We investigate the origin of mid-ocean ridge morphology with numerical models that successfully predict axial topographic highs, axial valleys, and the transition between the two. The models are time-dependent, simulating alternating tectonic and magmatic periods where far-field extension is accommodated by faulting and by magmatism, respectively. During tectonic phases, models predict faults to grow on either side of the ridge axis and axial height to decrease. During magmatic phases, models simulate magmatic extension by allowing the axial lithosphere to open freely in response to extension. Results show that fault size and spacing decreases with increasing time fraction spent in the magmatic phase \( F_M \). Magmatic phases also simulate the growth of topography in response to local buoyancy forces. The fundamental variable that controls the transition between axial highs and valleys is the “rise-sink ratio,” \( (F_M/F_T)(\tau_T/\tau_M) \), where \( F_M/F_T \) is the ratio of the time spent in the magmatic and tectonic periods and \( \tau_T/\tau_M \) is the ratio of the characteristic rates for growing topography during magmatic phases (\( 1/\tau_M \)) and for reducing topography during tectonic phases (\( 1/\tau_T \)). Models predict the tallest axial highs when \( (F_M/F_T)(\tau_T/\tau_M) > 1 \), faulted topography without a high or valley when \( (F_M/F_T)(\tau_T/\tau_M) \sim 1 \), and the deepest median valleys when \( (F_M/F_T)(\tau_T/\tau_M) < 1 \). New scaling laws explain a global negative correlation between axial topography and lithosphere thickness as measured by the depths of axial magma lenses and microearthquakes. Exceptions to this trend reveal the importance of other behaviors such as a predicted inverse relation between axial topography and spreading rate as evident along the Lau Spreading Center. Still other factors related to the frequency and spatial pervasiveness of magmatic intrusions and eruptions, as evident at the Mid-Atlantic and Juan de Fuca ridges, influence the rise-sink-ratio \( (F_M/F_T)(\tau_T/\tau_M) \) and thus axial morphology.

Components: 12,871 words, 10 figures, 2 tables.

Keywords: mid-ocean ridge; magmatism; seafloor spreading; faulting; rifting.

Index Terms: 1032 Geochemistry: Mid-oceanic ridge processes (3614, 8416); 8178 Tectonophysics: Tectonics and magmatism; 8004 Structural Geology: Dynamics and mechanics of faulting (8118).

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1. Introduction

The contrast in mid-ocean ridge morphology between deep (1–2 km) axial valleys with heavily faulted flanks, and highstanding, axial ridges with minimal fault relief (<~102 m) has been a long-standing topic of interest [e.g., Macdonald, 1982, 1986]. Slow spreading (20–30 km/Ma, full rate) ridges such as the Mid-Atlantic Ridge (MAR) typically form axial valleys; fast spreading (>70 km/Ma) ridges typically form axial highs; and ridges spreading at intermediate rates such as the Galapagos Spreading Center (GSC) and Southeast Indian Ridge (SEIR) display the full range of morphologies from valleys, to faulted flat topography, faulted highs, and minimally faulted highs (Figure 1). Axial high topography probably originates from the buoyancy of low-density, partially molten material below the axis and relatively small nonlithostatic stresses [Madsen et al., 1984; Wang and Cochran, 1993; Magde et al., 1995; Buck, 2001; Shah and Buck, 2001]. Axial valleys most likely reflect large nonlithostatic stress that both depresses the axis and causes the valley flanks to be uplifted [e.g., Tapponnier and Francheteau, 1978; Lin and Parmentier, 1989; Chen and Morgan, 1990a, 1990b; Lin and Parmentier, 1990].

Seminal work by Chen and Morgan [1990a, 1990b] and Phipps Morgan and Chen [1993a, 1993b] led to the widely believed concept that axis morphology is a direct consequence of the thickness of the axial lithosphere, which in turn is controlled by the balance of heat added primarily by magma delivered to the axis and heat lost to the seafloor. The large magma flux at fast spreading ridges and hot spot-influenced, intermediate-spreading ridges [Chen and Phipps Morgan, 1996; Chen and Lin, 2004] is predicted to produce a thin lithosphere and hot lower crust. These researchers proposed that the hot, partially molten lower crust supplies the buoyancy to uplift the axis but is too weak to transmit stress from the stronger, peridotite mantle and cause the overlying lithosphere to break. In contrast, the smaller magma flux at slow-spreading ridges or some intermediate-spreading ridges promotes a thick lithosphere that is well coupled to the strong upper mantle. A thick section of lithosphere is therefore hypothesized to be subject to large stresses, allowing it to support a deep axial valley and large faults. Poliakov and Buck [1998] provide a nice review of this framework of understanding and extend it by examining some of the first numerical models that simulated spontaneous formation and evolution of normal faults.

While extension at the ridge axis is known to occur both tectonically and magmatically [e.g., see Carbotte and Macdonald, 1994; Escartín et al., 1999, and references therein] only recently has this tectonomagmatic cycle been examined in numerical models. Buck et al. [2005], for example, simulated the injection of dikes at the axis, which accommodate a fraction $M$ of the total rate of seafloor spreading. They predicted faulted axial valleys when $M < 1$ and axial highs when $M = 1$. Behn et al. [2006] simulated the accretion of “blade-like” dikes (widths that increase with depth) and showed that there is likely a feedback between axial valley topography and the frequency of dike injection.

Our companion paper [Behn and Ito, 2008, hereafter referred to as Paper 1] presents an in-depth study of the mechanics of faulting at mid-ocean ridges. Supporting the results of Buck et al. [2005], Paper 1 shows that $M$ is a key parameter controlling fault evolution, with lower values leading to larger more widely spaced faults. Lithosphere structure also influences faulting. Contrary to some previous views, we found that thicker lithosphere (with the same fractional rate of off-axis thickening) actually leads to smaller, not larger, fault size and spacing. Lower rates of thickening away from the axis, however, have the opposite effect. The inference made is that large $M$ and rapid off-axis cooling of the crust promotes small, closely spaced faults at magmatically robust ridges (e.g., fast to intermediate spreading) whereas lower $M$ with more subdued off-axis lithosphere thickening, since it is already thick beneath the axis, leads to the larger, more widely spaced faults.
at magmatically starved ridges (e.g., intermediate to slow spreading).

[6] Paper 1 also reveals a key limitation of simulating the tectonomagmatic cycle with the single parameter $M$, which assumes an infinitesimally small time interval between tectonic and magmatic events. Specifically, this approach results in non-lithostatic tension at the ridge axis sufficient to continually fault the lithosphere for all cases in which $M < 1$. Consequently, all simulations with $M < 1$ in Paper 1 (and in the work of Buck et al. [2005]) predict axial valleys of varying depths and no model predicts faulted transitional or faulted axial high topography such as that seen in Figure 1.

[7] The purpose of this paper is to define the primary factors that influence faulting as well as the characteristic morphologies of mid-ocean ridges: axial valleys, axial highs, and the transition between them. Our calculations simulate time-dependent crustal accretion and tectonism by specifying a time fraction $F_M$ that the axis is in a magmatic phase compared to the whole tectonomagmatic cycle. Individual effects of various factors including $F_M$, the duration of the tectonomagmatic cycle, lithosphere thickness, and spreading rate are quantified. We will show that the transition between axial high and median valley morphology does not depend on the amount of lithospheric coupling with the stiff mantle [Chen and Morgan, 1990a, 1990b] but rather on the “rise-sink ratio” of the crust, $(F_M/F_T)(\tau_T/\tau_M)$, where $F_T = 1 - F_M$ is the time fraction in the amagmatic or tectonic period and $\tau_T/\tau_M$ is the ratio of the rates at which the ridge axis rises during magmatic phases and sinks during tectonic phases. Variables such as spreading rate and axial lithosphere thickness are shown to affect axial morphology through their influence on the “rise-sink ratio.”

