

Recap for SEISMIC RISK: SEISMIC WAVES

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Some basic definitions - 1



Principles of mechanics applied to bulk matter:

Mechanics of fluids

Mechanics of solids

Continuum Mechanics

A material can be called **solid** (rather than -perfect- fluid) if it can support a **shearing force** over the time scale of some natural process.

Shearing forces are directed parallel, rather than perpendicular, to the material surface on which they act.



Some basic definitions - 2



When a material is loaded at sufficiently low temperature, and/or short time scale, and with sufficiently limited stress magnitude, its deformation is fully recovered upon unloading:
the material is **elastic**

If there is a permanent (plastic) deformation due to exposition to large stresses:
the material is **elastic-plastic**

If there is a permanent deformation (viscous or creep) due to time exposure to a stress, and that increases with time:
the material is **viscoelastic** (with elastic response), or
the material is **visco-plastic** (with partial elastic response)

Stress as a measure of Force

Normal stress acts perpendicular to the surface

(F =normal force)



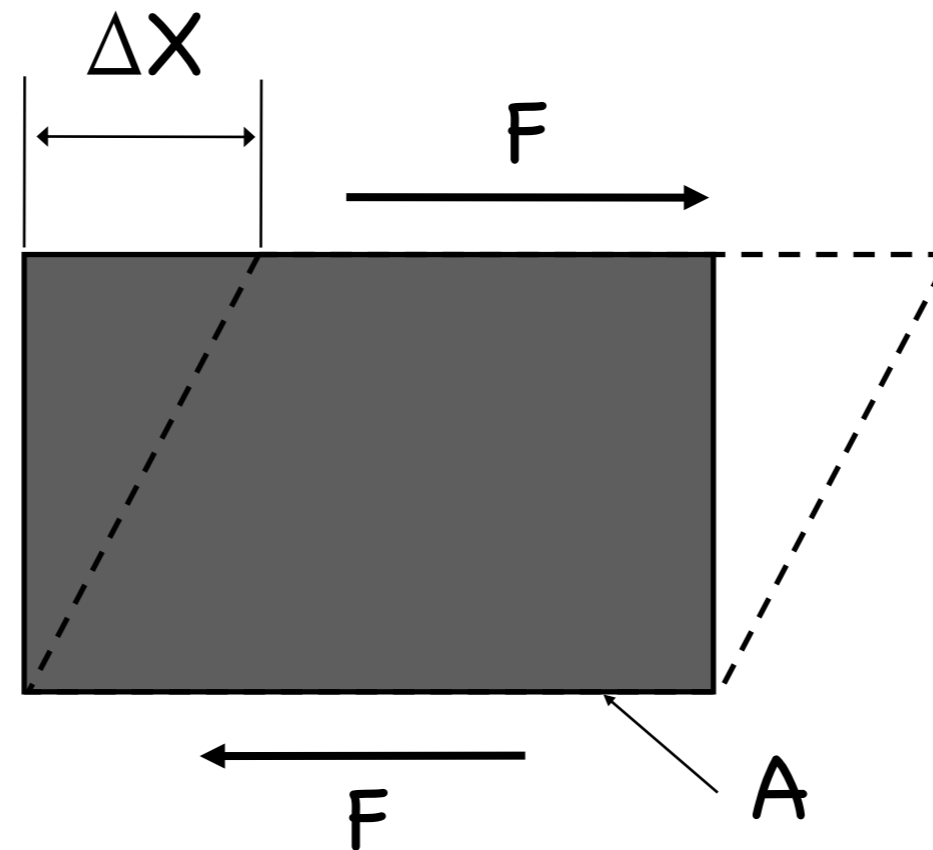
Tensile causes elongation



Compressive causes shrinkage

$$\sigma = \frac{\text{stretching force}}{\text{cross sectional area}}$$

Shear Stress as a measure of Force



$$\tau = \frac{\text{shear force}}{\text{tangential area}}$$

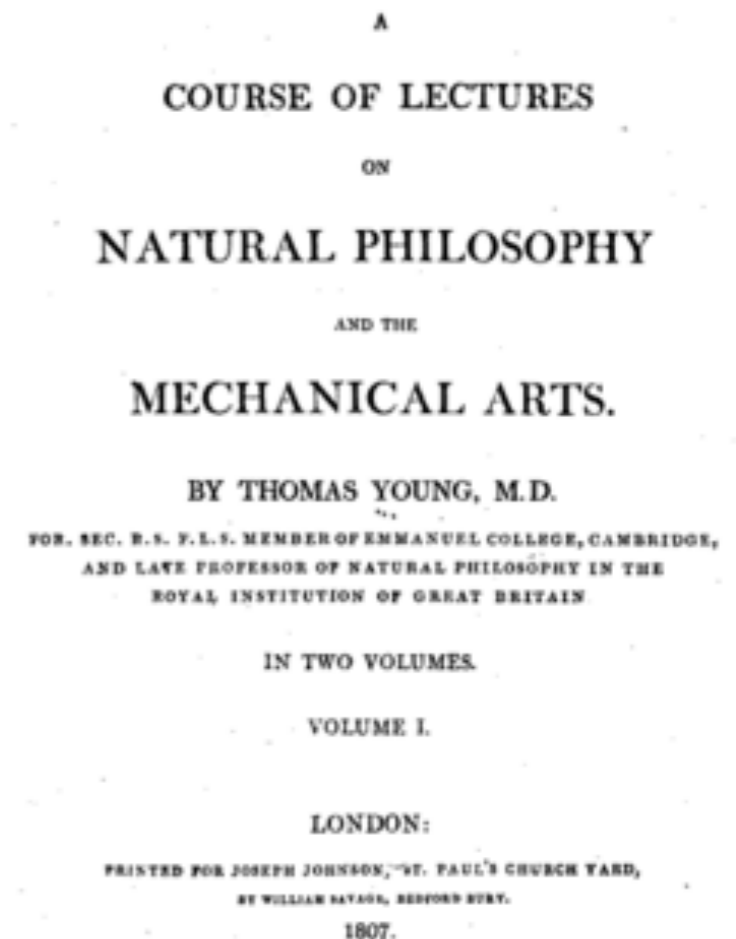
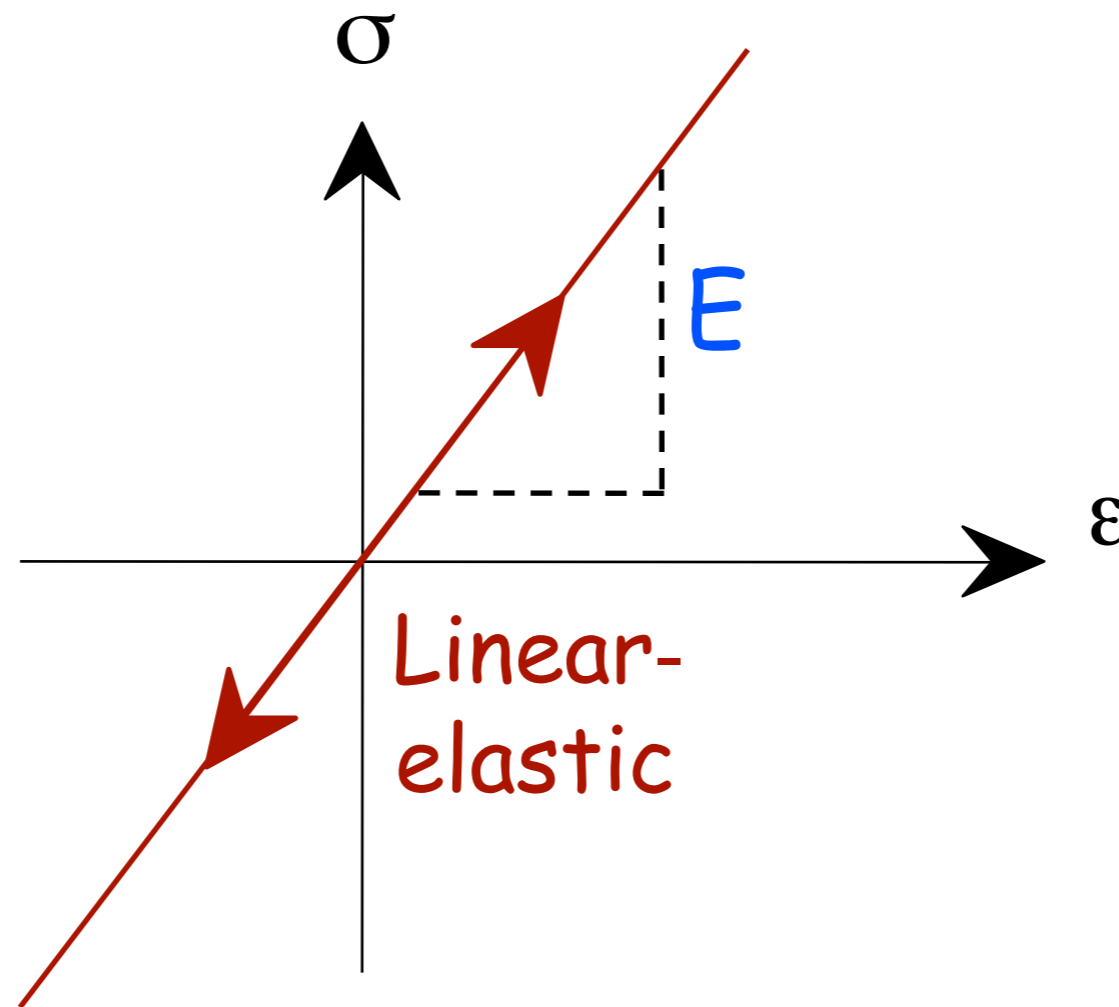
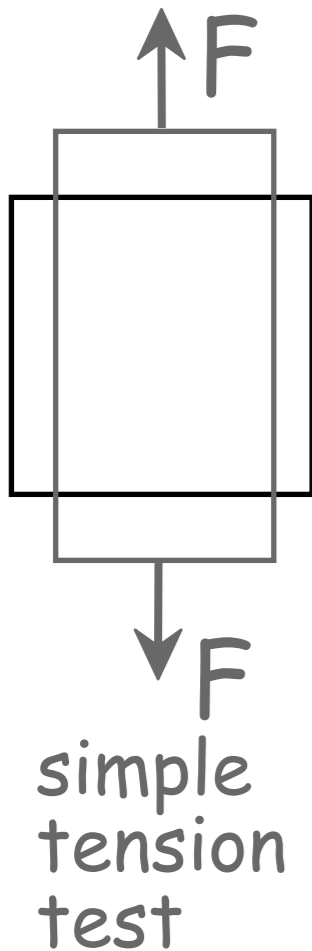
Linear Elastic Properties

Modulus of Elasticity, E :
(also known as Young's modulus)

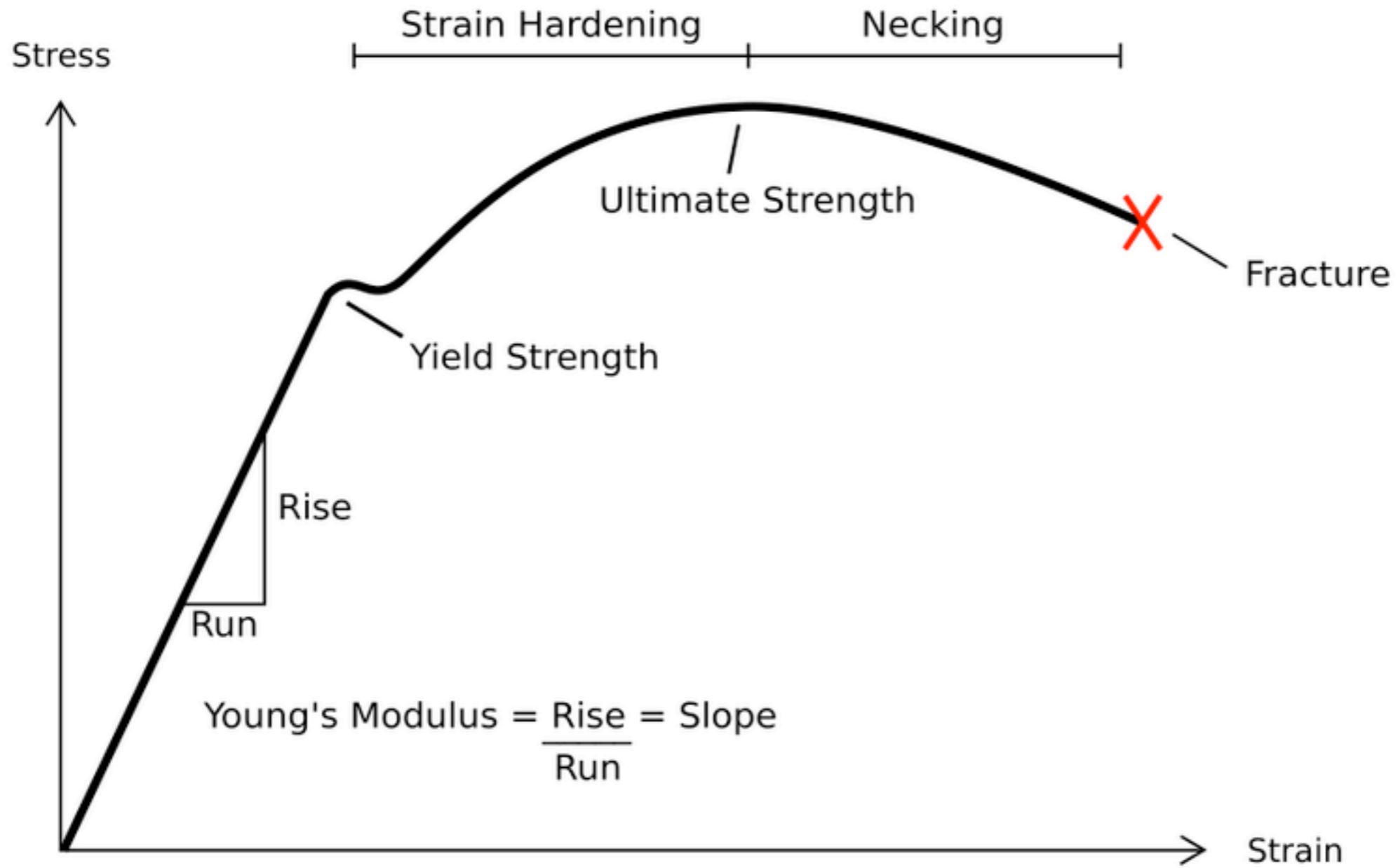
• **Hooke's Law:**

$$\sigma = E \varepsilon$$

E : stiffness (material's resistance to elastic deformation)



Young's modulus





Elasticity Theory



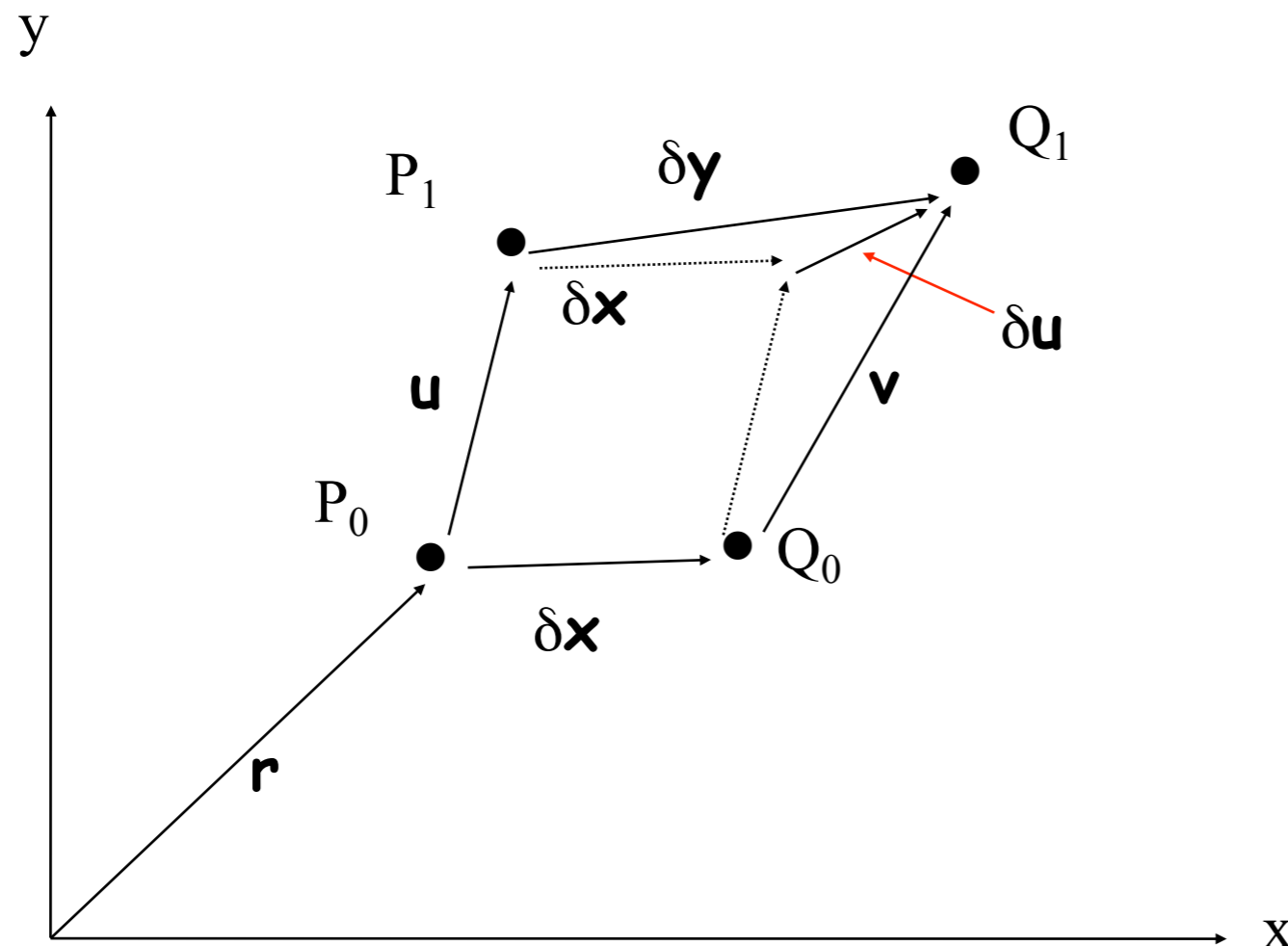
A time-dependent perturbation of an elastic medium (e.g. a rupture, an earthquake, a meteorite impact, a nuclear explosion etc.) generates elastic waves emanating from the source region. These disturbances produce local changes in **stress** and **strain**.

To understand the propagation of elastic waves we need to describe kinematically the **deformation** of our medium and the resulting forces (**stress**). The relation between **deformation** and **stress** is governed by **elastic constants**.

The time-dependence of these disturbances will lead us to the **elastic wave equation** as a consequence of conservation of energy and momentum.

Deformation

Let us consider a point P_0 at position r relative to some fixed origin and a second point Q_0 displaced from P_0 by δx



Unstrained state:

Relative position of point P_0 w.r.t. Q_0 is δx .

Strained state:

Relative position of point P_0 has been displaced a distance u to P_1 and point Q_0 a distance v to Q_1 .

Relative position of point P_1 w.r.t. Q_1 is $\delta \gamma = \delta x + \delta u$. The change in relative position between Q and P is just δu .

Linear Elasticity

The relative displacement in the **unstrained** state is $u(\mathbf{r})$. The relative displacement in the **strained** state is $\mathbf{v} = u(\mathbf{r} + \delta\mathbf{x})$.

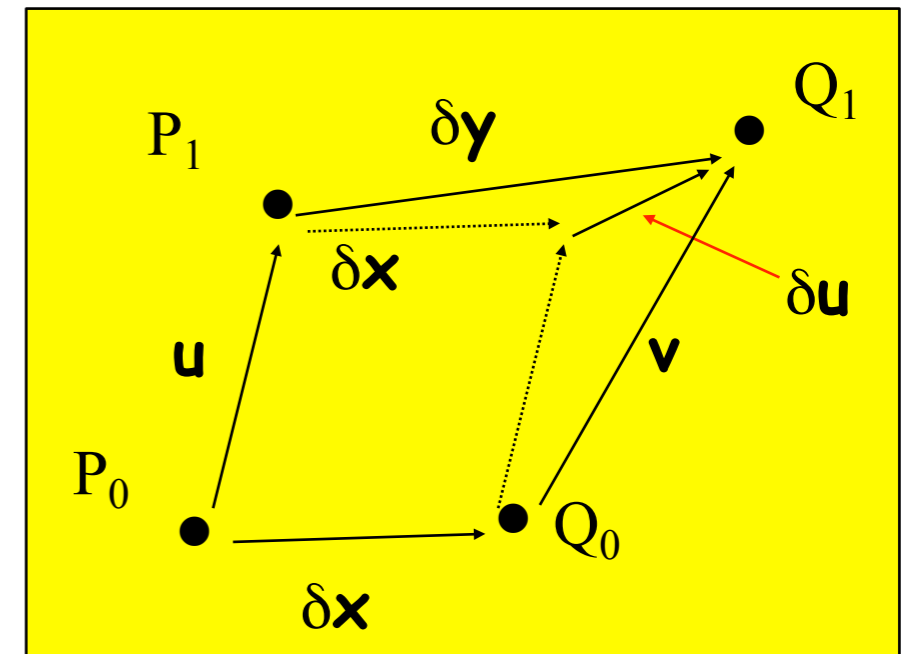
So finally we arrive at expressing the **relative displacement** due to strain:

$$\delta u = u(\mathbf{r} + \delta\mathbf{x}) - u(\mathbf{r})$$

We now apply Taylor's theorem in 3-D to arrive at:

$$\delta u_i = \sum_{k=1,3} \frac{\partial u_i}{\partial x_k} \delta x_k \equiv \frac{\partial u_i}{\partial x_k} \delta x_k$$

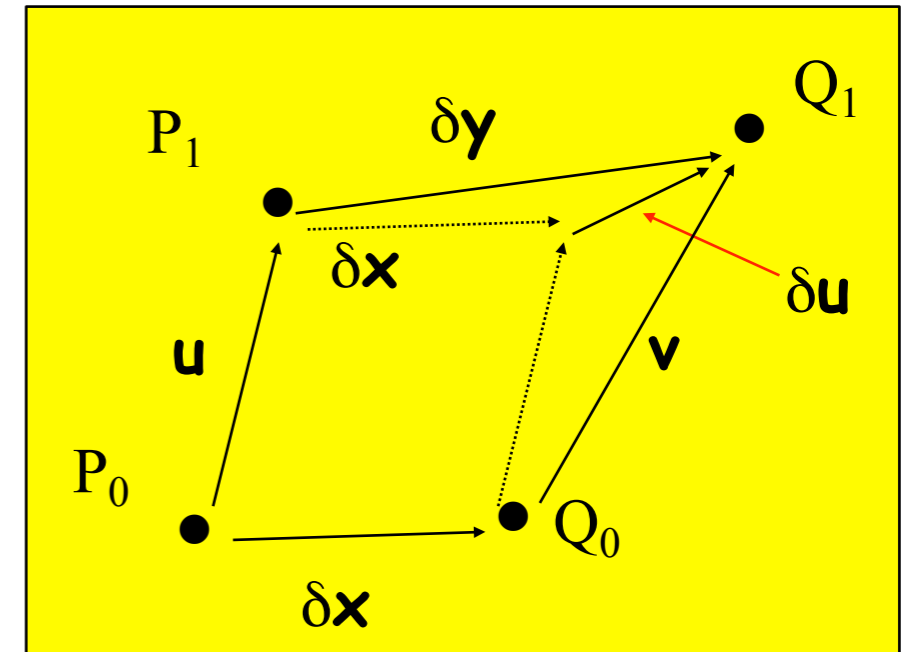
What does this equation mean?



