Geoscience Canada

Oceanic Lithosphere 4. The origin and evolution of oceanic lithosphere: Magmatic processes at oceanic spreading centres

John Malpas and Paul T. Robinson

Volume 27, Number 3, September 2000

URI: https://id.erudit.org/iderudit/geocan27_3ser01

See table of contents

Publisher(s)

The Geological Association of Canada

ISSN

0315-0941 (print)
1911-4850 (digital)

Explore this journal

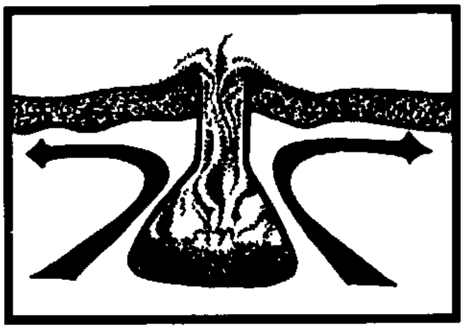
Cite this article

Malpas, J. & Robinson, P. T. (2000). Oceanic Lithosphere 4. The origin and evolution of oceanic lithosphere: Magmatic processes at oceanic spreading centres. *Geoscience Canada*, *27*(3), 131–146.

Article abstract

Primary melts generated in the oceanic mantle migrate upward and pond at major discontinuities to form magma chambers. Such chambers have thus far been identified only on fast- and intermediate-spreading ridges but probably also exist at slow-spreading ridges. The size, shape and longevity of subrift chambers reflect the magma supply rate, the extent of hydrothermal cooling, and the regional stress field. Ophiolite studies suggest small, ephemeral chambers rather than large, long-lived bodies. At fast-spreading ridges the chambers probably consist largely of crystalline mush, possibly with some melt sills, and a thin melt zone at the top. At slower-spreading ridges, magmatic activity is more episodic and seafloor spreading may be punctuated by periods of mainly tectonic extension. Fractionating melts inthe chambers are buffered by injections of more primitive melt from depth to produce the relatively uniform composition of MORB. The gross structural uniformity of the ocean crust must reflect extensive interplay and feedback of magmatic, hydrothermal and tectonic processes, resulting in a self-ordered system.

SERIES



Oceanic Lithosphere 4. The origin and evolution of oceanic lithosphere: Magmatic processes at oceanic spreading centres

John Malpas
Department of Earth Sciences
The University of Hong Kong
Pokfulam Road
Hong Kong, China

Paul T. Robinson[1]
Institute of Geochemistry
GeoForschungZentrum,
Telegrafenberg, D-14473
Potsdam, Germany

SUMMARY
Primary melts generated in the oceanic mantle migrate upward and pond at major discontinuities to form magma chambers. Such chambers have thus far been identified only on fast- and intermediate-spreading ridges but probably also exist at slow-spreading ridges. The size, shape and longevity of subrift chambers reflect the magma supply rate, the extent of hydrothermal cooling, and the regional stress field. Ophiolite studies suggest small, ephemeral chambers rather than large, long-lived bodies. At fast-spreading ridges the chambers probably consist largely of crystalline mush, possibly with some melt sills, and a thin melt zone at the top. At slower-spreading ridges, magmatic activity is more episodic and seafloor spreading may be punctuated by periods of mainly tectonic extension. Fractionating melts in the chambers are buffered by injections of more primitive melt from depth to produce the relatively uniform composition of MORB. The gross structural uniformity of the ocean crust must reflect extensive interplay and feedback of magmatic, hydrothermal and tectonic processes, resulting in a self-ordered system.

RÉSUMÉ
Les fluides créés lors des fusions initiales dans le manteau océanique migrent vers l'extérieur et s'accumulent à l'emplacement d'importantes discontinuités pour constituer des chambres magmatiques. Bien que de telles chambres n'est été observées qu'aux lieux de crêtes d'expansion rapide et intermédiaire, il est probable qu'elles existent aussi à l'endroit de crêtes à expansion lente. Le volume, la forme et la longévité de telles chambres dépendent de leur taux de d'alimentation en magma, de la vigueur du refroidissement hydrothermal, ainsi que des paramètres du champ de contraintes local. Les études sur des ophiolites nous portent à croire qu'il s'agirait de chambres de petites dimensions et de courte longévité. À l'endroit de crêtes à expansion rapide, il s'agit probablement de chambres constituées de mélanges de cristaux et, peut-être de sills en fusion, sur lesquelles surnage une couche fondue mince. À l'endroit de crête d'expansion plus lente, l'activité magmatique est plus discontinue et, les épisodes d'expansion du plancher océanique peuvent n'être que tectoniques. Le fractionnement de la phase liquide dans ces chambres magmatique se produit au gré des injections de matériaux primitifs issus des profondeurs, ce qui explique la composition assez uniforme des BCMO. L'uniformité structurale de la croûte océanique est sans doute le reflet de l'interaction généralisée entre des processus magmatiques, hydrothermaux et tectoniques et, qui constitue un système auto-régulé.

INTRODUCTION
The mantle constitutes 67.2% of the mass and 90% of the volume of Earth. The crust, hydrosphere and atmosphere, so geologically significant, comprise a mere 5% by weight of the whole Earth, and have essentially been derived by processes operating in the mantle throughout Earth history. It is not surprising therefore, that petrologists have devoted enormous effort to deducing the composition of the mantle and the nature of the physical and chemical processes that have occurred within it.

A variety of approaches may be used to decipher the composition of Earth's upper mantle. Given that basaltic melts are derived by partial melting at mantle depths, examination of their petrogenesis has been one of the most important and rewarding of these. The concept of a peridotitic mantle as a source of basaltic magmas was first suggested by Bowen (1928), and elaborated in detail by Ringwood (1962), who stated that the mantle must be "defined by the property that on fractional melting it would yield a typical basaltic magma and leave behind a residual refractory dunite-peridotite of alpine-type." This hypothetical mantle

[1]Permanent Address: Department of Earth Sciences, Dalhousie University, Halifax, Nova Scotia B3H 3J5

composition, called "pyrolite" or pyroxene-olivine rock, was subsequently modelled by Green and Ringwood (1967, and references therein), using various proportions of anhydrous dunite (the residuum) and an average transitional basalt composition (the melt). Although this approach provides a useful indication of the composition of undepleted upper mantle, its validity depends upon identification of basaltic compositions in equilibrium with the residuum at the time of melting. However, a simple calculation of the Mg/Fe ratios between residuum and melt used in pyrolite models shows that the two cannot be in equilibrium, implying that the melt was modified during its ascent to the surface.

The key then is to identify a basaltic composition that represents a primary melt from the upper mantle, unmodified since its generation. The uniform, highly depleted composition of seafloor basalts described by Engel and Engel (1963, 1964) led to the suggestion that they might represent such primary magmas. However, it is now clear that oceanic basalts have diverse compositions, reflecting variations in the nature of the source material and the degree to which they have been modified during ascent to the seafloor.

In the absence of clearly identifiable primary magmas, we must turn to ophiolites in which upper mantle rocks are abundantly exposed, and to sparse occurrences on the seafloor. Although most ophiolites are formed in suprasubduction zone environments rather than at mid-ocean ridges, it is likely that the dynamic processes of melt generation, melt migration, and crustal accretion are similar regardless of the tectonic environment in which the spreading axis is located. Detailed studies of large, well-exposed ophiolites such as Troodos, Semail and the Bay of Islands have shed much light on the structure and composition of the lower crust and upper mantle, as well as on the magmatic processes occurring in spreading environments. Likewise, peridotites dredged and drilled from the seafloor have provided insights into melting processes and melt migration in the upper mantle (*e.g.*, Dick, 1989; Johnson and Dick, 1992; Dick and Natland, 1996; Niu, 1997; Niu and Hekinian, 1997).

MANTLE MELTING BENEATH OCEAN RIDGES

In its simplest form, the generation of mid-ocean ridge basalt (MORB) is considered to occur through partial melting of depleted mantle material as it rises and decompresses adiabatically beneath spreading ridges (Allegre and Bottinga, 1974; Langmuir *et al.*, 1992). However, as mentioned above, MORB lavas erupted on the seafloor are no longer in equilibrium with residual mantle harzburgite and lherzolite. Whereas orthopyroxene is found ubiquitously in ophiolitic and abyssal peridotites, it is not a liquidus phase in MORB magmas at pressures below 8 kb, or some 25 km depth (O'Hara, 1965; Elthon and Scarfe, 1980). However, MORB lavas may retain some evidence of their high-pressure origin, *e.g.*, rare earth element patterns indicative of melting at depth (Shen and Forsyth, 1995) and Lu/Hf ratios (Salters and Hart, 1989). The preservation of such high-pressure features suggests that the melts must have separated from the solid residuum at depth and then been transported rapidly to the surface. The high degree of fractionation between light rare earth elements (LREE) and heavy rare earth elements (HREE) and between other trace elements in residual peridotites further requires that such magmas form by small degrees of partial melting and that they be extracted in small batches. Likewise, the extensive heterogeneity discovered in melt inclusions in olivine and plagioclase within MORB provides further evidence that these melts are produced in small batches at various depths (Sobolev and Shimizu, 1993;1994; Nielsen *et al.*, 1995; Shimizu, 1998), and that MORB lavas are mixtures of these melts. It is unlikely that these features would be preserved if the melt extraction process was entirely one of diffuse flow along grain boundaries, allowing for low-pressure chemical re-equilibration with the surrounding peridotite.

The melting regime beneath mid-ocean ridges differs according to the geothermal gradient, *i.e.*, lateral variations in mantle temperature lead to differences in the extent and depth of melting (Fig. 1). Hot mantle intersects the solidus at deeper levels, melts at greater average pressures, and melts more upon ascent than cooler mantle. Thus, melting of hot mantle produces relatively thick oceanic crust and, by isostatic adjustment, relatively shallow axial depths (Langmuir *et al.*, 1992). Such processes account, in part, for the variations in bathymetry shown by along-axis depth profiles and the correlation between basalt chemistry and axial depth (Klein and Langmuir, 1987) (Fig. 2). The extent of melting, which is reflected by the amount of modal olivine in a residual peridotite, can be correlated directly to the mineral chemistry of residual phases (Dick *et al.*, 1984 and Fig. 3). However, Niu (1997) and Niu and Hekinian (1997) have shown that there is excess olivine in some residual peridotites, which they attribute to precipitation during melt extraction. Crystallization within the mantle is also suggested by the relatively undepleted nature of some peridotites and their impregnation by plagioclase at slow-spreading ridges (Dick, 1989).

The compositions of primary melts reflect not only differing degrees of partial melting but also the nature of the melting process. Most melting models have emphasized either fractional or equilibrium melting. During fractional melting the liquids that are produced are separated continuously from the enclosing host and rise rapidly through the overlying mantle. Equilibrium or "batch" melting occurs when the melt remains in equilibrium with the residual solid until the batch is removed.

Much of the geochemical data obtained from abyssal peridotites has been used to support models of fractional melting (Fig. 4). MORB lavas are considered to be the result of accumulation and mixing of such melts produced at different depths. Primitive MORB compositions would reflect almost no mixing, and normal, or N-MORB compositions, would result from extensive mixing of primitive and more evolved melts. Johnson and Dick (1992), however, argue that simple fractional melting cannot account for the chemistry of clinopyroxenes in residual peridotites, which have high La/Sm and Zr/Ti. Thus, various models of "near-fractional" melting have been proposed (Maaloe and Scheie, 1982; Johnson *et al.*, 1990; Johnson and Dick, 1992; Kelemen *et al.*, 1997), the most prevalent of which includes some form of "equilibrium porous flow" (Elthon, 1992; Lundstrom *et al.,* 1995).

As mentioned above, equilibrium or batch melting generally refers to the production of a melt, which remains in equilibrium with the residual solid until the batch is removed. However, silicate melts are more buoyant than their surrounding matrix and, given the appropriate permeability, will rise through the solid residuum. In this case, if re-equilibrium is established continuously between the melt and surrounding bulk solid, the process is known as equilibrium porous flow. Lundstrom *et al.* (1995) have used the disequilibria in Uranium series isotopes to show not only that MORB magmas are derived from heterogeneous mantle sources but that the melting processes within enriched and depleted sources are significantly different. They suggest that equilibrium porous flow can better explain magma production at depths where garnet is present than can fractional melting. A similar argument can be made for shallower level melting, where garnet is absent, using Ra/Th disequilibrium data (Iwamori, 1994 and Fig. 5).

