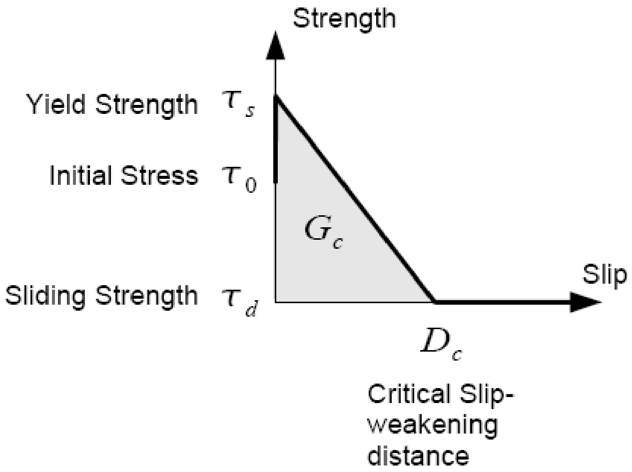


Large scale experiment
Dieterich (1980)



Slip weakening friction





Slip weakening (e.g. Ohnaka) is the main effect during fast dynamic rupture.

Linear slip weakening is a simplified model.

Important parameters:

- $\mathbf{D_c}$ = characteristic slip, associated to micro-contact evolution or grain rearrangement.

  Without gouge $D_c \approx 0.1$ mm.

  With gouge D_c >10 cm

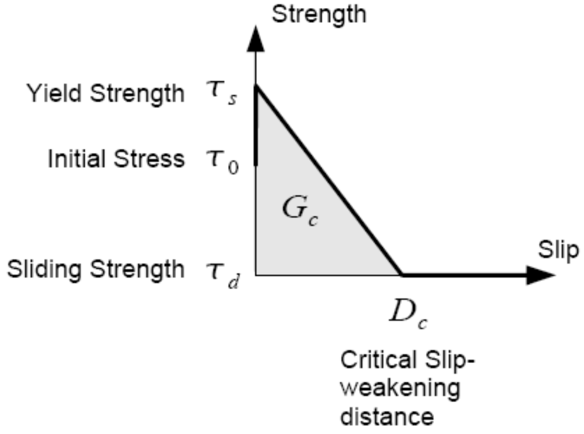
- Strength drop: $\tau_s - \tau_d$

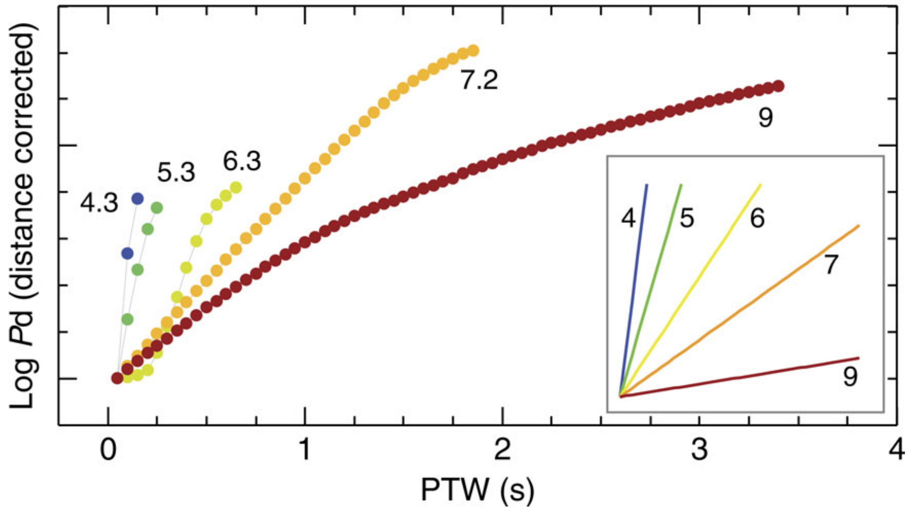
  Usually a small fraction of normal stress $\sim 0.1\ \sigma$

- Fracture energy of a linear slip weakening model :

$$\boldsymbol{G_c = \frac{1}{2}(\tau_s - \tau_d) D_c}$$

Exponential initiation





Linear slip-weakening:

$$\Delta\tau = (\tau_s - \tau_d)D/D_c$$

If there is some viscosity in the fault behavior:

$$\Delta\tau = \eta\ \dot{D}$$

Equating both:

$$\dot{D} = sD$$

Hence

$$D(t) \sim \exp(st)$$

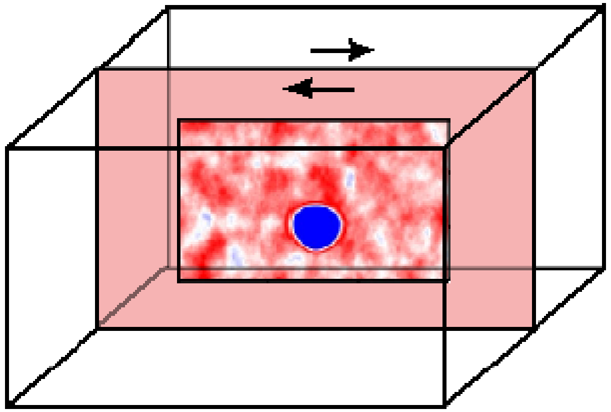
where $s = (\tau_s - \tau_d)/\eta D_c$

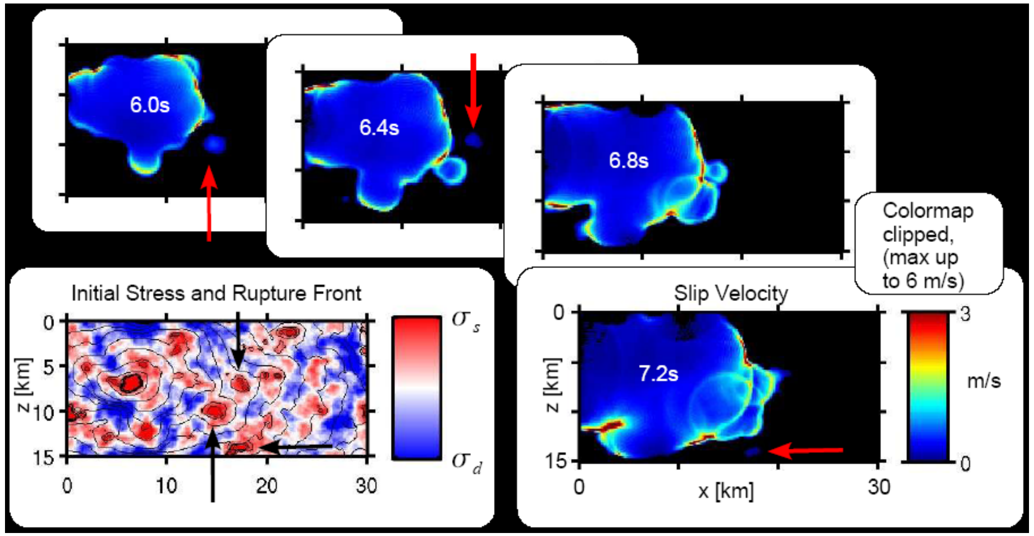
One form of viscosity is radiation damping, $\eta = \mu/2c_s$

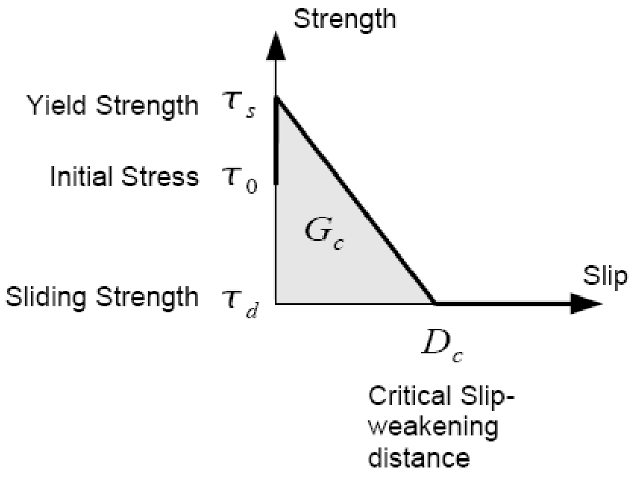
Dynamic Rupture Simulation

Setup:

- Planar fault embedded in homogeneous elastic full space
- Boundary Integral Equation Method (*Dunham, 2005*)



