Coupled crust-mantle dynamics and intraplate tectonics:
Two-dimensional numerical and three-dimensional analogue modeling

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[1] Tectonic deformation of some intraplate regions may be caused by the Rayleigh-Taylor (RT) instability of dense subcrustal lithosphere (mantle lithosphere) as it descends into the mantle. We report on a series of scaled three-dimensional (3-D) analogue and 2-D numerical experiments of coupled crust-mantle dynamics that study such a process. The models investigate the effects of two geometries of RT instability (linear versus axisymmetric) and use different rheological stratifications of the crust. The growth of the RT instability is strongly influenced by crustal rheology. While RT growth rates are not greatly influenced by the geometry of the mantle lithosphere instability itself, the surface is. Experiments having a viscous-only crust are characterized by significant crustal contraction and thickening above the mantle downwelling and extension and thinning in adjacent regions. In experiments with a brittle upper crust the surface deformation is much more subdued. However, strong mantle flow-induced deformation still occurs in the ductile lower crust beneath the quiescent brittle upper crust. When the lower crust is relatively strong, localized styles of extensional and contractional structure develop in the brittle upper crust owing to the greater degree of coupling between upper crust and the mantle instability. In all the experiments, a portion of the ductile lower crustal material is entrained within the downgoing RT instability deep into the mantle. The interplay between dynamic topography and crustal thickening/thinning induced by the underlying mantle flow governs various time-dependent phases of subsidence and uplift of the model surface. The experiments help to account for a number of first-order tectonic behaviors of the lithospheres of Earth and other terrestrial planets. For example, intraplate orogens and basins, complex localized tectonic structure in intraplate settings, and deep crustal seismic fabrics may arise as the variable crustal response to underlying mantle lithosphere RT instabilities.

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1. Introduction

[2] A recent challenge in geodynamics research has been to improve understanding of the complex interactions between Earth’s crust and mantle. It is clear that the convecting mantle acts as a thermal engine which drives plate tectonic surface motions, but details regarding the dynamic coupling between these regions remain uncertain.

[3] As a fundamental example of such coupling, creeping flow in the mantle can induce significant normal stresses at the surface, which may lead to transient events of surface uplift/depression or “dynamic topography” [McKenzie et al., 1974; McKenzie, 1977]. The link between surface topography and mantle dynamics has provided an important mechanism that explains various episodes of observed large-scale continental epeirogeny. Mantle flow induced lithospheric subsidence, for example, has been invoked to explain the large-scale subsidence patterns within sedimentary basins and on continental platforms [Mitrovica et al., 1989; Gurnis, 1992; Russell and Gurnis, 1994; Stern and Holt, 1994; Pysklywec and Mitrovica, 1999].

[4] While these studies are successful in explaining the long wavelength vertical motions of a lithospheric plate, they neglect the internal deformation of the crust driven by the mantle flow. However, there is geologic evidence of episodes of significant tectonic deformation of the crust within various intracontinental regions. As an example, portions of the present North American craton may have been prone to pulses of intraplate deformation in the Archean and Proterozoic [West, 1980; Aspler et al., 2002]. In addition, Australia represents a continental plate that has experienced a significant number of episodes of intraplate tectonic activity since the Paleoproterozoic [e.g., Shaw et al., 1991; O’Dea et al., 1997; Betts, 1999; Scott et al., 2000; Smithies, 2002]. It is useful to consider processes other than conventional plate boundary tectonics which may responsible for these deformational events in a continental interior. Furthermore, other terrestrial planets such as Mars and Venus show evidence of significant tectonic surface deformation within what may be regarded as intraplate environments [e.g., Solomon et al., 1992].

[5] One of the mechanisms that has been put forth to account for intraplate tectonics is gravitational instability of the mantle lithosphere (i.e., subcrustal lithosphere). Houseman et al. [1981] first proposed that during continental plate convergence, the mantle lithosphere is accommodated by distributed lithospheric thickening followed by convective removal of a viscous subcrustal root. They argued that the lower part of a dense thickened lithosphere in a convergent zone is gravitationally unstable, and under certain conditions may develop as a Rayleigh-Taylor-type mantle downwelling. This type of convective removal of the mantle lithosphere has been called upon to account for large-scale lithospheric dynamics in a range of tectonic problems [England and Houseman, 1989; Platt and Vissers, 1989; Marotta et al., 1998; Schott and Schmeling, 1998; Houseman et al., 2000; Levin et al., 2000; Saleeby et al., 2003; Billen and Houseman, 2004].

[6] The development of perturbations at the base of the lithosphere into gravitational instabilities requires that the rate of growth of the instabilities exceeds the dissipative effects of thermal diffusion. The growth rates of RT instabilities can be strongly affected by rheological variations such as temperature dependence [Buck and Toksoz, 1983; Lenardic and Kaula, 1995] and nonlinear viscosity [Houseman and Molnar, 1997; Molnar et al., 1998]. In addition, chemical alterations of the mantle lithosphere and associated buoyancy effects may increase the stability of the lithospheric roots [Lenardic and Moresi, 1999; Lee et al., 2001] and inhibit the growth of RT instabilities at the base of the lithosphere. An inherent assumption of RT instability of the mantle lithosphere is that this layer is more dense than the underlying mantle. Recent studies of mantle xenoliths/xenocrysts have been interpreted to imply that it may only be since the Phanerozoic that this density condition has been met, and thus the lithosphere may not have always been prone to gravitational instability [Djomani et al., 2001; O’Reilly et al., 2001]. In general, the propensity of the mantle lithosphere to develop as a RT instability is uncertain. Because of these uncertainties, we consider instead the crustal response when conditions are favorable for the development of a RT-type mantle downwelling. Such responses, quantified in terms of observables such as surface topography and patterns of crustal deformation, can be compared to the geological records of Earth and the terrestrial planets in order to evaluate the relative importance of mantle instabilities in their tectonic evolution. Furthermore, our models are not intended to simulate large-scale instability of long-lived cratonic roots, but instead consider smaller, localized instabilities of mantle lithosphere in certain intraplate environments.
Previous studies have demonstrated that significant intraplate crustal tectonics may be driven by gravitational instability of the mantle lithosphere or deeper thermal mantle downwellings. Subcrustal dynamics have been invoked to explain inferred tectonic deformation at the surface of Venus as a “single-plate” planet [Bindschadler and Parmentier, 1990; Lenardic et al., 1993; Smrekar and Parmentier, 1996]. Lenardic et al. [1993], for example, showed that mantle downwelling may drive crustal thickening (and inferred plateau formation) depending on the rheology of the basaltic crust. Coupled crust-mantle modeling also demonstrates that the deforming crust may in turn alter the style of the underlying mantle flow, as dynamically thickened crust can introduce significant changes in the thermal/mechanical state at the crust-mantle interface [Lenardic and Kaula, 1995, 1996]. For an Earth-like lithosphere, Neil and Houseman [1999] calculate the influence of RT instabilities of layered linear viscous lithosphere models on crustal deformation. They show that mantle lithosphere downwellings can cause continental crust to contract and thicken, and that such events may be interpreted as intraplate orogenies which may occur without any horizontal plate boundary forcing. Dynamical uplift/subsidence of the surface due to viscous mantle flow associated with the downwelling may work in concert with this effect, resulting in a time-dependent interplay between (1) dynamic topography and (2) flow-induced crustal thickening and uplift in contributing to surface topography during an episode of mantle downwelling.

