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Development of ophiolitic perspectives on models of oceanic magma chambers beneath active spreading centers

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ABSTRACT

Ophiolites played a minor role in the formulation of the ideas of the history and evolution of the ocean floor because initially they were not interpreted as oceanic in origin. Early researchers working on ophiolites were late in claiming their oceanic origin. Ophiolites were universally considered to be large sill intrusions or thick lava flows emplaced from an abyssal substratum into subsiding, marginal sedimentary basins. Such large pools of basaltic magma were proposed to have differentiated internally as a result of crystallization within an outer veneer of a giant lava pillow. It was first with the emergence of the theory of plate tectonics that it became possible to distinguish between mantle rocks and crustal intrusive plutons. This development led to the recognition of fossil magma chambers in ophiolites. Plate tectonic models implied that basaltic magma either erupted directly on the seafloor or accumulated in crustal chambers beneath spreading centers. The first models developed in the late 1960s for the structure of modern oceanic crust did not include lower crustal magma chambers; however, subsequent models formulated in the early 1970s incorporated the existence of magma chambers, but with little reference to ophiolite studies. Oceanic magma chambers were seen as large, continuously growing entities keeping pace with seafloor spreading and steady-state magma supply. This magma chamber interpretation was very influential and contained the key elements soon adapted by many of the subsequent studies of ophiolites. Such ophiolitic magma chamber models were extended to the oceans in the 1980s, inferring the size, shape, and evolution of magma chambers beneath active spreading centers.

The challenge to the steady-state magma chamber model came from geophysical and theoretical modeling of modern oceanic crust. These studies revealed that large magma chambers could not exist beneath slow-spreading centers, and that they might be present beneath fast-spreading centers only as a thin magma lens on top of a large crystal mush. Other predictions suggested the presence of small magma chambers in broad zones under the axial valley and emphasized the intermittent nature of crustal chambers as a result of the interplay between tectonic and magmatic processes. This modeling of oceanic crust made little use of observations from ophiolites. The general image that emerged in the 1990s was that magma chambers beneath fast-spreading centers were largely filled with crystal mush containing only small percentages of melt. The large molten chambers envisioned during the proceeding three decades have thus vanished and been replaced by crystal mush chambers. The most important implication

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of this interpretation for the origin of ophiolites is that layering commonly observed in gabbros is likely to have formed as a result of crustal strain, not from direct deposition or crystallization on chamber walls.

Deep drilling during the last ten years into in situ modern oceanic crust near active spreading centers has supported the existence of crystal mush zones. A surprising discovery has been the preservation of evolved gabbros within primitive host gabbro sequences. The evolved gabbros were interpreted to represent the interstitial melt that migrated upward and laterally as a result of compaction and/or tectonic deformation. The interaction between new magma seeping up from beneath the chamber and the lower strata of the mush makes the mush zone grow. Detailed field studies of ophiolites in recent years have demonstrated that new magma may form sill complexes in crystal mushes. The mush and sill complexes act as a reactive filter, modifying the migrating residual melt and controlling the composition of the residual melt, which eventually reaches the magma lens or gets erupted as lavas on the seafloor.

Our understanding of oceanic magma chambers has been much influenced by direct observations and careful fieldwork in plutonic complexes of ophiolites. The revival of systematic field and petrographic studies in ophiolites is a necessary approach to further advance our understanding of oceanic magma chambers and the magmatic and tectonic processes associated with the evolution of oceanic crust.

Keywords: ideas of magma chambers, conceptual models, ophiolite studies, history of ocean floor, plate tectonics, crustal accretion, ophiolites, oceanic crust, active spreading centers, , abyssal substratum, magma differentiation, steady-state magma chambers, crystal-mush magma chambers, melt lens, slow-spreading magma chambers, fast-spreading magma chambers, sill complexes, magmatic processes, revival of field and petrographic studies

INTRODUCTION

Development of the plate tectonic theory and the early models for the accretion of oceanic crust predicted the presence of magma chambers beneath the rift axes of mid-ocean ridge systems (Cann, 1970; Moore et al., 1974; Christensen and Salisbury, 1975; Dewey and Kidd, 1977; Bryan and Moore, 1977; Macdonald, 1989). Because magma chambers cannot directly be observed (Macdonald, 1982; Orcutt, 1987), ophiolite studies have become one of our principal sources of information about magma chambers and magma chamber processes. Fossil magma chambers have been identified in many ophiolite complexes (Gass, 1968) and ophiolitic plutons have been, and still are, central to our understanding of crustal accretion and magmatic processes in active oceanic spreading settings, although the ideas and conceptual models for magma chamber evolution and associated magmatic processes have varied considerably over the years since the structural views of ophiolitic and oceanic crust formation first emerged (Fig. 1). Studies of continental layered intrusions (Wager and Brown, 1967) have been another important source for understanding both ophiolitic and oceanic magma chambers and their solidification processes.

In this paper, we explore how ideas of ophiolitic magma chambers have developed through time and have been influenced by our geochemical, petrological, tectonic, and geophysical understanding of modern oceanic crust. We show that early models of ophiolitic magma chambers were influenced by studies of layered intrusions and other stratiform continental plutons. Development of the theory of plate tectonics resulted in a widely accepted idea of continually replenished magma chambers, an idea that, for years, strongly influenced ophiolite investigations. Studies of core samples drilled from gabbros near active oceanic spreading centers and their results have provided the critical information that stimulated a re-examination of oceanic magma chamber models. Thus, current models of oceanic magma chambers are fundamentally different than the early models, reflecting major changes and shifts in our geological thought.

After a brief introduction to oceanic magma chambers, we review the development of ideas on ophiolitic magma chambers, using observations from the Troodos (Cyprus), Bay of Islands (Newfoundland, Canada), and Oman (Sultanate of Oman) ophiolites. Next, we discuss the ophiolitic chamber models in the context of ideas developed concurrently for magma chambers beneath modern oceanic spreading centers. We then assess the mutual influence of ophiolitic and oceanic studies on the development of ideas for magma chamber evolution at active spreading centers. In our discussion, we use some of the most important original illustrations of ophiolitic and oceanic magma chambers (only slightly edited or redrawn for the purpose of our presentation).

MAGMA CHAMBERS

Magma forms by partial melting in adiabatically upwelling mantle (e.g., McKenzie and Bickle, 1988); it then migrates upward and concentrates beneath ridge crests (Morgan, 1987).

Figure 1. Modern artist's drawing of continental rifting and magmatism as inspired by eighteenth century illustrations. This illustration was used in connection with a conference, convened by M.A. Menzies at Royal Holloway, University of London in 1987, on *Oceanic and Continental Lithosphere: Similarities and Differences*. The conference resulted in a special volume of Journal of Petrology (1998). Reproduced with the permission of Martin Menzies.

The differential buoyancy between melt and the residual mantle peridotite, or the host crustal rocks, drives the upward movement to a level of neutral buoyancy (Turcotte and Morgan, 1992; Bryan, 1993). The melt is thought to migrate by channeled porous flow along grain boundaries and along fractures ranging in width from meters to kilometers (see review by Turcotte and Morgan, 1992). Magma migration may also be controlled by hydraulic overpressure and dike formation above a plume head (e.g., Maaløe, 2002). The melt differentiates in transient dikes and sills or in chambers of various shapes and sizes as a result of gradual cooling, and leaves behind solid crystallization and reaction products that may significantly modify host crustal and mantle properties, as well as the composition of the melt. The storage and the flux of magma through oceanic crust depend on the spreading rate (Dilek et al., 1998; Karson, 1998, 2002). The magma that migrates through the crustal "plumbing system" can change composition to an extent that the derived magmas may have little resemblance to the parental magma that originally separated from the host peridotite (Grove et al., 1992). Magma trapped in high-level plutons or erupted as lava flows on the seafloor therefore may not easily be "fingerprinted" to infer the source of melting and the mode of differentiation processes (Langmuir et al., 1992; Grove et al., 1992; Bédard, 1993).

It was mainly petrographic studies of gabbroic and ultramafic intercontinental-layered plutons (McBirney, 1993; Marsh, 2000) that shaped our early views and resulted in widely publicized models for spreading-related magma chambers (e.g., Bryan and Moore, 1977; Brown and Musset, 1981; Best, 1982; Bott, 1982; Fig. 2). The early ophiolitic magma chamber models were strongly influenced by studies of layered intrusions and other continental plutons, which were believed to represent trapped magma chambers solidified at crustal levels (Hopson and Frano, 1977; Pallister and Hopson, 1981; Nicolas, 1989, p. 261; McBirney, 1993; Marsh, 2000). The internal constitution of plutons records evidence for crystallization and consolidation processes (Marsh, 1996) that can only be unraveled by careful field mapping and detailed petrographic and geochemical studies (Wager and Brown, 1967; Turner and Campbell, 1986; Parsons, 1987; Cawthorn, 1996; Irvine et al., 1998). Such studies of stratigraphically well constrained sequences may not always be feasible at active spreading centers, where complex thermal and tectonic processes may have shaped or modified the internal constitution of plutons (Sleep, 1975; Lister, 1977) and where perhaps only a few dredges and drill core are available.

Development of the plate tectonic theory and the direct sampling of lava flows near active ridges have revealed systematic variations in lava compositions that are interpreted as related to the bathymetry of the ridge crest (Klein and Langmuir, 1987). Temporal and lateral ridge segmentations are likely to be controlled by the lack and/or the presence of the underlying axial magma chambers, magma supply rates, and the depth to the crust-mantle transition zone (Macdonald, 1982; Klein and Langmuir, 1987; Solomon and Toomey, 1992; Batiza, 1996). Such views, together with theoretical and geophysical studies, have markedly changed the classic view of oceanic magma chambers and the processes that control magma transport and differentiation. Most recently, major advances have resulted from direct sampling by drilling into in situ gabbroic complexes exposed near active oceanic spreading centers. The most complete section of oceanic gabbros has been sampled in Ocean Drilling Program Holes 735B and 1105A, near the very slow-spreading Southwest Indian Ridge (Dick et al., 2000; Natland and Dick, 2001; Thy, 2002). These results have highlighted the important role of spreading rate on magma chamber development and the interplay between the tectonic extension and magma chamber processes on the evolution of oceanic crust.

Despite the advances made in our understanding of the evolution of the plutonic foundation of modern oceanic crust, it is still important to realize that direct information on the nature and sizes of magma chambers and on the processes that control magma crystallization and differentiation in them is limited. Because magma chambers and magma solidification processes cannot be

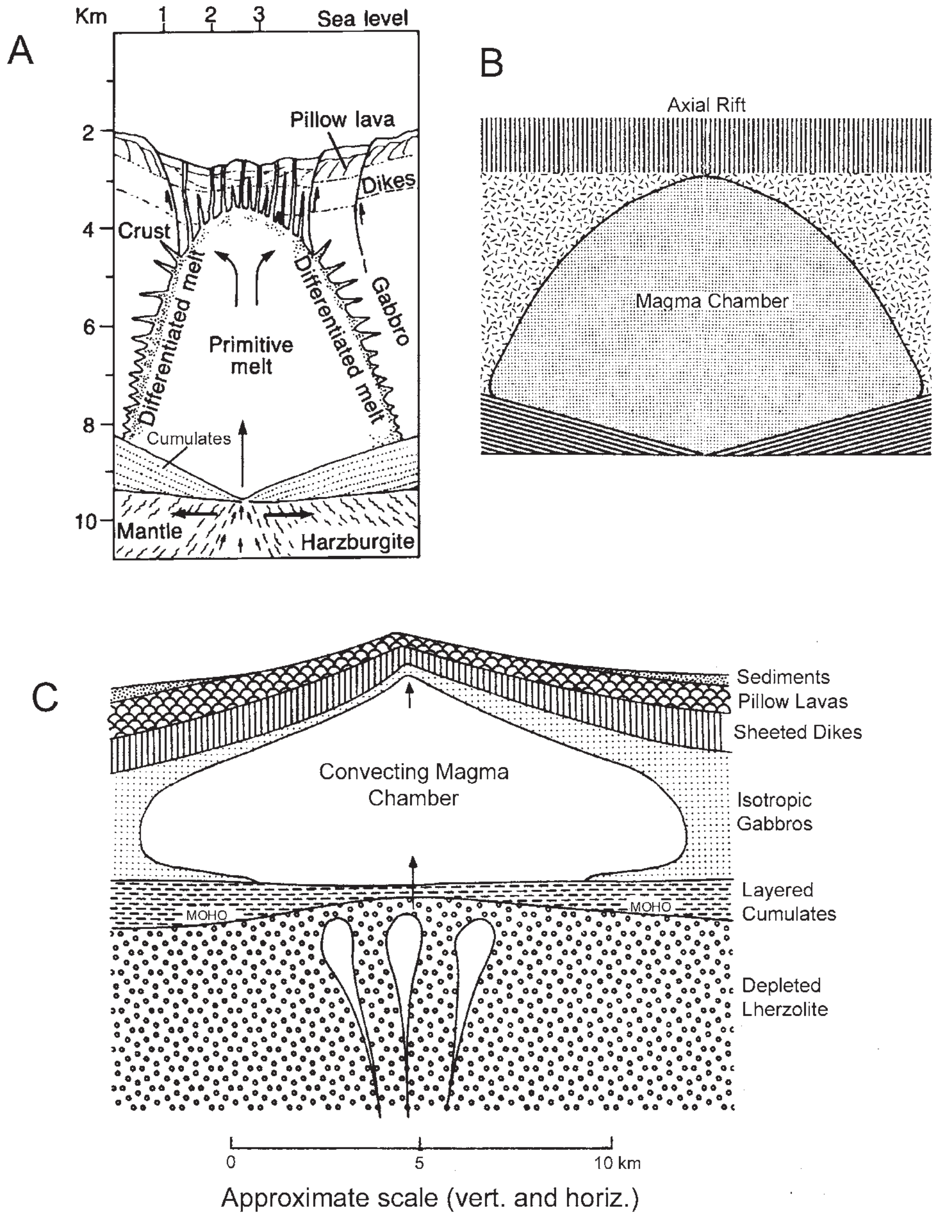

Figure 2. Schematic illustrations of oceanic spreading chambers. A: Spreading magma chamber after Bryan and Moore (1977, Fig. 16). The model is based on the petrology and geochemistry of dredge samples and was drawn to scale. B: Schematic spreading center magma chamber redrawn by Pallister and Hopson (1981, Fig. 17) after Cann (1974, Fig. 6). Note the similarity with Bryan and Moore's (1977) chamber model in Figure 2A. Reproduced by permission of Royal Astronomical Society. C: Textbook cross-section of a mid-ocean spreading center magma chamber (Brown and Mussett, 1981, Fig. 7.8).

observed directly, the formulated models often combine limited observations with current theoretical views of crustal accretion and mid-ocean ridge processes. As a result, magma chamber models naturally change as our conceptual ideas of ocean crust formation and magma crystallization processes evolve.

OPHIOLITE STUDIES

The term "ophiolite" was originally used to describe serpentinites in orogenic belts (cf., Coleman, 1977). After Steinmann (1927), the term ophiolite was used to identify a special association of rocks that included serpentinite, gabbro, diabase, spilite, and chert (see Bernoulli et al., this volume, Chapter 7). Ophiolites were believed to have formed as massive mafic and ultramafic submarine outpourings of magma onto the floor of eugeosynclinal basins during early stages of their development (Kay, 1951; Brunn, 1960; Aubouin, 1965). Differentiation in such thick outpourings was thought to have resulted in a variation of rocks ranging from peridotite to gabbro within an outer selvage of diabase and pillow lavas (like a gigantic basalt pillow; cf., Aubouin, 1965, p. 148–159). Ophiolitic peridotites were considered to have formed by accumulation of early crystallizing minerals, such as olivine and pyroxene, in the lower part of these in situ crystallizing massifs.

Although it was well known that the Earth's mantle was of peridotitic composition (MacGregor, 1967) and that melting of this peridotite would produce typical oceanic basalts (Bowen, 1928; Daly, 1933; Hess, 1955; Yoder and Tilley, 1962; Wyllie, 1967), the connection between these observations and the basaltic oceanic layer was not always made (Hess, 1955). Hess (1962) hypothesized that the oceanic layer formed by serpentinization of peridotite beneath ridge crests located above the rising limbs of mantle convection cells. This view was so influential that it actually might have held back the development of modern views of ocean ridges and the generation of basaltic crust (cf., Heezen, 1962). For this reason, ophiolite studies initially struggled with the origin of basaltic lavas and their relations with tectonized peridotites and serpentinites (Gass and Masson-Smith, 1963; Gass, 1967). It was first with the emergence of plate tectonics in the mid- to late 1960s that it became possible to distinguish between mantle tectonites and intrusive plutonic rocks as separate components of oceanic lithosphere. This development further made it clear that ophiolites were mostly basaltic in composition and that the internal variation in their pseudostratigraphy was caused by fractional crystallization forming solid cumulates and differentiated melts (Brunn, 1960; Dewey and Bird, 1971; Moores and Vine, 1971; Church, 1972). Thus, it became possible to identify fossil magma chambers in ophiolitic sequences. This was achieved convincingly, for the first time, in the Troodos ophiolite (e.g., Gass, 1968).

Troodos Ophiolite (Cyprus)

The Troodos ophiolite is part of a mountain range extending east-west across the island of Cyprus, and is composed of serpentinites, peridotites, mafic and ultramafic plutons, sheeted dikes, and lavas (Fig. 3). The general geology and the historical background of studies in the Troodos ophiolite can be found in several review papers (Gass, 1980, 1989, 1990; Gass and Smewing, 1981; Robertson and Xenophontos, 1993). Although the first geological maps of the Troodos Mountains were published in 1906, these early maps and accompanying reports contained little information on the plutonic complex. It was first with the systematic mapping initiated in the early 1950s by the Cyprus Geological Survey (Bishopp, 1952, 1954) that a detailed stratigraphy of the massif was achieved (e.g., Wilson, 1957; Bear, 1960; Gass, 1960). The result of this mapping project guided the development of the early models for the origin of the ophiolite (Gass and Masson-Smith, 1963; Gass, 1967) and laid the foundation for our current understanding of the Troodos massif.

The early models for the origin of the Troodos ophiolite were developed in the context of the geosynclinal theory, which dominated the geodynamic thinking of that time (i.e., Aubouin, 1965). The contact between the volcanic pile and the peridotitic rocks and the apparent lack of sialic crust beneath the ophiolite led to the

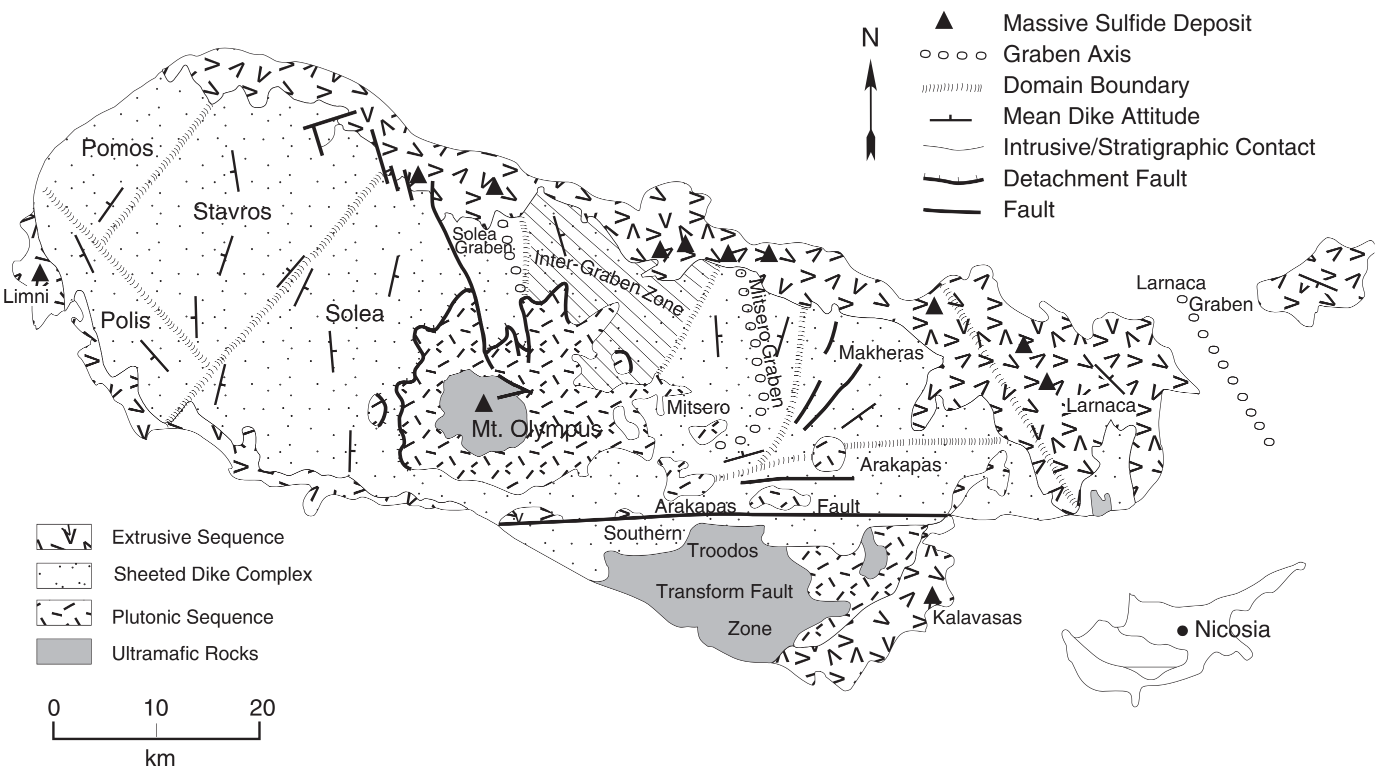

Figure 3. Simplified geological and structural map of the Troodos massif from Eddy et al. (1998, Fig. 1). Shown are the main lithological and tectonic features and orientations of structural domains within the sheeted dike complex. Locations of the structural grabens and graben axes are after Moores et al. (1990). Reproduced by permission of Elsevier.

suggestion that the volcanic outpouring occurred in a subaqueous oceanic setting (Gass and Masson-Smith, 1963) between the continents of Africa and Eurasia (Fig. 4A). The plutonic complex was considered to have formed by internal differentiation in a large, laccolith-shaped outpour (Fig. 4B) of submarine lava in a subsiding marginal sedimentary basin. The model suggested that this volcanic pile and a slice of the upper mantle were uplifted and thrust over the Eurasian continent during the Alpine orogeny.

