On the thermal structure of oceanic transform faults

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Abstract:

We use 3-D finite element simulations to investigate the upper mantle temperature structure beneath oceanic transform faults. We show that using a rheology that incorporates brittle weakening of the lithosphere generates a region of enhanced mantle upwelling and elevated temperatures along the transform, with the warmest temperatures and thinnest lithosphere predicted near the center of the transform. In contrast, previous studies that examined 3-D advective and conductive heat transport found that oceanic transform faults are characterized by anomalously cold upper mantle relative to adjacent intra-plate regions, with the thickest lithosphere at the center of the transform. These earlier studies used simplified rheologic laws to simulate the behavior of the lithosphere and underlying asthenosphere. Here, we show that the warmer thermal structure predicted by our calculations is directly attributed to the inclusion of a more realistic brittle rheology. This warmer upper mantle temperature structure is consistent with a wide range of geophysical and geochemical observations from ridge-transform environments, including the depth of transform fault seismicity, geochemical anomalies along adjacent ridge segments, and the tendency for long transforms to break into a series of small intra-transform spreading centers during changes in plate motion.

Key Words: Oceanic transform faults, mid-ocean ridges, fault rheology, intra-transform spreading centers

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1. Introduction

Oceanic transform faults are an ideal environment for studying the mechanical behavior of strike-slip faults because of the relatively simple thermal, kinematic, and compositional structure of the oceanic lithosphere. In continental regions fault zone rheology is influenced by a combination of mantle thermal structure, variations in crustal thickness, and heterogeneous crustal and mantle composition. By contrast, the rheology of the oceanic upper mantle is primarily controlled by temperature. In the ocean basins effective elastic plate thickness (e.g., Watts, 1978) and the maximum depth of intra-plate earthquakes (Chen and Molnar, 1983; McKenzie et al., 2005; Wiens and Stein, 1983) correlate closely with the location of the 600ºC isotherm as calculated from a half-space cooling model. Similarly, recent studies show that the maximum depth of transform fault earthquakes corresponds to the location of the 600ºC isotherm derived by averaging the half-space thermal structures on either side of the fault (Abercrombie and Ekström, 2001; Boettcher, 2005). These observations are consistent with extrapolations from laboratory studies on olivine that indicate the transition from stable to unstable frictional sliding occurs around 600ºC at geologic strain-rates (Boettcher et al., 2006). Furthermore, microstructures observed in peridotite mylonites from oceanic transforms show that localized viscous deformation occurs at temperatures around 600–800ºC (e.g., Jaroslow et al., 1996; Warren and Hirth, 2006).

While a half-space cooling model does a good job of predicting the maximum depth of transform earthquakes, it neglects many important physical processes that occur in the Earth’s crust and upper mantle (e.g., advective heat transport resulting from temperature-dependent viscous flow, hydrothermal circulation, and viscous dissipation). Numerical models that incorporate 3-D advective and conductive heat transport indicate that the upper mantle beneath oceanic transform faults is anomalously cold relative to a half-space model (Forsyth and Wilson, 1984; Furlong et al., 2001; Phipps Morgan and Forsyth, 1988; Shen and Forsyth, 1992). This reduction in mantle temperature results from a combination of two effects: 1) conductive cooling from the adjacent old, cold lithosphere across the transform fault, and 2) decreased mantle upwelling beneath the transform. Together these effects can result in up to a ~75% increase in lithospheric thickness beneath the center of a transform fault relative to a half-space cooling model,
well as significant cooling of the upper mantle beneath the ends of the adjacent spreading centers. This characteristic ridge-transform thermal structure has been invoked to explain the focusing of crustal production toward the centers of ridge segments (Magde and Sparks, 1997; Phipps Morgan and Forsyth, 1988; Sparks et al., 1993), geochemical evidence for colder upper mantle temperatures near segment ends (Ghose et al., 1996; Niu and Batiza, 1994; Reynolds and Langmuir, 1997), increased fault throw and wider fault spacing near segment ends (Shaw, 1992; Shaw and Lin, 1993), and the blockage of along-axis flow of plume material (Georgen and Lin, 2003).