Here, we extend the investigation of intraplate crustal deformation driven by lithospheric RT instabilities using scaled analogue and numerical models of the crust-mantle system. Specifically, the objective of the work is to consider how an underlying mantle driving force beneath a plate interior will manifest in varying styles of tectonic structure at the surface. The surface response will depend strongly on the degree of coupling between the mantle lithosphere and upper crust. This coupling can be expected to drive a time-dependent crustal response resulting from the relative contributions of dynamic subsidence caused by the mantle RT instability and surface uplift due to crustal thickening and isostasy (Figure 1). A suite of analogue and numerical experiments was conducted in which we modify the configuration and rheology of the crust as a method for varying the effective coupling between the mantle and surface. The evolution of surface strain and surface topography are examined as primary observables of the model crust deformation.

We study this geodynamical problem in both a 3-D and 2-D framework. In particular, in a series of analogue experiments we initiate the mantle lithosphere instability as a linear (i.e., approximately two-dimensional) and point (i.e., three-dimensional, approximately axisymmetric) feature. The experiments allow for a 3-D crustal response to the underlying mantle dynamics and illustrate the influence of geometry on the system. The series of approximately 2-D analogue experiments was compared with plane-strain numerical models to extend the quantitative analysis of the modeling and serve as an additional benchmark for the modeling results. In turn, the models also provide insight into how the nature of the crust controls the evolution of mantle lithosphere instability.

2. Experimental Setup

The basic configuration of both the analogue and numerical experiments is summarized in Figure 2. The starting condition for all models consists of asthenospheric mantle overlain by
denser mantle lithosphere, which is covered by a layer of buoyant crust. The experiments are confined by no-slip boundaries at the base and sides (i.e., zero tangential velocity, no material flux at each boundary), but the upper boundary is a free surface. A RT instability at the mantle lithosphere-asthenosphere interface is seeded in the center each model by a hemicylindrical (2-D numerical and linear instability laboratory experiments; Figure 2b) or hemispherical (point instability laboratory experiments; Figure 2c) perturbation. The principal variables are the rheological properties and thickness of the crust; aside from the geometry of the initial perturbation, the thickness and physical properties of the mantle lithosphere and asthenosphere are kept constant. Scaling arguments that set these properties and details of the laboratory and numerical model setups are outlined below.

2.1. Analogue Models

Scaling and the general construction of the laboratory experiments (Figure 2) follows the approach for analogue modeling of lithospheric deformation at one-gravity (1g) described by Davy and Cobbold [1988, 1991]. The experiments are approximately scaled to nature by selection of appropriate dimensions (geometric similarity) and
laboratory materials (dynamic and kinematic similarity). By necessity the scaling is only approximate because of practical considerations in the laboratory (limitations of model size, time and suitable materials) and uncertainties in the natural system (densities, rheologies, time dependence, etc.).

2.1.1. Materials

[12] In designing the laboratory experiments we used as our principal constraints the length scale and the physical properties of the asthenospheric mantle material. For practical purposes we define a length scale based on a model lithospheric thickness, \( l_m = 2.5 \text{ cm} \), which for a natural continental lithospheric thickness \( l_p = 150 \text{ km} \) gives a length scale ratio \( L = l_m/l_p = 1.67 \times 10^{-7} \) (where the subscripts \( m \) and \( p \) refer to laboratory and nature scales, respectively). As our working fluid to model asthenospheric mantle we use polydimethylsiloxane (PDMS), which is a transparent, high viscosity, high molecular weight silicone polymer that is frequently used in analogue tectonic modeling studies [Weijermars, 1986; Cruden et al., 1995]. PDMS has a density \( \rho_m = 960 \text{ kg/m}^3 \), which for a natural asthenospheric density \( \rho_p = 3100 \text{ kg/m}^3 \) sets a density scale of \( P = \rho_m/\rho_p = 0.31 \). In these gravity-driven experiments at \( 1g \) the gravity scale ratio is \( G = g_m/g_p = 1.0 \).

[13] The rheological properties of all ductile materials used in this study were measured with a rotary (Couette) viscometer over a range of strain rates relevant to the experiments. PDMS has a slightly non-Newtonian rheology defined by the flow law:

\[
\sigma^n = \eta \dot{\varepsilon},
\]

where \( \sigma \) is stress, \( \dot{\varepsilon} \) is strain rate and viscosity \( \eta \) and the power law exponent \( n \) are material constants. For scaling purposes we define an effective dynamic viscosity \( \eta_{\text{eff}} = \sigma/\dot{\varepsilon} \), which for PDMS at a laboratory strain rate of \( 10^{-5} \text{ s}^{-1} \) is \( 7.74 \times 10^4 \text{ Pa} \cdot \text{s} \). Assuming an effective viscosity of \( [\eta_{\text{eff}}]_p = 10^{21} \text{ Pa} \cdot \text{s} \) for the natural asthenosphere [Mitrovica and Forte, 1997] defines a viscosity scale ratio \( M = [\eta_{\text{eff}}]_m/[\eta_{\text{eff}}]_p = 7.74 \times 10^{-17} \). The time scale ratio for the experiments can now be defined as \( T = M/PLG = t_m/t_p = 1.50 \times 10^{-9} \).

