2 Earthquakes

2.1 Earthquake magnitude

Almost all earthquakes, and certainly all major earthquakes, are related to the slow but steady motion of the earth's plates.

Plate boundaries are locked

Although the plates are constantly in motion (at rates of a few cm per year, or approximately 0.1 mm per day) the rocks present at the places where plates meet (eg. at convergent boundaries and transform faults) are not necessarily sliding past each other. In most cases, particularly within a few tens of kilometres of surface, the friction between rocks at a boundary is great enough so that the two plates will be locked together. As the plates themselves gradually move, the rocks close to the boundary are elastically deformed. According the **elastic rebound theory**, energy is being stored in the rocks as they are deformed. Eventually the strain will become so great that the friction that is preventing movement between the plates will be overcome. The plates will suddenly slide past each other, and the stored energy will be released, producing an earthquake [Keller: Figure 7.18].

Energy released as body waves

A huge amount of energy radiates away from the focus of the earthquake in the form of deformation waves in the rock. These include **P-waves** (primary or compression waves) and **S-waves** (secondary or shear waves) - both of which propagate within the rock and are also known as **body waves.** When the body waves reach surface some of their energy will be transferred into surface waves, including **Love waves** (horizontal motion) and **Rayleigh waves** (vertical motion), which propagate along the interface between the rock and the surface [Keller: Figure 7.8].

Surface waves

Stress transfer

Once the first displacement takes place along the earthquake fault much of the stress which had been built up will be released, and much of the rock in the area will return back to an unstressed state. On the other hand, it has been shown recently that while an earthquake will release stress in some areas, it is likely to add to the stress in other areas. In the seconds, minutes, hours, days, weeks, years and even decades following the initial earthquake some of that added stress will be released in the form of aftershocks, both along the same fault plane, and along other fault planes in the region. (see below under *Stress transfer and aftershocks*)

Where earthquakes happen

A majority of the world's earthquakes take place along the transform faults associated with spreading ridges, but most of these are small because the rocks in these areas are relatively warm, and almost all are insignificant because they occur well away from populated areas. Some large earthquakes (> M 7) occur along larger transform boundaries - such as the San Andreas and Queen Charlotte Faults, the Anatolian Fault in Turkey or the Alpine Fault in New Zealand - however most very large earthquakes (> M 8) are generated at convergent boundaries within subduction zones.

The severity of earthquakes generated in any area will depend partly on the degree to which the two plates are locked together in different parts of the fault zone. The degree of locking will depend on several factors, including the rock types, the level of water saturation along the boundary, and, most importantly, the temperature.

Factors affecting the degree of locking

- The rock type is important for two reasons. The presence of weak or poorly consolidated rocks will prevent a fault zone from becoming locked - or at least it will reduce the extent of the locking. The presence of certain minerals, particularly clays or graphite, will lubricate a fault zone and also reduce the locking tendency.

Rock type

- Water will lubricate a fault zone, and thus will reduce the tendency for locking to take place.

Water

- Temperature, which increases systematically with depth, can affect the tendency for locking. As the temperature increases beyond 450º C most rocks begin to behave plastically and the locking tendency is reduced dramatically.

Temperature

The actual location of an earthquake within the earth, is called its "focus" (a.k.a. hypocenter) [Keller: Figure 7.8]. The point on the surface directly above the focus is known as the "epicentre". The location of an earthquake is determined from seismic records, specifically by measuring the time interval between the first P wave and the first S wave [Keller: Figure 7.16]. By comparing distances from several seismic stations it is possible to determine the location of the epicentre. If enough accurate data are available it is also possible to determine the depth, or the location of the focus. Most earthquakes occur within the range of 5 to 100 km depth, and all very strong earthquakes originate within this range. The maximum depth for earthquakes is 700 km because at greater depths the rocks are too hot and plastic to behave in a brittle manner.

Focus and epicentre

Earthquake magnitude - which is an indication of the amount of energy released by an earthquake - is expressed on a number of different scales, all of which provide slightly different types of information. Some of the methods for calculating magnitudes are summarized on the table below.

Magnitude

The first widely applied magnitude scale was the **Richter Scale** (after the American seismologist Charles F. Richter). Richter magnitude (M_L) (the L refers to *local* because this technique is only applicable for quakes within a few hundred km of the seismic station) is determined from the maximum amplitude of the S waves recorded on a **Wood-Anderson** seismograph, and is corrected for the distance between the epicentre and the seismic station. There

Richter magnitude

are very few operating Wood-Anderson seismographs today, but seismologists have adapted the algorithm so that comparable results can be obtained from other seismometers.

Magnitude type	**Applicable magnitude range**	**Applicable distance range**	**Comments**
Local or Richter (M_L)	2 to 6	0-400 km	The original magnitude relationship defined in 1935 by Richter and Gutenberg for Californian earthquakes. It is based on the maximum amplitude of S waves recorded on a Wood-Anderson torsion seismograph. Although these instruments are no longer widely in use, M_L values can be calculated using data from modern instruments.
Moment (M_W)	> 3.5	All	Based on the **seismic moment** of the earthquake, which is equal to the strength of the rock times the average amount of slip on the fault times the fault area that slipped.
Surface wave (M_S)	5 to 8	20 to 180°	A magnitude for distant earthquakes based on the amplitude of Rayleigh surface waves measured at a period near 20 sec.
Body (M_B)	4 to 7	16 to 100° (only deep quakes)	Based on the amplitude of P body-waves. This scale is most appropriate for deep-focus earthquakes.
Duration (M_D)	<4	0-400 km	Based on the duration of shaking as measured by the time decay of the amplitude of the seismogram. Commonly used to compute magnitude from seismograms with "clipped" waveforms due to limited dynamic recording range of analog instrumentation.
Tsunami (M_T)	large		Only applicable to large subduction-zone earthquakes which produce under-sea displacement. Based on wave height.

Partly derived from the USGS glossary of terms for earthquakes.

Summary of methods used for calculation of earthquake magnitude

(Partly derived from the USGS glossary of terms for earthquakes)

Moment magnitude

Moment magnitude is an expression of the energy released by an earthquake based on the size (area) of the rupture surface, the average displacement or slip distance, and the strength of the rock that broke. Of these values, the area of the rupture surface is the most variable[1], and hence the most significant to the final result. In fact, because the strength of the rocks being broken doesn't vary all that much, seismologists normally substitute a constant for this variable, a value of 3×10^{11} dyne/cm^2. The first part of the calculation of moment magnitude involves estimation of what is known as the **seismic moment**, as follows:

$$M_O = GLWu$$

where G is the rock strength value (3×10^{11} dyne/cm^2), L and W are the length and width of the fault rupture surface (in cm), and u is the average amount of displacement (in cm)[2]. Using some typical values, it is easy to see that seismic moments are very big numbers. For example, consider an earthquake for which the rupture surface is **10 km long** and **2 km wide** (ie. *deep* for a vertical fault), and if the average displacement is **50 cm**, the seismic moment value is as follows:

$$M_O = (3 \times 10^{11} \text{ dyne/cm}^2) \times (\mathbf{10 \times 10^5 \text{ cm}}) \times (\mathbf{2 \times 10^5 \text{ cm}}) \times (\mathbf{50 \text{ cm}}) = 3.0 \times 10^{24} \text{ dyne}\cdot\text{cm}$$

