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Origin of fayalite granitoids: new insights from the Cobquecura pluton, Chile, and its metapelitic xenoliths

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Abstract

Fayalite granitoids of the Cobquecura Pluton, Central Chile, occur in an active continental margin setting, which is unusual for such rocks. They are closely associated with gabbros, and both lithologies contain hercynite-rich, Fe-rich metapelitic xenoliths. The primitive magmas of the Cobquecura Pluton are derived from a mafic source, similar to the regional subarc subcontinental mantle. Sr-Nd-Pb isotope data demonstrate that crustal contributions were important in the petrogenesis of the fayalite granitoids. The crustal contribution, however, does not consist of rocks similar to the regionally exposed Paleozoic basement rocks, but rather resembles partial melts derived from metapelitic rocks at depth. The corresponding restites are the hercynite-rich xenoliths. A relatively high content of rare earth elements and high field strength elements in the fayalite granitoids is linked to selective assimilation of crust. This is evident from the abundant xenoliths that are entrained in the fayalite granitoids, and whose leucosomes are rich in accessory minerals carrying these elements. Such Fe-rich rocks similar to the hercynite-rich xenoliths could have acted as a chemical buffer keeping the oxygen fugacity low, which is crucial for the generation of fayalite bearing granitoids.

Keywords: Fayalite granitoid; Sr-Nd-Pb isotopes; restitic metapelitic xenoliths; selective assimilation; Chile; petrogenesis.

1. Introduction

Granitoids are the most abundant plutonic rocks in the Earth's upper continental crust and their formation is linked to a wide spectrum of different tectonic and geodynamic processes. Granitoids may either form by differentiation of mafic, mantle-derived melts, or by partial melting of crustal rocks. In many continental settings, granitoid petrogenesis involves both fractional crystallization of mantle-derived melts, and assimilation of crustal material. Fayalite-bearing granitoids are fairly rare. Their occurrences are mainly related to anorogenic or alkaline magmatic complexes, usually of intraplate setting (e.g., Njonfang and Moreau, 2000; Frost et al., 2002, and references therein).

Most fayalite granitoids occur in Precambrian terrains (e.g., Pikes Peak Batholith, Central Colorado, USA, Smith et al., 2001; Sherman Batholith, Wyoming, USA, Frost et al., 1999; Padthaway Ridge, South Australia, Turner et al., 1992). Only few fayalite granitoids are known from Phanerozoic terrains, e.g., the anorogenic province of Corsica (France) (Poitrasson et al., 1995) and the Tibchi anorogenic ring complex of Nigeria (Mücke, 2003). Fayalite granitoids are characterized by comparatively high overall Fe contents, high FeO/(FeO+MgO) ratios, the occurrence of Fe-rich anhydrous mafic minerals such as fayalite

and hedenbergite, and the late crystallization of hydrous mafic minerals like annite or Fe-rich amphibole (Turner et al., 1992; Mücke, 2003). These characteristics indicate distinct crystallization conditions from H₂O-poor magma at low fO_2 (Stephenson and Hensel, 1978; Frost et al., 1988).

The petrogenesis of fayalite granitoids has been a longstanding subject of discussion. Some authors proposed that these granites were generated by fractionation from mafic magmas (e.g., Loiselle and Wones, 1979; Frost et al., 2002). Other authors favored fayalite granitoid formation by partial melting of dehydrated lower crustal sources like granulites (Collins et al., 1982; Clemens et al., 1986) or tonalites (Creaser et al., 1991).

At the southwestern margin of Triassic Gondwana, along the South-Central Chilean coast, A-type granitoids (cf. Whalen et al., 1987) have been identified as an important constituent of the regional Triassic magmatism (Gana, 1991; Parada et al., 1991; Vásquez and Franz, 2008). Fayalite granites were discovered in the Triassic Cobquecura Pluton (Coastal Cordillera, ~36°S; Vásquez and Franz, 2008). Here we discuss the petrogenesis of these fayalite granitoids in the light of their petrographic, geochemical and Sr-Nd-Pb isotopic characteristics, with particular attention to the metapelitic xenoliths entrained in the magmas.

2. Geological setting and petrography

The southwestern margin of South America has been a continental margin of Gondwana and pre-Gondwana at least since the Late Proterozoic-Early Cambrian (e.g., Coira et al., 1982). In the southern part of Central Chile, the subduction regime between the Late Carboniferous and the Early Triassic generated a magmatic arc and an accretionary prism (Hervé et al., 1981; Parada, 1990; Martin et al., 1999). The metamorphic basement in south-central Chile is subdivided in two metamorphic complexes: the Western and the Eastern Series (Aguirre et al., 1972 and Fig. 1a). The Eastern Series consists of metagreywacke and metapelite, and is almost devoid of metabasites. It shows metamorphic gradients from very low-grade rocks to amphibolite-granulite gneisses, related to the proximity of Late Paleozoic calcalkaline granitoids (Hervé et al., 1988; Beck et al., 1991; Martin et al., 1999). Based on petrological, structural, and age characteristics, the Eastern Series is interpreted as resulting from subduction-related accretion and deformation of former passive margin sediments, followed by intrusion of arc granitoids (Hervé et al., 1988; Willner et al., 2005; Glodny et al., 2006). The Western Series comprises low-grade metapsammite, with metabasite intercalations and serpentinite, in transitional greenschist to blueschist facies. The Western Series is interpreted as a paleoaccretionary complex built by basal accretion, and the Eastern Series shows all characteristics of frontally accreted sediments (Glodny et al., 2005; Richter et al., 2007). According to this interpretation, the Western Series most probably represents the composition of the lower crust of the forearc in this region. Triassic granitoids with minor gabbros occur as scattered epizonal intrusive bodies in today's Cordillera de la Costa of Central- and South-Central Chile, between 29° and ~37°S (Fig. 1a). They intrude into the Paleozoic metamorphic and granitoid rocks. Their composition is mainly granitic, with isolated occurrences of gabbros and intermediate rocks. Vásquez et al. (2008) propose, based on the study of the magmatism of the Chilean Cordillera de la Costa (34-37°S), that the Triassic magmatism in today's forearc region could have been generated in a suprasubduction setting with slab rollback, which favors the melting of the subcontinental mantle. The coincidence of the Triassic plutonic bodies with NW-SE trending lineaments suggests that, previous to the steepening of the slab, there was oblique subduction generating extension-related NW-SE structures, which in turn facilitated the ascent of magmas (Vásquez et al., 2008).

2.1 The Cobquecura Pluton (~36°S, ~72°45'W)

The Cobquecura Pluton is an epizonal, complex intrusive body, cropping out at the western side of the Cordillera de la Costa (Fig. 1). Vásquez and Franz (2008) have previously

described it and we only give a short summary here. The Cobquecura Pluton displays a wide lithologic and textural variety compared to other Triassic intrusives of this crustal segment of the SW continental margin of Gondwana. The major rock types are gabbro and granite, and less commonly hybrid rocks. A peculiarity is that some granitoids are fayalite-bearing. The gabbros range from hornblende-bearing gabbros to olivine gabbro-norites. They are composed of plagioclase, clinopyroxene, amphibole and minor olivine, biotite and orthopyroxene. Fayalite granitoids are medium-grained and are composed of fayalite, Fe-rich clinopyroxene, Fe-rich edenite, quartz, plagioclase, annite, potassic feldspar and ilmenite. In both granitoids and gabbros, amphibole and biotite are late magmatic minerals. Medium-grained fayalite-free granitoids are composed of quartz, potassic feldspar, plagioclase and minor biotite. The suite of hybrid rocks consists of tonalite with magma mixing features (resorbed cores of plagioclase) and mingled rocks (composite dikes, mafic pillow-shaped enclaves and net-vein complexes) that contain mafic-igneous and metamorphic enclaves. Hybrid rocks generally are crosscut by both synplutonic and postplutonic dikes with acidic and basic compositions. The suite is subalkaline and transitional between metaluminous and peraluminous. The granitoids are also transitional between S- & I-type and anorogenic using the classification diagrams of Whalen et al. (1987).

Magmatic breccia occurs at the La Iglesia de la Piedra locality (Fig. 2) and is composed of matrix-embedded igneous and metapelitic xenoliths (Fig. 3 a, b). The matrix has a variable composition from amphibole-bearing tonalite to amphibole and biotite-bearing granodiorite. Magma mingling-related mafic enclaves, present as composite dikes and mafic pillow-shaped enclaves, display intergranular texture and contain plagioclase, clinopyroxene and amphibole. These enclaves result from injection of mafic magma into felsic melts (Vásquez and Franz, 2008).

The age of the Cobquecura magmatism is fairly well constrained. A K-Ar age of 212 ± 5 Ma (2σ) on ferroedenitic amphibole was obtained for a fayalite granodiorite from La Iglesia de la Piedra (Fig. 1b), and a K-Ar Mg-hornblende age of 213 ± 5 Ma (2σ) for a gabbro from Buchupureo (Fig. 1b) (Vásquez et al., 2005). LA-ICP-MS zircon U-Pb data gave similar Late Triassic ages for these intrusives, namely an age of 209 ± 2 Ma (2σ) for a gabbro from Buchupureo, while Pullay gabbro zircons yielded 203 ± 2 Ma (2σ) (Vásquez et al., 2008). Isotopic ages thus cluster around 210 Ma. Field observation, namely the evidence for magma mingling, shows that the magmas of the Cobquecura Pluton were strictly contemporaneous. This inference is backed up by major element correlations for fayalite granitoids and gabbro, suggesting a comagmatic evolution for all Cobquecura igneous rocks (Vásquez and Franz, 2008). However, the above U-Pb zircon age dates with their nominally small analytical uncertainties may incorrectly be used to infer separate magmatic events. Vásquez et al. (2008) pointed out that this discrepancy is most likely due to minor Pb loss associated with the nominally ‘youngest’ U-Pb zircon ages, which is not resolvable in LA-ICP-MS dating, and that Cobquecura magmatism actually occurred at around 210 Ma.

2.2 Country rocks

The Eastern Series (Fig. 1a) forms the low-grade metamorphic basement in the Cobquecura area. In the area around the Cobquecura Pluton, the Eastern Series is mainly composed of phyllite, with subordinate quartz-rich metapsammite. The foliation in the phyllite is defined by white mica. Other constituents are granoblastic quartz, albite and scarce tourmaline. Quartz-rich metapsammite with alternating quartz- and phyllosilicate-rich layers of white mica, chlorite and minor biotite crops out at the Iglesia de la Piedra, and is located directly above the roof of the Cobquecura Pluton. The metamorphic rocks of the area belong to the Paleozoic to Mesozoic accretionary wedge, which is exposed in many outcrops along the coast of Central and Southern Chile (e.g., Richter et al., 2007). The available isotopic age data

to this unit indicate that metamorphism occurred mainly in the Late Paleozoic (Permian-Carboniferous; Willner et al., 2005, Glodny et al., 2006, 2008, and references therein).

