

Continental magmatism, volatile recycling, and a heterogeneous mantle caused by lithospheric gravitational instabilities

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[1] Removal of the lower lithosphere (mantle lithosphere with or without portions of the crust) through ductile gravitational instabilities can produce magma under continents. Using numerical experiments approximating the rheology of continental crust and lithosphere and underlying asthenosphere and using phase equilibria from the literature, I investigate the topographic and magmatic results of lithospheric gravitational instabilities, along with the fate of the sinking material. Lithospheric removal, commonly referred to as delamination regardless of the mechanism, may allow asthenospheric material to rise and to melt adiabatically, and this asthenosphere can conductively heat portions of the lithosphere previously at lower temperatures. The size and rheology of the sinking material greatly influence the resulting surface topography as well as whether or not melting occurs. The sinking material may devolatilize as it reaches higher temperatures and pressures, just as a subducting slab does, triggering further melting. Gravitational instabilities are possible causes of nonmagmatic basins, continental magmatism of varying volume and composition in the absence of subduction, areas of high heat flow and uplift, and creation of an upper mantle heterogeneous in major and trace elements and volatiles. Magmatism can be simultaneous with topographic subsidence and possibly with subsequent uplift. In those cases that produce magma, melting can occur over a depth range from ~30 to 200 km, though anomalously hot mantle is required to reach the volume of flood basalts.

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1. Introduction

[2] Loss of the lower lithosphere by sinking into the mantle is commonly called delamination in the literature [e.g., Kay and Kay, 1993]. The term delamination, however, implies loss of a laminar structure, for example, a brittle oceanic plate falling away from beneath a collisional orogen (the “flake tectonics” of Oxburgh [1972], and see Bird [1979], Moore and Wiltschko [2004], Morency and Doin [2004]). Here loss of the lower lithosphere is hypothesized to occur in a ductile manner, first by flowing horizontally to create a growing boundary perturbation and then sinking vertically as a gravitational instability [e.g., Houseman *et al.*, 1981; Fleitout and Froidevaux, 1982; Neil and Houseman, 1999; Conrad and Molnar, 1999; Conrad, 2000; Schott *et al.*, 2000; Elkins-Tanton, 2005; Hoogenboom and Houseman, 2006]. This mechanism requires no specific structural weakness beyond a dense region in the lithosphere that is gravitationally unstable with respect to the underlying mantle

and that possesses a rheology conducive to flow [Schott *et al.*, 2000; Elkins-Tanton and Hager, 2000].

[3] Lithospheric thinning has been inferred from increases in crustal heat flow in specific regions, rapid regional uplift, and from the appearance of signature high-potassium magmas such as lamprophyres, leucitites, and absarokites [e.g., Kay and Kay, 1993; Ducea and Saleeby, 1998; Schott and Schmeling, 1998; Farmer *et al.*, 2002; Elkins-Tanton and Grove, 2003; Gao *et al.*, 2004; Lustrino, 2005; Hoernle *et al.*, 2006]. Seismic studies also support ductile delamination in specific areas [e.g., Jones and Phinney, 1998]. (For a list of candidate locations and references, see Elkins-Tanton [2005].)

[4] Geochemical arguments appear to require foundering of crustal and mantle lithospheric materials to balance elemental budgets. Though continental crust and mantle are complementary reservoirs with respect to most trace elements, Lee *et al.* [2006] observes that the continental crust is too felsic to be derived directly from the mantle [see also Gromet and Silver, 1987]. A possible solution is the loss from the continental lithosphere through delamination of mafic residues from fractionation of mantle melts. Further, Plank [2005] finds that the elements thorium and lanthanum are unfractionated during the processes of subducting sediments and producing arc magmatism, while their significant fractionation in continental crust can best be explained by the loss of mafic residues from igneous differentiation back into the mantle through gravitational

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instability. *Plank* [2005] finds that foundering of 25 to 60% of lower crustal cumulates is required to have occurred throughout geologic history to balance thorium and lanthanum budgets.

[5] Gravitational instabilities form in regions with higher density than the material horizontally adjacent to them. Many researchers have focused on instabilities formed through crustal thickening, for example, during a convergent orogen [e.g., *Houseman et al.*, 1981; *Fleitout and Froidevaux*, 1982; *Conrad and Molnar*, 1997; *Schott and Schmeling*, 1998; *Schott et al.*, 2000; *Morency et al.*, 2002]. *Kay and Kay* [1993] suggest that phase changes in the deepest regions of lithospheric roots may add to density instabilities: If a lithospheric root is pushed into the eclogite stability field it may then become denser than its surroundings due to phase changes. *Schott and Schmeling* [1998] conclude that a lithospheric root must be at least 100 to 170 km deep to create the negative buoyancy necessary for delamination.

[6] In a lithosphere lying within the eclogite stability field but itself gravitationally stable, intruding melts may freeze as eclogites and thus create a dense lower lithospheric region. Melt intrusion and freezing may occur in an arc setting [*Jull and Kelemen*, 2001], or above a mantle upwelling [*Elkins-Tanton and Hager*, 2000]. In addition, the mafic residua created by fractionation of erupted melts may create a dense lower lithosphere. *Kay and Kay* [1993] estimate that every addition of 10% eclogite to the lower lithosphere increases its density by about 1%. *Jull and Kelemen* [2001] find that lower crustal and mantle compositions that result from arc magmatism are likely to exceed the mantle density by 50 to 250 kg m⁻³, corresponding to about 1 to 5% density contrast. Density contrasts in this range are fully sufficient to drive gravitational instabilities.

[7] Because peridotite that has experienced partial melting is gravitationally buoyant with respect to fertile peridotite but cratons appear to be neutrally buoyant, *Kelly et al.* [2003] were drawn to conclude that cratonic crust or mantle lithosphere must contain denser regions to offset the highly buoyant depleted peridotite. Both *Lee et al.* [2005] and *Sleep* [2003] present geochemical and geophysical evidence that the lowest mantle lithosphere is enriched with respect to the upper mantle lithosphere, supporting the hypothesis that portions of the lithosphere may be gravitationally unstable, and also consistent with the hypothesis of melt injection, refertilization, and density addition of the lowest lithosphere. *Poudjom Djomani et al.* [2001] calculated likely density profiles in Archaen, Proterozoic, and Phanerozoic cratons based on heat flow and xenolithic compositions, and found that Phanerozoic lithosphere thinner than about 100 km is gravitationally unstable with respect to adjacent asthenosphere. These results imply that younger lithosphere is most likely to sink via gravitational instability and support the lithospheric thicknesses used in the numerical experiments in this paper.

[8] Timescales of formation and fall of instabilities have been discussed in detail by a variety of researchers [e.g., *Lord Rayleigh*, 1883; *Chandrasekhar*, 1961; *Whitehead*, 1988; *Hess and Parmentier*, 1995; *Conrad and Molnar*, 1997; *Houseman and Molnar*, 1997; *Neil and Houseman*, 1999; *Morency et al.*, 2002; *Elkins-Tanton et al.*, 2002; *Zaraneek and Parmentier*, 2004; *Hoogenboom and Houseman*, 2006]. Here, the great significances of the time required

for an instability to form and fall are the resulting size and temperature of the instability and the resulting topographies on both the bottom and top of the lithosphere. This paper specifically presents predictions for topography and melt production, quantities that can be measured on the Earth and other planets.

