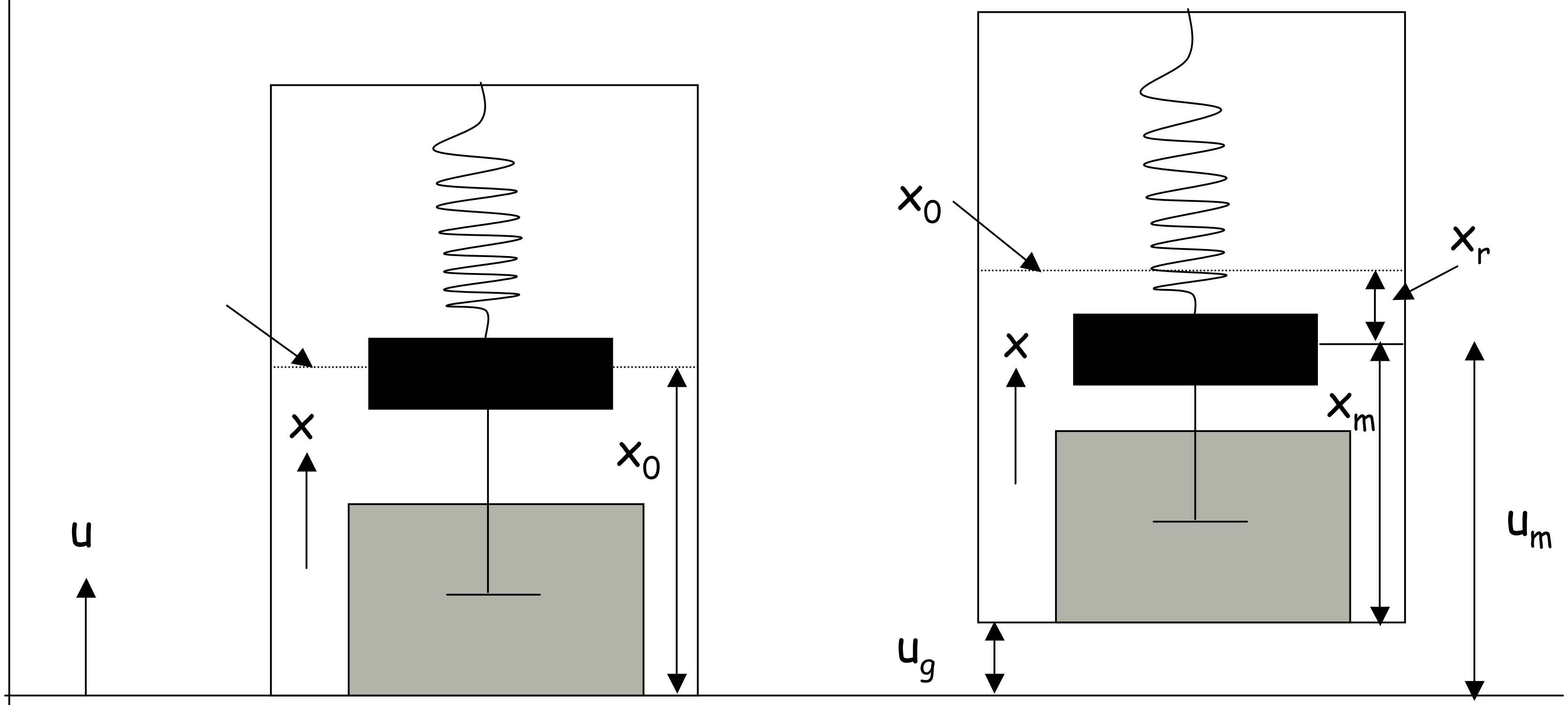
The tasks of an EARTHQUAKE SERVICE

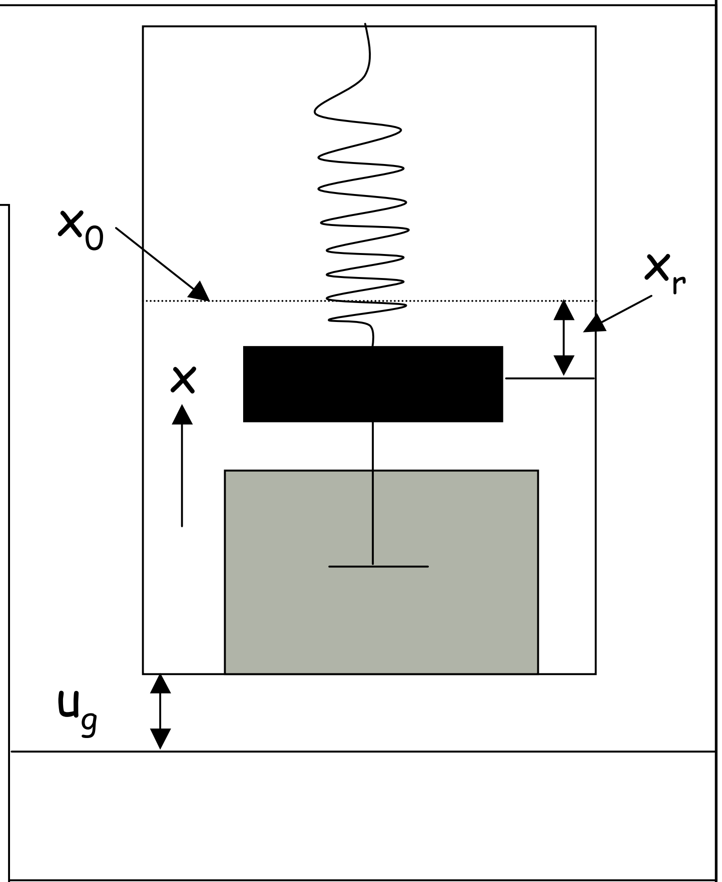
- Seismometers
- Locating earthquake
- Earthquake source - magnitudes
- Earthquake source characteristics
  - Energy
  - Source mechanism
  - Fault/rupture area
- Earthquake hazard
- Earthquake prediction?

Seismometer – The basic Principles



u	ground displacement
x_r	displacement of seismometer mass
x_0	mass equilibrium position

Seismometer – The basic Principles



The motion of the **seismometer** mass as a function of the ground displacement is given through a differential equation resulting from the equilibrium of forces (in rest):

$F_{spring} + F_{friction} + F_{gravity} = 0$

for example

$F_{spring} = -kx$ k spring constant

$F_{friction} = -D\dot{x}$ D friction coefficient

$F_{gravity} = -m\ddot{x}$ m seismometer mass

Seismometer – The basic Principles

using the notation introduced the equation of motion for the mass is (forced damped oscillator)

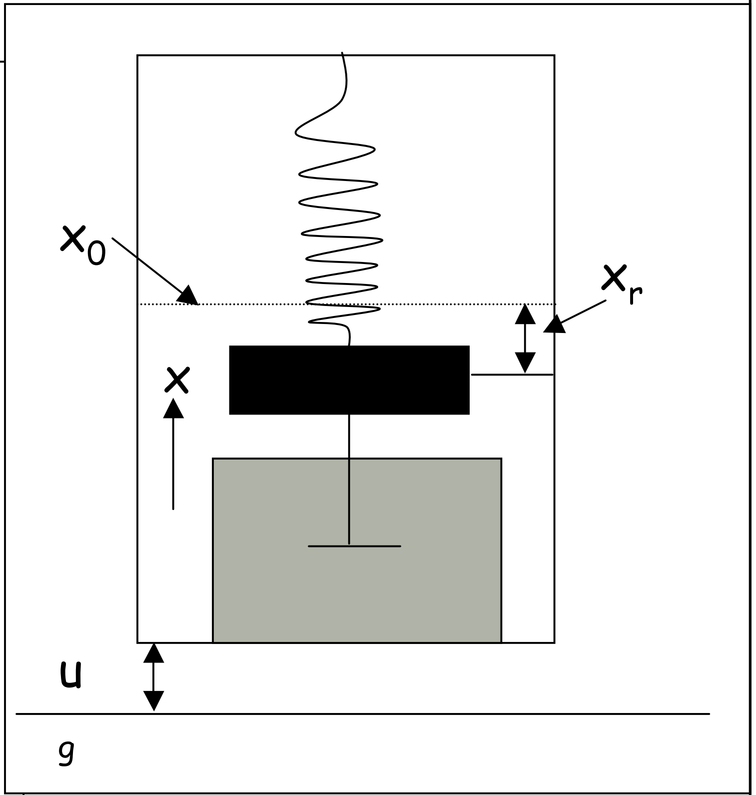
$$\ddot{x}_r(t) + 2\varepsilon\dot{x}_r(t) + \varpi_0^2 x_r(t) = -\ddot{u}_g(t)$$

$$\varepsilon = \frac{D}{2m} = h\varpi_0, \qquad \varpi_0^2 = \frac{k}{m}$$

From this we learn that:

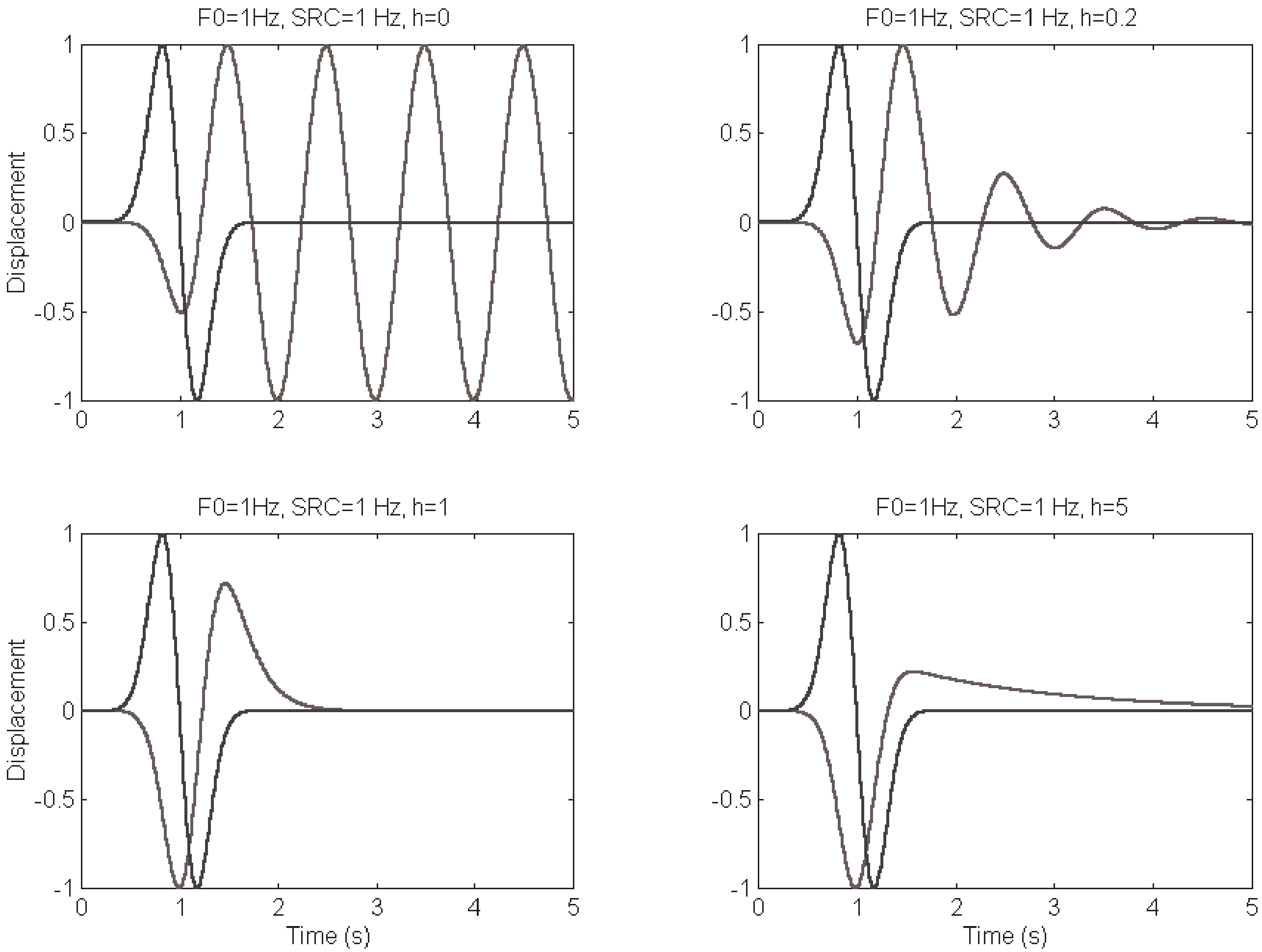
- for slow movements the acceleration and velocity becomes negligible, the seismometer records ground acceleration

- for fast movements the acceleration of the mass dominates and the seismometer records ground displacement



Seismometer – response

… varying damping …



Ground motion (blue) – seismometer mass (red)

Seismometer – Response Function

How does the seismometer amplify the ground motion? Is this amplification frequency dependent?

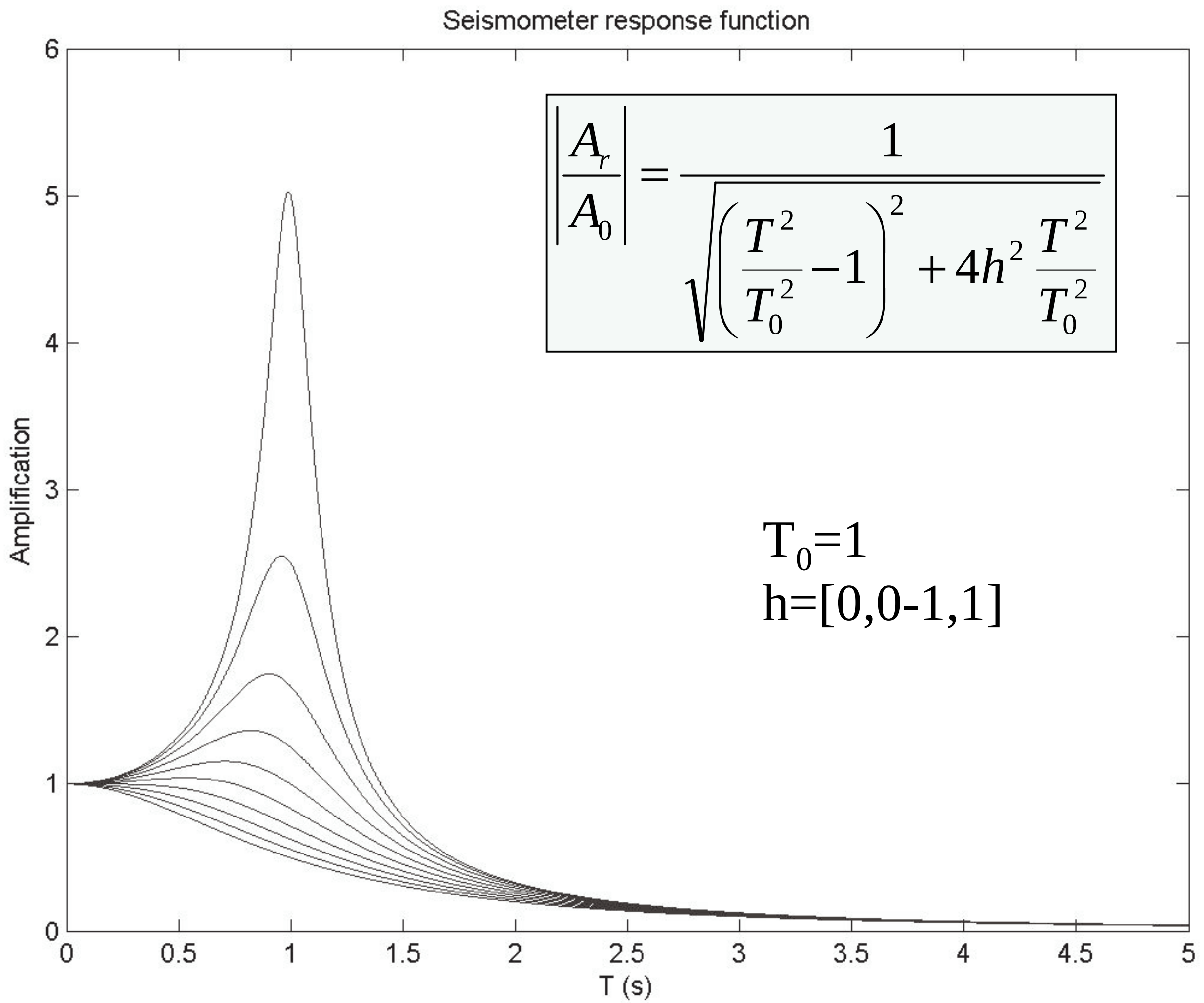
To answer this question we excite our seismometer with a mono-frequent signal and record the response of the seismometer:

$$\ddot{x}_r(t) + h\varpi_0 \dot{x}_r(t) + \varpi_0^2 x_r(t) = \varpi^2 A_0 e^{i\varpi t}$$

the amplitude response A_r of the seismometer depends on the eigenperiod of the seismometer T_0, the period of the excitation T and the damping constant h:

$$\left|\frac{A_r}{A_0}\right| = \frac{1}{\sqrt{\left(\frac{T^2}{T_0^2} - 1\right)^2 + 4h^2 \frac{T^2}{T_0^2}}}$$

Seismometer - Response Function



Seismometer – what to remember!

- A seismometer can be described as a mass-spring system that is forced by the ground motion -> forced damped oscillator.
- The output of a seismometer is descriptive of the motion of the mass (and not the ground). To recover the ground motion the seismogram needs to be corrected. This „deconvolution“ is called restitution in seismology.
- Seismometers amplify (or dampen) the ground motion.This ampflification is frequency dependent. The eigenfrequency (period) of the seismometer spring determines its domain of applicability.
- Today most seismometers are broadband sensors that measure accurately from T=500s to f=50 Hz.

Earthquake location

What do we have?

- Arrival times of P and S waves at various seismic stations

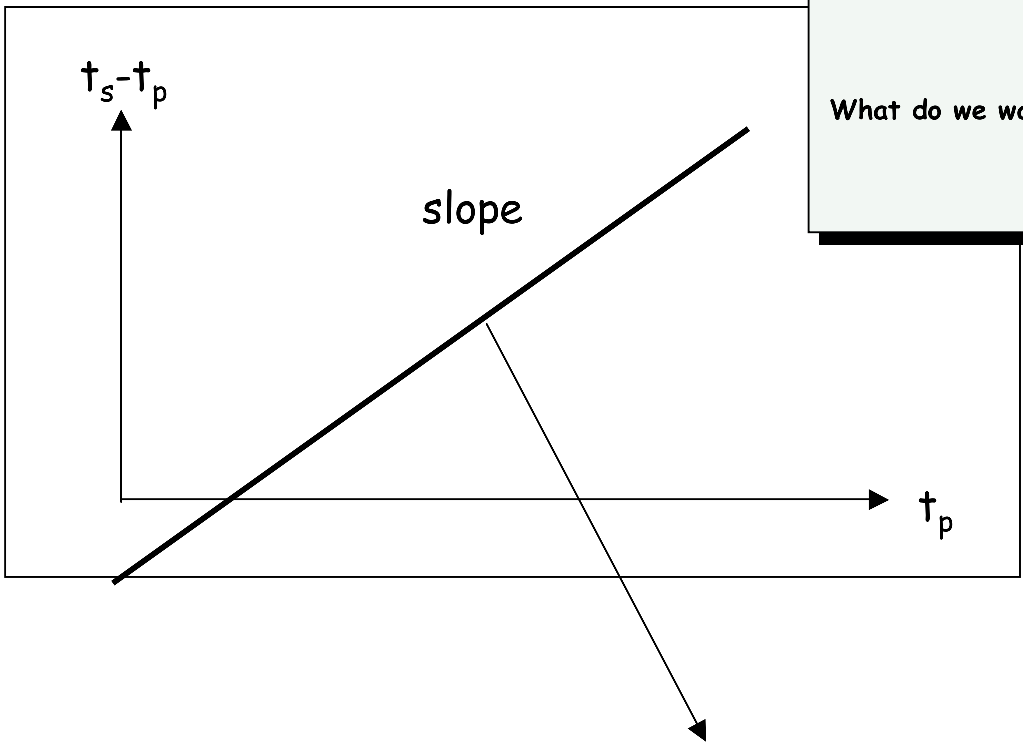
What do we want to know?

- origin time of the earthquake
- the epicenter and depth

Let us assume the earthquake happened at time t_0 and we know the seismic wave velocities of the ground to be v_p (P-waves) and v_S (S-waves). When we record an earthquake at a distance Δ we have

$$v_P = \frac{\Delta}{t_P - t_0}, \qquad v_S = \frac{\Delta}{t_S - t_0}$$

Wadati diagram



What do we have?

- Arrival times of P and S waves at various seismic stations

What do we want to know?

