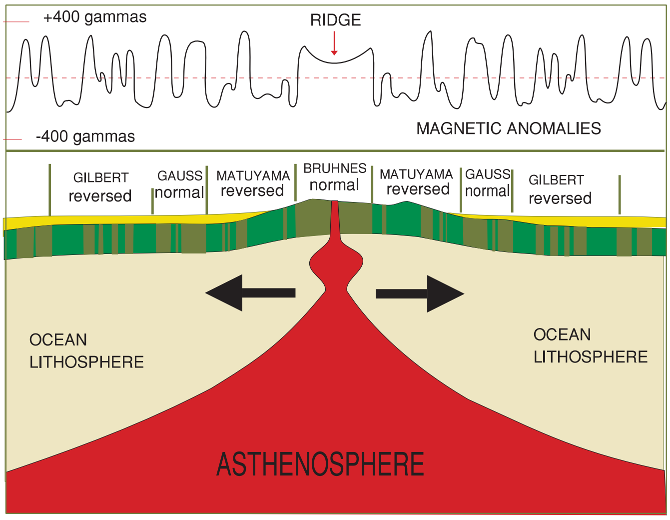
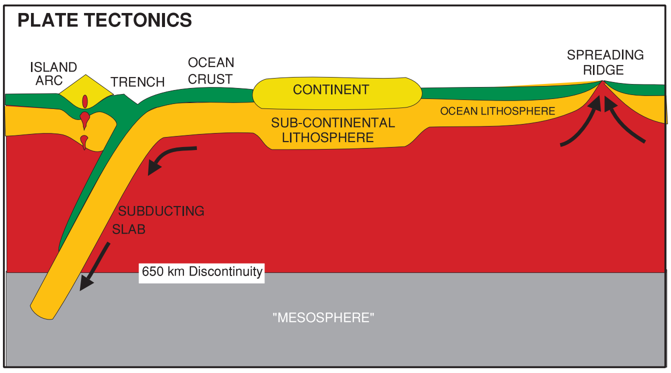
PLATE TECTONICS: Lecture 2

OCEAN RIDGE MAGMATISM

Magma production at the Earth's mid-ocean ridge system far exceeds that in any other tectonic environment, and this has been so since the early Precambrian. It is the dominant way in which internal heat is dissipated. The structure of a mid-ocean ridge is shown below:



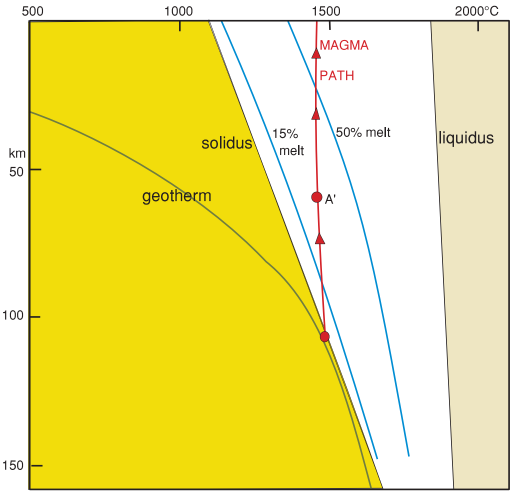
Note how the lithosphere thickens as it moves away from the ridge. Because the Earth's magnetic field oscillates between north and south at intervals of a few hundred thousand (or the odd million) years the basalts erupted then take on the current magnetisation, and so give rise to the seafloor magnetic lineations (patterns shown above) that can be used to date the ocean floor. Melting of pyrolite mantle extracts basaltic liquids to form the ocean crust, leaving a residue of harzburgite (ol+opyx) forming the underlying lithosphere. The ocean lithosphere suffers extensive hydrothermal alteration at the ridge (see below), but the rocks eventually finish up subducting back into the mantle:



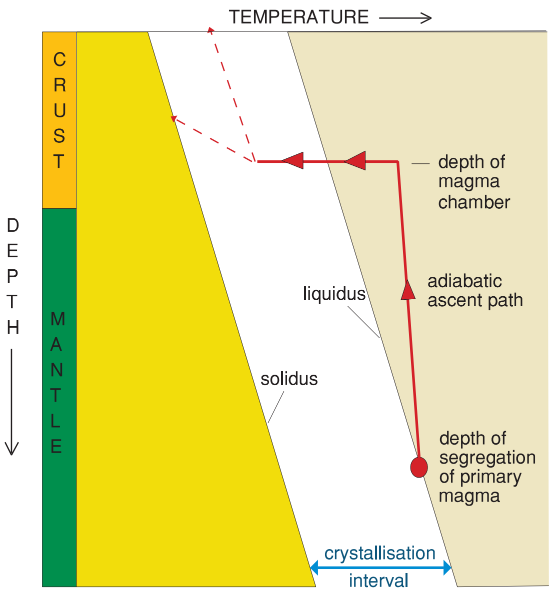
It is because these fluids are released in the Benioff Zone as the slab is subducted that magmas are able to be generated in the mantle wedge above the subduction zone. It is fluid, not friction, which is responsible for active margin magmatism. But it is ridge processes which make it all possible. So we need to look at these.

Why does melting occur? Melting temperatures of most silicate minerals increase with increasing pressures. So temperatures of solid mantle material at depth may be higher than the melting point of mantle near the earth's surface. As hot deep mantle rises beneath spreading ridges it will, as pressure falls, rise above its solidus, and begin melting.

