1

Introduction to Earthquakes

1.1 A Historical Perspective

The earthquake is among the most dreaded of all natural disasters, exacting a devastating toll on human life. In the last 100 years alone there have been many major earthquakes, including San Francisco (1906), Tokyo (1923), Alaska (1964), Iran (1968), Mexico (1985) (Figure 1.3), Kobe (1995) (Figure 1.4) and Turkey (1998), to name but a few. The devastating tsunami of 26 December 2004, that struck the coastlines of the Indian Ocean, was caused by an underwater earthquake. Most recently the destructive force of the earthquake was felt in Sichuan Province, China where in May 2008 an estimated 87,000 were killed and 5 million lost their homes.

During these earthquakes hundreds of thousands of lives were lost and billions of dollars of damage sustained to property, and the physical suffering and mental anguish of earthquake survivors are beyond contemplation.

Several earthquakes, of the many that have occurred in the past century, have featured more prominently in terms of the development of the seismic theories and hazard mitigation procedures that we now have at this time. One of these earthquakes, which triggered a great interest in the scientific community, was the 'Great' 1906 San Francisco earthquake. The San Francisco earthquake (18 April 1906) measured between VII and IX on the Modified Mercalli Intensity scale and confounded geologists with its large horizontal displacements and great rupture length, along the northernmost 296 miles (477 km) of the San Andreas fault. The San Francisco earthquake was a momentous event of its era. More than 28,000 buildings were destroyed, damaged or affected by the earthquake (Figure 1.1) and casualties totalled approximately 3000. The earthquake is still noted for the raging fire it caused, which burned for almost three days (Figure 1.2). In its immediate aftermath a State Earthquake Investigation Commission, consisting of some of the most distinguished US geologists of the time, was appointed to bring together the work of scientific investigations and observations following the San Francisco earthquake and its final report is seen as a landmark document in terms of geological and seismological research. Henry Fielding Reid, Professor of Geology at John Hopkins University, Baltimore developed the 'Elastic Rebound Theory' (discussed later) from his studies of displacements and strain in the surrounding crust following the

Fundamentals of Seismic Loading on Structures Tapan Sen

Figure 1.1 Aftermath of the San Francisco earthquake (1906). Wreckage of the Emporium and James Flood Building on Market Street. Reproduced from http://www.sfmuseum.org 18/07/08, the Virtual Museum of the City of San Francisco.

Figure 1.2 San Francisco (1906) Huge crowds watch Market Street fire fight. Reproduced from http://www.sfmuseum.org 18/07/08, the Virtual Museum of the City of San Francisco.

earthquake. For the first time people began to realize the knock-on effects earthquakes have on our planet.

On average 17,000 persons per year were killed as a result of earthquake activity in the twentieth century. Among the deadliest earthquakes of the past are those occurring in China, including the Gansu (1920) and Xining (1927) whose death tolls were 200,000 each and the

Figure 1.3 Image of damage during the Mexico City (1985) earthquake. Reproduced from http://www.drgeorgepc.com/Tsunami1985Mexico.html 21/07/08 by permission of Dr George Pararas-Carayaniss.

Figure 1.4 A scene of destruction caused by Kobe earthquake (1995). Reproduced from http://.bristol.ac.uk/civilengineering/research/structures/eerc/realearthquakes/kobe-fire.gif 13/12/07 by permission of Dr A.J. Crewe, University of Bristol.

Tangshan (1976) which resulted in 255,000 deaths. The European earthquake with the highest loss of life was the Messina earthquake in Italy (1908), which resulted in 70,000 deaths.

The enormity of the scale of these earthquakes is justification enough to understand them and therefore to design structures that can most effectively resist them. First we must understand where and why earthquakes happen. Most severe earthquakes occur where the Earth's tectonic plates meet along plate boundaries. For example as two plates move towards each other, one plate can be pushed down under the other into the Earth's mantle. This is a destructive plate boundary and if the plates become locked together immense pressure builds up in the surrounding rocks. When this pressure is released shock waves are produced. These are called seismic waves and radiate outward from the source of the earthquake known as the epicentre, causing the ground to vibrate. Earthquakes are also very common on conservative plate boundaries, where the two plates slide past each other. So-called *intraplate* earthquakes occur away from plate boundaries, along fault zones in the interior of the plate.

The effects of earthquakes vary greatly due to a range of different factors, including the magnitude of the earthquake, and the level of population and economic development in the affected area. Earthquakes in seismic prone areas have an impact on our lives, on our buildings and the very landscape of our planet.

The severity of an earthquake is described both in terms of its *magnitude* and its *intensity*. These two frequently confused concepts refer to different, but related, observations. Magnitude characterizes the size of an earthquake by measuring indirectly the energy released. By contrast, intensity indicates the local effects and potential for damage produced by an earthquake on the Earth's surface as it affects population, structures, and natural features.

Basic consequences of earthquakes include collapsed buildings, fires (San Francisco, 1906), tsunamis (Indonesia, 2004) and landslides. Again these impacts depend on magnitude and intensity. Smaller tremors, that can occur either before or after the main shock, accompany most large earthquakes; these are termed foreshocks and aftershocks. Aftershocks can be felt from halfway around the world. While almost all earthquakes have aftershocks, foreshocks occur in only about 10 % of events. The force of an earthquake is usually distributed over a small area, but in large earthquakes it can spread out over the entire planet.

Since seismologists cannot directly predict when the next earthquake will happen, they rely on numerical experiments to analyse seismic waves and to accurately assess the magnitude and intensity of earthquakes. Such analyses allow scientists to estimate the locations and likelihoods of future earthquakes, helping to identify areas of greatest risk and to ensure the safety of people and buildings located in such hazardous areas.

1.1.1 Seismic Areas of the World

Figure 1.5 shows the epicentres of the 358,214 earthquakes that occurred in the years 1963–1998. Many places at which the earthquakes occurred, match with the boundaries of the tectonic plates. There are 15 major tectonic plates on Earth:

1. African Plate
2. Antarctic Plate
3. Arabian Plate
4. Australian Plate
5. Caribbean Plate
6. Cocos Plate
7. Eurasian Plate
8. Indian Plate
9. Juan de Fuca Plate

10. Nazca Plate
11. North American Plate
12. Pacific Plate
13. Philippine Plate
14. Scotia Plate
15. South American Plate

Figure 1.5 illustrates clearly that the Oceanic regions of the world have well-revealed seismicity following the lines of the plate boundaries. The major tectonic plates of the earth are shown in Figure 1.14.

As well as the major tectonic plates, there are many smaller sub plates known as *platelets*. These smaller plates often move which could be due to movement of the larger plates. The presence of platelets means that small but nonetheless damaging earthquakes can potentially occur almost anywhere in the world. These unexpected earthquakes, though less likely to occur, can be extremely destructive as in the 1960 Agadir earthquake in Africa, where none of the buildings in the affected area had been designed to be earthquake-resistant.

1.1.2 Types of Failure

Normally infrastructure damage during earthquakes is a result of structural inadequacy (Northridge, 1994), foundation failure (Mexico, 1985; Kobe, 1995), or a combination of both. Where foundation failure occurs the soil supporting the foundation plays an important role. The behaviour of foundations often depends on the soil's response to the shaking of the ground. (Figures 1.6 to 1.8)

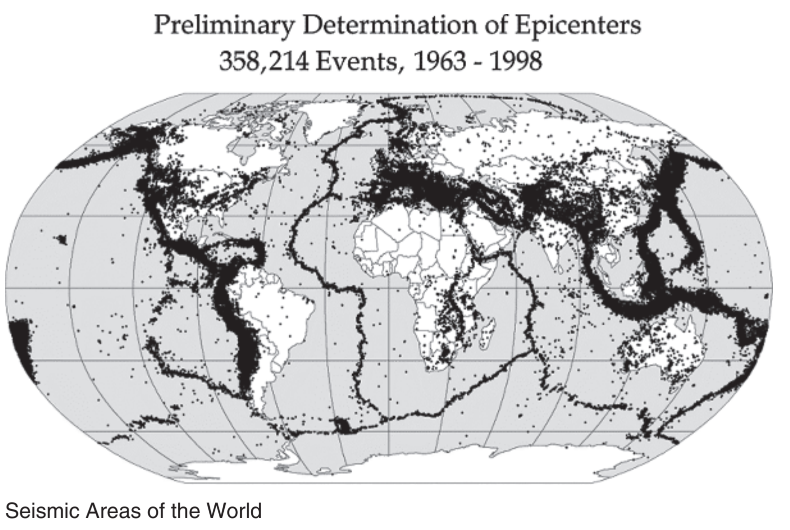


Figure 1.5 Epicentral Locations around the World (1963–1998). Courtesy of NASA. The NASA home page is http://www.nasa.gov 01/12/07.

The following examples of geotechnical damage to the built environment can be cited:

1. Failures of earth structures such as dams, embankments, landfill and waste sites. Failure of dams often causes flooding. During the Bhuj (India) earthquake of 2001 four dams (Fatehgarh, Kaswati, Suvi and Tapar) suffered severe damage. Fortunately their reservoirs were practically empty at the time of the earthquake and as a result there were no flood disasters.
2. Soil liquefaction resulting in widespread destruction to road networks and foundations. It is often observed that raft and piled foundations collapse without any damage to the superstructure.
3. Damage to underground or buried structures such as tunnels, box-culverts, underground storage facilities, buckling of pipelines, lifting of manholes etc. During the Turkish earthquake of 1999, the Bolu Tunnel suffered severe damage.
4. Damage to foundations due to large ground displacements owing to liquefaction-induced lateral spreading. Such foundations include underground retaining walls, quay walls and pile-supported wharfs.
5. Damage to foundations due to fault movement.

Some examples of damage to our environment due to geotechnical effects are shown in Figures 1.6 to 1.8.

