Crustal architecture across Southern California and its implications on San Andreas Fault development

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Coupled influence of tectonics, climate, and surface processes on landscape evolution in southwestern North America

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The Cenozoic landscape evolution in southwestern North America is ascribed to crustal isostasy, dynamic topography, or lithosphere tectonics, but their relative contributions remain controversial. Here we reconstruct landscape history since the late Eocene by investigating the interplay between mantle convection, lithosphere dynamics, climate, and surface processes using fully coupled four-dimensional numerical models. Our quantified depth-dependent strain rate and stress history within the lithosphere, under the influence of gravitational collapse and sub-lithospheric mantle flow, show that high gravitational potential energy of a mountain chain relative to a lower Colorado Plateau can explain extension directions and stress magnitudes in the belt of metamorphic core complexes during topographic collapse. Profound lithospheric weakening through heating and partial melting, following slab rollback, promoted this extensional collapse. Landscape evolution guided northeast drainage onto the Colorado Plateau during the late Eocene-late Oligocene, south-southwest drainage reversal during the late Oligocene-middle Miocene, and southwest drainage following the late Miocene.

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Crust and mantle rheology is key for linking geologic history to dynamic processes.
The role of gravitational body forces in development of metamorphic core complexes

Extensional collapse of topography and thickened crust produces metamorphic core complexes.

Crustal rheology (composition, temperature, melt) is key.
271 broadband stations from Southern California Seismic Network (CI) and 102 stations from other seismic arrays deployed in/near Southern California.

Apply 2-layer H-κ stacking method to teleseismic waves

Sui et al. (2023) GRL (under revision)
From Vp/Vs to SiO₂ wt%

By combining the Vp/Vs ratios with a high-resolution Vs model (CVM-H version 6.2), a 1-D SiO₂ wt% model beneath each station in this study can be constructed based on relationship revealed from the petrology database (Hacker et al. 2015, Christensen 1996, Fig. 4).

Sui et al., (2022)
\[ \eta_{\text{viscous}}(T, P, \dot{\varepsilon}) = f A^n \dot{\varepsilon}^{-(1-n)/n} \exp\left(\frac{E+PV}{nRT}\right), \]

\[ \eta_{\text{plastic}} = \frac{\sigma_{\text{yield}}}{2\dot{\varepsilon}} \]

Test hypothesis using a 3-D Thermomechanical Finite Element Approach for Southern California


Satisfy force-balance, conservation of energy, and conservation of mass. Also accounts for erosion and mass redistribution given a climate model.

3-D Cartesian domain with maximum lithosphere depth of 100 km, solved by 80 x 40 x 100 Eulerian nodes on a finite-element mesh with a resolution of 0.2° x 0.2° x 1km.

Top = free slip
Sides = Pacific-North America velocity
Thermal state of mantle from Shen and Ritzwoller (2016)
Thermal state of crust from Shinevar et al. (2018).
Flow law parameters applied to layers of variable thickness obtained from seismic model

<table>
<thead>
<tr>
<th>Material</th>
<th>Density (kg/m³)</th>
<th>Heat capacity (J/K/°K)</th>
<th>Heat diffusivity (m²/s)</th>
<th>Radiogenic heat production (µW/m³)</th>
<th>Flow law (dislocation creep)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Topography</td>
<td>2500</td>
<td>1000</td>
<td>$8.3 \times 10^{-7}$</td>
<td>0.5</td>
<td>Wet Quartz (Gleason and Tullis, 1995)</td>
</tr>
<tr>
<td>Oceanic crust</td>
<td>2950</td>
<td>1000</td>
<td>$1 \times 10^{-6}$</td>
<td>0.5</td>
<td>Dry Maryland Diabase (strong) (Burov, 2011)</td>
</tr>
<tr>
<td>Upper crust</td>
<td>2700</td>
<td>1000</td>
<td>$8.3 \times 10^{-7}$</td>
<td>0.9</td>
<td>Wet Quartz (Gleason and Tullis, 1995)</td>
</tr>
<tr>
<td>Middle crust</td>
<td>2900</td>
<td>1000</td>
<td>$6.7 \times 10^{-7}$</td>
<td>0.9</td>
<td>Wet Diorite (Carter and Tsen, 1987)</td>
</tr>
<tr>
<td>Lower crust</td>
<td>2900</td>
<td>1000</td>
<td>$6.3 \times 10^{-7}$</td>
<td>0.9</td>
<td>Dry Maryland Diabase (strong) (Burov, 2011)</td>
</tr>
<tr>
<td>Mantle lithosphere</td>
<td>3330</td>
<td>1000</td>
<td>$1 \times 10^{-6}$</td>
<td>0.022</td>
<td>Dry Olivine (Burov, 2011)</td>
</tr>
<tr>
<td>Air</td>
<td>1</td>
<td>100</td>
<td>$1 \times 10^{-6}$</td>
<td>0.0</td>
<td>$1.0 \times 10^{30}$ Pas</td>
</tr>
<tr>
<td>Sticky air</td>
<td>1</td>
<td>100</td>
<td>$1 \times 10^{-6}$</td>
<td>0.0</td>
<td>$5.0 \times 10^{18}$ Pas</td>
</tr>
<tr>
<td>Water</td>
<td>997</td>
<td>1000</td>
<td>$1 \times 10^{-6}$</td>
<td>0.0</td>
<td>$5.0 \times 10^{18}$ Pas</td>
</tr>
</tbody>
</table>
Strain rates are controlled by compositional anomalies, with highest crustal strain rates over the San Andreas region where there is a thick intermediate composition layer.

