Diverging Plates: The Underlying Story

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Much of the geological history of the Earth is characterized by the dynamic interaction of semi-rigid plates which collide, slide by one another, or are pulled apart. The plates, or lithosphere, overlie an asthenosphere that responds like a viscous fluid to the plate motions. As a result of the intensive focus on plate tectonics over the past few decades, we have a relatively good understanding of the structure and interactions of the plates, particularly within the upper 30 to 50 km. However, we are only beginning to address important questions concerning the nature of the lowermost lithosphere and the relationship between the lithosphere and asthenosphere. Simple concepts of a sharp boundary, for example, between the lithosphere and the asthenosphere have been useful in making first-order predictions from plate tectonic theory. However, we have now advanced past the stage where such simple ideas are sufficient, and we need to employ more realistic models.

One of the areas of study where our current simple concepts limit our ability to correctly predict observed geological structures is at divergent plate boundaries (Keen and Beaumont, 1990; Keen, 1985). Divergence starts with rupture of a lithospheric plate during continental rifting. This allows hot mantle material (asthenosphere) to well up into the space created. As the mantle rises, the confining pressure is reduced; part of it melts, yielding a relatively low viscosity fluid. This can seep relatively quickly upwards into the crust where it solidifies as igneous rock. As rifting proceeds, seafloor spreading occurs and an up-welling melt creates the oceanic crust. The centre of up-welling and divergence is termed a mid-ocean ridge. This is illustrated in Fig. 1.

Igneous crust formed at the mid-ocean ridge is typically 8 km thick, although the thickness depends on the spreading rate (White, 1992). At very slow spreading rates, for example less than 10 mm/yr, time is available for significant cooling of the rising mantle which results in the formation

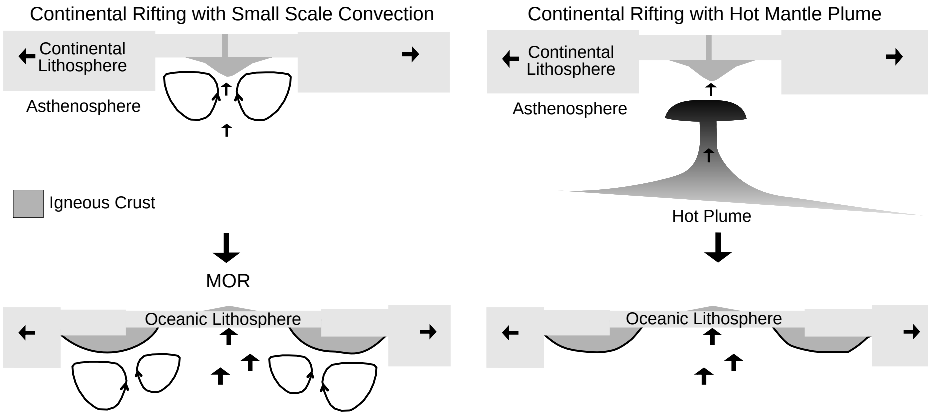

Figure 1. A cartoon showing two proposed explanations for the existence of large volcanic features associated with rifted continental margins. On the left, small scale convection, driven by the thermal perturbation created by continental rifting, delivers melt that forms igneous crust to the rifting centre. This is followed by a sea floor spreading stage (lower left) where convection and melt delivery can continue under the old continental margin, and new oceanic crust is created by upwelling at the mid ocean ridge (MOR). On the right is the hot spot explanation. The high temperature of the hot spot, arriving during the continental rifting stage, induces the large volume of melt that creates the thick igneous crust. In the sea floor spreading stage (lower right) the mantle has been cooled somewhat, or the hot spot material has been displaced, and normal temperature mantle wells up to create ordinary oceanic crust.

of a thinner oceanic crust (ca. 4 km). These low spreading rates are relatively uncommon in the world's oceans. More common is oceanic crust which is relatively uniform in thickness, varying between 6 and 8km (White, 1992).

Along much of the divergent continental margin of eastern North America, as far north as Nova Scotia, anomalously large thicknesses of igneous crust (as much as 20 km) are observed (e.g. Holbrook and Kelemen, 1993; Holbrook *et al.* 1994). This extreme thickness has been associated with a major magnetic feature called the East Coast Magnetic Anomaly (ECMA), although the exact relationship between the anomaly and the thick crust is a matter of considerable debate (e.g. Hutchinson *et al.* 1990; McBride and Nelson, 1990; Holbrook and Keleman, 1993; Keen and Potter, 1995). Figure 2 illustrates the position of the ECMA and shows observations of crustal thickness in several sections across this margin. These cross-sections are typical of observations from around the world at divergent plate margins.

