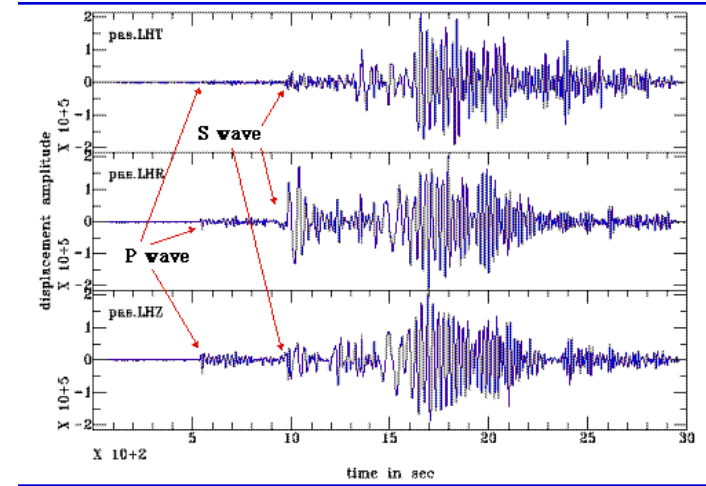
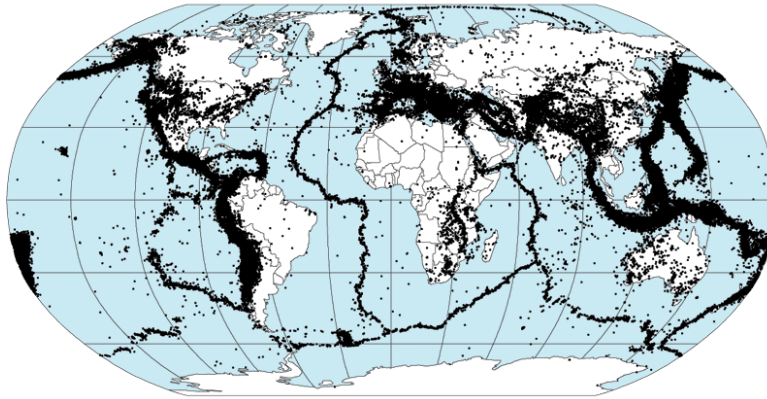
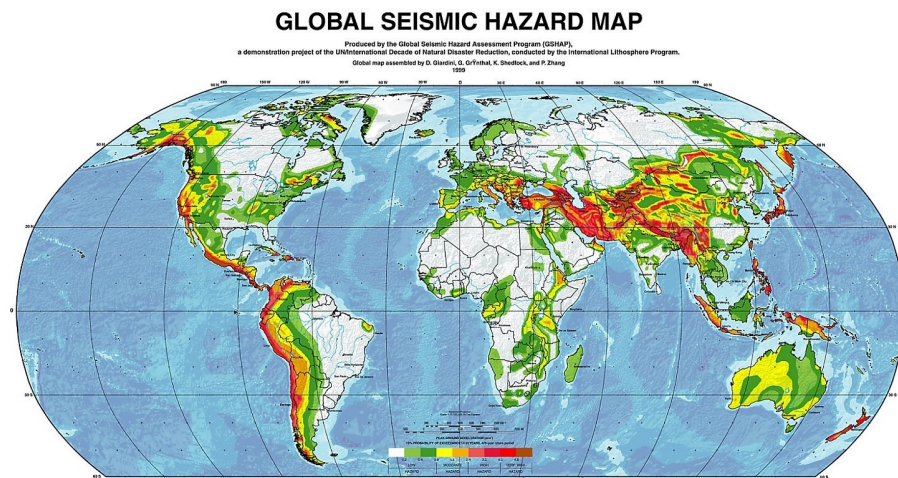
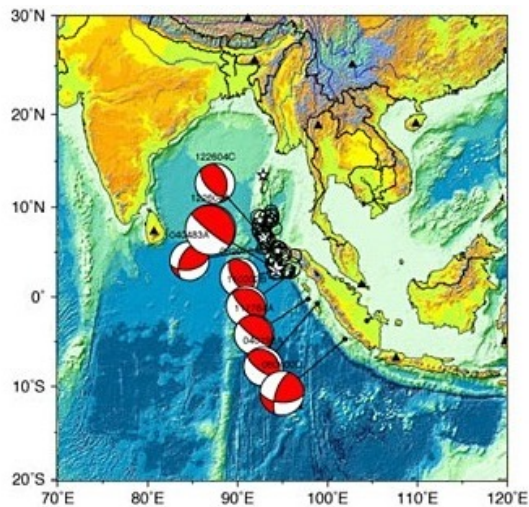


Preliminary Determination of Epicenters
358,214 Events, 1963 - 1998



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 $\times$ 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

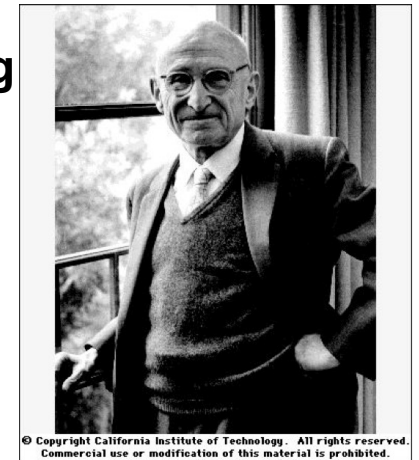
By 1909 - Existence of the Crust- Mantle Boundary by A. Mohorovicic (from Yugoslavia)

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.

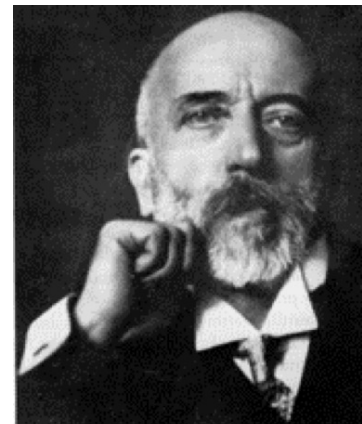
Beno Gutenberg



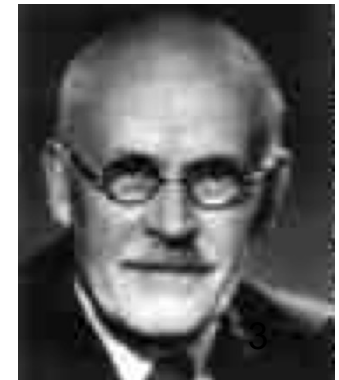
Inge Lehmann



Andrija Mohorovicic



Sir Harold Jeffreys



!1960-70: Validation of Plate Tectonics



Alfred Wagner

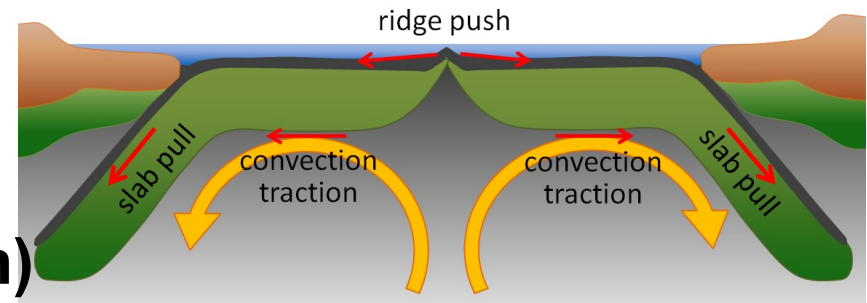
Continental Drift (1915):
One of the 4 ideas that never got
Nobel Prizes they deserved.



Tuzo Wilson (Canadian)

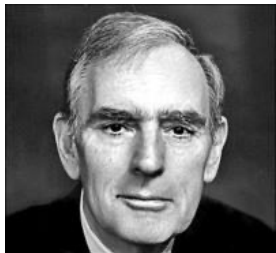
Multiple stages of ocean basin
development and closure,
Concept of Plate Tectonics (1968)

The 'Wilson cycle'



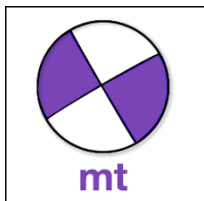
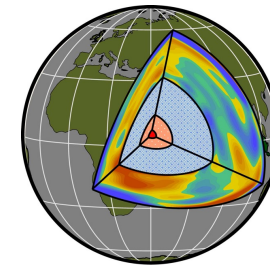
Ridge push- slab pull:
Recycling of oceanic crust and mantle

Validation of Plate Tectonics: 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)

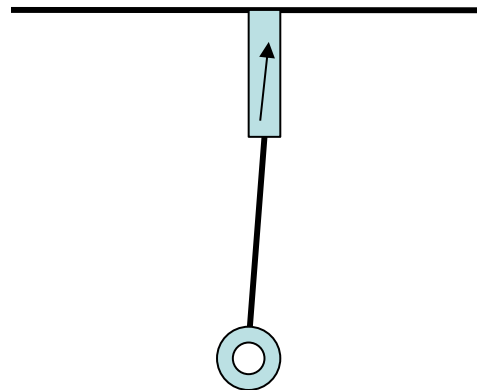


Similar idea

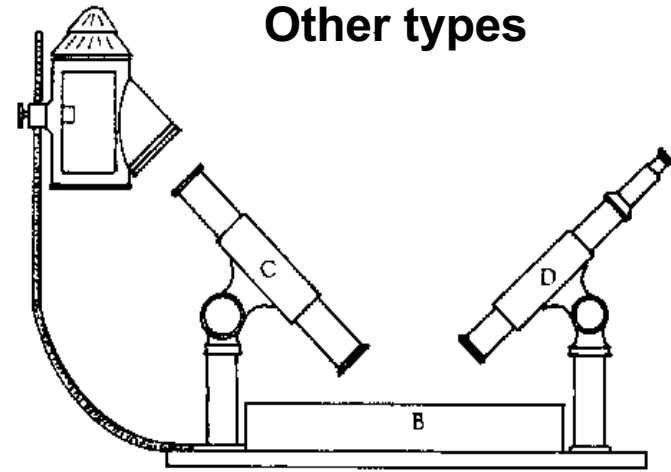
In 1700s, the invention of *spill-meter*
De la Haute Feuille



