Earthquakes: Supplementary information

This document provides material to supplement the chapter on Earthquakes in *Natural Hazards: An Australasian Perspective*.

<u>Supplement to Introduction: Notable Australasian earthquakes</u>

The differences in the seismic hazards of Australia, New Zealand and Indonesia are reflected in their earthquake statistics. Australia's seismicity is nearly two orders of magnitude lower than that of the New Zealand to Tonga margin, which experiences about half the seismicity of Indonesia. Consequently, Australia has suffered only 14, New Zealand 477, and Indonesia >26,000 earthquake-related fatalities (not including the 2004 tsunami) since the early 19th century (Table 1). These numbers do not provide a basis for complacency in Australia; despite this disparity in the seismicity and casualty numbers, every state of Australia and every region of New Zealand has experienced earthquake shaking during this time. Mainland Australia has experienced six earthquakes M_w >6 and one M_w >7. In both Australia and New Zealand, casualties and financial losses have been more strongly correlated with earthquake proximity to major urban centres than with Mw. The 1855 M_w 8.2 Wairarapa earthquake caused only 9 deaths, compared with 256 during the M_w 7.8 Napier earthquake in in 1931. The 1968 Meckering (Aus) and 2010 Darfield (NZ) earthquakes ruptured to the surface along >25km in rural areas and caused no deaths. In contrast, even relatively small events such as the 2011 M_w 6.2 Christchurch (NZ) and 1989 M_w 5.7 Newcastle (Aus) earthquakes occurred close to major urban centres and caused >billion dollar losses. Although remote earthquakes are less hazardous, some such as the 2010 Darfield (NZ), 1988 Tenant Creek (Aus) and 1946 Lake Coleridge (NZ) earthquakes have impacted lifeline infrastructure. Offshore earthquakes have so far caused only minor losses in Australia and New Zealand, despite causing shaking in cities including Perth (1906), Gisborne (2007), Invercargill (2003/9), Darwin (2012) and Wellington (2013). However, the 26/12/2010 earthquake off the coast of Sumatra demonstrated that offshore earthquakes have great tsunamogenic potential, particularly along the subduction and collisional margins of the Australian and Pacific Plates. This is an aspect of the earthquake hazard that has yet to be severely experienced in New Zealand and Australia since European occupation.

<u>Supplement to C. The Origin of Earthquakes</u>

C.1. Plate tectonics

The seismicity that gives rise to earthquake hazards arises from the movement of tectonic plates, which is driven primarily by subducting lithospheric slabs. These impart a **slab pull** force (Figure 1A in textbook) that drags the trailing plate forward. The over-riding plate may also be **suctioned** towards the subduction zone. Another driver of plate motion relates to the emergence and cooling of new oceanic crust at divergent plate boundaries along mid ocean ridges, which results in outward-directed gravitational sliding termed the **ridge push** force (Figure 1A,B). **Convection currents** in the underlying asthenospheric mantle may impart frictional forces on the base of the plates that encourage them to move along. Slab pull is thought to be the most important driving force because tectonic plates that are fringed by large subduction zones move the fastest.

If the plate driving forces such as slab pull and ridge push are both oriented in the same direction, as is the case for the Australian Plate (Figure 1A, B), why don't these plates continue to accelerate in accordance with Newton's Law (Acceleration = Force / Mass)? The reason is that plate motion is also opposed by **resisting forces** along plate boundaries, most commonly with continental collision zones, such as those present on New Zealand's South Island or in Papua New Guinea (Figure 1A). Tectonic plates could be personified as having complex feelings; being pushed this way, pulled that way, occasionally speeding up or slowing down, and even fragmenting into smaller plates when things get tough. As a result of the complex interplay between these forces, and including the additional forces that arise due to differences in the gravitational potential of the lithosphere due to elevation, thickness, and temperature variations, tectonic plates are highly stressed. **Stress**[1] is equivalent to Force / Area.

C.2. Rock fracturing

Major continental faults are not simple planar features confined to one surface, rather they are zones of highly fractured rock that are commonly wider than 100 m, and even wider than 1 km (Figure 3B in textbook). They typically consist of a thin (up to 10s of cm wide) **principal slip zone** consisting of **fault gouge** or **cataclasite**, upon which most of the fault slip occurs, flanked by a wider zone of highly fractured and deformed rock that in places may be a **fault breccia**, flanked by a wider zone of damaged rock (Figure 3B). Seismic waves travel more slowly though these zones of damaged rock, enabling scientists to characterise the fault geometry at depth using **seismic wave velocities** (Figure 3B). For faults that rupture the surface, a complex array of fractures may be produced in addition to the principal slip zone within a **surface fracturing zone**, and additional folding or **surface deformation** may occur over a broader wavelength outside of the zone of discrete fracturing (Figure 3B). The epicentres of smaller earthquakes (**aftershocks**) following the main earthquake on the major fault (**mainshock**) may be 100s of meters to several kilometres outside of the main fault deformation zone, indicating that even the **wall rock** is riddled with fractures that may be unrelated to the main fault zone (Figure 3B).

C.3. *Tectonic habitat of earthquakes*

The nature of the tectonic plates and their relative movement exerts the strongest influence on the types and rates of earthquake activity that occur at the plate boundary. Most earthquakes at **divergent plate boundaries**, such as the boundary between the Australian and Antarctic Plate south of the Australian continent (Figure 1A) involve displacement on normal and transform strike-slip faults. **Mid ocean ridges** tend to have the shallowest and smallest earthquakes. Plate divergence may also occur within continents (e.g., Taupo Rift in New Zealand's North Island), often causing thinning of the crust, mantle upwelling and volcanic activity.

The largest faults on Earth are the subduction zones at **convergent plate boundaries**. Some of these faults are more than 1000 kilometres long and 250 kilometres deep. Subduction zones have the deepest and largest earthquakes on Earth. Subduction zone earthquakes occur to depths below 200 km and in some cases can be as deep as 700 km. Earthquakes on subduction zones may involve displacement of over 60 metres in a single earthquake (Ito et al., 2011). The largest subduction zone earthquakes are called **megathrusts** (Figure 1B) and may involve fault rupture that extends from more than 50 to 60 km deep right up through the ocean floor. Subduction megathrusts may be accompanied by large tsunami, such as the 2004 Indonesian Boxing Day tsunami and the 2011 Tohuku Japan tsunami. Parts of descending slabs may initially break along normal faults like staircases on an escalator (Figure 1C). As slabs descend deeper into the Earth, they may begin to tear apart, causing extensional earthquakes within the slab (Figure 1B).

