The Genesis of Oceanic Crust:
Magma Injection, Hydrothermal Circulation, and Crustal Flow

JASON PHIPPS MORGAN

Institute of Geophysics and Planetary Physics, Scripps Institution of Oceanography, La Jolla, California

Y. JOHN CHEN

College of Oceanography, Oregon State University, Corvallis, Oregon

In this study we construct a thermal and mechanical model for the genesis of oceanic crust. Magma is hailed in its ascent within the oceanic crust when it reaches a freezing horizon, where the dilational volume change associated with magma freezing leads to viscous stresses that favor magma ponding near the freezing horizon. To model the steady state thermal impact of crustal accretion via dike injection and pillow flows, we treat all crustal accretion in rocks cooler than a magma “solidus” to occur in a narrow 250-m-wide dike-like region centered about the ridge axis. The rest of the oceanic crust is modeled to be emplaced as a steady state magma lens directly beneath the “solidus” freezing horizon where the steady state emplacement rate is determined by the constraint that this lens supply all crust that is not emplaced through dike/extension above the magma lens. If hydrothermal heat transport within crustal rocks cooler than 600°C removes heat 8 times as efficiently as heat conduction, then we find that a steady state magma lens will only exist within the crust for ridges spreading faster than a 25 mm/yr half rate. The depth dependence of the magma lens with spreading rate is in good agreement with seismic observations. These results suggest that a fairly delicate balance between magmatic heat injection during crustal accretion and hydrothermal heat removal leads to a strongly different crustal thermal structure at fast and slow spreading ridge axes. Our results support the hypothesis that median valley topography is due to extension of strong ridge axis lithosphere; it is the difference in thermal regime that is directly responsible for the striking difference between the typical median valley seen at slow spreading ridges (e.g., Mid-Atlantic Ridge) and the axial high seen at fast spreading ridges (e.g., East Pacific Rise). This paradigm for the origin of a median valley at a slow spreading ridge predicts that along-axis variations in median valley topography of a slow spreading center reflect variations in recent magmatic heat input along a segment, that is, that the axial topography is a good time-averaged indicator of the relative importance of hydrothermal cooling and magmatic injection along a given section of a ridge segment.

We determine the accumulated crustal strain associated with lower crustal flow which supports the hypothesis that the Oman Ophiolite crust was created at a paleo-analogue to a fast spreading ridge and also suggests that crustal strain, and not cumulative layering, may be the dominant physical process that generates “layered gabbros” within the Oman Ophiolite.

INTRODUCTION

Recent seismic studies of fast and slow spreading ridges suggest that a quasi steady state magma chamber exists at shallow ~1.5 km depths beneath the axis of a fast spreading ridge but is not present within the crust at a slow spreading center. This observation may be a fundamental clue to the physical origin for the characteristic difference in axial topography between the median valley topography of a slow spreading ridge and the axial high topography of a fast spreading ridge. Observations of crustal layering in the Oman Ophiolite can also be used to constrain the pattern of magma injection and crustal flow at a spreading center. After reviewing these observational constraints we present a simple model for heat and mass transport within oceanic crust that suggests that a fairly delicate balance between magmatic heat injection during crustal accretion and hydrothermal heat removal leads to a strongly different crustal thermal structure at fast and slow spreading ridge axes, that is that a difference in thermal regime is directly responsible for the observed differences in axial morphology.

SEISMIC OBSERVATIONS

Multichannel seismic (MCS) studies along the East Pacific Rise (EPR) have yielded intriguing clues to the depth, thickness, and along-strike continuity of the axial magma chamber at a fast spreading ridge [Detrick et al., 1987; Mutter et al., 1988; Harding et al., 1989; Vera et al., 1990; Kent et al., 1990; Detrick, 1991]. MCS data collected on the northern EPR [Detrick et al., 1987; Mutter et al., 1988] show a bright reflector interpreted as the top of an axial magma chamber along much of the spreading center. Figure 1 shows a typical across-strike MCS profile. The reflector is continuous where the axial depth and morphology point to strong magmatic activity, and it disappears as the axis deepens and narrows. The magma chamber reflector is found along roughly 60% of the rise axis between the Siqueiros transform and 13°30'N. After correction for the diffraction effects associated with the magma chamber reflector, the inferred across-strike width of the magma chamber reflector is 1-1.5 km [Kent et al., 1990].

Analysis of expanding spread profile data confirms that the reflector coincides with a sharp interface between high velocities above and low velocities below [Harding et al., 1989; Vera et al., 1990]. Analysis of the expanding spread profiles [Harding et al., 1989; Vera et al., 1990] and tomographic inversion of seismic travel time data [Burnett et al., 1989; Caress et al., 1990; Toomey et al., 1990] reveal a broader, lower-amplitude, low-velocity zone underlying the thin (~100-500 m) axial magma chamber reflector (see Figure 1). This axial low-velocity zone is roughly 6 km wide, extends from a depth of 1.5-2.0 km down to Moho, and consists of velocities de-
pressed by 0.5 to 1.0 km/s. This small reduction in velocity is consistent with mafic or ultramafic rocks containing less than a 3% melt fraction [Caress et al., 1992].

The picture of axial structure that emerges from the northern EPR seismic work includes a 6-km-wide zone of hot plutonic rocks with a small or negligible amount of melt capped by a thin (<100-500 m thick), narrow (0-2 km wide) lens of high melt fraction. The narrowness of the actual magma chamber and its further constriction or interruption at small axial discontinuities are consistent with the petrological observation that even the smallest discontinuities form boundaries between different magma sources. However, the combination of a surprisingly small volume of subaxial melt and the along-axis continuity of the magma chamber for tens of kilometers suggests that melt production must be quite steady and that the magma must generally cycle through the axial system rapidly.

This is a radically different magma chamber structure than that proposed in many previous ideas of a crustal–thickness–size chamber, although it is essentially the magma chamber structure envisioned by Sleep [1975; 1978]. It is consistent
with ophiolite evidence for 100 m scale cryptic geochemical variation within the crustal gabbros of the Oman Ophiolite [Browning, 1984], and a magma lens of the size seen at the EPR is large enough to easily generate the major extrusives recently mapped along the Cocos-Nazca Ridge [Macdonald et al., 1989]. Gravity data along the EPR are also consistent with a small magma chamber [Madsen et al., 1990], as is along-axis petrologic segmentation [Sinton and Detrick, 1992]. Sinton and Detrick [1992] present an excellent recent overview of this emerging paradigm about the axial structure of a fast spreading ridge axis, and Phipps Morgan [1991] also summarizes recent work on this subject. J.E. Quick and R.P. Denlinger (Ductile deformation and the origin of layered gabbro in ophiolites, submitted to Journal of Geophysical Research, 1992) (hereinafter referred to as Quick and Denlinger, 1992) present a petrological investigation and Henstock et al. [1992] present a geophysical investigation of this type of magma chamber structure that strongly complement the results of the present study.

