Mantle Flow, Lithospheric Structure and Surface Deformation

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Contributions to Topography

Factors:

- Isostatic balance of crust
- Orogenesis
  - short $\lambda$ uncompensated
- Epeirogeny
  - Long $\lambda$
  - Tectonic uplift; post-glacial rebound; \textit{dynamic topography} [Mitrovica \textit{et al.}, 1989; Gurnis, 1993]

\[ h = -\frac{q_r}{\delta \rho g} \]

Mantle Flow
Earth’s Geoid: Dynamic Topography

Observed Geoid (Degrees 2-15)

Predicted Geoid (Slab model, Degree 25)

Geoid (m)
Dynamic Topography: Physical Deflection

OCEAN   CONTINENT   OCEAN

Continental flooding

SUBDUCTING SLAB

410

660

CMB

CORE

UPPER MANTLE

TRANSITION ZONE

LOWER MANTLE

SEDIMENTARY BASIN

CRUST

LITHOSPHERE
Observations of Lithospheric Stress
Contributions: Mantle Stresses and Lithospheric Structure

The stress maps display the maximum horizontal compressional stress $S_H$

**Method**
- focal mechanism
- breakouts
- drill. induced frac.
- overcoring
- hydro. fractures
- geol. indicators

**Quality**
- A: $S_H$ is within $\pm 15^\circ$
- B: $S_H$ is within $\pm 20^\circ$
- C: $S_H$ is within $\pm 25^\circ$

**Stress Regime**
- Normal faulting
- Strike-slip faulting
- Thrust faulting
- Unknown regime

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**Data depth range**
- 0-40 km

**The stress maps display the maximum horizontal compressional stress $S_H$**

**Focal mechanism**
- $S_V > S_H > S_H$
- normal faulting regime

**Strike-slip regime**
- $S_H > S_V > S_H$
- thrust faulting regime

**Thrust faulting regime**
- $S_H > S_V$

[Heidbach et al., 2009, WSM release 2008]
Sources of Stress

Inhomogeneity
Topography
Edge Tractions
Basal Tractions
**Governing Equations**

**Momentum**

\[
\frac{\partial}{\partial t} (\rho v_i) + v_j \frac{\partial (\rho v_i)}{\partial x_j} = -\frac{\partial p}{\partial x_i} + \frac{\partial^2}{\partial x_i^2} (\eta_{ij} v_j) + f_i
\]

**Energy**

\[
\frac{\partial T}{\partial t} + v_i \frac{\partial T}{\partial x_i} = \kappa \frac{\partial^2 T}{\partial x_i^2} - H
\]

**Mass**

\[
\frac{\partial \rho}{\partial t} + v_i \frac{\partial \rho}{\partial x_i} = \frac{\partial (\rho v_i)}{\partial x_i}
\]

**Non-linear**

What is right Constitutive Relation?

FAULTS!

Large range of Time- & Length-Scales

\[\Delta \rho g = \rho \alpha \Delta T g\]

\[\nabla \cdot v = 0\]

[**Tackley, 1999**]
Rheology

Function of \((X, P, T, \sigma)\)
Computing Mantle Flow

CitComS Finite Element code

Internal density heterogeneity from S20RTSb

Laterally-Varying Viscosity: (T-dependence ~ age)

Lithospheric thickness from seismology.

Use tractions at base of lithosphere

[Conrad and Lithgow-Bertelloni, 2006; Naliboff et al., 2009; van Summeren et al., 2012]
Plate Driving Forces

Slab Pull from Upper Mantle Slabs
Slab Suction from Lower Mantle Slabs
Shallow Roots and Global Asthenosphere

(a) NUVEL–1A Plate Velocity

V_{\text{subd.}}/V_{\text{non-subd.}} = 3.4
V_{\text{aver}} = 3.7 \text{ cm/yr}

(b) No Asthenosphere

V_{\text{subd.}}/V_{\text{non-subd.}} = 3.2
V_{\text{aver}} = 7.1 \text{ cm/yr}
f_{\text{sp}} = 100\%

(c) Shallow Continental Roots

V_{\text{subd.}}/V_{\text{non-subd.}} = 3.7
V_{\text{aver}} = 3.6 \text{ cm/yr}
f_{\text{sp}} = 60\%

(d) Deep Continental Roots

V_{\text{subd.}}/V_{\text{non-subd.}} = 3.7
V_{\text{aver}} = 1.4 \text{ cm/yr}
f_{\text{sp}} = 20\%

[van Summeren et al., 2012]
Horizontal Traction Regimes

- **Normal**
- **Strike-slip**
- **Thrust**

**SH(max)** & **SH(min)**

[Figure from Naliboff et al., 2009]