Tectonic plate convergence may also involve **continental collision**. In this case, crustal thickening occurs, as neither side of the buoyant continental material is easily submerged. Large strike-slip faults accommodate oblique collision and enable geometric and lithologic irregularities and variations in convergence rate to be accommodated within the convergence zone. The **Southern Alps of New Zealand** (Figure 1D) provide a classic example of an obliquely converging continent-continent plate boundary collision zone. The relatively rapid northerly movement of the Australian Plate and relatively rapid westerly movement of the Pacific Plate (Figure 1A) causes thrusting, reverse faulting, and dextral strike-slip faulting throughout the South Island. About 75% of the total relative motion between these plates is accommodated by oblique slip (dextral strike-slip and reverse) on the largest and most active faults (e.g., the Alpine Fault; Figure 1D) (Norris and Cooper, 2001).

Strike-slip plate boundaries occur where plates primarily move past each other laterally (i.e. **translation**), with lesser (or absent) components of divergence or convergence. The San Andreas Fault in California is one of the most famous strike-slip plate boundaries in a continental setting. Examples of major strike-slip plate boundary faults in New Zealand include the Marlborough Faults (Figure 2) although some oblique motion also occurs along these faults. Strike-slip plate boundary faults in oceanic crust (i.e. **transform faults**) were first identified along the Mid-Atlantic Ridge, where they accommodate different geometries and rates of sea floor spreading. Where strike-slip plate boundaries also incur plate convergence they are classified as **transpressional plate boundaries** and where they incur divergence they are classified as **transtensional plate boundaries**.

Although classical models of plate tectonics treat the interior of tectonic plates as comparably rigid entities, earthquakes also occur in plate interiors on **intraplate faults**. Intraplate earthquakes only account for < 1% of the total global earthquake energy release (Johnston, 1989). Many continental interiors, including Australia, are dominated by compressive stresses. In general, intraplate earthquakes tend to be most concentrated in **zones of crustal weakness**, where past crustal deformation has been most abundant (e.g. crustal terrain boundaries, former continental rifts, failed rifts) and thus where faults capable of being reactivated are likely to be most densely concentrated (Figure 1B) (Johnston, 1989). Intraplate earthquakes may also occur in oceanic crust; some of the largest of these (e.g., 2012 M_w 8.7 and 8.2 Indian Ocean earthquakes near Sumatra) have occurred within the Indo-Australian Plate.

The five fault types described in Section B.2 may all be present and seismically active within the same plate boundary setting. For instance, even though it occurred in a convergent plate boundary setting, the 4 September 2010 Darfield earthquake in New Zealand’s South Island was caused by the successive rupture of reverse, strike-slip, and normal faults (with oblique slip components) all within about 40 seconds. Faults with different kinematics may interact and rupture together over timescales ranging from a single earthquake to millions of years.

<u>Supplement to D: Earthquake behaviour and triggering</u>

D.1. Models for describing earthquake behaviour

As discussed above, the stress acting on a fault plane increases over the **interseismic period** (i.e. the time period between successive major earthquakes) due to tectonic plate movement, which causes slow, continuous differential movement of rocks on either side of the fault (Figure 4A in textbook).

Once the stress on the fault increases to a level in excess of the **frictional strength of the fault** (i.e. **failure stress**), fault rupture occurs. During the earthquake rupture (i.e. **coseismic**) the stress state on the fault drops rapidly (i.e. **stress drop**; Figure 4B), and the elastic strain is converted into permanent strain in the form of **fault displacement**, resulting in an elastic strain decrease (Figure 4C) and total fault displacement increase (Figure 4D). The time encompassing the coseismic, postseismic, and interseismic intervals between successive earthquakes is termed the **earthquake cycle** and it is underpinned by the **elastic rebound theory** (Reid, 1910) (Figure 4A).

If the rate of stress accumulation on the fault (i.e **stress loading rate;** σ_{Lr} in Figure 4E), the frictional strength of the fault (equivalent to failure stress, σ_F in Figure 4E), and the stress drop ($\Delta\sigma$ in Figure 4E) associated with each earthquake is constant throughout the earthquake cycle, then earthquakes should occur at regular time intervals (Δt) and with consistent fault slip. In this case, one could **predict the timing, slip, and M_w of future earthquakes** based on the timing and slip of previous ones. This model for earthquake recurrence is called the **periodic earthquake model** and is shown by earthquakes 1-3 in Figure 4E. Earthquake 3 was preceded by some small earthquakes ('i' in Figure 4E; termed **foreshocks**), but these made little difference in the recurrence interval. Sometimes earthquakes may occur at lower (e.g. earthquake 4) or higher than expected failure stresses. In the case of earthquake 4, Δt was shorter than prior events, however the stress drop returned the stress state on the fault back to the same **residual stress** level (σ_R) as earthquakes 1 to 3. This represents an example of a **slip-predictable earthquake**; the amount of stress drop and associated fault slip of the forthcoming earthquake can be predicted based on the time elapsed since the last earthquake, however the timing of the forthcoming earthquake is unknown. Sometimes earthquakes fail at the expected failure stress but have lower than expected stress drop (e.g. earthquake 5). This is an example of a **time-predictable earthquake**; the timing of forthcoming earthquake can be predicted from the timing and slip of the last earthquake because the seismic loading rate is constant, however the slip in the forthcoming earthquake is unknown.

If the stress loading rate between earthquakes varies with time (e.g., ii, iii) then the timing of the next earthquake may not be predicted, even if the failure stress is similar to preceding earthquakes (e.g. earthquake 6). Increases or decreases in stress loading rate could reduce or increase the time to the forthcoming earthquake. Furthermore, continued slip after the expected coseismic stress drop (e.g. iv) could increase the time until the forthcoming earthquake. An overall stress loading rate might experience step changes (e.g., v) due to, for instance, earthquakes on other nearby faults (open stars) that increase the stress on the fault of interest; such step change increases bring forward the time of the forthcoming earthquake (earthquake 7). A stress drop following an earthquake might be similar to preceding earthquakes but accomplished in a different fashion, for instance by a cluster of smaller earthquakes, rather than a single event (small stars, vi).

