From Wilson, M. 1989
Igneous Petrogenesis

CHAPTER FIVE

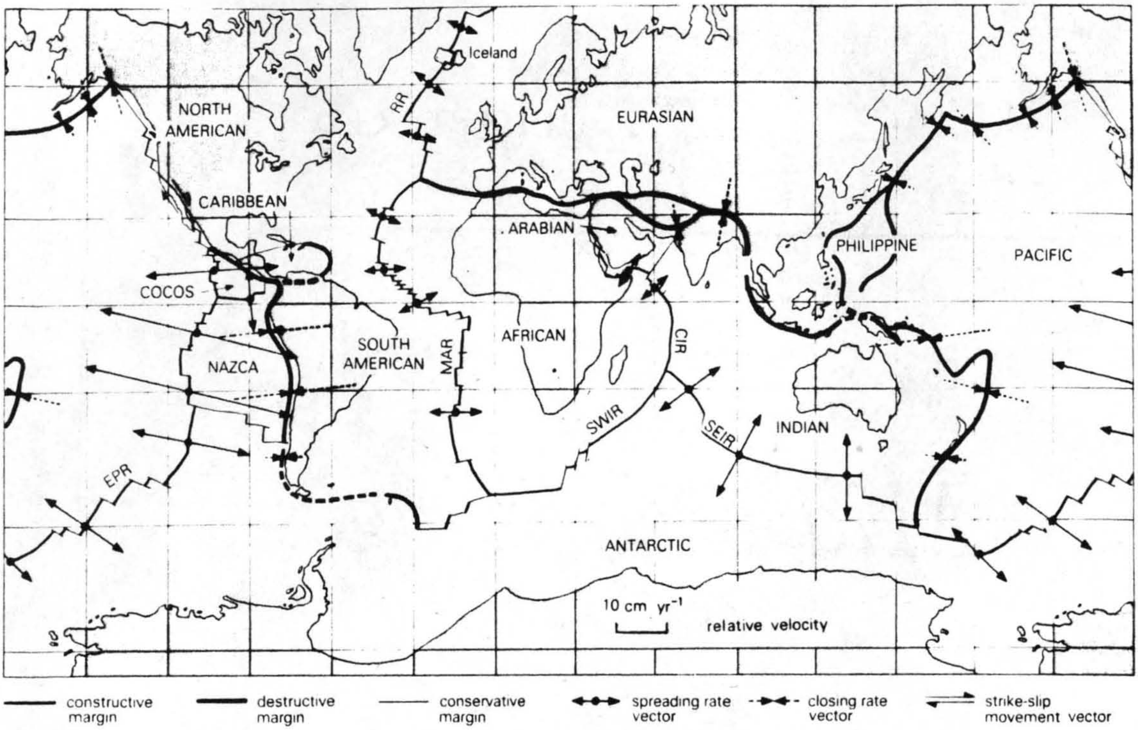
Mid-ocean ridges

5.1 Introduction

A map of the ocean basins (Fig. 5.1) shows that their most conspicuous topographic feature is the system of mid-oceanic ridges, the crests of which rise on average 1000–3000 m above the adjacent ocean floor. Such ridges extend through all the major ocean basins, with a total length in excess of 60 000 km. With the exception of the East Pacific Rise, they occur in the middle part of the oceans and essentially form a submarine mountain range, which rises to its highest elevation at the ridge crest and slopes away symmetrically on either flank. Topographically, they vary throughout their length, the East Pacific Rise being much broader and less rugged than the other ridges (Section 5.4). A rare terrestrial expression of the mid-oceanic ridge system occurs in Iceland, whose central graben is the extension of the Mid-Atlantic Ridge.

The ocean floor is cut by hundreds of fracture zones (Section 5.4.3) that, on an ocean-floor map, form a pattern of semi-parallel stripes cutting across and frequently offsetting the ridge axes. Such fracture zones are remarkably continuous features, normally extending for large distances across the flanks of the ridge, in some cases extending across the entire ocean floor to the continental margin.

Until the 1960s most geologists believed in the permanence of the continents and ocean basins. However, in 1962 H. H. Hess, in a classic paper, revolutionized thinking about the nature and origin of the oceanic crust and established the basic concepts of seafloor spreading, or plate tectonics, as we now know it. According to plate-tectonic theory, a mid-ocean ridge (or constructive plate margin) is a boundary between plates at which new



EPR	East Pacific Rise	SEIR	South East Indian Ocean Ridge
MAR	Mid-Atlantic Ridge	CIR	Central Indian Ocean Ridge
SWIR	South West Indian Ocean Ridge	RR	Reykjanes Ridge

Figure 5.1 Tectonic map of the ocean basins showing the system of mid-oceanic ridges. The length of the spreading rate vector arrows is proportional to the spreading rate (after Brown & Mussett 1981, Fig. 8.16, p. 153).

oceanic lithosphere (crust + mantle) is generated, in response to partial melting of mantle lherzolite undergoing adiabatic decompression in a narrow zone of upwelling (Fig. 5.2). Partial melting results in the formation of basaltic magma, which is injected through tensional fissures into a narrow zone only a few kilometres wide at the ridge axis. Surface volcanism, sometimes in the form of pillow lava, occurs but most of the magma solidifies within dykes and layered intrusives at greater depths. The new rocks thus generated are then transported away from the ridge axis by the continuous process of seafloor spreading at half rates of 1–10 cm yr.$^{-1}$

Since the size of the Earth is essentially constant, new lithosphere can only be created at mid-oceanic ridges if an equivalent amount of material is consumed elsewhere, at subduction zones (Part 3). Throughout geological time, a succession of ocean basins have been born and have grown, diminished and closed again. The present episode of continental drift and seafloor spreading began about 200 Ma ago with the opening of the Atlantic and Indian oceans, which are still growing in size with respect to the Pacific, which is decreasing.

Hess' theory of seafloor spreading was confirmed by the discovery that periodic reversals of the Earth's magnetic field are recorded in the oceanic crust symmetrically about the ridge axis (Section 5.3.1), and the lateral migration of new crust away from mid-oceanic ridges is now well documented. Nevertheless, our understanding of the actual processes involved in the generation of new oceanic crust is still incomplete.

Essentially, the oceanic crust can be divided into two major domains:

(1) the accreting plate boundary zone (mid-ocean ridge) at which new oceanic crust is created;

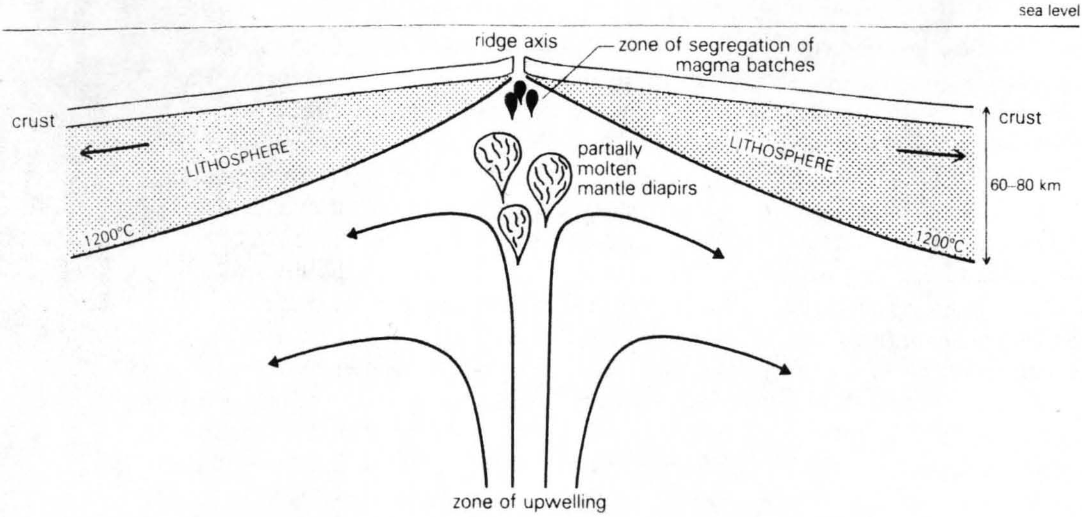


Figure 5.2 Schematic cross section of a constructive plate margin and its relation to the zone of upwelling of deep-mantle material. The oceanic lithosphere (crust + mantle) is generated at the ridge axis and increases progressively in thickness away from the ridge. The base of the lithosphere is represented as a thermal boundary layer (1200°C isotherm).

(2) the passive crust, which after creation at the ridge axis has moved away.

In this chapter attention will be focused on the processes operating at the accreting plate boundary zone.

Information about the character of the oceanic crust has been obtained from a number of sources: dredging; deep-sea drilling; marine geophysical studies (especially seismic surveys); direct observation of the ocean floor from submersibles; and studies of rock sequences on land thought to represent uplifted segments of oceanic crust (ophiolite complexes). Until comparatively recently, seismic refraction studies and dredging provided most of the data.

Basalt was first dredged from the ocean floor at the turn of the century and since then extensive sampling programmes have shown that basaltic lavas with distinctive chemical characteristics are the major component of the oceanic crust (Section 5.10.1). Such basalts have been variously termed submarine basalts, ocean-floor basalts (OFB), abyssal basalts and mid-ocean ridge basalts (MORB). Initially, based on rather limited sampling, it was considered that ocean-floor basalts had rather restricted chemical compositions, with tholeiitic characteristics, constant SiO_2, low K and low incompatible element contents. However, since the mid-1970s, detailed sampling programmes involving drilling, dredging and the use of submersibles have revealed significant chemical and petrological diversity, often within a single site. In particular, basalts erupted along topographically 'normal' segments of ridges have different isotopic and trace element characteristics from those erupted along topographic highs or platforms associated with islands astride the ridge axis (e.g. Iceland, Azores, Galapagos, Bouvet and Reunion). In this respect the latter have more affinities with oceanic-island basalts but nevertheless are indistinguishable from normal MORB in terms of petrography, mineralogy and major element chemistry (Section 5.10).

In general, MORB are olivine tholeiites with a narrow range of major element compositions, indicating a relative constancy of sources and processes operating along most spreading ridges. They are the most voluminous eruptive rocks on Earth, and their generation has been a significant process in the differentiation of the upper mantle throughout geological time. The trace element variability of MORB is mainly attributable to source heterogeneities and to shallow-level processes in open system steady-state magma chambers (Section 5.8). Notable exceptions to the general compositional uniformity include localized occurrences of Fe–Ti-rich basalt and rare silicic differentiates along the East Pacific Rise, Galapagos, Juan de Fuca, SW Indian and SE Indian oceanic ridges. In some instances these are associated with propagating rifts (Section 5.4.4), but this is not true for all ferrobasalt localities. Anomalous ridge segments such as Iceland, which have become emergent due to high magma production rates, exhibit exceptionally high volumes of silicic differentiates compared to normal ridge segments.

5.2 Simplified petrogenetic model

Basaltic magma generation at constructive plate margins should theoretically represent the simplest type of terrestrial magmatism. Nevertheless, as we shall see in the following sections, the petrogenesis of MORB is by no means simple. The apparent regularity of the oceanic crustal layer over hundreds of millions of square kilometres (Section 5.3) attests to the continuity of magmatic processes over a timespan of the order of 100 Ma (the average life of an ocean basin). In contrast, detailed geochemical studies of MORB (Section 5.10) reveal significant source heterogeneities and a diversity of petrogenetic processes.

Figure 5.2 presents a summary of the processes responsible for the generation of magma at mid-oceanic ridges. They are the site of localized upwellings of deep mantle material (Section 5.6) which undergoes adiabatic decompression and, in doing so, partially melts to produce basaltic magma. Initiation of such an upwelling seems a necessary precursor to the fragmentation of a continent (Ch. 10) and the subsequent generation of a new ocean basin by seafloor spreading.

The oceanic lithosphere is generated at the ridge and increases symmetrically in thickness away from the axis, due to progressive cooling, to a maximum of about 60–80 km. The base of the lithosphere, marked by the 1200°C isotherm in Figure 5.2, is a thermal boundary layer reflecting a marked change in mechanical, but not necessarily chemical, properties from the underlying asthenosphere. The lower parts of the lithosphere are probably composed of fairly fertile mantle lherzolite (Ch. 3), indistinguishable from that of the asthenosphere, whereas the upper parts must be variably depleted due to magma extraction at the ridge axis.

The oceanic crust, comprising the upper 8–10 km of the lithosphere, also has its origins at the ridge axis due to the extrusion and intrusion of basaltic magma. It has a well layered layer structure (deduced from seismic studies) comprising a surface layer of basalts underlain by a variety of intrusive rocks, including dykes, cumulate gabbros and ultramafics (Section 5.3). Many models have been proposed to explain this layered structure, mostly involving high-level magma chamber processes (Fig. 5.3; see also Section 5.8 and Cann 1970, 1974, Christensen & Salisbury 1975, Rosendahl *et al.* 1976, Rosendahl 1976, Nisbet & Fowler 1978).

The chemical composition of basalts generated at mid-oceanic ridges must depend upon a variety of factors, including the following:

(1) The composition and mineralogy of the source mantle.
(2) The degree of partial melting of the source and to a lesser extent the mechanism of partial melting (see Ch. 3).
(3) The depth of magma segregation.
(4) The extent of fractional crystallization and magma mixing processes during storage of magma in high-level sub-axial magma chambers.

With so many factors involved, the apparent worldwide compositional uniformity of MORB appears rather intriguing. Early workers used this to argue that MORB compositions are those of primary magmas unmodified by near-surface processes, but O'Hara (1968) showed fairly conclusively that most MORB are in fact highly fractionated. This will be considered further in Section 5.10. If MORB are considered in terms of their major element chemistry alone (Section 5.10.2), they do indeed represent an incredibly uniform magma

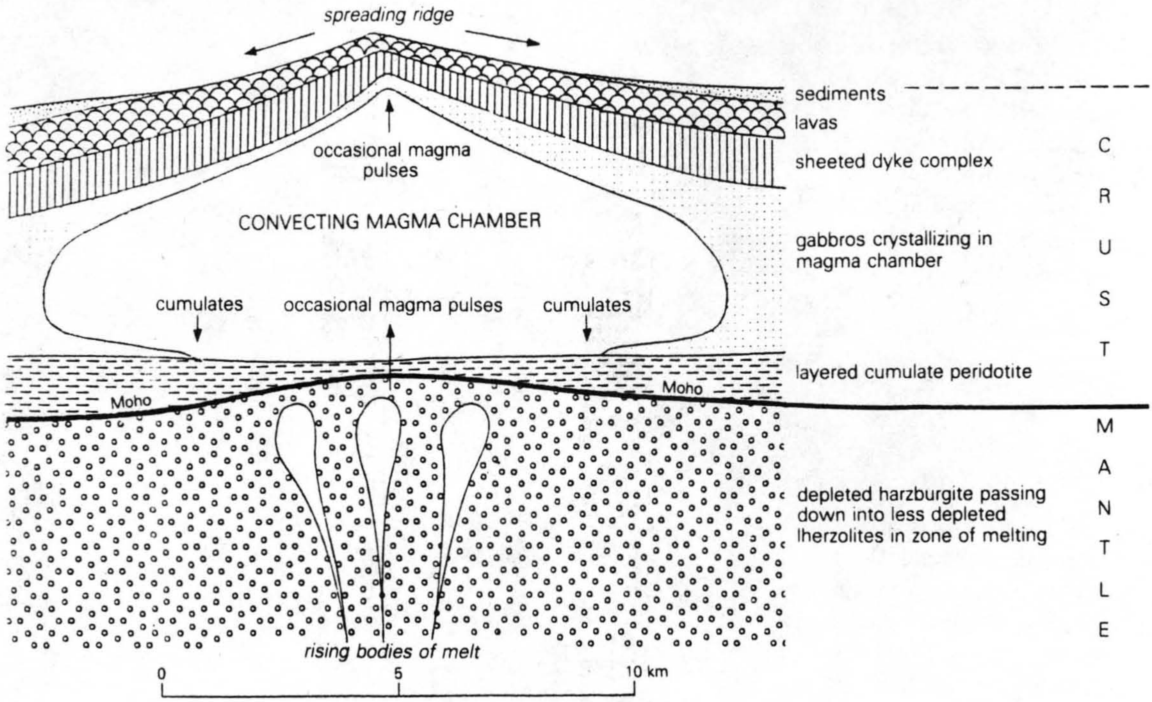


Figure 5.3 Hypothetical section through a mid-oceanic ridge, showing the development of the structure of the oceanic crust in response to magmatic processes at the ridge axis (after Brown & Mussett 1981, Fig. 7.8, p. 119).

type, the origins of which could be modelled in terms of rather simplistic processes. However, studies of their trace element and Sr, Nd and Pb isotopic composition (Sections 5.10.3 & 5) reveal the need for much more complex models.

Detailed discussion of the various parameters which control MORB chemistry is, of necessity, deferred until the appropriate sections, but the general conclusions to be drawn are presented here:

(a) *Source*. Generally depleted lherzolite in the spinel or possibly even plagioclase lherzolite facies. Some MORB associated with topographic highs along the ridge axis are apparently generated by partial melting of more enriched mantle sources (see Ch. 3 for explanation of the terms depleted and enriched). The trace element geochemistry of MORB shows no evidence for the existence of residual garnet in the source (Section 5.10.3).

(b) *Degree of partial melting*. As deduced from a variety of partial melting experiments on synthetic and natural lherzolites (Ch. 3 & Section 5.7) a minimum of about 20% partial melting appears to be required to generate the most primitive (MgO-rich) MORB compositions. If primary MORB are more picritic then the degree of melting would be somewhat greater (see Ch. 3).

(c) *Depth of partial melting and magma segregation*. Geophysical studies of the attenuation of P and S waves, although only available for a few ridge segments, suggest initiation of significant partial melting in rising mantle diapirs at depths of 60–80 km. Segregation of magma batches probably occurs at depths of about 20 km, feeding magma directly into high-level reservoirs (Hekinian 1982).

(d) *Fractional crystallization*. A major controversy in the study of MORB petrogenesis centres on the role of magma chamber processes in the control of their geochemistry and in the generation of the layered structure of the oceanic crust (Section 5.8). Large sub-axial magma chambers cannot persist on thermal grounds at slow-spreading (Atlantic-type) ridges and therefore high-level processes may be significantly different from those at fast-spreading (Pacific-type) ridges, where large magma chambers have been shown to exist.

5.3 Nature of the oceanic crust

5.3.1 Geophysical data

Palaeomagnetic studies in the early 1960s revealed the existence of magnetic stripes on the ocean floor and laid the foundation for modern theories of plate tectonics and seafloor spreading. Subsequently, detailed geophysical investigations (seismic, gravity and magnetic) of the world's ocean basins have provided important constraints on the thickness and structure of the oceanic crust. Seismic reflection methods have been mainly used to investigate the nature of the upper crust, particularly the veneer of oceanic sediments, whereas elucidation of the deep structure of the oceanic crust and upper mantle comes from seismic refraction work.

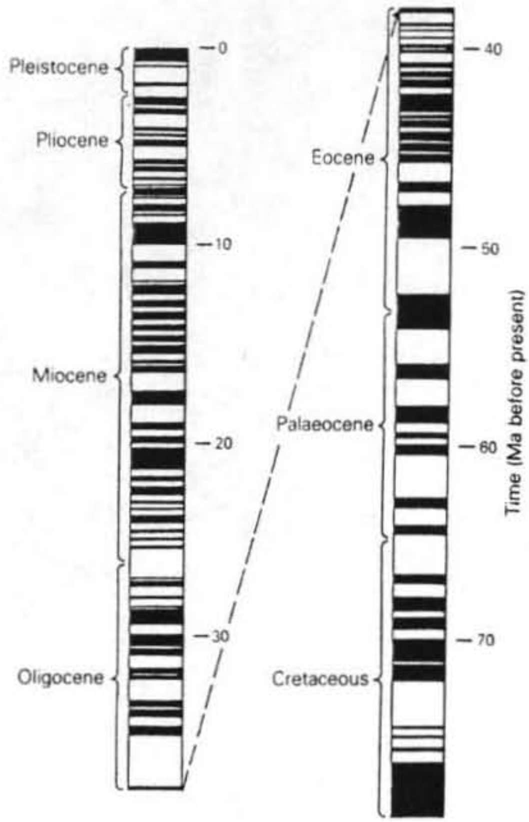


Figure 5.4 Palaeomagnetic record of reversals of the Earth's magnetic field over the past 80 Ma. The black intervals are normal (i.e. in the same sense as the present field) and the white intervals are reversed (after Brown & Mussett 1981, Fig. 6.4, p. 98).

Paleomagnetic studies

The Earth's magnetic field is highly variable and has reversed its direction many times throughout geological time, i.e. north palaeopoles become south palaeopoles and vice versa. Figure 5.4 shows the palaeomagnetic record of reversals in the past 80 Ma. In the early 1960s it was discovered that the ocean floor off the west coast of North America exhibited a regular pattern of magnetic stripes of alternating normal and reversed polarity (Mason & Raff 1961). This led Vine & Matthews (1963) to propose a model for the generation of the oceanic crust in which molten magma injected at the axis of the mid-oceanic ridge becomes magnetized in the direction of the Earth's prevailing magnetic field as it cools. This newly cooled material is subsequently pushed away from the ridge by the injection of new magma, forming stripes of alternately normal and reversed magnetism, depending on the polarity of the Earth's magnetic field when the magma solidified.

Magnetic anomaly patterns subsequently recorded in all the world's oceans (Fig. 5.5) reveal the same succession of magnetic stripes parallel to the mid-ocean ridge. The widths of individual stripes are different within each ocean due to the different spreading rates, but each ocean has the same sequence of reversals extending back to at least 76 Ma. Magnetic quiet zones occur beyond the oldest correlated anomalies, and in the Atlantic these lie near the margins of the continents. They are related to periods of infrequent polarity reversals in the late Mesozoic.

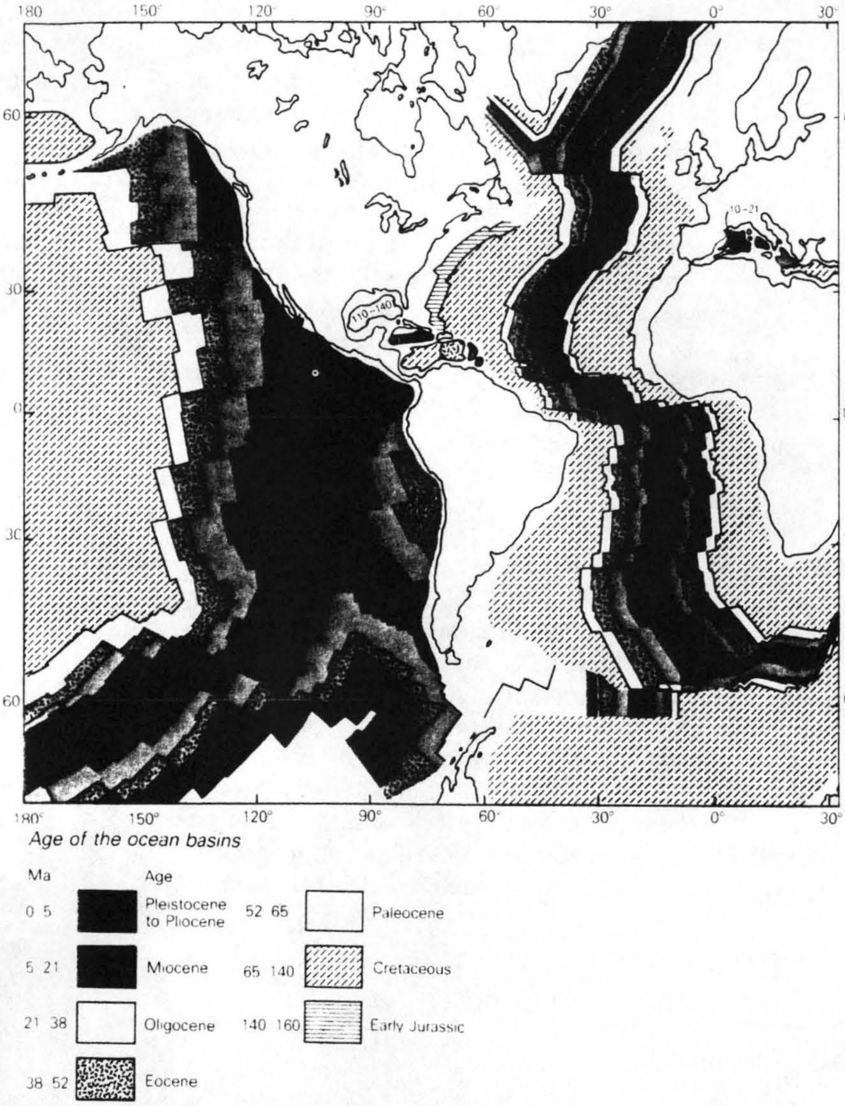


Figure 5.5 The magnetic anomaly pattern in the Atlantic and Pacific oceans. Note that the pattern in the Atlantic is symmetrical about the Mid-Atlantic Ridge, whereas that in the Pacific is highly asymmetrical (after Press & Siever 1982, Fig. 18.21, p. 432).