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



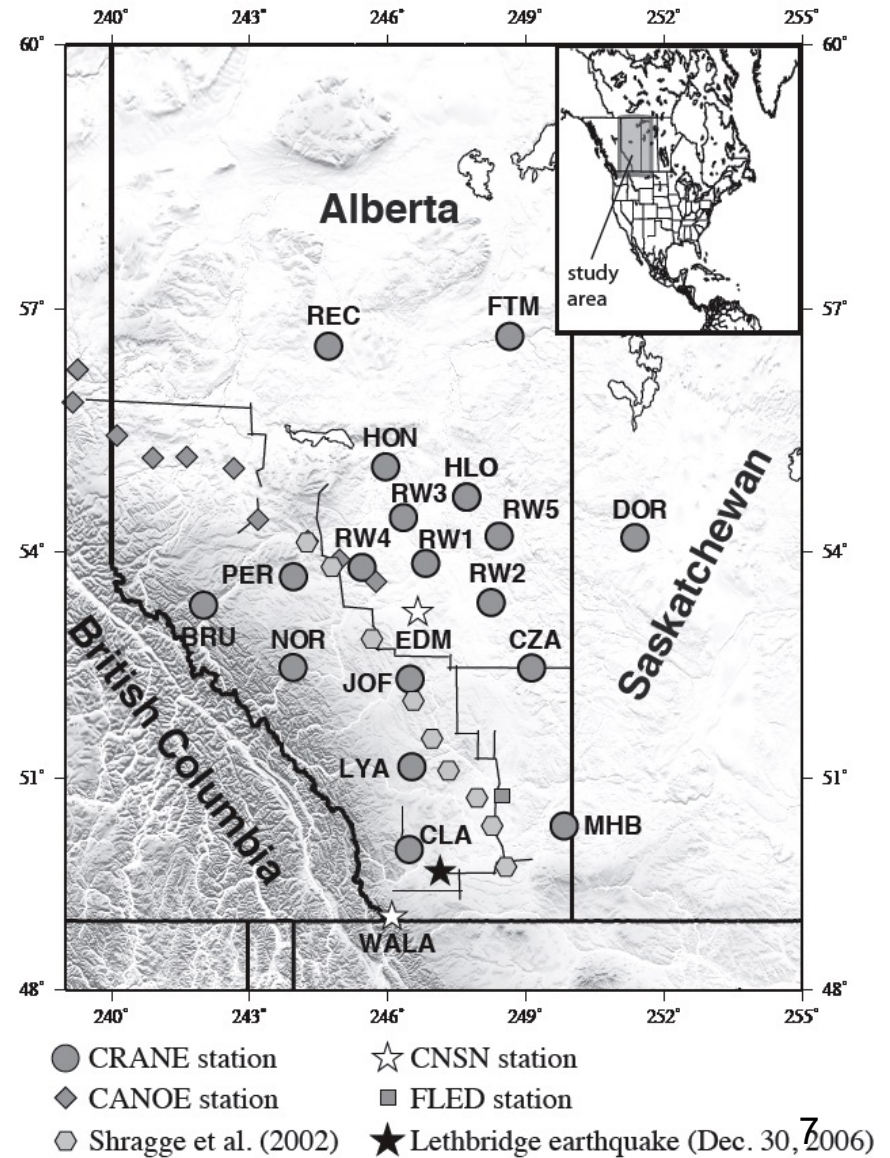
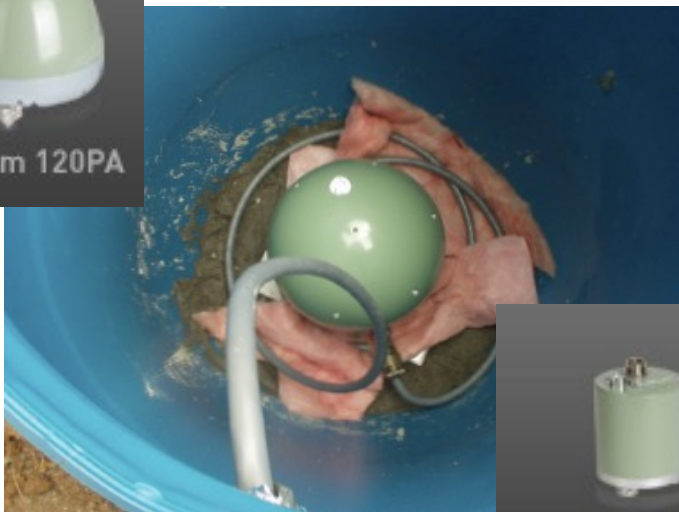
Other types

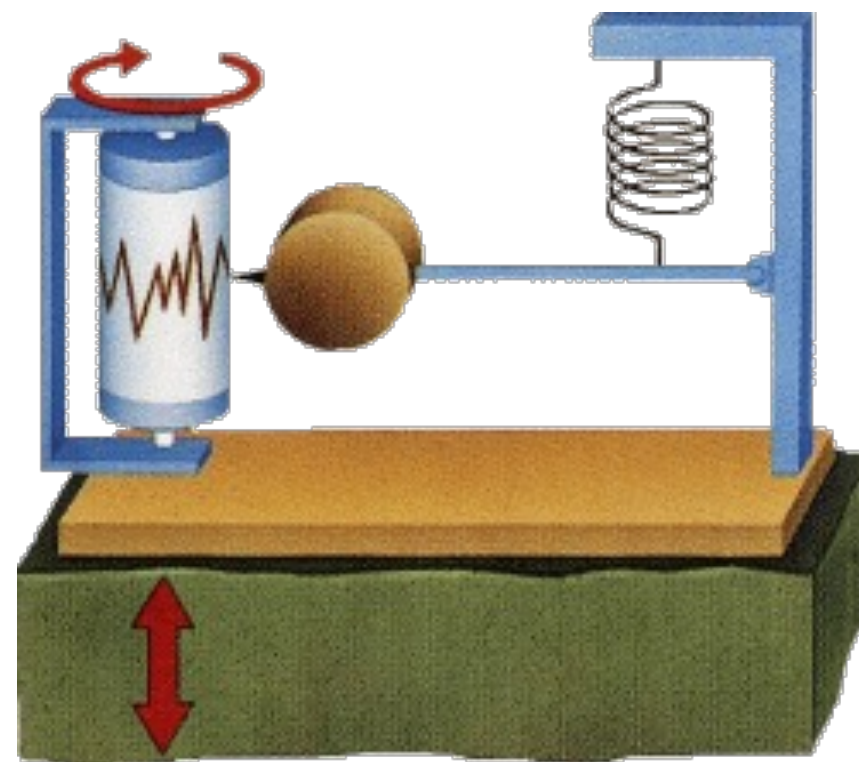
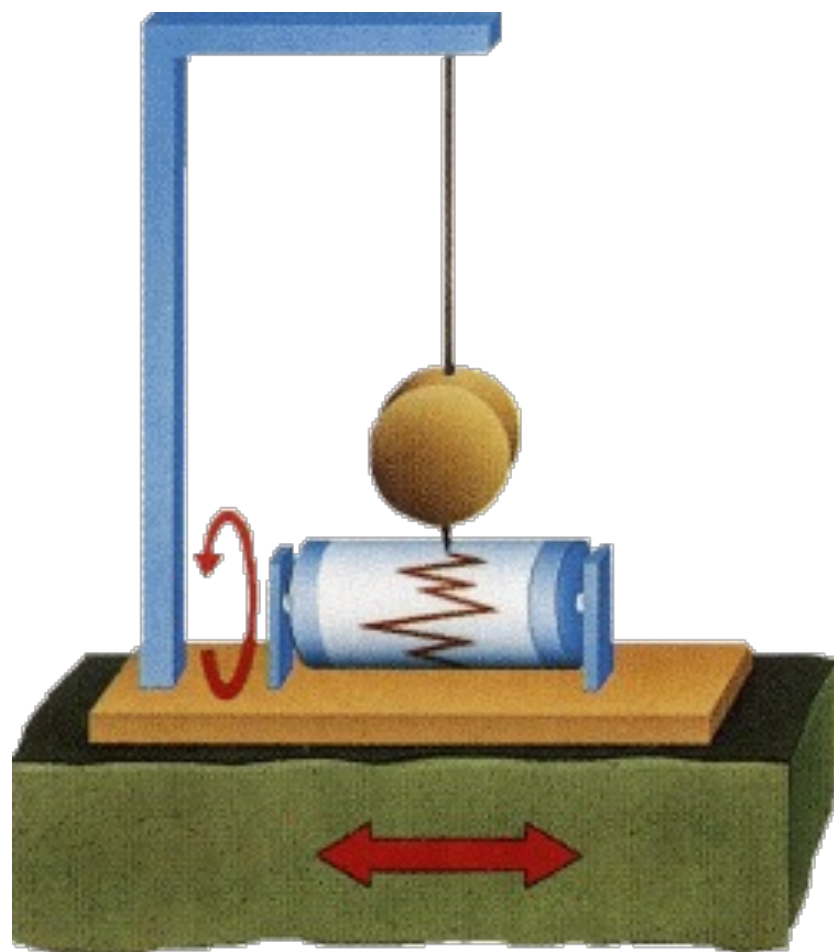


A remake of the seismoscopes of the Gu household, 1976, successfully mitigated hazards associated with the Tangshan earthquake (magnitude = 7.8, 180 km from my Beijing apartment building).



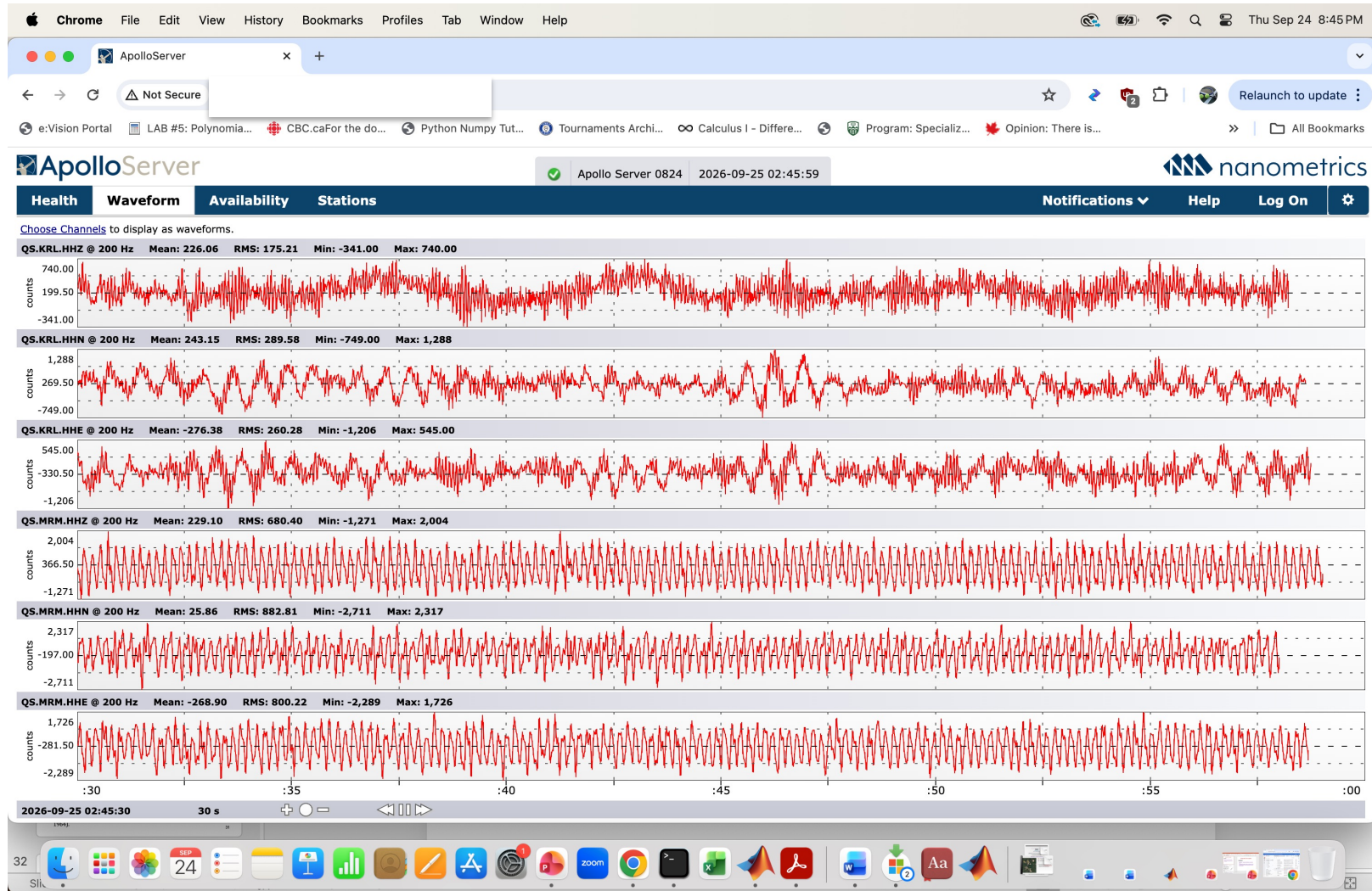
Our own deployment at UofA: (CRANE seismic array)





