The relative importance of plate-driven and buoyancy-driven flow at mid-ocean ridges

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Abstract. The dynamical interaction between three-dimensional (3-D) buoyant flow and plate-driven mantle flow beneath a mid-ocean ridge is examined using a combination of laboratory and numerical experiments. In a unique laboratory setup, a layer of strongly temperature-dependent viscous fluid is heated from below and cooled from above to drive thermal convection. Forced, plate-driven flow is modeled by dragging mylar sheeting in opposite directions across the fluid surface. Uniform viscosity 3-D numerical models have been designed to simulate the laboratory runs to provide additional information on the flow, and to attempt to isolate the effects of variable viscosity. In one type of experiment we modeled buoyancy on the scale of the partial melting region with a linear heat source beneath the spreading axis. In a second set of experiments the entire base of the tank is heated in order to investigate the interaction between plate-driven flow and larger-scale (upper mantle) convection. The pattern of segment-scale (100-150 km) convection beneath a spreading center is found to be a strong function of the spreading rate and the Rayleigh number of the buoyant flow. Purely two-dimensional (2-D) flow exists only in the case of low Rayleigh number (10^5) and faster spreading rates (>4.6 cm/yr half rate). Three dimensionality is strongly enhanced during transient pulses of upwelling. Convection on the scale of the upper mantle can contribute significant long-wavelength spatial (~600-1000 km) and temporal (20-40 m.y.) variability in upwelling rates and temperatures at spreading centers and may provide an alternative model for plume-ridge interaction. For a given Rayleigh the upwelling near the spreading axis can essentially be described by three regimes: weakly 3-D at slow spreading rates, strongly 3-D at slow to intermediate rates, and 2-D at fast spreading rates. The regime boundaries shift toward higher plate velocities with increasing Rayleigh number. Comparison of the laboratory and numerical experiments indicates that temperature-dependent viscosity may strengthen the position of focused upwelling centers and cause a sharper transition between 2-D and 3-D upwelling patterns. Taken together, these results suggest that buoyancy-driven, dynamic flow is an important element in the geodynamics of mid-ocean ridges.

Introduction

The pattern of mantle flow beneath mid ocean ridges is still a poorly understood aspect of seafloor spreading. The original, and simplest, model of mantle flow beneath spreading centers consists of a forced flow driven by viscous coupling between the diverging plates and the underlying asthenospheric mantle [Reid and Jackson, 1981]. Plate-driven flow will be essentially two-dimensional (2-D), except near ridge offsets, with upwelling velocities proportional to spreading rate. This mantle flow model has been widely used to investigate melting processes beneath oceanic spreading centers and to predict the relationship between mantle temperature, crustal thickness and the geochemistry of the basaltic crust [e.g., Phipps Morgan and Forsyth, 1988; Langmuir et al., 1993]. However, there is growing evidence from both geophysical observations of along-axis variations in ridge crest structure and theoretical studies of mantle geodynamics for the existence of a strong dynamic component to mantle flow beneath mid-ocean ridges that segments and focuses the broad mantle upwelling predicted by simple plate-driven flow models. This dynamic flow is inherently three-dimensional (3-D) and can result in significant along-axis variations in upwelling velocities and melt production and may also contribute to the fundamental along-axis segmentation of mid-ocean ridges [e.g., Whitehead et al., 1984; Crater, 1985].

Mid-ocean ridges are characterized by significant along-axis variations in depth at wavelengths ranging from a few kilometers to thousands of kilometers (Figure 1). The longest-wavelength depth anomalies appear to be related to
Figure 1. (a) Axial gravity and bathymetry data for slow, intermediate, and fast spreading ridges showing the increase in variability with decreasing spreading rate [from Lin and Phipps Morgan, 1992]. (b) mantle Bouguer anomaly (MBA) gradient versus spreading rate [from Wang and Cochlain, 1993].
regional variations in upper mantle temperature, typically associated with the existence of near-axis hotspots (e.g., Azores or Iceland) [Le Douaran and Francheteau, 1981; Schilling, 1986]. However, significant axial depth variations (tens to hundreds of meters) also occur at the much shorter, segment-scale wavelengths related to transform and nontransform offsets [Francheteau and Ballard, 1983; Macdonald et al., 1988]. The deepest ridge crest depths are located near segment boundaries, with shallower depths near the segment midpoints. These along-axis depth variations are greatest along slow spreading ridges and near major transform offsets.

The gravity field along slow spreading ridges also displays a distinctive "bells-eye" pattern with negative anomalies centered over segment midpoints and positive anomalies near segment boundaries [e.g., Kuo and Forsyth, 1988; Lin et al., 1990; Neumann and Forsyth, 1993; Detrick et al., 1995]. These gravity anomalies arise largely from along-axis variations in crustal thickness [Lin and Phipps Morgan, 1992; Tolstoy et al., 1993] which reflect significant variations in the delivery of melt to the base of the crust, consistent with 3-D mantle upwelling. There is a systematic spreading rate dependence to this gravity signature, with higher along-axis gravity gradients at the slow spreading Mid-Atlantic Ridge than at the intermediate to fast spreading Cocos-Nazca Ridge and East Pacific Rise [Lin and Phipps Morgan, 1992, Wang and Cochran, 1993] (Figure 1).

These contrasting gravity and topographic signatures have been interpreted as indicating that upwelling is intrinsically plume-like (3-D) beneath slow spreading ridges but more sheet-like (2-D) beneath fast spreading ridges due to the increasing importance of plate-driven flow over buoyant upwelling as spreading rate increases [Lin and Phipps Morgan, 1992]. However, the smaller along-axis gravity gradients observed at fast spreading ridges may not require a 2-D mantle upwelling pattern but instead may reflect a masking of an inherently 3-D flow pattern by a more efficient along-axis redistribution of magma or ductile deformation of the hot, lower crust at fast spreading ridges [Lin and Phipps Morgan, 1992; Bell and Buck, 1992; Wang and Cochran, 1993].

Convective upwelling may also play a role in creating and maintaining ridge segmentation. Dynamic mantle flow generates stresses on the base of the lithosphere that may affect the evolution of the geometry of a spreading center [Rabinowicz et al., 1993]. There is an excellent correlation between mantle Bouguer anomaly (MBA) amplitude and segment length along parts of the Mid Atlantic Ridge [Lin et al., 1990; Detrick et al., 1995], which suggests a link between mantle flow and ridge segmentation, but the physical mechanism responsible for this correlation is poorly understood.