The Baltimore Canyon and Carolina Trough cross sections (Fig. 2) show the thick igneous crust. Such margins are termed volcanic margins. Seaward of the large igneous structure the oceanic crust returns to a normal, uniform thickness (about 6 km). The volcanic margins of Figure 1 contrast sharply with the Nova Scotian Margin cross section, which represents an "avolcanic" margin, where the transition from continental to oceanic crust does not show evidence of large igneous crustal thicknesses.

There is currently a debate on the origin of the thick igneous crust, which forms a welt about 70 km wide and can be traced along the ECMA southward for ca. 2000 km along the axis of the ancient rift (Keen, 1969; Emery *et al.* 1970). One possibility is that the mantle source was excessively hot (hotter by 150 to 200 °C) during rifting and thus created a much larger volume of melt during upwelling. However, there is little evidence to indicate the former presence of such a long, thin strip of hot mantle. An alternative explanation is that small scale convection in the asthenosphere under the rift acted as a "conveyor belt" to enhance the supply of melted mantle to the lithosphere (e.g. Mutter *et al.* 1988). This suggestion is supported by the work of Anderson (1994, 1995) who proposes that the source of large igneous provinces may be the relatively shallow, low-viscosity sub-lithospheric mantle, which may undergo convection as a consequence of rift processes within the overlying plate. These two possible explanations are shown in cartoon form in Figure 1.

The thick igneous crust at volcanic margins is thought to have been created during and shortly after rifting (Austin *et al*., 1990; Sheridan *et al*., 1993). If thick igneous crust can be explained by the activation of small scale convection during the rift stage, it is evident that melt-producing convection must have been spatially limited and confined to a relatively short interval of the margin's history.

Our work involves the use of numerical models to test in a quantitative way the physical circumstances under which small scale convection would occur below a rift zone. The numerical models provide predictions of the timing and amount of melting in the mantle. This in turn predicts variations in crustal thickness that can be compared with observations. A major challenge in our modelling is to demonstrate process-driven factors which can explain both volcanic and avolcanic divergent plate margins.

Tools and Methods

The methods used build on those described in Keen and Boutilier, 1995. The models are two dimensional and solve for the temperature structure and flow velocities in the asthenosphere that result from rifting. The mantle is assumed to be a viscous fluid. The vis cosity depends in part on temperature. The solid-like lithosphere is thus simulated simply as a cold region within which viscosity is so large that flow is imperceptible during the model evolution. The viscosity depends as well on pressure, increasing slowly with depth. The viscosity also depends on strain rate (i.e. rate of change of flow). This results in a highly nonlinear behavior, where by flow reduces viscosity, leading to more and more localized flow. This in turn affects cooling rates, and creates other time dependent behavior. We must consider all of these factors in determining how realistic each model might be. These material properties are based on laboratory measurements on ultrabasic rocks and wet dunite (e.g. Kirby and Kronenberg, 1987; Chopra and Patterson, 1981).

