3-27-2009

Along-Strike Differences in the Southern Alps of New Zealand: Consequences of Inherited Variation in Rheology

Phaedra Upton

Peter O. Koons
University of Maine - Main, peter.koons@maine.edu

Dave Craw

C. Mark Henderson

Ray Enlow

Follow this and additional works at: https://digitalcommons.library.umaine.edu/ers_facpub

Part of the Earth Sciences Commons

Repository Citation
Upton, Phaedra; Koons, Peter O.; Craw, Dave; Henderson, C. Mark; and Enlow, Ray, "Along-Strike Differences in the Southern Alps of New Zealand: Consequences of Inherited Variation in Rheology" (2009). Earth Science Faculty Scholarship. 9.
https://digitalcommons.library.umaine.edu/ers_facpub/9

This Article is brought to you for free and open access by DigitalCommons@UMaine. It has been accepted for inclusion in Earth Science Faculty Scholarship by an authorized administrator of DigitalCommons@UMaine. For more information, please contact um.library.technical.services@maine.edu.
Along-strike differences in the Southern Alps of New Zealand:
Consequences of inherited variation in rheology

Phaedra Upton,1,2,3 Peter O. Koons,1 Dave Craw,2 C. Mark Henderson,4 and Ray Enlow5

Received 1 July 2008; revised 11 December 2008; accepted 30 January 2009; published 27 March 2009.

[1] Long- and short-term strain variations along the Australian-Pacific plate boundary through the South Island of New Zealand, including a 300% increase in orogen width, coexistence of oblique thrusting on orthogonal structures, and variability in the locus of orogenic gold deposits, coincide with rheologically relevant geological variation. Our model investigates the consequences of thin, strong lower crust in the north and thick, weak lower crust in the south. Solution of the full 3-D mechanical equations reproduces the larger wavelength strain patterns of the orogen. A 3-D perturbation-based analytical solution leads to the identification of the sensitivity of displacement type to minor stress changes. Transition from boundary-normal thrusting to boundary-parallel thrusting occurs at the transition from strong to weak lower crust and is related to an increase in either \(\tau_{yz}\) (shear stress in the \(yz\) plane) or the ratio of the coordinate normal stresses, \((\sigma_{yz}/\sigma_{xx})\), where \(x\) and \(y\) are in the horizontal and \(z\) is vertical. Both mechanisms are compatible with the geologically dependent rheological variation employed in our model. Citation: Upton, P., P. O. Koons, D. Craw, C. M. Henderson, and R. Enlow (2009), Along-strike differences in the Southern Alps of New Zealand: Consequences of inherited variation in rheology, Tectonics, 28, TC2007, doi:10.1029/2008TC002353.

1. Introduction

[2] The Southern Alps of New Zealand are a highly three-dimensional mountain range and yet have generally been described using two-dimensional cross-sectional models [e.g., Wellman, 1979; Koons, 1990; Beaumont et al., 1992; Willett et al., 1993; Upton, 1998; Koons et al., 2003]. These models are insufficient to describe deformation fields resulting from oblique plate motions and/or spatially variable rheological behavior and boundary conditions [Gerbauld et al., 2003; Upton and Koons, 2007]. Several examples of plate boundary parallel variation exist in the Southern Alps, including significant orogen width variation (Figure 1a), variation in orientation of principal horizontal strain axes, the existence of a region of anomalous displacement partitioning (coeval thrusts oriented parallel and perpendicular to the plate boundary), an increase in crustal thickness from north to south (Figure 1c) and changes in the location of the locus of fluid flow and orogenic gold deposition (Craw et al., 2007). We suggest that the increase in crustal thickness from north to south, and an associated variation in lower crustal strength, leads to the other characteristics mentioned and in this study we use numerical and analytical calculations to test this theory. By comparing strain and displacement fields from three-dimensional models with surface, geochemical and geophysical observations, we can impose bounds upon the dynamics compatible with real systems.

[3] Within the framework of our three-dimensional numerical models, conditioned by the geometry and far-field plate driving forces of the South Island, we investigate the influence lower crustal strength and thickness has on the nature of the developing orogen. The horizontal distribution of strain within a zone of oblique convergence reflects tectonic boundary conditions as transmitted through the rheological structure of the deforming lithosphere. In this case the crust that we are considering is thin and strong in one region and weak and thick in another.

[4] We also present an analytical approximation of our rheological model which uses perturbation theory to determine whether or not our model can produce the distinctive displacement partitioning observed in the eastern Southern Alps (Figure 1a). This method has previously been described by Enlow and Koons [1998] and used by Upton et al. [2008].

2. Southern Alps of New Zealand

2.1. Tectonic Setting

[5] The microcontinent Zealandia, of which New Zealand is the subaerial portion, was largely formed off the edge of Gondwana in the Mesozoic [Wandres and Bradshaw, 2005]. A distinct plate boundary began to develop through Zealandia at circa 45 Ma [Sutherland, 1999]. Since that time, 850 km of dextral shear is inferred to have occurred across the South Island [Sutherland, 1999]; 460 km of dextral strike slip is evident along the present-day plate boundary, the Alpine Fault, an oblique reverse structure which dips SE beneath the Southern Alps [Norris et al., 1990]. The remaining 400 km have been taken up by distributed shear across the South Island [Coombs et al., 1976; Molnar et al., 1999; Cox and Sutherland, 2007]. Today, the relative far-field plate vector driving fault movement and associated deformation is directed
ENE-WSW with relative motion of \( \sim 37 \text{ mm a}^{-1} \) [DeMets et al., 1994], resolved into 35.5 mm a\(^{-1}\) parallel to the plate boundary and 10 mm a\(^{-1}\) perpendicular, in the central South Island. Estimates of exhumation rates are on the order of 6–9 mm a\(^{-1}\) [Wellman, 1979; Simpson et al., 1994; Norris and Cooper, 2001; Little et al., 2005] in the central Southern Alps, immediately adjacent to the Alpine Fault and decrease eastward to \( \sim 1 \text{ mm a}^{-1}\) or less east of the Main Divide [Blick et al., 1985]. Uplift rates adjacent to the Alpine Fault decrease to 5–6 mm a\(^{-1}\) to the northeast, as much of the relative motion is transferred to the Hope Fault [Norris and Cooper, 2001]. South of Hasst, uplift rates decrease to \( \sim 3 \text{ mm a}^{-1}\) as motion is partitioned into the Fiordland subduction zone [Norris and Cooper, 2001].

