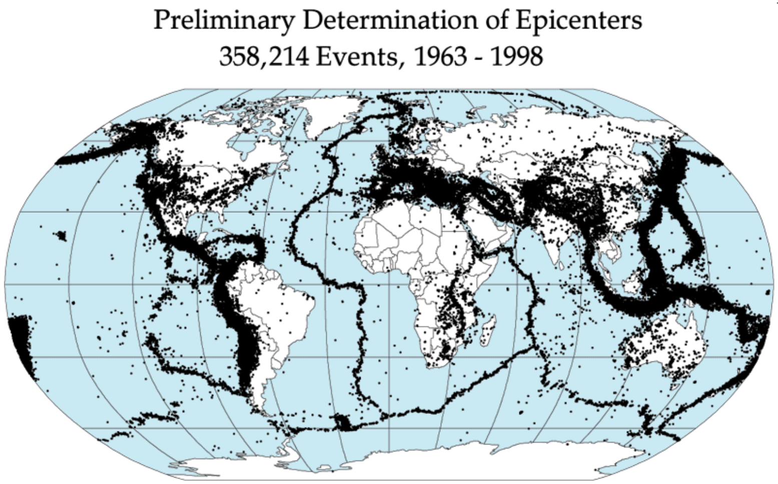
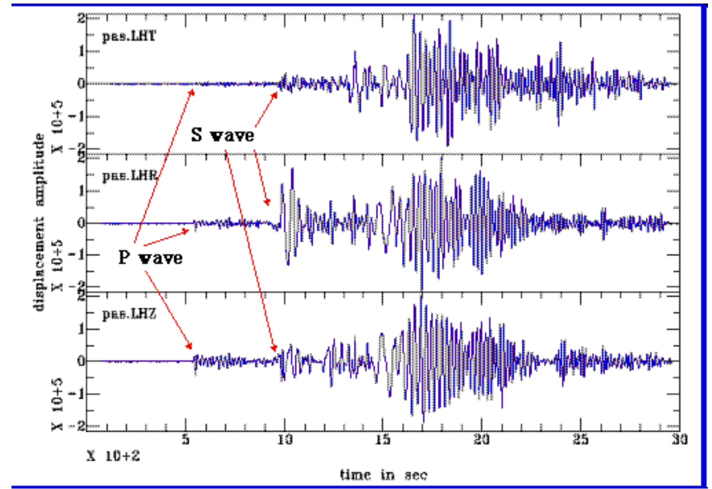
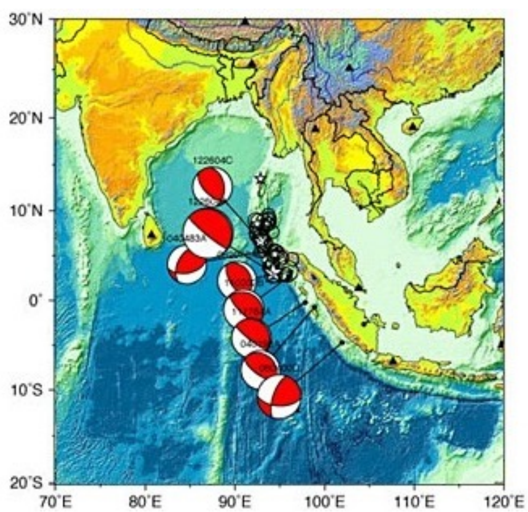
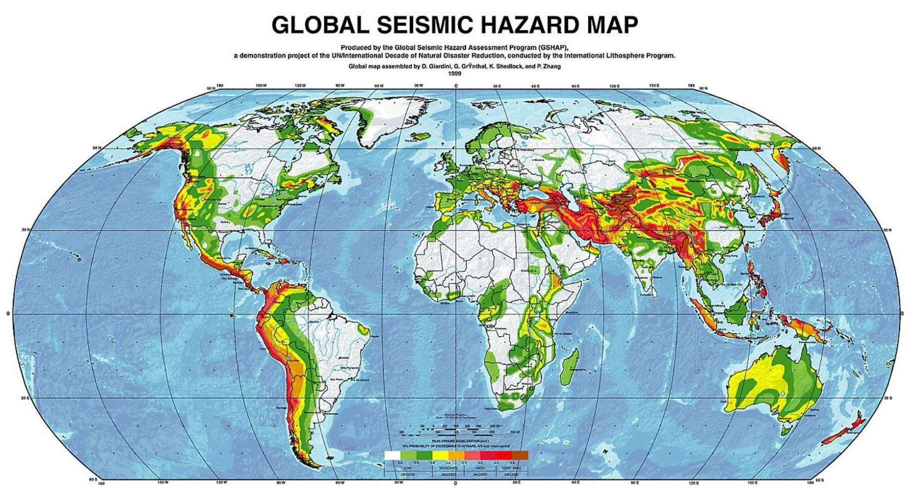




CHAPTER 3
Seismology & Earth's Interior





This file covers much of the material up to section 3.5 of Lowrie book

Theoretical Developments that were necessary

•*Galileo - Responses of Materials to Loading (1564-1642)*
•*Robert Hooke - Law of the Spring - Elasticity (1660)*
(Stress = something x Strain)
•*Navier - Theory of Elasticity followed by Cochy and Poisson. (1700s-1800s)*
•*Rayleigh(1885) and Love(1911) - Surface Waves.*

Development of first 'Seismometer'

•*First seismometer was built in 1875. Allowed detection of far earthquakes.*
•*Observation of P and S waves - predicted theoretically by Poisson.*
•*Rapid Progress - Building of early data base*

The heavy lifters of modern seismology

By 1906 - Oldham suggested presence of large fluid core

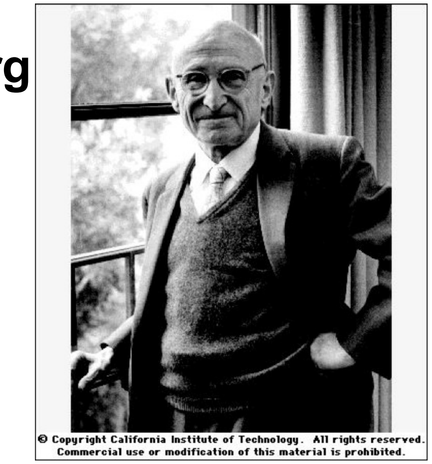
By 1913 - Size of core determined by Gutenberg

By 1909 - Existence of the Crust- Mantle Boundary by A. Mohorovicic (from Yugoslavia)

By 1936 - Existence of the inner core (solid?) by Inge Lehmann. A discussion of Inge Lehmann from the Jeffrey's point of view can be found at http://www.physics.ucla.edu/~cwp/articles/jeffreys/jeffreys_obituary.html

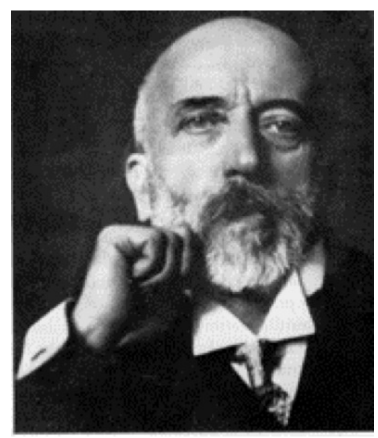
By 1940 - Continued development of 'global' data base of seismograms acquired from large earthquakes - compilation into extensive set of travel time tables by Jeffreys.

Beno Gutenberg

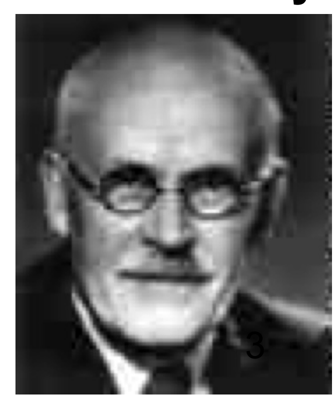


Inge Lehmann

Andrija Mohorovicic



Sir Harold Jeffreys



!960-70: Validation of Plate Tectonics

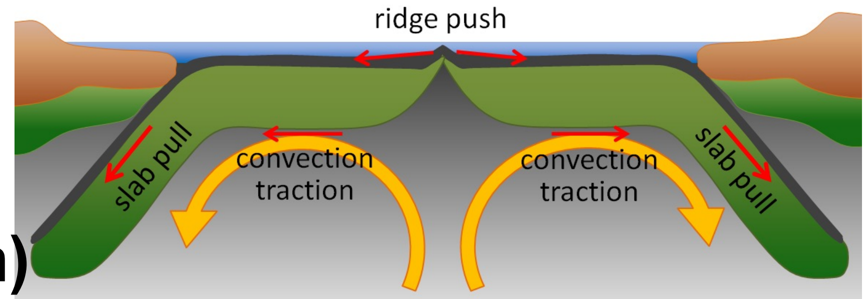
Alfred Wagner

Continental Drift (1915):
One of the 4 ideas that never got Nobel Prizes they deserved.

Tuzo Wilson (Canadian)

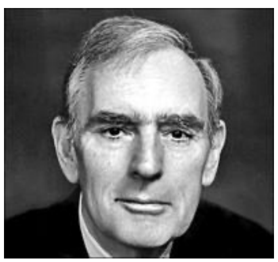
Multiple stages of ocean basin development and closure,
Concept of Plate Tectonics (1968)

The 'Wilson cycle'



Ridge push- slab pull:
Recycling of oceanic crust and mantle

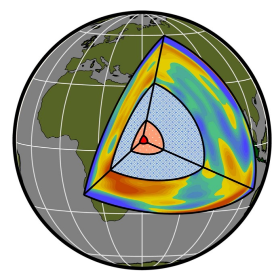
Validation of Plate Tecotnics: Discovery of ocean magnetic stripes, 1960-1970



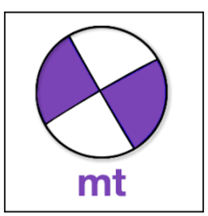
Adam Dziewonski (1936-2016)

1. **Preliminary Reference Earth Model (Dziewonski & Anderson, 1981):** First 1D model of seismic wave speeds and density vs. depth
2. **Seismic Tomography**

(Woodhouse & Dziewonski, 1984)
mantle is heterogeneous with different seismic properties at different locations



3. CMT: Earthquake Source Determination using Centroid Moment Tensors (***Dziewonski, Chou and Woodhouse, 1989***)



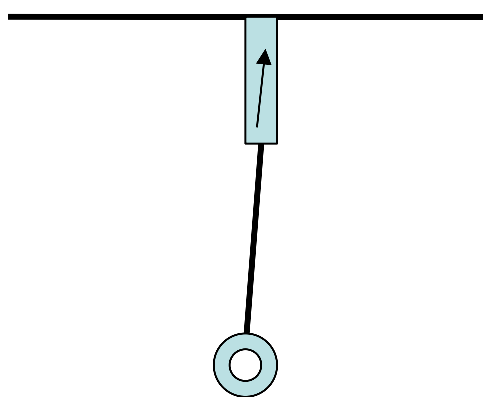
Observational Seismology (a young science)

Earliest known *seismoscope*, by Zhang Heng (132 AD)

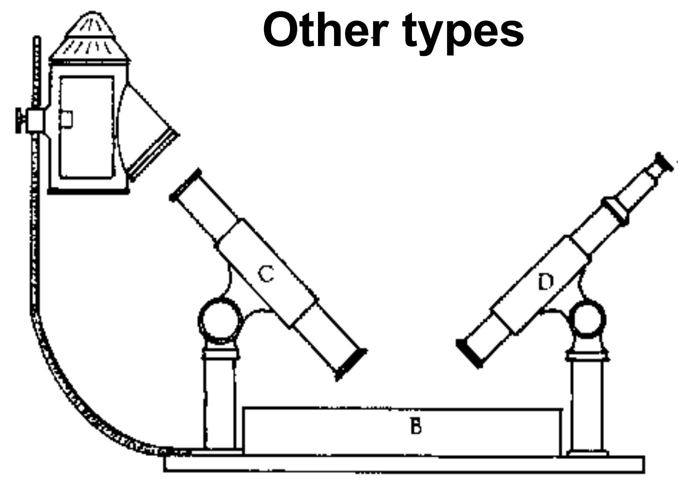
Similar idea

In 1700s, the invention of *spill-meter*
De la Haute Feulle

In 1840s, James Forbes designed the first inverted pendulum "seismometer"

