Readings in Natural Hazards

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Acknowledgements

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Chapter 1: Introduction to Hazards

The term **natural hazards** refers to any natural process that poses a potential threat to human life and property. Natural hazards tend to be repetitive events and are predictable. Some of the most devastating natural hazards include:

- Earthquakes
- Volcanic eruptions
- Landslides
- Hurricanes
- Heat waves
- Tsunamis
- Wildfires
- Floods
- Droughts

Earthquakes are perhaps the most dangerous of all natural hazards. During the 20th century more than a million lives were lost worldwide due to earthquakes. Each year more than 50 earthquakes occur that are strong enough to be felt in Canada. As discussed in the section on geology, BC is subject to frequent and violent earthquake activity because of the presence of an active boundary between tectonic plates.

Periodic disturbances from hazards such as earthquakes, volcanic eruptions and flooding adversely affect:

- Fertile soil
- Available water
- Diverse land and life
- Aesthetic beauty

Fortunately, some hazards can be predicted and much can be learned from past events. With appropriate data collection and analysis, scientists can use statistical models to identify patterns and evaluate the frequency of a particular event. For example, certain soil conditions may lead to the likelihood of landslides. Learning more about natural hazards and communicating this knowledge to the public can encourage hazard mitigation measures that may reduce the damage that could potentially be incurred by disaster. The predictability of hazards allows events to be forecasted and officials to implement warning systems, such as the tsunami warning system that is in place along the Pacific coast.

Natural hazards and their effects are closely linked to our environment, and one hazardous event can set off another. For example, an area with weak soils may be at risk of a landslide. A volcanic eruption can cause a landslide, and a subduction earthquake may cause a tsunami. Hazards often influence or disrupt ecosystems: for example, human-caused landslides in the past have cut off Hell’s Gate in the Fraser River leading to massive declines of the Fraser River sockeye population and impacts across the ecosystem.

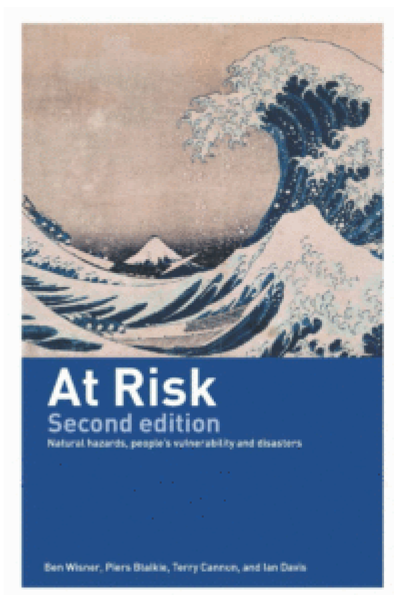
Risk from hazards can be estimated and adverse effects of hazards can be minimized through efforts such as Emergency Management BC.[1] Communities that have a more active versus reactive response to hazards stand a better chances of being prepared when adverse events occur. Emergency preparedness teams and communities need to consider land use planning, hazard-resistant construction and the protection of ecosystems.

The Canadian Disaster Database[2] is an interactive geospatial map that allows users to define their search of the disaster database by using a spatially defined area.

1. Emergency Management BC http://www.embc.gov.bc.ca/index.htm

2. Canadian Disaster Database http://cdd.publicsafety.gc.ca/srchpg-eng.aspx

Chapter 2: Introduction to Vulnerability and Risk (Excerpt from - At Risk: natural hazards, people's vulnerability and disasters (2nd edition))



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Royalties for the print versions of the book are donated to three disaster reduction networks in the South: *La Red* (Latin America), *Duryog Nivaran* (South Asia) and *Peri-Peri* (Southern Africa)

What is vulnerability?

The basic idea and some variations

We have already used the term *vulnerability* a number of times. It has a commonplace meaning: being prone to or susceptible to damage or injury. Our book is an attempt to refine this common-sense meaning in relation to natural hazards. To begin, we offer a simple working definition. By vulnerability we mean *the characteristics of a person or group and their situation that influence their capacity to anticipate, cope with, resist and recover from the impact of a natural hazard* (an extreme natural event or process). It involves a combination of factors that determine the degree to which someone's life, livelihood, property and other assets are put at risk by a discrete and identifiable event (or series or 'cascade' of such events) in nature and in society.

Some groups are more prone to damage, loss and suffering in the context of differing hazards. Key variables explaining variations of impact include class (which includes differences in wealth), occupation, caste, ethnicity, gender, disability and health status, age and immigration status (whether 'legal' or 'illegal'), and the nature and extent of social networks. The concept of vulnerability clearly involves varying magnitudes: some people experience higher levels than others. But we use the term to mean those who are more at risk: when we talk of vulnerable people, it is clear that we mean those who are at the 'worse' end of the spectrum. When used in this sense, the implied opposite of being vulnerable is sometimes indicated by our use of the term 'secure'. Other authors complement the discussions of vulnerability with the notion of 'capacity' – the ability of a group or household to resist a hazard's harmful effects and to recover easily (Anderson and Woodrow 1998; Eade 1998; IFRC 1999b; Wisner 2003a).

It should also be clear that our definition of vulnerability has a time dimension built into it: vulnerability can be measured in terms of the damage to future livelihoods, and not just as what happens to life and property at the time of the hazard event. Vulnerable groups are also those that also find it hardest to reconstruct their livelihoods following disaster, and this in turn makes them more vulnerable to the effects of subsequent hazard events. The word 'livelihood' is important in the definition. We mean by this the command an individual, family or other social group has over an income and/or bundles of resources that can be used or exchanged to satisfy its needs. This may involve information, cultural knowledge, social networks and legal rights as well as tools, land or other physical resources. Later we develop this livelihood aspect of vulnerability in an ‘Access model’. The Access model analyses the ability of people to deal with the impact of the hazards they face in terms of what level of access they have (or do not have) to the resources needed for their livelihoods before and after a hazard’s impact .

Our focus on vulnerable people leads us to give secondary consideration to natural events as determinants of disasters. Normally, vulnerability is closely correlated with socio-economic position (assuming that this incorporates race, gender, age, etc.). Although we make a number of distinctions that show it to be too simplistic to explain all disasters, in general the poor suffer more from hazards than do the rich. Although vulnerability cannot be read directly off from poverty, the two are often very highly correlated. The key point is that even a straightforward analysis on the basis of poverty and wealth as determinants of vulnerability illustrates the significance we want to attach to social forms of disaster explanation. For example, heavy rainfall may wash away the homes in wealthy hillside residential areas of California, such as Topanga Canyon (in greater Los Angeles) or the Oakland–Berkeley hills (near San Francisco), just as it does those of the poor in Rio de Janeiro (Brazil) or Caracas (Venezuela).

There are three important differences, however, between the vulnerability of the rich and the poor in such cases. Firstly, few rich people are affected if we compare the number of victims of landslides in various cities around the world. Money can buy design and engineering that minimises (but of course does not eliminate) the frequency of such events for the rich, even if they are living on an exposed slope.

Secondly, living in the hazardous canyon environment is a choice made by some of the rich in California, but not by the poor Brazilian or Philippine job seekers who live in hillside slums or on the edge of waste dumps. Without entering the psychological or philosophical definitions of ‘voluntary’ versus ‘involuntary’ risk taking (see Sjöberg 1987; Adams 1995; Caplan 2000), it should be clear that slum dwellers’ occupancy of hillsides is less voluntary than that of the corporate executive who lives in Topanga Canyon ‘for the view’. The urban poor use their location as the base for organising livelihood activities (e.g. casual labour, street trading, crafts, crime, prostitution). If the structure of urban land ownership and rent means that the closest they can get to economic opportunities is a hillside slum, people will locate there almost regardless of the landslide risk (Hardoy and Satterthwaite 1989; Fernandes and Varley 1998). This, we will argue, is a situation in which neither ‘voluntary choice’ models nor the notion of ‘bounded rationality’ (Burton et al. 1993: 61–65) are applicable.

Thirdly, the consequences of a landslide for the rich are far less severe than for the surviving poor. The homes and possessions of the rich are usually insured, and they can more easily find alternative shelter and continue with income earning activities after the hazard impact. They often also have reserves and credit. The poor, by contrast, frequently have their entire stock of capital (home, clothing, tools for artisan handicraft production, etc.) assembled at the site of the disaster. They have few if any cash reserves and are generally not considered creditworthy (despite the rapid development of 'micro-credit' schemes in a number of countries). Moreover, as emphasised above, the location of a residence itself is a livelihood resource for the urban poor. In places where workers have to commute to work over distances similar to those habitually covered by the middle class, transport can absorb a large proportion of the budget for a low income household. The poor self employed or casually employed underclass finds such transport expenses onerous. It is therefore not surprising that large numbers of working class Mexicans affected by the 1985 earthquake refused to be relocated to the outskirts of Mexico City (Robinson et al. 1986; Poniatowska 1998; da Cruz 1993; Olson et al. 1999; Olson 2000).

The nature of vulnerability

In evaluating disaster risk, the social production of vulnerability needs to be considered with at least the same degree of importance that is devoted to understanding and addressing natural hazards. Expressed schematically, our view is that the risk faced by people must be seen as a cross-cutting combination of vulnerability and hazard. Disasters are a result of the interaction of both; there cannot be a disaster if there are hazards but vulnerability is (theoretically) nil, or if there is a vulnerable population but no hazard event.1

'Hazard' refers to the natural events that may affect different places singly or in combination (coastlines, hillsides, earthquake faults, savannahs, rainforests, etc.) at different times (season of the year, time of day, over return periods of different duration). The hazard has varying degrees of intensity and severity.2 Although our knowledge of physical causal mechanisms is incomplete, some long accumulations of records (for example of hurricanes, earthquakes, snow avalanches or droughts) allows us to specify the statistical likelihood of many hazards in time and space. But such knowledge, while necessary, is far from sufficient for calculating the actual level of risk.

What we are arguing is that the risk of disaster is a compound function of the natural hazard and the number of people, characterised by their varying degrees of vulnerability to that specific hazard, who occupy the space and time of exposure to the hazard event. There are three elements here: risk (disaster), vulnerability, and hazard, whose relations we find it convenient to schematise in a pseudo-equation:

$$R = H \times V.$$

Alexander (2000: 13) distinguished between risk and vulnerability, noting that 'vulnerability refers to the potential for casualty, destruction, damage, disruption or other form of loss in a particular element: risk combines this with the probable level of loss to be expected from a predictable magnitude of hazard (which can be considered as the manifestation of the agent that produces the loss).'

A disaster occurs when a significant number of vulnerable people experience a hazard and suffer severe damage and/or disruption of their livelihood system in such a way that recovery is unlikely without external aid.3 By 'recovery' we mean the psychological and physical recovery of the victims, and the replacement of physical resources and the social relations required to use them.

Although there is still a serious lack of analysis of the linkages between vulnerability and major global processes, it is encouraging that during the last ten years many more authors and institutions have begun asking such questions. For example, it is now possible to identify more precisely how urbanisation increases hazard impact (Mitchell 1999a; Fernandez 1999; Velasquez et al. 1999) (see below).

There is a general consensus in research on disasters that the number of natural hazard events (earthquakes, eruptions, floods or cyclones) has not increased in recent decades.15 If this is true, then we need to look at the social factors that increase vulnerability (including, but not only, rising population) to explain the apparent increases in the number of disasters (as opposed to hazard events) in terms of the value of losses and the numbers of victims.

Figure 2 2 shows the number of great disasters during the second half of the twentieth century. Some of the increase may be a result of better reporting and improved communications, or the incentive for governments to declare a disaster in an attempt to win foreign aid. But the rising trend seems to be too rapid for these explanations alone (see Box 2.3 below).

Disasters are also becoming more expensive. Economic losses, and especially the share composed of insured losses, are increasing (Figure 2.3).

At this stage, it is important to review in very broad terms how certain of these various dynamic pressures contribute to the increase in disasters. We have chosen seven global processes for further attention: population change, urbanisation, war, global economic pressures (especially foreign debt), natural resource degradation, global environmental change and adverse agrarian trends. These processes are not independent of each other. They are intricately connected in a series of mutually influencing relationships that obscure causes and consequences. Also, it should be remembered that some of these processes appear both as root causes and dynamic pressures: for example, past urbanisation and past war may set up patterns that influence vulnerability hundreds of years later (the decision by the Spanish in 1521 to locate what became Mexico City on the bed of a lake they had drained once their Aztec opponents were conquered; the Second World War that resulted in a new map of Europe). In these cases urbanisation and war can be considered root causes. However, recent or current urban growth and violent conflict should be seen as dynamic pressures.

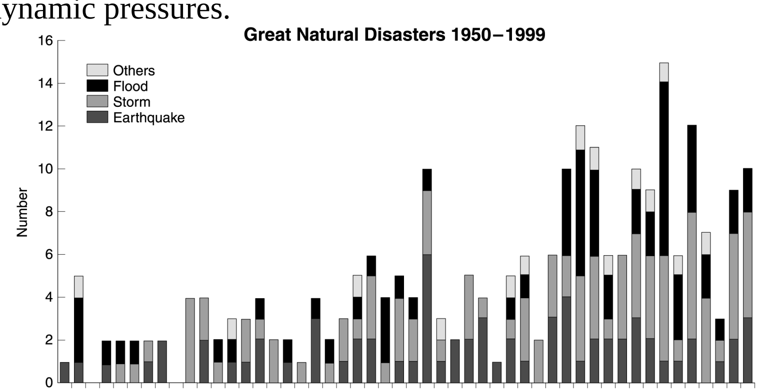

Figure 2.2 Numbers of great natural disasters 1950–1999

Note: The chart shows for each year the number of events defined as great natural catastrophes, divided up by type of event

Source: Munich Re. 2000. Great natural catastrophes – long-term statistics. Available online at http://www.munichre.com./pdf/ pm_2000_02_29_anhang3_e.pdf Adapted by kind permission of Munich Re

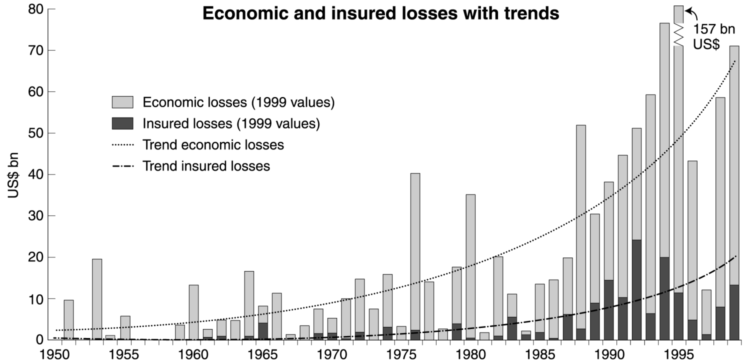

Figure 2.3 Economic and insured losses (with trends) for 1950–1999

Note: The chart presents the economic losses and insured losses – adjusted to 1999 values. The trend curves illustrate the alarming increase in catastrophic losses at the turn of the century

Source: Munich Re. 2000. Great natural catastrophes – long-term statistics. Available online at http://www.munichre.com./pdf/ pm_2000_02_29_anhang3_e.pdf Adapted by kind permission of Munich Re

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Box 2.3: Problems with disaster statistics

Where do disaster statistics come from?

As with worldwide health and population statistics, disaster statistics are reported by governments to United Nations agencies. These 'official' numbers are supplemented and cross-checked by some groups using the reports of non-governmental organisations (NGOs) and journalists. The preeminent of such institutions is the Centre for the Epidemiology of Disaster

(CRED) in Belgium (http://www.cred.be). Large reinsurance companies such as Munich Re and Swiss Re also compile international statistics on disasters. The World Bank and some of the UN regional economic commissions, such as the Economic Commission for Latin America (ECLA), have conducted studies of disaster loss and costs. Regional banks such as the Inter American Development Bank (IADB) and Asian Development Bank also study disaster statistics, but from the point of view of economic loss. The World Health Organisation (WHO) and Pan American Health Organisation (PAHO) do not maintain permanent registers of death, injury and post-disaster health consequences, but they do, on occasion, analyse and interpret such numbers.

How good are disaster statistics?

Like all numbers, disaster statistics are as good or bad as the methods used to collect them. Also disaster statistics have other specific weaknesses. Firstly, despite a large academic literature on the subject, there are no universally agreed definitions of the word ‘disaster’ (Quarantelli 1998) or other critical terms. One of the imprecise statistics often used by governments and aid organisations is the number of people ‘affected’ by a disaster. Since definitions of what it is to be ‘affected’ can vary so much, we do not use the number ‘affected’ at all in our book. ‘Injury’ is also a term that can have many meanings (Shoaf 2002; Benson 2002). The term ‘death’, too, can be problematic. For example, in the USA the death toll of the Northridge earthquake varies from 33 to 150+ depending on who defines what an earthquake-related death is: 33 died of direct or indirect earthquake injuries, 57 were defined by the LA County Coroner as dying of causes either directly or indirectly related to the earthquake; FEMA paid death benefits to survivors of more than 150 (Shoaf 2002).

Also, many extreme events that take only a few lives and affect only a local economy go completely unreported. This is an issue that a regional network of disaster researchers in Latin America have recognised by producing free, bilingual (English and Spanish) accounting software to be used to keep track of these ‘small’ disasters that could well have a highly erosive effect on development (http://www.desinventar.org/desinventar.html).

Secondly, there may be deficiencies in the reporting system itself. Many injuries may go unreported or simply are not recorded by health workers who are too busy because of the volume of care demanded in an emergency. In some countries, or regions of a country, even in ‘normal’ times there may be poor coverage of vital statistics, with many births and deaths going unrecorded. This could happen in isolated rural areas as well as densely populated squatter settlements in cities. So, some people may die in an extreme natural event whose lives were not even officially recognised as existing. Others are never found, and are ‘missing’, but are never recorded as ‘dead’, even after a considerable period of time. There is also wide historical variability in disaster data. Davidson observes (2002), ‘This is because of changes in the methods of reporting, the number of people in an affected place, systems and facilities for storing records. This all makes efforts to track historical trends in disasters even more problematic than trying to account for impacts in a single event today. Plus, of course, most records are short compared to the return period of events.'

Thirdly, there can be political pressures either to overstate or to understate casualties. If a government wishes to 'talk up' the level of relief assistance, it might exaggerate the lives lost, homes destroyed, people injured. On the other hand, if a government believes it will be criticised by its citizens for not protecting them, there may be a tendency to understate the impacts of a disaster, or to remain silent about it altogether . However, in fairness, it is very difficult to collect data on losses and damage in a timely way when undergoing the stress of the disaster itself, especially if a country has limited transport and communications. The sheer difficulty of drawing up reliable estimates should therefore be considered a fourth reason why disaster statistics should be handled with care.

Finally, when it comes to economic loss and long-term effects on development, the problem is even murkier (Benson 2003). The longer term 'knock on' effects of a disaster are conceptually difficult to model, and in most cases governments are not set up to study them (Benson and Clay 1998). Davidson (2002) puts the problem this way: '[W]ith economic effects it's difficult to assess which changes are caused by the disaster and which would have happened anyway. That is, there's always the problem that it's easier to compare before and after the disaster, but what we really should be comparing is with and without the disaster'.

Chapter 3: Fundamentals of Plate Tectonics (Excerpt from Earle, Physical Geology)

Plate tectonics is the model or theory that has been used for the past 60 years to understand Earth's development and structure — more specifically the origins of continents and oceans, of folded rocks and mountain ranges, of earthquakes and volcanoes, and of continental drift. It is explained in some detail in Chapter 10, but is introduced here because it includes concepts that are important to many of the topics covered in the next few chapters.

Key to understanding plate tectonics is an understanding of Earth's internal structure, which is illustrated in Figure 1.6. Earth's **core** consists mostly of iron. The outer core is hot enough for the iron to be liquid. The inner core, although even hotter, is under so much pressure that it is solid. The **mantle** is made up of iron and magnesium **silicate** minerals. The bulk of the mantle, surrounding the outer core, is solid rock, but is plastic enough to be able to flow slowly. Surrounding that part of the mantle is a partially molten layer (the **asthenosphere**), and the outermost part of the mantle is rigid. The **crust** — composed mostly of granite on the continents and mostly of basalt beneath the oceans — is also rigid. The crust and outermost rigid mantle together make up the **lithosphere**. The lithosphere is divided into about 20 **tectonic plates** that move in different directions on Earth's surface. (For a more accurate depiction of the components of the Earth's interior, please see Figure 9.2.)

An important property of Earth (and other planets) is that the temperature increases with depth, from close to 0°C at the surface to about 7000°C at the centre of the core. In the crust, the rate of temperature increase is about 30°C/km. This is known as the **geothermal gradient**.

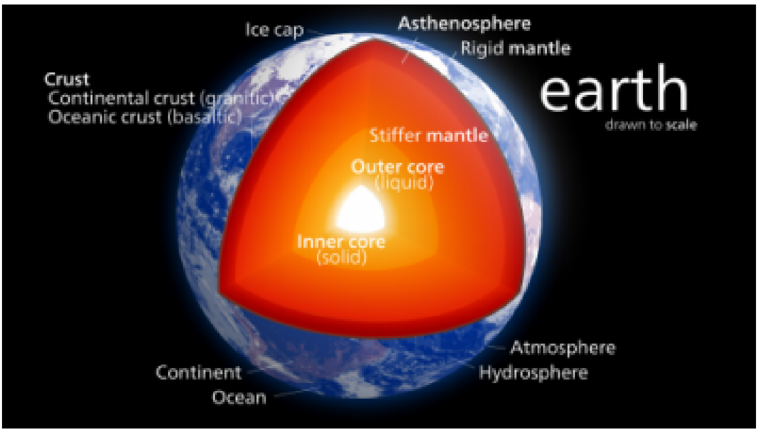


Figure 1.6 The structure of Earth's interior showing the inner and outer core, the different layers of the mantle, and the crust [Wikipedia]

Heat is continuously flowing outward from Earth's interior, and the transfer of heat from the core to the mantle causes convection in the mantle (Figure 1.7). This convection is the primary driving force for the movement of tectonic plates. At places where convection currents in the mantle are moving upward, new lithosphere forms (at ocean ridges), and the plates move apart (diverge). Where two plates are converging (and the convective flow is downward), one plate will be **subducted** (pushed down) into the mantle beneath the other. Many of Earth's major earthquakes and volcanoes are associated with convergent boundaries.

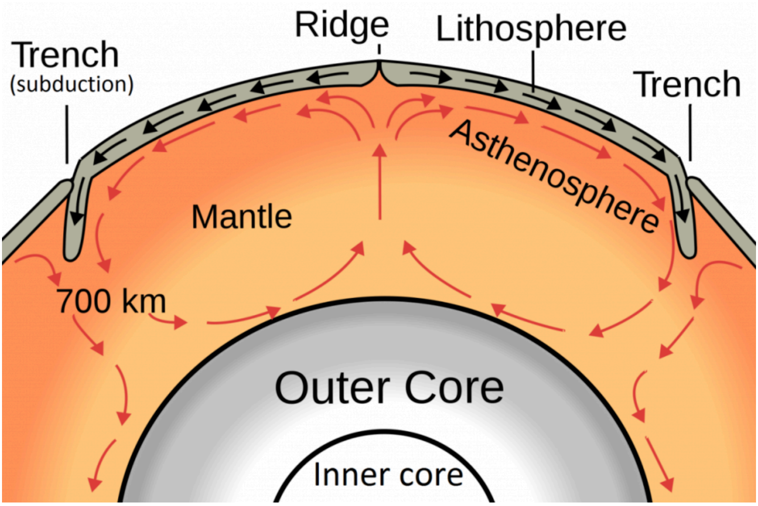


Figure 1.7 A model of convection within Earth's mantle [http://upload.wikimedia.org/wikipedia/commons/thumb/2/27/Oceanic_spreading.svg/1280px-Oceanic_spreading.svg.png]

Earth's major tectonic plates and the directions and rates at which they are diverging at sea-floor ridges, are shown in Figure 1.8.

Exercises

Exercise 1.2 Plate Motion During Your Lifetime

Using either a map of the tectonic plates from the Internet or Figure 1.8, determine which tectonic plate you are on right now, approximately how fast it is moving, and in what direction. How far has that plate moved relative to Earth's core since you were born?

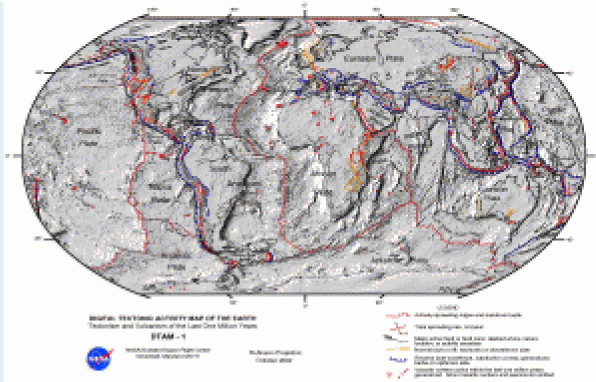


Figure 1.8 Earth's tectonic plates and tectonic features that have been active over the past 1 million years [http://commons.wikimedia.org/wiki/File:Plate_tectonics_map.gif]

Chapter 4: Earthquakes

Introduction

Learning Objectives

After carefully reading this chapter, completing the exercises within it, and answering the questions at the end, you should be able to:

- Explain how the principle of elastic deformation applies to earthquakes
- Describe how the main shock and the immediate aftershocks define the rupture surface of an earthquake, and explain how stress transfer is related to aftershocks
- Explain the process of episodic tremor and slip
- Describe the relationship between earthquakes and plate tectonics, including where we should expect earthquakes to happen at different types of plate boundaries and at what depths
- Distinguish between earthquake magnitude and intensity, and explain some of the ways of estimating magnitude
- Explain the importance of collecting intensity data following an earthquake

- Describe how earthquakes lead to the destruction of buildings and other infrastructure, fires, slope failures, liquefaction, and tsunami
- Discuss the value of earthquake predictions, and describe some of the steps that governments and individuals can take to minimize the impacts of large earthquakes

Earthquakes scare people … a lot! That’s not surprising because time and time again earthquakes have caused massive damage and many, many casualties. Anyone who has lived through a damaging earthquake cannot forget the experience (Figure 11.1). But geoscientists and engineers are getting better at understanding earthquakes, minimizing the amount of damage they cause, and reducing the number of people affected. People living in western Canada don’t need to be frightened by earthquakes, but they do need to be prepared.

Figure 11.1 A schoolroom in Courtenay damaged by the 1946 Vancouver Island earthquake. If the earthquake had not happened on a Sunday, the casualties would have been much greater. [from Earthquakes Canada, http://www.earthquakes canada. nrcan.gc.ca/historic-historique/ events/ images/19460623_1946. school.inside.jpg]

4.1 What Is an Earthquake?

An earthquake is the shaking caused by the **rupture** (breaking) and subsequent displacement of rocks (one body of rock moving with respect to another) beneath Earth's surface.

A body of rock that is under stress becomes deformed. When the rock can no longer withstand the deformation, it breaks and the two sides slide past each other. Most earthquakes take place near plate boundaries, but not necessarily right on a boundary, and not necessarily even on a pre-existing fault.

The engineering principle of **elastic deformation**, which can be used to understand earthquakes, is illustrated in Figure 11.2. The stress applied to a rock — typically because of ongoing plate movement — results in strain or deformation of the rock (Figure 11.2b). Because most rock is strong (unlike loose sand, for example), it can withstand a significant amount of deformation without breaking. But every rock has a deformation limit and will rupture (break) once that limit is reached. At that point, in the case of rocks within the crust, the rock breaks and there is displacement along the **rupture surface** (Figure 11.2c). The magnitude of the earthquake depends on the extent of the area that breaks (the area of the rupture surface) and the average amount of displacement (sliding).

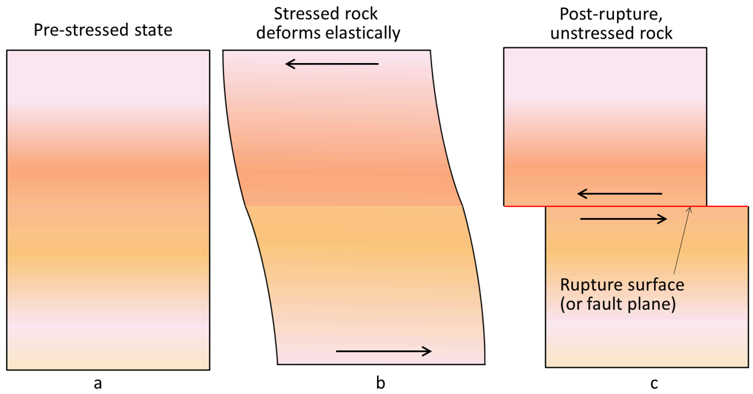


Figure 11.2 Depiction of the concept of elastic deformation and rupture, looking down. [SE]

The concept of a rupture surface, which is critical to understanding earthquakes, is illustrated in Figure 11.3. An earthquake does not happen at a point, it happens over an area within a plane, although not necessarily a *flat* plane. Within the area of the rupture surface, the amount of displacement is variable (Figure 11.3), and, by definition, it decreases to zero at the edges of the rupture surface because the rock beyond that point isn't displaced at all. The extent of a rupture surface and the amount of displacement will depend on a number of factors, including the type and strength of the rock, and the degree to which it was stressed beforehand.

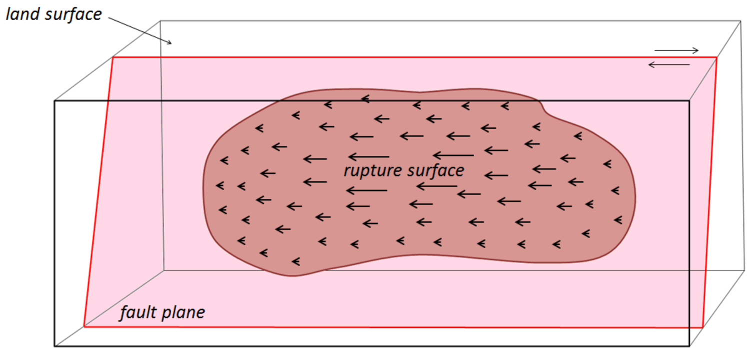


Figure 11.3 A rupture surface (dark pink), on a steeply dipping fault plane (light pink). The diagram represents a part of the crust that may be up to tens or hundreds of kilometres long. The rupture surface is the part of the fault plane along which displacement occurred. In this example, the near side of the fault is moving to the left, and the lengths of the arrows within the rupture surface represent relative amounts of displacement. [SE]

Earthquake rupture doesn't happen all at once; it starts at a single point and spreads rapidly from there. Depending on the extent of the rupture surface, the propagation of failures out from the point of initiation is typically completed within seconds to several tens of seconds (Figure 11.4). The initiation point isn't necessarily in the centre of the rupture surface; it may be close to one end, near the top, or near the bottom.