Gass and Masson-Smith (1963) deviated from the pure eugeosynclinal view by suggesting that the Troodos complex represented a partly or completely fused upper mantle fragment, and that this resultant melt pocket differentiated into a stratiform complex composed of dunites, peridotites, gabbros, and granophyres. This view might have been influenced by a contemporaneous paper of Hess (1962) that dealt with the relationship between mantle convection cells and the mid-ocean ridge systems. It might also have been influenced by Heezen's (1962) paper on the evolution of the deep-sea floor. Both authors had assumed an oceanic layer composed of serpentinites. Gass (1967) elaborated on the idea of peridotite melting, suggesting that the Troodos plutonic rocks belonged to a "differentiated ultrabasic mass of batholithic dimensions" (p. 126). Gass (1967), thus, subscribed to the widely held view that ultramafic melts were primary melts (Hess, 1938). Gass went further by pointing out that the "structure, particularly of the sheeted intrusive complex, is of the type that could be formed on the crest of a mid-ocean ridge"

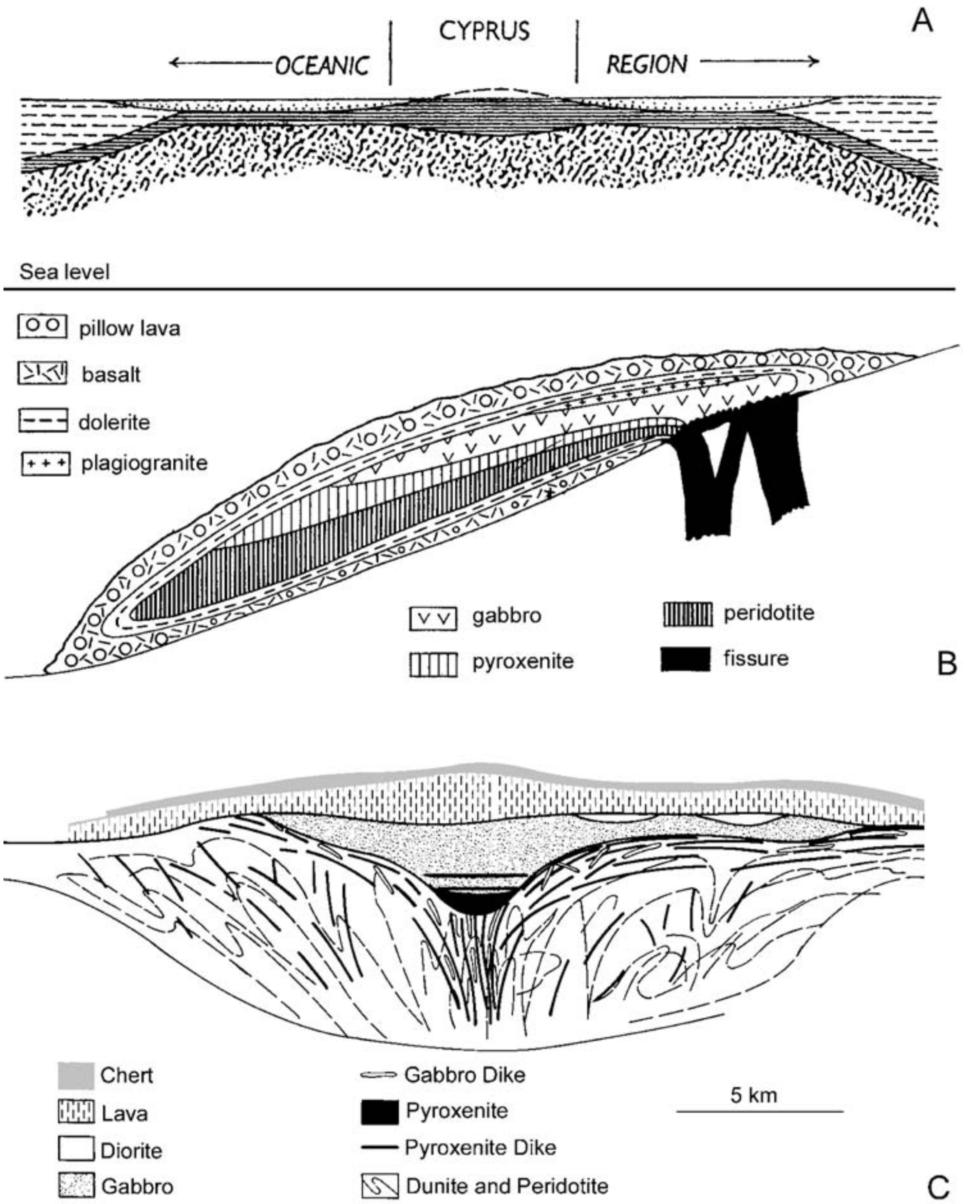

Figure 4. Early Troodos ophiolite models. A: Formation of the Troodos ophiolite as a volcanic pile on ocean floor (Gass and Masson-Smith, 1963, Fig. 14). Modified by permission of Royal Society of London. B: Troodos ophiolite formed from a massive volcanic outpouring in a geosynclinal basin (Aubouin, 1965, Fig. 37). C: Reconstructed cross-section of the Vourinos ophiolite in Greece (Moores, 1969, Fig 13).

(p. 133). Finally, he noted that "it is possible that the Troodos massif represents a volcanic edifice formed on the median ridge of Tethys" (p. 133). Thus, Gass (1967) was the first to see the similarities between the global mid-ocean ridge system and the Troodos massif, despite expressing at the same time his reservations for an ophiolite-ocean crust analogy because of the lack of direct evidence. Only a year later, Gass (1968) again expressed his reservation in a paper titled: "Is the Troodos Massif of Cyprus a Fragment of Mesozoic Ocean Floor?" Despite this skepticism, he proceeded with a comparison of the Troodos ophiolite with new oceanic crust generated at mid-ocean ridges. Gass (1968) proposed that the Troodos was indeed an uplifted fragment of fossil oceanic crust, and that its plutonic complex might represent oceanic layer 3 formed beneath a former spreading axis.

This idea was subsequently more fully developed by Moores and Vine (1971) and Vine and Moores (1972) in now-classic Troodos papers. Moores and Vine (1971) concurred with Gass (1968) in that the Troodos "represents a slice of oceanic crust and uppermost mantle" (p. 446). They inferred that the harzburgites and dunites represent depleted mantle, whereas the pyroxenites and gabbros are intrusive bodies that locally contained cumulates. This was the first time that fossil magma chambers were recognized within the plutonic sequence of the Troodos massif. Moores and Vine (1971) suggested that the lower part of the sheeted dike complex, isotropic gabbros, and the gabbroic and pyroxenitic cumulates were equivalent to ocean layer 3 (Fig. 5). They further advocated the high-level emplacement of multiple magma chambers in a slow-spreading oceanic setting characterized by varied rates of magma supply and extension (Moores and Vine, 1971, p. 461). Kidd (1977) modeled the zonation, subsidence, and rotation of the sheeted dike complex and extrusive lavas, and noted a good correspondence between the field observations and the theoretical predictions. The magma chamber modeled by Kidd (1977) was narrow, with steeply dipping walls and a general shape of a dike, broadening downward toward the base of the crust.

The significance of the Troodos plutons subsequently came into focus again with the pioneering work of Greenbaum (1972). Greenbaum (1972) accepted the proposals of Gass (1968) and Moores and Vine (1971) by suggesting that the Troodos was a slice of fossil ocean floor. Both the gabbros and pyroxenites were interpreted to represent cumulates similar to rocks found in layered intrusions (Hess, 1960; Wager et al., 1960) and also dredged from the ocean floor (Muir and Tilley, 1964; Engel and Fisher, 1969). The magma chamber shape proposed by Greenbaum (1972) was that of a large, infinite, steady-state magma chamber continuously replenished from below by undifferentiated magma (Fig. 6A). The chamber was envisioned to have a trough-like shape, 20-km wide and 1-km deep, and to be continuous along the entire length of the spreading axis. The internal constitution of this chamber was seen to contain a series of concentric shells of cumulates as referred to in "infinite onion" or "infinite leek" models (Cann, 1991). Greenbaum (1972) argued that new magma was injected at the bottom center of the chamber and that crystallization occurred along the margins in response to a lateral temperature gradient. Crystals accumulating on the outward moving chamber floor were in time being buried beneath lower temperature minerals. The Troodos observations argued for similar, laterally continuous chambers beneath active spreading centers of the ocean floor. The magma chamber model of Greenbaum (1972) was widely advocated for the Troodos, as well as for the Oman ophiolites, in the years to follow (Fig. 6B; Gass, 1980; Pallister and Hopson, 1981).

Several problems immediately became apparent with the large, steady-state chamber model of Greenbaum (1972). The principal concern was that the large single-chamber model proposed by Greenbaum (1972) required a non-convective magma, although thermal considerations predicted convection in such chambers (Bartlett, 1969; Irvine, 1970). Sleep (1975) brought up another problematic issue by pointing out that the roof over large magma chambers of the dimensions envisioned by Greenbaum (1972) would be mechanically unstable and would thus be prone to collapse. For this reason, strengthening of the roof structure by underplating and chilling of gabbro was considered necessary to prevent collapse (Dewey and Kidd, 1977). The diverse composition of the extruded lavas also appeared more consistent with multiple-chamber models, instead of large, long-lived, and zoned single-chamber models (Allen, 1975; Smewing et al., 1975; Greenbaum, 1977); however, it was still widely believed that the thick plutonic rock sequence of the Troodos ophiolite and other ophiolites was the product of cumulus processes that occurred on the floor and along the walls in a melt column that moved away from the hot axial rift zone into progressively cooler flanks (Gass and Smewing, 1981).

In the mid 1970s, Cameron Allen at Cambridge University (U.K.) proposed an episodic chamber model for the Troodos crust in an unpublished, but widely circulated Ph.D. thesis (Allen, 1975). He observed repeated cyclic variations in the cumulate sequence and inferred this phenomenon to reflect multiple episodes of replenishments of the chamber. He further suggested the existence of axial plutons, 4-km long, 2-km wide, and with an estimated depth of 4 km (Allen, 1975, p. 109). The plutons contained layering that could be traced for distances of >1.5 km. Allen (1975, p. 110) suggested that at any time, a number of discrete, possibly interconnected magma cells of variable stages of differentiation existed along the spreading axis (Fig. 7A). The emplacement and formation of new axial cells were episodic and led to splitting of the preexisting and perhaps partially solidified cells and to their migration away from the active rift axis (Fig. 7B). Allen's (1975) magma chamber model is significant because it assumes episodic and discontinuous chambers along the spreading axis. The model further allows chamber solidification by convection and formation of bottom and marginal cumulates. A somewhat different axial structure was proposed by Smewing et al. (1975), principally based on the compositional variability in the extruded lavas of the Troodos ophiolite. Smewing et al. (1975) inferred magma eruptions from discrete chambers and mantle diapirs over a broad zone, laterally extending ~30 km on both sides of the spreading axis

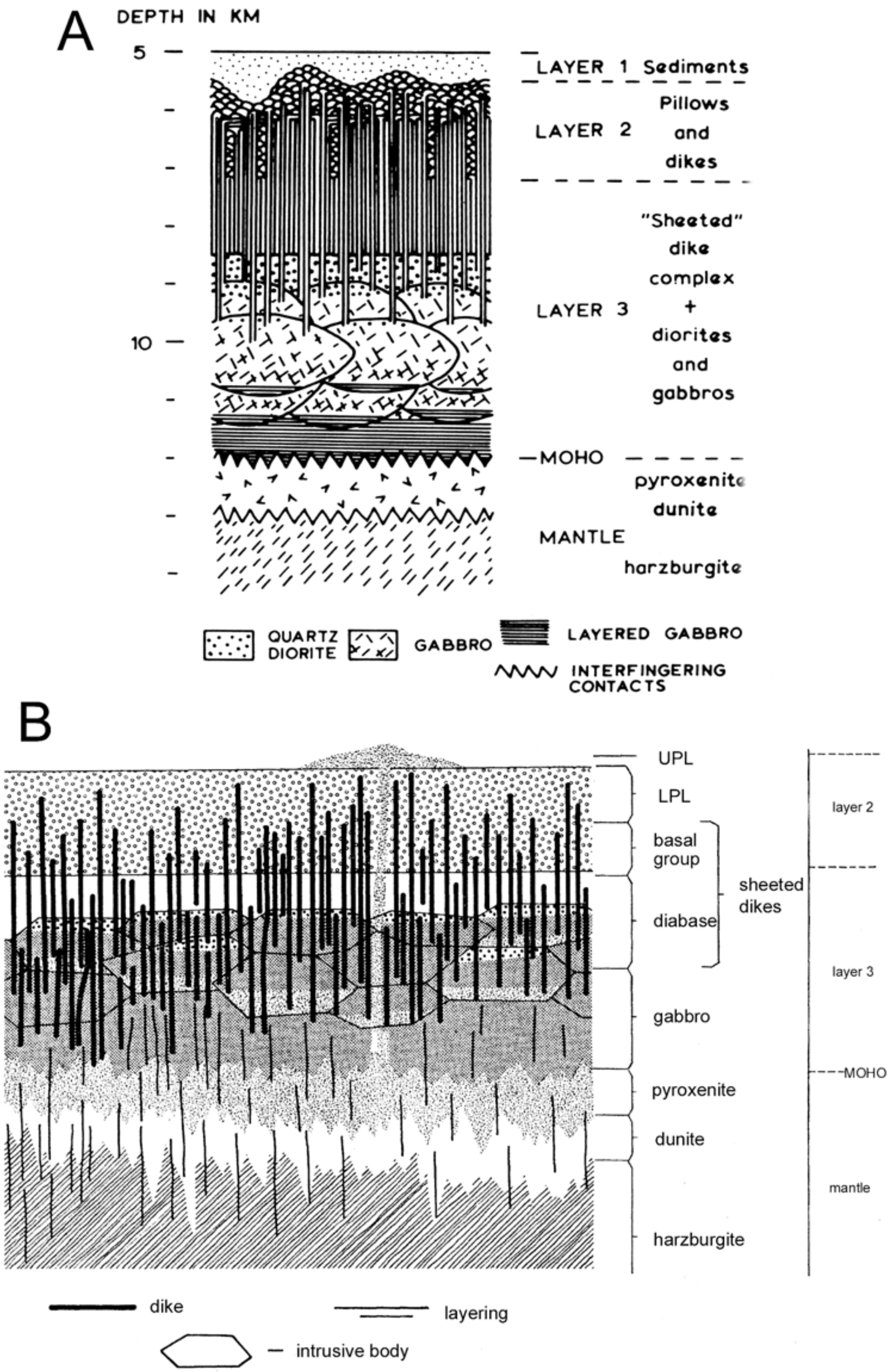

Figure 5. Cross-sections of a multi-chamber-spreading center, based on the Troodos ophiolite. A: Vine and Moores (1972, Fig. 1). Similar to the section drawn by Moores and Vine (1971), except that the multiple intrusions and the nested nature of the upper gabbro cells are marked and the lower gabbros are shown as layered. B: Moores and Vine (1971, Fig. 20). The upper gabbros are drawn as individual magma cells and the lower gabbros as stratiform gabbros. The dotted horizons on top of the magma cells are diorite/granophyre segregations; the fine dotted base of some cells is composed of ultramafic cumulates. Heavy vertical lines represent dikes, and thin vertical lines depict banding of lower gabbros and mantle tectonites. LPL—lower pillow lavas; UPL—upper pillow lavas. Reproduced by permission of Royal Society of London.

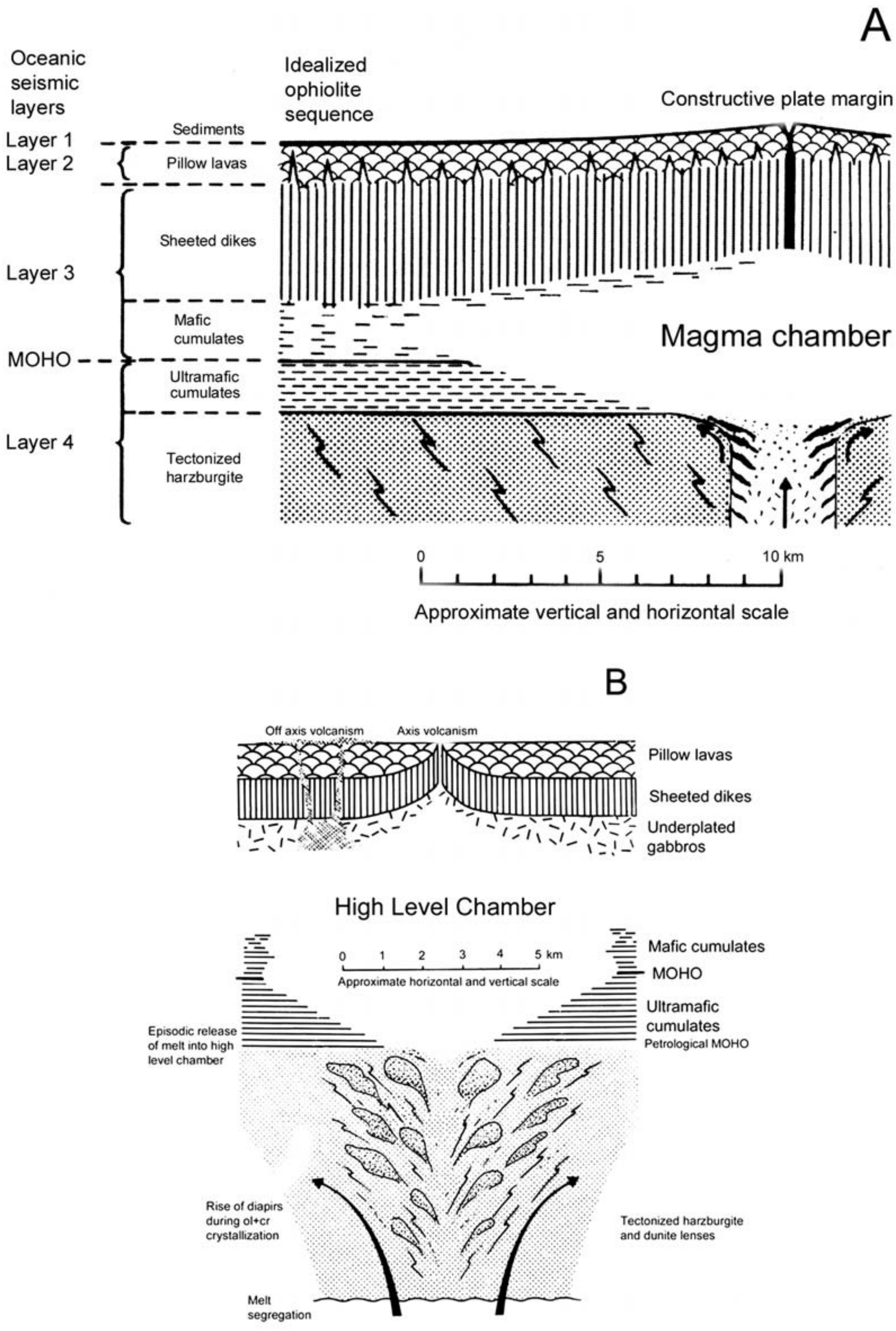

Figure 6. Cross-sections of a single-chamber-spreading center, based on the Troodos ophiolite. Reproduced by permission of the Geological Survey Department of Cyprus. A: After Gass (1980, Fig. 3), drawn from an illustration in Greenbaum (1972, Fig. 3). B: Chamber cross-section, based on a composite of ophiolites (simplified from Gass, 1980, Fig. 4).

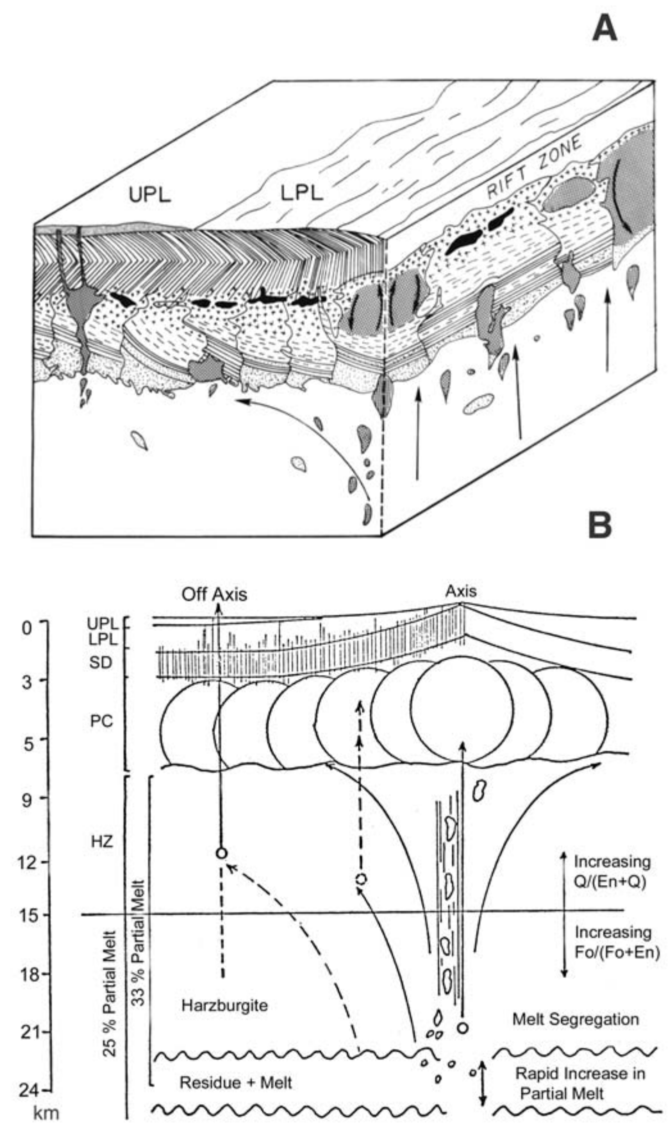

Figure 7. Cross-sections of a multi-chamber-spreading center, based on the Troodos ophiolite. A: Allen (1975, Fig. 49). Individual magma cell boundaries are illustrated to emphasize the periodic, along-axis construction of the crust. Also shown are off-axis intrusions and volcanics. B: Schematic model of spreading center. Simplified and redrawn from Allen (1975, Fig. 51).

(Fig. 8). The axial rift zone model of Smewing et al. (1975) is in essence similar to the subsequent models by Eldridge Moores and coworkers (Varga and Moores, 1985; Moores et al., 1990), suggesting several successive rift jump events during the formation of the Troodos ophiolite. This structure added an additional "degree of freedom" to the axial structure of magma chambers depicted by Allen (1975).

The plutonic section of the Troodos ophiolite was sampled in 1986 by drill hole CY4 of the Cyprus Crustal Study Project. The hole penetrated the transition between the sheeted dike complex and the gabbros at 630 m below the surface, continued through gabbroic cumulates to a depth of 1750 m, and was terminated in ultramafic cumulates at a depth of 2260 m, without reaching tectonized peridotites (Fig. 9A; Thy et al., 1989; Malpas et al., 1989). This 1630-m continuous core of the gabbroic and ultramafic cumulates provides a direct look into the nature of the fossil upper Troodos magma chambers and their crystallization processes. The lithologic units defined in the CY4 drill core are mainly based on the observed petrography and the compositional variation of pyroxene (Fig. 9B). The upper gabbros are isotropic and are composed of plagioclase, augite, and orthopyroxene cumulus phases. Fe-Ti oxides are not early crystallization phases, except in a short interval between the 1196 and 1213 m (Thy et al., 1989). The contact at 1300 m below the surface (between Units 3 and 4, lower and upper gabbros, respectively) was described as intrusive with the lower gabbro showing grain size fining upward against the upper gabbro (Thy et al., 1989). The lower gabbros, down to the 1750-m depth, commonly show grain-size and modal layering of gabbronorite, gabbro, olivine gabbro, and websterite on a scale ranging from 1 to 10 m. Olivine only locally joins the cumulus assemblage. The lowermost part of the core is composed dominantly of websterite with minor amounts of clinopyroxenite, wehrlite, lherzolite, and gabbronorite. Olivine is consistently a minor cumulus phase in the websterite. The lower cumulates display well developed modal and grain-size layering on scales ranging from a few centimeters to several meters. Medium-grained to pegmatitic gabbroic rocks are locally present as layers and patches in the websteritic cumulates (Thy et al., 1989; Browning et al., 1989).

