DETAILS ABOUT THE TECHNIQUE

We use a global mantle convection model (Bunge et al., 1997) in conjunction with a global model of the lithosphere (Kong and Bird, 1995) to compute plate motions consistent with shear tractions arising from realistic mantle convection calculations. The 3-D spherical mantle convection model solves the conservation equations of mass (a), momentum (b) and energy (c) for a highly viscous (Stokes) fluid in order to compute temperature and velocity throughout the mantle.

\begin{align*}
\text{a) } & \quad \text{div}(\bar{v}) = 0 \\
\text{b) } & \quad -\nabla p + \mu \nabla^2 \bar{v} - \rho g \hat{e}_r = 0 \\
\text{c) } & \quad \frac{\partial T}{\partial t} = -v \nabla T + k \nabla^2 T + H
\end{align*}

In the above equations \( \bar{v} \) = velocity, \( p \) = pressure, \( \mu \) = viscosity, \( \rho \) = density, \( g \) = gravitational acceleration, \( T \) = temperature, \( t \) = time, \( k \) = thermal conductivity and \( H \) = radiogenic heat production rate. A multigrid technique is used to solve the elliptic problem arising from the Stokes flow in a highly efficient manner. The use of local finite elements allows us furthermore to take advantage of modern parallel computers.
so that high numerical resolution can be achieved sufficient to resolve the dynamically
important lengthscales (thermal boundary layers) of convective motion in the Earth’s
mantle. Mantle buoyancy forces are derived from a history of plate motion and
subduction that spans the past 120 Myrs (Lithgow-Bertelloni and Richards, 1998).
The lithosphere model (SHELLS) is based on conservation of momentum (stress-
equilibrium) and uses non-linear material flow laws. In order to compute plate
velocities, the code solves for the momentum (d) and mass (see above) conservation
equations

\[- \rho g \hat{e}_r + \frac{\partial \sigma_{ij}}{\partial x_j} = 0\]

in the thin-sheet approximation (Bird, 1999), where the 3-D force balance is vertically
integrated along depth in order to reduce the 3-D problem to 2-D. In the above
equation \( \rho \) = density, \( g \) = gravitational acceleration and \( \sigma \) = stress. The lithosphere
model uses finite elements with a computational grid that accounts explicitly for
geological faults by means of interfaces between contact elements (figure DR1).
At each node of the computational grid, the stresses involved in the dynamics of the
lithosphere include the ones coming from large topography regions on the surface of
the upper plate (which contribute both through horizontal deviatoric stresses and vertical overburden pressure), and the shear stresses from the mantle.

No vertical shear traction is assumed on vertical planes so that vertical normal stresses are lithostatic at all nodes, and equal to the weight of the overburden per unit area. In the quasi-steady state of the lithospheric force balance considered in our formulation, elasticity contributes a negligible fraction of the strain rate in viscoelastic solutions. Elastic strain is then entirely neglected to eliminate arbitrary initial conditions and time-steps. Temperature plays an important role in defining the rheological properties of the deep crust and lithosphere; in SHELLS thermal conductivity and heat productivity are assumed to be constant laterally. Moreover the vertical heat conduction is assumed to be steady state.

Faults and plate boundaries are represented in the computational grid through contact elements. The fault dip angle is allowed to vary laterally and is constrained directly from seismological observations. The rheological properties at faults differ from the ones of continuum (intraplate) elements and are such that faults are weaker with respect to non-faulted material. Specifically at each fault three rheologic laws are evaluated: frictional (Mohr-Coulomb) faulting, dislocation and Newtonian creep. At each depth along the fault plane and depending on temperature, pressure and strain rate, the mechanism giving the lowest shear stress is the one presumed to dominate. As for the frictional regime, our tectonic model (Shells) accounts for a fault friction value
equal to 0.03, much lower than suggested by Byerlee’s low for non-faulted material (0.85), and in agreement with a number of observations suggesting low fault friction coefficients, including heatflow measurements and stress directions in transform faults. Concerning the dislocation creep regime (e), stress (\(\sigma\)) is proportional to strain rate (\(\dot{\varepsilon}\))

\[
\sigma = A \cdot \dot{\varepsilon}^{1/n} \cdot \exp\left(\frac{B + Cz}{T}\right)
\]

Where \(\sigma\) = stress, \(\dot{\varepsilon}\) = strain rate and \(T\) = temperature at depth \(z\). We use the following values for crust and mantle layers: \(A\) (Pa s\(^{1/n}\)) = 3.2\(\times\)10\(^7\) (crust) / 3.4\(\times\)10\(^4\) (lithosphere), \(B\) (activation energy in ° K) = 10048 (crust) / 21340 (lithosphere), \(C\) (° K m\(^{-1}\)) = 0 (crust) / 0.0223 (lithosphere) and \(n = 3\) (both crust and lithosphere). Once all the stresses have been computed, SHELLS determines plate velocities which equilibrate those stresses.

The link between the mantle and lithosphere model is performed by using the asthenosphere velocities derived from our MCMs as a velocity boundary condition at the base of plates in SHELLS, such that realistic mantle buoyancy forces are allowed to drive plate motion. Plate driving tractions are computed in SHELLS through a dislocation olivine creep rheology that depends on temperature, pressure and strain.
rate, where the strain rate is equal to the vertical gradient of the asthenosphere velocity pattern from our MCMs.

REFERENCES


Figure DR1. Finite element grid for the Nazca-South America region. Triangular elements are in gray, plate boundaries are in bold gray, coastline in red.