



3. Earthquakes

3.1. Elastic rebound theory

Rocks at the edges of tectonic plates are subject to tremendous forces resulting in intense deformation. The force per unit area acting on a rock is called **stress**. The three types of directional stress experienced by rocks are compressional, tensional, and shear stress (Fig. 26).



Figure 26: Types of stress and associated types of fault. Arrows indicate the direction of forces applied to the rocks.

If the stress is large enough, rocks undergo deformation, i.e. a change of shape and/or volume. The amount of deformation experienced by a rock is called **strain**. The behavior of a rock in response to stress can be **elastic**, **brittle** or **ductile**. A rock behaves in an elastic manner when it recovers its original shape after the stress is removed. When the stress exceeds a value called the rock **strength**, the rock experiences a permanent deformation. The deformation can be either brittle in which case the rock breaks, producing a fracture called a **fault** (Fig. 27A), or ductile (plastic) in which case the rock is deformed without breaking. An example of ductile deformation is a **fold** (Fig. 27B). Note that rocks may first be folded and then fractured as stress increases.



Figure 27: Examples of rock deformations. (A) Brittle deformation (faults in northwestern Australia, source: Google Earth: lat.: -18.0556, long.: 126.5107) and (B) ductile deformation (folds in the Appalachian Mountains, USA, source: Google Earth, lat.: 40.562241, long.: -76.716128).

In the upper part of the lithosphere, rocks tend to behave in a brittle manner and the rock strength increases with increasing pressure (depth). Ductile behavior is promoted by high temperature and high pressure, which is why rocks of the asthenosphere can flow in convective currents. Decreasing the strain rate also promotes ductile deformation (think about the behavior of modeling clay as you stretch it slowly or fast). Another factor influencing the behavior of rocks is their mineral composition (e.g. peridotite becomes ductile at a higher temperature than granite –remember the high melting temperature of olivine–).

Earthquakes are produced by the brittle deformation of rocks. They are confined to the cold rigid lithosphere, mostly the crust, where rocks behave in a brittle manner. Rocks of the asthenosphere under conditions of high temperature and high pressure display a ductile behavior.

Prior to a rupture event producing an earthquake, a body of rock experiences a certain amount of elastic deformation during which energy is stored (**strain energy**). Once the stress applied to the rocks exceeds the rock strength, rupture occurs along a fault plane and the rocks on both sides of



the fault slip past each other and recover their original shape (= elastic rebound, Fig. 28). During the rupture event, strain energy is released in the form of frictional heat and **seismic waves** (see section 7.3. for more details about the nature of seismic waves). The slipping motion of the two blocks starts at a point called the **focus** of the earthquake. The **epicenter** of the earthquake is the point on the Earth's surface directly above the focus. After the rupture event, stress buildup resumes until the next earthquake occurs. This is called the **elastic rebound theory** which explains the recurrence of earthquakes along active faults and is illustrated by the saw-tooth shape of the stress vs. time plot in figure 28.



Figure 28: Elastic rebound theory. Between t_1 and t_2 , strain energy accumulates and the rocks undergo elastic deformation until rock strength is reached. Between t_2 and t_3 , the rocks break along a fault plane and energy is released as heat and seismic waves. After t_3 , strain energy starts to build up again. The recurrence interval of earthquakes in this model is the time interval between t_1 and t_3 . Source: Understanding Earth 6th edition.

In reality, the frequency of earthquakes is not as regular as figure 28 suggests. If it was, it would be much easier to predict their timing. Irregularity can be attributed to the following factors (Fig. 29):

1. incomplete stress release during earthquakes
2. change in stress intensity (stress partially released by movement along a nearby fault)
3. change in rock strength (e.g. the infiltration of fluids in faults may decrease frictional strength due to the lubrication effect of fluids)



Figure 29: Factors influencing the frequency of earthquakes: (1) incomplete stress release, (2) change in stress intensity, and (3) change in rock strength. Blue lines represent the release of stress during successive earthquakes. Source: adapted from Understanding Earth 6th edition.