These different models for melt production have a direct influence on concepts of melt migration beneath mid-ocean ridges. Because the asthenosphere essentially lies immediately beneath the crust at a ridge axis, it is easy to envisage melts migrating directly to a crustal magma chamber. However, seismic studies, for example the Mantle Electromagnetic and Tomography (MELT) experiment (Forsyth, 1997; Toomey *et al.,* 1996; 1998; Scheirer *et al.,* 1998) show that the volume of mantle in which melting occurs is possibly several hundred kilometres wide and as much as 100 km thick and yet the neo-volcanic zone on the seafloor is at most a few kilometres wide. On the fast-spreading East Pacific Rise, the broad melt zone is asymmetrical and is centred west of the ridge axis. The problem is how to concentrate the upwelling magma into narrowly focussed magma chambers. Two end member scenarios have been proposed to explain this focussing of magmatic accretion to within a few kilometres of the ridge axis: focussed or channelized melt flow (Kelemen *et al.,* 1997) and diapiric uprise of melt-bearing mantle. Focussed or channelized flow may be along hydrofractures (Maaloe, 1981; Nicolas, 1986), or a result of restricted local melt reaction with wallrock. Nicolas (1986, 1990) has argued that channels produced by hydrofracturing would occur as sub-vertical dykes at depths less than 15 km. According to Nicolas, the orientation and location of these dykes would be controlled by the pressure drop associated with upwelling mantle undergoing corner flow. However, Phipps Morgan (1987) and Turcotte and Phipps Morgan (1992) argue that mantle viscosity is too low to influence magma migration in this way. Others (Spiegelman *et al.,* 1996; Kelemen *et al.,* 1997) suggest that melt conduits are produced by reactive flow. Such melt-wallrock reaction would obviously lead to some modification of the melt and changes in wall-rock mineralogy (Kelemen *et al.,* 1992; Edwards and Malpas, 1995; Zhou *et al.,* 1996), *e.g.,* the dissolution of pyroxene leaving a residue of dunite. The second model envisages a concentration of melt-bearing mantle directly beneath the ridge axis, so that melt focussing is effectively buoyancy driven. Hirth and Kolstedt (1996) argued that mantle viscosities might be too high to allow for focussed buoyancy-driven flow, but their calculations support a

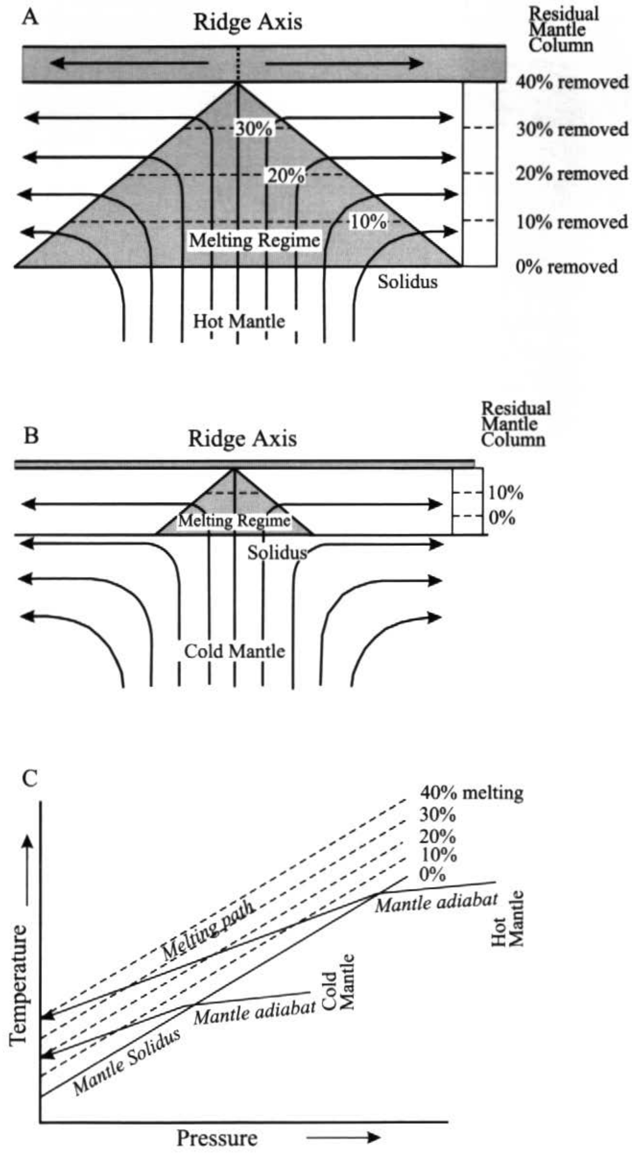


Figure 1 Steady state melting scenarios over **(A)** hot mantle and **(B)** cold mantle. **(C)** Hot mantle begins to melt deeper and melts to a greater extent while rising than cold mantle (after Langmuir *et al.*, 1992.)

model proposed by Phipps Morgan (1987) and Spiegelman and McKenzie (1987) in which pressure gradients in viscously deforming peridotite undergoing corner flow concentrate porous flow of melt beneath the ridge axis. Other mechanisms for focussing melt extraction have been proposed, *e.g.*, the formation of a permeability barrier that constricts upward flow of magma by partial crystallization on the margins of the upward sloping base of the lithosphere (Sparks and Parmentier, 1991), but the general consensus is that focussed porous mantle upwelling combined with channelized flow, is the most likely mechanism. The many pods, dykes and veins of dunite, commonly associated with chromitite in the mantle sequences of many ophiolites, are taken as direct evidence of this focussed flow involving melt-rock interaction (Leblanc and Ceuleneer, 1992; Zhou *et al.*, 1996; Kelemen *et al.*, 1997).

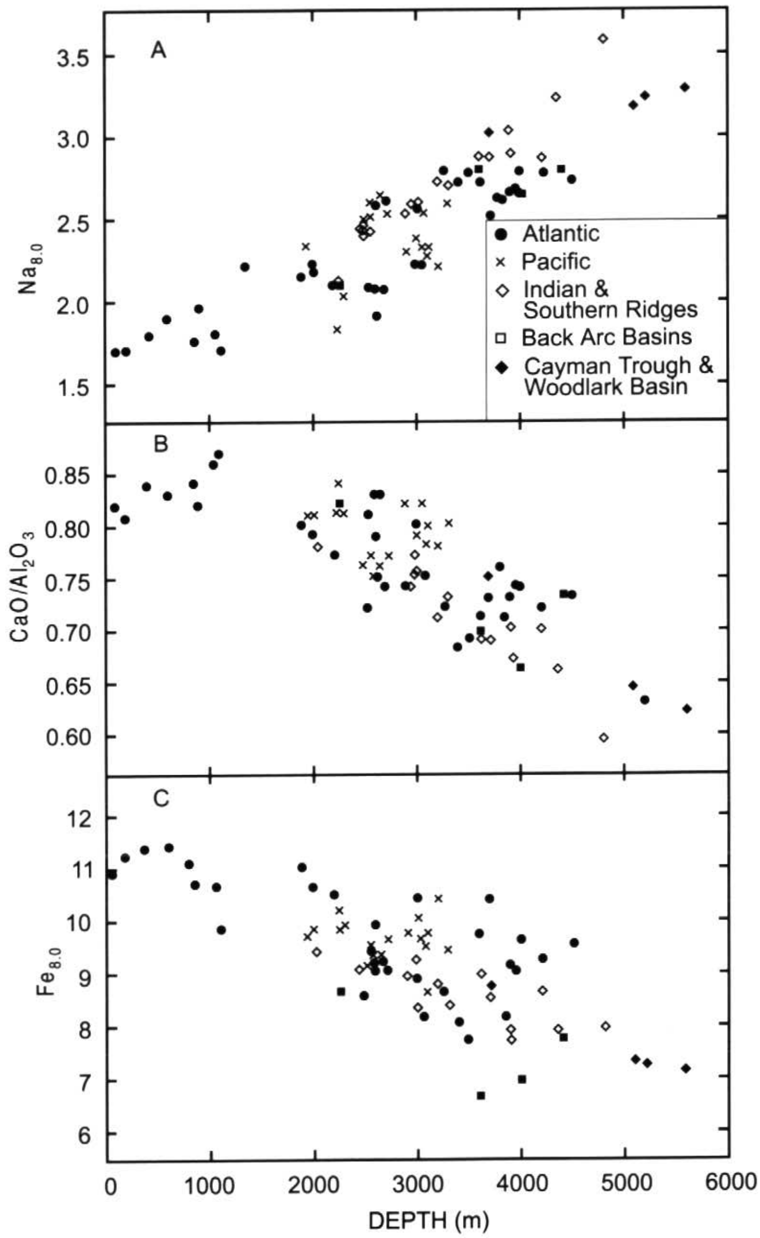


Figure 2 Variation of Na, CaO/Al_2O_3 and Fe in ridge axis lavas from the Atlantic, Pacific and Indian Oceans, from back arc basins and from the Cayman Trough and Woodlark Basin with spreading centre axial depth. Na and Fe are corrected to 8 wt% MgO in order to remove variations due to crystal fractionation. The chemistry is related to extent and pressure of melting; high degrees of melting produce a robust magmatic regime which is reflected in shallower ridge crest depths (from Klein and Langmuir, 1987).

SPREADING CENTRE MAGMA CHAMBERS

Geochemical evidence shows that primary MORB magmas undergo differentiation before eruption onto the seafloor. The processes of differentiation include crystal fractionation and magma mixing and mingling, which are considered to take place in shallow-level magma chambers beneath the spreading axis (Fig. 6). Evidence for such differentiation includes the fact that Mg numbers [100Mg/(Mg+Fe)] of MORB lavas are too low to be in equilibrium with mantle ferromagnesian minerals and that many MORB suites follow liquid lines of descent consistent with the fractionation of olivine, plagioclase and sometimes clinopyroxene at pressures less than 2 kb (6 km).

The nature of sub-rift magma chambers has been widely debated using mainly structural evidence from well-exposed ophiolites and seismic evidence from *in situ* lithosphere. Are they large, long-lived features that undergo continuous evolution or are they small, ephem-

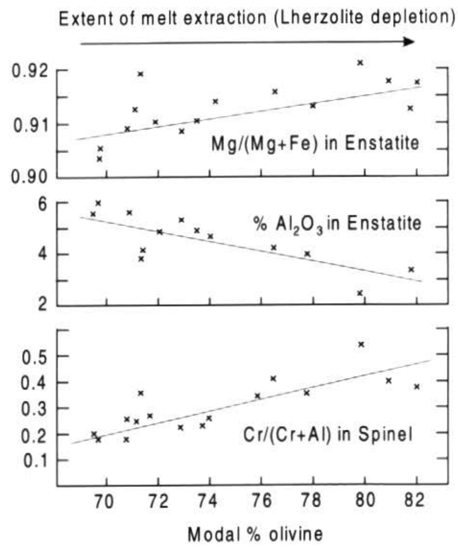


Figure 3 Modal percentage olivine in residual peridotites *versus* Mg# (100Mg/Mg+Fe), weight% Al_2O_3 in enstatite, and Cr# (100Cr/Cr+Al) in spinel, showing that with increasing melting, the compositions of residual phases change in a consistent manner (after Dick *et al.*, 1984.)

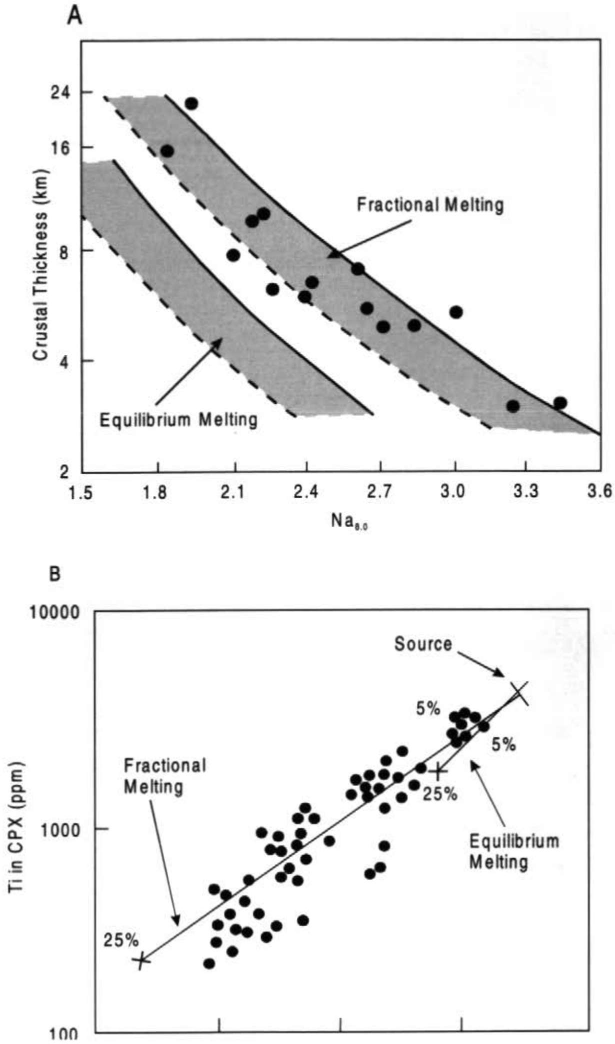


Figure 4 (**A**) Average Na contents of MORB (normalized to 8 wt% MgO) *versus* crustal thickness determined seismically, showing modelled results for different melting processes (After Langmuir *et al.,* 1992). (**B**). Clinopyroxene (Cpx) compositions from abyssal peridotites compared to trends from different melting processes (Johnson *et al.,* 1990). These plots suggest that fractional or "near fractional" melting is the dominant process in generating MORB melts.