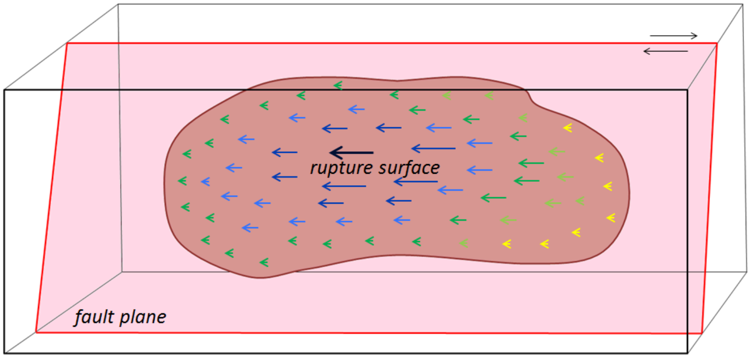


Figure 11.4 Propagation of failure on a rupture surface. In this case, the failure starts at the dark blue heavy arrow and propagates outward, reaching the left side first (green arrows) and the right side last (yellow arrows). [SE]

Figure 11.5 shows the distribution of immediate **aftershocks** associated with the 1989 Loma Prieta earthquake. Panel (b) is a section along the San Andreas Fault; this view is equivalent to what is shown in Figures 11.3 and 11.4. The area of red dots is the rupture surface; each red dot is a specific aftershock that was recorded on a seismometer. The hexagon labelled “main earthquake” represents the first or main shock. When that happened, the rock at that location broke and was displaced. That released the stress on that particular part of the fault, but it resulted in an increase of the stress on other nearby parts of the fault, and contributed to a cascade of smaller ruptures (aftershocks), in this case, over an area about 60 km long and 15 km wide.

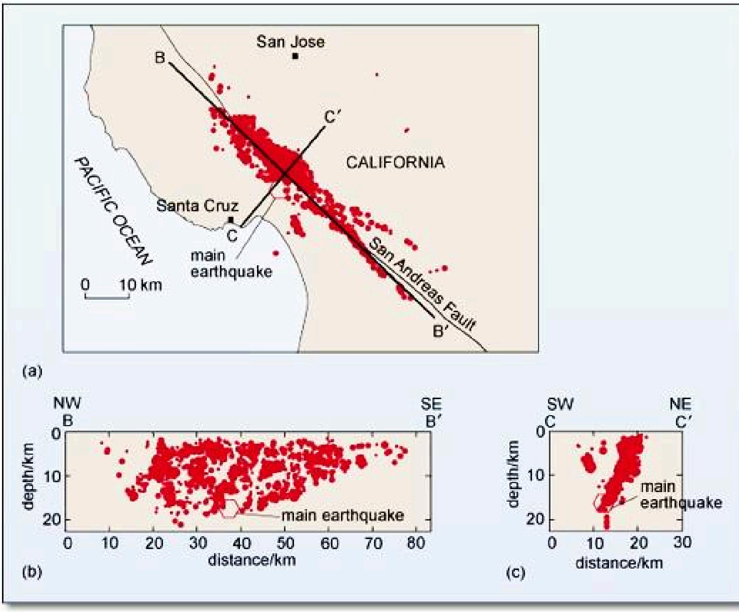


Figure 11.5 Distribution of the aftershocks of the 1989 M 6.9 Loma Prieta earthquake (a: plan view, b: section along the fault, c: section across the fault.) [from Open University under CC Sharealike, http://www.open.edu/openlearnworks/mod/page/view.php?id=45426]

So, what exactly is an aftershock then? An aftershock is an earthquake just like any other, but it is one that can be shown to have been triggered by **stress transfer** from a preceding earthquake. Within a few tens of seconds of the main Loma Prieta earthquake, there were hundreds of smaller aftershocks; their distribution defines the area of the rupture surface.

Aftershocks can be of any magnitude. Most are smaller than the earthquake that triggered them, but they can be bigger. The aftershocks shown in Figure 11.5 all happened within seconds or minutes of the main shock, but aftershocks can be delayed for hours, days, weeks, or even years. As already noted, aftershocks are related to stress transfer. For example, the main shock of the Loma Prieta earthquake triggered aftershocks in the immediate area, which triggered more in the surrounding area, eventually extending for 30 km along the fault in each direction and for 15 km toward the surface. But the earthquake as a whole also changed the stress on adjacent parts of the San Andreas Fault. This effect, which has been modelled for numerous earthquakes and active faults around the world, is depicted in Figure 11.6. Stress was reduced in the area of the rupture (blue), but was increased at either end of the rupture surface (red and yellow).

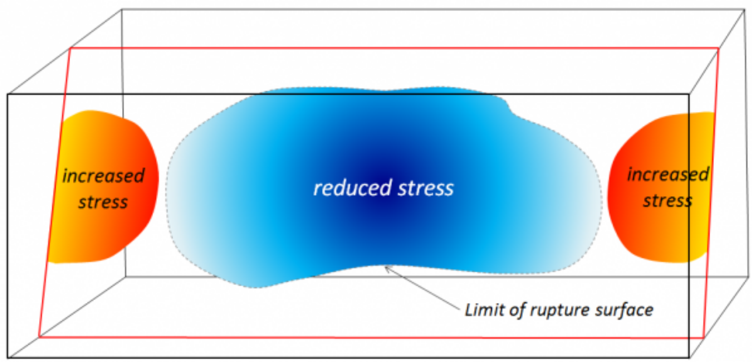


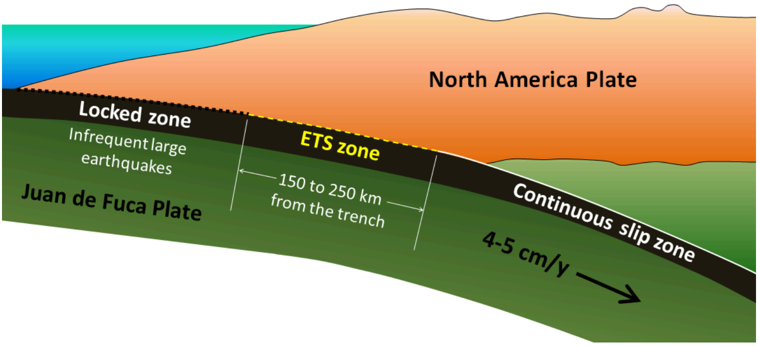
Figure 11.6 Depiction of stress changes related to an earthquake. Stress decreases in the area of the rupture surface, but increases on adjacent parts of the fault. [by SE based on data from 2010 Laguna Salada earthquake by Stein and Toda at: http://supersites.earthobservations.org/Baja_stress.png]

Stress transfer isn't necessarily restricted to the fault along which an earthquake happened. It will affect the rocks in general around the site of the earthquake and may lead to increased stress on other faults in the region. The effects of stress transfer don't necessarily show up right away. Segments of faults are typically in some state of stress, and the transfer of stress from another area is only rarely enough to push a fault segment beyond its limits to the point of rupture. The stress that is added by stress transfer accumulates along with the ongoing buildup of stress from plate motion and eventually leads to another earthquake.

Episodic tremor and slip

Episodic tremor and slip (ETS) is periodic slow sliding along part of a subduction boundary. It does not produce recognizable earthquakes, but does produce seismic tremor (rapid seismic vibrations on a seismometer). It was first discovered on the Vancouver Island part of the Cascadia subduction zone by Geological Survey of Canada geologists Herb Dragert and Gary Rogers.*

The boundary between the subducting Juan de Fuca Plate and the North America Plate can be divided into three segments, as shown below. The cold upper part of the boundary is locked. The plates are stuck and don’t move, except with very large earthquakes that happen *approximately* every 500 years (the last one was M8.5+ in January 26, 1700). The warm lower part of the boundary is sliding continuously because the warm rock is weaker. The central part of the boundary isn’t cold enough to be stuck, but isn’t warm enough to slide continuously. Instead it slips episodically, approximately every 14 months for about 2 weeks, moving a few centimetres each time.



You might be inclined to think that it's a good thing that there is periodic slip on this part of the plate because it releases some of the tension and reduces the risk of a large earthquake. In fact, the opposite is likely the case. The movement along the ETS part of the plate boundary acts like a medium-sized earthquake and leads to stress transfer to the adjacent locked part of the plate. Approximately every 14 months, during the two-week ETS period, there is a transfer of stress to the shallow locked part of the Cascadia subduction zone, and therefore an increased chance of a large earthquake.

Since 2003, ETS processes have also been observed on subduction zones in Mexico and Japan. [SE drawing]

*Rogers, G. and Dragert, H., 2003, Episodic tremor and slip on the Cascadia subduction zone: the chatter of silent slip, Science, V. 300, p. 1942-1943.

4.2 Earthquakes and Plate Tectonics

The distribution of earthquakes across the globe is shown in Figure 11.7. It is relatively easy to see the relationships between earthquakes and the plate boundaries. Along divergent boundaries like the mid-Atlantic ridge and the East Pacific Rise, earthquakes are common, but restricted to a narrow zone close to the ridge, and consistently at less than 30 km depth. Shallow earthquakes are also common along transform faults, such as the San Andreas Fault. Along subduction zones, as we saw in Chapter 10, earthquakes are very abundant, and they are increasingly deep on the landward side of the subduction zone.

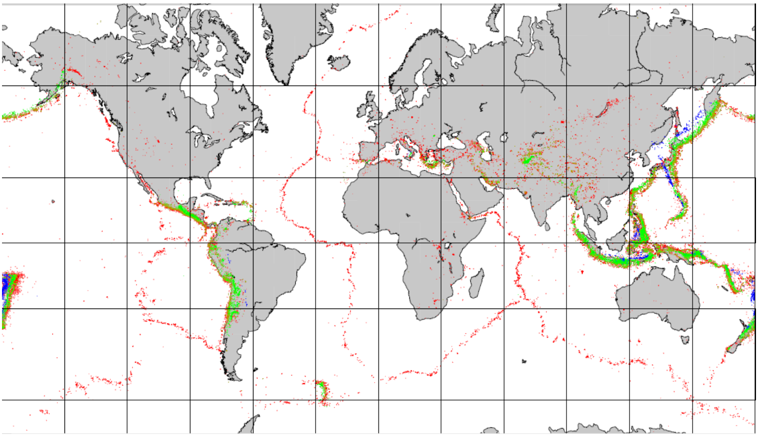


Figure 11.7 General distribution of global earthquakes of magnitude 4 and greater from 2004 to 2011, colour coded by depth (red: 0-33 km, orange 33-70 km, green: 70-300 km, blue: 300-700 km) [from Dale Sawyer, Rice University, http://plateboundary.rice.edu ,used with permission]

Earthquakes are also relatively common at a few intraplate locations. Some are related to the buildup of stress due to continental rifting or the transfer of stress from other regions, and some are not well understood. Examples of intraplate earthquake regions include the Great Rift Valley area of Africa, the Tibet region of China, and the Lake Baikal area of Russia.

Earthquakes at Divergent and Transform Boundaries

Figure 11.8 provides a closer look at magnitude (M) 4 and larger earthquakes in an area of divergent boundaries in the mid-Atlantic region near the equator. Here, as we saw in Chapter 10, the segments of the mid-Atlantic ridge are offset by some long transform faults. Most of the earthquakes are located along the transform faults, rather than along the spreading segments, although there are clusters of earthquakes at some of the ridge-transform boundaries. Some earthquakes do occur on spreading ridges, but they tend to be small and infrequent because of the relatively high rock temperatures in the areas where spreading is taking place.

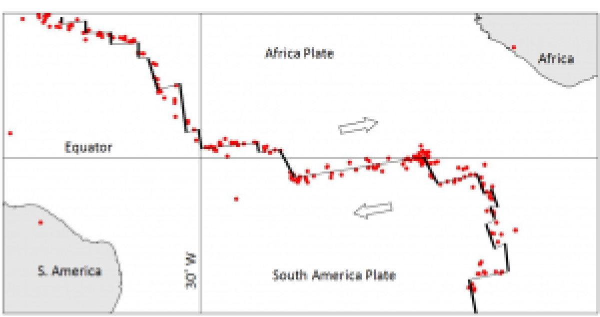


Figure 11.8 Distribution of earthquakes of M4 and greater in the area of the mid-Atlantic ridge near the equator from 1990 to 1996. All are at a depth of 0 to 33 km [SE after Dale Sawyer, Rice University, http://plateboundary.rice.edu]

Earthquakes at Convergent Boundaries

The distribution and depths of earthquakes in the Caribbean and Central America area are shown in Figure 11.9. In this region, the Cocos Plate is subducting beneath the North America and Caribbean Plates (ocean-continent convergence), and the South and North America Plates are subducting beneath the Caribbean Plate (ocean-ocean convergence). In both cases, the earthquakes get deeper with distance from the trench. In Figure 11.9, the South America Plate is shown as being subducted beneath the Caribbean Plate in the area north of Colombia, but since there is almost no earthquake activity along this zone, it is questionable whether subduction is actually taking place.

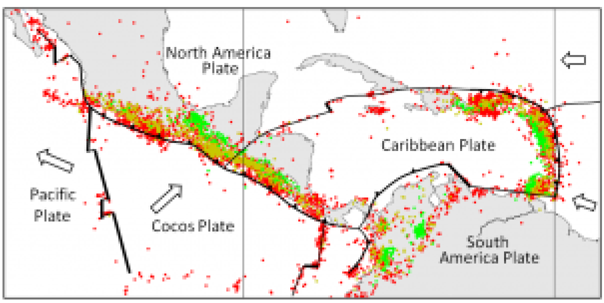


Figure 11.9 Distribution of earthquakes of M4 and greater in the Central America region from 1990 to 1996 (red: 0-33 km, orange: 33-70 km, green: 70-300 km, blue: 300-700 km) (Spreading ridges are heavy lines, subduction zones are toothed lines, and transform faults are light lines.) [SE after Dale Sawyer, Rice University, http://plateboundary.rice.edu]

There are also various divergent and transform boundaries in the area shown in Figure 11.9, and as we've seen in the mid-Atlantic area, most of these earthquakes occur along the transform faults.

The distribution of earthquakes with depth in the Kuril

Islands of Russia in the northwest Pacific is shown in Figure 11.10. This is an ocean-ocean convergent boundary. The small red and yellow dots show background seismicity over a number of years, while the larger white dots are individual shocks associated with a M6.9 earthquake in April 2009. The relatively large earthquake took place on the upper part of the plate boundary between 60 km and 140 km inland from the trench. As we saw for the Cascadia subduction zone, this is where large subduction earthquakes are expected to occur.

In fact, all of the very large earthquakes — M9 or higher — take place at subduction boundaries because there is the potential for a greater width of rupture zone on a gently dipping boundary than on a steep transform boundary. The largest earthquakes on transform boundaries are in the order of M8.

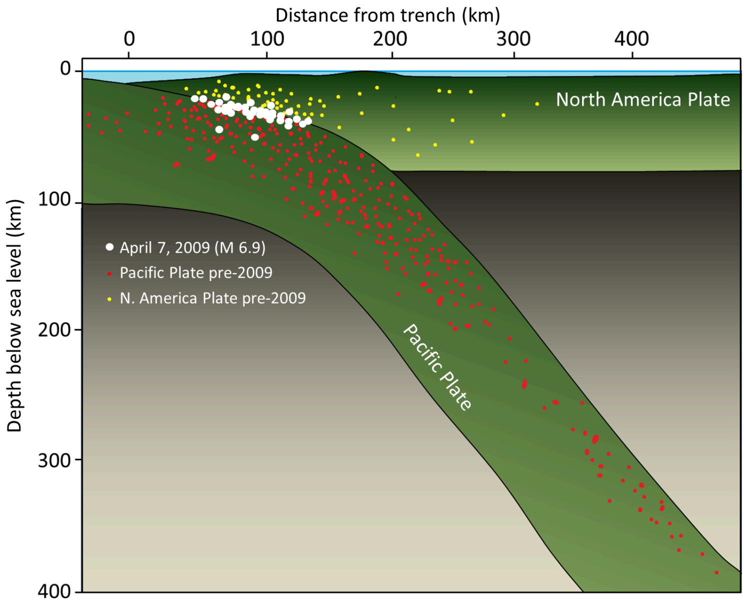


Figure 11.10 Distribution of earthquakes in the area of the Kuril Islands, Russia (just north of Japan) (White dots represent the April 2009 M6.9 earthquake. Red and yellow dots are from background seismicity over several years prior to 2009.) [SE after Gavin Hayes, from data at http://earthquake.usgs.gov/earthquakes/eqarchives/subduction_zone/us2009fdak/szgc/ku6_trench.pdf]

The background seismicity at this convergent boundary, and on other similar ones, is predominantly near the upper side of the subducting plate. The frequency of earthquakes is greatest near the surface and especially around the area where large subduction quakes happen, but it extends to at least 400 km depth. There is also significant seismic activity in the overriding North America Plate, again most commonly near the region of large quakes, but also extending for a few hundred kilometres away from the plate boundary.

The distribution of earthquakes in the area of the

India-Eurasia plate boundary is shown in Figure 11.11. This is a continent-continent convergent boundary, and it is generally assumed that although the India Plate continues to move north toward the Asia Plate, there is no actual subduction taking place. There are transform faults on either side of the India Plate in this area.

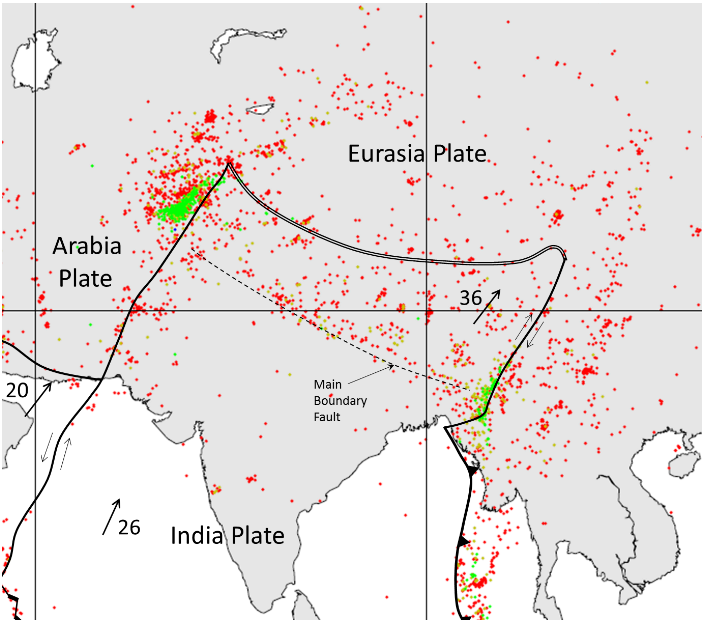


Figure 11.11 Distribution of earthquakes in the area where the India Plate is converging with the Asia Plate (data from 1990 to 1996, red: 0-33 km, orange: 33-70 km, green: 70-300 km). (Spreading ridges are heavy lines, subduction zones are toothed lines, and transform faults are light lines. The double line along the northern edge of the India Plate indicates convergence, but not subduction. Plate motions are shown in mm/y.) [SE after Dale Sawyer, Rice University, http://plateboundary.rice.edu]

The entire northern India and southern Asia region is very seismically active. Earthquakes are common in northern

India, Nepal, Bhutan, Bangladesh and adjacent parts of China, and throughout Pakistan and Afghanistan. Many of the earthquakes are related to the transform faults on either side of the India Plate, and most of the others are related to the significant tectonic squeezing caused by the continued convergence of the India and Asia Plates. That squeezing has caused the Asia Plate to be thrust over top of the India Plate, building the Himalayas and the Tibet Plateau to enormous heights. Most of the earthquakes of Figure 11.11 are related to the thrust faults shown in Figure 11.12 (and to hundreds of other similar ones that cannot be shown at this scale). The southernmost thrust fault in Figure 11.12 is equivalent to the Main Boundary Fault in Figure 11.11.

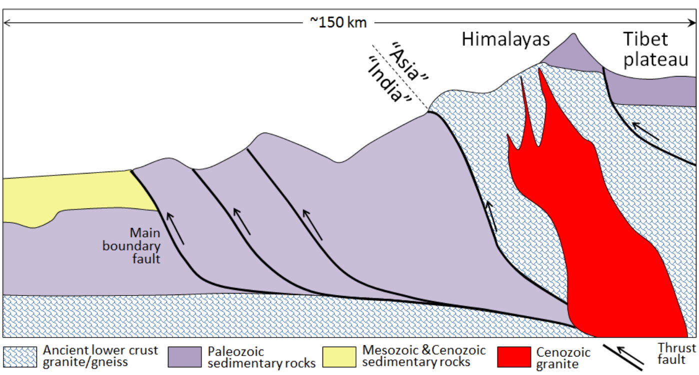


Figure 11.12 Schematic diagram of the India-Asia convergent boundary, showing examples of the types of faults along which earthquakes are focussed. The devastating Nepal earthquake of May 2015 took place along one of these thrust faults. [SE after D. Vouichard, from a United Nations University document at: http://archive.unu.edu/unupress/unupbooks/80a02e/80A02E05.htm]

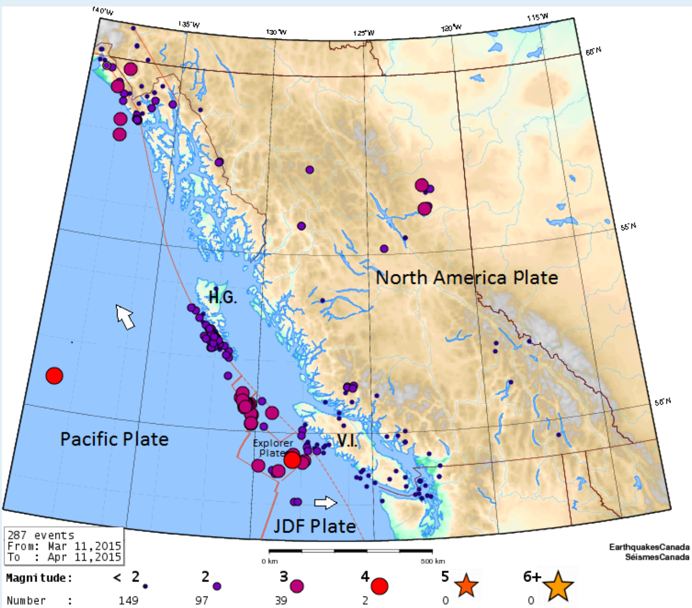
There is a very significant concentration of both shallow and deep (greater than 70 km) earthquakes in the northwestern part of Figure 11.11. This is northern

Afghanistan, and at depths of more than 70 km, many of these earthquakes are within the mantle as opposed to the crust. It is interpreted that these deep earthquakes are caused by northwestward subduction of part of the India Plate beneath the Asia Plate in this area.

Exercises

Exercise 11.1 Earthquakes in British Columbia

This map shows the incidence and magnitude of earthquakes in British Columbia over a one-month period in March and April 2015.



1. What is the likely origin of the earthquakes between the Juan de Fuca (JDF) and Explorer Plates?

2. The string of small earthquakes adjacent to Haida Gwaii (H.G.) coincides closely with the rupture surface of

the 2012 M7.7 earthquake in that area. How might these earthquakes be related to that one?

3. Most of the earthquakes around Vancouver Island (V.I.) are relatively shallow. What is their likely origin?

4. Some of the earthquakes in B.C. are interpreted as being caused by natural gas extraction (including fracking). Which of the earthquakes here could fall into this category?

4.3 Measuring Earthquakes

There are two main ways to measure earthquakes. The first of these is an estimate of the energy released, and the value is referred to as **magnitude**. This is the number that is typically used by the press when a big earthquake happens. It is often referred to as “Richter magnitude,” but that is a misnomer, and it should be just “magnitude.” There are many ways to measure magnitude — including Charles Richter’s method developed in 1935 — but they are all ways to estimate the same number: the amount of energy released.

The other way of assessing the impact of an earthquake is to assess what people felt and how much damage was done. This is known as **intensity.** Intensity values are assigned to locations, rather than to the earthquake itself, and therefore intensity can vary widely, depending on the proximity to the earthquake and the types of materials and conditions of the subsurface.

Earthquake Magnitude

Before we look more closely at magnitude we need to review what we know about body waves, and look at surface waves. Body waves are of two types, P-waves, or primary or compression waves (like the compression of the coils of a spring), and S-waves, or secondary or shear waves (like the flick of a rope). An example of P and S seismic wave records is shown in Figure 11.13. The critical parameters for the measurement of Richter magnitude are labelled, including the time interval between the arrival of the P- and S-waves — which is used to determine the distance from the earthquake to the seismic station, and the amplitude of the S waves — which is used to estimate the magnitude of the earthquake.

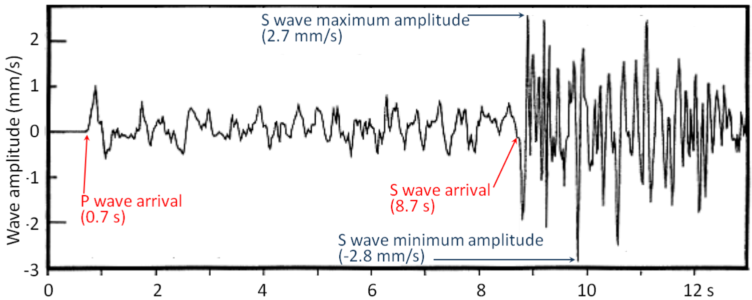


Figure 11.13 P-waves and S-waves from a small (M4) earthquake that took place near Vancouver Island in 1997. [SE]

When body waves (P or S) reach Earth's surface, some of their energy is transformed into surface waves, of which there are two main types, as illustrated in Figure 11.14. **Rayleigh waves** are characterized by vertical motion of the ground surface, like waves on water, while **Love waves** are characterized by horizontal motion. Both Rayleigh and Love waves are about 10% slower than S-waves (so they arrive later at a seismic station). Surface waves typically have greater amplitudes than body waves, and they do more damage.

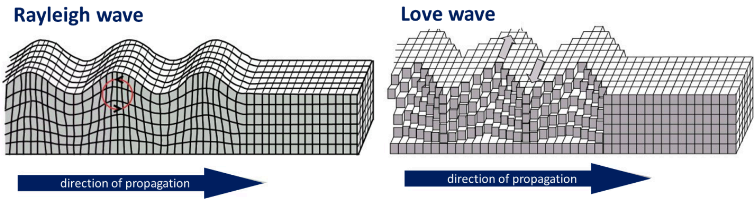


Figure 11.14 Depiction of seismic surface waves [SE after: https://en.wikipedia.org/wiki/Rayleigh_wave#/media/File:Rayleigh_wave.jpg and https://en.wikipedia.org/wiki/Love_wave#/media/File:Love_wave.jpg]

Other important terms for describing earthquakes are **hypocentre** (or **focus**) and **epicentre**. The hypocentre is the actual location of an individual earthquake shock at depth in the ground, and the epicentre is the point on the land surface directly above the hypocentre (Figure 11.15).

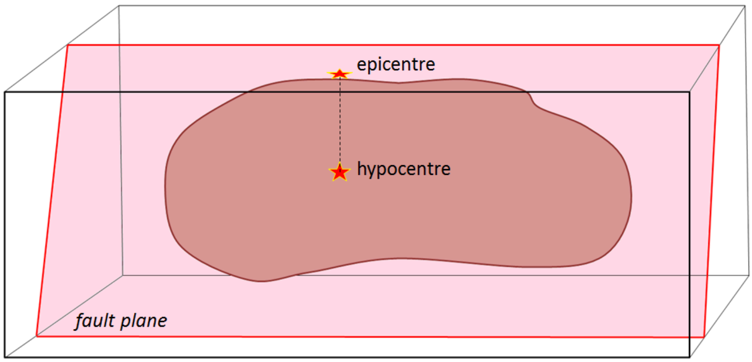


Figure 11.15 Epicentre and hypocentre [SE]

A number of methods for estimating magnitude are listed in Table 11.1. Local magnitude (ML) was widely used until late in the 20th century, but **moment magnitude** (MW) is now more commonly used because it gives more accurate estimates (especially with larger earthquakes) and can be applied to earthquakes at any distance from a seismometer. Surface-wave magnitudes can also be applied to measure distant large earthquakes.

Because of the increasing size of cities in earthquake-prone areas (e.g., China, Japan, California) and the increasing sophistication of infrastructure, it is becoming important to have very rapid warnings and magnitude estimates of earthquakes that have already happened. This can be achieved by using P-wave data to determine magnitude because P-waves arrive first at seismic stations, in many cases several seconds ahead of the more damaging S-waves and surface waves. Operators of electrical grids, pipelines, trains, and other infrastructure can use the information to automatically shut down systems so that damage and casualties can be limited.

Type	**M Range**	**Dist. Range**	**Comments**
Local or Richter (ML)	2 to 6	0-400 km	The original magnitude relationship defined in 1935 by Richter and Gutenberg. It is based on the maximum amplitude of S-waves recorded on a Wood-Anderson torsion seismograph. ML values can be calculated using data from modern instruments. L stands for local because it only applies to earthquakes relatively close to the seismic station.
Moment (MW)	> 3.5	All	Based on the **seismic moment** of the earthquake, which is equal to the average amount of displacement on the fault times the fault area that slipped. It can also be estimated from seismic data if the seismometer is tuned to detect long-period body waves.
Surface wave (MS)	5 to 8	20 to 180°	A magnitude for distant earthquakes based on the amplitude of surface waves measured at a period near 20 s.
P-wave	2 to 8	Local	Based on the amplitude of P-waves. This technique is being increasingly used to provide very rapid magnitude estimates so that early warnings can be sent to utility and transportation operators to shut down equipment before the larger (but slower) S-waves and surface waves arrive.