[a] To the north and south of the central South Island the tectonics of Zealandia reflect the changing character of the Australian-Pacific plate boundary in the south Pacific (Figure 1b). To the north, the oceanic Pacific plate subducts beneath the Australian plate and to the south, the oceanic Australian plate is obliquely overridden by the Pacific plate. The South Island of New Zealand is dominated by the actively rising Southern Alps along the continent-continent portion of the plate boundary (Figure 1a). Transition from continental collision to subduction occurs across the Northern South Island. Here five major dextral slip faults strike at a low angle to the present Australia-Pacific plate convergence vector. These are known as the Marlborough Fault Zone, and their strike-slip motion accommodates most of the plate motion with underthrusting of the Pacific plate and uplift and shortening within the Australian crust taking up most of the rest. The Marlborough faults young to the south, with the oldest and westernmost, the Wairau Fault, exhibiting the greatest total offset [Little and Jones, 1998]. To the south of the central Southern Alps, the Alpine Fault steepens and becomes dominantly strike-slip as relative plate motion is partitioned onto a series of offshore thrust faults which link into the northern end of the Puyseduc Trench [Barnes et al., 2005].

2.2. Three-Dimensional Nature of the Orogen

[7] Oblique collision between the Australian and Pacific plates in the South Island of New Zealand is often treated as varying only in the direction normal to the plate boundary, the Alpine Fault [e.g., Wellman, 1979; Koons, 1990; Beaumont et al., 1992; Willett et al., 1993; Upton, 1998]. Two-dimensional
cross sections through the central South Island, normal to the Alpine Fault, have commonly been used to represent the mechanical behavior of the oblique convergence in either two-dimensional [e.g., Koons, 1990; Beaumont et al., 1992; Willett et al., 1993; Upton, 1998], or three-dimensional formulations [e.g., Koons, 1994; Koons et al., 1998]. In reality, the width and character of the zone of active deformation associated with oblique collision varies significantly along strike of the plate boundary (Figure 1a) as mentioned in previous geodynamic descriptions [Gerbault et al., 2003; Upton and Koons, 2007]. To the south of the central Southern Alps, the zone of modern deformation widens by ~100%, the basement geology is dominated by midercristal schists as opposed to graywacke and an orthogonal pattern of actively growing ranges dominates the topography of the transition zone (Figure 1a). Consequently, a single cross-sectional view of the South Island provides an incomplete representation of the deformation along this portion of the plate boundary.

2.3. Canterbury and Otago Strain Regimes

The regions we consider in this study are within the central and southern South Island, which lie to the south of the Marlborough Fault System, and are known by the regional names Canterbury and Otago, respectively (Figure 1). The eastward extent of modern deformation associated with oblique Australian-Pacific collision can be broadly correlated with the width of elevated topography in the South Island.

In the Canterbury region, which extends for ~100 km along strike, from the Porter’s Pass Fault south to the Hunter Hills and Brothers faults, the Southern Alps are narrow, ~70 km wide, reflecting deformation concentrated close to the plate boundary and strain rates on the order of 5 \( \times 10^{-7} \) a \(^{-1}\) [Beavan et al., 1999, 2007; Henderson, 2003] (Figure 1a). The topography northwest of the Porter’s Pass Fault is associated with the Marlborough Fault System. Note also that the topography east of this region, Banks Peninsula, is volcanic in origin, dated at ~8–11 Ma [Sewell, 1988] and is not associated with the present-day collision. In Otago and South Canterbury, deformation associated with the current tectonic regime extends ~200 km to the east coast, including the Southern Alps and the Central Otago Ranges, in a broad zone of lower strain rates (<5 \( \times 10^{-8} \) a \(^{-1}\) (P. Denys, personal communication, 2008)) and distributed deformation (Figure 1a).

Active deformation in central Canterbury is dominated by oblique reverse faulting with minor folding. Structures predominately fall into two distinct patterns: those oriented NE, parallel to the Alpine Fault, and those oriented north (Figure 1). The first category includes the Forest Creek faults, Irishman Creek faults, the Range Front faults and the now inactive Main Divide Fault Zone [Cox and Findlay, 1995; Upton et al., 2003a, 2004]. The second category includes the Ostler Fault [Blick et al., 1989; Ghisetti et al., 2007; Amos et al., 2007] and the Fox Peak faults [James, 1998; Upton et al., 2003a, 2004]. Minor folding is associated with the Fox Peak faults [James, 1998; Upton et al., 2003a, 2004]. Active deformation in Otago occurs on NE trending folds and faults such as the Hyde, Dunstan, Blackstone, and Pisa faults. The folds in this area have been shown to be active fault propagation folds [Markley and Norris, 1999; Jackson et al., 1996].

The region between Canterbury and Otago is characterized by the coexistence of orthogonal thrust faults striking NE and NW (Figure 1a). Note these are not conjugate thrust faults which would be parallel but oppositely dipping but are orthogonal with transport directions at 90 degrees to one another. The NW striking faults are the Waihemo/Hawkdun faults and Waitaki Fault System; both are associated with normal faults formed during Cretaceous extension [Laird, 1993; Sutherland et al., 2001]. The NE striking faults include the Dansey Pass Fault and other parallel unnamed faults [Forsyth, 2002]. Both NE and NW striking fault sets are active implying a stress regime in which \( \sigma_2 \) and \( \sigma_3 \) (maximum and intermediate principal stresses) are close and easily switched [Enlow and Koons, 1998].

Another apparent anomaly exists in the transition between Canterbury and Otago. Some faults in this region exhibit sinistral strike-slip motion, kinematics rare in New Zealand which is dominated by dextral deformation. For example, a late Quaternary fault trace along the Wharekuri Fault in the Waitaki Valley records sinistral strike-slip motion [Gair and Gregg, 1960; Forsyth, 2002] (Figure 2a). Likewise, post-Pliocene offset on the Ostler at the edge of the mountains along the Canterbury-Otago transition is sinistral [Templeton et al., 1999] (Figure 2a).