The cryptic variation of the cumulus minerals is illustrated in Figure 9B. The lower gabbroic to websteritic cumulates (Units 4 and 5) show a systematic upward increase in the iron content for all three mafic phases. In the upper part of the lower gabbros (Unit 4), a reversal to more primitive compositions (and olivine saturation) is followed by a decrease in the iron content of pyroxene. Plagioclase shows a weak compositional variation in the albite content that mimics the variation of the mafic phases in the gabbroic cumulates. The unsystematic variation of plagioclase composition in the websteritic cumulates is related to the intercumulus origin of plagioclase in these rocks. These first order observations from the lower cumulate series led Thy et al. (1989) to suggest that the variations were in general consistent with fractional crystallization in a mostly closed magma chamber.

The intrusive contact at the 1300-m depth is a major compositional boundary in the core. The gabbros below this boundary are more primitive in terms of their major and minor element ratios than the gabbros above. The pyroxenes, for example, have markedly lower Al and Ti contents in the lower gabbros compared to the upper gabbros (Thy et al., 1989; Malpas et al., 1989). The upper gabbros show systematic variations that allow three units to be defined (Units 1–3), each having a systematic upward regression in magnesium contents of the pyroxenes, followed by a trend toward decreasing iron contents. Fe-Ti-oxide minerals occur

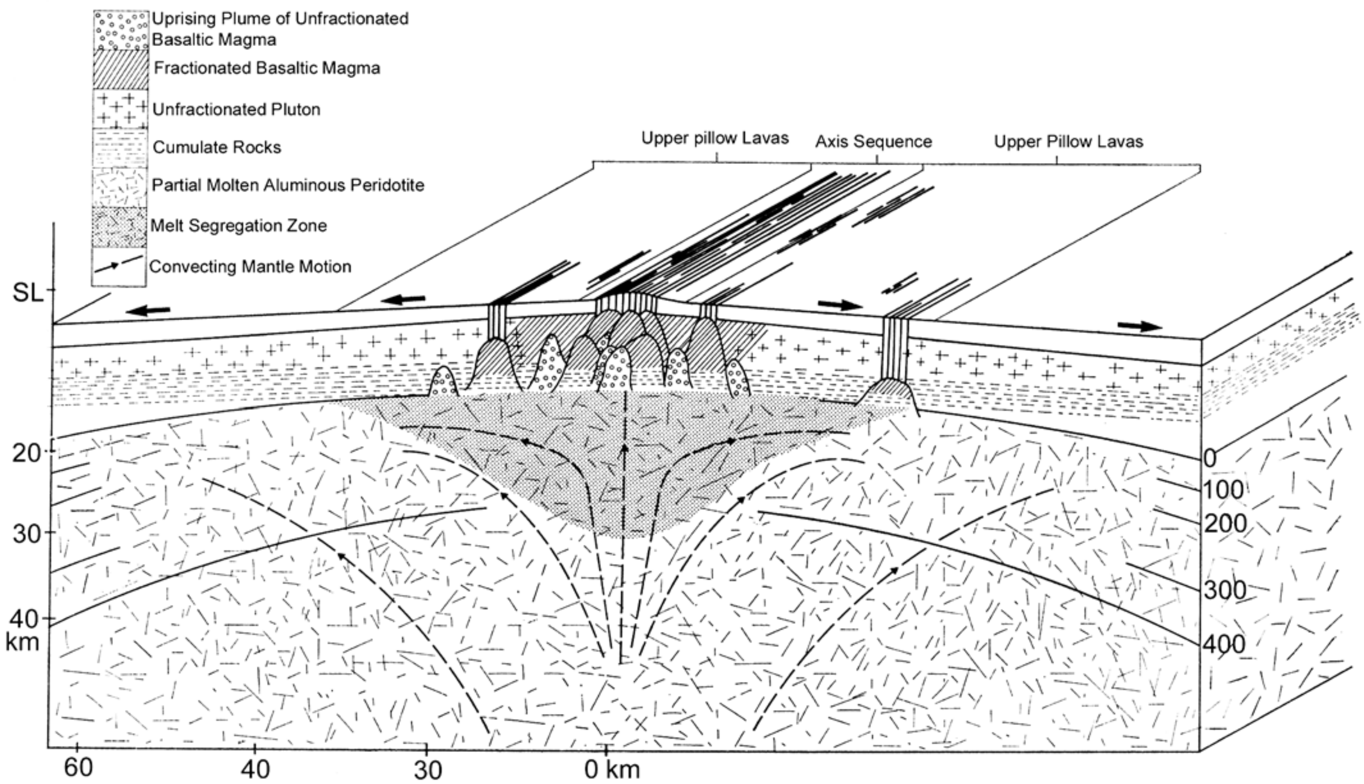

Figure 8. Spreading processes forming the Troodos ophiolite (after Smewing et al., 1975, Fig. 4). Shown are mantle flow lines and zone of partial melting (shaded). The vertical scale on the right-hand side of the diagram shows the excess temperatures relative to the fusion of olivine tholeiite. Reproduced by permission of Springer-Verlag.

intermittently in the most evolved gabbros at the top of Unit 3, just prior to initiation of the regression, marking the beginning of Unit 2. These units were interpreted (Thy et al., 1989) to reflect an intermittently open magma chamber system with three cycles, each showing gradational filling and mixing between resident magma and a new influx of a relatively more primitive magma before closed system fractionation was resumed.

Field studies have elaborated on the spatial and temporal relationships between intrusive episodes and the high temperature deformation in the Troodos plutonic complex (Malpas et al., 1989; Malpas, 1990; Dilek and Eddy, 1992). Malpas et al. (1989) mapped structural details in the Mount Olympus area of the Troodos complex and observed two major plutonic suites on the basis of relative ages of deformation and magmatism. An early suite consists mainly of harzburgite, dunite, layered olivine pyroxenite, and gabbro and displays a penetrative high-temperature deformation fabric. This tectonic fabric is generally parallel to the compositional layering in the plutonic suite. Malpas (1990) suggested that these structures of the early suite reflect the intrusion of plutons in an upwelling mantle column at, or near, a spreading center, and that they might have resulted from upward and lateral solid flow. A late suite of plutons includes gabbroic and websteritic cumulates and displays locally well preserved primary cumulus textures containing xenolithic inclusions of the early suite (Benn and Laurent, 1987; Malpas, 1990). These field observations thus illustrate the complex relations between the gabbroic and ultramafic plutons of the Troodos ophiolite and strongly support the multiple magma chamber evolution model (e.g., Allen, 1975) at different crustal levels associated with intermittent episodes of magmatic and tectonic spreading (Malpas, 1990). The upper parts of the CY4 drill core sampled the early plutonic suite, which records a relatively open magmatic system (Thy et al., 1989; Malpas, 1990). The lower part of the drill core sampled a late-stage pluton, which may represent one of several episodes of plutonism that constituted the late suite (Malpas, 1990). Geochemical studies of cumulus clinopyroxene and melt inclusions in plagioclase from the southern flank of the Troodos massif have suggested that the latest plutons originated from the most depleted melts (Batanova et al., 1995). This multiple and intermittently replenished type of magma chambers, seen in the Troodos ophiolite, are not found in the plutonic complexes of the Bay of Islands and the Oman ophiolites.

Bay of Islands Ophiolite (Newfoundland)

The Bay of Islands ophiolite complex is a 100-km long discontinuous belt of mafic and ultramafic rocks emplaced onto

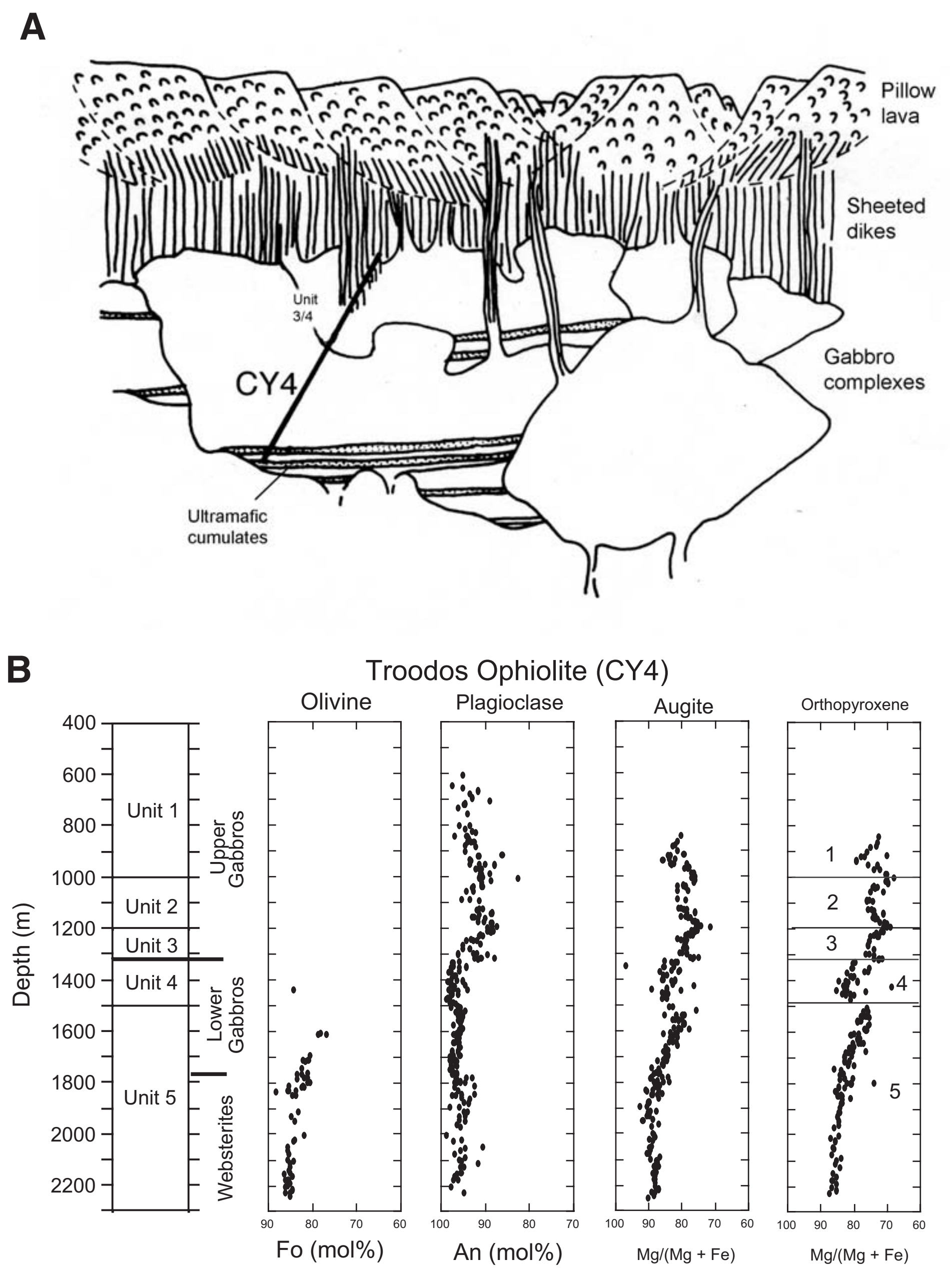

Figure 9. Cross-section and compositional variation in the multiple intrusive gabbros and ultramafic cumulates of drill core CY4 through the Troodos ophiolite. A: Schematic cross-section illustrates the inferred location and position of drill Hole CY4 (Malpas, 1990, Fig. 12). Reproduced by permission of the Geological Survey Department of Cyprus. B: Cryptic variation in the plutonic complex penetrated by CY4. The units are based on the systematic variations and discontinuities in the cryptic variation. The contact between Units 3 and 4 is interpreted to be intrusive. See Thy et al. (1989) and Thy and Dilek (2000) for details.

the Paleozoic continental margin of southwestern Newfoundland during the early stages of the tectonic evolution of the Appalachian orogen (Fig. 10). Structural and petrological studies (Cooper, 1937; Smith, 1958; Church and Stevens, 1971; Dewey and Bird, 1971; Church and Riccio, 1977) revealed the existence of a typical ophiolite succession composed of tectonized peridotite, ultramafic and mafic plutons, sheeted dikes, extrusive rocks, and overlying sedimentary rocks. The early studies of the Bay of Islands complex interpreted this ophiolite as part of the western Newfoundland eugeosynclinal belt (Fig. 10) and suggested that the internal ophiolite differentiation mostly resulted from accumulation in a syn-orogenic pluton (Smith, 1958). Our present understanding of this ophiolite is based mainly on the results of the systematic mapping programs in the late 1970s (Dewey and Kidd, 1977; Casey and Karson, 1981; Casey et al., 1981), well after the formulation of the general plate tectonic models (e.g., Casey and Karson, 1981; Casey et al., 1981).

In 1977, Dewey and Kidd proposed a general steady-state magma chamber model for accreting plate margins (Fig. 11) and based this model mainly on field evidence from the Bay of Islands ophiolite. Their model depicted five components that included, from top to base: (1) a lid composed of extruded lavas and an

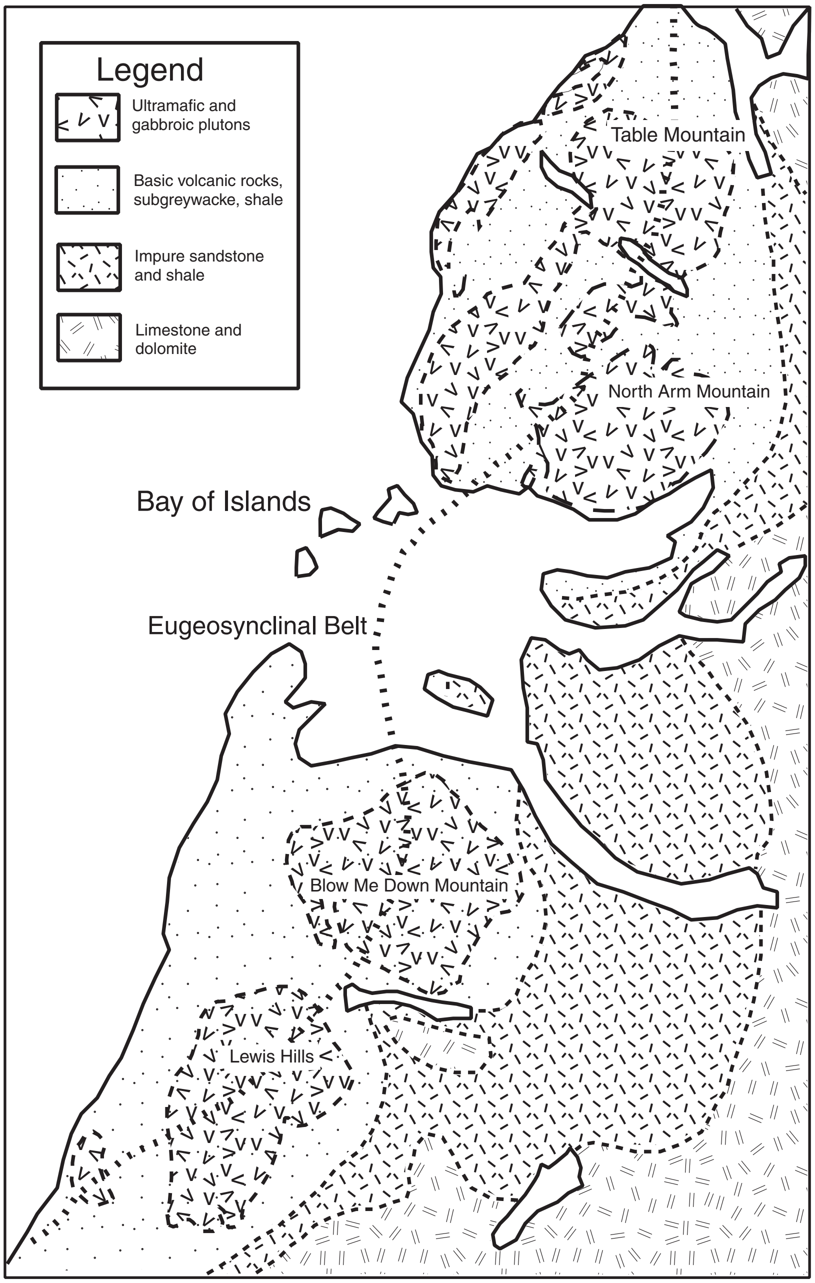

Figure 10. Simplified geological map of the Bay of Islands ophiolite complex, based on Smith (1958). Dotted line marked as the "eugeosynclinal belt" is the inferred axis of basaltic magma outpour into the Appalachian eugeosynclinal marginal belt of western Newfoundland.

underlying sheeted dike complex; (2) isotropic gabbros accreted to the lid and progressively thickened away from the axial rift; (3) a tent-shaped magma chamber with a flat floor; (4) differentially subsided cumulates beneath the chamber floor; and (5) a narrow axial dome of upwelling lherzolite, from which magma was segregated into the magma chamber within the crust, and from which a harzburgitic mantle was accreted beneath the lower crust. Girardeau and Nicolas (1981) expressed very similar views on the Bay of Islands ophiolite. In this model the depth of the magma chamber controls the thickness of the underplated upper gabbros; the width of the chamber (<10 km) is determined by the cooling and spreading rates that similarly control the thickness of the cumulates and the orientation of banding (Dewey and Kidd, 1977). Hydrothermal circulation is responsible for the thickness of the sheeted dike complex and for the accretion of the gabbros along the roof of the axial chamber. These gabbros contribute to the stabilization of the axial lid over the chamber (Rosencrantz and Nelson, 1982; Rosencrantz, 1983). The model of Dewey and Kidd (1977) assumes a near crystal-free magma chamber and infers that crystal mushes are confined to the underplated gabbros and to the subsiding mafic and ultramafic cumulates. Perhaps one of the most interesting aspects of Dewey and Kidd's axial cham-

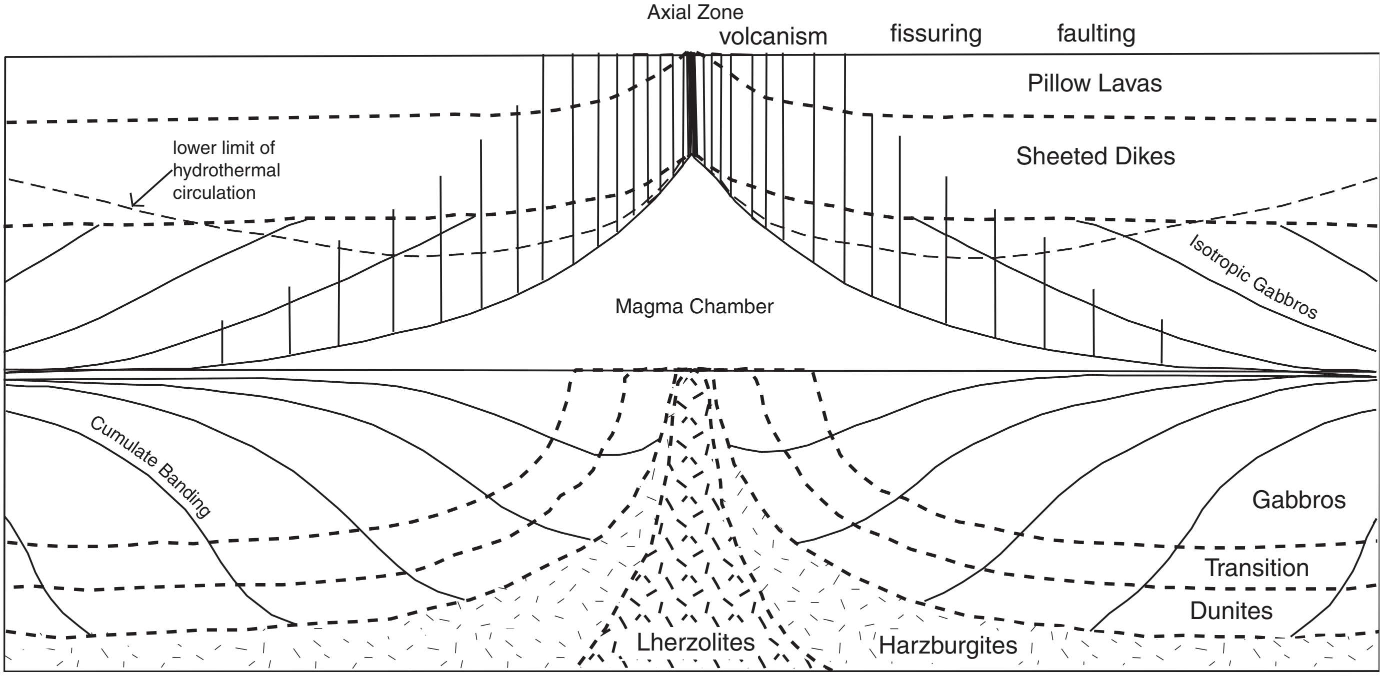

Figure 11. Constructive plate accretion model for axial spreading center, formulated from observations in the Bay of Islands ophiolite. Simplified from Dewey and Kidd (1977, Fig. 1).

ber model is that it differentiates between the upper, underplated gabbros and the lower, gabbroic and ultramafic cumulates.

Casey and Karson (1981) and Casey et al. (1981) subsequently reported the results of their field mapping in the Bay of Islands ophiolite. The axial magma chamber they envisioned differed significantly from that of Dewey and Kidd (1977). They proposed a large, steady-state chamber extending along the ridge axis and only terminating against the transform fault intersections. They inferred a chamber depth of 6–10 km and a width of 14–16 km. Banding in the gabbros was inferred to be parallel to the inclined margins of the chamber and was attributed to primary igneous layering that outlined the original shape of the axial chamber (Fig. 12). Steepening of the cumulate banding away from the axial center, inferred by Dewey and Kidd (1977) to have resulted from differential subsidence, was interpreted as the primary layering formed along the chamber margins. This interpretation required that the cumulates formed principally by in situ crystallization with little interstitial trapped melt and that they were unaffected by postcumulus deformation. The magma chamber suggested by Casey and Karson was very similar to the large continuous, steady-state chamber envisioned by Greenbaum (1972) for the Troodos ophiolite; however, Casey and Karson (1981) found no indication of a flat chamber floor, and suggested instead a large chamber keel extending into the tectonized mantle sequence (Fig. 12). Elthon et al. (1982, 1984) and Casey et al. (1983) expanded this keel to a depth of 30–40 km (Fig. 13) to account for the early appearance of clinopyroxene (Church and Riccio, 1977), which they interpreted to record high-pressure fractionation. With the exception of a suggestion in a short note by Strong and Malpas (1975), the occurrence of multiple magma chambers was not proposed for the Bay of Islands ophiolite.

