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MID-OCEAN RIDGES: Fine Scale Tectonic, Volcanic and Hydrothermal Processes Within the Plate Boundary Zone

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INTRODUCTION

A first order model of spreading centers as idealized linear boundaries of crustal and lithospheric generation provides only a gross understanding of global scale plate kinematics. As we attempt to understand the complexity of crustal and lithospheric structure of two thirds of the earth's surface, it is becoming increasingly necessary to study the tectonic, volcanic, and hydrothermal processes within the spreading center plate boundary zone. All oceanic crust bears the imprint of these processes. This review focuses on a few selected topics concerning the fine scale tectonics and geophysics of the active axial zone of mid-ocean ridges with reference to associated volcanic and hydrothermal processes. It draws heavily on recent studies that use deeply towed instrument packages, multi-beam bathymetric mapping, ocean bottom instruments, and ALVIN (e.g. the Famous, AMAR, RISE, and Galapagos expeditions).

We begin with a review of the large-scale structure of spreading centers. We then take a close look at the axial neovolcanic zone and progress away from the axis through the active tectonic zones. Next we consider the characteristics of the axial magma chamber and associated hydrothermal actvity, as well as the generation of magnetic anomaly "stripes" and their implications for crustal generation. One of our findings is that the initial two-dimensional model of volcanic and tectonic zones must be expanded upon to allow for variations

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along strike and for episodicity. While this is primarily a review paper, in the process of synthesis I hazard a number of suggestions and speculations, some of which are new, and most of which require further work for verification or disposal.

STRUCTURE AND TECTONICS OF THE AXIAL ZONE

Spreading rates, which vary from 1–18 cm/yr, seem to control the gross morphology of spreading centers (Figure 1; Menard 1967). At slow total opening rates of 1–5 cm/yr, a 1.5–3.0 km deep rift valley marks the axis. Rough and faulted topography created in the rift valley is largely preserved in the older ocean basin. The Mid-Atlantic Ridge (MAR) is a classic example (Figure 2; Van Andel & Bowin 1968, Loncarevic et al 1966).

At intermediate rates of 5–9 cm/yr, the rift valley is only 50–200 m deep. This shallow rift is superposed on a broad axial high, and the flanking topography is relatively smooth. The East Pacific Rise (EPR) at 21°N (RISE study area) and the Galapagos spreading center are examples (Larson 1971, Klitgord & Mudie 1974). At fast spreading rates (greater than 9 cm/yr) there is no rift valley, but a triangular-shaped axial high is observed (e.g. EPR south of 15°N, Rea 1978). The topography is relatively smooth with a fine scale horst and graben structure. A category of "ultra-fast" spreading centers has been pro-

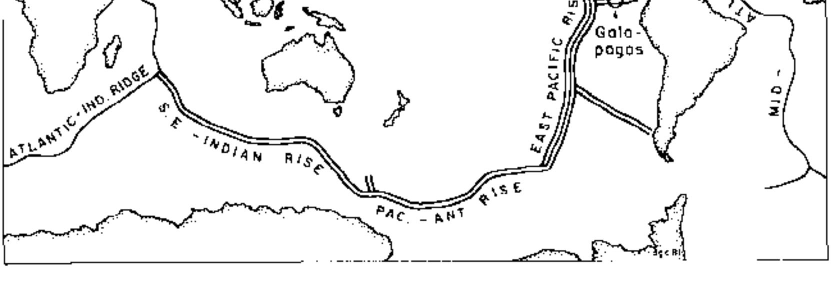


Figure 1 Major seafloor spreading centers shown schematically; transform faults, back-arc spreading centers, and subduction zones omitted. Slow spreading rates, 1.0–5.0 cm/yr; intermediate rates, 5.0–9.0 cm/yr; fast rates, 9.0–18.0 cm/yr. Mid-ocean ridge diving-expeditions indicated.

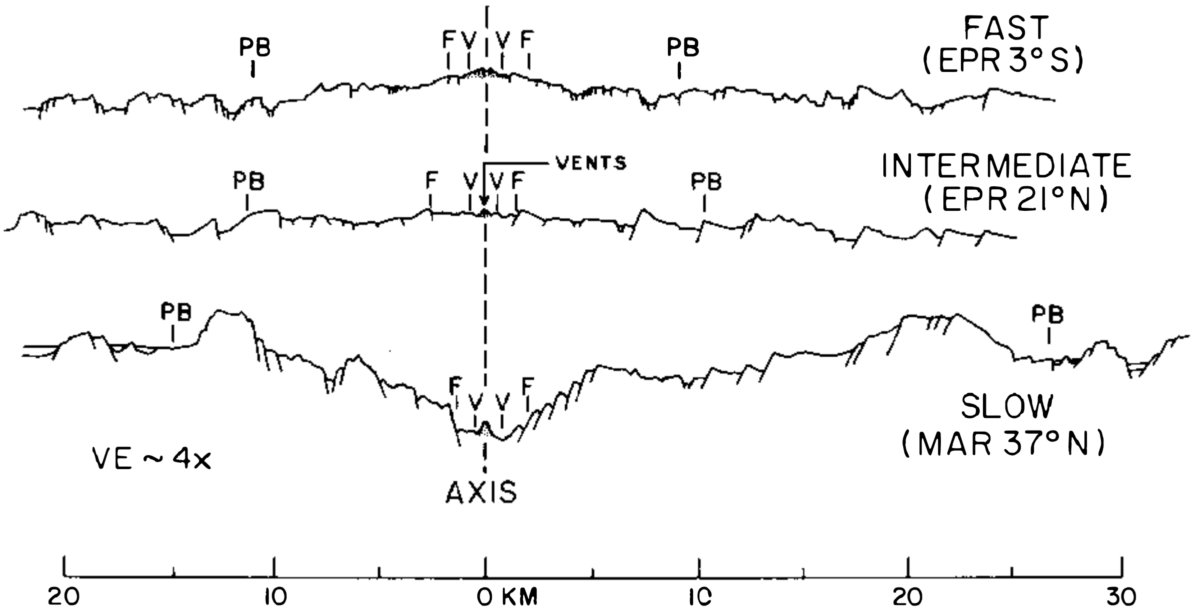


Figure 2 High resolution deep-tow profiles of mid
zone of fissuring by Fs, plate boundary zone (width
Active faulting occurs up to 10–30 km off axis. Data from Lonsdale (1977), Normark (1976), Shih (1980), Macdonald et al (1975), and Macdonald & Luyendyk (1977).

posed (Lonsdale 1977) but is omitted here as there is insufficient evidence for any significant morphologic or structural distinction between the fast and "ultra-fast" ridges.

Several hypotheses have been proposed to explain why a deep rift valley occurs at slow spreading rates but not at fast rates (an exception being the hot spot–dominated Reykjanes Ridge). One is the "hydraulic head loss" model proposed by Sleep (1969) with subsequent elaborations (Lachenbruch 1973, 1976, Sleep & Rosendahl 1979). Along the sides of an idealized conduit tapping deep magma sources, viscous forces are sufficient to cause a significant loss of hydraulic head, resulting in a topographic depression over the spreading center. To conserve energy, this head loss is regained by uplift of the rift valley walls relative to the valley floor. The head loss is directly proportional to the upwelling velocity and inversely proportional to the cube of the conduit width. A deep rift valley is present at slow spreading rates because material upwells through a narrow conduit formed by cool, old lithosphere. At faster spreading rates there is much less head loss because a wider conduit overwhelms the dependence on flow rate. The rift valley is replaced by flat topography or a crestal peak, which may be caused by the presence of a low density crustal magma chamber. Thus the cross section of fast spreading ridges closely approaches that expected from buoyancy in a relaxed state, while slow spreading ridges exhibit significant dynamic effects.

A relatively new hypothesis invokes "steady-state necking" of the lithosphere as the cause of the median valley (Tapponnier & Francheteau 1978). In this model, the rift valley is caused by necking or thinning in a ductile layer beneath the rift valley. The analogy is that of a beam plastically necking out under tensional stress. The layer does not actually break like a necking beam, because new material is added constantly from below (maintaining steady state), while the entire region is continually uplifted by buoyancy forces. At slow spreading rates, the strength of the crust in the axial zone is presumed great enough for necking to be significant, creating a rift valley. At faster rates, the young crust is too hot and weak at shallow levels for this process to be significant. Other models explain rift valleys and central peaks as caused by imbalances in the supply of new material versus crustal acceleration within the axial zone (Deffeyes 1970, Anderson & Noltimier 1973, Reid & Jackson 1981, Nelson 1981). These models are not mutually exclusive and some combination of them may apply.

Beyond these first order differences, high resolution instruments reveal that fine scale crustal structure and tectonics evolve in a pattern that is largely independent of spreading rate. This is surprising considering the magnitude of the first-order differences discussed above. On this scale, spreading rate seems to influence primarily the continuity in time and in space of volcanic, magmatic, and tectonic processes, and the amplitude of faulting, but the fundamental processes and structural evolution change little.

The axis of spreading is characterized by a narrow zone of recent volcanism, which is flanked by zones of crustal fissuring (Figure 2). Away from the axis, a zone of active normal faulting is characterized by significant vertical disruption of the crust (Needham & Francheteau 1974, Macdonald et al 1975, CYAMEX 1981). At some distance off-axis the crust becomes essentially stable and rigid, as assumed by plate tectonics. The zone in which 95% of crust created by volcanic and plutonic activity is of Holocene age is termed the "crustal accretion zone" (Luyendyk & Macdonald 1976), and the surface volcanic component of this zone is called here the "neovolcanic zone." The region in which active faulting and deformation occurs is the "plate boundary zone," and is subdivided into zones of fissuring and faulting. Let us consider each of these volcano-tectonic zones in greater deail.

The Neovolcanic Zone

The zone of recent and ongoing volcanism at the spreading center is usually remarkably narrow, on the order of 1–2 km wide (Figure 2). The narrowness of the central zone of volcanism has been verified by submersible observations in regions where total opening rates are 2–7 cm/yr (Ballard & van Andel 1977, Bryan & Moore 1977, Corliss et al 1979, Spiess et al 1980, CYAMEX 1981) and by deeply towed camera vehicles at rates up to 18 cm/yr (Lonsdale 1977, Ballard & Francheteau personal communication). The neovolcanic zone is characterized by fresh, glassy lava flows and an almost complete lack of sediment cover. Volcanic processes account for essentially all of the topography in this zone.

