Marginal stability of thick continental lithosphere

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Laboratory and numerical experiments show that when a chemically different layer overlies a hotter but otherwise denser layer, analogous to continental mantle lithosphere over asthenosphere, convective stability depends strongly on both the critical Rayleigh number and the buoyancy number, $B$, of the lithosphere-like layer. Sufficient cooling at low buoyancy number results in an oscillatory convective instability whereby the colder, more viscous, but chemically lighter layer is drawn into zones of downwelling flow adjacent to laterally extensive zones of upwelling. The critical Rayleigh number for instability increases with the buoyancy number from as little as $\approx 30$ for $B = 0$ to $\approx 1000$ for $B \approx 0.5$. Applied to continental lithosphere in a thermal and mechanical state near the instability threshold, this relationship implies that the lithospheric thickness must decrease as the mean density of the lithospheric mantle increases, consistent with the geological record.


1. Introduction

Low heat flux [Jaupart and Mareschal, 1999; Nyblade and Pollack, 1993], mantle xenolith geobarometry [Boyd, 1989; Rudnick et al., 1998], and seismic wave speeds [Grand, 1987; Jordan, 1981] all imply that Archean and Proterozoic lithosphere is thicker and colder than Phanerozoic lithosphere. To compensate for the low temperature, and resulting thermal contraction, such material must be chemically less dense than mantle material elsewhere at the same depths [Jordan, 1988], an inference corroborated by studies of lithospheric mantle xenoliths [e.g., Boyd and Gurney, 1986; Boyd, 1989; Poudjom Djomani et al., 2001; Griffin et al., 1999; Walter, 1999]. Xenoliths also reveal that both the characteristic thickness and mean lithospheric mantle density of stable cratons vary progressively with the age at which that lithosphere formed (see summary by O’Reilly et al. [2001]). Lithosphere created in the Archean is significantly thicker and chemically more buoyant than lithosphere created in the Proterozoic. Proterozoic lithosphere in turn is thicker and more buoyant that Phanerozoic-aged lithosphere. These differences pose the question of what mechanism controls lithosphere thickness.

2. Laboratory Experiments

Experiments began with two isothermal miscible fluids, stratified by their intrinsic chemical density, in a
level Plexiglas tank (Figure 1). To initiate an experiment, we imposed a temperature gradient on the more viscous layer, which we monitored every 60 seconds on each of 20 thermocouples located throughout the tank.

[6] We discuss results in subsequent paragraphs as if a viscous, low-density layer was cooled from above (Figure 1d). In reality, however, we heated a chemically more dense, lower layer from below. Double spiraled copper tubing carried a basal water bath heat source that remained within 0.1°C of the set point throughout an experiment. To make the two miscible aqueous fluid layers differ in density and viscosity, we added various amounts of NH4Cl and Natrosol brand cellulose. For the small strain rates of these experiments, the fluids behave as Newtonian fluids [Tait and Jaupart, 1989], with viscosity varying by less than a factor of 2 over the temperature range considered. Isothermal density changes were prevented because the time scale of salt diffusion was much greater than the duration of the experiments.

[7] Cooling of the lithosphere-like layer propagates into the asthenosphere-like layer and eventually induces small-scale convection in that layer. In some cases, the lithosphere-like layer remained stable with an almost constant temperature gradient across the layer (Figure 1a). In other cases, the system evolved into an unstable state evidenced by deformation of the interface between the chemically distinct layers (Figures 1b and 1c).

[8] Five dimensionless numbers characterize the flow. In all cases, however, the asthenosphere-like layer was sufficiently thick such that the ratio of the thicknesses of the layers has no impact (a result confirmed by the analysis of marginal stability), and the Prandtl number was sufficiently high (≥10) that the experiments could be studied in the infinite Prandtl number limit. The remaining three key numbers are: the viscosity ratio between the two layers, \( \gamma = \mu_l/\mu_a \), using the subscripts for the lithosphere-like and asthenosphere-like layers respectively; the Rayleigh number defined by the properties of the more viscous lithosphere-like layer (\( Ra_c = \frac{\gamma \Delta T h_l}{\nu_a \mu a} \) where \( \Delta T \) is the temperature difference across the pair of layers); and the buoyancy number \( B = \frac{\Delta \rho c}{\rho_a \Delta T} \) where \( \Delta \rho c \) is the intrinsic density difference between the two layers due to composition.