The pattern of magnetic anomalies shows that spreading rates are not constant throughout the length of the mid-oceanic ridge system, but vary from region to region. The spreading rate is normally quoted as a half-rate of separation from a bilaterally symmetrical axis. Rates vary from a few mm yr^{-1} to 8 cm yr^{-1} along parts of the East Pacific Rise. Differences in spreading rate appear to influence MOR topography (Section 5.4).

Seimic refraction data

Seismic refraction studies have revealed that, although structurally complex on a local scale, the oceanic crust in general possesses a rather consistent downward seismic velocity gradient which can be modelled in terms of three major layers, 1, 2 and 3, which grade into one another. Table 5.1 and Figure 5.6 show this layered structure in greater detail (Christensen & Salisbury 1975, Houtz & Ewing 1976, Rosendahl *et al.* 1976, Kennett 1982).

Table 5.1 The layered structure of the oceanic crust.

	V_p (kms^{-1})[a]	*Average thickness (km)*[b]	Average density (g cm^{-3})[b]
water	1.5	4.5	1.0
layer 1 – sediment	1.7–2.0	0.5	2.3
layer 2 – basalt	2.0–5.6	1.75	2.7
layer 3 – gabbroic/ ultramafic cumulate layer	6.5–7.5	4.7	3.0
		Moho	
upper mantle	7.4–8.6		3.4

Data sources: [a] Basaltic Volcanism Study Project (1981), Table 1.2.5.1., p. 133; [b] Kennett (1982), Table 7–1, p.207.

Layer 1: comprises a thin veneer of oceanic sediments.
Layer 2: is composed of basalt and can be subdivided into two sub-layers. 2A is a layer the seismic velocities of which are less than those predicted by laboratory measurements on basalts. This has been interpreted as indicating high porosities and low densities caused by the presence of cavities and fractures in the upper crust (Kirkpatrick 1979). The layer thins away from the ridge crest and is often absent in crust older than 20–60 Ma (Houtz & Ewing 1976). This is probably due to cementation effects accompanying diagenesis as the crust ages. *2B* is a layer with higher seismic velocities than 2A, more appropriate to basalt and metabasalt.
Layer 3: the precise nature of this layer is less obvious on the basis of seismic velocity data alone. Petrological arguments combined with ophiolite studies suggest that it is composed of gabbros and cumulate ultramafic rocks formed in high-level magma chambers (Cann 1974, Dewey & Kidd 1977). As with layer 2 this layer can also be subdivided into an upper layer, *3A*, and a lower layer, *3B*. Layer 3B apparently increases in thickness away from the ridge axis (Christensen & Salisbury 1975), growing at the expense of underlying asthenosphere either by off-axis intrusions or by underplating with mantle material (Clague & Straley 1977). Layer 3A is well developed and is characterized by relatively uniform velocities and thicknesses away from the ridge axis, although its presence at the ridge axis is a subject of debate (Rosendahl 1976).

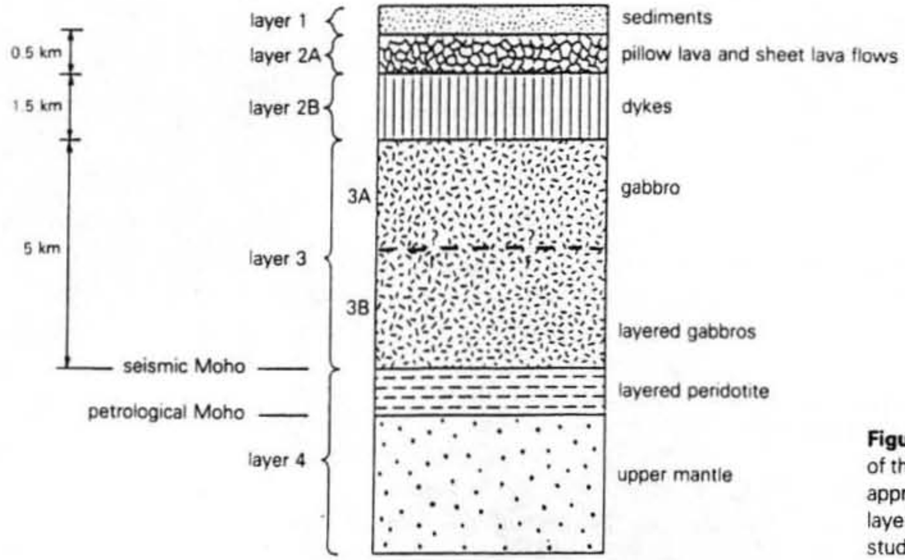


Figure 5.6 The layered structure of the oceanic crust, showing the approximate thickness of the layers as determined by seismic studies.

Mid-ocean ridges are elevated structures because they are composed of hotter and therefore less dense material than the surrounding plate. As the newly formed oceanic lithosphere moves away from the ridge axis it cools, becoming more dense, and subsides. A simple relation thus exists between the depth of the ocean and the age of the oceanic crust; mean ocean depth being proportional to the square root of the age. This is shown schematically in Figure 5.7a.

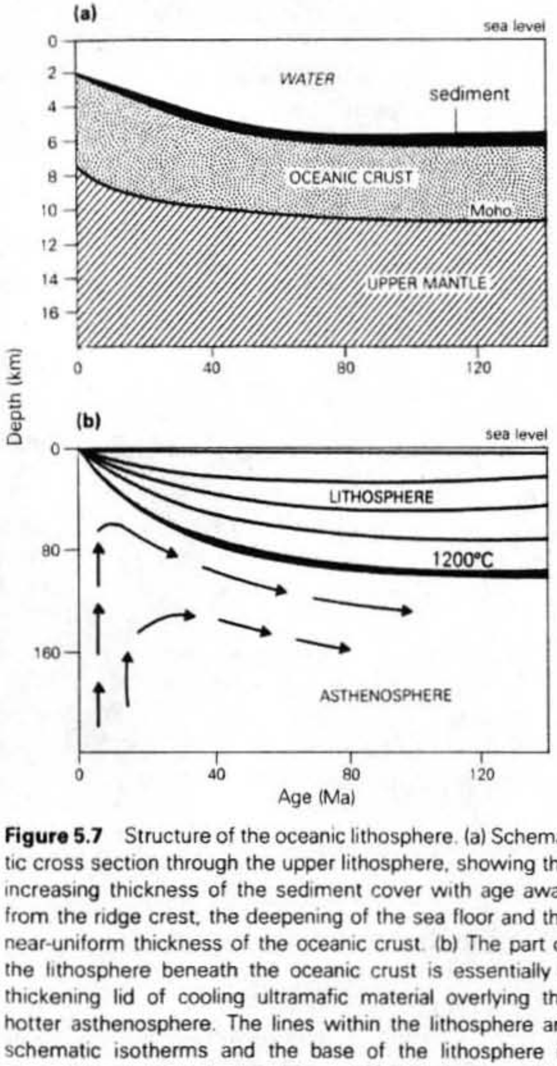


Figure 5.7 Structure of the oceanic lithosphere. (a) Schematic cross section through the upper lithosphere, showing the increasing thickness of the sediment cover with age away from the ridge crest, the deepening of the sea floor and the near-uniform thickness of the oceanic crust. (b) The part of the lithosphere beneath the oceanic crust is essentially a thickening lid of cooling ultramafic material overlying the hotter asthenosphere. The lines within the lithosphere are schematic isotherms and the base of the lithosphere is essentially an isothermal surface (~1200°C). Arrows within the asthenosphere indicate mantle flow lines away from the zone of upwelling at the ridge axis. (After Best 1982, Fig. 5.40, p. 183).

The thickness of the subcrustal lherzolitic lithosphere beneath layer 3 has been found to thicken exponentially with distance and therefore age away from the crest of mid-oceanic ridges (Fig. 5.7b; and see Leeds 1975). In the vicinity of the ridge crest it is virtually absent, but at distances from the axis where the crust exceeds 100 Ma in age its thickness becomes approximately 100 km. This thickening of the lithosphere as it ages is intimately related to its thermal evolution (Section 5.5.1).

5.3.2 Direct sampling of ocean-floor rocks

The first samples of ocean-floor rocks were obtained by dredging, and this technique is still in use largely because of its relatively low cost. Such samples obviously have no stratigraphic control and frequently come from anomalous areas such as fault scarps. Thus they can give us little information on spatial and temporal variations in MORB chemistry and mineralogy. The Deep Sea Drilling Project (DSDP), commencing in the late 1960s, gave a new dimension to the study of the oceanic crust, making it possible to define the vertical distribution of the various lithologies previously obtained by dredging. Unfortunately, it is still impossible to drill deep into the oceanic crust, maximum penetration being less than 1500 m. Additionally, most drill holes have to be sited on relatively old crust because a sediment cover of at least 80 m is required to hold the drilling units in place.

Most deep-crustal drilling has taken place in the North Atlantic, where it has been established that layer 2 is composed predominantly of pillow lavas with minor intercalated biogenic sediment to a depth of at least 600 m. However, until deeper penetration of the crust becomes possible the extact nature of the rocks at greater depths will remain unknown, and analogies with the lower parts of ophiolite complexes must be relied upon (Section 5.3.3). Deep-sea drilling has revealed a marked lack of lithologic and stratigraphic continuity in the crust, even between holes drilled only a few hundred metres apart. This indicates that seafloor eruptions are highly localized due to rapid chilling of the erupted magmas.

Technological advances since the 1970s have made possible the direct observation and sampling of the mid-ocean ridges by the use of submersibles. The FAMOUS (French American Mid-Ocean Undersea Study) project initiated in 1971 involved a detailed study of the Mid-Atlantic Ridge near the Azores at 37°N, using submersibles and remote-controlled instruments to collect samples. New navigation techniques allowed the mapping of topographic features to a scale of a few tens of metres and, combined with the detailed sampling, provided the first comprehensive description of a section of an active mid-ocean ridge (Moore *et al.* 1974, Ballard & Van Andel 1977, Ballard *et al.* 1975).

The FAMOUS project was followed in the late 1970s by similar large-scale projects in the Pacific (Van Andel & Ballard 1979, Ballard *et al.* 1981), concentrating on the Galapagos spreading centre and the East Pacific Rise at 21°N. Detailed surveys employing submersibles and bottom photography allowed mapping of the ridge topography and the distribution of various different morphological types of lava. Additionally, these studies led to spectacular discoveries of hydrothermal fields venting onto the ocean floor (Section 5.5.2; see also Corliss *et al.* (1979).

5.3.3 Ophiolites

Plate-tectonic theory provides an explanation for the transience of the ocean floors, being derived from the mantle and returning to it within a timescale of 100 Ma. Occasionally, portions of oceanic plates escape destruction at subduction zones and during the final stages of ocean closure become obducted onto one of the colliding continental forelands to form an *ophiolite*. Many ophiolite sequences have now been documented in detail (Coleman 1977, Gass *et al.* 1984). Their study theoretically allows elucidation of the deep structure of the oceanic crust which cannot yet be sampled by direct drilling. However, not all ophiolites need necessarily represent sections of normal oceanic crust; some may be fragments of marginal basin crust (Ch. 8).

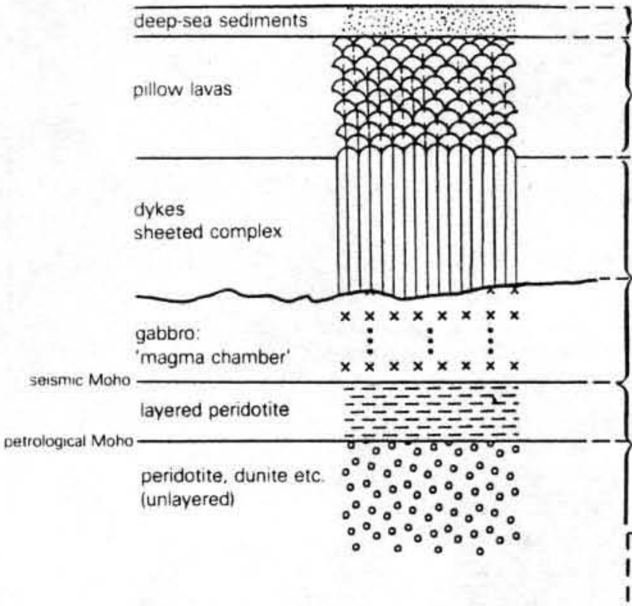
Figure 5.8 shows the theoretical structure of an ophiolite sequence and the correlation of the observed lithologies with the measured P-wave velocity profile of typical oceanic crust. Characteristically, seismic wave velocities are much lower in ophiolites than in normal oceanic crust. This may be related to the mineralogical changes attending their pervasive hydrothermal metamorphism. The upper portion of the model sequence comprises fine-grained Fe–Mn-rich mudstones, cherts, shales and limestones, i.e. typical deep-sea sediments. Both these sediments and the underlying pillow lava sequence contain Cu–Zn sulphides deposited by circulating hydrothermal solutions. The pillow lavas are the product of chilling of lava by sea water. In a typical succession they are apparently fed by a sequence of multiple dykes, the sheeted dyke complex, emanating from a basic magma chamber, now represented by coarse-grained gabbroic and ultramafic cumulates. A remarkable feature of such dykes is that they consistently exhibit one-way chilling, which is interpreted as the result of continued splitting in the axial zone.

The seismic distinction between ocean crust and upper mantle occurs where gabbros grade downwards into layered peridotites at the layer 3/layer 4 junction. However, the petrological Moho is in a different position, where layered peridotites, produced by crystal settling in magma chambers, change to massive peridotites that are part of the upper mantle.

Models for the formation of a typical ophiolite stratigraphy have had a profound influence upon models for magma chamber processes at mid-oceanic ridges (Cann 1970,1974, Robson & Cann 1982). These will be considered further in Section 5.8.

5.4 Structure of mid-ocean ridges

Mid-ocean ridges are made up of elevated volcanic mountains and valleys, in general rising about 2000 m from the adjacent ocean floor. Such ridges occupy some 33% of the total seafloor area and occur in all the major ocean basins. The system of ridges is essentially an oceanic phenomenon but in



	Typical ophiolite		Normal ocean crust	
	Thickness (km)	P-wave velocity (km s^{-1})	Thickness (km)	P-wave velocity (km s^{-1})
layer 1	c. 0.3	—	0.5	2.0
layer 2	0.3–0.7	3.3	1.7	5.0
layer 3	1.0–1.5	4.1	1.8	6.7
	2–5	5.1	3.0	7.1
layer 4	up to 7.0	2.5–4.2	—	8.1

Figure 5.8 Petrological, seismic and thickness data for a typical ophiolite sequence compared with the layered structure of the oceanic crust, as deduced from seismic studies (after Brown & Mussett 1981, Fig. 7.4, p. 114).

places it passes laterally into continental rift zones (Ch. 11). For example, recent volcanism in the Afar and Ethiopian rift system is the continental continuation of spreading processes operating at the Mid-Indian Oceanic Ridge. Rarely, a mid-oceanic ridge segment may become emergent, due to volcanic overproduction, as in the case of Iceland.

Topographically, the mid-oceanic ridges are very variable and this has been shown to correlate well with spreading rate. Fast-spreading ridges have rather smooth profiles, whereas slow-spreading ridges have jagged profiles and an axial rift valley (Table 5.2).

The nature and extent of volcanic activity in the deep ocean environment has been recognized only recently due to photographic surveys and detailed sampling programmes made from manned submersibles (Section 5.3.2). Unfortunately, these are restricted to a few areas of the mid-ocean ridge system, none of which are necessarily typical ridge segments (e.g. 37°N – FAMOUS, Galapagos Rise). The principal effects of the deep-sea environment are the almost complete elimination of explosive volcanism and a more rapid cooling rate due to quenching in water. Thus flows tend to be shorter and thicker than their subaerial counterparts. The major volcanic features are small volcanic cones, lava domes, flows and lava lakes.

5.4.1 Slow-spreading ridges

The Mid-Atlantic Ridge (MAR) is a typical example of a slow-spreading ridge (half-rate of 1–2 cm yr^{-1}; Hekinian 1982). Information about the structure of the ridge crest is limited to a few areas such as that between 36 and 37°N, known as the FAMOUS area (Section 5.3.2). The FAMOUS project commenced in 1971 using manned submersibles to document in detail the bathymetry, petrographic and geochemical variation of volcanic rocks from a small segment of the ridge crest. This segment is characterized by a broad 25–30 km

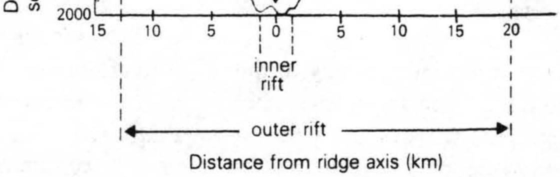
Table 5.2 Classification of mid-ocean ridges according to spreading rate.

		Latitude	Half-spreading rate (cm yr^{-1})	Source of data	
Fast	East Pacific Rise	21–23°N	3	a	No median valley; smooth topographic profile; basalts have higher FeO_T and TiO_2
		13°N	5.3		
		11°N	5.6	a	
		8–9°N	6	a	
		2°N	6.3	a	
		20–21°S	8	a	
		33°S	5.5	a	
		54°S	4	a	
		56°S	4.6	a	
Slow	Indian Ocean	south-west	1	b	Median valley; rugged topographic profile; basalts have lower FeO_T and TiO_2
		south-east	3–3.7	c	
		central	0.9	c	
	Mid-Atlantic Ridge	85°N	0.6	c	
		45°N	1–3	a	
		36°N	2.2	a	
		23°N	1.3	c	
		48°S	1.8	a	

Data sources: [a]Hekinian (1982); [b]Sclater *et al.* (1976); [c]Jackson & Reid (1983).

wide axial valley bounded by rift mountains (Fig. 5.9). Within this broad valley is a well defined narrow inner valley, 3–9 km wide, in which present volcanic activity appears to be concentrated. The flanks of this inner rift are fault controlled and have a veneer of thin lava flows originating on the valley walls and flowing inwards towards the rift axis. Small isolated volcanic hills, less than 300 m high, occur within the inner rift, defining the main focus of current activity. These are generally not split centrally as spreading continues but are carried as distinct units to one side of the valley (Ballard & Van Andel 1977), where they become dismembered by faulting. The fact that these volcanic hills are physically distinct features shows that volcanic activity is neither spatially nor temporally continuous.

The FAMOUS project documented marked compositional variations of basalts erupted within the inner rift zone. In general, the most primitive basalts (with high MgO, Ni and Cr contents) are associated with the central volcanic hills, whereas more evolved basalts are associated with eruptions at the margin of the rift (Fig. 5.10). Such compositional variations suggest the operation of low-pressure crystal fractionation processes and the



5.9 Schematic topographic profile across the central axis of the Mid-Atlantic Ridge in the FAMOUS area (36°N). The inset shows topographic profiles across the East Pacific Rise at 21°N and 13°N, drawn to the same vertical scale for comparison (after Francheteau & Ballard 1983, Fig.1).

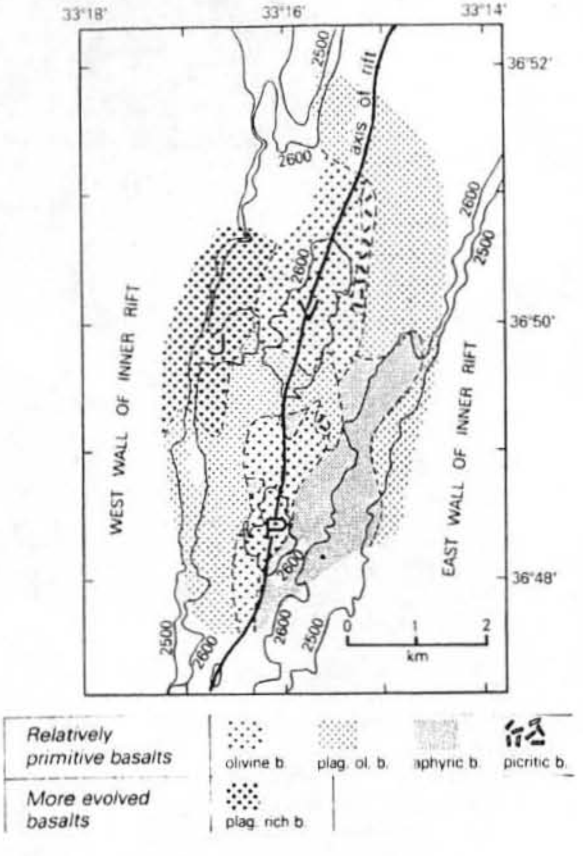


Figure 5.10 Distribution of the different types of basaltic rocks present in the inner rift valley of the Mid-Atlantic Ridge near 36° 50'N. Jupiter (J), Venus (V) and Pluto (P) are small volcanic hills defining the present focus of igneous activity along the ridge axis (after Hekinian 1982, Fig. 2.8, p. 73).

presence of some sort of high-level magma body beneath the axis of the inner rift. Two models for the nature of such magma bodies have been proposed (Hekinian 1982):

(a) A magma chamber the width of the inner floor (~ 3 km), in which case the central volcanic hills would simply be adventive cones.
(b) Each volcanic hill could have a small (<1.5 km wide) magma chamber beneath it, acting as an independent source for the volcanism.

The second model is most consistent with the geophysical data (Section 5.8), as a large magma chamber 3–4 km in diameter would attenuate S waves and this is not observed in the FAMOUS area.

5.4.2 Fast-spreading ridges

Information on the nature of fast-spreading ridges comes mainly from the Pacific, specifically the East Pacific Rise at 21°N and the Galapagos Rift, whose half-spreading rates are 6-7 cm yr $^{-1}$ (Crane & Ballard 1980, Spiess *et al.* 1980, Cyamex 1978, 1981, Francheteau & Ballard 1983). Unlike the Mid-Atlantic Ridge the ridge system of the Pacific is not symmetrical with respect to the bordering continents, most of the ridge being in southern latitudes and in the eastern part of the ocean basin (Fig. 5.1). Parts of the East Pacific Rise spread at half-rates of 8–9 cm yr $^{-1}$, greater than in any other part of the world.

Compared to slow-spreading ridges, the major difference in morphology is the lack of a well defined central rift valley (Fig. 5.11). In other respects there is a remarkable similarity in the dimensions and sequence of extrusive forms and the manner in which the ridge morphology changes from constructional volcanic along the axis to destructional tectonic along the margins. In both ridge types the central active extrusion zone is marked by a discontinuous line of small volcanic hills composed of sediment-free pillow lava (Fig. 5.12). In fast-spreading ridges these are flanked by plains of low relief made up of ponded lava lakes. These lava plains constitute the principal difference in volcanic morphology between fast- and slow-spreading ridges, and their formation is attributed to relatively high extrusion rates.

The total width of the extrusion zone along the East Pacific Rise is 2.5–3.0 km. Beyond this fissures and normal faults dominate, generating horst-like marginal highs. However, these faults have much less displacement than those bounding the inner rift in the FAMOUS area of the Mid-Atlantic Ridge. Active hydrothermal fields, 400–4000 m^2 in area, are associated with the ridge. These have only recently been discovered (Section 5.5.2) and may actually be a common feature of constructive plate margins, even slow-spreading ones, but have yet to be identified.

