Causes for axial high topography at mid-ocean ridges and the role of crustal thermal structure

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Abstract. Mid-ocean ridge topography is modeled as the flexural response to loads using a thin plate approximation and setting thermal structure of the lithosphere to allow, but not require, a region of rapid cooling near the axis. Loads on the lithosphere arise from the presence of lowdensity melt, densification due to cooling with distance from the ridge axis, and thermal contraction stresses. We find two end-member classes of temperature and melt structure that can produce axial high topography and gravity observed at the East Pacific Rise (EPR). One class is very similar to previous models, requiring a narrow column of melt extending to at least 30 km depth within the mantle and lithosphere which cools and thickens very gradually with distance from the ridge axis. The other is a new class, predicting lithosphere which cools rapidly within a few kilometers of the axis and then slowly farther from the axis, with melt which is contained primarily within the crust. The latter solution is consistent with tomography and compliance studies at the EPR which predict rapid crustal cooling within a few kilometers of the axis that is attributed to hydrothermal circulation. This solution also allows the melt region to be coupled to crustal thermal structure and requires no melt anomaly within the mantle. Model fits predict 0-30% melt in the lower crust, depending on how temperatures are distributed within the lithosphere and the degree to which thermal contraction stresses are assumed to contribute to topography. The model generally predicts a wider axial high for lithosphere which is thin over a wider region near the axis. This is consistent with previous correlations between large cross-sectional area of the high and indicators of higher melt presence or a warmer crustal thermal regime. For a slightly slower rate of lithospheric cooling at distances more than \sim 5 km from the axis the model predicts a trough at the base of the axial high. Such troughs have been previously observed at the base of the high on the western flank of the southern EPR, where subsidence rates are anomalously low. Finally, thick axial lithosphere reduces the amplitude of the high, making it sometimes difficult to distinguish from long-wavelength subsidence. This morphology is comparable to that of some intermediate spreading ridges, where topography is relatively flat, suggesting a transition from fast to intermediate style morphology.

1. Introduction

For a distance spanning over 4000 km of the fast spreading East Pacific Rise (EPR) and parts of other mid-ocean ridges, an elongate topographic high 3-20 km wide and rising 200-400 m above surrounding topography marks the ridge axis (Plate 1). The physical phenomena that produce this topography are still debated. The high has long been modeled as the result of a narrow region of low-density mantle material which buoyantly pushes up lithosphere, originally proposed by Madsen et al. [1984]. Wang and Cochran [1993] and Magde et al. [1995] determined that such a region must extend to depths well within the mantle, between depths of 20-30 km and 50-70 km respectively, as gravity data do not exhibit a sizable corresponding low which would be present for a shallower region (once topographic relief has been accounted for). However, analyses of the viscosity and density structure required for such an anomaly [Eberle et al., 1998] showed that the presence of such a narrow region would require extremely large

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Paper number 2000JB000079. 0148-0227/01/2000JB000079\$09.00 contrasts in viscosity. Furthermore, recent seismic tomography results from the Mantle Electromagnetic and Tomography (MELT) Experiment [*Forsyth et al.*, 1998; *Hung et al.*, 2000] do not show evidence of any low-velocity zone such as would be expected in a narrow region of high melt fraction.

Previous models which consider the flexural response of the lithosphere [Kuo et al., 1986; Wang and Cochran, 1993] require elastic lithosphere which thickens gradually with distance from the ridge axis (for example, the best fits by Wang and Cochran [1993] require lithosphere not more than 1 km thick at a distance of 10 km from the ridge axis). In these models the mechanical structure of the lithosphere is linked to the thermal structure of the crust when fitting bathymetry and gravity observations, so that the crust must cool slowly with distance from the ridge axis. This type of thermal structure may be inconsistent with the rapid cooling inferred from seismic velocity structures determined with tomography experiments [Vera et al., 1990; Toomey et al., 1994], compliance studies [Crawford et al., 1999], and studies of the response of near-axial lithosphere to topographic loads [e.g., Cochran, 1979; McNutt, 1979]. Models of axial thermal structure, such as those by Phipps Morgan and Chen [1993] and Henstock et al. [1993], also predict rapid cooling away from an axial magma chamber (AMC), due to cooling effects of hydrothermal circulation.

Eberle and Forsyth [1998a] proposed a mechanism for axial high topography which does not require low-density mantle or

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Plate 1. East Pacific Rise topography, $15^{\circ}-17^{\circ}$ S, looking southward. Data are from *Scheirer et al.* [1996] (anonymous FTP). Note some segments (such as near 17° S) exhibit a wide axial high, whereas others (such as near 15° S) exhibit a narrow axial high. Vertical exaggeration is 5X.

crustal material below the axis. They suggest that large stresses maintained in a high-viscosity lower crust, coupled with low stresses maintained in upper crust near the ridge axis (where extensional stress is relieved by dike injection), create a plate bending moment, which in turn creates an axial high. They also assume lithosphere which thickens slowly with distance from the ridge axis, combined with constant viscosity aesthenosphere beneath it (their best fits require elastic lithosphere which is 2 km thick at a distance of 10 km from the ridge axis). This type of mechanical structure also suggests thermal lithosphere coincident with the mechanical lithosphere and thus crust which cools slowly with distance from the ridge axis.

We propose that alternate thermal and magmatic scenarios can produce axial high topography. To examine possible scenarios, we use a simple one-dimensional flexural model to predict ridge profile topography and consider a greater range of lithosphere and melt structures than previous studies. We parameterize twodimensional temperature and melt structures and examine the range of parameters which can fit both bathymetry and gravity data. Once a set of fit parameters is found, we examine the effects of changing them slightly. The resulting changes in topography are compared to observed intersegment and intrasegment variations in axial high morphology and other geologic indicators of thermal and magmatic structure.

2. Model Formulation

2.1. Flexure of a Moving Plate

Observations such as the presence of moats around seamounts and hot spot islands and the bending of plates as they approach subduction zones [Cochran, 1979; McNutt, 1979] strongly suggest that the lithosphere responds flexurally to applied loads. This deflection has frequently been modeled using a thin plate approximation. We take this approach, modeling profiles of topography of the ocean floor due to the densification of buoyant material below the lithosphere or thermal bending stresses. The fact that the ocean lithosphere is created at a ridge axis and accreted onto the plate and stays in continuous motion away from this axis plays a significant role in the model formulation.

Kuo et al. [1986] determined that a plate in motion will respond differently than one which is still, i.e., static, if the plate rigidity varies with distance. We develop an approach to the same problem that differs only in the way material is accreted at the axis. To model the deflection of a moving plate, *Kuo et al.* [1986] assumed that the plate is initially stress-free at the axis. They then formulated an expression for the change in stress with each step of accretion, assuming that elastic equilibrium is maintained at each step. This expression was integrated to find a relationship between moment and deflection, yielding a fourth-order differential equation for deflection. Additional boundary conditions were zero curvature and shear stress at the axis $(d^3w/dx^3 = d^2w/dx^2 = 0 \text{ at } x = 0$, where w is deflection). It is important to note that no restrictions were placed on the deflection at the axis.