2. Method

2.1. Mechanical Model

[8] Full details of the numerical methods are described in Paper 1; here we outline the most salient similarities and specify the particular differences incorporated in this study. The mathematical variables are listed in Table 1. To simulate two-dimensional (2-D) lithosphere accretion, faulting, and topography we numerically solved the conservation equations of mass and momentum for a visco-elasto-plastic material (Figure 2). We used the same hybrid finite element method (FLAC) and rheological formulations as in...
Table 1. Summary of Variables Used

<table>
<thead>
<tr>
<th>Variable</th>
<th>Meaning</th>
<th>Value(s)</th>
<th>Units</th>
</tr>
</thead>
<tbody>
<tr>
<td>$C$</td>
<td>scaling constant in equation (7)</td>
<td>$7.92 \times 10^{14}$</td>
<td>Nm</td>
</tr>
<tr>
<td>$F_M$</td>
<td>$P_M/P$, time fraction of magmatic period</td>
<td>0.7–1</td>
<td>—</td>
</tr>
<tr>
<td>$F_T$</td>
<td>$P_T/P = (1 - F_M)$, time fraction of tectonic period</td>
<td>0–0.3</td>
<td>—</td>
</tr>
<tr>
<td>$H_{400}$</td>
<td>depth of 400°C-isotherm beneath the axial seafloor</td>
<td>1.3–4.0</td>
<td>km</td>
</tr>
<tr>
<td>$h$</td>
<td>long wavelength axial height relative to adjacent seafloor</td>
<td>—</td>
<td>km</td>
</tr>
<tr>
<td>$h_M$</td>
<td>axial height in isostatic equilibrium</td>
<td>—</td>
<td>km</td>
</tr>
<tr>
<td>$M$</td>
<td>fraction magmatic extension</td>
<td>—</td>
<td>—</td>
</tr>
<tr>
<td>$P_M$</td>
<td>time period of magmatic phase</td>
<td>—</td>
<td>ka</td>
</tr>
<tr>
<td>$P_T$</td>
<td>time period of tectonic phase</td>
<td>—</td>
<td>ka</td>
</tr>
<tr>
<td>$P_{M+T}$</td>
<td>duration of tectonomagmatic cycle</td>
<td>10–502.5</td>
<td>ka</td>
</tr>
<tr>
<td>$w_T$</td>
<td>depth of deepest axial valley</td>
<td>—</td>
<td>km</td>
</tr>
<tr>
<td>$\Delta x$</td>
<td>fault heave</td>
<td>—</td>
<td>km</td>
</tr>
<tr>
<td>$\Delta S$</td>
<td>average fault spacing</td>
<td>—</td>
<td>km</td>
</tr>
<tr>
<td>$U$</td>
<td>half spreading rate</td>
<td>—</td>
<td>km/Ma</td>
</tr>
<tr>
<td>$x_{AF}$</td>
<td>furthest distance from axis a fault remains active</td>
<td>10–75</td>
<td>km</td>
</tr>
<tr>
<td>$z_{400}$</td>
<td>depth of 400°C isotherm below seafloor</td>
<td>—</td>
<td>km</td>
</tr>
<tr>
<td>$\mu_M$</td>
<td>model maximum viscosity in accretion zone</td>
<td>—</td>
<td>Pa s</td>
</tr>
<tr>
<td>$\tau_T$</td>
<td>timescale for reducing topography during tectonic episodes</td>
<td>—</td>
<td>ka</td>
</tr>
<tr>
<td>$\tau_M$</td>
<td>timescale for topographic growth during magmatic periods</td>
<td>—</td>
<td>ka</td>
</tr>
</tbody>
</table>

Figure 2. (a) Model geometry and boundary conditions and example calculations for cases in which thermal structure is (b)–(d) imposed and (e)–(g) computed. In each case, depth to the 400°C isotherm is $H_{400} = 1.3$ km, time fraction of the magmatic phase is $F_M = 0.8$, and half spreading rate is $U = 25$ km/Ma. Figures 2b and 2e show surface topography (solid) and isostatic topography (dashed); Figures 2c and 2f show accumulated plastic strain (colors) and flow direction (arrows); and Figures 2d and 2g panels show density structure (colors) and temperature (contoured every 200°C).
Paper 1 and developed recently by Buck, Lavier, and Poliakov [e.g., Poliakov and Buck, 1998; Lavier and Buck, 2002; Buck et al., 2005]. The model domain was rectangular and centered on the ridge axis (Figure 2). In most cases the domain was 60 km wide by 15 km deep with the smallest grid elements (0.2 km by 0.2 km) located within a distance of 1 km by 1 km of the axis and a depth z ≤ 7 km. Outside of this region, grid size increased toward the bottom corners of the model box to a maximum size of 0.8 km by 0.8 km. For calculations with the thickest lithosphere (subaxial depth of 400 km isotherm being \(H_{400} = 4\) km, see section 2.3), we required less resolution (i.e., faults were more widely spaced and extended deeper) as well as a larger box to encompass the lithosphere’s larger flexural wavelength and full temperature structure. For these calculations, the box was 100 km by 20 km with minimum grid dimensions of 0.33 km by 0.33 km within x ≤ 12 km of the ridge and z ≤ 9 km.

Faulting was simulated with elastoplasticity, using Mohr-Coulomb failure criteria. Strain localization was promoted by reducing cohesion (shear stress for brittle failure at zero pressure) as a linear function of plastic strain. Cohesion was reduced linearly from 44 MPa to 2 MPa over a time that is well less than the duration individual faults remained active. For most of our calculations (i.e., \(F_M ≤ 0.9\)) this condition was met by reducing the cohesion over a total fault displacement of ~0.3 km (i.e., plastic strain of 0.39 for a grid spacing of 0.2 km). Calculations for higher \(F_M\), required a cohesion weakening over 0.1–0.2 km of slip (plastic strain of 0.10–0.25). Our results were insensitive to the rate of weakening with plastic strain, provided that faults weakened completely and over a time that was shorter than the specified fault healing time (320 ka).

Seafloor spreading was simulated by imposing horizontal motion on the left and right sides of the model box at a rate of \(U\) which defines the half-spreading rate. The top boundary was stress free and the base was maintained at the hydrostatic stress (which changes as the base moves up and down with the Lagrangian grid). The density contrast at the surface was set to that between mantle and water (2300 kg/m³). This treatment is equivalent to the simplifying assumption that the crust, which is less dense than mantle, is of uniform thickness everywhere at each point beneath the seafloor.

2.2. Method of Simulating the Tectonomagmatic Cycle

In contrast to Paper 1 in which a fraction \(M\) of the total spreading rate was accommodated by magmatism at all times, in this study, we explicitly simulated temporal variations in magmatism. Amagmatic or tectonic periods involved zero magmatic accretion (i.e., \(M = 0\)) for a time period \(T\). During tectonic periods, tension built in the lithosphere, the ridge axis was drawn downward, and faults formed and grew (Figures 3a–3b). Magmatic periods were simulated by allowing the two columns of elements that define the ridge axis to open freely within the crust (z ≤ 7 km) in response to plate spreading (Figures 3c–3d). This was done by removing all nonlithostatic stresses in these elements at each time step. In addition, magmatic

Figure 3. (a and c) Horizontal speed of the surface; (b and d) strain rate (as second invariant of the strain rate tensor). (left) During an amagmatic phase, extension is accommodated along a single fault. (right) During the magmatic phase extension is accommodated within the accretion zone. Figures 3a–3d are different time steps of the same model run with \(H_{400} = 1.3\) km, \(F_M = 0.8\), and \(U = 25\) km/Ma (same as in Figures 2b–2d).
periods involved limiting viscosity to a maximum of \( \eta_M \) within the magmatic zone of \( 1 \times 1 \leq 2 \) km. When the parameter \( \eta_M \) was large (10\(^{27}\) Pa s), the rheology was not perturbed compared to the pressure-temperature dependence that defined the basic rheology (i.e., the rheology everywhere else and in the whole model during tectonic phases). When \( \eta_M \leq 10^{20} \) Pa s, the whole magmatic zone deformed viscously and axial topography tended to rise in response to local buoyancy forces. Parameter \( \eta_M \) does not represent a laboratory-based rheological property but rather is a parameter that accounts for the time- and space-integrated effects of magmatic intrusions and eruptions, which both weaken and thicken the crust. Here \( \eta_M \) controls the characteristic timescale \( \tau_M \) or the rate \( 1/\tau_M \) that it takes magmatism to build topography toward the level of isostatic equilibrium (e.g., Figure 2).

