

5. The Internal Structure of the Earth

The Solid Earth

Time-scales for changes in the crust

- Short-term: earthquakes, volcanic eruptions, landslides
- Medium term (10^3 years): melting of glaciers, uplift
- Long-term (10^7 years): appreciable horizontal movement of the crust. Sea floor spreading.
- Very long term (10^8 - 10^9 years): Production of crustal motions, mountains and continents formed.

Below the crust the density is seen to vary with depth, but also with longitude. We see the crust subject to external forces like levelling down/production of sediments, and to internal forces which produce, for example, uplift (the evidence being sea-shells on the tops of mountains). There has been a long history of debate between those who believe in "catastrophic Processes" and those who believe that geological change has happened gradually. The latter seem most likely to be correct the majority of times. This philosophy can be summarised by the the "Principle of Uniformity of Process" as described by James Hutton (1726-1797).

Crust rocks and soil

The rocks and material of the crust are of four types:

- sedimentary
- igneous
- metamorphic
- magma

Metamorphic rocks (e.g. sl Metamorphic rocks (e.g. slate) are usually found deep inside mountains, and not too often at the surface. Most common rocks are igneous, and much of the surface is sedimentary.

Sediments are:

- Clastic (consisting of broken up pieces) in order of decreasing particle size: gravel, sand (sandstone), silt and clay (silt and clay constituting shales).
- Non-Clastic - mainly chemical deposits - salt, gypsum and coal for example.
- By biological deposition, e.g. limestone CaCO_3 and dolostone $\text{CaMg}(\text{CO}_3)_2$

These sediments are sometimes modified by pressure and temperature to become metamorphic rocks with different characteristics - e.g marble is limestone or dolostone after high T, p.

Elemental abundances in the crust.

	O	Si	Al	Fe	Ca	Na	K	Mg	H
% wt	46.6	27.7	8.1	5.0	3.6	0.8	2.6	2.1	0.14

Classifying minerals

A Mineral is a naturally occurring substance which consists of a conglomeration of crystals - individual crystals being called "grains". The size of the grains tells you much about the environment in which the crystals formed: small grains indicate cooling was rapid (e.g lava - surface cooled magma - has no evidence of structure at all), large that it cooled slowly, giving time for the crystals to build in size.

Classifying minerals can be done on erals can be done on a number of ways. One is by the number of oxygen atoms contained per molecule. That is, take ratio:

$$(\text{No. of metal atoms}) / (\text{No. of oxygen atoms})$$

Four ratios cover most of the common minerals: 12/24; 15/24; 16/24 and 18/24, that is, 1/2, 5/8, 2/3 and 3/4.

The 1/2 ratio is represented largely by Silica, SiO_2

The 5/8 minerals include Feldspars $\text{MM}'\text{AlSi}_2\text{O}_8$ where M can be K, Na or Ca and M' Si or Al (Ca and Al tend to go together).

The 2/3 minerals include Pyroxenes $\text{MM}'\text{Si}_2\text{O}_6$ where M or M' are Ca, Mg or Fe generally, though you also get Al, Tg, Na and K.

The 3/4 minerals include Olivines $\text{MM}'\text{SiO}_4$ where M.M' are Fe or Mg. Mica also is included here.

Note that the density of these minerals increases as the ratio gets larger. Which ions go into which compounds depends on ion sizes, crystal structures etc.

Feldspar

M = K	M' = Si	Potassium Feldspar or Orthoclase
M = Na	M' = Si	Albite
M = Ca	M' = Al	Anorthite

You also get mixtures of the first two. Mixtures of albite and anorthite are called soda-lime feldspar or plagioclase

Ion	K ⁺	Na ⁺	Ca ²⁺	Si ⁴⁺	Al ³⁺	
radius	1.35	0.97	0.99	0.42	0.51	(10 ⁻⁹ m)

Pyroxenes

Continuous spectrum of mixture

Diopside spectrum of mixture

Diopside	$\text{CaMgSi}_2\text{O}_6$	$\text{Mg}_2\text{Si}_2\text{O}_6$	Enstatite
Heidenbergite	$\text{CaFeSi}_2\text{O}_6$	$\text{Fe}_2\text{Si}_2\text{O}_6$	Ferrosilite

Olivines

Another continuous spectrum of mixture

Forsterite	Mg_2SiO_4	Fe_2SiO_4	Faryalite
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density $3.21 \times 10^3 \text{ kg m}^{-3}$ $4.39 \times 10^3 \text{ kg m}^{-3}$

Density at the lower end of the table is typified by the Silica group - e.g. Quartz $2.65 \times 10^3 \text{ kg m}^{-3}$. The Pyroxenes and Olivines are more important with depth.

