ABSTRACT

Newly discovered xenoliths within Pliocene and Quaternary intermediate volcanic rocks from southern Peru permit examination of lithospheric processes by which thick crust (60–70 km) and high average elevations (3–4 km) resulted within the Altiplano, the second most extensive orogenic plateau on Earth. The most common petrographic groups of xenoliths studied here are igneous or meta-igneous rocks with radiogenic isotopic ratios consistent with recent derivation from asthenospheric mantle (87Sr/86Sr = 0.704–0.709, 143Nd/144Nd = 0.5126–0.5129). A second group, consisting of felsic granulite xenoliths exhibiting more radiogenic compositions (87Sr/86Sr = 0.711–0.782, 143Nd/144Nd = 0.5121–0.5126), is interpreted as supra-crustal rocks that underwent metamorphism at ~9 kbar (~30–35 km paleodepth, assuming a mean crustal density of 2.8 g/cm³) and ~750 °C. These rocks are correlated with nonmetamorphosed rocks of the Mitu Group and assigned a Mesozoic (Upper Triassic or younger) age based on detrital zircon U-Pb ages. A felsic granulite Sm-Nd garnet whole-rock isochron of 42 ± 2 Ma demonstrates that these rocks were entrained as xenoliths in volcanic host magmas and transported toward the surface. Mafic granulites and peridotites from the same xenolith suite comprise the basement of the metasedimentary sequence, exhibiting isotopic characteristics of Central Andean crust. Calculated equilibrium pressures for these basement rocks are >11 kbar, suggesting that the basement-cover interface lies beneath the northernmost Altiplano at ~30–40 km below the surface. Together, these results indicate that crustal thickening under the northernmost Altiplano started earlier than major latest Oligocene and Miocene uplift episodes affecting the region and was coeval with a flat slab–related regional episode of deformation. Total shortening must have been at least 20% more than previous estimates in order to satisfy the basement to cover depth constraints provided by the xenolith data. Sedimentary rocks at >30 km paleodepth require that Andean basement thrusts decapitated earlier Triassic normal faults, trapping Paleozoic and Mesozoic rocks below the main décollement. Magma loading from intense Cenozoic plutonism within the plateau probably played an additional role in transporting Mesozoic cover rocks to >30 km and thickening the crust beneath the northern Altiplano.

INTRODUCTION

The formation of large orogenic plateaus remains a central topic of research in continental tectonics. High-temperature (igneous and metamorphic) petrology plays a large role in answering critical questions such as: (1) When and how did plateaus achieve their elevation (Saylor and Horton, 2014)? (2) How much compression and partial melting ensued as these rocks were entrained as xenoliths in volcanic host magmas and transported toward the surface? Mafic granulites and peridotites from the same xenolith suite comprise the basement of the metasedimentary sequence, exhibiting isotopic characteristics of Central Andean crust. Calculated equilibrium pressures for these basement rocks are >11 kbar, suggesting that the basement-cover interface lies beneath the northernmost Altiplano at ~30–40 km below the surface. Together, these results indicate that crustal thickening under the northernmost Altiplano started earlier than major latest Oligocene and Miocene uplift episodes affecting the region and was coeval with a flat slab–related regional episode of deformation. Total shortening must have been at least 20% more than previous estimates in order to satisfy the basement to cover depth constraints provided by the xenolith data. Sedimentary rocks at >30 km paleodepth require that Andean basement thrusts decapitated earlier Triassic normal faults, trapping Paleozoic and Mesozoic rocks below the main décollement. Magma loading from intense Cenozoic plutonism within the plateau probably played an additional role in transporting Mesozoic cover rocks to >30 km and thickening the crust beneath the northern Altiplano.

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The Altiplano-Puna Plateau in the Central Andes, one of the world’s largest modern plateaus, has been magmatically active since its birth to present (Isacks, 1988; Allmendinger et al., 1997). Magmatic rocks associated with the Altiplano-Puna resemble compositions found in subduction-related arcs, but they are not confined to linear “arcs”; instead, they are distributed across the width of the plateau. Magma compositions range from basaltic to
ryholitic (e.g., de Silva, 1989; Kay and Kay, 1993; Kay et al., 1994; Francis and Hawkesworth, 1994; Carlier et al., 2005; Mamani et al., 2008, 2010; Drew et al., 2009), and the volume of magmatic material ranges from <0.01 km³, for example, in the volcanic centers investigated here, to ~300 km³ at Cerro Galan, one of the largest felsic calderas on the planet (Francis et al., 1983). The plateau also hosts the Altiplano-Puna magma body, one of the largest (>10,000 km³ in volume) zones of active partial melting in a continental setting (Chmielowski et al., 1999; Babeyko et al., 2002; Zandt et al., 2003; Schilling et al., 2006).

Lower-crustal and deeper xenoliths are extraordinarily rare in volcanic rocks from the Altiplano-Puna, perhaps due to the thick crust beneath the region (locally >70 km; e.g., Beck and Zandt, 2002). To date, mantle peridotites have not been described from the plateau region, despite the presence of numerous mafic rocks within various plateau volcanic fields (e.g., Kay et al., 1994; Duca et al., 2013). In this study, we provide new major- and trace-element chemistry, Sr, Nd, and Pb whole-rock isotopic ratios, U-Pb zircon and monazite geochronology, and Sm-Nd garnet geochronology from six newly discovered xenolith localities from the northern part of the Altiplano in southern Peru. These localities feature crustal and mantle xenoliths hosted in small-volume Phiocene and Quaternary back-arc volcanic centers (Carlier and Lorand, 1997; Carlier et al., 2005). The purpose of this investigation is: (1) to characterize the rock types and estimate their distribution within the sub-Altiplano lithospheric column; (2) to compare Peruvian xenoliths with recently described crustal xenoliths from the Bolivian Altiplano (Davidson and de Silva, 1995; McLeod et al., 2012, 2013); (3) to place geochemical and geochronologic constraints on how the crustal structure beneath the northern Altiplano has evolved since the onset of thickening; and (4) to evaluate the relative importance of magmatic additions versus tectonic shortening in thickening the crust beneath the northern Altiplano.

BACKGROUND
Regional Geology
Southern Peru (13°S–16°S latitude) is located above a subduction zone in the Central Andes and is composed of, from west to east, a deep trench, a forearc, a frontal magmatic arc (the Western Cordillera), a high, internally drained plateau (the Altiplano), and a foreland fold-and-thrust belt (the Eastern Cordillera; Isacks, 1988; Allmendinger et al., 1997). Crustal thickness increases from 30 km in the forearc to over 70 km under the Cordilleras and the Altiplano, and decreases eastward beneath the sub-Andean foreland and the craton to ~35–40 km (James and Sacks, 1999; Beck and Zandt, 2002; Phillips and Clayton, 2014). Average elevations are above 4 km in the Western Cordillera and the Altiplano.

The Central Andes have resided in the upper plate of an active convergent margin for much of the time following the breakup of Rodinia (Ramos, 2008, and references therein). Over the course of this protracted convergent history, the locus of calc-alkaline arc magmatism has shifted several times due to either changes in slab dip and/or migrations of the subduction trench relative to the upper plate (e.g., Mamani et al., 2010). Arc magmas were emplaced into a framework comprising: (1) ca. 1.0–1.5 Ga crust intruded by younger Paleozoic (0.4–0.55 Ga) arcs along the southwestern edge of the Amazon craton (Tassinari et al., 2000; Loewy et al., 2004), (2) the Paleoproterozoic to early Paleozoic Arequipa terrane within the modern forearc (Wasteney et al., 1995; Loewy et al., 2004; Ramos, 2008; Casquet et al., 2010), and (3) ~5–7 km of Paleozoic low-grade metamorphic rocks and another ~5 km of Mesozoic and Tertiary cover rocks (Gotberg et al., 2010; Reimann et al., 2010; Bahlgur et al., 2011; Reitsma, 2012). The boundary between the Amazonia and Arequipa domains is obscured by Paleozoic and younger cover rocks and various magmatic arcs. These rocks include Carboniferous to Early Jurassic “Gondwanide” granitoids and related Triassic rifting basins of the Mitu Group, both produced during the assembly and breakup of western Gondwana, (Sempere et al., 2002; Ramos, 2008; Mišković et al., 2009; Reitsma, 2012; Perez and Horton, 2014).