[9] Because of the large temperature difference between lithosphere and asthenosphere in the Earth, \( \gamma \) takes very large values; but for \( \gamma > 10 \), the viscosity ratio has little influence on the regime diagram. Thus, in practice, only \( Ra \) and \( B \) distinguish stable and unstable regimes.

[10] The critical Rayleigh number, \( Ra_c \), which separates the stable regime of no flow within the lithosphere-like layer from the unstable regime in which the interface between the layers deforms, depends strongly on \( B \) (Figures 2a and 2b). For \( B < 0.5 \), \( Ra_c \) increases with increasing \( B \). For larger values of \( B \), \( Ra_c \) is approximately constant and equal to 1100.

[11] The unstable regime took two forms (Figures 1b, 1c, and 2b). At relatively large \( B \) (≈0.5), and \( Ra > Ra_c \), the interface formed small amplitude domes and basins due to internal buoyancy differences associated with upwellings and downwellings, but the deformation was virtually time-independent. In contrast, when \( B \approx 0.5 \), domes and basins on the interface rose and fell, oscillating in time and with greater amplitude. For some cases, however, it was difficult to distinguish between these two forms of instability. Hence, we have categorized all experiments with deformation at the interface as unstable to avoid ambiguity. For both unstable regimes, the characteristic spacing between basins and domes was approximately proportional to the thickness of the upper layer.

3. Analysis of Marginal Stability

[12] We analyzed numerically marginal stability of a similarly layered system. Following [Currie, 1967], we approximate this using a constant temperature gradient across the upper layer and a negligible gradient across the lower layer (Figure 1d). This simple structure omits a thin transition zone at the interface between the layers (Figure 1d), where small-scale convection could develop in the asthenosphere-like layer.
Figure 2. (a) Experimental Results: The critical Rayleigh number, $Ra_c$, for the upper layer is plotted vs. $B(=\Delta \rho c/\Delta \rho g)$. Diamonds = unstable regime; Open Squares = stable regime. The hand drawn dashed line delineates the boundary between the two regimes. (b) Stability Analysis: $Ra_c$ versus $B$ for $\gamma = 1, 10$, and 100. Note that for $\gamma > 10$, the precise value of $\gamma$ has little effect on the values of $Ra_c$ as a function of $B$. (c) The relationship between total thickness, $h$, of the sub-continental lithosphere and corresponding average xenolith densities [Poudjom Djomani et al., 2001].

[13] Like [LeBars and Davaille, 2002], who considered marginal stability of a two-layered system with a constant temperature gradient across both layers, we found two unstable regimes in addition to a stable one at low $Ra$ (Figure 2b). For $0 < B < 0.45$, the growth rate of the instability includes an imaginary part, corresponding to oscillatory convection [Richter and Johnson, 1974]. As with the laboratory experiments, $Ra_c$ depends strongly on $B$, increasing from $Ra_c = 28–32$ at $B = 0$ to $Ra_c = 1017$ at $B \approx 0.45$. We reproduced Currie et al.’s [1967] result that $Ra_c = 32$ for $B = 0$ and $\gamma = 1$.

[14] For $B > 0.45$, we traced the oscillatory convection branch to Rayleigh numbers higher than $\approx1100$, but it is of no practical interest because the upper layer becomes unstable in a layered regime at $Ra_c = 1017$ (where $Ra_c$ depends weakly on $\gamma$). Finite-amplitude flow should render the oscillatory branch, which is defined only at marginal stability, irrelevant. For $B > 0.45$ and $Ra$ somewhat larger than $Ra_c = 1017$, convection occurs within the upper layer in a form similar to Rayleigh-Bénard convection within a layer of constant temperature gradient. Not surprisingly, $Ra_c$ is close to 1100.65 obtained by Pellet and Southwell [1940] for Rayleigh-Bénard convection with one rigid boundary and the other with free slip.

[15] Original arguments on lithosphere stability relied exclusively on Rayleigh-Bénard convection [e.g., Jordan, 1988]. In the oscillatory regime at low buoyancy numbers, however, $Ra_c$ can be much smaller than that for Rayleigh-Bénard convection, as small as 28 depending on the buoyancy ratio $B$ and $\gamma$. Thus, the lithosphere should be much more unstable than previously thought. Note that in the layered regime (analogous to Rayleigh-Bénard convection), perturbations to the lithosphere are weak.

