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What is an earthquake?
In ancient times earthquakes were thought to be caused by restless gods or giant creatures slumbering beneath the Earth.

In Japan earthquakes were thought to be caused by a monster catfish (Namazu) that lived under Japan. In this picture people are punishing the catfish for causing a large earthquake in 1855.

The great Lisbon earthquake and subsequent tsunami of 1755 caused massive destruction and had a huge effect on European scientific and philosophical development.

The early Greek philosophers developed a theory that earthquakes were caused by movements of gases trying to escape from underground. Up until the 18th century western scientists (including Newton) thought they were caused by explosions of flammable material deep underground.

In 1760 Rev J. Mitchell proposed that earthquakes were caused by rock movements and related the shaking to the propagation of elastic waves within the earth.
By examination of the displacement of the ground surface caused by the 1906 San Francisco earthquake, (photo shows a fence offset by 2.5 m by the earthquake) Henry Fielding Reid, Professor of Geology at Johns Hopkins University, concluded that the earthquake must have involved an ‘elastic rebound’ of previously stored elastic stress.

Scientists can now measure how the movement caused by an earthquake is distributed in space. By comparing two radar satellite images before and after a large earthquake we can determine how much each pixel has moved. In this interferogram of the area around the 1999 M7.3 earthquake in Izmit, Turkey, each colour band (red-blue) represents a movement of 28 mm (one radar wavelength).
Elastic rebound

The mechanisms and processes involved when earthquakes occur are extremely complex. However, some of the characteristics of earthquakes can be explained by using a simple elastic rebound theory.

- Over time, stresses in the Earth build up (often caused by the slow movements of tectonic plates)
- At some point, the stresses become so great that the Earth breaks ... an earthquake rupture occurs and relieves some of the stresses (but generally not all)

In earthquakes, these ruptures generally happen along fault planes, or lines of weakness in the Earth’s crust. There are three basic types of fault.

Earthquake rupture occurs in three stages:
1) initiation of sliding along a small portion of the fault,
2) growth of the slip surface and
3) termination of the slip.

The amount of slip will vary in different places along the fault, and will grow with time.
This sequence of images shows the growth of an earthquake rupture, modelled using data from the 1994 M6.7 Northridge earthquake in California (a movie of this process can be found at www.data.scec.org/Module/links/northrup.html images used with permission of David Wald)

The size of the fault plane in these images is 18 km wide and 14 km deep.

Aerial view of offset streams along the San Andreas Fault, caused by many earthquakes (the average movement along this boundary is ~3 cm per year) © Photograph by (Wallace, R.E.), reproduced courtesy of U.S. Geological Survey

Small offset from a single earthquake in California (M6.9 strike slip event in 1979) © Photograph by (Cavit, D) reproduced courtesy of U.S. Geological Survey

Usually with earthquakes the movement on these faults does not intersect the surface. However when it does the results are easy to see.
A histogram of all the earthquakes measured worldwide over a 6 year period shows large events are far less frequent than smaller ones.

This is more usually plotted as a cumulative plot on a log axis. This way the data plots as a straight line. \( Y = ax + b \)

This is referred to as the Gutenberg-Richter relationship after Charles Richter (1900–1985) and Beno Gutenberg (1889–1960) who both worked at the California Institute of Technology (CALTECH). The slope of the line is approximately -1 for all earthquakes (Because this is a Log 10 graph that means that for each increase in magnitude there are 10 x fewer events). This relationship holds true even when you consider earthquakes in smaller areas.

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Where do earthquakes happen?

Although the frequency and timing of earthquakes when considered on a global scale can be treated as random their location certainly cannot. A simple look at the locations of earthquakes indicates that most events occur at or near the boundaries of tectonic plates.

A good online resource for earthquake maps is at http://www.iris.edu/quakes/maps.htm

© Reproduced courtesy of IRIS
Earthquake distribution along plate boundaries is not uniform

By plotting only earthquakes of certain magnitudes or depths we can distinguish between different types of tectonic plate boundary.

- **Transform boundary** e.g. California
- **Divergent boundary** e.g. mid-ocean ridges
- **Convergent (subduction) boundary** e.g. Sumatra and Chile
- **Convergent (diffuse) boundary** e.g. Tibet and Pakistan

Large events (>M6.5) are rare on divergent boundaries.

Deep events (>100 km) only occur on subducting plates.

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SeismicEruption
A free program written by Alan Jones to download and run on your own PC called SeismicEruption allows you to playback sequences of earthquakes and volcanic eruptions plotted on a map.

The program comes with a database of events from 1960 to present (updateable via the internet).
You can select which events to plot and zoom in on different world regions.

Some 20 days of aftershocks define the fault plane of the Mw9.3 event in Sumatra.

© Reproduced courtesy of Alan Jones
The program has a feature which allows you to plot 3D wire-frame images and cross-sectional views of earthquakes.

An earthquake sequence from Alaska plotted with events colour coded by depth.

The same sequence plotted as a 3D wireframe shows the events occurring on a plane, dipping surface.

A cross-sectional view of the sequence.

© Reproduced courtesy of Alan Jones

Earthquakes in Cook Inlet, Alaska, occur on a subducting plate. By looking at the location of the events in 3D and as a cross-sectional view it is possible to visualise the geometry of the descending slab.
Recording earthquakes in the UK

Over 80% of large earthquakes occur around the edges of the Pacific ocean, the ‘Ring of Fire’ where the Pacific plate is being subducted beneath the surrounding plates.

If we select all earthquakes with magnitudes greater than 6.5 from 2000–2006 and see where they occur relative to the UK we find that only about 2% occur within 30 degrees of the UK (about 3000km).

The signals from distant large earthquakes, or ‘teleseismic’ events, can be clearly recorded in the UK with simple seismometers.

A typical signal recorded in the UK from north of Japan (M8.3, 76 degrees (8400 km) distance).

For this event both the P wave and S wave arrivals are visible. P waves show up more clearly on vertical seismometers and the S waves more clearly on horizontal seismometers.

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UK Earthquakes

About 200 earthquakes a year are recorded by BGS in the UK. However on average only a few of these are strong enough to be felt by people and damaging events (>M5.0) occur less than once per decade.

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How frequent are earthquakes?

A question always asked by members of the public (and reporters) to seismologists is can you predict earthquakes, and if you cannot then what is the point of seismology. Once you have a basic understanding of plate tectonics and the fundamental causes for earthquakes it becomes obvious that they are not random events but rather the logical end product of a long series of movements and stresses building up in the Earth at specific points. They should be predictable. However despite an enormous amount of money being spent on researching the problem and the dubious claims of some scientists there is as yet no successful method for predicting the exact timing and magnitude of earthquakes.