The simplified situation is as follows:



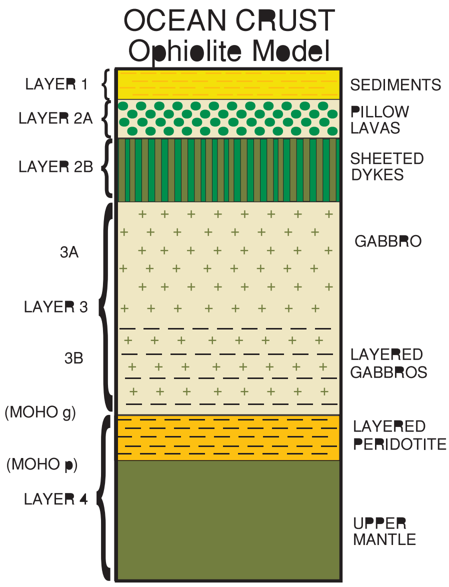
As the uprising mantle crosses the geotherm it begins to melt, and as the solidus temperature of mantle falls with decreasing pressure, the temperature of the melt increases relative to this solidus, thus effectively giving higher degress of melting with decompression, as shown. The amount of melt generated will be limited by the latent heat of fusion (which is high for silicates), and as the melting range of mantle peridotite lies between ca. 1100°C and ca. 1700°C, it is likely that most ridge basalts are partial (rather than complete) melts of mantle. The magma may enter a chamber in the ocean crust and begin crystallising, giving the following P-T path:



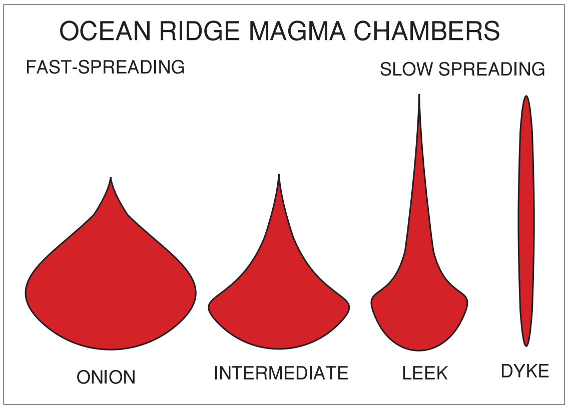
There is the possibility of superheat (i.e. temperature above the liquidus) if the magma can rise quickly, but it is apparent that most magmas are erupted or emplaced without superheat (a possible exception are ultramafic lavas called komatiites).

Because we haven't yet been able to drill very far down into oceanic crust, the only way we can begin to understand what happens to the

basaltic magma as it rises up at the ridge is to look at ophiolite complexes. There are many of these in the Alpine belt, although we are not always sure that these mafic slivers represent true ocean basin crust or whether some (or all) may represent marginal basin crust or the roots of island arcs.
Nonetheless, by putting together information from a number of ophiolite complexes, particularly Troodos on Cyprus, we come up with the following idealised section:

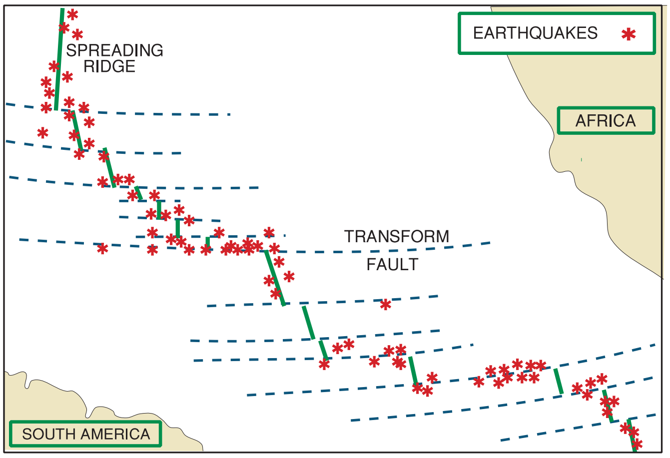


Not every ophiolite has all these components complete, and it is not always for tectonic reasons. Often the gabbro is missing, or the sheeted dykes, and in some cases the dykes may intrude the harzburgite. Of course sheeted dykes can only be formed if there is a continuously extending magma chamber (try doing it without!). So if sheeted dykes are missing it may mean that there has not been such a magma chamber. In fact there is a lot of debate on this issue. Some geophysical studies indicate a possible continuous magma chamber beneath the East Pacific Rise. However, the EPR is a smooth fast-spreading ridge, and maybe there is enough thermal input to keep a continuous magma chamber going. On the other hand in the slow-spreading Atlantic with its central rift valley and irregular topography, there is no direct evidence for a continuous magma chamber. Some workers, including those at Leicester, suggest that with slow-spreading ridges, each eruption may be a distinct event, and that any magma chamber is only short-lived. Some sections of the Atlantic ridge, like the FAMOUS area (south of the Azores) have numerous small volcanic cones, and this is now being recognised all over the Atlantic.
A consequence is that that there may be a variety of magma chamber profiles, with those from fast-spreading ridges having fat "onion" shapes, those from rather slower-spreading ridges having "leek" shapes. Very slow spreading ridges (e.g. SW Indian Ridge) may just have dykes feeding lavas which directly overly peridotite. There are ophiolites with this profile, where the dykes cut harzburgite tectonite and gabbro is only locally developed. Even with the type Troodos ophiolite, which has a moderatley thick gabbro section, geochemical studies have shown that the gabbros are in fact a compound of a number of small bodies.



