On the conditions for lower crustal convective instability

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Abstract. Continental crust is thought to be formed as a result of arc magmatism, but many of the lavas produced in these settings are basaltic, while those that are silicic are commonly evolved, with lower Mg #s than the continental crust. The bulk composition of continental crust can be produced by mixing of end-member basaltic and silicic compositions, via magma mixing or in mechanical, tectonic juxtaposition, but some process is required to remove the cumulates and residues formed during generation of the silicic, “granitic” end-member. We consider convective instability of dense mafic and ultramafic lower crust as a means to remove mafic residues of basalt differentiation in order to produce end-member compositions that can mix to form the bulk composition of the continental crust. Using a range of lower crustal and mantle bulk compositions, ranging from mafic and ultramafic cumulates to primary liquid compositions, we calculated the subsolidus phase assemblage and resulting density. The results show that densities of likely lower crustal lithologies can exceed those of the mantle (by ~50–250 kg m$^{-3}$), but the density contrast is a strong function of composition, temperature, and pressure. For a “cold” geotherm with a Moho temperature of 300 °C, relevant to cratonic settings, densities of all of the lower crustal compositions that we considered, except granulite, exceed the density of the underlying mantle at pressures as low as 0.8 GPa. For a “hot” geotherm with a Moho temperature in the range of 800-1000°C, the density of the lower crust is much more variable, with gabbroic and granulite compositions having lower densities than the mantle, while “arc gabbro-norite” and ultramafic cumulate compositions having higher densities than the mantle at pressures similar to that for the cold geotherm. Instability times calculated for a two-dimensional Rayleigh-Taylor convective instability, where a dense lower crustal layer sinks into a lower-density mantle, show that high temperatures (>700°C, or >500°C with a background strain rate) are required for this process to occur on a timescale of 10 Myr with rheological parameters expected for the crust and mantle. The high temperature required for dense lower crustal mafic-ultramafic cumulates to sink into the mantle suggests that this process is restricted to arcs, volcanic rifted margins, and continental regions that are undergoing extension, are underlain by a mantle plume, or have had part of the conductive upper mantle removed.

1. Introduction

A fundamental problem in the formation of continental crust is that the majority of magmas erupted on Earth are basaltic and yet the continents do not have a basaltic bulk composition. Continental crust has an average composition that can be characterized as high Mg # (Mg/(Mg + Fe)) andesite (i.e. molar Mg # > 0.5, SiO$_2$ > 55 wt %, and <10 wt % MgO; see Kelemen, 1995; Rudnick, 1995; and references therein). High Mg # andesite lavas with major and trace element compositions similar to the continental crust are found at arcs, suggesting that continental crust formed as a result of arc magmatism [e.g., Taylor, 1967]. However, of the magmas erupted at arcs, a substantial portion are basaltic [e.g., Kay and Kay, 1994], and many of the dactitic to andesitic lavas are evolved, with lower Mg #s than the continental crust.

Fractionation of a hydrous basaltic parent magma at oxygen fugacities observed in arc magmas has not been shown by melting experiments (along a fractionation trend, e.g., “basalt” to “granite”) to produce high Mg # andesite [Kelemen, 1995]. Mixing of end-member compositions, via magma mixing, assimilation, and/or in mechanical, tectonic juxtaposition, however, can produce the high Mg # andesite composition of the bulk continental crust [e.g., Grove et al., 1982; McBirney et al., 1987]. For a mixing mechanism to generate bulk crust with a high Mg # andesite composition, some process is required to remove the cumulates and residues...
formed during generation of the silicic, “granitic” endmember (Figure 1). “Delamination” of the lower crust has been suggested as a possible mechanism for the removal of the mafic residues of basalt differentiation [Kay and Kay, 1985, 1988, 1990, 1991; Arndt and Goldstein, 1989; Turcotte, 1989]. Kay and Kay [1991, 1993] proposed that a mafic lower crust, if it is thickened and cooled sufficiently, will convert to a high-density mineral assemblage, leading to a gravitationally unstable configuration in which the lower crust can sink into the underlying lower-density mantle.

Until now there have been few efforts to quantitatively assess the dynamics of dense lower crust. Convective instability of cold, conductively cooled upper mantle beneath orogenic belts has been the subject of a considerable amount of attention [e.g., Houseman and Molnar, 1997]. In this case, crustal shortening enhances the small density contrast that has been proposed to exist between the conductively cooled and thickened conductive upper mantle and the underlying convecting mantle. However, there has not yet been any attempt to quantitatively assess the consequences of a potentially much larger density contrast between the mafic/ultramafic lower crust and the conductive underlying mantle. In this paper we consider the development of convective instabilities in dense mafic and ultramafic cumulates in the lower crust. This process is distinct from delamination because the lower crust does not “peel off” (as proposed by Bird [1978, 1979]), but rather forms “blobs” that drip off the base of the crust. However, the result of delamination and convective instability is essentially the same thing: sinking of dense lower crust into the lower-density mantle. By calculating the equilibrium subsolidus phase assemblage for a suite of lower crustal and mantle compositions, we determine the density contrast that is expected between the lower crust and underlying mantle. We then use this information, along with rheological parameters for the lower crust and mantle, to determine the timescales for the development of two-dimensional lower crustal convective instability using both a linear analysis and finite element code developed by G. A. Houseman.

In initiating this study, we reasoned that convective instability of the lower crust might be dynamically impossible. In this view, the density contrast between lower crust and conductively cooled upper mantle would be controlled primarily by temperature. At high temperature, even though the lower crust is compositionally distinct from the mantle, it might be buoyant [Kay and Kay, 1993]. At lower temperature, where lower crust might be denser than mantle, viscosities might be so high that a convective instability could not occur in geologically relevant times. At intermediate temperatures we reasoned, there might be a “window” in which lower crust is denser than upper mantle and viscosities remain low enough to permit an instability to develop. We show in this study that this window of instability is wide and that lower crustal instabilities are likely at pressures higher than 1.0 GPa, even at near-solidus conditions.
2. Lower Crustal Compositions and Subsolidus Phase Equilibria

To become convectively unstable, the lower crust must be denser than the mantle. To determine the density contrast, we chose a number of bulk compositions that span the likely compositional range of the lower crust and mantle and calculated their densities from a subsolidus equilibrium phase assemblage program developed by Connolly [1990].

The bulk compositions that we considered for the lower crust and mantle are listed in Table 1. All lower crustal lithologies considered are “mafic,” as indicated by the majority of seismic refraction data from continental crust [e.g., Rudnick and Fountain, 1995; Christiansen and Mooney, 1995, and references therein] from volcanic arcs [e.g., Gill, 1981; Sugihiro et al., 1996; Parsons et al., 1998; Holbrook et al., 1999; Fledner and Klemperer, 1999, and references therein] and from volcanic rifted margins [e.g., White et al., 1987; Zehnder et al., 1990; Barton and White, 1997, and references therein]. These include relatively refractory cumulate or restite compositions that range from a lower crustal gabbro of the Wadi Tayin Massif in the Oman ophiolite (low-P gabbro), an inferred high-pressure gabbro from beneath the Faeroes Islands flood basalt section (high-P gabbro), a typical ultramafic cumulate xenolith from the Aleutian arc (olivine clinopyroxenite), a lower crustal average composition exposed in the Talkeetna arc section of south central Alaska (arc gabbro-noritite), a median composition from a worldwide compilation of granulite xenoliths (“average” granulite), and a primary basalt composition from East Greenland (basalt). Though it is unlikely that a basalt “liquid” composition would be present in the lower crust, we include it here for comparison with the results for the cumulate compositions. The mantle was considered in terms of depleted and primitive end-members, ranging from a highly depleted mantle harzburgite from the Archean craton of East Greenland (“cratonic” peridotite), through a typical composition of residual mantle peridotite dredged from the southern Mid-Atlantic Ridge, American-Antarctic Ridge and Southwest Indian Ridge (abyssal peridotite), to an inferred primitive mantle composition (pyrolite).

