Lecture 8. Physics of Earthquakes

Outline

- Some basic facts and questions
- Recent great earthquakes in Chili (2010, Mw=8.8) and in Japan (2011, Mw=9.0)
- Megathrust earthquakes and structure of the upper plate
- Cross-scale dynamic models

Some basic facts

The cause of larger earthquakes is the plate tectonics and most of them happen at plate boundaries

About 80% of relative plate motion on continental boundaries is accommodated in rapid earthquakes

With few exceptions, earthquakes do not generally occur at regular intervals in time or space.

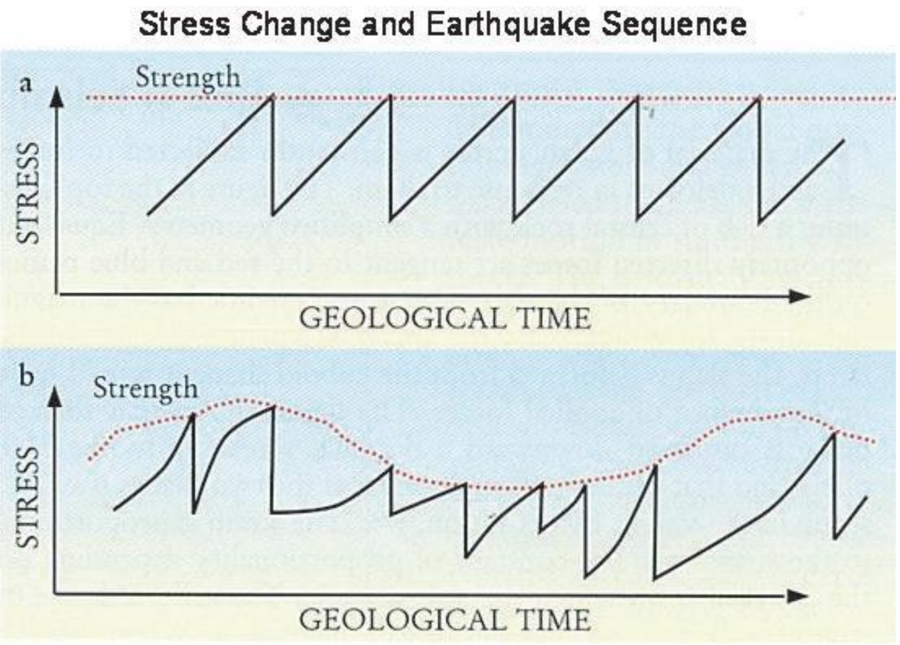
Some basic facts

The shear strain change associated with large earthquakes (*i.e. coseismic strain drop*) is of the order of 10^{-5}– 10^{-4}. This corresponds to a change in shear stress (*i.e. static* stress drop) of about 1–10 MPa.

The repeat times of major earthquakes at a given place are about 100–1000 years on plate boundaries, and 1000–10 000 years within plates.

The rupture velocity for large earthquakes is typically 75–95% of the S-wave velocity

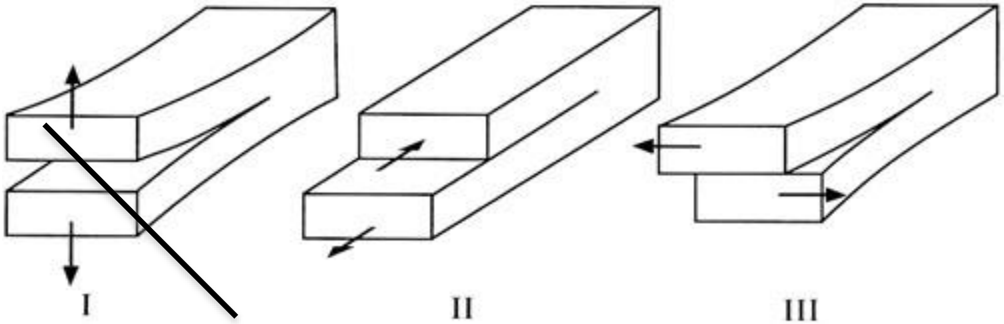
Some basic facts



Ideal

Real

Deformation modes



rare

Some basic facts

Definitions and scaling

Seismic moment: $M_0 = G \cdot D \cdot S$, G-shear modulus, D-average displacement, S-rupture area

Average stress drop $$\overline{\Delta\sigma_s} = \frac{1}{S}\int_S \Delta\sigma_s \, \mathrm{d}S.$$

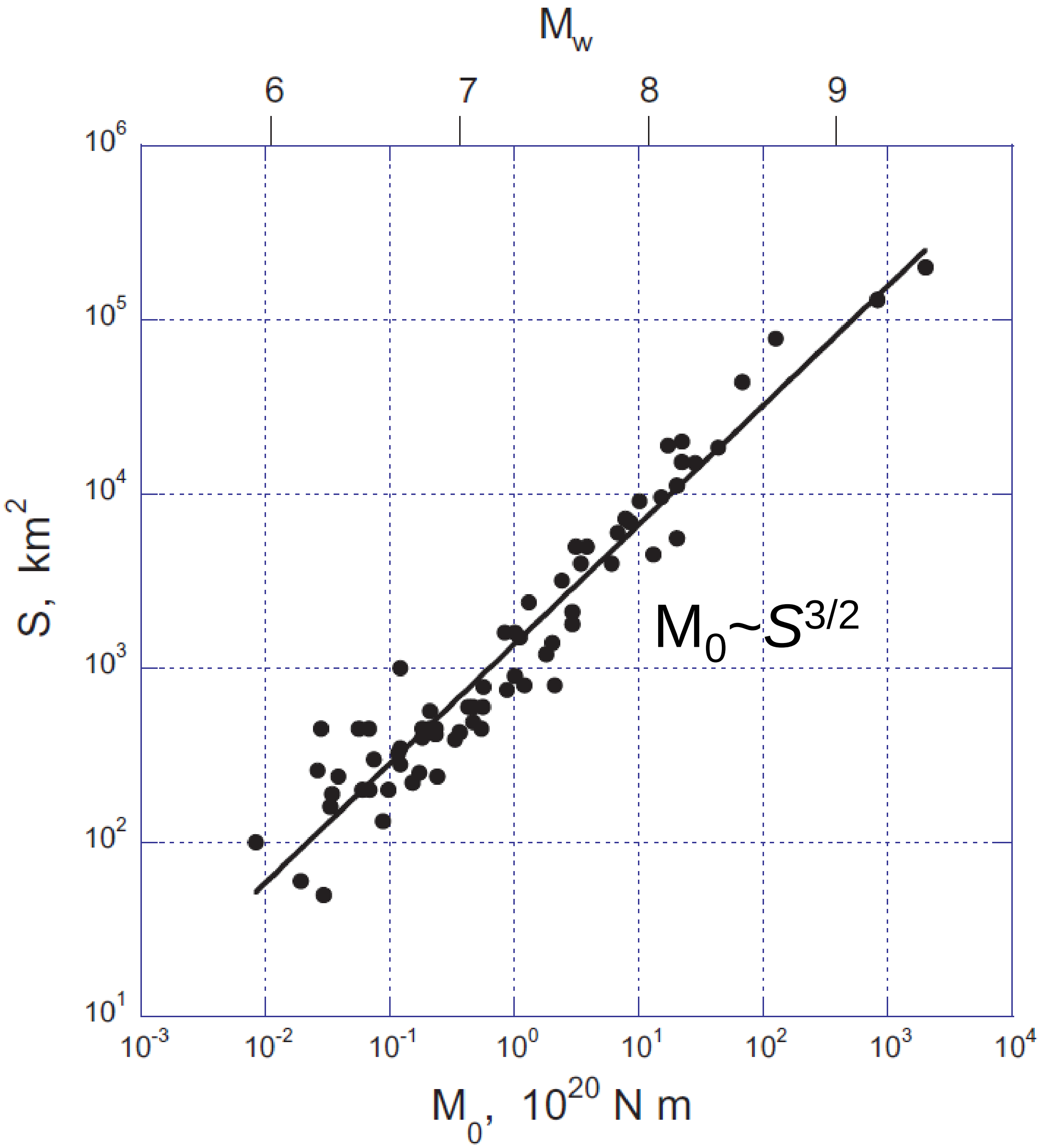
$\overline{\Delta\sigma_s} \approx C \cdot G \cdot D / L$, L-characteristic rupture length $L \approx S^{1/2}$

$\overline{\Delta\sigma_s} \approx C \cdot M_0 \cdot S^{-3/2}$ or

$M_0 \approx \overline{\Delta\sigma_s} \cdot S^{3/2}$; $D \approx S^{1/2} \, \overline{\Delta\sigma_s} / G$

Moment magnitude: $M_w = 2/3 \log_{10}(M_0) - 6.07$

Some basic facts



Kanamori and Brodsky, 2004

That means

$$\overline{\Delta\sigma_s} \approx const$$

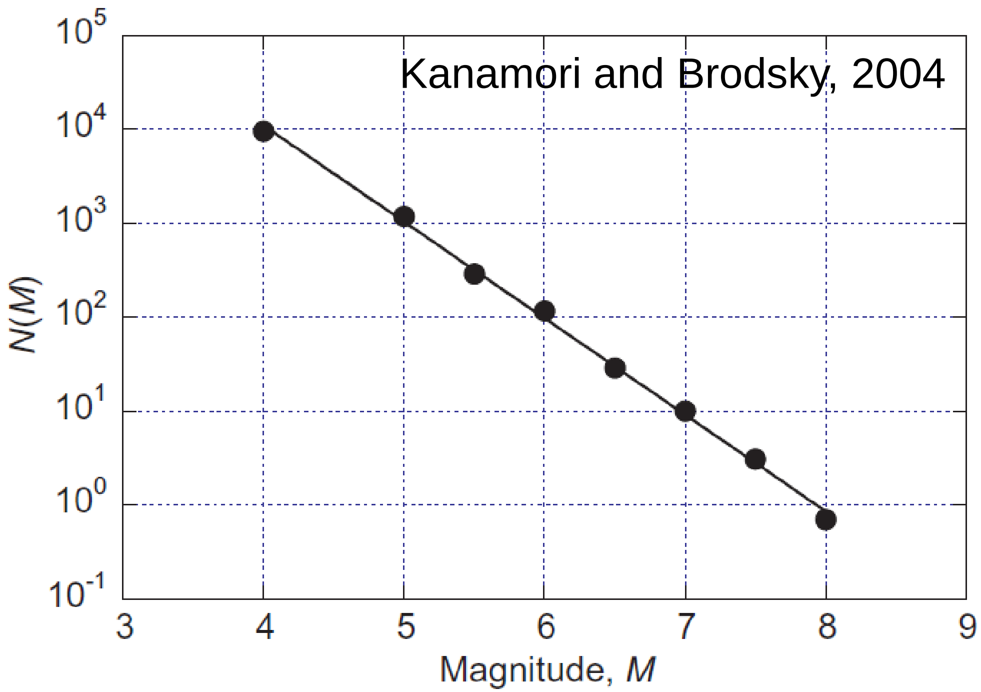
Mean value of $\overline{\Delta\sigma_s}$ is about 3 MPa

$$M_0 \approx \overline{\Delta\sigma_s} \cdot S^{3/2};$$

$$D \approx S^{1/2}\, \overline{\Delta\sigma_s} / G$$

Some basic facts

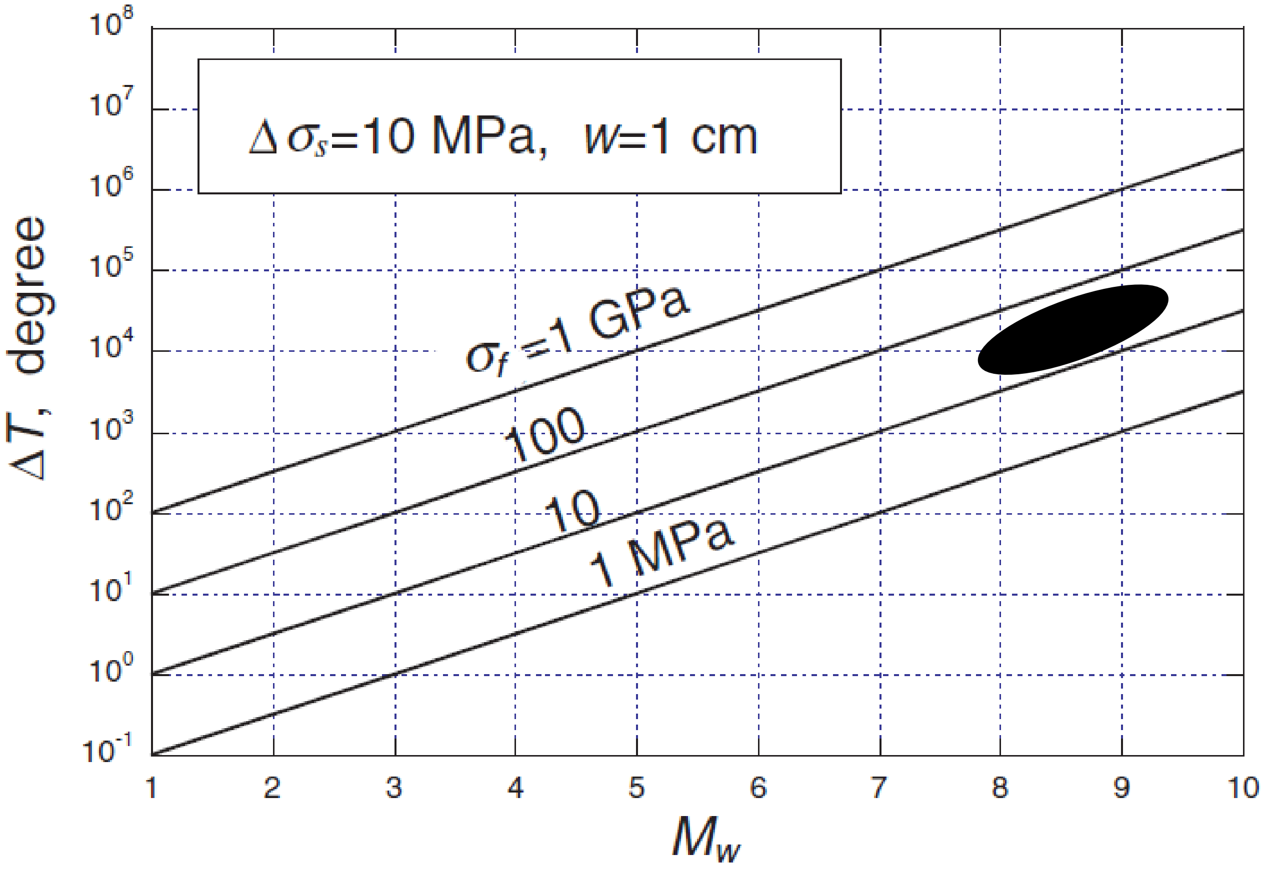
The magnitude–frequency relationship (the Gutenberg–Richter relation)



$\log N(M) = a - bM$, b is about 1

Thermal effect of Eq.

$$\Delta T = \frac{Q}{C\rho S w} = \frac{\sigma_{\mathrm{f}} D}{C\rho w},$$



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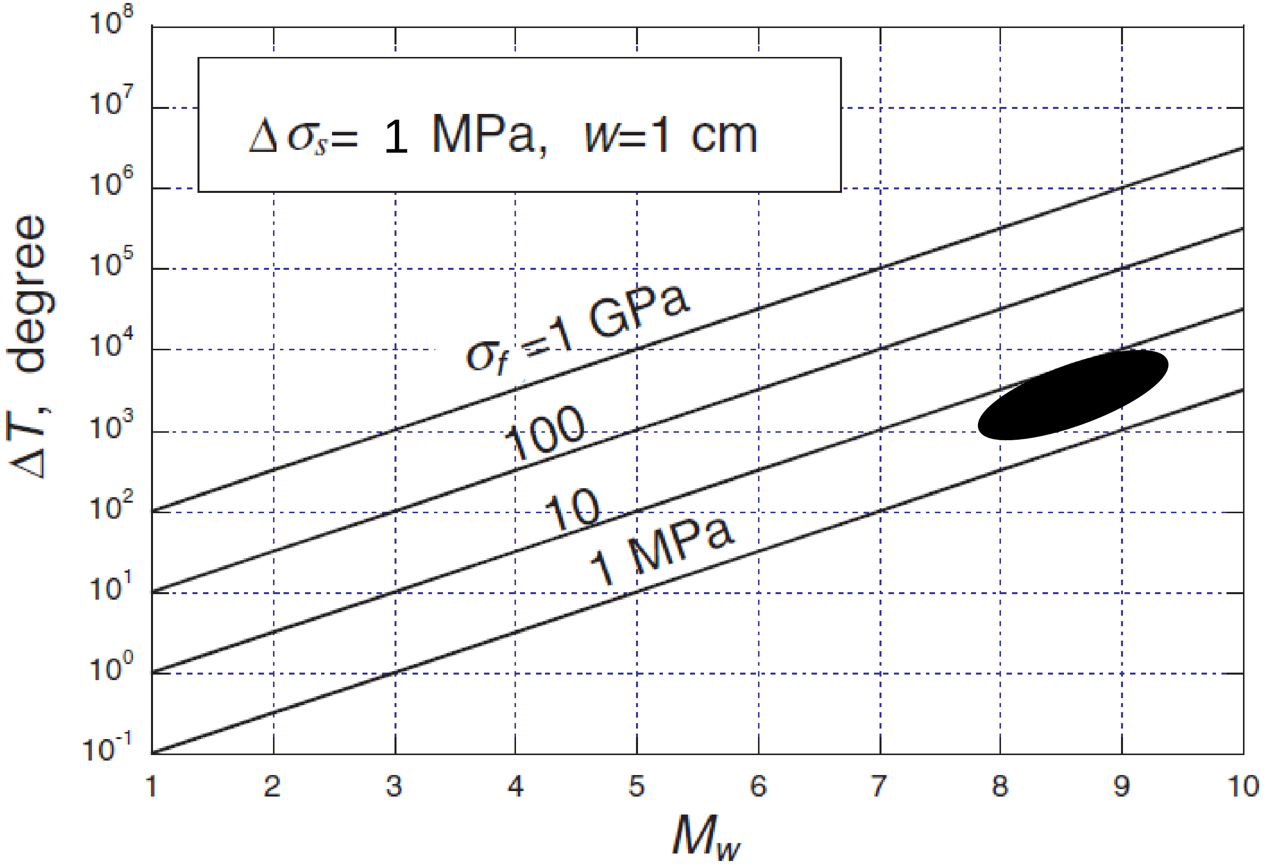


Some basic questions

Why some plate boundaries glide past each other smoothly, while others are punctuated by catastrophic failures?

Why do some earthquakes stop after only a few hundred meters while others continue rupturing for a thousand kilometers?

How do nearby earthquakes interact?

Why are earthquakes sometimes triggered by other large earthquakes thousands of kilometers away?

Great Earthquakes challenges

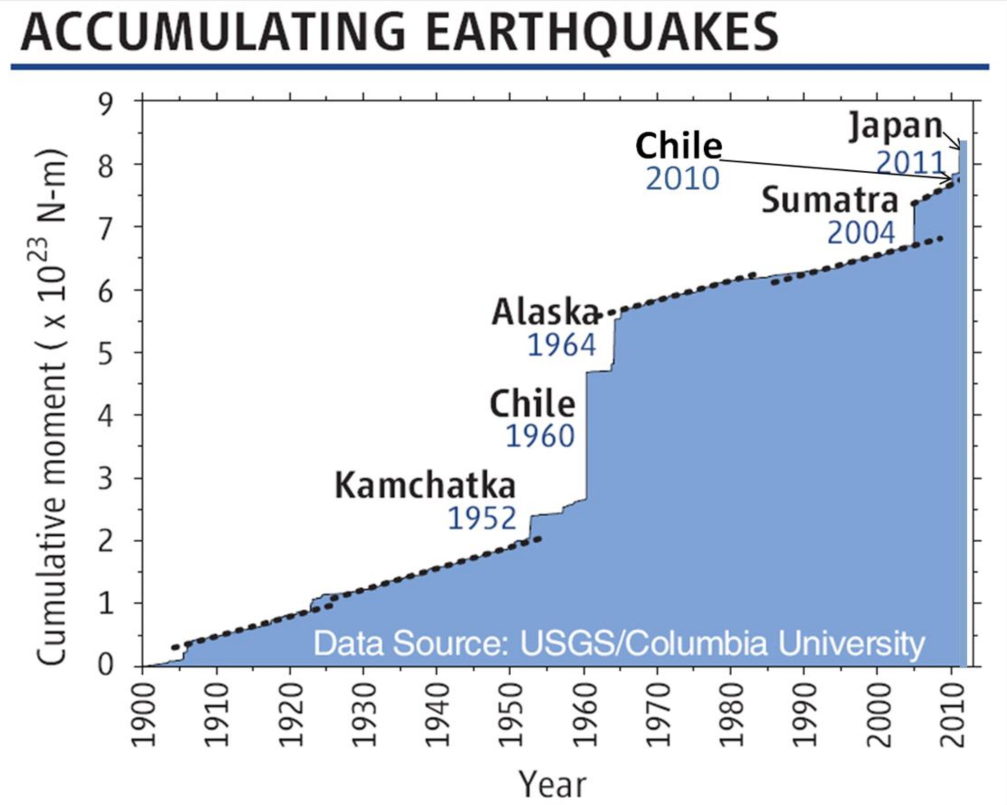
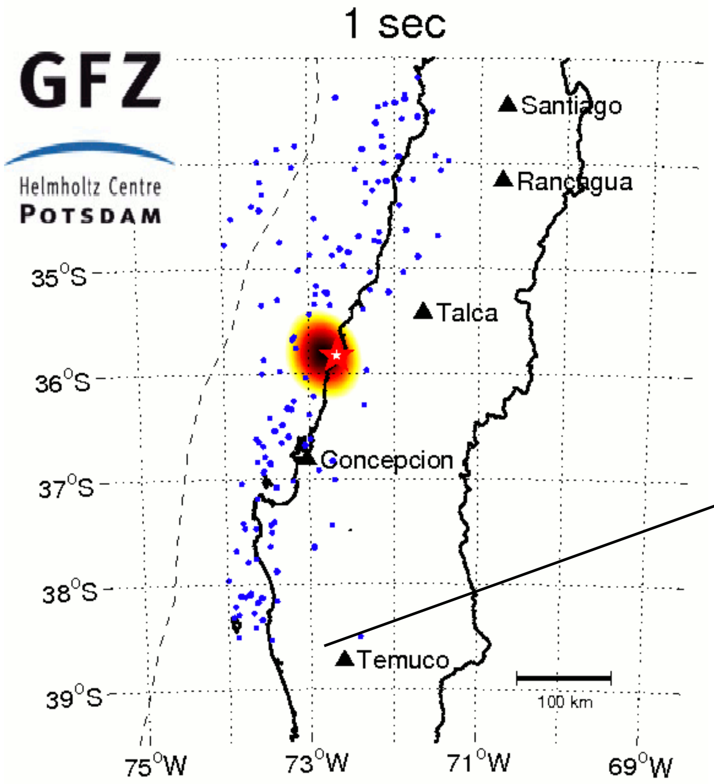


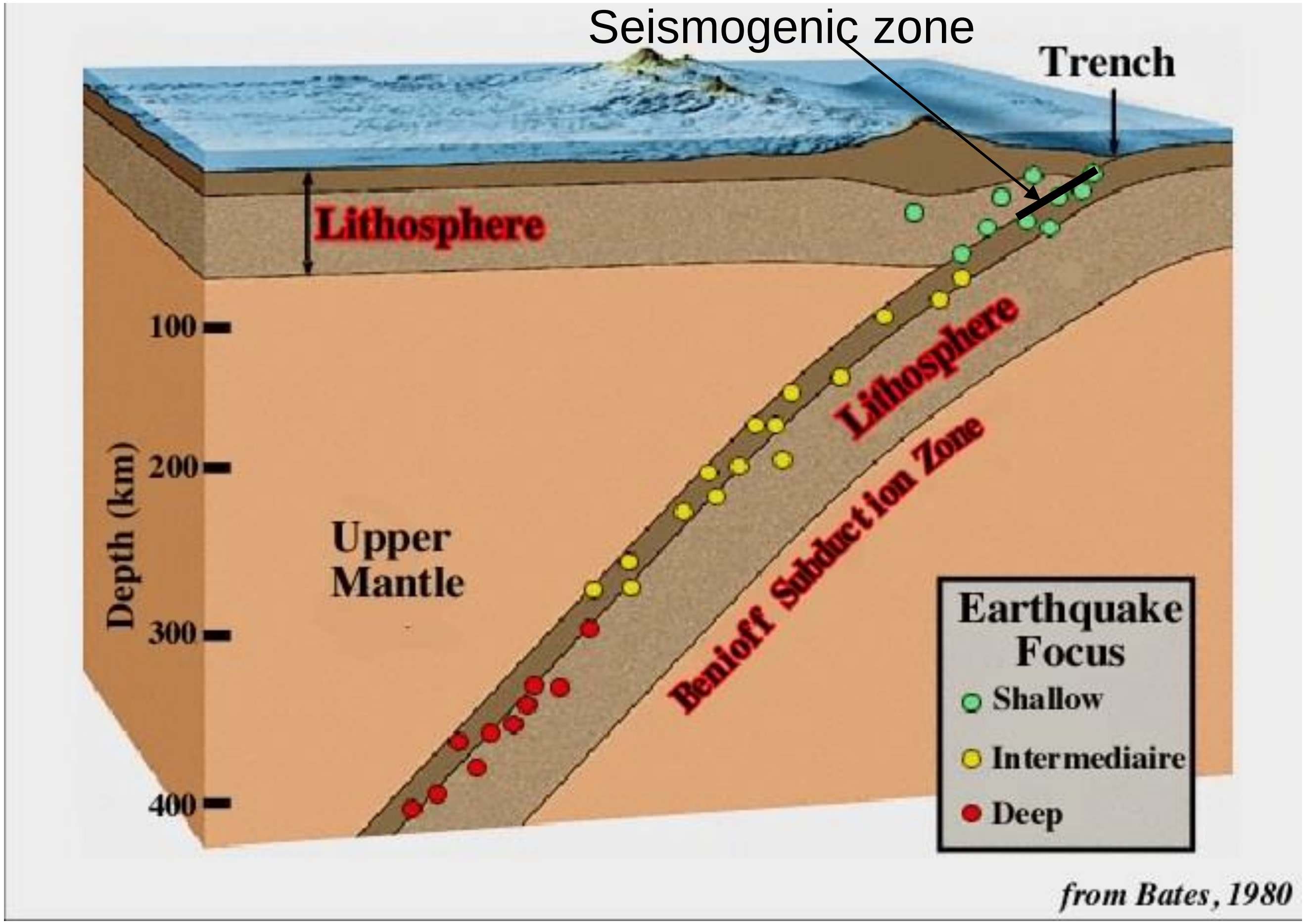
Figure 1. Cumulative seismic moment versus time. Modified from [Kerr, 2011]

Why the greatest earthquakes occur in the weakest zones? Do they indeed cluster?



