Mid-ocean ridge eruptive vent morphology and substructure: Evidence for dike widths, eruption rates, and evolution of eruptions and axial volcanic ridges

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Abstract. High-resolution side scan sonar data of Mid-Atlantic Ridge (MAR) inner valley floor axial volcanic ridges reveal details of their architectural elements. We develop quantitative models for basaltic eruptions from dikes and compare the predicted products of these eruptions with the observed morphologic features. Inhibition of gas exsolution, lack of magma disruption, and hydrothermal effects combine to decrease the rise speed of magma in submarine dikes and enhance cooling, leading to more rapid centralization of eruptions along the widest places in the dike. Dikes are predicted to initially feed eruptions from fissure vents, producing lines of hummocky ridges; centralization of activity to several adjacent vents produces chains of hummocky bulbous mounds and can enhance the effusion rate at a single vent to produce small seamounts. The widest dikes are predicted to produce smooth flows up to several kilometers in length, which should pond in adjacent lows. Edifice sizes predicted on the basis of volume fluxes, and flow lengths implied by the widths of dikes observed on the seafloor, in Iceland and in ophiolite complexes, are in quantitative agreement with the dimensions of observed features. Each ridge is made up of the products of a variety of these individual dike-emplACEMENT and extrusive events involving various dike widths, cooling times, and eruption durations. On the basis of typical MAR spreading rates, about one dike emplacement event would be expected every 40 years; since not all dikes reach the surface, the actual mean interval between eruptions will exceed this. MAR dikes should solidify between average dike emplacement events. Dike emplacement events are more frequent on the East Pacific Rise (EPR) and thus new dikes are more likely to reoccupy the sites of incompletely solidified older dikes. Differences between the morphology of the MAR axial volcanic ridge and the sheet flow dominated EPR are attributed to on average wider dikes erupting with greater frequency along the EPR.

Introduction

Exploration of the seafloor with multibeam echo sounders has provided a detailed view of the topography of slow spreading centers at relatively high spatial resolution (~150 m) and has been essential in defining their basic morphology and structure [e.g., Ballard and van Andel, 1977; Purdy et al., 1990; Grindlay et al., 1991; Sempere et al., 1993, and references therein]. Side scan sonar data obtained with deep-towed vehicles have been used to investigate the origin of this morphology and structure and have illustrated the importance of both volcanism and tectonism in shaping the Mid-Atlantic Ridge (MAR) [e.g., Macdonald and Luyendyk, 1977; Kong et al., 1988; Parson et al., 1993, and references therein]. Recently, high (a few tens of meters) horizontal resolution side scan sonar data were obtained for the complete width of the inner valley floor of the MAR median valley for major portions of several spreading segments between 25° and 29°N [Smith and Cann, 1990, 1992]. These data provided the basis for the classification and characterization of a wide range of volcanic features occurring in the inner valley floor [Smith et al., 1995]. The diversity and associations of volcanic features revealed by these data provide the opportunity to link models of the emplacement of magma in the oceanic crust to the morphological features representing the surface extrusion associated with this emplacement [Bonatti and Harrison, 1989]. Interestingly, seafloor eruption conditions bear some similarities to those typical of Venus (high ambient environmental pressures and high density of the surrounding medium). On Venus, these factors lead to inhibition of gas exsolution and explosive eruptions, and decreasing magma rise rates. These factors in turn lead to enhanced near-surface cooling of dikes, and when lava emerges to the surface, more efficient convective heat loss and more rapid initial cooling of lava flows, even in the very hot Venus environment [Head and Wilson, 1986, 1992].
Figure 1. Bathymetry illustrates the context of the axial volcanic ridge in the median valley of the Mid-Atlantic Ridge in the vicinity of 29° N. Perspective view looking north along the ridge shows the steep scarps bounding the inner valley floor and the position of the bathymetric cross sections. Bathymetric contour map of the same area of the Mid-Atlantic Ridge and four profiles across the median valley shows the location of the axial volcanic ridge (AVR), whose crest is indicated by an arrow in each profile. Width of each profile is about 9 kilometers. AVR is 100-300 m above surrounding valley floor and is often not symmetrically developed relative to position in valley or cross-sectional shape.
In this paper we start with the observation from ophiolites [e.g., Kidd, 1977; Nicholas, 1989; Baragar et al., 1987, 1989; Rochette et al., 1991; Schmincke and Bednars, 1990], Iceland [e.g., Gudmundsson, 1995a,b], and the seafloor [e.g., Azis et al., 1989; Hurd et al., 1994] that dikes of certain widths are extremely common in the upper part of the ocean crust, and we test the hypothesis that most of the observed surface morphologies could be produced by a relatively simple pattern of intrusion of dikes, penetration of some dikes to the surface, eruption on the seafloor, and subsequent evolution of the dikes and eruptions.

Observations

The topography of the slow spreading (25 mm year⁻¹, full rate) MAR between 20° and 40°N is characterized by a major rift valley and is composed of discrete spreading segments each tens of kilometers in length. The rift valley walls are defined by distinctive fault scarps, and the inner floor (Figure 1), structurally defined by the first of these major scarps, contains a variety of volcanic features commonly including a prominent axial volcanic ridge (AVR), which is the main site of lava extrusion in the axial zone [Ballard and van Andel, 1977; Ramberg and van Andel, 1977; Sempeere et al., 1993; Smith and Cannon, 1993; Smith et al., 1995]. Axial volcanic ridges are diverse in their morphologic characteristics and can vary in width and even be discontinuous along the strike of individual segments. Typical dimensions of AVRs in this study area [Smith et al., 1995] are 35-45 km long, a few kilometers wide, and about 150-250 m high, but with some ridges up to 500 m in elevation. Individual features and structures associated with AVRs and making up their architectural elements include (Figures 2 and 3) the following.

1. Hummocks and hummocky mounds are individual rounded mounds 50-500 m in diameter and less than 50 m high (Figures 3a and 3b).