To date, seismic reflection work on the Mid-Atlantic Ridge (MAR) shows no evidence for magma chamber reflectors, suggesting that magma chambers are transient features on slow spreading ridges [Detrick et al., 1990]. Along-axis variations in velocity structure near the TAG hydrothermal mounds area of the MAR does, however, suggest a recent intrusion of hot rock in this region [Kong et al., 1992]. Petrological data also indicate that long-lived magma chambers are not a common feature of slow spreading ridges [cf. Nattland, 1980]. In fact, Toomey et al. [1988] and Kong et al. [1992] show that microseismic activity beneath the MAR axis extends to depths of 6-10 km, implying that the brittle layer is present, at least episodically, to this depth. Huang and Solomon [1988] present teleseismic data which supports this conclusion. Their data, shown in Figure 8, also show that the thickness of the seismogenic ridge axis lithosphere decreases with increasing spreading rate. Since the depth to the seismic brittle-ductile transition is thought to roughly correspond to the 750°C isotherm [Bergman and Solomon, 1984; Wiens and Stein, 1983], this observation implies generally cooler crust at slower spreading ridges. Purdy et al. [1992] show a complementary relationship: the depth to the top of the axial low-velocity zone gets shallower with increasing spreading rate. Similarly, microearthquake observations [Riedesel et al., 1982] suggest that the brittle layer extends to a depth of only ~2 km depth at a fast spreading ridge where large amplitude ridge topography is absent.

The strong correlation between an axial seismic "magma chamber" and a surficial "fast spreading" axial morphology extends to intermediate spreading ridges. MCS lines across the Juan de Fuca Ridge [Morton et al., 1987; Rohr et al., 1988] and the Valu Fa Ridge in the Lau Basin [Morton and Sleep, 1985b; Collier and Sinha, 1990] also show an axial magma chamber reflector where "a fast spreading" axial high morphology exists. There is a slight spreading rate dependence of the depth of the magma lens with the shallowest lens depths at the fastest spreading rates [cf. Purdy et al., 1992]. These data and uncertainties are plotted in Figure 7. Unlike Purdy et al. [1992] we do not mix measurements of the depth to a magma lens with measurements of the top of the axial low-velocity zone at slower spreading rates. We feel that the depth to an axial low-velocity zone is a different thermal measurement than the depth to (or existence of) a magma lens and thus should not be directly compared with observational magma lens constraints.

**Ophiolite Observations**

As noted above, the Oman ophiolite, thought to be created at a fast spreading center [cf. Nicolas, 1989], shows cryptic chemical variation which suggests that only a small magma body was present at any time during crustal accretion [Browning, 1984]. In addition this ophiolite shows a structural sequence through the gabbro section from isotropic gabbros directly beneath the sheeted dike/gabbro contact grading downwards to more strongly "layered" gabbros near the gabbro/mantle peridotite interface [e.g. Pallister and Hopson, 1981; Nicolas et al., 1988]. Changing layer dips within the gabbro section have been used to infer the shape of a nearridge magma chamber in which these "cumulate layers" are deposited [Smeing, 1981]. In particular, this fabric development shown in Figure 2 was cited by Smeing [1981] and Pallister and Hopson [1981] as evidence for a broad, gently dipping magma chamber that deposited tilted cumulate layers on its floor.

Ophiolite crustal rocks also record evidence of near-ridge hydrothermal alteration [Harper et al., 1988; Nehlig and Juteau, 1988]. Below the volcanics, sheeted dikes, and upper gabbros of the Oman ophiolite, hydrous activity appears to be concentrated near faults or within narrow, apparently tonized zones [Nehlig and Juteau, 1988]. This suggests that there may be a link between hydrothermal activity and the style of extensional deformation.

**Heat and Mass Transport Within Oceanic Crust**

In this section we will construct a thermal and mechanical model for the genesis of oceanic crust that we feel is a good candidate model to integrate the above observations into a coherent synthesis. The basic mechanical model is not particularly new, and its conceptual foundation has already been presented by Phipps Morgan et al. [1987], Lin and Parmentier [1989], and Chen and Morgan [1990]. Its most important conceptual ingredient is that it is the interplay between magmatic crustal injection and hydrothermal cooling that is responsible for the presence of a quasi steady state magma lens beneath a fast spreading ridge and the dramatic change in axial morphology between a fast and slow spreading ridge axis (because axial morphology directly reflects the lithosphere strength or thickness across a ridge axis which is quite sensitive to this thermal balance). The resulting magma injection and crustal flow structure that we envision is extremely similar to the qualitative crustal accretion scenario sketched by Sleep [1975; 1978] and also similar to the subsiding magma chamber floor model of Dewey and Kidd [1977].

What is new in this study is that we consider here that crustal accretion at a fast spreading ridge is strongly shaped by the maximum height to which magma can rise in a steady state manner before freezing. We also extend this model to calculate the strain associated with lower crustal flow, so we may compare model predictions of crustal strain associated with a given magma injection geometry with observations from the Oman Ophiolite.

**Magma Ascent and Emplacement**

Magma ascent within the oceanic crust is probably controlled by two complementary processes: (1) magma will ascend only if buoyancy forces (or, if large enough, viscous pres-
Gabbro Layering & Genesis -- Oman Ophiolite [Smewing, 1981]

![Diagram of Diking and Magma Chamber]

**Feeder Zone**

Fig. 2. Schematic structural cross section of the crustal section of the Oman Ophiolite. Figure is redrawn from Smewing [1981]. The stratigraphic sequence below the pillow and sheeted dike complexes is shown on the right side of the figure. Isotopic gabbros just beneath the sheeted dike complex grade into gabbros with a (weakly developed) near vertical dip which becomes more developed [Nicolai et al., 1988; Nicolai, 1989] and more shallowly dipping as one moves deeper into the gabbro section. (The layering is best developed and parallel to the Moho directly above the gabbro-peridotite "petrologic Moho" contact.) This dip structure in particular was used by Smewing [1981] to infer that gabbro layering reflects cumulative deposition on the floor of the large magma chamber sketched here. We propose that these structures are equally well explained as the by-product of crustal strain shown in Figure 4, that is, that the structures are the result of crustal flow from a magma lens intrusive zone much like that imaged seismically along a large fraction of the present-day East Pacific Rise (see Figure 1).

