THE STRUCTURE OF THE EARTH FROM SEISMIC WAVES

OUTLINE

Main Features of Seismic Velocities in the Earth

Velocity–depth profile

Seismic phases and their nomenclature: seismic ray paths and travel–time curves

Constraints from long–period surface waves and free oscillations

Velocities in the Mantle

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The transition zone: velocity jumps and phase changes

Increase with depth in the lower mantle

Reflections, with mode conversion, from the core

Velocities in the Core

P– and S–wave shadow zones from the core

Core phases, evidence for a solid inner core

Background reading: Fowler §4.2.7 & 4.3, Lowrie §3.7

MAIN FEATURES OF SEISMIC VELOCITIES IN THE EARTH

Decades of observations of seismic travel times from earthquakes and large explosions have established reliable travel time tables for an "average" Earth. Regional variations in these travel times are also well established. The variation of seismic body wave velocities with depth in the Earth has been derived by modelling and inverting these travel times. Because of mode conversion and the existence of sharp boundaries in the Earth, there is large variety of seismic body waves, or *seismic phases* as they are called. Although this variety at first sight complicates the interpretation of seismic travel times, at the same time it greatly enhances the amount of information that can be inferred from them. Data from surface wave dispersion and Earth oscillations complement the travel time data in important ways, for example in estimating S-wave velocities in the mantle and in validating the existence of a solid inner core.

More information than is available from seismic velocities is needed to estimate the Earth's density distribution. Seismology gives us P- and S-wave velocities:

$$V_P = \sqrt{\frac{K + \frac{4}{3}\mu}{\rho}} \qquad V_S = \sqrt{\frac{\mu}{\rho}}$$

These equations contain three unknowns, the elastic moduli K and μ and the density ρ, which cannot be found from just two measurements. Constraints from the mass and moment of inertia of the Earth, the expected compression from gravitational loading and the likely temperatures in the Earth allow a reasonable estimate to be made of the density profile.

MAIN BOUNDARIES WITHIN THE EARTH

Distinct jumps in the Earth's velocity–depth profile mark the boundaries between the three major subdivisions of the interior: the crust, mantle and core.

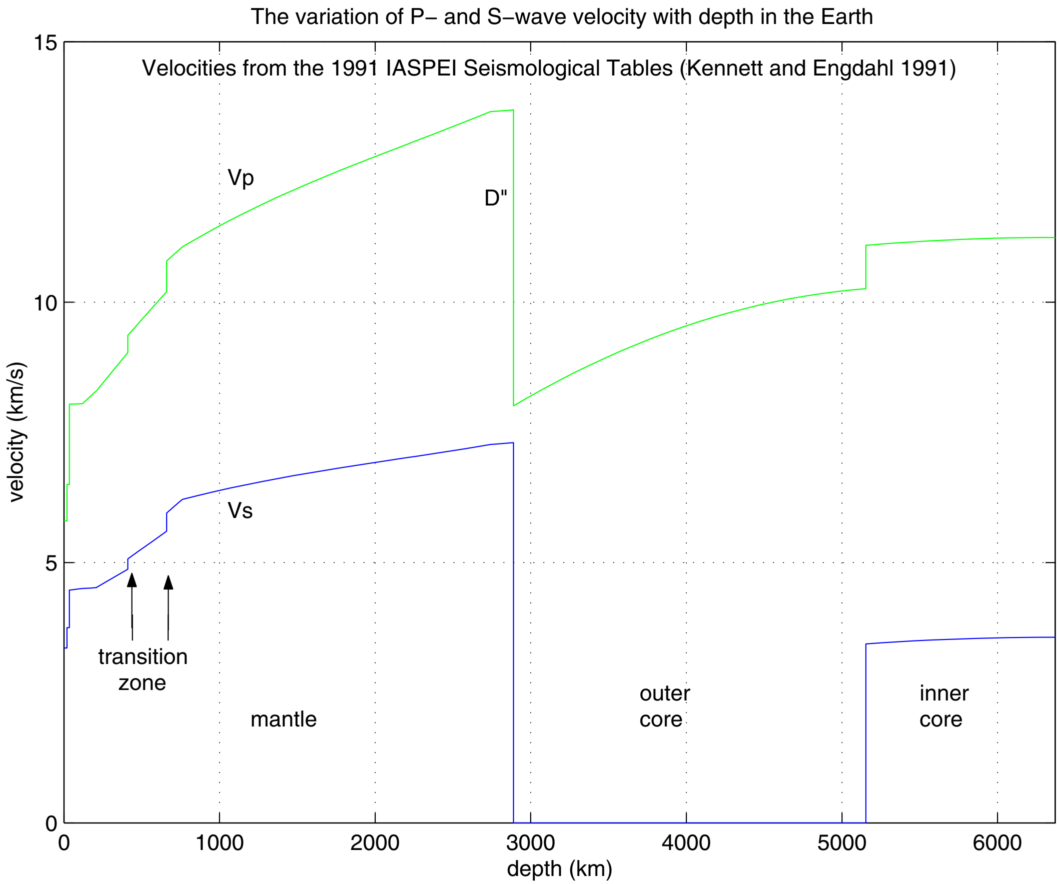
1. The crust–mantle boundary (the 'Moho') is at ~ 35 km depth under much of the continents and $\sim 5 - 10$ km under oceans. It varies considerably under continents, reaching ~ 90 km under the Himalayas for example.
2. The core–mantle boundary is at ~ 2890 km depth (i.e. the radius of the core is ~ 3481 km).

