Is small scale convection responsible for the formation of thick igneous crust along volcanic passive margins?

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[1] *Boutilier and Keen* [1999] showed that small scale convection may produce the observed thick igneous crust along volcanic passive margins provided that a non-linear rheology is adopted and the viscosity of the upper mantle is increased suddenly after 132 km of continental crust extension. The sudden viscosity increase is justified by the melt induced dewatering of the upper mantle rocks. Through this comparative study, it is shown that inclusion of the pressure of the newly formed igneous crust into the melting model, and/or adopting a more realistic viscosity law, significantly reduces the thickness of igneous crust produced by Boutilier and Keen's models. Furthermore, it is shown that melt migration would narrow the width of the thick igneous crust. Results of this study show that small scale convection may not play a significant role in the formation of the observed thick igneous crust along volcanic passive margins. *INDEX TERMS:* 8109 Tectonophysics: Continental tectonics—extensional (0905); 8159 Evolution of the Earth: Rheology—crust and lithosphere; 8120 Tectonophysics: Dynamics of lithosphere and mantle—general; 3230 Mathematical Geophysics: Numerical solutions; 3220 Mathematical Geophysics: Nonlinear dynamics

1. Introduction

[2] Evolution of continental rifting to sea-floor spreading in volcanic passive margins is accompanied by formation of thick igneous rocks within a narrow ocean-continent transition zone (50–100 km). Unusually thick igneous crust of 20–30 km has been detected beneath the narrow ocean-continent transition zone of volcanic passive margins [e.g., *White and McKenzie*, 1989; *White*, 1992]. Northern North Atlantic and U.S. East Coast Margin are among the typical examples of volcanic passive margins where during a short period of 2–3 m.y. over a few million km^3 of igneous rocks was produced [*White et al.*, 1987; *Holbrook*, 1993]. The sheer volume of igneous rocks rivals that of continental flood basalts and thus is considered to be in the category of the large igneous provinces [*Coffin and Eldholm*, 1994].

[3] During the last two decades, there has been many efforts to explain the observation of the thick igneous crust along volcanic passive margins. One school of thought considers an anomalously hot upper mantle as a responsible mechanism for formation of the thick igneous crust. The source of such temperature anomaly is postulated to be either related to the presence of hot mantle plumes [*White and McKenzie*, 1989; *White*, 1988] or the thermal gradient created by insulating effect of the large continents [*Anderson*, 1982, 1992; *Gurnis*, 1988]. The other school of thought holds the small scale convection happening at the initial stages of continental rifting responsible for the formation of the thick igneous crust [*Mutter and Wernicke*, 1988; *Zehnder et al.*, 1990]. *Mutter and Wernicke* [1988] hypothesized that the initial sharp thermal contrast between the rifting thinned continent and the asthenosphere induces small scale convection that increases the flux of material crossing the solidus and in turn increases the total volume of melt. By progress of the sea-floor spreading and cooling of the rifted continent, the sharp thermal contrast and its associated small scale convection fades away after initiation of sea-floor spreading, and thus a normal oceanic crust forms afterward. Recently, *Boutilier and Keen* [1999] (hereafter called B&K) tested the small-scale convection scenario and showed that considering a non-linear rheology leads to a small scale convection which successfully produces the observed thick igneous crust with a width close to that of observed. In their successful model (e.g., following the authors I call the successful model the wet model) the non-linear rheology also depends on melting. The viscosity of the 80 km of the upper mantle is increased suddenly after 132 km of continental crust extension as an attempt to simulate the melt induced viscosity increase of the upper mantle rocks. The authors ignore the effect of melt depletion buoyancy and pressure of the newly formed igneous crust. In computing the thickness of igneous crust, the authors assume that melt is instantly underplated directly above where it is formed.

