Plume-transform interactions at ultra-slow spreading ridges: Implications for the Southwest Indian Ridge

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[1] We explore a potentially important variable in controlling ridge-hot spot interaction, the effect of transform offsets in limiting along-axis flow of plume material. We focus on the Southwest Indian Ridge (SWIR), where the “transform damming” effect is likely to be pronounced because of both long offset lengths and large contrasts in lithospheric thickness across the transform faults due to the ultra-slow spreading rate. We investigate the degree to which transform faults affect axial asthenospheric flow by performing a series of three-dimensional (3-D) numerical experiments with simplified channel-flow geometry and extrapolating their results to the SWIR. 3-D mantle viscosity structure for a ridge-transform-ridge system is determined based on temperature- and pressure-dependent viscosity laws. We consider six transform lengths, spanning 0 to 250 km in increments of 50 km. We then calculate the 3-D viscous flow in response to an along-axis pressure gradient corresponding to a ridge-centered hot spot. Modeling results predict that transform faults affect along-axis asthenospheric flow in two important ways. First, transforms reduce along-axis flux. The longer the transform offset, the greater the reduction in across-transform flux relative to the zero-offset case. Second, transforms deflect shallow asthenospheric along-axis flow. The predicted transform damming effect is most pronounced for a viscosity structure that is strictly pressure- and temperature-dependent. Flux reduction effects could be less significant for viscosity laws that additionally consider dehydration, melting, and change in deformation mechanism. This model predicts that the waist width of an on-axis plume is dependent not only on such previously explored factors as buoyancy and spreading rate, but also on the geometry of ridge segmentation. Along the SWIR, axial flow driven by the Marion plume is likely curtailed by the long-offset Andrew Bain and Discovery II fracture zones, severely limiting its lateral extent.

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1. Introduction

[2] A significant portion of the mid-ocean ridge system is influenced by mantle plumes [Schilling, 1991]. Near- or on-ridge plumes thicken oceanic crust, alter mid-ocean ridge geochemistry, and impart a strong signal to seafloor bathymetry, gravity, and geoid. Numerous investigations have suggested that transform faults influence the distribution of plume material, including the Charlie Gibbs FZ for the Iceland hot spot [Vogt and Johnson, 1975] and the Agulhas FZ for the Discovery hot spot [Douglass et al., 1995]. The influence of ridge offsets on plume dispersal was first explored analytically by Vogt and Johnson [1975] and Vogt [1976]. However, the interaction between plume-driven flow and transform faults has only recently been addressed by numerical geodynamic models [e.g., Sleep, 1996]. While most of the recent plume-ridge numerical experiments generally consider straight ridges, the modeling study of Yale and Phipps Morgan [1998] predicts strong focusing of Kerguelen plume flow toward a segment of the Southeast Indian Ridge that is offset in the direction of Kerguelen, suggesting that ridge segmentation is important in plume-ridge interaction.

[3] Segmentation is a first-order feature of mid-ocean ridges. Ridges of all spreading rates are segmented, although segmentation patterns and the associated mantle dynamics vary with spreading rate [e.g., Macdonald, 1982; Schouten et al., 1985; Lin and Phipps Morgan, 1992]. Figure 1 shows the high degree of segmentation along the central portion of the ultra-slow spreading Southwest Indian Ridge (SWIR) near the Marion hot spot. Along the SWIR between 25°–45°E, the Andrew Bain and other transform faults cumulatively offset the SWIR by ~1500 km. The transform effect on along-axis plume flow is likely to be pronounced along the SWIR because of both particularly long transform offset lengths and large lithospheric thickness contrasts across transforms due to the ultra-slow spreading rate.

[4] In this study we use a hybrid finite difference/finite element 3-D numerical model to quantify the role of transform offsets in limiting the along-axis asthenospheric flow driven by a ridge-centered plume. We focus on the SWIR as an end-member case where such a transform effect is likely to be pronounced. We quantify the transform damming effect for six offset lengths, which are relevant to the range of transform lengths observed along the SWIR. We then examine the sensitivity of the transform effect to mantle viscosity structure and discuss the implications of model results for plume-ridge interactions.