Table 11.1 A summary of some of the different methods for estimating earthquake magnitude. [SE]

Exercises

Exercise 11.2 Moment Magnitude Estimates from Earthquake Parameters

A moment magnitude calculation tool is available at: http://solr.bccampus.ca:8001/bcc/items/24da5970-c4f3-4ab9-98ed-089a7caca242/1/. You can use it to estimate the moment magnitude based on the approximate length, width, and displacement values provided in the following table:

Length (km)	Width (km)	Displacement (m)	Comments	MW?
60	15	4	The 1946 Vancouver Island earthquake	
0.4	0.2	.5	The small Vancouver Island earthquake shown in Figure 11.13	
20	8	4	The 2001 Nisqually earthquake described in Exercise 11.3	
1,100	120	10	The 2004 Indian Ocean earthquake	
30	11	4	The 2010 Haiti earthquake	

The largest recorded earthquake had a magnitude of 9.5. Could there be a 10? You can answer that question using this tool. See what numbers are needed to make MW = 10. Are they reasonable?

The magnitude scale is logarithmic; in fact, the amount of energy released by an earthquake of M4 is 32 times higher than that released by one of M3, and this ratio applies to all intervals in the scale. If we assign an arbitrary energy level of 1 unit to a M1 earthquake the energy for quakes up to M8 will be as shown on the following chart:

Magnitude	Energy
1	1
2	32
3	1,024
4	32,768
5	1,048,576
6	33,554,432
7	1,073,741,824
8	34,359,738,368

In any given year, when there is a large earthquake on Earth (M8 or M9), the amount of energy released by that one event will likely exceed the energy released by all smaller earthquake events combined.

Earthquake Intensity

The intensity of earthquake shaking at any location is determined not only by the magnitude of the earthquake and its distance, but also by the type of underlying rock or unconsolidated materials. If buildings are present, the size and type of buildings (and their inherent natural vibrations) are also important.

Intensity scales were first used in the late 19th century, and then adapted in the early 20th century by Giuseppe Mercalli and modified later by others to form what we know call the modified Mercalli intensity scale (Table 11.2). Intensity estimates are important because they allow us to characterize parts of any region into areas that are especially prone to strong shaking versus those that are not. The key factor in this regard is the nature of the underlying geological materials, and the weaker those are, the more likely it is that there will be strong shaking. Areas underlain by strong solid bedrock tend to experience much less shaking than those underlain by unconsolidated river or lake sediments.

I Not felt	Not felt except by a very few under especially favourable conditions
II Weak	Felt only by a few persons at rest, especially on upper floors of buildings
III Weak	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 Light	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 Moderate	Felt by nearly everyone; many awakened; some dishes, windows broken; unstable objects overturned; pendulum clocks may stop
VI Strong	Felt by all, many frightened; some heavy furniture moved; a few instances of fallen plaster; damage slight
VII Very Strong	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
VIII Severe	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 Violent	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 Extreme	Some well-built wooden structures destroyed; most masonry and frame structures destroyed with foundations; rails bent
XI Extreme	Few, if any (masonry), structures remain standing; bridges destroyed; broad fissures in ground; underground pipelines completely out of service; earth slumps and land slips in soft ground; rails bent greatly
XII Extreme	Damage total; waves seen on ground surfaces; lines of sight and level distorted; objects thrown upward into the air

Table 11.2 The modified Mercalli intensity scale. [from http://en.wikipedia.org/wiki/Mercalli_intensity_scale]

An example of this effect is the 1985 M8 earthquake that struck the Michoacán region of western Mexico, southwest of Mexico City. There was relatively little damage in the area around the epicentre, but there was tremendous damage and about 5,000 deaths in heavily populated Mexico City some 350 km from the epicentre. The key reason for this is that Mexico City was built largely on the unconsolidated and water-saturated sediment of former Lake Texcoco. These sediments resonate at a frequency of about two seconds, which was similar to the frequency of the body waves that reached the city. For the same reason that a powerful opera singer can break a wine glass by singing the right note, the amplitude of the seismic waves was amplified by the lake sediments. Survivors of the disaster recounted that the ground in some areas moved up and down by about 20 cm every two seconds for over two minutes. Damage was greatest to buildings between 5 and 15 storeys tall, because they also resonated at around two seconds, which amplified the shaking.

Exercises

Exercise 11.3 Estimating Intensity from Personal Observations

The following observations were made by residents of the Nanaimo area during the M6.8 Nisqually earthquake near Olympia, Washington, in 2001. Estimate the Mercalli intensities using Table 11.2.

Building Type	Floor	Shaking Felt	Lasted (seconds)	Description of Motion	Intensity?
House	1	no	10	Heard a large rumble lasting not even 10 s, mirror swayed	
House	2	moderate	60	Candles, pictures and CDs on bookshelf moved, towels fell off racks	
House	1	no		Pots hanging over stove moved and crashed together	
House	1	weak		Rolling feeling with a sudden stop, picture fell off mantle, chair moved	
Apartment	1	weak	10	Sounded like a big truck then everything shook for a short period	

House	1	moderate	30	20-30	Teacups rattled but didn't fall off
Institution	2	moderate		15	Creaking sounds, swaying movement of shelving
House	1	moderate	30	15-30	Bed banging against the wall with me in it, dog barking aggressively

An intensity map for the 1946 M7.3 Vancouver Island earthquake is shown in Figure 11.16. The intensity was greatest in the central island region where, in some communities, chimneys were damaged on more than 75% of buildings, some roads were made impassable, and a major rock slide occurred. The earthquake was felt as far north as Prince Rupert, as far south as Portland Oregon, and as far east as the Rockies

Figure 11.16 Intensity map for the 1946 M7.3 Vancouver Island earthquake. [from: http://www.earthquakescanada.nrcan.gc.ca/historic-historique/events/19460623-eng.php]

4.4 The Impacts of Earthquakes

Some of the common impacts of earthquakes include structural damage to buildings, fires, damage to bridges and highways, initiation of slope failures, liquefaction, and tsunami. The types of impacts depend to a large degree on where the earthquake is located: whether it is predominantly urban or rural, densely or sparsely populated, highly developed or underdeveloped, and of course on the ability of the infrastructure to withstand shaking.

As we've seen from the example of the 1985 Mexico earthquake, the geological foundations on which structures are built can have a significant impact on earthquake shaking. When an earthquake happens, the seismic waves produced have a wide range of frequencies. The energy of the higher frequency waves tends to be absorbed by solid rock, while the lower frequency waves (with periods slower than one second) pass through the solid rock without being absorbed, but are eventually absorbed and amplified by soft sediments. It is therefore very common to see much worse earthquake damage in areas underlain by soft sediments than in areas of solid rock. A good example of this is in the Oakland area near San Francisco, where parts of a two-layer highway built on soft sediments collapsed during the 1989 Loma Prieta earthquake (Figure 11.17).

Figure 11.17 A part of the Cypress Freeway in Oakland California that collapsed during the 1989 Loma Prieta earthquake. [from: http://upload.wikimedia.org/wikipedia/ commons/9/91/ Cypress_collapsed.jpg]

Building damage is also greatest in areas of soft sediments, and multi-storey buildings tend to be more seriously damaged than smaller ones. Buildings can be designed to withstand most earthquakes, and this practice is increasingly applied in earthquake-prone regions. Turkey is one such region, and even though Turkey had a relatively strong building code in the 1990s, adherence to the code was poor, as builders did whatever they could to save costs, including using inappropriate materials in concrete and reducing the amount of steel reinforcing. The result was that there were over 17,000 deaths in the 1999 M7.6 Izmit earthquake (Figure 11.18). After two devastating earthquakes that year, Turkish authorities strengthened the building code further, but the new code has been applied only in a few regions, and enforcement of the code is still weak, as revealed by the amount of damage from a M7.1 earthquake in eastern Turkey in 2011.

Figure 11.18 Buildings damaged by the 1999 earthquake in the Izmit area, Turkey. [from U.S. Geological Survey at: http://gallery.usgs.gov/sets/1999_Izmit,_Turkey_Earthquake/thumb/_/1]

Fires are commonly associated with earthquakes because fuel pipelines rupture and electrical lines are damaged when the ground shakes (Figure 11.19). Most of the damage in the great 1906 San Francisco earthquake was caused by massive fires in the downtown area of the city (Figure 11.20). Some 25,000 buildings were destroyed by those fires, which were fuelled by broken gas pipes. Fighting the fires was difficult because water mains had also ruptured. The risk of fires can be reduced through P-wave early warning systems if utility operators can reduce pipeline pressure and close electrical circuits.

Figure 11.19 Some of the effects of the 2011 Tohoku earthquake in the Sendai area of Japan. An oil refinery is on fire, and a vast area has been flooded by a tsunami. [from: http://en.wikipedia.org/wiki/2011_T%C5%8Dhoku_earthquake_and_tsunami#/media/File:SH-60B_helicopter_flies_over_Sendai.jpg]

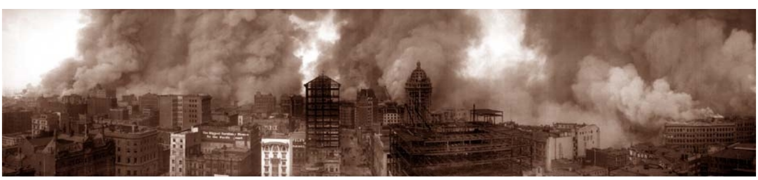


Figure 11.20 Fires in San Francisco following the 1906 earthquake. [from: http://upload.wikimedia.org/wikipedia/commons/3/3e/San_francisco_fire_1906.jpg]

Earthquakes are important triggers for failures on slopes that are already weak. An example is the Las Colinas slide in the city of Santa Tecla, El Salvador, which was triggered by a M7.6 offshore earthquake in January 2001 (Figure 11.21).

Figure 11.21 The Las Colinas debris flow at Santa Tecla (a suburb of the capital San Salvador) triggered by the January 2001 El Salvador earthquake. This is just one of many hundreds of slope failures that resulted from that earthquake. Over 500 people died in the area affected by this slide. [from: http://landslides.usgs.gov/ learning/ images/foreign/ElSalvadorslide.jpg]

Ground shaking during an earthquake can be enough to weaken rock and unconsolidated materials to the point of failure, but in many cases the shaking also contributes to a process known as **liquefaction**, in which an otherwise solid body of sediment is transformed into a liquid mass that can flow. When water-saturated sediments are shaken, the grains become rearranged to the point where they are no longer supporting one another. Instead, the water between the grains is holding them apart and the material can flow. Liquefaction can lead to the collapse of buildings and other structures that might be otherwise undamaged. A good example is the collapse of apartment buildings during the 1964 Niigata earthquake (M7.6) in Japan (Figure 11.22). Liquefaction can also contribute to slope failures and to fountains of sandy mud (sand volcanoes) in areas where there is loose saturated sand beneath a layer of more cohesive clay.

Figure 11.22 Collapsed apartment buildings in the Niigata area of Japan. The material beneath the buildings was liquefied to varying degrees by the 1964 earthquake. http://en.wikipedia.org/wiki/1964_ Niigata_earthquake#/media/File: Liquefaction_at_Niigata.JPG

Parts of the Fraser River delta are prone to liquefaction-related damage because the region is characterized by a 2 m to 3 m thick layer of fluvial silt and clay over top of at least 10 m of water-saturated fluvial sand (Figure 11.23). Under these conditions, it can be expected that seismic shaking will be amplified and, the sandy sediments will liquefy. This could lead to subsidence and tilting of buildings, and to failure and sliding of the silt and clay layer. Current building-code regulations in the Fraser delta area require that measures be taken to strengthen the ground underneath multi-storey buildings prior to construction.

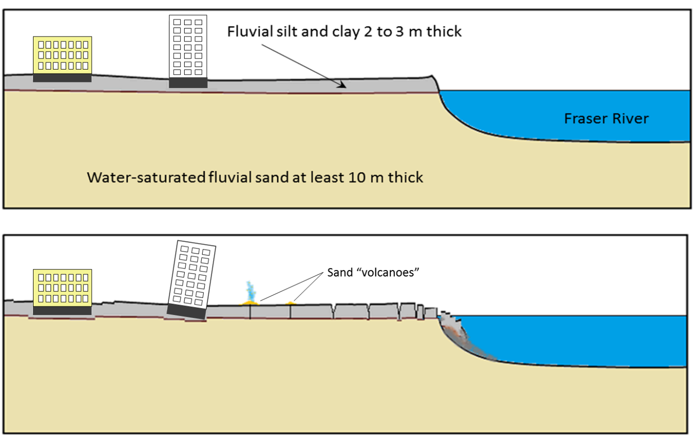


Figure 11.23 Recent unconsolidated sedimentary layers in the Fraser River delta area (top) and the potential consequences in the event of a damaging earthquake. [SE]

Exercises

Exercise 11.4 Creating Liquefaction and Discovering the Harmonic Frequency

There are a few ways that you can demonstrate the process of liquefaction for yourself. The simplest is to go to a sandy beach (lake, ocean, or river) and find a place near the water's edge where the sand is wet. This is best done with your shoes off, so let's hope it's not too cold! While standing in one place on a wet part of the beach, start moving your feet up and down at a frequency of about once per second. Within a few seconds the previously firm sand will start to lose strength, and you'll gradually sink in up to your ankles.

If you can't get to a beach, or if the weather isn't cooperating, put some sand (sandbox sand will do) into a small container, saturate it with water, and then pour the excess water off. You can shake it gently to get the water to separate and then pour the excess water away, and you may have to do that more than once. Place a small rock on the surface of the sand; it should sit there for hours without sinking in. Now, holding the container in one hand gently thump the side or the bottom with your other hand, about twice a second. The rock should gradually sink in as the sand around it becomes liquefied.

As you were moving your feet up and down or thumping the pot, it's likely that you soon discovered the most effective rate for getting the sand to liquefy; this would have been close to the natural harmonic frequency for that body of material. Stepping up and down as fast as you can (several times per second) on the wet beach would not have been effective, nor would you have achieved much by stepping once every several seconds. The body of sand vibrates most readily in response to shaking that is close to its natural harmonic frequency, and liquefaction is also most likely to occur at that frequency.

Earthquakes that take place beneath the ocean have the potential to generate **tsunami.[**footnote]Tsunami is the Japanese word for harbour wave. It is the same in both singular and plural.[/footnote] The most likely situation for a significant tsunami is a large (M7 or greater) subduction-related earthquake. As shown in Figure 11.24, during the time between earthquakes the overriding plate becomes distorted by elastic deformation; it is squeezed laterally (Figure 11.24B) and pushed up.

When an earthquake happens (Figure 11.24C), the plate rebounds and there is both uplift and subsidence on the sea floor, in some cases by as much as several metres vertically over an area of thousands of square kilometres. This vertical motion is transmitted through the water column where it generates a wave that then spreads across the ocean.

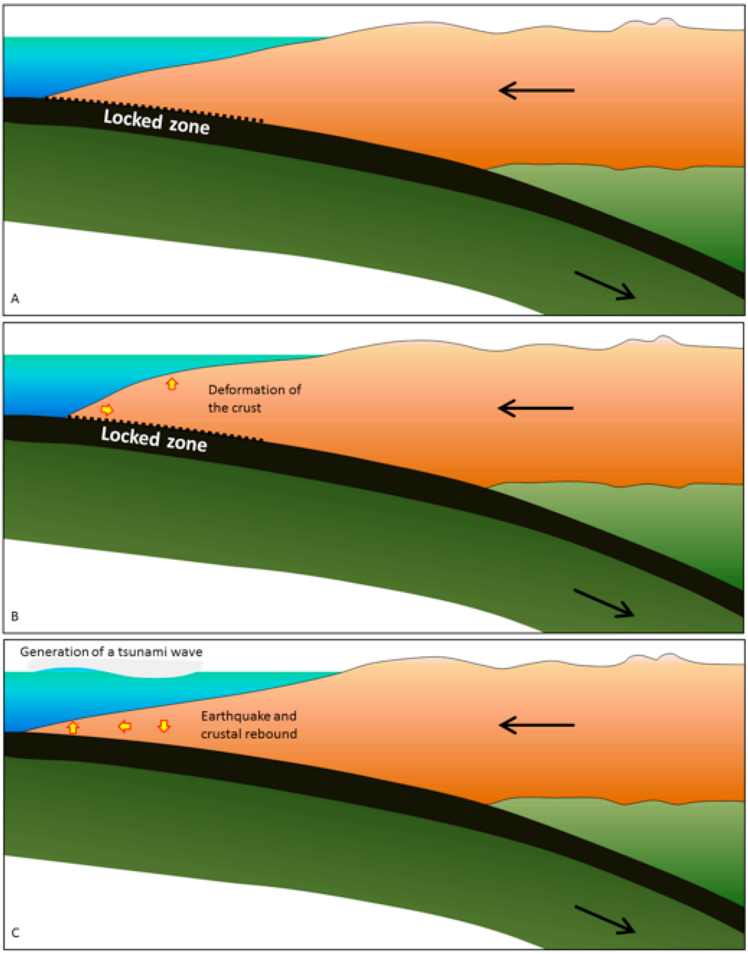


Figure 11.24 Elastic deformation and rebound of overriding plate at a subduction setting (B). The release of the locked zone during an earthquake (C) results in both uplift and subsidence on the sea floor, and this is transmitted to the water overhead, resulting in a tsunami. [SE]

Subduction earthquakes with magnitude less than 7 do not typically generate significant tsunami because the amount of vertical displacement of the sea floor is minimal. Sea-floor transform earthquakes, even large ones (M7 to M8), don't typically generate tsunami either, because the motion is mostly side to side, not vertical.

Tsunami waves travel at velocities of several hundred kilometres per hour and easily make it to the far side of an ocean in about the same time as a passenger jet. The simulated one shown in Figure 11.25 is similar to that created by the 1700 Cascadia earthquake off the coast of British Columbia, Washington, and Oregon, which was recorded in Japan nine hours later.

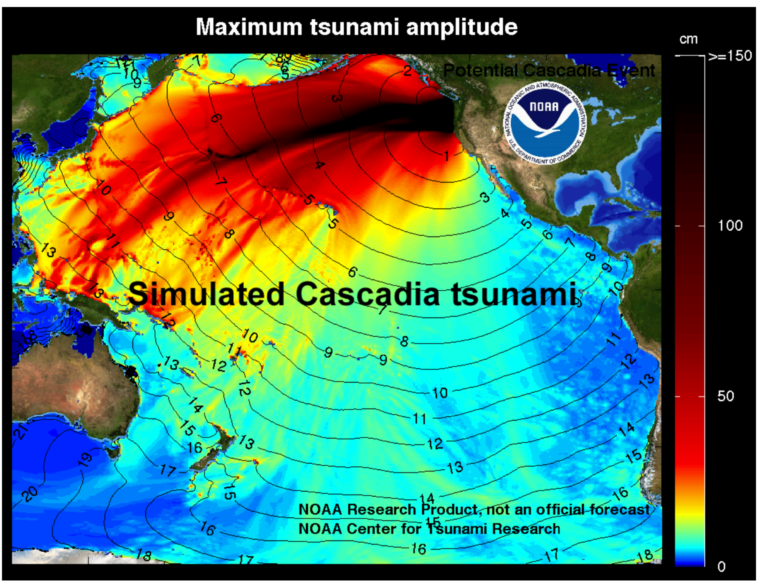


Figure 11.25 Model of the tsunami from the 1700 Cascadia earthquake (~M9) showing open-ocean wave heights (colours) and travel time contours. Tsunami wave amplitudes typically increase in shallow water. [from NOAA/PMEL/Center for Tsunami Research, at: http://nctr.pmel.noaa.gov/cascadia_simulated/]

Tsunami are discussed further in Chapter 17 under the topic of waves and coasts.

4.5 Forecasting Earthquakes and Minimizing Damage and Casualties

It has long been a dream of seismologists, geologists, and public safety officials to be able to accurately predict the location, magnitude, and timing of earthquakes on time scales that would be useful for minimizing danger to the public and damage to infrastructure (e.g., weeks, days, hours). Many different avenues of prediction have been explored, such as using observations of warning foreshocks, changes in magnetic fields, seismic tremor, changing groundwater levels, strange animal behaviour, observed periodicity, stress transfer considerations, and others. So far, none of the research into earthquake prediction has provided a reliable method. Although there are some reports of successful earthquake predictions, they are rare, and many are surrounded by doubtful circumstances.

The problem with earthquake predictions, as with any other type of prediction, is that they have to be accurate *most* of the time, not just *some* of the time. We have come to rely on weather predictions because they are generally (and increasingly) accurate. But if we try to predict earthquakes and are only accurate 10% of the time (and even that isn't possible with the current state of knowledge), the public will lose faith in the process very quickly, and then will ignore all of the predictions. Efforts are currently focused on forecasting earthquake probabilities, rather than predicting their occurrence.

There was great hope for earthquake predictions late in the 1980s when attention was focused on part of the San Andreas Fault at Parkfield, about 200 km south of San Francisco. Between 1881 and 1965 there were five earthquakes at Parkfield, most spaced at approximately 20-year intervals, all confined to the same 20 km-long segment of the fault, and all very close to M6 (Figure 11.26). Both the 1934 and 1966 earthquakes were preceded by small foreshocks exactly 17 minutes before the main quake.

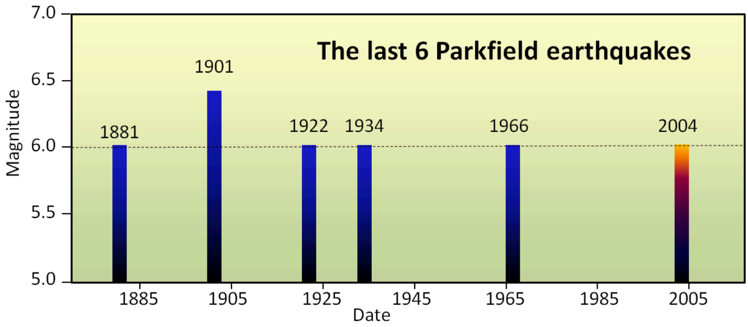


Figure 11.26 Earthquakes on the Parkfield segment of the San Andreas Fault between 1881 and 2004. [SE]

The U.S. Geological Survey recognized this as an excellent opportunity to understand earthquakes and earthquake prediction, so they armed the Parkfield area with a huge array of geophysical instruments and waited for the next quake, which was expected to happen around 1987. Nothing happened! The “1987 Parkfield earthquake” finally struck in September 2004. Fortunately all of the equipment was still there, but it was no help from the perspective of earthquake prediction. There were no significant precursors to the 2004 Parkfield earthquake in any of the parameters measured, including seismicity, harmonic tremor, strain (rock deformation), magnetic field, the conductivity of the rock, or creep, and there was no foreshock. In other words, even though every available technique was used to monitor it, the 2004 earthquake came as a complete surprise, with no warning whatsoever.

The hope for earthquake prediction is not dead, but it was hit hard by the Parkfield experiment. The current focus in earthquake-prone regions is to provide forecasts of the probability of an earthquake of a certain magnitude within a certain time period — typically a number of decades — while officials focus on ensuring that the population is educated about earthquake risks and that buildings and other infrastructure are as safe as can be. An example of this approach for the San Francisco Bay region of California is shown in Figure 11.27. Based on a wide range of information, including past earthquake history, accumulated stress from plate movement, and known stress transfer, seismologists and geologists have predicted the likelihood of a M6.7 or greater earthquake on each of eight major faults that cut through the region. The greatest probabilities are on the San Andreas, Rogers Creek, and Hayward Faults. As shown in Figure 11.27, there is a 63% chance that a major and damaging earthquake will take place somewhere in the region prior to 2036.

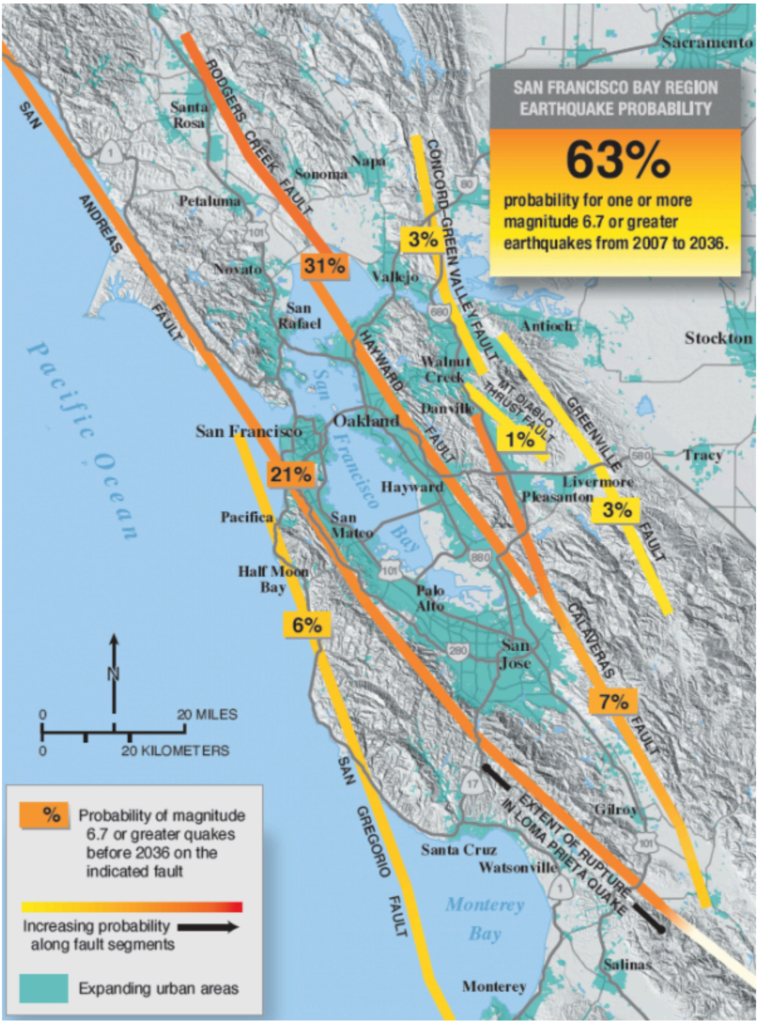


Figure 11.27 Probabilities of a M6.7 or larger earthquake over the period 2007 to 2036 on various faults in the San Francisco Bay region of California. [from USGS at: http://earthquake.usgs.gov/regional/nca/ucerf/]

As we've discussed already, it's not sufficient to have strong building codes, they have to be enforced. Building code compliance is quite robust in most developed countries, but is sadly inadequate in many developing countries.

It’s also not enough just to focus on new buildings; we have to make sure that existing buildings — especially schools and hospitals — and other structures such as bridges and dams, are as safe as they can be. An example of how this is applied to schools in B.C. is described in Box 11.2.

Making the seismic upgrade in B.C.’s schools

British Columbia is in the middle of a multi-billion-dollar program to make schools safer for students. The program is focused on older schools, because, according to the government, those built since 1992 already comply with modern seismic codes. Some schools would require too much work to make upgrading economically feasible and they are replaced. Where upgrading is feasible, the school is assessed carefully before any upgrade work is initiated.

An example is Sangster Elementary in Colwood on southern Vancouver Island. The school was originally built in 1957, with a major addition in 1973. Ironically, the newer part of the school, built of concrete blocks, required strengthening with the addition of a steel framework, while the 1957 part, which is a wood-frame building, did not require seismic upgrading. The work was completed in 2014.

[Sangster Elementary image from Google Maps – street view]

As of January 2015, upgrades had been completed at 145 B.C. schools, 11 were underway, and an additional 57 were ready to proceed with funding identified.* Another 129 schools were listed as needing upgrades. In May 2015, the provincial government announced that the target date for completion of the upgrades, originally set for 2020, had been delayed to 2030.* http://www2.gov.bc.ca/gov/topic.page?id=00C5FFBE51C44325A845819C007A01E7

Exercises

Exercise 11.5 Is Your Local School on the Seismic Upgrade List?

The B.C. Ministry of Education's list of schools in the seismic mitigation program as of January 2015 is available here: seismic-mitigation-progress-report.pdf. If you live in B.C., you can check to see if any of the schools in your area are on the list. If so, you might be able to find out, either from the school or on the Internet, what type of work has been done or is planned.

The seismic mitigation program has a strong focus on the Lower Mainland and Vancouver Island. Why do you think that is the case, and is it reasonable?

The final part of earthquake preparedness involves the formulation of public emergency plans, including escape routes, medical facilities, shelters, and food and water supplies. It also includes personal planning, such as emergency supplies (food, water, shelter, and warmth), escape routes from houses and offices, and communication strategies (with a focus on ones that don't involve the cellular network).

Chapter 5: Tsunami

Read an excerpt from **chapter 8** of Abbott's "Natural Disasters": Tsunami versus Wind-Caused Waves". You will have to log in with your SFU CANVAS credentials to view the link.

Abbott, Patrick L. *Natural Disasters*. Tenth edition. New York, NY: McGraw-Hill, 2017.

Tsunami Chapter (higher quality if you download and open as a .pdf rather than viewing in your browser window).

Then read **chapter 2** of the book: Playing against nature: Integrating science and economics to mitigate natural hazards in an uncertain world. You may have to log-in to SFU's library system using your SFU ID.

Stein, Seth, and Jerome L. Stein. *Playing against Nature: Integrating Science and Economics to Mitigate Natural Hazards in an Uncertain World*. Chichester: Wiley, 2014.

5.1 - Case Study: 1964 Port Alberni Tsunami

On the afternoon of Good Friday, March 27, 1964, the strongest earthquake recorded in North America, and the second strongest ever recorded, occurred in Alaska. The Great Alaskan Earthquake was a 9.2 magnitude subduction zone (megathrust) earthquake located at a depth of approximately 25 kilometres. It lasted four minutes and 38 seconds. 131 people died in the earthquake and ensuing tsunami's (USGS, 2012).

The epicentre of the earthquake was 125 kilometres east of Anchorage, Alaska, where many inadequately engineered houses, buildings and infrastructure were damaged or destroyed. Three hundred kilometers southwest, some areas near Kodiak were permanently raised by 9.1 metres. Southeast of Anchorage, areas around the head of Turnagain Arm near Girdwood and Portage dropped as much as 2.4 metres. A massive underwater slide at Port Valdez in Prince William Sound created an 8.2 metre tsunami that destroyed the village of Chenega, killing 23 of the 68 people who lived there. Post-quake tsunamis severely affected Whittier, Seward, Kodiak and other Alaskan communities, as well as people and property in Oregon, California and British Columbia.