In essence the Earth is too complex a system for us to successfully model its behaviour to such a fine detail, in fact some scientists think that it is a fundamentally unsolvable problem. Before an earthquake, the stresses on a faults system can remain critically balanced for some considerable time. Knowing when the fault will slip (i.e. an earthquake will happen) depends on so many additional factors (some microscopic in nature) in addition to the stress that it becomes unpredictable.

An analogy is with predicting the behaviour of sand in an egg timer. We can predict with great accuracy how long the sand will take to fall through to the lower chamber. We can predict what shape the pile of sand will make in the lower chamber. However we cannot predict where an individual grain of sand will end up when it falls through the hole.

Although predicting the exact timing of an earthquake is too difficult for scientists at present it is possible to calculate some statistics about the likelihood of an earthquake striking at a particular location. This then enables engineers to design buildings to withstand the likely ground shaking for a particular area (and they will use more stringent rules for designing a hospital or a nuclear power station than a warehouse).

A simple seismometer installed in a school in the UK should detect signals from any earthquake in the world greater than magnitude 7.0 and should detect any global event greater than magnitude 6.5 if it occurs at night time (when there are fewer road traffic vibrations and vibrations in the school caused by pupils). If we are going to conduct an experiment designed to record vibrations from an earthquake we ought to know how long the experiment will last. How long would you have to wait before you record an earthquake?
Global monitoring

Scientists have been monitoring earthquakes with a worldwide network of standardised seismometers since the 1970s and accurate records of all earthquakes that have happened globally (greater than magnitude 5.0) since 1990 are available online.

Seismic hazard is usually calculated as a peak ground acceleration (in units of m/s\(^2\)) with a 10% probability of being exceeded within a 50 year period. This information about the likely intensity of shaking is what structural engineers need in order to calculate how strong buildings within a particular area need to be. The UK generally has a value of less than 0.4 m/s\(^2\) whereas parts of Greece and Turkey have values as high as 5 m/s\(^2\). Building regulations are far stricter in areas of greatest seismic hazard.
**Simple probability**

A simple observation is that worldwide on average there are 18 earthquakes each year of magnitude greater than 7.0. So the average recurrence time for these earthquakes should be 365/18 or about 20 days. However this does not mean that major earthquakes happen every 20 days, a more meaningful calculation would be to calculate the probability of a major earthquake occurring.

<table>
<thead>
<tr>
<th>Descriptor</th>
<th>Magnitude</th>
<th>Average annually</th>
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</thead>
<tbody>
<tr>
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<td>8 and higher</td>
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<tr>
<td>Major</td>
<td>7–7.9</td>
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<tr>
<td>Strong</td>
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<tr>
<td>Light</td>
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<td>13,000 (estimated)</td>
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<td>Minor</td>
<td>3–3.9</td>
<td>130,000 (estimated)</td>
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<tr>
<td>Very minor</td>
<td>2–2.9</td>
<td>1,300,000 (estimated)</td>
</tr>
</tbody>
</table>

If we assume that globally the timing of earthquakes is random and earthquake events are independent then we can say that the chance of such an event happening today is 18/365 or 0.05 i.e. there is a 5% chance of a major earthquake occurring today. (Note this simple approach only works if the probability of two earthquakes happening on the same day is very small. For other situations you have to consider the probability distribution of events.)

To calculate how long we would have to wait before a >M7.0 event it is easier to consider the probability of such an event not occurring, which is 347/365 or about 0.95. i.e. there is a 95% chance of a major earthquake not occurring today. Using the multiplication rule for probabilities we can calculate that the probability of an event not happening for two consecutive days is 0.95*0.95 =0.9025. i.e. the chance of a major earthquake not occurring for two days on the run drops to about 90%.

If we tabulate the probability of an event not occurring with increasing time we find that to reach a 50% probability of a >M7.0 event not occurring (and hence a 50% probability of an event occurring) we need to wait for 14 days. For a 90% probability of detecting a >M7.0 event we would have to wait 45 days though. (For 100% probability we would have to wait for ever!)

You can download lists of earthquakes from a number of internet sources. One of the easiest to use is the hosted by the United States Geological Survey (USGS) [http://neic.usgs.gov/neis/eqlists/eqstats.html](http://neic.usgs.gov/neis/eqlists/eqstats.html)
**Probability using a Poisson distribution (advanced)**

A more accurate approach would be to assume that the distribution of global earthquakes follows a Poisson distribution. This distribution was discovered by the French scientist Siméon-Denis Poisson in 1837 and it allows the calculation of the probability of a number of events occurring in a fixed period of time if they occur with a known average rate and are independent of the time since the previous event. It is used widely throughout science and engineering.

If we assume that the probability \( P \) of \( n \) earthquakes during a time interval \( t \) is given by

\[
P(n, t, \tau) = \left(\frac{t}{\tau}\right)^n e^{-\frac{t}{\tau}} / n!
\]

Where \( \tau \) is the average recurrence time for the event.

Then the probability that no events will occur in a time interval \( t \) is given by

\[
P(0, t, \tau) = e^{-\frac{t}{\tau}}
\]

Or the probability that at least one event will occur is

\[
P(n \geq 1, t, \tau) = 1 - e^{-\frac{t}{\tau}}
\]

Using a Poisson distribution for events greater than M7 and an average recurrence time of 20 days then we reach a 50% probability of there being an event in 14 days and 90% probability in 46 days

This is a slightly more accurate method of calculating the probability of an event, however it does rely on earthquake events not being related to each other.

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**Internet search activity**

Use the USGS earthquake search facility to discover more details about the frequency of earthquakes of different magnitudes and in different areas.


**Some suggested searches**

Calculate the average recurrence time for:

- Earthquakes >M7.0 in different 2 year periods
- Earthquakes >M6.5 in different 2 year periods
- Earthquakes >M7.0 in a hemisphere (9990 km radius) centred on lat 0 N lon 30 E
- Earthquakes >M7.0 in a hemisphere (9990 km radius) centred on lat 0 N lon 150 W
- Earthquakes >M6.0 within 2000 km of lat 3 N lon 96 E Dec 1–25 2004
- Earthquakes >M6.0 within 2000 km of lat 3 N lon 96 E Dec 26 2004–Jan 19 2005

You can select either a global search or search a rectangular or circular area.
For this exercise any output format is OK

This is the most complete catalogue of events

The global earthquake database maintained by the USGS contains earthquakes as small as magnitudes 1–2, although only for regions of very dense seismograph coverage (e.g. Southern California). The completeness of the catalogue depends on the minimum magnitude event that could occur anywhere and still be detected. Since the majority of seismograph stations are located on land this means that mid-ocean events (usually on tectonic spreading centres) which might be thousands of kilometres from the nearest seismograph stations (and remember we need at least 3–4 stations to determine an event location and time) limit the catalogue completeness. The USGS catalogue is complete for earthquakes of magnitude greater than 5.1

You can select different areas to search. For this circular area search we are selecting an entire hemisphere (180 degrees=9990 km) centred on Uganda. This will include the whole of Africa, Europe and mainland Asia.