2. Hummocky ridges (<50 m high, ≤500 m wide, and about 1000-3500 m in length), ranging from volcanic ridges with individual mounds of comparable size, grading into more beaded rows of small mounds, and finally grading into smoother linear ridges where individual mounds are not easily distinguished (Figures 3c and 3e).

3. Near-circular volcanic edifices or seamounts located along the strike of, and continuous with, the AVR (Figures 2, 3d and 3e). There are abundant seamounts seen across the valley floor and most of these are on the axial volcanic ridges; 83% have a bulbous or hummocky morphology and the rest have a smooth sonar backscatter texture at these scales. Seamounts range in diameter from 0.5 to 3 km and in height from 50 to 350 m, with a median diameter of ~600 m and a median height of ~63 m. Smooth-textured seamounts include flat-topped and hat-shaped structures, often with a summit crater [Smith et al., 1995]. Construction of seamount morphology by extrusion on top of preexisting ridges and features means that the total extruded volume of the edifice is somewhat less than that implied by simple height-radius relationships.

4. Smooth flows occur in topographic lows between these other features and adjacent to AVRs (Figures 3e and 3f); their sources are not as visible as they occasionally are on the East Pacific Rise (EPR) [e.g., Macdonald et al., 1989] and the Juan de Fuca Ridge [e.g., Embley and Chadwick, 1994]. In some segments they appear to dominate the valley floor and in one segment of the study area they occur to the exclusion of an AVR [Smith et al., 1995]. Although these may be similar to the sheet flows of the EPR that are observed during submersible dives to be smooth-textured at submeter scales, observations during a dive on at least one volcano (Sorecki) topped by smooth flows revealed that they are characterized by pillow texture [Bryan et al., 1994]. We refer to them here as smooth flows because of their characteristics at the side scan sonar wavelength scale, but also use sheet flows to denote their often broad extent and general similarity to flows seen on the EPR.

5. Linear fractures, scarps, and graben occur on the AVRs and are typically oriented parallel to the strike of the segment (Figures 3c and 3e). Scarps range up to several kilometers in length and face in both directions; in some cases, facing scarps appear to make up graben, which typically are up to a few hundred meters in width, ranging down to narrow fissures in which a flat floor cannot be detected at the available resolution. In some cases, flows and edifices appear to be superposed on these features; in others, the tectonic features appear to be superposed.

These typical AVR edifice characteristics and relationships (Figures 1-3) [Smith et al., 1995] provide a basis on which to test the hypothesis that they can be explained by dike emplacement mechanisms and associated eruptions.

Model of Dike Emplacement and Eruption Evolution

Recent theoretical and laboratory modeling work has suggested that there should be a strong correlation between
magma eruption rate and the morphology of consequent effusive deposits [Griffiths and Fink, 1993; Pinkerton and Wilson, 1994; Whitehead and Kelemen, 1994]. In addition, observations and monitoring in the subaerial Earth and planetary environments have provided a new paradigm for the emplacement of dikes and their relation to eruptive phenomena. Here we summarize this paradigm in four stages and apply it to the seafloor environment in order to provide a basis for the interpretation of the landforms described above.

Dike Emplacement and Eruption Initiation

At shallow lithosphere depths, the thermal state, density state, stress state, and consequent rheological properties of lithospheric magma and rocks ensure that magma motion in this region almost always takes place through elongate, elastic cracks (dikes). This is in contrast to deeper levels, where melt transfer may take place through the rise of positively buoyant, relatively equant, diapiric bodies [Marsh, 1982; Rubin, 1993a]. Wherever magma segregation takes place at shallow levels beneath the seafloor, dike formation must be the typical mechanism by which magma leaves regions of accumulation (e.g., reservoirs) and travels upward to feed intrusions or eruptions [Rubin, 1993a; Ryan, 1993], whether by direct vertical rise or by a combination of lateral and vertical motion [e.g., Nicholas, 1989; Embley and Chadwick, 1994; Ryan, 1994; Dziak et al., 1995]. Indeed, observations of faulted margins and rifted areas of the seafloor have revealed the presence of dikes [Auzende et al., 1989; Francheteau et al., 1990; Hurst et al., 1994] and seismic reflection data [e.g., Sinton and Detrick, 1992] show that shallow reservoirs exist at depths of 1 to 1.5 km beneath the axial ridge of parts of the East Pacific Rise.

At slow spreading centers, magma segregation from regions of partial melting should take place in a similar way, but it is not clear that melt accumulation rates will be high enough to allow shallow reservoirs to exist at all, let alone to be as long lived as those at faster spreading centers. Nisbet and Fowler

Figure 3. High-resolution TOBI images of individual features on MAR axial volcanic ridges between 25° and 29° N. See Figure 8 for interpreted modes of emplacement of each of these features. (a) Hummocks and hummocky mounds (north toward top right; insonification from top of image). (b) Hummocky mound (north toward top right; insonification from top of image). (c) Hummocky ridge (north toward left; insonification from top of image). (d) Small seamount or shield (north toward top right; insonification from top of image). (e) Hummocky ridge and flat-topped seamount with central depression (north toward right; insonification from top of image). (f) Smooth or sheet-flow-like area (north toward top right; insonification from top of image). Note how the higher standing area in the bottom right serves as a barrier to the flow and how small hummocky mounds on the top appear partly embayed.
Figure 3.

[1978] suggested from seismic and geochemical data that, if they exist, MAR magma reservoirs must be <2 km in diameter. However, there is no direct evidence to date of presently existing magma bodies beneath the MAR; seismic data have not revealed reflectors thought to be the tops of melt zones, as they have at the EPR, and the detailed geometry of magma movement and dike emplacement is not well constrained. Where long-lived reservoirs are present, melt is likely to be organized into elongate 10 to 15-km-long bodies extending laterally from above mantle diapirs, themselves about 15 km in length; these together are separated by intervals of about 50 km, the typical ridge segmentation length [Nicholas, 1989]. Given these dimensions, MAR axial volcanic ridges with lengths of 35-45 km should be formed by magma reaching the surface from the shallow reservoirs by a combination of both lateral and vertical dike propagation.