[12] The model magmatic periods represent times when the liquid magma volume that is present within or transported through the crust is high enough to minimize nonlithostatic stresses and to cause positive growth of topographic. These periods include but are not limited to times of active eruptive episodes. Correspondingly, tectonic periods represent times in which magma volume and/or flux are relatively low such that nonlithostatic tension increases and topographic growth is negative. Magmatic periods last a time \( P_M \), tectonic periods last \( P_T \). Thus, the whole tectonomagmatic cycle lasts \( P = (P_M + P_T) \), and the corresponding time fractions of magmatic and tectonic periods are \( F_M \equiv P_M/P \) and \( F_T \equiv P_T/P = (1 - F_M) \), respectively. We examined values of \( F_M \) ranging from 0.7 to 1, which is consistent with estimates of spreading rate fractions of \( M = 0.7 - 0.9 \) at heavily faulted ridges from studies of seismic moment release [Solomon et al., 1998] and fault relief [Escartín et al., 1999].

[13] These strongly time-dependent calculations required dynamic adjustments of model time steps to maintain quasi-static equilibrium. This was done by reducing time steps when the maximum normalized acceleration (acceleration of a node divided by the acceleration that would be caused by the sum of the forces on the node if they acted in the same direction) exceeded 0.03. This adjustment was typically needed after remeshing steps and during the transition between tectonic and magmatic phases. In so doing, time steps typically varied between \( \sim 10^{-5} \) and \( \sim 10^{-4} \) ka with a total of \( 30 - 40 \times 10^6 \) steps needed for a typical calculation simulating 1.3 Ma. Another measure that was taken to stabilize solutions was to impose the transition from the tectonic to magmatic phase gradually over 1 ka in model time. By reducing both the nonlithostatic stresses in the central dike elements and the log of the viscosity cutoff in the magmatic zone (toward \( \eta_M \) linearly over this time, the large axial stresses that had built up during the tectonic phase decayed gradually toward those present during the magmatic phase. The transition from the magmatic back to the tectonic phase occurred over a single time step.

2.3. Temperature and Density Structure

[14] Temperatures are important in controlling thermal buoyancy and rheology as described in Paper 1. The thermal structure of our models includes a thin lithosphere beneath the axis that rapidly thickens on either side of the axis, consistent with geophysical studies of fast-intermediate spreading ridges [Toomey et al., 1994; Vera and Diebold, 1994; Dunn et al., 2000; Crawford and Webb, 2002; Canales et al., 2006], and segment centers of slow-spreading ridges [Dunn et al., 2005]. We used two ad hoc approaches to simulate such a thermal structure. The first approach was to impose temperatures as a function of distance \( x \) from the axis and depth \( z \) below the local seafloor (Figure 2d). In this case, temperature increases linearly from 0°C at \( z = 0 \) to 400°C at \( z = z_{400} \) (i.e., depth of 400°C isotherm), then to a maximum of 1300°C between \( z = z_{400} \) and \( z = 2.5z_{400} \) and remains uniform at 1300°C at greater depths. Depth \( z_{400} \) varies with distance \( x \) from the axis, being shallowest for \( 1 \times 1 \leq 1 \) km, increasing by 2.33 km at \( 1 \times 1 = 5 \) km, and then remaining at a constant, maximum value for \( 1 \times 1 > 5 \) km. The minimum of \( z_{400} \) for \( 1 \times 1 \leq 1 \) km defines the parameter \( H_{400} \), axial lithosphere thickness, for which we consider values of 1.3, 1.8, 2.5, and 4.0 km. This range implies lithosphere (i.e., the material that faults) with a minimum thickness slightly larger than the \(~1 \) km values determined from the shallowest axial magma lens at fast spreading ridges [e.g., Detrick et al., 1987] and a maximum thickness within the range of the slow-spreading Mid-Atlantic Ridge based on the deepest of microearthquakes (3–9 km) [Toomey et al., 1988; Kong et al., 1992; Wolfe et al., 1995; Barclay et al., 2001]. We will refer to these models as cases with imposed thermal structure.

[15] The second approach simulated a more realistic structure by solving the time-dependent, advection-diffusion equation for temperature with heat sources (Figure 2g and equation (2) of Paper 1).
Temperatures at the surface were maintained at 0°C and set equal to the steady state solution for a cooling half-space at the depths coinciding to the model base, subject to a maximum of 1300°C. Insulating boundaries were imposed on the left and right sides of the box. The initial conditions for cases with \( F_M = 1.0 \) was the steady state solution of a cooling half-space. The initial condition for cases with \( F_M < 1.0 \) was the final solution of the appropriate case with \( F_M = 1.0 \). As in Paper 1, we parameterized the effects of hydrothermal circulation by scaling the thermal diffusivity by a factor \( N_u \) [e.g., Phipps Morgan and Chen, 1993a, 1993b]. To achieve the values of \( H_{400} \) and similar variations in \( z_{400} \) as in our fixed temperature cases above we set \( N_u = N_{u0} \) for \( 1 \times 1 \leq 0.5 \text{ km}, N_u = 10 \) for \( 1 \times 1 = 4–5 \text{ km}, N_u = 1.0 \) (no hydrothermal circulation) for \( 1 \times 1 > 8 \text{ km}, \) and changed \( N_u \) linearly between the specified ranges of \( x \). Values for \( N_{u0} \) of 5.6 and 8.5 combined with the thermal and latent heat flux associated with 7 km of crust accreting at the spreading rate yielded \( H_{400} \) of 1.3 and 1.8 km, similar to the first two cases with imposed thermal structure. For simplicity, the thermal effects of magmatism where held constant in time; they did not vary between tectonic and magmatic phases. We will refer to these thermal models as cases with computed thermal structure. [18] From the predicted topography (example shown in Figure 2), we had to separate the short-wavelength component due to faulting and the long-wavelength component associated with axial morphology. Fault scarps were identified as the steepest features that dip toward the axis in Figure 2. The two key fault characteristics are fault spacing \( \Delta S \), which was measured by the distance between the bases of the faults, and fault heave \( \Delta h \), which was measured by the horizontal offset between the fault bases and peaks. To quantify the long-wavelength topography associated with axial highs or valleys, we measured the height of the axis above the surrounding seafloor, \( h \), using two methods. Method 1 measured axis height \( h_1 \) as the difference between the axis depth and the mean depth of the two fault peaks that were most distal from the axis on each plate. Method 2 measured axis height \( h_2 \) as the difference between the axis depth and the mean depth of the shallowest fault peaks on each side of the axis. Axial height \( h \) was defined as \( h = h_1 \) when \( h_1 > 0 \) and \( h = h_2 \) when \( h_1 < 0 \). Thus, \( h \) measures the height of axial topographic highs when \( h = h_1 > 0 \) and the depth of axial valleys when \( h = h_2 < 0 \) (\( h_2 \) is always less than \( h_1 \)).

