



4. A look at Earth's interior using seismic waves

The bulk of the Earth's interior cannot be studied directly. The deepest scientific boreholes are only a few kilometers deep. Tectonic forces sometimes bring deep crustal rocks to the Earth's surface where they can be studied by geologists. Rocks from the mantle may also be transported to the surface during volcanic eruptions. Another source of information comes from lab experiments trying to recreate the very high pressure and temperature conditions occurring deep inside the Earth. Meteorites also provide information on the structure, composition and formation process of planetary bodies and, by analogy, the Earth as well.

Indirect methods of exploring the Earth's interior are the domain of geophysics and use a variety of techniques related to:

- The behavior of seismic waves travelling inside the Earth
- Earth's gravitational field
- Earth's magnetic field
- The thermal properties of rocks

This chapter deals with the seismic waves as a tool to reconstruct the Earth's interior.

4.1. Behavior of seismic waves traveling inside Earth

There are two types of seismic waves traveling through Earth's interior: P-waves (compressional, "push-pull" motion) and S-waves (shear waves, "sideways" motion). As we saw in the previous chapter, P-waves are faster than S-waves. S-waves cannot propagate in liquids or gases. Seismic waves travel faster in denser materials. Hence, their speed generally increases with increasing depth (pressure) and is influenced by the chemical and mineral compositions of rocks. For example, P-waves travel more slowly in granite (6 km/s) than in basalt (7 km/s) or peridotite (8 km/s). Note that peridotite (upper mantle rock) is denser than basalt which is denser than granite.

P-waves and S-waves propagate in all directions. As they encounter the boundary between two layers, they can be either reflected or refracted. These reflected and refracted waves can be recorded by seismographs around the world. Knowing the distance from the epicenter and the arrival times of the different waves, information on the structure of the Earth's interior can be obtained. The relationship between the angle of incidence and the angle of refraction is given by Snell's Law of refraction (Fig. 45).

But let's consider first a simple model with two layers –layer 1 and layer 2– and a seismic wave (P or S) generated by an earthquake near the surface (Fig. 46). The interface between these two layers is smooth and horizontal. In addition, we assume that layer 2 is denser than layer 1. Consequently, seismic waves travel faster in layer 2 than in layer 1 ($v_2 > v_1$).



Figure 45: Representation of Snell's law of refraction. If $V_1 < V_2$, when the incident wave reaches the interface between layer 1 and 2 at the critical angle (i_c), the wave is refracted along the interface itself. The wave is reflected back into layer 1 for angles of incidence greater than the critical angle.

In this model, layer 1 could be the crust and layer 2 could be the mantle. We can distinguish 4 types of waves:

- **Direct wave:** seismic wave traveling at the surface —therefore never reaching the interface between layer 1 and 2— (red lines in Fig. 46). This wave travels at speed v_1 .
- **Refracted wave:** seismic wave refracted at the interface between layer 1 and 2 when the angle of incidence is smaller than the critical angle (magenta lines in Fig. 46).
- **Head wave:** seismic wave propagating along the interface between layer 1 and 2 at speed v_2 (critical refraction) and refracted upward into layer 1 where its speed is equal to v_1 (blue lines in Fig. 46). This wave is produced when the incident wave reaches the boundary at the critical angle. The angle at which the wave returns into layer 1 is equal to the critical angle.
- **Reflected wave:** seismic wave reflected back into layer 1 when the angle of incidence is larger than the critical angle (green lines in Fig. 46).

Seismographs located at various distances from an earthquake's epicenter can be used to reconstruct the travel-time curves of the different waves (Fig. 46). They will detect the direct wave, the head wave, and the reflected wave at different times. Seismographs close to the epicenter will record the direct wave first. Beyond a certain distance, seismographs will start to "see" the head wave first. That is because the head wave travels along the interface between layer 1 and 2 at speed v_2 , which is faster than the direct wave. The point on the travel-time curve from which the head wave is recorded prior to the direct wave depends on the thickness of layer 1 and on V_2 . Knowing the slope of the travel-time curve of the head wave, we can determine v_2 (Fig. 46). By recording the arrival times of the first seismic wave reaching seismographs located at various distances from an earthquake's epicenter, it is therefore possible to calculate the depth of the boundary between layer 1 and 2. The reasoning for a 2-layer model is the same for models with more layers (Fig. 47).