Detailed descriptions of earthquake physics are rupture behaviour are provided by Scholz (1992) and Kanamori and Brodsky (2004) amongst many others. Earthquakes are clearly not so simple beasts, although some faults may behave more regularly than others (Berryman et al., 2012). Perhaps one of the reasons for why earthquake recurrence doesn't seem to keep on a tight schedule relates to the myriad of ways in which earthquakes may be triggered.

D.2. Earthquake forecasting

Hazard managers need to clearly distinguish the concepts of **earthquake forecasting** and **earthquake prediction**. Spatial and temporal earthquake statistics, and the transfer of shear stress between faults (**Coulomb static stress change – section D.3.**) during a seismic sequence are parameters that can be mapped with some confidence. These factors relate closely to variations in the seismicity within a given area and give rise to earthquake forecasting. An **earthquake forecast** is a probabilistic statement about future earthquakes of defined magnitudes in a specified spatial-temporal window. A time-independent forecast is one in which the subdomain probabilities depend only on the long-term rates of target events; the events are assumed to be randomly distributed in time, and the probabilities of future events are thus independent of earthquake history. In a time-dependent forecast, the probabilities depend on the information available at time when the forecast is made (Jordan et al., 2011).

One field of earthquake forecasting relies upon the use of **statistical analysis of historical or prehistoric seismicity**. Earthquake forecasts using statistical models based on historical seismicity are presently being utilized in several countries, including New Zealand, for a variety of future time intervals ranging from the 24 hours to decades (Gerstenberger et al., 2014). Such forecasts typically report the probability of an earthquake of a given magnitude occurring in a given region over a given time interval, similar to weather forecasts. Stress change modelling may also be added to gain additional insights into the spatial distribution of future earthquakes. Earthquake forecasts for individual faults typically combine **paleoseismic** data (data about prehistoric earthqakes), which enables estimation of fault slip, earthquake recurrence interval and magnitude for preceding earthquakes, with geodetic measurements to forecast the timing and magnitude of future earthquakes on the target fault. Forecasts of this nature are often limited by the apparent variability in slip, magnitude, and inter-event timing of prehistoric earthquakes that is commonly revealed in geologic studies. Stress change modelling may also be used to estimate changes in earthquake probabilities on specific faults from preceding ruptures.

One field of earthquake forecasting utilizes the detection and monitoring of **earthquake precursors**, defined as physical, chemical, or biological anomalies that are interpreted to signify an impending earthquake. Notable precursors include anomalous animal behaviour, changes in electric and magnetic fields, changes in gas emission (flux and chemistry), changes in groundwater or spring temperature, gas concentration, and level, geodetic changes (e.g. strain rate), changes in ground temperature, changes in crustal structure (e.g. seismic wave velocities), changes in rock electrical conductivities, changes in thermal infrared radiation, and changes in the location, rate, and frequency-magnitude distributions of seismicity. The reliable use of earthquake precursors for earthquake prediction is challenging because (1) observed anomalies are commonly claimed as precursors only after the earthquake has occurred, (2) 'background' behaviour including non-precursory anomalies (that could lead to false alarms) are typically unquantified and thus the inherent non-precursory spatial and temporal variability of the phenomenon is unknown, (3) the causal mechanism linking precursory phenomenon to pre-earthquake processes may be unknown, debated, and/or not diagnostic of future earthquake magnitude, timing, or location, (4) apparent spatial and temporal relationships between precursory activity and earthquakes are commonly not statistically valid, (5) fault behaviour may be inherently complex in time and space (see section B.4), limiting the resolution with which future earthquakes can be predicted. As a consequence, the search for a single reliable precursor of a large earthquake has been unsuccessful to this point. Given the challenges outlined above, many scientists think that the search for a ubiquitously diagnostic earthquake precursor that fulfils the above-stated prediction criteria is less important than monitoring a set of precursors that could be perhaps be used to estimate changes in relative probabilities of future earthquakes. An approach of this nature is commonly used in the prediction of volcanic eruptions.

Another field of earthquake forecasting relies upon the use of **statistical analysis of historical or prehistoric seismicity**. Earthquake forecasts using statistical models based on historical seismicity are presently being utilized in several countries, including New Zealand, for a variety of future time intervals ranging from the 24 hours to decades (Gerstenberger et al., 2014). Such forecasts typically report the probability of an earthquake of a given magnitude occurring in a given region over a given time interval, similar to weather forecasts. Stress change modelling may also be added to gain additional insights into the spatial distribution of future earthquakes. Earthquake forecasts for individual faults typically combine **paleoseismic** data (data about prehistoric earthqakes), which enables estimation of fault slip, earthquake recurrence interval and magnitude for preceding earthquakes, with geodetic measurements to forecast the timing and magnitude of future earthquakes on the target fault. Forecasts of this nature are often limited by the apparent variability in slip, magnitude, and inter-event timing of prehistoric earthquakes that is commonly revealed in geologic studies. Stress change modelling may also be used to estimate changes in earthquake probabilities on specific faults from preceding ruptures.

D.3. Earthquake prediction

The discussion above concerns forecasting as opposed to prediction. An **earthquake prediction** is a statement that specifies the expected magnitude range, geographic location (approximate location and depth), and time interval of a future earthquake. These attributes should be stated with enough precision that the success or failure of the prediction can be assessed, and a quantitative measure of the confidence level associated with a given predictive technique can be cited. If an earthquake satisfying all of the specified attributes occurs, then the prediction can be considered successful, although a single success may be coincidental and unrepeatable, especially in highly seismically active areas. For instance, predictions made during the early part of an aftershock sequence have a greater chance of being correct, regardless of their scientific worth. If no earthquake occurs, then the prediction should be considered a **false alarm**. If an earthquake occurs without a prediction, or occurs but does not satisfy the stated attributes, it can be considered as a **failure to predict the earthquake**. The use of imprecise or incompletely defined magnitude, spatial and/or temporal ranges of a prediction increase the probability of a predictive statement being 'correct' but also reduce the usefulness of a prediction. For example, the occurrence of a magnitude 5.0 to 5.9 earthquake occurring on Earth in the next 24 hours is extremely likely, given that roughly 1300 of these events occur each year; a predictive statement specifying this occurrence would have a high success rate, irrespective of the validity of the proposed predictive method. Advantageous to any earthquake prediction is thus a confined spatial and temporal domain applicable to considering the seismic hazard for a given site or region, and a clearly defined mechanism that relates the predictive methodology to earthquake generation.