2. Numerical Method

[5] We model mantle flow along a segmented ridge driven by a plume-related pressure gradient. The model box is comprised of two segments, each of length \( L_s \), and an intervening transform of length \( L_t \) (Figure 2; see Table 1 for definition of model variables). We examine values of \( L_t \) ranging from 0 to 250 km in increments of 50 km, overlapping a wide range of offset lengths observed along the SWIR. Computational limitations and model resolution considerations prohibit exploration of offsets significantly greater than 250 km, for which only qualitative results are inferred from extrapolation. The model box is 750 km in the across-axis direction (X), 500 km in the along-axis direction (Y), and 660 km in depth (Z) to coincide with the 660 km mantle discontinuity. A half-spreading rate, \( U \), of 0.75 cm/yr was chosen to reflect the opening rate of the ultra-slow spreading SWIR near the Marion plume.

[6] Numerical calculations are performed in two steps: (1) We calculate three-dimensional (3-D)
viscosity structure for a given $L_t$ using the approach of Shen and Forsyth [1992]; and (2) we use this 3-D viscosity structure as input to a fluid dynamical calculation, in which a pressure gradient $\frac{dP}{dY}$, which simulates an on-axis plume, drives flow along-ridge in the Y-direction. This two-step approach significantly simplifies modeling complexity and enables us to isolate the purely geometrical effects of viscosity structure on the along-axis mantle flow. Our solutions are for the instantaneous interaction between along-axis flow and a transform offset and thus are not designed to investigate the time-dependent evolution of the plume-ridge system. Feedbacks between temperature, viscosity, and velocity fields, which were neglected in the present models, are expected to change the results quantitatively. For example, flow away from a thermal plume will advect high-temperature material, reducing viscosities and thinning the lithosphere. As the anomalously warm plume material approaches a transform, it is likely that relatively high viscosities near the offset will be reduced, and the transform will experience a form of “thermal erosion” [Vogt and Johnson, 1975] that was not considered in the present model. In this way, the transform would present less of a barrier to along-axis flow than in the present study.

[7] Pressure- and temperature-dependent viscosity is calculated using a hybrid, iterative finite element/finite difference approach [Shen and Forsyth, 1992]. First, the velocity field for incompressible mantle flow driven by passive plate separation is determined using a finite element code with successive overrelaxation. Then, upwind finite differences are used to solve for the mantle temperature field in the model box, assuming $T_{Z=200 \text{ km}} = 1350^\circ \text{C}$ and $T_{Z=0 \text{ km}} = 0^\circ \text{C}$. Viscosity is calculated at each node according to pressure and temperature. Finally, velocity, temperature, and viscosity calculations are iterated until a stable steady state solution is reached. The Shen and Forsyth [1992] code has been benchmarked against the analytical flow solutions of Reid and Jackson [1981] (Y. Shen, personal communication, 2003). The average upwelling rate predicted by Shen and Forsyth [1992] is 97% of the theoretical value $(2/\pi)*U$, with the most significant deviation from the analytical solutions occurring within one grid node of the ridge axis.

[8] The governing equation for viscosity is given by

$$\eta = A \sigma^{\alpha - \beta} \exp\left(\frac{E + PV}{RT}\right)$$

where $A$ is a pre-exponential constant, $\alpha$ is the stress exponent, $E$ is activation energy, $P$ is pressure, $V$ is volume, and $T$ is temperature.
is activation volume, $R$ is the universal gas constant, and $T$ is temperature (Table 1). We set $n = 1$ (Newtonian fluid) for all calculations, yielding a reference minimum viscosity $\eta_{\text{min}} = 10^{19}$ Pa s. Viscosity varies by $\sim 5$ orders of magnitude over the model space. Low viscosity values are predicted for a broad depth range with viscosity minima occurring at a depth of approximately 70–75 km (Figure 3a). In plan view, the region of lowest viscosity forms a continuous meandering band along-axis (Figure 3b). Figure 3c shows that the region of lowest viscosity is predicted to take a roughly triangular shape in across-axis section. Further details of the viscosity modeling calculations can be found in Shen and Forsyth [1992].

The 3-D viscosity field for each $L_t$ is then used as input to a finite element fluid dynamical code (ADINA [Bathe, 1996]) that solves the equation of continuity for an incompressible fluid

$$\nabla \cdot \mathbf{v} = 0$$  \hspace{1cm} (2)

and the equation of momentum balance

$$\nabla P = \nabla \cdot (\eta \nabla \mathbf{v}) + \rho g$$  \hspace{1cm} (3)