Crystallography

In any detailed study we would have to consider the crystallography of the rocks in detail as this gives insights into where and how they were formed. That is a different course, but we can illustrate some general principles with some simple examples. Thus, for example, we can look at the basic Silica Tetrahedron:



This has Silicon and Oxygen ions - Si^{4+} , O^{2-} - and so is overall $(\text{SiO}_4)^{4-}$.

In Olivines this silicate tetrahedron is found with Mg and Fe fitting between the tetrahedra. This gives a very compact structure.

In Pyroxenes we have the $(\text{Si}_2\text{O}_6)^{4-}$ grouping with two oxygen atoms shared between two tetrahedra - this gives chains of tetrahedra.

In Feldspars e.g. $\text{KAl}(\text{Si}_3\text{O}_8)$ we have groups of 2 double-tetrahedra (Si_4O_8) with one Si replaced by Al.

We can classify types of rock, as distinct from minerals, by i) density or ii) fabric. With density we would get quartz and feldspars as low density and pyroxenes and olivines as high density - i.e. silicon-rich and silicon-poor. We see that this fits in with our oxygen proportion and crystallographic classifications above. Similarly the "fabric" is equivalent to the grain-size/cooling history grading. If we plot % SiO_2 against ratio volcanic/plutonic origin then we get a familiar-looking chart with quartz in the top left-hand side and olivine in the bottom right-hand side.

The minerals in the earth's crust have been subject to fractionation due to the recycling of surface material in subduction caused by the processes of continental drift (see below). Thus crystals formed previously sink and re-melt when the crust is forced down into the mantle. This interchange of crust and mantle material causes in time the heavier material to sink and the lighter to rise: this is the reason the crust is lighter (e.g. more silicic (e.g. more silica, less Fe/Mg) than the material beneath it.

Phase transitions

e.g.	Quartz	--30kbar-->	coecite	--130kbar-->	stishovite
density	2.65		3.02		4.35
	(10 ³ kg m ⁻³)				

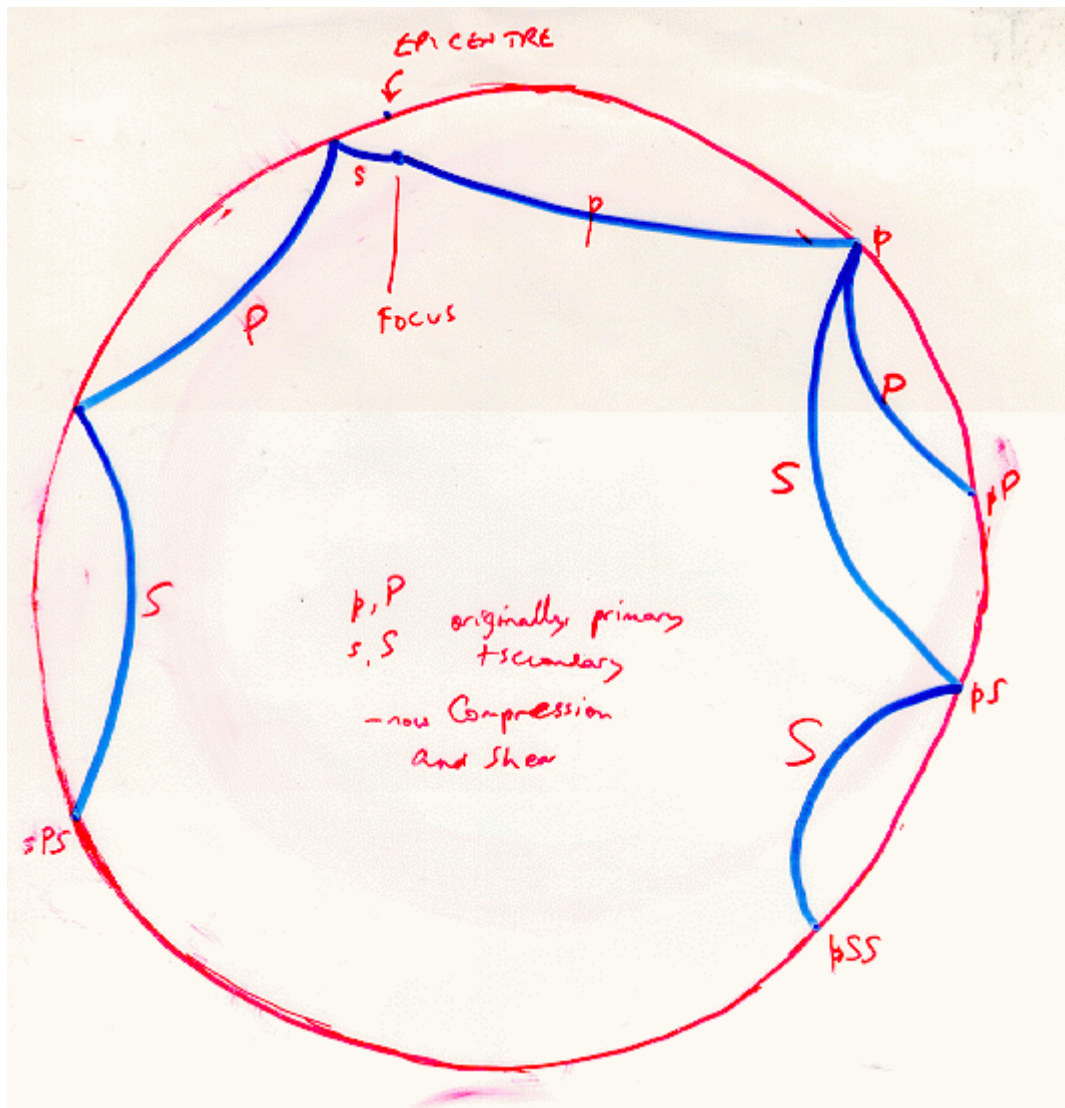
Close-packed oxygen is of two types of crystal structure - tetrahedral and octohedral. Oxygen atoms are larger than the other common atoms and so it is packing of O which tends to dominate the structure.