2.3 Characteristics of the xenoliths in the Cobquecura Pluton

The Cobquecura Pluton displays different types of xenoliths related to diverse structures within the pluton (Table 1, Figs. 3 and 4). Igneous, metapelitic and meta-igneous fragments are common in the magmatic breccia. In gabbros and fayalite granitoids, only metapelitic xenoliths were observed.

2.3.1 Metapelitic xenoliths

Different types of metapelitic xenoliths are present in the gabbros, with either upper crustal (low-grade metamorphic) or lower crustal (high-temperature metamorphic) characteristics. A first group of xenoliths is considered here as upper crustal xenoliths, as it is very similar to the local metamorphic basement at the outcrop level. These xenoliths are colored brown to gray, display a strong foliation, and show folds with segregated quartz (Fig. 3c, d). The upper crustal xenoliths contain white mica, chlorite, quartz and opaque minerals. Pinitized cordierite is found at the contact with the intrusive rocks.

A second group of metapelitic xenoliths is present both in the gabbro and the fayalite granitoids. Their mineralogy and their leucosome content indicate that they are restitic xenoliths, which, while entrained into the gabbro, preserved their original foliation and some relict leucosome. The shape of these xenoliths is stretched, indicating plastic deformation. In places there are quartz-feldspar-rich aureoles in the surrounding gabbro, indicating mass transport from the xenoliths into the gabbro (Fig. 3e, f). Regarding the leucosome content, two types of restitic xenoliths can be distinguished, one with high leucosome content (high-leucosome restitic xenoliths, HLRX) and another with low leucosome content (low-leucosome restitic xenoliths, LLRX) (see Table 1). Both show similar melanosomes with granolepidoblastic textures, abundant hercynite (ca. 30 vol.%), cordierite, corundum, plagioclase, scarce sillimanite and opaque minerals together with late biotite; the opaque minerals are mainly ilmenite with minor magnetite and Fe-sulphides (Fig. 4a, b, c). This mineralogy is typical for high temperature-low pressure metamorphism in metapelites. The HLRX occasionally display a layered structure with coarse-grained granitic leucosomes. The LLRX display a leucosome with high concentrations of zircon, with minor apatite, monazite and xenotime.

2.3.2 U-Pb data: zircon of a metapelitic xenolith

Abundant zircon in the leucosome of a low-leucosome restitic xenolith (LLRX; Fig. 4d) allowed us to date zircon in-situ by U-Pb methods using LA-ICPMS (see electronic Appendix A for methodology). Some zircon grains display inherited rounded cores, mantled by apparently magmatic overgrowths. Using data from zircon cores and mantle, a discordia line is defined with an upper intercept at $651 \pm 88/-91$ Ma and a lower intercept at $202 \pm 16/-19$ Ma (Fig. 5; analytical data in electronic Appendix B). We interpret the cluster of concordant age data at about 200 to 210 Ma (Fig. 5, inset) as the crystallization age of the last magmatic zircon generation. Obviously, the zircon population of the metasedimentary precursor rock of the xenolith was strongly affected by the magmatic processes of Cobquecura plutonism at around 210 Ma. The upper intercept age suggests presence of inherited Neoproterozoic zircon. For metapelitic rocks comparable to the country rocks of the Cobquecura pluton, presence of Neoproterozoic detrital zircon is common for Coastal Central Chile (Willner et al., 2008).

2.3.3 Xenoliths of the magmatic breccia

The magmatic breccia contains fragments of metapelite, quartzite, monzogabbroic xenoliths (MX) and meta-microgabbroic xenoliths (MMX), mostly with cusped forms, (Table 1, Fig. 3a, b), in a tonalitic to granodioritic matrix. MMX show ophitic to subophitic texture with plagioclase and clinopyroxene, together with recrystallized amphibole, orthopyroxene and biotite (Table 1). Metapelitic xenoliths with lepidogranoblastic texture (Fig. 3a, b) are similar to high-leucosome restitic xenoliths (HLRX, Table 1).

3. Geochemical characteristics

3.1 Intrusive rocks

The Cobquecura Pluton comprises a subalkaline suite with a SiO₂ content between 48 and 76 wt.% and a small compositional gap between 57 % and 63 % wt.% SiO₂ (Vásquez and Franz, 2008). The most differentiated rocks are leucocratic dikes and granitoids without fayalite. Compositional continuity exists between metaluminous mafic rocks and slightly peraluminous, more differentiated rocks richer in K₂O and with a higher FeO*/(FeO*+MgO) (total iron as FeO; Vásquez and Franz, 2008).

The different rocks of the Cobquecura Pluton all have gently sloped rare-earth element (REE) patterns, with a steeper slope in the LREE (La_N/Sm_N = 1.1 - 3.4) than in the HREE (Gd_N/Lu_N = 1.2 - 1.7; Fig. 6). The HREE patterns are almost parallel, whereas LREE patterns gradually increase their slopes with increasing SiO₂-content, which is typical for fractional crystallization processes (Fig. 6a). Consequently, REE patterns of the mafic rocks are more gently sloped (La_N/Yb_N = 1.9 - 3.7) than those of the felsic rocks (La_N/Yb_N = 4 - 8).

The total REE-contents of the fayalite granitoids are higher than those of the fayalite-free granitoids, despite their lower SiO₂-content compared to the fayalite-free granitoids (Fig. 6a, b).

All felsic rocks consistently have negative Eu anomalies (Eu/Eu* = 0.3 - 0.9), which are markedly less pronounced in the fayalite granitoids despite of their higher overall REE-contents (Fig. 6a, b). Gabbros have no or only a weakly negative Eu anomaly (Eu/Eu* = 0.9 - 1.0; Fig. 6a), whereas gabbro-norite displays a positive Eu anomaly (Eu/Eu* = 1.4 - 1.6).

In primitive-mantle normalized multielement diagrams, all rocks of the Cobquecura Pluton show positive Pb and negative P anomalies and display only a weak Nb-Ta trough (Fig. 7). The felsic rocks are richer in incompatible elements (e.g., Rb, Th, and U) than the mafic rocks and display strong negative Sr- and Ti-anomalies. Complementary Eu-, Sr- and Ba-anomalies point to feldspar fractionation, whereas negative Ti-anomalies point to ilmenite fractionation (Fig. 7a, b). The mafic rocks have weak positive Ba-, Sr- and Ti-anomalies that may indicate accumulation processes of feldspar and ilmenite in the mafic rocks. Among the felsic rocks, the fayalite granitoids show the smallest Eu-, Sr- and Ti-anomalies and among the intrusive rocks they are the richest in compatible elements (e.g., HREE) and high field strength elements, HFSE (Zr, Nb and Y).

Most of the variation diagrams of trace elements show well-defined trends for the whole suite of magmatic rocks (Fig. 8). U, Th, and Pb display a positive correlation with SiO₂ content. Nb increases with increasing SiO₂ content up to 57 wt.-% (compositional gap) and decreases from 63 wt.% SiO₂ content. Zr vs. Nb and Y vs. Nb show positively correlated trends with Y-, Zr-, and Nb-poor gabbroic rocks and Y-, Zr-, and Nb-rich felsic rocks (Fig. 8).

3.2 Metamorphic basement and xenoliths

The metamorphic basement in the Cobquecura area consists of metapelite and quartz-metapsammite (which are present and easy to recognize in the intrusive rocks as upper crustal xenoliths) with SiO₂ contents between 58 and 73 wt.%, Al₂O₃ of 13 - 17 wt.%, Fe₂O₃ of 3.9 - 8.5 wt.%, TiO₂ of 0.7 - 1.0 wt.% and K₂O of 3 - 3.5 wt.% (Table 2). The REE concentrations range between 163 and 203 ppm with uniform REE patterns. They display La_N/Yb_N = 7.9 - 8.5, combined with negative Eu anomalies (Eu/Eu* = 0.6 - 0.7). The trace element patterns

normalized to the average composition of continental crust are also rather uniform for most samples and show negative anomalies of Ba, Nb, Ta, Sr, and Ti (Fig. 7c). The trace element signatures of the basement around Cobquecura (36°S), of the regional metamorphic basement south of 37°S (Figs. 6c, 7c), and of North American Shale Composite (NASC, Gromet et al., 1984, not shown) are generally similar.

Restitic xenoliths in gabbro and fayalite granitoid (HLRX and LLRX) contain 46 to 56 wt. % SiO₂, with notably higher content of Al₂O₃ (20 to 27 wt. %), Fe₂O₃ (7.7 to 11.1 wt. %), TiO₂ (0.9 to 1.6 wt. %) and K₂O (1.2 to 6.3 wt. %) than the metamorphic basement around Cobquecura (Table 2). These metapelitic-restitic xenoliths also are richer in compatible elements (e.g., HREE), in HFSE (Y, Zr, and Nb), and in Th and Pb (14 - 22 and 12 - 26 ppm, respectively) than the regionally exposed basement. Restitic xenoliths have high REE contents (218 - 378 ppm), with steeper slopes in the REE patterns ($La_N/Yb_N = 7.4 - 11.4$) than the exposed metamorphic basement, and a small negative Eu anomaly ($Eu/Eu^* = 0.71 - 0.72$). Their contents of U, Pb, Nb, various compatible elements, and HFSE are similar to those of fayalite granitoid (Figs. 7 and 8). The LLRX have less SiO₂, K₂O, and Rb, and are enriched in Al₂O₃, Fe₂O₃, TiO₂, REE, and HFSE compared to the HLRX.

4 Pb, Nd, and Sr isotopic composition

We analyzed 22 whole rock samples from the Cobquecura Pluton, its different enclaves and xenoliths, and the surrounding metamorphic basement for Nd-Sr-Pb isotopic composition. Results are presented in Table 3 and in Figs. 9 and 10. Detailed analytical procedures are presented in the electronic appendix A.