Chile earthquake (2010, Mw=8.8)

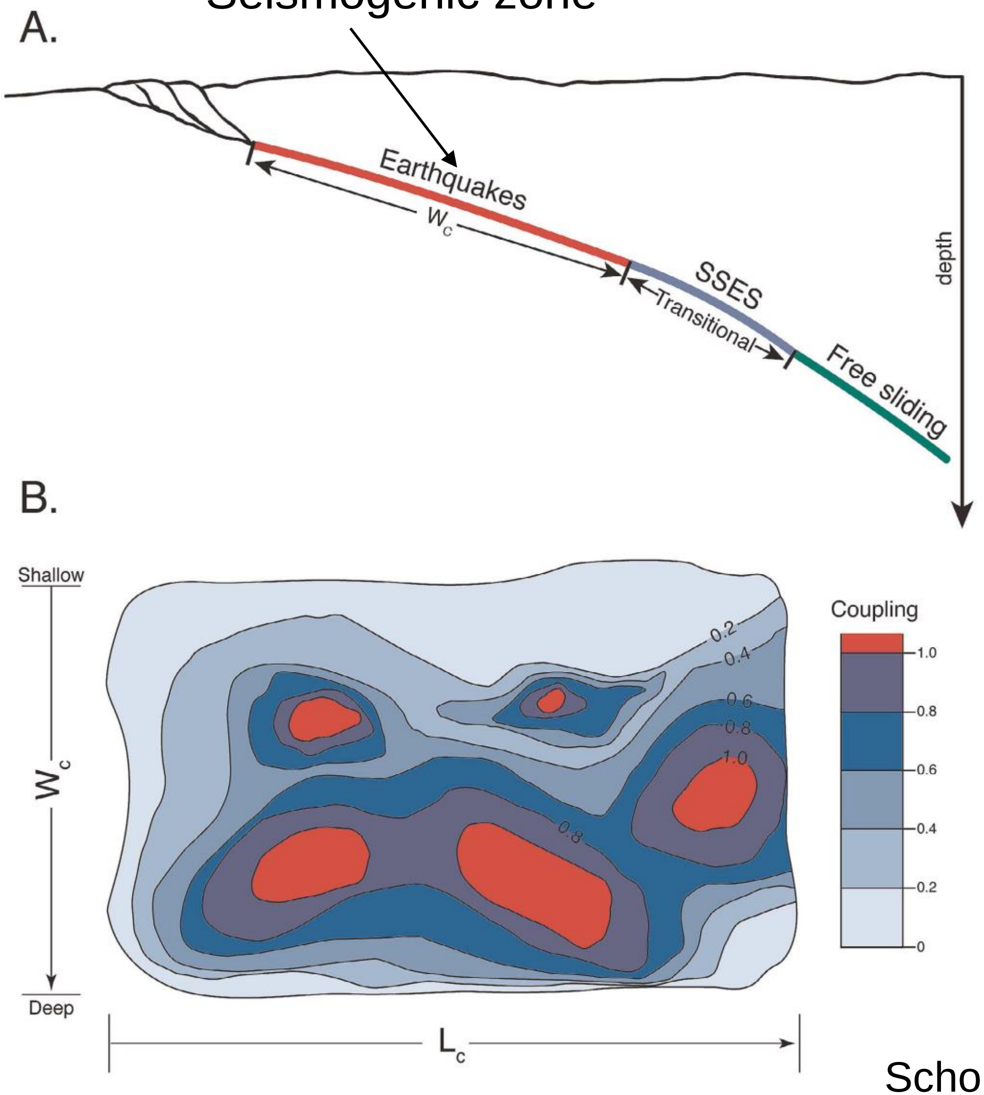
Subduction zone earthquakes



from Bates, 1980

Subduction zone earthquakes

Seismogenic zone



Scholz and Campos, 2012

Coupling paradox

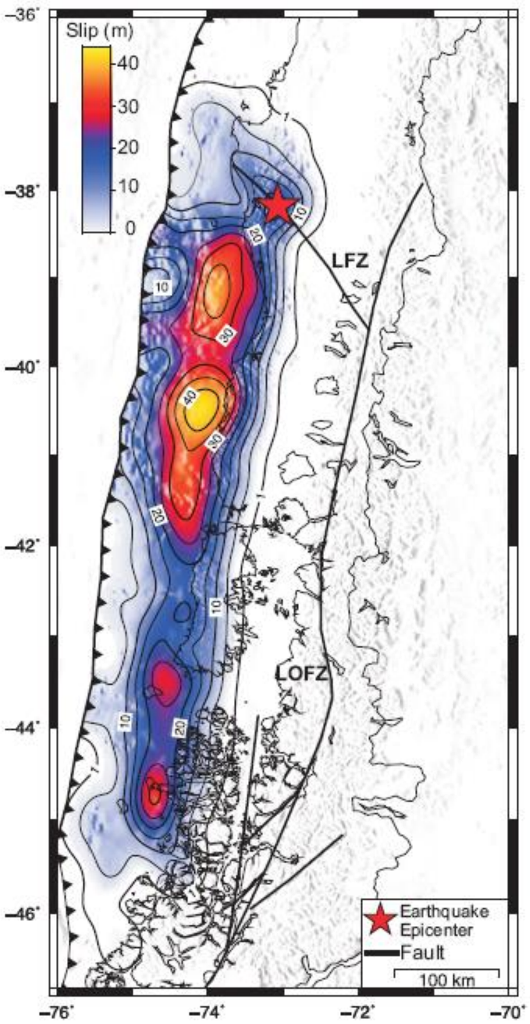
Is the idea about low mechanical coupling at subduction zones consistent with the occurrence there great earthquakes?

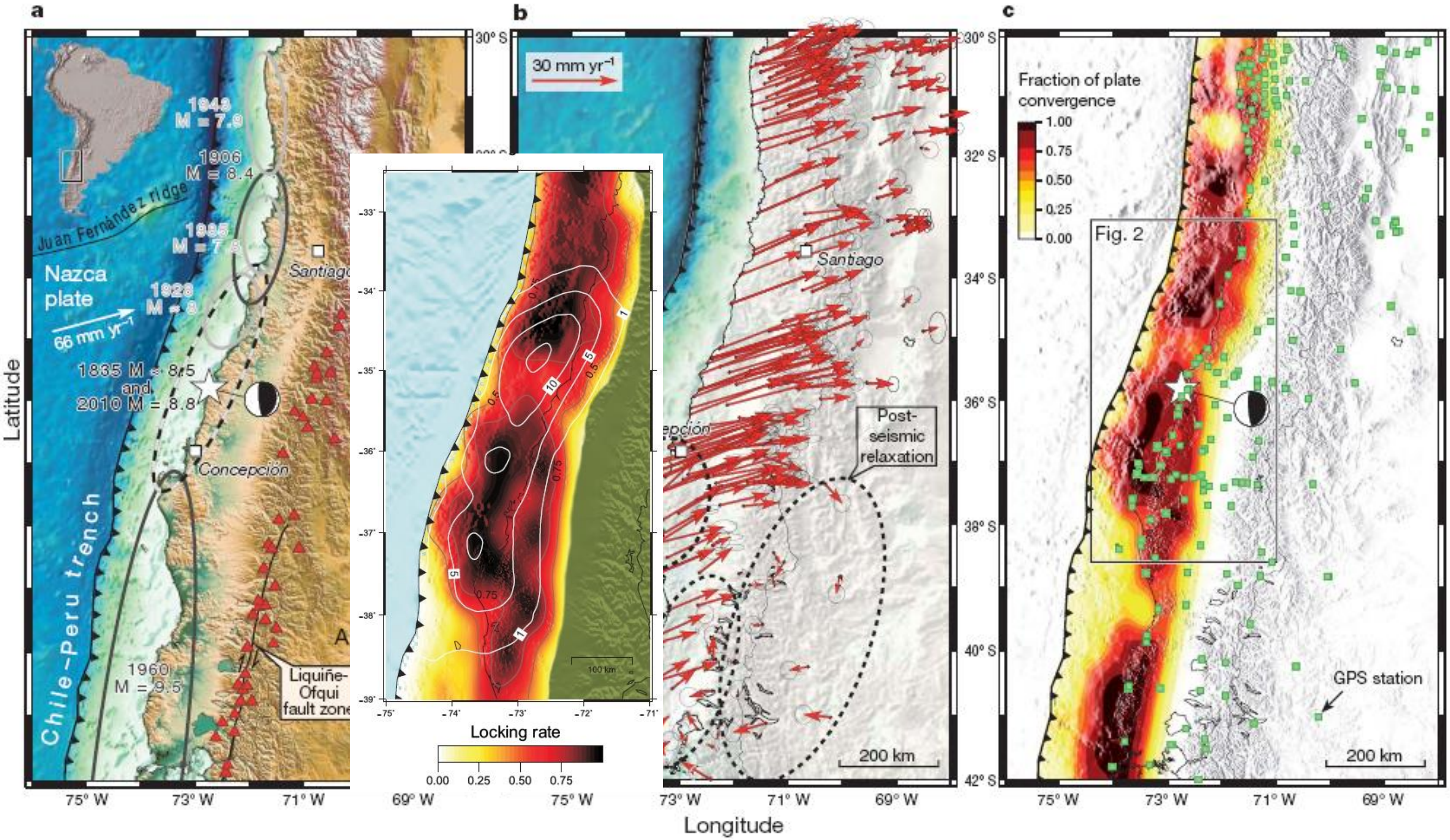
Great earthquakes may well happen within the very weak fault zones (subduction channels) with static friction about 0.01-0.05 due to the friction drop of about 0.005-0.01.

What makes earthquake great is not large stress drop, but rupturing at large area (homogeneous channel structure, no barriers).

Valdivia earthquake (1960)

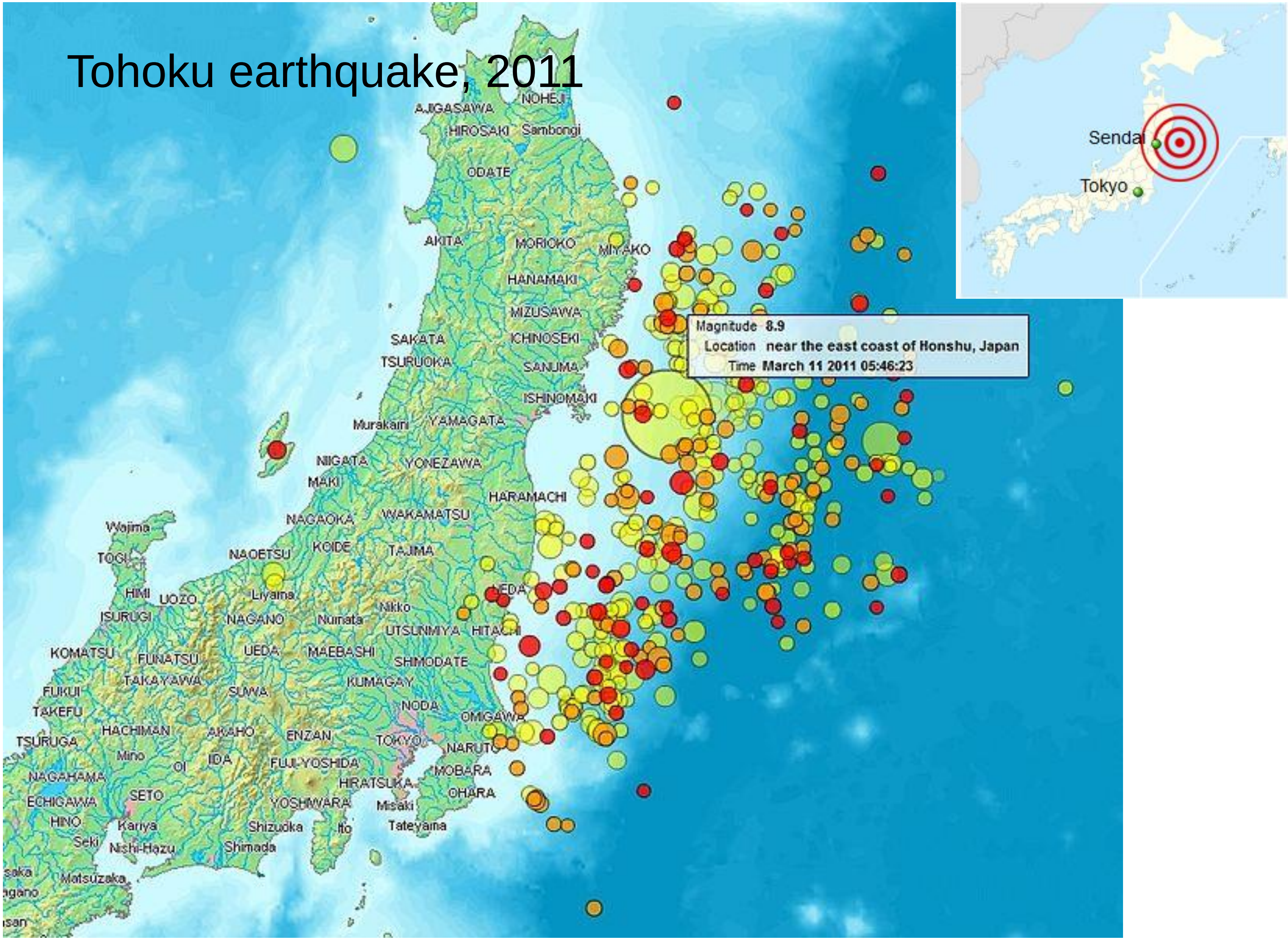
Slip distribution





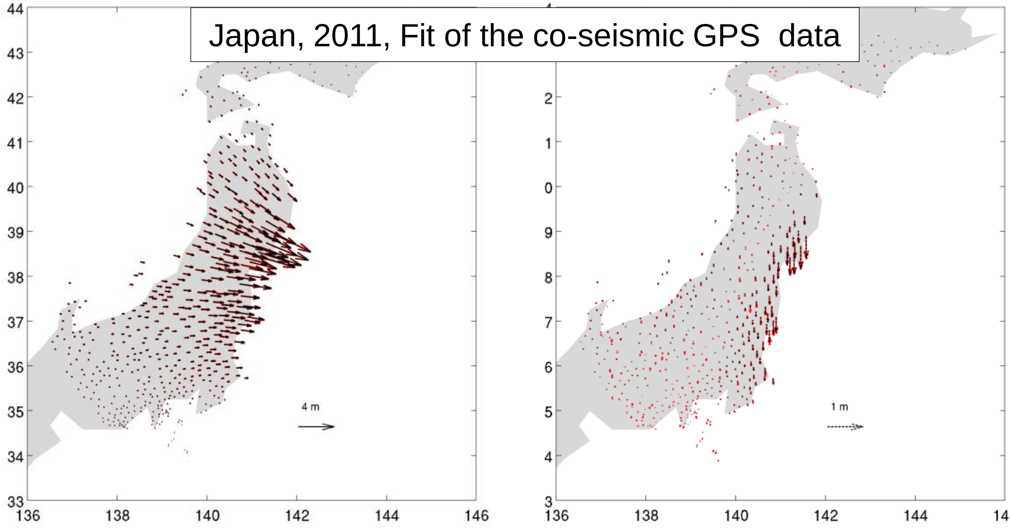
Moreno et al., 2010

Tohoku earthquake, 2011

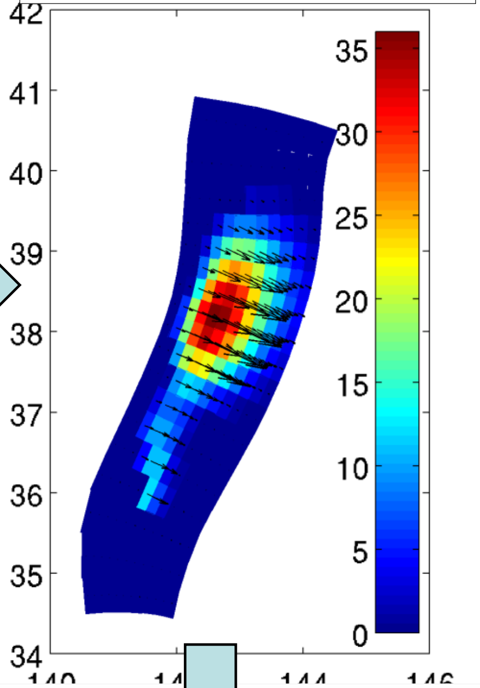

Magnitude 8.9
Location near the east coast of Honshu, Japan
Time March 11 2011 05:46:23

Tohoku Great Earthquake, 2011 (Mw=9.0)

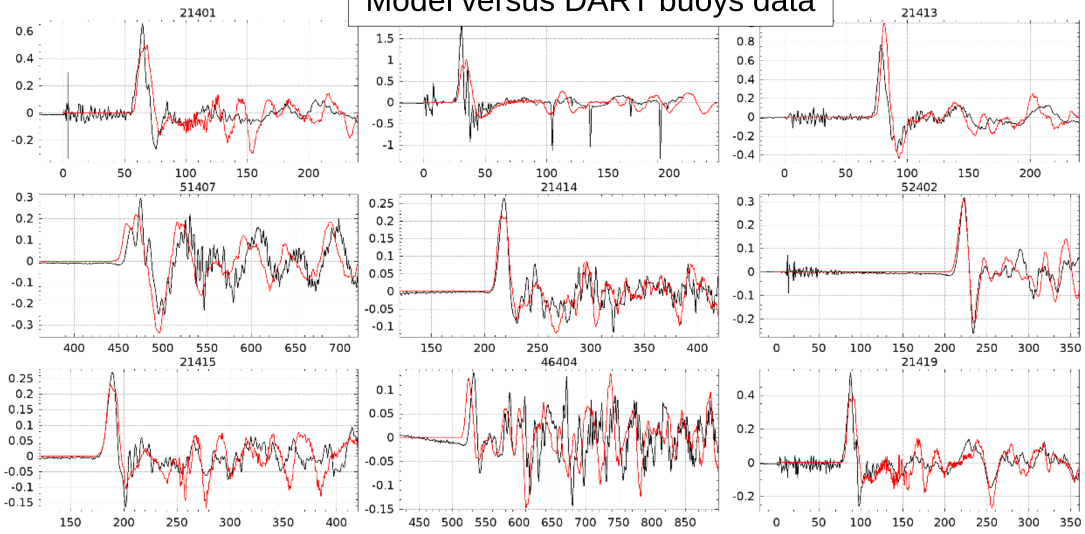
Japan, 2011, Fit of the co-seismic GPS data

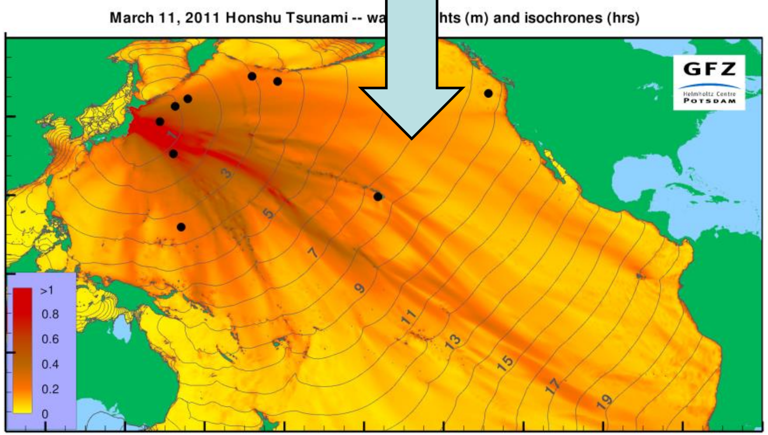


Japan, 2011, Inverted Slip, m



Model versus DART buoys data

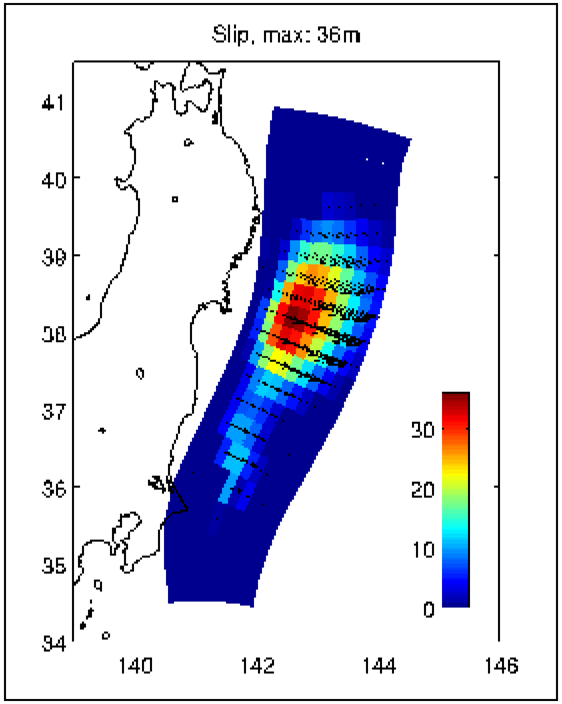


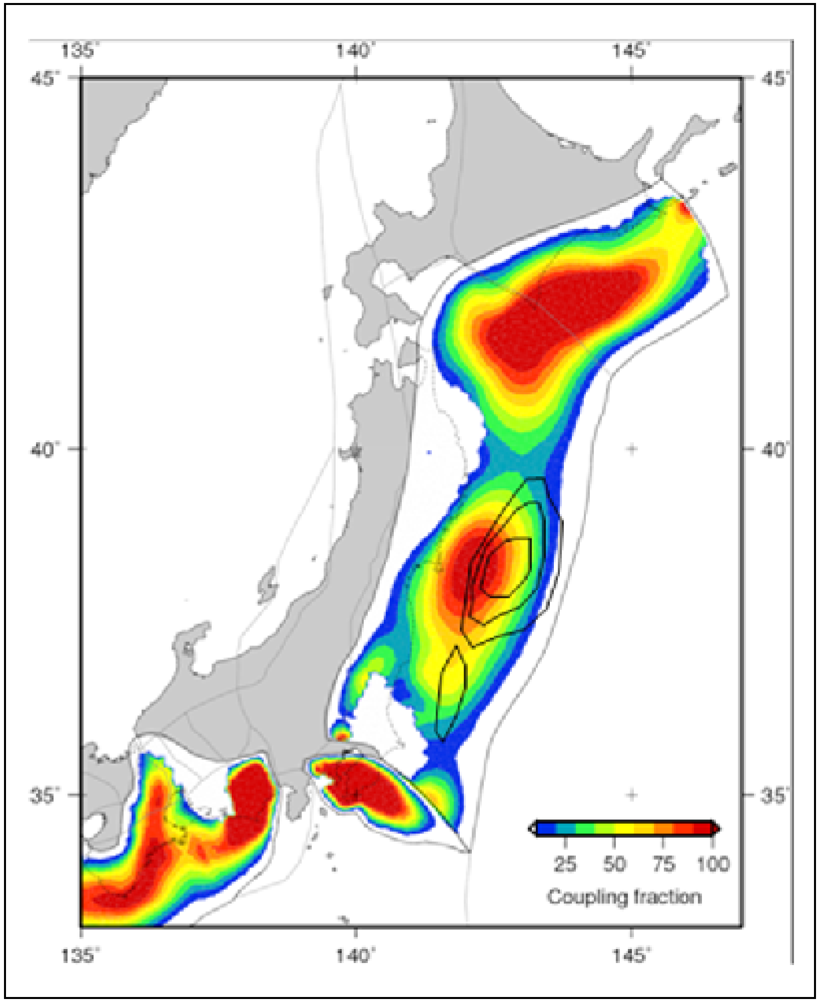


Tsunami based on source from GPS data inversion

Hoechner et al. 2013

Locking of plates





Loveless and Meade, J. Geophys. Res. 2010

Perspectives: Cross-scale dynamic models

Elastic deformation is included in our geological-time-scale (mln years) Andes model