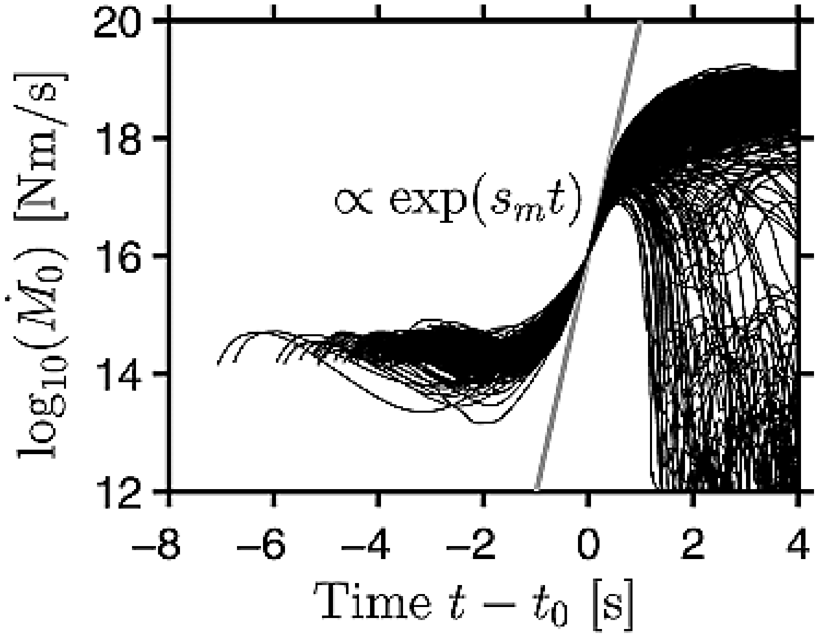




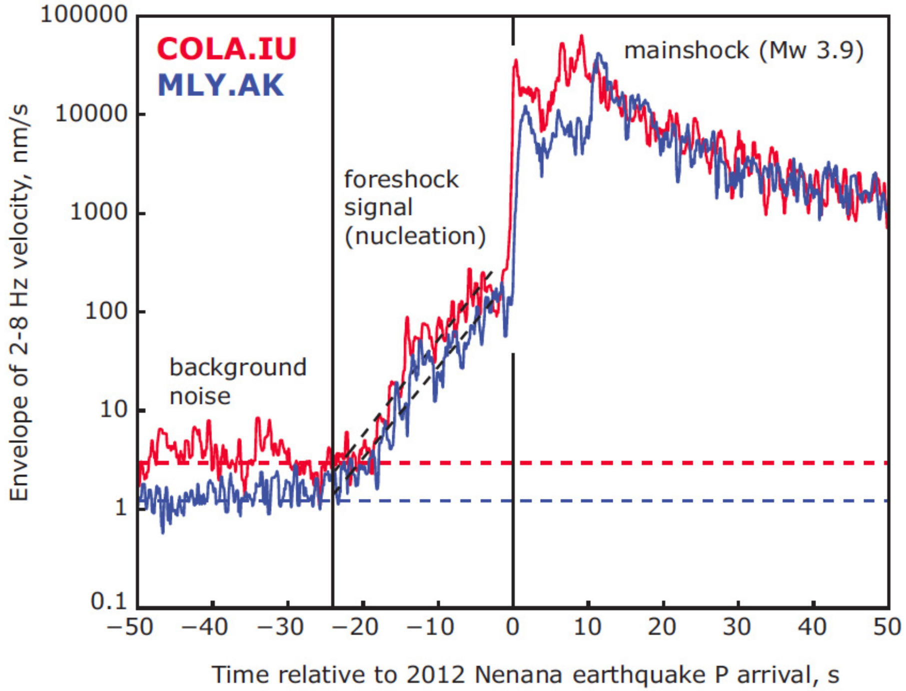
Exponential initiation

$$s_m = 2c_s(\tau_s - \tau_d)/\mu D_c$$

$$s = (\tau_s - \tau_d)/\eta D_c$$

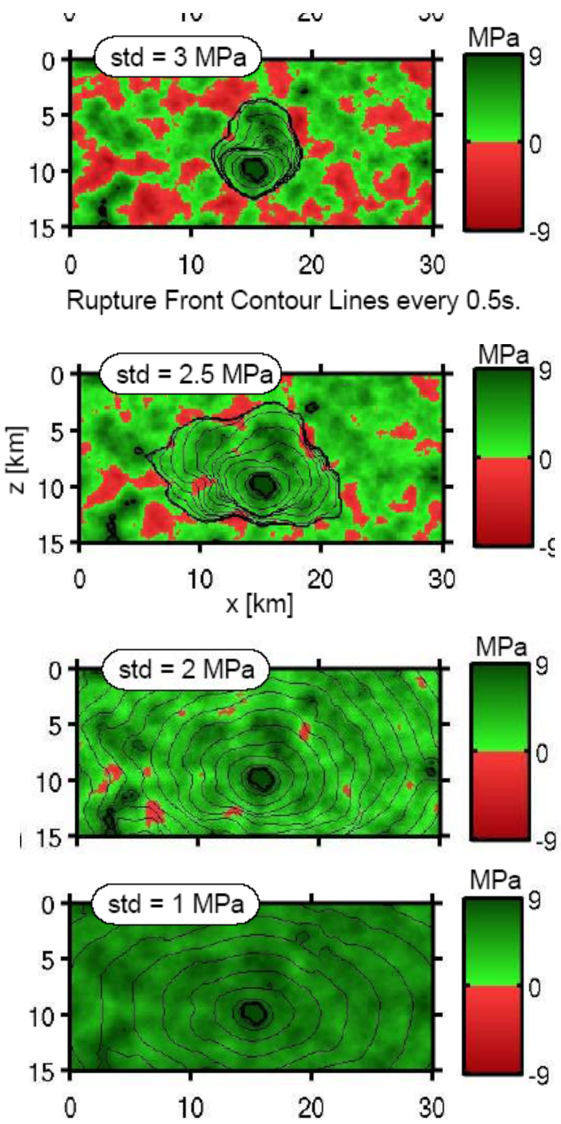


Simulations
Ripperger et al (2007)

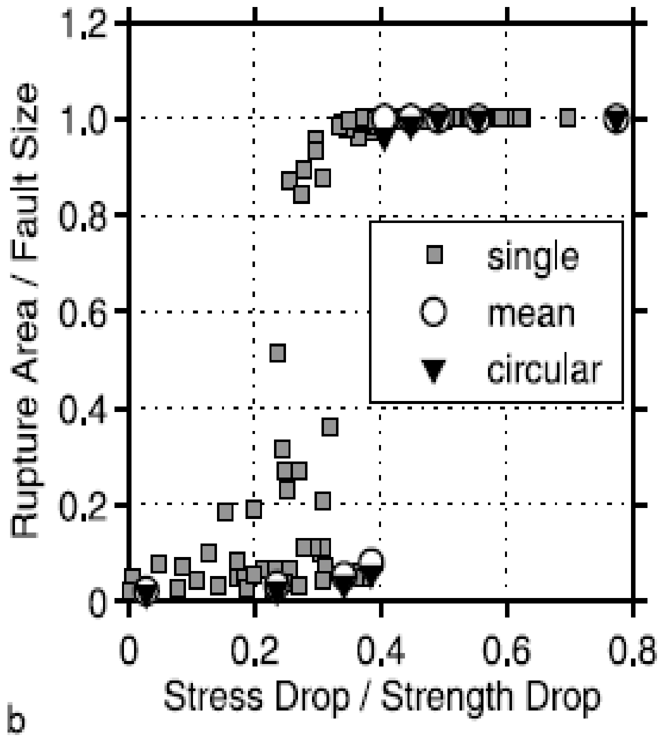


Observations
Tape et al (2013)

Rupture arrest



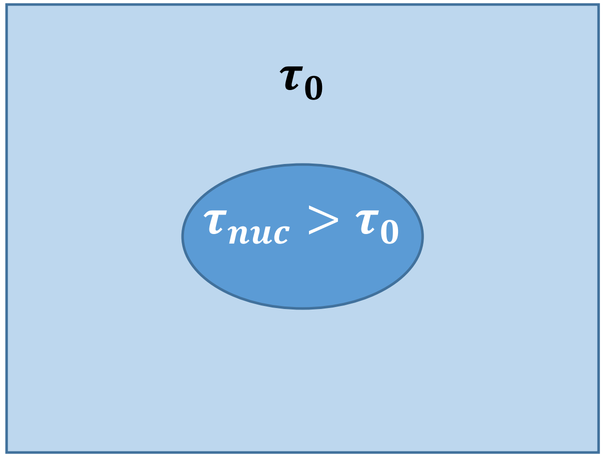
Rupture "percolation" transition



Ripperger et al (2007)

Rupture arrest

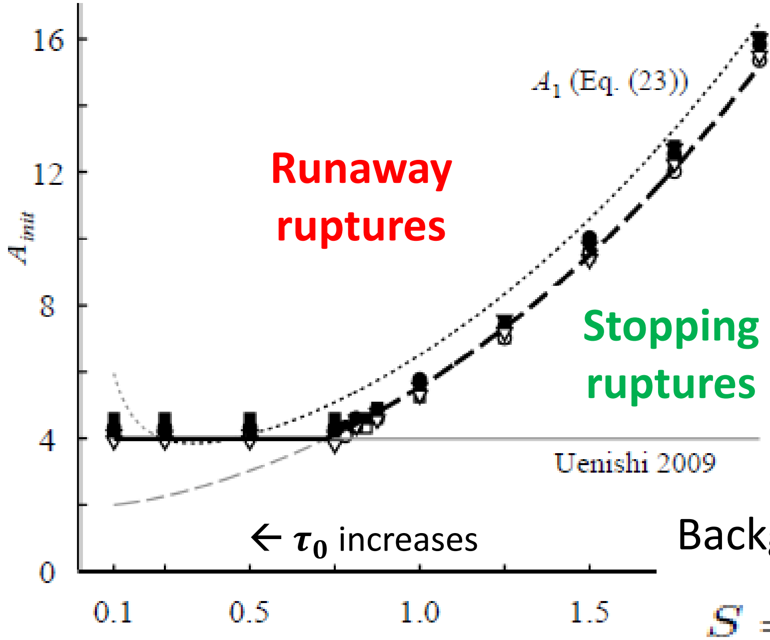
Fault plane



Rupture nucleated at a highly stressed patch.

- Will it stop spontaneously?
- How does the rupture outcome depend on patch size and overstress?