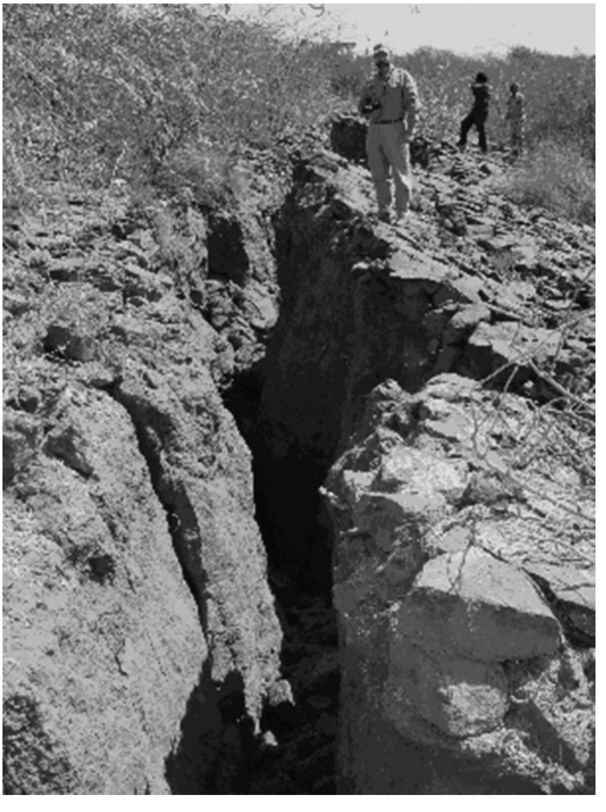


Figure 1.6 Damage to Fatehgarh dam. Reproduced from http://gees.usc.edu/GEER/Bhuj/image15.gif 18/07/08 by permission of NSF-sponsored GEER (Geo-Engineering Earthquake Reconnaissance) Association.

Figure 1.7 Liquefaction-induced damage. Image of leaning apartment houses following Niigata Earthquake, Japan 1964. Reproduced from http://www.ngdc.noaa.gov/seg/hazard/slideset/1/1_25_slide.shtml 01/06/08 by permission of the National Geophysical Data Center.

Figure 1.8 Subsidence of a running track by more than three metres. Reproduced from http://www.whfreeman.com/bolt/content/bt00/figure2.jpg 08/04/08 by kind permission of Beverley Bolt.

1.1.3 Fault Movement and its Destructive Action

Fault movements, which may occur, especially at plate boundaries are shown in Figure 1.9. Fault movements can be very destructive if there are structures passing through them, as

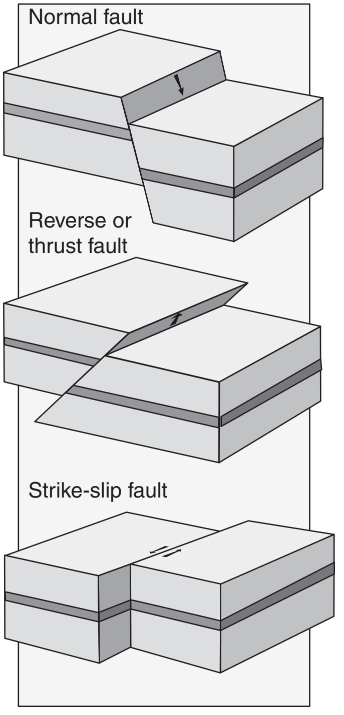


Figure 1.9 Fault categories. Courtesy of the U.S. Geological Survey. The USGS home page is http://www.usgs.gov 18/07/2007.

occurred in the 1999 Taiwanese and Turkish earthquakes. Faults are usually defined as a form of discontinuity in the bedrock and are associated with relative displacement of two large blocks of rock masses. Faults are broadly subdivided into three categories depending on their relative movement (Figure 1.9).

1. *Normal fault*: in normal faulting, one block (often termed as the hanging wall block) moves down relative to the other block (often termed as footwall block). The fault plane usually makes a high angle with the surface.
2. *Reverse fault* (also known as *thrust fault*): in reverse faulting, the hanging wall block moves up relative to the footwall block. The fault plane usually makes a low angle with the surface.
3. *Strike-slip fault*: in this fault, the two blocks move either to the left or to the right relative to one another.

1.2 The Nature of Earthquakes

As with many natural phenomena the origins of an earthquake are uncertain. In simplistic terms, earthquakes are caused by vibrations of the Earth's surface due to a spasm of ground shaking caused by a sudden release of energy in its interior. Seismologists have carried out

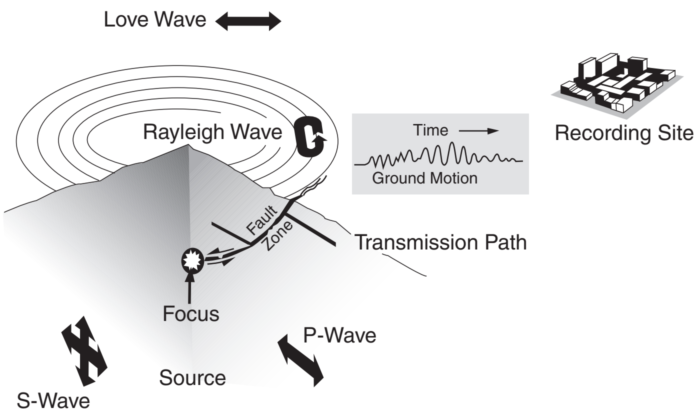


Figure 1.10 A schematic of the elements that affect ground motion.

studies of the geological aspects and changes that that have occurred inside the Earth over millions of years from which a broad picture of how earthquakes originate has emerged.

It is perhaps necessary to mention that tremors may be caused by other events, such as nuclear explosions or volcanic activity. However this book is concerned with earthquakes caused by disturbances deep inside the crust.

A schematic of the elements that affect ground motion is shown in Figure 1.10.

In the following sections some of the features of seismicity, the elements that affect ground motion, recording devices and evolution of the most important concepts of geology are explored.

1.3 Plate Tectonics

A cross section of the Earth is shown in Figure 1.11, illustrating the inner and outer core, mantle and crust, the crust itself being only 25–40 km in depth.

Various observations led the German meteorologist, Alfred Wegener, to expound the theory of *continental drift* in 1915. It was a revolutionary new idea and one of the founding principles of modern-day plate tectonics.

According to Wegener's theory, some 200–300 million years ago, the continents had formed a single land mass which slowly broke up and drifted apart. Wegener noticed that large-scale geological features on separated continents often matched very closely and that coastlines (for example, those of South America and Africa) seemed to fit together. Wegener also theorized that when continental land masses drifted together, crumpling and folding as they met, spectacular 'fold mountain' ranges such as the Himalayas and Andes were formed.

However, his fellow scientists found Wegener's theory hard to accept. This was mainly because he could not explain at that time *how* the continents had moved. His theory was finally shown to be right almost 50 years later. Scientists found (aided by modern recording instruments) that the sea floor is spreading apart in some places where molten rock is spewing out between two continents.

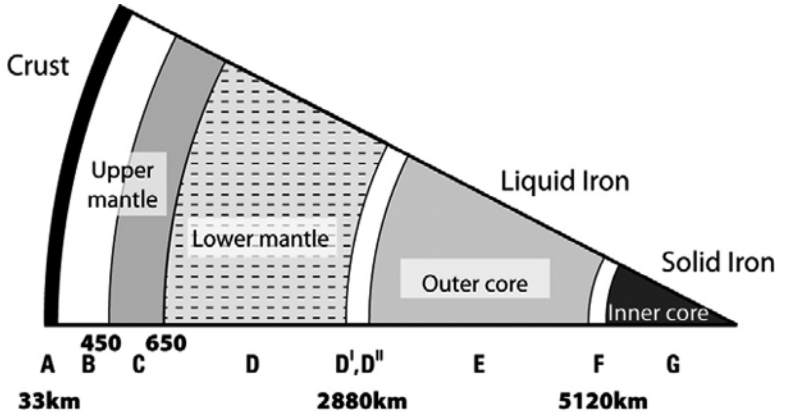


Figure 1.11 The Earth's structure, inner and outer core, mantle and crust. After Gubbins (1990) by permission of Cambridge University Press.

The theory thus gained acceptance by the scientific community and was hailed as one of the most revolutionary advances ever made in the field of geology from which the discipline of plate tectonics emerged. It is simple and elegant and has broad predictive power.

The basic principle of this discipline is that the Earth's crust consists of a number of plates that move with respect to each other floating on the underlying molten mantle. The movement is exceedingly slow on a geological scale but gives rise to enormous forces at the plate boundaries.

The movement of the plate boundaries is directly related to changes or reactions taking place inside the Earth. An accepted explanation of the source of plate movements is that the movement is governed by the requirement of thermo mechanical equilibrium of the mass inside the Earth. The upper portion of the mantle is in contact with the cooler crust whilst the lower portion is in contact with the hot outer core (Figure 1.12). A temperature gradient exists within the mantle, and leads to a situation where the cooler and dense crust rests on a less dense but warmer material. It gives rise to a situation where convection currents are set up within the mantle (Figure 1.12) which drives the plate movements. The cooler, denser material at the top begins to sink and the warmer less dense materials underneath rise. The sinking cooler material warms, becomes less dense, moves laterally and tries to rise, paving the way for subsequently cooled material to sink again; and this cycle would continue.

1.3.1 *Types of Plate Boundaries*

Plate tectonics is conceptually quite simple. The outer shell consists of 15 major plates about 100 km thick. The plates move relative to each other at very slow speeds (a few cm per year). The plates are considered rigid; there is little or no deformation within them. The deformation occurs at the boundaries. The outer strong shell forms the Earth's *lithosphere* and the movement takes place over the weaker layer called the *asthenosphere*. The types of movement at the plate boundaries are shown in Figure 1.13.