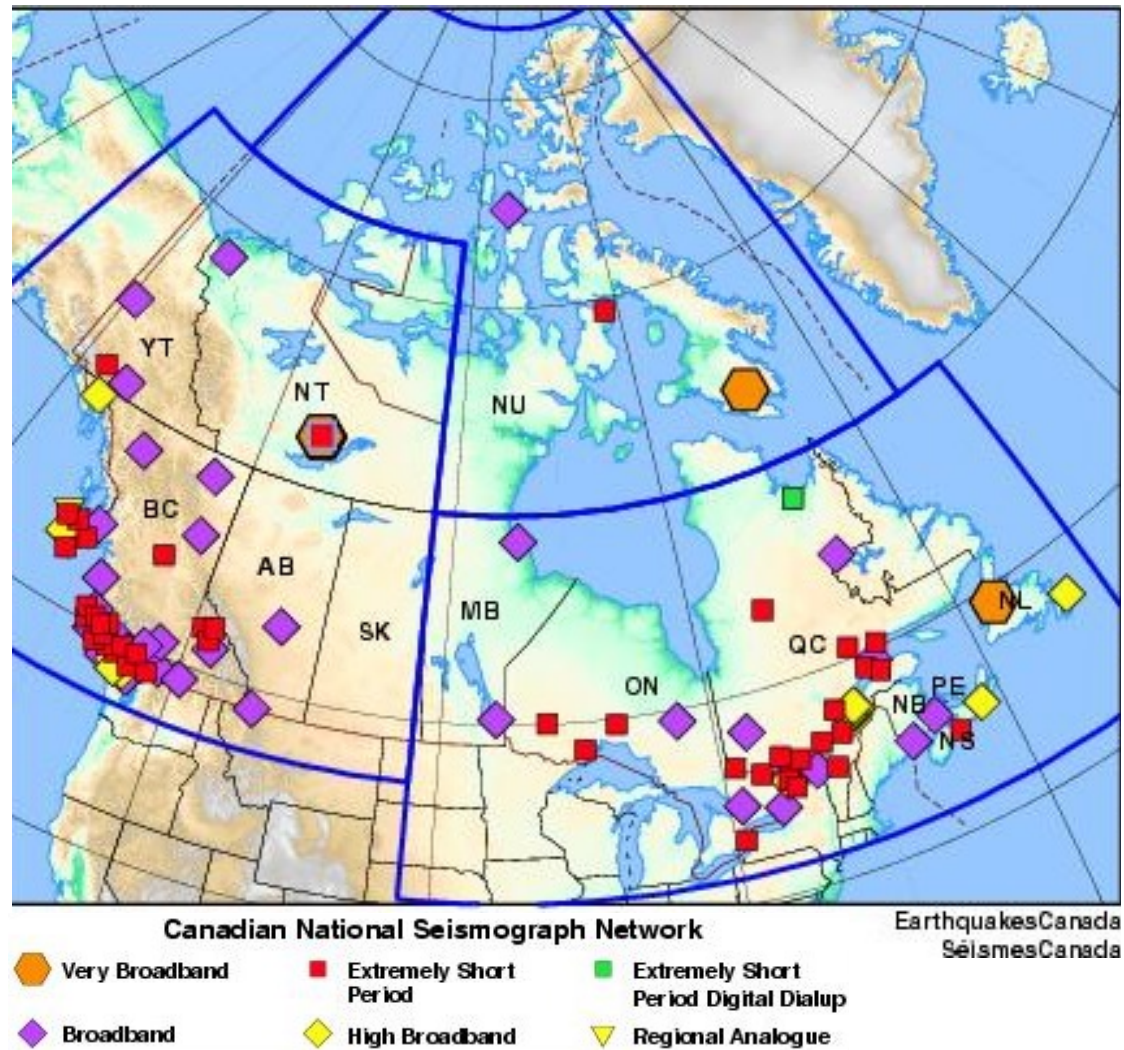


Telemetered real-time seismic data



Screenshot of QS network on Sep 24, 2026. This network is installed near Fort McMurray (University of Alberta) in August 2026, for monitoring Tailings ponds (industry wastewater 'lakes' from oilsands operations).

Canadian National Seismograph Network

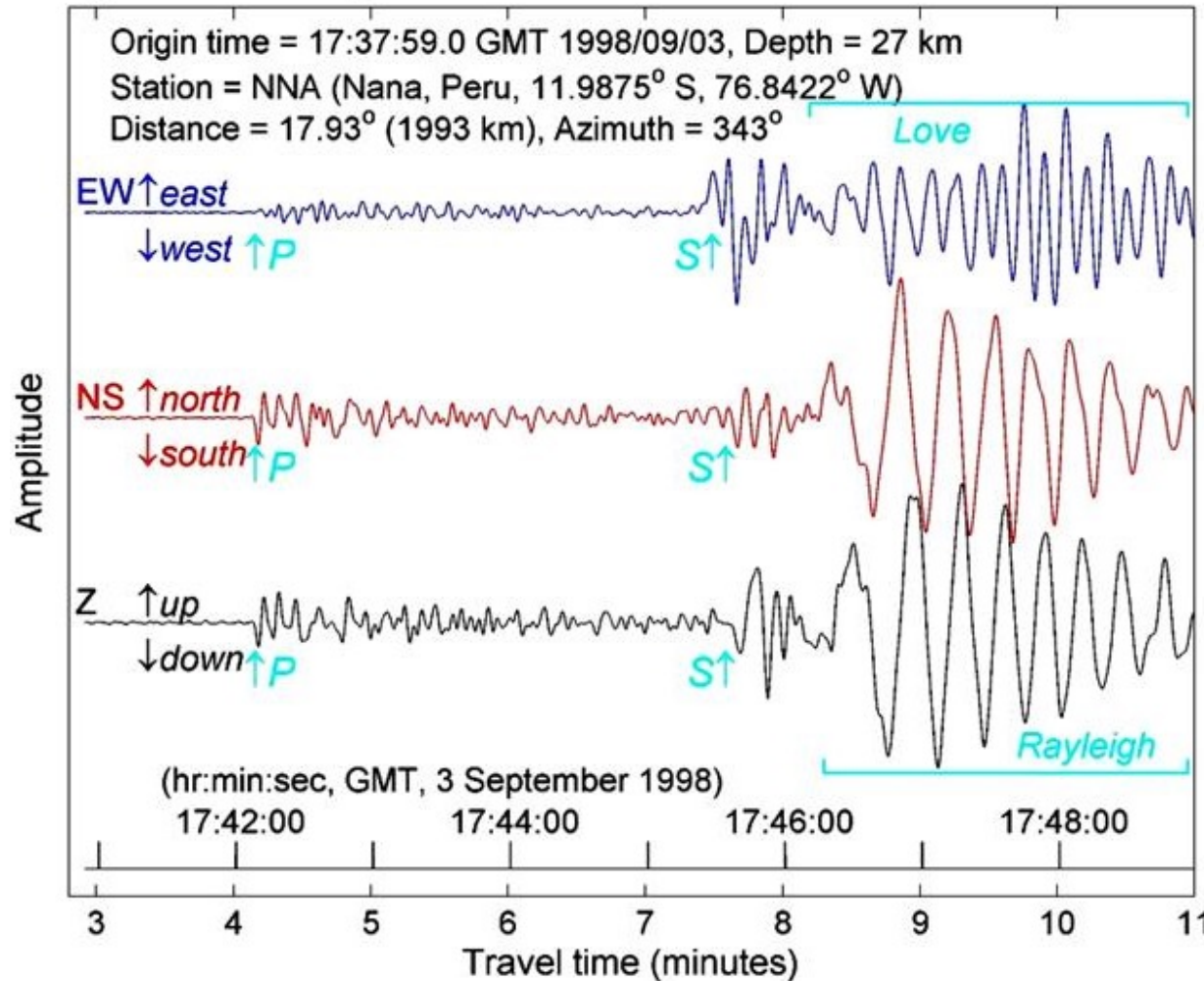


Data from this network can be requested by researchers worldwide in real-time.
It is independent from IRIS.

(earthquakescanada.nrcan.gc.ca)

Seismic recording (Seismogram)

Magnitude 6.5 earthquake, near coast of central Chile, 29.2934° S, 71.5471° W



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

Seismic waves – basic concepts

A wave:

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

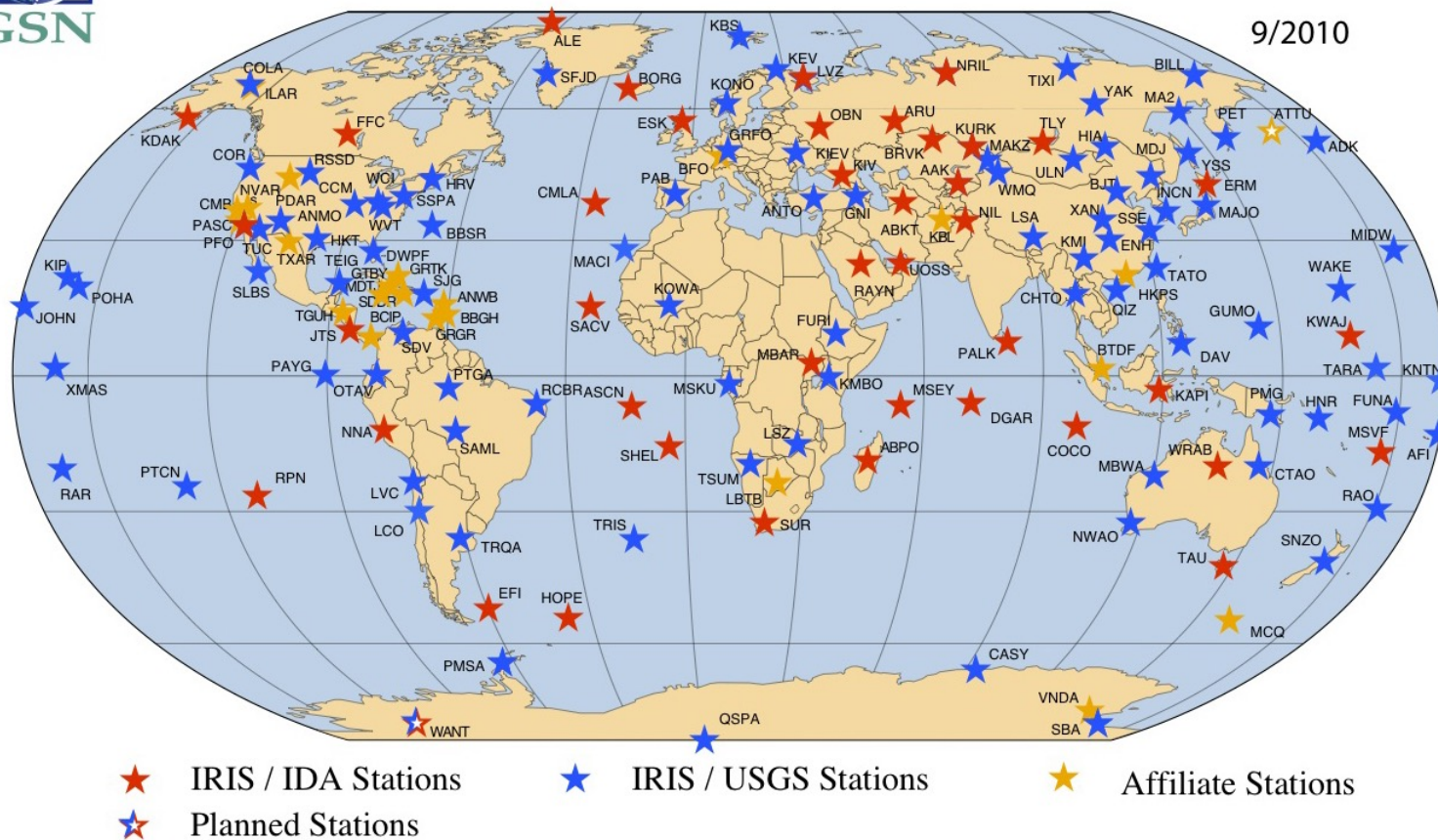
Seismic waves:

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

Seismograph = seismometer + recorder



GLOBAL SEISMOGRAPHIC NETWORK



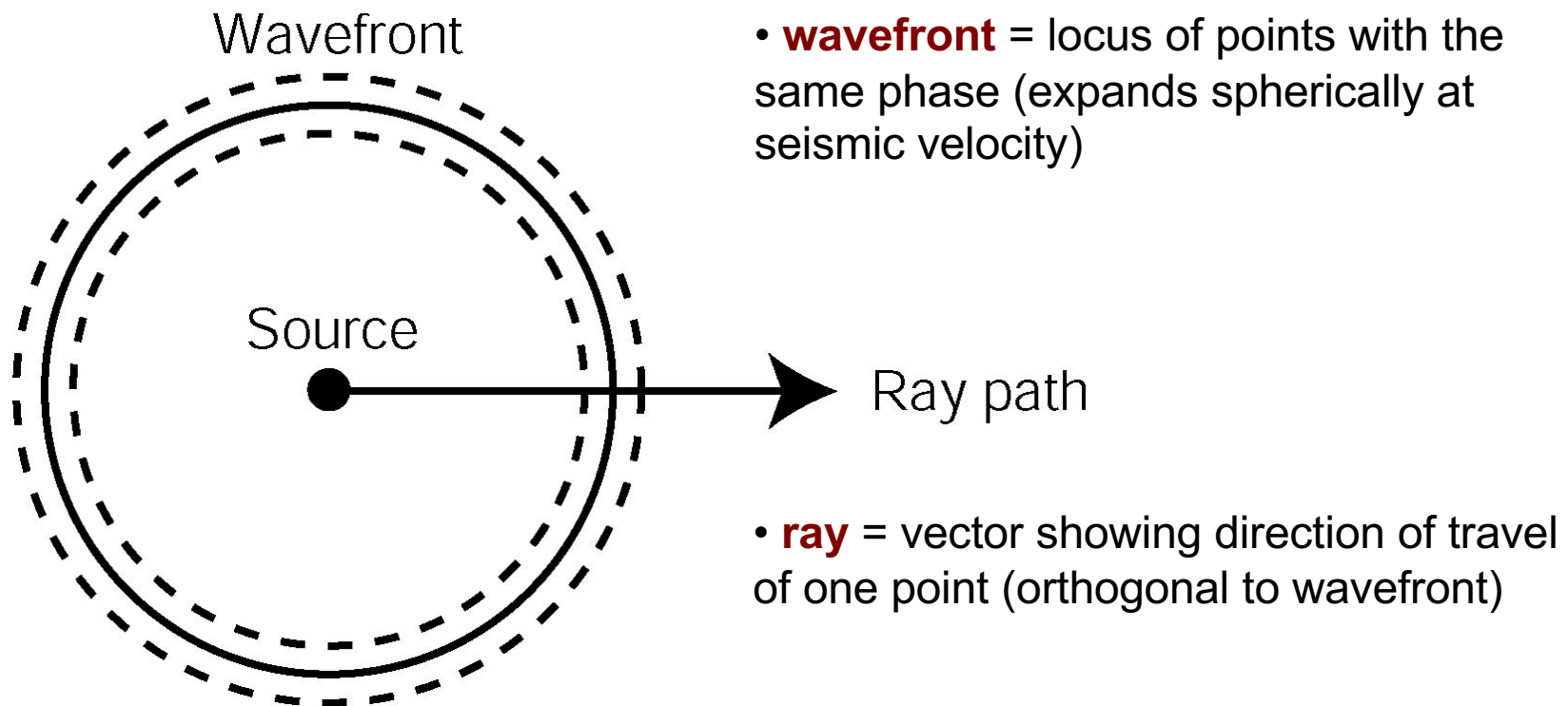
This network reports data continuously to Incorporated Research Institutions in Seismology (IRIS) for data dissemination to researchers worldwide in real-time.

(www.iris.edu)

Seismic wave propagation

Seismology – propagation of waves through the Earth

- governed by the wave equation (complex)
- can also be addressed through **visualization**



Plane wave assumption: when the ripples grow to large sizes (away from source, the curvature becomes small and looks planar,

Seismic wave is a 'wave' after all!

v = velocity (m/s)

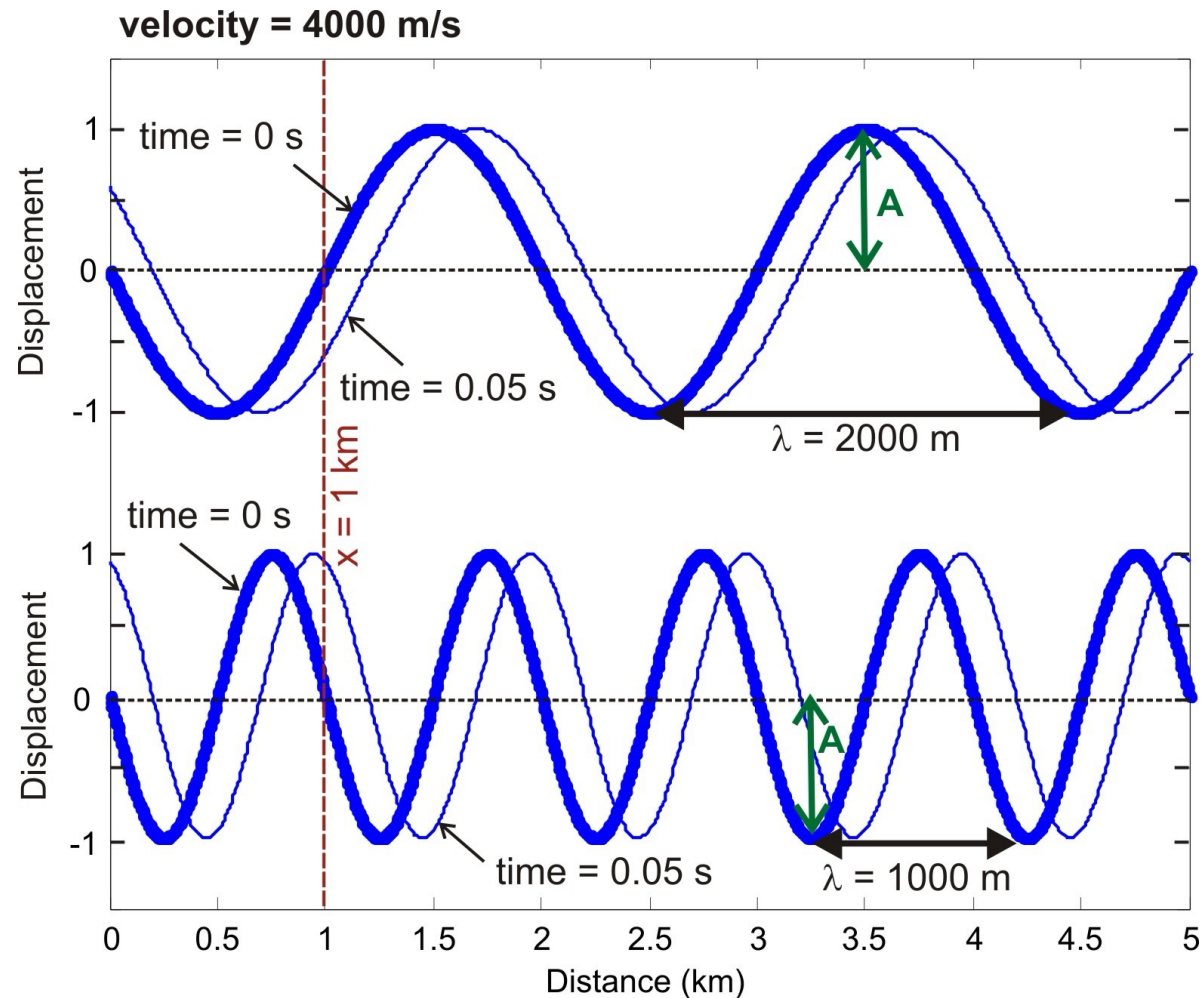
- speed at which the wave travels

A = amplitude (m)

- maximum displacement from rest position

λ = wavelength (m)

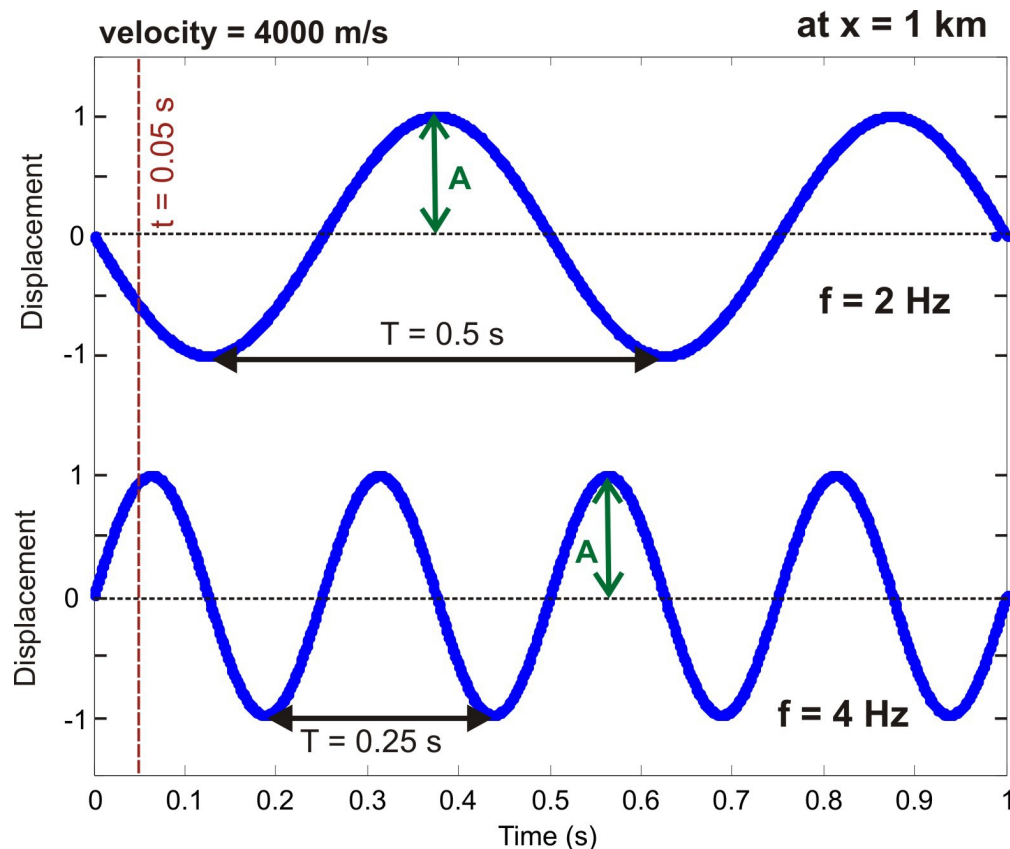
- distance between two points with same phase (e.g., peaks, troughs)



