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FOCUSED MANTLE UPWELLING BELOW MID-OCEAN RIDGES DUE TO FEEDBACK BETWEEN VISCOSITY AND MELTING

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Abstract. We present the first internally consistent calculation which leads to a narrow 'conduit' of rapid vertical advection and melting of mantle under a spreading center. In this model, mantle flow is driven by plate separation and compositional buoyancy. Melt segregation is described as flow through a permeable media. The major new feature is that the viscosity of the mantle is considered to be a strong function of the amount of partial melt present. Experiments show that the bulk viscosity of a partially molten rock is sharply reduced when the melt fraction exceeds a critical value. In the model, the viscosity is reduced as the critical melt fraction is approached. Whether or not a critical melt fraction can be reached under a spreading center depends on the mantle permeability for melt flow. The width of the upwelling area is controlled by the magnitude of the melt related viscosity reduction. Crust should be formed above the focused upwelling. Seismic observations show that the region of crustal accretion is only a few kilometers wide at fast spreading centers. With a viscosity reduction of three orders of magnitude the model predicts a zone of crustal accretion of this width.

Introduction

Several observations indicate that the region of melt emplacement and crustal thickening is very narrow at a mid-ocean ridge. Seismic data show that the crust attains its full thickness within less than 5 km of some fast spreading centers [Detrick et al., 1986]. Here, it is assumed that the seismic thickness equates with the thickness of basaltic crust. Topography at slow spreading ridges makes it more difficult to constrain crustal thickness using seismic data. However, the width of the neo-volcanic zone on both fast and slow spreading mid-ocean ridges, as determined from the morphology of the sea floor, is generally less than a few kilometers [e.g. Macdonald, 1982].

It has long been accepted that the creation of oceanic crust at mid-ocean ridges is a consequence of pressure release melting of mantle that is drawn up by the separation of lithospheric plates. Petrological studies indicate that the depth at which melting begins is between about 40 and 100 km below a ridge and the maximum degree of partial melting is between about 20 and 40 percent, depending on the temperature of the mantle [Klein and Langmuir, 1987]. Basaltic melt has a lower density than the residual solid from which it formed. This density contrast should lead to segregation of the melt from the solid. If some of the melt between solid crystalline grains forms interconnected channels, then it may be appropriate to model the melt migration as porous media flow and use Darcy's law to derive flow rates. Many workers have done such calculations and several have applied the results to the problem of melt segregation under mid-ocean ridges [Ahern and Turcotte, 1979; Reid and Jackson, 1981]. It has been generally assumed that the rate of melt flow is large relative to typical mantle flow rates, which are thought to be on the order of plate velocities. This results in very little melt being retained in the mantle even though the degree of melting may be large.

Combining melting relations with models of mantle flow driven by plate separation allows calculation of the location of significant melting and the width of the region of crustal accretion under a spreading center. A major result of this type of model is that the distance over which significant crustal thickening occurs is directly proportional to the spreading rate. For a slow spreading ridge, with a half spreading rate less than 1 cm/yr, the crust would be largely formed within 10-20 km of the ridge crest. However, for a half spreading rate of 5 cm/yr the crust would still be thickening 50 km from the ridge crest.

There is a large difference between observations of the width of the zone of melt emplacement at a spreading center and what is predicted by simple models. Several mechanisms have been suggested in attempts to resolve this paradox, each involving either concentration of the flow of melt toward a mid-ocean ridge or concentration of the mantle upwelling there. In the first class of models, partial melting occurs over a wide region of the mantle, but the melt is drawn to the ridge by non-hydrostatic pressure gradients, as first proposed by Sleep [1974]. Spiegelman and McKenzie [1987] have studied this mechanism, treating constant viscosity mantle flow driven by plate spreading. They find that to pull melt out of a wide area requires a mantle viscosity greater than 10^{21} Pa-s. Such large viscosity values at shallow depths in the mantle are not consistent with several geophysical observations [see discussion in Buck and Parmentier, 1986]. Another suggestion is that strain within the mantle flowing below a ridge would lower the permeability for melt flow in the direction of the ridge crest [Phipps Morgan, 1987]. This model requires an isotropic distribution of veins in the mantle either to exist before melting begins or to be created just as melting begins. Mantle flow then deforms the veins into a preferred orientation. This model must be considered speculative since the physics controlling the one-time creation of a vein network is not specified.