In order to provide new insight into these questions we have carried out a series of laboratory experiments and numerical studies focusing on the relative importance of plate-versus buoyancy-driven flow beneath oceanic spreading centers. Laboratory experiments have been an underutilized approach for studying the geodynamics of mid-ocean ridges and offer the potential for investigating complex 3-D flow regimes using fluids with spatially variable material properties (temperature, composition, viscosity). Numerical studies yield more quantitative information on the flow and its thermal and chemical attributes but are computationally limited to simplified models or idealized rheologies. The combined laboratory and numerical approach utilized in this study exploits the strengths of both techniques to better understand the dynamics of mantle flow beneath mid-ocean ridges and its affect on the crustal accretion process.

We have carried out approximately 30 laboratory experiments which simulate plate-driven flow using an apparatus that drags mylar sheets across a viscous fluid surface at a fixed speed. In addition, we have conducted numerical experiments which match laboratory parameters as closely as possible. We report here on two subsets of experiments which differ in the mode of heating used to model buoyancy-driven flow and the scale of mantle processes investigated. In the first set of experiments we use a line source of thermal buoyancy, similar to Whitehead et al. [1984], in order to focus on localized buoyant instabilities near the spreading axis, on the scale of ridge segmentation. In the second set of experiments the entire base of the tank is heated in order to investigate the interaction between plate-driven flow and larger-scale regional (upper mantle) convection. In these experiments we find purely 2-D flow at ridges only in the case of slow Ra (~10^5) and fast spreading rates (~4-6 cm/yr). For all other combinations we see a strong buoyant component which is sampled at the ridge due to the large-scale flow. These results suggest that buoyancy-driven, dynamic flow is an important element in the geodynamics of mid-ocean ridges.

Previous Laboratory and Numerical Studies

Relatively few laboratory studies have been carried out to investigate mantle flow in an experimental setting that is representative of a mid-ocean ridge. Richter and Parsons [1975] conducted a series of experiments using a layer of viscous silicon oil that is sheared by the motion of the top boundary. Initial convective patterns that consist of polygons, transverse rolls (aligned perpendicular to shear), and a mixture of transverse and longitudinal rolls (aligned parallel to shear) all transform into a longitudinal roll pattern. The time required for the rolls to reach a steady amplitude decreases with increasing Rayleigh number and increasing plate velocity, becoming relatively invariant at high plate velocities. Probably the most influential laboratory experiment on mid ocean ridges was the study by Whitehead et al. [1984]. In this experiment a water-glycerine mixture was injected into a glycerine-filled tank along a horizontal line leading to the development of gravitational instabilities as widely spaced, diapir-like bodies. This simple experiment helped to crystallize in the minds of many ridge workers the potential importance of buoyancy-driven, 3-D flow at mid-ocean ridges and the possible connection between this type of flow and the along-axis segmentation of mid-ocean ridges. The work reported here differs from the Whitehead et al. [1984] study in that we include the effects of plate-driven flow and examine the interaction between plate-driven and buoyancy flow as a function of spreading rate. More recent laboratory studies have considered the dynamical interaction between a spreading ridge and on-axis [Feighner and Richards, 1995] and off-axis [Kincaid et al., 1995] plumes created by a localized heat source. In this study we do not a priori impose a plume heat source, but rather we look at the natural growth and evolution of plume-like upwellings from heat sources that are uniform along-axis.

Significant progress has been made over the past decade in the development of more realistic 2-D and 3-D numerical
models of mantle flow and melt transport beneath mid-ocean ridges. Early 2-D studies examined the relative strengths of passive, plate-driven flow and buoyant upwelling by modeling flow in axis-perpendicular planes [Buck and Su, 1989; Scott and Stevenson, 1989; Sotin and Parmentier, 1989; Cordery and Phipps Morgan, 1993]. The sources of buoyancy in these studies were thermal expansion, the presence of melt, and compositional density variations due to melt extraction. All of these sources of buoyancy result in enhanced and focused upwelling beneath the ridge crest. The reduction in bulk density and viscosity of the mantle due to melt-filled porosity can have a large focusing effect on the upwelling [Buck and Su, 1989]. However, the magnitude of retained melt fractions is sensitive to the permeability of the mantle, which is poorly known. Therefore, the importance of melt buoyancy is still unknown. Compositional density variations are caused by the decrease in density of peridotite during the formation and segregation of partial melt. During partial melting, iron is preferentially partitioned into the melt phase, relative to magnesium, leaving the residual mantle progressively less dense as melt is extracted [Oxburgh and Parmentier, 1977]. As a result of flow through the melting region, the mantle far from the spreading axis becomes stably stratified, with the degree of depletion decreasing with depth. This stable stratification tends to suppress thermally driven convective overturn. The interaction of thermal and compositional buoyancy was found to often give rise to time-dependent, oscillatory flow patterns [Scott and Stevenson, 1989; Sotin and Parmentier, 1989]. Additionally, buoyant upwelling tends to reduce the spreading rate dependence of crustal production that is characteristic of passive flow models [e.g., Reid and Jackson, 1981].

More recent 3-D numerical studies have found that buoyant upwelling is inherently 3-D. Compositional density variations due to melt extraction cause the upwelling beneath an unsegmented plate boundary to become focused in the along-axis direction into broad centers of enhanced upwelling and melting, separated by narrow regions of reduced upwelling [Parmentier and Phipps Morgan, 1990]. This 3-D pattern occurs at slow spreading rates and low mantle viscosities. The additional buoyancy due to small amounts of retained melt enhances this 3-D instability and can further focus the upwelling into discrete centers [Jha et al., 1994a]. Thermal buoyancy also enhances three-dimensionality by giving rise to off-axis longitudinal rolls which form near the spreading axis at slower spreading rates and low mantle viscosities [Sparks and Parmentier, 1993].

The numerical experiments presented here are designed to duplicate as much as possible the experimental tank apparatus. The rationale for a combined laboratory and numerical approach is that both techniques have inherent strengths and weaknesses, and a combined approach will lead to a fuller understanding of the dynamical processes under investigation. Laboratory experiments offer the potential of investigating complex, time-dependent, 3-D flow regimes in fluids with spatially variable viscosity. The numerical studies yield quantitative 3-D information on the flow and temperature fields. However, the bulk of previous 3-D numerical studies of convection beneath ridges, while including important effects such as melt generation and transport, have assumed idealized rheologies due to computational limitations. In this work we compare the results of the laboratory and numerical work to begin to explore the role of temperature-dependent rheology on mantle dynamics beneath ridges, to assess the results of previous work, and to guide future observational and numerical ridge studies.

**Experimental Apparatus and Procedure**

A laboratory apparatus and an efficient experimental operating procedure have been developed for modeling a wide range of mantle dynamics beneath oceanic spreading centers. The first-order features of mid-ocean ridge systems which are essential to the laboratory models are (1) large-scale mantle flow driven by coupling with surface plates, (2) a spreading axis created by divergence of two plates, and (3) local sources of buoyancy.