3.2. Earthquakes and plate tectonics

Most earthquakes are distributed along active plate boundaries. The depth at which they occur (**focal depth**) varies according to the tectonic setting (Figs. 30 & 31).

Earthquakes occurring along divergent margins are shallow and associated with tensional forces (Fig. 31A). Shallow earthquakes also occur along transform faults. Transform faults associated with mid-ocean ridges produce earthquakes along the portion of the fault where the two plates move in opposite directions (Fig. 31B). The deepest earthquakes are related to convergent margins, particularly subduction zones (Fig. 31B). In this setting, the cold, brittle oceanic plate plunges into the asthenospheric mantle. The subducting plate remains brittle until it reaches a depth where temperature and pressure are too high for brittle deformation to occur. The deepest earthquakes are generated at around 700 km. Earthquakes associated with a continental collision may be deep as well but not as deep as those related to subduction zones. The maximum focal



depth recorded in the Himalayas is approximately 100 km. Below that depth, continental crustal rocks tend to deform in a ductile manner. Remember also that the brittle strength of crustal rocks increases with increasing pressure (depth).

Earthquakes may also occur far from plate boundaries (Fig. 30). These ***intraplate earthquakes*** are linked to hot spots or old faults related to ancient plate boundaries (e.g. the 2011 Virginia earthquake in the United States occurred in an area which used to be part of a collision zone between two ancient continents).



Figure 30: Global distribution of earthquakes. Here earthquakes are classified in three categories according to their focal depth: shallow (< or = 50 km, blue dots), deep (50-300 km, green dots), and very deep earthquakes (>300 km, red dots). Each dots represent a epicenter. Source: Understanding Earth 6th edition (M. Boettcher and T. Jordan).



Figure 31: Focal depth of earthquakes in different tectonic settings. The orange line delineates the lower limit of the seismogenic zone. (A) Mid-ocean ridge, (B) transform fault, (C) subduction zone, and (D) continental collision.

3.3. Seismic waves

During an earthquake, the rocks are affected by three types of movements corresponding to three distinct groups of seismic waves (Fig. 32):

1. ***P-waves*** (primary waves): compressional waves (push and pull motion, much like sound) traveling at 6 km/s in solid rocks and more slowly in liquids and gases.
2. ***S-waves*** (secondary waves): shear waves (sideways motion –at right angle to the direction of wave propagation–) traveling more slowly than P-waves and unable to propagate in liquids and gases.
3. ***Surface waves****: confined to the surface of the Earth with two components:
 - (1) the ***Rayleigh wave***: an elliptical motion decreasing with depth (similar to ocean waves).
 - (2) the ***Love wave***: a lateral motion (sideways shaking).



Figure 32: Seismic waves. (A) Primary wave, (B) secondary wave, and (C) surface wave (Love and Rayleigh waves). Source: adapted from USGS.

* Surface waves are responsible for most of the ground shaking during an earthquake.



When an earthquake occurs, P-waves and S-waves travel in all directions within the Earth. Surface waves travel exclusively at the surface of the Earth and are slower than P- and S-waves. S-waves travel more slowly than P-waves (Fig. 34). The speed of P- and S-waves depends on the physical properties of rocks. The harder the rock, the faster the wave. Therefore, wave speed depends on rock density –hence pressure, temperature and the rock mineral composition–. The relationship between the speed of seismic waves and rock density combined with the properties of wave refraction and reflection at the interface between Earth's layers can be used to reconstruct the structure of Earth's interior (see Chapter 7).

3.4. Earthquake characterization: localization, magnitude, and intensity

The measure of seismic waves amplitude and frequency is achieved by using an instrument called a **seismograph** (Fig. 33). A seismograph is built to record vertical and horizontal ground movements. The basic principle is very simple: a mass is loosely attached to a frame which is anchored to earth. The fixed mass is equipped with a pen which records the difference in motion between the mass and the frame on a revolving drum –the drum moves with the frame–. The mass must be heavy enough to remain at rest relative to the frame. A mass attached to a spring is used to record vertical ground movements (Fig. 33, right). A mass attached to a hinge is used to detect horizontal ground movements (N-S and E-W components, Fig. 33, left). The graphs obtained are representations of seismic waves and called **seismograms** (Fig. 34).



