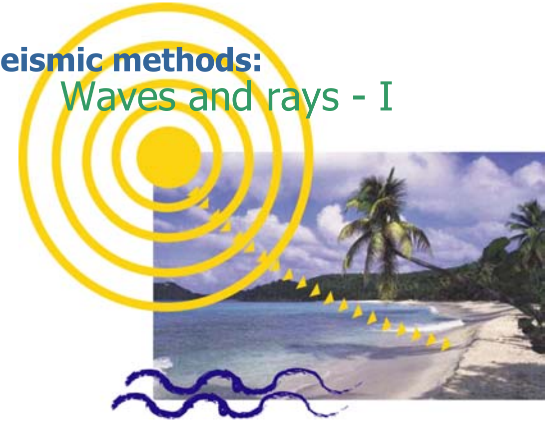
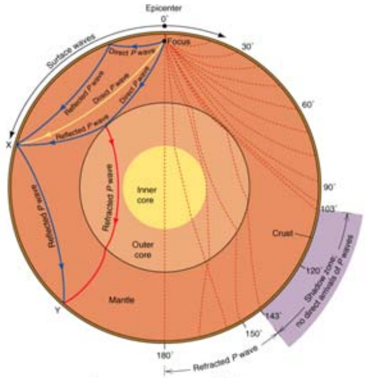
Seismic methods:
Waves and rays - I

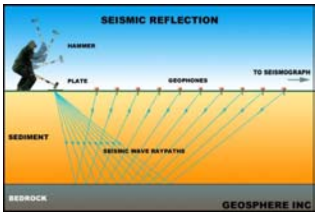


Reading:
Today: p113-123
Next Lecture: p123-133

Seismic surveying
...think sources and targets



Global seismology
- Earthquake sources
- Global ray paths
- Imaging 3D structure of the Earth's interior



Refraction & Reflection seismology
- Controlled sources
- Crustal and uppermost mantle ray paths
- Crustal structure

Elastic waves

When a stress is applied (or released) the corresponding strain propagates out from the source.

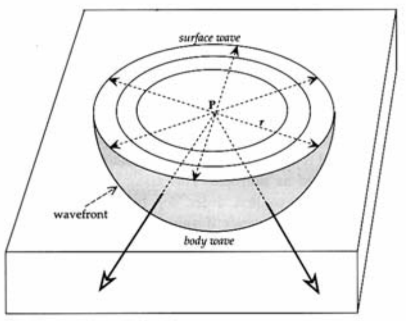
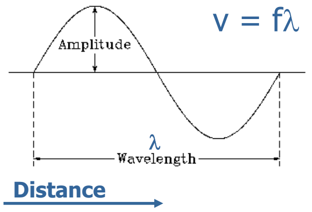


Fig. 3.9 Propagation of a seismic disturbance from a point source P near the surface of a homogeneous medium; the disturbance travels as a body wave through the medium and as a surface wave along the free surface.

Point source seismic disturbance:

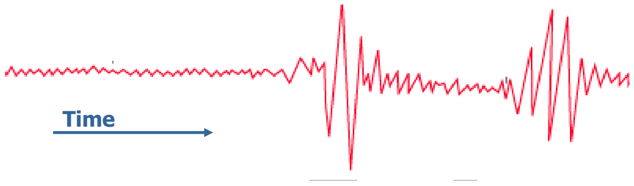
- Wavefront expands out from the point: **Huygen's Principle**
- Body waves: sphere
- Surface waves: circle
- **Rays**: perpendicular to wavefront

Waves – a reminder



$v = f\lambda$

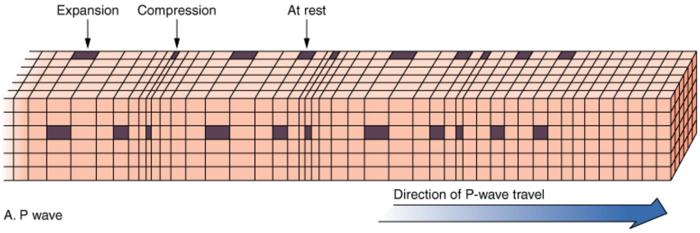
- Velocity, v
- Wavelength, λ
- Frequency, f
- Period, T = 1/f