Select a representative time period of at least a few years. This database does not contain records of earthquakes before 1973.

There are over ½ million earthquake records in this database. In order to display the results on your computer screen you must choose which ones to look at. Since 1974 there have been less than 500 earthquakes larger than magnitude 7.0

© Reproduced courtesy of USGS
You should get back a list of events. To calculate probabilities and average recurrence rates you just need to know to total number of earthquakes that your search produced.

This global search produced 84 earthquakes >M7.0 in a 5 year period. Giving an average recurrence time of $365 \times 5 / 84 = 21.7$ days.

**Suggested activities**

Use the USGS earthquake catalogue to get the data.

Calculate how many days you would have to wait for a 50% probability of there being an earthquake >M6.5 globally, then a 90% probability.

Advanced users
You can cut and paste the tables from here into a spreadsheet program to calculate histograms or to plot locations.
Complications

In practice the timing of earthquakes is far more complex than this.

Seismologists noticed a regular pattern to earthquakes at Parkfield in California with an apparent 20 year cycle. They predicted with 95% probability that an earthquake M5–6 would occur near Parkfield in the period 1985–1993. There was a magnitude 5.9 event in 2004, over 10 years ‘late’.

Intuitively you would expect there to be less probability of a large earthquake happening immediately following an initial large event. A simple elastic rebound model suggests that time is required for stress levels to build up to a critical level before an earthquake can be triggered. However there are usually a large number of smaller earthquakes on the same fault zone immediately following a large event, the frequency of these aftershocks decays with time.

In the 5 days after the M9.3 event in Sumatra on Boxing Day 2004, there were 17 events >M6.0 in the same region. Globally the average recurrence time for events >M6.0 is about 2 days.

Sometimes smaller earthquakes appear in swarms often associated with volcanic activity or sometimes with no apparent reason. In October –November 2002 a swarm of over 100 small (<M3.9) earthquakes occurred in Manchester, no-one knows why these earthquakes happened there at that time.
**Gutenberg-Richter relationship (advanced)**

If we plot earthquake magnitude against frequency of occurrence we see an interesting pattern.

Plotting the number of earthquakes against magnitude seems to show a linear relationship between the Log10 of the number of events and the magnitude.

If we reformulate this as a cumulative plot, the number of earthquakes per year of magnitude M or greater, N(M) is given by

\[ \log_{10} N(M) = a - bM \]

where \( b \approx 1 \). A ‘b’ value of 1.0 means that there are 10 times as many earthquakes >M5.0 as there are >M6.0 and 100 times as many earthquakes >M4.0 as >6.0.

This relationship seems to hold true for earthquakes globally with a b value of about 1. This b value will vary a little when you consider subsets of events from different regions but is remarkably consistent. Although the ‘b’ value on this plot (The ‘slope’ of the line ) is generally constant the intercept value will vary greatly for different regions.
This means that although earthquakes \(>M5.0\) are 10 times less frequent than earthquakes \(>M4.0\) in both the UK and California the frequency of these events is much higher in California.

**Aftershocks**

However when you look at catalogues of aftershocks (i.e. events that occur shortly after a large event that are caused by the change in stress regime brought about by a large event) you find this relationship changes to one where smaller events are more frequent (i.e. a higher ‘b’ value.) Other variations in ‘b’ value between regions are also thought to relate to differences in crustal stress. But be careful – ‘b’ values can’t be determined reliably for samples of less than about 100 earthquakes.

**Suggested activity**

Using the USGS earthquake catalogue search facility:

Try and find out the scaling laws for different areas (e.g. a high seismicity area and a medium seismicity area)

\[ y = -1.2x + 8.7 \]
Seismic waves

When an earthquake happens deep underground a crack will start to open on a pre-existing line of weakness in the Earth’s brittle crust. This crack will then grow larger and larger, relieving built up stress as it goes. The speed at which the crack propagates or grows is 2–3 km/sec. Eventually the rupture will cease to grow and will slow down and stop. The size or magnitude of the earthquake depends upon how much the fault has ruptured (the slip) and also the area over which the rupture has occurred.

This rupturing process creates elastic waves in the Earth which propagate away from the rupture front at a much faster speed than the rupture propagates, the exact speed depends upon the nature of the wave (a longitudinal or P wave is faster than a transverse or S wave), and on the elastic properties of the Earth. As you go deeper into the Earth, the density and pressure increases and so do the velocities of seismic waves.

Seismic waves are fundamentally of two types, compressional, longitudinal waves or shear, transverse waves. Through the body of the Earth these are called P waves (for primary because they are fastest) and S waves (for secondary since they are slower). However where a free surface is present (like the Earth–air interface) these two types of motion can combine to form complex surface waves. Although often ignored in introductory texts, surface waves are very important since they propagate along the surface of the Earth (where all the buildings and people are) and usually have much higher amplitudes than the P waves and S waves. It is usually surface waves which knock down buildings.

Seismic waves, like all waves, transfer energy from one place to another without moving material.
Wave propagation through a grid through a grid representing a volume of material. The directions X and Y are parallel to the Earth's surface and the Z direction is depth. T = 0 through T = 3 indicate successive times. The material returns to its original shape after the wave has passed. Animations of these images can be found at http://web.ics.purdue.edu/~braile/edumod/waves/WaveDemo.htm

Surface waves have a complex motion that decreases in amplitude with depth, the material returns to its original shape after the wave has passed. Animations of these images can be found at http://web.ics.purdue.edu/~braile/edumod/waves/WaveDemo.htm

P-waves are a compression followed by a dilatation. The particle motion is in the direction of propagation. Sound waves are P-waves.

S-waves have an up motion followed by a down motion. The particle motion is perpendicular to the direction of propagation.
Rayleigh waves have an elliptical motion similar to that of water waves.

Love waves have a motion that is horizontal and perpendicular to the direction of propagation.

“© L. Braile. 2000-2006”
P waves and S waves can be easily demonstrated in the classroom with a slinky.

Hold the slinky stretched across a classroom. Compress 4–5 coils at one end, wait for the slinky to come to a rest and then release the compressed coils.

Hold the slinky stretched across a classroom. Pull the slinky 10 cm perpendicular to the line 4–5 coils from one end. Wait for the slinky to come to a rest and then release the pulled coils.