Figure 33: The seismograph. Instrument recording horizontal ground movements (left) and vertical ground movements (right). See text for explanations. Source: USGS.



Figure 34: Seismogram of a distant earthquake recorded in United Kingdom. Source: British Geological Survey.

How can we determine the location of an earthquake?

Since P-waves travel faster than S-waves, the greater the distance from an earthquake, the greater the difference between the arrival times of P-waves and S-waves (P-S intervals). The relationship between P-S intervals and the distance from the epicenter of an earthquake is predictable over thousands of kilometers. Seismologists have established travel-time curves that can be used to pinpoint the distance from the epicenter of an earthquake (Fig. 35). These travel-time curves are based on data from a great number of earthquakes and many seismographs around the world. Providing you have at least three seismographs in different geographic locations recording the same waves, you can locate precisely the epicenter of the earthquake. The focal depth can be determined based on further analyses of seismograms.



Figure 35: Travel-time curves of seismic waves and three seismograms from three stations recording seismic waves from the same earthquake. Matching these seismograms with the travel-time curves allows to determine the distance between each stations and the earthquake focus. Source: USGS.

How can we determine the size of an earthquake?

The size of an earthquake can be determined based on two different techniques:

1. Measure of the amount of energy released (**magnitude scales**).
2. Estimate of the destructive effects of ground shaking (**intensity scales**).

The first widely used magnitude scale was the **Richter scale**. The Richter magnitude of an earthquake depends on the logarithm of the amplitude of the largest wave and the distance from the epicenter* –the time lag between the P-wave and S-wave– (Fig. 36). Magnitudes are expressed as whole numbers and decimal fractions (e.g. 3.5, 5.2, 6.4...). An increase of one unit of the Richter magnitude corresponds to a tenfold increase in wave amplitude (and 32 times more energy released*!)



Figure 36: Graphic representation of the relationship between the Richter magnitude of an earthquake, the amplitude of the largest wave (in this example, 23 mm measured on the seismogram) and the difference in arrival time between the P wave and S wave (in this example, 24 s). Note that for a given distance from the epicenter, an increase in one unit of magnitude corresponds to a tenfold increase in the wave amplitude. Source: USGS.

A scale that is more widely used nowadays is the **moment magnitude** scale. The moment magnitude is a function of the seismic moment. The seismic moment is proportional to the area of faulting and the average fault offset**. Like the Richter magnitude, the moment magnitude can be calculated based on the analysis of seismograms. In the moment magnitude scale, an increase of one unit of magnitude corresponds to a tenfold increase in the area of faulting. The values of moment magnitudes are roughly similar to those of the Richter scale. For example, if an earthquake measures 5.5 on the Richter scale, its moment magnitude will be approximately 5.5 too.

The other way to evaluate the size of an earthquake is to establish a scale related to the destructiveness of earthquakes. This is the **intensity scale**. An intensity scale that is commonly used today is the **modified Mercalli scale** (Fig. 37).

* Richter's magnitude (M_L) = $\log A + 2.56 \log D - 1.67$, where A is the amplitude of ground motion (in μm) and D is the distance from the epicenter (in km) (source: British Geological Survey). Hence for a given value of D, each increment of 1 unit of magnitude M corresponds to a tenfold increase of the amplitude A.

* The energy released during an earthquake (E) is proportional to $10^{(1.5 \cdot M)}$. If E1 is the energy of an earthquake of magnitude M and E2 is the energy of an earthquake of magnitude M+1, $E_2/E_1 = 10^{1.5} = 32$ (source: British Geological Survey).

** Moment magnitude (M_W) = $2/3 \log M_0 - 6.06$. The seismic moment (M_0) = $\mu \cdot \text{rupture area} \cdot \text{slip length}$, where μ is the shear modulus of the crust ($3 \times 10^{10} \text{ N/m}^2$). The shear modulus is the measure of the resistance of a material to shear strain (source: British Geological Survey).