Figure 2 illustrates the essential features of the apparatus used for modeling mantle geodynamics beneath ridges. This apparatus consists of a rectangular glass tank of inner dimensions 96x72 cm and a depth of 18 cm resulting in a tank aspect ratio of roughly 6x4x1. The working fluid is a concentrated sucrose solution manufactured by Archer Daniels Midland. Relevant material properties of the solution are listed in Table 1. The sucrose fluid is transparent and may be mixed with organic dyes to facilitate flow visualization and is easy to work with in the lab because it is nontoxic and readily dissolves in hot water. The fluid's large viscosity dependence on temperature is an important characteristic for modeling mantle convective processes. The relationship between dynamic viscosity (μ in Pascal seconds) and temperature (T, in degrees Celsius) for this fluid is similar to thermally activated creep in the mantle and can be approximated by an Arrhenius law, where viscosity varies exponentially with inverse temperature [Olson and Kincaid, 1991]:

$$\mu = \exp \left( \frac{1888}{T + 9.3} \right) \cdot 11.48.$$  \hspace{1cm} (1)

For the range in temperatures explored in these experiments, 15-55°C, the viscosity varies from 380 to 3 Pa s.

The plate-driven or forced flow is simulated by dragging two mylar sheets in opposite directions over the surface of the fluid. Each mylar sheet is threaded around a set of bars which insure contact with the working fluid. The sheets are pulled around the spreading axis bars, across the fluid to the take-up bars at the tank walls, and up to a receiving roller (Figure 2). Heating elements are bolted to the take-up bars so that the mylar and viscously coupled fluid are heated slightly before passing through a scraper system to separate mylar and fluid. The driving force is supplied by a synchronous, high-torque DC motor coupled to a gear reducer, enabling us to model a wide range in plate spreading rates. The support structure for the ridge drive system is constructed of lexan and consists of two subcomponents: a main support structure and a spreading ridge module. The support structure closely fits inside the tank and is supported by leveling posts. This structure holds the take-up bars, the scraper system, the take-up spools, and a set of gears for coupling to the drive system. The ridge module is a lexan car which holds the mylar spool reels and the ridge bars and attaches to the main lexan support structure. The car can be replaced by multiple sets of offset cars for future studies of segmented ridges.

Thermal buoyancy is supplied by a combination of cooling the fluid surface and heating the base of the tank. The upper surface of the fluid is cooled by circulating water from a refrigerated bath through a series of (70x10x2 cm) metal
The figure shows a schematic of apparatus highlighting the range of heating mechanisms and showing ridge-perpendicular and ridge-parallel views. Mylar is threaded from spool reels down through the ridge bars and across the fluid surface. The mylar is then passed through a take-up bar and scraper system to separate the mylar and fluid, up to a take-up reel. This system is driven by a DC motor to provide plate-driven flow in the fluid. The strip-heater and base-heater setups are also shown.

The tank wall and an outer 0.5-cm-thick Plexiglas wall, sealed with foam insulation. The fluid is allowed to thermally equilibrate for several days at room temperature prior to all experiments.

The temperature of the fluid can be continuously monitored at a number of points on and within the apparatus using small (0.2 x 0.2 x 0.1 cm) platinum resistance temperature detectors (RTDs). RTDs are placed on the base of the tank (upper and lower surfaces), along the surface of the mylar, and just below the surface of the fluid along the ridge axis. Temperature information from all RTDs is automatically logged, and temperatures are later corrected to remove any bias in particular RTDs. Fluid circulation/temperature patterns are recorded through the sides of the tank using still and video cameras. Polarized light is passed through the tank onto a diffusing screen. Focusing and defocusing of light is related to the curvature of the temperature field such that hot thermal boundary layers (e.g., around plume heads) and cold thermal boundary layers appear bright and dark, respectively. These images (shadowgraphs) provide integrated views through the tank of planes oriented perpendicular ($K_L$) and parallel ($K_T$) to the ridge axis (Figure 2).

The experiments are scaled to the mantle flows through the dimensionless Peclet number ($Pe$) and Rayleigh number ($Ra$):

$$Pe = U_p D / \kappa$$

$$Ra = \rho g \alpha \Delta T D^3 / \kappa \mu$$

where $\rho$, $g$, $\alpha$, $\kappa$, and $\mu$ are fluid density, gravity, coefficient of thermal expansion, thermal diffusivity, and dynamic viscosity, respectively. Fluid thickness $D$, plate velocity $U_p$, and vertical temperature drop across the fluid, $\Delta T$, are chosen to achieve a realistic range in mantle $Pe$ and $Ra$.
Numerical Model

Three-dimensional, uniform-viscosity numerical models, based on models of spreading centers [Sparks et al., 1993], were used to simulate the laboratory experiments. Finite difference approximations were used to solve the Stokes equations for viscous fluid flow and the time-dependent energy balance for temperature. The flow was decomposed into two components, a passive flow driven by the moving top boundary and a buoyancy-driven flow with no-slip top and bottom boundaries. The two-dimensional plate-driven flow was calculated using a propagator matrix method [see Sparks et al., 1993, Appendix B] for a uniform viscosity layer over a half-space. A viscosity increase between the layer and half-space of 1000 confines most of the flow to the layer, approximating the no-slip boundary at the base of the tank. The stream function-vorticity formulation for the buoyant flow was solved using a multigrid iterative technique. For simplicity, the vertical side walls in the numerical experiments were stress-free boundaries. The numerical techniques used to solve for both the flows and temperature are described by Sparks et al. [1993]. All of the numerical experiments were conducted in boxes that contained only half of the tank geometry, with the spreading axis placed at one end of the box. While this halves the computational work, it also constrains the plane of the spreading axis to be a symmetry plane. This was observed to be true of the laboratory experiments almost all of the time.

A fixed temperature was used for the top and bottom boundaries, and the side walls were insulating. For the strip-heated numerical experiments, all of the fluid in the tank is initially at a uniform temperature, and a higher temperature is imposed along the region of strip heating. The temperature of the surface of the laboratory fluid is somewhat variable, because, even when the mylar is cooled from above, it is not possible to cool the fluid in the space between the ridge-axis bars. The numerical temperatures we refer to are calculated just below the surface of the fluid. These temperatures follow roughly the same spatial and temporal patterns as the temperatures measured in the lab and have the same dependencies on \( Ra \) and \( Pe \). However, the absolute values of the variations are not exactly the same.