Body waves

P-waves

- P for “primary” or “push-pull”
- Compression and rarefaction, no rotation
- Causes volume change as the wave propagates
- Similar to sound waves traveling through air

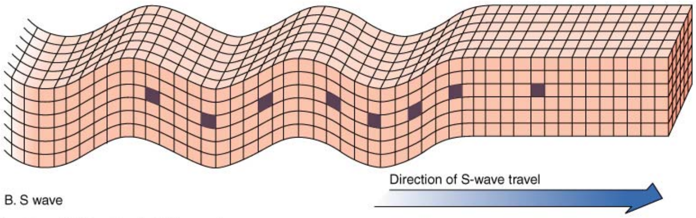


A. P wave

Body waves

S-waves

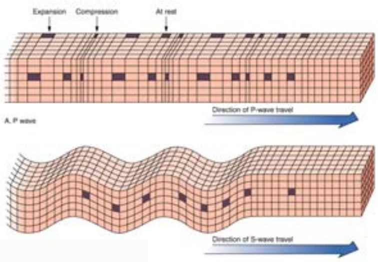
- S for “secondary” or “shear” and “shake”
- Shearing and rotation
- No volume change as the wave propagates

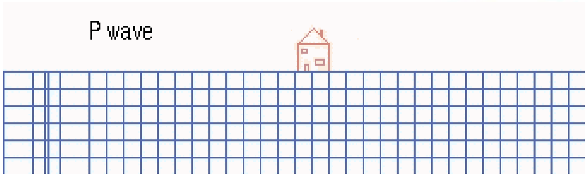


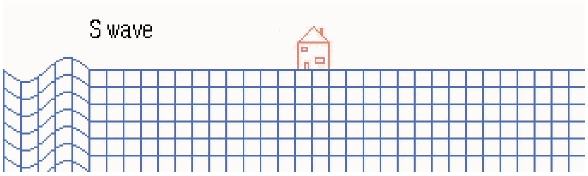
B. S wave

Body waves

P and S-waves







Elastic moduli

describe the physical properties of the rock …and determine the seismic velocity

Young's modulus

Poisson's ratio

Shear modulus, μ

- Force per unit area to change the shape of the material

Bulk modulus, κ

- Ratio of increase in pressure to associated volume change
- Always positive

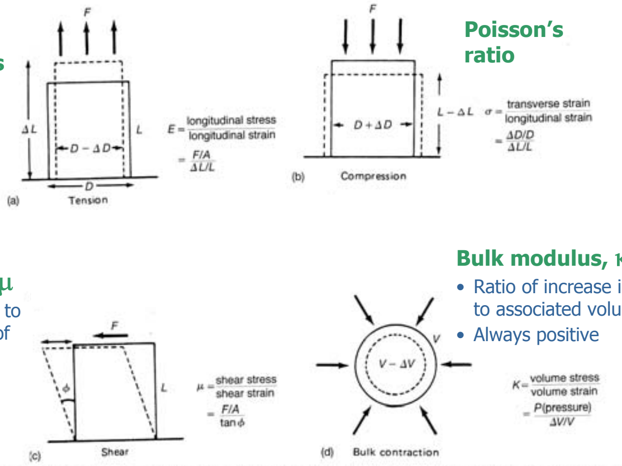


Fig. 4.2 Common types of elastic stress and strain. Cross-sections of bodies shown before strain (solid line) and after strain (dashed line). Directions of stress are shown by thick arrows. The related elastic moduli are defined. (a,b) Young's modulus, *E*, and Poisson's ratio, σ; (c) shear (or rigidity) modulus, μ; (d) bulk modulus, *K*: application of uniform pressure shown by thick arrows around the body. Poisson's ratio is a measure of the relative deformation of the body in two perpendicular directions. *F* denotes the force acting on a cross-sectional area *A*.

