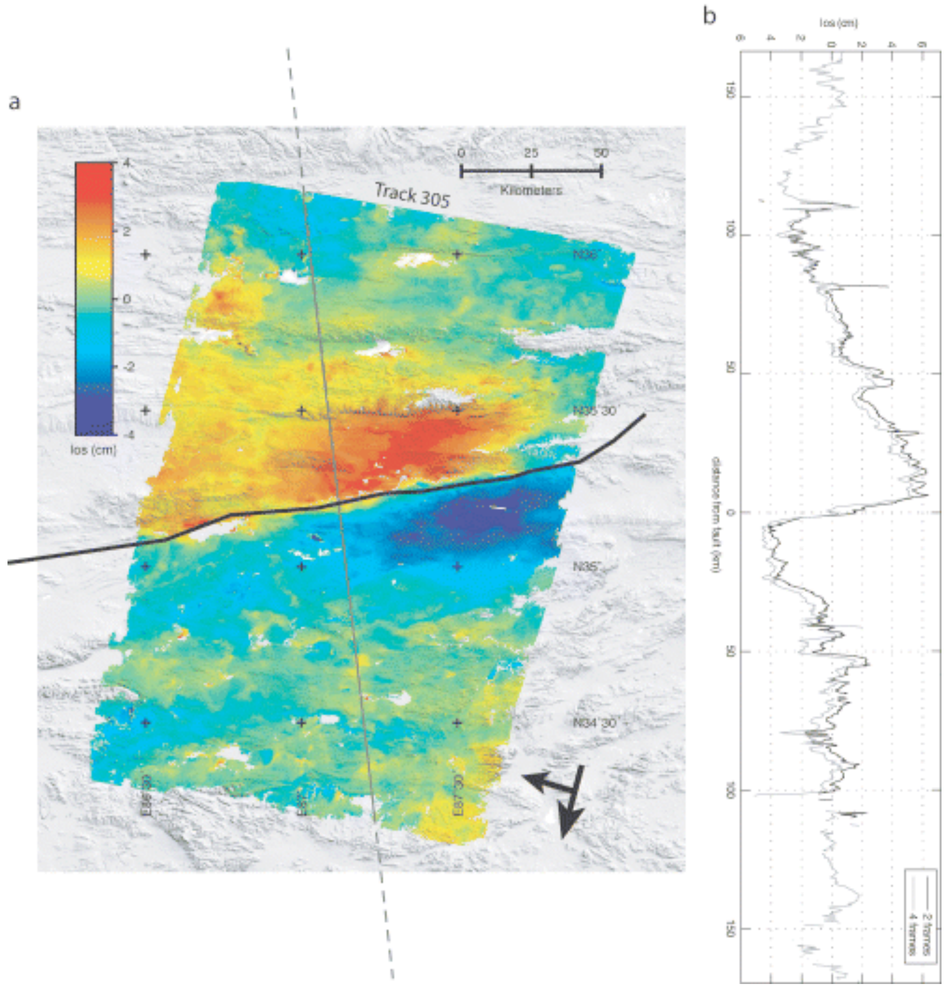
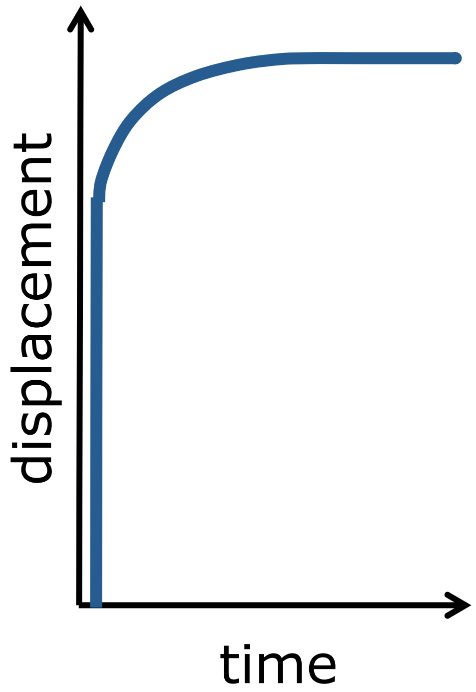
Postseismic Deformation



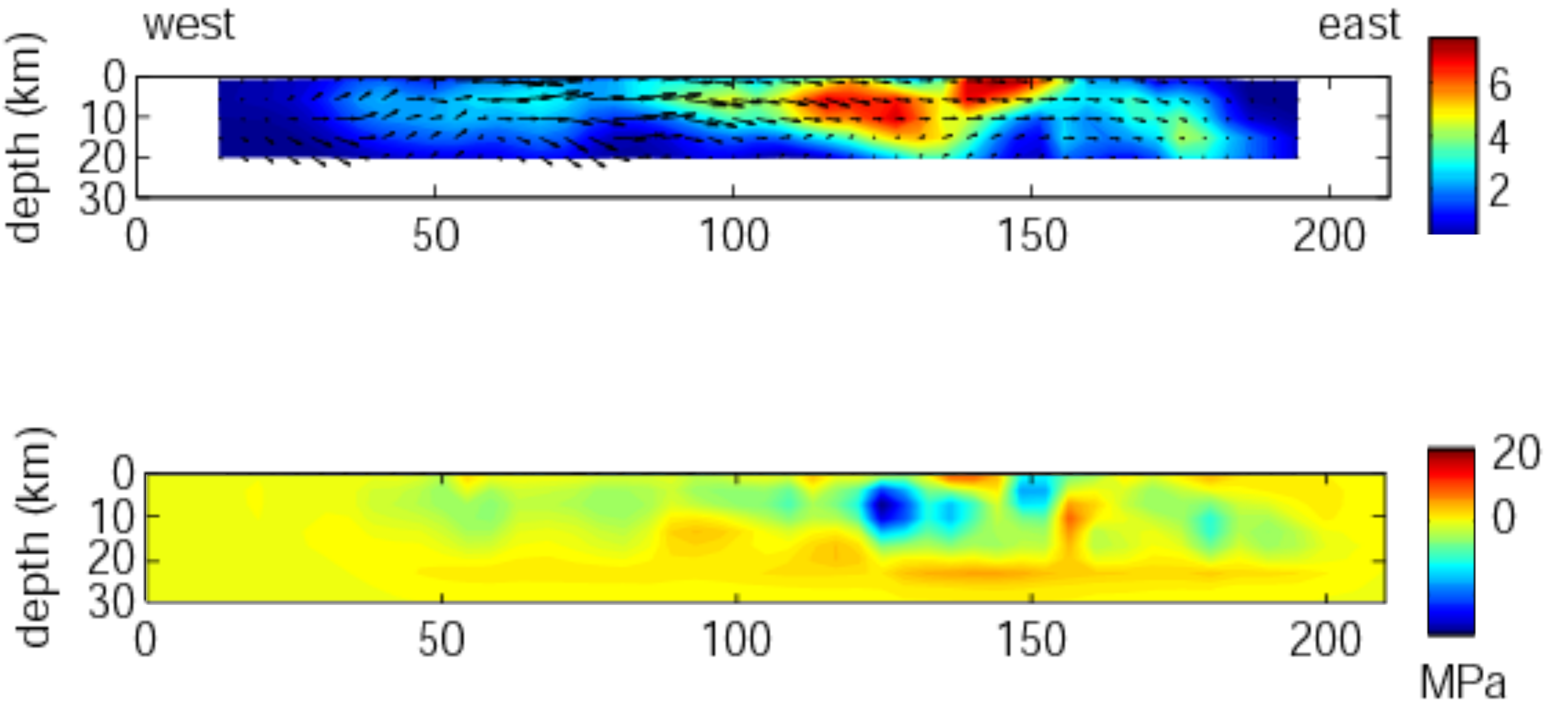
Juliet Biggs, Tim Wright
COMET+ [1]University of Bristol; [2] University of Leeds

Postseismic motion

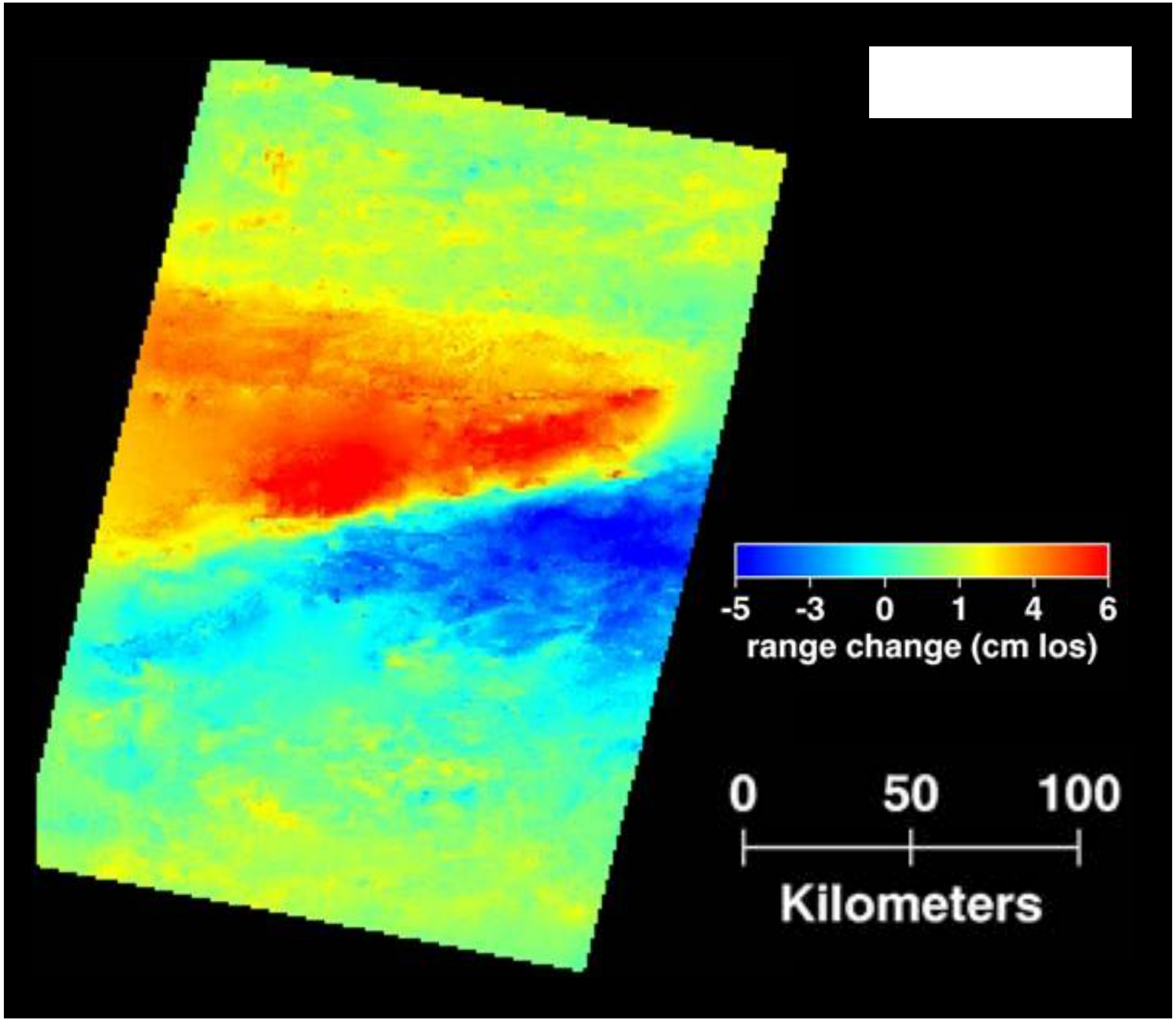


- **What is it?**
  - **Transient motion following an earthquake**
  - **Response to stress changes**
- **Why study it?**
  - **Stress relaxation**
  - **Rheology**

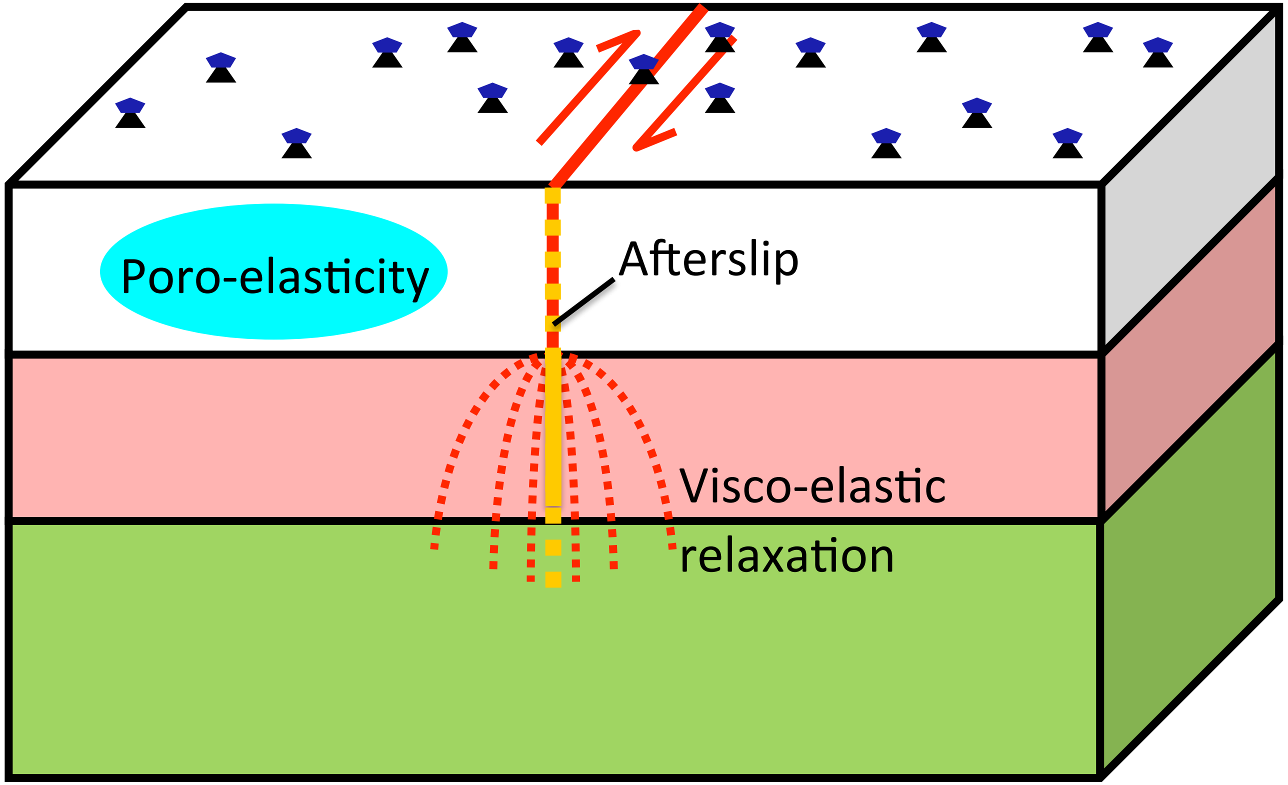
Coseismic slip induces stress



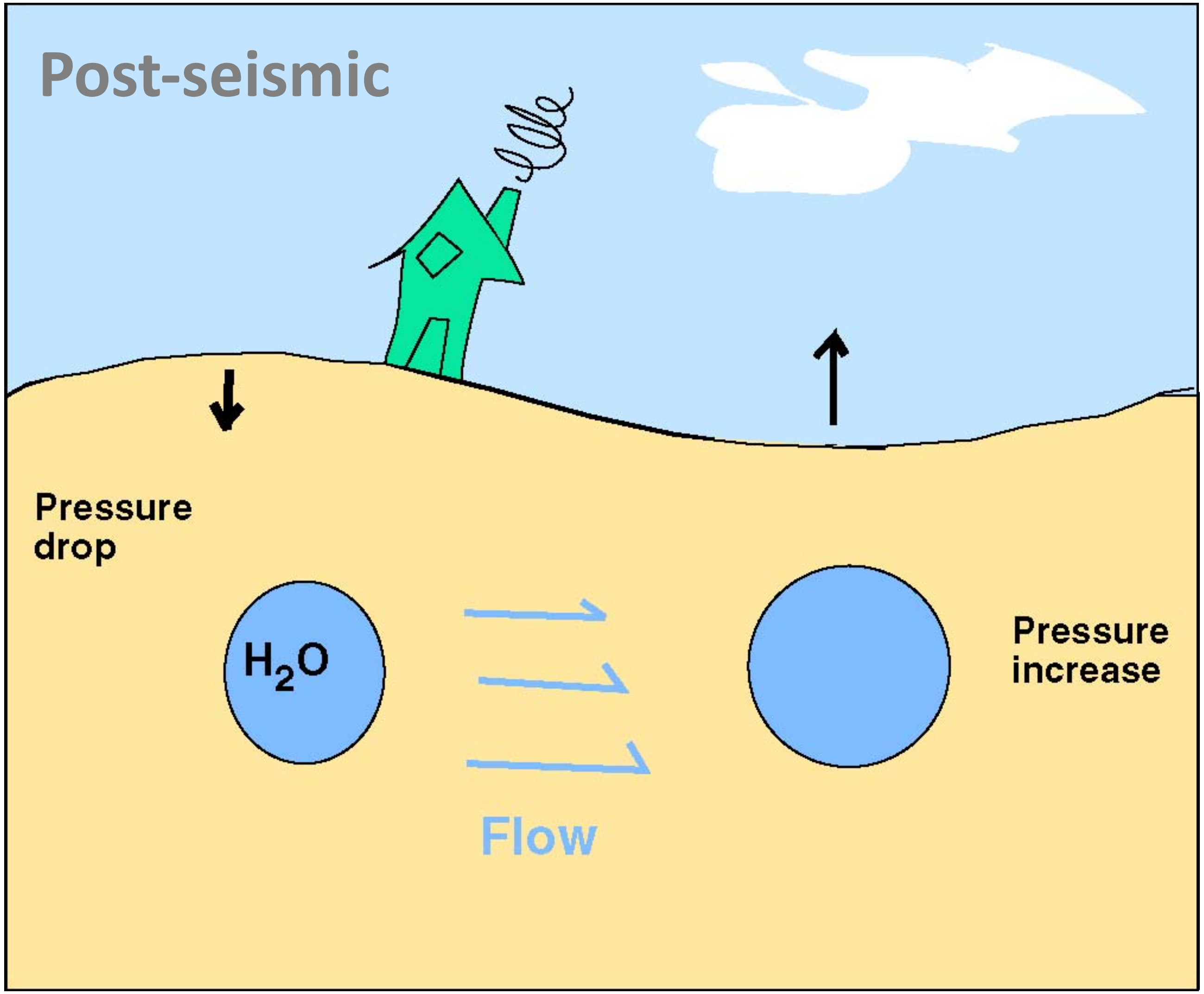
Causes deformation



Which mechanism?