Full set of equations

$$\frac{1}{K}\frac{DP}{Dt}-\alpha\frac{DT}{Dt}+\frac{\partial v_i}{\partial x_i}=0$$

mass

$$-\frac{\partial P}{\partial x_i}+\frac{\partial \tau_{ij}}{\partial x_j}+\rho g_i=\rho\frac{Dv_i}{Dt}$$

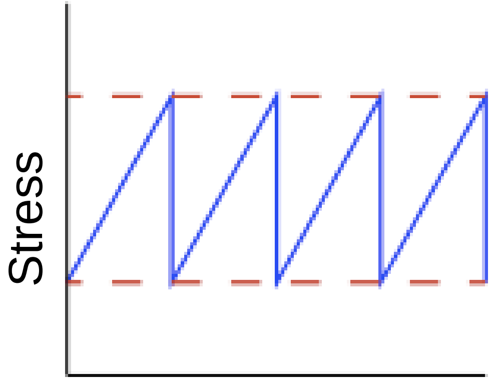
momentum

$$\rho C_p\frac{DT}{Dt}=\frac{\partial}{\partial x_i}(\lambda\frac{\partial T}{\partial x_i})+\tau_{II}\dot{\varepsilon}_{II}+\rho A$$

energy

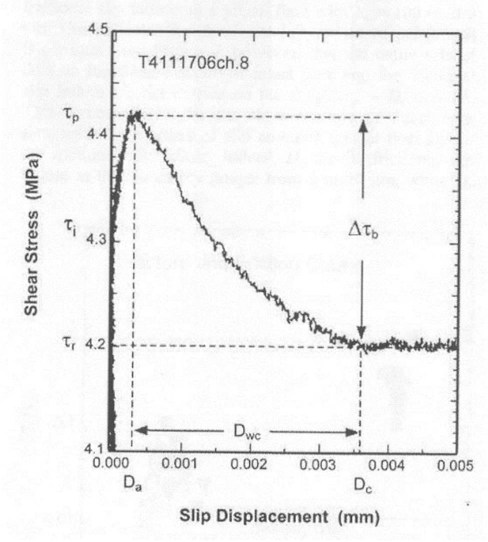
$$\dot{\varepsilon}_{ij}=\frac{1}{2}(\frac{\partial v_i}{\partial x_j}+\frac{\partial v_j}{\partial x_i})=\underline{\frac{1}{2G}\frac{D\tau_{ij}}{Dt}}+\frac{1}{2\eta_{eff}}\tau_{ij}$$

Frictional instabilities governed by static-kinetic friction

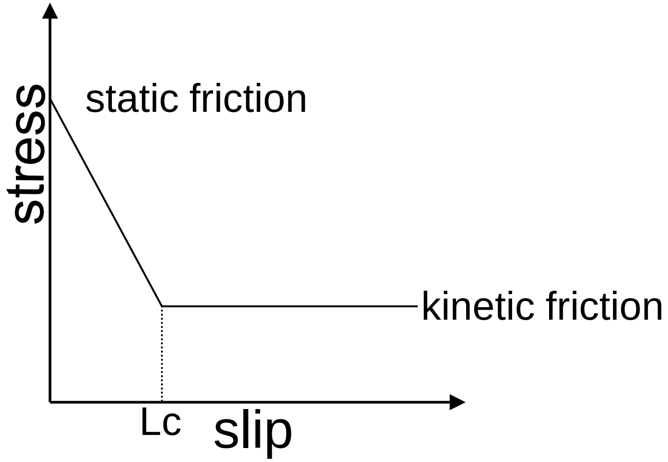


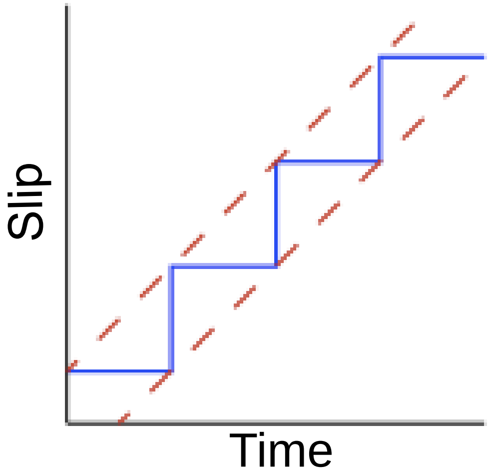
The static-kinetic (or slip-weakening) friction:

experiment → Constitutive law



Ohnaka (2003)





Frictional instabilities governed by rate- and state-dependent friction

Dieterich-Ruina friction:

$$\frac{\tau}{\sigma_n} = \mu = \mu^* + a\ln\left(\frac{V}{V^*}\right) + b\ln\left(\frac{\theta V^*}{D_C}\right)$$

and

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

At steady state:

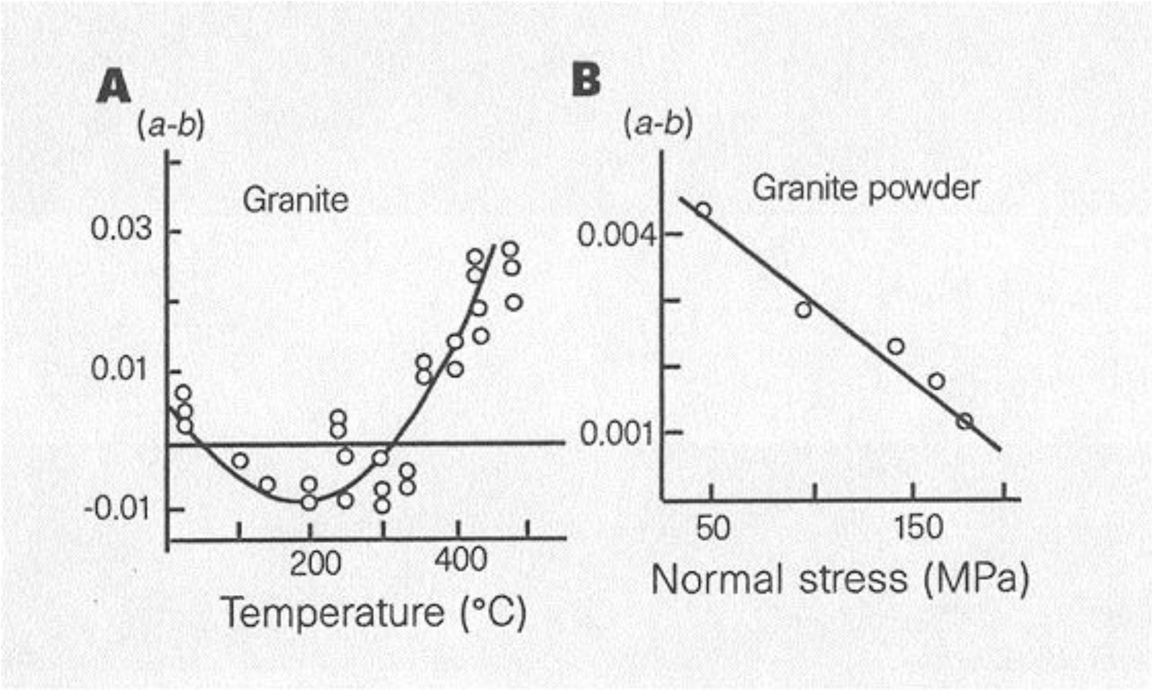
$$\mu = \mu^* + (a-b)\ln\left(\frac{V}{V^*}\right)$$

were:

- V and θ are sliding speed and contact state, respectively.
- a, b and α are non-dimensional empirical parameters.
- D_c is a characteristic sliding distance.
- The * stands for a reference value.

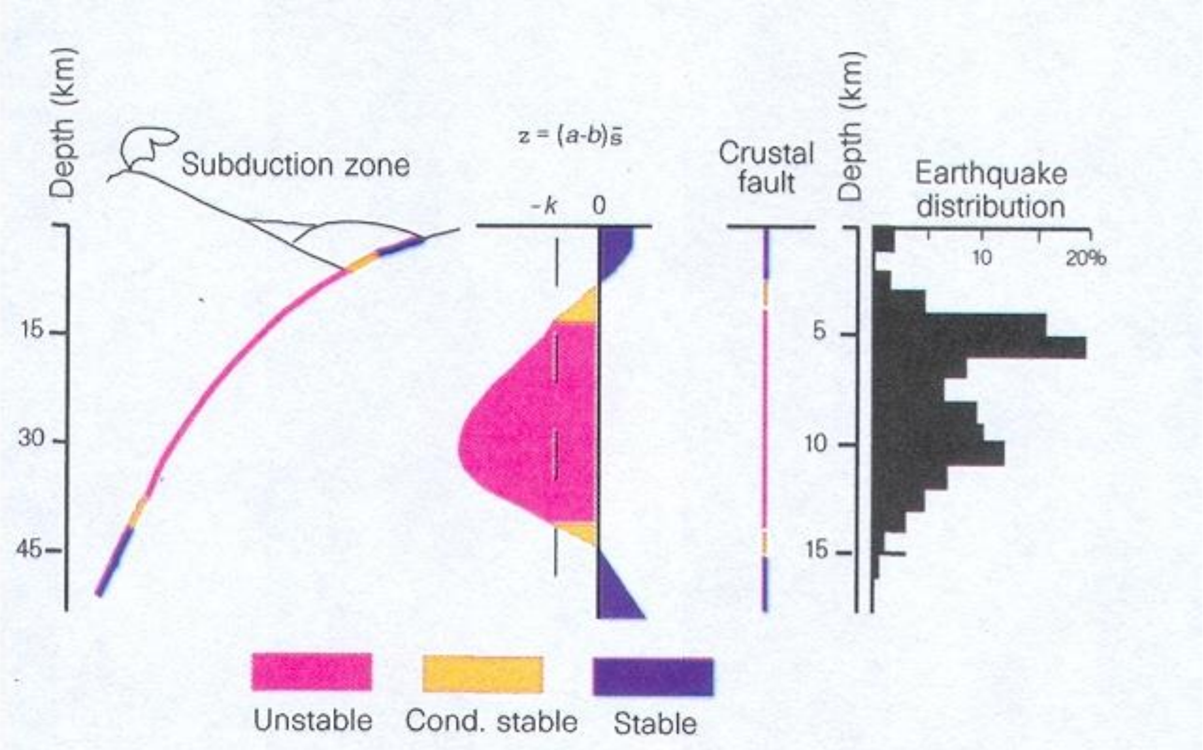
How b-a changes with depth ?

- Note the smallness of b-a.



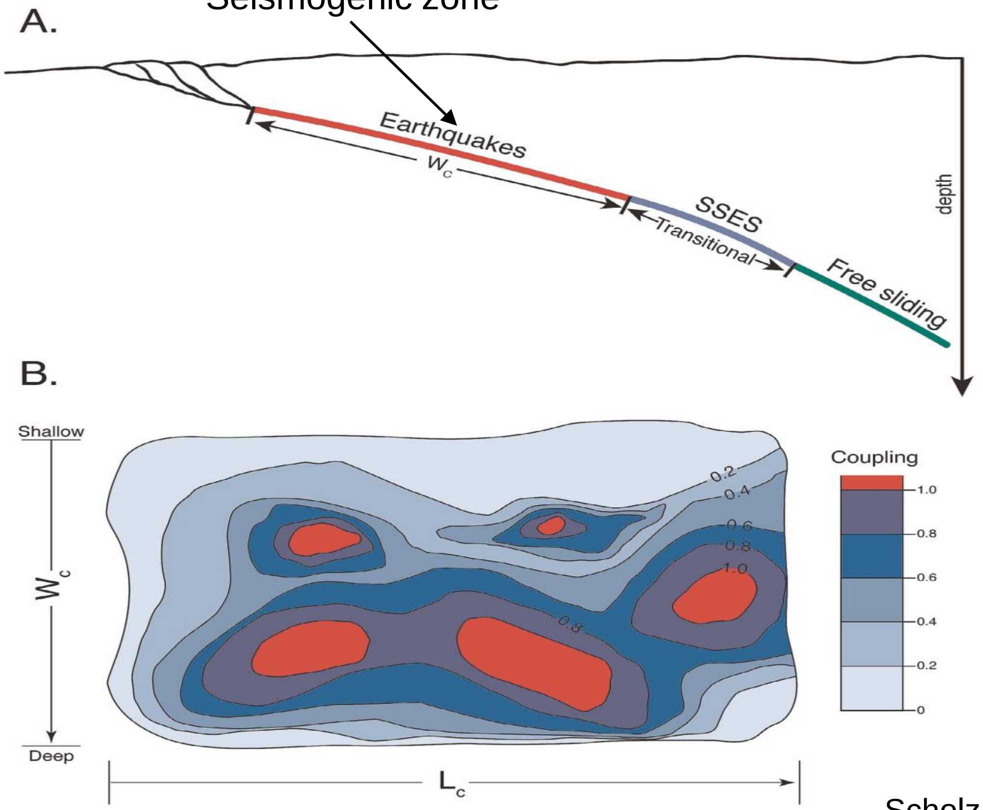
Scholz, Nature 1998 and references therein

The depth dependence of b-a may explain the seismicity depth distribution



Scholz (1998) and references therein

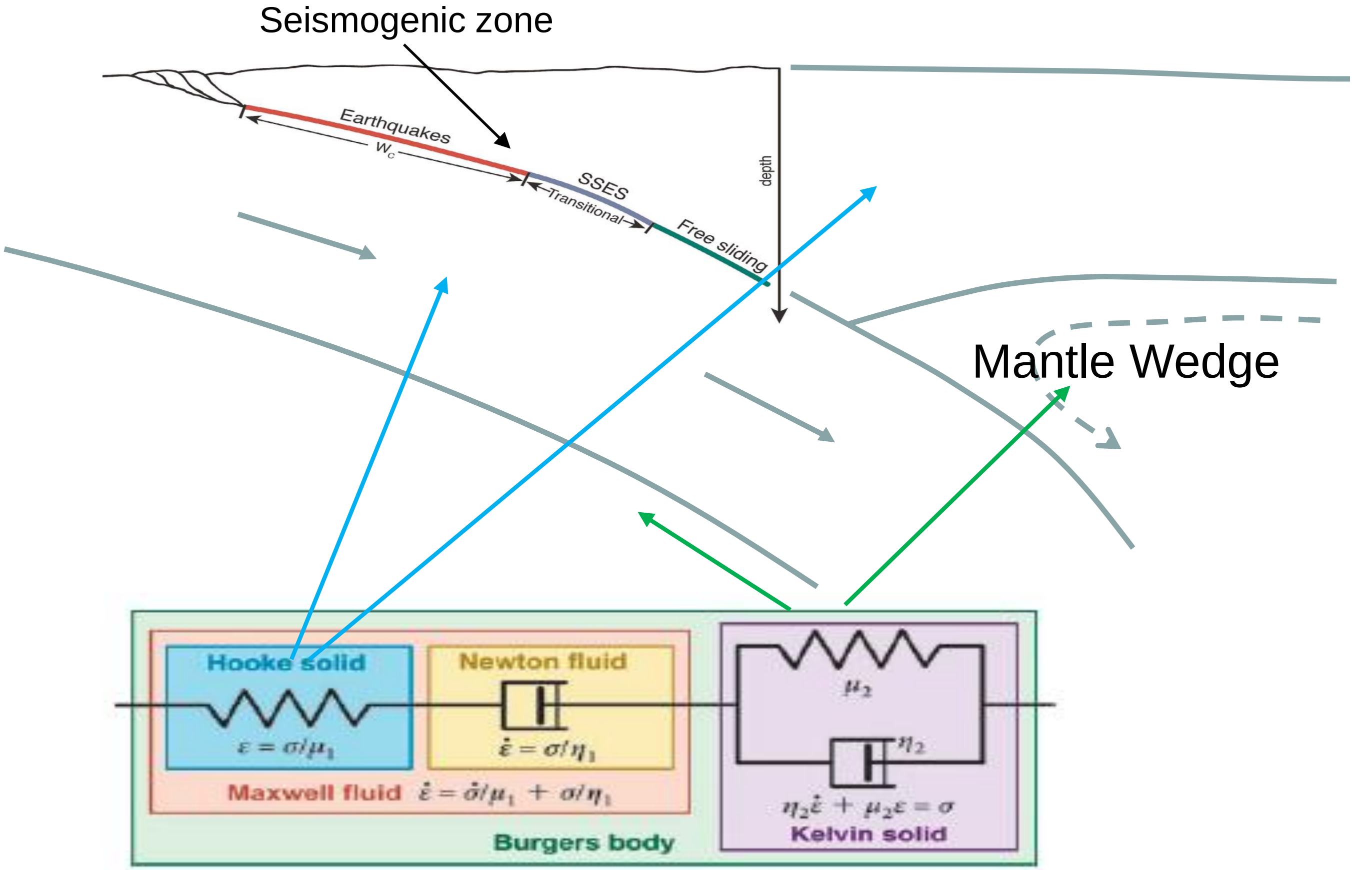
Subduction zone earthquakes



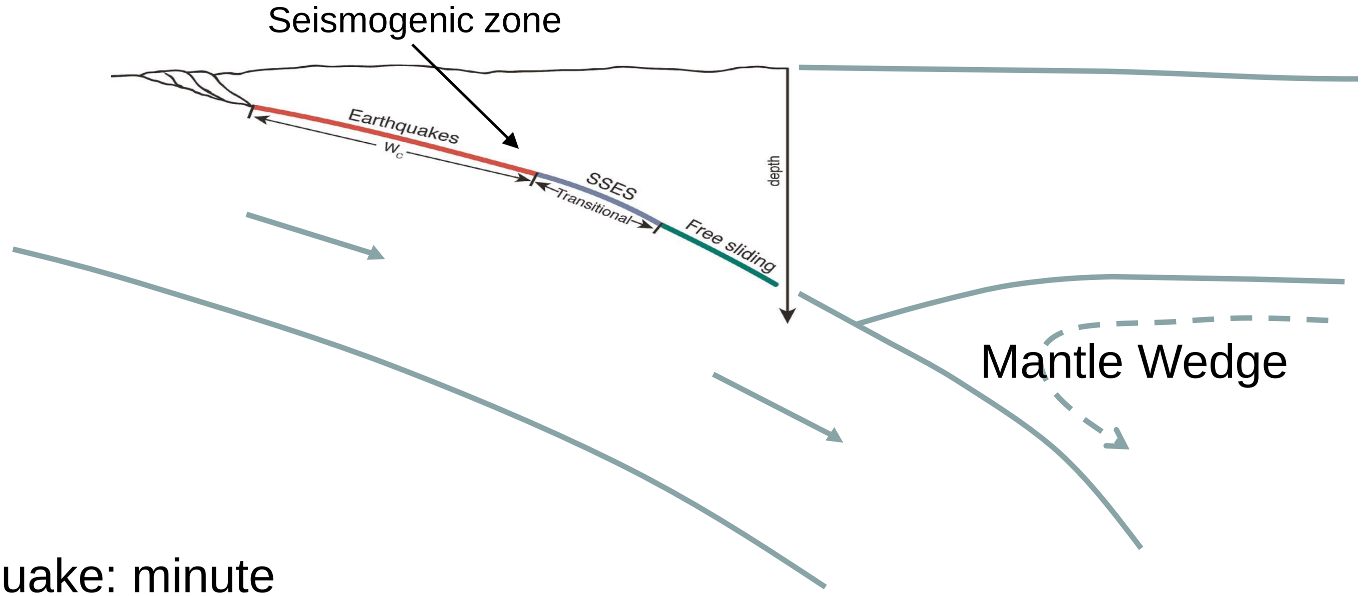
Seismogenic zone

Scholz and Campos, 2012

Subduction zone earthquakes



Subduction zone earthquakes



1. Earthquake: minute
2. Aftertslip (fault control) **<u>hours-1 year,</u>** $V \approx 1/t$
3. Visco-elastic relaxation (wedge control) **<u>year-decades</u>**

Our aim was to develop the thermo-mechanical model able to:

- Replicate long-term (10^6yr) evolution of subduction zone
- Generate earthquakes as spontaneous mechanical instabilities
- Replicate all stages of seismic cycle and multiple cycles in time scale range from minute to 10^4yr

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- Replicate long-term (10^6yr) evolution of subduction zone
- Generate earthquakes as spontaneous mechanical instabilities
- Replicate all stages of seismic cycle and multiple cycles in time scale range from minute to 10^4yr
- **And all that with mineral-physics-based rheology**

Technique

FEM code SLIM3D
(Popov and Sobolev
PEPI, 2008)

Balance equations

Momentum: $$\frac{\partial \sigma_{ij}}{\partial x_j} + \Delta\rho g\, z_i = 0$$

Energy: $$\frac{DU}{Dt} = -\frac{\partial q_i}{\partial x_i} + r$$

Deformation mechanisms

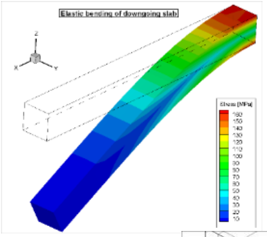
$$\dot{\varepsilon}_{ij} = \dot{\varepsilon}_{ij}^{el} + \dot{\varepsilon}_{ij}^{vs} + \dot{\varepsilon}_{ij}^{pl}$$

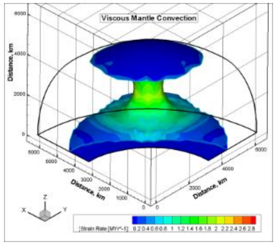
Elastic strain: $$\dot{\varepsilon}_{ij}^{el} = \frac{1}{2G}\hat{\tau}_{ij}$$

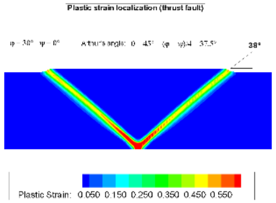
Viscous strain: $$\dot{\varepsilon}_{ij}^{vs} = \frac{1}{2\eta_{eff}}\tau_{ij}$$

Plastic strain: $$\dot{\varepsilon}_{ij}^{pl} = \dot{\gamma}\frac{\partial Q}{\partial \tau_{ij}}$$

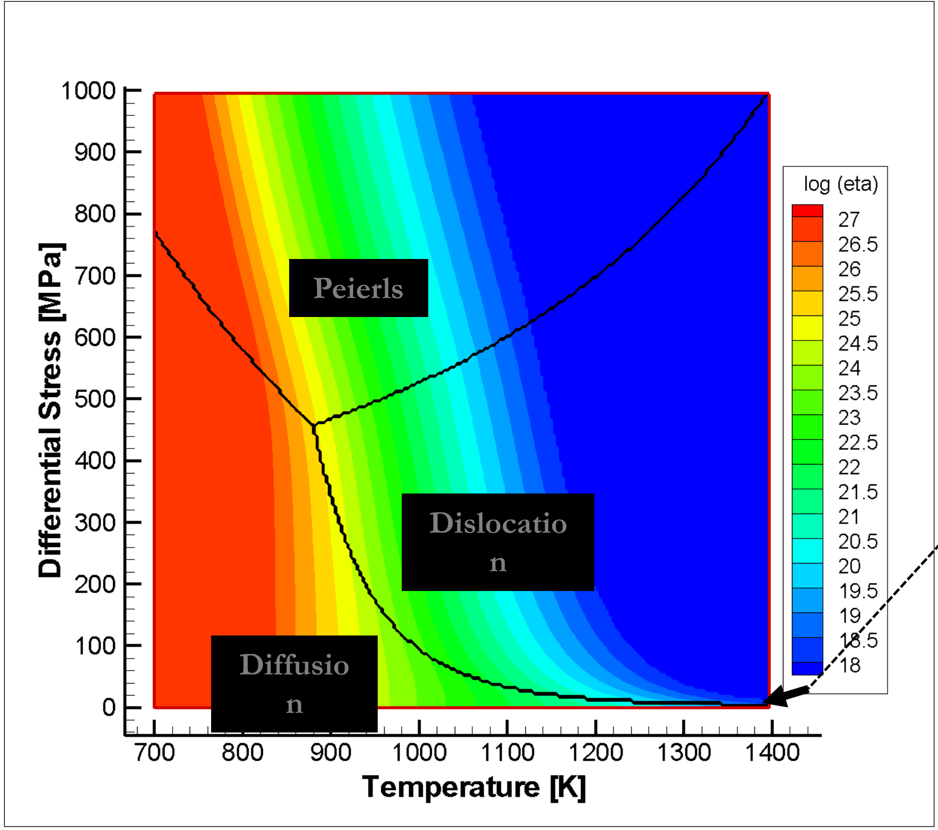
Mohr-Coulomb







Three creep processes



$$\eta_{eff} = \tfrac{1}{2}\tau_{II}\left(\dot{\varepsilon}_L + \dot{\varepsilon}_N + \dot{\varepsilon}_P\right)^{-1}$$

Diffusion creep

$$\dot{\varepsilon}_L = B_L\,\tau_{II}\exp\left(-\frac{E_L}{RT}\right)$$

Dislocation

$$\dot{\varepsilon}_N = B_N\left(\tau_{II}\right)^n\exp\left(-\frac{E_N}{RT}\right)$$

Peierls creep

$$\dot{\varepsilon}_P = B_P\exp\left[-\frac{E_P}{RT}\left(1-\frac{\tau_{II}}{\tau_P}\right)^2\right]$$

(Kameyama *et al.* 1999)