We can compare the proportions of minerals in the crust to that in chondritic meteorites as representatives of presumably undifferentiated (unfractionated) material from the relatively homogeneous formation of the original solar system material. We would expect to see more lighter material in the crust because of the fractionation:

In the interior of the earth not only are Fe and Mg more abundant, but so are ions of the same size (c. 0.08nm radius) such as Co, Ni and Mn. Elements which do not substitute (e.g. radii 0.1-0.15nm, such as K, Cs, Ba, Th, and U) tend to be more abundant near the surface. [Note the Th and U - does this mean we would only expect to see these radioactive elements now near the surface?]

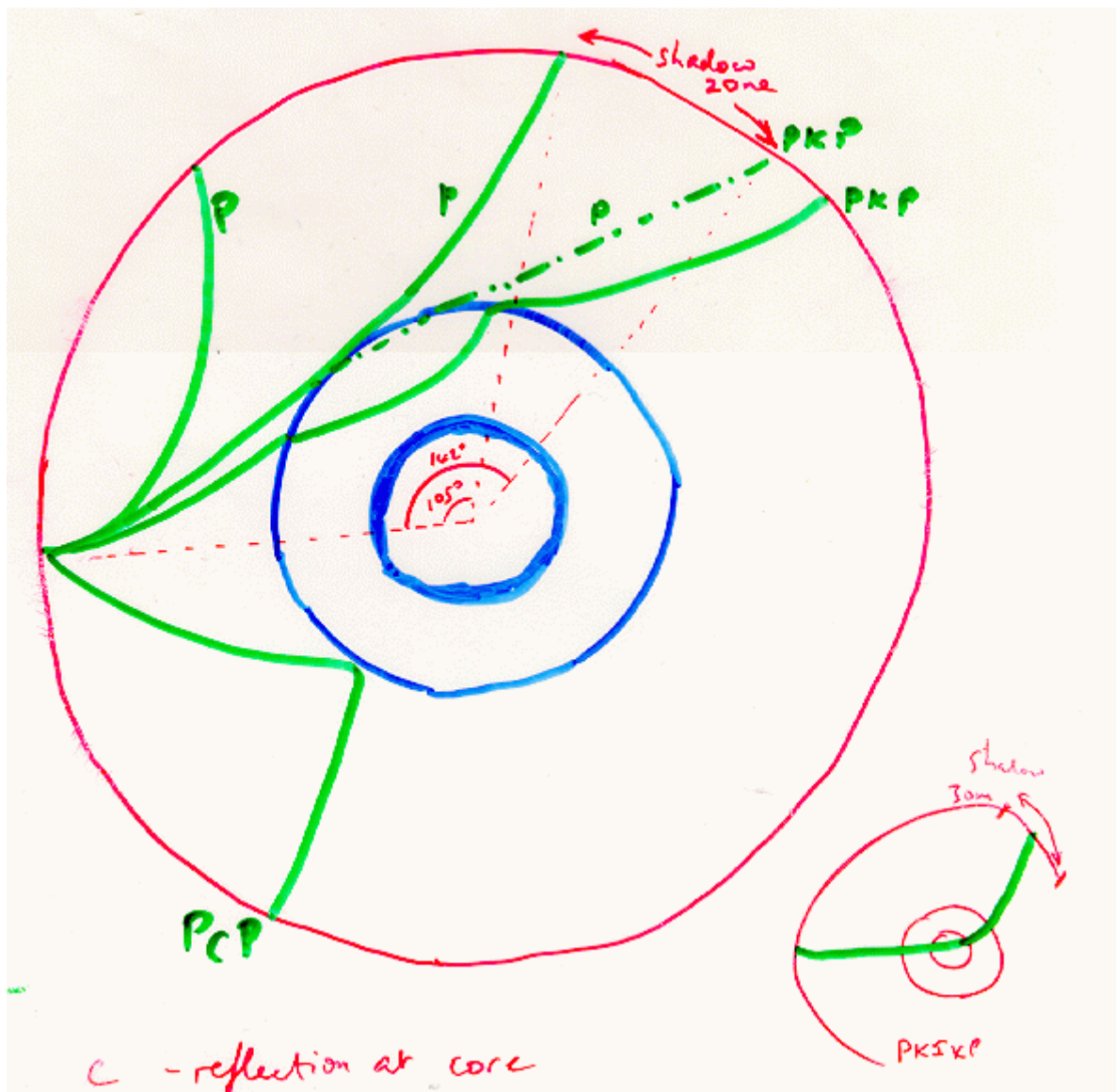
We can use seismic waves to probe deep into the earth. These are controlled by wave equations so they are of more use than potential functions like gravity for "inversion" to probe the earth's interior.