However, correlating the maximum depth of earthquakes on transform faults with this colder thermal structure implies that the transition from stable to unstable frictional sliding occurs at temperatures closer to ~350ºC, which is inconsistent with both laboratory studies and the depth of intra-plate earthquakes. Moreover, if oceanic lithosphere is anomalously cold and thick beneath oceanic transform faults, it is difficult to explain the tendency for long transform faults to break into a series of en echelon transform zones separated by small intra-transform spreading centers during changes in plate motion (Fox and Gallo, 1984; Lonsdale, 1989; Menard and Atwater, 1969; Searle, 1983).

To address these discrepancies between the geophysical observations and the predictions of previous numerical modeling studies, we investigate the importance of fault rheology on the thermo-mechanical behavior of oceanic transform faults. Earlier studies that incorporated 3-D conductive and advective heat transport used simplified rheologic laws to simulate the behavior of the lithosphere and underlying asthenosphere. Using a series of 3-D finite element models, we show that brittle weakening of the lithosphere strongly reduces the effective viscosity beneath the transform, resulting in enhanced upwelling and thinning of the lithosphere. Our calculations suggest that the thermal structure of oceanic transform faults is more similar to that predicted from half-space cooling, but with the warmest temperatures located near the center of the transform. These results have important implications for the mechanical behavior of oceanic transforms, melt generation and migration at mid-ocean ridges, and the long-term response of transform faults to changes in plate motion.
2. Model Setup

To solve for coupled 3-D incompressible mantle flow and thermal structure surrounding an oceanic transform fault we use the COMSOL 3.2 finite element software package (Figure 1). In all simulations, flow is driven by imposing horizontal velocities parallel to a 150-km transform fault along the top boundaries of the model space assuming a full spreading rate of 6 cm/yr. The base of the model is open to convective flux without resistance from the underlying mantle. Symmetric boundary conditions are imposed on the sides of the model space parallel to the spreading direction, and the boundaries perpendicular to spreading are open to convective flux. The temperature across the top and bottom of the model space is set to $T_s = 0^\circ C$ and $T_m = 1300^\circ C$, respectively. Flow associated with temperature and compositional buoyancy is ignored.

To investigate the importance of rheology on the pattern of flow and thermal structure at oceanic transform faults we examined four scenarios with increasingly realistic descriptions of mantle rheology: 1) constant viscosity, 2) temperature-dependent viscosity, 3) temperature-dependent viscosity with an pre-defined weak zone around the transform, and 4) temperature-dependent viscosity with a visco-plastic approximation for brittle weakening. In all models we assume a Newtonian mantle rheology. In Models 2–4 the effect of temperature on viscosity is calculated by:

$$
\eta = \eta_0 \left[ \frac{\exp(Q_o / RT)}{\exp(Q_o / RT_m)} \right]
$$

where $\eta_0$ is the reference viscosity of $10^{19}$ Pa·s, $Q_o$ is the activation energy, and $R$ is the gas constant. In all simulations we assume an activation energy of 250 kJ/mol. This value represents a reduction of a factor of two relative to the laboratory value as a linear approximation for non-linear rheology (Christensen, 1983). The maximum viscosity is not allowed to exceed $10^{23}$ Pa·s.

