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## INTERNAL STRUCTURE OF THE EARTH

The delineation of internal structure of the Earth, different discontinuities and nature of material between two major discontinuities is mainly based on the analysis of the recorded reflected and refracted seismic waves. In broad sense, the internal structure of the earth is divided into three concentric cells, namely crust, mantle and core according to the chemical property of the materials (Figure 1.16). Further, crust is divided as upper and lower crust, mantle as upper and lower mantle and core as inner core and outer core. The crust and mantle together are also classified as lithosphere, asthenosphere and mesosphere, on the basis of physical property of the materials. Following subheadings describe crust, mantle and core in brief along with the discovery of major discontinuities.

**Internal structure of theearth based on P- and S-waves velocity variations (after Kennett and Engdahl, 1991).**

### Crust

Andria Mohorovicic (1909) found only direct P-wave (Pg) arrivals near the epicentre during the analysis of an earthquake in Croatia. But beyond 100 km two P-wave arrivals were recorded and direct P-wave was overtaken by the second P-wave (Pn). He concluded that it is only possible when Pn has travelled at greater speed. Mohorovicic identified Pn as a refracted wave from the upper mantle. According to his calculations, the velocity of direct P-wave and refracted P-wave was 5.6 km/s and 7.9 km/s, respectively; and the estimated depth, at which sudden increase of velocity occurred was 54 km. Now, this seismic discontinuity between crust and mantle, where there is sudden increase of seismic wave velocity, is called as Mohorovicic discontinuity, or simply Moho. V. Conrad (1925) found faster P-wave (P*) and S-wave (S*) as compared to Pg and Sg waves during the analysis of Tauern earthquake of 1923 (Eastern Alps) in upper crustal layer. The estimated velocities of P* and S* waves (6.29 km/s and 3.57 km/s, respectively) were lesser than the velocities of Pn and Sn waves refracted from the Moho. Conrad inferred the existence of a lower crustal layer with higher velocity as compared to the upper crustal layer. The interface separating the crustal mass into upper and lower crust is called as Conrad discontinuity, in honour of V. Conrad. Worldwide analysis of recorded reflected and refracted seismic waves reveals that the structures of the crust and upper mantle are very complex. The thickness of crust is highly laterally variable. It is 5–10 km in oceanic region, below the mean water-depth of about 4.5 km. The vertical structure of continental crust is more complicated than that of oceanic crust. The thickness of continental crust varies from 35 to 40 km under stable continental areas and 50 to 60 km under young mountain ranges.

### Upper Mantle

The Mohorovicic discontinuity defines the top of the mantle. The average depth of Moho is 35 km, although it is highly variable laterally. Several discontinuities of seismic wave velocity and velocity gradients exist in the upper mantle. The uppermost mantle, 80–120 km thick, is rigid in nature in which velocity of seismic wave increases with depth. This rigid part of uppermost mantle together with crust forms the lithosphere. The lithosphere play an important role in plate tectonics.

There is an abrupt increase of seismic wave velocity (3 – 4%) at depth of around 220 ±

30 km. This interface is called as the Lehmann discontinuity. Between the base of lithosphere and the Lehmann discontinuity, there is low velocity layer (LVL) with negative velocity gradients.

The average thickness of LVL is around 150 km. This LVL is known as asthenosphere, which also plays an important role in plate tectonics. Asthenosphere behaves as viscous fluid in long term and thus decouples the lithosphere from the deeper mantle. The travel-time versus epicentral-distance curves of body wave show a distinct change in slope at epicentral distance of about 20°. This is attributed to a discontinuity in mantle velocities at a depth of around 400 km. This is interpreted as due to a petrological change from an olivinetype lattice to a more closely packed spinel-type lattice. A further seismic discontinuity occurs at a depth of 650–670 km. This is a major feature of mantle structure that has been observed world-wide. In the transition zone between the 400 km and 670 km discontinuities there is a further change in structure from b-spinel to g-spinel, but this is not accompanied by appreciable changes in physical properties.