Further small jumps or zones of rapid change occur within these three zones.

1. In many areas, velocities in the lower crust (e.g. $V_P \sim 6.7$ km/s) are markedly higher than those in the upper crust ($V_P \sim 6.0$ km/s).
2. The mantle contains a transition zone that extends from 400 km depth to 700 km or possibly 1050 km, and in most areas exhibits a low–velocity layer between ~ 70 and ~ 200 km depth.
3. There is a boundary layer called D" just above the core–mantle boundary that is of great current interest since it is speculated that mantle plumes may originate there and it may be the eventual "sink" for subducted slabs.
4. There is an solid inner and a fluid outer core. The radius of the inner core is ~ 1217 km.

There are uncertainties about the finer details of the Earth's velocity–depth distribution but the major features are very firmly established.

SEISMIC VELOCITIES IN THE EARTH



Reference:

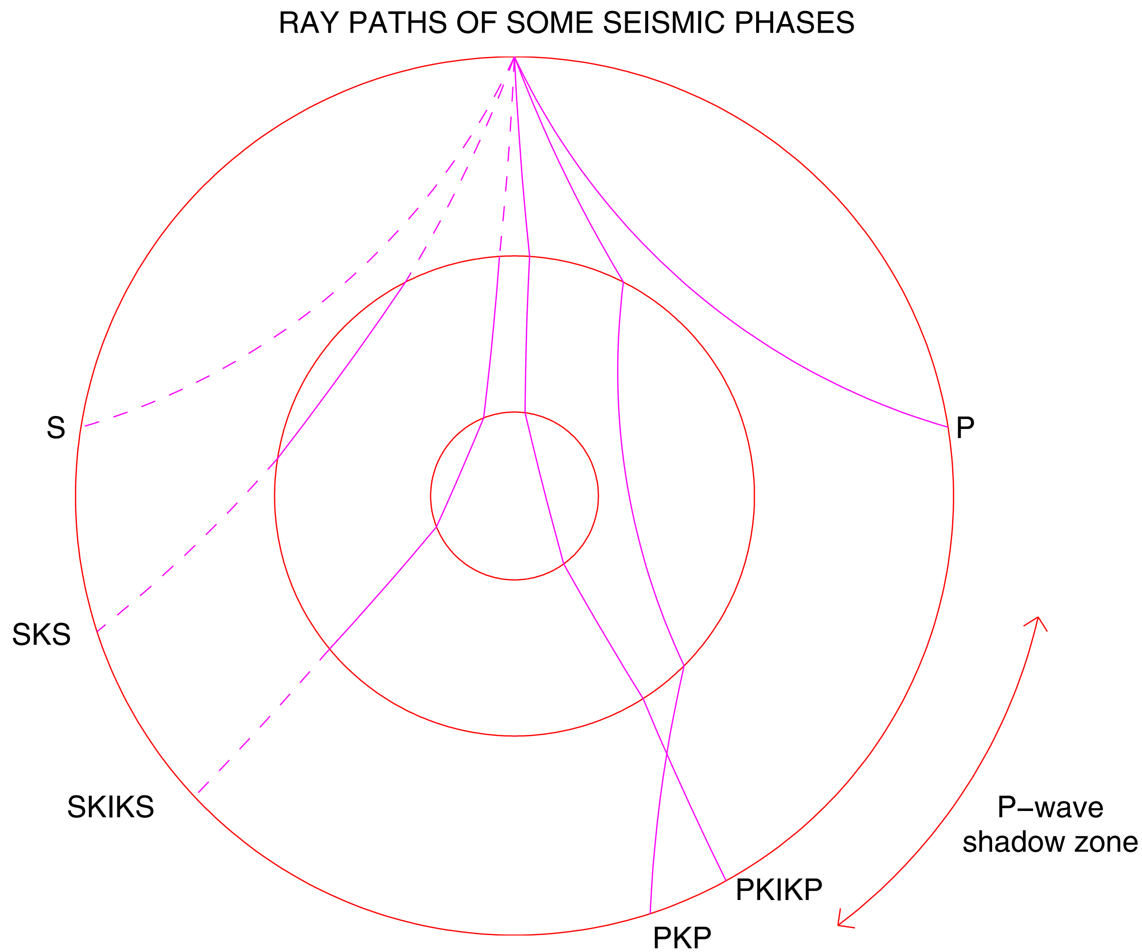
Kennett,B.L.N., and Engdahl, E.R., 1991. Traveltimes for global earthquake location and phase identification, Geophys.J.Int. **105**, 429-465.