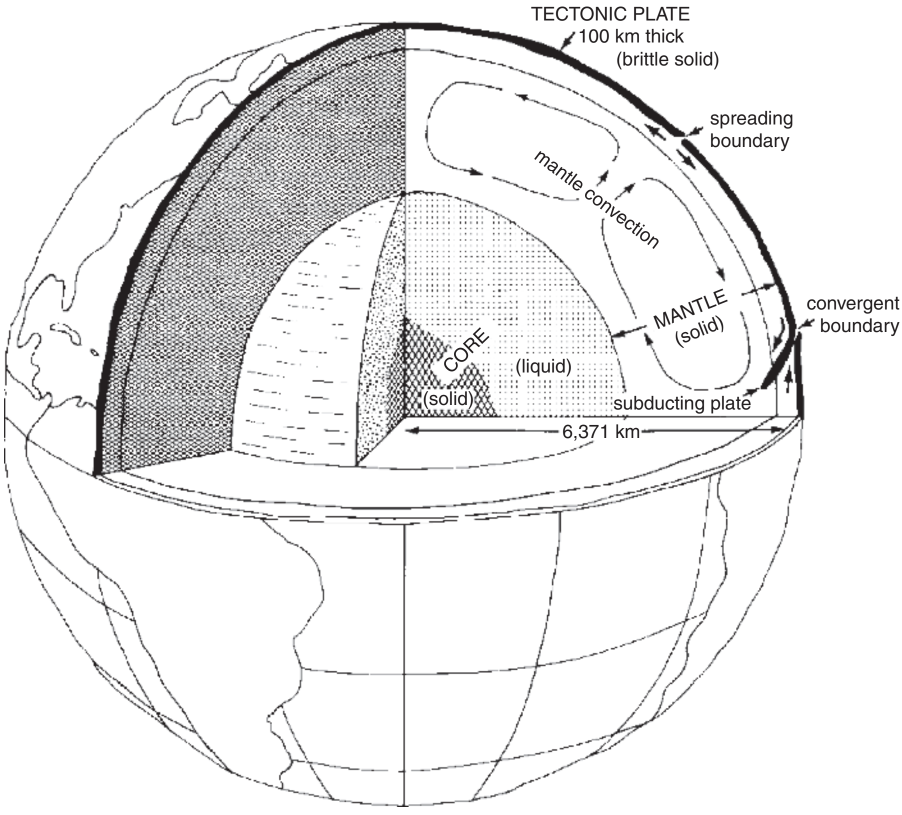


Figure 1.12 Convection currents in mantle. Near the bottom of the crust, horizontal components of convection currents impose shear stresses on bottom of crust, causing movement of plates on the Earth's surface. The movement causes the plates to move apart in some places and to converge in others. Reproduced from Noson *et al*. (1988) by permission of the Washington Division of Geology and Earth Resources.

1.3.2 Convergent and Divergent Boundaries

Three basic types of plate boundaries are as follows:

1. *Divergent* boundaries: where two plates are moving apart and new lithosphere is produced or old lithosphere is thinned. *Mid-oceanic ridges* (also known as *spreading centres*).
2. *Convergent* boundaries: where lithosphere is thickened or consumed by sinking into the mantle. *Subduction zones* and *alpine belts* are examples of convergent plate boundaries.
3. *Transcurrent* boundaries: where plates move past one another without either convergence or divergence. *Transform faults* and other strike-slip faults are examples of transcurrent boundaries.

At spreading centres both plates move away from the boundary. At subduction zones the subducting plates move away from the boundary. In general, divergent and transcurrent

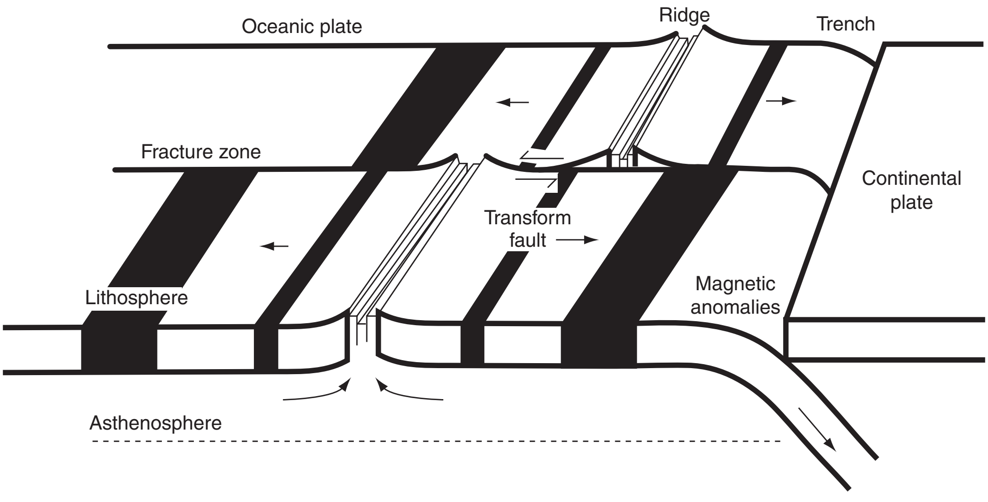


Figure 1.13 Types of movement at plate boundaries. Oceanic lithosphere is formed at ridges and subducted at trenches. At transform faults, plate motion is parallel to the boundaries. Each boundary has typical earthquakes. Reproduced from Stein and Wysession (2003) by permission of Blackwell Publishing.

boundaries are associated with shallow seismicity (focal depth less than 30 km). Subduction zones and regions of continental collision can have much deeper seismicity (Lay and Wallace, 1995).

Seismologists' study of plate tectonics points towards the following nature of plate behaviour and characteristics. Warm parts of the mantle material rise at *spreading centres* or mid-ocean ridges, and then cool. The cooling material forms strong plates of new oceanic lithosphere. The newly formed oceanic lithosphere while cooling down moves away from the ridges and eventually reaches the subduction zones or trenches, where it descends in down-going slabs back into the mantle, reheating during the process. At a common point on the two boundaries, it is the direction of relative motion that will determine the nature of the boundary.

1.3.3 Seismicity and Plate Tectonics

Crustal deformations occur largely at plate boundaries; when the stresses at these boundaries exceed the strength of the plate boundary material, strain energy is released, causing a tremor. If strain energy accumulates within the plate the consequent release of energy results in a full-blown earthquake.

The theory of plate tectonics has established that there are movements at plate boundaries. The movement between the portions of the crust occurs at new or pre-existing discontinuities in the geological structure known as faults. These may vary in length from several metres to hundreds of kilometres. Generally the longer the fault the larger the earthquake it is likely to generate. Figure 1.14 shows the major tectonic plates, mid-ocean ridges, trenches, and transform faults of the earth.

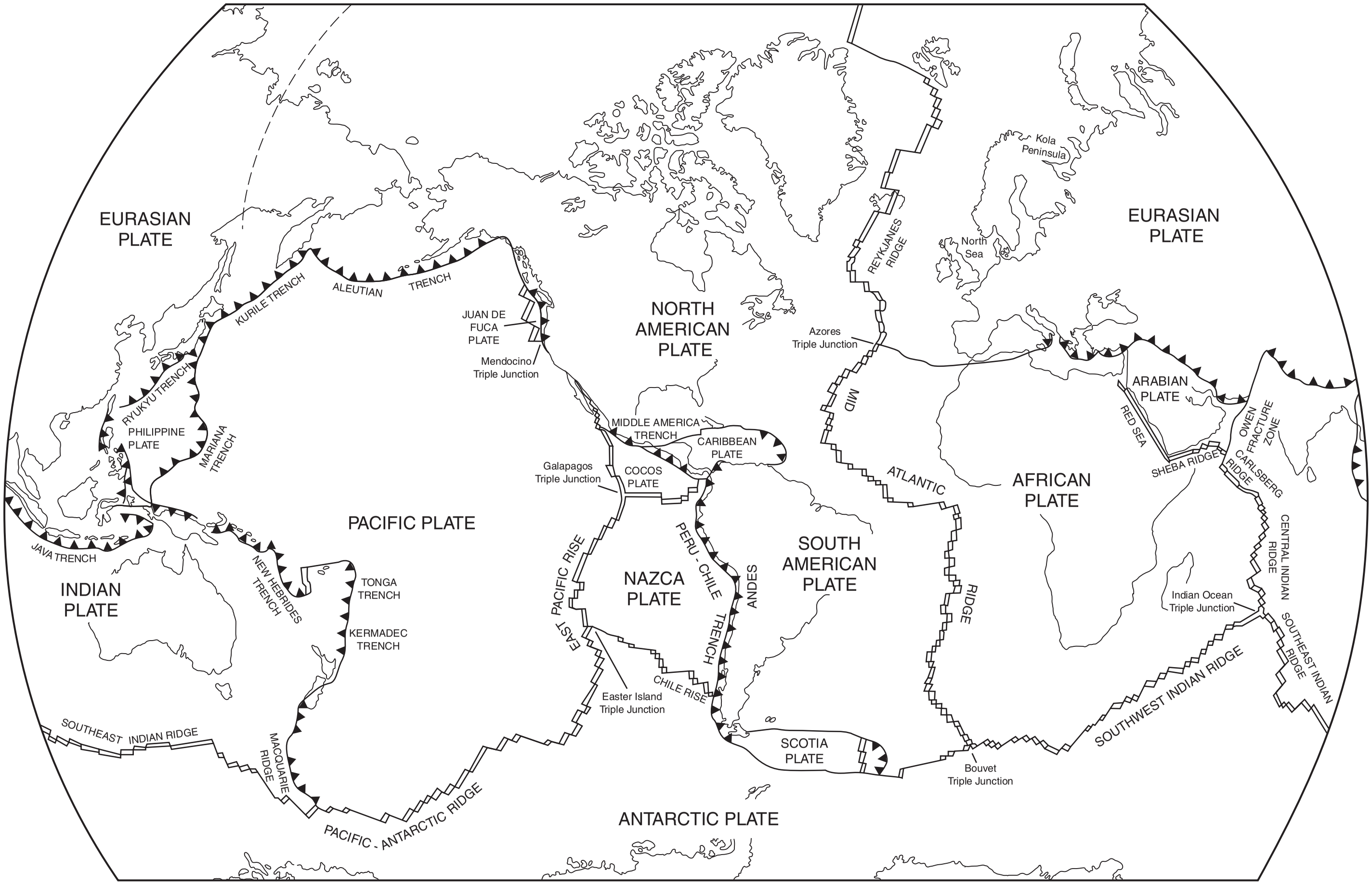


Figure 1.14 The major tectonic plates, mid-ocean ridges, trenches, and transform faults of the earth. Arrows indicate direction of plate movements. Reproduced from Fowler (1990) by permission of Cambridge University Press.