Since the numerical experiments have uniform viscosity, it is necessary to choose a \( Ra \) that models the behavior of the fluid in the tank. For the full basal-heating experiments, we use a \( Ra \) based on the average temperature in the fluid, which increases slightly over the duration of the experiment as the tank warms up. For the strip-heated experiments we found that the \( Ra \) needed to be based on the cold ambient fluid. Viscosity values of 50 and 100 Pa s produced results that roughly bracket the behavior of the lab experiments.

Strip-Heated Experiments

Experimental Design

The strip-heating setup examines the interaction of buoyancy-driven and plate-driven flow in the shallow, subaxial mantle. Assuming the working fluid thickness represents the uppermost 100 km of mantle in (2), a half-spreading rate of 0.5 cm/min in the laboratory corresponds to 3 cm/yr in the mantle. A series of strip heating experiments was run with the drive system operating at a range of plate velocities. In the majority of cases, the mylar spreading system was turned on 720 s after the heating unit was turned on (when the thermal boundary layer is approximately halfway to the plume onset temperature of 55°C) and left on for the duration of the experiment. A few experiments were run in which the drive was turned on closer to the plume onset time; however, the morphology of the flow was very similar to previous counterparts. Also, some experiments were run without the drive system to characterize plume onset times, rise times, and morphologies without the added complexity of a large-scale shear flow. Plate velocities and additional physical parameters for these experiments are given in Table 2.

A 10-cm-wide strip heater is used to introduce thermal energy into a linear region below the spreading axis. The resulting density deficit at the base of the fluid is meant to approximate a buoyancy source local to the mantle immediately beneath the spreading axis. In most of these experiments we input 60 V to the resistance heater, which produces a thermal boundary layer with an inside basal temperature approaching a roughly steady value of 55°C after

<table>
<thead>
<tr>
<th>Experiment</th>
<th>Mylar Speed, cm/min</th>
<th>Scaled Plate Velocity, cm/yr</th>
<th>Drive Initiation Time* ( \text{min/psi} )</th>
<th>Inside Basal</th>
<th>Maximum Axial Variation</th>
<th>Steady Axial Variation</th>
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<tr>
<td>1</td>
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*Mylar drive on at \( t_{on} \) seconds; plumes initiate at \( t_{pi} \) seconds
1300 s. The density contrast due to thermal expansion between the thermal boundary layer and room temperature fluid is 1.5%, while the average viscosity of the fluid is, from (1), about 100 Pa s. The resulting $Ra$ of about $10^9$ corresponds to a mantle with a reference density of 3300 kg m$^{-3}$, thermal expansion of $3 \times 10^{-5}$ $^\circ$C$^{-1}$, and viscosity in the upper 100 km of mantle of $10^{19}$ Pa s. This density contrast is maintained in the majority of strip-heated experiments, and spreading rate is the primary variable.

Results

In experiments without plate spreading, the temporal pattern of the purely buoyantly driven flow is similar to previous plume initiation laboratory studies [e.g., Whitehead et al., 1984; Olson et al., 1988]: a series of plumes, or Rayleigh-Taylor instabilities, develop when the growing thermal boundary layer goes unstable. The plumes are made up of large leading plume heads with narrow trailing conduits, as is expected for a temperature-dependent viscous fluid. The boundary layer produces five plumes spaced at intervals ranging between 10 and 16.5 cm, or roughly 1.5 times the fluid thickness. The plume heads rise through the fluid, with ascent rates ranging from 0.9 to 1.3 cm/min, and flatten and spread laterally as they impact the fluid surface. At this point the plume heads are no longer visible in shadowgraph; however, focused convective heat transfer continues through the established trailing conduits which remain connected to the basal thermal boundary layer.

The large-scale shear flow due to plate spreading dramatically modifies the pattern of convection in the fluid beneath the spreading axis. The results from the line source heating experiments show a relationship between spreading rate, upwelling morphology and evolution, and axial temperature variability and allow the definition of 2-D and 3-D upwelling regimes. The relationship between plate velocity and both axial temperature variability and plume morphology is illustrated by the sequence of shadowgraph images in Figure 3 taken at three times: just after plumes lift off from the thermal boundary layer (Figure 3a), when the plumes begin to impact the surface (Figure 3b), and just after impact (Figure 3c). Plate velocity in this experiment is 0.5 cm/min, corresponding to a mantle half-spreading rate of 3 cm/yr. The upwelling flow patterns in this experiment represent a transitional case, exhibiting both 2-D (right side of $R_d$ views) and 3-D styles of upwelling (e.g., left and center).

Figure 4 shows $R_d$ shadowgraph images for experiments with a range of spreading velocities at roughly the same evolutionary stage as the experiment in Figure 3b. At slower spreading rates (0.26 cm/min), plumes form and impact the ridge as discrete upwelling centers (Figure 4a). However, at faster half spreading rates (0.73 cm/min) thermal plumes rising beneath the ridge axis are diffused in the along-axis direction, and near-surface flow is essentially 2-D (Figure 4b). The initial upwelling in these cases resembles a sheet with bumps along the leading edge providing a small and short-lived degree of axial variability. The upwelling is completely sheet-like at the fastest rate (1.46 cm/min; Figure 4c).

This upwelling flow is quantified by monitoring temperature variations at the fluid surface along the ridge axis (Figures 5-7). Both laboratory and numerical experiments show a similar evolution from an initial transient associated with the plume heads, followed by a rapid decay to reduced temperature variability due to the trailing conduits embedded in the plate-driven upwelling (Figure 5). The onset and duration of the transient depends on magnitude of buoyancy forcing. In the low-buoyancy numerical cases (Figure 5c) the transient lasts roughly 900 s. Doubling the buoyancy (Figure 5b) roughly doubles the duration of the transient to 1800 s (4.5 m.y. when scaled to mantle values at the same $Pe$). Axial variability in the transient and long-term regimes is strongly related to the relative magnitudes of the plate- and buoyancy-driven flow. Maximum axial temperature variations during the transient are

Figure 3. Spreading axis (right) parallel and (left) perpendicular shadowgraph images from a strip heating experiment showing the form of three dimensionality during the initial transient pulse of upwelling. Views of flow for an experiment with moderate buoyancy input rate and intermediate spreading rate for three times: (a) plume initiation, (b) plumes impacting surface, and (c) 1800 s after plumes hit surface. The bright, scalloped line (e.g., in axis parallel views) marks the leading edges of hot, buoyant upwellings.
from 1 to 10°C (roughly 3-30% of the total temperature drop across the fluid layer ΔT). The wavelength of the upwelling is roughly 1-1.5 times the fluid thickness (Figure 6).