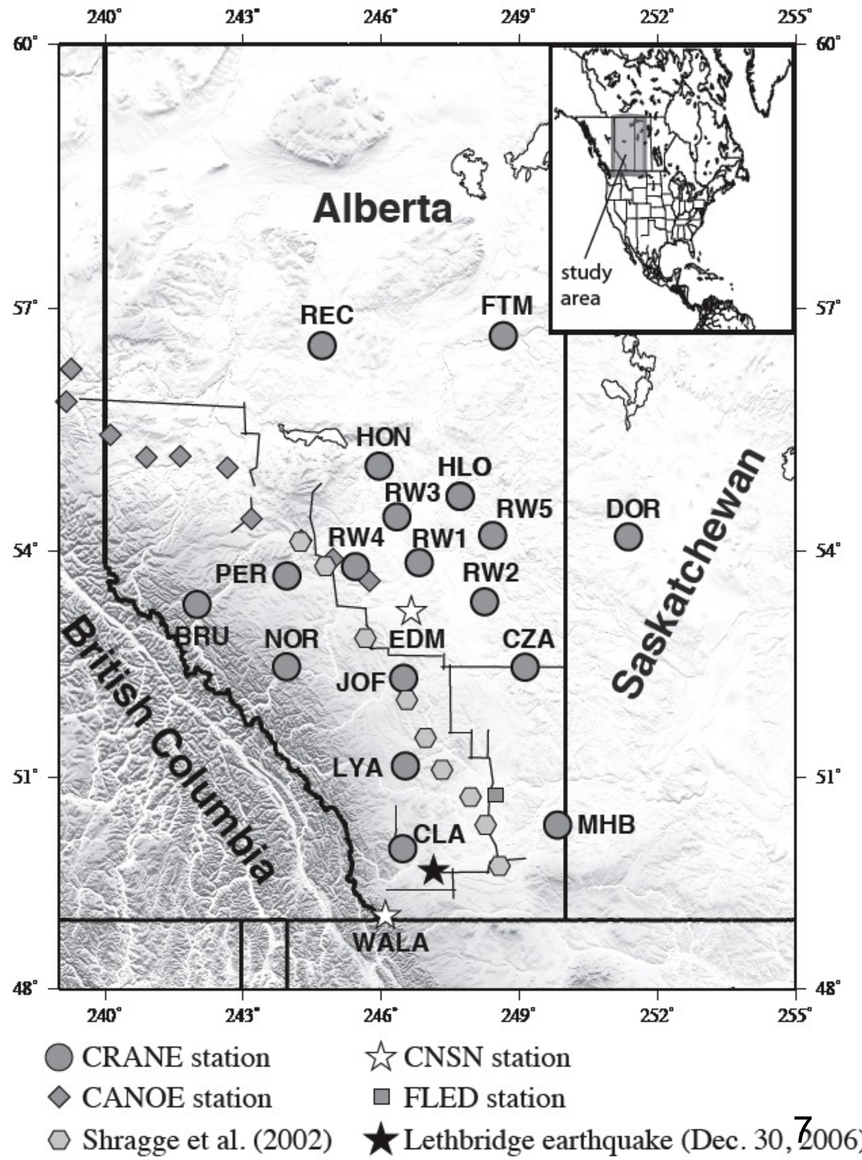


Other types



A remake of the seismoscopes of theGu household, 1976, successfully mitigated hazards associated with the Tangshan earthquake (magnitude = 7.8, 180 km from my Beijing apartment building).

Our own deployment at UofA: (CRANE seismic array)



Seismic waves – basic concepts

A wave:

- is a periodic disturbance
- transmits energy through a material
- no permanent deformation

Seismic waves:

- transmits elastic strain energy
- generated by a seismic source (explosion, earthquake…)

Seismic wave is a ‘wave’ after all!

v = velocity (m/s)

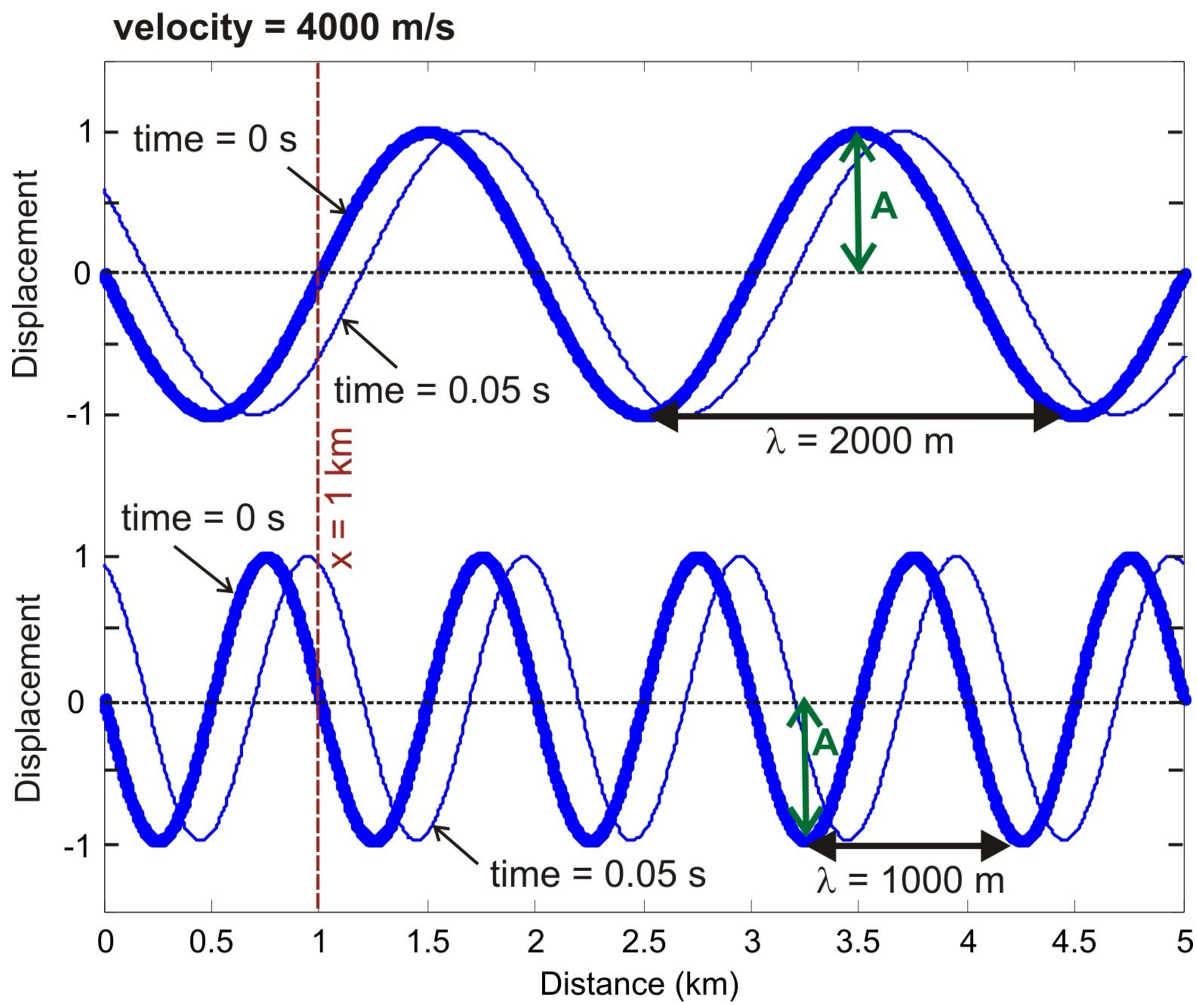
• speed at which the wave travels

A = amplitude (m)

• maximum displacement from rest position

λ = wavelength (m)

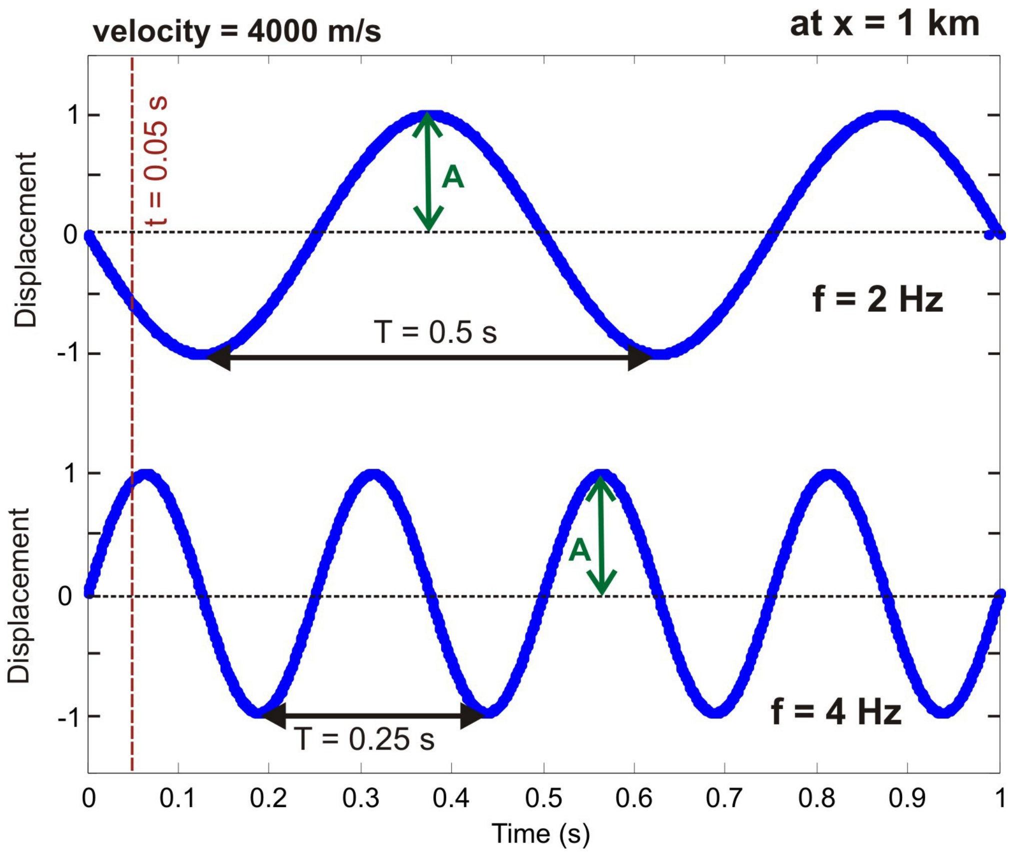
• distance between two points with same phase (e.g., peaks, troughs)



Displacement of one particle over time →

• particle moves up and down while wave goes left or right **= Shear Wave**

Particle motion (individual) does not equal to Wave (collective behavior)



T = period (s)

→ time for one complete cycle (oscillation)

f = frequency (s^{-1}, Hz)

→ number of cycles per second (=1/T)