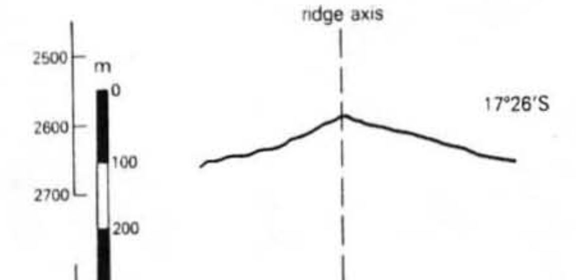


Figure 5.11 Bathymetric profiles across the fast-spreading East Pacific Rise at 17° 26'S, 18° 31'S and 21° 26'S, showing the essentially smooth topography of the ridge in comparison to the slow-spreading Mid-Atlantic Ridge (Fig. 5.9). Note the much expanded vertical scale in comparison to Figure 5.9. The vertical scale bar shows, for comparison, the relative depth of the inner rift valley (1000 m from floor to marginal highs) of the Mid-Atlantic Ridge in the FAMOUS area (after Renard *et al.* 1985, Fig. 3).

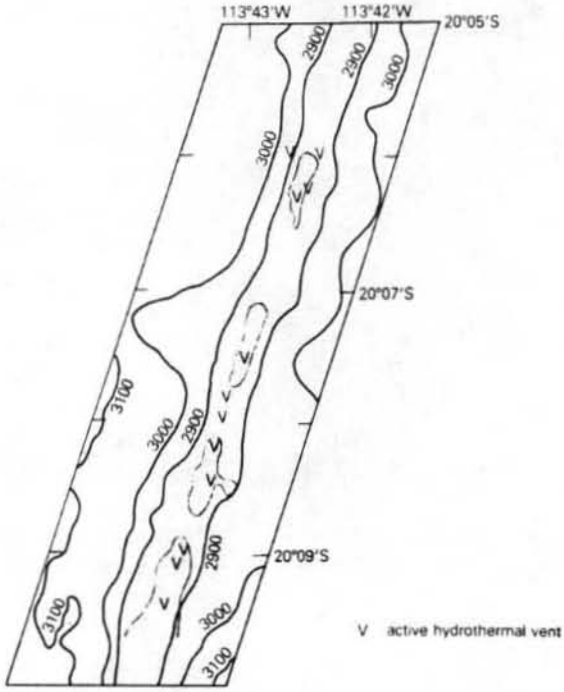


Figure 5.12 East Pacific Rise at 20°S, showing the bathymetry of the ridge axis (in metres). Shaded areas represent the youngest volcanic flows and these are clearly concentrated along the axial zone. Associated with these volcanics are active hydrothermal vents (after Francheteau & Ballard 1983, Fig. 13).

Petrographic and geochemical studies have shown the existence of a spectrum of primitive and more evolved basaltic magma types (e.g. Hekinian & Walker 1987). These are spatially distributed in a manner similar to that in the FAMOUS area, with the most primitive compositions erupted close to the ridge axis. However, fractionated basalts are the most common type sampled from the East Pacific Rise and this may be related to a fundamental difference in magma chamber processes beneath slow- and fast-spreading ridges. Certainly magma production rates at the EPR must be greater to sustain the high spreading rate, and this may lead to the establishment of much larger magma reservoirs beneath the ridge axis in which fractional crystallization can occur. This will be considered further in Section 5.8.

5.4.3 Transform faults and fracture zones

The dominant grain of the ocean floor is produced by the magnetic lineations that reflect the location of the spreading centre at the time of formation of the oceanic crust. This magnetic anomaly pattern is frequently offset by transform faults (fracture zones), often forming arrays subparallel to the direction of spreading (Fig. 5.13). The fracture zones are remarkably continuous features, normally extending for large distances across the flanks of the ridge and in some cases extending across the entire

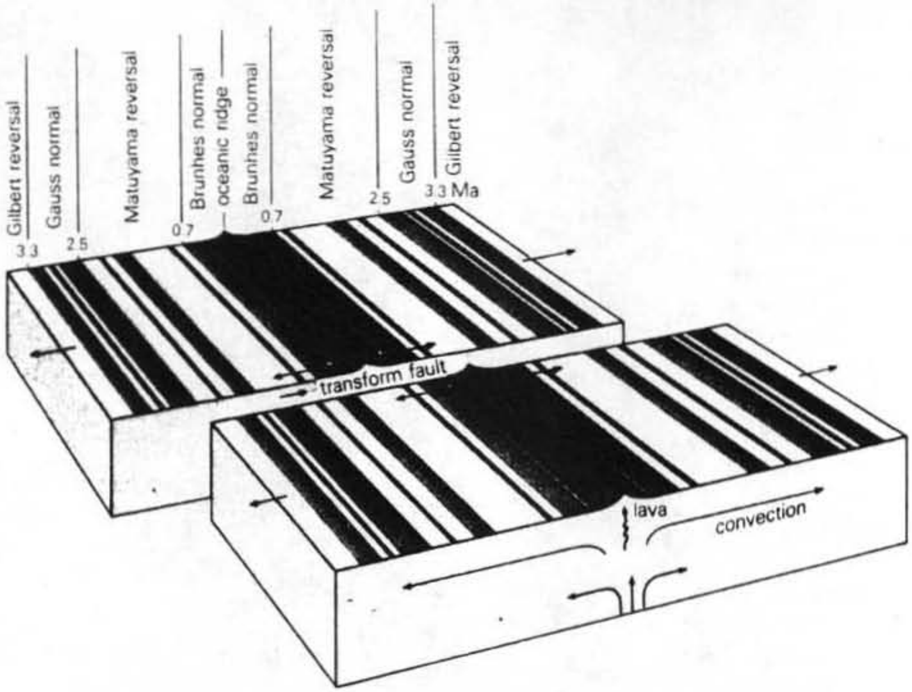


Figure 5.13 Offset of the magnetic anomaly pattern of the oceanic crust by a left lateral transform fault. Rocks of normal polarity are shown in black and of reversed polarity in white (after Press & Siever 1982, Fig. 18.17, p. 429).

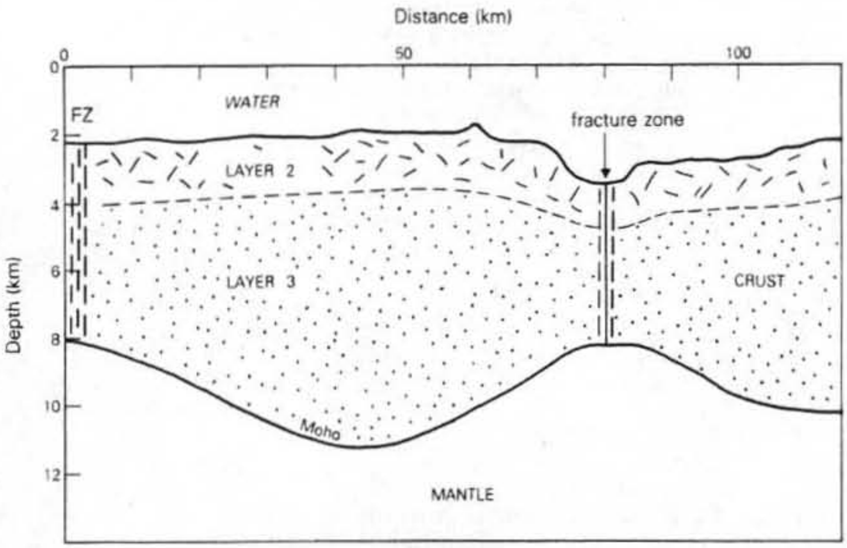


Figure 5.14 Schematic longitudinal section along the axis of a mid-oceanic ridge showing the variation in seafloor depth and crustal thickness in the vicinity of a fracture zone (FZ) (after White 1984, Fig. 4, p. 109).

ocean floor to the continental margin. This suggests that the locations of the active transform faults, from which the fracture zones originate, remain fixed with respect to the accreting plate boundary.

Fracture zones are usually marked by some irregularity in the topography of the ocean floor and are associated with shallow earthquakes generated by the lateral sliding of the adjacent segments of plate

(Fig. 5.14). Additionally, the oceanic crust thins adjacent to ridge–transform intersections, indicating a diminished rate of magma supply. Some fracture zones have a component of dip-slip movement and this may expose vertical sections through the oceanic crust. Dredge hauls from North Atlantic fracture zones have sampled a variety of lithologies including harzburgite, lherzolite, gabbro and amphibolite. A considerable amount of geophysical data from the North Atlantic ocean allows more detailed discussion of the nature of fracture zones (Johnson & Vogt 1973, Schouten & Klitgord 1982, Hekinian 1982). Here the average fracture zone spacing is ~50 km and many such fracture zones are traceable away from active transform faults at the ridge axis. Lateral offsets along the transforms vary from more than 20 km to effectively zero. However, even where the offset is small, the structural anomalies and discontinuities characteristic of the fracture zone persist.

A simplified model of the ocean floor (Fig. 5.15) suggests a pattern of strips of normal oceanic crust separated by narrow zones of anomalous crust. The normal oceanic crustal structure is formed in a string of individual spreading centre cells, whereas the anomalous crust is formed in the fracture zones. As more high-quality geophysical data become available, such a cellular pattern is becoming evident in all of the major oceanic ridge systems.

In some instances volcanism is associated with the transform, e.g. St Paul's Rocks and the Romanche Fracture Zone in the Atlantic and the Tamayo Fracture Zone at 23°N on the EPR. Basalts sampled from transform fault zones are characteristically more fractionated than average MORB. Bender *et al.* (1984) initiated a systematic study of the Tamayo zone to evaluate the effects of truncation of a normal spreading ridge by a large transform. For this particular section of the EPR, the normally broad swell morphology of the ridge changes to an axial rift valley morphology more characteristic of slow-spreading ridges within 20 km of the transform. Close to the transform a wider spectrum of magma compositions than normal MORB is observed, with a preponderance of differentiated ferrobasalts. Bender *et al.* (op. cit.) suggest that parental magmas close to the transform

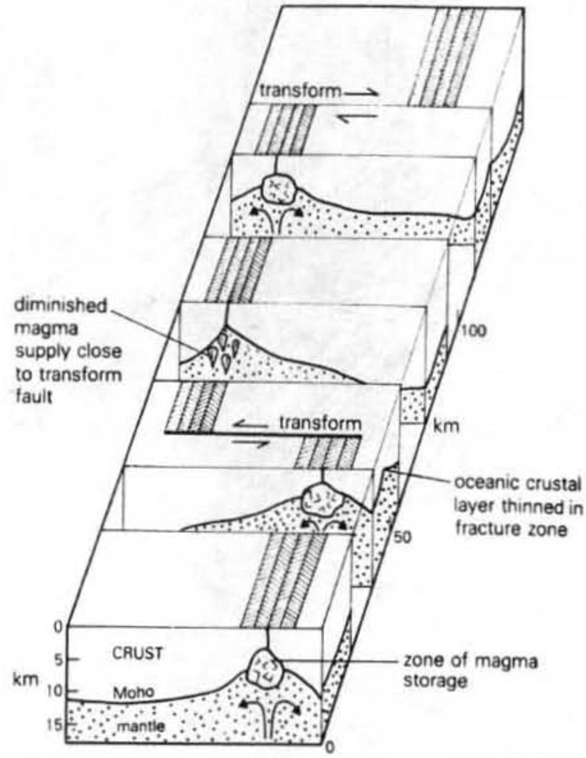


Figure 5.15 Simplified model of the ocean floor, showing the normal oceanic crustal structure developing in individual spreading centre cells separated by zones of anomalous crust generated in fracture zones (transform fault zones). Note the diminished rate of magma supply close to the fracture zone and the consequent thinning of the oceanic crustal layer. The zone of magma storage will only approximate to a large sub-axial magma chamber for fast-spreading ridge segments.

are derived from lower degrees of partial melting than those further away and, additionally, undergo more extensive fractional crystallization.

5.4.4 Propagating rifts

Propagating rifts are a special case of ridge–transform intersections in which one of the ridge segments begins to propagate into the older lithosphere adjacent to the transform, growing at the expense of the displaced ridge segment, which ceases to be active (Christie & Sinton 1981, Sinton *et al.* 1983). They are characterized by a diversity of erupted magma types extending from basalt through to rhyodacite, the latter representing the most differentiated magma type known from an oceanic spreading centre.

The process of propagation results in a failed rift that is offset from the propagating rift by a transform fault and a 'V'-shaped pair of magnetic anomaly offsets called pseudofaults, (Fig. 5.16), which strike obliquely to the spreading ridge and point in the direction of propagation. The pseudofaults mark the past positions of the propagating rift tip and separate crust formed at the propagating rift from the older crust through which the ridge propagated. In the vicinity of propagating rift–transform intersections, old and cold lithosphere bounds the active rift across the pseudofaults as well as across the transform. This unusual thermal regime may have a profound effect on the evolution of the propagating rift magmas.

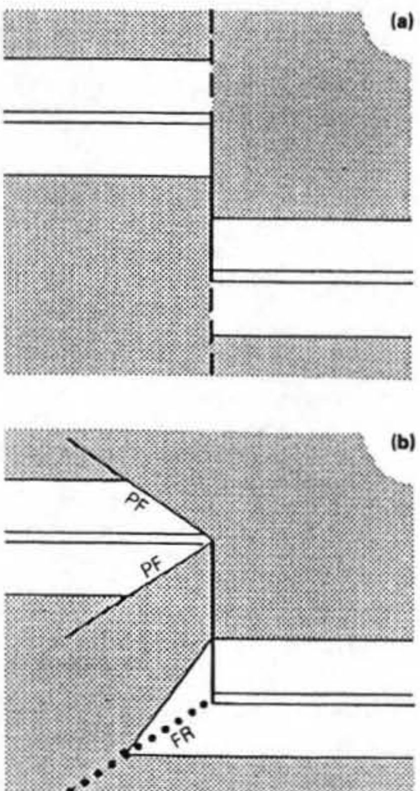


Figure 5.16 Schematic representation of normal ridge (a) and propagating ridge (b) transform intersections. Stippled areas denote all crust older than the axial anomaly which is unpatterned. Tectonic features specific to propagating rift systems include pseudofaults (PF) and a failed rift (FR) (after Sinton *et al.* 1983, Fig. 3).

The near uniformity of normal MORB compositions indicates that most ridges are characterized by steady-state magmatic processes involving balanced rates of supply and eruption. The association of highly differentiated lavas with propagating rifts suggests that this steady-state situation is not attained for some distance behind their tips. Propagating rift magmatic processes can be viewed as variations of those at normal ridge–transform intersections, in which the thermal environment is more conducive to high degrees of differentiation and thus the anomalous effects extend over an expanded length of ridge.

5.4.5 Anomalous ridge segments

Volcanics erupted along the strike of a mid-ocean ridge can show important compositional changes which appear to correlate with topographic and structural features (Le Douaran & Francheteau 1981, Francheteau & Ballard 1983, Schilling *et al.* 1983, Klein & Langmuir 1987). Topographic highs or volcanic platforms such as the Azores and the Galapagos have been termed 'hot spot' locations on their respective constructional plate boundaries. Such areas have crustal thicknesses intermediate between typical oceanic and continental values.

Figure 5.17 shows the variation of zero-age depth (i.e. depth to the floor of the inner rift) with latitude along the Mid-Atlantic Ridge (Le Douaran & Francheteau 1981). Sclater *et al.* (1975) demonstrated that these axial depth variations are not a young phenomenon, and have apparently persisted throughout the opening of the Atlantic. For example, the high between 26 and 31°N corresponds to the Canary Islands, and that between 13 and 22°N to the Cape Verde Islands. The major high between 34 and 42°N corresponds to the anomalous topography along the Azores–Gibraltar fracture zone, marking the triple junction between the European, North American and African plates. Superimposed upon the larger-scale anomaly pattern are relatively short-wavelength variations which are clearly associated with fracture zones. In general, an

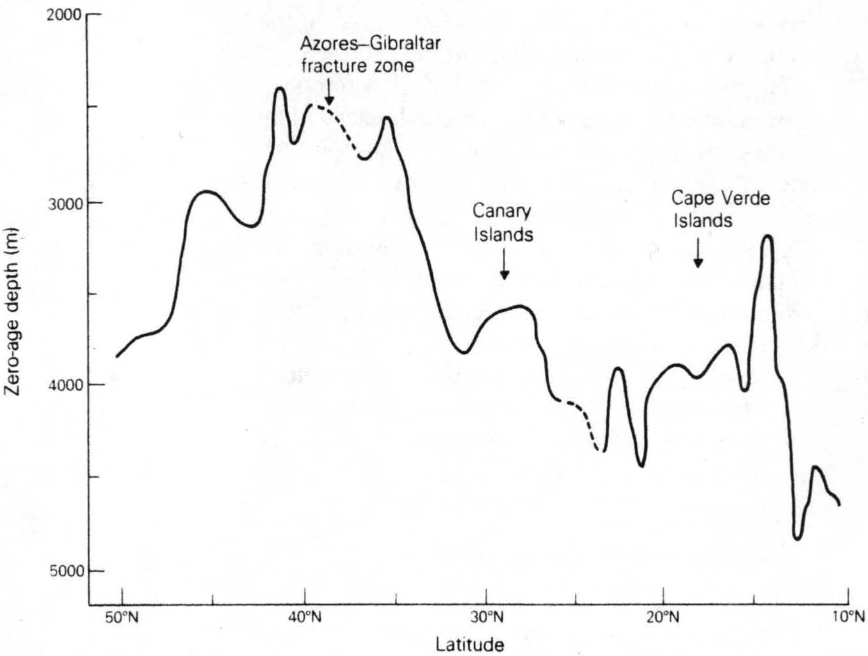


Figure 5.17 Zero-age depth variation with latitude along the axis of the Mid-Atlantic Ridge (after Le Douaran & Francheteau 1981, Fig. 3).

individual ridge segment displays a monotonic increase in depth towards the two adjoining transform faults from a single topographic high. The larger-scale depth anomalies correlate closely with geochemical anomalies in trace element and isotopic characteristics (Section 5.10) indicative of source heterogeneity. Such geochemical anomalies also seem to have been persistent throughout the opening of the Atlantic and testify to the relatively static nature of the mantle hot spots with respect to the accreting plate boundary. In comparison with the Atlantic, along-strike topographic variations along the East Pacific Rise are much smaller, of the order of 200–300 m, and the fracture zones are much more widely spaced.

The volcanism of Iceland (Imsland 1983, Gibson & Gibbs 1987), astride the Mid-Atlantic Ridge, can be taken as an extreme example of an anomalous ridge segment. This is the only large island that lies astride a spreading ridge and it is the most active volcanic region in the world. For these reasons Iceland has frequently been chosen as an ideal location for studying volcanic processes at constructional plate margins. However, its anomalous characteristics compared to the rest of the MAR in terms of topography, gravity, tectonic activity, crustal and mantle structure, seismicity and geochemistry must negate any generalizations.

The active plate boundary on Iceland is expressed as a series of tectonically and volcanically active rift zones, classified into axial and lateral types based on their tectonic structure and the chemical characteristics of the erupted magmas (Fig. 5.18). The axial rift zones are 30–60 km wide and erupt voluminous tholeiitic basalts from fissure swarms. These fissure swarms tend to be associated with central volcanic complexes in which volcanic activity is higher than normal, and subvolcanic magma chambers allow the fractionation of parental basalts to produce silicic differentiates. High-temperature geothermal fields are usually associated with the central complexes. In the south of Iceland two such axial zones run parallel to each other for a distance of approximately 100 km. In contrast, the lateral zones are characterized by the eruption of transitional and alkali basalts. Clearly, the constructive plate boundary on Iceland is considerably more complex than its submarine counterpart.

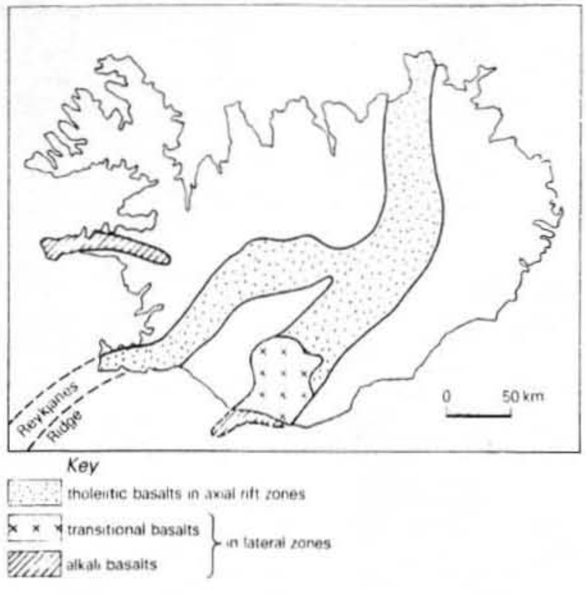


Figure 5.18 Outline of the active volcanic zones of Iceland, showing the distribution of the different basalt types (after Thy 1983, Fig. 1).

5.4.6 Aseismic ridges

Aseismic ridges are linear elevated volcanic structures rising 2000–4000 m above the surrounding ocean floor, varying from 250 to 400 km in width and 700 to 5000 km in length (Hekinian 1982). They comprise about 25% of the ocean floor, but are very poorly sampled and thus comparatively poorly understood features. They appear to represent chains of volcanic islands or seamounts that have subsided during their evolution.

The major aseismic ridges of the ocean basins are:

Atlantic	Iceland – Faeroe, Walvis Ridge – Rio Grande Rise
Pacific	Cocos, Carnegie
Indian	Ninety-East Ridge

Most are attached to a continental margin and they sometimes terminate in a volcanic island which generally forms a continuous structural feature with the ridge. For example, the Walvis Ridge in the South Atlantic Ocean extends from the volcanic island of Tristan da Cunha on the flanks of the Mid-Atlantic Ridge to the continental margin of Africa (Fig. 5.19). The ridges lack seismic activity and yet many are fractured perpendicularly to their axis in a closely similar manner to mid-ocean ridge fracture zones which are seismically active. All of the aseismic ridges are quite old and represent features which were formed during the early history of opening of the present ocean basins. The bathymetry across the ridges and along their length is smooth in comparison with mid-ocean ridge topography, and they are essentially made up of basaltic and more differentiated volcanics.

Morgan (1971, 1972a,b, 1983) suggested that aseismic ridges are the surface expression of plates moving over fixed hot spots in the mantle, analogous to the model proposed for linear chains of islands and seamounts in the Pacific (Fig. 5.20; see also Ch. 9). If this is correct then the age of the volcanic rocks in the aseismic ridge should decrease in the direction of the adjacent mid-oceanic ridge. This appears to be broadly correct for the Ninety-East and Walvis ridges.

From both gravity and seismic studies it is inferred that the ridges have a thicker crustal structure (15–30 km) than young oceanic crust (7–10 km; Figure 5.21). They are, in general, in isostatic equilibrium and their structure may contain large volumes of mafic and ultramafic cumulates. Chemically and mineralogically the volcanic rocks differ from MORB, though still dominantly tholeiitic, more closely resembling oceanic-island and seamount eruptives.

5.5 Heat flow and metamorphism

5.5.1 Oceanic heat flow

Measurements of oceanic heat flow (Fig. 5.22; see also Parsons & Sclater 1977) show that heat flow through mid-oceanic ridges is several times greater than through average ocean floor. This is precisely what would be predicted from plate-tectonic theory, as the ridge crest is the focus for intrusion and extrusion of high-temperature basaltic magma.

