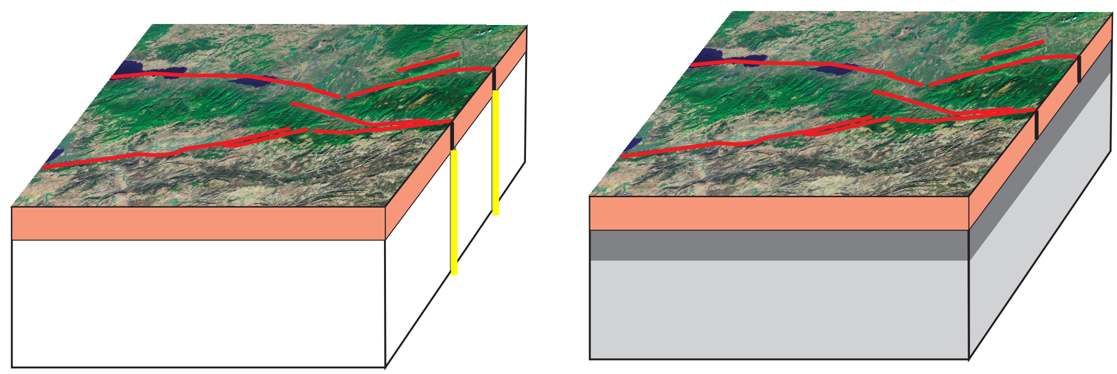
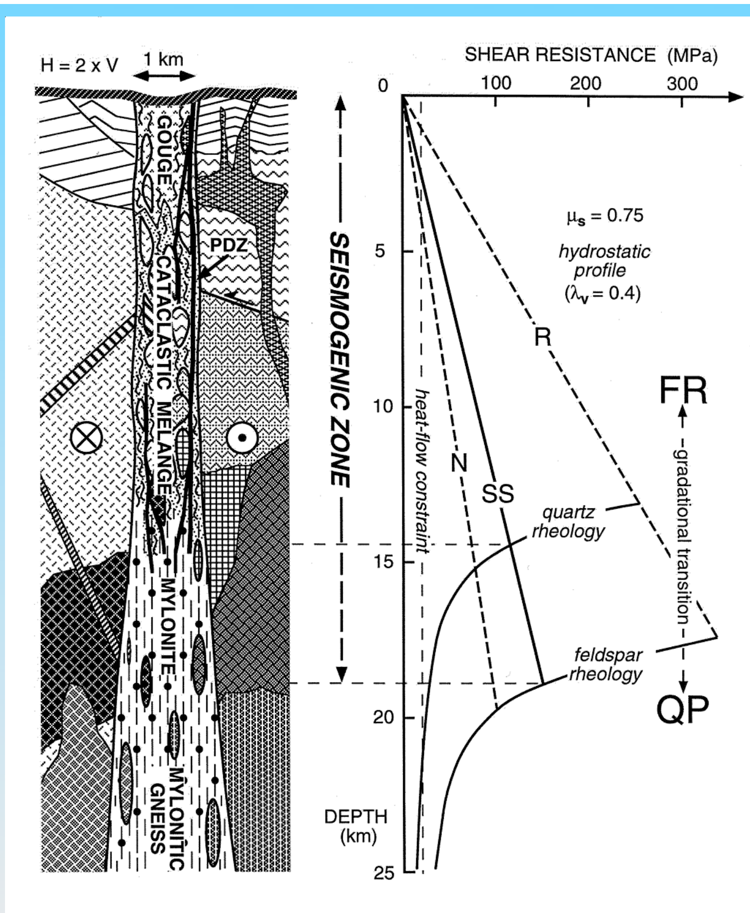
Below the seismogenic zone, how does relative motion of plates occur?



Recall Savage and Burford (1973) "arctangent" elastic dislocation model

Nur and Mavko viscoelastic model (postseismic only)
Savage and Prescott 1978
(we did not get to this)



Fault zone anatomy: rocks and rheology vs. depth

Shear stress versus depth from Byerlee friction and Anderson fault theory

Effective normal stress

Brittle vs ductile rheology

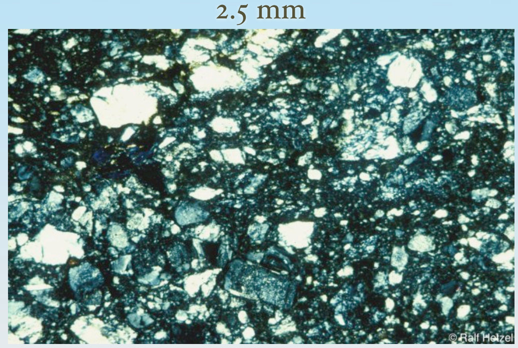
Sibson 2002

Shallow fault rocks: Gouge and cataclasite

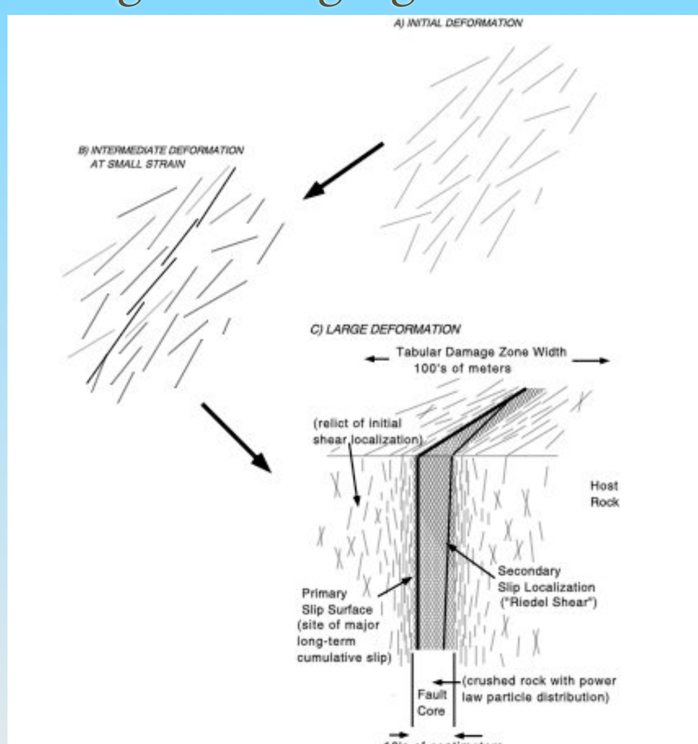
http://virtual.yosemite.cc.ca.us/ghayes/images/Dsc00086_Closeup_of_fault_gouge.jpg

http://earth.boisestate.edu/home/cjnorth/more/photogallery.html

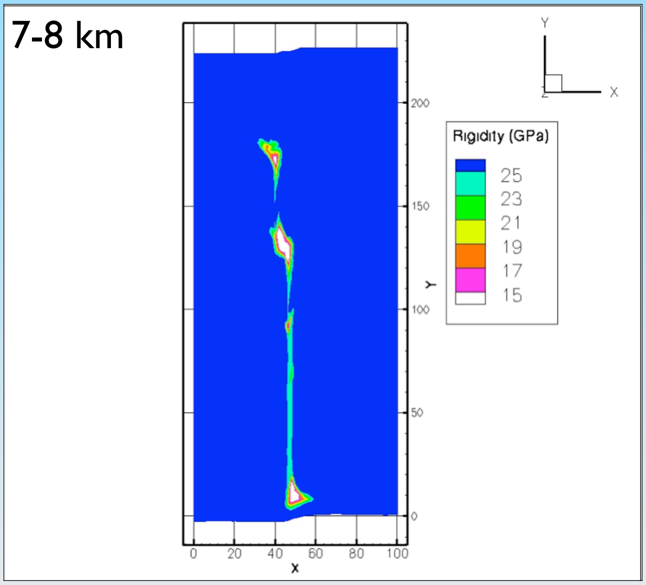
2.5 mm



damage zone > gouge zone > core



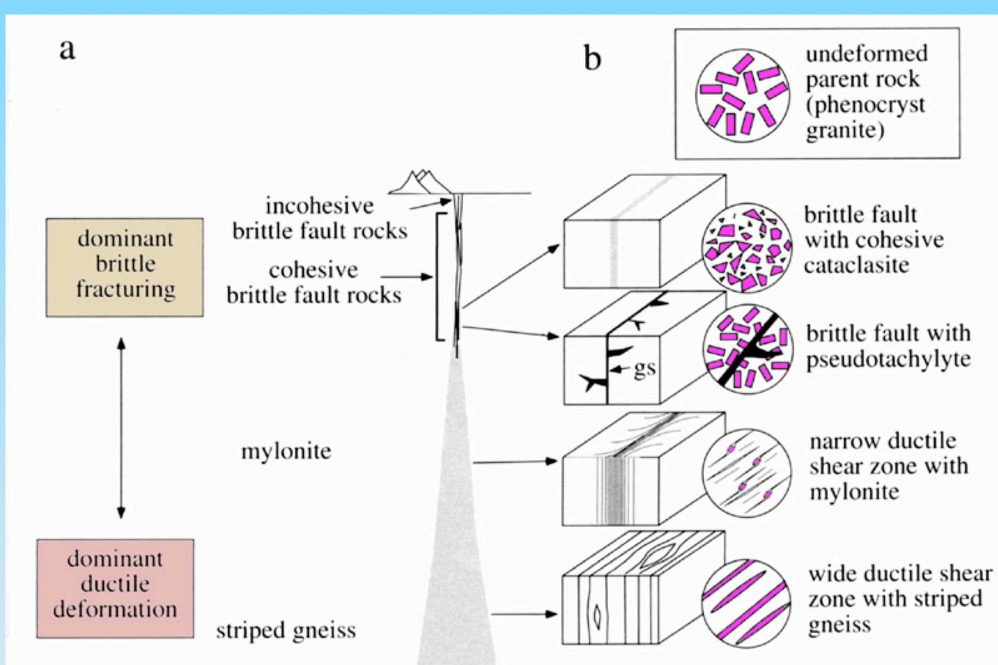
Yaron Finzi's estimates of shear modulus G in an evolving fault zone: shallow damage zone gets narrow below a few km



Damage rheology: elastic strength degraded by microcracks, strain dependent, localization results

Lyakhovsky et al. damage code, 1997 and subsequent references

Stepovers look permanently weak. (Is there observational evidence?)

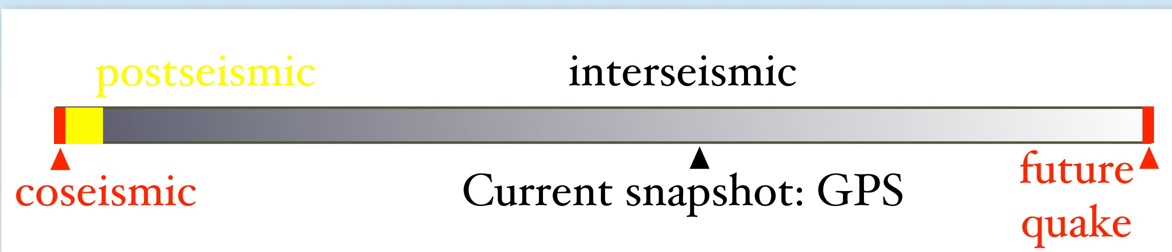


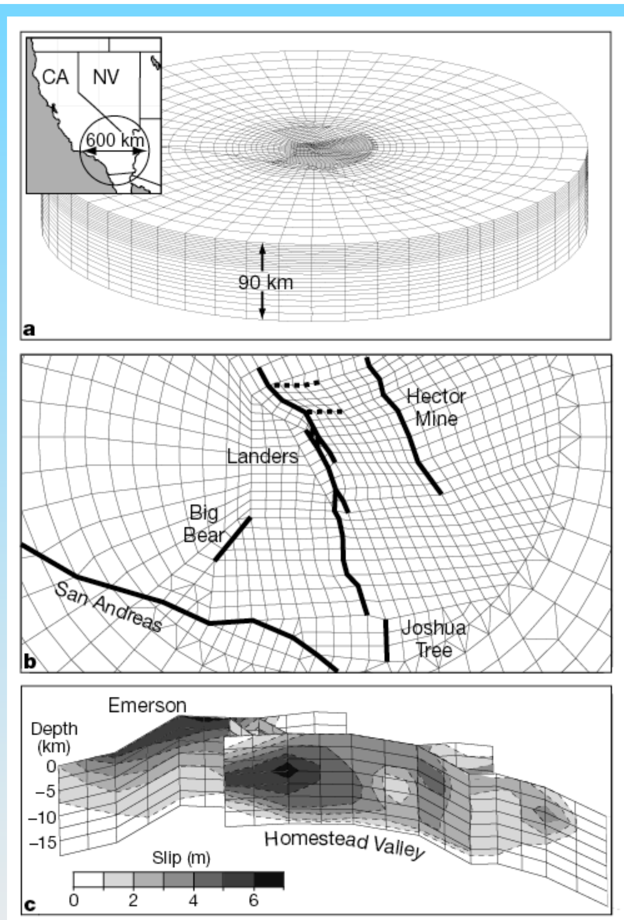
Darrell Henry, 2006

What do *postseismic* and *interseismic* surface deformation tell us about how faults work at depth?

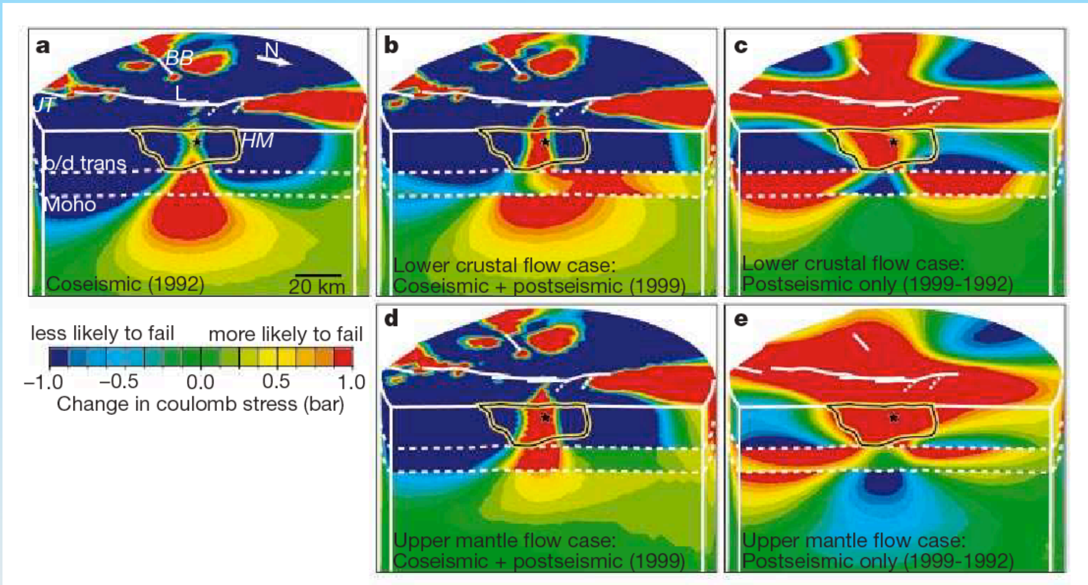
Postseismic deformation: *perturbation* to GPS velocities caused by large earthquakes

Interseismic deformation: GPS velocities around fault between quakes (not immediately after one). Usually, this is what GPS shows.





Freed and Lin, 2000 (I think)



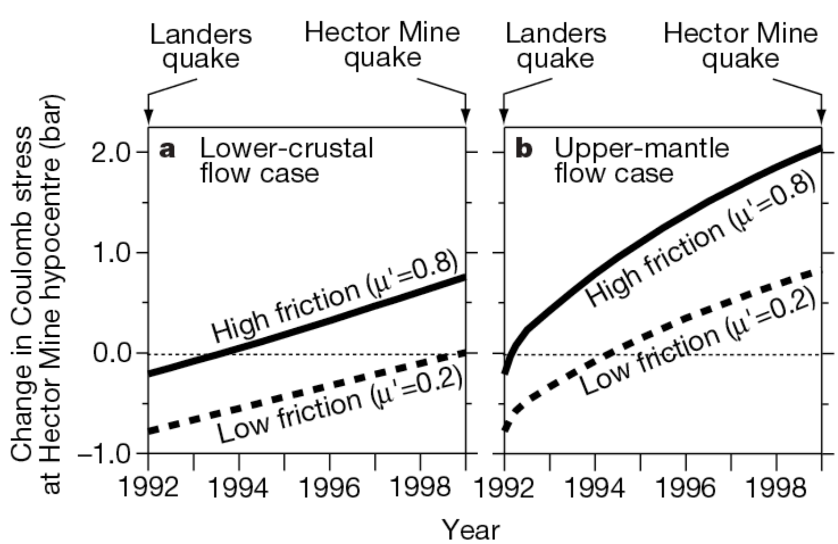


Figure 4 Calculated Coulomb stress changes at the Hector Mine hypocentre as a function of year and assumed friction. **a**, For a model that assumes viscous flow occurs only in the lower crust. **b**, For a model that assumes viscous flow occurs predominantly in the upper mantle. The calculated positive postseismic Coulomb stress changes suggest that the Hector Mine hypocentre was brought closer to failure by post-Landers relaxation processes regardless of the apparent friction or whether viscous flow occurs predominantly in the lower crust or upper mantle.

Figure 1. (*Continued.*)

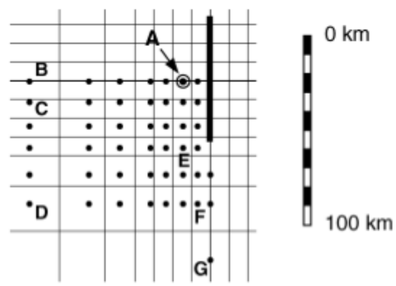


Figure 2. Modelled strike-slip rupture (plan view, one quadrant) and locations of hypothetical GPS stations A–G where displacements are calculated.