The "Blow Me Down" and "North Arm" cumulate sections in the Bay of Islands ophiolite (Fig. 10) are composed of interlayered dunite, wehrlite, harzburgite, pyroxenite, and gabbro. A thick gabbro sequence overlies the ultramafics and contains layers of troctolite, anorthosite, and gabbronorite. Various domains of porphyritic, pegmatitic and poorly layered gabbros, hornblendite, and evolved rocks also exist. These plutonic rocks collectively occur beneath the sheeted dike complex and pillow lavas (Church and Riccio, 1977; Casey and Karson, 1981; Bédard 1993). The cumulate section displays cumulus phase assemblages and cumulus compositions that apparently evolved upward due to fractionation processes (Fig. 14). The compositions recorded by the cumulates correspond to the total variation in the sheeted dike complex and the extrusive rocks (Church and Riccio, 1977). This observation was used to suggest an open nature of the Bay of Islands magma chamber. Despite the systematic upward cumulus evolution toward more evolved compositions, the ultramafic and gabbroic cumulates show reversals to more primitive compositions on various scales (Church and Riccio, 1977; Elthon et al., 1982, 1984; Komor and Elthon, 1990; Komor et al., 1985, 1987). These small-scale variations may encompass the near total compositional variation seen in the complete stratigraphic section (Fig. 14). Elthon et al. (1982) attributed this phenomenon to in situ crystallization processes along vertical chamber margins or conduits (Fig. 13).

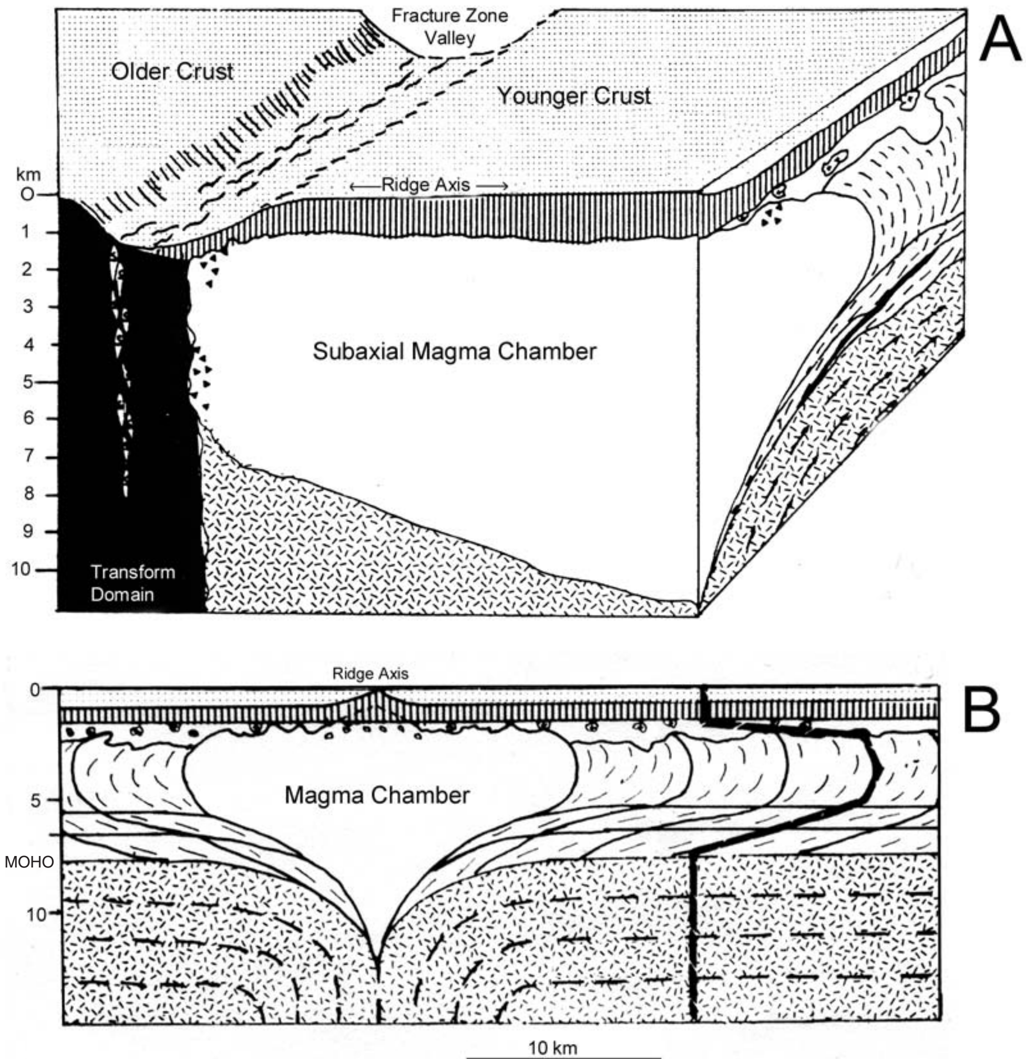

Figure 12. Magma chamber model based on the Bay of Islands ophiolite (from Casey and Karson, 1981, Figs. 4 and 5). Reproduced by permission of *Nature*. A: Block diagram illustrating the axial chamber and its intersection with a transform zone (Casey and Karson, 1981, Fig. 4). B: Cross-section through an idealized axial magma chamber for the Bay of Islands ophiolite. The thick line is an isochron at 1.25 m.y. for a 10 cm/yr half-spreading rate. Note the deep keel of the chamber.

A detailed field and petrographic study of the ultramafic and mafic cumulates of the North Arm massif by Jean Bédard and coworkers did not find strong evidence for large magma chambers, as were previously reported. Bédard (1991, 1993) and Bédard and Hébert (1996) recognized near-pervasive syn-magmatic deformation and proposed that the entire Bay of Islands crust be formed as a synkinematic sill complex. Residual magma from the sills and from the host mush/gabbro was expelled laterally or obliquely along shear zones. The ultramafic cumulate section was proposed to be an underplating boninitic sill complex genetically unrelated to the overlying gabbros (Bédard and Hébert, 1996). Upward regressions in cryptic mineral variations in the gabbroic section were interpreted not as replenishments into an open chamber, but either as tectonic inversion produced by deformation, or by injection of primitive sills into the cumulate mush (Bédard, 1993). The petrographic and compositional variations were thus attributed to assimilation and reaction processes between primitive intrusive sills and host cumulates, and not to an open, frequently replenished magma chamber, as previously argued. These observations have significant implications for how primitive magma migrates through cumulate mushes and how this migration may affect the residual solid product and the liquid line of descent of primitive melts beneath oceanic spreading centers. This theme was subsequently taken up by other researchers for the Oman ophiolite.

Oman Ophiolite (Sultanate of Oman)

The Oman ophiolite constitutes a significant component of a nearly 800-km long, arcuate mountain belt that borders the

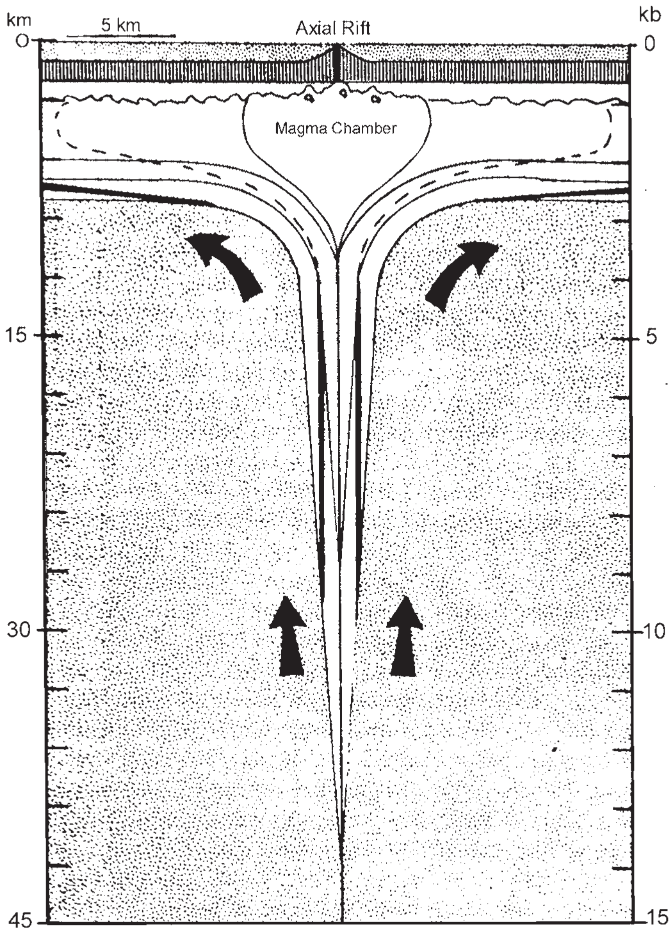

Figure 13. Bay of Islands magma chamber modified from Elthon et al. (1982, Fig. 8). Elthon et al. (1982) observed the effects of high-pressure fractionation in the North Arms massif and suggested that crystallization occurred at deep levels in the mantle. A chamber keel extending deep down into the mantle was proposed. Reproduced by permission of American Geophysical Union.

southern coast of the Gulf of Oman (Fig. 15). The ophiolite represents a fragment of the fossil Tethyan oceanic crust emplaced as a nappe system onto the passive continental margin of the Arabian Peninsula (Lees, 1928; Glennie et al., 1974; Coleman, 1981). The Oman ophiolite contains tectonized peridotites and subordinate gabbroic cumulates with lesser amounts of dunite and wehrlite (Reinhardt, 1969; Allemann and Peters, 1972; Glennie et al., 1974). Overlying the plutonic sequence are the basaltic sheeted dike complex and pillow lavas. Early studies culminated in typical geosynclinal views for the origin of the ophiolite (Wilson, 1969; Reinhardt, 1969). Wilson (1969) saw the igneous complex as being autochthonous and suggested that the peridotites and gabbros formed from a slowly cooling, large, abyssal ultrabasic magma flood erupted in a eugeosynclinal marginal basin. Other studies recognized an age hiatus between the host tectonized peridotites and basaltic intrusive and extrusive rocks. Of particular note is the model of Reinhardt (1969), illustrated in Figure 16. This model depicted the ophiolite as a large basaltic intrusion or a lava lake emplaced into the host tectonized peridotite and extruded on the sea bottom, forming sheeted dikes and extrusive spilitic pillow lavas. The internal constitution of the massif was believed to have resulted from differentiation of the magma into fine- and coarse-grained gabbros with a marginal gabbroic-peridotitic zone.

A Dutch group mapping in the Oman Mountains in the late 1960s contributed significantly to our understanding of the internal structure and stratigraphy of this ophiolite. The first comprehensive geological map as a result of this project was published by Glennie et al. (1974). This study demonstrated the similarities between the Oman ophiolite and the general structure of modern oceanic crust. The gabbroic sequence (or layer 3) was interpreted to have formed from a large dike-like intrusion that had extended beneath an oceanic ridge to a depth of 50 km above an upwelling mantle. The layers of oceanic lower crust formed on the walls of this dike and were transported upward and laterally as if on large conveyer belts driven by mantle convection (Fig. 17).

More detailed mapping in the 1970s by American and British research groups followed this early reconnaissance mapping. These international efforts resulted in detailed maps and produced significant petrological and geochemical information that permitted detailed magma chamber models to be formulated (Coleman and Hopson, 1981; Lippard et al., 1996). The most influential chamber models were published by Pallister and Hopson (1981) and Smewing (1981) as shown in Figure 18. Pallister and Hopson (1981) made several important observations that they used to infer magma chamber processes and the shape of the magma chamber of the Oman ophiolite. In this model, the typical cumulate sequence includes a basal section composed of dunite that grades downward into a transitional zone made of harzburgitic tectonites. The main part of the cumulates consists of interlayered wehrlite, melanogabbro, and gabbro rocks, followed upward by layered gabbro with recurring zones of wehrlite and melanogabbro. The uppermost plutonic rocks are made of laminated to non-laminated, cumulus gabbros and isotropic hornblende-bearing gabbros. The cumulates, composed dominantly of adcumulates (Wager et al., 1960), were accumulated on the floor of the magma chamber (Pallister and Hopson, 1981). Downward crystallization from the roof of the chamber was minor, as indicated by a rather thin unit of isotropic gabbros below the chamber roof and the overlying sheeted dike complex. Magmatic cyclicity in the chamber was seen as petrographic repetitions in the occurrence of olivine-rich intervals, commonly related to reversals in the cryptic variation. Despite such reversals, the gabbro sequence shows very limited cryptic variation (Fig. 19). This was taken to indicate frequent replenishments restricting differentiation of magma in the chamber. Pallister and Hopson (1981), thus, inferred that the Oman gabbros were the product of crystallization in a large, long-lived, frequently replenished magma chamber (Fig. 18A), as opposed to the small transient chambers inferred for the Troodos ophiolite (Smewing et al., 1975). Smewing (1981) suggested

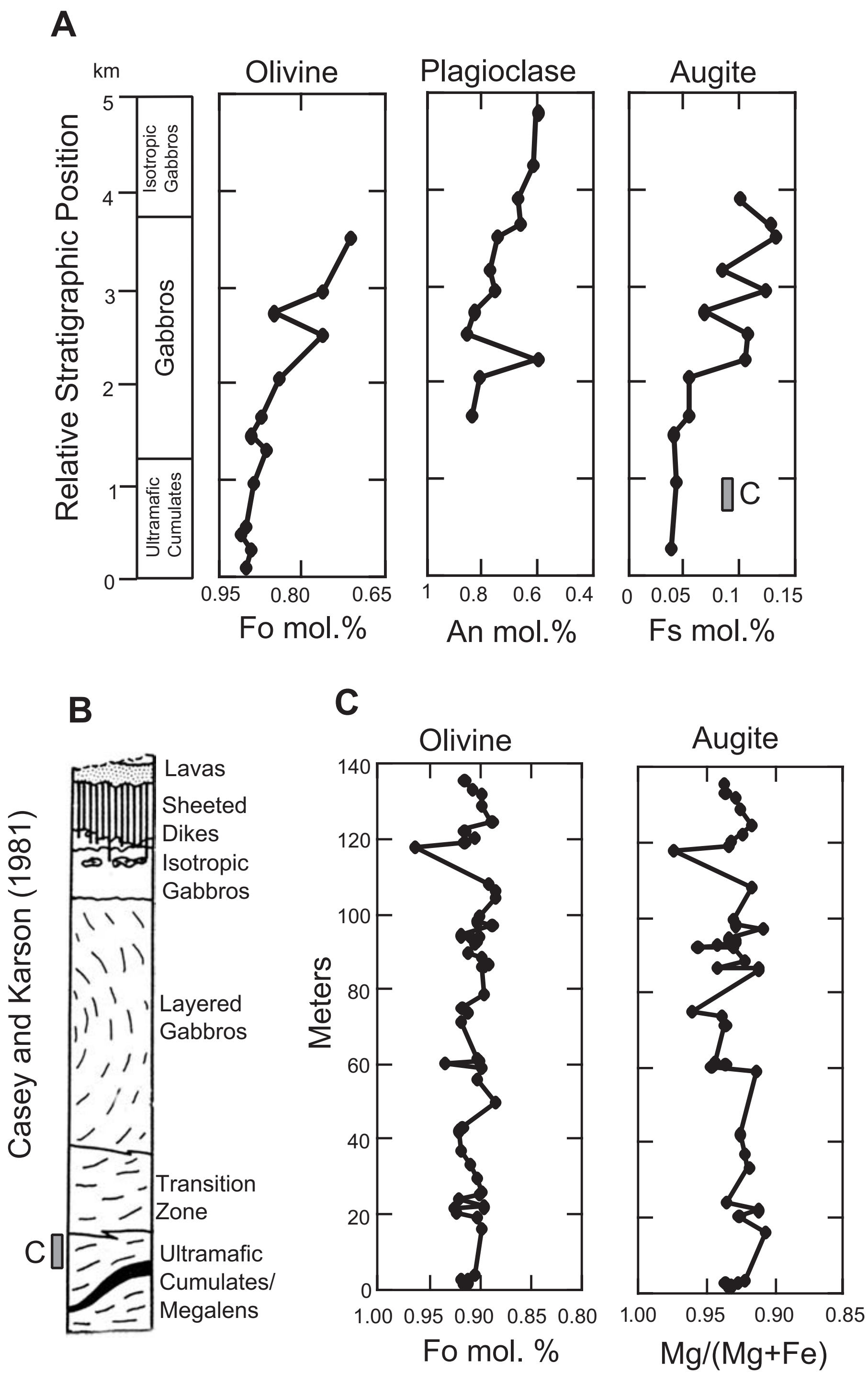

Figure 14. Cryptic variation of the main minerals of the Bay of Islands ultramafic and gabbroic cumulates. A: Composite of the "Blow Me Down" and "North Arm" sections, from Church and Riccio (1977). "Fs" is ferrosilite component of augite (Fe/(Fe+Mg+Ca)). B: Generalized stratigraphic section through the Bay of Islands ophiolite from Casey and Karson (1981, Fig. 2). The vertical bar marked as "C" shows the position of the section illustrated in greater details. C: Small-scale cryptic variation in a 140-meter layered section of ultramafic cumulates (Elthon et al., 1984, Table 2).

that the cyclicity in the Oman cumulates resulted from influx of relatively undifferentiated picritic melt entering the chamber at the base and being retained along the floor because of a density contrast with the residual, relatively evolved melt in the magma chamber (Fig. 18B). In Smewing's model, olivine-rich cumulates formed along the floor and away from the axial center until the bottom layer approached the composition and density of the overlying magma. Eventually, mixing between the bottom layer and the main chamber reservoir homogenized the remaining melt in the chamber.

The chamber shape was interpreted to have been controlled mainly by upward growth from the floor and to a lesser extent by

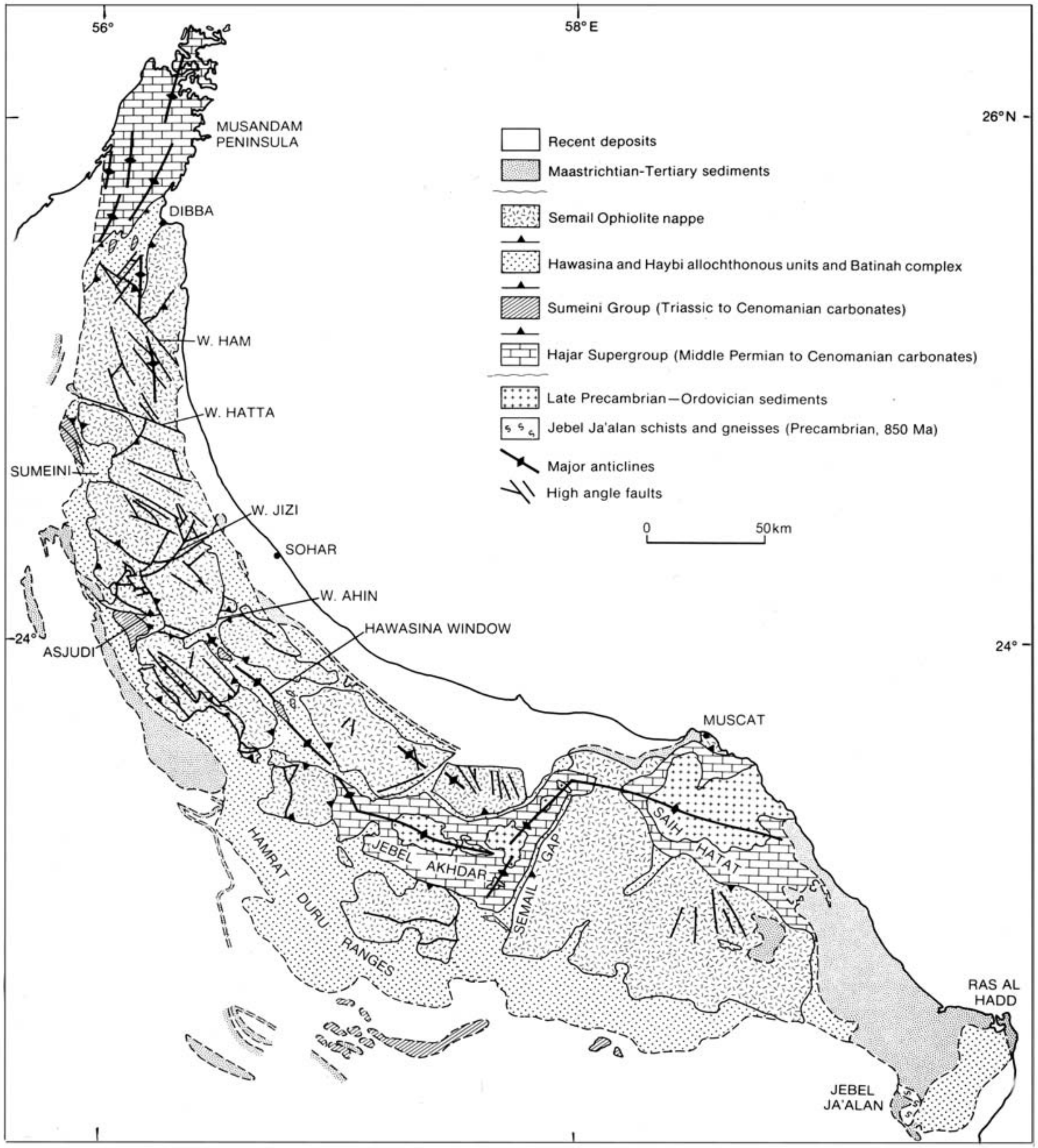

Figure 15. Geological and structural map of the Oman ophiolite (Lippard et al., 1986). Reproduced by permission of Geological Society of London.

downward growth from the ceiling (Fig. 18A). The layering in the gabbros was observed to approach banding in the peridotitic tectonites at a 20° angle. Pallister and Hopson (1981) envisioned a broad, funnel-shaped chamber with floors sloping gently inward toward the axial center and laterally terminating in a sandwiched position against and within the upper gabbros (Fig. 18A). They estimated a minimum width of 30 km for this funnel-shaped chamber by using a 15° slope of the layering in the ~5-km-thick cumulate package. These dimensions are relatively similar to those inferred by Smewing (1981; Figure 18B) and Lippard et al. (1986; Figure 20) from the northern part of the Oman ophiolite (see also Browning, 1984; Ernewein et al., 1988).