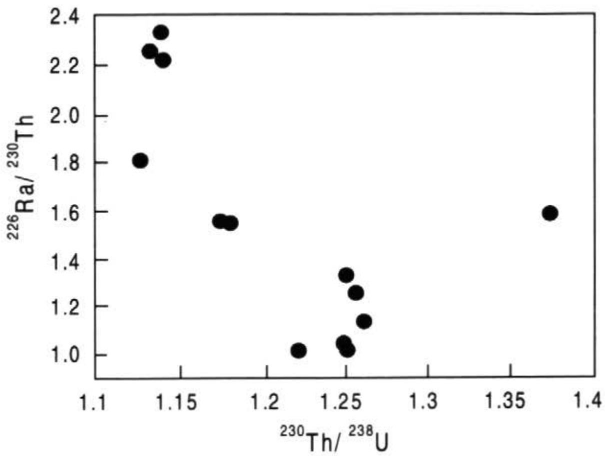


Figure 5 ^{226}Ra and ^{230}Th excesses in MORB. The negative correlation suggests that disequilibria are established at different depths (from Kelemen *et al.,* 1997).

eral features, which evolve independently? It can be reasonably argued that the sizes, shapes and longevity of magma chambers are likely to be very different at ridges with different spreading rates. Large sub-axial magma chambers are unlikely to exist beneath slow-spreading ridges on thermal grounds (Sleep, 1975) and thus, shallow-level magmatic processes are likely to be very different from those at fast-spreading ridges, where axial magma chambers have been seismically identified. On purely theoretical grounds, a magma chamber might be present but insignificant at half-spreading rates of about 1 cm per year, and as large as 20 km wide where half-spreading rates are as much as 6 cm per year (Kuznir, 1980). Nevertheless, every model of formation of oceanic crust by seafloor spreading invokes the presence, at some time, of some form of magma body beneath the spreading centre, not only to explain the diversity of MORB, but also the layered structure of the oceanic crust as deduced from seismic evidence and comparisons with ophiolite complexes (Cann, 1970, 1974; Christensen and Salisbury, 1975; Nisbet and Fowler, 1978). Yet the evidence from *in situ* oceanic lithosphere is sparse and, where available, is from scattered ridge segments. Thus, the available data may not be representative of any given spreading system as a whole.

Although most ophiolites are not typical pieces of oceanic lithosphere formed at mid-ocean ridges, they allow direct observation of lower ocean crust and upper mantle formed in spreading environments and thus provide many clues to the magmatic processes operating in these environments. Thus, current models for subrift magma chambers draw heavily on data from both *in situ* ocean lithosphere and from ophiolites.

EVIDENCE FROM THE OCEANS

Because very little is known about *in situ* sections of lower ocean crust, most data from the ocean basins come from geophysical studies of spreading ridges and petrological studies of the volcanic carapace.

Geophysical Studies

Geophysical studies can tell us little of the processes that take place in sub-rift magma chambers, but do place constraints on their size and shape. They also help to elucidate what exactly constitutes a magma chamber.

There have been a number of detailed seismic reflection and refraction studies of the northern portion of the East Pacific rise. Reflection experiments by Herron *et al.* (1978), Hale (1982) and Detrick *et al.* (1987) reported a bright, subhorizontal reflector in the upper crust (Fig. 7). This high-amplitude feature can be tied laterally to a broader low-velocity zone (LVZ) beneath the ridge crest, extending from 1-2 km below the seafloor to the base of the crust (McClain *et al.*, 1985; Vera *et al.*, 1990; Kent *et al.*, 1990). Within this zone, the lowest velocities are confined to a mid-crustal layer less than 1 km thick. Seismic tomography experiments by Toomey *et al.* (1990) indicate that the LVZ continues along the ridge axis, although its thickness is variable, generally reaching a minimum across axial discontinuities between ridge segments. Most of the LVZ has only a relatively small velocity anomaly, suggesting that it consists of effectively solid but still hot material. Only the lowest velocities are likely to represent any significant melt fraction (Sinton and Detrick, 1992). However, melt might be concentrated beneath the axial reflector, which is best modelled with zero or near-zero shear wave velocities and interpreted as the roof of an axial magma chamber (Harding *et al.*, 1989; Vera *et al.*, 1990). The depth to the top of this magma chamber varies from 1.2 km to 2.4 km below the seafloor (Detrick *et al.*, 1987) and its width across the ridge is no more than 3 km, and likely less (Kent *et al.*, 1990). Although highly interpretative, seismic evidence suggests that the molten part of the magma body is less than a few hundred metres thick, although it may extend for several kilometres off axis (Garmany,1989).

Although mid-crustal reflectors, interpreted as the top of an axial magma chamber, have also been recorded on the Juan de Fuca Ridge (Rohr *et al.*, 1988) and the Valu Fa Ridge in the Lau Basin (Collier and Sinha, 1990), such features are rare on the Mid-Atlantic Ridge. The best documented example on the Mid-Atlantic Ridge is from the Reykjanes segment near the Iceland hot spot, where a variety of geophysical data provides evidence for a crustal body of melt at a depth of 2.0 km to 2.5 km (Sinha *et al.*, 1999). The Reykjanes Ridge may be atypical of the Mid-Atlantic Ridge because of its proximity to the Iceland plume but Sinha *et al.* (1999) argue that the rate of melt production in this area is no higher than for other parts of the ridge. At other locations along the ridge, low seismic velocities associated with

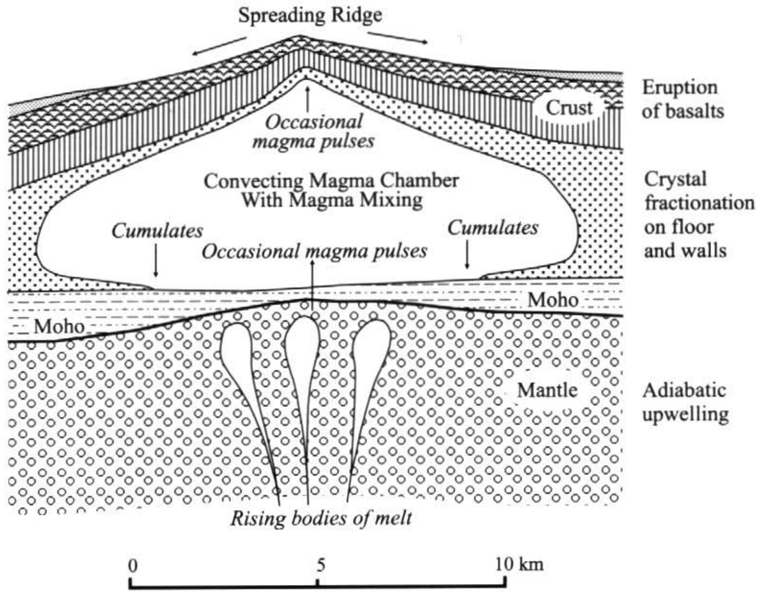


Figure 6 Schematic diagram showing the processes occurring in a sub-rift magma chamber at a mid-ocean ridge (after Brown and Mussett, 1981). Note that the size and shape of individual magma chambers are highly variable.

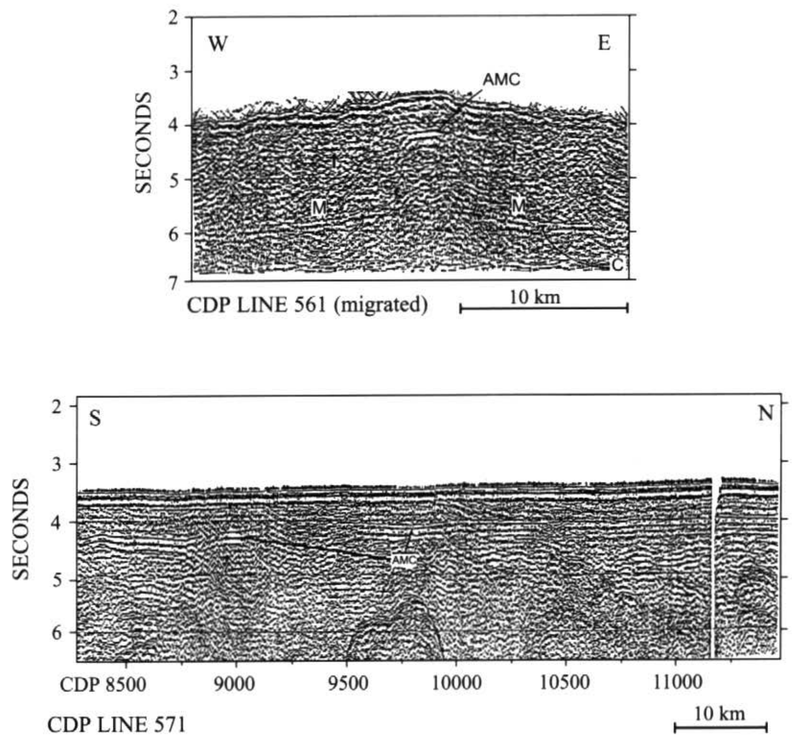


Figure 7 Multichannel seismic reflection profiles across the EPR near 9°30'N. The bright reflector (AMC) representing the top of the axial magma chamber can be seen in mid-crust (from Detrick *et al.*, 1987).

gravity lows beneath topographic highs, have been interpreted as volumes of upwelling hot rock possibly containing small isolated melt pockets (Schouten *et al.*, 1985; Lin *et al.*, 1990). Nevertheless, it is generally accepted that magma budgets at intermediate- to slow-spreading ridges like the Mid-Atlantic Ridge or Southwest Indian Ridge, are lower than at fast-spreading ridges and that extension can involve both magmatic and tectonic processes (Tucholke *et al.*, 1997; Dick *et al.*, 1992).

Petrological Evidence

Studies of the mineralogy of MORB lavas suggest that magma mixing at shallow levels is as important as crystal fractionation in controlling the composition of erupted magmas (Rhodes *et al.*, 1979; Walker *et al.*, 1979; Bryan, 1983; Langmuir, 1989). Such magma mixing can explain the apparent enrichment in incompatible elements in MORB beyond that expected from normal Rayleigh fractionation. However, the fact that a range of magma types derived from distinct mantle sources can be found side by side, or even interlayered, along ridge axes suggests that magma mixing in relatively large chambers is limited, and/or that magma chambers are small and generally isolated (Flower and Robinson, 1981a, 1981b). Certainly, many studies of MORB show that compositions can vary over distances of a few hundred metres, irrespective of tectonic setting or spreading rate (Bryan, 1979; Thompson *et al.*, 1989; Sinton *et al.*, 1991; Hawkins *et al.*, 1990). Flower (1980) and Langmuir *et al.* (1992) have suggested that the relative importance of fractionation *versus* magma mixing might be related to spreading rate. They argued that the variations in lava composition from fast-spreading ridges such as the EPR can be explained by shallow-level crystal fractionation, whereas lavas from slow-spreading ridges, such as the Mid-Atlantic Ridge, provide more evidence for magma mixing and mingling, particularly in the form of disequilibrium phenocrysts. Basalts erupted at fast-spreading ridges tend to be aphyric to weakly phyric whereas those erupted at slower-spreading ridges tend to be highly olivine- and plagioclase-phyric. Pyroxene phenocrysts are rare in MORB, and where they do occur are usually rounded and corroded, indicating a lack of equilibrium with the melt. Sinton and Detrick (1992) note that the processes of magma mixing and crystal fractionation effectively oppose one another during melt differentiation, the former buffering compositional variations by replenishment with more primitive magma from the mantle. The relative importance of the two processes at ridges of different spreading rate might therefore be indicated by the range of compositions of erupted lavas at each. Small-scale temporal and spatial heterogeneities of MORB have been well documented on the East Pacific Rise (Perfit *et al.*, 1994).