On a Seismograph we see the arrival of a wave triggered by an explosive event elsewhere. Or rather we see the arrival of a sequence of waves. Typically the **surface waves** arrive last. These have an amplitude inversely proportional to the distance from the source and attenuated by a factor of e for every approximately 5000km. Before these we get P and S waves - compressional and shear (originally Primary and Secondary) typically in a ten minute spread and with periods < 7 seconds. The P and S waves have arrived after travelling through the body of the earth and though the body of the earth and curve because the speed of waves increase towards the centre:



Shear waves cannot propagate through a liquid. Fermat's Principle can be used to estimate the minimum time of passage for a given structure.

There is seen to be a "shadow zone", from 105 degrees to 142 degrees from the source (epicentre), within which few waves arrive, and beyond which one sees no S waves, indicating there is an inner core which is liquid:



The nomenclature used is for P and S to be the compressional and shear waves, c to represent reflection at the core, K propagation in the outer core, and I propagation in the inner core. It is when there is propagation through the inner core that we sometimes get a signal returned in the shadow zone (PKIKP above). There is some evidence that the central core can support shear waves and so may be solid.

We have expressions for v_p and v_s in terms of the Bulk Modulus and Rigidity Modulus:

Bulk Modulus K and rigidity modulus μ .

$$v_p^2 = \frac{(K + \frac{4}{3}\mu)}{\rho} \quad v_s^2 = \frac{\mu}{\rho} \quad \therefore v_p^2 = \frac{4}{3} v_s^2 = \frac{K}{\rho}$$

ρ = density

$$K = \frac{\text{volume stress (pressure)}}{\text{volume strain}} = \frac{dp}{-\frac{dv}{v}} = \frac{dp}{d\rho/\rho} = \rho \frac{dp}{d\rho}$$

$$\therefore \frac{K}{\rho} = \frac{dp}{d\rho} \quad \frac{K}{\rho} \propto \rho$$

Poisson's ratio

The time of arrival of t_s

The time of arrival of t_s and t_p is of the order of Poisson's ratio - $t_s = \text{approx } 1.7 t_p$ on average.

$$\frac{dp}{d\rho} = \phi$$

(a known function of depth)

$$\frac{dp}{dz} = \rho g$$

(note positive, z is depth - $z+r = R_0$)