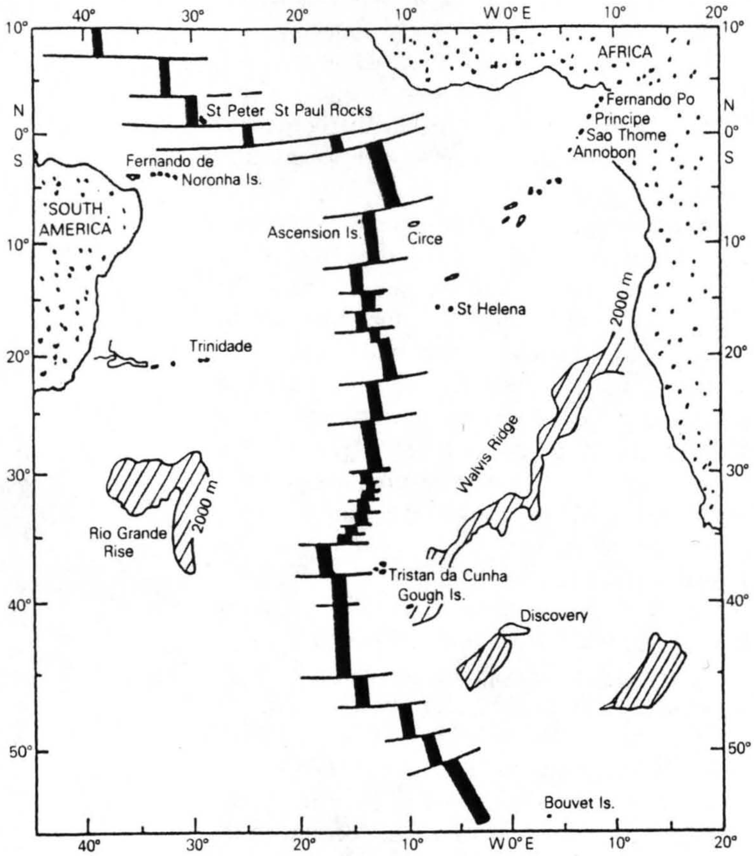


Figure 5.19 Paired aseismic ridges of the South Atlantic; Walvis Ridge – Rio Grande Rise (after Schilling *et al.* 1985, Fig. 1).

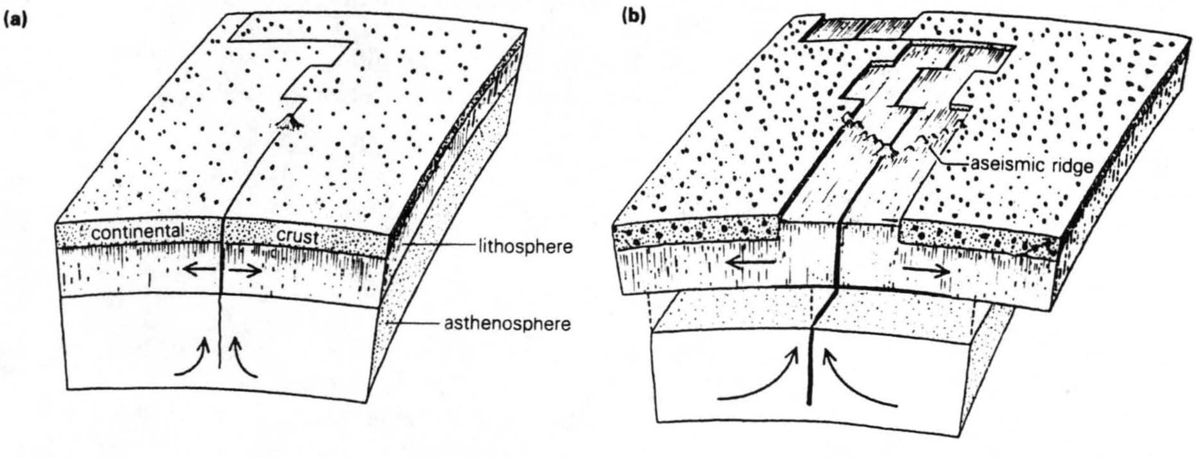


Figure 5.20 Hot spot model for the generation of paired aseismic ridges. (a) A hot spot exists at a continental rift zone, which becomes a site of successful rifting and the generation of a new ocean basin. (b) The hot spot remains a feature of the ridge axis in the growing ocean basin and is a focus of volcanic overproduction. Paired aseismic ridges develop, representing the trace of the hot spot in the newly formed oceanic crust.

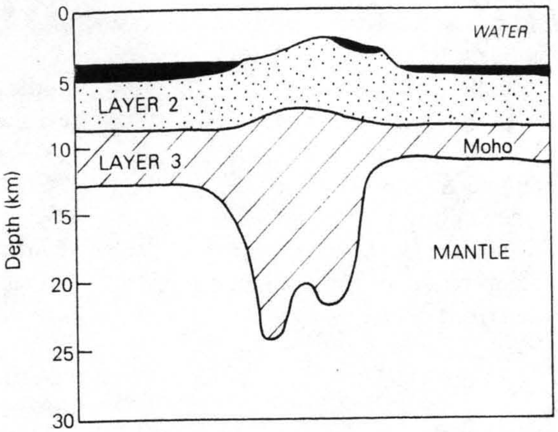


Figure 5.21 Crustal structure of the Walvis Ridge, showing the relative thickness of oceanic crustal layers 2 and 3 and the sedimentary layer (1), shaded black (after Hekinian 1982, Fig. 3.9, p. 163).

The oceanic plate is generated at the ridge and as heat flows out of its upper surface it cools and thickens (Fig. 5.7). Theoretical models of conductive heat flow give a simple equation to predict the heat flow through the plate (Sclater *et al.* 1980):

$$q = 11.3/\sqrt{t}$$

where q is the heat flow in HFUs (10^{-6} cal cm^{-2} s^{-1}) and t is the age of the crust in Ma. This equation is depicted as the solid curve in Figure 5.22 and clearly fits the measured heat flow data from the Pacific, Atlantic and Indian oceans quite well.

However, since the first oceanic heat flow measurements were made in the early 1950s it has become increasingly apparent that there is a very large scatter in the data obtained from mid-oceanic rift zones. Additionally, the mean of these data is consistently lower than the heat flow predicted by thermal models of seafloor spreading. This scatter

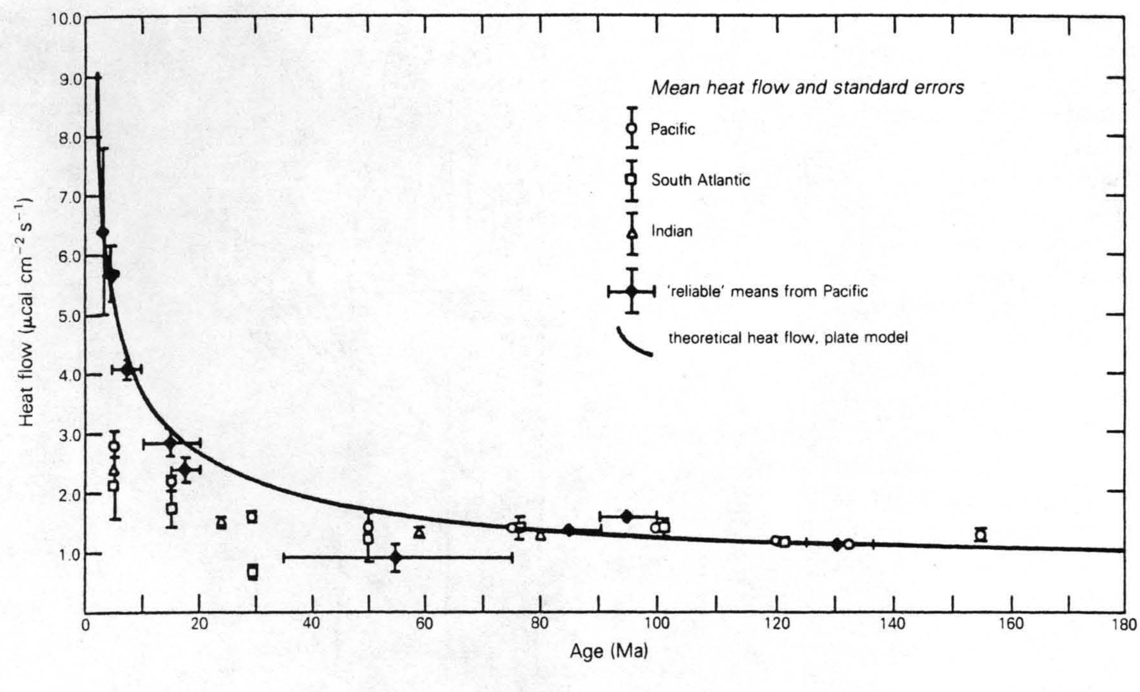


Figure 5.22 Changes in mean oceanic heat flow as a function of increasing crustal age away from the mid-oceanic ridge crest (after Parsons & Sclater 1977).

in the heat flow data can be attributed to permeation of sea water through fractures and cracks at the ridge crest, with the result that most of the heat loss is actually due to the advection of water through the crust. Evidence for this was provided during the 1970s by the discovery of extensive hydrothermal fields on the Galapagos ridge crest (Section 5.5.2). The scatter in the data decreases dramatically as the age of the crust increases and the heat flow values agree with the theoretical curves. This is due to the combined effects of the sealing of fractures due to ocean-floor metamorphism and the development of a sedimentary blanket over the crust.

5.5.2 Hydrothermal systems

The study of hydrothermal circulation in the oceanic crust is of primary importance in understanding its effect on the alteration of ocean-floor rocks. Exchange reactions between basaltic layer 2 and circulating hydrothermal solutions buffer the chemical and isotopic composition of seawater and can lead to the formation of metalliferous ore deposits near mid-ocean ridge crests. The amount of water incorporated into the crust by this process is a critical parameter in models for the generation of subduction-zone magmas (Ch. 6).

Water circulating through fissures and fractures in the oceanic crust near the ridge crest becomes heated, and ultimately re-emerges as hot springs on the ocean floor carrying various metals in solution. Initially, evidence for such hydrothermal activity came from studies of ophiolite complexes, particularly Troodos (Cyprus) and Oman (Section 5.3.3). In most ophiolites the pillow basalts are covered by several metres of metalliferous sediment called *umber*. These are often associated with lenticular iron sulphide ore bodies occupying depressions in the surface of the basalt, underlain by pipes of ore minerals. These pipes were clearly the conduits for ascending hydrothermal solutions.

Direct evidence for hydrothermal activity associated with mid-ocean ridge crests came in 1977 during a manned submersible study of the Galapagos spreading centre off the coast of Ecuador (Corliss *et al.* 1978). Subsequently, further north on the crest of the East Pacific Rise a second hydrothermal field was identified in which 350°C fluids, blackened by sulphide precipitates, were blasting upward through chimney-like vents as much as 10 m tall – the so-called 'black smokers' (Fig. 5.23; see also Edmond & Von Damm 1984). These chimneys protrude in clusters from mounds of sulphide precipitates and provide direct evidence for the mechanism of generation of the ophiolite complex ore bodies described above.

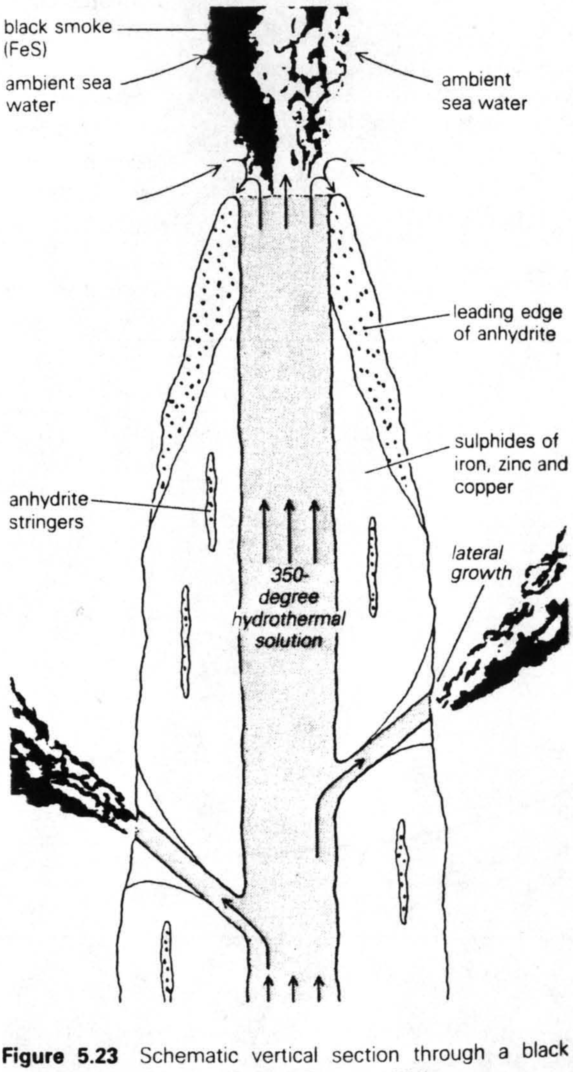


Figure 5.23 Schematic vertical section through a black smoker (after Edmond & Von Damm 1984).

5.5.3 Ocean-floor metamorphism

Studies of the oceanic crust in the 1960s and 1970s revealed that significant amounts of the crust are metamorphosed (Melson & Van Andel 1986, Cann 1969, Miyashiro *et al.* 1971, Hekinian & Aumento 1973). Dredge samples, especially those from the vicinity of fault scarps and transform faults, include greenstones, sepentinites and rare tectonized amphibolites. Such rocks have higher $^{87}Sr/^{86}Sr$ ratios than normal MORB, indicating that the water responsible for the growth of hydrous minerals was sea water. Much of this metamorphism has been ascribed to the circulation of hydrothermal brines through the oceanic crust in the vicinity of the ridge crest (Fig. 5.24).

Studies of the metamorphic mineral assemblages, field relations and tectonic fabrics of ophiolite complexes further support a model of dynamic hydrothermal metamorphism of the oceanic crust. Rapid increase in metamorphic grade stratigraphically downward into ophiolite complexes indicates geothermal gradients as much as several hundred degrees centigrade per kilometre, comparable with some of the highest heat flows measured at mid-oceanic ridges. However, the evidence for metamorphism within ophiolites must be viewed with some caution as, in many cases, it can be shown that they may have suffered several phases of metamorphism prior to and during obduction.

In marked contrast to the samples obtained by dredging, those from drill cores through the oceanic crust show rather limited alteration. This may in part be due to the restricted depth range of penetration of the crust (<600 m). Drill core samples in general show only zeolitic alteration along cracks and in vugs, with little evidence for major recrystallization.

The upper oceanic crust gains water by interaction with sea water through a series of metamorphic reactions involving the growth of chlorite, serpentine, smectite, illite and ultimately amphibole. The formation of such minerals is sequential and can be subdivided into three major stages (Staudigel *et al.* 1981):

- I formation of palagonite
- II formation of smectite
- III formation of carbonates

Stages I and II represent the stages of seawater–basalt interaction and involve large volumes of water, producing major chemical fluxes between the upper parts of layer 2 and the ocean reservoir. Both stages strongly deplete circulating solutions of alkalis and appear to end within a few Ma of formation of the crust. Stage III has a somewhat longer lifespan (< 10 Ma) and involves

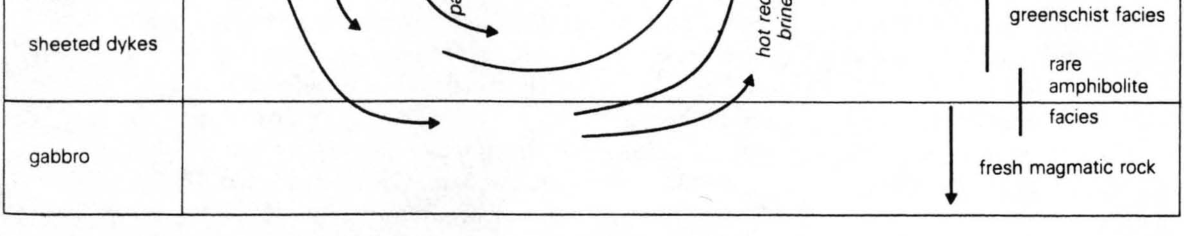


Figure 5.24 Convective circulation of sea water through the hot oceanic crust in the vicinity of a mid-ocean ridge crest (after Best 1982, Fig. 12.17, p. 430).

the release of Ca from basalt to the circulating solutions, culminating in carbonate precipitation to a depth of at least 500 m.

The basaltic layer 2 of young oceanic crust can be subdivided into an upper (low-velocity) layer 2A and a lower (high-velocity) layer 2B (Section 5.3.1). The depth of the boundary between 2A and 2B becomes shallower with increasing age, and layer 2A disappears completely at an age of approximately 70 Ma in the Atlantic. This disappearance is most readily explained by the filling of cracks and voids with secondary minerals.

Theoretically, the pervasiveness of ocean-floor metamorphism should decrease with depth, being directly related to the permeability of the oceanic crust. Generally, dredged samples are extensively altered because of biased sampling towards fault scarps and fracture zones. In contrast, cored samples show significantly less alteration, as considered previously. Figure 5.25 shows an estimate of the proportions of the various secondary minerals within the oceanic crust (Ito *et al.* 1983). This implies that perhaps only 15% of the oceanic crust is actually hydrothermally altered. This has profound implications for petrogenetic models for subduction-zone magmatism, which imply that subduction of altered oceanic crust recycles H_2O and Cl from the surface reservoir (atmosphere, hydrosphere and crust) back into the mantle (Ch. 6).

Ocean-floor metamorphism can produce significant compositional changes in the basaltic rocks of the oceanic crust. Some greenschist facies metabasalts preserve their original chemistry, apart from the addition of H_2O, but most show a marked decrease in CaO and significant variations in alkali and silica content. Thus caution must be exercised when considering the geochemistry of ocean-floor basalts, and fresh glassy material should be chosen whenever possible.

Albarede & Michard (1986) consider that the U content of the oceanic crust may be increased by 20% during hydrothermal alteration. In contrast, Pb is leached from the crust, resulting in subduction of oceanic crustal layers with high μ which may later contribute to the genesis of oceanic basalts with distinctive Pb isotopic compositions (Ch. 9).

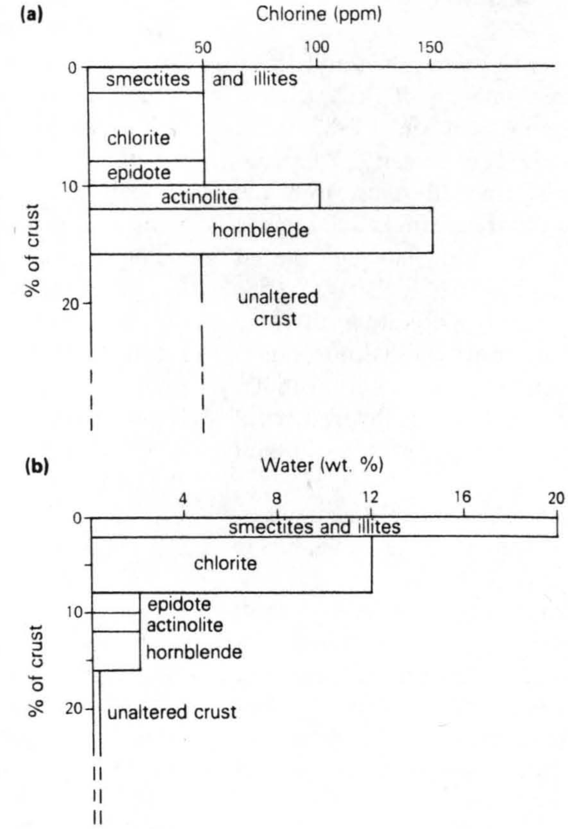


Figure 5.25 Proportions of various hydrous minerals in altered oceanic crust and the average concentration of (a) chlorine and (b) water in each mineral (after Ito *et al.* 1983, Figs. 2 & 3).

5.6 Convection systems at constructive plate margins

It is generally agreed that the generation of basaltic magma at constructive plate margins occurs in response to the upwelling of hot upper mantle material, which partially melts due to adiabatic decompression (Section 5.2). This upwelling is normally considered part of the large-scale convective motion of the asthenosphere. Convection is the principal form of heat transfer from the interior of the Earth, where heat is mainly generated by radioactive decay, and is linked via plate tectonic theory to the motions of the lithospheric plates (Ch.

3). However, it is still unclear as to how closely the shape, area and velocity of plates reflects the pattern of convection cells beneath. Additionally, the question of whether the convection extends in a single layer through the entire depth of the mantle (~3000 km) or is separated into two or more discrete layers is also undecided (Carrigan 1982, Houseman 1983a,b). One view is that convection in the upper mantle is separated from that in the lower mantle at a depth of about 650–700 km. This corresponds to the depth of significant phase changes in the upper mantle (marked by the 670 km seismic discontinuity) and also to the maximum depth to which subducted lithospheric plates can be traced. If the upper convecting layer is only 700 km deep the horizontal dimensions of the lithospheric plates imply large aspect ratios for upper mantle convection cells (Houseman 1983a).

Plastic flow in the upper mantle at the ridge crest must be driven by both large-scale phenomena related to plate drifting and by local phenomena related to partial fusion in the ascending mantle. The flow associated with the tectonic process is mainly driven by the cooling and sliding of the lithosphere and the buoyancy forces of the upwelling. However, superimposed upon this large-scale flow is a pattern of rapidly ascending partially molten mantle diapirs, which progressively expel their magma via dykes, lose their buoyancy and then flow away with the general circulation (Figure 5.26; see also Rabinowicz *et al.* 1984). In this model at shallow depths the upward flow is divided into two parts; one part closest to the axial plane is channelled into a narrow conduit with a 5–10 km half-width, while the remaining flow is deflected away more or less horizontally. In between, matter circulates on closed trajectories and thus stays permanently close to the ridge. The major asthenospheric upwelling is thus channelled into a narrow zone, at most 20 km wide in cross section, beneath the ridge axis.

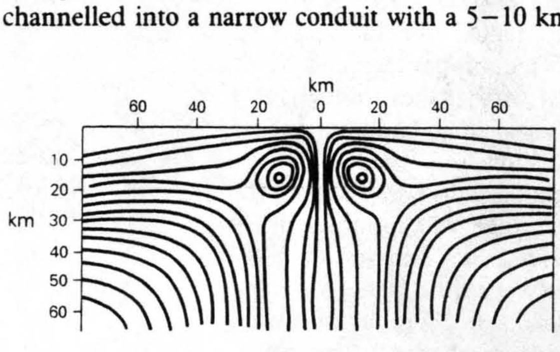


Figure 5.26 Theoretical flow pattern for mantle material rising beneath the axis of a fast-spreading oceanic ridge (after Rabinowicz *et al.* 1984, Fig. 5).

5.7 Partial melting processes

Partial melting beneath mid-oceanic ridges occurs in response to adiabatic decompression of ascending mantle lherzolite in the zone of upwelling (Ch. 3). In order to understand the processes involved, several important factors must be considered:

(a) The composition of primary mid-ocean ridge basalt magmas.
(b) The mineralogy of the source: plagioclase, spinel or garnet lherzolite.
(c) The degree of partial melting.
(d) The mechanism of partial melting (e.g. batch, fractional etc.).
(e) The depth of beginning of melting in the rising mantle material and the depth of segregation of the magma.

The enormous volume and apparent compositional uniformity of MORB (at least in terms of major element chemistry; see Section 5.10.2) led early workers to suppose that they were primary magmas, derived directly from the mantle without subsequent modification en route to the surface (Engel *et al.* 1965). However, O'Hara (1968) showed that this was unlikely and it is now accepted that the bulk of MORB are evolved magmas whose compositions have been modified by a variety of high-level processes including fractional crystallization, magma mixing and crustal contamination. It is therefore essential to characterize the compositions of primary MORB in order to provide constraints on the chemical and mineralogical composition of their source and on the extent of subsequent fractionation, contamination and mixing processes. Unfortunately, the nature of primary MORB compositions is still equivocal and detailed discussion will be deferred to Section 5.10, as many of the arguments are based on geochemical data. Essentially, the problem revolves around whether the most MgO-rich MORB compositions (10 wt.% MgO) are the primary magmas or whether the true parental basalts are in fact picrites.

Experimental melting studies provide some support for the picritic parental magma hypothesis, on the assumption that harzburgite (olivine + orthopyroxene) is the residue of the partial melting process (Ch. 3). If this is correct, then olivine and orthopyroxene should be on the MORB liquidus at pressures corresponding to the depth of segregation (O'Hara 1968). Green *et al.* (1979) and Stolper (1980) have shown that picritic liquids with 13–17% MgO are saturated with both olivine and orthopyroxene at 10–12 kbar pressure, whereas magmas with lower MgO contents do not have orthopyroxene on their liquidi at any pressure. Additionally, Presnall *et al.* (1979) predicted, by analogy with the CMAS system, that primitive MORB compositions should be saturated with olivine and orthopyroxene at about 9 kbar. These data, if directly applicable, would thus tend to suggest rather shallow depths of magma segregation of 30–40 km, within the stability field of spinel lherzolite. This is consistent with the trace element geochemistry of MORB (Section 5.10.3), which does not provide any evidence for the presence of residual garnet in the source.