Angular frequency: **$\omega = 2\pi f$ (rad s^{-1})**

$\boxed{v=d/t \;\&\; v=f\lambda}$

k **= spatial frequency= wave number = $2\pi/\lambda$**

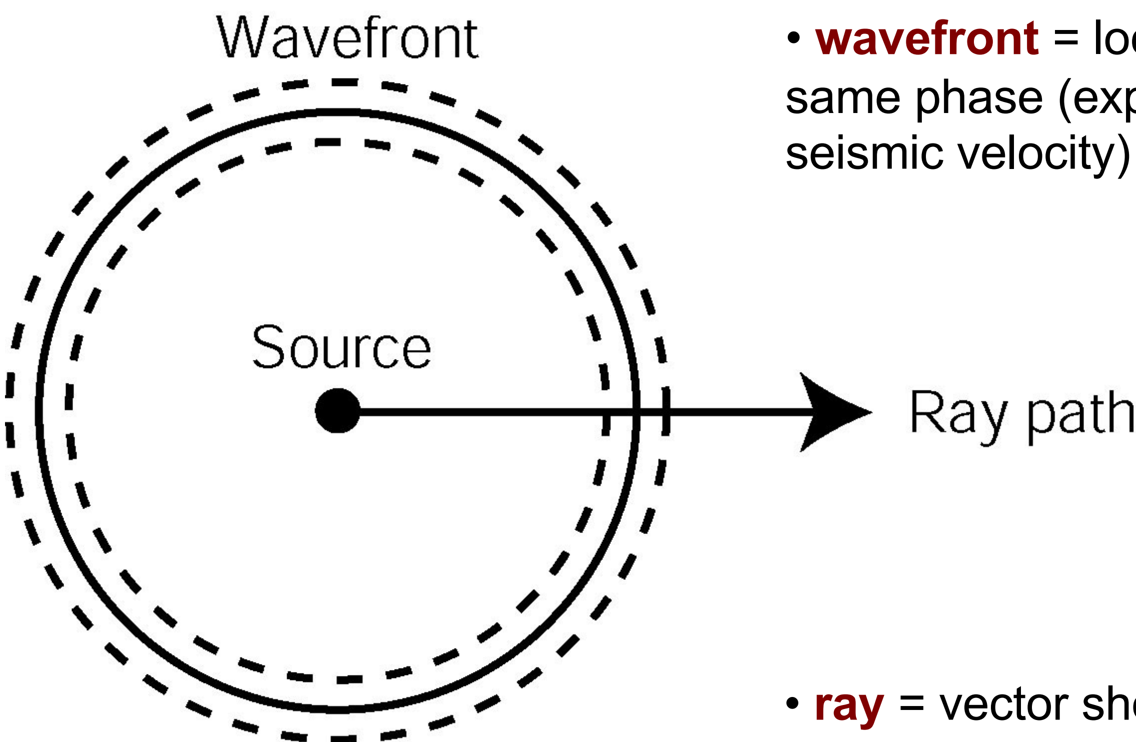
k **= ω/v**

Seismic wave propagation

Seismology – propagation of waves through the Earth

• governed by the wave equation (complex)

• can also be addressed through **visualization**



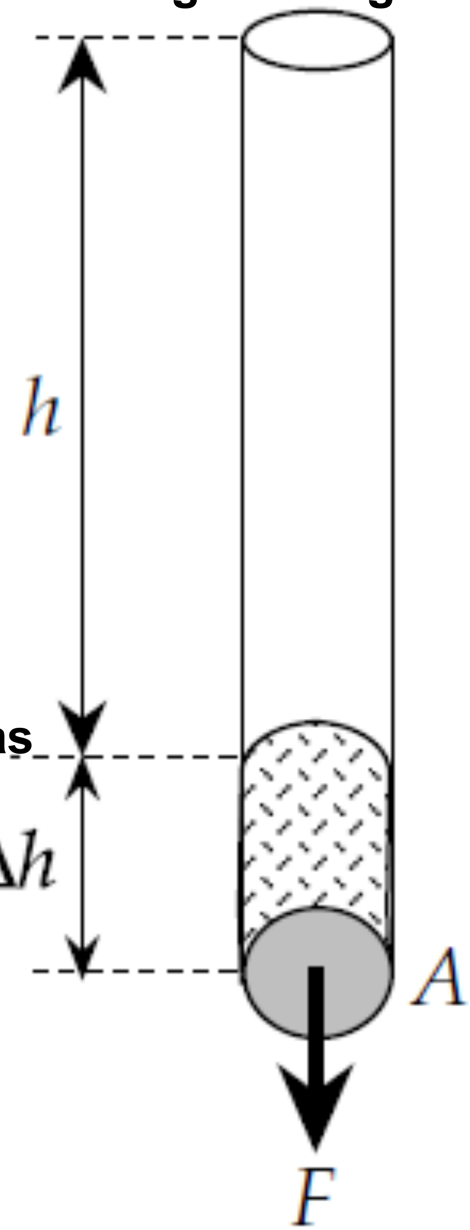
• **wavefront** = locus of points with the same phase (expands spherically at seismic velocity)

• **ray** = vector showing direction of travel of one point (orthogonal to wavefront)

Imagine that a bar is elongated along the length direction. L

$$\varepsilon = \frac{\Delta h}{h}$$

Remark:
h here is commonly defined as h+Δh (i.e., original length!)



ε **is strain, equals to the fractional change in volume (or height in this problem)**

σ **is stress (‘pressure’), which is force/area**

$$\sigma = \frac{F}{A}$$

(Lowrie, 2007)

Stress and strain

Seismic energy – exerts a **stress** on the material that it passes through

– material deforms slightly (**strain**)

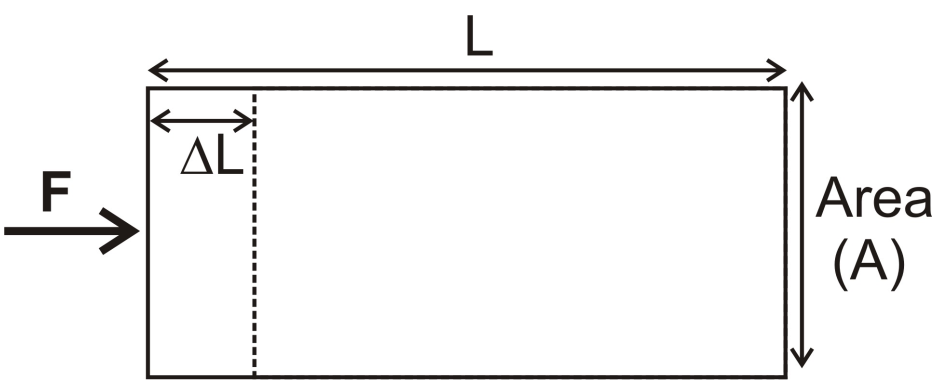
– **elastic** → once stress is removed, material returns to its original state (no permanent deformation)

stress = force per unit area (pressure) **N/m² or Pa**

strain = fractional change in shape (normalized deformation) **dimensionless**

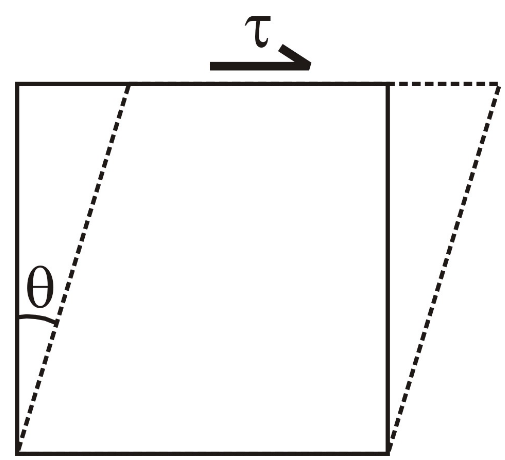
NORMAL STRESS
- force perpendicular to surface (A)

→ **longitudinal strain**: $\Delta L/L$



SHEAR STRESS
- force parallel to surface

→ **shear strain**: $\tan\theta$



Elastic deformation and Hooke's Law

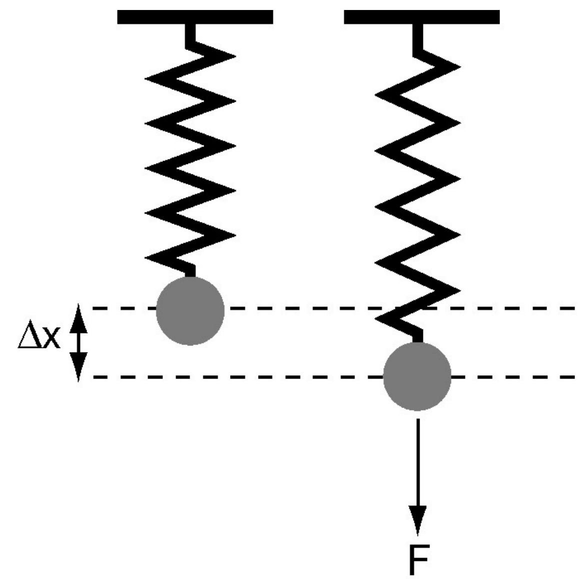
Strain is proportional to the stress that produced it (linear elasticity)

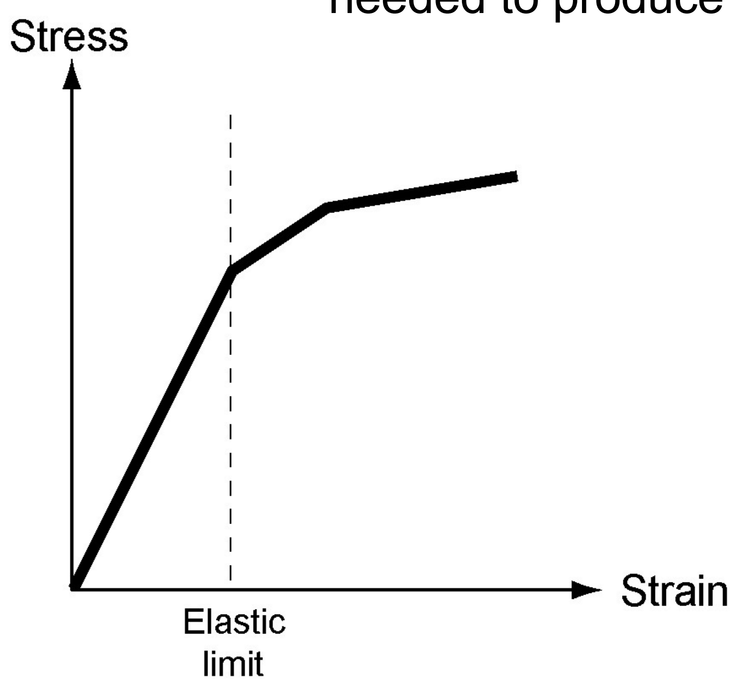
<u>For a spring:</u> $F = k\,\Delta x$

where k is the spring constant (material property)

Can rewrite as: $k = F / \Delta x$

→ k governs how much **stress** is needed to produce a given **strain**





- stress and strain are linearly related
- elastic material – strain goes to 0 when stress is removed
- above the elastic limit, deformation is permanent (plastic)
  - elastic limit for rocks is ~10^{-4}

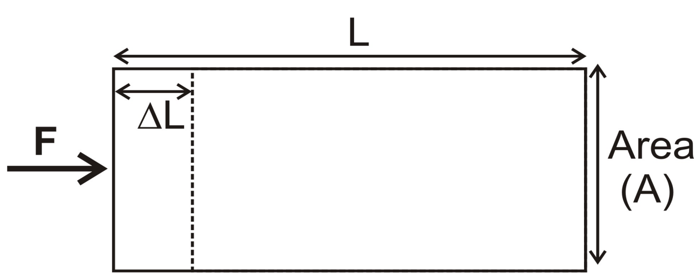
    (seismic wave strains are ~10^{-6})

Elastic moduli

For rocks: will deform when a stress is applied

• deformation (strain) is governed by the elastic moduli (material strength)