A second type of model examines ways to focus upwelling mantle into a narrow zone. Rabinowitz et al. [1987] suggest that the buoyancy of the melt/residuum combination compared to unmelted mantle will focus flow under a ridge. Scott and Stevenson [1989] have studied the effect of buoyancy driven circulation in a numerical model of plate driven constant viscosity flow and melt migration. They find some focusing of the upwelling region and therefore of the melt production, particularly at slow spreading rates. We have investigated this mechanism for a temperature dependent viscosity which naturally gives rise to a thickening lithosphere. We find that the upwelling and significant melting under a fast spreading

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ridge is not concentrated strongly due to the buoyancy of the melt/residuum combination; crust thickens appreciably out to 50 km away from the spreading center. This is a far wider crustal accretion zone than required to match the observations at fast spreading centers. In this paper, we consider a new mechanism which, when combined with the effects of buoyancy, can lead to an extremely narrow zone of melting below a spreading center.

Viscosity and Partial Melting

The new mechanism we incorporate into our models is that the viscosity of the mantle is a strong function of the melt fraction present, when melt fractions are large. The presence of small amounts of partial melt (<10%) may enhance diffusion controlled creep by allowing more rapid pressure solution of the crystalline matrix and precipitation in areas of low stress. Laboratory measurements on peridotites indicate that the viscosity is reduced by less than an order of magnitude by this effect [Cooper and Kohlstedt, 1984]. When melt fractions are larger, the viscosity can become dominated by melt lubricated grain sliding [Arzi, 1978]. No laboratory work has been done on ultramafics, but studies of granite partial melts indicated that the critical melt fraction for a large viscosity reduction may be as low as 20% [Auer et al., 1981]. The viscosity of a crystalline mush with greater than critical melt fraction could be many orders of magnitude lower than for material with viscosity that is dominated by diffusion or dislocation creep. Figure 1 shows how the viscosity of the mantle is related to melt fraction in our calculations. The critical melt fraction is labeled ϕ_c and the large change in viscosity is termed a "viscosity breakdown".

One-Dimensional Melt Flow

To give some insight into ways that large melt fractions could be produced under a mid-ocean ridge, we first treat the problem of melt generation and transport in one dimension. Consider a region of mantle where mantle upwelling with a constant velocity v_u undergoes pressure-release melting. As in previous studies [e.g. Ahern and Turcotte, 1979, Klein and Langmuir, 1987] the degree of melting, F, is taken to be a linear function of the distance above D_s, the depth of intersection of a mantle adiabat and the solidus. Thus, the rate of melt generation is $v_u \cdot F_{max}/D_s$. From a review of thermodynamic data on mantle minerals McKenzie [1984] estimates that the slope (F_{max}/D_s) of this melting relation is between 0.7 and 1.5 %/kbar. We use a value of 1.2 %/kbar or 0.4 %/km.

Partial melts form along grain boundaries and must migrate relative to the matrix to form concentrated magmas. Since melt is lighter than the matrix residuum bouyancy will drive it

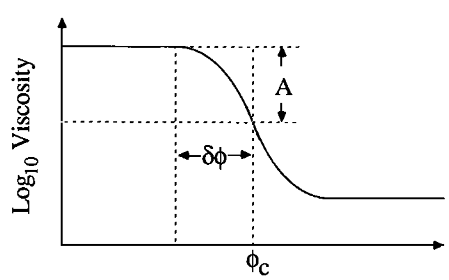

Fig. 1. The dependence of viscosity on melt fraction used in these calculations.