The moment magnitude is derived from the seismic moment using the following equation:

$$M_W = 2/3 \times (\log_{10}(M_O)) - 10.73$$

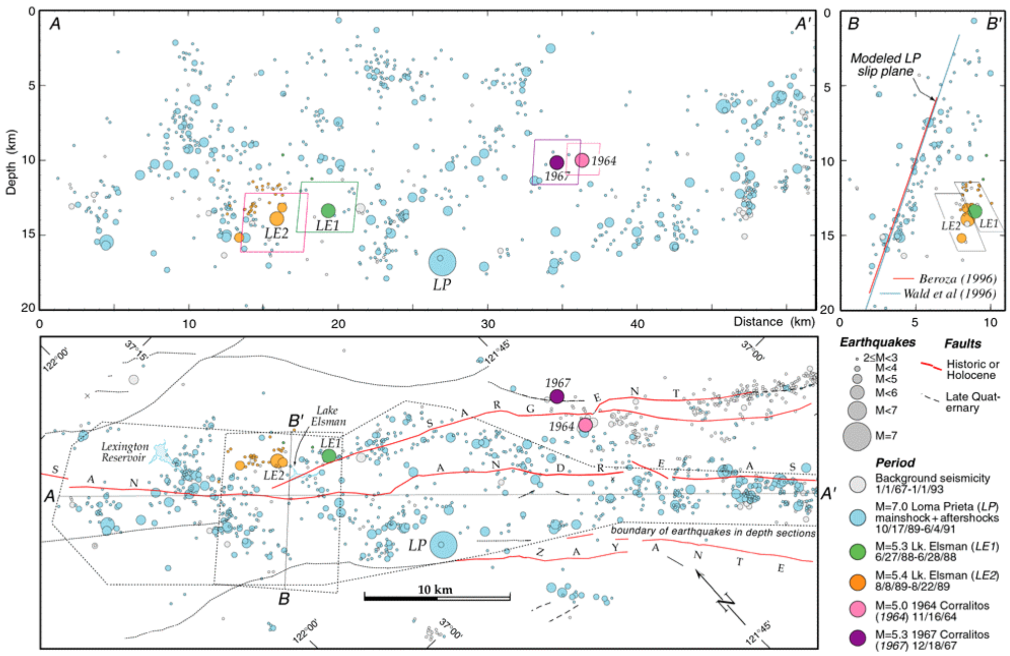
For the earthquake described above the moment magnitude is **5.59**.

Rupture zone

Seismologists acquire the information needed for moment magnitude calculations from seismic records for the main shock and the immediate aftershocks of an earthquake. In the case of a large earthquake there may be thousands of aftershocks, and these would be distributed around the area of the rupture surface. By carefully plotting the locations all of the events it is possible to estimate the length and width of the rupture zone. An example of the extent of the rupture surface for the 1989 Loma Prieta earthquake is shown on Figure 2.1. The length of the fault plane defined by the blue dots is around 50 km and the width (depth) is close to 20 km. The displacement was around 2 m and hence M_W=7.12.

[1] The length of a rupture zone may vary from a few tens metres to hundreds of kilometres, and the width from a few metres to over 100 kilometres. The displacement distance may vary from centimetres to tens of metres

[2] The amount of slip will vary quite widely in different parts of the rupture surface. It will be greatest near to the centre of the surface, and least near to the edges.



Distribution of shocks (blue circles) which define the rupture surface of the 1989 Loma Prieta ("World Series") earthquake

(The upper-left panel is a section along the fault, the upper right panel is a cross-section, the lower panel is a plan view.)

Estimation of displacement

If the rupture plane of the earthquake reached surface, and the displacement can be measured, this information can be used to help estimate the average slip distance. Other information, such as the amounts of subsidence and uplift in different places can also be used to help with this estimate. In many cases there is no surface manifestation of the earthquake, and the amount of slip can only be guessed, or estimated from other information, such as the rate of plate motion and the time interval since the last earthquake.

Moment magnitude is a much more accurate and relevant way of expressing earthquake energy than Richter magnitude - especially for large earthquakes - and it is being used increasingly by seismologists. When an earthquake occurs the Richter magnitude can be calculated within minutes or hours, and this is the number that is quickly released to the press, and hence is the number that sticks in people's minds. It can take days or even weeks to assemble enough data to calculate the moment magnitude.

Other magnitude estimations

Some other ways of expressing earthquake energy, as summarized in the table above, are as follows:

Surface-wave magnitude (Ms): based on the amplitude of surface waves and the distance to the epicentre (only applicable to earthquakes at least 20 (approx. 1600 km) away from the seismic station)[3].

[3] Surface-wave magnitude can also be used to estimate moment magnitude where the parameters of the rupture surface are not known.

Tsunami magnitude (Mt): based on the size of tsunami waves recorded at various locations (applicable only to subduction-zone earthquakes where there is underwater vertical displacement).

Earthquke incidence

Each year there are about 20 earthquakes over magnitude 7, and more than 100 over magnitude 6. Earthquakes over magnitude 8 are much less frequent, and those over 9 are typically decades apart. An earthquake over magnitude 3 can only be felt in the area immediately around the epicentre, and there are tens of thousands of these each year. The largest earthquakes occur along subduction zones in areas of converging plates in places like Chile, Ecuador, Alaska, Japan and Indonesia. Earthquakes over magnitude 8.5 have been measured in all of these areas.

Magnitude scales are logarithmic

All magnitude scales are logarithmic. A difference of 1 unit is equivalent a 10-fold difference in the amplitude of the waves on a seismograph, but to a 32-fold difference in the amount of energy released. The recent (Dec. 2004) M9 earthquake in Indonesia released the equivalent amount of energy as 32 M8 earthquakes and over 1000 M7 earthquakes. In other words, this earthquake released as much energy as all other earthquakes combined over approximately the past decade.

2.2 Earthquake intensity, harmonic amplification and liquefaction

Earthquake intensity

Earthquake intensity, or the amount of damage likely to be caused by an earthquake, is measured on the Mercalli Scale, [Keller: Table 7.2] which was developed in the late 1800's, and later modified by Giuseppe Mercalli. The scale ranges from I to XII, and values are assigned based on interviews with witnesses and assessment of damage to structures. The intensity assigned to the area immediately around the epicentre is normally much higher than that assigned to areas tens or hundreds of kilometres away. Events with intensities up to around III won't even be noticed by most people. Where the intensity is around VI there will be minor damage to buildings. If the intensity is X or more many buildings will be destroyed, and there will be serious damage to dams and bridges and triggering of landslides.

Regional variations in intensity

Variations in intensity with respect to distance from the epicentre will be affected by regional geology. In Eastern North America, for example, where the underlying rocks are generally old, hard and strong, the effects of an earthquake are felt over a much wider area than in western North America, where the rocks are generally younger, softer, weaker and more variable. Variations in Mercalli intensity around the 1994 Northridge (California) earthquake are shown on Keller's Figure 7.17.