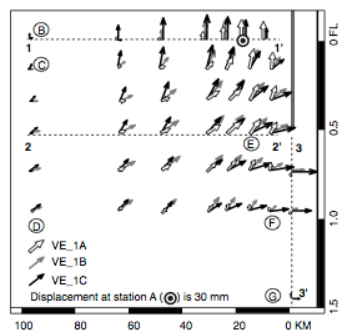
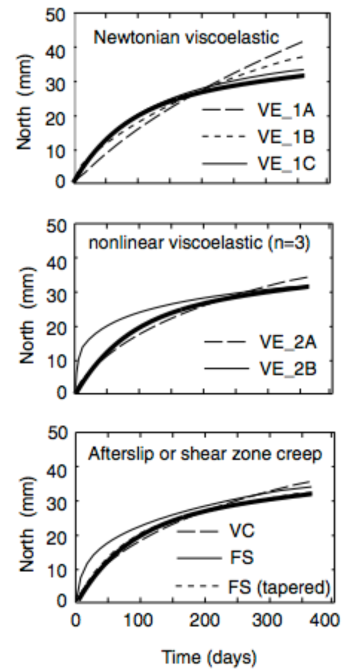
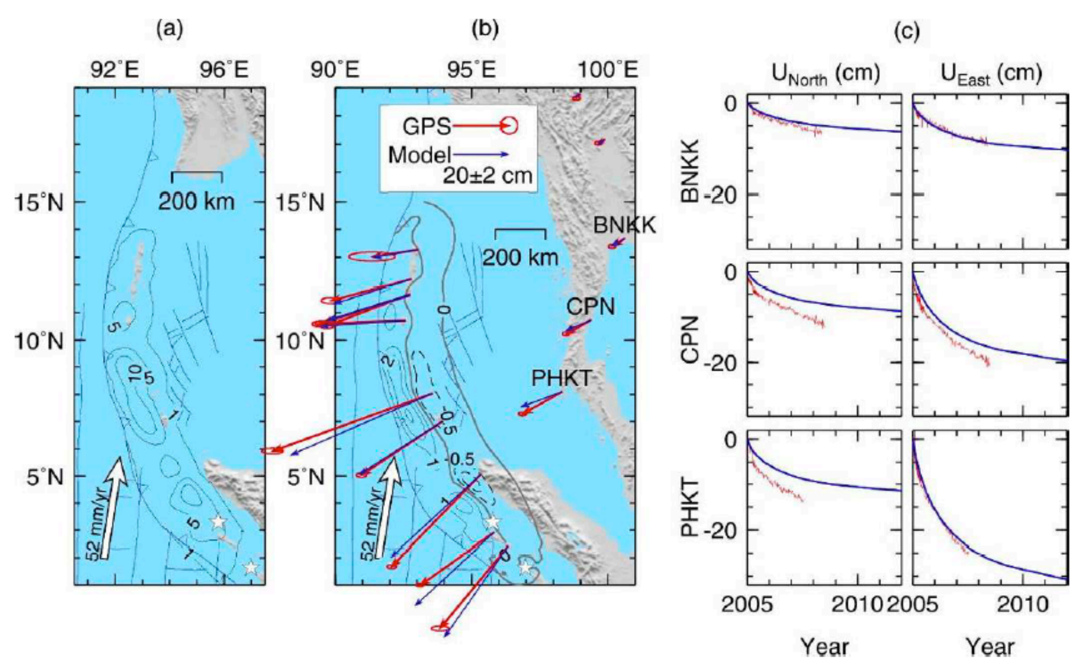


Figure 8. Horizontal displacements 80 d after the hypothetical $M_w = 7.4$ strike-slip earthquake. For this plot, group 1A, 1B and 1C model parameters are chosen to yield a displacement of 30 mm at Station A 80 d after the earthquake. Amplitudes and azimuths at other locations differ significantly for the three models.

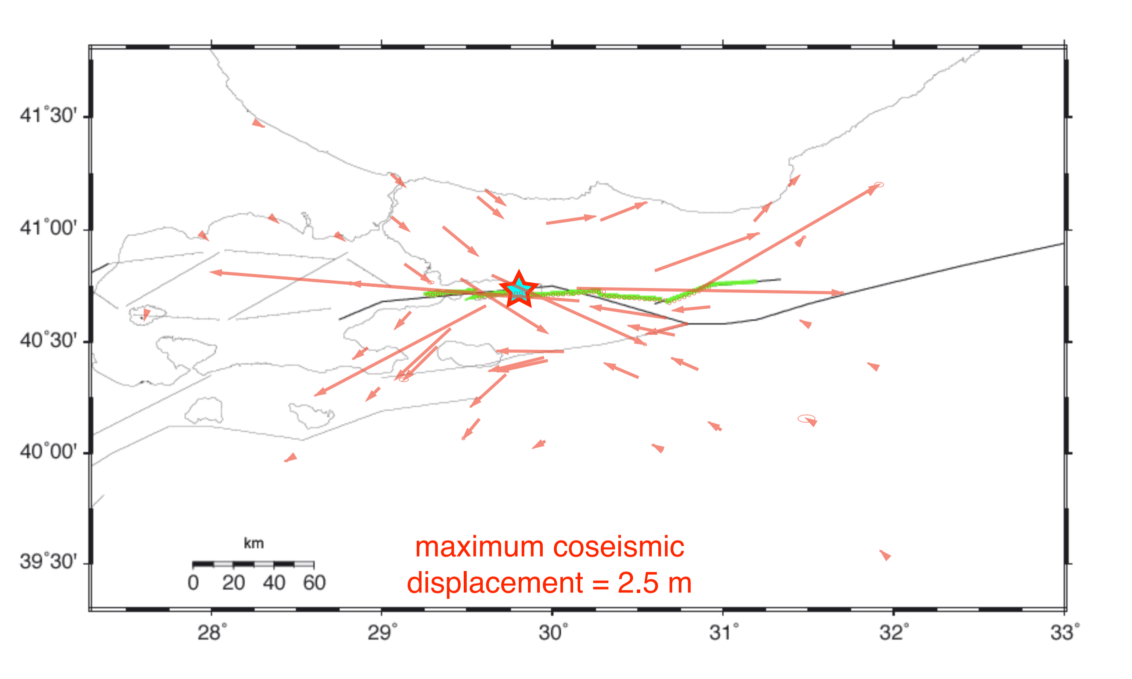
Dynamics of post-seismic deformation 761



Hearn 2003

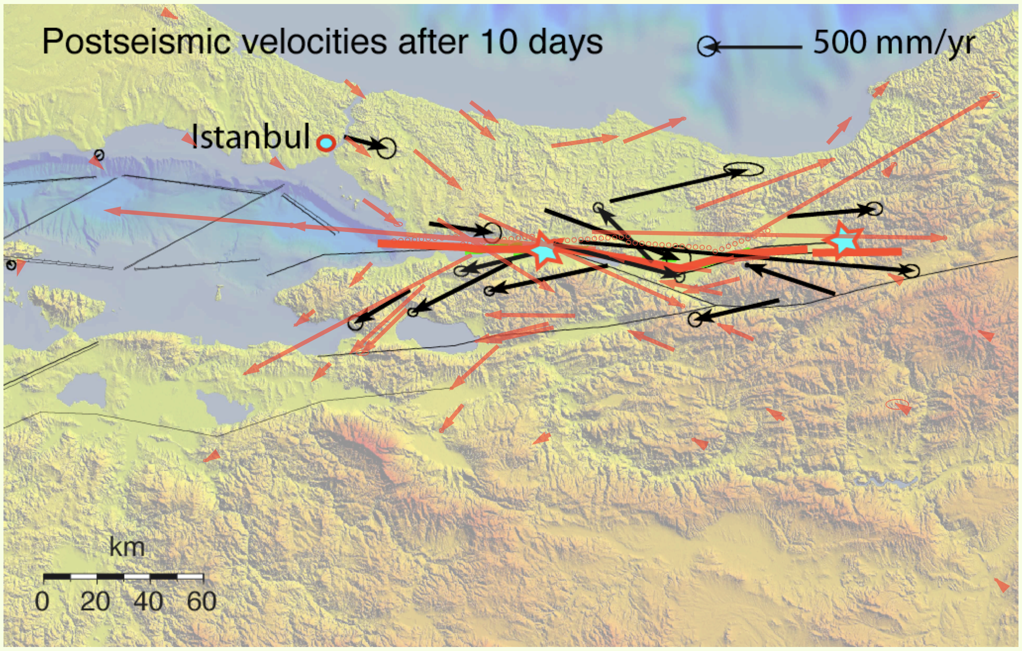


Postseismic deformation example: the 1999 M 7.5 and M 7.1 Izmit-Düzce earthquake sequence in Turkey



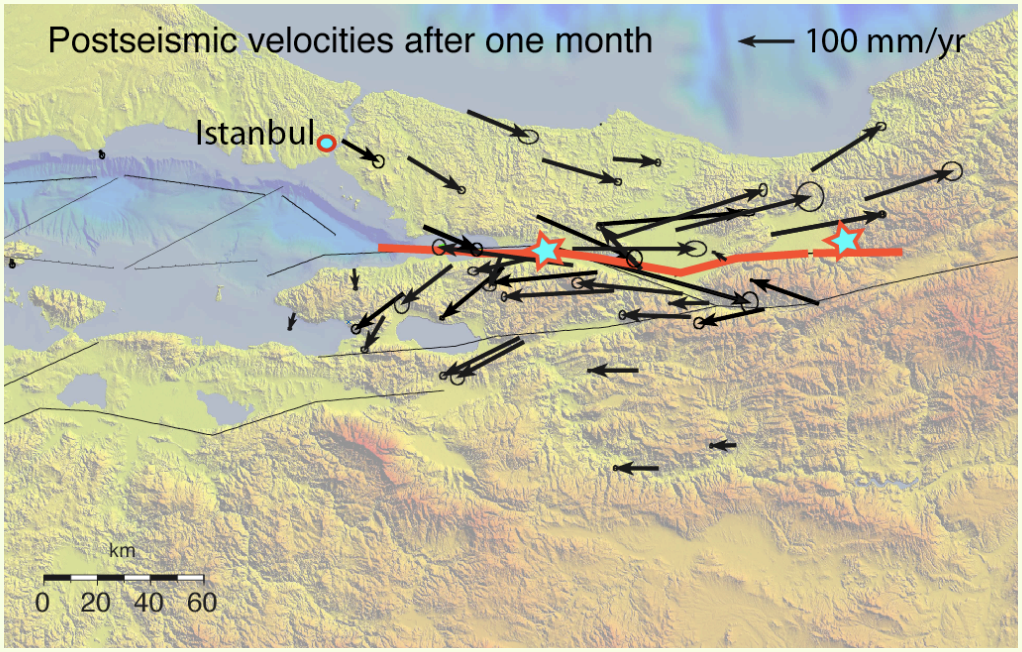
data: Reilinger et al., 2000

Postseismic GPS velocity field



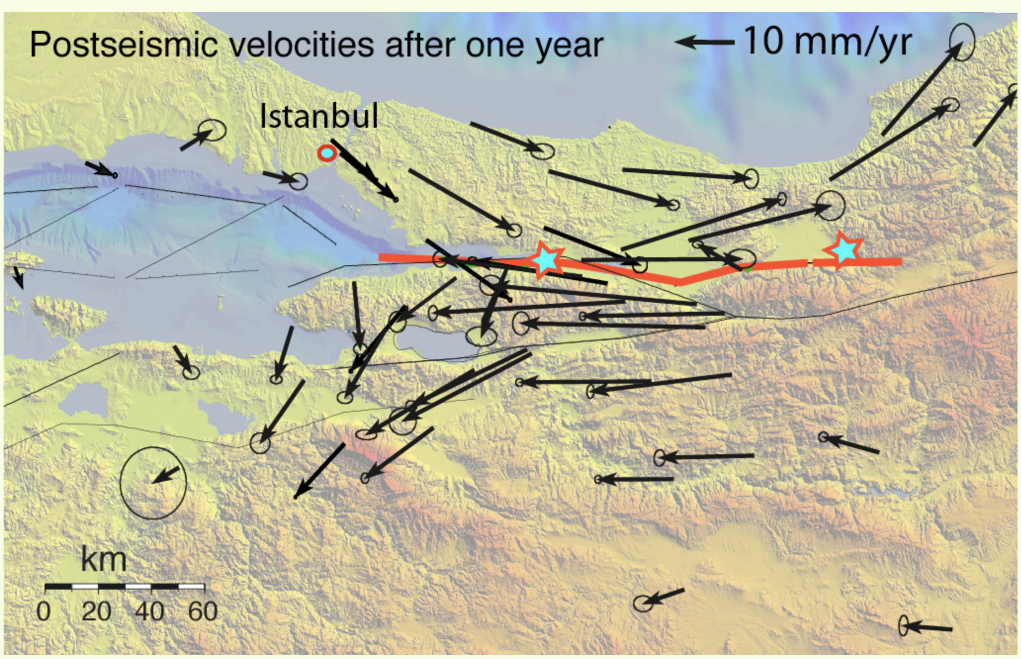
Ergintav et al., 2009

Postseismic GPS velocity field



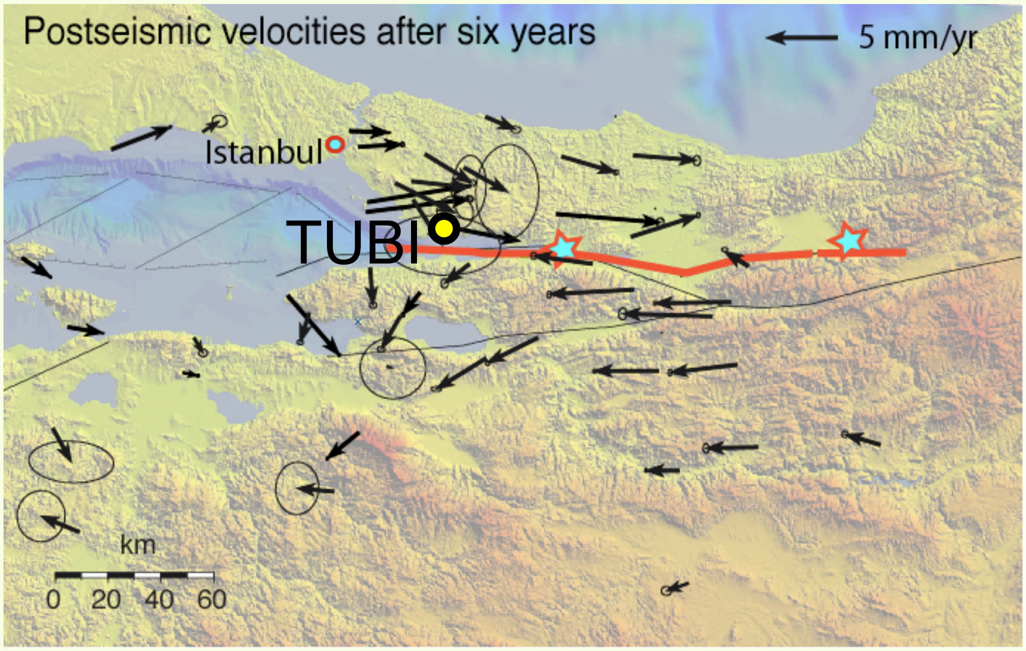
Ergintav et al., 2009

Postseismic GPS velocity field



Ergintav et al., 2009

Postseismic GPS velocity field



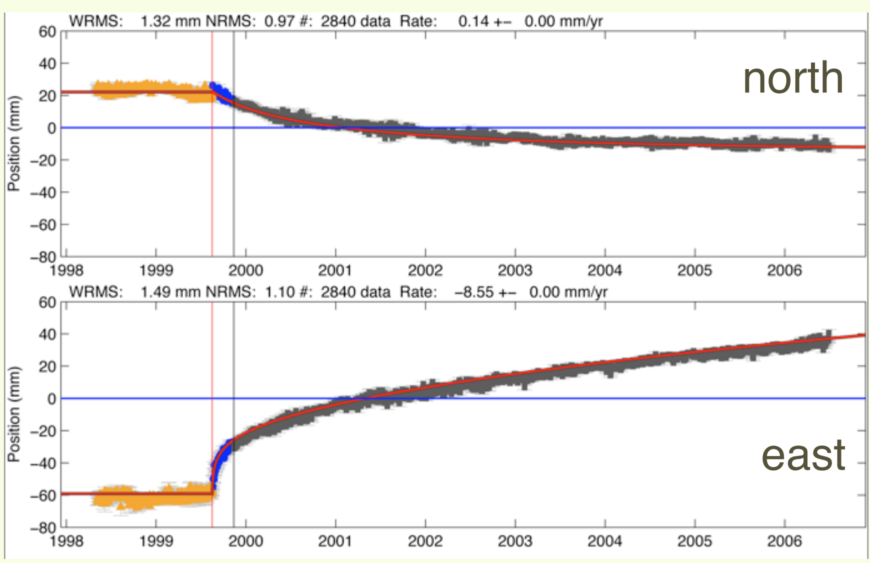
Ergintav et al., 2009

Postseismic motion at GPS site TUBI

moving southeast

(pre-earthquake velocity has been removed)

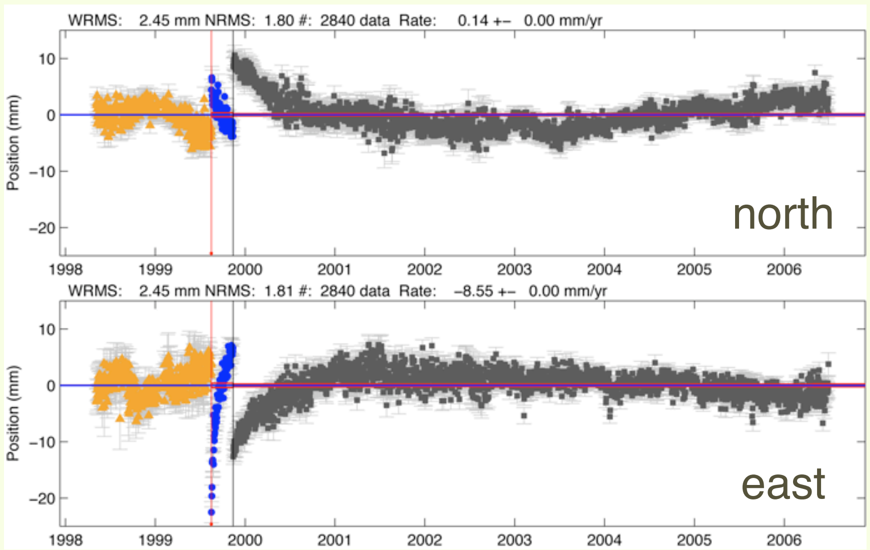
$$A_1 \, ln(1+\frac{t}{\tau_1}) + A_2 \, ln(1+\frac{t}{\tau_2}) + A_3 \, ln(1+\frac{t}{\tau_3})$$



