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Melt Generation at Volcanic Continental Margins: no Need for a Mantle Plume?

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Abstract. Melt generation in a rifting environment is studied using a dynamic 2-D finite element model. The lithosphere is extended to large, realistic thinning factors assuming a mantle temperature of 1333°C. The focussing of deformation results in a distribution of thinning factors along the margin at breakup time consistent with observations. The timing of melt production (late synrift) and the amounts of melt are consistent with observations at volcanic margins. The dynamical processes related to lithospheric rifting enhance the produced melt volumes sufficiently to explain the sometimes enigmatic melt volumes found at volcanic margins.

Volcanic continental margins have been the subject of numerous modeling studies in which the large production of intrusive and extrusive melt was simulated [e.g. *McKenzie and Bickle*, 1988; *Pedersen and Ro*, 1992; *Bown and White*, 1995]. A main conclusion of most of these studies was that a higher than normal potential mantle temperature is required to explain observed melt volumes at volcanic rifted margins. A temperature anomaly in the range of 50-200°C has been suggested by e.g. *Pedersen and Skogseid* [1989], *White and McKenzie* [1989], *Bown and White* [1995].

These increased mantle temperatures are generally explained by the influence of mantle plumes [e.g. *White and McKenzie*, 1989; *Skogseid et al.*, 1992; *White*, 1992]. There are several observations that require modifications of the plume model [*King and Anderson*, 1998]. For example, some volcanic margins including the northwestern of Australia and the eastern U.S. margin [*Hopper et al.*, 1992; *Holbrook and Kelemen*, 1993], cannot be directly linked to mantle plumes or hotspots. Also, predicted plume head dimensions [*Griffiths and Campbell*, 1991; *Bijwaard and Spakman*, 1999] are not sufficient to affect entire volcanic provinces. Recently, *Anderson* [2000] proposed that these upper mantle temperature variations can also be caused by other processes, like small scale convection. Studies of small scale convection induced by either rifting or discontinuities in lithosphere thickness [*Buck*, 1986; *Anderson*, 1994; *Boutilier and Keen*, 1999; *Keen and Boutilier*, 1995; 2000; *Mutter et al.*, 1988; *King and Anderson*, 1995; 1998] indeed show that the amount of melt produced is increased substantially by an enhanced upwelling of mantle material into the melting zone.

In this study, a dynamic 2-D finite element model is used to study melt production in a rifting environment on lithospheric scale. The lithosphere is extended to breakup using various mantle temperatures and extension rates. First order predictions of melt volumes that are generated using a mantle temperature of 1333°C are consistent with observations at volcanic continental margins.

Modeling approach

2-D numerical calculations are performed using the finite element method to solve the equations for elastic-plastic non-linear viscous flow with a power-law rheology, and the heat flow equation. A linear equation of state is used for the temperature dependence of the density. Yield conditions are described by the Mohr-Coulomb criterion. Sedimentation and erosion are not included. We used the code developed by [*Huismans et al.*, 2001], and refer to them for a detailed description. Parameters used are specified in Table 1.

Melt volumes resulting from decompressional melting are calculated using the empirical relations for dry mantle peridotite of *McKenzie and Bickle* [1988]. Neither the influence of melt on the density nor on the viscosity is included in the model, which means that only first-order melt volumes are predicted. We implemented a remeshing routine which enables deformation of the lithosphere to large thinning factors. Hereby, the boundaries defining earth's surface, upper and lower crust and mantle lithosphere are maintained, so they can be tracked properly (as well as thinning factors of crust and mantle lithosphere) through time.

A layered model with different rheologies for the upper crust, lower crust and mantle lithosphere, respectively granite, diabase and olivine, is used; see Figure 1 and Table 1. The lithosphere is stretched with uniform velocities (Table 2) that fall within the range of present day plate velocities [*Argus and Heflin*, 1995], until a crustal thinning factor of 20 is reached. This thinning factor defines continental breakup in our model. It results in realistic continental crust thicknesses close to the centre of the rift zone. At this time, the stretching of the lithosphere ceases. For the influence of rheology on

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Table 1. Parameter values. From *Ranalli* [1995] and *Carter and Tsenn* [1987]. u.c.=upper crust, l.c.=lower crust, m.l.=mantle lithosphere.

parameter	u.c., l.c., m.l.
density [kgm^{-3}]	2700, 2800, 3300
density liquid basalt[1][kgm^{-3}]	2600
thermal expansion [K^{-1}]	$1{\cdot}10^{-5}$
crustal heat production [Wm^{-3}]	$1{\cdot}10^{-6}$
specific heat [Jkg^{-1}K^{-1}]	1050
conductivity [Wm^{-1} K^{-1}]	2.6, 2.6, 3.1
bulk modulus [10^{10} Pa]	3.3, 3.3, 12.5
shear modulus [10^{10} Pa]	2.0, 2.0, 6.3
powerlaw exponent n	3.3, 3.05, 3.0
activation energy Q [kJmole^{-1}]	186.5, 276, 510
material constant A [Pa^{-n} s^{-1}]	$3.16{\cdot}10^{-26}$, $3.2{\cdot}10^{-20}$, $7.0{\cdot}10^{-14}$
friction angle	30°
dilatation angle	0°
cohesion factor [Pa]	$20{\cdot}10^{6}$

[1] for calculation melt volume

rifting we refer to *Bassi* [1991]. The initial lithosphere configuration is the same for all the tests performed.