SEISMIC PHASES

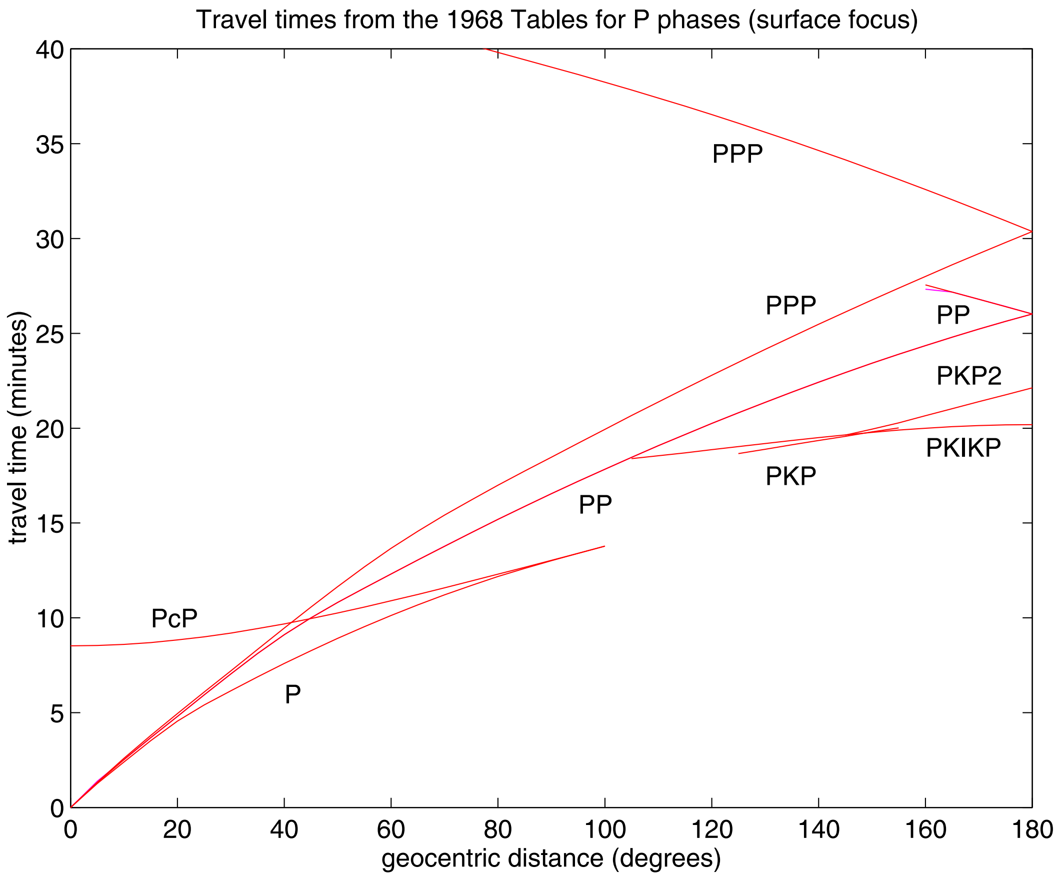
Nomenclature

$P_1, P_g (P_d)$	a P wave in the upper crust
P_2	a P wave in the lower crust or in a second crustal layer
P_n	a P wave in the upper mantle just beneath the Moho
P	a P wave in the mantle
S_1, S_g, S_2	crustal S waves analogous to P_1, P_g, and P_2
S_n, S	mantle S waves analogous to P_n and P
p	a P wave travelling upwards from an earthquake focus and reflected down from the Earth's surface
s	an S wave travelling upwards from an earthquake focus and reflected down from the Earth's surface