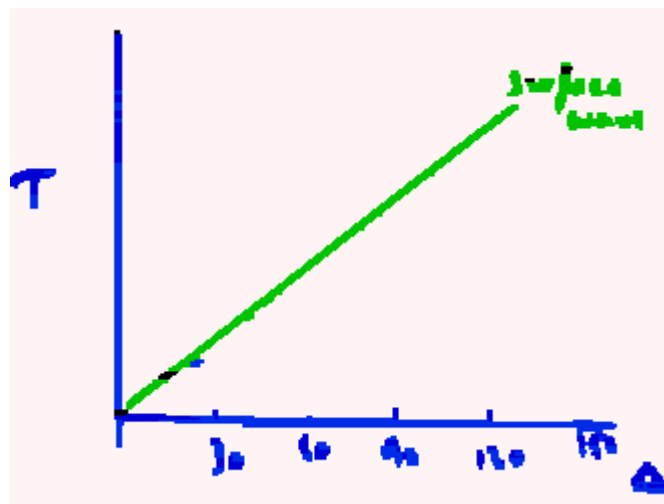
$g(r) = G m(r)/r^2$ if spherically symmetric

$$\therefore \frac{d\rho}{dz} = \frac{\rho z}{\phi} = \frac{\zeta_0(r)\rho}{\phi r^2}$$

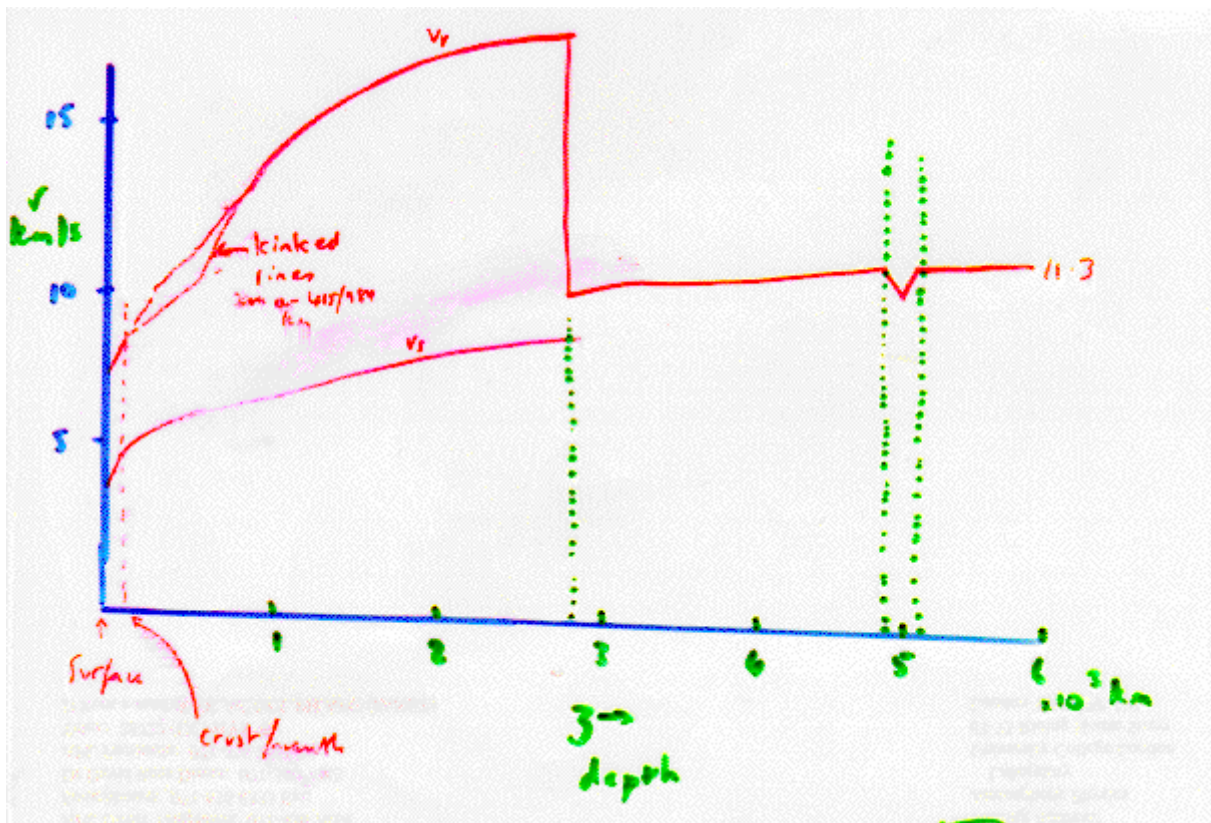
(The 1923 Adams-Williamson equation)