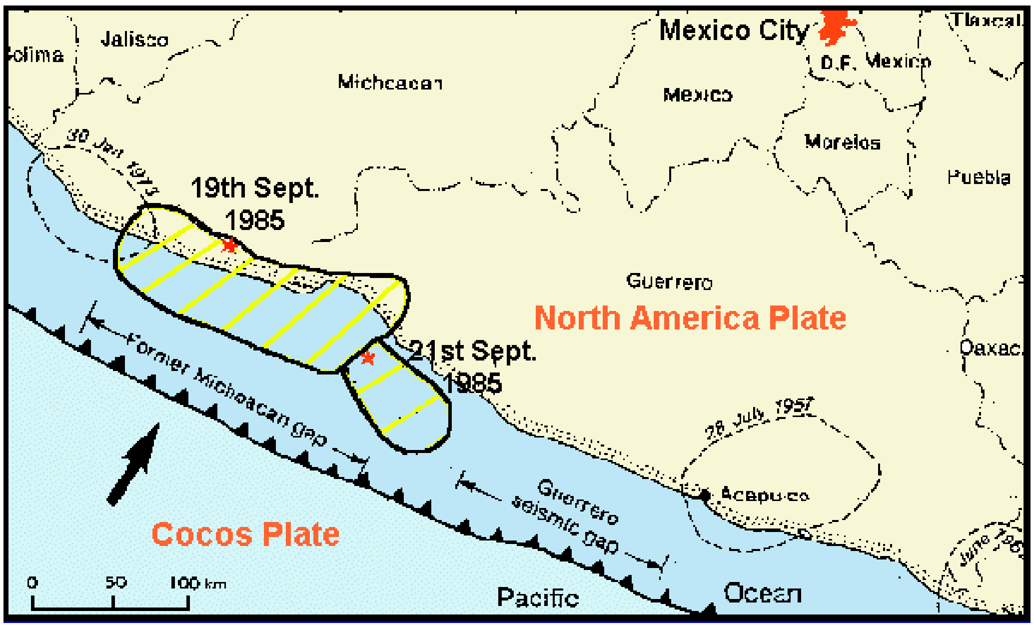
Damage

In fact the amount of damage caused by an earthquake will normally be more closely related to the type of geological material on which buildings are built than to the magnitude of the earthquake, or the Mercalli intensity felt in the general area. Buildings on solid rock will generally be damaged less severely than those on loose rock or unconsolidated sediments. Particularly severe damage can occur in areas where sediments are saturated with water.

related to foundation type

The M 8.1 Michoacan earthquake that struck Mexico in 1985 was situated very close to the coast, along the subduction zone where the Cocos Plate descends beneath the North American Plate. Several coastal cities situated within 100 km of the epicentre suffered relatively little damage, but Mexico City, which is approximately 400 km from the epicentre, suffered severe damage, including the destruction of around 1000 buildings, and the loss of 10,000 lives.

1985 Michoacan earthquake



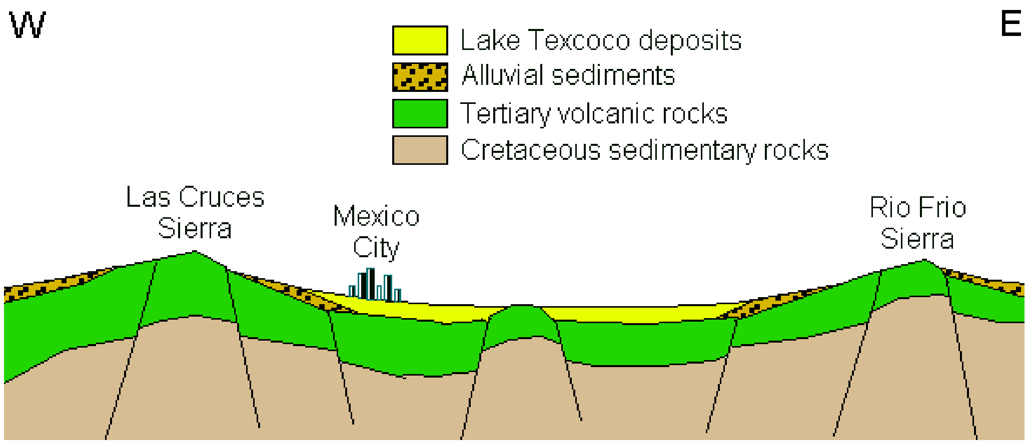
Epicentre locations (stars) and rupture surfaces of the 19th & 21st Sept. 1985 Michoacan earthquakes

Mexico City is built on largely unconsolidated 100,000 year old lake sediments which overly a series of volcanic rocks (see figure below). Seismic waves of a range of different frequencies were generated by the quake, but only those with relatively long-wave vibrations (with periods of 1 to 2 seconds) had enough energy (and were not absorbed by other geological materials) to make it all the way to Mexico City[4]. When these waves struck the lake sediments the amplitude of their motion increased by several times because the body of sediments beneath the city has a natural period of vibration of close to 2 seconds. The ground beneath the city started swaying back and forth - through a distance of about 40 cm - every two seconds, for nearly two minutes. This amount of shaking was too much for many of the

Mexico City lake sediments

[4] Seismic waves moving through rock behave in the same sort of way as sound waves in air. The low “beat” notes produced by a rock band take a lot of energy to generate, but they travel a long way. If you are listening to rock music from a significant distance you may only hear those low notes.

poorly-designed and poorly constructed buildings, and they just collapsed. Interestingly, it was the buildings between 5 and 15 stories in height that were affected most severely, because they also have a natural resonance of around 2 seconds [Keller: Figure 7.11].



Schematic geological section across Mexico City showing the lake sediments on which much of the city is constructed

Another significant factor in the damage of buildings in Mexico City was the **liquefaction** of the lake sediments.

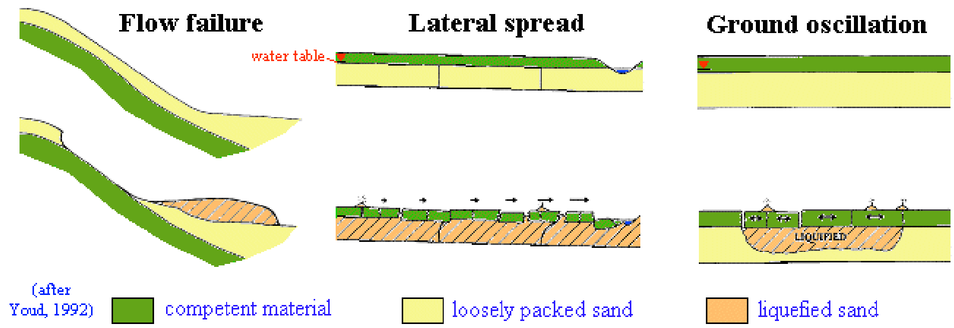
Liquefaction

Liquefaction is a phenomenon which occurs in loose, water-saturated sandy sediment. Under shaking conditions, commonly resulting from an earthquake, the sand grains are induced to move into a more closely packed configuration. This causes the water pressure to increase, which has the affect that the body of sand is no longer supported by grain-to-grain contact, but instead is supported by water. Forces between grains are dramatically reduced, and the body of sand loses strength and behaves as a thick liquid rather than a wet solid.

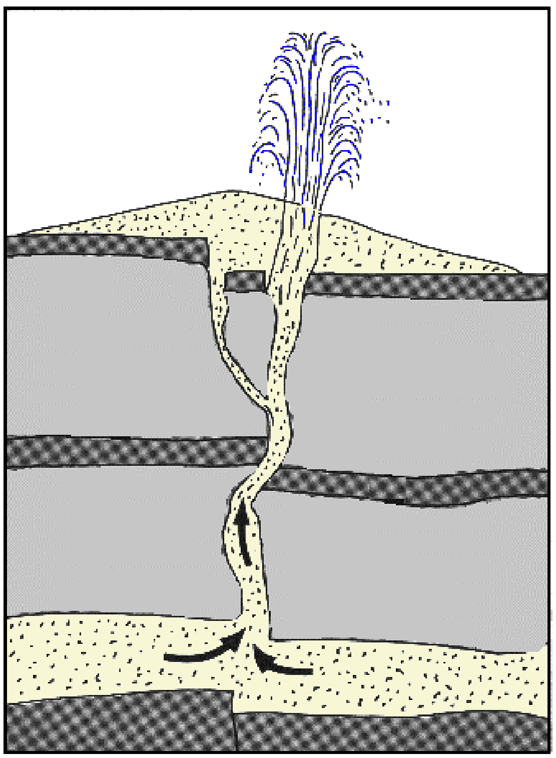
Liquefaction can have numerous consequences as illustrated in the figure below. Some of these include:

- the complete failure of a slope in what is known as a "flow failure".
- lateral spread of a competent layer above the layer which loses strength,
- liquefaction of a confined volume, and displacement of the overlying competent material, and

loss of bearing strength, which leads either to equal or to differential settling.