4.1 Intrusive rocks

⁸⁷Sr/⁸⁶Sr_i (at 210 Ma) ratios of Cobquecura Pluton igneous rocks range from 0.7035 to 0.708 with distinctly positive to slightly negative ϵNd_i between 6.8 and -1.9 (Fig. 9, Tab. 3). ⁸⁷Sr/⁸⁶Sr_i ratios of gabbros (0.7035 - 0.7055) are well correlated with the SiO₂-content (Fig. 11a); their ϵNd_i values range from 6.8 to -0.2 (Fig. 9). Fayalite granitoids have ⁸⁷Sr/⁸⁶Sr_i near 0.707 and ϵNd_i from -0.6 to -0.7, similar to the more differentiated gabbroic rocks (Fig. 9), whereas the fayalite-free granitoids have slightly more negative ϵNd_i values (-1.0 to -1.9), but display similar ⁸⁷Sr/⁸⁶Sr_i ratios (0.707 - 0.708). ⁸⁷Sr/⁸⁶Sr_i and ϵNd_i ratios in fayalite granitoids and fayalite-free granitoids display no correlation with SiO₂ (Fig. 11a, b).

Pb initial ratios of the magmatic rocks (Tab. 3) show an overall range from 18.26 - 18.70 (²⁰⁶Pb/²⁰⁴Pb_i), 15.60 - 15.66 (²⁰⁷Pb/²⁰⁴Pb_i), and 38.15 - 38.76 (²⁰⁸Pb/²⁰⁴Pb_i), respectively. They plot below the Andean Pb-line as defined by Lucassen et al. (2002) that shows the evolution of the average Andean crust (Fig. 10). Gabbros, fayalite granitoids, and fayalite-free granites display similar ²⁰⁷Pb/²⁰⁴Pb_i of 15.60 - 15.66, and ²⁰⁸Pb/²⁰⁴Pb_i of 38.15 - 38.76. Gabbros, fayalite-free granites and fayalite-free granitic dikes show more radiogenic ²⁰⁶Pb/²⁰⁴Pb_i ratios (18.37-18.70) than the fayalite granitoids (²⁰⁶Pb/²⁰⁴Pb_i = 18.26-18.28). Pb isotopes of the entire suite are not correlated with the SiO₂-content (not shown).

4.2 Metamorphic basement and xenoliths

The metamorphic basement in the Cobquecura area displays high ⁸⁷Sr/⁸⁶Sr_(210 Ma) ratios near 0.729, combined with strongly negative $\epsilon Nd_{(210 Ma)}$ values (-7.9 to -8.1) (Fig. 9). It shows a range of ²⁰⁶Pb/²⁰⁴Pb_(210 Ma) ratios from 18.33 to 18.38, of ²⁰⁷Pb/²⁰⁴Pb_(210 Ma) ratios from 15.62 to 15.63, and of ²⁰⁸Pb/²⁰⁴Pb_(210 Ma) ratios from 38.33 to 38.47 (Fig. 10, Tab. 3).

The two types of restitic xenoliths show initial Sr isotopic compositions, which are similar to each other. ⁸⁷Sr/⁸⁶Sr_i at 210 Ma are 0.7177 for low-leucosome restitic xenolith (LLRX) and 0.7185 for a high-leucosome restitic xenolith (HLRX). These values are significantly higher than those of Cobquecura plutonic rocks. Both types of restitic xenoliths show strongly negative $\epsilon Nd_{(210 Ma)}$ values of -6.8 to -6.3, which is only slightly less negative than in the

surrounding regional metamorphic basement. The Pb isotopic composition of HLRX is very similar to that of the regional basement, whereas the LLRX present a slightly lower initial $^{206}\text{Pb}/^{204}\text{Pb}$ ratio of 18.32 and a markedly lower $^{208}\text{Pb}/^{204}\text{Pb}$ ratio (37.78) compared to the regional basement (Fig. 10).

The igneous microgabbroic xenoliths enclosed by the magmatic breccia present Sr-Nd-Pb compositions that fall in the range of those of the gabbroic rocks ($^{87}\text{Sr}/^{86}\text{Sr}_{(210\text{ Ma})} \sim 0.706$ to 0.707 , $\epsilon\text{Nd}_{(210\text{ Ma})} = 1.4 - 2.4$, $^{206}\text{Pb}/^{204}\text{Pb}_{(210\text{ Ma})} = 18.45 - 18.48$, $^{207}\text{Pb}/^{204}\text{Pb}_{(210\text{ Ma})} \sim 15.63$, and $^{208}\text{Pb}/^{204}\text{Pb}_{(210\text{ Ma})} = 38.40 - 38.43$).

5 Discussion

5.1 Nature of the continental crust

Late Paleozoic granitoids and metapelites form the upper crust at the present outcrop level of the Central Chilean continental margin. The formation of the upper crust in this segment is mainly related to recycling of Proterozoic continental basement in a continental margin setting, with a resulting composition close to the average of the old continental crust of SW Gondwana (as described by Lucassen et al., 2004). This upper crust is isotopically and lithologically relatively homogeneous along the Chilean continental margin (Lucassen et al., 2004). The regional granitoid crust displays on average higher ϵNd_i and lower $^{87}\text{Sr}/^{86}\text{Sr}_i$ compared to values of the metapelitic crust (Fig. 9). Initial Pb isotopic ratios (at 210 Ma) of the regional granitoid crust are generally similar to those of the regional metapelitic crust but show more restricted ranges (Fig 10). The regional metamorphic metapelitic crust as constrained by Lucassen et al. (2004) displays a homogeneous and narrow composition in Pb isotopes ($^{206}\text{Pb}/^{204}\text{Pb}_{(210\text{ Ma})} = 18.32 - 18.70$, $^{207}\text{Pb}/^{204}\text{Pb}_{(210\text{ Ma})} = 15.65 - 15.73$, and $^{208}\text{Pb}/^{204}\text{Pb}_{(210\text{ Ma})} = 38.2 - 38.8$), shows distinctive $\epsilon\text{Nd}_{(210\text{ Ma})}$ ratios of -5 to -9, and displays a wide range of $^{87}\text{Sr}/^{86}\text{Sr}_i$ ratios (Figs. 9 and 10). The local metapelitic crust near the Cobquecura pluton has similar petrologic, geochemical, Sr-Nd, $^{208}\text{Pb}/^{204}\text{Pb}_i$ and $^{206}\text{Pb}/^{204}\text{Pb}_i$ characteristics and displays slightly lower $^{207}\text{Pb}/^{204}\text{Pb}_i$ ratios compared to the rest of the metamorphic basement along the Chilean margin south of 37°S, although the isotopic variability of the local samples is much smaller (Fig. 10).

The restitic xenoliths (both high-leucosome HLRX and low-leucosome LLRX) of the Cobquecura Pluton have distinctive petrographical, geochemical and Sr and Nd isotopic characteristics with respect to the metamorphic basement neighboring the Cobquecura Pluton as well as to the xenoliths that come from this basement. Differences in the Pb isotopic composition among these samples are not significant, with the exception of a remarkably low initial $^{208}\text{Pb}/^{204}\text{Pb}$ ratio of LLRX (Fig. 10). The restitic xenoliths (HLRX and LLRX) are rich in hercynite and cordierite, a typical paragenesis for metapelites at upper amphibolite to granulite facies conditions. The association cordierite – hercynite – quartz, without garnet, is stable at high temperatures at pressures of 2 to 8 kbar (Bucher and Frey, 2002; Spear, 1993). Some of these restitic xenoliths preserve layers of granitic leucosome, reflecting a stage of anatectic melting. The compositional layering with both hercynite and cordierite, which follow the main foliation, indicates that these minerals were formed by regional rather than by intrusion-related contact metamorphism. The low leucosome content in LLRX suggests that these xenoliths experienced a stage of melt extraction, with a melt composition similar to the leucosome of the HLRX. Decreasing leucosome-content from the HLRX to the LLRX is correlated with increasing REE, Th, and HFSE contents, and decreasing Rb, K₂O, and Pb. This pattern is probably related to the early feldspar melting during leucosome formation, and subsequent extraction of felsic melt. Furthermore, the high contents of Th, HFSE and REE in the LLRX with scarce leucosome are related to the presence of zircon, apatite, monazite and xenotime in the leucosome.

The restitic xenoliths (HLRX and LLRX) have, compared to the average value for metapelitic xenoliths metamorphosed at granulite or amphibolite facies conditions (Rudnick and Fountain, 1995), similar compatible element and Rb contents, but are enriched in Th. The xenoliths display Sr-Nd-Pb initial isotopic ratios similar to those of the local and regional Cobquecura crust (Figs. 9 and 10). However, the xenoliths display high $\text{La}_\text{N}/\text{Yb}_\text{N}$ ratios (Figs. 6 and 11), and are richer in FeO, Al_2O_3 , TiO_2 , Th, REE, and HFSE than the surrounding metapelitic basement. These geochemical differences, together with the higher metamorphic grade and the restitic nature of the xenoliths, suggest that they were not derived from the surrounding metamorphic basement, but from a “deeper level” with a different history, probably from the middle crust. The enrichment combined with low initial $^{208}\text{Pb}/^{204}\text{Pb}_{(210\text{ Ma})}$ ratio particularly of LLRX xenoliths (Fig. 10) indicates that the restitic xenoliths do not represent deep crust that had experienced Proterozoic or older high-grade metamorphism but that the present geochemical signatures were acquired in the Phanerozoic.

5.2 Mantle-derived melts and crustal influence

The most primitive rocks in the Cobquecura Pluton are gabbro-norites. They display the highest ϵNd_i , the lowest $^{87}\text{Sr}/^{86}\text{Sr}_i$ ratios and show the most depleted incompatible element signatures of the entire suite. Their Nd-Sr isotope signature is similar to that of other primitive magmatic rocks along the Chilean Mesozoic arc, which are considered to reflect the composition of the subarc subcontinental mantle including additions from the subduction zone (Lucassen et al., 2006). The subarc subcontinental mantle shows lower ϵNd_i and higher $^{87}\text{Sr}/^{86}\text{Sr}_i$ values than the asthenospheric Pacific oceanic mantle, the source of MORB (source: <http://www.petdb.org>) at 210 Ma (Fig. 9). The increase of $^{87}\text{Sr}/^{86}\text{Sr}_i$ ratios from 0.703 to 0.706 with ϵNd_i decreasing from 6.4 to -0.2 indicates contamination of the mantle-derived gabbroic melts by significant amounts of continental crust. Accordingly, the gabbroic rocks display higher initial $^{206}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$ ratios than the subarc subcontinental mantle and Pacific MORB at 210 Ma. The absence of negative Eu- and Sr-anomalies and the low ratios of $\text{La}_\text{N}/\text{Yb}_\text{N}$ (1.7 - 3.3) in the mafic rocks suggest that neither plagioclase nor garnet were residual phases in their source (Vásquez et al., 2008). The LILE enrichment, the positive Pb anomaly and the negative Nb-Ta trough present in both gabbros and granitoids (Fig. 7) are typical of magmatic rocks generated in continental magmatic arcs (e.g., Hawkesworth et al., 1993).