- origin time of the earthquake
- the epicenter and depth

$$(t_S - t_P) = (\frac{v_P}{v_S} - 1)(t_P - t_0)$$

$$y = ax + y_0$$

Earthquake location

With the slope v_p/v_s-1 of the diagram we can get the v_p/v_s ratio, again after rearranging we can calculate the origin time t_0 of the earthquake

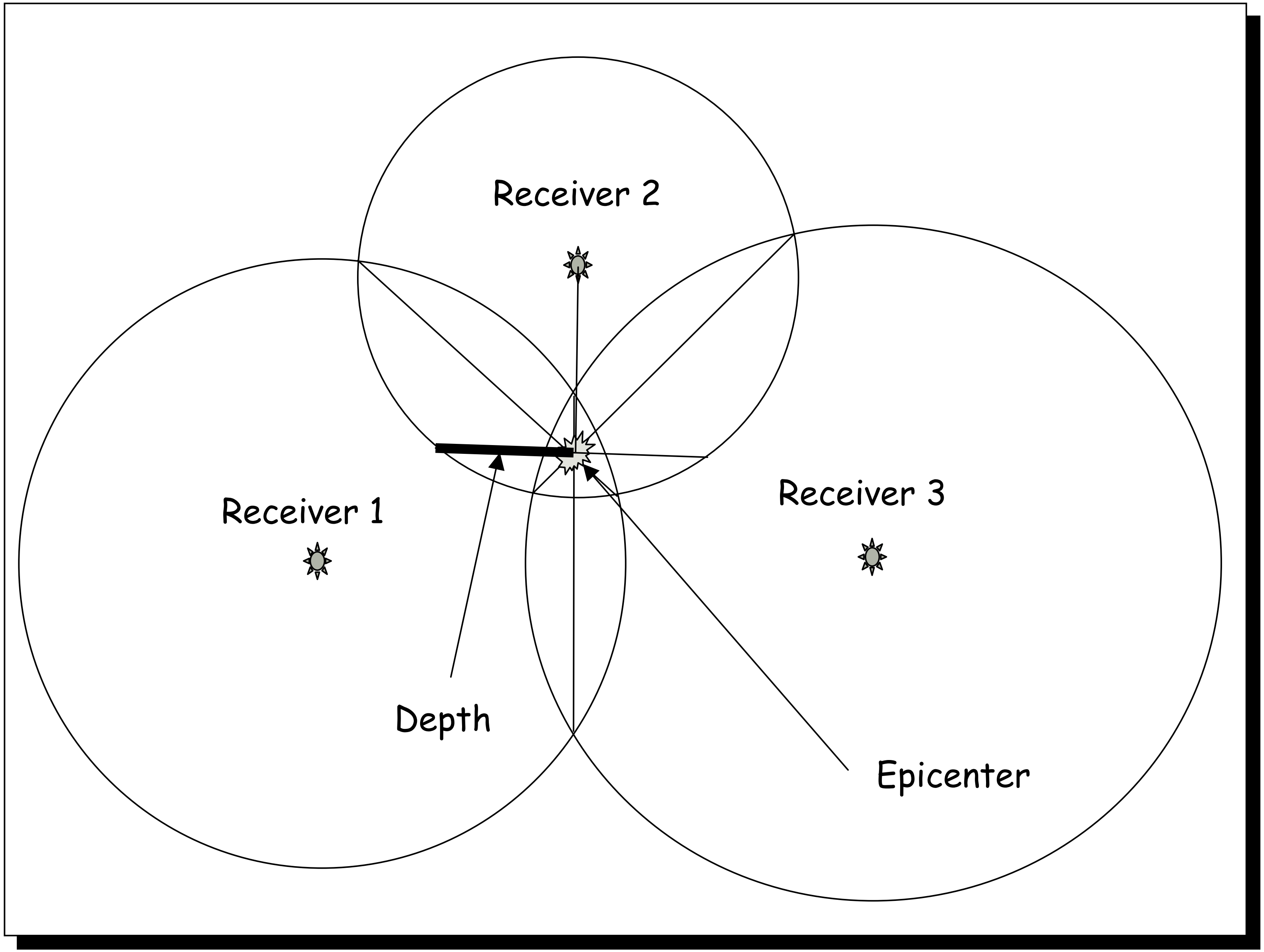
$$t_0 = t_P - \frac{t_S - t_P}{\frac{v_P}{v_S} - 1}$$

and the distance of the earthquake from each receiver i with P arrival time t_{Pi}

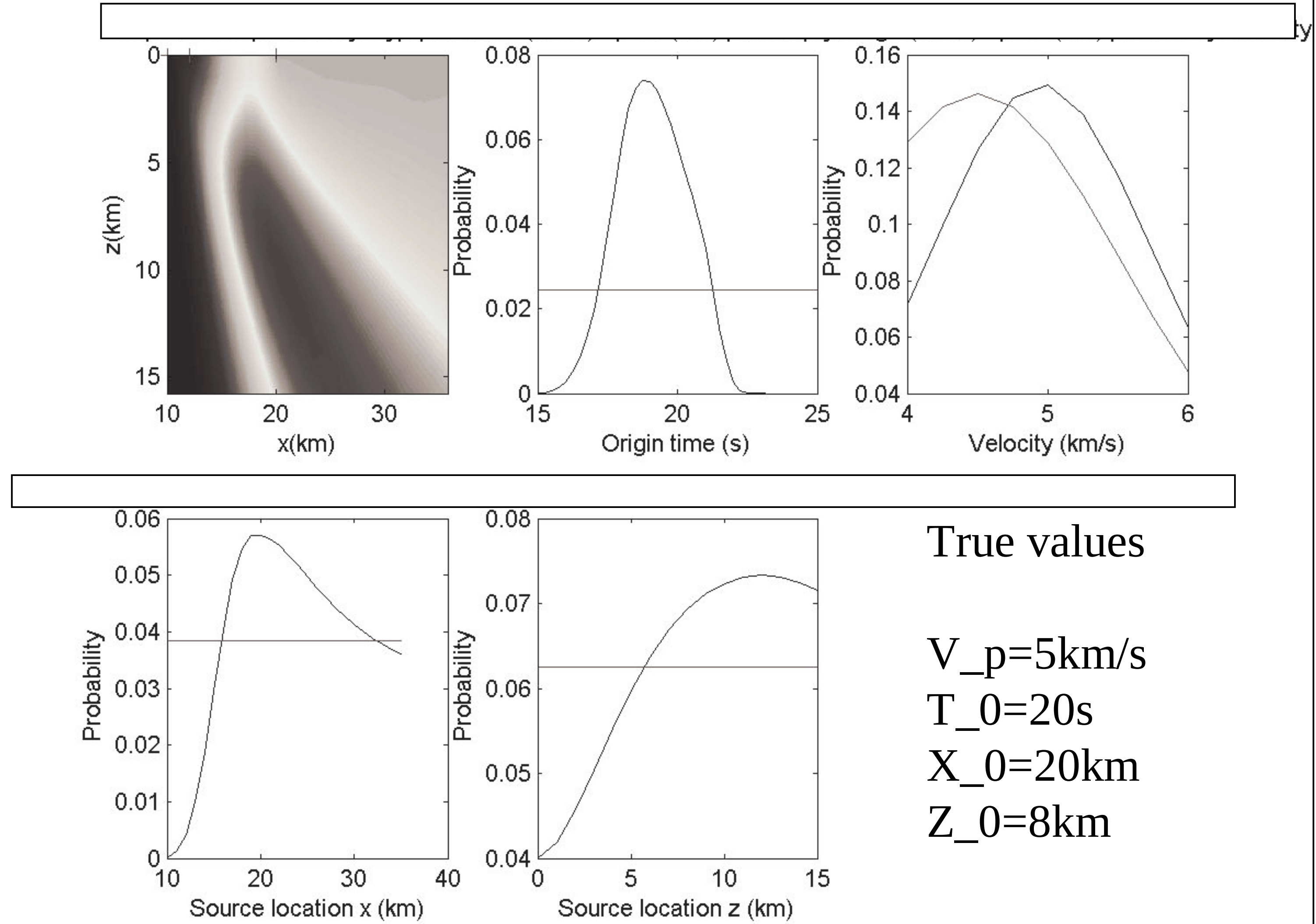
$$\Delta_i = v_P (t_{Pi} - t_0)$$

But how can we determine epicenter and depth?

Earthquake location



Earthquake location - uncertainties



True values

V_p=5km/s
T_0=20s
X_0=20km
Z_0=8km

Earthquake location – what to remember

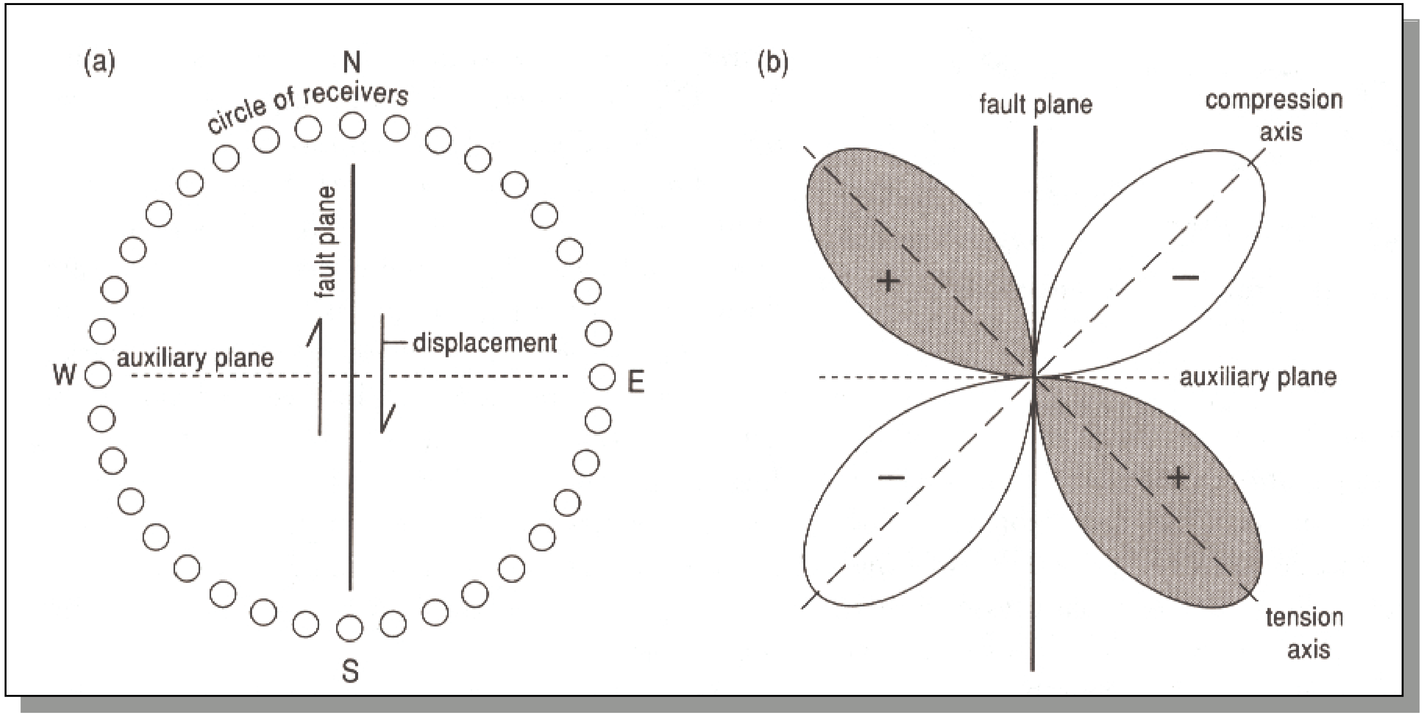
- To determine the location of earthquakes we need to know the origin time and we need to have an estimate of the seismic velocities in the Earth

- In theory, the observation of P and S arrival times at one station and the assumption of a homogenous velocity model allows the estimation of the origin time (Wadati diagram)

- The location of the earthquake with coordinates (x,y,z) or (r, θ, λ) is called its hypocenter, the projection to the Earth's surface epicenter.

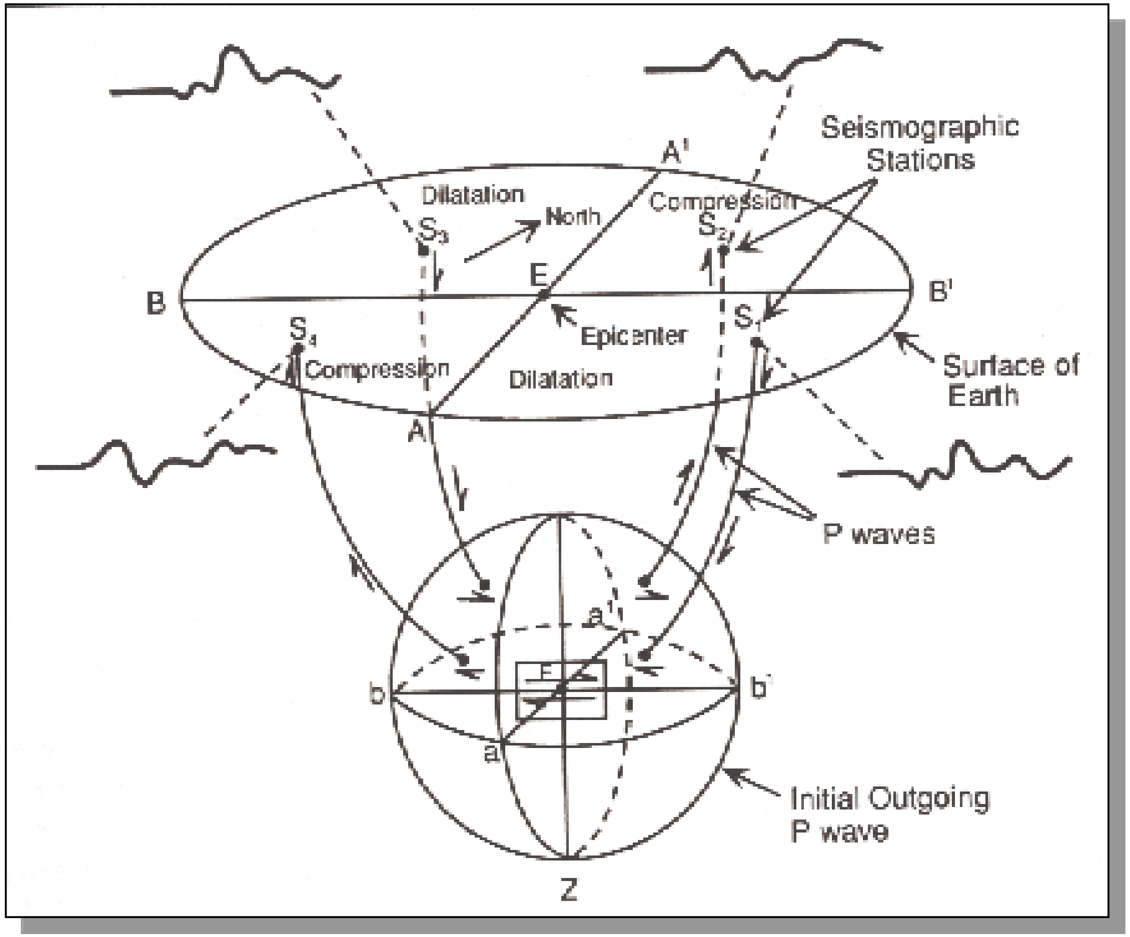
- The depth of earthquakes is usually much less well resolved than the horizontal coordinates (similar to GPS).

The double-couple point source



The basic physical (and extremely successful) model for a source is two fault planes slipping in opposite directions

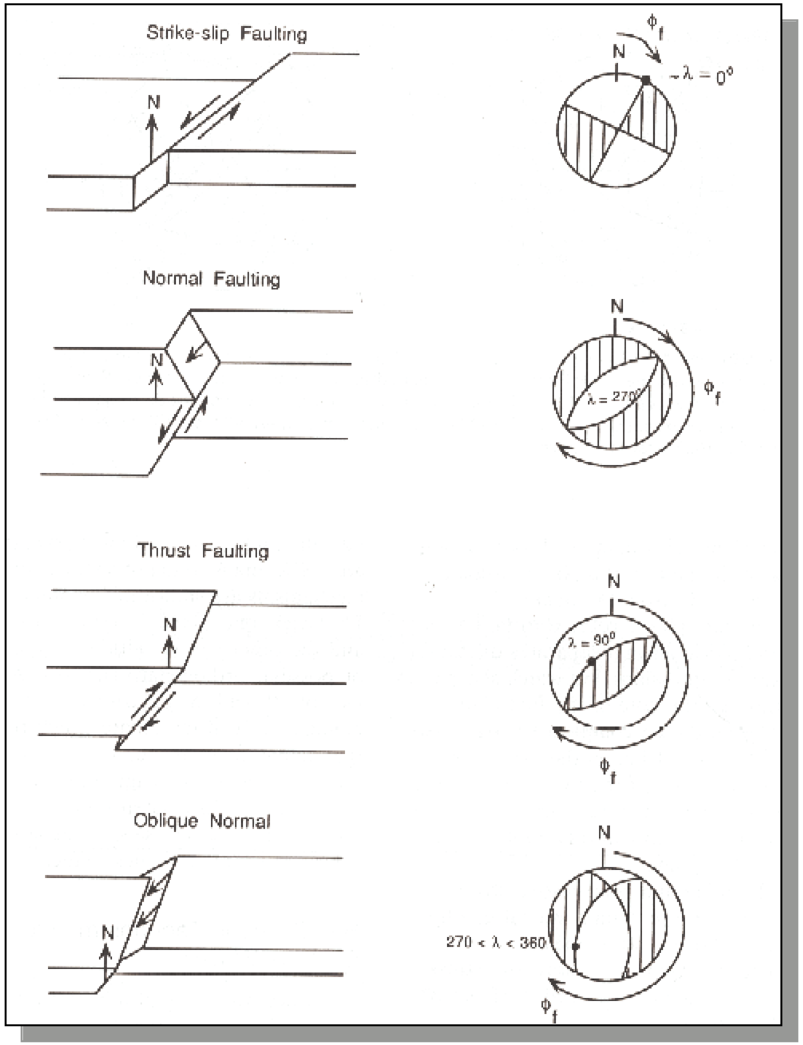
Radiation from shear dislocation



First motion of P waves at seismometers in various directions.

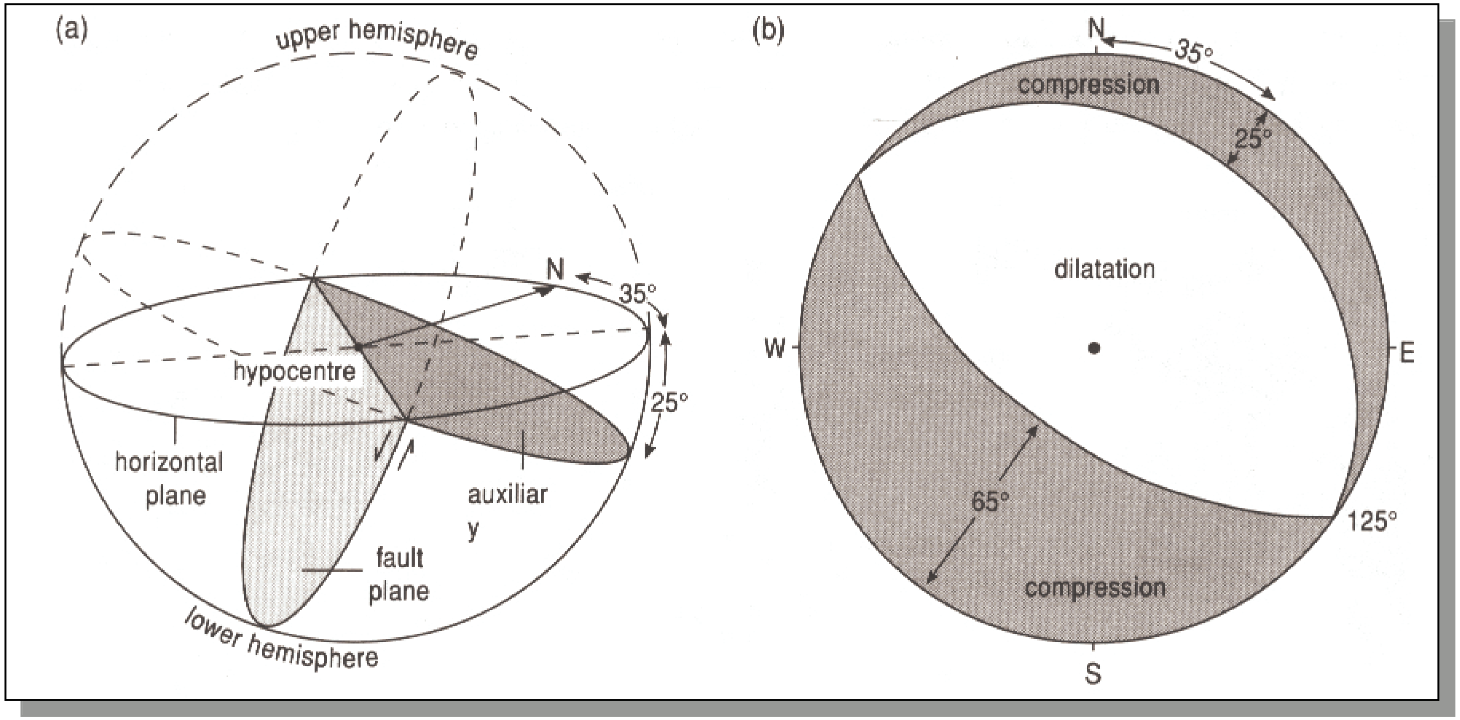
The polarities of the observed motion is used to determine the point source characteristics.

Fault types



Basic fault types and their appearance in the focal mechanisms. Dark regions indicate compressional P-wave motion.

Seismic sources



Our goal: find the **fault plane** and the **slip direction** from seismic observations

Radiation from a point source

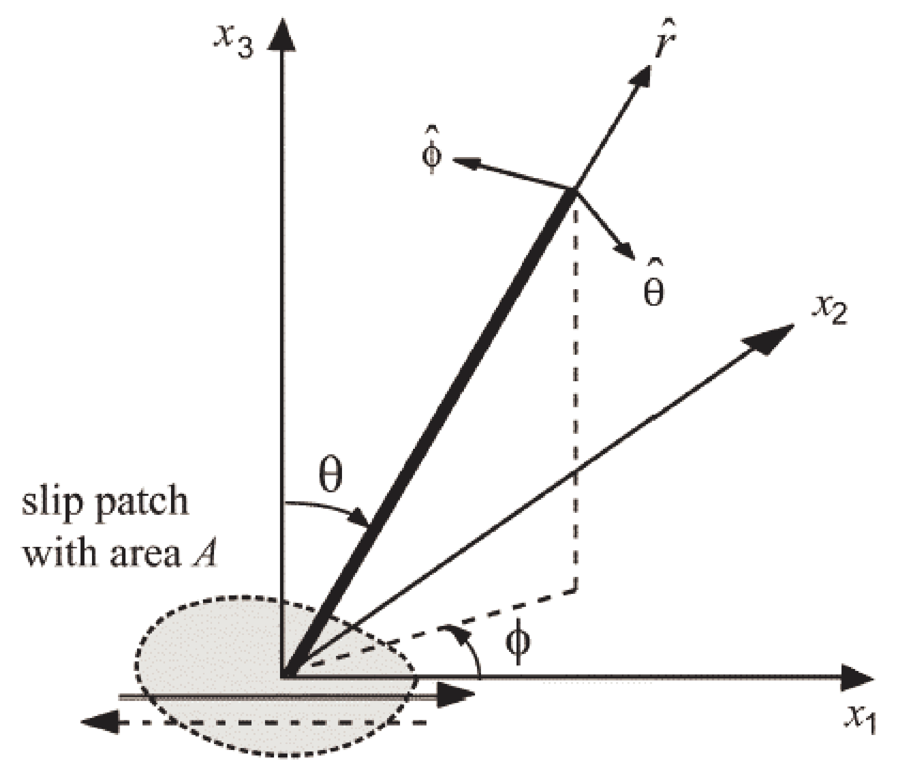


FIGURE 5 Cartesian and polar coordinate systems for analysis of radiation by a slip patch with area A and average slip $\langle \Delta u(t) \rangle$.