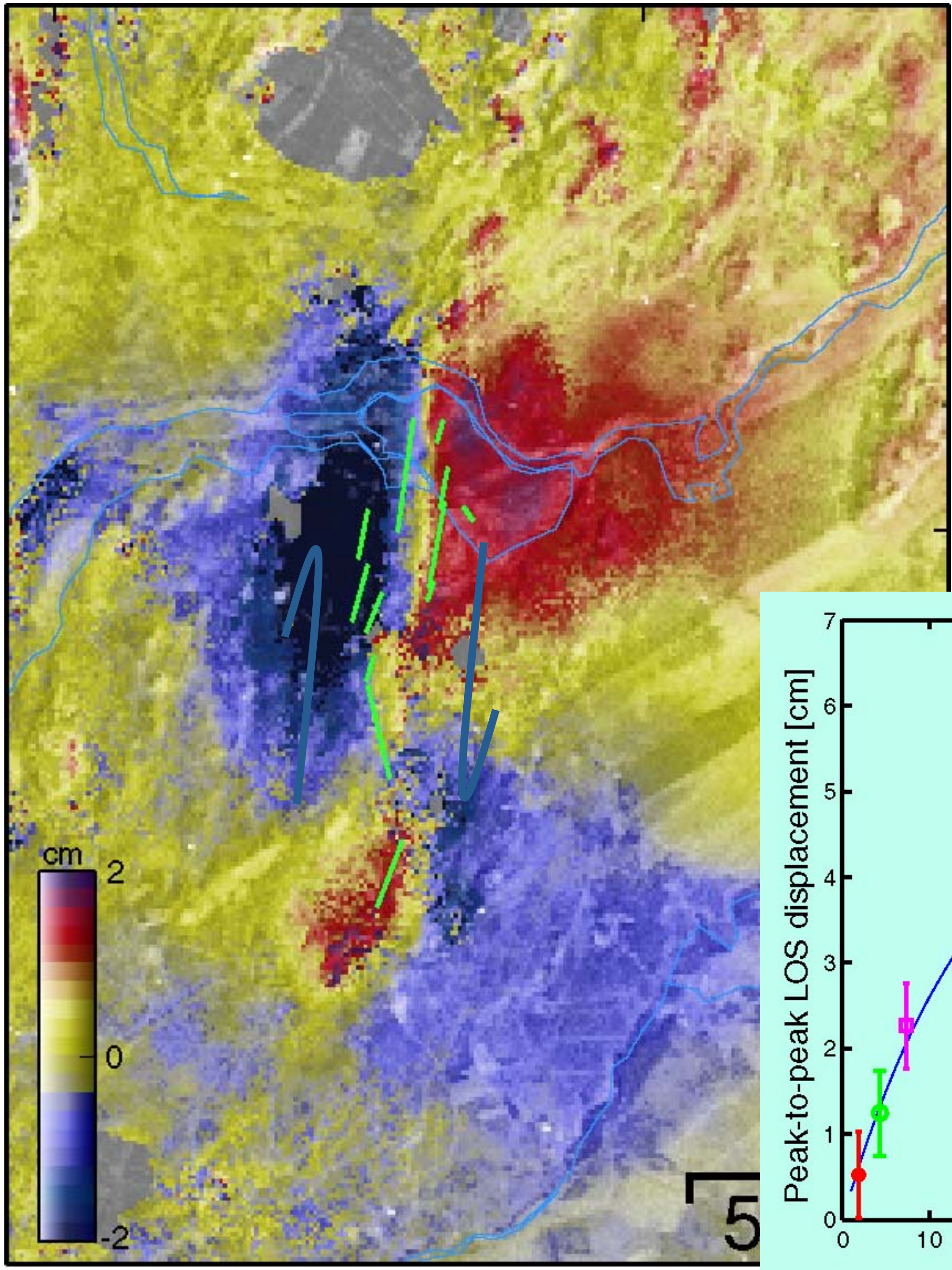


Poro-elastic rebound

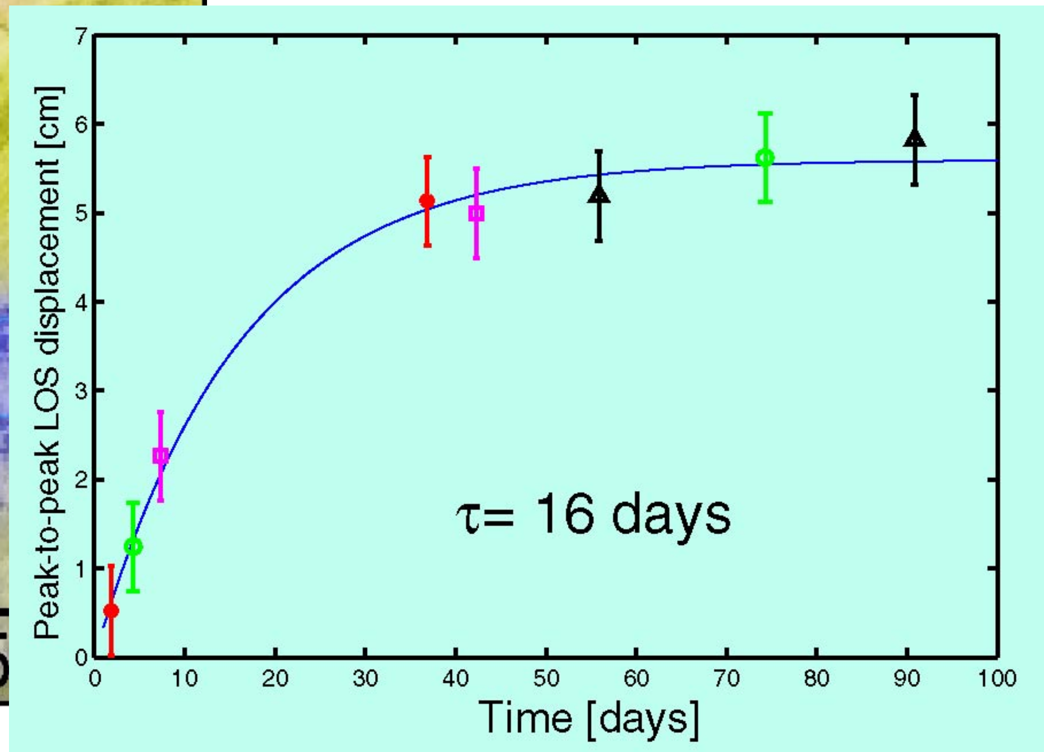


Post-seismic deformation during June 19 – July 24

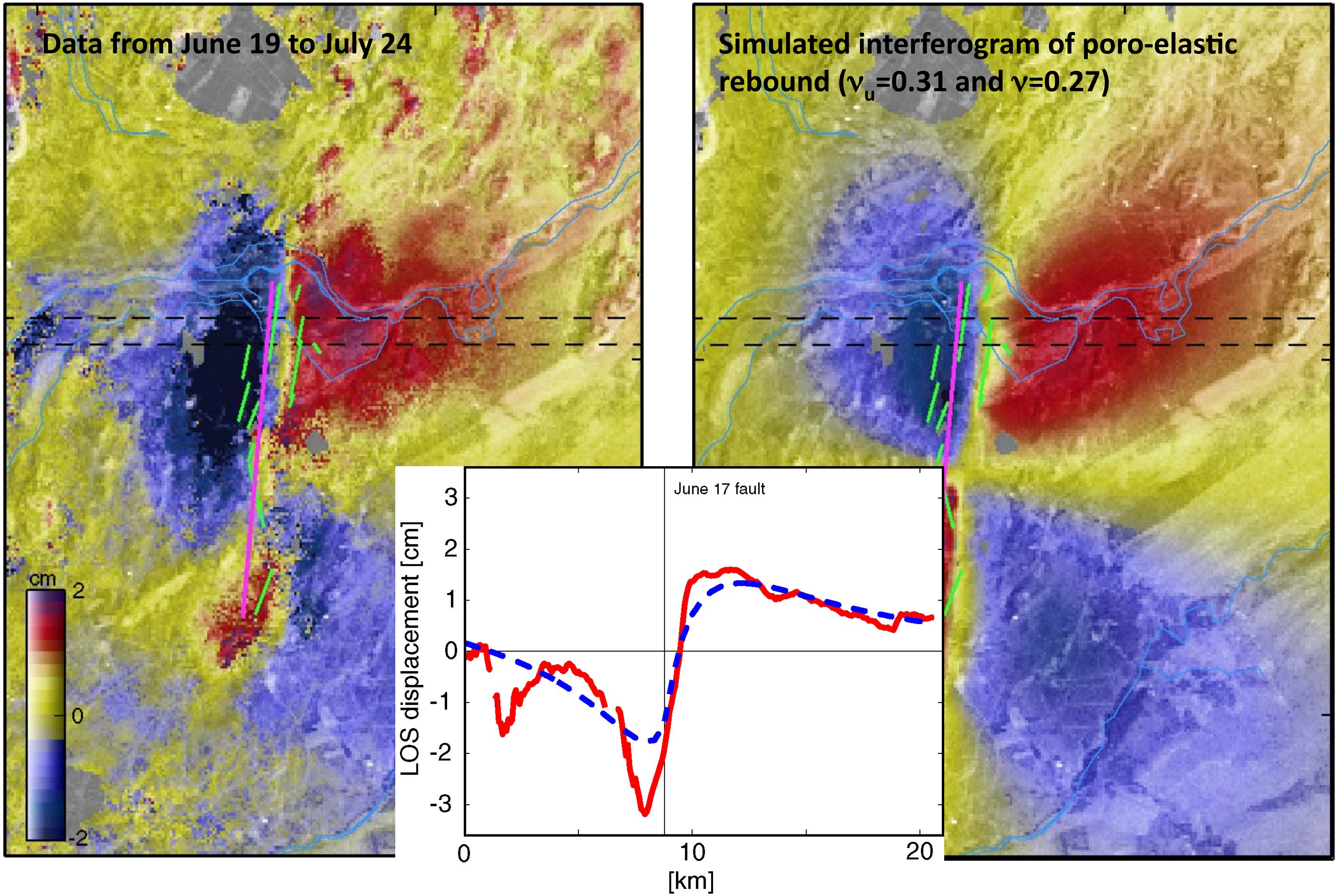
2-37 days after earthquake



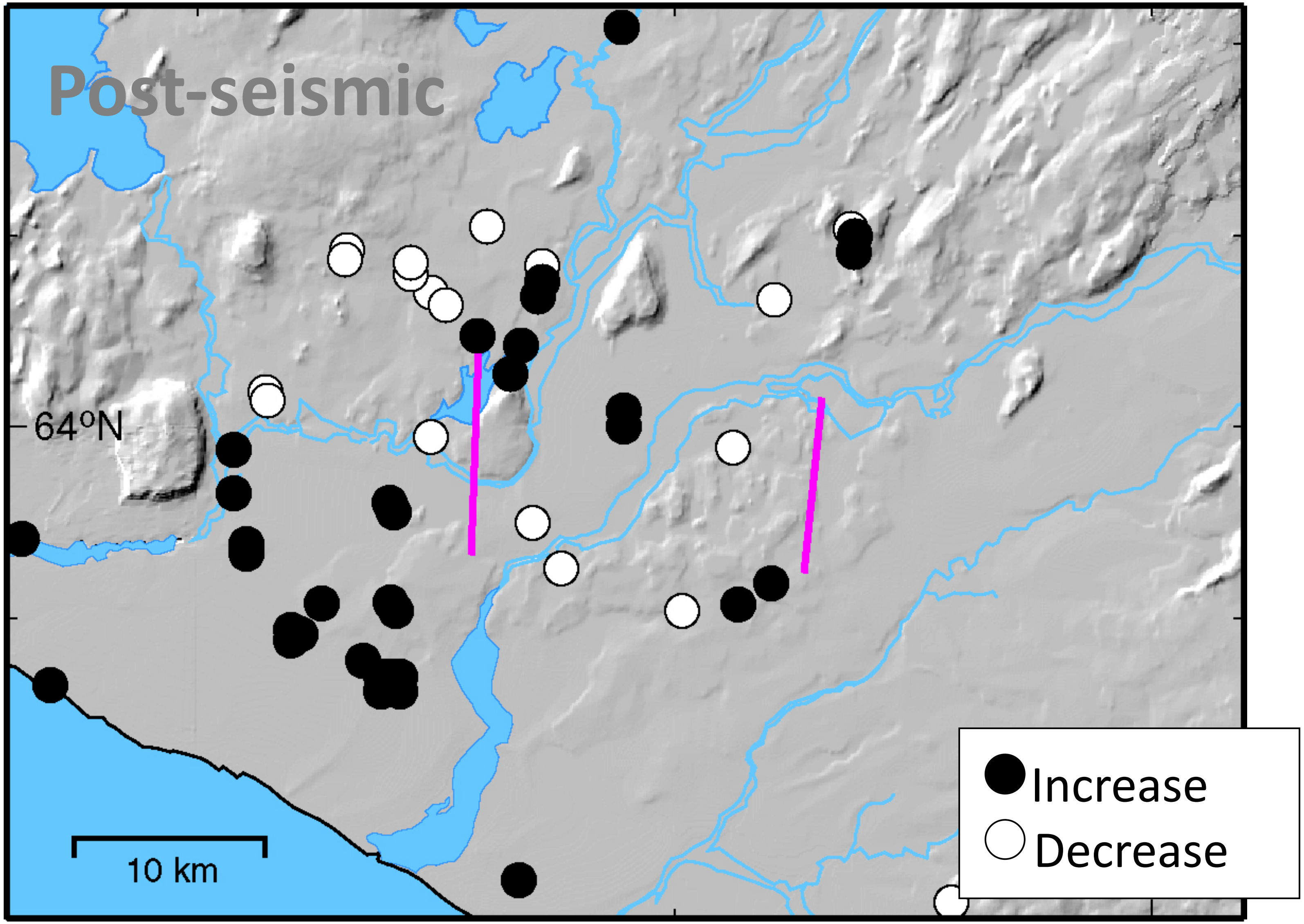
Duration of transient



Poro-elastic rebound near the June 17 fault

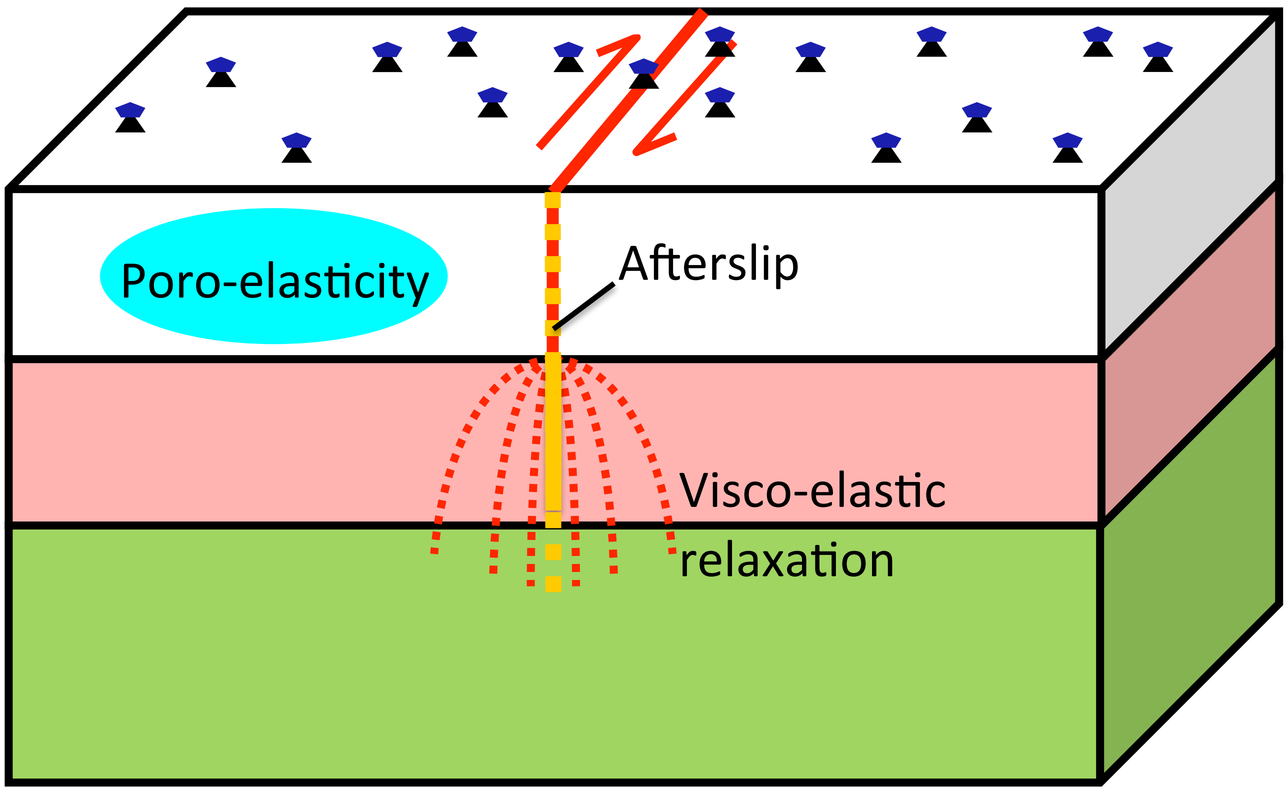


Water Level Changes in Geothermal Wells

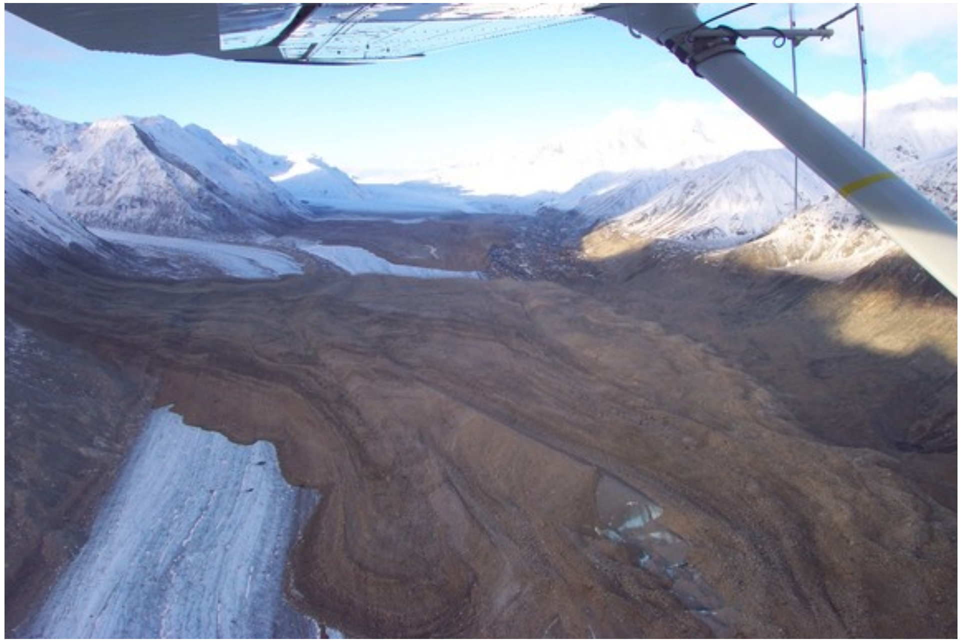


Jónsson et al., Nature, 2003

Postseismic Deformation



Denali Earthquake



USGS Press Release

3 November 2002. M_w 7.9.