Until recently, the products of magma crystallization at depth in oceanic layer 3 were little known because of the difficulty of sampling *in situ* sections. However, drilling at ODP Sites 735, 894, 895 and 1105 has provided many new insights into the origin and evolution of oceanic gabbros. Hole 735B on the Southwest Indian Ridge penetrated 1508 m into the lower crust with 87% recovery, providing a unique sample of oceanic Layer 3 (Robinson, Von Herzen *et al.*, 1989; Shipboard Scientific Party, 1999). During Leg 176, Hole 1105A, near the centre of Atlantis Bank was drilled to a depth of 158 m with 75% recovery (Pettigrew, Casey, Miller *et al.*, 1999). The thick plutonic section sampled in these holes consists mainly of olivine gabbro with lesser amounts of troctolite, gabbro, gabbronorite and Fe-Ti oxide gabbro. Five major intrusive bodies can be recognized, each constructed from many small magmatic units (Shipboard Scientific Party, 1999). Some of these small units appear to be sill-like bodies (Dick *et al.*, 1991; Bloomer *et al.*, 1991) whereas others have diffuse boundaries and may represent conduits through crystal mushes for upwardly-migrating melts. Each major intrusive body grades upward from troctolite and olivine gabbro at the base, through gabbro and gabbro norite, to highly fractionated iron- and titanium-rich gabbro at the top. The oxide-rich gabbros and gabbronorites are commonly associated with high-temperaure, ductile shear zones ranging from a few millimetres to many metres thick, not unlike those features which severely modify cumulate layering in the Bay of Islands ophiolite (Dunsworth *et al.*, 1986; Malpas, 1987). Mass-balance calculations for the rocks of Hole 735B suggest that the Fe-Ti oxide gabbros must have migrated into the section, either laterally or from depths below the bottom of the hole. In this ultraslow-spreading environment, both magmatic and amagmatic extension seem to have taken place, and crustal construction involved dynamic processes of tectonism and hydrothermal circulation as well as magmatism (Dick *et al.*, 1991, 1992).

Gabbroic rocks recovered from the Hess Deep (ODP Sites 894 and 895 are primarily gabbronorite and olivine gabbronorite with less amounts of gabbro, olivine gabbro and oxide gabbronorite (Gillis, Mevel, Allan *et al.*, 1993). The section lacks the abundant oxide gabbros and well-developed ductile shear zones characteristic of Hole 735B. Layering is absent in these rocks and they are all adcumulates (Pedersen *et al.*, 1996; Natland and Dick, 1996). The average residual melt porosity was about 4.5% and the rocks show strong evidence for migration of melts through the cumulates.

Based on the available geophysical and petrologic data, Sinton and Detrick (1992) have proposed two end-member models for magma chambers beneath fast- and slow-spreading ridges (Fig. 8). Beneath fast-spreading ridges, with high rates of magma supply, narrow sills of melt are postulated to intrude a mush zone, 1-2 km beneath the ridge axis. The crystal mush is surrounded by a transition zone with a more rigid framework, which passes outward into hot, solid rock. The LVZ, which includes the volume contained within the outer margin of the transition zone, defines a region in which melt might exist, albeit as small isolated patches except for the central melt lens. The relative amounts of melt and mush vary along the ridge axis and the melt lens may not exist at ridge discontinuities. What actually constitutes the "magma chamber" in such a model, is equivocal; does it comprise simply melt, melt and mush, or in addition, part of the transition zone? If the magma chamber is defined as that portion of the system that lies above the solidus temperature then it must be remembered that the solidus within this zone is likely not isothermal and will depend upon local melt compositions.

Beneath most slow-spreading ridges, with restricted magma supply, there is no steady state magma lens. Rather, small intrusive mush zones are thought to form layer 3, and some of the crustal extension is amagmatic. The listric normal faults bordering the central rift valley of slow-spreading ridges likely bottom out at the brittle-ductile boundary within the transition zone (Fig. 8).

These end-member models successfully explain a number of features observed at ocean ridges, including:

1. Most MORB lavas result from melts that migrate through the transition zone and crystal mush;
2. Separation of melt into isolated lenses, as beneath fast- and intermediate-spreading centres, promotes the local eruption of ferro-basalts and the products of advanced differentiation (Sinton and Detrick, 1992); and
3. Magma mixing and disequilibrium phenocrysts are more evident at slow-spreading ridges where eruptions are related to the injection of new primitive magma from the mantle into the higher-viscosity crystalline mush beneath the spreading axis.

Evidence from Ophiolites

Early studies of ophiolites proposed the existence of large axial magma chambers, several kilometres wide and deep, particularly beneath fast-spreading ridges (Cann, 1970; 1974; Smewing, 1981; Pallister and Hopson, 1981; Robson and Cann, 1982). These single magma chamber models showed the magma occupying the total thickness of the oceanic crust, as much as 5 km. The various models differed basically only in proposing chambers of different shape (Fig. 9). Crystallization along the margins of the chamber was thought to produce the upper isotropic gabbros of layer 3 (Fig. 9a), and settling of crystals to the bottom of the chamber was invoked to produce the lower layered gabbros. Layer 2 pillow lavas and sheeted dykes were thought to be produced by eruptions from the chamber.

The presence of continuous, flat-lying cumulate layers, several kilometres in length, in the Oman ophiolite (Pallister and Hopson, 1981), led to the concept of a wide, wing-shaped magma chamber with significant temperature differences between the central feeder zone and the extremities (Fig. 9b). Casey and Karson (1981) observed inclined cumulate layering in the plutonic section of the Bay of Islands ophiolite and suggested crystallization on the walls of a large, rounded magma chamber rather than on the floor (Fig. 9c).

All of these models considered that the magma chamber underwent open system fractionation, erupting lavas onto the seafloor and being replenished by primitive melts from a central, mantle-tapping conduit. Browning (1984) recognized that, in an open system, the cryptic variation displayed by the layered gabbros in the Oman ophiolite required crystallization from a magma body much thinner than the overall thickness of the Layer 3 gabbros. Two possible models were used to illustrate this (Fig. 10), one requiring a layered magma chamber undergoing crystallization by double-diffusive convection, the other a smaller

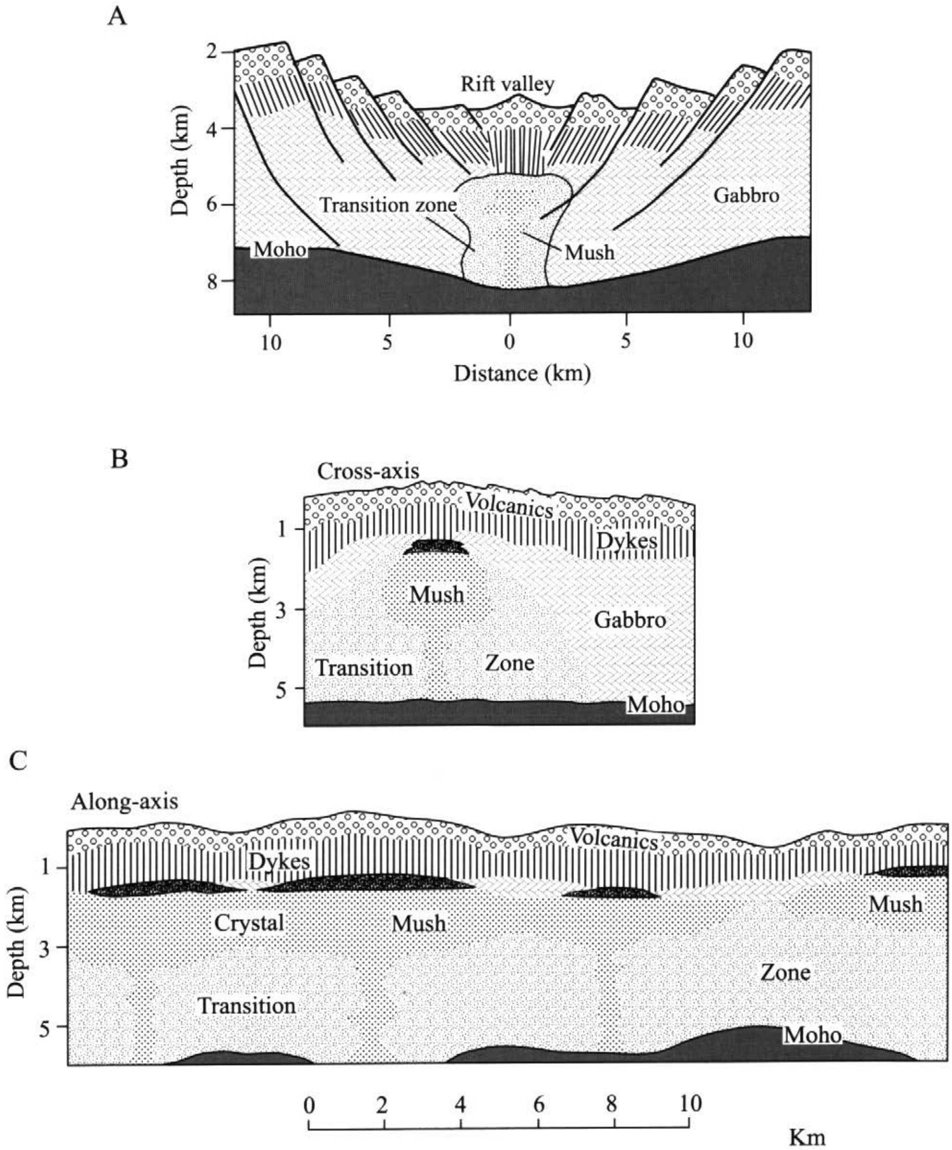


Figure 8 (**A**) Interpretative model of a "magma chamber" below a slow-spreading ridge. Note the preponderance of listric normal faults. After Sinton and Detrick (1992). (**B**) Interpretative model of "magma chambers" beneath a fast- or intermediate- spreading ridge. (**C**) The thin melt lens is seen at the top of a crystal mush zone, which is continuous along the strike of the ridge. Eruptions of basalts occur mainly from this melt lens (after Sinton and Detrick, 1992).

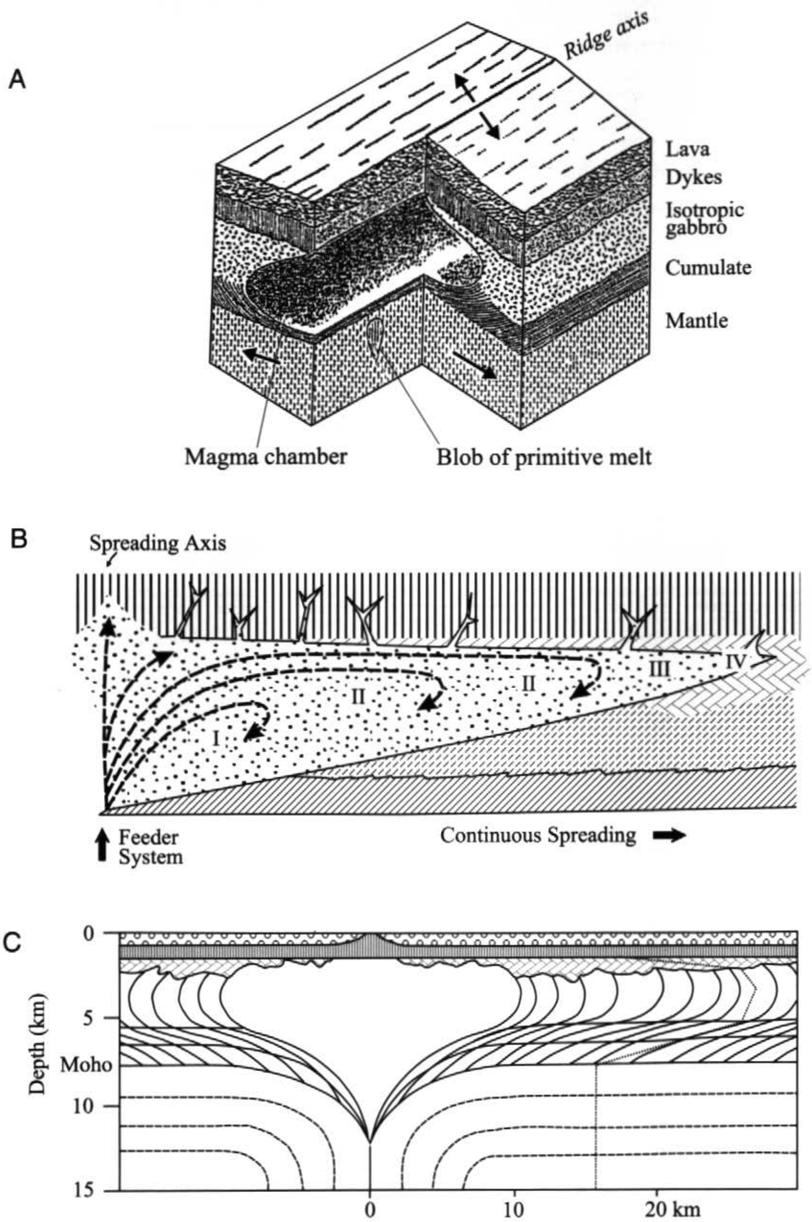


Figure 9 Various concepts of large, single magma chambers derived from ophiolite studies. (**A**) the "infinite onion" model; (**B**) the "wing-shaped" model; and (**C**) the "axial trough model" (references in text).