5.3 Petrogenesis of the Cobquecura Pluton

Previous models have explained fayalite granitoids as granitic residual melts from the fractional crystallization of mafic magmas (i.e., Pitcher, 1997; Frost and Frost, 1997; Turner et al., 1992). Another hypothesis is that fayalite granitoids originate from partial melting of a dehydrated source, like granulites of the lower continental crust (e.g., Collins et al., 1982). Possible regional sources for the formation of the fayalite granitoids of the Cobquecura Pluton are the subarc-subcontinental mantle, Late Paleozoic upper continental crust (granitic and metamorphic) or a deeper level crust represented by the restitic xenoliths described above. It is noteworthy that the restitic xenoliths (HLRX and LLRX) were not found in fayalite-free granitoids. Based on the field, petrographic and geochemical characteristics, possible processes contributing to fayalite granitoid's petrogenesis are fractional crystallization and assimilation of continental crust. These processes and the role of the possible sources in the generation of the fayalite granitoids magmas are evaluated in the next sections.

5.3.1 Evidence for fractional crystallization

Major and trace element arrays demonstrate the importance of fractional crystallization and accumulation processes. CaO, K_2O and MgO display coherent arrays in Harker diagrams and X_Mg of mafic minerals displays an excellent correlation with X_Mg of the corresponding whole

rock (Vásquez and Franz, 2008). Fractional crystallization processes are also indicated by the correlations of trace element contents such as Y and Nb (Fig. 8g) and the complementary Ba-Eu- and Sr-anomalies displayed by the gabbros and granites in the normalized multielement diagrams and the REE patterns, respectively (Figs. 6 and 7). An exclusive origin of the fayalite granitoids by fractional crystallization from the primitive mafic magmas can be ruled out by their different Sr-Nd isotope signatures (Fig. 9). However, it is remarkable that the fayalite granitoid and the more contaminated gabbros display similar ϵNd_i (Fig. 9), consistent with an origin of this granitoid from the more contaminated gabbros. For pure fractional crystallization processes, parallel patterns with increasing contents of REE and more pronounced negative Eu-anomalies with increasing differentiation could be expected. However, although the REE-patterns are fairly parallel (Fig. 6), fayalite-bearing granitoids (63 - 71 wt. % SiO_2) have higher REE contents and smaller negative Eu-anomalies than the fayalite-free granitoids (67 - 76 wt. % SiO_2) (Fig. 6). Therefore, the REE patterns disagree with an origin of the fayalite granitoids from closed-system fractional crystallization from a mafic magma. Instead, the fayalite granitoids must have originated from open-system fractional crystallization from a gabbroic, mantle-derived magma with associated assimilation processes.

5.3.2 Assimilation processes

The Cobquecura Pluton displays abundant field evidence for assimilation of metapelitic crust, such as metapelitic xenoliths with quartz-feldspar rich assimilation aureoles (Fig. 3). The average Sr-Nd isotopic composition of the local upper crust at the outcrop level is ϵNd_i of ~ -8 , and $^{87}\text{Sr}/^{86}\text{Sr}_i$ of 0.729. The ϵNd_i values near -6.5 and $^{87}\text{Sr}/^{86}\text{Sr}_i$ of ~ 0.718 of the restitic xenoliths are intermediate between those of the local upper crust and those of Cobquecura igneous rocks ($\epsilon\text{Nd}_i > -2$, $^{87}\text{Sr}/^{86}\text{Sr}_i < 0.707$) (Fig. 9, Table 3). This indicates that the above crustal lithologies are not the only and apparently not even the dominant sources for Cobquecura granitoids. The most primitive mafic rocks of the Cobquecura Pluton have compositions of ϵNd_i of 6.8, $^{87}\text{Sr}/^{86}\text{Sr}_i$ of 0.7035, close to the regional sub-arc mantle. In contrast, initial Pb isotopic ratios are similar to those of the regional continental crust (Fig. 10). These characteristics point to a mantle source of the gabbros, with modification of the magmas during ascent by assimilation of minor amounts of crust characterized by more radiogenic lead. The resulting signatures in Sr, Nd, and Pb are due to the contrasting contents of these elements in the mantle and the crust. Minor assimilation of crustal material will only slightly alter the Nd and Sr isotopic signatures of mantle-derived melts, whereas Pb isotopic signature will entirely be dominated by the Pb-rich crustal material. The Pb-Nd-Sr isotopic compositions of the Cobquecura Pluton demonstrate that none of the reservoirs upper crust, deeper crust and mantle could serve as sole source for the magmas.

The good correlation between the $^{87}\text{Sr}/^{86}\text{Sr}_i$ ratios and SiO_2 contents (and La_N/Yb_N) for the gabbroic rocks (Fig. 11) also suggests that fractional crystallization processes in the magma chamber were accompanied by assimilation of ambient crust, with a composition similar to the LLRX leucosome.

As to the more differentiated lithologies, the fayalite bearing granitoids and the fayalite-free granitoids display obvious differences. Among these are the modal composition and the content of Al- and Fe-rich restitic xenoliths. Further, fayalite granitoids show higher REE and HFSE contents (Figs. 6, 7, and 8), and their H_2O -poor, Fe-rich mineralogy points to lower $f\text{O}_2$ and H_2O activity. This shows that assimilation of Fe-rich rocks, which are compositionally similar to the Al- and Fe-rich restitic xenoliths, in the middle or lower continental crust could be an important process in the petrogenesis of the fayalite granitoids. In contrast, the relatively elevated $a_{\text{H}_2\text{O}}$ and $f\text{O}_2$ of the fayalite-free granitoids suggest an assimilation of a more hydrated crust, similar to the upper crust at the outcrop level. The fayalite granitoids crystallized at oxygen fugacity below the fayalite-magnetite-quartz buffer (Vásquez and

Franz, 2008), which suggests a major influence of reduced sources. Both the primitive mafic rocks (with their presence of sulfide phases) and restitic xenoliths (HLRX and LLRX, with their occasional presence of sulfide, and with the Fe²⁺-rich minerals hercynite and cordierite) display indications for low oxygen fugacity, which is consistent with the low oxygen fugacity recorded by the fayalite granitoids (this work, and Vásquez and Franz, 2008).

5.3.3 Selective assimilation processes

Anatectic processes are indicated by the presence of leucosomes in the restitic xenoliths, and may facilitate selective assimilation processes (c.f. Hogan and Sinha, 1991, Zeng et al., 2005). Selective assimilation of ambient crust is suggested by field evidence (Fig. 3), and is also indicated by the chemical and isotopic behavior of the Cobquecura rocks. The differences between leucosome compositions of the different restitic xenoliths in the Cobquecura Pluton define two types of leucosome: a feldspar-rich, accessory minerals-poor leucosome of the HLRX, and a feldspar-poor, accessory mineral-rich leucosome of the LLRX.

The leucosome of the HLRX is generated in an anatectic process and represents an initial and feldspar-quartz rich melt, with a zircon- and monazite-rich residuum. Fayalite-free granitoids could have been generated through the assimilation of a melt similar to the leucosome of the HLRX. The paragenesis of potassic feldspar, plagioclase and quartz suggests a low melt temperature similar to the temperature of eutectic melting in water-rich granitic systems, i.e. ca. 650°C (e.g., Holtz et al. 2001). After this first-stage melt extraction processes and with an increasing degree of melting, the ‘later’ melt will carry less feldspar and more zircon and monazite, as it is observed in the leucosome of the LLRX. The modal differences in feldspar, zircon, apatite, monazite and xenotime between the leucosomes of the HLRX and LLRX lead to differences in the Rb, K₂O, Pb, Th, HFSE, and REE contents in the different restitic xenoliths included in gabbros and fayalite granitoids. Progressive anatectic melting also has the potential to induce changes in Pb, Sr and Nd isotopic ratios in the leucosomes and their corresponding restites, as outlined below.

The initial Pb isotopic compositions of anatectic melts are defined by a mixture of Pb derived from major minerals with characteristic low ²³⁸U/²⁰⁴Pb and ²³²Th/²⁰⁴Pb ratios (i.e. feldspar) with Pb contributed from dissolved accessory minerals with characteristically high ²³⁸U/²⁰⁴Pb and ²³²Th/²⁰⁴Pb ratios (i.e. zircon, monazite) (Hogan and Sinha, 1991). Similarly, the Sr isotopic composition of the leucosomes may also change with progressive melting. Radiogenic Sr is associated with Rb-rich phases, mainly with micas and alkali feldspar. Dehydration melting, resulting in the decomposition of the micas and feldspar would result in a relatively radiogenic Sr isotopic composition in the leucosome and a less radiogenic Sr isotopic composition in the restite. With progressive melting, the bulk composition of the leucosome will become less radiogenic once the major Rb carriers have been melted, and low-Rb/Sr phases increasingly contribute to the leucosome. In a similar way, the preferential melting of minerals with a steep REE pattern (e.g., feldspars) will transfer Nd with lower εNd_(210 Ma) values into the melt than that of the bulk rock or the residue. Thus, the restite will have higher εNd_(210 Ma) values than the original bulk rock, possibly depending on the degree of melting.

The removal of leucocratic melts will affect both the chemical and isotopic character of the residue. Thus, the contrasting chemical and isotopic fingerprints of the metapelitic low-grade basement and the restites do not necessarily imply that different sources were involved. Instead, it implies that the restitic xenoliths have been modified by melt extraction. Furthermore, the component added to the mantle-derived melts by selective melting from metapelites carries an isotopic and geochemical signature that is complementary to that of the restites. In the Cobquecura Pluton, LLRX with low leucosome content and higher accessory mineral contents display slightly less radiogenic initial Pb isotopic compositions (with respect to ²⁰⁸Pb/²⁰⁴Pb_(210 Ma) and ²⁰⁶Pb/²⁰⁴Pb_(210 Ma); Fig. 10) than the HLRX with high leucosome

content and lower accessory mineral contents. Furthermore, the restites have less radiogenic Sr isotopic compositions than the metapelites of the basement.

The above considerations show that the elemental and isotopic composition of the restitic xenoliths does not necessarily represent the composition of the crustal contaminant of the Cobquecura mafic magmas. This poses difficulties to identify the exact nature and composition of the lower crustal domain from which the restitic xenoliths originate, and is a severe complication in quantitative modeling of the differentiation-assimilation history.