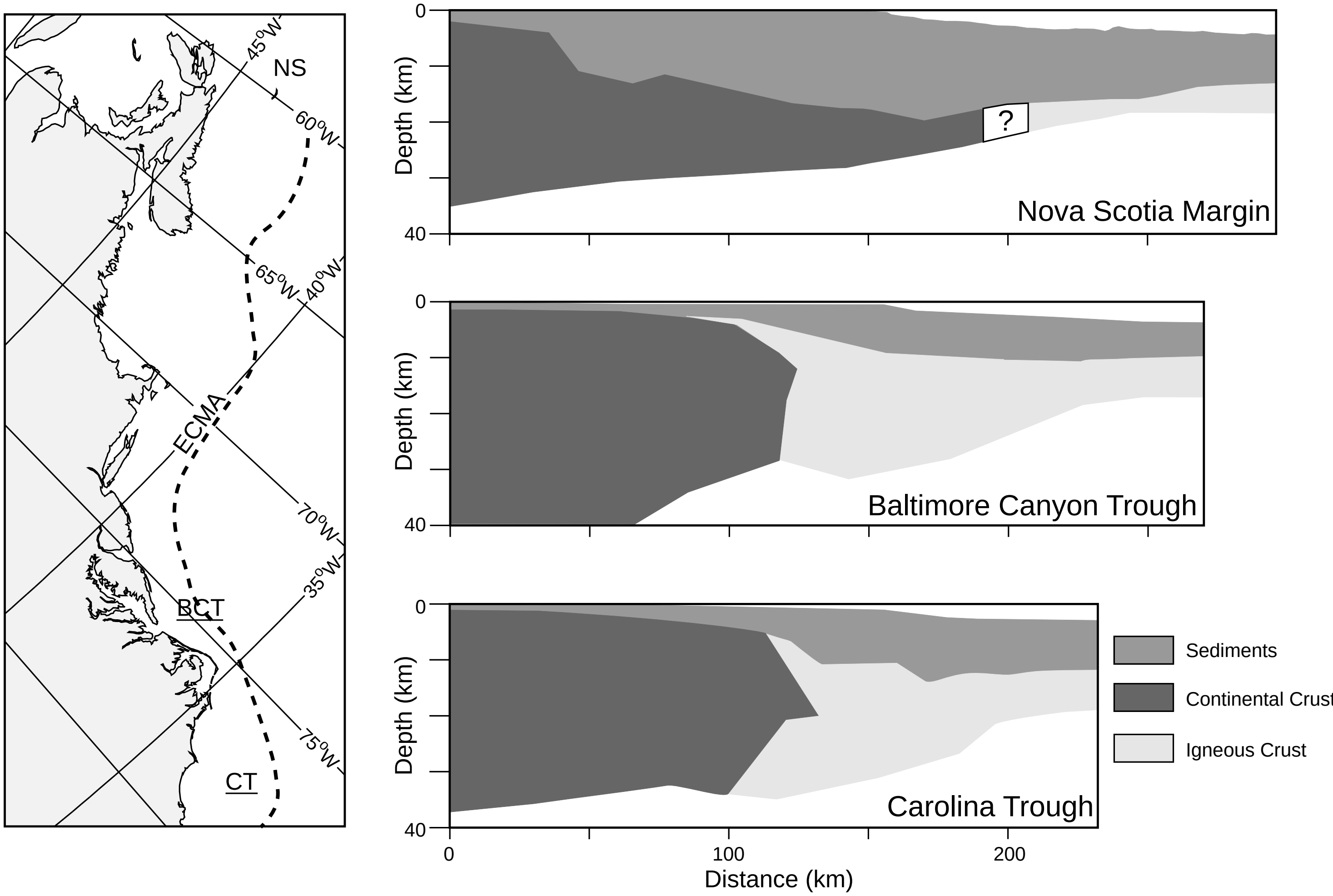

Figure 2. Position of the East Coast Magnetic Anomaly and cross sections of crustal structure at three locations: NS - Nova Scotia margin (Kay et al., 1991); BCT - Baltimore Canyon trough (Holbrook and Kelemen, 1993); CT - Carolina Trough (Holbrook et al., 1994). The ECMA is shown as a dashed line (Holbrook and Keleman, 1993) and marks the position of the 70 km wide welt of thick igneous crust below the margins. Simplified from Keen and Potter (1995).

Model calculations are performed at time steps, stepping forward in time through the continental rifting stage (see Fig.1). An upwelling zone analogous to a widening oceanic rift system is created, with the ridge axis at the upwelling centre. As the mantle rises at the rift axis, pressure is reduced. At about 80 km depth, which we call the base of the melt window, the reduced pressure allows decompression melting to occur. Part (10-20%) of the relatively solid mantle rock melts and a very low viscosity fluid is released. The melt may migrate upwards underplating or intruding the crust, or emerging on the surface through volcanism. As mantle material continues to rise, a larger fraction is able to melt. We used a compilation of laboratory measurements to determine the solidus for this decompression melting process (McKenzie and Bickle, 1988) and from this are able to calculate the volume of melt which would be produced by our models. Mantle material that flows upwards through the base of the melt window, provided it is sufficiently hot, can deliver melt. Increasing the flow rate increases the melt delivered, as does increasing the temperature of the mantle flowing upwards. Cooling of the uprising material by conduction tends to reduce melt delivery.

Evolution of a Model

Figure 3 shows the evolution of one model at four different times and illustrates most of the physical characteristics of our models. The extension and spreading rate (V0) are held constant in this model. The flow in the mantle throughout the evolution is highly time dependent; the frames were chosen at times when the velocities in the asthenosphere reached maximum values.

In Figure 3a, almost 7 Ma after the initiation of rifting, the flow in the asthenosphere is well behaved and directed into the developing rift. A thermal anomaly with a half-width of about 100 km has developed. Since the upward velocities are relatively high, and there is little concurrent cooling, the melt which is delivered to the surface is approximately equal to that needed to form normal oceanic crust (6 to 8 km).