2. Numerical Experiments

[9] The models include a lithosphere with a flat lower boundary (no lithospheric root) into which a region of higher-density material has been injected. This denser material is an analog of magmatic flux that has frozen as an eclogite, or has left behind dense mafic cumulates. This material is therefore both denser and warmer than the surrounding lithosphere, and it has a different composition. The denser composition is modeled as a half cosine wave in x with a maximum at the left side of the axisymmetric model box, falling to zero in the positive x direction, and a half cosine wave in z , with a maximum at the bottom of the mantle lithosphere, falling to zero vertically in z . The width of the dense regions is twice the height, except as noted in Table 1. Temperature is also added to the lithosphere in the same pattern, bringing the region of intrusion closer to the mantle temperature beneath and correspondingly lowering its viscosity. In some numerical experiments the density and temperature is added throughout the thickness of the lithosphere, and in others through only half the thickness of the lithosphere.

[10] The effect of a buoyant crust is examined by varying the relative thicknesses of a crustal layer and its underlying mantle lithosphere [see also *Neil and Houseman*, 1999; *Houseman and Hoogenboom*, 2006]. The crust differs from the mantle lithosphere in these models only in its buoyancy. The modeled crust and mantle lithospheric material both possess the same temperature dependence of viscosity and the same thermal diffusivity as the underlying convecting mantle, though the cooler temperatures in the lithosphere and crust produce higher viscosities.

[11] In some models a weak lower crustal layer is added, modeled after example H1 of *Ranalli and Murphy* [1986], which uses a hot geotherm in a granitic crust and shows a steep decrease in strength at 30 km depth, interpreted as the Moho. *Lamontagne and Ranalli* [1995] further show that depending upon its composition and the composition of the surrounding materials, warm lower crust can be up to 3 orders of magnitude lower in viscosity than the cool upper crust and lower lithosphere which surround it. When a weak lower crustal layer is added in these models, it has an initial viscosity 3 orders of magnitude lower than the background value. Starting conditions for several numerical experiments are shown in Figure 1. For a list of values used in the experiments, see Table 2, and for a list of all models, see Table 1.

[12] This series of numerical experiments was conducted using the axisymmetric two-dimensional finite element numerical code ConMan [*King et al.*, 1990]. The model box consists of a 100 by 100 grid of nodes. The left-hand side of the model box is an axis of symmetry. The bottom and right sides have flow-through boundary conditions, and the top is a free-slip boundary with its temperature set to the nondimensional equivalent of 20°C. The code solves non-

Table 1. Dimensionalized Characteristics of Numerical Models of Gravitational Instabilities^a

Experiment	Lith, km	Crust, km	Comp, km	η_0 , Pa s	E^* , kJ mol ⁻¹	$\Delta\rho$, %	Melt, km ³	Time to Sink, 250 km, Myr	Length of Experiment, Myr	Topography	Comments
d5050	100	50	50	10^{19}	250	5	0	0.4	0.9	stays near -8 m throughout	
d5050T	100	50	50	10^{19}	500	5	2	0.7	1.4	stays near -8 m throughout	as d5050 but higher E^*
d5050L	100	50	50	10^{19}	250	1	0	2.2	4.2	stays near -6 m throughout	as d5050 but less density contrast
d5050V	100	50	50	10^{21}	250	5	0	no fall	14.0	stays near -9 m throughout	as d5050 but $\eta_0=10^{21}$ Pa s
d2525	50	25	25	10^{19}	250	5	na	1.8	3.6	stays near -6 m throughout	as d5050 but thinner lithosphere
d2575	100	25	75	10^{19}	250	5	24	0.3	7.0	rises from -20 to -10 m by 1.3 Myr; no further by 7.0 Myr	as d5050 but thinner crust
d2575D	100	25	100	10^{19}	250	5	123	0.3	0.4	rises from -32 to -25 m	as d2575 but dense layer extends into crust
d2575W	100	25	50	10^{19}	250	5	1	0.5	1.9	rises from -10 to 0 m	as d2575 but dense layer is wide: 100 km in radius and 50 km high
dnc	100	0	50	10^{19}	250	5	0	0.4	0.9	rises from -9 to -6 m	as d5050 but no crust
dncD	100	0	100	10^{19}	250	5	200	0.3	1.0	rises from -33 to -20 m	as dnc but high-density material to surface

^aLith, lithospheric thickness, from surface to adiabatic mantle; crust, kilometers of crust within lithosphere; comp, maximum height of added composition in lithosphere, measured from adiabatic mantle upward; E^* , activation energy for viscosity law; $\Delta\rho$, density difference between added composition in lithosphere and mantle reference density; melt vol, cubic kilometers of melt produced during instability fall, assuming 0.1 wt % melt per degree above the dry mantle solidus, appropriate for fractional melting (no hydrous melting is included), at the mantle potential temperature indicated; sinking time, time required for main delaminating mass to sink below 250 km, the bottom of the model box; and topo, topographic movement at surface above center of instability (at axis of symmetry of model box).

dimensional equations for flow as given by *van Keken et al.* [1997], composed of an equation of motion,

$$\nabla \bullet (\eta \dot{\epsilon}) - \nabla P = (RaT - Ra_c \Gamma) \hat{z}, \quad (1)$$

a statement of incompressibility,

$$\nabla \bullet v = 0, \quad (2)$$

an advection-diffusion equation for temperature,

$$\frac{\partial T}{\partial t} + (v \bullet \nabla)T = \nabla^2 T, \quad (3)$$

and an advection equation for composition,

$$\frac{\partial \Gamma}{\partial t} + (v \bullet \nabla)\Gamma = 0, \quad (4)$$

depending upon viscosity (η), the deviatoric strain rate tensor ($\dot{\epsilon}$), dynamic pressure (P), thermal and compositional Rayleigh numbers (Ra and Ra_c , given below), composition (Γ), velocity (v), temperature (T), time (t), and the unit vector in the direction of gravity ($\hat{z}$).

[13] Both temperature and composition contribute to buoyancy. Thermal buoyancy is determined by the Raleigh number, containing terms for density (ρ), gravity (g), thermal expansivity (α), temperature range across the model box (ΔT), height of the model box (h), reference viscosity (η_0), and thermal diffusivity (κ):

$$Ra = \frac{\rho g \alpha \Delta T h^3}{\eta_0 \kappa} = 2 \times 10^6. \quad (5)$$

Both the negative buoyancy of the dense lower lithosphere and the positive buoyancy of the crust are measured by the

compositional Rayleigh number,

$$Ra_c = \frac{\Delta \rho g h^3}{\eta_0 \kappa} = Ra \left[\frac{\Delta \rho}{\rho \alpha \Delta T} \right], \quad (6)$$

in which $\Delta \rho$ is the difference between the densities of the lower lithospheric composition and the adjoining asthenosphere. The dense material in the lithosphere is set to either a 1 or 5% density increase over adjacent mantle lithosphere, and its maximum temperature addition is 10% above the ambient asthenospheric temperature, consistent with heats of fusion. The buoyant crust is set to a 12% density decrease when compared to reference mantle density.