A useful form of earthquake prediction concerns the use of **earthquake early warning systems** to provide a few seconds to tens of seconds of warning of oncoming ground shaking, allowing for short-term mitigation. In this sense, it is the shaking at a specific area induced from a preceding earthquake elsewhere, rather than the initial occurrence of that earthquake, that is being predicted. Regional seismic networks that are located proximal to the earthquake source could use the characteristics of seismic waves to send network alarms to areas where earthquake shaking has not yet commenced. Tasks such as power plant shutdowns, stopping trains, stopping medical procedures, opening fire station garages so that they do not get stuck, and alerting the public can be accomplished prior to earthquake shaking. Several countries around the world, but none in Australasia, have currently operating earthquake alarm systems, or are in the development stage of such systems.

Treatment of the terms prediction and forecast as synonyms by some workers sparks considerable debate in the scientific literature; operational earthquake forecasting has been effectively deployed in some countries around the world (Jordan et al., 2011), some scientists believe earthquake prediction *senso stricto* may never be possible (e.g. Geller et al., 1997, 2011).

Supplement to E: Measurement and characterisation of earthquake shaking and faulting

Earthquake shaking is caused by the radiation of seismic energy away from the earthquake source, and provides valuable information about the earthquake. Here, we introduce the basic concepts of seismic waves, the controls on their propagation, and how they aid in documenting the location, style of movement and size of the earthquake faulting.

E.1. Seismic waves

When fault rupture occurs, seismic waves propagate out in all directions from the propagating rupture, like an expanding balloon of energy. When they reach the surface of the earth, they cause vibration of the ground, and are classified by the type of particle motion that they cause. There are two main types of waves. **Body waves** are directly radiated out from the fault rupture and travel through the Earth's interior; they include compressional waves (**P waves**) and shear waves (**S waves**). P waves cause particle movement in the same direction as direction of wave propagation, whereas S waves cause particles to move perpendicular to the direction in which the wave is travelling. **Surface waves** are caused by the interaction of the P and S waves with the earth's surface. Surface waves travel along the earth's surface, and include **Rayleigh waves**, which make the ground roll, or shear perpendicular to the surface, and **Love waves**, which shear the ground parallel to the surface.

Seismic waves travel at different speeds; P waves travel fastest, followed by S waves and surface waves. Because of the different particle motions involved, which amount to small distortions of the earth, the local velocity of a seismic wave is dependent on the Earth's properties. Most important are the compressibility, rigidity (resistance to shear), and density of the earth. Unconsolidated sediments that are loosely packed and have low density, high compressibility and low rigidity have lower seismic wave velocities than more cemented sediments and rocks.

The P, S and surface waves that radiate out from an earthquake source are recorded by a global network of **seismometers**, which are instruments that provide a continuous record of ground movement, tied to a precise clock. The instruments respond predictably to a very broad range of frequencies. Generally, closer seismographs provide better records of high frequencies and distant instruments record low frequencies. Because of the difference between the velocities of P, S and surface waves, these waves arrive in sequence at a seismometer. Seismologists use these waveforms to derive critical information about the earthquake's size, orientation, slip sense, and location.

E.2. Controls on earthquake shaking: the Influence of geology and topography on earthquake shaking characteristics

The rock through which a seismic wave passes between the point where it is generated and the point where it is recorded exerts a primary control on the arrival time of seismic waves at that station, and also on the character of earthquake shaking at a given site. In order to understand seismic hazards and predict ground motions in a region, it is necessary to understand the spatial distribution of earthquake sources, the way in which they rupture, and how source and site parameters are correlated. Earthquake shaking generally decreases with distance from the earthquake source. However, regional and local geology can greatly influence the way an earthquake is felt, because the duration, amplitude, velocity and frequency of seismic waves depend on the character of the rupturing fault (**source effects**), the distance and geology of the path from source to the felt area (**path effects**), and the geology on which the effects are observed (**site effects**).

Earthquake shaking is strongly influenced by the properties of the fault and the earthquake process. Shaking declines with distance from a source, and for a given distance from the rupture, the shaking will scale with magnitude. However, magnitude alone does not account for all the **source effects**. The recurrence interval of a fault is important because faults weaken during earthquakes and strengthen interseismically. Interseismic healing occurs over years to decades and allows the fault to support increased shear stresses. For this reason, fast-slipping interplate faults are weak, slip often, and have lower coseismic stress drops; slow-slipping intraplate faults with rupture intervals greater than 10^3 to 10^4 years are typically much stronger and exhibit larger coseismic stress drops. The displacement, and hence the seismic moment M_0, are proportional to stress drop. High stress drops are also correlated with high rupture speeds, both in terms of the velocity at which slip accumulates, and the velocity at which the rupture front moves. During faster ruptures, less shear wave energy is able to escape ahead of the rupture front. In the case of a supershear earthquake, in which the rupture front propagates faster than the shear wave velocity, the entire packet of shear wave energy is carried at the rupture front, leading to an effect analogous to a sonic boom.

The way a fault ruptures (rupture directivity, Figure 3A) gives rise to **directivity effects**. For a given rupture velocity, bilateral ruptures may rupture the entire patch up to twice as fast as a unilateral rupture, thus halving the duration of the earthquake and hence of strong ground motion. Also, bilaterally propagating earthquakes direct the rupture phases in both directions, effectively splitting the seismic energy, whereas unilateral rupture propagation focuses most of the energy in one direction (this asymmetry may help to determine which nodal plane on a CMT solution represents the fault plane). Strong ground motion is enhanced forward of the rupture front as shear waves generated by fault rupture pile up. This causes increased amplitudes but, because all the shear waves arrive at a similar time, the duration of strong ground motion is reduced. Conversely, shaking at locations behind a unilaterally propagating rupture may have a lower amplitude but longer duration. Directed ruptures therefore cause major differences in intensity and duration of shaking, which are important predictors of damage.