At slow spreading-rates the axial zone is marked by a highly discontinuous chain of central volcanoes (Figure 3; Needham & Francheteau 1974, Macdonald et al 1975). Central volcanoes are elongate parallel to the spreading axis and appear to be accumulations of fresh, sediment-free pillow basalts (Ballard & van Andel 1977, Luyendyk & Macdonald 1977). Mt Venus in the Famous area is a typical example with dimensions of 1 by 4 km and a height of 250 m (ARCYANA 1975). At intermediate spreading-rates, the volcanoes are more continuous along strike, except where interrupted by small (< 1 km) *en echelon* offsets. Sheet flow basalts, similar in morphology to pahoehoe, are more frequently observed (Ballard et al 1979, Normark 1976). The central volcanoes seem to reach a maximum height of only 50 m on axis (Klitgord & Mudie 1974, Crane 1979). At fast spreading rates the nature of the central volcano changes considerably. It resembles a very elongate Hawaiian-type shield volcano with gently sloping sides and a summit rift zone caused by keystone collapse (Lonsdale 1977). The central volcano is 1–2 km wide, as at slower spreading rates, but is remarkably continuous (up to 100 km along strike) and is interrupted only by transform faults (Searle et al 1981). Both pillow lavas and sheet flows are observed.

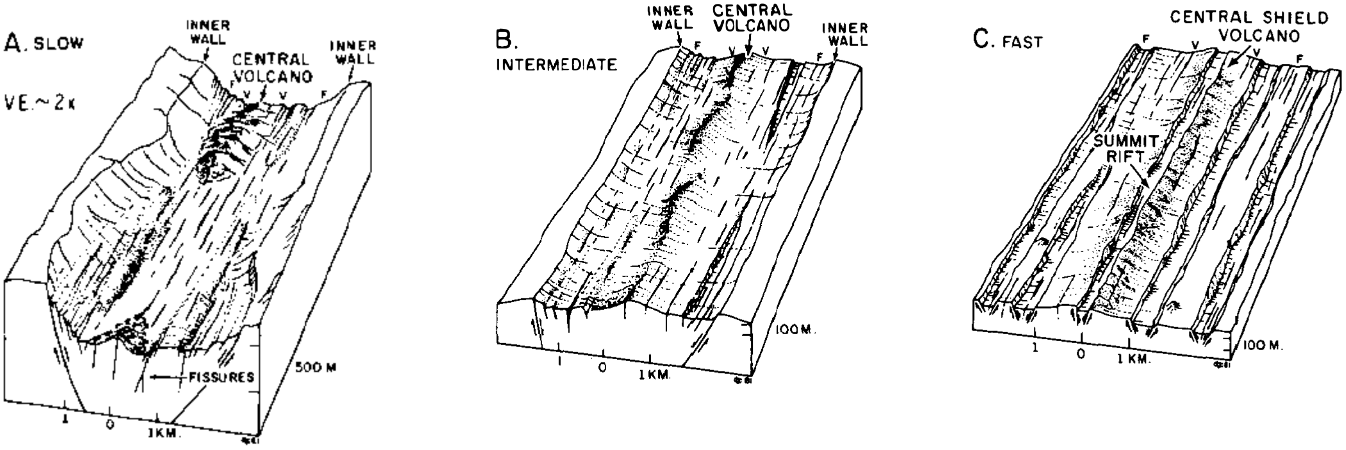


Figure 3 Schematic illustrations of the neovolcanic zone at different spreading rates. The central volcano is highly discontinuous at slow rates (*A*), moderately continuous with *en echelon* offsets at intermediate rates (*B*), and often almost perfectly continuous at fast rates (*C*). At fast rates the volcano resembles a Hawaiian shield volcano with a summit rift. At slow to intermediate rates it is a volcanic construction of pillow lavas. Fissuring of the crust appears to be greatest adjacent to the neovolcanic zone but may occur within it as well. Labels V and F as in Figure 2. (Sketch *A* modified after Moore et al 1974.)

PILLOWS VS. SHEET FLOWS: VOLCANIC CYCLES Careful analysis of the occurrence of sheet flows and pillow lavas provides important information about possible volcanic cycles along the spreading axis. It has been suggested that pillow lavas and sheet flows are the submarine equivalents of tube-fed and surface-fed pahoehoe, respectively (Ballard et al 1979). If so, then sheet flows erupt from a new volcanic vent at very high effusion rates during the initial stage of the volcanic cycle. As the volcanic edifice builds, the lava starts to flow through a volcanic catacomb of tunnels and tubes rather than erupting directly from fissures. Channeling of lava through the volcanic plumbing system and diminishing effusion rates produce pillow lavas rather than sheet flows (Figure 4). Pillow lavas and sheet flows should occur on ridge crests regardless of the spreading rate. However, at a slow spreading rate it is more likely that the late-stage pillow lavas will mask the sheet flows, while at fast spreading rates, there is a greater chance that sheet flows will remain exposed

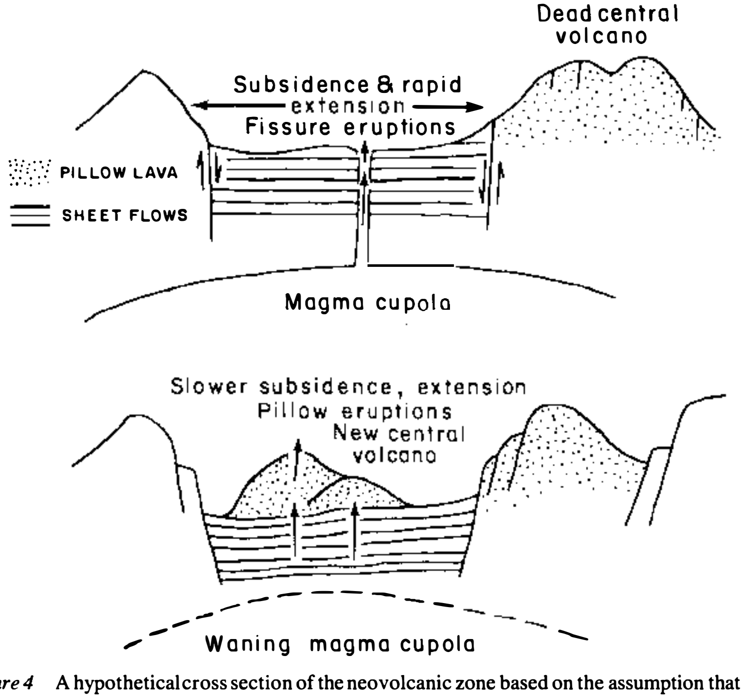


Figure 4 A hypothetical cross section of the neovolcanic zone based on the assumption that sheet flows erupt first during early stage fissure eruptions, followed by pillow lavas, which form steep volcanic constructions (modified from van Andel & Ballard 1979). Several eruptive cycles will create an approximately 1 km thick lava section that is a complex melange of pillow lavas and sheet flows (not shown). Here (*top*) the new eruptive phase is shown occurring in crust adjacent to the most recent central volcano. It is also possible for the new eruption to occur along the axis as the previous central volcano is split in two.

in many places. So far, field observations from ALVIN of pillow/sheet flow spatial relationships support this concept of volcanism (Atwater 1979, van Andel & Ballard 1979, Ballard et al 1981). It should be noted that field mapping in ophiolites indicates episodes of volcanism, but not necessarily the orderly sequence of sheet flows to pillow lavas summarized here (Hopson personal communication). A chaotic mix of sheet flows and pillow lavas is observed instead. Episodes of tectonism and hydrothermal activity may be linked to these volcanic episodes. This is discussed in later sections.

FREQUENCY OF VOLCANIC ERUPTIONS If volcanic cycles occur at spreading centers, what is their periodicity? Consider first the slowly spreading MAR. Since the approximate thickness of the volcanic section (1.0 km) and the dimensions and spacing of volcanoes are known, the spatial density of volcanoes in the crustal section can be estimated. When combined with the spreading rate, the frequency of major eruptive cycles can be estimated to be once every 5,000–10,000 years (Bryan & Moore 1977, Atwater 1979). This result is supported by deep-sea drilling results from the Atlantic (Hall 1976). In a similar exercise I arrive at an eruption frequency of once every 300–600 years for the intermediate spreading rate RISE area. For the fast spreading EPR, a slightly different analysis yields an eruption interval of approximately 50 years (Lonsdale 1977). While these calculations are all quite crude, it appears that the frequency of major volcanic eruptions increases as approximately the square of the spreading rate. We see later that fine scale studies of magnetic anomalies support this relationship.

Do axial eruptions occur rapidly with long periods of intervening quiescence, or is activity fairly continuous at a slow rate? Analogy with terrestrial eruptions suggests the former. Perhaps the best evidence is from Deep-Sea Drilling Project (DSDP) holes. For deep holes (> 500 m), thick crustal units occur in which magnetic, petrologic, and geochemical properties are nearly uniform and are significantly different from crustal sections above or below. In the presence of secular variation of the magnetic field, this suggests eruption episodes of short (1–100 yr) duration (Hall 1976) separated by long periods of quiescence.

STABILITY OF THE NEOVOLCANIC ZONE The neovolcanic zone is usually restricted to a zone only 1–2 km wide and appears to remain so for long periods. Submersible and photographic studies have delineated the narrowness of this zone in the Famous, AMAR, RISE and Galapagos study areas (Bryan & Moore 1977, Luyendyk & Macdonald 1977, Normark 1976, Spiess et al 1980, van Andel & Ballard 1979). Observations of fresh lavas and sparse sediment cover, however, give only an instantaneous view of the volcanic zone.

One way to observe the width and stability in time of the neovolcanic zone is through high resolution deep-tow studies of magnetic anomaly transitions.