Transform Fault Effects

It has long been known that the ocean crust is much thinner in the vicinity of oceanic transform faults. Also that a greater variety of rock types can be drilled or dredged in the vicinity of transforms, and that there is usually a significant topographic difference between the two sides of a transform fault (esp. the larger ones).



The latter effect arises because the ocean crust sinks as much as 3 km over the first 50 m.y. of its existence. So the greater the age difference of adjacent bits of ocean crust across a transform, then the greater the height of the transform wall. Obviously if the wall is 1 km high, then a large amount of rubble will fall down onto the lower plate, and deeper parts will become exposed. Moreover as the transform fault moves, the movement can deform the basalts into hornblende schists.
The thinner crust arises from the cold-wall effect, i.e. that the mantle rising up adjacent to the transform fault are actually in contact with older, and therefore cooler, oceanic crust on the other side. Cooler conditions give less melt and therefore thinner crust. Thinner crust also means there is more likelihood of mantle being exposed in the transform wall, again increasing the variety of rock types.

METAMORPHISM OF OCEANIC CRUST

There has been a very great deal of interest worldwide in the metamorphism and hydrothermal alteration of oceanic crust. After all there are few geological situations where you have a large red-hot magma chamber below and a 3 km column of ocean water on top trying to dowse it. There are a number of important questions that could be asked:

(a) how extensive is the metamorphism, and how far distant from the ridge do the metamorphic effects extend?
(b) if there is extensive hydrothermal activity, does this lead to equally extensive mineral deposits which could be mined?
(c) would the metamorphism affect the magnetic anomaly patterns that are so useful for dating ocean crust?
(d) is the ocean crust so hydrated that this represents an important source of fluids at subduction zones?
(e) does the hydrothermal interchange influence the chemical budget of the oceans?

There is no doubt that ophiolite complexes (obducted bits of ocean crust in mountain belts) are usually >90% altered, and there was debate during the 60's and 70's whether this was a result of metamorphism in the mountain belt during orogenesis, or resulted from ocean floor metamorphism. The latter is now the favoured explanation. Many of the samples of ocean crust recovered from the ocean floor by drillling or dredging are altered.

Metamorphism

Cann (1979) recognises 5 different mineral assemblage facies in oceanic basalts recovered by dredging, drilling etc. The rocks characteristically preserve igneous textures.

(1) Brownstone Facies

Low temperature ocean floor weathering or cool hydrothermal alteration. Products usually have yellowish brown tint due to oxidising conditions (bluish grey if reducing). Mineral assemblages not in equilibrium; just replace specific primary phases.
Olivine replaced by Celadonite (K-rich dioctahedral Fe-illite) under more extreme alteration. This fills vesicles and replaces glass. Under reducing conditions this is a Saponite (Mg-rich trioctahedral smectite). Pyrite common. Thus basalt has clay alteraton products. Plagioclase remains fresh, though under extreme alteration may be partly replaced by K-feldspar (Adularia).
Glass: Where basalt glass is common, Palagonite (orange coloured disordered illite) occurs, often associated with a low temperature zeolite (Phillipsite) and Calcite.

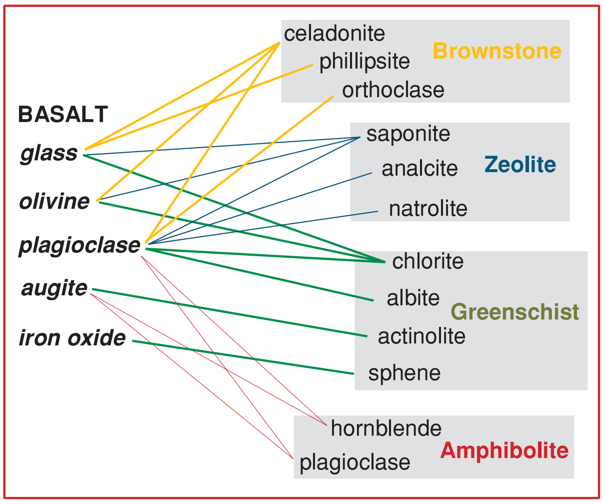


Fig. 8. Diagram from Cann (1979) tries to indicate how the minerals in a basalt affected by hydrothermal activity contribute to the secondary phases. At low temperatures it is mainly the basalt glass and olivine which are unstable and contribute to brownstone facies minerals, but plagioclase and then augite and iron oxide become progressively involved at higher grades until the whole rock recrystallises.

(2) Zeolite facies (Temperature above 50-100°C.)

Here Phillipsite is replaced by higher temperature zeolites - Analcite and Natrolite. Distinct zones of zeolites occur on Iceland.
Mafic minerals replaced by Saponite or Saponite-Chlorite mixed layer minerals, coarser grained than in Brownstone Facies.
Plagioclase may also be partly replaced by saponite, but augite stays fresh. Upper limit of facies (250-300°C) marked by disappearance of zeolites and saponite and incoming of albite and chlorite.

(3) Greenschist Facies

Albite ± chlorite ± actinolite ± epidote ± sphene. Degree of alteration variable, primary assemblages may be completely replaced. Augite is commonly a relic, veins are common, often quartz-bearing. Assemblages may or may not be equilibrium ones. Upper limit of facies marked by disappearance of albite, chlorite and actinolite and the appearance of green aluminous hornblende associated with more calcic plagioclase (An_{20-30}).

