Seismicity models based on static stress triggering

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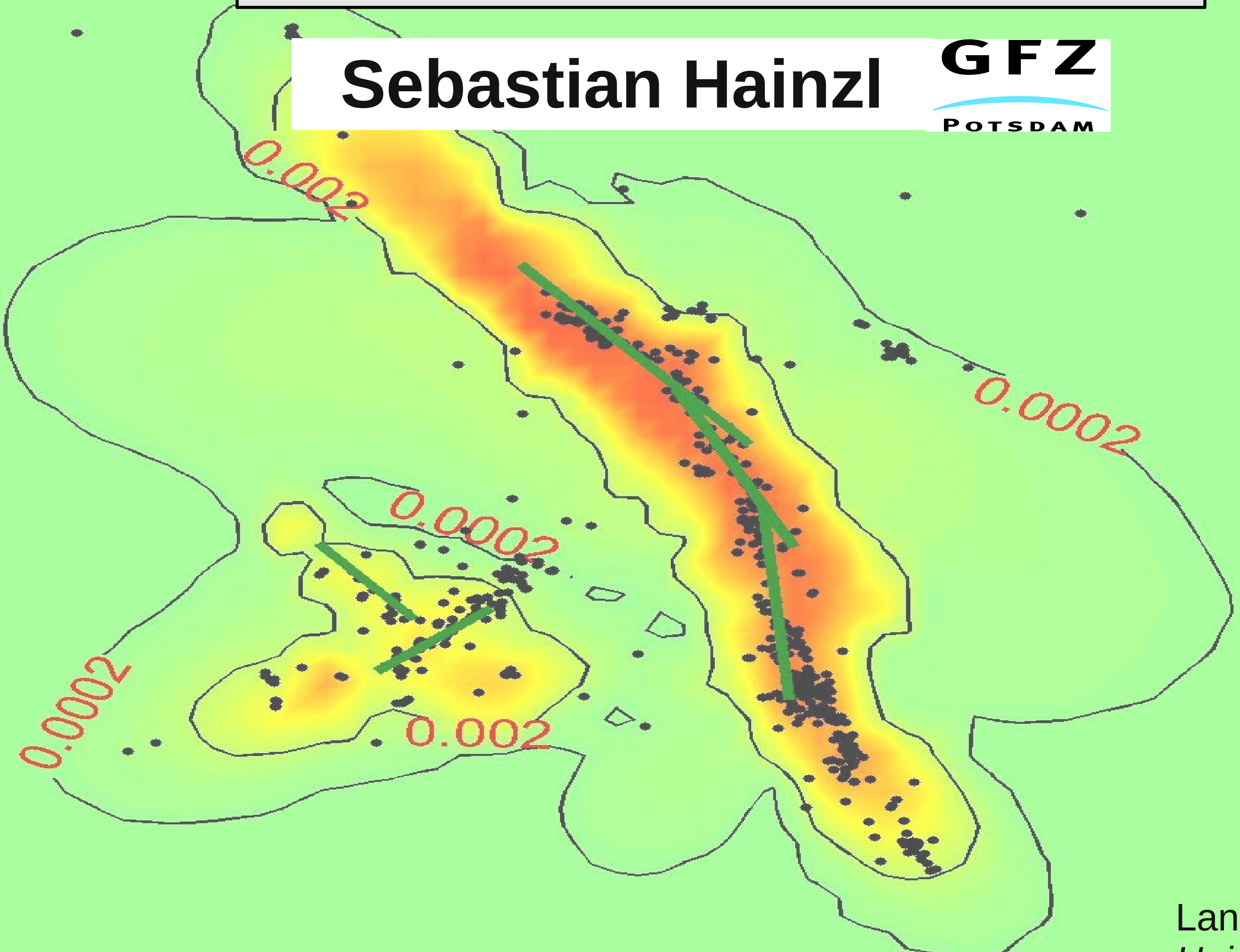
GFZ
POTSDAM



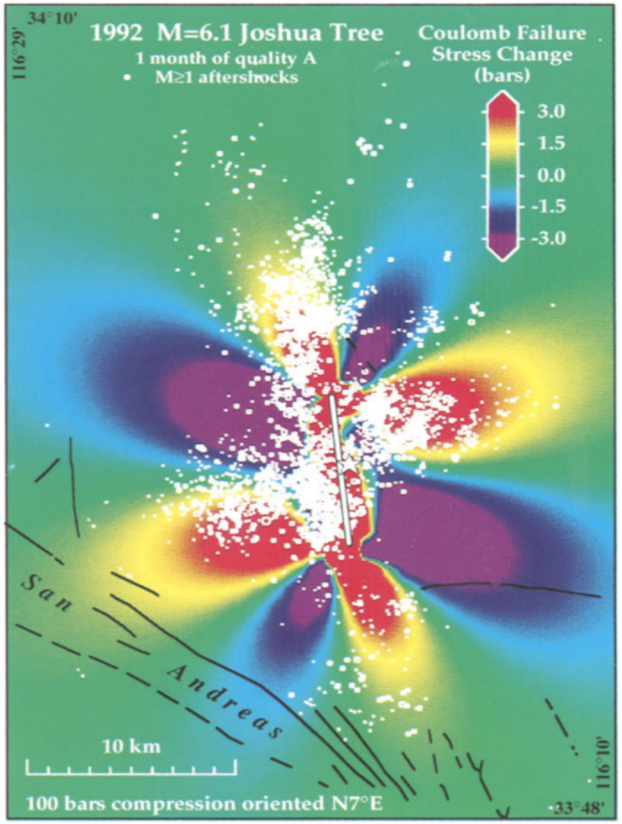
Landers sequence
Hainzl et al. JGR 2009

… what everybody knows:

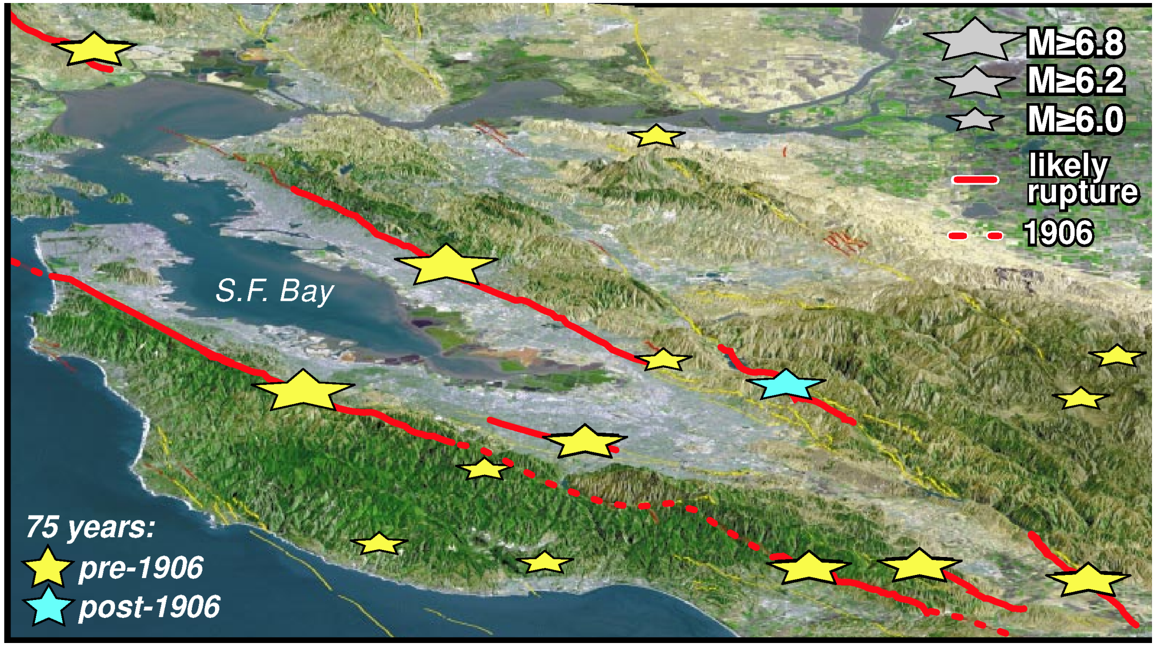
Colorful static stress maps showing generally some correlation between

- aftershock locations and positive stress lobes
- reduced activity and stress shadows

King et al. BSSA 1994



Stein Nature 1999



Goal of this lecture is to understand and discuss quantitative predictions of the static stress triggering model:

(1) Underlying assumptions

(2) Predictions of the model

(3) Potentials & Limitations (due to unkowns/uncertainties) of model applications

Outline:

(1) Neglect timing of earthquakes - A simple clock-advance model:

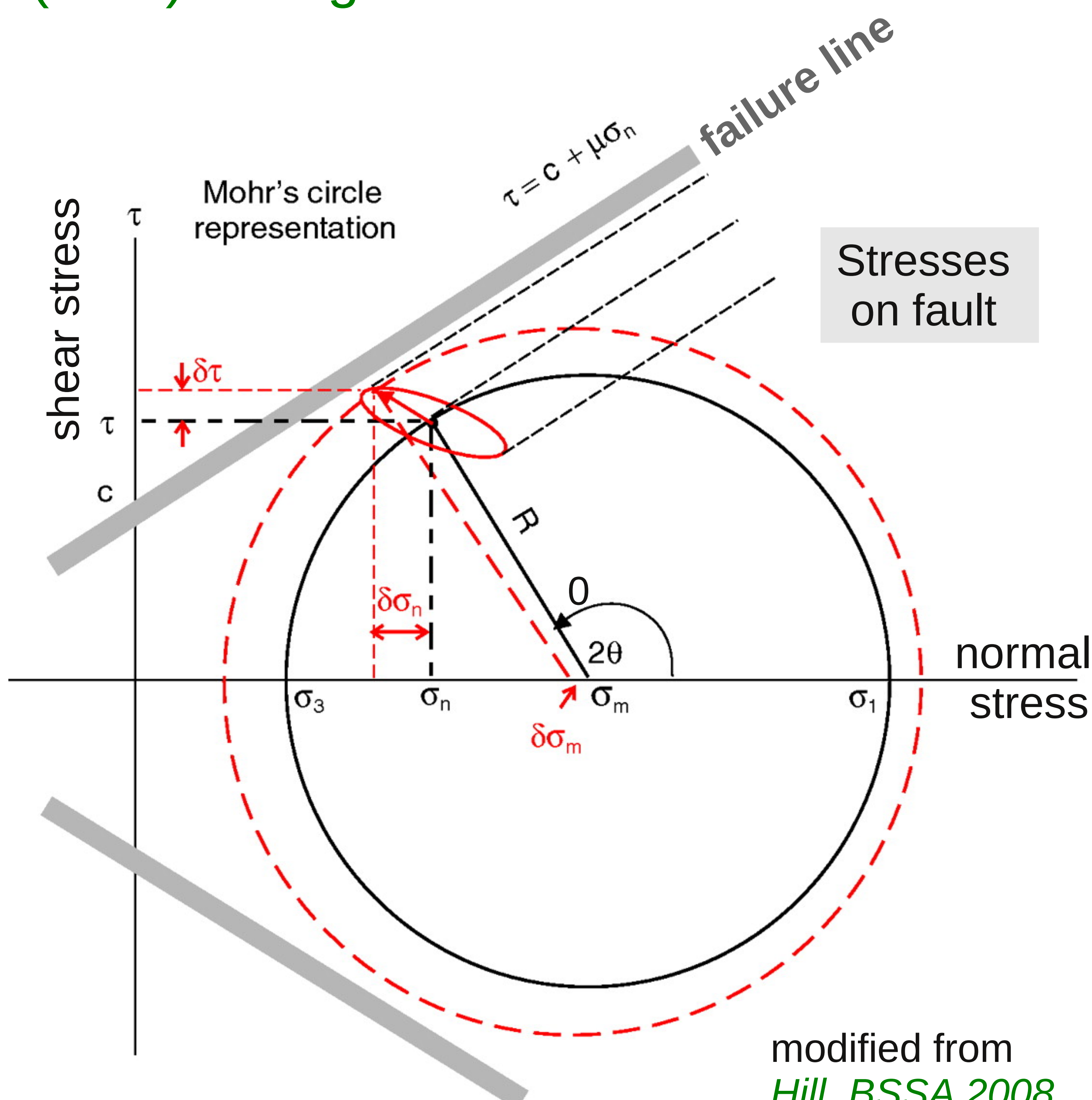
→ spatial aftershock distribution & decay
→ aftershock productivity

(2) Time-evolution assuming rate-state-dependent friction:

→ simple considerations
→ examples of model applications

Coulomb Failure Stress (CFS) change:

$$\Delta CFS = \Delta\tau - \mu\,\Delta\sigma_n$$



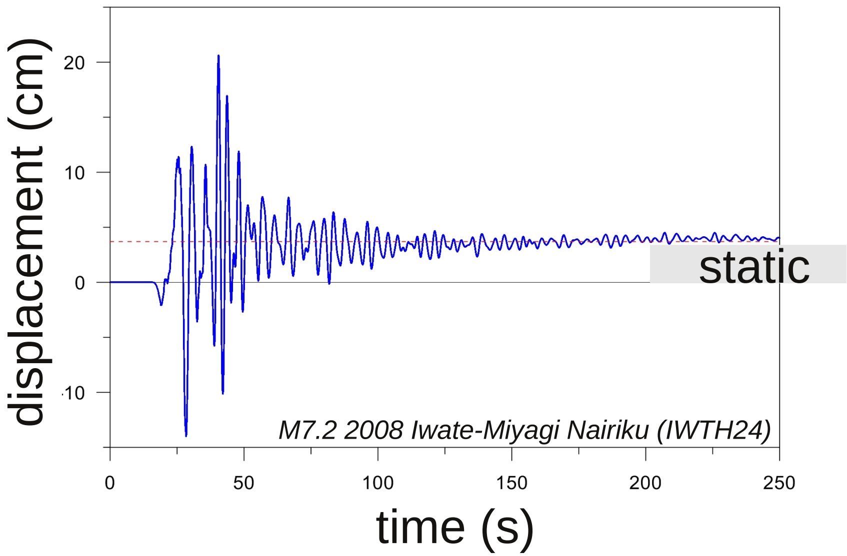
τ shear stress

σ_n normal stress (including pore pressure)

μ friction coefficient

modified from *Hill, BSSA 2008*

Origin of stress changes can be:



(1) **coseismic**:

- dynamic (seismic waves)
- static (permanent)

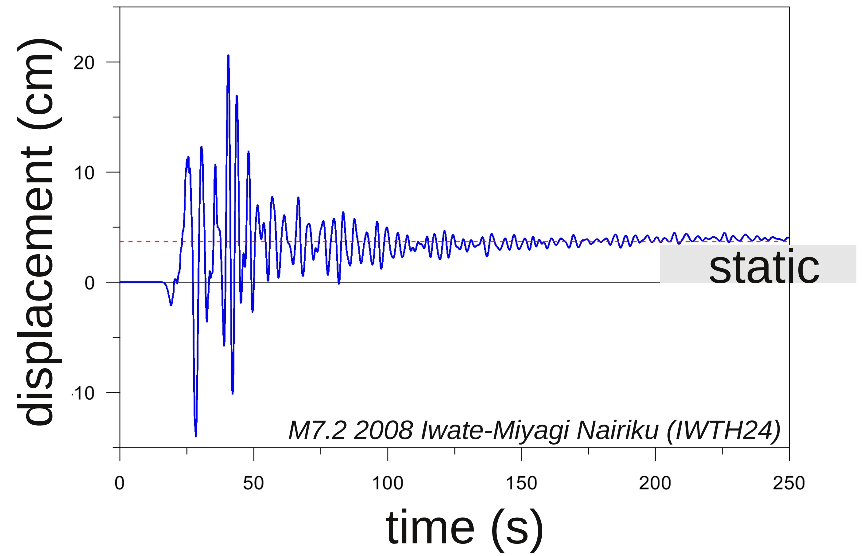
(2) **aseismic**:

- pore pressure changes
- continuous or transient creep
- visco-elastic deformation
- dike intrusion
- …

Origin of stress changes can be:

(1) **coseismic**:

- ~~dynamic (seismic waves)~~
- static (permanent)