$$dm = 4\pi r^2 \rho dr \quad \frac{dm}{dz} = -4\pi \rho (R_0 - z)^2$$

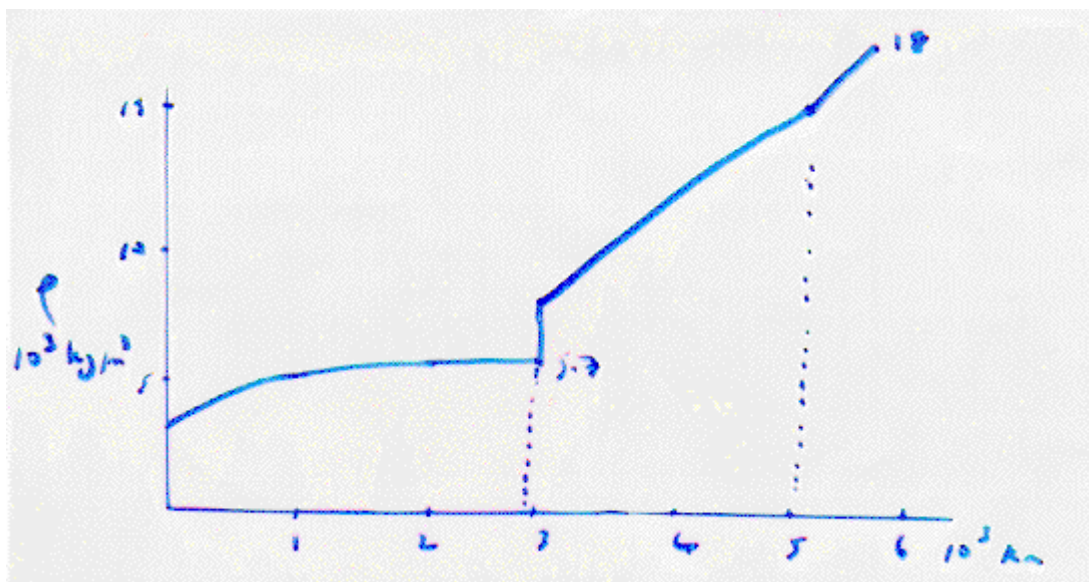
We can plot the time of arrival of the different types of wave on a distance v time graph:



where the lower axis is the angular distance from the source. Note the kink or discontinuity at 20 degrees and the cut-off at 105 degrees. These observations give v_p and v_s with depth:



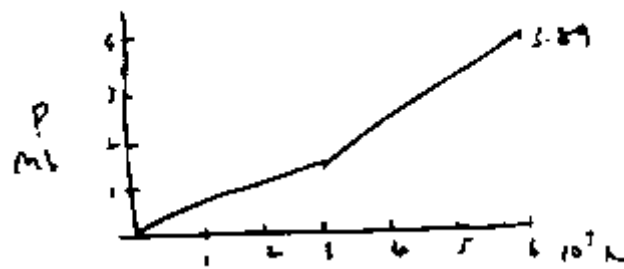
which in turn, from the equations above, enable us to derive the density distribution with depth:



Reasons for density increase with depth

The increase in density is partly due to the increase in pressure with depth.

i) Pressure increase with Depth



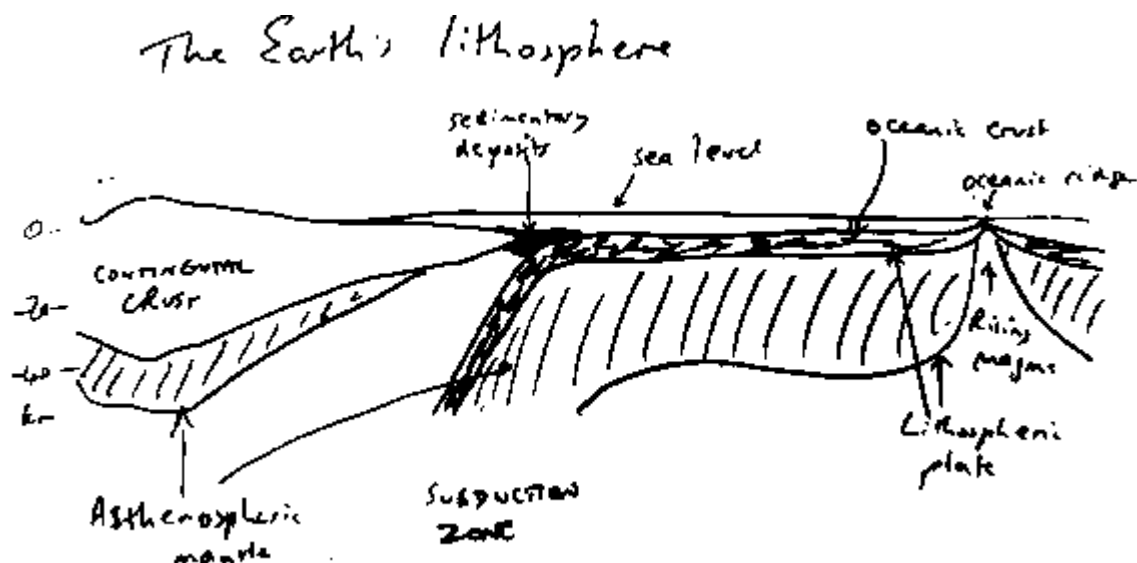
ii) Then there are phase transitions which give discontinuities or phase transitions which give discontinuities often at particular boundaries - thus at 300-1000km olivine changes to spinel. We have a sequence with the olivine Forsterite (Mg_2SiO_4) which at 110kbar turns to spinel with 11% greater density: this in turn will change to Stishovite at 321kbar (essentially $2\text{MgO} + \text{SiO}_2$). Forsterite can be turned directly into Stishovite at 199 kbar.