3. Influence of 3-D Mantle Flow on Transform Thermal Structure

Figure 2 illustrates the thermal structure calculated at the center of the transform fault from Models 1–4, as well as the thermal structure determined by averaging half-spacing cooling models on either side of the transform. Because the thermal structure calculated from half-space cooling is equal on the adjacent plates, the averaging approach predicts
the same temperature at the center of the transform fault as for the adjacent intra-plate regions. This model, therefore, provides a good reference for evaluating whether our 3-D numerical calculations predict excess cooling or excess heating below the transform. The temperature solution for a constant viscosity mantle (Model 1) was determined previously by Phipps Morgan and Forsyth (1988); our results agree with theirs to within 5% throughout the model space. The coupled temperature, mantle flow solution predicts a significantly colder thermal structure at the center of the transform than does half-space cooling, with the depth of the 600°C isotherm increasing from ~7 km for the half-space model to ~12 km for the constant viscosity flow solution (Figure 2A). As noted by Phipps Morgan and Forsyth (1988) this reduction in temperature is primarily the result of decreased mantle upwelling beneath the transform fault relative to enhanced upwelling under the ridge axis. Shen and Forsyth (1992) showed that incorporating temperature-dependent viscosity (e.g., Model 2) produces enhanced upwelling and warmer temperatures beneath the ridge axis relative to a constant viscosity mantle (Figure 3). However, away from the ridge axis the two solutions are quite similar and result in almost identical temperature-depth profiles at the center of the transform (Figures 2A & 3).

4. Influence of Fault Zone Rheology on Transform Thermal Structure

Several lines of evidence indicate that oceanic transform faults are significantly weaker than the surrounding lithosphere. Comparisons of abyssal hill fabric observed near transforms to predictions of fault patterns from numerical modeling suggest that the mechanical coupling across the fault is very weak on geologic time scales (Behn et al., 2002; Phipps Morgan and Parmentier, 1984). Furthermore, dredging in transform valleys and valley walls frequently returns serpentinized peridotites (Cannat et al., 1991; Dick et al., 1991), which may promote considerable frictional weakening along the transform (Escartín et al., 2001; Moore et al., 1996; Rutter and Brodie, 1987). Finally, in comparison to continental strike-slip faults, seismic moment studies show that oceanic transforms have high seismic deficits (Boettcher and Jordan, 2004; Okal and Langenhorst, 2000), suggesting that oceanic transforms may be characterized by large amounts of aseismic slip.
In Model 3 we simulate the effect of a weak fault zone, by setting the viscosity to \(10^{19}\) Pa\,s in a 5-km wide region surrounding the transform that extends downward to a depth of 20 km. This results in a narrow fault zone that is 3–4 orders of magnitude lower viscosity than the surrounding regions. Our approach is similar to that used by Furlong et al. (2001) and van Wijk and Blackman (2004), though these earlier studies modeled deformation in a visco-elastic system in which the transform fault was simulated as a shear-stress-free plane using the slippery node technique of Melosh and Williams (1989).

Our models show that the incorporation of a weak fault zone produces slightly warmer conditions along the transform than for either a constant viscosity mantle (Model 1) or temperature-dependent viscosity without an imposed fault zone (Model 2). However, the predicted temperatures from the fault zone model remain colder than those calculated by the half-space cooling model (Figures 2A & 3). Varying the maximum depth of the weak zone does not significantly influence the predicted thermal structure.

Representing the transform as a pre-defined zone of uniform weakness clearly over-simplifies the brittle processes occurring within the lithosphere. In Model 4, we incorporate a more realistic formulation for fault zone behavior by using a visco-plastic rheology to simulate brittle weakening (Chen and Morgan, 1990). In this formulation, brittle strength is approximated by defining a frictional resistance law (e.g., Byerlee, 1978):

\[
\tau_{\text{max}} = C_o + \mu \rho g z
\]

in which \(C_o\) is cohesion (10 MPa), \(\mu\) is the friction coefficient (0.6), \(\rho\) is density (3300 kg/m\(^3\)), \(g\) is the gravitational acceleration, and \(z\) is depth. Following Chen and Morgan (1990), the maximum effective viscosity is then limited by:

\[
\eta = \frac{\tau_{\text{max}}}{\sqrt{2\dot{e}_{\|}}}
\]

where \(\dot{e}_{\|}\) is the second-invariant of the strain-rate tensor. The effect of adding this brittle failure law is to limit viscosity near the surface where the temperature dependence of Equation 1 produces unrealistically high mantle viscosities and stresses (Figure 2B).