340 km rupture on Denali, Susitna Glacier and Totshunda Faults.

Max. offset = 9m.

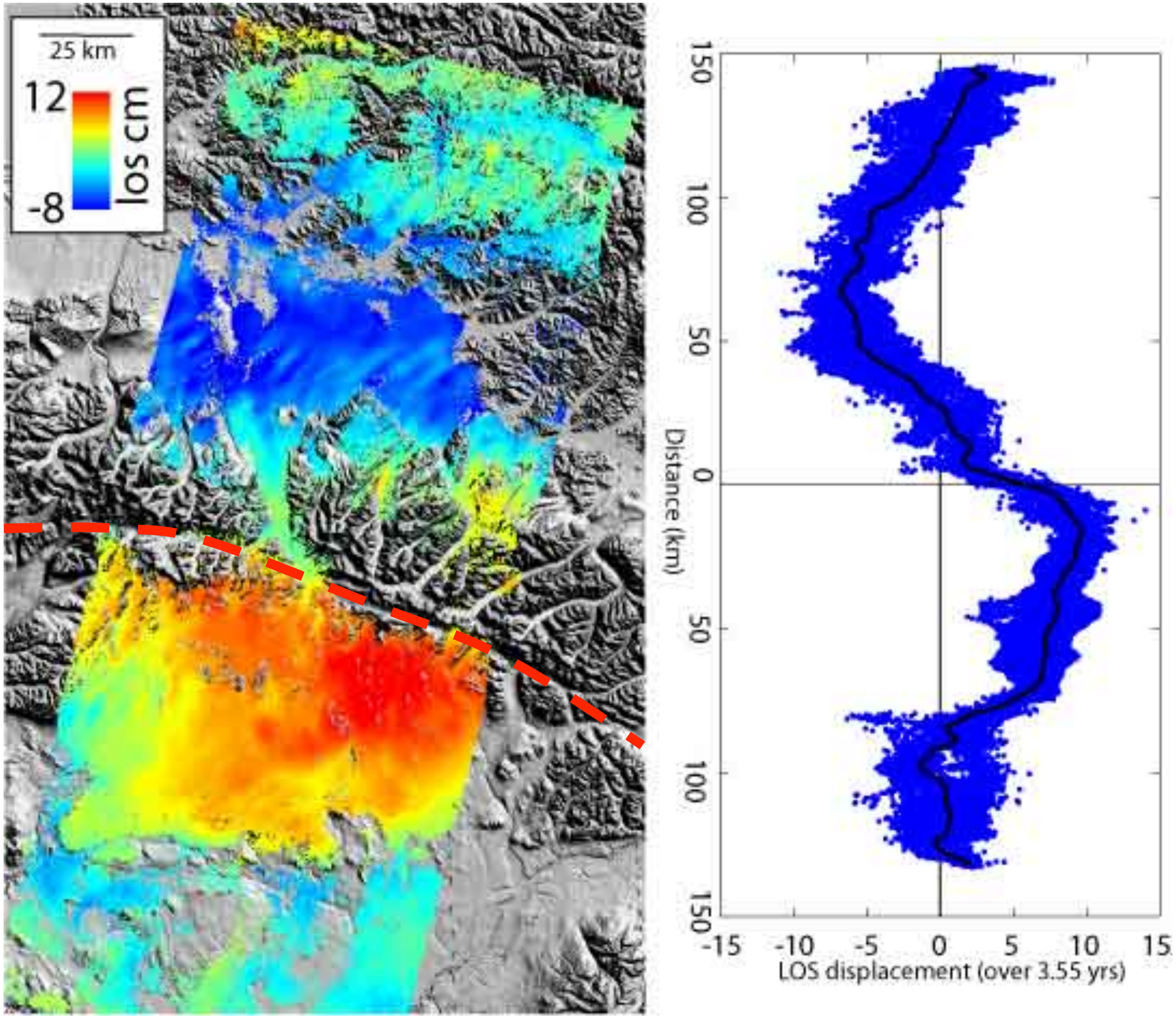
Denali Earthquake

Trans Alaska Pipeline and Richardson Highway

Denali Earthquake

Trans Alaska Pipeline

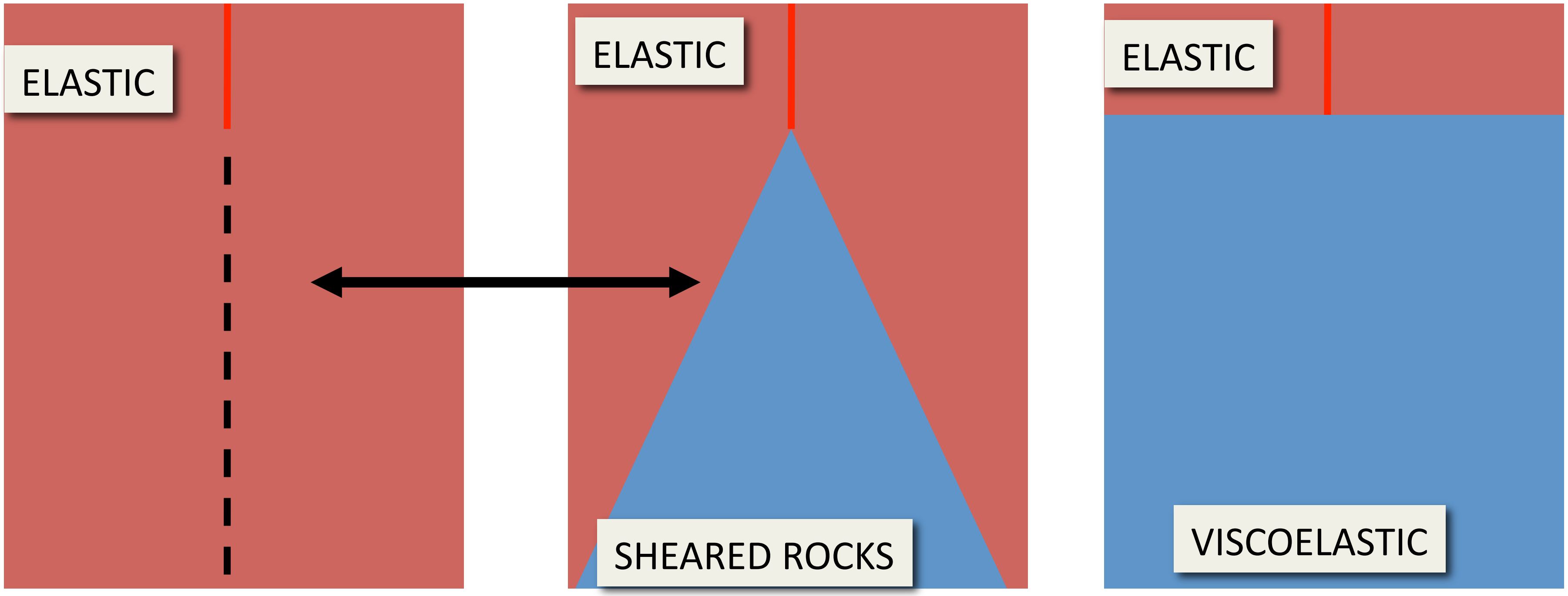
Richardson Highway



Summer 2003 (8 months) - Summer 2004 (20 months)

Biggs et al, 2009

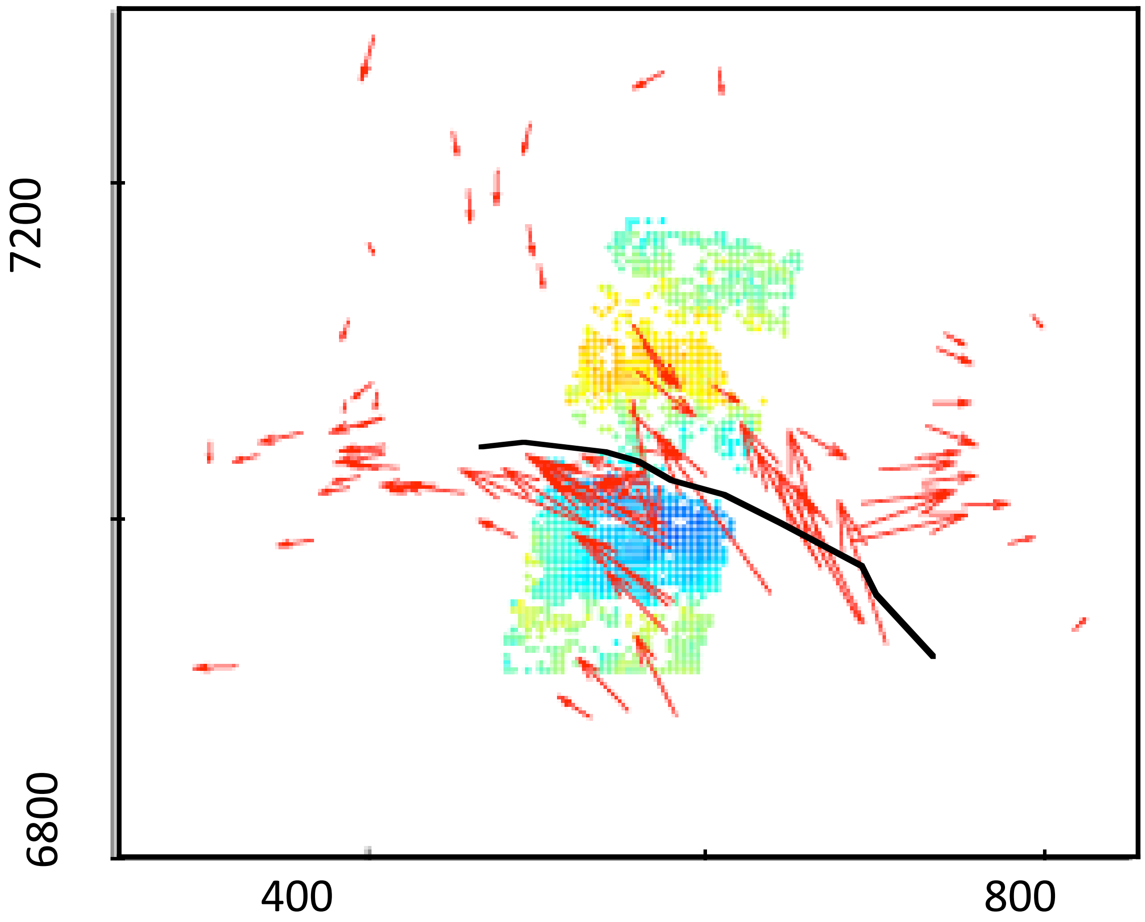
What happens under the seismogenic layer?



Single Fault,
Continuously Sliding

Shear Zone

Viscoelastic
Rheology

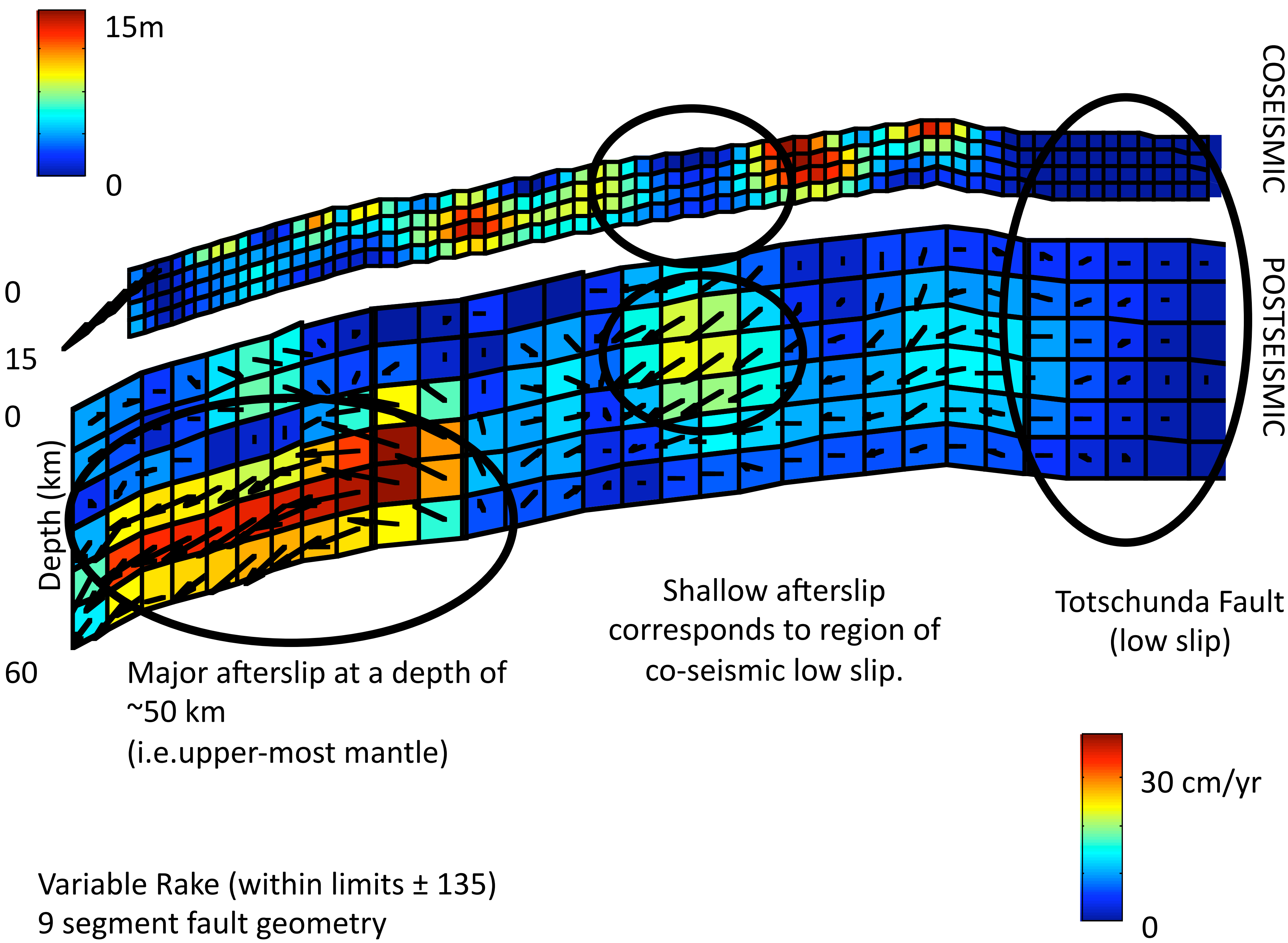


INSAR:

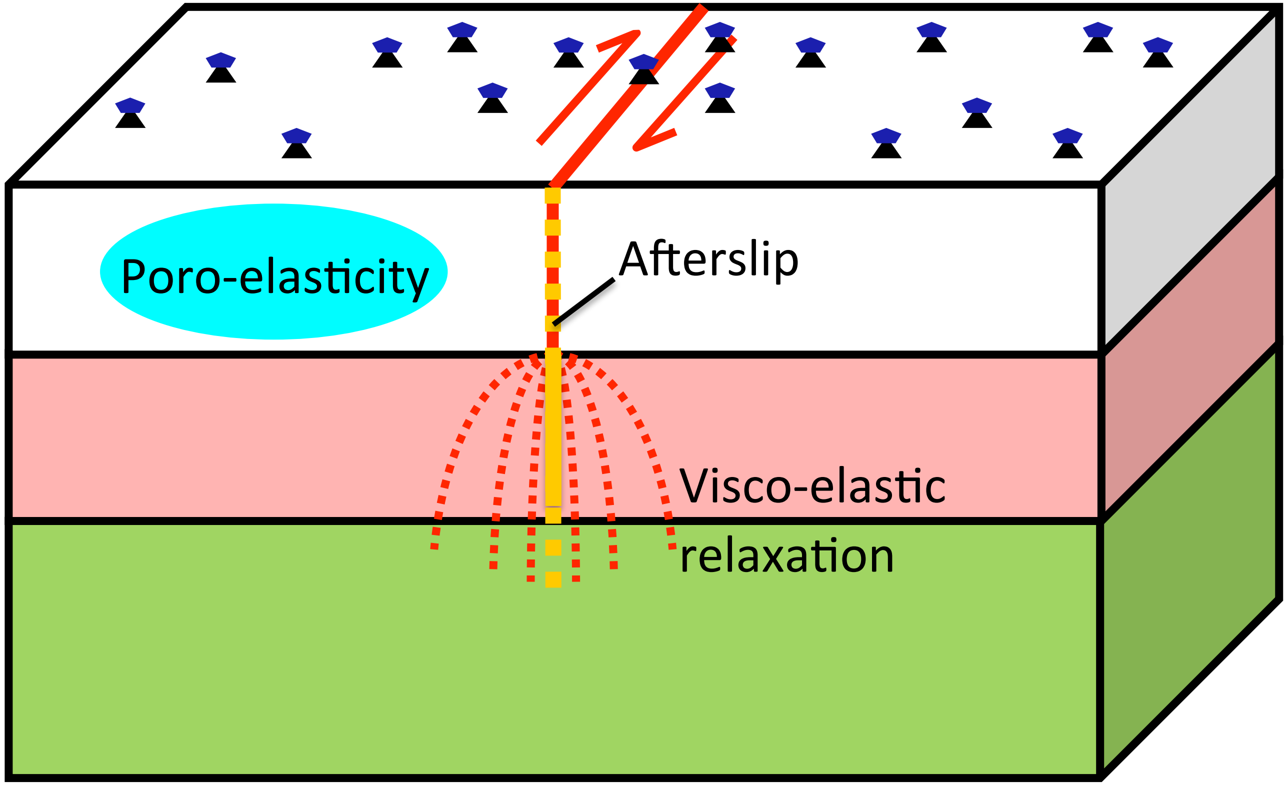
- Sub-sampled (1185 pts)
- Satellite line of sight (~23^{o} to vertical)

GPS:

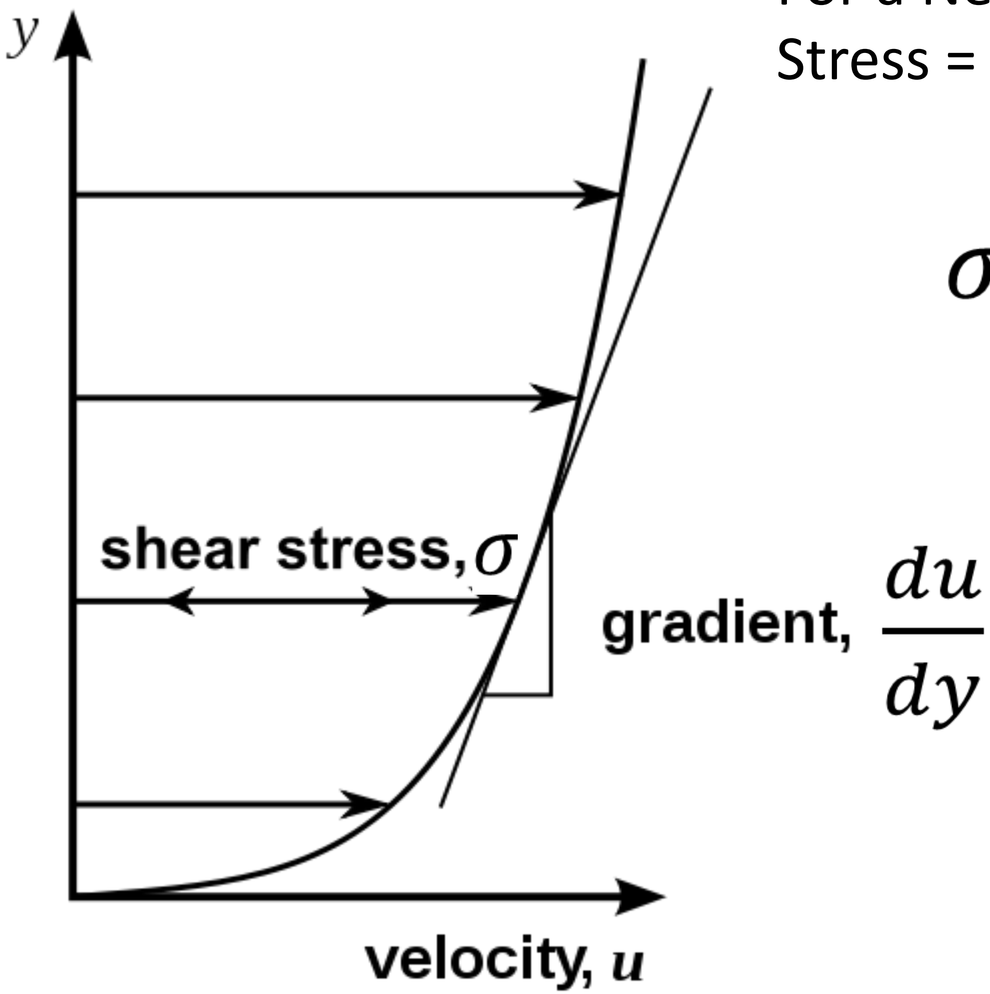
- 86 sites (errors < 1cm/yr)
- East and North components
- Corrected for interseismic vel. (Elliott et al)


15m
0
COSEISMIC
POSTSEISMIC
0
15
0
Depth (km)
60
Major afterslip at a depth of
~50 km
(i.e.upper-most mantle)
Shallow afterslip
corresponds to region of
co-seismic low slip.
Totschunda Fault
(low slip)
30 cm/yr
0
Variable Rake (within limits ± 135)
9 segment fault geometry

Postseismic Deformation



Viscosity

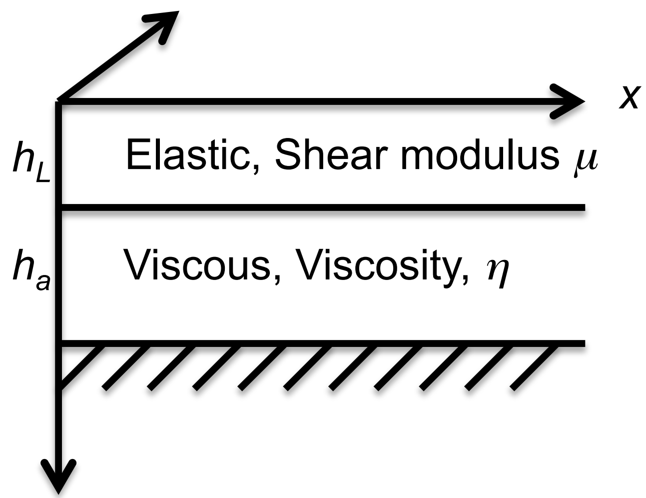


For a Newtonian (linear) viscous fluid,
Stress = viscosity x strain rate

$$\sigma = \eta\dot{\varepsilon} = \eta\frac{du}{dy}$$

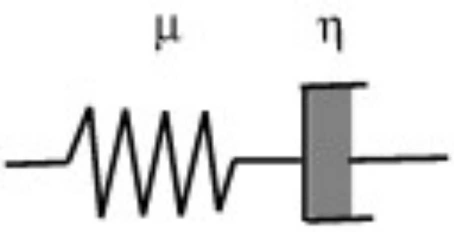
Stress Diffusion

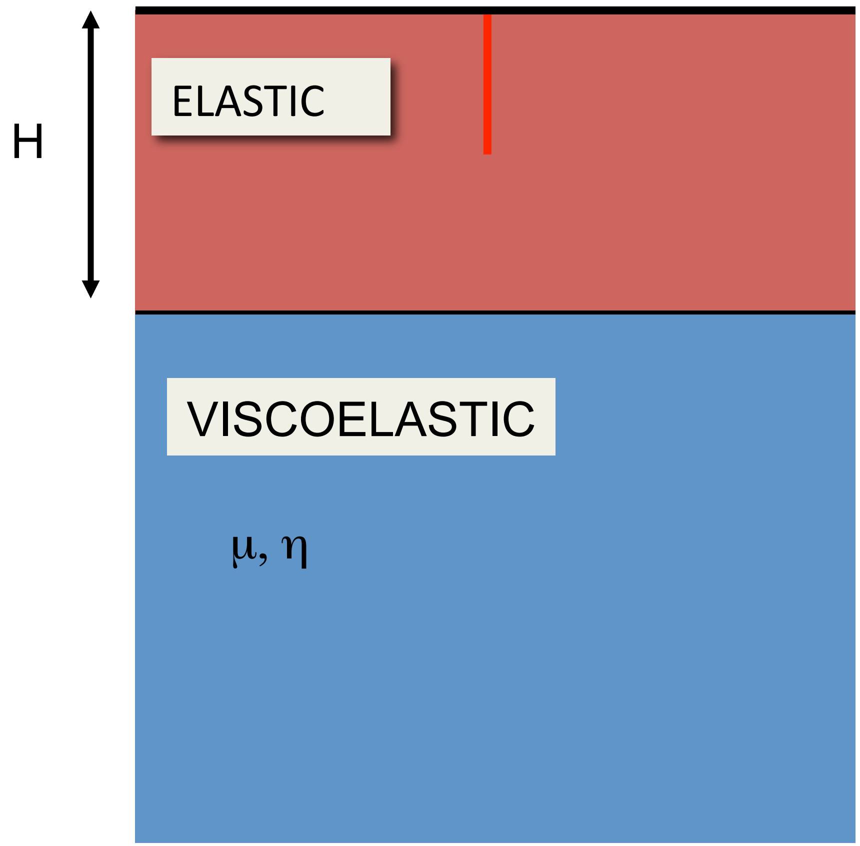
- For an elastic lid over a viscous channel, we can show that the deformation at the surface obeys the diffusion equation.
- Hence solution is identical to heat flow in the oceans.

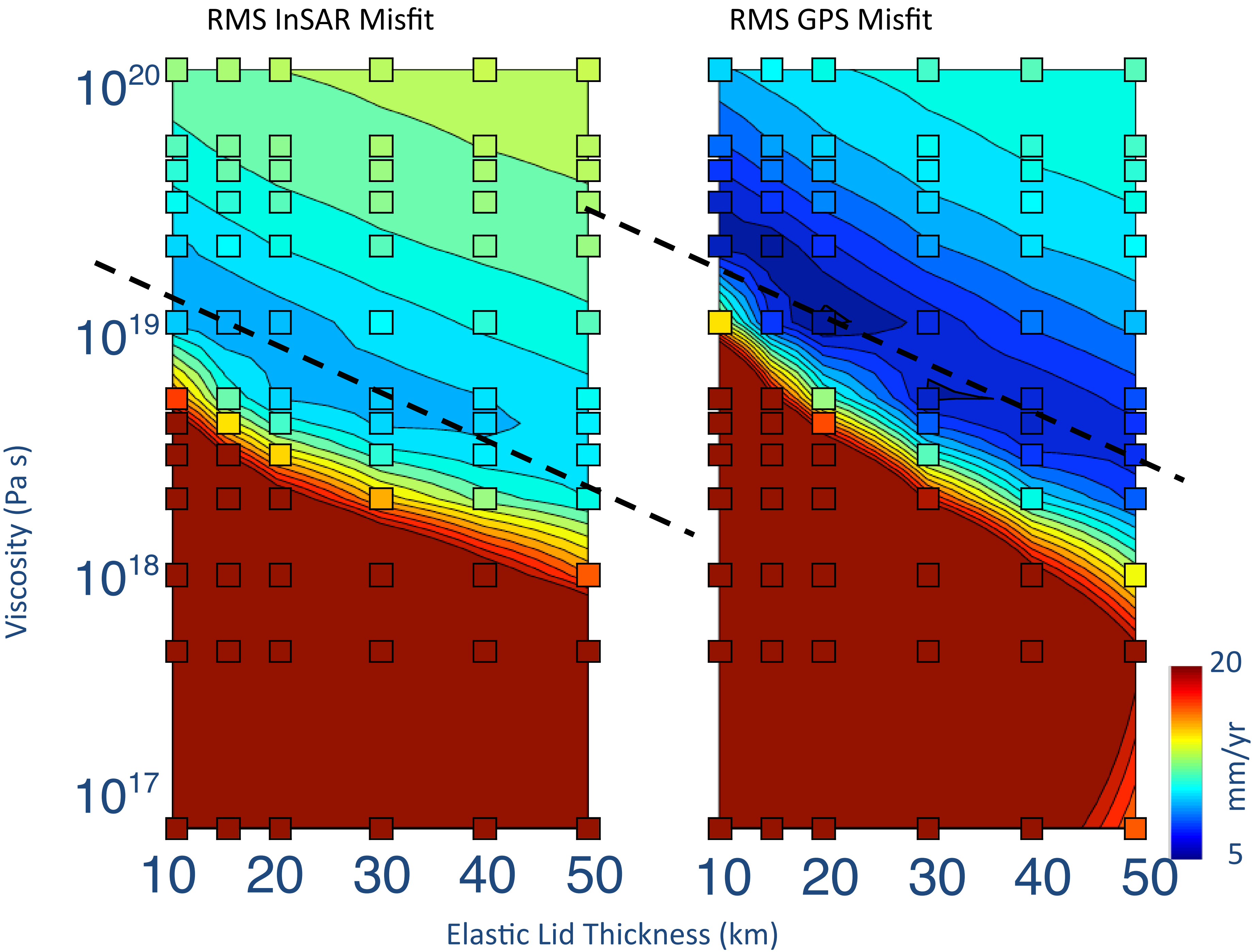


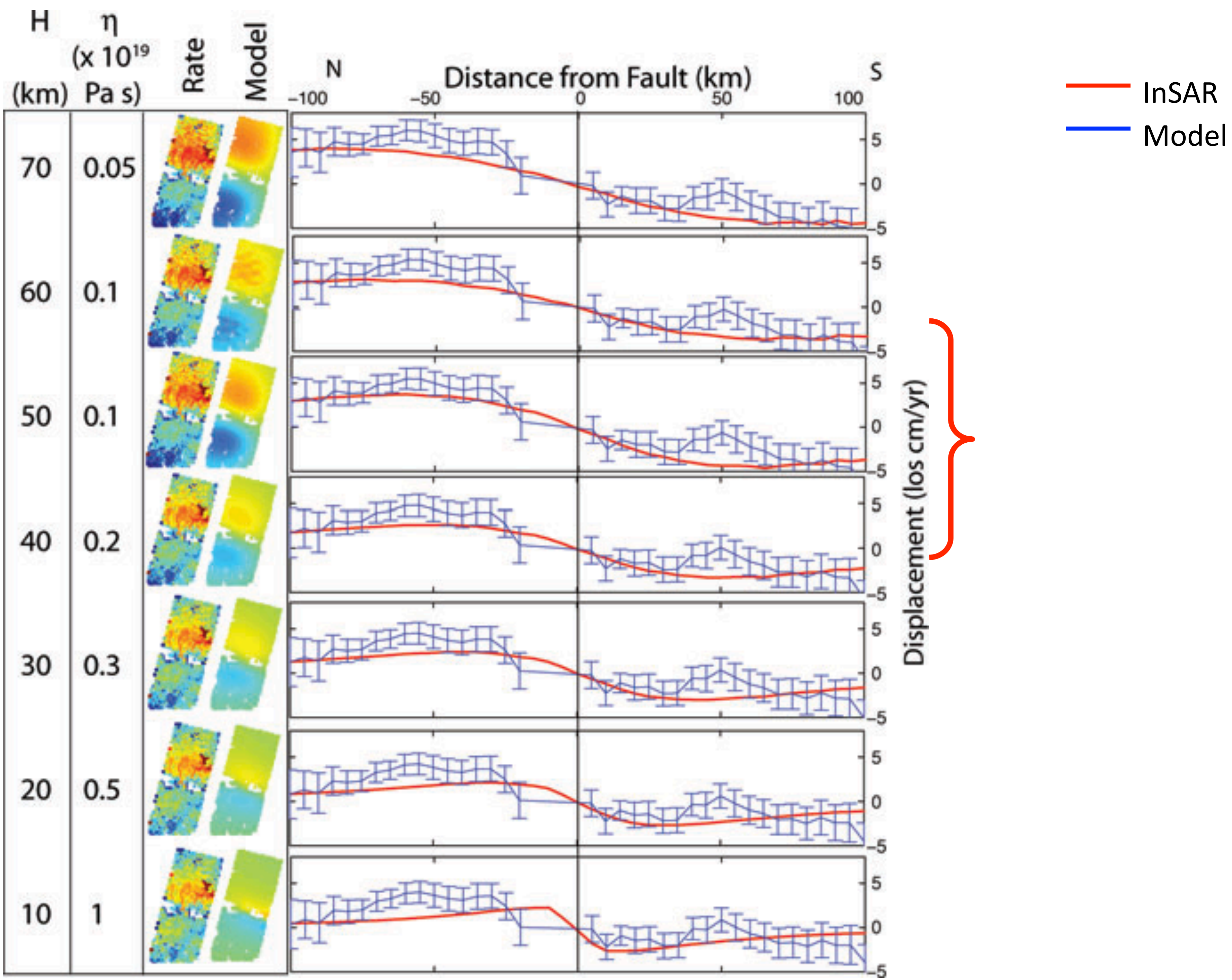
Viscoelastic Relaxation

- Earthquakes are not infinitely long
- Below the elastic lid, the earth is not viscous, it is **viscoelastic** (it transmits elastic waves on short time scales)