Ergintav et al., 2009

Postseismic deformation at one GPS site

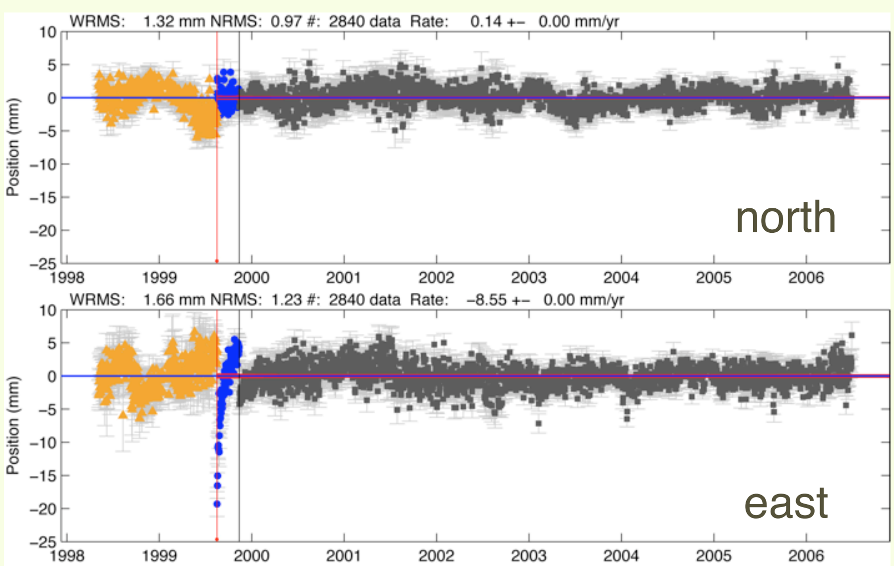
subtract $A_1 \; ln(1 + \frac{t}{\tau_1})$ $\quad \tau_1 = 3500 \; days$



Ergintav et al., 2009

Postseismic deformation at one GPS site

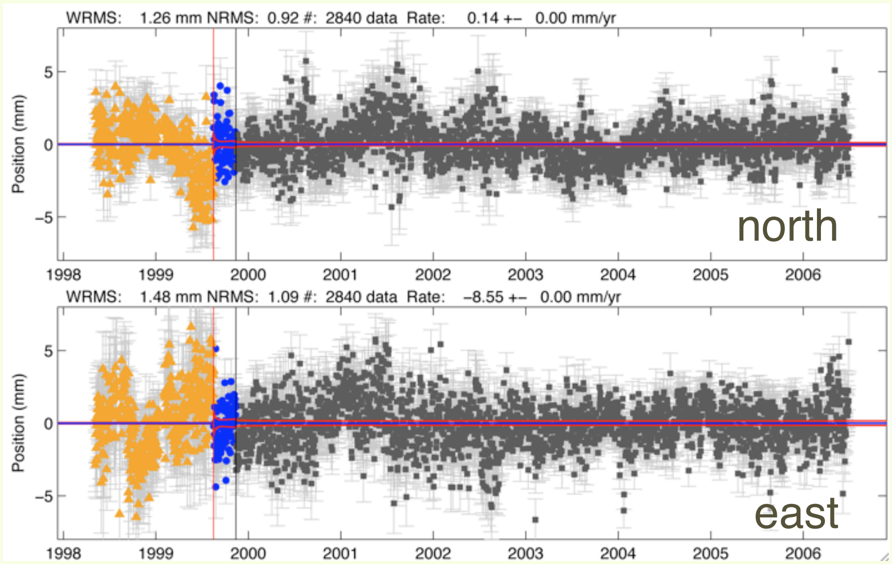
subtract $A_1 \; ln(1 + \frac{t}{\tau_1}) + A_2 \; ln(1 + \frac{t}{\tau_2})$ $\quad \tau_1 = 3500 \; days$, $\tau_2 = 150 \; days$



Ergintav et al., 2009

Postseismic deformation at one GPS site

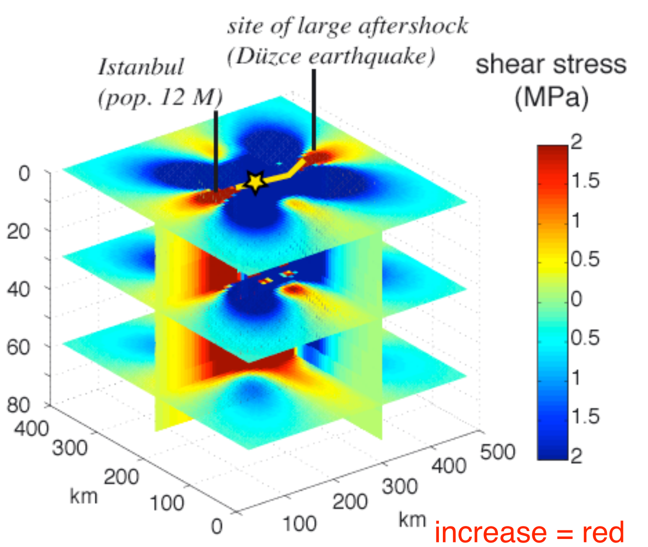
subtract $A_1\ ln(1+\frac{t}{\tau_1}) + A_2\ ln(1+\frac{t}{\tau_2}) + A_3\ ln(1+\frac{t}{\tau_3})$



$\tau_1 = 3500\ days \qquad \tau_2 = 150\ days \qquad \tau_3 = 1\ day$

Ergintav et al., 2009

Postseismic deformation provides clues



A major earthquake changes stress in the lithosphere by a known amount.

Result: accelerated and complicated **post-seismic deformation**, segueing to steadier **interseismic deformation.**

Modeling this deformation can tell us about the rheology and structure of the plate boundary lithosphere.

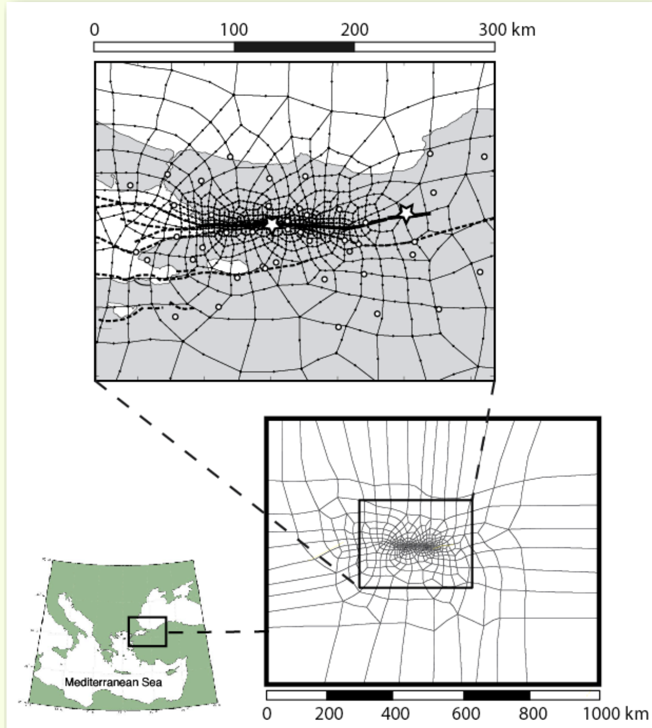
Modeling postseismic deformation

1. Build a numerical representation (FEM) of the lithosphere.

2. Apply earthquake slip. Model mesh side and bottom boundaries are fixed. Optional: pre-earthquake stress field.

3. Adjust the most poorly constrained parameters* until model fits the time-dependent GPS surface deformation data.

* rheology and distribution of viscoelastic material and fault zone rheology



Hearn et al., 2009

Velocity-strengthening friction and postseismic slip rate

- at some depths (~0 to 2 and 10+ km), the friction coefficient μ increases with sliding velocity V

$$\mu = \mu_o + (a-b)\ln(\frac{V}{V_o})$$

(a-b) = velocity-strengthening parameter

- this leads to stable sliding along the fault instead of earthquakes, and accelerated postseismic slip

$$V = V_o \exp(\frac{d\tau}{(a-b)\sigma_o})$$

σ_o = effective normal stress

$d\tau$ = shear stress change

$(a-b)\sigma_o = A - B$

equation applies after some small threshold slip distance

Find parameters that minimize misfit of modeled GPS site velocities to observations

$$\mathrm{WRSS} = \sum_{t=1}^{900} \sum_{stn=1}^{56} \sum_{dof=1}^{3} [(\frac{\Delta}{\sigma})^2]_{t,stn,dof}$$

Δ = residual σ = data error

WRSS at a particular time is WRSS*

$$\mathrm{WRSS}^* = \sum_{stn=1}^{56} \sum_{dof=1}^{3} [(\frac{\Delta}{\sigma})^2]_{stn,dof}$$

Izmit postseismic deformation prior to the Düzce earthquake: three hypotheses

afterslip: velocity-strengthening friction

$$V = V_0 \exp(\frac{d\tau}{A - B})$$ ✓

afterslip: viscous creep along shear zone (Newtonian)

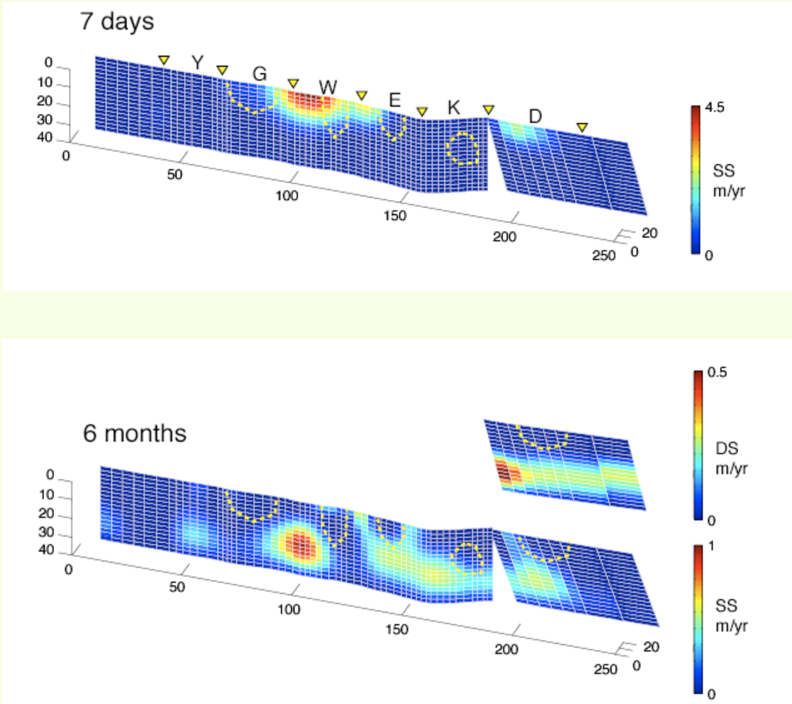
$$V = \frac{w d\tau}{\eta} + V_0$$ ✗

viscoelastic relaxation: lower crust (Newtonian)

$$\dot{\epsilon} = \dot{\epsilon}_o + \frac{d\sigma}{\eta}$$ ✗

(Hearn et al., 2002. 1st 80 days postseismic data)

Earliest postseismic deformation: mostly frictional afterslip



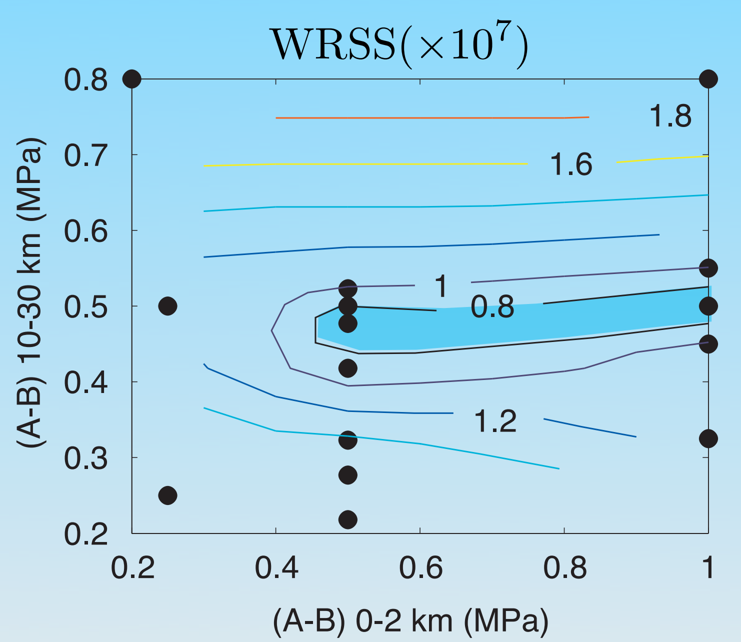
Early postseismic GPS velocities are well explained by afterslip.

Shallow afterslip is anti-correlated with coseismic slip.

Weakly velocity-strengthening friction.

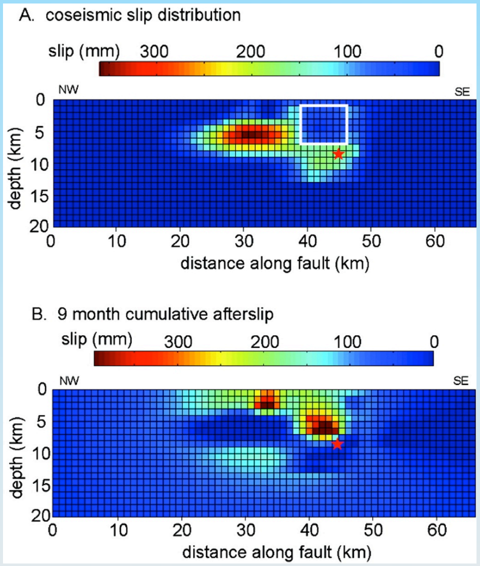
Ergintav et al., 2009 figures. Also Hearn et al. 2002 and 2009 (afterslip).

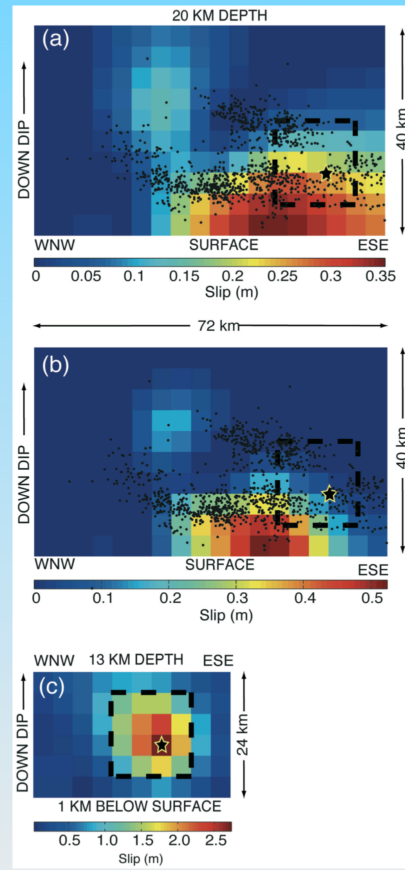
Which parameters worked best?



A-B = about 0.5 MPa in mid-crust: small (Hearn et al., 2009)

Parkfield (M6), Racha (M 6.8), other moderate earthquakes: postseismic deformation is only (velocity-strengthening) afterslip

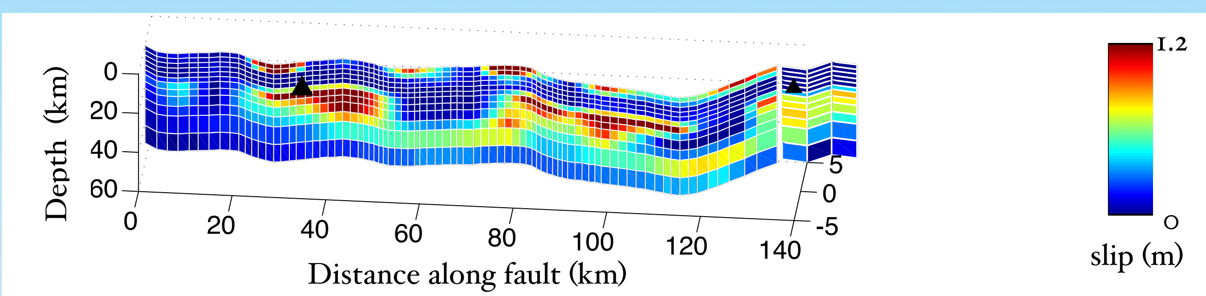




These quakes do not stress the lower crust / upper mantle enough to cause detectable viscoelastic relaxation

M 7.5 Izmit quake:Afterslip is insufficient to explain the GPS site velocities after 3 months

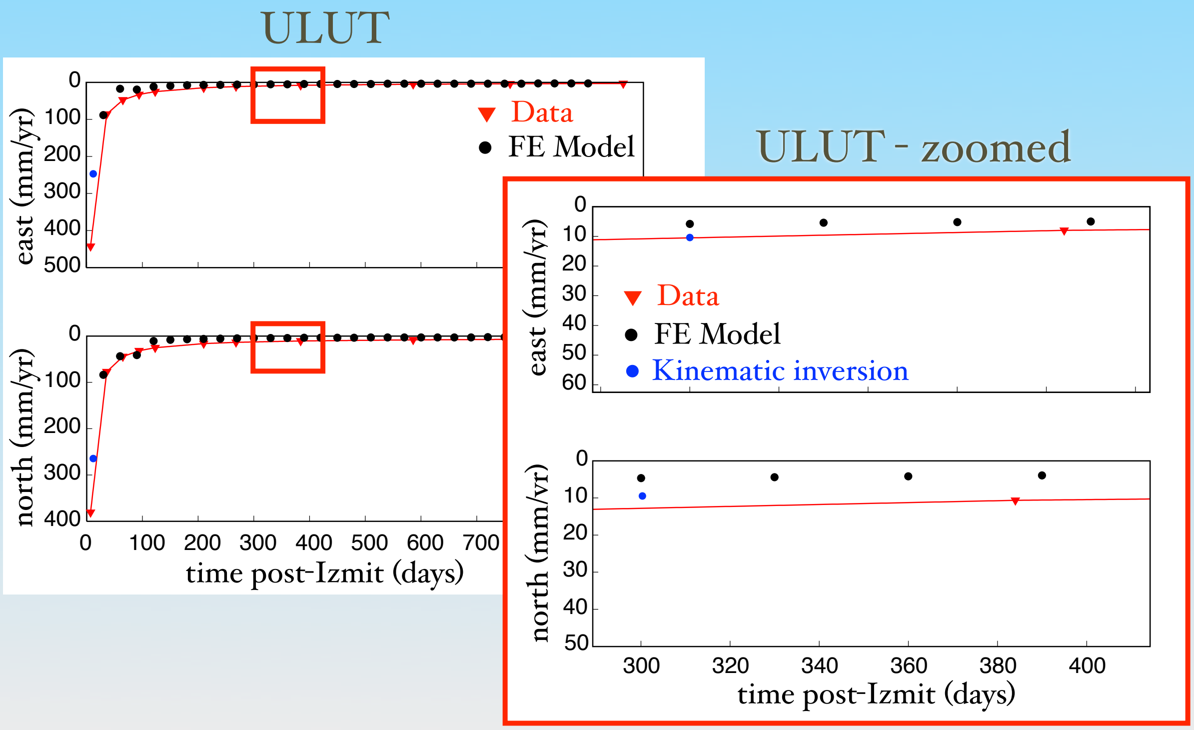
Total modeled afterslip after a year



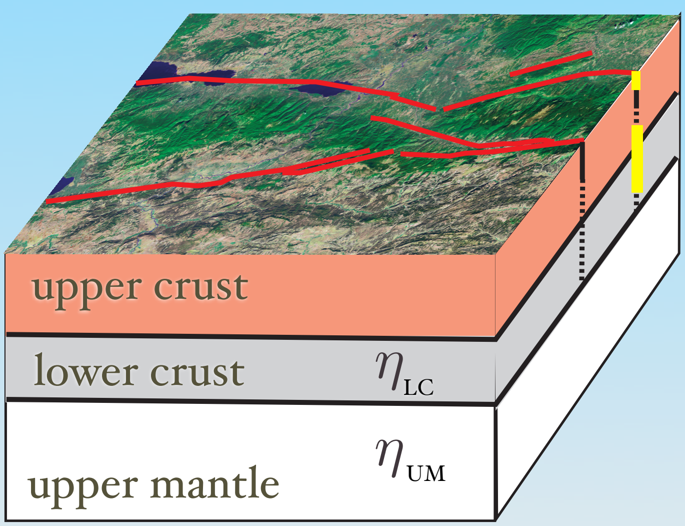
$M_o = 1.07 \times 10^{20}$ Nm

Not enough! About *twice* this slip would be required. But all coseismic shear stress on the fault has been spent.