The ambient pressures in magma reservoirs below the ocean floor should lie in the range 80 to 130 MPa. These values consist of a minimum pressure equal to the hydrostatic load of a 1000-to 1500-m thickness of crystalline rocks and a 3 to 4 km depth of overlying ocean, combined with a likely excess pressure of order 10 MPa inherited from the consequences of volume changes on melt formation in the magma source region. Magma entering such a reservoir from the deeper mantle is likely to contain -0.65 wt % CO₂, 0.3 wt % H₂O, 0.1 wt % S, 0.03 wt % F, and 0.01 wt % Cl [Gerlach and Graeber, 1985; Pan et al., 1991]. If the magma resides in the reservoir for a long enough period (at least a few hours [Mangan et al., 1993]), CO₂ is exsolved and a new equilibrium is reached with 450 to 700 ppm remaining in the magma [Bottinga and Javoy, 1990], this range reflecting the pressure variation with depth within the reservoir. Bubbles of exsolved gas drift toward the top of the reservoir and may be lost through cracks in the overlying rocks. However, if magma leaves the reservoir to form an intrusion or feed an eruption before significant bubble loss has occurred, we calculate (treating the -0.6 wt% of CO₂ as a near-perfect gas) that the bubbles will occupy 5.9 to 6.1% of the magma volume, leading to a ~6% reduction in its bulk density. Analyses of lateral dike emplacement and magma motion at subaerial basaltic volcanoes which possess linear rift zones, such as Kilauea in Hawaii, show that stable lateral propagation for appreciable distances requires that the dike be confined to a finite range of depths by the integrated effects of the spatial variations of the magma density, country rock density, and country rock stresses. A number of processes [Rubin and
removal of melt from the reservoir reduces its internal pressure is being fed from mantle source, at a rate greater than that at eventually lateral dike growth must cease unless the reservoir pressure is greater than some value (commonly of the order of depends mainly on its initial volume [Blake, 1981]. Thus as it relaxes (elastically or inelastically) by an amount which of the growing dike [Parfitt and Head, 1993]. Progressive of the magma in the tip of the dike while the dike is still growing upward. As soon as an eruption begins, a new pressure gradient is established along the dike, buffered by the reservoir pressure at one end and the ocean floor pressure at the other. The system adjusts to the new gradient on a timescale controlled by the passage of acoustic waves through the dike magma. These waves travel at speeds of ~100 m/s when the vesicularity is ~6% [Kieffer, 1977], the value found above for the reservoir magma, so that the relaxation time is a few tens of seconds for dikes a few kilometers long. Only the magma erupted during the early part of this period will have the above enhanced (~12%) vesicularity; the remainder of the magma should all be equilibrated with the ambient ocean-floor pressure, and thus be less vesicular, as commonly appears to be the case on the basis of samples from the sea floor at these depths [Moore, 1979].

Horizontal motion of magma and lateral dike emplacement from a reservoir can only be initiated if the pressure in the melt within the reservoir exceeds the local minimum compressive stress in the rocks forming the reservoir wall. The amount of excess pressure is a function of the size of irregularities on the reservoir wall and the tensile strength and grain size of the rocks, these being the factors which control the effective fracture toughness of the wall rocks [Sartoris et al., 1990; Parfitt et al., 1993; Rubin, 1993b]. Furthermore, dike propagation will continue only so long as the reservoir pressure is greater than some value (commonly of the order of several megapascals) which is a function of the size and shape of the growing dike [Parfitt and Head, 1993]. Progressive removal of melt from the reservoir reduces its internal pressure as it relaxes (elastically or inelastically) by an amount which depends mainly on its initial volume [Blake, 1981]. Thus eventually lateral dike growth must cease unless the reservoir is being fed from mantle sources at a rate greater than that at which it is being depleted by dike growth [Parfitt and Head, 1993]. The interaction between the rates of lateral growth driven by changing reservoir pressure and vertical growth driven by increasing amounts of gas exsolution is very complex. For example, rift zone dikes which feed eruptions at Kilauea may sometimes open to the surface first at the end furthest from the reservoir, sometimes open first at the proximal end, and sometimes begin to erupt at some intermediate distance with the active fissure, then enlarging in either or both directions [e.g., Wolfe, 1988].

Initial Eruptions from Linear Fissure Vents

Once an eruption breaks out at the surface, the pressure and velocity distributions in the rising magma (and the width of the feeding dike as a function of depth) adjust quickly until the pressure in the vent reaches an equilibrium value which, for submarine eruptions, will be the ambient water pressure on the ocean floor. Subaerial eruptions can take place under conditions in which the pressure in the vent is higher than the ambient value because the fissure system does not flare outward toward the surface rapidly enough to accommodate the expansion of gas being exsolved from the magma [Kieffer, 1984; Giberti and Wilson, 1990]. However, computation of the characteristics of eruptions on the ocean floor using the numerical models of magma motion described by Wilson and Head [1981] shows that, for the estimates of the amounts of exsolved gas in erupting magmas given earlier, above-ambient pressures in the vent are extremely unlikely. We saw earlier that the bulk density of a lava erupting on the ocean floor will not change rapidly with depth below the vent given the high ambient pressure. We can therefore use standard fluid-mechanical relationships for incompressible fluids [e.g., Knudsen and Katz, 1958] to express the magma rise speed, \( u \), and the volume flux per unit length along strike of the active fissure, \( Q \), in terms of the absolute pressure gradient driving the motion through the dike, \( dP/dz \), the dike width, \( W \), and the magma viscosity, \( \eta \):

\[
\begin{align*}
u &= \frac{(W^2 dP/dz)}{(12 \eta)} \tag{1} \\
Q &= \frac{u W}{(W^3 dP/dz)} \frac{(12 \eta)}{} \tag{2}
\end{align*}
\]