We instead assume an initial condition for the deflection at the axis and then consider changes in deflection, in response to the changes in loads, as the plate accretes and moves away from the axis. The initial condition that we assume is local isostasy at the axis, following *Buck* [2001]. This is suggested by an expectation that eruptions at the East Pacific Rise are likely to occur often over scales of several thousands of years. With frequent diking, the cut lithosphere is likely to frequently settle to a position of local isostasy.

For a moving, accreting plate acted on by a fixed distribution of loads q(x), we determine the change in deflection of the plate as it moves a distance Δx (Figure 1) by assuming elastic equilibrium is maintained at each step. Thus

$$\frac{d^2 \Delta M_B}{dx^2} + \Delta \rho g \Delta w = \Delta q; \qquad (1)$$

$$\Delta M_B = D \frac{d^2 \Delta w}{dx^2}, \qquad (2)$$

where $\Delta q(x) = q(x + \Delta x) - q(x)$, $\Delta w(x)$ is the change in plate deflection, $\Delta M_B(x)$ is the change in bending moment, D(x) is the plate rigidity, and $\Delta \rho$ is the density contrast between the lithosphere and water. The initial condition of local isostasy at the axis implies $\Delta \rho g w(0) = q(0)$. We can then shift w(0) and q(0) to an arbitrary reference level of zero. The total deflection is found by summing the changes in deflection at each accretionary step and adding these to the initial deflection:

$$w(x) = \int_0^x \frac{dw}{dx'} dx'.$$
 (3)

Level of local isostasy





Figure 1. Illustration of how incremental loads can contribute to axial high topography. The plate is assumed to be broken at the axis. (top) At the axis, melt rises to a height constant over time. (middle) As new material moves away from the ridge axis, it densifies because of cooling and solidification, producing a downward load. Thermal stresses arise because material near the seafloor usually remains near water temperatures but material at depth continually cools. The greater change in temperature at depth causes the bottom of the lithosphere to contract more than the top, causing the plate to bend down. (bottom) At the same time that these loads act, the lithosphere spreads away from the axis, and new material is accreted at the same height as before. The result is a high at the axis.

Changes in deflection are determined from (1) to (3) via a fourthorder finite difference scheme.

To assure local isostasy at the axis, we assume boundary conditions $d^2\Delta w/dx^2 = d^3\Delta w/dx^3 = 0$ at x = 0. The fact that these imply local isostasy at the axis is most easily illustrated by integrating (1) and (2) into the form of *Kuo et al.* [1986, equation (9)]:

$$\frac{d(Dd^3w/dx^3)}{dx} + \Delta\rho gw = q. \tag{4}$$

In the integrated form the boundary conditions become $d^3w/dx^3 = d^4w/dx^4 = 0$ at x = 0, yielding $\Delta \rho gw(0) = q(0)$ when applied to (4). The key difference between these boundary conditions and those of *Kuo et al.* [1986] is that they allow curvature to develop at the axis, since no restrictions are placed on d^2w/dx^2 . The development of

curvature at the axis can significantly affect the resulting deflection [*Buck*, 2001].

The difference in boundary conditions may or may not have a strong effect on the resulting deflections, depending on the rigidity at the axis. The models are equivalent if zero rigidity is assumed at the axis, since this implies local isostasy there for both formulations. However, the formulation used in this study predicts the same depth at the axis for different lithosphere rigidities because of the assumption of local isostasy there, whereas the formulation by *Kuo et al.* [1986] predicts that the depth of the axis depends on the strength of the lithosphere at the axis and the magnitude of the applied load. This arises because their formulation places constraints on the curvature and shear stress at the axis, but none on deflection; the deflection must be set accordingly so that curvature and shear stress are zero. The formulation that we use fixes the depth of the axis. Curvature and shear stress may

Table 1. Values of Material Properties Used

| Property | Symbol | Value |
|------------------------------------|----------------|--|
| Young's modulus | Ε | 9×10^{10} Pa |
| Poisson's ratio | ν | 0.25 |
| Density of water | ρ _w | 1000 kg m ⁻³ |
| Density of crust (bulk) | ρ_c | 3000 kg m^{-3} |
| Density of melt | ρ_m | 2700 kg m^{-3} |
| Density of mantle | | 3300 kg m^{-3} |
| Coefficient of linear expansion | α | $1 \times 10^{-5} ^{\circ}\mathrm{C}^{-1}$ |
| Maximum aesthenosphere temperature | T_m | 1200°C |

develop accordingly. Both models achieve a plate which is "broken," or discontinuous at the axis, yet the axial depth still depends on plate thickness for the *Kuo et al.* [1986] formulation. This formulation may be more appropriate for regions with less frequent eruptions, such as at intermediate and slow spreading ridges.

2.2. Loads on the Lithosphere

We consider three types of loads which are likely to affect lithospheric deflection. Two have been considered for most previous models of axial high topography: that of the solidification of melt below the ridge axis (q_m) and the thermal densification of the lithosphere with distance from the ridge axis (q_c) (the latter has been used to model subsidence of the ocean basins over scales of thousands of kilometers [*Parsons and Schlater*, 1977]). The third is bending due to thermal contraction stresses (q_t) . There are some difficulties with modeling thermal stresses using a thin plate approximation, discussed below, so we will consider cases both with and without this load. The final load distribution used in (2) is

$$q(x) = q_c(x) + q_m(x) + q_l(x).$$
 (5)

2.2.1. Density changes. As the lithosphere cools and becomes denser, its increasing weight will contribute to its deflection. The associated load is

$$q_c(x) = \int_0^{h_m(x)} g \rho \alpha_{\nu} [T(x,z) - T_m] dz, \qquad (6)$$

where α_{ν} is the volumetric coefficient of thermal expansion and h_m is the depth at which material reaches T_m , an estimate for melting temperature, in this study, 1200°C (Table 1). We assume that below this depth, temperatures do not vary considerably from melting temperatures and will not contribute significantly to the load so that T_m represents the maximum aesthenosphere temperature. We also consider the effects of melt solidification and densifi-

We also consider the effects of melt solidification and densification within the lower crust and mantle. In this case,

$$q_m(x) = \phi \Delta \rho_m g H(x), \tag{7}$$

where ϕ is the volume fraction of melt, with a density contrast $\Delta \rho_m$ relative to surrounding rock at 1200°C, distributed over a column of height H(x). The melt fraction is, in fact, likely to vary both vertically and horizontally within the crust, but this variability is poorly constrained. We simplify the situation by considering ϕ as an average melt fraction below a region defined by $h_m(x)$, the 1200°C isotherm.