Nucleation area

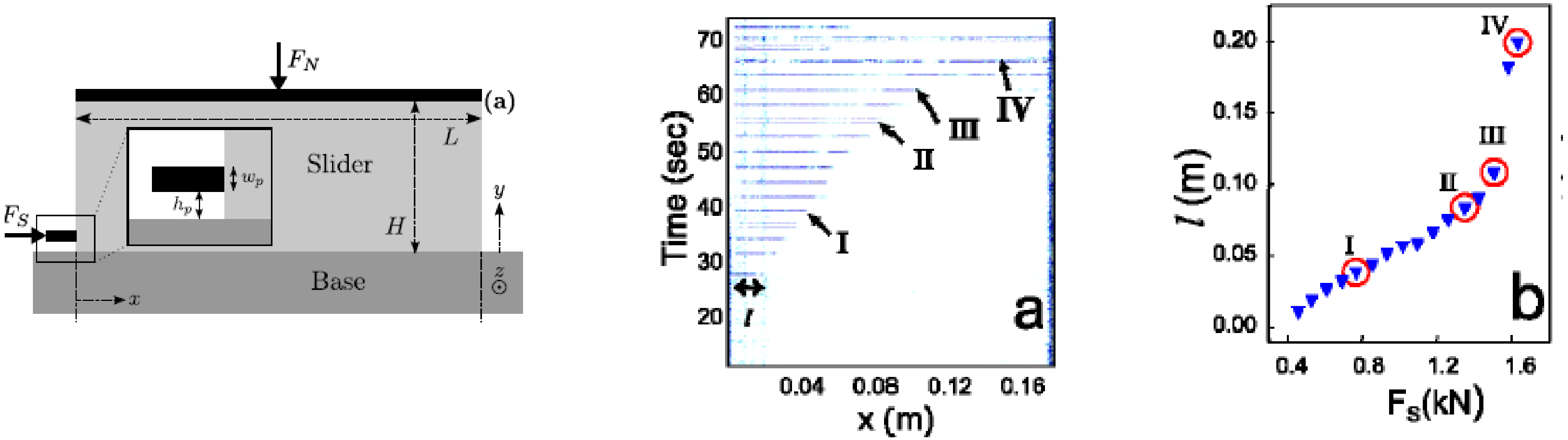


Background stress

$$S = \frac{\tau_s - \tau_0}{\tau_0 - \tau_d}$$

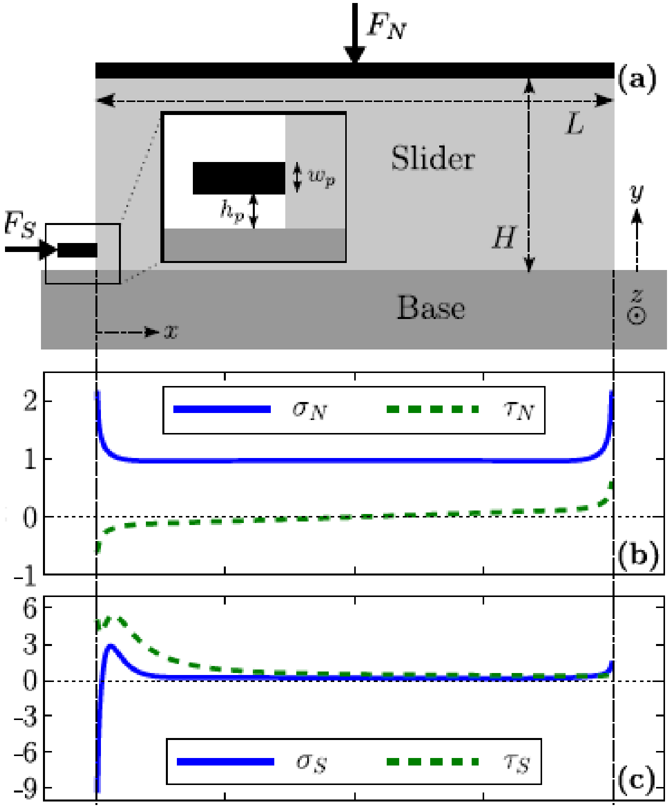
Numerical simulations compared to fracture mechanics predictions (Galis et al, 2014)

Laboratory foreshocks

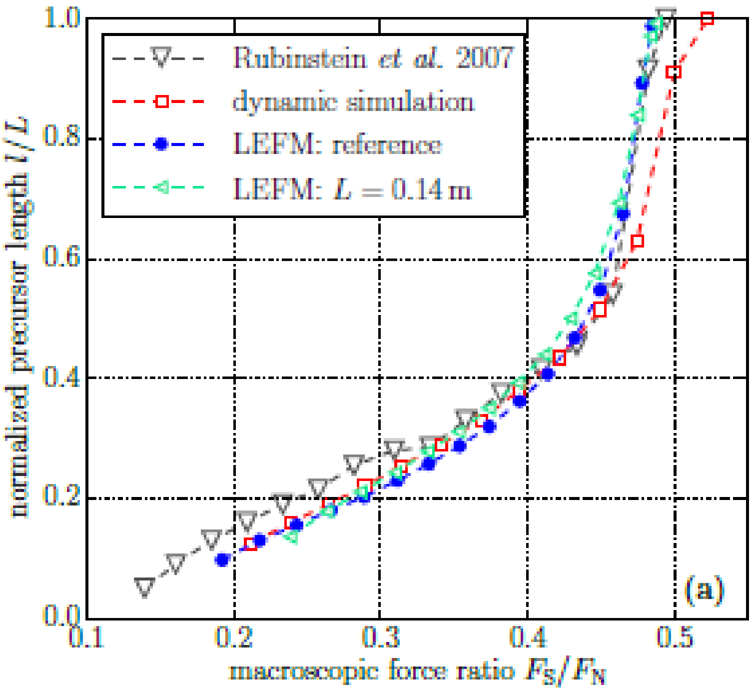


Rubinstein et al (2007)

Laboratory foreshocks

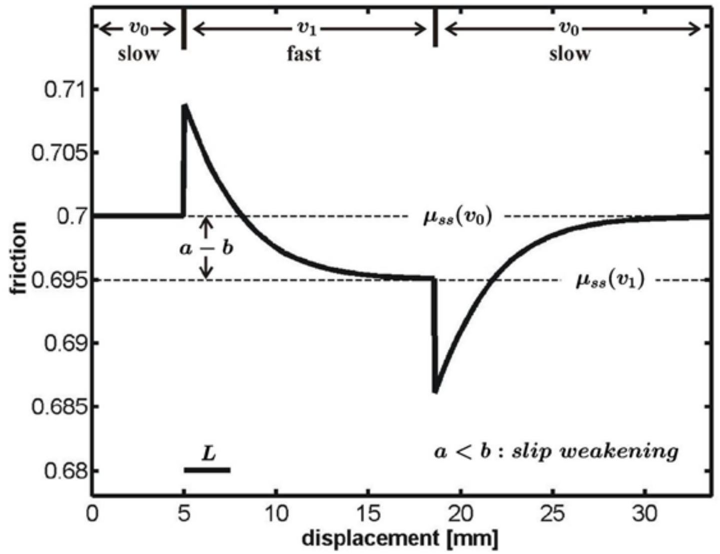


Kammer et al (2014)



Lab results reproduced by slip-weakening and fracture mechanics models

Rate-and-state friction



$$\mu = \mu^* + \boxed{a \ln\left(\frac{V}{V^*}\right)} + \boxed{b \ln\left(\frac{V^*\theta}{L}\right)}$$

$$\dot{\theta} = 1 - \frac{V\theta}{L}$$

V = slip velocity, θ = state variable

Second order effects: logarithmic healing (micro-contact creep) and velocity-weakening

→Phenomenological rate-and-state friction law introduced by Dieterich and Ruina in the early 1980s

Essential ingredients:

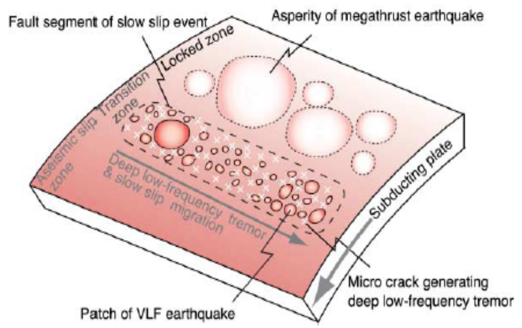
- non-linear viscosity
- evolution effect

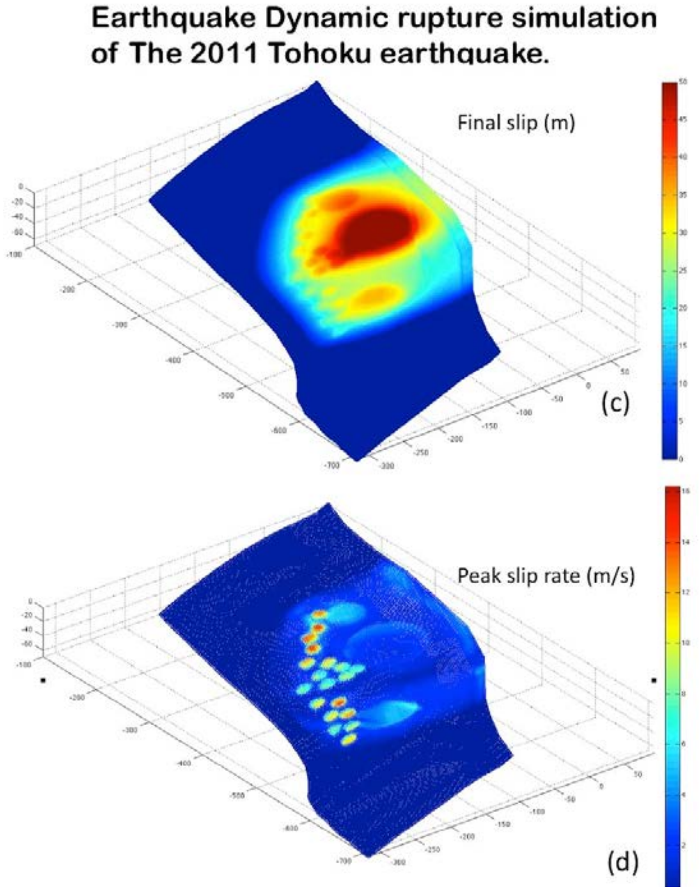
Most important during slow slip (nucleation and post-seismic)

During fast dynamic rupture, an equivalent D_c can be estimated:

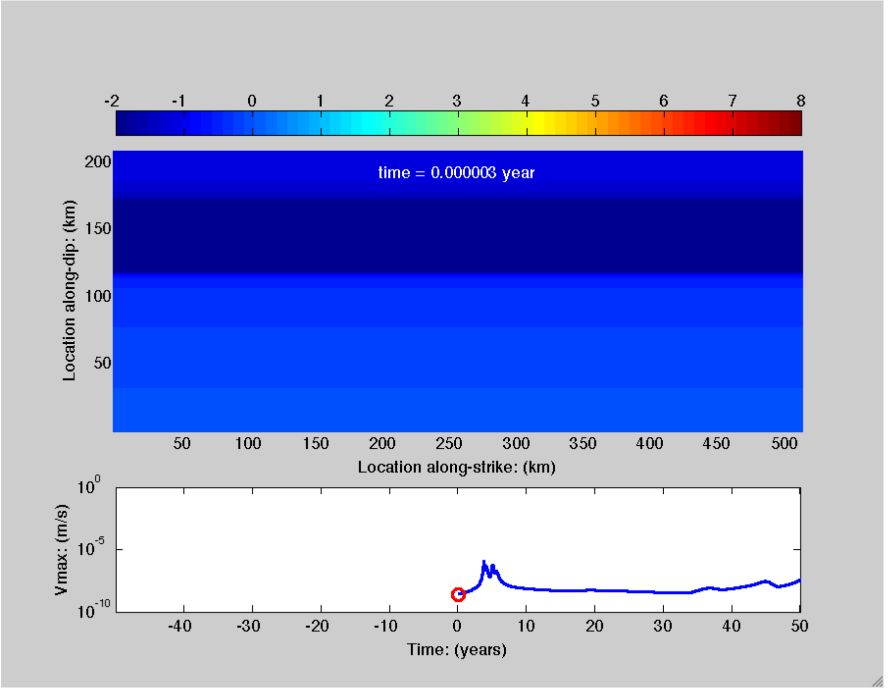
$D_c \approx 20$ L

Integration between dynamic rupture and earthquake cycle modeling



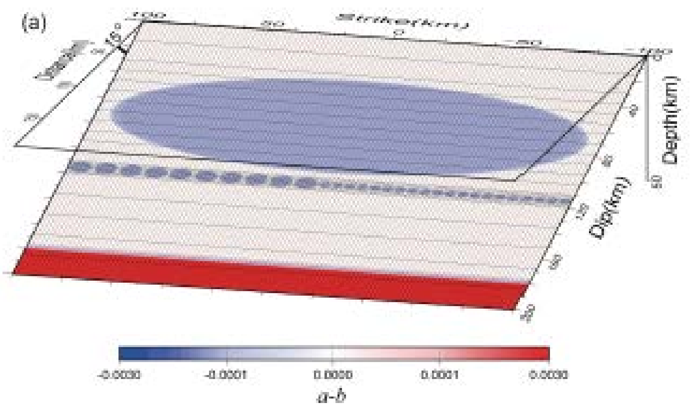


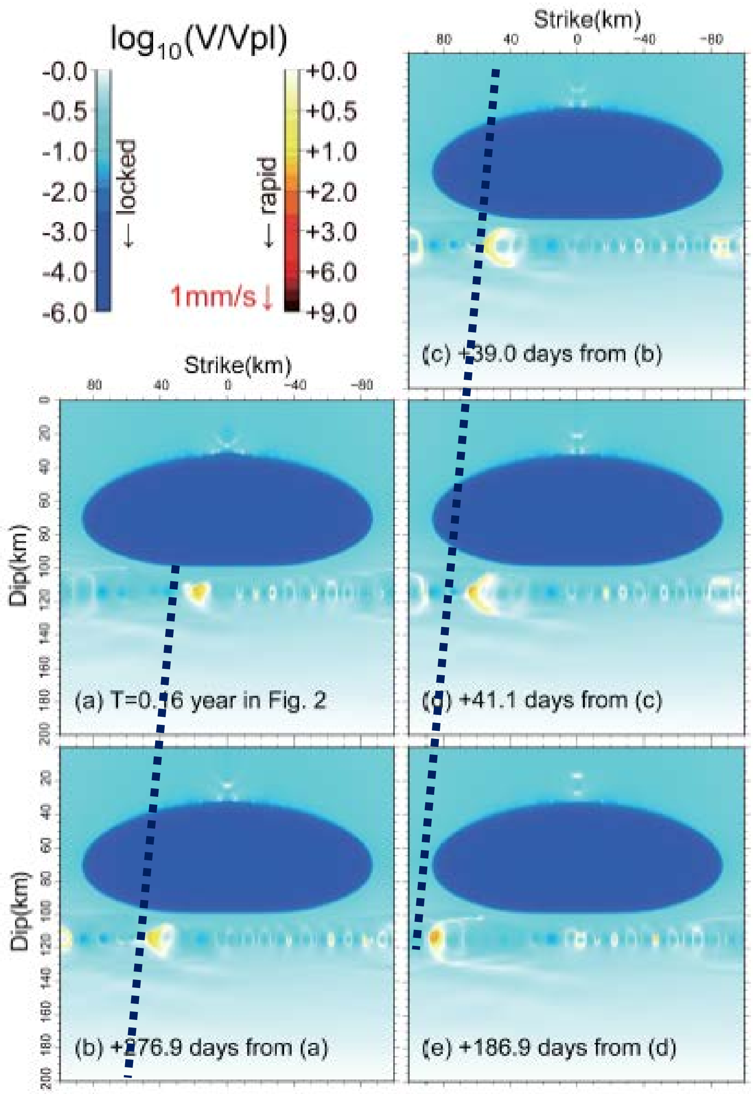
Galvez et al (2014)



Yingdi Luo, earthquake cycle simulations

Rate-and-state simulations of deep tremor swarms





Tremor swarms result from a cascade of triggering between brittle asperities **mediated by creep transients**:

Asperity failure

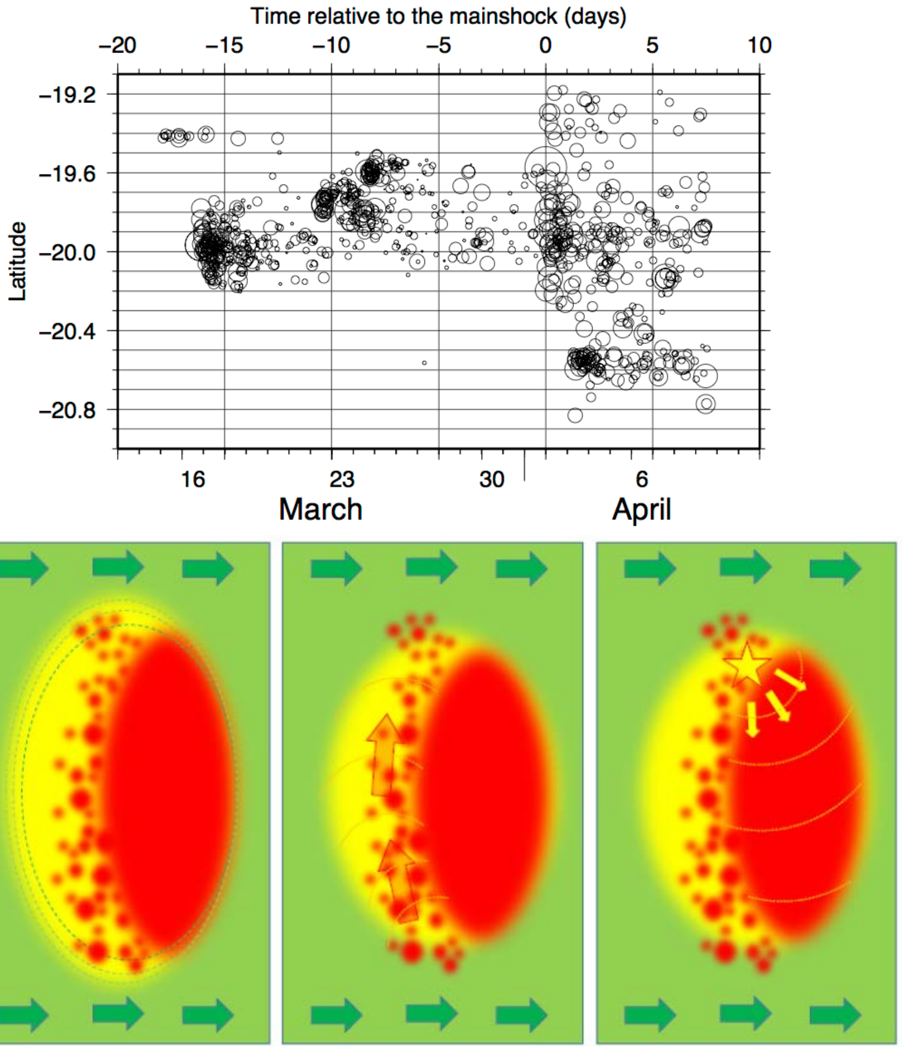
→ propagating creep perturbation

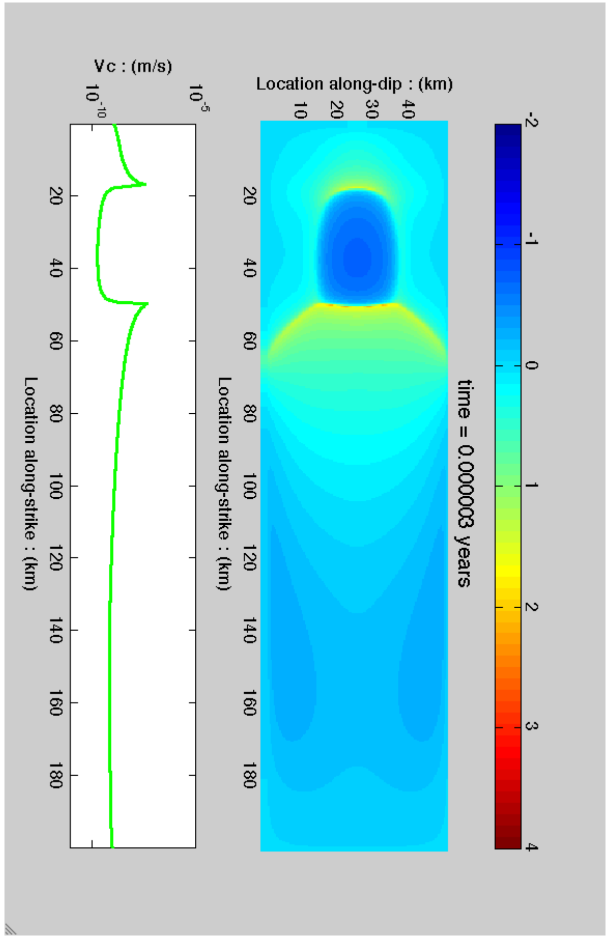
→ loading and failure of next asperity

Quasi-dynamic 3D simulations in collaboration with K. Ariyoshi (Earth Simulator, JAMSTEC, Japan)