(4) Amphibolite Facies

Hornblende+Ca-plagioclase + titanomagnetite±epidote. This assemblage is most commonly developed in coarser grained rocks - dykes and gabbros - obviously of deeper origin. Degree of metamorphism variable. Some primary hornblende occurs in gabbros or diorites, but it is clear that amphibolite facies metamorphic assemblages are superimposed on this. So metamorphism closely follows magmatic activity.

The results can be summarised in the following table. Note that it is not just the mafic ocean crust (basalt or dolerite) that is altered. The mantle itself is often brought up along faults, transforms and fracture zones, and this is frequently altered to serpentine (ca. 13% water) at temperatures below 450°C. But other hydrous minerals such as talc, tremolite and chlorite are possible at higher temperatures. There are also 3 different varieties of serpentine: antigorite, chrysotile (the glossy variety) and lizardite.

Summary of Mineral Assemblages in Altered Crust

Facies	***PERIDOTITE***	***BASALT***
Brownstone	Celadonite +	Phillipsite + Palagonite + Saponite ?
Zeolite	Saponite + mixed layers + analcite + natrolite	?
Greenschist	Chlorite + Albite + Actinolite + Epidote + sphene	Lizardite Chrysotile Magnetite
Amphibolite	Hornblende + Plagioclase + Iron Oxide	Tremolite + Olivine + Enstatite
Gabbro	Augite + Plagioclase + Hypersthene + Iron Oxide	Olivine Enstatite + Diopside + Chromite

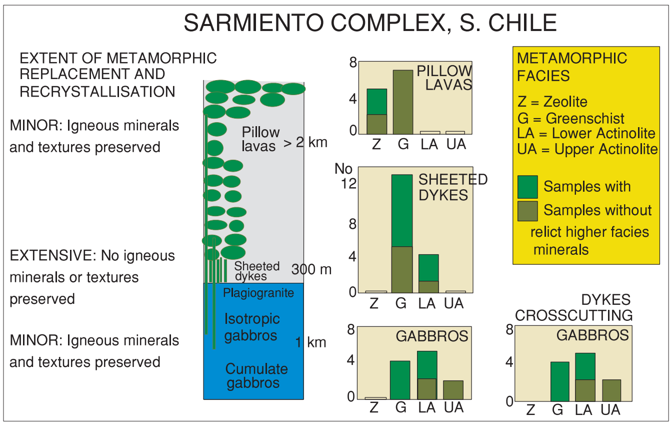
General Comments:
Greenschist and amphibolite facies metamorphism of the ocean floor differs from these facies in normal regional metamorphism in that:

(a) The thermal gradient is very high: can be several-100°C per km compared with 30 - 50°C/km in regional metamorphism.
(b) No garnet is developed in mafic rocks (pressures are not high enough).
(c) The rocks lack deformation textures (except in samples recovered from transform fracture zones)
(d) Very variable degree of recrystallization, because lower grade metamorphic assemblages are superimposed on earlier higher grade ones. This is because hydrothermal activity continues under cooler conditions as crust progressively moves away from ridge. (In regional metamorphism it is more common for the rocks to equilibrate at one set of P-T conditions)

Despite ca. 20 years of drilling, the deepest drill holes in the ocean floor (several hundred metres) have still only penetrated brownstone- and zeolite-facies rocks. No greenschist facies or amphibolite facies rocks penetrated. To see what happens deeper down, we really need to examine ophiolite complexes.

Ocean floor metamorphism - Sarmiento Ophiolite, Chile

The Sarmiento ophiolite (Saunders et al. 1979) is one of a series of discontinuous mafic lenses that represent the mafic floor of an extensional back-arc basin that was closed and uplifted in mid-Cretaceous times in the southern Patagonian Andes. Excellent vertical exposures have enabled the distribution of metamorphic zones resulting from the hydrothermal metamorphism to be established (Stern & Elthon, 1982?).



Lithological sequence at Sarmiento consists of:

2 km pillow lavas
300 m sheeted dykes
1 km gabbros with plagiogranite

Four main metamorphic equilibrium mineral facies can be recognised:

(1) Zeolite Facies. Zeolites, palagonitized glass ± smectites ± calcite ± quartz ± pyrite ± sphene ± albite.
(2) Greenschist facies. Chlorite, epidote, Na-plagioclase, sphene, ± quartz ± calcite ± biotite ± pyrite.
(3) Lower Actinolite Facies. Low-Al (2-5% Al_2O_3) fibrous green amphiboles, Ca-plagioclase, sphene ± biotite ± calcite.
(4) Upper Actinolite Facies. Higher-Al (5-8% Al_2O_3) brown-green amphiboles, Ca-plag. (>An50), titanomagnetite, ± ilmenite ± biotite.

These are arranged vertically in the complex, with metamorphic grade increasing downwards. However, intensity of metamorphism also varies and is at a maximum within the sheeted dyke unit. Moreover, the lower temperature facies may be superimposed on the higher grade ones. The histograms show the intensity of metamorphism in each of the components of the complex for each of the four metamorphic grades.

