Thermal history and origin of the Tanzanian Craton from Pb isotope thermochronology of feldspars from lower crustal xenoliths

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Common and radiogenic Pb isotopic compositions of plagioclase and anti-perthitic feldspars from granulite-facies lower crustal xenoliths from the Labait Volcano on the eastern margin of the Tanzanian Craton have been measured via laser ablation MC-ICP-MS. Common Pb in plagioclase and a single stage Pb evolution model indicate that the lower crust of the Tanzanian Craton was extracted from mantle having a 238U/204Pb of 8.1±0.3 and a 232Th/238U of 4.3±0.1 at 2.71±0.09 Ga (all uncertainties are 2σ). Since 2.4 Ga, some orthoclase domains within anti-perthites have evolved with a maximum 238U/204Pb of 6 and 232Th/238U of 4.3. The spread in Pb isotopic composition in the anti-perthitic feldspars yields single crystal Pb–Pb isochrons of ∼2.4 Ga, within uncertainty of U–Pb zircon ages from the same sample suite. The Pb isotopic heterogeneities imply that these granulites resided at temperatures <600 °C in the lower crust of the Tanzanian Craton from ca. 2.4 Ga to the present. In concert with the chemistry of surface samples, mantle xenoliths, and lower crustal xenoliths, our data imply that the cratonic lithosphere in Tanzania formed ca. 2.7 Ga, in a convergent margin setting, and has remained undisturbed since 2.7 Ga.

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1. Introduction

Insights into the origin of the lithosphere and its thermal history can be gained through the study of deep-seated xenoliths carried in basalts and kimberlites (e.g., Carlson et al., 2005; Rudnick, 1992; Rudnick et al., 1998; Schmitz and Bowring, 2003; and references therein). In particular, Pb isotope studies of lower crustal xenoliths can provide insights into the age, composition, cooling history and origin of the lower crust (e.g., Bolhar et al., 2007; Rudnick and Goldstein, 1990; Schmitz and Bowring, 2003).

The oldest pieces of the Earth’s crust, and therefore, any evidence for early Earth processes, are located in stable lithospheric blocks defined as cratons. Cratons formed in the Archean and have persisted through geologic time. Despite numerous geochemical and geophysical studies, however, debate continues regarding the origin of cratonic lithosphere and the reasons for its longevity. A defining characteristic of cratons is low surface heat flow (∼40 mW/m²) compared to crust that has been tectonically active since the Archean (∼50 mW/m²) (Nyblande and Pollack, 1993). Cratonic geotherms can be modeled as functions of surface heat flow, conductive heating from the mantle and the distribution of heat producing elements (K, Th, and U) through the lithosphere (e.g., Chapman and Pollack, 1977; Rudnick et al., 1998). A more direct approach to defining cratonic geotherms is through the application of experimentally calibrated geothermometers and geobarometers to the minerals of mantle xenoliths (e.g., Boyd, 1973; Rudnick and Nyblade, 1999). However, this approach fails to capture the time-dependent cooling of cratons and relies upon the assumption that the conditions recorded in the xenoliths reflect equilibration to present-day conditions (e.g., Michua et al., 2007; Rudnick et al., 1998).

Temporal constraints on the formation of cratonic geotherms, which correspond to lower crustal temperatures of 400–500 °C (Chapman and Pollack, 1977; Rudnick et al., 1998), can be determined through the use of radiogenic isotopic systems with different closure temperatures applied to minerals from lower crustal xenoliths (Schmitz and Bowring, 2003). For example, U–Pb thermochronology of accessory phases (e.g., apatite, monazite, zircon, rutile) from lower crustal xenoliths of the Kaapvaal Craton indicate a slow cooling rate of 1 °C/Ma for the lower crust of that craton followed by thermal perturbations in the Proterozoic and Mesozoic (Schmitz and Bowring, 2003).

In addition to insights into thermal history, the Pb isotopic composition of xenoliths can be used to determine the time-integrated U/Pb and Th/Pb composition of the lower crust. If the common Pb isotopic composition of a rock lies on a geochron with identical age as the sample suite, then the Pb isotopic composition and, therefore, the 238U/204Pb, 232Th/204Pb, and 238U/232Th of the mantle source from which the crust derives may also be inferred (e.g., Bolhar et al., 2007; Möller et al., 1998; Rudnick and Goldstein, 1990).