[14] Thicknesses and properties of the crustal and mantle components of the model lithospheres follow from these scaling parameters and are summarized in Table 1 and Figure 3. Mantle lithosphere in the experiments is made up of a 2 cm thick layer of a mixture of PDMS and Harbutt’s Blue plasticine. Addition of plasticine to PDMS results in increases in density and effective viscosity by factors of 1.08 and 6.4, respectively. Scaled up, the density difference between mantle lithosphere and asthenosphere in nature would be 260 \text{ kg/m}^3.

[15] Four crustal configurations were employed in the experiments (Table 1). Experiments A1 and P1 had a 4 mm thick layer of relatively strong ductile crust composed of PDMS + Harbutt’s White plasticine + low density ceramic microspheres. Experiments A2 and P2 employed 4 mm of relatively weak ductile crust composed of PDMS + low density glass microspheres. Addition of ceramic or glass microspheres results in a lowering of the density of the material and an increase in effective viscosity. In both cases the density of the ductile crust scales to a typical lower-mid crustal density of 2840 \text{ kg/m}^3.

[16] The other two crustal configurations had an additional 1 mm (equivalent to 6 km in nature) of granular material added to the top to simulate a brittle upper crust. The macroscopic behaviour of fractured upper crust is commonly modeled using a Coulomb or frictional sliding criterion, with material constants being the dimensionless angle of internal friction (\( \phi \)) and cohesion (\( \sigma_0 \)), with dimensions of stress. The scaling factor for stress can be defined as \( S = PL = 5 \times 10^{-8} \). Given that natural upper crustal rocks have cohesions \( \sigma_0 \sim 50 \text{ MPa} \), a suitable frictional material in the laboratory should therefore have a cohesion on the order of a few Pa. Because of their low cohesion and internal friction angles similar to rocks (\( \phi \sim 23^\circ \) to \( 42^\circ \)) dry granular materials such as sand and manufactured microspheres have suitable properties to model brittle upper crustal behaviour in the laboratory [Davy and Cobbold, 1988, 1991; Rossi and Storti, 2003]. In order to achieve the desired density, we employed ceramic microspheres with a bulk density of 820 \text{ kg/m}^3, which scales to a typical upper crustal density of 2650 \text{ kg/m}^3.

[17] Experiments A3 and P3 had a layer of brittle crust underlain by the same higher viscosity ductile crust as A1/P1, whereas experiments A4 and P4 had a lower crust composed of the lower viscosity ductile material used in A2/P2. Strength profiles for the model lithospheres of A3/P3 and A4/P4 are shown in Figure 3. Ductile strength values are computed for effective viscosities assuming a strain rate of \( 10^{-5} \text{ s}^{-1} \), which is a reasonable approximation of deformation rates observed in the experiments. The strength curve for the brittle crust assumes \( \sigma_0 = 20 \text{ Pa} \) and \( \phi = 35^\circ \). The cohesion
values of dry granular materials such as ceramic microspheres are notoriously difficult to measure accurately at the very low normal stress levels employed in analogue experiments. The finite strength of the brittle model crust at the surface is therefore only approximate (20 Pa). Likewise, the effective strength of the ductile crust, mantle lithosphere and asthenosphere will vary in the experiments because of variations in local strain rate. Despite these limitations, the strength profiles of the two model lithospheres are significantly different (Figure 3). Experiments A3/P3 and A4/P4 have “strong” and “weak” crusts, respectively. Because their brittle-ductile transition depth and

Table 1. Experiments and Associated Modeling Parameters

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<th>Expt.</th>
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\(^a\)See text for definition of variables. The values scaled to the natural system are given in parentheses. For calculation of the effective viscosity (see text) we assume a laboratory strain rate of 10^{-5} 1/s.

Figure 3. Strength profiles for models (a) A3/N3/P3 and (b) A4/N4/P4 and the (c) “Earth-like” lithosphere and upper mantle (see text for explanation of the generation of these profiles).
mantle lithosphere strengths are the same, in nature this would correspond to lithospheres with similar geothermal gradients, but significantly different lower crustal compositions.

2.1.2. Model Construction

[18] All laboratory experiments were run in a 15 cm (width) × 20 cm (length) × 15 cm (depth) Plexiglass tank. Models were constructed by sequential addition of asthenospheric and lithospheric layers, with sufficient time allowed to minimize the incorporation of air bubbles. Before adding the lithospheric mantle layer a linear (experiments A1-4) or axisymmetric point (experiments P1-4) perturbation was inserted into the top of the asthenosphere material (Figures 2b and 2c). This was done by first creating a suitably shaped indentation with a rigid die that was subsequently filled with lithospheric mantle material. Because creation of this perturbation marks the onset of the RT instability, all times reported for the experiments are measured from this stage onward.

2.1.3. Observation Scheme

[19] Progress in all experiments was monitored by digital photography from the side (measuring lowest point of descending RT-instability) and top (characterizes surface deformation field) as well as by linear scanning of the surface topography (Figure 4). Visualization of surface deformation is aided by imprinting a passive marker grid on the top of the model. In experiments with a ductile-only crust this is done using a transfer method [Dixon and Summers, 1983]. In experiments with a granular upper layer a grid is created by placing a rectilinear mask over the top of the model and sprinkling a thin layer of fine grained white sand through it. Digital photographs of the model side and top surface are taken, when possible, every 0.5–1 hour.

[20] Surface topography measurement is made using a laser triangulation measuring device (Acculux LaserRanger) that is translated linearly over the model surface by an XY-gantry system at a standoff distance of ~13 cm (Figure 4). Eleven scan lines with a spacing of 1.3 cm are made on each pass, which takes about 7 minutes. Individual topographic measurements are made with a spacing of 1 mm along each scan line and the positional accuracy of the XY gantry is ±0.1 mm. The distance resolution of the laser device is ±0.2 mm. Surface scans are repeated at 0.5 hour intervals for the duration of the experiment.

2.2. Numerical Model

[21] A series of two-dimensional numerical experiments (N1-N4) was conducted as a parallel comparison set to the analogue models initiated with a linear mantle lithosphere instability (A1-A4). Neglecting inertial terms and compressibility, the governing equations for the model are

$$\nabla \rho + \nabla \cdot \sigma' + \rho g = 0, \quad (2)$$

$$\nabla \cdot \mathbf{u} = 0, \quad (3)$$

where \( \mathbf{u} \) is the velocity field, \( \sigma' \) is deviatoric stress, and \( \rho \) is pressure. As with the analogue materials, for the fluid materials we assume a power law constitutive relation of the form, \( \sigma^\prime(t) = \eta f(t) \). For the brittle crust, a Coulomb yield criterion of the form \( \sigma_y = \rho \sin \phi + \sigma_0 \), was assumed.