[14] Viscosity is calculated in most experiments using the following Newtonian law:

$$\eta_{Newtonian} = \eta_0 \exp \left(\frac{E + Vz}{T + T_o} - \frac{E + Vz_o}{1 + T_o} \right), \quad (7)$$

where η_0 , z_o , and T_o are reference values for viscosity, depth, and temperature, respectively; E is the activation energy, and V is the reference volume (Table 2). The importance of the temperature dependence of viscosity is investigated by using values for the activation energy E equivalent to either 250 or 500 kJ mol⁻¹. The reference mantle temperature T_o is 1300°C.

[15] A starting condition for each numerical experiment was created by using a complementary error function cooling law to make a cooled lithosphere of the desired depth, employing the temperature at the surface (T_S), the temperature of the convecting mantle (T_M), thermal diffusivity (κ), and the time period of thermal diffusion (τ) [*Turcotte and Schubert, 2002*]:

$$T(z) = (T_S - T_M) \operatorname{erfc} \left[\frac{z}{2(\kappa \tau)^{0.5}} \right] + T_M \quad (8)$$

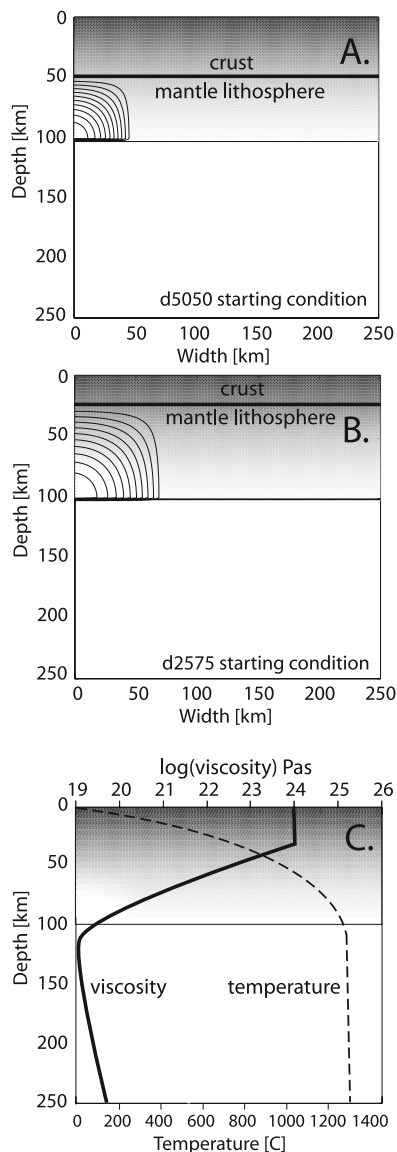


Figure 1. Example starting conditions for numerical experiments. Images are two-dimensional cross sections through the lithosphere and mantle, with an axis of symmetry at the left side. (a and b) Temperature (shades of gray and thin line), from 20°C at the surface to the mantle potential temperature (in white). Added material with density and temperature contrast is shown in contours of 10% of $\Delta\rho$ each, with the maximum addition at the axis of symmetry on the interface between the mantle and lithosphere. The maximum addition in each model corresponds to 1 or 5% density contrast with the mantle. (c) Viscosity (bold line) and temperature (dashed line).

to create a mantle lithosphere of 100, or in one case, 50 km thickness. There is no imposed initial mantle flow field.

[16] Any melt produced by dry adiabatic melting in convection currents associated with the gravitational instability is calculated with a postprocessor routine using the parameters listed in Table 2. This postprocessor routine uses the mantle flow fields to calculate the volume of astheno-

spheric material moving above its solidus during each time step of the numerical calculations. Melt is produced at a rate of 0.1% per degree rise above the solidus, which is fitted to experimental data on the continental fertile peridotite KLB-1 from *Herzberg et al.* [2000] and *Takahashi et al.* [1993]. The surface topography resulting from stresses imposed on the lithosphere by the density anomaly and its subsequent gravitational instability are calculated from differential stress values at the surface produced by the numerical calculations. The magnitude of topography calculated from deviatoric stress output from numerical models is highly dependent upon the scaling parameters used. Here I use the same scaling parameters used for the numerical experiments themselves. If other scaling rules are used, the magnitude of the topographic expression will change, but not its sense.

3. Results

[17] A gravitational instability forms through a vertical perturbation in a boundary that initiates lateral flow, allowing mass to gather at the site of the perturbation and causing the instability to enlarge. The growing instability begins to sink into the underlying mantle material as a drip, exactly analogous to but reversed in the sense of growth from a plume head initiating ascent from a deeper mantle layer. These numerical models follow the results of *Conrad* [2000]: the instability initially grows exponentially (if possessing stress-dependent viscosity, it grows superexponentially). A density contrast of only 1% is sufficient to drive a gravitational instability (see experiment d5050L in Tables 1 and 3), so the predominant controlling factor in the development of an instability is the viscosity of the dense material and of its adjacent materials. A reference viscosity of 10^{19} Pa s allow material to sink through the dimensional equivalent of 250 km within a few million years, but a reference viscosity of 10^{21} Pa s the instability just begins to grow, and is then frozen in place by conductive cooling (experiment d5050V in Tables 1 and 3).

[18] If at the time when the instability accelerates downward lateral flow cannot replenish material at the base of the instability fast enough, the material at the base of the instability separates at its edges from the laterally adjacent lithosphere. The result is an annulus of thinned lithosphere, centered on the instability (Figure 2, top left). Asthenospheric material can be drawn upward into this relatively shallow annulus by convection currents initiated by the moving instability, during which process the asthenosphere may melt adiabatically, depending upon the solidus and temperature of the asthenosphere and the height of the topography in the annulus (Figure 3). If the lithosphere is initially stiff, or if the instability falls slowly enough to allow significant conductive cooling, the lowest lithosphere will not have a viscosity low enough to allow formation of this annulus and no adiabatic melting will occur (Figure 4).

[19] In cases where melting occurs the volume of melt produced is dependent upon the composition and temperature of the asthenosphere, the range of depressurization allowed by flow created by the instability, and the time period over which such flow is induced. Using a solidus for fertile mantle peridotite KLB-1 derived from *Herzberg et al.* [2000], *Hirth and Kohlstedt* [1996], and *Takahashi et al.* [1993], no adiabatic melting in convection currents is

Table 2. Parameters Common to Numerical Models of Delaminating Arc Lithosphere

Parameter	Symbol	Value
Height and width of model box	h	250 km
Number of nodes in each dimension of the model box		100
Reference mantle density	ρ	3,300 kg m ⁻³
Crustal density		2,900 kg m ⁻³
Reference viscosity	η_0	10 ¹⁹ or 10 ²¹ Pa s
Thermal expansivity	α	3 × 10 ⁻⁵ deg ⁻¹
Thermal diffusivity	κ	1 × 10 ⁻⁶ m ² s ⁻¹
Rayleigh number	Ra	2 × 10 ⁶
Density contrast between altered lower lithosphere $\Delta\rho$ and asthenosphere		1 or 5%
Compositional Rayleigh number	Ra_c	5 × 10 ⁵ or 2.5 × 10 ⁶
Mantle potential temperature	T_p	1,300°C for dynamic models; up to 1450°C for postprocessor melt calculations
Heat capacity of silicates	C_p	1,256.1 J deg ⁻¹ kg ⁻¹
Heat of fusion of silicates	H_f	418,700 J kg ⁻¹
Fraction of melt produced per degree above solidus during fractional melting of dry lherzolite in convective currents	β or df/dT	0.1 wt % K ⁻¹

produced under a 100-km-thick lithosphere if mantle potential temperatures are below 1420°C (see Table 1).