Pressure gradients) cause it to; thus magmas may rest their ascent for a while at zones of neutral buoyancy until they have fractionated and reacted to the point where the magma's density is less than that of surrounding host rock (this idea has been explored in particular by Ryan [1987]) and (2) magma can also be halted in its ascent when it reaches a freezing horizon, where the dilational volume change associated with magma freezing leads to viscous stresses that favor magma ponding within roughly one viscous "compaction length" (term coined by McKenzie [1984]) of the freezing horizon. This idea has been recently proposed in a particularly simple form by Sparks and Parmentier [1991] to explain melt focusing to a narrow neovolcanic zone at a spreading center. In this study we will consider only the effects of magma freezing in limiting melt ascent and ignore density effects on lower crustal melt transport, reaction, and segregation; while magma neutral density effects in particular may play an important role in shaping the chemical stratification in oceanic crust we wish initially to explore the idea that it is the Sparks-Parmentier magma freezing effect that leads to a quasi steady state magma lens beneath a fast spreading ridge.

Since magma freezing must occur (by definition) within the sheeted dike and extrusive sections of oceanic crust, these accretion processes, unlike a magma lens, must be fundamentally transient in nature. To model the steady state thermal impact of crustal accretion via dike injection and pillow flows we treat all crustal accretion in rocks below a magma "solidus" to occur in a narrow 250-m-wide dike-like region centered about the ridge axis. Within this region, magma emplacement rates are taken to be equal to the spreading rate divided by the width of the diking region, that is, all shallow crust is modeled to be emplaced within this region. Although this correctly treats heat injection within the sheeted dike section, this approximation does somewhat overemphasize the importance of magma heat injection within the pillow section. Since pillow flows are extruded on the seafloor, burying previous flows, they rapidly cool to the ambient seawater temperature and essentially advect this cold boundary layer downward with subsequent pillow burial and subsidence [Sleep, 1991]; C. A. J. Harding et al. (A multichannel seismic investigation of upper crustal structure at 9°N of the East Pacific Rise: Implications for crustal accretion, submitted to Journal of Geophysical Research, 1992) (hereinafter referred to as Harding et al., 1992).

The maximum amount that we will underestimate the depth of the magma lens isotherm is given by the thickness of the pillow section above the magma lens. Recent MCS results constrain the axial thickness of the pillow section to be ~200 m along the axis of the EPR, thickening to ~600 m within 2 km of the ridge axis (Harding et al., 1992).

The rest of the oceanic crust is modeled to be emplaced as a steady state magma lens directly beneath the "solidus" freezing horizon. We take the ~1-km-wide ~250-m-thick prismatic shape inferred from recent EPR seismic studies cited above to be a kinematic constraint on the shape of this magma lens. Once we have this shape, the steady state emplacement rate is determined by the constraint that this lens supply all crust which is not emplaced through diking/extrusion above the magma lens. The depth of this lens is controlled by the depth of the magma "solidus" (here taken to be 1200°C) determined from a self-consistent thermal structure for a spreading center. Thus the lens will cease to exist in this model if the steady state thermal structure places this solidus isotherm...
beneath the crust. In addition, the injection rate within the lens will diminish as the lens moves deeper into the crust, since magma injection into the lens only supplies magma for crustal sections below the sheeted dike complex. Note that this "solidus" temperature should be viewed as the effective temperature at which the magma is sufficiently crystallized to behave mechanically as a strong, viscous fluid. While the freezing interval of a cooling basaltic liquid will have important effects on the chemical/petrologic evolution of magma and rock, it will only have a secondary effect on the thermal structure and flow away from a magma lens and will therefore be neglected within this initial study of the physics of crustal flow and heat transport. For the same reason we do not address convection and magma fractionation processes within the magma lens. See Quick and Denlinger (1992) for an analysis of some of the petrologic consequences of crustal genesis by predominant melt injection within a small magma lens.

The thermal structure at a spreading center is predominantly influenced by two factors in this model: (1) the depth and injection rate within a potential steady magma lens, and (2) the efficiency of hydrothermal circulation in removing heat through rocks that are cool enough to permit cracking and hydrothermal heat transport. To assess this conceptual model we have implemented it as a finite element code based on the code described by Chen and Morgan [1990]. We have modified the flow solver of this code to solve for a flow problem which includes the effects of magma injection or removal instead of purely incompressible viscous flow. Details of our solution algorithm are presented in the Appendix.

Hydrothermal Cooling

The primary remaining uncertainty in these numerical experiments is how to appropriately parameterize the form and magnitude of these effects of hydrothermal circulation on shaping heat transport within the crust. We choose the formulation developed by Phipps Morgan et al. [1987] who use the results of Combarroux [1978] to treat hydrothermal heat transport as an enhanced thermal conductivity within the temperature and depth range where hydrothermal activity occurs. Where hydrothermal cooling occurs, the ordinary thermal conductivity is enhanced by a factor \( N_u \), the Nusselt number or ratio of hydrothermal heat transport within a permeable layer to heat transport by heat conduction alone. Rock at temperatures greater than 600°C or a depth greater than 6 km is assumed to be impermeable. Since water as hot as 400°C discharges from vents on the seafloor, this is a minimum value of the maximum temperature through which water must circulate. Phipps Morgan et al. [1987] and Sleep [1991] use a value of 400°-450°C for this cutoff temperature, while Morton and Sleep [1985a] and Wilson et al. [1988] prefer a hotter 600°C cutoff for hydrothermal circulation. Morton and Sleep [1985a] suggest that the limit temperature should be at least that obtained by extrapolating the surface hydrothermal venting temperature down to 45-MPa pressures at a shallow magma lens which yields a 465°C cutoff [Bischoff and Rosenbauer, 1984]. In addition, they suggest that hydrothermal cooling will rapidly cool the near crack environs so that the average rock temperature can be \(-600°C\) with fluid present in locally cooler cracks.

For the purposes of this study we can sidestep this question to some degree because the hydrothermal heat loss will be governed by an approximate product of the hydrothermal heat transport enhancement factor \( N_u \) times the hydrothermal cutoff temperature, e.g., \( q_{\text{hydro}} \approx (T_{\text{cutoff}} - T_{\text{water}})N_u/\varepsilon_{\text{cutoff}} \).