[22] The system of equations is solved using the arbitrary Lagrangian-Eulerian (ALE) finite element method [Hirt et al., 1974; Fullsack, 1995]. This technique is useful for treating high strain materials (e.g., the convecting mantle) and for explicitly tracking moving material interfaces, such as the free surface and internal boundaries (details of the implementation of the method and application to general tectonic problems are described by...
Figure 5
Fullsack [1995]. The numerical routine has been benchmarked rigorously with previous studies [Poliakov and Podladchikov, 1992; Blankenbach et al., 1989; Houseman and Molnar, 1997; van Keken et al., 1997] to verify its accuracy in computing viscous flow and surface topography.

[23] The setup of the numerical model is based on a two-dimensional approximation of the scaled analogue model (Figure 2). The solution space is 14.5 cm deep and comprises a strong, dense mantle lithosphere of thickness 2.5 cm mantle above the mantle region (asthenosphere), overlain by a buoyant viscous crust of variable thickness (Table 1). The density of the computational elements has been concentrated in the topmost 4 cm of the models to increase the stability and accuracy of the lithospheric deformation and surface topography. The top boundary of the box is a free surface and all the other sides are no-slip. Implementation of perfectly free-slip boundary conditions proved to be uncertain in the analogue models, with various techniques for lubricating the sidewalls of the box. However, experimentation and previous laboratory results [Cruden et al., 1995] indicate that no-slip conditions closely represent the sidewall and bottom boundary conditions of the unlubricated box.

3. Results

3.1. Growth of the Mantle Lithosphere Instability

[24] The evolution of the mantle lithosphere instability for the analogue and numerical models, A2 and N2, is shown in Figure 5. The linear mantle lithosphere perturbation evident in the side view frames (t = 1 hour), seeds gravitational instability at the center of the box. Initial growth of the instability is slow, but by t = 16 hours, the downwelling is descending quickly. The downwelling flow entrains a significant amount of the surrounding mantle lithosphere toward the center of the box. This lateral entrainment, in turn, induces a shear stress at the base of the crust, resulting in convergence and gradual crustal thickening directly above the descending mantle lithosphere. Conversely, there is appreciable crustal extension in the off-axis regions from the downwelling. The crustal thickening is especially apparent in the 2-D numerical model.

[25] Between t = 19–23 hours, the downwelling reaches the bottom boundary of the solution space and begins to spread laterally at this depth (Figure 5). Descent of dense mantle lithosphere material continues at the center of the box, but this slows significantly after the head of the downwelling reaches the bottom of the box. Eventually, secondary mantle lithosphere instabilities begin to develop at the sides of the box, with a more symmetrical pattern in the numerical experiment than in the analogue model (Figure 4; t = 42 hours). By this stage there is almost complete accumulation of crustal material above the central and distal downwellings (although this is not apparent in the sideview photo of the analogue experiment). Consequently, the crust is highly thinned and essentially absent across the rest of the surface.

[26] It is interesting to note that by the end of the numerical experiment an appreciable portion of the ductile (lower) crustal material is entrained in the center of the mantle lithosphere downwelling to the bottom of the solution space. A similar phenomenon was observed in the other numerical experiments, and in the analogue models. Although this is not evident from the external profile photos of the experiment A2, sections cut through the center of the box at the end of the experiment found that lower crust was entrained to the base of the box within the mantle lithosphere downwelling.

[27] In Figure 6 we plot the normalized descent distance of the mantle lithosphere downwelling to illustrate the relative growth rates of the instability for the various rheological stratifications described in Table 1 and Figure 4. In all the models, the mantle lithosphere instability is characterized by an initial slow growth, followed by rapid descent. This first stage of growth is consistent with the exponential growth demonstrated for lithospheric Rayleigh-Taylor instabilities [Houseman and Molnar, 1997; Conrad and Molnar, 1997; Molnar et al., 1998]. Eventually the decelerating effects of the bottom boundary in the models cause the downwelling to slow down in the lower ~10–20% of its descent. This latter observation is consistent with 3-D
models of slab subduction where a modification of subduction behavior occurs as a descending slab reaches a strong viscosity interface at 660 km depth [Funicello et al., 2003].

[28] The results demonstrate that crustal rheology has a strong influence on the growth rate of the mantle lithosphere instability (Figure 6). For the viscous-only crust models, there is least impedance on the growth of the mantle lithosphere instability, promoting an earlier, and slightly faster, descent than for models having a brittle crust. Generally, with a stronger viscous crust (N1, P1) the growth of the instability is slightly delayed compared to the weak crust models (N2, P2), although this does not occur with the linear instability analogue experiment (A1 compared to A2). In the analogue experiments, variations in initial growth rate may be partly controlled by differences in the volume of material in the initial perturbation. However, in all experiments with ductile-only crust the descent rate (i.e., the slope of the curve during the main phase of descent) is essentially the same.

[29] Experiments with a brittle crust show the longest delay in development of mantle lithosphere mantle instability and significantly slower descent rates. For example, comparing experiments N1 to N3, which differ only by the presence of a brittle crust in N3, the downwelling takes 10 hours more (i.e., 40% longer) to reach a normalized distance of 0.8 (Figure 6b). Even with a weak lower crust which tends to decouple mantle flow from the brittle upper crust (A4, N4, P4) the growth of the instability is slower than in the strong ductile-only crust model. This suggests that the integrated strength of the crust, together with the mantle lithosphere-asthenosphere density contrast, controls the response of the mantle lithosphere downwelling.

[30] The geometry of the mantle lithosphere instability (i.e., linear versus axisymmetric; Figure 6a versus Figure 6c) does not have a major influence on the growth rate. In the out-of-plane direction to the analogue model frames shown in Figure 5, there is some variation in the development of the downwelling; that is, it is not a perfect 2-D sheet or axisymmetric cylinder. This may result in some of the detailed variation in the descent curves, such as the crossover in the curves for P1 and P2, which is due to a transition in the geometry of the downwelling in P2 from axisymmetric to quasi-planar. However, the growth of the maximum depth point in experiments with planar instabilities is consistent with those with axisymmetric downwellings.