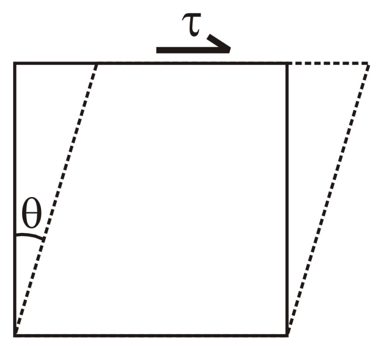
LONGITUDINAL MODULUS



• force needed to shorten by ΔL:

$$\nu = \frac{\text{long. stress } (F/A)}{\text{long. strain } (\Delta L/L)}$$

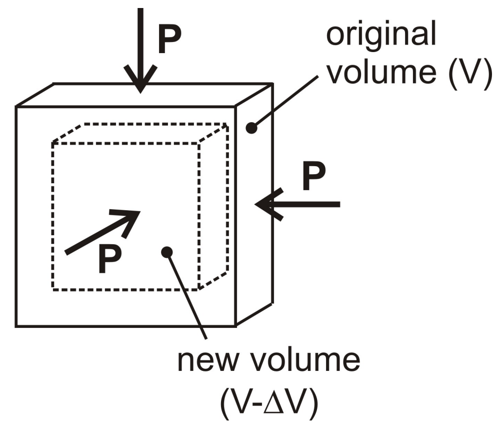
SHEAR MODULUS



• shear strain due to shearing:

$$\mu = \frac{\text{shear stress } (\tau)}{\text{shear strain } (\tan\theta)}$$

BULK MODULUS



• volume change under 3D pressure:

$$K = \frac{\text{volume stress } (P)}{\text{vol. change } (\Delta V/V)}$$

Can show: $\nu = K + \frac{4}{3}\mu$

stronger material is harder to deform → larger μ, K, and ν

Seismic waves

– disturbances in which energy is converted between elastic and kinetic

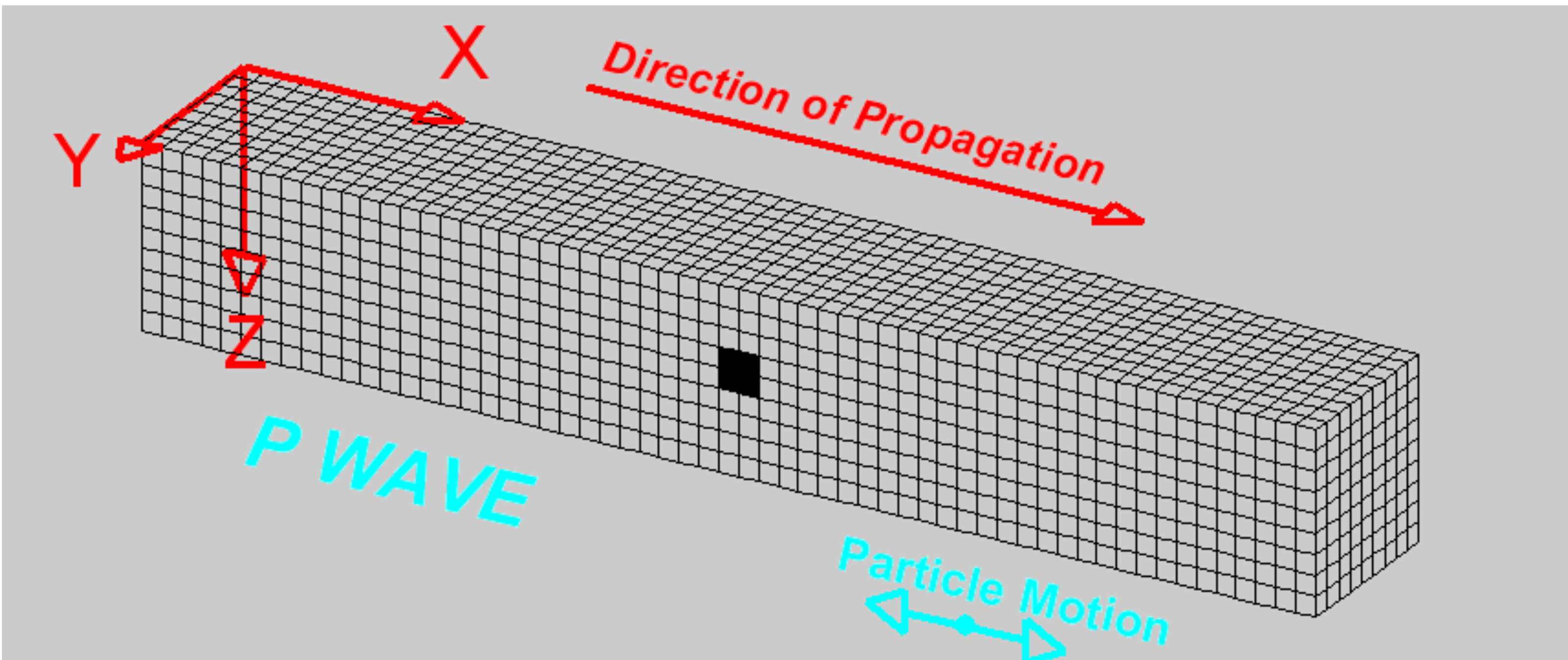
<u>TWO CATEGORIES:</u>

(1) **Body waves** – travel through the interior of the Earth

(2) **Surface waves** - travel along interfaces (e.g., the Earth's surface)

P-waves

- compressional waves (P = primary, pressure, push-pull)
- particle motion is in the direction of wave propagation
- sound waves are compressional waves



NOTE that the black square (particle) oscillates as wave passes, but returns to its original position

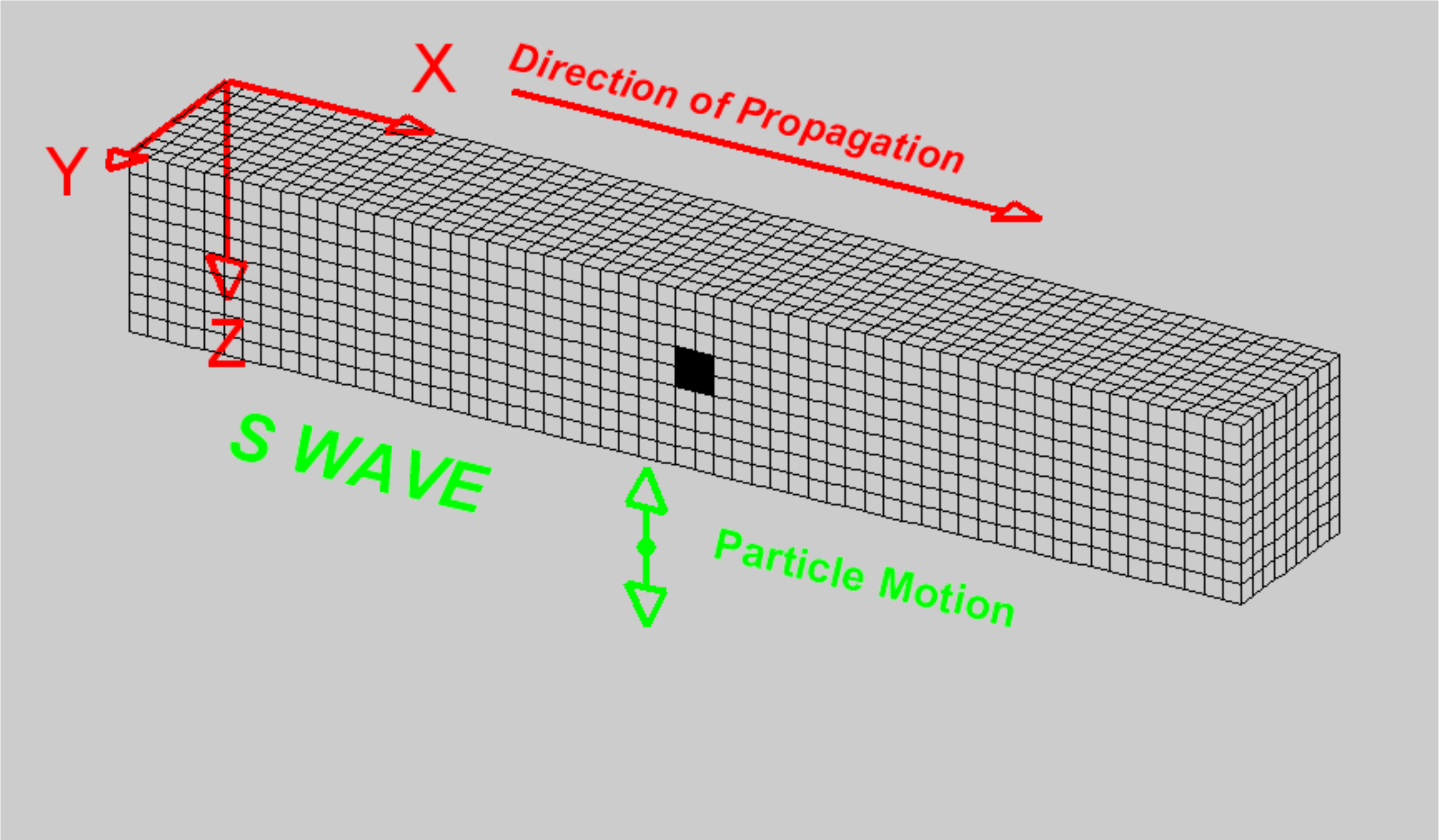
(http://web.ics.purdue.edu/~braile/edumod/waves/WaveDemo.htm)

S-waves

• shear waves (S = secondary, shear, shake….and transverse)

• particle motion is perpendicular to the direction of wave propagation

• two polarizations: SV and SH



(http://web.ics.purdue.edu/~braile/edumod/waves/WaveDemo.htm)

Seismic wave velocity

• determined by the elastic parameters and the density of the material

<u>Column of cross-sectional area A</u>

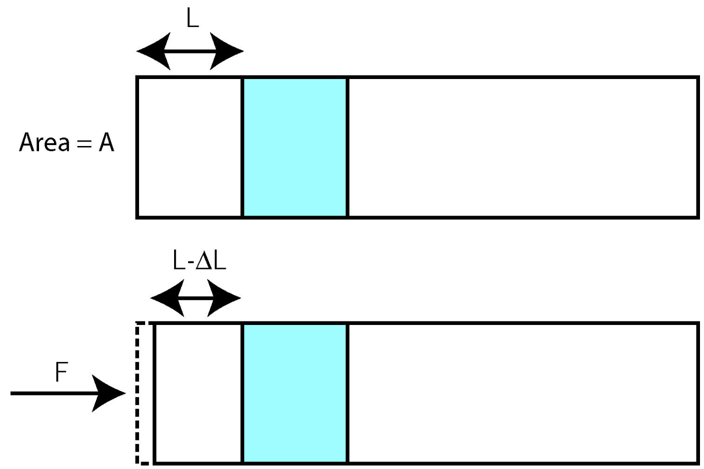
• apply force to left end

→ longitudinal stress (F/A)

• causes deformation of left side of white block

→ longitudinal strain (ΔL/L)

• stress and strain related by ψ



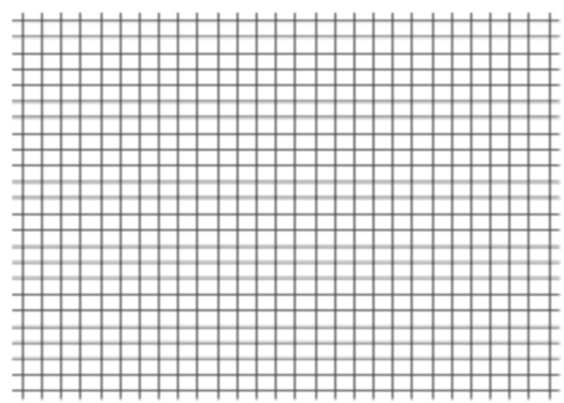
RESULTS IN:

• force exerted on shaded region

• lower stress in white region and increased stress in shaded region

• shaded region is strained – exerts stress on next block ….and so on…..

→ a **wave of deformation** travels down the block

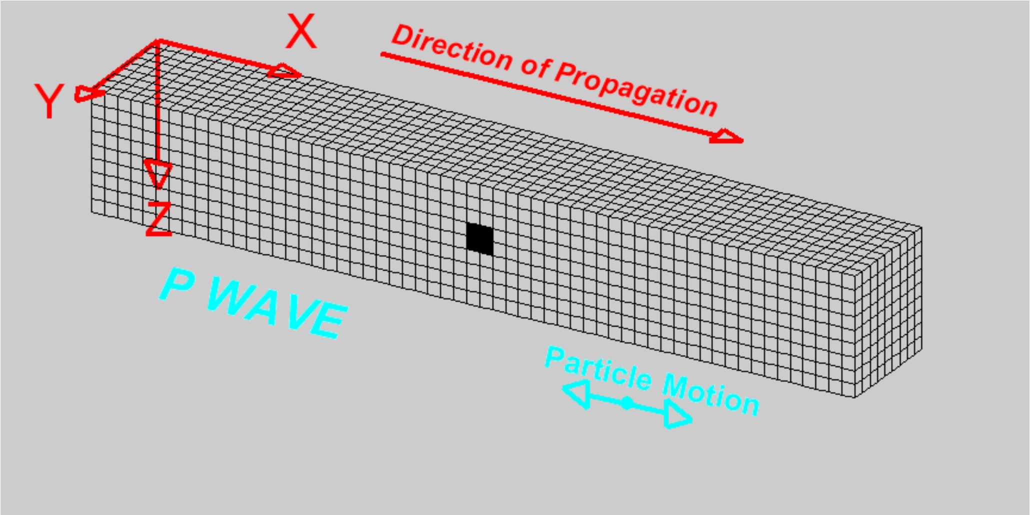


Can show that the wave velocity is:

$$V_P = \alpha = \left(\frac{\nu}{\rho}\right)^{\frac{1}{2}}$$

This is a P-wave!