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2. Pb isotope systematics of feldspars

The Pb isotopic composition of a rock or mineral is composed of two components: common Pb (initial Pb inherited from the source rock) and radiogenic Pb (Pb*: Pb that is produced in the rock or mineral due to in situ radioactive decay of U and Th). Common Pb dominates in minerals having low $^{208}$Pb/$^{206}$Pb and $^{232}$Th/$^{208}$Pb and, thus, these minerals may record the Pb isotopic composition at the time of a rock’s last equilibration, assuming the mineral has subsequently remained closed to Pb diffusion (e.g., Oversby, 1975; Zartman and Wasserburg, 1969). Common Pb can, therefore, be used to determine provenance (e.g., comparison of thorogenic Pb and uranogenic Pb in common Pb minerals can be used to distinguish these minerals may record the Pb isotopic composition at the time of a rock’s last equilibration, and model the $^{238}$U/$^{204}$Pb, $^{232}$Th/$^{204}$Pb, and $^{235}$U/$^{204}$U of the previous reservoir in which the Pb resided (e.g., Holmes, 1946; Houtermans, 1946; Stacey and Kramers, 1975).

If the rock in which the common Pb minerals are present is a mantle derivative, then common Pb should reflect the $^{238}$U/$^{204}$Pb, $^{232}$Th/$^{204}$Pb, and $^{235}$U/$^{204}$U of the mantle from which the rocks were derived, so long as there is no evidence for crustal contamination, mixing, or re-equilibration between minerals after initial cooling (e.g., Bolhar et al., 2007; Kamber et al., 2003; Möller et al., 1998). Oceanic basalts derive from the mantle but, due to recycling of oceanic crust at subduction zones, do not record mantle composition before ~180 Ma. In contrast, many continental basalts are contaminated by continental lithosphere (e.g., Farmer, 2003), and references therein) and may not be faithful recorders of the composition of the sublithospheric mantle, particularly for Pb, which is typically in low abundance in basalts (Farmer, 2003). Determining the $^{238}$U/$^{204}$Pb, $^{232}$Th/$^{204}$Pb, and $^{235}$U/$^{204}$U ratios in older mantle derivatives is important for tracking the changes in the mantle that have occurred through time, such as the apparent decrease in the Th/U from ~4 to ~2 from the Archean to the Proterozoic (Farmer, 2003). Determining the $^{238}$U/$^{204}$Pb, $^{232}$Th/$^{204}$Pb, and $^{235}$U/$^{204}$U ratios in older mantle derivatives is important for tracking the changes in the mantle that have occurred through time, such as the apparent decrease in the Th/U from ~4 to ~2 from the Archean to the Proterozoic (Farmer, 2003).

To the east of the Tanzanian Craton lies the western granulite section of the Mozambique Belt (Fig. 1) and to the southeast lies the Eastern granulite basement; G.T.: Galena Terrane. Map created by M. Blondes.

3. Geologic setting

The Tanzanian Craton comprises Archean terrains that amalgamated ca. 2.6 Ga (Maboko, 2000; Manya et al., 2006) (Fig. 1). The exposed surface of the Tanzanian Craton is comprised of tonalites, trondhjemites, and granodiorites (TTGs), gneisses, and greenstone belts (Schulter, 1997), all of which have Archean Nd model ages and U–Pb zircon ages of ~3.0 Ga and 2.6 Ga, respectively (Maboko, 2000; Manya et al., 2006; Möller et al., 1998). Both geophysical and xenolith evidence indicate that the Tanzanian Craton has a stable, deep lithospheric keel (Chesley et al., 1999; Lee and Rudnick, 1999; Ritsema et al., 1999). The lithosphere of the Tanzanian Craton is characterized by low surface heat flow ($34 \pm 4$ mW/m$^2$) (Nyblade et al., 1990) and high seismic velocities in the mantle to a depth of 150 ± 20 km (Weerarante et al., 2003).

To the east of the Tanzanian Craton lies the western granulite section of the Mozambique Belt (Fig. 1) and to the southeast lies the...
Usagaran Belt. The Mozambique Belt formed in a Himalayan-scale orogeny (Fritz et al., 2009) and marks the suture between east and west Gondwana (Cutten et al., 2006; Fritz et al., 2009). The western granulites have Archean Nd model ages (Möller et al., 1998) and U–Pb zircon ages of 2.6–2.8 Ga (Johnson et al., 2003) and were metamorphosed to granulate facies in the Acadian, 50–100 Ma after their emplacement into a continental arc (Johnson et al., 2003). Subsequently, the western granulites were then reworked during the Pan-African Orogeny ca. 560 Ma (Blondes et al., 2009; Cutten et al., 2006). The southeastern Tanzanian Craton was the overriding plate of a subduction zone ca. 2.0 Ga resulting in the Usagaran mountain belt and suture (e.g., Collins et al., 2004; Möller et al., 1995). There is no evidence for a 2.0 Ga event in northern Tanzania (Blondes et al., 2009; Cutten et al., 2006; Fritz et al., 2009; Johnson et al., 2003; Mansur, 2008).