Melting experiments on lherzolite bulk compositions can be used to deduce the degrees of partial melting necessary to generate typical MORB major element chemistry. Jaques & Green (1980; see also Ch. 3) have shown that tholeiitic basalt magmas can be produced by moderate degrees of batch partial melting (20–30%) of a lherzolite source at pressures below 15–20 kbar (50–60 km depth). At higher pressures, picritic liquids are generated at similar degrees of partial melting, while for greater degrees of melting (>35%) at all pressures the primary partial melts have komatiitic characteristics. It thus seems reasonable to assume degrees of partial melting somewhat in excess of 20% in the generation of MORB. However, this is apparently at variance with the much smaller degrees of partial melting predicted from studies of the trace element geochemistry of MORB (Section 5.10.3).

Delineation of the extent of partial melting beneath mid-oceanic ridges is theoretically possible by the study of seismic wave velocities, as both P and S waves are attenuated by regions of partial melt. Unfortunately, such studies are very few in number. A zone of unusually high attenuation has been discovered beneath the crest of the East Pacific Rise, extending from 20 to 60–70 km depth. The upper bound probably reflects the depth of segregation of the primary basalts from their mantle source, while the lower bound marks the onset of significant degrees of partial melting (Hekinian 1982).

5.8 Magma storage and release

Models of oceanic crustal generation by seafloor spreading usually assume the presence of a magma chamber beneath the ridge axis (Greenbaum 1972, Cann 1974, Robson & Cann 1982). This is required to explain the chemical and mineralogical diversity of MORB (Sections 5.10 & 5.11) via fractional crystallization and magma mixing processes and also the structure of ophiolite complexes (Section 5.3.3). Considerable effort has been made in recent years to search for such magma reservoirs, mostly using seismological techniques (McClain *et al.* 1985, McClain & Lewis 1980, Lewis & Garmony 1982). Evidence for the presence of high-level magma bodies is provided by attenuation of S waves and a marked lack of seismic activity in areas in which brittle deformation should be taking place. Conversely, efficient propagation of S waves and the occurrence of earthquakes throughout a crustal volume can be taken to indicate that no substantial body of magma is present.

It seems inevitable that magma rising beneath the axis of a mid-oceanic ridge will form stationary pools in the crust prior to its eruption onto the ocean floor. However, very little is known about the size and shape of such pools or chambers, despite a variety of theoretical models. Inferences about the nature of sub-axial magma reservoirs are based on a limited amount of data from scattered ridge segments, which need not necessarily be typical of their respective ridge systems. For example, geophysical studies of segments of the East Pacific Rise have revealed the existence of a low-velocity zone beneath the ridge axis (Orcutt *et al.* 1975, Rosendahl *et al.* 1976, McClain *et al.* 1985), which could represent a magma reservoir. In contrast, similar studies of the much slower spreading FAMOUS area of the Mid-Atlantic Ridge have failed to reveal any substantial axial magma chamber (Fowler 1976).

Axial magma chambers may exist as closed systems or periodically replenished open systems (Ch. 4), of which the latter is the most likely. Their size and persistence is intimately related to the spreading rate. Thermal modelling (Sleep 1975, Kuznir 1980) indicates the potential existence of a magma chamber beneath ridge crests spreading faster than $0.5-0.9$ cm yr^{-1} (half-rate). Below this critical spreading rate, no permanent magma chamber can exist on thermal grounds. As the spreading rate increases above the critical value the chamber must expand in size until at a half-rate of 6 cm yr^{-1} it will underlie 10 km of ocean floor on either side of the ridge axis (Kuznir 1980). Thus fast-spreading ridges are predicted to have large continuous magma chambers, whereas slow-spreading ridges will have small discontinuous magma reservoirs.

The nature of the sub-axial magma reservoir will obviously control processes of fractional crystallization and magma mixing. In general, a large magma chamber might be expected to erupt compositionally uniform, somewhat differentiated magma due to the efficiency of mixing processes. Small magma reservoirs, on the other hand, may undergo rather extensive differentiation and will thus tend to erupt a much wider compositional range from primitive basalt to more evolved ferrobasalts. Comparative studies of basalts from the Mid-Atlantic Ridge and East Pacific Rise appear to support this idea.

From seismic, petrological and theoretical thermal modelling studies, several models have been proposed for magma storage and release beneath the axes of mid-oceanic ridges. Two limiting cases will be considered here, appropriate to fast- and slow-spreading ridges respectively.

5.8.1 Large magma chamber: fast-spreading ridge segments

Cann (1970, 1974) proposed the existence of a large magma chamber beneath the axes of mid-oceanic ridges, partly on theoretical grounds but also to explain the layered structure of the oceanic crust (Section 5.3.1) and the structure of ophiolite complexes (Section 5.3.3). Such a chamber is shown schematically in Figure 5.27. As the ridge spreads, increments of lava are erupted through dykes feeding from the roof and, concurrently, magma crystallizes at the walls of the chamber to form isotropic gabbros. Crystal fractionation processes within the chamber produce a sequence of layered basic and ultrabasic rocks at the floor. Cann termed this model the 'infinite onion', as it continuously peels off layers at the edges.

Robson & Cann (1982) have considered in detail the influence of such a large open system magma reservoir on the chemical composition of erupted MORB. Magma feeding the high-level chamber should be close in composition to the primary basalt generated by partial melting of the mantle at depth, assuming that high-pressure crystal fractionation processes are not particularly significant in the evolution of MORB chemistry (Section 5.11). Although magma production beneath the ridge may be essentially continuous, magma batches will only rise up buoyantly through the asthenosphere on reaching a certain critical size. Thus the supply of magma to any high-level reservoir will be periodic, the rate of supply being proportional to spreading rate. In periods between the addition of new magma batches to the chamber, the chamber magma will undergo continuous open-system Rayleigh fractionation, producing more evolved liquids enriched in incompatible elements. As a new magma batch enters the chamber it can either mix with the chamber contents or form a temporary pool on the chamber floor. Here it may crystallize olivine cumulates before becoming sufficiently fractionated (and therefore less dense) to mix with the magma chamber contents. The chamber magma will be evolved (incompatible element enriched), and mixing with a primitive (relatively incompatible element depleted) magma will result

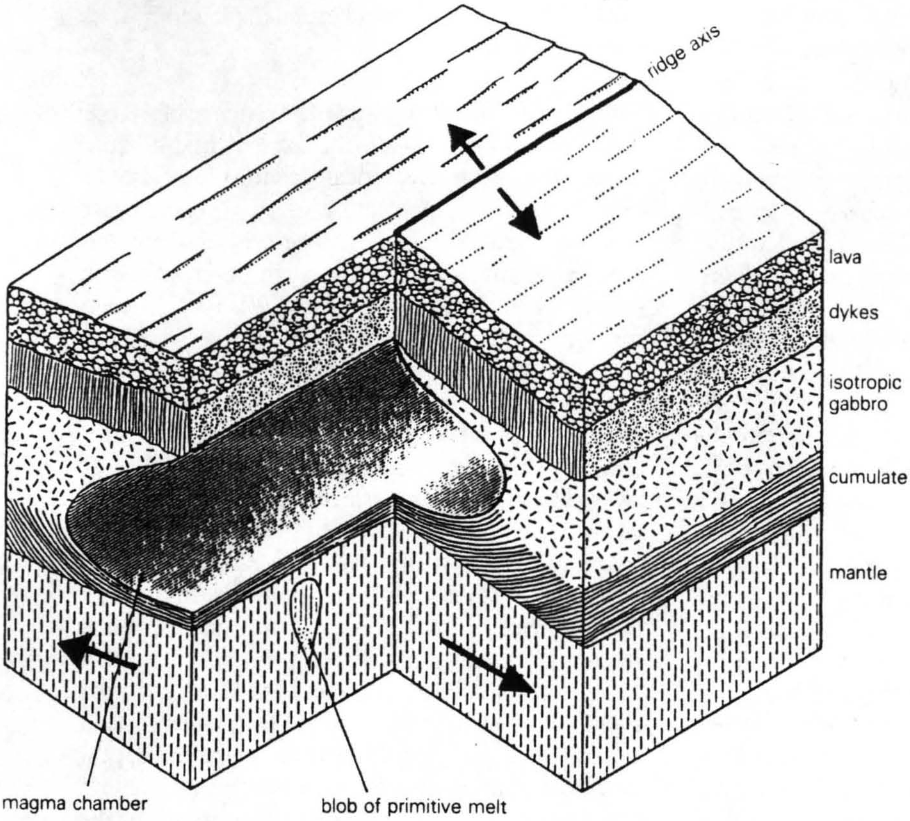


Figure 5.27 Schematic representation of the open-system magma chamber which may exist beneath all active mid-oceanic ridges (except those spreading extremely slowly). The chamber is periodically replenished by new magma batches, derived from partially molten mantle diapirs rising beneath the ridge axis, and is continually cooling and fractionating to produce new oceanic crust. The chamber size is directly proportional to the spreading rate; fast-spreading ridges have large continuous chambers while slow-spreading ridges have small discontinuous semi-permanent magma reservoirs (after Robson & Cann 1982, Fig. 1).

in its composition being reset to an intermediate composition more depleted in incompatible elements.

Lava will be erupted from the chamber when the magma pressure exceeds the lithostatic pressure and the strength of the chamber roof. This is most likely to be coincident, or nearly so, with the injection of a new batch of magma into the chamber. For small chambers this is likely to result in eruption of the fractionated chamber magma, followed closely by more primitive basalt. In a large magma chamber, appropriate to Cann's infinite onion model, magma mixing is likely to occur almost spontaneously and erupted magma will correspond to some fractionated 'perched state'.

A large axial magma chamber as proposed above should characteristically produce marked attenuation of S waves. Considerable support for such a model has come from seismic studies of the fast-spreading East Pacific Rise. However, the model cannot be considered generally applicable and, in particular, slow-spreading ridges appear to require a rather different model of magma storage and release (Nisbet & Fowler 1978).

5.8.2 Small ephemeral magma reservoir: slow-spreading ridge segments

Detailed studies of the FAMOUS area of the Mid-Atlantic Ridge in the late 1970s (Fowler 1976, 1978; Nisbet & Fowler 1978) cast serious doubts on the general applicability of Cann's infinite onion model to mid-oceanic ridge magmatic processes. In particular, there is no obvious attenuation of S waves beneath this segment of the MAR, suggesting that at present there can be no magma bodies larger than about 2 km across beneath the ridge axis. Additionally, thermal constraints (Sleep 1975) make the existence of a large axial chamber beneath such a slow-spreading ridge highly unlikely.

Nisbet & Fowler (1978) postulated an alternative model for crustal formation at slow-spreading ridges, assuming that no permanent sub-axial magma reservoir exists. This they termed the 'infinite leek', partly as a parody of Cann's 'infinite onion' model. In this model magma rises at high levels by a process of crack propagation through the brittle crust, allowing the existence of only very small storage reservoirs at very high levels. Figure 5.28 shows a comparison of the two models, emphasizing their capacity to produce a layered crustal structure. In general, the 'infinite leek' model will produce a poorly layered oceanic crust.

Closed-system fractional crystallization processes may dominate slow-spreading ridges if the magma is stored in very small ephemeral reservoirs. In contrast, open-system fractionation, combined with efficient magma mixing, must dominate the large axial reservoirs beneath fast-spreading ridges. This may partly account for the observed geochemical differences between MORB erupted at slow- and fast-spreading ridge segments (Section 5.10). The 'infinite leek' model may be regarded as a limiting case when the axial magma chamber is vanishingly small. As the spreading rate increases, small magma chambers may develop, eventually becoming large 'infinite onion' type chambers at half-rates in excess of a few centimetres per year.

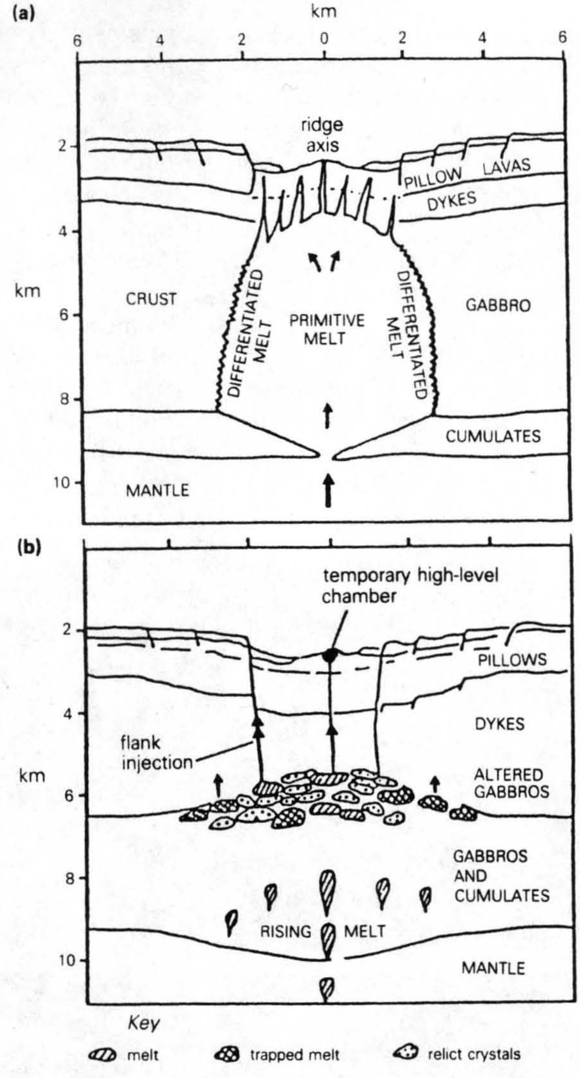


Figure 5.28 A comparison of the 'infinite onion' model (a) for magma chambers beneath fast-spreading ridge segments, and the 'infinite leek' model (b) for magma storage beneath slow-spreading ridge segments (after Nisbet & Fowler 1978, Fig. 6).

5.9 Petrography of mid-ocean ridge basalts

A detailed discussion of the petrography of the entire range of igneous rock types obtained from the oceanic crust by drilling and dredging is clearly beyond the scope of this work. For such detailed information the reader is referred to Hekinian (1982). Instead, attention is focused here on the petrography of mid-ocean ridge basalts (MORB), the most voluminous ocean-floor rocks sampled to date.

The petrographic characteristics of MORB reflect both the chemical composition of the magma and its cooling history. Fabrics reflect rapid cooling of near liquidus temperature magmas extruded into a cold submarine environment. Grain sizes are variable, from glassy to highly porphyritic types with 20–30% phenocrysts. Porphyritic basalts are common, and some highly phyric types are probably accumulative in origin.

The most commonly observed phenocryst assemblages are;

olivine ± Mg–Cr spinel
plagioclase + olivine ± Mg–Cr spinel
plagioclase + olivine + augite

Augite phenocrysts are rare and usually confined to rocks with abundant olivine and plagioclase. Olivine, spinel and calcic plagioclase are the first minerals to crystallize, followed by augite and then Fe–Ti oxides (Bender *et al.* 1978, Walker *et al.* 1979, Bryan 1983). The occurrence of olivine as the liquidus phase is consistent with models of MORB petrogenesis involving olivine fractionation from a more picritic primary magma (Section 5.11). Amphibole is exceedingly rare, being observed only in basalts with alkaline affinities and in cumulate gabbros. In the latter it is either a product of late-stage crystallization or hydrothermal alteration. Figure 5.29 shows a series of photomicrographs illustrating the petrographic variability of MORB. In some instances, the phenocryst minerals appear highly embayed and are clearly out of equilibrium with the host magma. This lends

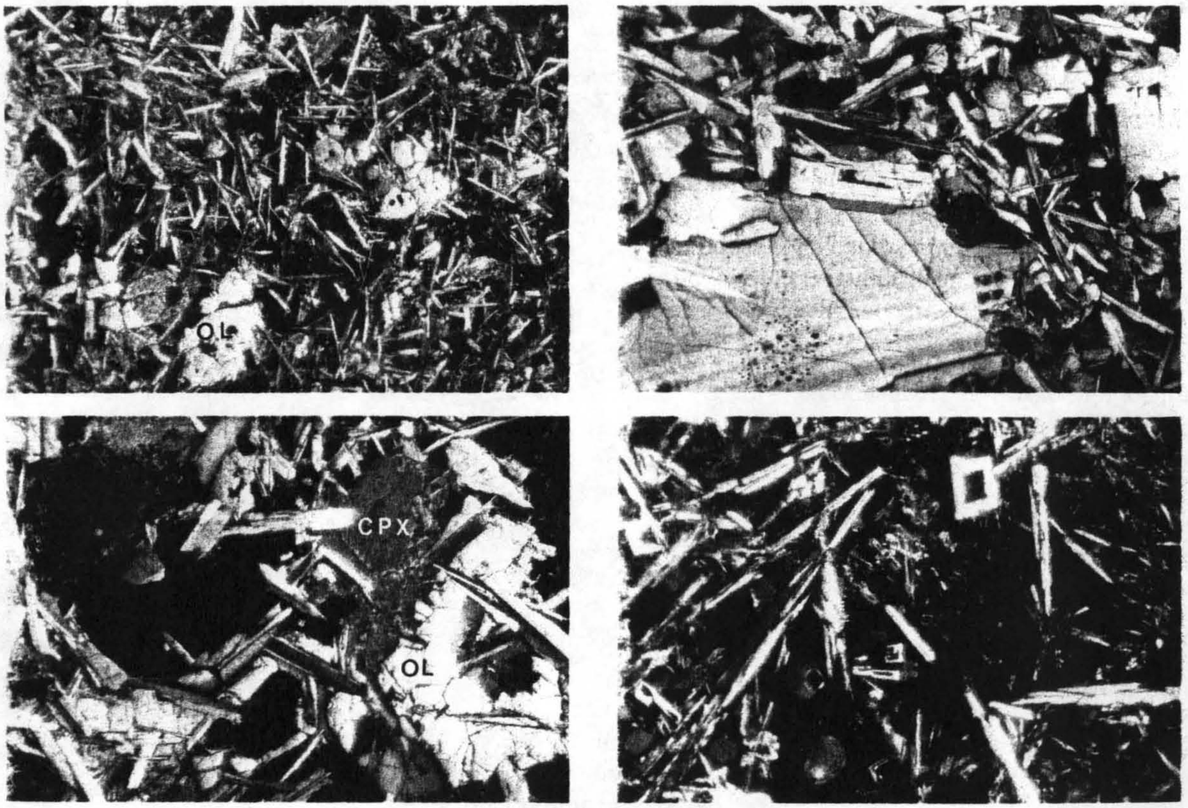


Figure 5.29 Photomicrographs illustrating the petrographic variability of MORB: (a) olivine phyric MORB from the FAMOUS area of the Mid-Atlantic Ridge (x40, crossed polars); (b) coarse-grained plagioclase phyric MORB from the FAMOUS area (x40, crossed polars); (c) olivine-clinopyroxene phyric MORB from Reykjanes Ridge (MAR) (x40, crossed polars); (d) fine-grained MORB from the Reykjanes Ridge, showing quench crystals of plagioclase (x100, crossed polars).

support to the importance of magma mixing in the evolution of MORB.

In general, basalts from normal and elevated (hot spot) ridge segments have different petrographic characteristics. In normal MORB, plagioclase is usually the dominant phenocryst phase, accompanied by olivine. Pyroxene is generally absent and bulk rock compositions may be modified by plagioclase and olivine accumulation (Section 5.10). In contrast, MORB from elevated ridge segments include both olivine and pyroxene phyric types. This may reflect both differing magma compositions and crystallization conditions in the two environments (Michael & Chase 1987).

The composition of olivine phenocrysts varies with the host magma composition, ranging from Fo_{73} in ferrobasalts to Fo_{91} in picrites (Fig. 5.30). It is generally euhedral in habit, becoming more anhedral in pyroxene-rich rocks. Moderate zoning of the olivine phenocrysts is common and in many cases the cores are too Mg-rich to be in equilibrium with the bulk rock, attesting to their derivation from a more mafic magma by magma mixing.

A spinel phase (Mg−chromite or Cr−spinel) is common in picritic and olivine-rich basalts, frequently occurring as tiny inclusions within olivine. This is rarely seen in plagioclase-rich basalts. Compositions of this spinel phase vary widely, with Al_2O_3 ranging from 12 to 30 wt.% and Cr_2O_3 from 25 to 45%. Extreme variations often occur within a single sample and are a consequence of the extreme sensitivity of the phase to fO_2 fluctuations (Fisk & Bence 1979).

Plagioclase compositions range from An_{88} to An_{40} and are uniformly orthoclase poor. As with the olivine phenocrysts they are freqently not in equilibrium with the bulk rock, again attesting to the importance of magma mixing processes. In general, there is a negative correlation between the An content of early formed plagioclase and the spreading rate. The fast-spreading East Pacific Rise has plagioclase compositions in the range An_{56-88} whereas the slower-spreading Mid-Atlantic Ridge has more calcic plagioclase phenocrysts (An_{75-92}) (Hekinian 1982). This reflects the fact that EPR basalts are more evolved than MAR basalts.

MORB clinopyroxene phenocrysts are colourless to pale green (in thin section) diopsidic augites, of generally very restricted chemical composition, clustering around Wo_{35-40} En_{50} Fs_{10-15} in the pyroxene quadrilateral (Fig. 5.31). Subcalcic augite and Mg−pigeonite are rare.

Gabbroic rocks dredged from the ocean floor show considerable overlap in composition with mid-ocean ridge basalts. Mineralogically, they comprise plagioclase, olivine, clinopyroxene, orthopyroxene and accessory minerals such as sphene, hornblende, apatite and titanomagnetite. Orthopyroxene-bearing gabbros are not common and are usually highly altered. However, their existence is significant as orthopyroxene is not observed as a phenocryst phase in the erupted basalts (Section 5.11). Textures and modal proportions of the above minerals vary widely, as might be expected in rocks of an essentially accumulative origin.

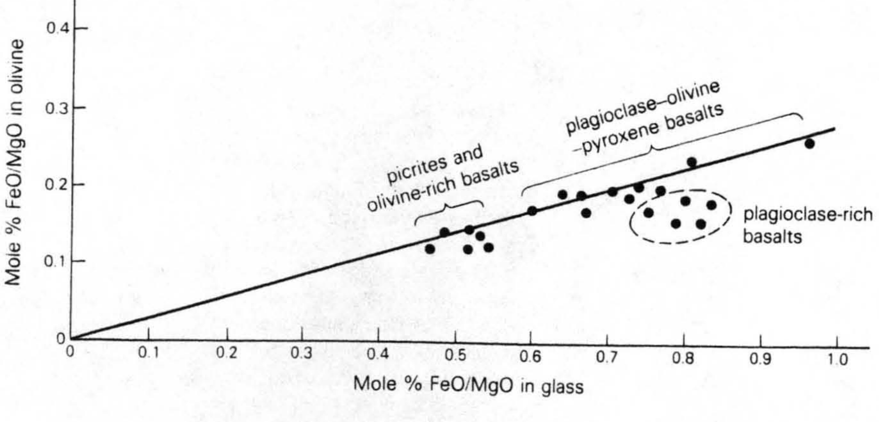


Figure 5.30 Correlation of molecular FeO/MgO ratio in olivine and coexisting basalt glass for MORB from the FAMOUS area of the Mid-Atlantic Ridge (after Hekinian 1982, Fig. 1.15, p. 42).