P and S-velocities

P-velocity

$$V_P = \sqrt{\frac{\kappa + \frac{4}{3}\mu}{\rho}}$$

change of shape and volume

S-velocity

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

change of shape only

For liquids and gases $\mu = 0$, therefore

➔ $V_S = 0$ and V_P is reduced in liquids and gases

➔ Highly fractured or porous rocks have significantly reduced V_P

The bulk modulus, κ is always positive, therefore $V_S < V_P$ always

P-waves are the most important for controlled source seismology

- They arrive first making them easier to observe
- It is difficult to create a shear source, explosions are compressional

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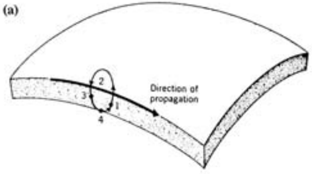
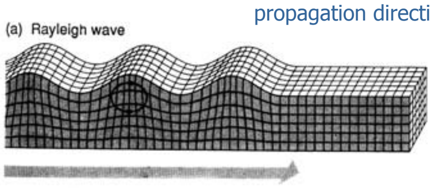
Surface waves

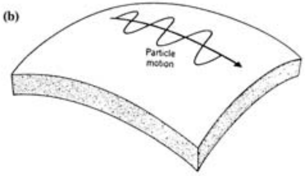
Rayleigh





Ground roll in plane of propagation direction

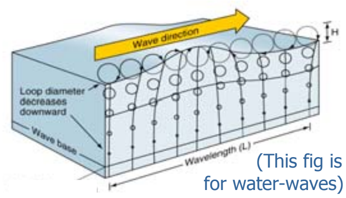
Love



Ground shake in a horizontal direction

- Surface waves have lower velocities than body waves
- They are mainly a source of noise for us

Velocity sensitivity

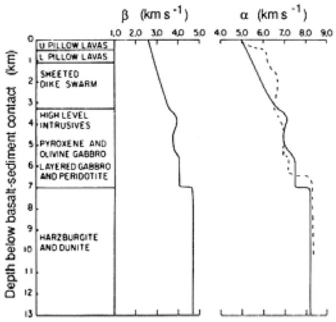


(This fig is for water-waves)

The amplitude of wave motion decreases with depth

➔ Related to depth/wavelength

➔ Longer wavelengths sample deeper



Oceanic lithosphere

Seismic velocity generally increases with depth.

Surface waves are **dispersive**, which means their velocity is dependent on their wavelength. This is because longer wavelength sample deeper where the velocity is greater.

Also, if velocity increases with depth, longer wavelengths arrive first

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Source spectrum

This is the range of frequencies within the source pulse

Try to avoid frequencies where there is significant noise e.g. 50 Hz

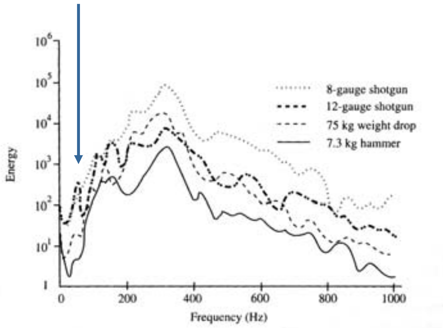


Figure 2-26 A comparison of relative energies and frequency content for shotgun and weight-drop seismic sources.

Sources:

Shotgun

Hammer

Explosive

Air gun

higher to lower frequency

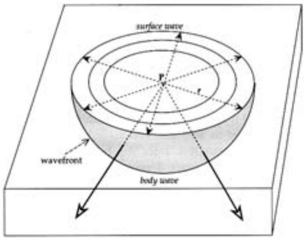
Attenuation

The amplitude of an arrival decreases with distance from the source

1. Geometric spreading

Energy spread over a sphere: $4\mu r^2$

Amplitude $\propto 1/r$



2. Intrinsic attenuation

Rocks are not perfectly elastic.

Some energy is lost as heat due to frictional dissipation.