“© L Braile. 2000-2006”
Seismic waves and the earth

P waves and S waves travel differently through the Earth., P waves travel faster and S waves cannot penetrate the liquid outer core.
Alan Jones has written a free program to illustrate how seismic waves propagate through the Earth called ‘Seismic Waves’ which is available to download at [www.geol.binghamton.edu/faculty/jones](http://www.geol.binghamton.edu/faculty/jones).

The program contains data from several real earthquakes and shows the wavefronts from these events travelling through and around the Earth, including refractions, reflections and conversions at the major velocity discontinuities in the Earth. The views are speeded up records of the wavefronts and every time a wavefront reaches one of the seismic stations around the world the seismogram recorded at that station is played (speeded up) as a sound file. You can choose which earthquakes and which views (cross-section, seismograms or surface) to see.

By default the program shows all possible seismic phases which can produce a cluttered image. To see what happens to the main P and S waves as they travel through the core click the Phases... button

And select only the main phases to view.

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## Summary of seismic wave types and properties

<table>
<thead>
<tr>
<th>Type (and names)</th>
<th>Particle motion</th>
<th>Typical velocity</th>
<th>Other characteristics</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>P</strong>&lt;br&gt;Compressional&lt;br&gt;Primary&lt;br&gt;Longitudinal</td>
<td>Alternating compressions ('pushes') and dilations ('pulls') in the same direction as the wave is propagating</td>
<td>$V_P \sim 5 - 7$ km/s in typical Earth's crust:&lt;br&gt;$\geq 8$ km/s in Earth's mantle and core;&lt;br&gt;1.5 km/s in water; 0.3 km/s in air</td>
<td>P motion travels fastest in materials, so the P-wave is the first-arriving energy on a seismogram. Generally smaller and higher frequency than the S and surface waves. P waves in a liquid or gas are pressure waves, including sound waves.</td>
</tr>
<tr>
<td><strong>S</strong>&lt;br&gt;Shear&lt;br&gt;Secondary&lt;br&gt;Transverse</td>
<td>Alternating transverse motions perpendicular to the direction of propagation.</td>
<td>$V_S \sim 3 - 4$ km/s in typical Earth's crust:&lt;br&gt;$\geq 4.5$ km/s in Earth's mantle; ~ 2.5-3.0 km/s in (solid) inner core</td>
<td>S-waves do not travel through fluids, so do not exist in Earth’s liquid outer core or in air or water or molten rock (magma). S waves travel slower than P waves in a solid and, therefore, arrive after the P wave.</td>
</tr>
<tr>
<td><strong>L</strong>&lt;br&gt;Love&lt;br&gt;Surface waves</td>
<td>Transverse horizontal motion, perpendicular to the direction of propagation and generally parallel to the Earth’s surface</td>
<td>$V_L \sim 2.0 - 4.5$ km/s in the Earth depending on frequency of the propagating wave</td>
<td>Love waves exist because of the Earth’s surface. They are largest at the surface and decrease in amplitude with depth. Love waves are dispersive, that is, the wave velocity is dependent on frequency, with low frequencies normally propagating at higher velocity. Depth of penetration of the Love waves is also dependent on frequency, with lower frequencies penetrating to greater depth.</td>
</tr>
<tr>
<td><strong>R</strong>&lt;br&gt;Rayleigh&lt;br&gt;Surface waves</td>
<td>Motion is both in the direction of propagation and perpendicular (in a vertical plane)</td>
<td>$V_R \sim 2.0 - 4.5$ km/s in the Earth depending on frequency of the propagating wave</td>
<td>Rayleigh waves are also dispersive and the amplitudes generally decrease with depth in the Earth. Appearance and particle motion are similar to water waves.</td>
</tr>
</tbody>
</table>

Stated images and text in this document are adapted from those of Larry Braile [web.ics.purdue.edu/~braile](http://web.ics.purdue.edu/~braile).

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[www.bgs.ac.uk/schoolseismology](http://www.bgs.ac.uk/schoolseismology) 29  Author Paul Denton
Building a simple seismometer

Seismometers operate on the principle of inertia, i.e. a body at rest will tend to remain that way unless a force is applied to make it move. An ideal seismometer would be a mass floating just above the ground. When the ground moves (due to the vibrations caused by an earthquake) the mass remains stationary (since it is experiencing no force to make it move), we then directly measure the relative motion between the ground and the floating mass which will tell us precisely the motion of the ground caused by the seismic waves. In the very earliest instruments (like this Chinese seismoscope) the motion of an inertial mass would cause a delicately balanced ball to drop from a dragon’s mouth, giving an indication that ground motion had occurred and from which direction.

However in practice masses do not float freely above the ground; they need to be held up by some mechanism which will transmit the motion from the ground to the mass and cause it to move with the ground. Seismometers are designed to ensure that this transmission of force is transmitted very weakly for the range of frequencies that we are interested in measuring.

Seismologists measure the motion of the ground in three separate directions (usually up–down, east–west and north–south) to get a complete record of how the ground has moved. However for most purposes measuring just one of these motions is usually sufficient.

A seismometer trying to detect seismic waves from large earthquakes thousands of km away needs to respond to signals with a frequency in the range 0.05 Hz (surface waves) to 2 Hz (P waves). For such small frequencies it is often easier to refer to the period of the wave (period=1/frequency) so 0.05 Hz = 20 seconds and 2 Hz = 0.5 seconds. Simple mechanical systems like pendulums or masses on springs will respond well to ground motion at frequencies at and above their natural frequency of oscillation.
Simple seismometers

For a simple pendulum of length $l$ the natural period of oscillation $T$

$$T = 2\pi \sqrt{\frac{l}{g}}$$

Where $g$ is the acceleration due to gravity (9.8m/s²)

For a pendulum of length 1m this gives $T \approx 2$ seconds (about the period of a pendulum in a grandfather clock). To design a simple pendulum with a natural period of 20 seconds we would need a pendulum length of ~100 m.

However if we adjust the suspension system to that of a ‘garden gate’ or ‘Lehman pattern’ where a horizontal boom rotates about a two point suspension we can adjust the effective length of the resulting system by adjusting the angle of suspension. In this case the effective length of the pendulum is $l/\sin(\alpha)$ so the natural period of the system $T$ becomes

$$T = 2\pi \sqrt{\frac{l}{g \sin(\alpha)}}$$

In this way it is relatively easy to design a seismometer with a reasonable sized boom (0.5 m) that has a natural period of 20 seconds by slightly tilting the system with an angle of 0.3 degrees (this corresponds to tilting one end of the seismometer by ~2 mm)

SEP school seismometer system

Simplified schematic of a simple seismometer (with the angle of tilt greatly exaggerated)
The SEP seismometer system is a simple horizontal motion seismometer. It is:
- simple in principle
- open in construction so that individual components can be identified
- adjustable in natural period (10–20 seconds)
- adjustable in damping factor
- sensitive enough to detect signals from M6.5 earthquakes anywhere in the world (20–30 per year)
- cheap enough for schools to purchase.