Linear Elasticity – symmetric part

The partial derivatives of the vector components

$$\frac{\partial u_i}{\partial x_k}$$



represent a **second-rank tensor** which can be resolved into a **symmetric** and anti-symmetric part:

$$\delta u_i = \frac{1}{2} \left(\frac{\partial u_i}{\partial x_k} + \frac{\partial u_k}{\partial x_i} \right) \delta x_k - \frac{1}{2} \left(\frac{\partial u_k}{\partial x_i} - \frac{\partial u_i}{\partial x_k} \right) \delta x_k$$

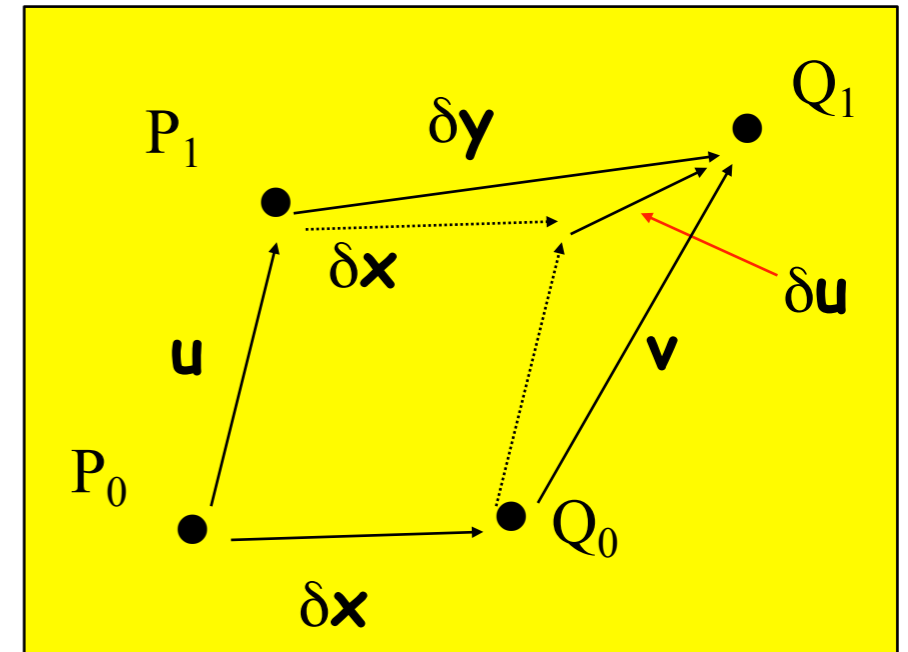
- **symmetric**
- **strain**

- **antisymmetric**
- **pure rotation**

Linear Elasticity – strain tensor

The symmetric part is called the **strain tensor**

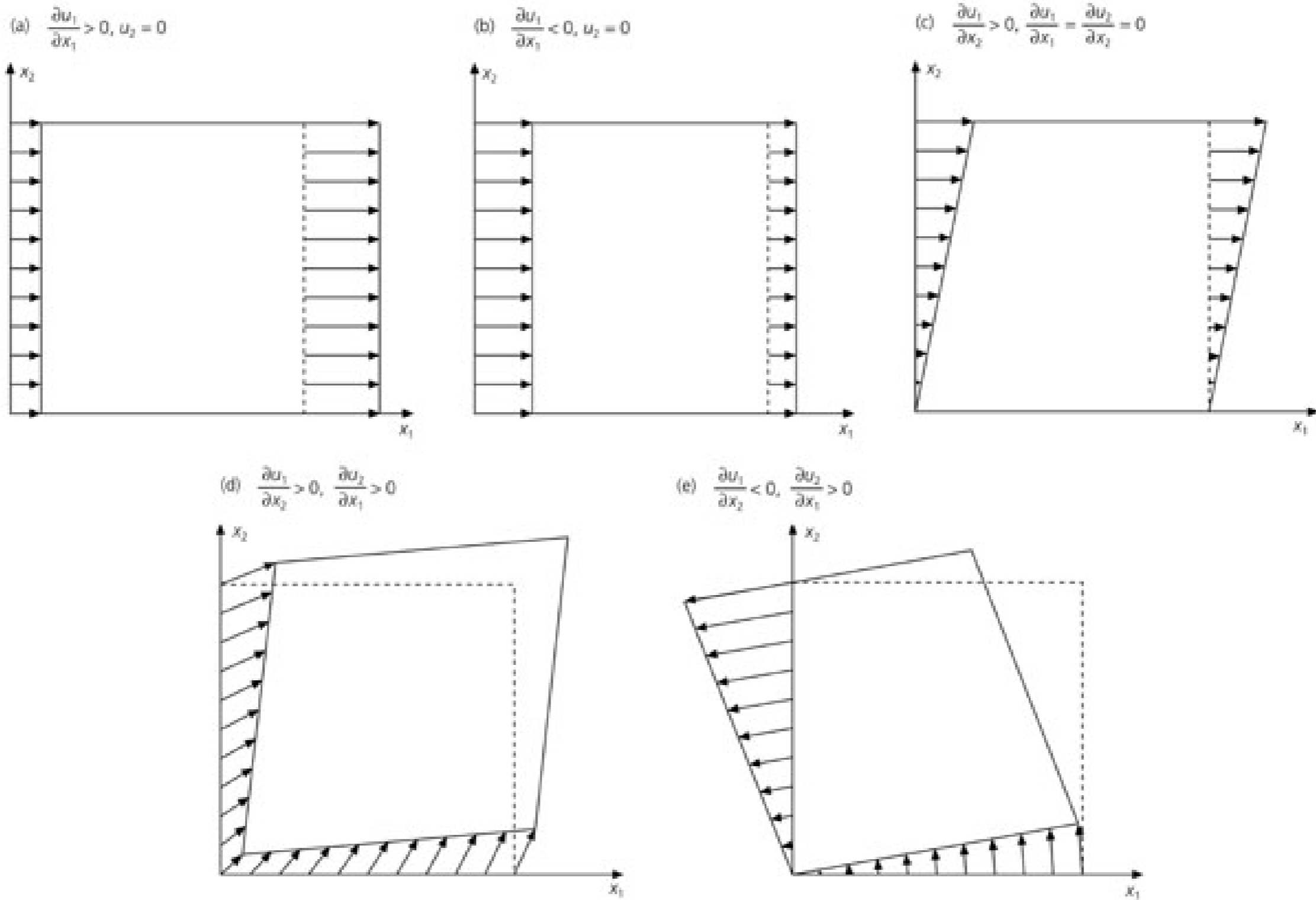
$$\varepsilon_{ij} = \frac{1}{2} \left(\frac{\partial u_i}{\partial x_j} + \frac{\partial u_j}{\partial x_i} \right)$$



and describes the relation between deformation and displacement in linear elasticity. In 2-D this tensor looks like

$$\varepsilon_{ij} = \begin{bmatrix} \frac{\partial u_1}{\partial x} & \frac{1}{2} \left(\frac{\partial u_1}{\partial y} + \frac{\partial u_2}{\partial x} \right) \\ \frac{1}{2} \left(\frac{\partial u_1}{\partial y} + \frac{\partial u_2}{\partial x} \right) & \frac{\partial u_2}{\partial y} \end{bmatrix}$$

Figure 2.3-12: Some possible strains for a two-dimensional element.



Deformation tensor – its elements

Thus

$$u_1 = \lambda_1 y_1$$

$$u_2 = \lambda_2 y_2$$

$$u_3 = \lambda_3 y_3$$

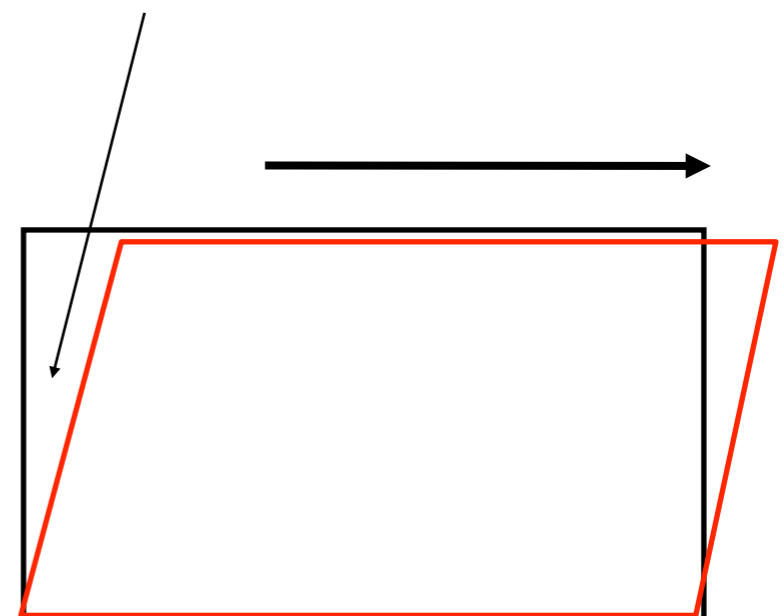
... in other words ...

the eigenvalues are the relative change of length along the three coordinate axes

$$\lambda_1 = \frac{u_1}{y_1}$$

In arbitrary coordinates the **diagonal** elements are the **relative change of length along the coordinate axes** and the **off-diagonal** elements are the **infinitesimal shear angles**.

shear angle



Deformation tensor – trace

The trace of a tensor is defined as the sum over the diagonal elements.

Thus:

$$\varepsilon_{ii} = \varepsilon_{11} + \varepsilon_{22} + \varepsilon_{33}$$

This trace is linked to the volumetric change after deformation. Before deformation the volume was V_0 . Because the diagonal elements are the relative change of lengths along each direction, the new volume after deformation is

$$V = L_1(1 + \varepsilon_{11})L_2(1 + \varepsilon_{22})L_3(1 + \varepsilon_{33})$$

... and neglecting higher-order terms ...

$$V = L_1L_2L_3(1 + \varepsilon_{ii}) \text{ or } V_0(1 + \varepsilon_{ii})$$

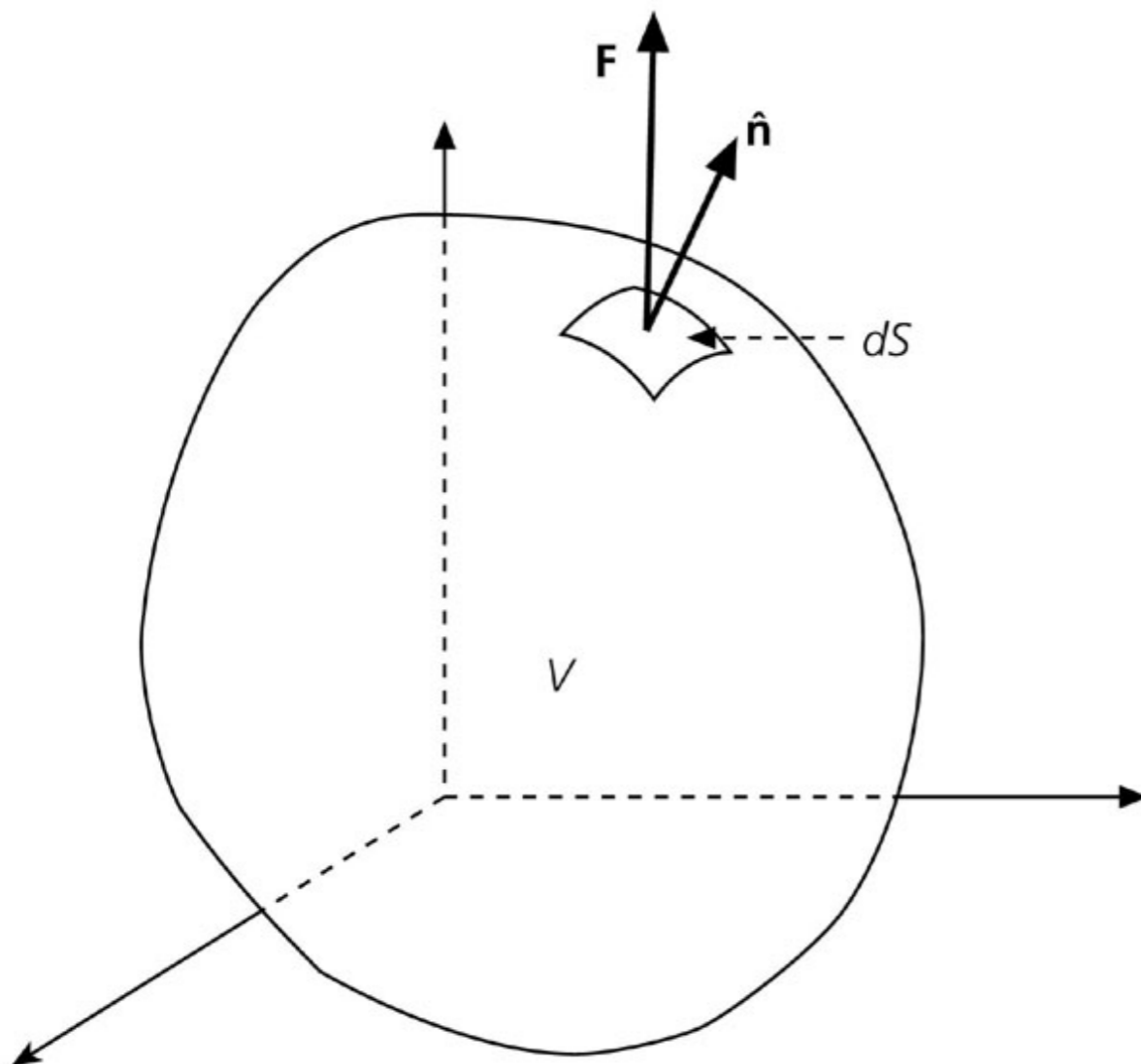
$$\theta = \frac{\Delta V}{V_0} = \varepsilon_{ii} = \frac{\partial u_i}{\partial x_i} = \frac{\partial u_1}{\partial x_1} + \frac{\partial u_2}{\partial x_2} + \frac{\partial u_3}{\partial x_3} = \text{div} \mathbf{u} = \nabla \cdot \mathbf{u}$$

Stress - Traction (vector)

In an elastic body there are restoring forces if deformation takes place. These forces can be seen as acting on planes inside the body. **Forces divided by an areas are called stresses.**

In order for the deformed body to remain deformed these forces have to compensate each other.

Figure 2.3-1: Surface force on a volume element.



Traction vector cannot be completely described without the specification of the force ($\Delta\mathbf{F}$) and the surface ($\Delta\mathbf{S}$) on which it acts:

$$\mathbf{T}(\mathbf{n}) = \lim_{\Delta S \rightarrow 0} \frac{\Delta\mathbf{F}}{\Delta S} = \frac{d\mathbf{F}}{dS}$$

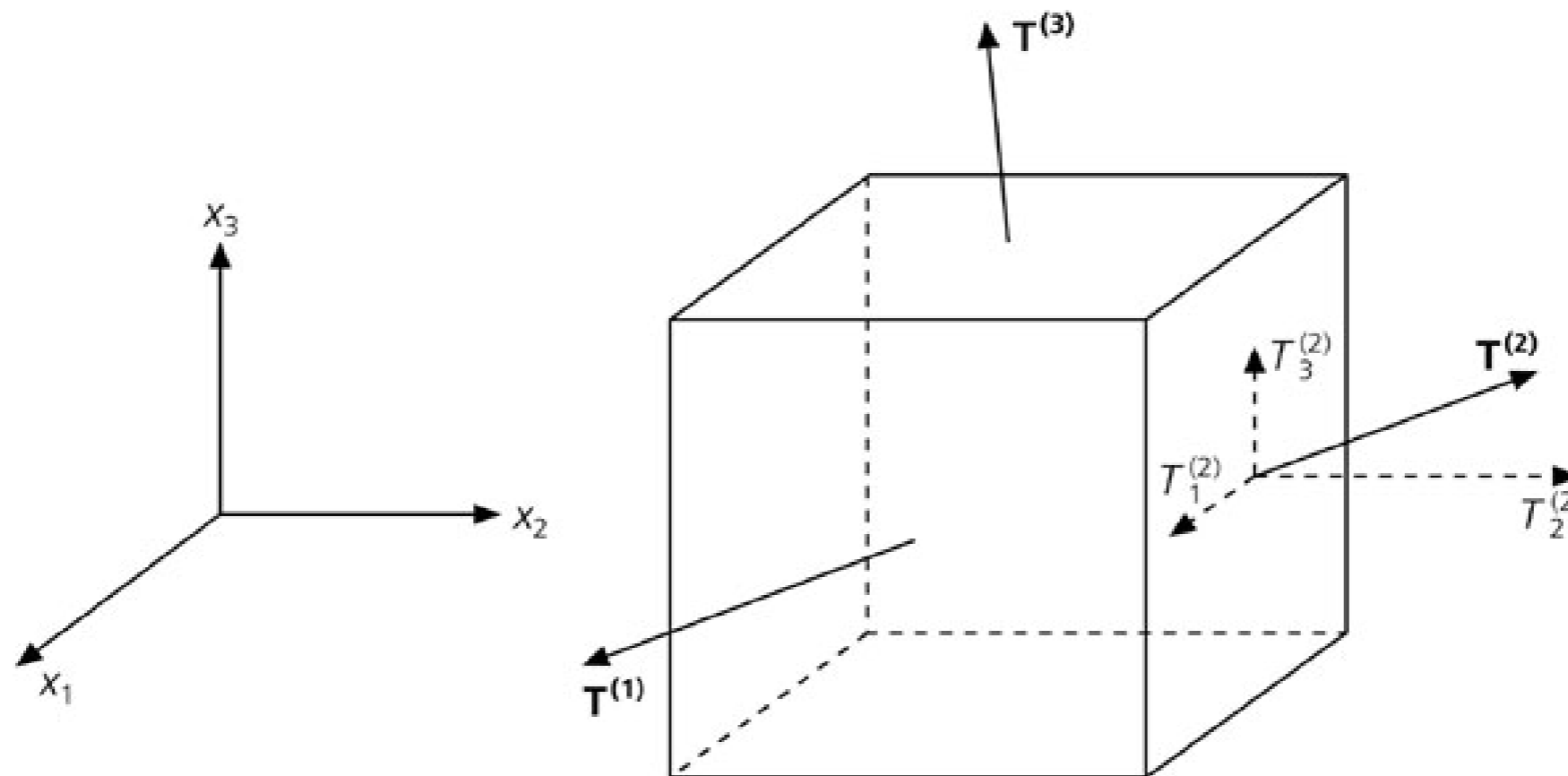
And from the linear momentum conservation, we can show that:

$$\mathbf{T}(-\mathbf{n}) = -\mathbf{T}(\mathbf{n})$$

Stress-Traction (cont.)