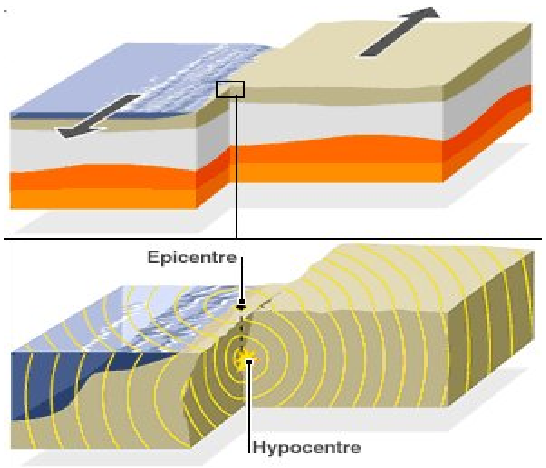


(2) **aseismic**:

- pore pressure changes
- continuous or transient creep
- visco-elastic deformation
- dike intrusion
- …

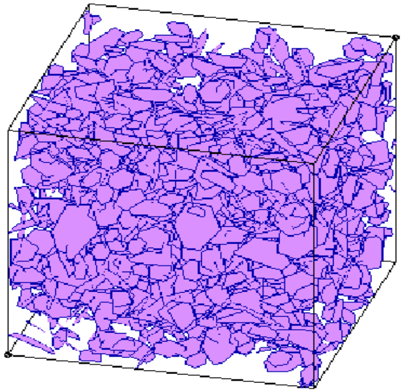
this talk

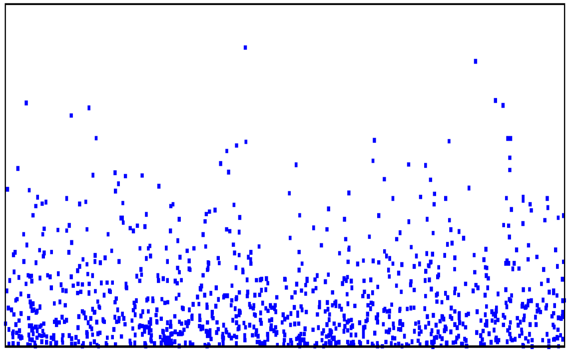
Let's start with simplest assumptions:



(1) Constant stress build-up
due to tectonic loading

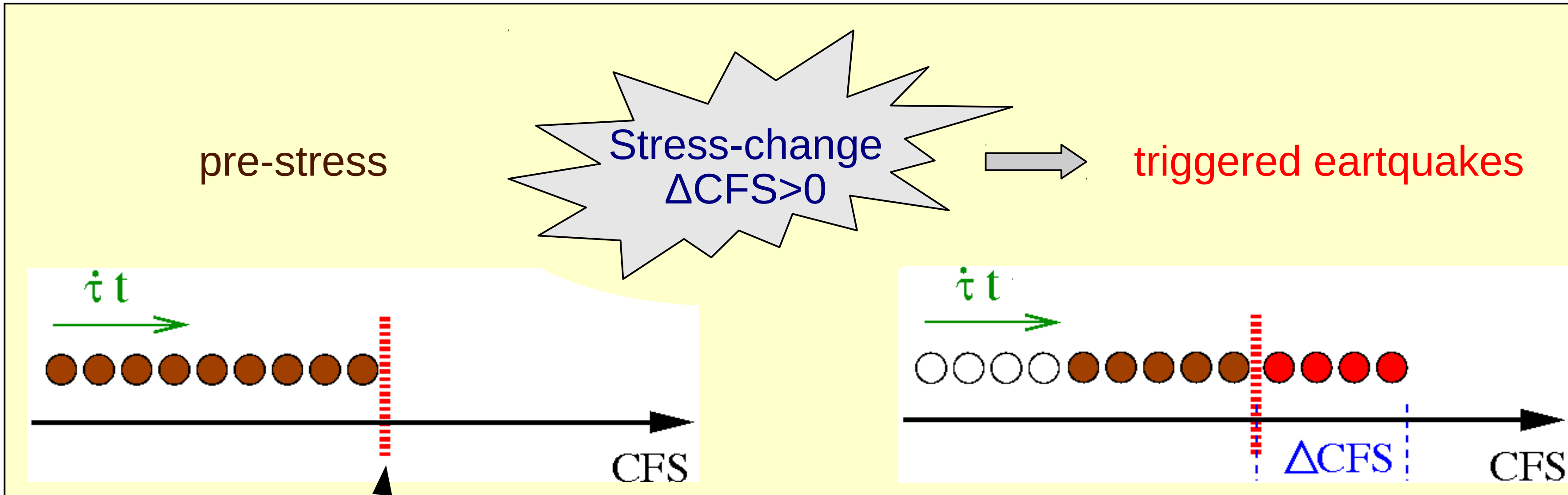
(2) Existence of a population of faults
where earthquake can occur





(3) **Without** any stress disturbance:
Earthquake rate is constant
→ Poisson model
(assumption of seismic hazard assessment)

Hainzl et al. JGR 2010



Critical stress at which earthquake nucleation starts

Clock advance - time needed to achieve ΔCFS by tectonic loading: $\Delta t = \Delta CFS / \dot{\tau}$

Number of triggered events: $N_a = r\,\Delta t = (r/\dot{\tau})\,\Delta CFS$ if ΔCFS>0
else $N_a = 0$

Number of triggered events ~ ΔCFS (if ΔCFS>0)

Proportionality factor: $c = \frac{r}{\dot{\tau}} \approx \frac{V}{\overline{M_0}}$

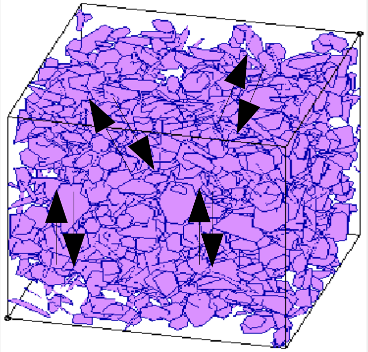
Ratio between seismogenic volume and the average seismic moment per earthquake

Kostrov 1974:

$$\dot{\tau} = 2G\dot{\epsilon}_{ij} = \lim_{\Delta t \to \infty} \frac{1}{V \Delta t} \sum_{k=1}^{N} M_{0ij}^{k}$$

$$= \lim_{\Delta t \to \infty} \frac{N}{V \Delta t} \overline{M_{0ij}} = \frac{r}{V} \overline{M_{0ij}}$$

$$\approx \frac{r}{V} \overline{M_0}$$



Assumption: Similar earthquake mechanisms

$N_a = (V / \overline{M}_0) \; \Delta CFS$

For Gutenberg-Richter distributed earthquakes within [M_{min}, M_{max}]:

$$pdf(M) = \ln(10)\, b \frac{10^{-b(M-M_{min})}}{1-10^{-b(M_{max}-M_{min})}}$$

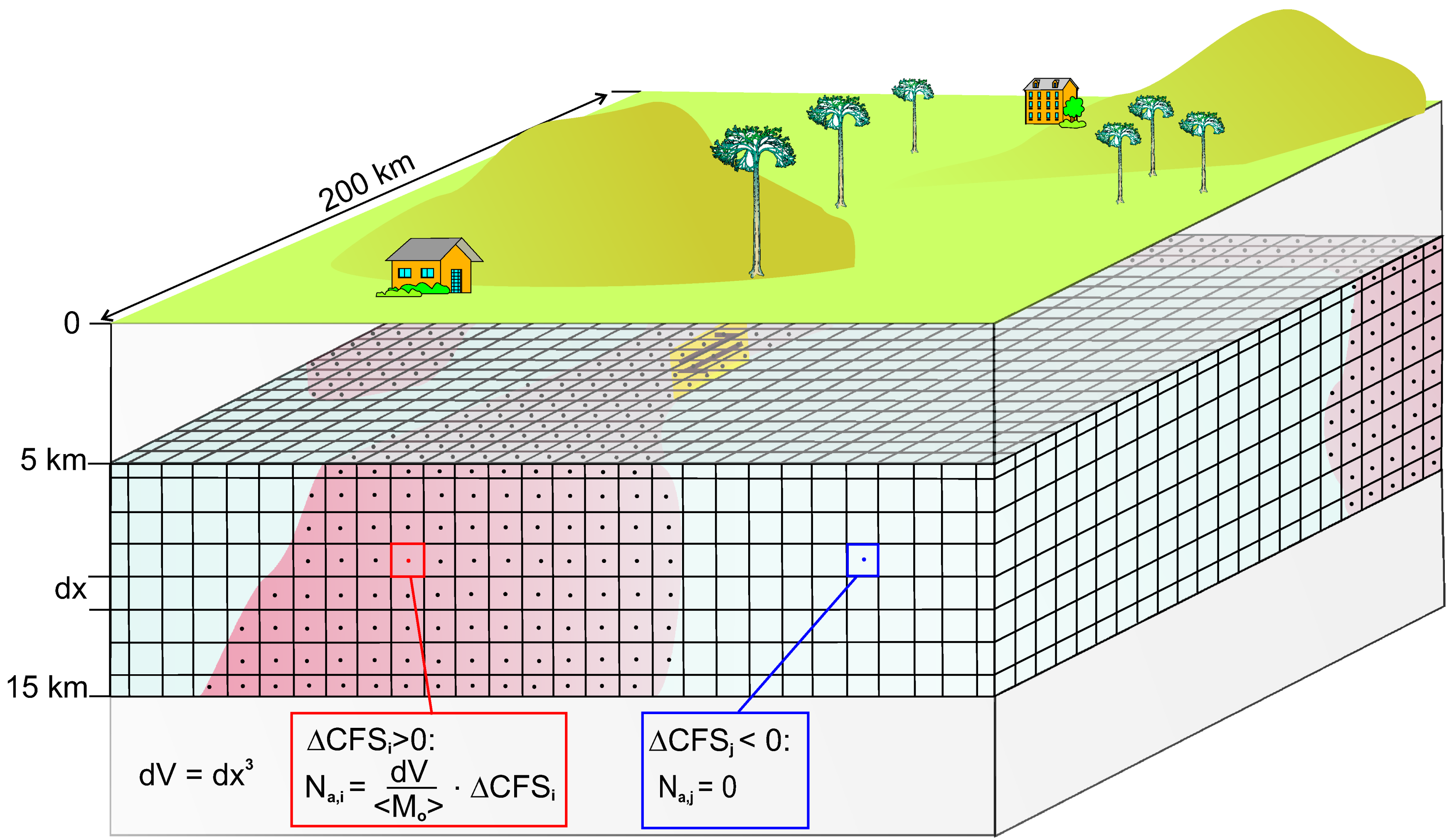
Average seismic moment per earthquake (NO time average!):

$$\overline{M}_0 = \int_{M_{min}}^{M_{max}} pdf(M)\, 10^{9.1+1.5M}\, dM$$

$$= 10^{9.1+1.5M_{min}} \frac{b}{1.5-b} \frac{10^{(1.5-b)(M_{max}-M_{min})}-1}{1-10^{-b(M_{max}-M_{min})}}$$

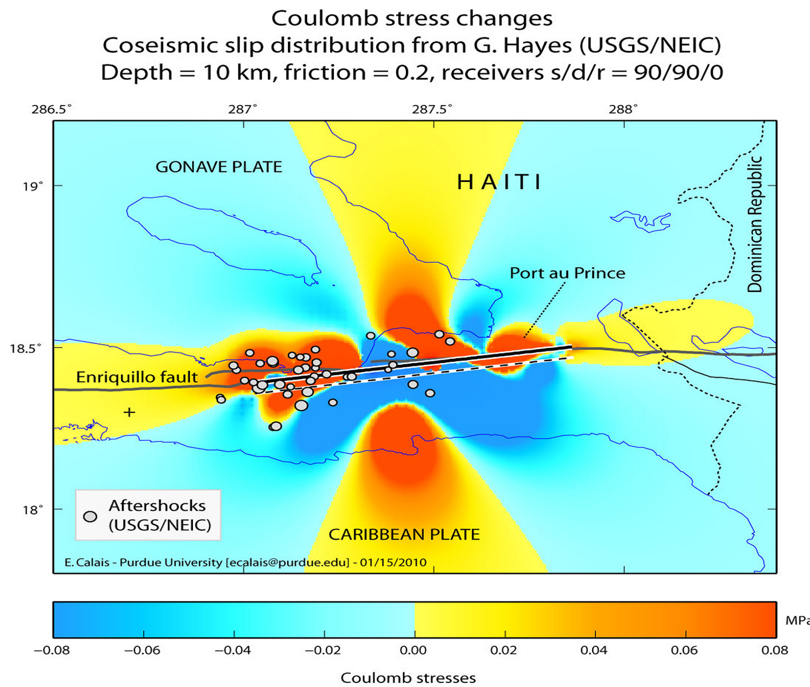
Proportionality factor $c = V/\overline{M}_0$
can be calculated for given b and M_{max} (M_{min})

$N_a(x) \sim \Delta CFS(x)$ (for $\Delta CFS>0$)



Model Prediction:

No earthquakes in regions
with ΔCFS<0 (stress shadows)

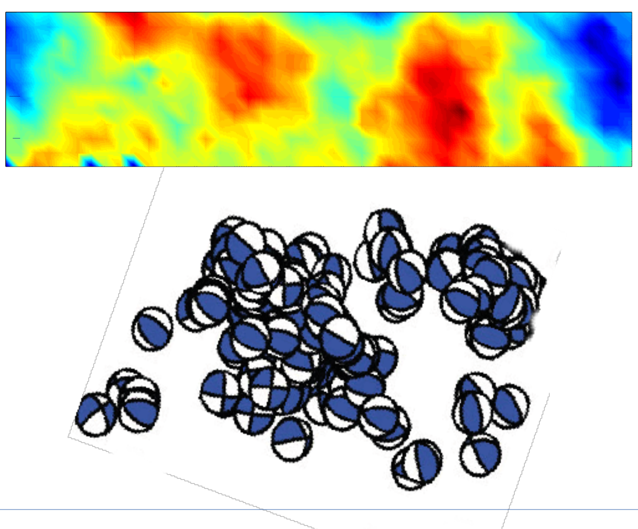


Observation:

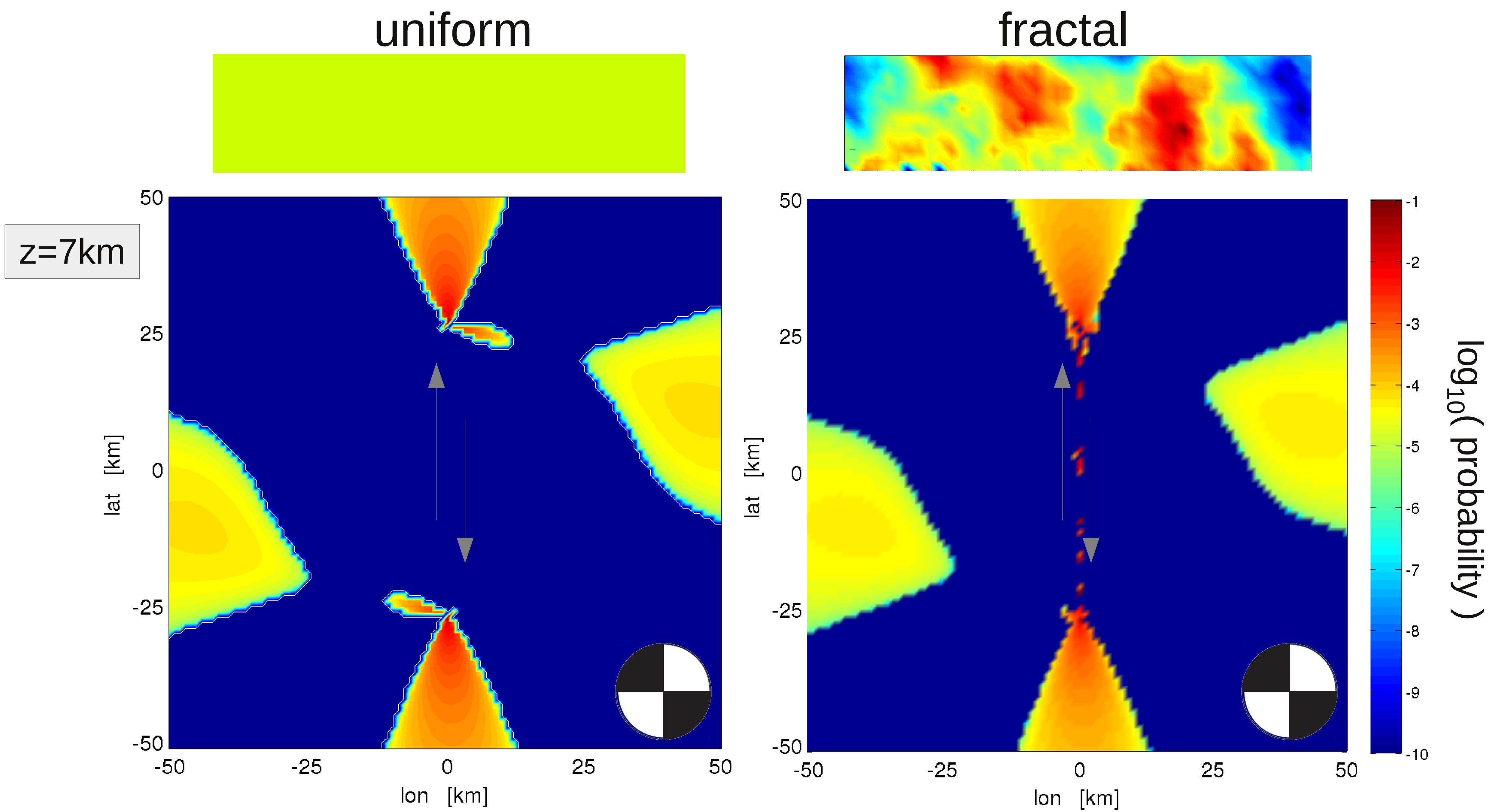
A significant fraction of earthquakes occur in stress shadows ...
is static stress triggering not working?