[20] Topographic uplift caused by the replacement of a dense lithospheric region with hot buoyant asthenosphere is commonly cited evidence for lithospheric delamination. A profound uplift, however, is not seen in these experiments (Figure 5). The use of spherical axisymmetry in these experiments is both more physically realistic, and suppresses topography in comparison to Cartesian, and even cylindrical coordinates. The buoyant crust significantly limits the topography otherwise created by the sinking and removal of dense lithospheric material. Without buoyant crust and in the presence of larger concentrations of dense material the topographic deviations from a zero datum can extend to hundreds of meters. The subtle topographies that result from these numerical experiments are given in Table 1 and shown in Figure 5.

[21] The original model surface, prior to the injection of the density anomaly in the lithosphere, is flat. The addition of the density anomaly (as described above in this section) pulls topography at the axis of symmetry down, and creates

a compensatory upward flexure at distance from the axis of symmetry. This initial basin is up to a few tens of meters deep at its center and rises to the zero datum of unaltered lithosphere over a lateral distance of about 80 to 120 km in these numerical experiments. In the topographic profiles shown in Figure 5 the original landscape, prior to density injection, would lie at the zero datum. Topography at the axis of symmetry can be further depressed by the initial growth and focusing of the instability, through viscous traction on the lithosphere as mass moves laterally to accumulate at the site of the perturbation. As the dense material sinks beneath the axis of symmetry, the neck of material attaching the drip to the lithosphere stretches and thins, gradually eliminating viscous traction between the drip and the lithosphere and allowing the lithosphere to rebound upward (Figures 5a, 5b, and 5c).

[22] In these experiments with thick, stiff, buoyant crusts, the topography subsides very little in response to the increased density, and it rebounds very little if at all with the sinking away of the dense material (experiments d5050 and d5050T, L, and W in Tables 1 and 3). The remainder of

Table 3. Effects of Changing Parameters on Development of the Gravitational Instabilities in Numerical Experiments

Parameter Change	Example	Effect on Instability Size	Effect on Sinking Time	Effect on Topography
Greater temperature dependence of viscosity	d5050T versus d5050	much smaller; leaves far more dense material frozen into the lithosphere	slows sinking rate	in thin lithospheres, lessens topographic expression
Greater density contrast	d5050 versus d5050L	little effect	slows sinking rate significantly	increases topographic expression
Greater initial viscosity	d5050V versus d5050	can eliminate instability	slows sinking rate significantly	lessens topographic expression
Thicker lithosphere	d5050 versus d2525	instabilities scale with layer thickness, except where upper portion of layer is too stiff to flow	larger instabilities sink more quickly	thinner lithospheres can allow greater flexure
Thinner crust	d2575 versus d5050 and d2575D versus dncD	instabilities scale with layer thickness, except where upper portion of layer is too stiff to flow	larger instabilities sink more quickly	increases topographic expression
Dense region much wider than its thickness	d2575W versus d2575	more dense material is removed	larger instabilities sink more quickly	lessens topographic expression

Experiment d2575

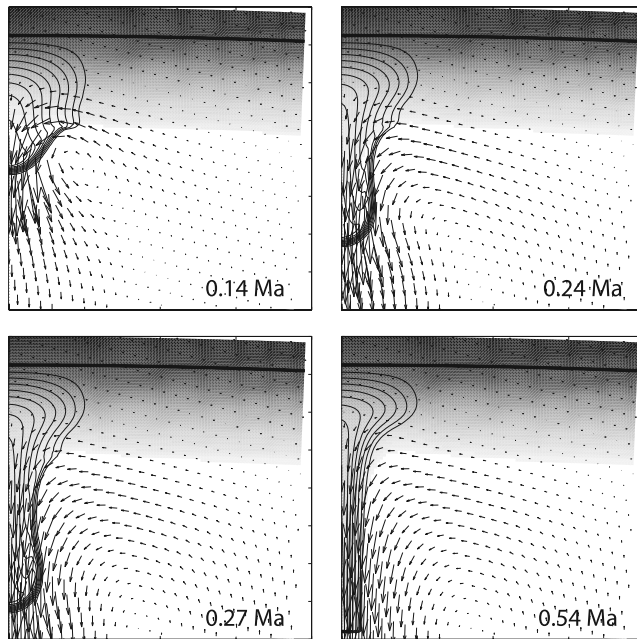


Figure 2. Four vector fields from numerical experiment d2575. The images show cross sections from the top of the crust into the mantle. The left boundary is an axis of symmetry, passing through the center of the gravitational instability. The scaled model box represents 250 km in width and height. Arrows represent the velocity vector field and shading represents temperature. Added material with density and temperature contrast is shown in contours of 10%, with the maximum addition at the axis of symmetry on the interface between the mantle and lithosphere. Material above the bold solid line is buoyant crust. A significant portion of the dense material remains frozen into the lithosphere after the sinking material detaches. Some upwelling is seen in an annulus around the growing instability, particularly in the first panel, and a small amount of adiabatic melt can be generated.

the experiments had thinner or nonexistent crusts, and their initial depression and subsequent rebound were larger. Both a thick lithosphere and conductive cooling, however, can retard or prevent uplift. None of these experiments rebounded to or above their original zero topographic datum, but topography is lifted back to the zero datum level by the addition of a warm upwelling in the mantle of as little as 1400°C.

[23] Though *Neil and Houseman* [1999] found that a buoyant crustal layer could thicken over the site of a gravitational instability such that it caused uplift, the temperature dependence in these experiments produced sufficiently high viscosity in the crust that no such effect occurred. The existence of a crust, however, limits both the range of topographic expression and the volume of dry adiabatic melting.

[24] Over the parameter ranges explored in these experiments, topography varied by a maximum of ~40 m from the zero datum that existed before the density anomaly was

introduced into the lithosphere. A few of the experiments produced mantle melt in convection currents in the lithosphere initiated by the sinking instability, and of those, the maximum melt volume was in the hundreds of cubic kilometers. For a reasonable viscosity structure in a 100 km lithosphere, intruded with the equivalent of several tens of percent of eclogite, therefore the topographic and dry melt signature of lower lithospheric gravitational instability may be relatively minor when viewed from the surface.