The maximum range in temperature can be used to define 2-D and 3-D upwelling regimes as a function of plate velocity for both the transient and long-term upwelling stages of these experiments (Figure 7). The transition from 2-D to 3-D style upwelling beneath the ridge axis occurs near 0.6 cm/min. This corresponds to a scaled mantle half spreading rate of roughly 4 cm/yr, which is in agreement with observational data and previous numerical experiments with local buoyancy sources (Parmentier and Phipps Morgan, 1990). Variations are significantly reduced during the period following the transient and are also a function of Ra (Figure 7b). For low Ra, essentially no variability is recorded using either laboratory or numerical methods at high spreading rates. For slow spreading rates the flow is 3-D for all Ra.

**Discussion**

The strip-heated experiments demonstrate that three-dimensional upwelling on the scale of ridge segmentation (~100 km) can be generated from local 2-D buoyancy sources. The strength of the three dimensionality depends on both the Ra of the convecting system and the spreading rate, in agreement with previous theoretical work [e.g., Parmentier and Phipps Morgan, 1990]. Variable fluid viscosity tends to sharpen the transition between 2-D and 3-D upwelling regimes.

The magnitude of temperature variations recorded in the laboratory experiments can be related to an equivalent mantle temperature variation, ΔT_M, through the Rayleigh number:

\[ ΔT_M = \Delta T_L \frac{PL}{PM} α_L \frac{α_M}{α_M} \frac{κ_L}{κ_M} \left( \frac{D_L}{D_M} \right)^{\frac{3}{2}} \]

The subscripts L and M denote values for laboratory and mantle, respectively. For the strip-heated experiments, the important viscosity is that of the ambient fluid averaged over the duration of the experiment, about 40 Pa s. Since these experiments are meant to investigate buoyant flow in the immediate vicinity of partial melting region beneath the ridge, appropriate values for D_M and μ_M are 100 km and 10^{19} Pa s. Assuming these values, each 1°C variation in the models corresponds to a mantle temperature variation of 22°C. Therefore temperature variations in the 0.26 cm/min laboratory experiment correspond to ~200°C during the transient and 33°C in steady state. These temperature variations also illustrate the strong dependence of the flow on Ra. For a spreading rate of 0.2 cm/min, the steady state scaled mantle temperature variations decrease from 70°C, for Ra=1x10^5, to 0.75°C (essentially a 2-D, sheet-like flow), for Ra=0.5x10^5.

Temperature in the upwelling mantle beneath spreading centers will be buffered by the loss of latent heat during partial melting. However, variations in temperature will be converted into variations in the degree of melting. A positive temperature anomaly of 20-40°C, or roughly the predicted range in steady state values for a linear buoyancy source and half-spreading rates of 4 cm/yr, would result in an increase in the depth of initial melting of ~6-12 km and in the extent of melting of about 2.0 4.0%, assuming a solidus slope of 4°C/km and 0.16% melting per 1°C above the solidus [Langmuir et al., 1993]. Additional variability is expected in
the amount of retained melt. All of these variations in upwelling and melting can translate into variations in crustal thickness, which is the main contributor to mantle Bouguer gravity anomalies [Tolstoy et al., 1993; Sparks et al., 1993].

Most theoretical models of permeable networks at either the grain scale or at a larger vein scale show a strong dependence of the permeability on the fraction of fluid present. As a result, faster upwelling and higher melting rates lead to greater amounts of retained melt and therefore greater melt buoyancy. This feedback, which cannot be simulated in the laboratory by thermal buoyancy, can drive significant changes to the buoyant flow when the melt fraction is more than about 1% [Scott, 1992; Su and Buck, 1993]. On the basis of numerical studies [Iha et al., 1994a; Sparks et al., 1995], this additional buoyancy source should enhance three-dimensionality at lower spreading rates. Because it is uncertain whether in situ melt fraction ever exceeds 1% in the mantle, it is not clear whether or not this feedback effect will be important.

Figure 5. Plots of maximum axial temperature difference versus time for a range in plate velocities shown by the labels on each curve. Cases include (a) 3-D variable viscosity laboratory experiments and (b) high Ra and (c) low Ra 3-D, constant viscosity numerical experiments. Plots show the transient pulses associated with plume heads which quickly decay to lower levels of axial temperature variation. Both transient and longer term variability depends on plate velocity. Temperature in Figures 5b and 5c is measured at 0.06 of the box depth. Differences in the time to peak variability in Figure 5a are attributed to minor fluctuations in tank initial conditions as well as choices for start time.
Figure 6. Plots of axial temperature perturbation versus distance for strip heating experiments at a range of values in scaled plate velocities. The plume source buoyancy contrast is the same for each experiment. Data are collected at nine RTDs spaced at 5-cm intervals. The temperatures are taken 3000 s into each experiment. Axial variability decreases with increasing spreading rate.

Figure 7. Maximum amplitude of axial temperature variations versus spreading rate from the laboratory and numerical strip heated experiments. (a) Maximum amplitude measured during the initial transient pulse of an experiment or (b) during the longer-term, posttransient stage of the experiments. Two- and three-dimensional upwelling regimes are apparent in the laboratory experiments, with a transition occurring near 0.6 cm/min (corresponding to 4 cm/yr in the analog mantle system).

There is good agreement in the flow patterns and the degree of variability measured at the spreading axis in the numerical and laboratory experiments. This indicates that temperature-dependent rheology does not play an important role in determining the basic form of the convection in these particular laboratory experiments. The major qualitative difference between the lab experiments and the numerical simulations is the relatively sharper transition between the 2-D and 3-D regimes seen in the laboratory results. The increase in along-axis temperature variations with decreasing spreading rate is greater below a transition plate speed of about 0.6 cm/min. While there is a similar increase in axial variations in the numerical experiments, there is no well-defined transition. As in previous numerical studies [Sparks and Parmentier, 1993], the effect of thermal convection on upwelling near the ridge is a smoothly increasing function of spreading rate. Compositional and melt buoyancies produce a sharp transition between 3-D flow at slow spreading rates and 2-D flow at fast spreading rates [Parmentier and Phipps Morgan, 1990, Jha et al., 1994a], but the addition of thermal buoyancy blur or eliminates that transition [Jha et al., 1994b]. These results indicate that temperature-dependent rheology tends to create a transition spreading rate for thermal buoyancy. Such a transition is in better agreement with observations of along-axis gravity gradients (Figure 1 and Wang and Cochran [1993]) and ridge morphology and chemistry [Phipps Morgan et al., 1994] which indicate that there is a fundamental difference between ridges spreading faster and slower than about 4 cm/yr, half rate.