We use a range of pressure gradients (\( dP/dz = 1000 \text{ to } 3000 \text{ Pa m}^{-1} \)) consistent with magma rising 1.5 km vertically while traveling 2 km horizontally to an eruption site as a result of some combination of a magma buoyancy (relative to the crustal rocks) of \( \sim 100 \text{ kg m}^{-3} \) and an excess reservoir pressure of 10 MPa. A range of dike widths was used (\( W = 0.2 \text{ to } 3 \text{ m} \)) comparable to those commonly found in ophiolite complexes [Kidd, 1977] reports an average width of 1.5 m for the Troodos ophiolite, an average for Betts Cove, Newfoundland, of 0.6 m) [Basaltic Volcanism Study Project, 1981; Nicholas, 1989; Rochette et al., 1991; Baragar et al., 1987, 1989], Iceland (average thickness in Pleistocene dike swarms is less than 2 m [Gudmundsson, 1995a]) and the seafloor [Francheteau et al., 1990; Hurst et al., 1994]. A magma viscosity of \( \eta = 300 \text{ Pa s} \) was used, somewhat greater than common basaltic magma viscosities below the vent to allow for enhanced cooling in the seafloor environment (unpublished calculations by the authors, 1995). These boundary conditions lead to magma rise speeds of \( u \sim 10^{-1} \text{ to } 1 \text{ m s}^{-1} \), a similar range to that of subaerial eruptions [Wilson and Head, 1981], and volume eruption rates \( Q \sim 10^{-2} \text{ to } 1 \text{ m}^3 \text{ s}^{-1} \text{ m}^{-1} \). The values of \( Q \) are given in detail as a function of \( W \) in Table 1.
that cooling fronts have penetrated it, has decreased from an initial high value to a critical value, $G_z \approx 300$. For ocean floor eruptions, this value requires slight revision upward, leading to slightly shorter flow lengths. This is due to the enhanced cooling that takes place in the submarine environment during the first few tens of minutes after eruption [Head and Wilson, 1986; Griffiths and Fink, 1993]. However, $G_z = 300$ will be appropriate for all longer flows (unpublished calculations by the authors, 1995). If it is assumed that a fissure vent eruption always gives rise to a sheet flow, that is, that the flow width is exactly equal to the length of active fissure, the Gratz number criterion leads to a maximum flow length $L_f$ given by

$$L_f = \left[ \frac{Q^{1/2}/(\kappa \ G_z^2)}{[3 \eta V/(\rho g \sin \alpha)]^{1/4}} \right],$$

where $Q$ is the volume eruption rate per unit horizontal length of active fissure, $\kappa$ is the thermal diffusivity of the lava ($\approx 10^{-6}$ m$^2$ s$^{-1}$) and $g$ is the acceleration due to gravity. The mean flow thickness corresponding to this length is $D_f = \left[ \frac{Q^{1/2}/(\kappa \ G_z^2)}{[3 \eta V/(\rho g \sin \alpha)]^{1/4}} \right]^{1/2}$, and the time $t_f$ required to emplace one cooling limited sheet flow unit is found from continuity requirements to be $(L_f D_f/Q)$.

Table 1 shows some expected flow properties for the range of dike widths and corresponding effusion rates mentioned earlier ($Q \approx 10^{-2}$ to $1 m^3 s^{-1}$). We use a mean substrate slope of $\sin \alpha = 0.186$ (estimated from bathymetric maps of the MAR AVR; [Purdy et al., 1999]), an apparent magma density of 1500 kg m$^{-3}$ to take account of buoyancy effects of the seawater (density 1000 kg m$^{-3}$) on the slightly vesicular lava (density $\approx 2500$ kg m$^{-3}$), and a magma viscosity of $3 \times 10^4$ Pa s, a value appropriate to a basaltic lava flow a few kilometers from its vent (unpublished calculations by the authors, 1995). A wide range of flow lengths and thicknesses is predicted, flow lengths being most strongly dependent on the feeder dike thickness. Equations (3) and (4) show that both length and thickness depend only on the cube root of the magma properties (density and viscosity) and the ground slope: an eightfold change in any of these parameters leads to only a twofold change in the flow dimension. Note that we assume that flow units are never fed by more than one active dike simultaneously. The propagation of a dike from a reservoir always changes the state of strain, and hence the stress conditions, in such a way as to minimize the chances of a second dike propagating in the vicinity of the first [Pollard, 1987; Rubin and Pollard, 1987]. At the lower end of the effusion rate range, individual flow units are relatively
equidimensional, being only a few tens of meters long and ~1 m thick. The distinctive morphology of pillow lavas [Walker, 1992; Griffiths and Fink, 1992; Gregg and Fink, 1995; Schmincke and Bednarz, 1990] is directly implied by these flow lengths and thicknesses.

Evolution of Eruption Conditions

During an eruption, the commonest circumstance in the magma reservoir is that the reservoir pressure decreases with time. In rare cases, the excess pressure in the magma reservoir feeding an eruption will remain at its initially high value as the eruption proceeds; this occurs when the supply rate from the mantle to the reservoir is unusually high [Parfitt and Head, 1993]. Eventually, if the eruption were able to continue long enough, the pressure would approach a near-constant value at which the resupply rate from the mantle on average equaled the volume eruption rate. This appears to be a relatively rare circumstance in subaerial eruptions, because any decline in pressure will mean that the flow speed of the magma through the dike system will decrease, and the effects of cooling through the dike walls will become progressively more important at any given depth below the vent [Bruce and Huppert, 1989, 1990; Carrigan et al., 1992]. The melt in contact with the wall will become more viscous as it approaches the surface; its rheology will become increasingly non-Newtonian and it may begin to develop a finite yield strength [Pinkerton and Stevenson, 1992]. As a result, the average flow speed and the erupted flux will decrease even further and eventually the eruption will stop even though there is still a finite excess pressure in the magma reservoir [Parfitt and Wilson, 1994]. This cessation due to cooling rather than relaxation of the reservoir should be more common in submarine than subaerial eruptions: overall dike cooling rates are relatively high in the oceanic environment because of the ubiquitous presence of hydrothermally circulating water in vein systems pervading the dike complex [Nehlig, 1993; Lowell et al., 1995].