The East-African Rift (EAR) surrounds the Tanzanian Craton, forming the eastern (Gregory) and western (Lake Albert) branches, which first developed at ca. 30–50 Ma, with volcanism continuing today (Dawson, 1992; Nyblade and Brazier, 2002). The Labait volcano lies on the eastern edge of the Tanzanian Craton (4°34′37.6″S, 35°26′1.60″E), where the rift splays into a wide region of extension and volcanism. Labait erupted tuft and olivine melilitite carrying a wide variety of minerals, including orthopyroxene, clinopyroxene, quartz, biotite, muscovite, and orthoclase (Fritz et al., 2009). The equilibration conditions for these xenoliths are similar to those of the felsic granulites found in the western granulite belt (Johnson et al., 2003). One sample, LB04-87, is a granite, which is distinct in both mineralogy and texture (Table 1), and likely derived from the shallow crust. Importantly, no radiogenic Pb is observed in mineral separates of apatite (closure temperature ~450 °C for 100 µm crystals cooled at 2 °C/Ma, Chemnák et al., 1991) from the granulite-facies xenoliths, demonstrating their derivation from the present-day lower crust, where the temperature must have exceeded 450 °C (Blondes et al., 2009). U–Pb zircon ages and Nd model ages for these samples, 2.64±0.002 Ga and 2.9–3.8 Ga, respectively (Blondes et al., 2009; Mansur, 2008), mark their formation during the Acadian. These ages coincide with the timing of greenstone belt volcanism in the northern part of the Tanzanian Craton (Manya et al., 2006) and overlap the 3.0–3.1 Ga Nd model ages from surface samples from the Tanzanian Craton (Möller et al., 1998).

Feldspars from the Labait xenoliths consist of both plagioclase and ortho-pyrite. The plagioclase composition is An32.55Ab17.54Or39.52 while the orthoclase, in anti-perthite, has a composition of An3.5Ab2.5Or75.91 (Mansur, 2008). One sample, LB04-27, is an anorthosite having plagioclase with a bytownite composition (An87) (Mansur, 2008).

Table 1: Samples and major mineralogy.

<table>
<thead>
<tr>
<th>Sample #</th>
<th>Rock type</th>
<th>Major mineralogy</th>
</tr>
</thead>
<tbody>
<tr>
<td>LB04-09</td>
<td>2 Px gran</td>
<td>Plagioclase orthopyroxene, clinopyroxene</td>
</tr>
<tr>
<td>LB04-19</td>
<td>2 Px gran</td>
<td>Anti-perthite, orthopyroxene, clinopyroxene</td>
</tr>
<tr>
<td>LB04-38</td>
<td>2 Px gran</td>
<td>Plagioclase, orthopyroxene, clinopyroxene</td>
</tr>
<tr>
<td>LB04-43</td>
<td>2 Px gran</td>
<td>Plagioclase, orthopyroxene, clinopyroxene</td>
</tr>
<tr>
<td>LB04-50</td>
<td>2 Px gran</td>
<td>Anti-perthite, orthopyroxene, garnet</td>
</tr>
<tr>
<td>LB04-65</td>
<td>2 Px gran</td>
<td>Anti-perthite, orthopyroxene, clinopyroxene</td>
</tr>
<tr>
<td>LB04-67</td>
<td>Granite</td>
<td>Anti-perthite, orthoclase, quartz, biotite, muscovite</td>
</tr>
<tr>
<td>LB04-48</td>
<td>2 Px Hbl gran</td>
<td>Plagioclase, clinopyroxene, orthopyroxene, hornblende</td>
</tr>
<tr>
<td>LB04-27</td>
<td>Anorthosite</td>
<td>Plagioclase, garnet, orthopyroxene</td>
</tr>
<tr>
<td>LB04-39</td>
<td>Gt Opx gran</td>
<td>Plagioclase, orthopyroxene, garnet</td>
</tr>
<tr>
<td>LB04-91</td>
<td>Gt Opx gran</td>
<td>Anti-perthite, orthopyroxene, garnet</td>
</tr>
</tbody>
</table>