Various effects of the liquefaction of wet sandy sediments



Depiction of a sand-blow caused by the liquefaction of sand overlain by clay

Liquefaction is also commonly associated with sand dykes and sand "blows" as shown in figure to the left.

Liquefaction and the 1989 Loma Prieta earthquake

Liquefaction contributed significantly to the destruction of buildings and loss of life in Mexico City in 1985, and also in the San Francisco area from the Loma Prieta earthquake in 1989. Areas immediately around San Francisco Bay are underlain by unconsolidated fine-grained sediments and "Bay Fill" - material that was dumped into the bay. (Ironically, part of the Bay Fill was derived from the clean-up operation from the devastating 1906 earthquake and fire). Most of the rest of San Francisco and Oakland are built on older alluvial sediments and bedrock. The worst damage from the 1989 earthquake was in the Marina District (adjacent to the Golden Gate Bridge) which is underlain by bay-fill and mud; and also in the shore area of Oakland, where those sections of a double-deck freeway which collapsed were built on bay-fill and mud. [Keller: pages 174 and 175, Figures 7.12 to 7.14].

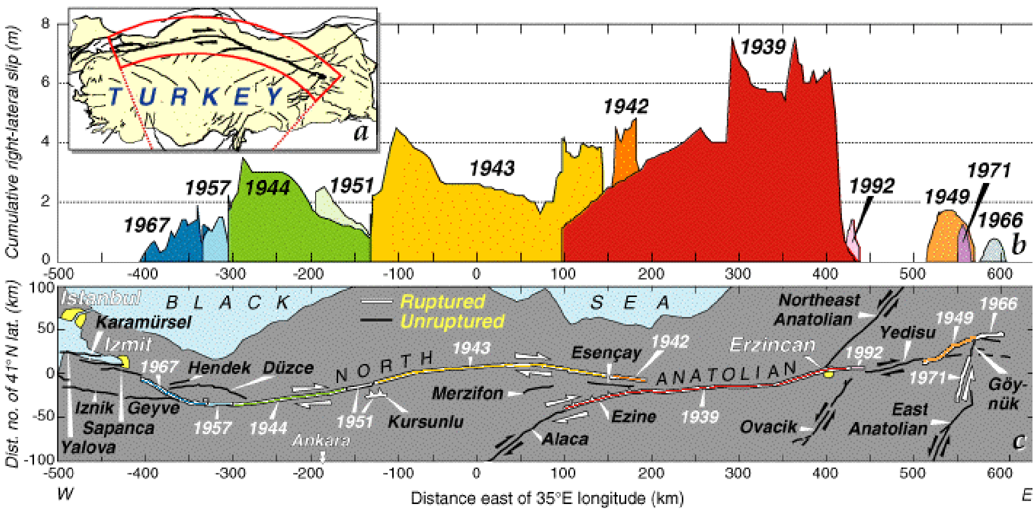
2.3 Stress transfer and aftershocks

Stress transfer

The role of stress transfer in earthquakes is summarized by Stein (1999). This phenomenon, which had been predicted by geophysicists, was first clearly shown by the 1992 Mw 7.3 Landers earthquake in southern California. The fault which slipped is concave to the west, and stress transfer theory showed that this quake should have produced an area of enhanced stress about 40 km to the west. The Mw 6.5 Big Bear earthquake struck within this area just 2.5 hours after the Landers quake.

20th century Turkey earthquakes

The earthquakes along the North Anatolian Fault from 1939 to 1999 provide another good example of stress transfer (figure below). As shown in a paper by Stein et al. (1997)[5] the steady westward progression of earthquakes along this fault is consistent with stress transfer calculations. It is argued that each of the earthquakes in 1939, 1942, 1943, 1944, 1957 and 1967 added to the stress along the adjacent westerly segment of the fault. The August 1999 Izmit quake occurred in an area that was stressed by the 1967 quake.



20th century earthquakes along the North Anatolian Fault prior to the 1999 Izmit and Duzce earthquakes

The next earthquake in Turkey

There is additional evidence that the November 1999 Duzce earthquake, which was centred to the east of the Izmit area, was a product of stress transfer from the August 1999 Izmit earthquake. It is also speculated that these recent earthquakes have resulted in increased stress on the North Anatolian Fault in the region south of Istanbul - a city of 12 million. There is additional information at: http://www.mala.bc.ca/~earles/anatolian-mar00.htm

The 1987 Parkfield earthquake

Earthquake zones don't necessarily behave in the way that they are expected to as predicted by stress transfer and stress accumulation considerations. The Parkfield segment of the San Andreas fault in California ruptured with a M 6 earthquake approximately every 20 years from 1880 to 1965, and was predicted to rupture again by 1987. See:
http://www.mala.bc.ca/~earles/parkfield-sep02.htm
In fact the "1987 Parkfield Earthquake" didn't happen until September 2004. For more information see: http://www.cisn.org/special/evt.04.09.28/

Aftershocks

An aftershock is traditionally defined as an earthquake caused by adjustments along a fault following a larger earthquake, however by any criteria we can use, an aftershock is still an earthquake in its own right. The number of

[5] http://quake.usgs.gov/research/deformation/modeling/papers/anatolia.html (see their Fig. 4),

aftershocks is partly but not strictly related to the size of the original shock, and, in most cases, the largest aftershock is at least a magnitude number smaller than the original shock. In general, the frequency of aftershocks decreases exponentially with time following the original quake.

Stress transfer and aftershocks

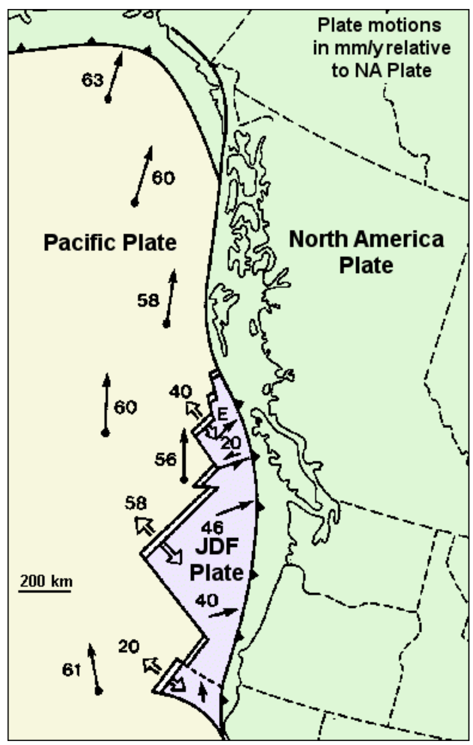
Stress transfer theory broadens the definition of aftershock to a considerable degree. Aftershocks take place in areas where the stress was increased rather than decreased by the original quake. While most occur within hours of the quake, and most occur along the same fault, this is by no means always the case. The Big Bear earthquake was clearly an aftershock of the Landers quake, and it was certainly on a different fault. Many of the North Anatolian Fault earthquakes of this century could be described as aftershocks of the Mw 7.9 1939 quake, and indeed all were smaller than that quake.