Stress acting on a given internal plane can be decomposed in 3 mutually orthogonal components: one normal (direct stress), tending to change the volume of the material, and two tangential (shear stress), tending to deform, to the surface. If we consider an infinitely small cube, aligned with a Cartesian reference system:

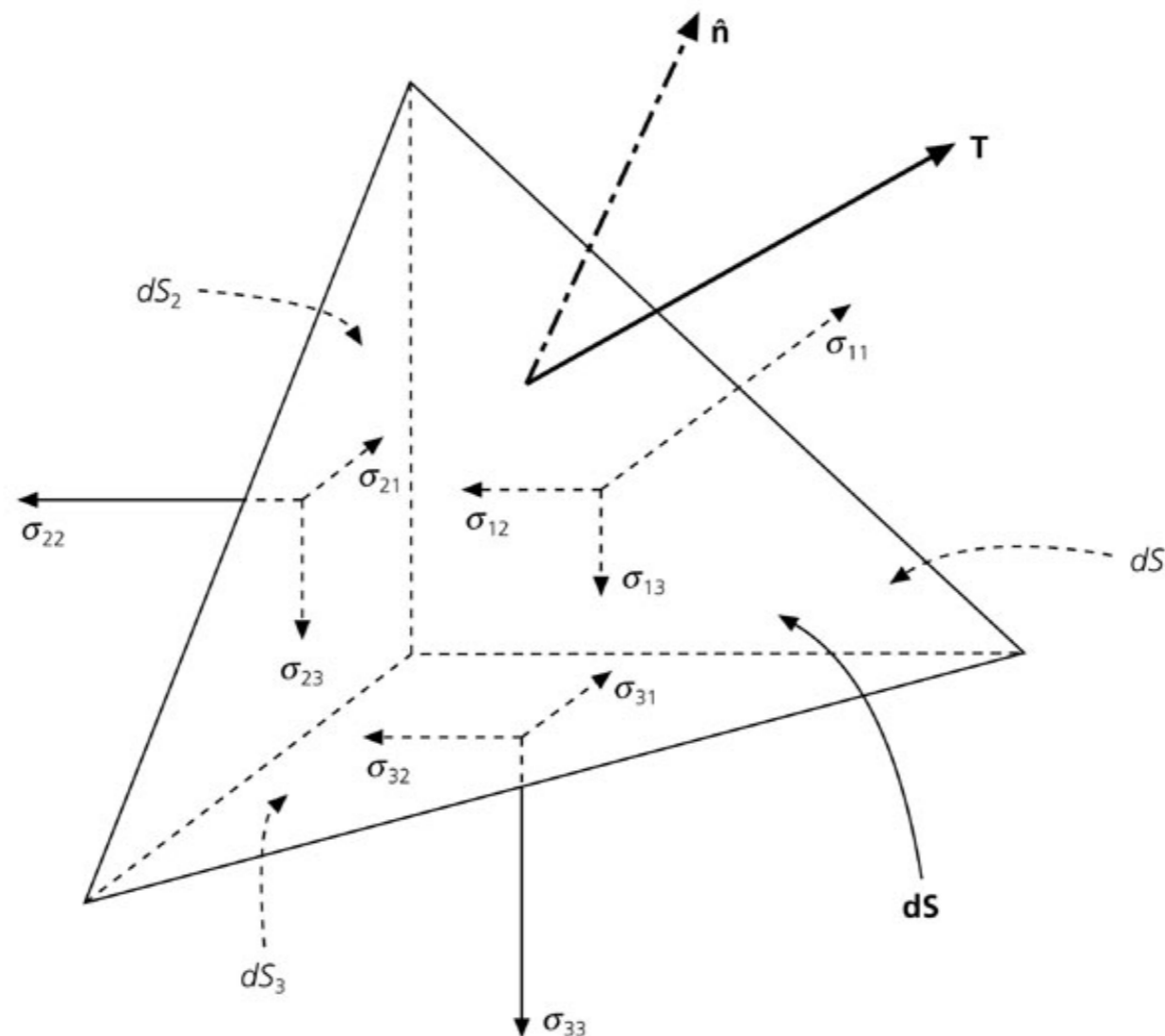
Figure 2.3-2: Traction vectors on the faces of a volume element.



$$\mathbf{T}^{(n)} = n_i \mathbf{T}^{(i)} = n_i T_j^{(i)} \mathbf{e}_j = n_i \sigma_{ij} \mathbf{e}_j$$

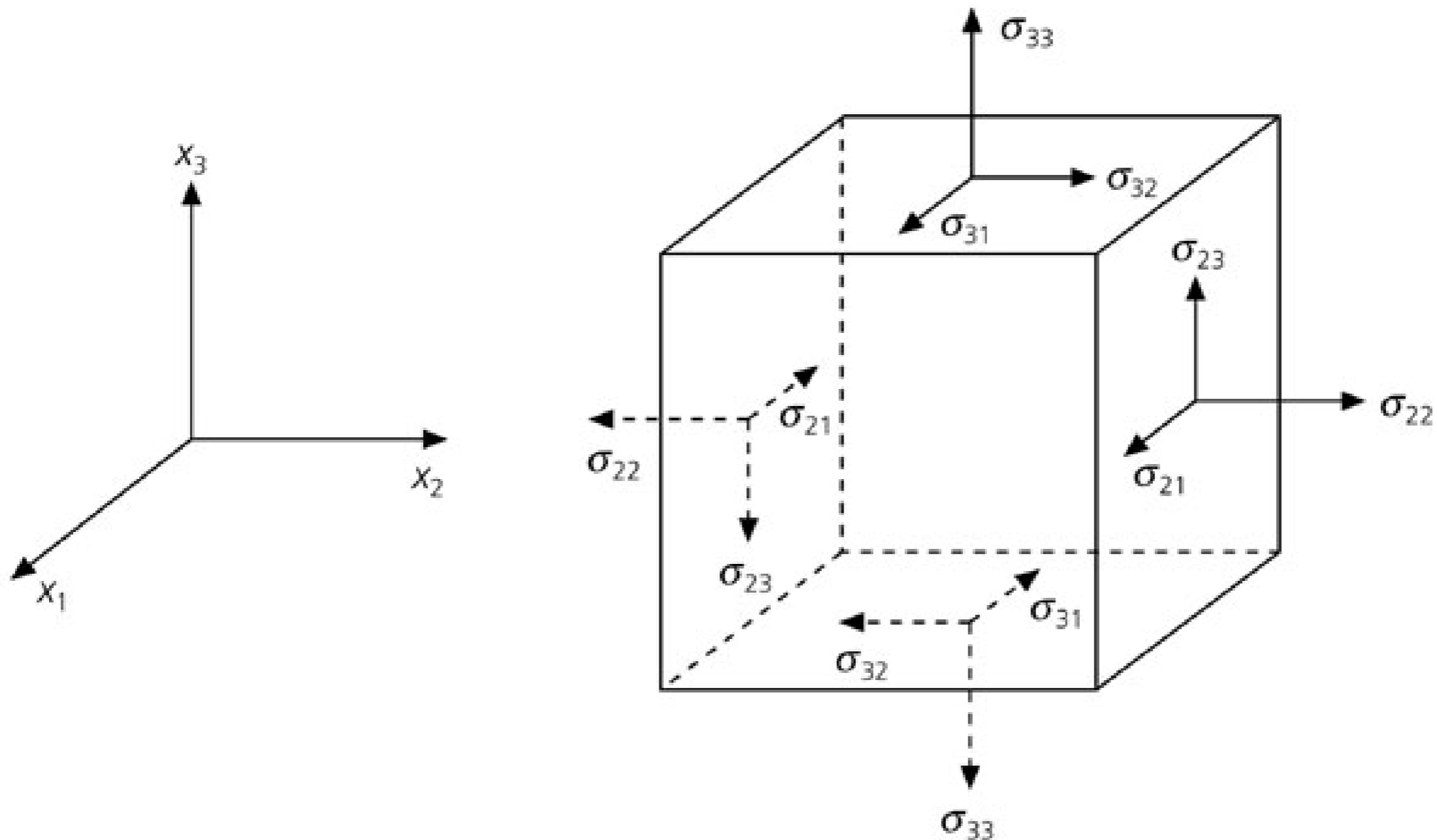
Consider an infinitively small tetraedrum, whose 3 faces are oriented normally to the reference axes. The components of traction \mathbf{T} , acting on the face whose normal is \mathbf{n} can be written using the directional cosines referred to versor system $\hat{\mathbf{e}}$

$$\mathbf{T}^{(n)} = n_i \mathbf{T}^{(i)} = n_i T_j^{(i)} \mathbf{e}_j = n_i \sigma_{ij} \mathbf{e}_j$$



...and the stress state in a point of the material can be expressed with:

Figure 2.3-4: Stress components on the faces of a volume element.



Stress-strain relation

The relation between stress and strain in general is described by the tensor of elastic constants c_{ijkl}

$$\sigma_{ij} = c_{ijkl} \varepsilon_{kl}$$

Generalised Hooke's Law

From the symmetry of the stress and strain tensor and a thermodynamic condition it follows that the maximum number of independent constants of c_{ijkl} is 21. In an isotropic body, where the properties do not depend on direction, the relation reduces to

$$\sigma_{ij} = \lambda \theta \delta_{ij} + 2\mu \varepsilon_{ij}$$

Hooke's Law

where λ and μ are the Lamé parameters, θ is the dilatation and δ_{ij} is the Kronecker delta.

$$\theta \delta_{ij} = \varepsilon_{kk} \delta_{ij} = \left(\varepsilon_{11} + \varepsilon_{22} + \varepsilon_{33} \right) \delta_{ij}$$

Stress-strain relation - 2

The complete stress tensor looks like

$$\sigma_{ij} = \begin{pmatrix} (\lambda + 2\mu)\varepsilon_{11} + \lambda(\varepsilon_{22} + \varepsilon_{33}) & 2\mu\varepsilon_{12} & 2\mu\varepsilon_{13} \\ 2\mu\varepsilon_{21} & (\lambda + 2\mu)\varepsilon_{22} + \lambda(\varepsilon_{11} + \varepsilon_{33}) & 2\mu\varepsilon_{23} \\ 2\mu\varepsilon_{31} & 2\mu\varepsilon_{32} & (\lambda + 2\mu)\varepsilon_{33} + \lambda(\varepsilon_{11} + \varepsilon_{22}) \end{pmatrix}$$

Mean stress (invariant respect to the coordinate system) $M = \frac{(\sigma_{11} + \sigma_{22} + \sigma_{33})}{3} = \frac{\sum_{n=1}^3 \lambda_n}{3}$

Deviatoric stress:

$$D_{ij} = \sigma_{ij} - M\delta_{ij}$$

In the Earth the mean stress is essentially due to **lithostatic load**:

$$P = -\int_0^z \rho(z) dz$$

Elastic parameters

Rigidity is the ratio of pure shear strain and the applied shear stress component

$$\mu = \frac{\sigma_{ij}}{2\varepsilon_{ij}}$$

Bulk modulus of incompressibility is defined the ratio of pressure to volume change. Ideal fluid means no rigidity ($\mu = 0$), thus λ is the incompressibility of a fluid.

$$K = -\frac{P}{\theta} = \lambda + \frac{2}{3}\mu$$

Consider a stretching experiment where tension is applied to an isotropic medium along a principal axis (say x).

$$\text{Poisson's ratio} \equiv \nu = -\frac{\varepsilon_{22}}{\varepsilon_{11}} = \frac{\lambda}{2(\lambda + 2\mu)} \quad \text{Young's modulus} \equiv E = -\frac{\sigma_{11}}{\varepsilon_{11}} = \frac{\mu(3\lambda + 2\mu)}{\lambda + \mu}$$

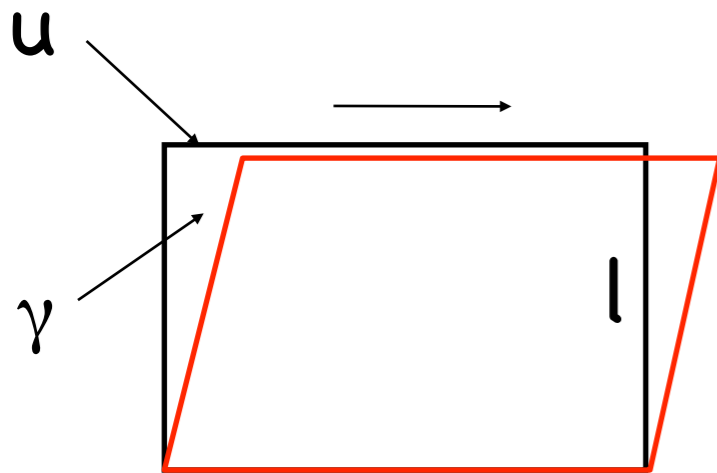
$$\lambda = \frac{\nu E}{(1 + \nu)(1 - 2\nu)} \quad \mu = \frac{E}{2(1 + \nu)}$$

For Poisson's ratio we have $0 < \nu < 0.5$.

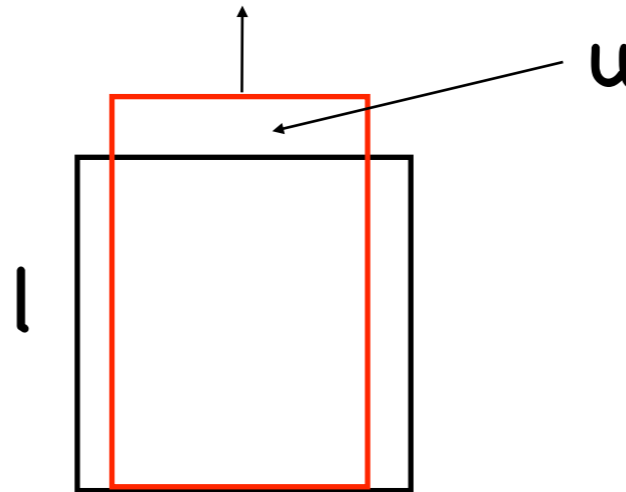
A useful approximation is $\lambda = \mu$ (Poisson's solid), then $\nu = 0.25$ and for fluids $\nu = 0.5$

Stress-strain - significance

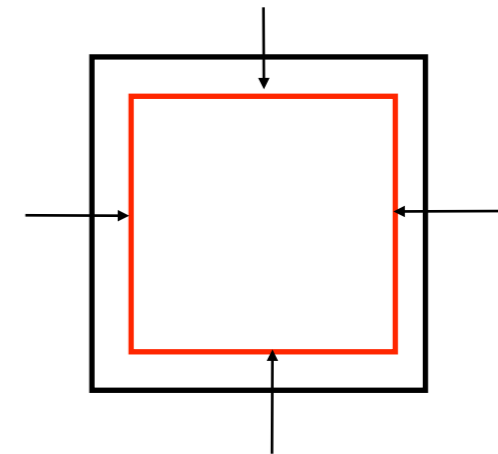
As in the case of deformation the stress-strain relation can be interpreted in simple geometric terms:



$$\sigma_{12} = \mu\gamma = \mu\varepsilon_{12}$$



$$\sigma_{22} = E \frac{u}{l} = K\varepsilon_{ii}$$



$$p = K \frac{\Delta V}{V} = K\varepsilon_{ii}$$

Remember that these relations are a generalization of Hooke's Law:

$$F = Kx$$

Elastic constants

Let us look at some examples for elastic constants:

Rock	K 10^{12} dynes/cm ²	E 10^{12} dynes/cm ²	μ 10^{12} dynes/cm ²	ν
Limestone		0.621	0.248	0.251
Granite	0.132	0.416	0.197	0.055
Gabbro	0.659	1.08	0.438	0.219
Dunite		1.52	0.6	0.27

Equations of elastic motion

We now have a complete description of the forces acting within an elastic body. Adding the inertia forces with opposite sign leads us from

$$f_i + \frac{\partial \sigma_{ij}}{\partial x_j} = 0$$

to

$$\rho \frac{\partial^2 u_i}{\partial t^2} = f_i + \frac{\partial \sigma_{ij}}{\partial x_j}$$

the equations of motion for dynamic elasticity

Equations of motion – P waves

$$\rho \partial_t^2 \mathbf{u} = \mathbf{f} + (\lambda + 2\mu) \nabla \nabla \cdot \mathbf{u} - \mu \nabla \times \nabla \times \mathbf{u}$$

Let us apply the **div** operator to this equation, we obtain

$$\rho \partial_t^2 \theta = (\lambda + 2\mu) \Delta \theta$$

where

$$\theta = \nabla \cdot \mathbf{u}$$

Acoustic wave
equation

or

P-wave velocity

$$\frac{1}{\alpha^2} \partial_t^2 \theta = \Delta \theta$$

$$\alpha = \sqrt{\frac{\lambda + 2\mu}{\rho}}$$

Equations of motion – S waves

$$\rho \partial_{\dagger}^2 \mathbf{u} = \mathbf{f} + (\lambda + 2\mu) \nabla \nabla \cdot \mathbf{u} - \mu \nabla \times \nabla \times \mathbf{u}$$

Let us apply the **curl** operator to this equation, we obtain

$$\rho \partial_{\dagger}^2 \nabla \times \mathbf{u} = (\lambda + \mu) \nabla \times \nabla \theta + \mu \Delta (\nabla \times \mathbf{u})$$

we now make use of $\nabla \times \nabla \theta = 0$ and define

$$\boldsymbol{\varphi} = \nabla \times \mathbf{u} \quad \text{to obtain}$$

Shear wave
equation

$$\frac{1}{\beta^2} \partial_{\dagger}^2 \boldsymbol{\varphi} = \Delta \boldsymbol{\varphi}$$

S-wave velocity

$$\beta = \sqrt{\frac{\mu}{\rho}}$$

Plane waves

... what can we say about the direction of displacement, the **polarization** of seismic waves?

$$\mathbf{u} = \nabla\Phi + \nabla \times \Psi \quad \Rightarrow \quad \mathbf{u} = \mathbf{P} + \mathbf{S}$$

$$\mathbf{P} = \nabla\Phi \quad \mathbf{S} = \nabla \times \Psi$$

... we now assume that the potentials have the well known form of plane harmonic waves

$$\Phi = A \exp[i(\mathbf{k} \cdot \mathbf{x} - \omega t)]$$



$$\mathbf{P} = A \mathbf{k} \exp[i(\mathbf{k} \cdot \mathbf{x} - \omega t)]$$

P waves are **longitudinal** as P is parallel to k

$$\Psi = B \exp[i(\mathbf{k} \cdot \mathbf{x} - \omega t)]$$

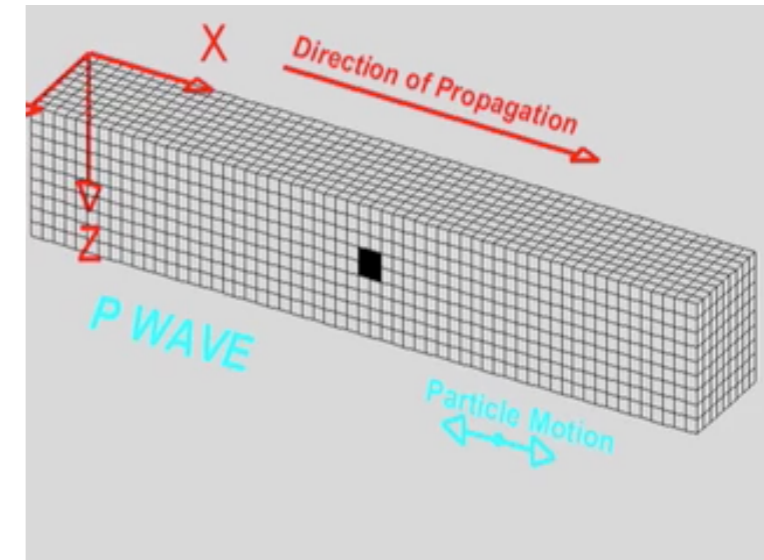
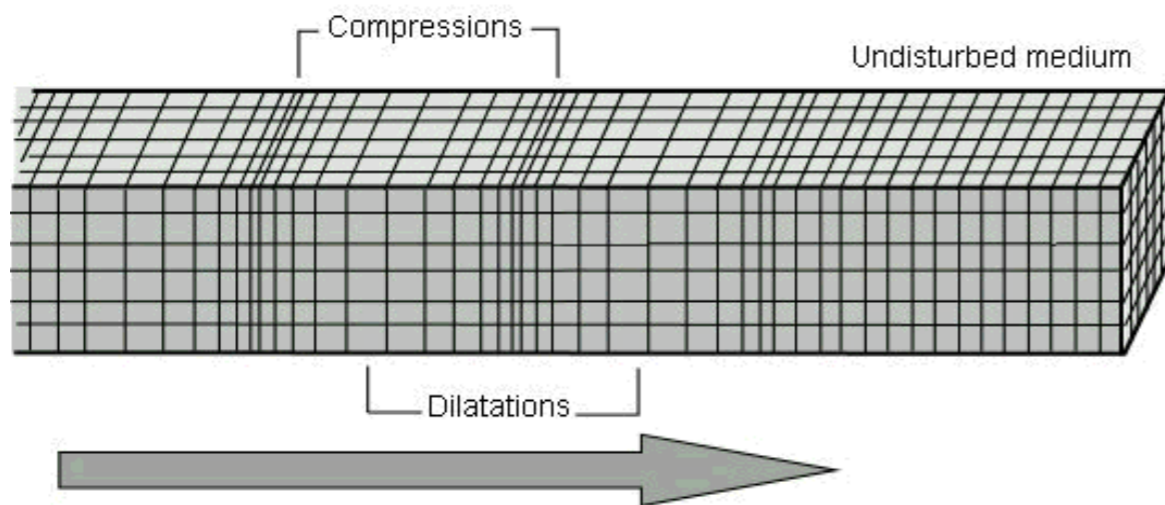


$$\mathbf{S} = \mathbf{k} \times B \exp[i(\mathbf{k} \cdot \mathbf{x} - \omega t)]$$