Zones of thicker intermediate composition are locations of predicted lower crustal flow.
Southern Salton Trough region – seismic constraints yield a thin intermediate felsic layer and thick mafic layer. Strain rates in southern Salton Trough are controlled by a thermal anomaly.

Everywhere else, strain rates controlled by composition
Hypothesis: This Rheologically Strong Peninsular Ranges forced the development of the Big Bend.

Future Directions

• We want to move in the direction of SiO₂ content as a function of depth from seismology ➔ mineral assemblages as a function of depth ➔ polyphase rock viscosities as a function of depth

Integrate with Geologic Framework – mineral assemblages, temperature model, etc.

This is on track from CRM – Montesi presentation
Boundary Conditions

Traction boundary conditions from mantle flow models (time-dependent)

Thermal boundary conditions at base of lithosphere

TX2008 Tomography Model
Simmons et al. (2009), Moucha et al. (2009)

Compted with ASPECT code Gassmoller et al., 2018; Heister et al., 2017; Kronbichler et al., 2012
The MCC mode (model 1). The model with a crustal root and non-uniform distribution of GPE at the late Eocene demonstrates the formation of a MCC. The imposed traction/uniFB01 field at the base of the lithosphere, derived from the mantle convection model (Supplementary Fig. 3), leads to the shearing and advection of materials underneath the North American lithosphere (Fig. 3). In our simulations, regions with a thickened crustal root develop weakening via conductive heating. Following this weakening, the extreme extension within zones of thick crustal welts results in free-boundary collapse of the thickened crust with substantial differential motion between brittle upper crust and weak lower crust, indicating lower crust/uniFB02 ow (Fig. 3). Based on our results, this extensional collapse is driven entirely by gravity acting on density differences created by topography of surface and Moho, along with internal density variations within the crust and upper mantle (Figs. 3 and 4).

As such, at the onset of the simulation (36 Ma), deformation is mainly concentrated in areas with high topography and thick crustal root. At this early stage of deformation, the model shows the formation of high angle (>38°) and high-strain-rate conjugate shear zones within the brittle upper crust that results in the formation of a necking zone there (Fig. 3a, g). The increase in local strain in the necking zone causes strain softening and significant plastic deformation in the upper crust (Fig. 3). Once necking has begun, it becomes the exclusive location of yielding and plastic deformation. After 2 Myr, the deformation is more concentrated at the high angle conjugate fault, and rheological contrast and body forces set up a shear zone and large offsets along the Moho (Figs. 3b, h and 4a). By 32.5 Ma, the right limb of the conjugated shear zone turns into a less-active normal fault, leading to the formation of an asymmetric shear zone with dominant slip along a low angle normal fault (Figs. 3c, i and 4b). Thinning of brittle crust below the highlands results in a reduction of the vertical load and consequently a horizontal pressure gradient at depth, which causes a lateral/uniFB02 ow of lower crustal material toward the zone of necking, initiating doming of the brittle-ductile transition (Fig. 3b, c). As such, at ~32.5 Ma, the main detachment is at a shallower dip angle (20°), and the deformation is concentrated on this asymmetric shear zone (Fig. 4b). With ongoing lower crust/uniFB02 ow and crustal extension, the strain rate continues to localize strongly on the main detachment of the asymmetric fault zone. Through time, the dip of the main detachment gradually rotates to lower angles as it approaches the surface. By

Model Set-Up for 2-D, 2.5-D, and 3-D Thermomechanical Simulations

Bahadori et al. (2022a, 2022b)
Embryonic core complexes in narrow continental rifts: The importance of low-angle normal faults in the Rio Grande rift of central New Mexico, Jason W. Ricketts, Karl E. Karlstrom, and Shari A. Kelley