Modification of viscous rheology

Steady power-law dislocation creep

$$\dot{\varepsilon}_{ss} = B \cdot \tau^{n} \exp(-H_{a} / RT)$$

Transient rheology (motivated by Karato (1998))

$$\dot{\varepsilon} = \dot{\varepsilon}_{ss}(1 + (\beta - 1)\exp(-\varepsilon_{visc}^{after\_eq} / \varepsilon_{el}^{eq}))$$

where:

- $\dot{\varepsilon}_{ss}$ is power-law steady state creep strain rate (lab data) ε_{el}^{eq}
- $\varepsilon_{visc}^{after\_eq}$ is elastic strain induced by earthquake
- β is viscous creep strain after the earthquake
- is a constant about 10

Modification of brittle rheology

Rate and state friction law

Dieterich-Ruina friction:

$$\frac{\partial \tau}{\partial \sigma_n} = \mu = \mu^* + a \ln\left(\frac{V}{V^*}\right) + b \ln\left(\frac{\theta V^*}{D_C}\right)$$

and

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

were:

- V and θ are sliding speed and contact state, respectively.
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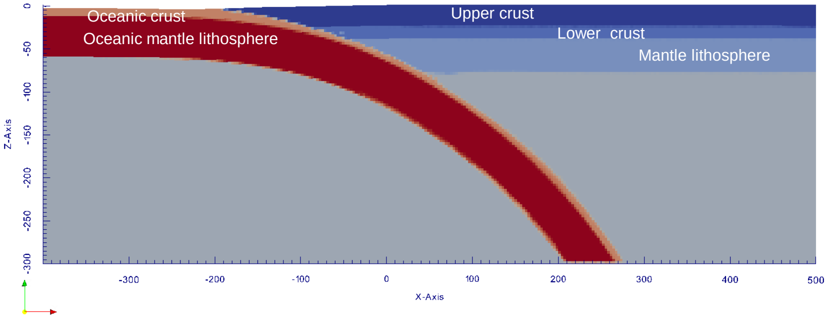
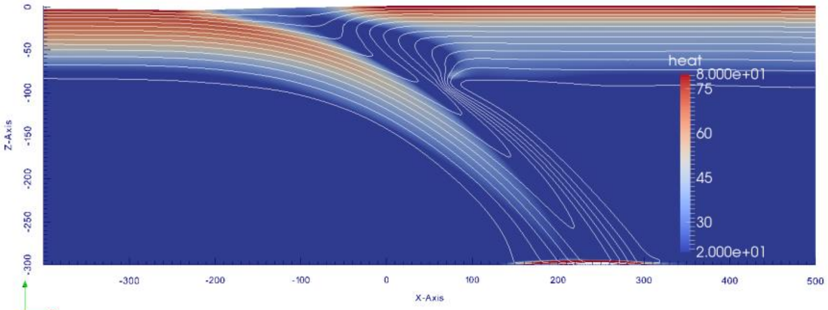
Model setup

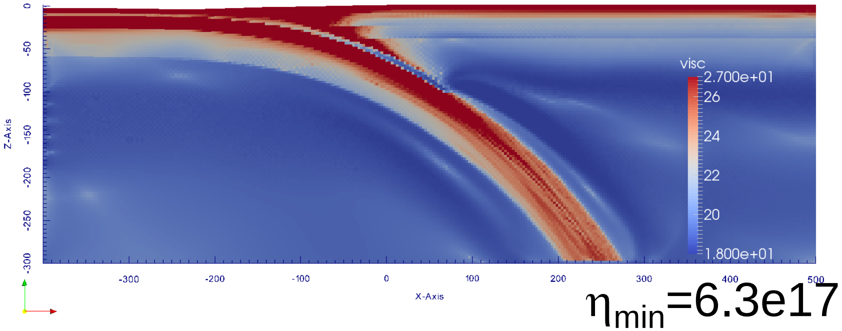
(long time scale)

Surface heat flow 70-80 MW/

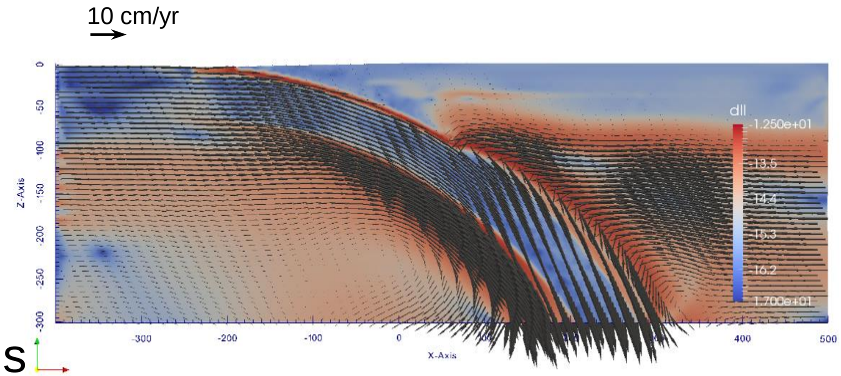




Temperature isolines 100-1300° C,



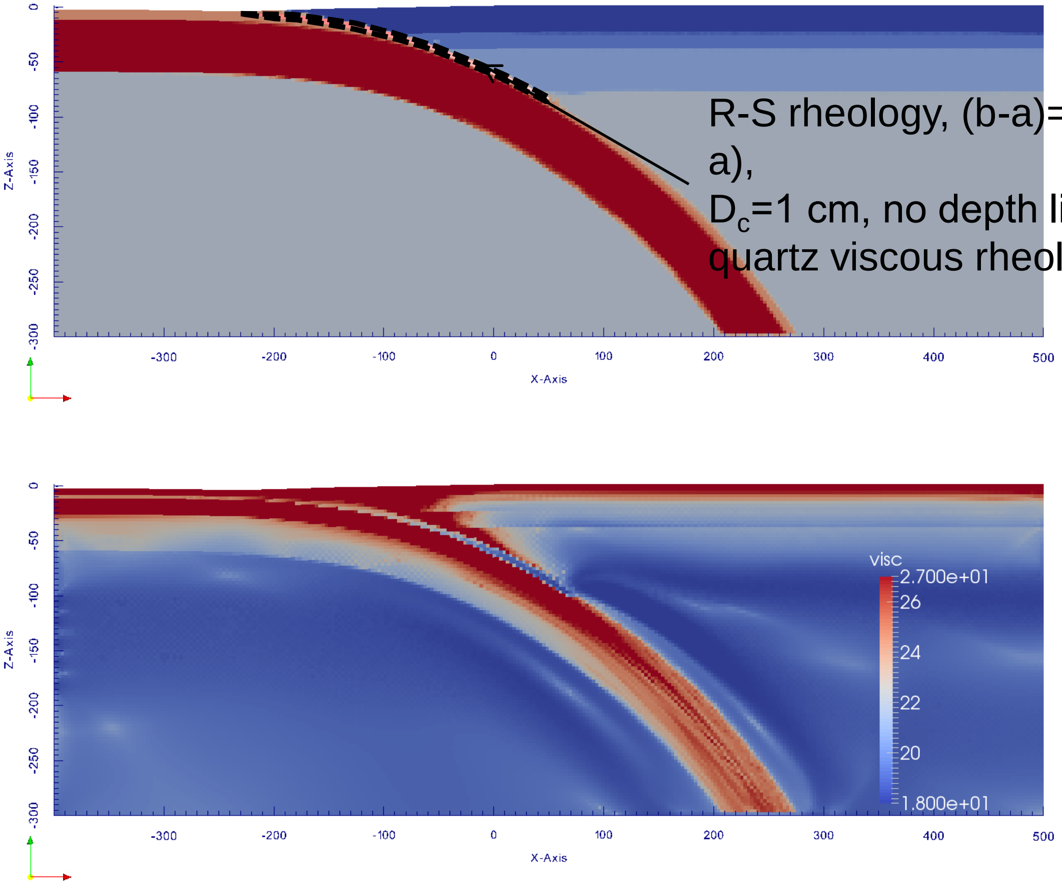
η_{min}=6.3e17 Pa.s



Before the rate-and-state rheology

Model setup

(short time scale)

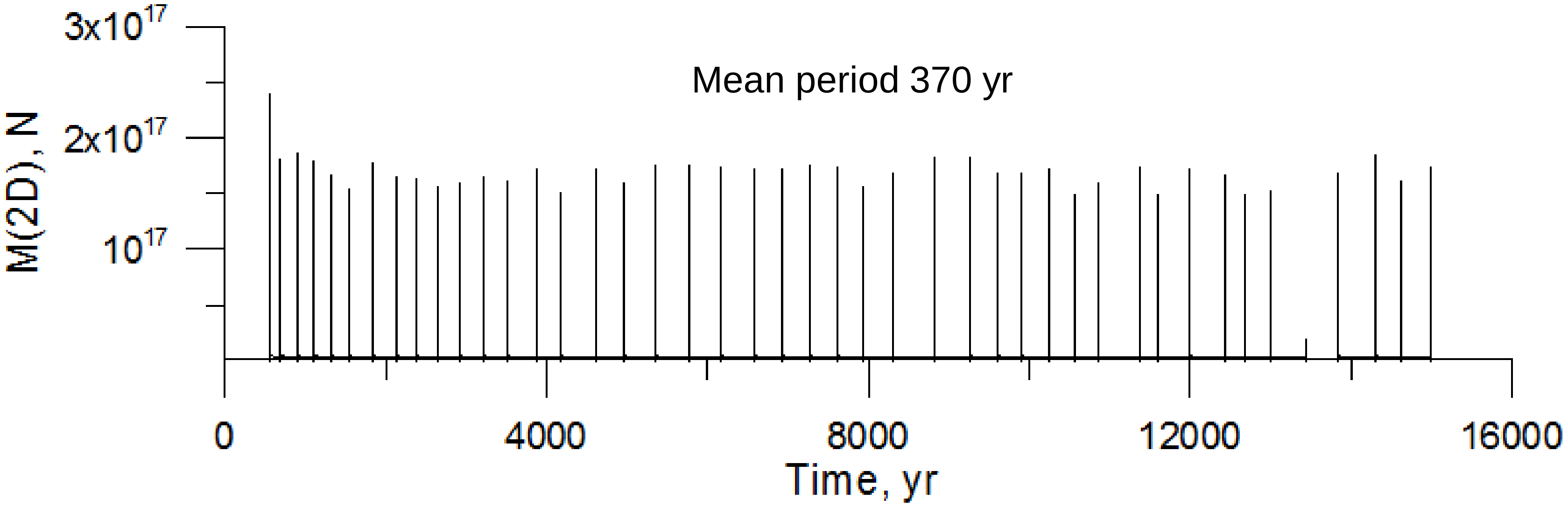


R-S rheology, $(b-a)=4.10^{-4}$, $b=2(b-a)$,

D_c=1 cm, no depth limit; weak „wet“ quartz viscous rheology

Earthquakes

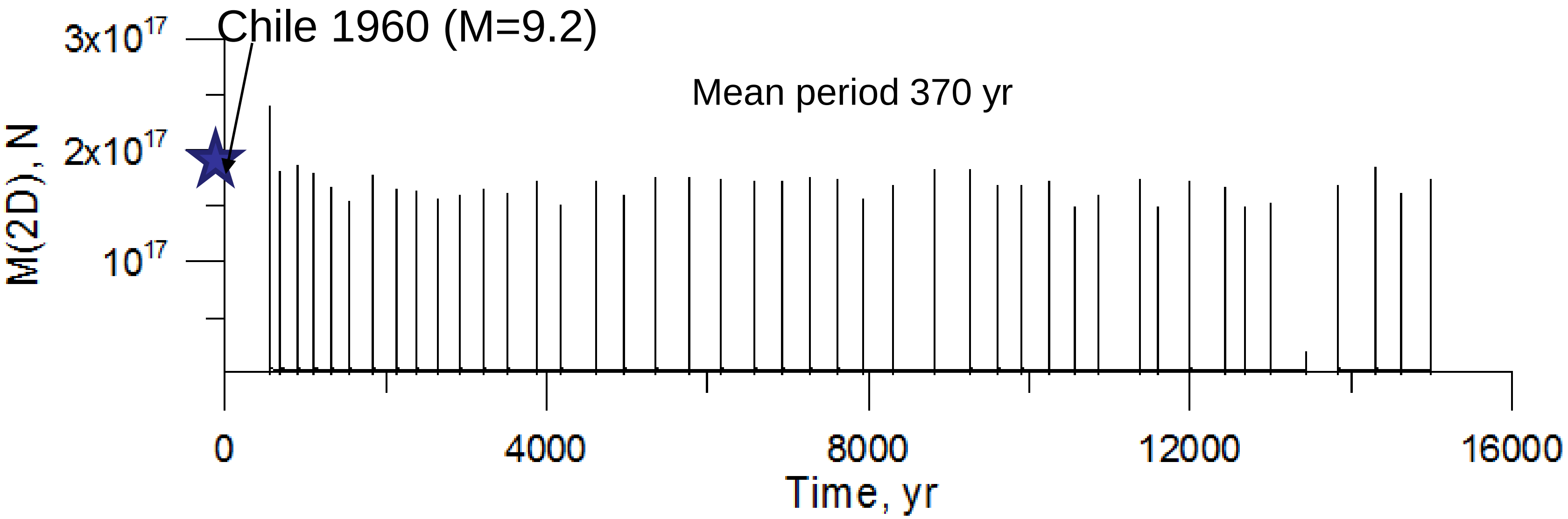
Adaptive time-step algorithm: from 5 yr step gradually multiplying by ½ to about 40 sec and back



Generated earthquakes sequence

Earthquakes

Adaptive time-step algorithm: from 5 yr step gradually multiplying by ½ to about 40 sec and back



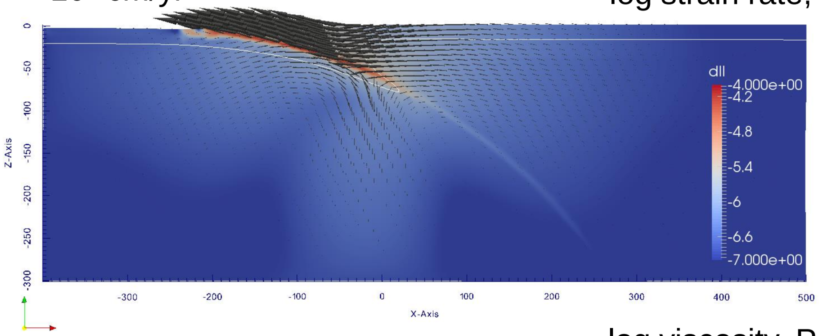
Generated earthquakes sequence

Zoom-in to earthquake

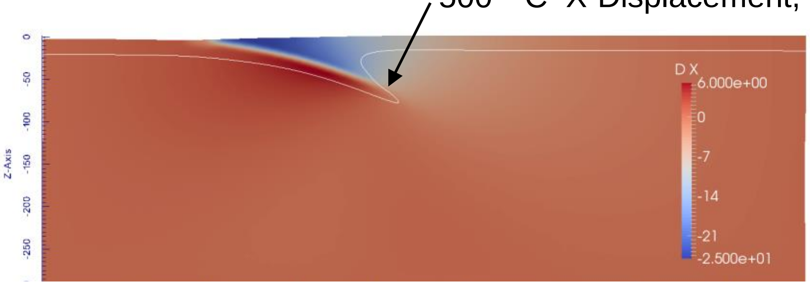
about 40 sec time-scale, M(2D)=1.8×10^{17}, mean slip at the fault 17 m, stress drop 6 MPa, rupture penetrates to about 500° C-isotherm depth

10^9 cm/yr

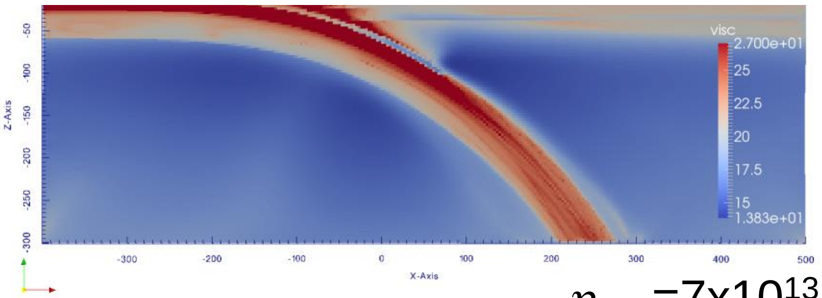
log strain rate, 1/s



500° C X-Displacement, m

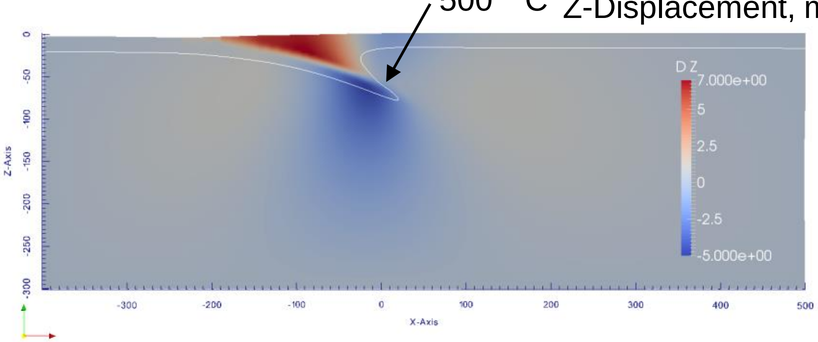


log viscosity, Pa.s



$\eta_{min}=7x10^{13}$ Pa

500° C Z-Displacement, m



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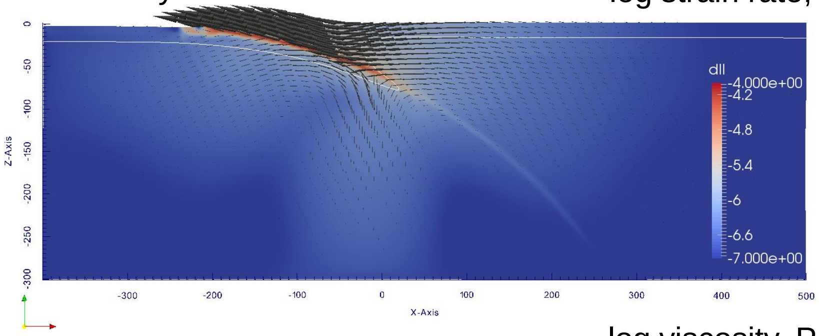
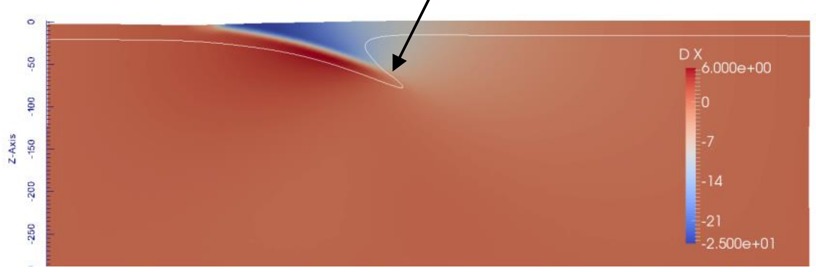
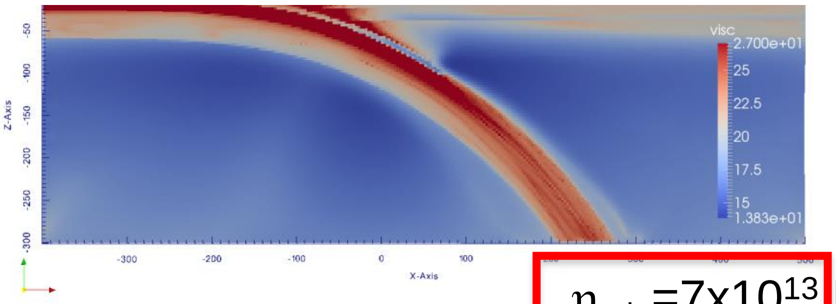