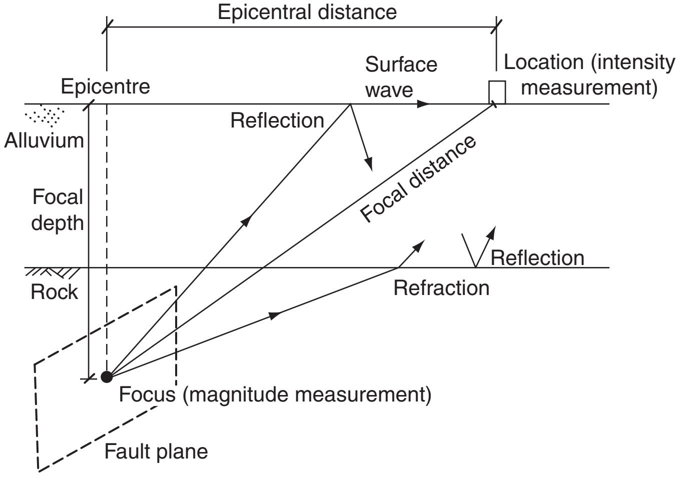


Figure 1.15 The focus and epicentre of an earthquake. Reproduced from Stein and Wysession (2003) by permission of Blackwell Publishing.

The striking similarity between Figure 1.5 and Figure 1.14 may be noted. Only in the 1960s were seismologists able to show that 'focal mechanism' (the type of faulting inferred from radiated seismic energy) of most global earthquakes is consistent with that expected from plate tectonic theory.

1.4 Focus and Epicentre

The source of an earthquake within the crust is commonly termed the *focus* or *hypocentre*. The point on the Earth's surface directly above the focus is the *epicentre*. The distance on the ground surface between *any site of interest* (for example a recording station) and the epicentre is the epicentral distance and the distance between the focus and the *site* is called the *focal distance* (shown in Figure 1.15). The distance between the focus and epicentre is called the *focal depth*.

1.5 Seismic Waves

Earthquakes generate elastic waves when one block of material slides against another, the break between the two blocks being called a 'fault'. Explosions generate elastic waves by an impulsive change in volume in the material.

If the equilibrium of a solid body like the earth is disturbed due to fault motion resulting from an earthquake or explosion seismic (elastic) waves are transmitted through the body in all directions from the focus. Earthquakes radiate waves with periods of tenths of seconds to several minutes. Rocks behave like elastic solids at these frequencies. Elastic solids allow a variety of wave types and this makes the ground motion after an earthquake or explosion quite complex.

1.5.1 Body Waves

Two categories of seismic body waves are produced during an earthquake:

- primary waves (P)
- secondary waves (S)

P and S waves travel through the interior of the earth from the focus to the surface. That is why they are called *body waves*. The velocities encountered depend upon the elastic constants and densities of the materials and other properties of the surrounding medium.

The first waves to arrive are the P (‘pressure’ or compression) waves (Figure 1.16a). The name P wave has its roots in the Latin *primus*. (‘first’), since they are the first waves to arrive, having the highest velocity.

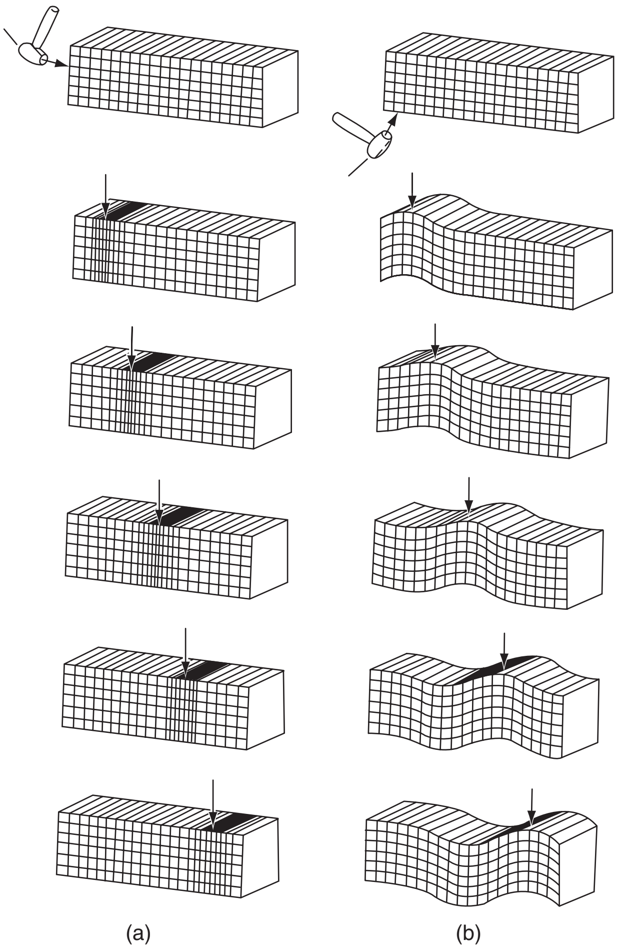


Figure 1.16 Primary (P) and Secondary (S) waves. Reproduced from Doyle (1996) by permission of John Wiley & Sons, Ltd.

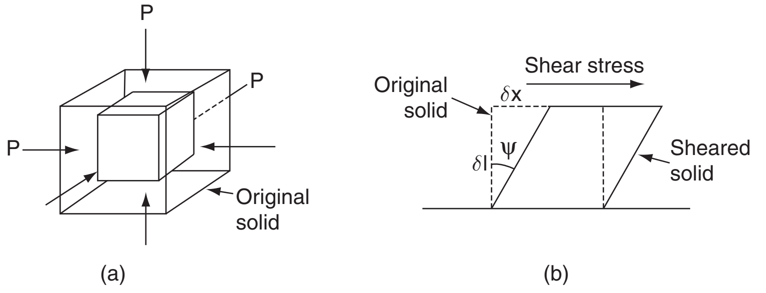


Figure 1.17 (a) Deformation by change of volume due to compression; (b) shear strain for case of block with lower side fixed. Reproduced from Doyle (1996) by permission of John Wiley & Sons, Ltd.

The second waves to arrive are the S (*secundus)* which have a transverse, shear vibration in a plane perpendicular to the direction of propagation (Figure 1.16b). The presence of two types of wave arises from the fact that there are two fundamental ways one can strain a solid body: (1) by volume change without change of shape (pure compression or expansion) Figure 1.17a; (2) by change of shape without change of volume (a shear distortion) (Figure 1.17b).

The P or compression waves transmit pressure changes through the Earth in a series of alternating compressions and rarefactions. Since they are the first waves to arrive, P waves can be recorded very accurately and are most commonly used in earthquake location and other related fields of seismic exploration. Their velocity is given by Doyle (1995):

$$V_p = \frac{(k + 4/3\mu)^{1/2}}{(\rho)}$$

Where

k is the bulk modulus
μ is the rigidity (Shear Modulus)
ρ is the density

(It may be noted that strains produced in the rock due to passage of seismic waves are normally very small (of the order of 10^{-6}), hence linear relationships between stress and strain may be assumed).

The S waves arrive later, having a velocity about 60 % of that of the P waves. The S (shear) wave velocity is given by:

$$V_s = (\mu/\rho)^{1/2}$$

S waves can only travel through solids, therefore cannot travel through the Earth's liquid outer core. Seismologists can extract valuable information about the makeup of the interior of the Earth and possible fluid content, as fluids have no shear strength. That is why it is believed that the outer core of the earth is fluid and almost certainly mainly liquid iron as has been deduced from geochemical and magnetic data (Doyle, 1995). P waves can travel through the core. An S_v wave is one in which the ground motion (vibration) is vertical and an S_h wave refers to one where the ground motion is horizontal (side to side).

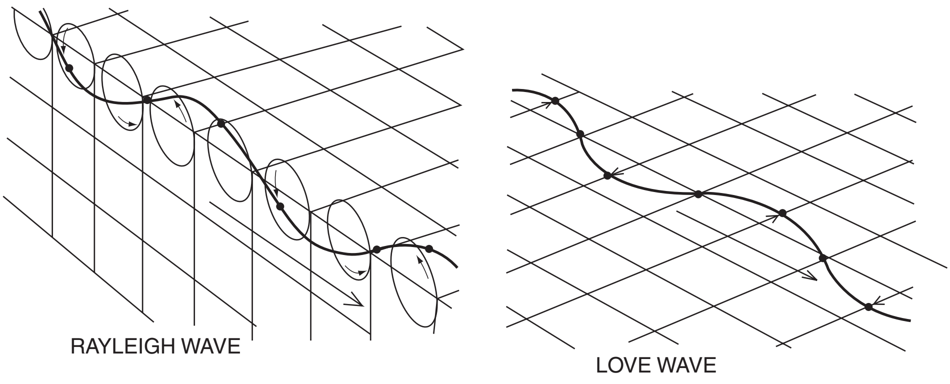


Figure 1.18 Surface waves: Love waves and Rayleigh waves. Reproduced from Doyle (1995) by permission of John Wiley & Sons, Ltd.

1.5.2 Surface Waves

When the two types of body wave reach the surface of the Earth, an interesting change occurs in the behaviour of the waves. The combination of the two types of wave in the presence of the surface leads to other types of waves, two of which are important for geophysics:

- Rayleigh waves
- Love waves

These are surface waves as distinct from body waves and produce large amplitude motions in the ground surface. They decay at a much slower rate than body waves and hence result in maximum damage. Surface waves are more destructive because of their low frequency, long duration and large amplitude.

The two types of surface waves are shown in Figure 1.18.

Rayleigh waves, also called ‘ground roll’, are the result of interaction between P and S_v waves and are analogous to ocean waves. The existence of these waves was first demonstrated by the English physicist, John William Strutt (Lord Rayleigh) in 1885. Rayleigh waves may be clearly visible in wide open spaces during an earthquake.

Love waves (Figure 1.18) are the result of interaction between P and S_h waves. Their existence was first deduced by the British mathematician, A.E.H. Love, in 1911 and they travel faster than Rayleigh waves.

1.6 Seismometers

Seismic waves are measured and recorded using the seismograph. The seismograph assists seismologists, geologists and scientists in the measurement and location of earthquakes. The instrumentation must above all be able to: (1) detect transient vibration within a moving reference frame (the pendulum of the instrument will be stationary as the Earth moves); (2) operate continuously with a detection device that is able to record accurately ground motion variation with time producing a seismogram; and (3) for instrument calibration purposes have a fully known linear response to ground motion; this would allow seismic recording to be accurately related to the amplitude and frequency content of the recorded ground motion. Such a recording system is called a *seismograph* and the actual ground motion sensor that converts ground motion into some form of signal is called a *seismometer*.