[31] The primary difference between the results from the numerical and analogue experiments is a shift in the onset of the rapid growth stage of the mantle lithosphere downwelling. In the numerical experiments, this rapid growth stage is delayed by ~5 hours. This discrepancy may be due to uncer-
tainty in the precise size and form of the initial mantle lithosphere perturbation introduced into the analogue models. For example, for the analogue models the mantle lithosphere perturbation was emplaced before several stages where the overlying mantle lithosphere and crustal layers were added and allowed to stabilize. During this final assembly, the perturbation would have begun to descend into the lower layer. Although we have tried to correct for this time lag, some offset in the data may still remain. As previous studies of the growth of Rayleigh-Taylor instabilities have shown [Houseman et al., 1981], the initial amplitude of the perturbation will strongly affect its growth rate.

3.2. Evolution of Surface Strain

[32] Figure 7 shows a series of top view photographs for four of the analogue experiments. For the experiments with a linear mantle lithosphere perturbation the orientation of the perturbation is vertical in these frames (Figures 7a and 7b).

[33] In experiment A2 the crust is purely viscous and is relatively weak (Table 1). The initial rectangular surface grid deforms as the instability begins to descend and laterally entrains mantle lithosphere material and crust toward the downwelling (Figure 7a; \( t = 1–16 \) hours). Crustal convergence above the downwelling is readily apparent from \( t = 16–23 \) hours as the marker grid lines increasingly concentrate at the center of the box. The character of the crustal deformation is approximately plane-strain, with the zone of convergence forming above the underlying linear downwelling in the mantle lithosphere. However, no-slip side wall boundary effects hinder crustal convergence close to the edges of the experiment. This is also reflected in the evolution of the initially linear mantle lithosphere downwelling, which amplifies slightly more quickly at the center of the box than toward the sidewalls. This results in a subtle secondary component of convergence toward the center of the box parallel to the orientation of the linear downwelling.

[34] By \( t = 40 \) hours, accumulation of crust in the central zone of convergence has ceased, and this crust begins to spread out (compare the spacing of the grid lines in this region between \( t = 23 \) hours and \( t = 40 \) hours; Figure 7a). This occurs due viscous relaxation of the thickened crust once the mantle instability reaches the bottom of the box, resulting in a cessation of the force driving lithospheric deformation. Extreme crustal thinning occurs on either side of the convergent zone and away from the sides of the box. By the end of the experiment at \( t = 40 \) hours almost complete thinning of the crust and mantle lithosphere in these regions has occurred and the clear PDMS asthenospheric mantle is visible. The general behaviour of the model, with an approximately plane-strain style of flow-induced high-strain crustal deformation, is also recorded in experiment A1. The stronger viscous crust in this experiment only alters the rate at which similar crustal deformation occurs.

[35] Several small anomalous light-colored features appear on the surface of A2 that disturb the marker grid in the model are caused by small air bubbles which rise to the model surface during the experiment (Figure 7a). These features cause only an aesthetic disturbance and have no influence on the general evolution of the experiments.

[36] In experiments with a brittle upper crust and ductile lower crust the surface deformation is markedly different than those with a viscous-only crust (Figure 7b). In experiment A3, no detectable surface strain is observed at the early stages of the experiment (\( t = 1–19 \) hours), despite the extreme deformation that must be occurring in the ductile lower crust (e.g., Figure 7a). Through \( t = 23–40 \) hours, two linear zones of localized extensional deformation appear at the surface over the regions of asthenospheric upwelling on either side of the mantle downwelling. Notably, there is still negligible shortening of the surface crust along the central portion of the model at this stage of the experiment. The dominance of extensional features and lack of compressional features is not surprising given the lower strength of frictional material in extension [e.g., Ranalli and Murphy, 1987]. As in the previous model, air bubbles cause minor deformation of the surface, resulting in small oval depressions at the periphery of the model.

[37] Models equivalent to A2 and A3 but with a central, axisymmetric point initial perturbation in the mantle lithosphere show significantly different crustal deformation patterns (Figures 7c and 7d). In experiment P2 the viscous crust is strongly deformed by the descending mantle lithosphere instability (Figure 7c). At \( t = 16 \) hours the pattern of axisymmetric deformation is clearly visible as a high concentration of marker grid lines develops in the center of the box above the developing cylindrical-shaped mantle lithosphere downwelling. A zone of particularly high surface marker strain delineates the diameter of the downwelling structure. Through \( t = 21–40 \) hours increasing convergence and accumulation of crust occurs directly...
above the developing instability and in four spoke-like features that trend toward the corners of the box. As in experiment A2 significant crustal thinning occurs in regions of active asthenospheric upwelling. With time the mantle lithosphere instability migrates toward the lower sidewall resulting in a slightly asymmetric pattern of crustal deformation. Once the downwelling reaches the base of the box the region of greatest crustal thickening undergoes gravitational collapse. Unlike A2, surface extensional strain is localized over the downwelling, which has been shifted toward the lower edge of the box.

[38] Experiment P3 illustrates deformation of brittle upper crust and strong viscous lower crust overlying an axisymmetric mantle lithosphere instability (Figure 7d). At $t = 16$ hours there is very subtle, but perceptible deformation of the surface as the marker grid lines are slightly

Figure 7. Plan view photos for four of the analogue experiments (as indicated). Marker grids (dark lines) are traced on the light colored crustal surface of each model, and deformation of these grids track surface strain as the experiments progress. In experiments with a viscous crust, A2 and P2, blue mantle lithosphere and clear mantle material become evident in regions of high extension ($t \geq 16$ hours). Labeled time is relative to the start of each experiment.
The pattern of brittle upper crustal deformation observed in P3 is consistent with the ductile crustal strain field recorded in P1 (not shown) and P2. Although sidewall boundary effects clearly influence the deformation pattern, the surface strain field and the geometry of the underlying RT instability strongly resemble so-called diapirc “spoke patterns” observed in previous laboratory studies [Talbot and Jackson, 1987; Talbot et al., 1991]. Spoke patterns develop during self-organization of multiple RT instabilities during overturn of a gravitationally unstable interface whose width is >> the dominant wavelength of the individual instabilities. Such patterns are characterized by polygonal arrays of buoyant material (negative or positive) which accumulate in linear zones (spokes) that radiate away from buoyant axisymmetric instabilities (or fingers). Similar patterns are also observed in laboratory experiments of thermal convection [McKenzie and Richter, 1976]. Hence the spoke-like geometry of convergent deformation and complementary annular pattern of extensional deformation observed in experiments P1, 2 and 3, although influenced here by boundary effects, may be a significant and characteristic feature of natural axisymmetric mantle RT downwellings.