[25] These numerical experiments indicate that the topographic expression of a gravitational instability is increased by low-viscosity, higher-density contrast, a thinner lithosphere, and a thinner crust (Figures 5a and 5c), but increased most of all by the existence of a low-viscosity layer in the lithosphere. This layer may be envisioned as a ductile lowest crust above a compositionally stiffer mantle lithosphere. This low-viscosity layer allows the crust and mantle lithosphere to slip past each other and therefore to display the maximum surface topography; existence of such a layer increases topographic expression by as much as an order of magnitude. The use of Cartesian geometry, which the dense material in these images represents the cross section of a long rod rather than an axially symmetric half spheroid, extends the topographic expression by as much as 2 orders of magnitude over spherical geometry without a low-viscosity layer. For maximum topographic expression, therefore the lithosphere should be hot and thin and contain a low-viscosity layer, and the dense material should have a high aspect ratio in the plane of the lithosphere-asthenosphere boundary. Nominally dry melting of asthenosphere in adia-

Experiment d2575D

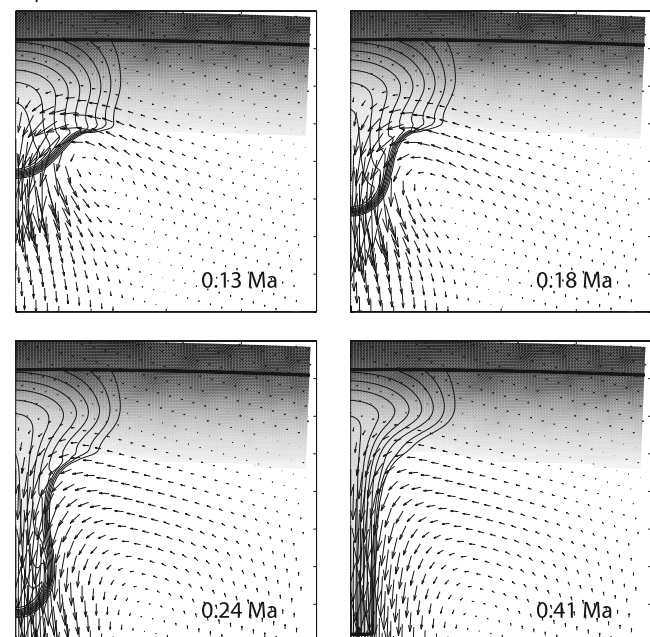


Figure 3. Four vector fields from numerical experiment d2575D. Symbols are as for Figure 2. Experiment d2575D had identical starting conditions to d2575 (Figure 2) except d2575D has dense material extending to crustal levels. Both melting and topographic expression are maximized; see Tables 2 and 3.

Experiment d2575W

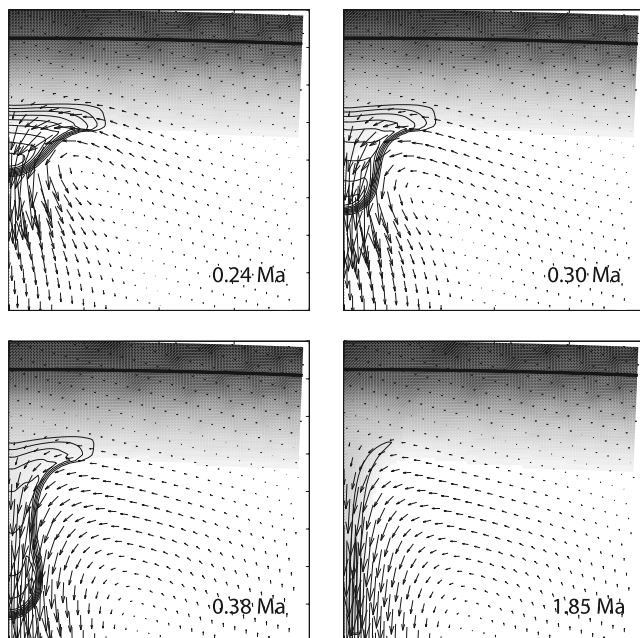


Figure 4. Four vector fields from numerical experiment d2575W. Symbols are as for Figure 2. Experiment d2575W had identical starting conditions to d2575 (Figure 2) except its dense region was twice as wide and half as thick. Virtually all of the dense material is removed from the lithosphere by the growing instability, but the topographic expression is minimized and virtually no dry melting occurs: Note the absence of upwelling around the growing instability as compared to d2575 in Figure 2.

batic convection currents is increased by a greater topographic rebound and by a larger volume of material sinking faster. These results are summarized in Table 3.

4. Discussion

[26] The suite of numerical experiments presented here demonstrate that it is possible for gravitational instabilities tens of kilometers in diameter sinking from under 100-km-thick lithosphere to create only modest surface signatures, including topographic deflections on the order of meters. The most extreme initial conditions in these experiments produce at most 200 km³ of magma, and topographic subsidence in the tens of meters. More extreme conditions, producing more extreme and more easily detectable topographic and magmatic signatures, can be hypothesized for special situations: for example, lithosphere over a hot upwelling will be softer and more topographically responsive, and any magmatism may be multiplied into the thousands of cubic kilometers by the extra heat of the mantle [see, e.g., *Elkins-Tanton*, 2005]. Even in the conservative models presented here, however, an additional and perhaps most important effect can be demonstrated.

4.1. Production of Hydrous Continental Magmas and a Heterogeneous Mantle

[27] If the lower lithosphere is volatile-rich, it may devolatilize as it sinks just as descending lithosphere in a

subduction zone devolatilizes as temperature and pressure increase. Mantle xenoliths containing hydrous minerals have been found in a variety of localities [e.g., *Lee and Rudnick*, 1999; *Buhlmann et al.*, 2000; *Powell et al.*, 2004], so assuming the possibility that volatiles may be found in the mantle lithosphere, the depths and temperatures of devolatilization can be calculated given phase stability diagrams of hydrated and carbonated mantle materials.

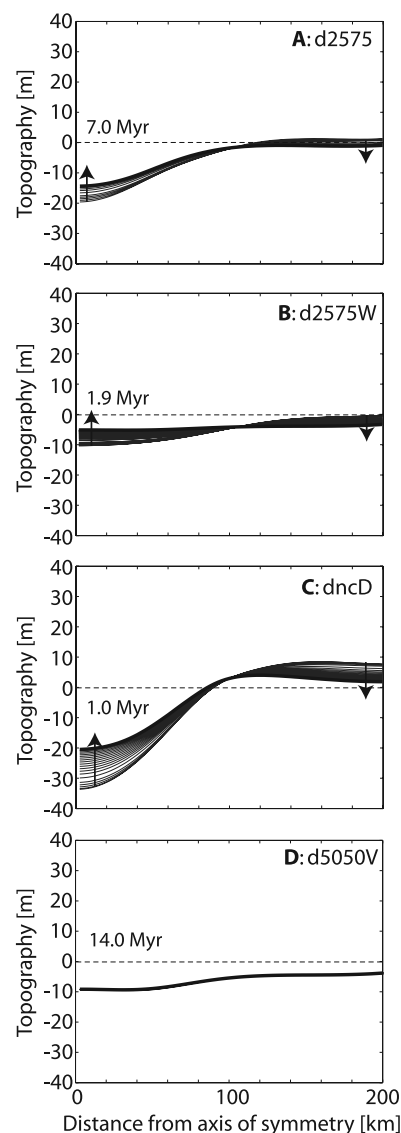


Figure 5. Surface topographic evolution of the gravitational instabilities from models d2575 (flow fields shown in Figure 2), d2575W (flow fields shown in Figure 4), dncD, and d5050V. Topography is given in meters, and distance from the axis of symmetry is given in kilometers. The first topographic profile shown indicates the topography with the density anomaly in place but before significant growth of the instability has occurred. The land surface prior to density injection would be like at the dashed zero datum. Arrows show the sense of time evolution of the topography (but not its magnitude); bold line is the final topographic expression at the end of the numerical experiment, which is given as a scaled time. Lines are not equally spaced in time.