Subduction-related magmatism started regionally during the Carboniferous and has been more or less continuous until today (Sandeman et al., 1995; Mamani et al., 2010). Several arc domains are recognized, based on composition, age, and location, in southern Peru (Fig. 1): the Chocolate arc (301–90 Ma), the Toquepala arc (90–45 Ma), the Anta arc (45–30 Ma), the Tacaca arc (30–24 Ma), the Huayllillas arc (24–10 Ma), the lower Barroso arc (10–3 Ma), the upper Barroso arc (3–1 Ma), and the active volcanic arc (<1 Ma). Of these, the Chocolate arc consists of submarine volcanics and volcanioclastic deposits, whereas all subsequent arcs starting with the Toquepala arc at ca. 90 Ma were formed in a subaerial environment (Mamani et al., 2008, 2010). The Chocolate arc consists primarily of basaltic rocks, with a thickness of 1 to >3 km, that overlie Paleozoic passive-margin (meta)sedimentary rocks (Pino et al., 2004). Calc-alkaline, dominantly mafic magmatism continued well into the Cretaceous over a >100 km width of the margin throughout the Central Andes, suggesting that these areas represent frontal arc, in contrast to back-arc, magmatic products within an attenuated upper plate (Oliveros et al., 2006; Charrier et al., 2007; Rossell et al., 2013). The Late Cretaceous marks a significant change in the character of magmatism, which switched from extension-related and mafic-dominated volcanic and intrusive rocks to a more typical Cordilleran intermediate and silicic arc (Mukasa, 1986). This also corresponds to a well-known shift in the sedimentary record from marine to continental deposition (Sempere et al., 2002). Together, these observations suggest that the upper plate (South America) changed from an extensional to compressional stress regime, marking the onset of crustal thickening at ca. 90 Ma.

Volcanic rocks provide additional support for significant Cenozoic crustal thickening in the Central Andes. Geochemical proxies for crustal thickness (e.g., Sr/Y, La/Yb, La/Sr, Sm/Yb, and Dy/Yb) increase with time in these rocks, particularly between 30 Ma and today (Mamani et al., 2010). The pattern of crustal thickening is complicated, however, as these arcs migrated during this time period. The Toquepala arc, the first continental arc sensu stricto, is manifested in the geologic record by the emplacement of the Coastal Peruvian batholith and related ignimbrite activity (Pitcher et al., 1985; Mukasa, 1986; Clark et al., 1990;Martínez and Cervantes, 2003). The Toquepala arc activity waned at ca. 50 Ma (Mamani et al., 2010), and the arc migrated inboard by >150 km. The 45–30 Ma Anta arc consists of the Andahuaylas-Yauri batholith and the Anta volcanic rocks. These rocks are interpreted as products of an inboard magmatic sweep, probably related to shallowing of the subducting slab (James and Sacks, 1999). Initiation of the Anta arc also corresponds to the Incaic shortening event that began at 45 Ma (Noble et al., 1979; Sandeman et al., 1997; Roperch et al., 2006) and is thought to represent an episode of flat slab–related crustal thickening. However, substantial crustal thickening did not begin until the mid-Oligocene (ca. 30 Ma) and was accompanied by formation of the Tacaca, Huayllillas, and lower Barroso and upper Barroso arcs and stratovolcanoes of the modern arc front. The volcanic products of these arcs record a progressive thickening of the crust regionally (Mamani et al., 2010) and are overall more silicic than older arc rocks in the area.

While most of the Quaternary arc products are located in the frontal arc and are defined by a linear array of stratovolcanoes (de Silva, 1989),
small-volume potassic volcanic rocks crop out in the back arc within the northernmost Altiplano and appear on regional geologic maps as the Quaternary Rumicolca formation (Carlier et al., 1997, 2005; Carlotto et al., 2011). They are high-K calc-alkaline lava flows and domes, ranging in composition from basalt to rhyolite, which erupted along the NW-striking Cuzco-Vilcanota fault system. The slip history of the Cuzco-Vilcanota fault system is poorly understood, most likely originating as a pre-Andean suture zone between two distinct lithospheric blocks (Carlier et al., 2005); it appears to have been most recently active as a left-lateral transtensional fault, based on surface morphology along the fault (field observations; Fig. 2; e.g., Sebrier et al., 1985). The volcanic rocks (described in more detail below) that crop out along the Cuzco-Vilcanota fault system host crustal and rare mantle xenoliths.

Xenolith Localities

This study focuses on six xenolith localities discovered during the 2010 and 2011 field seasons (Figs. 1 and 3). Xenolith-bearing volcanic rocks are concentrated in two areas: (1) in flows of Quaternary age found mostly along the Cuzco-Vilcanota fault system, near Cuzco, and (2) in Pliocene lava flows near Puno, along the western shore of Lake Titicaca (Fig. 1). Both the Cuzco and Puno suites are high-K calc-alkaline lavas that range from basalt to dacite. On average, the Puno flows are less silicic than the Cuzco field, although both suites have typical calc-alkaline major-element trends (Carlier et al., 2005; Chapman and Ducea, 2013). Below we provide a general summary of the petrography and geochemistry of xenolith-bearing volcanic rocks in southern Peru.

The Cuzco volcanic suite appears on geologic maps of Peru as the Quaternary Rumicolca formation (e.g., Carlotto et al., 2011), which is composed of high-K intermediate material, referred to as shoshonitic rocks (Carlier et al., 2005; Mamani et al., 2010). The Rumicolca formation crops out as lava flows and domal structures less than 1.5 km in diameter scattered along the Cuzco-Vilcanota fault system (Figs. 1 and 2; Carlier et al., 2005). In detail, exposures of the Rumicolca formation contain a variety of compositions from basaltic andesite to dacite and are best classified as lamprophyres with $K_2O/Na_2O = 1–2$. These flows contain a ubiquitous phenocryst assemblage of biotite, amphibole, clinopyroxene, Fe-Ti oxides, and moderately to strongly aligned plagioclase laths, with subordinate sanidine, orthopyroxene, and apatite phenocrysts (Fig. 4A). Nepheline phenocrysts are present in one locality (Raqchi, described later herein) in samples lacking orthopyroxene. Cuzco suite lavas are extremely enriched in incompatible trace elements, lack
Eu anomalies among rare earth elements (REEs), and have high Sr/Y (Chapman and Ducea, 2013). Carlier et al. (2005) interpreted these rocks as partial melts of clinopyroxene-rich veins in the uppermost mantle. Alternatively, they could be small-degree partial melts of the lowermost crust beneath the Altiplano, which is dominated by clinopyroxene-rich arc cumulates with or without amphibole and garnet. These rocks yield negative $\varepsilon_{Nd}$ values and Sr > 0.705, indicating a lithospheric origin (Carlier et al., 2005). Cuzco suite xenoliths, consisting of (in order of decreasing abundance) gabbros and diorites, granulites, clinopyroxenites, and peridotites, were recovered from Huarocondo, Oropesa, Raqchi, Taray, and Tipon (Fig. 1; Table 1). Of these localities, the flow east of Oropesa contains the highest xenolith abundance and the most variety in the composition and size of xenoliths (up to 10 cm). Most other localities contain smaller (<3 cm) nodules. Granulite xenoliths reported from a Pliocene–Quaternary lamproite dike ~15 km northeast of Cuzco (Carlier and Lorand, 1997) were not investigated here.