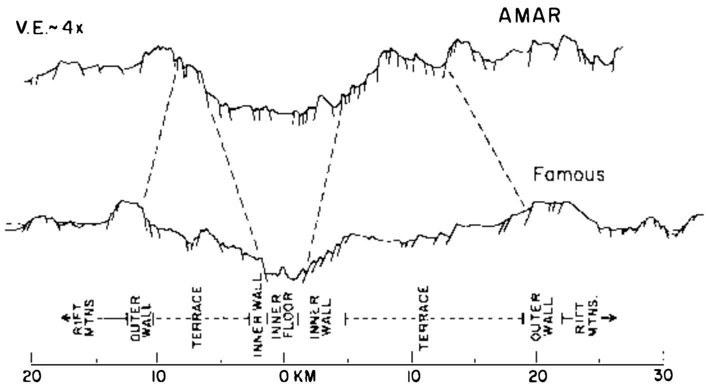


Figure 5 Deep-tow profiles of the Famous and AMAR rift valleys. Dashed lines connect the corresponding inner and outer walls (terminology at bottom after Macdonald et al 1975). These are typical end-members in the evolution of a rift valley which may have a narrow inner floor and wide terraces (Famous) or vice versa (AMAR). Magnetic anomalies may be clearly recorded when the inner floor is narrow, and "smeared out" with wide transitions when the inner floor is wide.

The magnetic anomaly polarity transition width is a measure of the total width of the crustal accretion zone, including the neovolcanic zone and magnetized plutonics (discussed in detail later) (Harrison 1968, Macdonald 1977). Transition widths at intermediate to fast spreading rates suggest that the crustal accretion zone has remained narrow (1–3 km) and localized for millions of years (Klitgord et al 1975, Macdonald et al 1980a). At slow spreading rates, the zone appears to vary between 1 and at least 8 km in width (Macdonald 1977). Thus for all spreading rates the crustal accretion zone is remarkably narrow compared to the lateral dimensions of the plates, but at slow spreading rates there may be a greater tendency for the crustal accretion zone to wander laterally, or to periodically widen to 5–10 km within the rift valley.[1]

The variation in transition width for slow spreading ridges, and thus the width of the zone of crustal accretion, may reflect a time-varying median valley structure (Macdonald 1977). The valley has either a wide inner floor and narrow terraces, in which case the neovolcanic zone is wide and magnetic anomalies are poorly recorded (wide transition zones), or it has a narrow inner floor and well-developed terraces (Figure 5). The neovolcanic zone is then narrow and anomalies are clearly recorded (narrow transition widths). The median valley of any slow spreading ridge may vary between these two extreme structures with time as well as along strike. On the MAR, the Famous (at 36°45′N) and AMAR (36°25′N) rifts are two end-member examples (Macdonald & Luyendyk 1977).

[1]There is some controversy here in that surface tow magnetic studies show a different variation in transition widths than the deep-tow results reviewed here (Blakely & Lynn 1977). The discrepancy may be due to resolution limits of surface tow magnetic studies (Miller 1977).

Another approach to the stability of the neovolcanic zone is to consider the mechanism by which volcanoes are transported out of the neovolcanic zone; do they move away as intact units or are they split along the axis? A model has been proposed for the Famous area in which volcanoes are transported out of the rift valley as whole units. The plate boundary shifts laterally up to 1 km, or 1 volcano width, and a new central volcano is born along the shifted axis (Ballard & van Andel 1977). However, at intermediate to fast spreading rates, there is growing evidence that central volcanoes may occasionally split along the axis. A portion of the volcano moves to each side and is rafted away on the diverging plates. In the RISE area approximately one seventh of the volcanoes appear to be split and in each case the rifted face is toward the spreading axis (Macdonald et al 1980a). A statistical model suggests that the zone of splitting must be less than 1.6 km wide on the average for this to occur (Macdonald 1982). This model agrees well with the narrowness and stability of the intrusion zone implied by the single-sided chilling of feeder dikes in ophiolites (Kidd 1977) and with the magnetic and observational data for the width of the neovolcanic zone.

Sparse observations suggest that the tendency for axial volcanoes to split is greater at intermediate to fast spreading rates than at slow. If so, this agrees well with the magnetic anomaly transition studies that suggest a wider or less stable neovolcanic zone at slow spreading rates. Consider the following model. At slow spreading centers, major volcanic eruptions occur only every 5,000–10,000 years. The crust cools during intervening quiescent periods, thickening and gaining brittle strength along the axial zone. This is enhanced by rapid deep cooling due to hydrothermal circulation. With the preexisting zone of weakness largely erased, the next episode of rifting and volcanism 10^4 years later may occur anywhere within the inner floor of the rift valley. (There may be a preference for rifting along the edge of the last volcano, where the crust may be thinnest.) Thermal models also suggest a thicker and stronger lid over the magma chamber at slow spreading rates (Sleep 1975). At fast spreading rates, the periods of quiescence between eruptions are shorter, only 50–600 years. Even in the presence of hydrothermal cooling, the system is likely to have a thermal memory locating the preexisting zone of weakness along the axial feeding system of the last volcano erupted (Macdonald 1982). As a result, the tendency for splitting of the central volcano will be greater and the neovolcanic zone will be narrower and more stable.

Needless to say, off-axis volcanism occurs, especially at slow spreading rates (Heirtzler & Ballard 1977, Luyendyk & Macdonald 1977). In the Famous area, deep-tow magnetic data indicate that up to 10% of the volcanism occurs outside the main neovolcanic zone in crust 0.5–2.0 m.y. old, i.e. 5–20 km off-axis (Macdonald 1977, Atwater 1979). At faster spreading rates sheet lavas may flow up to 4 km off-axis (Spiess et al 1980, Ballard et al 1979).

However, the volume of lava is usually small, and the basic model of a stable and centered neovolcanic zone seems to be valid.

Tectonic Zones

Intense crustal fissuring becomes apparent at the edges of the neovolcanic zone (Figures 2 and 3). Within 2–3 km of the axis some of these fissures develop significant vertical offset by normal faulting. In older crust, faulting diminishes significantly. We now discuss these three tectonic zones in some detail and consider the problems of steady-state maintenance of mid-ocean ridge structure.

ZONE OF CRUSTAL FISSURING Before the seafloor is ~100,000 years old, the crust becomes intensely fissured. At slow spreading rates, where the neovolcanic zone is discontinuous, fissures are also observed along the spreading axis. At intermediate to fast spreading rates, fissuring of the crust probably occurs along the axis but is obscured by continuous central volcanoes. At all spreading rates, the most intense observable fissuring occurs in bands 1–2 km wide flanking the central volcanoes. These fissures closely resemble the *gjar* in Iceland and are typically 1–3 m wide, extending 10 m to 2 km along strike (Luyendyk & Macdonald 1977, Ballard & van Andel 1977). Their azimuth closely parallels the strike of the ridge, which suggests that they are caused by tensional failure of the crust during spreading rather than cracking due to thermal contraction (Luyendyk & Macdonald 1977).

The tensional stresses producing this pervasive fissuring are caused by the horizontal acceleration of crust from zero at the idealized center of the neovolcanic zone to the full spreading rate at the edge of the plate boundary zone. With an overburden pressure of nearly 300 bars from the water layer, one might expect failure under shear rather than simple tension. However, the effective pressure is close to zero at the seabed because of high crustal permeability, and failure occurs under simple tension for several tens of meters into the young crust (Macdonald et al 1982a).

It is likely that these fissure fields provide access for cold seawater to penetrate the young, hot oceanic crust, creating the recharge system for hydrothermal convection (Lister 1972, Lowell 1975). The intensity of fissuring may control the vigor and exit temperature of hydrothermal convection. For example, the 350° hydrothermal vents ("black smokers") in the RISE area occur in a relatively unfissured portion of the spreading center (Macdonald et al 1980b, Ballard et al 1981). Areas that are intensely fissured are more efficiently cooled and have either ceased vigorous hydrothermal activity or have high temperature activity at depth with considerable subsurface mixing. Examples of the latter are the Famous and AMAR areas of the MAR (Fehn et al 1977), the TAG geothermal area (MAR 26°N, Scott et al 1974), the RISE area only a few km north of the black smokers (Crane & Normark 1977), and the Galapagos spreading center (Green et al 1981, Edmond et al 1979).

Crustal fissuring is the likely cause for the seismically defined "layer 2A" in the ocean basin. This layer is on the order of 500 m thick and has a bulk velocity of only 2.5–3.8 km/sec (e.g. Houtz & Ewing 1976), considerably lower than the velocities of 5.0–6.0 km/sec found in hand specimens. Cracking and other forms of porosity cause the very low velocities. Eventually the fissures fill with sediment and are sealed by low temperature diagenetic cementation. In addition, exiting of high temperature metalliferous solutions tends to fill cracks with hydrothermal minerals. As these processes continue, the seismic velocity of layer 2A will increase to that of layer 2 (approximately 5.5 km/sec) and the cracked layer will "disappear" as a seismically detectable entity (Christiansen & Salisbury 1975).

ACTIVE FAULT ZONE At a distance of 1–4 km from the spreading axis, some fissures develop large vertical offsets by normal faulting, with most of the faults dipping toward the spreading axis (Figures 2 and 3). At slow spreading centers, the individual faults have vertical throws of 200 m or greater. A series of these fault slivers creates scarps 600 m or greater in height (Macdonald & Luyendyk 1977), resulting in the ubiquitous deep rift valley characteristic of slow spreading ridges. The rift valley varies in depth from 1.2–3.0 km. Its shape may resemble a "U" with a single set of rift valley walls, or a nested "V" having an inner and outer set of walls with intervening horizontal areas, or "terraces" (Macdonald et al 1975) (Figure 5). These shapes represent end-members in the evolutionary stages of the rift valley in which the terraces are non-steady state. This variation in structure may control the width of the neovolcanic zone and the recording of magnetic anomalies as discussed earlier.

At intermediate spreading rates, the faults dipping toward the axis develop throws of only 50 m or less and the rift valley is shallower (Figures 2 and 3). The upfaulted relief is essentially cancelled by a gentle ($\sim 5°$) back-tilting of the fault blocks (Klitgord & Mudie 1974).