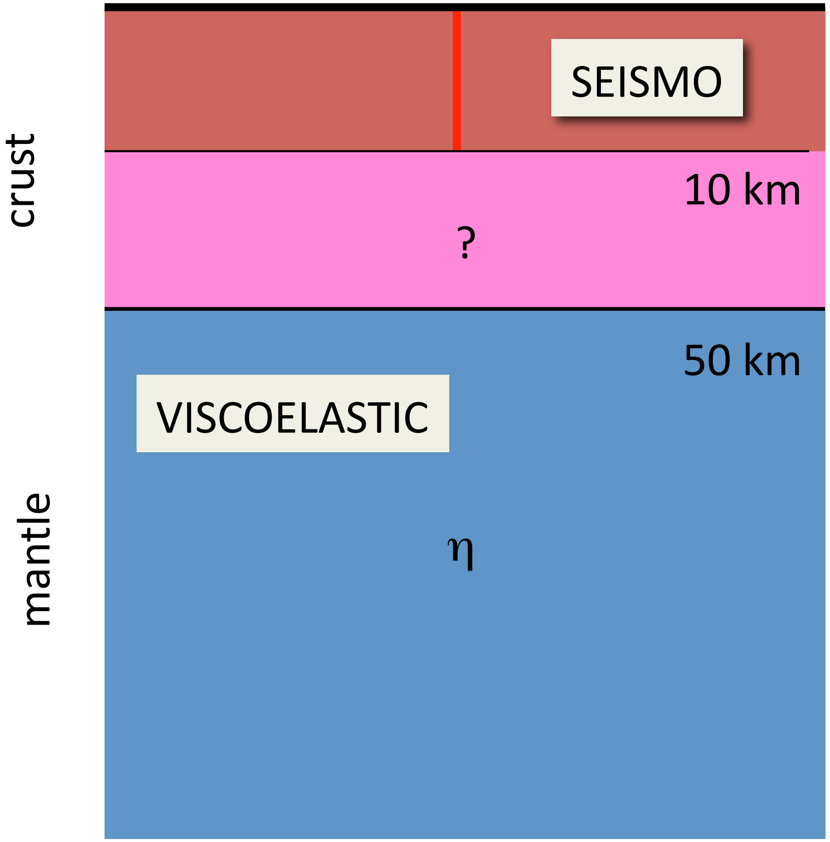


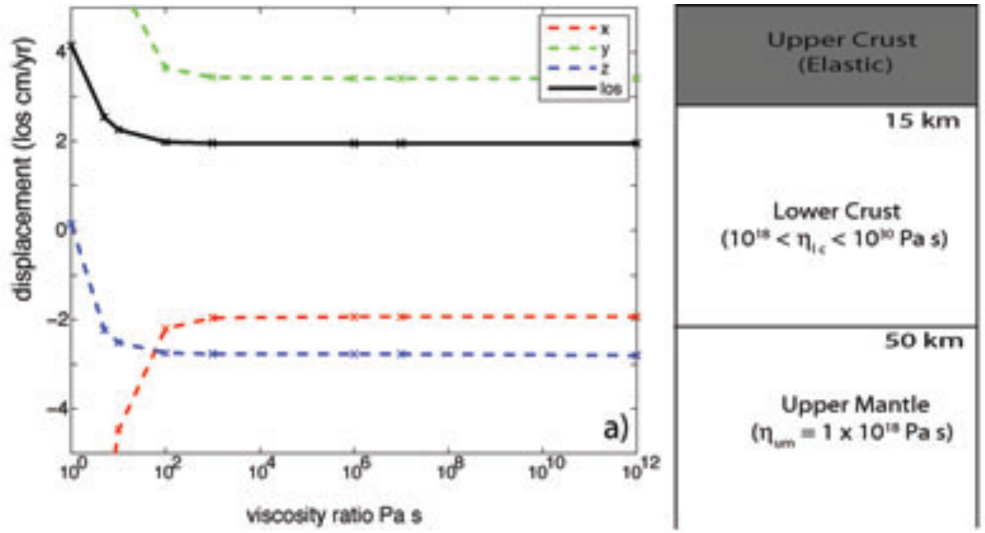


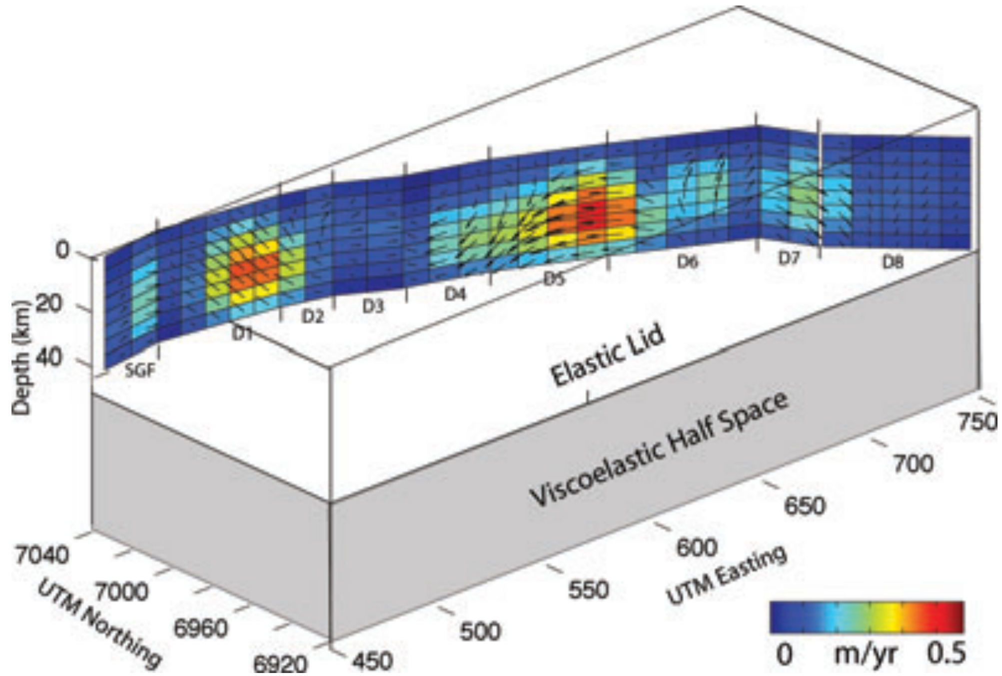

RMS InSAR Misfit
RMS GPS Misfit
10^{20}
10^{19}
10^{18}
10^{17}
Viscosity (Pa s)
10
20
30
40
50
Elastic Lid Thickness (km)
20
5
mm/yr


H (km)
η (x 10^19 Pa s)
Rate
Model
N
Distance from Fault (km)
S
−100
−50
0
50
100
70
0.05
60
0.1
50
0.1
40
0.2
30
0.3
20
0.5
10
1
5
0
−5
Displacement (los cm/yr)
InSAR
Model

What is the elastic lid?



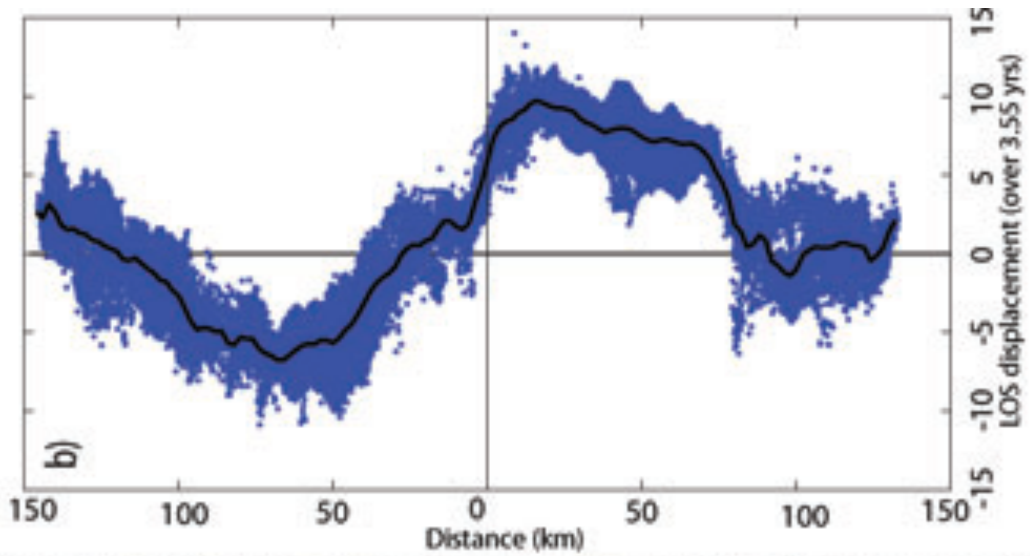


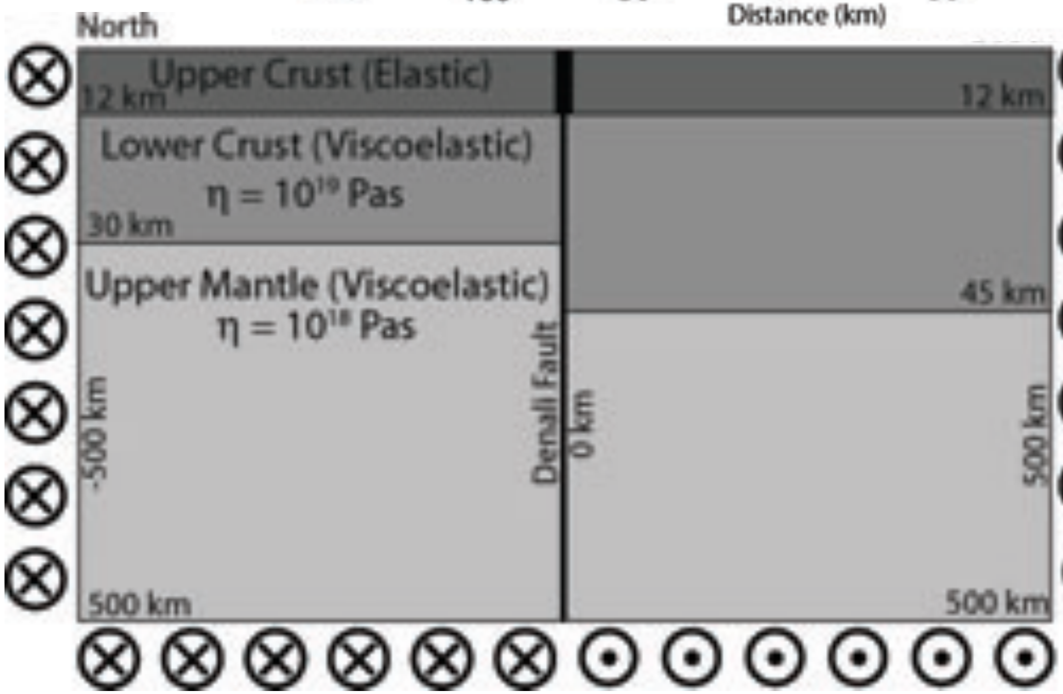


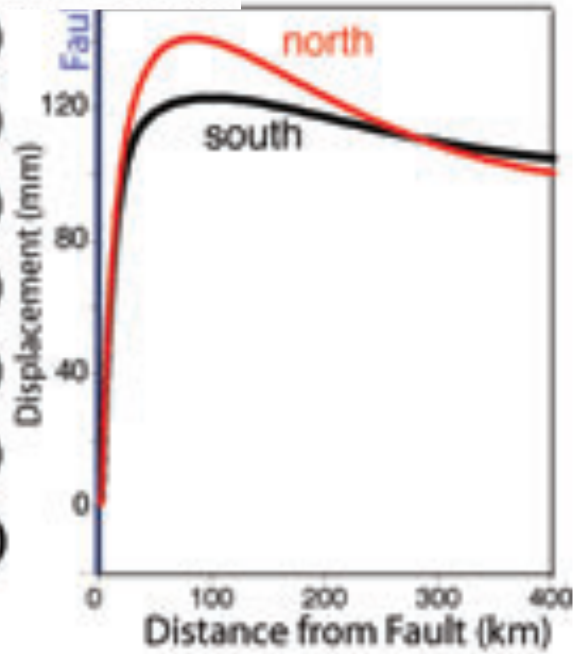
Elastic lid does not necessarily corresponds to seismogenic layer.

In this case, the lower crust is elastic, or at least has a higher viscosity than the upper mantle.