[28] The initial temperature, conductive temperature increase, and pressure change of the material at the center of each drip can be calculated from numerical experiment results, and are given in Figure 6. Material that sinks relatively slowly heats conductively to higher temperatures while at lower pressures; material that sinks quickly retain its initial temperature to greater depths. These two generalized paths result in different predictions: the drips that sink slowly cross solidii in pressure-temperature space as they heat, and can move into regions where damp peridotite will melt while dry peridotite will not (Figure 6b). The drips that sink quickly can descend into a region where free hydrous fluid exists. These latter drips therefore may release volatiles into the surrounding mantle without necessarily melting themselves (Figure 6c).

[29] To produce melt in the sinking material or in the surrounding asthenosphere, the source must move above its solidus. At a given pressure the solidus temperature depends upon the composition of the source. The higher the volatile content or the more fertile the major element composition, the lower the solidus temperature. Thus where melting will occur and to what extent it will proceed depend entirely upon the major element and volatile composition of the material melting. These models predict heating in slowly sinking material sufficient that some material would exceed the solidus of fertile peridotite bearing ~80 ppm water [Hirth and Kohlstedt, 1996, 2003] (Figure 6). This water content is suggested by Hirth and Kohlstedt [2003] to be typical of asthenosphere that has been processed through a subduction zone, and therefore may be considered a reasonable and even common value.

[30] The fate of volatile-rich fluids released from sinking material is likely to be complex, as shown in Figure 7. Fluids may move percolatively upward into a region of higher temperature where they can induce melting in the asthenosphere, or they may percolate into a region in which volatile-rich phases are stable, creating a volatile-rich region in the asthenosphere. Depending upon its composition and potential temperature, the asthenosphere may then be triggered to melt, just as portions of the mantle wedge in a subduction zone are thought to when reached by hydrous fluids fluxed from the descending slab. Fluid percolation is likely controlled by pathways of permeability in the solid matrix, which may or may not be altered by pressures produced by flow fields. Pressure gradients inside the neck of the sinking instability may be great enough in some cases to guide the volatiles upward toward the lithosphere rather than allowing them to rise straight into the asthenosphere, but in other cases low viscosity may reduce the pressure gradients to the point of ineffectuality.

[31] Any of these melting scenarios have the potential to create primitive hydrous basaltic magmas with high alkali contents and lithospheric trace element signatures, typical of small-volume continental magmas worldwide (see references for Figure 6 and Elkins-Tanton [2005, and references therein]) and consistent with the potassic, hydrous magmas previously attributed to delamination [e.g., Kay and Kay, 1993]. The pressures and temperatures of origin for a number of these hydrous, potassic magmas are shown in Figure 6c, and they lie within the pressure and temperature parameter ranges predicted for melting from volatile-bearing sinking material. These magmas are compositionally dis-

tinct from the relatively dry adiabatic melts that result at mid-ocean ridges, and which the dry adiabatic melting created by the movement of the sinking instability (discussed above) would more closely resemble. Volumes of volatile-driven melting are not calculated in these experiments.

[32] To investigate the fate of melt or volatile-rich fluids in the sinking material, a comparison of the speed of descent of the material with the speed of percolative ascent of the fluid is required: Can the fluid rise percolatively fast enough to escape the sinking material? First, in the literature the rate of fall of gravitationally unstable material is often calculated according to the rule of Stokes flow [Turcotte and Schubert, 2002]. Comparing predicted Stokes flow rates for a comparable spherical volume with the rates of sinking shown in numerical experiments (Figure 8) shows that for Newtonian viscosity laws the numerical experiments have far slower rates of fall, while for non-Newtonian (stress-dependent) laws the instabilities fall faster than predicted by Stokes flow. In the Newtonian case, the viscous neck of the instability retards its fall. In the case of non-Newtonian viscosity, the extensive shearing produced by instability growth and fall reduce viscosities, and therefore strength, sufficiently to allow rapid sinking. Non-Newtonian flow laws are probably most applicable to the asthenosphere [Karato and Wu, 1993; Hirth and Kohlstedt, 1996], and so faster rates of sinking must be considered a physical possibility.

[33] As the material sinks and devolatilizes, the volatiles will presumably rise upward buoyantly through grain boundary percolative flow, the rates of which can be calculated using Darcy flow rules [Turcotte and Schubert, 2002] (Figure 8). Permeability k of the solid material is expressed as

$$k = \frac{b^2 \Phi^2}{72\pi},$$

where b is grain size in meters and Φ is dimensionless porosity [Turcotte and Schubert, 2002]. When compared to numerical experiments with Newtonian rheologies, all fluids escape through buoyant flow into the asthenosphere. The magma density used in these calculations is 500 kg m^{-3} lower than the assumed surrounding solid (2800 kg m^{-3} versus 3300 kg m^{-3}). For each 100 kg m^{-3} added to the density of the magma, its velocity slows by a factor of 0.8. When compared to experiments with non-Newtonian experiments, fluids with low viscosity or those in material with high porosity will escape into the asthenosphere, but fluids with high viscosities or resident in material with low porosity will be carried to depth in the sinking material.

[34] Volatiles therefore can have two melting effects (they can be released from the sinking material and cause the asthenosphere melt, or they can allow parts of the delaminating material itself to melt) as well as two possible fates that do not involve melting (they can be retained by the falling material, or they can rise to create stable volatile-rich phases in the asthenosphere or mantle lithosphere). This provides a number of geochemically different source regions that can produce melt eventually erupted onto the surface: The volatile-rich delaminating material itself; volatile-enriched mantle material in the wake of the sinking material; and dry mantle material upwelling beneath the