Not necessarily because of:

1. uncertainty of slip inversions
2. unresolvable small scale slip
3. variable receiver mechanisms
4. secondary stress changes



Small scale slip variability

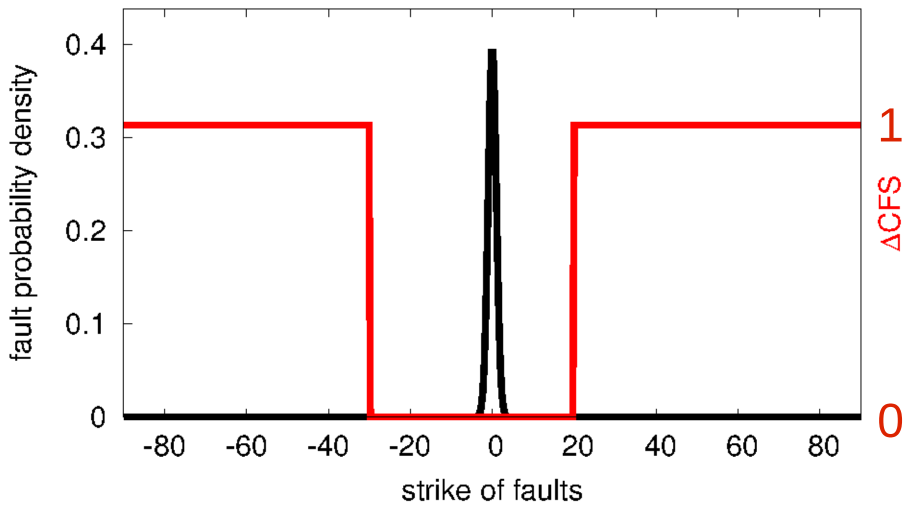


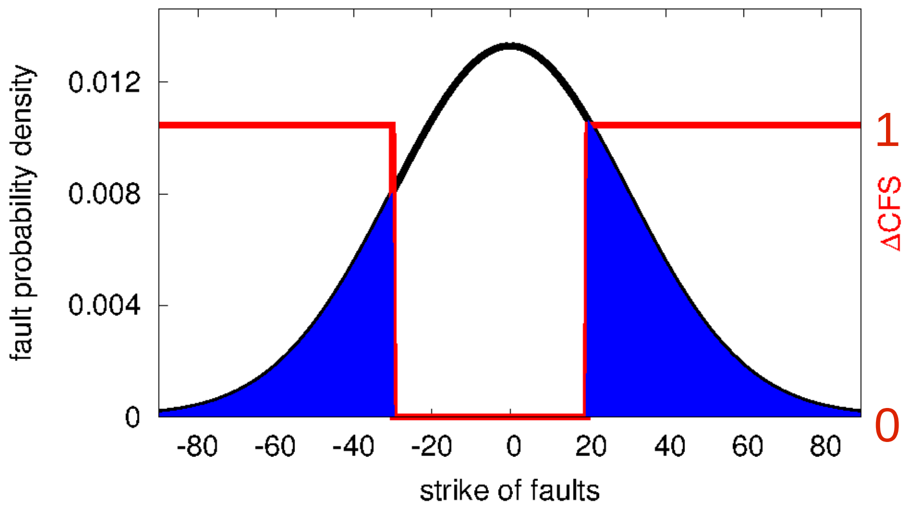
→ can explain on-fault activation

Variable receiver mechanisms

Faults with different orientations always exists

Examples:



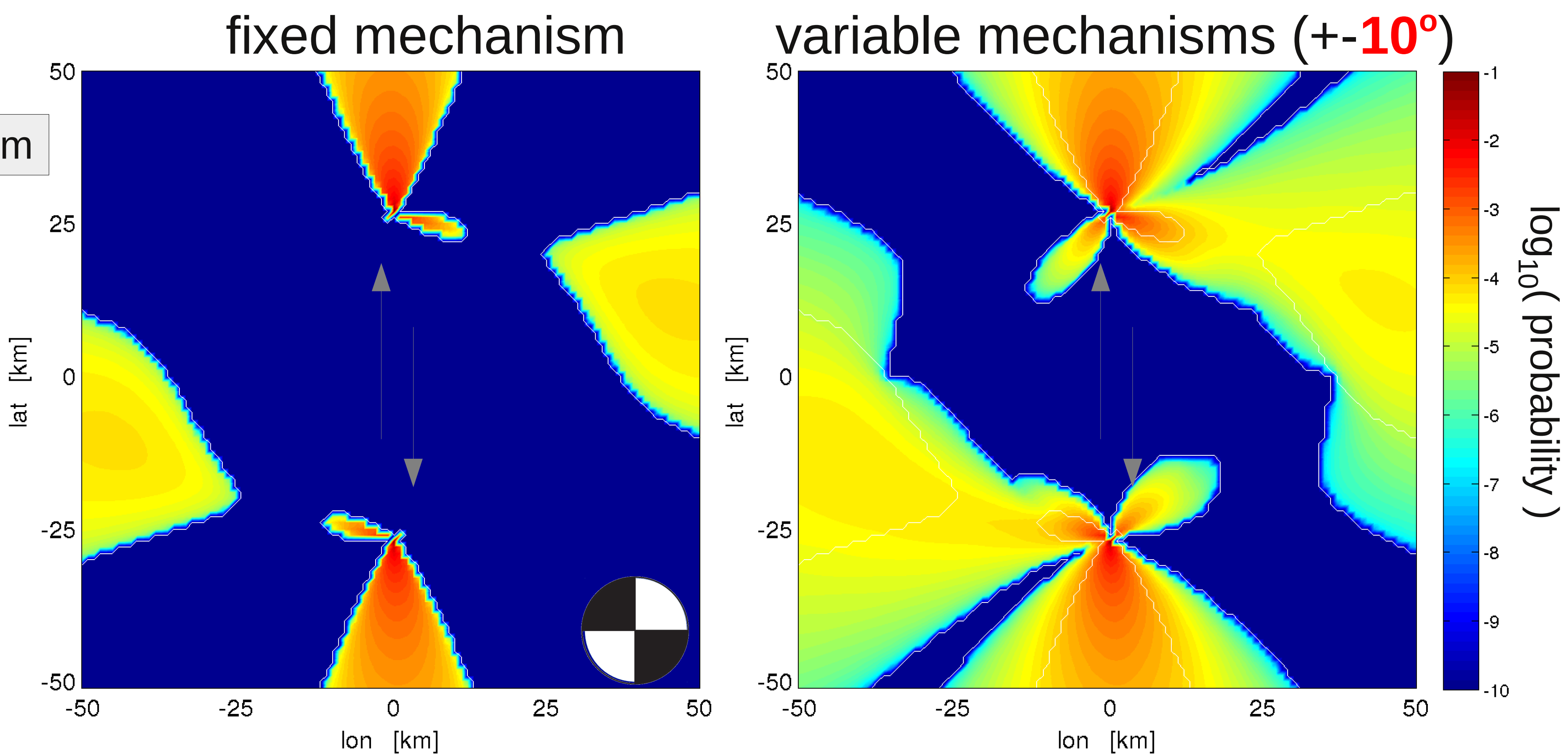


“unique” mechanism: ΔCFS<0
→ no aftershocks

mean mechanism: ΔCFS<0
but some faults have ΔCFS>0
→ aftershocks ~ blue area

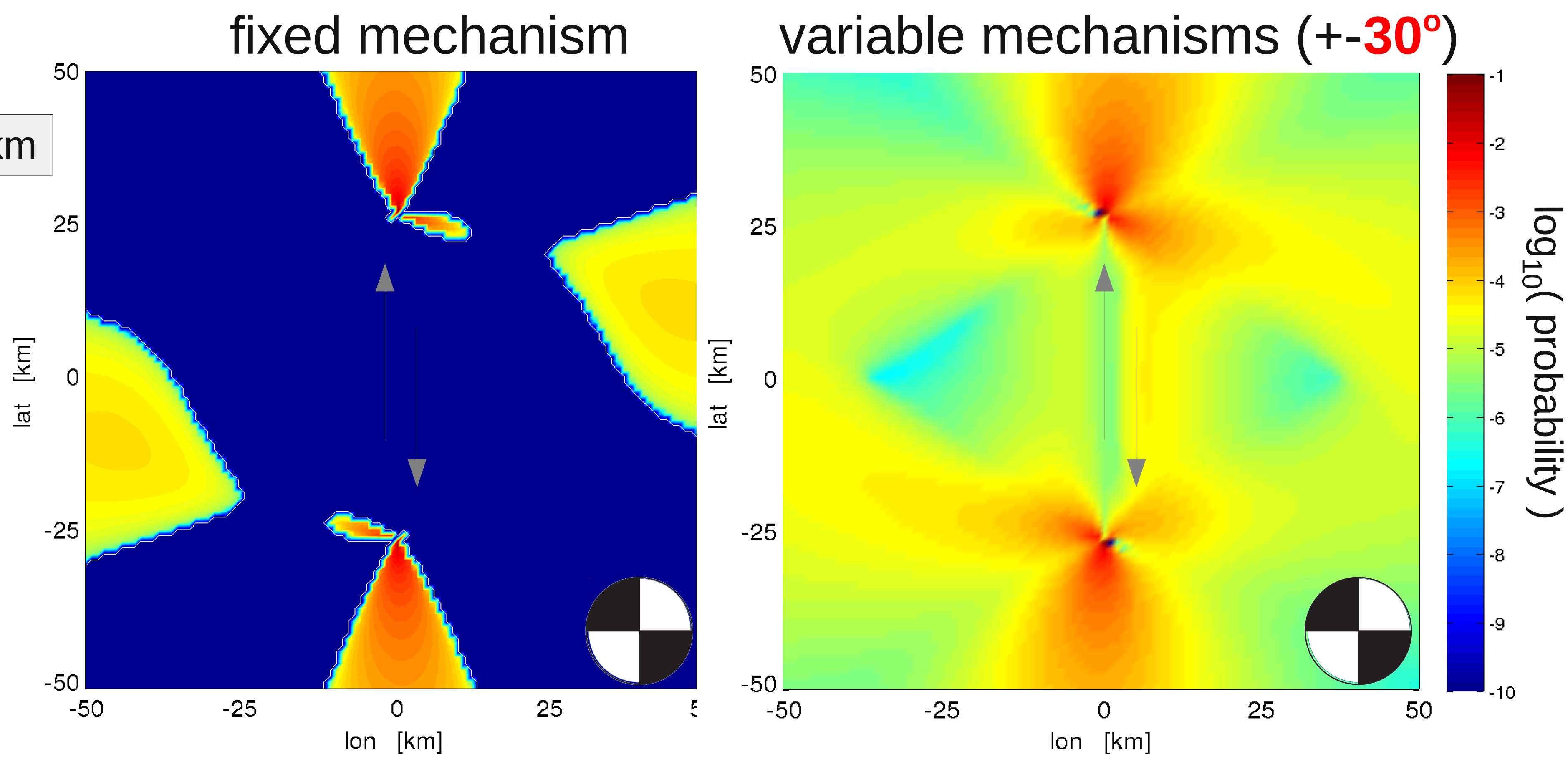
Variable receiver mechanisms

z=7km



Variable receiver mechanisms

z=7km



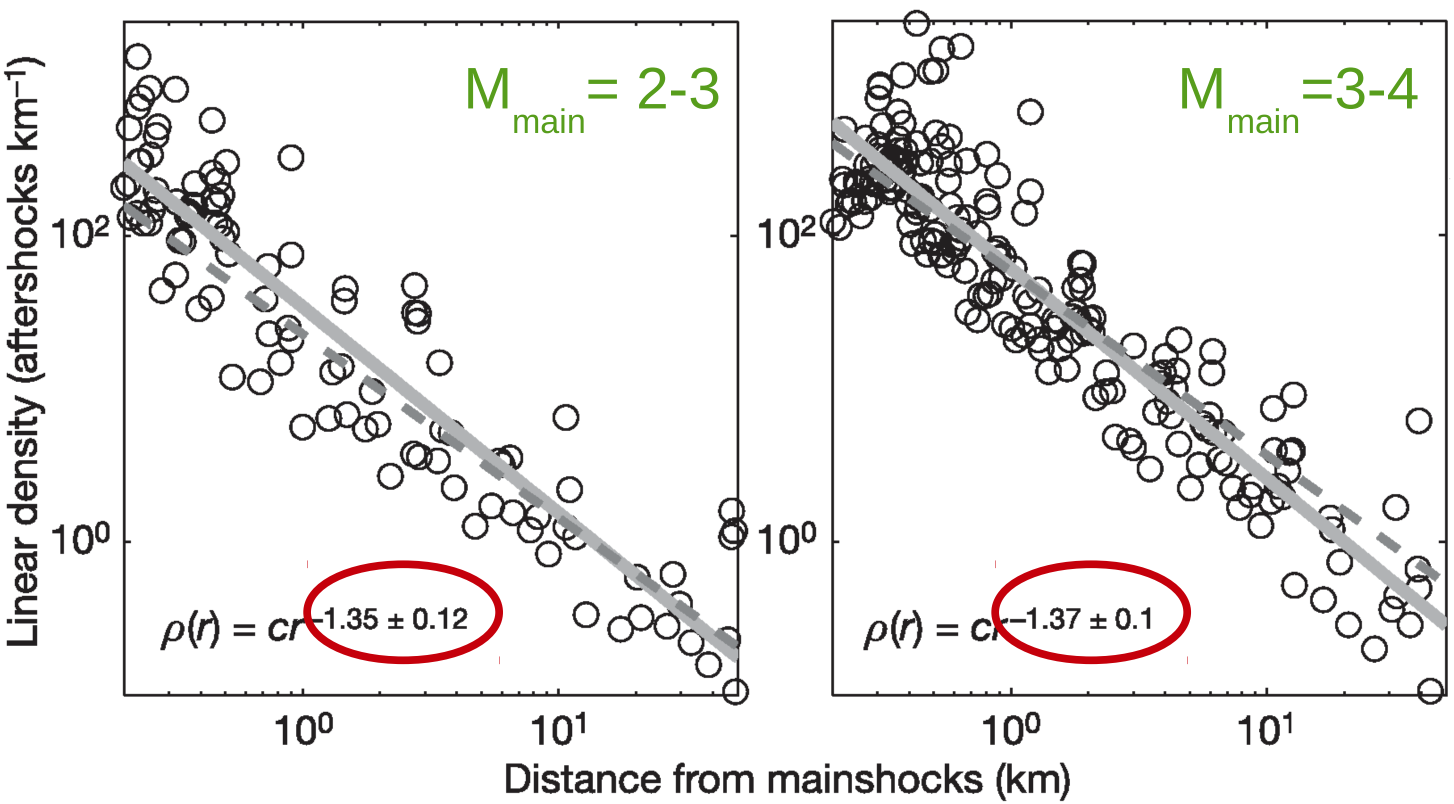
NO absolute stress shadows, only regions with reduced activation!