[16] The critical values of $B$ for the oscillatory regime are slightly smaller than those determined in the experiments (by about 20%; $B = 0.4$ instead of $B = 0.5$). We attribute this difference to small-scale convection at the interface between the fluids in the laboratory, which we excluded in the numerical analysis.

[17] Both the numerical analysis and the laboratory experiments ignore the strong dependence of viscosity on temperature in the earth, but Conrad’s scaling using “available buoyancy” can correct for this effect. Conrad and Molnar [1999] showed that the temperature dependence of viscosity can be approximately scaled using the dimensionless integral of the ratio of temperature anomaly to viscosity over the thickness of the layer, which they called “available buoyancy.” With that scaling they reported critical Rayleigh numbers of about 50 for $B = 0$ and a range of values of $\gamma$. The large uncertainty in their estimated value makes it indistinguishable from the values of 28–32 that we found for different values of $\gamma$.

4. Discussion and Conclusions

[18] Stability must be assessed not only as a function of buoyancy, but also as a function of Rayleigh number. Insofar as continental lithosphere is in a state close to the threshold of instability, more depleted, less dense lithosphere, corresponding to large $B$, requires a greater local Rayleigh number to become unstable.

[19] Lithospheric mantle probably formed as a residue of peridotite partial melting [e.g., Boyd and Gurney, 1986; Boyd, 1989; Herzberg, 1999; Jordan, 1981, 1988]. The evolution toward smaller depletion and smaller density contrasts from Archean to Proterozoic and then to Phanerozoic time (Figure 2c) may be due to secular cooling of the earth. The temperature difference across the lithosphere has changed by only $\approx200^\circ C$ since Archean time (20% change). Thus, the major factors contributing to differences in Rayleigh number are the thickness of the lithosphere, especially because $Ra$ varies with the cube of thickness, and lithosphere viscosity. A warmer Archean mantle would serve to lower the viscosity of the Archean asthenosphere, but lithospheric instability depends on the viscosity of the lithosphere, not that of the underlying fluid (Figure 2b). Present-day differences of lithosphere viscosity among continental regions of different ages cannot be reliably determined and we assume that they are small. The relationship between $Ra_c$ and $B$ then implies that chemically more depleted Archean lithosphere should be thicker than the younger less depleted Proterozoic and Phanerozoic lithosphere, as the geological record shows [e.g., Boyd and Gurney, 1986; Poudjom Djomani et al., 2001; Griffin et al., 1999].

[20] Direct determination of $Ra$ requires values for lithospheric viscosity, which are ill-constrained. Using $\Delta T \approx 1000 K$ and data from [Poudjom Djomani et al., 2001], $B$
for Phanerozoic, Proterozoic and Archean lithosphere is about 0.20, 0.40 and 0.50 respectively, corresponding to $Ra_c$ of 150, 500 and 1100 (Figure 2). Thus, experimentally-predicted $Ra_c$ for Proterozoic and Archean lithosphere is larger than that of Phanerozoic lithosphere by factors of about 3.3 and 7.3 respectively. Estimated average thicknesses are 120 km, 180 km and 240 km respectively [Poudjom Djomani et al., 2001], which translate into larger $Ra_c$ for Proterozoic and Archean lithosphere by factors of 3.5 and 8 respectively (assuming that all other variables are identical). The agreement between the two estimates shows that thickness variations are consistent with buoyancy differences.

[21] As most agree, the low temperature of continental lithosphere, which makes it strong, contributes to its longevity. Yet, as noted above, enough Archean lithosphere has been reactivated or altered, that strength can neither preserve Archean lithosphere indefinitely, nor maintain it against all possible perturbations [Gao et al., 2002; Kaminski and Jaupart, 2000; Lee et al., 2000; O'Reilly et al., 2001]. The hypothesis that the subcontinental mantle lithosphere persists in a thermal and mechanical state near the threshold of stability accounts for many different observations, including the existence of multiple stable states and the relationship between thickness and intrinsic density. Specifically, the lithosphere may be much less stable than previously thought based on Rayleigh-Bénard convection [e.g., Jordan, 1988].

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