S. V. Sobolev and A. Y. Babeyko, What drives orogeny in the Andes?

**Supplementary data**

### Model details

#### Basic equations

The deformation process is modelled by numerical integration of the fully coupled system of 2-D conservation equations for momentum (Eq. 1), mass (Eq. 2) and energy (Eq. 3). These equations are solved together with rheological relations (Eq. 4-5) including those for a Maxwell visco-elastic body with temperature and stress dependent viscosity (Eq. 4), and a Mohr-Coulomb failure criterion with non-associated (zero dilation angle) shear flow potential (Eq. 5). It is assumed that viscous deformation consists of competing dislocation, diffusion and Peierls creep mechanisms (Kameyama et al., 1999).

\[
- \frac{\partial p}{\partial x_i} + \frac{\partial \tau_{ij}}{\partial x_j} + \rho g_i = 0, \quad i = 1, 2, \tag{1}
\]

\[
\frac{1}{K} \frac{\partial \rho}{\partial t} + \frac{\partial \rho v_i}{\partial x_i} = \frac{\partial n_i}{\partial x_i}, \tag{2}
\]

\[
\rho C_v \frac{d T}{d t} = \frac{\partial}{\partial x_i} \left( \lambda (T, T) \frac{\partial T}{\partial x_i} \right) + \tau_{ij} \varepsilon_{ij} + \rho A \tag{3}
\]

\[
\frac{1}{2G} \frac{\partial \tau_{xx}}{d t} + \frac{1}{2\eta} \tau_{xx} = \varepsilon_{xx} + \frac{1}{\eta_d (\tau_d, T)} \frac{1}{\eta_v (\tau_v, T)} \frac{1}{\eta_c (\tau_c, T)} \tag{4}
\]

\[
\sigma_i - \sigma_\sigma \frac{1 + \sin \phi}{1 - \sin \phi} + 2c \frac{(1 + \sin \phi)}{1 - \sin \phi} = 0; \quad g_i = \sigma_i - \sigma_c, \tag{5}
\]

Here the Einstein summation convention applies and \( x_i \) are coordinates, \( t \) – time, \( v_i \) – velocities, \( p \) – pressure, \( \tau_{ij} \) and \( \varepsilon_{ij} \) – stress and strain deviators, \( \frac{d}{dt} \) – convective time derivative, \( \frac{\partial \tau_{xx}}{\partial x_i} \) – Jaumann co-rotational deviatoric stress rate, \( \rho \) – density, \( g_i \) – gravity vector, \( K \) and \( G \) – bulk and shear moduli, \( \eta \) – viscosity, \( \eta_d \) – diffusion creep viscosity, \( \eta_{dis} \) – dislocation creep viscosity, \( \eta_{pie} \) – Peierls creep viscosity, \( \tau \) – square root of second invariant of stress tensor, \( R \) – gas constant, \( T \) – temperature, \( \tau_0 \) – maximal and minimal principal stresses, \( \phi \) – angle of friction, \( c \) – cohesion, \( C_p \) – heat capacity, \( \lambda \) – heat conductivity, \( A \) – radioactive heat production. Dependencies of all creep mechanisms viscosities from temperature and stress in (Eq.4) are taken from (Kameyama et al., 1999), with original (non-asymptotic) form of the Peierls creep viscosity. Material parameters are listed in Table RD1.

#### Interplate interface

The interface between the slab and the upper plate is modeled as a 12 km thick (3 finite elements, 2 elements at oceanic slab side and 1 element at continental plate side) subduction channel with plastic rheology. The yield stress is defined as the smallest from (Mohr-Coulomb) frictional stress:

\[
\tau = c + \mu \sigma_n, \tag{6}
\]

and temperature-dependent viscous shear stress (Peacock, 1996):

\[
\tau = \tau_0 \exp (- (T - T_0) / \Delta T). \tag{7}
\]

In (6,7), \( \tau \) is the square root of the second invariant of the stress tensor, \( c \) is cohesion, \( \sigma_n \) is normal stress, \( \mu \) is the subduction channel friction coefficient, \( T \) and \( T_0 \) are local and reference temperatures, \( \tau_0 \) and \( \Delta T \) are parameters. Parameters of (Eq.7), \( T_0 \), \( \Delta T \) and \( \tau_0 \), are assumed to be 400°C, 75°C and 60 MPa, respectively, close to (Peacock, 1996).