A second suite is represented by more mafic trachyandesites and basaltic trachyandesites in Pliocene lava flows (Mamani et al., 2010) exposed ~15 km southeast of Puno (Fig. 1). These rocks are part of the western Altiplano volcanic suite of Carlier et al. (2005) and were classified as shoshonites by those authors. Puno lavas are best classified as high-K calc-alkaline basanites with K$_2$O/Na$_2$O ratios between 1 and 2. Phenocryst assemblages found in the Puno suite are distinct from those found in the Cuzco suite, consisting of plagioclase laths, clinopyroxene, olivine, nepheline, and opaques, and lacking biotite and amphibole. The Puno volcanic suite yields lower $\varepsilon_{Nd}$ (<–3) and higher Sr (>0.705) baseline values than the Cuzco suite (Carlier et al., 2005), also suggesting an origin by partial melting of the South American lithosphere (see following). The Puno suite is also characterized by elevated Sr/Y (>40) and La/Yb (>20) values with no Eu anomalies (Carlier et al., 2005), suggesting a source rich in clinopyroxene and garnet, but not in feldspar. Small (<5 cm) fresh xenoliths of clinopyroxenite and subordinate weathered granitic assemblages were recovered from a single Pliocene–Pleistocene flow in the Umayo volcanic complex (Kaneoka and Guevara, 1984) of the Puno suite (Table 1).

**RESULTS**

Details of sample preparation and the analytical techniques used for whole-rock (major and trace elements, and Sr-Nd-Pb isotopes) and mineral geochronology (garnet Sm-Nd, zircon U-Pb, and monazite U-Th-Pb) are provided in the GSA Data Repository.

**Xenolith Petrography**

Xenoliths recovered from the Cuzco suite are divided into five categories: spinel-bearing peridotites (Fig. 4B), clinopyroxenites (Fig. 4C), gabbros and diorites (Fig. 4D), mafic granulites (Fig. 4E), and felsic peraluminous granulites (Figs. 4F, 4G, and 4H). The main conclusions of this study are based on data from the lattermost group, but petrographic descriptions of all rock types are provided here, as it is important to note the relative abundance of igneous (peridotites

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1 GSA Data Repository item 2015205, analytical techniques, images of garnet mineral separates, and data tables, is available at http://www.geosociety .org/pubs/ft2015.htm or by request to editing@geosociety.org.
and gabbros) versus metasedimentary (clinopyroxenite skarns and both mafic and felsic granulites) xenolith types. For each lithology, average modal mineralogical abundances are given in the GSA Data Repository (see footnote 1), and key petrographic features are shown in Figure 4.

Two fresh peridotite inclusions, both ~1 cm in diameter, were collected from a single location in the Huarocondo flow. These nodules are mineralogically and texturally homogeneous, exhibiting a medium- to coarse-grained xenomorphic granular texture with common 120° grain boundary triple junctions (Fig. 4B) and containing mainly olivine (Fo89-Fo92), orthopyroxene, minor (<3% modal abundance) primary clinopyroxene and chromite-spinel, and secondary (presumably metasomatic in origin) phlogopite and vein calcite and quartz.

Gabbro and diorite xenoliths constitute a second, more abundant and widespread, igneous xenolith group recovered from the Huarocondo, Oropesa, Taray, and Tipon flows of the Cuzco suite. Plagioclase, and hornblende and/or clinopyroxene are the predominant minerals in these rocks, which contain varying amounts of orthopyroxene, olivine, biotite, K-feldspar, titanite, Fe-Ti oxides, apatite, quartz, rutile, and magmatic epidote. Gabbro and diorite xenoliths lacking biotite are medium to coarse grained with allotriomorphic granular textures, whereas biotite-rich varieties are medium grained with well-developed igneous foliation. Biotite- and K-feldspar–rich diorites probably represent fragments of a gradational contact zone between gabbro/diorite and metasedimentary xenolith types, described later. Quenched glass, ubiquitous in metasedimentary xenoliths, is not observed in gabbro and diorite nodules.

Clinopyroxene-rich assemblages were studied from the Cuzco suite (Oropesa and Raqchi flows) and in the Puno suite. Clinopyroxenites from the Oropesa flow contain 40%–90% diopсидic clinopyroxene, with plagioclase and quartz constituting the remainder of the major phases present, and with minor amounts (in order of decreasing abundance) of Fe-Ti oxide, kelyphitized grossular garnet, biotite, K-feldspar, calcite, titanite, apatite, and zircon, likely reflecting a metasedimentary (calc-silicate skarn) origin. Puno suite clinopyroxenites generally have a higher proportion of clinopyroxene (84%–92%), also diopsidic in composition, with minor amounts of Fe-Ti oxides, plagioclase, and quartz. All clinopyroxenite xenoliths are coarse grained with weakly to moderately foliated fabrics.

Mafic granulite xenoliths were recovered from the Oropesa and Raqchi flows in the Cuzco suite and are distinguished from the clinopyroxenite suite based on less abundant clinopyroxene, and a higher proportion of orthopyroxene, plagioclase, biotite, and K-feldspar. Minor phases of the granulite suite include olivine (< 1%), hornblende, garnet, titanite, Fe-Ti oxides, apatite, and up to 1% quenched felsic melt phase along grain boundaries. No quartz was detected in these nodules. Mafic granulite xenoliths are coarse grained and show well-developed gneissic foliation with respect to the clinopyroxenite suite, defined by aligned biotite-rich layers.

A more felsic suite of granulite xenoliths occurs as inclusions in the Oropesa flow, dominated by plagioclase, K-feldspar, and quartz (generally 80%–90% of studied assemblages), with lesser amounts of biotite, garnet, kyanite, clinopyroxene, titanite, Fe-Ti oxides, apatite, and zircon. Garnet-bearing granulites reported from ~15 km northeast of Cuzco (Carlier and Lorand, 1997) probably represent correlative assemblages. This suite of xenoliths exhibits anastomosing networks of rhyolitic glass.
Figure 4. Photographs of structural and petrologic features in xenoliths and volcanic host rocks. (A) Lamprophyre showing biotite, clinopyroxene, opaques, and plagioclase (~20-µm-long laths, flow aligned parallel to long dimension of photograph) phenocrysts set in a glassy matrix, Oropesa flow; crossed-polarized light (xpl). (B) Spinel harzburgite (~10–80 µm spinel grains not visible at magnification of image), Huarocondo flow; xpl. (C) Foliated clinopyroxenite, Puno flow; xpl. (D) Foliated hornblende-clinopyroxene diorite, Oropesa flow; xpl. (E) Mafic granulite, Raqchi flow; xpl. (F) Garnet-bearing felsic granulite, Oropesa flow; plane-polarized light (ppl). (G) Felsic granulite showing evidence for biotite breakdown by dehydration melting. Note melt pockets in textural equilibrium with monazite along biotite cleavage, Oropesa flow; backscattered electron image. (H) Felsic granulite showing neocrystallization of monazite, spinel, Fe-Ti oxides, orthopyroxene, orthoamphibole, and zircon from a pocket of quenched melt. Note radiating needles of orthoamphibole, Oropesa flow; backscattered electron image. Mineral abbreviations: Bt—biotite; Cpx—clinopyroxene; Ged—gedrite; Grt—garnet; Hbl—hornblende; Hc—hercynite; k—kelyphite; Kfs—K-feldspar; Mag—magnetite; Mnz—monazite; Plag—plagioclase; Ol—olivine; Opx—orthopyroxene; Qtz—quartz; Zrn—zircon.
TABLE 1. XENOLITH LOCALITIES AND HOST ROCK PETROGRAPHY