The subsolidus phase equilibria calculations were performed on initial compositions that include only the major element oxides SiO₂, Al₂O₃, FeO, MgO, CaO, and Na₂O. We excluded TiO₂, K₂O, and P₂O₅ because these elements introduce uncertainty in the results and are unlikely to significantly affect the calculated subsolidus model. Cr₂O₃ was excluded in the Connolly [1990] subsolidus code, even though it is important for the formation of spinel. We tested the effect of excluding Cr₂O₃ by suppressing garnet formation in our mantle compositions and found that density was lower by only ~10 kg m⁻³. Amphibole was not included because of the complexity of the solid solutions of this mineral. However, omitting this mineral phase would favor the formation of garnet at lower pressures. Thermodynamic databases of both Holland and Powell [1998] and Berman [1988] were used in the subsolidus calculations. Except for a difference in mantle densities calculated over a large pressure range (see discussion), the difference in densities calculated using these two databases was small, and so in all results presented here we use only Holland and Powell’s [1998] more recent database.

Table 1. Crustal and Mantle Compositions Used in Subsolidus Calculations

<table>
<thead>
<tr>
<th></th>
<th>SiO₂</th>
<th>TiO₂</th>
<th>Al₂O₃</th>
<th>FeO</th>
<th>MnO</th>
<th>MgO</th>
<th>CaO</th>
<th>Na₂O</th>
<th>K₂O</th>
<th>Cr₂O₃</th>
<th>NiO</th>
<th>Mg #</th>
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<td><strong>Crustal Compositions</strong></td>
<td></td>
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<tr>
<td>Gabbro (low P)</td>
<td>49.90</td>
<td>0.31</td>
<td>16.89</td>
<td>6.00</td>
<td></td>
<td>10.50</td>
<td>14.60</td>
<td>1.50</td>
<td>0.03</td>
<td>—</td>
<td>75.73</td>
<td>81.34</td>
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<tr>
<td>Gabbro (high P)</td>
<td>48.49</td>
<td>0.61</td>
<td>15.87</td>
<td>5.12</td>
<td>0.03</td>
<td>12.52</td>
<td>16.64</td>
<td>0.50</td>
<td>0.10</td>
<td>—</td>
<td>83.85</td>
<td>54.31</td>
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<tr>
<td>Olivine clinopyroxenite</td>
<td>44.55</td>
<td>0.61</td>
<td>6.76</td>
<td>7.88</td>
<td>0.17</td>
<td>22.95</td>
<td>12.22</td>
<td>1.26</td>
<td>0.43</td>
<td>—</td>
<td>—</td>
<td>70.00</td>
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<td>Arc gabbro-noritite</td>
<td>45.66</td>
<td>0.86</td>
<td>19.26</td>
<td>11.53</td>
<td>—</td>
<td>7.69</td>
<td>12.88</td>
<td>1.74</td>
<td>0.07</td>
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<td>59.78</td>
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<td>Granulite</td>
<td>49.36</td>
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<td>16.72</td>
<td>8.78</td>
<td>0.14</td>
<td>7.32</td>
<td>10.36</td>
<td>2.45</td>
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<td>—</td>
<td>—</td>
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<tr>
<td>Basalt</td>
<td>47.50</td>
<td>1.85</td>
<td>10.89</td>
<td>12.26</td>
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<td>9.26</td>
<td>1.67</td>
<td>0.15</td>
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<td><strong>Mantle Compositions</strong></td>
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<td>Pyrolite</td>
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<td>3.30</td>
<td>8.00</td>
<td>0.15</td>
<td>38.10</td>
<td>3.10</td>
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<td>Cratonic peridotite</td>
<td>42.63</td>
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<td>0.82</td>
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<td>0.19</td>
<td>48.43</td>
<td>0.19</td>
<td>0.03</td>
<td>0.72</td>
<td>0.33</td>
<td>92.84</td>
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<tr>
<td>Abyssal peridotite</td>
<td>43.65</td>
<td>0.03</td>
<td>1.96</td>
<td>8.43</td>
<td>0.14</td>
<td>42.87</td>
<td>1.99</td>
<td>0.05</td>
<td>0.01</td>
<td>0.36</td>
<td>0.22</td>
<td>90.06</td>
</tr>
</tbody>
</table>

aOman ophiolite low-pressure gabbro [Pullister, 1984].

bEast Greenland high-pressure gabbro [Bernstein, 1994].
cAleutian ultramafic xenolith [Conrad and Kay, 1984].
dTouina arc section gabbronorite [Burns, 1983].
eAverage granulite xenolith [Rudnick and Presper, 1990].
fEast Greenland (calculated) primary basalt [Fram and Lesher, 1997].
gInferred primitive mantle [Ringwood, 1979].
hEast Greenland harzburgite xenolith [Bernstein et al., 1998].
iSouthern Mid-Atlantic Ridge dredge sample of depleted peridotite [Dick, 1989].
Reaction rates for the transformation of gabbroic rocks to denser eclogite facies (i.e., garnet + omphacite) are poorly known. Hacker [1996] points out that most gabbroic rocks are transformed to eclogite at temperatures $\geq 550^\circ C$ and that examples of incomplete transformation above $800^\circ C$ are rare. Therefore, to approximate finite reaction times for phase equilibrium to be attained at subsolidus conditions, we fixed the lowest temperature of phase equilibration at $800^\circ C$. At temperatures below $800^\circ C$ we corrected the density at constant pressure using the bulk expansivity and compressibility but kept the equilibrium phase proportions (i.e., mode) constant.

Densities calculated for the six crustal compositions listed in Table 1 are shown in Figure 2 as a function of temperature and pressure. The effect of the imposed minimum equilibration temperature is that the density changes very little at a given pressure for temperatures $< 800^\circ C$. The density of all compositions increases with increasing pressure as garnet becomes a stable mineral phase, with this transition occurring at higher pressures with increasing temperature. At temperatures below $800^\circ C$ garnet is stable at and above $\sim 0.8$ GPa for all compositions except olivine clinopyroxenite (Figure 2e), in which very little garnet forms because of the low aluminum content.

Performing these calculations on the three mantle compositions listed in Table 1 allowed us to determine the density contrast between the lower crustal and mantle compositions at a given temperature and pressure. The density contrast calculated by taking the difference between the crustal compositions of Figure 2 and the density of pyrolite at the same P-T conditions ($\Delta \rho = \rho_{\text{crust}} - \rho_{\text{mantle}}$) is shown in Figure 3. Comparison of the pyrolite with the crustal densities gives us a minimum bound on the expected density contrast because pyrolite is the densest mantle composition. Contours are shaded everywhere that the crustal compositions are more dense than the mantle. The results show that at temperatures $< 800^\circ C$ the minimum pressure at which the lower crust is denser than the mantle is $\sim 0.25$ GPa for olivine clinopyroxenite, $\sim 0.8$ GPa for arc gabbronorite and basalt, $\sim 1.5$ GPa for high-P and low-P gabbro, and $\sim 2$ GPa for average granulite.