Afterslip model : GPS velocities too slow at some sites, especially after several months



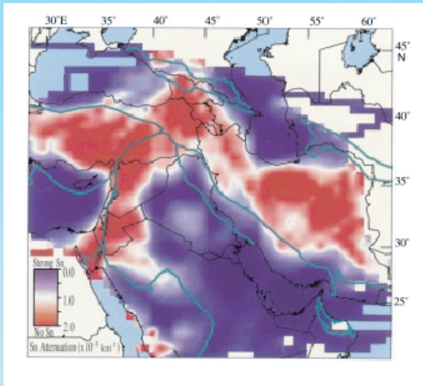
Simplest hybrid model: Afterslip plus Maxwell viscoelastic relaxation



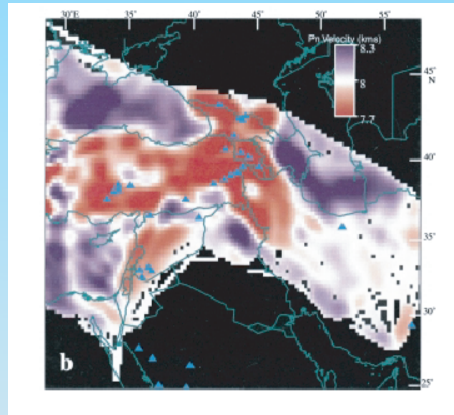
(mantle asthenosphere)

- (A-B): 0 to 2 km and 10+ km intervals held constant
- η_{LC}: lower crust - vary
- η_{UM}: upper mantle - vary

Geophysical evidence for moderate mantle viscosity?



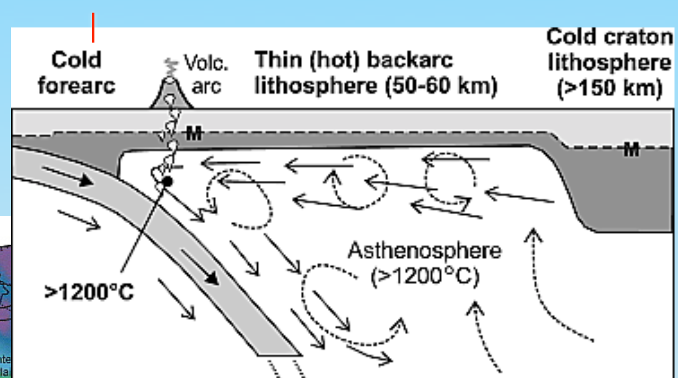
Sandvol et al., 2001: attenuated Sn (regional seismic phase) beneath Anatolia

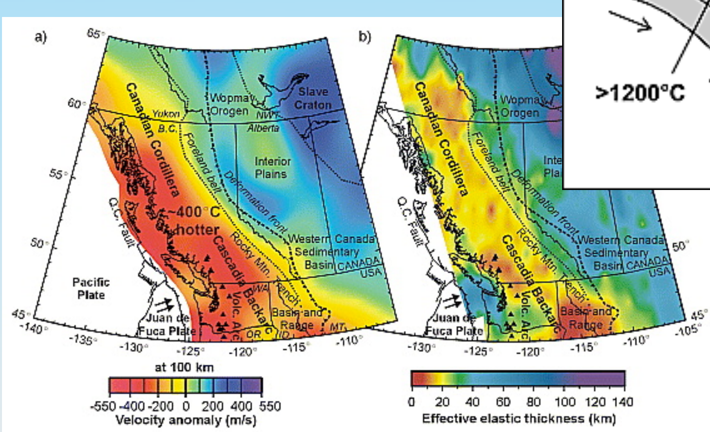


Hearn et al. 1994: (not me!) Slow Pn beneath Anatolia

- Several seismic studies suggest high T and/or melt
- This is consistent with moderate viscosities

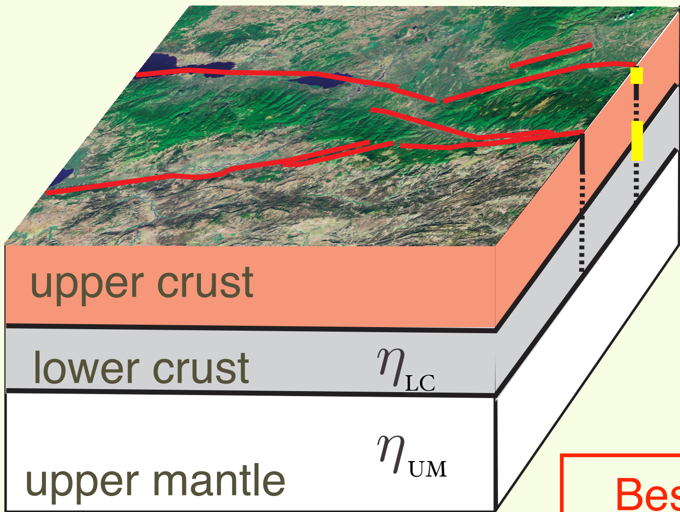
Regions with subduction zones nearby: moderate upper mantle viscosity agrees with heat flow models





Currie and Hyndman, 2006

Later postseismic deformation: relaxation of viscoelastic lower crust and upper mantle



(mantle asthenosphere)

A simple model

- η_{LC} : lower crust - vary
- η_{UM} : upper mantle - vary

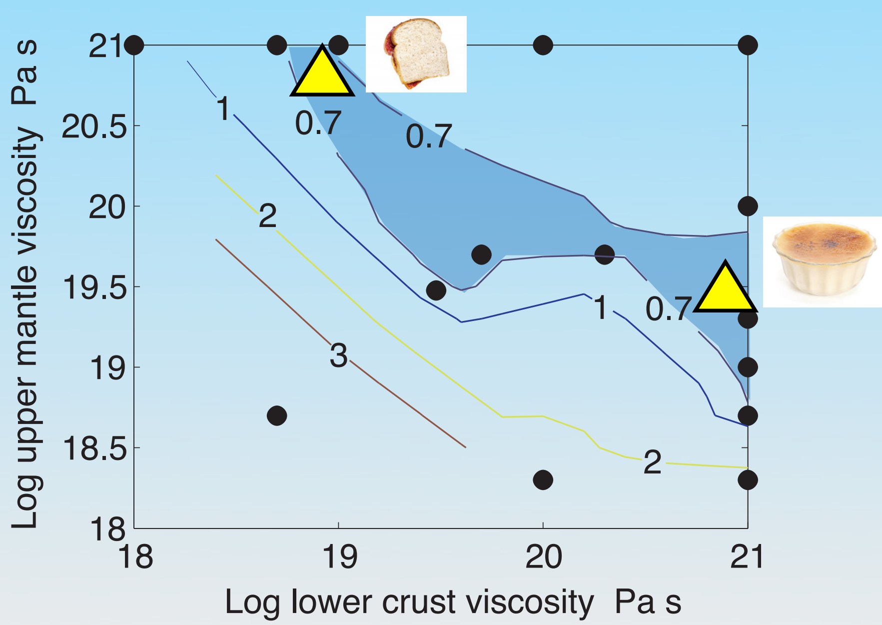
Best fit to postseismic data:

$\eta_{(LC\ or\ UM)} \approx\ 2\ to\ 5 \times 10^{19} Pa\ s$

or $\eta_{(both)} \approx 5 \times 10^{19} Pa\ s$

Which parameters worked best? Afterslip + viscoelastic relaxation models

Normalized WRSS, t = 0 to 900 days



Fit improves for lower crust or upper mantle viscosities of 2 - 5 x 10 ^{19} Pa s

Maxwell time is 40 to 80 years

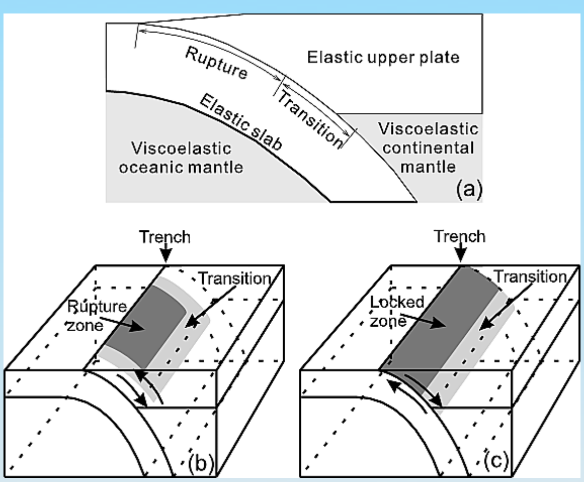
How my LC-UM viscosity estimates compare with others

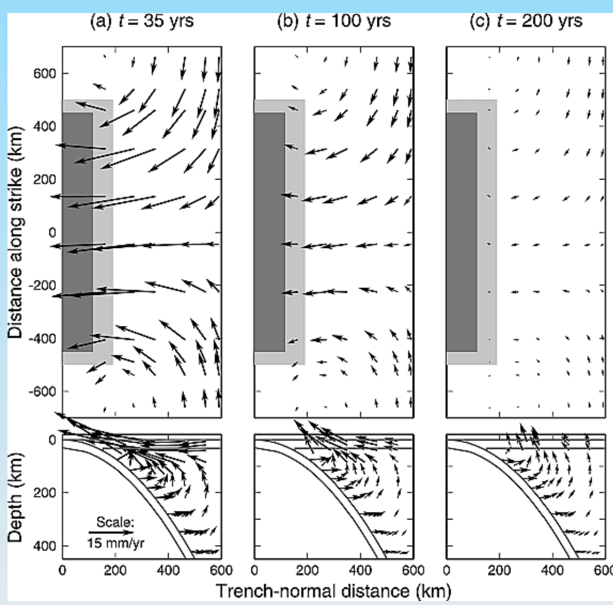
SUPPLEMENTAL TABLE 2 Viscosity estimates derived from geodetic measurements of post-loading deformation[a]

Source event	Mw	Slip[b]	Tectonics[c]	Viscosity (x 10^{18} Pa s) LC	UM	Reference
Earthquakes						
1915-54 Central Nevada	7.6	ss/ns	C-PBZ/BA	> 100	1-7	(Gourmelen & Amelung 2005)
1915-54 Central Nevada	7.6	ss/ns	C-PBZ/BA	100-300	10-30	(Hammond et al 2007)
1959 Hebken Lake,		ns	C-PBZ/BA	> 100	4	(Nishimura & Thatcher 2003)
1992 Landers[d]	7.4	ss	C-PBZ/BA	8-24	1-6	(Pollitz et al 2000)
1997 Manyi, Tibet	7.6	ss	C-PBZ	4-8	-	(Ryder et al 2007)
1999 Hector Mine[d]	7.1	ss	C-PBZ/BA	32	4.6	(Pollitz 2003)
1999 Izmit, Turkey	7.4	ss	C-PBZ	20-50	20-50	Hearn et al., 2007 unpublished
2000 South Iceland	6.5	ss	MOR-PBZ	10	3	(Árnadóttir et al 2005)
2002 Denali[d]	7.8	ss	C-PBZ/BA	> 10	2-4	(Freed et al 2006)

Burgmann and Dresen, 2008

Postseismic deformation following 1960 M 9.4 Chile earthquake: afterslip + (now) viscoelastic relaxation





Hu et al., 2004

Similar findings for Denali M 7.9 2002

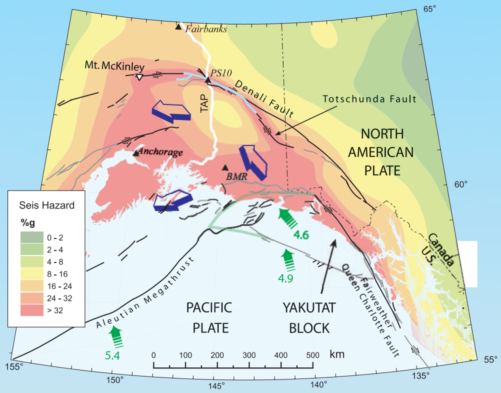
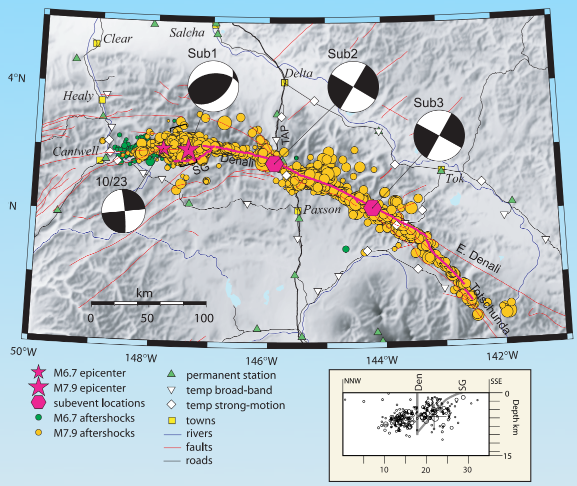
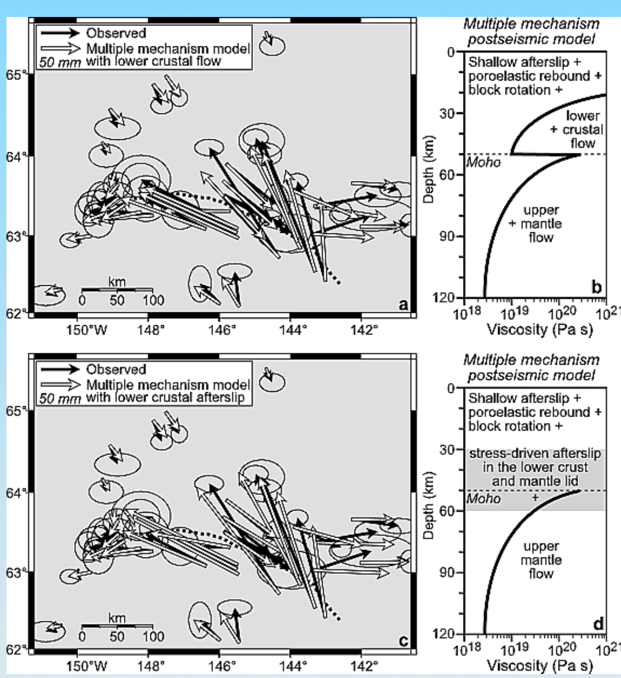


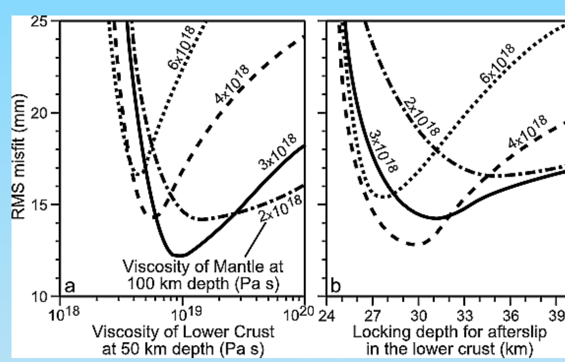
Fig. 1. Plate tectonic setting of southern Alaska and major tectonic elements. The Pacific and North American plates converge at 5.4 cm/year beneath Anchorage (*53*), and the Yakutat block collides with North America independently. Plate motion is indicated by green arrows. Blue open arrows schematically show lateral movement of broad region south of the Denali fault. Blue line, surface rupture. The probabilistic seismic hazard is shown by peak ground acceleration with 10% probability of exceedance in 50 years (*14*). TAP, Trans Alaska Pipeline. Triangles, station locations in Fig. 5. Black lines, Quaternary faults; gray lines, Neogene faults.