<table>
<thead>
<tr>
<th>Locality</th>
<th>Location</th>
<th>Age of flow*</th>
<th>Lithology†</th>
<th>Phenocrysts§</th>
<th>Xenoliths present#</th>
</tr>
</thead>
<tbody>
<tr>
<td>Huaraocondo</td>
<td>13.42671°S, 72.20319°W</td>
<td>Q</td>
<td>d; lamp</td>
<td>pl, san, cpx, hb, bt, opq</td>
<td>hz, gbd</td>
</tr>
<tr>
<td>Oropesa</td>
<td>13.59760°S, 71.74499°W</td>
<td>&lt;0.7 Ma†</td>
<td>b-a; lamp</td>
<td>bt, hb, cpx, pl, opq, cal, ap</td>
<td>gbd, fgr, cpx, mgr, qz</td>
</tr>
<tr>
<td>Puno</td>
<td>15.88321°S, 69.90642°W</td>
<td>5.6 Ma†</td>
<td>b-a; bsnt</td>
<td>pl, cpx, op, nh, opq</td>
<td>cpx</td>
</tr>
<tr>
<td>Riqui</td>
<td>14.15833°S, 71.38109°W</td>
<td>&lt;0.03 Ma†</td>
<td>b-a; lamp</td>
<td>cpx, op, nh, opq</td>
<td>mgbd</td>
</tr>
<tr>
<td>Taray</td>
<td>14.24040°S, 71.88109°W</td>
<td>Q</td>
<td>b-a; lamp</td>
<td>bt, hb, cpx, op, ap, sq</td>
<td>gbd</td>
</tr>
<tr>
<td>Titon</td>
<td>13.56919°S, 71.78333°W</td>
<td>Q</td>
<td>b-a; lamp</td>
<td>bt, pl, cpx, nh, ap, opq</td>
<td>gbd</td>
</tr>
</tbody>
</table>

*Quaternary (Q; age not available from studied exposure) adjacent flows younger than 2 Ma (Carlier et al., 2005); †—Kaneoka and Guevara (1984).
†Lithologic abbreviations: b-a—basaltic andesite; bsnt—basanite; d—dacite; lam—lamprophyre; teph—tephritite.
§Mineral abbreviations: ap—apatite; bt—biotite; cal—calcite; cpx—clinopyroxene; hb—hornblende; nph—nepheline; ol—olivine; op—orthopyroxene; pl—plagioclase; opq—Fe-Ti oxide; san—sanidine.
#Xenolith abbreviations: cpx—clinopyroxenite; fgr—felsic granulite; hz—harzburgite; gbd—gabbro/diorite; mgr—mafic granulite; qz—quartzite.

(average wt% Na₂O = 1.9, K₂O = 6.16, SiO₂ = 72.6, and Al₂O₃ = 13.6) along grain boundaries. These quenched melts are most voluminous in areas within and surrounding biotite grains. A secondary mineral assemblage of orthopyroxene, orthoamphibole (gedrite), hycertic spinel, corundum, and monazite (Figs. 4F, 4G, and 4H) is associated with quenched melts. Felsic granulite xenoliths are also referred to here as paragneisses, given the ubiquitous gneissose textures and the likely pelitic protolith for these rocks. The ~1–2 cm average diameter of these xenoliths is comparable to the spacing of alternating quartzofeldspathic leucosomes and biotite ± garnet-rich melanosomes; as a result, xenoliths of leucocratic and melanocratic material may be misinterpreted as having been derived from different rocks. Garnet grains are ~0.5–2 mm diameter, generally inclusion-poor, subidioblastic porphyroblasts enclosing mainly apatite and quartz grains <100 μm in diameter. Kelyphite coatings up to 50 μm thick containing mainly glass, ilmenite, and spinel are present on ~10% of garnet porphyroblasts studied here and appear to be restricted to contacts with biotite and/or melt (Figs. 4 and 5).

Thermobarometry and Petrogenetic Modeling

Representative mineral compositions used in thermobarometric calculations are reported in the GSA Data Repository (see footnote 1). Equilibrium pressure-temperature (P-T) conditions (Table 2) were calculated for garnet-bearing felsic granulite xenoliths using the internally consistent average P-T mode in THERMOCALC, version 3.26, assuming a water activity of unity, due to the abundance of hydrous phases (Fig. 6; Powell and Holland, 1994; Holland and Powell, 1998). The P-T conditions for felsic granulite xenoliths were calculated from garnet + kyanite + plagioclase + quartz assemblages interpreted, based on chemical and textural evidence, to have equilibrated during peak metamorphism. Garnet grains from these rocks are rich in almandine, with lesser amounts of pyrope, spessartine, and grossular components (X₃₉ ~ 0.69, X₄₀ ~ 0.17, X₄₀ ~ 0.09, and X₄₀ ~ 0.05; Fig. 5; see also GSA Data Repository material [footnote 1]), and are compositionally homogeneous with flat interior profiles (Fig. 5). Calculated temperatures and pressures and 1σ uncertainties for felsic paragneiss xenoliths are ~760 ± 45 °C at 9.1 ± 1.4–9.5 ± 1.3 kbar (Table 2).

The Gibbs free energy minimization software package THERIAK-DOMINO (de Capitani and Brown, 1987) and the thermodynamic end-member and solution models of the accompanying “cdb55c2d” database (THERMOCALC database as distributed in version 3.30) were used to construct a P-T pseudosection for the bulk composition of sample 11MA20A (Fig. 7). The peak mineral assemblage of garnet + plagioclase + K-feldspar + biotite + quartz present in this sample is quite common in other samples of the felsic granulite suite and is consistent with those predicted from the pseudosection. The P-T results from THERMOCALC plot within the field of the peak assemblage and are in close proximity to the biotite-consuming and orthopyroxene-producing dehydration melting reaction (Fig. 7).

Diorite xenoliths containing the equilibrium assemblage hornblende + plagioclase + K-feldspar + quartz + titanite + Fe-Ti oxide were analyzed for aluminum-in-hornblende (Al-in-hbl) igneous barometry (Hammarstrom and Zen, 1986; Hollister et al., 1987; Johnson and Rutherford, 1989; Schmidt, 1992; Anderson and Smith, 1995; Ague, 1997). We report hornblende-plagioclase temperatures as well as Al-in-hbl pressures for calibrations based on Hammarstrom and Zen (1986), Hollister et al. (1987), Johnson and Rutherford (1989), Schmidt (1992), and Anderson and Smith (1995) in the GSA Data Repository material (see footnote 1). Diorite xenoliths yield Anderson and Smith (1995) temperatures and pressures ranging from 710 °C to 839 °C and 8.6–13.7 kbar, respectively.

Two samples of mafic granulite yielded significantly higher P-T estimates than either felsic paragneiss or diorite xenolith suites, ranging from 978 °C to 1377 °C and from 10.9 to 11.7 kbar. These estimates are based on two-pyroxene thermometry (Wood and Banno, 1973) and clinopyroxene-plagioclase–quartz barometry (Ellis, 1980).

Whole-Rock Major- and Trace-Element Geochemistry

Felsic granulite xenoliths are the most silicic assemblages studied here, with SiO₂ ranging from 39.51 to 66.51 wt% (Fig. 8A; see footnote 1 for data). One sample containing abundant garnet (16.4 modal%) and biotite (28.5 modal%), most likely representing a nodule of melanosomic material, yielded lower silica values of 49.59 wt% (see footnote 1). Relative to the other xenoliths studied here, this suite of xenoliths shows: (1) the highest weight percent values of Al₂O₃ (16.60–20.60), Na₂O (2.13–5.19), and K₂O (3.94–7.09) and the lowest MgO (0.44–4.01), FeO* (1.20–13.65), and CaO (1.58–2.75); (2) the most pronounced enrichments in fluid-mobile large ion lithophile elements (LILEs) relative to high field strength elements (HFSEs), with notable trace-element peaks at K and Pb and depletion in Nb, Zr, and Ti, although the Zr deficit may be due to incomplete dissolution of zircon during sample preparation; and (3) negative Eu anomalies and heavy (H) REE depletion in samples with relatively low modal plagioclase (~30%) and garnet (~5%), respectively (Fig. 8B; see footnote 1). Compositionally, leucogneiss xenoliths are similar to the average composition of the North American shale composite (Gromet et al., 1984); they are modally similar to a granodiorite.