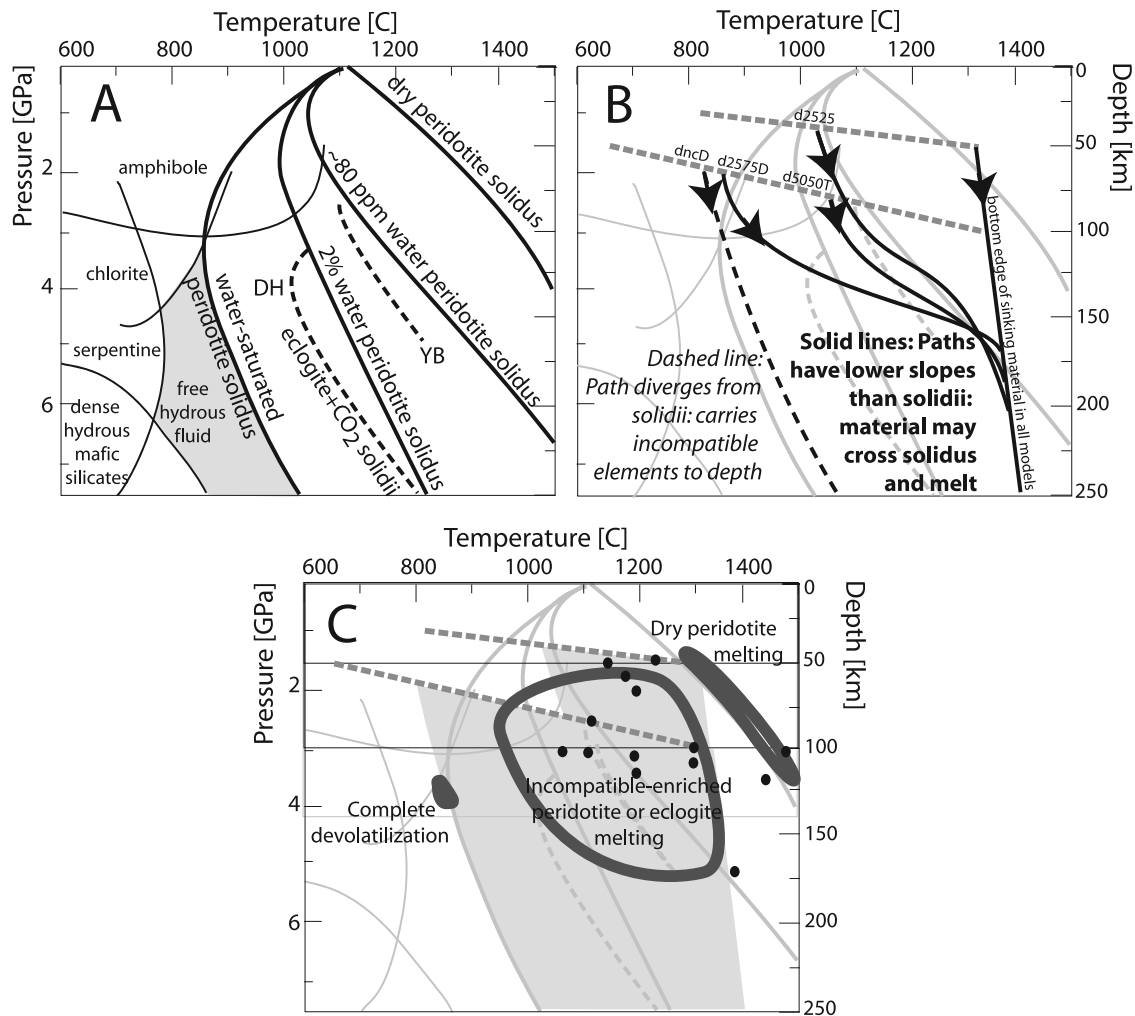


Figure 6. (a) Phase boundaries (solid lines) for peridotite with varying water content. Solidii are shown for a range of water contents from zero to saturation. Below the solidus stability limits for water-bearing minerals are shown, as is a shaded region in which no water-bearing minerals are stable, and so water will be released as a free fluid. Dashed lines show solidii for eclogite with two concentrations of CO₂ from Yaxley and Brey [2004] (YB) with 15 wt % CO₂ and Dasgupta et al. [2004] (DH) with only 5 wt % CO₂. (b) Example paths of material in sinking instabilities, showing paths in bold that heat sufficiently quickly that, depending upon their compositions, may cross their solidii and produce melt. The dark dashed line shows the path of a quickly sinking instability that will not cross solidii but may either devolatilize at depth and will contribute to a heterogeneous mantle. Subhorizontal dashed lines show the starting conditions of material in the dense regions for 100-km- and 50-km-thick lithospheres. (c) Regions covered by paths of material in instabilities from numerical experiments (shaded). Regions of dry adiabatic melt in the annulus around the sinking instability, incompatible-enriched melting of the sinking material or its neighboring asthenosphere, and complete devolatilization, are shown with bold outlines. The pressures and temperature of origin of hydrous, alkali-rich continental magmas from a variety of regions are shown as black dots, and fit well with the predicted melting regions from the numerical models. Phase boundaries from Kawamoto [2004], Herzberg et al. [2000], Takahashi et al. [1993], Trønnes and Frost [2002], Schmidt and Poli [1998]; dry peridotite solidus from Hirschmann [2000]; 2% water solidus from Litasov and Ohtani [2003] and Ohtani et al. [2004]; water-saturated solidus from Grove et al. [2006] and Mysen and Boettcher [1975]; carbonated eclogite solidii from Yaxley and Brey [2004] and Dasgupta et al. [2004]; multiple saturation points of lavas from Barton and Hamilton [1978, 1979], Edgar and Condiliffe [1978], Edgar et al. [1976, 1980], Elkins-Tanton and Grove [2003], Elkins-Tanton et al. [2006], Esperanca and Holloway [1987], Nichols and Whitford [1983], Richter and Carmichael [1996], and Sato [1997].

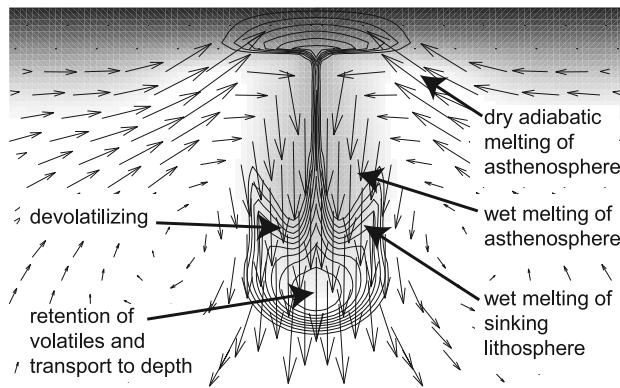


Figure 7. Schematic, exaggerated model for devolatilization and magma production during delamination by a gravitational instability. As the instability sinks, normal asthenosphere may flow into the resulting annulus of thinned lithosphere, and may melt adiabatically. If the instability is volatile-enriched it may devolatilize as it sinks and heats, triggering melting of the mantle or of itself. Fast sinking instabilities may carry volatiles to depth.

thinned lithosphere (Figure 7). Volatiles may also be carried to depth in rapidly sinking material, producing an important method for creating a heterogeneous, volatile-rich mantle, second in importance only to subduction zones, and most critical on one-plate planets such as Venus.

[35] The suggestion here of devolatilization and resulting melt, or of melt of the sinking material itself, is dependent upon the unstable material having a different composition to the rest of the lithospheric mantle, and in particular, to the underlying asthenosphere. The dense, unstable material is hypothesized to be mantle lithosphere in combination with eclogitized material produced by freezing melts into the cooler lithosphere. This melt injection would occur incrementally and freeze rapidly, and so would create very little positive buoyant force before freezing into its denser final form. *Hoogenboom and Houseman* [2006] note, however, than in the case of a buoyant density anomaly in a lithosphere, a central upwelling model of instability can form, creating a different pattern of lithospheric thinning and topographic response. This process is less likely to occur in the scenario hypothesized here than in a case with a more significant and long-lived positive anomaly, for example, a magma chamber.

4.2. Extreme End-Members: Nonmagmatic Basins and Flood Basalts

[36] In many of these experiments, no dry adiabatic melting occurs (or so little that no surface expression of the melting would be expected), and a topographic depression persists at the end of numerical modeling. If the volatile content of the unstable material is low enough to prevent any additional melt production, these conditions are candidates to form intracratonic basins. Basins have been proposed to form in response to a dense, eclogitized zone [e.g., *Ahern and Mrkvicka*, 1984; *Naimark and Ismail-Zadeh*, 1995; *Yamasaki and Nakada*, 1997]. Here I additionally propose that the dense region is likely to sink through density instability, but that the basin may not rebound even after the dense material is lost.