Geometry we use to express the seismic wavefield radiated by point double-couple source with area A and slip Δu

Here the fault plane is the x_1x_2-plane and the slip is in x_1-direction. Which stress components are affected?

Radiation pattern

$$\mathbf{A}^{N} = 9\sin 2\theta\cos\phi\hat{\mathbf{r}} - 6(\cos 2\theta\cos\phi\hat{\boldsymbol{\theta}} - \cos\theta\sin\phi\hat{\boldsymbol{\phi}}),$$

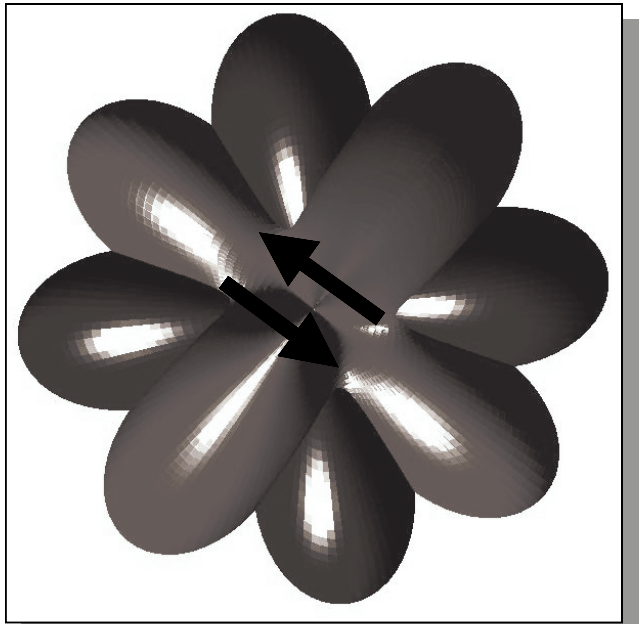
$$\mathbf{A}^{IP} = 4\sin 2\theta\cos\phi\hat{\mathbf{r}} - 2(\cos 2\theta\cos\phi\hat{\boldsymbol{\theta}} - \cos\theta\sin\phi\hat{\boldsymbol{\phi}}),$$

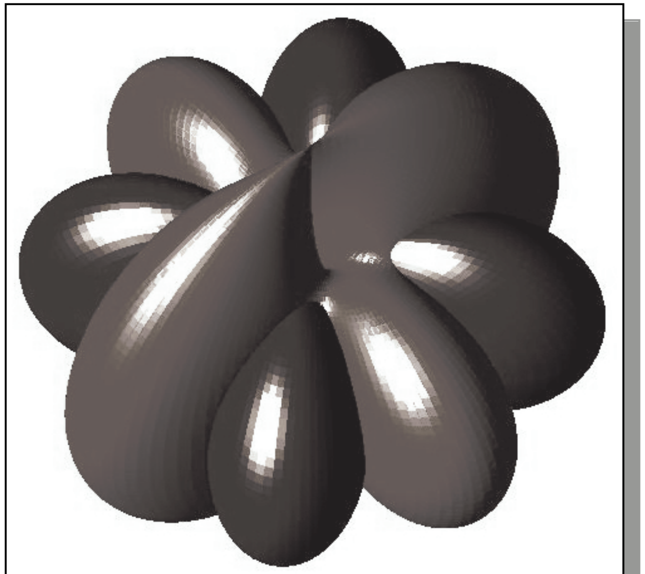
$$\mathbf{A}^{IS} = -3\sin 2\theta\cos\phi\hat{\mathbf{r}} + 3(\cos 2\theta\cos\phi\hat{\boldsymbol{\theta}} - \cos\theta\sin\phi\hat{\boldsymbol{\phi}}),$$

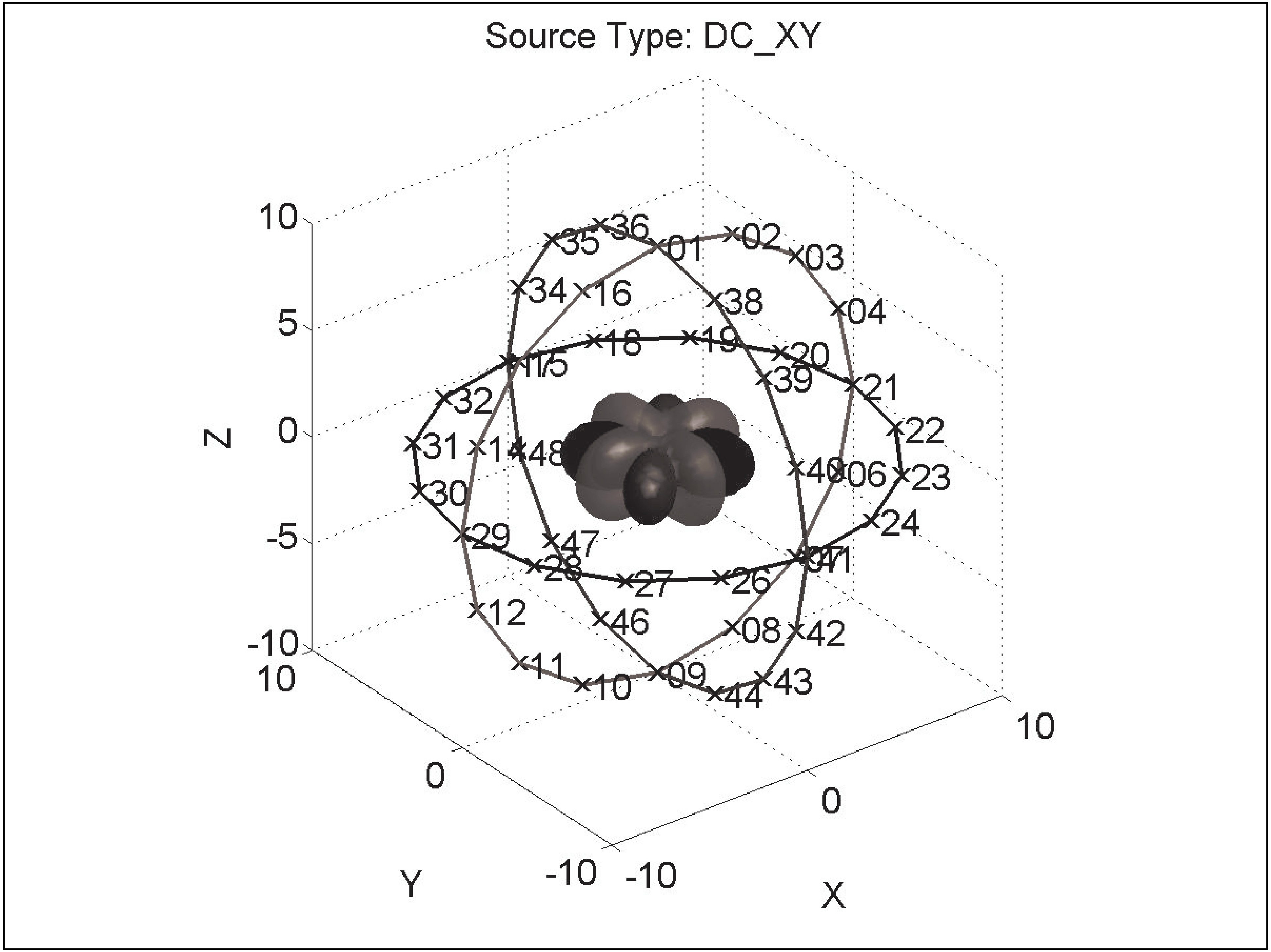
$$\mathbf{A}^{FP} = \sin 2\theta\cos\phi\hat{\mathbf{r}},$$

$$\mathbf{A}^{FS} = \cos 2\theta\cos\phi\hat{\boldsymbol{\theta}} - \cos\theta\sin\phi\hat{\boldsymbol{\phi}},$$

Far field P – blue
Far field S - red






Source Type: DC_XY
Z
Y
X
10
5
0
-5
-10
10
0
-10
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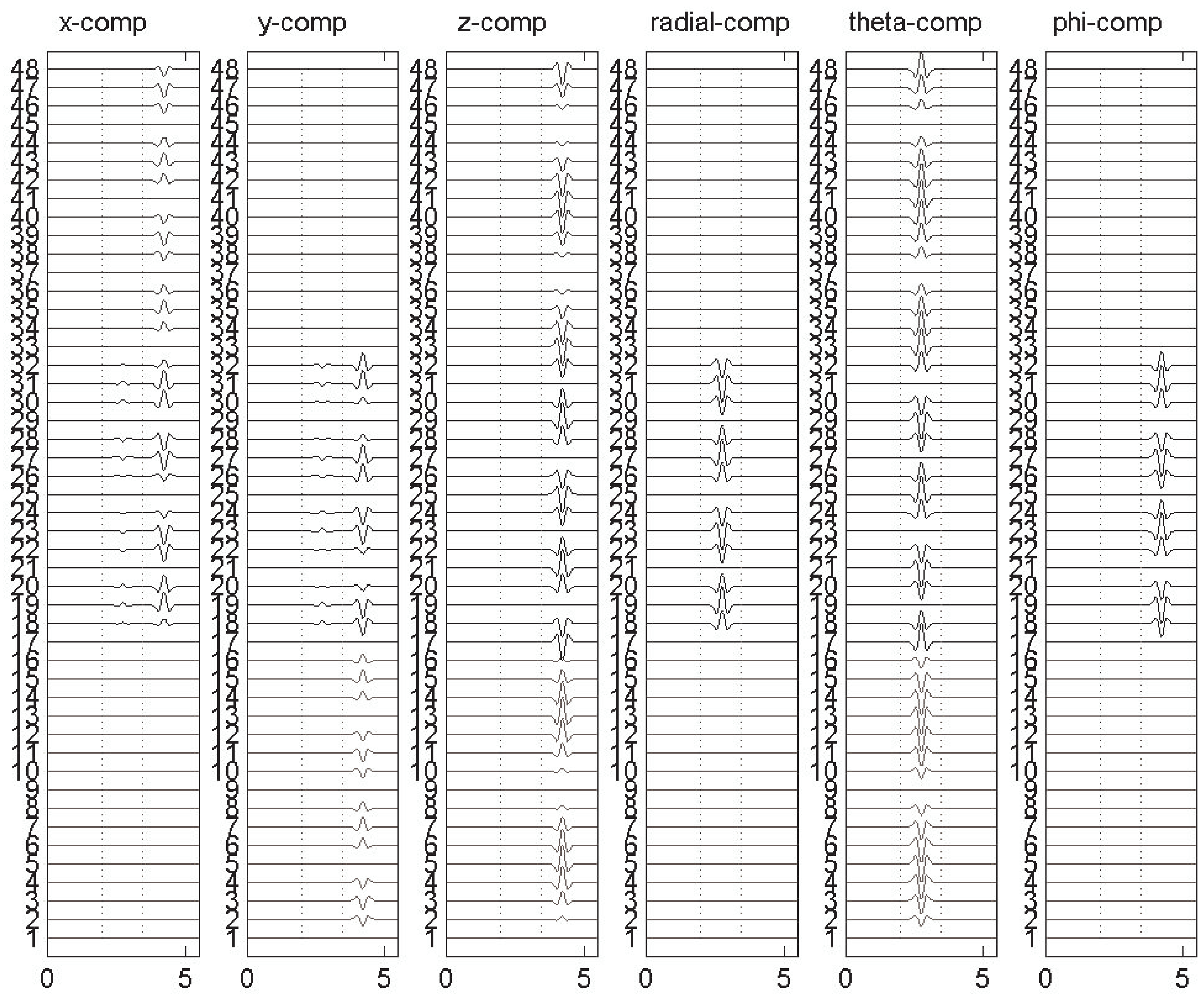

x-comp
y-comp
z-comp
radial-comp
theta-comp
phi-comp
0
5

Beachballs and moment tensor

Moment Tensor	Beachball	Moment Tensor	Beachball
$\frac{1}{\sqrt{3}}\begin{pmatrix}1&0&0\\0&1&0\\0&0&1\end{pmatrix}$		$-\frac{1}{\sqrt{3}}\begin{pmatrix}1&0&0\\0&1&0\\0&0&1\end{pmatrix}$	
$-\frac{1}{\sqrt{2}}\begin{pmatrix}0&0&0\\0&0&1\\0&1&0\end{pmatrix}$		$\frac{1}{\sqrt{2}}\begin{pmatrix}0&0&0\\0&1&0\\0&0&-1\end{pmatrix}$	
$\frac{1}{\sqrt{2}}\begin{pmatrix}0&1&0\\1&0&0\\0&0&0\end{pmatrix}$		$\frac{1}{\sqrt{2}}\begin{pmatrix}0&0&1\\0&0&0\\1&0&0\end{pmatrix}$	
$\frac{1}{\sqrt{2}}\begin{pmatrix}1&0&0\\0&-1&0\\0&0&0\end{pmatrix}$		$\frac{1}{\sqrt{2}}\begin{pmatrix}1&0&0\\0&0&0\\0&0&-1\end{pmatrix}$	
$\frac{1}{\sqrt{6}}\begin{pmatrix}1&0&0\\0&1&0\\0&0&-2\end{pmatrix}$		$\frac{1}{\sqrt{6}}\begin{pmatrix}1&0&0\\0&-2&0\\0&0&1\end{pmatrix}$	
$\frac{1}{\sqrt{6}}\begin{pmatrix}-2&0&0\\0&1&0\\0&0&1\end{pmatrix}$		$-\frac{1}{\sqrt{6}}\begin{pmatrix}-2&0&0\\0&1&0\\0&0&1\end{pmatrix}$	