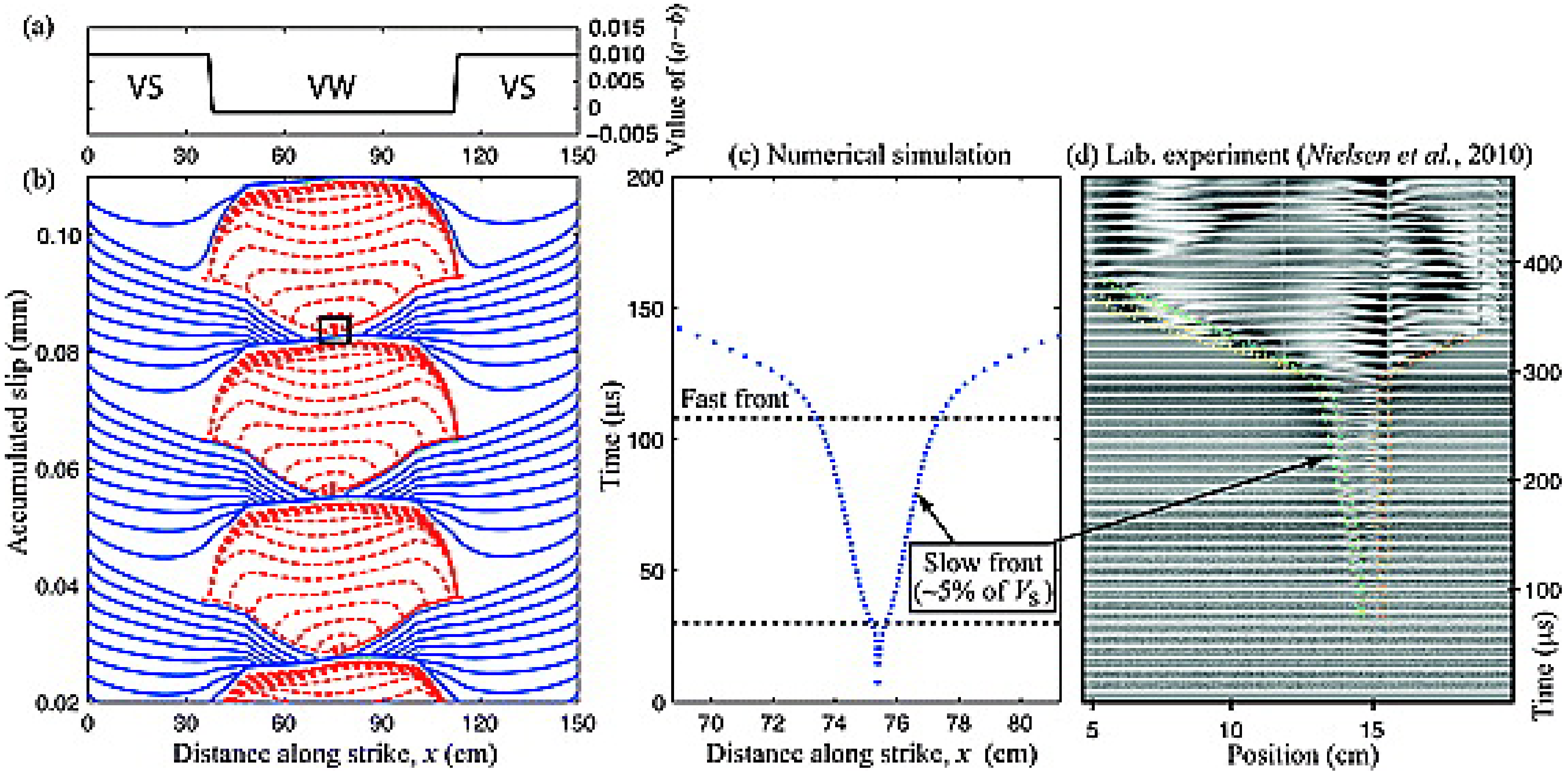
Rate-and-state simulations of foreshock swarms

Iquique
2014
foreshock
sequence

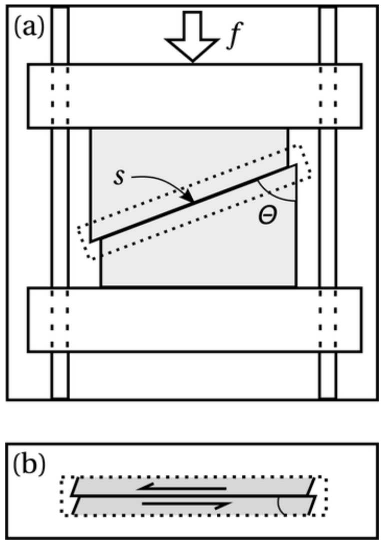




Slow fronts during nucleation



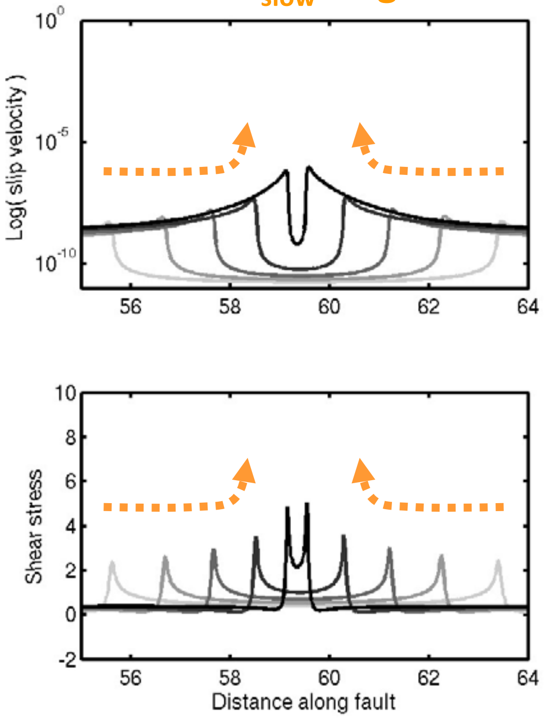
Kaneko and Ampuero (2011)



Nielsen et al (2010)

Slow fronts in rate-and-state earthquake models
(Kaneko and Ampuero 2011)

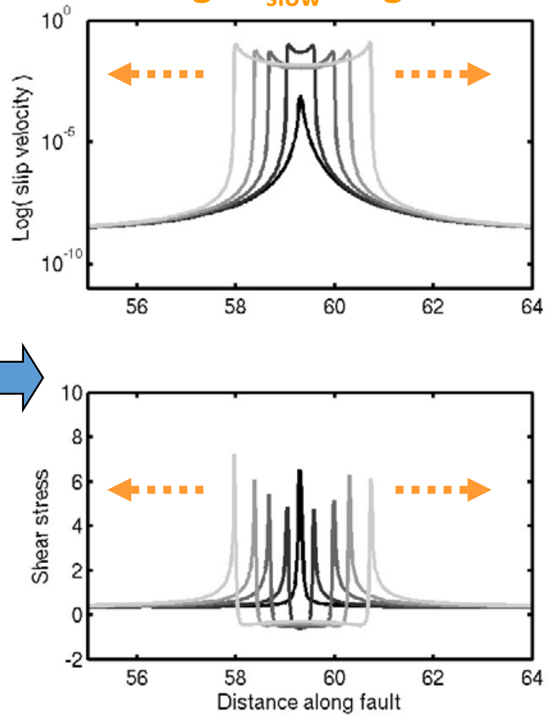
Before "V_{slow} stage"



Before the V_{slow} stage (interseismic stage): inward propagation of two very slow creep fronts

The V_{slow} stage starts on a **stress concentration** inherited from the coalescence of the creep fronts

During "V_{slow} stage"



Consider a stress concentration (=F×length) over a background stress drop ($\Delta\tau$).

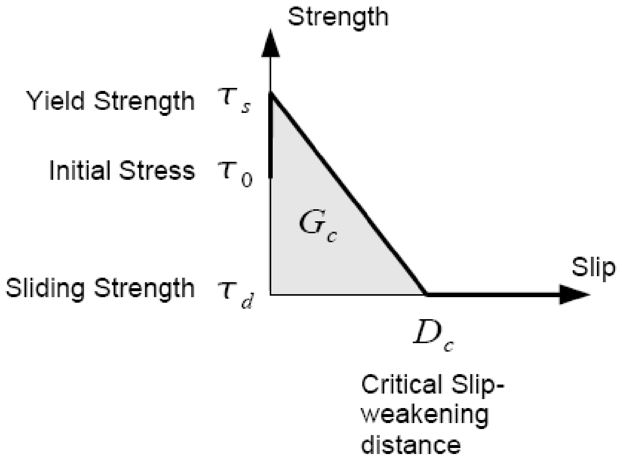
The static **energy release rate** as function of distance to the stress concentration (a) **reaches a minimum at some distance**.

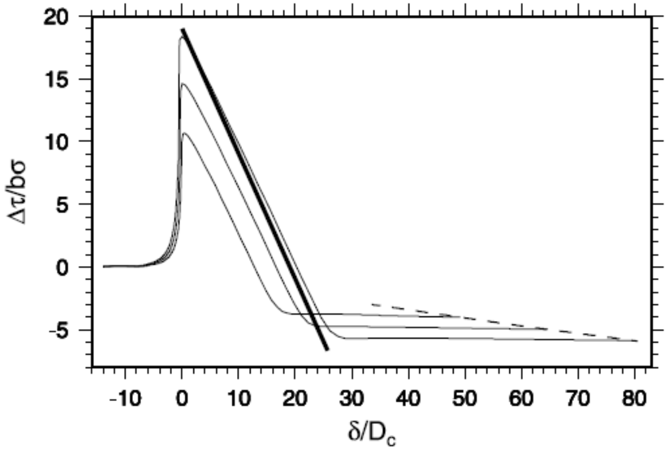
This implies a roughly constant rupture speed.

$$G_0 \doteq \frac{K_0^2}{2\mu}$$

$$\overline{K}_0(a) \approx \sqrt{\pi a}\left(\Delta\overline{\tau} + \frac{2}{\pi}\frac{F}{a}\right)$$

Process zone size





Rubin and Ampuero (2005)