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})**

$v = d/t$ & $v = f\lambda$

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

$k = \omega/v$

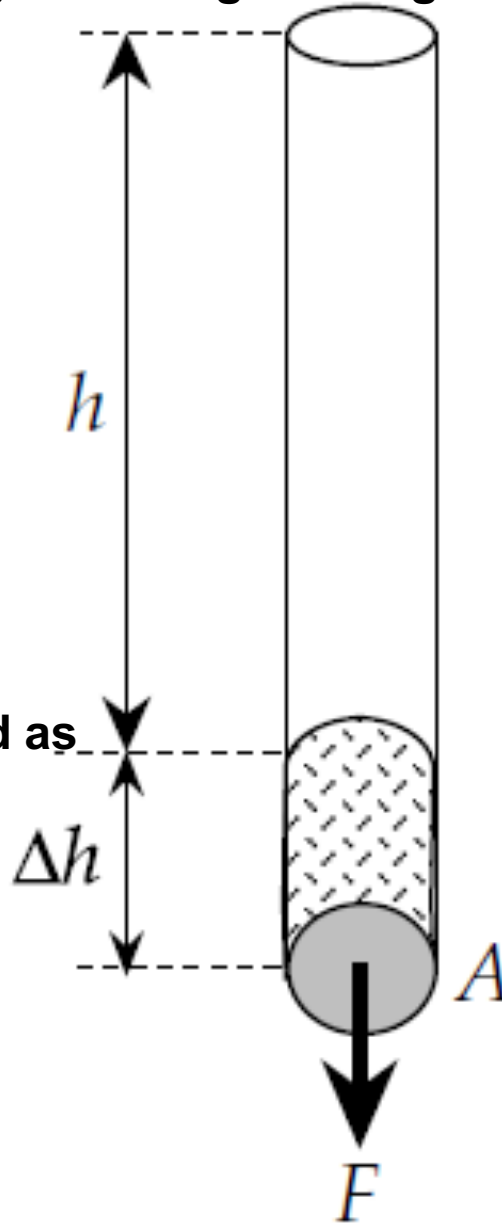
Imagine that a bar is elongated along the length direction. L

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

Remark:

h here is commonly defined as $h + \Delta h$ (i.e., original length!)

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



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

σ is stress ('pressure'), which is force/area

(Lowrie, 2007)

Stress and strain

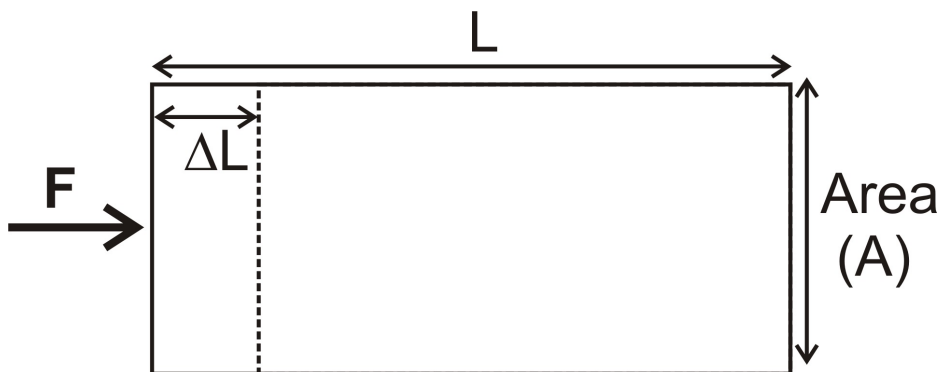
- Seismic energy** – exerts a **stress** on the material that it passes through
- material deforms slightly (**strain**)
 - **elastic** → once stress is removed, material returns to its original state (no permanent deformation)

stress = force per unit area (pressure) **N/m^2 or Pa**

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

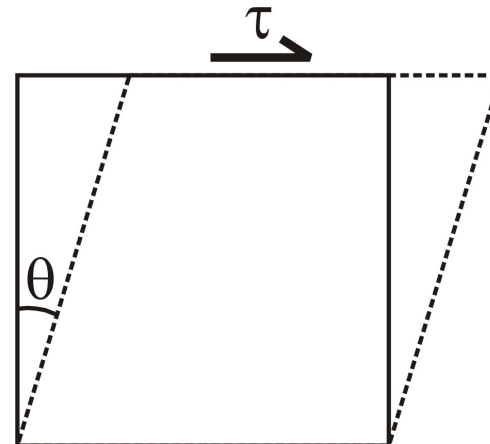
NORMAL STRESS

- 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

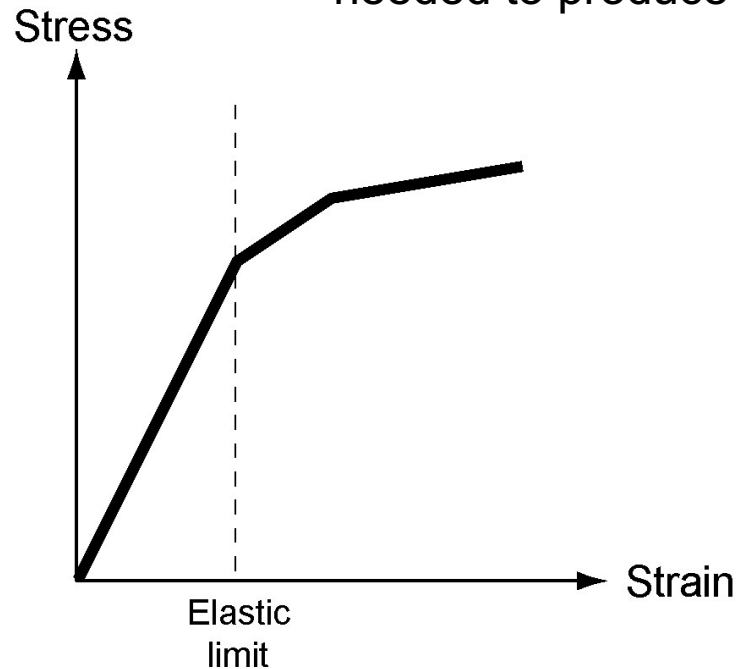
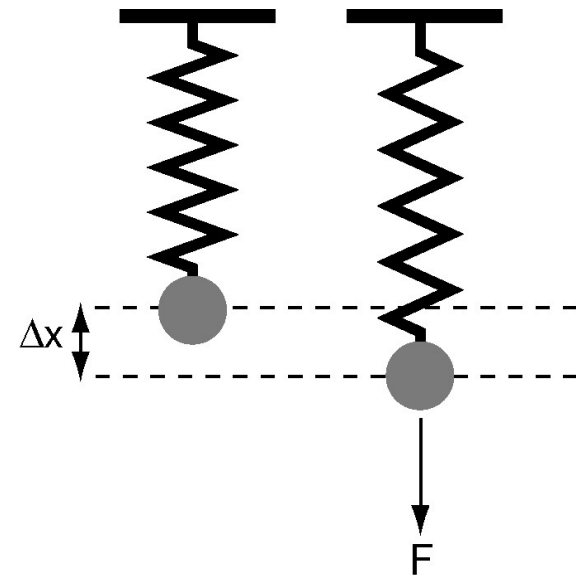
Strain is proportional to the stress that produced it (**linear elasticity**)

For a spring: $F = k \Delta x$

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 $\sim 10^{-4}$

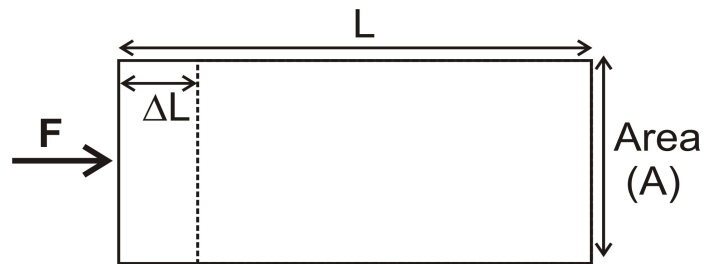
(seismic wave strains are $\sim 10^{-6}$)

Elastic moduli

For rocks: will deform when a stress is applied

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

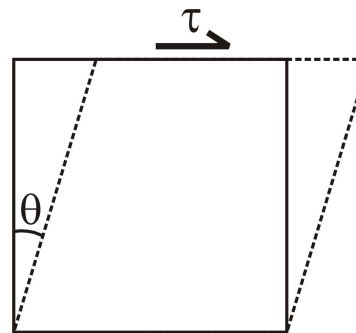
LONGITUDINAL MODULUS



- force needed to shorten by ΔL :