Figure 37: Modified Mercalli scale compared with the Richter magnitude scale (source: Missouri Department of Natural Resources) and a modified Mercalli intensity map of the 1906 San Francisco earthquake (source: USGS, authors: John Boatwright and Howard Bundock).

In Japan, it is the intensity scale established by the Japan Meteorological agency that is preferentially used (the JMA or Shindo scale). It is analogous to the Mercalli scale but comprises only 7 levels of intensity instead of 12 (<http://www.jma.go.jp/jma/en/Activities/inttable.html>).

3.4. Fault mechanisms

In order to fully characterize an earthquake, we not only need to know its focus, origin time, magnitude and intensity, but we also need to describe the characteristics of the fault which triggered the earthquake or **fault mechanism**: the direction of the **fault strike*** (e.g. north-south, east-west), the angle and direction of the **fault dip** (e.g. 60° south, 75° east), and the **type of fault** involved (normal, reverse, or strike-slip –right lateral or left-lateral–, Fig. 38).



Figure 38: The three basic types of fault. Note that strike-slip faults can be either right-lateral or left-lateral. In the former case, one observer standing on one side of the fault sees the other side moving to the right (i.e. right-lateral). In the latter case, the observer sees it moving to the left (i.e. left-lateral).

For faults that can be observed at the surface of the Earth, the fault mechanism can be determined based on field observations. But for faults that are too deep to break the surface or faults at the bottom of the ocean, the fault mechanism must be deduced from the analysis of seismograms. Seismologists look at the **first ground movement** recorded by seismographs located at different places around the earthquake's epicenter. The first ground movement is determined on a seismogram by looking at the shape of the first P-wave. For example, on a seismogram recording horizontal ground movements, if the line of the seismogram goes up first, it means that the ground has moved away from the earthquake's epicenter (extension). If the line goes down first, it means the ground has moved closer to the earthquake's epicenter (compression) (Fig. 39). This allows the identification of two perpendicular planes, one of which is the fault plane. This method alone does not allow distinguishing the fault plane from the other one. Additional information is needed to determine which one is the fault plane.

* the fault strike is the intersection of the fault plane and a horizontal surface.



Figure 39: Schematic relationship between the first ground motion (horizontal in this example) recorded on seismograms and the fault mechanism. The figure shows a bird's eye view of the ground. The center of the large circle represents the earthquake's epicenter. Black dots are seismographs recording a first ground displacement away from the epicenter (line going up first). White dots are seismographs recording a first ground displacement toward the epicenter (line going down first). One of the two perpendicular lines delineating the four quadrants is the fault plane. Source: adapted from USGS.

3.5. Seismic hazards and risks

Seismic hazards are the natural phenomena related to earthquakes that may be potentially harmful to human populations. Seismic hazards directly caused by an earthquakes are called primary hazards: faulting and ground shaking. These may in turn cause secondary hazards, such as landslides, soil liquefaction, tsunami (see below for more details about these secondary hazards). **Probabilistic seismic hazard maps** display the probability that a given site will experience ground shaking exceeding a given value within a given period of time (see the end of this chapter for more information about earthquake prediction). Figure 40 intends to show the difficulty to accurately predict that a large earthquake will occur at a particular place within a certain period of time.



Figure 40: Probabilistic seismic hazard map of Japan showing the probability (0-100% chance it happens) of ground shaking of intensity "6-lower" or higher (on the JMA seismic intensity scale) during a 30-year period starting in January 2010. Note that the large 2011 Tohoku earthquake was not predicted by this model. Source: Robert J. Geller (2011) in *Nature*, v. 472, 407-409.

The **Seismic risk** is a measure of the damage expected in a given interval of time in a particular region. The seismic risk in a city with buildings engineered to resist large earthquakes will be much lower than in a city in which building codes are not adapted even if these two cities are exposed to the same seismic hazards. Seismic hazards must be determined prior to the assessment of seismic risks.