Horizontal Traction Map

![Horizontal Traction Map]

25 MPa

**Compression**

**Extension**

**Strike-Slip**
Stresses due to Basal Traction

\[
\frac{\partial}{\partial \theta} (N_{\theta\theta} \sin \theta) + \frac{\partial N_{\theta\phi}}{\partial \phi} - N_{\phi\phi} \cos \theta + q_{\theta} R \sin \theta = 0
\]

\[
\frac{\partial}{\partial \theta} (N_{\theta\phi} \sin \theta) + \frac{\partial N_{\phi\phi}}{\partial \phi} + N_{\theta\phi} \cos \theta + q_{\phi} R \sin \theta = 0
\]

\[
N_{\theta\theta} + N_{\phi\phi} + q_{r} R = 0
\]
Radial Traction

REGIME

- Normal
- Strike-slip
- Thrust

[modified from Naliboff & Lithgow-Bertelloni, submitted]
Dynamic Uplift & Extension

Wednesday, 19/11/13
Dynamic Topography from S40RTS

[Lithgow-Bertelloni and de Koker, in revision]

[Forté et al., 2010; Moucha and Forte, 2011]
Model of lithospheric structure: TDL

Procedure

• Divide globe into regions (4 continental + oceans(age)
• Crustal structure (CRUST 2.0) + lithospheric mantle (depleted + undepleted)
  • Oceans half-space cooling based on isochrons
• Lithospheric mantle densities at P and T [Stixrude and Lithgow-Bertelloni, 2005; 2011]
• Thicknesses determined by matching spherically averaged P at 350 km to PREM

[Naliboff et al., 2012; Lithgow-Bertelloni and de Koker, in revision]
Best Guess at “Observed” i.e. Residual

Figure 1. Graphical comparison between four residual topography fields calculated independently. All four models account for the flattening of old ocean floor (“plate model”) with water-loaded oceans. Continents are air-loaded, and the mean residual topography is set to zero for comparison with dynamic topography models. In the continents, A was calculated by removing the mean continental elevation (529 m), B was obtained by removing the isostatic contribution of the crust only, updated from Steinberger (2007) using the age grids of Müller et al. (2008a), and C and D were obtained by removing the isostatic contribution of both the continental crust and the continental lithosphere. The thin and thick black lines are the coastlines and plate boundaries, respectively. In A, the white contours are the continent-ocean boundary from Müller et al. (2008a), the magenta contours are the edge of the continents (−200 m; Harrison et al., 1983), the green contours outline Phanerozoic large igneous provinces, and the yellow contours indicate continental crust thicker than 50 km. The red star is the location of well COST-B2 offshore New Jersey. Mollweide projection. Data in B, C, and D are courtesy B. Steinberger.

<table>
<thead>
<tr>
<th>Model</th>
<th>Minimum (m)</th>
<th>Maximum (m)</th>
<th>RMS* (m)</th>
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<tr>
<td>Residual topography (Fig. 1)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>A. This study</td>
<td>−1638</td>
<td>3158</td>
<td>556</td>
</tr>
<tr>
<td>B. Steinberger (2007)</td>
<td>−1557</td>
<td>2105</td>
<td>474</td>
</tr>
<tr>
<td>C. Panasyuk and Hager (2000)</td>
<td>−1704</td>
<td>1119</td>
<td>439</td>
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<tr>
<td>D. Kaban et al. (2003)</td>
<td>−2053</td>
<td>2144</td>
<td>613</td>
</tr>
<tr>
<td>Dynamic topography (Fig. 4)</td>
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<td></td>
<td></td>
</tr>
<tr>
<td>A. This study</td>
<td>−2235</td>
<td>1653</td>
<td>760</td>
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<tr>
<td>B. Steinberger (2007)</td>
<td>−2778</td>
<td>3039</td>
<td>909</td>
</tr>
<tr>
<td>C. Ricard et al. (1993)</td>
<td>−1605</td>
<td>611</td>
<td>430</td>
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<tr>
<td>D. Conrad and Husson (2009)</td>
<td>−1550</td>
<td>1450</td>
<td>480</td>
</tr>
<tr>
<td>E. Spasojevic and Gurnis (2012)</td>
<td>−3652</td>
<td>883</td>
<td>884</td>
</tr>
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*Root mean square amplitude.