Diorite and gabbro xenoliths are considerably more mafic than felsic granulite nodules, with lower weight percent values of SiO₂ (42.24–52.49) and K₂O (0.48–3.08), higher values of MgO (2.43–9.92), FeO* (4.48–15.51), and CaO (8.04–14.28), and comparable ranges of Al₂O₃ (2.13–11.57) and Na₂O (1.68–4.43). With respect to felsic granulite xenoliths, diorite and gabbro xenoliths are depleted in LILEs and show similarly variable HFSE concentrations (Fig. 8B). Hornblende-rich xenoliths show concave-downward middle rare earth element (MREE) trends with maxima between Gd and...
Ho, in contrast to clinopyroxene-rich gabbroic xenoliths, which show monotonically decreasing light (L) REE to HREE patterns. Samples lacking significant clinopyroxene and containing relatively low modal plagioclase (<30%) and relatively high modal hornblende (>60%) show positive Eu anomalies; samples containing abundant clinopyroxene (>20%) show negative Eu anomalies; and remaining diorite and gabbro xenoliths lack Eu anomalies.

Clinopyroxenite xenoliths show a wide range of silica values, with two distinct clusters from 41.57 to 55.66 and from 69.33 to 76.13 wt% SiO₂. The large variation in silica values derives from the former of the two clusters being clinopyroxene-dominated assemblages, whereas the latter of the two clusters contain >40% modal quartz. These xenoliths have relatively low Al₂O₃ (1.53–7.40), K₂O (0.03–0.51), and Na₂O (0.38–1.97), high values of MgO (6.94–16.68) and CaO (10.39–21.46), and a wide range of FeO* (2.19–18.68). This suite is depleted in LILEs, HFSEs, and REEs relative to diorite/gabbro and felsic granulite xenoliths. Two distinct REE patterns are apparent in this xenolith suite: (1) Nodules containing quartz and the lowest clinopyroxene abundances exhibit negative Eu anomalies and La/Yb values of 4–15, and (2) clinopyroxene-rich (>80%) assemblages yield La/Yb of <5 and show concave-downward patterns with maxima between Nd and Sm and lack significant Eu troughs.

Mafic granulite xenoliths generally overlap the major-element compositions of diorite/gabbro xenoliths but give lower CaO (4.18–8.13) and Na₂O (1.56–2.57) and higher K₂O (2.92–7.28) values. These assemblages display similarly “spiky” trace-element patterns to those of the felsic granulite suite and are depleted in HREEs relative to LREEs (La/Yb = 7–62).

Peridotite xenoliths are the most mafic of the assemblages studied here, with wt% SiO₂ of 40.90–45.52. These xenoliths show the lowest weight percent oxide values of Al₂O₃ (0.59–2.14) and CaO (0.03–1.24), with negligible Na₂O and K₂O, the highest values of...
MgO (42.99–46.23), and intermediate FeO* (7.92–10.34). These xenoliths show prominent Nb troughs and Pb peaks and LILE to HFSE depletion in trace-element variation diagrams, and they show slight enrichment in LREEs with respect to MREEs and HREEs (La/Yb ~ 5) and small negative Eu anomalies.

**Whole-Rock Isotope Data**

Sr and Nd isotopic data from 27 xenoliths and Pb isotopic data from 29 xenoliths are provided in the GSA Data Repository (see footnote 1). Felsic granulite xenoliths define distinct and enriched arrays in isotope plots (Fig. 9), showing elevated \(^{87}\text{Sr}/^{86}\text{Sr}\) spanning a wide range from 0.7109 to 0.7820, \(^{143}\text{Nd}/^{144}\text{Nd}\) values from 0.51258 to 0.51292, \(^{207}\text{Pb}/^{204}\text{Pb}\) from 15.618 to 15.679, \(^{208}\text{Pb}/^{204}\text{Pb}\) from 38.577 to 38.921, and \(^{208}\text{Pb}/^{204}\text{Pb}\) values from 18.731 to 19.679. Isotopic compositions of the felsic granulite suite overlap values reported from “garnet granulite” xenoliths from Bolivia (McLeod et al., 2013) and are dissimilar to gneisses of the Arequipa terrane, which exhibit higher \(^{208}\text{Pb}/^{204}\text{Pb}\) and \(^{208}\text{Pb}/^{206}\text{Pb}\) values (Fig. 9; Loewy et al., 2004). Cuzco suite metasedimentary clino.pyroxenites lie along the Sr-Nd and Pb isotope array defined by felsic granulite xenoliths at lower \(^{87}\text{Sr}/^{86}\text{Sr}\) (0.7086–0.7094) and Pb (i.e., \(^{207}\text{Pb}/^{206}\text{Pb}\), \(^{208}\text{Pb}/^{206}\text{Pb}\), and \(^{208}\text{Pb}/^{204}\text{Pb}\) values), and higher \(^{143}\text{Nd}/^{144}\text{Nd}\) (0.51261–0.51277). Puno suite clino.pyroxenites plot at lower \(^{87}\text{Sr}/^{86}\text{Sr}\) and Pb values and higher \(^{143}\text{Nd}/^{144}\text{Nd}\) than Cuzco suite clino.pyroxenite xenoliths. Diorite and gabbro xenoliths show the most radiogenic Nd isotope ratios (0.51258–0.51292) and the least radiogenic Sr values (0.7041–0.7089) over a wide range of Pb values. This suite of xenoliths shows a high degree of overlap in isotopic composition with the high-K intermediate volcanic rocks that entrain them (Carlier et al., 2005; Mamani et al., 2010). Peridotite and mafic granulite xenoliths exhibit intermediate Sr-Nd-Pb isotopic values (Fig. 9), approximating the average xenolith value for the entire suite.

**Geochronology**

**Zircon U-Pb**

Zircon grains from a single diorite xenolith from the Oropesa flow (11MA20E) are subhedral, 100–150 μm in length, and inclusion poor, and they exhibit simple oscillatory zoning patterns in cathodoluminescence (CL) images (Fig. 10A). In situ laser-ablation–inductively coupled plasma–mass spectrometry (LA-ICP-MS) analysis of nine zircon grains from this sample yielded a weighted mean age of 0.9 ± 0.4 Ma (2σ; Fig. 10A [see footnote 1 for data]), which is identical within uncertainty to published ages from the volcanic host rocks (Carlier et al., 2005).

![Figure 6. Calculated pressures and temperatures for felsic granulite, diorite/gabbro, and mafic granulite samples, showing 30 °C/km gradient and aluminosilicate stability fields for reference. 1σ uncertainty ellipses (felsic granulite samples) are based on propagation of uncertainties on thermodynamic data and activity-composition relationships through thermobarometric calculations. 1σ error bars (diorite/gabbro and mafic granulite) are from Table 2. Al—aluminum; Hbl—hornblende; Plag—plagioclase; PX—pyroxene; CPX—clino.pyroxene; Qtz—quartz; GASP—garnet-aluminosilicate-plagioclase-quartz; Ky—kyanite; And—andalusite; Sil—sillimanite.](image-url)
Forty-six concordant zircon grains were analyzed from a felsic granulite (11MA20A) within the same flow (Fig. 10B [see footnote 1]). Zircon grains from this and other felsic granulite samples are subrounded, ~50–200 µm in length, and inclusion poor, and they preserve a faint and broad zoning texture in approximately half of studied grains. The remainder show two distinct grain domains: (1) oscillatory zoned CL-dark cores overgrown by (2) oscillatory zoned to patchy and convolute zoned CL-bright rims up to ~50 µm wide. Targeted analyses of core and rim domains yielded exclusively Mesoproterozoic (1017–1529 Ma) and Neoproterozoic to Paleozoic (593–472 Ma), Mesoproterozoic (1017–1529 Ma) and Neo-

A normalized probability plot comparing detrital zircon age spectra from sample 11MA20A with Bolivian granulite xenoliths (McLeod et al., 2013) and with Paleozoic and Mesozoic cover assemblages of the Coastal Cordillera, Altiplano, and the Eastern Cordilleran of Peru (Reimann et al., 2010; Bahlburg et al., 2011; Reitsma, 2012) is shown in Figure 10B. The normalized probability plot was constructed using 207Pb/206Pb ages for grains younger than 800 Ma and 207Pb/206Pb ages for grains older than 800 Ma. Analyses with greater than 10% uncertainty, 30% discordance, and/or 5% reverse discordance (17 grains) were excluded.