Significant Points:

(1) The intensity of metamorphism (recrystallisation) is greatest in the sheeted dykes. This is because the vertical dyke margins permit easy access of circulating fluids, coupled with the fact that higher temperatures speed reaction rates. The high water-rock ratios in the sheeted dyke zone mean that the rocks are strongly leached. Any chemical elements not required by the newly forming minerals (hornblendes and chlorite) are removed upwards by the fluids. This means elements like Rb, U, Th, K, some Sr, Ba and chalcophile elements like Zn, Cu and Pb are removed from the deep dyke rocks. Some of the former group will be re-absorbed by the zeolites and clays in the uppermost 'Brownstone' part of the section, others are dissolved in seawater, whereas the chalcophile elements form valuable mounds of sulphide on the seafloor (black smokers). Overall, the hydrothermal activity achieves a major amount of vertical chemical redistribution within the ocean crust. The mobile elements are moved to the top of the ocean crust. These are the same

elements that become "mobile" when the ocean crust is subducted and arc magmas are generated.
(2) As thermal gradients fall (i.e. crust moves away from ridge axis) circulating water permits lower grade assemblages to form - superimposed on high grade ones. But this secondary metamorphism is of lower intensity because circulation channels become blocked with the growth of secondary minerals. The sedimentary cap that progressively builds up on top of the basalts will eventually block circulation. There is a good example of this in the "Megaleg"

Chemical fluxes in oceanic crust - the 'MEGALEG'

The Deep Sea Drilling Project Legs 51-53 drilled two deep 200 m holes in Cretaceous (110 m.y.) oceanic crust in the western Atlantic near Bermuda. The holes were only 450m apart, but one hole, Hole 417A, drilled some of the most altered basalts found on the ocean floor, whereas those in Hole 417D were relatively fresh. All alteration was at 'Brownstone' facies. Basalts in both holes petrographically similar. The differences just reflect the relative access by circulating fluids.

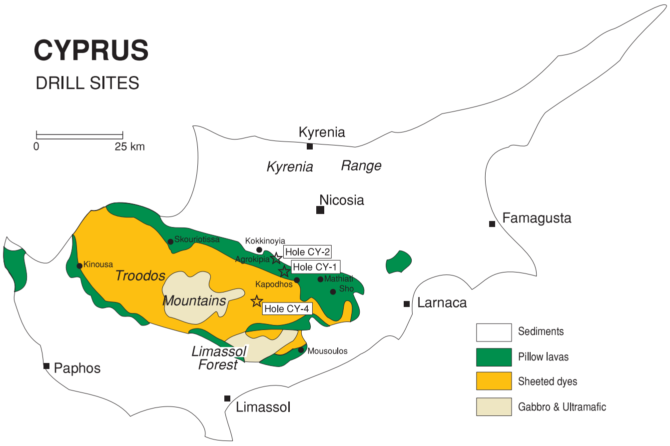
Compare compositions (water free):

	417A hyaloclastite	417A avge	417D avge
SiO_2	53.6	49.9	49.4
TiO_2	1.13	1.50	1.50
Al_2O_3	11.4	10.9	10.5
MgO	5.80	5.44	6.14
CaO	3.68	10.3	13.5
Na_2O	1.70	2.21	2.40
K_2O	4.36	1.79	0.12

Note significantly higher K_2O and lower CaO in altered rocks, particularly the most fragmentary hyaloclastite 417A, which is the most permeable of rock types.

Interpretation: Hole 417A is on basement 'high' which remained uncovered by sediment for a considerable time thus permitting long-term circulation of warm (30°C) water. Hole 417D was a located in a topographic depression which became quickly filled with sediment which blocked extensive water circulation.
The results were first interpreted to suggest that the ocean crust may be a tremendous 'sink' for K_2O and Rb transferred from sea water through hydrothermal circulation involving major volumes of seawater. However, how much of the Rb, K_2O, etc. is derived through leaching from more altered rocks (e.g. in sheeted dyke unit) at deeper levels?
To see what other chemical changes occur during alteration of oceanic crust it is best to look at the "type" ophiolite, the Troodos Complex on Cyprus. This has been intensively studied through field investigations, mining operations and by scientific drilling. Troodos formed at 91 Ma, but hydrothermal activity continued for a further 40 Ma after crust formation (Gallahan & Duncan 1994).

Chemical Changes in Oceanic Crust - Troodos Ophiolite



1. Strontium Isotopic Composition
Studies by Spooner et al. (1977) of zeolite- to amphibolite-facies altered basalts on Cyprus show that 87Sr/86Sr ratios are increased relative to fresh basalts and gabbros.

	$^{87}Sr/^{86}Sr$
Zeolite	0.80760 ± 3
Altered Basalt	0.7069
Fresh Gabbros	0.70338 ± 10
Cretaceous Seawater	0.7076

Similar results have been obtained on altered ocean basalts. High $^{87}Sr/^{86}Sr$ in altered mineral products can be leached away in dilute acid so that unaltered minerals yield the original magmatic Sr isotopic ratios.
Spooner suggested that interchange of seawater Sr with ocean crust Sr occurs during hydrothermal circulation and may buffer Sr isotope composition of seawater:

	$^{87}Sr/^{86}Sr$
Fresh Ocean Crust Av.	0.703
Seawater Average	0.709
Continental Rocks	> 0.712

There is a considerable variation in 87Sr/86Sr in seawater with time that can be linked to varying plate activity. This is discussed in more detail below.