2.4. Basement Geology

The basement for the Southern Alps consists of metasedimentary terranes that were amalgamated along the Gondwana margin in the Mesozoic [MacKinnon, 1983; Norris and Craw, 1987; Mortimer, 1993a, 1993b] (Figure 2c). The Gondwana margin was oriented approximately south southeast [Wandres and Bradshaw, 2005]. Accounting for Cenozoic rotations, it was parallel to the present-day boundary between Canterbury and Otago and was orthogonal to the present-day plate boundary (Figure 2c). The Torlesse Terrane is the most significant of these terranes as it underlies both Canterbury and Otago [MacKinnon, 1983; Mortimer, 1993a, 1993b]. The Canterbury portion of this terrane is an accretionary complex made up of variably deformed quartzofeldspathic metagraywackes [MacKinnon, 1983]. Exposed Torlesse rocks are dominated by unfoliated graywackes (prehnite-pumpellyite facies), with some kilometer-scale slices of weakly foliated schists (pumppellyite-actinolite facies) (Figures 2a and 2b). Jurassic collision between the Torlesse Terrane and the Aspiring and Caples terranes resulted in the thick metamorphic pile that became the Otago crustal block [Norris and Craw, 1987; Mortimer, 1993a, 1993b]. The metamorphic belt was unoofed erosionally and tectonically between the Jurassic and middle Cretaceous, exposing green-rich facies rocks (up to biotite-garnet zone) over most of Otago (Figures 2a and 2b) [Craw, 1998; Mortimer, 2000].

Extensive deformation in the middle Cretaceous resulted in steep and gently dipping normal faulting, oriented NE and NW, especially along the margins of the Otago schist belt where ~1–5 km scale rock slices of differing metamorphic grades were juxtaposed [Deckert et al., 2003; Gray and Foster,
Extension continued, accompanying the breakup of Gondwana, through to the Oligocene when Zealandia was essentially submerged below sea level. Marine planation accompanied this progressive submergence, leading to a widespread diachronous unconformity, the Waipounamu erosion surface, that cut into the basement rocks and remnants of Cretaceous–early Tertiary sediments.
Landis et al., 2008]. Zealandia reemerged from the sea, beginning in the late Oligocene–early Miocene, with the development of the current Pacific-Australian plate boundary through the region [Sutherland, 1999]. Deeper levels (greenschist facies and amphibolite facies) of the Canterbury portion of the Torlesse Terrane are now being exposed by Southern Alps uplift and exhumation immediately SE of the Alpine Fault (Figure 1a). Maximum uplift, maximum exhumation and highest relief are all focused near Mount Cook–Aoraki (Figure 1a).

2.5. Crustal Thickness

[14] Present-day crustal thickness varies both along and across the South Island (Figure 1c). A crustal root has developed along the central South Island, associated with the present-day continental collision. This crustal root reaches depths of ~44 km beneath western Otago [Scherwath et al., 2003; Melhuish et al., 2005; Bourguignon et al., 2007]. In this study, we are interested in the initial thickness of the Canterbury and Otago regions, prior to the start of Cenozoic collision. The best estimate we have for precollisional thicknesses is the present-day crustal thickness at the eastern edge of the continent, which appears to be unaffected by the present-day mountain building. Estimates of Moho depths in mid-Canterbury derived from seismic reflection (SIGHT transect 1), refraction and inversion of seismic velocities are 18 km in the Bounty Trough and ~25 km at the East Coast [Van Avendonk et al., 2004]. Sixty kilometers to the southwest, along SIGHT transect 2, Moho depth is estimated at ~27 km at the East Coast [Scherwath et al., 2003]. The Moho deepens parallel to the coast to ~32 km east of Otago [Kohler and Eberhart-Phillips, 2003]. Further evidence comes from velocity inversion which suggests that lower crustal rocks extend an additional 10 km downward in Otago relative to Canterbury [Eberhart-Phillips and Bannister, 2002; Stern et al., 2001].

[15] The pre-Cenozoic Otago crust was thicker than that in Canterbury as a result of collision between quartzofelspathic graywackes of the Torlesse Terrane and arc-derived graywackes of the Capeles terranes on the eastern edge of Gondwana [Mortimer, 1993a, 1993b; Bradshaw, 1989; Gray and Foster, 2004]. The rocks to the north of Otago were not thickened by the Mesozoic collision. The bulk of the metamorphism and deformation associated with collision occurred between ~160 and 140 Ma through to 120 Ma [Gray and Foster, 2004]. This was followed by very slow exhumation and cooling until ~110 Ma and more significant exhumation of the high-grade metamorphic core between 109 and 100 Ma [Gray and Foster, 2004]. During this time the crust underlying what is now Canterbury was further from the Mesozoic collision zone, closer to the subduction margin and underwent little if any deformation [Bradshaw, 1989] (Figure 2c).

3. Mechanical Modeling

3.1. Mechanical Models

[16] We model the oblique deformation that occurs along the South Island transpressional boundary using a two-layered crust overlying an elastic Pacific mantle (Figure 3). The solution domain is a three-dimensional numerical region extending 400 km normal to the plate boundary (x) by 800 km parallel to the plate boundary (y) by 50 km depth (z). The western edge of the model consists of an elastic block simulating the Australian plate which deforms minimally. The erosional regime of the Southern Alps is a result of an asymmetric precipitation pattern that efficiently removes material from the western inboard slope [Koons, 1990; Beaumont et al., 1992; Willett et al., 1993]. We have simulated this orographic effect by maintaining the western slope at a constant elevation [Koons et al., 1998, 2003; Upton and Koons, 2007].