While the frequency of aftershocks decreases with time, their size does not necessarily follow the same trend. The largest aftershocks of some Californian earthquakes of this century occurred months after the original shock.

2.4 The 1946 Vancouver Island and 1700 Juan de Fuca earthquakes

Plate motions relative to the NA plate in the NW Pacific region

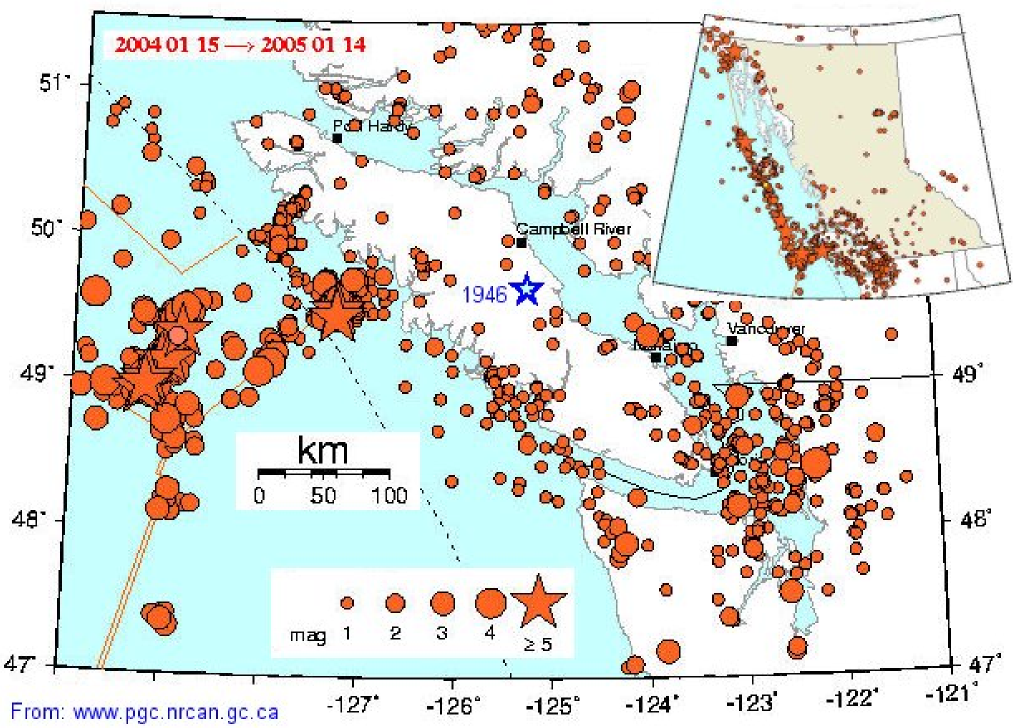
The Pacific Plate is moving north relative to the North America Plate, at a rate of about 6 cm per year. In the area north of Vancouver Island this motion takes place along the Queen Charlotte Fault (a transform fault), while to the south of Oregon, the motion is along the the Mendocino Fault and the San Andreas Fault (both also transform faults). In between these two areas there is spreading along the northeasterly trending Explorer, Juan de Fuca and Gorda ridges. The Juan de Fuca plate and related plates are moving easterly at 2 to 4.6 cm per year relative to North America, and are being subducted beneath southern British Columbia, Washington and Oregon Oregon .



Nootka Fault

The 2 cm/y difference in rates of movement of the Juan de Fuca plate and the adjacent Explorer plate is expressed in motion along the Nootka Fault (a transform fault), which extends easterly towards central Vancouver Island.

The seismicity of British Columbia and surrounding areas for the 12 months ending January 14, 2005 is shown on the figure below. Most of these are small earthquakes with magnitudes under 5. There are several main concentrations of seismic activity, including the Queen Charlotte Fault from northern Vancouver Island to the Alaska Panhandle (see inset), the Nootka Fault area extending off of the west coast of the Vancouver Island, and the area around Puget Sound, and Georgia Strait. The concentration of quakes in the Puget Sound and Georgia Strait area is assumed to be related to a flexure of the descending oceanic plate in two senses, firstly in the vertical sense, as it bends to a steeper angle (~30º), and secondly in a horizontal sense as it arches to accommodate the bend in the coastline.



Earthquakes in British Columbia and surrounding areas over a 1-year period from January 2004 to January 2005

The 1946 Vancouver Island earthquake

1946 Vancouver Island earthquake

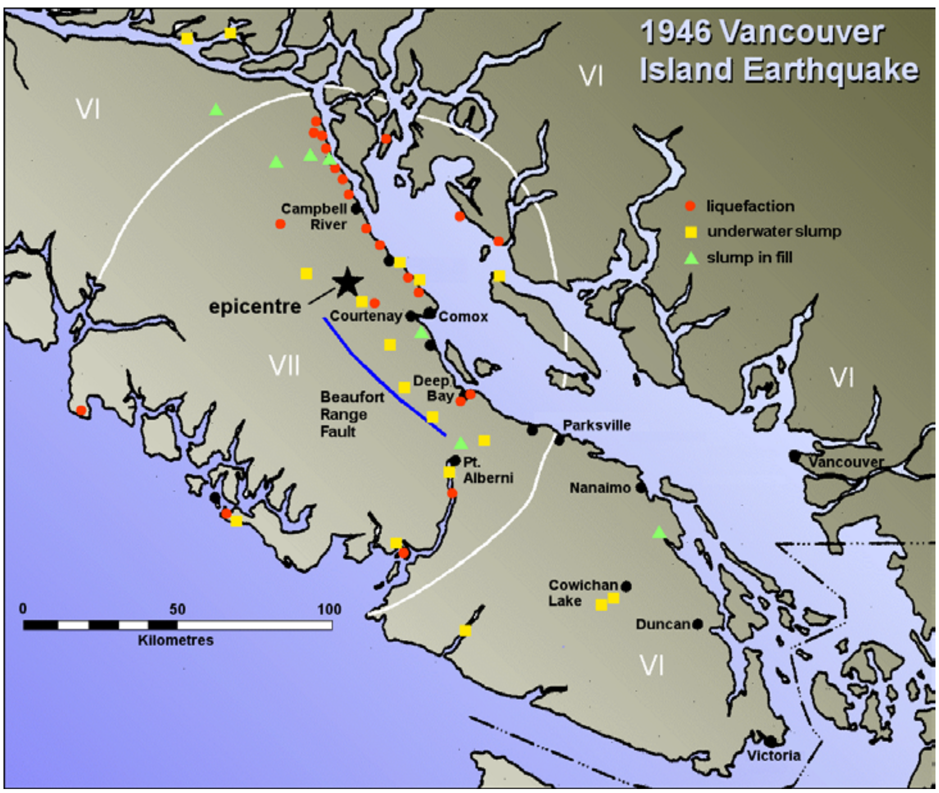
The large 1946 Vancouver Island earthquake (M_L 7.3) occurred at 10:15 in the morning on June 23rd, with its epicentre approximately 25 km west-northwest from Courtenay (Figure 2.8). The depth is estimated at 25 km, which means that the earthquake was within rocks of the North America Plate, not on the subduction zone, and not within the underlying Juan de Fuca Plate (Rogers and Hasegawa, 1978). There were very few significant aftershocks associated with this event

Focal mechanism

The actual geometry of the motion of the rocks (the ***focal mechanism***) that caused the earthquake is not clearly understood. No surface expression of the offset was observed, probably because the affected area is so remote, inaccessible and heavily forested. A detailed examination and computer interpretation of seismic data from over 50 stations has shown that the most likely explanation involves a fault parallel to the long axis of Vancouver Island - coinciding with a mapped fault known as the Beaufort Range Fault (Figure 2.9). A fault running across the island, parallel to the projection of the Nootka Fault is also a possibility, but this is rated lower because in this case there should have been evidence of offsets along the Island Highway and other roads between Courtenay and Campbell River. The estimated depth of the quake places it within the continental crust, not at the boundary with the subducting plate, and clearly not within the subducting plate itself.