In summary, geochemical and Sr-Nd-Pb isotopic evidence indicates that the fayalite granitoids formed by means of an open-system fractional crystallization from a gabbroic, mantle-derived magma, combined with selective assimilation of continental crust during differentiation. Fayalite granitoids have comparatively low $^{208}\text{Pb}/^{204}\text{Pb}_{(210\text{ Ma})}$ and $^{206}\text{Pb}/^{204}\text{Pb}_{(210\text{ Ma})}$ in common with the LLRX xenoliths. This may point to a major role of leucosome lost from the LLRX in the formation of the fayalite granitoids. This is consistent with the relatively high content of REE and HFSE elements in the fayalite granitoids, which corresponds to the high concentration of these elements in the preserved leucosomes of LLRX xenoliths.

5.4 Comparison with other fayalite granitoids worldwide

Fayalite granitoids from Cobquecura occur in an unusual extensional, subduction-related geodynamic setting (Vásquez et al., 2008). This tectonic setting differs from the typical intracratonic setting of Precambrian fayalite bearing granitoids and from the extensional setting of most Phanerozoic fayalite granitoids. Worldwide, the Precambrian fayalite-bearing granitoids are commonly closely associated with rapakivi granites, anorthosite – mangerite – charnockite - granite suites (AMCG), and the Phanerozoic fayalite granitoids to anorogenic ring complexes (e.g., Frost et al., 2002; Poitrasson et al., 1995).

The petrogenesis of AMCG suites has been subject of discussion for a long time and its relation with fayalite granitoids is still problematic owing to the suggested different origins for the fayalite granitoids in the different localities where they occur. One of the best-studied fayalite-bearing suites is the Laramie Anorthosite Complex (LAC) in Wyoming (Frost and Frost, 1997, Frost et al., 2002; Anderson et al., 2003). Anderson et al. (2003) conclude that these fayalite granitoids were formed by the interaction of a hot, dry ferrodioritic magma derived by fractional crystallization or partial melting from a tholeiitic parent, with assimilated felsic crust. This ferrodioritic magma had high $\text{FeO}^*/(\text{FeO}^*+\text{MgO})$ ratios and high REE and Zr contents.

Crystallization conditions, mineralogical association, trace element and Sr-Nd initial isotope ratios of fayalite granitoids from Cobquecura are very similar to those from the Sherman Batholith and Mule Creek fayalite granitoids (associated with the LAC, Frost et al., 2002). In the Cobquecura Pluton there is no evidence of a residual ferrodiorite, but restitic xenoliths also present high $\text{FeO}^*/(\text{FeO}^*+\text{MgO})$, are rich in REE and Zr, and display a signature of low oxygen fugacity. Our data show that a deep crust, depleted fragments of which are represented by the restitic xenoliths, has contributed to the magmas of the fayalite granitoids. Probably the selective assimilation of this metapelitic deep crust has enhanced the REE and HFSE contents and $\text{FeO}^*/(\text{FeO}^*+\text{MgO})$ ratios in the fayalite granitoids.

A common aspect of all models for fayalite granitoid's genesis is involvement of a lithology with low oxygen fugacity (mafic rocks, ferrodiorite, granulitic felsic lower crust). This lithology is required as a chemical buffer to keep the redox condition below or near to the FMQ buffer, which together with low $a_{\text{H}_2\text{O}}$ favor the generation of fayalite-bearing granitoids.

6 Concluding remarks

The Cobquecura Pluton displays comagmatic and genetically related fayalite granitoids and gabbroic rocks. Both lithologies contain restitic metapelitic xenoliths. Our data show that the

important processes involved in the generation of the fayalite granitoids were fractional crystallization combined with assimilation. Major element and trace element data show that an open-system fractional crystallization from a mafic magma with associated selective assimilation were the key processes in the generation of the granitoids of Cobquecura. The Sr-Nd isotope signature of the most primitive mafic magma is similar to that of the regional subarc-subcontinental mantle. The isotopic data show that the assimilated crustal material is strikingly similar to the deep-crustal restitic metapelitic xenoliths which are observed both within gabbros and fayalite granitoids. Different degrees and styles of assimilation may explain why there are fayalite-free and fayalite-bearing-granitoids. We propose that fayalite-free granitoids formed by assimilation of melts with overall characteristics of the granitic leucosome of the high-leucosome restitic xenoliths (HLRX), whereas the composition of the fayalite granitoids is best explained by assimilation of melts resembling the accessory-mineral-rich leucosome of the low-leucosome restitic xenoliths (LLRX).

Fayalite granitoids from Cobquecura occur in a subduction-related extensional geodynamic setting (Vásquez et al., 2008), which is atypical and differs from the usual intracratonic and extensional setting where most of the fayalite granitoids worldwide occur. From comparison of existing models for fayalite granite genesis with our observations it appears that the important and common feature critical for fayalite granitoids generation is the presence of a lithology with low oxygen fugacity (certain mafic rocks, ferrodiorite, or, as in our case, a certain type of lower crust), which have a role as a chemical buffer and keep the oxygen fugacities near to the FMQ buffer. In the case of the Cobquecura Pluton this low oxygen fugacity component most likely had a composition close to that of the restitic xenoliths present in the mafic rocks and the fayalite granitoids.

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Figures

Figure 1: a) Distribution of late Triassic-early Jurassic magmatism in the Cordillera de la Costa between 29° and 37°S; compiled after Godoy (1970), Hervé and Munizaga (1978), Dávila et al. (1979), Gajardo (1981), Gana and Hervé (1983), Mpodozis and Kay (1992), Parada et al. (1999) and Willner et al. (2004). LF: Lanalhue Fault (Glodny et al. 2008), BBFZ: Bío-Bío Fault Zone (Rehak et al. 2008); IL: Itata Lineament; MA: Melipilla Anomaly (Yañez et al. 1998). Western and Eastern Series distribution after Hervé et al. 2007. b) Geological sketch map of the Cobquecura Pluton (after Vásquez and Franz, 2008) and sample location. Isotopic ages are taken from Vásquez et al. (2005) and Vásquez et al. (2008). The arrows show the localities with fayalite granite.

Figure 2: a) Detailed map of the Punta Iglesia de la Piedra area. b) "La Iglesia de la Piedra Profile" that shows the contact relationship between the gabbros, magmatic breccia and fayalite granitoids.

Figure 3: Different types of enclaves and xenoliths. a) and b) image and drawing of different xenoliths in the magmatic breccia. Metapelitic xenoliths (similar to high-leucosome restitic xenoliths, HLRX) display abundant leucosome and a quartz-feldspar rich reaction rim. Igneous enclaves show indications for interaction with surrounding melts. c) and d) image and drawing of a large metapelitic xenolith (upper crustal xenolith, resembling the local near-surface metamorphic basement) included in gabbro, showing a strong foliation of the xenolith with folding and a thin quartz-feldspar rich – reaction rim (indicating partial assimilation). e) and f) image and drawing of a low leucosome restitic xenolith (LLRX) in gabbro, displaying a quartz-feldspar rich – reaction rim (assimilation), and relict leucosome in the xenolith. The line indicates the direction of the foliation.

Figure 4: Textures and minerals of Cobquecura Pluton xenoliths. a) Xenoliths in a granodioritic to tonalitic matrix (high leucosome restitic xenolith, HLRX) with a granitic leucosome of quartz, altered potassic feldspar, and biotite, side by side with restitic melanosomes with strong foliation with late biotite, hercynite and corundum (sample 04-31). b) Strongly foliated xenolith in gabbro (low leucosome restitic xenolith, LLRX, cordierite-rich restite with hercynite, sample 04-07). c) Foliated xenolith in gabbro (LLRX) composed of hercynite-rich restite with cordierite and late biotite (sample 07-44). d) Leucosome from the LLRX, with coarse-grained quartz, plagioclase, biotite, and zircon (sample 07-45). Abbreviation: Bt = biotite, Cd = cordierite, Hc = hercynite, Kfs = potassic feldspar, Pl = plagioclase, Qtz = quartz, Zrn = zircon.

Figure 5: Results of *in situ* LA-ICP-MS U-Pb age determinations on zircon from leucosome of a low-leucosome restitic xenolith (LLRX, sample 07-45). Age data for the youngest zircon generation cluster at around 210 to 200 Ma (see inset), which is the magmatic age of the Cobquecura pluton. See electronic Appendix B for data.

Figure 6: Chondrite-normalized rare earth element spectra (cf. McDonough and Sun, 1995), Cobquecura Pluton and environs: a) gabbros (48-55% wt. SiO_2) and fayalite-free granitoids (67-76% wt. SiO_2), b) fayalite granitoids (63-72% wt. SiO_2) and c) metamorphic basement in the Cobquecura area and metapelite xenoliths. The gray area represents the compositions of the Paleozoic metamorphic basement of the Cordillera de la Costa south of the 37°S (adapted from Lucassen et al. 2004).

Figure 7: Primitive mantle-normalized (McDonough and Sun, 1995) trace element diagrams for a) Gabbros (48-55% wt. SiO_2) and fayalite-free granites (67-76%wt. SiO_2) b) Fayalite granites and c) Continental crust-normalized (Taylor and McLennan, 1995) multielement diagrams for rocks from the metamorphic basement at the Cobquecura area and metapelite xenoliths. The gray area represents the compositions of the Paleozoic metamorphic basement at the Cordillera de la Costa south of the 37°S (adapted from Lucassen et al. 2004).

Figure 8: Correlations of trace element concentrations with SiO₂, and trace element intercorrelations in the Cobquecura intrusive suite.

Figure 9: Initial Nd and Sr isotopic compositions (at 210 Ma), Cobquecura Pluton and related rocks. Fields are shown for the regional metapelitic crust and the regional granitoid crust (Southern Coastal Batholith 30-38°S and metamorphic basement of the Chilean Coastal Cordillera 36°- 41°S; Parada et al. 1999, Lucassen et al. 2004), for the subarc subcontinental mantle of northern Chile (18°-27°S; Lucassen et al. 2006), and for average of the Pacific MORB (Petrological database of the ocean floor <http://petdb.org>); the Nd isotope composition is corrected for in-situ decay to 210 Ma using a $^{147}\text{Sm}/^{144}\text{Nd}$ ratio of 0.214). The Sr isotope composition of Bulk Silicate Earth (BSE) is calculated at 210 Ma assuming a Rb/Sr ratio of 0.03 (McDonough and Sun, 1995) and a present day $^{87}\text{Sr}/^{86}\text{Sr}$ of 0.7045 (Salters and Stracke, 2004). Mantle array field was taken from Dickin, 1997.