Distance decay

Theory (homogeneous elastic full-space):

Far-field: dynamic stress ~ $1/r^2$
static stress ~ $1/r^3$

Felzer & Brodsky, Nature 2006:



Static stress cannot be the driving force

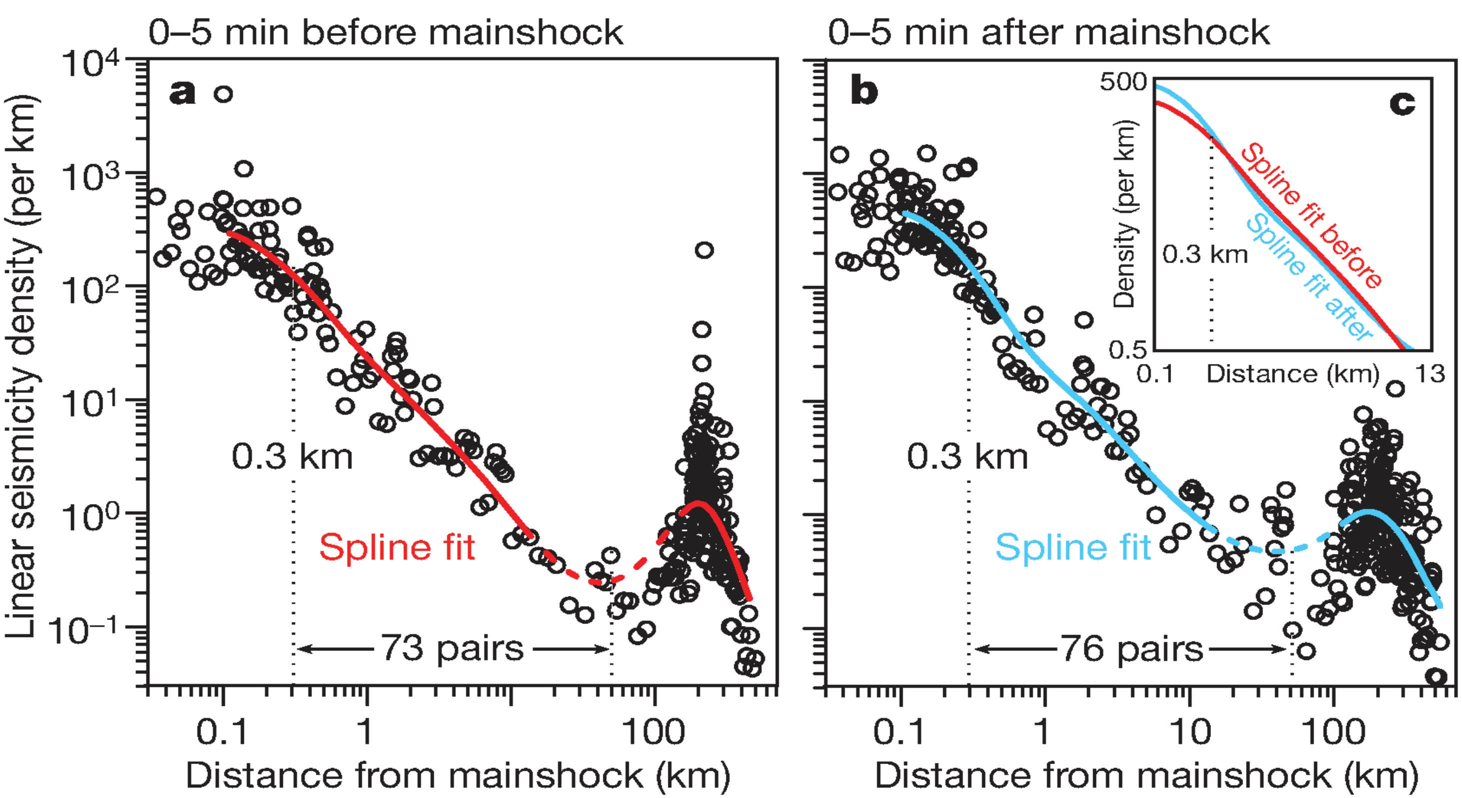
Distance decay

Theory for homogeneous elastic full-space:

Far-field: dynamic stress ~ $1/r^{2}$

static stress ~ $1/r^{3}$

Richards-Dinger et al., Nature 2010:



improper selection of “aftershocks” does not allow this conclusion

Distance decay

Theory for homogeneous elastic full-space:

Far-field: dynamic stress ~ $1/r^{2}$
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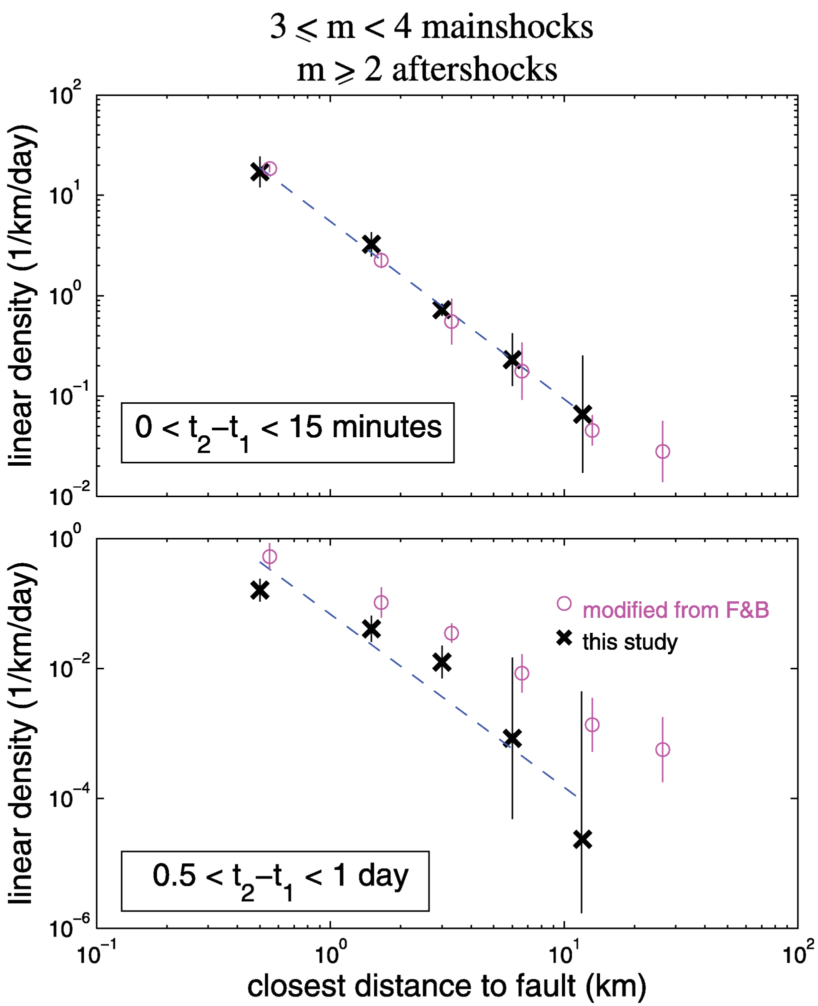


Marsan & Lengliné, JGR 2010:

… with an improved statistical attempt to isolate earthquakes causally related to the mainshock:

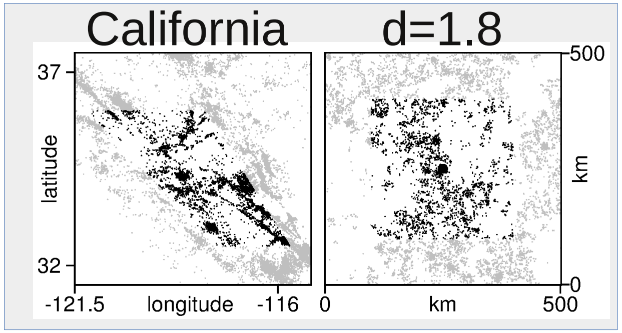
Exponent = 1.7-2.1

Moradpour et al. JGR 2014

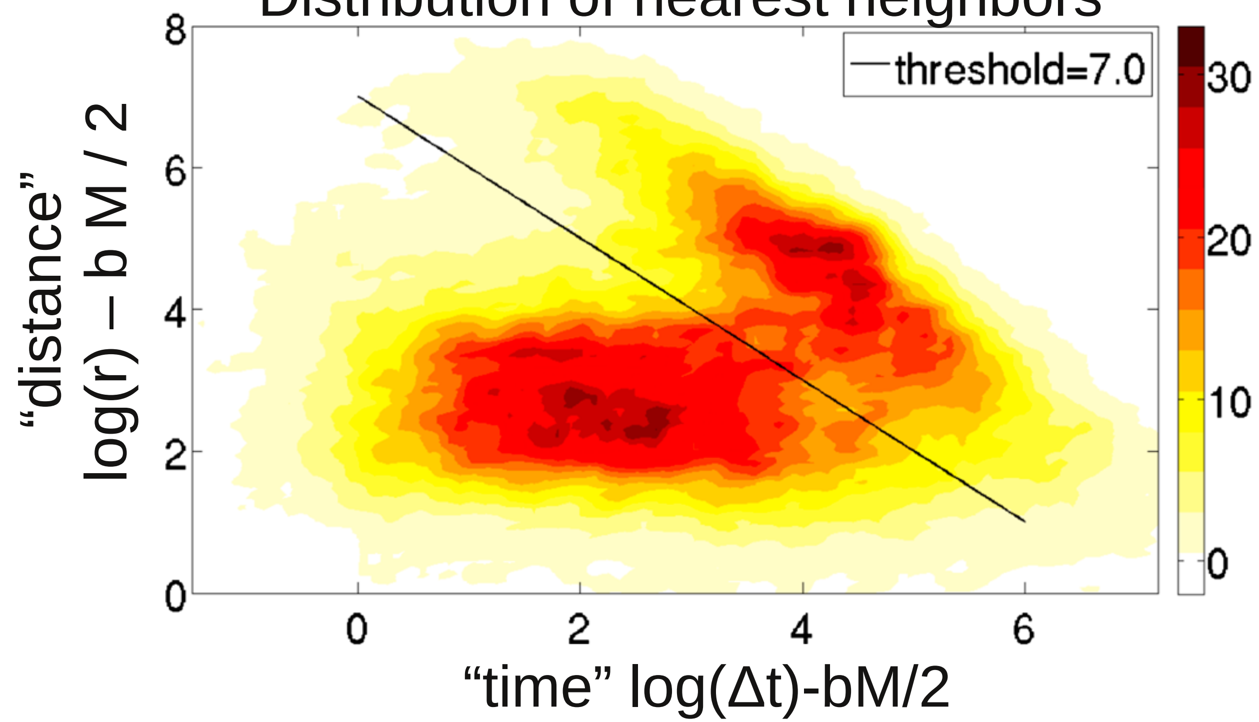
Separation based on the smallest space-time distance

$$n_{ij} \sim \Delta t\, r^{d}\, 10^{-bM}$$

between an event and all preceding events, where d is the assumed fractal dimension.



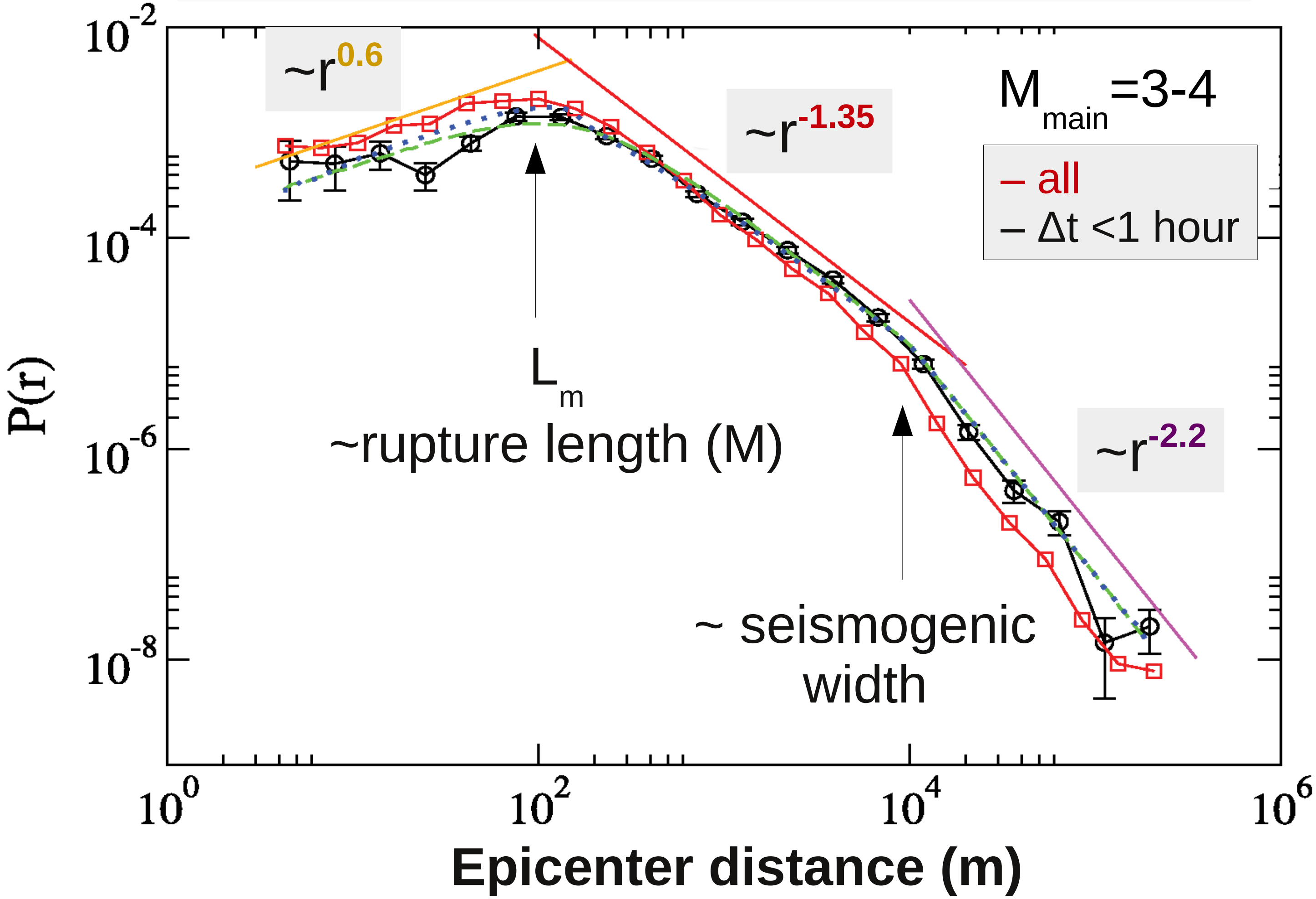
Distribution of nearest neighbors



Method based on
Biasi & Paczuski 2004
Zaliapin et al. 2008

$$P_m(r) = \begin{cases} \alpha \dfrac{q r^{\gamma}}{L_m^{\gamma+1}\left(\dfrac{r^{\gamma+1}}{L_m^{\gamma+1}}+1\right)^{1+\frac{q}{\gamma+1}}} & \text{if } r < 10\text{km}, \\ \beta \dfrac{d r^{\gamma}}{L_m^{\gamma+1}\left(\dfrac{r^{\gamma+1}}{L_m^{\gamma+1}}+1\right)^{1+\frac{d}{\gamma+1}}} & \text{if } r > 10\text{km}. \end{cases}$$