4. Samples

Petrographic, mineral chemistry, whole rock major and trace element geochemistry, Rb–Sr and Sm–Nd isotopic data, as well as some U–Pb zircon ages for the Labait granulate xenoliths are presented in Mansur (2008). U–Pb analyses of zircon and accessory phases from these samples are presented in Blondes et al. (2009). The xenoliths for this study fall into five different groups, based on major mineralogy and trace element chemistry: two pyroxene granulites (n = 7), garnet orthopyroxene granulites (n = 2), two pyroxene hornblende granulites (n = 2), and anorthosite (n = 1), all of which derive from the lower crust (Mansur, 2008). In addition, we studied a xenolithic twomica granite (n = 1). Plagioclase and/or anti-perthite (orthoclase exsolved from albite) from eleven granulate-facies xenoliths, spanning all varieties, were analyzed here, as well as anti-perthite from the granite xenolith. Table 1 lists the major mineralogy of these samples.

The Labait xenoliths, like most other granulate-facies lower crustal xenoliths, have mafic compositions with ≤54 wt.% SiO2 (Mansur, 2008). They are enriched in light rare-earth elements (REE), have negative Ti and Nb anomalies, and have high La/Nb ratios (between 1.3 and 7.1), all of which suggest formation as a crystallized basaltic melt in an arc setting (Mansur, 2008). Additionally, these samples show striking depletions in Rb, Cs, Th and U, which has been interpreted to have occurred during granulate facies metamorphism (Mansur, 2008). Based on two-pyroxyene Fe–Mg exchange thermobarometry, the Labait xenoliths record a broad range of equilibration temperatures from 480° to 850 °C, with most <670 °C (assuming a pressure of equilibration of 1 GPa or 33 km depth) (Mansur, 2008). The equilibration conditions for these xenoliths are similar to those of the felsic granulites found in the western granulite belt (Johnson et al., 2003). One sample, LB04-87, is a granite, which is distinct in both mineralogy and texture (Table 1), and likely derived from the shallow crust.

Analytical methods

5.1. Pb isotopic measurements

Samples were prepared either as polished rock slabs mounted in epoxy or 50 µm polished thick sections. The Pb isotopic compositions were determined in situ via LA–MC-ICP-MS using a frequency-quintupled solid-state Nd:YAG laser system (213 nm wavelength) and a Nu Plasma MC-ICP-MS (following Kent, 2008a,b; Paul et al., 2005). Full analytical details are provided in the Supplementary materials and a brief overview is given below.

Ion beams of 204Hg, 206Hg, and 208Pb/Hg were monitored using parallel ion counters, whereas isotopes 203Hg, 206Pb, 207Pb, and 208Pb were detected simultaneously as ion currents in parallel Faraday cups equipped with 1013-Ω resistors. In order to correct for the isobaric interference of 204Hg on 206Pb, 207Pb, and 208Pb, was monitored to allow subtraction of a fractionation-corrected abundance of 204Hg. The Hg fractionation factor was calculated using the exponential law and assuming natural isotopic abundances (de Laeter et al., 2003). The Hg fractionation factor was then applied to the 204Hg:206Hg ratio and subsequently the 206Hg signal was stripped from the 204Pb signal. The contribution of 204Hg on the 204Pb signal was typically <0.7%. Backgrounds were measured on-peak for 60 s prior to ablation with the laser on and shuttered. The average of each signal’s background was subtracted in real time from each individual measurement collected at intervals of 0.2 s using the Nu Plasma time resolved software. For accurate isotopic measurements, two corrections must be applied to the background-corrected data: 1) mass fractionation corrections and 2) ion counter-Faraday cup gain corrections. Mass
fractionation effects are corrected using standard sample bracketing using SRM NIST 612 as the reference standard and calculating the fractionation factor of \(208\text{Pb}/206\text{Pb}\) using the exponential fractionation law. Using the fractionation factor from \(208\text{Pb}/206\text{Pb}\), an expected \(206\text{Pb}/204\text{Pb}\) (\(x = 8, 7,\) or 6) can be calculated. Discrepancies between the Faraday cup and ion counter gains can be corrected by using the ratio between the measured and calculated \(208\text{Pb}/206\text{Pb}\) values of SRM 612 (Baker et al., 2004).