Thus for a lower \( T_{\text{cutoff}} \) we find that we will need a higher \( N_u \) value to produce a magma lens at a given depth for a fast spreading numerical experiment. Figure 3 shows the strong trade-off between \( T_{\text{cutoff}} \) and \( N_u \) in cooling off the axial upper crust, illustrating graphically that once we can determine \( T_{\text{cutoff}} \) from rock and water chemistry observations, then we can use this approach to determine the effective additional heat transport by hydrothermal circulation and so estimate the effective permeability within the hydrothermal system. Gregory and Taylor [1981] report that subsolidus oxygen isotope exchange occurred mainly within the upper 5-6 km of the ophiolite, thus giving an estimate of the maximum depth of water penetration. In these studies we generally choose to model hydrothermal heat transport as \( N_u = 8 \). We choose this value because it leads to a solution where a steady state magma lens can exist at 1.2- to 1.5-km depths beneath a fast spreading ridge and not exist beneath a slowly spreading ridge (see Figure 7).

As noted above, deep hydrothermal alteration within the Oman ophiolite seems to be concentrated near faults. This observation suggests that it may be a reasonable hypothesis that crustal extension through faulting opens channels for hydrothermal flow and these channels lead to a higher effective hydrothermal cooling enhancement in slow spreading environments where large median valley bounding normal faults and 6- to 10-km-deep seismically active faults are present (Figure 8a). In this study we will only briefly explore a spreading rate dependence to the efficiency of hydrothermal heat transport. If this dependence does exist (which we feel is likely), it is likely to be an enhancement from \( N_u = 8-10 \) at fast spreading ridges to \( N_u = 12-15 \) at median valley ridges. We choose not to include this effect because it will only enhance the already strong trends that are seen in the following suite of numerical experiments.

**MODEL RESULTS**

At a fast spreading ridge where hydrothermal circulation

![Fig. 3. Plot of the depth to the top of an axial magma lens described in the text for a half-spreading rate of 50 mm/yr and various hydrothermal heat transport enhancement factors \( N_u \) and cutoff isotherms \( T_{\text{cutoff}} \) above which hydrothermal flow ceases. The effectiveness of hydrothermal cooling will determine how deep a steady state magma lens will reside. Thus if we can determine \( T_{\text{cutoff}} \) by a geochemical means, then the depth of a steady state magma lens will directly constrain \( N_u \). See text for more explanation of this relationship.](image-url)
removes heat 8 times as efficiently as purely conductive crustal cooling when the crust is cooler than 600°C, magma freezes at 1200°C, and the latent heat of solidification is 334 kJ/kg, a steady state magma chamber exists 1.35 km below the seafloor 0°C bounding isotherm (see Figure 4). All magma that forms the lower crust rises to this level, freezes, and then flows to deeper crustal levels.

Strain and the Development of Gabbro Layering

Once we have solved for a steady state flow field, the accumulated crustal strain associated with crustal flow is found using the formulation and techniques developed by McKenzie [1979]. Figure 4 shows that at a fast spreading ridge the accumulated strain during this flow process is most intense at deeper crustal levels. Strain becomes progressively more intense and more flat lying as the Moho is approached; the strain within the lowermost kilometer of the crust is too strong to effectively show with with the "stretched ellipse" convention shown in Figure 4. A comparison of Figure 4 with Figure 2 clearly shows that crustal strain can produce layering with the dip and layer development seen in the Oman Ophiolite, thus providing an appealing explanation for the orientation and strength of lower crustal layering. In contrast, for a slow spreading ridge (Figure 5), no steady state magma lens exists within the crust for the same hydrothermal cooling enhancement, while for an intermediate spreading half rate of 30 mm/yr a steady state magma lens can exist at a deeper level within the crust (see Figure 6).

**Spreading Rate Dependence of a Magma Lens**

Figure 7 shows the depth to a magma injection lens plotted versus spreading rate where Nu and the solidus temperature are fixed and only the spreading rate is varied. In this case there is a fairly abrupt transition with spreading rate from a shallow steady state magma lens at a 30 mm/yr spreading half rate to no steady state magma lens within the crust at a 20 mm/yr spreading half rate. There is good agreement with seismic observations of the depth to a magma lens as a function of spreading rate which are also plotted in Figure 7. (Note, however, that the higher Nu=12 shown in Figure 7 is more consistent with the 30 – 35 mm/yr spreading half rate which is roughly observed [Macdonald, 1986; Small and Sandwell, 1989] as the transitional spreading rate between median valley and axial high relief, but it is less consistent with the depth of the magma lens at fast spreading ridges; these results support a slight hydrothermal enhancement associated with median valley extension.) We performed a suite of numerical experiments in which the magma lens was assumed to be a 2-km-wide, 500-m-thick body, that is twice the width and more than 4 times the volume of the lens in the above numerical experiments. We found that the depth of the lens is most strongly controlled by the balance between the rate of magma injection within the lens and hydrothermal cooling; to first order a bigger lens does not influence the net rate of magma injection and hence does not affect the depth of the lens. To second order, a wider lens is more efficiently cooled at the axis, resulting in a slightly (200-300 m) deeper 2-km-wide lens than a 1-km-wide lens for
the same $N_u$, hydrothermal cutoff temperature, and spreading rate.

We can use Figures 4 and 5 to assess the implications of a deeper magma injection lens beneath a fast spreading ridge. If magma injection is relatively uniform with depth (like in Figure 5) then there will not be major differences in accumulated strain with depth in a crustal or ophiolite section. In the accumulated strain hypothesis for the development of layering, it is where the flow “turns” a corner that the straining is most intense; thus if injection occurred at the bottom of the crust with crustal flow to shallower levels, then we would expect the opposite sense of layer development to that seen in the Oman Ophiolite. It is only a shallow-level intrusive center that leads to isotropic gabbrons at shallow stratigraphic levels underlain by progressively more deformed gabbro sections. Thus the layering development, intensity, and gabbro dips in the Oman Ophiolite support a scenario where crustal injection to form this crust occurred by predominant magma freezing within a narrow sill near the sheeted dike-isotropic gabbro contact. This is in agreement with previous assertions that this crust was created at an analogue to a fast spreading ridge like the present EPR spreading center [e.g. Nicolas, 1989].