c	a reflection at the mantle–core boundary
K	a P wave in the outer core
I	a P wave in the inner core
i	a reflection at the outer core–inner core boundary
J	an S wave in the inner core
LR	a Rayleigh surface wave
LQ	a Love surface wave

RAY PATHS OF SOME SEISMIC PHASES



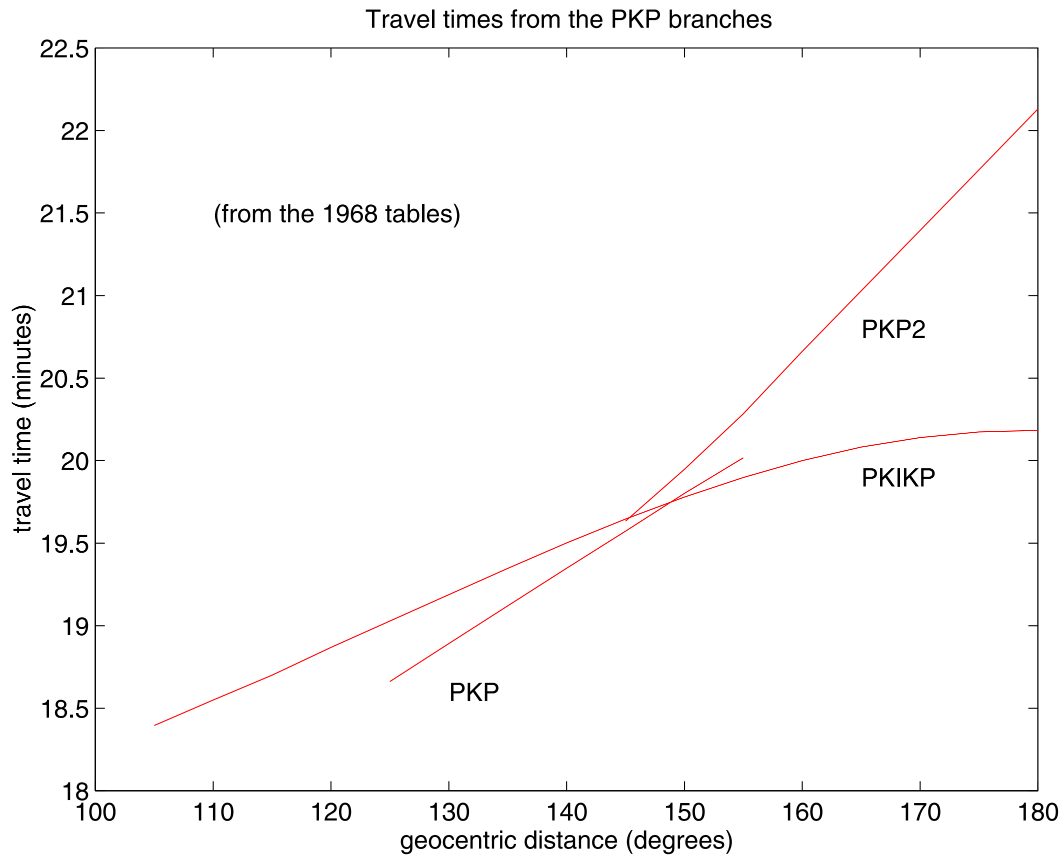
SEISMIC TRAVEL TIMES THROUGH THE EARTH



Reference:

Herrin, E., 1968. Introduction to '1968 seismological tables for P phases', Bull.seism.Soc.Am. **58**, 1193-1195.