[17] Our models are aimed at unraveling crustal deformation and the displacement of the mantle is imposed as a boundary condition. Previously, two-end-member kinematic models have been used to explain strain accommodation within the mantle lithosphere beneath the Southern Alps. One model evokes South Island subcrustal mantle deformation over ~200 km [Molnar et al., 1999; Moore et al., 2002; Baldock and Stern, 2005], while the other model assumes little or no deformation in the uppermost mantle and resembles a subduction type boundary condition [e.g., Beaumont et al., 1992; Koons et al., 2003; Ellis et al., 2006; Upton and Koons, 2007]. We have chosen a subduction boundary
condition where the Pacific plate is descending beneath the Australian plate without a large region of mantle deformation, based on a comparison between mechanical models and observations [e.g., Ellis et al., 2006; Liu and Bird, 2006; Savage et al., 2007; Beavan et al., 2007; Upton and Koons, 2007]. These boundary conditions are similar to previously published mechanical models of the Southern Alps [Koons et al., 1998, 2003; Upton and Koons, 2007]. The two-layered Pacific crust is dragged along its base, 25 km below sea level for Canterbury and 35 km below sea level for Otago, at a velocity of 35.5 mm a\(^{-1}\) parallel to (along y axis) and 10 mm a\(^{-1}\) normal to (along x axis) the plate boundary compatible with the calculated relative plate vectors from NUVEL-1A [DeMets et al., 1994] (Figure 1b). The northwestern edge of the model, representing the Australian plate, is held fixed. The velocity conditions on the lateral boundaries are allowed to vary through the model time. We do this by taking the velocity profiles from \(y = \pm 300 \text{ km}\) (internal to the model) and imposing those velocity condition on the lateral boundaries, that is \(y = \pm 400 \text{ km}\). These boundary conditions are continually updated as the model runs. This method allows these oblique boundaries to develop as a function of the rheology. Note that Figures 5 and 7 only show the central portion of the models, the solution domain, and do not extend out as far as the lateral boundaries.

### 3.2. Solution Methodology

[18] Models were developed using the numerical code FLAC\(^3\)D, a three-dimensional finite difference code, which we have modified to accommodate large strains and local erosion [Cundall and Board, 1988]. A large number of comparisons have been made between FLAC\(^3\)D results and exact solutions for plasticity problems (ITASCA, FLAC\(^3\)D, Fast Lagrangian Analysis of Continua in 3 Dimensions, 1997, available at http://www.itascag.com/flac3d/index.php). The results show uniformly good agreement with theoretical solutions and in modeling geometrical instabilities well past the collapse or failure limit. Materials are represented by polyhedral elements within a three-dimensional grid that uses an explicit, time-marching solution scheme and a form of dynamic relaxation. Each element behaves according to a prescribed linear or nonlinear stress/strain law in response to applied forces or boundary restraints. The inertial terms in Cauchy’s equations of motion

\[
\frac{\partial \sigma_{ij}}{\partial x_i} + \rho b_i = \rho \frac{dv_i}{dt}
\]

(1)

(where \(\sigma_{ij}\) is the stress tensor, \(x_i, v_i\) are the vector components of position and velocity, respectively, \(\rho\) is the density of the material, and \(b_i\) is the body force) are used as a numerical means to reach the equilibrium state of the system under consideration. The resulting system of ordinary differential equations is then solved numerically using an explicit finite difference approach in time. The drawbacks of the explicit formulation (i.e., small time step limitation and the question of required damping) are overcome by automatic inertia scaling and automatic damping that does not influence the mode of failure. The governing differential equations are solved alternately, with the output for the solutions of the equations of motion used as input to the constitutive equations for a progressive calculation. Solution is achieved by approximating first-order space and time derivatives of a variable using finite differences, assuming linear variations of the variable over finite space and time intervals, respectively. The continuous medium is replaced by a discrete equivalent with all forces involved concentrated at the nodes of a three-dimensional mesh used in the medium representation.

### 3.3. Model Crustal Rheology

[19] Numerical modeling requires approximating the rheology of the material being studied by a mathematical description of the stress/strain or stress/strain rate relationship. The rheological descriptions we use are generally derived from laboratory experiments. Crustal rheology is unlikely to be as simple as these experimentally derived flow laws would imply, however, a sharp reduction in flow stress at temperatures above 300–350°C due to the exponential dependence upon temperature appears to be a robust feature of quartz- and feldspar-dominated crust [Sibson, 1982]. We have chosen a rheological model similar to the “jelly sandwich” type [Chen and Molnar, 1983; Burov and Watts, 2006].

[20] Crustal velocities of 6.0–6.2 km s\(^{-1}\) determined for the upper and midcrust of the Southern Alps along SIGHT transects 1 and 2 are compatible with a quartzofeldspathic dominated crust [Scherwath et al., 2003; Van Avendonk et al., 2004]. At the base of the crust, crustal velocities of 7.0–7.1 km s\(^{-1}\) are consistent with a more mafic rheology [Scherwath et al., 2003; Van Avendonk et al., 2004]. Given these values, we represent crustal materials using an elastic/Mohr-Coulomb, pressure-dependent upper crust overlying a feldspar-dominated lower crust with flow parameters determined from the published creep laws at a reference strain rate of \(10^{-14}\) s\(^{-1}\) and an average geothermal gradient of 20°C km\(^{-1}\) (equation (2)) [e.g., Upton and Koons, 2007]:

\[
K_\varphi = 0.5 \Delta \sigma
\]

(2)

where

\[
\Delta \sigma = \left[ \left( \frac{\dot{\varepsilon}}{A} \right) \exp \left( \frac{Q}{RT} \right) \right]^{1/n}
\]

\(\Delta \sigma\) differential stress, \(\sigma_1 - \sigma_3\);  
\(K_\varphi\) shear strength of the material;  
\(A\) a preexponential factor;  
\(\dot{\varepsilon}\) shear strain rate;  
\(n\) power law exponent;  
\(Q\) activation energy;  
\(R\) universal gas constant;  
\(T\) temperature (K).
Bulk modulus (Pa) 1
Shear modulus (Pa) 3
Density (kg m$^{-3}$) 2800
Fricition angle (deg) 35
$K_r$ 1 x 10$^6$ to 1 x 10$^8$ 1 x 10$^6$ to 6 x 10$^7$