In slip-weakening friction:

$$\Lambda \approx \mu D_c/(\tau_s - \tau_d)$$

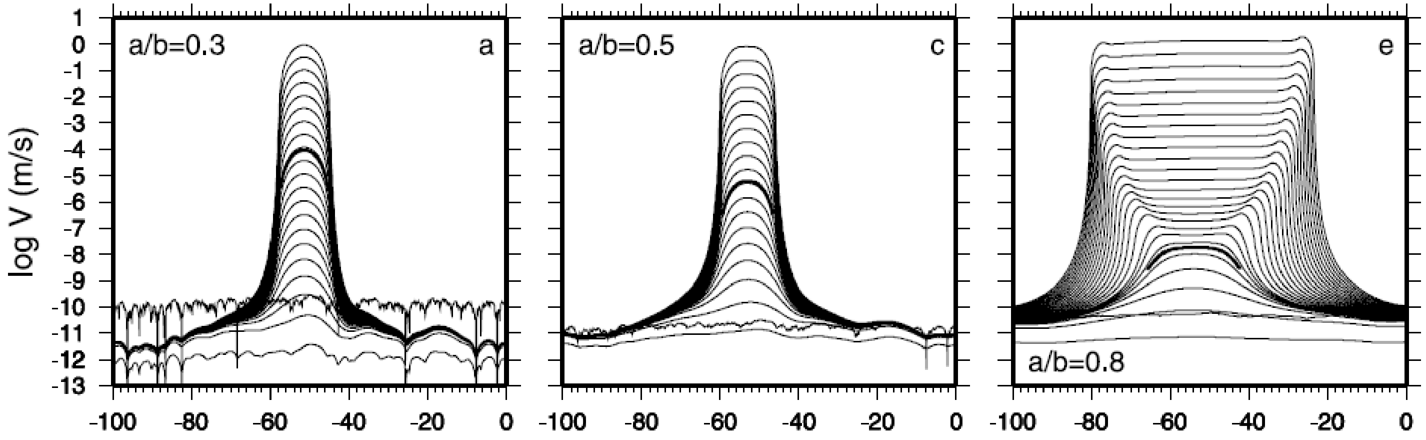
In rate-and-state friction:
Fast sliding far above steady-state ($V \gg L/\theta$) produces quasi-linear slip-weakening with:

$$D_c \approx L \ln \frac{V}{V^*}$$

$$\tau_s - \tau_d \approx b\sigma \ln \frac{V}{V^*}$$

$$\boldsymbol{\Lambda \approx \frac{\mu L}{b\sigma} = L_b}$$

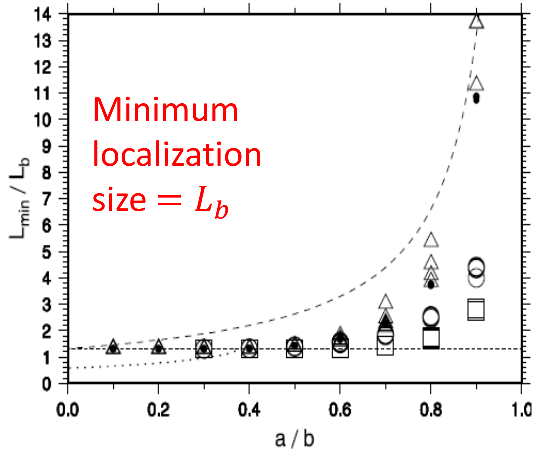
Nucleation size



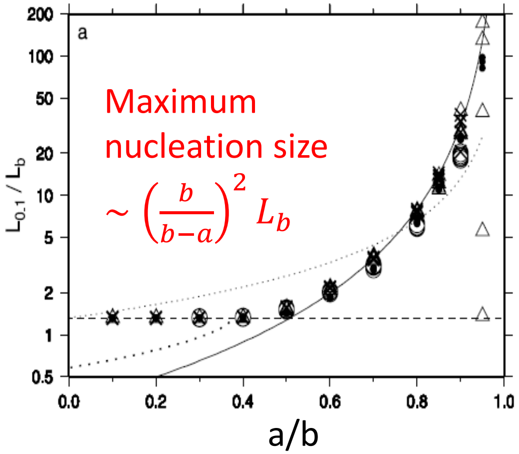
Different nucleation regimes depending on a/b (ratio of viscous to weakening effects)

Localized slip at low a/b

Expanding slip at high a/b

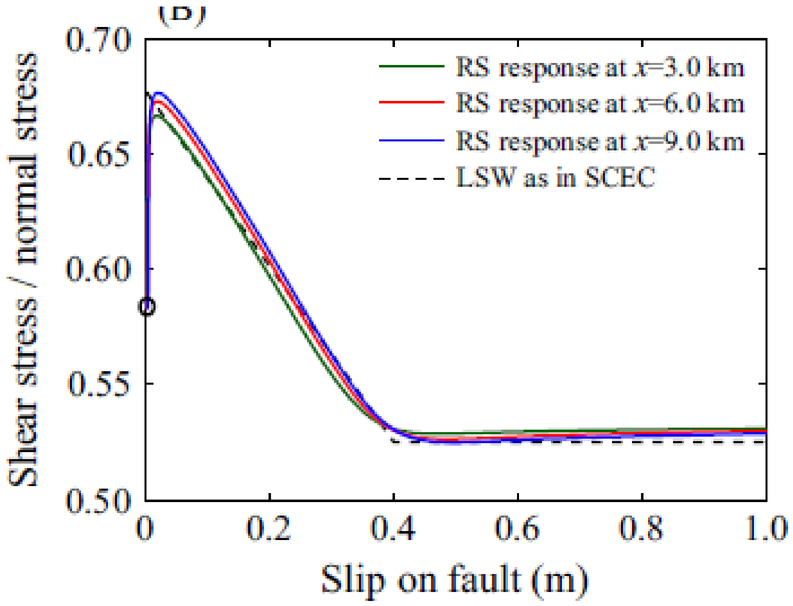


Different nucleation sizes



Rubin and Ampuero (2005)

Rate-and-state friction and fracture energy



Kaneko et al (2008)

Most important during slow slip (nucleation and post-seismic)

Rate-and-state behaves as slip-weakening during fast dynamic rupture

Equivalent :

$$D_c = L \ln\left(\frac{V}{V^*}\right) \approx 20\ L$$

$$G_c \approx \frac{1}{2} b\sigma L \ln\left(\frac{V}{V^*}\right)^2$$

Faults operating at low stress

How large is stre**ss** drop $\Delta\tau$ compared to stren**gth** drop $\tau_s - \tau_d$?

From seismological observations: $\Delta\tau = 1 - 10$ Mpa

From friction and lithostatic overburden:

$$\tau_s - \tau_d = \sigma(\mu_s - \mu_d) = O(100\ MPa)$$

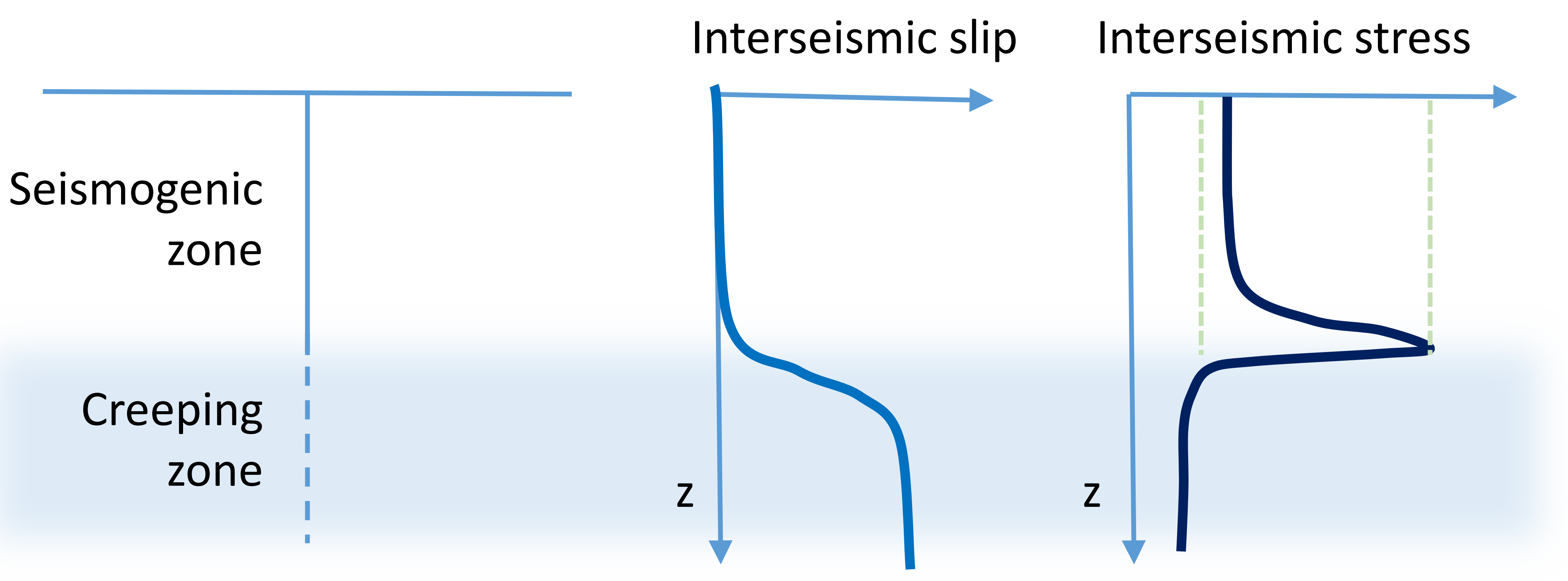
$\rightarrow \Delta\tau \ll \tau_s - \tau_d$

Why so small?