explosion - implosion

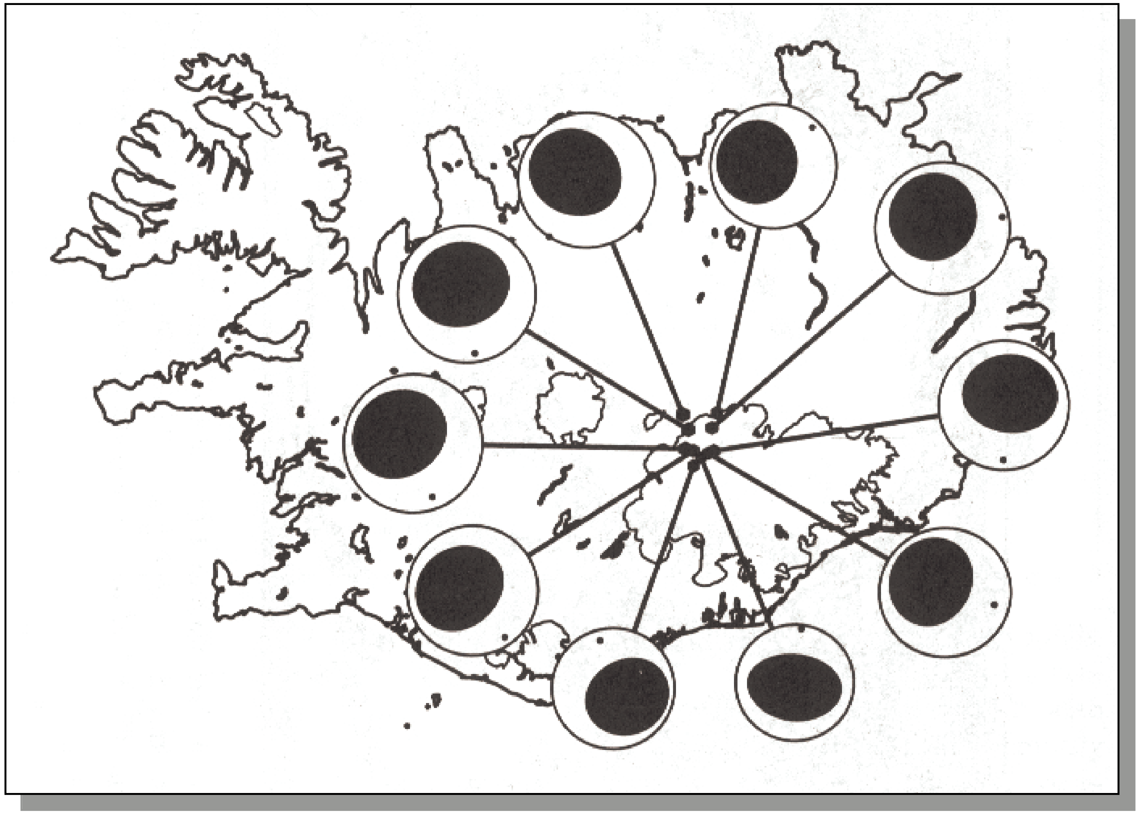
vertical strike slip fault

vertical dip slip fault

45° dip thrust fault

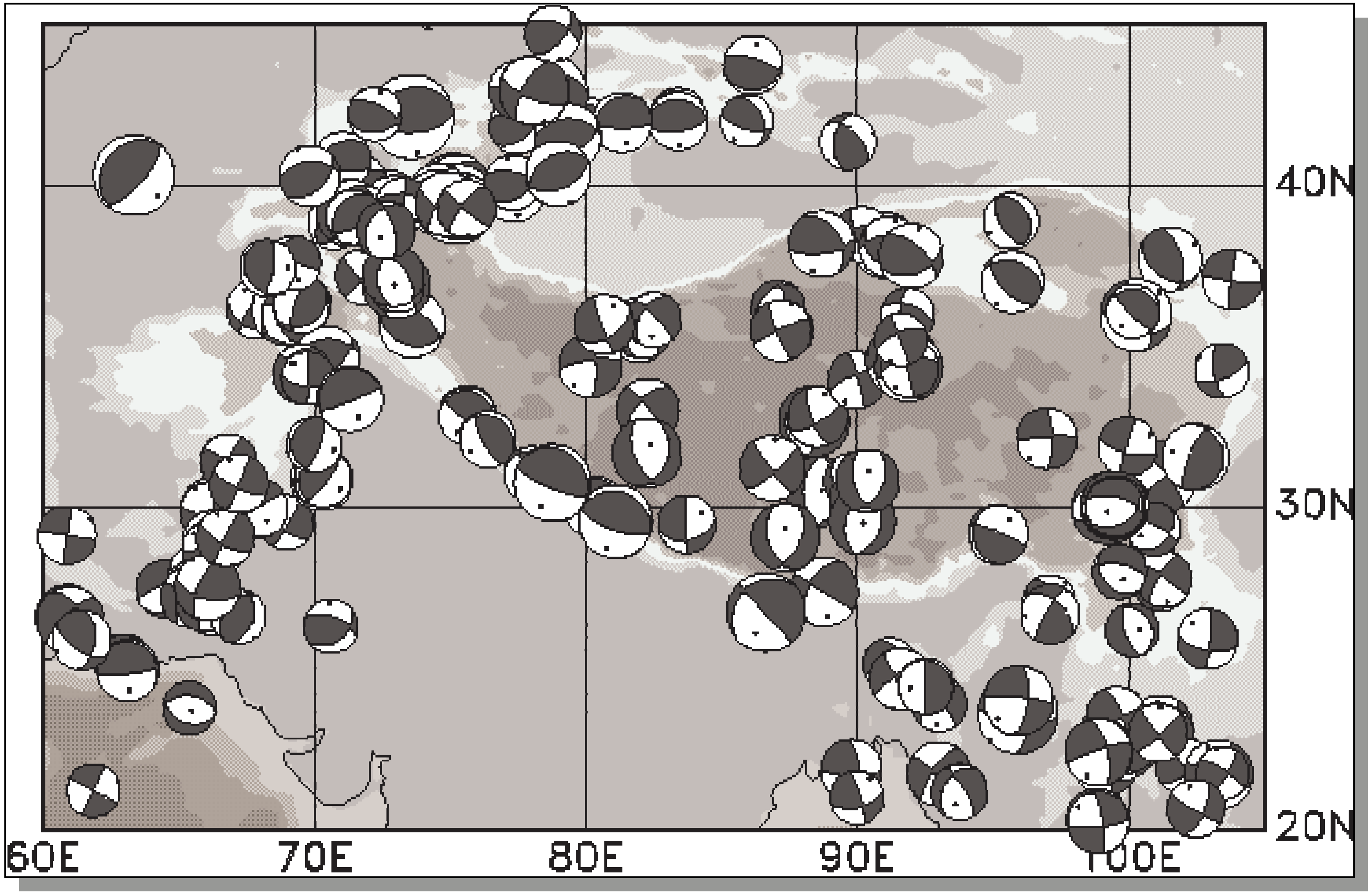
compensated linear vector dipoles

Beachballs - Iceland



Fried eggs: simulataneous vertical extension and horizontal compression

Beachballs - Himalaya

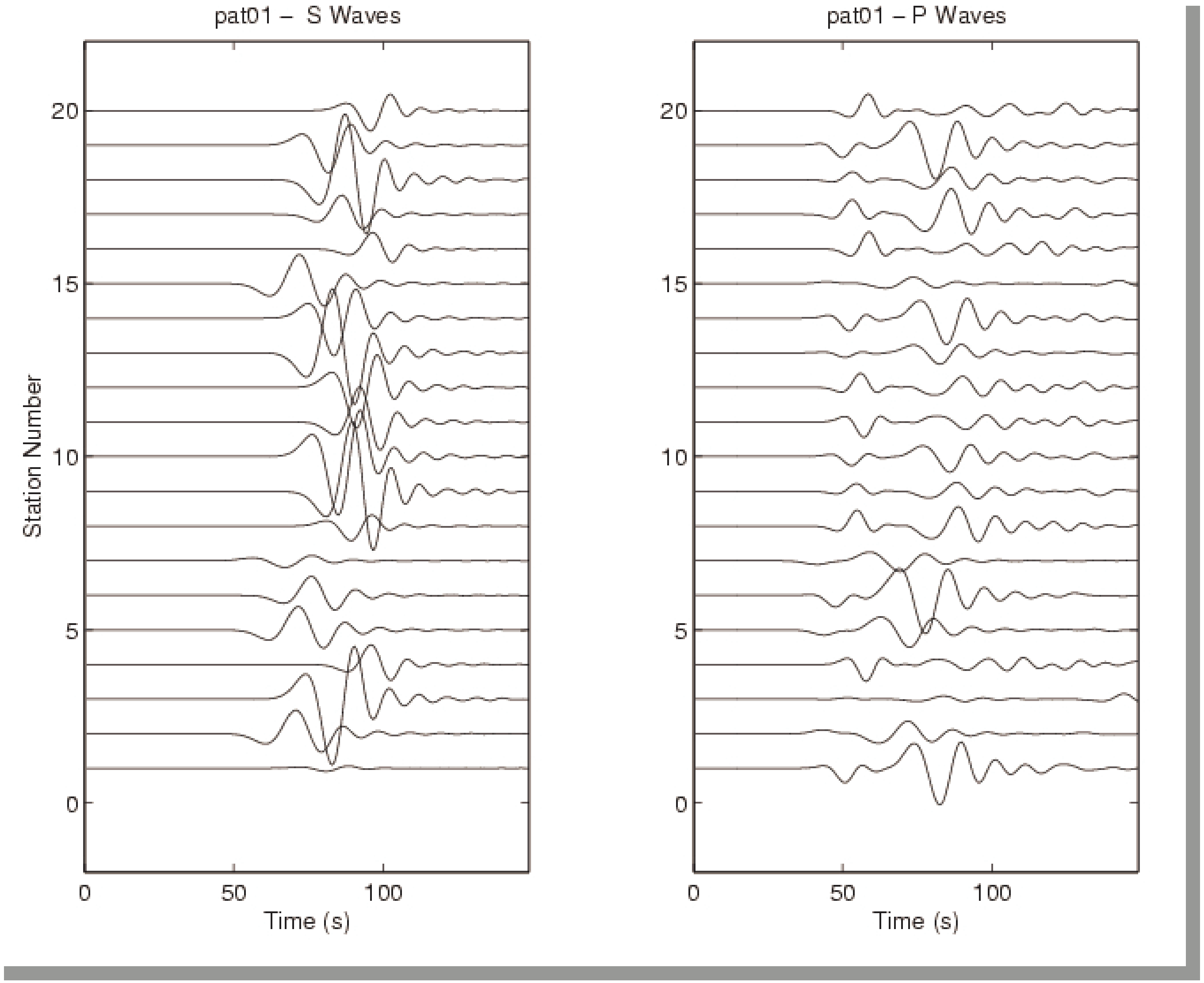


Seismic point sources

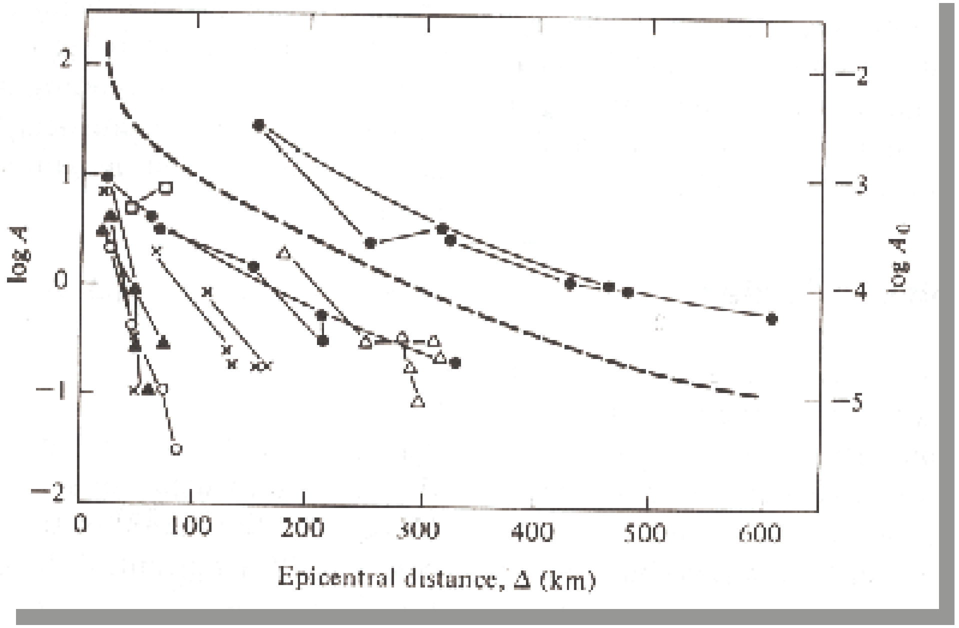
what to remember!

- Earthquake sources can be described by a double-couple (DC) force system (two force vectors slightly displaced acting in opposite directions).
- The DC source leads to a very anisotropic radiation pattern that carries information on the orientation of the fault plane and the direction of the force couple (slip direction)
- Mathematically the solution is ambiguous, there are always two orthogonal fault planes and slip directions that are possible
- The polarity of body waves (P and S) in seismic recordings is used to recover the earthquake source parameters (source mechanism, -> beachballs).

Exercise


pat01 – S Waves
pat01 – P Waves
Station Number
Time (s)
Time (s)
0
5
10
15
20
0
50
100

Magnitude Scales - Richter

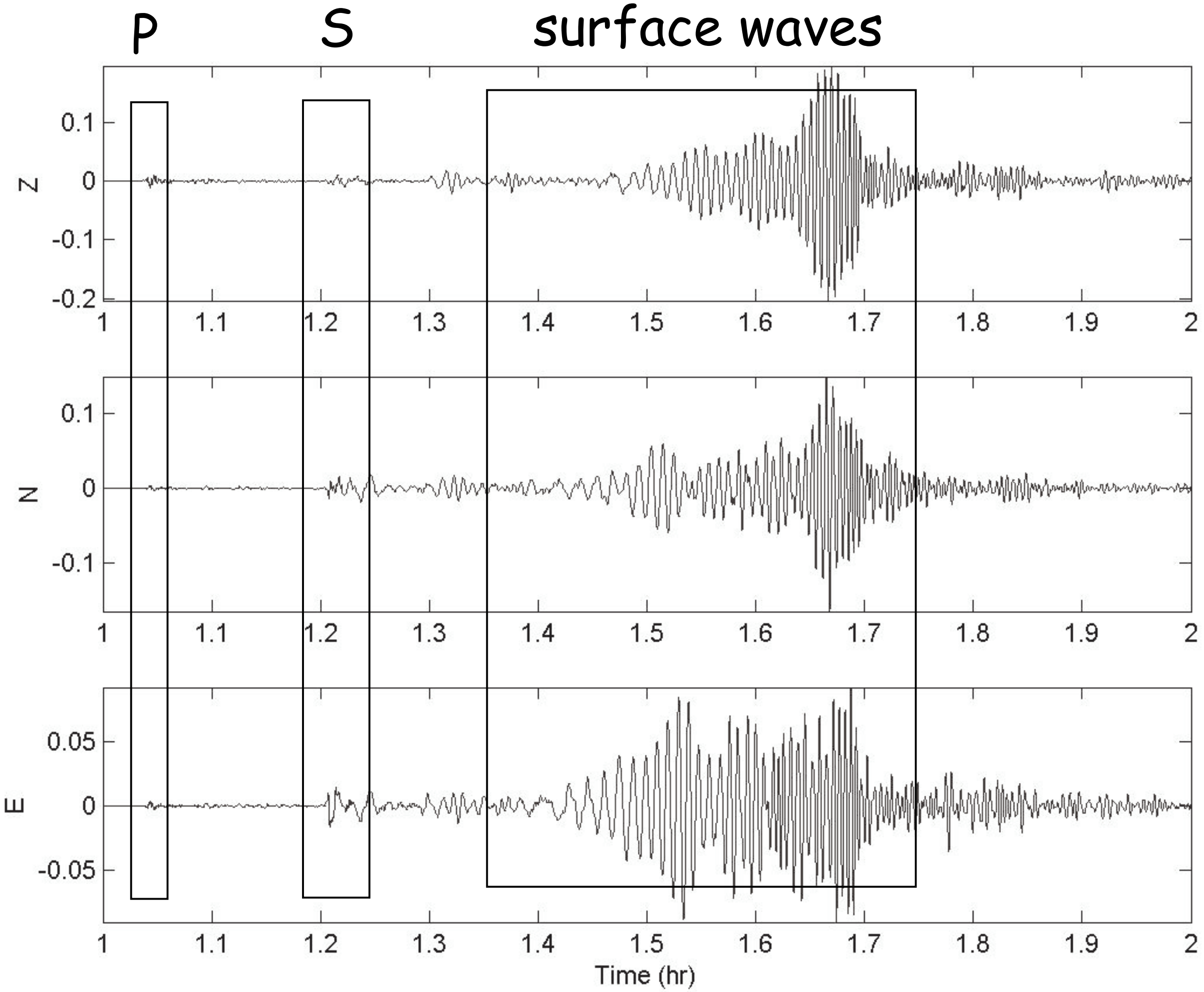


Data from local earthquakes in California

The original Richter scale was based on the observation that the amplitude of seismic waves systematically decreases with epicentral distance.

3C seismograms

Tokachi-oki, M8.3, 25.9.03, Δ=90°



Magnitude Scales

$$M = \log(A/T) + f(\Delta, h) + C_s + C_r$$

M seismic magnitude
A amplitude
T period
f correction for distance
C_s correction for site
C_r correction for receiver

M_L Local magnitude
M_b body-wave magnitude
M_s surface wave magnitude
M_w energy release

Magnitude Scales – Richter and others

Local Magnitude M_L

$M_L = \log A - \log A_0$ — $-\log A_0$ from tables or

$M_L = \log A + 0.003R + 0.7$ — R distance in km, A in mm

Domain: $R < 600$km

Surface wave magnitude M_S

$M_S = \log(A/T) + 1.66 \log D + 3.3$ — $T=18\text{-}22$s, $D=20\text{-}160^o$, $h < 50$km

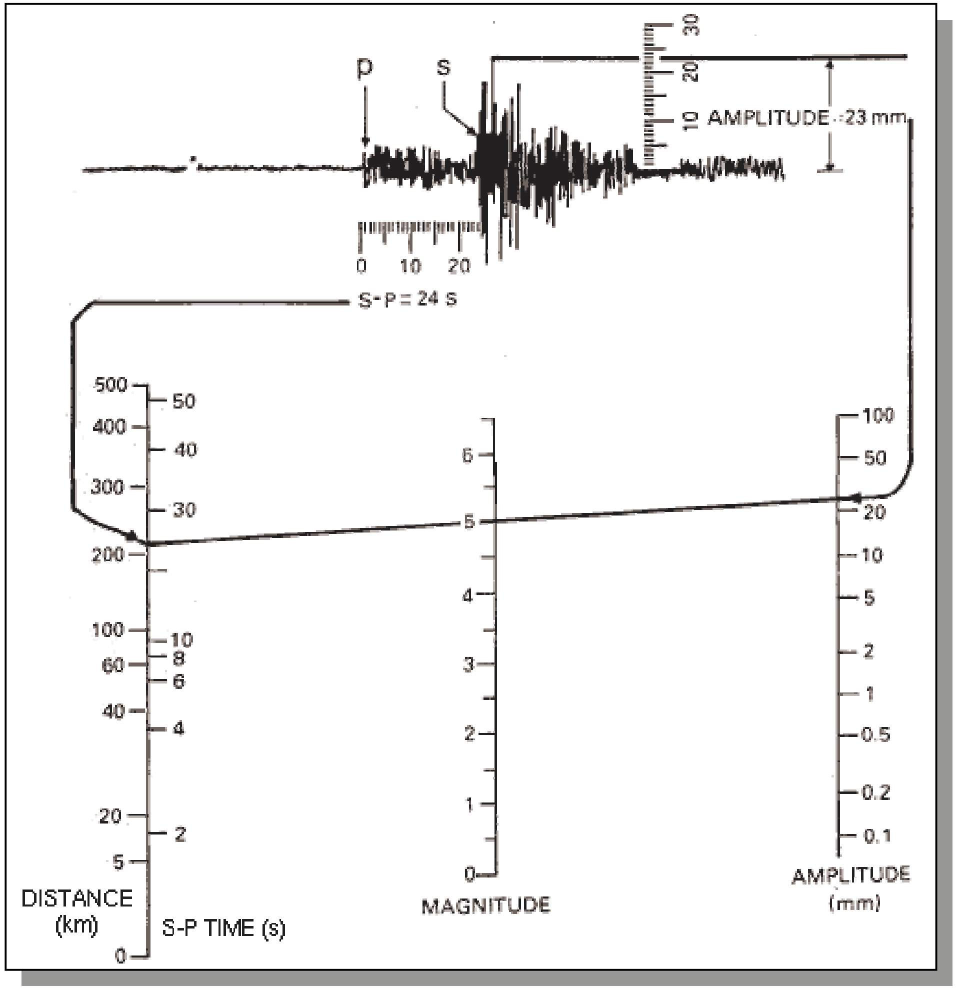
D in deg, A in micrometers

Body wave magnitude M_b

$M_b = \log(A/T) + Q(D,h)$ — $T=0.1\text{-}3.0$s

Definition: An earthquake recorded on a Wood-Anderson seismometer at a distance of 100km generating an amplitude of 1mm has Magnitude M_L=3.

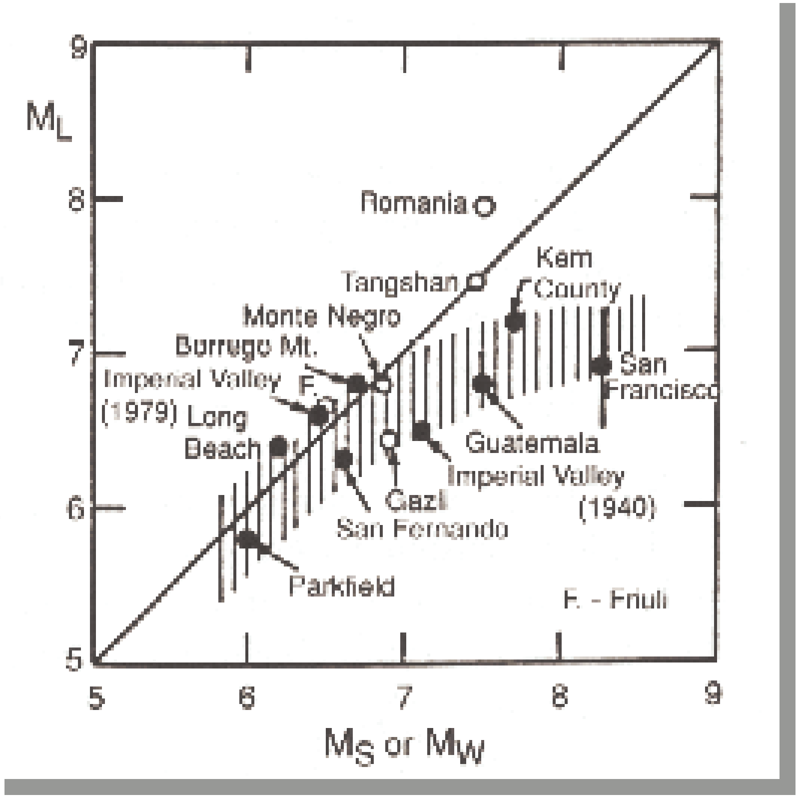
Richter Scale



Determination of the magnitude of an earthquake graphically.

$$M_L = \log_{10} A(mm) + (\text{Distance correction factor})$$

Saturation of Local Magnitude



For large earthquakes the originally defined Richter scale is not appropriate. Better indicators of the size of large earthquakes are the surface wave M_s scale or the energy scale M_w.

Seismic energy

Gutenberg-Richter developed a relationship between magnitude and energy (in ergs)

$$\log E_S = 11.8 + 1.5 M_{S,w}$$

Example: M8.3 earthquake

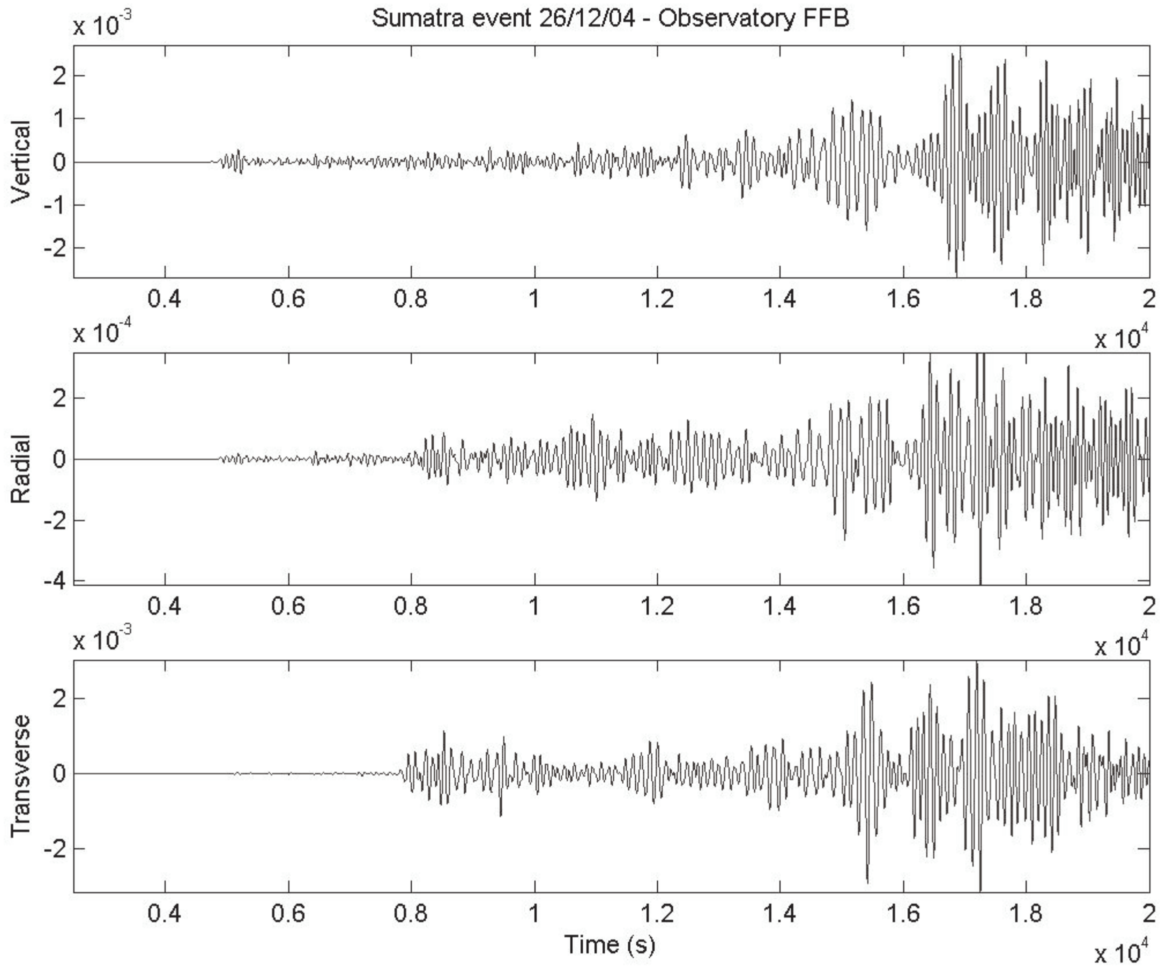
-> Energy $E_S = 10^{11.8+1.5*8.3}$ ergs

1 erg = 1 dyn-cm

Magnitudes and intensities
what to remember!