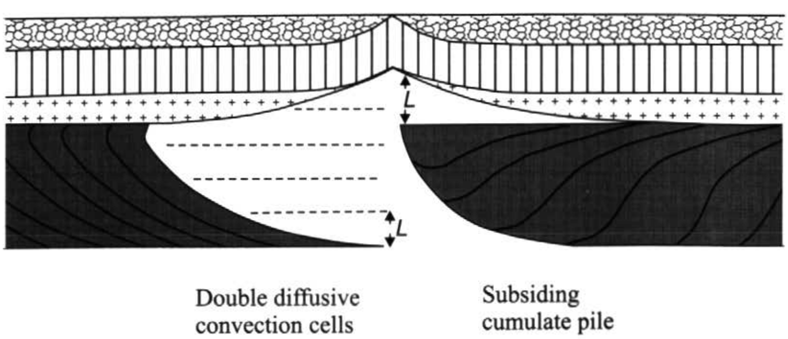


Figure 10 Browning's (1984) comparison of a layered magma chamber (**left**) *versus* a subsiding magma chamber (**right**) to explain the cryptic variation defining a thin magma body (*L*) in the Oman ophiolite.

magma lens high in the crust, giving rise to the cumulate gabbro section by subsidence, as originally proposed by Dewey and Kidd (1977).

In order to explain structural features observed in the Oman ophiolite, Nicolas *et al.* (1988) and Nicolas (1989) proposed a chamber with a flat floor, a narrow roof, and layering in the gabbros parallel to the solidifying walls (Fig. 11). They suggested that, at least close to its walls and floor, away from the ridge axis, such a magma chamber would be filled with a crystal mush sufficiently viscous to transmit the shear stress imparted by the lateral translation of the underlying asthenosphere. If pure melt was present at all, it would be confined to a very narrow central zone. In this model, focussing of the magmatic activity to a narrow neo-volcanic zone at the ridge axis is partly a result of magma chamber geometry.

At the same time as these models of single, large and long-lived sub-rift magma chambers were being proposed, others were arguing, also on the basis of field studies of ophiolites, that the plutonic sequence must be formed from multiple, small and ephemeral magma bodies. For example, Moores and Vine (1971) and Allen (1975) considered that at least the upper part of the plutonic series in the Troodos ophiolite had to be derived from a number of small magma bodies (Figs. 12, 13). Strong and Malpas (1975) put forward a similar model for the Newfoundland ophiolites.

The Troodos and Oman ophiolites have been the focus for studies of oceanic plutonic rocks for many years. During the Cyprus Crustal Study Project in the 1980s (Robinson and Malpas, 1990), extensive areas of the plutonic section were mapped, from the mantle peridotites to the base of the sheeted dyke complex. On the basis of petrology, state of deformation and cross-cutting features, two major plutonic suites were recognized with clear intrusive relationships between them (Malpas, 1990). The most widespread gabbros in Troodos are poorly layered, mesocratic rocks with variable grain size from fine-grained to pegmatitic, for which the names "segregation gabbros" (Aldiss, 1978) or "vari-textured gabbros" (Pedersen and Malpas, 1984) have been used. These occur throughout the plutonic section, in places associated with plagiogranites, and are thought to have formed close to the roofs of intrusive magma bodies (Pedersen and Malpas, 1984). Likewise, layered gabbros and associated ultramafic rocks are not restricted to the base of the plutonic sequence, but are found at various levels throughout the cumulate section. However, unlike the layering described for Oman by Pallister and Hopson (1981), the layering in Troodos is restricted laterally to distances of a few hundred metres. Thus, field relationships among the plutonic rocks of the Troodos ophiolite are complex and variable, as are those between plutonic rocks and the carapace of sheeted dykes.

In Troodos, the base of the sheeted dyke complex is most commonly intruded by plutonic rocks; less commonly, dykes cut the plutonic rocks. Least common is the case in which dykes arise directly from their plutonic equivalents. The relief on the sheeted dyke/plutonic contact is in excess of 500 m because dyke swarms cut deep into the plutonic section, and cupolas of plutonic rocks punch high into the dyke section (Fig. 14). These relationships can only be explained if the plutonic section was constructed from multiple bodies of magma, which propagated dykes laterally along the ridge axis as well as vertically to erupt pillow lavas (Baragar *et al.,* 1990).

The field relationships of the plutonic rocks observed in Troodos are not confined to this ophiolite alone. It is now clear that the simple layered structure of residual mantle peridotites overlain by cumulate ultramafic rocks, which are succeeded by layered gabbros, isotropic gabbros and plagiogranites, and finally, sheeted diabase dykes, is no longer valid. The relationships seen in most ophiolite massifs support the existence of multiple magma chambers.

The multiple magma chamber model proposed for the Troodos ophiolite by Malpas (1990), is reproduced in Figure 15. These diagrams are intended to illustrate concepts rather than particular geological features. The upper part of the plutonic sequence is derived from small, ephemeral magma chambers that originate by fractionation from larger, more mafic bodies, the remnants of which are found lower in the plutonic section. These deeper plutons are dominated by ultramafic rocks with lesser gabbros and few plagiogranitic differentiates. Melts accumulate at the base of the crust because of lithologic differences and density contrasts between crustal and mantle rocks. The shape of the magma body at this stage is controlled by the regional stress field. The magma body slowly rises through the crust as a result of buoyancy, but makes room for itself in the main through stoping and assimilation. Energy for this process is supplied both from the incipient heat of the magma and from the latent heat of crystallisation of ultramafic rocks at the base of the advancing chamber. The composition of the magma is effectively buffered at this stage by the steady input of fresh parental magma from below, and the assimilation of essentially gabbroic rocks along the walls and roof of the chamber. At a level in the crust controlled by its structural state, the magma chamber receives little fresh supply from below and is cooled by hydrothermal circulation from above. It shrinks through crystallization and undergoes more advanced fractionation, while the older parts of the crust undergo high-temperature deformation during spreading. A roof assemblage of varitextured gabbro is formed, which may produce a hornfelsed contact aureole in the adjacent rocks. Final consolidation of the magma body produces the most fractionated rock types: plagiogranites, which are trapped between isotropic gabbros and the roof assemblage. The repetition of this sequence of events, with magma chambers of different sizes and minor compositional variation, results in the complicated plutonic arrangement seen in Troodos.

Several points are worth emphasizing in this model. Magma in the chamber changes from an initial magnesium-rich composition to a more differentiated composition as a result of magma replenishment and mixing, crystal fractionation,

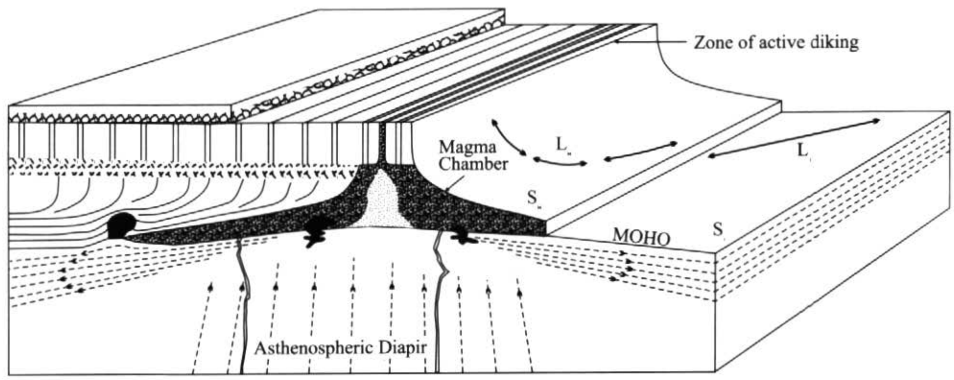


Figure 11 The "tent-shaped" model of Nicolas *et al.* (1988) based on structural evidence from the Oman ophiolite.

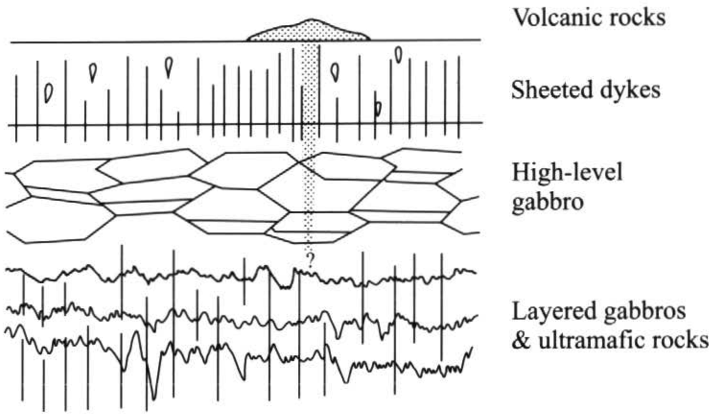


Figure 12 The model of Moores and Vine (1971) for the structure of the Troodos ophiolite. They recognized the presence of numerous small magma "cells" in the high-level gabbros.

and wall rock assimilation. The first process is dominant lower in the crust where periodic replenishment allows the accumulation of mafic phases at the lower levels of the nested plutons. Crystal fractionation is an important process throughout the magma evolution, providing layered cumulate rocks early on, and diorites and plagiogranites late in the sequence. Assimilation plays an important part in the evolution of the magma as it rises to upper crustal levels. Above the high-level plutonic rocks, hydrothermal circulation of seawater actively cools the crust (Spooner, 1980; Schiffman *et al.,* 1987). This circulation is driven by the heat of the crystallizing magma, and upwelling zones are focussed above high-level intrusions or cupolas in the crustal section. Seawater can come close to the magma bodies *via* fractures in the sheeted dyke complex produced by brittle failure of extending crust and the development of planar and listric normal faults. The hornfelses resulting from local thermal metamorphism of the dyke complex by intrusive gabbroic rocks have characteristics of a conductive boundary layer separating a convecting hydrothermal cell from the magmatic heat source (Gillis and Roberts, 1999). Water may actually enter the magma by assimilation of hydrated, greenschist-facies roof rocks leading to volatile transport of hygromagmatophile elements from the magma and the formation of plagiogranites (Malpas *et al.,* 1989).

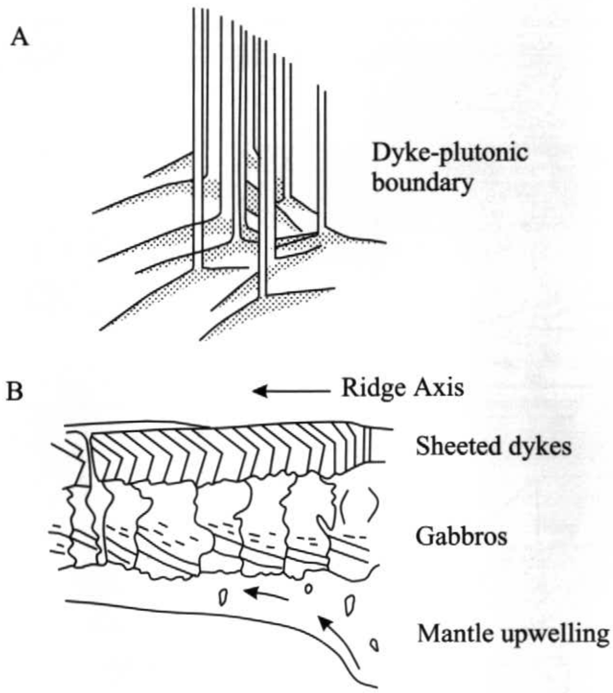


Figure 13 (**a**) Schematic illustration of the relationship between sheeted dykes and plutonic rocks in the Troodos ophiolite and (**b**) Model showing the formation of nested magma chambers and their relation to off-axis magmatism (after Allen, 1975).

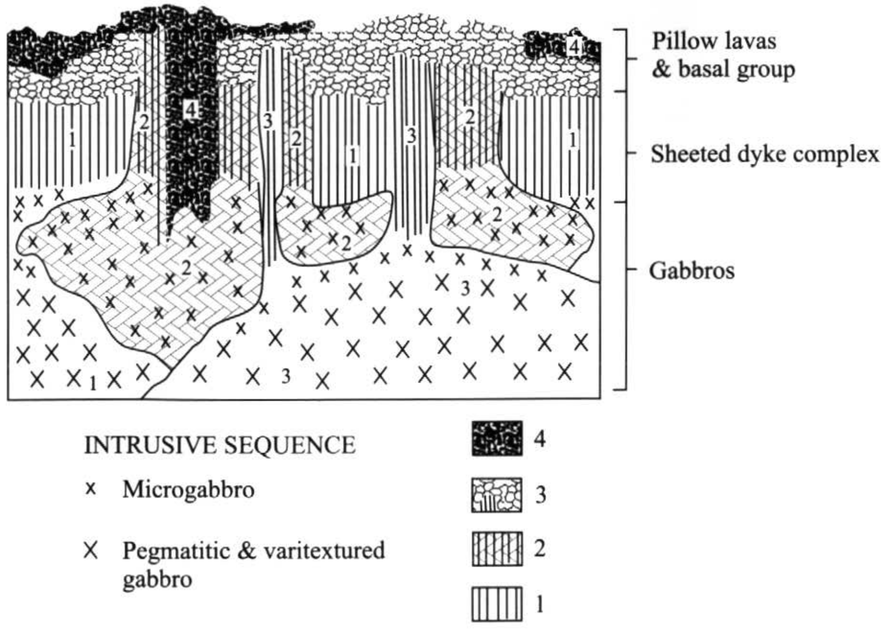


Figure 14 Schematic illustration of the relationships between the sheeted dyke complex and upper plutonic rocks of the Troodos ophiolite. Multiple chambers are intruded to different heights in the sequence 1-4, and give rise to their own portion of the overlying sheeted dyke complex and extrusive section (after Malpas, 1990).

