Data Repository Item

Model description

Model calculations are done with the finite element code Citcom (Moresi and Gurnis, 1996; Zhong et al., 2000, van Hunen and Zhong, 2003) for an incompressible infinite Prandtl number fluid with Boussinesq approximations. Unlike most numerical codes used for rift modeling, Citcom has a Eulerian framework, which means that finite elements do not move along with the flow, but instead material migrates through the static finite element grid. Such setup is ideally suitable for modeling mantle flow with large strain accumulation without the disadvantage of remeshing procedures. Conservation of mass, momentum and thermal energy are used to describe mantle and lithosphere flow and deformation and are solved with an iterative multigrid method for mass and momentum equations, and an explicit time integration scheme for the thermal energy equation. A visco-plastic rheology was used, which combines linear, viscous Arrhenius type rheology for diffusion creep

$$\eta = \eta_0 \exp\left[\frac{(E^* + pV^*)}{R} * (\frac{1}{T} - \frac{1}{T_m})\right]$$

with a stress-dependent pseudo-plastic rheology for Byerlee's law

$$\eta_y = \frac{\mu p}{\dot{\varepsilon}_{II}},$$

with $\eta_0 = 9.78 \cdot 10^{20}$ Pa s the reference viscosity at reference mantle temperature T_m (1350°C) and reference depth z_r=400 km, E^*=360 kJ/mol the activation energy, V*=5 cm^3/yr the activation volume, $p = \rho g z$ the lithostatic pressure (with $\rho = 3300$ kg/m^3 and g=9.8 m/s^2), R the universal gas constant, T the ambient temperature, μ=0.7 is the friction coefficient, and $\dot{\varepsilon}_{II}$ the 2nd invariant of the strain rate tensor. The effective viscosity η_{eff} is defined as the minimum of the diffusion creep viscosity η and the pseudo-plastic 'yield viscosity' η_y. η_{eff} is nonlinear (i.e. stress-dependent) due to the feedback of strain rate $\dot{\varepsilon}_{II}$ into the effective viscosity. A Rayleigh number Ra=10^5 is used in all calculations. Most model parameter values are adopted from van Hunen and Zhong (2003). E^* and V* are taken from Karato and Wu (1993). E* determines how quickly viscosity decreases with increasing temperature, and controls the dynamics of thermal boundary layer instabilities (Davaille and Jaupart, 1994). Van Hunen et al. (2005) studied the influence of a lower activation energy on the stability of the thermal boundary layer and found that lower E*

values increase thermal-mechanical erosion of the lithosphere. We performed a test where E* is lowered to study the effect on the high velocity structure below the Great Plains. This test showed an increase in downwelling of lithospheric material below the edge of the Great Plains and, as a result, a higher seismic wave velocity amplitude of this structure.

Rifting is to first order a 2-D process, and variations in the third dimension are of minor importance. In the specific case of our RGR model, the rift extends over more than 1000 km in the rift-parallel direction, and the tip of the rift is located in Colorado, hundreds of kilometers to the north of the area of interest. Therefore, we expect that also in our study the 3-D effects are of minor importance, and we used a 2-D model setup. Thermal boundary conditions of our vertical 2-D model are 0^{o}C temperature at the surface, 1350^{o}C at the bottom of the model domain, and a zero diffusive heat flux ($\frac{\partial T}{\partial x} = 0$) at the side boundaries. The surface is prescribed as a free-slip, impermeable boundary, the bottom boundary has a vertical flow prescribed with zero normal stress (also sometimes referred to as 'developed' flow boundary conditions), and side boundaries have an imposed spreading velocity at the surface that tapers to zero linearly towards their bottom. The velocity boundary is tapered for numerical convenience. We also performed the same tests with a constant velocity-depth distribution and found that this did not influence the modeling results. Due to the presence of a low viscosity asthenosphere, imposed velocities below the lithosphere are ineffective at driving any of the mantle flow. The small-scale convection in the models is caused by the step in lithosphere thickness (edge-driven convection) and local thinning of the lithosphere in the rift zone (convection cells within the rift). A no-stretching test in which we prescribed velocity boundary conditions with amplitude zero also resulted in edge driven convection around the step in lithosphere thickness.

All calculations are performed with a 1600 km wide model domain, covered by a grid of 64 vertical and 256 horizontal finite elements, with grid refinement in the area of interest (usually in the center of the domain).

Seismic conversion method

The method we use to convert a given thermal structure as a function of pressure/depth and composition into seismic velocity and density consists of three components: (1) calculation of the phase diagram which provides the mineral proportions and mineral composition, accounting for Fe, Al, Ca content as a function of pressure and temperature, (2) calculation of the mineral elastic parameters (K and G) and density at the desired pressure-temperature (P-T) conditions, and averaging of the mineral properties according to their proportions from step 1, (3) addition of the dispersive effects on velocity of anelasticity.

As an update (Cobden et al., in prep) of to our previous approach (Goes et al., 2000, Cammarano et al., 2003, Goes et al., 2004), we now use for step 1 the Gibbs free energy minimization program of James Connolly (Connolly, 1990; Connolly and Petrini, 2002; www.perplex.ethz.ch) for the Ca-Mg-Fe-Al-Si system, with the mantle mineral parameter database from Stixrude and Lithgow-Bertelloni (2005), which is appropriate down to 400 km depth. The compositions we use are given in Table DR1. The CMFAS system is a good approximation for pyrolitic and more melt-depleted compositions like harzburgite and those of subcontinental lithosphere. For eclogite it should be taken as a first-order approximation only, as other components like sodium may have a non-negligible influence on its properties.

Table DR1. Compositions in oxides in volume %

composition	MgO	FeO	CaO	Al_2O3	SiO2	Reference
Pyrolite	48.53	5.72	3.50	1.80	38.66	*Sun*, 1982
harzburgite	57.42	5.44	0.44	0.48	36.22	*Irifune & Ringwood*, 1987
Eclogite	16.31	6.70	14.58	9.88	52.54	*Perrillat, et al.*, 2006
Archon	56.48	4.46	0.52	0.49	38.05	*Griffin et al.*, 1999

For step 2, we use a third-order adiabatic finite-strain equation to extrapolate in pressure, plus a Grüneisen thermal extrapolation, with a Voigt-Reuss-Hill averaging scheme to combine mineral properties, as was done in Goes et al. (2000, 2004). Mineral parameters are from the compilation of Cammarano et al. (2003).

For step 3, we add anelasticity as in Cammarano et al. (2003), Goes et al. (2000, 2004). The expression for Q is $Q_S(T,z) = Q_0 \cdot \omega^a \cdot \exp(agT_m(z)/T)$. We use the parameters in Table DR2, which give a temperature-dependence similar to model Q_1 that provided compatible temperature estimates from V_P and V_S at lithospheric depths (Goes et al., 2000), but use the parameterization in terms of melting temperature to account for the pressure dependence. This Q model does quite well at reproducing the average velocity-depth gradients beneath the study region (Fig. 3) down to about 250 km, but deeper gradients are somewhat lower than those found seismically. Other Q parameters may be able to also reproduce the deeper gradients (Faul & Jackson, 2004), or compositional heterogeneity may be required (Cammarano & Romanowicz, 2007; Cobden et al., in prep).

Table DR2. Anelasticity model

depth	Q0	a	ω	g	Q_K
0-400 km	0.1	0.15	0.02 Hz	40	1000

Supplemental References

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