**Lower Mantle**

The lower mantle lies just below the important seismic discontinuity at 670 km. Its composition is rather poorly known, but it is thought to be consisting of oxides of iron and magnesium as well as iron-magnesium silicates with a perovskite structure. The uppermost part of the lower mantle between 670 and 770 km depth has a high positive velocity gradient. Beneath it, there is great thickness of normal mantle, characterized by smooth velocity gradients and the absence of seismic discontinuities. Just above the core-mantle boundary an anomalous layer, approximately 150–200 km thick, has been identified in which body-wave velocity gradients are very small and may even be negative.

**Core**

R.D. Oldham first detected the fluid nature of the outer core seismologically in 1906. He observed that, if the travel-times of P-waves observed at epicentral distances of less than 100° were extrapolated to greater distances, the expected travel-times were less than those observed. This meant that the P-waves recorded at large epicentral distances were delayed in their path. Oldham inferred from this the existence of a central core in which the P-wave velocity was reduced. Gutenberg (1914) verified the existence of a shadow zone for P-waves in the epicentral range between 105° and 143°. Gutenberg also located the depth of top of outer core at about 2900 km. A modern estimate for this depth is 2889 km. It is characterized by very large seismic velocity change and is the most sharply defined seismic discontinuity. In honour of Gutenberg, the core-mantle boundary is known as the Gutenberg seismic discontinuity.
Inga Lehmann (1936), a Danish seismologist, reported weak P-wave arrivals within the shadow zone. She interpreted this in terms of a rigid inner core with higher seismic velocity at depth of around 5154 km. Thus core has a radius of 3480 km and consists of a solid inner core surrounded by a liquid outer core.

**SEISMIC WAVES**

Seismic waves are classified into two groups: **body waves**, which travel through the earth in all directions and to all depths, and **surface waves**, whose propagation is limited to a volume of rock within a few seismic wavelengths of the earth's surface. There uses and analysis methods for the two types of waves are substantially different. Body waves are used for resource exploration purposes and for the study of earthquakes. Surface waves are used to delineate the layered-earth structure.

**Body Waves:-**

Types of body waves exist:

**Compression waves**

**Shear waves or S-wave**

**P-waves** are similar to sound waves. They obey all the physical laws of the science of acoustics. The mass particle motion of a P-wave is in the direction of the propagation of the wave. In addition P-waves cause a momentary volume change in the material through which they pass, but no concomitant momentary shape change occurs in the material.

**S-waves**, or **shear waves** as they are commonly called, move in a direction perpendicular to the direction of particle motion. Vertically and horizontally polarized S-waves are known as SV-wave and SH-wave, respectively. They are sometimes called secondary waves because they travel more slowly than P-waves in the same material. S-waves do not change the instantaneous volume of the materials through which they pass, but as they pass through materials, they distort the instantaneous shape of those materials. The velocity of S-wave is directly related to the shear strength of materials. S-waves do not propagate through fluids as those do not have any shear strength.

**Surface Waves** A disturbance at the free surface of a medium propagates away from its source partly as seismic surface waves. Surface waves, sometimes known as L-waves, are subdivided into Rayleigh (LR) and Love waves (LQ). These surface waves are distinguished from each other by the type of motion of particles on their wave fronts.

**Rayleigh waves** Lord Rayleigh (1885) described the propagation of Rayleigh wave along the free surface of semi-infinite elastic half-space. In the homogeneous half-space, vertical and horizontal components of particle motion are 90° out of phase in such a way that as the wave propagates the particle motion describes a retrograde ellipse in the vertical plane, with its major axis vertical and minor axis in the direction of wave propagation. The resulting particle motion can be regarded as a combination of P- and SV-vibrations.
In the case of a layered and Rayleigh wave ( ) LR. Dissipative medium, the path is always elliptical but not necessarily retrograde. Further, the axis of the ellipse may not be vertical and horizontal since the phase difference between vertical and horizontal displacement can be different from 90°. The velocity of Rayleigh wave is very much dependent on the Poisson's ratio and it is equal to 0.9194 times to that of S-wave in the Poisson's solid (Poisson's ratio = 0.25). The particle displacement is not confined entirely to the surface of the medium but the passes of the Rayleigh waves also displace the particle below the free surface up to a depth equal to the wavelength. In a uniform half space, the amplitude of particle displacement decreases exponentially with depth.

