THE SEISMICITY OF THE EARTH

OUTLINE

The earthquake source

Types of earthquake:

- from impacts and minor tremors to tectonic earthquakes, elastic rebound theory, deep earthquakes.

Location of earthquakes:

- definition of focus, epicentre, focal depth, origin time;
- determining the epicentre and focal depth of an earthquake.
- worked example

Earthquake magnitude:

- surface–wave, body–wave and local magnitude scales;
- magnitude and energy, magnitude and frequency of occurrence.

Source mechanism:

- faulting, the double couple source, and initial P–wave motion;
- fault–plane solutions – procedure, exercise and examples.

Geographical distribution of earthquakes

Circum–Pacific and Alpine–Himalayan belts;
Wadati–Benioff zones on subducting plates;
Mechanisms at mid–ocean ridges, rifts, transforms, subduction and collision zones.

Background reading: Fowler §4.2, Lowrie §3.5

EARTHQUAKES

Terminology

- Seismos = Greek for earthquake.
- Seismometer (also seismograph) = instrument designed to record seismic waves.
- Seismograph = recording of seismic waves from a seismometer.
- Seismicity = seismic (earthquake) activity.

Types of earthquake

Seismograph stations regularly record minor tremors from artificial explosions, impacts, landslides and rockbursts. Those near volcanoes record shocks as precursors to eruption and from volcanic eruptions and explosions.

Genuine earthquakes are tectonic in origin and are accompanied by a release of tectonic stress. They are distributed most densely near the boundaries of tectonic plates.

- Major earthquakes are sometimes preceded by *foreshocks*.
- They are often followed by a sequence of *aftershocks*.
- *Earthquake swarms* are longish series of shocks with no major event; they are frequently associated with volcanic activity.
- Although *intraplate* earthquakes are rare, they are generally accompanied by a larger than normal release of stress and can be surprisingly large.

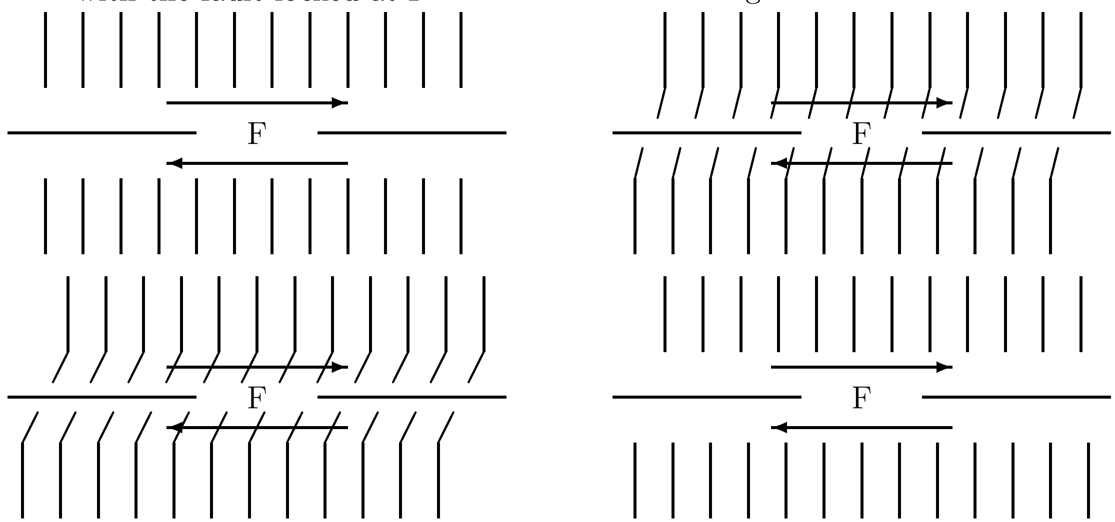
Earthquakes are not the sole means of relieving tectonic stresses; anelastic creep (plastic flow) is commonly observed in many areas. Although slip directions from earthquake focal mechanism studies provide estimates of the direction of tectonic stresses, earthquakes underestimate the amount of deformation from those stresses over a period of time.

ELASTIC REBOUND THEORY

This is the generally accepted mechanism for *shallow* earthquakes. It is driven by the build–up of strain energy from tectonic forces which try to move one crustal block relative to another. The blocks are separated by an active fault or fault system.

initial state of the two blocks with the fault locked at F

build up of strain from regional shear stress



stress exceeds strength of rock at F causing fracture

distorted blocks snap back towards stress–free state

The difficulty with elastic rebound theory is that, at the huge pressures found at the depth of the lower crust and beyond, the frictional forces on a fault would exceed the shear strength of the rock: the rock would shatter before it would slide.

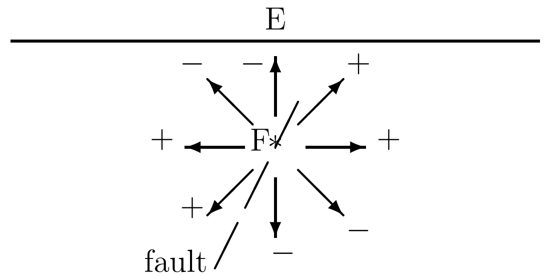
Mechanisms suggested for intermediate and deep earthquakes are:

1. fluid pore pressure: this reduces the effective normal stress on the fault and hence reduces the friction across the fault, i.e. lubrication of the fault by pore pressure allows sliding to occur at depth; pore pressures well in excess of hydrostatic pressure, from, say, confined fluids released by metamorphism, are necessary to achieve this;
2. unstable creep which generates heat and melting, and accelerates to sudden shearing;
3. collapse of a rock volume and release of tectonic stress due to a sudden change of phase.

EARTHQUAKE LOCATION

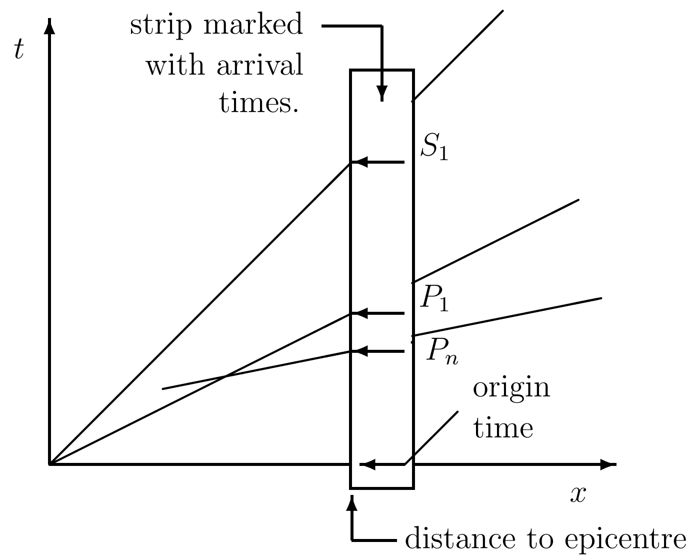
Definitions

The *focus* or *hypocentre* F is the point at which the fracture originates. The first seismic waves are radiated from F.
The *epicentre* is the point vertically above the focus F.
The focal depth is the vertical distance EF from the focus to the epicentre.
The origin time is the time at which fracture began.