Results

The different tests are summarized in Table 2. The results of the test with v_{ext}=32 mm/yr are discussed in more detail below. The rift duration of this test before continental breakup takes place agrees best with the duration of the final stretching phase of the mid-Norwegian margin [*Skogseid et al.*, 1992]. The time evolution of the stress and temperature fields, until just prior to breakup, is shown in Figure 2. A reduction in stress amplitudes is focussed in the centre region of the domain where localisation of deformation occurs, eventually resulting in breakup in this area. The development of the asthenospheric upwelling is clearly visible in the temperature field panels. About 15 m.y. after stretching started lithospheric breakup occurs.

Thinning factors of the crust and mantle lithosphere are shown in Figure 3. Large thinning factors of the

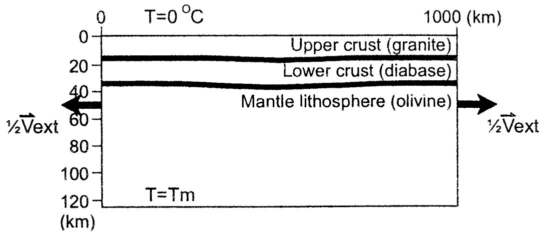

Figure 1. Initial model configuration. Crust is locally thickened with 3 km in centre of the domain to facilitate localisation of deformation. 2560 triangular, straight sided elements were used. After the breakup thinning factor is reached, $\vec{v}_{ext}$=0. The surface is free to move. Heat flow through sides of model domain is 0.

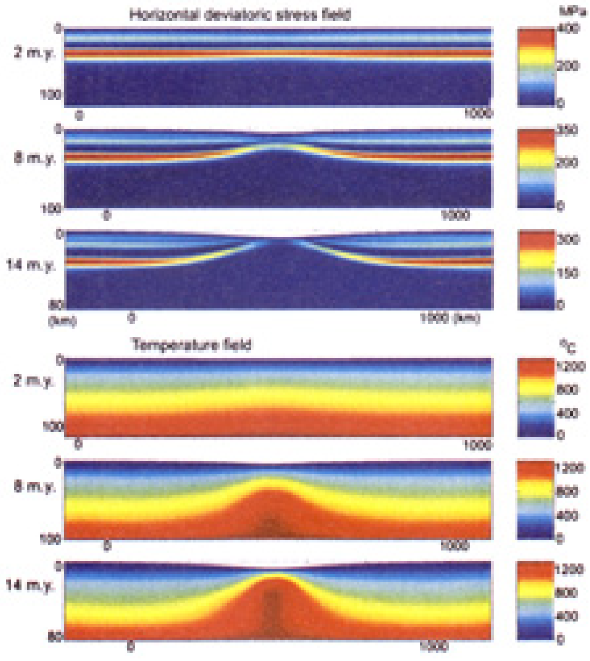

Figure 2. Evolution of horizontal deviatoric stress and temperature fields for v_{ext}=32 mm/yr. Times in m.y. after stretching started.

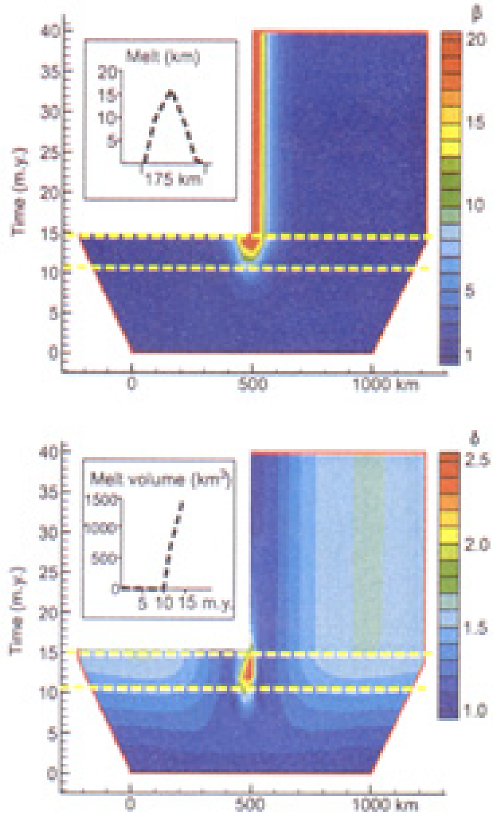

Figure 3. Thinning factors for upper and lower crust (β) and mantle lithosphere (δ) for v_{ext}= 32 mm/yr. After breakup one margin is shown. Thinning factors are defined by ratio of the original thickness of crust resp. mantle lithosphere, and their current thicknesses. Horizontal yellow broken lines bound the timewindow in which melt is produced. Upper panel inset: total melt production distribution at breakup time. Melt is generated in 175 km wide centre zone of the domain. Lower panel inset: evolution of melt production volume (per margin length).