In 1910, Andrija Mohorovičić (1857-1936), a Croatian seismologist, used travel-time curves of seismic waves to prove the existence of the crust-mantle boundary and to determine its depth. This boundary is called the Mohorovičić discontinuity in his honor (referred to as the "Moho").



Figure 46: Behavior of seismic waves (P- or S-waves) generated by an earthquake in an idealized 2-layer model. Dotted lines represent wavefronts and plain lines terminated by arrows represent the paths of seismic waves. Four types of waves are recognized (direct wave, reflected wave, head wave, and refracted wave). The arrival times measured by seismographs are used to construct travel-time curves (reflected wave in green, direct wave in red, and head wave in blue). In the present example, the five seismographs closest to the epicenter record the direct



wave first (A), whereas the others record the head wave first (B).



Figure 47: 3-layer model showing the path of a direct wave (red line) and two head waves (blue lines). Seismographs (black squares) are located at various distances from the earthquake's epicenter. The arrival times of the first wave reaching these seismographs are plotted against the distance from the epicenter. Three lines can be recognized. The first represents the direct wave (slope $1/v_1$). The second line is the head wave refracted along the interface between layer 1 and 2 (slope $1/v_2$). The third is the head wave refracted along the interface between layer 2 and 3 (slope $1/v_3$).

4.2. Reconstruction of Earth's structure based on seismic waves

Seismic waves generated by a large earthquake may travel through the entire Earth. The speed of seismic waves increases with increasing pressure. The speed of waves therefore generally increases gradually as seismic waves travel deeper into Earth's interior*. Since speed increases gradually, so does the angle of refraction (Snell's law, Fig. 45). Therefore, the trajectory of seismic waves traveling inside the Earth is not straight but curved (Fig. 48).



Figure 48: (A) multi-layer model representing the increase in wave speed with increasing depth. As a consequence, the wave is refracted at an angle that becomes progressively larger and therefore delineates a curved path. (B) Schematic representation of the curved path of a seismic wave traveling through the Earth.

Information on the structure of the Earth and the composition of Earth's layers can be inferred from travel-time curves of seismic waves. The method that was used for the discovery of the Moho (crust-mantle boundary) can be applied to other discontinuities of the Earth's interior. For example, the German seismologist Beno Gutenberg (1889-1960) identified for the first time seismic waves that had been reflected on the mantle-core boundary. He used their travel-time curves to calculate in 1914 the depth of this boundary. These waves were named PcP for the P-waves reflected on the mantle-core boundary and ScS for the S-waves. In 1936, Inge Lehmann (1888-1993), a Danish seismologist, suggested the existence of a solid inner core based on the way P-waves are refracted on its surface. P-waves are absent between 105 and 142 degrees from an earthquake's epicenter because of the way they are refracted downward on the surface of the core (P-waves' shadow zone, Fig. 49). S-waves do not propagate through the liquid outer core and therefore S-waves do not reappear beyond 105 degrees (S-waves' shadow zone, Fig. 49). Many more categories of waves can be identified according to their travelling path inside the Earth (Fig. 50).



Figure 49: Paths of P-waves (left) and S-waves (right) inside the Earth after they originate from the focus of an earthquake. The distance from the earthquake's focus is expressed in degrees.

* Notable exceptions to this trend are the Low Velocity Zone (LVZ) and the mantle-core boundary. See section 7.3 for more explanations.



Figure 50: Examples of seismic wave paths and their name (left) and examples of travel-time curves for different categories of waves (source: USGS).

4.3. The Earth's interior: the crust, the mantle, and the core

Based on the travel-time curves of seismic waves seen above, geologists constructed a model of the Earth's interior reflecting how the speed of P-waves and S-waves changes with depth (Fig. 51). The speed of seismic waves inferred from travel-time curves can be compared with the speed determined experimentally for different types of rocks. Using these data and other lines of evidence (e.g. field observations, meteorites...), geologists have established a model of the structure and composition of Earth's interior.



Figure 51: Variations of various properties of Earth's layers with increasing depth down to the center of the Earth. (A) Variations with depth of the temperature (geotherm, grey line) and melting temperature (pink line) of the mantle (down to 3000 km) and the core (from 3000 km to the center of the Earth), (B) variations with depth of the speed of seismic waves (P-wave in blue and S-wave in green) and density (black line), and (C) close-up of the variations with depth of S-wave speed from 0 to 900 km. Source: modified from Understanding Earth 6th edition.