3. Results

3.1. General Model Predictions

[19] Calculations were integrated in time so that all of the seafloor was created during the model runs (e.g., 1300 ka for \( U = 25 \text{ km/Ma} \)). Like in Paper 1, our simulations typically predict one fault to form on one side of the axis and remain active to a maximum distance \( x_{AF} \), at which point a new fault forms on the opposite plate. The axis tends to rise during magmatic phases and sink during tectonic phases in a saw-tooth pattern in time with amplitudes typically 0.01–0.02 km. Below we show how and why fault characteristics (i.e., spacing \( \Delta S \) and heave \( \Delta h \)) and long-wavelength topography \( h \) are controlled by five key parameters: (1) \( P \), duration of the tectonomagmatic cycle, (2) \( F_M \), magmatic time fraction, (3) \( H_{400} \), axial lithosphere thickness, (4) \( \eta_M \), which is proportional to the rate \( 1/\tau_M \) of topographic growth during magmatic phases, and (5) \( U \), half-spreading rate.

2.4. Measuring Long-Wavelength Height and Fault Characteristics

[17] The main output of our models is surface topography. Because repeated regridding of the Lagrangian mesh caused the sharp topography of faults to diffuse with time, we tracked topography using passive tracers. These tracers were initially placed along the surface of the model at 0.1 km spacing and then added at the axis and removed from the two sides of the model during the simulation of seafloor spreading. Velocities of each particle were computed by linearly interpolating the velocities between the finite element nodes; positions were computed using upwind differencing in time.
3.2. Varying the Duration of Tectonomagmatic Cycle $P = P_M + P_T$

[20] Figure 4 illustrates the key results of varying the duration of the tectonomagmatic period $P$ between 10.0 and 502.5 ka for $F_M = 0.8$, imposed thermal structure with $H_{400} = 1.3$ km, and in which tectonomagmatic period, $P = P_M + P_T$ is varied with shown values. Length of arrows indicates long-wavelength axial topography ($h$), pointing down to indicate depth of axial valley. (b) Long-wavelength axial topography and (c) fault spacing versus $P$. Dashed line is at 280 ka, which is the age contrast between successive faults at the preferred spacing of ~7.08 km. Solid curve is the distance of spreading during one tectonomagmatic cycle (i.e., $\Delta S = PU$). Error bars show maximum and minimum fault separation within each model run.

For $P > 162.5$ ka, $h$ becomes progressively more negative, and for $P \geq 160$ ka, $\Delta S$ increases linearly with $P$. For these longer cycles, fault spacing is controlled primarily by the temporal variability between tectonic and magmatic periods; i.e., faults form simultaneously on both plates, remain active for only one tectonic period, and then move away from the axis during the magmatic phase. Subsequent fault pairs form during the next tectonic period. The curve in Figure 4c is the product of $P$ and spreading rate $U$ and shows a good match with model fault spacing.

3.3. Importance of $F_M$ on Fault Spacing and Heave

[22] The remainder of the paper will focus on cases in which $P$ is small enough for fault spacing to be controlled mechanically by the force required to sustain active faults versus breaking new ones. Specifically, the tectonic period $P_T$ is held constant at 3 ka and we vary $P_M$ to change $F_M$; for $F_M = 0.7–0.975$, $P = 10–120$ ka, respectively. For $F_M = 1$, $P_T = 0.0$ ka.

3.3. Importance of $F_M$ on Fault Spacing and Heave

[23] Figure 5a summarizes our results for varying the magmatic time fraction $F_M$ in models with imposed thermal structure. Example topographic...
profiles are shown for cases in which $h_m = 5 \times 10^{18}$ Pa s and $H_{400} = 1.3$ km (Figure 5a). When the ridge remains in the magmatic phase at all times ($F_M = 1$) a distinct axial high forms without any faults (Figure 5a). Compared to the isostatic topography (dashed green), the calculation (solid blue) predicts a total variation in topography that is slightly greater and distributed over a broader distance from the axis. Both of these differences are consistent with the topography being controlled by the flexure of accreting lithosphere [Buck, 2001] the key cause for the dynamic topography is the resistance of initially curved lithosphere to unbend, which transfers some of the weight of the axis to its flanks, thus depressing the flanks much like “flexural moats”. The topographic profile for $F_M = 1$ shows good match with a separate calculation (red dotted curve) of an accreting elastic lithosphere, based on thin-plate theory [Shah and Buck, 2001] and using the same density structure and variation
in effectively elastic plate thickness (assumed to correspond to the depth of the 600°C isotherm).

[24] Faults form whenever \( F_M < 1 \). With \( \eta_M = 5 \times 10^{18} \) Pa s, axial height grows rapidly during magmatic phases and approaches the isostatic height. As \( F_M \) decreases, the amount of time spent in the magmatic phase decreases, and thus the total uplift is less. Also as \( F_M \) decreases, topography away from the axis is increasingly influenced by faulting. For the case shown in Figure 5a, long-wavelength topography can be characterized as faulted axial highs when \( 1.0 > F_M \geq 0.925 \), transitional when \( 0.9 \geq F_M \geq 0.85 \), and beginning to approach an axial valley when \( F_M = 0.8 \). Thus, \( F_M \) is one parameter that influences long-wavelength topography.

[25] The major influence of \( F_M \), however, is on the short-wavelength fault topography. Figures 5b and 5c show the results for our imposed temperature calculations with \( H_{400} = 1.3, 1.8, \) and \( 4.0 \) km for all values of \( \eta_M = 1.25 \times 10^{18} \) to \( 1 \times 10^{27} \) Pas. Models predict fault spacing \( \Delta S \) and heave \( \Delta x \) to increase with decreasing \( F_M \), in a manner that is reasonably well explained by the following scaling for fault spacing:

\[
\Delta S = \frac{F_M}{(F_M - 0.5)} x_{AF}, \quad (1)
\]

and heave

\[
\Delta x = \frac{(1 - F_M)}{(F_M - 0.5)} x_{AF}, \quad (2)
\]

where \( x_{AF} \) is the maximum distance that a fault moves away from the axis while remaining active (before a new fault forms on the opposite plate). These equations come from simple kinematic arguments formulated by Buck et al. [2005] and further developed in Paper 1. They have the same form as equations (3) and (4) of Paper 1 but with the temporal fraction \( F_M \) replacing the kinematic fraction \( M \) of the plate spreading rate that is accommodated magmatically. The ability to explain our model predictions with equations (1) and (2) implies that \( M \) is simply related to \( F_M \) (Figure 5d). We see no systematic deviation from these trends with \( \eta_M \), which indicates that the rate of axial topographic growth during magmatic phases does not strongly influence fault characteristics.

[26] Comparisons between model predictions and equations (1) and (2) are shown in Figures 5b and 5c. Each solid curve is for a constant, representative value of \( x_{AF} \) taken from the numerical model runs for a given value of \( H_{400} \). The force balance arguments developed in Paper 1 predict similar values of \( x_{AF} \) that vary by a small amount with \( F_M \) (dashed). Fault distance \( x_{AF} \), and thus \( \Delta S \) and \( \Delta x \) are all predicted to be larger for larger axial lithosphere thickness (Figures 5b and 5c). The cause can be understood in context the force balance arguments as follows. Off-axis distance \( x_{AF} \) is limited by the point at which the net tectonic pulling force \( f \) needed to keep the fault active exceeds the strength \( f_I \) of the intact lithosphere nearer to the axis (i.e., \( f > f_I \) at which point a new fault forms and takes over). The net pulling force \( f \) is the sum of those needed to bend the lithosphere near the active fault \( f_B \) and to overcome friction and strain-weakened cohesion on the fault \( f_F \) (i.e., \( f = f_B + f_F \)). The bending part \( f_B \) increases as the lithosphere thickens off axis, whereas \( f_F \) initially decreases due to cohesion loss with fault slip but then increases with lithosphere thickness once weakening is complete. Here \( x_{AF} \) is predicted to decrease with \( H_{400} \) if the rate of off-axis thickening is proportional to \( H_{400} \) because \( (f_B + f_F)/f_I \) increases toward 1 more rapidly with off-axis distance for larger \( H_{400} \) (Paper 1). However, in the cases considered here, the rate of off-axis thickening is not proportional to \( H_{400} \); it is held constant. Consequently, the fractional rates that lithosphere thickness and \( (f_B + f_F)/f_I \) increase off axis are smaller for larger \( H_{400} \), thus allowing \( (f_B + f_F)/f_I \) to remain <1 and the fault to remain active to a greater distance \( x_{AF} \) away from the axis. This result further illustrates our findings in Paper 1.