Considering all analyzed grain domains, sample 11MA20A has major age peaks at ca. 415 Ma and 470 Ma, scattered ages between 215 and 400 Ma and 1000 and 1550 Ma, and one Cretaceous grain. The spectrum of ages from this sample suggests a detrital origin for zircon grains within it. A maximum protolith depositional age is difficult to assign to this sample, given the presence of a single Cretaceous grain. Based on a cluster of three grains that overlap in age within analytical error, a conservative maximum depositional age of 229.2 ± 7.3 (2σ) Ma (Late Triassic) was calculated for this sample.

**Garnet Sm-Nd**

Two-point Sm-Nd garnet and whole-rock isochrons were calculated from felsic granulite samples 13 and 11MA20A from the Oropesa flow, yielding ages and 2σ uncertainties of 42 ± 2 and 17 ± 39 Ma, respectively (Fig. 11). Taken together, garnet and whole-rock Sm-Nd data from both analyzed samples yield a four-point isochron age of 41 ± 7 Ma (2σ). Mineral data and whole-rock Sm and Nd isotope data are provided in the GSA Data Repository (see footnote 1). Given an average garnet grain size of ~1 mm and assuming a slow cooling rate of 3 °C/m.y. (typical of orogenic plateaus), a Sm-Nd closure temperature of ~760 °C was calculated, using available diffusion parameters of Ganguly and Tirone (1999). This calculated closure temperature is similar to peak metamorphic temperature estimates from these rocks (Fig. 6; Table 2), which may provide an explanation for the wide range in precision in our two Sm-Nd ages. In other words, elevated metamorphic temperatures may have disturbed the Sm-Nd systematics in sample 11MA20A relative to sample 13.

**Monazite U-Th-Pb**

Monazites in samples 11MA20A and 11MA20C show two distinct compositional domains, based on electron microprobe X-ray mapping and spot analyses (Fig. 12 [see footnote 1]): grain interiors with average wt%
Figure 8 (on this and following page). (A) Major-element variation diagrams for Peruvian xenoliths compared to back-arc lavas (Chapman and Ducea, 2013) exposed along the Cuzco-Vilcanota fault system (CVFS). FeO* denotes total Fe as FeO.
ThO$_2$ = 5.43%, UO$_2$ = 1.08%, and Y$_2$O$_3$ = 3.46% and rims with ThO$_2$ = 5.08%, UO$_2$ = 1.04%, and Y$_2$O$_3$ = 2.93%. Monazites show sensitive high-resolution ion microprobe (SHRIMP)–determined REE patterns with strong enrichment of the LREEs relative to the HREEs (average La/Yb = 605) and show strong negative Eu anomalies (Eu/Eu* = 0.06; Fig. 13).

Weighted mean monazite ages of 4.4 ± 0.3 Ma (mean square of weighted deviates [MSWD] = 6.3, 2σ) and 3.2 ± 0.2 Ma (MSWD = 4.6, 2σ) were calculated from samples 11MA20A and 11MA20C, respectively (Fig. 13 [see footnote 1]). Calculated weighted mean ages do not include three analyses in which both monazite and adjacent minerals were overlapped by the ion beam. Uncertainties were multiplied by the square root of the MSWD to reflect the higher degree of scatter for individual analyses. Monazite U-Pb ages from the two compositional domains recognized in monazite from each sample overlap within analytical error. Monazite U-Th-Pb ages may reflect either the timing of growth or cooling through ~800–900 °C (Cherniak et al., 2004; Gardes et al., 2006).

DISCUSSION

Xenolith Classification Based on Radiogenic Isotopes

There are two groups of crustal xenolith samples based on their measured Nd and Sr isotopes: Group 1 rocks show $^{143}$Nd/$^{144}$Nd > 0.5125 and $^{87}$Sr/$^{86}$Sr < 0.710, and group 2 rocks yield $^{143}$Nd/$^{144}$Nd < 0.5125 and $^{87}$Sr/$^{86}$Sr > 0.710 (Fig. 9). Group 1 rocks generally have igneous textures, with mostly mafic to intermediate compositions (SiO$_2$ < 55 wt%), and are interpreted to represent Cenozoic additions to the crust. Group 2 assemblages show metamorphic textures and are interpreted here to represent basement and metamorphosed cover samples of the Altiplano, e.g., Paleozoic and Mesozoic rocks from the Central Andes, including the Famatinian arc, the Ollantaytambo formation, the Mitu Group, and other regionally extensive units. Similar distributions of isotopes exist among igneous and metamorphic basement rocks from the Central Andes (Wörner et al., 2000; Lucassen et al., 2001; Loewy et al., 2003, 2004; Fig. 9).

The majority of group 1 xenoliths are ultramafic, mafic, or intermediate in composition, and they were most likely derived from relatively primitive mantle magmas. The extent to which some of them represent asthenosphere-derived additions to the crust or mixtures of asthenospheric and lithospheric materials is debatable, but it is clear that they are more narrowly grouped in terms of radiogenic isotopes and have younger model ages that the second group of crustal xenoliths (generally >1.0 Ga vs. >1.0 Ga [see footnote 1]). Next, we will focus on the significance of group 2 xenoliths, which we interpret to represent basement and metamorphosed cover rocks of central South America.

Origin of Crustal Xenoliths

Metasediments

Garnet leucogneisses (i.e., felsic granulites) are interpreted as the metasedimentary cover to the South American basement, similar to high-grade intermediate to felsic granulite facies xenoliths recovered from the Bolivian Altiplano (Davidson and de Silva, 1995; McLeod et al., 2012, 2013). These deposits must be younger than the 550 Ma Rb-Sr whole-rock error.
Xenolith constraints on Altiplano architecture and assembly

The detrital distributions of age peaks corroborated by the maximum depositional age of 229.2 ± 7.3 Ma from zircon U-Pb constraints on sample 11MA20A (Fig. 10B) suggest that the age is Late Triassic or younger, and possibly as young as Cretaceous, based on the presence of a single grain of this age.

Mesozoic volcanic and sedimentary rift and postrift cover rocks of the mid- to Late Triassic Mitu Group, the Early Cretaceous Huancané Formation, and the mid- to Late Cretaceous Yuncaypata Group commonly comprise the country rocks into which Rumicolca back-arc volcanic rocks intruded (e.g., Kontak et al., 1990a; Sempere et al., 2002; Mišković et al., 2009; Reitsma, 2012). Two key similarities between Mesozoic strata of SE Peru, in particular the Mitu Group and Huancané Formation, and leucogneiss xenoliths lead us to suggest that the latter are the highly metamorphosed equivalents of the former. First, detrital zircon age spectra from Mesozoic strata (Reitsma, 2012; Pérez and Horton, 2014; Fig. 10B) overlap the spectrum of leucogneiss sample 11MA20A, each containing distinct late Paleozoic to Late Triassic, early Paleozoic, and Mesoproterozoic age peaks. These age peaks are associated with 0.16–0.35 Ga Gondwanide (e.g., Kontak et al., 1990a; Mišković et al., 2009), 0.44–0.53 Ga Famatinian-Pampean, and 0.9–1.3 Ga Sunsas/Grenville magmatic-orogenic events. Mesozoic strata, as well as leucogneiss xenoliths studied here, lack significant proportions of the Neo- oproterozoic and Paleoproterozoic and older zircon grains that are abundant in South American cratonic crust. Grains of these ages are present in Paleozoic strata overlying the Altiplano, Eastern Cordillera, and Arequipa terrane (Reimann et al., 2010; Bahlburg et al., 2011).