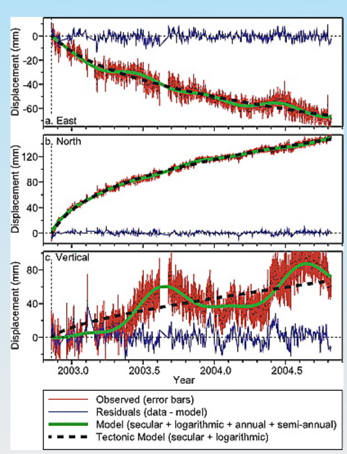


A. Freed

Similar findings for Denali M 7.9 2002

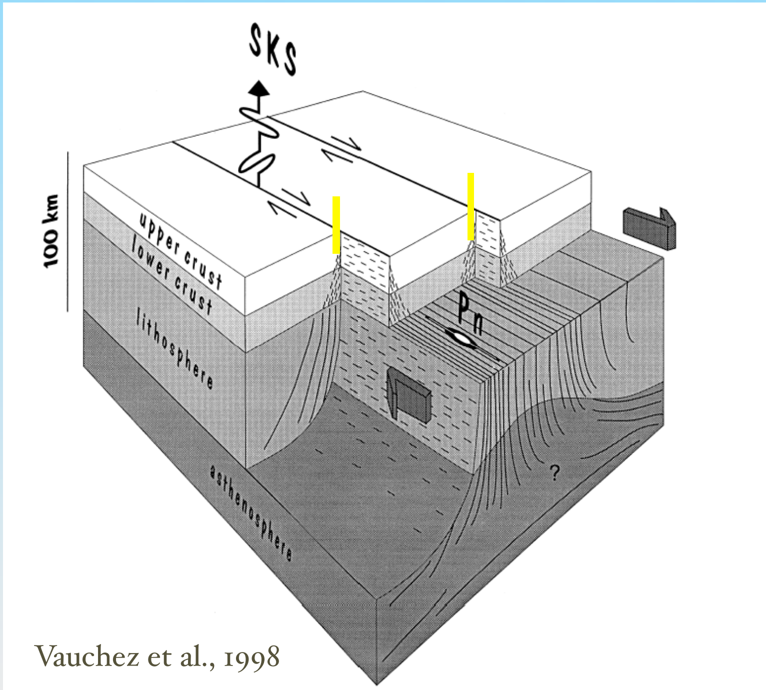






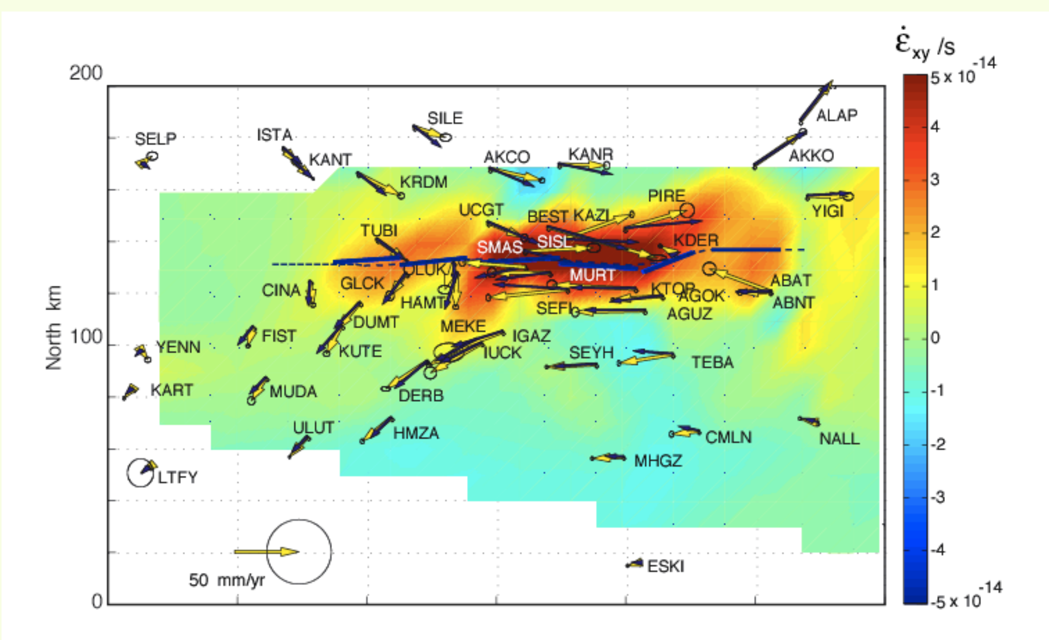
Small quakes - just afterslip

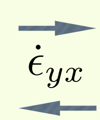
Large quakes - afterslip plus viscoelastic relaxation of lower crust and upper mantle (halfspace? broad shear zone? rheology?)



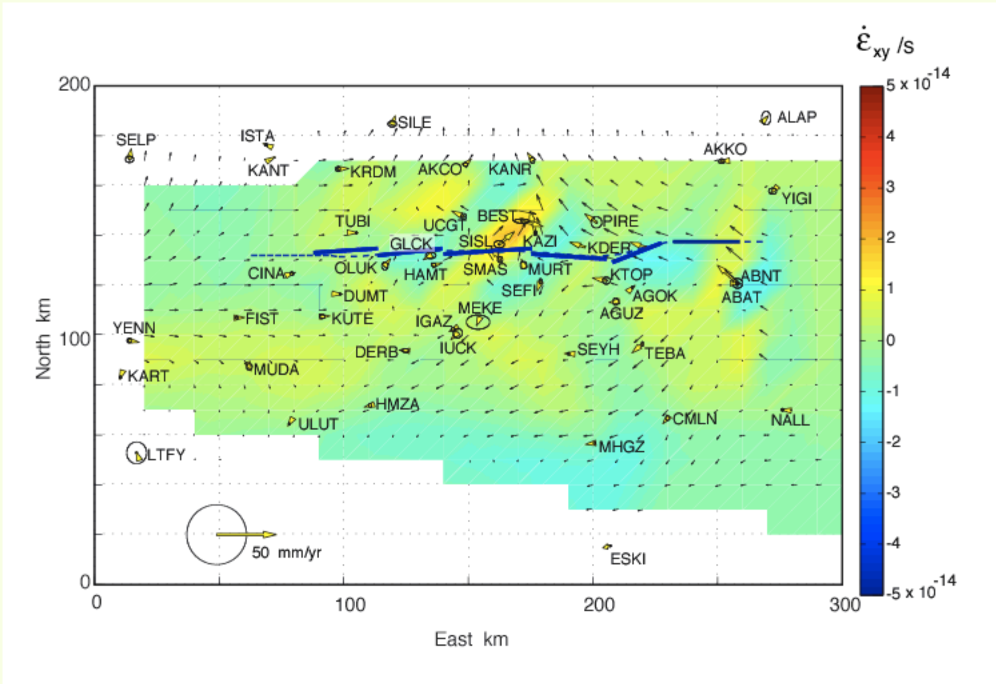
Vauchez et al., 1998

Modeled and GPS velocities six months after the Izmit earthquake



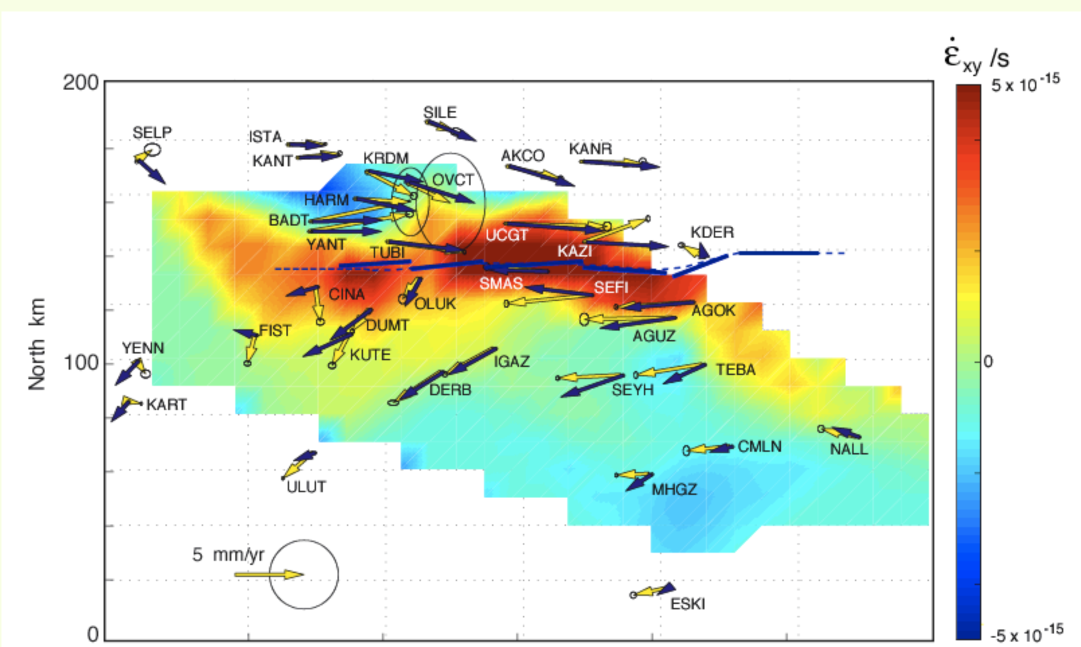


Velocity residuals six months after the Izmit earthquake



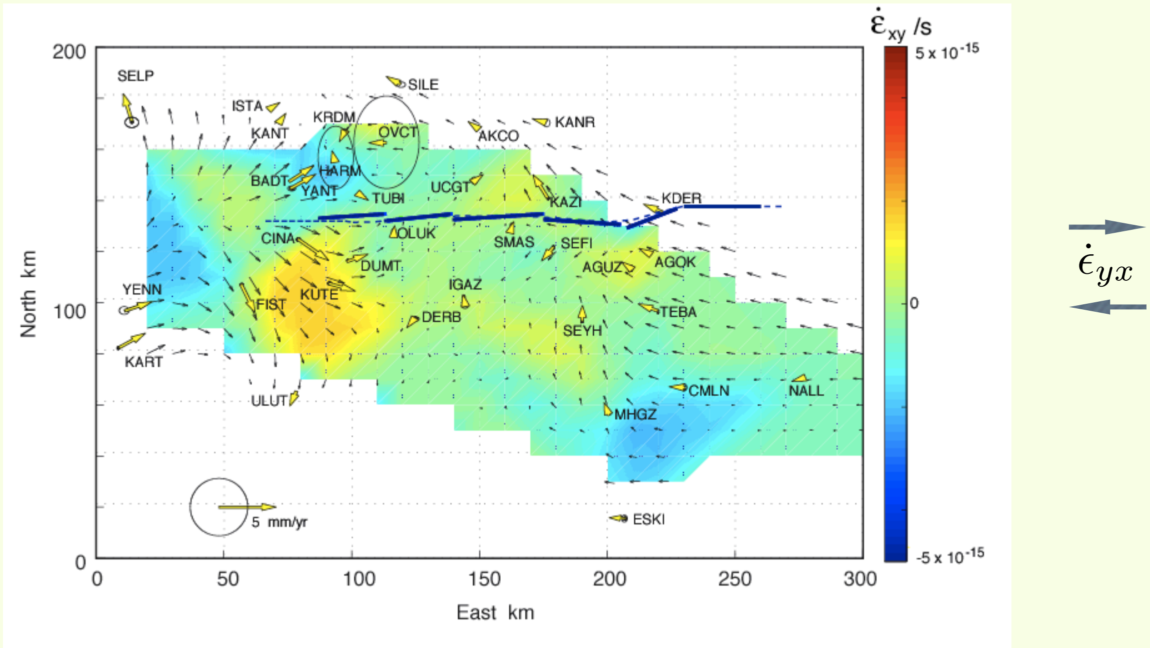
$\dot{\epsilon}_{yx}$

Modeled and GPS velocities six years after the Izmit earthquake



$\dot{\epsilon}_{yx}$

Residuals six years after the Izmit earthquake



But we have more information!

1. interseismic (pre-Izmit) GPS velocities
2. geological observations:
   - nonlinear viscoelasticity
   - shear zones in the lower crust and upper mantle
   - geologic slip rate and earthquake chronology

Interseismic deformation: before 1999

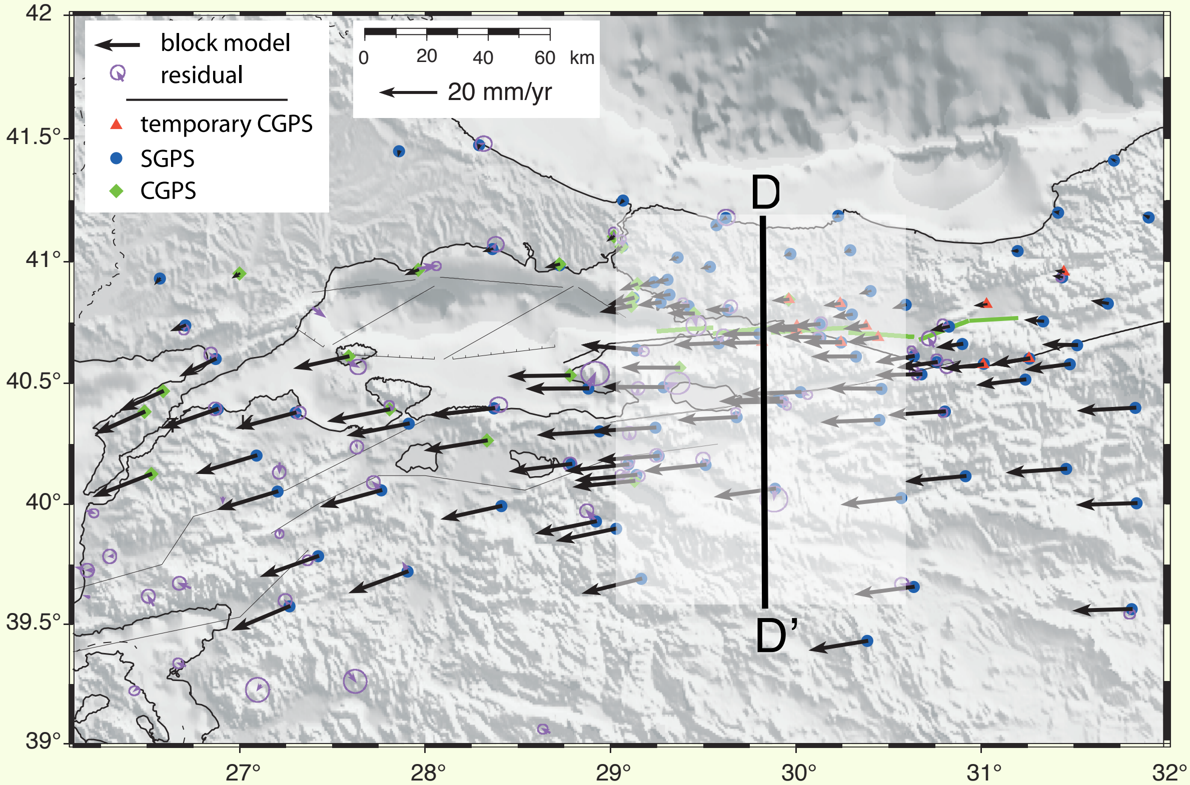
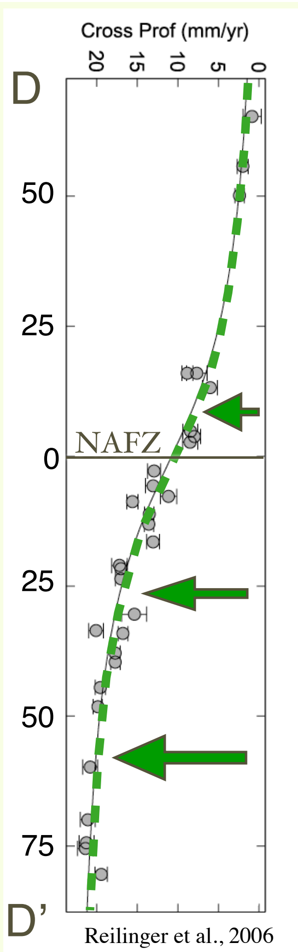


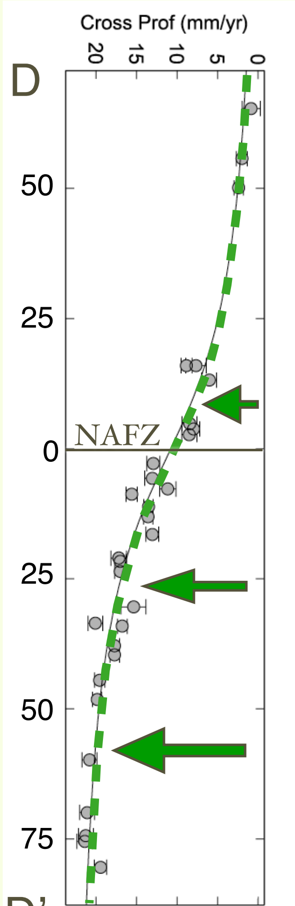
figure from Ergintav et al., 2009



NAFZ interseismic deformation

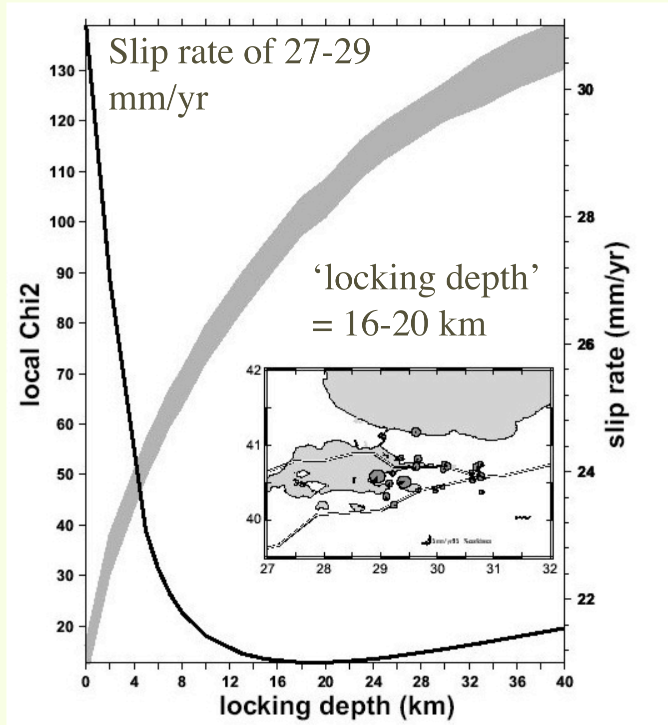
- localized strain around NAFZ
- insensitive to time since previous major earthquake (profiles across various NAFZ segments look similar)

Reilinger et al., 2006

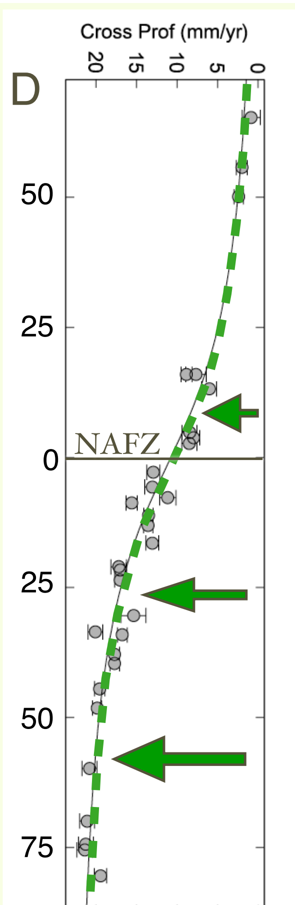


Reilinger et al., 2006

NAFZ interseismic deformation



Reilinger et al., 2006



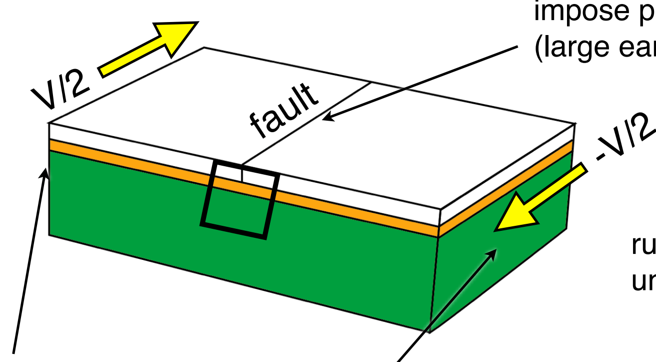
Reilinger et al., 2006

NAFZ interseismic deformation

Is a lower crust / upper mantle viscosity of 5×10^{19} Pa s consistent with this stationary, localized deformation?