500° C X-Displacement, m

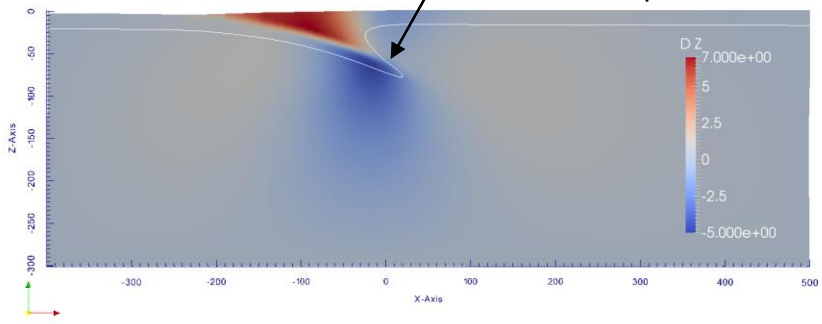


log viscosity, Pa.s



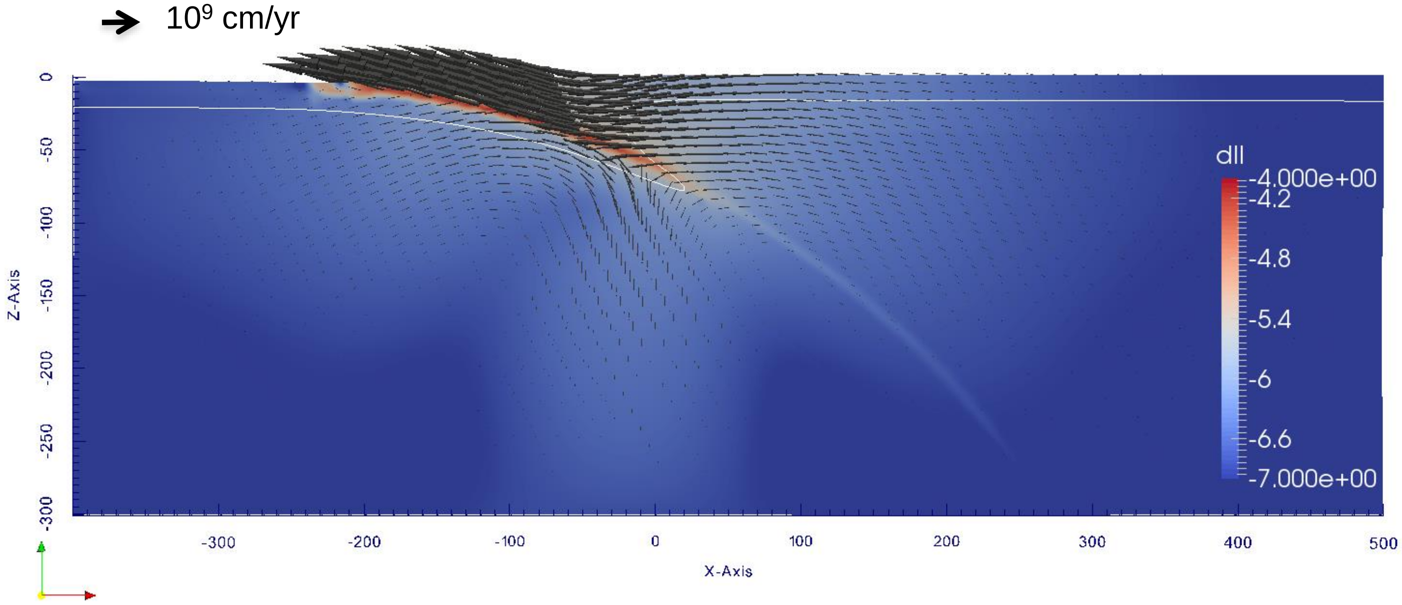
η_{min}=7×10^{13} Pa

500° C Z-Displacement, m



Seismic-cycle tour

40 sec

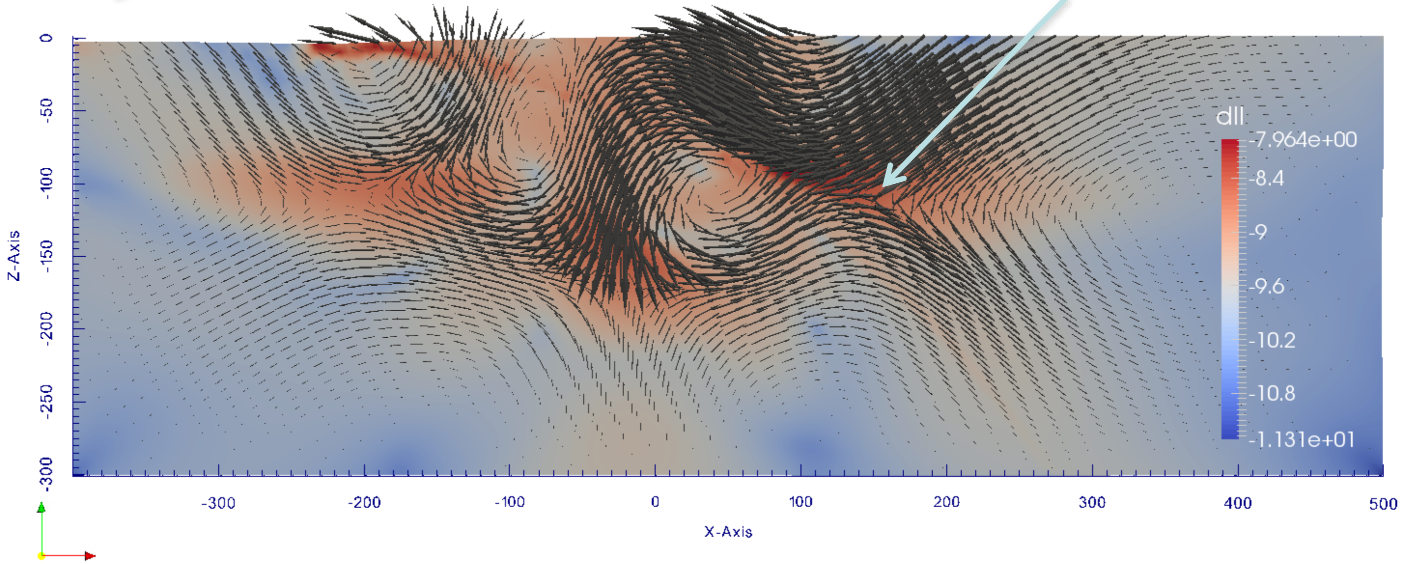


Seismic-cycle tour

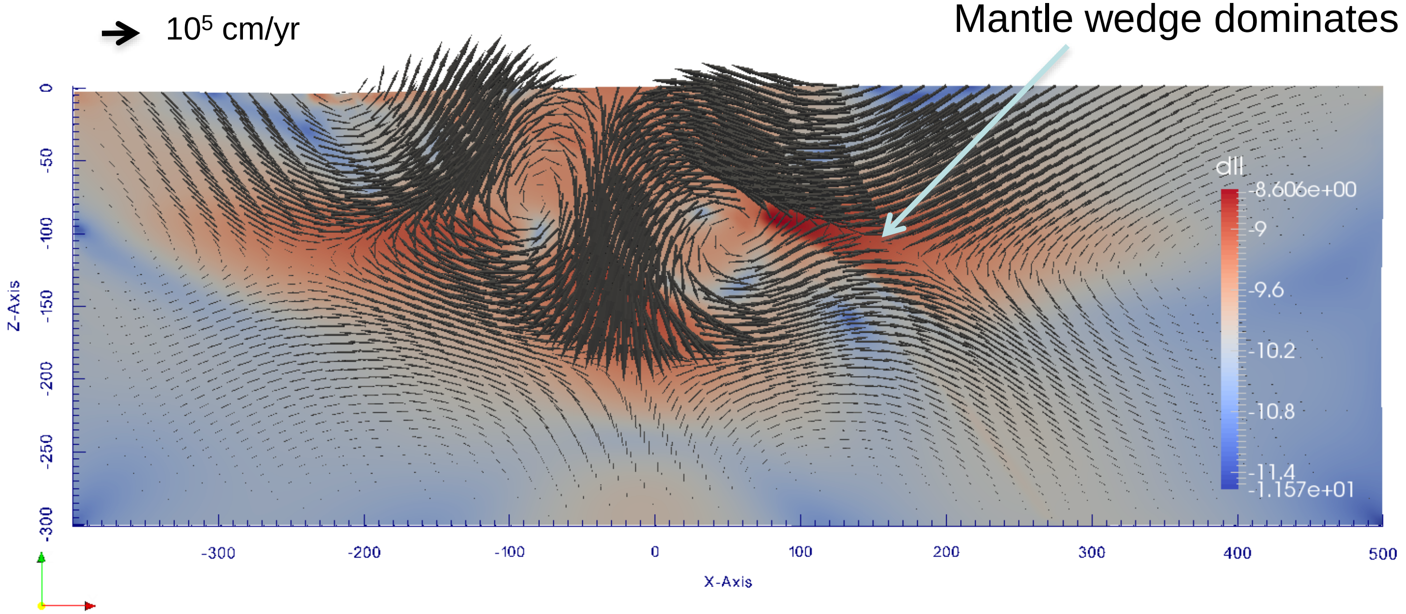
7 min

$\rightarrow$ 2x10^5 cm/yr

Mantle wedge is active



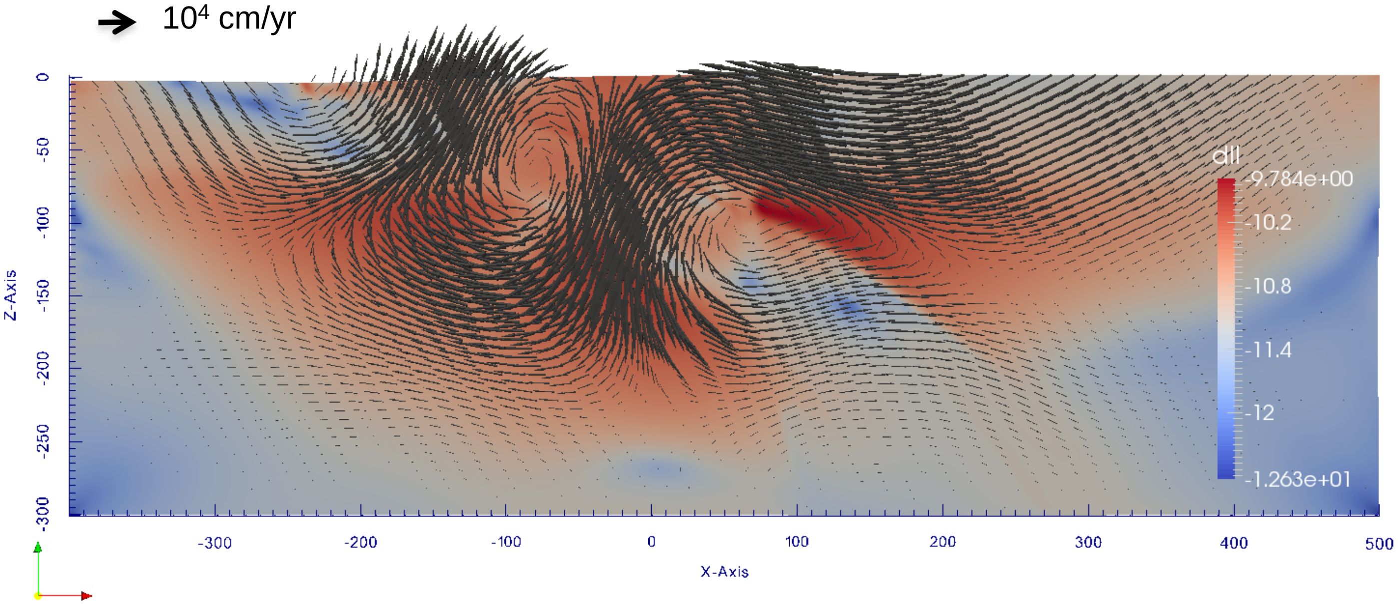
1 hour



10^5 cm/yr

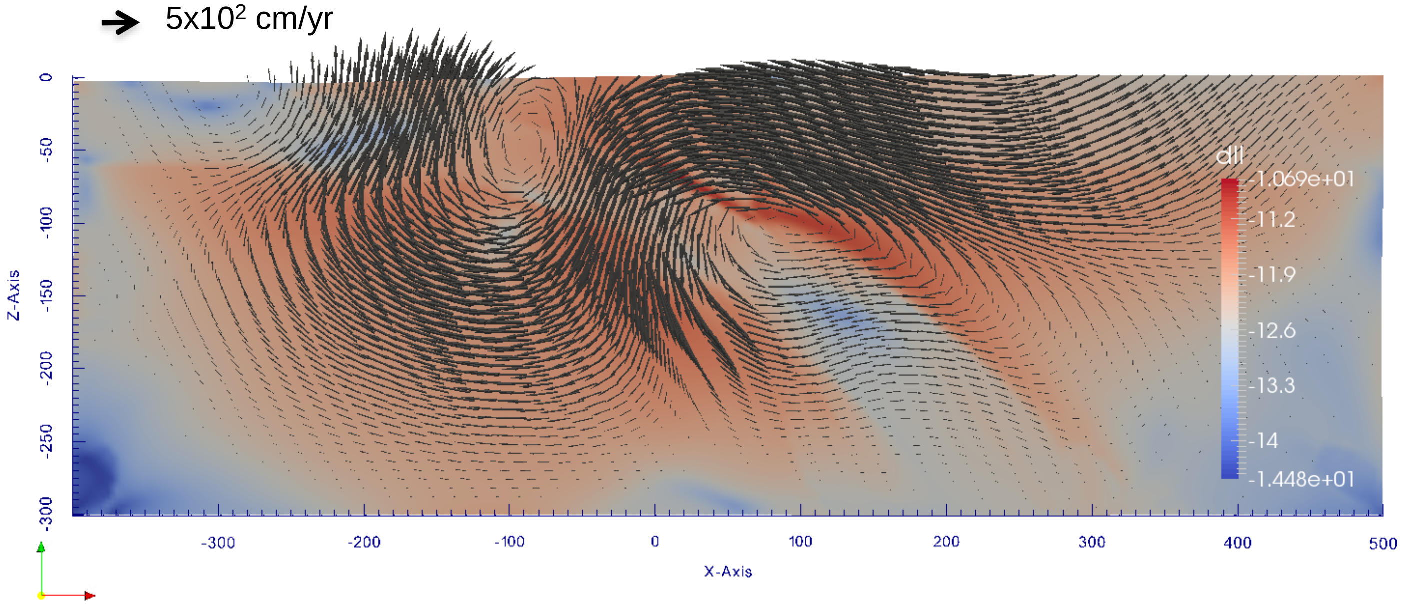
Mantle wedge dominates

1 day

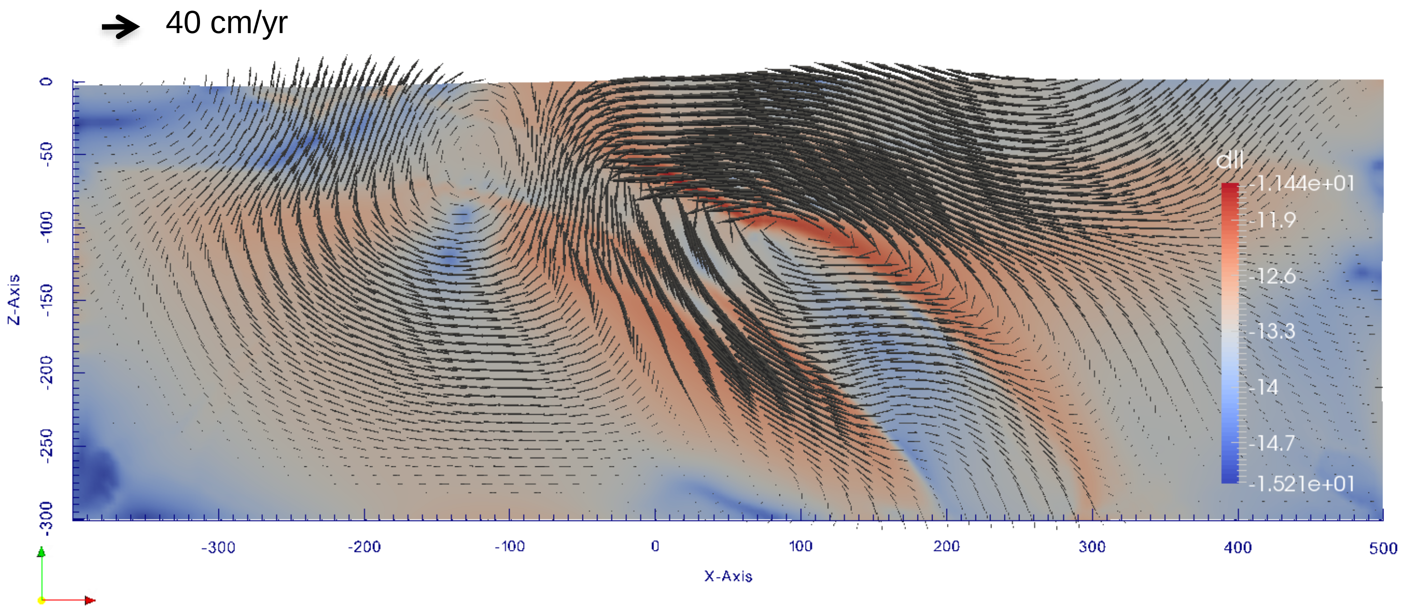


10^4 cm/yr
Z-Axis
0
-50
-100
-150
-200
-250
-300
-300
-200
-100
0
100
200
300
400
500
X-Axis
dll
-9.784e+00
-10.2
-10.8
-11.4
-12
-1.263e+01

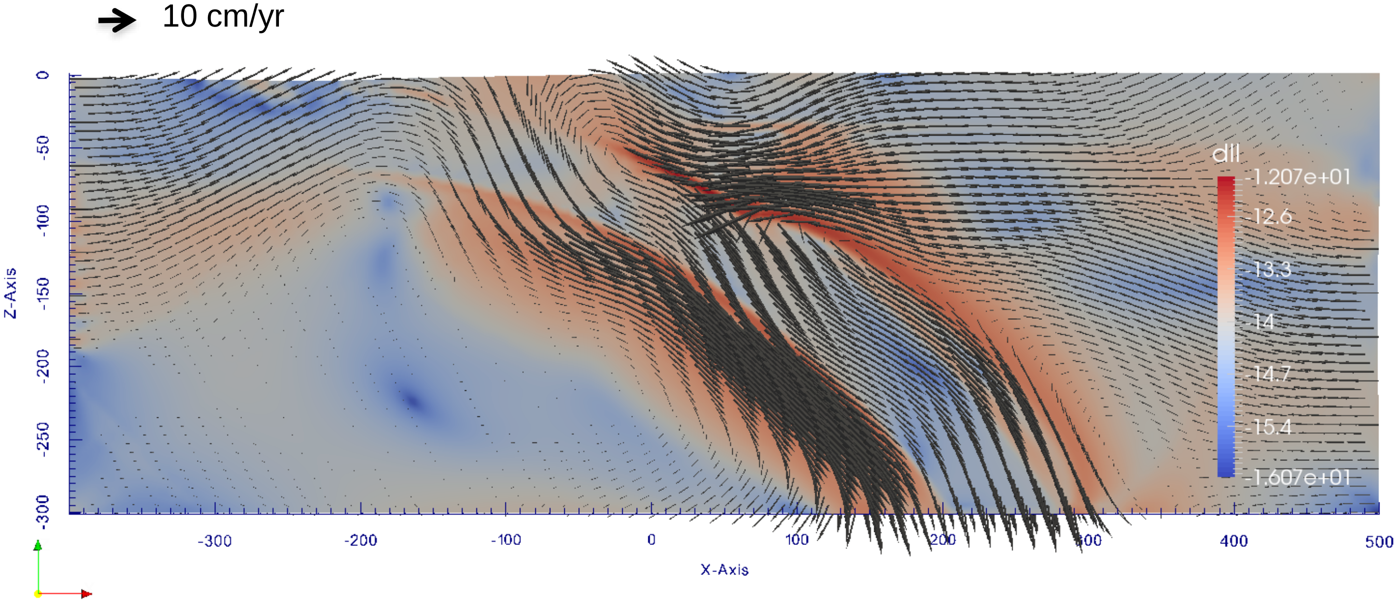
1 month


→ 5x10² cm/yr
dll
-1.069e+01
-11.2
-11.9
-12.6
-13.3
-14
-1.448e+01
Z-Axis
0
-50
-100
-150
-200
-250
-300
X-Axis
-300
-200
-100
0
100
200
300
400
500

1 year

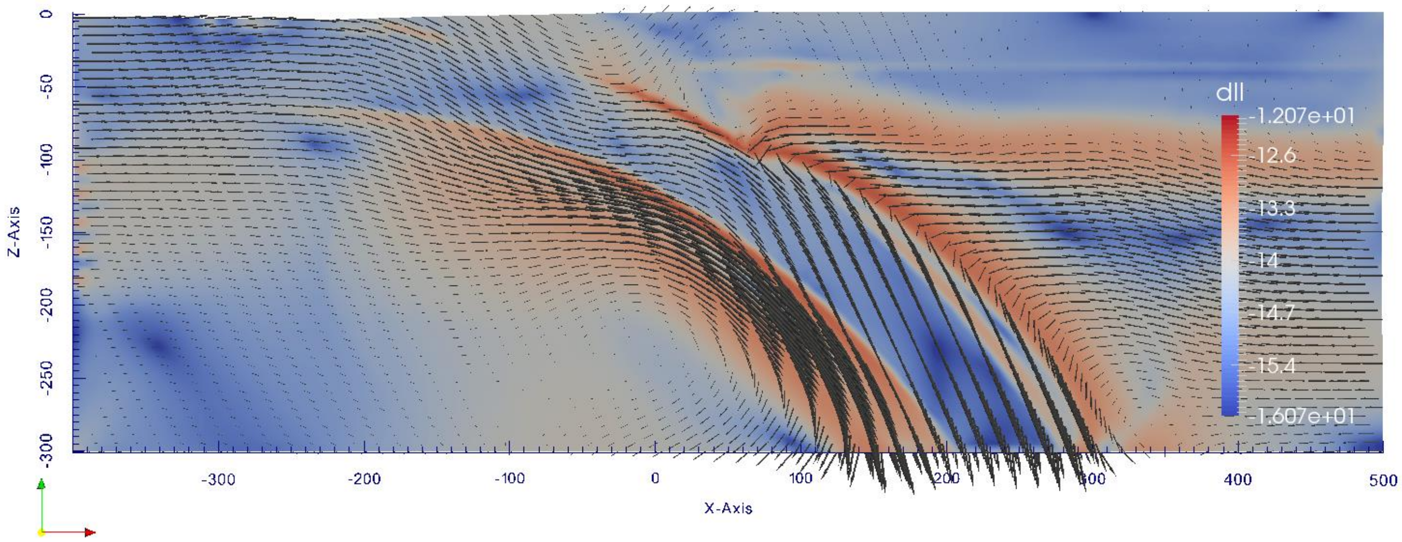

40 cm/yr
Z-Axis
0
-50
-100
-150
-200
-250
-300
X-Axis
-300
-200
-100
0
100
200
300
400
500
dll
-1.144e+01
-11.9
-12.6
-13.3
-14
-14.7
-1.521e+01

10 years


10 cm/yr
dll
-1.207e+01
-12.6
-13.3
-14
-14.7
-15.4
-1.607e+01
Z-Axis
0
-50
-100
-150
-200
-250
-300
X-Axis
-300
-200
-100
0
100
200
300
400
500

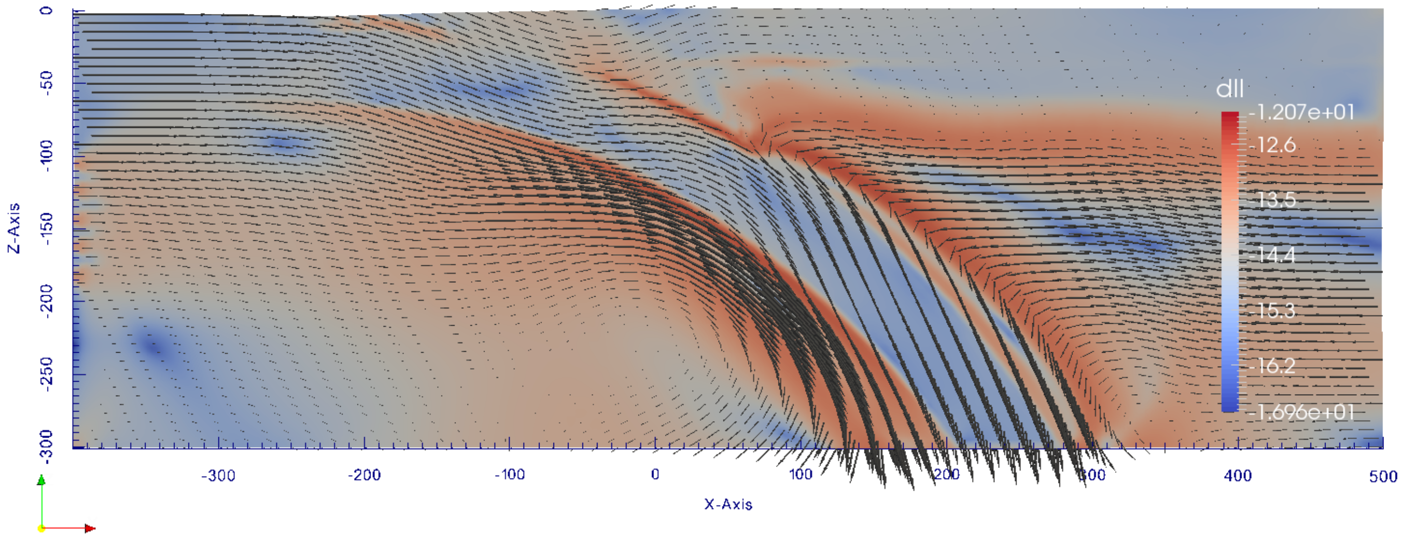
50 years

→ 10 cm/yr

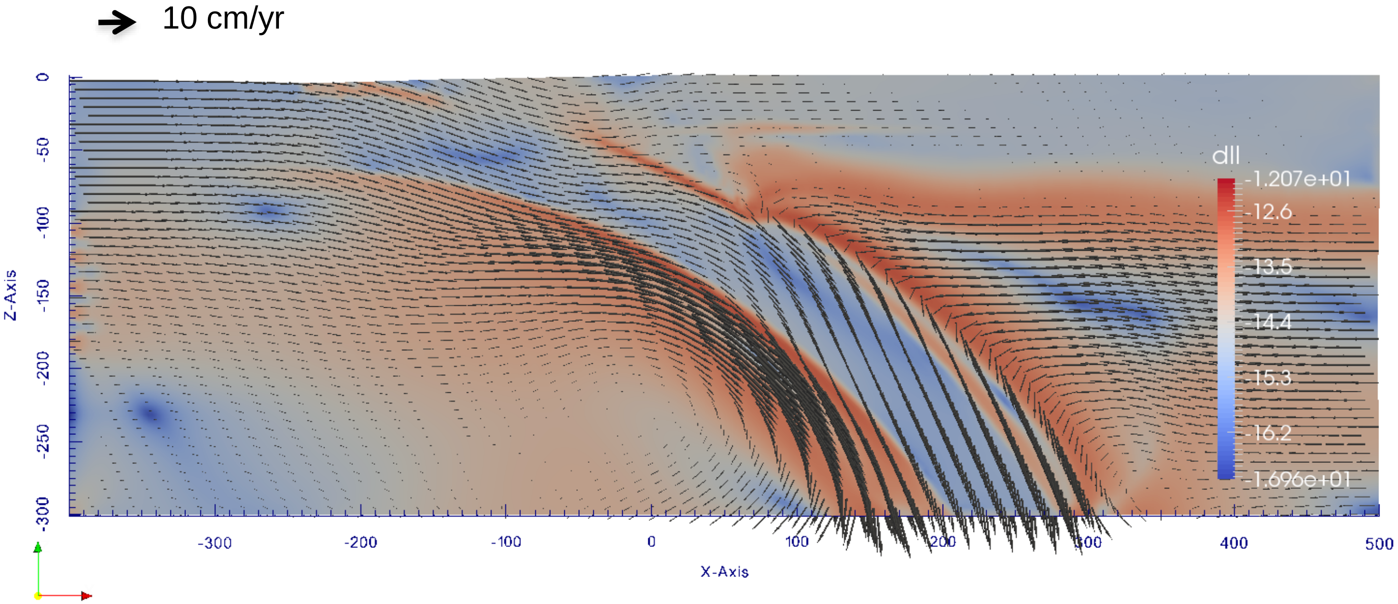


100 years

➔ 10 cm/yr

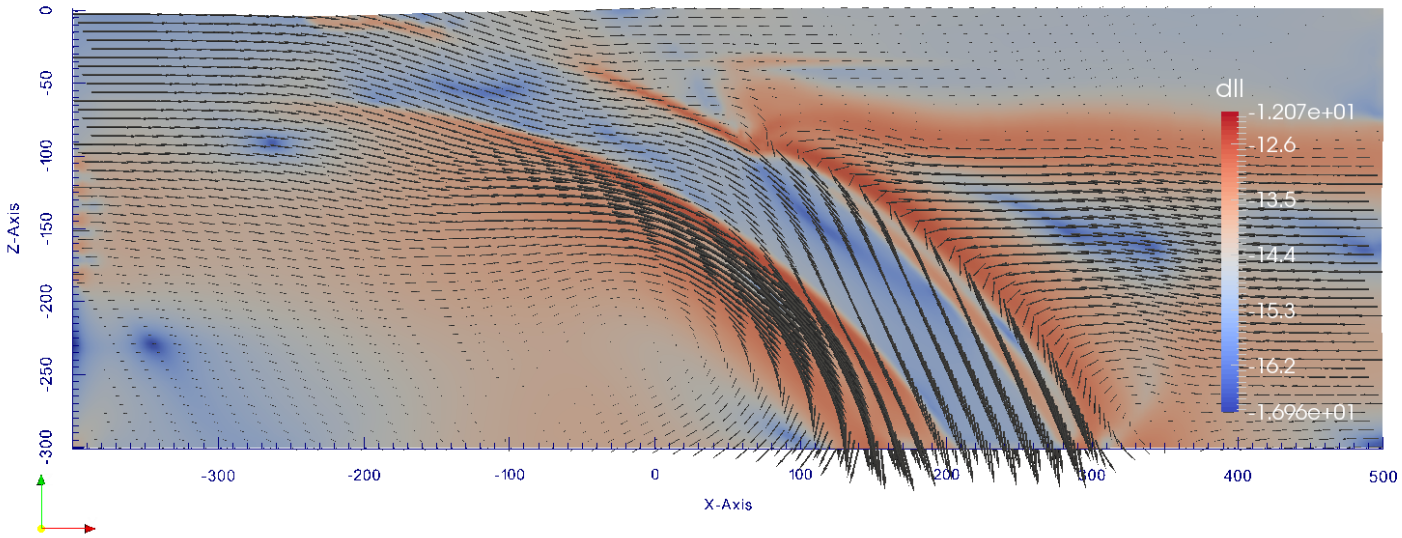


150 years

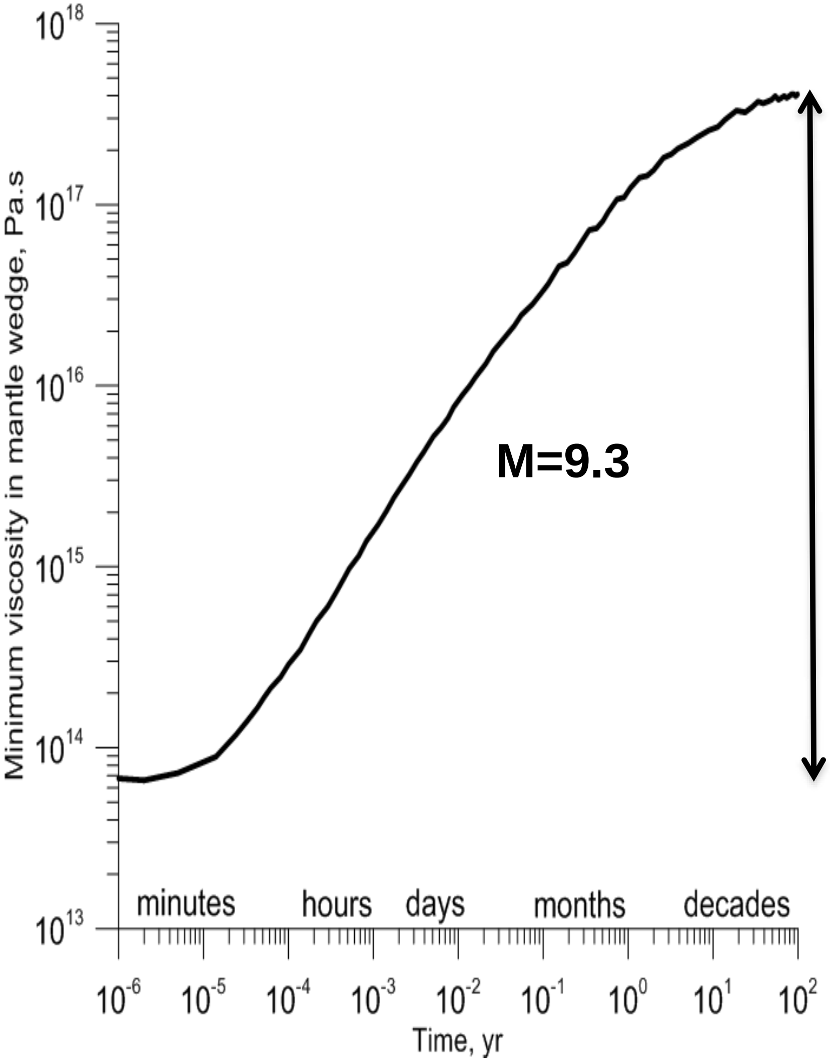

10 cm/yr
dll
-1.207e+01
-12.6
-13.5
-14.4
-15.3
-16.2
-1.696e+01
Z-Axis
0
-50
-100
-150
-200
-250
-300
X-Axis
-300
-200
-100
0
100
200
300
400
500