3.4. Influence of \( H_{400} \), \( \eta_M \), and \( U \) on the Transition Between Axial Highs and Median Valleys

[27] Figure 6 shows predicted bathymetry profiles in which \( U = 25 \) km/\( \)Ma, \( H_{400} = 1.3 \) km, \( F_M = 0.9 \), and the maximum viscosity parameter \( \eta_M \) (see section 2.2) in the magma zone is varied from \( 1.25 \times 10^{18} \) to \( 1 \times 10^{27} \) Pa s. Again, \( \eta_M \) does not change fault characteristics in a systematic manner. The major influence of \( \eta_M \) is on the long-wavelength topography. When the maximum viscosity is not altered from the basic temperature- and stress-dependent rheology (i.e., \( \eta_M = 1 \times 10^{27} \) Pa s), the growth of topography during the magmatic phases is suppressed by the strong axial lithosphere, even though the nonlithostatic stresses associated with extension are completely removed. The result is a prominent axial valley (lowermost profiles in Figure 6). As \( \eta_M \) is decreased to \( 1 \times 10^{19} \) Pa s, so does the characteristic timescale
for topographic growth during the magmatic phases ($t_M$ is proportional to $h_M$), and the axial valley is replaced by flat, faulted topography, which often characterizes the “transitional” morphology at intermediate spreading rates (e.g., Figure 1). With further reduction of $\eta_M$ below $1 \times 10^{19}$ Pa s, a faulted axial high emerges. Cases with imposed (Figure 6a) and computed (Figure 6b) thermal structure predict the same gross morphology but the latter predicts slightly greater axial heights overall. The main reason for this difference is that lithosphere continues to thicken away from the axis when temperatures are computed but remains constant in the imposed thermal models (see Figures 2d and 2g).

Figure 7 summarizes our findings of long-wavelength axial heights, $h$, for all calculations in which $\eta_M$ was varied. The results reveal an exponential dependence of $h$ on the shown function of $F_M$ and $\eta_M$. A conceptual way of understanding these results is to consider the isostatic adjustment of topography during a single magmatic period. At the beginning of this period (time $t = 0$), axial topography is fully depressed to a depth $w_T$ from the previous tectonic period. Then as magmatism occurs, the height of the ridge axis relaxes toward the height of isostatic equilibrium ($h_M$). Assuming only viscous relaxation (as opposed to the combined effects of viscoelastic plasticity, which is simulated in models), the height of the ridge axis evolves in time according to [e.g., Turcotte and Schubert, 2002]

$$h = -w_T + \left(w_T + h_M\right)\left(1 - \exp\left[-\frac{t}{t_M}\right]\right). \quad (3)$$

The relaxation timescale $t_M$ depends on the density contrast between mantle and water $\Delta \rho$, gravitational acceleration $g$, and the wavelength of the topography $\lambda$, and on viscosity $\eta_M$, according to

$$t_M = \frac{4\pi \eta_M}{\Delta \rho g \lambda}. \quad (4)$$
During a single magmatic cycle, we find that the $U_{ht} = 2.5 \cdot H_{h} = 1$ is on the right of $= (\text{different symbols}).$ Best fits to the model $P_{c0}/C_{0}/C_{2}$ varies from an empirical term is included to illustrate the $F = 2 H/C_{0}(H/C_{2})$ in equation (7). $F_{c0}/C_{2}$ are the characteristics rates that topography rises during magmatic phases and sinks during tectonic phases, respectively. Thus $P_{c0}/\tau_{M}$ and $P_{cT}/\tau_{T}$ represent the total (fractional) amounts that the axis can rise and sink, respectively. We therefore refer to equation (5) as the topographic “rise-sink ratio.”

The numerical results are well explained by a scaling law of the same form as (3),

$$h = -w_{T} + \left(\frac{w_{T} + h_{M}}{1 - \exp \left[ -\left(\frac{F_{M}}{F_{T}}\frac{\tau_{T}}{\tau_{M}}\right)\right]}\right),$$  \hspace{1cm} (6)$$

in which,

$$\left(\frac{F_{M}}{F_{T}}\right)\left(\frac{\tau_{T}}{\tau_{M}}\right) = \left(\frac{F_{M}}{F_{T}}\right)\left(\frac{C}{\eta_{M}H_{400}^{2}U}\right),$$  \hspace{1cm} (7)$$

where $U$ is half spreading rate and $C$ is an empirical constant, $C = 7.92 \times 10^{14} \text{ Nm}.$ In Figure 7, axial height $h$ is plotted versus $(F_{M}/F_{T})(\tau_{T}/\tau_{M})H_{400},$ which was varied by changing $\eta_{M}$ in equation (7). The $H_{400}$ term is included to illustrate the differences between cases with different values of $H_{400}$ (different symbols). Best fits to the model results are achieved when the maximum valley depth $w_{T}$ is estimated as the average of axial heights for all cases with $\eta_{M} = 1 \times 10^{12} \text{ Pa s},$ and

![Figure 7](image73x712to120x756)

**Figure 7.** Predicted axial height, $h$, for cases in which thermal structure is (a) imposed and (b) computed and $U = 25 \text{ km/Ma},$ each on different horizontal scales; Horizontal axis is rise-sink ratio times $H_{400};$ (equations (6) and (7)). Symbols distinguish results of calculations with different $H_{400}: 1.3 \text{ km}$ (circles), $1.8 \text{ km}$ (up-pointing triangles), $2.5 \text{ km}$ (down-pointing triangles), and $4.0 \text{ km}$ (squares). Color shading denotes groupings according to viscosity: (white) $\eta_{M} = 1 \times 10^{17}$ and $1 \times 10^{19} \text{ Pa s},$ (light) $\eta_{M} = 5 \times 10^{18} \text{ Pa s},$ and (dark) $\eta_{M} = 2.5 \times 10^{18}$ and $1.25 \times 10^{18} \text{ Pa s},$ except for the one square at $(F_{M}/F_{T})(\tau_{T}/\tau_{M})H_{400} \sim 78 \text{ km}^{2}$ for which $\eta_{M} = 5 \times 10^{18} \text{ Pa s}.$ Results for $F_{M} = 1$ are plotted at arbitrary locations on the right side of the diagram; they lie above the theoretical curves (equations (6) and (7)) as predicted by the effects of lithosphere flexure (see text). Asterisks show results for models with smaller faults (see text) shown in Figure 10 in which $\eta_{M}$ varies from $1 \times 10^{21}$ to $1.25 \times 10^{18} \text{ Pa s},$ from left to right.

While equation (3) describes the evolution of $h$ during a single magmatic cycle, we find that the time-integrated results of many tectonomagmatic cycles can be explained by replacing the dimensionless time in the exponential of (3) with

$$\frac{P_{M}/\tau_{M}}{P_{cT}/\tau_{T}} = \left(\frac{F_{M}}{F_{T}}\right)\left(\frac{\tau_{T}}{\tau_{M}}\right).$$  \hspace{1cm} (5)$$

This is the ratio of the time fractions for topography to rise, $P_{M}/\tau_{M},$ and sink, $P_{cT}/\tau_{T},$ where $\tau_{T}$ is the characteristic time for topography to sink during the tectonic phase and recalling that $F_{T} = 1 - F_{M}.$ The inverse of the timescales, $1/\tau_{M}$ and $1/\tau_{T}$ are the characteristics rates that topography rises during magmatic phases and sinks during tectonic phases, respectively. Thus $P_{M}/\tau_{M}$ and $P_{cT}/\tau_{T}$ represent the total (fractional) amounts that the axis can rise and sink, respectively. We therefore refer to equation (5) as the topographic “rise-sink ratio.”