These loads are comparable to that proposed by *Buck* [2001], except that Buck simplified the situation by applying the loads due to solidification and cooling at one point, at the end of the plate (this simplification assumes that most cooling and solid-ification occur over a short spatial scale). In this study, we trace the solidification of melt with distance from the axis according to thermal structure (equation (7)), with cooling accounted for in (6).

2.2.2. Thermal contraction stresses. Thermal stresses arise when different depths of a material are cooled by different amounts. If a material is cooled at a bottom surface, the cooled area will contract, causing the material to bend downward if it is allowed to deflect freely or to crack if it is held fixed. Thermal stresses have been used to explain deflections at transform faults near ridge-transform intersections of up to a kilometer, due to sizable differences in temperatures and cooling rates across a transform fault [e.g., Parmentier and Haxby, 1986]. They may also contribute to deflections near the ridge axis: Lithosphere at depth will continue to cool as it moves away from the axis, whereas lithosphere near the seafloor has already cooled to nearwater temperatures, and experiences relatively little subsequent change in temperature (Figure 1). This difference will cause the lower part of the lithosphere to contract more than the top, creating a moment, so that the plate will bend downward while it is moving away from the ridge axis. Downward deflection with motion away from the ridge axis will leave a high remaining at the axis.

Deflection created by thermal stresses can be estimated as follows. As an elastic material undergoes a change in temperature ΔT , the corresponding change in stress is, for a plate which is allowed to deflect freely,

$$\Delta \sigma_T(x,z) = \frac{E}{1-\nu^2} \alpha \Delta T(x,z) - \int_0^h \frac{E}{1-\nu^2} \alpha \Delta T(x,z') dz', \quad (8)$$

where α is the linear coefficient of thermal expansion, h is the thickness of the lithosphere, and z is the depth below the seafloor. The first term of (8) arises from strain associated with thermal contraction. The second term is needed if the plate is allowed to deflect freely in the vertical direction at one end, as it forces the integrated stress over the thickness of the plate to be zero. For accreting lithosphere, temperatures change as lithosphere cools and the plate moves a distance Δx , so

$$\Delta T(x,z) = T(x,z) - T(x - \Delta x, z) \tag{9}$$

assuming a steady state temperature field fixed in space.

Thermal bending stresses act from within the plate but can be incorporated into the one-dimensional flexure equation via a load term, following *Parmentier and Haxby* [1986]. Thermal bending stresses contribute to the moment term on the left-hand side of (1), so deflections can determined by incorporating this contribution as a load:

$$q_T(x) = \frac{-d^2 M_T(x)}{dx^2},$$
 (10)

where

$$M_T(x) = \int_0^h \sigma_T(x,z) z dz = \int_0^h \left(\int_0^x \Delta \sigma_T(x',z) dz \right) z dx', \quad (11)$$

| Table 2. | Variable | Parameters |
|----------|----------|------------|
|----------|----------|------------|

| Parameter | Symbol | |
|--|--------|--|
| Thermal structure | | |
| Depth to 1200°C isotherm near axis | h_0 | |
| Depth (of 1200°C isotherm) to which distance-squared cooling is used | h_1 | |
| Width over which 6-km crust would reach temperatures below 1200°C, if distance-squared cooling continued below h_1 | W | |
| Coefficient for square root of distance cooling | S | |
| Melt | | |
| Percent of lower crust which is melt | φ | |
| Strength | | |
| Strength reduction factor | ω | |



Figure 2. (left) Model parameters. The region is horizontally divided into three sections, with different types of cooling for each. Within 1 km of the axis, the depth of the 1200°C isotherm is held at a constant value h_0 . Starting at 1 km from the axis, the 1200°C isotherm deepens as distance from axis squared, to a depth h_1 . W is the distance from the axis at which quadratic cooling would reach 6 km depth. Away from the axis, isotherms deepen as the square root of age $(S^*\sqrt{x} + c)$, where c is used to make the isotherms continuous). (right) Associated thermal structure. For profile A, temperatures increase linearly with depth, from 0°C at the top to 1200°C. For profile B a percentage of the plate is held at a constant temperature of 200°C near the surface, simulating hydrothermal cooling, with temperatures increasing linearly with depth below that.

the moment created by thermal stresses. We note that if there were no buoyancy term in (1) (e.g., the plate floated in air, rather than on a viscous mantle), the bending moment would match the thermal moment exactly.

Using a thin plate approximation assumes that thermal contraction in a direction horizontally parallel to the plate is much greater than that vertically perpendicular to the plate. For lithosphere which thickens rapidly over a short distance, this assumption may not hold, and the actual deflection due to thermal stresses will be overestimated by the thin plate approximation. We will thus examine cases with a thermal stress contribution and without, recognizing that reality should lie somewhere between the two.

Some degree of thermal bending stress should naturally develop within the ocean lithosphere, depending on the lithospheric thermal structure, yet the effect on axial topography has not previously been studied in detail. This may be due to the fact that for lithosphere that cools as a function of the square root of age, there would be minimal deflection due to thermal bending stresses [*Parmentier and Haxby*, 1986]. An assumption that the lithosphere cools at a rate other than proportional to the square root of age is necessary for thermal bending stresses to contribute to a high at the axis.

2.3. Parameters Controlling Deflection

2.3.1. Thermal structure. Model deflections will depend strongly on the geometry of lithospheric thickening, which, in turn, depends on the thermal structure of the crust. Thermal and lithosphere structures are only somewhat constrained by currently available data and models, so we consider a group of lithosphere thickness distributions defined by a set of four parameters. The two-dimensional temperature structure is then defined as a function of the plate thickness. We consider two vertical temperature distributions: one where temperatures simply increase linearly with depth from 0°C at the seafloor to 1200°C at depth (assuming crustal melt temperatures are near

1200°C), and a second where a fraction of the upper part of the plate is assumed to cool to 200° C but then increases linearly to 1200°C at depth. The latter scenario is intended to mimic hydrothermal cooling of the crust. These distributions are illustrated in Figure 2.

Depth to the 1200°C isotherm is defined as a function of distance from the axis. The plate is divided horizontally into three sections, with different types of cooling for each (Figure 2). Within 1 km of the axis (where a magma chamber is likely to be present) a constant 1200°C isotherm depth h_0 is assumed. Beginning 1 km from the axis (where hydrothermal circulation might quickly cool the crust), isotherms deepen as the square of distance, up to a depth h_1 for 1200°C. Beyond that distance, isotherms deepen more slowly, as the square root of distance (approximating conductive cooling). Associated parameters for the outer two sections are S, the coefficient of the square root of distance cooling, and W, defined as the distance from the axis where quadratic deepening would place the 1200°C isotherm at 6 km depth. W can be roughly interpreted as the distance from that axis at which the entire crust has cooled to submelt temperatures. These parameters are summarized in Table 2.