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

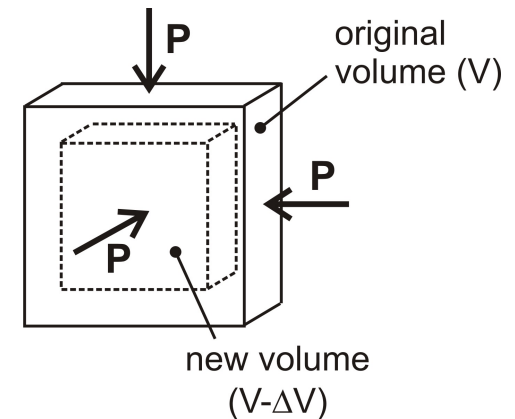
SHEAR MODULUS



- 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

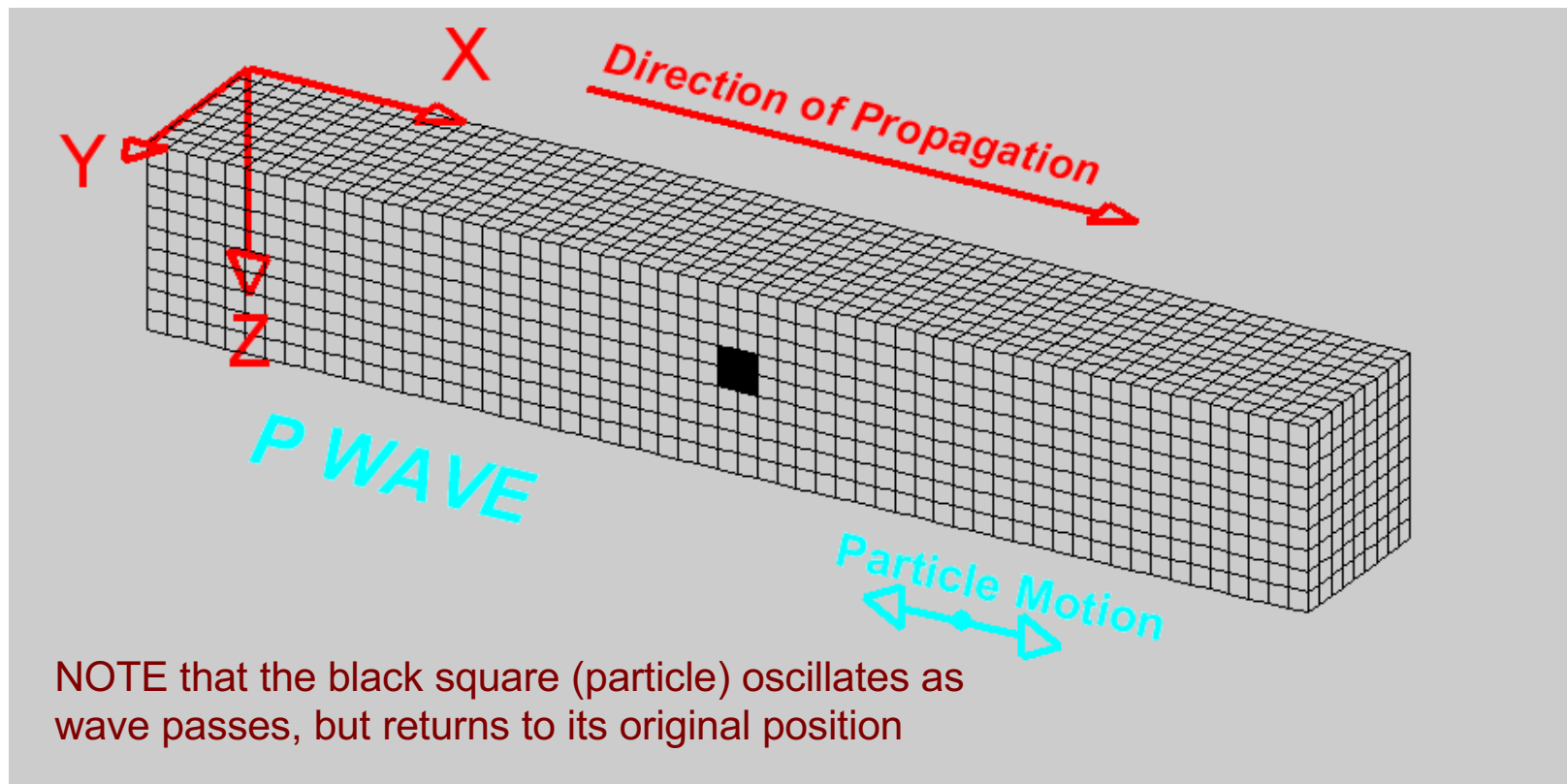
TWO CATEGORIES:

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

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

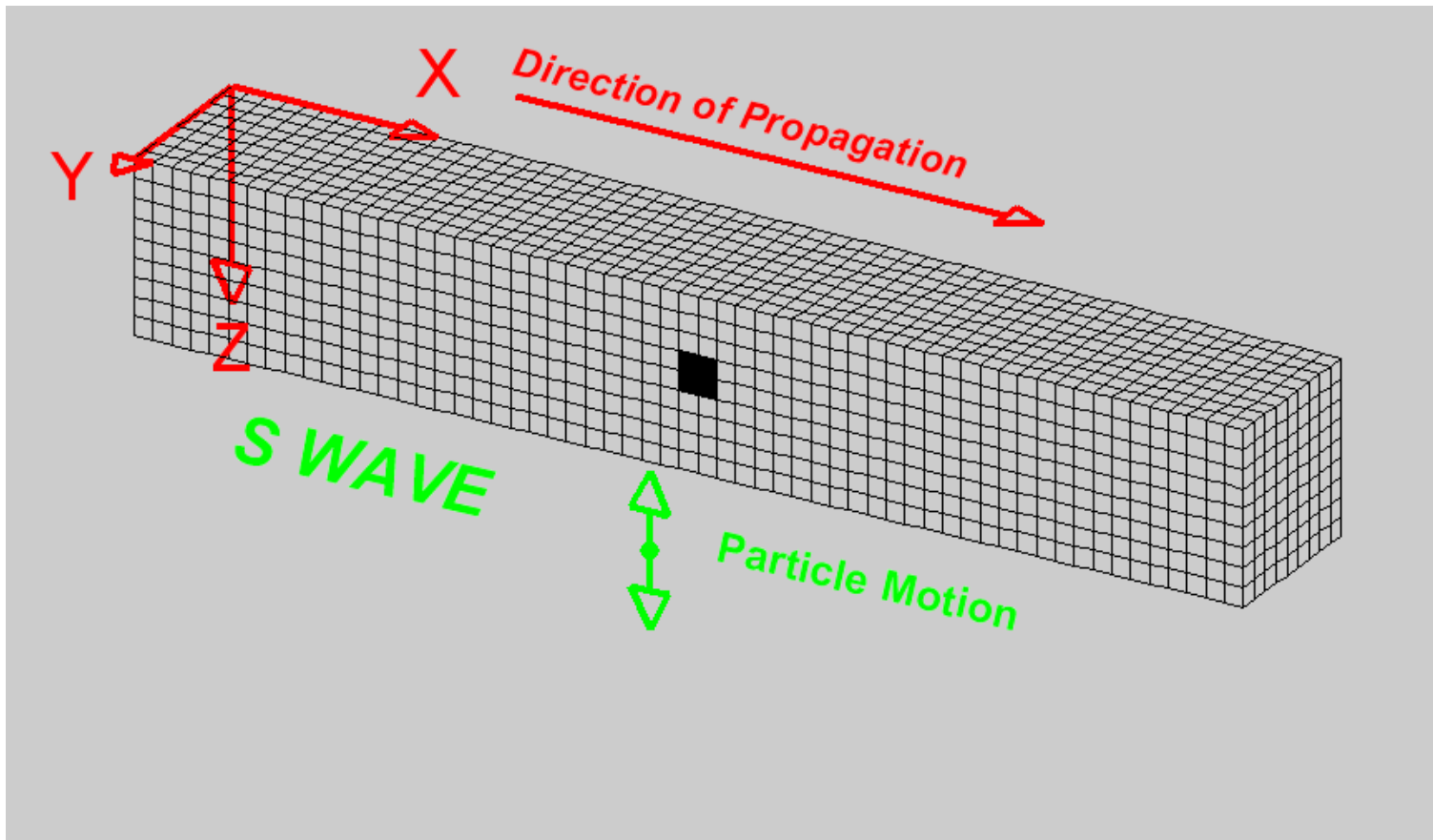
P-waves (body wave)

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



S-waves (body wave)

- shear waves (S = secondary, shear, shake....and transverse)
- particle motion is perpendicular to the direction of wave propagation
- two polarizations: SV and SH



Seismic wave velocity

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

Column of cross-sectional area A

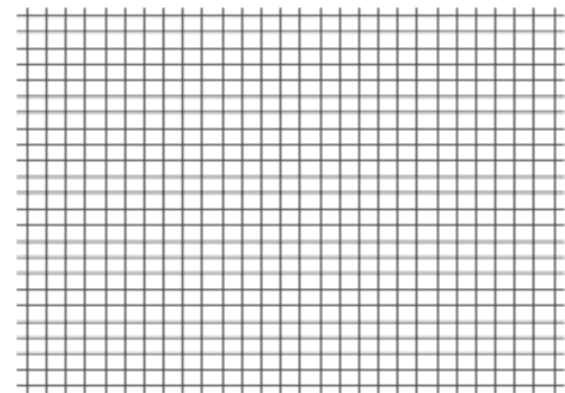
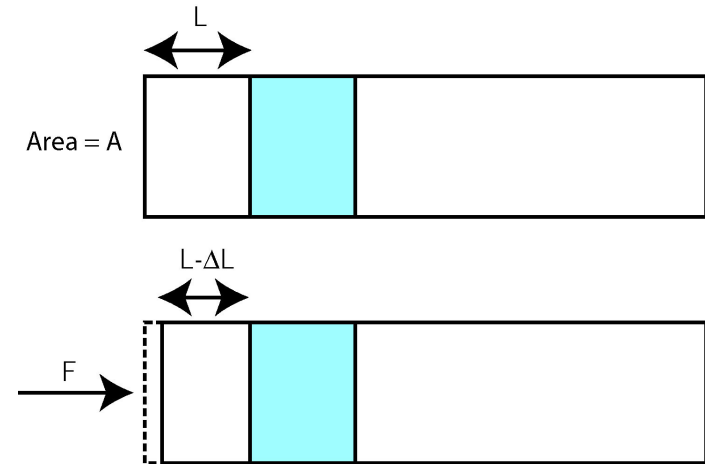
- apply force to left end
→ longitudinal stress (F/A)
- causes deformation of left side of white block
→ longitudinal strain ($\Delta L/L$)
- stress and strain related by ψ

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 blockand so on.....
→ a **wave of deformation** travels down the block

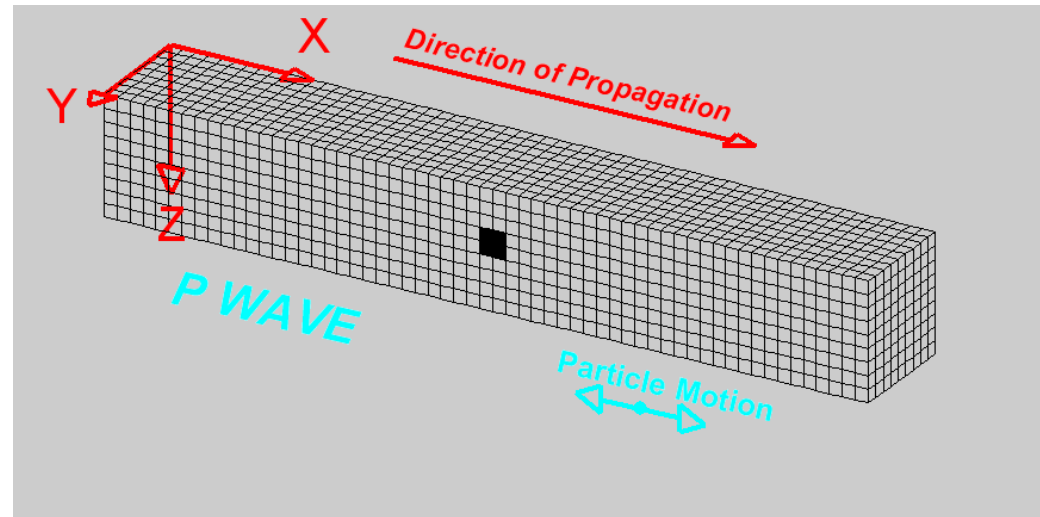
Can show that the wave velocity is:

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



This is a P-wave!