Similarly Pyroxene (2MgSiO_3) will go to Mg_2SiO_4 (spinel) + SiO_2 (stishovite) at 133 kbar, or at 182 kbar it will go to $\text{MgO} + \text{SiO}_2$ (stishovite)

Note that at the highest pressures we often end up with the production of oxides. This will be very important at great depths. At 1000-2700km (near mantle-core boundary) there will be mainly oxides of Mg, Si and Fe (same composition as olivine). At 2700-2900 km there will be diffusion of Si and Ni from the core leading to precipitation of Fe, though there are regional discrepancies (seen inside the shadow zone).

There seems little doubt the outer core is liquid (Ni, Fe and Si). The inner core could easily be solid - there is certainly a jump in the velocity of the p waves there. Either the bulk modulus goes up sharply or it goes solid and the rigidity modulus changes sharply. It is thought that the bulk modulus cannot change anything but steadily where the pressure is over 1Mbar. There is only sparse evidence of shear waves in the shear waves in the central core, though.

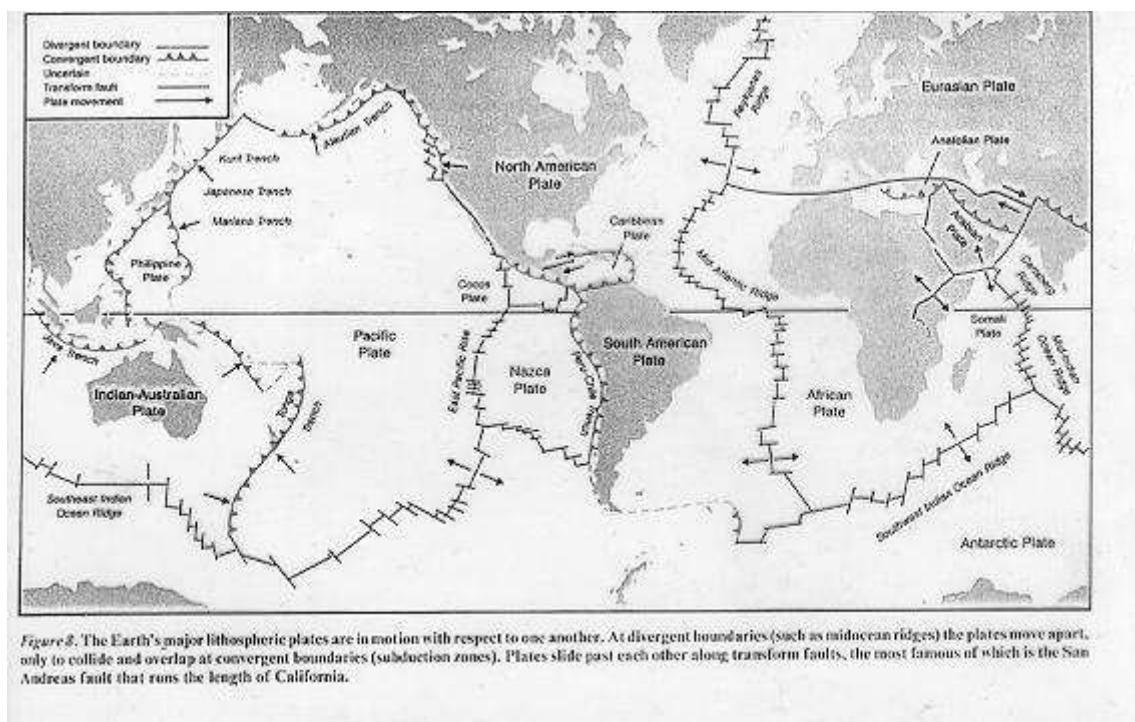
The Earth's lithosphere



Note the crust is thicker under the land than under the oceans. The crust-mantle division is a compositional change, the crust-asthenosphere division is a change in plasticity. The crust-mantle interface is called the Moho (after Mohorovicic), a transition from lower density silicates to higher density silicates 0.1 to 0.5 km thick, 5-10 km below the ocean floor or 20-90 km beneath the continents.