The basic components which make up most seismographs are:

- a frame lodged in the Earth, sometimes the most expensive part of the device;
- an inertial mass suspended in the frame, using springs or gravity to create a stable reference position;
- a damper system to prevent long-term fluctuation after reading an earthquake;
- a way of recording the motion or force of the mass, in relation to the frame.

1.6.1 Early Seismographs

John Milne is credited with the invention of the modern seismograph in 1880 while based at the Imperial College of Engineering in Japan. A surviving picture of Milne's device is shown below in Figure 1.19.

In the late nineteenth century most seismographs were developed by a team consisting of John Milne, T.A. Ewing and others working in Japan (1880–1895). These instruments consisted of a large stationary pendulum with a stylus. During an earthquake, as the ground

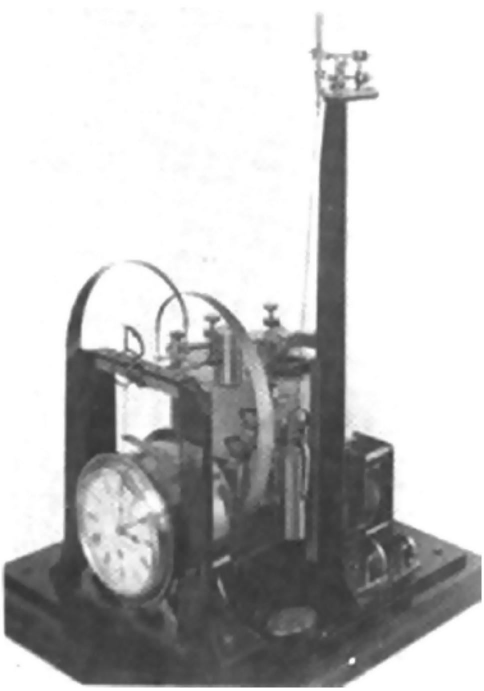


Figure 1.19 John Milne's device. http:/tremordeterra.blogspot.com 01/08/08

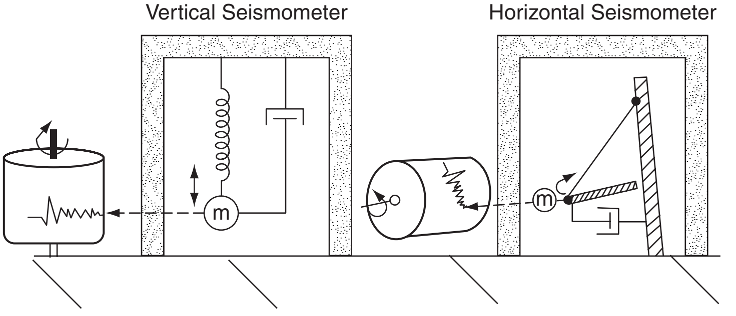


Figure 1.20 A schematic of vertical and horizontal seismometers. Actual ground motions displace the pendulums from their equilibrium positions, inducing relative motions of the pendulum masses. The dashpots represent a variety of possible damping mechanisms. Mechanical or optical recording systems with accurate clocks are used to produce the seisograms. Reproduced from Lay and Wallace (1995), Academic Press imprint by permission of Elsevier.

moved, the heavy mass of the pendulum remained stationary due to the inertia. The stylus at the bottom records markings which correspond to the Earth's movement.

Almost all seismographs are based on damped inertial-pendulum systems of one form or another. A schematic of simple vertical and horizontal instruments developed is shown in Figure 1.20. Passing seismic waves move the frame, while the suspended mass tends to stay in a fixed position. The seismometer measures the relative motion between the frame and the mass.

Two seismic instruments in the classic mould, developed around the turn of the earlier twentieth century are shown in Figure 1.21.

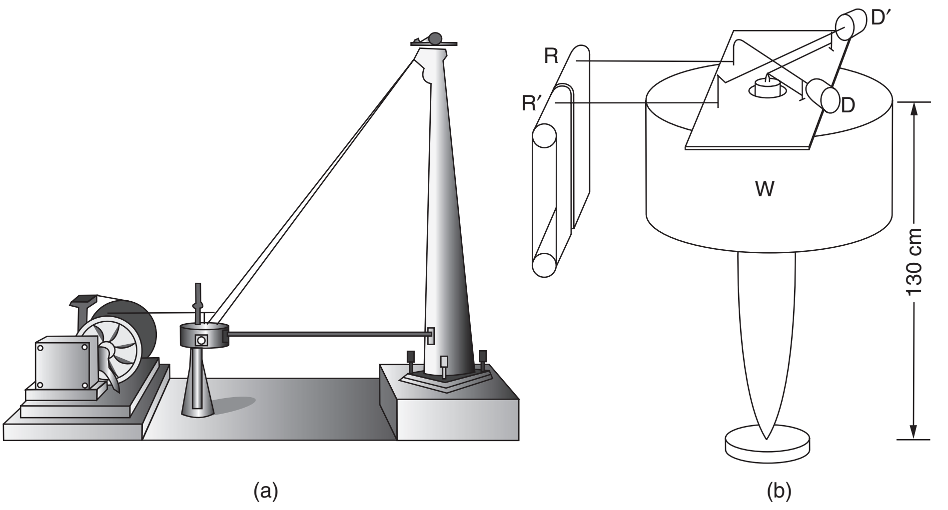


Figure 1.21 Early mechanical horizontal-motion seismographs: (a) the 1905 Omori 60-s horizontal-pendulum seismograph and (b) the 1904 1000-kg Wiechert inverted pendulum seismograph. Reproduced from Lay and Wallace (1995), Academic Press imprint by permission of Elsevier.

The Omori seismograph was developed by a student of John Milne in Japan. The instrument had direct response to ground displacement for periods of less than 60 s. The only damping in the system was due to the stylus friction in the hinges. The restoring force acting on the mass was gravity. It is interesting to note that the Omori instrument recorded the 1906 San Francisco earthquake which provided valuable data for further scientific research.

In 1898 Emil Wiechert, Professor of Geophysics at Gottingen University introduced viscous damping in his horizontal pendulum instrument. Figure 1.21 shows the version introduced in 1904 with 1000 kg mass at the top. Mechanical levers magnified the signal up to 200 times and pistons provide the damping. The Wiechert inverted-pendulum system has been in operation for more than 90 years.

Both instruments etched a record on smoked-paper recorders. Friction on the stylus provided the only damping in the Omori system, while air pistons (D and D′) damped the Wiechert instrument. Restoring springs connected to the mass, W, kept the inverted pendulum in equilibrium, with a special joint at the base of the mass permitting horizontal motion in any direction.

If the ground motion frequency is much higher than the natural frequency of the instrument ($\omega. >> \omega_0$) displacement on the seismometer is directly proportional to the ground displacement. Many of the early efforts in the development of seismometers focused on reducing ω_0 to yield displacement recordings for regional-distance seismographs.

1.6.2 Modern Developments

Earthquake design practices did not progress significantly until strong motion recording instruments were invented. Historical records can provide us with written descriptions of earthquake damage dating back 2000 years. However they are of little help in the development of appropriate methods of design or quantitative assessment of hazard.

There were basic forms of seismograph in use which were able to record the Great San Francisco earthquake in 1906 which was picked up as far away as Japan. The first measurements on a relatively modern seismograph of strong ground motion were recorded during the Long Beach (California) earthquake in 1933 and measurement techniques have advanced exponentially since then.

A recent development is the teleseismometer, a broadband instrument for registering seismographs which can record a broad range of frequencies. A servo feedback system keeps the mass motionless and the relative movement of the frame is measured by electronic displacement transducer with minimum force applied to the mass.

Different types of seismograph have been developed, including a digital strong-motion seismograph, also known as an *accelerograph*. The data produced from these instruments is essential to understanding how an earthquake can impact on built structures. A strong motion seismograph measures acceleration and this can be mathematically integrated to give velocity and displacement.

Strong-motion seismometers are not as sensitive to ground motion as teleseismometers but they stay on scale during the strongest seismic shaking.

A seismic recording centre is shown in Figure 1.22 and an example of a modern seismic recording in Figure 1.23.

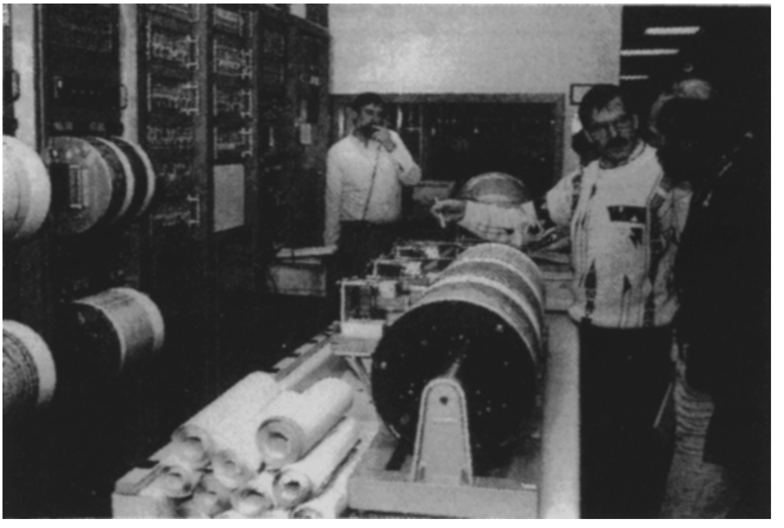


Figure 1.22 A seismic recording centre at the University of Alaska. Geophysical Institute in Fairbanks, Alaska. The data is recorded digitally on a mainframe and on a PC and displayed on the recording drums. Photo by James Kocia, Courtesy of the US Geological Survey. The USGS home page is http://www.usgs.gov 18/07/2007.