with no heat transfer and no buoyancy-driven flow, where $\mathbf{v}$ is velocity vector, $P$ is fluid pressure, $\eta$ is viscosity, $\rho$ is mantle density, and $g$ is the acceleration of gravity. The driving force for fluid flow is an along-axis pressure gradient $dP/dY$ created by imposing pressure with a Gaussian spatial distribution at $Y = L_s$ and zero pressure at $Y = -L_s$ (Figure 2). The pressure at $Y = L_s$ is...
assumed to be maximum at the ridge axis with an across-axis distribution given by

\[ P(X) = P_o \exp \left( -0.5 \frac{X}{X_o} \right)^2 \],

where \( P_o \) is the maximum pressure and \( X_o \) is the Gaussian distribution standard deviation width. Other boundary conditions are zero slip at \( Z = 0 \) km and \( Z = Z_{\text{max}} \), \( v_z = 0 \) at \( Z = Z_{\text{max}} \), and zero normal stress at \( Z = 0 \) km and the ridge-plane-parallel surfaces of the model box. Along-axis flux \( Q \) of mantle material is defined as

\[ Q(Y) = \int v_y \, dX \, dZ \]

where \( v_y \) is along-axis velocity (Table 1) and the integration is taken over the entire \( X-Z \) plane at constant \( Y \).

Since the along-axis viscous flow scales with \( dP/dY \), it is important to examine in more detail the selection of \( P_o \). The selection of \( P_o \) follows two different arguments, relating to plume buoyancy and topographic loading, respectively. First, following Conder [2000], we assume that plume-induced differential pressure \( \Delta P \) scales with buoyancy force:

\[ \Delta P \sim \rho \Delta T g \Delta L, \quad \text{or} \quad \frac{\Delta P}{\Delta L} \sim \rho \alpha \Delta T g \]

where \( \rho \) is reference mantle density, \( \Delta T \) is plume thermal anomaly, \( \Delta L \) is upwelling length, and \( \alpha \) is the coefficient of thermal expansion. If we assume that \( \rho = 3300 \, \text{kg/m}^3 \), \( \alpha = 3 \times 10^{-5} \, \text{K}^{-1} \), and \( g = 9.8 \, \text{m/s}^2 \), the scaling relation of Equation (6) becomes \( \Delta P/\Delta L \sim \Delta T \). For the on-axis Iceland plume, models suggest \( \Delta T \sim 50–200 \, \text{K} \) [Ito et al., 1999], yielding \( \Delta P/\Delta L \sim 50–200 \, \text{kPa/km} \). This is probably an upper limit for plume-related along-axis pressure gradients. \( \Delta P/\Delta L \) will be smaller for off-axis plumes and plumes with smaller \( \Delta T \). Two recent studies have modeled asymmetric flow across the axis of the East Pacific Rise, assuming that mantle flow is driven by the distant Pacific Superswell [Phipps Morgan et al., 1995]. Conder et al. [2002], for example, assume \( \Delta P/\Delta L \sim 1–10 \)

Table 1. Model Parameters

<table>
<thead>
<tr>
<th>Variable</th>
<th>Meaning</th>
<th>Value</th>
<th>Units</th>
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<tr>
<td>E</td>
<td>Activation energy</td>
<td>520</td>
<td>kJ/mol</td>
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<tr>
<td>V</td>
<td>Activation volume</td>
<td>10 × 10^{-6}</td>
<td>m^3/mol</td>
</tr>
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<td>n</td>
<td>Stress exponent</td>
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</tr>
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<td>( \eta_{\text{min}} )</td>
<td>Minimum viscosity</td>
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<td>Pa s</td>
</tr>
<tr>
<td>( \rho )</td>
<td>Mantle density</td>
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<td>kg/m^3</td>
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<td>g</td>
<td>Gravitational acceleration</td>
<td>9.8</td>
<td>m/s^2</td>
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<td>g</td>
<td>Gravity vector</td>
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<td>R</td>
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<td>Thermal diffusivity</td>
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<td>( Y_{\text{max}} )</td>
<td>Box length, along-axis direction</td>
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<tr>
<td>( \Delta \text{stag} )</td>
<td>Distance of velocity stagnation point from plume origin</td>
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<td>km</td>
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<tr>
<td>( Z )</td>
<td>Depth below surface</td>
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<td>km</td>
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<tr>
<td>( Z_{\text{max}} )</td>
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<td>Transform length</td>
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<td>( U )</td>
<td>Half-spreading rate</td>
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<td>v</td>
<td>Velocity vector</td>
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<td>cm/yr</td>
</tr>
<tr>
<td>( v_x )</td>
<td>Velocity in spreading direction</td>
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<td>cm/yr</td>
</tr>
<tr>
<td>( v_y )</td>
<td>Along-axis velocity</td>
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<td>cm/yr</td>
</tr>
<tr>
<td>( v_z )</td>
<td>Vertical velocity</td>
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<td>cm/yr</td>
</tr>
<tr>
<td>Q</td>
<td>Along-axis volumetric flux</td>
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<td>km^3/yr</td>
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<tr>
<td>P</td>
<td>Pressure</td>
<td></td>
<td>Pa</td>
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<td>( P_o )</td>
<td>Maximum pressure at ( Y = L_s )</td>
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<td>MPa</td>
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<tr>
<td>( X_o )</td>
<td>Pressure Gaussian width</td>
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<td>km</td>
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kPa/km. Likewise, Toomey et al. [2002] use a pressure gradient of 2 kPa/km. If mantle flow is driven by topographic loading of a lithospheric plate by emplacement of a volcanic edifice, \( \Delta P \) should scale as