In a highly influential paper published in 1984, Browning described the occurrence of cyclic variations in the cumulus phase compositions of the cumulate sequence in the Oman ophiolite. His results were used to estimate the height of the melt column for each cyclic unit as ~100 m. Browning (1984) discussed two ways of achieving this cyclic variation (Fig. 21). He briefly

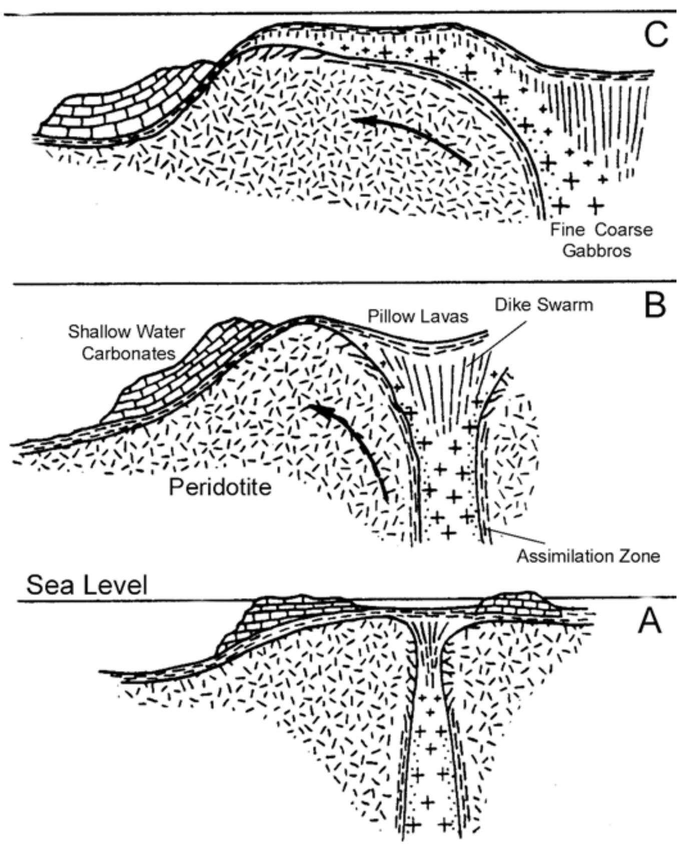

Figure 16. Eugeosynclinal emplacement and structural development (A to C) of the Oman ophiolite, as suggested by Reinhardt (1969, Fig. 17).

mentioned the possibility of a subsiding cumulate pile and stated that the true depth of the subaxial chamber was no more than the 100 m estimated for each cyclic cumulate unit (e.g., Dewey and Kidd, 1977); however, Browning (1984) indicated his preference for the idea of a zoned magma chamber and the breakdown of the melt column into double-diffusive convection cells, each with approximately a 100-m depth. This latter idea had been formulated from scaled tank experiments (e.g., Turner and Chen, 1974; Turner and Campbell, 1986) and applied to explain some features of layered intrusions (McBirney and Noyes, 1979; Irvine et al., 1983; Wilson and Larsen, 1985). The important outcome of Browning's (1984) observations is that magma cells of very limited depths can be mapped in the layered gabbroic cumulates in Oman, and that they may similarly exist in magma chambers beneath oceanic spreading centers.

Structural mapping by French groups in the Oman Mountains during the early 1980s led to a detailed understanding of the mantle-gabbro transition that markedly changed magma chamber models (Nicolas, 1989). Nicolas et al. (1988) showed that the crustal section was affected by magmatic flow and that this flow was coupled with a solid-state mantle flow. They also showed that the orientation of gabbro layering resulted in part from tectonic deformation, and that it, therefore, did not necessarily reflect primary magmatic layering. The chamber model proposed by Nicolas et al. (1988) is shown in Figure 22. The principal component of this model is a small central, tent-shaped chamber constrained by the upwelling mantle at the mantle-crust transition. The chamber is also drawn to concur in size and shape with the upper crustal seismic results from the East Pacific Rise (Detrick et al., 1987). The magma chamber and the surrounding mush zone are controlled in this model by cooling, which results in a mush below a basalt fraction of 0.70 (~1185 °C), and by marked changes in the rheological properties near the basalt solidus (~1100 °C). An additional observation, also made by Nicolas et al. (1988), is that the chamber is fed from two main sources: (1) the dominant axial basaltic melt injected from greater depths, and (2) subordinate wehrlitic mush released by compaction of material within the transitional zone at the mantle-crust boundary.

In the model of Nicolas et al. (1988), the layering of the gabbro section dips away from the axial center in a concave-up geometry as a result of the outward drag induced by deeper mantle flow. This latter observation is in sharp contrast with the previous chamber models for the Oman and other ophiolites (Greenbaum, 1972; Pallister and Hopson, 1981), as well as with the subsiding cumulate pile explanation of Dewey and Kidd (1977). This Oman chamber model has changed over the years since first proposed by Nicolas et al. (1988). Its essential features are preserved, however, in latter models (Nicolas et al., 1993; Nicolas, 1992, 1994; Boudier et al., 1996; Nicolas and Ildefonse, 1996; Chenevez et al., 1998; Nicolas and Poliakov, 2001).

Nicolas (1994) points out that the differences between the Oman magma chamber model and the oceanic model proposed by Phipps Morgan et al. (1994) can be explained by whether the mantle is upwelling in an active or passive manner (Fig. 23C). The observations and modeling for the Oman chamber suggest active upwelling (Nicolas, 1994), which results in diversion of the mantle below the axial zone. This mantle flow causes a drag in the overlying lower crust that, in turn, deforms the gabbroic rocks along planes parallel to the mantle foliation. Another important point is that the chamber is filled by a mush with perhaps 10%–20% melt, or less, and hence is able to support magmatic suspension flow (Nicolas, 1992; Nicolas and Poliakov, 2001). The magma chamber, thus, is defined as a combination of melt lens and thick mush (Nicolas et al., 1993; Nicolas and Ildefonse, 1996).

A further development in the magma chamber models for the Oman ophiolite was presented in a paper by Boudier et al. (1996). Previous models were based on the assumption that the lower crustal section was built from a subsiding floor of a perched magma lens beneath the axial rift (Fig. 23A). With an actively upwelling mantle (Nicolas, 1994), the gabbro mush pile subsides while being dragged out of the low-velocity zone, and then freezes parallel to the chamber margin. Previous observations had noted the existence of sills (Benn et al., 1988) and intrusive wehrlite bodies (Nicolas et al., 1988) at or near the mantle-

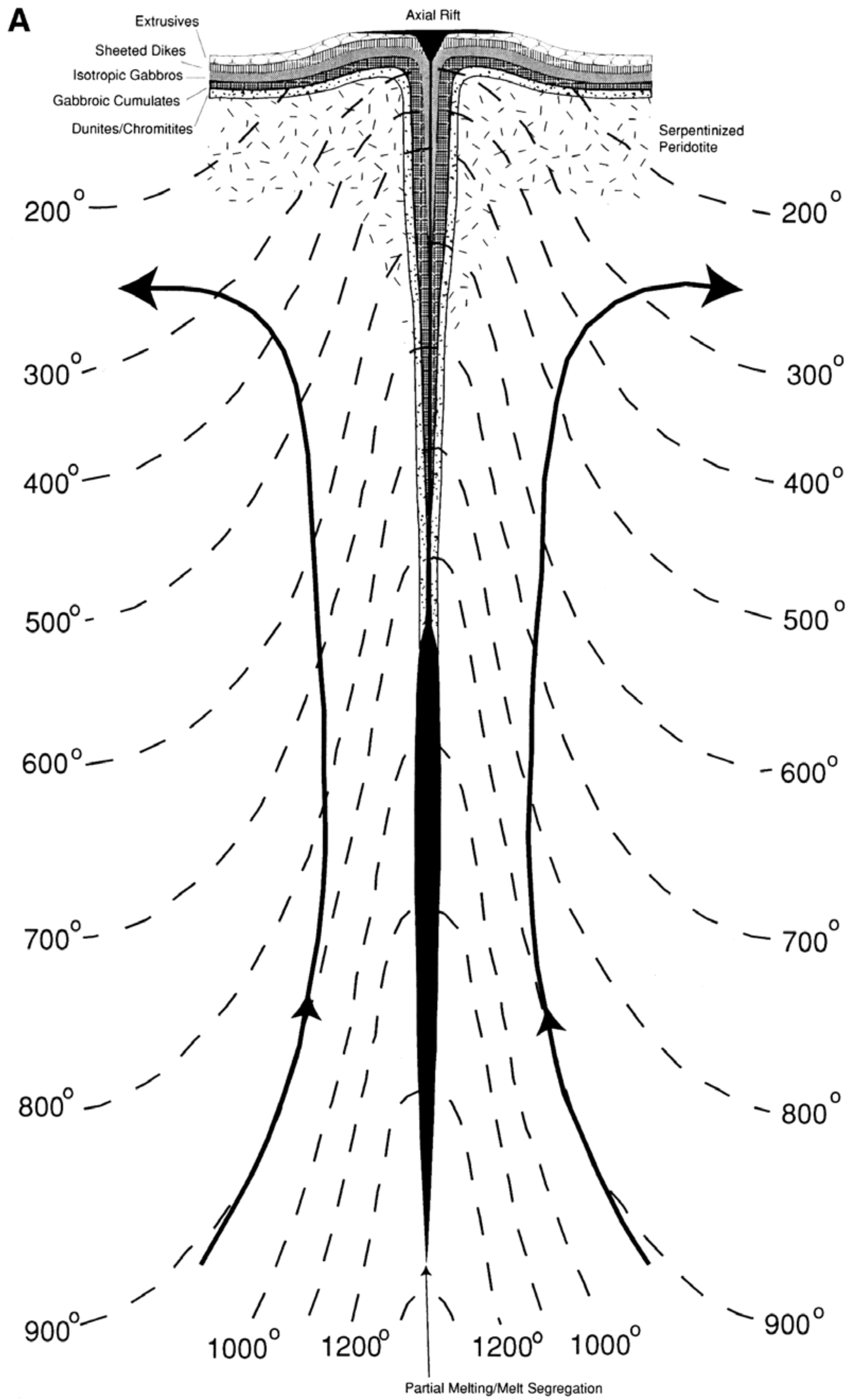

Figure 17. Crustal construction of the Oman ophiolite, as interpreted by Glennie et al. (1974, Fig. 6.6.3).

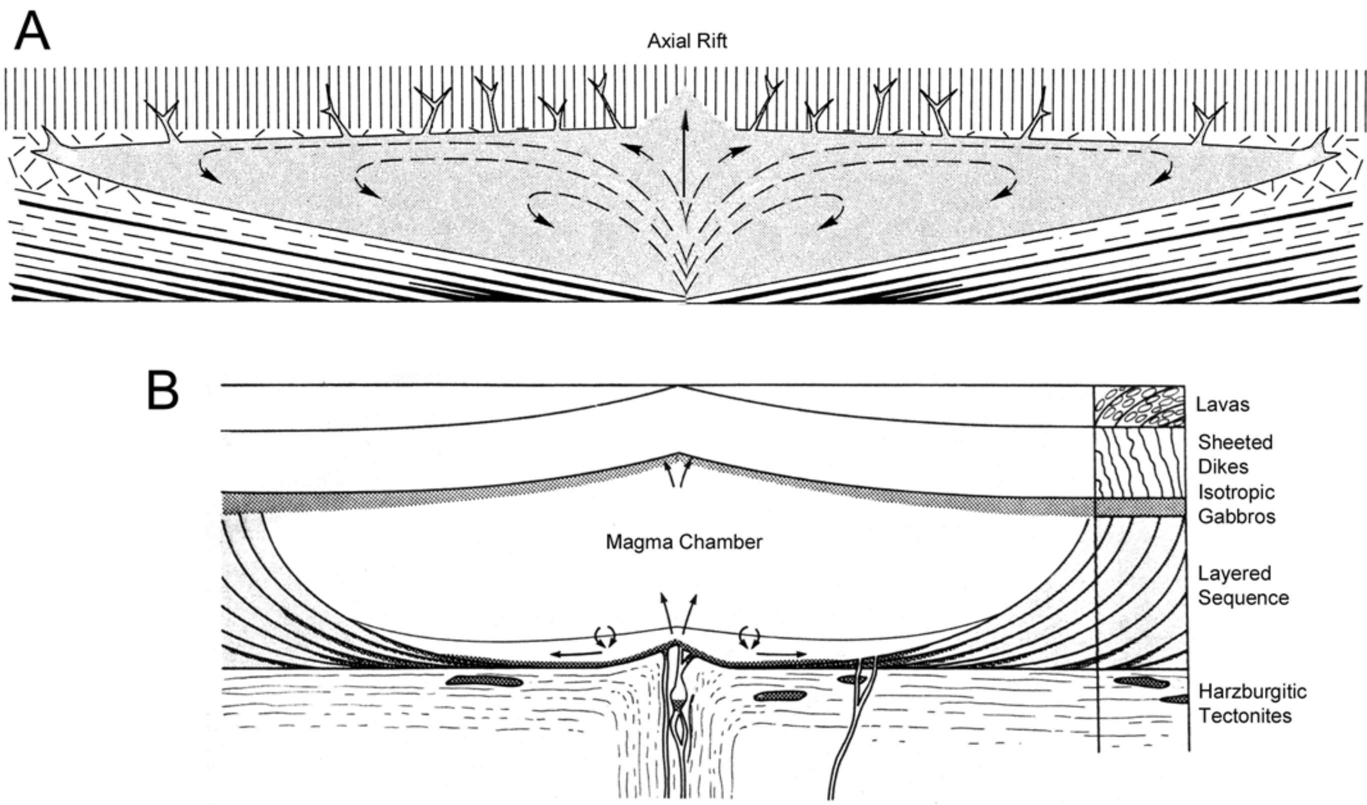

Figure 18. Steady-state magma chambers of the Oman ophiolite. A: Pallister and Hopson (1981, Fig. 19). See Hopson and Franco (1977) for an earlier version of this model. Modified by permission of American Geophysical Union. B: Smewing (1981, Fig. 9). Reproduced by permission of American Geophysical Union.

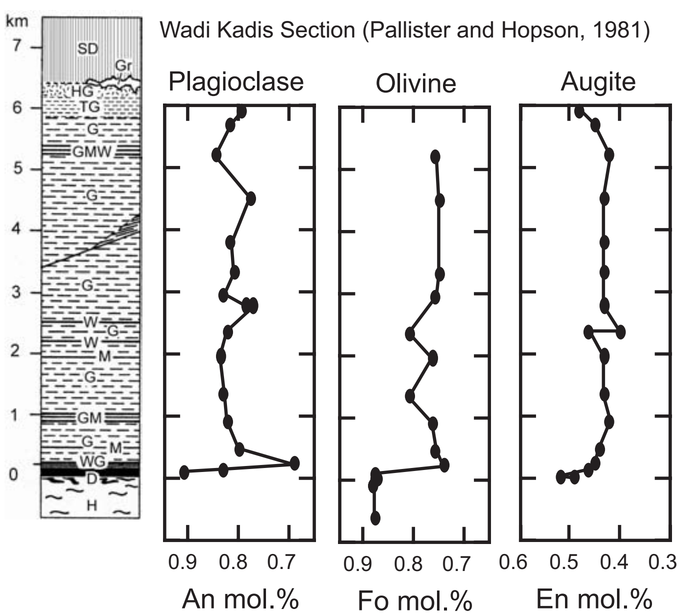

Figure 19. Cryptic variation in the Wadi Kadis section of the Oman ophiolite. Stratigraphic section and mineral compositions from Pallister and Hopson (1981, Fig. 5; Tables 2 and 3). Figure 19. Stratigraphic column reproduced by permission of American Geophysical Union.

crust transition. Boudier et al. (1996) provided further evidence for sill intrusions in the deeper parts of the crust and presented the implications of this phenomenon for magma chamber models. They suggested that the crustal section was built up both from the top via crystallization on the floor and the walls of a perched magma lens, and from the bottom through sill intrusions in the lower crustal section and within the transitional zone (Fig. 23B). In this model, the upper gabbros crystallized on the floor of the subsiding lens as a magmatic mush and obtained their foliated nature from subsidence and outward flow. The lower banded gabbros were interpreted as a composite of foliated gabbros and gabbroic or wehrlitic sill intrusions. The suggestion was based in part on the observation that graded layers were present in both sills intruded into the mantle, as well as in the transitional zone and within the lower gabbros. The strong deformation was suggested to have transformed gabbro layers into bands and lenses and to have obliterated the evidence for intrusive contacts and igneous banding (including graded layers). Numerical modeling (Chenevez et al., 1998) has supported field evidence for actively upwelling mantle and has shown that whether the magma was fed from the top or the bottom is of little consequence for the shape of the Oman magma chamber. The shape of the chamber is mainly controlled by the thermal boundary conditions.

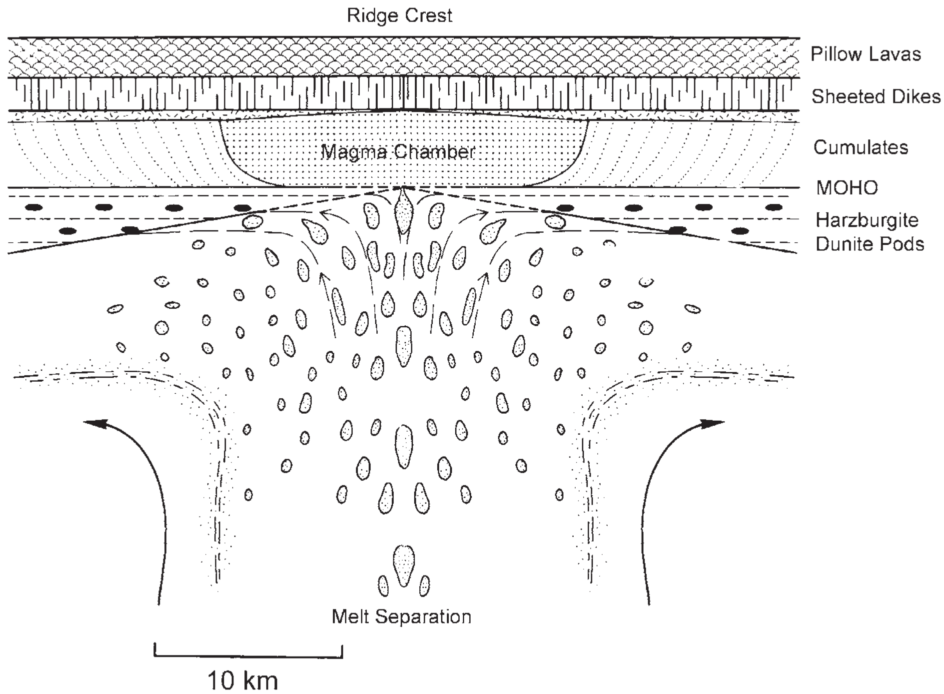

Figure 20. Schematic cross-section of the northern Oman spreading center (from Lippard et al., 1986). Reproduced by permission of the Geological Society of London.

The most recent development in magma chamber models signals the disappearance of the classic spreading center magma chambers from our conceptual thinking (Fig. 24). Kelemen et al. (1997) studied the petrography and geochemistry of gabbroic sills at the crust-mantle transition zone in the Oman ophiolite and extended the observation of Boudier et al. (1996) to the main crustal section. They proposed that primitive and residual melts migrated upward through a combination of porous flow and hydrofracturing to form the upper gabbros, the sheeted dike complex, and the lava flows (Fig. 24). The upper gabbros appear to have crystallized largely in situ (cf., MacLeod and Yaouancq, 2000), without contributing to the construction of the sheeted dike complex. The melt migration is seen in these recent models as controlled by porous flow, permeability barriers, and hydrofracturing (Korenaga and Kelemen, 1997, 1998; Kelemen and Aharonov, 1998).

DISCUSSION

Pre–Plate Tectonic Models on Oceanic Magma Chambers

Our earliest understanding of the nature of modern oceanic crust came mostly from studies of ocean islands. The rocks of these islands were found to be compositionally similar to flood basalts and other continental basaltic eruptions (Daly, 1933). The submerged ocean ridges were sampled for the first time in 1947 on an expedition to the Mid-Atlantic Ridge led by Maurice Ewing of the Woods Hole Oceanographic Institution. A petrographic examination of the dredged lavas, gabbros, and serpentinite rocks confirmed the relatively global uniformity of basalts (Shand, 1949). During the 1960s, the collection of dredge samples significantly increased our knowledge of the petrology of the ocean floor (e.g., Muir and Tilley, 1964; Christensen and Salisbury, 1975; Fox and Stroup, 1981). This advance made it possible to show that the oceanic crust of gabbroic and peridotitic compositions was associated with escarpments on the seafloor, where deeper parts of the lithosphere were interpreted to have been uplifted to the surface by transform faulting.

Marine gravimetric and seismic surveys had established very early that the earth's mantle was composed of peridotite (Hess, 1955). The oceanic crust was assumed to be composed of gabbros and lavas with a thin veneer of marine sediments (Daly, 1942). It was also well understood that melting of peridotite would produce basalts and gabbros, and that pressure release could result in decompressional melting of peridotite (Bowen, 1928; Daly, 1933; Barth, 1962). Daly (1933) believed that the lower part of the gabbroic crust and the mantle peridotites was vitreous or glassy. The mid-ocean ridges were seen as compressional zones, where the oceanic crust was bent downward into the vitreous layer, causing melting. "If…the depressed crust was broken into blocks, forced away from one another, the rising melt would occupy large chambers within the broken crust and then solidify there as crystalline peridotite or its derivatives" (Daly, 1942, p. 96).