- The size of an earthquake (at the source) is described by its magnitude
- Magnitudes are determined from peak amplitudes of body waves (P,S) or surface waves
- Different magnitude definitions have to be used depending on the epicentral distance
- Earthquake magnitudes can be related empirically to the total physical energy that is released through an earthquake (in Nm or dyne-cm)
- The damage of an earthquake at a certain location is described by seismic intensities (in grades I-XII like wind speeds) -> Mercalli scale

Exercise (Sumatra quake)
Distance 84°, 9300km



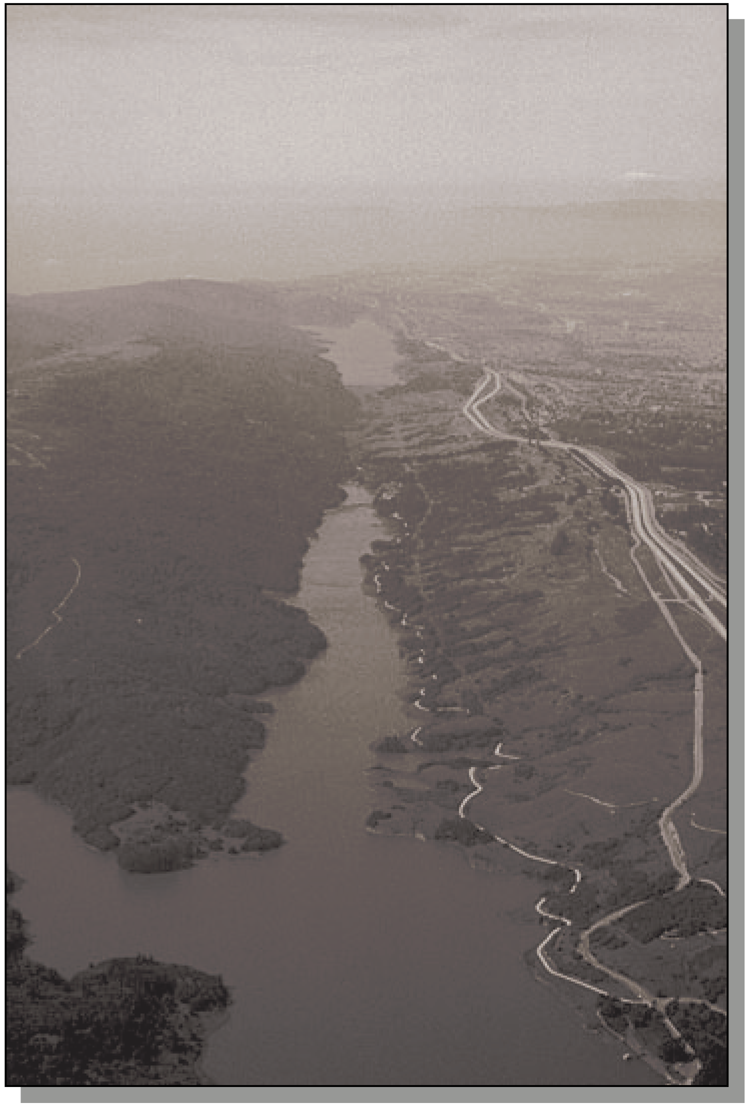
Seismic energy (Examples)

Richter Magnitude	TNT for Seismic Energy Yield	Example (approximate)
-1.5	6 ounces	Breaking a rock on a lab table
1.0	30 pounds	Large Blast at a Construction Site
1.5	320 pounds	
2.0	1 ton	Large Quarry or Mine Blast
2.5	4.6 tons	
3.0	29 tons	
3.5	73 tons	
4.0	1,000 tons	Small Nuclear Weapon
4.5	5,100 tons	Average Tornado (total energy)
5.0	32,000 tons	
5.5	80,000 tons	Little Skull Mtn., NV Quake, 1992
6.0	1 million tons	Double Spring Flat, NV Quake, 1994
6.5	5 million tons	Northridge, CA Quake, 1994
7.0	32 million tons	Hyogo-Ken Nanbu, Japan Quake, 1995; Largest Thermonuclear Weapon
7.5	160 million tons	Landers, CA Quake, 1992
8.0	1 billion tons	San Francisco, CA Quake, 1906
8.5	5 billion tons	Anchorage, AK Quake, 1964
9.0	32 billion tons	Chilean Quake, 1960
10.0	1 trillion tons	(San-Andreas type fault circling Earth)
12.0	160 trillion tons	(Fault Earth in half through center, OR Earth's daily receipt of solar energy)

Surface expressions of earthquakes: fault scarps

Fault scarps

California



Fault scarps

California

Fault scarps

Taiwan, Chi-Chi earthquake 1999

Fault scarps

Taiwan, Chi-Chi earthquake 1999

Seismic moment

... from point source to finite source ...

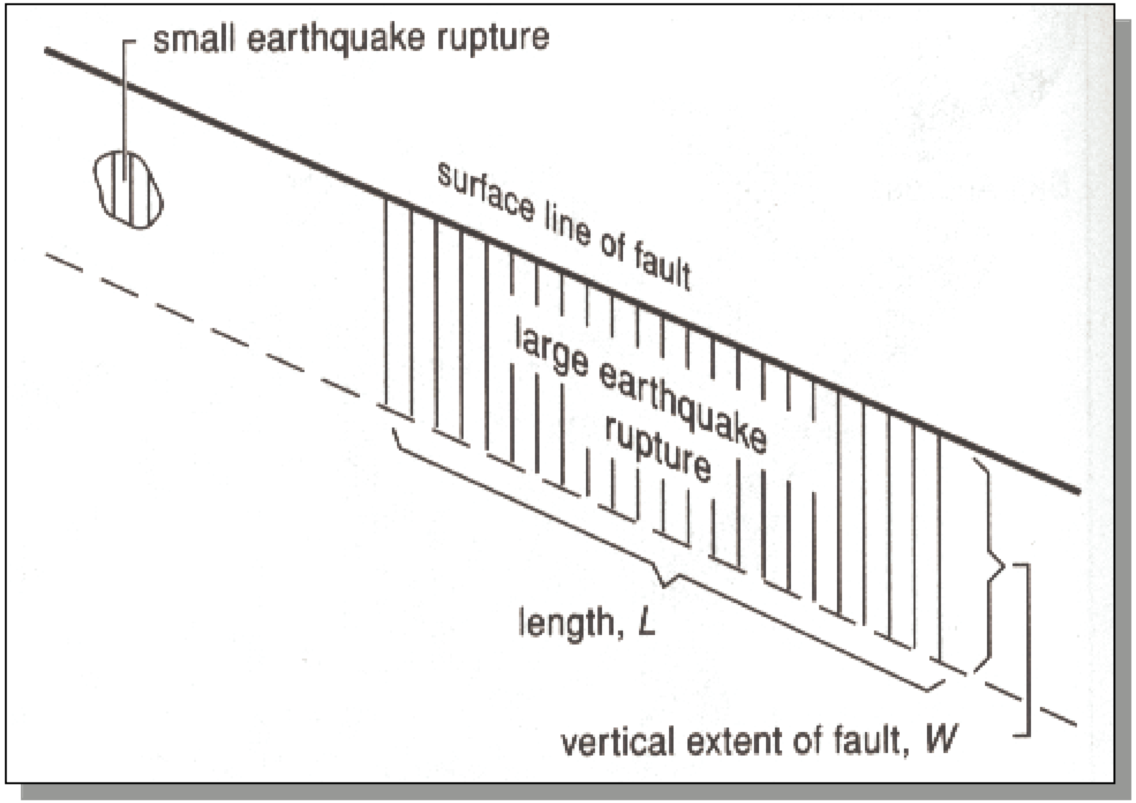
Seismologists measure the size of an earthquake using the concept of seismic moment. It is defined as the force times the distance from the center of rotation (torque). The moment can be expressed suprisingly simple as:

$$M_0 = \mu A d$$

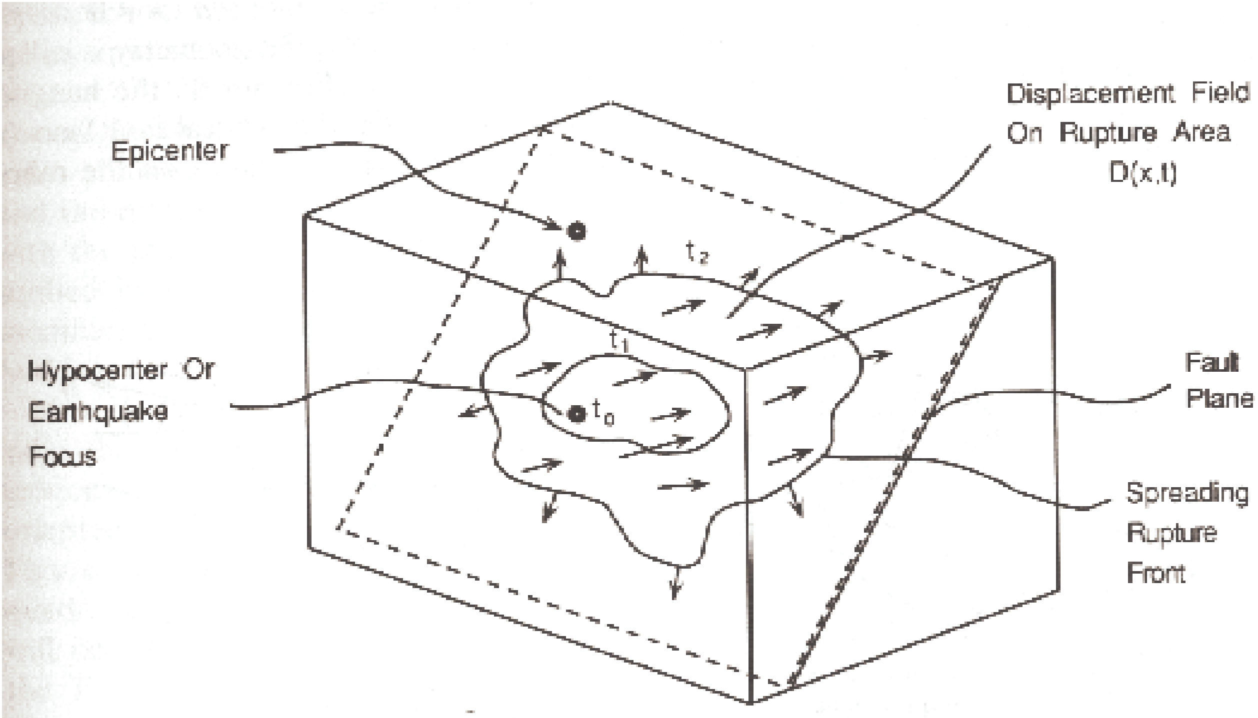
M_0	seismic moment
μ	Rigidity
A	fault area
d	slip/displacement

Seismic moment

$$M_0 = \mu A d$$

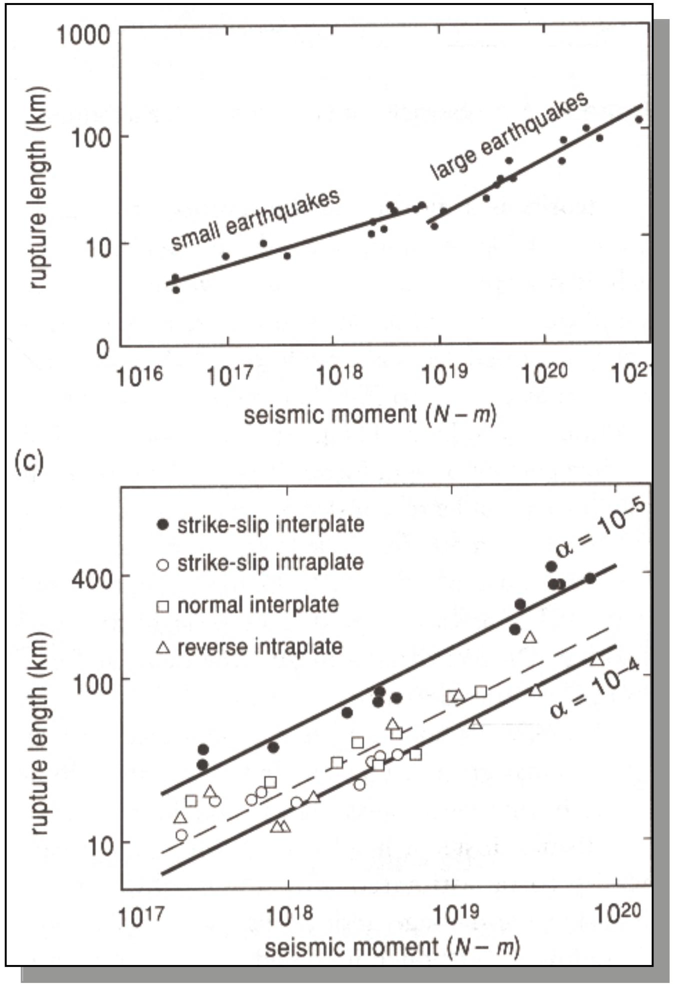


Fault Slip



Schematic diagram of rupture on a fault. All regions sliding radiate outgoing P- and shear waves. Note that the direction of rupture propagation is not in general parallel to the slip direction.

Seismic moment



$$M_0 = \mu A d$$

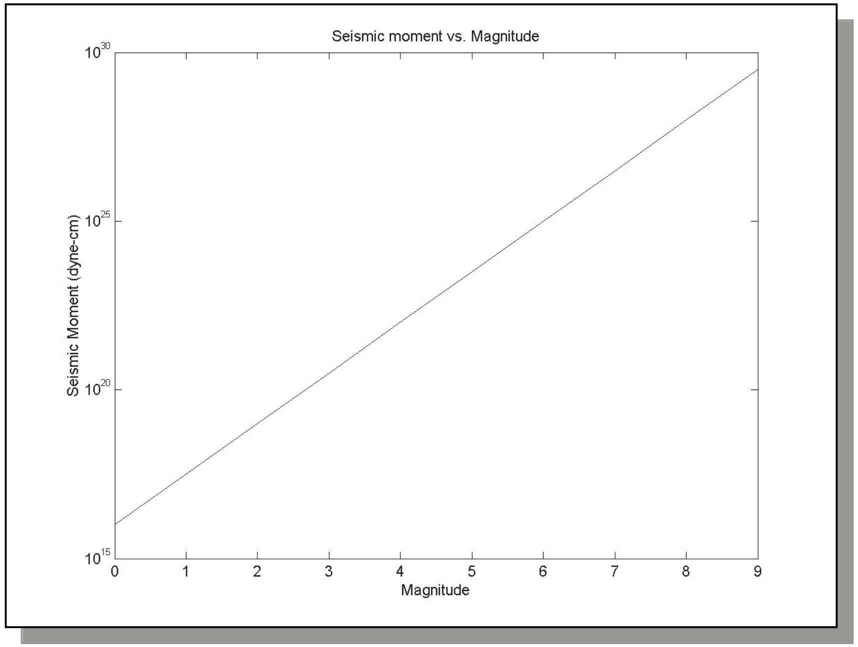
There are differences in the scaling of large and small earthquakes

Stress drop is larger in non-plate boundary (low deformation regimes)

Seismic moment - magnitude

There is a standard way of converting the seismic moment to magnitude M_w:

$$M_w = \frac{2}{3}\left[\log_{10} M_0(dyne - cm) - 16.0\right]$$



Source kinematics

Point source characteristics (source moment tensor, rise time, source moment, rupture dimensions) give us some estimate on what happened at the fault. However we need to take a closer look. We are interested in the space-time evolution of the rupture.

Here is the fundamental concept:

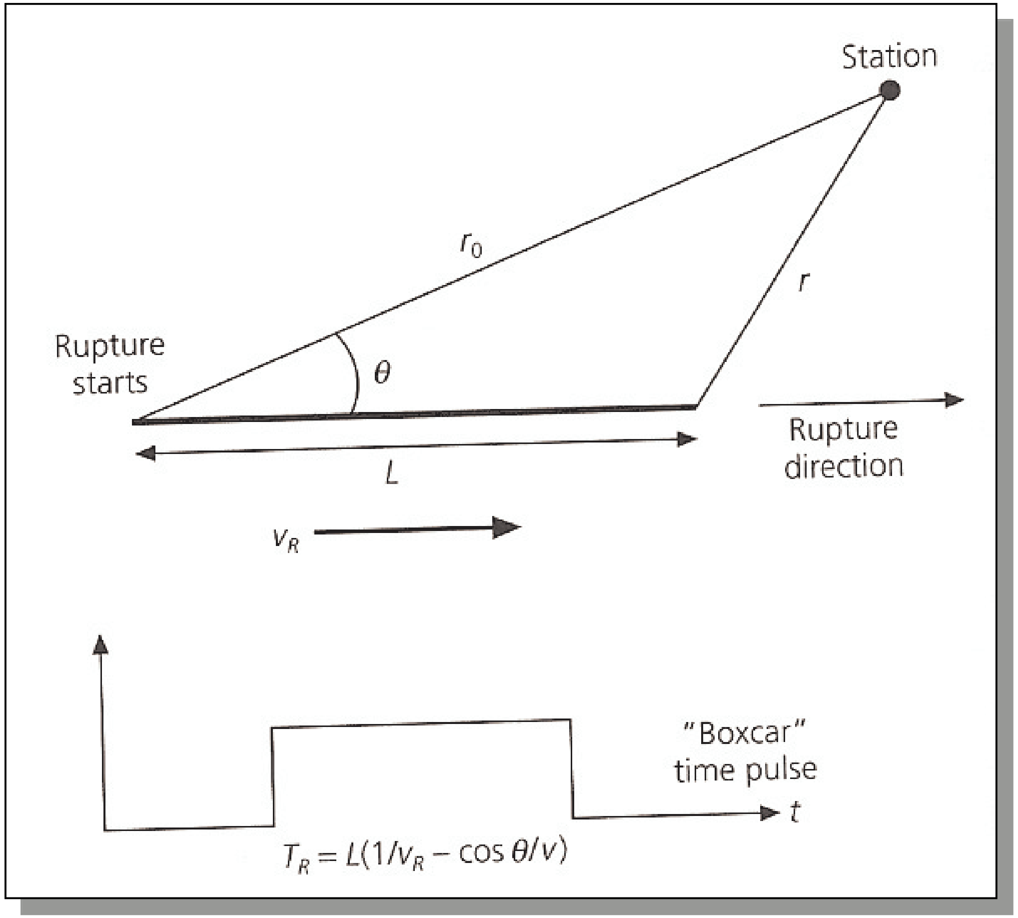
The recorded seismic waves are a superpositions of many individual double-couple point sources.

This leads to the problem of estimating this space-time behavior from observed (near fault) seismograms. The result is a kinematic description of the source.

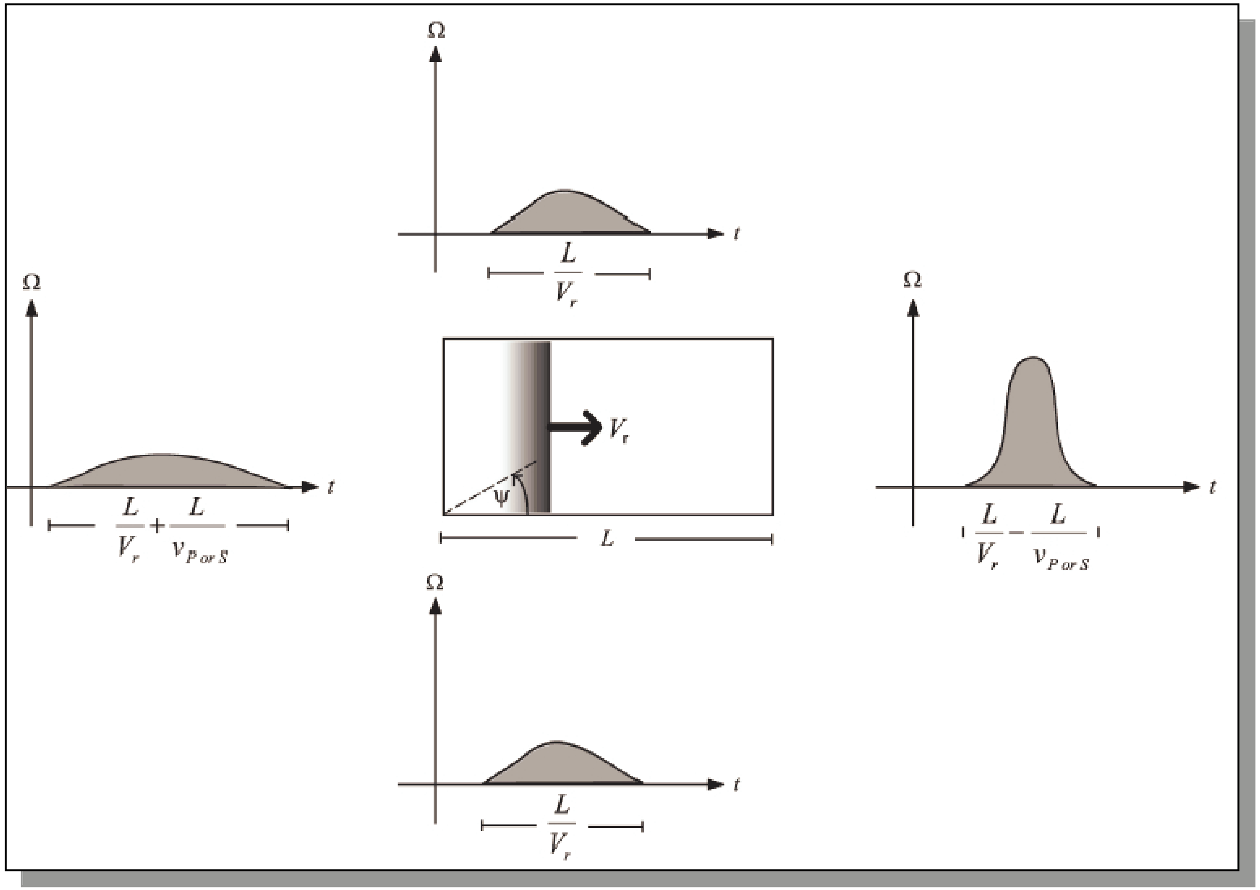
Source directivity

When a finite fault ruptures with velocity v_r, the time pulse is a boxcar with duration

$T_R = L(1/v_r - \cos(\theta/v))$



Source directivity



The energy radiation becomes strongly anisotropy (Doppler effect). In the direction of rupture propagation the energy arrives within a short time window.

Finite sources – seismo-tectonics

what to remember!