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



The expression with K and μ applies only to a material is 'isotropic' (uniform property in all directions)

$\nu = E =$ Young's modulus = Longitudinal modulus = *'a stiffness parameter that governs how a material resists stretching or compression along one direction'*

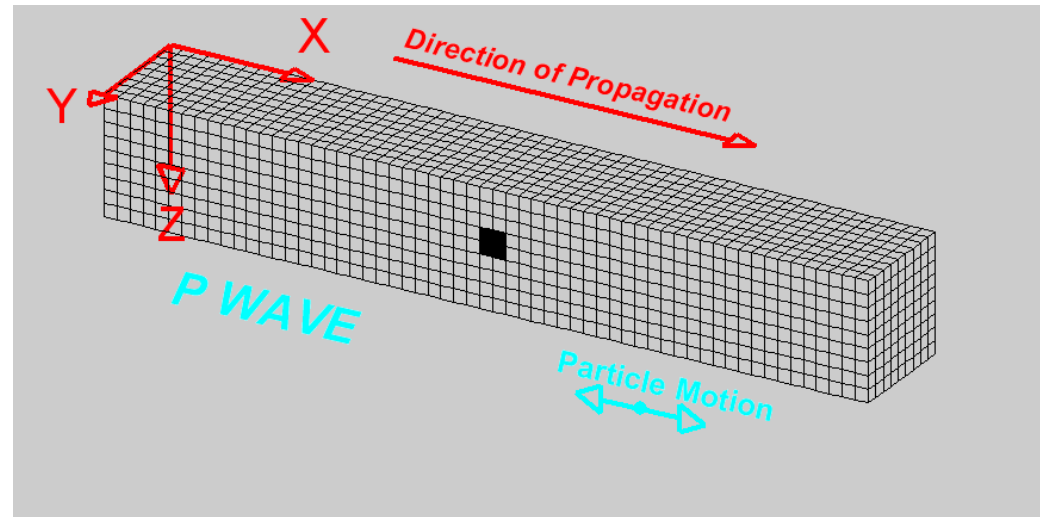
$$\nu = E = K + \frac{4}{3}\mu$$

ν = Greek letter for 'n' (nu), it is not velocity of the wave v

K = **bulk modulus** = *'also called compressibility, a material property that defines how easily a material resists being squeezed from all sides'*

This is a P-wave!

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

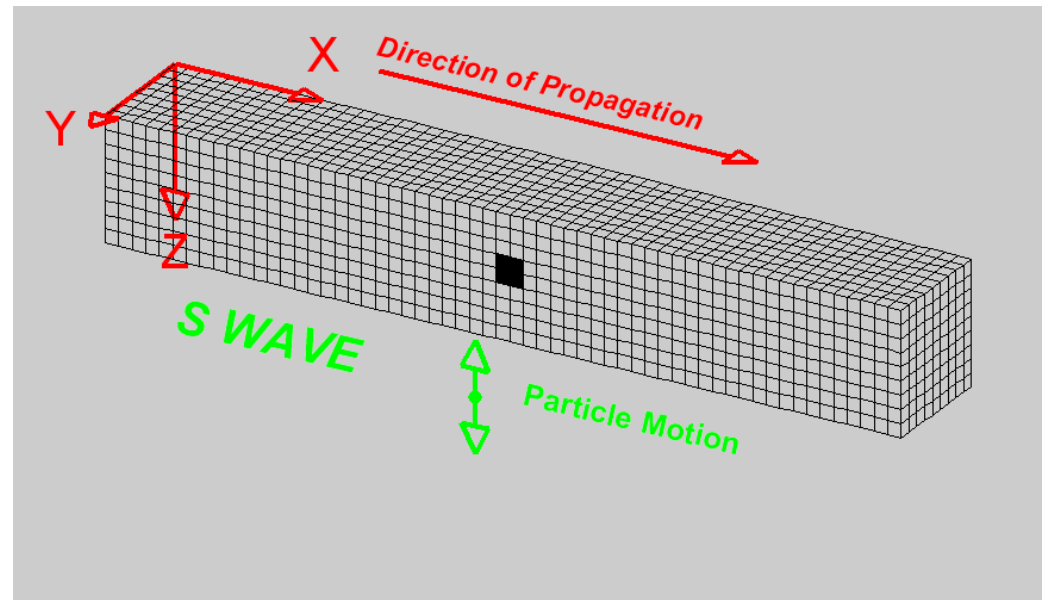


S-waves

– 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 $\rightarrow$ higher V_P and V_S
- V_P and V_S do not depend on frequency $\rightarrow$ **non-dispersive**
- in fluids, $\mu = 0 \rightarrow$ 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
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Air	~350
Water	1400 – 1500
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SEDIMENTARY ROCKS

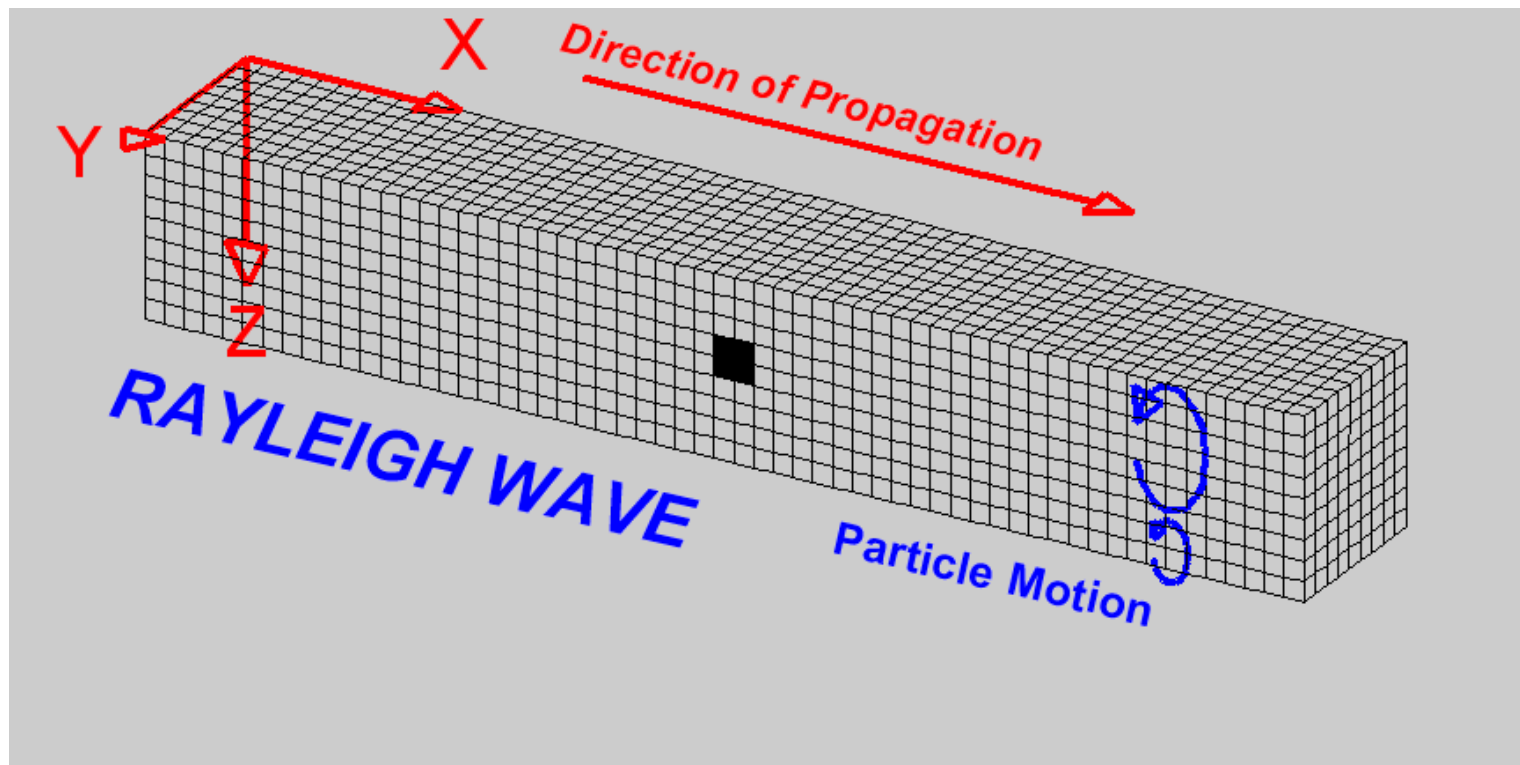
- low velocity due to **porosity**
 - low velocity material in pores
 - velocity decreases as porosity increases

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

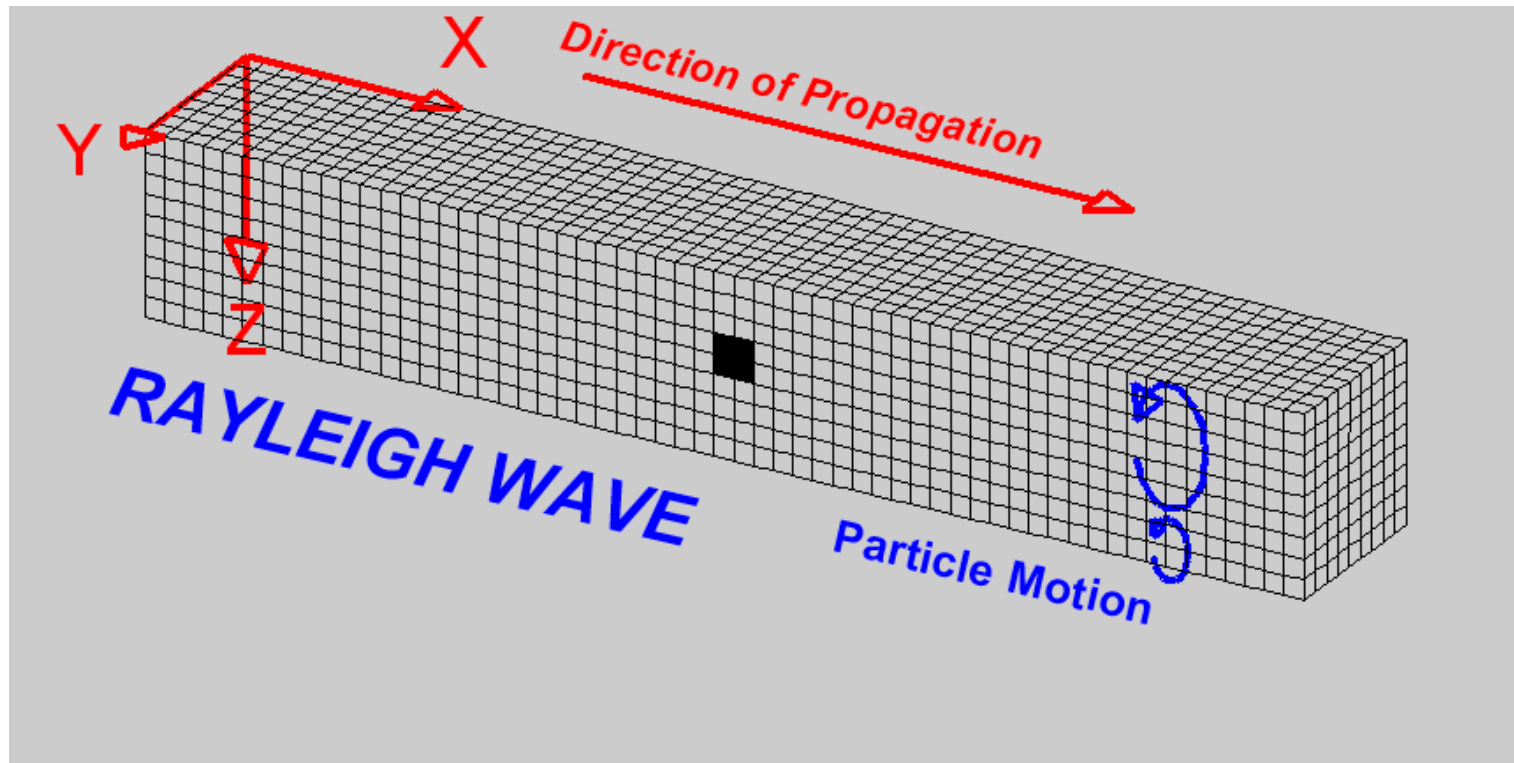
- coupling between P-waves and SV-waves at an interface
- velocity is less than V_s
- Special case for a Poisson solid when $\nu = 0.25$, then: $V_r = 0.92 V_s$
- elliptical retrograde particle motion (“ground roll”)
- particle motion decreases exponentially with depth