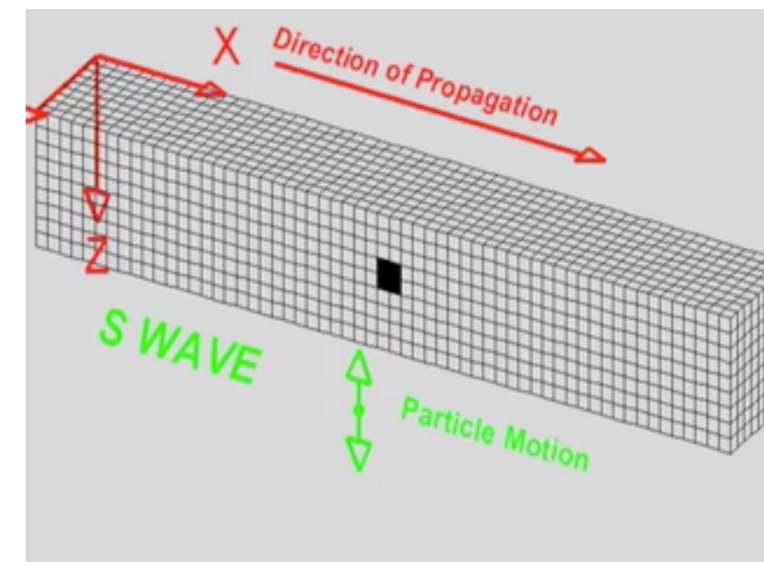
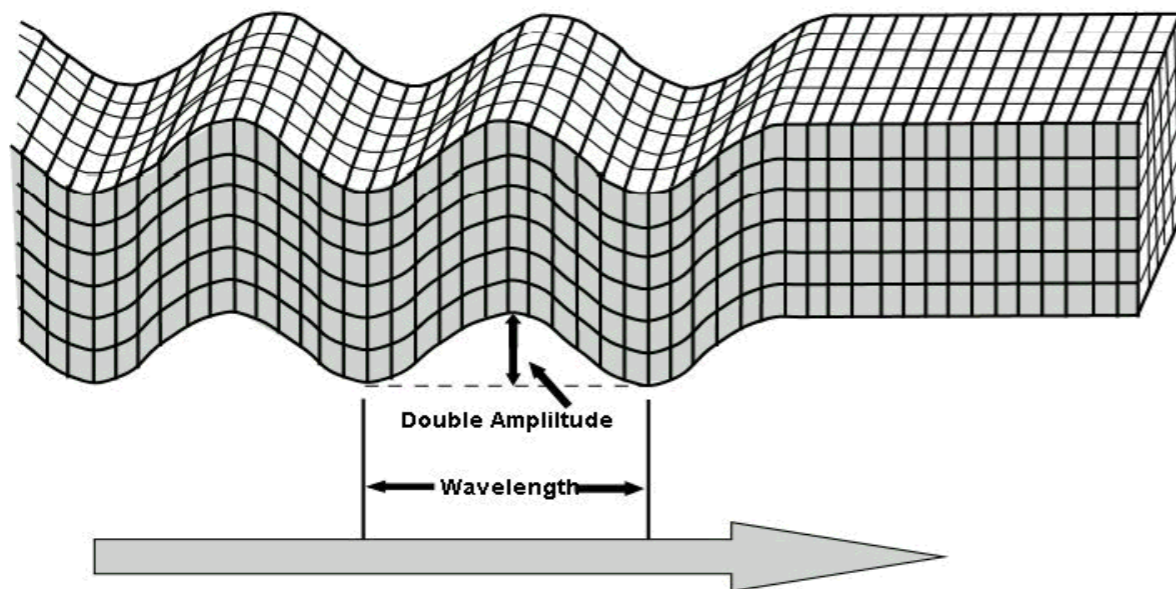
S waves are **transverse** because S is normal to the wave vector k

Wavefields visualization - body waves

P Wave



S Wave



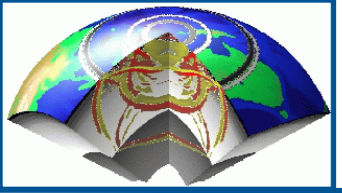
They are **spherical** waves
and decay as $(r)^{-1}$

Seismic Velocities

Material	P-wave velocity (m/s)	shear wave velocity (m/s)
Water	1500	0
Loose sand	1800	500
Clay	1100-2500	
Sandstone	1400-4300	
Anhydrite, Gulf Coast	4100	
Conglomerate	2400	
Limestone	6030	3030
Granite	5640	2870
Granodiorite	4780	3100
Diorite	5780	3060
Basalt	6400	3200
Dunite	8000	4370
Gabbro	6450	3420

Seismic Velocities

Material	V_p (km/s)
Unconsolidated material	
Sand (dry)	0.2-1.0
Sand (wet)	1.5-2.0
Sediments	
Sandstones	2.0-6.0
Limestones	2.0-6.0
Igneous rocks	
Granite	5.5-6.0
Gabbro	6.5-8.5
Pore fluids	
Air	0.3
Water	1.4-1.5
Oil	1.3-1.4
Other material	
Steel	6.1
Concrete	3.6



Ray paths inside the Earth

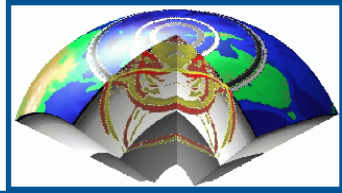
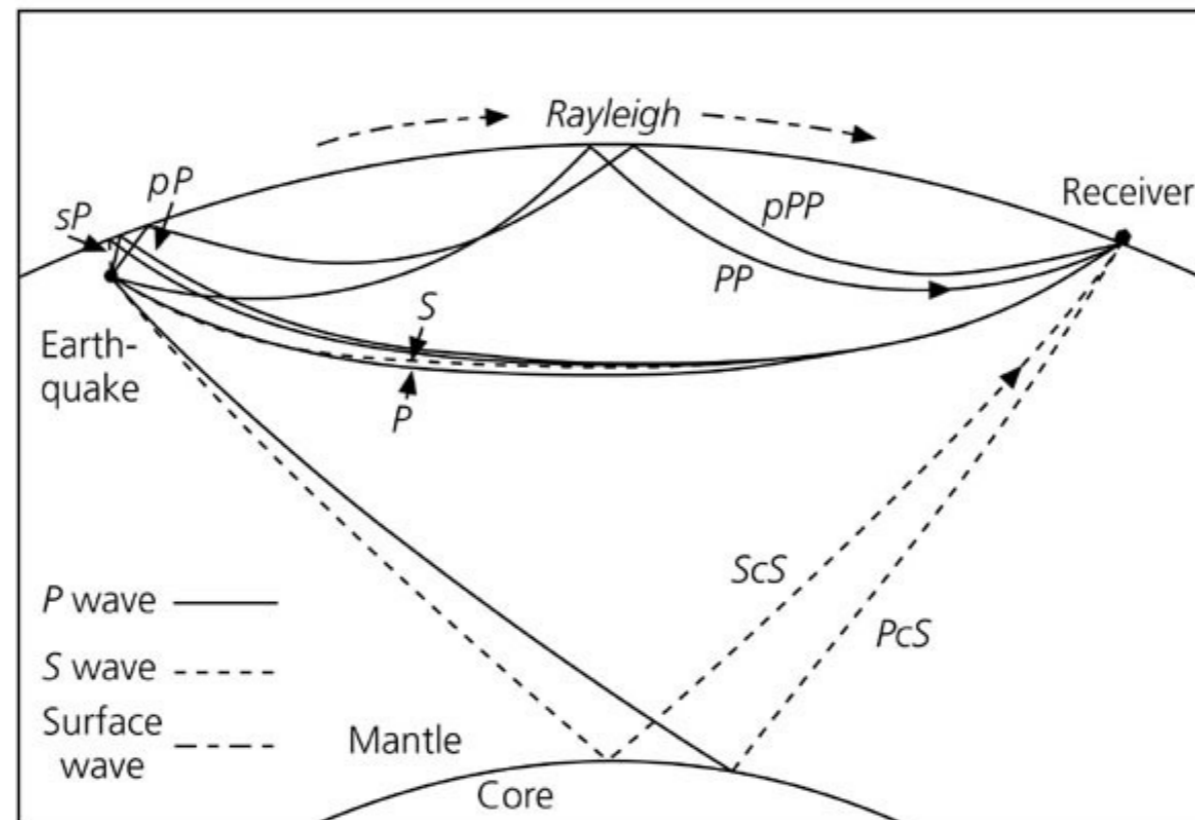
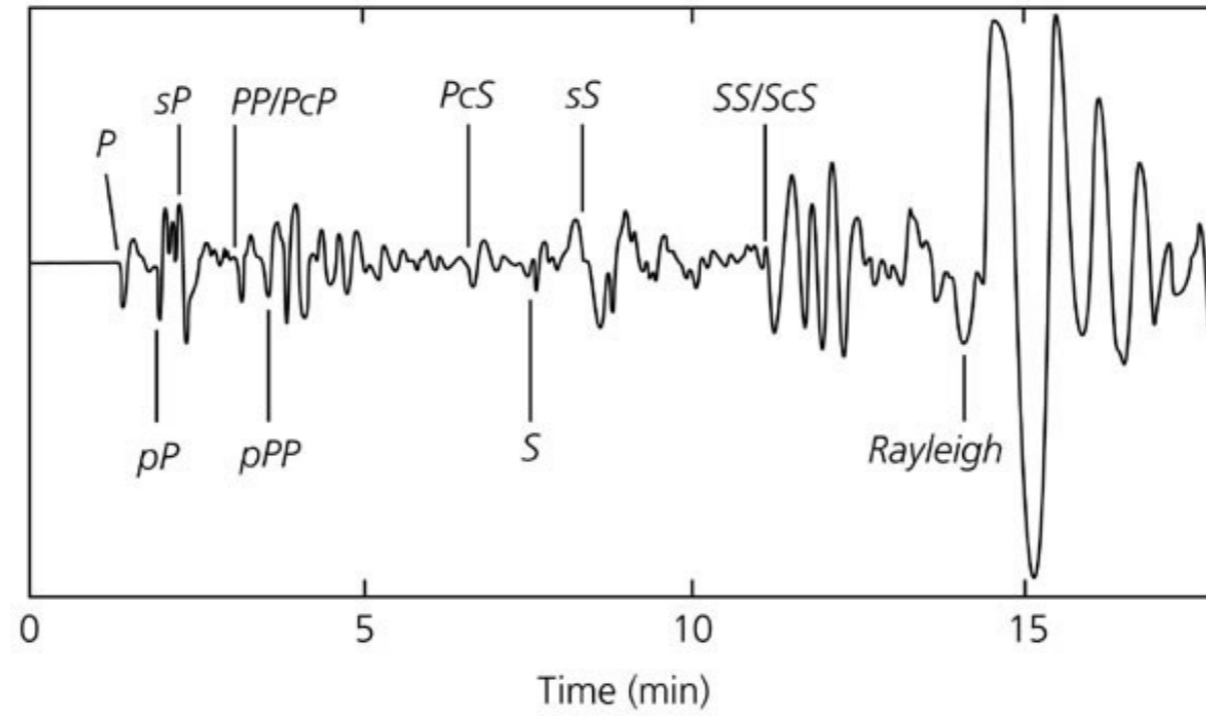
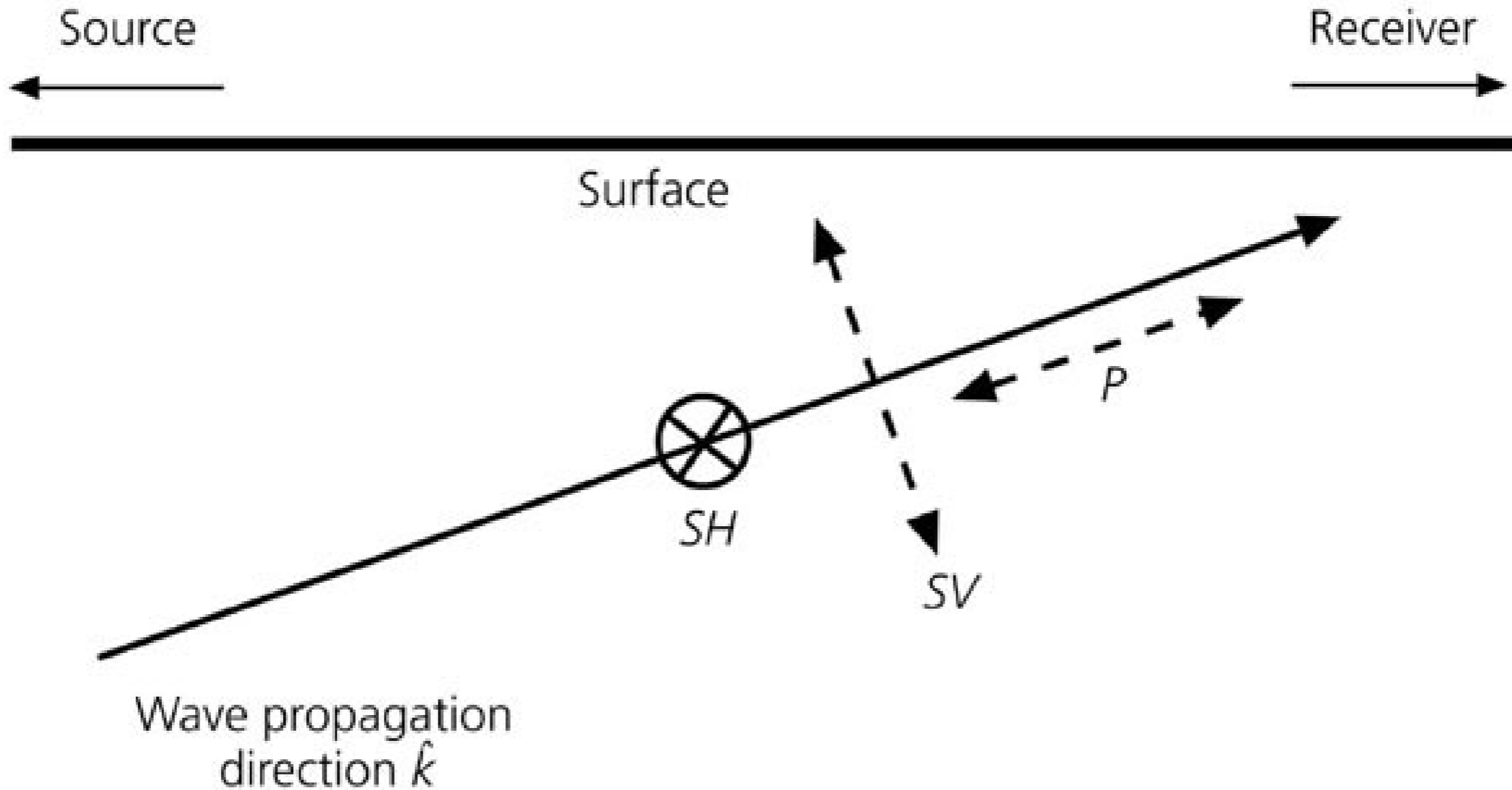


Figure 3.5-2: Selection of body phases and their ray paths.



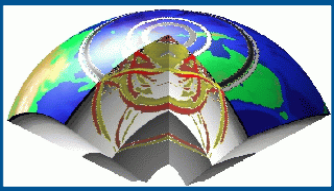
Free surface: P-SV-SH

Figure 2.4-4: Displacements for P, SV, and SH.





The Inverse Problem



It seems that now we have the means to predict arrival times T_{pre} at a given the travel distance of a ray with a given emergence angle (ray parameter) and given structure. This is also termed a **forward (or direct) problem**.

We have recorded a set of travel times, T_{obs} , and we want to determine the structure of the Earth. Thus, what we really want is to solve the **inverse problem**.

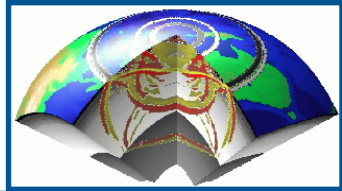
In a very general sense we are looking for an Earth model that **minimizes** the difference between a theoretical prediction and the observed data:

$$\sum \left(T_{\text{obs}} - T_{\text{pre}}(m) \right)$$

where m is an Earth model.

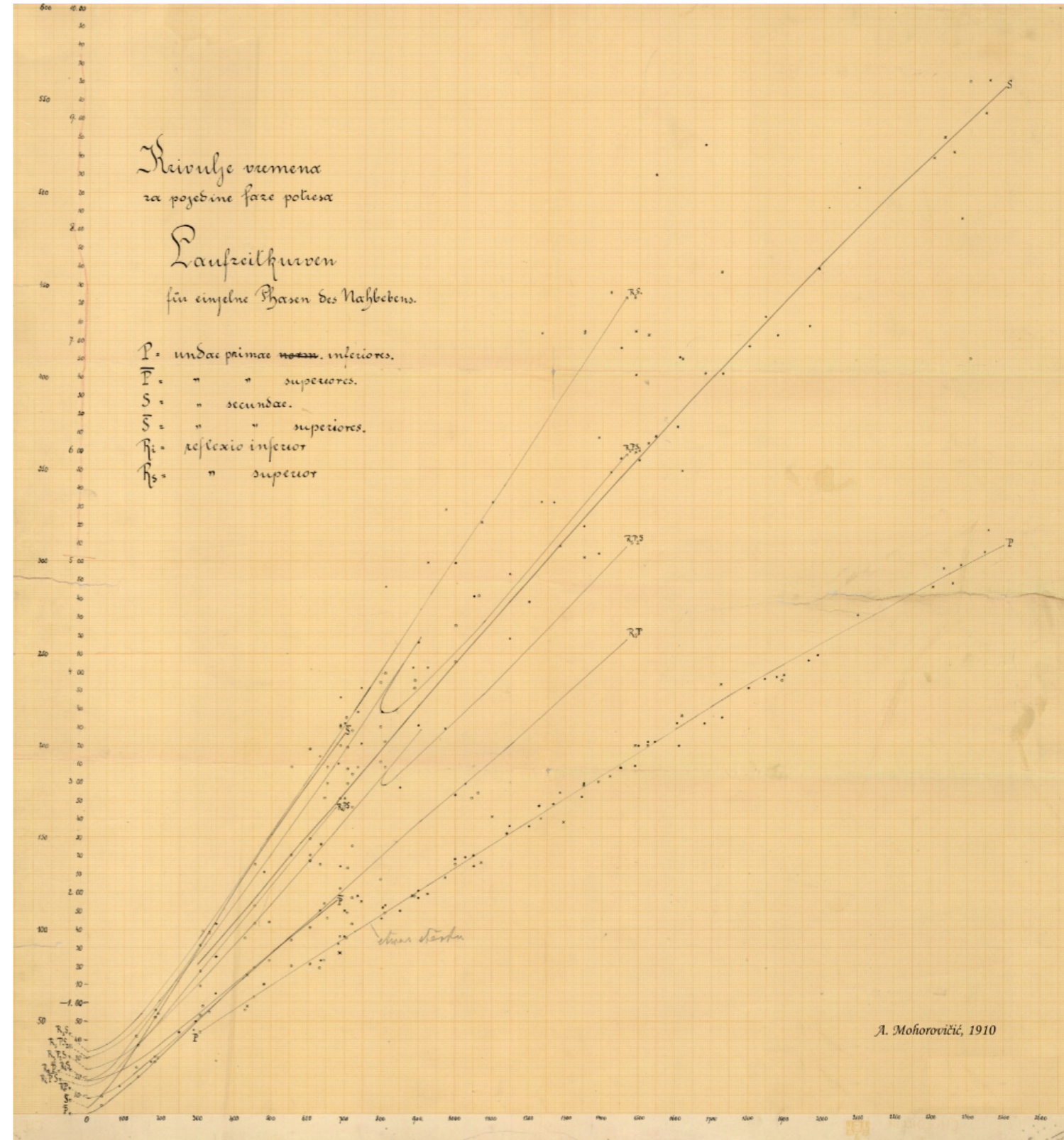
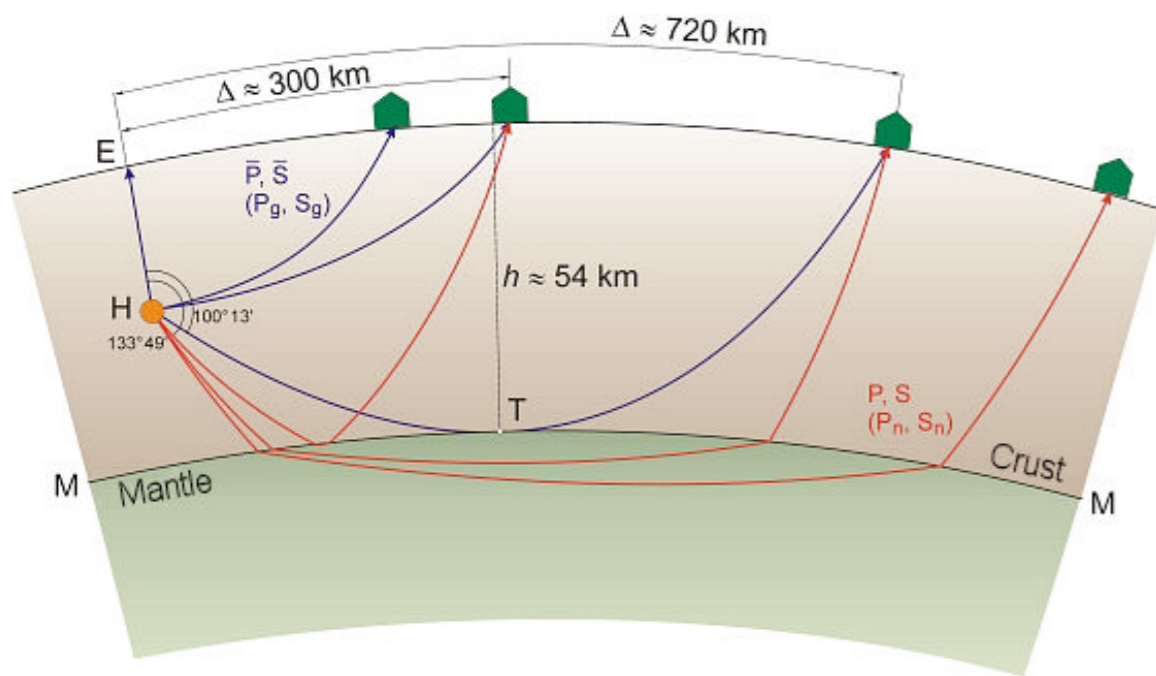