Finding the epicentre of a local earthquake

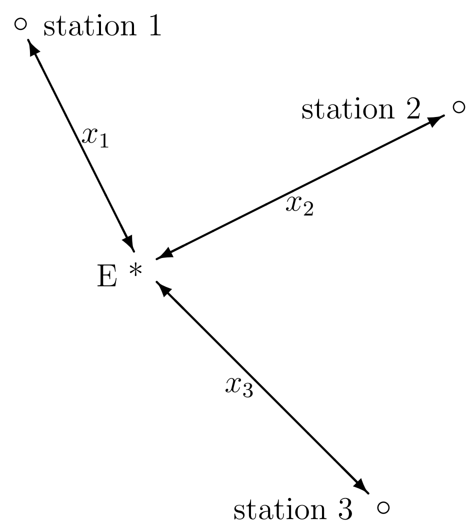
1. Read the arrival times of events on the seismograms of the earthquake.
2. Fit the arrival times to the region's travel time curves.
3. Estimate the origin time and and distance x_i of the recording station from the epicentre from the best fit of the observed times and the travel time curves.
4. Repeat for seismograms from the other stations.
5. Draw a circle of radius x_i around each station.



EARTHQUAKE LOCATION (CTD)

Finding the epicentre of a local earthquake (ctd)

The intersection of the circles gives an initial estimate of the epicentre. The average of the origin times from each station gives an initial estimate of the earthquake's origin time. The origin time and coordinates of the epicentre can be refined by computer. Focal depth can be included in the calculations. More weight is given to initial P–wave readings than to later phases in locating the epicentre, although later phases are often important in establishing focal depth.



The worked example that follows illustrates the application of this procedure.

WORKED EXAMPLE: LOCATING A NEAR EARTHQUAKE

On 24 February 1965 a magnitude 3 earthquake occurred in the vicinity of Robertstown in South Australia. It was recorded at the seismograph stations at Adelaide (ADE), Hallet (HTT) and Cleve (CLV). The table below gives the arrival times of the first arrival and clear later arrivals at these stations. The direct crustal P-wave is denoted by P_1, the direct crustal S wave by S_1, and the phase P_n is the sub-Mohorovicic P head wave.

Station		P_1 arrival time	S_1 arrival time
HTT		$16^h36^m57.5^s$	$16^h37^m04.3^s$
ADE		$16^h37^m06.6^s$	$16^h37^m21.0^s$
CLV		$16^h37^m25.0^s$	$16^h37^m52.3^s$
	(P_n)	$16^h37^m22.0^s$	

WORKED EXAMPLE: P- AND S-WAVE TRAVEL TIMES

The map on the following page gives the locations of the three seismograph stations. The map should plot at a scale of 5 cm $\equiv$ 100km. To check that it has, measure the length of one degree of latitude; it should measure 55.6 mm (1° latitude = 111.2 km).

The velocity of the P_1 phase in this part of South Australia is 6.23 km/s and the velocity of the S_1 phase is 3.58 km/s. The velocity of the P_n phase is 8.05 km/s and the depth to the Moho is 38 km. The travel time (in s) of the P_n phase is given by the equation:

$$t(P_n) = 7.7 - 0.10h + \frac{\Delta}{8.05}$$

where h is the focal depth in km and Δ is the epicentral distance in km.

The following pages show

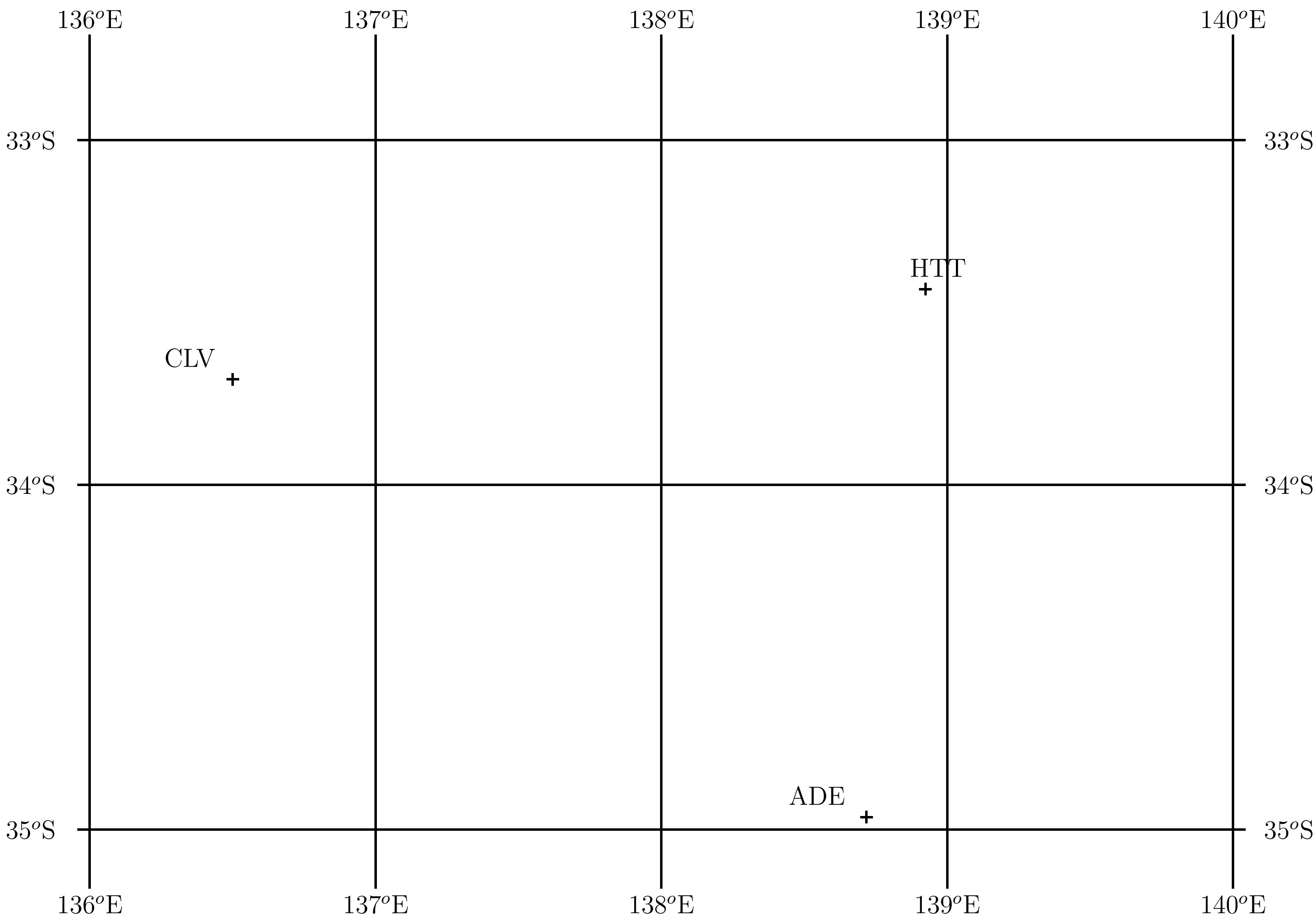
- how to find the origin time and the epicentre of the earthquake,