To determine the density contrast between the crust and mantle in different tectonic settings, we chose “hot” and “cold” end-member geotherms and then calculated the resulting density profile. We will refer to the geotherms by their temperatures at 1 GPa, which is approximately the Moho pressure in typical continental crust [Rudnick and Fountain, 1995; Christensen and Mooney, 1995], arcs [Gill, 1981; Suizh Xie et al., 1996; Parsons et al., 1998; Holbrook et al., 1999; Fleder and Klemperer, 1999], and volcanic rifted margins [White et al., 1987; Zehnder et al., 1990; Barton and White, 1997]. Notable exceptions are crustal thicknesses in excess of 60 km in the Andes [e.g., Isacks, 1988] and Himalaya and Tibet [e.g., Chen and Molnar, 1981]. Profiles of the calculated density and modal composition corresponding to a cold $300^\circ C$ Moho (surface heat flow of $\sim 40$ mW m$^{-2}$) for four of the crustal compositions and compared to the density of pyrolite are shown in Figure 4. For all compositions an increase in density at $\sim 0.8$ GPa corresponds to the formation of garnet according to the reaction plagioclase + olivine = garnet, as, for example,

$$\text{CaAl}_{2}\text{Si}_{2}\text{O}_{8} + \text{Mg}_2\text{SiO}_4 = \text{CaMg}_2\text{Al}_2\text{Si}_3\text{O}_{12}. \quad (1)$$

Despite solid solutions in all of these phases, the “olivine-out” isograd is calculated to be sharp for all of the mafic compositions.

Throughout almost the entire P-T range, the average granulite composition is less dense than the gabbroic composition, while the olivine clinopyroxenite is denser than the mantle even without garnet present. Densities of the three gabbroic compositions in Figure 2 differ owing to (1) high Mg/Fe in the gabbro and granulite compositions compared to the arc gabbronorite and (2) the unusually high Al$_2$O$_3$ content of the arc gabbronorite (19.26 wt %). Both of these factors lead to formation of larger quantities of garnet at lower pressures in the arc gabbronorite composition. It should be emphasized that the arc gabbronorite we have chosen is an average of 16 analyses of lower crustal samples from the Talkeetna arc section [Burns, 1983], all of which show high Al$_2$O$_3$ from 16.18 to 23.28 wt %, and is very similar to another average of lower crustal gabbronorite compositions from the Talkeetna section [Debari and Sleep, 1991] with 18.69 wt % Al$_2$O$_3$. Furthermore, this composition is remarkably similar to aluminous gabbronorite from the Kohistan arc section [Miller and Christensen, 1994]. Thus the presence of aluminous gabbronorite, with relatively low Mg/Fe compared to ophiolites and layered gabbro intrusions, is well documented for arc lower crust.

The calculated density profiles for all of the crustal and mantle compositions listed in Table 1 is shown in Figure 5 for the $300^\circ C$ cold geotherm. The density of cratonic peridotite is less than pyrolite by $\sim 75$ kg m$^{-3}$, and the abyssal peridotite has a density similar to that of the pyrolite. When compared to the crustal compositions, the lower density of the cratonic peridotite does not change the pressure at which the crustal densities exceed that of the mantle. All compositions show an increase in density at $\sim 0.8$ GPa when garnet forms, but the amount of garnet that forms is dependent on the proportions of olivine and plagioclase in the equilibrium mineral assemblage at low pressures.

To explore the effects of a hot geotherm on the density contrast between the crust and mantle, we calculated density profiles along geotherms with Moho temperatures of $800^\circ C$ and $1000^\circ C$ (Figure 6). The conductive
Figure 2. Calculated densities (kg m$^{-3}$) for crustal compositions of (a) low-pressure gabbro, (b) high-pressure gabbro, (c) average granulite, (d) basalt, (e) olivine clinopyroxenite, and (f) arc gabbro-norite using Connelly’s [1990] subsolidus phase equilibria program (VERTEX) and Holland and Powell’s [1998] thermodynamic database. Major element chemistry for the compositions shown is listed in Table 1. The lowest temperature of equilibration was set at 800°C.
Figure 3. Calculated density contrast ($\text{kg m}^{-3}$) between the crustal compositions shown in Figure 2 and a pyrolite bulk composition calculated at the same P-T conditions for (a) low-pressure gabbro, (b) high-pressure gabbro, (c) average granulite, (d) basalt, (e) olivine clinopyroxenite, and (f) arc gabbro-norite. Shaded contours are shown only where the density contrast is positive, indicating the P-T conditions at which the crustal compositions are denser than the mantle.
Figure 4. Calculated density and mode for (a) olivine clinopyroxenite, (b) arc gabbronorite, (c) low-pressure gabbro, and (d) average granulite overlying a pyrolite mantle composition for a “cold” 300°C geotherm. The temperature profile used in the density calculations is shown along with the mantle adiabat (dashed line) extending to the surface, corresponding to a mantle potential temperature of 1300°C. The bold solid curves show the density profile for the crust extending down to 2.0 GPa. The light dotted curve shows the pyrolite density extended to the surface. The calculated mode for the crust and mantle compositions are abbreviated: (ol) olivine, (cpx) clinopyroxene, (opx) orthopyroxene, (plg) plagioclase, (sp) spinel, (gt) garnet, (ky) kyanite, and (qrtz) quartz.

"Cold " 300 °C Geotherm

Figure 5. Calculated densities of all crustal and mantle compositions along a 300°C geotherm. Crustal densities are shown extending down to a pressure of 2.0 GPa. Densities for pyrolite (thin dashed curve), abyssal peridotite (thin dotted curve), and cratonic peridotite (thin solid curve) are shown over the entire pressure range.
Figure 6. Calculated densities for crustal and mantle compositions corresponding to hot 800°C and 1000°C geotherms. The gray horizontal line in the density plots marks the pressure at which the conductive temperature profile intersects the adiabat. The density of the high- and low-P gabbros are significantly reduced at these higher temperatures, while the arc gabbronorite and basalt compositions have densities that are very sensitive to small changes in temperature and pressure.

Part of these geotherms extends to the pressure at which the geotherm intersects the mantle adiabat. This transition depth is shown as a gray horizontal line in the density plots. At high temperatures, both the “high-P” and “low-P” gabbro compositions are less dense than the mantle over almost the entire pressure range considered. The arc gabbronorite and basalt compositions have densities that differ markedly between the 800°C and 1000°C geotherms. For the 800°C geotherm, both compositions are denser than the mantle, but for the 1000°C geotherm, both have densities that are less than that of the mantle down to the base of the conductive geotherm. The reason for this density decrease is that the P-T boundary of the garnet stability field lies between the 800°C and 1000°C geotherms for these compositions.

(Note that, in fact, the arc gabbronorite composition would be partially molten at 1300°C and 1.2 GPa, and this effect has not been incorporated into our density calculations. As well, the time taken for phase transformations to occur would be lowered if melting occurred, and if the melt migrates upward, the density of the residues from the melting would be greater than that of the unmelted rock [Kay and Kay, 1993].)

The evolution of temperature and density for the arc gabbronorite starting from an initial 1000°C geotherm and cooling conductively to an 800°C geotherm is shown in Figure 7. The depth to the base of the conductive geotherm increases from 1.2 to 1.6 GPa in this calculation. The result shows a transition from an initially stable configuration of positively buoyant arc lower crust at high temperature to a dense and unstable lower crust as a result of relatively small amounts of cooling on a timescale of ~10.2 Myr.

In summary, subsolidus phase equilibria calculations show that density of mafic and ultramafic lower crustal compositions can exceed the density of the mantle, but the pressure at which this occurs is a strong function of composition and temperature. With a cold geotherm, relevant to cratonic settings, the density of five of the six lower crustal compositions exceeds that of the mantle at a minimum pressure between 0.25 and 1.5 GPa. Olivine clinopyroxenite, arc gabbronorite, and basalt have densities that are greater than that of the mantle at minimum pressures <1.0 GPa. The density contrast between the crust and mantle varies from 50 to 250 kg m⁻³, depending on the compositions chosen. For a hot geotherm, densities are lower, with only olivine clinopy-
roxenite, arc gabbronorite, and basalt being denser than the mantle if the Moho temperature is 800°C. For a 1000°C Moho, olivine clinopyroxenite is the only composition denser than the mantle because its density is not dependent on the formation of garnet.