Implications of Along-Axis Magma Transport and Injection

Currently, there is a debate about whether magma emplacement along a fast spreading ridge is fairly continuous along-axis or confined to a few discrete volcanic centers that are foci of axial accretion processes [e.g. Macdonald et al., 1988; Phipps Morgan, 1991]. Lin and Phipps Morgan [1992] note that while gravity and topography data show that slow-spreading ridges have a clear along-axis variation in crustal thickness (i.e., integrated magma supply varies along-axis), the much smaller along-axis gravity and topography variation at a fast spreading ridge can be explained by either a more two-dimensional (2-D) pattern of upwelling and melting beneath a fast spreading ridge or by a well-connected, temporally persistent magma lens (or low-viscosity zone) that smooths the along-axis crustal structure at a fast spreading ridge. Within this latter scenario we can interpret the above model results as modeling crustal flow due to along-axis feeding of the sheeted dike section and the magma lens; to first order the model will not be different if magma supply comes from the side instead of from locally beneath the crust. For slow spreading ridges, the model assumption of a 2-D crustal flow structure away from a diking region is not as strong. Again, along-axis crustal flow will preferentially occur in the hottest, weakest regions since plate extension tends to confine flow of stronger regions to the plate-spreading direction. However, the crustal thickness is not strongly smoothed by along-axis flow resulting in ~3 km along-axis crustal thickness variation [e.g. Blackman and Forsyth, 1991]. The resulting heat transport can be partially approximated by considering 2-D vertical slices in the plate spreading direction which vary in crustal injection rate (or crustal thickness) from a thick segment-center slice to a thinner crust segment-end slice. However, a quantitative
comparison of the along-axis implications of this model for a slow spreading ridge really demands a full three-dimensional (3-D) model treatment which is beyond the scope of the current study. In addition, transient magma injection effects will become more important in regions where a quasi-continuous along-axis magma lens does not exist.

Implications for Axial Topography

A viscous flow-induced pressure drop due to asthenosphere ascending in a narrow subridge channel or conduit is one frequently cited explanation for forces which lead to a median valley [Sleep, 1969; Lachenbruch, 1973, 1976; Sleep and Rosendahl, 1979]. However, the strength variations required to form a conduit are difficult to explain solely on the basis of feasible near-ridge thermal structures like those shown in this study (compare Figure A1) as well as in previous studies [Phipps Morgan et al., 1987; Lin and Parmentier, 1989; Chen and Morgan, 1990]. Phipps Morgan et al. [1987] note an additional significant difficulty for viscous "conduit" theories for median valley topography. How can they explain the existence of fossil median valley topography on 10-100 Ma

Fig. 6. Temperature, flow, and crustal strain for a model spreading center with an opening half-rate of 30 mm/yr, $N_u = 8$ and $T_{outoff} = 600^\circ$C. Thermal structure is shown by levels of grey shading; crustal flow by arrows, and crustal flow lines by light lines emanating from the intrusion region. The extent of the magma intrusion zone is shown by the heavy contour; in this case intrusion within the crust is deeper than in Figure 4, reflecting the reduction in magma emplacement compared with a faster spreading ridge. Accumulated strain is shown at 35,000-year time steps along the crustal flow lines. In this figure there is a similar but deeper and weaker gradient in accumulated crustal strain compared to that in Figure 4.

Fig. 7. Depth to the top of the magma lens as a function of spreading rate, all other parameters being held constant (the same as in Figures 4–6, which plot as the 50, 10, and 30 mm/yr points on this curve). Solid line shows results from a suite of numerical experiments with $N_u = 8$ and $T_{outoff} = 600^\circ$C. For these hydrothermal cooling parameters a steady state magma lens can exist within the crust only at spreading half rates greater than about 20 mm/yr. A well-developed shallow magma lens exists only for spreading rates greater than 30 mm/yr. Dashed line shows results from a suite of numerical experiments with $N_u = 12$ and $T_{outoff} = 600^\circ$C. For these hydrothermal cooling parameters a steady state magma lens can exist within the crust only at spreading half rates greater than about 30 mm/yr. A well-developed shallow magma lens exists only for spreading half rates greater than 40 mm/yr. Solid squares (and associated uncertainties from Purdy et al. [1992]) show multichannel seismic observations of the depth of the magma lens along intermediate and fast spreading ridges (JdF is Juan de Fuca [Morton et al., 1987; Rohr et al., 1988]; Lau is Lau Basin [Collier and Sinha, 1990]; and EPR is East Pacific Rise [Detrick et al., 1987; Purdy et al., 1992]). There is good agreement between model predictions of the depth dependence of the magma lens with spreading rate and multichannel seismic observations.
old abandoned spreading centers when the decay time for the viscous stress supported topography above an axial conduit is of the order of only 30,000 years.

Phipps Morgan et al. [1987] proposed that the extension of a strong ridge axis lithosphere layer may be the origin of median valley topography as originally suggested by Tapponier and Francheteau [1978]. (Note that Tapponier and Francheteau [1978] actually develop an elastic–isostatic model for the formation of median valley topography that is a different mechanism from their original “necking” suggestion and that wrongly predicts an extremely thin crustal “hole” at a slow spreading ridge axis.) Phipps Morgan et al. [1987] show that moments due to lithospheric stresses within a brittle plate that is 8 km thick at the ridge axis and thickens by only a few kilometers within the 30 km halfwidth of the axial valley can produce the typical axial topography of a slow spreading ridge. Lin and Parmentier [1990] developed an elastic/plastic idealization of plate extension which allows them to directly explore the transient development of extensional rift valley topography. They found that the form of the rift valley depends on the thickness and thickness variations in the stretching lithosphere and that the lithosphere strength-supported topography remains after extension stops, successfully explaining the persistence of failed rift topography. In this scenario, the axial high typically seen at fast spreading ridges is not supported by lithosphere stretching stresses which are very small across the thin, weak, axial lithosphere of a fast spreading ridge. Instead the axial high is isostatically supported by the low density melt lens and region of “hot rock” directly beneath the ridge axis.