The structural state of the crust, *i.e.,* whether it deforms by brittle or ductile processes, is directly related to its temperature profile. The intrusion of hot magma raises the geothermal gradient, but the level of intrusion depends upon the presence of fractures and faults, which enable cooling of the magma body by hydrothermal circulation. Hydrothermal circulation is likewise dependent upon a heat source and crustal permeability. Thus, magmatic, tectonic and hydrothermal processes are intimately tied together in a complex feedback mechanism, which leads to self-ordering of the system. This self-ordering provides the consistency of crustal structure and can ultimately be switched off only by the lack of magma supply, leading to cooling of the crust.

Recent studies of the Oman ophiolite (Boudier *et al.,* 1996; Kelemen *et al.,* 1997; Korenaga and Kelemen, 1997) have emphasized the differences between lower gabbro cumulates and upper isotropic and varitextured gabbros, and the importance of "sheeted" gabbro sills in crustal construction These sills are compositionally similar to the lower gabbros, but intrude dunite in the crust-mantle transition zone, and are believed to make up much of the lower oceanic crust (Fig. 16). They are considered to form beneath permeability barriers within crystallizing gabbroic crust and to be emplaced close to their final crystallization depths. One such barrier is the mantle-crust boundary, the site of initial magma pooling in the model for Troodos. However, in preference to having the sills stope and assimilate their way upward through the crust, Kelemen *et al.* (1997) consider that individual sills crystallize in place, with those higher in the crust being replenished periodically through hydrofractures from below. Similarly, these sills are themselves tapped by periodic hydrofractures to transport more fractionated liquids to the uppermost parts of the crust, providing an overall upward trend to more evolved compositions.

Some of the small, high-level magma chambers in the Troodos ophiolite are sill-like in shape and may have formed by this mechanism. However, these are composed of mafic and ultramafic rocks derived from primitive melts presumably associated with the depleted lavas in the extrusive section rather than from evolved magmas.

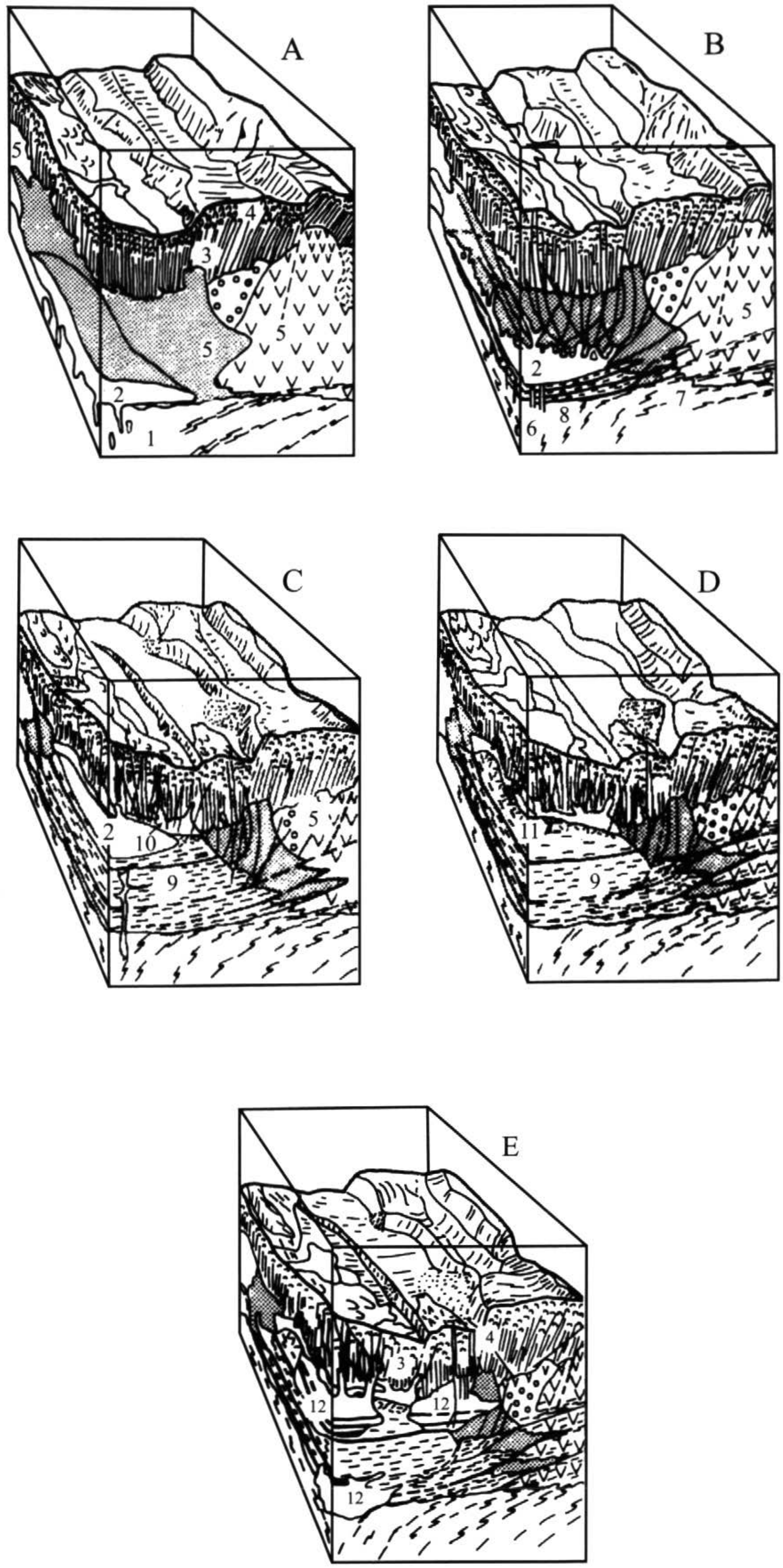


Figure 15 Cartoon of the sequential development of the Troodos plutonic complex by the intrusion of multiple magma bodies and penecontemporaneous ductile and brittle deformation associated with spreading. (**A**) Accumulation of magma at the base of the crust. (**B**) Upward progression of the magma chamber by stoping and assimilation. Heat is supplied in part by crystallization of cumulate rocks. (**C**) Cut off from source supply and cooled from above, the chamber shrinks and differentiates rapidly. Older parts of the crust undergo deformation at high temperatures. (**D**) Final consolidation of the magma body and crystallization of plagiogranites in the roof assemblage. (**E**) Repetition of the sequence with chambers of different sizes results in the complex of nested magma chambers.
Legend: **1**, Mantle tectonites; **2**, Magma body (mush and melt); **3**, Sheeted dykes; **4**, Pillow lavas; **5**, Older plutonic bodies; **6**, Input of primitive magma; 7, High temperature deformation; **8**, Ultramafic layers; **9**, Layered gabbros; **10**, Roof assemblage of varitextured gabbros; **11**, Plagiogranite and dioritic bodies; **12**, Later plutonic bodies (after Malpas, 1990).

CONCLUSIONS

Ever since the recognition of seafloor spreading there has been much debate over subrift magma chamber processes and their control on the chemistry of eruptive magmas at ocean ridges. In the last 10 years, geophysical investigations of spreading ridges, combined with detailed investigations of portions of *in situ* oceanic crust accessible through drilling and well-exposed ophiolites, have led to a rapid evolution of our ideas on the magmatic processes that underpin crustal accretion processes. The older models derived from ophiolite studies, of large, long-lived chambers, have been superceded by models of small, ephemeral, possibly nested chambers, whose dimensions are constrained by seismic data from present-day ocean ridges. The properties of magma chambers are clearly dependent upon magma supply rate and thus can be correlated on a large scale with spreading rate. Magma supply rates, *i.e.,* the rates of melt production and extraction, ultimately depend upon the nature and geometry of mantle flow beneath spreading centres. Yet this is one of the least understood aspects of ridge processes. Mantle flow beneath the ocean basins has been interpreted as passive flow driven by plate separation or active flow driven by the buoyancy of mantle containing partial melt (Phipps Morgan and Chen, 1993). High rates of spreading and high mantle viscosities appear to favour passive flow, but mantle properties can be greatly modified by the presence of water (Hirth and Kohlstedt, 1996).

Another question inadequately addressed to date is the actual process of accretion of lower crust from a body of magma. The two most popular scenarios are the "gabbro glacier'" model (Quick and Denlinger, 1993) and the "sheeted sill" model (Henstock *et al.,* 1993; Boudier *et al.,* 1996; Kelemen *et al.,* 1997). In the former, the complete lower crust is produced by crystallization at a thin magma lens directly beneath the sheeted dykes, with subsequent subsidence and lateral flow. In the latter, the crust is produced by a series of nested sills. The viability of these different models depends on the viscosity and rheology of partially molten gabbro, for which laboratory data are scarce. One of the most difficult factors to model is the amount of melt present in the mush at each stage of deformation.

Another somewhat puzzling feature of the ocean crust is that, whatever the nature of spreading, it develops an overall gross uniformity in structure. The seismically defined crustal layers are relatively constant in thickness and in overall lithology. Once the lithosphere ages, the metamorphic profile is generally predictable. Given the variety of processes involved in the generation of oceanic crust, this could only be achieved through extensive interplay and feedback resulting in a self-ordered system. This is clearly the case at the interface between the lithosphere and hydrosphere, where magmatism, hydrothermal circulation, and deformation style are implicitly interdependent. We are now beginning to realize that similar interactions operate between the asthenosphere and lithosphere where oceanic crust is generated.

ACKNOWLEDGMENTS

We thank Andrew Hynes and Kathyrn Gillis for perceptive and thoughtful reviews of the manuscript, and Xu-Feng Hu for assistance in preparation of the illustrations.

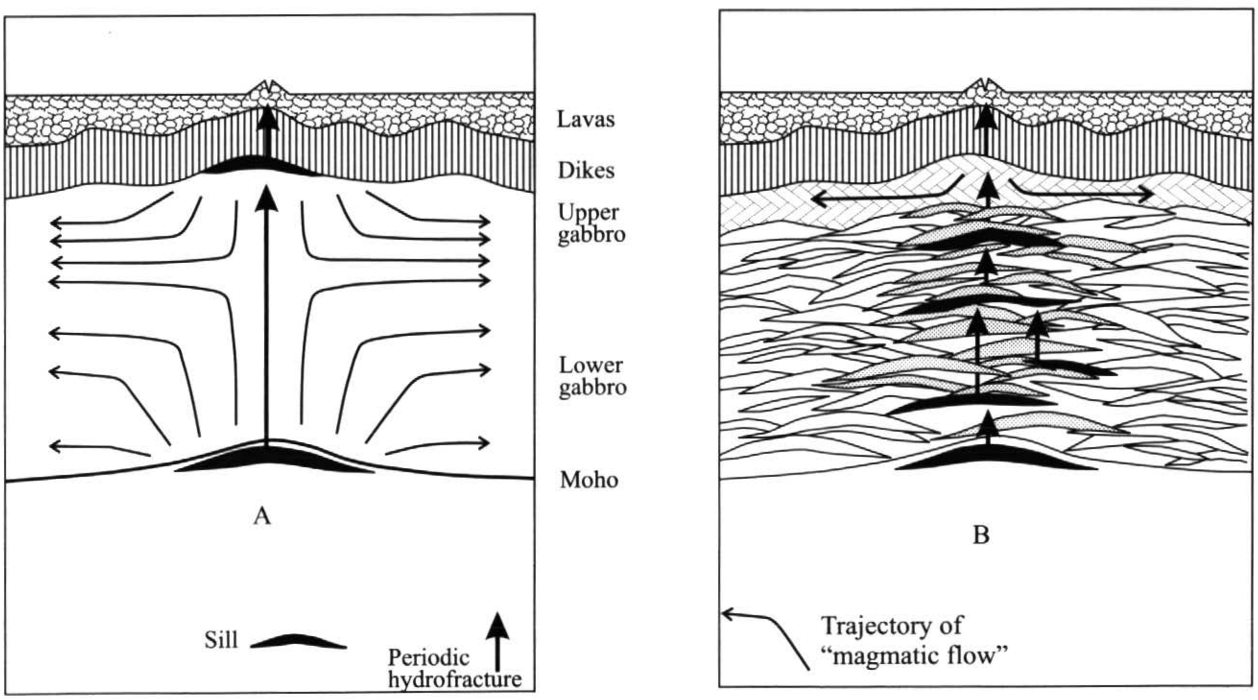


Figure 16 Models showing formation of layer 3 gabbros from sill-like bodies. (**A**) shows crystallization within two sills followed by ductile flow (after Schouten and Denham, 1995). (**B**) "sheeted-sill" model of Kelemen *et al.* (1997). The heavy arrows indicate movement of melt by local hydrofracturing.