\[
\Delta P \sim \rho g \Delta h, \text{ or } \Delta P/\Delta L \sim \rho g \Delta h/\Delta L.
\]

where \( \Delta h \) is elevation of the topographic anomaly relative to the surrounding seafloor and \( \rho = 2700 \text{ kg/m}^3 \). On the basis of the plateau heights of the Galapagos, Azores, and Iceland hot spots, \( \Delta h/\Delta L \) is approximately 0.8/500, 4/1300, and 1.8/1100 km/km [Ito and Lin, 1995], respectively, yielding \( \Delta P/\Delta L \sim 42, 43, \) and 81 kPa/km. The topographic anomaly for Marion is less straightforward to define than for these other hot spots [Georgen et al., 2001], but a best estimate for \( \Delta P/\Delta L \) for Marion is approximately 30–40 kPa/km. Thus in this study we assign \( P_o = 15 \text{ MPa} \), yielding an along-axis pressure gradient \( \Delta P/\Delta L = P_o/Y_{\text{max}} = 15 \text{ MPa}/500 \text{ km} = 30 \text{ kPa/km} \). Note that the resultant along-axis plume velocity is scaled proportionally to the assumed \( \Delta P/\Delta L \).

Figure 3. (a) Viscosity-depth profiles for the top 150 km of the model space. Black lines show profiles from this study, taken at the midpoint of a ridge segment (solid line) and the midpoint of a transform with \( L_t = 150 \text{ km} \) (dashed line). All of our models in panels a–c are for \( U = 0.75 \text{ cm/yr} \). Additional profiles (colored lines) show viscosity-depth profiles from other modeling studies for reference. The profile labeled “Braun et al. 3.0 cm/yr” (blue line) shows a mid-segment profile for \( U = 3.0 \text{ cm/yr} \) when considering the effects of not only pressure and temperature but also dehydration, melting, and deformation mechanism [Braun et al., 2000]. Braun et al. [2000] assume a mantle potential temperature of 1350°C at 660 km. Note that the Braun et al. [2000] solution predicts a high-viscosity lid extending to depths of \( \sim 60 \text{ km} \), the transition from dry to damp melting. In contrast, Ito et al. [1999], who also consider the effects of dehydration on viscosity, place the boundary of wet and dry melting at approximately 110 km (cyan curve). The Ito et al. [1999] viscosity-depth profile is taken through the center of a ridge-centered plume. The maximum hot spot mantle potential temperature was assumed to be 1530°C for \( Z > 240 \text{ km} \), half-spreading rate was 0.95 cm/yr, and viscosity was \( 3.5 \times 10^{19} \text{ Pa s} \) at 200 km depth. For comparison, the Ito et al. [1999] viscosity-depth profile without dehydration is plotted in green. (b) Viscosity slice at \( Z = 42 \text{ km} \) for \( L_t = 150 \text{ km} \). Viscosity contours are in Pa s. (c) Across-axis viscosity slice at \( Y = 125 \text{ km} \).
decreases with depth may reflect a hot spot whose origin is a relatively shallow geochemical anomaly with vertical extent much less than 660 km. Alternatively, along-axis flow may be driven by upwelling and dispersal of a buoyant plume. However, such buoyancy driving forces are unlikely to yield qualitatively significantly different flow results than the pressure configuration used here. For example, Ito et al. [1996] modeled the dispersion of a buoyant Iceland plume along the Mid-Atlantic Ridge. Qualitative comparison of the along-axis velocity results from Ito et al. [1996] and this study shows that the depths and across-axis extents of the region of fast along-axis flow are quite similar, despite differences in the magnitude of the flow because of the unusually large flux of the Iceland plume. This confirms that the shallowly rooted, buoyancy-driven plumes should yield the same qualitative flow structure as the depth-independent pressure gradient considered here because in both cases nearly all flow occurs in the low-viscosity region in the upper 200 km of the model space.