While the analytical precision using this method is largely dependent on the Pb concentration in the sample, which can range between 10 and 20 \(\mu\text{g/g}\), repeat measurements (\(n = 85\)) over a several month period of the basaltic glass standard BCR 2-g, which contains \(\sim 10 \, \mu\text{g/g}\) Pb, yielded accuracies of \(\leq 0.1\%\) for \(208\text{Pb}/204\text{Pb}\) and an external reproducibility of \(\leq 0.5\%\) (2\(\sigma\)) for \(208\text{Pb}/204\text{Pb}\). A detailed discussion of the accuracy and precision of this method is given in the Supplementary materials.

5.2. U–Pb

The U–Pb ratio in a feldspar from one sample (gt-opx granulite LB04-91) was measured by LA–ICP-MS utilizing the same laser ablation system described above coupled to a ThermoFinnigan Element 2 ICP-MS. Ten spots, 40 \(\mu\text{m}\) in diameter, were analyzed along a Pb-isotope laser ablation track (Fig. 5). Further details of the analytical conditions and results are given in the Supplementary materials.

5.3. Solution analyses

The Pb isotopic composition of the host basalt was analyzed by both solution MC-ICP-MS and LA–MC-ICP-MS. The solution measurements were performed using parallel Faraday cups. Pb was purified from dissolved rock using standard HBr anion chromatography. Mass fractionation was accounted for using the exponential fractionation law through the addition of Tl (e.g., Baker et al., 2004). The lab blank from dissolved rock using standard HBr anion chromatography. Mass fractionation effects are corrected using standard sample bracketing using SRM NIST 612 as the reference standard and calculating the fractionation factor of \(208\text{Pb}/206\text{Pb}\) using the exponential fractionation law. Using the fractionation factor from \(208\text{Pb}/206\text{Pb}\), an expected \(206\text{Pb}/204\text{Pb}\) (\(x = 8, 7,\) or 6) can be calculated. Discrepancies between the Faraday cup and ion counter gains can be corrected by using the ratio between the measured and calculated \(208\text{Pb}/206\text{Pb}\) values of SRM 612 (Baker et al., 2004).

While the analytical precision using this method is largely dependent on the Pb concentration in the sample, which can range between 10 and 20 \(\mu\text{g/g}\), repeat measurements (\(n = 85\)) over a several month period of the basaltic glass standard BCR 2-g, which contains \(\sim 10 \, \mu\text{g/g}\) Pb, yielded accuracies of \(\leq 0.1\%\) for \(208\text{Pb}/204\text{Pb}\) and an external reproducibility of \(\leq 0.5\%\) (2\(\sigma\)) for \(208\text{Pb}/204\text{Pb}\). A detailed discussion of the accuracy and precision of this method is given in the Supplementary materials.

6. Results

The Labait host melilitite has a Pb isotopic composition similar to that of other rift basalts from Northern Tanzania (e.g., Paslick et al., 1995); see the Supplementary materials for comparison.

The Pb isotopic composition of plagioclase and anti-perthite from the Labait lower crustal xenoliths (\(n = 137\)) are presented in Figure 2 (data tables are provided in the Supplemental materials). Anti-perthites in six samples (four two-pyroxene granulites LB04-19, LB04-50, LB04-82, one gt-opx granulite, LB04-91 and the granite, LB04-87) display a significant spread in Pb isotopic compositions. Plagioclase in a further six samples (which include specimens from all four groups of granulites) show significant spread in their Pb isotopic compositions. Plagioclase in six samples (which include specimens from all four groups of granulites) show significant spread in their Pb isotopic compositions (all analyses are within analytical uncertainty (\(\leq 0.2\%\)) of the LA–MC-ICP-MS. This uncertainty is similar to that obtained for measurements of standard materials. Data are provided in the Supplementary materials.