This study supports the idea that it is the balance between magmatic heat input and hydrothermal heat removal which determines the thickness (≈ yield strength) of the axial lithosphere, which in turn controls the axial morphology associated with plate boundary extension. The axial “strength” of the lithosphere is a qualitative measure of the magnitude of the horizontal extensional stress that can be supported during ridge axis extension. Figure 8 shows the axial yield strength as a function of spreading rate for two crustal and mantle rheologies: (1) where the crust and mantle rheology are both described by an olivine brittle-ductile rheology, and (2) where the crust is described by a weaker diabase rheology while the mantle is described by an olivine rheology. (See Chen and Morgan [1990] for more discussion of an appropriate ridge axis rheological structure.) Again, in all numerical experiments the only physical parameter that varies is the spreading rate. Independent of detailed crustal rheology, there is a strong increase in axial yield strength once the half spreading rate drops below ≈ 20 mm/yr. The large variation in integrated axial yield strength with spreading rate shown in Figure 8 is a likely reason for the typical presence of a strong lithosphere extension-generated median valley at a slow spreading ridge and its absence at a fast spreading ridge where a shallow melt lens is commonly seen. Figure 8a shows that the depth to the 750°C isotherm found in these experiments correlates well with the spreading rate dependent maximum earthquake slip depths inferred from teleseismic and microseismic studies.

Axial Variability Along a Single Ridge Segment

This paradigm for the origin of a median valley at a slow spreading ridge would predict that along-axis variations in median valley topography of a slow spreading center reflect variations in recent magmatic heat input along a segment, that is, that the axial topography is a good time-averaged indicator of the relative importance of hydrothermal cooling and magmatic injection along a given section of a ridge segment. Thus along-axis variations in crustal thickness should be correlated with variations in median valley relief, with thicker (= hotter) axial crust at relative lows in median valley relief along a segment [Phipps Morgan, 1991]. This hypothesis is supported by current evidence about along-axis variations in the temperature structure of a slow spreading ridge found from seismic velocity variation and along-axis variations in maximal microearthquake hypocentral depths. Figure 9 shows a recent hypocenter distribution along a segment of the Mid-Atlantic Ridge [Kong et al., 1992]. There is a strong positive correlation between maximum hypocentral depths and ridge axis bathymetry and a similar relationship between along-axis velocity structure and ridge axis bathymetry. These correlations imply that the shallowest sections of a ridge segment are underlain by a hotter, weaker crust; at a given depth, a hotter temperature leads to slower seismic velocities. The axial depth/maximum hypocentral depth relationship shown in Figure 9 also implies that the shallowest sections of a ridge segment are underlain by a thinner, strong, brittle lithosphere, suggesting that the volume of crustal injection (magmatic heat input) shapes the lithosphere thickness along a spreading center.

DISCUSSION

This study has focused on marshaling further evidence for two aspects of an emerging paradigm for the origin of axial topography at a mid-ocean ridge: (1) stretching of strong axial lithosphere is responsible for the origin of a median valley along a spreading center, and (2) the thickness of this strong lithosphere is controlled by a delicate balance between heat input by magma injection and removal by hydrothermal circulation so that the presence or absence of axial topography directly reflects the axial thermal structure along a ridge. This paradigm is shown to be consistent with a thermally controlled magma “solidus” temperature which limits the height at which magma rises to and can exist as a quasi steady state feature at a spreading center. It is consistent with the melt lenses recently imaged along fast spreading ridges but which have not been found where an axial valley morphology is present. We have also shown that the progressive development and orientation of gabbro layering with increasing depth within the crustal section of the Oman Ophiolite support this emerging paradigm if the development of layer-structures within the crust reflects how much viscous strain a gabbro has experienced rather than reflecting a cumulative “settling” mechanism.

While we feel these results are robust, there are several aspects of this study that should be improved and explored in future studies. The first is time dependence. We chose in this study to examine as simple a model as possible and thus parameterize dike injection above the gabbro sections of the crust as a time-averaged steady state process despite the inherently time-dependent nature of individual dike intrusion episodes. Thus this study cannot address to what extent and over what time scales a time-dependent intrusion process would change the temperature structure of oceanic crust. See Henstock et al. [1992] for an initial analysis of these effects.

We also have ignored the possibility of crustal magmas assimilating and reacting with gabbros during their ascent through the crustal section; an investigation of these processes will require a better description of the effects of chemical reaction on the density of both magma and host rock. Fortunately, one
(a) Depth (km) vs. temperature with lines for 750°C and 1200°C.

Nu = 8

(b) Depth (km) vs. yield strength at different spreading rates (10, 30, and 50 mm/yr).

(c) Total yield strength vs. half-spreading rate for olivine rheology and olivine-diabase.

Nu = 8
is becoming rapidly available [cf. Ghiorso, 1987]. Quick and Denlinger (1992) have made an initial analysis of these effects. Finally, we have neglected the effects of a three-dimensional along-axis variation in axial lithosphere strength in shaping along-axis variations in ridge morphology. While simpler 2-D models yield intuition into how lithosphere stretching can be related to the amplitude and shape of the resulting axial valley, a 3-D stretching model is needed to accurately predict the magnitude of topographic variations associated with along-axis variations in magma injection and the resulting variations in thermal lithosphere structure. These further studies are needed to test the conclusions and implications of this work. The task is clearly doable now; let us do it!

APPENDIX: PROBLEM FORMULATION AND SOLUTION

The problem we consider is steady state heat and mass transport beneath a spreading center. The temperature structure of the ridge satisfies a conservation equation for energy due to thermal diffusion (Fourier's Law), heat advection due to viscous fluid motions, and a latent heat release on cooling from a magma to “rock” phase.

Conservation of Energy

\[ L \psi + \nabla \cdot (\bar{u} T) = \kappa \nabla^2 T \]

where \( \psi \) is the magma injection rate, \( L \) is the magma's latent heat of cooling (334 kJ/kg) converted into an effective “superheat” of 320°C, \( \kappa = 10^{-6} \text{ m}^2/\text{s} \), \( \bar{u} \) is the velocity, and \( T \) is the temperature. Viscous flow within the crust and mantle is treated to satisfy momentum and mass conservation for a variable viscosity fluid that is incompressible except in dilational regions of magma injection.

Conservation of Mass

\[ \nabla \cdot \bar{u} = \psi \]

Conservation of Momentum

\[ \tau_{ij} = \mu \frac{\partial u_i}{\partial x_j} + \frac{\partial u_j}{\partial x_i} + \lambda \delta_{ij} \nabla \cdot \bar{u}. \]

\[ \rho s + \tau_{ij,} = 0. \]

where \( p \) is the pressure, and \( \mu \) and \( \lambda \) are the shear and bulk viscosity, respectively [cf. Tritton, 1977]. In the above equations we use a comma to represent differentiation by the subsequent index and the summation convention applies to an implicit sum of repeated indices. While the shear viscosity of mantle and crustal rocks is poorly known, the bulk viscosity is even less well known, so for this suite of numerical experiments we choose \( \lambda = \mu \). In addition, we consider in this suite of experiments that mantle upwelling is solely driven by plate spreading and we do not treat buoyant contributions to mantle upwelling that are discussed by Phipps Morgan [1991]. In this suite of experiments we are trying to limit our initial model complexity to the oceanic crust.