Earthquake cycle modeling is required.

Earthquake cycle modeling



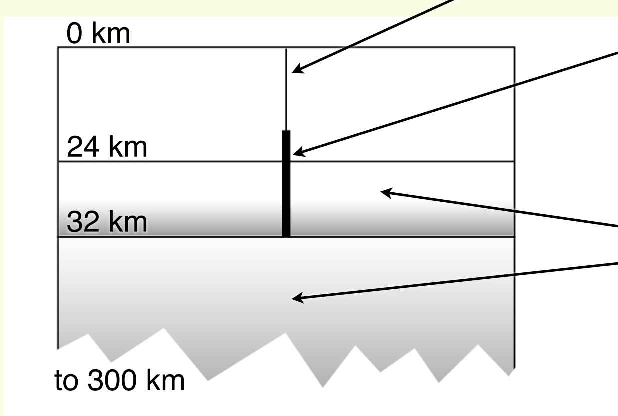
impose periodic coseismic slip
(large earthquakes)

run for several earthquake cycles,
until cycle invariant status attained

impose regional strain via
model boundary velocities

compare predicted interseismic
velocities with observed GPS profiles

Earthquake cycle modeling



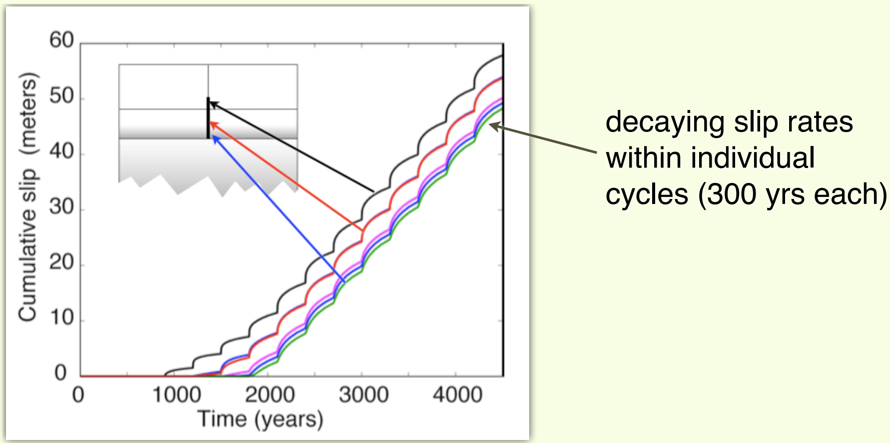
periodic coseismic slip and far-
field velocity boundary conditions
(20 mm/yr, quakes with 6 m of
slip every 300 years)

frictional or viscous afterslip:
allowed to 24 km
(same parameters as Izmit
postseismic model)

viscoelastic layers:
to start, same as the Izmit
postseismic model

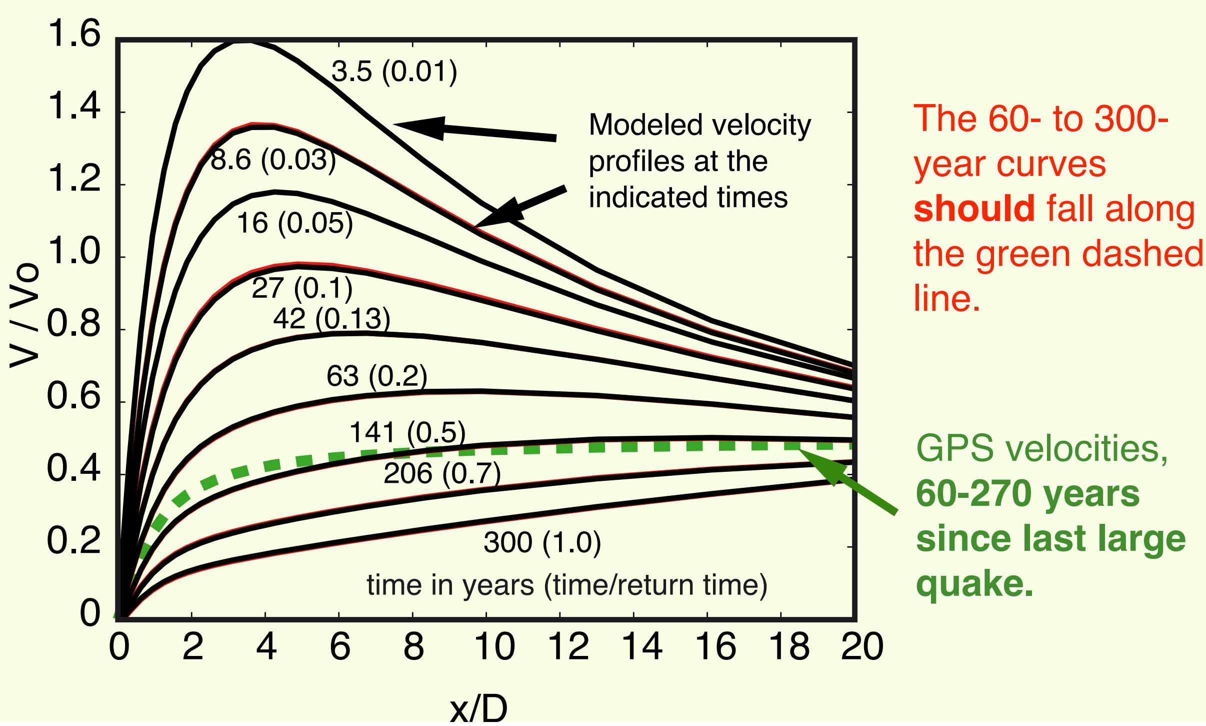
Earthquake cycle modeling

- cycle-invariant status is attained when results from one cycle look like those from the last one: slip rates, surface velocities, stresses...



- we can compare modeled surface velocities at appropriate time in the earthquake cycle to GPS velocities

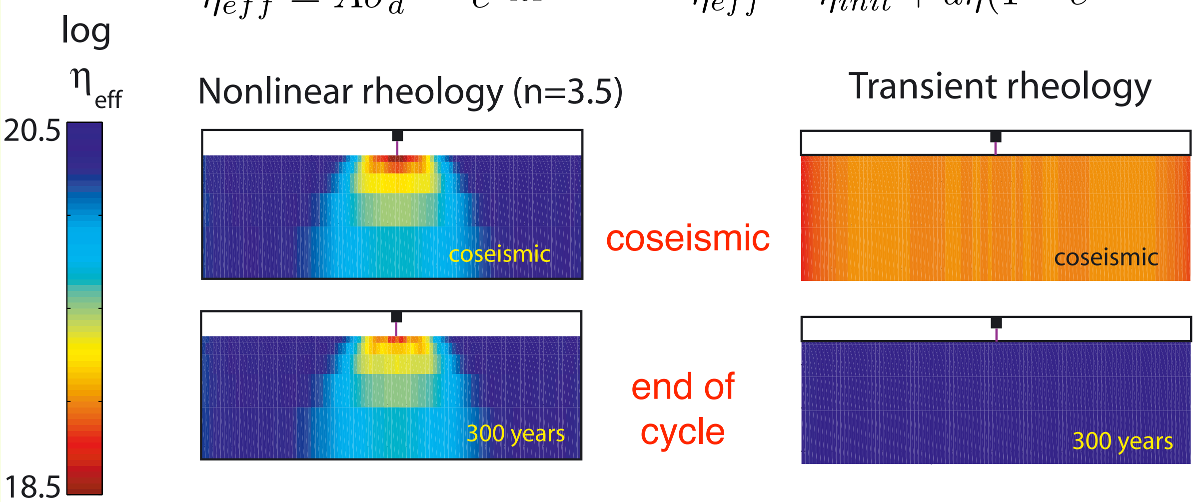
Interseismic velocity profiles from the best postseismic model



Try other viscoelastic rheologies

$$\eta_{eff} = A{\sigma_d}^{1-n} e^{(\frac{Q}{RT})}$$

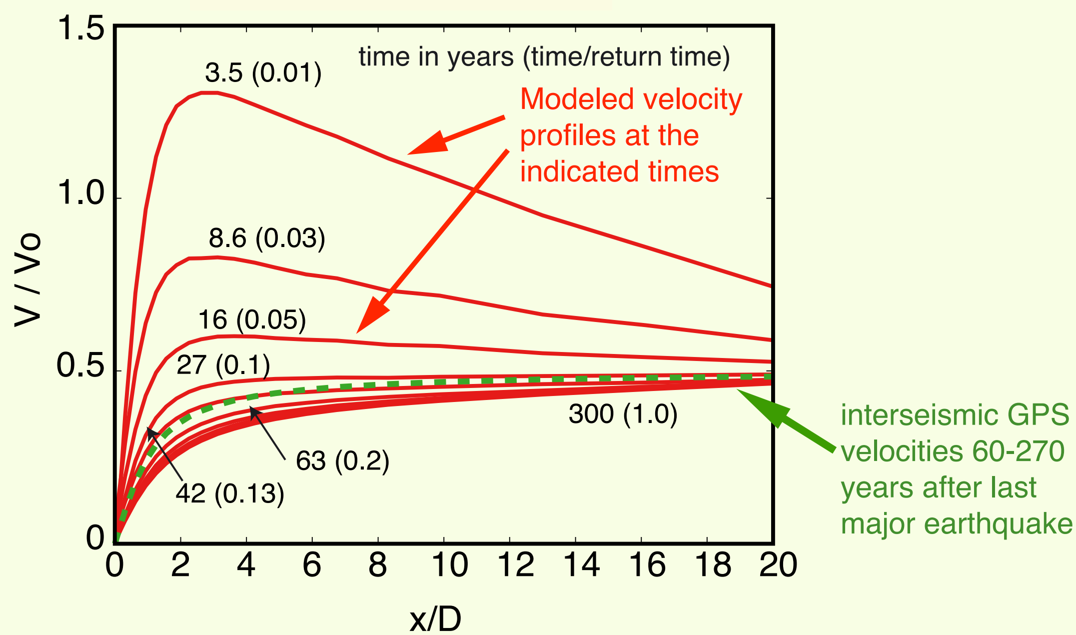
$$\eta_{eff} = \eta_{init} + d\eta(1 - e^{-t/t_{char}})$$



A is a constant
R is the gas constant
Q is the activation energy
σ_d is differential stress

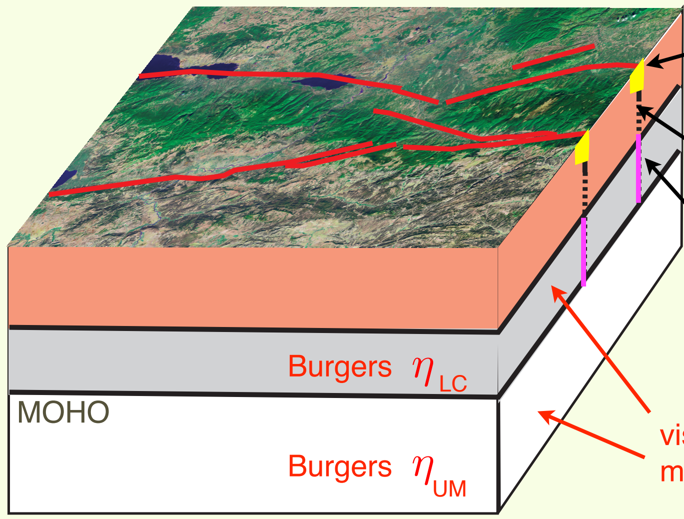
t_{char} is a characteristic evolution time
η_{init} is the initial viscosity (low)
$d\eta/\eta_{init}$ depends on change in stress rate, and temperature

Burgers (transient) rheology works best



$d\eta/\eta_{init}$ = 10 η_{init} = 2 to 5 x 10^{19} Pa s t_{char} = 10 years

A model that can explain postseismic and interseismic, central NAFZ deformation



- $\eta_o = 2 - 5 \times 10^{19} Pa\ s$
- $\eta_s > 5 \times 10^{20} Pa\ s$
- $t_c =$ decades

Hearn et al., 2009, Burgers body with similar parameters also required by Hetland et al. (2009, 2005)

No experimental justification for such a dramatic evolution of effective viscosity

Based on the experiments of Post (1977) and Chopra (1997):

- $\frac{\eta_s}{\eta_o} = 2 - 7$

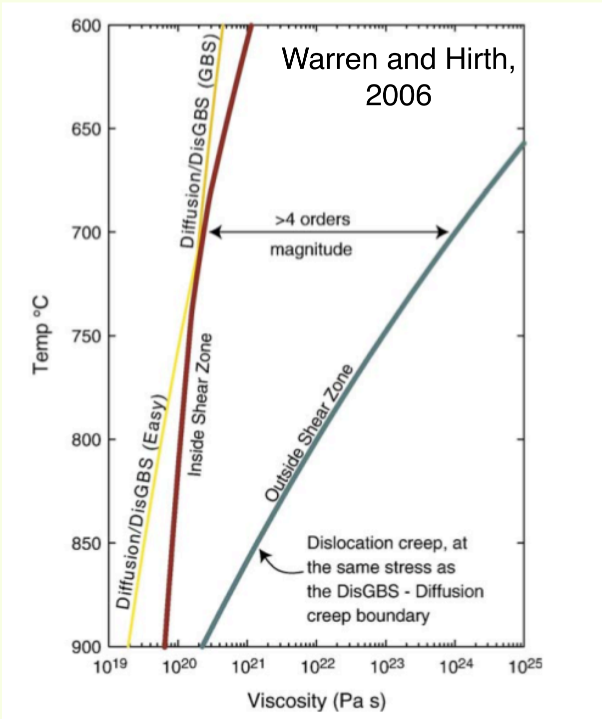
We (and Hetland 2005 and 2009) require at least 10.

- η evolution rate t_c depends on $\dot{\epsilon}$ as shown here: SLOW →

$\dot{\epsilon}$	t_c
$10^{-14}\ /s$	20,000 years
$10^{-13}\ /s$	2,650 years
$10^{-12}\ /s$	304 years
$10^{-11}\ /s$	35 years

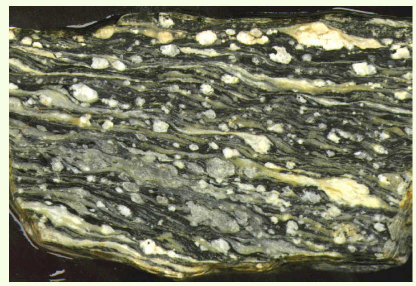
We (and Hetland 2005 and 2009) require decades. The data above are for dry dunite. Evolution for wet peridotite would be faster but I cannot find any experimental data.

Grain-size sensitive creep makes viscous shear zones that extend down to the Moho



upper mantle shear zones: much lower viscosity than host rock (to 950° C)

same with lower crustal shear zones (e.g., Mehl and Hirth, 2008)



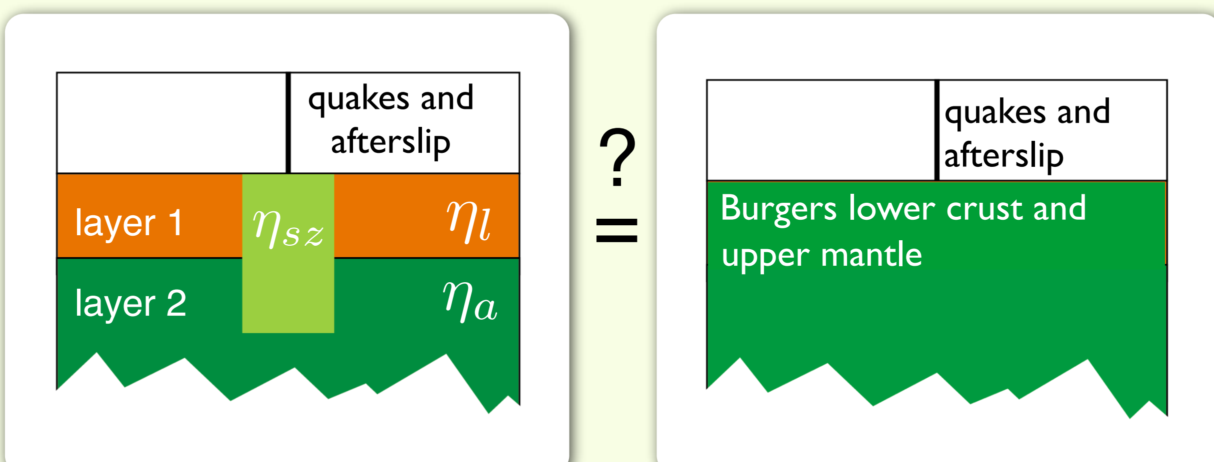
Is transient rheology required if we model lithosphere a bit more realistically?

Test models:

(Maxwell or power-law layers, Maxwell channel)

Target:

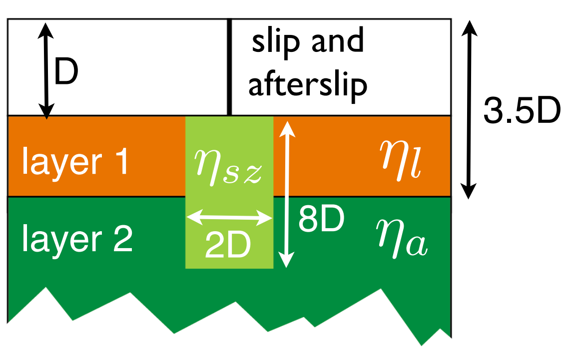
“reference model” which worked for the NAFZ



Conclusions so far (work in progress)

- A lithosphere-scale shear zone can **help** explain high postseismic velocities while preserving localized, stationary interseismic deformation
- Models with a moderate asthenosphere η and a high shear zone (and lithosphere) η work best
- Burgers body material still needed, but required η change may be more consistent with available lab values
- Experimental constraints on transient rheology for more rock types at high P and T would be nice.