Most damage caused by earthquakes is off-fault, so the seismic waves must travel through a volume of rock prior to reaching the damage site. During this travel, the seismic waves will usually **attenuate** (i.e. lose energy), by 1) geometric dissipation of energy as the volume of earth within the wavefront grows larger, 2) conversion of seismic energy to heat, and 3) scattering of the energy by discontinuities in a heterogeneous rock mass. In general higher frequencies attenuate more quickly and lower frequencies more slowly. However, the rate of attenuation depends on the properties of the wave path. An exception to typical attenuation occurs when fast moving seismic waves bounce off a high velocity contrast in the lithosphere (e.g. **Moho bounce**) and return to the surface relatively unattenuated, in which case amplitudes can briefly increase at a critical distance from the source. Attenuation is estimated using attenuation equations, which describe the proportionality of amplitude and magnitude and how that changes with distance from the source. These equations are generally derived based on empirical data and attempt to account for source, path and site effects.

The properties of rock in the seismic wave travel path are important because they directly influence the amplitudes of seismic waves. The wavelength of a seismic wave of a given frequency is proportional to its velocity.

$$\lambda = \frac{V}{f}$$

If a shear wave passes into a lower velocity medium, it will reduce its wavelength. To conserve energy, the wave is amplified. Sedimentary basins have lower velocities than bedrock and can therefore produce intense amplification of incoming seismic waves. The change in velocity also changes the rate at which seismic energy is attenuated.

Site effects are the focus of an entire engineering discipline aimed at quantifying and predicting the way in which the soil under a site influences the local intensity of ground shaking. The Canterbury earthquakes provided many clear examples of this. Sites separated by 10^1-10^2 m experienced extremely different levels of shaking (Bradley et al., 2012, 2014), even though the source characteristics of the seismic waves would have been similar. The difference in acceleration between sites depends on the **amplification factor** at the site. The amplification factor describes the amplification between the top and bottom of the soil column and is widely considered to be at least partly attributable to differences in the shear wave velocity of the upper 30 m of soil underlying the site (**V_s30**). Sites with lower V_s30 will amplify (increase the acceleration of) low frequency (long period) inputs, while sites with stiffer soils (higher V_s30) will amplify relatively higher frequency (short period) inputs.

Sites with low V_s30 and low fines contents generally have very loose soils, which are susceptible to the **trampoline effect**. In this phenomenon, loose material accelerating upwards is compressed by gravitational forces and deforms elastically, a little like a trampoline mat. When the wave reaches its max amplitude, the loose grains of the upper soil column decouple from the substrate and continue to move up as the substrate accelerates downwards. The detached soil freefalls under the influence of gravity only, only to encounter the upward accelerating, stiffer sediments below the loose soil column. This produces a characteristic asymmetric acceleration with upward acceleration greater than downward acceleration.

The amplification factor can be estimated using 1D site response software that accounts for the physical and dynamic properties of the soil, and its thickness. Given the importance of spectral acceleration as a predictor of the fate of engineered structures, it is arguably more important to define site amplification characteristics than to understand seismic sources and path effects.

Engineered structures are commonly located, not only on flat land in alluvial basins, but on elevated, bedrock topography. The shape of this topography, and its orientation relative to the incoming seismic waves, can give rise to an effect called **topographic amplification**. Narrow, steep ridges located transverse to incoming seismic waves amplify seismic waves relative to other bedrock sites and sometimes even relative to adjacent valley fill sites. This may cause damage and/or collapse of even relatively well-built structures

E.3. Seismic moment, moment magnitude and energy release

The network of seismometers dotted around on the surface of the crust provide important information about the physical size and strength of the fault plane in the earthquake. Seismologists commonly use two separate parameters to describe the effects of an earthquake on the crust: 1) **seismic moment** and 2) **radiated energy**. A detailed account of the physics underlying seismology is beyond the scope of this chapter, but it is useful to basically understand the relationships that describe the relationship between seismic moment and radiated energy, in terms of the rock properties and fault geometry.

The amplitudes of earthquake waves recorded by seismometers provide a measure of the seismic moment M_O, which is used by seismologists to quantify the size of an earthquake. M_O can be directly estimated from the amplitude and duration of shear waves recorded by a seismogram using a relationship of the form:

$$logM_O = a + b \cdot log(C \times D \times \Delta^p)$$

where C is the maximum peak-to-peak amplitude of S waves on the seismogram, D is the duration between the S wave arrival and their onset with amplitude *C/d*, Δ is the epicentral distance, and $a=16.74\pm0.2$; $b=1.22\pm0.14$; $p\approx1$; $d\approx1$ are all constants (e.g. Bolt and Herraiz, 1983).

In physical terms, this seismologically-derived M_O is a product of the area A of the earthquake rupture patch in m^2, the average displacement *D in m*, and the rigidity or shear modulus of the fault (μ – its resistance to shearing) in Pa, i.e.:

$$M_O = \mu AD$$

Seismic moment therefore has the units of energy (joules), and provides the most consistent measure of earthquake size because it accounts for the fault geometry. For this reason, a magnitude scale was developed based on seismic moment, with values that are approximately consistent with the older **Richter magnitude** scale but scale with the size of the rupture. **Moment magnitude**, M_w, is calculated as:

$$M_W = 2/3log10(M_O)$$

The average displacement, D, which is incorporated in M_O, is a measure of the strain associated with the earthquake, and is proportional to the stress drop Δσ, such that:

$$\Delta\sigma \sim c\mu \frac{D}{A^{1/2}}$$

Where c is a constant typically ≈1. The seismic moment is therefore also proportional to the stress drop (represented by the average displacement) and the area of the rupture:

$$M_O = \Delta\sigma A^{3/2}$$

During an earthquake, potential energy in the crust is transformed. Seismic moment is a measure of the total energy transformed, but is not equivalent to the energy radiated by an earthquake, because it includes the energy that results in fracturing, heating, and displacement. In order to calculate the small proportion of radiated seismic energy, it is necessary to integrate the energy radiated across the full spectrum of (mostly high) radiated frequencies. A **slow slip earthquake** radiates very little high frequency energy. At the opposite end of the spectrum, a **super-shear earthquake**, rupturing faster than the shear wave velocity, radiates very large amounts of high frequency energy. For a typical rupture at slightly below shear wave velocity, the energy radiated per unit volume of ruptured fault is pretty consistent. This means that the high frequency content is relatively constantly related to the seismic moment, and so energy release can be estimated from M_O.