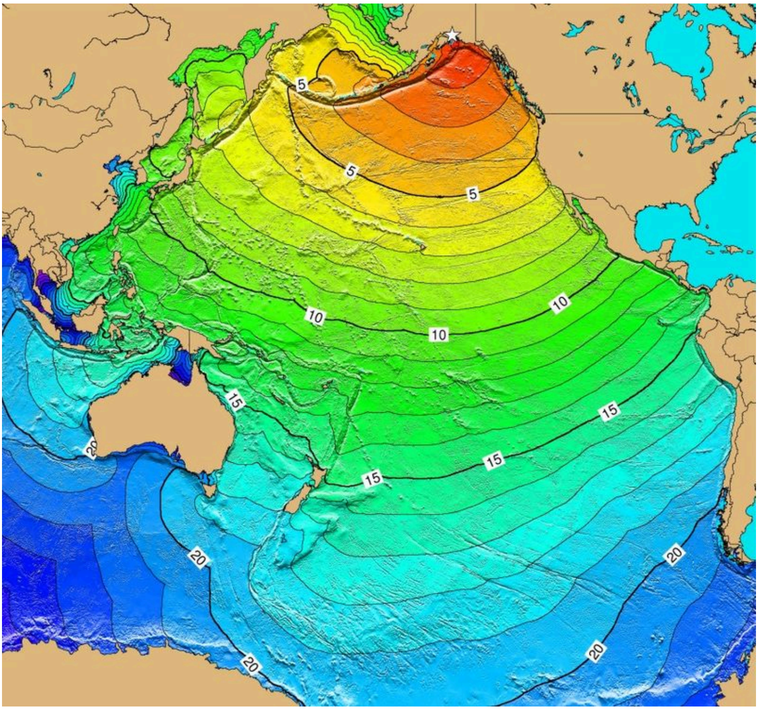


Figure 8.11 Calculated travel time map for the tectonic tsunami produced by the 1964 earthquake in Alaska. Map does not show the height or strength of the waves, only the calculated travel times. Number represents time in hours for the wave to reach the destination.

Two types of tsunami were produced as a result of the earthquake: a **tectonic tsunami** caused by the movement of the tectonic plates, and **subaerial landslide (or submarine) tsunami** caused by underwater landslides. About 20 of these smaller tsunamis were responsible for the majority of the tsunami damage in over 20 countries, including Canada, Peru, New Zealand, Papua New Guinea and Japan. The largest recorded wave was at Shoup Bay, Alaska, at height of about 67 metres.

Three hours after the quake, a 1.4 metre tsunami wave reached Prince Rupert, BC, and did little damage. Then, around midnight, the first of two waves began travelling up the 40 kilometre Alberni Inlet toward the towns of Alberni and Port Alberni. The narrow inlet amplified the size and intensity of the wave, and when it struck the two towns it had a height of 2.44 metres. One hour later, a second, larger wave of 3.05 metres hit. It was the second wave that caused most of the damage, lifting houses off their foundations and sweeping log booms on shore. The second wave was followed by four more waves ranging in height from 1.52 metres and 1.83 metres and ocurring at roughly 90 minute intervals.

In total, the tsunami washed away away 55 homes and damaged 375 others. It caused $5 million in damage in Port Alberni and Alberni. The mill at Port Alberni, which employed 4,000 people, was temporarily closed. Incredibly, there were no fatalities in either town. The tsunami lead directly to the amalgamation of Alberni and Port Alberni into a single town in 1966.

Animation of Tsunami Generated by the Earthquake

If you are reading this in print, you can see the animation at https://www.youtube.com/watch?v=rJOGJApz1M0.

Attributions

Figure 8.11 Calculated Travel Time Map for 1964 Alaska Tsunami (http://en.wikipedia.org/wiki/File:Calculated_Travel_Time_Map_for_1964_Alaska_Tsunami.jpg) by Ngdchazards used under CC BY SA 3.0 (http://creativecommons.org/licenses/by-sa/3.0/deed.en)

Chapter 6: Volcanism

Introduction

Learning Objectives

After carefully reading this chapter, completing the exercises within it, and answering the questions at the end, you should be able to:

- Explain the relationships between plate tectonics, the formation of magma, and volcanism
- Describe the range of magma compositions formed in differing tectonic environments, and discuss the relationship between magma composition (and gas content) and eruption style
- Explain the geological and eruption-style differences between different types of volcanoes, especially shield volcanoes, composite volcanoes, and cinder cones
- Understand the types of hazards posed to people and to infrastructure by the different types of volcanic eruptions
- Describe the behaviours that we can expect to observe when a volcano is ready to erupt, and the techniques that we can use to monitor those behaviours and predict eruptions

- Summarize the types of volcanoes that have erupted in British Columbia over the past 2.6 Ma, and the characteristics of some of those eruptions

A volcano is any location where magma comes to the surface, or has done so within the past several million years. This can include eruptions on the ocean floor (or even under the water of lake), where they are called **subaqueous eruptions**, or on land, where they are called **subaerial eruptions**. Not all volcanic eruptions produce the volcanic mountains with which we are familiar; in fact most of Earth's volcanism takes place along the spreading ridges on the sea floor and does not produce volcanic mountains at all — not even sea-floor mountains.

Canada has a great deal of volcanic rock, but most of it is old, some of it billions of years old. Only in B.C. and the Yukon are there volcanoes that have been active within the past 2.6 Ma (Pleistocene or younger), and the vast majority of these are in B.C. We'll look at those in some detail toward the end of this chapter, but a few of them are shown on Figures 4.1 and 4.2.

The study of volcanoes is critical to our understanding of the geological evolution of Earth, and to our understanding of significant changes in climate. But, most important of all, understanding volcanic eruptions allows us to save lives and property. Over the past few decades, volcanologists have made great strides in their ability to forecast volcanic eruptions and predict the consequences — this has already saved thousands of lives.

Figure 4.1 Mt. Garibaldi, near Squamish B.C., is one of Canada's tallest (2,678 m) and most recently active volcanoes. It last erupted approximately 10,000 years ago. [SE photo]

Figure 4.2 Mt. Garibaldi (background left, looking from the north) with Garibaldi Lake in the foreground. The volcanic peak in the centre is Mt. Price and the dark flat–topped peak is The Table. All three of these volcanoes were active during the last glaciation. [SE photo]

6.1 Plate Tectonics and Volcanism

The relationships between plate tectonics and volcanism are shown on Figure 4.3. As summarized in Chapter 3, magma is formed at three main plate-tectonic settings: divergent boundaries (decompression melting), convergent boundaries (flux melting), and mantle plumes (decompression melting).

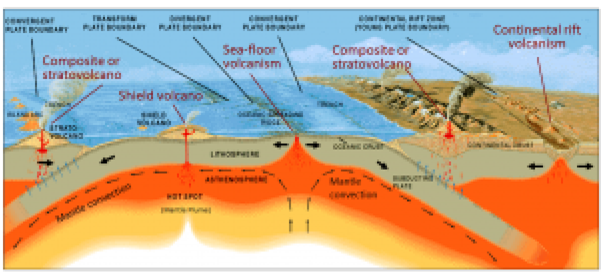


Figure 4.3 The plate-tectonic settings of common types of volcanism. Composite volcanoes form at subduction zones, either on ocean-ocean convergent boundaries (left) or ocean-continent convergent boundaries (right). Both shield volcanoes and cinder cones form in areas of continental rifting. Shield volcanoes form above mantle plumes, but can also form at other tectonic settings. Sea-floor volcanism can take place at divergent boundaries, mantle plumes and ocean-ocean-convergent boundaries. [SE, after USGS (http://pubs.usgs.gov/gip/dynamic/Vigil.html)]

The mantle and crustal processes that take place in areas of volcanism are illustrated in Figure 4.4. At a spreading ridge, hot mantle rock moves slowly upward by convection (cm/year), and within about 60 km of the surface, partial melting starts because of decompression. Over the triangular area shown in Figure 4.4a, about 10% of the ultramafic mantle rock melts, producing mafic magma that moves upward toward the axis of spreading (where the two plates are moving away from each other). The magma fills vertical fractures produced by the spreading and spills out onto the sea floor to form basaltic **pillows** (more on that later) and lava flows. There is spreading-ridge volcanism taking place about 200 km offshore from the west coast of Vancouver Island.

Exercises

Exercise 4.1 How Thick Is the Oceanic Crust?

Figure 4.4a shows a triangular zone about 60 km thick; within this zone, approximately 10% of the mantle rock melts to form oceanic crust. Based on this information, approximately how thick do you think the resulting oceanic crust should be?

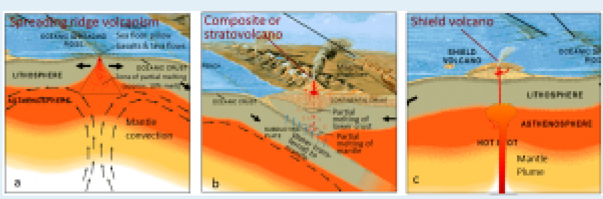


Figure 4.4 The processes that lead to volcanism in the three main volcanic settings on Earth: (a) volcanism related to plate divergence, (b) volcanism at an ocean-continent boundary, and (c) volcanism related to a mantle plume. [SE, after USGS (http://pubs.usgs.gov/gip/dynamic/Vigil.html)]*

**Similar processes take place at an ocean-ocean convergent boundary.*

At an ocean-continent or ocean-ocean[1] convergent boundary, oceanic crust is pushed far down into the mantle (Figure 4.4b). It is heated up, and while there isn't enough heat to melt the subducting crust, there is enough to force the water out of some of its minerals. This water rises into the overlying mantle where it contributes to flux melting of the mantle rock. The mafic magma produced rises through the mantle to the base of the crust. There it contributes to partial melting of crustal rock, and thus it assimilates much more felsic material. That magma, now intermediate in composition, continues to rise and assimilate crustal material; in the upper part of the crust, it accumulates into plutons. From time to time, the magma from the plutons rises toward surface, leading to volcanic eruptions. Mt. Garibaldi (Figures 4.1 and 4.2) is an example of subduction-related volcanism.

A mantle plume is an ascending column of hot rock (not magma) that originates deep in the mantle, possibly just above the core-mantle boundary. Mantle plumes are thought to rise at approximately 10 times the rate of mantle convection. The ascending column may be on the order of kilometres to tens of kilometres across, but near the surface it spreads out to create a mushroom-style head that is several tens to over 100 kilometres across. Near the base of the lithosphere (the rigid part of the mantle), the mantle plume (and possibly some of the surrounding mantle material) partially melts to form mafic magma that rises to feed volcanoes. Since most mantle plumes are beneath the oceans, the early stages of volcanism typically

1. At an ocean-continent convergent boundary, part of a plate that is made up of oceanic crust is subducting beneath part of another plate made up of continental crust. At an ocean-ocean convergent boundary, oceanic crust is being subducted beneath another oceanic-crust plate.

take place on the sea floor. Over time, islands may form like those in Hawaii.

Volcanism in northwestern B.C. (Figures 4.5 and 4.6) is related to continental rifting. This area is not at a divergent or convergent boundary, and there is no evidence of an underlying mantle plume. The crust of northwestern B.C. is being stressed by the northward movement of the Pacific Plate against the North America Plate, and the resulting crustal fracturing provides a conduit for the flow of magma from the mantle. This may be an early stage of continental rifting, such as that found in eastern Africa.

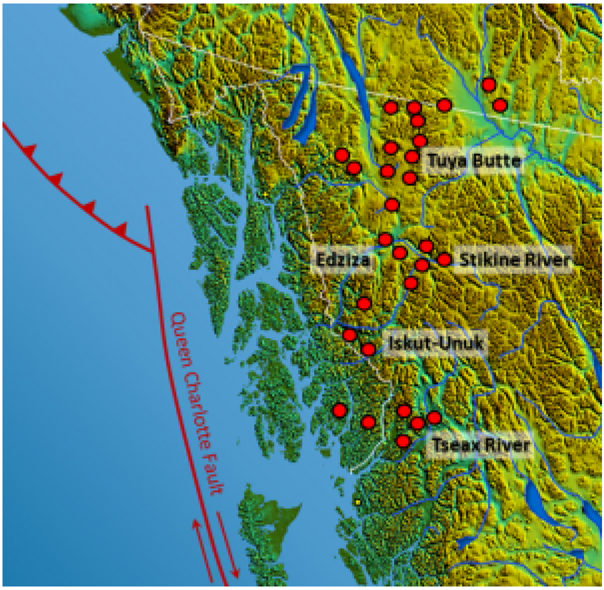


Figure 4.5 Volcanoes and volcanic fields in the Northern Cordillera Volcanic Province, B.C. (base map from Wikipedia (http://commons.wikimedia.org/wiki/File:South-West_Canada.jpg). Volcanic locations from Edwards, B. & Russell, J. (2000). Distribution, nature, and origin of Neogene-Quaternary magmatism in the northern Cordilleran volcanic province, Canada. Geological Society of America Bulletin. pp. 1280-1293[SE]Cordillera Volcanic Province, B.C.

Figure 4.6 Volcanic rock at the Tseax River area, northwestern B.C. [SE]

6.2 Magma Composition and Eruption Style

As noted in the previous section, the types of magma produced in the various volcanic settings can differ significantly. At divergent boundaries and oceanic mantle plumes, where there is little interaction with crustal materials and magma fractionation to create felsic melts does not take place, the magma tends to be consistently mafic. At subduction zones, where the magma ascends through significant thicknesses of crust, interaction between the magma and the crustal rock — some of which is quite felsic — leads to increases in the felsic character of the magma.

As shown in Figure 4.7, several processes can make magma that is stored in a chamber within the crust more felsic, and can also contribute to development of vertical zonation from more mafic at the bottom to more felsic at the top. Partial melting of country rock and country-rock xenoliths increases the overall felsic character of the magma; first, because the country rocks tends to be more felsic than the magma, and second, because the more felsic components of the country rock melt preferentially. Settling of ferromagnesian crystals from the upper part of the magma, and possible remelting of those crystals in the lower part can both contribute to the vertical zonation from relatively mafic at the bottom to more felsic at the top.

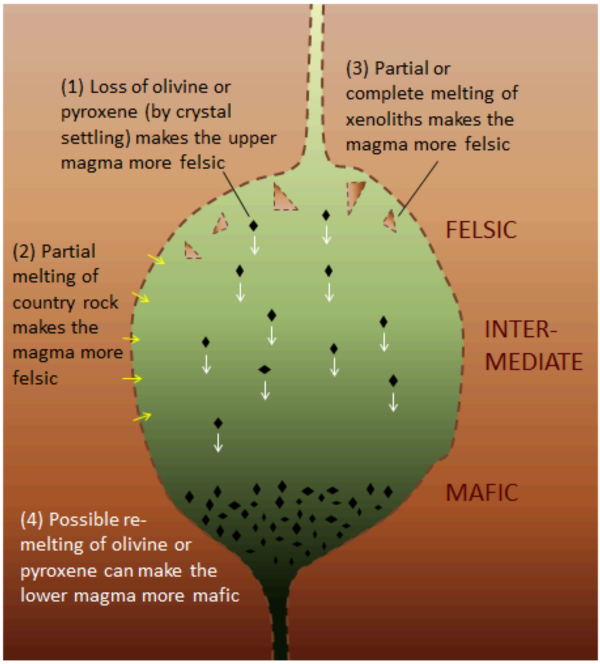


Figure 4.7 The important processes that lead to changes in the composition of magmas stored within magma chambers within relatively felsic rocks of the crust. [SE]

From the perspective of volcanism there are some important differences between felsic and mafic magmas. First, as we've already discussed, felsic magmas tend to be more viscous because they have more silica, and hence more polymerization. Second, felsic magmas tend to have higher levels of volatiles; that is, components that behave as gases during volcanic eruptions. The most abundant volatile in magma is water (H_2O), followed typically by carbon dioxide (CO_2), and then by sulphur dioxide (SO_2). The general relationship between the SiO_2 content of magma and the amount of volatiles is shown in Figure 4.8. Although there are many exceptions to this trend, mafic magmas typically have 1% to 3% volatiles, intermediate magmas have 3% to 4% volatiles, and felsic magmas have 4% to 7% volatiles.

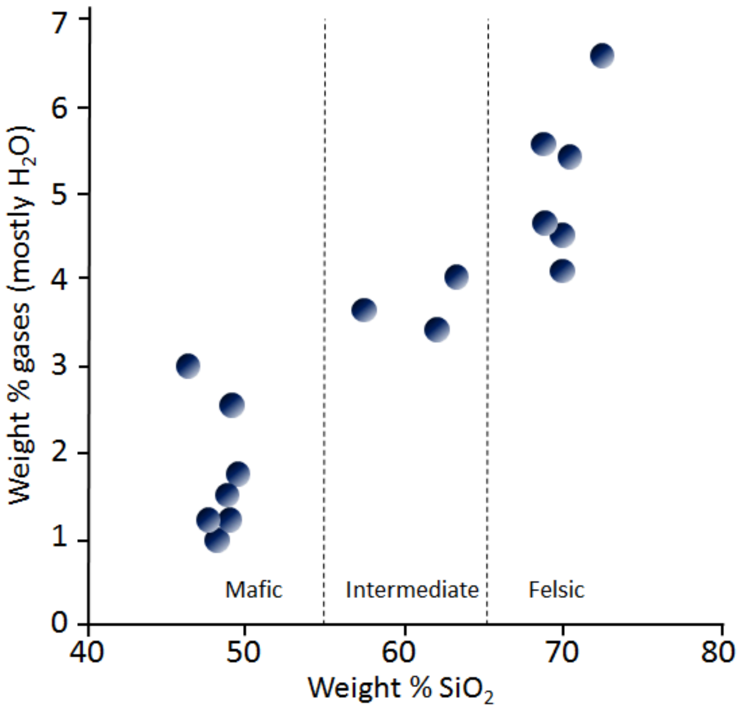


Figure 4.8 Variations in the volatile compositions of magmas as a function of silica content [SE after Schminke, 2004, (Schminke, H-U., 2004, Volcanism, Springer-Verlag, Heidelberg)]

Differences in viscosity and volatile level have significant implications for the nature of volcanic eruptions. When magma is deep beneath the surface and under high pressure from the surrounding rocks, the gases remain dissolved. As magma approaches the surface, the pressure exerted on it decreases. Gas bubbles start to form, and the more gas there is in the magma, the more bubbles form. If the gas content is low or the magma is runny enough for gases to rise up through it and escape to surface, the pressure will not become excessive. Assuming that it can break through to the surface, the magma will flow out relatively gently. An eruption that involves a steady non-violent flow of magma is called **effusive**.

Exercises

Exercise 4.2 Under Pressure!

A good analogy for a magma chamber in the upper crust is a plastic bottle of pop on the supermarket shelf. Go to a supermarket and pick one up off the shelf (something not too dark). You'll find that the bottle is hard because it was bottled under pressure, and you should be able to see that there are no gas bubbles inside.

Buy a small bottle of pop (you don't have to drink it!) and open it. The bottle will become soft because the pressure is released, and small bubbles will start forming. If you put the lid back on and shake the bottle (best to do this outside!), you'll enhance the processes of bubble formation, and when you open the lid, the pop will come gushing out, just like an explosive volcanic eruption.

A pop bottle is a better analogue for a volcano than the old baking soda and vinegar experiment that you did in elementary school, because pop bottles, like volcanoes, come pre-charged with gas pressure. All we need to do is release the confining pressure and the gases come bubbling out.

[Wikipedia image: http://upload.wikimedia.org /wikipedia/commons/6/64/ Champagne_uncorking_photographed_with_a_high_speed _air-gap_flash.jpg]

If the magma is felsic, and therefore too viscous for gases to escape easily, or if it has a particularly high gas content, it is likely to be under high pressure. Viscous magma doesn't flow easily, so even if there is a way for it to move out, it may not flow out. Under these circumstances pressure will continue to build as more magma moves up from beneath and gases continue to exsolve. Eventually some part of the volcano will break and then all of that pent-up pressure will lead to an explosive eruption.

Mantle plume and spreading-ridge magmas tend to be consistently mafic, so effusive eruptions are the norm. At subduction zones, the average magma composition is likely to be close to intermediate, but as we've seen, magma chambers can become zoned and so compositions ranging from felsic to mafic are possible. Eruption styles can be correspondingly variable.

6.3 Types of Volcanoes

There are numerous types of volcanoes or volcanic sources; some of the more common ones are summarized in Table 4.1.

Type	Tectonic Setting	Size and Shape	Magma and Eruption Characteristics	Example
Cinder cone	Various; some form on the flanks of larger volcanoes	Small (10s to 100s of m) and steep (>20°)	Most are mafic and form from the gas-rich early stages of a shield- or rift-associated eruption	Eve Cone, northern B.C.
Composite volcano	Almost all are at subduction zones	Medium size (1000s of m) and moderate steepness (10° to 30°)	Magma composition varies from felsic to mafic, and from explosive to effusive	Mt. St. Helens
Shield volcano	Most are at mantle plumes; some are on spreading ridges	Large (up to several 1,000 m high and 200 km across), not steep (typically 2° to 10°)	Magma is almost always mafic, and eruptions are typically effusive, although cinder cones are common on the flanks of shield volcanoes	Kilauea, Hawaii
Large igneous provinces	Associated with “super” mantle plumes	Enormous (up to millions of km^2) and 100s of m thick	Magma is always mafic and individual flows can be 10s of m thick	Columbia River basalts

Sea-floor volcanism	Generally associated with spreading ridges but also with mantle plumes	Large areas of the sea floor associated with spreading ridges	At typical eruption rates, pillows form; at faster rates, lava flows develop	Juan de Fuca ridge
Kimberlite	Upper-mantle sourced	The remnants are typically 10s to 100s of m across	Most appear to have had explosive eruptions forming cinder cones; the youngest one is over 10 ka old, and all others are over 30 Ma old.	Lac de Gras Kimberlite Field, N.W.T.

Table 4.1 A summary of the important types of volcanism

The sizes and shapes of typical shield, composite, and cinder-cone volcanoes are compared in Figure 4.9, although, to be fair, Mauna Loa is the largest shield volcano on Earth; all others are smaller. Mauna Loa rises from the surrounding flat sea floor, and its diameter is in the order of 200 km. Its elevation is 4,169 m above sea level. Mt. St. Helens, a composite volcano, rises above the surrounding hills of the Cascade Range. Its diameter is about 6 km, and its height is 2,550 m above sea level. Cinder cones are much smaller. On this drawing, even a large cinder cone is just a dot.

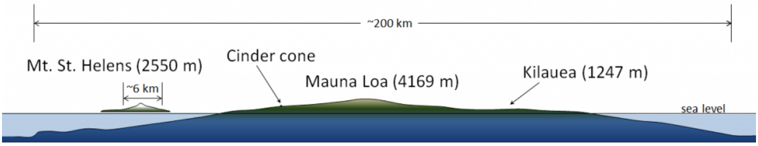


Figure 4.9 Profiles of Mauna Loa shield volcano, Mt. St. Helens composite volcano, and a large cinder cone [SE]

Cinder Cones

Cinder cones, like Eve Cone in northern B.C. (Figure 4.10), are typically only a few hundred metres in diameter, and few are more than 200 m high. Most are made up of fragments of **vesicular** mafic rock (scoria) that were expelled as the magma boiled when it approached the surface, creating fire fountains. In many cases, these later became effusive (lava flows) when the gases were depleted. Most cinder cones are **monogenetic**, meaning that they formed during a single eruptive phase that might have lasted weeks or months. Because cinder cones are made up almost exclusively of loose fragments, they have very little strength. They can be easily, and relatively quickly, eroded away.

Figure 4.10 Eve Cone, situated near to Mt. Edziza in northern B.C., formed approximately 700 years ago [Wikipedia, http://en.wikipedia.org/wiki/Eve_Cone# mediaviewer/ File:Symmetrical_Eve_Cone.jpg]

Composite Volcanoes

Composite volcanoes, like Mt. St. Helens in Washington State (Figure 4.11), are almost all associated with subduction at convergent plate boundaries — either ocean-continent or ocean-ocean boundaries (Figure 4.4b). They can extend up to several thousand metres from the surrounding terrain, and, with slopes ranging up to 30˚, are typically up to 10 km across. At many such volcanoes, magma is stored in a magma chamber in the upper part of the crust. For example, at Mt. St. Helens, there is evidence of a magma chamber that is approximately 1 km wide and extends from about 6 km to 14 km below the surface (Figure 4.12). Systematic variations in the composition of volcanism over the past several thousand years at Mt. St. Helens imply that the magma chamber is zoned, from more felsic at the top to more mafic at the bottom.

Figure 4.11 The north side of Mt. St. Helens in southwestern Washington State, 2003 [SE photo]. The large 1980 eruption reduced the height of the volcano by 400 m, and a sector collapse removed a large part of the northern flank. Between 1980 and 1986 the slow eruption of more mafic and less viscous lava led to construction of a dome inside the crater.

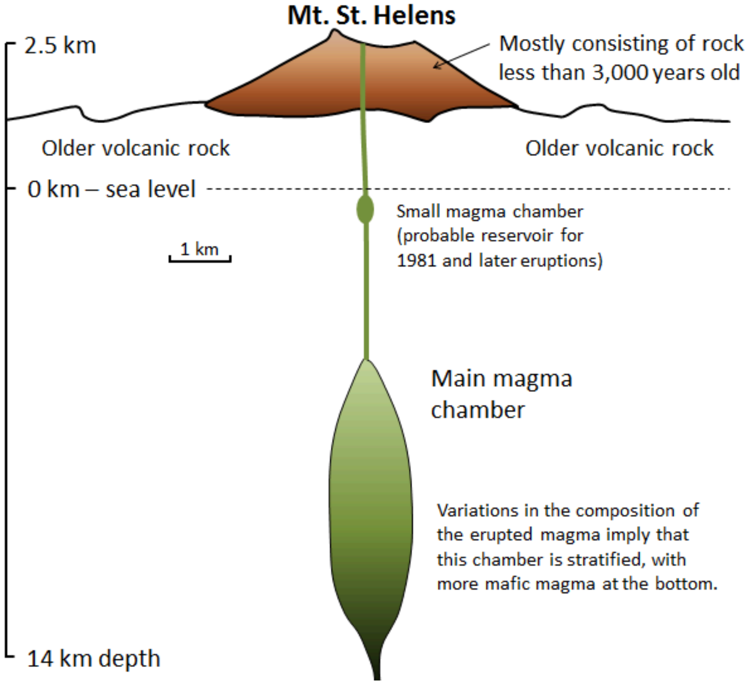


Figure 4.12 A cross-section through the upper part of the crust at Mt. St. Helens showing the zoned magma chamber. [SE, after Pringle, 1993]

Mafic eruptions (and some intermediate eruptions), on the other hand, produce lava flows; the one shown in Figure 4.13b is thick enough (about 10 m in total) to have cooled in a **columnar jointing** pattern (Figure 4.14). Lava flows both flatten the profile of the volcano (because the lava typically flows farther than pyroclastic debris falls) and protect the fragmental deposits from erosion. Even so, composite volcanoes tend to erode quickly. Patrick Pringle, a volcanologist with the Washington State Department of Natural Resources, describes Mt. St. Helens as a "pile of junk." The rock that makes up Mt. St. Helens ranges in composition from rhyolite (Figure 4.13a) to basalt (Figure

4.13b); this implies that the types of past eruptions have varied widely in character. As already noted, felsic magma doesn’t flow easily and doesn’t allow gases to escape easily. Under these circumstances, pressure builds up until a conduit opens, and then an explosive eruption results from the gas-rich upper part of the magma chamber, producing **pyroclastic** debris, as shown on Figure 4.13a. This type of eruption can also lead to rapid melting of ice and snow on a volcano, which typically triggers large mudflows known as **lahars** (Figure 4.13a). Hot, fast-moving pyroclastic flows and lahars are the two main causes of casualties in volcanic eruptions. Pyroclastic flows killed approximately 30,000 people during the 1902 eruption of Mt. Pelée on the Caribbean island of Martinique. Most were incinerated in their homes. In 1985 a massive lahar, triggered by the eruption of Nevado del Ruiz, killed 23,000 people in the Colombian town of Armero, about 50 km from the volcano.

In a geological context, composite volcanoes tend to form relatively quickly and do not last very long. Mt. St. Helens, for example, is made up of rock that is all younger than 40,000 years; most of it is younger than 3,000 years. If its volcanic activity ceases, it might erode away within a few tens of thousands of years. This is largely because of the presence of pyroclastic eruptive material, which is not strong.

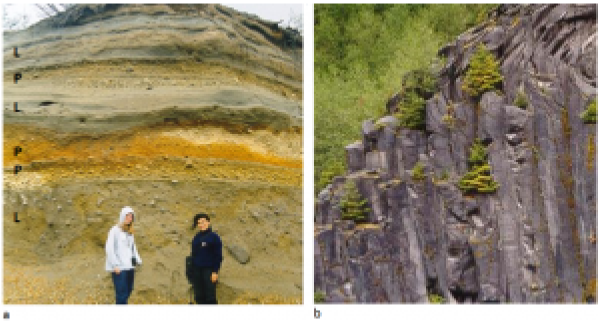
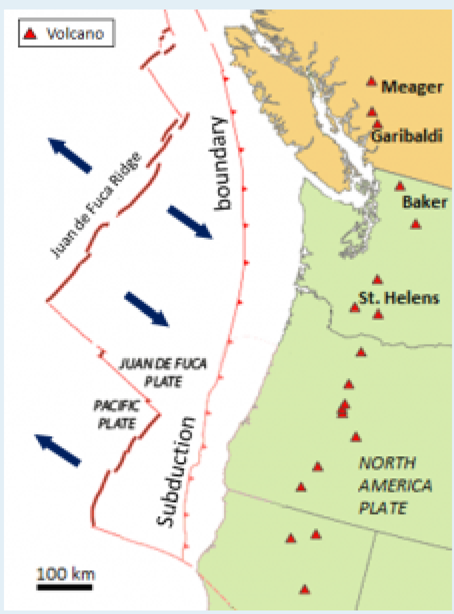


Figure 4.13 Mt. St. Helens volcanic deposits: (a) lahar deposits (L) and felsic pyroclastic deposits (P) and (b) a columnar basalt lava flow. The two photos were taken at locations only about 500 m apart. [SE]

Exercises

Exercise 4.3 Volcanoes and Subduction



The map shown here illustrates the interactions between the North America, Juan de Fuca, and Pacific Plates off the west coast of Canada and the United States. The Juan de Fuca Plate is forming along the Juan de Fuca ridge, and is then subducted beneath the North America Plate along the red line with teeth on it ("Subduction boundary").

1. Using the scale bar in the lower left of the map, estimate the average distance between the subduction boundary and the Cascadia composite volcanoes.

2. If the subducting Juan de Fuca Plate descends 40 km for every 100 km that it moves inland, what is its likely depth in the area where volcanoes are forming?