4.3.1. The crust

The upper part of the continental crust is made primarily of felsic rocks (granite) whereas the oceanic crust is composed mainly of mafic rocks (basalt and gabbro). Felsic rocks are less dense than mafic rocks (see section 5.4). Differences in density and resistance to compression affect the speed of P-waves. As a result, seismic waves travel more slowly in felsic rocks than in mafic rocks. Seismic wave speed increases across the crust-mantle boundary (Moho) because the upper mantle is composed of peridotite (ultramafic rock) which is denser and more resistant to compression than the rocks of the continental and oceanic crusts.

The crust is thin under the ocean (7 km), thicker under continents (33 km) and thickest under large mountains (70 km). The variations of crust thickness relative to Earth's topography can be explained by the principle of **isostasy**. Isostasy explains why the oceanic crust lies at low elevations below sea level (ocean basins) whereas the continental crust rises above sea level and, at places, reaches several kilometers in altitude (mountains).

A nice and easy way to explain this is to assimilate the mantle to a fluid and the crust to a solid floating on it. Although the mantle is solid in reality, it may "flow" (deform without breaking) on the very long term (over millions of years). In a static liquid, the pressure at a given depth is everywhere the same. The pressure is calculated by determining the weight of the material above the given depth (pressure = weight/area = $mg/A = \rho Vg/A = \rho gh$). As a result, the changing elevation of the crust floating on the mantle involves either changes in the crust thickness (Airy's model, Fig. 52A) or changes in the crust density (Pratt's model, Fig. 52B).



Figure 52: Two models explaining how the Earth's topography relates either to (A) lateral changes in the crust thickness (Airy's model) or (B) lateral changes in the crust density (Pratt's model). g = gravitational acceleration (in m/s^2), considered constant), ρ_m = density of the mantle (in kg/m^3), ρ_{cr} = density of the crust, ρ_w = density of seawater, w = ocean depth.

4.3.2. The mantle

The mantle down to 410 km is made of peridotite, a rock depleted in silica and enriched in iron and magnesium compared to crustal rocks (Fig. 51C). The most abundant mineral of the upper mantle is olivine. At a depth of around 410 km, the pressure forces olivine to adopt a more compact crystal structure. This change in mineralogy is responsible for a jump in wave speed at the same depth (Fig. 51C). A second —larger— jump occurs at around 660 km below the surface (Fig. 51C). This second jump corresponds to another change in the crystal structure of olivine with atoms becoming even more closely packed. The 660-km discontinuity defines the limit between the upper and lower mantle.

The outer rigid shell of the Earth is the lithosphere (100 km thick on average). The speed of seismic waves increase with increasing pressure (depth) in this layer but a remarkable thing happens near its base and in the upper part of the asthenosphere between 100 and 200 km below the surface: a decrease of wave speed defining a zone of low wave velocity (**Low Velocity Zone** or LVZ, Fig. 51C). The slow down of seismic waves in this zone is due to the partial melting of the peridotite composing the upper mantle (see the geotherm crossing the mantle melting curve in Fig. 51A). This is the weak part of the asthenosphere on which the lithosphere (tectonic plates) may slide (see section 2 on plate tectonics).

4.3.3. The core

A complete change in composition and physical properties occurs across the mantle-core boundary. Above this boundary, the mantle is composed of solid silicate rocks. Below this boundary, the outer core is composed of a liquid iron and nickel alloy. This greatly affects the behavior of seismic waves. P-waves slow down abruptly whereas the speed of S-waves drops to zero (Fig. 51B). This mantle-core boundary also delineates a zone where temperature increases greatly. Scientists have suggested that part of the mantle directly above the mantle-core boundary could be melted and provide a source of magma that could be linked to hot spot volcanism.

Note that the core is probably composed mostly of iron and nickel* but its precise composition is still debated. The core could possibly also contain lighter elements like sulfur and oxygen.

* The composition of the core cannot be observed directly. Indirect evidence comes from the study of seismic waves, astronomical data (relative abundance of elements in the solar system, theory of planetary formation), and our knowledge of the composition of the mantle and the crust and the mass of the Earth. Another important piece of information comes from meteorites. One particular type of meteorites is composed almost entirely of iron and nickel. These meteorites are believed to be fragments of the core of other planetary bodies.