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)

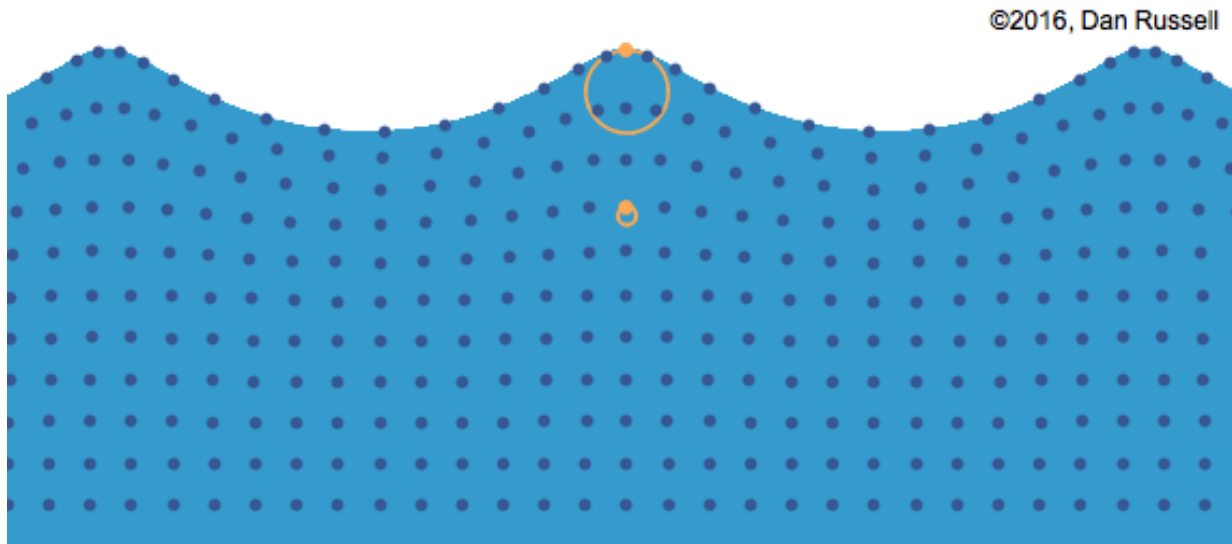


Water waves, which are very similar to Rayleigh waves in nature.

- Key difference:

- water waves (particles rotate clockwise, i.e., *prograde*, see below)

- Rayleigh waves (particles rotate counter-clockwise, i.e., *retrograde*)



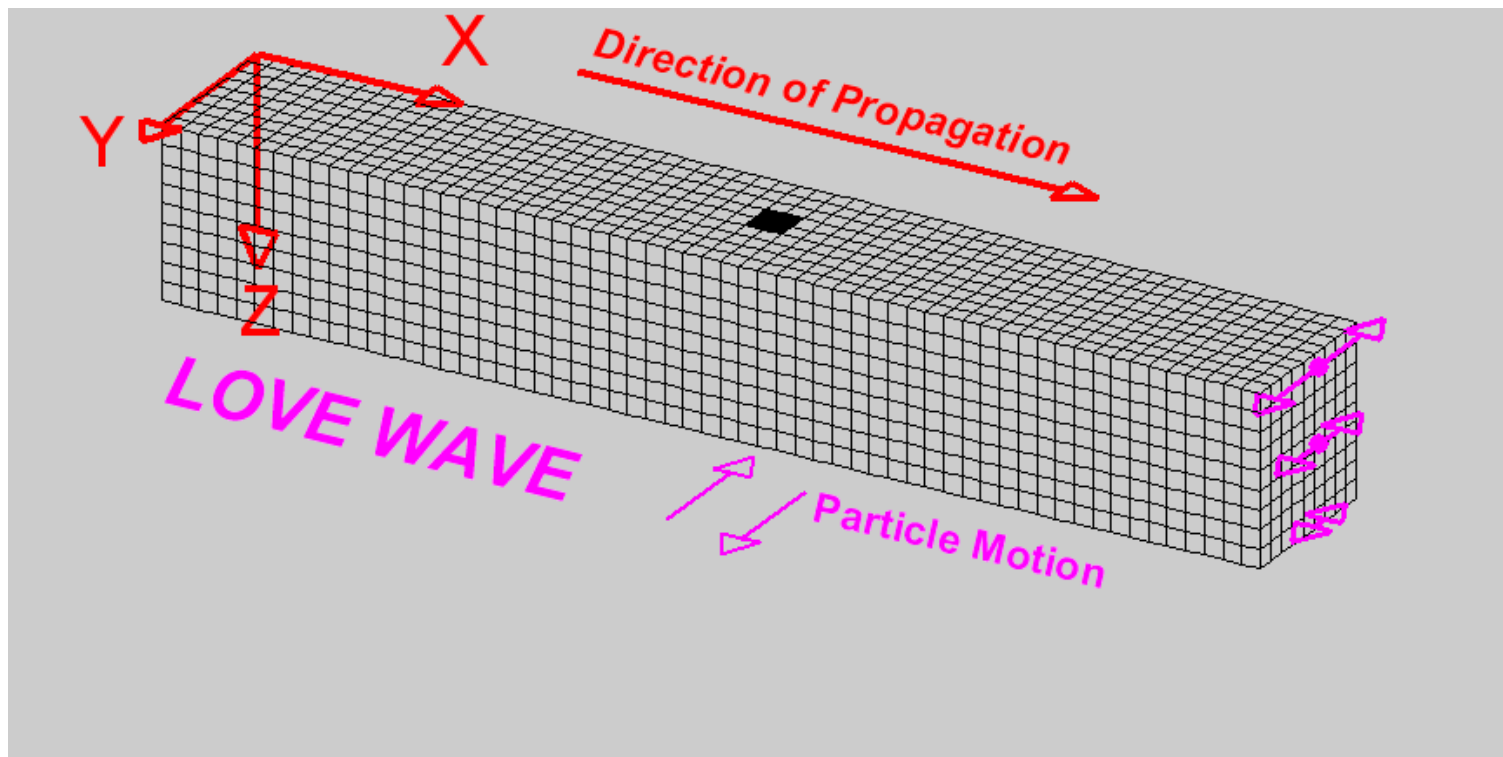
Key observations:

- circles, but in reality, ellipses that are placed vertically
- larger ellipses near the surface, suggests bigger movements (i.e., larger vertical displacement amplitude vs. horizontal)

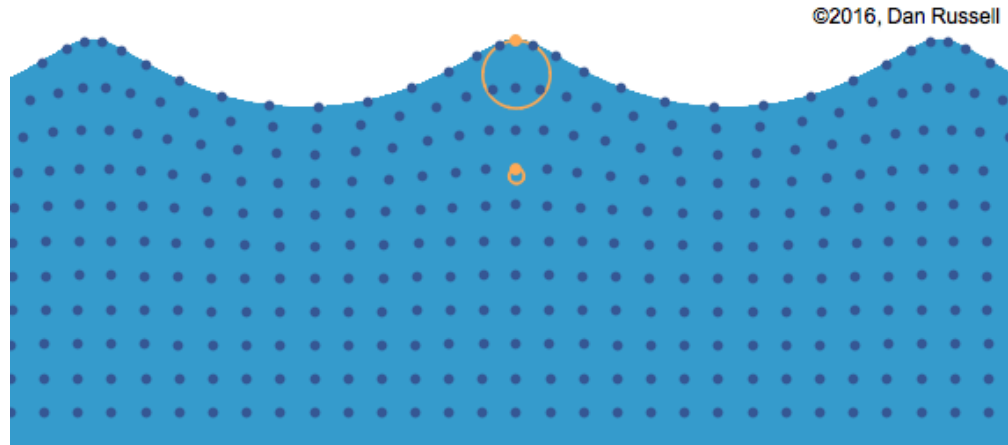
Love waves

- 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 $\rightarrow$ dispersive

$$V_{S_1} < V_L < V_{S_2}$$



**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

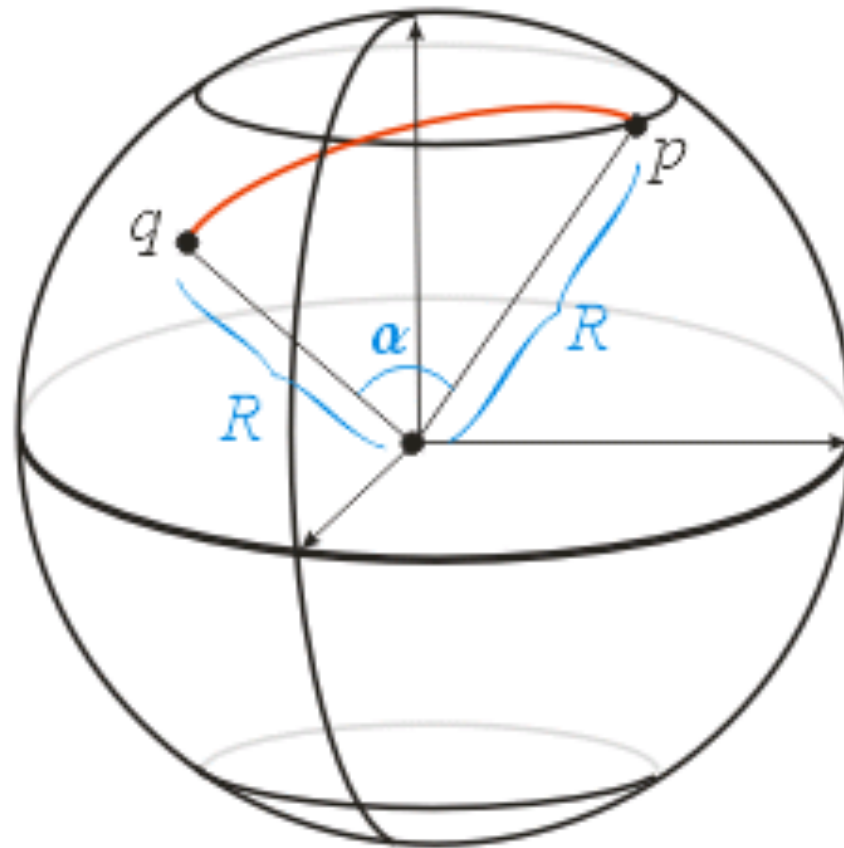
- * Travels at a speed (V) ($g = 9.81 \text{ m/s}^2$, d = water depth):

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

- * Deep water, wavelength $\sim 200 \text{ km}$, amplitude $\sim \text{few cm}$.

- * Shallow water - much higher (20 m! in Japan 1986, 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 $d=R\alpha$

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