7. Discussion

7.1. Origin of the linear trends

Linear trends in isotopic data can either reflect mixing between two isotopically distinct components or result from the in-growth of the radiogenic daughter product. Only the latter has any age significance.
One possible source of radiogenic Pb that may have been mixed into the xenoliths is the host lava. Figure 4 shows a mixing trajectory between the least radiogenic plagioclases (~20 μg/g Pb, from samples LB04-48 and LB04-38) and the Labait melilitite (~8 μg/g). Many of the feldspar analyses plot beyond analytical uncertainty of this mixing line, although some overlap it. To achieve this range in Pb isotopic compositions via host basalt mixing, up to ~80% of the Pb in the most radiogenic samples must have originated from the basalt. This level of mixing is untenable on two accounts. First, the whole-rock Nd isotopic compositions and trace element compositions of the samples do not record any such mixing (Mansur, 2008). Second, in order for Pb from the basalt to enter the feldspars, the samples would need to be heated above 600 °C (Tc for Pb in feldspar) during the xenolith’s entrainment within the basalt. Such heating is precluded on the basis of U–Pb thermochronology of titanite (Tc > 550 °C) and apatite (Tc > 450 °C) from lower crustal xenoliths hosted in a similar alkali basalt from a near-by locality (Lashaine, Blondes et al., 2009). Apatite in these samples was open to Pb diffusion at the time of eruption, like those in the Labait xenoliths, but titanite records concordant Paleozoic 206Pb/238U ages, reflecting slow cooling following the Pan-African Orogeny. Collectively, these results indicate that the samples remained above apatite closure temperature but below titanite (and feldspar) closure temperature for hundreds of millions of years—more
than two orders of magnitude longer than the duration of rift volcanism in northern Tanzania. Similar results were obtained for kimberlite-hosted lower crustal xenoliths from the Kaapvaal Craton (Schmitz and Bowring, 2003). We conclude that mixing with the host basalt did not generate the linear correlation seen in the plot of \( \frac{206\text{Pb}}{204\text{Pb}} \) vs. \( \frac{207\text{Pb}}{204\text{Pb}} \).

Because the U/Pb ratio of anti-perthite from gt-opx granulite LB04-91 shows over an order of magnitude change over a distance of \( \sim 800 \mu m \), and correlates with the \( \frac{206\text{Pb}}{204\text{Pb}} \) ratio (Fig. 5), we suggest that the linear correlation reflects an isochron, rather than mixing between two ancient Pb components. Below we explore the thermochronological implications of the feldspar isochrons.

### 7.2. Thermochronometry

Individual feldspars in five samples (2 px granulites LB04-19, LB04-50, LB04-82, gt-opx granulite LB04-91 and the granite, LB04-87) are heterogeneous with respect to their Pb isotopic compositions, indicating that they have resided at temperatures below \( \sim 600 °C \), given the closure temperature of Pb in feldspar calculated above. The Pb isotopic heterogeneities reflect several reservoirs of radiogenic Pb, similar to other U-bearing silicate minerals (Frei and Kamber, 1995) and yield isochron ages of \( \sim 2.4 \) Ga. This result demonstrates that the Pb–Pb system in feldspar is a viable thermochronometer in rocks having coarse-grained (mm) feldspars, significant age, and U and Th concentrations. Because feldspars from the Labait xenoliths record a range of Pb isotopic compositions, the temperature of the lower crust of the Tanzanian Craton must not have been elevated above 600 °C at any time following 2.4 Ga, including during the Pan-African Orogeny (ca. 560 Ma) and the development of the East African Rift.

Given the spread in Pb isotopic values and the isochron age, a maximum \( \frac{238\text{U}}{204\text{Pb}} \) can be calculated for the feldspar grains. The

### Table 2
Summary of Pb–Pb ages from Labait xenoliths.

<table>
<thead>
<tr>
<th>Sample suite</th>
<th>Sample(s)</th>
<th>Minerals</th>
<th>Age</th>
<th>2σ</th>
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</thead>
<tbody>
<tr>
<td>2Px granulite</td>
<td>LB04-19</td>
<td>Multiple plagioclase</td>
<td>2.96</td>
<td>1.50</td>
</tr>
<tr>
<td>2Px granulite</td>
<td>LB04-43</td>
<td>Multiple plagioclase</td>
<td>3.54</td>
<td>0.50</td>
</tr>
<tr>
<td>2Px granulite</td>
<td>LB04-50</td>
<td>Single anti-perthite</td>
<td>2.74</td>
<td>0.43</td>
</tr>
<tr>
<td>2Px granulite</td>
<td>LB04-50</td>
<td>Single anti-perthite</td>
<td>2.30</td>
<td>0.60</td>
</tr>
<tr>
<td>2Px granulite</td>
<td>LB04-50</td>
<td>Anti-perthite, ilmenite</td>
<td>2.51</td>
<td>0.35</td>
</tr>
<tr>
<td>2Px granulite</td>
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<td>Anti-perthite, apatite</td>
<td>2.42</td>
<td>0.00</td>
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<td>2Px granulite</td>
<td>LB04-65</td>
<td>Anti-perthite, apatite</td>
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<td>LB04-65</td>
<td>Multiple anti-perthite, multiple apatite</td>
<td>2.32</td>
<td>0.68</td>
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<tr>
<td>2Px granulite</td>
<td>LB04-09, LB04-19, LB04-38, LB04-43, LB04-50, LB04-65</td>
<td></td>
<td>2.71</td>
<td>0.24</td>
</tr>
</tbody>
</table>