and

- how to estimate the focal depth of the earthquake.

GEOPHYSICS (08/430/0012)

LOCATING A NEAR EARTHQUAKE: MAP SHOWING LOCATION OF SEISMOGRAPH STATIONS



LOCATING A NEAR EARTHQUAKE: METHOD

There are various ways of locating the epicentre. In this case it can be estimated from the P_1 and S_1 arrival times.

1. A graphical method:
   - Draw the travel-time graphs for the P_1 and S_1 phases;
   - Mark the P_1 and S_1 times at each station on a strip of paper, using the same time scale as the travel time graphs;
   - Match the marked times with the graph to obtain estimates of epicentral distance and origin time.
2. A numerical method:
   - Calculate the relation between epicentral distance and the difference in travel-time $t(P_1) - t(S_1)$;
   - Calculate the relation between P_1 travel time and the difference in travel-time $t(P_1) - t(S_1)$;
   - Estimate epicentral distance and origin time from the P_1 and S_1 times.

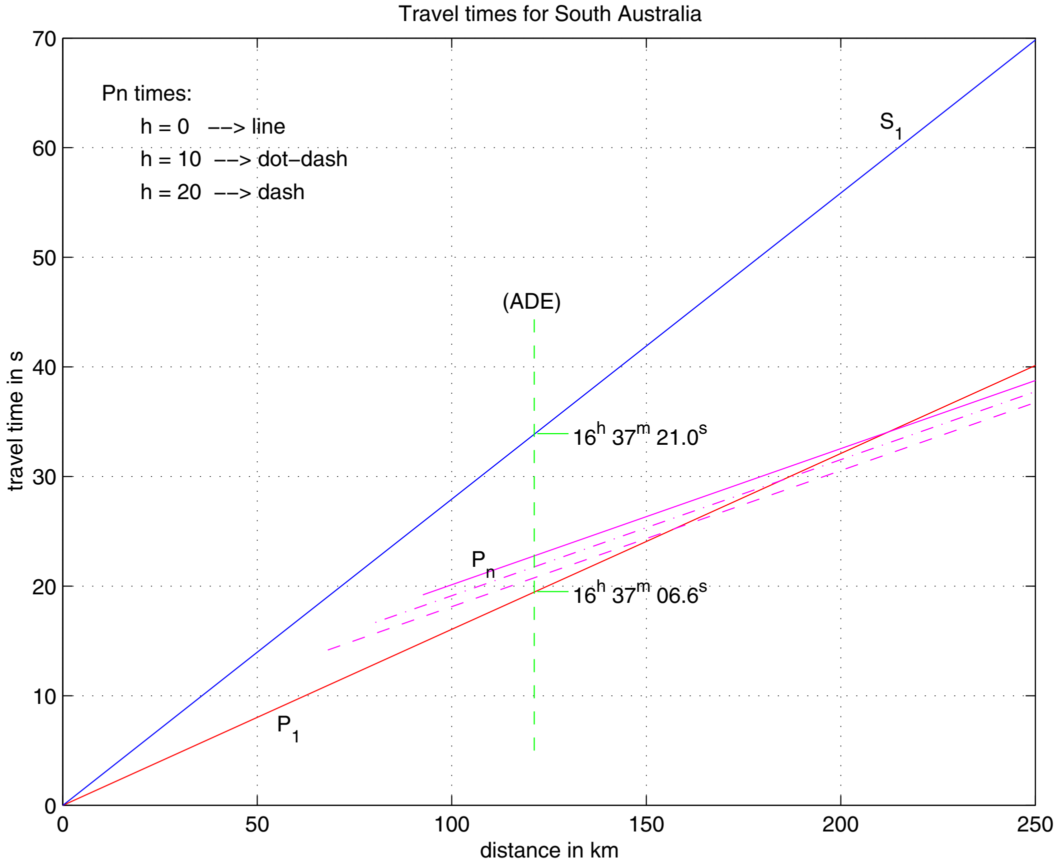
The epicentre is found by drawing circles of radius equal to the epicentral distance around each seismograph station. These are unlikely to intersect at a point. The epicentre is estimated as the centre of the area formed by the intersecting circles.

A refinement of method 2 would be to average the estimates of origin time from the three stations. The P_1 travel times could then be calculated from the observed arrival times and the average origin time and the epicentral distances calculated from these P_1 travel times. The advantage from doing this is that the P-wave arrival times can usually be measured more accurately than S-wave arrival times. In practice the epicentre and origin time is normally found by least squares fit to the P-wave arrival times. Additional phases are used for focal depth estimation.

LOCATING A NEAR EARTHQUAKE: TRAVEL TIME PLOTS

Travel times for P_1, S_1 and P_n

The P and S times at ADE are shown indicating the distance where their difference equals the travel time difference $t(S_1) - t(P_1)$ from the computed travel times. The travel times of P_n are plotted for focal depths of 0, 10 and 20 km. Why do the P_n plots start at different distances?



LOCATING A NEAR EARTHQUAKE: RESULTS

The travel time equation for P_1 is:

$$t(P_1) = \frac{\Delta}{6.23} = 0.1605\Delta$$

where Δ is epicentral distance (distance from source to receiver). The travel time equation for S_1 is:

$$t(S_1) = \frac{\Delta}{3.58} = 0.2793\Delta$$

Hence

$$t(S_1) - t(P_1) = 0.1188\Delta$$

or

$$\Delta = 8.416[t(S_1) - t(P_1)]$$

and

$$t(P_1) = 1.351[t(S_1) - t(P_1)]$$

The following table can be drawn up using these equations.