γ=0.6
q=0.35
d=1.2

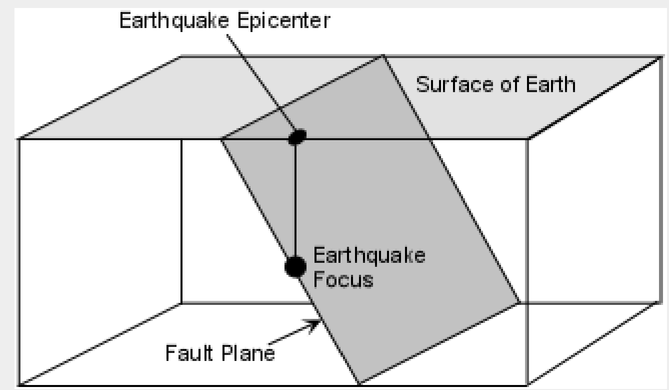


3 scaling regimes

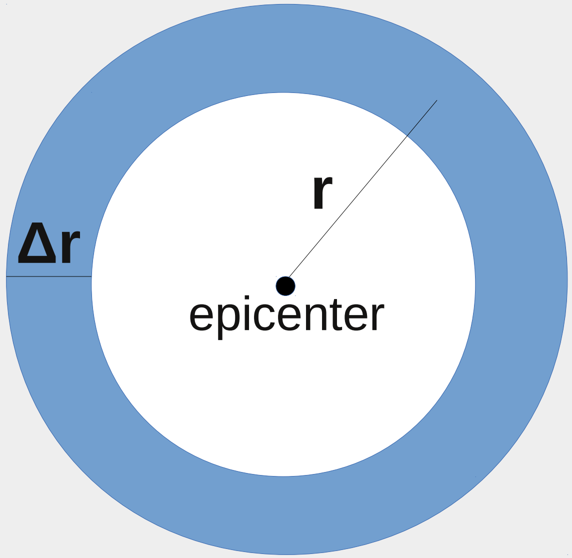
Southern California

… from empirical observations back to static stress-triggering:

Analysis for synthetic mainshock ruptures:



- select magnitude M
- uniform (or fractal) slip
- empirical relations between M and slip/area
- epicenter randomly chosen within rupture area
- effective friction coefficient: $\mu = 0.5$
- elastic half-space (or layered half-space)



$A \sim (r+\Delta r/2)^d - (r-\Delta r/2)^d$

A: seismogenic area at distance r
d: fractal fault dimension (uniform: d=2)

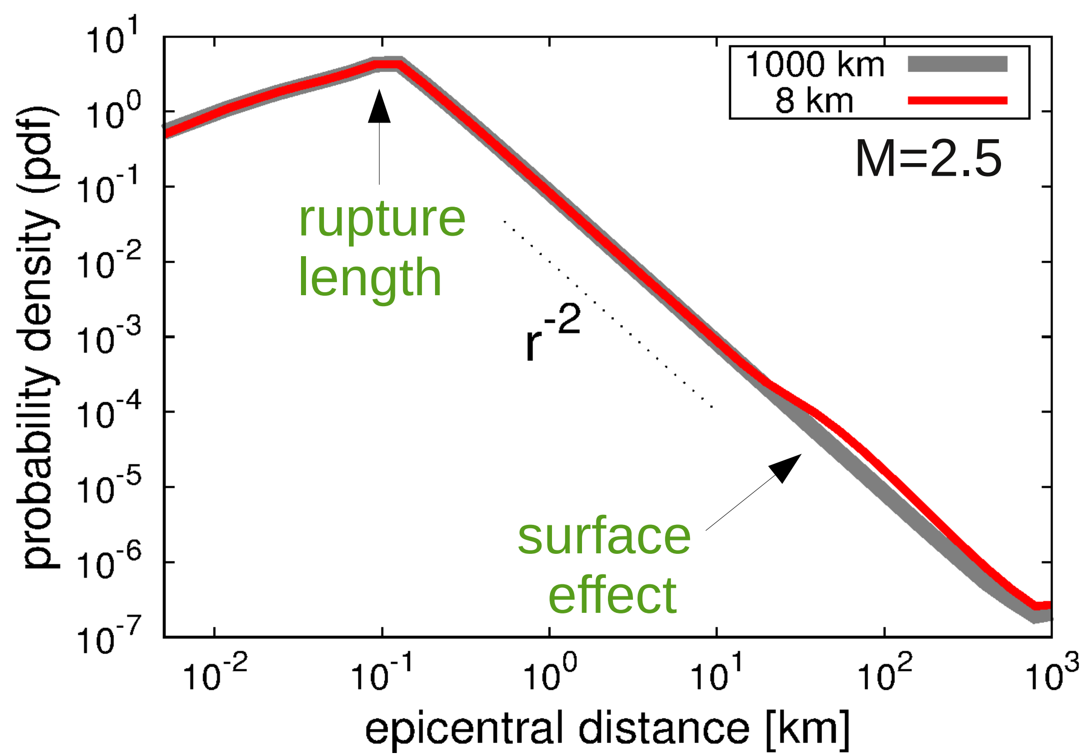
Linear aftershock density:

$P(r) \sim \overline{\Delta CFS(r)\, H(\Delta CFS)}\;\; A / \Delta r$

Aftershocks in the <u>hypocenter</u> depth layer:

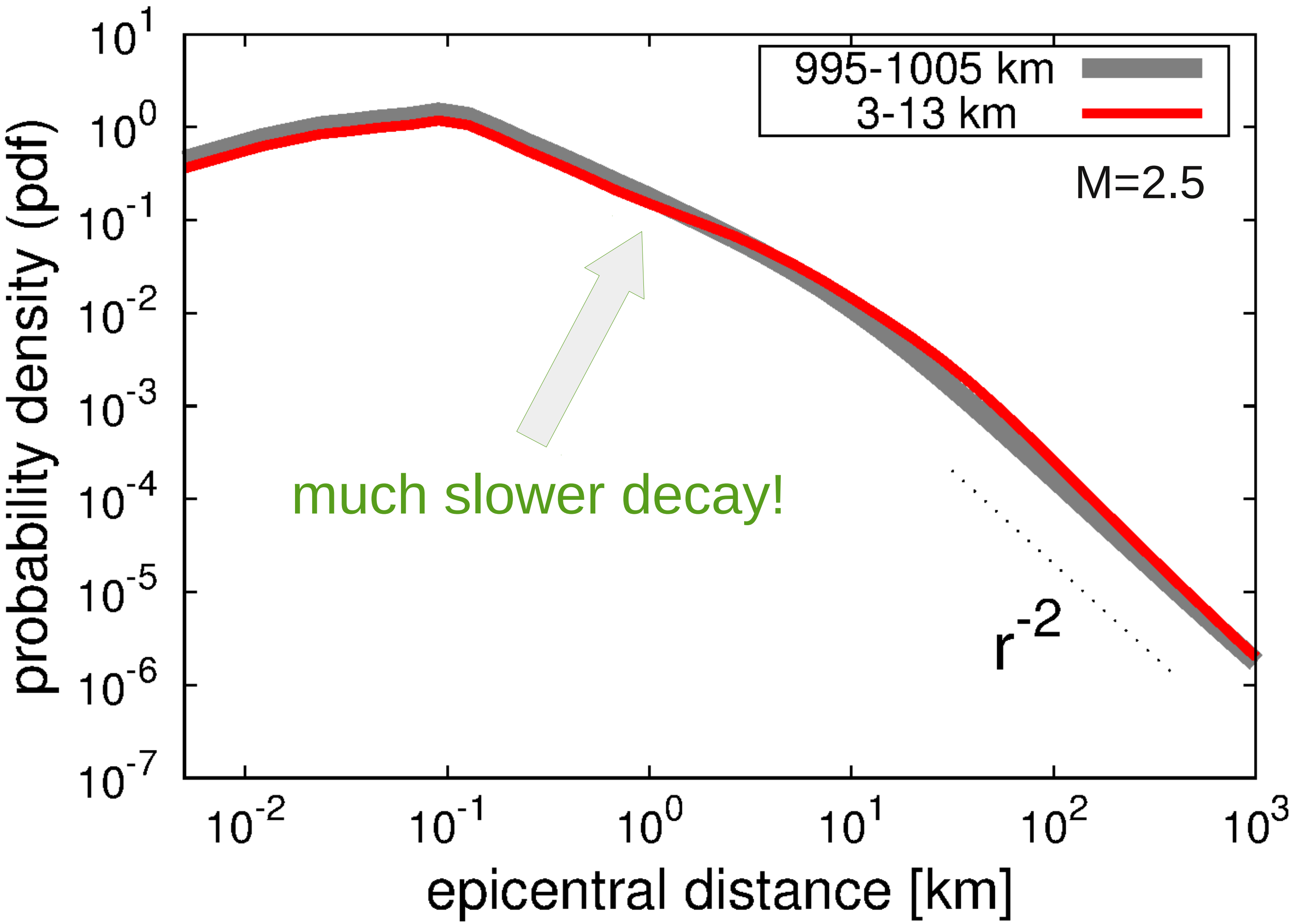
Far field: $\Delta CFS \sim r^{-3}$
$A \sim r$

$\rightarrow \;\; N_a \sim r^{-2}$

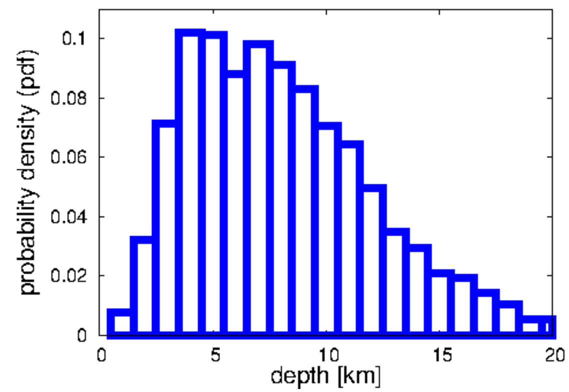


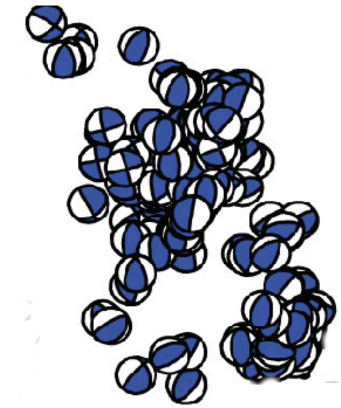
… but aftershocks occur not only at the hypocenter depth layer:

Integration over depth interval

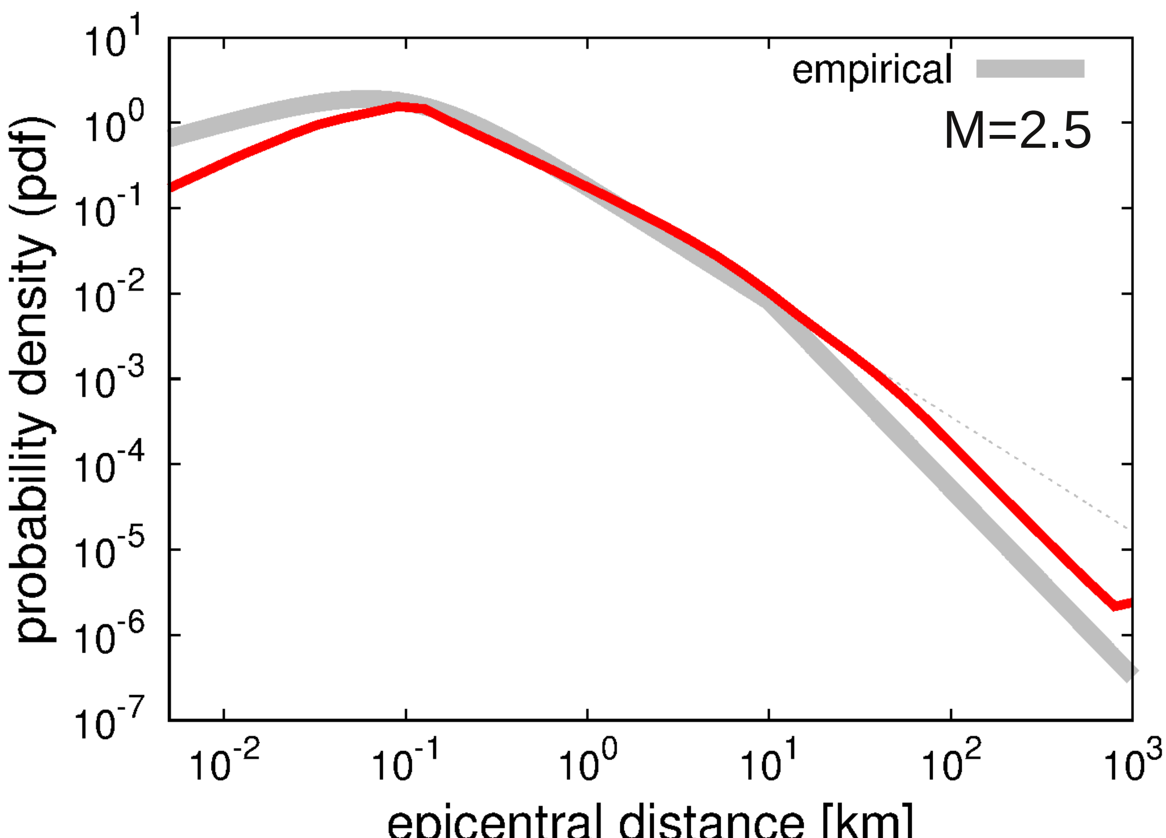


Adaption to Southern California seismicity by random selection of depth & focal mechanism



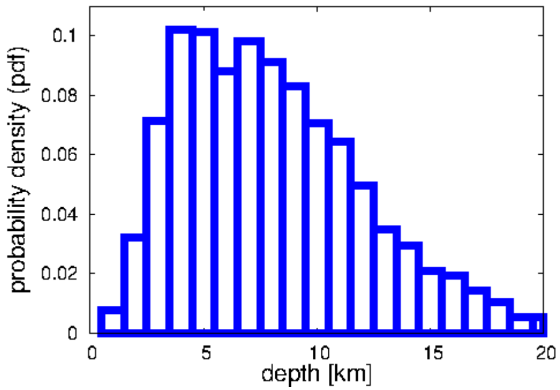


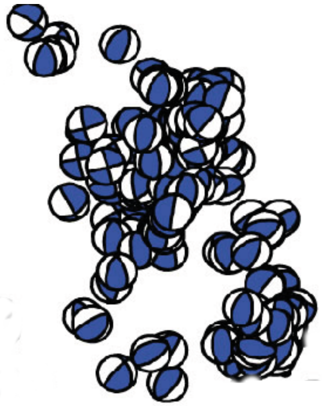
Yang, Hauksson & Shearer (2012)



adapted model forecast
d=2, uniform

Adaption to Southern California seismicity by random selection of depth & focal mechanism