200 years

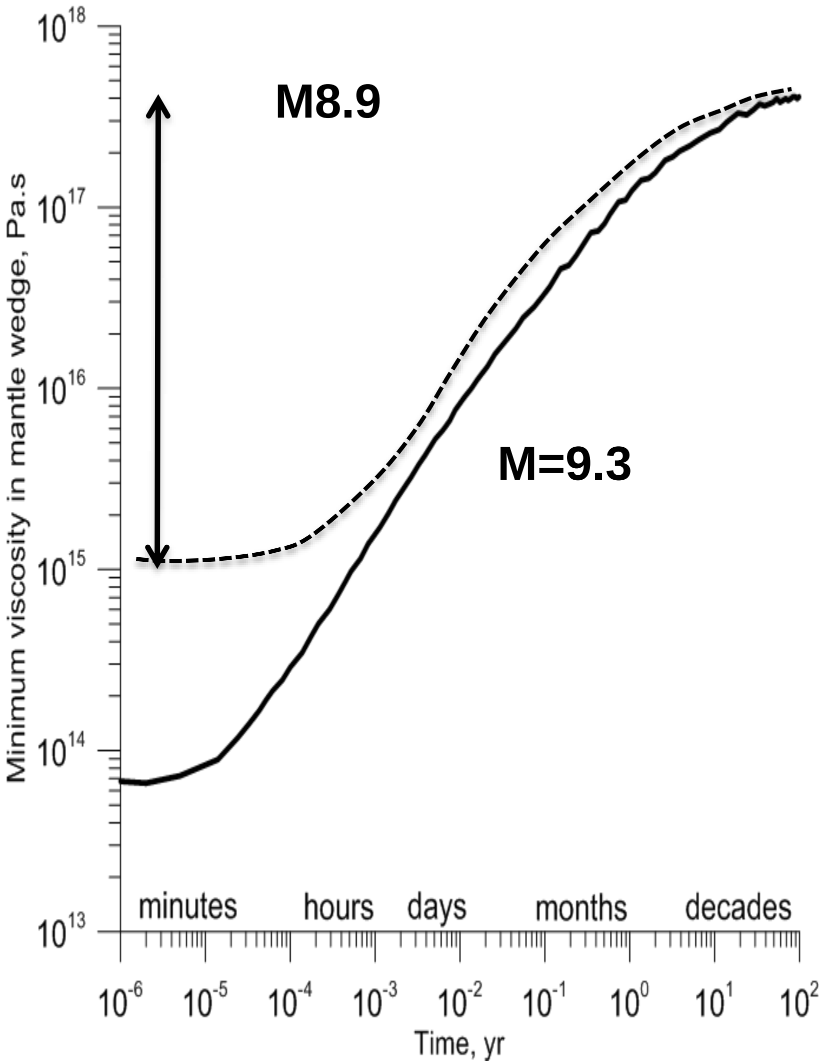
→ 10 cm/yr



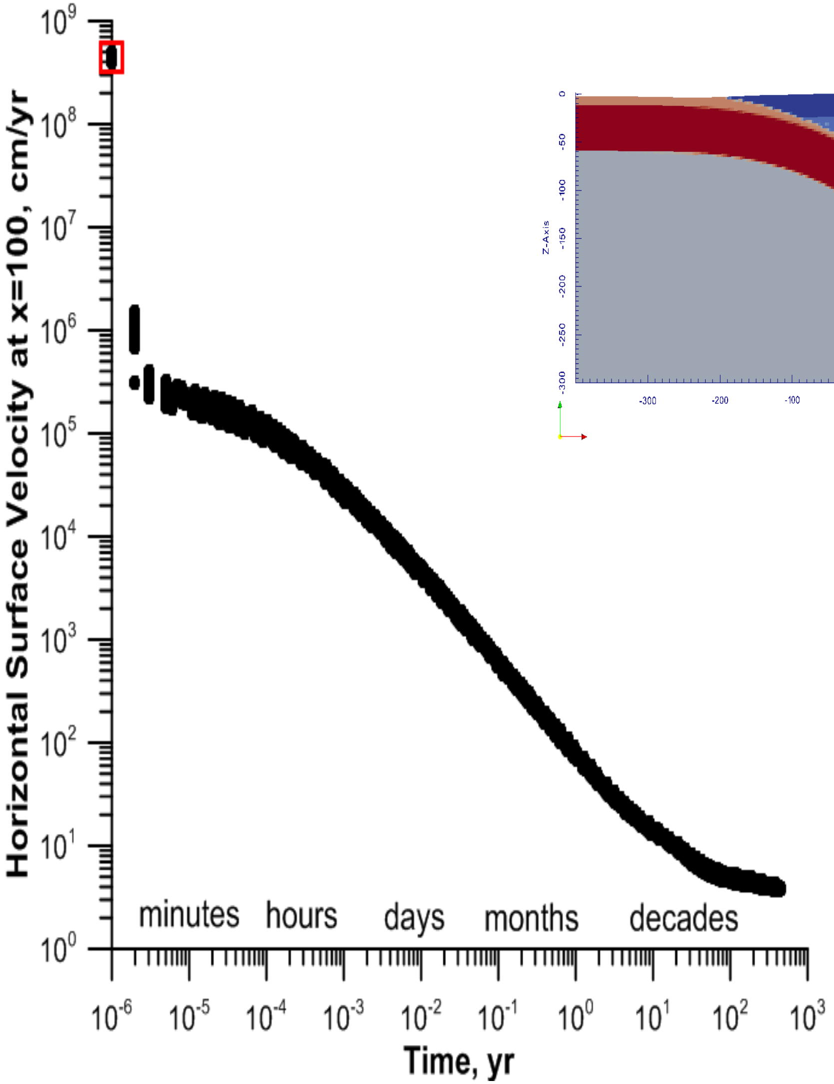
Evolution of viscosity in mantle w

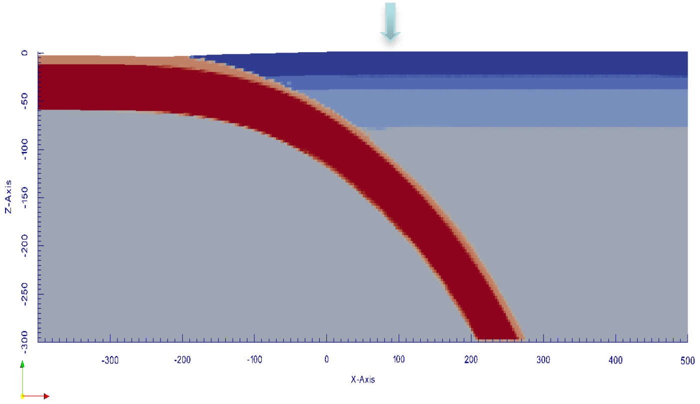


Evolution of viscosity in mantle w

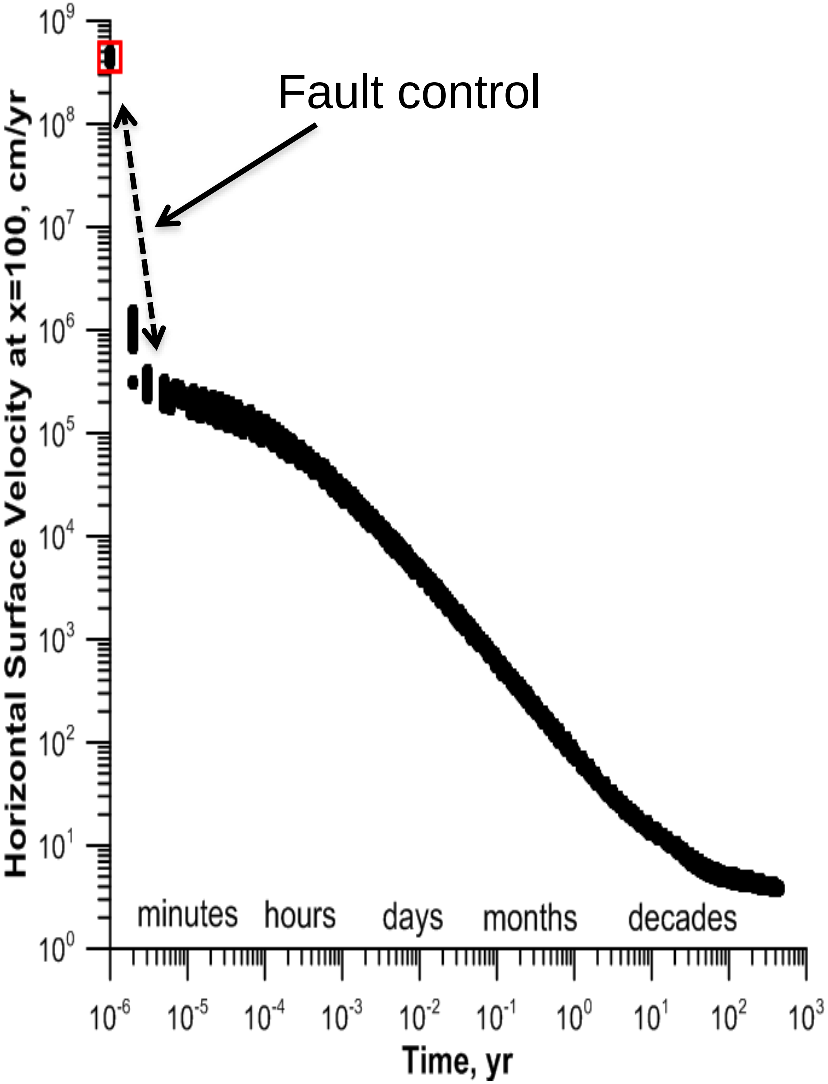


Surface X-velocity vs time

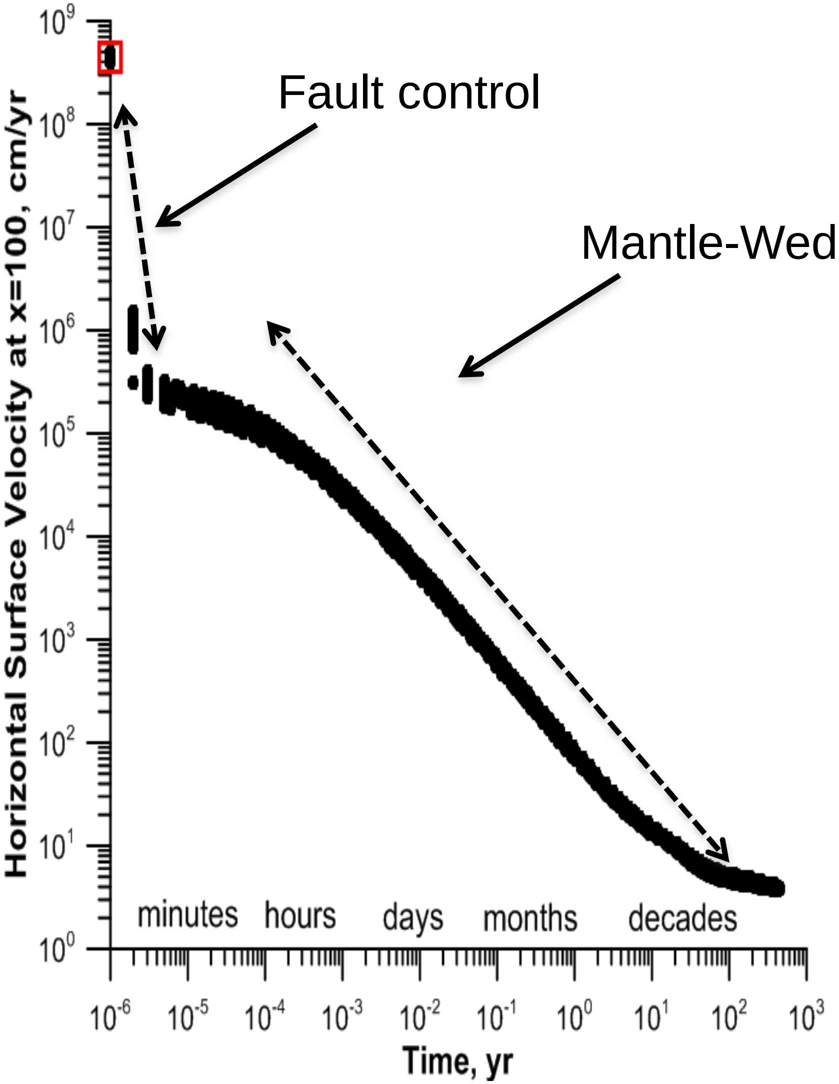




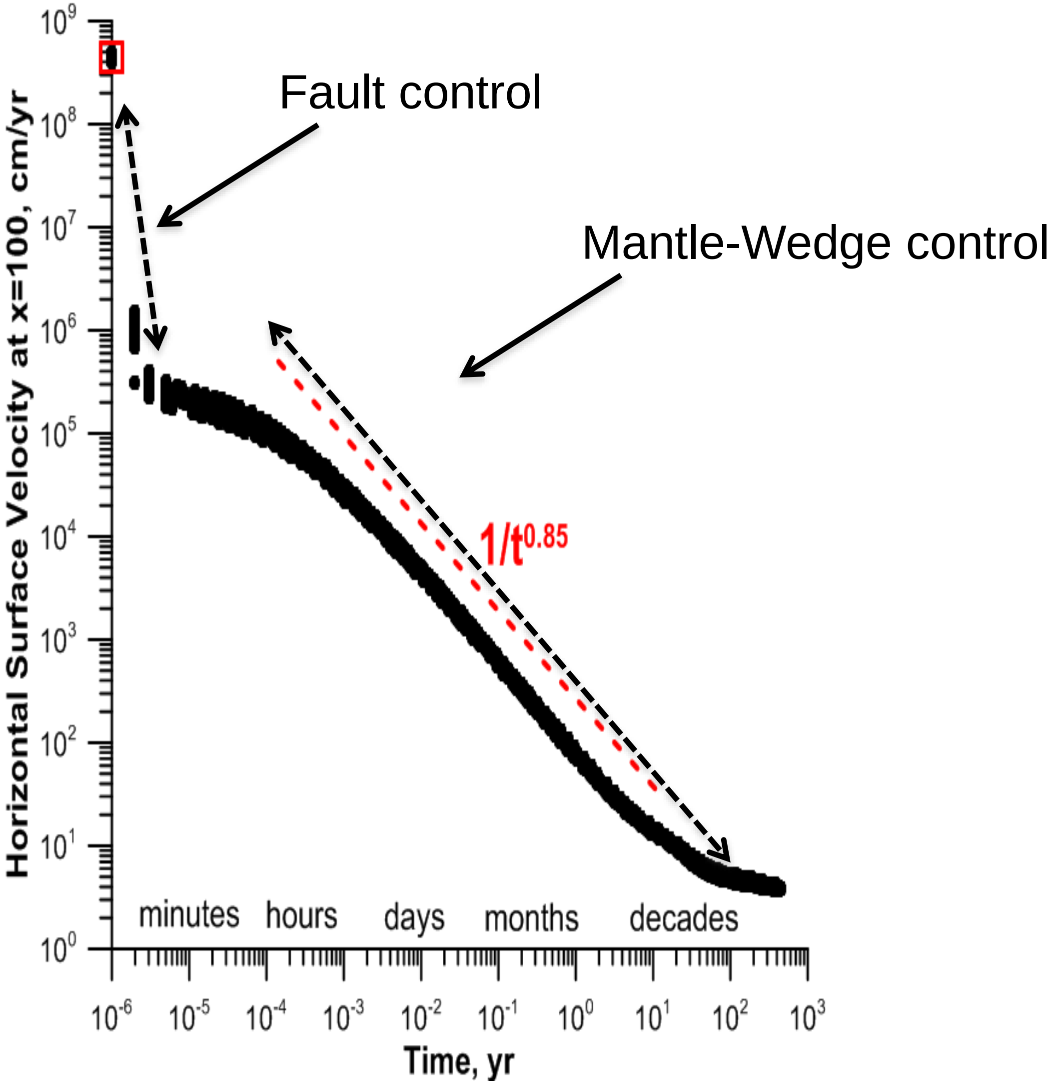
Visco-elastic relaxation



Visco-elastic relaxation



Visco-elastic relaxation

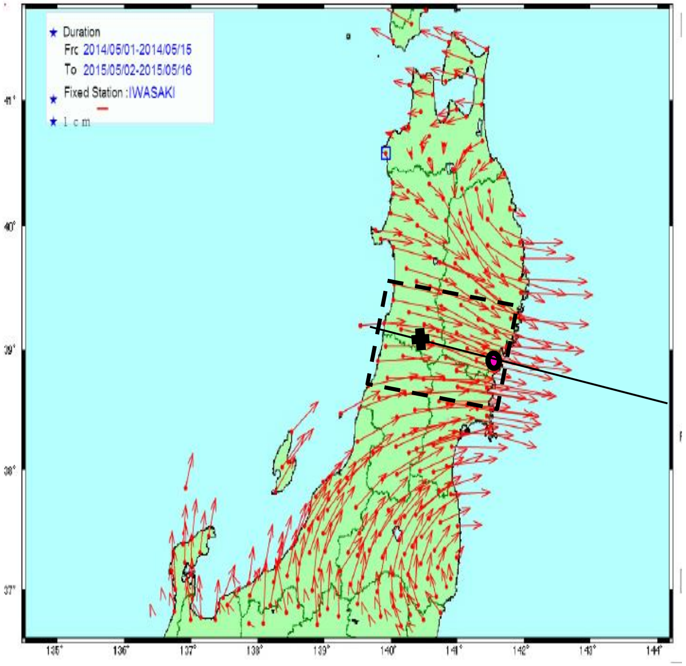


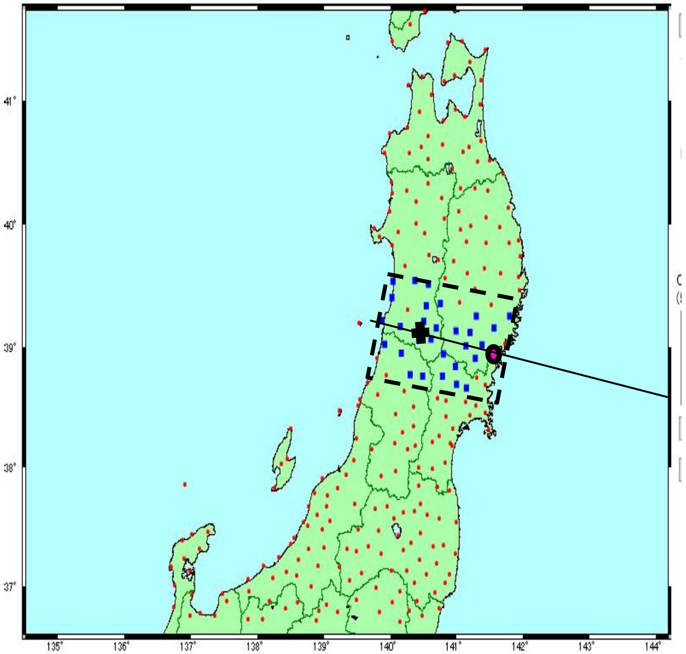
Fault control

Mantle-Wedge control

Model verification

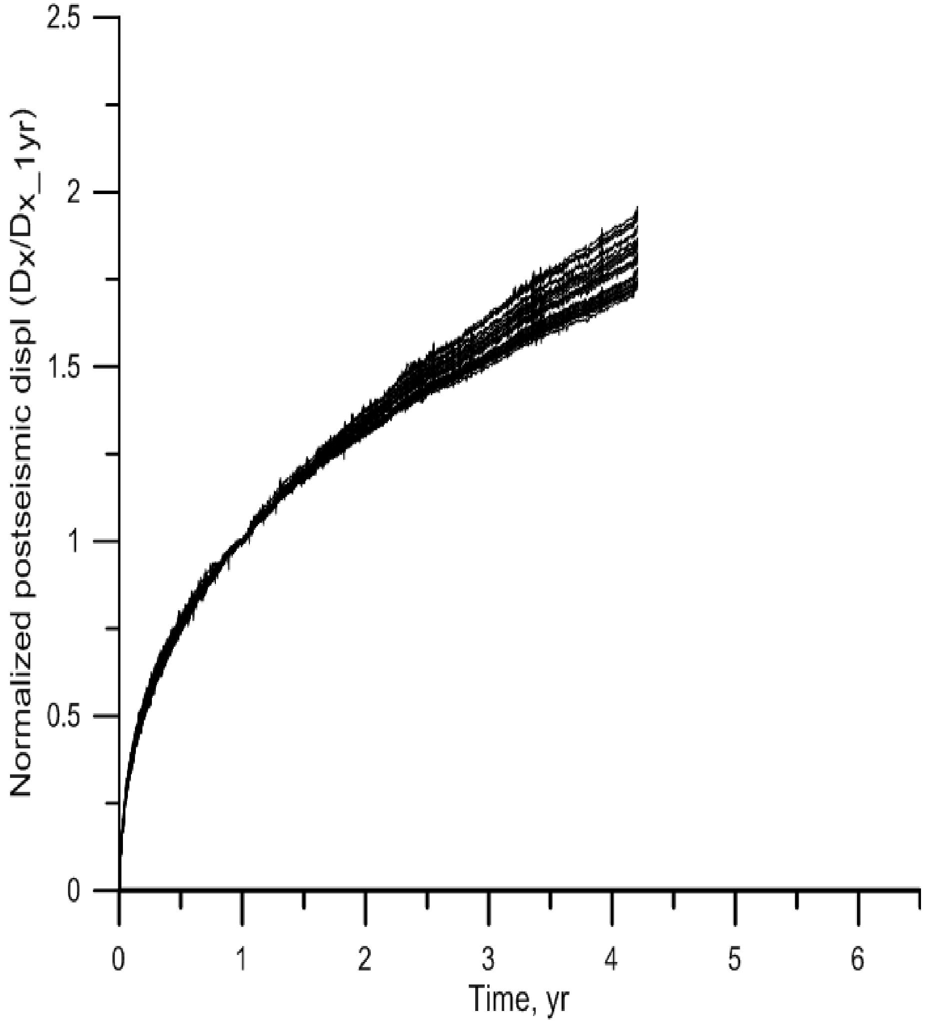
Comparison with GPS observations for Tohoku 2011 earthquake