[4] In this paper, I will report the results of a comparative study of my models with B&K's. By investigating the effect of the melt depletion buoyancy, the pressure of newly formed igneous crust and a more realistic melt-dependent rheological model on the volume of melt production, I will show that some of the assumptions used in B&K's models are unrealistic. As it will be shown later, considering melt depletion buoyancy for the upper mantle rocks will not significantly change the results obtained by the authors. However, including the effect of the pressure of the newly formed igneous crust and/or adopting of a more realistic viscosity law will significantly reduces the thickness of the igneous crust. The new viscosity law simulates the melt induced viscosity increase of the solid upper mantle rocks by a gradual increase of viscosity upon progress of melting. Finally, I show that ignoring melt migration in the partially molten upper mantle plays an important role in redistributing melt production to a wider area and thus avoiding creation of a localized narrow thick oceanic crust in the B&K wet model.

2. Numerical Model

[5] I model the corner flow induced by continental rifting in a two dimensional (2D) computational frame of 2100 km wide and 700 km deep. My base model has the same settings of the B&K wet model. The top boundary moves horizontally with a specified spreading velocity of 5 km/m.y. in all of my models. The bottom boundary is free-slip. Mass is allowed to only enter/exit horizontally from the right boundary. Temperature is fixed at the top and bottom boundaries to zero and 1350°C, respectively. The initially assigned temperature field changes linearly within the continental lithosphere and is constant at depths beneath. Viscosity depends on strain rate ($\dot{\varepsilon}$), temperature (T) and pressure (P) according to following equation of state,

$$\eta_s = A\dot{\varepsilon}_r^{(1-n)/n}\exp\left(\frac{E+PV}{R(T+273)}\right) \qquad (1)$$

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Table 1 Physical Parameters Used in this Paper

Parameter	Value
Mantle density (ρ_s^o)	3300 kg m^{-3}
Reference shear viscosity (η_o)	1.29×10^{19} Pa s
Viscosity activation energy (E)	444 *kJ*/mol
Viscosity activation volume (V)	15×10^{-6} m^3/mol
Viscosity strain rate component (n)	3.356
Thermal diffusivity (κ)	8.0×10^{-7} m^2/*s*
Latent heat of melting (H)	$1.2 \times 10^9 J$ m^{-3}
Thermal expansion coefficient (α)	3.2×10^{-5} °C^{-1}
Melting expansion coefficient (β)	0.024[a]

[a] *Sparks and Parmentier*, 1994.

values of E, V and n are the same as those B&K' models (Table 1). A reference strain rate ($\dot{\varepsilon}_r$) of 1.67×10^{-17} and a reference viscosity of 3.6×10^{23} are assumed at 700 km depth, the bottom of my computational domain, to calculate value of A. In order to simulate the effect of melt induced dewatering on viscosity, B&K increased viscosity by a factor of 140 at depths above 80 km when 132 km of lateral extension is completed. Following B&K, the initial thickness of continental lithosphere is reduced by an extension factor of five over a narrow region of 16 km at the rift axis. This is an attempt to simulate a sharp initial necking of continental lithosphere.

[6] An iterative finite volume method [*Patankar*, 1980] is adopted to solve equations of mass, momentum and energy. Following B&K, the melt phase is completely ignored in the governing equations. The numerical results presented in this paper are calculated in a non-uniform rectangular grid with 106 × 81 nodes with high resolution for an area of 87.5 km depth from the surface and 262 km width from the ridge axis, where melting usually occurs. Vertical grid-spacing within the high resolution area is 1.8 km and varies between about 6 to 27 km outside of the region. Horizontal grid-spacing within the high-resolution area varies between 1.5 to 4.2 km and increases from 16 km to a maximum of 88 km outside the high-resolution area. The smooth horizontal and vertical gradients of velocity and temperature fields at the edge and outside of the high-resolution area justifies the use of the rather large grid-spacings within the low resolution area.

3. Melting Model

[7] I use a fractional melting model [*Ghods and Arkani-Hamed*, 2000] where melt is extracted completely. The fractional melting model is a modified version of *McKenzie and Bickle*'s [1988] batch melting model and is constructed by replacing the liquidus temperature (T_l^o) of peridotite by the liquidus temperature of pure forsterite (i.e., $T_l^o = 2164 + 4.77\ P_H$, [*Presnall and Walter*, 1993]). Furthermore, to account for the dynamic nature of melting, the depletion field $\mathcal{M}$ is advected in each time step using the following pure advective transport equation,