2.3.2. Lithosphere strength. The thickness of the effective elastic lithosphere is frequently defined as an isotherm roughly describing the brittle-ductile transition. The temperature range over which this transition occurs, however depends on various conditions: *Hirth et al.* [1998], for example, summarize estimates of the brittle-plastic transition of shallow oceanic lithosphere which can vary up to 200° C, depending on whether there is a dry or wet diabase rheology. Other factors complicate estimates of lithospheric strength, such as plastic weakening and faulting. Determining the precise effects of the latter requires knowledge of the magnitude and distribution of weakening, which is beyond the scope of this study. We combine the effects of weakening and imprecision in estimating a brittle-ductile transition by assuming that the



Figure 3. Contours of RMS misfit to stacked profiles of the East Pacific Rise, $8^{\circ}-8^{\circ}15'S$ (top) for bathymetry plus gravity (scaled by range of each data set before summing), (middle) for topography only, and (bottom) for gravity. The lowest RMS value over a range of trapezoid widths was used at each point. Other parameters used are $h_0 = 1$ km, $h_1 = 6$ km, S = 0 m^{1/2}, and $\omega = 0.33$. The width of at the base of the melt zone was taken at 4 km, since this value achieved the lowest RMS values over all values of W and melt zone depth. The bulk density of the zone simulates between ~ 2 and 5% melt over 70 km, equivalent to 30 and 70% melt over 5 km, for densities shown in Tables 1 and 2. Data are from the RIDGE Multibeam Synthesis and GEODAS Worldwide Geophysics Databases.

effective elastic thickness is a fraction of the depth to the 1200° C isotherm. We have

where E is Young's modulus and ν is Poisson's ratio, h is the effective elastic thickness, and ω is a parameter between 0 and 1 [e.g., see *Turcotte and Schubert*, 1982]. Then, for example, a value of $\omega = 0.5$ suggests that either the lithosphere has weakened to half

$$D = E[\omega h(x)]^3 / [12(1 - \nu^2)], \qquad (12)$$

Previous Models

New Model



Plate 2. Illustration of the two classes of solutions yielding best fits to observations. One class is similar to previous models, requiring a deep, narrow column of melt in the mantle and temperatures which decrease very slowly with distance from the axis. The other requires less melt in the mantle and temperatures which decrease rapidly over a short distance from the axis.

its elastic thickness or that the brittle-ductile transition lies at the 600°C isotherm for the case where temperatures increase linearly with depth, or a combination of the two (the last scenario being the most likely).

3. Results

3.1. Forward Models Fitting Observations

We examine a general range of lithosphere and melt region geometries which will fit observed topography. Since the implied thermal structure will also affect the gravity anomaly via density variations, gravity data are also used to further constrain the range of possible structures. To model a "typical" axial high, we considered stacked profiles of the southern East Pacific Rise at $8^{\circ}-8^{\circ}15'S$ [Cochran et al., 1993]. This region is particularly favorable for model comparisons because along much of its length, there are few offsets, large seamounts within 20 km of the axis, or near-axis features associated with recent ridge jumps or propagation. For this initial study we focus on lithosphere cooling width and the dimensions of the melt region.

We assumed that the melt region has a trapezoidal shape with variable dimensions and allowed its depth, width, and bulk density to vary. The bulk density of the region was scaled with vertical length of the region in order to produce a high of 400 m at the axis; changing the depth of the region will then have no effect on topography for a given bulk density. We compare the dependence of topography and gravity on the dimensions of the melt region to effects of changing the width of lithospheric cooling W. For a closer comparison to previous models, effects of thermal contraction stresses were not included (recall that the model formulation is equivalent to that of Kuo et al. [1986] for lithosphere with zero thickness at the axis). We assume constant values of h_0 , h_1 , S, and ω and temperatures that increase linearly with depth. Synthetic gravity anomalies were calculated using the method of Talwani et al. [1959], using observed bathymetry.

Contours of RMS misfit to gravity and topography for a range of lithosphere cooling widths and melt region depths are shown in Figure 3 (the best fit melt region width and bulk density for

each point were used). These contours reveal two broad parameter-space regions which can fit the data: one where the lithosphere cools over a long distance from the axis (W large), and melt extends to at least 40 km depth, comparable to the results of Kuo et al. [1986] and Wang and Cochran [1993]; and a second where the lithosphere cools over a shorter distance from the axis (W small), and a greater range of melt depths is feasible. These different scenarios are illustrated in Plate 2. The large widths of lithospheric cooling for first case imply conditions similar to local isostasy for a significant distance from the ridge axis so that a large degree of melt is needed to support the high. The second case illustrates conditions where the flexural response of a plate which is relatively thick near the axis contributes to topography. This, in addition to a greater degree of densification due to nearaxis lithospheric cooling, implies much less melt is required. The lithosphere cools over a distance comparable to that over which melt solidifies, and melt may even be entirely contained within the crust. We note that for both scenarios the range of parameters which fit observations is broad, particularly for the depth of melting.

The shape of RMS misfit can be understood by considering gravity and bathymetry RMS misfit separately, shown in Figures 3 (middle) and 3 (bottom). Changing the depth of the melt region will not affect bathymetry, since the bulk density of the melt region changes accordingly. There is a slight increase in bathemetry RMS for lithosphere cooling widths centered near 10 km. Other model runs assuming a different value of lithosphere weakening (ω), not shown, reveal a similar pattern, but with the RMS increase shifted slightly where the worst fit W decreases as ω increases. RMS contours of gravity misfit show a high centered near 18 km. These patterns can be understood by considering the residual gravity anomaly for several crustal thermal structures (the residual anomaly is defined as the observed free-air anomaly less the mantle Bouguer correction for topography and less the contribution of crustal thermal structure). Figure 4 shows the mantle Bouguer anomaly (MBA) and residual anomaly for three values of W. For W < 5 km, nearly all the MBA is accounted for by crustal thermal structure. For W > 5 km the contribution from crustal thermal structure is so great that a dense object would need to be present



Figure 4. (top) Mantle Bouguer anomaly (dashed lines) for the EPR $8^\circ - 8^\circ 15'S$ assuming a 6-km-thick, constant density crust. Residual gravity anomaly (solid lines) (free-air anomaly less topography and thermal contributions) assuming various thermal structures. (bottom) Depth to 1200°C isotherm for thermal structures corresponding to above residual anomaly. *W* as shown, $h_0 = 1$ km, $h_1 = 6$ km, and S = 0 m^{1/2}.

below the axis in order to match observations. For W > 30 km the crust cools much more gradually with distance from the axis, so that not all the observed gravity anomaly is accounted for, and additional low-density material is required below the axis. These differences in the residual anomaly illustrate the importance of gravity as a tool which can constrain thermal structure below the seafloor.