3. Rheology of the Lower Crust and Mantle

The timescale over which a dense lower crust can become convectively unstable is dependent not only on the density contrast but also on the strength of the lower crust and mantle. To assess the effect of rheology on the development of convective instabilities, both dislocation creep and diffusion creep were considered as end-member deformation mechanisms. The general form of the flow law is given by

$$\dot{\varepsilon} = A \sigma^n \exp(-Q/RT) d^m,$$

where $\dot{\varepsilon}$ is the strain rate, $A$ is an empirical constant, $n$ is the differential stress ($\sigma_3 - \sigma_3$), $Q$ is the activation energy, $R$ is the universal gas constant, $T$ is the absolute temperature, $d$ is the grain size, and $m$ and $n$ are exponents that depend on the mechanism controlling deformation. For dislocation creep ($m = 0$; $n > 1$) the strain rate is independent of the grain size but varies nonlinearly with the stress, while for diffusion creep ($m < 0$; $n = 1$) the strain rate varies linearly with the stress but nonlinearly with the grain size. In general, the transition from diffusion creep to dislocation creep occurs when the differential stress is large ($\approx 100$ MPa for dry olivine of Hirth and Kohlstedt [1996]). In this study, we examine the effect of the end-member deformation mechanisms separately and do not take into account the transition from diffusion to dislocation creep.

Experimentally determined rheological parameters can be related to those used for viscous flow by assuming a general constitutive relation between deviatoric stress $\tau_{ij}$ and strain rate $\dot{\varepsilon}_{ij}$,

$$\tau_{ij} = B \dot{\varepsilon}_{ij}^{1/n-1} \dot{\varepsilon}_{ij},$$

where $\sigma_{ij} = \rho \dot{\varepsilon}_{ij} + \tau_{ij}$, $B$ is a rheological coefficient, $\dot{\varepsilon}$ is the second invariant of the strain rate tensor, and $n$ is the stress exponent. The effective viscosity $\eta$ can then be written as

$$\eta = \frac{B}{2} \dot{\varepsilon}_{ij}^{1/n-1},$$

where the rheological coefficient $B$ is a function of the rheological parameters and is defined as

$$B = A^{-1/n} \exp\left(\frac{Q}{nRT}\right).$$

For diffusion creep, $\eta = B/2$ and the strain rate varies linearly with the applied stress. For dislocation creep,

<table>
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<tr>
<th>Table 2. Rheological Parameters</th>
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<tr>
<td>$n$</td>
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<tr>
<td>Clinopyroxenite$^a$</td>
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<tr>
<td>Jadeite$^b$</td>
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<td>Garnet$^c$</td>
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<td>Olivine (dry)$^d$</td>
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</tbody>
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$^a$Boland and Tullis [1986]  
$^b$Stockhart and Rener [1998]  
$^c$Ji and Martignole [1994]  
$^d$Hirth and Kohlstedt [1996]  

Figure 7. Density profile of the arc gabbronorite composition during cooling from a 1000°C geotherm to a 800°C geotherm. Time intervals are shown in Myr, starting at 0 Myr for the initial 1000°C profile, with cooling complete in 10.2 Myr. The arc gabbronorite composition extends down to the pressure where the 800°C geotherm intersects the adiabat (dashed gray line). Densities for pyrolite (thin solid curve) and cratonic peridotite (thin gray curve) mantle compositions are shown for reference.
η varies non-linearly with the strain rate, so that increasing the strain rate decreases the effective viscosity.

The values of rheological constants relevant to the lower crust and mantle are listed in Table 2. For the mantle it is generally agreed that olivine is the main phase controlling deformation [e.g., Hirth and Kohlstedt, 1996]. For the lower crust, garnet and omphacite (sodium-rich clinopyroxene) have been proposed as the main phases controlling deformation. Experimental results on the deformation of omphacite are not available. However, the measured flow laws for diopside (clinopyroxenite) and jadeite are possible end-members. Effective viscosities of clinopyroxenite, jadeite, garnet, dry olivine, and wet olivine calculated from

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

are shown in Figure 8 as a function of differential stress at temperatures of 600°C, 900°C and 1200°C. These mineral phases have effective viscosities that span nearly 10 orders of magnitude at low temperatures and ~5 orders of magnitude at high temperatures. We therefore chose two different rheological models for the crust and mantle: one where the crust is strong and has the same rheological coefficient \( B_1/B_0 = 1 \) as the underlying mantle, and the other where the lower crust is weak and has a rheological coefficient that is 2 orders of magnitude less than the mantle \( B_1/B_0 = 0.01 \). This choice of rheological coefficients is based on the results shown in Figure 8 and the fact that at a given differential stress, \( B \propto \eta^{1/n} \). We used a constant stress exponent of \( n = 3.5 \) for the crust and mantle; the results differed very little for \( n \) ranging from 3.3 to 3.6.

4. Convective Instability of Dense Lower Crust

To investigate the removal of dense mafic cumulates from the base of the crust, we assume that this process occurs as a convective instability, resulting from a dense fluid overlying a less dense fluid in a gravitational field. This problem has been studied extensively [e.g., Rayleigh, 1883; Taylor, 1950; Chandrasekhar, 1961; Whitehead and Luther, 1975; Canright and Morris, 1993; Houseman and Molnar, 1997]. We use a linear analysis that follows the development of Conrad and Molnar [1997] and Bassi and Bonnin [1988], and the two-dimensional finite element code of G. A. Houseman in order to determine growth rates and timescales of this process. Linear analysis of Rayleigh-Taylor instability allows for the solution of the equation of motion for infinitesimal deflections of the interface between the dense layer and underlying low-density layer for both linear and non-linear viscosity. However, use of finite element code to characterize this process for non-Newtonian viscosity is important for accurately determining the growth of an instability at large amplitudes because the viscosity depends on the strain rate, which varies spatially. Houseman and Molnar, [1997]. We use both a linear analysis and finite element code to study the growth of convective instabilities with a linear viscosity but use only the finite element code to study the growth of instabilities for a nonlinear viscosity. The linear analysis is not presented here in detail because it is outlined extensively by Conrad and Molnar [1997] and because with multiple layers (i.e., more than 2) it is not possible to obtain simple analytic expressions.

A schematic depiction of the layout and parameters chosen for a dense lower crustal layer overlying a less
pertainring to the interface an instability would not develop.

Solving the Navier-Stokes equations with inertial terms neglected (i.e., Stoke's flow) is facilitated by non-dimensionalization. Doing so aids in determining the fundamental behavior of the solution and leads to the definition of characteristic scaling values for length, density, and viscosity. For length scales we chose either the dense layer thickness $h$ or viscous decay length $L$ of the mantle, depending on whether the viscosity profile of the mantle is constant, or exponentially decreasing. The density contrast between the dense layer and mantle $\Delta \rho$ and the rheological coefficient $B$ at the top of the mantle were chosen as the other characteristic scaling values, leading to a timescale given by

$$T = \left( \frac{B}{2 \Delta \rho g [h, L]} \right)^n,$$

where $[h, L]$ implies that either $h$ or $L$ is used, depending on the whether the dense layer thickness or viscous decay length of the mantle are used. This scaling equation contains the fundamental interdependency of the various parameters which affect the timescale for the growth of an instability. In the following sections, nondimensional (primed) quantities refer to their dimensional counterparts with the following relations:

$$Z = Z'[h, L]; \quad \lambda = \lambda'[h, L]; \quad t_b = t_b' T,$$

$$q = q'/T; \quad u = u'[h, L]/T; \quad \dot{e} = \dot{e}'/T.$$

Exponential decrease in viscosity in the mantle was characterized in terms of the ratio between the length scale of viscous decay and thickness of the dense layer ($L/h$). Values of $L/h$ of 3.474, 1.158, and 0.8681 were chosen in order to approximate the decrease in viscosity over 1, 5, and 10 km length scales in the conductively cooling mantle. Exponentially decreasing viscosity approximates the effect of a conductive geotherm, where the temperature changes significantly with depth over the thickness of the conductively cooled upper mantle, while a constant viscosity profile approximates the effect of an adiabat, where the temperature changes very little over the depth range considered.

