**Text to Electronic archive, representing supplementary data for the article “Crust first/mantle second and mantle first/crust second lithospheric breakup scenarios along the Indian margins” by Nemčok et al.**

***Finite-element modeling***

Finite-element modeling (**Fig. 4**) was used to visualize the thermal control of the fault development in the last layer rheology.

It uses coupled thermo-mechanical 2D ANSYS® code (Ansys Inc., Houston, USA) for a system of 3˟3˟3 scenarios including:

1. initial crustal thicknesses of 30, 40 and 50 km;
2. initial surface heat flow values of 40, 60 and 80 mWm-2; and
3. extension rates of 10, 30 and 50 mmy-1.

All 27 models were done for 200 km long three-layer models representing the upper crust, lower crust and the mechanically strong portion of the lithospheric mantle. Their specific thermo-physical properties were temperature-dependent (Zoth and Hänel, 1988; Clauser and Huenges, 1995). Properties include density, thermal conductivity, specific heat and radiogenic heat production. Depth of the mantle base varied among models, being controlled by the brittle-ductile transition zone. Crustal input included radiogenic heat.

Thermal conductivity variations with temperature were described by (Clauser and Huenges, 1995):

k(T) = A + B/(350+T), (1)

where A and B are lithology-dependent empirical constants (e.g., A= 0.64, B=807 and A=1.18, B=474 for felsic and mafic rocks, in the case of crustal rocks (Clauser and Huenges, 1995)).

Density variations with temperature were described by:

(T) = (T0) / (1 + (T-T0)), (2)

where (T0) is the density at the reference temperature T0 = 273 K and  = 3 x 10-5 K-1 is the volumetric thermal expansion coefficient.

Lagrangian formulation was used to calculate the 2D temperature field changing with time. The heat transport by conduction and advection is described as (Turcotte and Schubert, 2002):

⋅c {∂T/∂t + vx∂T/∂x + vy∂T/∂y} = ∂2T/∂x2 + ∂2T/∂y2 + A, (3)

where x, y are the coordinate directions, vx, vy are the displacements in coordinate directions, T is the temperature, t is the time, is the density, c is the specific heat,  is the thermal diffusivity expressed as = k/c, k is the thermal conductivity and A is the radiogenic heat production.

The Lagrangian approach was applied to simulate the 2D, plane strain deformation using mainly four-node isotropic elements to represent the individual lithological layers. Their mechanical behavior in the elastic domain was described by the generalized Hooke’s law, relating strains ε to stresses σ via Young’s modulus E and Poisson’s ratio ν under plane strain conditions (ε2 = 0) according to (Turcotte and Schubert, 2002):

  (4)

  (5)

The Drucker-Prager yield (failure) criterion was used (Drucker and Prager, 1952; Fjaer et al., 2008). It is an expansion of the von Mises yield criterion to account for the pressure- (mean or hydrostatic pressure) dependence of the yield surface and is frequently used for rocks where normal and shear stresses can result in material failure. Expressed in terms of cohesion C and angle of internal friction ϕ, the Drucker-Prager yield criterion can be written as:

 (6)

with

  (7)

and

  (8)

It thus represents a smooth version of the Mohr-Coulomb failure criterion, i.e. the Drucker-Prager yield surface inscribes the Mohr-Coulomb yield surface.

The spatial and temporal temperature distribution derived from the aforementioned approach was used to infer information about the ductility of the lower crust. The rheology of the continental lower crustal rocks was assumed to be governed by power-law dislocation creep (Ranalli, 1995; Kohlstedt et al., 1997). The corresponding material law was expressed in terms of strain rate according to:

ne(-Q/RT), (9)

where  is the strain rate,  is the stress, n is the stress exponent, Q is the creep activation energy, R is the universal gas constant and T is the temperature.

In the following step, an effective viscosity was calculated according to (Williams and Richardson, 1991; Turcotte and Schubert, 2002):

eff = σ/2 (10)

where eff is the effective viscosity.

The stress in equation (10) is considered as differential stress σ1 – σ3, which is provided by the numerical model and varies with depth.

1018 Pa s as threshold value was assumed for the effective viscosity of the lower crust to deform at appreciable geological strain rates, justified in previous studies (Henk and Nemčok, 2016). Our intention was to identify the lower crustal regions, which are most likely to undergo a ductile flow.

The prescription of the crustal normal fault was done using contact elements, defined at opposite sides of the prescribed fault. This approach helped to monitor large differential movements of juxtaposed blocks without description of the fault propagation. Contact elements were given stiffness values, in a way similar to the Young’s moduli of the contacting rocks, to enforce compatibility between juxtaposed blocks. A single listric normal fault deforming the upper crust was detached along the top of the lower crust.

Boundary conditions of the thermal model were:

1. surface temperature at the surface; and
2. fixed heat flow at lithospheric base.

Boundary conditions for the mechanical model were:

1. stresses defined by displacements;
2. isostasy;
3. gravity forces; and
4. a load of sedimentary fill of the developing half-graben.

**References**

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