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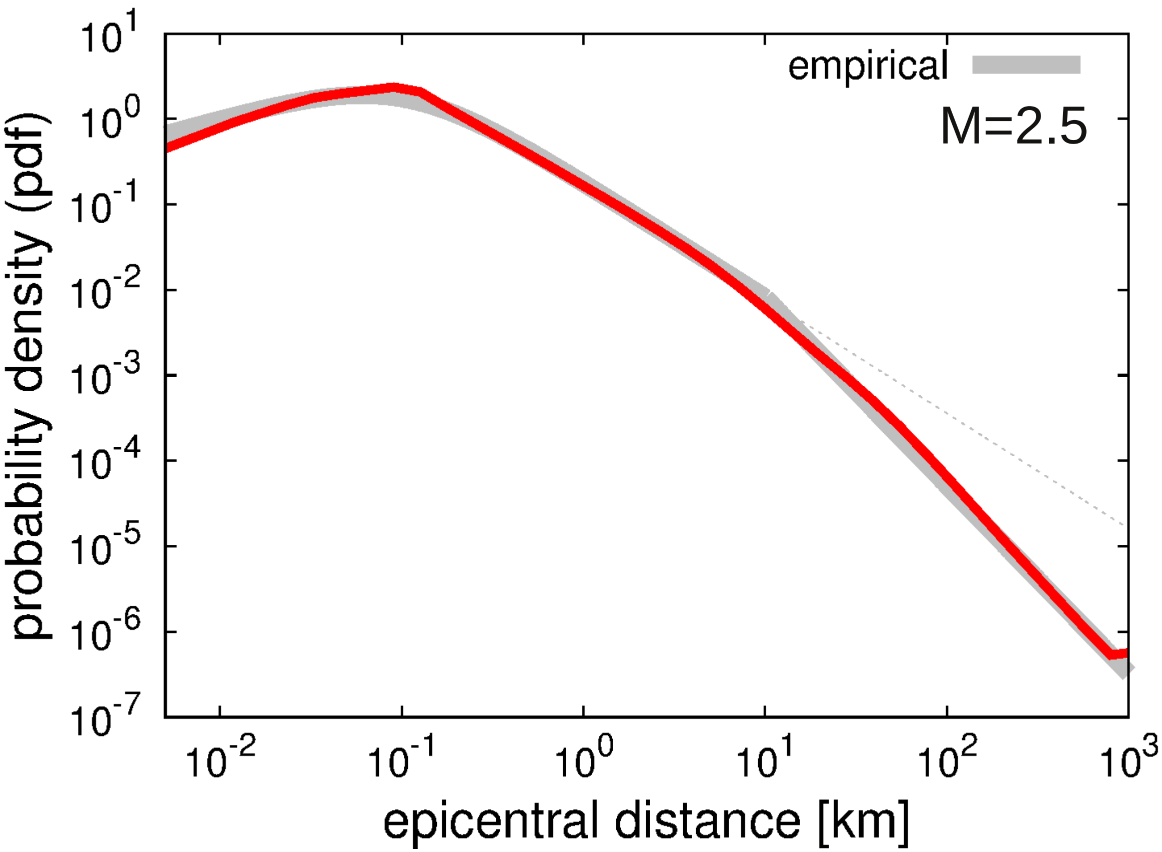
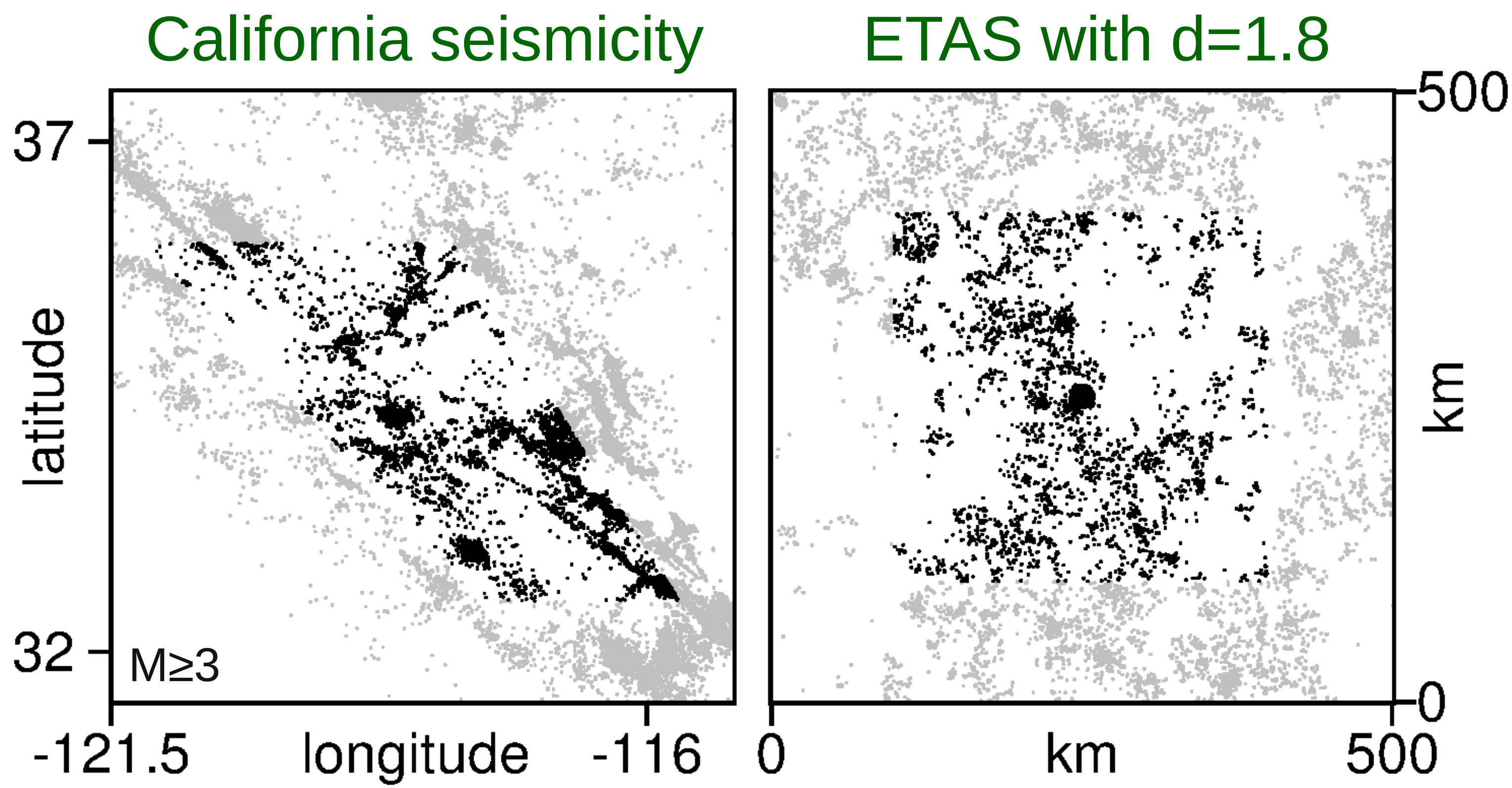


adapted model forecast
d=1.8, fractal

Independent evaluation by means of the ETAS model:

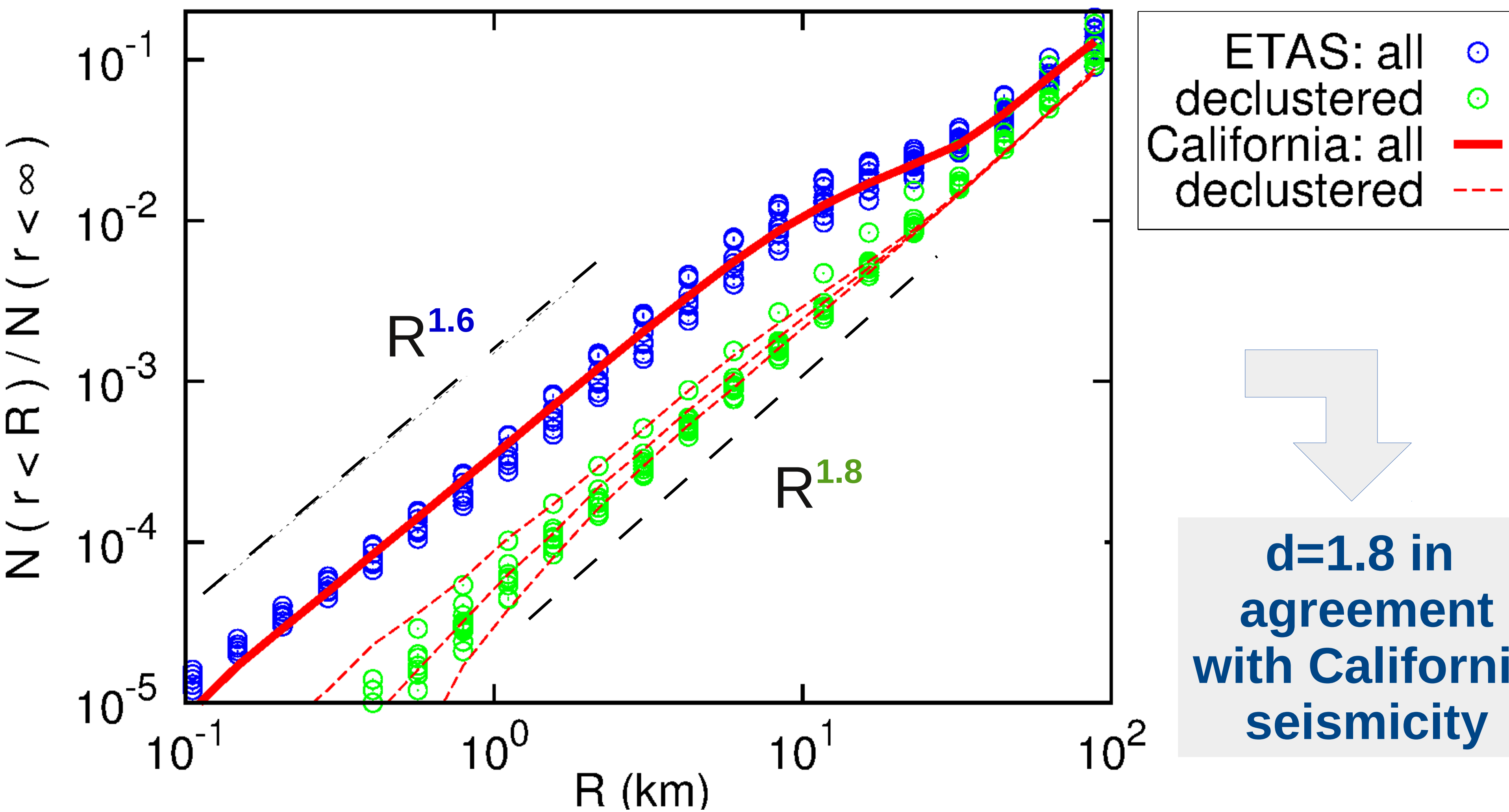
Is d=1.8 a reasonable value for background activity?



Method based on
Beauval et al., BSSA 2006

Scaling of the correlation integral:

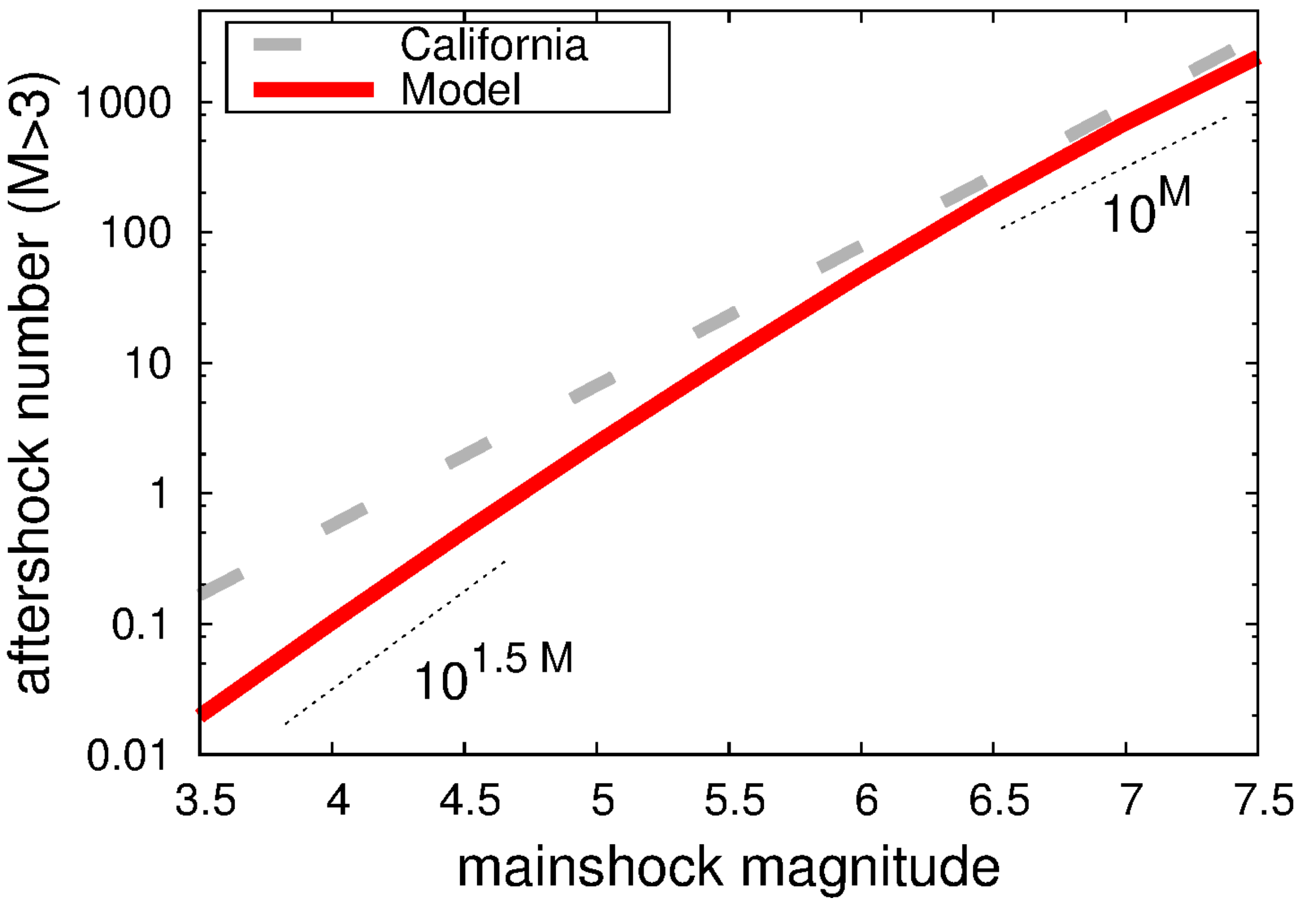
scaling exponent = fractal dimension



d=1.8 in agreement with California seismicity

Productivity of aftershocks

$$N_a = \int_0^\infty \overline{\Delta CFS(r) H(\Delta CFS)} \; A(r) \; W \; / \; \overline{M_0} \, dr$$

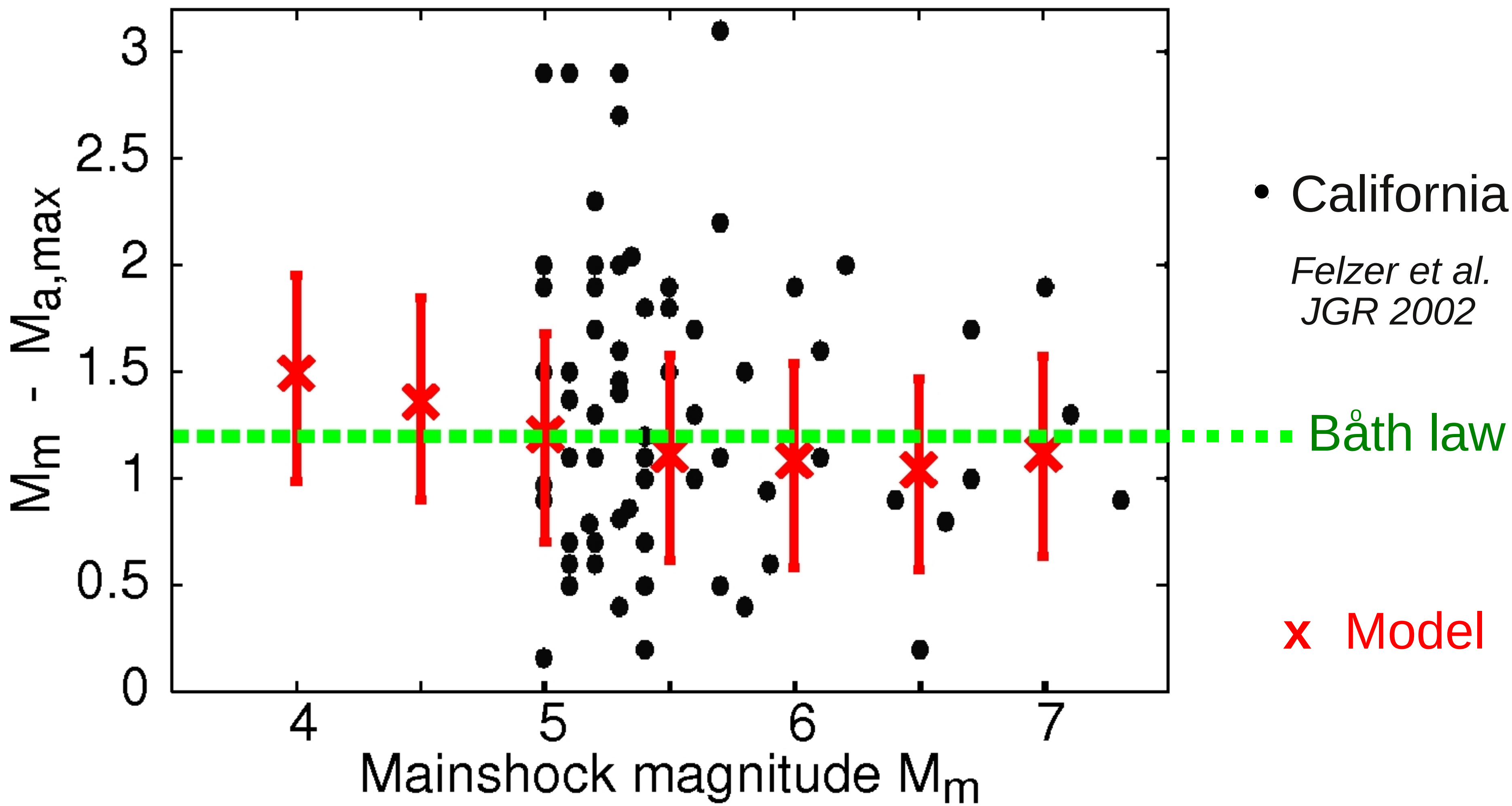


W = 10 km
b = 1.0
M_{min} =-2.0
M_{max}= 7.5

California: *Helmstetter et al., JGR 2005*

Largest aftershock magnitude as another way to compare with observations:

Empirical observation: **$\langle M_{a,max} \rangle = M_m - 1.2$** (Bath, 1965)



… thus static stress triggering can explain first-order aftershock characteristics regarding total numbers and spatial distribution.