While an outcome of this work is that plate spreading rate modulates the magnitude, morphology, and evolution of upwelling beneath ridges, an important point is that the presence of 3-D upwellings is not simply a function of spreading rate. A number of mechanisms exist for producing 3-D flow even at fast ridges. These experiments show that transient pulses of upwelling from a shallow, subaxial buoyancy source are much more likely to be three-dimensional, even for fast spreading rates. This implies that fast spreading centers with steady, two-dimensional mantle upwelling could experience periods of three dimensionality due to a change in the relative amounts of buoyancy production or plate-driven flow (e.g., from a change in spreading rate or thermal or chemical heterogeneities in the upwelling mantle). One potential source of imposed heterogeneity results from the natural interaction between upper mantle scale buoyant flow and plate-scale flow which can cause the advection of either thermal heterogeneity (e.g., mantle plumes) or chemical heterogeneity (e.g., mantle plumes) into the melt region beneath ridges. The next section describes a set of experiments that addresses the interaction of such larger-scale convection with the shallow mantle under the ridge.
Base-Heated Experiments

Experimental Design

In this set of experiments the tank is scaled to the entire upper mantle, and the entire base is heated uniformly. The fluid layer in these experiments is 1 cm thick, so from (2), a mylar speed of 1 cm/min corresponds to a plate velocity of 1.4 cm/yr. We explored a range of both Pe and Ra (Table 3). The corresponding plate velocities modeled range from very slow (0.3 cm/yr) to superfast (>8 cm/yr) half spreading rates. The range in Ra (0.5x10^5 to 2.25x10^6) is reasonable for characterizing convection within the upper mantle, assuming an average dynamic viscosity of 10^{20} to 10^{21} Pa s and a potential temperature drop in the convecting part of the upper mantle of 500°C.

To achieve the desired Ra numbers in these experiments, we both heated the bottom of the tank and cooled the surface to produce lower and upper thermal boundary layers. As in the previous experiments, the growing thermal boundary layers go unstable into a number of plumes which trigger a series of upwellings and downwellings. If the mylar drive is not turned on, the convection evolves into a pattern, as viewed from above, of hexagonal convection cells, with central upwelling plumes surrounded by downwelling sheets. In the experiments listed in Table 3, the mylar drive was turned on when the basal and surface temperatures were approximately halfway toward the desired steady state values, prior to the development of any convective patterns, and was left on for the duration of the experiment. The experimental parameters which are varied are the basal buoyancy flux, the surface temperature (e.g., upper boundary layer structure), mylar spreading rates, and, to a lesser extent, the mean fluid temperature and viscosity.

Results

The flow in these experiments is a superposition of boundary layer instabilities (BLIs), which form on the hot bottom of the tank, and the overall plate-driven circulation, which can sweep these BLIs toward the axis. As in the strip-heated experiments, 2-D and 3-D dynamical regimes (numbered I and II, respectively, in Figure 8) are defined based on the relative magnitudes Ra and Pe. However, three subregimes (II.1-II.3) are also defined for the 3-D flow with respect to the signal at the ridge axis (Figure 8). In regime I, at the lowest Ra values (<10^5) and highest spreading rates (>6 cm/min), 3-D flow is suppressed. Plate-driven flow dominates over buoyant flow, and no BLIs form. In the opposite end-member regime (II.3), where plate flow is weak and plumes dominate, the flow forms a classical hexagonal convective pattern, characterized by plume-like upwellings and sheet-like downwellings [Ogawa et al., 1991]. Plate-driven flow is sufficiently weak in these cases that sampling of plumes at the ridge is determined by the hexagonal arrangement of upwellings, not the sweeping of plumes to the ridge. The two intermediate 3-D regimes, II.1 and II.2, are more interesting with regard to plume-plate interactions and the resultant axial signal. In the latter case, buoyant flow is largely dominant such that BLIs form discreet plumes, characterized by a large leading head and a trailing conduit. The plume heads surface off axis. At “midplate” positions. The trailing conduits may be swept into the large central upwelling where they serve as a source of focused subaxial upwelling and mild along-axis temperature variations (Figures 9d and 9e). It is interesting to note that the plume tails which feed the midplate “swell” remain connected to the basal thermal boundary layer even as the deep conduit is swept into the large central upwelling. A portion of this original

Table 3. Parameters for Base-Heated Experiments

<table>
<thead>
<tr>
<th>Experiment</th>
<th>Mylar Speed Up, cm/min</th>
<th>Peclet Number Pe</th>
<th>Scaled Plate Velocity, cm/yr</th>
<th>ΔT, °C</th>
<th>Viscosity at Mean Temperature, Pa s</th>
<th>Rayleigh Number Ra, x10^6</th>
<th>Mean σ², x10^3</th>
<th>Axial Temperature Variance, °C</th>
<th>Average Number of Axial Upwellings</th>
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<td>1.5</td>
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<td>0.9</td>
<td>22</td>
<td>32</td>
<td>2.25</td>
<td>2.2</td>
<td>5</td>
<td>-</td>
</tr>
<tr>
<td>4</td>
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<td>820</td>
<td>3.5</td>
<td>22</td>
<td>32</td>
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<td>14</td>
<td>8</td>
<td>4.2</td>
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<td>32</td>
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<td>9</td>
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<td>5.8</td>
<td>1640</td>
<td>7.0</td>
<td>18</td>
<td>41</td>
<td>1.4 (3.4)</td>
<td>5.5</td>
<td>6</td>
<td>2</td>
</tr>
<tr>
<td>7</td>
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<td>11</td>
<td>72</td>
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<td>0</td>
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<td>760</td>
<td>3.3</td>
<td>11</td>
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<td>1.4</td>
<td>3.5</td>
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</table>

a ΔT is defined as inside basal temperature minus surface temperature.

b Values are determined using viscosity at the mean fluid temperature. Values in parentheses are determined using viscosity at mean basal boundary layer temperature.