Second, volcano-sedimentary assemblages of the Mitu Group (interbedded sandstones, mudstones, conglomerates, and intermediate lavas) represent suitable protoliths for leucogneiss xenoliths.

The inference that felsic xenoliths in the Rumicolca Formation represent the metamorphic equivalents of the Mitu Group implies that portions of the Mitu Group must somehow have been transferred to ~30–35 km depth. As mentioned previously, the volcano-sedimentary strata of the Mitu Group are thought to represent rift deposits produced during the initial stages of Gondwana disassembly (Kontak et al., 1990a; Sempere et al., 2002; Pérez and Horton, 2014). Inversion of basins containing Mitu Group strata is thought to have commenced at a slow pace in Late Jurassic–Early Cretaceous time and culminated during the Andean orogeny (Sempere et al., 2002), although the precise timing of
uplift is not well understood. Our garnet whole-rock Sm-Nd age of 42 ± 2 Ma from leucogneiss sample 13 indicates that garnet growth in this sample, and hence a major pulse of Mitu basin inversion and accompanying crustal thickening beneath the plateau, took place in the Eocene. Burial of the Mitu basin through thrust inversion requires that the vertical throw on Triassic rift structures was in places significant (greater than the ~7-km-thick Paleozoic section) and that the normal fault basins were not completely inverted. Complete reactivation of some rift structures, with limited reactivation of others, provides a mechanism for burial of Mitu rocks preserved in the hanging walls of normal faults (Fig. 14). Burial by basement thrust structures can produce Mitu paleodepths no greater than 25–26 km, which is below the lower limit of 10 km uncertainty in calculated peak metamorphic pressures of 9.1 ± 1.4–9.5 ± 1.3 kbar (Table 2), assuming a mean crustal density of 2.8 g/cm³. Therefore, it is considered highly likely that additional processes, discussed in the following sections, were responsible for conveying Mesozoic sediments to ~9 kbar conditions.

Following burial of Mesozoic sediments to midcrustal levels, peak temperatures at 750 °C and higher (and a thermal gradient of 30 °C/km; Fig. 6) were likely achieved through some combination of thermal relaxation in the crust and Cenozoic magma input. The time lag between Eocene garnet Sm-Nd ages and Pliocene monazite U-Th-Pb ages provides insight into the relative importance of thermal relaxation versus magmatism in the metamorphism of these rocks. Pliocene monazite U-Th-Pb ages probably represent the timing of mineral growth rather than cooling ages, considering textural evidence for monazite growth during partial melting (Figs. 4G and 4H), calculated peak metamorphic temperatures below ~800–900 °C U-Th-Pb closure, and the fact that monazite must spend considerable periods of time at or above 800 °C to affect U, Th, and Pb systematics (e.g., Cherniak et al., 2004; Gardes et al., 2006; Williams et al., 2007).

Monazite growth during Pliocene time indicates that ~40 m.y. had elapsed since garnet growth before partial melting of felsic leucogneisses took place. Recent thermochronologic results from plinths in the Eastern Cordillera of SE Peru suggest similarly protracted shortening beginning in the middle Eocene and continuing until at least middle Miocene time (Perez et al., 2013).

Monazite chemistry, specifically the observed decline in Y concentration from monazite core to rim, reveals additional information regarding garnet stability and equilibration with monazite. Recall that garnet grains in felsic leucogneiss xenoliths exhibit thin, melt-containing, kelyphitic rims in the vicinity of biotite grains that are breaking down to glass, orthopyroxene, spinel, monazite, Fe-Ti oxides, orthoamphibole, and zircon. Similar textural and phase relationships are commonly observed in granulite-grade metapelitic rocks in which garnet-involved incongruent melting of biotite has taken place (e.g., Cesare, 2000; McLeod et al., 2012). The biotite-out reaction has a positive slope in the vicinity of the calculated peak conditions in P-T space (Fig. 7), indicating that dehydration partial melting is expected under decreasing pressure and/or increasing temperature conditions. We suggest that biotite dehydration melting with garnet as a reactant and monazite, spinel, and orthopyroxene as products occurred during decompression, perhaps accompanied by a thermal pulse, due to entrainment of these rocks in back-arc magmas as xenoliths. We further speculate that garnet decomposition contributed Y to the reaction volume, which was sequestered into monazite at the outset of the reaction, with diminishing Y as the reaction progressed. Given that the volcanic field containing these xenoliths is ~3 m.y. younger than monazite U-Pb ages from xenoliths entrained within it (Table 1; Kaneoka and Guevara, 1984), these xenoliths were likely plucked from ~30–35 km depth and resided in the mid- to upper crust prior to ultimate eruption.

**Meta-Igneous Assemblages**

Mafic granulite samples (e.g., 11MA2A) have measured Sr isotopic ratios of 0.707–0.708, εNd of ~5, and equilibration temperatures exceeding 1000 °C and pressures of over 11 kbar (Table 2; Fig. 6). These rocks have mafic chemistry, have metamorphic textures, and are characterized by isotopic ratios similar to the Famatinian arc (e.g., Otamendi et al., 2009, 2012). Equivalent mafic granulites were not described from the Bolivian xenolith suites (McLeod et al., 2013). We interpret these rocks to represent fragments of the South American lithosphere upon which Mesozoic strata were deposited. Peridotite nodules from the Huarococondo flow also probably represent xenoliths of the sub-Altiplano mantle lithosphere or pieces of peridotite lenses that crop out within the Mitu rift zone (Jacay et al., 1999; Sempere et al., 2002). It could be argued, based on the small size of peridotite fragments studied here, that these nodules represent glomerocrysts. However, foam textures and the lack of glomeroporphyritic textures preclude the possibility that these nodules formed by synneusis (Hogan, 1993).

Nd ratios and depleted mantle model ages (TDM) values of 0.46–1.4 Ga in mafic granulite and peridotite xenoliths (see footnote 1) suggest that these rocks represent mafic fragments of the Famatinian arc. An Amazonian origin is ruled out because Amazonian basement rocks are expected to have lower εNd ratios (<–15) and older TDM values (>1.5 Ga). Another possibility is that mafic granulites represent relatively young (Permian to Cenozoic) mafic magmatic additions to the lower crust that were subsequently metamorphosed. However, the majority of mafic magmatic products with late Paleozoic

![Figure 10](image-url)
Figure 10 (continued). (B) Normalized probability plot comparing the spectrum of concordant zircon ages from felsic granulite sample 11MA20A, plotted alongside spectra from Bolivian granulite xenoliths (McLeod et al., 2013) and composite curves from Paleozoic and Mesozoic cover assemblages of the Coastal Cordillera, Altiplano, and the Eastern Cordillera of Peru (Reimann et al., 2010; Bahlburg et al., 2011; Reitsma, 2012). Cathodoluminescence image (inset) shows analyzed core and rim age domains (1σ ages of each spot given in Ma).
Figure 11. Garnet (Grt) whole-rock (wr) Sm-Nd isochron for samples 11MA20A (labeled wr-11) and 13 (wr-13). Error bars and calculated two-point isochron ages (Ludwig, 2003) are 2\(\sigma\).