2. Oxygen and Hydrogen Isotopic Compositions

Spooner et al. (1977) showed that oxygen isotope ratio values in Troodos and other Mediterranean ophiolites were higher (d18O = ca. 9) than in fresh 'mantle-derived' basalts (d18O = 6) and were consistent with alteration by seawater at high temperatures of ca. 350°C. The interesting point about this is that because oxygen is the most abundant element in any rock, it is necessary to exchange almost all the oxygen in the rock to significantly change the isotopic ratio. In other words, water-rock ratios are high, or large volumes of seawater interact with ocean crust at spreading centres. The implication is that if oxygen can be exchanged on this scale then many other elements can be changed too.
Heaton and Sheppard (1977) showed that the isotopic composition of hydrogen in water in equilibrium with chlorite and amphibole from

altered dykes from Cyprus was indistinguishable from that of seawater.

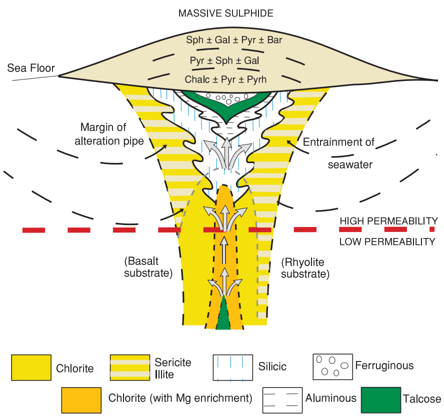
Comment: Altered oceanic crust (now with higher 87Sr, 18O and 2H contents) which is subducted at Benioff Zones may modify the isotopic composition of island arc magmas from "normal" mantle values. The hydrous fluids driven off as the subducting slab heats up as it goes down subduction zones will be enriched in the heavy isotope of these elements. So it is not surprising that island arc magmas differ in their isotopic ratios from other mantle-derived igneous rocks.

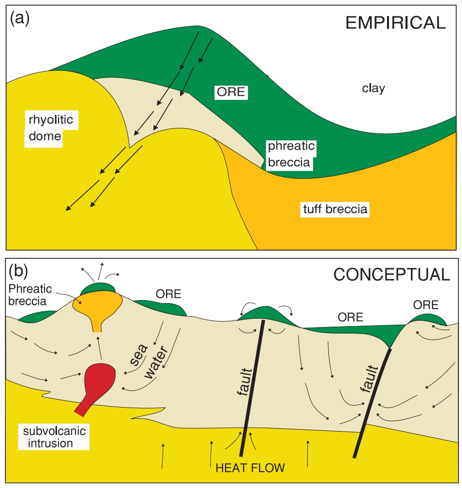
Sulphide Ore Deposits in Oceanic Crust

Sulphide deposits have been found on East Pacific Rise. They occur at positions of discharge of hydrothermal systems ("black smokers"). On Cyprus ore bodies are 500 m x 350 m x 50 m, and consist of pyrite and chalcopyrite with accessory marcasite, sphalerite and galena. Chemical and isotopic data suggest that the sulphide deposits mostly formed on seafloor:

(a) Fluid inclusions in ore material have composition of seawater.
(b) Ore material has $^{87}Sr/^{86}Sr = 0.7075$.
(c) Hydrogen isotope composition same as seawater.

Calculations suggest ore bodies may have formed in 100,000 yrs. Fluid inclusion studies suggest that the temperature of the plume of rising hydrothermal fluid was 300 - 350°C. Spooner suggests that contained sulphur has isotopic composition of seawater sulphate. So ocean crust ore sulphide may be largely of reduced seawater sulphate origin. The following diagrams illustrate some of the processes of convective seawater circulation and the respective mineral zones in the formation of hydrothermal mounds on the ocean floor (l.h. side) and other submarine envirnments (r.h. side of 1st picture):





These diagrams illustrate the importance of fault-control on the location of the discharge zones of hydrothermal activity, of permeability in allowing the hydrothermal solutions to circulate, of P-T-pH-Eh in controlling which minerals are stable and thus which elements are leached and which are deposited.

Comment: Large amounts of sulphide in adddition to chloride and hydroxyl are added to ocean crust as a result of hydrothermal activity. A lot of pre-concentration of potential ore metals already occurs in the ocean crust. So what happens when ocean crust goes down subduction zones?
At subduction zones chlorine- and sulphide-rich fluids are released during dehydration. Could this give us a possible explanation for porphyry copper deposits that occur commonly at continental margins like the Andes? Spooner has stressed that water is needed as transport medium, chloride for metal complexing and sulphur for fixing the metals as solid phases. All these are present in ocean crust as it is subducted.

Metal Deposits on Ocean Floor

Sulphide ores common in ophiolites. Could we locate them on ocean floor? Many hydrothermal discharge zones have now been found at ocean ridges (mostly in the East Pacific, but also now in the Atlantic) by submersible investigations. These are potentially important ore reserves in terms of total volume, but individual deposits are too small to mine economically, even by remote techniques. Discharge areas may be located by trace element profiles in seawater near the ocean bottom. So we can find them, but to exploit them is another matter.