Surface deformation in experiment P4 (not shown), which had brittle crust overlying weak ductile crust, was minimal as in A3 and A4. We attribute this to the decoupling effect of the weak ductile crustal layer, which prohibited the transfer of sufficient shear stress to the base of the brittle layer to allow it deformation to occur.

There are several potential reasons why deformation of the brittle upper crust is more apparent in experiment P3 compared to A3. Firstly, the inherent geometry of the axisymmetric instability means that the mantle flow-induced stresses in the crust are more localized at the region of central convergence than is the case with the linear instability. Secondly, the mantle lithosphere instability in P3 may be larger and more energetic than the instability in A3, which may be caused, for example, by uncertainties in the initial size of the perturbation. Finally, we speculate that short-wavelength deformation of the brittle crust may also be facilitated by the presence of preexisting heterogeneities. In experiment P3 the central convergent structures and extensional system at the right side of the box both develop early and near preexisting locations (from \(t = 16\) hours) of crustal deformation associated with air bubble transit through the crust (Figure 7d).

In comparing this set of laboratory experiments to the numerical models, we note that a similar pattern of high surface strain and crustal deformation occurred in the numerical experiments when there was a purely viscous crust. For example, Figure 8a shows the top portion of the 2-D numerical experiment N2 at \(t = 19\) hours. The highly deformed Lagrangian mesh near the surface demonstrates the strong shortening of crust above the mantle lithosphere instability and adjacent extension. However, in all the numerical experiments with a brittle upper crust, the top surface experienced relatively little deformation (Figure 8b). Notably, localized crustal structure, such as is evident in Figures 7b and 7d did not develop. This is likely the result of the absence of heterogeneity in the numerical crust, except for extremely small computational round-off error. Furthermore, without an explicit mechanism to account for strain localization in the rheological parameterization of the numerical models, crustal deformation tends to remain distributed.

### 3.3. Surface Topography

Although deformation of the upper is crust is subtle in most models having brittle crust, flow occurring in the ductile lower crust has an appreciable effect on the topographic evolution of the surface. In Figure 9 wire frame plots of model surfaces illustrate the evolution of the top surface of analogue experiments A4 and P4.

The initial surface in experiment A4 reflects the essentially random, low amplitude roughness of the poured granular material of the upper crust layer (Figure 9b; \(t = 2\) hours). By \(t = 16\) hours, a distinct surface subsidence trough develops,
coincident with the strike of the linear mantle lithosphere downwelling. The maximum amplitude of the subsidence is $\approx 0.5 \text{ mm}$, at the center of the trough. Subsequently, the negative topography through this region inverts, and a linear belt of positive topography is established by $t = 32–41 \text{ hours}$. The coherent pattern of linear subsidence and inversion of topography demonstrates that these events are related to the evolution of the underlying linear mantle lithosphere downwelling.

[45] The general response of the model surface in experiment P4 is similar, except that the geometry of the surface depression and subsequent uplift is approximately circular in plan reflecting the axial symmetry of the mantle downwelling (Figure 9b). As noted in the surface strain field for P2, the locus of surface depression and uplift in P4 migrates toward one side of the box with time due to changes in the shape and locus of the downwelling during the experiment.

[46] In both examples, corresponding areas of surface uplift are observed in the regions adjacent to the mantle downwelling during the first 24 hours of the experiment, above regions of asthenospheric upwelling (Figure 9). Shorter wavelength topographic features form due to the crustal response to developing secondary mantle instabilities. Very short wavelength spikes are air bubble-induced surface features, and the majority of these tend to occur near the model edges. Initial small variations in topography tend to persist during the progress of the experiments and these also contribute to the medium to long wavelength topography of the models. The topographic complexity of the model surfaces provides some insight into how a natural planetary surface might respond to a three-dimensional mantle RT instability over geological timescales.

[47] The variation in surface topography at a fixed point on the surface is plotted as a time series for each of the experiments in Figure 10. For the case of the analogue models, the point was chosen at the center of the surface of the box, directly above the initial mantle lithosphere downwelling. The topography was visually determined from surface profiles generated during the laser scanning process and the error bars ($\pm 0.2 \text{ mm}$) reflect the uncertainty in the measurements themselves, the visual fit of the line scanning data and the likelihood that the true locus of maximum surface subsidence or uplift will lie between scan lines. For the numerical models, the point is at the middle of the upper surface of the solution space. In general, the trends of evolving surface topography are consistent between the analogue and numerical models. In each experiment, the topographic evolution is characterized by two primary phases: (1) Initially

Figure 8. Numerical experiments (a) N2 at $t = 19 \text{ hours}$ and (b) N3 at $t = 30 \text{ hours}$ showing material regions and Lagrangian mesh. Lagrangian mesh initially consists of regular, rectangular cells; distorted cells indicate progressive deformation of materials in model (mesh plotted at one tenth of actual resolution; highly deformed cells are not plotted).
Figure 9. Wire frame plots of the evolving top surface for analogue experiments A4 and P4. These plots were created by linear interpolation between the 11 topographic profiles generated during periodic scanning of the model surfaces using the laser-XY gantry setup described previously. The interpolation method used was kriging with no adjustments for anisotropy, and the resulting gridded data were filtered using a $3 \times 3$ Gaussian low-pass filter.
the crust subsides above the mantle lithosphere instability. This is caused by the fluid normal stresses acting on the base of the crust/lithosphere (Figure 1) during the vigorous initial growth of the mantle lithosphere downwelling. (2) Gradually, there is an inversion from subsidence to uplift.

[48] The inversion is a consequence of both a decrease in the fluid dynamic subsidence and intrinsic buoyancy produced by crustal thickening above the mantle lithosphere downwelling. Although dense mantle lithosphere material is entrained into the downwelling throughout the duration of the experiment, the vigor of fluid flow decreases substantially when the initial mantle lithosphere downwelling reaches the bottom of the box or solution space, which results in a partial recovery of the dynamic subsidence. This is illustrated in Figure 11, where a time series of the growth of the mantle lithosphere instability is superimposed on a plot of the topographic evolution of experiment A1. Clearly, the maximum subsidence correlates with the inflection point of the growth curve for the instability. This indicates that subsidence progresses as long as the downwelling is accelerating, but dynamic topography is recovered as the descent of the downwelling slows. The additional uplift, caused by the isostatic response of the surface to flow-induced crustal thickening (Figure 1), is strongly dependent on the rheology of the crust.