Station	$t(S_1) - t(P_1)$	Δ (km)	Calc. $t(P_1)$	origin time	Obs. $t(P_1)$	Δ (km)
HTT	6.8 s	57.2	9.2 s	$16^h36^m48.3^s$	9.7 s	60.4
ADE	14.4 s	121.2	19.5 s	$16^h36^m47.1^s$	18.8 s	117.1
CLV	27.3 s	229.8	36.9 s	$16^h36^m48.1^s$	37.2 s	231.8

LOCATING A NEAR EARTHQUAKE: RESULTS (CTD)

Epicentre and origin time

The observed P_1 times were calculated using the average origin time $16^h 36^m 47.8^s$. The epicentre is found from the intersection of the circles drawn around the seismograph stations with radii equal to the epicentral distances estimated in the table above. The circles intersect in a small triangle. The epicentre estimated from the centre of this triangle is 33.95^o S , 138.95^o E . The degree of agreement between the intersections gives some idea of the accuracy of the epicentre location.

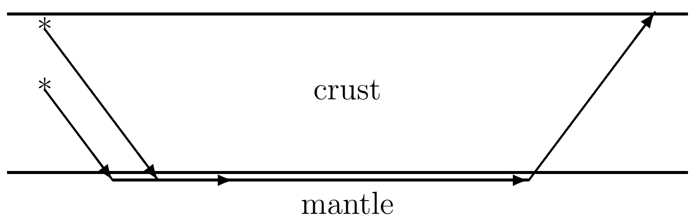
Focal depth

The computed P_n travel time (in s) to CLV is:

$$t(P_n) = 7.7 - 0.10h + \frac{231.8}{8.05} = 36.5 - 0.10h$$

The observed P_n travel time is 34.2s, 2.3s less than 36.5s. Hence $h = 2.3/0.10 = 23$ km.

Ray diagram



Waves from the deeper earthquake spend less time in the crust and have more of their travel path in higher velocity material.

The P_n travel time decreases with focal depth because the closer the focus is to the Mohorovicic discontinuity, the shorter P_n ray path is overall. Moreover the extra path in the mantle is more than compensated by the shorter crustal segment from the focus.

EARTHQUAKE MAGNITUDE (1)

Earthquake magnitude is a rough but useful overall measure of the size of an earthquake. It is related to the energy released by the earthquake. An earthquake of magnitude 7 releases about a million (10^6) times as much energy as an earthquake of magnitude 3. To accommodate the huge range of energy release, a logarithmic scale is used.

Definitions

• The original scale was developed by Richter (1935) to provide a standard rating for magnitudes of local earthquakes in southern California. It was:

$$M_L = \log_{10} A_{WA}$$

where A_{WA} is the maximum amplitude in microns ($1\mu = 10^{-6}$m) recorded on a Wood–Anderson torsion seismometer at a distance of 100 km from the epicentre.

• More generally, the magnitude M of an earthquake is defined as:

$$M = \log_{10} \frac{A}{T} + f(\Delta, h) + C$$

where A is the maximum amplitude in microns, usually of a specified seismic phase, whose period at the maximum amplitude is T. $f(\Delta, h)$ is a function of epicentral distance Δ and focal depth h which compensates the reduction in amplitude with distance due to geometrical spreading and absorption of the seismic wave. C is a constant that includes a station correction, which adjusts for localised effects of the recording site on amplitudes.

The epicentral distance Δ is normally measured in geocentric degrees; i.e. Δ is the angle subtended by the epicentre and recording station at the centre of the Earth.

EARTHQUAKE MAGNITUDE (2)

DEFINITIONS (CTD)

• *Body wave magnitude* m_b uses the maximum amplitude in microns of P–waves with period T of about 12 s:

$$m_b = \log_{10}\frac{A}{T} + 0.01\Delta + 5.9 \quad (\Delta \text{ in degrees})$$

m_b rarely exceeds 6.0 because the 12 s period is outside the dominant range of P–wave periods radiated by earthquakes larger than this.

• *Surface wave magnitude* M_S is based on the maximum amplitude in microns of Rayleigh waves with period T of about 20 s:

$$M_S = \log_{10}\frac{A}{T} + 1.66\log_{10}\Delta + 3.3 \quad (\Delta \text{ in degrees})$$

It is not suitable for deep earthquakes, large epicentral distances, and very large earthquakes. M_S saturates around 8.0 because the main surface waves radiated from very large earthquakes have periods longer than 20 s.

• *Moment magnitude* M_W is based on the *seismic moment* M_0 of an earthquake:

$$M_W = \frac{2}{3}\log_{10} M_0 - 10.73$$

M_0 is directly proportional to the seismic energy radiated by an earthquake. It is also related to the *stress drop* during the earthquake.

The advantage of M_W is that, unlike M_S and m_b, it does not saturate. It is therefore the best measure of magnitude of large earthquakes. The calculation of M_W requires an analysis of the mechanism of the earthquake.

• *Local magnitude* scales can be set up in earthquake–prone areas specifically for those small earthquakes observed only on the local seismograph network.

M_S and m_b require teleseismic recordings and are intended for earthquakes large enough to be observed world–wide.

EARTHQUAKE MAGNITUDE RELATIONSHIPS

Relation betwen magnitude scales

Gutenburg & Richter (1954) give $m_b = 2.9 + 0.56M_S$.
Richter (1958) and Bath (1966) give $m_b = 2.5 + 0.63M_S$.
Local magnitude scales are usually designed to correspond with m_b or M_S. They are tied to m_b or M_S by means of those earthquakes in the area that are recorded at least on a wide regional scale.

Magnitude and energy

Gutenburg & Richter (1954) give $\log_{10} E = 4.8 + 1.5M_S$.
Bath (1966) gives $\log_{10} E = 5.24 + 1.44M_S$.
where E is the energy in joules released as seismic radiation.
According to the first equation an earthquake of magnitude $M_S = 8.8$ releases 10^{18} joules as seismic wave energy. For comparison:

- underground nuclear explosions have magnitudes of about 7, corresponding to a seismic energy release of 2×10^{15} joules;
- the annual energy consumption of the U.S.A. is about 10^{19} joules;
- the annual flow of heat from the Earth is about 10^{21} joules.

Only a small fraction of the energy released in an earthquake goes into seismic waves.