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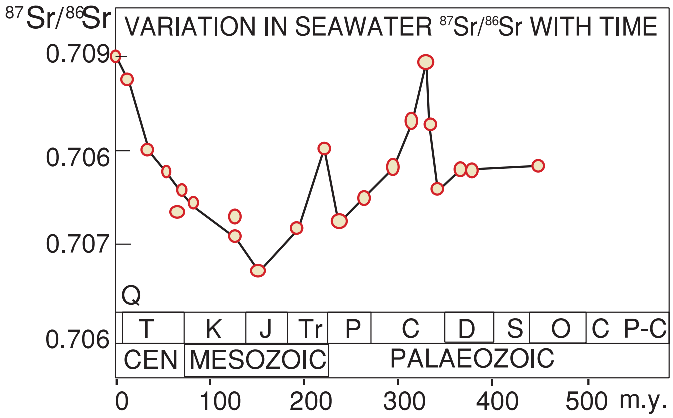
VARIATION IN Sr ISOTOPIC COMPOSITION of SEAWATER WITH TIME: the plate tectonics connexion

You may wonder what the strontium isotopic composition of seawater has to do with plate tectonics? Surprisingly the variation in the Sr isotopic composition of ocean seawater with time it is turning out to be an excellent monitor of past plate tectonic activity. But we are only just beginning to understand why. For instance the present day seawater Sr isotopic composition (expressed as $^{87}Sr/^{86}Sr$) is 0.709, and because ocean water is globally well mixed, all modern shells and limestones that incorporate seawater Sr have this ratio. By measuring the Sr isotopic ratios in limestones or shells from older geological formations it is possible to log the Sr isotopic variations in seawater with time. The early work by Peterman et al. (1970) and others produced the following curve:

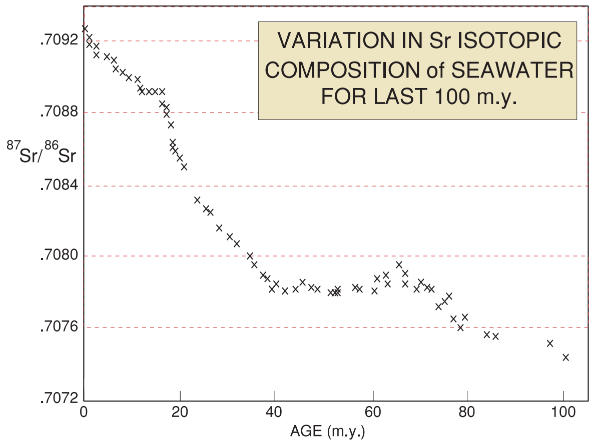


It can be seen that there is a progressive increase from the late Jurassic to the present day. However there is also a decrease from the early Carboniferous to the late Jurassic. So there must be some geological process or processes which produce a decrease in $^{87}Sr/^{86}Sr$ as well as those which produce an increase. So what are these processes?

Rubidium-87 (^{87}Rb) is radioactive and decays to ^{87}Sr, so that the ratio $^{87}Sr/^{86}Sr$ must increase in the Earth with time (^{86}Sr is unradiogenic so stays constant). However the rate at which the $^{87}Sr/^{86}Sr$ ratio will increase depends on the elemental Rb/Sr ratio of the rock. The Earth's mantle is low in Rb relative to Sr, so mantle derived rocks, which have a very low Rb/Sr ratio, tend to have low $^{87}Sr/^{86}Sr$ (the ratio has increased from only 0.699 to 0.703 over the last 4500 m.y.!). However, crustal rocks such as granites and shales are rich in Rb and have a high Rb/Sr ratio so, given time, become enriched in ^{87}Sr and have a high $^{87}Sr/^{86}Sr$. So if seawater interchanges chemically with crustal rocks its ratio of $^{87}Sr/^{86}Sr$ will increase, wheareas if it interchanges with mantle rocks this ratio will be pulled down again. Clearly to change the ratios in a reservoir the size of an ocean must indicate (at least) two very big process, one pushing the ratio up and the other dragging it down. What are they?

Significance of Increasing Sr Isotope Ratios

Clearly, enhanced erosion of high $^{87}Sr/^{86}Sr$ continental material will drive the ratio up. As feldspars and micas weather and breakdown to clays their radiogenic Sr is released. This is very soluble in river water and finishes up in the sea. But enhancement can only come if the proportion of mountain belts is increased. And this will be brought about by continental collision at the end of the Wilson Cycle. So does the sharply increasing curve since the late Jurassic then monitor the development of our recent mountain belts such as the Alps, Himalayas and Andes? There have been a number of attempts to model this most recent 100 m.y. growth, since the seawater variation curve for this period is very well known (after Richter *et al*. 1992):



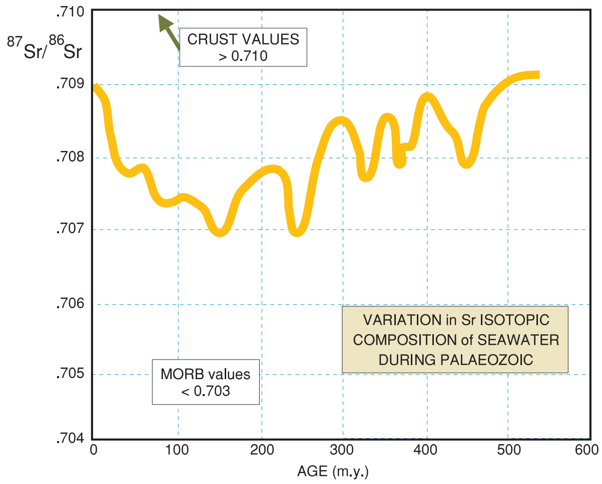
The Sr flux into the oceans can be estimated from the large rivers, e.g.,

River	Sr Flux (mol/yr)	$^{87}Sr/^{86}Sr$
Amazon	2.2 × 109	0.7109
Orinoco	0.2 × 109	0.7183
Himalayan-Rivers total	7.7 × 109	0.7127
Global Total	3.3 × 1010	0.711

The growth is most rapid between 40 m.y. to the present, and particularly in the period 20 - 15 m.y. Richter et al. (1992) find that they can model this rapid growth mainly as a consequence of uplift of the Himalayan-Tibetan plateau, following the collision of India with Asia. See the high values for the Sr flux in the Himalayan Rivers (Indus, Ganges, Irrawady, Yangtze, Mekong, etc.). So can we detect mountain belt formation in the geological record by rapid increases in the Sr isotopic ratio of seawater?