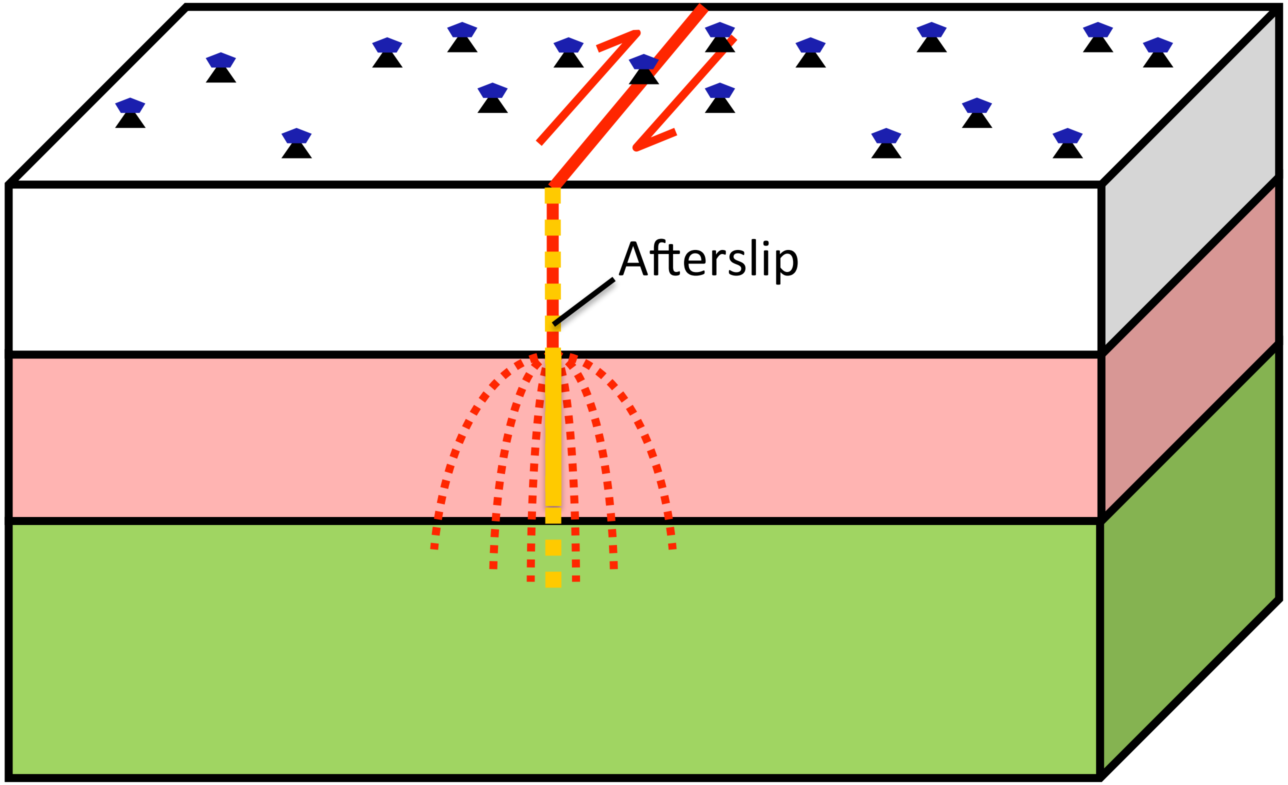
Origin of Asymmetry



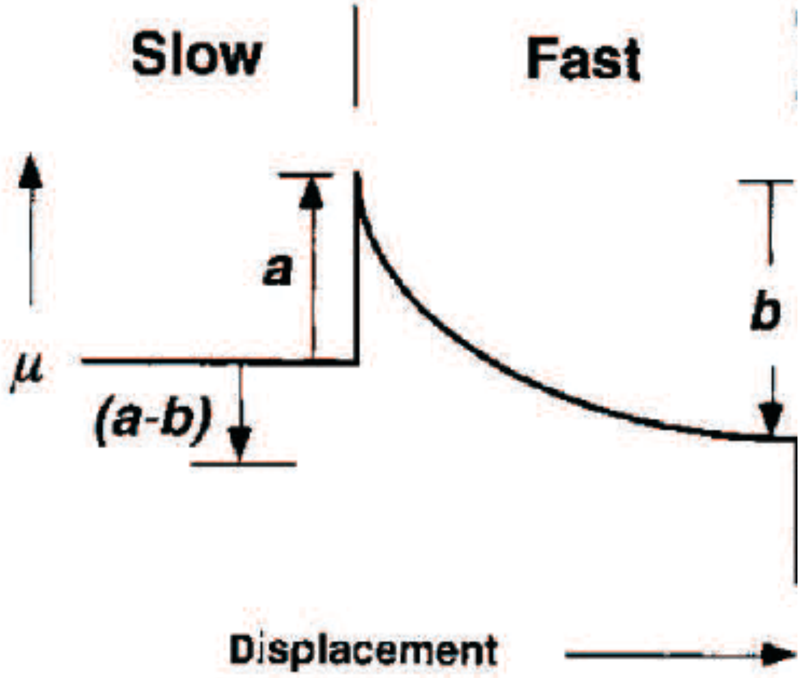




Postseismic Deformation



Rate and State Frictional Behaviour

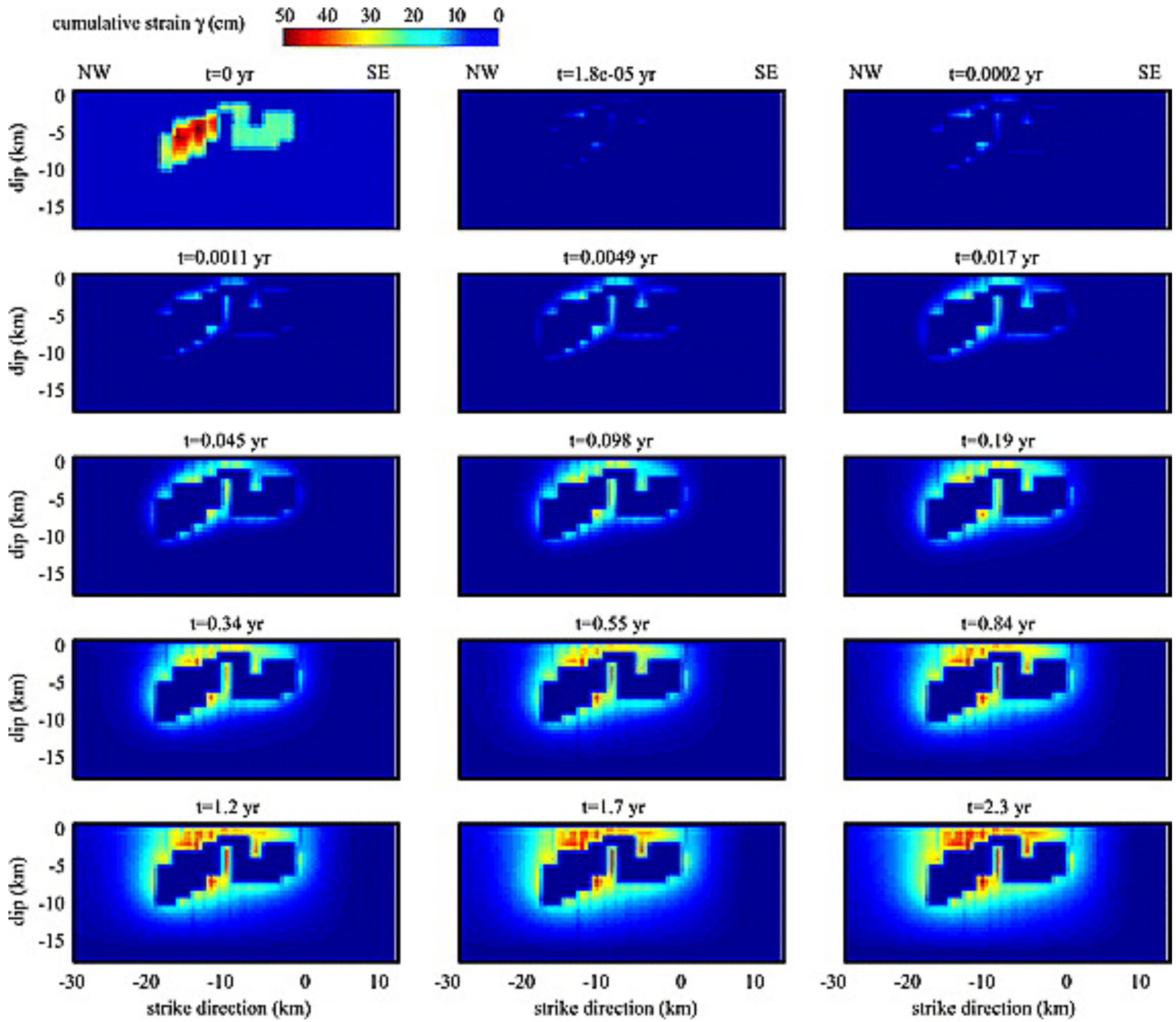


$$\tau = \underbrace{\left[\mu_0 + a \ln\left(\frac{V}{V_0}\right) + b \ln\left(\frac{V_0\,\theta}{L}\right)\right]}_{\mu} \bar{\sigma}$$

Upper crust (T<300°C)
(a-b) < 0
⇒ velocity weakening
⇒ stick-slip behaviour
⇒ earthquakes in seismogenic zone.

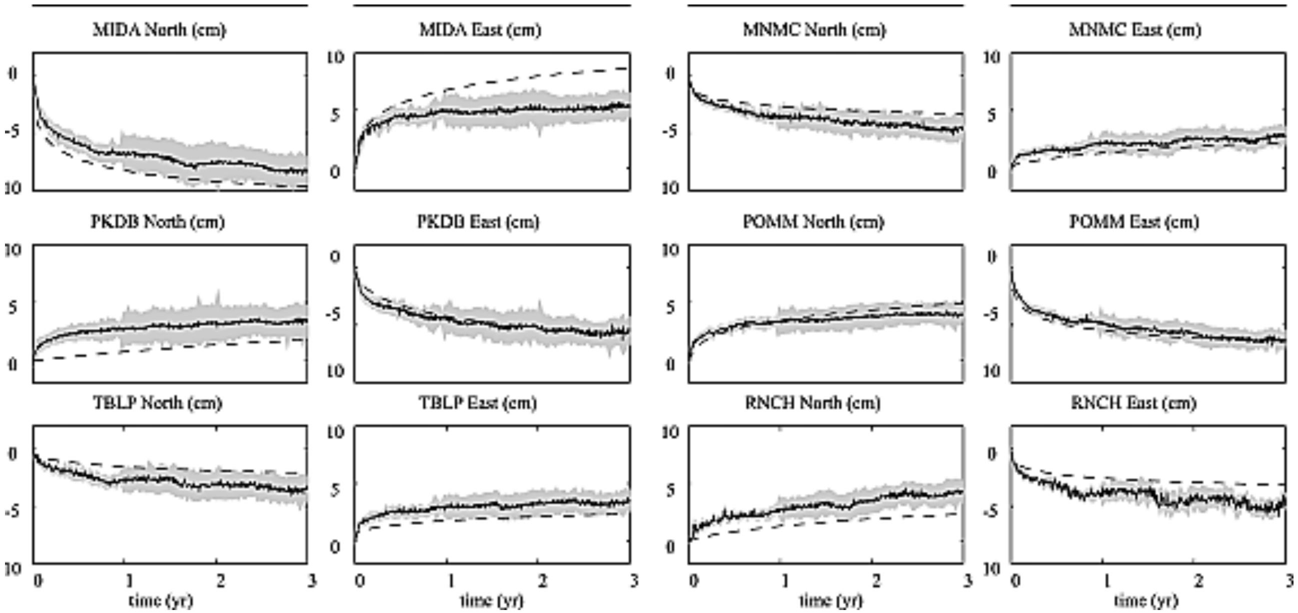
Lower crust/upper mantle
(a-b) > 0
⇒ velocity strengthening
⇒ stable-sliding
⇒ aseismic creep in ductile shear zones.

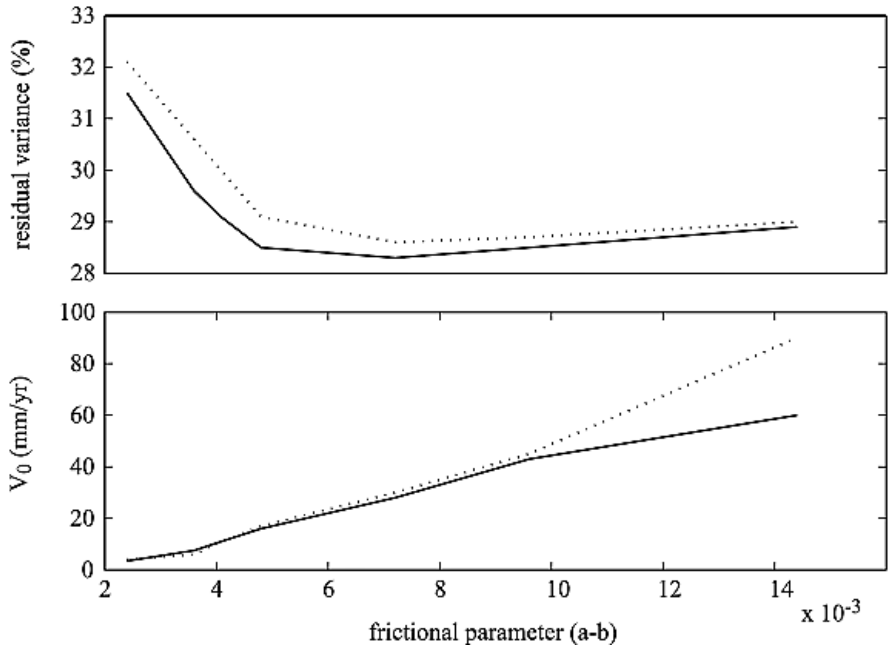
Parkfield Earthquake



Coseismic Model drives afterslip using rate and state friction law.
Solve for (a-b)