#### Gabbro-eclogite transformation

For simplicity the same gabbro-eclogite phase diagram for the oceanic crust and continental lower crust is used, calculated for the average gabbroic composition using free Gibbs energy minimization technique (Sobolev and Babeyko, 1994). Density of the eclogite (at room conditions) is 3450 kg/m³. In all models kinetic blocking temperature for the gabbro-eclogite transformation is 800°C for the oceanic crust and 700°C for the lower continental crust.
Table DR1. Material parameters.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Sediments</th>
<th>Felsic crust</th>
<th>Gabbro continent/ocean</th>
<th>Mantle lith-re of slab and shield</th>
<th>Mantle lith-re/asth-re</th>
</tr>
</thead>
<tbody>
<tr>
<td>Density, $\rho$, [kg/m$^3$]**</td>
<td>2670</td>
<td>2800</td>
<td>3000</td>
<td>3280</td>
<td>3280/3300</td>
</tr>
<tr>
<td>Thermal expansion, $\alpha$, [K$^{-1}$]**</td>
<td>3.7·10$^{-5}$</td>
<td>3.7·10$^{-5}$</td>
<td>2.7·10$^{-5}$</td>
<td>3.0·10$^{-5}$</td>
<td>3.0·10$^{-5}$</td>
</tr>
<tr>
<td>Elastic moduli, $K$, $G$, [GPa]**</td>
<td>55, 36</td>
<td>55, 36</td>
<td>63, 40</td>
<td>122, 74</td>
<td>122, 74</td>
</tr>
<tr>
<td>Heat capacity, $C_p$, [J/kg/K]**</td>
<td>1200</td>
<td>1200</td>
<td>1200</td>
<td>1200</td>
<td>1200</td>
</tr>
<tr>
<td>Heat conductivity, $\lambda$, [W/K/m]</td>
<td>2.5</td>
<td>2.5</td>
<td>2.5</td>
<td>3.3</td>
<td>3.3</td>
</tr>
<tr>
<td>Heat productivity, $A$, [mW/m$^2$]</td>
<td>1.3</td>
<td>1.3</td>
<td>0.2</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>Initial friction angle, $\phi$, [degree]</td>
<td>30</td>
<td>30</td>
<td>30</td>
<td>30</td>
<td>30</td>
</tr>
<tr>
<td>Initial cohesion, $C_h$, [MPa]</td>
<td>2</td>
<td>20</td>
<td>40</td>
<td>40</td>
<td>40</td>
</tr>
<tr>
<td>Diffusion creep, log($A$), [Pa$^{-1}$s$^{-1}$]</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-10.59</td>
<td>-10.59</td>
</tr>
<tr>
<td>Diffusion creep activation energy, [kJ/mol]</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>300</td>
<td>300</td>
</tr>
<tr>
<td>Grain size [mm] Grain size exponent</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>0.1 2.5</td>
<td>0.1 2.5</td>
</tr>
<tr>
<td>Dislocation creep, log($A$), [Pa$^{-1}$s$^{-1}$]</td>
<td>-28.0</td>
<td>-28.0</td>
<td>-15.4/-25.9</td>
<td>-16.3</td>
<td>-14.3</td>
</tr>
<tr>
<td>Dislocation creep activation energy, [kJ/mol]</td>
<td>223</td>
<td>223</td>
<td>356/485</td>
<td>535</td>
<td>515</td>
</tr>
<tr>
<td>Power law exponent, $n$</td>
<td>4.0</td>
<td>4.0</td>
<td>3.0/4.7</td>
<td>3.5</td>
<td>3.5</td>
</tr>
<tr>
<td>Peierls creep, log($A$), [Pa$^{-1}$s$^{-1}$]</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>8.5</td>
<td>8.5</td>
</tr>
<tr>
<td>Peierls creep activation energy, [kJ/mol]</td>
<td>223</td>
<td>223</td>
<td>445</td>
<td>535</td>
<td>535</td>
</tr>
</tbody>
</table>

*(Lukassen et al., 2001), ** (Sobolev and Babeyko, 1994)


Strain softening

For felsic and mafic crust continental crust we assume that cohesion and friction angles linearly decrease by factor of 3 when accumulated plastic strain changes from 1 to 2. For sediments the softening is assumed to be faster; by factor 10 when accumulated plastic strain changes from 0 to 0.5. Viscosity of all continental crustal materials is assumed to decrease by factor 10 (log linearly) at finite strain 0.5-1.0.

References

Fig. DR1. Time snapshots of the temperature (left) and density (right) distributions for the Central Andes model.