This poses an interesting problem as to why, if mountain belts are repeatedly formed throughout history, the Sr ratio does not just go on up and up. After all, if the seawater ratio can rise by 0.002 in the last 40 m.y., why has it risen by only 0.010 in the last 4500 m.y. Some equally powerful process must be bringing it down again.

The curve for the whole of the Phanerozoic (Cambrian to present) was established by Burke et al. (1982) by very careful work on limestones of various ages, taking care to avoid rocks that experienced later diagenetic effects:



It can be seen that we actually have to go back to the Cambrian before the seawater Sr ratio was as high as it is at the present day. In fact there has been a general decline for a period of almost 400 m.y. between the Cambrian and the late Jurassic, and superimposed on this is at least five sharp falls.

Significance of Decreasing Sr Isotope Ratios

The main reason for the decrease must be hydrothermal exchange of seawater with hot basalt at mid-ocean spreading centres. As we have seen from ophiolites, the breakdown of feldspars with low "mantle" Sr ratios releases Sr into the seawater. At the same time, zeolites and clays growing in the low-temperature altered basalts take Sr out of seawater with a high "continental" component. Because ocean crust is eventually subducted at trenches, the net effect is to remove some of this "continental" Sr component located in the hyrothermally-altered ocean crust deep into the Earth's mantle.

So can we correlate these periods of rapidly falling Sr isotope composition with periods of faster spreading? Or with the breakup of major continents like Gondwanaland? Note that if there is enhanced ridge activity, this probably means that the uprising mantle is hotter and less dense, and that most likely it will displace ocean water volume, like Iceland, and flood the continental shelves. Of course the more the continents are flooded, the less active will be the rivers, and the lower their content of "continental" Sr. But, in contrast, once there is a major continental collision, like that following the closure of Tethys and the formation of the Alps and the Himalayas, then ridge activity must stop, or slow, for a while until new plate configurations are established. So is this why the curve oscillates around. It is advisable to try to model it. This is what Richter et al. (1992) have tried to do.

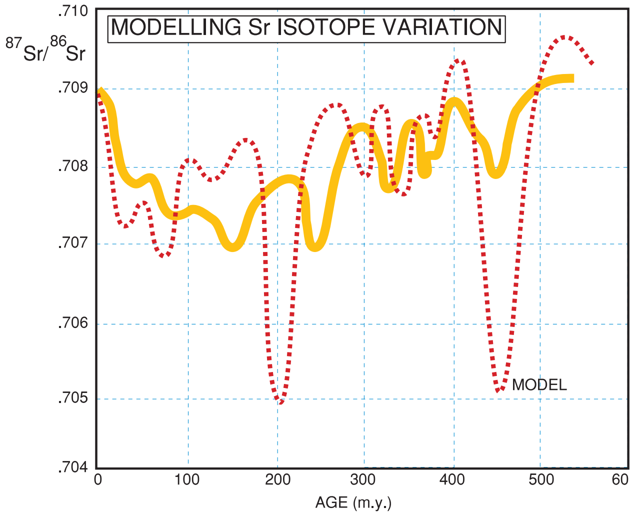
Modelling the Phanerozoic Curve

Richter et al. (1992) used the following numbers:

Total Sr in Oceans:	1.25 × 1017 (mol)	0.7092
River flux:	3.3 × 1010 (mol/yr)	0.711
Hydrothermal flux:	0.82 × 1010(mol/yr)	0.7030
Diagenetic flux:	0.3 × 1010 (mol/yr)	0.7084

The latter flux is that due to carbonate diagenesis, but is not an important controlling factor.

They then tried to model the Sr isotopic variation throughout the Phanerozoic by making various assumptions about plate motions, collisions, spreading rates, etc. Obviously things get more uncertain the further back in time, and because there are parts of the Earth where the geology is not very well known. The result is shown below:



The model accounts reasonably well for the large-scale structure of the sewater Sr isotopic curve, but fails to reproduce several of the local maxima and minima, especially in the period 300 - 100 m.y. However the "highs" in the Cambrian, Devonian and present day do correlate with extensive mountain building.

The interesting point about such graphs is that they tell us that there may be something missing from our current plate tectonic models. For instance, Richter et al. assume that all mountain building is due to continent collision. But there was no major continent collision as such involved in the uprise of the Andes, which is also a major contributor to seawater Sr. The Andean uplift may have resulted in part from the docking and attempted subduction of major ocean plateaus. This would add a new dimension to the story because the formation of hot ocean plateaus would substantially enhance the ridge hydrothermal Sr flux, to be followed later by an enhanced "continental" flux as mountain belts were formed as these hot thick plateaus tried to subduct.
For further discussion of long-term changes in geological features, and possible implications, see Moores (1993).

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