$$E_S = \frac{\Delta\sigma}{2\mu} M_O$$

or using typical values of stress drop (3MPa) and crustal rigidity

$$E_S = M_O \cdot 1.6 \times 10^{-5}$$

E.4. Earthquake kinematics

The slip on a fault can be defined uniquely if the strike of the fault, its dip, the magnitude of slip, and the rake (or inclination) of the slip along the fault plane, measured from the horizontal, are known. A vector can be defined by two angles and a displacement, whereas three angles (strike, dip and rake) and a displacement define a tensor. The tensor used to define earthquake slip is called the **centroid moment tensor**. A force tensor in a Cartesian coordinate system consists of nine force couples, but in practice, the dynamic deformation generated by slip on the fault, and expressed by seismic waves, is represented by a simplification of the system of forces known as a **double couple**. A double couple consists of two pairs of forces that operate in a) the fault plane parallel to the slip, and b) the normal to the fault plane perpendicular to the slip direction (called the **auxiliary plane**). This representation assumes that the waves leaving the source are 1) produced by slip on a single fault plane, and 2) propagating through a medium of uniform velocity. Complex earthquakes involving the rupture of several fault planes, and / or earthquakes in heterogeneous crust or on heterogeneous faults may violate create more complex forces termed ‘**non-double-couple**’ earthquakes.

The double couple simplification of the centroid moment tensor can be solved graphically by analysis of seismic waves. Imagine a strike-slip fault, surrounded by a number of seismometers (Figure 6). The first motion of particles in the wall rocks of the fault is sub-parallel to the slip direction of that wall. This means that seismometers on either side of the fault centroid (Figure 3a), but in the same wall, will record opposite P wave first motions (i.e. they break in opposite directions on the seismogram). Based on this behaviour, four quadrants can be defined relative to the centroid. Two diametrically opposite quadrants have first motions towards the instrument (compressive P waves), and two quadrants have first motions away from the instrument (dilational P waves). If the fault orientation or slip sense is changed (e.g. to reverse or normal, Figure 2 in textbook), the first arrival polarity will change accordingly.

Now imagine a 3-dimensional fault plane slipping within the spherical earth, and surrounded by a global network of seismometers on the earth’s surface. Each seismometer records waves propagated away from the rupture in a unique direction. Once the earthquake has been located, the 3-dimensional **take-off direction** (the direction in which the waves propagated) can be calculated based on an understanding of seismic wave propagation through the earth, and on the location of the seismometer relative to the centroid. When the take-off directions of a suitable number of waves recorded by the global network of seismometers are plotted on a lower hemisphere stereographic projection, and colour-coded by their first motion (toward = dark; away = light), dark and light quadrants are revealed that record the kinematics of the earthquake. The light and dark regions are separated by two orthogonal great circles that are the traces of the **nodal planes**. One of these nodal planes represents the fault plane, the other the auxiliary plane. It is not possible to distinguish which is which without further data.

These graphical solutions representing the earthquake tensor from seismic wave orientations and polarity are referred to as centroid moment tensor solutions (**CMT solutions**). Because they also reveal the orientation of the slipping plane and the slip vector, they are also referred to as **fault plane solutions**, or earthquake **focal mechanisms**. These solutions are now routinely derived by automated analyses of earthquake waveforms recorded by seismometer networks. For a rigorous encounter with the radiated seismic wavefield see Kennett (2002).

E.5. Earthquake location

The earth is approximately spherical, and that the line connecting two points on the earth’s surface is an arc. When an earthquake occurs, the seismic waves that arrive at seismometers located within 50 to 500 km of the earthquake will have travelled along an arc, roughly parallel to the earth’s surface. Waves arriving at seismometers located closer (further) have paths that are steeper (more complex) so for the purposes of this chapter we only consider the 50-500km range. The travel time, *t*, of the P or S wave, between the earthquake hypocentre and the seismometer, is a function of the arc distance, *d*, and of the speed (V) of the wave:

$$t_{P\ (or\ S)} = d/V_{P\ (or\ S)}$$

Because P and S waves travel at different speeds, they become increasingly separated from each other with distance from the earthquake. The time difference, Δt, between the arrival of P and S waves at a seismometer, can be measured on a seismogram and tells us how far the earthquake is located from the seismometer. Δt is given by

$$\Delta t = d\left(\frac{1}{V_S} - \frac{1}{V_P}\right)$$

The distance to an earthquake can therefore be calculated by rewriting the time delay equation in terms of distance, and employing reasonable estimates of P and S wave velocities.

$$d = \frac{\Delta t}{{}^{1}/_{V_S} - {}^{1}/_{V_P}}$$

Upper crustal **P wave velocities** are typically ~8 km s^{-1} and **S wave velocities** ~3. 5 km s^{-1}, so the lower part of the equation is slightly greater than 1/8 km s^{-1}. Using this value as a rule of thumb, the distance from earthquake to seismometer, is typically ≈8 × Δt.

This calculation gives us distance but not direction. By repeating the process for three or more seismograms that record the event, and using the earthquake-seismometer distance as a radius around each seismograph, the epicentral location can quickly be estimated. In practice, it is necessary to account for other factors including velocity variations along the travel path between the earthquake and the seismometer, the depth of the earthquake, and measurements at seismometers that are not on an upper crustal travel path.

Supplement to G: Earthquake hazards

G.1. Techniques for assessing earthquake hazards

Techniques for assessing earthquake hazards can be broken down into geological, geophysical and geotechnical investigations. The breadth and depth of seismic hazard assessments is dependent on the end user, but ultimately, the key inputs to hazard analysis are magnitude and age data for paleoseismic earthquakes, with robust chronologies to allow meaningful probabilistic analysis. The cornerstone of probabilistic seismic hazard analysis is traditional paleoseismic studies of exposed prehistoric and historic surface ruptures (see Section C7). In order to be assessed, faults must be recognised and located. However, faults are commonly not exposed and can only be located within a relatively wide zone, within which there may be distributed deformation. **Ground-penetrating radar** is commonly used to target paleoseismic sites and is useful on a scale of meters to a few tens of meters, dependent on frequency. Carefully-interpreted **Seismic reflection surveys** penetrate deeper and can illuminate both fault location and useful long-term fault activity rates but provide no data about recent seismicity (e.g. Dorn et al., 2010). **Shear wave seismic surveys** provide similar information in terms of locating the fault in the near surface, and have been shown to be sensitive to the distribution of ground deformation (e.g. Duffy et al., 2014). Furthermore, strong ground motion polarization and frequency is controlled by fracture orientation, which imparts directionality to shear wave velocities (e.g. Panzera et al., 2014). Shear wave velocity studies of directional site effects can therefore help to mitigate the damaging effects of strong ground motion on near-fault infrastructure. **Cone penetrometer testing** (CPT) is also used to locate buried faults, by defining the progressive tilt and thickness changes in layers deposited during and between episodes of fault activity. The technique is especially useful in loose sediments where trenching is difficult (e.g. Grant et al., 1997).