Cooling will be most important where a dike is narrowest. The three-dimensional shapes of ideal dikes [Pollard, 1987] are such that the smallest widths are expected to be found near the ends of the fissure along which the dike emerges onto the seafloor. However, in practice significant changes in width occur along the fissure vents due to random fluctuations in the elastic properties of the host rocks. Another factor is dike splitting and rotation of dike segments at shallow levels in response to changes in orientation of the least compressive stress, leading to en echelon outcrops. Enhanced cooling, especially near the surface, in the narrower parts of such systems leads to a progressive concentration of eruptive activity into the wider parts of the dike (Figures 4a and 4b) [Wilson and Head, 1988], where the higher magma rise speed coupled with the greater width leads to a much greater volume flux per unit length (Table 1). We anticipate that this concentration of activity due to near-surface cooling will be expressed much more strongly on the seafloor than on land: the rise speed of a typical basaltic magma over the last hundred meters of its ascent on land will be of order 50 m/s, because the melt will have undergone fragmentation to feed a lava fountain. The undisrupted and only slightly vesicular magma of a corresponding seafloor eruption will rise at less than 1 m/s, as shown above, and cooling processes will have 2 orders of magnitude more time to operate.

When eruptions occur from the active segments of partially cooled dikes, the flows which form are likely to be wider than the length of the active part of the fissure, and it is not appropriate to treat them as sheet flows as was done in the previous section. Instead we use formulae which treat the vent as a point source from which lava spreads laterally and down slope to form a channelized flow which has a slowly varying width determined by the effusion rate and the temperature-dependent rheological properties of the lava. The analysis of Pinkerton and Wilson [1994] then leads to a predicted maximum cooling-limited flow length \( L_p \) given by

\[
L_p = K [FD_p^{2/3} \omega^{9/11}] \tag{5}
\]

where \( F \) is the total volume flux from the vent; \( K \) is a constant weakly dependent on the rheological properties of the lava and having a value close to 1900 s^{9/11} m^{-2}. Here \( D_p \) is the flow thickness given by

\[
D_p = \frac{\sigma (\rho g \sin \alpha)}{\pi} \tag{6}
\]

where \( \sigma \) is the apparent yield strength of the lava, ~5000 Pa for intermediate length basaltic flows (unpublished calculations by the authors, 1995). To obtain values of \( L_p \) from (5) we require values of \( F \) and hence some assumption about how the total flux from the vent is related to the effusion rate per unit length of fissure. We assume that the active fissure length is 10 times the width of the dike, a ratio consistent with commonly observed fissure vent geometries on subaerial volcanoes once vent localization due to cooling has occurred [Wilson and Head, 1988; Heslop et al., 1989], and hence use \( F = (10 W Q) \). Note that (6) implies a constant flow thickness (~1.8 m using the present magma properties and substrate slope), the consequence of our having neglected the progressive increase of the apparent yield strength and hence flow thickness of the lava as it travels away from the vent and cools. Table 1 contains the results of modeling localized vents and shows that the flow lengths calculated from (5) (point sources) are similar to those implied by (3) (fissure sources) for short flows but are smaller by a factor of up to 3 for long flows.

Continued Eruption from Central Vents

Once activity becomes concentrated into a few locations along a fissure system, each behaves more nearly as a point source (Figures 4a and 4b) than a line source of lava. The mechanics of lava flow formation therefore change significantly. Continuity requires that flows spreading radially (e.g., in pie-piece-like segments) become thinner as they spread, and so they are much more susceptible to cooling than sheet or channeled flows. For purely radial motion, the treatment by Wilson and Parfitt [1993] of flows spreading out from a central point source (Figures 4a and 4b) [Wilson and Head, 1988], where the higher magma rise speed coupled with the greater width leads to a much greater volume flux per unit length (Table 1). We anticipate that this concentration of activity due to near-surface cooling will be expressed much more strongly on the seafloor than on land: the rise speed of a typical basaltic magma over the last hundred meters of its ascent on land will be of order 50 m/s, because the melt will have undergone fragmentation to feed a lava fountain. The undisrupted and only slightly vesicular magma of a corresponding seafloor eruption will rise at less than 1 m/s, as shown above, and cooling processes will have 2 orders of magnitude more time to operate.

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Continued Eruption from Central Vents

Once activity becomes concentrated into a few locations along a fissure system, each behaves more nearly as a point source (Figures 4a and 4b) than a line source of lava. The mechanics of lava flow formation therefore change significantly. Continuity requires that flows spreading radially (e.g., in pie-piece-like segments) become thinner as they spread, and so they are much more susceptible to cooling than sheet or channeled flows. For purely radial motion, the treatment by Wilson and Parfitt [1993] of flows spreading out from a central point source (Figures 4a and 4b) [Wilson and Head, 1988], where the higher magma rise speed coupled with the greater width leads to a much greater volume flux per unit length (Table 1). We anticipate that this concentration of activity due to near-surface cooling will be expressed much more strongly on the seafloor than on land: the rise speed of a typical basaltic magma over the last hundred meters of its ascent on land will be of order 50 m/s, because the melt will have undergone fragmentation to feed a lava fountain. The undisrupted and only slightly vesicular magma of a corresponding seafloor eruption will rise at less than 1 m/s, as shown above, and cooling processes will have 2 orders of magnitude more time to operate.