REFERENCES

Aldiss, C.R., 1978, Granitic Rocks of Ophiolites: unpublished Ph.D. Thesis, Open University, UK, 198 p.

Allegre, C.J. and Bottinga, Y., 1974, Tholeiite, alkali basalt and ascent velocity: Nature, v. 252, p. 31-32.

Allen, C.R., 1975, The Petrology of a Portion of the Troodos Plutonic Complex, Cyprus: unpublished Ph.D. Thesis, Cambridge University, UK, 335 p.

Baragar, W.R.A., Lambert, M.B., Baglow, N. and Gibson, I.L., 1990, The sheeted dyke zone in the Troodos ophiolite, *in* Malpas, J., Moores, E.M., Panayiotou, A. and Xenophontos, C., eds., Ophiolites. Oceanic Crustal Analogues: Geological Survey of Cyprus, p. 37-52.

Bloomer, S.H., Meyer, P.S., Dick, H.J.B., Ozawa, K. and Natland, J.H., 1991, Textural and mineralogical variations in gabbroic rocks from Hole 735B, *in* Von Herzen, R.P., Robinson, P.T., *et al.*, eds., Proceedings: Ocean Drilling Program, Scientific Results, v. 118, p. 21-40.

Boudier, F., Nicolas, A. and Ildefonse, B., 1996, Magma chambers in the Oman ophiolite: fed from the top and the bottom: Earth and Planetary Science Letters, v. 144, p. 239-250.

Bowen, N.L., 1928, The Evolution of the Igneous Rocks: Princeton University Press.

Brown, G.C. and Mussett, A.E., 1981, The Inaccessible Earth: Allen and Unwin, London.

Browning, P., 1984, Cryptic variations within the cumulate sequence of the Oman ophiolite: magma chamber depth and petrological implications: Geological Society of London, Special Paper 13, p. 71-82.

Bryan, W.B., 1979, Regional variation and petrogenesis of basalt glasses from the FAMOUS area, MAR: Journal of Petrology, v. 20, p. 293-325.

Bryan, W.B., 1983, Systematics of modal phenocryst assemblages in submarine basalts: petrologic implications: Contributions to Mineralogy and Petrology, v. 83, p. 62-74.

Cann, J.R., 1970, New model for the structure of the ocean crust: Nature, v. 226, p. 928-930.

Cann, J.R., 1974, A model for oceanic crustal structure developed: Geophysical Journal of the Royal Astronomical Society, v. 39, p. 169-187.

Casey, J.F. and Karson, J.A., 1981, Magma chamber profiles from the Bay of Islands ophiolite complex: Nature, v. 298, p. 295-301.

Christensen, N.I. and Salisbury, M.H., 1975, Structure and constitution of the lower oceanic crust: Reviews of Geophysics and Space Physics, v. 13, p. 57-86.

Collier, J. and Sinha, M., 1990, Seismic images of a magma chamber beneath the Lau Basin back-arc spreading centre: Nature, v. 346, p. 646-648.

Detrick, R.S., Buhl, P., Vera, E., Mutter, J., Orcutt, J., Madsen, J. and Brocher, T., 1987, Multichannel seismic imaging of a crustal magma chamber along the East Pacific Rise: Nature, v. 326, p. 35-41.

Dewey, J.F. and Kidd, W.S.F., 1977, Geometry of plate accretion: Geological Society of America, Bulletin, v. 88, p. 960-968.

Dick, H.J.B., 1989, Abyssal peridotites, very slow spreading ridges and ocean ridge magmatism, *in* Saunders, A.D. and Norry, M.J., eds., Magmatism in the Ocean Basins: Geological Society, Special Publication 42, p. 71-105.

Dick, H.J.B., Fisher, R.L. and Bryan, W.B., 1984, Mineralogic variation of the uppermost mantle along mid-ocean ridges: Earth and Planetary Science Letters, v. 69, p. 88-106.

Dick, H.J.B., Meyer, P., Bloomer, S., Kirby, S., Stakes, D. and Mawer, C., 1991, Lithostratigraphic evolution of an in situ section of oceanic layer 3 *in* Von Herzen, R.P., Robinson, P.T., *et al.*, eds., Proceedings: Ocean Drilling Program, Scientific Results, v. 118, p. 439-515.

Dick, H.J.B. and Natland, J.H., 1996, Late-stage melt evolution and transport in the shallow mantle beneath the East Pacific Rise, *in* Mevel, C., Gillis, K.M., Allan, J.F. and Meyer, P.S., eds., Proceedings: Ocean Drilling Program, Scientific Results, v. 147, p. 103-134.

Dick, H.J.B., Robinson, P.T. and Meyer, P.S., 1992, The plutonic foundation of a slow-spreading ridge, *in* Duncan, R.A., *et al.*, eds., Synthesis of Results from Scientific Drilling the Indian Ocean: American Geophysical Union, Geophysical Monograph 70, p. 1-39.

Dunsworth, S., Calon, T. and Malpas, J., 1986, Structural and magmatic controls on the internal geometry of the plutonic complex and its chromite occurrences in the Bay of Islands ophiolite, Newfoundland: Current Research: Geological Survey of Canada, 86-1A, p. 471-482.

Edwards, S.J. and Malpas, J., 1995, Multiple origins for mantle harzburgites: examples from the Bay of Islands ophiolite, Nfld: Canadian Journal of Earth Sciences, v. 32, p. 1046-1057.

Elthon, D., 1992, Chemical trends in abyssal peridotites: refertilisation of depleted suboceanic mantle: Journal of Geophysical Research, v. 97, p. 9015-9025.

Elthon, D. and Scarfe, C.M., 1980, High pressure phase equilibria of a high-magnesium basalt: implications for the origin of mid-ocean ridge basalts: Carnegie Institute of Washington Yearbook, p. 277-281.

Engel, A.E.J. and Engel, C.G., 1964, Igneous rocks of the East Pacific rise: Science, v. 146, p. 477-485.

Engel, C.G. and Engel, A.E.J., 1963, Basalts dredged from the northeastern Pacific Ocean: Science, v. 140, p. 1321-1324.

Flower, M.J.F., 1980, Thermal and kinematic control on ocean ridge magma fractionation: contrasts between Atlantic and Pacific spreading axes:Geological Society London, Journal, v. 138, p. 695-712.

Flower, M.F.J. and Robinson, P.T., 1981a, Basement drilling in the western Atlantic Ocean 1. Magma fractionation and its relation to eruptive chronology: Journal of Geophysical Research, v. 86, p. 6273-6298.

Flower, M.F.J. and Robinson, P.T., 1981b, Basement drilling in the western Atlantic Ocean 2. A synthesis of construction processes at the Cretaceous ridge axis: Journal of Geophysical Research, v. 86, p. 6299-6309.

Forsyth, D.W., 1997, Melting and mantle flow beneath mid-ocean ridges; constraints from the seismological component of the MELT experiment: EOS, v. 78, p. 17, 325.

Garmany, J., 1989, Accumulations of melt at the base of young oceanic crust: Nature, v. 340, p. 628-632.

Gillis, K.M., Mevel, C., Allan, J. *et al.*, 1993, Proceedings Ocean Drilling Program, Initial Reports, 147.

Gillis, K.M and Roberts, M., 1999, Cracking at the magma-hydrothermal transition: evidence from the Troodos ophiolite: Earth and Planetary Science Letters, v. 169, p. 227-244.

Green, D.H. and Ringwood, A.E., 1967, The genesis of basaltic magmas: Contributions to Mineralogy and Petrology, v. 15, p. 103-190.

Hale, L.D., Morton, C.J. and Sleep, N.H., 1982, Reinterpretation of seismic reflection data over the East Pacific Rise: Journal of Geophysical Research, v. 87, p. 7707-7717.

Harding, A.J., Orcutt, J., Kappus, M., Vera, E., Mutter, J., Buhl, P., Detrick, R. and Brocher, T., 1989, The structure of young oceanic crust at 13°N on the East Pacific Rise from expanding spread profiles: Journal of Geophysical Research, v. 94, p. 12,163-12,196.

Hawkins, J., Lonsdale, P.F., Macdougall, J.D. and Volpe, A.M., 1990, Petrology of the axial ridge of the Mariana Trough backarc spreading centre: Earth and Planetary Science Letters, v. 100, p. 226-250.

Henstock, T.J., Woods, A.W. and White, R.S., 1993, the accretion of oceanic crust by episodic sill intrusion: Journal of Geophysical Research, v. 98, p. 4143-4161.

Herron, T.J., Ludwig, W.J., Stoffa, P.L., Kan, T.K. and Buhl, P., 1978, Structure of the East Pacific Rise from multichannel seismic reflection data: Journal of Geophysical Research, v. 83, p. 798-804.

Hirth, G. and Kohlstedt, D.L., 1996, Water in the oceanic mantle: implications for rheology, melt extraction and the evolution of the lithosphere: Earth and Planetary Science Letters, v. 144, p. 93-108.

Iwamori, H., 1994, $^{238}U/^{230}Th/^{226}Ra$ and $^{235}U/^{231}Pa$ disequilibria produced by mantle melting with porous and channel flows: Earth and Planetary Science Letters, v. 125, p. 1-16.

Johnson, K.T.M. and Dick, H.J.B., 1992, Open system melting and temporal and spatial variation of basalt and peridotite at the Atlantis II fracture zone: Journal of Geophysical Research, v. 97, p. 9219-9241.

Johnson, K.T.M., Dick, H.J.B. and Shimizu, N., 1990, Melting in the oceanic upper mantle: an ion microprobe study of diopsides in abyssal peridotites: Journal of Geophysical Research, v. 95, p. 2661-2678.

Kelemen, P.B., Dick, H.J.B. and Quick, J.E., 1992, Formation of harzburgite by pervasive melt/rock reaction in the upper mantle: Nature, v. 358, p. 635-641.

Kelemen, P.B., Hirth, G., Shimizu, N., Spiegelman, M. and Dick, H.J.B., 1997, A review of melt migration processes in the adiabatically upwelling mantle beneath oceanic spreading ridges: Royal Society London, Philosophical Transactions, A, v. 355, p. 283-318.

Kent, G.M., Harding, A.J. and Orcutt, J.A., 1990, Evidence for a smaller magma chamber beneath the East Pacific Rise at 9°30'N: Nature, v. 344, p. 650-653.

Klein, E.M. and Langmuir, C.H., 1987, Global correlations of ocean ridge basalt geochemistry with axial depth and crustal thickness: Journal of Geophysical Research, v. 92, p. 8089-8115.

Korenaga, J. and Kelemen, P.B., 1997, Origin of gabbro sills in the Moho transition zone of the Oman ophiolite: Implications for magma transport in the oceanic crust: Journal of Geophysical Research, v. 102, p. 27729-27749.

Kuznir, N.J., 1980, Thermal evolution of the oceanic crust: its dependence on spreading rate and effect on crustal structure: Geophysical Journal Royal Astronomical Society, v. 61, p. 167-181.

Langmuir, C.H., 1989, Geochemical consequence of in situ crystallisation: Nature, v. 340, p. 199-205.

Langmuir, C.H., Klein, E.M. and Plank, T., 1992, Petrological systematics of mid-ocean ridge basalts: constraints on melt generation beneath ocean ridges: Geophysical Monographs, v. 71, p. 183-281.

Leblanc, M. and Ceuleneer, G., 1992, Chromite crystallisation in a multicellular magma flow: evidence from a chromitite dike in the Oman ophiolite: Lithos, v. 27, p. 231-257.

Lin, J., Purdy, G.M., Schouten, H., Sempere, J-C. and Zervas, C., 1990, Evidence for focussed magmatic accretion along the Mid-Atlantic Ridge: Nature, v. 344, p. 627-632.

Lundstrom, C.C., Gill, J., Williams, Q. and Perfit, M.R., 1995, Mantle melting and basalt extraction by equilibrium porous flow: Science, v. 270, p. 1958-1961.

Maaloe, S., 1981, Magma accumulation in the ascending mantle: Geological Society London, Journal, v. 138, p. 223-236.

Maaloe, S. and Scheie, A., 1982, The permeability controlled accumulation of primary magma: Contributions to Mineralogy and Petrology, v. 81, p. 350-357.