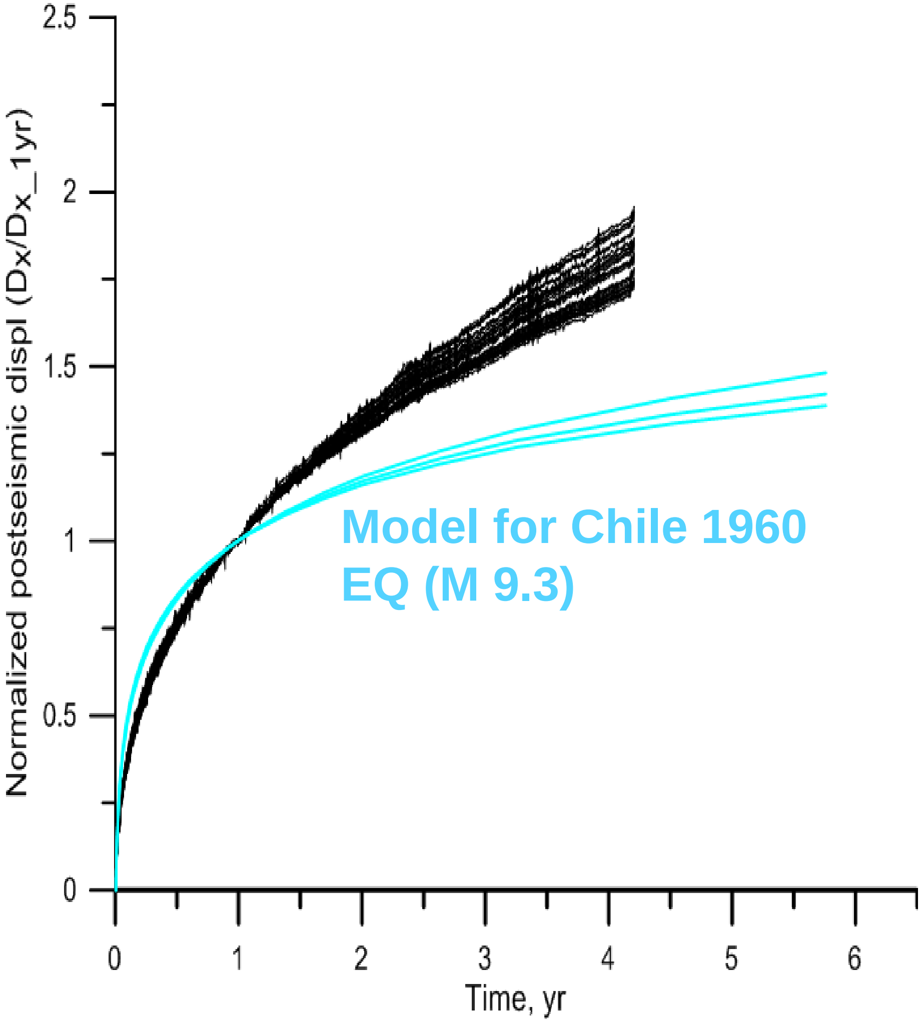


From GPS coordinates for each station we calculate EW displacement relative to the **2nd day** after the earthquake, and then normalize it by 1 year displacement

Application for Tohoku 2011

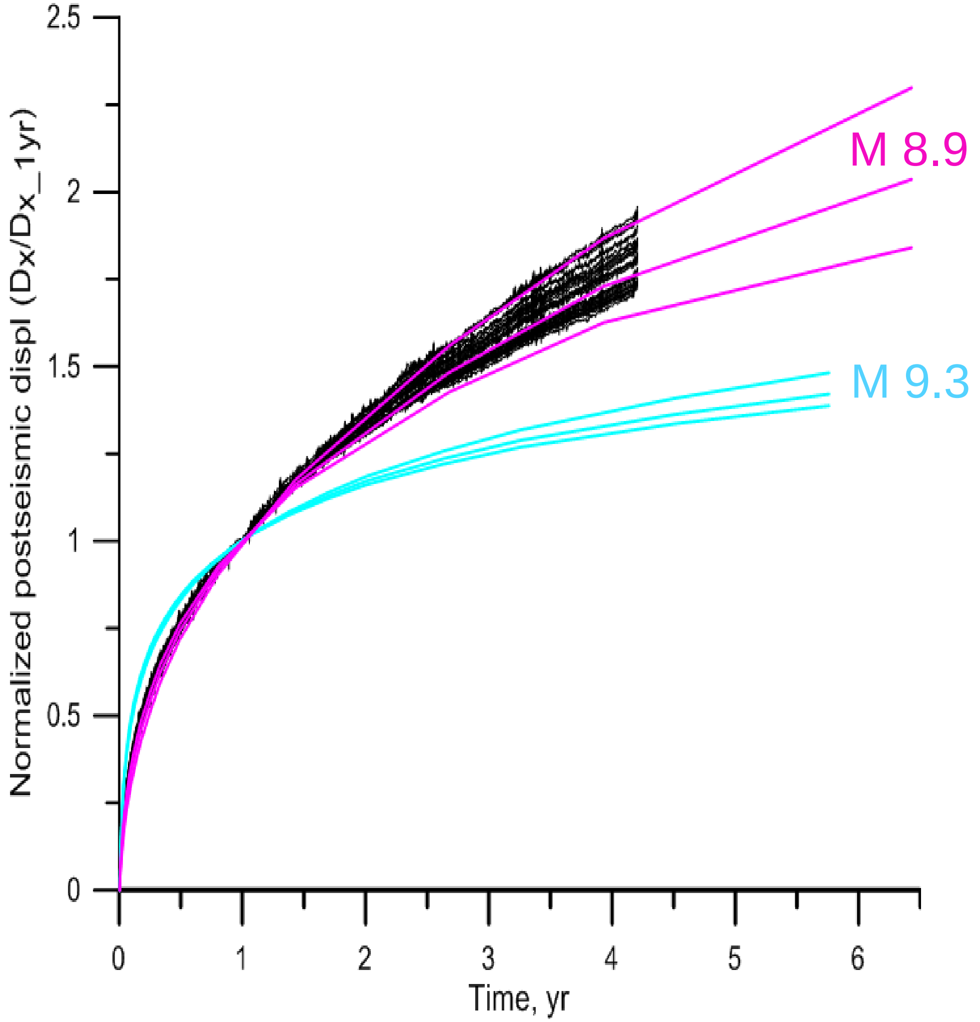


Application for Tohoku 2011

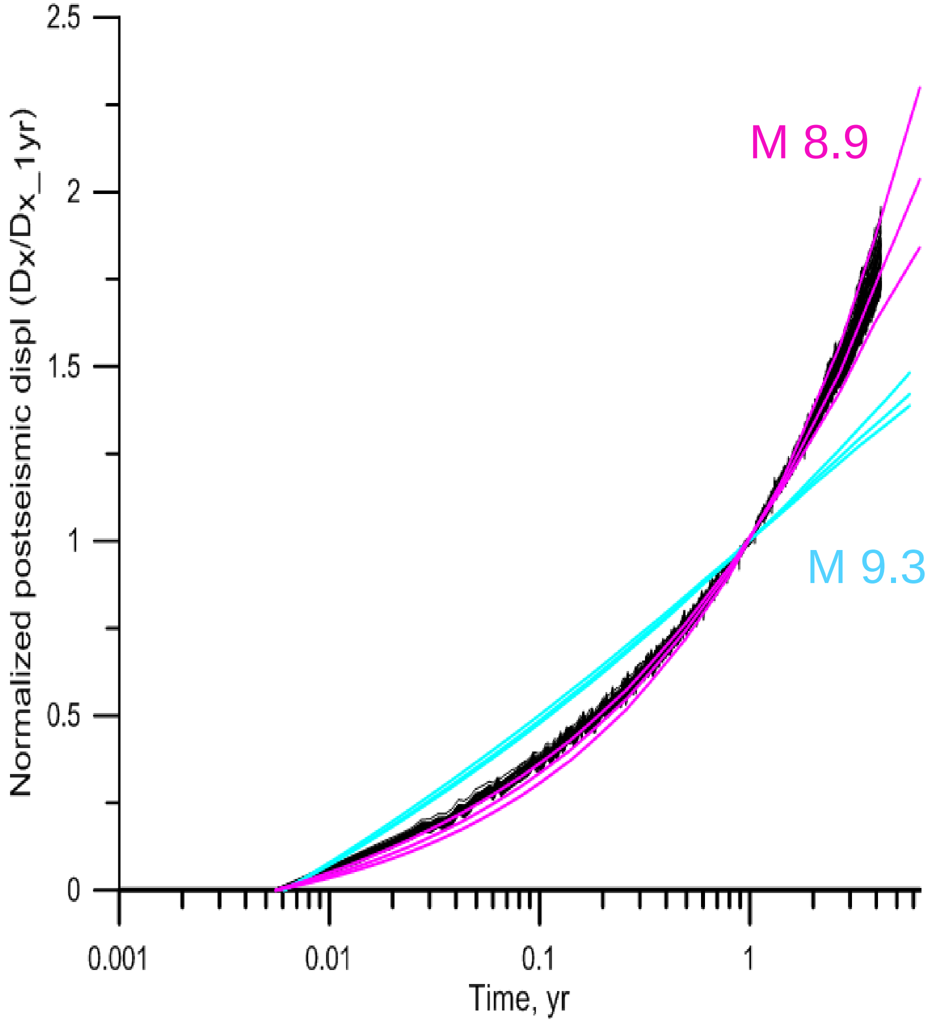


Model for Chile 1960 EQ (M 9.3)

Application for Tohoku 2011

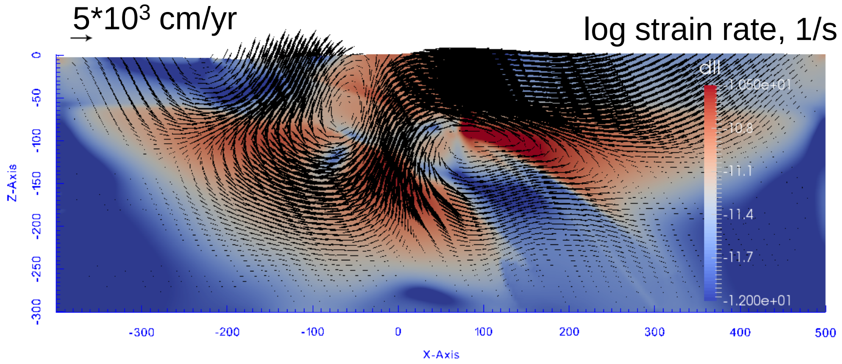


Application for Tohoku 2011

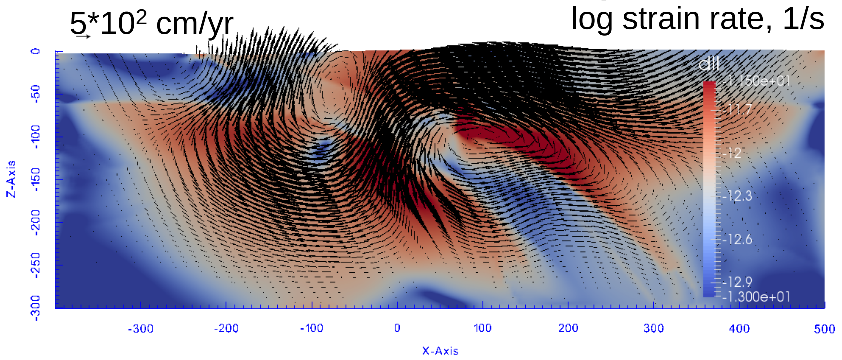


Postseismic relaxation

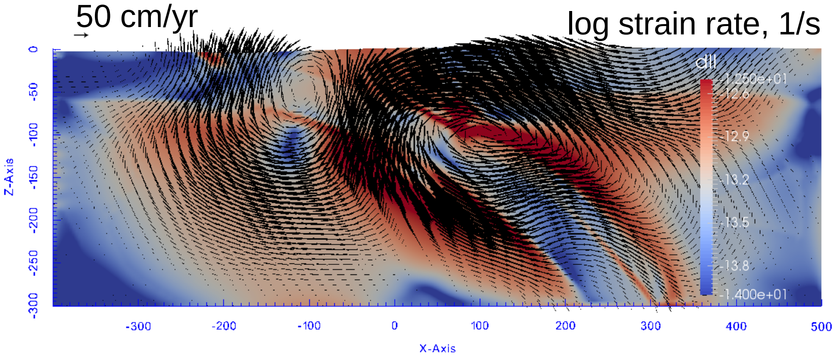
4 days after eq.



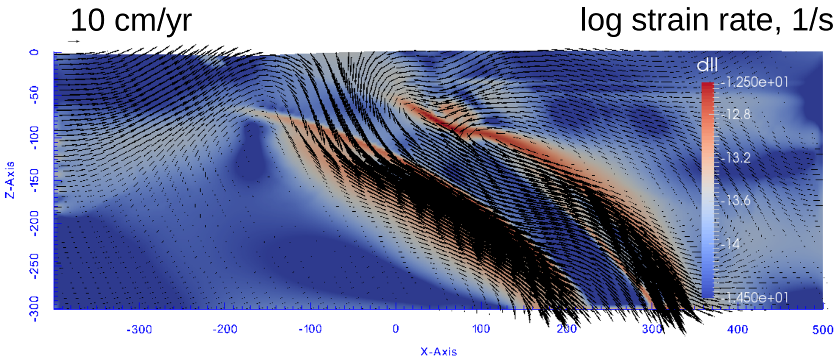
1 month after eq.



1 year after eq.



10 years after eq.

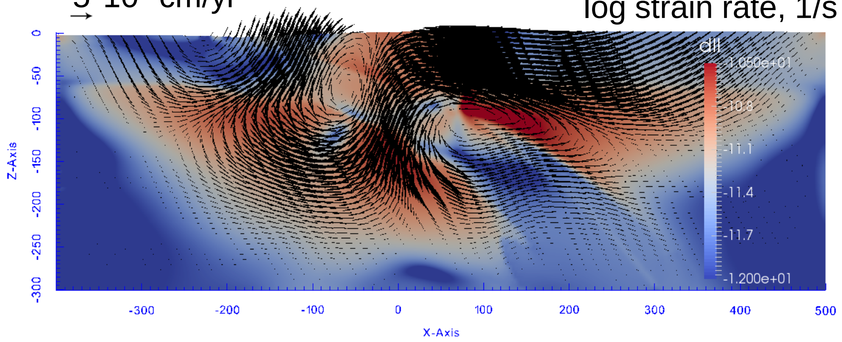


Postseismic relaxation

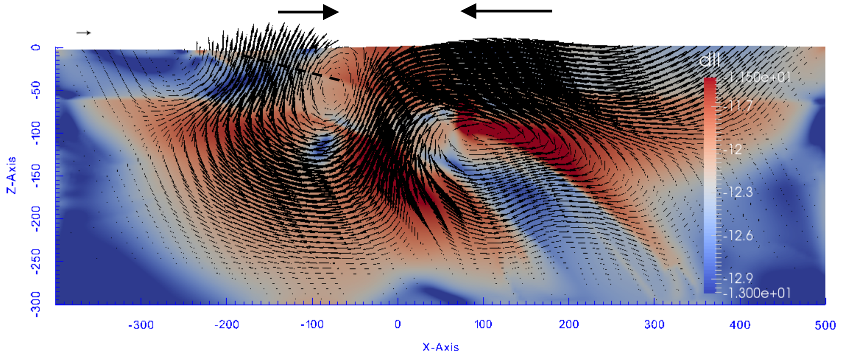
4 days after eq.

$5*10^3$ cm/yr

log strain rate, 1/s



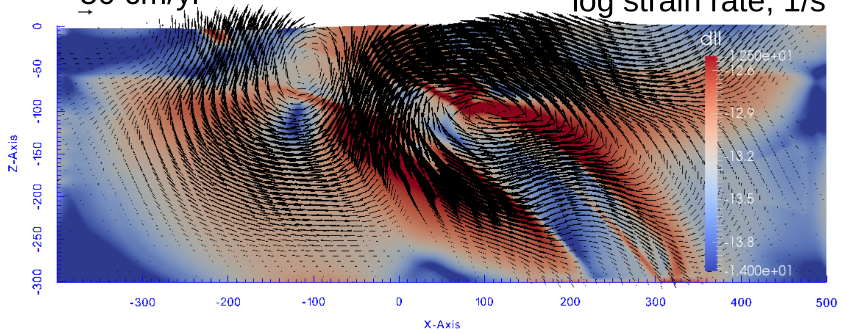
1 month after eq.



1 year after eq.

50 cm/yr

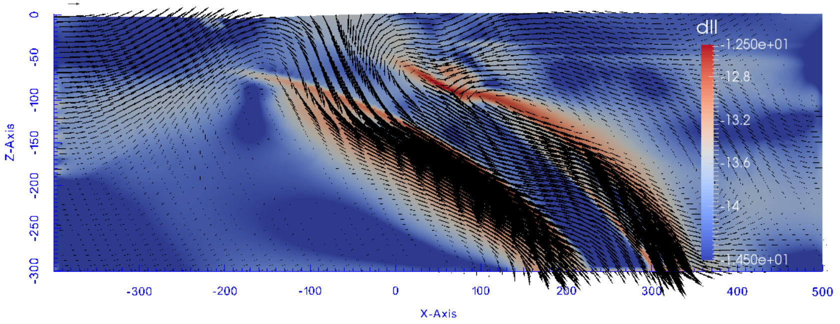
log strain rate, 1/s



10 years after eq.

10 cm/yr

log strain rate, 1/s

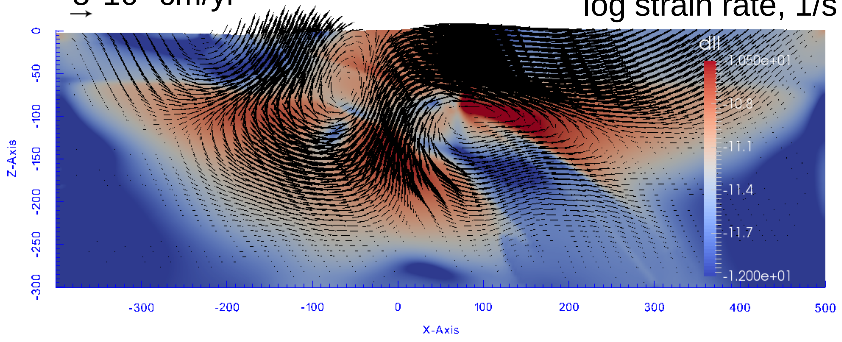


Postseismic relaxation

4 days after eq.

5*10^3 cm/yr

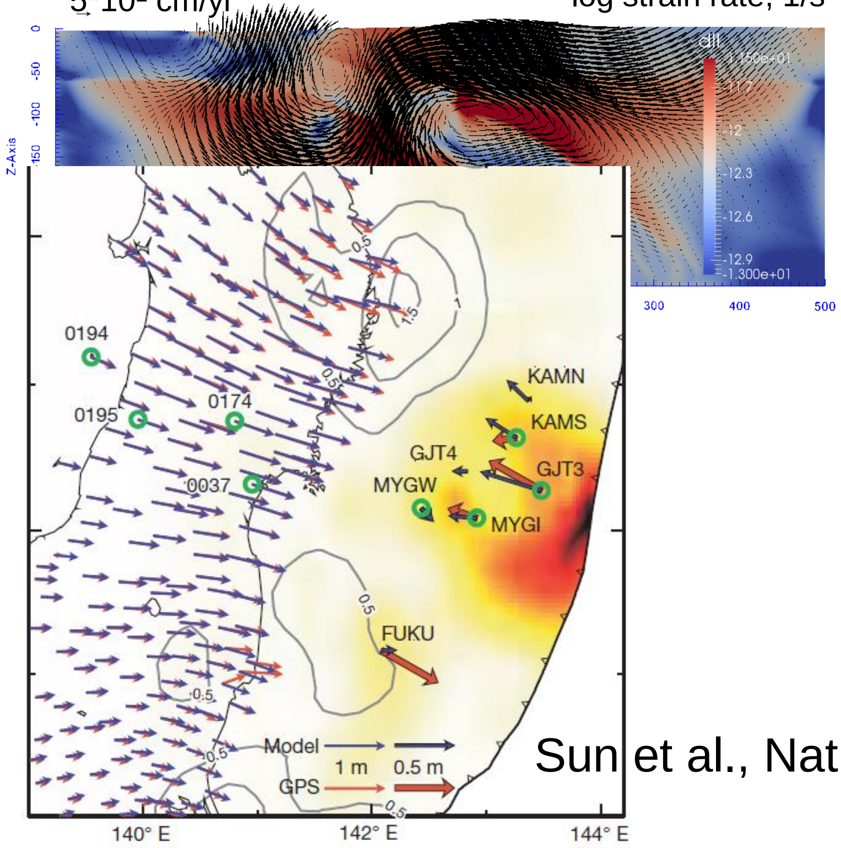
log strain rate, 1/s



1 month after eq.

5*10^2 cm/yr

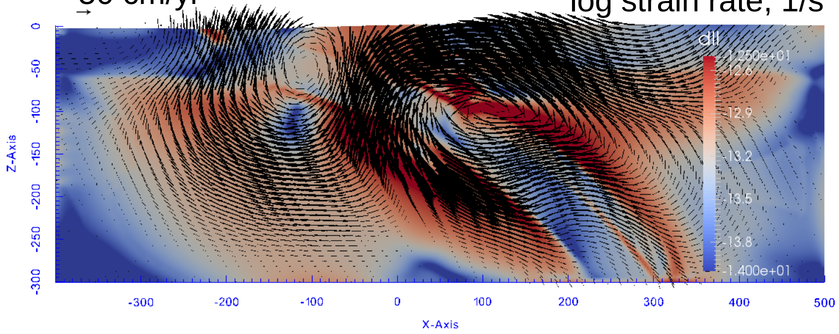
log strain rate, 1/s



1 year after eq.

50 cm/yr

log strain rate, 1/s



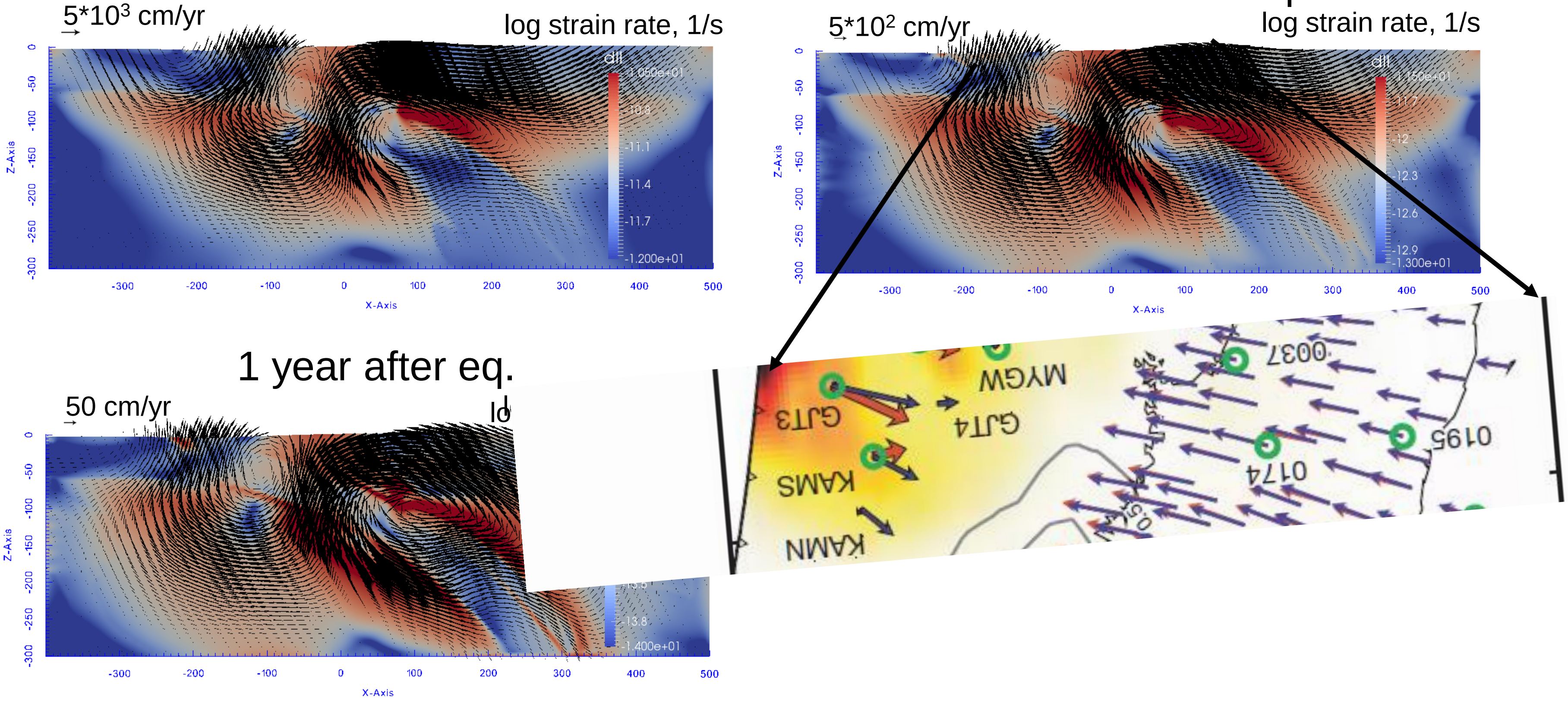
Sun et al., Nature,

Postseismic relaxation

4 days after eq.