When eruptions occur from the active segments of partially cooled dikes, the flows which form are likely to be wider than the length of the active part of the fissure, and it is not appropriate to treat them as sheet flows as was done in the previous section. Instead we use formulae which treat the vent as a point source from which lava spreads laterally and down slope to form a channelized flow which has a slowly varying width determined by the effusion rate and the temperature-dependent rheological properties of the lava. The analysis of Pinkerton and Wilson [1994] then leads to a predicted maximum cooling-limited flow length \( L_p \) given by

\[
L_p = K [FD_p^{2/3} \omega^{9/11}] \tag{5}
\]

where \( F \) is the total volume flux from the vent; \( K \) is a constant weakly dependent on the rheological properties of the lava and having a value close to 1900 s^{9/11} m^{-2}. Here \( D_p \) is the flow thickness given by

\[
D_p = \frac{\sigma (\rho g \sin \alpha)}{\pi} \tag{6}
\]
Figure 4. Geometry of a typical dike intruding the Kilauea East Rift Zone showing the connection from a dike approaching the surface to several areas on the surface where surface vents and related deposits have been produced. (a) View normal to the strike of the dike and rift axis. (b) View generally along rift zone and normal to dike. Magma penetrates to the surface, forms a curtain of fire, the upper narrowest parts of the dike solidify rapidly, the eruption centralizes along the widest part of the dike and builds a local cone. During edifice building the dike serves as a "planar magma storage zone," not a large local reservoir. Dike height averages -1-3 km and width is <3 m; gas exsolution begins at about 230 m depth. From Wilson and Head [1988].

3000 m³ s⁻¹. Using the same values for η, p and α as before, (7) predicts values of R in the range 160 to 2200 m, and the corresponding flow unit thicknesses T lie between about 1.0 and 2.0 m. If eruptive activity continues for a long enough time under these conditions, a stack of radially spreading piece-like cooling-limited flow units could build up to form a roughly circular edifice. As the eruption rate decreases toward the end of the activity, flow unit lengths will be smaller and the width of the growing edifice will decrease. The evolution of the topographic profile will be a very nonlinear function of the decreasing effusion rate as cooling narrows the effective dike width, the magma reservoir pressure decreases, and the edifice flank slope changes.

Fractures, Fissures, and Graben

Near-surface stress fields associated with dike emplacement can produce a wide range of linear structures, including simple fractures associated with dikes propagating to the surface, graben with horizontal extensions of the order of 2/3 the mean dike width for dikes stalling near the surface [Mastin and
Figure 5. Width of a planar dike that can just freeze (i.e., reach the solidus temperature along its center line) in a given time when it is intruded at a temperature close to the magma liquidus into country rocks that are cooler by 800, 200, and 100 K. From Wilson and Head [1988].

Pollard, 1988; Rubin, 1992, Head and Wilson, 1993], and larger normal faults and graben reflecting subsurface volume changes in the underlying upper crust and shallow magma reservoir (the linear equivalent of usually more equidimensional subaerial calderas) (Figures 6 and 7). Care must be taken to identify these and distinguish them from the abundant adjacent linear faults and fractures related to spreading and rift valley margin deformation.

Discussion and Interpretation

On the basis of these theoretical considerations, we now compare predictions (Figures 4-8) to observed landforms and structures (Figures 1-3). We assume that initial dike emplacement occurs through a magma-filled crack approaching the surface with a range of widths similar to those observed in several analogous settings (0.2-3 m). Some dikes will intrude only to the upper part of the crust and create a near-surface stress field that results in broad small-scale topography [Ryan, 1988, Figure 6] and/or fractures and graben. Other dikes may penetrate to the surface only in several places, creating fractures or open fissures in a zone above the top of the dike (Figure 8a). Dikes in the range of widths considered here would create surface fracturing only if they penetrated to depths of less than a few tens of meters.

When magma does reach the surface, theory predicts that the narrowest parts of the dike (< 0.3 m) will be characterized by very low effusion rates and flows less than about 50 m in length and thicknesses of up to ~0.8 m (Table 1, Figure 8b). These narrowest portions of the dike will be the first to cool; before cessation, eruptions would produce thin flow veneers or perhaps small hummocky edifices a few meters high and several tens of meters wide (Figure 8b). These resulting "walls of pillow mounds" are the submarine analog of the products of subaerial "curtains of fire." Using conductive cooling models for the dike [e.g., Wilson and Head, 1988], we estimate that the typical maximum duration of activity along such a segment would be of the order of 9-18 hours (Figure 5).

Other parts of the dike characterized by narrow-intermediate dike widths (~0.3-0.5 m) would produce low effusion rate flows up to about 1 m thick and with a range of flow lengths up to about 300 m; these flows should accumulate to produce edifices up to 600 m in width, with heights depending on eruption duration and the total number of accumulated flow units. Eventually, due to waning magma supply or to general heat...
controlling factor, this number is close enough to the 40 units

cooling-limited sheet flow unit from a 0.5-m-wide dike is 1.6

emplaced in one eruption. If dike cooling is the only

hours (Figure 5), so about 32 successive flow units could be

Table 1) flows, they would only do so if the eruption continued

produce very long (many tens of kilometers - see L values in

shape seamounts, and a large negative gravity anomaly. This

configuration would be consistent with a larger reservoir,

relatively larger dike widths, and higher effusion rates

characterized by relatively high effusion rates producing flows

m), the model predicts that fissure eruptions should be

required to explain the individual flow dimensions [Wilson and

At the widest end of the range of dike widths treated here (1-

m), the model predicts that fissure eruptions should be

characterized by relatively high effusion rates producing flows

a few meters thick and up to many tens of kilometers in length

(Figure 8d). These are almost certainly equivalent to the

smooth sheet-like flows that are commonly observed in the
topographic lows adjacent to the axial volcanic ridges (Figure
3f). Indeed, Smith et al. [1995] have noted that one segment
(6) of their study area has no axial volcanic ridge but instead is
characterized by smooth flows, smooth flat-topped and hat-
shaped seamounts, and a large negative gravity anomaly. This
configuration would be consistent with a larger reservoir,
relatively larger dike widths, and higher effusion rates
producing distributed and centralized sheet-like flows rather
than a distinctive axial volcanic ridge (Figure 9).