upwards. Numerous workers have treated melt migration as flow through a porous media, where the porosity equals the melt fraction. Ribe [1985] and several other workers argue that the compaction of the matrix is not the rate limiting factor in melt migration because the length scale for rapid compaction in the mantle (10-100m) is small compared to the length scale for major changes in melt concentration (1-10 km). The rate of melt flow is limited by the resistance between the viscous melt and intergrain channels. The average resistance can be quantified by a permeability k. Following Richter and McKenzie [1984] and others we adopt a power law relation between porosity ϕ and permeability k, namely $k = a^2 \phi^2/b$, where a is the grain size and b depends on the geometry of the partial melt distribution and is approximately 10^4. The velocity of melt, v_m, relative to the crystalline matrix can then be expressed as:

$$v_m - v_u = k\, g\, \Delta\rho / \mu_m \phi = v_r \phi \qquad (1)$$

where g is the acceleration of gravity, $\Delta\rho$ is the density contrast between melt and solid and μ_m is the viscosity of the melt. We take $\Delta\rho$ =500 kg/m and the viscosity of basaltic melt μ_m = 10 Pa-s [Kushiro, 1982]. We lump together several parameters into one term $v_r = a^2\, g\, \Delta\rho / b\, \mu_m$ which we call the reference velocity for Darcy flow. Taking a= 0.1 mm gives a value of v_r of 1 cm/yr. Previous workers assumed larger grain sizes and thus much larger values of the rate of percolative flow [e.g. Ahern and Turcotte [1979] assume a grain size of 2 mm which would give a value of v_r of order 10^2 cm/yr]. The reference velocity times the porosity gives the rate of melt flow relative to the solid matrix. For example, if v_r = 10 cm/yr and ϕ = 0.1 then the relative velocity of melt flow is 1 cm/yr.

An expression for the steady-state melt fraction can be derived for the case of constant velocity upwelling within a region of partial melting:

$$\partial\phi/\partial t = 0 = v_u\, F_{max}/D_s - v_u\, \partial\phi/\partial z - v_r\, \partial(\phi^2)/\partial z \qquad (2)$$

Solving this equation numerically gives the melt fraction present as a function of depth within the melting region. Figure 2 shows solutions of this equation for different values of the ratio v_r/v_u and for the depth D_s. The plots show that when the v_r is small compared to the upwelling velocity then the melt fraction present at shallow depth can approach the total amount of melt produced by the upwelling. Obviously, if no melt migration takes place then $\phi(z) = F(z)$. If v_r is large relative to the upwelling velocity then little melt is retained within the matrix. If the critical value of melt fraction for the breakdown of viscosity is less than F_{max} then for a range of reference velocities melt fractions will exceed ϕ_c at shallow depths.

The upwelling velocity under a ridge for simple plate driven flow models [e.g. Spiegelman and McKenzie, 1987] is roughly equal to the plate spreading velocity. Spreading velocities range from just under 1 cm/yr to nearly 10 cm/yr [Macdonald, 1982]. Thus for our estimated value of v_r of about 1 cm/yr our one dimensional model predicts that melt fractions greater than 20% would be produced under ridges. We, therefore, expect that the viscosity of the sub-ridge partially molten mantle may be sharply reduced relative to unmelted mantle.

Two-Dimensional Flow Model

To show the effect of viscosity breakdown on the pattern of flow under a mid-ocean ridge, we carried out the following

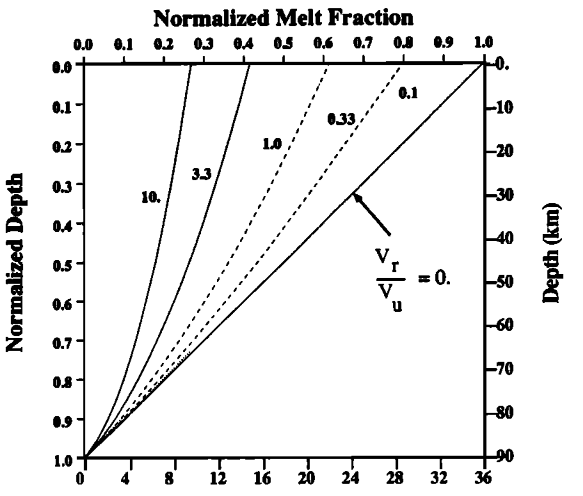

Fig. 2. Relation between the melt fraction f and depth as a function of the ratio between the reference velocity for Darcy flow, v_r, and the upwelling velocity v_u. Dimensional scale assumes that the maximum fraction of melting at the surface F_{max}=36% and the depth where melting begins D_s = 90 km.