| Gt-Opx-granulate      | LB04-91                     | Single anti-perthite              | 2.56  | 0.24 |
| Gt-Opx-granulate      | LB04-91                     | Single anti-perthite              | 2.15  | 0.08 |
| Gt-Opx-granulate      | LB04-91                     | Multiple anti-perthite            | 2.42  | 0.26 |
| 2Px Hbl gran          | LB04-82                     | Single anti-perthite              | 2.48  | 0.59 |
| 2Px Hbl gran          | LB04-82                     | Multiple anti-perthite, pyroxene  | 2.61  | 0.76 |
| Granite               | LB04-87                     | Anti-perthite–apatite inclusion   | 2.68  | 0.15 |
| Granite               | LB04-87                     | Anti-perthite–apatite inclusion   | 2.61  | 0.05 |
| Granite               | LB04-87                     | Multiple anti-perthite, apatite   | 2.62  | 0.25 |

Average age of anti-perthite: 2.44 ± 0.41
Average of non anti-perthite ages: 2.71 ± 0.73
Isochron of the entire sample suite: 2.71 ± 0.09

![Fig. 6. Isochrons calculated for each sample suite, two pyroxene granulites, gt-opx granulites, granite, and Hbl granulites.](image-url)
largest range in Pb isotopic values for a single feldspar is seen in
gt-opx granulite LB04-91, which has $^{206}\text{Pb}^{204}\text{Pb}$ varying from 14.87
to 15.38, corresponding to a maximum $^{238}\text{U}^{204}\text{Pb}$ of 6. Most of the
samples fall on a linear trend on the $^{206}\text{Pb}^{204}\text{Pb}$ vs. $^{206}\text{Pb}^{204}\text{Pb}$
diagram, corresponding to a $^{232}\text{Th}^{238}\text{U}$ of 4.3. Three samples, one
from each of three different lithological groups, fall off this trend: 
gt-opx granulite LB04-39, the single anorthosite, LB04-27 and the
granite LB04-87. Assuming these samples were extracted at the same
time and from the same mantle source as the other Labait xenoliths
(because they follow the same trend as the other samples in the
$^{207}\text{Pb}^{204}\text{Pb}$ vs. $^{208}\text{Pb}^{204}\text{Pb}$ plot), their Pb isotopic composition can be
modeled with a $^{232}\text{Th}^{238}\text{U}$ of 2, 3.6 and 2, respectively (Fig. 2). All of
the calculated $^{232}\text{Th}^{238}\text{U}$ ratios are within the range of the whole rock
Th/U (0.7–5.6, Mansur, 2008). The variable $^{232}\text{Th}^{238}\text{U}$ ratios in the
feldspars of these three samples illustrate the potential of the
$^{208}\text{Pb}^{204}\text{Pb}$ vs. $^{206}\text{Pb}^{204}\text{Pb}$ space (except hornblende granulite LB04-48), we
interpret them to have formed from the same mantle reservoir at the
same time. Therefore, we group all of the Pb isotopic compositions from
this suite of xenoliths to get an age of lower crust extraction from the
mantle of 2.71 ± 0.09 Ga (Table 2). This mantle extraction age is similar
to that of the western granulites (Johnson et al., 2003). In addition,
the age recorded by the anti-perthitic feldspars (~2.4 Ga) could date the age
of metamorphism of the lower crust to granite facies, which is similar
to the metamorphic event seen by the western granulites, occurring
50–100 Ma after extraction from the mantle.