Amplitude $\propto e^{-\alpha r}$

where α is the absorption coefficient (dependent on wavelength)

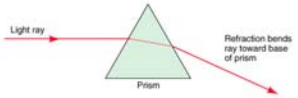
Total attenuation

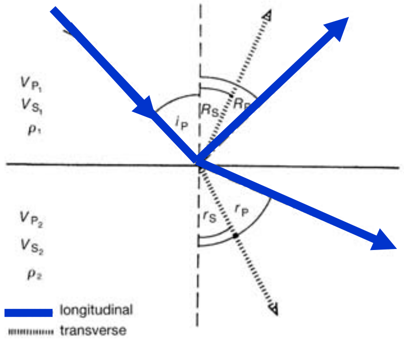
$$A = (A_0\, e^{-\alpha r})/r$$

Higher frequencies attenuate over shorter distances due to their shorter wavelengths.

Therefore, high frequencies decay first leaving a low frequency signal remaining.

Reflection and transmission





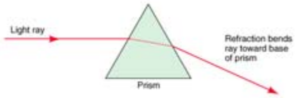
Seismic rays obey Snell's Law

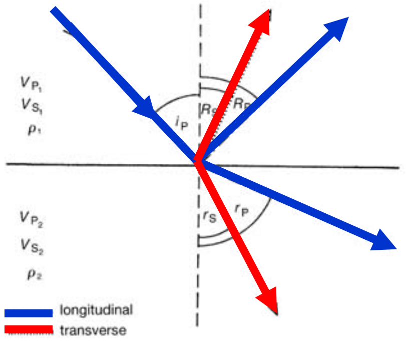
(just like in optics)

The angle of incidence equals the angle of reflection, and the angle of transmission is related to the angle of incidence through the velocity ratio.

$$\frac{\sin i_P}{V_{P1}} = \frac{\sin R_P}{V_{P1}} = \frac{\sin r_P}{V_{P2}}$$

Reflection and transmission





Seismic rays obey Snell's Law

(just like in optics)

The angle of incidence equals the angle of reflection, and the angle of transmission is related to the angle of incidence through the velocity ratio.

But a conversion from P to S or vice versa can also occur. Still, the angles are determined by the velocity ratios.

$$\frac{\sin i_P}{V_{P1}} = \frac{\sin R_P}{V_{P1}} = \frac{\sin r_P}{V_{P2}} = \frac{\sin R_S}{V_{S1}} = \frac{\sin r_S}{V_{S2}} = p$$

where p is the **ray parameter** and is constant along each ray.

Amplitudes reflected and transmitted

The amplitude of the reflected, transmitted and converted phases can be calculated as a function of the incidence angle using **Zoeppritz's equations**.

Simple case: **Normal incidence**

Reflection coefficient

$$R_C = \frac{A_R}{A_i} = \frac{\rho_2 V_2 - \rho_1 V_1}{\rho_2 V_2 + \rho_1 V_1}$$

Transmission coefficient

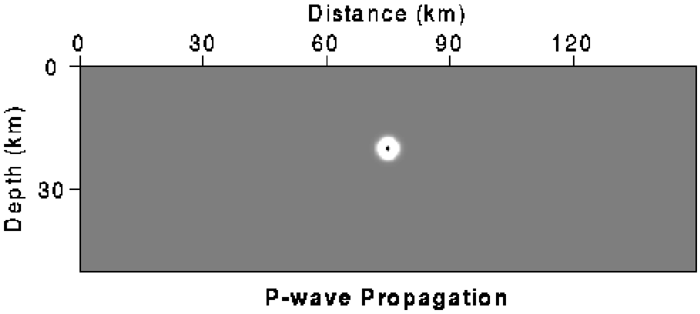
$$T_C = \frac{A_T}{A_i} = 1 - R_C = \frac{2\rho_1 V_1}{\rho_2 V_2 + \rho_1 V_1}$$

Reflection and transmission coefficients for a specific impedance contrast

These coefficients are determined by from the product of velocity and density – the **impedance** of the material.

R_C usually small – typically 1% of energy is reflected.