![Figure 8](image345x209to535x324)

**Figure 8.** Red circles shows predicted long-wavelength topography for different spreading rates, $U.$ From left to right, $U = 75, 50, 25, 17, 10$ and km/Ma, thus illustrating the predicted increase in rise-sink ratio with decreasing spreading rate. Changing rise-sink ratio by changing $U$ is indistinguishable from changing rise-sink ratio by changing $\eta_{M}, F_{M},$ and $H_{400}$ as illustrated by comparing with all cases with $U = 2.5 \text{ km/Ma}$ shown in Figure 7. Symbols denote differences in $H_{400}$ as in Figure 7. Curve is full scaling law, equations (6) and (7) for $H_{400} = 1.3 \text{ km}.$ Case with $F_{M} = 1$ is on far right of plot.
shallowest axial high topography \( h_M \) as the height of the calculation with the largest rise-sink ratio \( (F_M/F_T)(\tau_T/\tau_M) \) for \( F_M < 1 \). All cases with \( F_M = 1 \) predict \( h > h_M \) due to the effects of flexure to produce a total relief that exceeds the isostatic topography (Figure 5 and Buck [2001]). Figure 8 shows results of all cases with imposed thermal structure plotted versus rise-sink ratio. They closely follow the theoretical curve for \( H_{400} = 1.3 \) km; some of the scatter is due to differences in maximum and minimum height, \( h_M \) and \( w_p \) respectively.

[31] The form of (7) is mostly intuitive. First, the proportionality between \( \tau_M \) and \( h_M \) in equation (4) accounts for \( h_M \) being in the denominator of (7). Second, the inverse dependence of rise-sink ratio on \( H_{400} \) probably reflects the fact that increasing \( H_{400} \) decreases the vertical pressure gradient that drives upwelling by both increasing the thickness of the lithosphere channel through which the low-viscosity material rises and by decreasing the non-lithostatic pressure below this channel due to partial melt in the crust. Decreasing melt pressure with lithosphere thickness also reduces axial topography for \( (F_M/F_T)(\tau_T/\tau_M) = \infty \) (i.e., \( F_M = 1 \)) as evident in Figures 7 and 8. Last, the proportionality between the characteristic sinking rate \( 1/\tau_T \) and half spreading, \( U \) can be predicted because, in our models, extension-driven faulting is the main process that depresses the ridge axis [see also Qin and Buck, 2005]. This dependence is revealed by comparing the results of cases with different \( U \). Figure 8 shows that when the rise sink-ratio is varied by changing \( U \) between 10 and 75 km/Ma, the heights \( h \) lie along the same trend as the results presented in Figure 7a, also following the exponential curve given by (6).

4. Discussion

4.1. Geologic Meaning of \( P \)

[32] The tectonomagmatic cycle, \( P \), simulated in our numerical experiments involves times when extension is accommodated entirely magmatically (\( M = 1 \)) and entirely amagmatically (\( M = 0 \)). We therefore envision \( P \) to represent natural time spans less than the \( \sim 10^3 \) ka episodicity inferred from variations in axial topography [Pockalby et al., 1988] or gravity [Tucholke and Lin, 1994], which probably represent longer period fluctuations in magma flux. \( P \) is likely to be more comparable to the age contrast between individual eruptive units, i.e., \( 10^{-3} \)–\( 10^1 \) ka [Curewitz and Karson, 1998; Perfit and Chadwick, 1998; Sinton et al., 2002]. If, in extreme cases, the full tectonomagmatic time period is large enough to control the spacing of faults (see Figure 4c), as has been suggested for the Juan de Fuca Ridge [Kappel and Ryan, 1986], then it would be \( \sim 10^7 \) ka [Canales et al., 2005].

4.2. Importance of \( F_M \)

[33] The importance of the magmatic time fraction \( F_M \) on faulting and axial height can be examined with observations. From bathymetry profiles across different mid-ocean ridges, we have estimated fault characteristics, axial height, and the kinematic fraction \( M \) of tectonic extension (see Figure 9 caption and Paper 1 and Table 2). If we assume \( F_M \) equals \( M \), we can compare observations with our predictions. First, we see that the observed fault spacing \( \Delta S \) increases with decreasing \( M \) as predicted by equation (1) (Figure 9a). Nearly the full range of observations is encompassed by the theoretical curves with \( x_{AF} \) ranging 0.5–1.7 km.

[34] Figure 9b compares observed axial height (\( h \), computed in the same way as the model topography) versus \( M \) with predictions of equation (6), except with the rise-sink ratio in the exponent described as \( M/(1-M)(\tau_T/\tau_M) \). Indeed the observations show decreasing axial height with decreasing \( M \) consistent with the predicted decrease in \( h \) with \( F_M \). The above consistencies between observations and predictions supports (6) and our assumption that \( F_M \sim M \). Our study therefore predicts that \( F_M \) controls both fault characteristics and axial height and does so for distinct reasons: \( F_M \) controls faulting through its relation with the kinematics of extension (i.e., \( M \)) [Buck et al., 2005; Paper 1], and influences axial height through its control on the relative time for building versus reducing topography \( F_M(1-F_M) (=F_M/F_T) \).

4.3. Importance and Geologic Meaning of \( \tau_T/\tau_M \)

[35] The positive relationship between \( F_M \) and \( h \) that is predicted by equation (6) is also exhibited by the data, however, the full range of axial heights can only be explained if \( \tau_T/\tau_M \) changes from relatively high values (\( \sim 0.6 \)) for the tallest axial highs at the Juan de Fuca and Galapagos spreading centers to lower values (\( \sim 0.01 \)) for the deepest valleys along the Mid-Atlantic Ridge (Figure 9b). The inverse relationship between \( \tau_T \) and spreading rate is opposite the global tendency for axial valleys (low \( \tau_T/\tau_M \)) to occur at slow spreading.
rates and highs (high $\tau_f/\tau_M$) at fast rates. The most logical explanation is therefore that there is a corresponding global tendency for $\tau_M$ to decrease more rapidly than $\tau_f$ with spreading rate. Also, the relatively good correlation between $h$ with $M$ seen in data for $M/C_{21}/C_{24}$ is much steeper than the trend predicted by (6) for any single value of $\tau_f/\tau_M$. We can thus interpret this discrepancy as indicating that $\tau_f/\tau_M$ and $F_M$ often change together in nature.

[36] One possible explanation for this correlation has to do with the predicted influence of lithosphere thickness on $\tau_f/\tau_M$ in equation (7). Figure 9c shows estimates of axial topography at various ridges versus the depth to the 400°C isotherm, which is based the depths of earthquakes (MAR other than Lucky Strike) and seismically imaged magma lenses. The observations show a dominant trend of decreasing axial height with increasing $H_{400}$, which is well explained by equations (6) and (7). A simple explanation for $\tau_f/\tau_M$ varying naturally with $F_M$ (and $M$) is that, if $F_M$ is closely related to magma flux, then changes in magma flux and $F_M$ lead to changes in lithosphere thickness as...
predicted by models of crustal accretion and cooling by Phipps Morgan and Chen [1993a, 1993b].