- (Large) earthquakes can be described as a superposition of DC point source on a fault plane
- The earthquake moment (thus magnitude and energy) is related to the fault area and the (average) slip of both fault planes
- The cumulative slip of earthquakes at major faults can be related to the tectonic deformation (observable at the surface with GPS)
- The observation of surface deformation (and the seismicity) is today the most important key to montior stress increase and thus predict seismic hazard

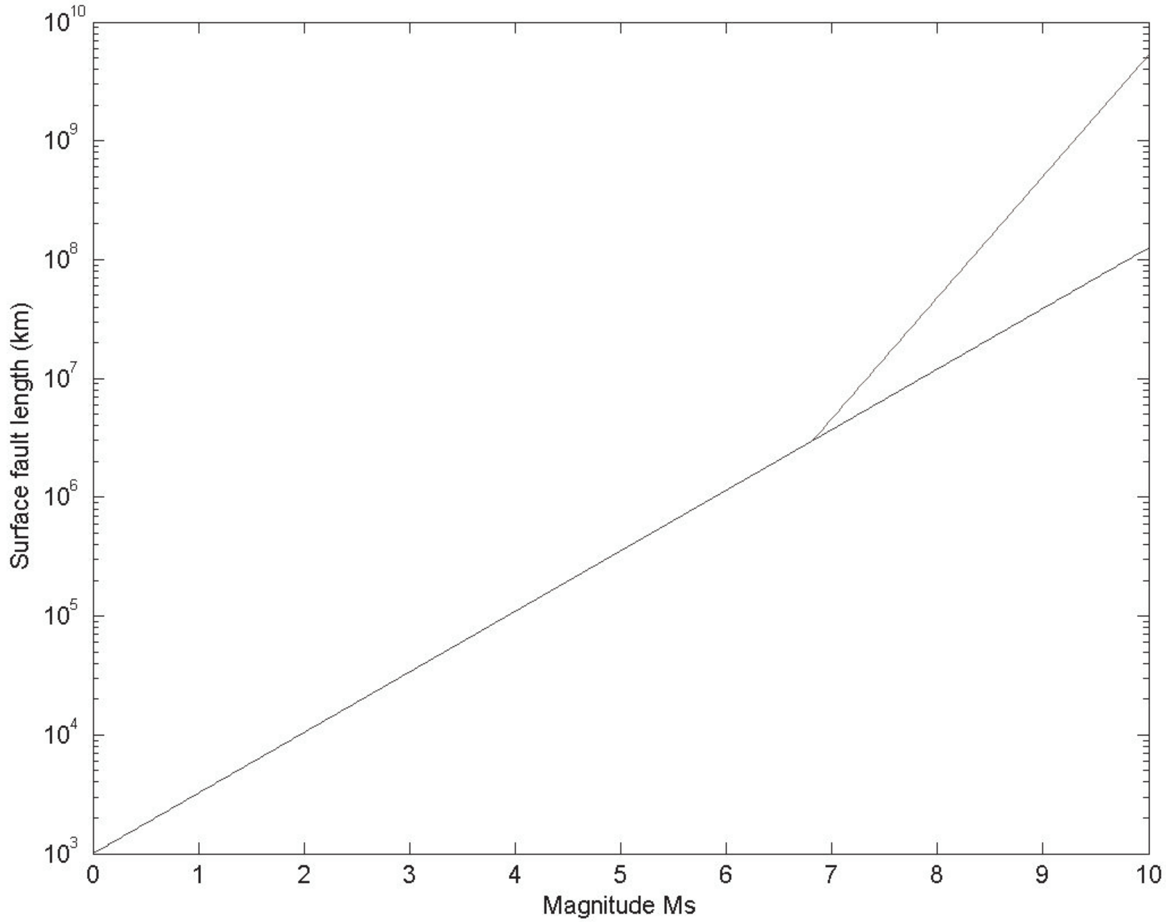
Exercise

Zwischen der Oberflächenwellenmagnitude M_S und der Bruchfläche A gibt es einen empirischen Zusammenhang der Form

$$\log A = 1.02 M_S + 6.0$$

wobei A in cm^2 angegeben wird. Skizzieren Sie die Länge des Bruchs (in logarithmischer Form) unter der Annahme einer quadratischen Bruchfläche für das Magnitudenintervall [0,10] (Geradengleichung). Wie ändert sich diese Grafik wenn die maximale vertikale Bruchausdehnung 30km beträgt? Leiten Sie zuerst die entsprechende Geradengleichung her. Welchen Bruchlänge erhalten Sie für das Sumatrabeben mit M=9.3? Woran könnte es liegen, dass die Werte nicht übereinstimmen? Welche Magnitude hätte ein Beben, welches die Erde in zwei Hälften teilt (Bruchfläche wäre Ebene durch Erdmittelpunkt).

Solution



Surface fault length (km)
Magnitude Ms
10^{10}
10^{9}
10^{8}
10^{7}
10^{6}
10^{5}
10^{4}
10^{3}
0
1
2
3
4
5
6
7
8
9
10

Exercise

Die seismische Energie eines Erdbebens kann als Funktion von $M_{S,w}$ berechnet werden.

$$\log E = 11.8 + 1.5 M_{S,w}$$

E ist hier in erg (dyn-cm) gegeben. Welche Konsequenz hat diese Beziehung für das Verhältnis der Energien zweier benachbarter Magnituden (zB. M5 und M6)? Berechnen Sie aus der beigefügten Tabelle aller Beben>M8 seit 1975 die kumulative Energie der Jahre 2000-2005 in erg. Welchen Anteil daran hatten die beiden Sumatra Beben M9.0 und 8.7 Beben der letzten Monate?

Mercalli Intensity and Richter Magnitude

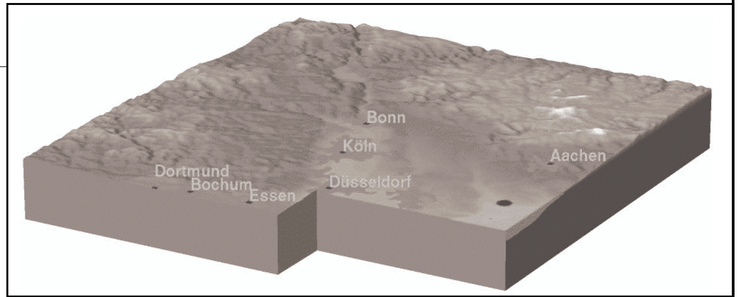
Magnitude	Intensity	Description
1.0-3.0	I	**I.** Not felt except by a very few under especially favorable conditions.
3.0 - 3.9	II - III	**II.** Felt only by a few persons at rest, especially on upper floors of buildings. **III.** Felt quite noticeably by persons indoors, especially on upper floors of buildings. Many people do not recognize it as an earthquake. Standing motor cars may rock slightly. Vibrations similar to the passing of a truck. Duration estimated.
4.0 - 4.9	IV - V	**IV.** Felt indoors by many, outdoors by few during the day. At night, some awakened. Dishes, windows, doors disturbed; walls make cracking sound. Sensation like heavy truck striking building. Standing motor cars rocked noticeably. **V.** Felt by nearly everyone; many awakened. Some dishes, windows broken. Unstable objects overturned. Pendulum clocks may stop.
5.0 - 5.9	VI - VII	**VI.** Felt by all, many frightened. Some heavy furniture moved; a few instances of fallen plaster. Damage slight. **VII.** Damage negligible in buildings of good design and construction; slight to moderate in well-built ordinary structures; considerable damage in poorly built or badly designed structures; some chimneys broken.
6.0 - 6.9	VII - IX	**VIII**. Damage slight in specially designed structures; considerable damage in ordinary substantial buildings with partial collapse. Damage great in poorly built structures. Fall of chimneys, factory stacks, columns, monuments, walls. Heavy furniture overturned. **IX.** Damage considerable in specially designed structures; well-designed frame structures thrown out of plumb. Damage great in substantial buildings, with partial collapse. Buildings shifted off foundations.
7.0 and higher	VIII or higher	**X.** Some well-built wooden structures destroyed; most masonry and frame structures destroyed with foundations. Rails bent. **XI.** Few, if any (masonry) structures remain standing. Bridges destroyed. Rails bent greatly. **XII.** Damage total. Lines of sight and level are distorted. Objects thrown into the air.

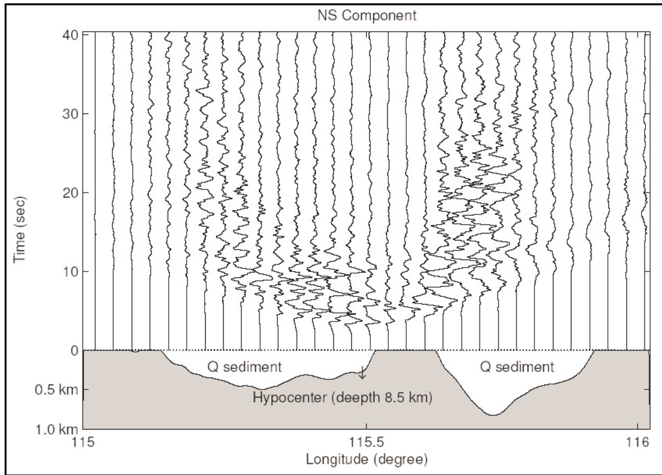
Erdbebengefährdung/Risiken

Fragestellungen:
- Kennen wir die Erdbebengefährdung-Risiken spezifischer Regionen?
- Können wir die Bodenbewegungen möglicher Erdbebenszenarien zuverlässig vorhersagen?
- Welche Information/Methodik ist notwendig, um diese Berechnungen zuverlässig zu machen?
- Wie kann diese Information an risiko-relevante Regionen vermittelt werden?

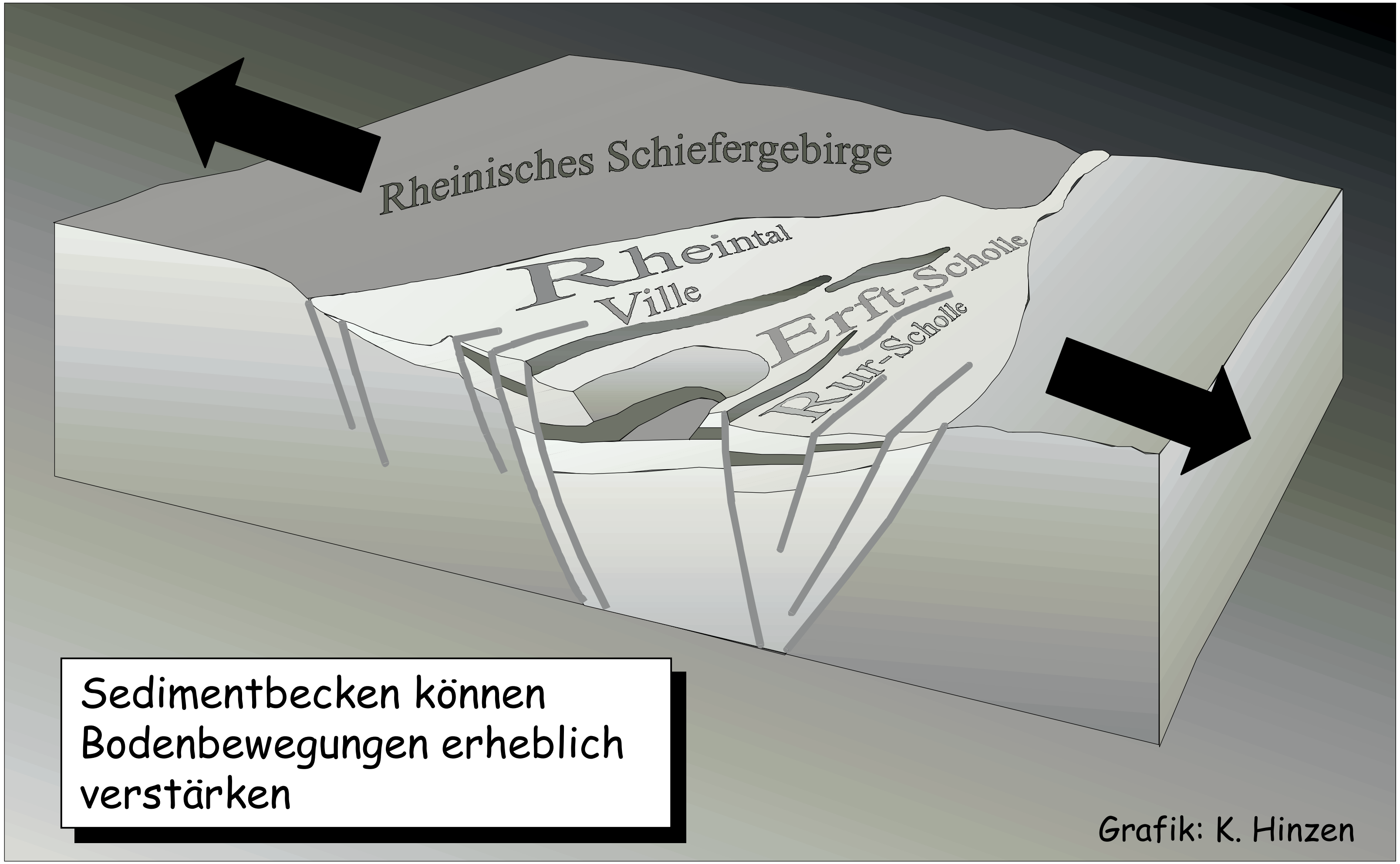
Disziplinen:
- Seismologie – Erdbebeningenieurwesen – Geologie – (Satelliten-)Geodäsie – (Neo-) Tektonik – Paläoseismologie – Mechanik-Geomorphologie - Statistik

Supercomputing:
- Berechnung von Erdbebenszenarien bis zu Frequenzen, die für Gebäudeschäden relevant sind (ca. 1-3 Hz)





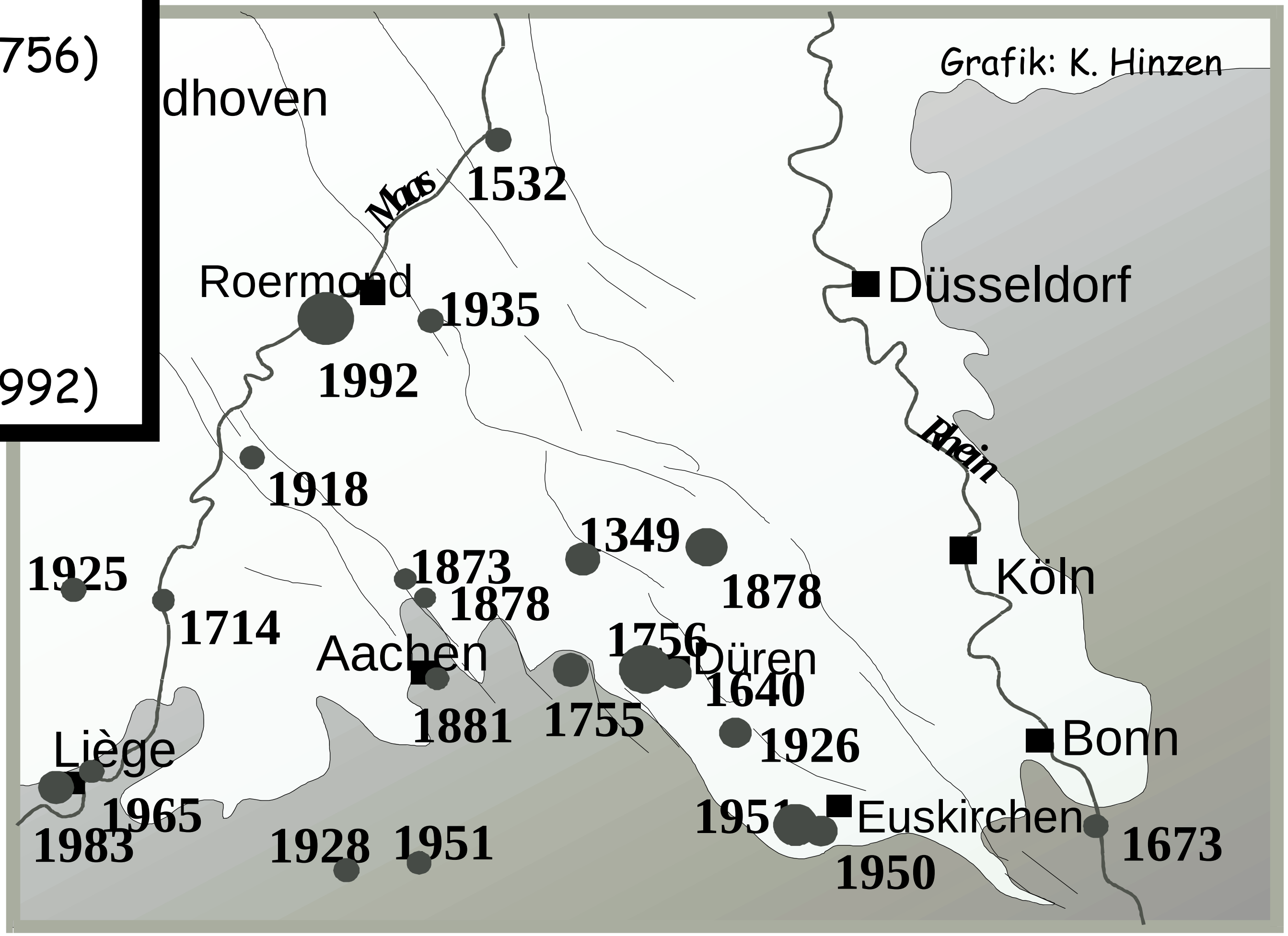
Tektonik des Kölner Beckens



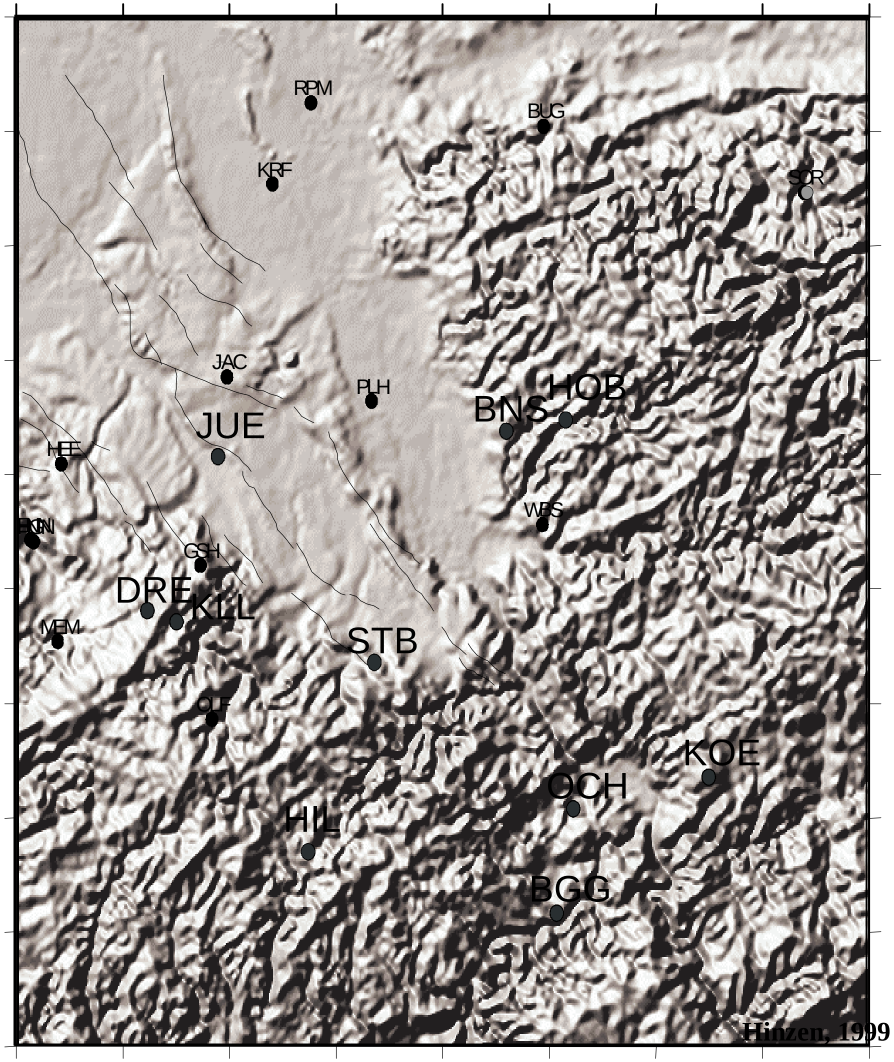
Erdbeben im Kölner Becken

Größtes Ereignis
M_L=6.4 (1756)

Letztes größeres Beben:
M_L=5.9 (1992)



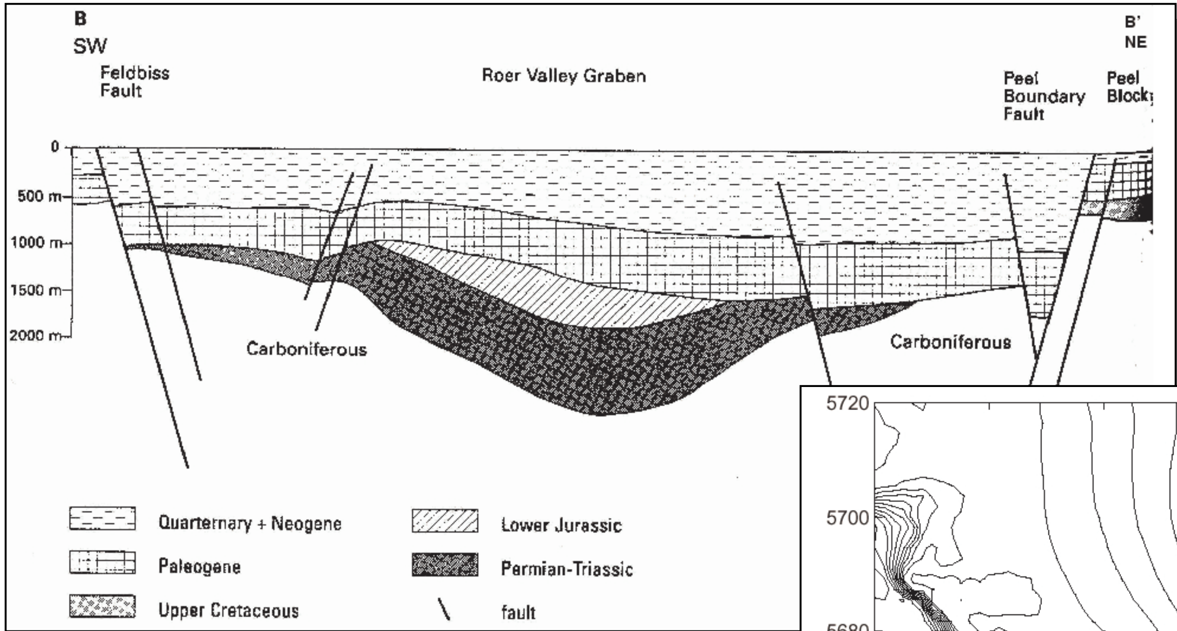
Bekannte Verwerfungen

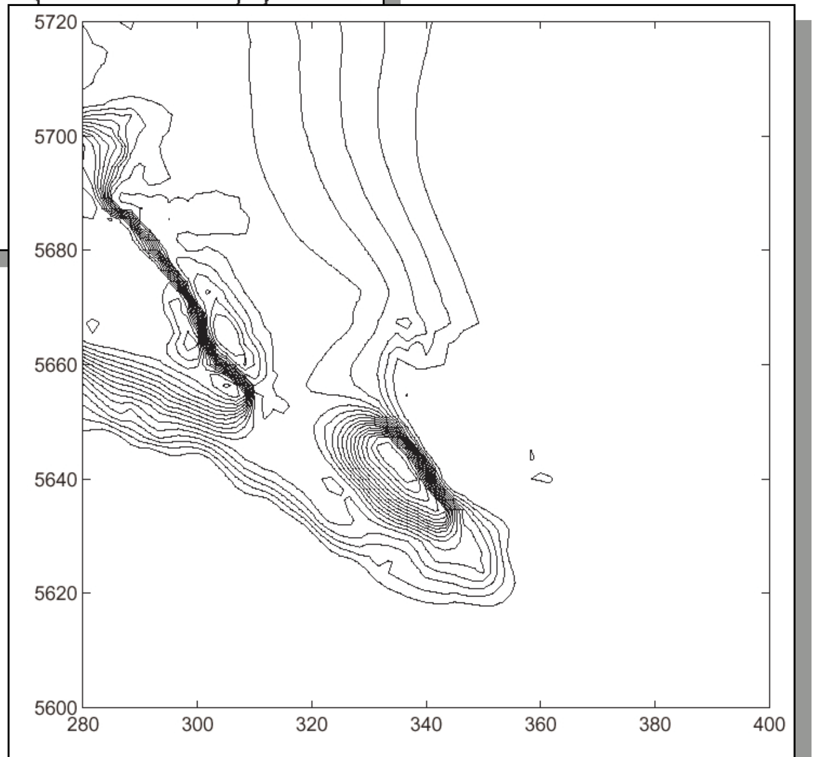


Die Paläoseismologie, sowie die Untersuchung oberflächennaher Strukturen ermöglichen die Kartierung (aktiver) Verwerfungen:

Damit wird die Simulation realistischer Erdbebenszenarien möglich.