Besides ground shaking and faulting, three important secondary seismic hazards that can cause great damages are **landslides**, **soil liquefaction**, and **tsunamis**. Landslides and soil liquefaction are due to ground instabilities resulting from ground shaking. The former occur when material on a slope becomes unstable and moves down the slope as a result of this instability (Fig. 41). Soil liquefaction refers to a layer of water-saturated sediment losing its cohesion and behaving as a liquid when shaken (Fig. 42). The shaking causes the grains to loose contact between each other, therefore allowing the sediment to flow like a liquid (Fig. 42A).



Figure 41: Landslides in El Salvador caused by a large earthquake in 2001 (left, source: USGS) and in Fukushima, Japan, caused by the 2011 Tohoku earthquake (right, source: Public Works Research Institute, Landslide Research Team, Erosion and Sediment Control Research Group).



Figure 42: Schematic representation of sediment liquefaction (left) and example of the damages caused by liquefaction during the 1964 Niigata earthquake, Japan (right, source: Japan National Committee on Earthquake Engineering, Proceedings of the 3rd World Conference in Earthquake Engineering, Volume III, pp s.78-s.105).

A tsunami represents another important secondary seismic hazard. The effect of a tsunami can be a lot more disastrous than ground shaking. During an earthquake, a tsunami forms when a portion of the seafloor moves vertically along a fault plane. The above water mass is displaced and creates a set of waves which propagate in all directions. Large tsunami can be generated at subduction zones where the overriding plate gets squeezed by the subducting plate and a considerable amount of stress accumulates. Stress is released when a portion of the crust springs seaward, raising the seafloor and pushing a large water mass up (Fig. 43). Tsunami waves travel very fast (up to 800 km/h) and have a very long wavelength (hundreds of kilometers between two successive waves). As they approach the shoreline, waves slow down by friction against the seafloor and their height increase (conversion of kinetic energy to potential energy).



Figure 43: Process of tsunami formation at subduction zone. (A) a portion of the overriding plate (1) is pushed landward by the subducting plate (2) and stress accumulates. (B) Stress is suddenly released as the portion of the overriding plate springs seaward along a thrust fault (low-angle reverse fault), generating an earthquake and pushing a water mass up. (C) The displacement of the water mass creates a set of tsunami waves at the surface which propagates in all directions.

Tsunami can travel over a considerable distance and can still be detected thousands of kilometers away from their origin (Fig. 44).



Figure 44: Simulation of the maximum wave amplitude related to the 2011 Tohoku tsunami based on data from buoys and sea-floor sensors. Source: NOAA.

When it comes to improving public safety in areas prone to earthquakes, there are two fundamental questions which have to be considered: (1) how can seismic risk be reduced? and (2) can earthquakes be predicted?



How can seismic risk be reduced?

- Hazard characterization
- Land-use policies
- Earthquake-proof engineering
- Emergency preparedness and response
- Warning systems (for earthquakes and tsunamis)

Can earthquakes be predicted?

With our current knowledge and technology, geologists cannot yet predict an earthquake early enough to evacuate the area. As we have seen earlier, probabilistic models exist but they make predictions over the long term (30 or 50 years) and sometimes fail to predict large earthquakes. In order to make predictions, geologists need to estimate the recurrence interval of earthquakes in a particular region. There are two ways to do it:

- (1) One is to study historical data to infer the probability of future earthquakes. One can also go back further in time by looking at the geological record and dating earthquake-induced deformations of sediment layers or tsunami-lain sediments.
- (2) Another way is to measure how much strain (deformation) accumulates along a fault every year (e.g. using GPS). Knowing when this fault slipped for the last time and how much strain was released, it is possible to estimate the timing of the next earthquake based on strain accumulation measurements. This kind of data is still relatively rare because it requires monitoring fault systems for a long time.

Earthquake prediction is complicated by the fact that faults are often not isolated but form complex networks. The release of stress in one fault segment may decrease or increase stress in another fault segment. It is therefore essential to understand how stress is distributed in these fault systems and how individual faults influence each other in order to improve the reliability of earthquake predictions. Geologists have been trying to use small earthquakes (foreshocks) occurring before larger seismic events to predict large earthquakes days or weeks before they occur. However, it is very difficult to distinguish foreshocks from the seismic background and the method is not yet reliable for most quake-prone areas.