At fast spreading rates, no rift valley is evident (Figure 2). Normal faulting creates axially dipping fault scarps with throws of 50 m or less. Here however, this relief may be cancelled by both back-tilting and outward-dipping normal faults (Lonsdale 1977). The result is an undulating horst and graben terrain of typical Pacific abyssal hills. An alternative hypothesis is that back-tilting of fault blocks is dominant at fast rates as well (Rea 1975). At present, the high resolution data base for fast spreading centers is too sparse to conclude which model is generally correct.

Linearity On the fast-spreading East Pacific Rise, the major fault scarps continue uninterrupted for tens of kilometers from transform fault to transform fault (Searle et al 1981). On the slow spreading ridges, the gross inner and outer walls continue from one transform fault to the next, but individual faults as large as 200 m in throw may disappear in only 1–2 km, or merge with other faults along strike (Macdonald et al 1975, Macdonald & Luyendyk 1977). The spreading rate dependence of faulting continuity is probably a result of two factors: tectonics at slower spreading rates may be more episodic, and greater crustal thicknesses may result in less uniform stress fields and resulting strain.

Fault dips and crustal tilt Many of the faults are nearly vertical at the top indicating failure under tension. Larger throw scarps have dips of approximately 50°–60°, indicating a transition to failure under shear at approximately 20–100 m depth in the crust (Macdonald & Luyendyk 1977). First motion solutions for earthquakes on spreading centers have tension axes perpendicular to the spreading axis, appropriate for normal faulting, and also yield dips of 50° to 60° (Sykes 1967).

Fault blocks on slow-spreading centers are typically tilted 5°–15° away from the spreading axis (Atwater & Mudie 1973, Macdonald & Luyendyk 1977). Observations from ALVIN indicate that some major blocks such as those bounding the outer walls of the rift valley may be back-tilted in excess of 30° (Macdonald & Atwater 1982). Shallow magnetic inclinations measured in DSDP holes have led Verosub & Moores (1981) to hypothesize tilts in excess of 60° by movement along listric normal faults whose dips shallow with depth. Tilts of 3°–5° and occasionally up to 10° appear to be typical of intermediate-to-fast-spreading crust (Klitgord & Mudie 1974). While the data set from faster spreading centers is too sparse for a reliable comparison, it appears that crust at slow-spreading centers suffers greater tectonic disruption through tilting than crust at faster spreading centers. Radical disruption of slow-spreading crust may be caused by creation of the 1.5–3.0 km deep rift valley and its subsequent transformation into the relatively horizontal, undulating rift mountains. A spreading rate dependence in tectonic tilt may be one of several reasons why magnetic anomalies are clearer in the Pacific than in the Atlantic (Macdonald et al 1982b).

Antithetic faults Small throw faults that dip away from the axis frequently occur at the base of large axially dipping faults, especially at slow spreading centers. These small antithetic faults create an apparent "reverse drag" on the parent fault. They are likely to be caused by a shallowing of fault dip with depth on the listric parent fault and a filling in of the resulting gap with antithetically down-dropped fault blocks (see Figure 8, Macdonald & Luyendyk 1977). These antithetic faults may play a major role in exposing deeper crustal units by repeated chopping and uplifting of the crust. This mechanism may be particularly important at spreading center/transform fault intersections where crust may be anomalously thin (Stroup & Fox 1981).

Horizontal extension The faults and fissures result in a significant horizontal extension of the crust. In the Famous area the extension is 11% to the west and 18% to the east, which is in the same sense as the spreading rate asymmetry of 0.7 cm/yr to the west and 1.3 cm/yr to the east (Macdonald & Luyendyk 1977). This suggests that asymmetric spreading is accomplished by asymmetric crustal extension as well as by asymmetric crustal accretion. Crustal extension also implies crustal thinning to a degree that could be a significant fraction of the thickness of the crust. Extension of the crust due to faulting is less at faster spreading rates and is on the order of 5% (Shih 1980).

Width of the active fault zone Unfortunately earthquake locations are neither numerous enough nor accurate enough to determine the width of the plate boundary zone. With precise bathymetric data it is possible to evaluate the cumulative fault displacement as a function of distance. Where the slope becomes constant, active faulting has presumably diminished. In such a way, Macdonald & Atwater (1978a) find that most of the active faults on the MAR occur within ±5 km of the rift axis and dip toward the axis. The faulting is characterized by significant microearthquake activity (Reid & Macdonald 1973, Spindel et al 1974). Faulting on planes dipping away from the axis and tilting of faulted blocks continues in crust up to ±30 km off-axis in the rift mountains (Figures 5 and 6). Using a similar analysis for intermediate to fast spreading centers, Shih (1980) finds that active faulting occurs within ±4 to ±10 km of the spreading axis, apparently narrower than for slowly spreading ridges. A fault zone width of ±10 km was verified during submersible investigations in the RISE area (CYAMEX 1981). Beyond the active fault zone, the oceanic lithosphere is relatively stable and rigid as assumed in plate tectonics. This marks the edge of the plate boundary zone and the beginning of the rigid plate, although mid-plate earthquakes attest to some continued activity in this stable region.

EVOLUTION OF THE RIFT VALLEY The ubiquity of a deeply rifted valley at slow spreading centers has lead to the hypothesis that the rift valley is a steady-state feature (Deffeyes 1970). If this is true, a problem arises in explaining the "disappearance" of the axially dipping regional slope of the rift valley as it is transformed into the horizontal, undulating relief of the rift mountains (Figure 5). One model is that the rift valley walls are tilted back to approximately horizontal in the rift mountains (Figure 6*A*). The rift valley/rift mountain transition may be accomplished by a modest rotation (5–9°) of the entire rift valley half-section as it passes into the rift mountains. A second model is that the rift valley relief is undone by "unfaulting" along the preexisting inward-facing faults (Figure 6*B*). Thus, as new normal faults are created near the center of the valley, the relict normal faults are collapsed by reverse faulting at the valley edges. A third hypothesis is that the rift valley staircase is

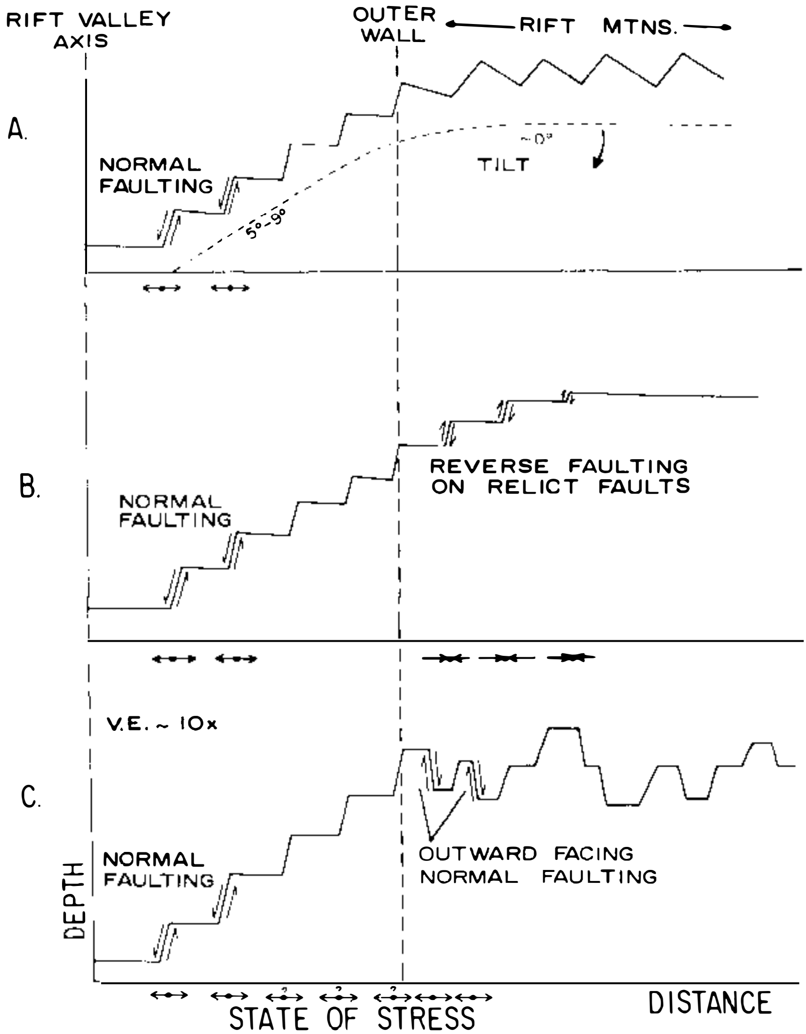


Figure 6 Three idealized models for transformation of the rift valley into the rift mountains (after Macdonald & Atwater 1978a). *A*. Regional tilting of the fossil rift valley walls in the rift mountains. *B* Reverse faulting at the outer walls and in the rift mountains. *C* Outward-facing normal faulting in the rift mountains just beyond the outer walls. Relative motion on presumed active faults shown by arrows. State of stress indicated; diverging arrows—tension, converging arrows—compression.

effectively overprinted by normal faults, which dip away from the valley axis (Figure 6*C*). These may be new fault planes or reactivated faults that previously had small offsets. While at first I favored the outward-faulting model, it now seems that all three processes must be acting in the rift mountains to maintain a steady-state rift valley (Macdonald & Atwater 1978a, b, Harrison

& Stieltjes 1977). The dominant mechanisms appear to be related to a significant tilting of fault blocks away from the rift axis by as much as 5°–50°. This tilting is often manifested by flexure and resulting failure of the crust by outward-dipping normal faulting, as well as by simple tilting (Macdonald & Atwater 1982). These processes occur as far as 30 km off-axis, which accounts for the greater width of the plate boundary zone on slow spreading ridges (~60 km) compared with fast spreading centers (~8–20 km). The relief that results from these processes is the ubiquitous abyssal hills of the Atlantic Ocean basin (Heezen et al 1959).