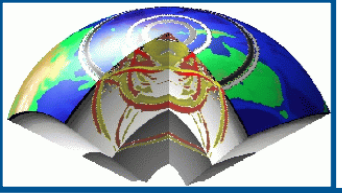
Earth's crust



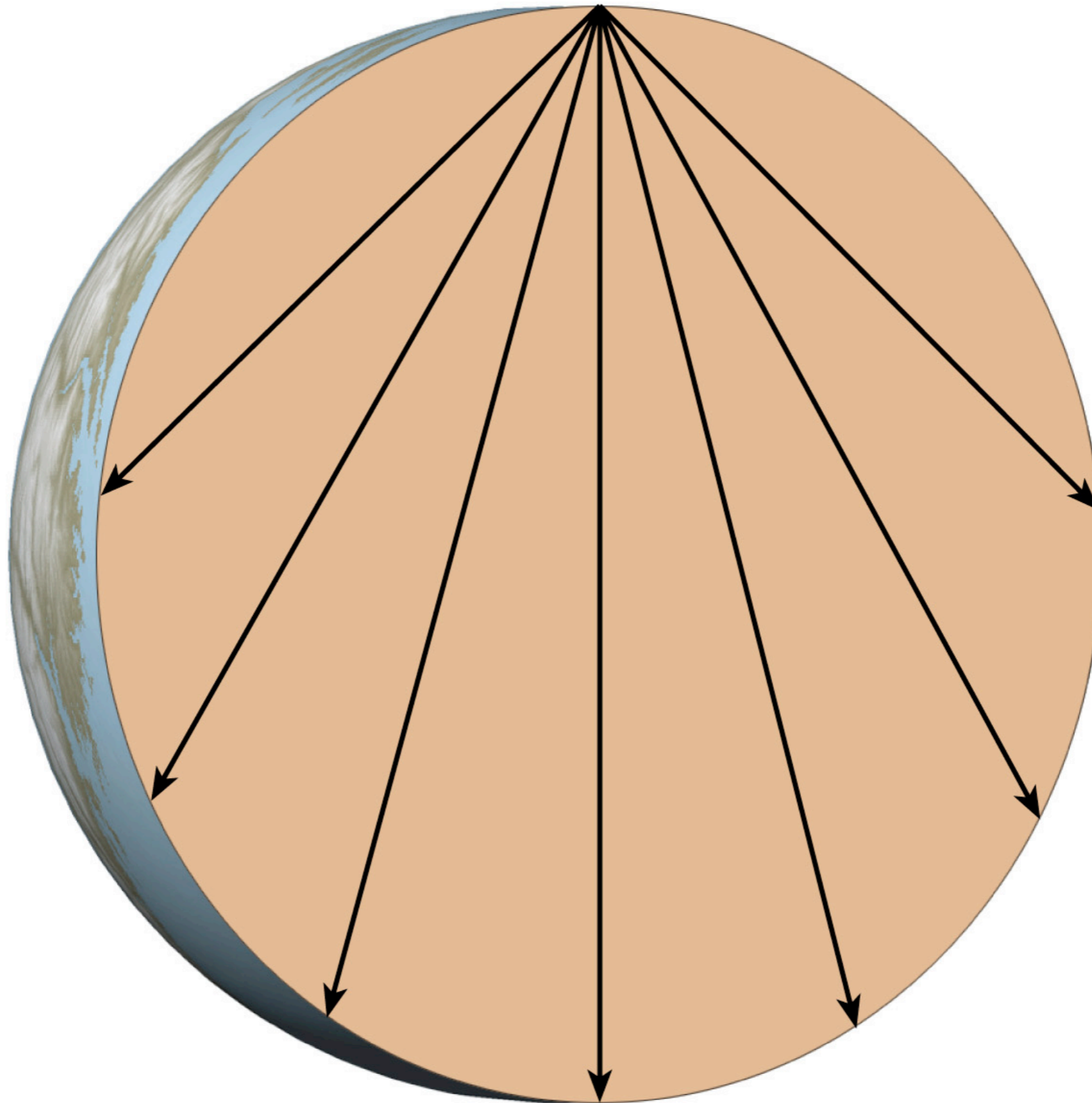
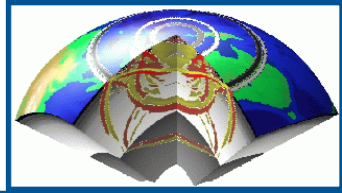
Andrija MOHOROVIČIĆ

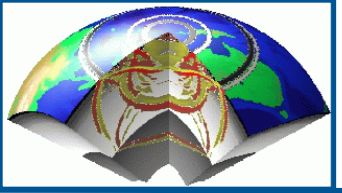
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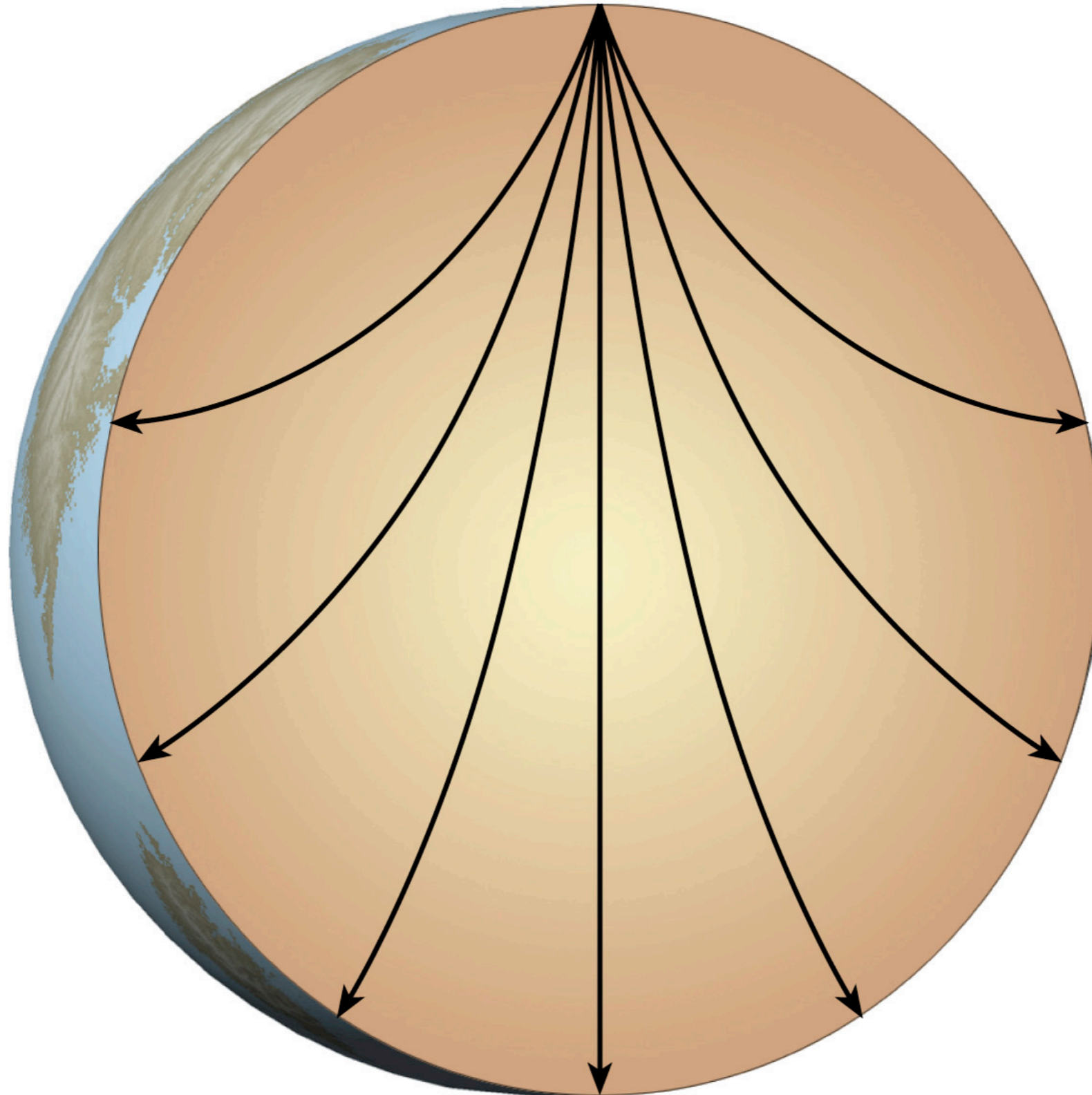
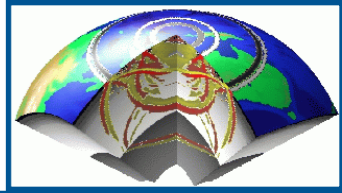


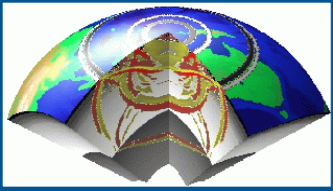
Rays in homogeneous sphere



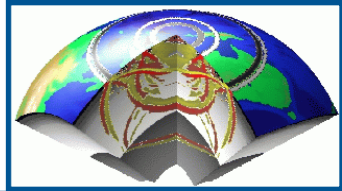


Sphere with increasing velocity...



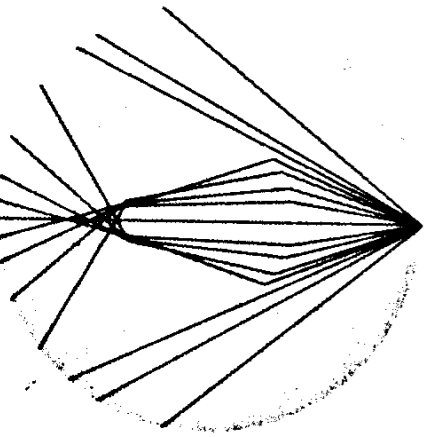


Earth's core

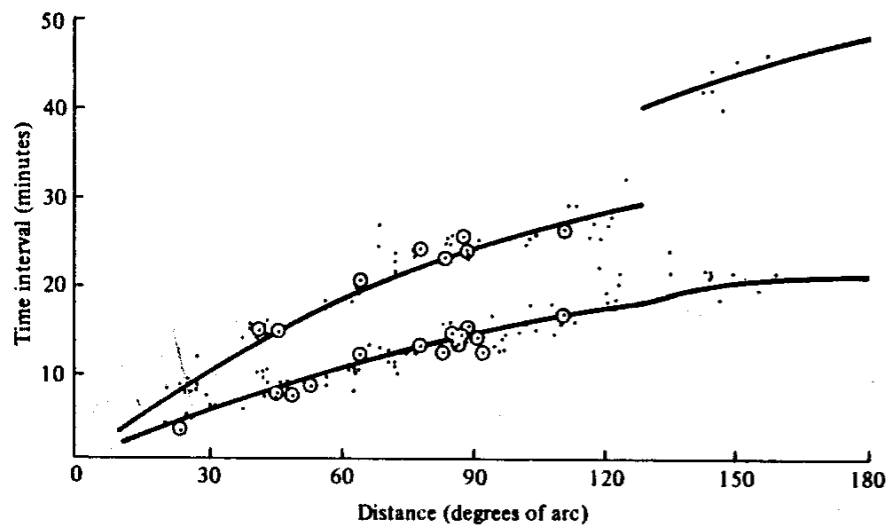


Richard Dixon Oldham

The Constitution of the Earth as revealed by earthquakes, Quart. J. Geological Soc. Lond., 62, 456-475, 1906



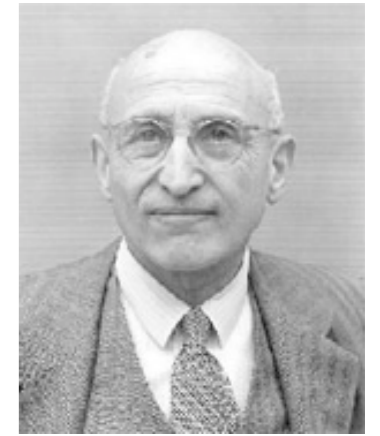
Paths of seismic waves through the Earth assuming a core of radius $0.4R$, in which the speed is 3 km/sec, while the speed outside it is 6 km/sec. [From Oldham, 1906.]



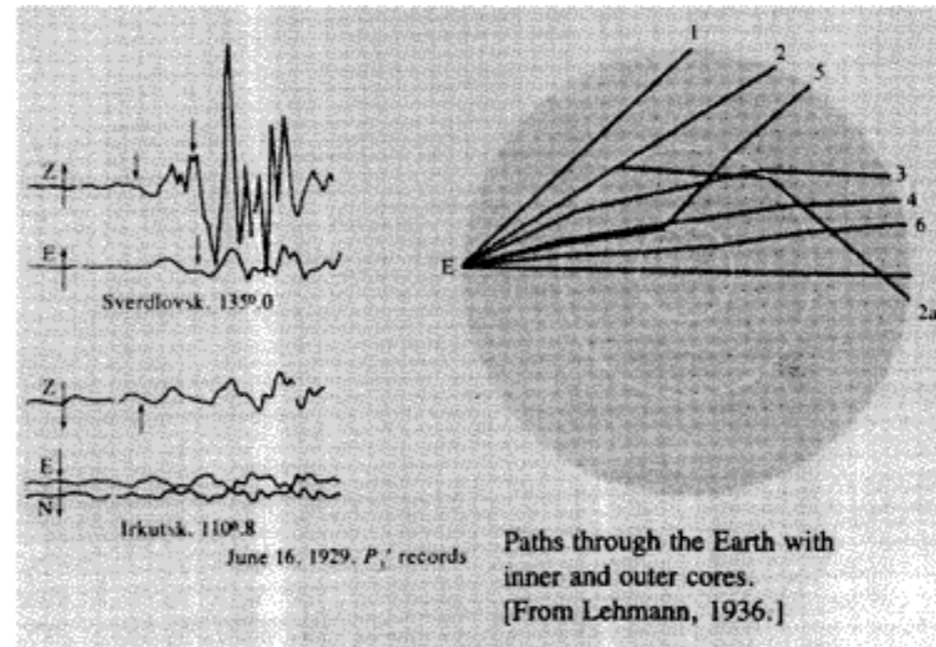
Time curves of first and second phases of preliminary tremors. The marks surrounded by circles are averages. [From Oldham, 1906.]

Beno Gutenberg

1914 Über Erdbebenwellen VIIA. Nachr. Ges. Wiss. Göttingen Math. Physik. Kl, 166.



who calculated depth of the core as 2900km or $0.545R$



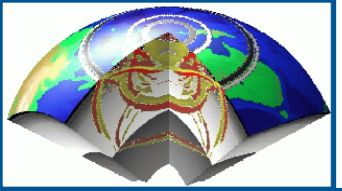
Paths through the Earth with inner and outer cores. [From Lehmann, 1936.]

Inge Lehmann

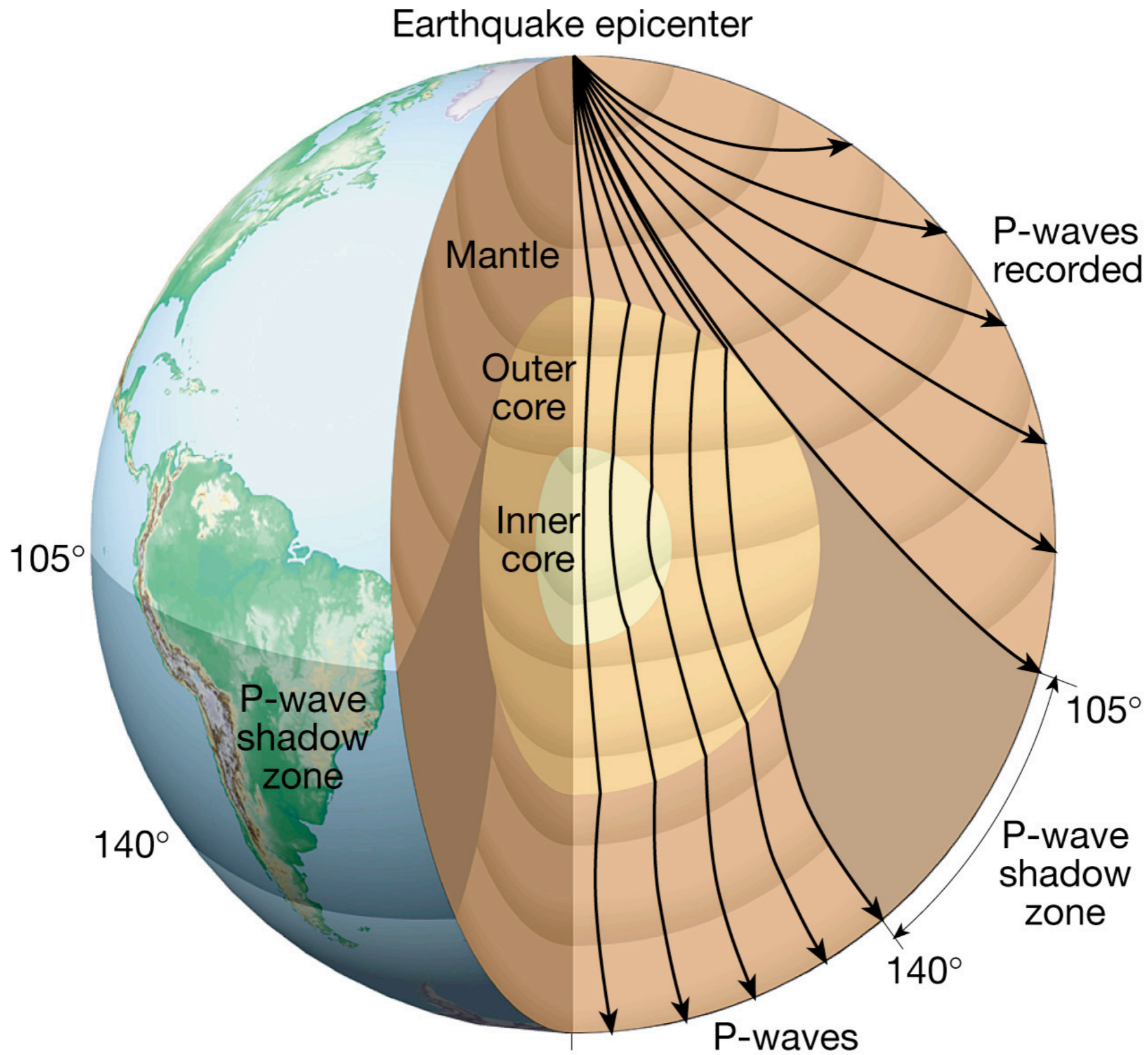
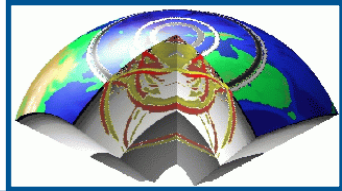
Bureau Central Seismologique International, Series A, Travaux Scientifiques, 14, 88, 1936.

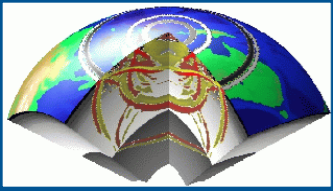


who discovered of the earth's inner core.

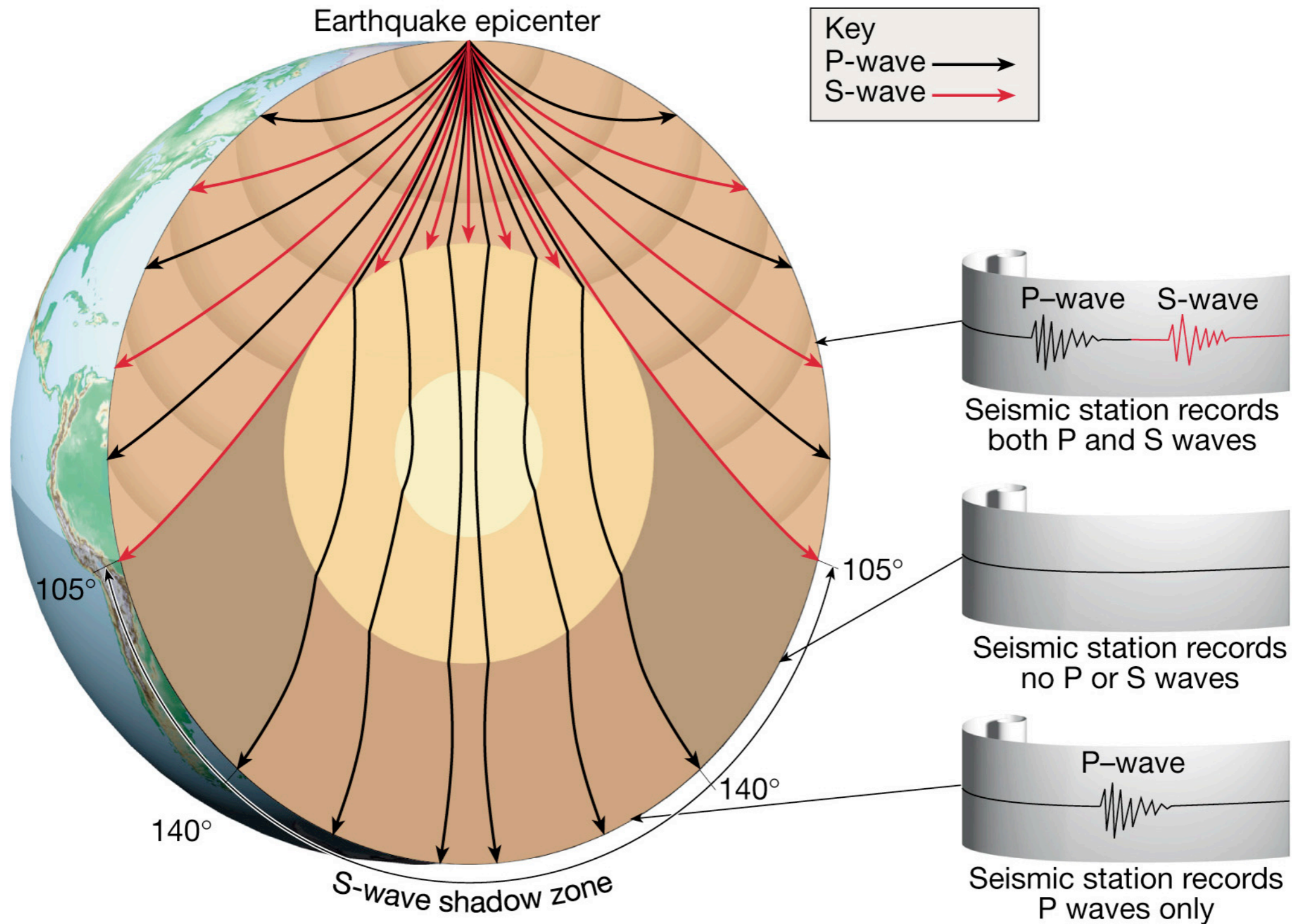
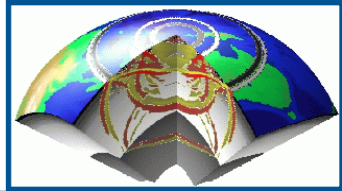