numerical experiment. Consider a model box filled with viscous fluid representing an area of the mantle under a spreading center. The right side of the box is a symmetry boundary representing a vertical surface passing through a linear ridge crest. The fluid at the surface is made to move horizontally at a constant rate, except within 5 km of the symmetry boundary where the velocity tapers smoothly to zero. The bottom and left side allow flow that is normal to the boundaries. These boundaries are far (200 km) from the model spreading center. The temperature of the material which fills the box from below is 1300 °C and the top boundary is kept at 0 °C. The viscosity μ within the box is related to temperature T and pressure P by a standard relation : $\mu(T,P) = \mu_0 \exp[(E+PV)/RT]$, where R is the universal gas constant, the activation energy, E, is set at 420 Kj/mole and the activation volume,V, equals 20 cm^3/mole. μ_0 is set so that the viscosity at zero pressure and 1300°C is 10^{18} Pa-s. The viscosity changes by a factor of 10 for a 100° C change in temperature; so regions with temperatures more than 250° C colder than the upwelling region are essentially frozen and move as a rigid plate (Figure 3). This model lithosphere thickens with distance from the ridge crest due to conductive cooling. To avoid numerical difficulties, the viscosity is made to drop smoothly as a function of melt fraction as in Figure 1. When ϕ is within $\delta\phi$ of ϕ_c the viscosity varies as $\log[\mu(T,P,\phi)] = \log[\mu(T,P)]\ A \cos[\pi(\phi+\delta\phi-\phi_c)/2\,\delta\phi]$. Here, we take $\phi_c = 0.20$ and $\delta\phi = 0.10$ and A = 2. Thus, the viscosity drops by 2 orders of magnitude when $\phi = \phi_c$ and by a maximum of 4 orders of magnitude when ϕ=0.3.

We solve the Navier-Stokes equations for mass, momentum and energy conservation within the model box, making the usual Boussinesq approximations. Equation 2, with $v_u\,\partial\phi/\partial z$ replaced by $v\cdot\nabla\phi$, is used to determine the melt fraction present. We have included the buoyancy due to low density melt and to low density residual mantle. The density of the melt is taken to be 600 kg/m^3 less than the unmelted mantle and the density of the residual mantle depends linearly on the degree of depletion. Mantle depleted by 30% is taken to be 100 kg/m^3 less than unmelted mantle. A finite difference technique is used to solve the equations for points on an irregularly spaced grid. We decreased the grid size until further reduction made a negligible difference to the solutions. A uniform grid spacing of .3 km was used in the region of most rapid flow. 110 grid points were used in each direction.

Several models with different spreading rates and reference velocities for Darcy flow have been run to steady state, but we only illustrate a fast spreading case here. Figure 3 shows results of a calculation with a plate half-spreading rate of 5 cm/yr, a Darcy velocity of 1 cm/yr. Here, D_s = 90 km and F_{max} = 0.36 for a temperature of 1300°C. Melting depends exponentially on temperature, but freezing is not allowed. The viscosity is 10^{18} Pa-s for zero pressure, T=1300°C and for zero melt fraction. The figure shows the streamlines for the flow, the position of the base of the effectively rigid lithosphere and contours of the melt fraction in one quarter of the region of calculation. The gradual concentration of streamlines over the entire depth range of the plot is a result of the pressure dependence of viscosity. The melt fraction is greater than 20% up to 25 km below the ridge crest. The flow streamlines are tightly focused as a result of the large viscosity reduction there. This flow pattern causes much of the melting and melt flow to be concentrated under the ridge.

To crudely show the rate of crustal accretion, we assume that all melt delivered to a depth of 10 km can efficiently segregate from the mantle and be added to the crust. The top part of Figure 3 shows the result of this calculation: the crust is largely formed within one to two kilometers of the spreading center. The width of the zone of crustal accretion depends strongly on the magnitude of the viscosity reduction assumed. If the maximum melt related viscosity reduction is half that used for the calculation shown in Figure 3 the zone of significant crustal accretion extends more than 10 km from the ridge crest. At lower spreading rates a smaller viscosity reduction is required to achieve the same width of the crustal accretion zone.