Figure 12. Monazite X-ray maps. Location of two analyzed spots is shown in backscattered electron image (BSE); monazite core and rim regions overlap in age (1\(\sigma\) uncertainties shown).
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pre-Eocene ages that are found on the plateau within 250 km of the xenolith localities have more depleted isotopic signatures (Mamani et al., 2010). Specifically, their $e_{Nd}$ values range from $-2$ to $+4$, which are distinctive from the western South American orogenic margin range of $e_{Nd}$ ($\sim -10$ in the Central Andes). In conclusion, we interpret mafic granulites as basement rocks to the Mesozoic and older cover sequence, represented by garnet leucogneisses.

A corollary to this study is that all igneous-textured basement rocks (hornblende diorites, gabbros, etc.), which are the predominant xenolith populations among the localities from southern Peru and Bolivia (McLeod et al., 2013), are magmatic additions newer than the mid-Cenozoic metamorphic event described here. In fact, there is U-Pb zircon evidence that at least one of the diorites from Oropesa is Quaternary in age (Fig. 10A).

Figure 13. Sensitive high-resolution ion microprobe (SHRIMP) data: (A) U-Pb geochronologic and (B) chondrite-normalized (e.g., Sun and McDonough, 1989) rare earth element (REE) data for selected monazite grains from samples 11MA20A and 11MA20C. Error ellipses are 2σ. MSWD—mean square of weighted deviates.

Implications for Plateau Uplift and Crustal Flow

The history of surface uplift of the Central Andean orogen is debated. Some workers have suggested that major (>2 km over 2-3 m.y.) surface uplift of the Altiplano occurred during the mid- to late Miocene based on temperature-sensitive paleoaltimeters (Ghosh et al., 2006; Garzione et al., 2006; Hoke and Garzione, 2008). In contrast to the hypotheses of rapid pulses of uplift, others have favored long-lived and steady isostatic rise in topography due to protracted crustal shortening (e.g., Barnes and Ehlers, 2009; Isacks, 1988).

Our results demonstrate that low-density sedimentary cover rocks and their metamorphosed equivalents have occupied the upper
35 km of the northern Altiplano crust since the Eocene. This is consistent with models for early Cenozoic shortening and crustal thickening and accompanying rock uplift from beneath the northern Altiplano. However, our results do not preclude the possibility that regional uplifts were driven by short-lived lithospheric delamination events, as Eocene–Oligocene crustal thickening and magmatism could have preconditioned the lower crust and/or mantle beneath the northern Altiplano to foundering and renewed uplift during and following Miocene time. A major influx of primitive magmas represented by diorites, gabbros, and cumulate xenoliths in these suites could be related to a younger convective removal event that took place in the Miocene or more recently.

Monazite U-Pb data and textural observations from metasedimentary assemblages, as well as zircon U-Pb data from dioritic rocks, indicate that a significant proportion of the middle crust (~20–25 km depth) beneath the northern Altiplano has either undergone some degree of partial melting or has been added to the crust as a magma since ca. 5 Ma. This observation is pertinent to discussions of the Bolivian orocline, which possibly resulted from along-strike variations in shortening due to variations in thickness of the colliding Nazca plate (Isacks, 1988; Sempere et al., 2002; Capitanio et al., 2011). The orocline is also characterized by extremely thick (locally ~75 km; Beck et al., 1996; Beck and Zandt, 2002), mechanically decoupled crust. While the crustal thickness is fairly uniform along the orocline, shortening estimates are not. The total amount of crustal shortening accumulated since Eocene time is estimated to have been 115 km (this study) to 177 km (gotberg et al., 2010) greater in central Bolivia compared to southern Peru. The Bolivian shortening estimates can account for crustal thickness in that portion of the orocline; however, shortening estimates from southern Peru cannot account for the extreme crustal thickness beneath the northern Altiplano. A possible explanation for

Figure 14. Balanced cross-section A-A’ from the eastern edge of the Western Cordillera, across the Altiplano, and into the Eastern Cordillera. Cross-section trace is shown in Figure 1. Deformed length is 185 km, and restored undeformed length is 370 km (185 km, or 50%, shortening). Additional burial of Mitu Group sediments from ~25 km depth, perhaps by magma loading, to >30 km depth is necessary to explain thermobarometric data. Approximate location of xenoliths recovered in this study is shown on deformed section.
The shortening to thickness disparity is lateral flow of ductile mid- to lower-crustal material from Bolivian to Peruvian sections of the Altiplano (Yang and Liu, 2003; Ouimet and Cook, 2010; Gotberg et al., 2010). Peruvian xenoliths provide evidence for a zone of partial melt situated between 35 km (based on leucogneiss pressure estimates) and 20 km (based on diorite estimates) depth beneath the northern Altiplano. However, the degree to which this relatively weak crustal zone has been mobilized parallel to the strike of the orogen is uncertain. Similarities between garnet leucogneisses and Mesozoic sediments in the vicinity of Cuzco suggest that, aside from being brought from the surface to ~30–35 km depth, the protoliths for garnet leucogneisses were deposited proximal to their current position (i.e., it is unlikely that these rocks flowed northward in the mid- to deep crust). Hence, partially molten material beneath the northern Altiplano was more likely generated during a thermal pulse associated with the Pliocene and later influx of primitive magmatic additions. Such an influx may have played a significant role in thickening the crust under the northern Altiplano through magmatic additions and possibly convective overturning of the crust, thereby transporting supracrustal rocks downward as has been suggested beneath the Altiplano-Puna volcanic complex (Babayko et al., 2002).

SUMMARY AND CONCLUSIONS

Crustal and probable mantle xenoliths discovered in Pliocene and Quaternary high-K intermediate and mantle xenoliths from the southern Peruvian Altiplano show that the Mesozoic regional base-cover contact is located ~30–40 km below the modern surface. Equilibrium P-T estimates of >1000 °C and >11 kbar from mafic granulite xenoliths indicate that basement assemblages below this interface are predominantly mafic and are isotopically similar to Central Andean crust. Granulite grade metasedimentary rocks with radiogenic compositions (60Sr/88Sr >0.711 and 184Nd/144Nd <0.5126) comprise a cover sequence—correlative to Triassic Mitu Group rift basin deposits—that has been transferred to >30 km depth beneath the plateau at temperatures >750 °C. Based on our new shortening estimates, a maximum of ~25 km of Mitu basin tectonic burial could have been accomplished through basin-involved thrusting, falling short of thermobarometric estimates requiring that these rocks equilibrated at ~30–35 km paleodepth. Therefore, additional processes such as magma loading must have been in part responsible for conveying Mesozoic sediments to ~9 kbar conditions.

A garnet Sm-Nd age of 42 ± 2 Ma, interpreted to reflect the timing of garnet growth in Mitu Group protoliths, suggests that metamorphism associated with burial of these rocks and accompanying major crustal thickening and uplift of the northern Altiplano began in the Eocene, probably during the Incaic regional shortening episode. Following burial, a protracted phase (~40 m.y.) of thermal relaxation–related heating and lower-crustal residence with limited unroofing occurred. Monazite neoblasts, yielding U-Pb ages of 3.2 ± 0.2–4.4 ± 0.3 Ma, associated with incongruent melting of biotite and involving garnet as a reactant, probably resulted from near-isothermal decompression resulting from xenolith entrainment and rapid transport toward the surface. Significant primitive magmatism with 60Sr/88Sr <0.709 and 184Nd/144Nd <0.5126, as evidenced by Pliocene–Quaternary dioritic gabbroic xenoliths emplaced at ~7 kbar and their extrusive equivalent volcanic host rocks, is inferred to be responsible for further regional crustal thickening by magma loading. The presence of these primitive (asthenospheric) mafic melts in the crustal section may reflect later removal of parts of the lithosphere.

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Xenolith constraints on Altiplano architecture and assembly


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