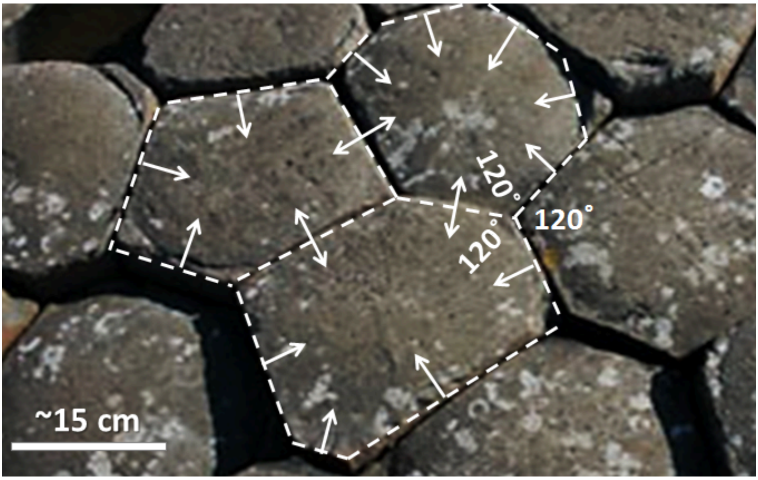


Figure 4.14 Figure 4.14 The development of columnar jointing in basalt, here seen from the top looking down. As the rock cools it shrinks, and because it is very homogenous it shrinks in a systematic way. When the rock breaks it does so with approximately 120˚ angles between the fracture planes. The resulting columns tend to be 6-sided but 5- and 7-sided columns also form. [SE]

Shield Volcanoes

Most shield volcanoes are associated with mantle plumes, although some form at divergent boundaries, either on land or on the sea floor. Because of their non-viscous mafic magma they tend to have relatively gentle slopes (2 to 10°) and the larger ones can be over 100 km in diameter. The best-known shield volcanoes are those that make up the Hawaiian Islands, and of these, the only active ones are on the big island of Hawaii. Mauna Loa, the world's largest volcano and the world's largest mountain (by volume) last erupted in 1984. Kilauea, arguably the world's most active volcano, has been erupting, virtually without interruption, since 1983. Loihi is an underwater volcano on the southeastern side of Hawaii. It is last known to have erupted in 1996, but may have erupted since then without being detected.

All of the Hawaiian volcanoes are related to the mantle plume that currently lies beneath Mauna Loa, Kilauea, and Loihi (Figure 4.15). In this area, the Pacific Plate is moving northwest at a rate of about 7 cm/year. This means that the earlier formed — and now extinct — volcanoes have now moved well away from the mantle plume. As shown on Figure 4.15, there is evidence of crustal magma chambers beneath all three active Hawaiian volcanoes. At Kilauea, the magma chamber appears to be several kilometres in diameter, and is situated between 8 km and 11 km below surface.[1]

1. Lin, G, Amelung, F, Lavallee, Y, and Okubo, P, 2014, Seismic evidence for a crustal magma reservoir beneath the upper east rift zone of Kilauea volcano, Hawaii. Geology. V.

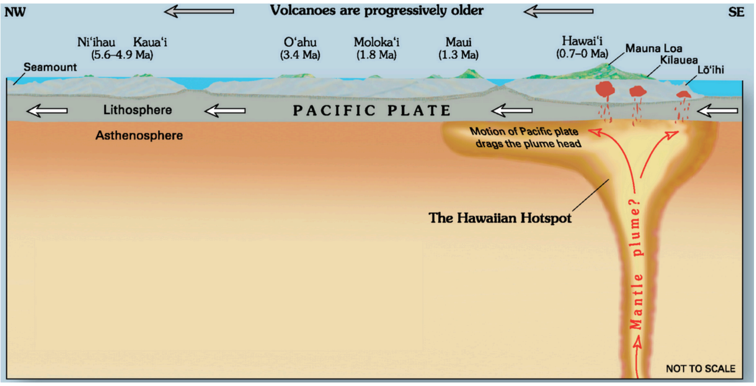


Figure 4.15 Mauna Kea from near to the summit of Mauna Loa, Hawaii [http://upload.wikimedia.org/wikipedia/commons/f/f1/Hawaii_hotspot_cross-sectional_diagram.jpg]

Although it is not a prominent mountain (Figure 4.9), Kilauea volcano has a large **caldera** in its summit area (Figure 4.16). A caldera is a volcanic **crater** that is more than 2 km in diameter; this one is 4 km long and 3 km wide. It contains a smaller feature called Halema'uma'u crater, which has a total depth of over 200 m below the surrounding area. Most volcanic craters and calderas are formed above magma chambers, and the level of the crater floor is influenced by the amount of pressure exerted by the magma body. During historical times, the floors of both Kilauea caldera and Halema'uma'u crater have moved up during expansion of the magma chamber and down during deflation of the chamber.

Figure 4.16 Aerial view of the Kilauea caldera. The caldera is about 4 km across, and up to 120 m deep. It encloses a smaller and deeper crater known as Halema'uma'u. [http://upload.wikimedia.org/wikipedia/commons/b/b4/Kilauea_ali_2012_01_28.jpg]

One of the conspicuous features of Kilauea caldera is rising water vapour (the white cloud in Figure 4.16) and a strong smell of sulphur (Figure 4.17). As is typical in magmatic regions, water is the main volatile component, followed by carbon dioxide and sulphur dioxide. These, and some minor gases, originate from the magma chamber at depth and rise up through cracks in the overlying rock. This degassing of the magma is critical to the style of eruption at Kilauea, which, for most of the past 30 years, has been effusive, not explosive.

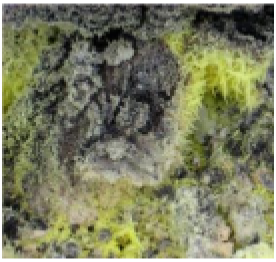



Figure 4.17 A gas-composition monitoring station (left) within the Kilauea caldera and at the edge of Halema'uma'u crater. The rising clouds are mostly composed of water vapour, but also include carbon dioxide and sulphur dioxide. Sulphur crystals (right) have formed around a gas vent in the caldera. [SE photos]

The Kilauea eruption that began in 1983 started with the formation of a cinder cone at Pu'u 'O'o, approximately 15 km east of the caldera (Figure 4.18). The magma feeding this eruption flowed along a major conduit system known as the East Rift, which extends for about 20 km from the caldera, first southeast and then east. Lava fountaining and construction of the Pu'u 'O'o cinder cone (Figure 4.19a) continued until 1986 at which time the flow became effusive. From 1986 to 2014, lava flowed from a gap in the southern flank of Pu'u 'O'o down the slope of Kilauea through a **lava tube** (Figure 4.19d), emerging at or near the ocean. Since June 2014, the lava has flowed northeast (see Exercise 4.4).

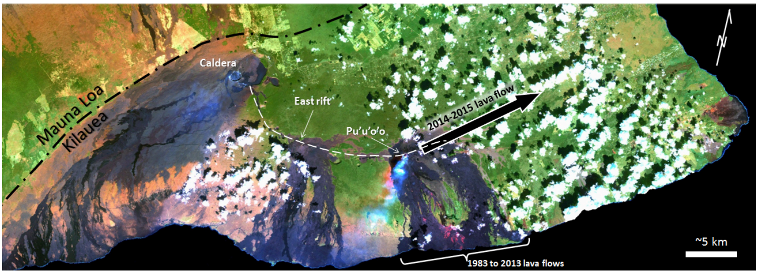


Figure 4.18 Satellite image of Kilauea volcano showing the East rift and Pu'u 'O'o, the site of the eruption that started in 1983. The puffy white blobs are clouds. [SE after, http://en.wikipedia.org/wiki/Hawaii_(island)#mediaviewer/File:Island_of_Hawai%27i_-_Landsat_mosaic.jpg]

The two main types of textures created during effusive subaerial eruptions are pahoehoe and aa. **Pahoehoe**, ropy lava that forms as non-viscous lava, flows gently, forming a skin that gels and then wrinkles because of ongoing flow of the lava below the surface (Figure 4.19b, and "lava flow video"). **Aa**, or blocky lava, forms when magma is forced to flow faster than it is able to (down a slope for example) (Figure 4.19c). **Tephra** (lava fragments) is produced during explosive eruptions, and accumulates in the vicinity of cinder cones.

Figure 4.19d is a view into an active lava tube on the southern edge of Kilauea. The red glow is from a stream of very hot lava (~1200°C) that has flowed underground for most of the 8 km from the Pu'u 'O'o vent. Lava tubes form naturally and readily on both shield and composite volcanoes because flowing mafic lava preferentially cools near its margins, forming solid **lava levées** that eventually close over the top of the flow. The magma within a lava tube is not exposed to the air, so it remains hot and fluid and can flow for tens of kilometres, thus contributing to the large size and low slopes of shield volcanoes. The

Hawaiian volcanoes are riddled with thousands of old lava tubes, some as long as 50 km.

Figure 4.19 Images of Kilauea volcano taken in 2002 (b & c) and 2007 (a & d) [SE photos] (a) Pu'u'O'o cinder cone in the background with tephra in the foreground and aa lava in the middle, (b) Formation of pahoehoe on the southern edge of Kilauea, (c) Formation of aa on a steep slope on Kilauea, (d) Skylight in an active lava tube, Kilauea.

Kilauea is approximately 300 ka old, while neighbouring Mauna Loa is over 700 ka and Mauna Kea is over 1 Ma. If volcanism continues above the Hawaii mantle plume in the same manner that it has for the past 85 Ma, it is likely that Kilauea will continue to erupt for at least another 500,000 years. By that time, its neighbour, Loihi, will have emerged from the sea floor, and its other neighbours, Mauna Loa and Mauna Kea, will have become significantly eroded, like their cousins, the islands to the northwest (Figure 4.15).

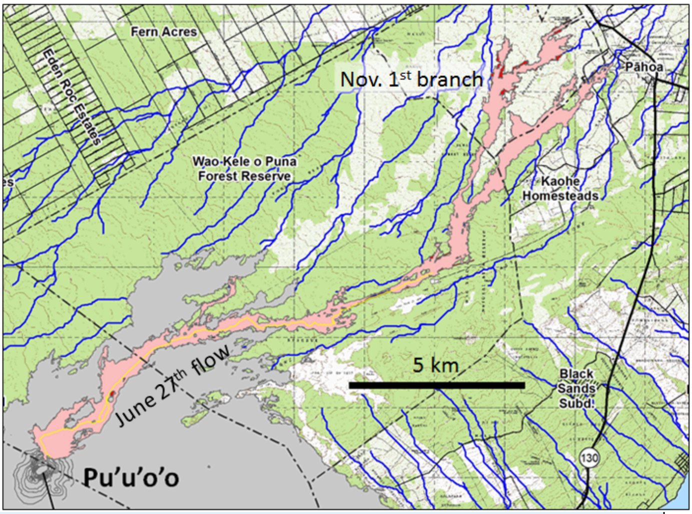
Exercises

Exercise 4.4 Kilauea's June 27th Lava Flow

The U.S. Geological Survey Hawaii Volcano Observatory (HVO) map shown here, dated January 29, 2015, shows the outline of lava that started flowing northeast from Pu'u 'O'o on June 27, 2004 (the "June 27th Lava flow," a.k.a. the "East Rift Lava Flow"). The flow reached the nearest settlement, Pahoa, on October 29, after covering a distance of 20 km in 124 days. After damaging some infrastructure west of Pahoa, the flow stopped advancing. A new outbreak occurred November 1, branching out to the north from the main flow about 6 km southwest of Pahoa.

1. What is the average rate of advance of the flow front from June 27 to October 29, 2014, in m/day and m/hour?

2. Go to the Kilauea page of the HVO website at: http://hvo.wr.usgs.gov/activity/kilaueastatus.php to compare the current status of the June 27th (or East Rift) lava flow with that shown on the map below.



[From USGS HVO: http://hvo.wr.usgs.gov/maps/]

Large Igneous Provinces

While the Hawaii mantle plume has produced a relatively low volume of magma for a very long time (~85 Ma), other mantle plumes are less consistent, and some generate massive volumes of magma over relatively short time periods. Although their origin is still controversial, it is thought that the volcanism leading to **large igneous provinces** (LIP) is related to very high volume but relatively short duration bursts of magma from mantle plumes. An example of an LIP is the Columbia River Basalt Group (CRGB), which extends across Washington, Oregon, and Idaho (Figure 4.20). This volcanism, which covered an area of about 160,000 km2 with basaltic rock up to several hundred metres thick, took place between 17 and 14 Ma.

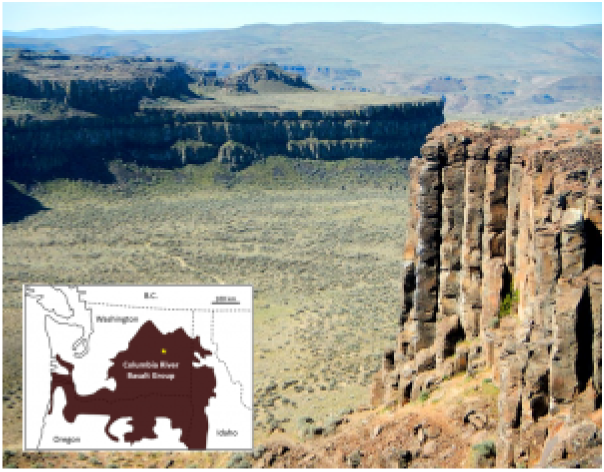


Figure 4.20 A part of the Columbia River Basalt Group at Frenchman Coulee, eastern Washington. All of the flows visible here have formed large (up to two metres in diameter) columnar basalts, a result of relatively slow cooling of flows that are tens of m thick. The inset map shows the approximate extent of the 17 to 14 Ma Columbia River Basalts, with the location of the photo shown as a star. [SE – photo and drawing]

Most other LIP eruptions are much bigger. The Siberian Traps (also basalt), which erupted at the end of the Permian period at 250 Ma, are estimated to have produced approximately 40 times as much lava as the CRBG.

The mantle plume that is assumed to be responsible for the CRBG is now situated beneath the Yellowstone area, where it leads to felsic volcanism. Over the past 2 Ma three very large explosive eruptions at Yellowstone have yielded approximately 900 km^3 of felsic magma, about 900 times the volume of the 1980 eruption of Mt. St. Helens, but only 5% of the volume of mafic magma in the CRBG.

Sea-Floor Volcanism

Some LIP eruptions occur on the sea floor, the largest being the one that created the Ontong Java plateau in the western Pacific Ocean at around 122 Ma. But most sea-floor volcanism originates at divergent boundaries and involves relatively low-volume eruptions. Under these conditions, hot lava that oozes out into the cold seawater quickly cools on the outside and then behaves a little like toothpaste. The resulting blobs of lava are known as **pillows**, and they tend to form piles around a sea-floor lava vent (Figure 4.21). In terms of area, there is very likely more pillow basalt on the sea floor than any other type of rock on Earth.

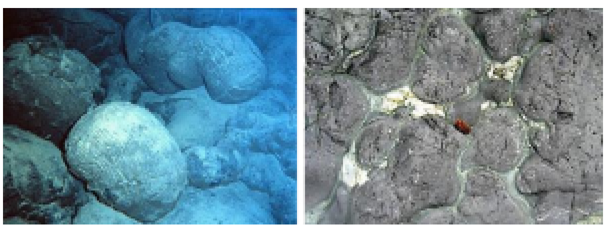


Figure 4.21 Modern and ancient sea-floor pillow basalts (left) Modern sea-floor pillows in the south Pacific [NOAA, from http://en.wikipedia.org/wiki/Basalt#mediaviewer/File:Pillow_basalt_crop_l.jpg] (right) Eroded 40 to 50 Ma pillows on the shore of Vancouver Island, near to Sooke. The pillows are 30 to 40 cm in diameter. [SE]

Kimberlites

While all of the volcanism discussed so far is thought to originate from partial melting in the upper mantle or within the crust, there is a special class of volcanoes called **kimberlites** that have their origins much deeper in the mantle, at depths of 150 km to 450 km. During a kimberlite eruption, material from this depth may make its way to surface quickly (hours to days) with little interaction with the surrounding rocks. As a result, kimberlite eruptive material is representative of mantle compositions: it is ultramafic.

Kimberlite eruptions that originate at depths greater than 200 km, within areas beneath old thick crust (**shields**), traverse the region of stability of diamond in the mantle, and in some cases, bring diamond-bearing material to the surface. All of the diamond deposits on Earth are assumed to have formed in this way; an example is the rich Ekati Mine in the Northwest Territories (Figure 4.22).

Figure 4.22 Ekati diamond mine, Northwest Territories, part of the Lac de Gras kimberlite field [http://upload.wikimedia.org/wikipedia/commons/8/88/Ekati_mine_640px.jpg]

The kimberlites at Ekati erupted between 45 and 60 Ma. Many kimberlites are older, some much older. There have been no kimberlite eruptions in historic times. The youngest known kimberlites are in the Igwisi Hills in Tanzania and are only about 10,000 years old. The next youngest known are around 30 Ma old.

6.4 Volcanic Hazards

There are two classes of volcanic hazards, direct and indirect. Direct hazards are forces that directly kill or injure people, or destroy property or wildlife habitat. Indirect hazards are volcanism-induced environmental changes that lead to distress, famine, or habitat destruction. Indirect effects of volcanism have accounted for approximately 8 million deaths during historical times, while direct effects have accounted for fewer than 200,000, or 2.5% of the total. Some of the more important types of volcanic hazards are summarized in Table 4.2.

Type	Description	Risk
Tephra emissions	Small particles of volcanic rock emitted into the atmosphere	Respiration problems for some individuals Significant climate cooling and famine Damage to aircraft
Gas emissions	The emission of gases before, during, and after an eruption	Climate cooling leading to crop failure and famine In some cases, widespread poisoning
Pyroclastic density current	A very hot (several 100°C) mixture of gases and volcanic tephra that flows rapidly (up to 100s of km/h) down the side of a volcano	Extreme hazard — destroys anything in the way
Pyroclastic fall	Vertical fall of tephra in the area surrounding an eruption	Thick tephra coverage of areas close to the eruption (km to 10s of km) Collapsed roofs
Lahar	A flow of mud and debris down a channel leading away from a volcano, triggered either by an eruption or a severe rain event	Severe risk of destruction for anything within the channel — lahar mud flows can move at 10s of km/h

Sector collapse/ debris avalanche	The failure of part of a volcano, either due to an eruption or for some other reason, leading to the failure of a large portion of the volcano	Severe risk of destruction for anything in the path of the debris avalanche
Lava flow	The flow of lava away from a volcanic vent	People and infrastructure at risk, but lava flows tend to be slow (km/h) and are relatively easy to avoid

Table 4.2 A summary of the important volcanic hazards

Volcanic Gas and Tephra Emissions

Large volumes of **tephra** (rock fragments, mostly **pumice**) and gases are emitted during major **plinian eruptions** (large explosive eruptions with hot gas a tephra columns extending into the stratosphere) at composite volcanoes, and a large volume of gas is released during some very high-volume effusive eruptions. One of the major effects is cooling of the climate by 1° to 2°C for several months to a few years because the dust particles and tiny droplets and particles of sulphur compounds block the sun. The last significant event of this type was in 1991 and 1992 following the large eruption of Mt. Pinatubo in the Philippines. A drop of 1° to 2°C may not seem like very much, but that is the global average amount of cooling, and cooling was much more severe in some regions and at some times.

Over an eight-month period in 1783 and 1784, a massive effusive eruption took place at the Laki volcano in Iceland. Although there was relatively little volcanic ash involved, a massive amount of sulphur dioxide was released into the atmosphere, along with a significant volume of hydrofluoric acid (HF). The sulphate **aerosols** that formed in the atmosphere led to dramatic cooling in the northern hemisphere. There were serious crop failures in Europe and North America, and a total of 6 million people are estimated to have died from famine and respiratory complications. In Iceland, poisoning from the HF resulted in the death of 80% of sheep, 50% of cattle, and the ensuing famine, along with HF poisoning, resulted in more than 10,000 human deaths, about 25% of the population.

Volcanic ash can also have serious implications for aircraft because it can destroy jet engines. For example, over 5 million airline passengers had their travel disrupted by the 2010 Eyjafjallajökull volcanic eruption in Iceland.

Pyroclastic Density Currents

In a typical explosive eruption at a composite volcano, the tephra and gases are ejected with explosive force and are hot enough to be forced high up into the atmosphere. As the eruption proceeds, and the amount of gas in the rising magma starts to decrease, parts will become heavier than air, and they can then flow downward along the flanks of the volcano (Figure 4.23). As they descend, they cool more and flow faster, reaching speeds up to several hundred km/h. A **pyroclastic density current** (PDC) consists of tephra ranging in size from boulders to microscopic shards of glass (made up of the edges and junctions of the bubbles of shattered pumice), plus gases (dominated by water vapour, but also including other gases). The temperature of this material can be as high as 1000°C. Among the most famous PDCs are the one that destroyed Pompeii in the year 79 CE, killing an estimated 18,000 people, and the one that destroyed the town of St. Pierre, Martinique, in 1902, killing an estimated 30,000.

The buoyant upper parts of pyroclastic density currents can flow over water, in some cases for several kilometres. The 1902 St. Pierre PDC flowed out into the harbour and destroyed several wooden ships anchored there.

Figure 4.23 The plinian eruption of Mt. Mayon, Philippines. in 1984. Although most of the eruption column is ascending into the atmosphere, there are pyroclastic density currents flowing down the sides of the volcano in several places. [USGS photo from: http://upload.wikimedia.org/wikipedia/commons/7/73/ Pyroclastic_flows_at_Mayon_Volcano.jpg]

Pyroclastic Fall

Most of the tephra from an explosive eruption ascends high into the atmosphere, and some of it is distributed around Earth by high-altitude winds. The larger components (larger than 0.1 mm) tend to fall relatively close to the volcano, and the amount produced by large eruptions can cause serious damage and casualties. The large 1991 eruption of Mt. Pinatubo in the Philippines resulted in the accumulation of tens of centimetres of ash in fields and on rooftops in the surrounding populated region.

Heavy typhoon rains that hit the island at the same time added to the weight of the tephra, leading to the collapse of thousands of roofs and to at least 300 of the 700 deaths attributed to the eruption.

Lahar

A **lahar** is any mudflow or **debris flow** that is related to a volcano. Most are caused by melting snow and ice during an eruption, as was the case with the lahar that destroyed the Colombian town of Armero in 1985 (described earlier). Lahars can also happen when there is no volcanic eruption, and one of the reasons is that, as we’ve seen, composite volcanoes tend to be weak and easily eroded.

In October 1998, category 5 hurricane Mitch slammed into the coast of central America. Damage was extensive and 19,000 people died, not so much because of high winds but because of intense rainfall — some regions received almost 2 m of rain over a few days! Mudflows and debris flows occurred in many areas, especially in Honduras and Nicaragua. An example is Casita Volcano in Nicaragua, where the heavy rains weakened rock and volcanic debris on the upper slopes, resulting in a debris flow that rapidly built in volume as it raced down the steep slope, and then ripped through the towns of El Porvenir and Rolando Rodriguez killing more than 2,000 people (Figure 4.24). El Porvenir and Rolando Rodriguez were new towns that had been built without planning approval in an area that was known to be at risk of lahars.

Figure 4.24 Part of the path of the lahar from Casita Volcano, October 30, 1998. [USGS photo from: http://volcanoes.usgs.gov/hazards/lahar/casita.php]

Sector Collapse and Debris Avalanche

In the context of volcanoes, **sector collapse** or flank collapse is the catastrophic failure of a significant part of an existing volcano, creating a large debris avalanche. This hazard was first recognized with the failure of the north side of Mt. St. Helens immediately prior to the large eruption on May 18, 1980. In the weeks before the eruption, a large bulge had formed on the side of the volcano, the result of magma transfer from depth into a satellite magma body within the mountain itself. Early on the morning of May 18, a moderate earthquake struck nearby; this is thought to have destabilized the bulge, leading to Earth’s largest observed landslide in historical times. The failure of this part of the volcano exposed the underlying satellite magma chamber, causing it to explode sideways, which exposed the conduit leading to the magma chamber below. The resulting plinian eruption — with a 24 km high eruption column — lasted for nine hours.

In August 2010, a massive part of the flank of B.C.'s Mt. Meager gave way and about 48 million cubic metres of rock rushed down the valley, one of the largest slope failures in Canada in historical times (Figure 4.25). More than 25 slope failures have taken place at Mt. Meager in the past 8,000 years, some of them more than 10 times larger than the 2010 failure.

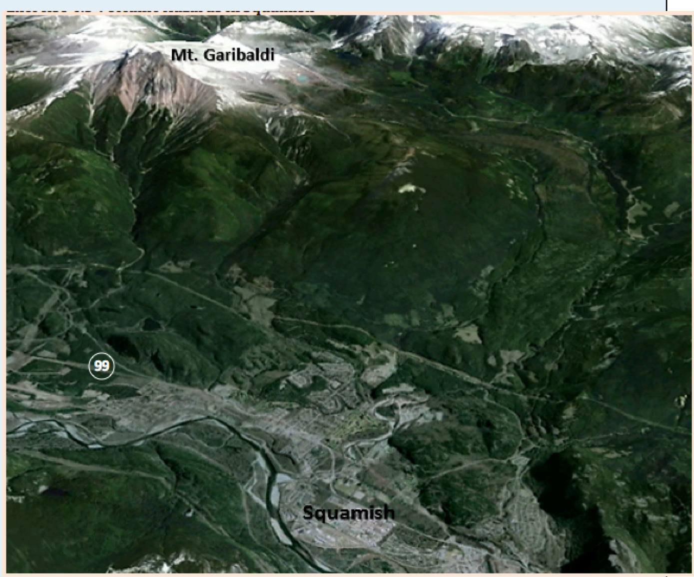
Figure 4.25 The August 2010 Mt. Meager rock avalanche, showing where the slide originated (arrow, 4 km upstream), its path down a steep narrow valley, and the debris field (and the stream that eventually cut through it) in the foreground. (Mika McKinnon photo, http://www.geomika.com/blog/2011/ 01/05/ the-trouble-with-landslides/ Used with permission (mika@geomika.com)

Lava Flows

As we saw in Exercise 4.4, lava flows at volcanoes like Kilauea do not advance very quickly, and in most cases, people can get out of the way. Of course, it is more difficult to move infrastructure, and so buildings and roads are typically the main casualties of lava flows.

Exercises

Exercise 4.5 Volcanic Hazards in Squamish



The town of Squamish is situated approximately 10 km from Mt. Garibaldi, as shown in the photo. In the event of a major eruption of Mt. Garibaldi, which of the following hazards has the potential to be an issue for the residents of Squamish or for those passing through on Highway 99? [SE after Google Earth]

Hazard	**Yes or No, and Brief Explanation**
Tephra emission	
Gas emission	
Pyroclastic density current	
Pyroclastic fall	
Lahar	
Sector collapse	
Lava flow	

6.5 Monitoring Volcanoes and Predicting Eruptions

In 2005 USGS geologist Chris Newhall made a list of the six most important signs of an imminent volcanic eruption. They are as follows:

1. *Gas leaks* — the release of gases (mostly H_2O, CO_2, and SO_2) from the magma into the atmosphere through cracks in the overlying rock
2. *Bit of a bulge* — the deformation of part of the volcano, indicating that a magma chamber at depth is swelling or becoming more pressurized
3. *Getting shaky* — many (hundreds to thousands) of small earthquakes, indicating that magma is on the move. The quakes may be the result of the magma forcing the surrounding rocks to crack, or a harmonic vibration that is evidence of magmatic fluids moving underground.
4. *Dropping fast* — a sudden decrease in the rate of seismicity, which may indicate that magma has stalled, which could mean that something is about to give way
5. *Big bump* — a pronounced bulge on the side of the volcano (like the one at Mt. St. Helens in 1980), which may indicate that magma has moved close to surface
6. *Blowing off steam* — steam eruptions (a.k.a.

> **phreatic eruptions**) that happen when magma near the surface heats groundwater to the boiling point. The water eventually explodes, sending fragments of the overlying rock far into the air.

With these signs in mind, we can make a list of the equipment we should have and the actions we can take to monitor a volcano and predict when it might erupt.

Assessing seismicity: The simplest and cheapest way to monitor a volcano is with seismometers. In an area with several volcanoes that have the potential to erupt (e.g., the Squamish-Pemberton area), a few well-placed seismometers can provide us with an early warning that something is changing beneath one of the volcanoes, and that we need to take a closer look. There are currently enough seismometers in the Lower Mainland and on Vancouver Island to provide this information.[1]

If there is seismic evidence that a volcano is coming to life, more seismometers should be placed in locations within a few tens of kilometres of the source of the activity (Figure 4.26). This will allow geologists to determine the exact location and depth of the seismic activity so that they can see where the magma is moving.

1. See: http://www.earthquakescanada.nrcan.gc.ca/stndon/CNSN-RNSC/stnbook-cahierstn/index-eng.php?tpl_sorting=map&CHIS_SZ=west

Figure 4.26 A seismometer installed in 2007 in the vicinity of the Nazco Cone, B.C. [photo Cathie Hickson, used with permission]

Detecting gases: Water vapour quickly turns into clouds of liquid water droplets and is relatively easy to detect just by looking, but CO_2 and SO_2 are not as obvious. It's important to be able to monitor changes in the composition of volcanic gases, and we need instruments to do that. Some can be monitored from a distance (from the ground or even from the air) using infrared devices, but to obtain more accurate data, we need to sample the air and do chemical analysis. This can be achieved with instruments placed on the ground close to the source of the gases (see Figure 4.17), or by collecting samples of the air and analyzing them in a lab.

Measuring deformation: There are two main ways to measure ground deformation at a volcano. One is known as a **tiltmeter**, which is a sensitive three-directional level that can sense small changes in the tilt of the ground at a specific location. Another is through the use of GPS (global positioning system) technology (Figure 4.27). GPS is more effective than a tiltmeter because it provides information on how far the ground has actually moved — east-west, north-south, and up-down.