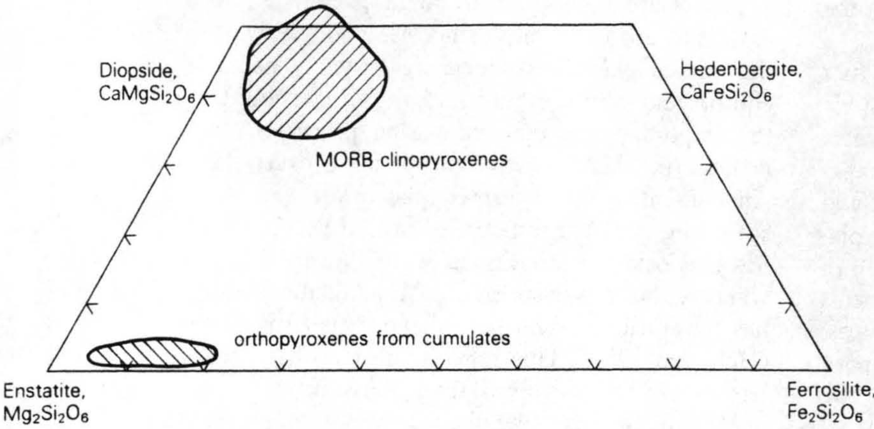


Figure 5.31 MORB pyroxene compositions projected into the pyroxene quadrilateral (after Basaltic Volcanism Study Project 1981, Fig. 1.2.5.5, p. 137).

5.10 Chemical composition of erupted magmas

5.10.1 Characteristic magma series

The majority of mid-ocean ridge basalts (MORB) are sub-alkaline and tholeiitic, according to the classification schemes given in Chapter 1 (see also Fig. 5.32). Alkali and transitional basalts occur only rarely, associated with seamounts, aseismic ridges and fracture zones. In terms of their major element chemistry, MORB are broadly similar to oceanic-island tholeiites (Ch. 9), island-arc tholeiites (Ch. 6) and continental flood tholeiites (Ch. 10) (Table 5.3). However, compared to such basalts they show characteristically low concentrations of incompatible elements, including Ti and P and LIL elements (K, Rb, Ba) (Section 5.10.3). Low K_2O contents appear to be a particularly useful discriminant in distinguishing MORB from basalts erupted in other tectonic settings (Pearce 1976). Hawaiian (ocean-island) tholeiites typically have lower Al_2O_3 contents than MORB at comparable degrees of differentiation, suggesting a difference in the Al_2O_3 contents of the primary magmas. This could be related to differing source compositions, degrees of partial melting and residual source mineralogies or to the high-pressure fractionation of an Al-rich mineral in the evolution of oceanic-island tholeiites (Ch. 9). Available data on the geochemistry of MORB are strongly biased towards the Atlantic Ocean. Much of the discussion in the

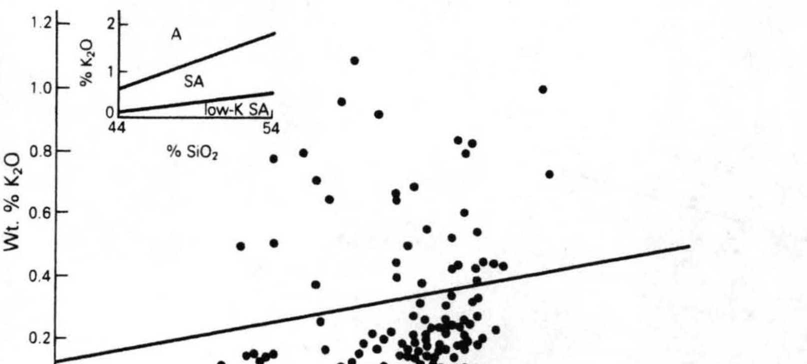


Figure 5.32 The variation of wt% K_2O versus wt% SiO_2 for basalts from 29°N to 73°N along the Mid-Atlantic Ridge. Most of the basalts fall in the low-K subalkalic field, although basalts from elevated ridge segments are richer in K_2O (data from Schilling *et al.* 1983). Boundaries between the alkalic (A), sub-alkalic (SA) and low-K subalkalic (low-K SA) fields from Middlemost (1975).

Table 5.3 Comparison of the major element goechemistry of MORB with that of a typical oceanic-island tholeiite, island-arc tholeiite and continental flood tholeiite.

	MORB[a]			OIT[b]	IAT[c]	CFT[b]
	MAR	EPR	IOR			
SiO_2	50.68	50.19	50.93	50.51	51.90	50.01
TiO_2	1.49	1.77	1.19	2.63	0.80	1.00
Al_2O_3	15.60	14.86	15.15	13.45	16.00	17.08
FeO	9.85	11.33	10.32	9.59	9.56	10.01
Fe_2O_3	—	—	—	1.78	—	—
MnO	—	—	—	0.17	0.17	0.14
MgO	7.69	7.10	7.69	7.41	6.77	7.84
CaO	11.44	11.44	11.84	11.18	11.80	11.01
Na_2O	2.66	2.66	2.32	2.28	2.42	2.44
K_2O	0.17	0.16	0.14	0.49	0.44	0.27
P_2O_5	0.12	0.14	0.10	0.28	0.11	0.19

Data sources: [a] Melson *et al.* (1976); [b] Basaltic Volcanism Study Project (1981), Tables 1.2.6.2 & 1.2.3.2; [c] Jakes & White (1972).

MAR, Mid-Atlantic Ridge; EPR, East Pacific Rise; IOR Indian Ocean Ridge; OIT, oceanic-island tholeiite; IAT, island-arc tholeiite; CFT, continental flood tholeiite.

following sections thus relies heavily on Atlantic data sets, comparison with Pacific and Indian ocean MORB being made whenever possible.

5.10.2 Major elements

Early studies of MORB emphasized their compositional uniformity, but since the 1970s significant geochemical variations have been observed, suggesting that a variety of magmatic processes and a heterogeneous mantle source region are involved in their genesis (Le Roex *et al.* 1983, Schilling *et al.* 1983, Le Roex 1987).

SiO_2, TiO_2, Al_2O_3, Fe_2O_3, FeO, MnO, MgO, CaO, Na_2O, K_2O, P_2O_5 and H_2O can all be considered as major element oxides in the description of MORB geochemistry. For most mid-oceanic ridge segments, SiO_2 shows a remarkably narrow range of variation, from 47 to 51%, and therefore it cannot be used successfully as an index of differentiation. Only in propagating rift segments (e.g. Galapagos), transform fault zones and anomalous ridge segments (e.g. Iceland) does extensive fractionation produce more silicic differentiates.

Table 5.3 shows a representative range of MORB glass compositions from the Atlantic, Pacific and Indian oceans. When comparing analyses of MORB from different provinces it is important to use glass compositions whenever possible, as bulk rock analyses can differ from the original magmatic liquid compositions due to the accumulation of olivine and plagioclase. There appear to be no significant inter-ocean differences in major element chemistry, apart from a tendency for Pacific MORB to be slightly more Fe- and Ti-rich.

As SiO_2 is unsuitable as an index of differentiation, MgO content or M value ($100\ Mg/(Mg + Fe^{2+})$) is used instead to illustrate differentiation from primitive to more evolved compositions. Figure 5.33 shows the frequency of occurrence of M values for MORB glasses from the major ocean basins. Although there is a wide range of values there is a very obvious maximum in the data between 55 and 65. A value of 70 defines a basaltic magma in equilibrium with mantle olivine (Section 2.4), and thus the diagram reveals the rarity of primitive glass compositions amongst the spectrum of erupted MORB. Relatively fractionated magma compositions appear to be dominant, indicating that the primary MORB magmas must have

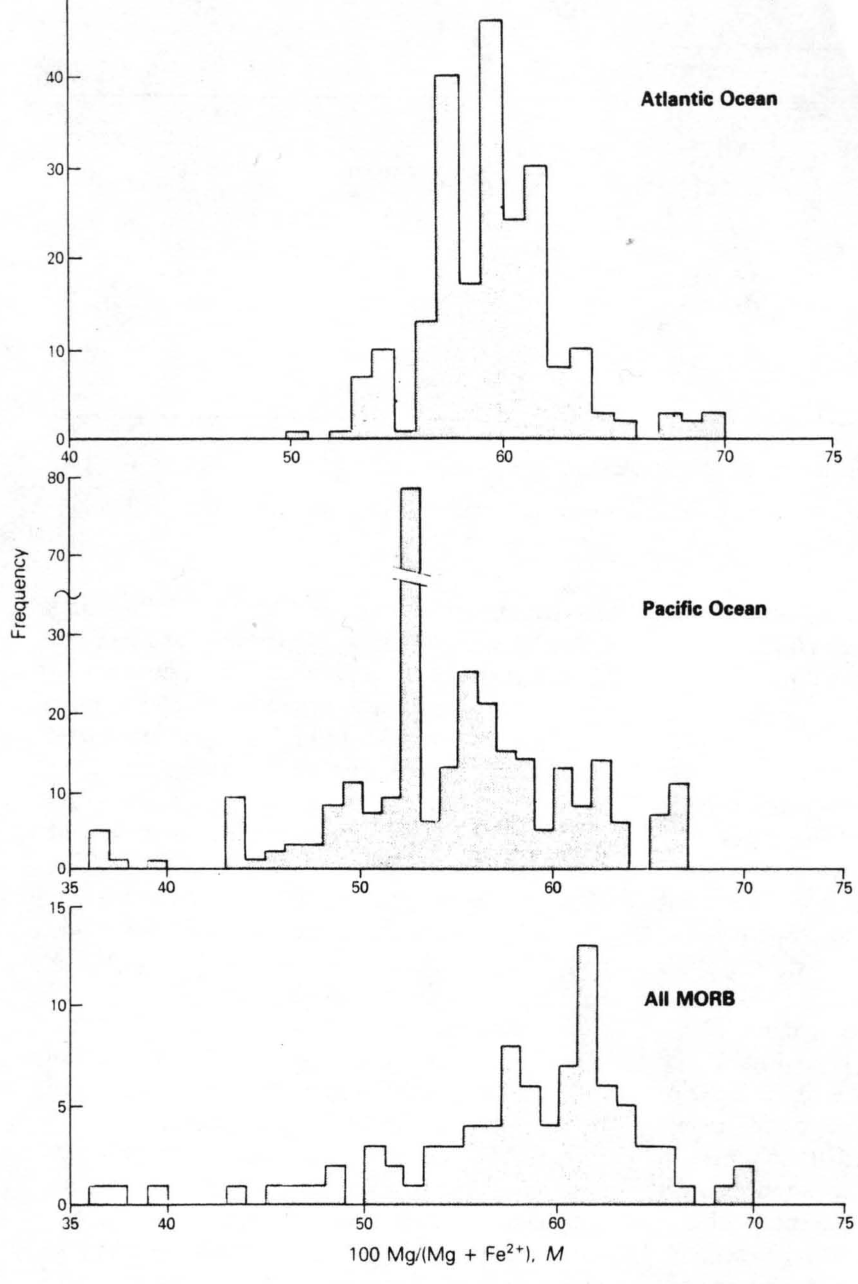


Figure 5.33 Frequency of occurrence of M values for MORB glasses from the major ocean basins: $M = 100\mathrm{Mg}/(\mathrm{Mg} + \mathrm{Fe}^{2+})$ (after Wilkinson 1982, Figs 1–3).

undergone high-level fractionation after segregation from their mantle source.

Volcanics erupted along the strike of mid-ocean ridges can show important compositional changes which appear to correlate with topographic and structural features that characterize particular ridge segments (Hekinian 1982, Klein & Langmuir 1987). Topographic highs and volcanic platforms appear to be associated with 'hot spots' and have positive free-air gravity anomalies, high geothermal gradients and crustal thicknesses intermediate between oceanic and continental values, in addition to erupting MORB of a rather distinctive trace element geochemistry (Section 5.10.3). On this basis MORB have been classified as normal (N-type, depleted), plume (P-type, enriched) and transitional (T-type) (Bryan *et al.* 1976, Sun *et al.* 1979, Schilling *et al.* 1983). However, despite significant variations in the trace element concentrations of basalts erupted along these different types of ridge segment, major element compositions remain remarkably uniform (Table 5.4). N-type basalts are recovered mostly from the Pacific and from the Atlantic south of 30°N, whereas P-type basalts

Table 5.4 Major (a) and trace element (b) geochemistry of average primitive (M= 60–70) normal, plume and transitional type MORB from the Mid-Atlantic Ridge (data from Schilling *et al.* 1983, Table 3).

	Normal MORB		Plume MORB		Transitional MORB	
	28–34°N	49–52°N	Azores	Iceland	34–38°N	61–63°N
(a)						
SiO_2	48.77	50.55	49.72	47.74	50.30	49.29
Al_2O_3	15.90	16.38	15.81	15.12	15.31	14.69
Fe_2O_3	1.33	1.27	1.66	2.31	1.69	1.84
FeO	8.62	7.76	7.62	9.74	8.23	9.11
MgO	9.67	7.80	7.90	8.99	7.79	9.09
CaO	11.16	11.62	11.84	11.61	12.12	12.17
Na_2O	2.43	2.79	2.35	2.04	2.24	1.93
K_2O	0.08	0.09	0.50	0.19	0.20	0.09
TiO_2	1.15	1.31	1.46	1.59	1.21	1.08
P_2O_5	0.09	0.13	0.22	0.18	0.14	0.12
MnO	0.17	0.16	0.16	0.20	0.17	0.19
H_2O	0.30	0.29	0.42	0.42	0.26	0.31
M value	66.5	64.1	64.9	62.2	62.8	63.9
(b)						
La	2.10	2.73	13.39	6.55	5.37	2.91
Sm	2.74	3.23	3.93	3.56	3.02	2.36
Eu	1.06	1.12	1.30	1.29	1.07	0.92
Yb	3.20	3.01	2.37	2.31	2.91	2.33
K	691	822	4443	1179	1559	572
Rb	0.56	0.96	9.57	2.35	3.50	1.02
Cs	0.007	0.012	0.123	0.025	0.042	0.013
Sr	88.7	106.4	243.6	152.5	95.9	86.0
Ba	4.2	10.7	149.6	36.0	39.8	14.3
Sc	40.02	36.47	36.15	39.49	42.59	41.04
V	262	257	250	320	281	309
Cr	528	278	318	330	383	374
Co	49.78	40.97	44.78	57.73	45.70	54.94
Ni	214	132	104	143	94	146
$(La/Sm)_N$	0.50	0.60	2.29	1.28	1.27	0.85
K/Rb	1547	869	475	498	465	560

come mostly from the Atlantic north of 30°N and from the Galapagos spreading centre.

The major element chemistry of suites of volcanic rocks is commonly used to model genetic relationships in terms of fractional crystallization processes. Several authors (Bryan *et al.* 1976, Bender *et al.* 1978, Rhodes & Dungan 1979) have pointed out the importance of cotectic crystallization of olivine and plagioclase at low pressures in controlling the bulk rock chemistry of MORB. This is clearly demonstrated in Figures 5.34 and 5.35, in which Atlantic MORB data exhibit a marked covariance between Al_2O_3 and MgO and TiO_2 and MgO, MgO being used as an index of differentiation. There is obviously a certain amount of scatter in the data attributable to phenocryst accumulation and magma mixing processes, but nevertheless the trends are clear. On the basis of such Harker diagrams the compositional variability of Atlantic MORB can be interpreted in terms of olivine plus plagioclase fractionation from primitive magmas with 10–11% MgO and 16% Al_2O_3. Figure 5.35 shows the dominance of plagioclase over olivine in the fractionating assemblage, and both diagrams demonstrate the non-involvement of clinopyroxene as a major fractionating phase.

There is apparently a greater compositional variability of basalts erupted along the East Pacific Rise relative to those erupted along the Mid-Atlantic Ridge (Natland 1978), suggesting that Pacific MORB are in general more evolved. This has been attributed to the higher spreading rates along the EPR and the consequent existence of larger sub-axial magma reservoirs in which low-pressure fractional crystallization of magmas can occur.

One of the diagnostic characteristics of the tholeiitic magma series as defined in Chapter 1 is the marked trend of iron enrichment in the early stages of fractionation. Figure 5.36 shows the variation of FeO + Fe_2O_3 versus MgO for basalt data from 29 to 73°N along the MAR (Schilling *et al.* 1983). These data clearly show a pronounced trend of progressive iron enrichment with frac-

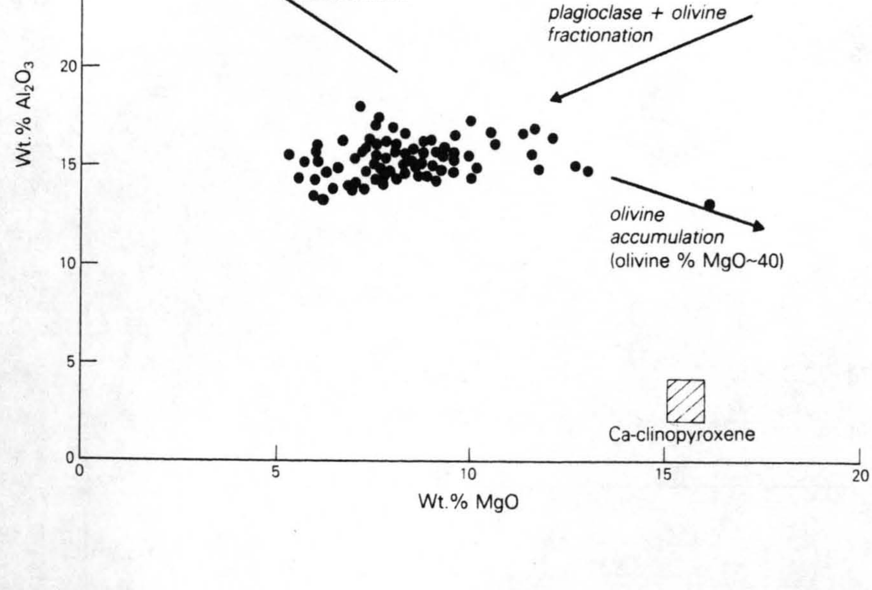


Figure 5.34 The variation of wt.% Al_2O_3 versus wt.% MgO for basalts from 29°N to 73°N along the Mid-Atlantic Ridge. The data can clearly be explained in terms of olivine + plagioclase fractionation. Clinopyroxene does not appear to be involved (data from Schilling *et al.* 1983).

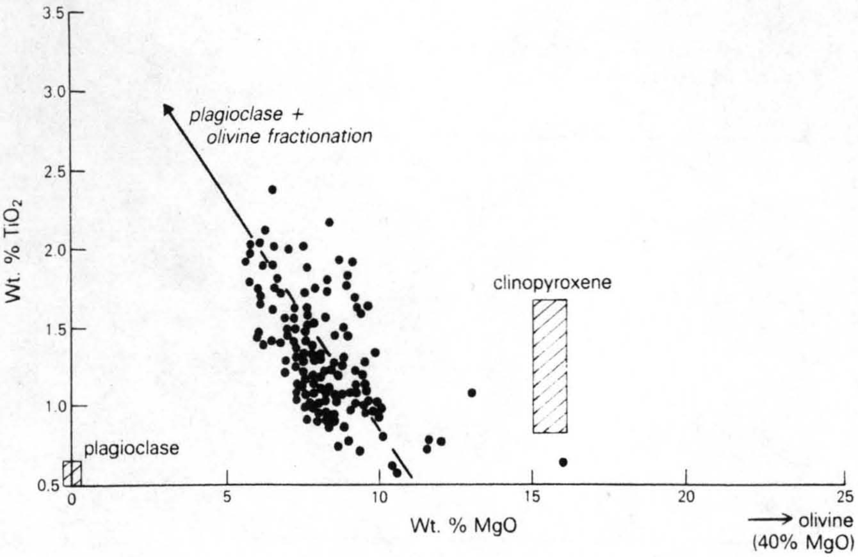


Figure 5.35 Variation of wt.% TiO_2 versus wt.% MgO for basalts from 29°N to 73°N along the Mid-Atlantic Ridge. The data array demonstrates the importance of plagioclase + olivine fractionation. Clinopyroxene is not a major fractionating phase (data from Schilling *et al.* 1983).

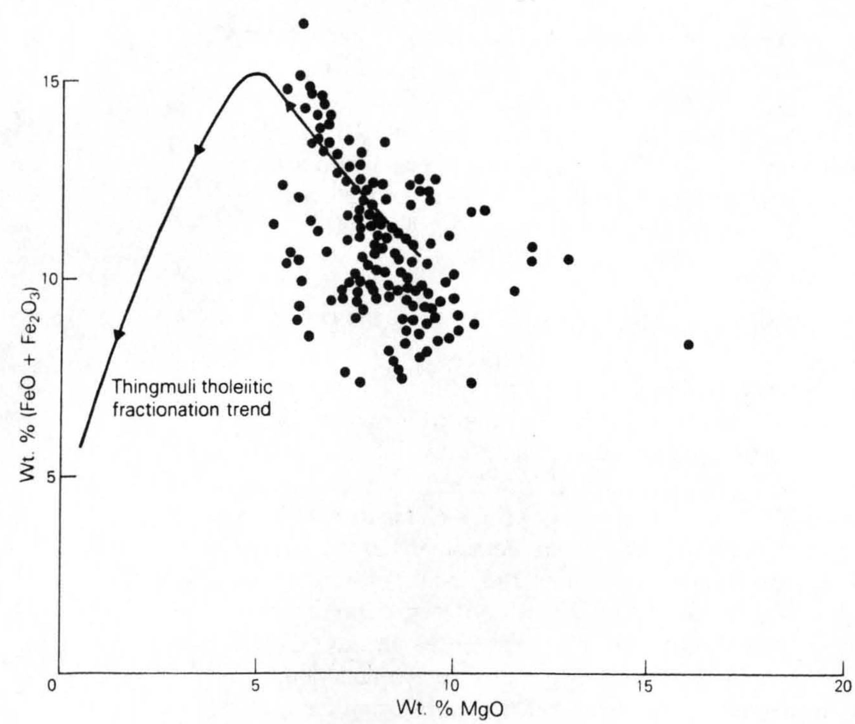


Figure 5.36 The variation of wt.% ($FeO + Fe_2O_3$) versus wt.% MgO for basalts from 29°N to 73°N along the Mid-Atlantic Ridge. Shown for comparison is the characteristic tholeiitic differentiation trend of increasing iron enrichment in the early stages of fractionation displayed by the volcanics of Thingmuli, Iceland (Carmichael 1964). Data from Schilling *et al.* (1983).

tionation, although there is considerable scatter. Shown for comparison is the differentiation trend for the Iceland tholeiitic volcano, Thingmuli (Carmichael 1964), which shows the complete evolutionary spectrum from basalt to rhyolite.

5.10.3 Trace elements

Large low-valency cations

Most MORB are depleted in large low-valency cations (Cs, Rb, K, Ba, Pb and Sr) relative to oceanic island and continental tholeiites (Basaltic Volcanism Study Project 1981). Additionally, for basalts erupted along topographically normal ridge segments, the larger ions are depleted to a greater extent than the smaller ions such that element ratios K/Rb, K/Ba and Sr/Rb are characteristically higher for MORB than for tholeiitic basalts generated in other tectonic environments (Table 5.5). Basalts erupted along elevated ridge sections adjacent to volcanic platforms associated with oceanic islands (e.g. Iceland, Azores and Galapagos) typically contain higher abundances of large cations, with the exception of Sr, and have much lower small/large cation ratios (Table 5.5). In this respect they have much closer affinities with oceanic island tholeiites.

With the exception of Sr, which is partitioned into plagioclase, most of the large low-valency cations are incompatible. Thus their abundance ratios should be essentially independent of the source mineralogy, the degree of partial melting and the extent of high-level fractional crystallization. The ratios of these elements in MORB should therefore reflect the ratios in their mantle source. The marked differences in K/Rb and K/Ba between normal and plume-type MORB shown in Table 5.5 must reflect significant differences in their source compositions, at least with respect to trace elements. However, it must be remembered that this group of elements is particularly susceptible to seawater alteration, and therefore only analyses of fresh glassy basalts should be used for comparative purposes.

Large high-valency cations

The group of large high-valency cations (Th, U, Zr, Hf, Nb and Ta) are called 'immobile elements', and these have been widely used in conjunction with other alteration-resistant elements (Ti, Y, P and Sr) to discriminate amongst basalts from different tectonic settings (Ch. 2). These elements tend to be depleted in N-type MORB relative to P-type and oceanic-island tholeiites (Basaltic Volcanism Study Project 1981). The Zr/Nb ratio serves as a particularly useful discriminant; N–type MORB have high ratios (>30) whereas P-type MORB have low ratios (~10), similar to oceanic-island tholeiites. Zr/Nb ratios have been used to investigate regional heterogeneities in basalts erupted along the strike of the Mid-Atlantic Ridge (Wood *et al.* 1979).