Magnitude and earthquake frequency

Earthquake frequency is measured by $N(M)$, the number of earthquakes within a fixed time interval in a given area whose magnitude exceeds M. It is found that

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

where a and b vary with locality and type of sequence (normal seismicity, swarm, aftershocks). Typically $b \sim 0.6$ to 1.0.

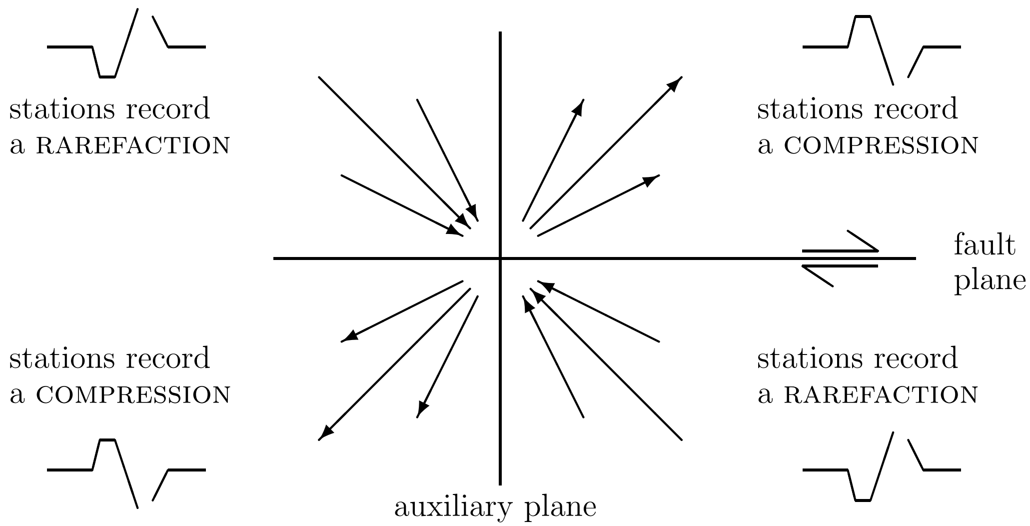
EARTHQUAKE FOCAL MECHANISM (1)

Initial P–wave motion from an earthquake

The first wave recorded from an earthquake is a P–wave:

- if the ray path to the recorder leaves the focus in a direction along which the initial ground motion is outwards, then the initial recorded motion is a *compression*;
- if the ray path leaves the focus in a direction along which the initial ground motion is inwards, then the initial recorded motion is a *rarefaction (or dilatation)*.

Thus the initial P–wave motions follow a four–lobed pattern with alternating quadrants of compressions and dilatations.



The plane perpendicular to the *fault plane* and perpendicular to the fault motion is called the *auxiliary plane*. The *slip vectors* are perpendicular to the auxiliary plane: the slip vectors give the initial motion on each side of the fault plane.

EARTHQUAKE FOCAL MECHANISM (2)

P–wave radiation pattern

The P–wave amplitudes are largest at $\sim 45°$ to the fault plane. There are P–wave nulls on the fault plane and auxiliary plane where equal and opposite motions cancel. These nulls separate the compressional and dilatational quadrants.
Note: An earthquake from a fault along the auxiliary plane, with shear motion in the opposite rotational sense, would generate the same pattern of first P–wave (and S–wave) motions.

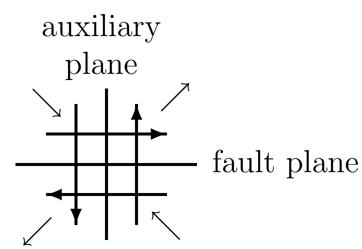
S–wave motion from an earthquake

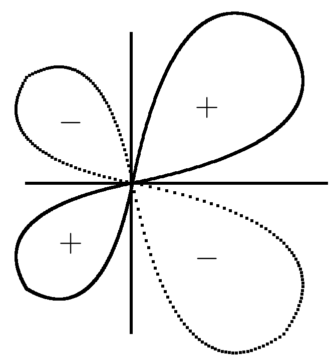
The amplitudes of S–waves also follow a quadrantal pattern. They have nulls at $\sim 45°$ to the fault plane and are largest on the the fault plane and auxiliary plane. The initial S–wave motion is usually obscured by earlier wavetrains.

Equivalent stress field

The stress field from the shearing motion on a fault corresponds to a double couple (or double torque): there is no net rotation of rock at the fault. The stress field is also equivalent to a compression and extension perpendicular to each other and at $\sim 45°$ to the fault.

The strongest positive amplitudes are radiated along the tensional axis of the stress field





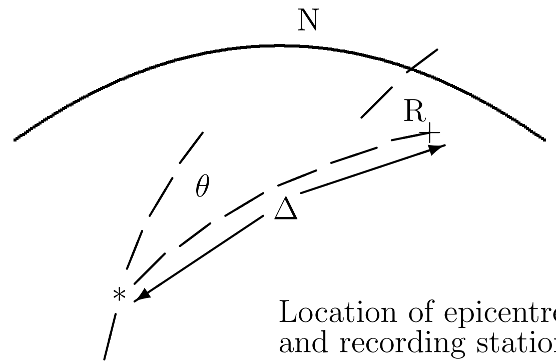
P–wave radiation pattern

FAULT PLANE SOLUTION FROM FIRST MOTION ANALYSIS

The fault plane and auxiliary plane can be inferred from the first P–wave motions of an earthquake recorded at a good spread of seismograph stations. Vertical seismographs are calibrated such that a compression appears as an upward ground motion and a dilatation as a downward motion. Knowledge of the Earth's velocity–depth profile allows the motion recorded at any station to be tracked back to the angle at which it left the source. The usual procedure is to track the first motions back to a hypothetical sphere, called the focal sphere, and to project the motions on the sphere (stereographically) onto the horizontal plane through the focus.

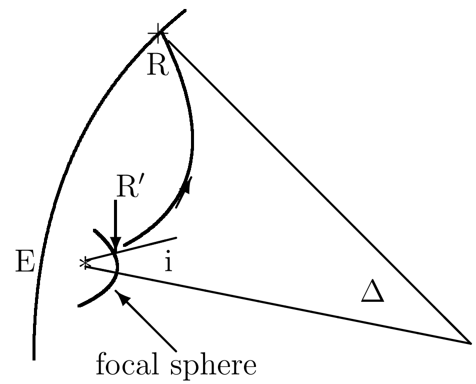
PROCEDURE FOR FINDING A FAULT PLANE SOLUTION FROM P-WAVE FIRST MOTIONS

1. The epicentre E and focal depth of the earthquake are determined from travel time data. The geocentric distance Δ and azimuth θ of each station R from the epicentre are calculated.