Tectonic explanation

The most favoured hypothesis (by Rogers and Hasegawa) for the tectonic origin of the earthquake is that it is related to subduction motion on the descending slab of either the Juan de Fuca or the Explorer plates – and to the resulting compression and uplift of the North America plate. On the other hand, differential motion on the Nootka Fault is not ruled out.



Epicentral location, Mercalli intensity and sites of soil failure related to the June 23rd 1946 Vancouver Island earthquake

Damage

There was relatively little property damage from the 1946 quake, in spite of the high magnitude, primarily because of the low population density of central Vancouver Island in 1946, and the fact that most buildings were one and two-story wood-frame structures. Most of the damage was to brick chimneys. Reports indicate that in Campbell River 75% of chimneys were damaged, at Williams Beach (½-way to Courtenay) all 20 chimneys were damaged. In both Courtenay and Comox approximately 50% of chimneys were damaged, while at Union Bay (south of Courtenay) "most" chimneys were damaged. The area classified as Mercalli intensity VII and VIII is restricted to the central part of the Island - from Campbell River to Union Bay - and over to the west coast, including Tofino and Pt. Alberni. Most of the rest of the island, and much of the adjacent mainland (including Vancouver), had intensities of VI. Much of southwestern British Columbia, and northwestern Washington State, including the Seattle area, had intensities of V.

Slope failure

Apart from the property damage, there were many examples of soil-failure and liquefaction, and there were numerous landslides caused by the 1946 quake (Rogers, 1980). There was a spectacular landslide at Mt. Colonel Foster, just east of Gold River (Figure 2.10), and there were many other landslides in the western part of the island. In many places clay-rich soils were liquified, causing dramatic sinking of the ground, mud boiling up to the surface, and water spouts. On Reid Island, near to Campbell River, the soil slumped by nearly 10 m, and trees were splattered with mud. At Comox geysers of black mud shot 3 m out of the water. At Courtenay water came gushing out of the ground, flooding a field. In the ocean between Denman Island and Vancouver Island a 9 m high water spout was reported. At Buttle Lake about 75 m of beach simply disappeared.

Path of the 1946 landslide at Mt. Colonel Foster (Strathcona Park)

(Rock was dislodged from the bare patch on the upper right, and then slid down into the lake in the foreground.)

Most of the ground failure damage associated with the earthquake was restricted to a radius of about 50 km of the epicentre. A significant exception is the failure of near-shore sand deposits (mainly beaches) at Cowichan Lake, which is approximately 130 km from the epicentre. All around the lake beach deposits simply disappeared into the deep water, and a large wave, reported to be as high as 18 feet (5 m), caused considerable damage around the lake.

There was no major tsunami associated with the 1946 quake, although there were minor waves in areas where material slumped into the water. The only fatality directly attributed to the earthquake was a drowning when a boat was swamped by one such wave.

The next Vancouver

Because we still don't know exactly what caused the 1946 earthquake it is difficult to even speculate whether there will be another similar event in the near future. All that we can say is that whatever caused this quake is probably still happening, and that we should be prepared for another one. What is absolutely clear is that a quake of a similar magnitude would cause a great deal more damage now than it did in 1946. In recent years earthquakes with magnitudes of less than 7.3 have killed many people and caused billions of dollars of damage in places like Kobe (M 6.9 in 1995), Northridge California (M 6.7 in 1994) and Loma Prieta (M 7.1 in 1989) .

Island earthquake?

The 1700 Juan de Fuca Earthquake

1700 Juan de Fuca or Cascadia earthquake

Major earthquakes, of magnitude 8 or more, have occurred along virtually all of the world's subduction zones within the last century, but in this area of British Columbia, Washington and Oregon there has been no major earthquake within recorded history (ie. for at least 250 years). The possible explanations for the lack of a major subduction zone earthquake in historical times are as follows:

(a) the plates are not converging,
(b) the plates are converging but the subducting Juan de Fuca Plate is sliding smoothly under the North America plate, or
(c) the time interval between large events is more than 200 years.

Evidence for past earthquakes

All of the available data show that there is convergence on this boundary - between the Juan de Fuca and North American Plates - in the order of 4.5 cm per year (south of the Nootka fault). Furthermore, there is a great deal of evidence, from studies of soil profiles, that large earthquakes have occurred in this area within the past several hundred years. Geodetic studies of Vancouver Island and the Olympic Peninsula have shown that the plates are locked. For example, careful elevation measurements show that, whereas there is little change in elevation in the eastern part of Vancouver Island, there is increasing uplift towards the centre of Vancouver Island, to a maximum of several centimetres per year. This type of deformation is consistent with a locked plate boundary.

Extent of the locked zone

Based on estimated temperature gradients along the subducting Juan de Fuca Plate it is evident that there should be a locked zone between the Juan de Fuca Plate and the North America Plate extending over a width of 50 to 100 km and lying just off the coast. This width is less than that observed in most other similar situations. For example in South America the locked zone extends well underneath the continent. One reason for the restricted width is that subduction is taking place relatively near to the spreading centre (the Juan de Fuca Ridge). The subducting oceanic crustal material is only between 4 and 8 m.y. old, and thus has not cooled down as much as the 50 m.y. old crustal material subducting beneath South America, or the 100 m.y. old crustal material subducting beneath Japan.

The restricted width of the Juan de Fuca locked zone may limit the potential magnitude and damaging effects of a subduction zone earthquake in this area, however it is still believed that an earthquake of over magnitude 8 will occur at some time in the future.

Evidence for the 1700 earthquake

Although there are no direct historical records of a major subduction-zone earthquake in our area, we now know that there was such an event on January 26th of the year 1700. The story behind our understanding of the timing of this event is a fascinating one that spans two continents and three cultures. The evidence is summarized at the following Malaspina website: http://www.mala.bc.ca/~earles/1700quake. Please review the information presented at this site.

The next Juan de Fuca earthquake?

We now have convincing evidence to suggest that there will be another large earthquake in our area sometime in the future - although we don't know when. The evidence from studies of coastal soil profiles suggests an earthquake frequency of around 500 years (although the pattern is not regular and gaps can be as little as 200 years and as long as 1000 years).

All of the available data show that there is convergence on this boundary in the order of 4 cm per year. Geodetic studies of Vancouver Island and the Olympic Peninsula have shown that the plates are locked. For example, careful elevation measurements show that whereas there is little change in elevation in the eastern part of Vancouver Island, there is increasing uplift towards the centre of Vancouver Island, to a maximum of several centimetres per year. Other types of measurements show that the coastal parts of British Columbia (especially Vancouver Island) and being pushed towards the rest of the province. (These data are described in detail at the web-site of the Pacific Geoscience Centre: http://www.pgc.nrcan.gc.ca).