Table 5.5 Typical large cation abundances (ppm) and ratios for oceanic tholeiites (Basaltic Volcanism Study Project 1981, Table 1.2.5.4, p. 144).

	MORB		Ocean-island tholeiite
	Normal	Plume	
K	1064	1854	1600–8300
Rb	1.0	4.5	5–12
Ba	12.2	55	70–200
Sr	127	105	150–400
K/Rb	1046	414	400
K/Ba	109	34	25–40
Sr/Rb	127	23	20–70

Ferromagnesian elements (Cr, V, Sc, Ni and Co)

Crystal–liquid distribution coefficient data indicate that Ni and Co will partition into olivine during partial melting and fractional crystallization processes, while Sc, Cr and V will enter clinopyroxene. Thus the abundances of these elements should be useful indicators of petrogenetic processes. Despite controversy concerning the choice of appropriate olivine/basaltic melt partition coefficients, there is little doubt that Ni abundances in MORB are strongly controlled by olivine fractionation. Ni contents range from >300 ppm in primitive glassy basalts to 25 ppm in highly evolved basalts, and correlate well with MgO content (Fig. 5.37). Cr contents similarly show a marked reduction from 700 to 100 ppm with progressive fractionation. However, this is not considered to reflect significant clinopyroxene fractionation (which has already been negated on major element grounds) but the simultaneous crystallization of olivine- and Cr-rich spinel (Basaltic Volcanism Study Project 1981).

Rare earth elements

Figure 5.38 shows the range of chondrite-normalized REE patterns displayed by MORB.

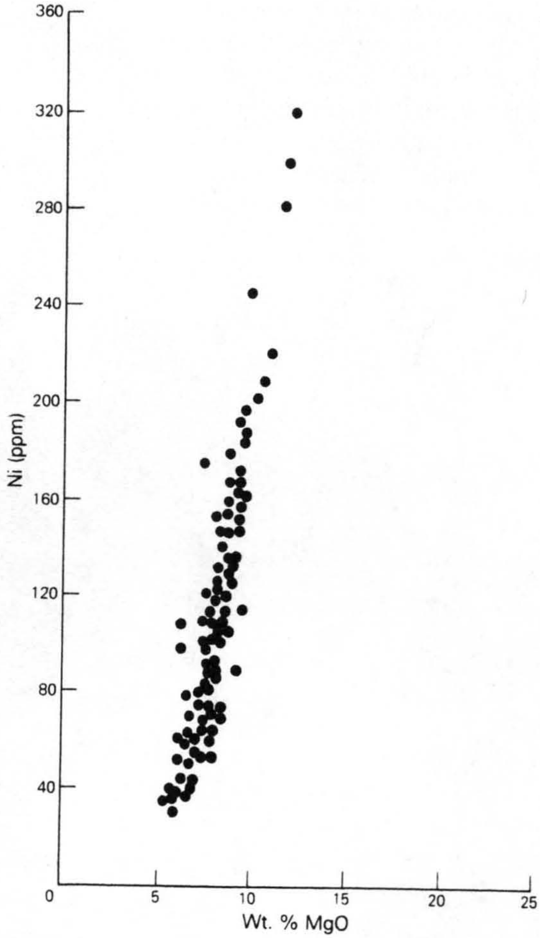


Figure 5.37 The variation of Ni content (ppm) versus wt.% MgO for basalts from 29°N to 73°N along the Mid-Atlantic Ridge. Ni content is clearly controlled by olivine fractionation (data from Schilling *et al.* 1983).

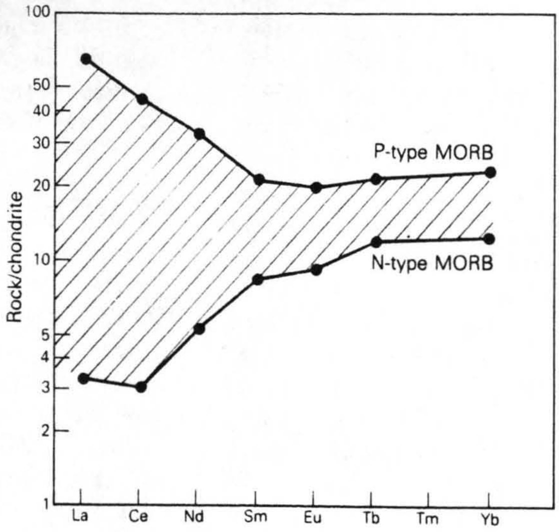


Figure 5.38 The range of chondrite-normalized REE patterns displayed by basalts from normal and plume ridge segments along the Mid-Atlantic Ridge. P-type MORB from 71° 33′N, $(La/Sm)_N = 3.04$; N-type MORB from 66° 51′N, $(La/Sm)_N = 0.4$ (data from Schilling *et al.* 1983).

Typical N-type MORB have unfractionated heavy REE abundances and are strongly depleted in light REE. Primitive basalts have REE concentrations of 10× chondrite or less, whereas extremely differentiated basalts may contain up to 50× chondrite. Fractional crystallization involving olivine, plagioclase, clinopyroxene and spinel increases the total REE content of more evolved MORB, but does not produce any significant inter-element fractionations. Thus the characteristic shape of the primary basalt REE pattern will be maintained in the more evolved basalts. There is, however, a tendency for a negative Eu anomaly to develop as fractionation proceeds, because Eu is preferentially partitioned into plagioclase. In contrast, P-type MORB show relatively little tendency for light-REE depletion and in some instances are light-REE enriched. Generally, N-type MORB have $(La/Sm_N < 1$ whereas P-type have $(La/Sm)_N > 1$.

If partial melting is fairly extensive (>10%) the REE should not be fractionated from each other during partial melting and therefore ratios of REE (e.g. La/Sm, La/Yb and La/Ce) should reflect the ratios in the mantle source of the magmas. However, only the very light REE are truly incompatible and thus, of the above ratios, only La/Ce is likely to be diagnostic of source composition. Figure 5.39 shows the variation of $(La/Sm)_N$ with latitude along the Mid-Atlantic Ridge (Schilling *et al.* 1983). This shows that the division into N- and P-type MORB is clearly an artificial one. Lavas with the highest $(La/Sm)_N$, and hence the greatest degree of light-REE enrichment, are associated with the volcanic platform areas of the Azores, Iceland and Jan Mayen. Figure 5.40 shows that there is a good correlation between $(La/Sm)_N$ and the Zr/Nb ratio for Atlantic, Pacific and Indian ocean MORB, suggesting that binary mixing of end-member source components may be significant in determining their geochemical characteristics. Figure 5.41 shows the variation of La versus Ce for basalts sampled along a traverse along the Mid-Atlantic Ridge from 29 to 73°N. Also plotted for comparison are the data of Humphris & Thompson (1983) for the Walvis Ridge. These data define a remarkably coherent trend which must reflect the La/Ce ratio of the Atlantic MORB source mantle.

Spiderdiagrams

Following Sun (1980), Figure 5.42 compares incompatible element abundances, normalized to primordial mantle values, for N- and P-type MORB and a typical oceanic-island tholeiite. The elements are ordered in a sequence of decreasing incompatibility, from the left to right, in a four-phase lherzolite undergoing partial fusion. Assuming that MORB are indeed produced by comparatively large degrees of partial melting, then their relative incompatible element abundances should be similar to those of their source. Partial melting of a chondritic mantle will produce magmas with spiderdiagram patterns variably enriched in the elements Rb to Nd, depending upon the degree of

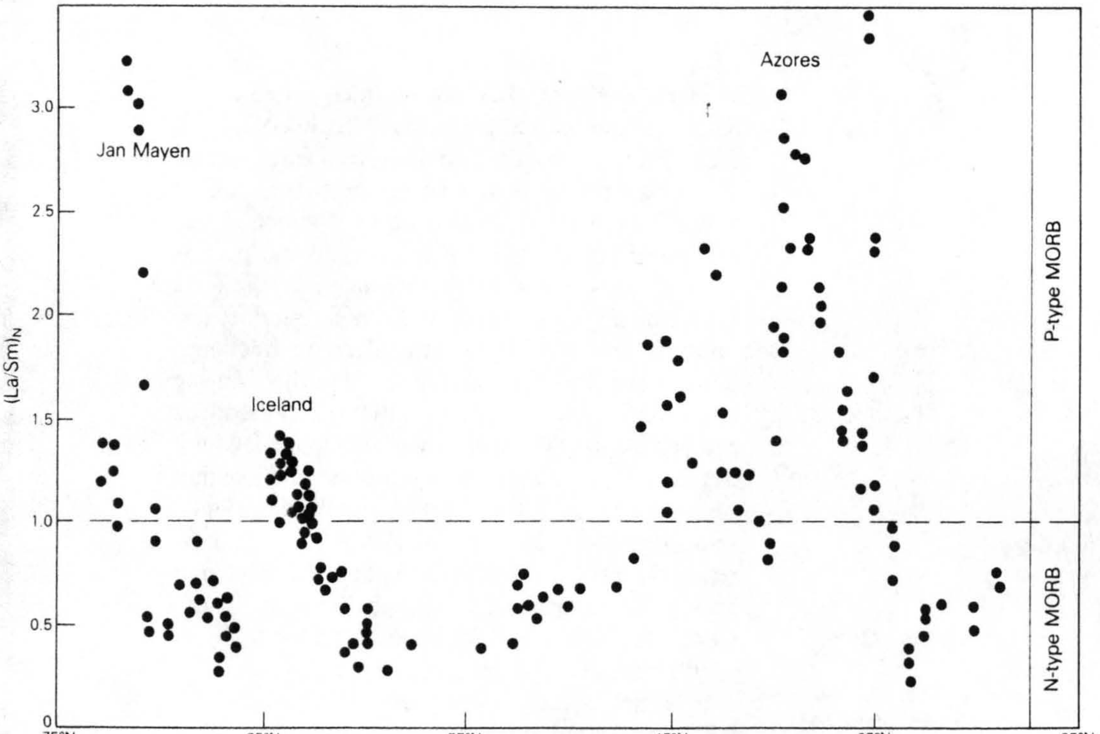


Figure 5.39 Variation in $(La/Sm)_N$ with latitude along the Mid-Atlantic Ridge (data from Schilling *et al.* 1983).

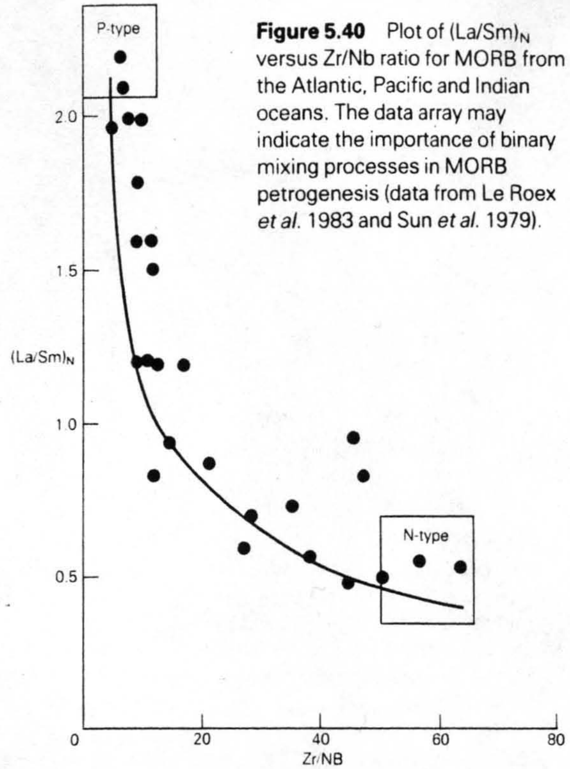


Figure 5.40 Plot of $(La/Sm)_N$ versus Zr/Nb ratio for MORB from the Atlantic, Pacific and Indian oceans. The data array may indicate the importance of binary mixing processes in MORB petrogenesis (data from Le Roex *et al.* 1983 and Sun *et al.* 1979).

melting. Thus the very obvious depletions in the most incompatible elements in N-type MORB must reflect similar depletions in their source mantle. This may be a long term phenomenon related to the continued extraction of continental crustal materials from the upper mantle throughout geological time. In contrast, MORB from plume ridge segments have a distinctive pattern enriched in the most incompatible elements, similar to oceanic-island tholeiites. This would seem to suggest that the sources of P-type MORB and oceanic-island tholeiites have similar geochemical characteristics. The implications of this for mantle dynamics will be considered further in Section 5.11.

5.10.4 Volatile contents

Estimates of the proportions of juvenile volatiles in any basalt are always subject to a certain degree of uncertainty due to the effects of degassing. Moore (1965) showed that vesicle formation and gas release in ocean-floor basalts is directly related to the depth of water in which extrusion takes place. Degassing appears to be inhibited by the hydrostatic pressure of the water at depths greater than

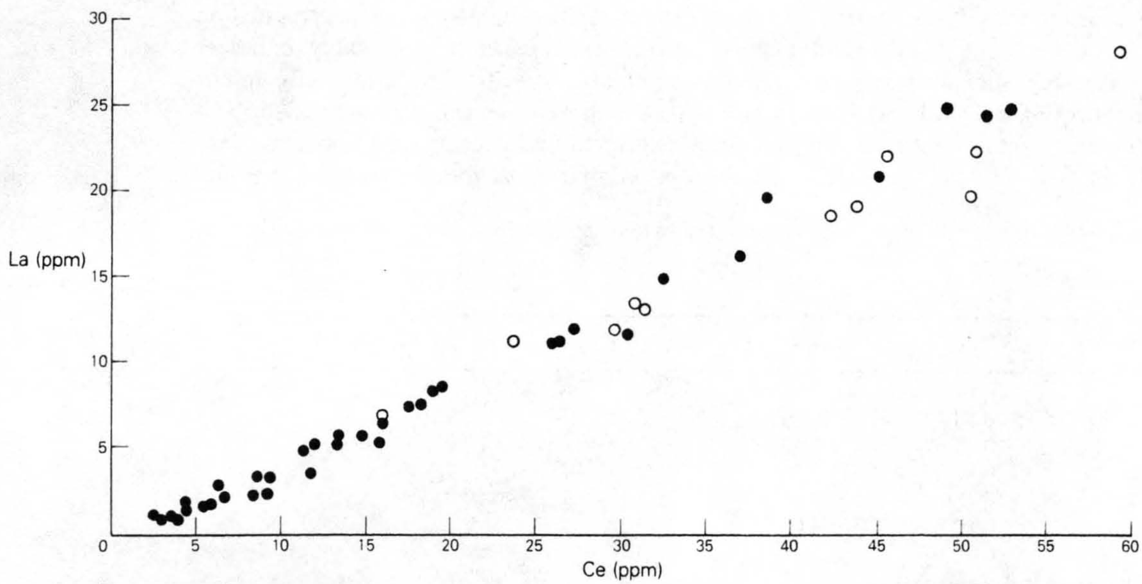


Figure 5.41 Graph of La (ppm) versus Ce (ppm) for basalts from 29°N to 73°N along the Mid-Atlantic Ridge (closed circles) and the Walvis Ridge (open circles) (data from Schilling *et al.* 1983 and Humphris & Thompson 1983).

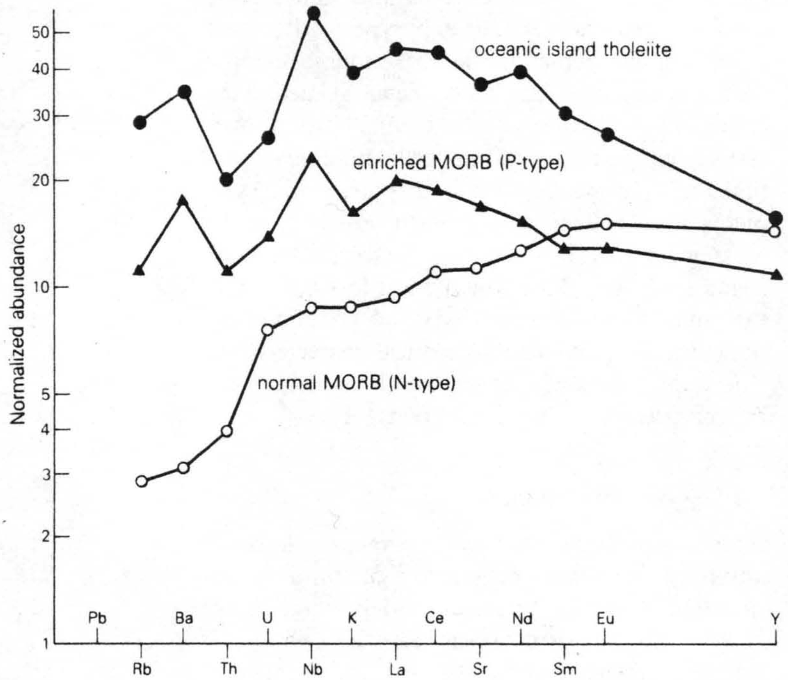


Figure 5.42 Spiderdiagrams showing the difference between normal and P-type MORB, and the similarity between P-type MORB and oceanic-island tholeiites (data from Sun *et al.* 1979).

200 m and therefore most MORB should retain near-primary volatile component characteristics. Concentrations of 0.2–1.0 wt.% H_2O appear to be typical and correlate broadly with K_2O content. Byers *et al.*(1983), in a study of basalts and andesites from the Galapagos spreading centre, have shown that abundances of H_2O, Cl and F in rapidly quenched glasses increase progressively with fractionation (Table 5.6).

5.10.5 Radiogenic isotopes

Sr, Nd and Pb

Geochemical studies of MORB have resulted in their characterization as a remarkably coherent group of oceanic basalts, displaying only minor petrological and geochemical variations along and between different mid-oceanic ridge spreading centres when compared to the spectrum of oceanic-

Table 5.6 Volatile contents (wt.%) in the glassy rims of basalts and andesites from the Galapagos spreading centre (Byers *et al.* 1983).

M-value	Total volatiles	H_2O	CO_2	Cl	F
55	0.31	0.13	0.09	0.04	0.01
52	0.58	0.33	0.10	0.10	0.02
51	0.65	0.20	0.16	0.11	0.02
50	0.52	0.27	0.07	0.13	0.01
35	1.12	0.77	0.07	0.22	0.04
33	1.41	0.87	0.15	0.34	0.02
24	1.50	0.94	0.11	0.34	0.09
16	1.90	1.27	0.10	0.37	0.13

island basalts (Ch. 9). Nevertheless, MORB do show a significant range in $^{87}Sr/^{86}Sr$, $^{208}Pb/^{204}Pb$, $^{207}Pb/^{204}Pb$, $^{206}Pb/^{204}Pb$ and $^{143}Nd/^{144}Nd$ isotopic ratios, indicating that they are in fact derived from a heterogeneous mantle source. Variations in isotopic and trace element geochemistry have been variously attributed to the proximity of nearby hot spots, variations in magma-chamber dynamics or melt generation processes, fracture zone effects and large-scale mantle heterogeneities (Cohen & O'Nions 1982a, Zindler *et al.* 1982, Dupré & Allègre 1983, Allègre *et al.* 1984, Hamelin *et al.* 1984, Schilling 1985).

Some of the spread in the observed $^{87}Sr/^{86}Sr$ ratios may be attributable to seawater-alteration effects, but the Nd and Pb isotopic ratios are unlikely to be modified significantly. In general, N-type MORB have very restricted $^{87}Sr/^{86}Sr$ in the range 0.7024–O.7030, whereas P-type MORB are characterized by slightly more radiogenic compositions (0.7030–0.7035), overlapping with the oceanic-island basalt range (0.7030–0.7050).

Figures 5.43, 5.44 and 5.45 are plots of $^{143}Nd/^{144}Nd$ versus $^{87}Sr/^{86}Sr$, $^{207}Pb/^{204}Pb$ versus $^{206}Pb/^{204}Pb$ and $^{87}Sr/^{86}Sr$ versus $^{206}Pb/^{204}Pb$ for MORB from the Atlantic, Pacific and Indian oceans compared to the spectrum of oceanic-island basalts (OIB). The Nd-Sr diagram (Fig. 5.43) clearly shows the restricted range in isotopic composition of MORB when compared to the entire oceanic basalt data array. Simplistic interpretations of the Nd–Sr 'mantle array' involve binary mixing between depleted and enriched regions of the Earth's mantle (Cohen & O'Nions 1982a), with MORB source mantle representing the depleted end-member. However, interpretation of the combined Pb–Sr–Nd data arrays requires more complex models (Dupré & Allègre 1980, Zindler *et al.* 1982). The MORB-source mantle has clearly been depleted in Nd with respect to Sm and Rb with respect to Sr over a large part of Earth history (>1Ga), and has conventionally been regarded as the geochemical complement of the incompatible enriched upper crust.

Hamelin *et al.* (1984), in a detailed study of the Pb–Sr isotopic variations in Atlantic and Pacific MORB, revealed good positive correlations between $^{207}Pb/^{204}Pb$ versus $^{206}Pb/^{204}Pb$ and $^{87}Sr/^{86}Sr$ versus $^{206}Pb/^{204}Pb$ (Figs 5.44 & 45). The trend on the Pb–Pb plot is subparallel to the oceanic basalt

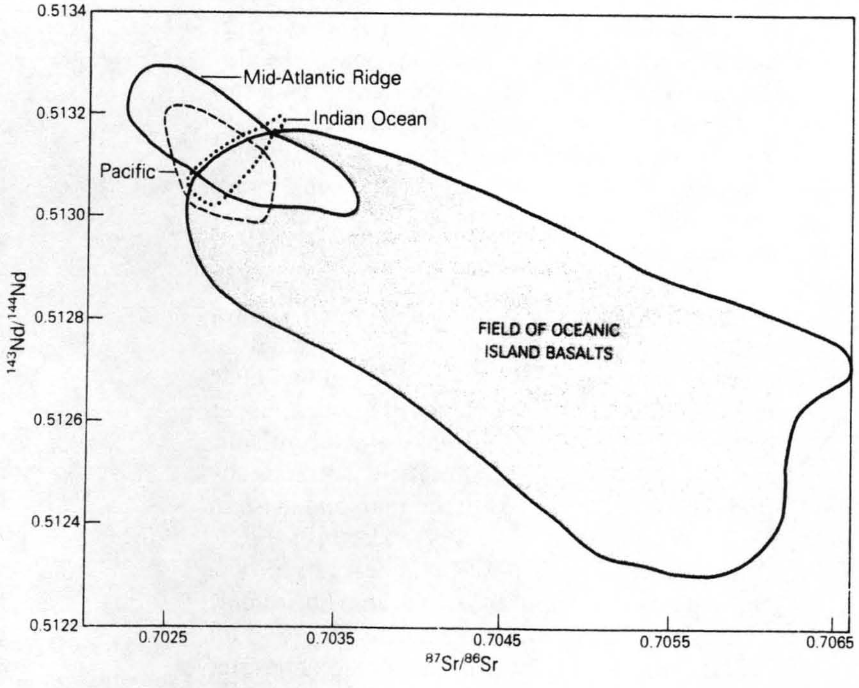


Figure 5.43 $^{143}Nd/^{144}Nd$ versus $^{87}Sr/^{86}Sr$ for MORB from the Atlantic, Pacific and Indian oceans, compared to the spectrum of oceanic-island basalts (after Staudigel *et al.* 1984, Fig. 4).