[17] Other factors that could contribute to the above relation are more directly related to magma transport and primarily influence \(1/\tau_M\). For example, topography can grow internally by crustal deformation associated with repeated intrusions of magma or externally by eruptions themselves. Longer magmatic periods (i.e., larger \(F_M\)) could be associated with more frequent events and higher \(1/\tau_M\). Evidence for a positive correlation between eruption frequency and spreading rate (and thus \(1/\tau_M\)) could be found in the

\[ \text{Table 2. Ridge-Perpendicular Bathymetry Profiles Used in Our Analysis Cross the Ridges at the Shown Locations} \]

<table>
<thead>
<tr>
<th>On-Axis Location of Ridge-Perpendicular Transect (Lat. N, Lon. E)</th>
<th>Nontectonic Strain ( M )</th>
<th>Mean Fault Spacing ( \Delta S ) (km)</th>
<th>Long-Wavelength Axial Height ( h ) (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Mid-Atlantic Ridge (MAR)</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>28.95, -43.22</td>
<td>0.73 ± 0.05</td>
<td>5.8 ± 3.6</td>
<td>-0.82 ± 0.30</td>
</tr>
<tr>
<td>29.00, -43.21</td>
<td>0.69 ± 0.06</td>
<td>5.3 ± 2.6</td>
<td>-1.19 ± 0.42</td>
</tr>
<tr>
<td>29.05, -43.20</td>
<td>0.69 ± 0.06</td>
<td>3.3 ± 1.6</td>
<td>-1.15 ± 0.43</td>
</tr>
<tr>
<td>29.11, -43.18</td>
<td>0.71 ± 0.05</td>
<td>3.0 ± 1.8</td>
<td>-1.27 ± 0.15</td>
</tr>
<tr>
<td>29.15, -43.17</td>
<td>0.70 ± 0.06</td>
<td>2.9 ± 1.1</td>
<td>-0.77 ± 0.19</td>
</tr>
<tr>
<td>29.19, -43.15</td>
<td>0.73 ± 0.05</td>
<td>3.2 ± 1.7</td>
<td>-0.76 ± 0.31</td>
</tr>
<tr>
<td>26.72, -44.55</td>
<td>0.58 ± 0.08</td>
<td>7.6 ± 6.3</td>
<td>-0.78 ± 0.25</td>
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<tr>
<td>26.66, -44.57</td>
<td>0.56 ± 0.08</td>
<td>4.9 ± 5.9</td>
<td>-0.48 ± 0.23</td>
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<td>23.37, -44.91</td>
<td>0.45 ± 0.11</td>
<td>8.5 ± 2.5</td>
<td>-0.99 ± 0.30</td>
</tr>
<tr>
<td>23.45, -44.91</td>
<td>0.50 ± 0.10</td>
<td>8.5 ± 4.8</td>
<td>-1.41 ± 0.82</td>
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<tr>
<td><strong>Southeast Indian Ridge (SEIR)</strong></td>
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<tr>
<td>-47.38, 100.75</td>
<td>0.83 ± 0.03</td>
<td>2.6 ± 1.7</td>
<td>0.52 ± 0.04</td>
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<tr>
<td>-47.45, 100.95</td>
<td>0.80 ± 0.04</td>
<td>2.6 ± 1.3</td>
<td>0.45 ± 0.05</td>
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<tr>
<td>-47.73, 101.75</td>
<td>0.79 ± 0.04</td>
<td>2.7 ± 1.0</td>
<td>0.07 ± 0.05</td>
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<td>3.0 ± 2.1</td>
<td>0.10 ± 0.05</td>
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<tr>
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<td>3.3 ± 2.7</td>
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<td><strong>Galapagos Spreading Center (GSC)</strong></td>
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<tr>
<td>2.14, -96.99</td>
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<td>3.4 ± 0.6</td>
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<td>0.84 ± 0.03</td>
<td>3.1 ± 2.5</td>
<td>0.31 ± 0.11</td>
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<td><strong>Eastern Lau Spreading Center</strong></td>
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</tr>
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<td>-21.61, 183.58</td>
<td>0.82 ± 0.03</td>
<td>2.5 ± 0.2</td>
<td>0.40 ± 0.09</td>
</tr>
<tr>
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<td>0.77 ± 0.05</td>
<td>1.0 ± 0.7</td>
<td>0.01 ± 0.04</td>
</tr>
<tr>
<td>-20.36, 183.86</td>
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<td>2.6 ± 1.3</td>
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</tr>
<tr>
<td>-19.85, 0183.95</td>
<td>0.79 ± 0.04</td>
<td>2.8 ± 1.1</td>
<td>-0.15 ± 0.04</td>
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<tr>
<td><strong>Juan de Fuca</strong></td>
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<tr>
<td>44.70, -130.34</td>
<td>0.82 ± 0.04</td>
<td>2.7 ± 1.0</td>
<td>0.42 ± 0.05</td>
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<td>44.82, -130.29</td>
<td>0.84 ± 0.03</td>
<td>3.0 ± 1.3</td>
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<td>45.30, -130.08</td>
<td>0.83 ± 0.03</td>
<td>2.6 ± 1.2</td>
<td>-0.03 ± 0.13</td>
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<tr>
<td>45.39, -130.03</td>
<td>0.82 ± 0.04</td>
<td>2.1 ± 1.1</td>
<td>-0.00 ± 0.05</td>
</tr>
</tbody>
</table>

\(^a\) Fault tops and bottoms were picked based on along-axis continuity of bathymetric lineaments (examples given in Figure 1), assuming all faults are inward facing. \( M \) is estimated from the best fitting slope of cumulative fault heave versus distance on both sides of the axis; errors estimated to be ±20% the slope. Faults spacing is the mean distance between adjacent fault bases with errors being one standard deviation. Values of \( \Delta S \) are consistent with characteristic widths of abyssal hills near the flanks of the MAR and SEIR as measured by Goff et al. [1997]. For the MAR near 29°N, our averages of \( \Delta S \) and tectonic strain (1-M) are 2–3 times larger than those of Escartin et al. [1999]. Escartin et al.’s [1999] high-resolution side-scan sonar data enabled them to identify more, smaller faults than we did, which led to their smaller spacing estimates. They also determined heave based only exposed (high backscatter) faults surfaces, whereas we have assumed the full length of the steepest slopes are part of the fault surface, whether overlain (by sediment, talus, lava) or exposed. We therefore anticipate actual fault heaves being between our and Escartin et al.’s [1999] estimates. Long-wavelength topography is computed the same way as we did for model predictions (see section 2.4); errors in \( h \) are the difference in heights of the faults on either side of the axis that define \( h \). See also Paper 1 for more details.
literature [Curewitz and Karson, 1998; Perfit and Chadwick, 1998; Sinton et al., 2002] and includes findings of sparse magmatic rocks in areas with deep axial valleys [e.g., Cannat, 1993, 1996; Blackman et al., 2002; Dick and Schouten, 2003; Michael et al., 2003]. The spatial pervasiveness of eruptions may also influence $1/\tau_M$ as suggested by observations of expansive sheet flows at ridges with topographic highs (high $1/\tau_M$) versus more confined pillow flows and point-source eruptions at axial-valley-dominated ridges (low $1/\tau_M$) [Smith and Cann, 1990; Lawson et al., 1996; Perfit and Chadwick, 1998; Sinton et al., 2002; Behn et al., 2004]. Ultimately, the behavior of $\tau_T/\tau_M$ in conditions of different $F_M$ thermal structure, and magma supply is governed by the physics of melt storage, intrusions, and eruptions.