The development of a convective instability caused by an initially small displacement to the base of the dense layer is shown in Figure 10. In this finite element calculation the initial displacement is in the form of a sinusoidal perturbation with a wavelength equal to the horizontal extent of the dense layer, but it is clear that the fastest growing wavelength is much shorter and dominates the behavior of the instability at later times. For the fastest growing wavelength, Figure 10b shows a series of snapshots at different times during the development of an instability. Starting initially from an
undeformed state, the time series shows the formation of an instability and the dense layer sinking into the underlying mantle. In this example the horizontal extent is half the total wavelength (i.e., mirror symmetry is assumed at the edge of the layers). The initial displacement of the dense layer is greatest on the left, and it is clear that this is where the instability initiates. Note that the entire dense layer participates in the instability. Growth of the instability is determined by tracking the depth of maximum displacement of the interface as a function of time.

For a Newtonian viscosity a convective instability grows exponentially with a growth rate \( q' \), starting from an initial displacement \( Z'_0 \) according to

\[
Z' = Z'_0 \exp(q't').
\]

The maximum downward displacement of the instability \( Z' \) as a function of time \( t' \) is shown in Figure 11a. Also shown is the result of plotting \( \ln(Z') \) as a function of time, which yields a nearly straight line and indicates that the growth is exponential. It is convenient to define the instability time \( t'_b \) for a Newtonian viscosity as the time at which the interface is displaced downward by an amount equal to the initial dense layer thickness

\[
t'_b = -\frac{1}{q'} \ln \left( \frac{Z'_0}{Z'} \right) T.
\]

For non-Newtonian viscosity the dependence of viscosity on strain rate causes fundamentally different behavior in the growth of an instability. Growth is super-exponential but is initially slow at small deflections of the interface. With time, growth rates increase as the deflection of the interface increases, leading to the rapid and almost instantaneous deflection of the interface (Figure 11b). *Houseman and Molnar [1997]* derived a simplified theory that approximates this behavior, where the vertical displacement of the boundary is given by

\[
Z' = \left[ (n - 1) \left( \frac{C'}{n} \right) \left( t'_b - t' \right) \right]^{1/n}
\]

and the instability time is given by
4.1. Newtonian Viscosity (Diffusion Creep)

Growth rates for a Newtonian viscosity calculated from both a linear analysis and finite element experiments are shown as a function of wavenumber in Figure 12. The results show that the finite element calculations (points) are consistent with the linear analysis (curves). Determining the fastest growing wavelength of the instability is important for characterizing the instability, since this wavelength will eventually dominate the growth, regardless of the wavelength of the initial perturbation. In Figure 12a the dense lower crustal layer is strong and has a rheological coefficient equal to that of the mantle ($B_1/B_0 = 1$) and in Figure 12b the dense lower crustal layer is weak and has a rheological coefficient that is 2 orders of magnitude less than that at the top of the mantle ($B_1/B_0 = 0.01$). The effect of exponentially decreasing $B_0$ in the mantle with depth ($L/h = 3.474, 0.8681$) is also shown and compared to a case where $B_0$ is constant ($L >> h$). For a strong dense layer, maximum (nondimensional) growth rates ($q_{\text{max}}$) are $\sim 0.15-0.25$, while for a weak dense layer, maximum growth rates are $\sim 4-12$, depending on the decay length of $B_0$ in the mantle. The wavenumber of maximum growth rate ($k_{\text{max}}'$) is $\sim 0.9$ for a strong dense layer and 0.5 for a weak dense layer, corresponding to wavelengths of 7 and 12.5 times the dense layer thickness, respectively. Limiting values of maximum growth rate ($q_{\text{max}}$) and wavenumber ($k_{\text{max}}'$) as a function of $L/h$ calculated from the linear analysis are shown in Figure 13. It is clear that the wavenumber of fastest growth reaches values of 0.9 and 0.3 (for strong and weak dense layers, respectively) when there is a constant rheological coefficient in the mantle $L >> h$, and 1.1 and 1.5 when there is a strongly depth dependent rheological coefficient ($L << h$). These results show that the rheological coefficient of the dense layer and the decay length of rheological coefficient in the mantle can strongly influence the growth rate of the instability. Furthermore, combining a weak dense layer with a strongly depth-dependent mantle rheological coefficient produces the fastest growth rates: the weak dense layer experiences very little viscous resistance from the mantle as the decay length ($L/h$) in the mantle decreases.

To determine dimensional instability times from the maximum growth rates shown in Figure 12, we scale the results using equations (7) and (10). For the fastest growing wavenumber, Figure 14 shows the calculated instability time as a function of Moho temperature for 1 and 10 km thick strong and weak dense lower crustal layers and density contrasts of 50, 100, and $200 \text{ kg m}^{-3}$. The Moho temperature required for a 10 Myr instability is shown by the vertical shaded bars, and is related to the rheological coefficient at the top of the mantle by equation (5). For a 10 km thick, strong dense layer (Figure 14a), a 950-1000°C Moho temperature is required for an instability to develop with a density contrast of 50-200 kg m$^{-3}$ and a constant rheological coefficient.

$$t_b = \left( \frac{n}{C'} \right)^n \frac{Z_0^{(1-n)}}{(n-1)} T,$$

where $C'$ is a growth rate factor and is analogous to $q'$. Plotting $Z(t^{1-n})$ as a function of time yields a linear relationship (Figure 11b), verifying that that deflection increases according to equation (11), and allows calculation of the growth rate $C'$. The abrupt initiation and completion of the instability for non-Newtonian rheologies occurs at a small fraction of the initial displacement.

![Figure 11](image-url)
mantle. A 1 km thick dense layer requires Moho temperatures ~100°C greater for the same instability time. It is clear that the density contrast has a smaller effect on the instability time than the rheological coefficient (i.e., Moho temperature) because the rheological coefficient varies by orders of magnitude over the temperature range considered, but the density contrast varies by < 1 order of magnitude. An exponentially decreasing rheological coefficient for the mantle decreases the required Moho temperature for an instability to develop by an almost insignificant amount (Figure 14c), indicating that the strength of the dense layer is important in controlling the time scale for the formation of a convective instability. For a weak dense layer overlying a constant rheological coefficient mantle (Figure 14b) convective instability of the dense layer occurs in 10 Myr if the Moho temperature is ~900°C for a 10 km thick dense layer and 1000°C for a 1 km thick dense layer. The lowest Moho temperature for a 10 Myr instability occurs with an exponential decaying rheological coefficient in the mantle and a dense layer that is weak (Figure 14d). Moho temperatures in this case are ~800°C for an instability time of 10 Myr for a 10 km thick dense layer and ~950°C for a 1 km thick dense layer. These instability times refer to an initial displacement of Z₀ = 1%. A larger initial displacement does not change the exponential growth rate of the instability.