The inclusion of the visco-plastic rheology results in significantly warmer thermal conditions beneath the transform than predicted by Models 1–3 or half-space cooling (Figures 2A & 3). The higher temperatures result from the brittle weakening of the
lithosphere, which reduces the effective viscosity by up to 2 orders of magnitude in a 10-km wide region surrounding the fault zone. Unlike Model 3 the width of this region is not predefined and develops as a function of the rheology and applied boundary conditions. The zone of decreased viscosity enhances passive upwelling beneath the transform, which in turn increases upward heat transport, warming the fault zone and further reducing viscosity (Figure 4). The result is a characteristic thermal structure in which the transform fault is warmest at its center and cools towards the adjacent ridge segments (Figures 2C). Moreover, rather than the transform being a region of anomalously cold lithosphere relative to a half-space cooling model and the surrounding intra-plate mantle, the center of the transform is warmer than adjacent lithosphere of the same age.

Although we have not explicitly modeled the effects of non-linear rheology, several previous studies have examined the importance of a non-linear viscosity law on mantle flow and thermal structure in a segmented ridge-transform system (Furlong et al., 2001; Shen and Forsyth, 1992; van Wijk and Blackman, 2004). Without the effects of brittle weakening, these earlier studies predicted temperatures below the transform that were significantly colder than a half-space cooling model. Thus, we conclude that the inclusion of a visco-plastic rheology is the key factor for producing the warmer transform fault thermal structure illustrated in Model 4.

5. Implications for the Behavior of Oceanic Transform Faults

Our numerical simulations indicate that brittle weakening plays an important role in controlling the thermal structure beneath oceanic transform faults. Specifically, incorporating a more realistic treatment of brittle rheology (as shown in Model 4), results in an upper mantle temperature structure that is consistent with a wide range of geophysical and geochemical observations from ridge-transform environments. The temperatures below the transform fault predicted in Model 4 are similar to the half-space cooling model, indicating that the maximum depth of transform fault seismicity is indeed limited by the ~600°C isotherm as shown by Abercrombie and Ekström (2001). This temperature is consistent with the transition from velocity-weakening to velocity-strengthening frictional behavior extrapolated from laboratory experiments (Boettcher et
al., 2006) and the depth of oceanic intra-plate earthquakes (McKenzie et al., 2005). In addition, a combination of microstructural and petrological observations indicate that the transition from brittle to ductile processes occurs at a temperature of ~600°C in oceanic transform faults. Microstructural analyses of peridotite mylonites recovered from oceanic fracture zones indicate that strain localization results from the combined effects of grain size reduction, grain boundary sliding and second phase pinning (Warren and Hirth, 2006). Jaroslow et al. (1996) estimated a minimum temperature for mylonite deformation of ~600°C, based on olivine-spinel geothermometry. Furthermore, the randomization of pre-existing lattice preferred orientation (LPO) in the finest grained areas of these mylonites indicates the grain size reduction promotes a transition from dislocation creep processes to diffusion creep, consistent with extrapolation of experimental olivine flow laws to temperatures of 600–800°C.

While the inclusion of a visco-plastic rheology results in significant warming beneath the transform fault, the temperature structure at the ends of the adjacent ridge segments changes only slightly relative to the solution for a constant viscosity mantle. In particular, both models predict a region of cooling along the adjacent ridge segments that extends 15–20 km from the transform fault (Figure 2C). This transform “edge effect” has been invoked to explain segment scale variations in basalt chemistry (Ghose et al., 1996; Niu and Batiza, 1994; Reynolds and Langmuir, 1997) and increased fault throw and fault spacing toward the ends of slow-spreading ridge segments (Shaw, 1992; Shaw and Lin, 1993). In addition, this along-axis temperature gradient provides an efficient mechanism for focusing crustal production toward the centers of ridge segments (Magde and Sparks, 1997; Phipps Morgan and Forsyth, 1988; Sparks et al., 1993).