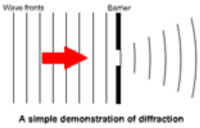
Reflection and transmission



You can see:
 a direct wave, reflected and transmitted waves, plus multiples…

Diffraction

From optics



A simple demonstration of diffraction

A sharp break in a reflector acts as a secondary source of a spherical wavefront

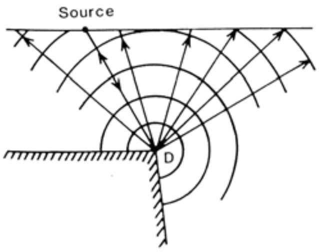
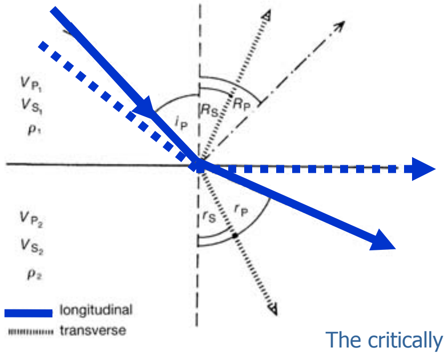


Fig. 4.5 Diffracted wavefronts from a sharp edge, D, which has been set into oscillation by waves coming from a seismic source.

Critical incidence



$$\frac{\sin i_P}{V_{P1}} = \frac{\sin r_P}{V_{P2}}$$

when $V_2 > V_1$, $r_P > i_P$

therefore, we can increase i_P until $r_P = 90°$

When $r_P = 90°$ $i_P = i_C$ the critical angle

$$\sin i_C = \frac{V_{P1}}{V_{P2}}$$

The critically refracted energy travels along the velocity interface at V_2 continually refracting energy back into the upper medium at an angle i_C

➔ a **head wave**

Head wave

- Occurs due to a low to high velocity interface
- Energy travels along the boundary at the higher velocity
- Energy is continually refracted back into the upper medium at an angle i_C
- Provides constraints on the boundary depth e.g. Moho depth

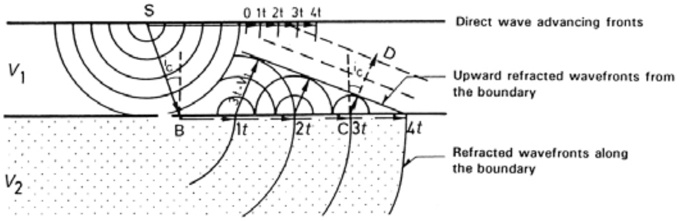
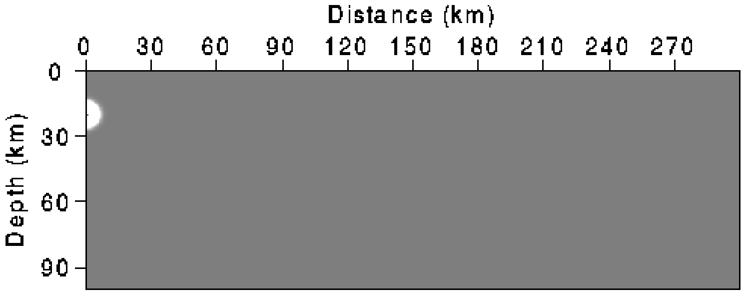


Fig. 4.6 Schematic illustration showing the ray paths of the incident wave (SB) striking the boundary at critical angle (i_c), and the refracted wave (BC) traveling along the boundary with velocity V_2 ($>V_1$). The latter is refracted back to the first medium (V_1) at the same angle (i_c) and re-emerges with a ray path such as CD. Advancement of the wavefronts is shown from the instant ($t=0$) when the incident ray strikes the boundary at B. (Modified from Klitten, 1987.)

Head wave



Reflection and Refraction at a Boundary

You can see:

a head wave, trapped surface wave, diving body wave

Factors affecting velocity

$$V_P = \sqrt{\frac{\kappa + \frac{4}{3}\mu}{\rho}}$$

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

Density – velocity typically increases with density

(κ and μ are dependant on ρ and increase more rapidly than ρ)

Porosity and **fluid saturation**

Increasing porosity reduces velocity.