The Pb isotopic compositions of the all of the feldspars from the
Labait lower crustal xenoliths, except those in hornblende granulite
LB04-48, lie within error of a 2.71 Ga isochron (Fig. 2). The 2.71 Ga
isochron intersects the 2.71 geochron (Holmes, 1946) at
$^{206}\text{Pb}^{204}\text{Pb} = 13.4 ± 0.4$ and $^{208}\text{Pb}^{204}\text{Pb} = 14.7 ± 0.4$ (2σ), near the
common Pb isotopic composition of feldspars in the two least-
radiogenic samples, LB04-38 and LB04-48. Using a single-stage Pb
evolution model, the primordial Pb isotopic composition of Canyon
Diablo troilite (Chen and Wasserburg, 1983), and assuming the
intersection between the Labait isochron and geochron represents the
isotopic composition of the Tanzanian Craton crust at the time of its
extraction from the mantle, this crust derived from a mantle source with
a $^{238}\text{U}^{204}\text{Pb}$ of 8.1 ± 0.3 (Fig. 2). Similarly, the $^{238}\text{U}^{204}\text{Pb}$ value for the
mantle source was 35 ± 1, with a $^{232}\text{Th}^{238}\text{U}$ of 4.3 ± 0.1 (Fig. 2). This
$^{238}\text{U}^{204}\text{Pb}$ is similar to that found for Archean komatites from Canada,
Australia, and Africa that indicate mantle $^{238}\text{U}^{204}\text{Pb}$ values at ~2.7 Ga
of 7.8–8.5 (Chauvel et al., 1993 and Dupré and Arndt, 1990). The inferred
$^{232}\text{Th}^{238}\text{U}$ is consistent with previous observations that the Archean
mantle had a $^{232}\text{Th}^{238}\text{U}$ of ~4 (e.g., Zartman and Haines, 1988; Zartman
and Richardson, 2005, and references therein).

Two samples do not follow the general trends defined by the rest
of the suite. The anorthosite (LB04-27) has unusually radiogenic
plagioclase, which lies close to the present-day geochron. The
anorthosite does not fall on the mixing trend with the Labait basalt,
suggesting that it may have derived from a deeper (hotter) level than
the other samples and may have equilibrated with U-bearing phases
more recently. In contrast, plagioclase in hornblende granulite LB04-
48 has the least radiogenic Pb isotopic composition of the entire suite
and lies below the linear trend defined by the other samples in both
$^{208}\text{Pb}^{204}\text{Pb}$ vs. $^{206}\text{Pb}^{204}\text{Pb}$ and $^{207}\text{Pb}^{204}\text{Pb}$ vs. $^{206}\text{Pb}^{204}\text{Pb}$ plots
(Fig. 2). The non-radiogenic composition of the common Pb in this
sample could reflect its derivation from somewhat older material
present in the cratonic lower crust.

The common Pb isotopic composition of the Tanzanian lower crust
overlaps with the Pb isotopic composition of feldspars and galenas from
outcrops within the Tanzanian Craton, in a similar, slightly curved array
(Fig. 7) (Coomer and Robertson, 1974; Möller et al., 1998). The similarity
between Pb isotopic compositions of surface samples and lower crustal
xenoliths indicates a similar provenance, and therefore, similar model
ages for the upper and lower crust. The linear relationship of the Pb
isotopic composition of lower crustal feldspars and surface samples in
both $^{208}\text{Pb}^{204}\text{Pb}$ vs. $^{206}\text{Pb}^{204}\text{Pb}$ and $^{207}\text{Pb}^{204}\text{Pb}$ vs. $^{206}\text{Pb}^{204}\text{Pb}$
plots indicates that there have been no additions or major fractionation
of U/Pb or Th/Pb in the crust since the amalgamation of the Tanzanian
Craton in a convergent margin setting ca. 2.7 Ga.

8. Conclusions

The Pb isotopic composition of feldspars from lower crustal
xenoliths from the Tanzanian Craton, in combination with previous
studies on surface samples and mantle xenoliths indicate that the
crust and lithospheric mantle comprising the Tanzanian Craton
formed in the Archean in a convergent margin setting at approxi-
mately the same time, ca. ~2.7 Ga. Using a single-stage Pb evolution
model, the crust of the Tanzanian Craton was extracted from a mantle
with a $^{238}\text{U}^{204}\text{Pb}$ of 8.1 ± 0.3 and $^{232}\text{Th}^{238}\text{U}$ of 4.3 ± 0.1. Single anti-
perthitic feldspars have isotopes that have an average age of 2.4 Ga,
which indicates the time at which the lower crust cooled to a

![Fig. 7. Pb isotopic data for feldspars in lower crustal xenoliths compared to surface feldspars (Möller et al., 1998) and galenas/pyrites (Coomer and Robertson, 1974) from the Tanzanian Craton. The granitic sample (LB04-87) is shown in open symbols.](image-url)
temperature < 600 °C. After 2.4 Ga, the temperature of the Tanzanian Craton lower crust, as represented by the xenoliths examined here, did not rise above 600 °C during either the Pan-African Orogeny or the present-day rifting. The identical common Pb isotopic compositions of the surface samples and lower crust indicate that there was no addition, subtraction, or fractionation of U, Pb, or Th in the crust of the Tanzanian Craton since its formation ca. 2.7 Ga.

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Appendix A. Supplementary data

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References