→ static stress changes seem to be the major driving force for aftershock nucleation!

However, the model misses so far the
timing of aftershocks
and the contribution of aftershock induced
secondary stress changes

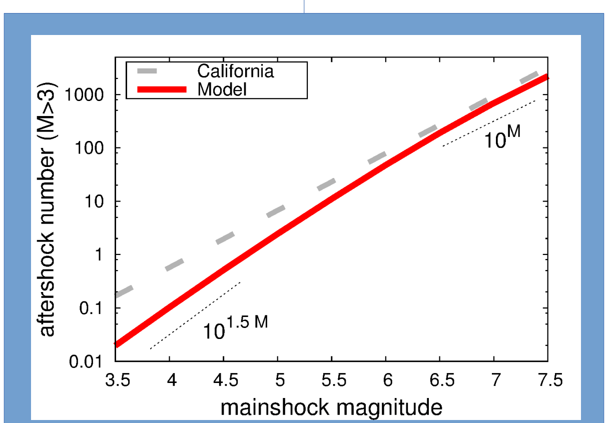
One possibility: A statistical model extension

modified ETAS-model:

… where productivity scaling and spatial kernel are replaced

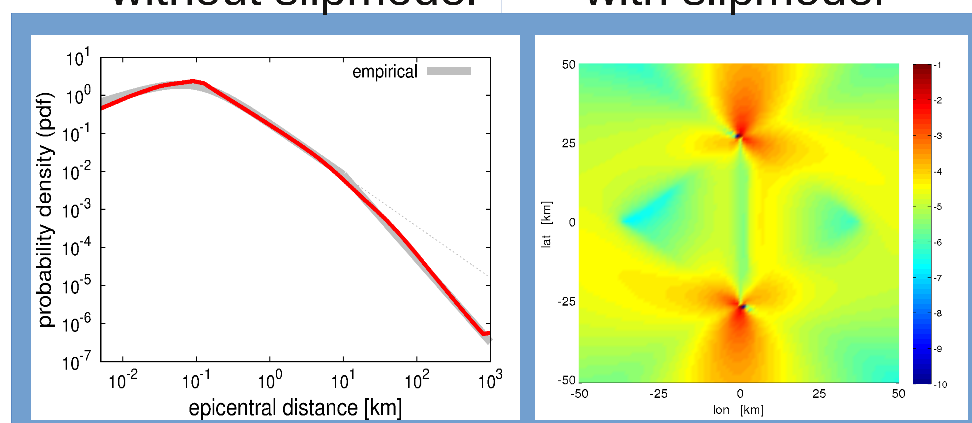
$$\boldsymbol{R}(t,\vec{x}) = \mu + \sum_{i:t_i<t} K_0\, e^{\alpha M_i} \cdot \frac{1}{(c+t-t_i)^p} \cdot \frac{c_{norm}}{(r^2+d^2)^{\gamma/2}}$$

Ogata, 1988



without slipmodel

with slipmodel



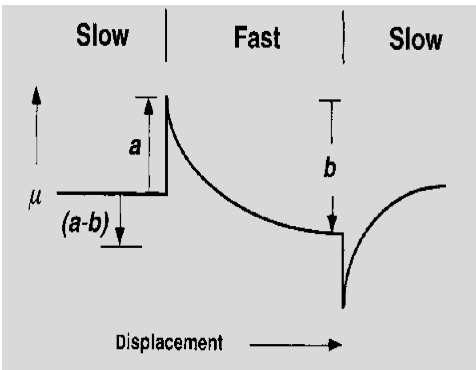
Bach & Hainzl, JGR 2012; Zakharova et al., JGR 2013

… currently tested in CSEP New Zealand

Another possibility:

Rate- and state-dependent frictional nucleation model

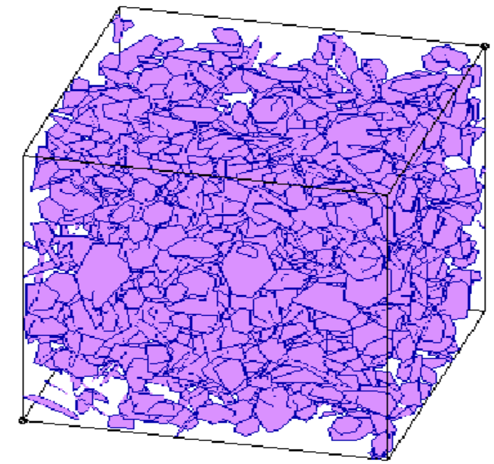
Dieterich, JGR 1994



from lab-derived friction law

→

to statistical response of fault populations



$R = r/\dot{\tau}\gamma$ & $d\gamma = (dt - \gamma\, dS)/A\sigma_n$

stress step: ΔS

$$R(t, \Delta S) = \frac{r}{1 + \left(e^{-\frac{\Delta S}{A\sigma_n}} - 1\right) e^{-\frac{t}{t_a}}}$$

r background rate

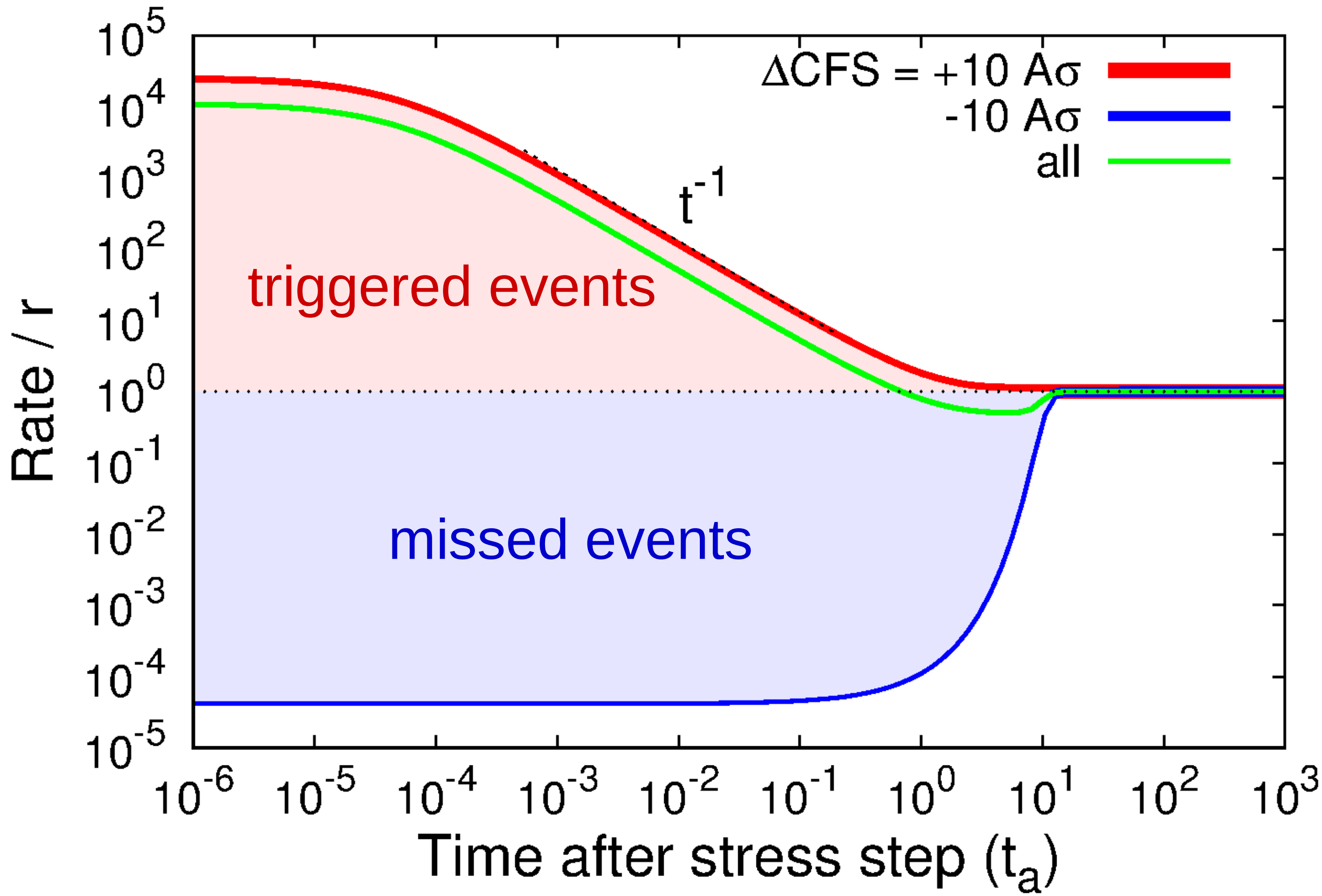
A friction parameter

$\dot{\tau}$ tectonic stressing rate

→ relaxation time:

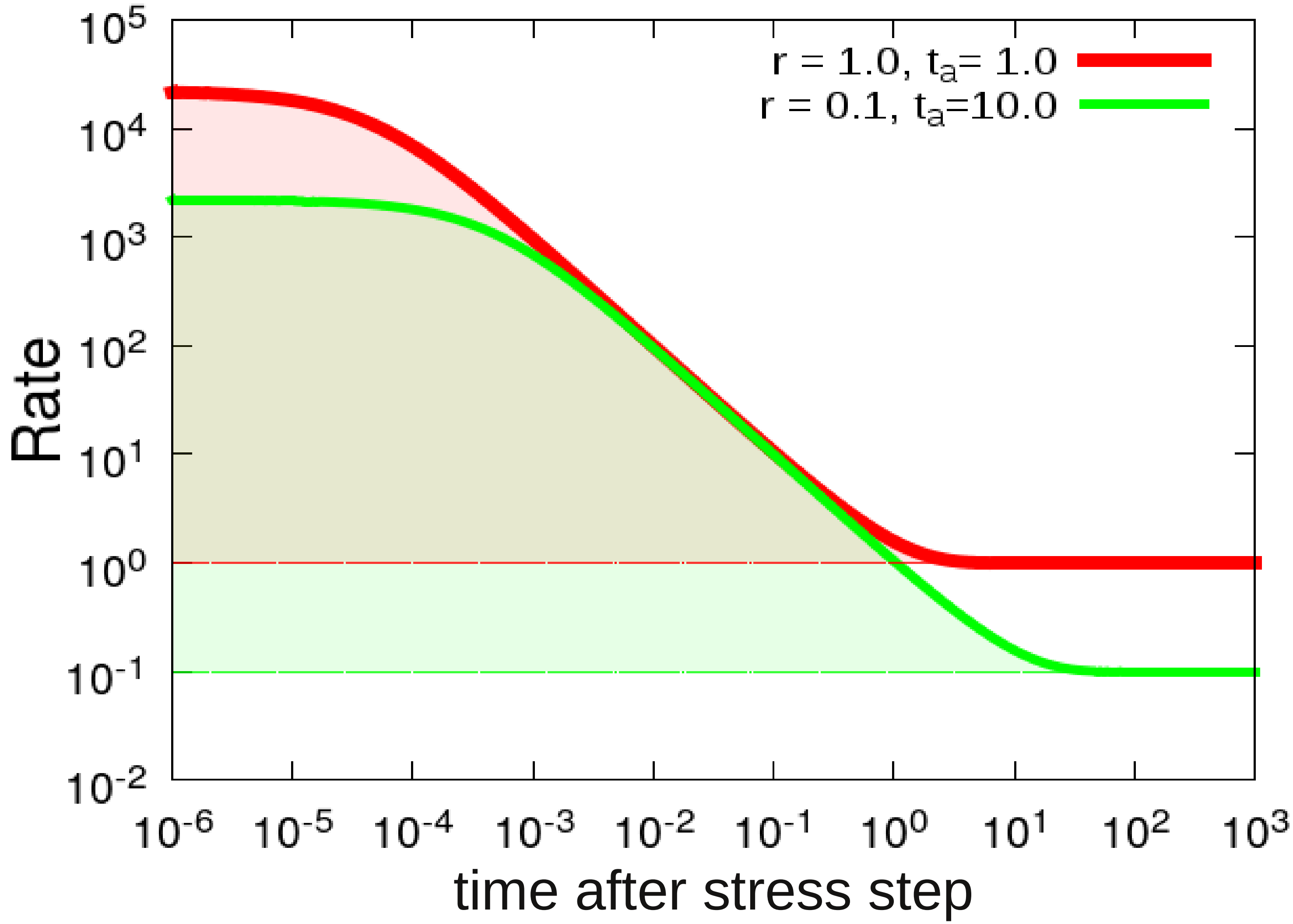
$$t_a = \frac{A\sigma_n}{\dot{\tau}}$$

Response to a stress step of ±ΔCFS:



triggered (+ΔCFS) = # missed (–ΔCFS)

In the case of different stressing rates:



$N_{tot} = (r/\dot{\tau})\ \Delta CFS$ (same as in the clock advance model!)