4.2. Non-Newtonian Viscosity (Dislocation Creep)

The growth rate factor \( C' \) for a non-Newtonian viscosity with \( n = 3.5 \) for all layers is shown in Figure 15. An additional decay length for the rheological coefficient of the mantle was included in these results (\( L/h = 1.1581 \)) in order to help clarify the dependence of growth rate factor on the decay length of \( B₀ \) because an accurate linear analysis cannot be made. When the dense lower crustal layer is strong \( (B₁/B₀ = 1) \), the wavenumber of fastest growth is about \( k'_{\text{max}} = 1.5 \), and when the dense lower crustal layer is weak \( (B₁/B₀ = 0.01) \), the fastest growing wavenumber is about \( k'_{\text{max}} = 0.5 \). Compared to the results for a Newtonian viscosity, \( k'_{\text{max}} \) is shifted to slightly higher values and corresponds to wavenumbers of maximum growth rate factor of ~4.2 and 12.5 times the dense layer thickness, respectively. It is also clear from the results that the growth rate factor \( C' \) depends only weakly on \( L \) for \( L/h < 1.158 \).

Instability times for the fastest growing wavenumbers for strong and weak dense layers are shown in Figure 16. These results all assume an initial perturbation of the interface between the dense layer and the mantle of 100% of the initial dense layer thickness. Houseman and Molnar [1997] showed that for non-Newtonian viscosity the instability time for an instability with different initial deflections is related by...
\[
\frac{1}{T_{\text{Moho}_1}} - \frac{1}{T_{\text{Moho}_2}} = \frac{nR}{Q} \ln \left( \frac{b_1}{b_2} \right)^{1/n} 
\]

and is \(\sim 150^\circ\text{C}\) if the Moho is 600\(^\circ\text{C}\), and \(\sim 350^\circ\text{C}\) if the Moho is 1000\(^\circ\text{C}\).

For a strong dense layer overlying a constant rheological coefficient mantle (Figure 16a) the Moho temperature required for an instability to develop in 10 Myr is \(\sim 900-1000^\circ\text{C}\) for a 10 km thick dense layer and \(\sim 1100-1300^\circ\text{C}\) for a 1 km thick dense layer, depending on the density contrast. Note that since the instability time depends in part on \(\Delta \rho^{-3.5}\), the effect of the density contrast on Moho temperature is larger for a non-Newtonian viscosity than for a Newtonian viscosity where the instability time depends on \(\Delta \rho^{-1}\). An exponentially decreasing rheological coefficient in the mantle requires Moho temperatures \(\sim 50^\circ\text{C}\) less than if the rheological coefficient is constant. A weak dense layer over a constant \(B_0\) mantle requires a Moho temperature of \(\sim 800-900^\circ\text{C}\) for a 10 km thick dense layer and \(\sim 1000-1150^\circ\text{C}\) for a 1 km thick dense layer to become unstable in 10 Myr (Figure 16b). As with a Newtonian viscosity, an exponentially decreasing \(B_0\) in the mantle and a weak dense layer results in the lowest Moho temperatures. These are \(\sim 650-700^\circ\text{C}\) for a 10 km thick dense layer and \(\sim 900-1000^\circ\text{C}\) for a 1 km thick dense layer to become unstable in 10 Myr. For “wet” olivine the rheological coefficient is \(\sim 1\) order of magnitude less (see Figure 8), resulting in a decrease in Moho temperatures of 100-150\(^\circ\text{C}\) for an instability to develop in 10 Myr.

### 4.3. Horizontal Shortening and Non-Newtonian Viscosity

Imposing an initial background strain rate when the viscosity is non-Newtonian reduces the initial effective viscosity and therefore decreases the instability time. Molnar et al. [1998] show that the deformation can be divided into two separate stages: an initial period of exponential growth followed by a period of superexponential growth. The growth rate for the initial stage is determined by the effective viscosity caused by the initial background strain rate. To quantify this effect for the dense lower crustal layer, we performed numerical experiments with different initial horizontal strain rates. Figure 17 shows the instability time \(t_{\text{in}}^t\) as a function of horizontal strain rate \((\dot{e}_{\text{xx}}^t)\) for strong and weak lower crust. After \(\dot{e}_{\text{xx}}^t\) exceeds a threshold value, the instability time decreases in proportion to the increase in strain rate (as viewed on a log-log plot). Higher initial horizontal strain rates cause shorter instability times because of lower initial effective viscosities. In practice it is difficult to explore this effect numerically because large initial horizontal strain rates change the dimensions of the dense lower crustal layer, leading to a change in the wavelength of the instability. For the
results shown in Figure 17, the maximum horizontal strain rate in our numerical experiments reduces the horizontal dimension of the dense layer by ~20%, but this does not affect our estimates of maximum growth rate factor significantly.

We can exploit the linear relationship between instability time and horizontal strain rate (shown in Figure 17) to determine the change in instability time for larger strain rates, assuming that this linear relationship still holds. The change in instability time for a given initial horizontal strain rate, after simplification, is given by

$$ t_B' = \begin{cases} 
  t_{B0}' \left( \frac{\epsilon''_{zz0}}{\epsilon''_{zz0}} \right) & \epsilon''_{zz} \geq \epsilon''_{zz0} \\
  t_{B0}' & \epsilon''_{zz} < \epsilon''_{zz0}, 
\end{cases} $$

(15)

where $t_{B0}'$ is the nondimensional instability time with no initial horizontal strain rate, $\epsilon''_{zz0}$ is the minimum initial strain rate at which the instability time is affected by horizontal shortening, and $\epsilon''_{zz}/\epsilon''_{zz0}$ is determined from Figure 17.

The effect of imposing initial horizontal strain rates of $10^{-18}$ s$^{-1}$ and $10^{-14}$ s$^{-1}$ on the growth of an instability with a non-Newtonian viscosity is shown in Figure 18. It is clear from the results that an initial strain rate can significantly reduce the Moho temperature required for an instability to develop. At high Moho temperatures, where instabilities develop quickly in response to the already small $B_0$ values, instability times are relatively unchanged by the presence of an initial background strain rate. With a strain rate of $10^{-14}$ s$^{-1}$, Moho temperatures for a 10 Myr convective instability of 1 km thick dense layer are reduced to as low as 700-800°C for a strong dense layer and 650-750°C for a weak dense layer. For a 10 km thick dense layer the same initial horizontal strain rate reduces Moho temperatures to 600-650°C for a strong dense layer and to 550-600°C for a weak dense layer.
Figure 15. Calculated growth rate factor ($C'$) as a function of wavenumber ($k' = 2\pi/\lambda'$) for (a) strong ($B_1/B_0 = 1$) and (b) weak ($B_1/B_0 = 0.01$) dense lower crustal layers with a non-Newtonian viscosity and a mantle with several different viscous decay lengths ($L/h$). If $L >> h$, the viscosity of the mantle is nearly constant, while if $L/h > 1.158$, the viscosity of dense layer dominates the growth of an instability.