The elevated temperatures near the center of the transform in Model 4 may also account for the tendency of long transform faults to break into a series of intra-transform spreading centers during changes in plate motion (Fox and Gallo, 1984; Lonsdale, 1989; Menard and Atwater, 1969). This “leaky transform” phenomenon has been attributed to the weakness of oceanic transform faults relative to the surrounding lithosphere (Fox and Gallo, 1984; Lowrie et al., 1986; Searle, 1983). However, the leaky transform hypothesis is in direct conflict with the thermal structure predicted from Models 1–3, which show the transform to be a region of anomalously cold, thick lithosphere. In contrast, the
thermal structure predicted by Model 4 indicates that transforms are hottest and weakest near their centers (Figure 2C). Thus, perturbations in plate motion, which generate extension across the transform, should result in rifting and enhanced melting in these regions.

The incorporation of the brittle rheology also promotes strain localization on the plate scale. In particular, if transforms were regions of thick, cold lithosphere (as predicted by Models 1–3) then over time deformation would tend to migrate outward from the transform zone into the adjacent regions of thinner lithosphere. However, the warmer thermal structure that results from the incorporation of a visco-plastic rheology will tend to keep deformation localized within the transform zone on time-scales corresponding to the age of ocean basins.

In summary, brittle weakening of the lithosphere along oceanic transform faults generates a region of enhanced mantle upwelling and elevated temperatures relative to adjacent intra-plate regions. The thermal structure is similar to that predicted by a half-space cooling model, but with the warmest temperatures located at the center of the transform. This characteristic upper mantle temperature structure is consistent with a wide range of geophysical and geochemical observations, and provides important constraints on the future interpretation of microseismicity data, heat flow, and basalt and peridotite geochemistry in ridge-transform environments.

Acknowledgements

We thank Jeff McGuire, Laurent Montési, Trish Gregg, Jian Lin, Henry Dick, and Don Forsyth for fruitful discussions that helped motivate this work. Funding was provided by NSF grants EAR-0405709 and OCE-0443246.
Figure Captions

Figure 1: Model setup for numerical simulations of mantle flow and thermal structure at oceanic transform faults. All calculations are performed for a 150-km long transform and a full spreading rate of 6 cm/yr. Locations of cross-sections used in Figures 2–4 are shown in grey.

Figure 2: (A) Thermal structure and (B) stress calculated versus depth calculated at the center of a 150-km long transform fault assuming a full spreading rate of 6 cm/yr. (C) Location of the 600°C and 1200°C isotherms along the plate boundary for the half-space model (grey), Model 1 (black), and Model 4 (red).

Figure 3: Cross-sections of mantle temperature at a depth of 20 km for (A) Model 1: constant viscosity of $10^{19}$ Pa·s, (B) Model 2: temperature-dependent viscosity, (C) Model 3: temperature-dependent viscosity with a weak fault zone, and (D) Model 4: temperature-dependent viscosity with a frictional failure law. Black arrows indicate horizontal flow velocities. Grey lines show position of plate boundary. Location of horizontal cross-section is indicated in Figure 1. Note that Model 4 incorporating frictional resistance predicts significantly warmer temperatures along the transform than Models 1–3.

Figure 4: Vertical cross-sections through the center of the transform fault showing (left) strain-rate, and (right) temperature and mantle flow for Models 1–4. The location of the cross-sections is indicated in Figure 1. Note the enhanced upwelling below the transform results in warmer thermal structure for Model 4 compared to Models 1–3.
References


Behn et al: Thermal Structure of Oceanic Transform Faults, Submitted to Geology March 2006


Figure 1
Figure 2
Along-Ridge Distance (km)

A. Model 1: Constant $\eta$

B. Model 2: $\eta(T)$

C. Model 3: $\eta(T) + \text{Fault}$

D. Model 4: $\eta(T, \text{friction})$

Temperature (°C)

Figure 3
Figure 4