The idea that the newly formed oceanic crust moved apart at mid-ocean ridges can be attributed to Hess (1962) and Dietz (1961, 1962) both of whom proposed that upwelling mantle beneath the ridges would laterally spread away. In contrast to his

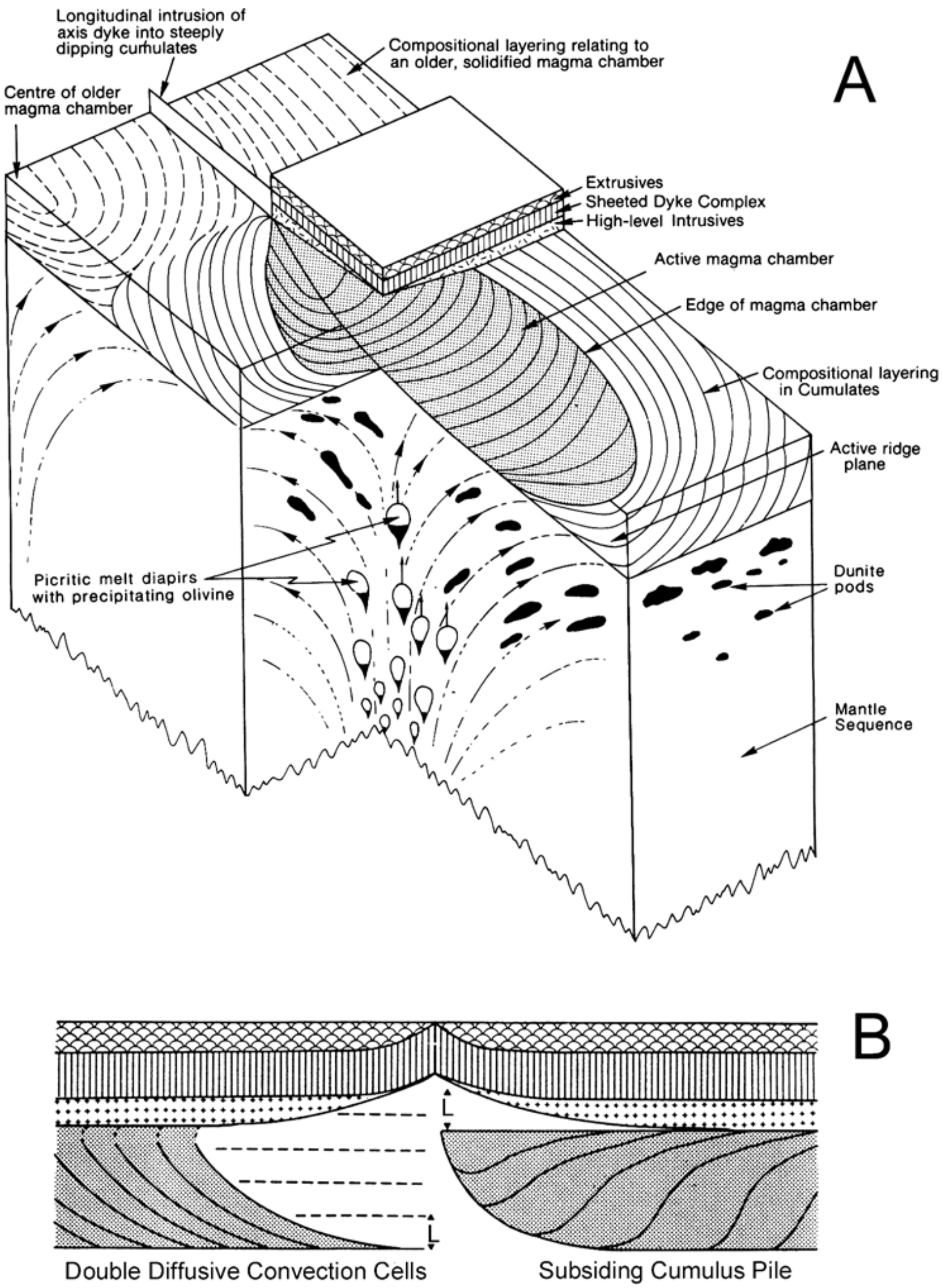

Figure 21. The Oman spreading center and axial magma chamber. After Browning (1984). Reproduced by permission of the Geological Society of London. A: Block diagram illustrating the spreading center and the overlying magma chamber (from Gass, 1989, Fig. 5). B: Cross-section through an axial chamber, illustrating alternative models to explain the calculated magma column heights: (1) double diffusive convection and (2) subsiding cumulate pile (Browning, 1984, Fig. 5).

earlier views, Hess (1962) assumed that the crust was composed of serpentinites produced by high-level hydration of peridotite. Although it was known for a long time that the oceanic crust was basaltic in composition and that melting of the mantle peridotite would produce basalt (Daly, 1914; Bowen, 1928; Hess, 1955; Erwing and Erwing, 1959; Barth, 1962; Holmes, 1965; MacGregor, 1967; Green, 1968), basalts or gabbros were not mentioned in the ocean-spreading model of Hess (1962). The view of Harry Hess that oceanic crust was composed of serpentinites was so influential at the time that when in 1962 Bruce Heezen summarized the results of seismic studies of the Mid-Atlantic Ridge, he based his interpretation on Hess' views and attributed the anomalous crest seismicity to serpentinization above the upwelling limbs of mantle convection cells (Fig. 25). Although it was accepted that melting of peridotite would produce basalts, the connection between the seismically distinct oceanic layer (layer 3) and peridotite melting was not made until much later. Alpine peridotites, serpentinites, and gabbros were thought to have first been intruded into their hosts as crystal mushes (the Alpine Mafic Magma Stem of Thayer, 1967), and then to have

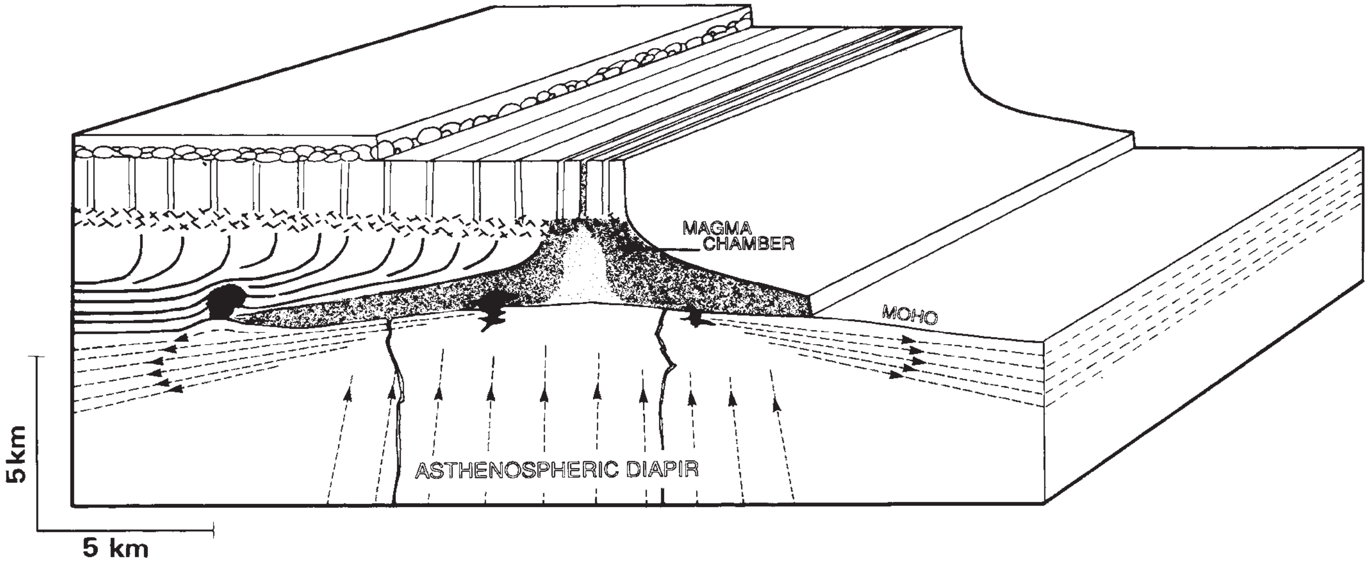

Figure 22. Magma chamber model based on structural studies of the Oman ophiolite (Nicolas et al., 1988, Fig. 10). Reproduced by permission of Elsevier.

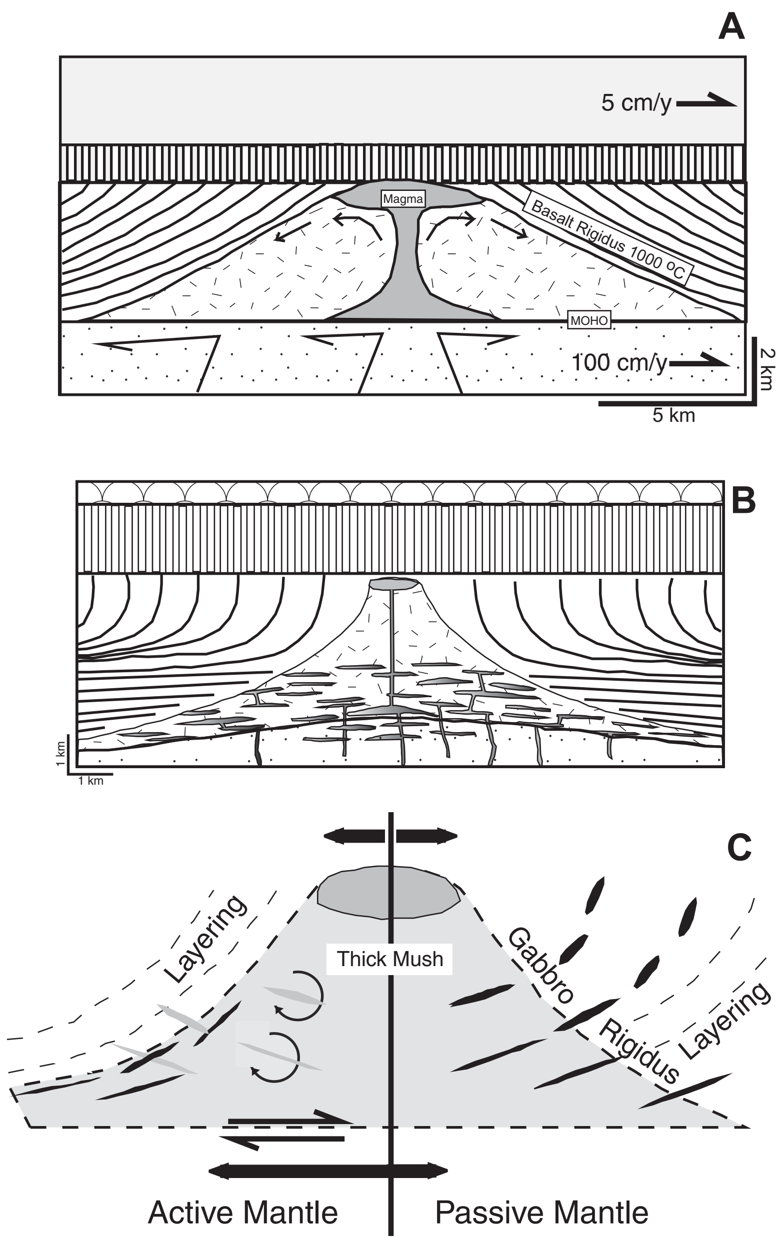

Figure 23. Models of Oman magma chambers. A: After Nicolas (1992, Fig. 15). Modified by permission of Oxford University Press. B: After Boudier et al. (1996, Fig. 7). Modified by permission of Elsevier. C: Illustration of the effects of active versus passive magma upwelling on subaxial magma chamber and gabbro banding (Nicolas, 1994, Fig. 1). Modified by permission of American Geophysical Union.

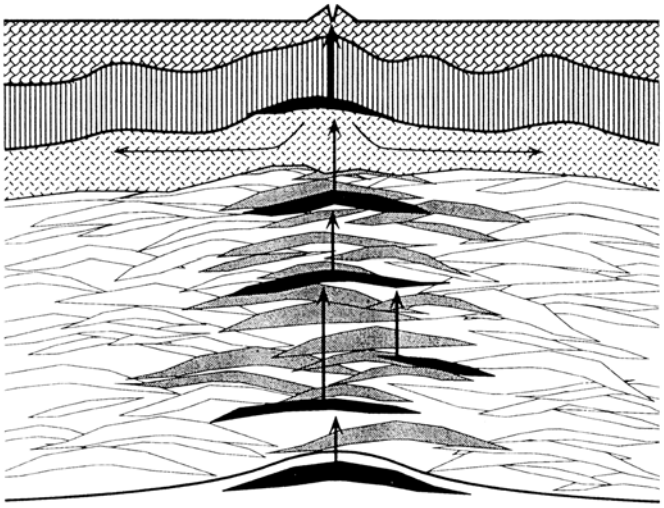

Figure 24. Schematic illustration of sill complex formation in the lower crustal section of the Oman crust. (After Kelemen et al., 1997, Fig. 6). Reproduced by permission of Elsevier.

been tectonized. Hence, tectonized peridotitic complexes in the continental crust were not always interpreted as being slices of oceanic mantle rocks.

Ophiolites had minor role, if any, in the early formulation of the history of the ocean floor during the 1950s and 1960s. They were universally considered to be thick lava flows erupted into subsiding marginal basins (geosynclines; see also Moores, this volume, Chapter 2; Shallo and Dilek, this volume, Chapter 20; Smith and Rassios, this volume, Chapter 19). This is perhaps best illustrated by Kay (1951) in a paleogeographical section of the Caledonian geosyncline of eastern North America (Fig. 26). The basaltic magma was viewed as having been injected from an abyssal substratum as a result of geosynclinal down warping (Daly, 1914, 1942). The presence of sedimentary rocks associated with many ophiolitic complexes was taken to confirm the suggestion that ophiolitic complexes were intruded into developing sedimentary basins. Thus, these views saw ophiolitic complexes as having formed from large intrusive laccoliths or sill intrusions emplaced into subsiding sedimentary basins (Steinmann, 1927, and this volume, Chapter 6). The ideas that subsequently dominated the evolving ophiolite concept considered ophiolites as products of fissure eruption-generated flood basalts (e.g., Aubouin, 1965). The internal differentiation in ophiolites was seen as having occurred in situ by crystal accumulation in large basaltic outpours (Gass and Masson-Smith, 1963; Aubouin, 1965).

Tectonized harzburgites in many ophiolite complexes were observed to have been intruded by ultramafic and gabbroic plutons that were part of these ophiolite sequences. This observation clearly was a problem for the traditional view of a eugeosynclinal eruptive environment suggested by earlier field studies in ophiolites. Ophiolite researchers dealt with this problem in various ways. In a paper published relatively late in 1969, Reinhardt suggested that the Oman ophiolite was emplaced through tectonized peridotite and onto the seafloor at a mid-ocean ridge. Gass and Masson-Smith (1963), perhaps having completed their manuscript just before the publication of Harry Hess' influential 1962 paper, considered the Troodos ophiolite as an oceanic pond of ultramafic magma. Moores (1969) interpreted the Vourinos ophiolite, northern Greece, as being a result of partial fusion of lherzolite. A reconstructed Vourinos cross-section showed ultramafic to mafic cumulates as a stratiform complex nested within a structural trough in the lherzolite tectonites (Fig. 4C; Moores, 1969, and this volume, Chapter 2). In a 1967 review paper, Ian Gass suggested that low-pressure partial fusion of the peridotite produced a liquid fraction of tholeiitic composition. Without quoting any of the papers supporting ocean-spreading models, Gass (1967) pointed out that it was possible that the Troodos formed as a "volcanic edifice" at a Tethyan median ridge, similar to the situation in modern ocean basins. Gass cited several of the early ocean-spreading and continental drift papers, published since 1958, for the first time in his 1968 paper, suggesting that the Troodos complex likely represented a slice of a Mesozoic ocean floor.

It is clear from these early developments that ophiolitic studies had little direct influence on the formulation of ocean floor

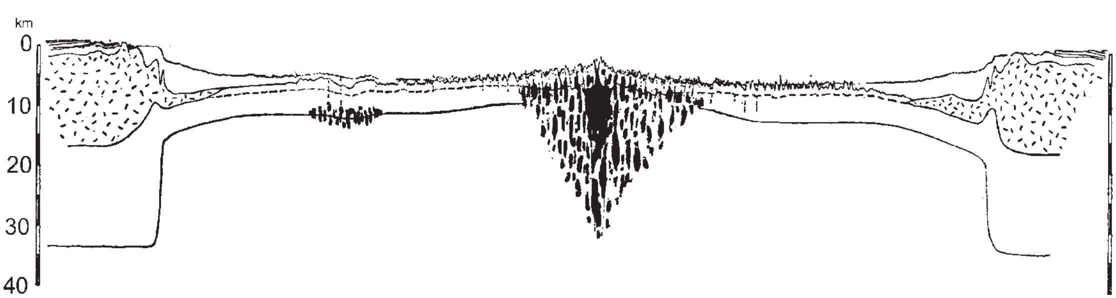

Figure 25. Cross-section through the Atlantic Ocean basin (after Heezen, 1962, Fig. 23). The dark shaded areas and dots are the 7.2 to 7.4 km/s material beneath the Mid-Atlantic Ridge. This material is interpreted as serpentinized upper mantle. The figure is constructed to conform to Hess' (1962) view of mantle convection and serpentinite composition of the lower crust (layer 3). Reproduced by permission of Elsevier.

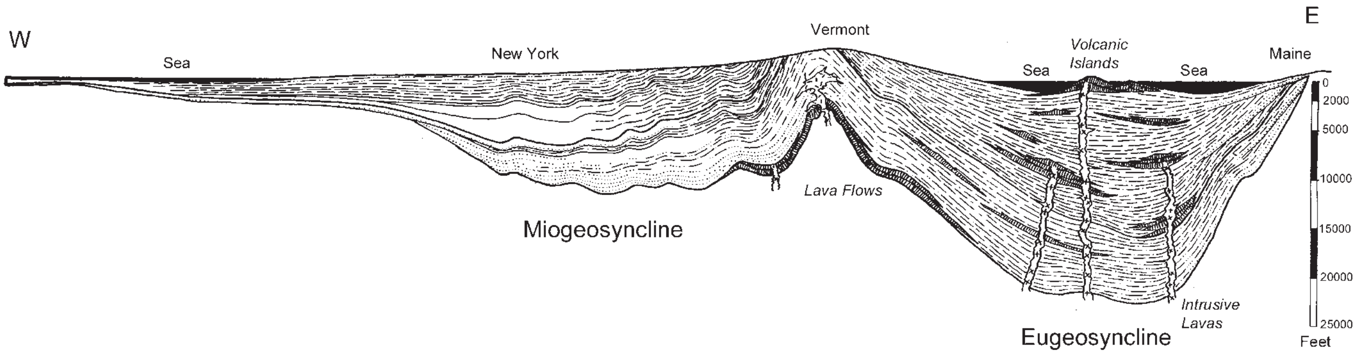

Figure 26. Paleogeographic reconstruction of the Maine (east) to New York (west) geosynclinal basins (after Kay, 1951, plate 9).

spreading models. Neither Dietz (1961, 1962) nor Hess (1962) based their theories on ophiolite studies for the simple reason that ophiolites were not seen as oceanic in origin. Similarly, ophiolite researchers were late in claiming the oceanic origin of ophiolite complexes. This concept emerged first in the late 1960s.

Development of Steady-State Magma Chambers

The plate tectonic theory was formulated during the late 1960s, principally on the basis of geophysical evidence (cf., Glen, 1982). Ophiolite studies had little direct impact on the development of plate tectonic ideas. Oxburgh and Turcotte (1968) modeled the thermal structure at mid-ocean ridges to be result of a convecting mantle due to thermal conduction. They proposed that basaltic melt was separated from upwelling zones of partially molten peridotite beneath the ridge crest, and that it was then either erupted on the seafloor or accumulated in magma chambers situated at neutral buoyancy levels within the crust and/or in the upper mantle. A consequence of this phenomenon was that a basaltic oceanic layer 3 was being continuously generated at a mid-ocean ridge and then laterally transported away.

In 1968, Cann proposed a model for the structure of oceanic crust that did not include lower crustal magma chambers. Only two years later, however, Cann (1970) proposed a "new model for the structure of ocean crust," now involving the presence of lower crustal magma chambers and gabbros. Cann illustrated this concept of ocean floor spreading by a remarkably simple and elegant drawing that in a few strokes gave the fundamental aspects of mid-ocean ridge spreading and subaxial magma chambers (Fig. 27A). It is quite illuminating to observe the impact of ophiolite studies on oceanic crust models during this time, as reflected in the three papers published by Joe Cann between 1968 and 1974. His 1968 paper contains no references to ophiolites. In the 1970 paper, mentioned above, the discussion of ophiolites is restricted to a short paragraph stating the similarities between the ocean crust model based on geophysical and kinematic reasoning, and observations from ophiolites. "The fact that (ophiolitic) complexes very much like the (ocean crust) model exist reinforces both the model and the interpretation of ophiolite complexes as pieces of ocean crust" (Cann, 1970, p. 930). In a following paper in 1974, Cann proposed an elaborate model for the genesis of oceanic crust involving an axial magma chamber. This chamber was seen with cooling along the roof to form upper isotropic gabbros and lower cumulates with layers dipping toward the chamber center (Fig. 27B). A three-dimensional version of this chamber model was subsequently developed by Robson and Cann (1982) (Fig. 27C). This chamber model was very influential and contained the key elements used in many later studies of ophiolites and modern oceanic crust (Kusznir and Bott, 1976; Pallister and Hopson, 1981; Casey and Karson, 1981; Smewing, 1981; Browning, 1984). The principal difference between the oceanic model of Cann (1974) and that developed for the Troodos ophiolite by Greenbaum (1972) is, respectively, the inward dipping versus the horizontal layering in the lower cumulates. Greenbaum (1972) assumed a zoned chamber, whereas Cann (1974) assumed magma convection.

Cann's (1974) oceanic crust model, principally derived from marine geological and geophysical information, was used to infer the origin of ophiolites, but ophiolite-derived models were not readily accepted by marine geologists and geophysicists to explain the origin of modern oceanic crust. Cann (1974) pointed out that although ophiolites contained all the fundamental elements inferred for modern ocean crust, it would not be possible to determine the origin of ophiolites without documenting the dip of slumping directions in layered gabbros, dike chilling statistics, dip of lava flows, paleo-horizontal, and orientation and kinematics of conjugate faults, among other key elements. Christensen and Salisbury (1975) expressed a similar view in a survey of the modern lower oceanic crust, concluding that ophiolites could not be fossil remnants of normal oceanic crust. They rejected the ophiolitic view that ocean crust would form in a narrow zone under the median valley of a mid-ocean ridge, as expressed by Greenbaum (1972). Instead, they proposed that the lower levels of layer 3 are thickened under the ridge flanks by

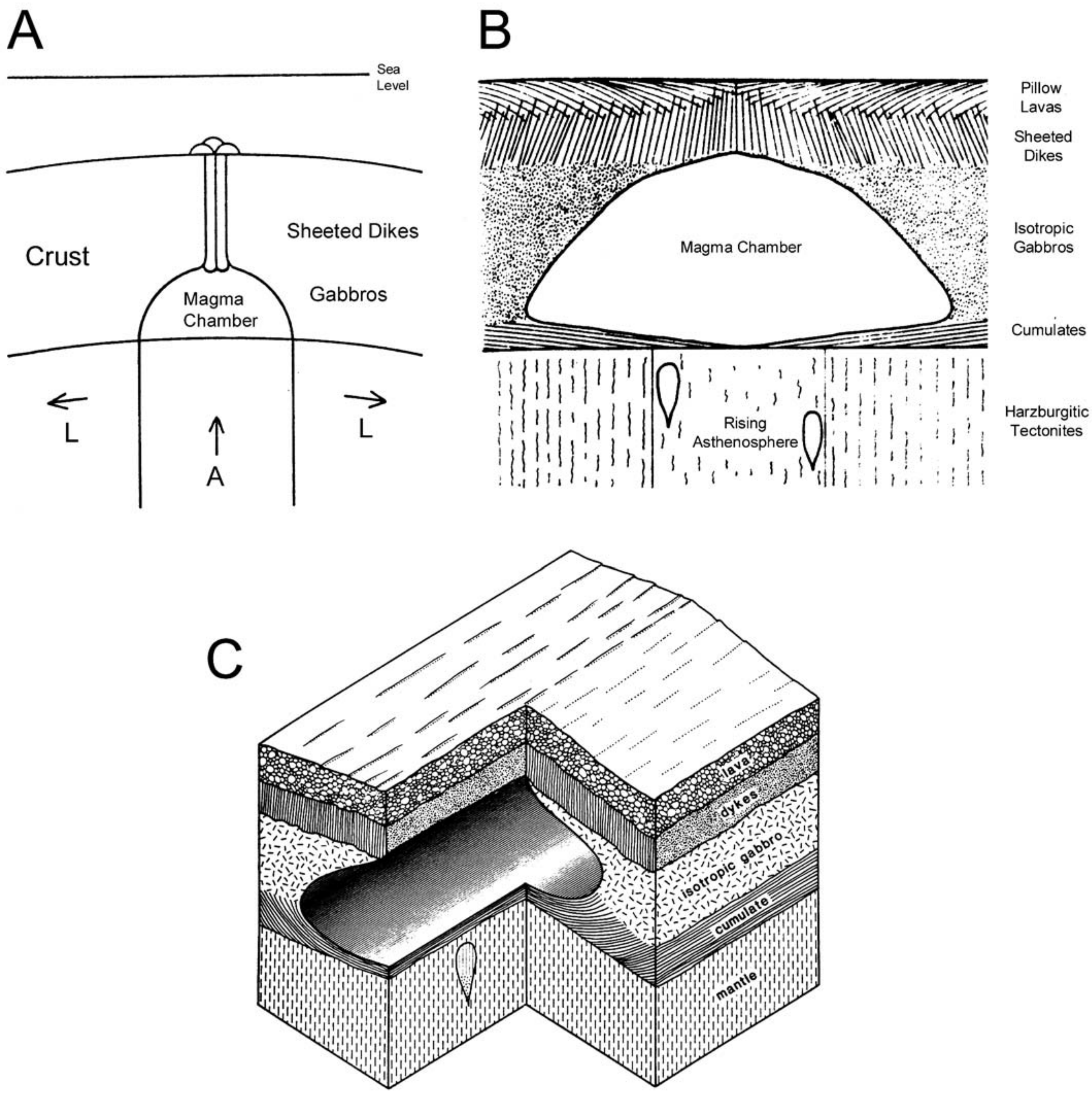

Figure 27. The ocean crust according to Joe Cann. A: Schematic sketch published in 1970 (Fig. 2) Reproduced by permission of *Nature*. B: Schematic cross-section of a spreading center chamber (Cann, 1974, Fig. 6). The illustration was used by Pallister and Hopson (1981) to develop their chamber model for the Oman ophiolite. Reproduced by permission of Royal Astronomical Society. C: Spreading chamber model based on ophiolite studies (Robson and Cann, 1982, Fig. 1). Reproduced by permission of Elsevier.

gabbro intrusions added to the crust from the thick anomalous mantle, inferred to exist beneath the ridge areas. The ophiolitic layer 3 was attributed, in this model, to development of immature oceanic crust lacking the effect of off-axis thickening.