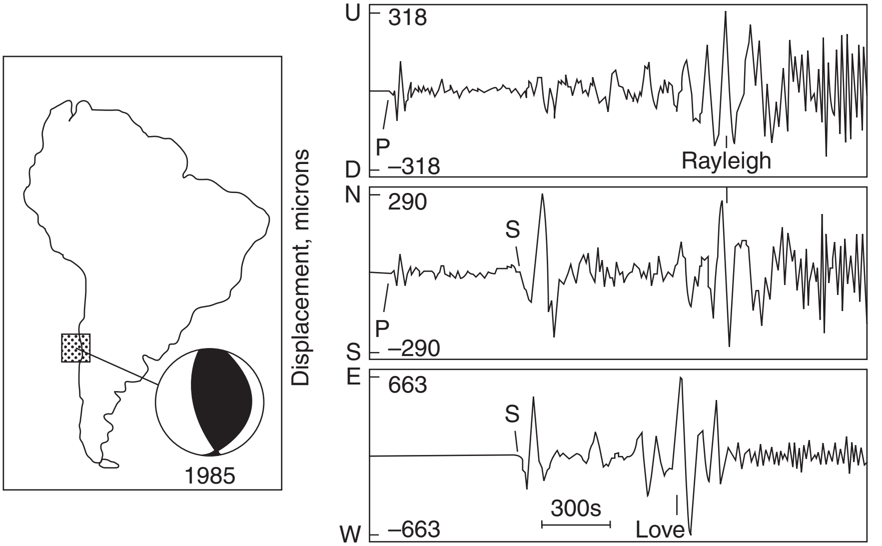


Figure 1.23 Recordings of the ground displacement history at station HRV (Harvard Massachusetts) produced by seismic waves from the 3 March 1985 Chilean earthquake, which had the location shown in the inset. Reproduced from Lay and Wallace (1995), Academic Press imprint by permission of Elsevier.

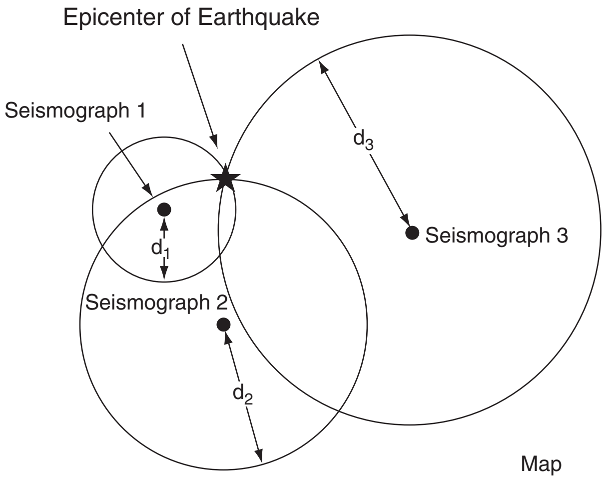


Figure 1.24 The intersection point between three seismographs, the location of the epicentre. Reproduced from http://www.tulane.edu/~sanelson 17/01/08 by permission of Professor Stephen A. Nelson, Department of Earth and Environmental Sciences, Tulane University, USA.

The three seismic traces correspond to vertical [U–D], north–south [N–S], and east–west [E–W] displacements. The direction to the source is almost due south; so all horizontal displacements transverse to the ray path appear on the east–west component. The first arrival is a *P* wave that produces ground motion along the direction of wave propagation. The Love wave occurs only on the transverse motions of the E–W component, and the Rayleigh wave occurs only on the vertical and north–south components.

1.6.3 Locating the Epicentre

The location of an earthquake is usually reported in terms of its epicentre. P (primary) waves travel at a faster rate than S (secondary) waves and are consequently the first waves to arrive at a seismic recording station (where the seismograph is located) followed by S waves. Preliminary identification of the epicentre location is based on the relative arrival times of P and S waves at the recording station. Readings from at least three recording stations are obtained to calculate the exact location of the epicentre (Figure 1.24).

1.7 Magnitude and Intensity

To be of use for engineering it is critical to know the size of an earthquake. The size of the earthquake can be measured in terms of *magnitude* and *intensity*.

- *Magnitude* is the amount of energy released from the source or focus of the earthquake.
- *Intensity* is the impact of ground-shaking on population, structures and the natural landscape; the impact will be greater nearer the site and less far away.

To measure these characteristics seismologists use two fundamentally different but equally important types of scale, the Magnitude scale and the Intensity scale.

1.7.1 Magnitude Scales

The magnitude of an earthquake is related to the amount of energy released by the geological rupture causing it and is therefore a measure of the absolute size of the earthquake without reference to the distance from the epicentre.

The best-known measure of earthquake magnitude was introduced by Charles Richter in the 1930s and became known as the *Richter Scale*, now referred to as the *local magnitude* (M_L). Richter was motivated by his desire to compile the first catalogue of Californian earthquakes and also saw the need for an objective size measurement to assess earthquakes' significance. Richter observed that the logarithm of maximum ground motion decayed with distance along parallel curves for many earthquakes. All the observations were from the same type of seismometer, a simple Wood-Anderson torsion instrument.

From his original recordings Richter expressed the magnitude scale as:

$$M_L = \log_{10} A(\Delta) - \log_{10} A_0(\Delta),$$

where A is the maximum trace amplitude for a given earthquake at a given distance as recorded by a Wood-Anderson instrument and A_0 is that for a particular earthquake selected as reference.

As may be seen it is on a logarithmic scale expressed in ordinary numbers and decimals. For example, the magnitude could be expressed as 4.3 on the Richter scale and the higher the number, the greater the damage.

Though M_L in its original form is rarely used today as Wood-Anderson torsion instruments are uncommon, it remains a very important magnitude scale because it was the first widely used 'size measure' and all other magnitude scales are tied to it. Further, M_L is useful for engineering. Many structures have natural periods close to that of a Wood-Anderson instrument (0.8 s) and the extent of earthquake damage is closely related to M_L (Lay and Wallace, 1995).

Although useful in establishing *local magnitude*, Wood-Anderson seismometers cease to be helpful for shocks at distances beyond 1000 km, hence the term *local magnitude*. It helps to distinguish it from magnitude measured in the same way, but from recordings on long-period instruments, which are suitable for more distant events. Beyond about 600 km the long-period seismographs of shallow earthquakes are dominated by surface waves (M_s) usually with a period of approximately 20 s. Surface wave amplitudes are strongly dependent on the source depth. Deep earthquakes (> 45 km) do not generate much surface-wave amplitude and there is no appropriate correction for source depth.

Partly to overcome this problem and also to be applicable for shallow and deep earthquakes, Gutenberg proposed what he called the 'unified magnitude' denoted by m or m_b, which is dependent on body waves and is now generally referred to as *body wave magnitude* m_b. This magnitude scale is particularly appropriate for seismic events with a focal depth greater than 45 km.

The three magnitude scales (M_L, m_B *and* M_S) are interrelated by empirical formulae. Gutenberg and Richter (1956) reported the following empirical equations:

$$m_b = 0.63M_s + 2.5$$
$$M_s = 1.27(M_L - 1) - 0.016M_L^2.$$

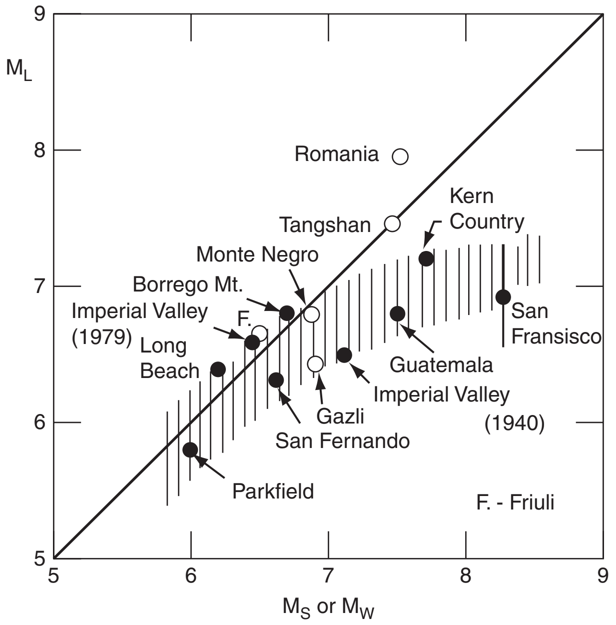


Figure 1.25 Effects of magnitude saturation for high frequency magnitude M_L versus the lower frequency magnitudes M_s or M_w (Note M_w is tied to M_s). Reproduced from Kanamori (1977), courtesy of the American Geophysical Union (AGU).

Since then several such equations have been published relating the different magnitudes (Utsu, 2002).

Both M_s and m_b were designed to be as compatible as possible with M_L; thus at times all three magnitudes give the same value for an earthquake. However this is rare. In all three cases we are making frequency dependent measurement of amplitudes at about 1.2, 1.0 and 0.05 Hz for M_L, m_b and M_s. Because of the frequency ω (or period) at which m_b is measured, earthquakes above a certain size will have a constant m_b (Lay and Wallace, 1995). This is referred to as *magnitude saturation*. Figure 1.25 shows measured values of M_L and M_s for several different earthquakes. It is clear that M_L begins to saturate at about magnitude 6.5, but for rare examples such as Tangshan and Romania, M_L does not saturate. A magnitude measure that does not suffer from this saturation deficiency is preferable.

1.7.2 Seismic Moment

Seismic moment, which characterizes the overall deformation at the fault, is considered by many researchers to be more accurate in calculating the earthquake's source strength. The advantage of seismic moment is that it can be interpreted simply in terms of the ground deformation. The moment is proportional to the product of the area of dislocation and the displacement across the fault. It is possible to relate seismic moment M_0, to seismic energy released. Also, surface wave magnitude M_s has been expressed (empirical relationship) in terms of seismic energy and thus M_0 and M_s are tied (Lay and Wallace, 1995).

The seismic moment, is estimated from the expression:

$$M_0 = G \cdot A \cdot d$$

where:

G is the shear modulus of the medium,
A is the area of the dislocation or fault surface,
d is the average displacement of slip on that surface.

It is possible at times to estimate d and A from field data and aftershock areas, but M_0 is usually estimated from the amplitudes of long period waves at large distances, corrected for attenuation, directional effects etc. (Doyle, 1995).

Hanks and Kanamori (1979) and Kanamori (1983) have, using certain assumptions, suggested a new magnitude scale, *moment magnitude* M_w, based on seismic moment:

$$M_w = \log_{10}(M_0/1.5) - 10.7$$
$$(M_0 \text{ in dyne-cm; } 10^5 \text{dyne} = 1\text{N, thus} 10^7 \text{dyne} - \text{cm} = 1\text{N} - \text{m}).$$

This scale has the important advantage that it does not saturate and the same formula can be used for shallow and deep earthquakes.

The range of seismic phenomenon in the earth is indicated in Figure 1.26.