Figure 10: Initial Pb isotopic composition of the Cobquecura Pluton. Average of the continental crust (Coastal Cordillera of Chile 36°-41°S) calculated at 210 Ma using data of Lucassen et al 2004. Average of the Mesozoic magmatic arc of northern Chile (18-27°S) calculated at 210 Ma (data of Lucassen et al. 2006). Pacific MORB at 210 Ma (Petrological database of the ocean floor <http://petdb.org>). Northern hemisphere reference line (NHRL; Hart, 1984).

Figure 11: Sr-Nd initial isotopic ratios vs. SiO_2 content and $\text{La}_\text{N}/\text{Yb}_\text{N}$ ratios: a) Initial $^{87}\text{Sr}/^{86}\text{Sr}$ vs. SiO_2 content. The correlation line shows the good correlation present in the gabbroids ($r = 0.8$), b) $\epsilon_{\text{Nd(T)}}$ vs. SiO_2 content, with no obvious correlation in gabbroic rocks. c) $^{87}\text{Sr}/^{86}\text{Sr}$ vs. $\text{La}_\text{N}/\text{Yb}_\text{N}$ ratios (chondrite-normalized), d) ϵ_{Nd} vs. $\text{La}_\text{N}/\text{Yb}_\text{N}$ ratios.

Tables

Table 1: Petrographic characteristics of the different xenoliths included in the Cobquecura Pluton, Cordillera de la Costa, Chile (~36°S, 72°45'W)

Type of xenolith	Texture	Metamorphic grade	Mineral assemblage	Leucosome	Occurrence	Chemical features
Upper crustal xenoliths (UCX)	with strong foliation	low-grade metamorphism up to greenschist facies	white mica, quartz, albite, with pinitized cordierite near xenolith rims	absent	in gabbro	similar to the regional upper crust
High-leucosome restitic xenoliths (HLRX)	with foliation, presence of leucosome and melanosome	high-temperature metamorphism (granulite facies)	melanosome: hercynite, cordierite, corundum leucosome: quartz, K-feldspar, plagioclase, biotite	abundant (~30 vol. %)	in magmatic breccia	extremely rich in FeO and Al ₂ O ₃
Low-leucosome restitic xenoliths (LLRX)	with foliation, presence of leucosome and melanosome	high-temperature metamorphism (granulite facies)	melanosome: hercynite, cordierite leucosome: quartz, plagioclase, biotite, zircon, monazite	scarce (~5 vol. %)	in gabbro, fayalite granitoid	extremely rich in FeO and Al ₂ O ₃ , rich in REE and HFSE
Monzogabbro xenoliths (MX)	without foliation	magmatic without recrystallization texture	clinopyroxene, plagioclase, K-feldspar, amphibole	absent	in magmatic breccia	similar to the gabbroic rocks
Metamicrogabbroic xenoliths (MMX)	without foliation	magmatic with local recrystallization texture	clinopyroxene, plagioclase, opaque minerals, late amphibole, biotite	absent	in magmatic breccia	similar to the gabbroic rocks

Table 2: Major and trace elements: composition of granitoids, xenoliths and country rocks from the Cobquecura Pluton, Cordillera de la Costa, Chile (~36°S, 72°45'W)

Sample	Mafic rocks				Fayalite granitoids		Fa-free Granites		Granitic dike	Mafic dike	Metapelitic xenoliths		Metamorphic basement		Igneous xenoliths (microgabbroic)	
No	PV-03-11*	PV-03-30*	PV-05-19*	Mafic enclave PV-03-17*	PV-04-10*	PV-04-19*	PV-03-18*	PV 04-28*	PV-03-10*	PV-03-15*	PV 04-31 HLRX	PV 04-07 LLRX	PV 05-11	PV 05-13	PV 05-06	PV 05-03
SiO ₂	52.9	48.9	48.3	51.2	63.1	66.8	67.5	75.6	76.0	52.1	56.1	46.0	73.3	57.7	49.5	56.7
Al ₂ O ₃	17.8	18.0	18.7	16.2	15.1	14.4	15.2	13.6	12.5	14.0	20.0	26.6	13.1	17.0	16.8	15.7
Fe ₂ O ₃ *	7.74	7.62	6.06	8.65	9.40	7.55	4.67	1.39	1.74	12.4	7.68	11.1	3.91	8.54	10.4	7.95
MnO	0.11	0.12	0.11	0.13	0.15	0.13	0.06	0.03	0.03	0.19	0.08	0.11	0.03	0.06	0.19	0.13
MgO	5.28	9.92	7.76	6.50	0.87	0.50	0.83	0.27	0.25	3.30	2.29	3.69	1.05	2.15	8.55	3.92
CaO	8.35	9.72	13.0	8.54	4.24	2.94	2.56	0.99	0.64	5.44	1.35	2.95	0.19	4.20	9.57	7.18
Na ₂ O	3.41	2.34	2.33	3.47	4.06	4.02	3.69	3.63	3.62	4.01	3.6	3.31	0.81	1.26	2.11	3.35
K ₂ O	1.17	0.61	0.38	1.28	2.40	2.96	2.39	4.19	4.14	1.63	6.34	1.17	2.95	3.53	1.15	1.90
TiO ₂	1.23	0.82	0.54	1.32	0.88	0.65	0.62	0.21	0.12	2.53	0.90	1.58	0.67	0.95	1.30	1.31
P ₂ O ₅	0.16	0.09	0.06	0.16	0.27	0.18	0.13	0.06	0.05	0.38	0.14	0.25	0.10	0.18	0.16	0.22
LOI	1.33	1.47	2.30	1.65	0.41	0.61	1.05	0.55	0.33	2.51	1.85	2.90	3.70	4.05	0.83	1.63
<i>Total</i>	<i>99.5</i>	<i>99.6</i>	<i>99.5</i>	<i>99.1</i>	<i>100.9</i>	<i>100.7</i>	<i>98.7</i>	<i>100.5</i>	<i>99.4</i>	<i>98.5</i>	<i>100.3</i>	<i>99.7</i>	<i>99.8</i>	<i>99.6</i>	<i>100.6</i>	<i>100</i>
Y	22	16	9	24	37	37	33	21	39	43	43	54	26	35	29	31
Rb	64	29	16	69	120	130	118	211	176	60	205	43	142	159	65	106
Sr	281	225	300	232	205	157	171	82	28	203	168	323	74	97	192	216
Co	33	43	41	32	6	3	7	2	0	24	16	25	2	7	39	21
Ga	20	16	16	18	24	24	23	17	21	24	28	40	20	26	18	20
Nb	7	4	2	6	17	14	13	7	10	13	16	25	12	15	4	7
Pb	7	4	2	6	17	18	20	26	31	15	26	12	14	11	7	14
Th	3	2	0	3	9	11	16	13	17	4	14	22	12	12	1	5
U	0.9	0	0	0.7	2	3	3	5	4	1	4	3	3	3	0	2
Sc	25	21	36	29	28	25	11	4	4	29	22	34	16	18	30	27
Zr	139	80	44	130	226	278	401	107	173	263	171	271	150	162	103	158
La	12	8	2	12	29	37	42	23	24	22	45	76	33	40	7	15
Ce	28	18	6.1	29	64	77	83	46	48	53	89	159	72	84	19	33
Pr	4	2	1	4	8	9	11	5	7	7	10	18	8	10	3	4
Nd	16	11	5	17	35	36	41	18	27	32	38	70	30	37	14	18

Sm	4	3	1	4	8	8	8	4	7	8	7	13	6	7	4	4
Eu	1.4	0.9	0.8	1.4	2.3	2.0	1.7	0.6	0.7	2.4	1.9	3.6	1.7	1.6	1.2	1.5
Gd	4.3	3.0	1.7	4.7	8.2	7.9	7.8	3.4	7.1	8.3	6.7	11.0	4.3	6.3	4.4	4.8
Tb	0.7	0.5	0.3	0.8	1.3	1.2	1.1	0.6	1.3	1.4	1.3	1.9	0.8	1.1	0.9	0.9
Dy	4.5	3.2	1.9	4.9	8.0	7.7	6.8	3.8	7.8	8.6	7.5	10.0	4.6	6.4	5.3	5.5
Ho	0.9	0.6	0.4	1.0	1.6	1.6	1.4	0.7	1.6	1.8	1.4	1.9	0.9	1.2	1.0	1.1
Er	2.6	1.9	1.1	3.0	4.5	4.5	3.9	2.1	4.6	5.0	4.2	5.4	2.8	3.6	2.9	3.1
Tm	0.38	0.28	0.13	0.43	0.67	0.65	0.58	0.33	0.71	0.75	0.66	0.80	0.43	0.53	0.43	0.46
Yb	2.5	1.8	1.0	2.8	4.2	4.4	3.7	2.1	4.2	4.7	4.3	5.2	2.9	3.5	2.8	2.9
Lu	0.38	0.27	0.15	0.43	0.64	0.66	0.57	0.31	0.59	0.72	0.63	0.77	0.45	0.52	0.40	0.44

(*) Representative RFA major element analyses from magmatic rocks of the Cobquecura Pluton taken from Vásquez and Franz (2008). Major elements in wt-%; trace elements in ppm.