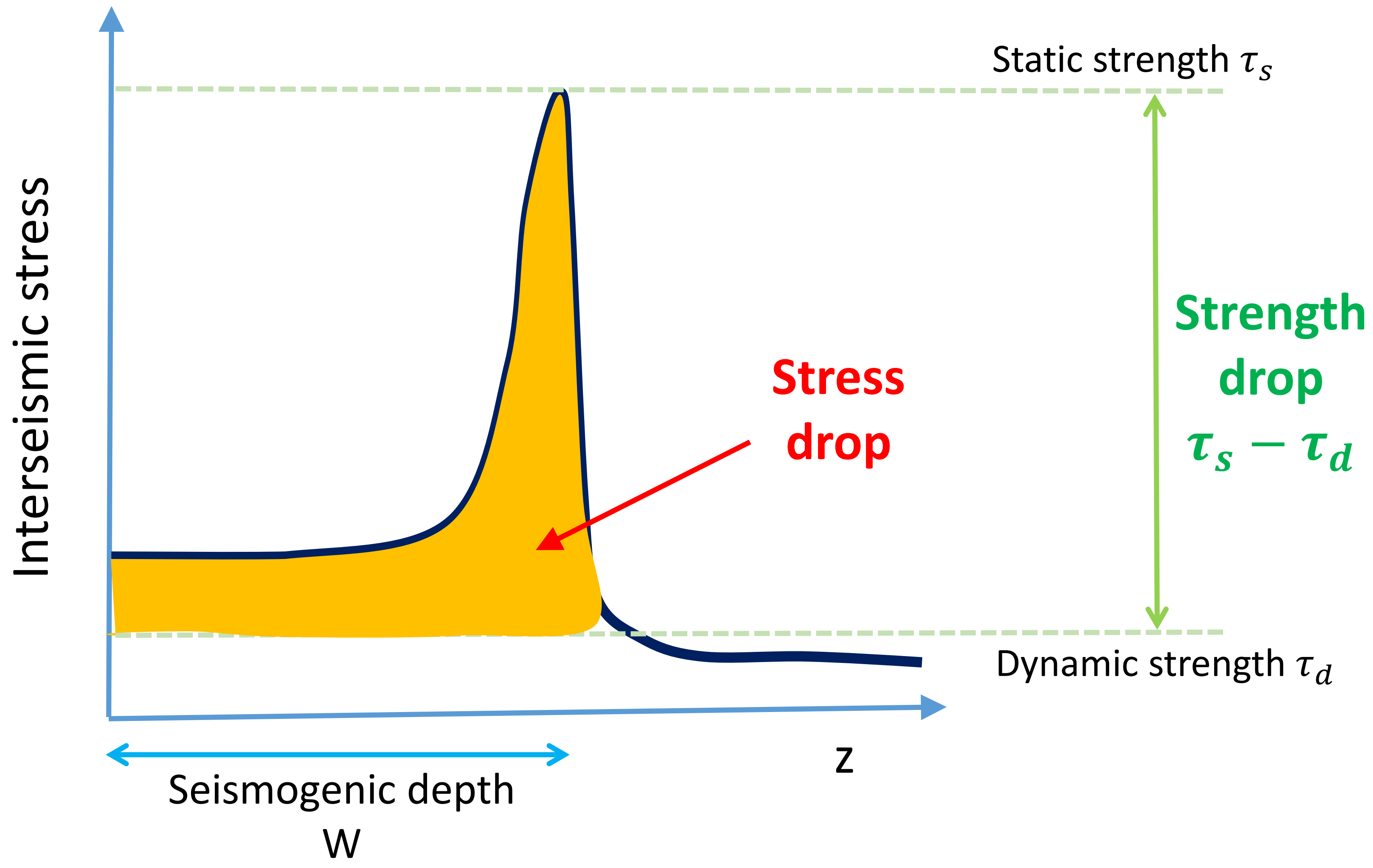
Faults operating at low stress

Fault loaded by deep creep

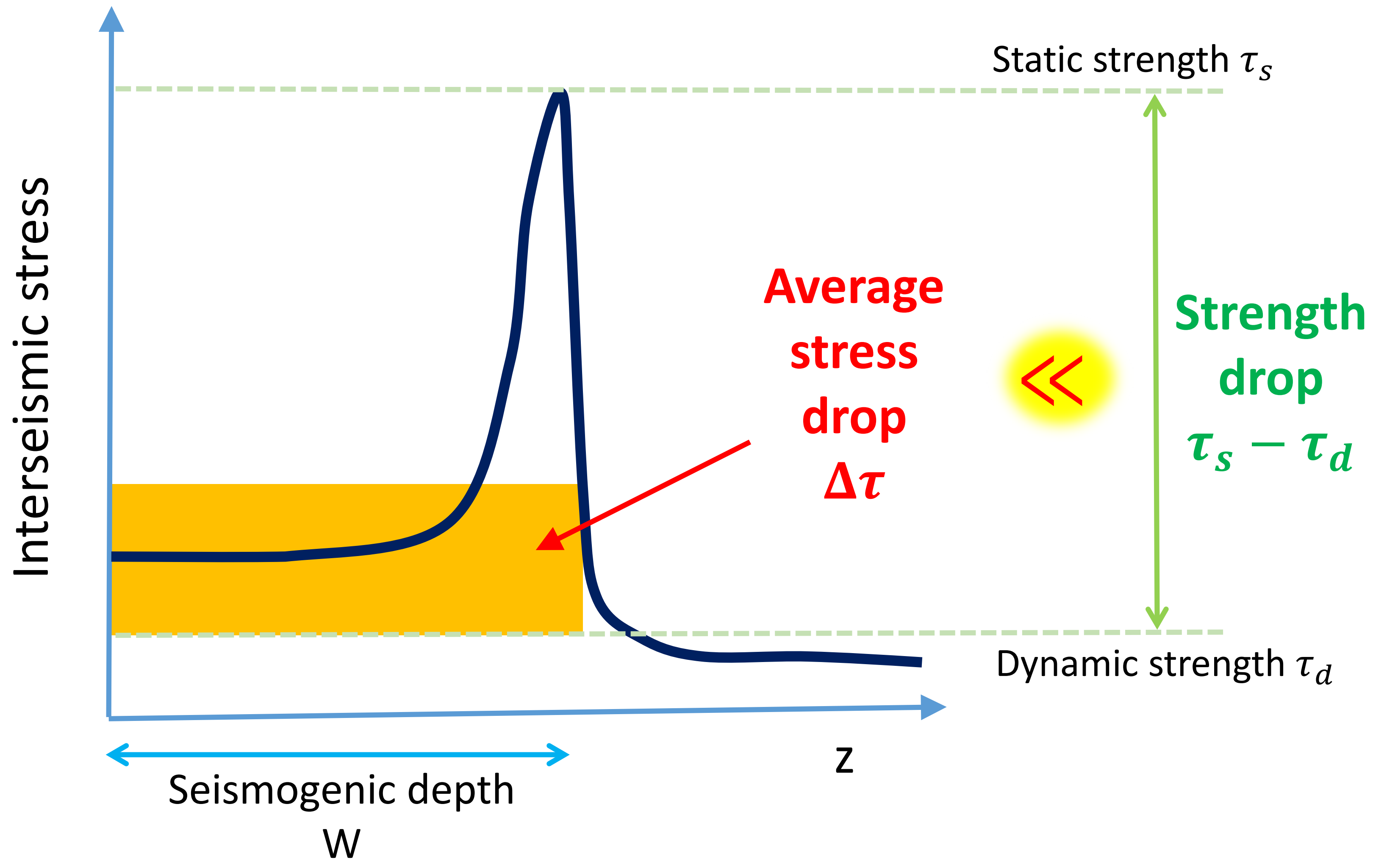
→ stress concentration at the base of the seismogenic zone



Faults operating at low stress



Faults operating at low stress



Faults operating at low stress

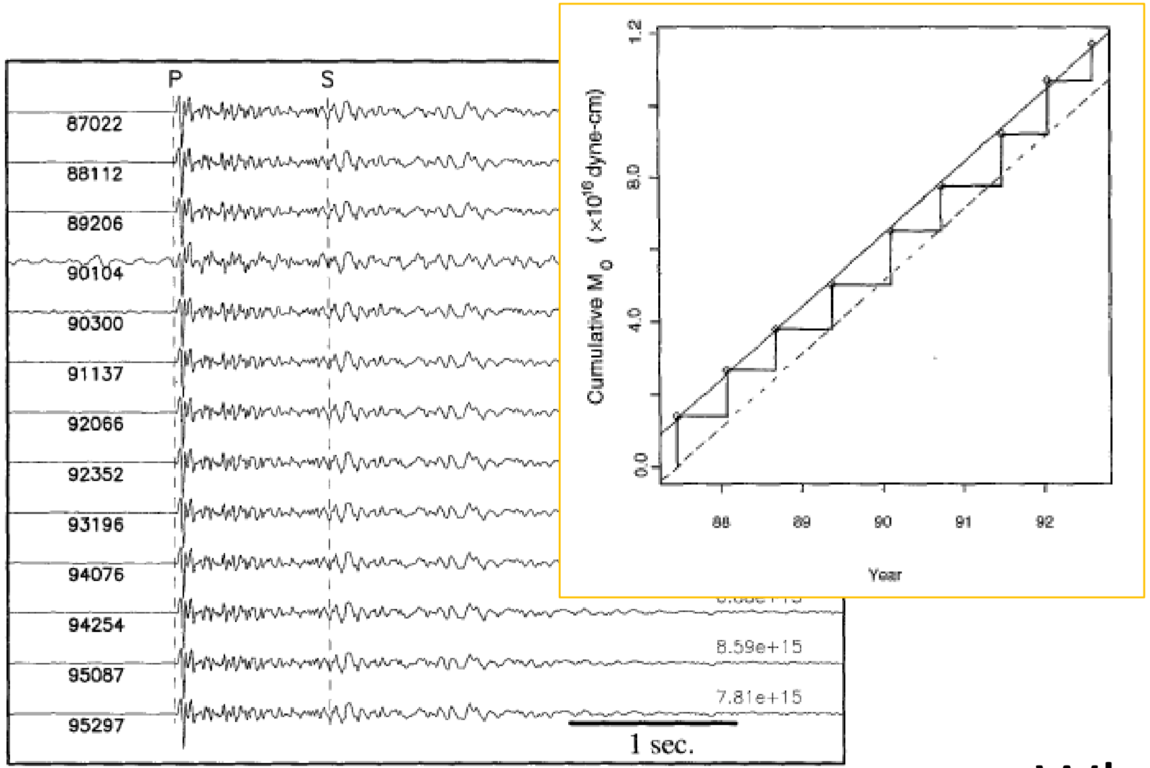
Fracture energy balance: $G_c = \frac{K^2}{2\mu} \sim \frac{\Delta\tau^2 W}{2\mu}$