It comes complete with an electronics box:
- wideband amplifier (x 100, x 200 or x 500)
- bandpass filters between 0.0166 Hz and 5Hz
- 16-bit digitiser at 20 samples/sec

This unit can then be connected to a PC to record and analyse the data (see datasheet on Amaseis software).
**Vertical seismometers**

At its simplest a vertical seismometer consists of a mass suspended on a spring. The natural frequency of such a system depends on the size of the mass and the spring constant \( k \) of the spring, (where \( k \) is determined by plotting a graph of force \( F \) required to stretch the spring against the extension \( x \) produced and measuring the slope of that graph). Lower frequencies can be made by using larger masses or springs with lower spring constants.

In practice such a system is very susceptible to horizontal motion (it acts like a simple pendulum). Usually a pivot or hinge system is employed to constrain the mass to move only in the vertical plane, this then also gives the possibility of using angled spring systems to create longer period systems.

One such design is the ‘Lacoste’ suspension system which can easily be made to have a natural period of 2 seconds or more.

In order to faithfully record the actual motion of the ground in 3D seismologists need to use three separate sensors each recording the vibrations in different direction. These are usually orientated:
- **Z** component measuring up–down motion
- **E** component measuring east–west motion
- **N** component measuring north–out motion.

Vector addition of these data streams allows reconstruction of the motion in any direction.
Detecting the motion

Once we have a mechanical system in which the motion of an inertial mass moves relative to the ground (and hence the seismometer frame) when a seismic wave passes we need some method of detecting and recording this relative motion.

Historically this motion was recorded by pens moving against a rotating drum. There would be a simple lever system connecting the mass to a pen which would mechanically amplify the motion of the mass and the paper drum would rotate with a clockwork mechanism that would enable users to determine the time at which a seismic wave had arrived (to within a few tens of seconds).

However the sensitivity and dynamic range of these systems was very limited and they were quickly superseded by electromechanical systems in which the motion of the mass was converted to a voltage and the voltage was amplified before causing a pen to make marks on a rotating drum. This also allowed the recording system to be located away from the sensor, by using radio or telephone line links this analogue voltage signal could be transmitted hundreds or thousands of kilometres to a central laboratory where seismic data from a whole network of seismic stations could be recorded and analysed together.

Nowadays the seismic signal is digitised (usually into 16 bit or 24 bit binary code) with the data being recorded and analysed by computer. This enables seismologists to use the internet to exchange data in realtime and create vast ‘virtual networks’ of seismic stations that span the globe and be automatically analysed to provide near instantaneous warnings after large events.
Detecting the motion

The principle of electromagnetic induction used in a seismometer is the same as that of a loudspeaker (in reverse). The relative motion between a magnet and a coil (one of which is attached to the inertial mass and one is attached to the frame) generates a voltage in the coil that is proportional to the velocity of the relative motion.

The magnitude of the voltage is also proportional to the strength of the magnet used and the number of turns in the coil.

In practice either the magnet or the coil can be attached to the inertial mass (in commercial systems the magnet is itself often used as the inertial mass)
**Damping**

Lightly damped seismometer signal

- Simple mechanical seismometers will have a natural frequency of oscillation at which they resonate.
- If we try to record a signal that contains components close to this natural frequency then the seismometer will resonate.
- We can prevent this resonance by applying a damping force to the motion of the inertial mass.
- This can be done with a simple dashpot (an oil-filled pot with a paddle in attached to the boom). This produces a force due to viscous resistance that opposes the motion and is proportional to the speed of the motion.
- We can also use a magnetic damping system where a conducting plate is attached to the moving boom which moves in a static magnetic field, this produces strong eddy currents in the conducting plate when the boom moves, these eddy currents produce their own magnetic field which opposes the static field. The end result is a force that opposes the motion and is proportional to the speed of the motion.

Near critically damped signal
**Weight drop test**

A simple vertical seismometer was constructed using a wooden boom, a Stanley knife blade hinge, an magnet and a home wound coil. The coil is connected directly to a PC soundcard for recording. Suitable sound recording software is AUDACITY which is free and can be downloaded from [http://audacity.sourceforge.net/](http://audacity.sourceforge.net/)

This sensor has a natural period of about 2 seconds. It is unsuitable for recording earthquake signals since the wooden boom will experience shrinkage and warping with changes in temperature and humidity.

Recording seismic signals with soundcards can be problematic, soundcards generally have a high-pass filter set at 20 Hz. Seismic signals generated by dropping a weight on the floor (try using a 4 kg bag of sand tightly bound) will have a dominant frequency of about 100 Hz. A simple seismometer like the one illustrated will respond to these high frequencies without any damping required (the 20 Hz high pass filter on the soundcard will remove any signals close to the natural period of the system at about 0.5 Hz).
**Simple experiments with simple seismometers**

A simple homemade seismometer connected to a PC soundcard can be used to carry out a number of simple experiments to investigate how the amplitude and timing of signals recorded varies with different sources. For these experiments a standard weight (4 kg of sand tightly bound in a strong bag) was used and the data was recorded using the free Audacity sound recording software.

A weight dropped 1m at different distances from the sensor shows how the amplitude of the signal decays with distance in a non-linear way. Another simple experiment is to keep the distance between the weight drop and the sensor fixed and to vary the height of the weight-drop (and hence the energy)}
The results of this experiment indicate that the amplitude of the recorded signal is proportional to the height of the weight-drop.
Amaseis seismic recording and analysis software

The Amaseis program written by Alan Jones is a freely available software package that can be used to monitor and record data from a school seismometer system but can also be used as a stand-alone program to view and analyse seismic data downloaded from the internet.

The installation package *AmaSeisSetup.exe* should be downloaded from [http://www.geol.binghamton.edu/faculty/jones/AmaSeis.html](http://www.geol.binghamton.edu/faculty/jones/AmaSeis.html).

Save the setup program to your hard-disk and then run the program by double clicking (you might need to get your system administrator to install the software for you).

You might get a security warning when you try to install the software, click *Run*.. and then *Yes* to continue.

A standard setup sequence will then start, prompting you for your name and institution, default values will be OK at this stage.
If possible try to install the software in the default destination (c:\AmaSeis) and assign it to the default Seismology Program Folder.