Sedimentstruktur

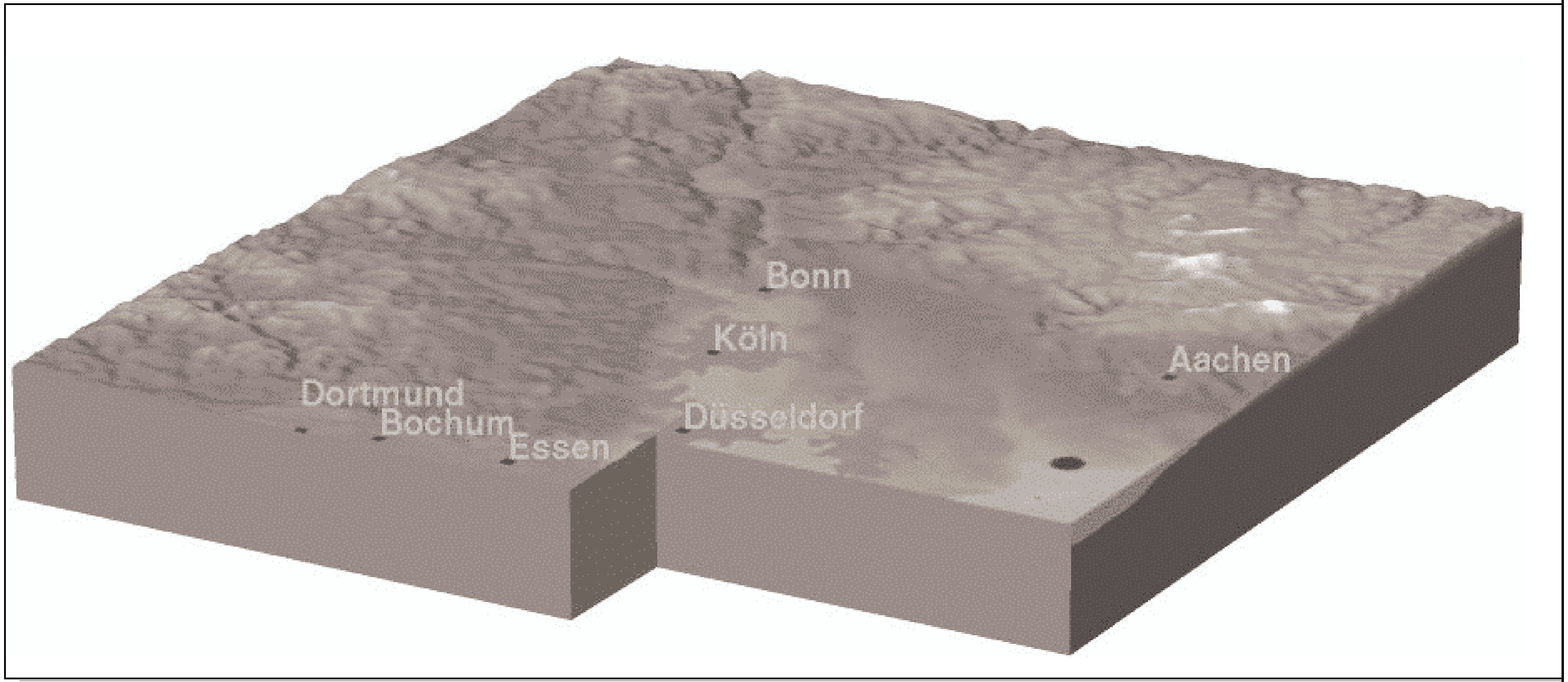




Topographie der
Sedimentunterkante

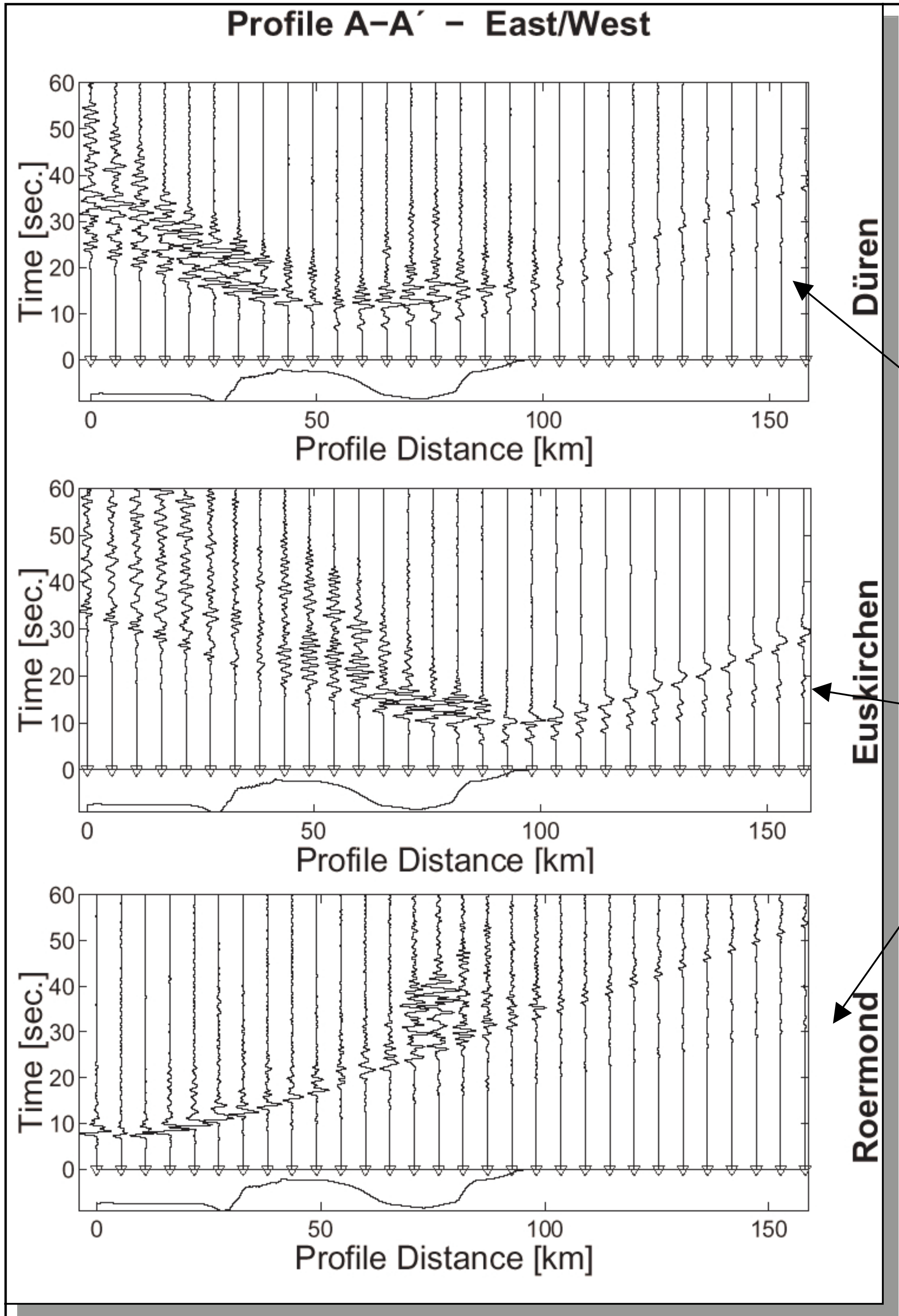
Erdbeben-Szenarien

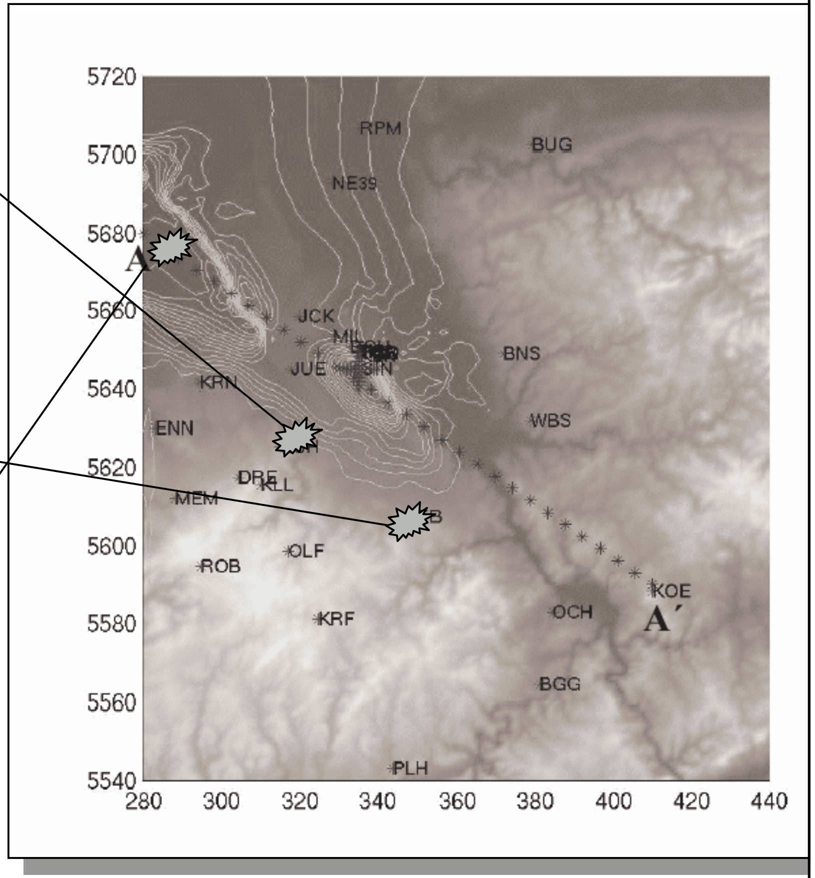
Roermond – Erdbeben M5.9, 1992



Positive (red) and negative (blue) horizontal ground velocity

Bodenbewegungen im Kölner Becken

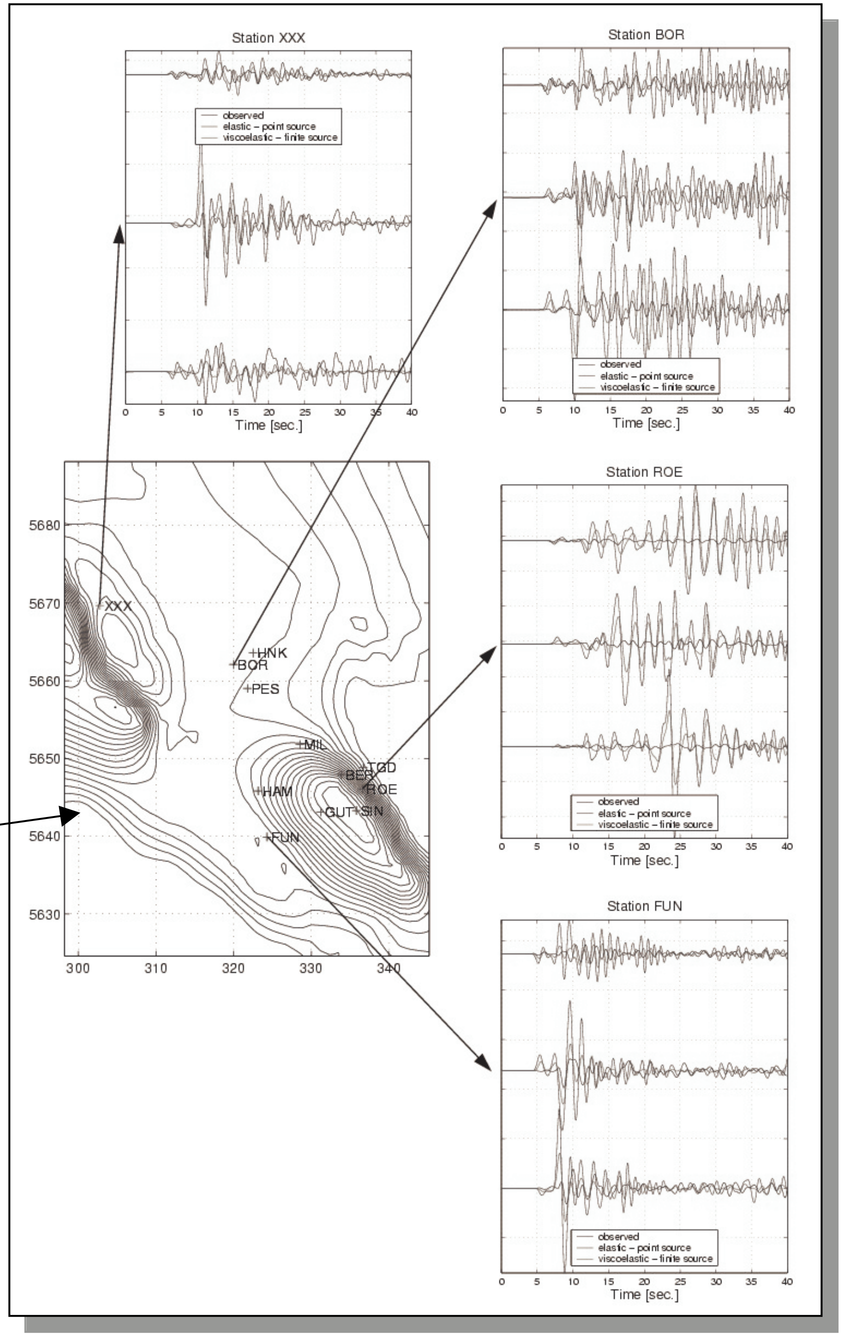




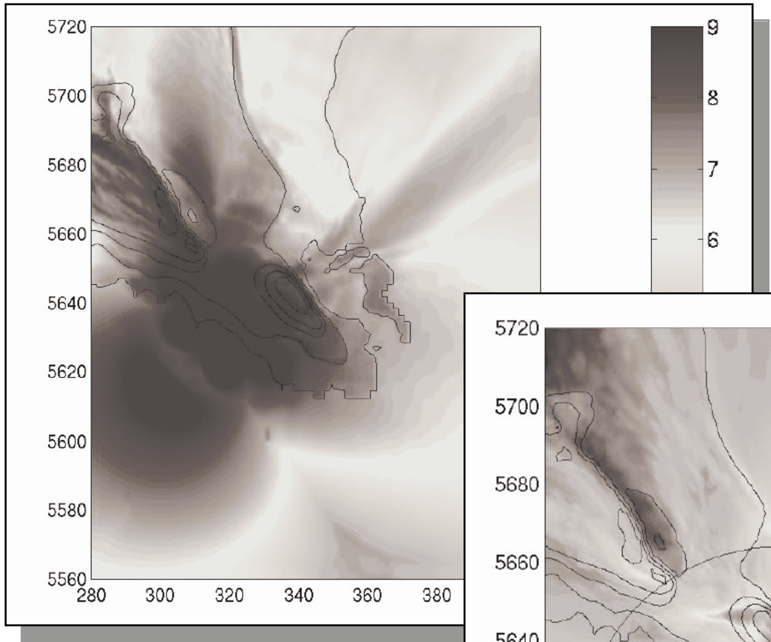
Vergleich mit Beobachtungen

M4.9, Juli 2002
Kölner Becken, Germany

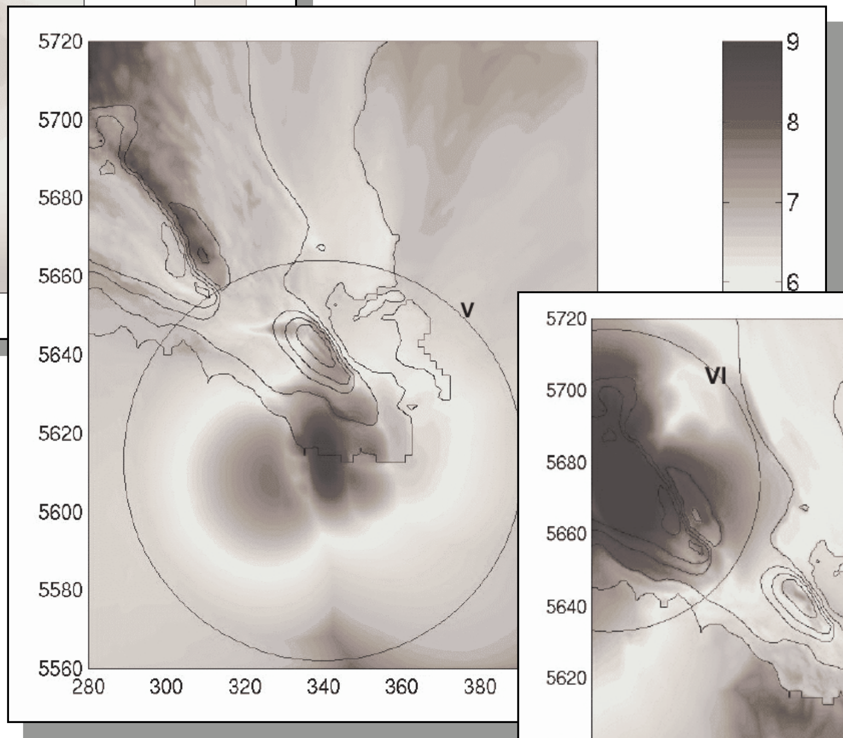
Beckentopographie und
seismische Stationen



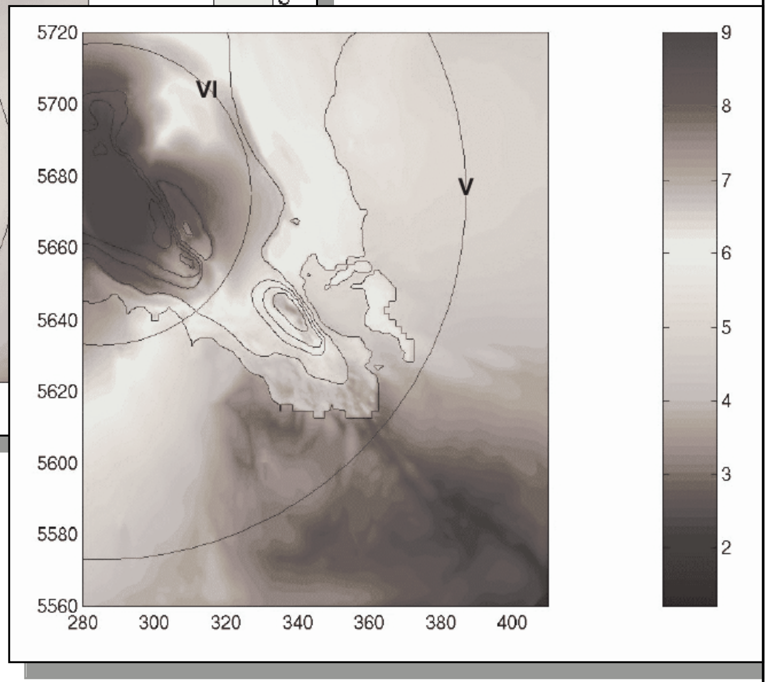
Seismische Intensitäten
(Mercalli scale)