Test model parameters (today)



channel + 2 layers

w_{sz} = 2D

η_{sz} vary

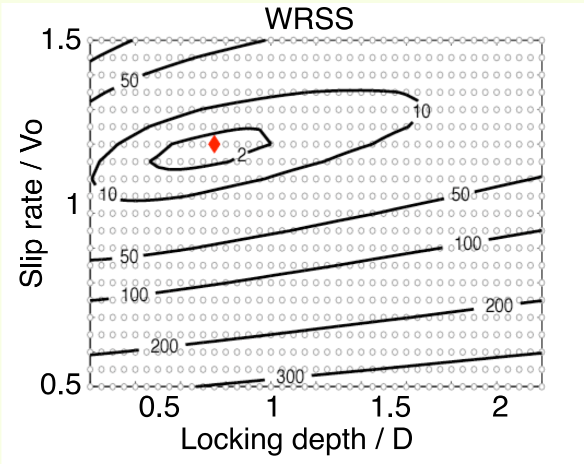
η_a vary*

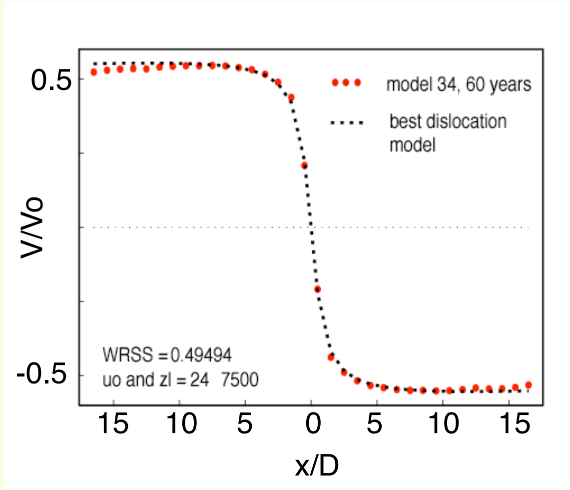
η_l = 10^{21} Pa s

*or depth-dependent power-law rheology from Hirth and Kohlstedt, 2003

Do any of the channel models produce stationary interseismic deformation?

For each model:
Estimate locking depth (z_l) and slip rate (u_o),
60 and **200** years into a 300-year earthquake cycle

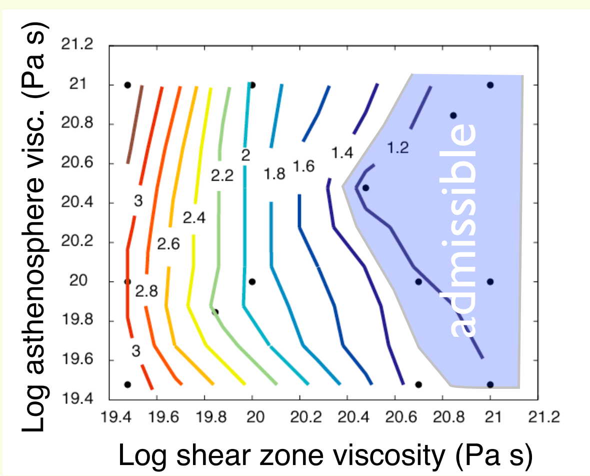




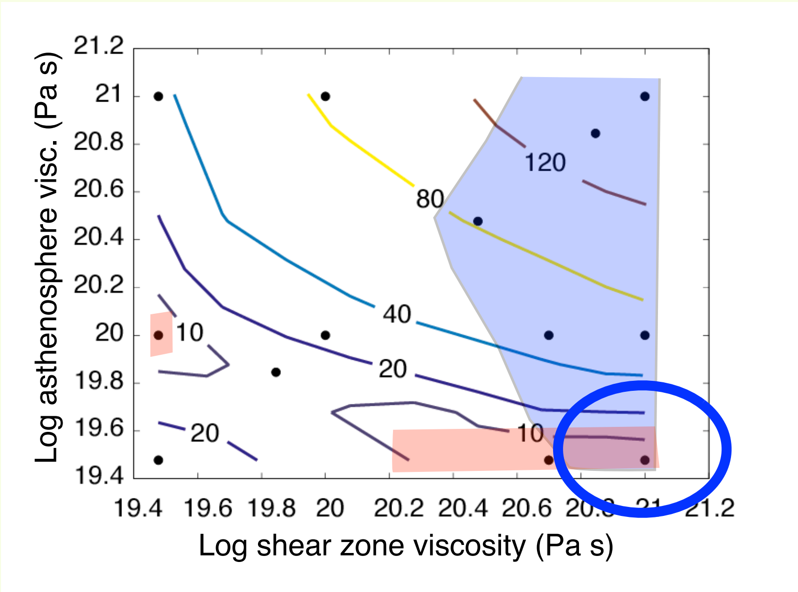
Models with a stiff shear zone produce stationary interseismic deformation

apparent change in locking depth

$$\frac{z_{l(200\ yr)}}{z_{l(60\ yr)}}$$



Are any of the admissible channel models also consistent with Izmit postseismic deformation?



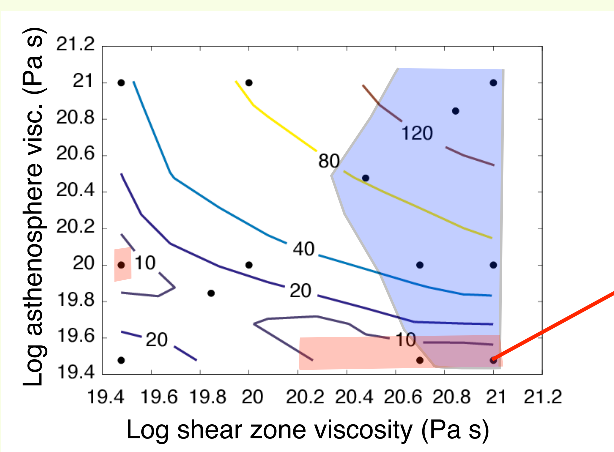
Contours of misfit* to velocities from the reference Burger's Body model with $\eta_o = 5 \times 10^{19} Pa\ s$

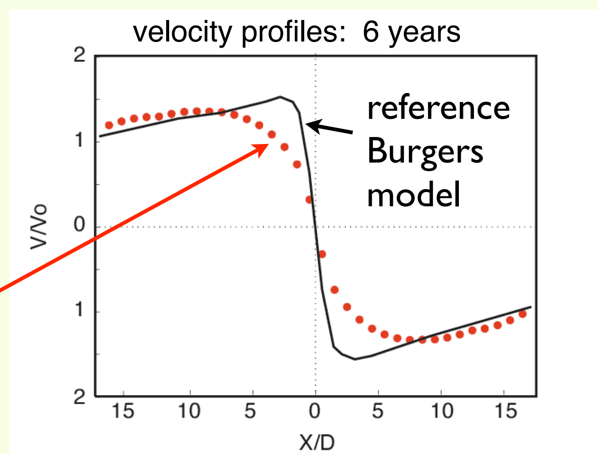
6 years into a 300-year cycle

Models with low asthenosphere viscosity work best.

*squared mean velocity residuals

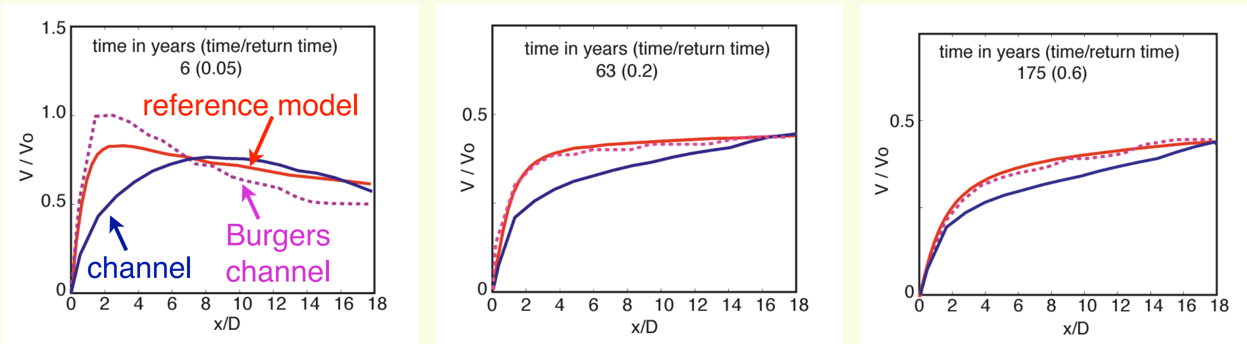
Fit of best channel model to postseismic velocities: not so good





- velocity profile is fit poorly in the near field.
- transient rheology in the channel (shear zone)?

Best model with transient rheology for the channel material (so far)



Required channel parameters:

$$\eta_o = 2 \times 10^{19}\ Pa\ s \qquad \eta_s = 10^{20}\ Pa\ s \qquad t_c = 10\ years$$

$$\frac{\eta_s}{\eta_o} = 5+$$

Conclusions so far
(work in progress)

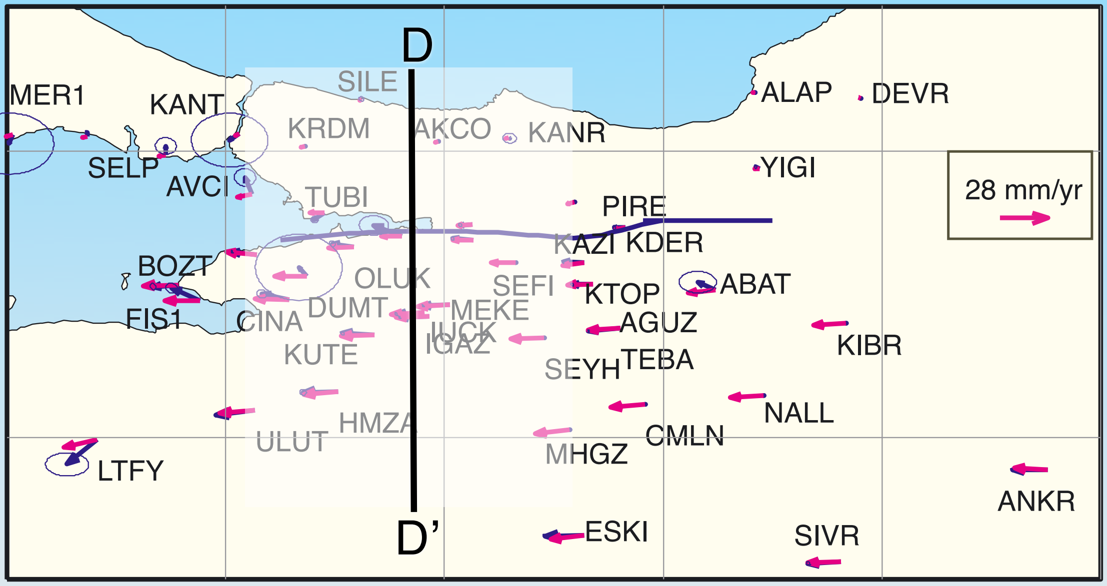
- A lithosphere-scale shear zone can **help** explain high postseismic velocities while preserving localized, stationary interseismic deformation

- Models with a moderate asthenosphere η and a high shear zone (and lithosphere) η work best

- Burgers body material still needed, but required η change may be more consistent with available lab values

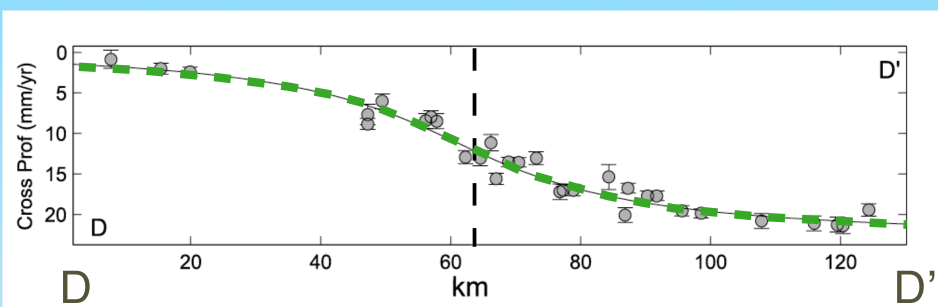
- Experimental constraints on transient rheology for more rock types at high P and T would be nice.

Izmit: Is a broad zone of Maxwell viscoelastic upper mantle compatible with interseismic GPS velocities?



Blue = pre-Izmit GPS velocities, 1-sigma errors
Pink = block model velocities (Reilinger et al., 2006)

Interseismic GPS velocities

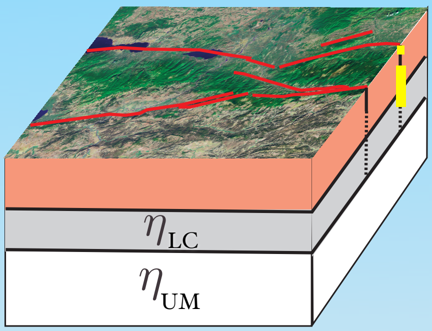


Reilinger et al., 2006

- *localized* strain around NAFZ: like a 20 km locking depth
- *insensitive to time* since previous major earthquake (profiles across other NAFZ segments look similar)

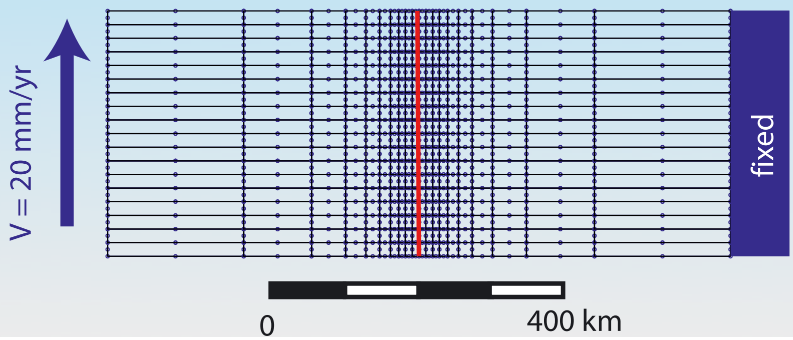
Can lower crust or upper mantle with a viscosity of 5×10^{19} Pa s produce this?

Can the postseismic deformation model explain the observed interseismic deformation?

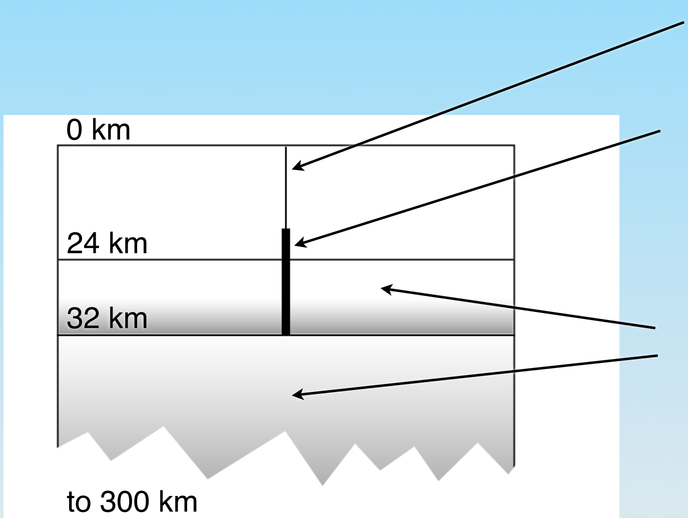


Earthquake cycle modeling is required

- impose periodic earthquakes and velocity boundary conditions on 3D finite-element model of NAFZ and lithosphere
- model several cycles, until cycle invariant status attained
- compare absolute velocities at appropriate time in the earthquake cycle to GPS velocities



Earthquake cycle models



periodic coseismic slip

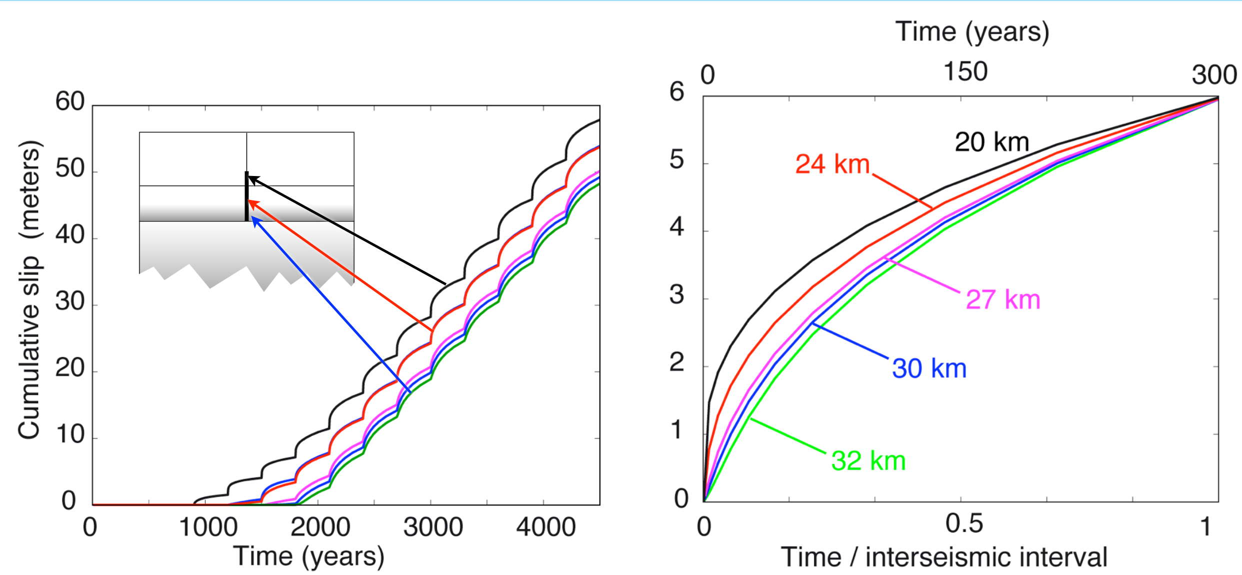
frictional afterslip:
all models to 24 km
(some to 32 km)

viscoelastic layers:
- linear
- nonlinear
- transient

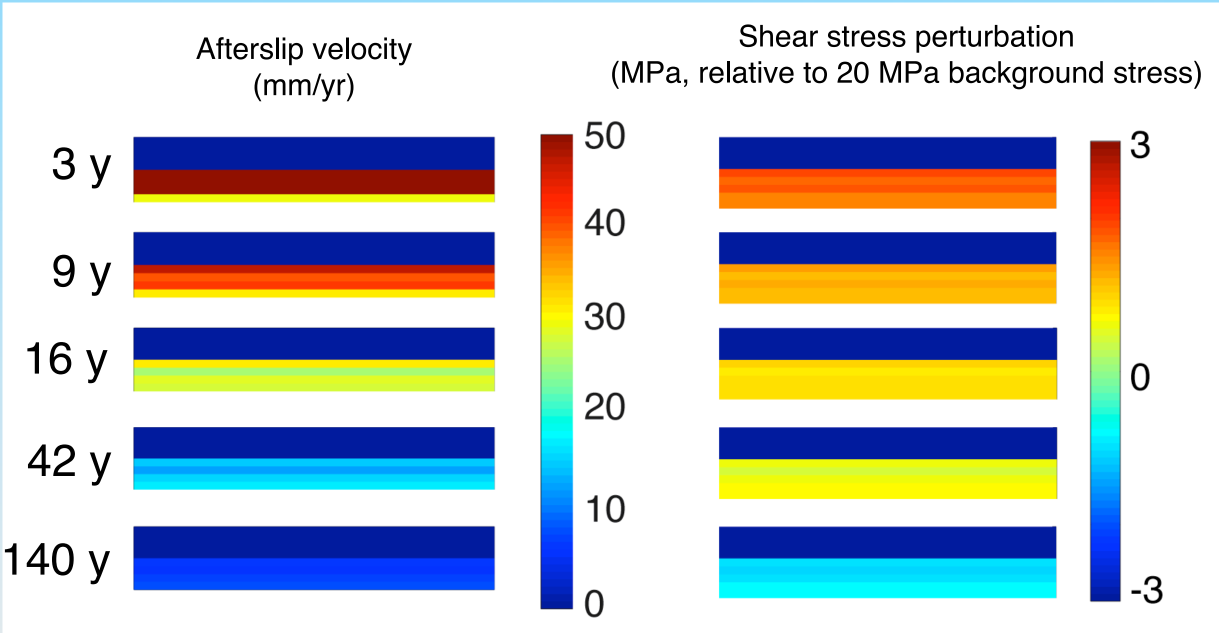
How do we model interseismic deformation?