Evaluating seismic hazard beyond faults and across a region requires broad-based, multidisciplinary geological studies. Off-fault, geological information can be gleaned by building empirical relationships between earthquake effects and shaking intensity. Seismic shaking decreases with distance from an earthquake and effects decrease accordingly. An example is liquefaction, which occurs within an epicentral radius that depends on earthquake magnitude (e.g. Pirotta et al. 2007). Where the **liquefaction threshold** can be instrumentally established (e.g. Quigley et al., 2013), seismic hazard at a point can be evaluated using a paleoliquefaction record (e.g. Bastin et al., 2015). Similarly, dating of co-seismic rock-slope failures (e.g. Mackey and Quigley, 2014) or of **precariously balanced boulders** (Stirling et al., 2002) can both provide constraints on seismic hazard. Many of the off-fault proxies for paleoseismicity reflect site response to an earthquake, and therefore provide not only an earthquake chronology but a strong motion proxy.

G.2. Fault rupture hazards

G.2.1. Avoiding fault rupture hazards

Structures that are built on faults usually suffer the worst damage in an earthquake, followed by structures in close proximity, especially within a zone of distributed deformation. Developed nations with active fault hazards, including New Zealand, therefor provide guidelines or regulations for planning for development adjacent to active faults (Kerr et al., 2003). The New Zealand guidelines require planners to quantify the hazard posed by earthquake ground rupture in terms of fault recurrence interval, location and **fault complexity** of deformation. Where a fault trace or traces are confined within a narrow zone of metres to 10s of metres, the complexity of deformation may be classified as **well-defined**. Well-defined in this sense does not necessarily mean a single linear feature and may include several metres of width. A simple example is a linear scarp, where the well-defined trace includes everything from the top to the bottom of the scarp. If the fault scarp has degraded, the zone may be even wider as it needs to incorporate any uncertainty about the fault location. Another example of a wide zone with a well-defined complexity classification is the zone of shearing created by a strike slip fault. Well-defined simply means that discrete ground ruptures are probable within this zone.

Deformation that encompasses faulting and/or folding distributed over 10s to 100s of metres is classified as having **distributed** complexity. In many instances, well-defined complexity zones are adjoined by distributed complexity zones, particularly on the hanging wall of thrust faults and in step over zones between strike-slip segments. Poorly mapped or hidden faults are classified as **uncertain**. Under the New Zealand guidelines, minimum 20 m wide buffers are placed around the perimeter of areas of well-defined, distributed and uncertain fault complexity. These buffers encompass **fault avoidance zones** (Figure 10).

Faults are additionally classified on their surface rupturing earthquake recurrence interval. Six classes are recognised, ranging from >2 kyr to <125 kyr. During the consenting process for development in these areas, buildings are assigned importance categories, ranging from 1 (typically isolated farm buildings) to 4 (Hospitals and other buildings requiring post-disaster functionality).

Decisions regarding limitations on building within fault avoidance zones are thus based on a matrix of the fault complexity, building importance and surface rupture recurrence interval. In most instances, the building activity matrix treats zones of uncertain fault complexity as zones of distributed deformation. This means that robust reclassification of the fault as distributed or well defined offers a planning advantage by reducing either or both of the size of the fault avoidance zone and the extent of restricted building activity within the zone.

G.2.2 Mitigating fault rupture hazards
In an active tectonic environment, faults cannot be completely avoided, particularly by lifelines and major infrastructure. Some components of infrastructure, such as power, pipelines and roads, simply have to cross faults (Figure 10). In these situations, mitigation measures typically focus on maintaining post-rupture functionality. As with fault avoidance, mitigation requires some knowledge of the orientation of the fault, its kinematics, and the amount, sense and width of the zone of surface deformation. With strike slip faults in particular, the damage to linear services will include components of shear, lengthening or shortening, depending on the orientation of the service relative to the fault.

Perhaps the best example of a successful engineering mitigation of fault rupture is the Alaskan oil pipeline, which crosses the trace of the surface rupture associated with the Denali earthquake that struck Alaska in November 2002. The pipeline was built in the 1970s with considerable design input from paleoseismologists, who had located the fault within a 500 m corridor and predicted up to 6 m of horizontal and 1.5 m of vertical coseismic slip during a M_w 8 earthquake. Thirty years later, the fault slipped 4.3 m horizontally and 0.8 m vertically during a M_w 7.9 earthquake. The coseismic displacements were absorbed by movement of the pipeline on Teflon sleds, which were placed at intervals across the width of the zone of deformation. No oil was spilled, even though a subsequent survey of the pipeline joints showed that horizontal flexure extended almost 1 km back from the fault trace. A detailed summary of this nice example of fault rupture mitigation is provided by Honegger et al. (2004).

G.3. Shaking hazards

G.3.1. Avoiding shaking hazards
Many shaking hazards are mass-movement and landslide related. These hazards are best avoided by enforcement of setbacks at both top and bottom of vulnerable topography such as cliffs and gulleys. Implementation of such setbacks, however, is not trivial and requires extensive investigation of source area geometry and 3D modelling of rockfall dynamics and runout. Additional topographic hazards arising from topographic amplification of seismic waves can be avoided by establishing amplification factors prior to development using background seismicity (e.g. Buech et al., 2010). Avoidance of liquefaction and related hazards such as lateral spreading and sand blows requires detailed geotechnical investigations of liquefaction susceptibility prior to development of a site. In general, these hazards are concentrated in young alluvial sediments that are poorly consolidated and saturated to shallow depths. Sites such as this have low shear wave velocities and low resistance to penetration and can be quickly identified using borehole, CPT and shear wave surveys (e.g. Andrus and Stokoe, 2000). The length of lateral spreading fissures is inversely related to distance from a free edge, and fissuring generally accommodates extension towards a convex edge. These

factors suggest that setback distances from rivers, and avoidance of young meander point bars, provide an effective means of hazard avoidance.