[13] The computational grid for viscosity structure is $41 \times 30 \times 19$ nodes, yielding grid spacing of $18.8 \times 17.2 \times 10.5$ km. Viscosity is calculated for a box with dimensions $X_{\text{max}} \times Y_{\text{max}} \times 200$ km, and padded to a depth of 660 km by extrapolation according to pressure-depth relationships. A slightly different nodal grid of $41 \times 25 \times 21$ is used for the flow velocity calculations. This results in spatial resolution of $18.8 \times 20.8$ km in the X and Y directions, respectively, and variable spacing, from 6 to 60 km, in Z with highest resolution near the surface where vertical gradients in viscosity are greatest. Benchmark calculations for an isoviscous channel-flow problem yield numerical results that differ by <1% from analytical solutions.

3. Results

[14] The presence of a transform offset is calculated to have two general effects on along-axis flow, (1) deflection of shallow mantle flow toward the offset direction and (2) reduction in mantle flux across the transform. We discuss these two effects below.

3.1. Deflection of Shallow Mantle Flow

[15] In across-axis section, the high-velocity region forms the shape of a flattened triangle (Figure 4), following viscosity contours (Figure 3c). For the prescribed along-axis pressure gradient of 15 MPa over 500 km ($\Delta P/\Delta L = 30$ kPa/km), the maximum
Along-axis velocity, $v_{y\text{ max}}$, is $\sim 10$ cm/yr for the zero transform offset case, significantly faster than the spreading rate. These high velocities are achieved at a depth of approximately 75 km, where minima in viscosity-depth profiles are reached (Figure 3a).

[16] In plan view, the region of high along-axis velocity forms a band around the ridge axis, again following viscosity contours (Figure 5). Owing to flow in the spreading (X) and vertical (Z) directions, the along-axis flux $Q$ decreases as a function of decreasing $Y$ even in the absence of a transform fault (i.e., for $L_t = 0$ km) (Figures 5a and 5d). For the zero transform offset case, for example, the calculated flux decreases from $Q = 2.5$ km$^3$/yr at $Y = L_s$ to $Q = \leq 1$ km$^3$/yr at $Y = 0$, corresponding to $\sim 60\%$ reduction in $Q$ over a distance of $L_s$ (Figure 6). Within the zone of fastest along-axis flow at the depth of 72 km, the along-axis velocity decreases from $v_{y\text{ max}}$ at $Y = L_s$ to $0.185 \cdot v_{y\text{ max}}$ at $Y = 0$.

[17] Along-axis flow stagnates farther upstream from the transform as the transform offset increases (Figure 5). Comparison of Figures 5a and 5c shows that high values of $v_y$ are more restricted for $L_t = 150$ km than for $L_t = 0$ km. Similarly, Figure 7 shows that $v_y$ at 72 km depth for $L_t = 150$ km is reduced relative to the case of $L_t = 0$ km. Along-axis velocity for $L_t = 150$ km begins to significantly decrease, compared to $L_t = 0$ km, approximately 100 km from the transform offset.

[18] Figure 8 quantifies decreases in along-axis velocity for all $L_t$. We define $D_{stag}$ as the distance from the edge of the model box at $Y = L_s$ to where the along-axis velocity is reduced to $v_y = 0.185 \cdot v_{y\text{ max}}$ at a depth of $Z = 72$ km. For zero transform offset ($L_t = 0$ km), $D_{stag} = L_s$. Figure 8 suggests that the longer the offset, the farther upstream from the transform fault the high-velocity flow is predicted to stop. For example, $D_{stag} \sim 0.83 \cdot L_s$ for $L_t = 50$ km, but $D_{stag}$ reduces to $0.67 \cdot L_s$ for $L_t \geq 150$ km.
In addition to impeding along-axis flow, transform offsets may also deflect it in the direction of the next ridge segment. Such deflection can be seen for the case of $L_t = 50$ km (Figures 5b and 5e), where the region delineated by $v_y = 0.125$ cm/yr persists across the transform. This effect is not clearly visible for greater $L_t$, however, because of the stagnation effect described above.