The installation program will have created a desktop icon for AmaSeis, double click to start.
If you try to run Amaseis without a sensor attached it will complain that a COM port could not be opened, you can safely ignore this message and click **OK**.

The opening screen of Amaseis will again remind you that no data is being received from a seismometer, if you are not using a seismometer you can safely ignore this message and click **OK**.
By default AmaSeis will show 24 hours of data as a “helicorder plot”. In this plot each line of data represents one hour’s worth of data with time along the X axis. The position of the plot down the page is offset for each successive hour of the day (earliest at the top and latest at the bottom). Along each trace the amplitude of the wiggles would correspond to the amplitude of the received signal from the seismometer (all zero in this case since no seismometer is attached).

Using the File… open tabs or the File open icon, you can review data files downloaded from the internet in SAC Binary format.
The SAC data files downloaded from IRIS will contain information in their headers about the seismic station which recorded them and the event.

This header information is printed on the display screen.

In this case Station ESK (Eskdalemuir in Scotland) recorded an event located at Lat 46.6 and Lon 153.2 depth 27.7km. This event was 76 degrees distance from ESK.
Earthquakes generate vibrations over a very wide of frequencies, for large earthquakes these can range from as low as 0.01Hz (100 seconds period) to 100Hz (0.01 seconds period). However the earth will rapidly attenuate vibrations at the higher frequencies so when we look at data from earthquakes that occurred thousands of kilometres away we generally only see signals with frequencies less than 1 Hz (1 second period).

Microseisms are ever present background vibrations in the ground with a dominant period of six seconds. They are caused by the action of oceanic water waves on the coast. When there are large storms in the nearest ocean to a seismic monitoring station the amplitude of these microseismic vibrations increases.

The data can be improved by applying a frequency filter to remove unwanted noise from the data. For large distant events the best filter to apply is one which attenuates all frequencies above 0.1Hz (10 second period), this is called a low-pass filter.
Simple frequency filters can be accessed by using the control filter tabs

For smaller events, or to emphasis the P and S wave arrivals it is better to concentrate on the higher frequencies by applying a high-pass filter.

By clicking on the travel time icon you can proceed to a view of this seismogram overlaid by travel time curves. Enter the earthquake depth in the Depth Dialog box.
The seismogram will then be displayed on a graph of travel time curves.

The seismogram can be dragged around the display area with the mouse.

Drag the seismogram until the P and S arrivals match with the Curves labelled P and S. This should occur at the great circle distance of 76 degrees for this example.

The curves represent travel times for raypaths identified by the phase codes P, PP, S, SS etc (see appendix)

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Finding out what earthquakes have happened recently

Subscribe to an email alert system

Hosted by the United States Geological Survey
http://earthquake.usgs.gov/eqcenter/ens/

Once registered you can select which earthquakes you would like to be notified of. Selecting global earthquakes greater than magnitude 5.5 will usually result in one or two messages per day.

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**UK Earthquakes on the Web**

The British Geological Survey (BGS) record about 200 earthquakes per year in the UK, although only one or two each year are large enough to be felt by people (> M4.0). Information about UK earthquakes can be found viewed at the BGS website http://www.earthquakes.bgs.ac.uk.

The BGS website also lets you look at live seismic data from 9 seismic stations around the UK.

The data is presented as “helicorder plots” where each line of data is a time record of consecutive 30minute time periods. The data can be viewed as

- Short period (3-5Hz) for local events (as shown here)
- Long period (0-1Hz) for regional events
- Very Long period (0-0.1Hz) for distant events
**Worldwide earthquakes on the web**

Go to [www.iris.edu](http://www.iris.edu) and click on Seismic Monitor, this will give you a graphical presentation of recent earthquakes. To see more details about a particular event you can either click on the event on the map or click on last 30 days earthquakes.

**Other internet sites for earthquakes**

- [www.orfeus-eu.org](http://www.orfeus-eu.org) - European centre for archiving broadband seismic data
- [www.emsc-csem.org](http://www.emsc-csem.org) - European centre for locating earthquakes
- [www.isc.ac.uk](http://www.isc.ac.uk) - International Seismological Commission, the definitive archive of earthquake locations (not realtime)
Looking at real data from a recent event

IRIS maintains an online database of seismic data from over 8000 seismic stations around the world. In 2006 this database held 55 terrabytes of data and was growing at about 12 terrabytes per year. Realtime data for recent large earthquakes from hundreds of seismic stations is easily available through their website. Go to the WilberII interface directly at http://www.iris.edu/cgi-bin/wilberII_page1.pl

Get a list of events by choosing the relevant time period (e.g. last 90 days) and clicking on list all events (You can also choose an event by clicking on the map)

© Reproduced courtesy of IRIS
The underlined date (some events are listed with two entries FARM and SPYDER, either will allow you to access the data although FARM should be quicker).

Choose the seismic network you want to see data from. You will see a list of all the regional networks of seismometers that have recorded this event. One good seismic station to start with is at Eskdalemuir in Scotland (called station ESK in network II the Global Seismic Network.)

Nearby seismic stations

ESK in the II network is a station at Eskdalemuir in Scotland that forms part of the Global Seismic Network and produces high quality data.

STED in the UK network is the first UK school to be able to relay live data from their own seismic station to IRIS, St Edwards School in Poole has a Guralp Instruments vertical component broadband seismometer.

Others

QSPA in network IU is at the South Pole and has very low noise.

Network PN has data from 7 high schools in Indiana

Other data centres.

There is a European equivalent of IRIS called ORFEUS www.orfeus-eu.org which archives data from many European stations (including another 5–6 stations in the UK) ORFEUS also operate a WILBER system for retrieving data.

Select the event you are interested in and click on the underlined date (some events are listed with two entries FARM and SPYDER, either will allow you to access the data although FARM should be quicker).
A large event will be recorded by thousands of seismic stations worldwide. Each seismogram from each station is a record that can never be repeated and contains unique information about both the event and the nature of the Earth in between the event and the recording station. Scientists studying the structure of the earth use data from as many different stations and events as they can in order to build up a detailed 3D view of the interior of the earth using a technique called tomography.

For this reason wherever possible seismic data is permanently archived in databases at large data centres. It might be analysed several times years or even decades after it was recorded.

Once you have chosen the network or networks you want scroll down to the bottom of the page and click proceed.
Downloading data files, to download data files from any of these stations for a direct comparison with your own data recorded in Amaseis. Your data request will be processed and made available for download. If you selected all components then you will see a number of different files for the same event.

Choose network code (ALL)

Choose channel BHZ for vertical or BHE for East or BHN for north data

You can automatically select (filter) only some data sets

Choose SAC Binary individual files

You can produce simple record sections from this point (see box) . Only for advanced users

Choose 5 minutes before P to start and 60 minutes after P to end

Identify yourself here and give this dataset a name

The requested dataset can then be retrieved from the database.