The above equations are solved using a finite element formulation which extends the formulation reported by Chen and Morgan [1990]. Since the viscosity is a function of temperature which is a priori unknown we must solve this problem by iterating to a steady state solution. For speed and computer memory savings we treat the tempera re and flow subproblems separately within a larger iteration loop.

Weak Form of Energy Conservation

For temperature we solve the streamline-upwind Petrov-Galerkin weak form of this equation using finite element techniques discussed by Brooks and Hughes [1982]:

\[ 0 = \int_{\Omega} (w + q)(-L \psi - T \psi - u_T d\Omega - \kappa \int_{\Omega} w T d\Omega \]

where \( w \) is the standard Galerkin weighting or basis function (here the basic functions of a bilinear quadrilateral element) and \( q \) is an additional discontinuous streamline upwind weighting function given by \( q = \kappa u x y / \|u\|^2 \), where \( \kappa = (\xi u_x h_{\text{ms}} + \eta u_y h_{\text{ms}}) / 2 \) and \( \xi, \eta \) in terms of a local element Peclet number \( Pe \) in the local \( \xi, \eta \) direction are found from \( \xi = (\cosh(Pe_\xi) - Pe_\xi)^{-1} \) with a similar expression for \( \eta \). There are several advantages to this formulation as an “upwind” finite element advection solver. It has no crosswind artificial diffusion because \( q \) acts only in the direction of flow. It also provides a numerically consistent weighting for the energy source term \( \nabla \psi \) in the equation.

Numerical Flow Problem: Solution Algorithm

The finite element solution for viscous flow within a region with a prescribed dilatation is a new feature of this study. The problem is extremely similar to that for an incompressible fluid which is well summarized by Reddy [1984] and Hughes [1987]; the difference is that we now prescribe the dilatation to have a potentially nonzero value. In what follows we will present a penalty finite element formulation for this problem. A standard approach to developing a finite element formulation is to pose the governing equations in terms of a variational principle. The standard penalty formulation for slow viscous flow (Stokes’ Equation) is presented in this way by Reddy [1984]. The variational form

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Fig. 8. Integrated axial lithosphere strength for these experiments plotted as a function of spreading rate. (a) Depths to the 750°C and 1200°C isotherm are shown from a suite of numerical experiments with \( Nu = 8 \) and \( \text{T cutoff} = 600°C \). The 750°C isotherm correlates well with the spreading rate dependence inferred from centroid depths (shown by small dots and associated bars representing inferred total rupture depth) of Huang and Solomon's [1988] teleseismic study. Triangles are microearthquake focal depths beneath the axis of the northern Mid-Atlantic Ridge [Toomey et al., 1988; Kong et al., 1992]. There is good agreement between model isotherms and the spreading rate dependence of the depth of the seismically observed brittle/ductile transition. (b) Mantle yield strength envelopes used in the calculations for 10 mm/yr, 30 mm/yr, and 50 mm/yr. Solid lines show the yield strength envelope for a crust and mantle with an assumed brittle/ductile olivine rheology. Dashed lines show the yield strength envelope where the crust has a diabase rheology while the mantle has an olivine rheology (see Chen and Morgan [1990] for more details of this rheological structure). (c) Depth-integrated total axial yield strength as a function of spreading rate. There is a strong increase in total axial yield strength at spreading half rates below 20 mm/yr which suggests that extension of strong axial lithosphere is a likely mechanism for the presence of a median valley at slow-spreadings ridges and the absence of a median valley relief along fast-spreadings ridges.
Along-Axis Cross section of Seismic Velocity and Microearthquake Hypocenters (Northern MAR)

Fig. 9. Along-axis variations in seismic velocity and microearthquake focal depths along the trans-Atlantic geotraverse (TAG) median valley near 26°N on the Mid-Atlantic Ridge. Heavy labeled contours show $P$ wave velocity in kilometers per second. Filled circles show microearthquake hypocenters constrained with $P$ and $S$ wave arrivals; open circles show more poorly constrained hypocenters constrained solely by $P$ wave arrivals. (Events within 2 km of the along-axis vertical plane are shown here.) Axial seafloor topography is shown as the solid line. Note the along-axis deep is roughly 800 m deeper than the along-axis high. Solid squares show the locations of the ocean bottom hydrophones that were located along the ridge axis (two ocean bottom seismographs and seven ocean bottom hydrophones were used in this experiment). A strong correlation exists between maximum microearthquake focal depths (equal to thickness of the brittle lithosphere?) and ridge axis bathymetry. A shallower microearthquake depth cutoff is found beneath this along axis depth minimum. The along-axis depth minimum also appears to be seismically slower than deeper sections of the ridge axis, suggesting that this is the site of hotter as well as thinner lithosphere. This evidence suggests that the lithospheric thickness at the ridge axis may be thermally controlled and that thicker axial lithosphere correlates with a deeper median valley (figure modified from Kong et al. [1992]).

$$I(u, v) = \int \left\{ \mu \left[ \left( \frac{\partial u}{\partial x} \right)^2 + \left( \frac{\partial v}{\partial y} \right)^2 + \frac{1}{2} \left( \frac{\partial u}{\partial y} + \frac{\partial v}{\partial x} \right)^2 \right] \right. $$

$$+ \frac{\lambda}{2} \left( \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right)^2 + f_s u + f_y v \right\} \, dx \, dy$$

with $\vec{u} = (u, v)$ satisfies momentum conservation. This can be seen by direct construction of the Euler-Lagrange equations. If we now modify the variational form by adding an additional Lagrange multiplier

$$L(u, v, \gamma) = I(u, v) + \frac{\gamma}{2} \int_0^L \left[ \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} - \Delta(x, y) \right] \, dx \, dy$$

then this variational form will satisfy both mass conservation

$$\left( \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right) = \Delta(x, y)$$

and momentum conservation

$$\frac{\partial}{\partial x} \left[ 2\mu \frac{\partial u}{\partial x} + \lambda \left( \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right) + \frac{\partial}{\partial y} \left( \frac{\partial u}{\partial y} + \frac{\partial v}{\partial x} \right) \right] - \frac{\partial p}{\partial x} = f_x$$