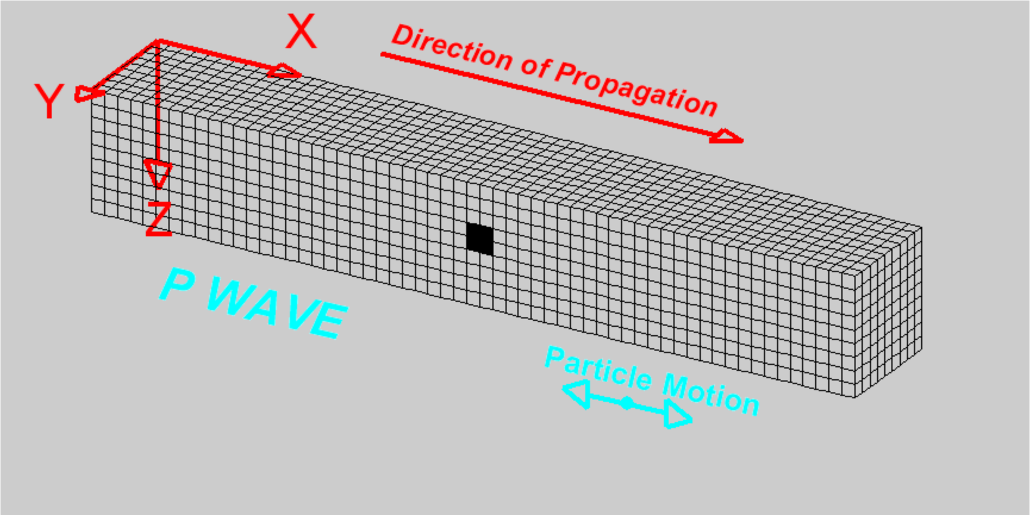
$$V_P = \left(\frac{\nu}{\rho}\right)^{\frac{1}{2}} = \left(\frac{K + \frac{4}{3}\mu}{\rho}\right)^{\frac{1}{2}}$$



ν = E = Young's modulus = Longitudinal modulus

This is a P-wave!

$$V_P = \left(\frac{\nu}{\rho}\right)^{\frac{1}{2}} = \left(\frac{K + \frac{4}{3}\mu}{\rho}\right)^{\frac{1}{2}}$$

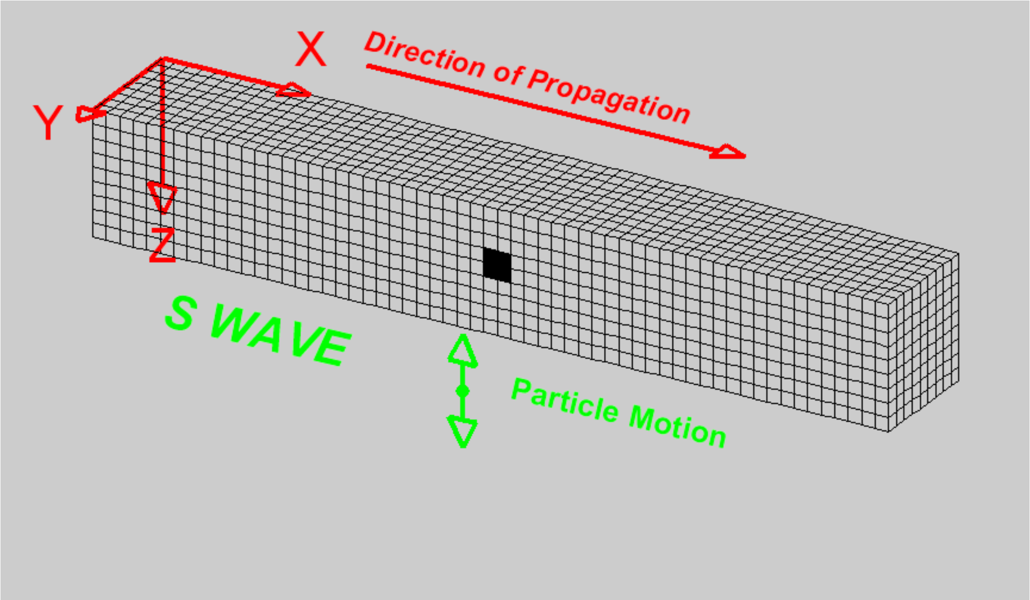


S-waves

– similar idea, but material is sheared, instead of compressed

→ shear modulus

$$V_S = \left(\frac{\mu}{\rho}\right)^{\frac{1}{2}}$$



P-waves (V_P or α)

$$V_P = \left(\frac{\nu}{\rho}\right)^{\frac{1}{2}} = \left(\frac{K + \frac{4}{3}\mu}{\rho}\right)^{\frac{1}{2}}$$

S-waves (V_S or β)

$$V_S = \left(\frac{\mu}{\rho}\right)^{\frac{1}{2}}$$

- V_P is always larger than V_S. For most crustal rocks: $V_P \approx 1.7\ V_S$
- elastic moduli are larger for stronger materials → higher V_P and V_S
- V_P and V_S do not depend on frequency → **non-dispersive**
- in fluids, $\mu = 0$ → only P-waves

Typical P-wave velocities

Dry sand	200 – 1000
Wet sand	1500 – 2000
Clay	1000 – 2500
Sandstones	
Tertiary	2000 – 2500
Carboniferous	4000 – 4500
Limestones	
Cretaceous	2000 – 2500
Jurassic	3000 – 4000
Carboniferous	5000 – 5500
Granite	5500 – 6000
Gabbro	6500 – 7000
Ultramafic rocks	7500 – 8500
Air	~350
Water	1400 – 1500
Petroleum	1300 –1400

IGNEOUS AND METAMORPHIC ROCKS

- **composition**
  mafic = higher V_P
- but mafic rocks have high ρ

$$V_P = \left(\frac{\nu}{\rho}\right)^{\frac{1}{2}} = \left(\frac{K + \frac{4}{3}\mu}{\rho}\right)^{\frac{1}{2}}$$

→ K and μ increase with ρ leading to overall increase in V_P with ρ

OTHER FACTORS

- velocity increases with increasing **pressure (depth)**

- velocity decreases with increasing **temperature**

Typical P-wave velocities

Dry sand	200 – 1000
Wet sand	1500 – 2000
Clay	1000 – 2500
Sandstones	
Tertiary	2000 – 2500
Carboniferous	4000 – 4500
Limestones	
Cretaceous	2000 – 2500
Jurassic	3000 – 4000
Carboniferous	5000 – 5500
Granite	5500 – 6000
Gabbro	6500 – 7000
Ultramafic rocks	7500 – 8500

Air	~350
Water	1400 – 1500
Petroleum	1300 –1400

SEDIMENTARY ROCKS

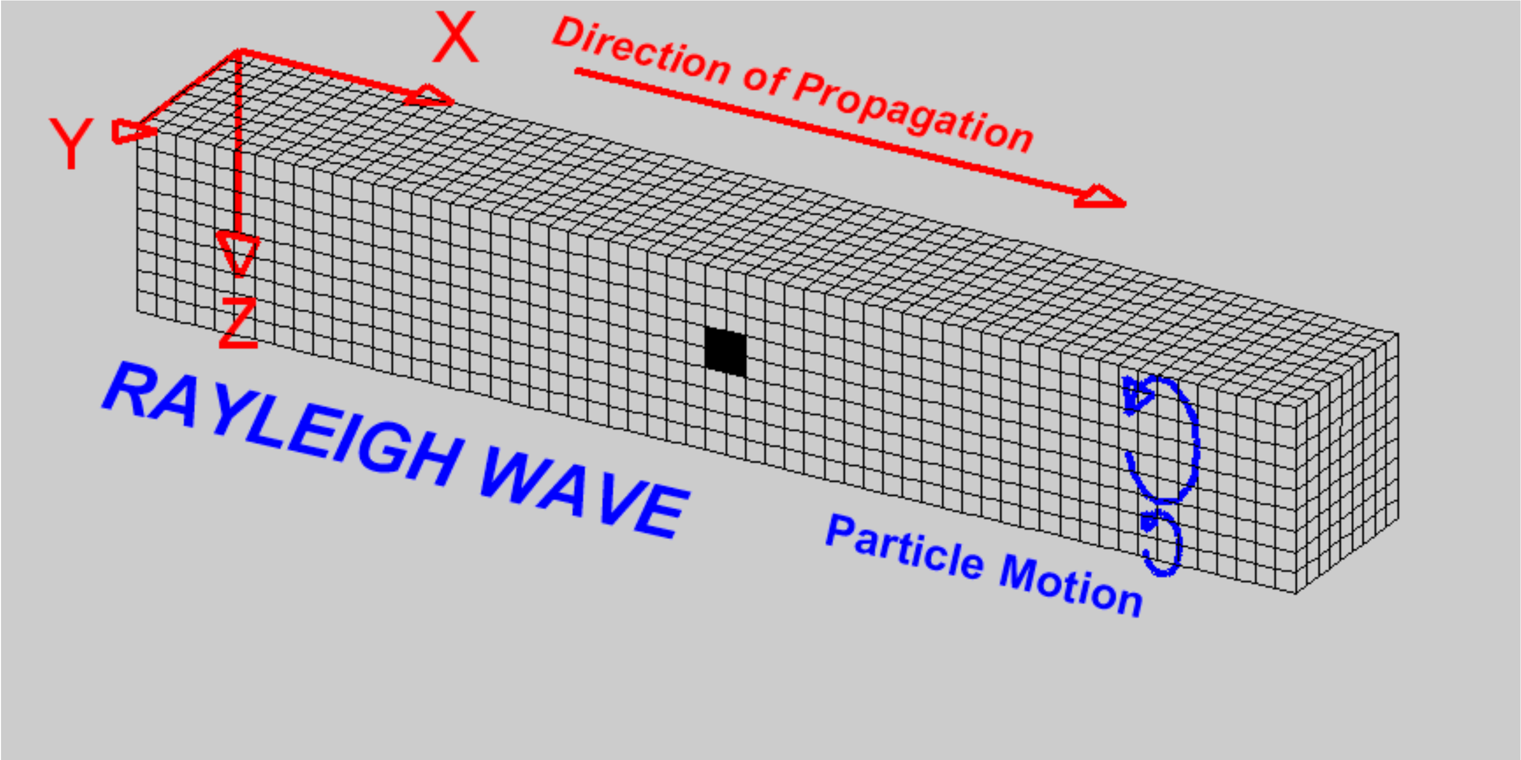
- low velocity due to **porosity**
  - → low velocity material in pores
  - → velocity decreases as porosity increases

OTHER FACTORS

- **composition** of rock matrix
- **depth / pressure** –
(1) compaction (reduces porosity)
(2) elastic moduli increase
- **age** – cementation increases rock rigidity

Surface waves: Rayleigh waves

- coupling between P-waves and SV-waves at an interface
- velocity is less than V_S
- Special case for a Poisson solid when $\nu = 0.25$, then: Vr =0.92 Vs
- elliptical retrograde particle motion ("ground roll")
- particle motion decreases exponentially with depth



(http://web.ics.purdue.edu/~braile/edumod/waves/WaveDemo.htm)