Figure 16. Instability times calculated from equation (12) as a function of Moho temperature for a non-Newtonian viscosity with four different profiles: (a) strong dense layer over a constant $B_0$ mantle, (b) weak dense layer over a constant $B_0$ mantle, (c) strong dense layer over a mantle with exponentially decreasing $B_0$, and (d) weak dense layer over a mantle with exponentially decreasing $B_0$. Results are shown for 10 and 1 km dense layer thickness (dark and light shaded regions) and with density contrasts of 50, 100, and 200 kg m$^{-3}$ (labeled curves). Intersection of the calculated instability times with the horizontal gray line gives the Moho temperature required for the instability to develop in 10 Myr (shown as light shaded bars).
5. Discussion

Calculations of subsolidus phase equilibria and convective instability times show, for likely rheologies for the lower crust and mantle, that dense mafic and ultramafic cumulates in the lower crust can initiate convective instabilities on geologic timescales. The low viscosity necessary for the development of an instability on a timescale of 10 Myr or less requires either high temperatures (>700°C), or large initial strain rates (\(\dot{e}_{\text{bulk}} > 10^{-14}\text{s}^{-1}\)). This suggests that lower crustal convective instability can be ruled out in old, tectonically stable cratonic settings where the geotherm is cold (\(\sim 300^\circ\text{C}\) at 1 GPa) and is restricted to regions where Moho temperatures are in excess of \(\sim 500^\circ\text{C}\) or, in the absence of a background strain rate, higher than \(\sim 700^\circ\text{C}\). Thus arcs, (young) passive margins, and continental regions in which the thermal boundary layer is thinned are likely settings for lower crustal convective instability. Large strain rates from continental extension (rifting) or shortening (orogenic zones) may help instabilities develop with colder geotherms.

In arc settings the combined effect of high Moho temperatures (800-900°C at 1.1 GPa) [e.g., Debari et al., 1987; Debari and Coleman, 1989] and decreased effective viscosities for wet olivine enhances the likelihood of lower crustal convective instability. Furthermore, the effects of arc magma composition favor the formation of dense lower crust. As noted by many workers [e.g., Kay and Kay, 1994], high H\(_2\)O contents in primitive arc basalts will generally suppress plagioclase stability, leading to crystallization of abundant olivine clinopyroxene cumulates. This, in turn, produces high-Al derivative liquids which produce Al-rich cumulate gabbros and gabbroidalites at lower pressures [e.g., Debari and Coleman, 1989]. Thus the dense lower crustal compositions used in our Figures 2c, 2f, 3e, 3f, 4a-4d, 5b, and 6b are taken from arc xenoliths [Conrad and Kay, 1984] and the base of an exposed arc crustal section in Alaska [Burns, 1993; Debari and Coleman, 1989] and may be typical of lower crustal rocks in arcs since they are similar to deep crustal rocks in the Kohistan arc section in Pakistan [Jan and Howe, 1981; Miller and Christensen, 1994].

As previously suggested by Arnald and Goldstein, olivine clinopyroxenites at depths \(\sim 8\) km are denser than the mantle by \(\sim 50\) kg m\(^{-3}\) because cumulate ultramafic rocks have higher Fe/Mg than residual mantle. Along a “steady state” arc geotherm they would probably have to reach thicknesses of up to 10 km to become convectively unstable on geological time scales. However, if Moho temperatures are higher, for example, if Moho temperatures are similar to liquidus temperatures for hydrous basaltic magmas, at around 1200°C, then ultramafic cumulate layers a few hundred meters thick will become unstable in very short times, perhaps as

![Figure 17. Nondimensional instability time (\(t_{\text{b}}^*\)) as a function of nondimensional horizontal strain rate (\(\dot{e}'\)) for strong and weak dense lower crustal layers with non-Newtonian viscosity. With small initial horizontal strain rates, instability times are unchanged. After reaching a critical initial horizontal strain rate that varies for the strong and weak dense layers, instability times decrease because the initial effective viscosity is lower.](image-url)

In summary, application of the scaling laws show that for a Newtonian viscosity, Moho temperatures must be greater than \(\sim 900^\circ\text{C}\) to produce an instability in 10 Myr if we assume that 10 km is representative of the thickness of the dense lower crustal layer. Instability at temperatures as low as 800°C is possible only if the dense lower crustal layer is weak and the decay length of the rheological coefficient of the mantle is small. Non-Newtonian viscosity requires Moho temperatures of 850-1000°C for a 10 Myr instability, and results show a much stronger dependency on the density contrast between the dense layer and mantle than for a Newtonian viscosity. Temperatures as low as \(\sim 700^\circ\text{C}\) can result in a 10 Myr instability if the dense layer is weak and the rheological coefficient of the mantle decays exponentially. Horizontal shortening can reduce the Moho temperature for a 10 Myr instability with a non-Newtonian viscosity to \(\sim 550-650^\circ\text{C}\) for an initial strain rate of \(10^{-14}\text{ s}^{-1}\), assuming a dense layer thickness of 10 km. A wet olivine rheology would reduce Moho temperatures required for a 10 Myr instability by an additional 100-150°C because of the lower effective viscosity of H\(_2\)O-rich olivine [Hirth and Kohlstedt, 1996].
Figure 18. Instability times with initial horizontal strain rates of $10^{-18}$ s$^{-1}$ and $10^{-14}$ s$^{-1}$ and without any initial horizontal strain rate. Results (a) for a 1 km thick dense layer and (b) for a 10 km thick dense layer with both constant and exponentially decreasing $B_0$. Intersection of the horizontal gray lines with the calculated instability times give Moho temperatures required for 10 Myr instabilities to develop from 1% and 100% initial deflections.
quickly as they form, for Moho pressures of 1.0 GPa or more.

Arc gabbronite compositions, on the other hand, have density contrasts of up to 150 kg m\(^{-3}\) for 800°C geotherm, and are less dense than the mantle at higher temperatures (i.e., 1000°C at 1.0 GPa). Therefore arc gabbronite lower crustal sections could initially form a stable lower crust and then become convectively unstable after garnet formation due to a small amount of cooling. Because they become dense at temperatures as high as 950°C (see Figure 7), the low viscosities of the mantle required for short instability times will be present. Their high densities would probably cause these garnet-rich lithologies to become convectively unstable at thicknesses <10 km.

For continental regions, gabbro and olivine clinopyroxene compositions are likely to represent the mafic and ultramafic component of the lower crust. Results from Figure 5 show that the gabbros and granulite xenolith compositions are generally stable at the high temperatures needed for convective instability, except if a composition similar to our high-pressure cumulate extends to depths >40 km (1.4 GPa), in which case it will be denser by ~50 kg m\(^{-3}\) with respect to our most depleted mantle composition. The olivine clinopyroxenite, on the other hand, is denser than the mantle by ~50 kg m\(^{-3}\) at depths as shallow as 8 km (0.25 GPa). Thus convective instability of ultramafic lower crustal cumulates is possible, but only if temperatures are hot due to rifting, the influence of a mantle plume, or thinning of the thermal boundary layer. Rifting will lower the effective viscosity by imposing a background strain rate on the crust and mantle as well as increasing the temperature by thinning the conductive upper mantle. In addition, the influence of a mantle plume could raise Moho temperatures significantly by increasing temperatures, by convective “erosion” of the conductively cooled upper mantle, and by advective heat transport due to both (1) ascent of melt into the crust and (2) the ascent of isotherms during thinning of the conductive upper mantle. A further possibility is that thickened continental crust could have a hotter geotherm due to increased heating from radiogenic elements [e.g., Le Pichon et al., 1997].

In cold continental regions, convective instability of the lower crust may be possible if the conductively cooled upper mantle becomes convectively unstable first. Convective instability of the upper mantle has been proposed as a process which might account for the observation that compressional thickening precedes extensional collapse of many orogens [e.g., England and Housman, 1989; Housman and Molnar, 1997; Molnar et al., 1998]. The combined effect of decreased effective viscosities due to horizontal shortening and entrainment of the lower crust with a convectively downwelling upper mantle may cause dense lower crust to participate in “mantle” convective downwelling, as suggested by Kay and Kay [1993, 1994]. If only a portion of the conductive upper mantle is convectively removed due to increased viscosities at shallow levels [Molnar et al., 1998], then the remaining thinner conductive upper mantle would have a higher temperature, potentially allowing dense lower crustal cumulates to subsequently become unstable on geologic timescales short compared with the time scale of orogeny.