From the above discussions it is clear how important it is to know the type of magnitude that is being used. In this book the local (i.e. Richter) magnitude (M_L) is used unless noted otherwise.

1.7.3 Intensity Scales

The intensity of an earthquake is a measurement of the observed damage at a particular location. This intensity will vary with distance from epicentre and depend on local ground conditions. It must be emphasized that intensity is a qualitative description of the effects of an earthquake at a particular site and this concept has proved very useful in interpreting historical data. Interpreting historical earthquake data can help in establishing the location, recurrence rates, and size of earthquakes in areas where no seismographs were installed at the time of the earthquake.

The Rossi-Forel (RF) scale of intensity (values ranging from I to X) was introduced in the late nineteenth century. The RF scale was in use for many years until its replacement in English-speaking countries by the Modified Mercalli Intensity (MMI) scale in 1931 (Table 1.1).

The MMI scale has twelve grades denoted by Roman numerals I–XII which describe the physical effects of earthquakes in a way similar to the Beaufort scale of wind strength.

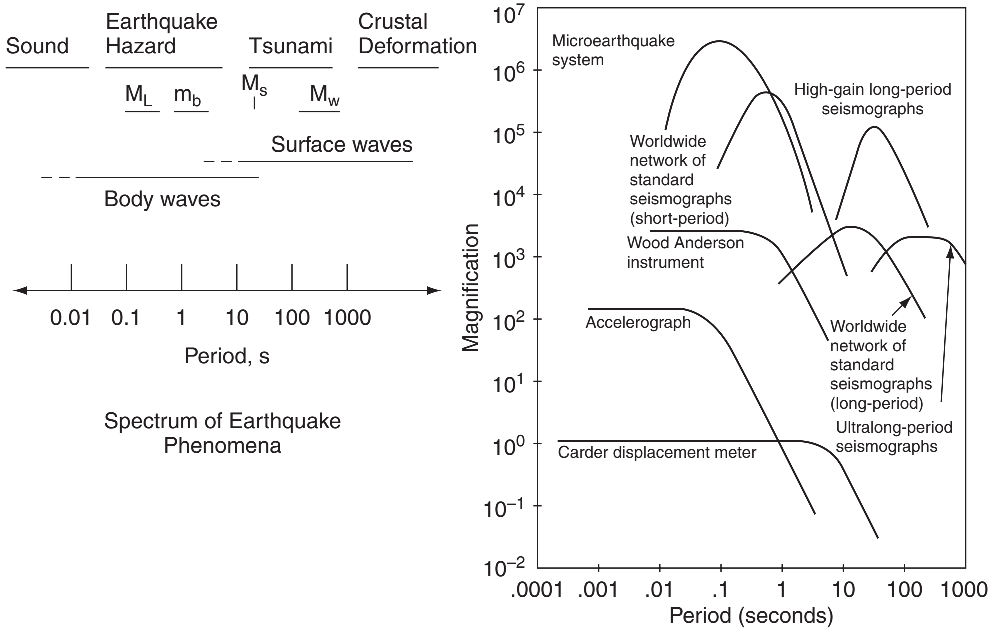


Figure 1.26 The range in period of seismic phenomenon in the Earth is shown on the left along with the characteristics: period of body waves; surface waves; and different seismic magnitude scales. On the right the amplitude responses of some major seismometer systems are shown. Each magnitude scale is associated with a particular instrument type; for example the Richter magnitude M_L is measured on the short period Wood-Anderson instrument. Reproduced from Kanamori (1988), Academic Press imprint by permission of Elsevier.

Table 1.1 Modified Mercalli Intensity Scale.

I. Not felt except by a very few under especially favourable conditions.
II. Felt only by a few persons at rest, especially on upper floors of buildings.
III. Felt quite noticeably by persons indoors, especially on upper floors of buildings. Many people do not recognize it as an earthquake. Standing motor cars may rock slightly. Vibrations similar to the passing of a truck. Duration estimated.
IV. Felt indoors by many, outdoors by few during the day. At night, some awakened. Dishes, windows, doors disturbed; walls make cracking sound. Sensation like heavy truck striking building. Standing motor cars rocked noticeably.
V. Felt by nearly everyone; many awakened. Some dishes, windows broken. Unstable objects overturned. Pendulum clocks may stop.
VI. Felt by all, many frightened. Some heavy furniture moved; a few instances of fallen plaster. Damage slight.
VII. Damage negligible in buildings of good design and construction; slight to moderate in well-built ordinary structures; considerable damage in poorly built or badly designed structures; some chimneys broken.

Table 1.1 (*continued*)

VIII. Damage slight in specially designed structures; considerable damage in ordinary substantial buildings with partial collapse. Damage great in poorly built structures. Fall of chimneys, factory stacks, columns, monuments, walls. Heavy furniture overturned.
IX. Damage considerable in specially designed structures; well-designed frame structures thrown out of plumb. Damage great in substantial buildings, with partial collapse. Buildings shifted off foundations.
X. Some well-built wooden structures destroyed; most masonry and frame structures destroyed with foundations. Rails bent.
XI. Few, if any (masonry) structures remain standing. Bridges destroyed. Rails bent greatly.
XII. Damage total. Lines of sight and level are distorted. Objects thrown into the air.

Source: Courtesy of the U.S. Geological Survey. The USGS home page is http://www.usgs.gov 18/07/2007.

1.8 Reid's Elastic Rebound Theory

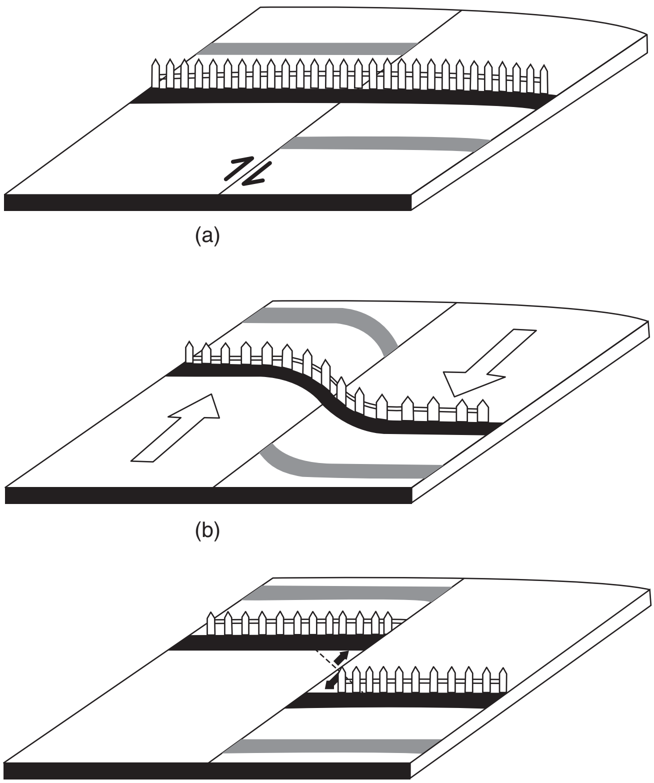
As mentioned earlier in the chapter a key development in the study of earthquakes is Reid's elastic rebound theory first advanced by Henry Fielding Reid and based on his observations following the Great 1906 San Francisco Earthquake. Following an in-depth study he believed *elastic rebound* to be the cause of earthquakes and his simple theory has been confirmed over the years.

Reid examined ground displacement around the San Andreas fault which led him to conclude that the San Francisco earthquake was due to the sudden release of elastic energy accumulated in rocks on both sides of the fault. Elastic rebound theory may be explained in the simplest possible terms by thinking about what happens when an elastic band is stretched then either broken or cut – the energy stored in the band during stretching is released in a sudden 'elastic rebound'. Like the rubber band, the more the fault is strained the more energy is stored in the rocks. When the fault ruptures, this elastic energy is released. The energy is dissipated partly as heat, partly in cracking under ground rocks and partly as elastic waves. These waves cause the actual earthquake. The mechanism is illustrated in Figure 1.27.

In his studies of the San Andreas fault Reid discovered that several metres of relative motion occurred along several hundred kilometres of the fault (Figure 1.28). The Pacific plate slid as much as 4.7 metres in a northerly direction past the adjacent North American plate. Reid concluded that due to the plates sliding past each other, the rocks at the fault zone were bending and storing up elastic energy. When the rocks released this elastic energy and returned to their original form, the result was the Great 1906 San Francisco Earthquake.

1.9 Significant Milestones in Earthquake Engineering

Historically, earthquakes have shown the shortcomings of contemporaneous design methodologies and construction practices, resulting in structural failures and loss of life. Post-earthquake investigations have led to improvements in engineering analysis, design and construction practices. A summary of the historical development of earthquake engineering practice, showing how earthquake engineers have learned from past failures, is shown in Table 1.2.



(a)

(b)

(c)

Figure 1.27 In his studies of the San Andreas fault, Reid discovered that several metres of relative motion occurred along several hundred kilometres of the San Andreas fault.

1.10 Seismic Tomography

Also known as seismic imaging, travel times along ray paths from different directions crossing the region of interest (Figure 1.29) are studied by seismologists to produce three-dimensional models of the interior, applying a process known as *mathematical inversion*. Recordings of ground motions as a function of time (Figure 1.23) provide the basic data that seismologists use to study elastic waves as they spread throughout the planet. Seismology is an observation-based science that tries to address the make-up of the Earth's interior by applying elastodynamic theory to interpret seismograms. There is also the physical constraint of being able to record seismic wave motions only at (or very near) the surface of the Earth. Hence seismologists draw heavily from mathematical methodologies for solving a system of equations commonly referred to as *inverse theory*. A good summary of state-of-the-art 'inverse theories' is provided by Romanowicz (2002).

The quest for mapping the interior of the Earth began from the time seismograms became available. Two eventful discoveries are of interest (a historical digression).