THE TRAVEL TIME BRANCHES OF PKP



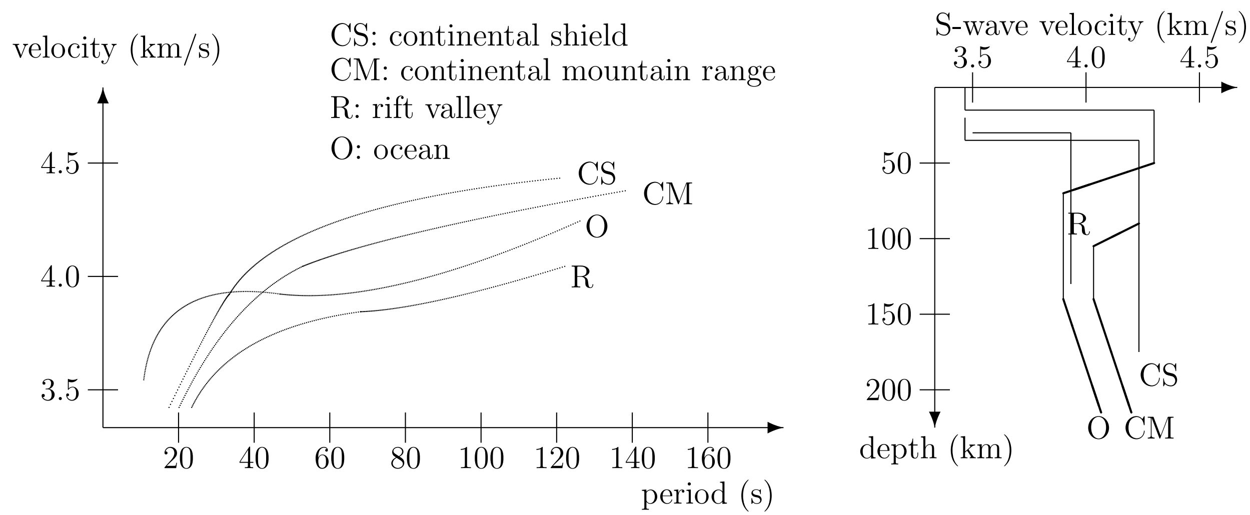
Reference:

Herrin, E., 1968. Introduction to '1968 seismological tables for P phases', Bull.seism.Soc.Am. **58**, 1193-1195.