Sometimes it is useful to look at data from lots of different stations at the same time. One way of doing this is by plotting a record section. On a record section the position of each trace up the page (the Y-axis) is governed by the distance of that recording station from the event. The x-axis (time) is common for all traces. On a record section it is easy to see how the arrival of certain seismic phases varies with distance from an event. In this case the direct P arrival (in green) disappears at about 105 degrees distance. Also the direct S arrival (orange) disappears. This is because of interference from the Earth’s liquid outer core which sharply bends (or refracts) P waves and does not allow S waves to pass through at all.
Depending on the time of day and the amount of data you have requested this will take a minute or two.

The processing system will process small data requests first, so if you only want data from a single station your data request will probably be processed first.

When your data is ready for download a link will be provided to an ftp site. Click on the current FTP directory link to access your data.

www.bgs.ac.uk/schoolseismology
Right click on the data file you wish to download

**FTP Directory:**
ftp://ftp.iris.washington.edu/pub/userdata/ Paul/ Kuril_event/

The database will automatically assign a filename to your data file which refers to the contents of the data file.

- **File name:** 2006.319.11.20.55.4733.II.ESK.00.BHZ.R.SAC

  - **II** is the network code
  - **ESK** is the Station Code
  - **00** is a code for the sample rate
  - **SAC** indicates the data format
  - **BHZ** indicates the data is Broadband, High sample rate, Vertical component data

The first digits are the start time of the data
yyyy.ddd.hh.mm.ss.ssss

**Three component data**

In order to faithfully record the actual motion of the ground in 3D seismologists need to use three separate sensors each recording the vibrations in different direction. These are usually orientated:
- **Z** component measuring up–down motion
- **E** component measuring east–west motion
- **N** component measuring north–out motion.

Vector addition of these data streams allows reconstruction of the motion in any direction.
You can now view the data file in Amaseis (Use the File > Open option, see the ‘Using Amaseis’ data sheet).
See resource ‘Decoding Seismograms’ to find out how to identify when the different wave types (called phases) arrive at your station.

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Decoding seismograms, theoretical travel times

Once you think you have located an earthquake on your record you can find out what the details of the seismogram are showing you by finding out the predicted arrival times of seismic waves from this event at your station. These predicted arrival times use an approximate velocity model for the Earth (called the IASP91 model) that should give arrival times at your station accurate to a second or two.

The website to use for this is [http://neic.usgs.gov/neis/travel_times/artim.html](http://neic.usgs.gov/neis/travel_times/artim.html)

Seismic waves can travel from an earthquake to a recording site by a large number of different paths, some direct and some involving reflections and refractions from velocity contrasts within the earth. Each different raypath is called a “phase” by seismologists (see appendix for how phases are named).

Simple velocity models of the Earth

To a first approximation the Earth behaves like a sphere where the seismic velocity varies only with depth. Such a model is referred to as a 1 dimensional or 1D model.

There are a number of 1D velocity models used by seismologists (IASP91, PREM, AK135) however the variations between them are too small to show on a simple plot like this. (Note that the S wave velocity in the liquid outer core is zero.)
You will then see the results for a number of recent earthquakes of the form:

```
DATE-(UTC)-TIME   LAT    LON    DEPTH   MAG   Q   COMMENTS
```

For example:

```
2006/08/24 21:50:36  51.16N 157.49E  43.0  6.5      US: NEAR EAST COAST OF KAMCH
```

**Expected 20s period surface wave amplitude**:

- $2.44 \times 10^1 \mu$m
- $7.68 \times 10^0 \mu$m/s

**Expected 1s period body wave amplitude**:

- $5.92 \times 10^{-1} \mu$m
- $3.72 \times 10^0 \mu$m/s

The first line gives the origin time, location, and magnitude of the event.

You can now decode the seismogram that you have recorded or downloaded.

Note that if you are looking at horizontal ground motions the initial P waves do not always show up well.

You will probably only see the main phases (highlighted). See later for phase naming conventions.

Epicentral distance in degrees is how seismologists refer to distance. This refers to the angle subtended at the centre of the earth by the great circle path on the surface linking earthquake and recording station. One degree of distance equals approximately 111km.
For this particular seismogram the event was approximately 75 degrees away (see box). For events greater than 103 degrees distance seismic rays have to pass through the earth’s core which is liquid. This means that S-waves (transverse waves) cannot penetrate at these distances, this gives rise to the **S-Wave shadow zone**, the region beyond 103 degrees where recording stations will not detect directly arriving S-waves.

P-waves (compressional waves) are transmitted by liquids but P-waves which pass through the liquid core are bent (refracted) quite strongly by the large velocity contrast. This gives rise to a zone between 104 degrees and 143 degrees where direct P waves cannot be seen which is referred to as the **P wave shadow zone**

You should now be able to identify the various wiggles on your seismogram with the appropriate phase label and hence know the raypath through the earth that this energy travelled by.
Amaseis has a useful feature that allows you to overlay a set of traveltime curves on your seismogram that can either be useful for identifying phases if you know how far away an event is, or for determining how far away an event is if you have managed to identify the phases. If you have downloaded data from another station from the IRIS website you can load it into Amaseis and analyse it alongside your own data for comparison (use file open and make sure that your downloaded file has a filename of the form ******.sac).
**Naming the main seismic phases**

This is a simplified list of the main seismic phases that you might see on a seismic record from a distant earthquake. Each leg of a seismic raypath is assigned a letter. P indicating that the leg is traversed as a longitudinal wave, S as a transverse wave, K that the wave has travelled through the core (by necessity as a longitudinal wave). Each interaction at the Earth’s surface or the core mantle boundary initiates a new leg.

### Phases of distant shallow earthquakes

**P, S** Direct longitudinal or transverse waves.

**PKP, SKS** Direct longitudinal or transverse waves traversing the Earth's core. S waves passing through the core as P waves, transformed back into S waves on emergence.

**PP, SS** P or S waves reflected at the Earth's surface.

**PcP, ScS** P or S waves reflected at the Earth's core boundary.

**PS, SP** P and S waves reflected and transformed at the Earth's surface.

**Pdiff, Sdiff** P or S waves diffracted around the Earth's core.

---

The speed of seismic waves through the Earth increases with depth (see the 1D velocity model box). This causes rays to be continuously refracted into curved ray-paths.
Phases of deep-focus earthquakes

The major branches of the travel-time curves carry the same descriptions as for shallow-focus events. Waves leaving the focus in an upward direction, and reflected at the surface are described by the letters p, s, as follows:

pP, sS etc  P or S waves reflected from the surface as P waves.