[49] The experiments with a viscous-only crust (Figures 10a–10c) produced qualitatively similar results to those having a brittle upper crust and ductile lower crust (Figures 10d–10f). However, we focus on the latter models, since the former models generally produced less reliable topographic results. This is a consequence of the ability of the laser beam to penetrate into the ductile upper

Figure 10. Plots of surface topography versus time for (a–d) viscous crust models and (d–f) brittle upper crust/viscous lower crust models. Surface topography is measured at a point directly above the mantle downwelling at the center of model box.
crustal materials under certain circumstances (incidence angle, material composition and thickness) thereby measuring the depth to internal reflections within the crust, rather than giving a precise measurement of the top surface.

For the strongest crust end-member experiments (strong lower crust, brittle upper crust; A3, N3, P3), the topographic evolution is dominated by subsidence. The strong upper crust undergoes relatively little deformation and the higher viscosity lower crust is more resistant to horizontal flow-induced thickening; hence the uplift phase is subdued and primarily represents recovery of the flow-induced subsidence. For instance in the analogue models A3 and P3, the maximum subsidence reaches 0.8 mm and 1.2 mm, respectively, and in both cases, this negative topography is not fully recovered by the end of the experiments. When scaled to actual Earth dimensions, these results correspond to significant subsidence events of magnitude 4.8 km and 7.2 km. Despite the uncertainties in topography measurement in ductile-only crust models and the tendency of the locus of the maximum downwelling to drift away from the model center there is a clear tendency for the topographic response (subsidence and uplift) to be muted in experiments that include a brittle crust. We attribute this to the damping effect that the strong brittle layer likely has on both lower crustal flow and the dynamics of the downwelling itself. Given the presence of a brittle upper crust on Earth and the terrestrial planets we therefore consider the brittle + viscous crust models to provide the most realistic insight into likely topographic responses to mantle RT instabilities, whereas the viscous-only crust experiments provide important information about the nature and contribution of lower crustal flow.

4. Summary of Experimental Results

The models demonstrate various aspects of the coupling between mantle instabilities and crustal tectonics:

- Downwelling of dense mantle lithosphere can drive various styles of crustal deformation including episodes of long-wavelength crustal shortening and extension, as well as complex structure over small length scales. The flow-induced deformation results in a time-dependent response of surface topography. The interplay between crustal thickening/thinning and dynamic topography supported by the underlying mantle flow governs various phases of subsidence and uplift of the crustal surface. The experiments also indicate that the crustal strength, in turn, can significantly affect the growth rate of the mantle lithosphere instability.

- The character of surface response is strongly influenced by the geometry of the mantle lithosphere downwelling. In particular, the laboratory experiments demonstrate that even in these simple systems, a 3-D geometry can allow for significantly enhanced complexities of the crust-mantle evolution. For example, in experiments A1–A4, the mantle lithosphere downwelling has a propensity to evolve away from the initially imposed linear geometry toward a mixed plume-sheet feature.
While this may be the consequence of the boundary conditions in the models, this observation is consistent with other studies of planar RT instabilities, which have a tendency to break up into fingers as they amplify [Kerr and Lister, 1988]. The transition in geometry of the mantle lithosphere forcing may manifest either in significant variations in the crustal surface response (Figure 7a), or with essentially no crustal signature (Figure 7b), depending on the crustal rheology.

- The surface tectonics are strongly controlled by the mechanical coupling between the brittle and ductile crustal layers, and the underlying mantle lithosphere and mantle. In both linear and point instability experiments, when coupling between the upper and lower crust is reduced due to a weaker lower crust (A2, A4) a relatively quiescent upper crust may be underlain by strong deformation and mobility of the lower crust and mantle lithosphere. In these cases, the surface responds by vertical motions only, while the ductile lower crust accumulates strong horizontal fabrics due to flow induced by the underlying mantle instability. When significant coupling occurs between the upper and lower crust (stronger ductile crust; A1, A3), both the brittle and ductile crust experience significant horizontal deformation, particularly when contractional deformation above the down-going mantle is localized by a point instability.

- Deformation of strong crust may be enhanced by local physical or thermal heterogeneities. In models with an effective coupling between the upper and lower crust, mantle flow-induced complex surface deformation may nucleate at anomalous crustal features.

- All experiments demonstrate viscous entrainment of the lower crust to depth within the dense mantle lithosphere downwelling.

5. Discussion

In general, the modeling results demonstrate how complex tectonic structure may develop within the lithosphere in the absence of large-scale plate boundary effects. In addition to plate interiors on Earth, this would include crustal deformation on relatively inactive planetary surfaces. Enigmatic, complex tectonic structure such as localized regions of contraction and extension, fault/fracture patterns, and other short-wavelength topographical features, have been interpreted for the surface of Venus [e.g., Bindschadler et al., 1992; Solomon et al., 1992]. As discussed, previous studies have invoked mantle flow mechanisms to account for large-scale topographic features on the surface [Bindschadler and Parmentier, 1990; Lenardic et al., 1993; Smrekar and Parmentier, 1996]. Our models indicate that even short wavelength tectonic deformation on the surface may be driven by mantle dynamics in the absence of planetary plate tectonics. The surface morphology of experiment P3 (Figure 7d), for example, shows that a broad and rather simple mantle downwelling event can be responsible for complex, small lengthscale structure of the crust. With the exception of erosional processes, which have not been modeled here, the effects of secondary mantle instabilities, initial crustal thickness and relief variations, and short wavelength features (related to volcanism?) can be expected to be superimposed on the first order response to a major mantle downwelling.