AXIAL MAGMA CHAMBER AND HYDROTHERMAL ACTIVITY

It has been proposed that crustal volcanic and plutonic rocks are formed by differentiation of mantle-derived parent magmas in a shallow crustal magma chamber, rather than by injection and eruption directly from the mantle (Figure 7, *top*). The magma chamber model has been strongly supported by analysis of ophiolites, petrologic studies (e.g. Cann 1974, Rhodes & Dungan 1979), thermal models (Sleep 1975), and seismic experiments (Orcutt et al 1975). Let us consider the geophysical evidence. The first strong seismic indication of a magma chamber came from ocean bottom seismometer refraction studies on the axis of the East Pacific Rise near 9°N. The primary evidence here is the presence of a substantial shadow zone beginning at ranges of 15 km,

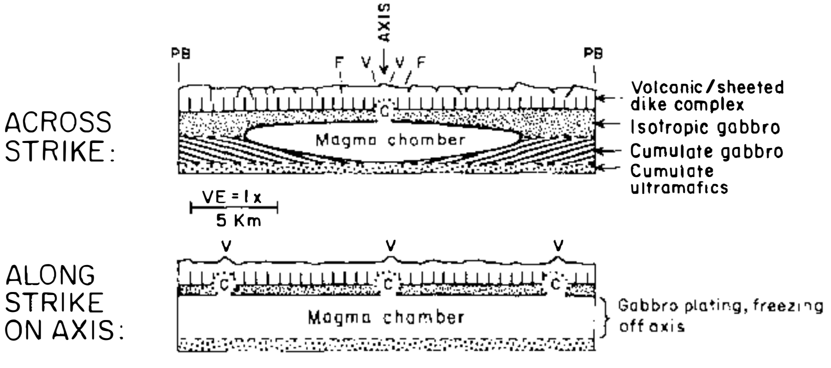


Figure 7 Hypothetical cross section of the axial magma chamber at an intermediate to fast spreading rate ridge. Topography from EPR 21°N (RISE), no vertical exaggeration. Labels F, V, PB as in Figure 2. Depth and width of chamber loosely constrained by seismic data (see text), precise shape unconstrained, C represents a possible non-steady-state magma cupola. Notice stratigraphic relation of crustal rock types to magma chamber and dependence of stratigraphy on chamber shape, width, and depth. (*Bottom*) Hypothetical section parallel to strike along the axis. Vs now correspond to active or very recent central volcanoes. Cupola spacing totally unconstrained. Along strike variation in magma chamber configuration would be difficult to detect.

in which wave amplitudes are significantly attenuated and travel times are delayed (Orcutt et al 1976, Rosendahl et al 1976). Such data require a low velocity zone, which is interpreted as a shallow magma chamber beneath the spreading axis. Subsequent refraction studies have indicated axial magma chambers beneath the EPR at 22°N (McClain & Lewis 1980), 21°N (Reid et al 1977), 12–13°N (ROSE in preparation), 10°S (Bibee et al 1981), and the GSC at 86°W (Bibee et al 1981). Additional evidence for a shallow magma chamber beneath the EPR has come from attenuation of shear waves from microearthquakes (Reid et al 1977) and from multichannel reflection work (Herron et al 1980).

Numerous seismic refraction studies over the slowly spreading MAR have not provided clear evidence for an axial magma chamber, with the continued exception of the Reykjanes Ridge (Keen & Tramontini 1970, Poehls 1974, Whitmarsh 1975, Fowler 1976). In addition, measurements of microearthquakes near the axis do not indicate attenuation of shear waves as would be expected if a magma chamber were present (Francis & Porter 1973, Reid & Macdonald 1973, Spindel et al 1974, Francis et al 1977). While it is not possible to rule out magma pockets less than a seismic wavelength in dimension (~1 km), magma chambers of the size proposed by Bryan & Moore (1977) and Hekinian et al (1976) conflict with the seismic evidence. Microearthquakes on the axis of the MAR at 45°N are up to 10 km deep (Lilwall et al 1977) and preclude a shallow steady-state magma chamber in this area.

Is the Magma Chamber Steady State?

We have seen that seismic evidence demonstrates the presence of a shallow axial magma chamber at the six carefully studied fast- and intermediate-rate spreading centers. Conclusive evidence for a magma chamber beneath the slow spreading MAR is still lacking, although closely spaced fracture zones and rough topography make seismic experiments difficult. Petrologically, the spectrum of basaltic glass compositions show different degrees of fractionation. This seems to require at least a temporary magma chamber at both fast and slow spreading rates (e.g. Juteau et al 1980, Spiess et al 1980, Bryan & Moore 1977). Furthermore, petrologic evidence suggests that the magma chambers fractionate as open systems with repeated replenishment and magma mixing, even at slow spreading rates. Thus, the seismic and petrologic data at slow spreading centers appears to be in conflict.

A reasonable hypothesis is that the magma chamber beneath intermediate to fast spreading centers is steady state while beneath slow spreading centers it is transitory. Two additional lines of reasoning support this hypothesis.

Thermal models suggest that a steady-state magma chamber cannot persist at opening rates less than 2.0 cm/yr (MAR rates), but should be present at intermediate and fast (EPR) rates (Sleep 1975). The case for occasional "freezing" of the magma chamber at slow spreading centers is particularly strong when the added cooling of hydrothermal circulation is considered. While apparently strong petrological arguments can be made for or against a steady-state chamber at slow spreading rates, the thermal arguments lead me to side with the "freezers." Nisbet & Fowler (1978) propose that if there is any magma chamber beneath the MAR, it is a narrow vertical slot. I suggest that the chamber may occasionally grow in size, but only as a non-steady-state feature during volcanic episodes. The cutoff spreading rate for a steady-state chamber is not well defined. It depends on the chamber shape, degree of partial melt, frequency of magma replenishment, and perhaps most importantly, on the depth and degree of hydrothermal circulation. There are likely to be cases in which a slow-spreading ridge segment may have a steady-state chamber either because hydrothermal penetration is lacking or unusual heat sources are present. In the latter case, Icelandic hot spot volcanism may support a steady-state magma chamber beneath the Reykjanes ridge as well as a non-rifted fast spreading center morphology (Sleep & Rosendahl 1979). Conversely, unusually extensive hydrothermal penetration may occasionally "freeze out" a normally steady-state magma chamber at intermediate rates, though probably not at fast rates.

The morphology of the neovolcanic zone outlined earlier supports this model (Figure 3). At fast spreading rates, the axial shield volcano is remarkably continuous, interrupted only by transform faults. Recent Gloria side-scan sonar data clearly shows 100 km segments of the axial shield volcano near EPR 3°S to be perfectly continuous complete with a narrow summit graben less than 500 m wide (R. Searle personal communication). It would be difficult to maintain such perfect two-dimensionality and uniform depth without a steady-state magma chamber and frequent eruptions. In contrast, slow spreading ridges have a highly discontinuous string of volcanoes of varying height and morphology in the neovolcanic zone. Only the gross tectonic structure of the rift valley shows transform-to-transform continuity. Here the magma chamber is likely to be transitory in time and discontinuous along strike. At intermediate rates, the chain of central volcanoes is frequently *en echelon*, but continuous enough to suggest an underlying continuity in the magma source. The case, however, is certainly not as strong as for fast spreading ridges.

It is possible that magma chambers at any spreading rate develop transitory peaks beneath the axis that periodically penetrate to the surface to feed lava flows (Figure 7). These "cupolas" to the magma chamber may only be feeder dikes, or may form significant shallow penetrations of the magma chamber into the dike complex (Macdonald & Luyendyk 1981, Sleep oral communication). On-bottom gravity measurements from ALVIN at EPR 21°N reveal a short-wavelength gravity minimum on the axis that is consistent with a shallow, narrow magma cupola at 1 km depth, approximately 1 km wide (Spiess et al 1980, Luyendyk in preparation). The narrowness of the neovolcanic zone has also been interpreted as suggesting a narrow, shallow magma reservoir only 1 km below the seafloor (Ballard et al 1981). Inspection of the dike-gabbro contact in ophiolites can apparently be used to either support (N. Sleep personal communication) or refute (C. Hopson personal communication) the possibility of non-steady-state cupolas. Further verification of shallow magma cupolas is difficult. If they are as small as the gravity data indicates, they are not resolvable from seismic refraction. In addition, it is likely that they are non-steady state, and disappear by magma drain-back and subsequent faulting. These cupolas may provide temporary zones of weakness at shallow levels, which spatially stabilize the zone of dike intrusion and the neovolcanic zone. Indeed, the single-sided chilling of dikes in ophiolites (Kidd 1977) and the occurrence of split volcanoes attest to this stability (Macdonald et al 1980a, Macdonald 1982).

Depth and Dimensions of the Magma Chamber

The width, shape,and depth of the axial magma chamber are critical parameters governing the petrology, structure, and stratigraphy of oceanic crust (Figure 7; Pallister & Hopson 1981). These parameters are also important boundary conditions for hydrothermal models. The total stratigraphic thickness of the volcanic section (sheet flows and pillow lavas) plus the sheeted dike complex is governed by the depth to the magma chamber. The extent of the plutonic gabbro section is determined by the thickness of the magma chamber. The uppermost zone of isotropic gabbro is thought to have crystallized downward from the chamber roof and is described as "plated gabbro" (Dewey & Kidd 1977). Magma chamber solidification apparently involves "plating" of gabbros down from the roof while gabbroic and ultramafic cumulates deposit progressively upward from the floor. The downward- and upward-solidifying parts of the chamber meet in a "sandwich zone." Here the fractionating magma reaches its most evolved composition, yielding quartz-bearing gabbros, diorites, and minor plagiogranite.

The shape of the magma chamber determines the relative thicknesses of the plated (isotropic) and cumulate gabbro sections, as well as the location of the highly fractionated sandwich zone (Figure 7, *top*). For example, a funnel-shaped chamber would yield a thick cumulate section, a thin plated gabbro section, and a highly fractionated zone at shallow levels at distances 5–15 km away from the spreading axis, depending on the size of the chamber. A bell-shaped model in which the chamber widens downward would yield a very different crustal stratigraphy (see Pallister & Hopson 1981, Figures 17 and 18). The width of the chamber will influence magma mixing, scale of crustal heterogeniety, off-axis volcanism, and overall structure and tectonics of the ridge axis. A qualified petrologist/geochemist could expound on many other implications; these are only meant to be examples.