3.2. Model Inversion

The forward models suggest a scenario in which lithosphere and melt structures are coupled and melt may be entirely contained within the crust. We thus assumed a new shape to the melt region, where the boundaries are defined by the 1200°C isotherm and the base of the crust at 6 km depth. We then examined the range of thermal structures which can produce observed topography and gravity by varying all parameters shown in Table 2. To search for a best fit to data, we considered the RMS difference between observed and predicted topography and gravity and minimized the scaled sum of these over a range of model parameters. The minimization, performed using a Nelder-Meade technique [e.g., Press et al., 1986], provided a set of parameters which can fit both types of data, shown in Figures 5a and 5b for the two types of temperature distributions in Figure 2. The contributions to topography from three different sources, thermal bending stresses (load q_t), lower crustal melt (load q_m), and cooling of the lithosphere (load q_c) for the best fit set of parameters and temperature structure assuming hydrothermal cooling are illustrated in Figure 6. We note that the broad range of data fits shown in Figure 3 suggests a fair margin of error for these best fit parameters.

For both temperature distributions the model temperatures decrease rapidly over a narrow range near the axis and then almost negligibly away from the axis. Fits without thermal contraction stresses require 30% melt when temperatures increase linearly with depth but only 12-15% melt when the lithosphere has been cooled by hydrothermal circulation. This reduction in required melt arises because a greater amount of the lithosphere is cooled with the latter temperature distribution, increasing the densification load q_c and allowing the melt load q_m to decrease. Thermal contraction stresses also reduce the amount of melt needed to negligible amounts for this temperature distribution. This reduction occurs because deflections associated with the contraction stresses add over 100 m of relief to topography, the same amount of relief created by the melt load (for cases with linear temperature profiles, thermal contraction stresses added at most 50-60 m relief, but the melt contributed \sim 200 m). Figures 5b and 6 show cases where we have assumed that 75% of the depth to the 1200°C isotherm has cooled to 200°C, which may be a large amount. Other model runs with less of the plate cooled (not shown) predicted 5-20% melt for 50% cooling and 17-26% for 25% cooling.

Thermal stresses predict a much greater effective elastic thickness for both temperature scenarios. For cases with hydrothermal temperature distributions this change is quite substantial, from $\omega = 20-22\%$ for no thermal stresses to $\omega = 70-74\%$. This occurs because the narrow peak near the axis of the thermal stress deflection profile allows the deflections due to other sources to be wider. We note that at distances far from the axis, thermal contraction stresses contribute little to the overall topography, as the square root of distance cooling produces no load contribution [*Parmentier and Haxby*, 1986].



Figure 5a. Best fit of stacked bathymetry and gravity for the EPR $8^{\circ} - 8^{\circ}15'S$, assuming temperatures increase linearly with depth. (top) The 1200°C isotherm and best fit parameters. Note that the east and west sides of the ridge were fit separately. For each flank, left column parameters are for models with thermal stresses, and right column parameters are for models without thermal stresses. (middle) Stacked profiles of bathymetry and best fit model topography. (bottom) Stacked gravity and best fit model gravity. Dashed lines indicate solution with thermal stresses, shaded lines indicate solution without thermal stresses, and solid lines indicate observations. For these model runs, h_0 was kept to a value near 1 km, since most AMCs of the EPR have been imaged at depths of ~1 km [Detrick et al., 1987; Carbotte et al., 1997].

3.3. Behavior With Small Changes in Parameter Values

Previous geologic and geophysical observations have suggested that variations in morphology may be directly linked to the thermal and magmatic structure of the crust. For a model primarily dependent on thermal structure, we can examine the changes in shape of the high with slight changes in the values of some parameters. For simplicity, we consider cases without thermal bending stress effects.

3.3.1. Thickness of axial lithosphere. There is an amplitude reduction of the high as the axial lithosphere thickens (h_0 increases) (Figure 7) because of a combination of two effects. With thicker axial lithosphere, there is a smaller region of solidifying melt to load the lithosphere. There is also less horizontal temperature contrast within the crust and thus a lower magnitude of thermal

stress. For $h_0 > 3$ km the deflection has very long wavelength, comparable to that due to plate cooling alone.

3.3.2. Long-wavelength cooling (subsidence). The effect of varying the subsidence rate S is illustrated in Figure 8. For S small a trough is visible at the base of the high, but with increasing S the trough disappears. The trough occurs naturally, as curvature develops at the axis, and its history is maintained off axis. However, as S increases, so does the load q_c owing to cooling of the lithosphere. If this load is great enough, it will dominate topography and counteract the curvature creating the trough.

3.3.3. Width of lithospheric cooling and thickening, effective plate thickness. Figure 9 shows model deflections for different cooling widths W and strength reduction factors ω .



Figure 5b. Best fit of stacked bathymetry and gravity for the EPR between 8°S and 8°15'S, assuming temperature structure in Figure 3 (middle), where a percentage of the plate is cooled to a constant temperature of 200°C.

Lithosphere that is weaker, or to a lesser degree thickens over a narrower distance from the axis, produces a narrower axial high. As ω decreases, there is a narrowing due to the decrease in flexural wavelength. For small ω , there is a slight narrowing of the high as W decreases, but there is little response to changes in W for larger ω . The narrowing with decreasing W occurs because the maximum change in temperature is closer to the axis and thus also the contribution from lithospheric solidification and cooling (as well as the maximum change in thermal bending moment), so that the greatest deflections are closer to the axis. This effect is less apparent with larger ω , since the increased flexural wavelength creates wider deflections, regardless of how narrow the load on the plate may be.

4. Discussion

4.1. Features of the New Model

The above studies show that a combination of the following characteristics can create an axial high with observed dimensions, which is consistent with gravity data, illustrated in Figure 10: (1) lithosphere which is weak enough to be considered locally supported at the axis, (2) temperatures which are near melting temperatures near the axis (except within the uppermost kilometer), which decrease rapidly within a few kilometers of the axis (lithosphere should have a corresponding structure), (3) partial melt within the lower crust, and (4) some weakening of the elastic lithosphere.