Discussion

The major new feature of this self-consistent model calculation is a narrow zone of focused melt rich upwelling

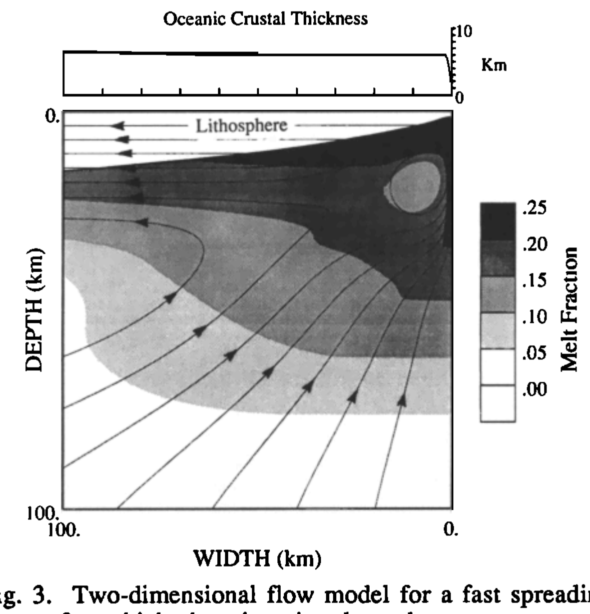

Fig. 3. Two-dimensional flow model for a fast spreading center for which the viscosity depends on temperature, pressure and melt fraction. Grey shades show the melt fraction present. The matrix streamlines are sharply focused in the area of large melt fraction and reduced viscosity.

which might be likened to a vertical conduit. It is easy to understand why this tight focusing occurs. Once some concentration of the flow begins to occur there is positive feedback for the flow to concentrate more. The ratio v_r/v_u decreases as the upwelling narrows and becomes faster (v_u increases). As the one-dimensional calculations showed, the melt fraction is increased and this reduces the viscosity further. The horizontal density difference between the upwelling and the slower moving asthenosphere depends on the lateral variation in melt fraction. Thus, the density contrast which drives the upwelling increases as the upward flow becomes faster. This feedback can explain the narrowness of the observed zone of crustal accretion at mid-ocean ridges.

For this viscosity/flow feedback to occur the viscosity of the mantle must be a function of the melt fraction present. Following Cooper and Kohlstedt (1984]) we considered that there would be little reduction in viscosity when the melt fractions are small (<10%). Thus, we were forced to use permeabilities which are lower than commonly assumed for the mantle. In support of low permeabilities, we note that field evidence indicates up to 10% melt retention in some ophiolite ultramafics (Nicolas, 1986). Alternatively, the presence of even a few percent melt may reduce the viscosity of the mantle by several orders of magnitude, as McKenzie [1984] asserts. If this is the case, a viscosity/flow feedback could occur at much lower melt fractions than shown in Figure 3. The model permeability required for this case could be in the range typically assumed for the mantle.

To make the crust at a spreading center, the segregation of melt from residual at shallow depths must be more rapid than for the melt flow rates considered here. Otherwise the divergence of mantle flow immediately below the ridge would carry much of the melt away from the ridge crest. Several mechanisms may act to increase the rate of segregation at shallow depths in the mantle (<10km). Where the deviatoric stresses are large, crack formation should occur, leading to a local increase in the permeability for melt flow. The mode of melt segregation also may change when viscosity breakdown leads to focused, essentially diapiric upwelling.

The flow lines in Figure 3 show that under the model spreading center the direction of flow changes 90° in just a few kilometers. Flow structures in the ultramafic section of the Oman ophiolite show such small-scale changes in flow direction and led Rabinowitz et al. [1987] to suggest that viscosities must be low under spreading centers and that the buoyancy of melt might drive diapiric mantle flow there. Our self-consistent model is based on their suggestions; so it is heartening that our results are consistent with the data that inspired their ideas, as well as producing a narrow zone of crustal accretion.

This mechanism of mantle flow and melt segregation may affect the topography of a ridge. A buoyant, narrow, low viscosity upwelling or conduit will produce normal stress variations at the top of the asthenosphere which should be reflected in the topography of a spreading center. Also, the conduit may not be the stable flowing region shown in these calculations. Local instabilities within the conduit may lead to the formation of diapirs which periodically sweep through the conduit. We will consider these effects in future work.

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