Surface waves

Surface waves travel around the free surface of the earth along great circle paths.

LQ  Love waves.  LR  Rayleigh waves.

A complete list of seismic phase names can be found at [www.iris.edu/data/vocab.htm](http://www.iris.edu/data/vocab.htm)
For a standard reference model of the Earth’s velocity structure it is possible to calculate the theoretical arrival time of any particular ray-path (or seismic phase) for any given earthquake. A commonly used velocity model is called IASP91, a radially symmetrical velocity model where velocity varies only with depth. On the left is a graph of how the time taken for different raypaths varies with distance. Note that not all phases are present for all distances.

The diagram on the right shows some of the many possible routes that a seismic wave can take through the Earth’s mantle, liquid outer core, and solid inner core.

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Earthquake magnitude

The magnitude of an earthquake is a number that relates to the amplitude of the earthquake. Earthquake magnitude scales are logarithmic (i.e. a 1 unit increase in magnitude corresponds to a 10 fold increase in amplitude). Scientists can only estimate the true magnitude of an earthquake by measuring its effects, this leads to earthquakes appearing to have different magnitudes depending on what method is used for estimating the magnitude and which datasets have been used to make this estimate. Press reporters love the Richter scale and will report any earthquake magnitude as a ‘magnitude on the Richter scale’, however for any large earthquake that has made the news it is very unlikely that a Richter (or Local magnitude) is an appropriate scale. Due to the complexities of the calculations the reported magnitude of events can change as more data gets analysed. For the devastating Dec 24th 2004 event in Sumatra the original magnitude of Mw 9.0 was later recalculated to Mw 9.3 as more data was analysed.

Richter magnitude (Local Magnitude \( M_L \))

Originally earthquake magnitudes were based on the amplitude of ground motion displacement as measured by a standard seismograph. The best known of these is the Richter Magnitude which was defined for local earthquakes in southern California

\[
M_L = \log A + 2.56 \log D - 1.67
\]

Where \( A \) is the measured ground motion (in micrometers) and \( D \) is the distance from the event (in km). This is still used for measuring the magnitude of shallow events at distances less than 600 km (today called the Local Magnitude). For events larger than magnitude 8 this scale saturates and gives magnitude estimates that are too small.

Body wave magnitude \( M_b \)

For earthquakes measured at distances greater than 600 km magnitude can be estimated from the formula.

\[
M_b = \log (A/T) + \sigma(D,h)
\]

Where \( A \) is the maximum amplitude (in micrometers) of the P waves measured at period \( T \) (generally about 1second) and \( \sigma \) is a calibration term (in the range 6–8) that depends on distance from the event \( D \) and depth of the event \( h \) (tables of \( \sigma \) are used).

Surface wave magnitude \( M_s \)

For shallow earthquakes (i.e. ones that generate surface waves) magnitudes can be estimated using the formula.

\[
M_s = \log (A/T) + 1.66 \log \Delta + 3.3
\]
Where $A$ is the maximum amplitude (in micrometers) of the Rayleigh waves, $T$ is the period (usually about 20 seconds) and $\Delta$ is the distance (in degrees).

**Moment magnitude**

Earthquakes occur when the ground ruptures. Stresses build up over time (usually caused by the slow movements of tectonic plates) and eventually a piece of the Earth’s brittle crust deep under ground breaks (the technical term is ruptures). This rupture then grows until eventually a large area has shifted (the rupture propagates at a velocity of 2–3km/sec). The magnitude of the earthquake is related to the size of the rupture.

Seismic moment ($Mo$) = $\mu$ * rupture area * slip length

where $\mu$ is the shear modulus of the crust (approx $3 \times 10^{10}$ N/m)

Moment magnitude ($Mw$) = $2/3\log(Mo) - 6.06$

Nowadays the moment magnitude scale is the one used by seismologists to measure large earthquakes. The historic Richter magnitude is calculated by measuring the deflection on a seismometer corrected for distance from the event. Richter magnitudes underestimate the size of large events and are no longer used. However the constants used in the definition of Moment magnitude ($Mw$) were chosen so that the magnitude numbers for Richter and Moment magnitudes match for smaller events. For the largest events (the $Mw$ 9.3 event on Dec 26th 2004) the rupture area can be 1200 km long by 100 km deep with a slip length of up to 15 m (it had a seismic moment of $1.1 \times 10^{23}$ Nm).

**How can an earthquake have a negative magnitude?**

Very small events (eg. If $2/3\log(Mo)<6.0$ or if $\log A + 2.56 \log D < 1.67$) will have a magnitude less than zero. In practice earthquakes this small, although quite numerous, are usually only recorded and located in very small scale studies (e.g. studying rockbursts in underground mines).

**How is energy related to magnitude?**

Seismologists have determined that the energy radiated by an earthquake is a function of both the amplitude of the waves and the duration of the earthquake. A very small earthquake is over in less than a second while for the largest events the fault may continue to slip for more than 5 minutes.

For each unit increase in magnitude $M$, the amplitude increases by a factor of 10.

Empirical studies have found that:

Energy is proportional to $10^{(1.5M)}$
Consider the energy (E1) from a magnitude M and from (E2) from magnitude M+1

\[
\frac{E_2}{E_1} = \frac{10^{1.5M + 1.5}}{10^{1.5M}}
\]

Thus, for each unit increase in magnitude, the energy increases by a factor of 32.

For two units of magnitude, the increase is a factor of 10^3 or one thousand.

**Seismic energy:**

Both the magnitude and the seismic moment are related to the amount of energy that is radiated by an earthquake. Richter, working with Dr. Beno Gutenberg, early on developed a relationship between magnitude and energy. Their relationship is:

\[
\log E = 4.8 + 1.5M
\]

giving the energy E in joules from the magnitude M.

<table>
<thead>
<tr>
<th>Earthquake energy as a function of magnitude</th>
</tr>
</thead>
<tbody>
<tr>
<td>Magnitude</td>
</tr>
<tr>
<td>---------</td>
</tr>
<tr>
<td>-3.0</td>
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<tr>
<td>-2.0</td>
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<td>8.0</td>
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<td>9.0</td>
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</tbody>
</table>

Note that E is not the total ‘intrinsic’ energy of the earthquake, transferred from sources such as gravitational energy or to sinks such as heat energy. It is only the amount radiated from the earthquake as seismic waves, which ought to be a small fraction of the total energy transferred during the earthquake process.

More recently, Dr. Hiroo Kanamori came up with a relationship between seismic moment and seismic wave energy.

\[
\text{Energy} = \frac{\text{Moment}}{20 000}
\]

For this moment is in units of Nm, and energy is in units of joules.