The various styles of crustal behaviour in the models have implications for the geometry and relative age of seismic reflectors in the crust beneath intraplate basins and orogens. For example, there is considerable debate as to whether strong subhorizontal fabrics observed in deep crustal sections and seismic reflection data in orogenic belts are related to early subduction-accretion processes, post accretionary crustal shortening, late orogenic crustal collapse and/or removal of the mantle lithosphere [Rey, 1993; Meissner and Tanner, 1993; Meissner and Mooney, 1998; Burg et al., 2002]. Our models indicate that late to post-orogenic gravitational removal of dense subcrustal lithosphere can result in significant lower crustal flow that will be recorded by strong subhorizontal fabrics, as well as time dependent change in surface topography. For example, RT instability of dense subcrustal material beneath the Sierra Nevada has been invoked to explain the absence of a lithospheric root beneath the batholith, extension and strong deep crustal seismic fabrics of the adjacent Basin and Range province and the uplift and heat flow of the Sierran mountain range [Meissner and Mooney, 1998; Saleeby et al., 2003]. Our linear instability models are generally consistent with this hypothesis in that they predict significant uplift above the mantle downwelling at a late stage of its development (e.g., Sierra Nevada) and crustal extension due to asthenospheric upwelling in the adjacent region (e.g., Basin and Range). Furthermore, shorter wavelength features such as the Pleistocene annular Tulare Lake basin (maximum subsidence 200 m) and related fluvial-alluvial flooding of topography at the margin the batholith, suggest that there is an active axisymmetric RT
instability beneath the western margin of the central Sierra Nevada [Saleeby and Foster, 2004]. On the basis of our experiments, this stage of axisymmetric downwelling may have evolved from an earlier planar instability that developed under the length of the batholith in the Neogene [Meissner and Mooney, 1998].

[63] In the models, flow-induced crustal deformation is favored for a relatively low bulk crustal strength. While the Earth’s present crust is strong [Kohlstedt et al., 1995], there may be localized regions or times in Earth’s past in which the crust is more susceptible to such tectonic forcing. For example, at inferred higher Archean mantle temperatures, the crust may have been less “plate-like” [West, 1980; Hamilton, 1998]. Modeling of the early Earth shows that deformational and melt processes may have been active in continental interiors and that intraplate events may account for various geologic observables [Vlaar et al., 1994; Zegers and van Keken, 2001].

[62] As described for the models, certain features of the crustal structure may nucleate at preexisting structural elements caused by the passage of an air bubble through the model lithosphere (Figure 7d). In a similar respect, it has been suggested that localized regions of high radioactive heat production in the crust may be especially vulnerable to intraplate deformation. The crust in parts of central Australia, for example, contains a relative abundance of highly radioactive granites and sediments. In these regions, there may be a causal relationship between the radioactive heating and consequent weakening of the crust, resulting in distinct phases of intracraticon deformation [Sandiford and Hand, 1998; McLaren, 1999; McLaren and Sandiford, 2001]. In addition, changes in the distribution of the radioactive materials during progressive evolution of the crust may result in various phases active/inactive tectonism [Sandiford and McLaren, 2002]. Thermal-mechanical models indicate that under the influence of RT-type instability of the mantle lithosphere, otherwise strong crust may experience significant episodes of deformation, if there is a certain magnitude and distribution of radioactive elements within the crust [Pyskylywec and Beaumont, 2004]. These findings suggest a possible explanation for what appears to be an anomalous preponderance of intraplate orogenesis in Australia during the Proterozoic and Paleozoic. In contrast, North America lacks such widespread highly radioactive crust and intraplate orogens, but has several intracraticon sedimentary basins. This would be more consistent with experiments A4/N4/P4 (Figure 10; brittle upper crust, weak lower crust), which demonstrate primarily surface subsidence, and relatively little deformation of the strong upper crust and decoupling from the highly deformed weak lower crust.

[63] Entrainment of viscous crust to depth within the mantle lithosphere downwelling is a common feature of the models, and is a behaviour predicted by previous laboratory and numerical investigations of three-layer RT instabilities [Cruden et al., 1995]. In the case of a buoyant RT instability the degree of entrainment is controlled by the gravitational resistance of the nonbuoyant underlying layer and the viscosity contrast between the layers. When applied to our experiments the results of Cruden et al. [1995] predict that about 15% of the original cross-sectional area of the viscous crust should be entrained into the mantle RT instability and that this material should be drawn down to the base of the box. This prediction is consistent with observations in both the laboratory and numerical experiments and suggests that significant amounts of crustal material can possibly be entrained to great depth in the mantle. These results suggest a greater depth of entrainment into the mantle than other analyses that considered subduction of continental lithosphere [Molnar and Gray, 1979]. The strong entrainment in our experiments compared to estimated continental subduction may be due to the larger driving buoyancy contrast between the mantle lithosphere and sublithospheric mantle for the development of the RT instability (Table 1).

[64] Limitations of the modeling include confinement of the dynamics within a limited modeling space and the omission of thermal effects, namely the temperature dependence of rheology. With respect to the former, boundary effects had a consequence on certain aspects of the evolution of the experiments. Patterns of surface strain at the margins of the model box (Figure 7), for example, clearly reflect interaction of the materials with the no-slip sidewalls. We have restricted the bottom boundary to a scaled equivalent depth of 870 km by assuming mantle convective flow is largely occurring within the upper mantle. This may be partially justified by estimates of mean mantle viscosity which suggest there is an increase in viscosity of up to two orders of magnitude from the upper to lower mantle through this region [Mitrovica and Forte, 1997]. Furthermore, the endothermic phase transformation of olivine at ~670 km depth may introduce buoyancy effects
that may significantly hinder mass flux across the boundary [Christensen and Yuen, 1984, 1985].

Temperature dependence of rheology can have an important influence on the growth of the mantle lithosphere instability [Buck and Toksöz, 1983; Lenardic and Kaula, 1995] and coupled mantle crust evolution [Pysklywec and Shahnas, 2003]. However, we assume the growth rate of the instability is higher than the rate that thermal diffusion would “erode” the mantle lithosphere perturbation. In the models, the vigorous descent of the mantle lithosphere instability occurs over timescales of ~20 hours. On the basis of the characteristic time scaling for the models, T = tp/tp = 1.50 × 10⁻⁹, this corresponds to 1.5 Myr, which is a short timescale compared to rates for effective heating/erosion of the lithospheric instability [e.g., Houseman and Molnar, 1997]. We must note, however, that we do not place a strong emphasis on an exact time scaling of the experiments to nature, since the growth rates of the model instabilities are probably short owing to rather large density contrasts and somewhat weak mantle lithosphere in the experiments.

Despite the approximations of the modeling, the experiments help to account for some first-order behaviours of the lithosphere for the Earth and other terrestrial planets. Namely, intraplate orogens and basins, localized tectonic structure, and deep crustal seismic fabrics may arise as the order behaviours of the lithosphere for the Earth of the original manuscript.

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