Figure 4.27 A GPS unit installed at Hualalai volcano, Hawaii. The dish-shaped antenna on the right is the GPS receiver. The antenna on the left is for communication with a base station. [from USGS at: http://hvo.wr.usgs.gov/volcanowatch/view.php?id=173]

By combining information from these types of sources, along with careful observations made on the ground and from the air, and a thorough knowledge of how volcanoes work, geologists can get a good idea of the potential for a volcano to erupt in the near future (months to weeks, but not days). They can then make recommendations to authorities about the need for evacuations and restricting transportation corridors. Our ability to predict volcanic eruptions has increased dramatically in recent decades because of advances in our understanding of how volcanoes behave and in monitoring technology. Providing that careful work is done, there is no longer a large risk of surprise eruptions, and providing that public warnings are issued and heeded, it is less and less likely that thousands will die from sector collapse, pyroclastic flows, ash falls, or lahars. Indirect hazards are still very real, however, and we can expect the next eruption like the one at Laki in 1783 to take an even greater toll than it did then, especially since there are now roughly eight times as many people on Earth.

Exercises

Exercise 4.6 Volcano Alert!

You're the chief volcanologist for the Geological Survey of Canada (GSC), based in Vancouver. At 10:30 a.m. on a Tuesday, you receive a report from a seismologist at the GSC in Sidney saying that there has been a sudden increase in the number of small earthquakes in the vicinity of Mt. Garibaldi. You have two technicians available, access to some monitoring equipment, and a four-wheel-drive vehicle. At noon, you meet with your technicians and a couple of other geologists. By the end of the day, you need to have a plan to implement, starting tomorrow morning, and a statement to release to the press. What should your first day's fieldwork include? What should you say later today in your press release?

6.6 Volcanoes in British Columbia

As shown on the Figure 4.28, three types of volcanic environments are represented in British Columbia:

- The Cascade Arc (a.k.a. the Garibaldi Volcanic Belt in Canada) is related to subduction of the Juan de Fuca Plate beneath the North America plate.
- The Anahim Volcanic Belt is assumed to be related to a mantle plume.
- The Stikine Volcanic Belt and the Wells Gray-Clearwater Volcanic Field are assumed to be related to crustal rifting.

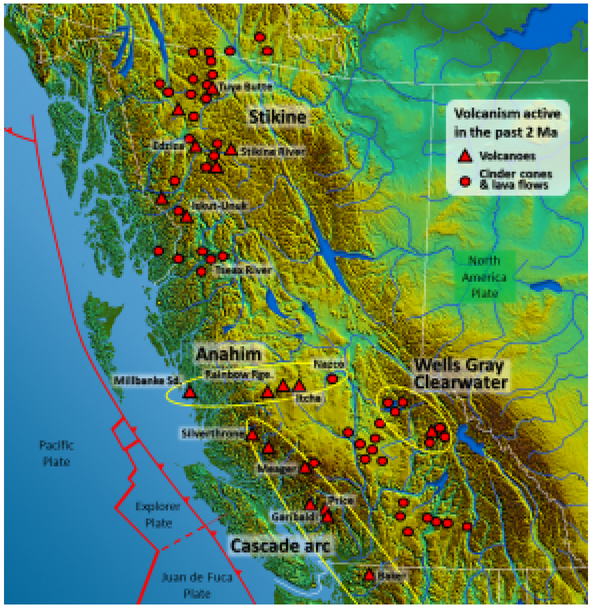


Figure 4.28 Major volcanic centres in British Columbia (base map from Wikipedia (http://commons.wikimedia.org/wiki/File:South-West_Canada.jpg). Volcanic locations from Wood, D., 1993, Waiting for another big blast – probing B.C.'s volcanoes, Canadian Geographic, based on the work of Cathie Hickson)

Subduction Volcanism

Southwestern British Columbia is at the northern end of the Juan de Fuca (Cascadia) subduction zone, and the volcanism there is related to magma generation by flux melting in the upper mantle above the subducting plate. In general, there has been a much lower rate and volume of volcanism in the B.C. part of this belt than in the U.S. part. One reason for this is that the northern part of the Juan de Fuca Plate (i.e., the Explorer Plate) is either not subducting, or is subducting at a slower rate than the rest of the plate. There are several volcanic centres in the Garibaldi Volcanic

Belt: the Garibaldi centre (including Mt. Garibaldi and the Black Tusk-Mt. Price area adjacent to Garibaldi Lake (Figures 4.1 and 4.2), Mt. Cayley, and Mt. Meager (Figure 4.25). The most recent volcanic activity in this area was at Mt. Meager. Approximately 2,400 years ago, an explosive eruption of about the same magnitude as the 1980 Mt. St. Helens eruption took place at Mt. Meager. Ash spread as far east as Alberta. There was also significant eruptive activity at Mts. Price and Garibaldi approximately 12,000 and 10,000 years ago during the last glaciation; in both cases, lava and tephra built up against glacial ice in the adjacent valley (Figure 4.29). The Table in Figure 4.2 at the beginning of this chapter is a **tuya**, a volcano that formed beneath glacial ice and had its top eroded by the lake that formed around it in the ice.

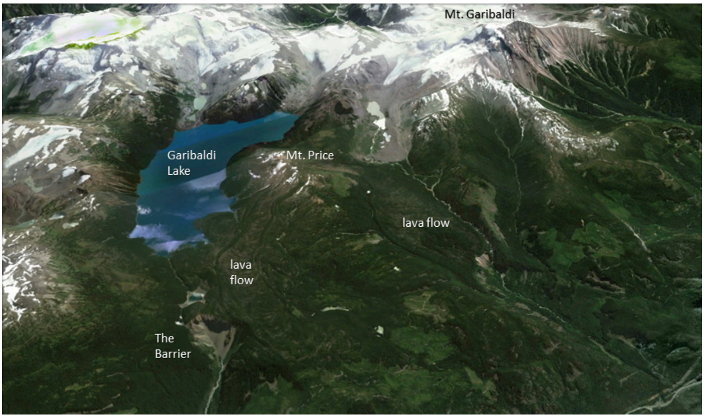


Figure 4.29 Perspective view of the Garibaldi region (looking east) showing the outlines of two lava flows from Mt. Price. Volcanism in this area last took place when the valley in the foreground was filled with glacial ice. The cliff known as the Barrier formed when part of the Mt. Price lava flow failed after deglaciation. The steep western face of Mt. Garibaldi formed by sector collapse, also because the rocks were no longer supported by glacial ice. [SE after Google Earth]

Mantle Plume Volcanism

The chain of volcanic complexes and cones extending from Milbanke Sound to Nazko Cone is interpreted as being related to a mantle plume currently situated close to the Nazko Cone, just west of Quesnel. The North America Plate is moving in a westerly direction at about 2 cm per year with respect to this plume, and the series of now partly eroded shield volcanoes between Nazco and the coast is interpreted to have been formed by the plume as the continent moved over it.

The Rainbow Range, which formed at approximately 8 Ma, is the largest of these older volcanoes. It has a diameter of about 30 km and an elevation of 2,495 m (Figure 4.30). The name "Rainbow" refers to the bright colours displayed by some of the volcanic rocks as they weather.

Figure 4.30 Rainbow Range, Chilcotin Plateau, B.C. (http://upload.wikimedia.org/wikipedia/commons/f/fd/Rainbow_Range_Colors.jpg).

Rift-Related Volcanism

While B.C. is not about to split into pieces, two areas of volcanism are related to rifting — or at least to stretching-related fractures that might extend through the crust. These are the Wells Gray-Clearwater volcanic field southeast of Quesnel, and the Northern Cordillera Volcanic Field, which ranges across the northwestern corner of the province (as already discussed in section 4.1). This area includes Canada’s most recent volcanic eruption, a cinder cone and mafic lava flow that formed around 250 years ago at the Tseax River Cone in the Nass River area north of Terrace. According to Nisga’a oral history, as many as 2,000 people died during that eruption, in which lava overran their village on the Nass River. Most of the deaths are attributed to asphyxiation from volcanic gases, probably carbon dioxide.

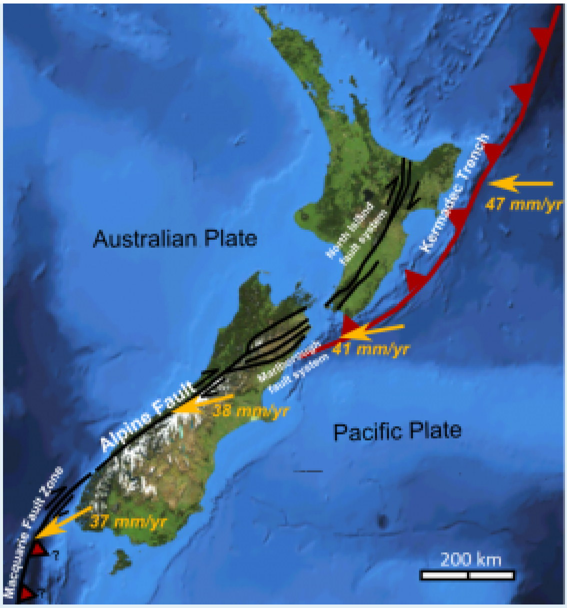
The Mount Edziza Volcanic Field near the Stikine River is a large area of lava flows, sulphurous ridges, and cinder cones. The most recent eruption in this area was about 1,000 years ago. While most of the other volcanism in the Edziza region is mafic and involves lava flows and cinder cones, Mt. Edziza itself (Figure 4.31) is a composite volcano with rock compositions ranging from rhyolite to basalt. A possible explanation for the presence of composite volcanism in an area dominated by mafic flows and cinder cones is that there is a magma chamber beneath this area, within which magma differentiation is taking place.

Figure 4.31 Mount Edziza, in the Stikine area, B.C., with Eve Cone in the foreground. (http://upload.wikimedia.org/wikipedia/commons/5/54/Mount_Edziza%2C_British_Columbia.jpg).

Exercises

Exercise 4.7 Volcanoes Down Under

This map shows the plate tectonic situation in the area around New Zealand.



1. Based on what you know about volcanoes in B.C., predict where you might expect to see volcanoes in and around New Zealand.

2. What type of volcanoes would you expect to find in and around New Zealand?

[from: http://upload.wikimedia.org/wikipedia/commons/8/8a/NZ_faults.png]

Chapter 6 Summary

The topics covered in this chapter can be summarized as follows:

6.1	Plate Tectonics and Volcanism	Volcanism is closely related to plate tectonics. Most volcanoes are associated with convergent plate boundaries (at subduction zones), and there is also a great deal of volcanic activity at divergent boundaries and areas of continental rifting. At convergent boundaries magma is formed where water from a subducting plate acts as a flux to lower the melting temperature of the adjacent mantle rock. At divergent boundaries magma forms because of decompression melting. Decompression melting also takes place within a mantle plume.
6.2	Magma Composition and Eruption Style	The initial magmas in most volcanic regions are mafic in composition, but they can evolve into more felsic types through interaction with crustal rock, and as a result of crystal settling within a magma chamber. Felsic magmas tend to have higher gas contents than mafic magmas, and they are also more viscous. The higher viscosity prevents gases from escaping from the magma, and so felsic magmas are more pressurized and more likely to erupt explosively.

6.3	Types of Volcanoes	Cinder cones, which can form in various volcanic settings, are relatively small volcanoes that are composed mostly of mafic rock fragments that were formed during a single eruptive event. Composite volcanoes are normally associated with subduction, and while their magma tends to be intermediate on average, it can range all the way from felsic to mafic. The corresponding differences in magma viscosity lead to significant differences in eruptions style. Most shield volcanoes are associated with mantle plumes, and have consistently mafic magma which generally erupts as lava flows.
6.4	Volcanic Hazards	Most direct volcanic hazards are related to volcanoes that erupt explosively, especially composite volcanoes. Pyroclastic density currents, some as hot as 1000˚C can move at hundreds of km/h and will kill anything in the way. Lahars, volcano-related mudflows, can be large enough to destroy entire towns. Lava flows will destroy anything in their paths, but tend to move slowly enough so that people can get to safety.
6.5	Monitoring Volcanoes and Predicting Eruptions	We have the understanding and technology to predict volcanic eruptions with some success, and to ensure that people are not harmed. The prediction techniques include monitoring seismicity in volcanic regions, detecting volcanic gases, and measuring deformation of the flanks of a volcano.

6.6	Volcanoes in British Columbia	There are examples of all of the important types of volcanoes in British Columbia, including subduction volcanism north of Vancouver, mantle-plume volcanism along the Nazco trend, and rift-related volcanism in the Wells Gray and Stikine regions.

Questions for Review

1. What are the three main tectonic settings for volcanism on Earth?
2. What is the primary mechanism for partial melting at a convergent plate boundary?
3. Why are the viscosity and gas content of a magma important in determining the type of volcanic rocks that will be formed when that magma is extruded?
4. Why do the gases in magma not form gas bubbles when the magma is deep within the crust?
5. Where do pillow lavas form? Why do they form and from what type of magma?
6. What two kinds of rock textures are typically found in a composite volcano?
7. What is a lahar, and why are lahars commonly associated with eruptions of composite volcanoes?
8. Under what other circumstances might a lahar form?
9. Explain why shield volcanoes have such

gentle slopes.

10. In very general terms, what is the lifespan difference between a composite volcano and a shield volcano?
11. Why is weak seismic activity (small earthquakes) typically associated with the early stages of a volcanic eruption?
12. How can GPS technology be used to help monitor a volcano in the lead-up to an eruption?
13. What type of eruption at Mt. St. Helens might have produced columnar basalts?
14. What is the likely geological origin of the Nazko Cone?
15. What might be the explanation for southwestern B.C. having much less subduction-related volcanism than adjacent Washington and Oregon?
16. What was the likely cause of most of the deaths from the most recent eruption at the Tseax River Cone?

Chapter 7: Mass Wasting

Learning Objectives

After reading this chapter, completing the exercises within it, and answering the questions at the end, you should be able to:

- Explain how slope stability is related to slope angle
- Summarize some of the factors that influence the strength of materials on slopes, including type of rock, presence and orientation of planes of weakness such as bedding or fractures, type of unconsolidated material, and the effects of water
- Explain what types of events can trigger mass wasting
- Summarize the types of motion that can happen during mass wasting
- Describe the main types of mass wasting — creep, slump, translational slide, rotational slide, fall, and debris flow or mudflow — in terms of the types of materials involved, the type of motion, and the likely rates of motion
- Explain what steps we can take to delay mass wasting, and why we cannot prevent it permanently
- Describe some of the measures that can be

taken to mitigate the risks associated with mass wasting

Figure 15.1 The site of the 1965 Hope Slide as seen in 2014. The initial failure is thought to have taken place along the foliation planes and sill within the area shown in the inset. [SE]

Early in the morning on January 9, 1965, 47 million cubic metres of rock broke away from the steep upper slopes of Johnson Peak (16 km southeast of Hope) and roared 2,000 m down the mountain, gouging out the contents of a small lake at the bottom, and continuing a few hundred metres up the other side (Figure 15.1). Four people, who had been stopped on the highway by a snow avalanche, were killed. Many more might have become victims, except that a Greyhound bus driver, en route to Vancouver, turned his bus around on seeing the avalanche. The rock failed along weakened foliation planes of the metamorphic rock on Johnson Peak, in an area that had been eroded into a steep slope by glacial ice. There is no evidence that it was triggered by any specific event, and there was no warning that it was about to happen. Even if there had been warning, nothing could have been done to prevent it. There are hundreds of similar situations throughout British Columbia.

What can we learn from the Hope Slide? In general, we cannot prevent most mass wasting, and significant effort is required if an event is to be predicted with any level of certainty. Understanding the geology is critical to understanding mass wasting. Although failures are inevitable in a region with steep slopes, larger ones happen less frequently than smaller ones, and the consequences vary depending on the downslope conditions, such as the presence of people, buildings, roads, or fish-bearing streams.

An important reason for learning about mass wasting is to understand the nature of the materials that fail, and how and why they fail so that we can minimize risks from similar events in the future. For this reason, we need to be able to classify mass-wasting events, and we need to know the terms that geologists, engineers, and others use to communicate about them.

Mass wasting, which is synonymous with "slope failure," is the failure and downslope movement of rock or unconsolidated materials in response to gravity. The term "landslide" is almost synonymous with mass wasting, but not quite because some people reserve "landslide" for relatively rapid slope failures, while others do not. Because of that ambiguity, we will avoid the use of "landslide" in this textbook.

7.1 Factors That Control Slope Stability

Mass wasting happens because tectonic processes have created uplift. Erosion, driven by gravity, is the inevitable response to that uplift, and various types of erosion, including mass wasting, have created slopes in the uplifted regions. Slope stability is ultimately determined by two factors: the angle of the slope and the strength of the materials on it.

In Figure 15.2 a block of rock situated on a rock slope is being pulled toward Earth's centre (vertically down) by gravity. We can split the vertical gravitational force into two components relative to the slope: one pushing the block down the slope (the **shear force**), and the other pushing into the slope (the **normal force**). The shear force, which wants to push the block down the slope, has to overcome the strength of the connection between the block and the slope, which may be quite weak if the block has split away from the main body of rock, or may be very strong if the block is still a part of the rock. This is the **shear strength**, and in Figure 15.2a, it greater than the shear force, so the block should not move. In Figure 15.2b the slope is steeper and the shear force is approximately equal to the shear strength. The block may or may not move under these circumstances. In Figure 15.2c, the slope is steeper still, so the shear force is considerably greater than the shear strength, and the block will very likely move.

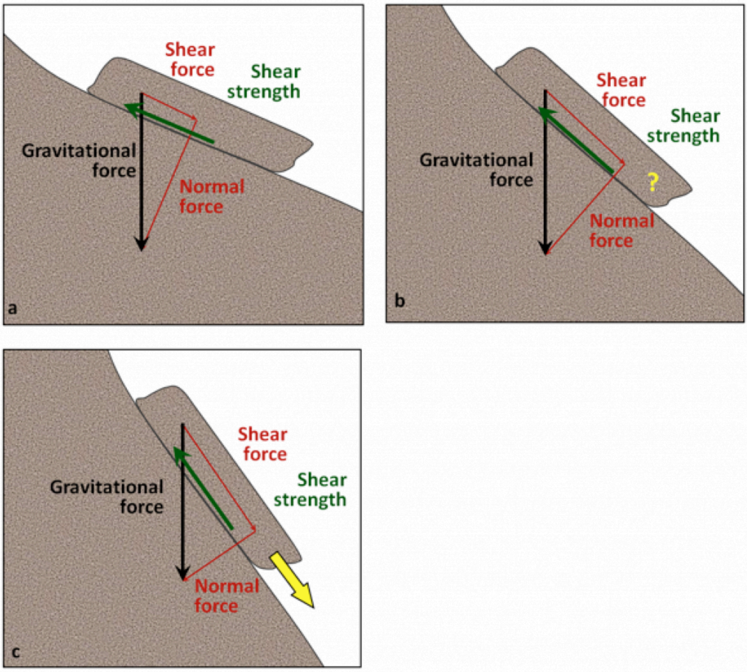


Figure 15.2 Differences in the shear and normal components of the gravitational force on slopes with differing steepness. The gravitational force is the same in all three cases. In (a) the shear force is substantially less than the shear strength, so the block should be stable. In (b) the shear force and shear strength are about equal, so the block may or may not move. In (c) the shear force is substantially greater than the shear strength, so the block is very likely to move. [SE]

As already noted, slopes are created by uplift followed by erosion. In areas with relatively recent uplift (such as most of British Columbia and the western part of Alberta), slopes tend to be quite steep. This is especially true where glaciation has taken place because glaciers in mountainous terrain create steep-sided valleys. In areas without recent uplift (such as central Canada), slopes are less steep because hundreds of millions of years of erosion (including mass wasting) has made them that way. However, as we'll see, some mass wasting can happen even on relatively gentle slopes.

The strength of the materials on slopes can vary widely. Solid rocks tend to be strong, but there is a very wide range of rock strength. If we consider just the strength of the rocks, and ignore issues like fracturing and layering, then most crystalline rocks — like granite, basalt, or gneiss — are very strong, while some metamorphic rocks — like schist — are moderately strong. Sedimentary rocks have variable strength. Dolostone and some limestone are strong, most sandstone and conglomerate are moderately strong, and some sandstone and all mudstones are quite weak.

Fractures, metamorphic foliation, or bedding can significantly reduce the strength of a body of rock, and in the context of mass wasting, this is most critical if the planes of weakness are parallel to the slope and least critical if they are perpendicular to the slope. This is illustrated in Figure 15.3. At locations A and B the bedding is nearly perpendicular to the slope and the situation is relatively stable. At location D the bedding is nearly parallel to the slope and the situation is quite unstable. At location C the bedding is nearly horizontal and the stability is intermediate between the other two extremes.

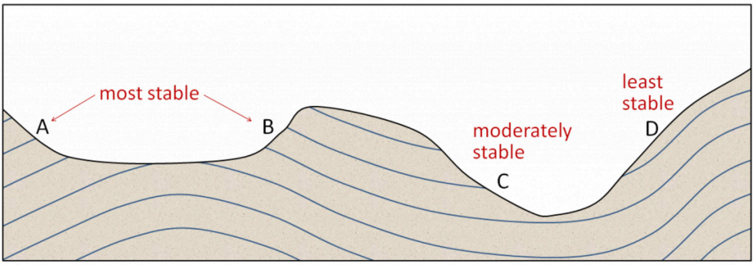


Figure 15.3 Relative stability of slopes as a function of the orientation of weaknesses (in this case bedding planes) relative to the slope orientations. [SE]

Internal variations in the composition and structure of rocks can significantly affect their strength. Schist, for example, may have layers that are rich in sheet silicates (mica or chlorite) and these will tend to be weaker than other layers. Some minerals tend to be more susceptible to weathering than others, and the weathered products are commonly quite weak (e.g., the clay formed from feldspar). The side of Johnson Peak that failed in 1965 (Hope Slide) is made up of chlorite schist (metamorphosed sea-floor basalt) that has feldspar-bearing sills within it (they are evident within the inset area of Figure 15.1). The foliation and the sills are parallel to the steep slope. The schist is relatively weak to begin with, and the feldspar in the sills, which has been altered to clay, makes it even weaker.

Unconsolidated sediments are generally weaker than sedimentary rocks because they are not cemented and, in most cases, have not been significantly compressed by overlying materials. This binding property of sediment is sometimes referred to as *cohesion*. Sand and silt tend to be particularly weak, clay is generally a little stronger, and sand mixed with clay can be stronger still. The deposits that make up the cliffs at Point Grey in Vancouver include sand, silt, and clay overlain by sand. As shown in Figure 15.4 (left) the finer deposits are relatively strong (they maintain a steep slope), while the overlying sand is relatively weak, and has a shallower slope that has recently failed. Glacial till — typically a mixture of clay, silt, sand, gravel, and larger clasts — forms and is compressed beneath tens to thousands of metres of glacial ice so it can be as strong as some sedimentary rock (Figure 15.4, right).

Figure 15.4 Left: Glacial outwash deposits at Point Grey, in Vancouver. The dark lower layer is made up of sand, silt, and clay. The light upper layer is well-sorted sand. Right: Glacial till on Quadra Island, B.C. The till is strong enough to have formed a near-vertical slope. [SE]

Apart from the type of material on a slope, the amount of water that the material contains is the most important factor controlling its strength. This is especially true for unconsolidated materials, like those shown in Figure 15.4, but it also applies to bodies of rock. Granular sediments, like the sand at Point Grey, have lots of spaces between the grains. Those spaces may be completely dry (filled only with air); or moist (often meaning that some spaces are water filled, some grains have a film of water around them, and small amounts of water are present where grains are touching each other); or completely saturated (Figure 15.5). Unconsolidated sediments tend to be strongest when they are moist because the small amounts of water at the grain boundaries hold the grains together with surface tension. Dry sediments are held together only by the friction between grains, and if they are well sorted or well rounded, or both, that cohesion is weak. Saturated sediments tend to be the weakest of all because the large amount of water actually pushes the grains apart, reducing the mount friction between grains. This is especially true if the water is under pressure.

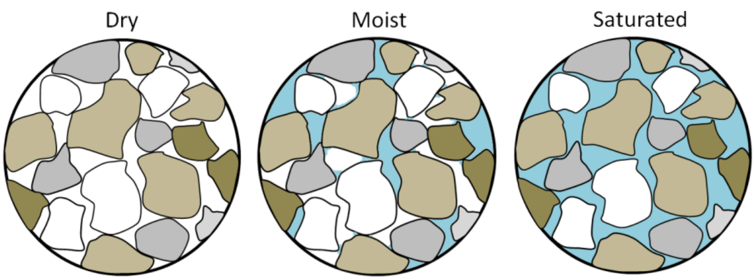


Figure 15.5 Depiction of dry, moist, and saturated sand [SE]

Exercises

Exercise 15.1 Sand and Water

If you've ever been to the beach, you'll already know that sand behaves differently when it's dry than it does when it's wet, but it's worth taking a systematic look at the differences in its behaviour. Find about half a cup of clean, dry sand (or get some wet sand and dry it out), and then pour it from your hand onto a piece of paper. You should be able to make a cone-shaped pile that has a slope of around 30°. If you pour more sand on the pile, it will get bigger, but the slope should remain the same. Now add some water to the sand so that it is moist. An easy way to do this is to make it completely wet and then let the water drain away for a minute. You should be able to form this moist sand into a steep pile (with slopes of around 80°). Finally, put the same sand into a cup and fill the cup with water so the sand is just covered. Swirl it around so that the sand remains in suspension, and then quickly tip it out onto a flat surface (best to do this outside). It should spread out over a wide area, forming a pile with a slope of only a few degrees. [SE]

Water will also reduce the strength of solid rock, especially if it has fractures, bedding planes, or clay-bearing zones. This effect is even more significant when the water is under pressure, which is why you'll often see holes drilled into rocks on road cuts to relieve this pressure. One of the hypotheses advanced to explain the 1965 Hope Slide is that the very cold conditions that winter caused small springs in the lower part of the slope to freeze over, preventing water from flowing out. It is possible that water pressure gradually built up within the slope, weakening the rock mass to the extent that the shear strength was no longer greater than the shear force.

Water also has a particular effect on clay-bearing materials. All clay minerals will absorb a little bit of water, and this reduces their strength. The **smectite** clays (such as the **bentonite** used in cat litter) can absorb a lot of water, and that water pushes the sheets apart at a molecular level and makes the mineral swell. Smectite that has expanded in this way has almost no strength; it is extremely slippery.

And finally, water can significantly increase the mass of the material on a slope, which increases the gravitational force pushing it down. A body of sediment that has 25% porosity and is saturated with water weighs approximately 13% more than it does when it is completely dry, so the gravitational shear force is also 13% higher. In the situation shown in Figure 15.2b, a 13% increase in the shear force could easily be enough to tip the balance between shear force and shear strength.

Mass-Wasting Triggers

In the previous section, we talked about the shear force and the shear strength of materials on slopes, and about factors that can reduce the shear strength. Shear force is primarily related to slope angle, and this does not change quickly. But shear strength can change quickly for a variety of reasons, and events that lead to a rapid reduction in shear strength are considered to be **triggers** for mass wasting.

An increase in water content is the most common mass-wasting trigger. This can result from rapid melting of snow or ice, heavy rain, or some type of event that changes the pattern of water flow on the surface. Rapid melting can be caused by a dramatic increase in temperature (e.g., in spring or early summer) or by a volcanic eruption. Heavy rains are typically related to storms. Changes in water flow patterns can be caused by earthquakes, previous slope failures that dam up streams, or human structures that interfere with runoff (e.g., buildings, roads, or parking lots). An example of this is the deadly 2005 debris flow in North Vancouver (Figure 15.6). The 2005 failure took place in an area that had failed previously, and a report written in 1980 recommended that the municipal authorities and residents take steps to address surface and slope drainage issues. Little was done to improve the situation.

Figure 15.6 The debris flow in the Riverside Drive area of North Vancouver in January, 2005 happened during a rainy period, but was likely triggered by excess runoff related to the roads at the top of this slope and by landscape features, including a pool, in the area surrounding the house visible here. [The Province, used with permission]

In some cases, a *decrease* in water content can lead to failure. This is most common with clean sand deposits (e.g., the upper layer in Figure 15.4 (left)), which lose strength when there is no more water around the grains.

Freezing and thawing can also trigger some forms of mass wasting. More specifically, the thawing can release a block of rock that was attached to a slope by a film of ice.

One other process that can weaken a body of rock or sediment is shaking. The most obvious source of shaking is an earthquake, but shaking from highway traffic, construction, or mining will also do the job. Several deadly mass-wasting events (including snow avalanches) were trigged by the M7.8 earthquake in Nepal in April 2015.

7.2 Classification of Mass Wasting

It's important to classify slope failures so that we can understand what causes them and learn how to mitigate their effects. The three criteria used to describe slope failures are:

- The type of material that failed (typically either bedrock or unconsolidated sediment)
- The mechanism of the failure (how the material moved)
- The rate at which it moved

The type of motion is the most important characteristic of a slope failure, and there are three different types of motion:

- If the material drops through the air, vertically or nearly vertically, it's known as a **fall.**
- If the material moves as a mass along a sloping surface (without internal motion within the mass), it's a **slide.**
- If the material has internal motion, like a fluid, it's a **flow**.

Unfortunately it's not normally that simple. Many slope failures involve two of these types of motion, some involve all three, and in many cases, it's not easy to tell how the material moved. The types of slope failure that we'll cover here are summarized in Table 15.1.

Failure Type	Type of Material	Type of Motion	Rate of Motion
Rock fall	Rock fragments	Vertical or near-vertical fall (plus bouncing in many cases)	Very fast (>10s m/s)
Rock slide	A large rock body	Motion as a unit along a planar surface (translational sliding)	Typically very slow (mm/y to cm/y), but some can be faster
Rock avalanche	A large rock body that slides and then breaks into small fragments	Flow (at high speeds, the mass of rock fragments is suspended on a cushion of air)	Very fast (>10s m/s)
Creep or solifluction	Soil or other overburden; in some cases, mixed with ice	Flow (although sliding motion may also occur)	Very slow (mm/y to cm/y)
Slump	Thick deposits (m to 10s of m) of unconsolidated sediment	Motion as a unit along a curved surface (rotational sliding)	Slow (cm/y to m/y)
Mudflow	Loose sediment with a significant component of silt and clay	Flow (a mixture of sediment and water moves down a channel)	Moderate to fast (cm/s to m/s)
Debris flow	Sand, gravel, and larger fragments	Flow (similar to a mudflow, but typically faster)	Fast (m/s)

Table 15.1 Classification of slope failures based on type of material and type of motion [SE]

Rock Fall

Rock fragments can break off relatively easily from steep bedrock slopes, most commonly due to frost-wedging in areas where there are many freeze-thaw cycles per year. If you've ever hiked along a steep mountain trail on a cool morning, you might have heard the occasional fall of rock fragments onto a **talus slope**. This happens because the water between cracks freezes and expands overnight, and then when that same water thaws in the morning sun, the fragments that had been pushed beyond their limit by the ice fall to the slope below (Figure 15.7).