$$\frac{\partial \mathcal{M}}{\partial t} + V \cdot \nabla \mathcal{M} = 0 \qquad (2)$$

where V is velocity of the upper mantle rocks. $\mathcal{M}$ quantifies the depletion of a partially molten rock and indicates the ratio of the pristine rock that should melt to produce the solid residue. The above equation is solved using monotonic second-order upwind method (*Sweby*, 1984). The fraction of melt produced in a given time step $\Delta\mathcal{M}$ is computed using the following coupled equations,

$$\rho_s C_P T_1 = \rho_s C_P T_2 + \Delta\mathcal{M} H,$$

$$\Delta\mathcal{M} = \frac{T_2 - T_s}{T_l^o - T_s}(1 - \mathcal{M}), \qquad (3)$$

By solving the above equations we can also calculate the temperature field corrected for the absorption of the latent heat of melting, T_2, and update the depletion field $\mathcal{M}$. T_1 is temperature before the correction. The value of latent heat of melting H is given in Table 1. T_s is the instantaneous solidus temperature of the solid residue at the end of the previous time step. To account for the increase of T_s through melting, T_s is updated at the end of each time step using $\mathcal{M} = \frac{T - T_s^o}{T_l^o - T_s^o}$ where $T_s{}^o$ is solidus temperature of pristine dry peridotite, and then setting $T_s = T$. The removal of the melt in a fractional melting model increases T_s because the residue becomes more refractory.

[8] There are two minor differences between B&K's melting model and that of mine. First, I correct temperature field to account for the latent heat of melting whereas B&K do not. Second, my models use a fractional melting model which corrects the original batch melting model of *McKenzie and Bickle* [1988] to take into account the effect of fractional melting. This has been accomplished by considering a liquidus curve whose temperature is generally larger than the batch melting liquidus curve used by B&K. A comparative study of the two melting models shows that the above mentioned differences have very slight effects on the results obtained by B&K's models. Therefore I use my fractional melting model when reproducing B&K' models.

4. Modeling

[9] In the first set of experiments, the effect of inclusion of melt depletion buoyancy into the B&K wet model is investigated. Addition of melt depletion buoyancy to mid-ocean ridge models increases melt production at very slow spreading rates (e.g., *Cordery and Phipps Morgan*, 1993), however melt depletion buoyancy would decrease the rate of recycling of the depleted upper mantle rocks into the melting zone of the upper mantle. Inclusion of melt buoyancy increases the peak crustal thickness by about 2 km (Figure 1). However, such increase of crustal thickness once compared with the observed thick igneous crust is marginal. This implies that the melt buoyancy can be safely ignored as was done in B&K's models.

[10] In the second set of experiments, the effect of the thickness of newly formed igneous rocks on the volume of melt production at later stages of rifting is investigated. Volcanism and underplating at early stages of rifting forms a new igneous layer at the edge of a rifting crust. Such a layer exerts extra pressure on the melt producing zone and thus may significantly reduce the thickness

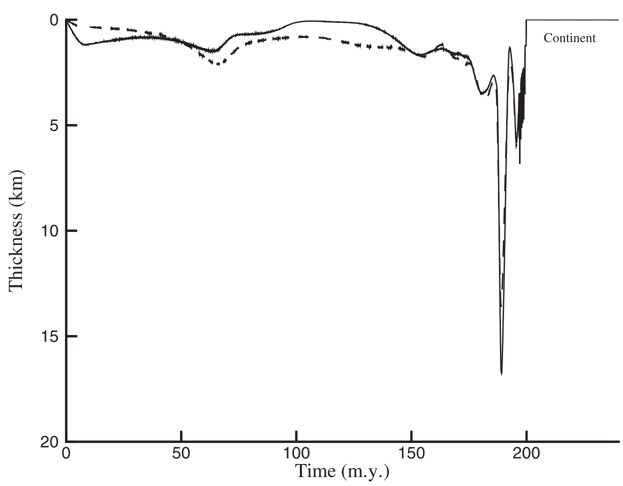

Figure 1. Crustal thickness versus time. The dashed line is the B&K wet model and the solid line is the model that consider melt depletion buoyancy. The horizontal axis is the time from the initiation of rifting.