$$\rightarrow \Delta\tau \sim \sqrt{2\mu G_c / W}$$

Uenishi and Rice's nucleation size: $L_c = \frac{\mu D_c}{\tau_s - \tau_d}$

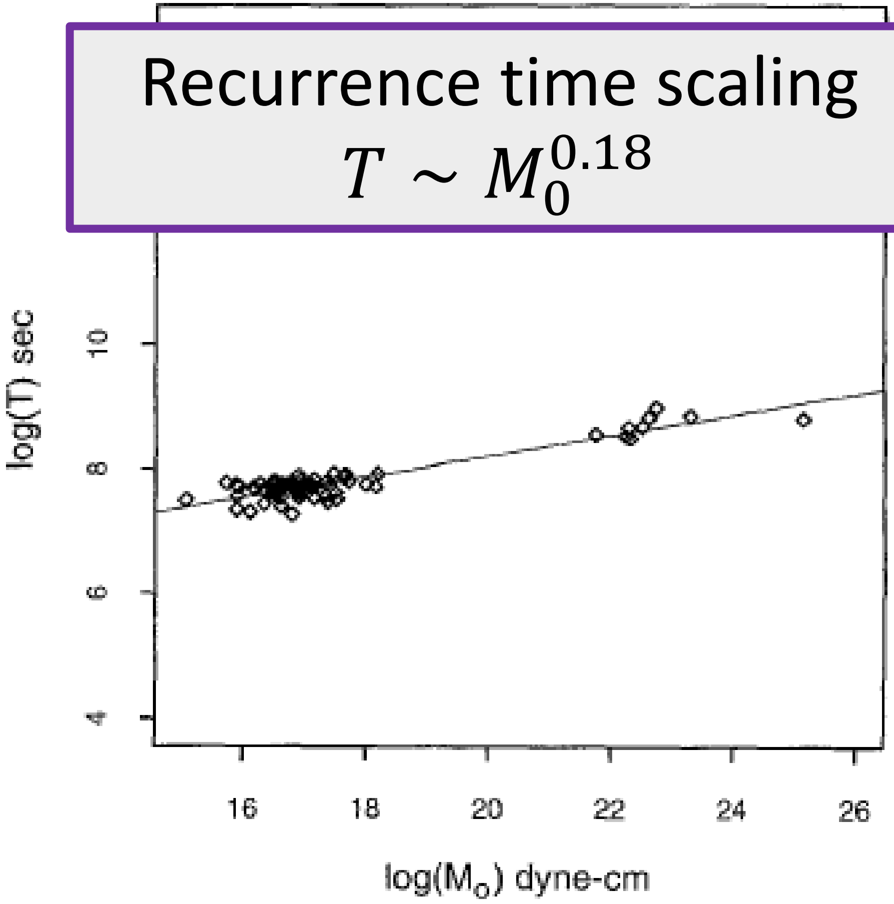
$$\boxed{\rightarrow \frac{\Delta\tau}{\tau_s - \tau_d} \sim \sqrt{\frac{L_c}{W}} \ll 1}$$

Recurrence time scaling of repeating earthquakes



Nadeau and Johnson (1989)

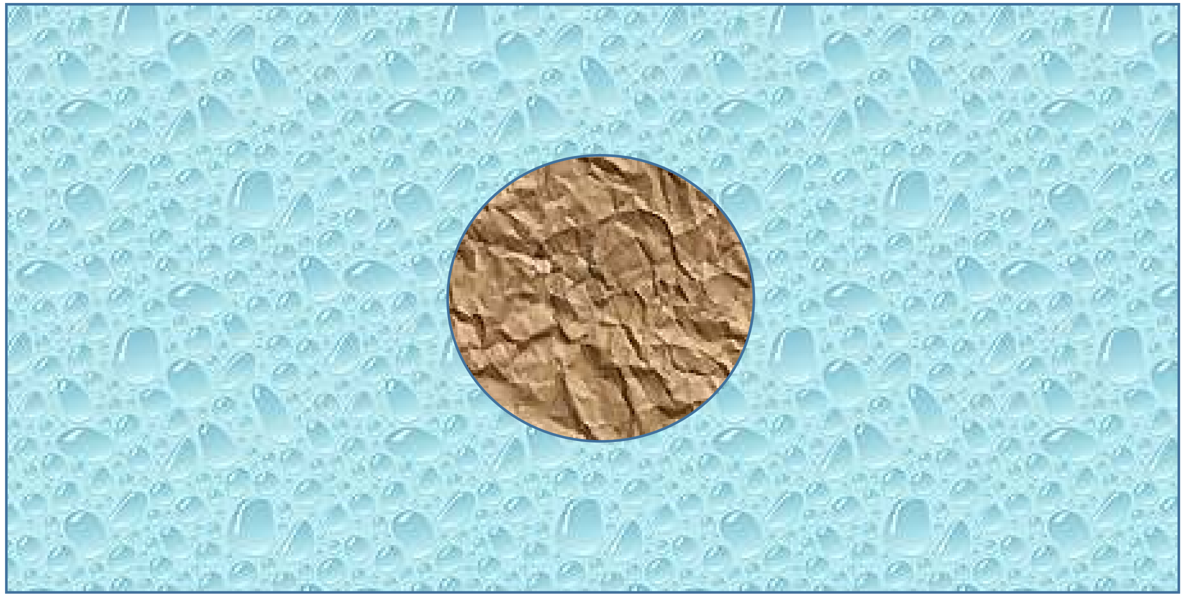
Recurrence time scaling
$T \sim M_0^{0.18}$



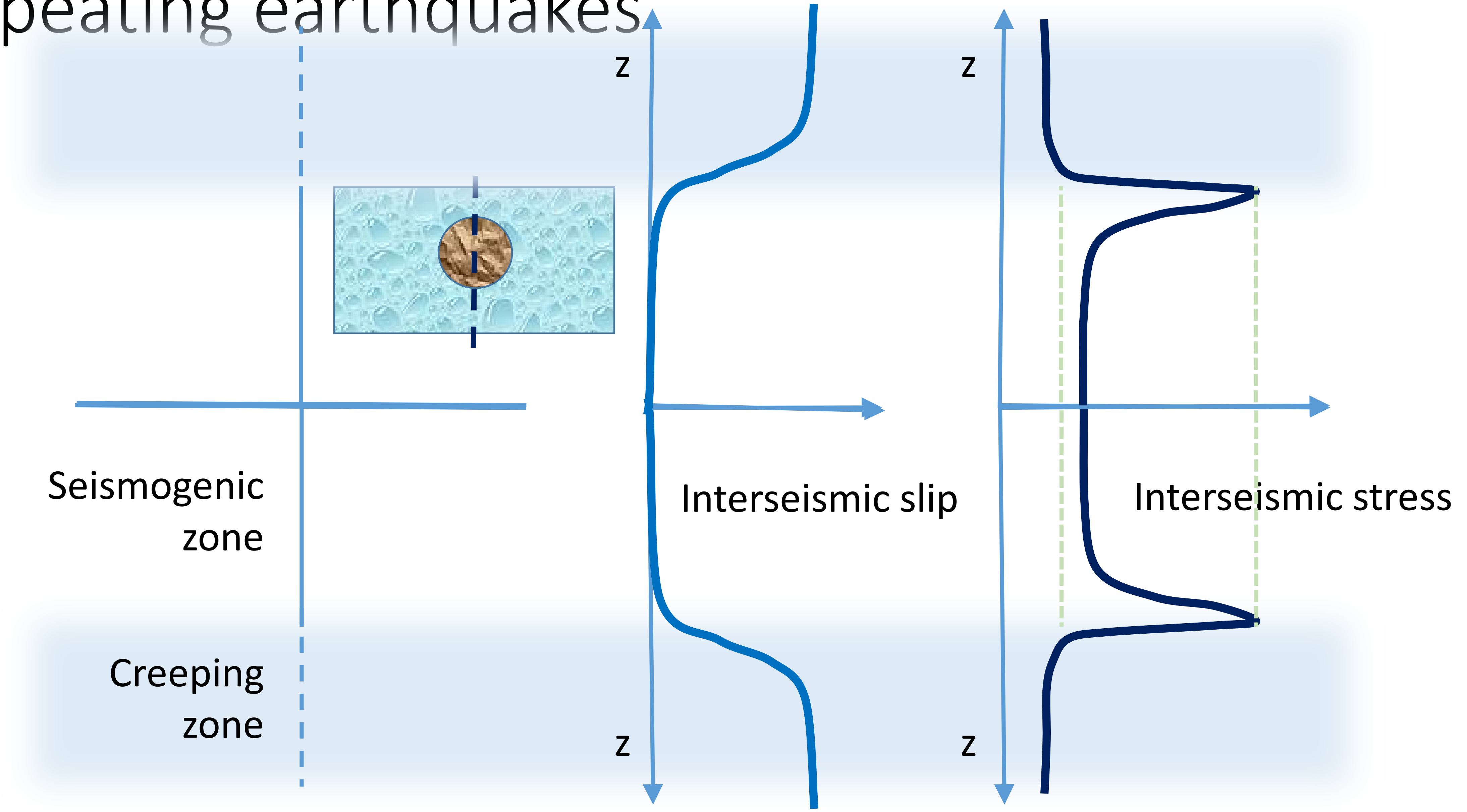
Whereas classical scaling is $T \sim M_0^{1/3}$

Repeating earthquakes

Model: a circular brittle patch (radius R) embedded in a creeping fault



Repeating earthquakes



Recurrence time scaling of repeating earthquakes

Repeating earthquake model: a circular brittle patch (radius R) embedded in a creeping fault (steady slip rate V_{creep})

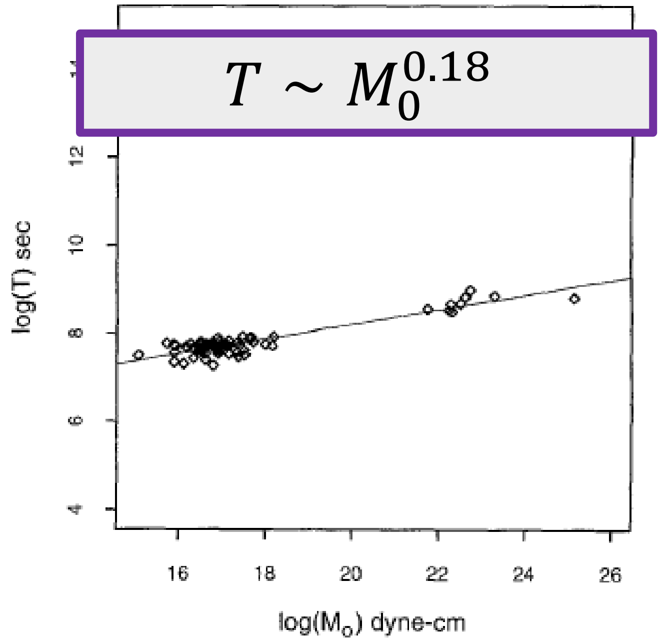
From fracture mechanics $\Delta\tau \sim \sqrt{2\mu G_c/R}$

From elasticity: $\Delta\tau \sim \mu D/R$

Slip budget: $D = V_{creep}T$ per event

Seismic moment: $M_0 = \mu\pi R^2 D$

$$\rightarrow T \sim \left(\frac{2G_c}{\mu}\right)^{\frac{2}{5}} \frac{1}{V_{creep}} M_0^{\frac{1}{5}}$$



How do earthquakes start?

Do small and large earthquakes start differently?

Predictive value of earthquake onset and foreshock sequences?

- Seismological observations: seismic nucleation phase, foreshock sequences
- Laboratory observations
- Friction laws and earthquake nucleation
- Further implications: stress drop, recurrence time, ~~seismicity rate~~, ~~tremors~~