Rayleigh waves – velocity dispersion

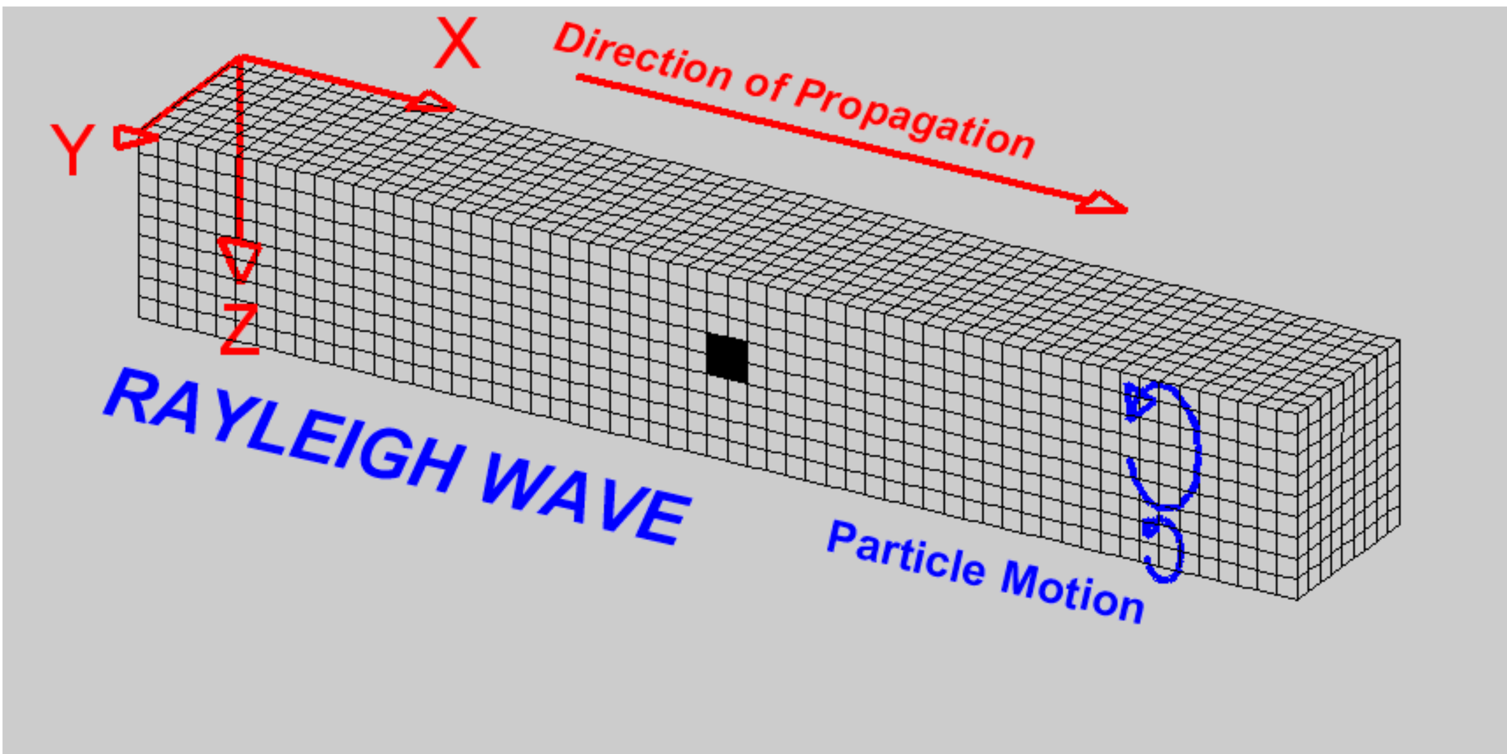
• low frequency Rayleigh wave extend to greater depth

• velocity increases with depth in Earth

• low frequency Rayleigh waves travel faster

→ **dispersion** (velocity dependent on frequency)

(like water ripples)

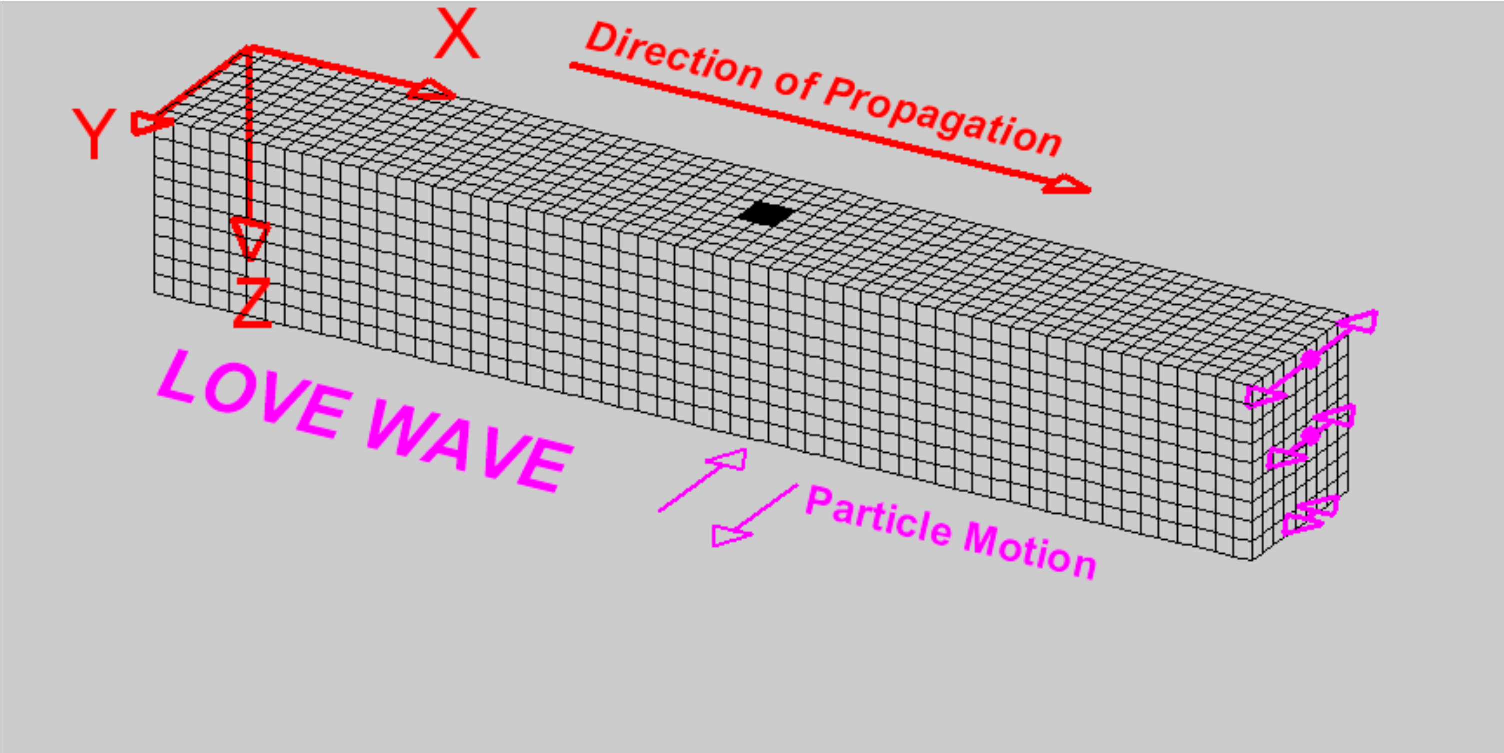


(http://web.ics.purdue.edu/~braile/edumod/waves/WaveDemo.htm)

Love waves

- occur when a near-surface layer has lower V_S than underlying material
- S-waves become trapped in near-surface layer
- horizontal particle motion (analogous to SH-waves)
- velocity is intermediate between V_S of two layers → dispersive

$$\mathbf{V}_{S_1} < \mathbf{V}_L < \mathbf{V}_{S_2}$$



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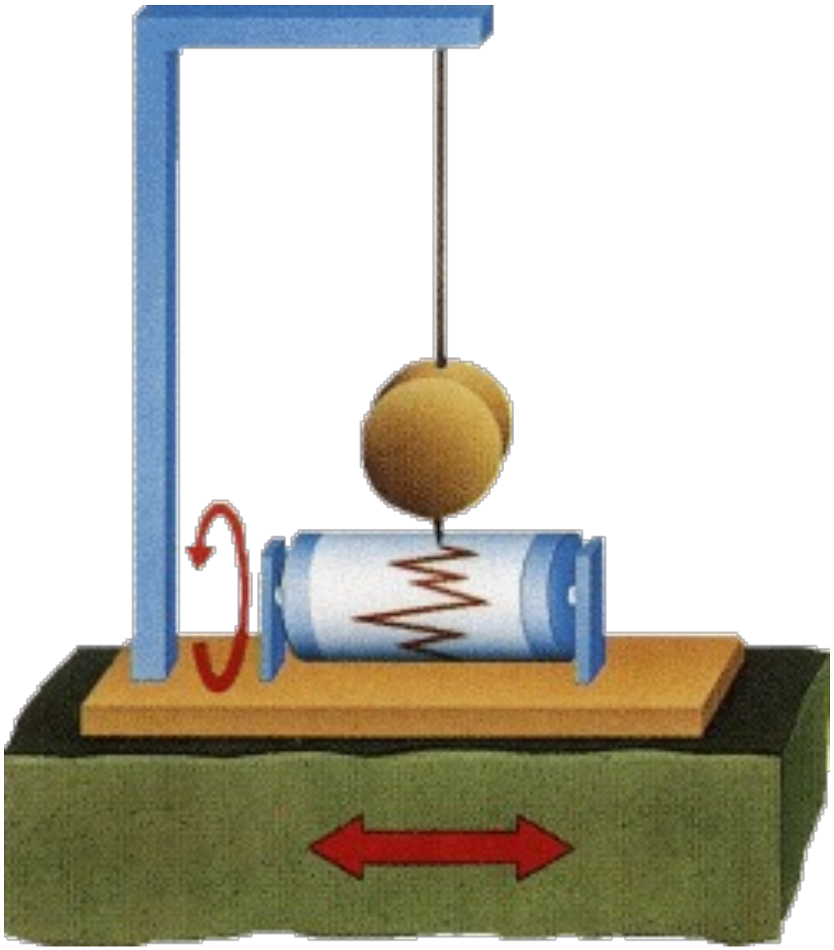
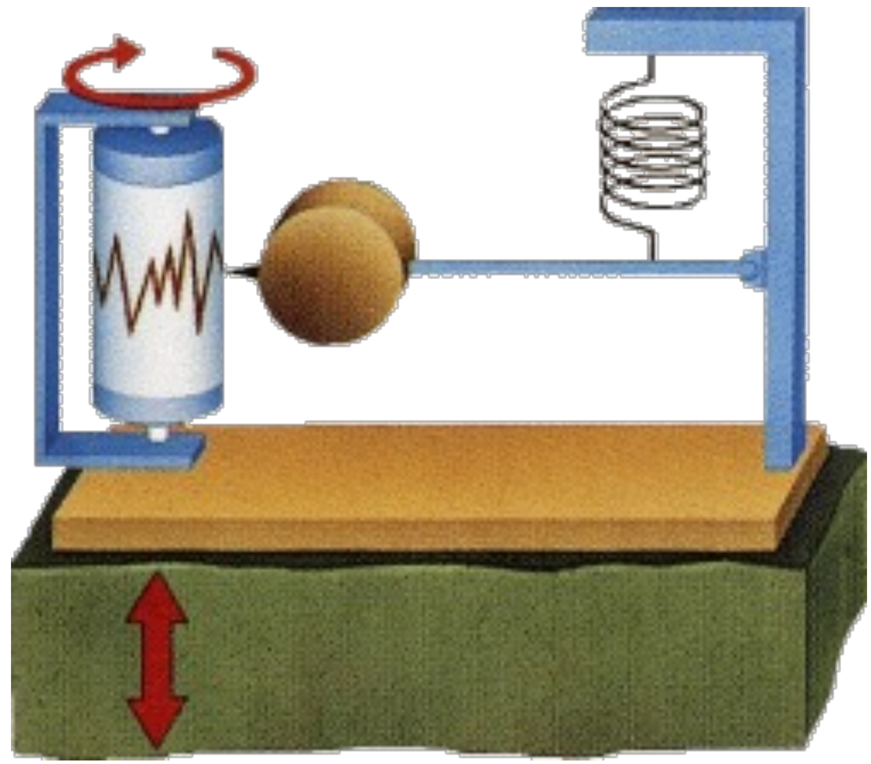
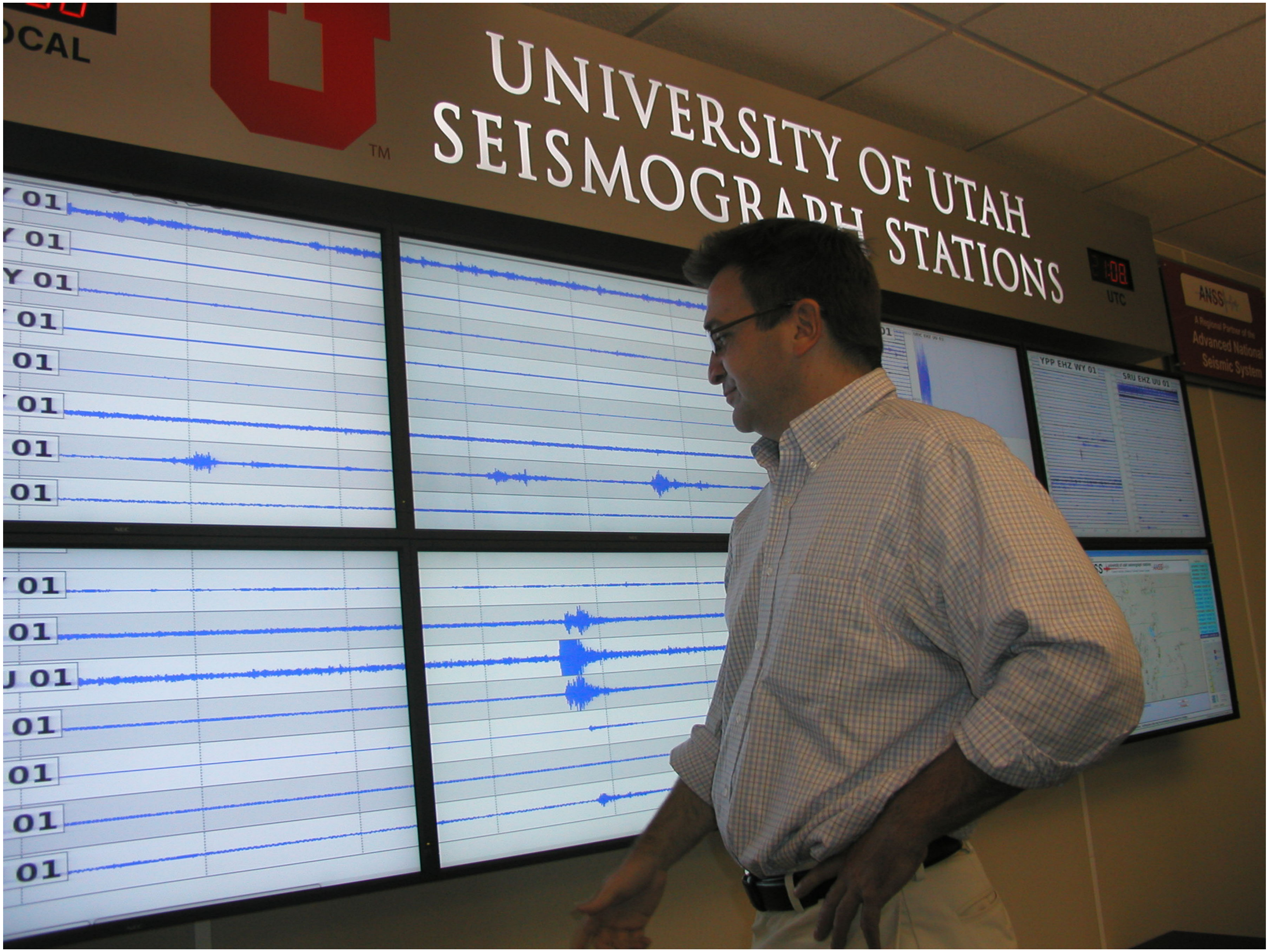