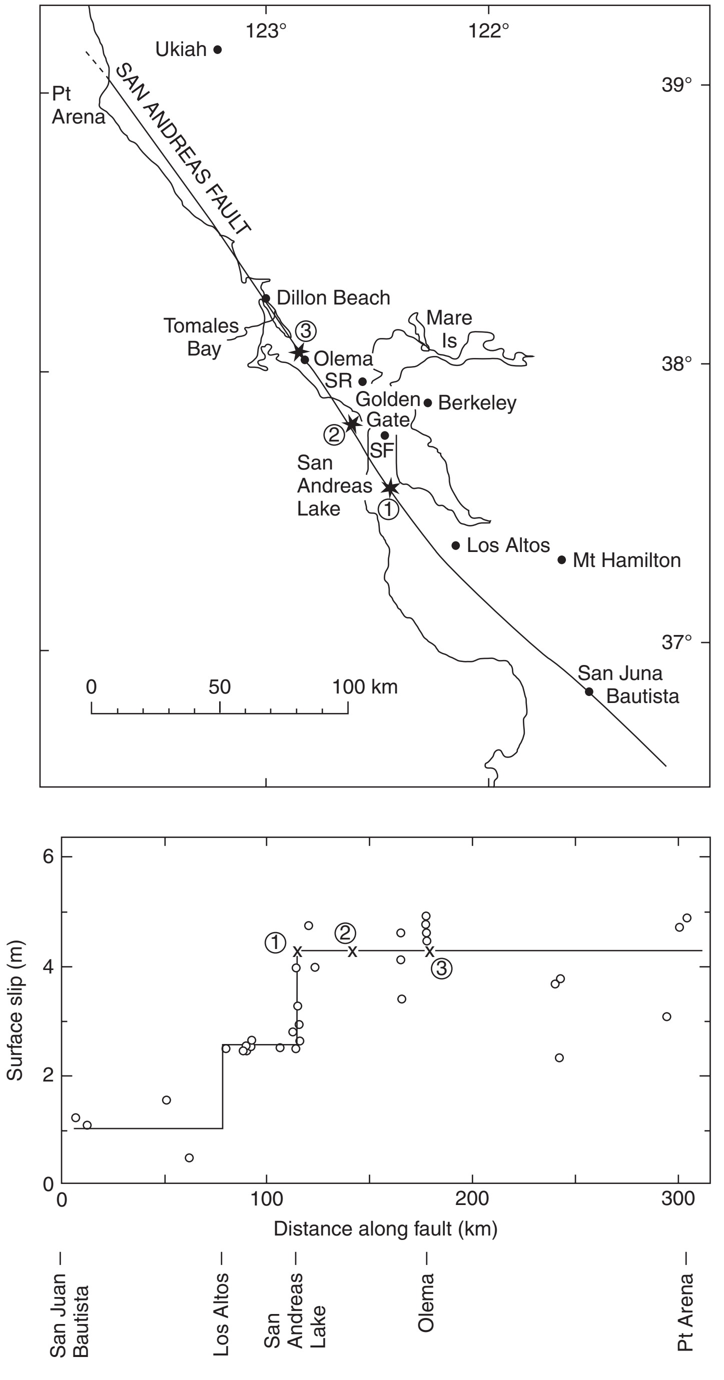


Figure 1.28 Map of the portion of the San Andreas fault that slipped in the 1906 San Francisco earthquake (top) and the amount of surface slip reported at various points along it (bottom) The slip is the distance by which the earthquake displaced originally adjacent features on the opposite side of the fault. Reproduced from Boore (1997), (c) Seismological Society of America (SSA).

Table 1.2 Key milestones in the development of earthquake engineering practice.

Year	Event	Comments
1906	San Francisco Earthquake	State Earthquake Investigation Commission appointed. Realization for the first time of the knock-on effects earthquakes have on our planet. A wealth of scientific knowledge derived.
1911	Reid introduces Elastic Rebound Theory	A significant step towards our understanding of the focal mechanism.
1915	Wegener expounds the theory of '*continental drift*'	Wegener's theory is one of the most significant advances in the twentieth century in the field of plate tectonics.
1927	UBC (Uniform Building Code) published in USA.	First modern code of practice.
1933	Long Beach Earthquake (USA).	Destruction of buildings, damage to schools especially severe. UBC 1927 revised. Field Act and Riley Act introduced. First earthquake for which records were obtained from the recently developed strong motion seismometers.
1964	Niigata Earthquake (Japan).	Showed that soil can also be a major contributor of damage. Soil liquefaction studies started.
1994	Northridge Earthquake (USA) Steel connections failed in bridges.	Importance of ductility in construction realized.
1995	Kobe Earthquake (Japan) Kawashima.	Massive foundation failure. Soil effects were the main cause of failure. Downward movement of a slope (lateral spreading) is said to be one of the main causes. JRA (Japanese Road Association) Code 1996 modified (based on lateral spreading mechanism) for design of bridges.
1995	Economic damages from Loma Prieta (1989) and Northridge (1994) earthquakes evaluated.	Performance based earthquake engineering initiated in USA.

Oldham in 1906 studied the travel times of P and S waves as they travelled from one side of the Earth to the other. He hypothesized for the first time that S waves had penetrated a central core where they travelled at a much lower rate. Seismologists now believe that the outer part of the core is fluid through which S waves do not propagate.

Gutenberg, a seismologist working in Germany in the early 20th century, fixed the depth to the boundary of the separate core to be about 2900 km. This value has remained virtually unchanged since then.

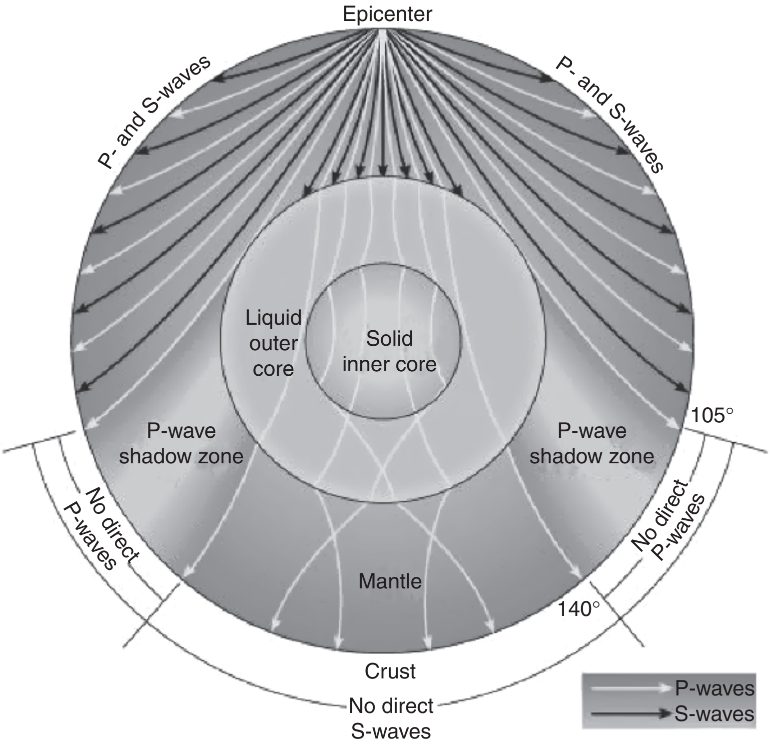


Figure 1.29 Travel paths of seismic waves [Note: S waves do not pass through the liquid outer core]. Levin, H.L. 2003, Reprinted by permission of John Wiley & Sons, Inc.

All high resolution methods for studying the Earth's interior are based on analysing the propagation of seismic body waves as they travel across the Earth. Recall that P waves are compressional waves and the nearest analogy is that of the sound waves in air. P waves can travel through solids and fluids. Shear waves (S waves) propagate only in solids.

Figure 1.29 shows the travel paths of seismic waves through the Earth's mantle from the source to seismographs at the surface. If travel times are available for enough ray paths in different directions crossing a small region, then it is possible to unscramble the recorded plots to reveal a three-dimensional wave speed distribution.

P and S waves penetrate though the Earth as X-rays do in the human body. The concept has an analogy from modern medicine. A common technique used by doctors to obtain internal images of the human body is a CAT (Computerized Axial Tomography) scan. By taking multiple X-rays, the variation in density of the body's tissues can be deduced mathematically and assembled into a three-dimensional image. Probing the interior of the Earth by P and S waves is known as *geophysical tomography*.

The data are treated in much the same way as in medical imaging. Usually a small block of different tectonic type (for example a block for tomography of UK) is probed. Numerical methods play a major role. The resolution of the block depends upon the number of rays passing through the block.

Figure 1.23 shows the recordings of the 1985 Chilean Earthquake. P, S, Love (L) and Rayleigh (R) waves are marked. From these recordings a seismologist can determine the location of the hypocentre, magnitude and source properties.

The surface waves are affected by the structure and the elasticity of the rocks through which they traverse. The measurement of the speeds and wave forms constitute what is referred to as 'tomographic' signals. Decoding these signals is the job of seismologists. Once decoded the signals can reveal the tectonic make-up of the upper part of the Earth. It is also important to establish consistency from one event to another, which is now an active area of research.

1.10.1 The Challenges Ahead

What are the challenges of our times? These are multi-faceted. There is a need, more than ever perhaps, for seismological and geotechnical engineers to work closely with earthquake engineers. The biggest challenge in the field of seismology appears to be understanding the fault rupturing process and predicting the travel path of the waves from the source to site. How are the waves modified during their travel? How does all of this affect the strong motion at the site?

Earthquake source dynamics provides key elements for the prediction of strong ground motion. Early studies in 1970s pioneered our understanding of friction and introduced simple models of dynamic fault rupture using homogeneous distributions of stress and friction parameters. Rapid progress has since been made in the understanding of dynamic rupture modelling and a very good review of the state-of-the-art has been provided by Madariaga (2006). A numerical model able to predict ground motion at a site is a development which would be of great value.

Understanding of soil behaviour and changes thereof during an earthquake continues to elude us. Sometimes, when liquefaction does not manifest itself on the ground surface, it is still very difficult to know; what changes have taken place beneath the surface and more importantly how the soil will respond in the future?

Perhaps the most exciting area of seismological research concerns movements deep inside the earth. Many of these movements can now be described but we know very little about the driving forces behind them.

At the other end of the seismic design spectrum, we note that structures need to be designed to withstand seismic forces. A key requirement for modelling the mechanical response of a structure is to have realistic estimates of how the ground beneath that structure will move during an earthquake.

The potential payoff for being able to perform accurate nonlinear analysis of soil structure behaviour is high. Accurate numerical simulations can provide a more realistic picture of how a structure actually responds during strong ground motion. With this information, engineers can be significantly more confident that a structure will survive a large-magnitude earthquake.

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