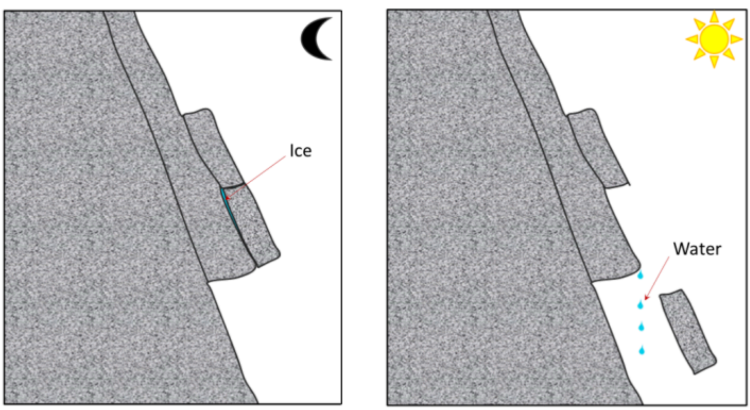


Figure 15.7 The contribution of freeze-thaw to rock fall [SE]

A typical talus slope, near Keremeos in southern B.C., is shown in Figure 15.8. In December 2014, a large block of rock split away from a cliff in this same area. It broke into smaller pieces that tumbled down the slope and crashed into the road, smashing the concrete barriers and gouging out large parts of the pavement. Luckily no one was hurt.

Figure 15.8 Left: A talus slope near Keremeos, B.C., formed by rock fall from the cliffs above. Right: The results of a rock fall onto a highway west of Keremeos in December 2014. [SE]

Rock Slide

A rock slide is the sliding motion of rock along a sloping surface. In most cases, the movement is parallel to a fracture, bedding, or metamorphic foliation plane, and it can range from very slow to moderately fast. The word **sackung** describes the very slow motion of a block of rock (mm/y to cm/y) on a slope. A good example is the Downie Slide north of Revelstoke, B.C., which is shown in Figure 15.9. In this case, a massive body of rock is very slowly sliding down a steep slope along a plane of weakness that is approximately parallel to the slope. The Downie Slide, which was recognized prior to the construction of the Revelstoke Dam, was moving very slowly at the time (a few cm/year). Geological engineers were concerned that the presence of water in the reservoir (visible in Figure 15.9) could further weaken the plane of failure, leading to an acceleration of the motion. The result would have been a catastrophic failure into the reservoir that would have sent a wall of water over the dam and into the community of Revelstoke. During the construction of the dam, they tunnelled into the rock at the base of the slide and drilled hundreds of drainage holes upward into the plane of failure. This allowed water to drain out so that the pressure was reduced, which reduced the rate of movement of the sliding block. BC Hydro monitors this site continuously; the slide block is currently moving more slowly than it was prior to the construction of the dam.

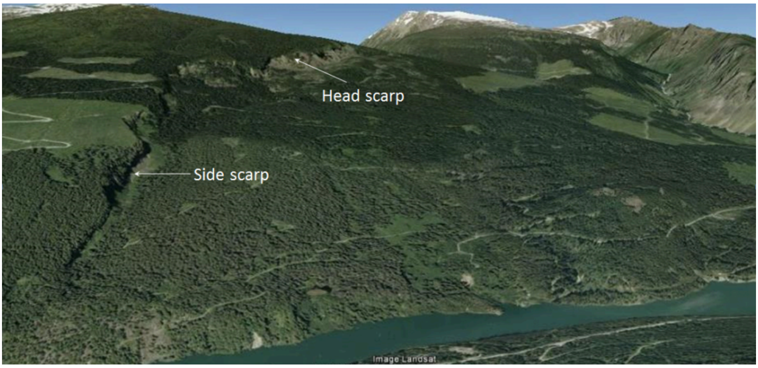


Figure 15.9 The Downie Slide, a sackung, on the shore of the Revelstoke Reservoir (above the Revelstoke Dam). The head scarp is visible at the top and a side-scarp along the left side. [from Google Earth]

In the summer of 2008, a large block of rock slid rapidly from a steep slope above Highway 99 near Porteau Cove (between Horseshoe Bay and Squamish). The block slammed into the highway and adjacent railway and broke into many pieces. The highway was closed for several days, and the slope was subsequently stabilized with rock bolts and drainage holes. As shown in Figure 15.10, the rock is fractured parallel to the slope, and this almost certainly contributed to the failure. However, it is not actually known what triggered this event as the weather was dry and warm during the preceding weeks, and there was no significant earthquake in the region.

Figure 15.10 Site of the 2008 rock slide at Porteau Cove. Notice the prominent fracture set parallel to the surface of the slope. The slope has been stabilized with rock bolts (top) and holes have been drilled into the rock to improve drainage (one is visible in the lower right). Risk to passing vehicles from rock fall has been reduced by hanging mesh curtains (background). [SE photo 2012]

Rock Avalanche

If a rock slides and then starts moving quickly (m/s), the rock is likely to break into many small pieces, and at that point it turns into a **rock avalanche**, in which the large and small fragments of rock move in a fluid manner supported by a cushion of air within and beneath the moving mass. The 1965 Hope Slide (Figure 15.1) was a rock avalanche, as was the famous 1903 Frank Slide in southwestern Alberta. The 2010 slide at Mt. Meager (west of Lillooet) was also a rock avalanche, and rivals the Hope Slide as the largest slope failure in Canada during historical times (Figure 15.11).

Figure 15.11 The 2010 Mt. Meager rock avalanche, showing where the slide originated (arrow, 4 km upstream). It then raced down a steep narrow valley and out into the wider valley in the foreground. [Mika McKinnon photo, http://www.geomika.com/blog/2011/01/05/the-trouble-with-landslides/ Used with permission.]

Creep or Solifluction

The very slow — mm/y to cm/y — movement of soil or other unconsolidated material on a slope is known as creep. **Creep**, which normally only affects the upper several centimetres of loose material, is typically a type of very slow flow, but in some cases, sliding may take place. Creep can be facilitated by freezing and thawing because, as shown in Figure 15.12, particles are lifted perpendicular to the surface by the growth of ice crystals within the soil, and then let down vertically by gravity when the ice melts. The same effect can be produced by frequent wetting and drying of the soil. In cold environments, **solifluction** is a more intense form of freeze-thaw-triggered creep.

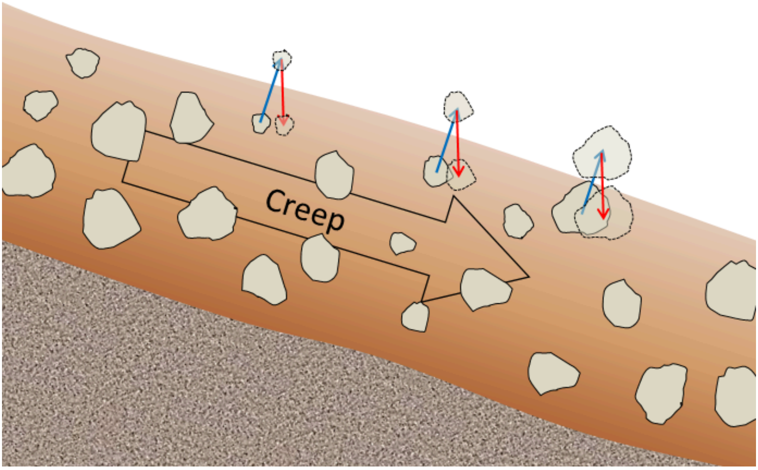


Figure 15.12 A depiction of the contribution of freeze-thaw to creep. The blue arrows represent uplift caused by freezing in the wet soil underneath, while the red arrows represent depression by gravity during thawing. The uplift is perpendicular to the slope, while the drop is vertical. [SE]

Creep is most noticeable on moderate-to-steep slopes where trees, fence posts, or grave markers are consistently leaning in a downhill direction (Figure 15.13). In the case of trees, they try to correct their lean by growing upright, and this leads to a curved lower trunk known as a "pistol butt."

Figure 15.13 Evidence of creep (tilted grave markers) at a cemetery in Nanaimo, B.C. [SE]

Slump

Slump is a type of slide (movement as a mass) that takes place within thick unconsolidated deposits (typically thicker than 10 m). Slumps involve movement along one or more curved failure surfaces, with downward motion near the top and outward motion toward the bottom (Figure 15.14). They are typically caused by an excess of water within these materials on a steep slope.

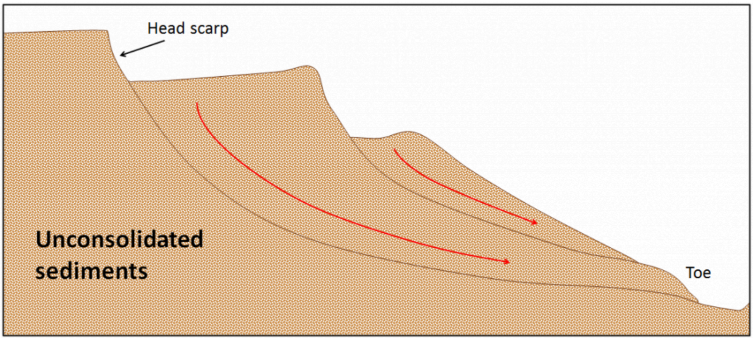


Figure 15.14 A depiction of the motion of unconsolidated sediments in an area of slumping [SE]

An example of a slump in the Lethbridge area of Alberta is shown in Figure 15.15. This feature has likely been active for many decades, and moves a little more whenever there are heavy spring rains and significant snowmelt runoff. The toe of the slump is failing because it has been eroded by the small stream at the bottom.

Figure 15.15 A slump along the banks of a small coulee near Lethbridge, Alberta. The main head-scarp is clearly visible at the top, and a second smaller one is visible about one-quarter of the way down. The toe of the slump is being eroded by the seasonal stream that created the coulee. [SE 2005]

Mudflows and Debris Flows

As you saw in Exercise 15.1, when a mass of sediment becomes completely saturated with water, the mass loses strength, to the extent that the grains are pushed apart, and it will flow, even on a gentle slope. This can happen during rapid spring snowmelt or heavy rains, and is also relatively common during volcanic eruptions because of the rapid melting of snow and ice. (A mudflow or debris flow on a volcano or during a volcanic eruption is a *lahar*.) If the material involved is primarily sand-sized or smaller, it is known as a mudflow, such as the one shown in Figure 15.16.

Figure 15.16 A slump (left) and an associated mudflow (centre) at the same location as Figure 15.15, near Lethbridge, Alberta. [SE]

If the material involved is gravel sized or larger, it is known as a debris flow. Because it takes more gravitational energy to move larger particles, a debris flow typically forms in an area with steeper slopes and more water than does a mudflow. In many cases, a debris flow takes place within a steep stream channel, and is triggered by the collapse of bank material into the stream. This creates a temporary dam, and then a major flow of water and debris when the dam breaks. This is the situation that led to the fatal debris flow at Johnsons Landing, B.C., in 2012. A typical west-coast debris flow is shown in Figure 15.17. This event took place in November 2006 in response to very heavy rainfall.

There was enough energy to move large boulders and to knock over large trees.

Figure 15.17 The lower part of debris flow within a steep stream channel near Buttle Lake, B.C., in November 2006. [SE]

Exercises

Exercise 15.2 Classifying Slope Failures

These four photos show some of the different types of slope failures described above. Try to identify each types and provide some criteria to support your choice. [SE]

7.3 Preventing, Delaying, Monitoring, and Mitigating Mass Wasting

As already noted, we cannot prevent mass wasting in the long term as it is a natural and ongoing process; however, in many situations there are actions that we can take to reduce or mitigate its damaging effects on people and infrastructure. Where we can neither delay nor mitigate mass wasting, we should consider moving out of the way.

Preventing and Delaying Mass Wasting

It is comforting to think that we can prevent some effects of mass wasting by mechanical means, such as the rock bolts in the road cut at Porteau Cove (Figure 15.10), or the drill holes used to drain water out of a slope, as was done at the Downie Slide (Figure 15.9), or the building of physical barriers, such as retaining walls. What we have to remember is that the works of humans are limited compared to the works of nature. The rock bolts in the road cut at Porteau Cove will slowly start to corrode after a few years, and within a few decades, many of them will begin to lose their strength. Unless they are replaced, they will no longer support that slope. Likewise, drainage holes at the Downie Slide will eventually become plugged with sediment and chemical precipitates, and unless they are periodically unplugged, their effectiveness will decrease. Eventually, unless new holes are drilled, the drainage will be so compromised that the slide will start to move again. This is why careful slope monitoring by geological and geotechnical engineers is important at these sites. The point here is that our efforts to “prevent” mass wasting are only as good as our resolve to maintain those preventive measures.

Delaying mass wasting is a worthy endeavour, of course, because during the time that the measures are still effective they can save lives and reduce damage to property and infrastructure. The other side of the coin is that we must be careful to avoid activities that could make mass wasting more likely. One of the most common anthropogenic causes of mass wasting is road construction, and this applies both to remote gravel roads built for forestry and mining and large urban and regional highways. Road construction is a potential problem for two reasons. First, creating a flat road surface on a slope inevitably involves creating a cut bank that is steeper than the original slope. This might also involve creating a filled bank that is both steeper and weaker than the original slope (Figure 15.18). Second, roadways typically cut across natural drainage features, and unless great care is taken to reroute the runoff water and prevent it from forming concentrated flows, oversaturating fill of materials can result. A specific example of the contribution of construction-related impeded drainage to slope instability was shown earlier in Figure 15.6.

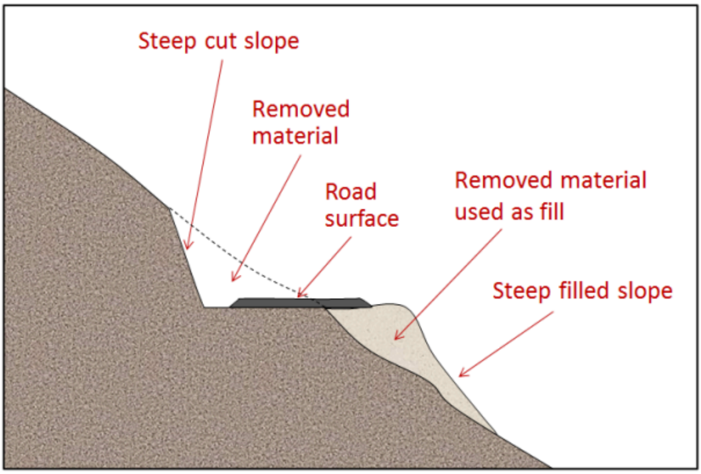


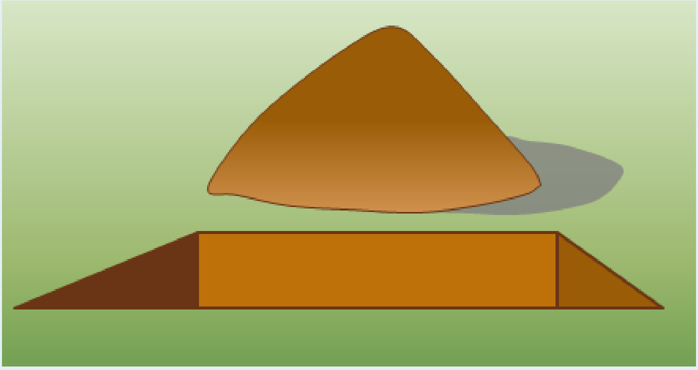
Figure 15.18 An example of a road constructed by cutting into a steep slope and the use of the cut material as fill. [SE]

Apart from water issues, engineers building roads and other infrastructure on bedrock slopes have to be acutely aware of the geology, and especially of any weaknesses or discontinuities in the rock related to bedding, fracturing, or foliation. If possible, situations like that at Porteau Cove (Figure 15.10) should be avoided — by building somewhere else — rather than trying to stitch the slope back together with rock bolts.

It is widely believed that construction of buildings on the tops of steep slopes can contribute to the instability of the slope. This is probably true, but not because of the weight of the building. As you'll see by completing Exercise 15.3, a typical house isn't usually heavier than the fill that was removed from the hole in the ground made to build it. A more likely contributor to instability of the slope around a building is the effect that it and the changes made to the surrounding area have on drainage.

Exercises

Exercise 15.3 How Much Does a House Weigh and Can It Contribute to a Slope Failure?



It is commonly believed that building a house (or some other building) at the top of a slope will add a lot of extra weight to the slope, which could contribute to slope failure. But what does a house actually weigh? A typical 150 m^2 (approximately 1,600 ft^2) wood-frame house with a basement and a concrete foundation weighs about 145 t (metric tonnes). But most houses are built on foundations that are excavated into the ground. This involves digging a hole and taking some material away, so we need to subtract what that excavated material weighs. Assuming our 150 m^2 house required an excavation that was 15 m by 11 m by 1 m deep, that's 165 m^3 of "dirt," which typically has a density of about 1.6 t per m^3.

Calculate the weight of the soil that was removed and compare that with the weight of the house and its foundation.

If you're thinking that building a bigger building is going to add more weight, consider that bigger buildings need bigger and deeper excavations, and in many cases the excavations will be into solid rock, which is much heavier than surficial materials.

You may wish to consider how a building might change the drainage on a slope. There are a number ways. Water can be collected by roofs, go into downspouts, and form concentrated flows that are directed onto or into the slope. Likewise drainage from nearby access roads, lawn irrigation, leaking pools, and septic systems can all alter the surface and groundwater flow in a slope.

Monitoring Mass Wasting

In some areas, it is necessary to establish warning systems so that we know if conditions have changed at a known slide area, or if a rapid failure, such as a debris flow, is actually on its way downslope. The Downie Slide above the Revelstoke Resevoir is monitored 24/7 with a range of devices, such as inclinometers (slope-change detectors), bore-hole motion sensors, and GPS survey instruments. A simple mechanical device for monitoring the nearby Checkerboard Slide (which is also above the Revelstoke Reservoir) is shown in Figure 15.19. Both of these are very slow-moving rock slides, but it’s very important to be able to detect changes in their rates of motion because at both of these locations a rapid failure would result in large bodies of rock plunging into the reservoir and sending a wall of water over the Revelstoke Dam, potentially destroying the nearby town of Revelstoke.

Figure 15.19 Part of a motion-monitoring device at the Checkerboard Slide near Revelstoke, B.C. The lower end of the cable is attached to a block of rock that is unstable. Any incremental motion of that block will move the cable, which will be detectable on this device. [SE]

Mt. Rainier, a glacier-covered volcano in Washington State, has the potential to produce massive mudflows or debris flows (lahars) with or without a volcanic eruption. Over 100,000 people in the Tacoma, Puyallup, and Sumner areas are in harm's way because they currently reside on deposits from past lahars (Figure 15.20). In 1998, a network of acoustic monitors was established around Mt.

Rainier. The monitors are embedded in the ground adjacent to expected lahar paths. They are intended to provide warnings to emergency officials, and when a lahar is detected, the residents of the area will have anywhere from 40 minutes to three hours to get to safe ground.

Figure 15.20 Mt. Rainier, Washington, from Tacoma. [By Lynn Topinka, US Geological Survey, http://en.wikipedia.org/wiki/Mount_ Rainier#/media/File:Mount_Rainier_over_Tacoma.jpg]

Mitigating the Impacts of Mass Wasting

In situations where we can't predict, prevent, or delay mass-wasting hazards, some effective measures can be taken to minimize the associated risk. For example, many highways in B.C. and western Alberta have avalanche shelters like that shown in Figure 15.21. In some parts of the world, similar features have been built to protect infrastructure from other types of mass wasting.

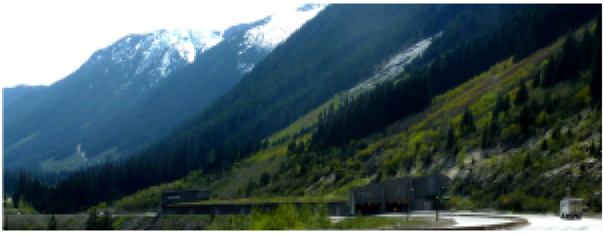


Figure 15.21 A snow avalanche shelter on the Coquihalla Highway. The expected path of the avalanche is the steep un-treed slope above. [SE]

Debris flows are inevitable, unpreventable, and unpredictable in many parts of B.C., but nowhere more so than along the Sea-to-Sky Highway between Horseshoe Bay and Squamish. The results have been deadly and expensive many times in the past. It would be very expensive to develop a new route in this region, so provincial authorities have taken steps to protect residents and traffic on the highway and the railway. Debris-flow defensive structures have been constructed in several drainage basins, as shown in Figure 15.22. One strategy is to allow the debris to flow quickly through to the ocean along a smooth channel. Another is to capture the debris within a constructed basin that allows the excess water to continue through, but catches the debris materials.

Figure 15.22 Two strategies for mitigating debris flows on the Sea-to-Sky Highway. Left: A concrete –lined channel on Alberta Creek allows debris to flow quickly through to the ocean. Right: A debris-flow catchment basin on Charles Creek. In 2010, a debris flow filled the basin to the level of the dotted white line. [SE]

Finally, in situations where we can't do anything to delay, predict, contain, or mitigate slope failures, we simply have to have the sense to stay away. There is a famous example of this in B.C. at a site known as Garibaldi, 25 km south of Whistler. In the early 1980s the village of Garibaldi had a population of about 100, with construction underway on some new homes, and plans for many more. In the months that followed the deadly 1980 eruption of Mt. St. Helens in Washington State, the B.C. Ministry of Transportation commissioned a geological study that revealed that a steep cliff known as The Barrier (Figure 15.23) had collapsed in 1855, leading to a large rock avalanche, and that it was likely to collapse again unpredictably, putting the village of Garibaldi at extreme risk. In an ensuing court case, it was ruled that the Garibaldi site was not a safe place for people to live. Those who already had homes there were compensated, and everyone else was ordered to leave.

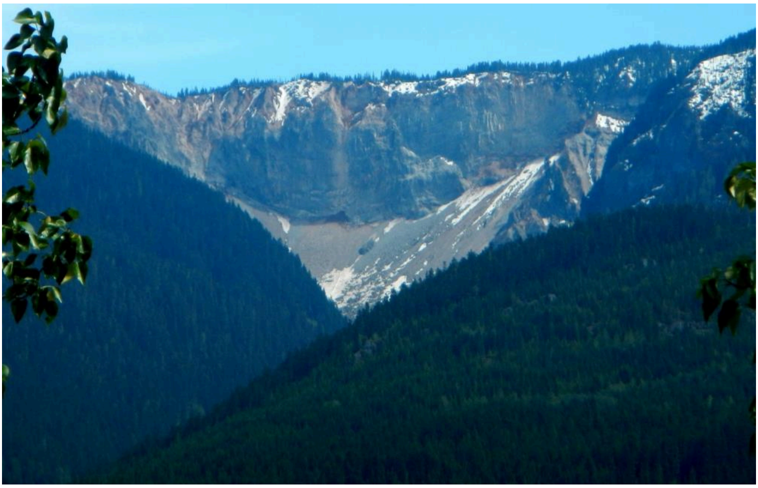


Figure 15.23 The Barrier, south of Whistler, B.C., was the site of a huge rock avalanche in 1855, which extended from the cliff visible here 4 km down the valley and across the current location of the Sea-to-Sky Highway and the Cheakamus River. [SE]

7.4 - Case Study: 1965 Hope Slide

A **landslide** is the general term that describes the movement of rock, soil and other debris down a slope as a result of gravitational pull. Landslides are most likely to occur in mountainous regions but may occur on any slope.

There are many factors that can contribute to the likelihood of a landslide, including geological causes such as weak or sensitive materials, morphological causes such as tectonic plate shifting or soil erosion, and human causes such as deforestation, water leakage, imposed vibrations or excavation.

One of the most tragic landslides in Canadian history was the Hope Slide, which occurred at 7 a.m. on January 9, 1965, in the Nicolum Valley in the Cascade Mountains near Hope, British Columbia.[1]

1. This Week in BC History video can be viewed at https://www.youtube.com/watch?v=Aigd-_En5z0

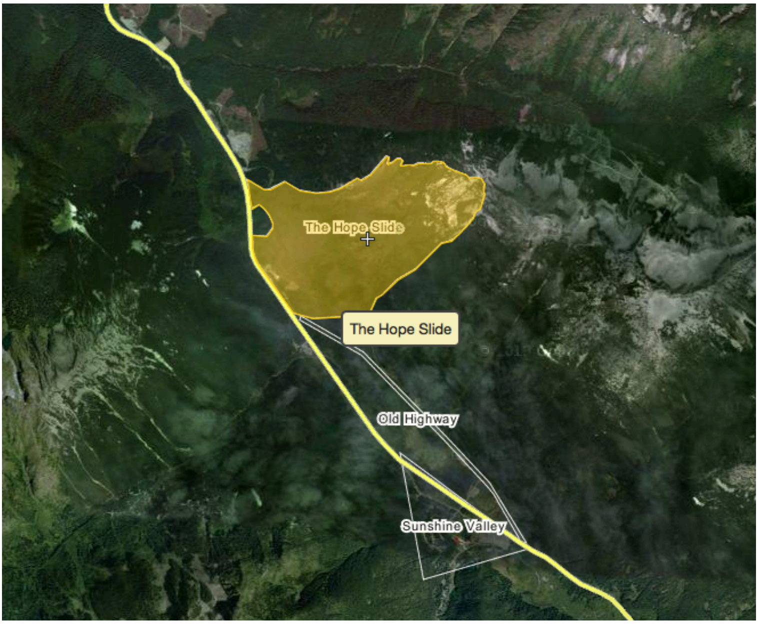


Figure 8.9 Location of Hope Slide

Shortly before the landslide, a small avalanche (different from a landslide in that an avalanche is made up of moving snow, not rock or sediment) occurred on a stretch of the Hope-Princeton Highway below Johnson Peak, approximately 150 kilometres from Vancouver, which forced four people to stop their vehicles. While they were waiting for the clearing crews, a second landslide occurred.

The landslide buried the vehicles and the occupants, and all four people died under 47 million cubic metres of pulverized rock, mud and debris that fell down the 2,000-metre mountainside. This mass of debris completely displaced the water and mud in Outram Lake with incredible force, throwing it against the opposite side of the valley, wiping all vegetation and trees down to the bare rock, then splashing it back up the original (now bare) slope before settling.

Recent research shows that these impacts against the opposite valley sides produced the seismic signatures interpreted as earthquakes.

Figure 8.10 Hope Slide; mountain scare is clearly visible from the side

The landslide was caused by the presence of pre-existing tectonic structures (faults and shear zones) within the southwestern slope of Johnson Ridge. The lower parts of the slide scar are underlain by **felsite sheets** (which may have failed first) while the upper parts of the slide scar are underlain by highly jointed **Paleozoic greenstone** beds. Ongoing weathering and tectonic activity weakened the slide mass to the point where it had reached limiting equilibrium.

Although we know the cause of the landslide, it remains unclear what the actual trigger was. A published analysis of seismograph records from January 9, 1965, suggested that the Hope Slide probably occurred as two rock avalanches separated by about 3 hours.

Changes in groundwater condition are often a trigger for landslides, though are not thought to have played a role in the Hope Slide as the slide occurred during a protracted period of sub-zero temperatures in the winter. Weather records for the 25 days prior to the slide indicate that the temperature was much colder than normal, with the average daily temperature not exceeding 0º C, and on most days it was below -10º C (Earle, 2006). Some have suggested that freezing of seepage exit points may have caused an increase in water pressure at the toe of the slide. It is also suggested that progressive long-term deformation of the slopes of the southwest flanks of Johnson Peak caused the stability of the slope to deteriorate. We can conclude that the Hope Slide was the catastrophic result of very long term non-episodic mountain slope deformation.

Since the landslide, the highway has been rerouted around and over the base of the slide's debris field 55 metres above the original ground level on the other side of valley. As a result of the slide, the mountainside remains bare rock, without significant growth of trees or other large vegetation. Visitors to the region can view the scar of the landslide along Highway 3 at a viewpoint.

Attributions

sa/2.0/ca/)

- **Figure 8.10** Hope Slide; mountain scare is clearly visible from the side. Hope Slide (http://en.wikipedia.org/wiki/Hope_Slide) by Fawcett5 (http://en.wikipedia.org/wiki/User:Fawcett5) is in the Public Domain.

Chapter 7: Summary

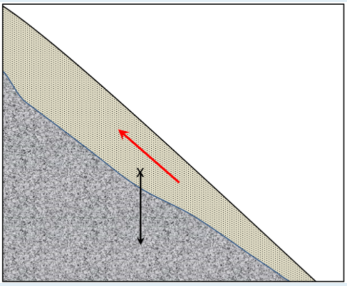
The topics covered in this chapter can be summarized as follows:

7.1	Factors That Control Stability on Slopes	Slope stability is controlled by the slope angle and the strength of the materials on the slope. Slopes are a product of tectonic uplift, and their strength is determined by the type of material on the slope and its water content. Rock strength varies widely and is determined by internal planes of weakness and their orientation with respect to the slope. In general, the more water, the greater the likelihood of failure. This is especially true for unconsolidated sediments, where excess water pushes the grains apart. Addition of water is the most common trigger of mass wasting, and can come from storms, rapid melting, or flooding.
7.2	Classification of Mass Wasting	The key criterion for classifying mass wasting is the nature of the movement that takes place. This may be a precipitous fall through the air, sliding as a solid mass along either a plane or a curved surface, or internal flow as a viscous fluid. The type of material that moves is also important — specifically whether it is solid rock or unconsolidated sediments. The important types of mass wasting are creep, slump, translational slide, rotational slide, fall, and debris flow or mudflow.

7.3	Preventing, Delaying, and Mitigating Mass Wasting	We cannot prevent mass wasting, but we can delay it through efforts to strengthen the materials on slopes. Strategies include adding mechanical devices such as rock bolts or ensuring that water can drain away. Such measures are never permanent, but may be effective for decades or even centuries. We can also avoid practices that make matters worse, such as cutting into steep slopes or impeding proper drainage. In some situations, the best approach is to mitigate the risks associated with mass wasting by constructing shelters or diversionary channels. And in other cases, where slope failure is inevitable, we should simply avoid building anything there.