3.2. Reduction in Flux Across a Transform Fault

Figure 9 shows the relationship between transform offset length and along-axis flux reduction, using normalized flux, $Q/Q_{L_t = 0}$. The presence of a transform fault is predicted to reduce $Q$ along the entire length of the ridge axis, both upstream ($Y > 0$ km) and downstream ($Y < 0$ km) of the transform fault. For example, for the case of $L_t = 250$ km (Figure 9), the calculated $Q/Q_{L_t = 0}$ at $Y = 100$ km is $\sim 0.8$, indicating that flux is already noticeably reduced upstream of the transform. This upstream flux reduction occurs because model viscosity increases gradually along-axis at all depths as the transform is approached. For large transforms, the increased viscosity due to thermal cooling around the transform fault is predicted to extend for a relatively long distance along-axis. Downstream from the transform fault, the spatial gradient in the reduction of $Q$ is even greater. As a result, the calculated along-axis material flux $Q$ is essentially zero within 100 km downstream of the transform for large $L_t$ (Figure 9). In general, the fluxes are predicted to decrease by a greater amount for larger $L_t$. For example, for $L_t = 250$ km, flux across the transform ($Y = 0$ km) is predicted to be reduced by $40\%$ relative to the no-transform case, whereas for $L_t = 50$ km, this flux reduction is only $\sim 10\%$.

4. Alternative Viscosity Models

The modeling results discussed so far are for viscosity that varies only as a function of temperature and pressure. However, viscosity may also be influenced by other factors including the presence of melt, dehydration during melting, and transitions in creep mechanism [Hirth and Kohlstedt, 1995a, 1995b, 1996; Phipps Morgan, 1997; Braun et al., 2000]. Numerical models of Braun et al. [2000] yield the following general predictions. (1) Latent
Figure 8. The effect of a transform fault on the calculated location of a velocity stagnation point at depth of $Z = 72$ km (within the zone of fastest along-axis flow). (a) $D_{stag}$ is the distance from the edge of the model box (at $Y = L_s$) to where the along-axis velocity is reduced to $v_y = 0.185 v_{y,max}$ for $L_t = 0$. For zero transform offset ($L_t = 0$ km), $D_{stag} = L_s$. (b) Normalized stagnation distance plotted as $D_{stag}/L_s$. Note that the longer the transform offset, the farther upstream from the transform is the stagnation point.

Figure 9. Normalized flux $Q/Q_{L_t=0}$ as a function of along-axis distance. Normalized flux curves are given for all six $L_t$, ranging from 0 km to 250 km. The transform offset is located at $Y = 0$ km. Note that flux at all locations decreases with increasing transform length $L_t$. Downstream from the transform fault ($Y < 0$), the reduction in the flux is even greater.
heat changes during melting affect viscosity relatively little, increasing it at the depths where dry melting occurs by less than an order of magnitude compared to strictly temperature- and pressure-dependent viscosity. (2) The retention of a small amount of melt in the mantle matrix (e.g., ~3%) also has a relatively small effect, decreasing the shallow mantle viscosity by less than an order of magnitude. (3) Dehydration during melting is suggested to increase viscosity by approximately two orders of magnitude in the shallow mantle. (4) Transition in creep mechanism results in a viscosity decrease of an order of magnitude. The predicted combined effect of all four of these processes is to cause a maximum increase in viscosity in the dry melting regime (shallow depths) of approximately an order of magnitude, and a maximum decrease in viscosity in the wet melting regime (greater depths) of approximately an order of magnitude. This effect can be seen in Figure 3a, where the Braun et al. [2000] viscosity-depth curve for 3.0 cm/yr has a pronounced step at a depth of ~60 km, approximately the depth of the transition from damp to dry melting.

To assess the potential effects of melting, dehydration, and transition in creep mechanism on along-axis velocity, we computed 3-D flow fields using the viscosity structure predicted by Braun et al. [2000] for a single ridge with U = 3.0 cm/yr, and compared the results to the flow fields predicted for a strictly pressure- and temperature-dependent viscosity structure, also with U = 3.0 cm/yr. It was necessary to use this intermediate spreading rate, rather than the ultra-slow rate used in the rest of the model runs, because Braun et al. [2000] did not model U = 0.75 cm/yr. Nevertheless, the results are qualitatively applicable to the U = 0.75 cm/yr case. Also, since Braun et al. [2000] only examined a 2D, axis-perpendicular geometry, we repeated their viscosity-depth solution along-axis to generate a 3-D viscosity structure with no transform offsets. We adjusted the minimum viscosity used by Braun et al. [2000] to match $\eta_{\text{min}} = 10^{19}$ Pa, the minimum viscosity in the present study.