[Lithgow-Bertelloni and de Koker, in revision]

[Flament et al., 2013]
Topography and Lithospheric Structure

\[
\frac{\partial \tau_{ij}}{\partial x_j} = \frac{\partial \tau_{zz}}{\partial x_i} - \frac{\partial \Omega}{\partial x_i}
\]

\[
\Omega = \sigma_{zz} = \frac{1}{L + h} \int_{-L}^{h} \sigma_{zz} \, dz = -\frac{1}{L + h} \int_{-L}^{h} \int \rho(z') \, dz'
\]

[England and McKenzie]
Effects of Lithospheric Structure

Figure 3. Variations in global mean lithostatic stress ($\Omega_1$) and the tectonic principal stresses balancing these variations for a 100 km base depth. Regions with large negative values of $\Omega_1$ often correspond with topographically high regions and are characterized by extensional principal stresses (white bars, compression $\sim$ black bars). Mean lithostatic stress and tectonic stress patterns are shown for models with isostatically adjusted (a) and TDL (b) mantle density structures. The globally averaged mean lithostatic stress values are (a) 1500 MPa and (b) 1496 MPa. To compare mean lithostatic stress values in (a) and (b) directly, a reference mean lithostatic stress value was removed from each field prior to plotting rather than subtracting the globally averaged values. The reference value was taken from a lithospheric column in the isostatically adjusted model corresponding geographically to the isostatic reference column in the TDL model.

Increasing the base depth reduces lateral variations in $\Omega_1$ thereby reducing stress magnitudes, in both continents and oceans by roughly the same amount. Increasing the base depth to 250 km (Fig. 5a) largely reproduces these trends. In TDL, increasing the base depth incorporates in the lithospheric column additional mantle density variations that have no assigned role in enforcing isostatic balance. As a result, the additional mantle in each column may drive the models towards or away from regional isostatic compensation, and increase or decrease regional mean lithostatic stress gradients. The mantle incorporated by increasing the base depth from 100 to 175 km leads to larger gradients in the mean lithostatic stress distribution (Fig. 4b), particularly across tectonic provinces where different mantle geotherms influence the density structure. Increasing $L$ from 100 to 175 km magnifies the tectonic stress magnitudes regionally while the orientations remain similar (i.e. Antarctica, Mediterranean, Ural Mountains, Western Australia) while in other regions the stress orientation is strongly modified as well (i.e. Western North America and Andes). Increasing the base depth to 250 km (Fig. 5b) generates the largest $\Omega_1$ gradients and resulting tectonic stress magnitudes despite the lowest averaged mantle density variations (Fig. 2f), where the tectonic stress field in many regions strongly deviates from the 100 km reference model and long-wavelength patterns in the world stress map. In many regions, the tectonic stress magnitudes are more than a factor of 2 larger than those in the isostatically compensated model with a 250 km base depth (Fig. 5).

3.3 Effects of strength variations within the lithosphere

Decreasing the base depth from 100 to 50 km for the TDL mantle density structure (Fig. 6) illustrates the development of large-scale stress patterns related to regions of high topography. As the base depth decreases the relative contribution of topography to the mean lithostatic stress increases, as shown in Tibet, the Western US and the Andes. The larger influence of the topographically highest regions for $L = 50$ km reveals a long-wavelength stress pattern where compressional stresses run parallel to a large percentage of the Pacific Plate boundary (Fig. 6b) and increase in magnitudes by up to a factor of 2.

To examine possible lateral variations in strength, we limit the sources of stress to the regional scale (Fig. 7). Stress fields are...
Effects of Lithospheric Structure

Isostasy enforced

TDL

[modified from Naliboff et al., 2012]
Horizontal and Radial Traction:

[Nablihoff et al. 2009; Naliboff & Lithgow-Bertelloni, submitted]
Effect of Lateral Viscosity Variations

[Naliboff et al., 2009]
Effect of Weak Asthenosphere

[Graph showing dynamic vs. residual topography with various viscosity contrasts (RTS) and global distributions (%)].

[Graph showing LAB viscosity contrast (RTS) with change in stress magnitude (MPa)].

[Naliboff and Lithgow-Bertelloni, submitted]

Structure and Dynamics of the Lithosphere/Asthenosphere System
Mantle-Lithospheric coupling INEVITABLE but VARIABLE
- horizontal mantle tractions are large..., match plate motions, but largely not stresses
- radial tractions (i.e DYNAMIC TOPOGRAPHY) determine regime and transmit efficiently

-Lithospheric structure assumptions CRUCIAL both in density and rheological structure!
- Choice of mantle density heterogeneity also matters

What do we need to do?
- Complete crustal, lithospheric structure needed
- Better representations of lithospheric and mantle rheology (crustal...)
- Temporal evolution of stress field