It should be noted that, although 2-3 m wide dikes could
produce very long (many tens of kilometers - see Lp values in
Table 1) flows, they would only do so if the eruption continued
for long enough (i.e., the magma reservoir volume was large
enough) to allow the flow fronts to reach such great lengths.
The fact that sheet flows on the ocean floor are rarely longer
than a few kilometers [e.g., Kidd, 1977; Klitgord and Mudie,
1974; Moore et al., 1974; Fornari et al., 1985; Fornari, 1986;
Macdonald et al., 1989; Applegate and Embley, 1992] implies

loss to the dike walls, these portions of the dike will solidify
on timescales no longer than 18-50 hours (Figure 5). The
somewhat greater widths of these narrow-intermediate dikes
favor longer dike segments with this geometry and this will
lead to adjacent along-strike eruptions with similar
characteristics, producing aligned edifices, overlapping
edifices, and continuous ridges (Figure 8c). Eruption products
with these characteristics are very similar to the observed
hummocky mounds, and when aligned, to hummocky ridges
(Figures 3c and 3d).

Hummocky ridges are up to 50 m in height and thus should
be built up of at least 40 flow units. These could in principle
be successive cooling-limited flow units emplaced during a
single eruptive event or the accumulated results of several
events. To explore how many events might be needed we
note, from Table 1, that the time required to emplace a single
cooling-limited sheet flow unit from a 0.5-m-wide dike is 1.6
hours. The dike will survive against cooling for about 50
hours (Figure 5), so about 32 successive flow units could be
emplaced in one eruption. If dike cooling is the only
controlling factor, this number is close enough to the 40 units

predicted to be required to build the largest ridges to suggest
that essentially all such ridges can be built in single
eruptions.

At wider parts of the dike (~0.5-1.0 m), magma flow will be
characterized by higher volume fluxes (Figure 8c) and flows
with thicknesses up to ~2.6 m and flow lengths in the range
250-5000 m. As dike cooling progresses, eruptions should
centralize to locations characterized by these latter (~0.5-1.0
m) dike widths. Continued eruptions of radially spreading,
cooling-limited flows from these centralized locations could
build edifices up to about 4 km in diameter (equation (7)), with
heights and shapes depending on the number and lengths of
accumulated flows. These properties are most similar to those
of the near-circular edifices or seamounts observed along
portions of the AVRs, features which are often continuous
with the smaller hummocky ridges (Figures 2, 3d, 3e, 8c, and
9). Continued eruptive activity produces a stack of radially
spreading, narrow pie-piece-like, cooling-limited flow units
which will build up to form a roughly circular edifice. A
general decrease in eruption rate toward the end of the activity
causes flow unit lengths to be smaller and the width of the
width is 1.5 times the total extension. (b) Relationship
between the depth to the top of the dike and the width of the
graben it produces. Total amount of horizontal surface strain

Figure 7. (a) Sketch illustrating the relationship between
the horizontal extension, across a graben (shown distributed
equally over the two bounding faults) and the width (W) of the
steeply dipping dike which gives rise to the graben. The dike
width is 1.5 times the total extension. (b) Relationship
between the depth to the top of the dike and the width of the
graben it produces. Total amount of horizontal surface strain

Figure 8. Block diagrams illustrating the model for the intrusion and eruption of dikes on the seafloor at a slow spreading MAR spreading center. (a) Shallow dike intrusion. (b) Narrow dike reaches surface. (c) Variable-width dike reaches surface. (d) Wide dike reaches surface. Many such flows are likely to be volume-limited because flows emerging from dikes of >1 m width are predicted to be of excessive length to be cooling-limited (see Table 1); for example, a single flow from a 2-m-wide dike would produce a volume approximately 16% of a typical AVR before its cooling-limited length was reached. Any of these shallow dike intrusion and extrusion events is sufficient to generate and maintain local hydrothermal circulation events in suitably porous crustal rocks [Delaney, 1987]. On the basis of likely dike widths and predicted cooling times, hydrothermal vents might remain active for a period of weeks to months following cessation of an eruption or intrusion.
Figure 9. Axial volcanic ridge (AVR) architectural elements. Axial volcanic ridges are made up of a number of individual dike emplacement events, each with different dike widths and surface morphologic manifestations. In this block diagram, the examples illustrate the basic building blocks of the AVRs. (Left) A narrow dike has produced a discontinuous wall of pillow mounds, which may appear to be discrete features unrelated to each other except for being aligned or possibly connected by fractures or fissures. (Middle) A dike of intermediate and variable width has been intruded and through the course of its cooling has produced pillow mounds, hummocky ridges, and finally, a central small shield at the widest place in the dike. In some places fractures and small grabens, the surface manifestation of the dike where it has not penetrated to the surface, connect the features. (Right) A wide dike has penetrated to the surface and become the source for high-effusion rate sheet flows which extend down the side of the AVR, covering previous flows and structures, and surrounding and burying preexisting volcanic edifices to form kipukas; as the sheet flows reach the low between the AVR and the inner valley walls, they move laterally and partially fill the floor. The axial volcanic ridge is built from a number of these events and is a complex mosaic of volcanic landforms and deposits and dike-intrusion-related deformation structures linked to individual events with different dike widths. Lateral variation in AVR characteristics within and between AVRs could be simply due to the relative proportions of dike widths, but the observed range of features requires no more than variation over a narrow range of dike widths (e.g., 0.2 to 3 m). For example, the portion of the ridge between 25° and 29° N which contains no axial volcanic ridge, some small shield volcanoes, smooth flows, and a large negative gravity anomaly [Smith et al., 1995; Lin et al., 1990] could be due to the presence of somewhat wider dikes over a region of upwelling, and the predominance of high-volume sheet flows dominating the inner floor and obscuring the AVR and its components.