Table 1. Parameters, Material Properties, and Initial Conditions Used in the Numerical Models

<table>
<thead>
<tr>
<th>Region</th>
<th>Canterbury</th>
<th>Otago</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bulk modulus (Pa)</td>
<td>1 x 10$^{10}$</td>
<td>1 x 10$^{10}$</td>
</tr>
<tr>
<td>Shear modulus (Pa)</td>
<td>3 x 10$^9$</td>
<td>3 x 10$^9$</td>
</tr>
<tr>
<td>Density (kg m$^{-3}$)</td>
<td>2800</td>
<td>2800</td>
</tr>
<tr>
<td>Fricition angle (deg)</td>
<td>35</td>
<td>35</td>
</tr>
<tr>
<td>$K_r$</td>
<td>1 x 10$^6$ to 1 x 10$^8$</td>
<td>1 x 10$^6$ to 6 x 10$^7$</td>
</tr>
</tbody>
</table>

[21] We approximate temperature-dependent weakening by decreasing the value of $K_r$ with increasing temperature in a manner constant with yield stress data for rheologies determined by laboratory creep experiments. $A$, $Q$, and $n$ are determined from extrapolation of experiments on wet synthetic quartzite [Paterson and Luan, 1990], $A = 6.5 \times 10^{18}$ MPa$^{-n_s}$ s$^{-1}$; $Q = 135$ kJ mol$^{-1}$; $n = 3.1$. In the lower crust the rheology is dominated by mafic minerals [Scherwath et al., 2003; Van Avendonk et al., 2004]. We use the diabase rheology of Shelton and Tullis [1981] as an estimate for lower crustal strength, $A = 2 \times 10^{-4}$ MPa$^{-n_s}$ s$^{-1}$; $Q = 260$ kJ mol$^{-1}$; $n = 3.4$ (Table 1).

[22] We have defined the lower crust in this way rather than using the power law rheology also available to us in FLAC3D as we have found that our solutions are more stable when running to higher strains. We recognize that we are unable to model time-dependent behavior or localization using this definition. The reference strain rate of 10$^{-14}$ s$^{-1}$ is likely to be too low for materials close to the Alpine Fault, but the scale of our models is such that we cannot localize deformation to the extent that occurs in nature. In the central Southern Alps (adjacent to the Canterbury region), ~75% of strain is focused onto the Alpine Fault at the surface [Norris and Cooper, 2001]. This leaves ~25% distributed across the rest of the orogen. We are using a large-scale model, with a resolution of 4 km per grid cell, across the >200 km width of the orogen. Our analysis is concentrated on the lower strain eastern portion of the orogen, and for these reasons, we feel that this simplifying assumption can be justified.

[23] The rheological properties are varied along strike of the boundary in a manner constrained by the geological arguments presented above (Figure 4). The northern half of the model represents the Canterbury region, the crust here is given a thickness of 25 km and the lower crust is strong (Figures 3 and 4). The southern half of the model represents the Otago region. It is given a crustal thickness of 35 km and the lower crust is weak relative to that of the Canterbury region (Figures 3 and 4).

[24] It is the relative difference in the strengths that is important for the modeling rather than the details of the actual values. Thus we believe that using an approximation based on a quartzite upper crust and a diabase lower crust for different crustal thicknesses is reasonable. The difference in strength implied between model Canterbury and Otago (Figure 4) is much more significant than the difference in actual values for either region if we were to alter the rheology to a granite-based upper crust and amphibolite-based lower crust rather than a quartzite/diabase crust for example.

4. Results

4.1. Model Surface Velocities

[25] The strength contrast in the model lower crust is reflected in the strain pattern at the surface as illustrated in the two components of displacement, parallel to and normal to the plate boundary (Figures 5a and 5b). In model Canterbury, plate-parallel motion is accommodated adjacent to the model plate boundary, while plate-normal displacement extends out to about 100 km east of the model boundary (Figures 5a and 5b). The net result is high strain rates in model Canterbury within ~60 km of the model Alpine Fault with the highest strain rates adjacent to the fault. In model Otago, both plate-parallel and plate-normal motion are taken up in a broad zone extending across much of the model region (Figures 5a and 5b). Some partitioning of deformation occurs with the plate-parallel motion gradually fanning out from the model Otago/Canterbury transition zone versus the sudden expansion of plate-normal motion across all of Otago at the transition.

[26] Model vertical uplift rates also reflects the along-strike rheological variation (Figure 5c). In model Canterbury, uplift is concentrated into a ~70 km wide band adjacent to the model Alpine Fault, with greatest uplift rates adjacent to the fault. Vertical motion in model Otago is more widespread with a region of moderate uplift rates adjacent to the model Alpine Fault with distributed low uplift rates extending eastward to the edge of the continent. A robust feature of this model is a region of high uplift rates where the transition from weak to strong lower crustal rheology intersects the model plate boundary.

4.2. Examination of Transition Between Displacement States Using Perturbation Theory

[27] The similarity of natural and model kinematics at a regional scale encourages some confidence in our chosen mechanical model. However, the surface displacement field
that accommodates the regional kinematics contains more information on earth dynamics than is visible at the regional scale. The coexistence of at least two distinct displacement states, thrusting normal to, and thrusting parallel to, the plate boundary as discussed above (Figure 1a), suggests coexisting stress states with a stress transition analogous to that associated with displacement partitioning between strike slip and oblique thrusting regimes [Enlow and Koons, 1998]. Numerous phenomena that alter any of the coordinate stress components can cause minor changes of displacement orientation and provide limited constraint on stress states, however, as discussed below the coexistence of distinct displacement states can only occur for a limited set of stress conditions and therefore severely restricts the degrees of freedom of the relative values of the coordinate stress tensor.

Using perturbation theory for 3-D deforming Mohr-Coulomb wedges, Enlow and Koons [1998] demonstrated the conditions under which very different displacement states can be separated by minor stress differences. This transition between displacement states occurs with rapid rotation of the principal stress axes as they change their association with the coordinate stresses (Figure 6). The requisite conditions for

---

**Figure 5.** Model results overlain onto an outline of the South Island. (a) Contours of plate-parallel velocity relative to a fixed Pacific plate. (b) Contours of plate-normal velocity relative to a fixed Pacific plate. (c) Absolute uplift rate. See text for explanation.