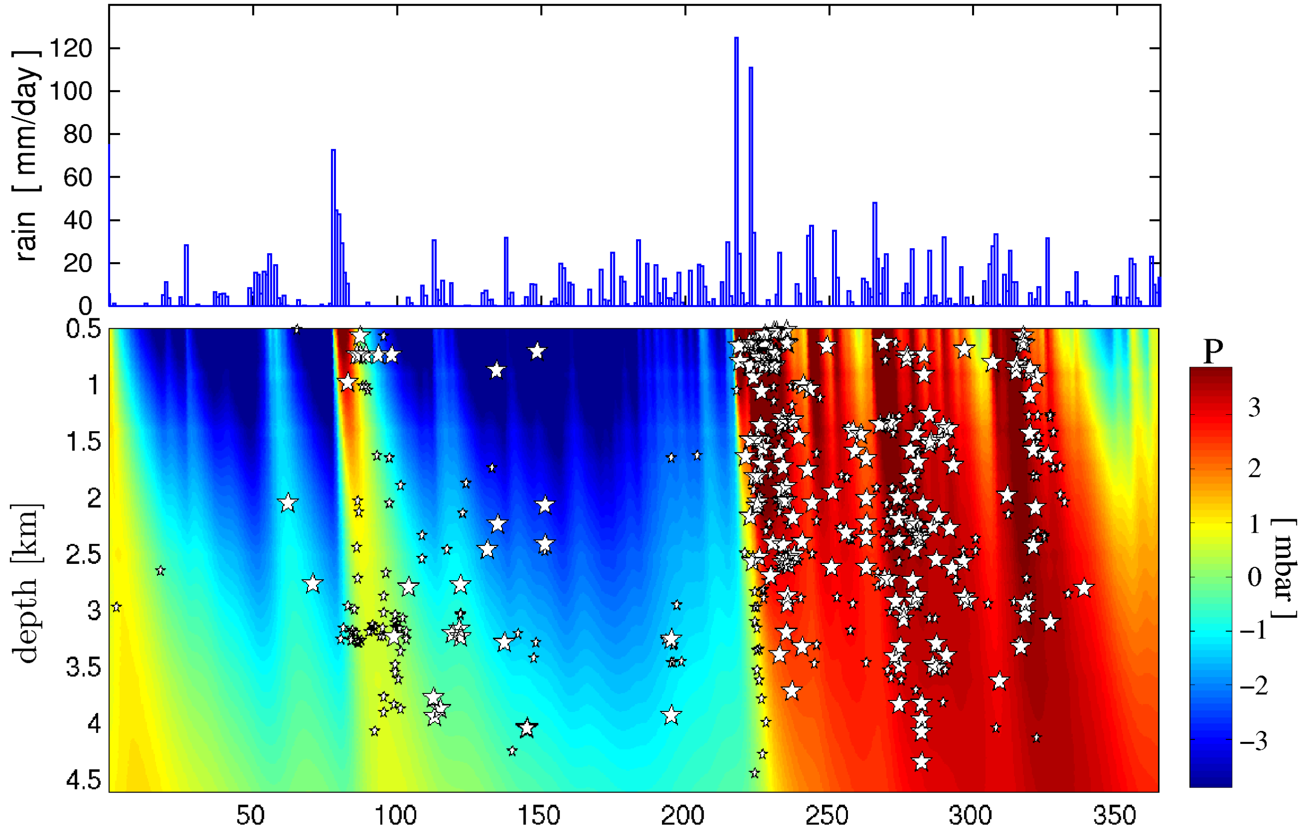
triggered = # triggered if $r/\dot{\tau}$ = constant

Advantage: Earthquake rate can be determined for complex stress histories resulting from aseismic & coseismic processes

Example:

rainfall

pore pressure diffusion



Earthquake activity at Mt. Hochstaufen SE Germany

Hainzl et al. GRL 2006
Hainzl et al. JGR 2013

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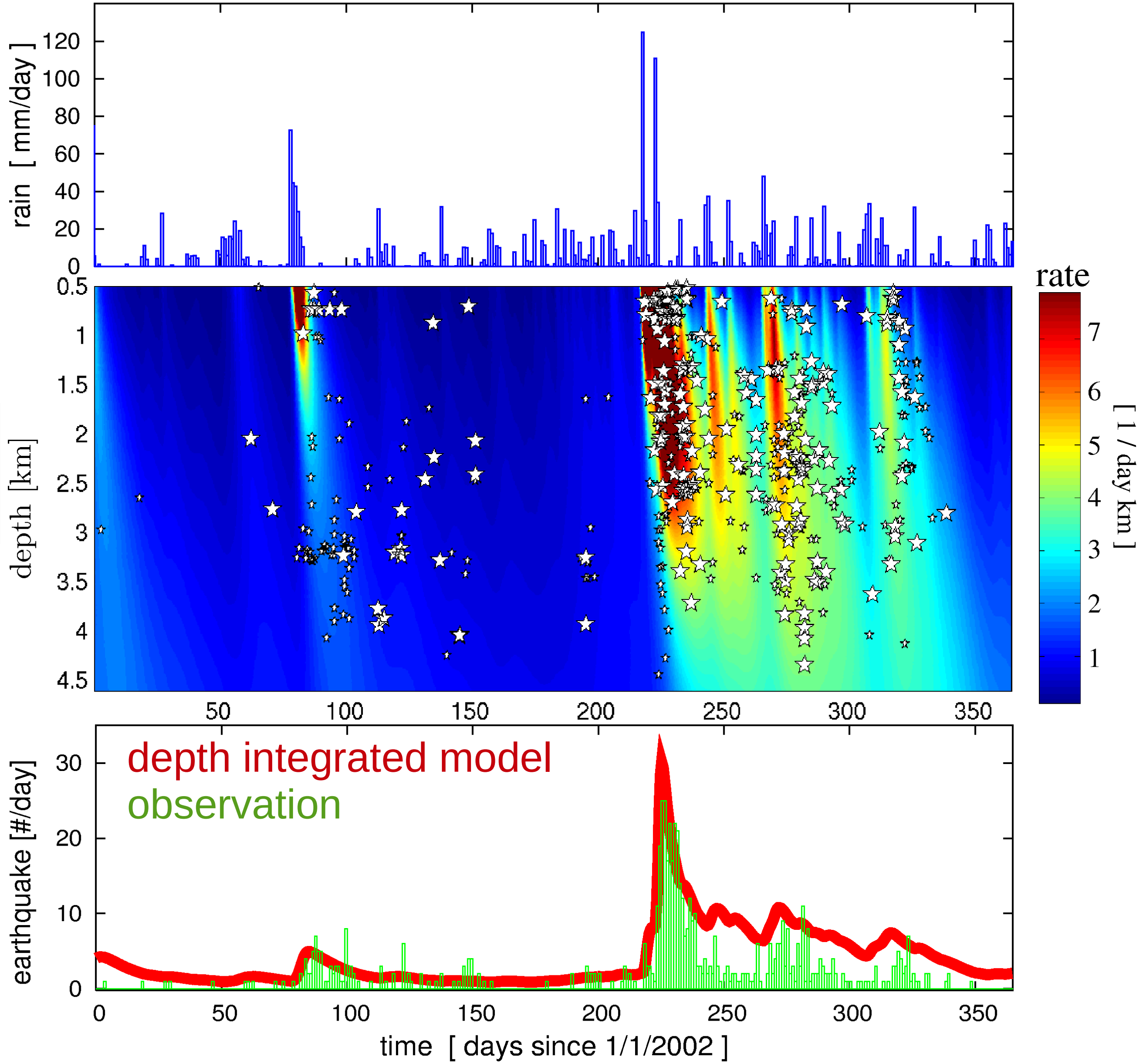
rainfall

earthquake rate



Earthquake activity at Mt. Hochstaufen SE Germany

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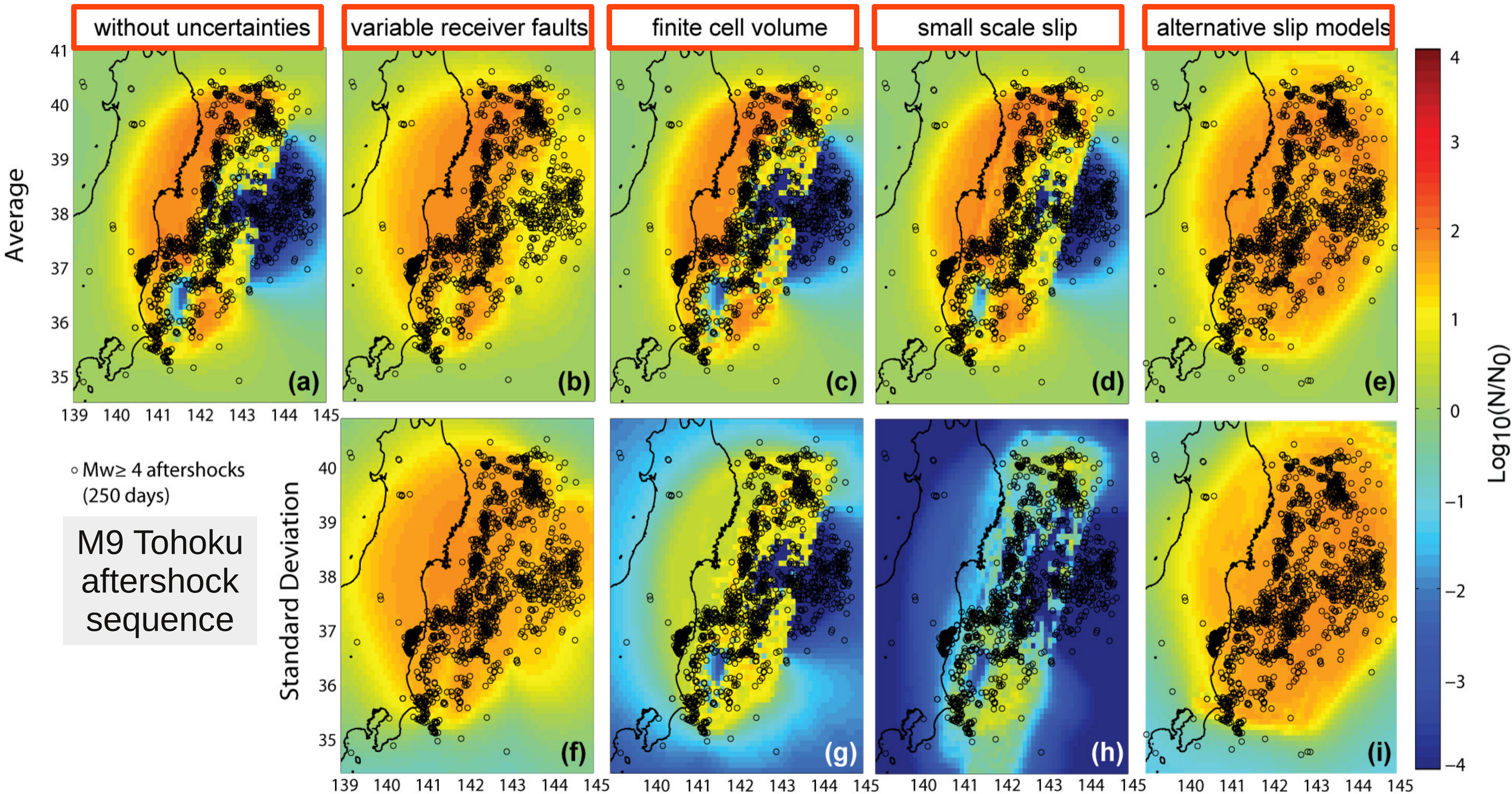
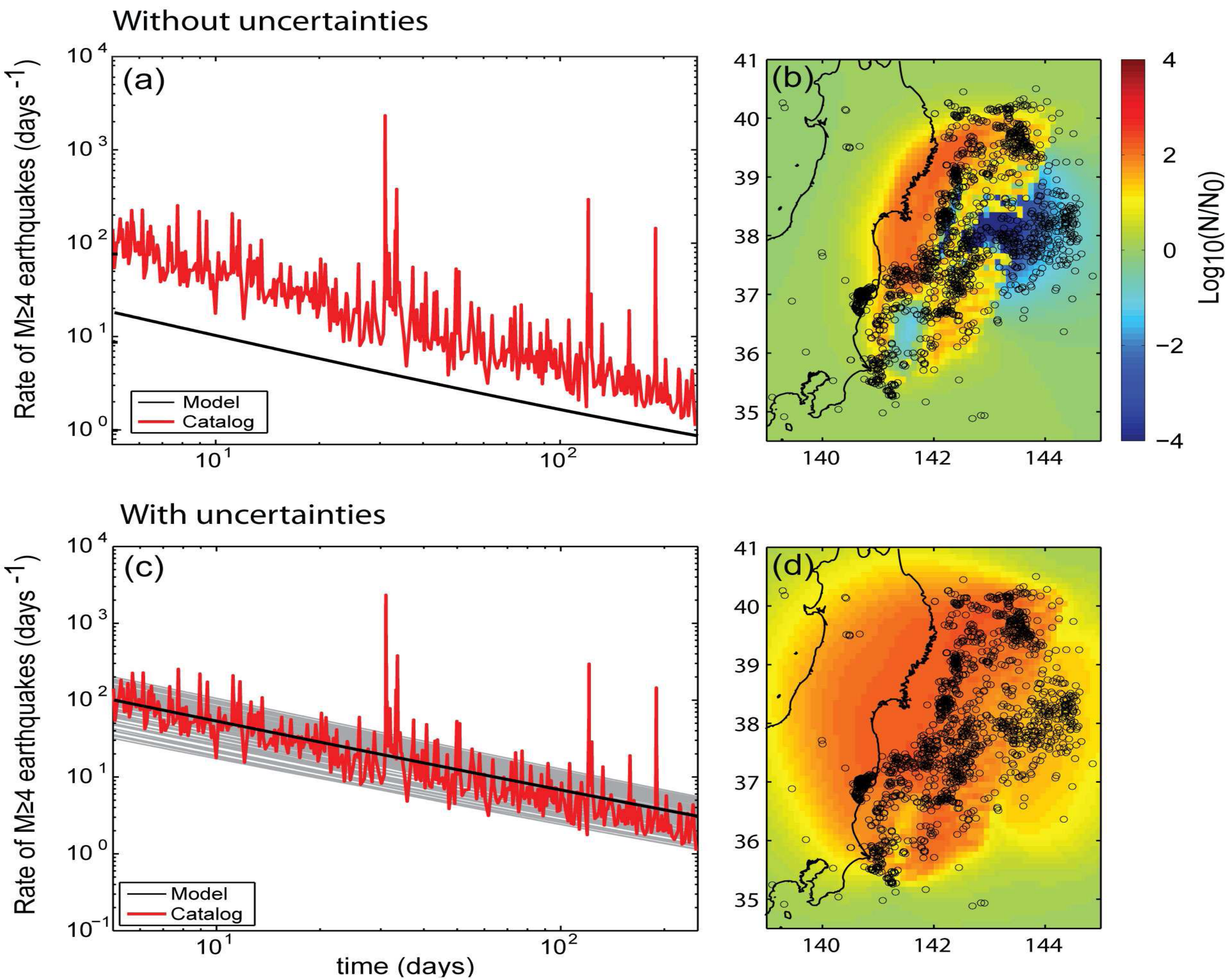
Challenge: Large uncertainties of ΔCFS-calculations



M9 Tohoku aftershock sequence

Model integration by Monte-Carlo sampling … *Cattania et al. JGR 2014*

Model performance improves significantly if uncertainties are systematically taken into account:



M9 Tohoku aftershock sequence

Cattania et al. JGR 2014

Summary

Static stress triggering can explain first-order characteristics:

- spatial distribution
- productivity
- Omori aftershock decay

However:

Uncertainties of the actual stress state (use of Poisson model for the initial state) and the ΔCFS-calculations are large and limit the forecast ability.

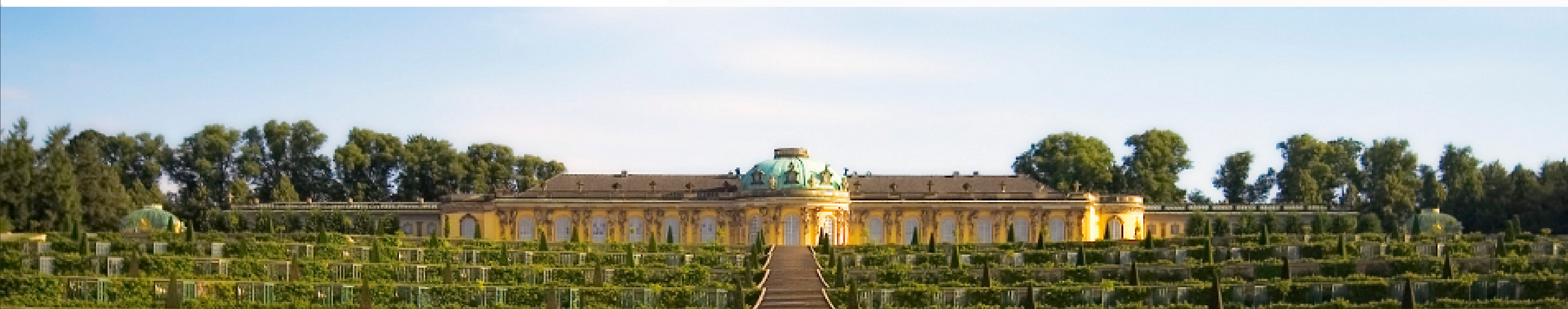
Proper accounting for uncertainties is necessary for hypothesis testing and improved forecasts.

Thank you!

Potsdam, June 14-19, 2015

StatSei9

https://statsei9.quake.gfz-potsdam.de



You are here: 9th International Workshop on Statistical Seismology

MAIN MENU

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9th International Workshop on Statistical Seismology

The International Statistical Seismology (StatSei) workshop is a well-established biannual meeting focusing on recent developments in statistical seismology. The meetings covered a series of topics, including statistical descriptions of earthquake occurrences, earthquake physics, earthquake forecasting and forecast evaluations, earthquake triggering and many more. Starting in 2001, it was held in China, Japan, New Zealand, Mexico, United States, Italy and Greece. The 9th StatSei workshop will be hosted by the GFZ Potsdam (German Research Center for Geosciences) and the University of Potsdam.

We invite statisticians and seismologists as well as geologists working on theoretical and applied questions of seismology to join the workshop in the beautiful historic environment of Potsdam.

We are looking forward to welcome and host you in Potsdam!

The organizing comittee