To increase the melt delivered to levels approaching those observed at volcanic margins (about 20 km), the upward flow through the base of the melt window would have to be increased. Convection at this stage might provide that increase, but the strongly focused upward flow does not allow enough room at this stage for convection to develop. Were the viscosity to be

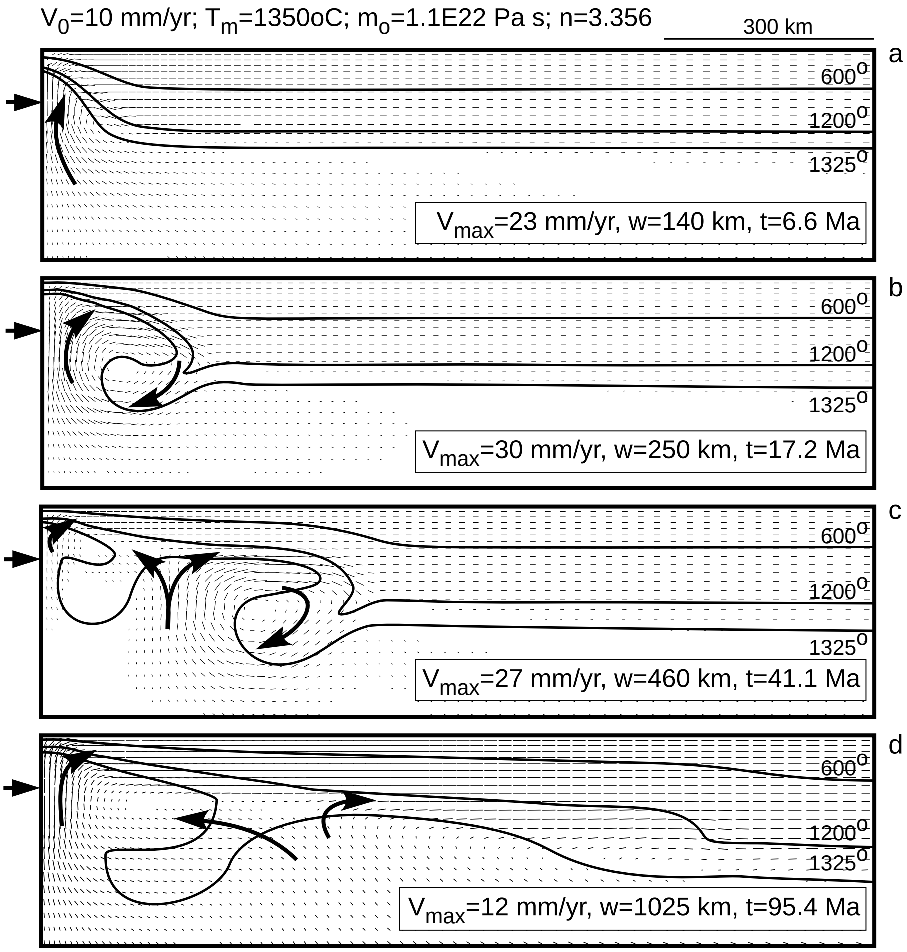

Figure 3. The evolution of one model shown at four times. The initial half-width of the rift zone for this model is about 70 km. In each frame bold arrows indicate the general direction of asthenospheric flow, with the short lines being the instantaneous direction of flow calculated by the numerical program. Selected temperature isotherms showing the perturbed thermal structure are indicated. At the top of the figure the extensional velocity, or spreading rate, the mantle temperature, a characteristic viscosity and degree of non-linearity are indicated. In each frame are indicated the maximum velocity of flow in the asthenosphere, the distance from the ridge axis to the continental margin, and the time since the beginning of rifting. Flow line vectors are scaled relative to the maximum velocity value. On each frame, a small arrow on the left edge indicates the base of the melt window (corresponding to about 80 km depth). Only flow penetrating upwards through this depth can contribute the melt which forms the crust.

lower or the thermal anomaly sharper, there might be a convective response.

Figure 3b shows the configuration at 17.2 Ma after the initiation of rifting. Here, convection has begun adjacent to the widening margin. A mass of colder material under the margin is falling, but this relatively cold material is also being entrained in the flow of the convection cell; the strong entraining flow is due in part to convection but is also driven by the need for mass to flow into the ridge axis. Thus, the two phenomena are linked by geometry.