Euskirchen



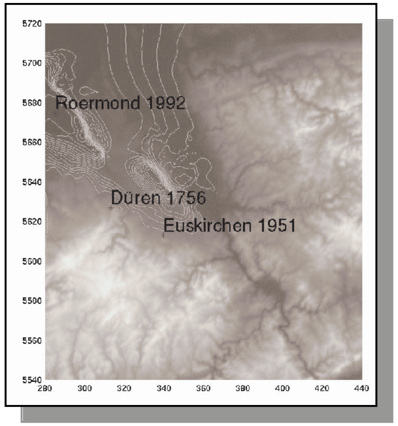
Dueren

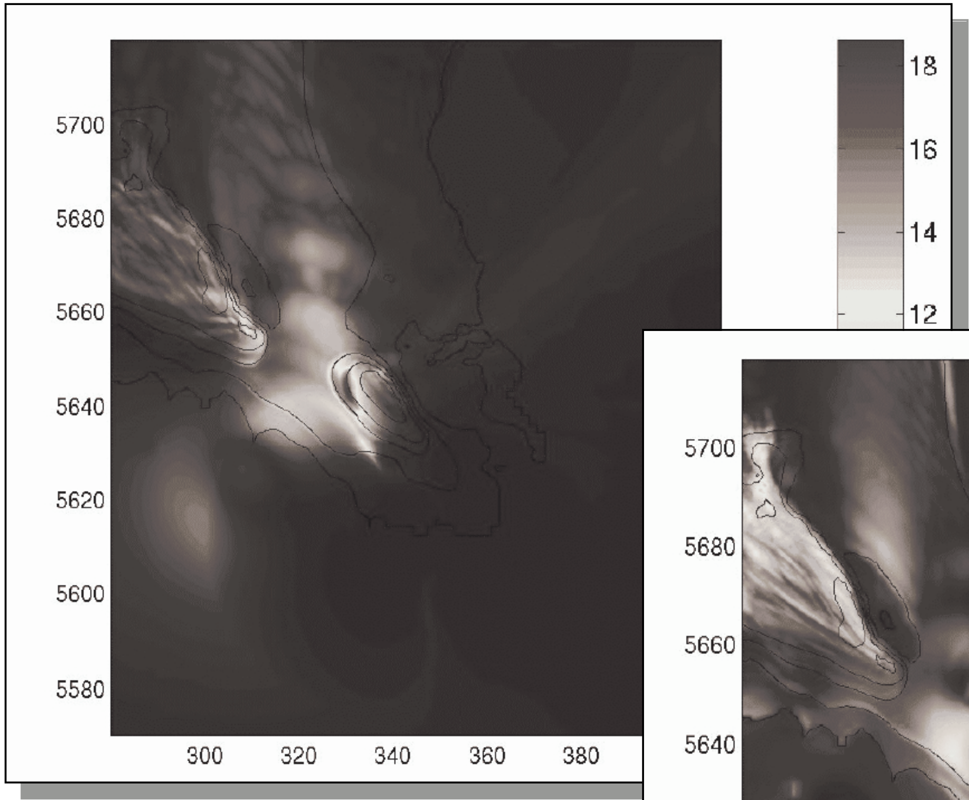


Roermond

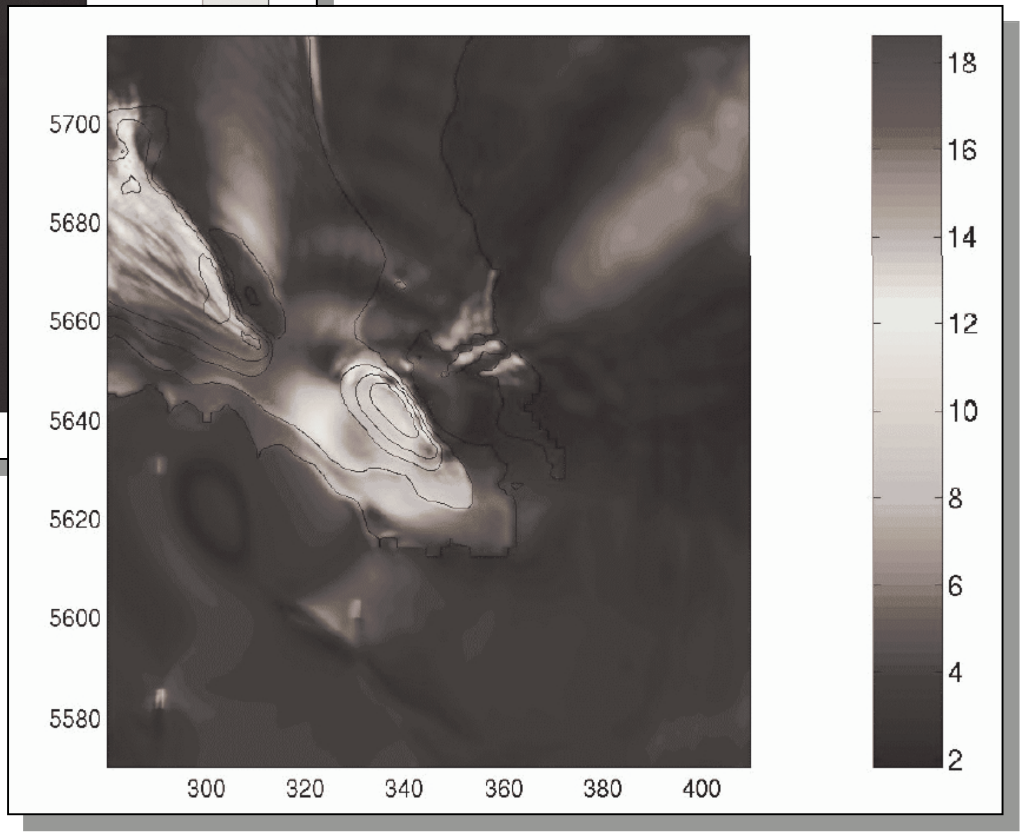
Die Intensitäten werden aus den maximalen Bodenbewegungen an jedem Ort berechnet

Verstärkung im Vergleich zu geschichtetem Modell



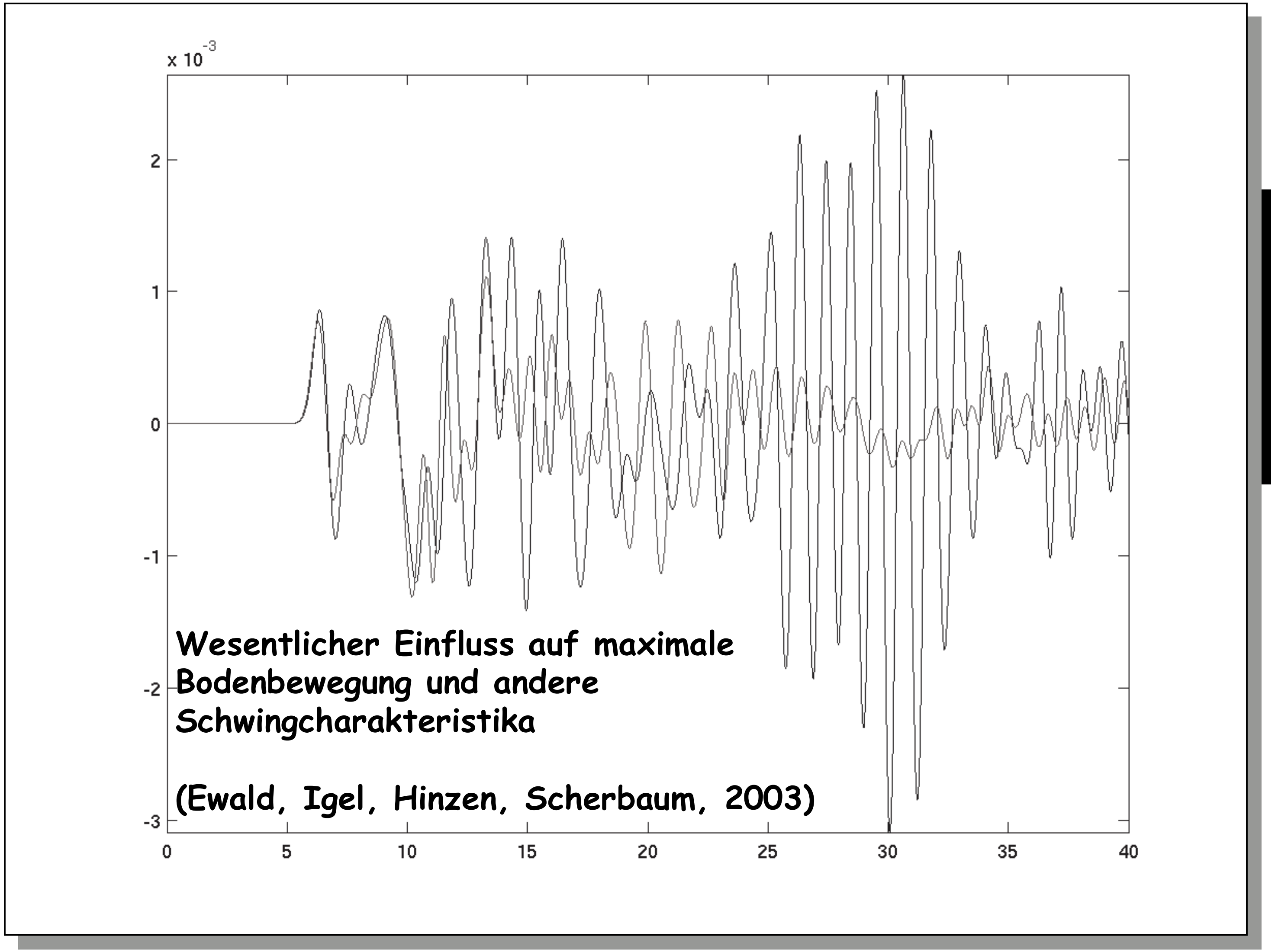


Dueren



Euskirchen

Was machen wir mit den Unsicherheiten in Modell und Quelle?



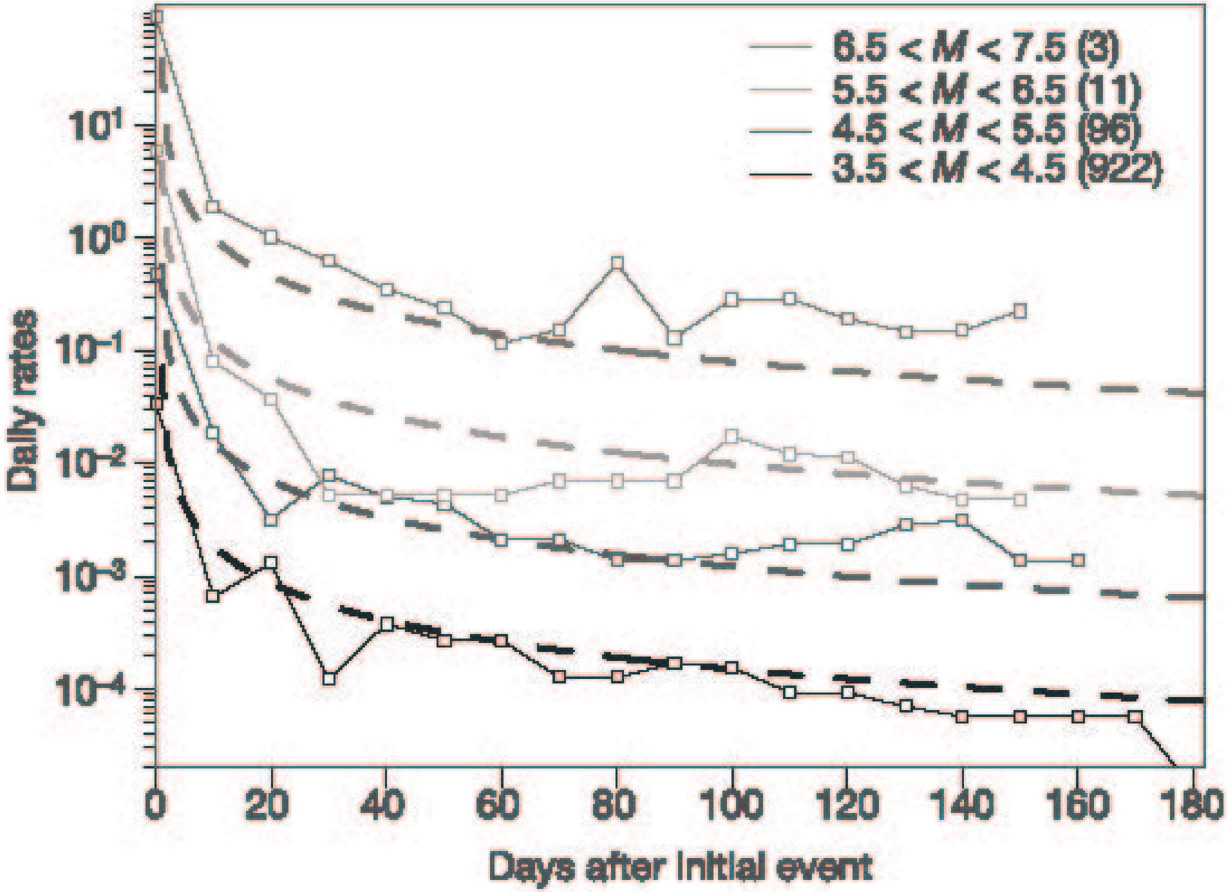


Figure 3 | Calculated and observed rates of events $M \geq 4$ in 24-hour intervals following mainshocks occurring between 1988 and 2002 in southern California. Dashed lines show the rates forecasted by the generic California clustering model (without cascades) for the mainshock magnitude (M) shown. For this test a simple circular aftershock zone implementation (solid lines) gives the observed rates of $M \geq 4.0$ aftershocks following all mainshocks with magnitude within 0.5 units of M. The aftershock zones are defined as the areas within one rupture length of the mainshock epicentre.

Earthquake prediction: recent probabilistic approach

Gerstenberger et al., Nature (2005)

Earthquake prediction: recent probabilistic approach

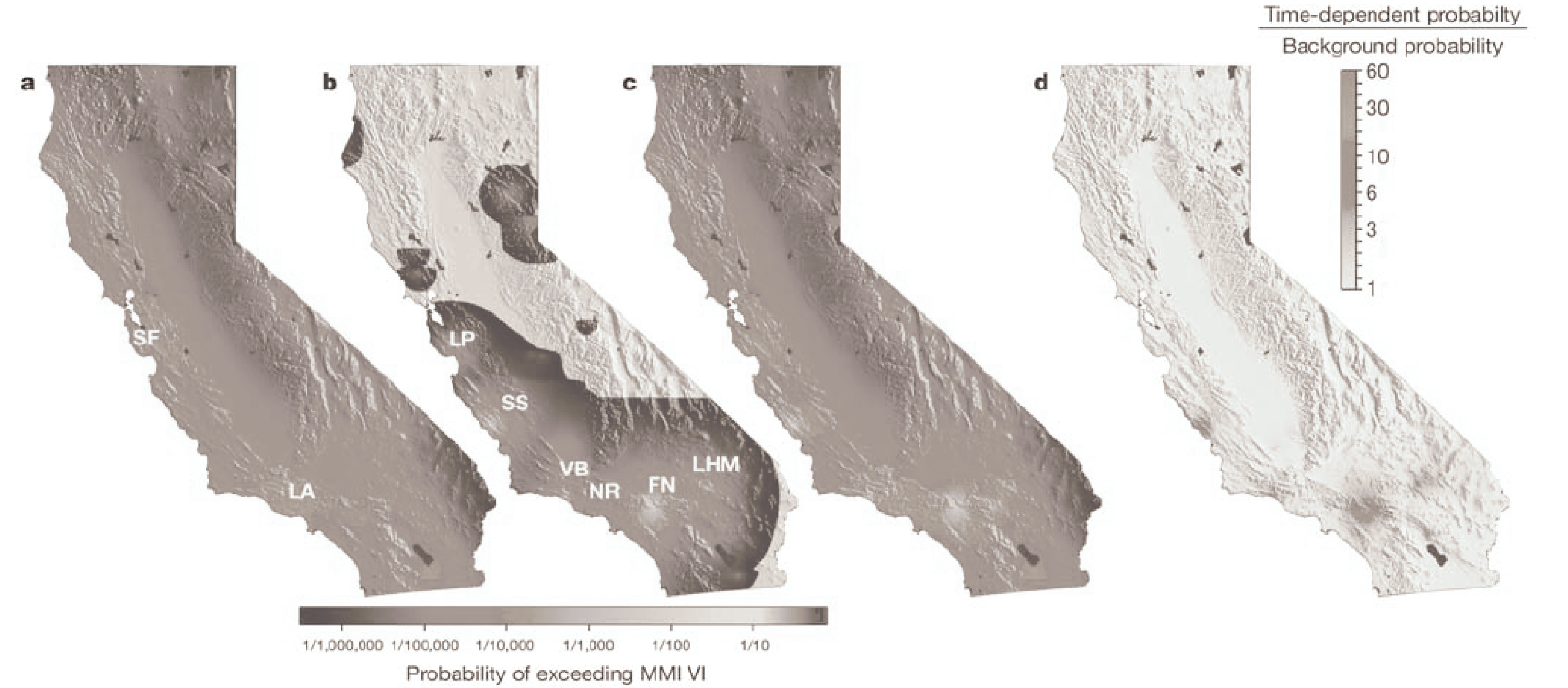


Figure 1 | Maps of California, showing the probability of exceeding MMI VI over the next 24-h period. The period starts at 14:07 Pacific Daylight Time on 28 July 2004. **a**, The time-independent hazard based on the 1996 USGS hazard maps for California. SF and LA are the locations of San Francisco and Los Angeles, respectively. **b**, The time-dependent hazard which exceeds the background including contributions from several events: 22 December 2003, San Simeon (SS, $M_w = 6.5$), a $M_l = 4.3$ earthquake 4 days earlier near Ventura (VB), a $M_l = 3.8$ event 30 min before the map was made near San Bernardino (FN), the 1999 Hector Mine $M_w = 7.1$ earthquake in the Mojave desert (LHM), and the 1989 $M_w = 6.9$ Loma Prieta (LP) earthquake. **c**, The combination of these two contributions, representing the total forecast of the likelihood of ground shaking in the next 24-hour period. **d**, The ratio of the time-dependent contribution to the background.

Gerstenberger et al., Nature (2005)

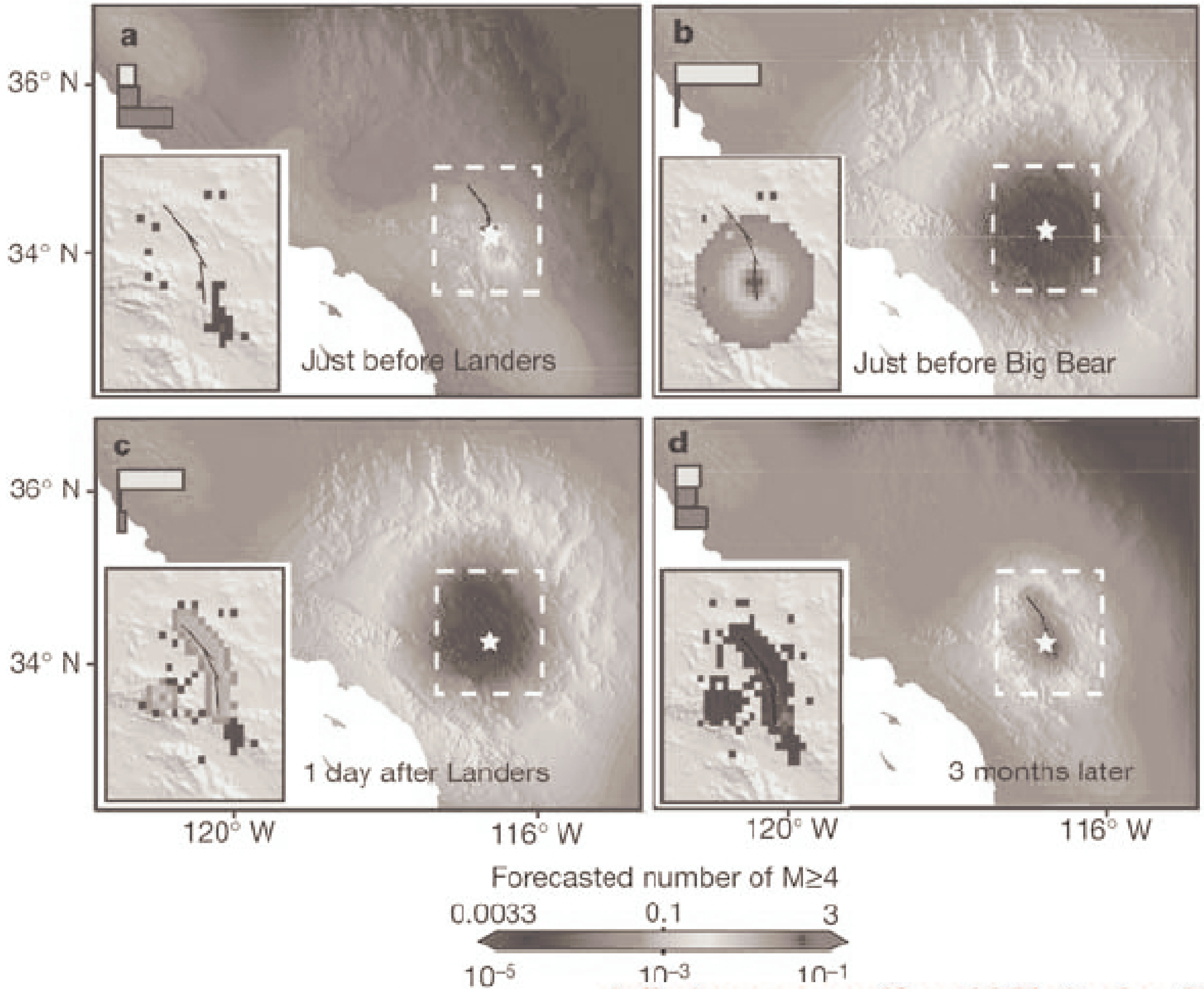


Figure 2 | Hazard maps calculated for four 24-hour intervals in the vicinity of the 1992 Landers ($M = 7.3$) earthquake. The colour shows the probability of exceeding MMI VI shaking intensity during interval (labels below colour scale). The black line indicates the Landers rupture and the white star is the Landers epicentre. The inset maps correspond to area within dashed white lines and indicate expected number of $M \geq 4.0$ aftershocks in the 24-hour interval (labels above colour scale). Histograms on upper left of figures indicate relative mean weight over inset area given to generic model (yellow), sequence-specific model (blue) and spatially heterogeneous mod (red). **a**, Interval beginning 1 min before Landers mainshock shows increased hazard south of the Landers epicentre associated with the foreshocks to Landers (largest $M_l = 3.0$) and aftershocks of the ($M_w = 6.$ Joshua Tree earthquake two months earlier. **b**, Interval beginning 2 hour after the mainshock and 1 hour before the ($M_w = 6.3$) Big Bear aftersho **c**, Interval beginning 24 h after Landers mainshock. **d**, Interval beginning three months after mainshock.