Flow fields calculated using the Braun et al. viscosity structure differ from those predicted using strictly pressure- and temperature-dependent viscosity in several important ways (Figure 10). For example, while maximum $v_y$ for the strictly pressure- and temperature-dependent viscosity solution occurs at a depth less than 100 km, the maximum $v_y$ for the

![Figure 10](image-url)
equalized Braun et al. viscosity structure is found at a depth slightly less than 200 km (Figure 10a). Moreover, maximum $v_y$ for the Braun et al. solution is greater (>30 cm/yr) than that for the strictly pressure- and temperature-dependent viscosity solution (~15 cm/yr) for the same imposed along-axis pressure gradient (Figure 10a). The calculated $Q$ for the Braun et al. viscosity structure is roughly twice that of the strictly pressure- and temperature-dependent solution upstream of the transform ($Y > 0$ km), but is more than 4 times that of the strictly pressure- and temperature-dependent solution downstream of the transform (at $Y \sim -100$ km) (Figure 10b), indicating that plume-driven flow can travel longer distances along-axis for the Braun et al. viscosity structure. Because more flow can be accommodated at greater depth for the Braun et al. model than the strictly pressure- and temperature-dependent model, the shallow transform damming effect is predicted to be considerably smaller for the Braun et al. viscosity structure.

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5. Discussion: Implications for Plume-Ridge Interactions

5.1. Marion Hot Spot and Its Interactions With Ultra-Slow Spreading SWIR

[25] The results of this modeling study can be used to explore how ridge segmentation affects the along-axis length of plume-generated geochemical/geophysical anomalies, or waist width $W$. The ultra-slow spreading SWIR is highly segmented in the vicinity of the Marion hot spot (Figure 1a). Georgen et al. [2001] use ship track bathymetry data, free-air gravity, residual gravity anomalies (RMBA), and limited available geochemical data to investigate Marion’s waist width. Although precise definition of the length of Marion-affected SWIR is made difficult by short-wavelength, segment-scale variations, Georgen et al. [2001] point out that the along-axis influence of the Marion hot spot is most prominent between the Andrew Bain and Discovery II FZs (Figure 1a, and additional figures in Georgen et al. [2001]). Andrew Bain FZ is among the world’s longest transform offsets, with a length of ~720 km, while Discovery II FZ is a dual fracture zone system with a combined offset of approximately ~350 km.

[26] To apply the numerical results obtained in this study to the SWIR, we assume that the Marion plume drives mantle flow from Marion Island to the SWIR axis near the Eric Simpson FZ, and the flow then disperses along-axis to both sides of the Eric Simpson FZ. To the east, the first fracture zone encountered by the hypothesized along-axis flow is Discovery II. Extrapolation of the results in Figure 9 predicts that the reduction in along-axis flux for the 350-km long Discovery II system should be much greater than 40%, which is the calculated flux reduction for the $L_t = 250$ km case (Table 2). Moreover, within a distance of <100 km beyond the transform, $Q$ is predicted to diminish to 0% (Figure 9). Therefore modeling results predict that the along-axis geophysical expression of the
Marion plume would terminate in the vicinity of the Discovery II FZ. A similar argument applies for the westernmost boundary of the Marion axial anomaly. The Andrew Bain FZ, with an offset length approximately twice as long as that of Discovery II, is predicted to effectively block all westward asthenospheric flow.

[27] Geophysical data qualitatively support the transform-limited Marion plume dispersal model. Bathymetric surveys of the segment between the Andrew Bain and Marion FZs, as well as the segment between the Marion and Prince Edward FZs, reveal more robust magmatism than would be expected for an ultra-slow spreading ridge [Grindlay et al., 1996, 1998, 2000]. This observation supports the hypothesis that the relatively short offsets of the Marion and Prince Edward FZs are not sufficient to curtail Marion-driven along-axis flow. In contrast, the large offset Andrew Bain is sufficient to limit the Marion bathymetric and gravity anomalies as noted above (Figure 1). Geochemical substantiation of Marion’s waist width is difficult at present because of wide sample spacing. However, as new geochemical data become available, they will provide valuable additional constraint on Marion’s interaction with the SWIR, and better illustrate the extent to which transforms control Marion dispersion.