**Love waves**

Love explained the mechanism of generation of Love waves in horizontal soil layer overlying the half-space .When the angle of reflection at the base of soil layer is more than the critical angle, SH-waves are trapped in the soil layer. The constructive interference of reflected SH-waves from the top and bottom of the soil layer generate horizontally travelling Love waves. The particle motion is in horizontal plane and transverse to the direction of wave propagation. The velocity of Love wave lies between the velocity of S-wave in the soil layer and in the half-space. The velocity of Love wave with short wavelength is close to the velocity S-wave in soil layer and velocity of longer wavelength Love wave is close to the S-wave velocity in half-space. This dependence of velocity on wavelength is termed dispersion. Love waves are always dispersive, because they can only propagate in a velocity layered medium.

**Earthquake Magnitude**

Earthquake magnitude is a measure of the amount of energy released during an earthquake. Depending on the size, nature, and location of an earthquake, seismologists use different methods to estimate magnitude. Since magnitude is the representative of the earthquake itself, there is thus only one magnitude per earthquake. But magnitude values given by different.

**Details of MMI intensity scale**

**Intensity Descriptions**

seismological observatories for an event may vary. The uncertainty in an estimate of the magnitude is about ± 0.3 unit. Seismologists often revise magnitude estimates as they obtain and analyze additional data.

**Richter magnitude (ML)**

One of Dr. Charles F. Richter's most valuable contributions was to recognize that the seismic waves radiated by earthquakes could provide good estimates of their magnitudes. Richter (1935) collected the recordings of seismic waves from a large number of earthquakes and constructed a diagram of peak ground motion versus distance (Figure 1.11). The logarithm of recorded amplitude was used due to enormous variability in amplitude. Richter inferred that the larger the intrinsic energy of the earthquake, the larger the amplitude of ground motion at a given distance.

**A plot of log of peak amplitude in mm versus epicentral distance of earthquakes recorded in Southern California (different symbols represent different earthquakes).**

The idea of a logarithmic earthquake magnitude scale struck into the mind of Richter after analysing the roughly parallel curves generated by different size earthquakes on the plot of log of the recorded amplitude at various epicentral distances. The parallel nature of curves for different earthquakes suggested that a single number could quantify the relative size of different earthquakes. He proposed zero magnitude for an earthquake that would produce a record with amplitude of 1.0 mm at a distance of 100 km from the epicentre on Wood-Anderson (WA) seismograph with 1.25 Hz natural frequency and 2800 magnification factor. The logarithmic form of Richter magnitude scale (ML) is given as: $ML = \log_{10} A - \log_{10} A_0$ (1.1) where A0 is the amplitude for zero magnitude earthquakes at different epicentral distances and A is the recorded amplitude in mm. The zero magnitude amplitude can be computed for different epicentral distances taking into account the effects of geometrical spreading and absorption of considered wave.

The Richter scale used in Southern California for different epicentral distances and 18 km fixed focal depth is as follows.
ML = log10 A (mm) + Distance correction factor 's' Distance correction factor 's' is log of inverse of zero magnitude amplitude measured in mm at an epicentral distance 'D' in km. The distance correction factors for different epicentral distances are given The distance correction factors given in cannot be used in other regions of the world since considered focal depth was constant. So, to compute ML in any other region like Himalayas, first zero magnitude amplitude at different epicentral distances should be determined according to the original definition of ML at 100 km and different focal depths taking into account the geometrical spreading and appropriate measure of absorption. Since, sufficient time resolution of high frequency records is no longer a problem, therefore, frequency dependent distance correction factors, matched with Richter scale at 100 km distance, have been developed based on epicentral as well as hypo-central distances (Hutton and Boore, 1987; Kim, 1998; Langston et al., 1998).