The observations from ophiolites might thus be interpreted in the light of observations from the ocean floors, but they would not necessarily serve to formulate models for normal oceanic crust (Christensen and Salisbury, 1975). The reason for this thought was that ophiolites might have formed either in an off-axis setting or in an unusual tectonic setting, such as back-arc basins. Geochemical studies of extrusive rocks in various ophiolite complexes have strongly substantiated this view and led to the formulation of the supra-subduction zone concept for the origin of ophiolites (e.g., Pearce, this volume, Chapter 15; Taylor and Nesbitt, 1988; Elthon, 1991). Although this concept certainly has a strong validity, it is not clear how magma chamber processes in supra-subduction zone environments would differ from those in mid-ocean ridge environments. It is expected that it principally would be active spreading, with its specific spreading rates and its interplay of tectonic and magmatic activities, that will control magma chamber formation and evolution, and not the presence or absence of an underlying subduction zone.

Kusznir and Bott (1976) used thermal and heat balance calculations to model magma chamber size and shape as a function of spreading rate (Fig. 28A). They assumed that layer 2 was cooled by water circulation and that layer 3 was formed from magma solidifying to gabbro as a result of thermal conduction. Cumulates forming along the floor would narrow the magma chamber and would result in a relatively flat roof. They further showed that the chamber width was a function of spreading rate (at 3 cm/yr, a total width of 40 km was predicted). For spreading rates below 0.5 cm/yr, the heat flux was too fast to develop cham-

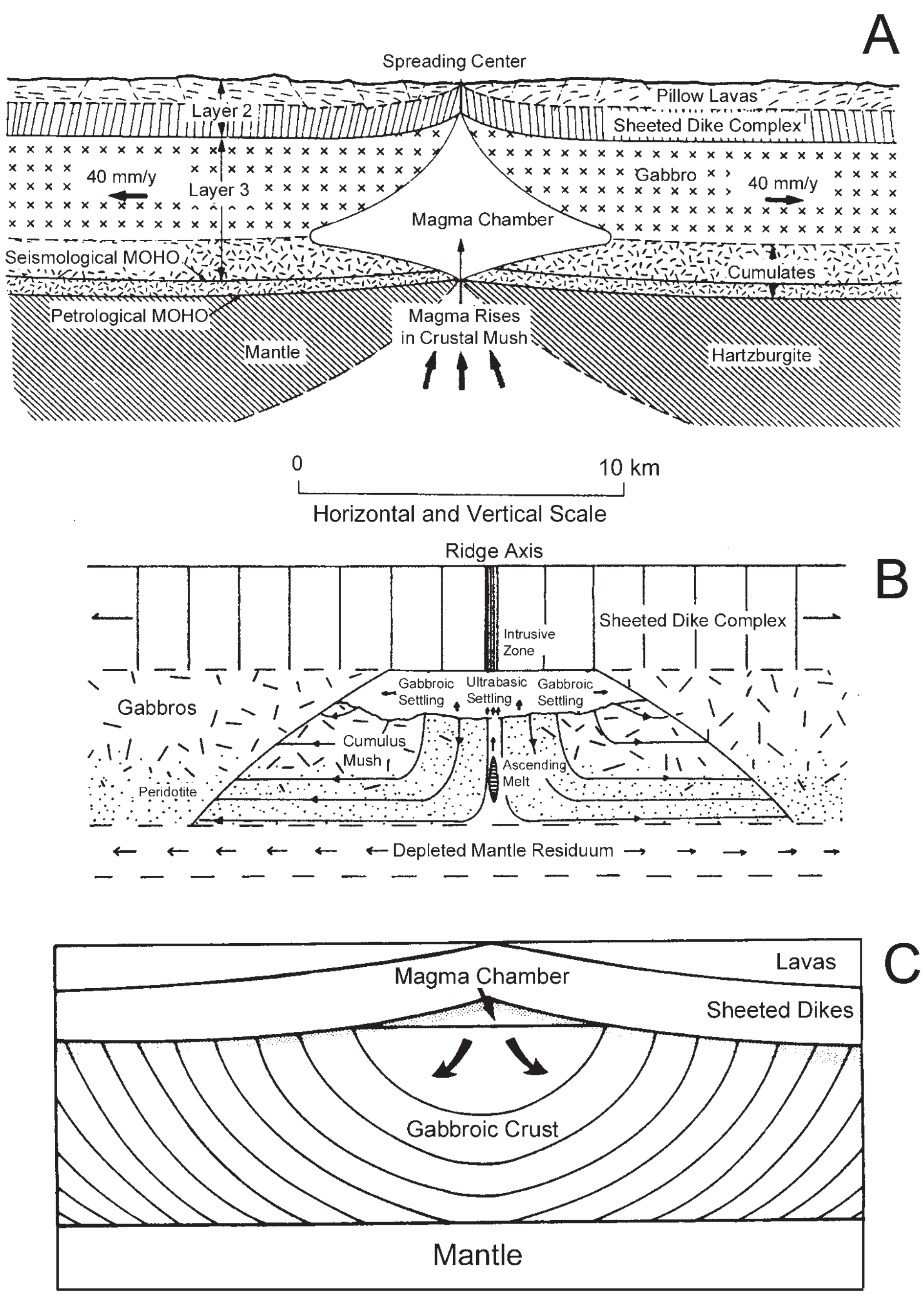

Figure 28. Thermal models for the constitution of ocean crust. A: Kusznir and Bott (1976) as illustrated by Bott (1982, Fig. 3.22). B: Sleep (1975, Fig. 4). Reproduced by permission of American Geophysical Union. C: Henstock et al. (1993, Fig. 6b). Reproduced by permission of American Geophysical Union.

bers and, instead, magma would solidify as dikes. The model of Kusznir and Bott (1976) shows similarities to the one proposed by Cann (1974), by reflecting a decreasing cooling rate toward the accretionary boundary. Other chamber models have been proposed more recently that differ mainly in their relative sizes and shapes (e.g., Lewis, 1983; Wilson et al., 1988).

In a similar fashion, but using heat budget constraints, Sleep (1975, 1978) deduced the size and shape of mid-ocean ridge magma chambers for relatively fast-spreading rates. Sleep concluded that magma chambers exist beneath the axial zone as a thin melt lens just below the sheeted dike complex, and that they extend downward into the mantle as a crystal-liquid mush (Fig. 28B). Fresh magma would have to penetrate this mush in order to enter the chamber, which is laterally zoned with ultramafic cumulates forming in the center and gabbroic cumulates forming near its margins, as a consequence of cooling and crystal settling. Cumulates formed on the floor of the melt lens would subside and flow laterally such that the centrally formed ultramafic cumulates eventually would constitute the base of the crust (Fig. 28B). The horizontal layering of the cumulate section would be dipping toward the axial rift as a result of differential flow outward from the axis, as had been proposed previously for oceanic crust (Cann, 1974; Kusznir and Bott, 1976; Dewey and Kidd, 1977). This explanation contrasted with the dominating view from ophiolite studies at the time, which suggested the occurrence of large magma chambers with little mush and with layering dipping toward the axial rift without strong indications for flow deformation (Smewing, 1981; Pallister and Hopson, 1981). Despite this major difference, it was the general geometry of magma chambers beneath the oceanic spreading centers, as proposed by Sleep (1975), that was adopted in the late 1980s by Adolphe Nicolas for the Oman ophiolite, with some important modifications necessitated by detailed petrographic and structural observations (Nicolas et al., 1988; Nicolas, 1992; Boudier et al., 1996). On the basis of detailed structural mapping, Nicolas et al. (1988) observed that the layering in the gabbros was dipping away from the inferred spreading center in the Oman ophiolite. This geometry with cumulate layering dipping away from the spreading axis had not been reported from the Oman ophiolite before, or from any other ophiolite, for that matter.

Only to a certain extent have ophiolite studies served as a test for the theoretical and geophysical modeling of oceanic crust. On the other hand, it is fair to say that many, if not all, ophiolite studies that present magma chamber models extended these ideas also to oceanic spreading centers (Greenbaum, 1972; Allen, 1975; Casey and Karson, 1981; Glennie et al., 1974; Pallister and Hopson, 1981; Smewing, 1981; Browning, 1984; Nicolas et al., 1988; Nicolas, 1992; Boudier et al., 1996). There are several outstanding studies that developed spreading center models based mainly on the observations derived from ophiolite studies. Most notably, Kidd (1977) used field data from the Troodos ophiolite to propose a crustal construction model involving a narrow zone of dike injection at a ridge axis, subsidence and faulting of accumulating lavas, and rotation of dikes and lava flows. The magma chamber envisioned by Kidd (1977) was narrow and had very steep contacts (<80°), such that its shape approached that of a downward broadening dike. While this model integrated the existing observations from the Troodos ophiolite effectively, it was also used successfully to discuss some oceanic features, such as hydrothermal convection cells, seafloor metamorphism, and magnetic reversals. Dewey and Kidd (1977) developed a comprehensive kinematic model for the Bay of Islands ophiolite. This model involved a subsiding floor of a centrally located chamber. Despite the simplicity and beauty of the model, it was rarely applied in subsequent studies, not even to the Bay of Islands ophiolite complex (Casey and Karson, 1981; Elthon et al., 1982).

The large, steady-state chamber idea was the dominating conceptual model for magma chambers beneath oceanic spreading centers until the early 1990s. Typically, it was based on the model of Cann (1970), and was in general applied to Troodos (Greenbaum, 1972), Bay of Islands (Casey and Karson, 1981), and Oman (Pallister and Hopson, 1981; Smewing, 1981) ophiolites. Suggestions involving the subsidence of magma chamber floor and flow in crystal mushes (Dewey and Kidd, 1977; Sleep, 1975) were not seriously considered (e.g., Browning, 1984). Solidification of the chamber was principally seen as having proceeded by crystal accumulation on the floors and along the walls of a commonly zoned magma chamber. Despite the fact that crystal mushes were known to have existed during formation of continental layered intrusions, this possibility was not considered in the formulation of oceanic magma chambers. The reason for this omission was simply that ophiolitic cumulates were believed to have solidified in equilibrium with the main melt volume, without retainment of any trapped melt (adcumulates; Pallister and Hopson, 1981).

Magma Lenses and Crystal Mushes

The conundrum between the two opposite views—the presence of large, steady-state magma chambers beneath the axial centers of mid-ocean ridges, as predicted by ophiolite studies, and the absence of such magma chambers, as suggested by the geophysical studies—was not easily resolved. One important reason was the strong influence among ophiolite researchers of early magma chamber solidification models developed from studies of layered continental intrusions. Seismic studies of the fast-spreading East Pacific Rise had shown the presence of small, shallow magma pockets and low-velocity zones along ridge segments (Fig. 29A) (Harding et al., 1989; Toomey et al., 1990; Vera et al., 1990; Kent et al., 1990, 1993). The existing evidence for axial magma chambers at fast-spreading centers has been reviewed by Sinton and Detrick (1992), Henstock et al. (1993), Quick and Denlinger (1993), Phipps Morgan and Chen (1993), and Phipps Morgan et al. (1994), among many others. The image that has emerged from these studies is illustrated in Figure 29B (Sinton and Detrick, 1992; Kent et al., 1990, 1993). The seismic results have suggested magma chambers beneath fast-spreading centers to be mainly filled with crystal mushes containing only

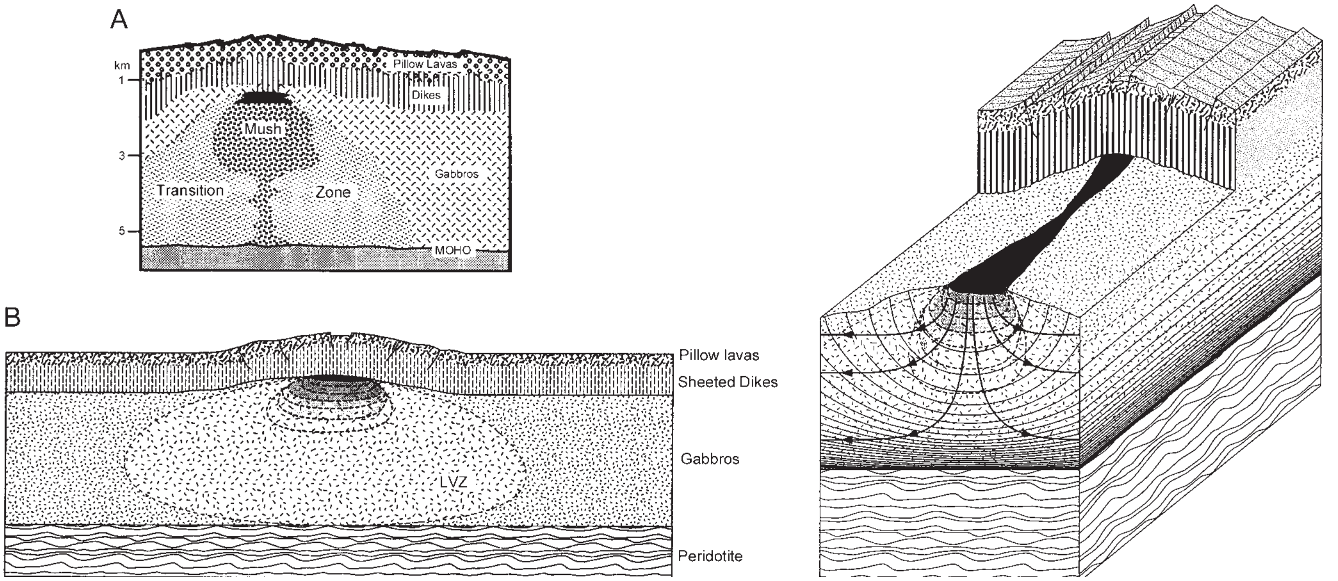

Figure 29. Geophysical models of ocean spreading center magma chambers. A: Sinton and Detrick (1992, Fig. 9). Reproduced by permission of American Geophysical Union. B: Phipps Morgan et al. (1994, Fig. 1b). Reproduced by permission of Elsevier. C: Kent et al. (1990) and Kent et al. (1993, Fig. 21). Reproduced by permission of American Geophysical Union.

small percentages of melt extending to the base of the crust. The large molten chambers, inferred from ophiolite studies, have thus vanished and have been replaced by chambers filled with crystal mushes with a thin melt lens (or sill) on top of the mush zone (Fig. 29C). Despite the fact that this view was anticipated by Sleep (1975) and Dewey and Kidd (1977), it was the work of Nicolas et al. (1988) that introduced and incorporated these ideas effectively into ophiolite studies (Nicolas, 1992, 1994; Boudier et al. 1996). MacLeod and Yaouancq (2000) have demonstrated the existence of a ferrobasaltic fossil magma chamber as a thin melt lens overlying the cumulates and perched just beneath the sheeted dike complex in the Oman ophiolite. They inferred that this lens was fed by melt seeping up from below through the underlying mush zone. The most important implication of the mush interpretation for the ophiolite concept has been that layering could have formed as a result of crustal strain, and not from direct deposition or crystallization on chamber walls.

Lately, these ophiolite studies-based views on magma chambers have been extended to continental layered intrusions. It has been documented convincingly that some of the classic layered intrusions, such as Skaergaard, consolidated from a mush by compaction and migration of interstitial melt toward an upper zone of largely molten chamber (Nicolas, 1992; McBirney, 1995; McBirney and Nicolas, 1997). It is worth noting at this point that crystal mushes should be viewed as part of a magma chamber, and that processes during their solidification are fundamental for the mode and nature of magma crystallization. Thus, there is no foundation for the view that studies of continental layered intrusions have outlasted their relevance for understanding the magma chamber processes at spreading centers (e.g., Sinton and Detrick, 1992). The existence of mushes has long been recognized in layered intrusions (Wager and Brown, 1967). The models for large, completely molten magma chambers beneath spreading centers were the outcome of marine studies of oceanic spreading centers and were readily adopted by the students of ophiolites.

Several important questions remain to be answered. How do fresh pulses of magma migrate through the mantle-crust transition and through the crystal mush, and to what extent does this migration affect the magma composition? More than anything, it has been the deep drilling into in situ oceanic crust near active spreading centers undertaken by various projects of the Ocean Drilling Program that has contributed some partial answers to these questions.

Slow-Spreading Magma Chambers

In contrast to the studies of the fast-spreading East Pacific Rise, similar studies of the slow-spreading Mid-Atlantic Ridge has provided little evidence for the presence of low-velocity zones and magma chambers (Purdy and Detrick, 1986). This difference in the amount of detectable melt fractions as a function of spreading rate is surprising, considering the similarities in the overall seismic structures (Henstock et al., 1993). The geophysical observations were supported by theoretical modeling that showed that steady-state magma chambers did not, and could not, exist beneath spreading centers with slow-spreading rates (Sleep, 1975). This observation was often in conflict with petrographic and geochemical studies of the rocks recovered from the slow-spreading centers, such as the Mid-Atlantic Ridge system (~1 cm/yr), which was assumed to have a central, steady-state magma chamber under the inner floor of its median valley (e.g., Bryan and Moore, 1977).

The reason for the difference between fast- and slow-spreading centers is most likely due to the dynamics of slow-spreading ridges, which make them more directly controlled by the availability and arrival of intermittent batches of magma from the mantle (Sinton and Detrick, 1992). A wide range of processes that are locally determined by thermal, magmatic, and tectonic regimes thus might control the crystallization and the evolution of magma chambers at slow-spreading centers. Nisbet and Fowler (1978) suggested, on the basis of the Mid-Atlantic Ridge seismic and petrographic data, that small melt pockets migrate upward by crack propagation and form magma chambers in a broad zone under the axial valley and at a horizon beneath the sheeted dike complex (Fig. 30A). They described the shape and the geometry of these inferred small chambers as “infinite leeks.” Other studies of the formation of oceanic crust have pointed out that magma chambers are intermittent features (Lister, 1983), and that magmatic and tectonic activities likely shift laterally and along-axis of the spreading centers (Pálmason, 1973; Ramberg and van Andel, 1977; Nisbet and Fowler, 1978). Smith and Cann (1992) used the distribution of seamounts on the Mid-Atlantic Ridge to propose that crustal accretion resulted from the “pile-up” of seamounts and lava flows. They related this eruption style to the buoyant rise of magma chambers with limited size and frequency and inferred that this phenomenon would form a gabbroic crust composed of multiple frozen chambers over a broad interval (Fig. 30B).

Such small intermittent and laterally limited chambers have certainly had their followers among the researchers of the Troodos ophiolite (Allen, 1975; Smewing et al., 1975; Varga and Moores, 1985; Thy et al., 1989; Malpas, 1990; Moores et al., 1990), but have rarely been considered for other ophiolites (i.e., Strong and Malpas, 1975).

In Situ Oceanic Crust

Our view of oceanic magma chambers has been strongly influenced by the results of deep drilling by the ODP into in situ lower oceanic crust. Of particular significance is Hole 735B, drilled ~1500 mbsf into a gabbroic massif that formed beneath the median valley of the slow-spreading Southwest Indian Ridge near the Atlantis II Fracture Zone (Dick et al., 2000; Natland and Dick, 2001). An association of evolved Fe-Ti-oxide gabbros with relatively primitive olivine gabbros characterizes the upper ~500 m of gabbros in Hole 735B. The Fe-Ti-oxide gabbros occur as the major constituent of some plutonic units or as a minor component of layers and vein networks in dominantly olivine gabbro units. The contacts between the gabbros and the Fe-Ti-oxide gabbros are, when preserved, mostly gradational and only rarely intrusive. The apparent bimodal variation in the core, indicated by all cumulus phases, suggests limited mixing and compositional re-equilibrium between the two magmatic components. This phenomenon may suggest that one cumulate component was solidified, or partially solidified, before the other component was formed or was introduced into a preexisting mush pile.

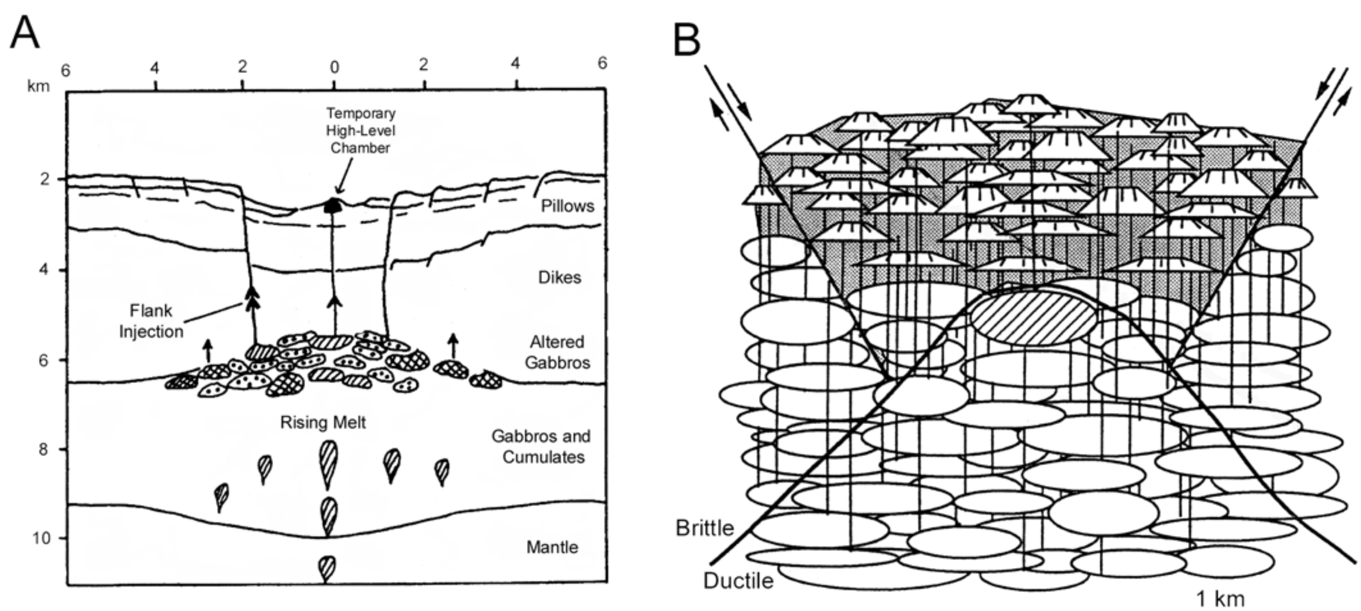

Figure 30. Magma chamber models depicted for the Mid-Atlantic Ridge. A: Multiple chamber model of Nisbet and Fowler (1978, Fig. 6b), which they referred to as the “infinite leek” model. Reproduced by permissions of the Royal Astronomical Society. B: Multiple chamber model of Smith and Cann (1992, Fig. 10). Reproduced by permission of American Geophysical Union.