The above conditions will be most often found at fast spreading ridges: A scenario of frequent diking is most likely to be associated with a region of discontinuous, isostatically supported lithosphere. In cases where magmatic activity occurs frequently but the axial lithosphere is thick (say for a deeper axial magma chamber), the amplitude of the high will decrease, suggesting a transition to flat, intermediate-style morphology. At slow spreading, and some intermediate spreading, ridges we expect lithosphere to be stronger and more continuous, so that stretching creates an axial valley. With the given model formulation the high can also be created assuming the presence of a narrow column of melt which extends deep into the mantle, and lithosphere cools very slowly with distance from the axis, similar to



Figure 6. Breakdown of contributions to topography for solutions in Figure 5b for (top) melt solidification contribution, (middle) lithospheric cooling contribution, and (bottom) thermal contraction stress contribution. Dashed line indicates solution with thermal stresses, shaded line indicates solution without thermal stresses, and solid line indicates observed topography.

that proposed by *Madsen et al.* [1984], *Kuo et al.* [1986], and *Wang and Cochran* [1993]. The amount of melt required depends on the type of temperature distribution assumed and the degree to which thermal contraction stresses are assumed to play a role.

4.2. Comparisons to Previous Models

The model formulation presented in this paper is equivalent to that used by *Kuo et al.* [1986] and *Wang and Cochran* [1993] if lithosphere thickness at the axis is zero, yet neither of these studies solved for the thermal structure preferred here. The primary difference between those and our approaches is the way that the lithosphere thickness and crustal thermal structure are parameterized. Both *Kuo et al.* [1986] and *Wang and Cochran* [1993] assumed lithosphere that thickens proportionately to the square root of age, and they associated the base of the lithosphere with an isotherm estimating the brittle-ductile boundary. To simulate a narrow zone of cooling, the square root of age coefficient would have to be very large, creating rapid thickening deep within the mantle, a somewhat extreme scenario. Best fit solutions thus required small square root of age coefficients. However, a greater melt load was then required to support topography, which had to reside deep within the mantle in order not to create a significant gravity anomaly. We note that these two models are based on that initially proposed by *Madsen et al.* [1984] and predict comparable density and melt structures.

Magde et al. [1995] did consider crust which cools significantly over a few kilometers from the ridge axis and found that most of the observed gravity anomaly could be attributed to the associated crustal density structure. However, they did not take into account the density and lithosphere structure when determining topography, either for the contribution of changing densities or the flexural effects of thickening lithosphere (they assume local isostasy for the entire plate). For this reason, they still required a significant melt load deep within the mantle to support the high at the axis.

The *Eberle and Forsyth* [1998a] viscous stress model does not require a region of buoyant material in the crust or mantle to



Figure 7. Model deflections for various values of h_0 with no thermal stresses. Parameters used are W = 5 km, $h_1 = 4$ km, S = 0 m^{1/2}, and $\omega = 0.33$.

produce an axial high and is thus consistent with results from the MELT experiment. However, they also require elastic lithosphere which thickens slowly with distance from the axis. The fact that the crust below the lithosphere has a uniform viscosity of value 10^{19} strongly suggests that the elastic lithosphere boundary is also a thermal boundary. This places it in a range of crustal thermal structures as solutions similar to those of *Madsen et al.* [1984], *Kuo et al.* [1986], and *Wang and Cochran* [1993]. This model also requires a relatively strong lower crust at the axis in order to maintain extensional stresses there. This is significantly different from the model we propose, where melt is present in the lower



Figure 8. Model deflections for various subsidence rates S with no thermal stresses. Parameters used are W = 5 km, $h_0 = 1 \text{ km}$, $h_1 = 4 \text{ km}$, and $\omega = 0.33$.



Figure 9. Model deflections for two values cooling widths W and two values of weakening ω with no thermal stresses. Parameters used are $h_0 = 1$ km, $h_1 = 4$ km, and S = 0 m^{1/2}.

crust near the axis, precluding the development of large stresses there.

4.3. Data Supporting the New Model

Model runs suggest that both the proposed thermal and magmatic structure proposed here and that suggested by Madsen et al. [1984] and others can fit both gravity and bathymetric data. However, seismic [Vera et al., 1990] and compliance data [Crawford et al., 1999] collected at the East Pacific Rise (EPR), as well as thermal models [Phipps Morgan and Chen, 1993; Henstock et al., 1993], suggest that the crust cools (and lithosphere thus thickens) rapidly within a few kilometers from the ridge axis. Furthermore, the presence of a narrow column of melt extending deep within the mantle is inconsistent with results from the MELT experiment. For these reasons, the previous models seem less preferable. We note that although the thermal structure of the model that we propose is quite different from that of the previous models, the effective elastic plate thickness is actually similar in cases where best fit values of ω are <0.5. An allowance for plastic weakening of the lithosphere may thus be an important component of the model. However, crust which cools over a narrow distance from the axis is also critical to the model. The associated density variations contribute significantly to the amplitude of the high and thus reduce the amount of melt needed. These density variations also explain much of the source of the observed mantle Bouguer anomaly [see also Magde et al., 1995].

The amount of melt required in the lower crust is $\sim 30\%$ for a purely linear temperature-depth dependence and 0-15% assuming that the upper part of the lithosphere is cooled to hydrothermal temperatures. Values near 0% require the full effect of thermal contraction stresses. Because of the model's one-dimensional formulation, effects of thermal bending stresses are most likely exaggerated, so that the amount of melt needed is probably somewhat greater than 0%. Recent tomography studies at the EPR near 9°30'N [Dunn et al., 2000] suggest the presence of melt within the lower crust, from amounts of 1% to 38% (the greatest percentages only within a few hundred meters depth from the axial magma chamber), with average estimates roughly ranging from ~5 to 18%. These amounts are consistent with model predictions for temperature profiles which assume some degree



Figure 10. Preferred models. (bottom) Shaded area shows depth to the 1200° C isotherm. Amount of required melt depends on degree to which thermal stresses contribute, only roughly estimated in this study. Dashed lines depict the effective elastic plate thickness, for cases with thermal stresses (deeper) or without (shallower). Fits to (middle) gravity and (top) bathymetry, assuming no thermal stress contribution. Solid line indicates stacked bathymetry and gravity (EPR, $8^{\circ}-8^{\circ}15'$ S). Shaded line indicates model prediction.

of hydrothermal cooling. The degree to which water cools the upper crust is poorly constrained, but some amount of cooling in combination with a contribution from thermal stresses yields melt amounts which are consistent with the tomography results. Incorporating a small degree of melt within the shallow mantle will somewhat reduce the amount modeled within the crust and should not change predicted topography and gravity dramatically. A small amount of melt within the mantle would be consistent with Moho melt lenses suggested by *Crawford et al.* [1999] and *Dunn et al.* [2000].