**Figure 6.** Relationship of coordinate stresses and principal stresses for a model orogen with increasing obliquity (Figures 6a to 6c) [Koons, 1994; Enlow and Koons, 1998]. As the relative motion becomes more oblique, the horizontal shear stress increases and the orientation of the principal stresses changes. (a) Initially, \( \sigma_1 \) is associated with margin-normal direction \((x)\), \( \sigma_2 \) is associated with margin-parallel direction \((y)\), and \( \sigma_3 \) is associated with the vertical \((z)\). (b) As the horizontal shear stress increases, \( \sigma_2 \) and \( \sigma_3 \) rotate about the margin-parallel direction \((y)\), staying close to the \(yz\) plane shown in light gray. (c) Eventually swapping such that \( \sigma_2 \) becomes associated with the vertical \((z)\) and \( \sigma_3 \) becomes associated with the margin-parallel direction \((y)\). The style of faulting changes from thrusting normal to the plate boundary (Figure 6a) through oblique thrusting (Figure 6b) to nearly pure strike slip (Figure 6c).
transition from predominantly reverse faulting to dominantly strike-slip faulting were shown to occur when

$$\tau_{xy} \rightarrow \tilde{\tau} \equiv \sqrt{\tau_{zz}^2 + (\sigma_{xx} - \sigma_{yy}) (\sigma_{yy} - \sigma_{zz})}$$

(4)

\(\tau_{xy}\)  horizontal shear stress;
\(\tilde{\tau}\)  characteristic stress \([Enlow and Koons, 1998];\)
\(\tau_{xz}\)  shear stress in the \(x - z\) plane;
\(\sigma_{xx}\)  normal stress in the \(x\) direction;
\(\sigma_{yy}\)  normal stress in the \(y\) direction;
\(\sigma_{zz}\)  normal stress in the \(z\) direction.

\([29]\) A dimensionless stress perturbation parameter, \(\varepsilon^{(x-y-z)} \equiv \tau_{xy}/\tilde{\tau}\) \([Enlow and Koons, 1998]\) is used to represent how close a system is to transition. The symbol \(\varepsilon\) reflects the origin of the parameter from perturbation theory. The notation of this parameter refers to \(\sigma_2\) (the intermediate principal stress) changing from being associated with the \(y\) coordinate axis to being associated with the \(z\) (vertical) coordinate axis.

\([30]\) With increasing horizontal shear stress \((\tau_{xy})\), the intermediate principal stress, \(\sigma_2\), switches from being associated with the plate-parallel direction \((y)\) to being associated with the vertical axis \((z)\). This switch is accompanied by the equivalent switch of \(\sigma_3\) from near vertical to near horizontal. These two minor stress axes rotate about the horizontal \(\sigma_1\).

\([31]\) An algebraic symmetry exists between this previously examined transition from thrusting to strike-slip (Figure 6), and the transition between oblique reverse faulting in approximately orthogonal directions (Figure 7). This symmetry allows transition to be expressed by an analogous dimensionless parameter, \(\varepsilon^{(x-y)} \equiv \tau_{xy}/\tilde{\tau}\). The notation of this parameter refers to \(\sigma_1\) (the maximum principal stress) changing from being associated with the \(x\) coordinate axis to being associated with the \(y\) coordinate axis. This transition indicates a switching of the major principal axes between the two horizontal coordinates as they rotate about a near vertical axis; i.e., \(\sigma_1\) switches from predominantly associated with the boundary-normal stress \((\sigma_{xx})\) to being predominantly associated with the boundary-parallel coordinate-normal stress \((\sigma_{yy})\) (Figure 7). Transition from boundary-normal thrusting to boundary-parallel thrusting occurs when \(\varepsilon^{(x-y)}\) approaches unity either through an increase in \(\tau_{yz}\) or by a reduction in \(\tilde{\tau}\) as

$$\tau_{yz} \rightarrow \tilde{\tau} \equiv \sqrt{\tau_{zz}^2 + (\sigma_{xx} - \sigma_{yy}) (\sigma_{yy} - \sigma_{zz})}$$

(5)

\(\tau_{yz}\)  vertical shear stress in the \(yz\) plane;
\(\tilde{\tau}\)  characteristic stress \([Enlow and Koons, 1998];\)
\(\tau_{xz}\)  shear stress in the \(xz\) plane;
\(\sigma_{xx}\)  normal stress in the \(x\) direction;
\(\sigma_{yy}\)  normal stress in the \(y\) direction;
\(\sigma_{zz}\)  normal stress in the \(z\) direction.

\([32]\) The dimensionless parameter \(\varepsilon^{(x-y)} \leq 1\) corresponds to \(\sigma_1\) being associated with \(x\) and \(\varepsilon^{(x-y)} > 1\) corresponds to \(\sigma_1\) being associated with \(y\); \(x\) is the convergence direction and thus \(\sigma_1\) is most often associated with \(x\), implying that \(\varepsilon^{(x-y)}\) is most commonly \(\leq 1\).

\([33]\) We extracted stress components from our three-dimensional mechanical model described above (Figure 5) and used them to calculate the spatial distribution of \(\varepsilon^{(x-y)}\). We were then able to identify those regions where transition among displacement states is likely to occur (Figure 7a). The stress diagrams, calculated along a horizontal slice \(\sim 8\) km below the surface, show that generally \(\varepsilon^{(x-y)} \leq 1\) over both the high and low strength regions where we predict that only boundary-normal thrusting is likely to occur (Figure 7a). However, \(\varepsilon^{(x-y)} \geq 1\) over the boundary between high strength and low strength regions (model Canterbury–model Otago) due to a local increase in both \(\tau_{yz}\) and the ratio of the coordinate-normal stresses, \(\gamma^{(x-y)} = \sigma_{yy}/\sigma_{xx}\) (Figures 7b and 7c). Consequently, transition from boundary-normal to boundary-parallel oblique thrusting is predicted over the model rheological boundary.