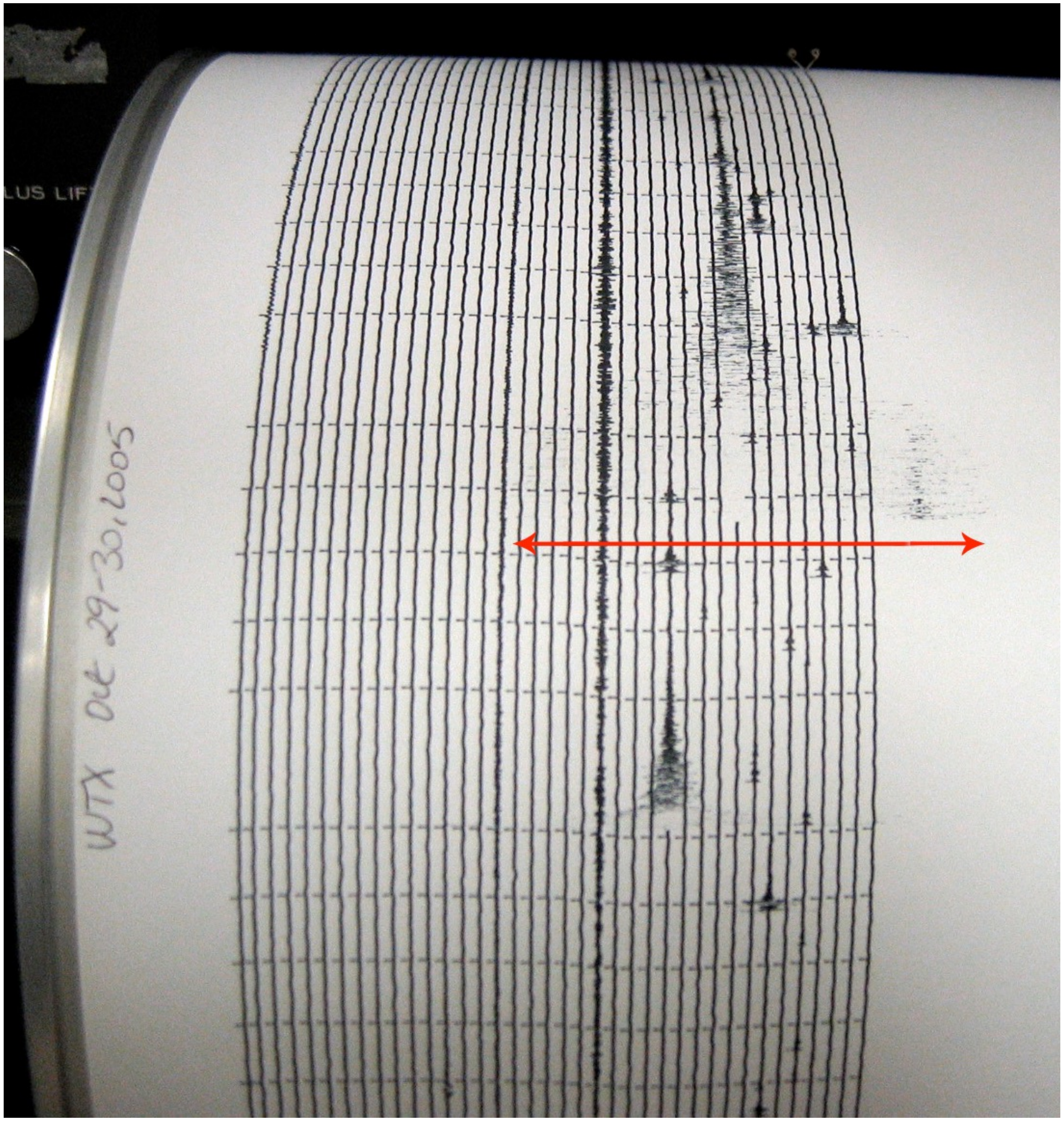


http://twisterrob.uw.hu/peq/eng/attek_figy.htm


university of utah
seismograph stations

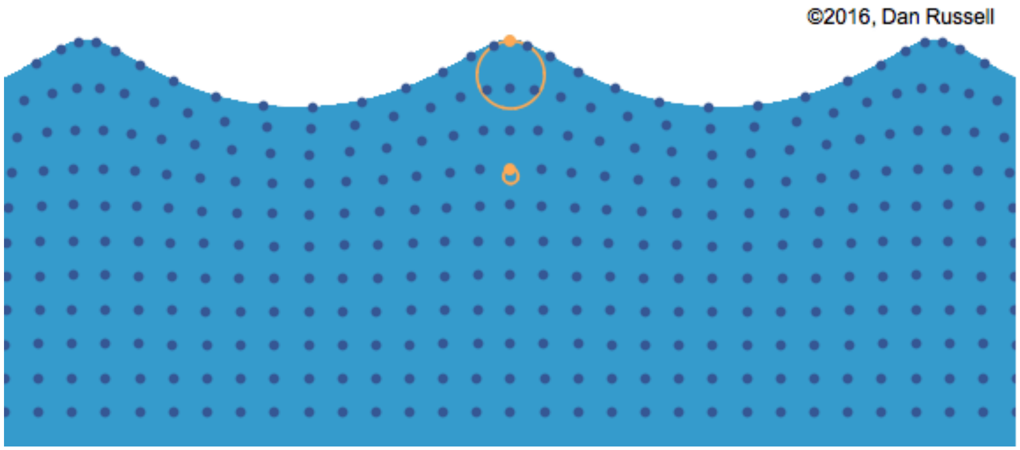


UNIVERSITY OF UTAH
SEISMOGRAPH STATIONS



http://www.ees.nmt.edu/outside/NMTSO/Images/paper_records.jpg

Water waves (opposite rotation from Rayleigh waves)



Animation from : https://www.acs.psu.edu/drussell/demos/waves/wavemotion.html

Tsunamis- triggered by rapid down drop or uplift of a section of the seafloor (especially near subduction zones).

* Long Period - 15 to 30 minutes

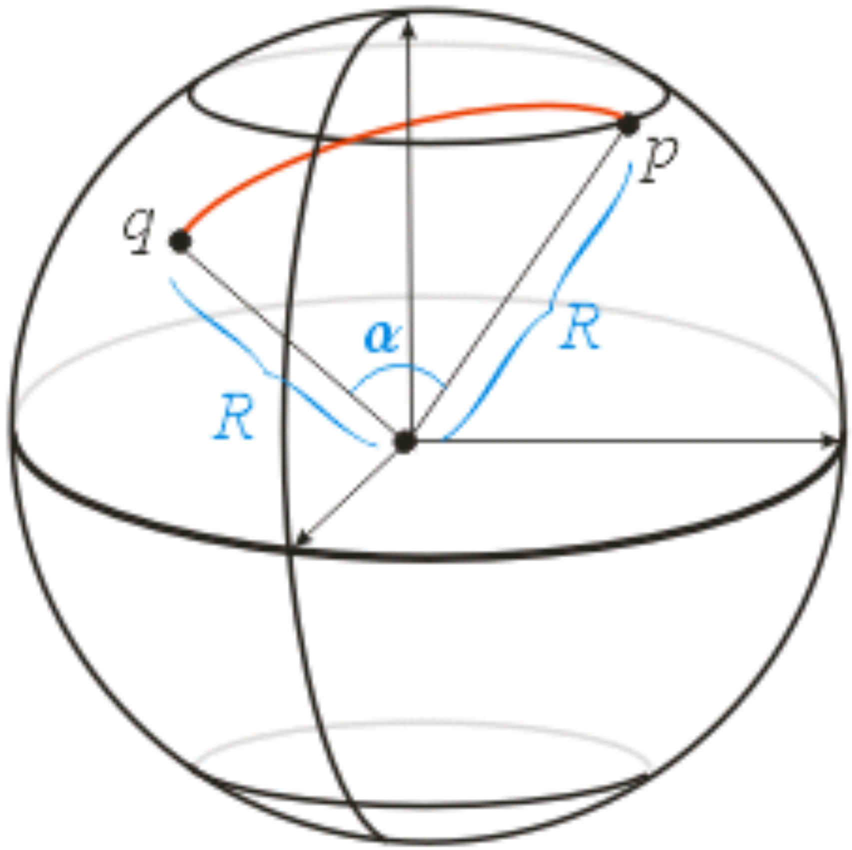
* Travels at a speed (V) (g = 9.81 m/s^2, d = water depth):

$$V = \sqrt{g * d}$$

* Deep water, wavelength ~200 km, amplitude ~few cm.

* Shallow water - much higher (20 m! in Japan1986, 7 m in Hawaii in 1964).