Barbot et al, 2009

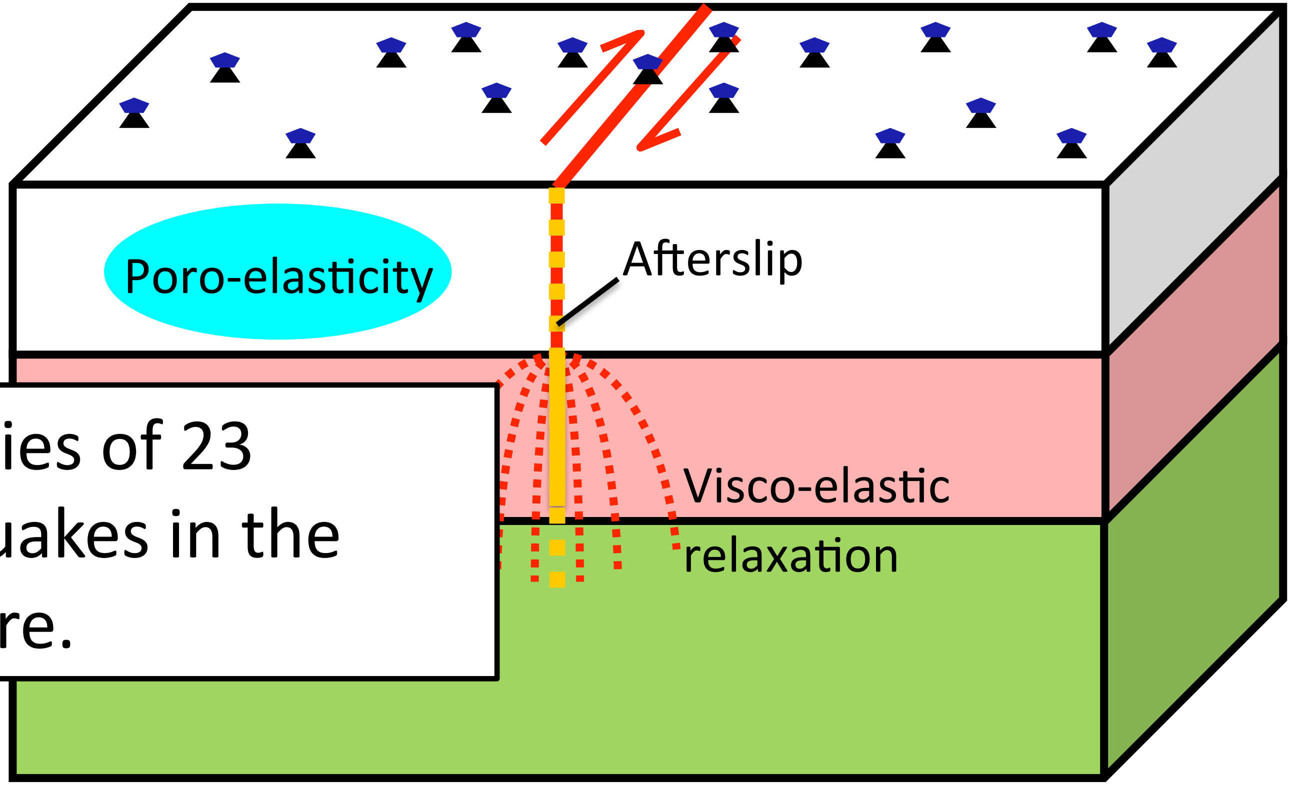




Use observations to solve for appropriate (a-b) and V_o parameters

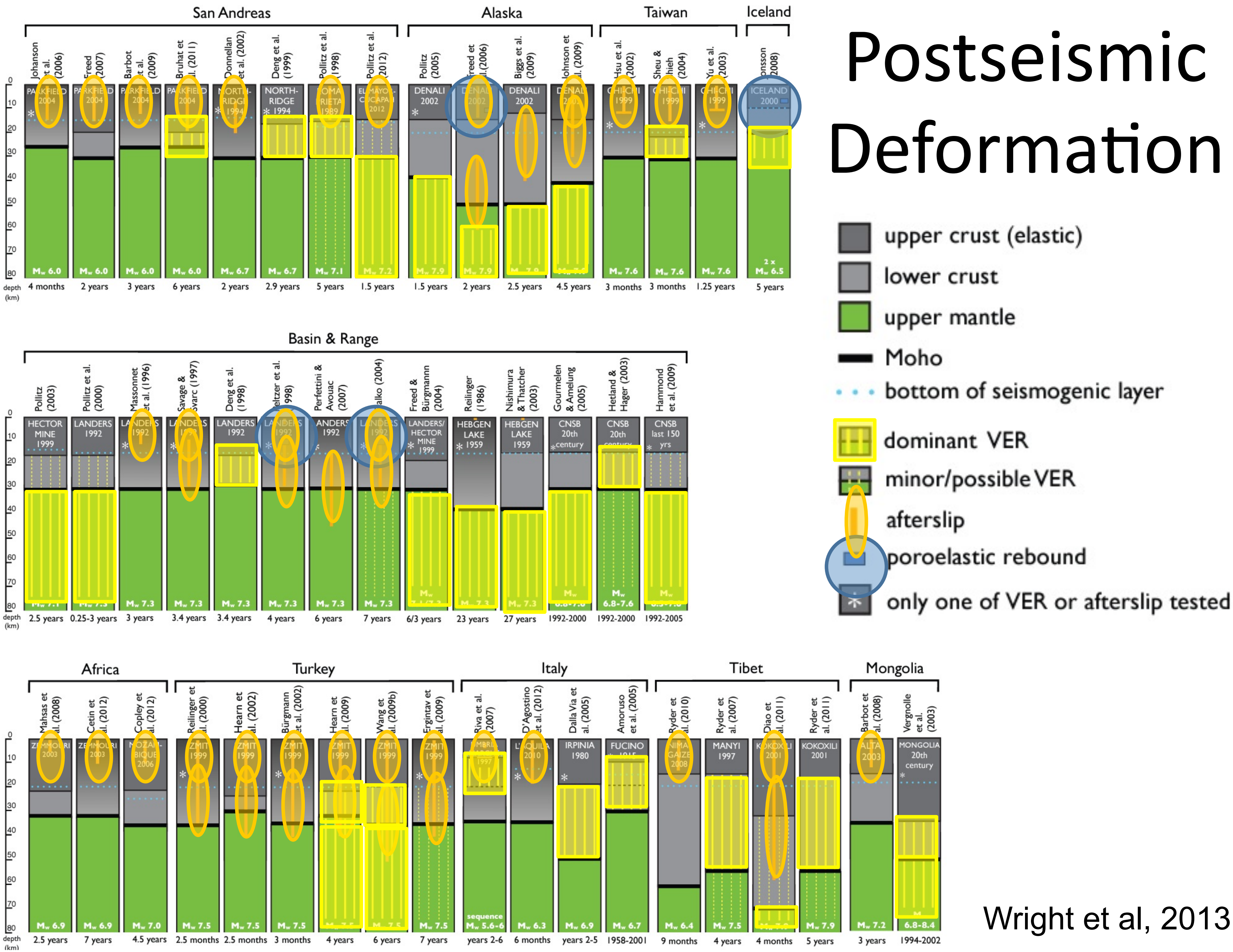
Barbot et al, 2009

Postseismic Deformation



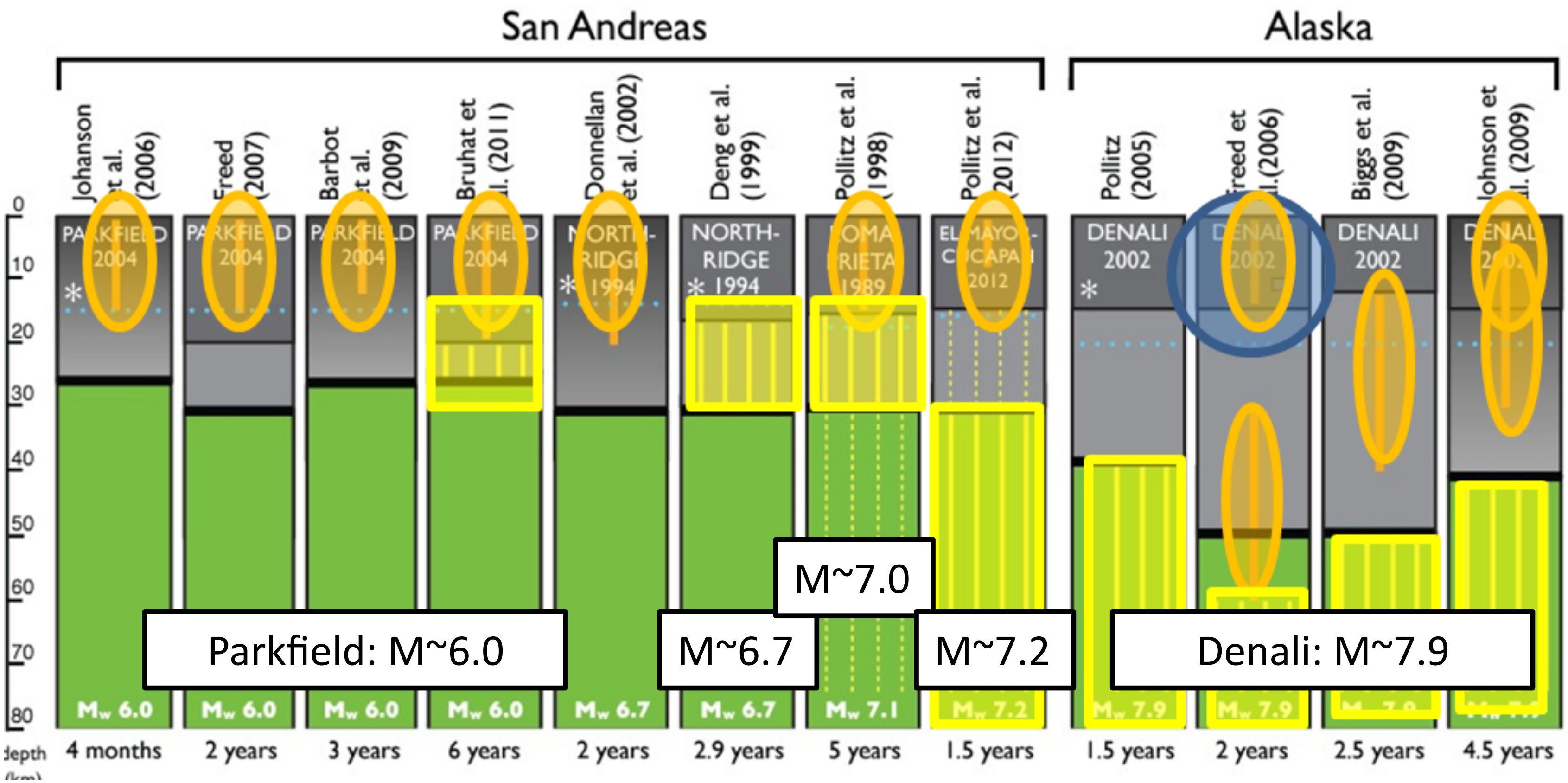
49 studies of 23 earthquakes in the literature.

Postseismic Deformation



Wright et al, 2013

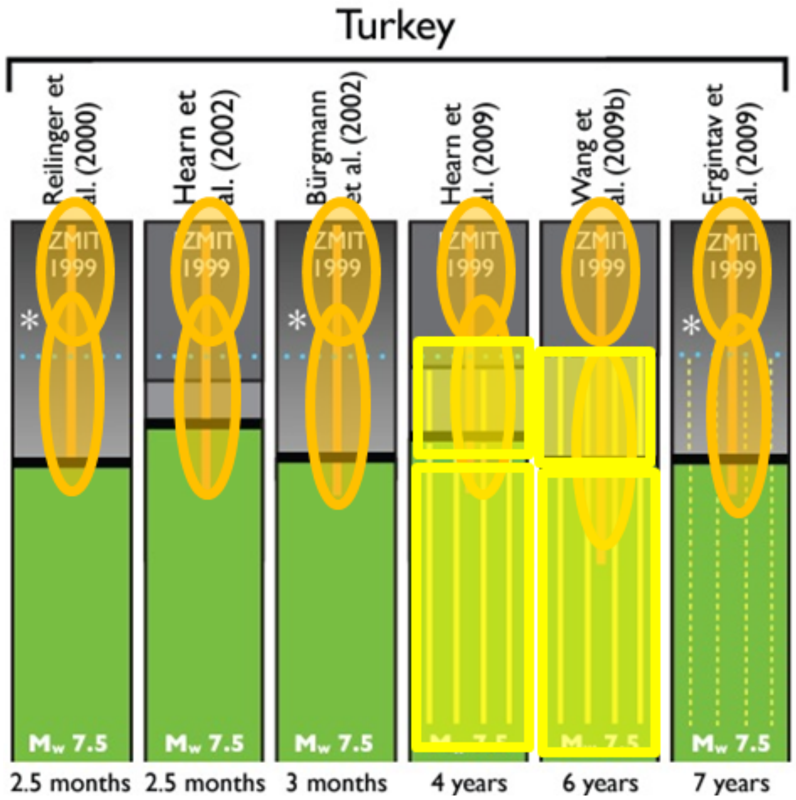
Postseismic Deformation



- Inferred mechanisms vary as a function of the size of the earthquake: small earthquakes do not cause deep flow

Wright et al, 2013

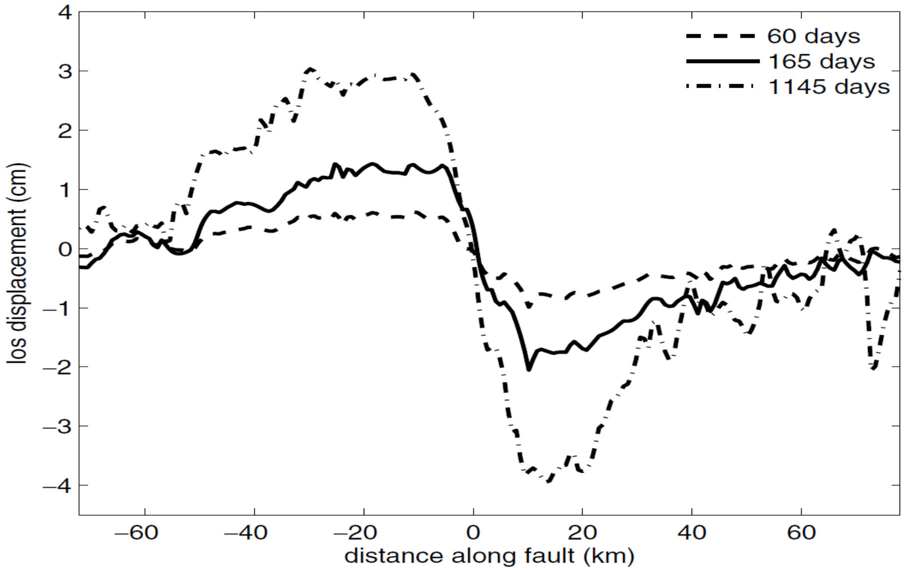
Postseismic Deformation

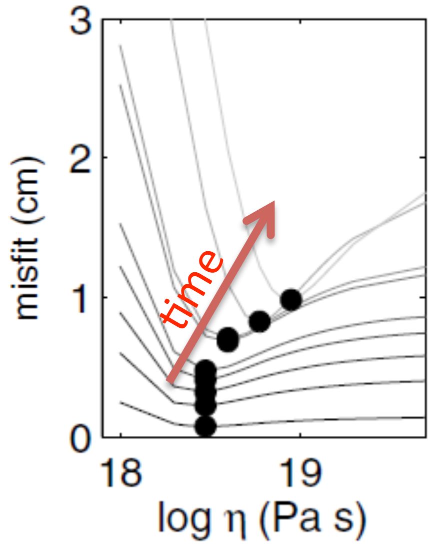


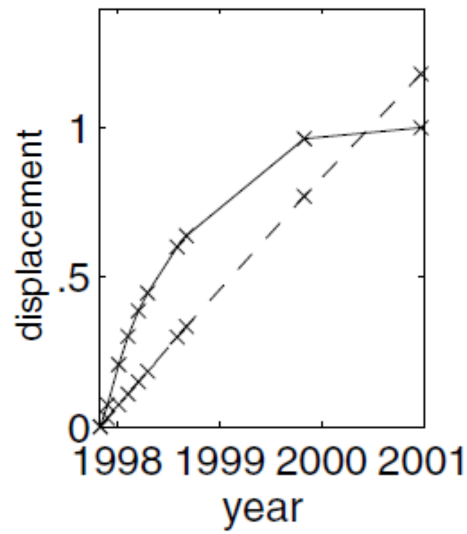
- Inferred mechanisms and timescales vary as a function of the time period of observation

Wright et al, 2013

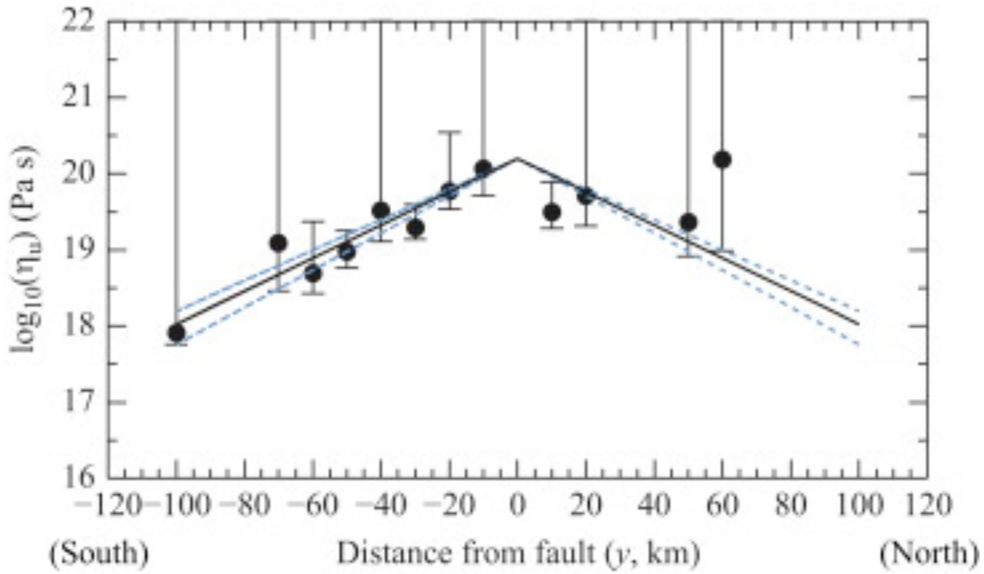
• Inferred mechanisms and timescales vary as a function of the time period of observation





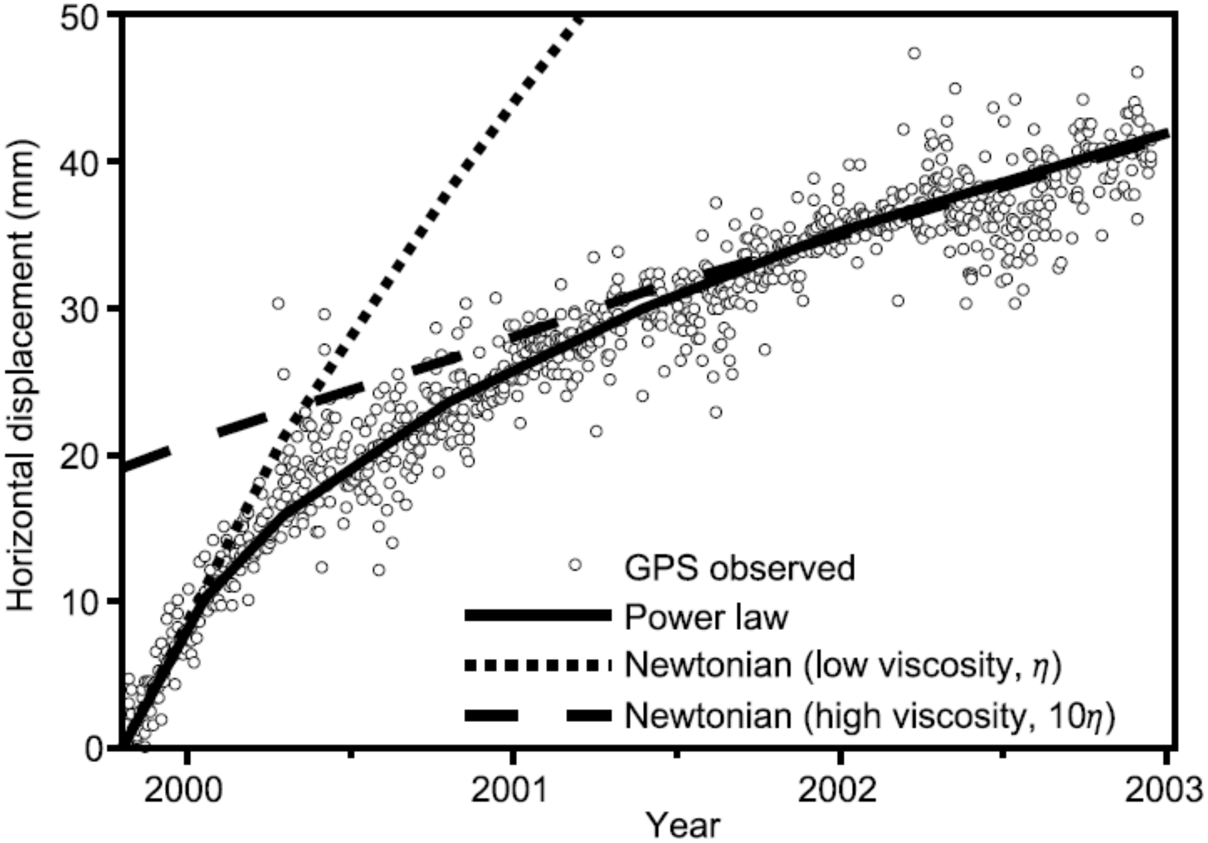


• and as a function of the distance of observation



Manyi Earthquake: Ryder et al, GJI 2007; Yamasaki and Houseman, 2012

Power-law



n~3-3.5: Freed and Burgmann, Nature 2004

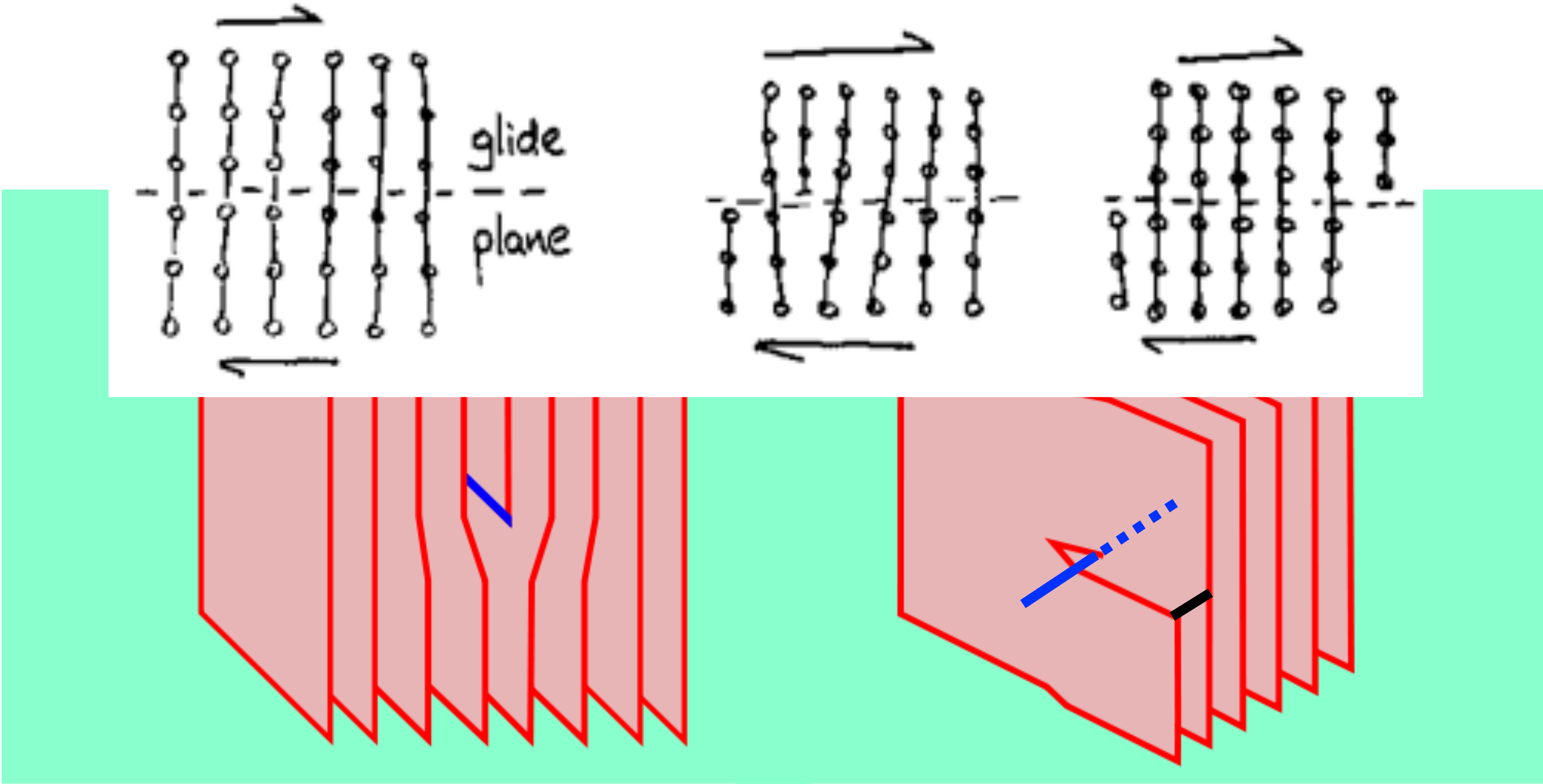
Laboratory experiments

Deviatoric strain rate, Deviatoric stress, Grain size, Water fugacity, Activation energy, Pressure, Activation volume, Temperature

$$\dot{\varepsilon} = A\sigma^{n} d^{-m} f_{H_2O}^{r} \exp\left(-\frac{Q + pV}{RT}\right)$$

$$\eta_{eff} = \frac{\sigma}{\dot{\varepsilon}} = A^{-1}\sigma^{1-n} d^{m} f_{H_2O}^{-r} \exp\left(\frac{Q + pV}{RT}\right)$$