Values highlight the result that increasing plate velocity, while holding Ra fixed, leads to fewer upwelling centers which tend to be more stationary than those at slower plate rates.
Figure 8. Regime plot for the laboratory experiments (numbered 2-11) with full basal heating showing the range in Raleigh numbers ($Ra$), governing buoyancy driven flow, and spreading velocities (actual in cm/min and scaled values in cm/yr). At intermediate to high $Ra$ 3-D subaxial upwelling was seen for all spreading rates. At lower $Ra$ a spreading rate dependence in upwelling style is observed, with strong, weak, and no plumes (and plume conduits) observed for cases of 1.0, 3.5, and 6.2 cm/yr scaled spreading rates, respectively. Positions of East Pacific Rise (EPR) and Mid-Atlantic Ridge (MAR) regimes are plotted for two values of upper mantle viscosity ($\mu$ in Pa s). Upwelling regime plots I (2-D) and II (3-D) are also plotted. Regime II is subdivided based on the signal sampled at the spreading axis: II.1, plate flow dominates plume upwellings, plume heads and tails swept to the spreading axis, moderate axial variability; II.2, plume heads rise off-axis and tails are swept to the ridge providing strong source of axial variability; and II.3, upwellings dominate flow, rise off-axis.

plume tail is advected toward the surface and eventually draped along the base of the plate. In this horizontal orientation the conduit connects the off-axis “swell” with the spreading axis at a point where the captured plume tail feeds the ridge. An intermediate case between 2-D and 3-D flow (II.1) is illustrated in Figures 9a-9c. Here the plate strength is sufficient to sweep BLs to the central upwelling as plumes are being established. In these cases, plume heads surface at the ridge, much like the transient instabilities in the prior set of experiments, producing very strong axial variations. Centers of focused

Figure 9. Shadowgraph images taken in the ridge-perpendicular orientation showing two distinct dynamical interaction regimes for plumes and plates at different times. Figures 9a-9c are for experiment 6, regime II.1. For cases of intermediate $Ra$ and fast spreading, the BLs (nascent plume heads) are swept into the large central upwelling and surface at the ridge axis. Conduits remain embedded within the main upwelling as regions of focused flow and heat transfer to the ridge. Figures 9d-9f are for experiment 4, regime II.2. For cases of higher $Ra$ and intermediate, or weaker, plate spreading, the plume head rises off-axis and surfaces at a midplate position. The ridge does not sample the plume head, but the trailing conduit gets swept into the upwelling. In this case, once the conduits are entrained, focused, subaxial upwelling persists for more than an overturn time, although conduit fluxes may vary in time.
3-D upwelling tend to exhibit greater fixity, both spatially and temporally, than in regime II.2.

A similar diversity of plume-plate interactions is seen in a set of constant viscosity numerical experiments which are run with similar physical parameters to the base-heated laboratory experiments. At intermediate spreading rates, axial variability is due to thermal plumes that form off-axis and are swept toward the ridge. At higher plate velocities, evolution of buoyant instabilities are suppressed, and no axial variability is observed. Figure 10 shows the 3-D temperature structure from a numerical experiment for a regime II.2-type flow. BLIs form and erupt into plumes which are swept into the upwelling beneath the axis. The variation in the width of the shallow hot region near the axis shows variability in heat transfer within the main upwelling due to a previous set of plume conduits that were swept into the central upwelling. Note that the positions of the previous plume signals are offset in the along-axis directions from the plumes that are in the process of forming.

Three interaction regimes (I-II.2) are evident on contour plots of laboratory temperature as a function of distance along the spreading axis and time in Plate 1. In these experiments, each minute scales to a time of 2.5 Myr in the corresponding mantle system. The greatest level of axial variability for this series of laboratory experiments occurs for high Ra (7.25x10^5), intermediate Pe (8210) conditions (Plate 1a, type II.2) where the number of upwellings at any one time varies between 3 and 5 with spacing between 10 and 30 cm or 0.6 and 2 times the fluid thickness. Moreover, upwelling centers vary significantly in both spatial and temporal character (e.g., plume pulses and drifting). In the intermediate regime (Plate 1b, type II.1) increased Pe (1600) and decreased Ra (1.4x10^5) results in increased spacing between upwellings. It is interesting to note that the rightmost of the two long-lived upwelling centers in Plate 1b remains relatively fixed spatially but pulsates with time. The other center of focused upwelling wanders back and forth along the ridge axis with time. Holding spreading rate fixed, and further reducing Ra by a factor of 3 leads to the plate flow dominated regime (Plate 1c, type I) where no plumes and no axial variability are recorded.

The numerical experiments allow the determination of along-axis variations in terms of the flow velocities. Plates 1d-1f show the vertically averaged upwelling velocity in the plane of the axis during the course of three experiments. The degree of axial variability clearly drops as plate velocity is systematically increased. At slow spreading rates, bursts of three-dimensional upwelling occur as waves of upwellings strike the axis at time intervals of about 900 s. In some cases, the off-axis plumes form in a roughly hexagonal pattern, so each successive wave is offset from the other. At the fastest spreading rate (Plate 1f), the flow remains two-dimensional, since perturbations in the hot boundary layer do not have enough time to grow. The along-axis wavelength in these experiments is roughly twice the layer depth, although it is somewhat variable in the laboratory experiments.

**Discussion**

These experiments indicate that interaction between plates and convection on the scale of the upper mantle can generate pulses of along-axis variability at a spreading center with wavelengths of 600-1000 km and on timescales of 20-40 Myr. The possible effect of such heterogeneity is summarized in Figure 8. The range of spreading rates along the East Pacific Rise (EPR) and Mid-Atlantic Ridge (MAR) is overlain on the plot of convective regimes found in the laboratory experiments. For high Ra a slow/intermediate spreading

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**Figure 10.** Three-dimensional plot showing the temperature isosurface at 0.64 of the maximum temperature in the box, for a numerical experiment with a mylar speed of 1.5 cm/min. The isosurface shows plumes forming on the bottom boundary layer and being swept into upwelling in the plane of the spreading axis. The remnants of a previous wave of plumes is shown by the large variations in the lateral width of the isosurface just below the axis.
Plate 1. Contour plots of axial temperature for laboratory experiments (Plates 1a-1c) and upwelling for numerical experiments (Plates 1d-1f) along the spreading axis versus time (minutes) showing the differences in the spatial and temporal evolution of focused upwelling as a function of spreading rate. (a) For a high Ra, intermediate spreading case, a large number of focused upwelling centers are recorded, which vary in both space and time. (b) When Ra number is decreased and Pe is increased, the number of upwellings drops to two, one of which remains fixed spatially but pulsates in time while the other wanders along the ridge axis. (c) The plate dominated end-member, with low Ra experiments and fast spreading, shows no focused upwelling or related axial variability. Similar patterns are seen in the vertically averaged upwelling velocity calculated during numerical experiments conducted at a single Ra and spreading rates of (d) 1 cm/min, (e) 2.3 cm/min, and (f) 5 cm/min.

Center sees moderate 3-D upwelling due to the tails of off-axis plumes being swept toward the axis, while a fast spreading center sees strong 3-D input from focused upwellings forming beneath the axis. For lower Ra the slow spreading center is in the regime of strong along-axis three dimensionality, while convection beneath the faster spreading center is two-dimensional. Since the latter condition is in better agreement with observations of ridges, this suggests that upper mantle Ra numbers are relatively low, and the form of convection, at least beneath intermediate to fast spreading ridges, is not dominated by well-developed detached plumes.