The subsolidus equilibria results show that the density contrast between lower crustal and mantle compositions is likely to exceed any density contrast expected for the mantle due to conductive cooling. Controversy exists as to whether the possible density inversion caused by conductive cooling in the mantle is counterbalanced by positive compositional buoyancy of the upper mantle caused by the extraction of melt [Oxburgh and Parmentier, 1977; Jordan, 1988]. In addition, we calculated the density profile for the three mantle compositions down to a depth of 200 km and found that for a conductive geotherm representative of a cratonic setting (300°C Moho temperature), densities increased with depth when using Holland and Powell’s [1998] database but decreased with depth using Berman’s [1988] database (Figure 19), because the values for the thermal expansivity and compressibility differ. This result shows that the association of a conductive geotherm with increasing density at decreasing depth in a constant composition mantle is not necessarily correct. Furthermore, since this does not take into account possible increased chemical buoyancy of the mantle at shallower depths [i.e., Jordan, 1988], the results using Holland and Powell’s [1998] database suggests a stable density structure for much of the conductive continental upper mantle.

A remaining question is What happens to dense lower crust after it becomes convectively unstable? Does it sink into the convecting mantle or does it remain trapped in the conductively cooled upper mantle? Our calculations do not take into account changes in density and temperature of the dense lower crustal layer after an instability is initiated. Competing effects of increasing temperature and pressure would likely change the density of a descending blob. However, from the results shown in Figure 3, the P-T trajectory of a descending lower crustal blob is likely to maintain the negative buoyancy contrast with the mantle. The depth to which descending blobs sink is likely to depend on the scale at which the instability forms and whether melting occurs. It has been suggested that subducting oceanic crust, after converting to eclogite (i.e., garnet and omphacite), can sink as far as the core-mantle boundary [e.g., van der Hilst et al., 1998].

In light of our results, it is interesting to speculate on the influence that lower crustal convective instability has on the thickness of continental crust, which ranges between 25 and 45 km away from orogenic belts [Hobrook et al., 1992; Rudnick and Fountain, 1995]. The results of the phase equilibrium calculations show a broad P-T range in which plagioclase coexists with garnet for
Figure 19. (a) Density profiles calculated for a pyrolite mantle composition using Holland and Powell's [1998] and Berman's [1988] thermodynamic databases along a 30°C geotherm extending down to a pressure of 6 GPa. (b) The calculated density profile using Holland and Powell's [1998] database with a positive slope (solid curve) and which is convectively “stable,” while the density profile calculated using Berman's [1988] database (dashed curve) has a negative slope and is “unstable”. The reason for this is evident from the different values for (c) bulk expansivity and (d) compressibility calculated from these two databases.

all except the ultramafic composition. If the Moho transition, which is defined by an increase in P wave velocities from ~7.2 to 8.0 km s⁻¹, is due to the gabbro to eclogite transition it should be gradual. However, seismic data indicate that for a large majority of cases the Moho transition is sharp and extends over only a few kilometers. Therefore the Moho is not due to a phase change but is more likely a break between mafic crustal compositions and underlying ultramafic mantle. We can then ask, If the Moho depth is not simply the result of a pressure sensitive phase transition from gabbro to eclogite, why does the thickness of the crust vary only between 25 and 40 km worldwide? One explanation is that for isostatically compensated crust, erosion rates at the surface are highest for thick crust, thus acting to reduce the total variation in crustal thickness. Another is that lateral flow in ductile lower crust could work to decrease lateral thickness variations. However, the results of this study offer an alternative and perhaps complementary explanation. At temperatures of 800°C or less, all the gabbro and gabbronorite compositions are denser than the underlying mantle at pressures ~1.0–1.5 GPa (~30–45 km). Though thicker crust clearly can form in collisional belts, it would seem that on timescales of 100–1000 Myr, crustal thickness is likely determined by the density contrast between gabbro and peridotite. Further, the fact that average lower crust granulite xenoliths are the least dense composition we considered (Figures 2 and 3) suggests that density sorting may have affected the composition of the remaining lower crust.

During igneous crustal formation, in arcs, along volcanic rifted margins, and probably beneath oceanic plateaus, Moho temperatures could conceivably be very high, close to basalt liquidus temperatures of 1200–1300°C. Under these circumstances, several of the compositions we considered would be buoyant at pressures exceeding 1.5 GPa. The absence of thick crust in a number of arc settings, such as the Aleutian [Holbrook et al., 1999; Fledner and Klemperer, 1999], Izu-Bonin [Suemitsu et al., 1996], and Cascadia arcs [Parsons et al., 1998], and along ancient volcanic rifted margins suggest that, even in Mesozoic and Cenozoic settings, crustal thickness may often be determined by density instabilities. This, in turn, implies that (1) steady state Moho temperatures are less than 1200°C in most arcs and (2) crustal thickness along rifted margins may have been modified by convective instability during cooling.
6. Conclusions

Lower crust that ranges from mafic and ultramafic cumulates to primary liquid compositions can be significantly denser than the mantle, but the density contrast is a strong function of composition, temperature, and pressure. For basaltic and gabbroic compositions, garnet must be a stable mineral phase in order for the density to exceed that of the mantle. As a result, these compositions have densities that are pressure-dependent and kinetically unfavorable at high temperatures and H₂O-free conditions. In contrast, ultramafic cumulates are denser than the mantle mainly because of their composition and not phase transformations.

For a cold geotherm, relevant to cratonic settings, densities of all lower crustal compositions we considered, except that of “average” granulite, exceed that of the underlying mantle by 50–250 kg m⁻³ at minimum pressures <1.5 GPa. Arc garnetite and olivine clinopyroxenite are denser than the mantle at pressures as low as 0.25–0.8 GPa. For a hot geotherm, relevant to arc settings and regions of crustal extension, the density of the lower crust is much more variable and the maximum density contrast ranges from ~25–100 kg m⁻³. Two of the gabbro compositions we considered (high-P and low-P gabbro) are less dense than the mantle. Arc garnetite and basalt have densities that are very sensitive to the temperature of the geotherm in the range of 800-1000°C at 1 GPa. For an 800°C geotherm, these compositions are denser than the mantle over a pressure range that is similar to that for a cold geotherm. However, with a 1000°C geotherm these compositions have densities that are less than the mantle density at pressures down to the base of the conductive geotherm. Calculations of the time required for cooling of a 1000°C geotherm to an 800°C geotherm indicates that ~10 Myr is required for cooling, allowing the low density assemblage to convert to their high density form. Olivine clinopyroxenite has a density contrast that is nearly unchanged from the cold geotherm.

Calculation of the instability times for a dense, lower crustal layer to sink into the mantle show that high temperatures (>700°C, or >500°C with an initial background strain rate) are required for this process to occur in ~10 Myr. The high temperatures required for lower crustal convective instability of dense mafic-ultramafic compositions suggest that this process is restricted to arcs, volcanic rifted margins, and continental regions that are either undergoing extension, are underlain by a mantle plume or have had part of the conductive upper mantle removed. At arcs, lower crustal convective instability of olivine clinopyroxenites and arc garnetite compositions can occur readily. The high densities of arc garnetite compositions are sensitive to small changes in temperature. This suggests that these compositions may initially form a stable lower crust at high temperatures, to be followed by subsequent cooling and densification of garnetite present at >1.0 GPa, leading to convective instability at Moho temperatures that are hot enough to permit short instability times. In continental regions, gabbroic compositions have densities that are generally less than the mantle at temperatures hot enough for convective instability. However, if significant quantities of ultramafic cumulates are present, it is likely that these components of the lower crust will become convectively unstable on short timescales during rifting, convective thinning of the conductively heated upper mantle, or heating of thickened crust by radioactive decay.

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