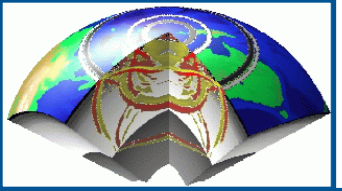
Ray Paths in the Earth (1)



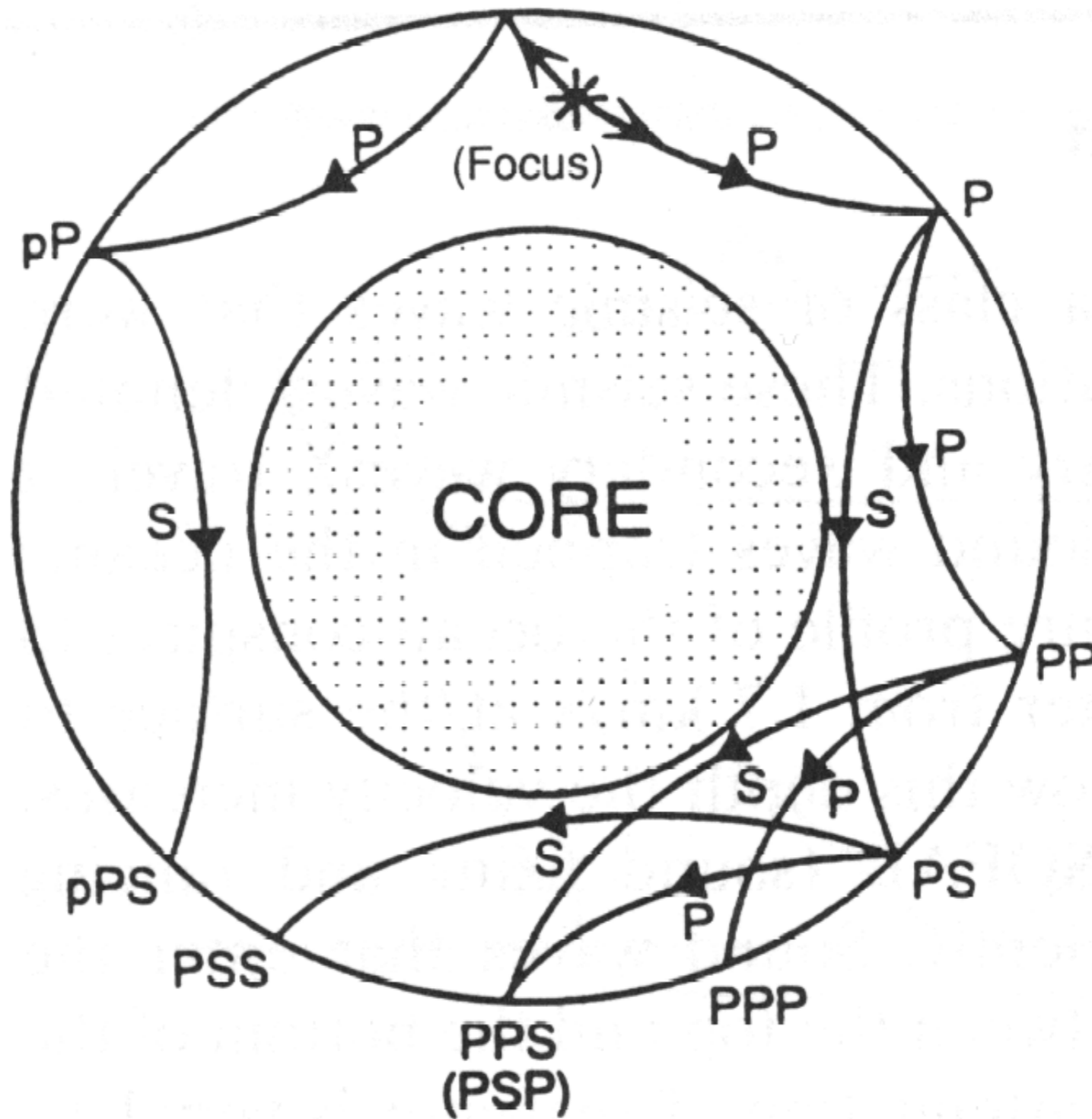
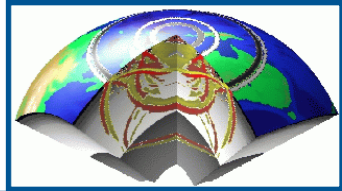


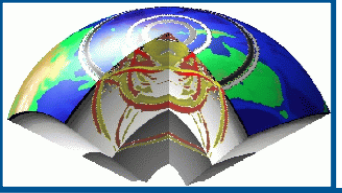
Ray Paths in the Earth (2)





Ray Paths in the Earth (3)





Ray Paths in the Earth (4)

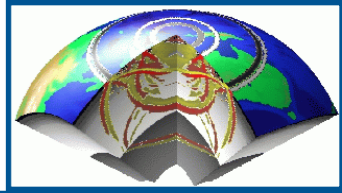
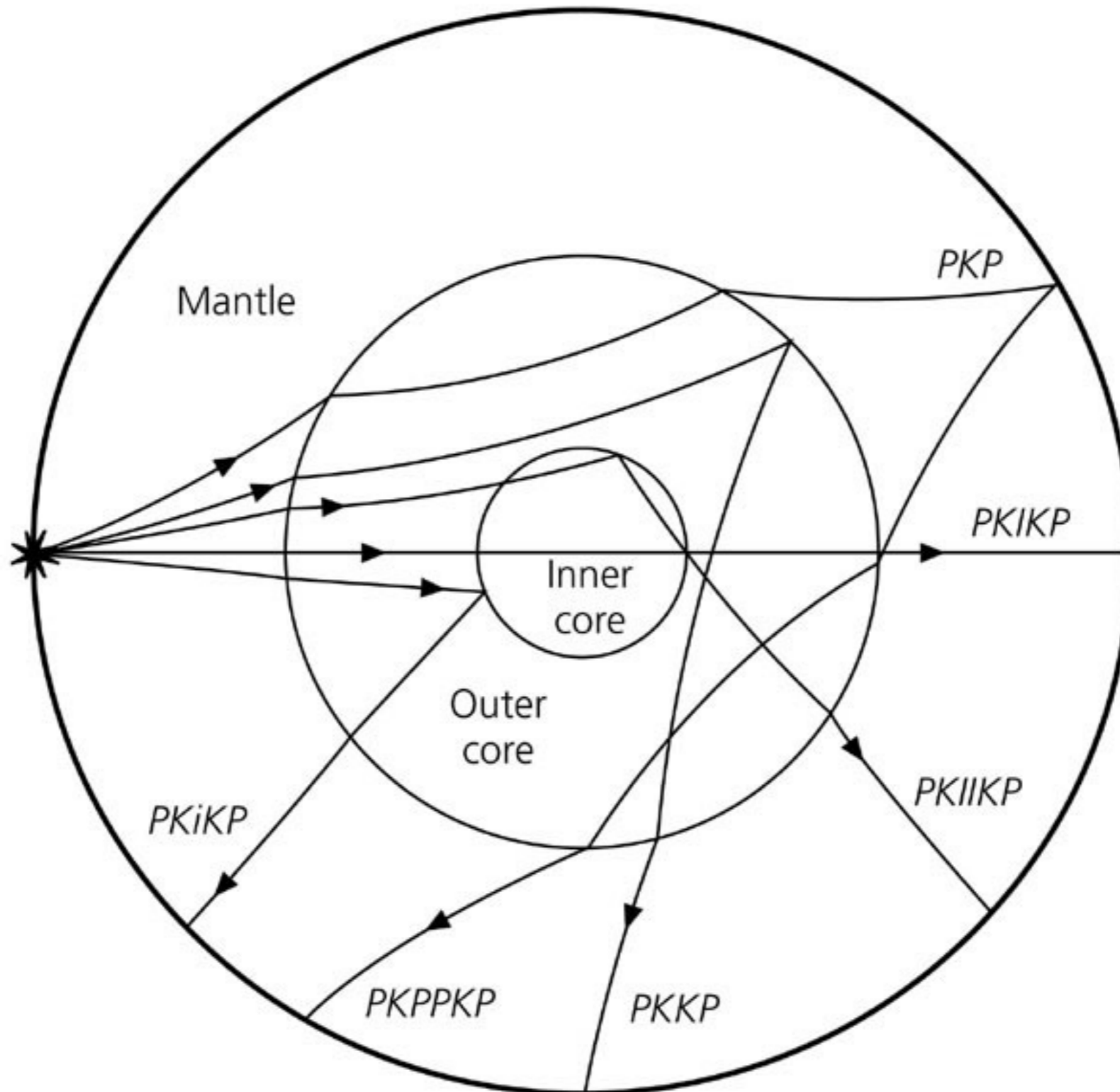
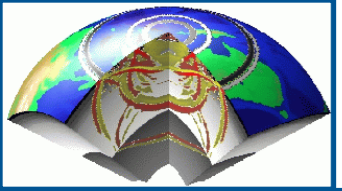
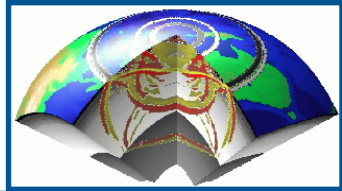


Figure 3.5-10: Ray paths for additional core phases.



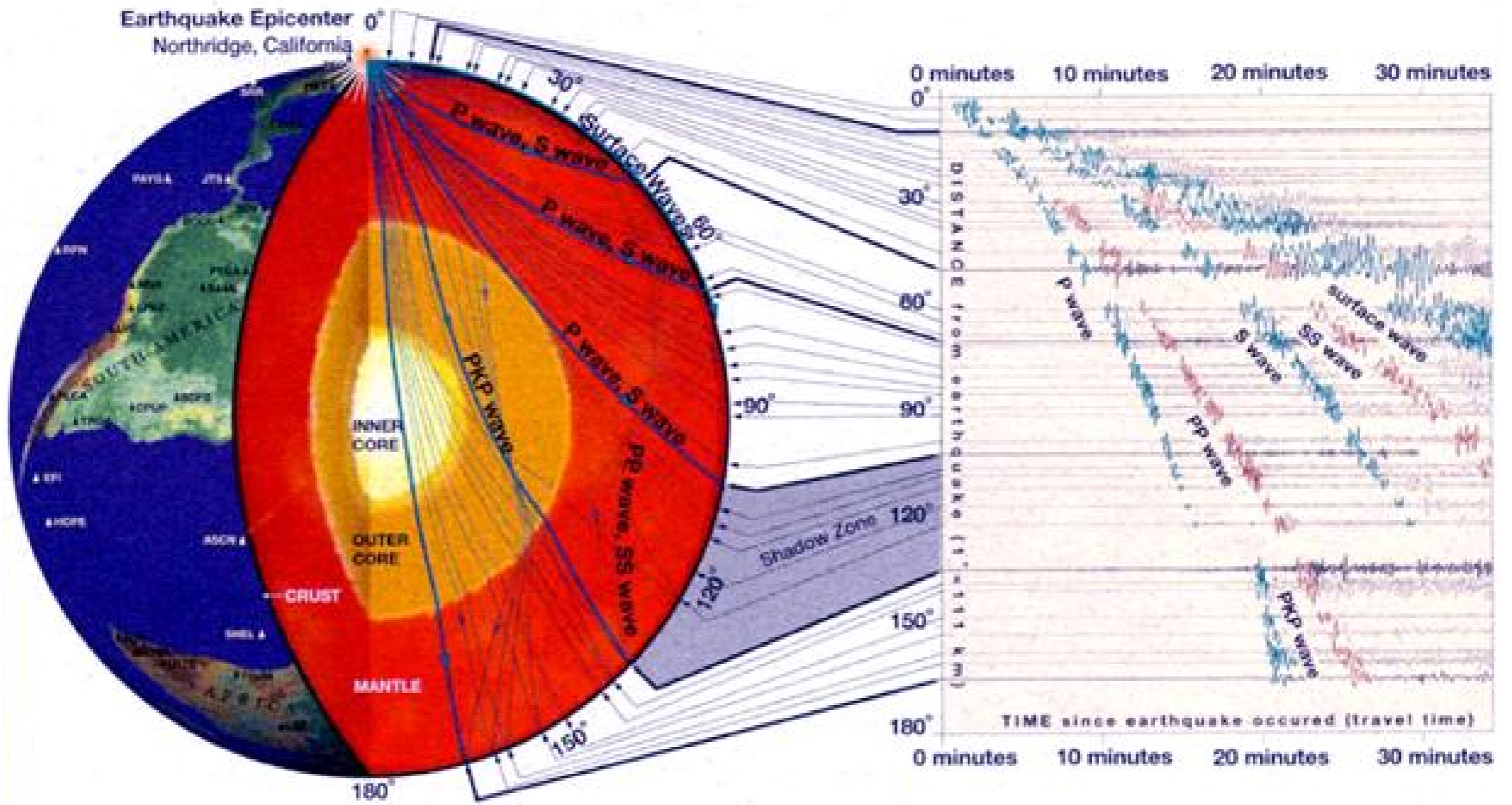
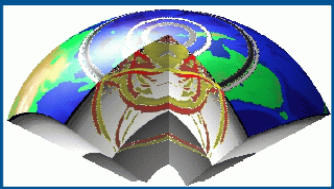
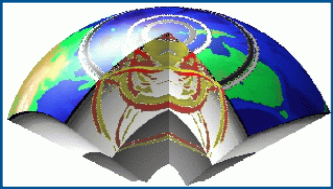


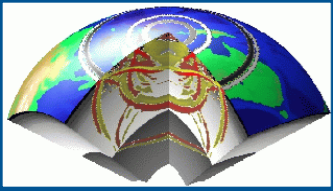
Ray Paths in the Earth - Names



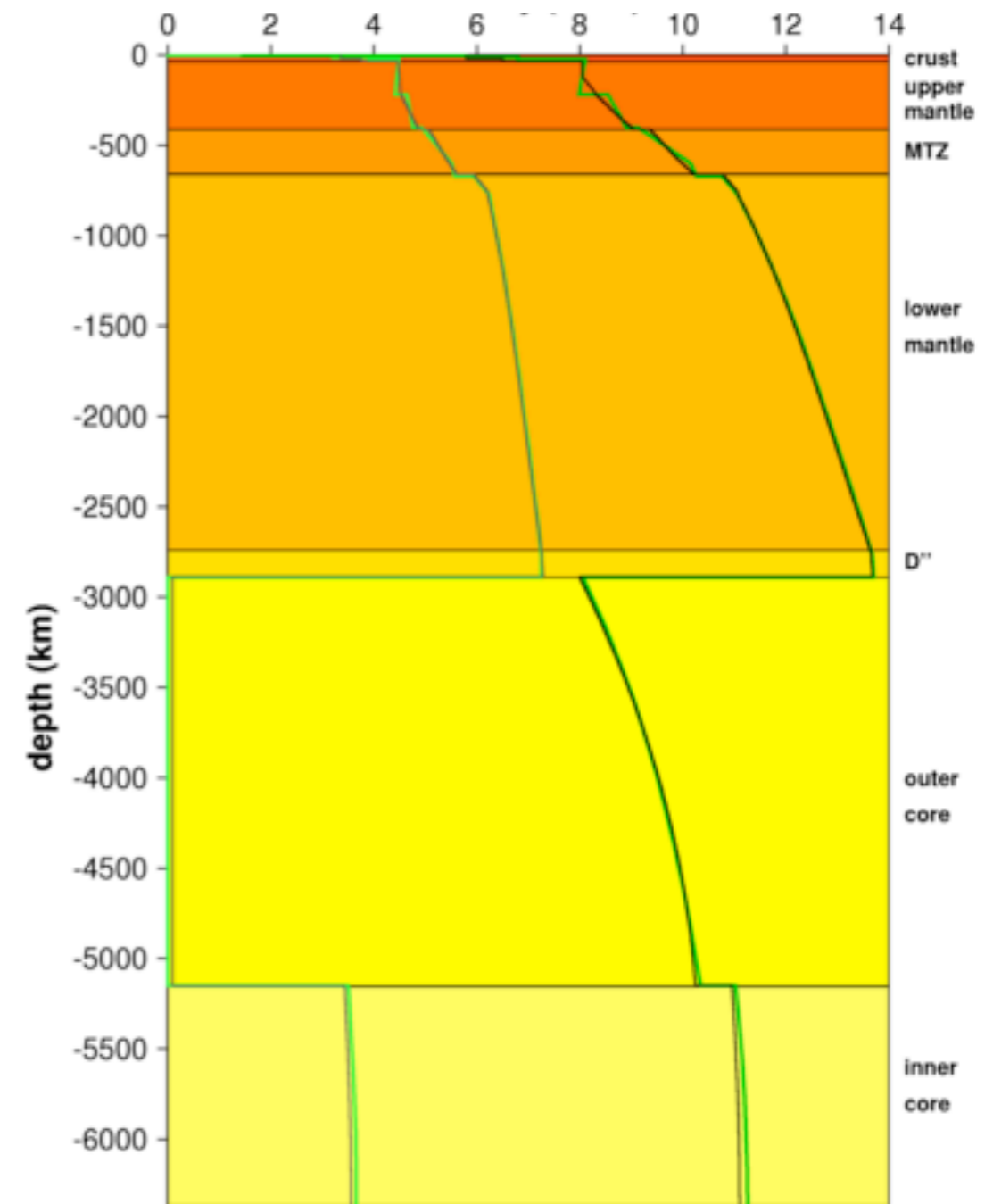
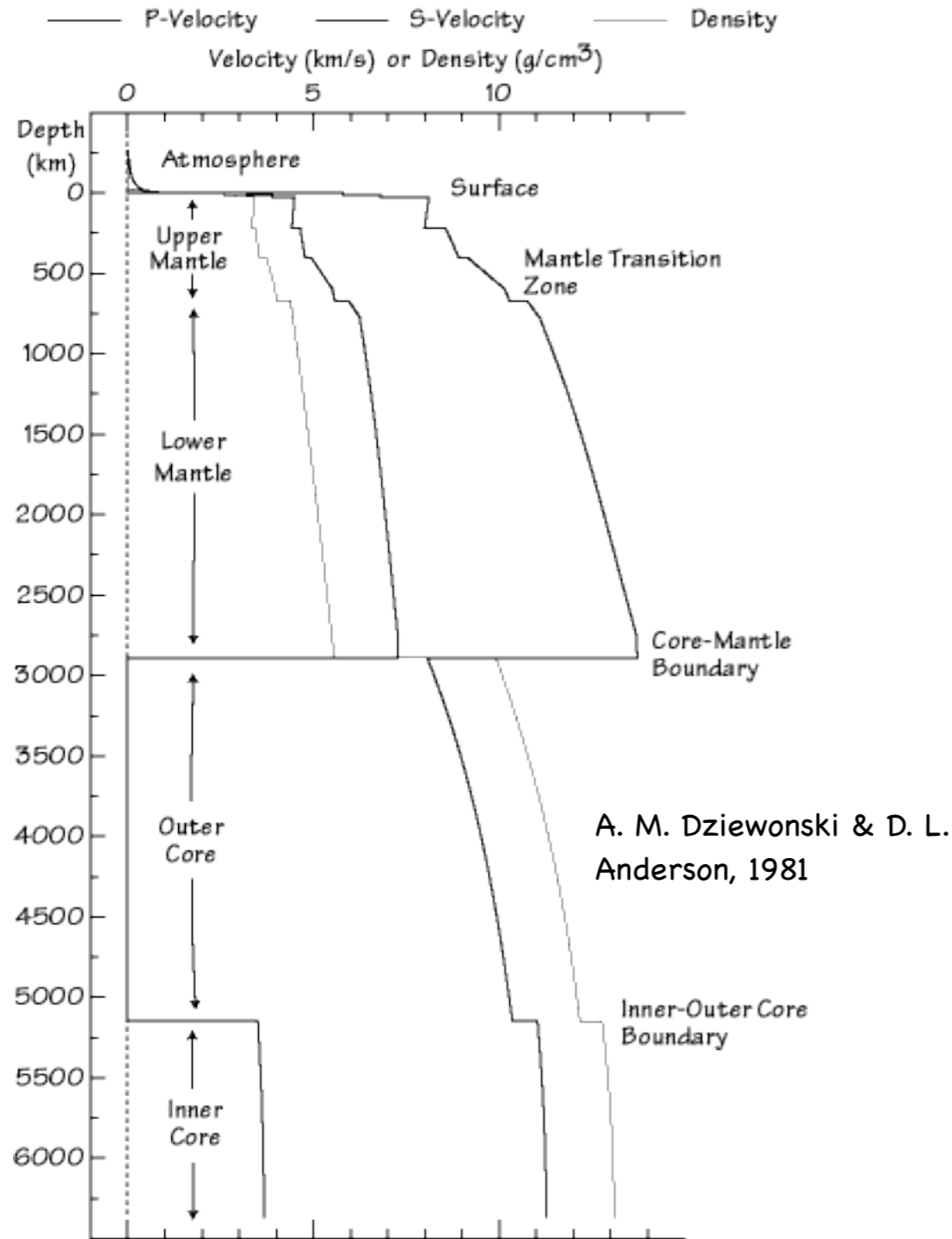
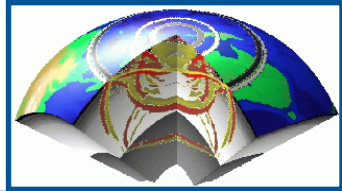
P	P waves
S	S waves
small p	depth phases (P)
small s	depth phases (S)
c	Reflection from CMB
K	wave inside core
i	Reflection from Inner core boundary
I	wave through inner core