RESULTS FROM SURFACE WAVE DISPERSION

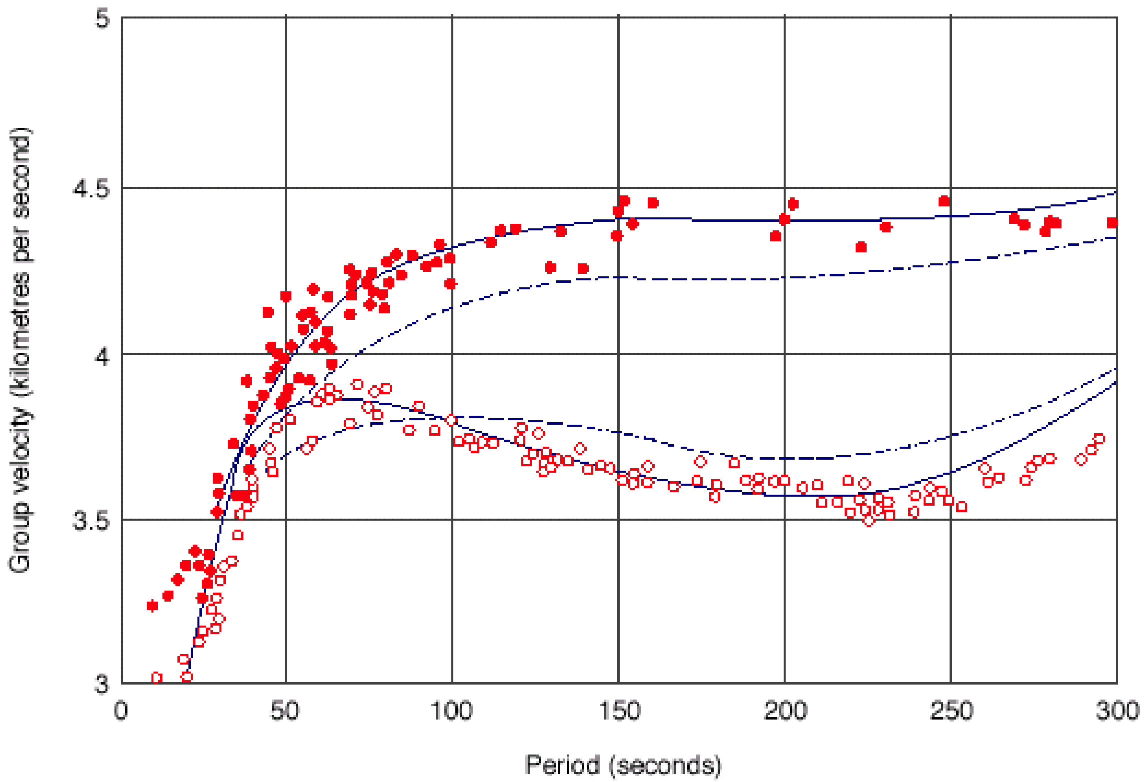
Surface wave dispersion: seismic surface waves of both Rayleigh and Love type, are dispersive; that is, their speed of propagation depends on their wavelength. Consequently surface wave trains spread out, or disperse, as they travel away from their source, the faster waves arriving first and the slower waves becoming progressively delayed.
Sensitivity to velocity structure: the ground motion from surface waves decays with depth: the shorter the wavelength, the more rapid the decay with depth. It follows that longer period surface waves 'see' further into the Earth than short period waves. Thus it is that surface wave velocities are sensitive to the variation of velocity with depth. Because of the general increase in velocity with depth, longer period waves tend to travel faster than short period waves.
Surface wave dispersion curves, shown schematically below, plot wave speed against period (or frequency or wavelength). They show an increase from crustal to mantle velocities with increasing period.
Long period surface waves: a surface wave with 100s period has a wavelength of $\sim$ 400m which is long enough to penetrate the upper mantle. Such waves give useful information on velocities, especially S-wave velocities, in the mantle. S-wave velocities in the low-velocity layer cannot be measured directly because no refraction arrivals are recorded from a low-velocity layer.

RAYLEIGH WAVE DISPERSION CURVES AND S-WAVE VELOCITY DISTRIBUTIONS



EFFECT OF A LOW-VELOCITY LAYER ON SURFACE WAVE DISPERSION

The figure below shows surface wave dispersion results for oceanic tarvel paths. Open circles are Rayleigh wave data and closed circles are Love wave data. The solid curves are the computed curves for a sub-oceanic mantle having a low-velocity layer; the dashed curves are from a sub-oceanic mantle having no low-velocity layer.



RESULTS FROM EARTH OSCILLATIONS

The periods of the Earth's free oscillations depend on the elastic moduli and densities of the interior. The longest period (lowest order) modes penetrate the core whereas higher order modes mainly sample the mantle. The differing sensitivities of the oscillation modes are used to refine models of the Earth's internal structure. The periods of oscillation range from several minutes to several hours.

The periods of the Earth's free oscillations establish that the inner core is solid. This is an important confirmation of results from body waves that penetrate the inner core as the identification of inner core seismic phases is not clear cut.