Temperature-dependent viscosity. The temperature dependence of viscosity may be the cause of two differences between the laboratory and numerical results in this set of experiments. First, the upwellings in the laboratory experiments are more focused and occur at roughly twice the spacing of those in the numerical experiments (Plate 1). This is shown by relative magnitudes of the variance in axial temperature, $\sigma^2$ (Figure 11). Dimensionless variance, normalized by the temperature drop across the box ($\Delta T$), is defined as a function of time during the run by

$$\sigma^2(t) = \frac{1}{N-1} \sum_{i=1}^{N} \left[ T(x_i, t) - \overline{T}(t) \right]^2 / \Delta T^2$$

where $x$, $N$, $T$, and $\overline{T}$ are distance along the spreading axis, number of data points along-axis, axial temperature, and mean axial temperature, respectively. For the numerical runs,
Figure 11. Plot of the time-averaged mean in axial temperature variance versus the ratio in strength of plate-driven \((Pe)\) to buoyancy-driven \((Ra)\) flow. The \(Pe/Ra\) ratio used here is defined with values for \(Ra\) scaled by \(10^{-3}\). Plots are shown for both numerical and laboratory data. Both plots show reduced variance for low \(Pe/Ra\) which does not reflect reduced three-dimensionality in buoyant flow. For higher values of \(Pe/Ra\), the general trend of decreasing variance with increasing \(Pe/Ra\) (or plate spreading if \(Ra\) is assumed fixed) recorded in both data sets is extremely similar to observed trends in MBA gradient data (Figure 1). Upwelling regime boundaries, defined in Figure 8, are also plotted.

Temperatures are measured at 0.031 of the box depth. In Figure 11, each run is defined by the ratio of the strengths of plate - \((Pe)\) to buoyancy - \((Ra)\) driven flow, and the variance plotted is time-averaged over the duration of the run. While both numerical and laboratory data sets show similar trends, the magnitudes in the laboratory cases are uniformly greater. In the constant viscosity numerical experiments the variations in upwelling velocity and temperature are roughly sinusoidal, whereas in the laboratory experiments with strongly temperature-dependent viscosity, the focused upwellings are narrower and stronger.

The second rheological difference is the existence of a "memory" in the temperature-dependent fluid. The initial group of plumes that rise beneath the axis establishes upwelling paths that are warmer and less viscous and therefore offer far less resistance than the surrounding cold fluid. As a result, it is easier for new BLIs to rise along the established pathways, resulting in surface expressions that remain relatively fixed, or migrate only slowly with time (Plate 1c). In contrast, the BLIs in the uniform viscosity numerical experiments do not exhibit the same degree of fixity, and in some cases each successive wave of BLIs arrives at the axis spatially out of phase with the previous pulse (Plate 1e).

If upper mantle \(Ra\) (which depends on the temperature and viscosity structure) is relatively uniform beneath the ocean basins, then the variance shown in Figure 11 is only a function of \(Pe\). In that case, these experiments suggest that the thermal input from upper mantle-scale convection to the melting region beneath spreading centers may be a function of spreading rate. Moreover, the general shape of these curves is strikingly similar to the plots of MBA gradient versus spreading rate shown in Figure 1b. While these results do not constrain the importance of enhanced along-axis magma transport at shallow levels, they do suggest that the upwelling dynamics of the deeper mantle can account for some of the along-axis variability that is observed.

Plume-ridge interaction. An interesting result of the global-scale plume-plate interaction models which has implications for plume-ridge interaction models is recorded for cases in which plume-driven and plate-driven flow magnitudes are similar. A characteristic pattern of plume evolution is noted in these cases in which plume heads escape the background flow to form an off-axis swell while the trailing conduits, with their bases still attached to the thermal source region, are swept to the ridge axis. Conduits are tilted into a horizontal orientation and are draped along the underside of the plate, forming a connection between off-axis swell and ridge axis. This type of plume-ridge interaction is distinctly different from plume channel models (e.g., Morgan, 1971; Schilling, 1991; Kincaid et al., 1995, 1996) where plume material rises off axis and is deflected toward the ridge and more closely resembles a "hot line" model for producing plume-like chemical signals at both ridge and off-axis locations. This type of hot line model predicts that "fresher" plume material should be sampled at the ridge with an age progression toward the swell, which is opposite from the plume channel model.

Conclusions

The conclusions of the combined laboratory and numerical experiments described in this paper can be summarized as follows:

1. While these experiments do not constrain the importance of shallow, along-axis magma transport processes, they do show that variability in dynamics of mantle upwelling is strongly modulated by the plate-driven flow and that the natural interaction between plate- and buoyancy-driven flow can account for the observed range in axial (thermal) variability for spreading ridges.

2. Local and upper mantle-scale buoyancy forcing contribute distinct length scales of variability to the ridge.

The pattern of segment-scale (100-150 km) convection beneath a spreading center is a strong function of both the spreading rate and the \(Ra\) of the buoyant flow. Three dimensionality is also strongly enhanced during transient pulses of upwelling.

Convection on the scale of the upper mantle can contribute significant long-wavelength (600-1000 km) spatial and temporal variability in upwelling rates and temperatures at spreading centers. While the amount of variability is sensitive to spreading rate, the direction of that sensitivity depends on \(Ra\).

3. Experiments suggest two mechanisms for maintaining significant, long-term variability in the dynamic upwelling beneath even fast ridges. These include (1) time-dependent pulsing of the buoyancy source in the shallow subaxial mantle to produce upwelling transients and (2) the advection of thermal heterogeneities (e.g., plumes) into the melt region due the plate-scale "mantle wind."

4. We find that for cases where plume and plate strengths are roughly equivalent, plume heads surface off axis and the conduit basalt are brought up at ridges in such a way that the conduits themselves form linear connections between the
ridge and the off-axis plume heads. This manner of plume conduit surfacing provides an alternative model for plume-ridge interaction to existing channel flow models.

5. Experiments predict that the degree of axial variability in mantle upwelling beneath the ridge may only be the same at fast and slow spreading ridges if upper mantle \( R_a \) is dependent on spreading rate. However, if one assumes that \( R_a \) does not vary from ocean basin to ocean basin, the conclusion is that plate spreading rates control axial variability. These dynamical results and observational evidence for fast versus slow spreading centers are best reconciled if \( R_a \) for the upper mantle is low (< 1x10^6).

6. The general pattern of flow observed in the laboratory and numerical experiments are qualitatively similar. Comparisons between variable viscosity laboratory and constant viscosity numerical experiments indicate that temperature-dependent viscosity may stabilize the position of focused upwelling centers and cause a sharper transition between 2-D and 3-D upwelling patterns relative to flows modeled in previous uniform-viscosity numerical studies.

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