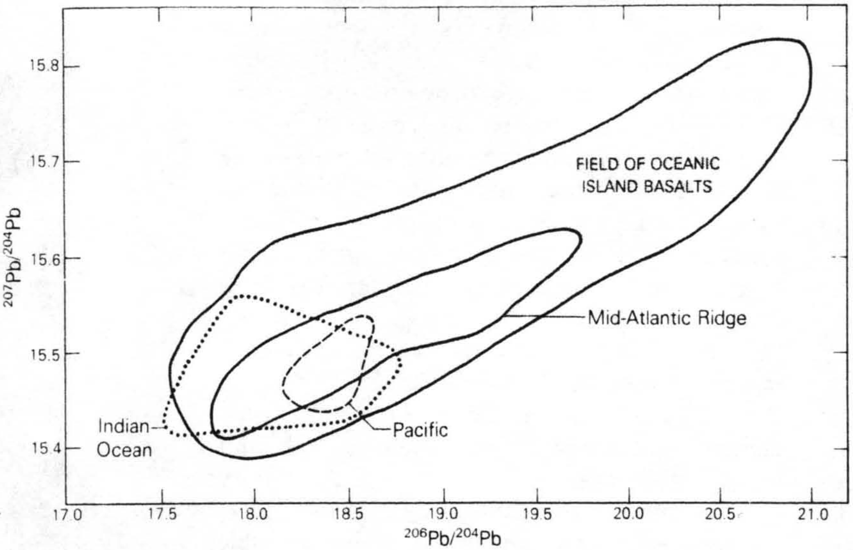


Figure 5.44 $^{207}Pb/^{204}Pb$ versus $^{206}Pb/^{204}Pb$ for MORB from the Atlantic, Pacific and Indian oceans compared to the spectrum of oceanic-island basalts (after Staudigel *et al.* 1984, Fig. 5).

data array, whereas the Sr–Pb MORB trend is mostly at variance with the oceanic-island basalt spectrum. The precise interpretation of these trends is still somewhat equivocal but they must represent the mixing of isotopically distinct components in the source region of MORB. This will be considered further in Ch. 9.

White & Schilling (1978) documented systematic variations in $^{87}Sr/^{86}Sr$ with latitude along the axis of

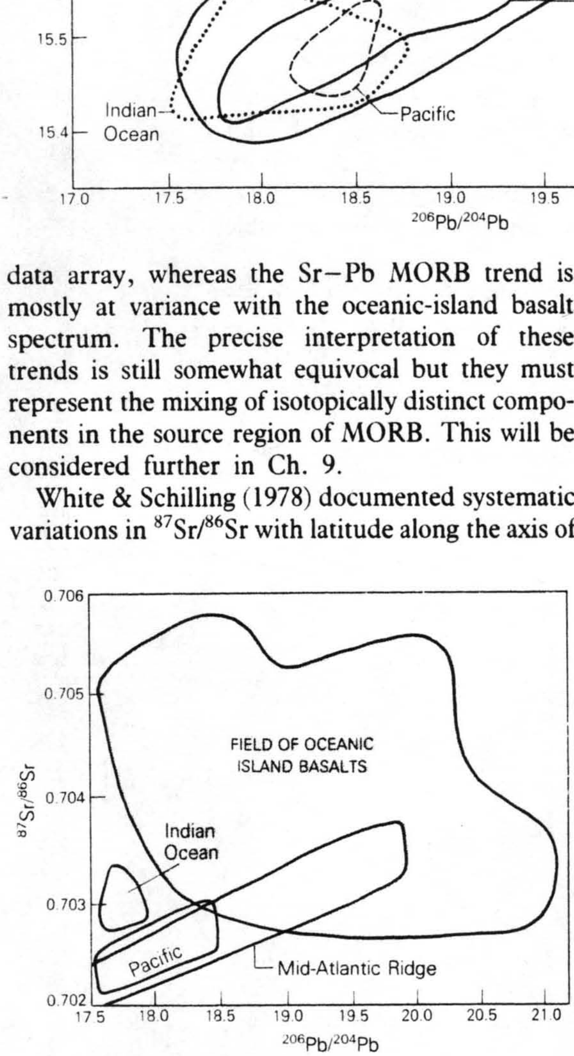


Figure 5.45 $^{87}Sr/^{86}Sr$ versus $^{206}Pb/^{204}Pb$ for MORB from the Atlantic, Pacific and Indian oceans compared to the spectrum of oceanic basalts (after Staudigel *et al.* 1984, Fig. 7; and Dupré & Allègre 1983).

the northern Mid-Atlantic Ridge which correlate well with topographic and other geochemical anomalies (Fig. 5.46). Such along-strike variations in isotopic characteristics can be explained in terms of mixing between an isotopically fairly homogeneous depleted upper mantle reservoir (the source of N-type MORB) and blobs of an isotopically heterogeneous more radiogenic mantle reservoir (the source of OIB and P-type MORB) injected at hotspot locations along the ridge axis. Detailed discussion of the geochemical characteristics of this latter reservoir will be deferred until Chapter 9. The isotopic heterogeneity of a particular ridge segment will depend upon internal blob heterogeneity, the proportions of blobs and depleted asthenosphere in the mixture and the efficiency of the mixing process. The 'blob' component appears to show distinct regional characteristics on a global scale, as evidenced by comparative studies of oceanic-island basalts which are inferred to sample blob material less diluted by MORB-source upper mantle. For example, OIB from the South Atlantic and Indian oceans are quite different isotopically from those from the North Atlantic and Eastern Pacific (Dupré & Allègre 1983, Hart 1984).

In general, Pacific MORB appear to show a much narrower range in Sr, Nd and Pb isotopic compositions than Atlantic MORB (White *et al.* 1987). This might be attributable to a greater

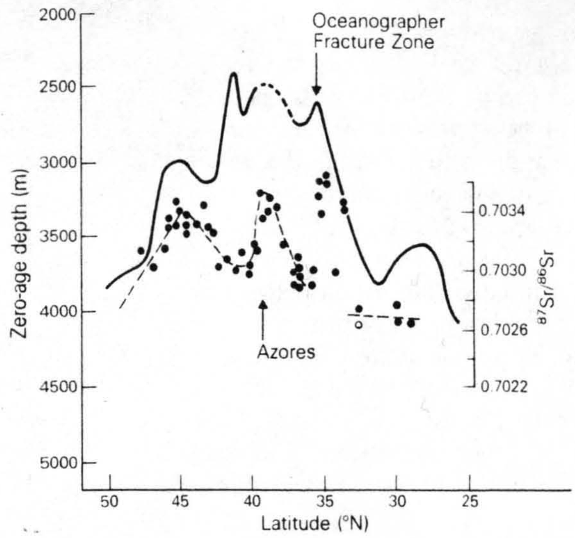


Figure 5.46 Superposition of the zero-age depth curve and variation in $^{87}Sr/^{86}Sr$ with latitude along the northern Mid-Atlantic Ridge (after Le Douaran & Francheteau 1981; $^{87}Sr/^{86}Sr$ data from White & Schilling 1978).

isotopic heterogeneity in the Atlantic MORB source compared to the Pacific. However, it seems more reasonable to postulate similar degrees of isotopic heterogeneity in all MORB source regions, with the greater degrees of melt production associated with the fast-spreading East Pacific Rise, effectively homogenizing the isotopic compositions.

5.11 Detailed petrogenetic model

The most primitive basalt compositions recorded amongst the MORB spectrum have 10% MgO, M values of 70, Ni contents of 300 ppm and highly magnesian olivine phenocrysts (Fo_{90-91}). Magmas with such chemical characteristics could be primary mantle partial melts (Bender *et al.* 1978) and thus there is no fundamental necessity to propose more picritic primary magma compositions. However, O'Hara (1968, 1973, 1982) has consistently argued for picritic primary magmas, which undergo extensive olivine fractionation en route to the surface, in his models of MORB petrogenesis. The high-MgO primary magma theory is supported by lherzolite melting experiments and by petrological studies of ophiolite complexes. However, it is inconsistent with the observation that no glasses or aphyric rocks with MgO contents greater than 11% have ever been sampled from the ocean floor. Huppert & Sparks (1980) support a picritic primary magma hypothesis and explain the lack of high-MgO erupted basalts as a consequence of the fluid dynamics of such high-density liquids.

The vast majority of MORB samples are, however, more fractionated (evolved), attesting to the importance of olivine + plagioclase ± clinopyroxene fractionation in producing their observed geochemical characteristics. O'Hara (1977) has stressed the importance of magma mixing and fractional crystallization in high-level open-system magma chambers in the production of magma batches whose composition corresponds to some 'perched state'. His model explains the rarity of both primitive basalts and highly evolved differentiates in the MORB spectrum. Mixing and fractionation processes will obviously be enhanced in large sub-axial magma reservoirs beneath fast-spreading ridge segments, perhaps explaining why Pacific MORB appear more evolved than Atlantic MORB.

Two extreme types of basalt are erupted along mid-oceanic ridges:

(1) *Normal (N–type)*. LREE and incompatible element depleted. High K/Ba, K/Rb, Zr/Nb and low $^{87}Sr/^{86}Sr$.
(2) *Plume (P-type)*. Less depleted than N-type in LREE and incompatible elements with higher $^{87}Sr/^{86}Sr$ ratios. K/Ba, K/Rb, La/Ce and Zr/Nb ratios are lower than those of N-type MORB and comparable to those of oceanic-island tholeiites.

A continuous spectrum of intermediate types exists between these two end-members. N-type MORB appear to be derived from a depleted asthenospheric upper-mantle source, whereas P-type MORB are derived from a more enriched plume or hotspot component. Source heterogeneity is thus an important parameter in MORB petrogenesis combined with fractional crystallization, magma mixing, variable degrees of partial melting and variable residual source mineralogies.

Mg-rich MORB are multiply saturated with olivine + clinopyroxene + orthopyroxene at pressures in excess of 8–10 kbar (Bender *et al.* 1978, Green *et al.* 1979, Stolper 1980), corresponding to minimum depths of segregation from the mantle of 25–30 km. Thus MORB parent magmas should have last equilibrated within the spinel lherzolite stability field, which is consistent with their observed trace element geochemistry. The primary magmas clearly evolve by polybaric partial melting processes in ascending mantle diapirs, commencing at considerably greater depths, perhaps as much as 60 km. However, it is the last point of equilibration within the mantle before segregation which constrains the geochemical characteristics of these magmas.

Sr, Nd and Pb isotopic studies of MORB (Section 5.10.5) have revealed important source heterogeneities which can be explained in terms of present-day mixing processes beneath the ridge axis between depleted mantle material from the asthenosphere and blobs of hotspot or plume material coming from deeper levels (Allègre *et al.* 1984, Dupré & Allègre 1983). Such mixing processes can explain the observed negative correlation between Nd and Sr isotopic ratios, and the positive correlation between Sr and Pb isotopic ratios for Atlantic MORB. Figure 5.47 is a cartoon depicting such a mixing model. N-type MORB are derived by partial melting of the isotopically fairly homogeneous, well mixed, depleted upper-mantle

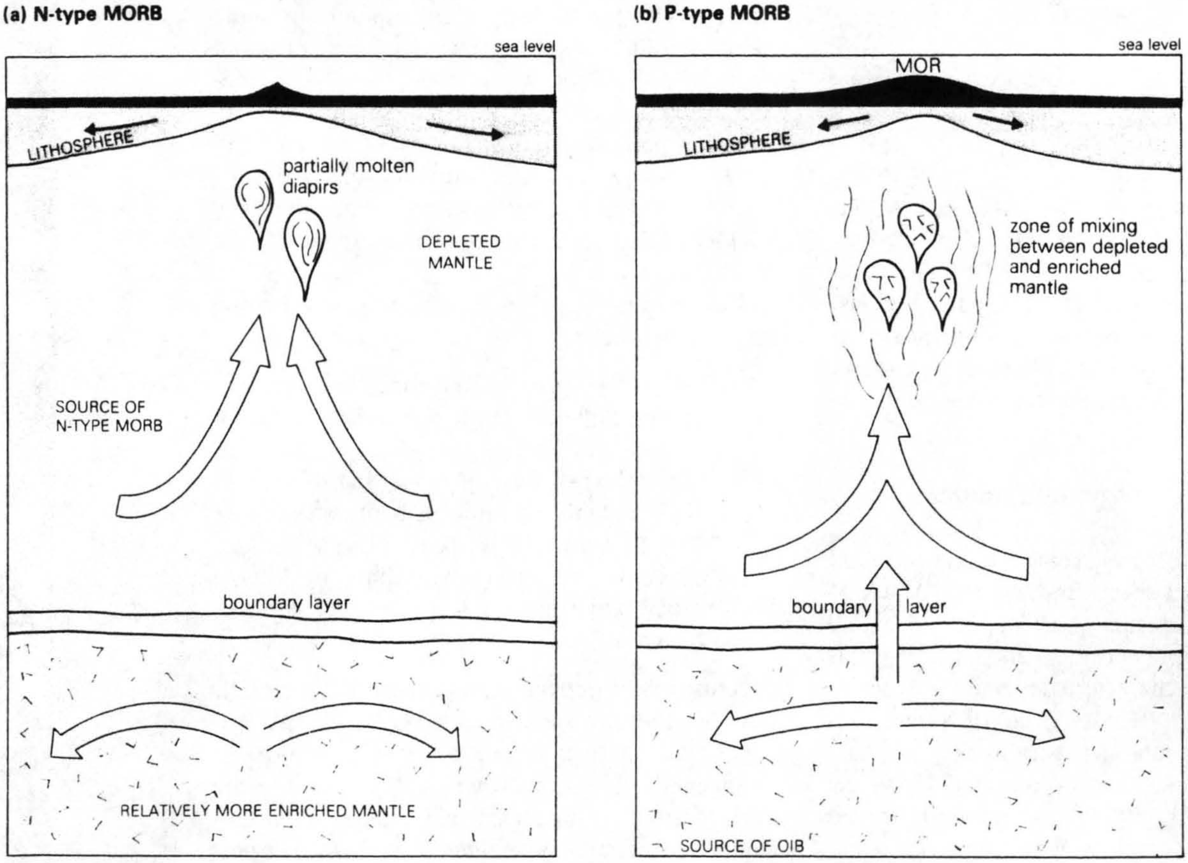


Figure 5.47 Models for the involvement of different source components in the origin of N- and P-type MORB. (a) N-type MORB are derived by partial melting of an isotopically fairly homogeneous, well mixed, depleted upper mantle reservoir. (b) P-type MORB are derived from sources containing variable amounts of a 'blob' component, derived from a lower isotopically heterogeneous reservoir mixed with the depleted N-type MORB source; this lower reservoir is also the source of OIB. (After Zindler *et al.* 1984, Fig. 12).

reservoir, whereas P-type MORB contain variable amounts of a blob component from the underlying isotopically heterogeneous reservoir, which is also the source of oceanic-island basalts (OIB). Parts of this lower reservoir may be almost primordial in composition, whereas other parts may represent recycled subducted slab components (Ch. 9).

The idea of hotspot injection of material beneath mid-ocean ridge axes originated in studies of the northern Mid-Atlantic Ridge by Schilling (1973). In the original model of J. T. Wilson (1973) hotspots were considered to be fixed, more or less continous, flows ascending from the lower mantle. However, the model shown in Figure 5.47 is more realistic, involving the ascent of discontinuous blobs from the lower-mantle source. On slow-spreading ridges (e.g. the Mid-Atlantic Ridge) the hotspot signature is clearly visible in both bathymetry and geochemical characteristics, whereas on fast-spreading ridges (e.g. the East Pacific Rise) it may be diluted by the rapid supply of asthenospheric material. In the model, blobs reach high levels more or less efficiently mixed with asthenospheric material. Such blobs may be expected to be larger beneath slow-spreading ridges, and smaller and more numerous beneath fast-spreading ridges (Allègre *et al.* 1984; see also Figure 5.48).

Many studies have shown a correlation between mid-ocean ridge bathymetry and MORB isotopic and trace element geochemistry (Hart *et al.* 1973, White & Schilling 1978, Dupré & Allègre 1980, le Douaran & Francheteau 1981, Hamelin *et al.* 1984, Humphris *et al.* 1985, Hanan *et al.* 1986). Ridge-

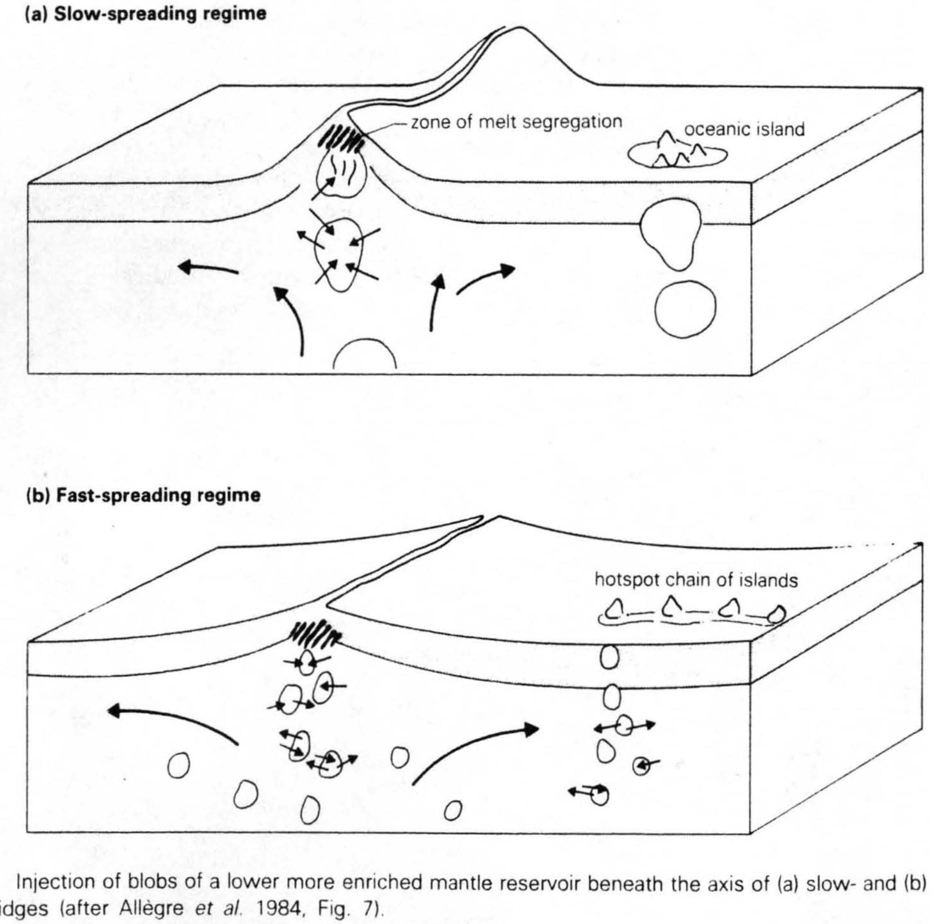


Figure 5.48 Injection of blobs of a lower more enriched mantle reservoir beneath the axis of (a) slow- and (b) fast-spreading mid-ocean ridges (after Allègre *et al.* 1984, Fig. 7).

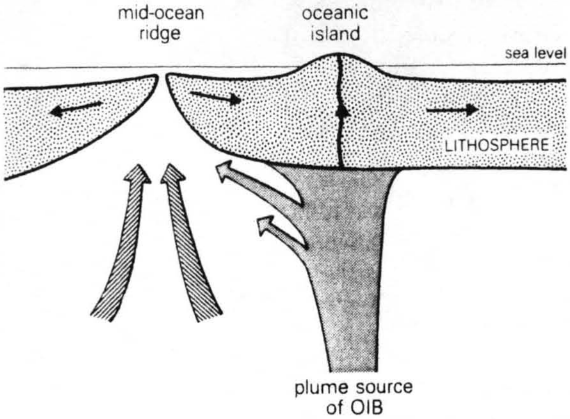


Figure 5.49 Migration of a mid-ocean ridge axis away from a hotspot, inducing a non-radial flow in the rising plume towards the ridge (after Schilling *et al.* 1985, Fig. 4).

centred hotspots appear to produce large-scale isotopic and trace elements gradients, unusually intense constructional volcanism and elevation anomalies (Schilling *et al.* 1985). If the ridge axis subsequently drifts away from the hot spot then the rising plume tends to develop a preferential non-radial flow towards the migrating ridge axis (Fig. 5.49). In such a case lava compositions are still gradational along the ridge axis, symmetrical about the hotspot and the connecting channel, but the gradients are not as pronounced and the elevation anomaly is more subdued. Eventually, with continued migration of the ridge axis, the plume supply will be cut off and will cease to influence axial magmatic processes.

Figure 5.50 is a schematic illustration of the hypabyssal environment existing at the axis of a mid-ocean ridge spreading sufficently rapidly to maintain a dynamic sub-axial magma chamber of moderate dimensions. Here magmas may pond, fractionate, mix and precipitate mafic and ultramafic cumulates before erupting. The primary magma entering this plumbing system is either a picrite or a highly magnesian basalt. Initally, this may remain as a high-density layer at the base of the chamber underlying more fractionated (less dense) chamber magma until fractionation proceeds to the point of chamber rollover, and homogenization of the two layers occurs. If the chamber is horizontally stratified in this way then the roughness of the fault-block-controlled chamber roof can lead to the sampling of more primitive magmas at the centre of the rift and more differentiated magmas at the margins. This has been observed in the FAMOUS area of the Mid-Atlantic Ridge. Lavas emerging from such a periodically replenished, periodically

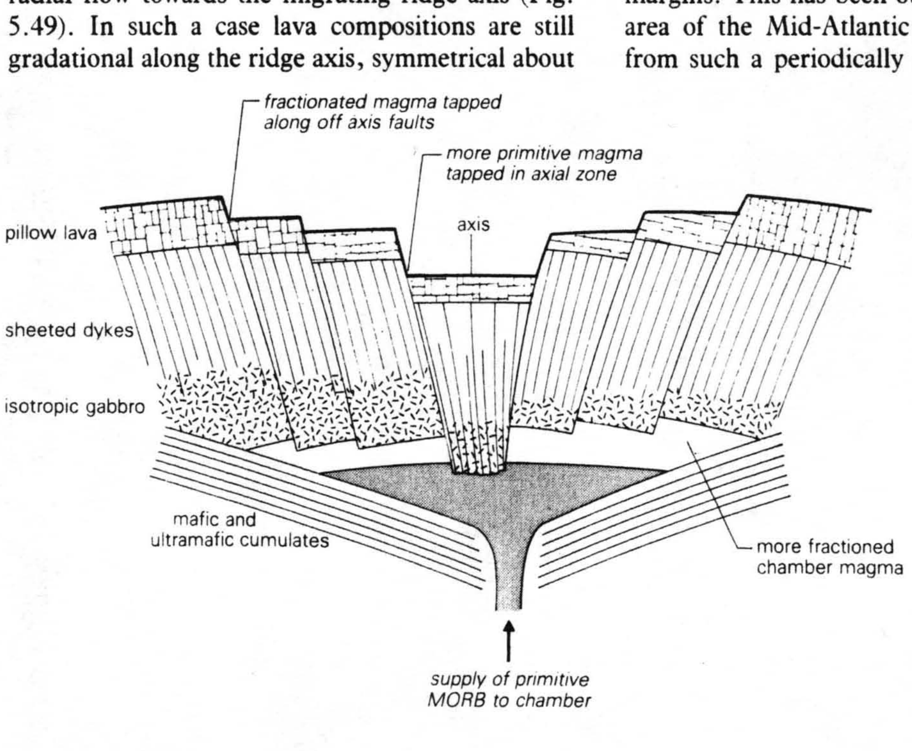


Figure 5.50 Hypabyssal environment at the axis of a mid-ocean ridge that is spreading sufficiently rapidly to maintain a dynamic sub-axial magma chamber.

tapped, continuously fractionating magma chamber may be far evolved from their parent magmas, displaying relatively constant major element compositions, but large variations in concentrations and concentration ratios of incompatible trace elements (O'Hara 1982). The chemical characteristics of such magmas obviously cannot be uniquely inverted to elucidate the nature of the parent magma and the geochemical characteristics of the source mantle. Thus, although MORB have conventionally been regarded as a 'window to the upper mantle', that window is by no means transparent.

Further reading

Basaltic Volcanism Study Project 1981. *Basaltic volcanism on the terrestrial planets*. New York: Pergamon Press.

Gass, I.G., S.J. Lippard & A.W. Shelton (eds) 1984. *Ophiolites and oceanic lithosphere*. Oxford: Blackwell Scientific.

Hekinian, R. 1982. *Petrology of the ocean floor*. Amsterdam: Elsevier.

Kennet, J.P. 1982. *Marine geology*. Englewood Cliffs, N.J: Prentice-Hall.

Le Roex, A.P. 1987. Source regions of mid-ocean ridge basalts: evidence for enrichment processes. In *Mantle metasomatism*, M.A. Menzies & C.J. Hawkesworth (eds). London: Academic Press.

Wilkinson, J.F.G. 1982. The genesis of mid-ocean ridge basalt. *Earth Sci. Rev.* **18**, 1–57.