\([34]\) In oblique convergence using the convention shown in Figure 3, \(\sigma_{xx}\) is associated with the normal to the plate boundary and, as indicated by the ratio of the two horizontal compressive stresses, \(\sigma_{yy}/\sigma_{xx}\) \((\gamma^{(x)})\) in the strong-based Canterbury region, is the dominant normal stress (Figure 7b). The magnitude of \(\sigma_2\) in a Mohr-Coulomb material is constrained only to lie between that of the other principal stresses and differs from the coordinate stress \(\sigma_{yy}\) as a function of the shear stresses, \(\tau_{xy}\) and \(\tau_{yz}\). An increase in the horizontal shear stress, \(\tau_{xy}\), reduces \(\sigma_2\), while increased shear stress in the vertical plane, \(\tau_{yz}\), increases \(\sigma_2\). When the magnitude of \(\sigma_{yy}\) approaches that of \(\sigma_{xx}\) then either an increase in \(\tau_{yz}\) or a reduction in \(\tau_{xz}\) can cause the magnitude of \(\sigma_2\) to reach that of \(\sigma_1\). Under these conditions, the coexistence of reverse faulting on orthogonal displacement planes is possible for minor shifts in shear stresses \([Enlow and Koons, 1998]\).

\([35]\) In summary, the coexistence of orthogonal reverse fault systems is favored by \(\sigma_{yy}\) approaching \(\sigma_{xx}\) and increasing vertical shear stress parallel to the plate boundary, \(\tau_{yz}\) so that \(\varepsilon^{(x-y)}\) approaches 1 (Figure 7). The vertical rheological boundary created by the change in crustal thickness and strength imposes an additional vertical shear stress in this region, which is transmitted up through the brittle crust. In the central South Island, coexistence of orthogonal thrusting directions associated with the schist to graywacke transition are compatible with a rheological transition from weak to strong lower crust.

\([36]\) Models were also run in which the width of the transition zone was varied (see Figure S1 in the auxiliary material).\(^{1}\) As the width of the transition zone was increased up to \(100\) km, the model results were largely unchanged. The zone of \(\varepsilon^{(x-y)} > 1\) becomes narrower and is shifted slightly toward to stronger crust (see Figure S1).

5. Discussion

\([37]\) Oblique convergence of the model crust with differing lower crustal strengths results in an increased width of deformation and lower strain rates over the weak zone relative to the strong region (Figure 5). This is to be expected because

\(^{1}\)Auxiliary materials are available in the HTML. doi:10.1029/2008TC002353.
the weak lower crust is unable to sustain high stresses and will consequently fail over a larger region while failure in the strong lower crust will be more confined adjacent to the plate boundary. However, the surface displacement field of the South Island contains more information than is visible at the regional scale. By using our three-dimensional mechanical models and examining the resulting stress states using perturbation theory we are able to extract stress components and to investigate the implications of this rheological variation for other processes such as surface faulting, timing and localization of uplift, fluid flow and gold mineralization.

5.1. Geological Consequences

The different crustal thicknesses and rheologies of Canterbury and Otago, that are a consequence of their mode of formation, have governed many features of their subsequent geological history. Comparison of geology and our modeling results allows us to offer an explanation for some of the features of the Southern Alps that vary along strike from Canterbury in the north and Otago in the south.

5.1.1. Partitioning and Width of Deformation

The model results compare favorably with observations from the South Island, where deformation and uplift
extend to the coast in Otago but are restricted to about 70 km east of the Alpine Fault in Canterbury. Partitioning of the horizontal velocity components is more pronounced where the lower crust is weaker as predicted from analytical analysis [Enlow and Koons, 1998] and observed in model results [Upont et al., 2003b]. Norris and Cooper [2001] show that there is clear evidence in the slip rate data for increasing partitioning toward the south of the fault-normal component onto structures other than the Alpine Fault, while the amount of fault-parallel displacement taken up on the Alpine Fault remains relatively constant along the length of the boundary. In the early states of model convergence while deformation in both Canterbury and Otago is expanding most rapidly, a zone of sinistral displacement develops between the weak and strong base blocks at high angle to the plate boundary (Figure 5b). Recent movement along the Wharekuri Fault, within the transition between Canterbury and Otago and approximately normal to the plate boundary, as well as geodetic analysis shows similar sinistral kinematics over the long term [Gair and Gregg, 1960; Forsyth, 2002] and anticlockwise rotation in the modern geodetic field [Pearson, 1990; Henderson, 2003].

5.1.5. Current Motion on the Canterbury-Otago Boundary

The exact geometry of the transition between thick and thin crust, constructed in the Mesozoic, is unknown because of subsequent modification. The thickness transition became the locus for normal faulting in the middle Cretaceous, and this has accentuated the Canterbury-Otago boundary. The most prominent feature of this boundary at the surface is the Waihemo-Hawkdun Fault system and related structures (Figure 1a). This fault system juxtaposes prehnite-pumpellyite facies metagraywackes of Canterbury against greenschist facies schists of Otago via a series of narrow northwest striking fault slices over a 5–10 km wide zone [Forsyth, 2002; MacKenzie and Craw, 2006]. Gently dipping normal faults subparallel to schist foliation immediately SW of the Waihemo-Hawkdun Fault system were related to this mid-Cretaceous extension, and these also juxtaposed rocks of differing metamorphic grade in the transition between schist and graywacke [Deckert et al., 2003; Craw et al., 2005].