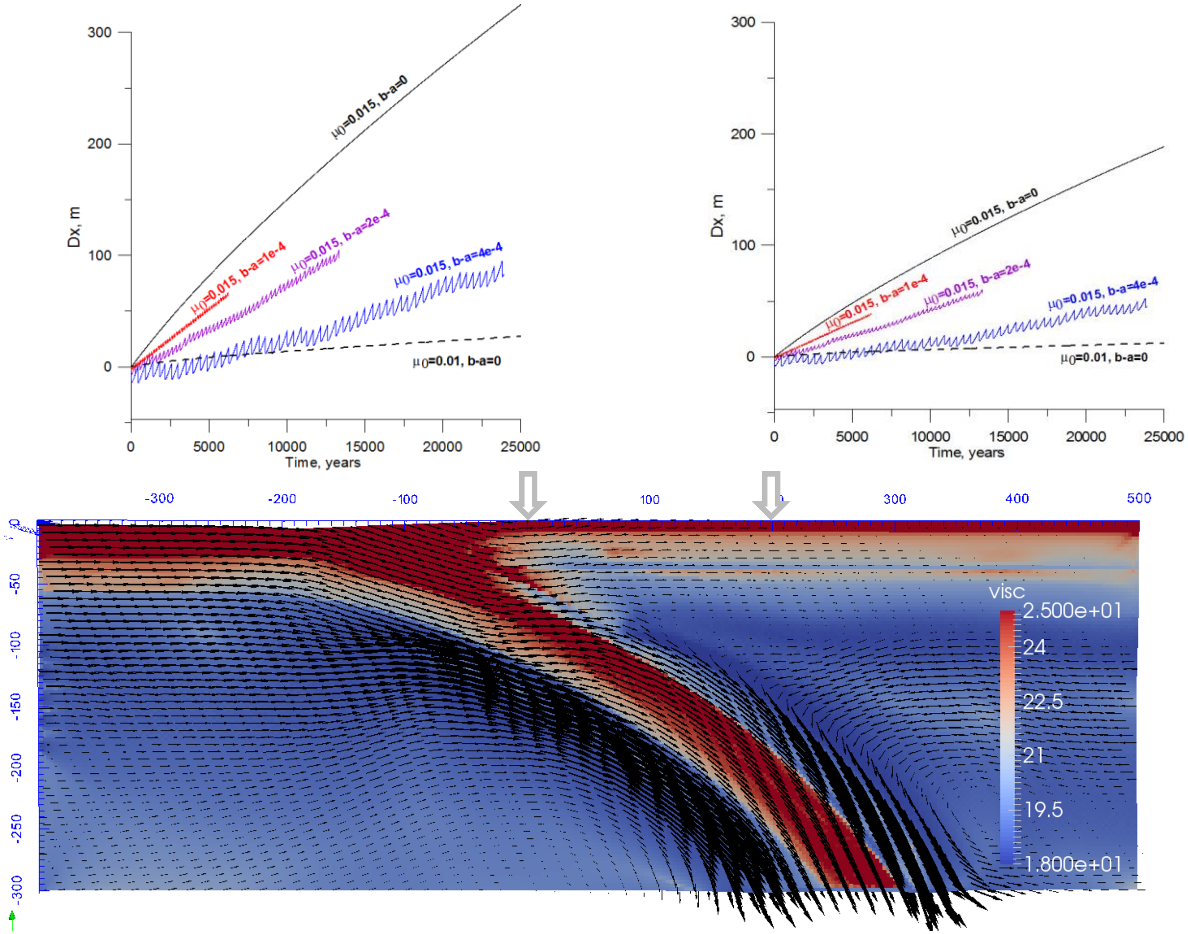
1 month after eq.

1 year after eq.

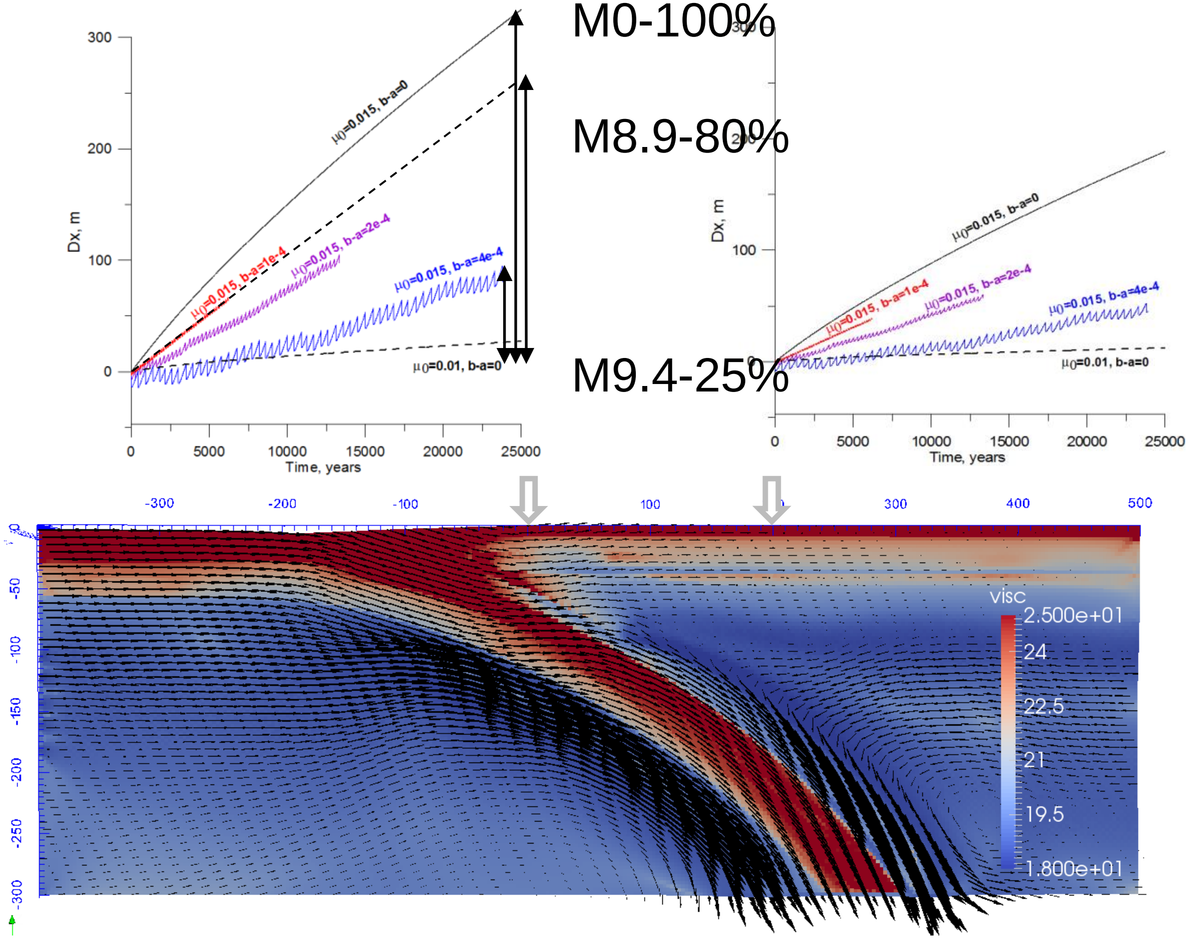


Interesting effects:
Upper plate deformation

Horizontal displacement



Horizontal displacement

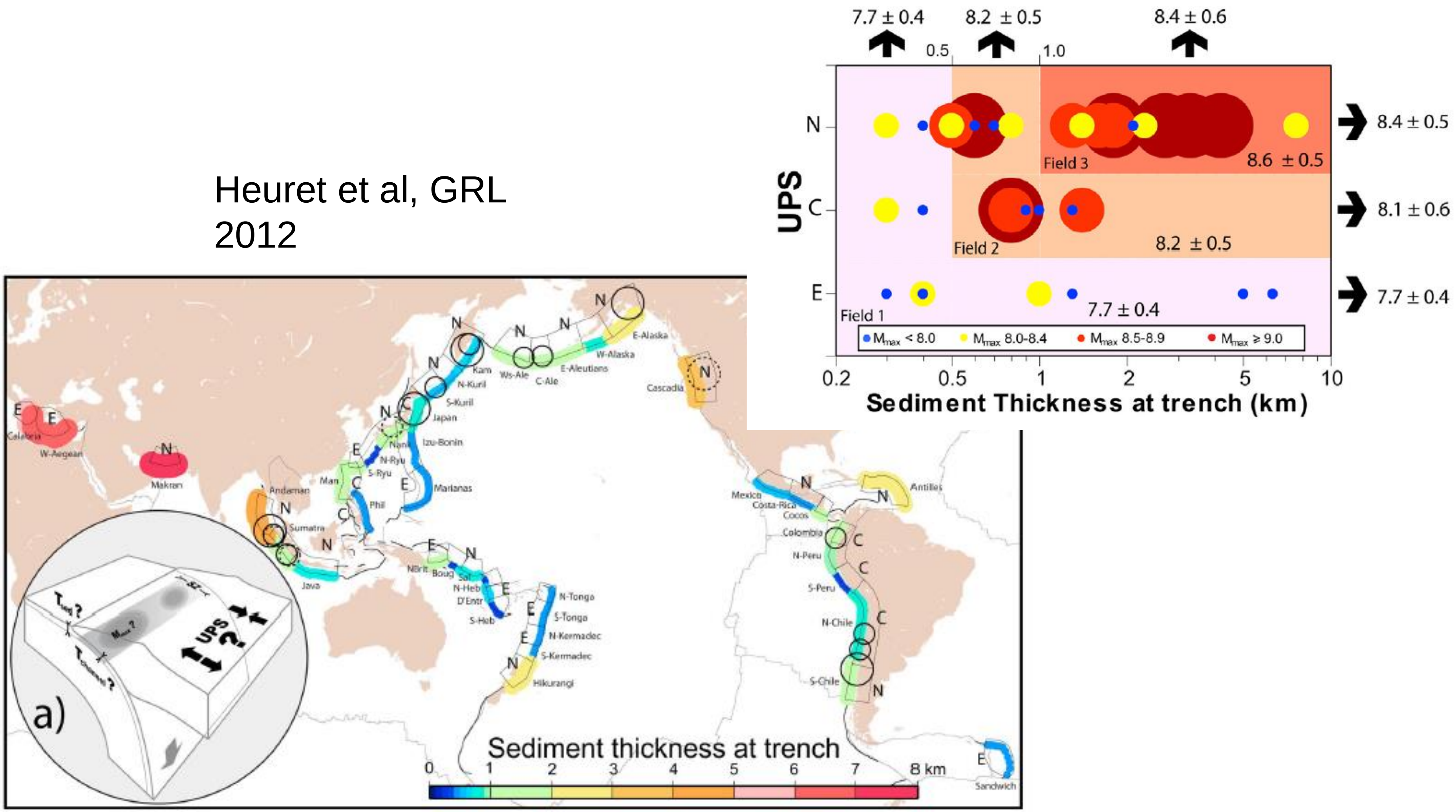


M0-100%

M8.9-80%

M9.4-25%

Heuret et al, GRL 2012



Conclusions (2D)

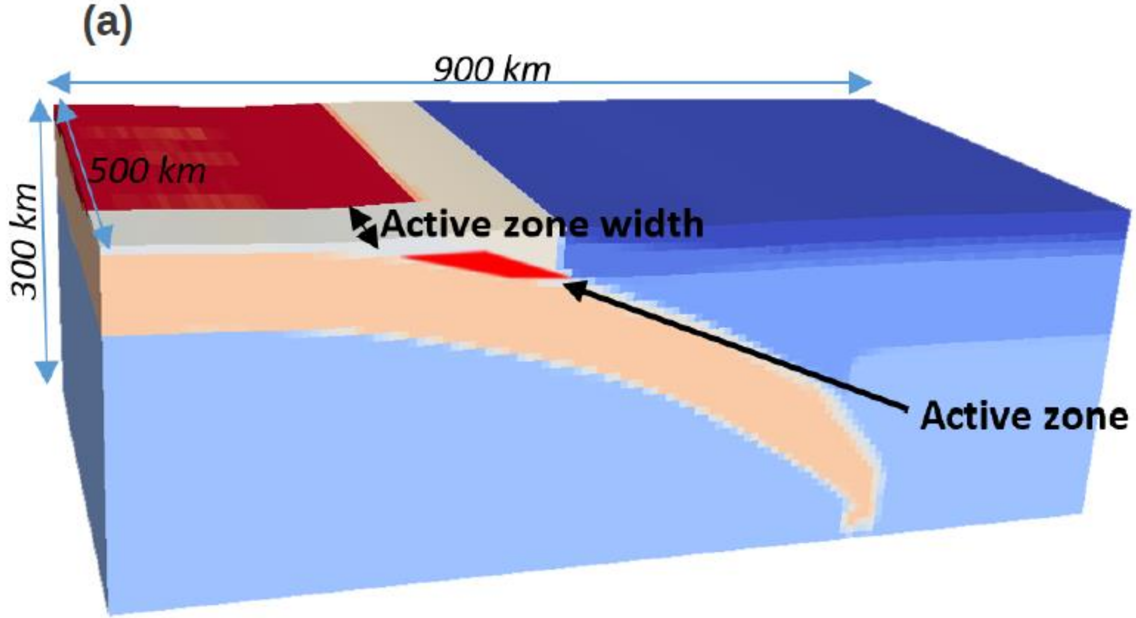
- We have developed the model able to simulate seismic cycle and subduction process in time scale range from rupture (minute) to geological time (Mln years)

- The model suggests that after the great (M>9) earthquake viscosity in the mantle wedge can drop by 4 orders of magnitude. As a result, surface displacements are controlled by the relaxation in mantle wedge already since 1 hour after the earthquake.

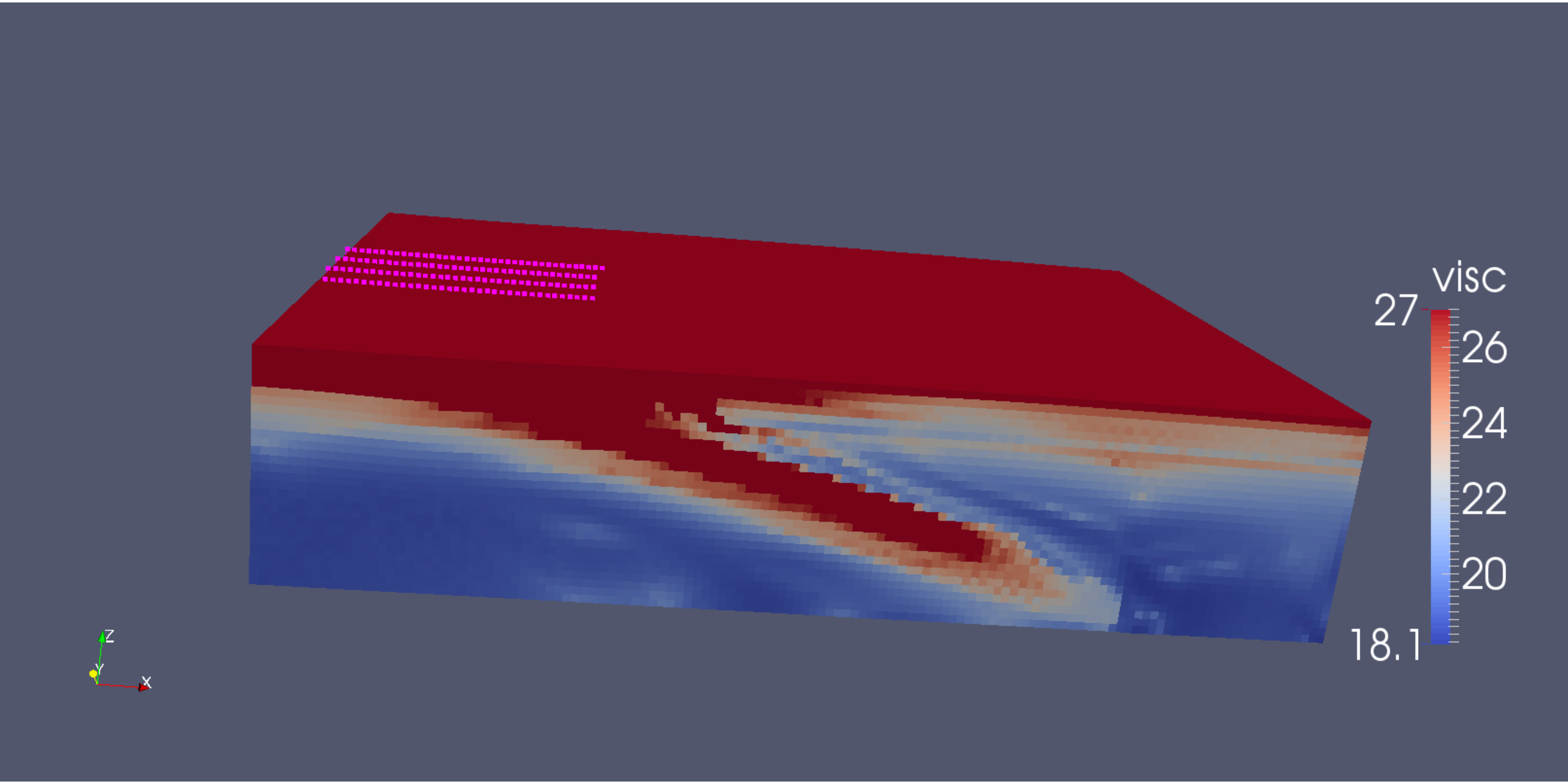
- The model is consistent with the short-time scale GPS data for Tohoku 2011 earthquake

- Many interesting effects show up in the models already but much more can be expected

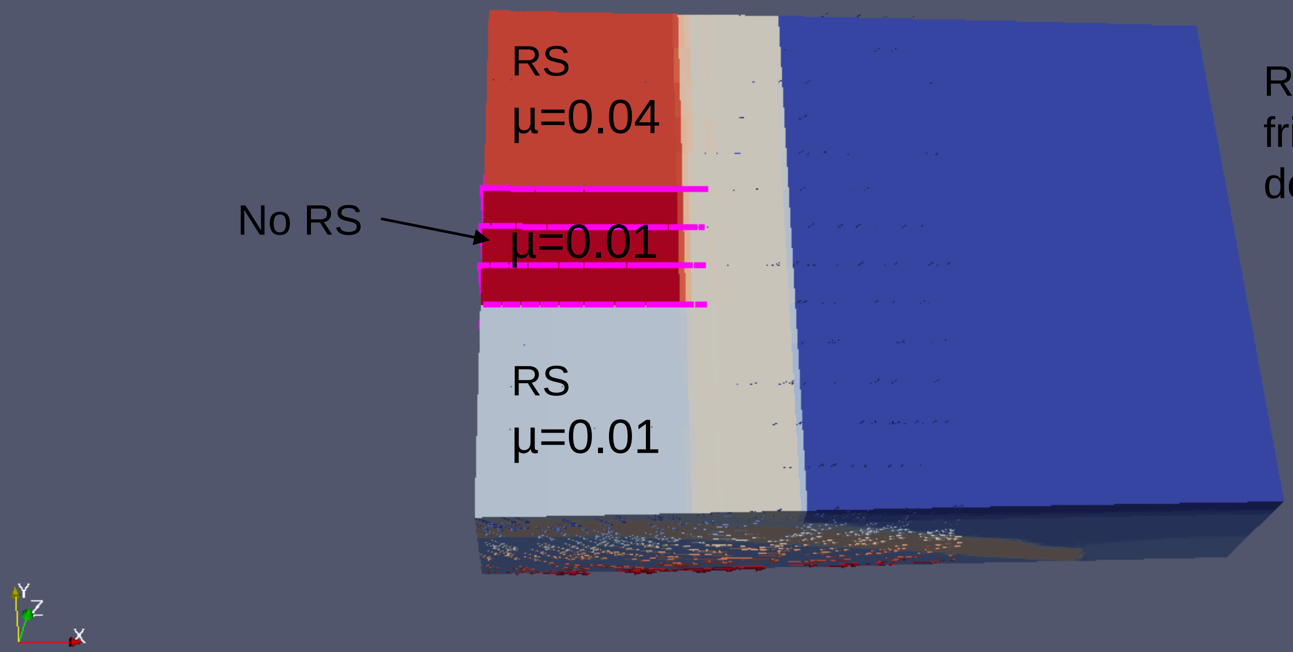
3D Modelling



SLIM3D Cross-scale model

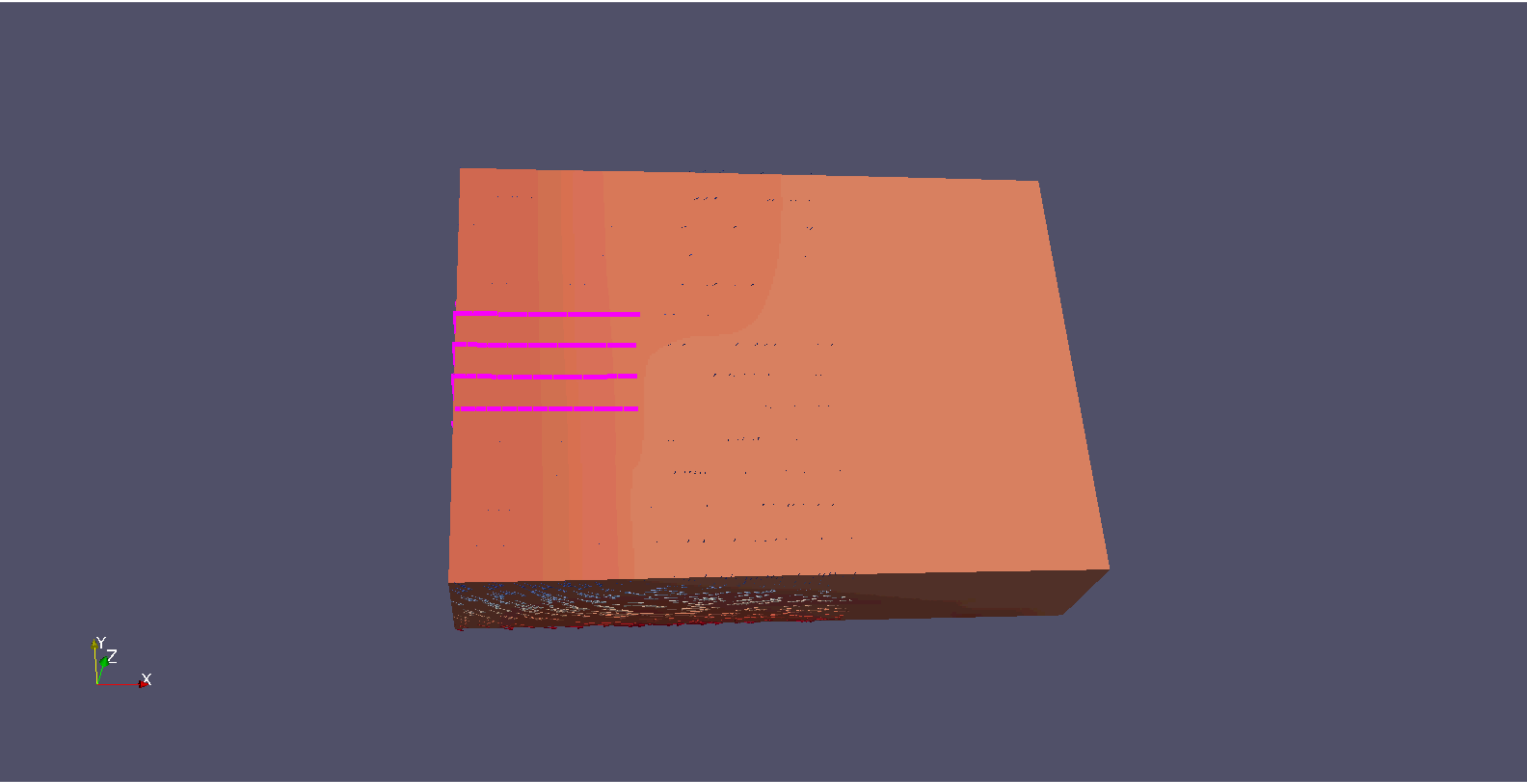


SLIM3D Cross-scale model



Rate and State (RS) friction is applied for the depth range 14-42 km

SLIM3D Cross-scale model



Conclusions

Great earthquakes may well happen within the very weak fault zones (subduction channels) with static friction about 0.01-0.05 due to the friction drop of about 0.005-0.01.

What makes earthquake great is not large stress drop, but rupturing at large area (homogeneous channel structure, no barriers).

Conclusions

Observed correlation with the structure of the upper plate (not subducting plate), in particular with presence of sedimentary basins above seismogenic zones, is surprising and intriguing.

The best (till now) explanation is stability (and low permeability) of the wedge (Fuller at al, 2005), but their model needs update

Interesting perspective is a cross-scale modeling allowing simulation of seismic cycle in the same model that explains geological-time-scale processes