Below 500 m and toward the bottom, coarse-grained olivine gabbro dominates, with subordinate amounts of Fe-Ti-oxide gabbro (Natland and Dick, 2001; Niu et al., 2002). ODP Hole 1105A was drilled in a location offset by 1 to 2 km NNW of Hole 735B, and recovered gabbro sequences quite similar to those from the upper part of Hole 735B (Shipboard Scientific Party, 1999; Thy, 2003). The important observation was the preservation in the olivine gabbro sequence of an evolved component of Fe-Ti-oxide gabbros believed to represent migrating interstitial melt. The distribution of this interstitial melt and the processes responsible for its preservation and migration have naturally been a main theme for development of our ideas for spreading-related magma chambers (e.g., Sinton and Detrick, 1992).

The compositional variation of the main cumulus phases in Hole 735B (Bloomer et al., 1991; Ozawa et al., 1991; Natland et al. 1991; Hébert et al., 1991; Niu et al., 2002) is illustrated in Figure 31. The main features shown by the cryptic variation are that the olivine gabbros are composed of at least three segments, each showing upward trends toward more primitive compositions (Niu et al., 2002). A more detailed stratigraphy has been constructed on the basis of the whole-rock composition (Natland and Dick, 2001), which has not yet been confirmed by the still limited, extant mineral composition data (Niu et al., 2002). The two upper segments of host gabbros recovered during Leg 118 are characterized by associated Fe-Ti-oxide gabbros that show steep downward evolution trends toward more evolved compositions, in the opposite direction of the upward evolution of the host gabbros (Fig. 31). It is also significant to note that the upper segments of host gabbros are relatively more primitive than the lower segment recovered during Leg 176. The segmentation of the host gabbros is in many respects similar to that seen from ophiolitic gabbros (Troodos; Thy et al., 1989) and layered intrusions (Wilson and Larsen, 1985). Such large-scale, vertical gabbro segmentation has been related to replenishment and influx of fresh magma into an existing and partially crystallized chamber. It is also possible that the segmentation results from the intrusion of multiple chambers at approximately the same crustal level.

The systematic upward increase in the amount of Fe-Ti-oxide gabbros suggests that the gabbro cells were connected, acted, and developed as one crystal mush, perhaps formed from a

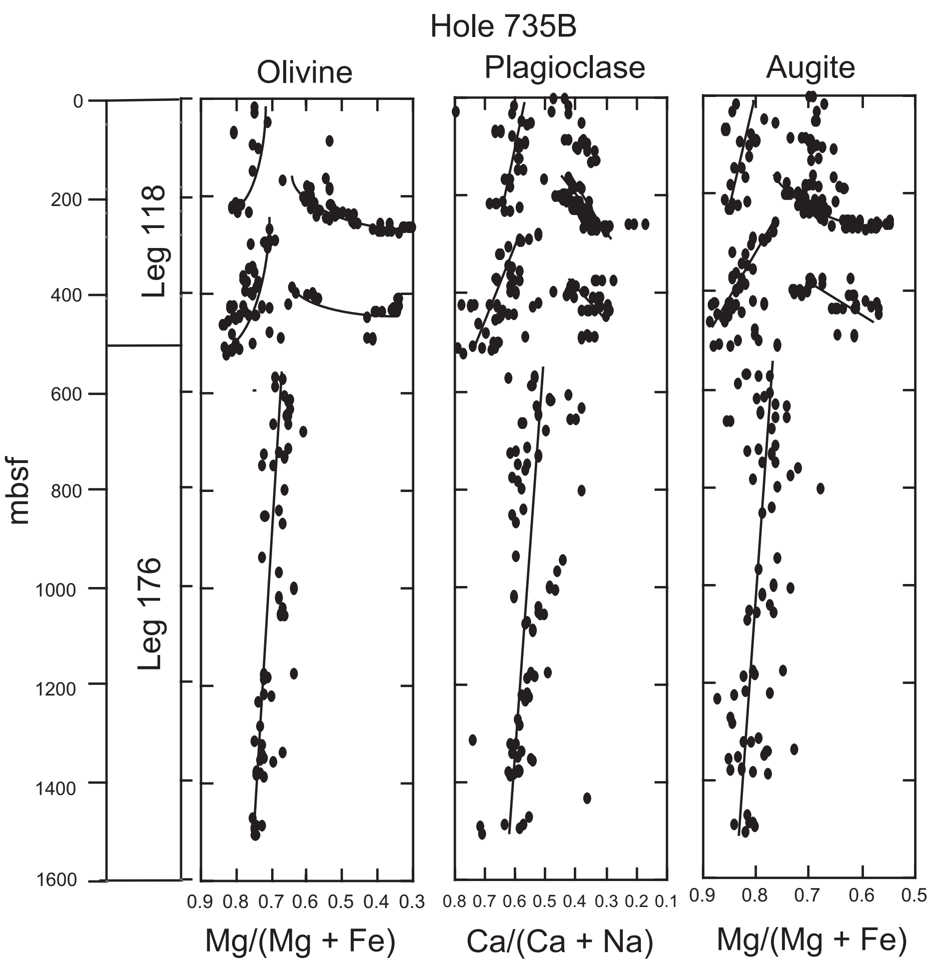

Figure 31. The cryptic variation in the composition of the main silicate minerals in the gabbros of Hole 735B (Bloomer et al., 1991; Ozawa et al., 1991; Natland et al. 1991; Hébert et al., 1991; Niu et al., 2002).

subsiding magma lens beneath the active ridge. The Fe-Ti-oxide gabbros have been inferred to represent interstitial melt that has migrated upward due to compaction, or laterally due to tectonic deformation and shear in a crystal mush. The observation that the evolved gabbro component appears to be concentrated upward in the gabbro section of Hole 735B has been interpreted to indicate that compaction was the dominating process on which the effects of ductile deformation and the lateral accumulation and flow of the interstitial melt were locally superimposed. These observations are not so different from the observations from continental layered intrusions (Thy and Dilek, 2000) and are predictable consequences of the solidification of any crystal mush pile at the bottom of a magma chamber (McBirney and Nicolas, 1997).

How the crystal mush grows and interacts with the new and fresh mantle-derived magma seeping up from beneath the chamber and entering the lower strata of the mush is an obvious and challenging question that has been addressed only most recently (e.g., Sinton and Detrick, 1992). It has been suggested from detailed field studies in the mantle-cumulate transitional zones of the Oman (Benn et al., 1988; Nicolas et al., 1988; Boudier et al., 1996) and the Bay of Islands (Bédard, 1991, 1993; Bédard and Hébert, 1996) ophiolites that new magma can form sill complexes in the crystal mush, perhaps in a similar fashion to sill intrusions in magmatically active, evolving sedimentary basins (Einsele, 1982). This concept has been developed by Bédard (1993) and more recently has been used by Kelemen et al. (1997) to suggest that the whole of the lower crust in Oman is built by a sill complex. It has also been proposed that the lower crust acts as a reactive filter modifying the residual melt migrating through the mush and sill complex, and, thus, buffering the composition of the melt that reaches the magma lens above the mush (Bédard, 1993; Bédard et al., 2000; MacLeod and Yaouancq, 2000).

These latest results have clearly demonstrated that the further development of our understanding of oceanic magma chambers has been strongly modified and improved by careful fieldwork of plutonic complexes in ophiolites accompanied by detailed petrographic and geochemical studies. This development substantiates the notion that the revival of systematic field studies in ophiolites is a forward approach to further advance our understanding of the workings of oceanic spreading centers and the related magmatic and tectonic processes during the evolution of oceanic crust.

EPILOGUE

It is tempting to incorporate near the conclusion of this review of the historical development of ophiolitic and oceanic magma chambers our own preferred idea of the "nature" of ophiolitic and oceanic magma chambers. We do this with the help of several simple diagrams depicted in Figure 32. From the outset, we would like to caution against a complete rejection of closed magma chambers and the traditional cumulus theories, and to point out the potential pitfalls for the adoption of a singular view of oceanic magma chambers. We think that this caution is well founded in studies of ophiolites and continental layered plutons. The problems with the poor resolution of seismic imaging and the difficulties of direct sampling of in situ oceanic crust contribute to our limited understanding of magma chamber processes.

The evolution of slow-spreading crust is characterized by small, ephemeral chambers that were emplaced at various crustal levels, and is controlled by localized and varied magma supply rates, sporadic episodes of tectonic extension, and possible rift axis relocations. Chambers emplaced by a single or a few pulses of magma may, therefore, be common features of slow-spreading centers. The relative volume of crystal mushes in such ephemeral magma chambers and the nature of crystallization processes in them may vary as a function of cooling rate. In many aspects, plutonic rocks that formed in such chambers may show many similarities to plutonic rocks in continental layered intrusions. Some continental layered intrusions seem to have been dominated by thick cumulate (or crystal mush) sections with significant amounts of trapped melt that were expelled slowly by compaction and migrated toward the main reservoir (Fig. 32A). Other intrusions might have been dominated by the inward migration of the crystallization front along a boundary layer, leaving behind a solid product in the bottom and along the margins of magma chambers (Fig. 32A). Similar processes can be envisioned for slow- to intermediate-spreading oceanic crust (Fig. 32B). This would account for the lack of deformation banding in some ophiolite complexes, such as in the Troodos, and for the evidence of multiple intrusive events that alternated with strong differentiation processes in the absence of steady-state equilibria. Because the actual mode of solidification of continental layered intrusions is still controversial, there is no objective reason to assume that this task should be easier for ophiolitic plutons or for oceanic magma chambers.

The high rates of magma flux at fast-spreading centers are likely to support development of large chambers with steady-state mushes (Fig. 32C) and crustal thermal gradients that are well above those observed from slow-spreading centers. It is important to emphasize at this point that because the seismically imaged melt lens is fed and maintained by the underlying mush zone, these two entities, the melt lens and the mush, should not be viewed as spatially isolated systems. The chambers underlying the intermediate- to fast-spreading centers thus may have substantial volumes of magma, far exceeding those beneath slow-spreading centers. Hence, the definition and discussion of a magma chamber should include the melt lens, as well as the surrounding and underlying mush zone.

Magma chambers beneath fast-spreading centers are likely controlled by the processes of compaction, subsidence, and near-solidus deformation in thick mush zones, which results in flow banding rather than the modally graded magmatic layering commonly observed in continental layered intrusions. At fast-spreading centers, fresh magma is continually added to the mush zone from the underlying mantle. The addition of this new magma to the mush and eventually to the melt lens above it helps the chamber attain its steady-state volume and makes the crust grow. The migration and the interaction of the fresh magma with the existing

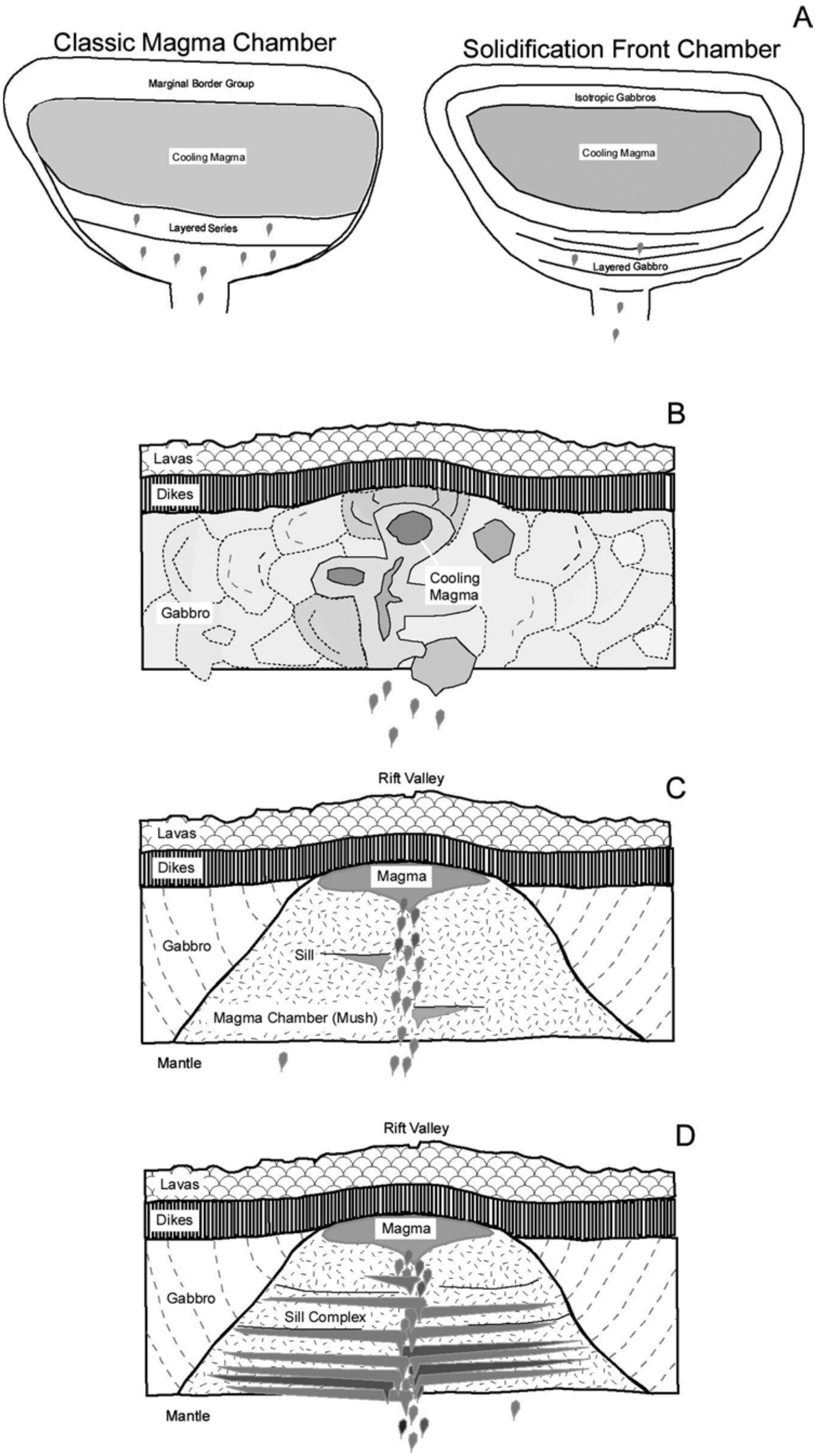

Figure 32. Tentative end-member models for magma chambers. A: Closed chamber crystallizing by either gravitational settling and formation of a crystal mush along the floor with little influx of new magma, or by crystallizing along the margins and inward without forming a mush. The latter model incorporates initial phenocrysts settling along the bottom of the chamber (after Marsh, 2000). B: Slow-spreading crust with intermittent magma supply and tectonic extension. Magma chamber formation and evolution are viewed as being very similar to what we envision for continental layered intrusions. C: Open, spreading-controlled magma chamber with thick mush on the floor and a thin magma lens on top. Compaction and migration of residual melt, as well as of fresh magma seeping up from the mantle into the mush, fill up the chamber. Local development of sills in the mush is determined by local porosity variation. D: Mush-filled, spreading-controlled magma chamber replenished with incoming fresh magma, which builds a sill complex in the mush concurrently with compaction and upward migration of the residual melt.

mush are critical processes during the evolution of fast-spreading chambers. In some cases, the new magma percolates rather freely through the existing mush while crystallizing and reacting on its way to the melt lens (Fig. 32C). In some other cases, the new influx of magma may form sill complexes in the existing mush (Fig. 32D), as a function of magma flux rate, permeability variation, and/or contrasting physical properties between the mush and the migrating magma. We see evidence for the artifacts of complex assimilation and percolation processes in some ophiolite complexes, such as in the Bay of Islands and Oman, whereas other ophiolites, e.g., Troodos, show no evidence for them.

Our three "end-member" types of magma chambers, as illustrated in Figure 32, are not always mutually exclusive. For example, whereas strong evidence for flow deformation exists in the Oman gabbros, modally graded layers more reminiscent of classic cumulus processes are also observed in these gabbros. Similarly, upward migrating ferrobasaltic melt appears to have existed in gabbros near the Atlantis Fracture Zone (Southwest Indian Ridge), that also contain evidence for magma replenishment and differentiation, as inferred for the Troodos gabbros and for some continental layered intrusions. Thus, we suggest that different aspects of these three end-members may occur in individual ophiolite complexes, depending on the mode and nature of solidification processes that dominated during their formation.

We shall always be limited in our interpretations of oceanic magma chambers because we cannot directly observe the responsible processes. We can, however, make significant advances in understanding oceanic magma chambers and processes by conducting detailed petrographic and geochemical studies of ophiolites, as well as of continental layered intrusions, combined with systematic structural field investigations. In the meantime, future deep drilling into in situ oceanic crust at and near modern spreading centers, and associated geophysical measurements and modeling, should still provide critical information, observations, and stimulating questions about oceanic crust formation and related processes.

SUMMARY

Conceptual models for oceanic magma chambers have evolved continually over the years since the first structural observations and interpretations from ophiolites and modern oceanic crust. We can examine this evolution of ideas in three overlapping time frames.

Prior to the advent of plate tectonics, mafic-ultramafic and sedimentary rock associations, which we today know as ophiolites, were considered generally as large submarine eruptions into subsiding marginal basins. The genetic connection between mantle peridotites and the commonly associated basaltic lavas, as in the Steinmann Trinity, was not yet made. This was the time when marginal basins (or geosynclines) were considered as the precursors to mountain belts.

Recognition of a sheeted dike complex in the Troodos ophiolite in the 1960s prompted the early researchers to think about seafloor spreading processes and to suggest a possible ophiolite-ocean crust analogy. The identification of harzburgites and dunites as mantle rocks, and gabbros and pyroxenites as crustal rocks in the Troodos, Bay of Islands, and Oman ophiolites (see Juteau, this volume, Chapter 3) led to the recognition of fossil magma chambers in the plutonic sections of ophiolite complexes. This major step in the formulation of the ophiolitic magma chamber concept was prompted by the newly established plate tectonics theory, which predicted that basaltic melt, separated from upwelling zones of partially molten peridotites beneath the crest of mid-ocean ridges, would accumulate in crustal magma chambers.

Early models of the structure and evolution of modern oceanic crust, based mainly on geophysical modeling, did not incorporate magma chambers and made little use of observations and ideas derived from ophiolites. The ruling geophysical interpretation of magma chambers was the classic, steady-state large chamber model that dominated the models of ocean crust up to the early 1990s. Ophiolite researchers adopted this chamber model and readily extended their ophiolite-derived interpretations of magma chambers to the oceans, suggesting various models for magma chamber evolution beneath active spreading centers throughout the 1970s and 1980s. Ophiolite-derived magma chamber models evolved from single-chamber to multiple-chamber configurations with various sizes and geometries during the 1980s to explain the thickness of gabbroic rocks and their layering, the observed fractionation patterns, and the distribution of cumulus phase assemblages.

Following the early ideas, geophysical modeling and interpretations of the late 1970s envisioned oceanic magma chambers as a thin melt lens located below the sheeted dike complex and extending downward to the mantle as a crystal-liquid mush zone. These models eventually predicted that steady-state magma chambers could not exist beneath slow-spreading centers, as thermal calculations had suggested, and that small crustal chambers beneath the axial valleys of these spreading centers were likely to be intermittent (ephemeral) in nature.

These geophysical interpretations of oceanic magma chambers were substantiated by seismic studies of the fast-spreading East Pacific Rise and the slow-spreading Mid-Atlantic Ridge in the 1990s. Whereas through these studies small, shallow magma pockets and low-velocity zones were detected along ridge segments of the East Pacific Rise, there was no indication of the presence of low-velocity zones and magma chambers beneath the Mid-Atlantic Ridge. The significance of the seafloor spreading and magma supply rates for magma chamber evolution and, hence, for oceanic crust generation was thus underscored by the mid-1990s.

Observations and interpretations from ophiolites and modern spreading centers started converging in the early 1990s, and this "two-way traffic" in the development of scientific concepts modified the magma chamber models significantly. This era in the evolution of geological thought regarding the magma chamber concept coincides with the productive attempts of the ODP to drill into in situ lower oceanic crust at or near active spreading centers.

Various projects of the ODP and the related workshops brought together the members of the ophiolite and marine geology and geophysics communities, facilitating the exchange and cross-pollination of ideas and observations on oceanic crust formation.

The new magma chamber model that emerged in the mid-1990s envisioned the magma chambers beneath fast-spreading centers to be filled mainly with crystal mush containing only small percentages of melt and to be capped by a thin melt lens perched on top of a mush zone. Thus, a magma chamber of a fast-spreading system would consist of a melt lens on top and the mush zone below that is made of a plastically deforming body of high viscosity material. The operation of mush zones during the crystallization of gabbros had long been predicted for the igneous development of continental layered intrusions; thus, the idea of magma chambers containing mush zones was not necessarily new. Core samples of gabbroic rocks recovered from deep drilling into in situ lower oceanic crust include evolved Fe-Ti-oxide gabbros representing the interstitial melt trapped in the host gabbro. Upward and lateral migration of this melt within the host gabbro network is facilitated by compaction and/or tectonic deformation occurring in the spreading system.

Detailed and systematic studies in recent years of ophiolitic plutons and plutonic complexes have highlighted the significance of syntexis processes, by which new magmas react with their host rocks, during the evolution of oceanic crust. New batches of magma seeping up from beneath the chamber form networks of sill-like bodies within the crystal mush that act as a reactive filter to the residual melt migrating upward through the mush. Thus, the composition of this melt changes as it passes through the mush zone on its way to the melt lens on top. This current thinking in melt and magma chamber evolution has been significantly improved by the revival of systematic petrographic and geochemical studies in ophiolites, and the synergy and mutual exchange between the ophiolite and marine geology and geophysics communities have helped further advance our understanding of oceanic magma chambers and related processes.

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