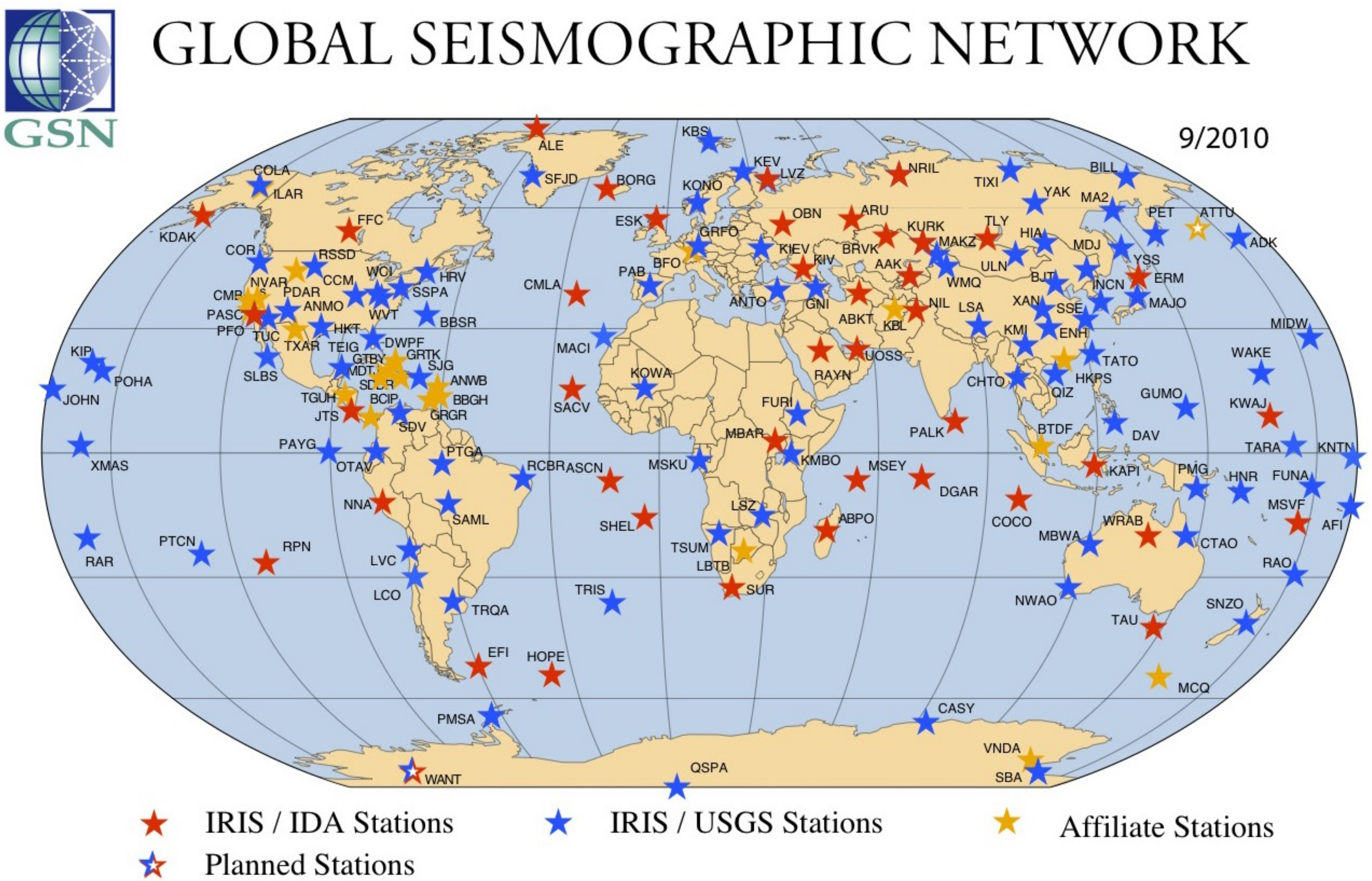
How far tsunami can travel?



The great circle distance is the portion of the circumference described by *α* hence *d*=*Rα*

Figure from https://users.cs.jmu.edu/bernstdh/web/common/lectures/slides_great-circle-distance_spherical.php

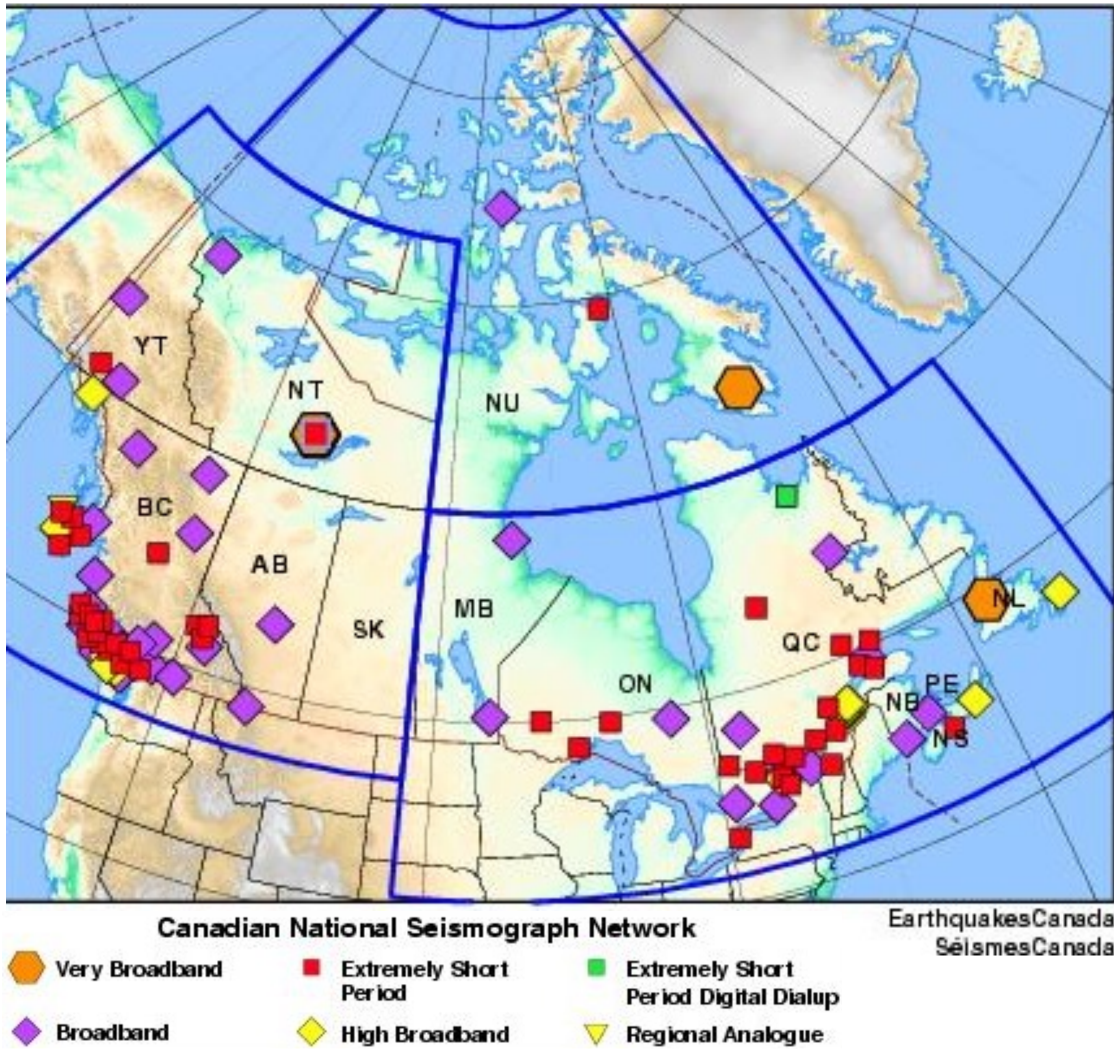
Seismograph = seismometer + recorder



This network reports data continuously to Incorporated Research Institutions in Seismology (IRIS) for data dissemination to researchers worldwide in real-time.

(www.iris.edu)

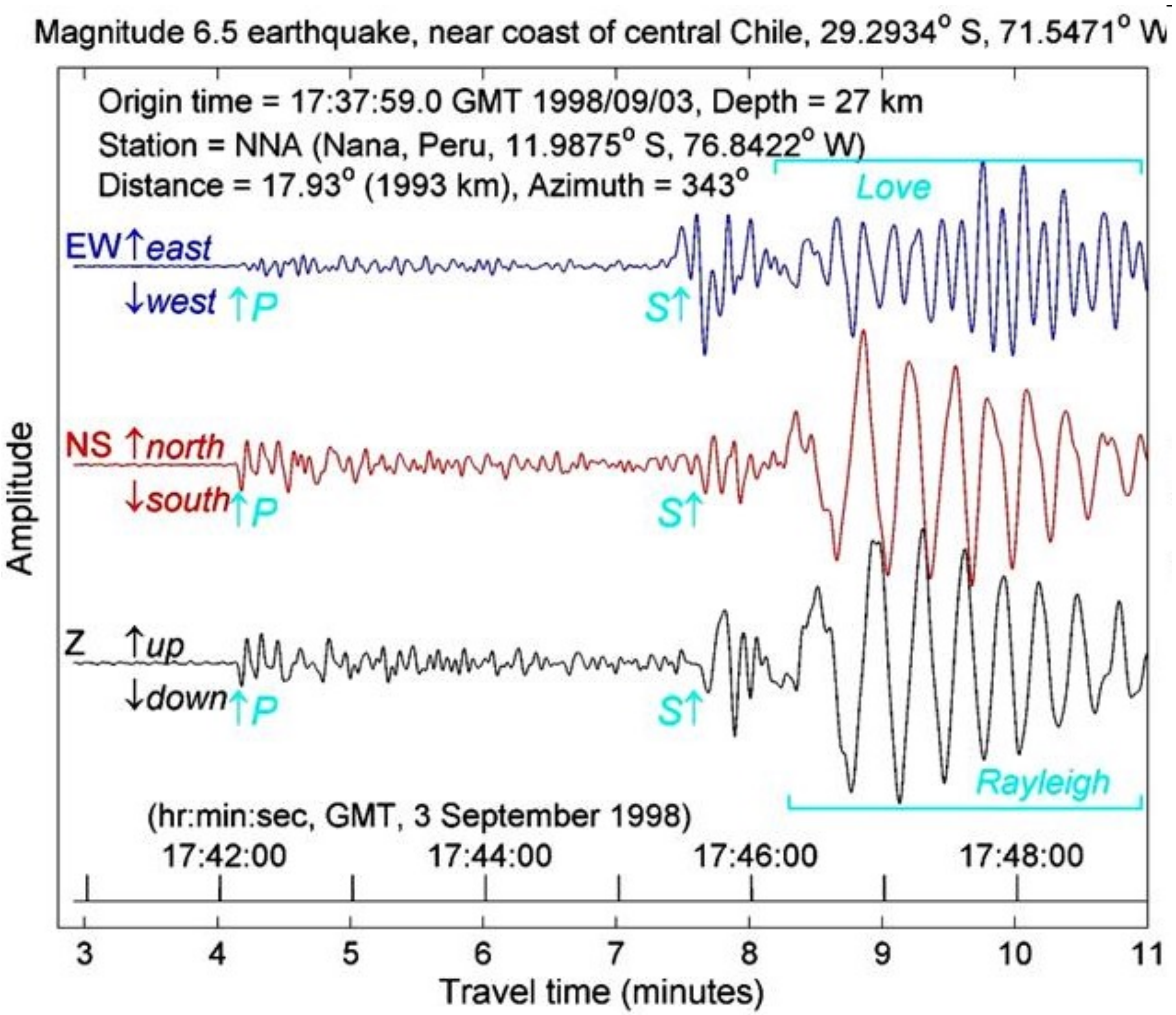
Canadian National Seismograph Network



Data from this network can be requested by researchers worldwide in real-time. It is independent from IRIS.

(earthquakescanada.nrcan.gc.ca)

Seismic recording (Seismogram)



Analysis of seismic waves:

- arrival time
- amplitude & direction of motion (polarity)

(1) Source of seismic waves

Earthquake – location, magnitude, type of fault, fault slip, etc.

(2) Structure of the Earth

- seismic waves carry information about all the material that they have travelled through from the source to the detector

(http://web.ics.purdue.edu/~braile/edumod/waves/WaveDemo.htm)