Location of epicentre and recording station

2. The angle i to the vertical at which the first arrival leaves the focus is found from seismological tables giving i as a function of focal depth h and distance Δ.



Cross section through the great circle joining E and S.

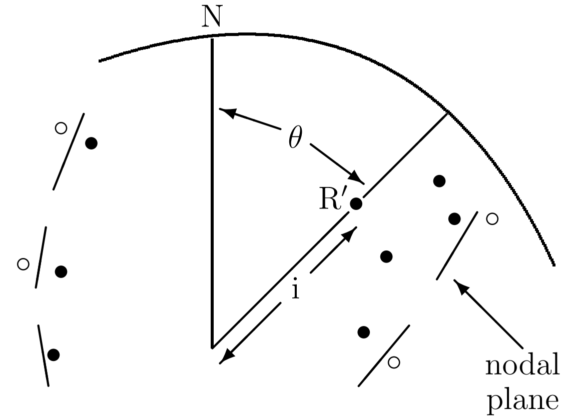
3. The azimuth θ and angle of incidence i define the point R′ at which the ray to the recording station leaves the focal sphere.

PROCEDURE FOR FINDING A FAULT PLANE SOLUTION FROM P-WAVE FIRST MOTIONS (CTD)

4. The first motion is plotted on a projection of the focal sphere onto a horizontal plane (e.g. using black dots for compressions and open circles for dilatations).

5. The nodal planes (fault plane and auxiliary plane) are constructed to give the best separation of compressions and dilatations. These nodal planes must be constucted to be at right angles to each other; i.e. each must pass through *pole* of the other. The pole of a plane is the point where the normal to the plane cuts the focal sphere; it is therefore 90° from the nodal plane along the great circle perpendicular to the nodal plane. (See the focal mechanism exercise).

6. The direction of the slip vectors can be determined provided the fault plane can be distinguished from the auxiliary plane. Extra information, from knowledge of the tectonics of the area, is needed to distinguish the fault plane from the auxiliary plane. The slip vector is then the normal to (the pole of) the auxiliary plane.



Lower hemisphere projection of the focal sphere

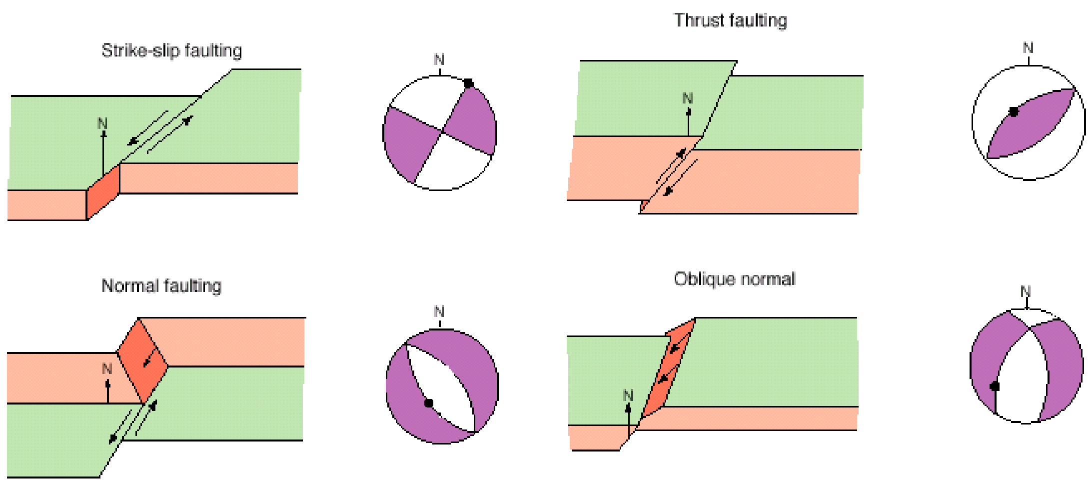
FAULT PLANE MECHANISMS

Earthquake fault plane mechanisms are plotted on maps as 'beach ball' diagrams. These are small stereographic projections of the lower focal sphere on which are drawn the two nodal planes. Segments of the focal sphere containing P-wave compressions are shaded in; segments containing P-wave dilatations are left blank. There are three basic types corresponding to normal, thrust and strike-slip faulting:

- alternate segments joining close to the centre = strike-slip faulting;
- blank segment through the centre = normal faulting;
- shaded segment through the centre = thrust faulting;

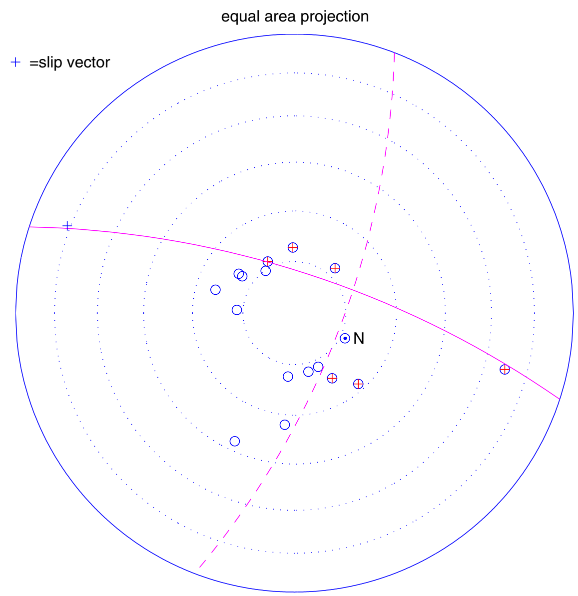
These three types are the end members of a range of fault plane mechanisms. Most fault plane mechanisms are dominantly but not purely one of these three types. The figure overleaf illustrates the three basic types and a normal fault with a strike-slip component.

SCHEMATIC DIAGRAM OF FAULT PLANE MECHANISMS



EXAMPLES OF FAULT PLANE SOLUTIONS (1)

Unless they are recorded at short distances ($<\sim 2^\circ$), first arrivals correspond to refracted P–waves that leave the focus in a *downward* direction. Consequently it is first motions on the *lower* half of the focal sphere that are displayed in fault plane solutions. It is helpful to remember that the projection looks down into the Earth when interpreting a fault plane solution: black dots represent compressions, and open circles dilatations, leaving the focus downwards along raypaths into the Earth. The figure below shows the P-wave first motions from the 1975 Haicheng earthquake plotted on an equal area projection of the lower focal hemisphere and the inferred fault plane solution. Open circles denote dilatations, + signs compressions, and N a point on a nodal plane. The data come from Cipar (1979).