Table 3: Whole-rock Sr, Nd and Pb isotope data of granitoids, xenoliths and of country rocks from the Cobquecura Pluton, Cordillera de la Costa, Chile (~36°S, 72°45'W)

Sample N ^{oa}	Lithologic unit	Rock type	Age Ma	Rb ^c ppm	Sr ^c ppm	⁸⁷ Sr / ⁸⁶ Sr ^b	⁸⁷ Sr _(T) / ⁸⁶ Sr ^c	Nd ppm	Sm ppm	¹⁴³ Nd / ¹⁴⁴ Nd ^b	εNd _(T) ^c	Pb ppm	U ppm	Th ppm	²⁰⁶ Pb ^d ²⁰⁴ Pb	²⁰⁷ Pb ^d ²⁰⁴ Pb	²⁰⁸ Pb ^d ²⁰⁴ Pb	²⁰⁶ Pb ^e ²⁰⁴ Pb	²⁰⁷ Pb ^e ²⁰⁴ Pb	²⁰⁸ Pb ^e ²⁰⁴ Pb
05-19	Cobquecura Pluton	gabbronorite	203±2	16	300	0.704156±7	0.70373	5	1.4	0.512948±8	6.8	2	0.1	0.3	18.728	15.652	38.737	18.63	15.65	38.64
05-17	Cobquecura Pluton	gabbronorite	203±2	11	297	0.703854±7	0.70355	6	1.6	0.512937±6	6.8	2	0.1	0.3	18.512	15.634	38.441	18.41	15.63	38.34
04-22	Cobquecura Pluton	gabbronorite	203±2	13	291	0.705344±9	0.70498	13	3.7	0.512671±6	1.3	4	0.1	0.7	18.525	15.636	38.577	18.47	15.63	38.46
04-19	Cobquecura Pluton	Fa-granite	209±2	130	157	0.714081±9	0.70694	36	8.0	0.512518±5	-0.7	18	2.8	11.0	18.603	15.661	38.659	18.28	15.64	38.24
04-10	Cobquecura Pluton	Fa-granite	209±2	120	205	0.712219±9	0.70717	35	8.2	0.512532±5	-0.6	17	2.5	8.9	18.564	15.640	38.512	18.26	15.62	38.15
03-30	Cobquecura Pluton	gabbro	209±2	29	224	0.705258±9	0.70414	11	2.7	0.512729±5	3.1	4	0.4	1.7	18.609	15.659	38.619	18.40	15.65	38.33
03-27	Cobquecura Pluton	gabbro	209±2	33	260	0.706037±7	0.70495	17	4.5	0.512576±7	-0.2	4	0.5	2.1	18.849	15.669	39.046	18.58	15.66	38.68
03-18	Cobquecura Pluton	Fa-free granite	209±2	118	171	0.713876±7	0.70793	41	8.4	0.512484±5	-1.1	20	3.3	16.4	18.721	15.643	38.718	18.37	15.63	38.15
03-17	Cobquecura enclave	microgabbro	209±2	69	232	0.707895±7	0.70533	17	4.3	0.512678±4	1.9	6	0.7	3.0	18.865	15.664	38.786	18.62	15.65	38.44
03-15	Cobquecura dike	microgabbro	209±2	60	202	0.707932±9	0.70537	32	7.9	0.512698±6	2.4	15	0.9	3.9	18.585	15.608	38.500	18.46	15.60	38.32
03-14	Cobquecura dike	granite	209±2	151	149	0.715335±9	0.70660	40	8.1	0.512483±4	-1.0	19	3.1	12.7	18.959	15.647	39.046	18.61	15.63	38.58
03-13	Cobquecura Pluton	monzogabbro	209±2	62	244	0.708149±9	0.70596	22	5.5	0.512649±4	1.4	11	1.1	4.6	18.875	15.666	38.907	18.66	15.66	38.62
03-11	Cobquecura Pluton	gabbro	209±2	64	281	0.707607±9	0.70564	16	4.0	0.512645±5	1.4	7	0.9	3.1	18.657	15.642	38.779	18.39	15.63	38.47
05-11	Country rock	metapelite*	at 210	71	132	0.733701±9	0.72905	30	5.7	0.512122±4	-7.9	14	3.9	12.2	18.924	15.662	39.072	18.33	15.63	38.47
05-13	Country rock	metapelite*	at 210	80	105	0.736040±7	0.72945	38	7.3	0.512114±5	-8.1	11	3.3	12.0	19.022	15.654	39.089	18.38	15.62	38.33
04-31	Xenoliths	HLRX	210	85	91	0.726557±9	0.71848	38	7.4	0.512182±5	-6.8	26	3.5	14.5	18.723	15.632	38.688	18.44	15.62	38.30
04-07	Xenoliths	LLRX	210	15	224	0.718326±9	0.71774	77	13	0.512183±5	-6.3	12	3.3	22.0	18.906	15.654	39.055	18.32	15.62	37.78
04-28	Cobquecura Pluton	Fa-free granite	209±2	200	73	0.730518±7	0.70692	18	3.8	0.512447±4	-1.9	26	5.0	13.4	18.892	15.679	38.932	18.48	15.66	38.58
05-06	Xenoliths	microgabbro	210	19	141	0.708049±8	0.70686	14	4.0	0.512729±5	2.4	7	0.3	1.0	18.539	15.635	38.498	18.45	15.63	38.40
04-21	Cobquecura dike	microgabbro	210	35	207	0.707559±8	0.70608	26	6.0	0.512689±4	2.5	5	1.8	4.4	19.483	15.684	39.380	18.70	15.64	38.76
05-03	Xenoliths	microgabbro	209±2	86	211	0.709643±9	0.70611	18	4.5	0.512645±5	1.4	14	1.6	5.0	18.720	15.645	38.678	18.48	15.63	38.43

a) Samples were dissolved with conc. HF for four days at 160°C on the hot plate. Digested samples were dried and taken up in 6N HCl. Sr and Nd were separated and purified using ion-exchange chromatography as described in Romer et al. (2005). Pb was separated using the HBr-HCl ion-exchange procedure as described in Romer et al. (2005)

b) ⁸⁷Sr/⁸⁶Sr and ¹⁴³Nd/¹⁴⁴Nd normalized to ⁸⁷Sr/⁸⁶Sr = 0.1194 and ¹⁴³Nd/¹⁴⁴Nd = 0.7219 respectively. Isotopic data for Sr and Nd were obtained on a VG54 and Finnigan MAT262 multi-collector TIMS mass-spectrometer using dynamic multicollection, respectively. Analytical uncertainties are given at 2σ level.

c) ⁸⁷Sr/⁸⁶Sr_(T) and εNd_(T) were calculated for the time of the intrusion using λ⁸⁷Rb = 1.42E-11 y⁻¹ and λ¹⁴⁷Sm = 6.54E-12 y⁻¹, (¹⁴⁷Sm/¹⁴⁴Nd)_{0CHUR} = 0.1967, (¹⁴³Nd/¹⁴⁴Nd)_{0CHUR} = 0.512638. Concentration data for Rb and Sr were determined by isotope dilution, for the aliquot used for ⁸⁷Sr/⁸⁶Sr determination.

- d) Lead isotope data were obtained on a Finnigan MAT262 multi-collector TIMS mass-spectrometer using static multicollection. Lead isotope data were corrected for mass discrimination with 0.1% / A.M.U. Reproducibility at 2 σ level is better than 0.1%.
- e) Lead isotope data recalculated to the age of intrusions using the contents of Pb, Th and U (Vásquez et al. 2005, Vásquez et al., 2008) and the constants recommended by IUGS (Steiger and Jäger. 1977). Abbreviations: HLRX, high-leucosome restitic xenoliths; LLRX, low-leucosome restitic xenoliths. *: Crystallization age of metapelite is ~300 Ma.

Electronic Appendix A

Analytical Methods

I. Whole rock, trace elements and isotope geochemistry methods

Macroscopically fresh samples have been selected for analysis. All samples have been checked under the petrographic microscope for absence of alteration by weathering. Whole rock XRF analyses for major elements have been carried out at Technical University Berlin. For selected samples, REE were determined by ICP-MS (cf. Dulski, 2001) at GeoForschungsZentrum Potsdam (GFZ Potsdam). For other samples, REE analyses were carried out together with analyses of Y, and Sc by ICP-AES at GFZ Potsdam (cf. Zuleger and Erzinger 1988). All other trace elements were analyzed either by ICP-MS methods (GFZ Potsdam / ACTLABS, Ancaster, Ontario) or by ID-TIMS (Rb, Sr; GFZ Potsdam).

U–Th–Pb, Sm–Nd, and Rb–Sr isotopic and Rb–Sr concentration analyses were performed on different splits of one solution of ~100 mg sample powder. Nd, Sr and Pb isotopic ratios have been determined at the GFZ Potsdam (for analytical procedures for Sr and Pb, see Romer et al., 2005; for Nd, see Lucassen et al., 2004, 2006). Sr has been analyzed on a VG Sector 54 TIMS instrument in dynamic multicollection mode. During the period of analytical work the NBS 987 Sr standard yielded $^{87}\text{Sr}/^{86}\text{Sr} = 0.710268 \pm 10$ (2σ ; $n = 9$). Nd was analyzed on a Finnigan MAT 262 TIMS in dynamic multicollection mode. The La Jolla Nd standard yielded $^{143}\text{Nd}/^{144}\text{Nd} = 0.511850 \pm 4$ (2σ , $n = 14$). Nd isotopic ratios have been normalized to $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$, Sr isotopic ratios to $^{86}\text{Sr}/^{88}\text{Sr} = 0.1194$. Pb isotopic ratios have been measured on a Finnigan MAT 262 TIMS using static multicollection at controlled temperatures between 1200–1250°C. Instrumental mass-fractionation has been corrected using 0.1% per a.m.u. as determined from the repeated analysis of NBS SRM 981 standard. The 2σ reproducibility of all Pb isotope ratios of the NBS SRM 981 standard ($^{206}\text{Pb}/^{204}\text{Pb} = 16.90 \pm 0.01$, $n = 52$) is better than 0.1% and a 2σ error of 0.1% is assumed for the measured samples. Procedural blanks were <30 pg for Pb, <50 pg Nd and <100 pg for Sr. No blank corrections have been applied to the measured ratios because blank contribution were insignificant in comparison with the amount of the respective elements in the sample.

II. In-situ U-Pb zircon geochronology by laser ablation – magnetic sector field - inductively coupled plasma mass spectrometry (LA-ICP-MS)

For LA-ICP-MS age determinations, we worked with 100 μm thick polished thick sections, to control the textural setting of the zircon crystals. BSE and CL imaging was utilized to characterize analyzed spots.

Uranium, thorium and lead isotopes were analyzed using a ThermoScientific Element 2 sector field ICP-MS coupled to a New Wave Research UP-213 ultraviolet laser system (Gerdes and Zeh, 2006, 2008). Data were acquired in time resolved – peak jumping – pulse counting mode over 810 mass scans during 19 second background measurement followed by 30 second sample ablation. Laser spot-size was 20 μm and depth penetration typically ~15-20 μm . Signal was tuned for maximum sensitivity for Pb and U while keeping oxide production, monitored as $^{254}\text{UO}/^{238}\text{U}$, well below 1%. A teardrop-shaped, low volume (< 2.5 cm^3) laser cell was used. With a response time of <1s and wash-out (< 99% of previous signal) time of ~5s this cell enables detection and sequential sampling of heterogeneous grains (e.g., growth zones) during time resolved data acquisition (cf., Janousek et al., 2006). Raw data were corrected offline for background signal, common Pb, laser induced elemental fractionation, instrumental mass

discrimination, and time-dependent elemental fractionation of Pb/U using an in-house MS Excel[®] spreadsheet program (Gerdes and Zeh, 2006). A common-Pb correction based on the interference- and background-corrected ^{204}Pb signal and a model Pb composition (Stacey and Kramers, 1975) was carried out, where necessary. The necessity of the correction was judged on whether the corrected $^{207}\text{Pb}/^{206}\text{Pb}$ lied outside of the internal errors of the measured ratios. Laser-induced elemental fractionation and instrumental mass discrimination were corrected by normalization to the reference zircon GJ-1 of each analytical session. Prior to this normalization, the drift in inter-elemental fractionation (Pb/U) during 30s sample ablation was corrected for the individual analysis. Reported uncertainties (2σ) were propagated by quadratic addition of the external reproducibility (2 SD) obtained from the standard zircon GJ-1 ($n = 12$; 2 SD $\sim 1.2\%$ and $\sim 1.4\%$ for the $^{207}\text{Pb}/^{206}\text{Pb}$ and $^{206}\text{Pb}/^{238}\text{U}$, respectively) during the analytical session and the within-run precision of each analysis (2 SE; standard error). See Gerdes and Zeh (2006, 2008) for more details.