[38] As a further illustration of the distinct effects of $\tau_T/\tau_M$ and $F_M$ in controlling axial height, we show a set of calculations designed to be more realistic in that fault spacing is more similar to those of Galapagos Spreading Center (GSC) and the South East Indian Ridge (SEIR) than our prior calculations (by imposing a slightly larger minimum cohesion of 11 MPa) (Figure 10). Thermal structure is computed, $U = 25$ km/Ma, and $H = 1.3$ km. The first case is for $F_M = 1$, and the other cases have $F_M = 0.9$ with $\tau_T/\tau_M$ changing between 0.45 and $10^{11}$ by varying $\eta_M$ between $1.25 \times 10^{18}$ Pa s and $1 \times 10^{27}$ Pa s. First, even though fault spacing and heave are much smaller than in the previous calculations, the long-wavelength topography closely matches those shown in Figure 7b, with the same parameters. This result illustrates the robustness of our method of separating long- and short-wavelength topography. Second, changing $\tau_T/\tau_M$ alone yields the full range of axial morphologies observed along the GSC and SEIR. Last, Figure 10 further illustrates that the rise-sink ratio ($F_M/F_T$) ($\tau_M/\tau_T$) is the primary quantity that controls the transition in ridge-axis morphology from axial topographic highs, to intermediate morphology lacking a high or valley, to deep median valleys.

4.4. Variations in Axial Morphology That Deviate From Global Trends

[39] A key prediction we make is that magma supply and axial thermal structure correlate with or control topography to the extent that they correlate with or control the rise-sink ratio ($F_M/F_T$) ($\tau_M/\tau_T$). At odds with the mechanical influence proposed by Chen and colleagues [e.g., Phipps Morgan and Chen, 1993a, 1993b; Chen and Morgan, 1990a, 1990b], Figure 10 as well as previous studies [Buck et al., 2005; Qin and Buck, 2005; Paper 1] show that rift valley-like morphology can occur independent of the amount of mechanical coupling between the crustal lithosphere and mantle peridotite and often beneath moderately thin lithosphere. Figure 9c shows good evidence for lithosphere thickness being a prominent factor but it also illustrates the importance of
other factors that influence the rise-sink ratio. For example, explaining the axial topography of the MAR near 29°N with $F_M \sim 0.7$ (Figure 9b, 9c) requires a factor of two larger $\eta_M$ than needed to explain the topographic trend of the GSC (green, Figure 9c).

[40] The Lucky Strike area, where Singh et al. [2006] image a magma lens beneath an axial valley, is a clear outlier to overall trend of the other observations. One explanation is a higher $\eta_M$, which again is likely to reflect distinct physical processes of magma storage, intrusions, and eruptions. That is, magma supply is sufficient to sustain a hot crust but $\eta_M$ is too high, and $(F_M/F_T)(\tau_T/\tau_M)$ is too low to fill in the axial valley at present day. Another explanation is that the Mid-Atlantic Ridge is an example where along-axis variations in structure are large [Lin and Phipps Morgan, 1992; Escartin and Lin, 1995; Escartin et al., 1999] and therefore where the morphology at any given location can be influenced heavily by structure or processes at other locations along-axis. This possibility would suggest that the effective values of $H_{400}$, $\eta_M$, or both are greater than the local values above the Lucky Strike magma lens. Ultimately, full 3-D calculations are need to more completely understanding how local topography responds to along-axis variations in the rise-sink ratio.

[41] The Juan de Fuca Ridge is another example of where the axial morphology does not follow a simple relationship with lithosphere thickness. Carbotte et al. [2006] emphasized that the height of a small but prominent axial graben seen on the ridge does not correlate positively with axial magma lens depth as is usually expected. Instead, in three to four of the five segments considered, the magma lens tends to shoal as graben height increases. Following Chadwick and Embley [1998], Carbotte et al. [2006] proposed that these axial grabens are driven not by seafloor spreading but instead by tension occurring above near-vertically intruding dikes. A seafloor spreading origin, however, is still possible if $(F_M/F_T)(\tau_T/\tau_M)$, and thus surface faulting, are sufficiently decoupled in time from lithosphere thickness (i.e., magma lens depth). Different spatial and timescales of magma storage, transport, and eruptions, (i.e., $\eta_M$) can cause such decoupling. Another point to consider is that if the duration of the tectonomagmatic cycle $P$ is long enough to control fault spacing [Kappel and Ryan, 1986; Canales et al., 2005] (i.e., $P$ is longer than age contrast of the mechanically preferred fault spacing, Figure 4), then the deepest graben would correspond to both maximum (end of tectonic period) and minimum (end of magmatic period prior to new faults forming) lithosphere thicknesses.

[42] Last, the Lau Spreading Center displays a clear exception to the overall tendency for a positive correlation between axial height and spreading rate [Martinez et al., 2006]. Here, axial topography decreases from a prominent high in the south (Valu Fa Ridge) to a faulted axial valley in the northern section of the Eastern Lau Spreading Center (ELSC) in the face of both a doubling in spreading rate (and magma supply) as well as a northward shoaling of the axial magma lens between Valu Fa and the central ELSC [Jacobs et al., 2007]. This is clearly opposite to global trends as emphasized by Martinez et al. [2006]. However, this trend is, in fact, consistent with our prediction that $\tau_T/\tau_M \sim 1/U$. The ELSC could be unusual from most other ridges in that the $1/U$-dependence is accentuated by the large change in spreading rate (39–97 km/Ma) over a relatively short distance, thus being the dominant factor in the northward reduction of the rise-sink ratio.

5. Conclusions

[43] Our numerical models simulate the time-dependent behavior of alternating magmatic and tectonic cycles at mid-ocean ridges to investigate the origin of different morphologies seen throughout the global ridge system. We find that if the duration of the full tectonomagmatic cycle $P$ is comparable to or greater than the seafloor age contrast represented by the mechanically preferred fault spacing, faults form and grow only during subsequent tectonic phases and thus fault spacing and size increase linearly with $P$. When $P$ is less than the age contrast of the preferred spacing, as is the case for the remaining results, the overall faulting pattern and long-wavelength morphology are independent of $P$. The mechanically preferred spacing is controlled by the net force required to keep an active fault slipping as it moves off-axis and the force needed to create a new fault by breaking the lithosphere near the axis.

[44] One key parameter that influences axial morphology is the time fraction spent in the magmatic phase $F_M$. Observations support our predictions that fault spacing and total heave are sensitive to $F_M$, both increasing with decreasing $F_M$ in much the same way as simple theory predicts spacing and heave to change with kinematic fraction $M$ of
migmatitic extension [Buck et al., 2005; Paper 1]. The key parameter that controls the global transition between axial topographic highs and median valleys is the “rise-sink ratio” \( \frac{F_M}{F_T} (\tau_\ell / \tau_M) \). A positive relation between axial height and rise-sink ratio reveals that axial height grows with larger times spent in a magmatic phase \( F_M \) relative to a tectonic phase \( F_T = (1 - F_M) \), and by a larger topographic rise rate during magmatic phases compared to the sinking rate during tectonic phases \( \tau_\ell / \tau_M \). Models predict deep axial valleys when \( \frac{F_M}{F_T} (\tau_\ell / \tau_M) < 1 \), tall axial highs when \( \frac{F_M}{F_T} (\tau_\ell / \tau_M) \gg 1 \), and faulted transitional morphology lacking a high or valley at \( \frac{F_M}{F_T} (\tau_\ell / \tau_M) \sim 1 \).

[45] A global inverse correlation between long-wavelength axial height and lithosphere thickness is explained by our new scaling laws that predict the rise-sink ratio \( \frac{F_M}{F_T} (\tau_\ell / \tau_M) \) to decrease with lithosphere thickness. Variations along the Galápagos Spreading Center are particularly well explained by this relation. Changes in axial morphology along the Lau Spreading Center, however, show a locally opposite correlation with lithosphere thickness (within the region that the magma lens was imaged) but are consistent with our predictions that \( \frac{F_M}{F_T} (\tau_\ell / \tau_M) \) is inversely proportional to spreading rate. Other processes more directly related to the frequency or spatial pervasiveness of magmatism also influence the rise sink ratio as evident at ridges such as the Mid-Atlantic and Juan de Fuca ridges. Understanding such processes requires a fuller understanding of the physics of magma storage and transport within or near the crust at mid-ocean ridges.

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References