The maximum flow velocity Vmax at t=17.2 Ma is 30 mm/yr, which is 30% higher than that at t=6.6 Ma. The flow impinges on the base of the melt window. Thus we expect more melt to be delivered at this time. However, if the colder material from the convection cell under the old margin were to get entrained into the upwelling at the ridge axis, the drop in temperature alone would decrease the melt delivered. Additional reduction in melt delivery can occur if material that had already experienced partial melting were re-circulated in like manner. The melt delivery calculations presented here include the first effect, but not the second.

Figure 3c and 3d show the continued evolution of the model. The convective response continues under the old margin but at a reduced intensity (the thermal gradient under the old margin is fading away), and the convective flow is now far enough away from the ridge axis to be relatively independent. Some interaction is still visible, however. Note that the convection is occurring just below the base of the melt window, so that no additional igneous activity will result. Additionally, cold structures resulting from convection are still present in the model and are clearly interacting with the ridge axis flow. However, it can be seen that at the ridge axis the flow evolves to a laminar form; this model will thus generate uniform oceanic crust.

We have run many other models with different parameters. They allow us to verify the following general statements. (1) The driving force for convection is the lateral thermal gradients near the base of the lithosphere, generated by rifting in the plate. Stronger gradients will produce more vigorous convection. (2) When the viscosity is relatively higher (or asthenospheric temperature lower), the convective response is lessened or suppressed. In the opposite case, more vigorous convection occurs, the time dependence of the system increases, and more complex interactions result. (3) Possible flow behaviors that are more nonlinear (i.e. have a stronger relationship between viscosity and strain rate) provide models that are less stable and more time dependent. Highly nonlinear models can convect vigorously for an interval of time and then stop, allowing the system to revert to a stable form. This might explain volcanic margins. However, there is a tendency for these highly nonlinear models to be unstable, thus preventing the generation of uniformly thick oceanic crust.

Figure 4 shows two typical crustal cross sections from our model calculations. Figure 4a shows the results from the model shown in Figure 3. At the old margin (900 km in the figure) the Moho has been thinned by extension at the early stages of the model. The melt that is first delivered is deposited there, and subsequent delivery of melt continues to build the oceanic crust.

The initial delivery of melt is irregular and time dependent, reflecting the convection that began in the early stages of the model. There is then a stage of relatively thin oceanic crust (the middle third) that reflects the presence of colder material rising to the ridge axis through time, resulting from the convection. This effect is seen in many models that were run, often consisting of a period of enhanced melt delivery followed by a period of relative melt starvation. In most cases this was used as a criterion to reject the model.

The long-term oceanic crust created in the model Figure 4a appears stable and of approximately the required thickness for oceanic crust (although somewhat more variable in thickness than desirable). We tentatively find this model acceptable, and suggest that it may approximate processes at an avolcanic divergent margin.

The bottom frame depicts our most successful model thus far for a volcanic margin. The model has the same extension rate as the model in Figure 4a but the rift geometry was localized to a 40 km half-width, creating asharp break in the lithosphere and a strong thermal gradient in the early rift stage. The model successfully provided a narrow wedge of melt over 20 km thick adjacent to the margin, and provides more or less uniform oceanic crust thereafter. The blip in the oceanic crust at the mid-left is problematic and suggests that a slightly more stable model is required.