Filling the porosity with fluid increases the velocity.

$$\frac{1}{V_{sat}} = \frac{\phi}{V_F} + \frac{1-\phi}{V_M}$$

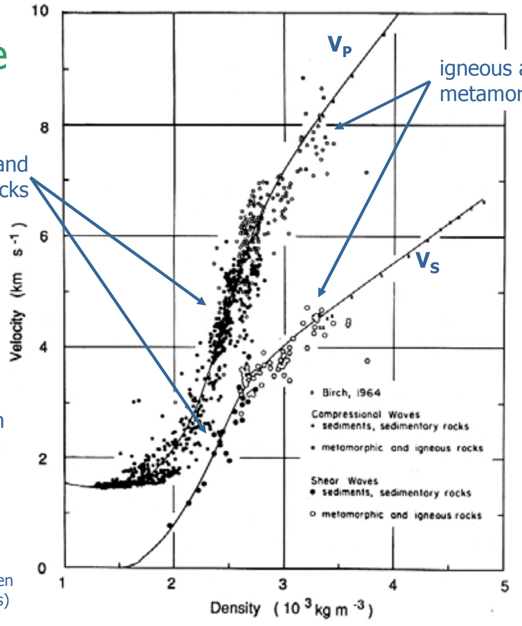
Poisson's ratio – related to V_P/V_S

This is used to distinguish between rock/sediment types. It is usually more sensitive than just V_P alone.

The significant variations in sediments are usually due to porosity variations and water saturation. Water saturation has no effect on V_S (for low porosities) but a significant effect on V_P.

Velocity and density

Nafe-Drake curve



V_P

V_S

sediments and
sedimentary rocks

igneous and
metamorphic rocks

This curve has been approximated using the expression

$$\rho = aV_P^{1/4}$$

(a is a constant: 1670 when ρ in km/m^3 and V_P in km/s)

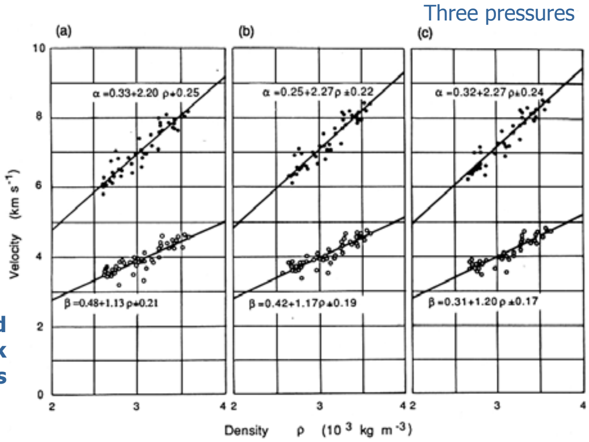
Velocity and density

Birch's Law

A linear relationship between velocity and density

$$v = a\rho + b$$

Three pressures



Crust and mantle rock observations

Typical rock velocity ranges

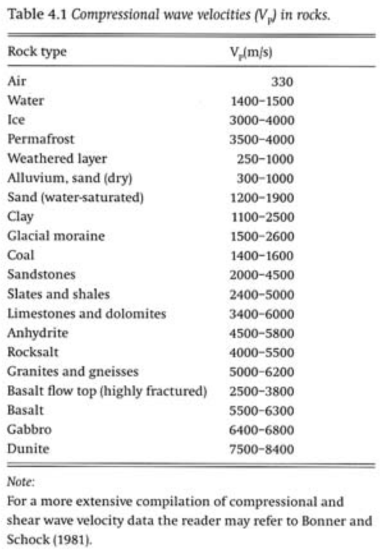


Table 4.1 *Compressional wave velocities (V_p) in rocks.*

Rock type	V_p(m/s)
Air	330
Water	1400–1500
Ice	3000–4000
Permafrost	3500–4000
Weathered layer	250–1000
Alluvium, sand (dry)	300–1000
Sand (water-saturated)	1200–1900
Clay	1100–2500
Glacial moraine	1500–2600
Coal	1400–1600
Sandstones	2000–4500
Slates and shales	2400–5000
Limestones and dolomites	3400–6000
Anhydrite	4500–5800
Rocksalt	4000–5500
Granites and gneisses	5000–6200
Basalt flow top (highly fractured)	2500–3800
Basalt	5500–6300
Gabbro	6400–6800
Dunite	7500–8400

Note:
For a more extensive compilation of compressional and shear wave velocity data the reader may refer to Bonner and Schock (1981).

Using velocity alone to determine rock type is problematic to impossible.