Table 2. Predicted melt volume and rift duration to continental breakup for different extension velocities and mantle temperatures. v_{ext} [mm/yr], mantle T [°C], melt [km^3] and rift duration [m.y.].

v_{ext}	Tm	melt	duration
6	1333	0	>70
11	1333	±100	40
16	1333	±490	28
20	1333	±640	20
28	1333	±975	16
32	1333	±1400	15
16	1283	±25	30
16	1383	±1550	27

crust ($\beta \geq 5$) are restricted to a ~ 150 km wide zone. Outside of this zone lies a transition zone with a width of 200 km, in which the thinning factor decreases from 5 to 1.2. On both sides of this transition zone the crust is almost not affected by the extension, and crustal thinning factors are less than 1.2. This distribution of thinning factors is in agreement with the stretching factors presented by *Reemst and Cloetingh* [2000] and *Skogseid et al.* [2000] for several transects of the North Atlantic volcanic margin. The crustal thinning accelerates after about 10 m.y. of stretching.

Mantle lithospheric thinning is considerably less, and shows a different pattern. The mantle lithosphere is confined by the Moho and the 1300 °C isotherm. In the centre, where the upwelling is strongest, δ approaches 2.5. The lateral varying pattern of δ is explained by the difference between the timing, width and gradient of the crustal thinning, and the rise of the 1300 °C isotherm. After breakup the evolution of one margin is shown. The crustal thinning factor remains constant. For the mantle part holds that the steep geothermal gradient on the margin edge decreases, causing a reduction in δ, to values that might eventually become less than 1.

Decompressional partial melting takes place in the head of the upwelling diapir in a ~ 175 km wide zone in the centre of the domain. It starts 5 m.y. before breakup (see insets Figure 3), but almost all melt is produced in the 4 m.y. prior to breakup. This is in agreement with e.g., the Vøring volcanic margin off mid-Norway, where most melt is generated in the late synrift [*Skogseid et al.*, 1992]. In the northern North Atlantic, the time gap between the initiation of extension and the formation of partial melt has been related to the timing (late synrift) of the impingement of the Iceland-plume head on the lithosphere. However, the results presented here indicate that late synrift melting may be a direct consequence of the lithospheric deformation evolution. The calculated depth of melt production lies between ~ 20 and 50 km. At the time of breakup, a total column of melt of almost 17 km is predicted in the centre of the domain where the thinning is largest, while substantial amounts of melt are also produced on both sides of this peak (see inset in upper panel in Figure 3). Melt volumes as estimated for 4 transects in the North Atlantic Volcanic Province (transects from *Eldholm and Grue* [1994]) range from ~900 km^3 per margin length (Lofoten), ~1150 km^3 (Møre), ~1550 km^3 (Vøring) to ~1700 km^3 (Hatton Bank margin). The melt volume predicted by this test (see Table 2) falls within the range of estimated melt volumes per unit length margin and observations of other volcanic margins. Table 2 shows the dependence of melt volume on extension velocity and mantle temperature. The increasing importance of syn-rift cooling explains the reduction in melt volume at lower extension rates. Raising the mantle temperature with only 50°C increases the melt production substantially, but this raise in temperature is not needed to explain the melt volumes at e.g. the North Atlantic Volcanic Province. This result reinforces the idea that a mantle plume is not a prerequisite to explain the melt volumes at volcanic continental margins.

Conclusion

In this study the lithosphere was stretched to realistic thinning factors; the focussing of deformation resulted in a lithosphere almost unaffected by extension further away from the rift centre, a zone characterized by larger thinning factors closer to the rift centre and a very small zone with very high thinning factors where breakup occurs. The amounts of melt generated are dependent on extension velocity and mantle temperature. For the parameters that fit best the situation at the mid-Norwegian margin, the timing of melt production (late synrift) and the amounts of melt are consistent with the observations. These results were obtained using a mantle temperature of 1333°C. This modeling thus suggests that a mantle plume is not always a prerequisite to generate a volcanic margin; dynamical processes related to lithospheric rifting may enhance the produced melt volumes sufficiently to explain the sometimes enigmatic amounts of melt observed at volcanic margins.

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