[29] The qualitative agreement between geophysical observations and model predictions in the vicinity of the Marion hot spot also has implications for the nature of upper mantle viscosity structure. As discussed above, models that employ a more complex viscosity structure, including dehydration, melting, and changes in deformation mechanism, predict that plume-driven flow should occur at relatively great depths as compared to models that use strictly temperature- and pressure-dependent viscosity. As a result, the transform damming effect is expected to be significantly less for the complex viscosity structure. Thus the apparent sensitivity of geophysical anomalies to ridge offsets in the Marion area is qualitatively consistent with the hypothesis of a relatively shallow low viscosity zone beneath the SWIR axis.

### 5.2. Plume Waist Width

[30] Results of this study predict that long transform offsets along ultra-slow spreading ridges may strongly localize axial plume anomalies and thus decrease the plume width W relative to the case of an unsegmented ridge. Numerical [Ribe et al., 1995; Ito et al., 1996] and laboratory modeling [Feighner and Richards, 1995] suggests that steady state waist width W scales as $W \sim C_0 (Q_v/2U)^{1/2}$, where $Q_v$ is plume volume flux and $C_0$ is a scaling coefficient between 1.77 and 2.12, predicting W increases with decreasing spreading rate U. These studies, however, considered only unsegmented ridges and did not take into account the effects of transform offsets. In contrast, our modeling results suggest that in the presence of a transform offset, W will decrease according to transform offset length. The above scaling relationship, therefore,
may be modified to $W \sim c_0 r_o (Q_v/2U)^{1/2}$, where $r_o$ accounts for waist width reduction due to ridge geometry. The value of $r_o$ should be highly dependent on specific ridge configuration with $r_o = 1$ for no transform offset and $r_o = 0$ for a transform fault of infinite offset length. Further numerical modeling work is required to calculate $r_o$ for cases where plume-driven asthenospheric flow crosses multiple transforms with intervening segments of variable length. For example, for the case of the SWIR to the west of the Marion plume, it is likely that reduction in flow occurs across the Marion and Prince Edward FZs prior to reaching the Andrew Bain FZ. Additional modeling work can estimate the cumulative value of $r_o$ for this three-transform system.

Figure 11. Schematic cartoons of plume-ridge interaction for (a) an unsegmented ridge and (b) a ridge with significant transform offsets. Red circles show plan view of a hypothesized vertical plume conduit. The conduit size and flux of plume 1 are assumed to equal to those of plume 2. The half-spreading rate $U$ is also assumed to be the same. Yellow shading depicts along-axis dispersal of plume material along a low-viscosity pipe, and the ridge axes with a plume signature are emphasized with blue lines. The lengths of the ridges affected by the plumes, or waist width, are $W_1$ and $W_2$. Because along-axis flux in b is limited by transform faults, $W_2 < W_1$. Thus flux inferred from waist width will be an underestimate for plume 2 compared to the case of an unsegmented ridge.

6. Conclusions

The results from this study indicate that transform faults affect plume-driven mantle flow in two important ways:

1. Transform faults reduce along-axis flux. The degree of flux reduction increases with increasing transform offset length. For example, relative to the case with no transform offset, the along-axis flux is calculated to be reduced by as much as 40% when crossing a transform with offset length of $L_t = 250$ km, for a half-spreading rate of 0.75 cm/yr and strictly pressure- and temperature-dependent viscosity structure. Furthermore, flux decreases rapidly along the segment downstream of the transform, such that for $L_t = 250$ km, the normalized flux is effectively zero within 100 km downstream of the transform.

2. Transforms deflect shallow asthenospheric along-axis flow toward the direction of the next ridge segment. Moreover, with longer transform offsets, the along-axis flow stops farther upstream of the transform.

The degree to which a transform fault affects plume-driven along-axis asthenospheric flux is sensitive to viscosity structure. The transform damming effect is most pronounced for strictly pressure- and temperature-dependent viscosity, because most flow occurs in a region of low viscosity at the relatively shallow depth of ~75 km, approximately the thickness of cold lithosphere. However, the transform damming is calculated to be less for viscosity structures that additionally include the effects of melting, dehydration, and change in deformation mechanism, since these additional effects result in a thick,
high-viscosity layer extending to depths greater than lithospheric thickness.

[36] Transform offsets in slow- and ultraslow-spreading, highly segmented ridge environments are likely to greatly limit the along-axis dispersion of plume material. Along the Southwest Indian Ridge, axial flow driven by the Marion plume is likely curtailed by the long-offset Andrew Bain and Discovery II fracture zones, severely limiting its lateral extent.

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