5.1.6. Displacement Partitioning at the Transition

Along the transition zone between Canterbury and Otago, both NE and NW striking thrusts are active, implying that \( \sigma_1 \) and \( \sigma_2 \) are of similar magnitude and easily switched [Enlow and Koons, 1998]. This switching is facilitated by an increase in \( \tau_{xy} \) and/or an increase in the ratio \( \sigma_{xy}/\sigma_{xx} \) (Figures 7b and 7c). Both occur over the transition from a strong to weak lower crust, such as occurs between Canterbury and Otago. A similar situation exists on the northern Colorado Plateau where kinematic analysis of Laramide basement-cored uplifts indicate that at least four major uplifts were formed by northeast directed shortening while at least two more were constructed by southeast directed shortening [Kelley and Clinton, 1960; Jamison and Stearns, 1982; Anderson and Barnhard, 1986; Davis, 1999; Tindall and Davis, 1999; Bump and Davis, 2003]. Available timing data for these uplifts are scarce but point to simultaneous rise of both sets of uplifts [Lawton, 1983; Goldstrand, 1994; Bump, 2004].
5.2. Gold Mineralization

Hydrothermal gold mineralization accompanied several stages of the Miocene-Recent tectonic activity described above. Gold-bearing fluids were generated by metamorphic dehydration reactions, particularly associated with greenschist facies metamorphism [Craw et al., 2002; Pitcairn et al., 2006]. These fluids were channeled to sites of gold deposition via structures formed during synchronous deformation [Craw et al., 2002]. Hence, formation of gold deposits was intimately linked to the tectonic evolution of the crust, and there are some clear controls on distribution of gold deposits related to the differences in crustal thicknesses and rheologies between Canterbury and Otago.

5.2.1. Miocene Gold

Deformation and uplift of thick Otago crust during initiation of the Alpine Fault and Southern Alps (above) were accompanied by gold mineralization [Craw et al., 2007]. Emplacement of gold-bearing veins was intimately related to deformation along the Moonlight Fault (Figure 1a). The veins occur in Aspiring Terrane schists that underlie the Torlesse Terrane in the thick Otago schist pile of NW Otago [Norris and Craw, 1987]. The Aspiring Terrane contains abundant chlorite-rich metabasites that may have generated fluids at depth during the Miocene deformation [Norris and Craw, 1987; Craw et al., 2007]. In contrast, there is no evidence for this style of deformation or gold mineralization in Canterbury during the Miocene. The thinner Canterbury crust, lack of fertile Aspiring Terrane source, and lack of focused Miocene deformation are the probable reasons for the lack of Canterbury Miocene gold, and these are all related to the differences in crustal structure and rheology between Canterbury and Otago.

5.2.2. Plio-Pleistocene Gold

A belt of small gold deposits occurs near the crest of the Canterbury portion of the Southern Alps, and these formed during Plio-Pleistocene uplift and deformation adjacent to the Alpine Fault [Craw et al., 2002]. Mineralizing fluids have been, and are being, generated in the middle crustal portion of the narrow thickened root beneath the Southern Alps, and these fluids rise buoyantly along fractures beneath the main mountain chain [Craw et al., 2002; Upton and Koons, 2007]. These processes operating in the Southern Alps of Canterbury are similar to those that operated at the Miocene inception of the Southern Alps in Otago (section 5.2.1), but the resultant veins are temporally and spatially distinct. Gold mineralization ceased in Otago when the orogen changed from a narrow focused zone in the Miocene to the present wide orogen, and the gold mineralization system stepped up to the narrow focused Canterbury portion of the orogen.

6. Conclusions

As with most natural materials, the Earth’s crust exhibits a path-dependent rheology, with a memory dependent upon deformation (including strain and thermal evolution) history [e.g., Koons et al., 2003; Jamieson et al., 2004]. The Southern Alps of New Zealand are a highly threedimensional mountain belt that cannot be adequately modeled using a two-dimensional description in which properties vary only normal to the plate boundary. We suggest that many features of the long-term displacement field of this region can be produced using a simple along-strike variation in the rheological model of the crust subjected to far-field driving velocities derived from plate tectonic observations. At the orogen scale, the increasing width of the orogen is compatible with reduced flow resistance at depth beneath the southern Otago portion of the plate boundary where geological studies suggest a deeper, weaker continental crust. At suborogen scale, the coexistence of contrasting displacement states in the transition region of eastern New Zealand is also compatible with the same rheological model. The perturbation-derived variable used here, $\varepsilon_x^\alpha$ and $\varepsilon_x^\beta$, provides a means of evaluating crustal stress states where displacement partitioning is observed.

Our successful application of experimentally derived crustal strength profiles within our continuum model framework of the central South Island by no means provides proof...
of our rheological model. Other combinations of rheology and varying boundary conditions may be capable of producing similar results. However, similarity of field observations and numerical results from our simple model based upon temperature-dependent crustal rheology indicates that the “jelly sandwich” rheological model is a permissible model in this instance. This model is sufficient to reproduce strain observations at both the lithospheric scale of the South Island and at the scale of the local structures where displacement partitioning occurs.

## Acknowledgments

P.O.K. acknowledges the enthusiasm and insight of the late Sarah Beanland for sparking our interest in the coexisting thrust faulting and strike-slip deformation states of the Otago-Canterbury transition. Richard Norris, Chuck Landis, Graham Bishop, and Chris Pearson have contributed both knowledge and advice, which we appreciate, as we do the financial support from NZ FRST and the University of Otago. Chuck Landis drew the sketch on which Figure 2 is based. Comments on the manuscript from Richard Norris helped clarify numerous issues. Mattel and two anonymous reviewers are thanked for thorough reviews which greatly improved the manuscript.

### References


Baldock, G., and T. Stern (2005), Width of mantle deforma-

## References


Baldock, G., and T. Stern (2005), Width of mantle deforma-

## References


Baldock, G., and T. Stern (2005), Width of mantle deforma-

## References


Baldock, G., and T. Stern (2005), Width of mantle deforma-

## References


Kelley, V. C., and N. J. Clinton (1960), Fracture systems and tectonic elements of the Colorado Plateau, Publ. Geol. 6, Univ. of N. M., Albuquerque.


Wannamaker, P. E., G. J. Jiracek, J. A. Stoth, G. Caldwell, V. M. Gonzalez, J. D. McKnight, and


C. M. Henderson, Department of Geological Sciences, School of Geography, Environment and Earth Sciences, Victoria University, P.O. Box 600, Wellington 6140, New Zealand.

P. O. Koons, Department of Earth Sciences, University of Maine, Orono, ME 04469, USA.

P. Upton, GNS Science, Private Bag 1930, Dunedin 9054, New Zealand. (p.upton@gns.cri.nz)