Malpas, J., 1987, The Bay of Islands ophiolite. A cross section through Paleozoic ocean crust and mantle in Western Newfoundland: Geological Society of America, DNAG field guides, p. 451-456.

Malpas, J., 1990, Accretionary processes in the Troodos ophiolite, Cyprus: evidence from field mapping and deep crustal drilling *in* Malpas, J., Moores, E., Panayiotou, A. and Xenophontos, C., eds., Ophiolites, oceanic crustal analogues: Geological Survey of Cyprus, p. 65-74.

Malpas, J., Case, G. and Moore, P., 1989, Geology of the area immediately surrounding the CY-4 borehole, Cyprus Crustal Study Project *in* CCSP Initial Report, Hole CY-4: Geological Survey of Canada, Paper 88-9, p. 31-37.

McClain, J.S., Orcutt, J.A. and Burnett, M., 1985, The East Pacific Rise in cross section; a seismic model: Journal of Geophysical Research, v. 90, p. 8627-8640.

Moores, E.M. and Vine, F.J., 1971, The Troodos massif, Cyprus and other ophiolites as oceanic crust: evaluation and implications: Royal Society London, Philosophical Transactions, A, v. 268, p. 443-466.

Natland, J.H. and Dick, H.J.B., 1996, Melt migration through high-level gabbroic cumulates of the East Pacific Rise at Hess Deep: The origin of magma lenses and the deep crustal structure of fast-spreading ridges *in* Mevel, C., Gillis, K.M., Allan, J. F. and Meyer, P.S., eds., Proceedings Ocean Drilling Program, Scientific Results, 147, p. 21-58.

Nicolas, A., 1986, A melt extraction model based on structural studies in mantle peridotites: Journal of Petrology, v. 27, p. 999-1022.

Nicolas, A., 1989, Structures of Ophiolites and Dynamics of Ocean Lithosphere: Kluwer Academic Publishers, Dortrecht, 367 p.

Nicolas, A., 1990, Melt extraction from mantle peridotites: hydrofracturing and porous flow, with consequences for oceanic ridge activity *in* Ryan, M.P., ed., Magma Transport and Storage: Wiley, New York, p. 159-174.

Nicolas, A., Reuber, I. and Benn, K., 1988, A new magma chamber model based on structural studies in the Oman ophiolite: Tectonophysics, v. 151, p. 87-105.

Nielsen, R.L., Crum, J, Bourgeois, R., Hascall, K., Forsythe, L.M., Fisk, M.R., and Christie, D.M., 1995, Melt inclusions in high-An plagioclase from the Gorda Ridge: an example of the local diversity of MORB parental magmas: Contributions to Mineralogy and Petrology, v. 122, p. 34-50.

Nisbet, E.G. and Fowler, C.M.R., 1978, The Mid-Atlantic ridge at 37° and 45° N: some geophysical and petrological constraints: Geophysical Journal Royal Astronomical Society, v. 54, p. 631-660.

Niu, Y., 1997, Mantle melting and melt extraction processes beneath ocean ridges: evidence from abyssal peridotites: Journal of Petrology, v. 38, p. 1047-1074.

Niu, Y. and Hekinian, R., 1997, Basaltic liquids and harzburgitic residues in the Garrettt Transform: a case study at fast spreading ridges: Earth and Planetary Science Letters, v. 146, p. 243-258.

O'Hara, M.J., 1965, Primary magmas and the origin of basalts, Scotland: Journal of Geology, v. 1, p.19-40.

Pallister, J.S. and Hopson, C.A., 1981, Semail ophiolite plutonic suite: field relations, phase variation, cryptic variation and layering, and a model for the spreading ridge magma chamber: Journal of Geophysical Research, v. 86, p. 2593-2644.

Pedersen, R.B. and Malpas, J., 1984, The origin of oceanic plagiogranites from the Karmoy ophiolite, Western Norway: Contributions to Mineralogy and Petrology, v. 88, p. 36-52.

Pedersen, R.B., Malpas, J., and Fallon, T., 1996, Petrology and geochemistry of gabbroic and related rocks from Site 894, Hess Deep, *in* Mevel, C., Gillis, K.M., Allan, J.F. and Meyer, P.S., eds., Proceedings: Ocean Drilling Program, Scientific Results, 147, p. 3-19.

Perfit, M.R., Fornari, D.J., Smith, M.C., Bender, J.F., Langmuir, C.H., and Haymon, R.M., 1994, Small scale spatial and temporal variations in mid-ocean ridge crest magmatic processes: Geology, v. 22, p. 375-379.

Pettigrew, T.L., Casey, J.F., Miller, D.J. *et al.*, 1999, Proceedings: Ocean Drilling Program, Scientific Results, 147.

Phipps Morgan, J., 1987, Melt segregation beneath mid ocean spreading centres: Geophysical Research Letters, v. 14, p. 1238-1241.

Phipps Morgan, J. and Chen, Y. J., 1993, The genesis of ocean crust: magma injection, hydrothermal circulation, and crustal flow: Journal of Geophysical Research, v. 98, p. 6283-6297.

Quick, J.E. and Denlinger, R.P., 1993, Ductile deformation and the origin of layered gabbros in ophiolites: Journal of Geophysical Research, v. 98, p. 14015-14027.
Rhodes, J.M., Dungan, M.A., Blanchard, D.P. and Long, P.E., 1979, Magma mixing at mid ocean ridges: evidence from basalts drilled near 22°N on the Mid-Atlantic Ridge: Tectonophysics, v. 55, p. 35-61.
Ringwood, A.E., 1962, A model for the Upper Mantle; Pt 1: Journal of Geophysical Research, v. 67, p. 857-867.
Robinson, P.T. and Malpas, J., 1990, The Troodos Ophiolite of Cyprus: new perspectives on its origin and emplacement *in* Malpas, J., Moores, E., Panayiotou, A. and Xenophontos, C., eds., Ophiolites, oceanic crustal analogues: Geological Survey of Cyprus, p. 13-26.
Robinson, P.T. and Von Herzen, R., *et al.*, 1989, Proceedings: Ocean Drilling Program, Initial Reports, 118, Part A, 826 p.
Robson, D. and Cann, J.R., 1982, A geochemical model of mid ocean ridge magma chambers: Earth and Planetary Science Letters, v. 60, p. 93-104.
Rohr, K.B., Mildreit, B. and Yorath, C.J., 1988, Asymmetric deep crustal structure across the Juan de Fuca Ridge: Geology, v. 16, p. 533-537.
Salters, V.J.M. and Hart, S.R., 1989, The Hafnium paradox and the role of garnet in the source of mid-ocean ridge basalts: Nature, v. 342, p. 420-422.
Scheirer, D.S., Forsyth, D.W., Cormier, M-H., and Macdonald, K.C., 1998, Shipboard Geophysical Indications of Asymmetry and Melt Production Beneath the East Pacific Rise Near the MELT Experiment: Science, v. 280, p. 1221-1224
Schiffman, P., Smith, B.M., Varga, R.J. and Moores, E.M., 1987, Geometry, conditions and timing of off-axis hydrothermal metamorphism and ore deposition in the Solea Graben: Nature, v. 325, p. 423-425.
Schouten, H., Klitgord, K.D. and Whitehead, J.A., 1985, Segmentation of mid-ocean ridges: Nature, v. 317, p. 225-229.
Shen, Y. and Forsyth, D.W., 1995, Geochemical constraints on initial and final depths of melting beneath mid-ocean ridges: Journal of Geophysical Research, v. 100, p. 2211-2237.
Shimizu, N., 1998, The geochemistry of olivine-hosted melt inclusions in FAMOUS basalt ALV519-4-1: Physics of Earth Planetary Interiors, v. 107, p. 183-201.
Shipboard Scientific Party, 1999, Leg 176 Summary *in* Dick, H.J.B., Natland, J.H., Miller, D.J., *et al.*, Proceedings: Ocean Drilling Program, Initial Reports, v. 176, p. 1-70.
Sinha, M.C., Navin, D.A., MacGregor, L.M., Constable, S., Peirce, C., White, A., Heinson, G. and Inglis, M.A., 1999, Evidence for accumulated melt beneath the slow-spreading Mid-Atlantic Ridge *in* Cann. J.R., Elderfield, H. and Laughton, A., eds., Mid-Ocean Ridges: Cambridge University Press, p. 17-38.
Sinton, J.M. and Detrick, R.S., 1992, Mid-ocean ridge magma chambers: Journal of Geophysical Research, v. 97, p. 197-216.
Sinton, J.M., Smaglik, S.M., Mahoney, J.J. and Macdonald, K.C., 1991, Magmatic processes at superfast spreading ocean ridges: glass compositional variations along the East Pacific Rise, 13° – 23°S: Journal of Geophysical Research, v. 96, p. 6133-6155.
Sleep, N.H., 1975, Formation of oceanic crust: some thermal constraints: Journal of Geophysical Research, v. 80, p. 4037-4042.
Smewing, J.D., 1981, Mixing characteristics and compositional differences in mantle-derived melts beneath spreading axes: evidence from cyclically layered rocks in the ophiolite of North Oman: Journal of Geophysical Research, v. 86, p. 2645-2660.
Sobolev, A.V. and Shimizu, N., 1993, Ultra-depleted primary melt inclusions in an olivine from the Mid-Atlantic Ridge: Nature, v. 363, p. 151-154.
Sobolev, A.V. and Shimizu, N., 1994, The origin of typical N-MORB: the evidence from a melt inclusion study: Mineralogical Magazine, v. A58, p. 862-863.
Sparks, D.W. and Parmentier, E.M., 1991, Melt extraction from the mantle beneath mid-ocean ridges: Earth and Planetary Science Letters, v. 105, p. 368-377.
Spiegelman, M., Aharonov, E. and Kelemen, P. B., 1996, The compaction reaction: magma channel formation by reactive flow in deformable media: EOS, v. 77, p. F783.
Spiegelman, M. and McKenzie, D., 1987, Simple 2D models for melt extraction at mid-ocean ridges and island arcs: Earth and Planetary Science Letters, v. 83, p. 137-152.
Spooner, E.T.C., 1980, Cu-mineralisation and seawater convection in ocean crust – the ophiolite ore deposits of Cyprus: Geological Society of America, Special Paper 20, p. 665-704.
Strong, D.F. and Malpas, J., 1975, The sheeted dyke layer of the Bett's Cove ophiolite does not represent spreading: further discussion: Canadian Journal of Earth Sciences, v. 12, p. 894-896.
Thompson, G., Bryan, W.B. and Humphris, S.E., 1989, Axial volcanism on the East Pacific Rise, 10° -12°N: Geological Society London, Special Publication 42, p. 181-200.
Toomey, D.R., Solomon, S.C., Wilcock, W.S. D., Blackman, D.K., Orcutt, J.A. and Hammond, W.C., 1996, Upper mantle structure beneath the MELT region of the East Pacific Rise from initial observations of P and S delay times: EOS, v. 77, p. 652.
Toomey, D.R., Purdy, G.M., Solomon, S. and Wilcox, W., 1990, The three dimensional seismic velocity structure of the East Pacific Rise near latitude 9°30'N: Nature, v. 347, p. 639-644.
Toomey, D.R., Wilcock, W.S., Solomon, S.C., Hammond, W.C., Orcutt, J.A., 1998, Mantle Seismic Structure Beneath the MELT Region of the East Pacific Rise from P and S Wave Tomography: Science, v. 280, p. 1224-1227
Tucholke, B.E., Lin, J., Kleinrock, M.C., Tivey, M.A., Reed, T.B., Goff, J. and Jaroslow, G.E., 1997, Segmentation and crustal structure of the western Mid-Atlantic Ridge flank, 25° 25'-27'N and 0-29 m.y.: Journal of Geophysical Research, v. 102, p. 10203-10223.
Turcotte, D.L. and Phipps Morgan, J., 1992, The physics of magma migration and mantle flow beneath a mid-ocean ridge: Geophysical Monograph 71, p.155-182.
Vera, E.E., Buhl, P., Mutter, J.C., Harding, A.J., Orcutt, J.A. and Detrick, R.S., 1990, The structure of 0-0.2 My old oceanic crust at 9°N in the East Pacific Rise from expanded spread profiles: Journal of Geophysical Research, v. 95, p. 15,529-15,556.
Walker, D., Shibata, T. and DeLong, S.E., 1979, Abyssal tholeiites from the Oceanographer Fracture Zone, II. Phase equilibria and mixing: Contributions to Mineralogy and Petrology, v. 70, p. 111-125.
Zhou, M-F., Robinson, P.T., Malpas, J. and Zijin, L., 1996, Podiform chromitites in the Luobusa ophiolite (Southern Tibet): Implications for melt-rock interaction and chromite segregation in the upper mantle: Journal of Petrology, v. 37, p. 3-21.

Accepted as revised 15 July 2000