Exercises

Questions for Review



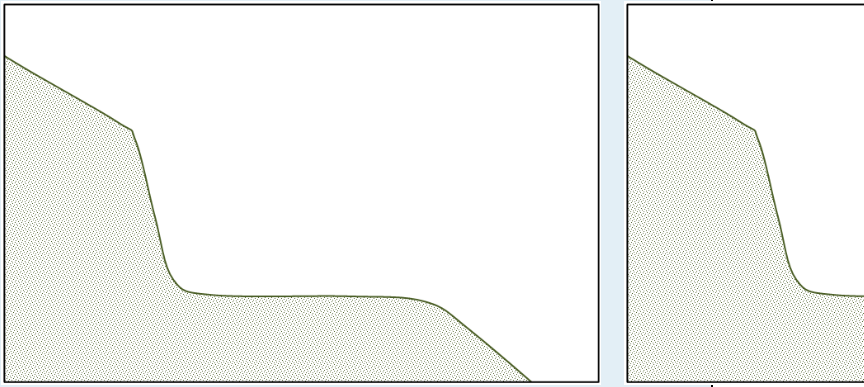
1. In the scenario shown here, the gravitational force on the unconsolidated sediment overlying the point marked with an X is depicted by the black arrow. Draw in the two arrows that show how this force can be resolved into the shear force (along the slope) and the normal force (into the slope).

2. The red arrow in the diagram depicts the shear strength of the sediment. Assuming that the relative lengths

of the shear force arrow (which you drew in question 1), and the shear strength arrow are indicative of the likelihood of failure, predict whether this material is likely to fail or not.

3. After several days of steady rain, the sediment becomes saturated with water and its strength is reduced by 25%. What are the likely implications for the stability of this slope?

4. In the diagrams shown here, a road cut is constructed in sedimentary rock with well-developed bedding. On the left, draw in the orientation of the bedding that would represent the greatest likelihood of slope failure. On the right, show the orientation that would represent the least likelihood of slope failure.



5. Explain why moist sand is typically stronger than either dry sand or saturated sand.

6. In the context of mass wasting, how does a flow differ from a slide?

7. If a large rock slide starts moving at a rate of several metres per second, what is likely to happen to the rock, and what would the resulting failure be called?

8. In what ways does a debris flow differ from a typical mudflow?

9. In the situation described in the chapter regarding lahar warnings at Mt. Rainier, the residents of the affected regions have to assume some responsibility and take precautions for their own safety. What sort of preparation should the residents make to ensure that they can respond appropriately when they hear lahar warnings?

10. What factors are likely to be important when considering the construction of a house near the crest of a slope that is underlain by glacial sediments?

Chapter 8: Avalanche

Read an excerpt from **chapter 7** of Keller et al. "Natural Hazards". You will have to log in with your SFU CANVAS credentials to view the link.

Keller, Edward A, Duane E DeVecchio, and John Clague. *Natural Hazards: Earth's Processes as Hazards, Disasters, and Catastrophes.*, 2010.

Avalanche Chapter (higher quality if you download and open as a .pdf rather than viewing in your browser window).

Read the article: GEODAR Data and the Flow Regimes of Snow Avalanches, available at the following link. You may have to sign-in using your SFU credentials in order to access the article.

Köhler, A., J. N. McElwaine, and B. Sovilla. "GEODAR Data and the Flow Regimes of Snow Avalanches" *Journal of Geophysical Research: Earth Surface*, June 14, 2018.

https://agupubs.onlinelibrary.wiley.com/doi/abs/10.1002/2017JF004375

Chapter 9: Tropical Storms/Flooding

9.1 The Hydrological Cycle

Water is constantly on the move. It is evaporated from the oceans, lakes, streams, the surface of the land, and plants (transpiration) by solar energy (Figure 13.2). It is moved through the atmosphere by winds and condenses to form clouds of water droplets or ice crystals. It comes back down as rain or snow and then flows through streams, into lakes, and eventually back to the oceans. Water on the surface and in streams and lakes infiltrates the ground to become groundwater. Groundwater slowly moves through the rock and surficial materials. Some groundwater returns to other streams and lakes, and some goes directly back to the oceans.

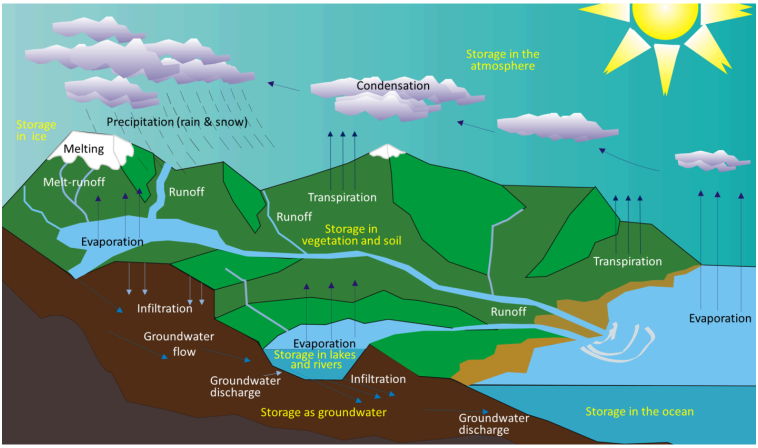


Figure 13.2 The various components of the water cycle. Black or white text indicates the movement or transfer of water from one reservoir to another. Yellow text indicates the storage of water. [SE after Wikipedia: http://upload.wikimedia.org/wikipedia/commons/5/54/Water_cycle_blank.svg]

Even while it's moving around, water is stored in various reservoirs. The largest, by far, is the oceans, accounting for 97% of the volume (Figure 13.3). Of course, that water is salty. The remaining 3% is fresh water. Two-thirds of our fresh water is stored in the ground and one-third is stored in ice. The remaining fresh water — about 0.03% of the total — is stored in lakes, streams, vegetation, and the atmosphere.

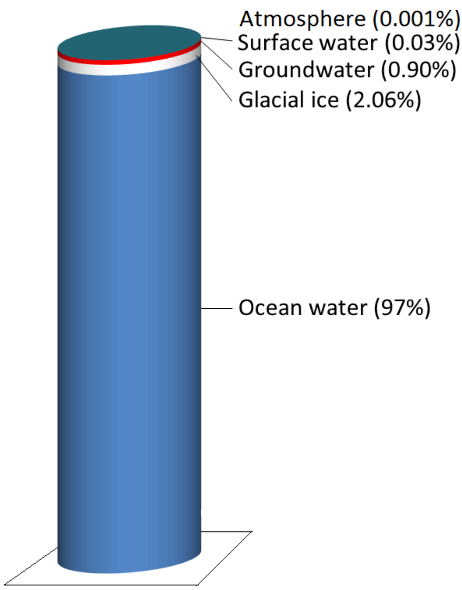


Figure 13.3a The storage reservoirs for water on Earth. Glacial ice is represented by the white band, groundwater the red band, and surface water the very thin blue band at the top. The 0.001% stored in the atmosphere is not shown. [SE using data from: https://water.usgs.gov/edu/watercyclefreshstorage.html]

To put that in perspective, let's think about putting all of Earth's water into a 1 L jug (Figure 13.3b). We start by almost filling the jug with 970 ml of water and 34 g of salt. Then we add one regular-sized (~20 mL) ice cube (representing glacial ice) and two teaspoons (~10 mL) of groundwater. All of the water that we see around us in lakes and streams and up in the sky can be represented by adding three more drops from an eyedropper.

Figure 13.3b Representation of the Earth's water as a 1 L jug. The three drops represent all of the fresh water in lakes, streams, and wetlands, plus all of the water in the atmosphere. [SE]

Although the proportion of Earth's water that is in the atmosphere is tiny, the actual volume is huge. At any given time, there is the equivalent of approximately 13,000 km^3 of water in the air in the form of water vapour and water droplets in clouds. Water is evaporated from the oceans, vegetation, and lakes at a rate of 1,580 km^3 per day, and just about exactly the same volume falls as rain and snow every day — over both the oceans and land. The precipitation that falls on land goes back to the ocean in the form of stream flow (117 km^3/day) and groundwater flow (6 km^3/day). Most of the rest of this chapter is about that 117 km^3/day of streamflow. The average discharge of the Fraser River into the ocean is approximately 0.31 km^3/day, or 0.26% of the world's total.

Exercises

Exercise 13.1 How Long Does Water Stay in the Atmosphere?

The residence time of a water molecule in the atmosphere (or any of the other reservoirs) can be estimated by dividing the amount that is there by the rate at which it is transferred in and out. For the atmosphere, we know that the reservoir size is 13,000 km^3, and the rate of flux is 1,580 km^3/day. If we divide 13,000 by 1,580, we get 8.22 days. This means that, *on average*, a molecule of water stays in the atmosphere for just over eight days. "Average" needs to be emphasized here because obviously some molecules stay in the air for only a few hours, while others may stay up there for weeks.

The volume of the oceans is 1,338,000,000 km^3 and the flux rate is approximately the same (1,580 km^3/day). What is the average residence time of a water molecule in the ocean?

9.2 - Introduction to Tropical Storms

Read sections 16.1-16.3 (pages 604-616) of the following textbook chapter:

Ch16-TropCycl-v102

9.3 Flooding

The discharge levels of streams are highly variable depending on the time of year and on specific variations in the weather from one year to the next. In Canada, most streams show discharge variability similar to that of the Stikine River in northwestern B.C., as illustrated in Figure 13.24. The Stikine River has its lowest discharge levels in the depths of winter when freezing conditions persist throughout most of its drainage basin. Discharge starts to rise slowly in May, and then rises dramatically through the late spring and early summer as a winter's worth of snow melts. For the year shown, the minimum discharge on the Stikine River was 56 m3/s in March, and the maximum was 37 times higher, 2,470 m3/s, in May.

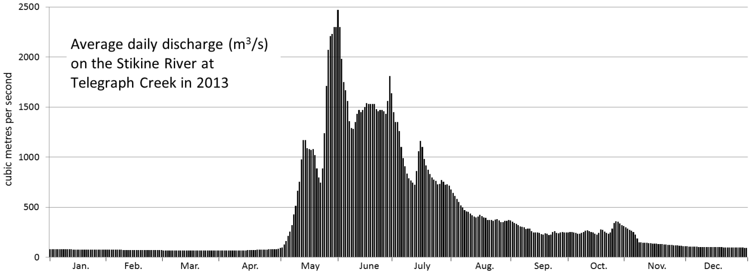


Figure 13.24 Variations in discharge of the Stikine River during 2013. [SE from data at Water Survey of Canada, Environment Canada, http://www.ec.gc.ca/rhc-wsc/]

Streams in coastal areas of southern British Columbia show a very different pattern from those in most of the rest of the country because their drainage basins do not remain entirely frozen and because they receive a lot of rain (rather than snow) during the winter. The Qualicum River on Vancouver Island typically has its highest discharge levels in January or February and its lowest levels in late summer (Figure 13.25). In 2013, the minimum discharge was 1.6 m3/s, in August, and the maximum was 34 times higher, 53 m3/s, in March.

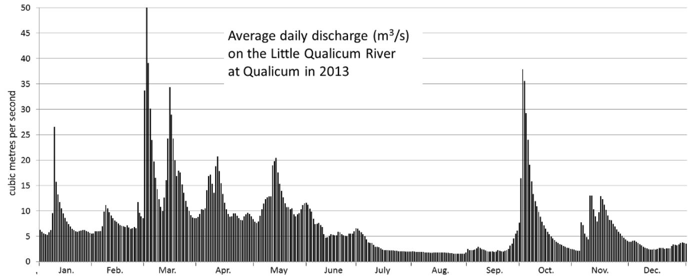


Figure 13.25 Variations in discharge of the Qualicum River during 2013. [SE from data at Water Survey of Canada, Environment Canada, http://www.ec.gc.ca/rhc-wsc/]

When a stream's discharge increases, both the water level (**stage**) and the velocity increase as well. Rapidly flowing streams become muddy and large volumes of sediment are transported both in suspension and along the stream bed. In extreme situations, the water level reaches the top of the stream's banks (the **bank-full stage**, see Figure 13.17), and if it rises any more, it floods the surrounding terrain. In the case of mature or old-age streams, this could include a vast area of relatively flat ground known as a flood plain, which is the area that is typically covered with water during a major flood. Because fine river sediments are deposited on flood plains, they are ideally suited for agriculture, and thus are typically occupied by farms and residences, and in many cases, by towns or cities. Such infrastructure is highly vulnerable to damage from flooding, and the people that live and work there are at risk.

Most streams in Canada have the greatest risk of flooding in the late spring and early summer when stream discharges rise in response to melting snow. In some cases, this is exacerbated by spring storms. In years when melting is especially fast and/or spring storms are particularly intense, flooding can be very severe.

One of the worst floods in Canadian history took place in the Fraser Valley in late May and early June of 1948. The early spring of that year had been cold, and a large snow pack in the interior was slow to melt. In mid-May, temperatures rose quickly and melting was accelerated by rainfall. Fraser River discharge levels rose rapidly over several days during late May, and the dykes built to protect the valley were breached in a dozen places. Approximately one-third of the flood plain was inundated and many homes and other buildings were destroyed, but there were no deaths. The Fraser River flood of 1948, which was the highest in the past century, was followed by very high river levels in 1950 and 1972 and by relatively high levels several times since then, the most recent being 2007 (Table 13.1). In the years following 1948, millions of dollars were spent repairing and raising the existing dykes and building new ones; since then damage from flooding in the Fraser Valley has been relatively limited.

Rank	Year	Month	Date	Stage (m)	Discharge (m3/s)
1	1948	May	31	11.0	15,200
2	1972	Jun	16	10.1	12,900
3	1950	Jun	20	9.9	12,500
4	1964	Jun	21	9.6	11,600
5	1997	Jun	5	9.5	11,300
6	1955	Jun	29	9.4	11,300
7	1999	Jun	22	9.4	11,000
8	2007	Jun	10	9.3	10,850
9	1974	Jun	22	9.3	10,800
10	2002	Jun	21	9.2	10,600

Table13.1 Ranking of the maximum stage and discharge values for the Fraser River at Hope between 1948 and 2008. Typical discharge levels are around 1,000 m3/s. [From date in Mannerstrom, 2008[1]]

Serious flooding happened in July in 1996 in the Saguenay-Lac St. Jean region of Quebec. In this case, the floods were caused by two weeks of heavy rainfall followed by one day of exceptional rainfall. July 19 saw 270 mm of rain, equivalent to the region's normal rainfall for the entire month of July. Ten deaths were attributed to the Saguenay floods, and the economic toll was estimated at $1.5 billion.

1. Mannerström, M, 2008, Comprehensive Review of Fraser River at Hope Flood Hydrology and Flows Scoping Study, Report prepared for the B.C. Ministry of the Environment. Available at: http://www.env.gov.bc.ca/wsd/public_safety/flood/pdfs_word/review_fraser_flood_flows_hope.pdf

Just a year after the Saguenay floods, the Red River in Minnesota, North Dakota, and Manitoba reached its highest level since 1826. As is typical for the Red River, the 1997 flooding was due to rapid snowmelt. Because of the south to north flow of the river, the flooding starts in Minnesota and North Dakota, where melting starts earlier, and builds toward the north. The residents of Manitoba had plenty of warning that the 1997 flood was coming because there was severe flooding at several locations on the U.S. side of the border.

After the 1950 Red River flood, the Manitoba government built a channel around the city of Winnipeg to reduce the potential of flooding in the city (Figure 13.26). Known as the Red River Floodway, the channel was completed in 1964 at a cost of $63 million. Since then it has been used many times to alleviate flooding in Winnipeg, and is estimated to have saved many billions of dollars in flood damage. The massive 1997 flood was almost too much for the floodway; in fact the amount of water diverted was greater than the designed capacity. The floodway has recently been expanded so that it can be used to divert more of the Red River’s flow away from Winnipeg.

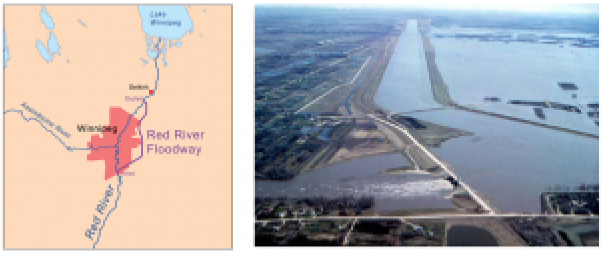


Figure 13.26 Map of the Red River Floodway around Winnipeg, Manitoba (left), and aerial view of the southern (inlet) end of the floodway (right). [Map from http://en.wikipedia.org/wiki/1997_Red_River_Flood#/media/File:Rednorthfloodwaymap.png and photo from Natural Resources Canada 2012, courtesy of the Geological Survey of Canada (Photo 2000-118 by G.R. Brooks).]

Canada’s most costly flood ever was the June 2013 flood in southern Alberta. The flooding was initiated by snowmelt and worsened by heavy rains in the Rockies due to an anomalous flow of moist air from the Pacific and the Caribbean. At Canmore, rainfall amounts exceeded 200 mm in 36 hours, and at High River, 325 mm of rain fell in 48 hours.

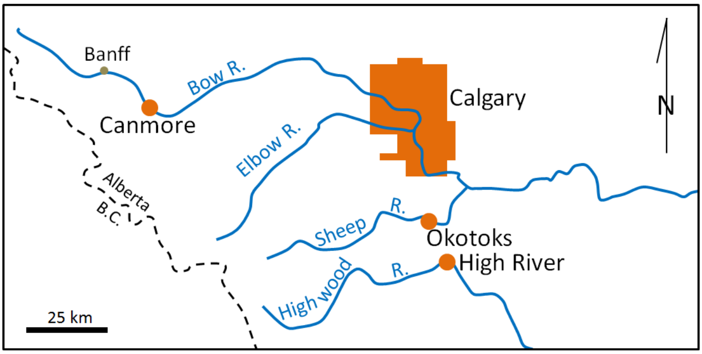


Figure 13.27 Map of the communities most affected by the 2013 Alberta floods (in orange) [SE]

In late June and early July, the discharges of several rivers in the area, including the Bow River in Banff, Canmore, and Exshaw, the Bow and Elbow Rivers in Calgary, the Sheep River in Okotoks, and the Highwood River in High River, reached levels that were 5 to 10 times higher than normal for the time of year (see Exercise 13.5). Large areas of Calgary, Okotoks, and High River were flooded and five people died (see Figures 13.27 and 13.28). The cost of the 2013 flood is estimated to be approximately $5 billion. For more about Alberta's flood of the century, visit: http://www.ec.gc.ca/meteo-weather/default.asp?lang=En&n=5BA5EAFC-1&offset=2&toc=hide.

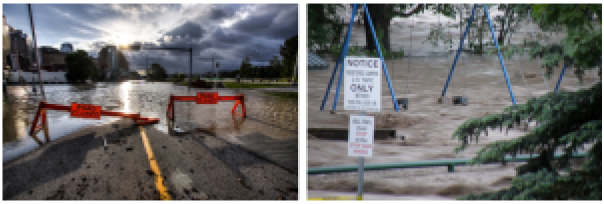


Figure 13.28 Flooding in Calgary (June 21, left) and Okotoks (June 20, right) during the 2013 southern Alberta flood [http://upload.wikimedia.org/wikipedia/commons/6/6a/Riverfront_Ave_Calgary_Flood_2013.jpg http://upload.wikimedia.org/wikipedia/en/9/9b/Okotoks_-_June_20%2C_2013_-_Flood_waters_in_local_campground_playground-03.JPG]

Exercises

Exercise 13.5 Flood Probability on the Bow River

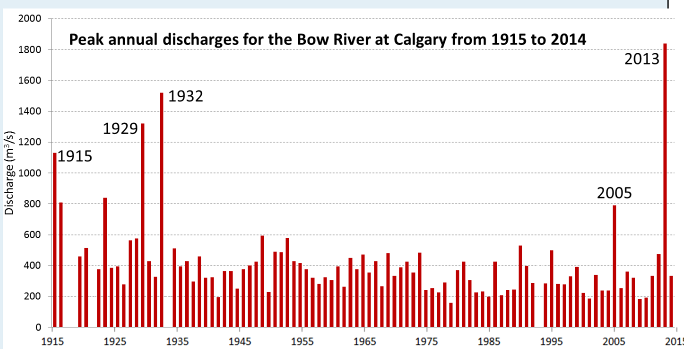
The graph below shows the highest discharge per year between 1915 and 2014 on the Bow River at Calgary. Using this data set, we can calculate the recurrence interval (Ri) for any particular flood magnitude using the equation: Ri = (n+1)/r (where n is the number of floods in the record being considered, and r is the rank of the particular flood). There are a few years missing in this record, and the actual number of data points is 95.

The largest flood recorded on the Bow River over that period was the one in 2013, 1,840 m^3/s on June 21. Ri for that flood is (95+1)/1 = 96 years. The probability of such a flood in any future year is 1/Ri, which is 1%. The fifth largest flood was just a few years earlier in 2005, at 791 m^3/s. Ri for that flood is (95+1)/5 = 19.2 years. The recurrence probability is 5%.

1. Calculate the recurrence interval for the second largest flood (1932, 1,520 m^3/s).

2. What is the probability that a flood of 1,520 m^3/s will happen next year?

3. Examine the 100-year trend for floods on the Bow River. If you ignore the major floods (the labelled ones), what is the general trend of peak discharges over that time?



[SE, from data at Water Surveys of Canada, Environment Canada, http://wateroffice.ec.gc.ca/search/searchDownload_e.html]

One of the things that the 2013 flood on the Bow River teaches us is that we can't predict when a flood will occur or how big it will be, so in order to minimize damage and casualties we need to be prepared. Some of the ways of doing that are as follows:

- Mapping flood plains and not building within them
- Building dykes or dams where necessary
- Monitoring the winter snowpack, the weather, and stream discharges
- Creating emergency plans

- Educating the public

9.4 Sea-Level Change

Sea-level change has been a feature on Earth for billions of years, and it has important implications for coastal processes and both erosional and depositional features. There are three main mechanisms of sea-level change, as described below.

Eustatic sea-level changes are global sea-level changes related either to changes in the volume of glacial ice on land or to changes in the shape of the sea floor caused by plate tectonic processes. For example, changes in the rate of mid-ocean spreading will change the shape of the sea floor near the ridges, and this affects sea level.

Over the past 20,000 years, there has been approximately 125 m of eustatic sea-level rise due to glacial melting. Most of that took place between 15,000 and 7,500 years ago during the major melting phase of the North American and Eurasian Ice Sheets (Figure 17.25). At around 7,500 years ago, the rate of glacial melting and sea-level rise decreased dramatically, and since that time, the average rate has been in the order of 0.7 mm/year. Anthropogenic climate change led to accelerating sea-level rise starting around 1870. Since that time, the average rate has been 1.1 mm/year, but it has been gradually increasing. Since 1992, the average rate has been 3.2 mm/year.

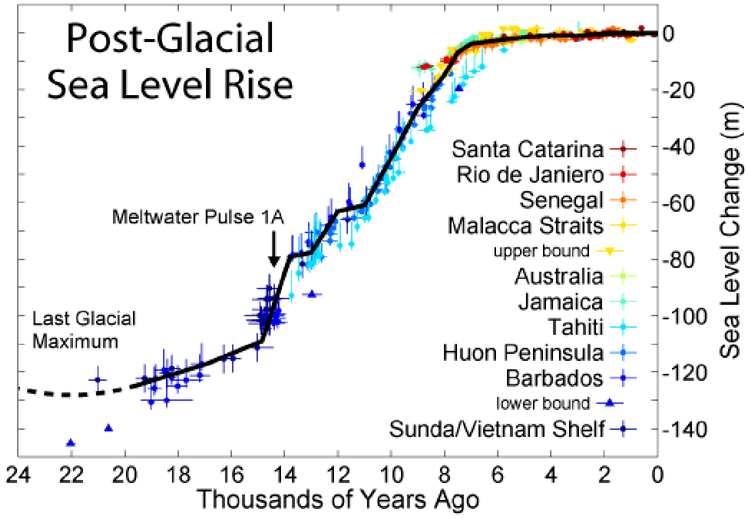


Figure 17.25 Eustatic sea-level curve for the past 24 ka (sea-level rise resulting from the melting of glacial ice). Sea-level rise is global; the locations listed in the caption are the places where data were acquired to create this diagram. [https://en.wikipedia.org/wiki/Sea_level_rise#/media/File:Post-Glacial_Sea_Level.png]

Isostatic sea-level changes are local changes caused by subsidence or uplift of the crust related either to changes in the amount of ice on the land, or to growth or erosion of mountains.

Almost all of Canada and parts of the northern United States were covered in thick ice sheets at the peak of the last glaciation. Following the melting of this ice, there has been an isostatic rebound of continental crust in many areas. This ranges from several hundred metres of rebound in the central part of the Laurentide Ice Sheet (around Hudson Bay) to 100 m to 200 m in the peripheral parts of the Laurentide and Cordilleran Ice Sheets — in places such as Vancouver Island and the mainland coast of B.C. In other words, although global sea level was about 130 m lower during the last glaciation, the glaciated regions were depressed at least that much in most places, and more than that in places where the ice was thickest.

There is evidence of isostatic rebound along the southwest coast of Vancouver Island, where a number of streams enter the ocean as 5 m high waterfalls, as shown in Figure 17.26.

Figure 17.26 This stream is on the southwest coast of Vancouver Island near Sooke. Like many other streams along this coast, it used to flow directly into the ocean, but the land has been uplifted by post-glacial isostatic rebound. [SE]

Tectonic sea-level changes are local changes caused by tectonic processes. The subduction of the Juan de Fuca Plate beneath British Columbia is creating tectonic uplift (about 1 mm/year) along the western edge of Vancouver Island, although much of this uplift is likely to be reversed when the next large subduction-zone earthquake strikes.

Coastlines in areas where there has been net sea-level rise in the geologically recent past are commonly characterized by estuaries and fiords. Howe Sound, north of Vancouver, is an example of a fiord (Figure 17.27). This valley was filled with ice during the last glaciation, and there has been a net rise in sea level here since that time. Coastlines in areas where there has been net sea-level drop in the geologically recent past are characterized by uplifted wave-cut platforms (or stream valleys as shown in Figure 17.26). Uplifted beach lines are another product of relative sea-level drop, although these are difficult to recognize in areas with vigorous vegetation. They are relatively common in Canada's far north.

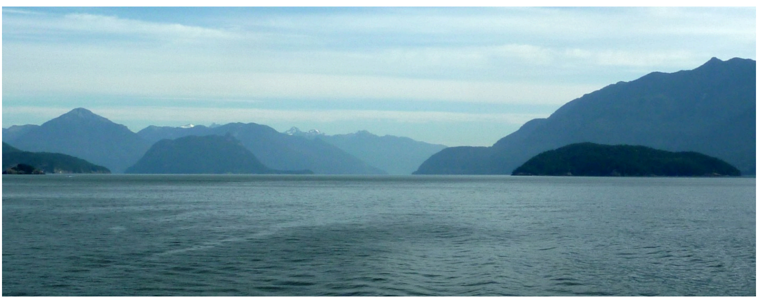


Figure 17.27 Howe Sound, north of Vancouver, is a fiord with well-defined glacial erosion features. [SE]

Exercises

Exercise 17.4 A Holocene Uplifted Shore

The blue-grey sediments in the photo contain marine fossils of early Holocene age (~12,500 years ago). Explain the role that melting ice might have played in raising this material 60 m above sea level. Is this change a result of isostatic or eustatic relative sea-level change?

Early Holocene marine sediments at 60 m elevation on Gabriola Island, B.C. [SE]

Chapter 10: Wildfire

Read the following article on adaptation to changes in wildfire regime for Western North America.

Schoennagel, Tania, Jennifer K. Balch, Hannah Brenkert-Smith, Philip E. Dennison, Brian J. Harvey, Meg A. Krawchuk, Nathan Mietkiewicz, et al. “Adapt to More Wildfire in Western North American Forests as Climate Changes.” *Proceedings of the National Academy of Sciences* 114, no. 18 (May 2, 2017): 4582–90. https://doi.org/10.1073/pnas.1617464114.

http://www.pnas.org/content/114/18/4582.full

10.1 Case Study: 2003 Okanagan Mountain Park Fire

On August 16, 2003, a wildland fire was started near Rattlesnake Island in Okanagan Mountain Provincial Park in British Columbia. The one-in-a-hundred-year inferno began with a lightning strike and was fuelled by a constant wind and one of the driest summers in a decade.

If you are reading this in print, you can view the following video at: https://www.youtube.com/watch?v=AsPG8zB-V1E

The fire grew northward and eastward, initially threatening a small number of lakeshore homes, but quickly became an **interface zone** fire as it approached the city of Kelowna. An **interface zone** is an area where urban development occurs on or near lands that have a significant fire risk.

Figure 8.12 Satellite image in natural colours of the fire location and the surrounding area. In white is the smoke from the fire, in red inside the white area are fire hotspots of intense heat.

The fire eventually grew to cover over 250 square kilometres and forced the evacuation of 27,000 residents. The firestorm challenged personnel from over 60 fire departments across BC: over 1,000 forestry fire fighters, contractors and loggers as well as 1,400 members of the Canadian armed forces.

Tourism and the local wine industry were heavily impacted by the fire. August is the peak tourist season in the Okanagan region and tourism numbers are estimated to have declined 30% to 50% below normal during the period of the fire. A popular tourist attraction for extreme mountain biking, the Crawford area trails were closed indefinitely. The Kelowna Crags, a rock-climbing area within the fire zone was compromised by the heat of the fires, and six marine campgrounds accessible only by boat on Okanagan Lake, were destroyed. The St. Hubertus Estate Winery, originally planted in 1928, was destroyed along with the the owner's home and the entire 2003 crop. Although St. Hubertus was the only winery damaged by fire, many others in the region also lost their crop that year due to smoke taint.

In the end, 239 buildings were destroyed. The fire also destroyed 12 historic wooden railway trestles and damaged two other steel trestles in the Myra Canyon**.** These trestles were part of the Myra Canyon Section of the Kettle Valley Railway, in use from 1914 to 1978, and were considered a historic engineering feat. The section was declared a national historic site of Canada in 2002.

Attributions

- **Figure 8.12** Satellite image in natural colours of the fire location and the surrounding area. 2003 Okanagan Mountain Park Fire (en.wikipedia.org/wiki/2003_Okanagan_Mountain_Park_Fire) by NASA/GSFC/METI/Japan Space Systems, and U.S./Japan ASTER Science Team (http://asterweb.jpl.nasa.gov/gallery-detail.asp?name=okanagan) is in the Public Domain.