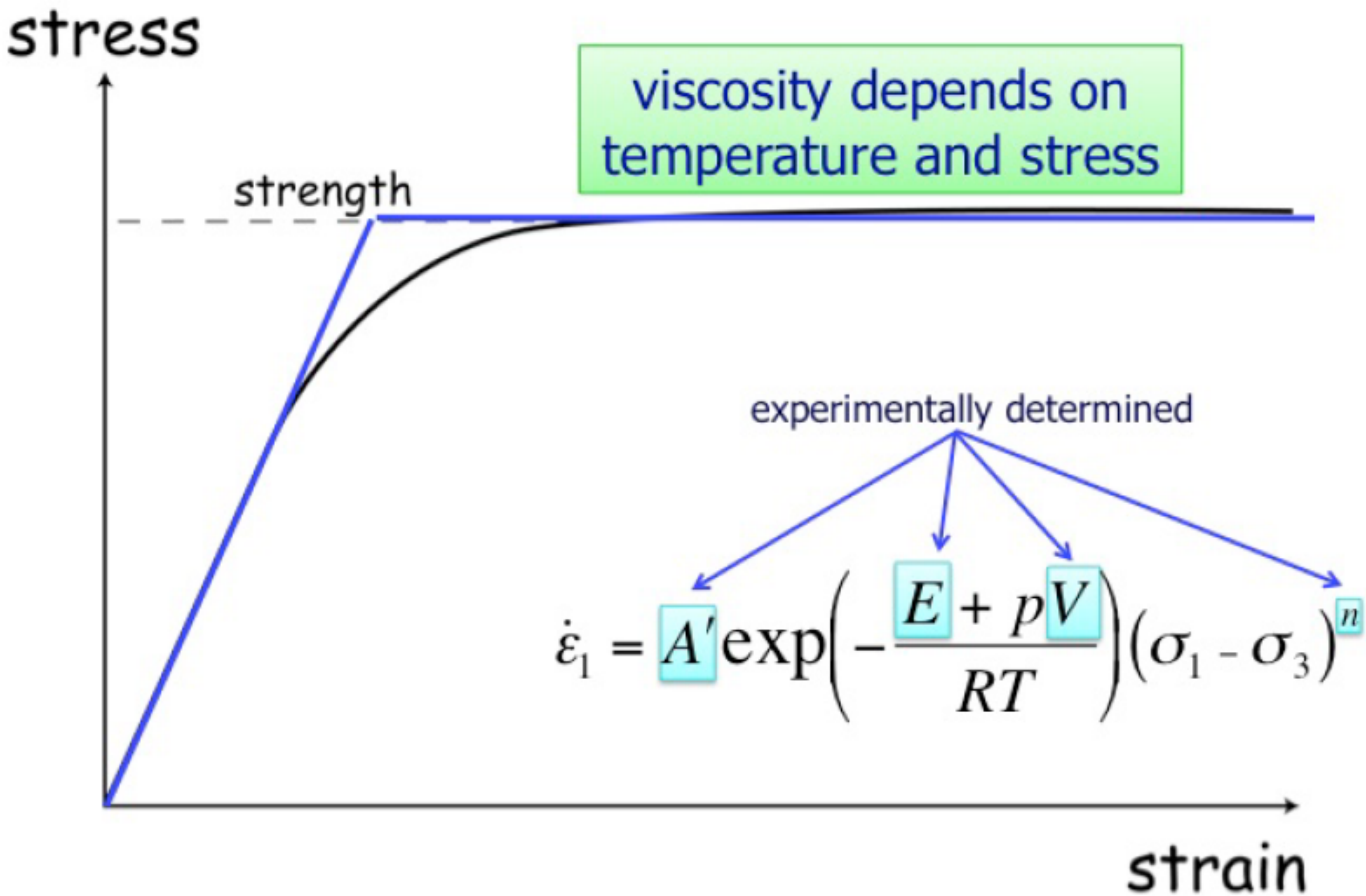
"Effective viscosity"



- Occurs when dislocation lines move through the crystal lattice
- Plane along which the movement takes place is called a glide plane
- Strain rate is dependent on $(\text{stress})^n$, hence sometimes called “power-law creep”

[Cartoon from http://ijolite.geology.uiuc.edu/07fallclass/geo411/Ductile/ductile.html]

Viscous flow by *power-law creep*



Diffusion Creep

- Point defects come in three basic forms:

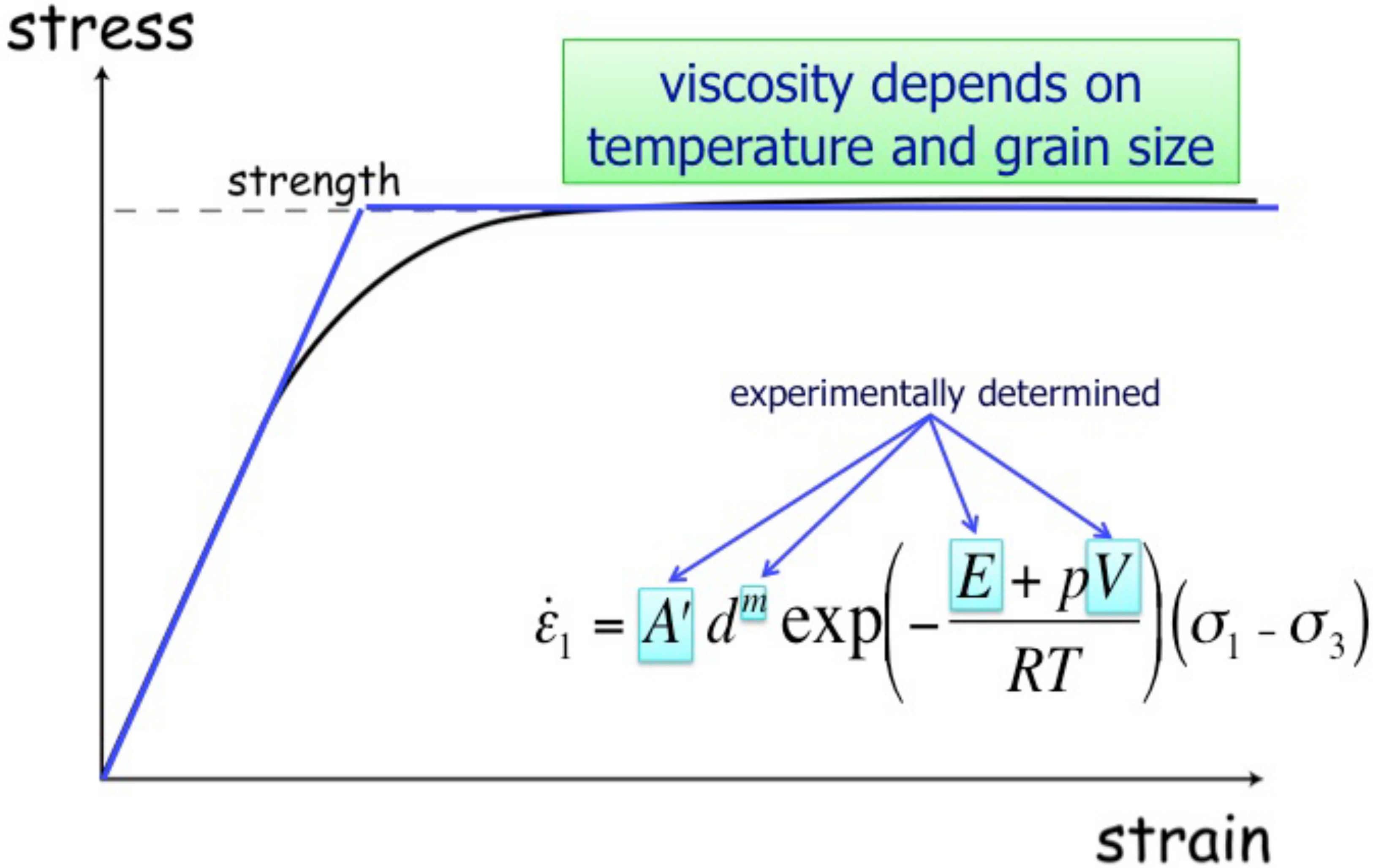
  (i) Vacancies (where an atom is missing from the lattice, leaving a hole)

  (ii) Interstitial defects (where an extra atom is inserted into the lattice)

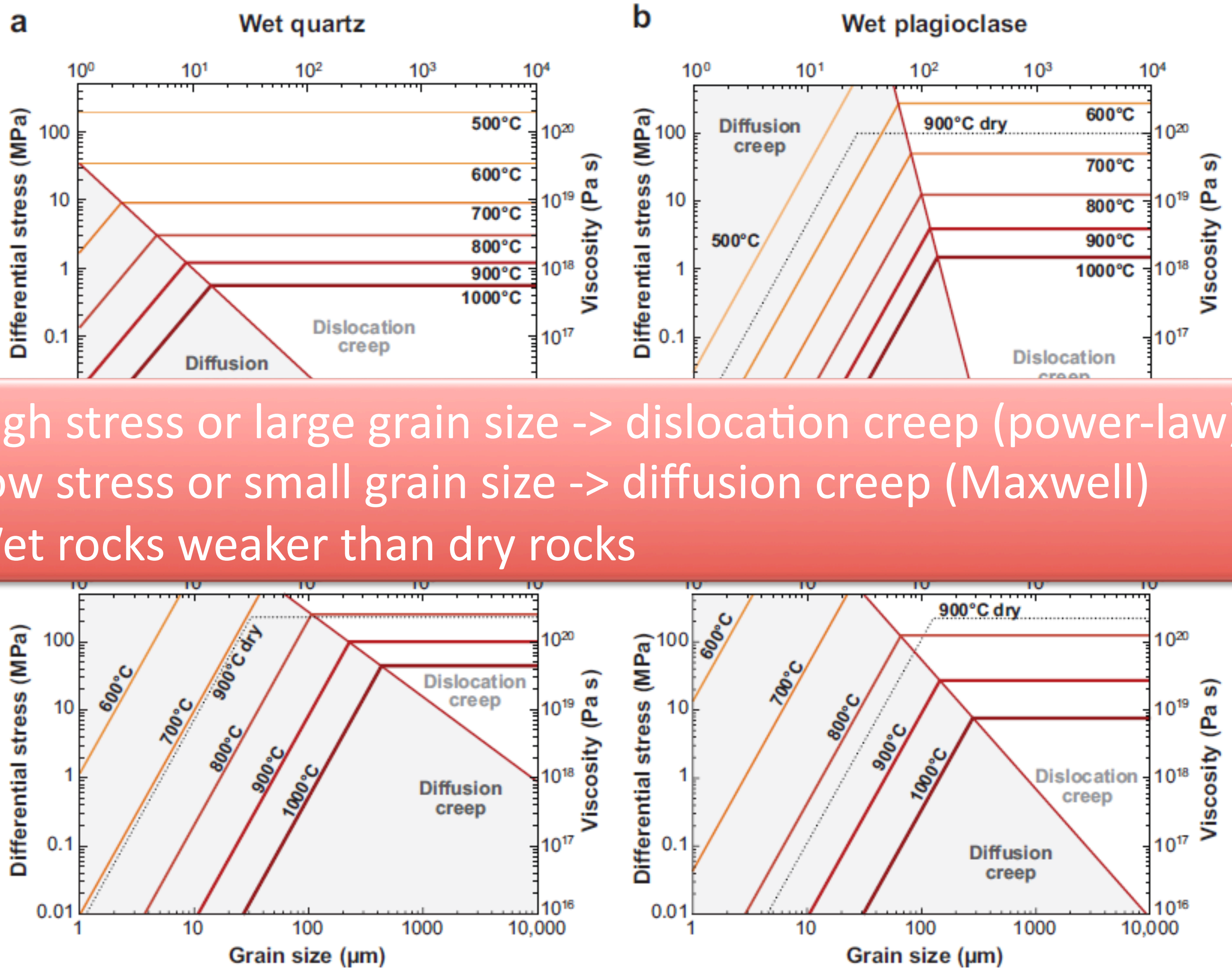
  (iii) Substitutional defects (where a different atom replaces what should be there, inducing strain in the crystal lattice)
- Defects move through crystal by diffusion – thermally activated process.
- Linearly dependent on stress, but grain size is important

[Cartoon from http://ijolite.geology.uiuc.edu/07fallclass/geo411/Ductile/ductile.html]

Viscous flow by *diffusion creep*



$$\dot{\varepsilon}_1 = A' d^m \exp\left(-\frac{E + pV}{RT}\right)(\sigma_1 - \sigma_3)$$



High stress or large grain size -> dislocation creep (power-law)

Low stress or small grain size -> diffusion creep (Maxwell)

Wet rocks weaker than dry rocks

Strain rates 10^{-12} s^{-1}; Burgmann and Dresen, Ann Rev 2008

Spatial variations

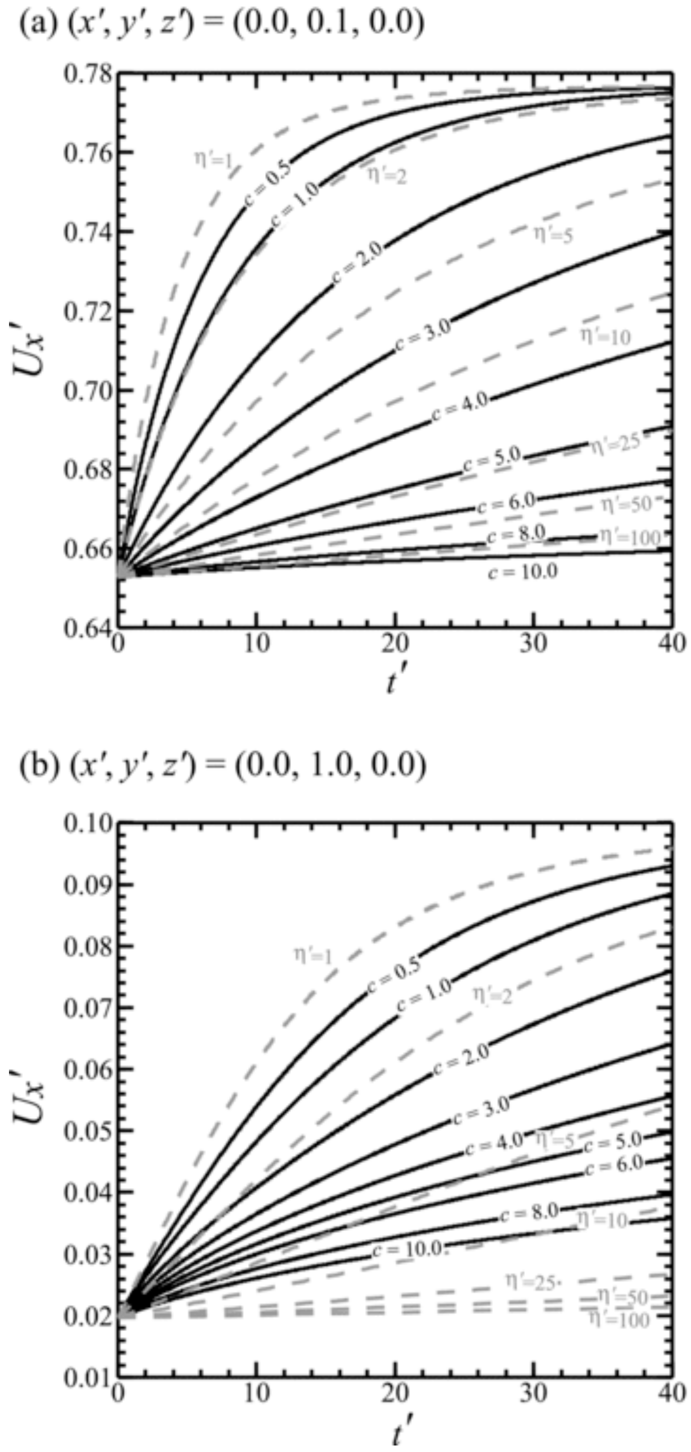
Another way to achieve multiple timescales is with spatially different material properties

Lateral Variations in Properties:

- Shear heating
- ‘Weak zone’ (e.g. smaller grain size).

Vertical Variations in Properties:

- Temperature dependence with depth => vertical viscosity gradient



Yamasaki and Houseman, 2012

Summary of Observations

Transient deformation on timescales of years to decades occurs after most earthquakes > M6

Four primary mechanisms: 1) poroelastic rebound, 2) afterslip, 3) lower crustal viscoelasticity, 4) upper mantle flow viscoelasticity operate under different circumstances.

Maxwell viscoelasticity of a single layer cannot match postseismic observations at all time periods. Requires, power-law rheology, responses of multiple layers, depth-dependence or lateral variations.