Figure 10. Models for the shape of small shield volcanoes. A range of shapes has been observed for the small shield volcanoes described in the text and seen in Figures 2, 3, 8c, and 9. Evolution of the topographic profile of these edifices will be very nonlinear and a function of (1) decreasing effusion rate as cooling narrows the effective dike width, (2) decreasing magma reservoir pressure, and (3) changes in the edifice flank slopes. In Figures 10a and 10b, flow through the dike has concentrated at the widest place in the dike and a series of lobate or relatively smoother flows have formed around the vent to produce, respectively, a hummocky shield or a smooth shield. In Figure 10b, systematically decreasing effusion rates due to dike cooling and/or decreasing magma reservoir pressure yield progressively shorter flows and a smooth topographic profile; the hummocky shield has a more nonlinear history of extrusion and its topography is also influenced by the morphology of preexisting pillow mounds. In Figure 10c, the production of a normal symmetrical shield is interrupted by a drop in magma reservoir pressure sufficient to cause cessation of transport of magma through the dike system; this produces a flat-topped shield. In Figure 10d, a hat-shaped shield is formed either by a nonlinear change in effusion rate in an ongoing eruption, or by the reoccupation of the vent region by a subsequent dike to produce a superposed edifice. Preexisting pillow mounds and ridges cause additional asymmetry in topographic profiles. In Figure 10e, a flat-topped shield with a central depression is formed by the rapid extrusion of flows to form a summit lava lake which subsequently breaches the rim and drains, or drains internally through near-surface fractures and the dike system. Thus a wide range of morphologies can be produced through simple variations in eruption parameters.
which extends this same distance along strike. For illustrative purposes the magma chamber is taken to have an elliptical cross section with height $X$ and width $2X$. Blake [1981] showed that there should be a range of relationships between the volume of magma erupted from the reservoir and the total reservoir volume, depending on whether the response of the country rocks was purely elastic or largely inelastic, and on this basis we postulate that the reservoir volume must typically be at least 30 times the volume of erupted magma. These relationships then imply that the reservoir dimension $X$ will be related to the product of the flow length $L_f$ and thickness $D_f$ (i.e., to the volume per unit length along strike of the fissure and reservoir) by

$$30 L_f D_f = \pi X (X/2), \text{ i.e. } X = 4.4 \left(\frac{L_f D_f}{30}\right)^{1/2} \tag{9}$$

The values of $X$ implied by this analysis are given in Table 1. For the smaller dike widths (up to 0.5-0.7 m) and corresponding shorter flow lengths (several hundred meters up to 1 km), the reservoir dimension is less than 100 m, which does not seem greatly at variance with the sizes of axial reflectors detected seismically below fast spreading ridges [e.g., Sinton and Patrick, 1992]. However, for wider dikes and hence longer flows, the reservoir cross section dimensions required to provide enough volume for cooling-limited flows escalate rapidly, reaching $\sim 7 \times 15$ km for a 3-m-wide dike. These large values make it clear that, even allowing for the simplified geometry we have assumed for the reservoir/dike/flow geometry in this illustration, cooling-limited flow lengths (which are the maximum lengths that can ever be reached) are not likely to exceed a few kilometers, consistent with these observations.

Summary and Implications

We conclude that the range of morphologic features observed along the axis of the Mid-Atlantic Ridge is consistent with the emplacement of dikes with widths similar to those observed in ophiolite complexes, on Iceland, and in seafloor examples. The consequent range of eruption conditions along a single dike and the behavior of the dike as the eruption evolves predict features which are similar in morphology and dimensions to observed hummocky mounds, hummocky ridges, bulbous and flat topped seamounts, and smooth flows. On the basis of the size of the axial volcanic ridge and the observed often parallel hummocky ridges, each AVR is the product of multiple dike intrusion and eruption events which vary both between events and along strike in a single event (Figure 9). Thus a tremendous diversity in AVR morphology, structure, and architecture should be observed along the Mid-Atlantic Ridge. With spreading rates typical of the MAR, about one dike emplacement event would be expected every 40 years. Since not all dikes reach the surface, the actual mean interval between eruptions will be longer than this. Thus, given the cooling time for dikes in the submarine environment (the solidification time for a 3-m-wide dike intruded into hot country rock is $\sim 1$ year; Figure 5 [Wilson and Head, 1988]), MAR dikes should solidify between average dike emplacement events, with new events typically producing a separate dike and extrusive ridge on the larger axial volcanic ridge (Figure 9).

Typically, the morphology of rise crests differs between the Mid-Atlantic Ridge and the faster spreading East Pacific Rise, which is characterized by an abundance of smooth sheet flows, many fewer hummocky ridges and, only very rarely, small central seamounts. In the context of the processes and relationship to dike emplacement events and dike widths outlined above, this difference may be due to the faster spreading rates, anticipated wider, more continuous shallow reservoirs, wider dikes, and consequent higher effusion rate eruptions that would be typical of the EPR (Figure 8d). In addition, the faster spreading rate of the EPR means that subsequent overpressurization events leading to dike emplacement are more common (about one every 7 years for a spreading rate of 150 mm yr$^{-1}$ and a mean dike width of 1.0 m) and thus new dikes are more likely to reoccupy the sites of incompletely solidified older dikes, rather than producing a new pathway and new eruptive ridge each time. Together, these characteristics favor the development of accumulations of extensive smooth sheet flows, rather than the development of a distinctive series of hummocky ridges and seamounts as appears to be typical of the Mid-Atlantic Ridge (compare Figures 9 and 8d). In addition, this implies that the percentage of reoccupied dikes observed in ophiolites might provide evidence for spreading rates, with a higher percentage of reoccupied dikes reflecting higher spreading rates.

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