Travel times in the real Earth





Spherically symmetric models

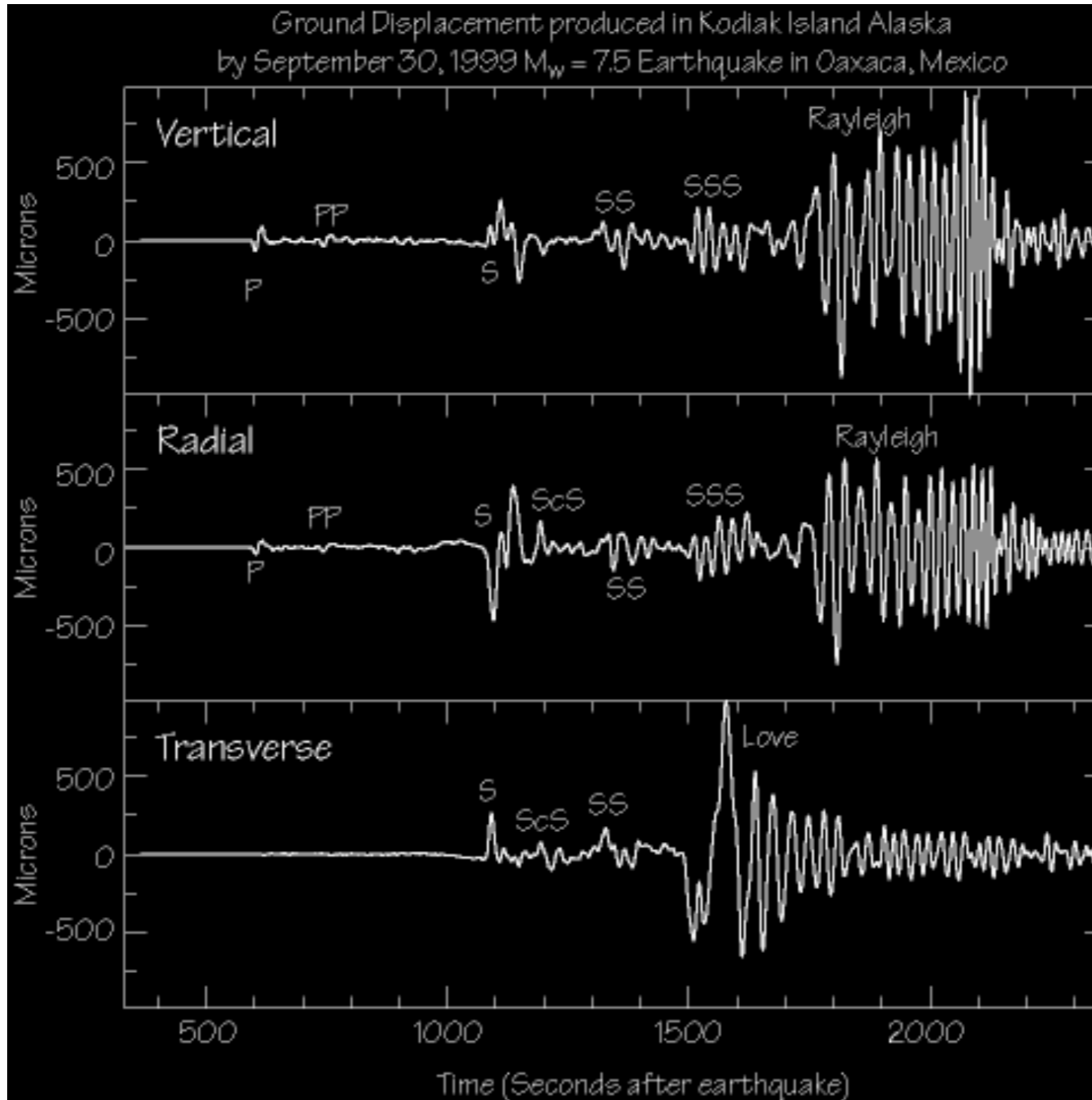


Velocity and density variations within Earth based on seismic observations. The main regions of Earth and important boundaries are labeled. This model was developed in the early 1980's and is called **PREM** for Preliminary Earth Reference Model.

Model **PREM** giving S and P wave velocities (light and dark green lines) in the earth's interior in comparison with the younger **IASP91** model (thin grey and black lines)

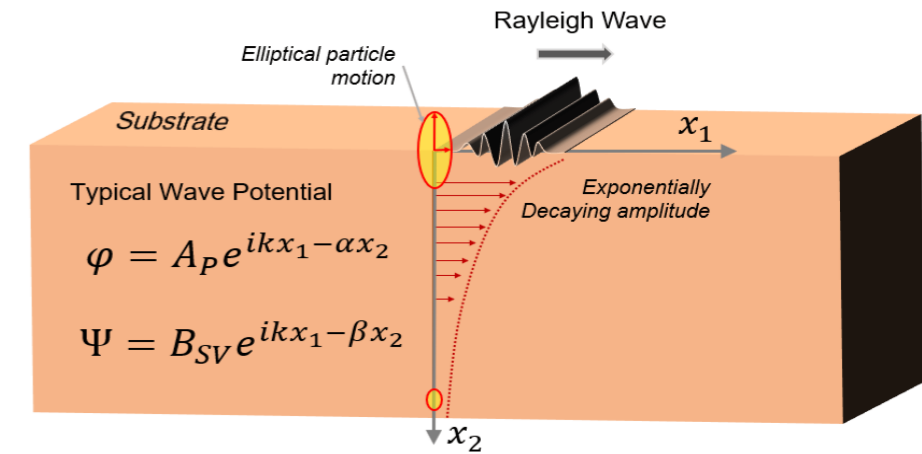
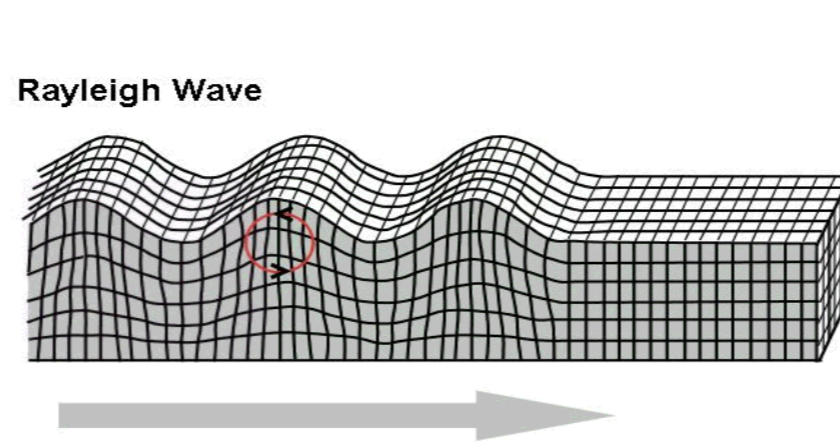
<http://ds.iris.edu/ds/products/emc-referencemodels/>

Data example

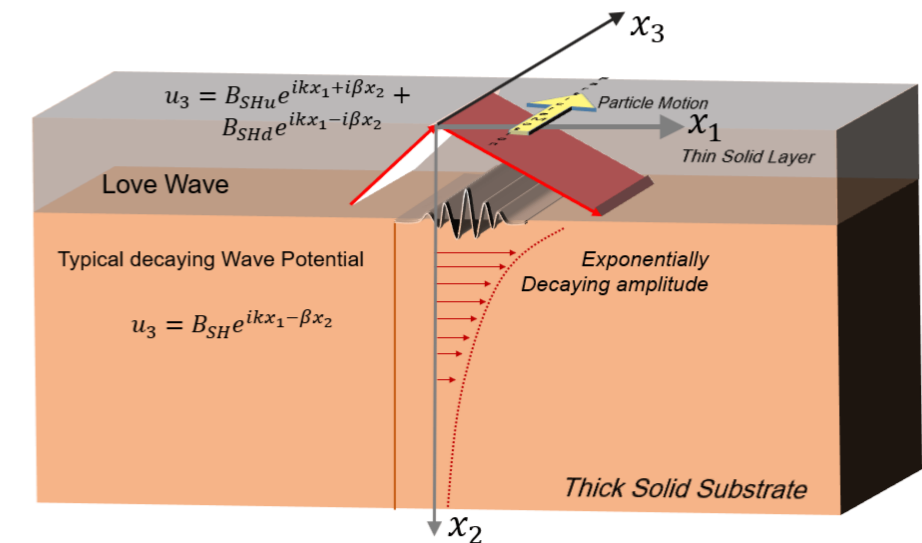
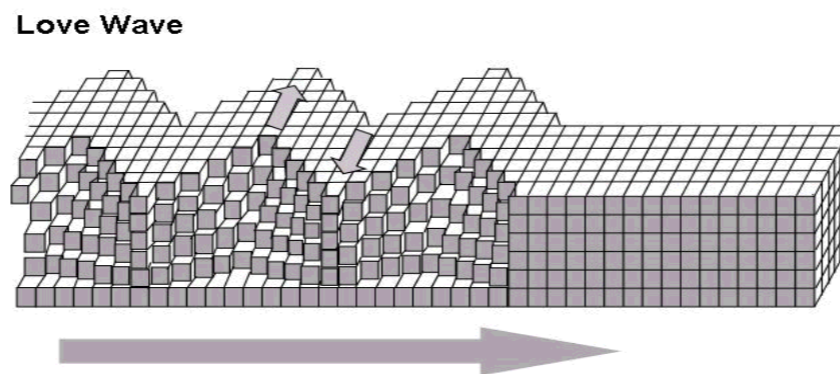


Wavefields visualization - surface waves

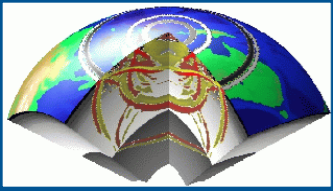
Interference of **P-SV** waves
at surfaces (e.g. free surface)
and velocity is roughly
92% of β



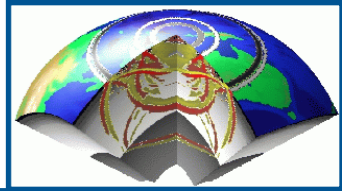
Interference of multiply reflected
SH waves
at surfaces (e.g. free surface)
and velocity depends on β



They are **cylindrical dispersed** waves and decay as $(r)^{-1/2}$



Effect of dispersion...



Demonstration: sum two harmonic waves with slightly different angular frequencies and wavenumbers:

$$u(x, t) = \cos(\omega_1 t - k_1 x) + \cos(\omega_2 t - k_2 x)$$

$$\omega_1 = \omega + \delta\omega \quad \omega_2 = \omega - \delta\omega \quad \omega \gg \delta\omega$$

$$k_1 = k + \delta k \quad k_2 = k - \delta k \quad k \gg \delta k$$

Add the two cosines:

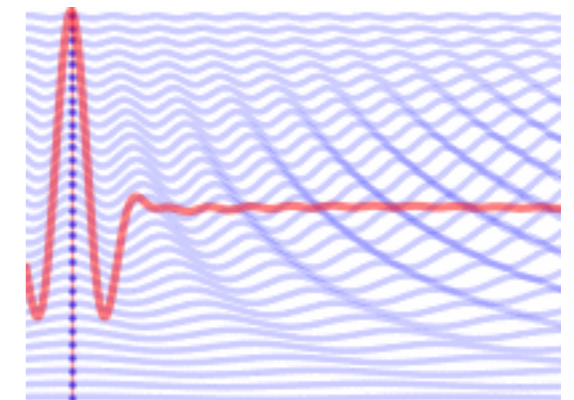
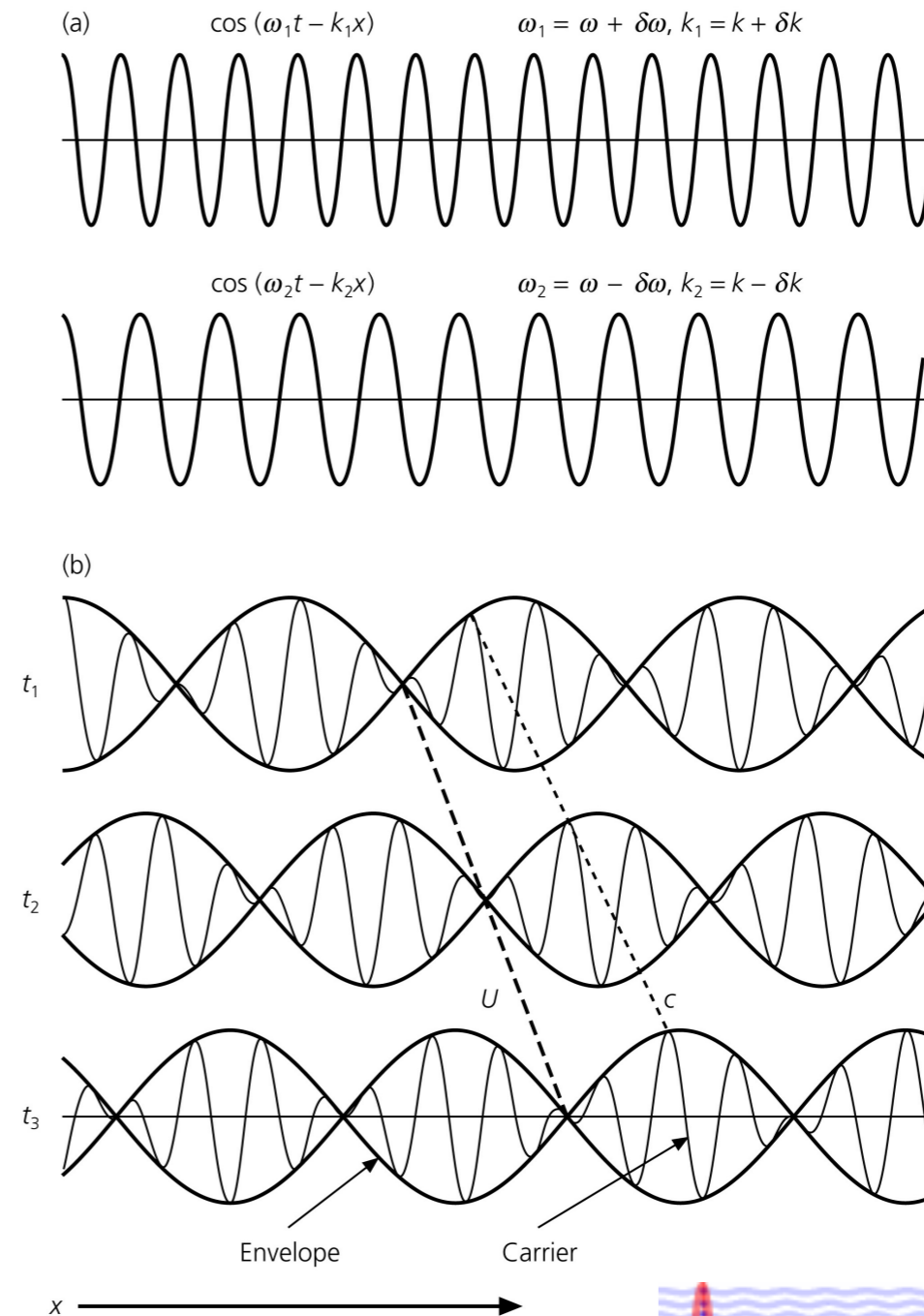
$$\begin{aligned} u(x, t) &= \cos(\omega t + \delta\omega t - kx - \delta kx) \\ &+ \cos(\omega t - \delta\omega t - kx + \delta kx) \\ &= 2 \cos(\omega t - kx) \cos(\delta\omega t - \delta kx) \end{aligned}$$

The envelope (beat) has a *group velocity*:

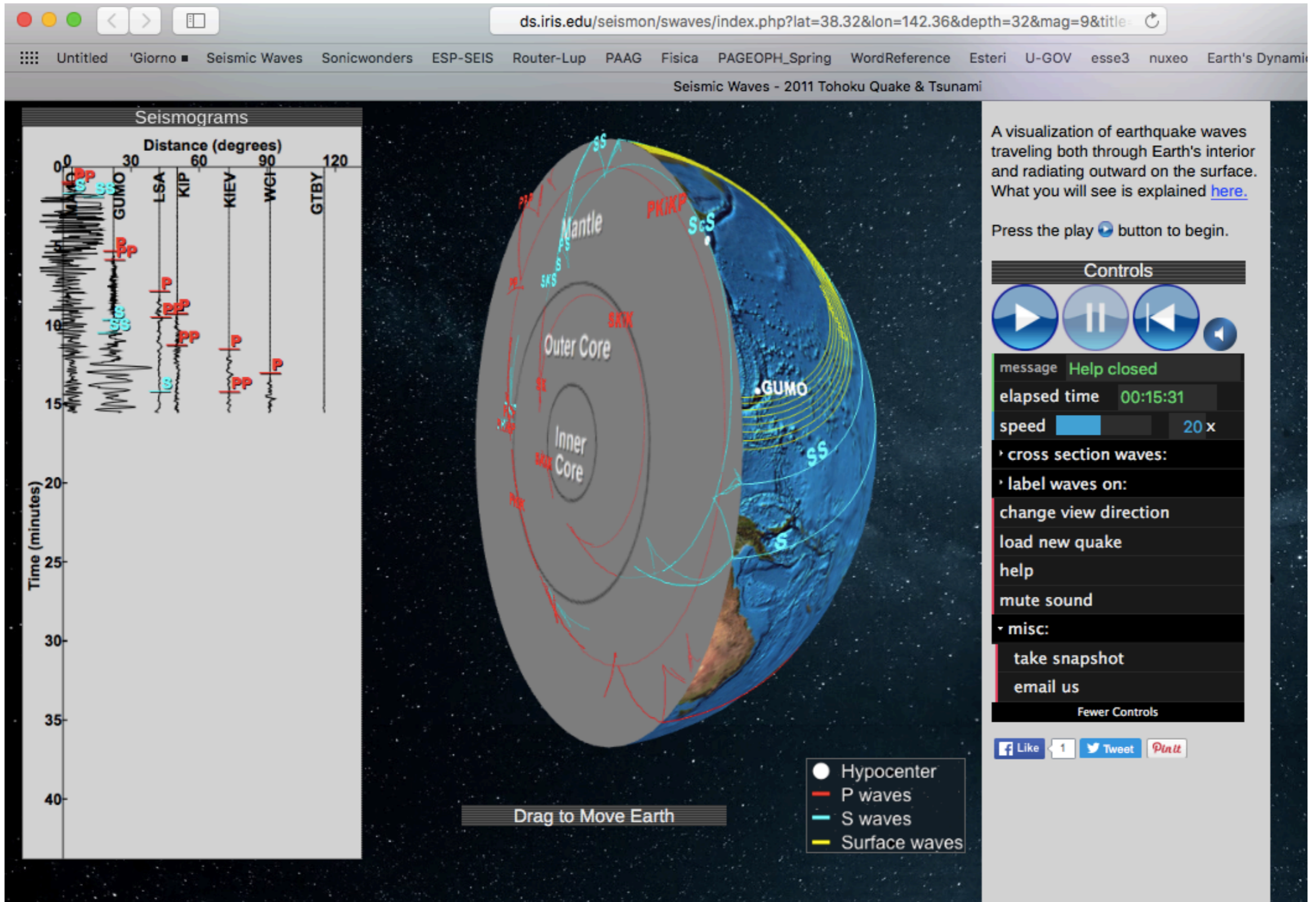
$$U = \delta\omega / \delta k$$

The individual peaks move with a *phase velocity*:

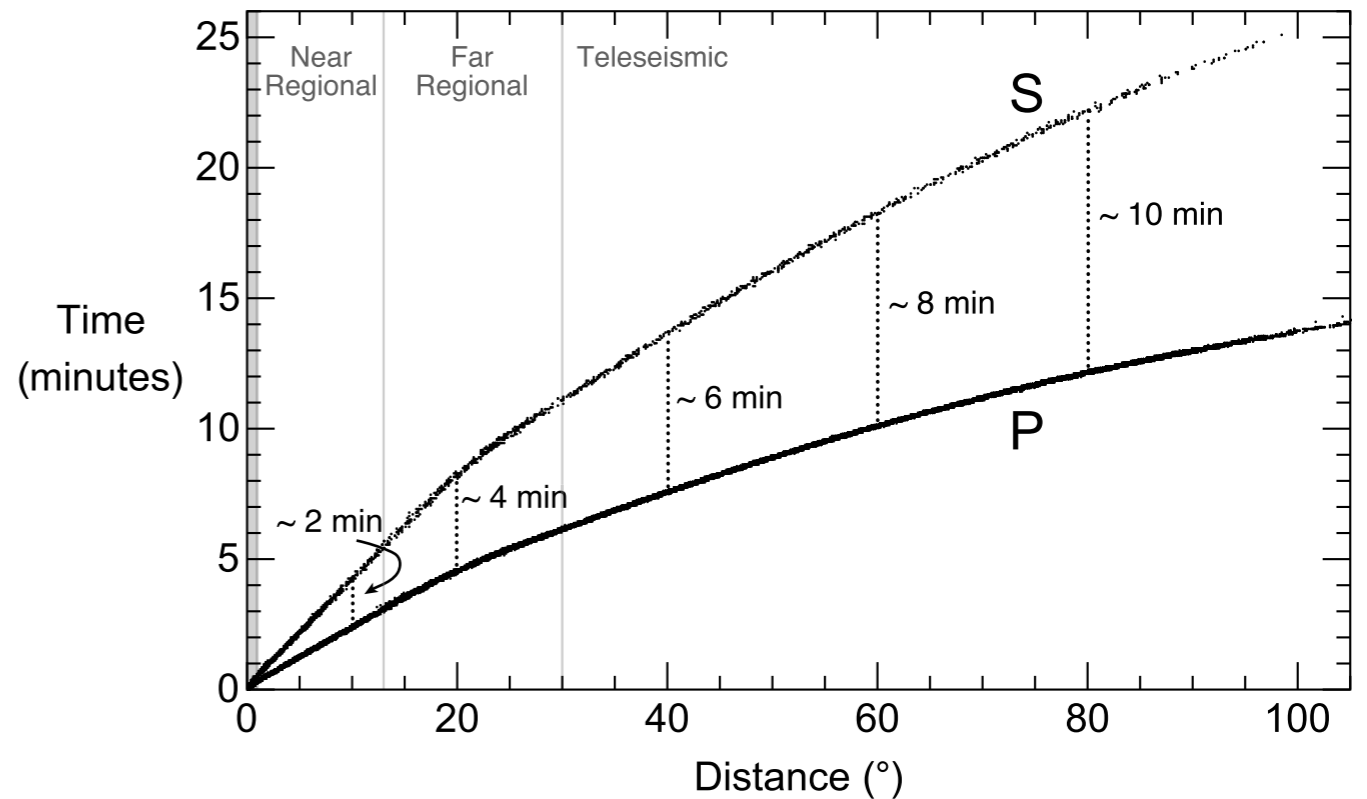
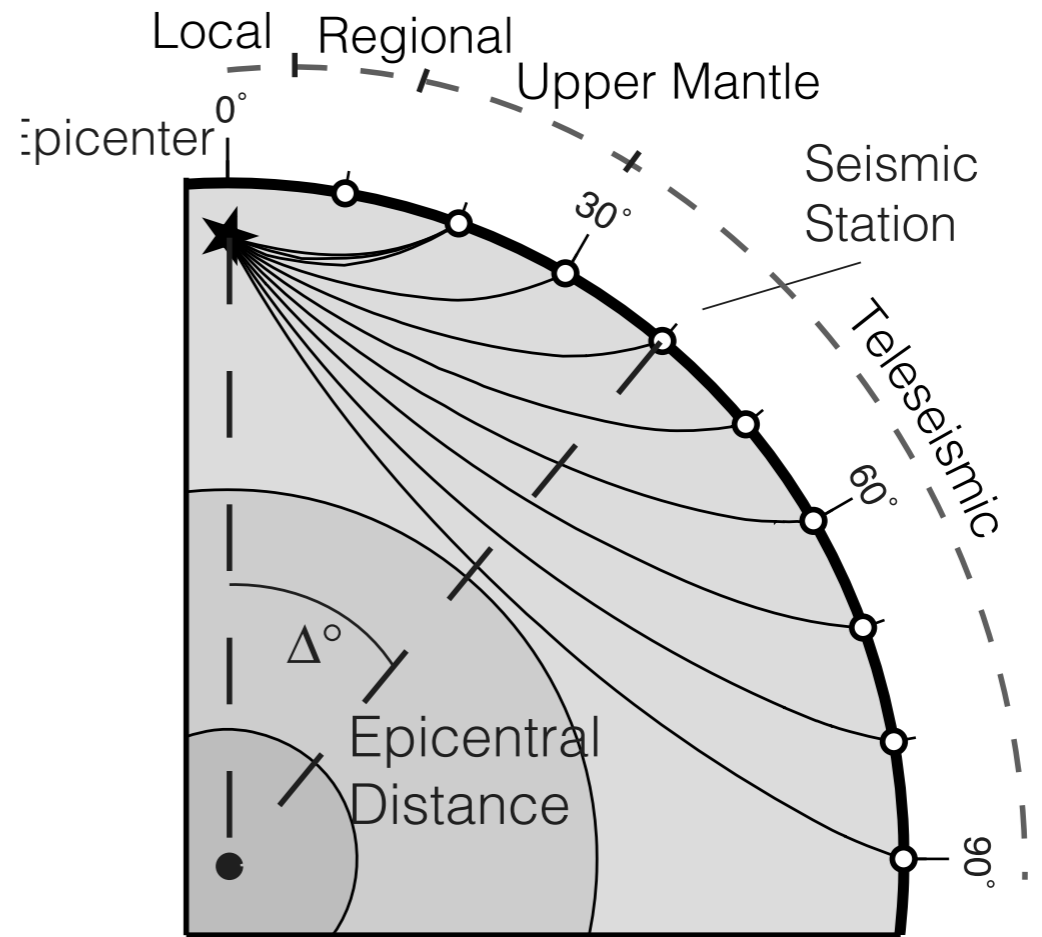
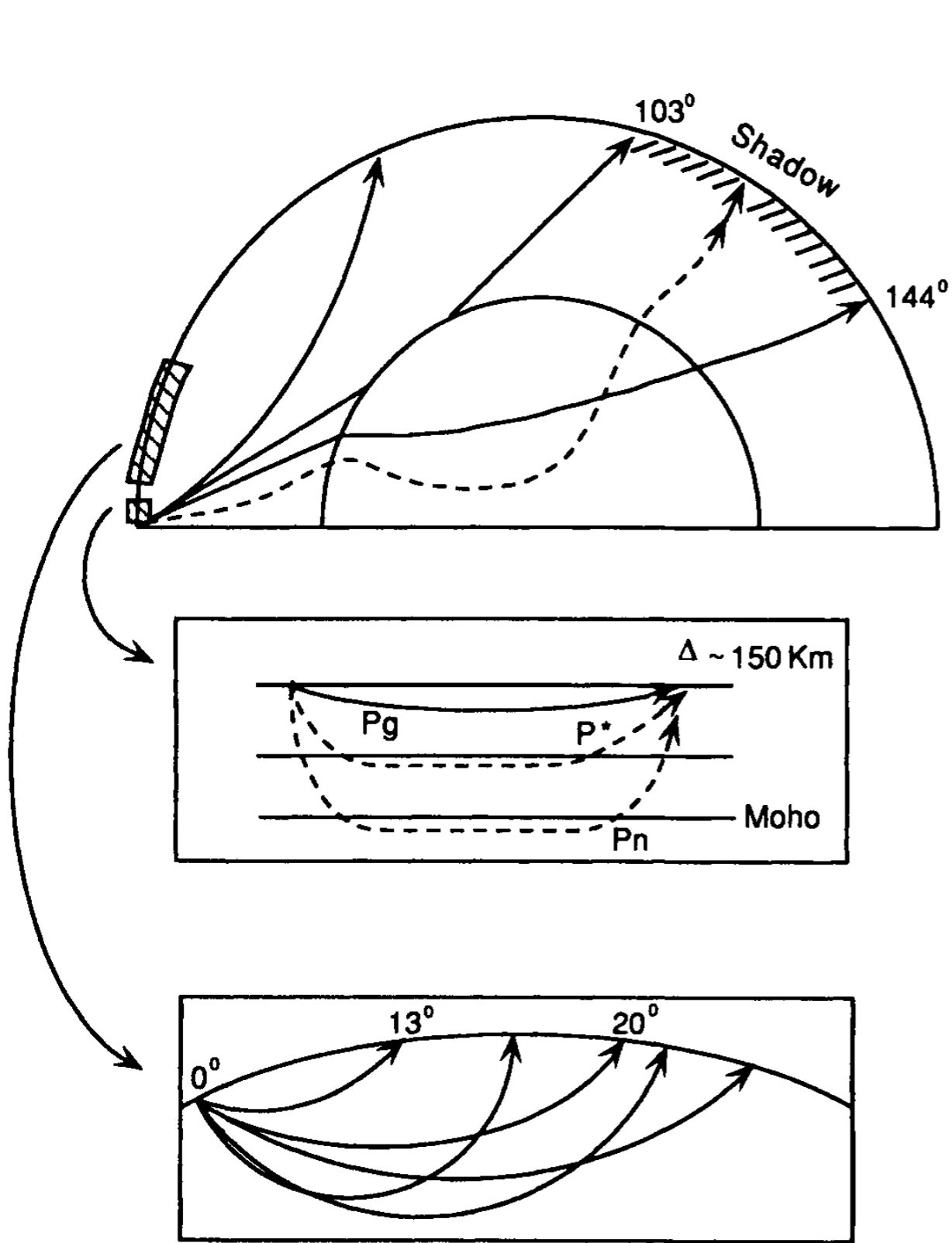
$$c = \omega / k$$



Data example - 2



Seismological (body waves) distance ranges



Strong motion seismology

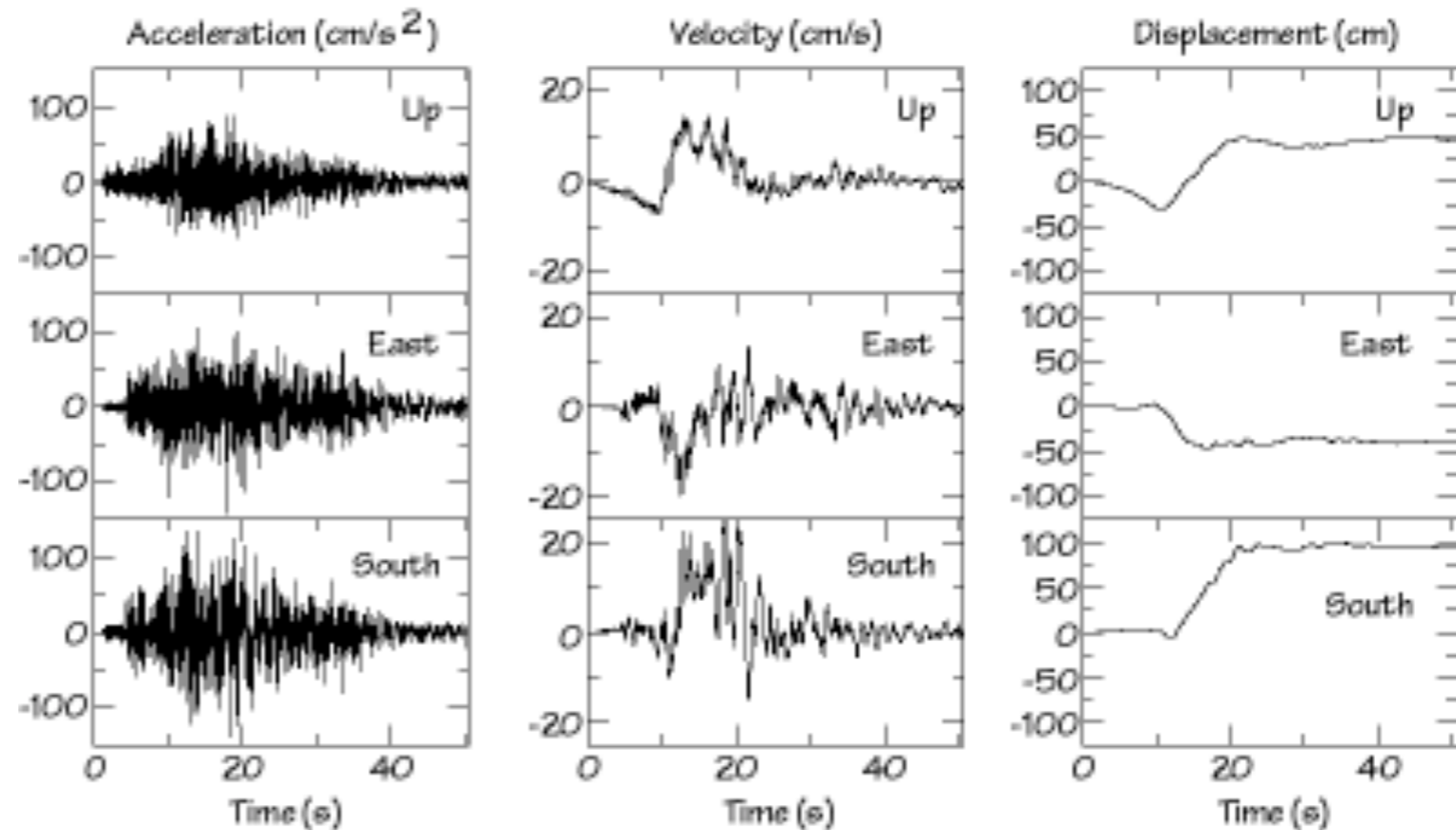
- Strong ground motion is an event in which an earthquake cause the ground to shake at least strongly enough for people to feel the motion or to damage or destroy man-made structures.
- The goal of strong motion seismology is to be able to understand and predict seismic motions sufficiently well that the predictions can be used for engineering applications
- The field of strong-motion seismology could initially be identified with a type of instrument, designed to remain on-scale and record the ground motion with fidelity under the conditions of the strongest ground motions experienced in earthquakes.

Strong motion seismology

- Early instruments were typically designed so that ground motions up to the acceleration of gravity (1g) would be on-scale.
- The lower limit of ground motion considered by the early strong motion seismology studies was roughly defined by the thickness of the light beam read until the edge of a recorded film. The minimum acceleration resolved is somewhat less than 0.01g, that approximately coincided with minimum ground motions that humans are able to feel.
- Since much smaller ground motions can be recorded on modern instruments, the distinction between strong-motion seismology and traditional seismology is blurred.

Example of Recordings

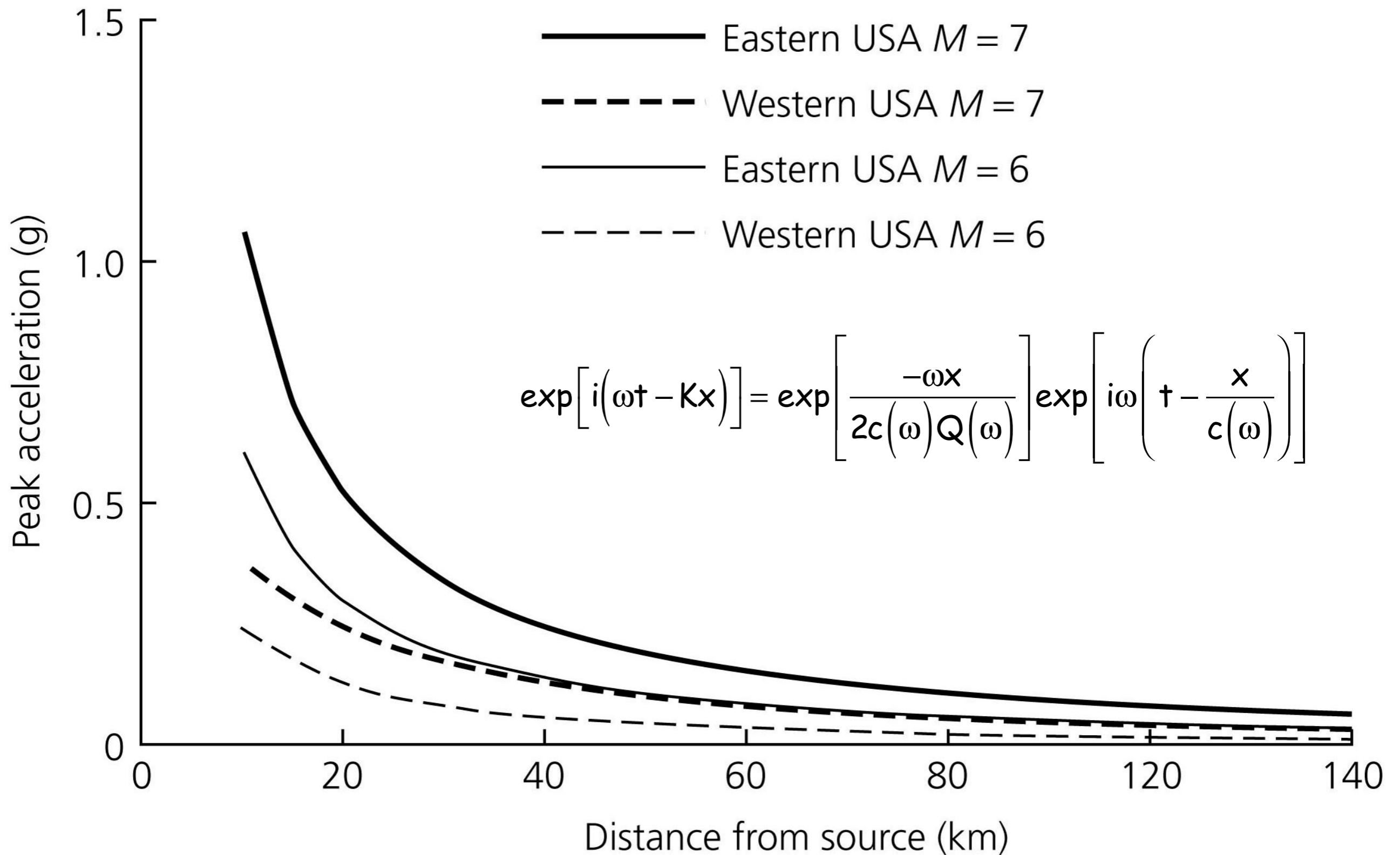
Ground acceleration, velocity and displacement, recorded at a strong-motion seismometer that was located directly above the part of a fault that ruptured during the 1985 Mw = 8.1, Michoacan, Mexico earthquake.



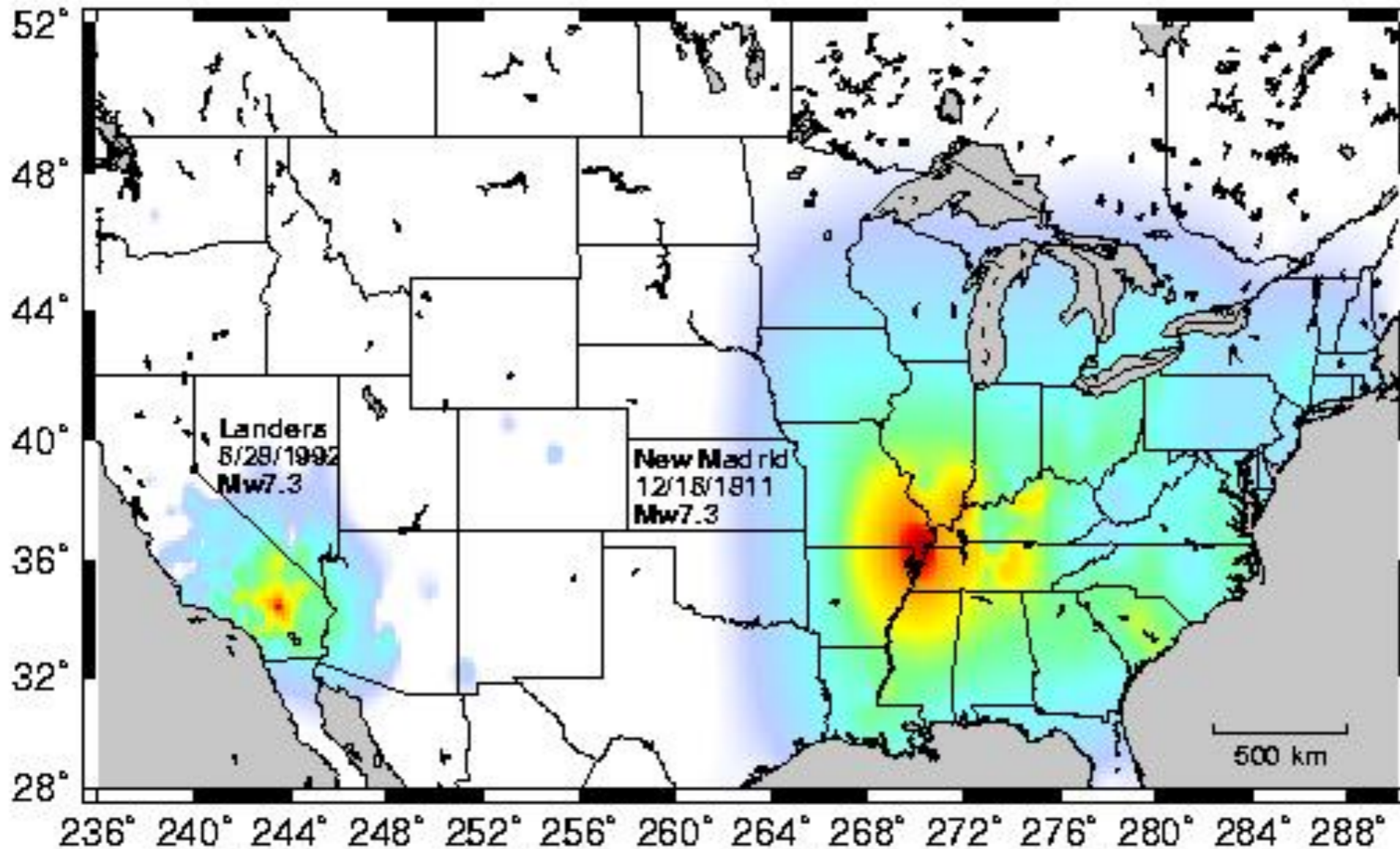
The left panel is a plot of the three components of acceleration: strong, high-frequency shaking lasted almost a minute and the peak acceleration was about 150 cm/s² (or about 0.15g). The middle panel shows the velocity of ground movement: the peak velocity for this site during that earthquake was about 20-25 cm/sec. Integrating the velocity, we can compute the displacement, which is shown in the right-most panel: the permanent offsets near the seismometer were up, west, and south, for a total distance of about 125 centimeters.

“Attenuation” laws

Figure 1.2-5: Predicted strong ground motion in eastern and western U.S.



Seismic hazard



PERCEIVED SHAKING	Not felt	Weak	Light	Moderate	Strong	Very strong	Severe	Violent	Extreme
POTENTIAL DAMAGE	none	none	none	Very light	Light	Moderate	Moderate/Heavy	Heavy	Very Heavy
PEAK ACC. (%g)	< 0.17	0.17-1.4	1.4-3.9	3.9-9.2	9.2-18	18-34	34-65	65-124	> 124
PEAK VEL. (cm/s)	< 0.1	0.1-1.1	1.1-3.4	3.4-8.1	8.1-16	16-31	31-60	60-116	> 116
INSTRUMENTAL INTENSITY	I	II-III	IV	V	VI	VII	VIII	IX	X+