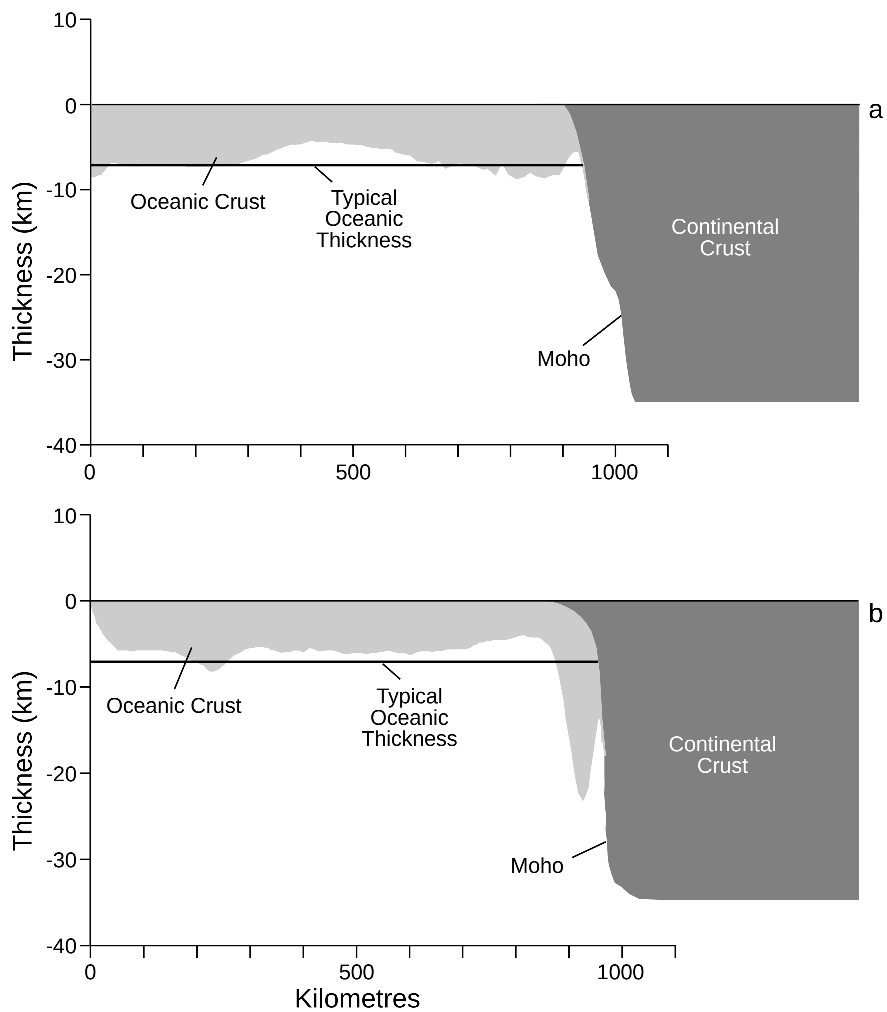

Figure 4. Predictions of oceanic crustal thickness for two models. Note the Moho near 1000 km in each frame, which was thinned by extension prior to the onset of sea-floor spreading. (a) is the melt delivery from the model shown in Figure 3 which had a relatively wide rift zone within the lithosphere. (b) is the melt delivery from a model with a sharp rift zone. At the top of each frame the mantle temperature, a characteristic viscosity and degree of non-linearity used for the model are indicated. Vertical exaggeration is approximately 16:1.

There remain significant problems with this otherwise encouraging result. The main problem is that the mantle physical properties of the two models shown in Figure 4 are too dissimilar (in background temperature, and in degree of nonlinearity). We are searching for a physical model that has just one viscosity relationship (i.e. dependence on pressure, temperature and strain rate, and characteristic viscosity), where natural variations in spreading rate and mantle temperatures produce acceptable results. The volcanic margin result, for example, does not provide this; slowing the spreading rate by a factor of two produces a model (not shown) whose melt delivery was unacceptably irregular. Changing the initial asthenospheric temperature by 25 °C also produced an unacceptable result. Since such variations obviously occur in the Earth, we conclude that we need to search further for an acceptable model. That model will work for both volcanic and avolcanic margins.

Discussion

These results are the first to quantitatively test the hypothesis that small scale convection in the asthenosphere can arise from plate divergence in the overlying lithosphere. They thus hold the promise for a viable alternative to explanations which require high temperature asthenosphere as a source of large volumes of igneous rocks in some continental rifts and rifted continental margins. We hope that the difficulties presented by the results of these models can be overcome with additional work.

One likely consequence of melting is that the creation and extraction of melt from the mantle will alter the viscosity and density of the residual mantle matrix. The viscosity is likely to be reduced, creating more vigorous flow. In contrast the density of residual mantle is decreased, and the pooling of lighter mantle material near the edges of the rift zone will tend to stabilize the system and impede convective flow (Su and Buck, 1993). These are properties that have not been incorporated into our models but that may be important in enhancing up-flow and melting, while concurrently providing the necessary stability. We are currently investigating this aspect of the problem.

Small scale convection in the asthenosphere could predict other important aspects of rifted margins which are otherwise difficult to explain. One of these is the history of vertical motions, recorded in the sedimentary record. The sediments show that after formation the margin undergoes a long term subsidence due mainly to cooling of the lithosphere. Superimposed on the long term cooling are shorter period oscillations (ca 10 to 40 Ma cycle time) with amplitudes of several hundred metres. These relatively short term variations may be linked with the time dependence of convective flow, as seen, for example, in Figure 3. Acceleration or deceleration in flow will change the vertical stress field acting on the base of the plate and thereby cause a change in the elevation of the plate.

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