With an axial high that is dependent on thermal and magmatic structure of the crust, we can examine the response of morphology to varying model parameters. Increasing the thickness of the lithosphere (h_0) at the axis has the effect of reducing the amplitude of the high. A proxy for axial lithosphere thickness may be depth to a melt lens. At ridges with intermediate spreading rates, axial highs are sometimes present but may show a decrease in height [Sempéré et al., 1997; Cochran et al., 1997]. Where axial magma chambers have imaged beneath intermediate spreading ridges, they generally appear at

greater depths than for fast spreading ridges [Purdy et al., 1992]. This suggests a link between thermal structure immediately at the axis and amplitude of the high. This sort of dependence also occurs for lithosphere the structure of Kuo et al. [1986]. Eberle and Forsyth [1998a], on the other hand, find that their viscous stress model actually predicts a taller axial high for a deeper AMC.

A noted feature present over much of the southern East Pacific Rise is a \sim 70-m-deep, 5- to 10-km-wide trough at the base of the western side of the axial high [Eberle and Forsyth, 1998b]. Observations such as numerous seamounts on the western flanks of the axis [Scheirer et al., 1996], an asymmetric low-velocity zone at shallow mantle depths [Forsyth et al., 1998], and lower best fitting subsidence rates on the western flanks versus the eastern flanks [Eberle and Forsyth, 1998b] suggest a relatively warm thermal regime over the west flank of the EPR in this region. For relatively small values of the rate at which the lithosphere cools away from the axis (i.e., reducing our parameter S), the models display such a trough, though over a longer distance. The model might be able to better match the

dimensions of this trough if highly localized weakening effects, such as those created by faulting, were taken into consideration. We note that *Eberle and Forsyth* [1998b] were also able to model this trough by reducing the viscosity within the lower crust for their viscous stress model. Model runs by *Kuo et al.* [1986] exhibit a small trough at the base of the axial high for a plate with minimal off-axis thickening (in the extreme, a uniform thickness plate).

Correlations between morphology and indicators of thermal and magmatic structure of the crust have been observed over much of the EPR: Regions with a wider axial high are generally associated with a lower mantle Bouguer anomaly, the presence of a melt lens, more recent lava flows, and less fractionated basalts [Macdonald and Fox, 1988; Scheirer and Macdonald, 1993; Hooft et al., 1997]. The models we present here predict a slightly wider axial high for crust which cools over a greater distance from the axis (W larger) and for which a wider region of melt is present. This effect is most pronounced for cases with significant plate weaking (ω small). A wider melt region seems more likely to produce more recent eruptions and an imageable melt lens. These variations are likely both from segment to segment and on a smaller scale along-axis, where a widening of a high toward a segment center suggests a warmer and more magmatic regime there. Models based on the work by Madsen et al. [1984] that assume that the source of the high resides primarily in the mantle do not implicitly reproduce this dependence. If the crust is assumed to cool over a narrow region (implying rapid lithospheric thickening), the models will predict a wider high due to the increase in the plate's strength and hence flexural wavelength. This effect could be offset by narrowing the width of the melt region in the mantle and allowing only slight variations in lithospheric cooling rate. However, this then suggests that processes in the mantle, rather than the crust, will control the width of the high. Effects of these types of lithospheric variations on predicted topography using the Eberle and Forsyth [1998a] model have not, to date, been examined in detail.

Because of the thin plate approximation, rough estimates of temperature variation with depth, and complex effects of lithospheric weakening, we do not expect to solve for a precise temperature structure but rather for a feasible range. With this in mind, the above parameter systematics suggest causes behind correlations between various morphologic characteristics and other variable ridge properties indicative of crustal thermal and magmatic structure.

5. Summary

Topography and gravity at the East Pacific Rise can be modeled assuming effects of melt solidification, cooling of the lithosphere, and deflection due to thermal bending stresses. Fits to bathymetry and gravity data suggest two end-member lithosphere and melt scenarios are possible: either lithosphere (and thus crust) which cools over a very wide distance from the axis, with melt that extends to at least 40 km depth in the mantle, or lithosphere which cools over a very narrow region, with melt contained mostly in the crust. The latter is more consistent with seismic and compliance constraints on crustal thermal structure and is our preferred scenario. The models predict the amount of melt in the lower crust to be anywhere from 0% to 30%, depending on how temperatures are assumed to vary within the lithosphere and the degree which thermal stresses contribute to topography. Lower amounts of melt are needed when temperature profiles representing hydrothermal cooling are used and when the maximum effect of thermal stresses is considered (the three-dimensional nature of the lithosphere suggests the actual thermal stress contribution is slightly less than modeled). The amount of crustal melt needed may be reduced by the presence

of small amounts of melt within the mantle, as suggested by previous seismic and compliance results.

Varying model parameters suggest that certain morphologic features are highly dependent on thermal structure, including width and height of the high and presence of a flanking trough. Each of the parameter dependencies is consistent with other geologic observations. A wider high is produced given a wider region of warm material, consistent with correlations between width of a high and indicators of a "robust" magmatic structure. Reducing the rate of long-wavelength cooling will produce a trough at the base of the high, consistent of such troughs observed on the western flank of the southern EPR between 15°S and 20°S, where the presence of abundant seamounts and other observations suggest thinner lithosphere some distance from the axis. For thick axial lithosphere the model predicts a reduction in height of the high, comparable to that observed at some intermediate spreading ridges and consistent with observations of deep magma chamber reflectors at intermediate ridges. Lithospheric thickness near the ridge axis can be considered a major component of the initial transition from fast- to intermediate-style morphology, until discontinuity and local isostasy at the axis become less likely, with less frequent diking events.

Several simplifying assumptions are implicit in the model, such as a steady state thermal structure, two-dimensionality, simplistic thermal structures, and simplistic lithospheric weakening regimes. However, the model can predict segment-scale topography, assuming a thermal and magmatic structure consistent with gravity observations and other data.