2.5 The 1964 Alaska earthquake

1964 Alaska earthquake

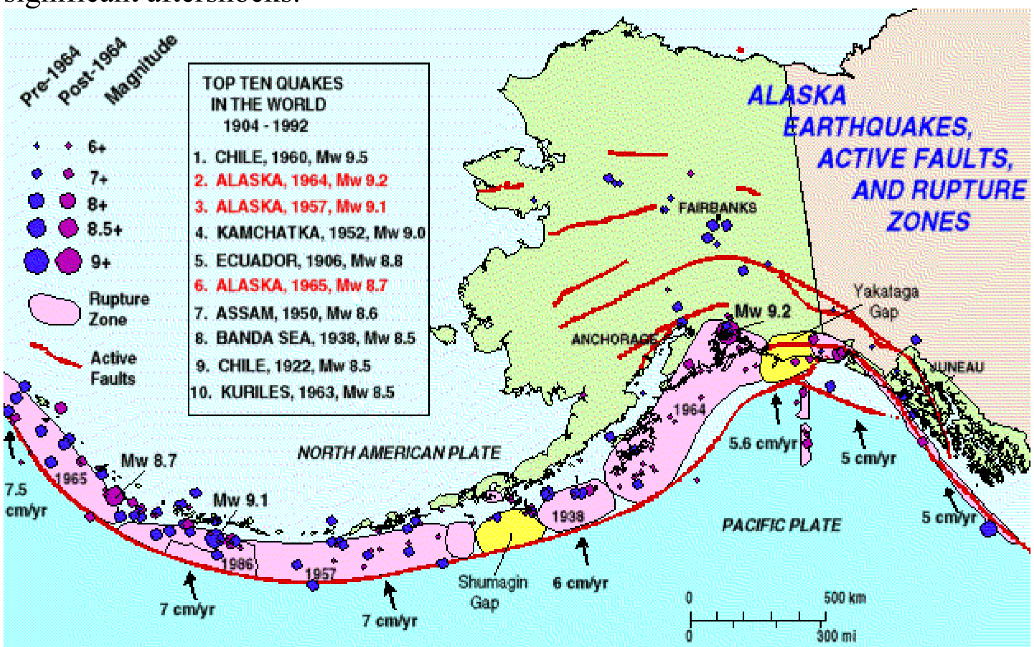
The 1964 M_W 9.2 earthquake which occurred near to Valdez, Alaska (east of Anchorage), is the largest recorded earthquake in North America, and second largest in the world after the 1960 M_W 9.5 earthquake in southern Chile). This quake is important for us to study because the geological environment of the quake, and the physical environment of the area affected, are both similar to those of the Juan de Fuca area.

Pacific Plate subduction

The Pacific Plate is subducting beneath the North America plate along an arc-shaped trend extending from the Alaska Panhandle to the end of the Aleutian Islands. The rate of subduction is high - close to 5 cm/y east of Anchorage, and over 7 cm/y at the western end of the Aleutians. As shown below, there have been many major earthquakes in this area in the past.

Main shock and aftershocks

The 1964 earthquake occurred at 5:36 in the afternoon on Good Friday - March 27th. The epicentre was approximately 120 km east of Anchorage, at a depth of 25 km. The shaking from the main shock lasted for about 4 minutes, but there were numerous large aftershocks. There were major aftershocks at 9, 19, 28, 29, 44 and 72 seconds after the initial event. In the first day there were 11 aftershocks greater than M 6, and in the next three weeks there were 9 more. In the four months following the quake there were a total of 1260 significant aftershocks.



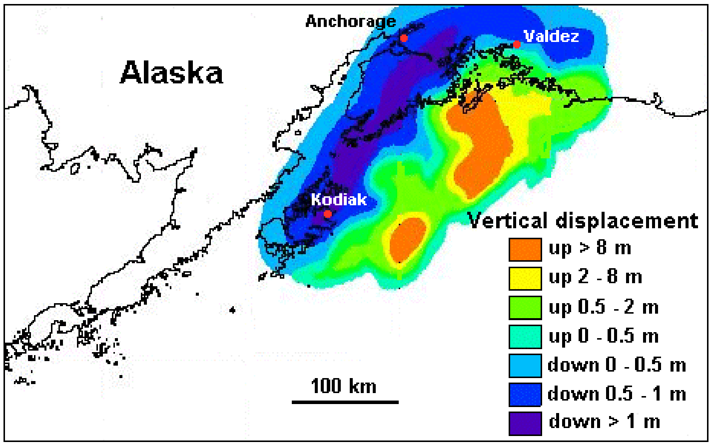
Earthquakes in Alaska during the 20th century

(for more information see: http://www.aeic.alaska.edu/)

Elastic deformation

In the decades prior to the 1964 quake subduction of the Pacific Plate had caused significant deformation of the North American Plate - including lateral compression of the crust, subsidence (largely in the area between the subduction zone and the coast), and uplift (over a wide area around Anchorage, the Kenai Peninsula and Kodiak Island). The overall extent of the deformation is not known because crustal deformation studies in this region have only been carried out in recent decades.

The 1964 earthquake resulted in a very sudden and dramatic rebound of the deformed crust. The area around Latouche (the farthest-south point of land south of Anchorage) moved 18 m laterally towards the southeast. The offshore region was uplifted by as much as 11 m, and the onshore region subsided by as much as 3 m. The so-called "hinge line" (the line of no vertical change), runs through Valdez, through the Kenai Peninsula, and along the southern shore of Kodiak Island. The greatest vertical displacements were within 50 to 75 km on either side of this line. Some areas were permanently flooded because of the extensive subsidence, while in other areas pre-existing sea-floor has been exposed. The immediate aftershocks from the 1964 quake extended for a distance of well over 800 km, from the epicentre of the first shock (near to Valdez) to the southwestern end of Kodiak Island. It is evident from this distribution, and from the displacement data, that the movement along the fault encompassed a huge area of the zone of contact between the Pacific and North American Plates, and that much of this movement took place within close to a minute of the main shock. The width of the rupture zone is close to 200 km. to Kodiak region. Further movement occurred over the following hours, days and weeks.



Vertical displacement of the crust resulting from the 1964 Alaska earthquake

The devastation of much of the area around Anchorage was severe, although the number of lives lost was limited - primarily because of the sparse population. The worst of the damage was caused by landslides and liquefactions of soils, rather than the collapse of buildings related to ground shaking. In the Turnagain Heights area, near to Anchorage, a 2600 by 365 m block of unconsolidated sediments dropped by an average of 11m and moved up to 600 m towards the sea, as a layer of clay underneath sand and gravel deposits lost strength. Some 75 homes in this area were destroyed or made uninhabitable.

Landslides

The deaths of 131 people are attributed to the 1964 quake, but of these only 9 resulted from the collapse of buildings and other structural failures. The rest were due to huge water waves generated by the deformation (uplift) in the off-shore zone. The largest wave reported was 70 m high in Valdez Arm. Whole towns in the quake region were destroyed, and in some places fishing boats were washed up into the streets. The tsunami was observed as far away as Antarctica, and 16 wave-related deaths were recorded in Oregon and California. As described below, there was major tsunami damage in Port Alberni.