$$\frac{\partial}{\partial y} \left[ 2\mu \frac{\partial v}{\partial y} + \lambda \left( \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right) + \frac{\partial}{\partial x} \left( \frac{\partial u}{\partial y} + \frac{\partial v}{\partial x} \right) \right] - \frac{\partial p}{\partial y} = f_y$$

with the definition $p = -\gamma \left( \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} - \Delta(x, y) \right)$. From the above relations we find the recipe for modifying a standard penalty formulation for incompressible viscous flow to solve the dilatational problem. We form the momentum conservation statement at an element level

$$k^e \vec{u}^e + g^e \vec{p}^e = f^e$$

where $f^e$ are the standard nodal loads due to body forces or surface tractions for viscous flow, $g^e$ is the finite element form of the gradient operator, and $k^e$ is the element stiffness matrix that is identical in form to that for a elastic problem with Lamé parameters $\lambda$ and $\mu$ as analogues to the bulk and shear viscosity, respectively (This notation follows that of Hughes [1987, chap. 4], who presents an excellent introduction to the penalty method for incompressible viscous flow.) We now use the previously found relation for pressure in terms of a penalty number $\gamma$, $p^e = -\gamma (\partial u/\partial x + \partial v/\partial y - \Delta(x, y))$ to substitute velocity unknowns for pressure unknowns in the momentum equation which yields

$$\vec{K}^e \vec{u}^e = \vec{f}^e$$

where $\vec{K}^e = k^e - \gamma g^e (g^e)^T$ is modified exactly as done for penalized incompressible viscous flow and the force vector
\( \tilde{f}^e = f^e - \gamma g^e \Delta^e \) is also modified during element formation. The terms involving \( \gamma \) are all evaluated using reduced integration. The resulting solution algorithm shares all of the advantages and potential pitfalls [Pelletier et al., 1989] of the penalty algorithm which is a current workhorse for solving 2-D variable viscosity incompressible flow. See Hughes [1987] for more information.

We implement a further modification to the classical penalty method to solve this problem. Pelletier et al. [1989] show that if we implement the penalty formulation as a single step of an Uzawa algorithm, then by iteration we can find a solution that satisfies the continuity equation to any desired precision. We solve an iterative sequence

\[
\tilde{k} \mathbf{u}_n = -g p_{n-1} + \tilde{f}
\]

where we update the pressure after each iteration by

\[
p_n = p_{n-1} + \gamma (\nabla \cdot \mathbf{u}^n_{n-1} - \Delta) \]

until the divergence of the velocity field is sufficiently small. For a large enough \( \gamma > O(10^6 \mu) \) this iteration converges extremely rapidly (2-5 iterations). The first iteration with a pressure guess \( p = 0 \) is simply the standard penalty solution, which is already quite good for large \( \gamma \). Note also that in the direct Gaussian elimination solution implementation of the finite element equations we only have to form and factor the stiffness matrix once for this problem; thus subsequent iterations are extremely cheap to compute.

**Rheology and Boundary Conditions**

Although we are primarily interested here in solving for crustal flow and thermal structure, we need to also solve for mantle flow to properly treat the thermal and mechanical effects of a growing lithosphere. Thus we solve for heat and mass transport within a 90 x 140 km region on one side of a symmetric ridge axis as shown in Figure A1. Each of the figures in the main text was solved in this larger computational domain. The viscous rheology that we use is a simplification of that presented by Chen and Morgan [1990]. For this study we assume a Newtonian lithosphere, mantle, and crustal viscosity structure with a lithosphere viscosity which is \( 10^6 \) greater than the asthenosphere (mantle) viscosity. The asthenosphere viscosity is, in turn, \( 10^7 \) times greater than the viscosity of "hot crust" above 750°C, resulting in a total viscosity range of \( 10^7 \). (The total strength range is limited by the dynamic viscosity range of a penalty algorithm which is about \( 10^7 \) for a double precision numerical code [Hughes, 1987].) The lithosphere-asthenosphere transition within the mantle is governed by the 750°C isotherm, and the dike injection region is set to have a strength 10 times less than that of the lithosphere. These

**Problem Solution Region and Boundary Conditions**

![Fig. A1. Problem geometry and boundary conditions for the numerical experiments performed in this study. The problem region that we consider is a 140-km-wide by 90-km-deep region on one side of a symmetric ridge axis. Mantle flow is driven solely by plate spreading, and crust is emplaced at the ridge axis according to thermal and geometrical criteria developed in the text. We solve the problem on a 63(x direction) x 80(y direction) variable spacing tensor-product grid with an x and y nodal spacing shown by tick marks along the top and right hand sides of the region. The problem boundary conditions are shown on each side of the box. The lightly shaded box is the subregion from which the solution is extracted to make the detail plots shown in Figures 4-6. This subregion contains 43 vertical by 30 horizontal grid points. The example temperature and flow field shown here is for a half spreading rate of 10 mm/yr where no steady state magma lens exists within the crust (the run from which Figure 5 was extracted). Solution isotherms are contoured at 200°C intervals.](image-url)
strength contrasts roughly approximate the rheology in \textit{Chen and Morgan}'s \cite{Chen:1990} study and were chosen to eliminate an additional nonlinear solution iteration needed to solve a non-Newtonian flow problem \citep{Chen:1990}. The temperature boundary conditions for this problem are that the mantle is flowing into this region at a constant temperature \(T_m = 1350^\circ\text{C}\), that heat is free to move out of the sides of the box, and that on the top surface \(T_0 = 0^\circ\text{C}\). For flow, the vertical velocity and horizontal shear stress are zero at the top of the box; on the sides of the box we assume that passive plate spreading flow occurs beneath the rigid part of the lithosphere, and we assume the bottom of the region to be a shear and normal stress free surface. See \textit{Chen and Morgan} \cite{Chen:1990} for further discussion of these boundary conditions. The boundary condition that is unique to this problem is that for the influx of magma in the crustal accretion zone. We treat this magma, for a geometry which is determined as discussed in the main body of the text, as inflowing at a steady state rate which exactly balances crust leaving the region with plate spreading. The effective temperature of this inflowing material is set to be higher than the magma solids by an amount equal to the energy released as latent heat of cooling (see \textit{Phipps Morgan et al.} \cite{Phipps:1987} for further discussion of this type of boundary condition). All of this latent heat of cooling is released at the top of the magma lens.

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