G.3.2 Mitigating shaking hazards

Mitigation of shaking hazards typically involves engineering solutions. Dynamic rockfall barriers can be deployed to catch boulders detached from cliffs and hillslopes. Rock slopes can be stabilized with rock bolts, shot-crete and/or high tensile wire mesh to prevent co-seismic movement of boulders out of the face. Drainage installation in rock faces can reduce the pore pressure in joints and increase the effective normal stress in the rockmass. This reduces the potential for destabilizing pore-pressure fluctuations in the rock mass during an earthquake. Many co-seismic landslides in steep terrain occur within the soil profile. These can be mitigated by appropriate drainage installation and construction of seismically-designed retaining walls. The importance of drainage to both rock and soil stability emphasises the importance of understanding fluid flow on and within steep slopes. For more information see Volkwein et al. (2011).

In young sediments, **containment structures,** such as grout curtains parallel with rivers, can contain lateral spreading on a limited scale, but do little for liquefaction. Liquefaction rarely occurs below ~3 m because of the overburden stress, so in cases where it is necessary to build on liquefaction prone ground, improvements can be made using techniques to **densify** or **solidify** the ground. **Weight drops** provide a means of densification. **Stone columns** can be constructed by driving a casing into the ground and filling it with compacted gravel. This increases both density and drainage. **Compaction grouting** can be achieved by injecting slow moving grout into the soil, gradually forming a bulb that displaces and densifies the host soil. Whatever solution is employed, it should be verified prior to proceeding with construction. In a similar way, a very liquid cement slurry can be injected that mixes with in-situ soils, and solidifies rather than compacts them

E.4. Cascading hazards

The hazards caused by earthquakes do not end with the cessation of shaking, or even of aftershock activity (e.g. Robinson and Davies, 2013). Following large earthquakes the cascading stream of geomorphic consequences continues over timescales that range up to decades. For example, extreme rainfall events may re-mobilize landslide debris, fuelling river aggradation over decadal timescales. Where earthquake or aftershock activity coincides with extreme weather events, the rates of geomorphic processes may be greatly increased, fuelling the hazard. Mitigation of cascading hazards needs to be underpinned by sound geomorphic analysis and probabilistic modelling of causative events. For a full review of geomorphic consequences of shaking at multiple timescales see Robinson and Davies (2013).

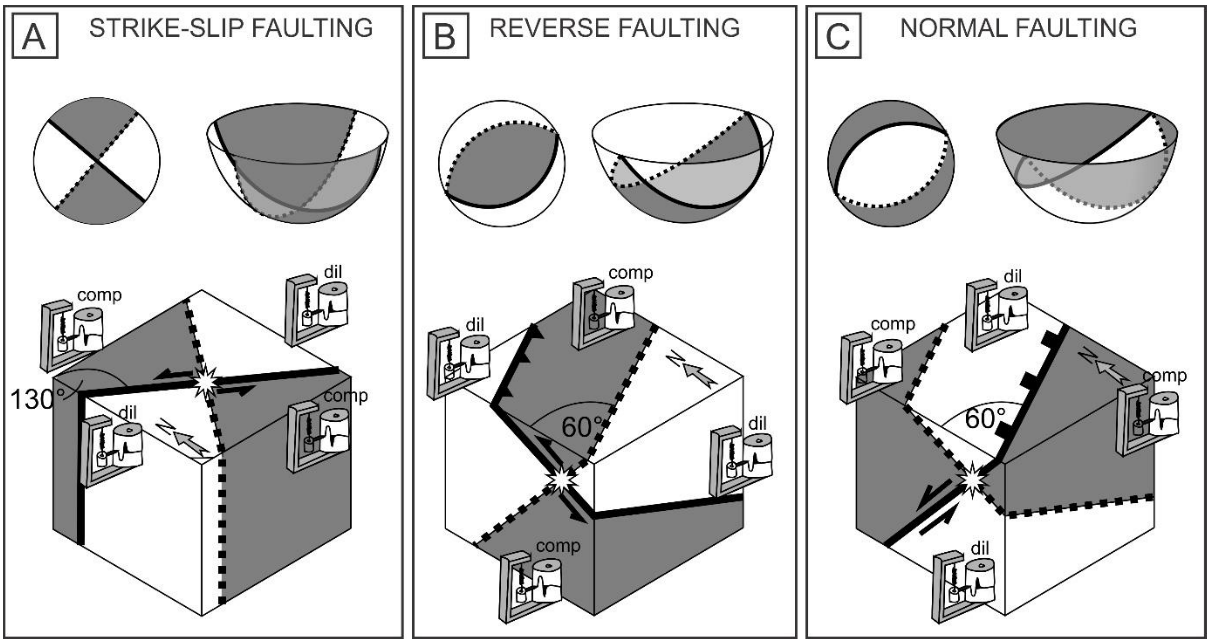

Figure S1. Fault plane solutions (beachballs) and block diagrams associated with each of the end-member types of faulting. Perspective views of fault plane solutions show how nodal planes are projected in a lower. Dark fill shows areas where the P-wave first motion recorded by the seismometers was compressive; light is dilational. Heavy solid lines on all diagrams show the fault plane, dashed lines show the auxiliary plane. A strike slip fault (A) slips parallel to the surface trace of the plane, so seismometers located around the fault trace record four quadrants, with compressive arrivals in the direction of slip of each wall of the fault. For a reverse fault (B), the hanging wall of the fault above the rupture is compressive and is represented by the dark panel in the centre of the beachball. A matching black panel can be imagined on the unseen side of the beachball, representing the down-dip sector of the footwall. For a normal fault of the same orientation (C), the hanging wall is compressive below the rupture and the central panel of the beachball, which corresponds with the hanging wall above the rupture, is light.

Table S1. Land zonation in Christchurch following the Canterbury earthquake sequence

Zone/Technical Category (TC)	Interpretation	Foundation requirements
Red	Residential Red Zone - land repair would be uncertain, costly and probably highly disruptive	Building not permitted
Green/TC1	Future land damage from liquefaction is unlikely	Standard foundations for concrete slabs or timber floors.
Green/TC2	Minor to moderate land damage from liquefaction is possible in future significant earthquakes.	Standard timber-piled foundations for houses with lightweight cladding/roofing and suspended timber floors. Otherwise, enhanced concrete foundations
Green/TC3	Moderate to significant land damage from liquefaction is possible in future significant earthquakes.	Site-specific geotechnical investigation and specific engineering foundation design is required.