References

- Buck, W. R., Accretional curvature of lithosphere at magmatic spreading centers and the flexural support of axial highs, J. Geophys. Res., 106, 3953-3960, 2001.
- Carbotte, S. M., J. C. Mutter, and L. Xu, Contribution of volcanism and tectonism to axial and flank morphology of the southern EPR, 17°10'-17°40'S, from a study of layer 2A geometry, J. Geophys. Res., 102, 10,165-10,184, 1997.
- Cochran, J. R., Analysis of isostasy in the world ocean, J. Geophys. Res., 84, 4713-4729, 1979.
- Cochran, J. R., J. A. Goff, A. Malinverno, D. J. Fornari, C. Keeley, and X. Wang, Morphology of a "superfast" mid-ocean ridge crest and flanks: The East Pacific Rise, 7°N-9°N, *Mar. Geophys. Res.*, 15, 65-75, 1993.
- Cochran, J. R., J.-C. Sempéré, and the SEIR Scientific Team, The Southeast Indian Ridge between 88°E and 118°E: Gravity anomalies and crustal accretion at intermediate spreading rates, J. Geophys. Res., 102, 15,463– 15,487, 1997.
- Crawford, W. C., S. C. Webb, and J. A. Hildebrand, Constraints on melt in the lower crust and Moho at the East Pacific Rise, 9°48'N, using seafloor compliance measurements, *J. Geophys. Res.*, 104, 2923–2939, 1999.
- Detrick, R. S., P. Buhl, E. Vera, J. Mutter, J. Orcutt, J. Madsen, and T. Brocher, Multi-channel seismic imaging of a crustal magma chamber along the East Pacific Rise, *Nature*, 326, 35-41, 1987.
- Dunn, R. A., D. T. Toomey, and S. C. Solomon, Three-dimensional seismic structure and physical properties of the crust and shallow mantle beneath the East Pacific Rise at 9°30'N, J. Geophys. Res., 105, 23,537-23,555, 2000.
- Eberle, M. A., and D. W. Forsyth, An alternative, dynamic model of the axial topographic high at fast spreading ridges, J. Geophys. Res., 103, 12,309-12,320, 1998a.
- Eberle, M. A., and D. W. Forsyth, Evidence from the asymmetry of fastspreading ridges that the axial topographic high is due to extensional stresses, *Nature*, 394, 360-363, 1998b.
- Eberle, M. A., D. W. Forsyth, and E. M. Parmentier, Constraints on a buoyant model for the formation of the axial topographic high on the East Pacific Rise, J. Geophys. Res., 103, 12,291-12,307, 1998.
- Forsyth, D. W., and D. S. Scheirer, MELT Seismic Team, Imaging the deep seismic structure beneath a mid-ocean ridge: The MELT Experiment, *Science*, 280, 1215-1218, 1998.
- Géli, L., et al., Evolution of the Pacific-Antarctic Ridge south of the Udintsev Fracture Zone, *Science*, 278, 1281–1284, 1997.
- Henstock, T. J., A. W. Woods, and R. S. White, The accretion of oceanic crust by episodic sill intrusion, J. Geophys. Res., 98, 4143-4161, 1993.
- Hirth, G., J. Escartin, and J. Lin, The rheology of the lower oceanic crust: Implications for lithospheric deformation at mid-ocean ridges, in *Faulting* and Magmatism at Mid-Ocean Ridges, Geophys. Monogr. Ser., vol. 106,

edited by W. R. Buck et al., pp. 291-323, AGU, Washington, D. C., 1998.

- Hooft, E. E. E., R. S. Detrick, and G. M. Kent, Seismic structure and indicators of magma budget along the East Pacific Rise, J. Geophys. Res., 102, 27,319-27,340, 1997.
- Hung, S.-H., D. W. Forsyth, and D. R. Toomey, Can a narrow, melt-rich, low-velocity zone of mantle upwelling be hidden beneath the East Pacific Rise? Limits from waveform modeling and the MELT experiment, J. Geophys. Res., 105, 7945-7960, 2000.
- Kuo, B. Y., D. W. Forsyth, and E. M. Parmentier, Flexure and thickening of the lithosphere at the East Pacific Rise, *Geophys. Res. Lett.*, 13, 681-684, 1986.
- Macdonald, K. C., and P. J. Fox, The axial summit graben and cross-sectional shape of the East Pacific Rise as indicators of axial magma chambers and recent volcanic eruptions, *Earth Planet. Sci. Lett.*, 88, 119-131, 1988.
- Madsen, J. A., D. W. Forsyth, and R. S. Detrick, A new isostatic model for the East Pacific Rise crest, J. Geophys. Res., 89, 9997-10015, 1984.
- Magde, L. S., R. S. Detrick, and the TERA Group, Crustal and upper mantle contribution to the axial gravity anomaly at the southern East Pacific Rise, J. Geophys. Res., 100, 3747-3766, 1995.
- McNutt, M., Compensation of Oceanic topography, an application of the response function technique to the Surveyor area, J. Geophys. Res., 84, 7589-7598, 1979.
- Parmentier, E. M., and W. F. Haxby, Thermal stresses in the oceanic lithosphere: Evidence from geoid anomalies at fracture zones, J. Geophys. Res., 91, 7193-7204, 1986.
- Parsons, B., and J. G. Schlater, Ocean floor bathymetry and heat flow, J. Geophys. Res., 82, 803-827, 1977.
- Phipps Morgan, J., and Y. J. Chen, The genesis of oceanic crust: Magma injection, hydrothermal circulation, and crustal flow, J. Geophys. Res., 98, 6283-6298, 1993.
- Press, W. H., B. P. Flannery, S. A. Teukolsky, and W. T. Vetterling, Numerical Recipes, The Art of Scientific Computing, Cambridge Univ. Press, New York, 1986.

- Purdy, G. M., G. Christeson, and S. Solomon, Relationship between spreading rate and the seismic structure of mid-ocean ridges, *Nature*, 355, 815-817, 1992.
- Scheirer, D. S., and K. C. Macdonald, Variation in cross-sectional area of the axial ridge along the East Pacific Rise: Evidence for the magmatic budget of a fast spreading center, J. Geophys. Res., 98, 7871-7885, 1993.
- Scheirer, D. S., K. C. Macdonald, D. W. Forsyth, S. P. Miller, D. J. Wright, M.-H. Cormier, and C. M. Weiland, A map series of the southern East Pacific Rise and its flanks, 15°S to 19°S, *Mar. Geophys. Res.*, 18, 1–12, 1996.
- Sempéré, J.-C., J. R. Cochran, and the SEIR Scientific Team, The Southeast Indian Ridge between 88°E-118°E: Variations in crustal accretion at constant spreading rate, J. Geophys Res., 102, 15,489-15,506, 1997.
- Talwani, M., J. L. Worzel, and M. Landisman, Rapid Gravity computations for two-dimensional bodies with application to the Mendocino submarine fracture zone, J. Geophys. Res., 64, 49-59, 1959.
- Toomey, D. R., S. C. Solomon, and G. M. Purdy, Tomographic imaging of the shallow crustal structure of the East Pacific Rise at 9°30'N, J. Geophys. Res., 99, 24135-24157, 1994.
- Turcotte, D., and G. Schubert, Geodynamics: Applications of Continuum Physics to Geological Problems, John Wiley, New York, 1982.
- Vera, E. E., J. C. Mutter, P. Buhl, J. A. Orcutt, A. J. Harding, M. E. Kappus, R. S. Detrick, and T. M. Brocher, The structure of 0- to 0.2-m.y.-old oceanic crust at 9°N on the East Pacific Rise from expanded spread profiles, J. Geophys. Res., 95, 15529-15556, 1990.
- Wang, X., and J. R. Cochran, Gravity anomalies, isostasy, and mantle flow at the East Pacific Rise crest, J. Geophys. Res., 98, 19,505-19,531, 1993.

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