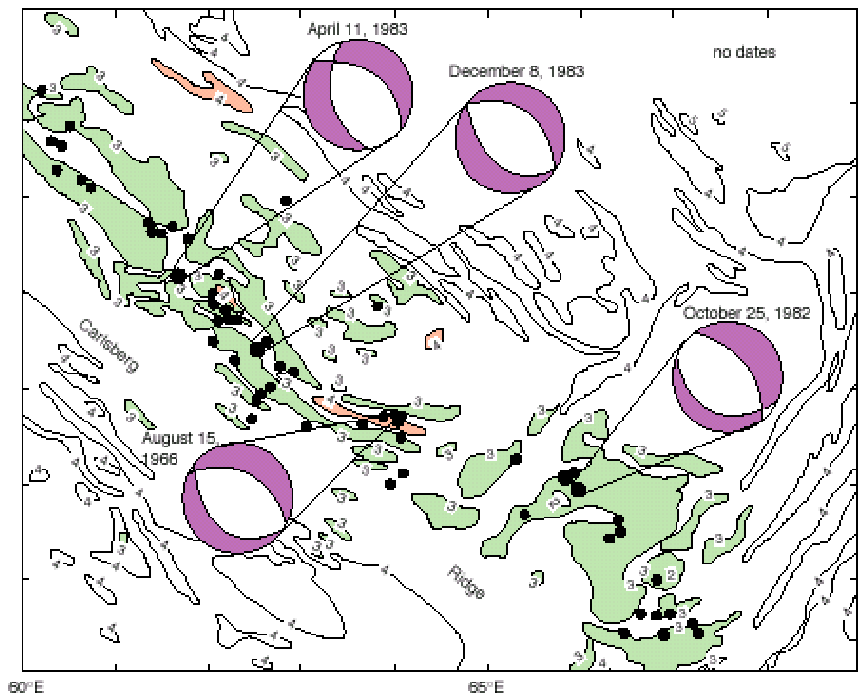
Reference: J.Cipar, 1979. Source processes of the Haicheng, China, earthquake from observations of P and S waves, Bull.Seism.Soc.Am. **69**, 1903–1916.

EXAMPLES OF FAULT PLANE SOLUTIONS (2)

Normal faulting

This is common at spreading ridges, in rift valleys, and in regions of crustal extension. There is a zone of dilatation through the centre of the projected first motions. The figure below shows the fault plane mechanisms from four earthquakes on the Carlsberg Ridge.

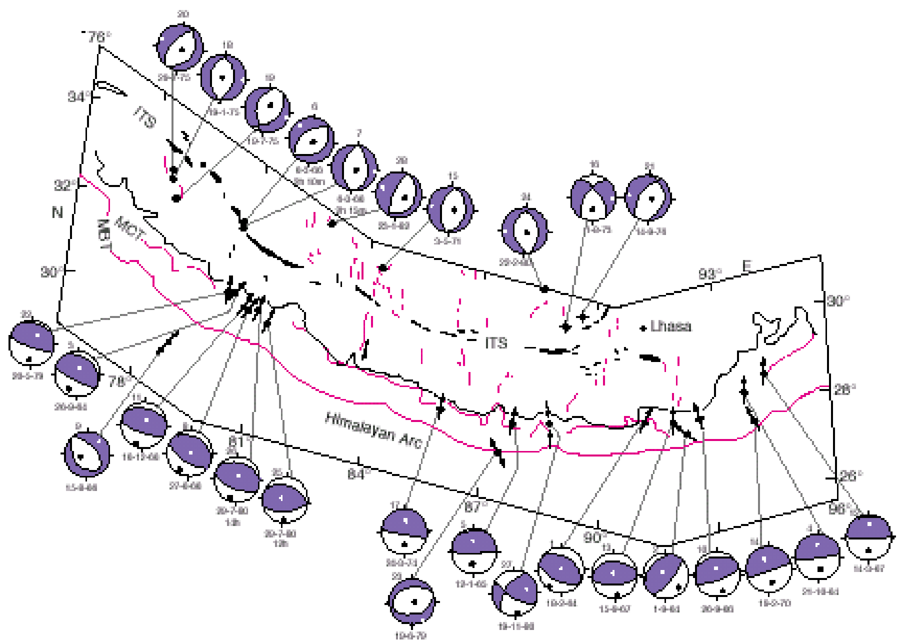
Reference: P.Huang & S.Solomon, 1987. Centroid depths and mechanisms of mid-oceanic ridge earthquakes in the Indian Ocean, Gulf of Aden and Red Sea, J.Geophys.Res. **92**, 1361-1382.



EXAMPLES OF FAULT PLANE SOLUTIONS (3)

Thrust (reverse) faulting

This is found at subduction and collision zones and in regions having a compressional stress regime. The projected first motions show a quadrant of compressions through their centre.

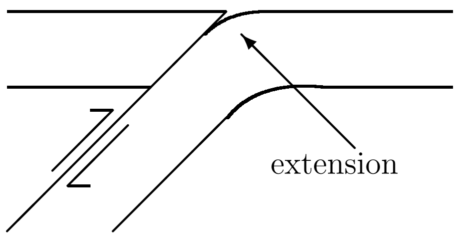


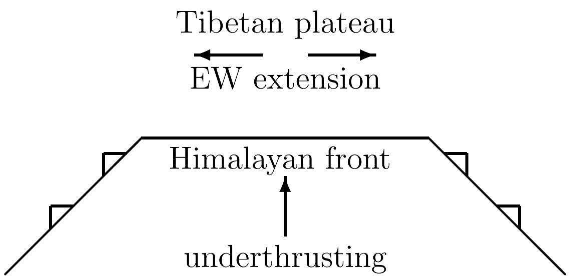
Reference: J.Ni & M.Barazangi, 1984. Seismotectonics of the Himalayan collision zone: Geometry of the underthrusting Indian plate beneath the Himalaya, J.Geophys.Res. **89**, 1147-1163.

EXAMPLES OF FAULT PLANE SOLUTIONS (3)

Thrust faulting (ctd)

The seismicity of subduction and collision zones is not straightforward. The outer bend of a subducting plate gives rise to extensional stresses which are characterised by normal faulting. The Himalayan thrust front is succeeded to the north by east–west extension in the Tibetan plateau.