[37] At the other end of the magmatic spectrum, some flood basalt provinces are produced on stable cratons, for example, the Deccan and the Siberian large igneous provinces. These provinces are almost certainly created by hot mantle upwellings. The initial magma injections into the lithosphere from the thermal upwelling, along with heating from conduction and from heats of fusion from melts stalled in the lithosphere, create an ideal environment for instabilities from the lithosphere, lowering its viscosity and adding density, the driving force for instability [*Elkins-Tanton and Hager*, 2000]. The gravitational instabilities will thin the lithosphere and enhance melt production. While the instabilities themselves are insufficient to create the volume of melt of a flood basalt province, their action may add to the volume of melt and may also produce unusual compositions, perhaps including the meimechites of Siberia [*Elkins-Tanton et al.*, 2006].

4.3. Detection of Gravitational Instabilities and Further Work

[38] When the material in the instability has sunk away from the lithospheric boundary and its connective neck stretched and thinned to insubstantiality, almost no topography remains at the bottom boundary of the lithosphere.

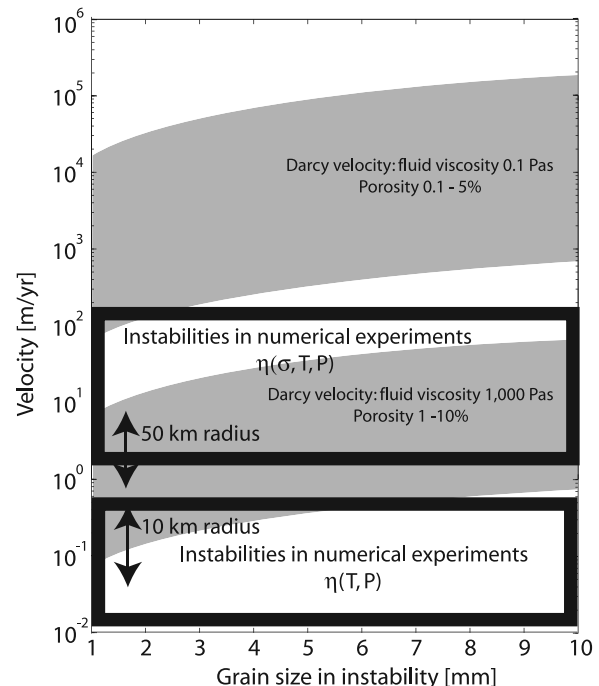


Figure 8. Comparisons of sinking velocities of solid lithospheric gravitational instabilities and rising velocities of buoyantly percolating fluids. Simple calculations of Stokes flow velocities for sinking instabilities are shown in arrows on the left. Numerical experiments with Newtonian rheologies (bottom black box) sink more slowly than Stokes predictions, while those with non-Newtonian rheologies (top black box) sink more quickly. Fluid velocities calculated using Darcy's law show that low-viscosity fluids will percolate out of falling instabilities and enter the asthenosphere, while fluids in low-porosity instabilities or those with higher viscosities will be carried to depth with the sinking instability.

Lateral flow to feed the growing instability has kept the bottom boundary of the lithosphere near horizontality throughout the process, the existence of a shallow annulus notwithstanding.

[39] Conductive cooling quickly erases any remaining perturbations and returns the bottom of the thermal lithosphere to horizontality (approached in Figures 2, 3, and 4). This implies that “scars” from gravitational instabilities at the bottom boundary of the lithosphere will not remain to be detected geophysically. Within a few million years at most any small perturbations to the bottom of the lithosphere left in their wake will be erased, and so looking seismically at the base of lithospheres for characteristic topography will not result in finding any hallmarks of gravitational instabilities.

[40] The instabilities themselves or their much longer-lived tails have been detected seismically [e.g., Jones and Phinney, 1998], and further searching in regions of volcanism on continents is the most promising way to detect active dripping from the lithosphere. Further petrologic study of magmas and xenoliths may also serve to test this hypothesis: for example, Elkins-Tanton and Grove [2003] found that a primitive hydrous magma from the Sierra Nevada originated from a source created by metasomatizing asthenospheric peridotite with a hydrous, alkali-rich fluid, consistent with the models presented here.

5. Conclusion

[41] Lithospheric gravitational instabilities may be a significant means of recycling both crustal and lithospheric materials and volatiles back into the mantle. Numerical experiments show that instabilities are likely to form in continental lithosphere and that extreme conditions are not required to allow them to grow and fall. The topographic expression of these instabilities begins with a depression, but that depression may grow or may diminish, even to the point of regaining a zero datum of unperturbed lithosphere, depending upon lithospheric rheology. However, in these conservative models, with 100-km-thick lithosphere and a stiff, buoyant crust, surface topographic changes are minimal and may be difficult to detect, particularly when trying to reconstitute ancient topographic movements.

[42] Uplift following the initial topographic subsidence is not predicted by these models with stiff buoyant crust. Uplift has been demonstrated with a hot upwelling under the site of the instability [Elkins-Tanton and Hager, 2000] and also can be the result of crustal thickening in the case of a more ductile crust [Neil and Houseman, 1999]. These processes may be differentiable through measurements of heat flow.

[43] As they sink the instabilities may devolatilize (as a descending slab in a subduction zone does), they may themselves melt, or they may carry volatiles to depth, depending upon their rate of fall. Volatiles released into the mantle may be stabilized in solid phases or may trigger melting, even in cases where no dry adiabatic melting is triggered by convection around the sinking instability. Kimberlites or other rapidly decompressing volatile-rich magmas from depth may be created through this process. In the upper mantle, where descending slabs are forced to move at plate rates and therefore devolatilize more com-

pletely, instabilities are able in some cases able to sink sufficiently fast that their volatiles are carried to depth and enrich the mantle. Volatiles carried to depth in this manner may be a primary means of creating mantle heterogeneity and of volatile recycling.

[44] Magmatism as a result of gravitational instabilities is possible but only when compositions, temperatures, and pressures coincide fortuitously. Gravitational instabilities themselves are a likely accompaniment to magmatism created by subduction zones or by thermal upwellings (for example, deep plumes). While lithospheric erosion from viscous traction of plume material is an inefficient process and unlikely to thin lithosphere to any significant extent, the gravitational instabilities that may be triggered by small amounts of decompression melting beneath a thick lithosphere can thin the lithosphere sufficiently to increase the adiabatic melting column in the upwelling.

[45] Ancient continental roots may be retained in part by preventing this process: if the continental root has grown through accretion of buoyant, depleted mantle material to a depth under which no partial melting through adiabatic decompression can occur, then a driving force for gravitational instabilities is removed, and the lithosphere may continue accreting without significant thinning from instabilities. Portions of eclogitic material emplaced into the lithosphere through earlier melting events, however, may remain frozen into the lithosphere indefinitely; these experiments indicate that some dense material is almost invariably left frozen in place as the mobile portions of the unstable material sinks away.

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