- Block models (kinematic)
- Earthquake cycle models

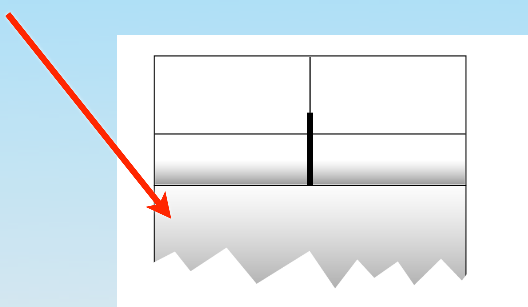
Depth distribution and rate of aseismic slip



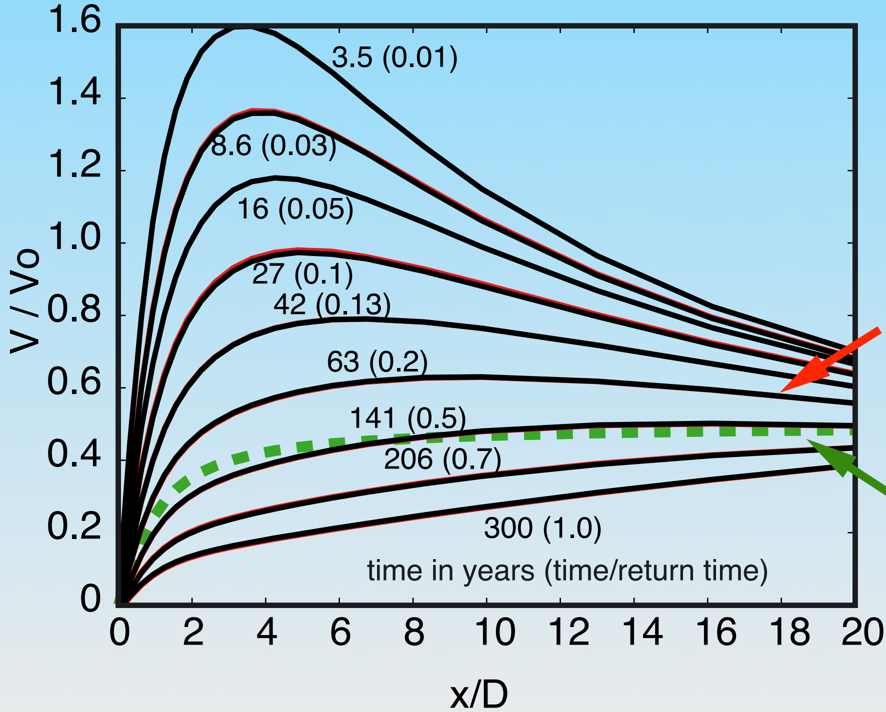
Aseismic slip rate and shear stress fluctuations over the interseismic interval



Meanwhile, viscoelastic relaxation is occurring in the upper mantle, and together these processes control interseismic velocities around the fault.



Best postseismic model is incompatible with interseismic GPS velocities around the NAFZ



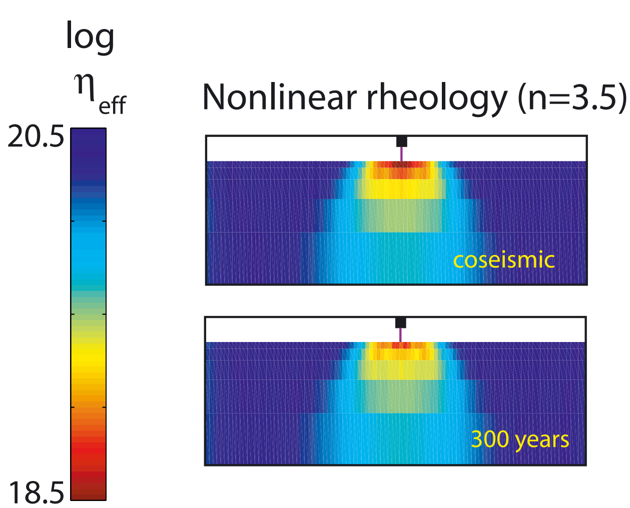
Modeled velocities around fault are very sensitive to time since the last earthquake

Model prediction for 1940's rupture segment

Reilinger et al. 2006 interseismic GPS data

No variation with time in EQ cycle!

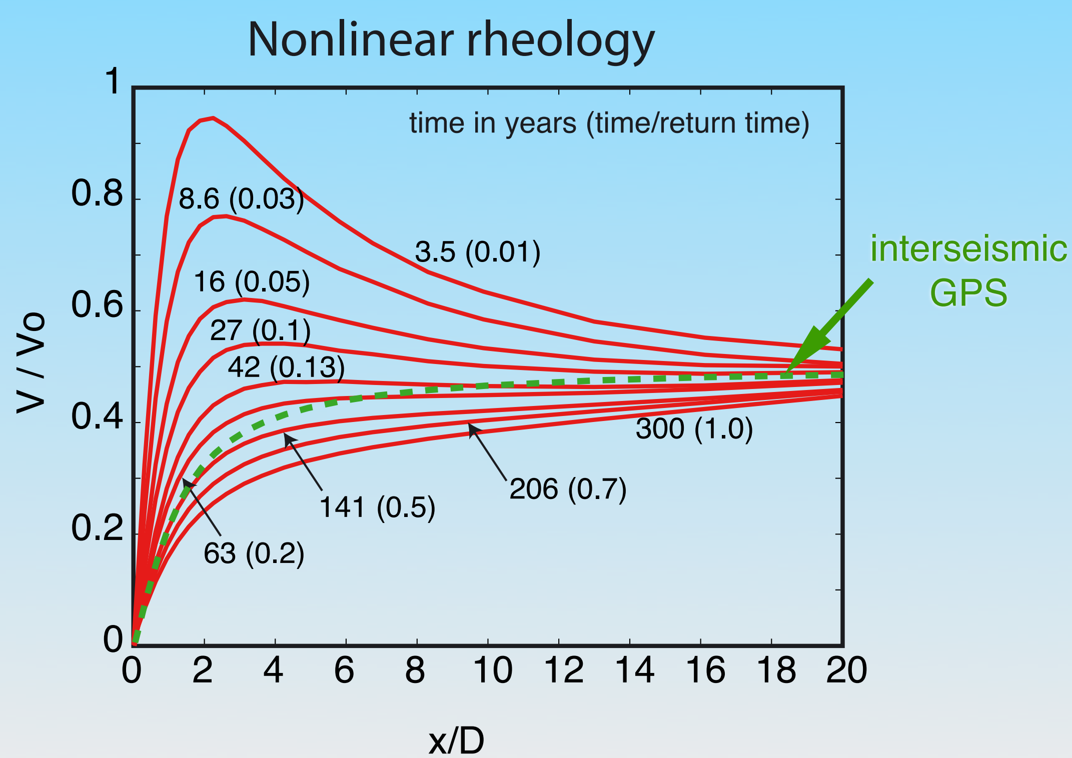
Explore other mantle rheologies: nonlinearly stress-dependent viscosity



$$\eta_{eff} = A\tau^{1-n}e^{\frac{Q}{RT}}$$

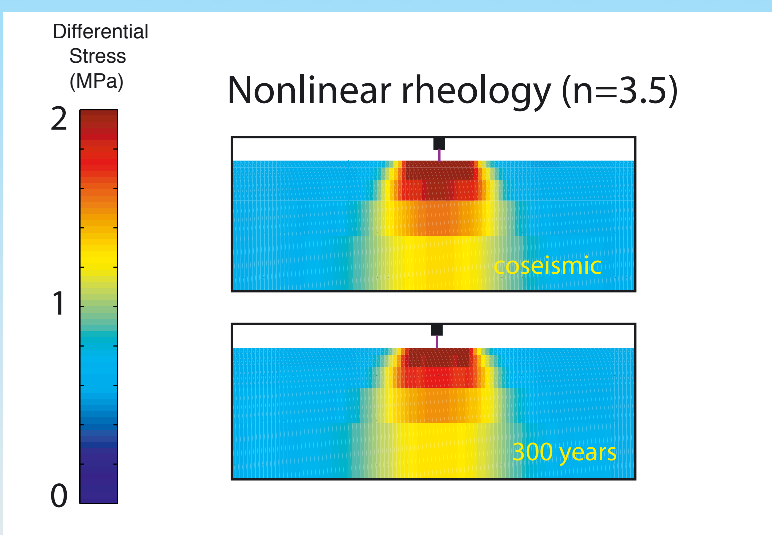
Q is activation energy
R is the gas constant
A is an experimentally determined constant*
τ is the differential stress
*sensitive to grain size, melt fraction etc

This is a bit better...

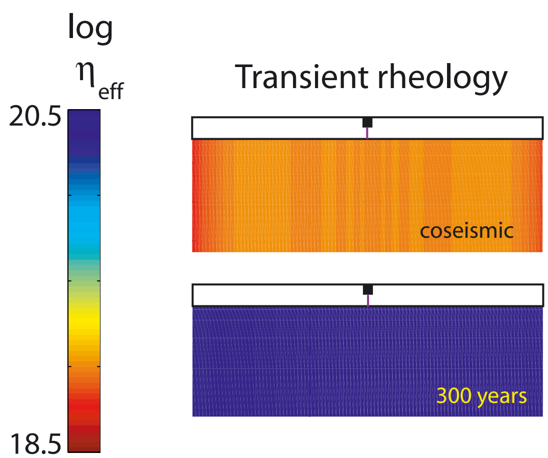


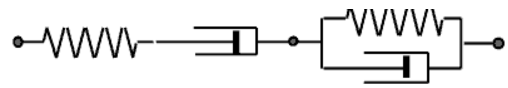
Problem

Differential stress is *too low* for dislocation creep
(nonlinear flow with n > 3)



Explore other mantle rheologies: Burgers Body rheology (two viscosities)





$$\eta_{eff} = \eta_{init} + d\eta(1 - e^{-t/t_{char}})$$

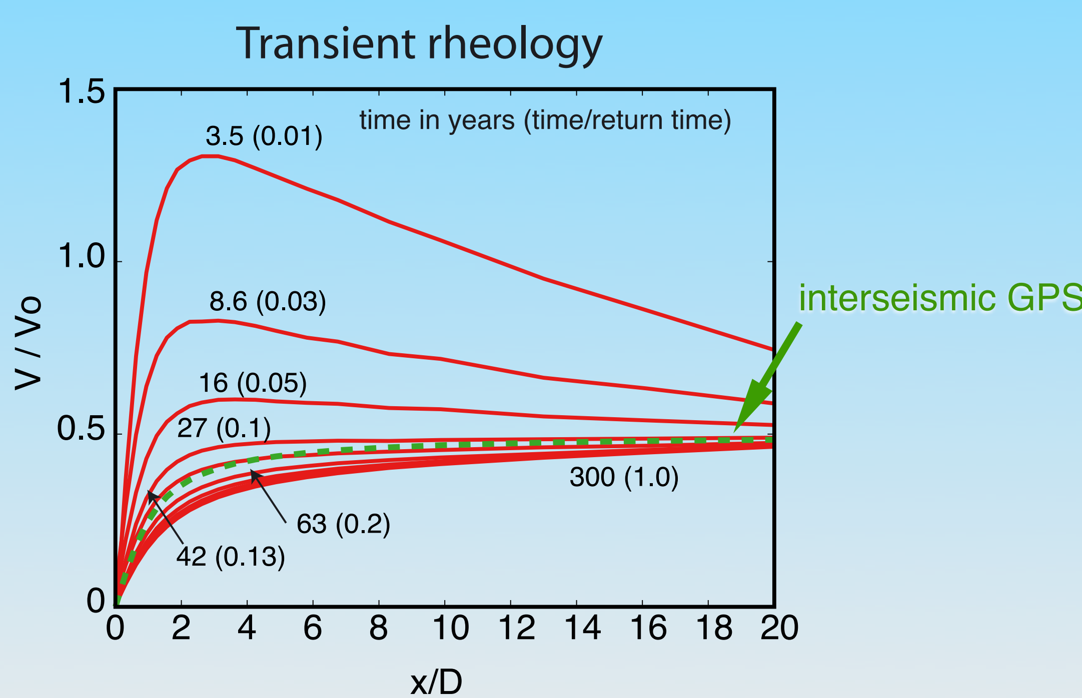
η_{eff} depends on time since a step in stress rate

t_{char} and

$d\eta/\eta_{init}$ depend on change in stress rate and on temperature (for dunite, Chopra, 1997)

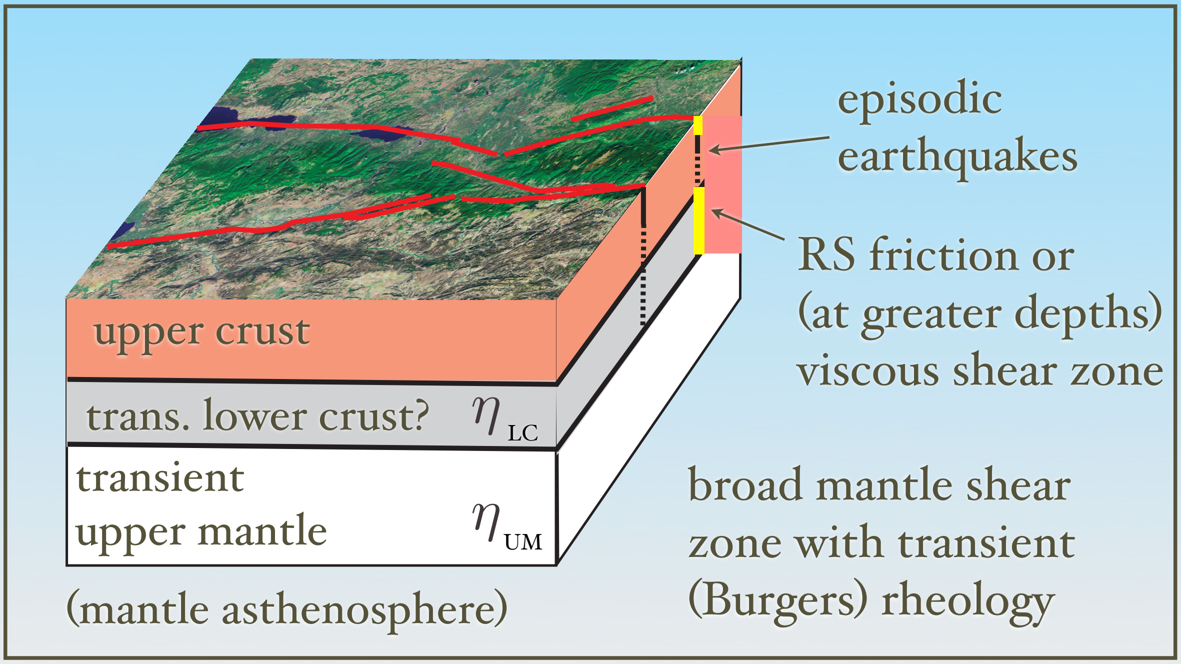
see also Hetland (2005) for 2D analytical models of the NAFZ with transient mantle rheology based on the *correspondence principle*!

This is good - little variation in strain rates for most of the interseismic interval



$d\eta/\eta_{init}$ = 10 $\quad$ η_{init} = 2 to 5 x 10^{19} Pa s $\quad$ t_{char} = 10 years

What do models of postseismic and interseismic deformation tell us about the NAFZ plate boundary?



Could the NAFZ model work for the SAF?

- from Parkfield: shallow frictional afterslip occurs first (Johnson et al. 2006).

- rich history of earthquake cycle models.

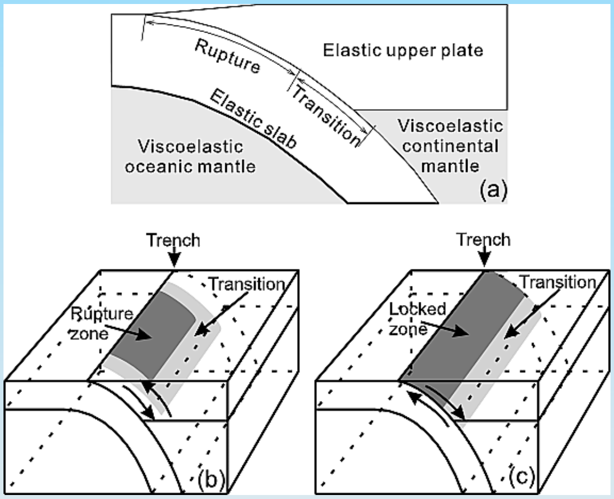
-- Rate-and-state frictional or viscous fault plus Maxwell substrate (Johnson et al., 2004; Li and Rice, 1986, and more).

---Models with nonlinear lower crust (Reches et al., 1993)

--- Many earlier / classic models (e.g. Savage and Prescott, 1978; Segall 2002, Thatcher 1983)

- no M 7.5's in the GPS era: if there is a transient or nonlinear response, it could be hard to see til one happens.

Subduction zones: similar models assumed though true viscoelastic earthquake cycles are lacking



Consensus that earthquakes, rapid afterslip and viscous flow in a wide viscoelastic shear zone at depth accommodate relative motion across fault zones

Unresolved:

Localized shear zone or broad scale viscoelastic flow in the lower crust?

Velocity-strengthening or viscous fault zone creep in the middle crust?

Rheology of the mantle?

Absolute stresses and friction along faults