Let us now consider the geophysical constraints from intermediate to fast spreading centers, taking depth to the top of the magma chamber first. Travel time inversion of first arrivals alone (such as τ-p inversion) provides bounds on velocity as a function of depth, but cannot confirm the presence of a low velocity zone (Kennett & Orcutt 1976). Synthetic seismograms, which model the travel time and amplitude behavior of the entire seismogram, can be used to resolve low velocity zones in a forward-modeling sense (Kennett 1976). Using these techniques Orcutt et al (1976) resolved the top of a magma chamber at only 2–3 km depth on the East Pacific Rise at 9°N. Very shallow depths of 2–3 km were found on the East Pacific Rise at 22°N and 21°N as well (Reid et al 1977, McClain & Lewis 1980). In addition a multichannel profile at 9°N revealed a reflector at 2 km depth tentatively identified as the top of the magma chamber (Herron et al 1980). However, an alternative interpretation of refraction experiments at the Galapagos spreading center and the East Pacific Rise at 10°S suggest a magma chamber at 6 km depth, which would place it beneath the moho (Bibee et al 1981). The difference in the models centers primarily on the interpretation of larger amplitude arrivals in the wave train between the 25 km to 40 km range. Both synthetic seismogram analyses are limited to a one-dimensional assumption for crustal structure, i.e. no lateral variations. One hopes that this is a plausible assumption for shooting parallel to the spreading axis, but it may not be. A very high density experiment with arrays and using three-dimensional ray tracing may be necessary to resolve this controversy. For example, it is possible (even likely at slow to intermediate spreading rates) that the magma chamber varies significantly in depth along strike (Figure 7, *bottom*). This cannot be resolved with existing data and certainly not with synthetic seismogram algorithms, which assume lateral homogeneity. Orcutt's interpretation of a shallow (2 km deep) crustal magma chamber is in far better agreement with ophiolite models and with thermal models, which predict a 1 km deep chamber at intermediate spreading rates (Sleep 1978). A steady state 6 km deep chamber is difficult to reconcile with ophiolite and thermal models.

The width (and, therefore, shape) of the magma chamber is also difficult to determine. Refraction across the East Pacific Rise at 22°N and along the rise at 9°N suggest a magma chamber no wider than 10 and 20 km, respectively (McClain & Lewis 1980, Rosendahl et al 1976). Attenuation of shear waves at 21°N also suggests a chamber less than 20 km wide (Reid et al 1977). Existing data therefore indicates a narrow axial magma chamber (<20 km), with a shallow roof (2–6 km) and with an unspecified shape. Clearly much remains to be learned about the stratigraphic location of the magma chamber within the oceanic crust and the resulting effects of petrology and structure (see Lewis 1978 for review).

Hydrothermal Activity in the Axial Zone

The large discrepancy between measured conductive heat flow and theoretical cooling-plate models suggests that at least 40% of the heat loss at mid-ocean ridges and 20% of the Earth's total heat loss is accomplished by hydrothermal circulation near spreading centers (Lister 1972, Williams & Von Herzen 1974, see Lister 1980 for review). This was first directly verified on the Galapagos spreading center (Weiss et al 1977). Hydrothermal fluids with temperatures of 20°C discharge from several sites along the Galapagos rift axis, creating a life-support system for an exotic benthic community (Corliss et al 1979). The vents were apparently associated with swarms (up to 80 per hour) of very shallow microearthquakes recorded in 1972 (Macdonald & Mudie 1974).

The discovery of the RISE hydrothermal field in 1979 allowed the first direct estimate to be made of axial hydrothermal heat flux (Macdonald et al 1980b). Here hydrothermal fluids flow from discrete chimneys in many cases, rather than diffusing around pillows. Waters blackened by sulfide precipitates jet out at rates of 1–5 m/sec at temperatures of up to 350°C. These are by far the hottest water temperatures and highest flow rates ever measured at a seafloor hydrothermal vent. These fluids are thought to be pristine and undiluted by subsurface mixing (Edmond 1980).

The entire RISE hydrothermal field is 6.2 km long with at least 12 separate vent clusters of chimneys and mineralized mounds. The field lies entirely within the neovolcanic zone in a band less than 500 m wide (Spiess et al 1980). The vents here are also characterized by shallow microearthquakes and possible harmonic tremors of the type associated with volcanic eruptions (Macdonald et al 1980c).

The resulting heat flux is approximately $(6 \pm 2) \times 10^7$ cal/sec for a single black smoker chimney. This is between three and six times the total theoretical heat flow along a 1 km segment out to a distance of 30 km on each side (crustal age 1 m.y.; see Figure 8). At least 12 major individual chimneys were found in a narrow linear zone less than 1 km long; obviously the total heat flux from the vents is overwhelming. The rate of heat loss is so great that it is highly unlikely that these vents are steady state. Macdonald et al (1980b) estimate that individual vents have a lifetime on the order of ten years. This is in keeping with the time estimated for the sulfide mound edifices to be built by precipitation (Haymon & Kastner 1981, Finkel et al 1980), and with the age distribution of clams near the vents (Killingley et al 1980, Turekian & Cochran 1981).

On the other hand, recent surveys in the Pacific indicate that high temperature vents are now commonly found on most intermediate to fast spreading centers (Spiess, et al 1980, Francheteau personal communication). This possibility raises serious questions regarding the size and longevity of the axial magma chamber once again. For example, given a simple cross-sectional model of a magma chamber beneath a single black smoker chimney, I calculate that in only 10^2–10^3 years the magma chamber will totally freeze. A shallow, 1 km diameter cupola would freeze in only 10^1–10^2 years.

It is unlikely that the spreading center hydrothermal system is either sufficiently two-dimensional or of sufficient duration to freeze the major chamber at intermediate to fast spreading rates. However, vigorous hydrothermal activity may periodically depress the depth of the chamber roof. If the chamber penetrates to shallow depths during eruptive cycles (e.g. creating a shallow magma cupola), then subsequent hydrothermal activity may act to drive the solidus back down. Magma withdrawal and subsequent faulting may also act to depress the solidus to the depth of the main chamber.

It is difficult to place bounds on the depth of hydrothermal circulation. Hydrothermal penetration should be slightly less deep than the depth of the magma chamber roof, where a chamber is present. In the RISE hydrothermal area shallow microearthquake focal depths (1–3 km) suggest that circulation on the axis is shallow (Macdonald et al 1980c). Off-axis, however, oxygen isotope studies in ophiolites suggest circulation to greater than 5 km depths, penetrating at least as deep as the moho locally (Gregory & Taylor 1981). Discrepancy between theoretical and measured conductive heat flow profiles suggests that hydrothermal activity may persist in crust up to 80 m.y. old

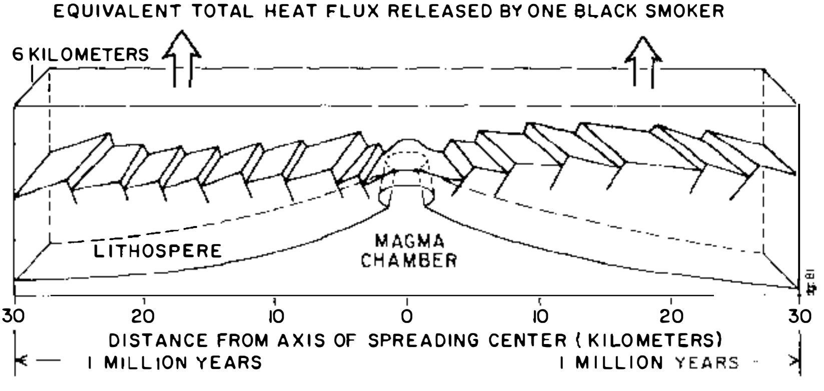


Figure 8 Heat flux from high-temperature axial hydrothermal vents make a significant contribution to the total heat budget of the Earth. The heat output of a single "black smoker" vent is equivalent to the total theoretical conductive heat flux for a spreading segment 6 km along axis out to 1 m.y. age to either side (±30 km at 6 cm/yr).

(Anderson et al 1977). However, high-temprature (~350°C) shallow-penetration (1–3 km) hydrothermal activity appears to be resticted to the neovolcanic zone and immediate surrounds.

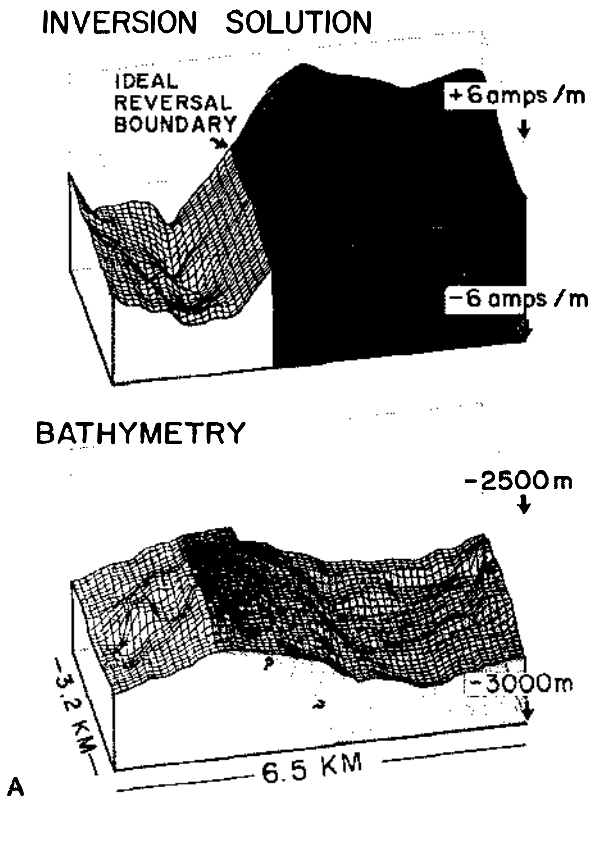
CRUSTAL MAGNETIZATION AND IMPLICATIONS

The study of marine magnetic anomalies on mid-ocean ridges has gone beyond the calculation of spreading rates to the unraveling of oceanic crustal emplacement processes, fine scale spreading behavior, and formation of the magnetic source layer. Previously, I alluded to the use of magnetic studies in constraining the width and stability of the neovolcanic zone, estimating the frequency and duration of volcanic eruptions at various spreading rates, and detecting off-axis volcanism. Here I summarize a detailed study of a typical magnetic reversal boundary, in the RISE area, (Figure 9) and consider the implications for crustal structure and evolution.