The crust sits in isostatic equilibrium on the magma. We know that the crust is divided into "plates" which move with respect to each other, gradually changing the alignment of the continents. Oceanic ridges generate new crust at the rate of 17 km^3 per year, whereas at the continental boundaries a similar amount of material is subducted and recycled.

Plate tectonics



This continual movement of the continents can be traced back in time: see for example Beatty and Chaikin or a Geology/Palaeontology text book for the development of the continents from Gondwanaland and Pangea 100s of millions of years ago.

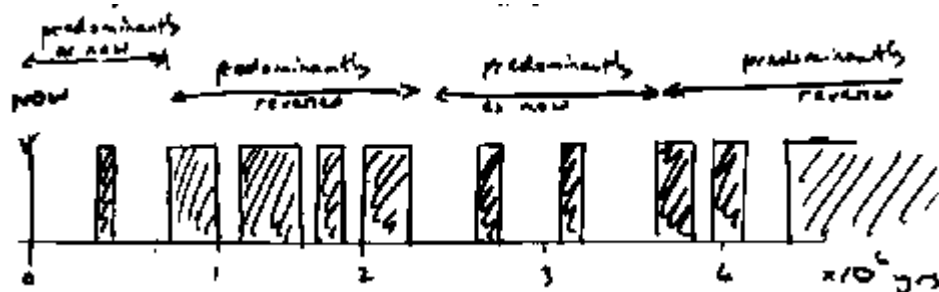
Paleomagnetism

materials show one of three types of magnetism:

- diamagnetism (quartz, salt, feldspar) - low susceptibilities
- paramagnetism (fayalite Fe_2SiO_4 , ferrosilite FeSiO_3) susceptibilities 100-200 times diamagnetic
- ferromagnetism (magnetite Fe_3O_4 , Hematite Fe_2O_3) This is **remanent** unless heated to the Curie temperature.

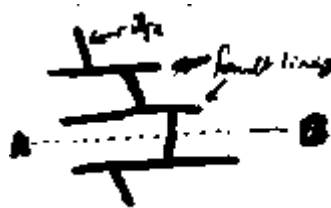
At constant T if an applied field is removed the magnetisation which is left is referred to as the Isothermal Remanent Magnetisation (IRM). If the material is cooled below the Curie temperature in a field, the field it retains is known as the Thermodynamic Remanent Magnetisation (TRM).

One can study of the Earth's magnetic field at the times rocks formed and calculate the direction of the field at the time of the cooling through the Curie Point, assuming a dipole field. You can then calculate where the dipole axis was at the time. We find the field has reversed several times:

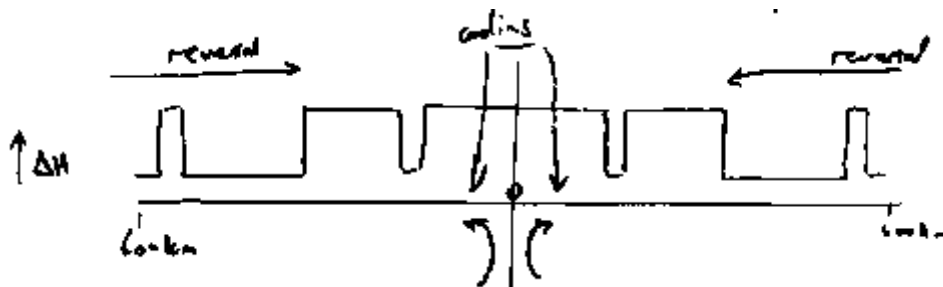


The reversals take about 5×10^3 years, with the main reversals about every 10^6 years.

One of the main uses of this has been to look at sea-floor spreading. If we look at the morphology of, say, the mid-Atlantic ridge, we see a complex fracture pattern following the line of the ridge:



If we cross the ridge from A to B and measure the magnetic field directions we find a pattern mirrored about the ridge line: the pattern is the sort of reversal pattern we saw above:



The time-scale is known from the previous diagram, which tells us the rate of sea-floor spreading. This has been crucial in developing the theory of continental drift. Typical rates of sea-floor spreading are: near Vancouver c. 2.9 metres/yr; off Peru c. 4.5 metres/year; these are typical values.