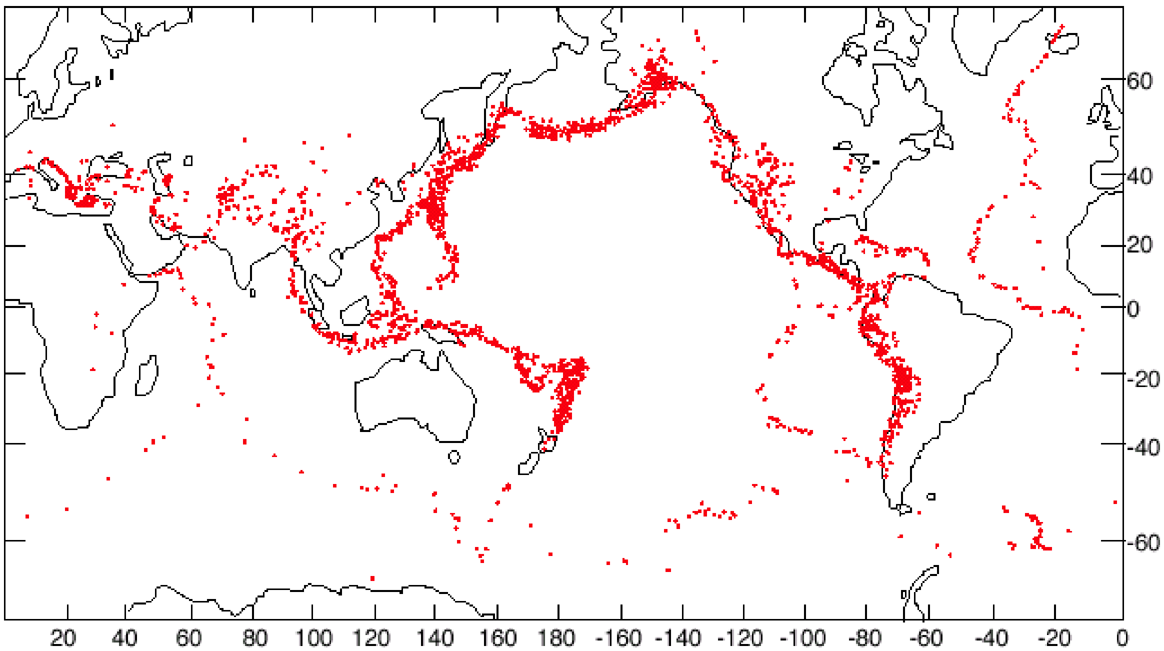
Strike–slip faulting

In this case the fault plane is close to vertical and the slip vector more or less horizontal. The quadrantal pattern is very clear since it is viewed along the axis of intersection of the fault and auxiliary planes. The fault plane solution from the 1979 Haicheng earthquake shown earlier was dominantly strike slip, with a small normal faulting component.
Earthquakes from transform faults provide obvious examples.
For examples see Figures 8.26 (mid-Atlantic) and 8.42 (Queen Charlotte Islands transform) in Fowler.

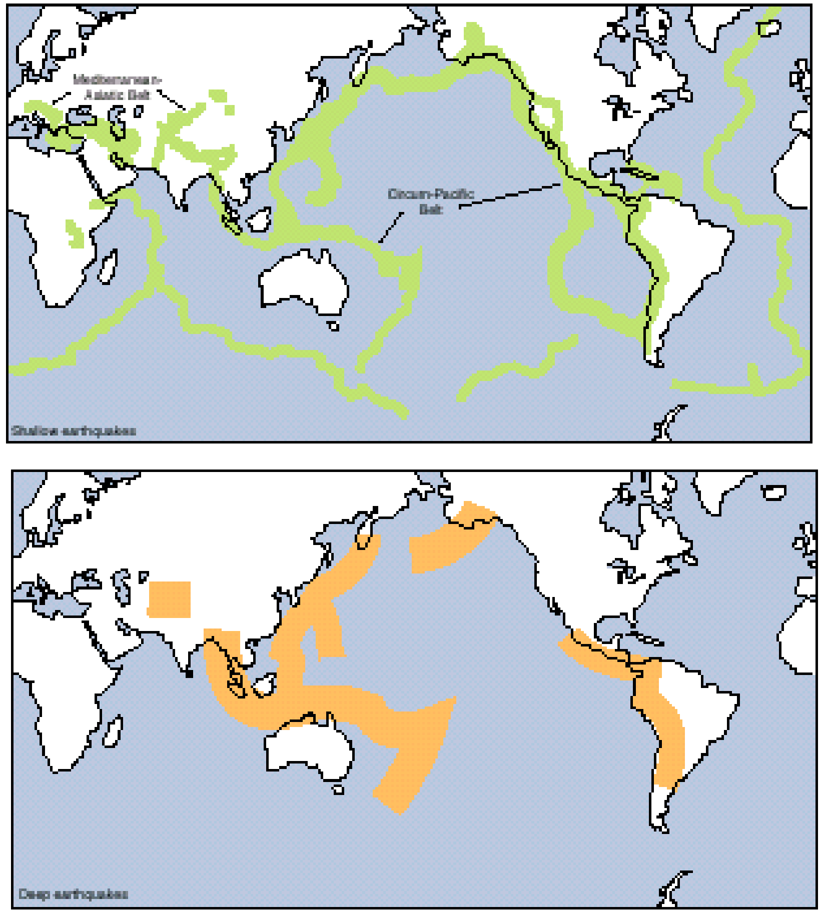
GEOGRAPHICAL DISTRIBUTION OF EARTHQUAKES

The energy release from shallow ($0-70$km) earthquakes is 75% of the total. That from intermediate ($70-300$km) earthquakes is 22% of the total. The energy release from deep ($300-700$km) earthquakes is 3% of the total.
Of the shallow earthquakes, 75% of the energy release comes from the circum–Pacific belt; 23% from the Alpine–Himalayan belt and just 2% from the rest of the world. Deep earthquakes occur mainly in the circum–Pacific belt and up through Burma and the Himalaya, although they have been found to occur, for example, beneath Italy.

Global distribution of seismicity



DISTRIBUTION OF SHALOW AND DEEP EARTHQUAKES



DEPTH DISTRIBUTION OF EARTHQUAKES IN SUBDUCTION ZONES

Almost all deep earthquakes are associated with subduction zones. Plots of earthquake foci on vertical cross sections show that the foci fall on narrow zones, called Wadati–Benioff zones. Wadati–Benioff zones dip at an angle of about 45° and coincide with velocity anomalies that mark the outline of relatively cold and brittle downgoing lithospheric slabs. Accurate hypocentre location shows that the earthquakes in Wadati–Benioff zones often follow two trends, the upper one being close to the top edge of the subducting plate, as illustrated in the cross section below of the subducting plate beneath Japan.