In the first stage of the study, we conducted a dense, gridded deep-tow magnetic survey of the Brunhes/Matuyama reversal transition (Macdonald et al 1980a). The data were analyzed using a Fourier inversion method including the effects of topography (Parker & Huestis 1974). The method was used in three dimensions, thus avoiding the usual assumption that the magnetized sources and bathymetry are perfectly lineated. In fact, we found that the transition is extremely linear on all scales down to hundreds of meters, and very narrow, only 1000–1400 m wide (Figure 9*A*). Encouraged and slightly surprised by the results, since the anomalies here are poorer than average for the Pacific, we extended the resolution of our study. A three component magnetometer and vertical gradiometer were mounted on ALVIN so that the magnetic polarity of individual lava flows and faulted outcrops could be determined. This gave us a spatial resolution of 0.5–1.0 m versus 200–400 m for deep-tow data (due to filtering required by inversion) and 2–4 km for surface tow data.

The 280 polarity determinations made in situ with ALVIN yielded several surprises (Figure 9*B*). Even on long traverses across both sides of the boundary, nearly every magnetic target had the correct polarity, i.e. the same polarity as that of the regional magnetic stripe defined by the deep-towed magnetometer (Macdonald et al 1982b). This observation was not too surprising for the younger side of the boundary, because newer positive-polarity crust, with the same polarity as the earth's present magnetic field, would be expected to overlie older, negatively polarized crust. What was surprising was that there were no outlying regions of new crust on the older side of the boundary. The transition in the outcropping volcanic section was sharp and linear, delineated in some cases by an actual geologic contact of opposing flow fronts, and no wider than 150 m in other crossings.

We found that the magnetic-reversal boundary surveyed with the ALVIN is displaced about 500 meters further away from the spreading axis than the boundary position calculated for the three-dimensional inversion studies. The calculated boundary marks the average position of the magnetic reversal for the entire source layer. The fact that the boundary mapped on the sea floor surface is displaced 500 meters from the calculated average position indicates a spillover of basalt away from the volcanic vents over older, negatively polarized crust. By this measure, the total width of the neovolcanic zone circa 0.7 m.y. ago was approximately 1000 m. This is in excellent agreement with submersible observations in the present neovolcanic zone, which vary from 400 m to 1200 m in width (Spiess et al 1980, CYAMEX 1981). Now consider again the 1000–1400 m width found for the magnetic transition from deep-tow. This is an indirect measure of the width of the *crustal accretion zone* (volcanics and magnetic plutonics). After the effects of crustal dilation by



faulting and field reversal time are removed, we find that the half-width of the zone of crustal accretion is only 600–1000 m. Thus the zone of crustal accretion and the neovolcanic zone both now and circa 0.7 m.y. ago have been very stable in space and generally less than 2 km and 1 km wide respectively. (Macdonald et al 1980a, 1982b).

The remarkably neat picture just derived for magnetic anomalies and polarity transitions in the Pacific is reassuring, and yet puzzling in the light of complex magnetic results from 3 DSDP holes which penetrate deeper than 500 m into Atlantic Ocean basaltic crust (332B, 395A, 418A). These DSDP results show anomalous inclination and numerous reversals in a single hole. In most cases the 500 m section represented is totally inadequate for generating the measured magnetic anomaly (Hall & Robinson 1979). This contradicts the assumption derived largely from EPR anomalies that the magnetic source lies within the ~1 km thick volcanic layer (Huestis & Parker 1977, Atwater &

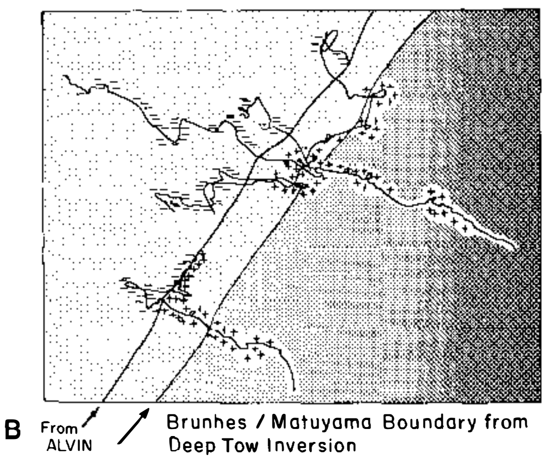


Figure 9 (*A*) *Top* Deep-tow magnetic inversion solution for crustal magnetization over a magnetic reversal boundary near the RISE area (from Macdonald et al 1982b). Notice how linear and uniformly narrow the transition boundary is. The transition width suggests that the half-width of the zone of crustal accretion was only 600–1000 m circa 0.7 m.y. ago. *Bottom* Corresponding (smoothed) bathymetry used in solution above. This is a typical abyssal hill, fault-bounded on the east side. The overlap zone of polarities marked by ? is based on ALVIN measurements shown in Figure 9*B*. (*B*) Magnetic polarity measurements made from ALVIN over the polarity transition shown in Figure 9*A*. Each symbol stands for one or more closely spaced mesurements. Underlying shading shows the inversion solution from Figure 9*A*. Note the ~500 m "spillover" of positive onto negative crust. This suggests a neovolcanic zone approximately 1000 m wide circa 0.7 m.y. ago, in excellent agreement with the width observed today from submersibles. Notice that on either side of this spillover zone, surface polarity is remarkably uniform.

Mudie 1973), and with the sharpness and clarity of the EPR magnetic stripe studied from ALVIN.

To reconcile the EPR and Atlantic results, I suggest that the crustal generating processes and resulting magnetic structure vary significantly with spreading rate (Macdonald et al 1982b). On the slow-spreading MAR, the major episodes of volcanism are likely to be infrequent (10^4 yr), the magma chamber is non-steady state, and the neovolcanic zone shifts or varies in width considerably. This sporadic, start-and-stop spreading process will contribute to a highly heterogeneous and complex crustal and magnetic structure. In addition, significant tilting of the crust may disrupt MAR crust (Hall & Robinson 1979). Models for both extreme tilting toward the spreading axis (Hall & Robinson 1979) and away from the axis (Verosub & Morres 1981) have been proposed. In some cases, extreme tilting of the crustal blocks may be sufficient to account for the non-dipole inclinations measured in Atlantic DSDP holes. For the fast-spreading EPR, more frequent volcanism (50–600 yr), a steady-state magma chamber, and a sharp, stable neovolcanic zone will create a less complex magnetic and crustal structure. This model is strongly supported by statistical studies of magnetic anomalies, which show that several reversals may be encountered in a single DSDP hole if the crustal accretion zone is 5–10 km wide (see Figure 5, Schouten & Denham 1979). Unfortunately most DSDP data is from the Atlantic, and most ALVIN magnetic data is from the Pacific. However, deep-tow and surface tow studies in both oceans verify that magnetic anomalies are far clearer and more lineated in the Pacific than in the Atlantic.

In addition, there is evidence that reversely magnetized rocks exist within the inner floor of the MAR (Ade-Hall et al 1973, Macdonald 1977, Johnson & Atwater 1977). This is quite startling, because the crust here should all be of positive polarity. Of the various possibilities, the most likely explanation for this is that blocks of older (>0.7 m.y.) crust have been left behind (Macdonald 1977, Atwater & Macdonald in preparation). Indeed, the sporadic tectonism within the MAR plate boundary zone renders approximately a 14% chance for old negative crust to become temporarily stranded near the axis (Macdonald 1977). The chances of this occurring drop off rapidly with spreading rate.

Another problem which arises is the rapid decay of the magnetic intensity of the volcanic layer away from the ridge axis. Intensities decrease to $1/e$ of their initial values in only 0.6 m.y. (Irving 1970, Macdonald 1977). This fact plus the DSDP results suggest that deep sources such as the gabbro section ("layer 3") carry a significant magnetic signal (Cande & Kent 1976, Harrison 1976, Blakely 1976). Magnetization of gabbros of 0.5–1.0 amp/m support this contention (Kent et al 1978, Day & Luyendyk 1981). On the other hand, the amplitudes of magnetic anomalies in old ocean basins still indicate a strong contribution from the volcanic layer, especially at fast spreading rates (Miller & Macdonald in preparation). It is still uncertain what percentage of the magnetic signal comes from different depths in the crust. This certainly varies with age, and may vary with spreading rate as well.

Two other magnetic curiosities from the MAR are worth mentioning. Oblique spreading patterns may be stable for most slow spreading ridges while an orthogonal pattern, as normally assumed, is preferred by intermediate to fast spreading centers (Atwater & Macdonald 1977). Asymmetric spreading with asymmetry up to a factor of 2 may persist for several m.y., but tends to reverse in sense so that over tens of m.y. the spreading is grossly symmetric (Macdonald 1977).

TOWARD A THREE-DIMENSIONAL MODEL OF SPREADING CENTERS

While it is tempting to think of a spreading center as a two-dimensional structure, there is evidence for considerable variation in the third dimension, along strike. For example, the neovolcanic zone is characterized by high-relief pillow lava constructions in some places and smooth sheet flows in others (Ballard et al 1978). Particularly on slow spreading ridges, the neovolcanic zone often has "gaps" in the sense that there is no central volcano and the crust on the axis is intensely fissured and faulted with no signs of volcanic activity in the last 50,000 years (Figure 3; e.g. AMAR rift, Ballard et al 1978). The neovolcanic zone may also be characterized by hydrothermal activity. It may be of the vigorous, high-temperature type (Spiess et al 1980), low temperature (Corliss et al 1979, Scott et al 1974), or dormant (Fehn et al 1977). Patterns of faulting also exhibit changes. For example, the MAR rift valley varies between a V-shaped valley with wide terraces and a narrow inner floor, and a U-shaped valley with narrow or no terraces and a wider inner floor (Figure 5). Magnetic anomalies also exhibit considerable variations along strike in clarity and linearity especially at slow spreading rates (compare Loncarevic & Parker 1971 with Macdonald 1977). These variations suggest episodicity in time. It is also possible that these episodes of tectonism, volcanism, and hydrothermal activity occur in some cyclic manner. This was suggested earlier for the sequencing of pillow lavas and sheet flows during volcanic eruptions (Figure 4; van Andel & Ballard 1979). Let us now consider a totally hypothetical cycle that can account for many of the observations.