**Surface wave magnitude**

As more seismograph stations were installed around the world, it became apparent that the method developed by Richter was strictly valid only for certain frequency and distance ranges. Further, at large epicentral distances, body waves are usually attenuated and scattered so that the resulting motion is dominated by surface waves. On the other hand, the amplitude of surface waves, in case of deep focus earthquakes is too small. So, in order to take advantage of the growing number of globally distributed seismograph stations, new magnitude scales that are an extension of Richter's original idea were developed. These include body-wave magnitude (mB) and surface-wave magnitude (MS). Each is valid for a particular period range and type of seismic wave.

A commonly used equation for computing MS of a shallow focus (< 50 km) earthquake from seismograph records between epicentral distances 20°< D < 160° is the following one proposed by Bath (1966). MS = log10 (As/T)max + 1.66log10 D + 3.3 (1.4)
Where AS is the amplitude of the horizontal ground motion in 'mm' deduced from the surface wave with period T (around 20 ± 2 seconds) and epicentral distance D is in degree.

**Body wave magnitude (mB)**
Gutenberg (1945) developed body wave magnitude mB for teleseismic body-waves such as P, PP and S in the period range 0.5 s to 12 s. It is based on theoretical amplitude calculations corrected for geometric spreading and attenuation and then adjusted to empirical observations from shallow and deep-focus earthquakes. mB =log10(A/T)max + s(D, h) Gutenberg and Richter (1956) published a table with distance correction factors s(D, h) for body waves, which enable magnitude determinations. These distance correction factors are used when ground motion trace amplitudes are measured in 'mm'.

**Duration magnitude (MD)**

Analogue paper and tape recordings have a very limited dynamic range of only about 40 dB and 60 dB, respectively. ML cannot be determined since these records are often clipped in case of strong and near earthquakes. Therefore, alternative duration magnitude scale MD has been developed. Duration from the P-wave onset to the end of the coda (back-scattered waves from bnumerous heterogeneities) is used in computations. Aki and Chouet (1975) reported that for a given local earthquake at epicentral distances lesser than 100 km the total duration of a signal is almost independent of distance, azimuth and property of materials along the path. This allows development of duration magnitude scales without a distance term.

MD = a0 + a1 log D (1.6)

a0 and a1 are constant and D is the duration in seconds. The values of these constants vary region to region according to crustal structure, scattering and attenuation conditions. They have to be determined locally for a region with the help of available ML

**Moment magnitude**

In case of large earthquakes, the various magnitude scales (ML, mB or MS) based on maximum amplitude and period of body waves or surface waves under estimate the energy released due to saturation.Recently, seismologists have developed a standard magnitude scale, known as moment magnitude. Moment magnitude is calculated using moment released during an earthquake rupture. The moment released depends on the physical dimension of the rupture (A), shear strength of the rock (m) and the average displacement on the fault plane (d). the strained fault just before the rupture. a couple of the shear forces acting on the either side of the fault are considered, '2b' distance apart. The moment of the couple (M0) is simply 'F.2b'. Now, if 'd' is the displacement, the strain (g) developed by the couple is 'd/2b'. The value of considered force can be obtained in terms of shear strength rock and area of rupture, using stress–strain relationship.

s = F/A = m◊g = m◊d/2b or F = m◊A◊d/2b

The rigidity 'm' is measured using samples of rock or is estimated from knowledge of the rocks in the area. Aftershocks are believed to reveal the rupture area because most of them lie on a plane. The simplest way to measure the length 'L' and average displacement'd' of a fault is to look at the newly faulted surface, or fault break. The seismic moment can also be estimated from the long period components of seismograms (Bullen and Bolt, 1985).

**References-**


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