VELOCITIES IN THE MANTLE

P–wave velocities

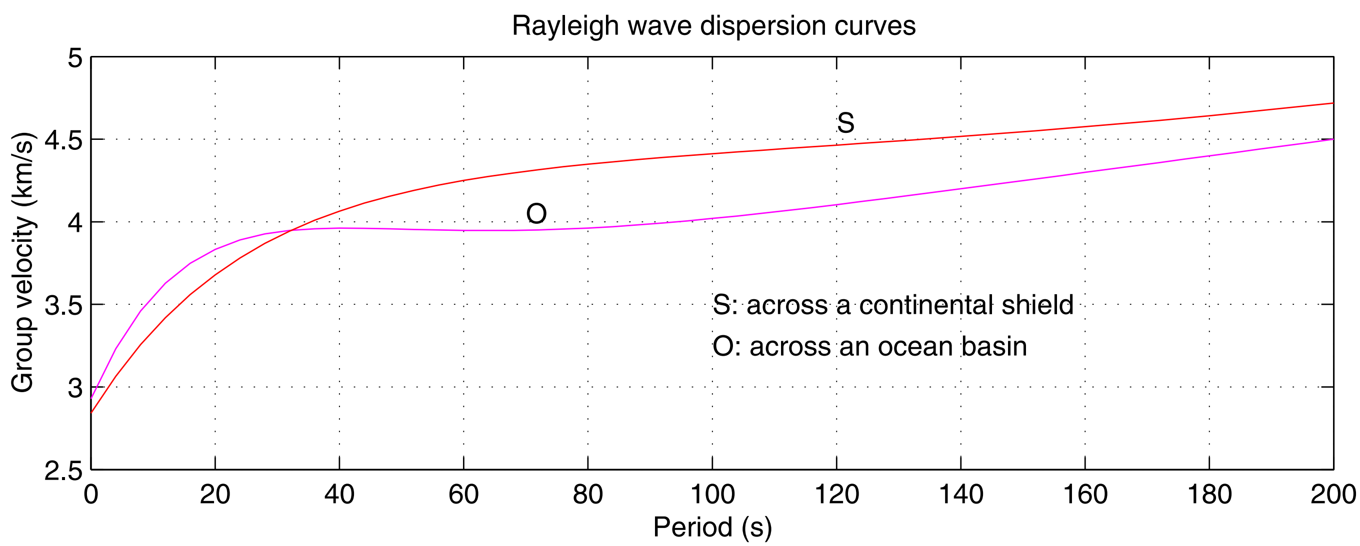
• Sub–Mohorovicic P–wave velocities mostly exceed 8.0 km/s; they fall below 8.0 km/s under ocean ridges and active continental margins.
• The properties of the upper mantle vary from region to region.
• There is a low–velocity zone at $\sim 70-200$ km depth beneath oceans and in many continental regions. A low–velocity zone has not been positively identified everywhere, notably beneath continental shields.
• Seismic waves are attenuated more strongly in the low–velocity layer.
• There are jumps in P–wave velocity at depths around $440, 670$ and possibly 1050 km. These jumps define a *transition zone* within the mantle. They are attributed to phase changes to higher density polymorphs in the peridotite model of the upper mantle. Current opinion favours $\sim$ 700 km as the base of the transition zone. No earthquakes have been observed beneath this depth.
• The P–wave velocity increases smoothly beneath 700 km through the lower mantle to $\sim$ 13.6 km/s at the core–mantle boundary. Just above this boundary there appears to be a layer in which the velocity flattens out or may even decrease slightly.

VELOCITIES IN THE MANTLE (CTD)

S–wave velocities

The S–wave velocity profile in the mantle is similar to the P–wave profile, S–wave velocities being $\sim 0.52 - 0.57$ of the corresponding P–wave velocities. The low–velocity layer tends to be more pronounced for S–waves than for P–waves.

S–wave travel times cannot be measured as accurately as P–wave travel times. Much useful information on S-wave velocities in the mantle comes from long period Love and Rayleigh waves and Earth oscillations. Surface wave dispersion data is particularly crucial in defining S–wave velocities in the low–velocity layer. The figure below illustrates the sensitivity of surface wave dispersion curves to the presence or absence of the low–velocity layer.