Seafloor spreading results in a mass deficit along the axis which must be balanced by upwelling of material from the asthenosphere. Crustal volcanic and plutonic rocks will form by differentiation of these parent magmas in a shallow magma chamber. As seafloor spreading proceeds, a magma chamber is either created if it is non-steady state (e.g. MAR) or replenished if it is steady state (e.g. EPR).

Eventually, feeder dikes will break through to the surface and feed lava flows. During the initial stages of magma penetration to the surface, a shallow magma cupola may form. If so, it may help to stabilize the neovolcanic and dike intrusion zone (Figure 7). The first stages of volcanism will be fissure eruptions (as in Hawaii, Iceland, or Afar) in which lava erupts at a rapid rate, forming sheet flows. As the volcanic plumbing system develops, coupled with a possible decrease in lava flux, pillow lavas develop and build edifices upon the sheet flows (Figure 4). As new vents open, a complex intercalation of pillows and sheet flows develops. After 1–100 years the volcanism will wane and the last gasps of pillow lava will cap the volcanic carapace.

Hydrothermal activity may accompany volcanism and persist for some time after the volcanic episode. Continued faulting and fissuring in the off-axis tectonic zones (Figure 2) will allow cold seawater to penetrate to the base of the brittle crust to where ductile deformation dominates at approximately 600°C (Lister 1977). The mode of hydrothermal activity will depend on the degree of subsurface dilution of hydrothermal fluids with cold seawater. If the crust is intensely fissured, considerable mixing and dilution may occur at depth in the crust. High temperature activity and metallogenesis will not be evident on the seafloor, but may occur at depth within the crust. Diffuse, low temperature activity may be detectable on the seafloor. If the structure is such that little dilution occurs at depth, then pristine hydrothermal fluids may jet out of the volcanic carapace. The RISE area represents the latter stage, while the Galapagos and TAG (MAR 26°N) areas represent the former (Green et al 1981, Scott et al 1974). As cooling proceeds the axial magma cupola, and perhaps even the main chamber (at slow spreading rates), may freeze. A waning heat source combined with continued opening of the crust by faulting and fissuring (increasing subsurface mixing) will lower the temperature and vigor of hydrothermal circulation. The Galapagos spreading center may be just entering this phase since old abandoned smoker chimneys are observed a short distance off-axis, while present discharges are low temperature (22°C) (van Andel & Ballard 1979, Atwater personal communication). High-temperature (350°C) hydrothermal activity may last 10–100 years after volcanism wanes, while low-temperature (<20°C) activity may persist much longer. For example, ~20°C hydrothermal fluids are detected in the Galapagos mounds hydrothermal field in crust nearly 1 m.y. old (Williams et al 1979). At slow spreading rates, 10,000 years may pass until the next volcanic cycle, while at intermediate to fast rates, only 50–600 years of quiescence may intervene. Tectonism is likely to continue as indicated by the broad distribution of mid-ocean ridge earthquakes. Central volcanoes will either split or move as intact units away from the plate boundary. In addition to spreading, subsidence through faulting will occur so that the next volcanic episode will add to the thickness of the volcanic carapace.

The hypothetical cycle outlined above is a composite of ideas from van Andel & Ballard (1979), Macdonald et al (1980b), Crane & Ballard (1980), and N. Sleep (personal communication), as well as from hallway discussions with M. Mottl, T. Atwater, C. Hopson, and others at the Ocean Lithosphere Chapman Conference (1981). This cycle may not in fact be a clearly defined repetitive sequence, but a more random series of episodes. The crucial tests will be from analysis of field relationships seen from ALVIN, the DSDP holes, and in ophiolites.

Regardless of the existence of particular cycles, it appears that spreading rate plays an important role in the frequency of volcanic and associated hydrothermal episodes. As I pointed out earlier, the period between episodes may be roughly proportional to the square of the spreading rate. The frequency influences the manifestations of these episodes. At fast spreading rates, the increased continuity in time and the probability of a steady-state magma chamber produce a continuity along strike. The neovolcanic and tectonic zones are quite linear as are the resulting magnetic anomalies. High-temperature venting is more likely to occur. Three-dimensionality occurs primarily because of site-to-site variation in the recency of volcanism and the style of hydrothermal activity (high vs low temperature). This in turn may be affected by the presence or absence of a shallow magma cupola.

On slow spreading ridges, long period tectonic cycles may also occur. The valley may have a narrow inner floor and wide terraces (Figure 5, *bottom*) or a wide inner floor and narrow terraces (Figure 5, *top*). The time represented between the development of these two extreme structures is approximately 0.2–0.4 m.y. Clearly this is a much longer time scale than the volcanic episodes discussed earlier. Several episodes of volcanism will occur with intervening periods of down-faulting, quiescence, and subsequent burial during the next volcanic episode. When the inner floor is narrow, the neovolcanic zone may be narrow, restricted by the inner walls of the rift valley. As a result magnetic anomalies should be clearly recorded and have narrow polarity transitions. When the inner floor is wide, magnetic anomalies may be poorly recorded, as in parts of the Famous area (Macdonald 1977) and the MAR at 45°N (Loncarevic & Parker 1971). This model is supported by magnetic (Macdonald 1977), petrologic (Stakes et al in preparation), and structural data (Atwater 1979).

Another important concept in the three-dimensional behavior of spreading centers is that of propagating rifts. While this concept has been employed in several tectonic contexts (Bowin 1974, Shih & Molnar 1975), Hey has developed it into an important corollary of plate tectonics (Hey 1977, Hey et al 1980). It explains the westward propagation of the Galapagos spreading center, and quite spectacularly unravels the oblique magnetic anomaly offsets of the northeast Pacific (see Figure 2 of Hey et al 1980 for a summary of propagating rift kinematics). Oblique fracture zone patterns ("pseudofaults"), V-shaped trends in seafloor morphology, and occurrences of ferro-basalts and high amplitude magnetic anomalies appear to be related to these peripatetic rifts. This model also provides a mechanism for spreading centers to change azimuth in response to changes in stress directions. The rifts propagate at approximately 1–5 times the local spreading rate (Hey et al 1980, Crane 1979).

As well as creating and destroying plate boundaries on a large scale, rifts may also propagate between transform faults on a smaller scale. The tectonic, volcanic, and hydrothermal episodes outlined earlier may propagate back and forth along strike in the wake of a propagating crack between transform faults. This hypothesis remains to be tested.

Other more obvious perturbations to spreading centers in the third dimension are transform faults. Besides offsetting the spreading center, it appears that transform faults profoundly affect the creation of oceanic crust near ridge crest intersections. The cold, rigid boundary at the transform/ridge intersection appears to cause a decrease in the volcanic and plutonic budget (Fox et al 1980). This results in local crustal thinning and contributes to the topographic depression at ridge/transform intersections. Hydraulic head loss at the intersection may also contribute to the topographic low (Sleep & Biehler 1970). Seismic refraction results show that the resulting volcanic carapace and plutonic section is anomalously thin, approximately one half normal thicknesses (~2 km) (Fox et al 1980). As a result, faults and antithetic faults may often expose plutonic gabbros in the faulted intersection walls (Stroup & Fox 1981).

Morphologic data suggest that transform faults may influence the structure and generation of oceanic crust throughout the entire length of spreading centers that are at most 50 km long (Macdonald & Luyendyk 1977). This may be due to both thermal edge effects and increased head loss at the intersection. This would affect most of the Atlantic oceanic crust, where spreading segments average 50 km in length (Fox et al 1969). Transform spacing is wider on the EPR, so crustal thinning may not alter the structure of Pacific crust as extensively. This is yet another complexity in the third dimension that may render the slow-spreading Atlantic crust extremely heterogeneous relative to the Pacific.

Epilogue

In the process of writing this review, I have been impressed both by how much we have learned about the fine scale tectonic, volcanic, and hydrothermal processes of mid-ocean ridges, as well as by how many important controversial issues remain unresolved. For example, evidence from seismic, petrologic, and ophiolite field studies underscores the importance of an axial magma chamber, yet several key questions remain unanswered. What is the stratigraphic relationship of the magma chamber to the oceanic crust and upper mantle? What are the dimensions and shape of the chamber? At what spreading rate does the chamber cease to be steady state? Concerning axial hydrothermal activity, what is the temporal behavior of high temperature springs of the "black smoker" variety? How deep does hydrothermal activity penetrate? What are the water/rock ratios? What is the nature of hydrothermal activity and metallogenesis at slow spreading centers? Concerning magnetic anomalies, how stable (in time and in space) is the crustal generation process? How deep in the crust are magnetic sources significant? How can the magnetic heterogeneity observed in DSDP holes be reconciled with clearly recorded magnetic anomalies documented for much of the ocean floor? Concerning faulting, how deep does active faulting penetrate on mid-ocean ridges? Does the oceanic crust undergo extreme tilting in its tectonic evolution, and if so, is the degree of tilting spreading-rate dependent? How far away from the spreading center does faulting persist? What is the state of stress in the crust? Finally, do mid-ocean ridges undergo rhythmic cycles linking tectonic, volcanic, and hydrothermal processes, or is the ridge more chaotically episodic?

The level of excitement in this field is exemplified by the degree of controversy which brewed in the hallways at UCSB shortly after I dropped a first draft of this paper on the desks of the reviewers. The plausibility of cycles and of shallow magma cupolas was hotly debated. Strong evidence was presented for both ephemeral and steady-state magma chambers at slow spreading centers. Many other issues were raised. I hope that this preliminary attempt at a synthesis catalyzes debate in your hallways as well.

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