Tsunami

2.6 Tsunami[6]

Sources of tsunami

Most tsunami are caused by undersea earthquakes, especially those in which there is a significant component of vertical motion. The 1964 Alaska quake is a prime example, as there was tremendous uplift of the sea-floor (to a maximum of around 11 m) over a wide area. Tsunami can also be caused by major volcanic eruptions, such as the massive 1883 eruption of Krakatau in Indonesia (which killed 36,500), and by both coastal and submarine landslides.[7]

Sea-floor displacement

The extent of the vertical motion on the sea-floor is controlled by a number of factors, including the magnitude of the earthquake, the type of quake and the depth of the quake relative to the sea floor. Concerning the type of earthquake, strike-slip or lateral-motion earthquakes like those on the San Andreas Fault do not produce much vertical motion, while subduction zone earthquakes like the Chile and Alaska quakes of 1960 and 1964 and the Juan de Fuca quake of 1700 can produce significant vertical motion. If the shelf is covered with a significant thickness of soft sediments the extent to which the vertical motion is propagated to the sea-floor may be reduced, but if the earthquake also causes collapse of these sediments, the effect may be worse.

Wave characteristics

Tsunami waves are characterized by very long wavelengths and high velocities. The distance between crests can be more than 200 km (compared with around 100 m for ordinary large ocean waves), and the wave velocity in deep water can exceed 900 km/h (compared with less than 100 km/h for ordinary waves). In the open ocean a tsunami wave may have an amplitude (still-water level to peak top) of no more than 1 metre, but as the wave approaches shallow water the velocity drops and the amplitude increases. Shoreline amplitudes of 5 to 10 m are common, and in some cases amplitudes of 30 to 40 m above ordinary sea level have been reported. Because of the very long wavelength, the time between successive wave peaks reaching shore may be as much as an hour. The first wave is not necessarily the largest in a series of tsunami waves. After the Alaska quake the citizens of Crescent City California were warned of a tsunami danger, and most moved inland. Four large waves arrived over a period of several hours, and then some people went back towards the shore to inspect the damage. They were hit with the largest wave of them all, and 12 people died.

Run-up

The height that a wave reaches on shore is called **run-up**, and is usually expressed relative to the normal high tide. The degree of run-up at any location is dependant on numerous factors, including the profile of the sea-floor near to shore, the shape of the coast and the presence of islands. The

[6] Tsunami is a Japanese word meaning "harbour wave". Technically its plural is "tsunami", but "tsunamis" is used more often.
[7] For example see: http://www.mala.bc.ca/~earles/kohala-tsunami-sep04.htm

eastern side of Vancouver Island is likely to be protected from a major tsunami because of the narrowness of the waterway and the presence of numerous islands. The western side is not protected, and places like Port Alberni are susceptible to the concentration of a tsunami wave, into what is known as a **bore**.

1964 tsunami

Tsunami from the 1964 Alaska quake reached the western coast of Vancouver Island around midnight, some 6 hours after the event. The wave heights along the open coast were small, but at Port Alberni the wave height was 2.5 times that at Tofino and Ucluelet. The maximum run-up was around 2.6 m, from the second wave. The third wave was even bigger, but by then (3:30 AM) the tide had fallen and the run-up was less. 260 homes were damaged, including two that drifted into Alberni Inlet and were never seen again. There were no fatalities, but there was about $10 million in damage at various places along the coast (Clague, 1996).

2.7 Predicting earthquakes and controlling losses

Earthquake prediction

While we do not have the knowledge or the technology to accurately predict the timing of an earthquake, we can make some predictions about the likelihood of an earthquake occurring in a certain area and within a certain period of time (usually no better than a period of decades).

The two main tools in this type of prediction are the observed deformation of the rocks, as determined from precise measurements of topography; and the history of recent earthquakes in an area where we understand the tectonic setting.

Crustal deformation

The observed vertical and horizontal rebounds associated with the 1964 Alaska quake were preceded by many decades of deformation in the opposite directions to the eventual rebound. In other words, near to the subduction zone the North American Plate had been laterally compressed by the force of the descending Pacific Plate. In addition, the offshore region had been depressed, and the onshore region had been uplifted, probably by amounts similar to the eventual rebound. Since 1964 the US Geological Survey, and other agencies, have been very active in making precise topographic measurements all along the southern coast of Alaska in order to map the magnitude of strain which is being experienced by the rocks. Using their knowledge of the magnitude of the movements, and the amount of energy released by the 1964 quake, they may be able to make some broad predictions concerning the amount of strain which the rocks of this area will sustain before the locked zone is broken again.

Crustal deformation on Vancouver Island

Similar studies are being carried out by the Geological Survey of Canada in southwestern British Columbia, including Vancouver Island[8]. It is evident that the North American plate crustal rocks are being laterally compressed in areas within one or two hundred km of the plate boundary, and also that there is uplift on parts of Vancouver Island. There is also subsidence in the western part of the island, and probably in offshore areas between the island and the plate boundary. This deformation is very similar to that which was in effect in the area of the 1964 Alaska quake, but because we have only been studying the deformation for a few years we don't know the extent to which the rocks are deformed, and we don't have any expectation that we'll be able to use this information to predict the timing of the next earthquake.

Earthquake history

Another method for predicting the likelihood of an earthquake is through the knowledge of past earthquakes along a known seismic region. In areas which have had historical large earthquakes it can be generally assumed that the stress has been released in the immediate area. Gaps in the historical quake pattern represent areas that are liable to be affected by significant earthquakes sooner than other areas.

Sesimic gaps

Along the Alaskan subduction zone there are two major gaps, one called the Shumagin gap, near to the western end of the Alaska Peninsula, and one called the Yakataga gap, situated between Valdez and border with Yukon Territory (Figure 2.11). All other regions of the subduction zone have experienced major earthquakes since 1938.

The September 1985 Mexico earthquake occurred within what was known as the **Michoacan gap**, but there still remains a significant gap to the southeast called the **Guerrero gap**. This zone lies adjacent to Acapulco, and is closer to Mexico City than the region of the 1985 quake.

Cascadia gap

There is a major seismic gap in the Cascadia subduction zone, in fact, as described above, the entire zone - from central Vancouver Island to the Mendocino Fault off California - has not had a major earthquake in 300 years. While the interval between earthquakes on this subduction zone is quite long (500 years on average), it is also quite variable, and we should be prepared for a large quake at any time.

Stress transfer

Understanding and application of stress transfer theory (as described above) should be useful for forecasting earthquakes in some areas. Where the geometry of a fault is well known it is possible to determine which nearby areas should have experienced an increase in stress, and which should have experienced a reduction in stress following an earthquake. This technique might be most applicable in areas such as California and Turkey, where many

[8] For a look at the work being done by the Geological Survey on deformation studies go to the following web site: http://www.pgc.nrcan.gc.ca/geodyn/wcda_res.html

of the faults are exposed at surface and the patterns of faulting have been mapped in some detail.

Short-term predictions

Shorter-term predictions have been made successfully in some areas on the basis of accelerated rock deformation or tilt, on the frequency of minor tremors (ie. "foreshocks"), on changes in water levels in wells, and even on animal behavior. Considerable work is being done in several such areas, but as yet no prediction method has been proven to be consistently effective.

Earthquake preparedness

While we cannot yet predict the timing of earthquakes, we do know which areas are prone to seismic activity, and we can take steps to limit the damage and loss of life. The most important of these is the design of buildings, bridges and dams. Most houses in western North America are wood-frame structures, and as such they have the flexibility to resist significant damage. New larger buildings in seismic areas of North America are all being built to strict codes, but in many areas (such as the Fraser River flood-plain) large buildings are still being constructed on unconsolidated water-saturated sediments, and thus they are still prone to the effects of liquefaction no matter how well they are designed.

[There is more on the topic of earthquake prediction and earthquake risk on pages 184 to 195 of Keller.]

Apart from ensuring that our surroundings are safe, we all have a responsibility to be personally prepared to deal with the effects of a major earthquake, including the provision of emergency equipment and supplies.

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