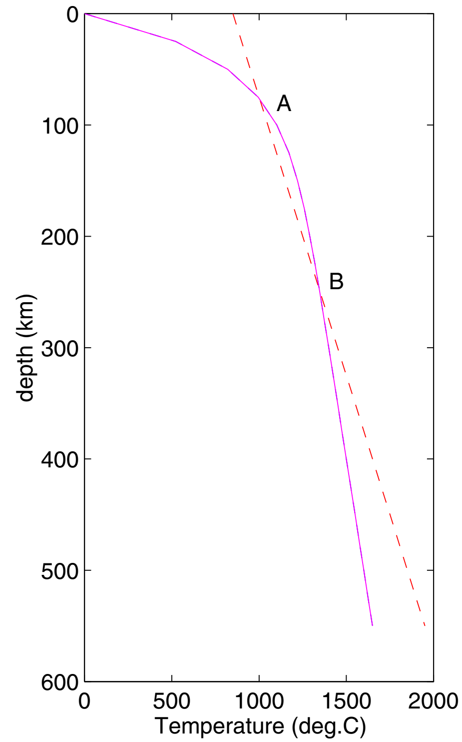
VELOCITIES IN THE MANTLE (CTD)

Anisotropy

There is clear evidence that the upper mantle at least is anisotropic; that is, the velocity of seismic waves depends on their direction of travel. A major cause is the layered or banded structure of the upper mantle.

Partial melting as an explanation of the low–velocity zone

The solid curve represents the increase in temperature with depth beneath an ocean basin. At high temperatures peridotite (mantle rock) begins to melt, forming a basaltic melt and leaving the mantle depleted in Rb, U and the rare earths (Sm, Lu, Re, etc.). The onset of this partial melting increases with increasing pressure (dashed line). The partial melt forms in the zone between A and B where the temperature exceeds the partial melting point. Only a small percentage of rock melts but this is enough to explain the inelastic properties of the low–velocity zone.



SEISMIC WAVE VELOCITIES IN THE CORE

- The P–wave velocity increases smoothly with depth from 8.1 km/s at the outer rim to 10.3 km/s above the inner core boundary.
- The boundary between the inner and outer core is fairly sharp.
- The P–wave velocity in the inner core is about 11.3 km/s.
- The outer core does not transmit S–waves and is therefore presumed to be fluid.

The P– and S–wave shadow zones from the core

Because the P–wave velocity in the core is less than that in the lower mantle, the core casts a shadow. P–waves incident on the core are refracted towards the centre of the Earth, leaving a gap in the P–wave travel time curve. The shadow appears at geocentric distances from the source between 103° and 143°. 103° is the distance at which the P–wave ray through the mantle just grazes the core; 143° is the distance at which the first PKP wave emerges. In practice some P–waves are diffracted a few degrees beyond 103° and some PKP waves diffract to distances less than 143°

Since the outer core cannot transmit S–waves, there is a complete shadow for S–waves. The only S–waves that are recorded in the shadow zone are waves that have bounced around between the core and surface and those that were mode converted to S on emerging from the outer core.

SEISMIC WAVE VELOCITIES IN THE CORE (CTD)

Evidence for a solid inner core

• P–waves transmitted to depths of 5154 km, the depth to the inner core, are sharply refracted; these represent the phase PKIKP (see below). P–waves corresponding to reflections upwards off the inner core (the phase PKiKP) are also regularly recorded. The sudden increase in P–wave velocity suggests a fluid–solid boundary.
• There is some evidence that P–waves incident on the inner core generate mode–converted refracted S–waves within the inner core (the phase PKJKP).
• Models of the periods of Earth oscillations employ a solid inner core.

Seismic core phases

Seismic waves that traverse a particular type of ray path are referred to as seismic phases. The following are some seismic phases associated with the core:

PcP: the phase travelling as a P–wave from the source and reflected as a P–wave at the core–mantle boundary.

PcS: the phase travelling as a P–wave from the source and and mode–converted to a reflected S–wave at the core–mantle boundary.

PKP: the phase travelling as a P–wave from the source through the mantle, outer core and mantle again in its path back to the surface. P–waves in the outer core are labelled K.

PKIKP: the phase travelling as a P–wave from the source through the mantle, outer core, inner core, outer core and mantle again in its path back to the surface. P–waves in the inner core are labelled I.

Many other combinations of travel path are possible.