Electronic Appendix B

U, Th and Pb isotope analyses for zircon from a low leucosome restitic xenolith (LLRX; sample 07-45), included in gabbro of the Cobquecura Pluton, Cordillera de la Costa, Chile (~36°S, 72°45'W)

Analysis No.	²⁰⁷ Pb ^a (cps)	U ^b (ppm)	Pb ^b (ppm)	Th ^b U	Age (Ma)													
					²⁰⁶ Pb	²⁰⁶ Pb ^c	±2σ	²⁰⁷ Pb ^c	±2σ	rho ^e	²⁰⁷ Pb ^c	±2σ	²⁰⁶ Pb	±2σ	²⁰⁷ Pb	±2σ	²⁰⁷ Pb	±2σ
					²⁰⁴ Pb	²³⁸ U		²³⁵ U			²⁰⁶ Pb		²³⁸ U		²³⁵ U		²⁰⁶ Pb	
48	3932	502	21	1.79	9851	0.0357	1.1	0.254	2.8	0.40	0.0515	2.6	226	3	230	6	261	60
49*	2103	274	199	128	6210	0.0380	1.3	0.285	3.6	0.36	0.0544	3.3	241	3	255	8	386	74
2b	1390	153	5.4	0.54	798	0.0334	1.3	0.234	4.7	0.28	0.0509	4.5	212	3	214	9	235	104
2c	2038	159	11	0.32	3476	0.0691	1.5	0.563	3.7	0.40	0.0591	3.4	431	6	453	14	570	74
10	3193	169	8.8	0.38	5060	0.0504	1.7	0.399	3.8	0.45	0.0574	3.4	317	5	341	11	507	75
51	1813	159	8.8	0.57	5714	0.0542	1.2	0.417	4.0	0.31	0.0558	3.8	340	4	354	12	444	85
52	2112	190	11	0.51	3078	0.0557	1.0	0.428	4.8	0.21	0.0557	4.7	350	3	362	15	439	105
53	1471	183	6.7	0.85	2822	0.0336	1.6	0.235	4.9	0.32	0.0506	4.7	213	3	214	10	222	108
54	607	79	2.6	0.50	1616	0.0328	3.4	0.235	6.9	0.50	0.0519	5.9	208	7	214	13	283	136
4	690	81	2.9	0.54	1235	0.0350	1.6	0.244	8.3	0.20	0.0505	8.1	222	4	222	17	218	188
55	1465	161	6.4	0.51	2836	0.0384	1.6	0.273	5.6	0.29	0.0515	5.4	243	4	245	12	265	124
6	572	98	3.2	0.79	866	0.0319	2.5	0.221	6.8	0.37	0.0502	6.3	203	5	203	13	206	147
56	1036	120	5.0	0.77	799	0.0406	2.6	0.294	5.5	0.48	0.0526	4.8	256	7	262	13	311	109
29	1600	210	7.7	0.52	1181	0.0342	2.4	0.236	7.7	0.32	0.0501	7.3	217	5	215	15	200	169
31b	2278	184	6.7	0.42	2093	0.0348	2.0	0.248	5.0	0.39	0.0517	4.6	220	4	225	10	271	106
31	1424	97	3.6	0.42	1845	0.0347	2.4	0.252	7.1	0.34	0.0526	6.7	220	5	228	15	313	151
32	1434	167	5.4	0.31	1880	0.0325	2.4	0.222	6.5	0.38	0.0497	6.0	206	5	204	12	180	141
33	1527	187	6.4	0.45	1017	0.0326	2.4	0.220	5.7	0.42	0.0489	5.2	207	5	202	11	144	122
35	2929	526	43	1.93	1674	0.0869	3.4	0.756	6.0	0.56	0.0631	4.9	537	18	572	26	711	105
34	1949	346	11	0.38	1785	0.0318	2.4	0.229	6.6	0.37	0.0522	6.2	202	5	209	13	294	141
28	1482	179	6.2	0.49	542	0.0336	2.4	0.231	6.7	0.36	0.0500	6.2	213	5	211	13	196	144
27	1555	257	8.6	0.44	480	0.0326	1.6	0.232	3.7	0.43	0.0516	3.4	207	3	212	7	266	77
26	2004	152	5.7	0.44	2384	0.0365	2.1	0.258	5.3	0.39	0.0514	4.9	231	5	233	11	258	112
38	2511	201	7.2	0.44	1121	0.0346	2.3	0.233	4.8	0.47	0.0487	4.2	219	5	212	9	134	99
37	1713	254	9.1	0.61	1869	0.0334	2.1	0.229	8.0	0.26	0.0497	7.7	212	4	209	15	181	180
40	1422	147	5.2	0.43	1510	0.0337	2.2	0.230	8.7	0.25	0.0495	8.4	214	5	210	17	170	196
39	1611	110	4.5	0.43	900	0.0351	2.6	0.251	9.9	0.26	0.0520	9.5	222	6	228	20	285	218
23	1413	160	5.4	0.45	1426	0.0325	2.3	0.215	7.1	0.32	0.0480	6.8	206	5	198	13	97	160

Analysis No.	$^{207}\text{Pb}^a$ (cps)	U^b (ppm)	Pb^b (ppm)	Th^b U	Age (Ma)													
					$\frac{^{206}\text{Pb}}{^{204}\text{Pb}}$	$\frac{^{206}\text{Pb}^c}{^{238}\text{U}}$	$\pm 2\sigma$ (%)	$\frac{^{207}\text{Pb}^c}{^{235}\text{U}}$	$\pm 2\sigma$ (%)	ρ^e	$\frac{^{207}\text{Pb}^c}{^{206}\text{Pb}}$	$\pm 2\sigma$ %	$\frac{^{206}\text{Pb}}{^{238}\text{U}}$	$\pm 2\sigma$	$\frac{^{207}\text{Pb}}{^{235}\text{U}}$	$\pm 2\sigma$	$\frac{^{207}\text{Pb}}{^{206}\text{Pb}}$	$\pm 2\sigma$
45	1585	210	7.6	0.51	967	0.0342	2.4	0.235	8.0	0.30	0.0499	7.6	217	5	215	15	189	177
19	1419	146	5.4	0.54	1629	0.0348	2.0	0.264	7.4	0.27	0.0551	7.1	220	4	238	16	418	159
18	1713	231	8.2	0.48	1258	0.0340	2.0	0.255	6.4	0.32	0.0545	6.0	215	4	231	13	391	135
20	1074	192	7.3	0.46	1384	0.0318	2.4	0.209	9.4	0.25	0.0478	9.1	202	5	193	17	88	215
21	1563	185	6.4	0.53	2271	0.0326	2.0	0.235	5.9	0.34	0.0523	5.5	207	4	214	11	297	126
15	1423	152	5.7	0.43	1439	0.0361	2.3	0.259	5.9	0.39	0.0520	5.4	228	5	234	12	288	123
14	1546	221	7.7	0.56	1280	0.0325	2.3	0.223	5.7	0.40	0.0497	5.2	206	5	204	11	182	122
13	1391	232	10	0.44	164	0.0318	2.3	0.208	9.4	0.24	0.0474	9.1	202	5	192	17	67	217
17	1609	223	7.6	0.48	1306	0.0324	2.3	0.226	6.0	0.38	0.0506	5.6	206	5	207	11	223	129
16	1687	192	6.8	0.54	1984	0.0320	2.4	0.229	5.5	0.44	0.0519	4.9	203	5	209	10	280	113
11	1947	297	12	0.66	616	0.0356	2.6	0.255	6.8	0.38	0.0520	6.3	226	6	231	14	286	145
10	1672	218	7.6	0.41	1712	0.0336	2.2	0.245	6.2	0.36	0.0529	5.8	213	5	223	12	325	131
12	1608	214	7.6	0.45	1315	0.0335	2.2	0.228	7.4	0.29	0.0493	7.1	212	5	209	14	164	165
8	1600	217	7.9	0.47	1724	0.0345	1.9	0.241	6.0	0.32	0.0506	5.7	219	4	219	12	222	131
9	1485	182	6.2	0.49	1923	0.0325	2.2	0.223	7.3	0.30	0.0496	7.0	206	5	204	14	178	163
4	1637	163	6.3	0.55	1669	0.0359	2.3	0.253	8.0	0.29	0.0510	7.6	227	5	229	16	242	176
5	34099	303	122	4.15	16937	0.0325	1.8	0.227	2.0	0.88	0.0508	1.0	206	4	208	4	231	23
7	2081	284	9.5	0.31	1599	0.0316	2.4	0.231	7.9	0.30	0.0529	7.5	201	5	211	15	323	171
3	2474	596	24	0.55	1454	0.0377	1.7	0.266	4.5	0.38	0.0512	4.1	239	4	240	10	249	95
1	1450	144	5.6	0.62	1912	0.0354	1.8	0.237	11.1	0.17	0.0485	11.0	224	4	216	22	122	258
2	1464	167	6.0	0.38	1239	0.0349	2.0	0.227	12.0	0.17	0.0472	11.8	221	4	208	23	61	281
44	1580	180	6.2	0.47	1324	0.0320	2.4	0.228	6.4	0.37	0.0516	6.0	203	5	208	12	270	137
43	1772	300	11	0.58	2204	0.0345	1.6	0.228	7.2	0.22	0.0478	7.0	219	3	208	14	88	166

^a Within-run background-corrected mean ^{207}Pb signal in counts per second (cps).

^b U and Pb concentrations and Th/U ratios were calculated relative to GJ-1 zircon reference standard (LA-ICP-MS values, Gerdes, unpubl.)

^c corrected for background and within-run Pb/U fractionation and subsequently normalised to GJ-1 reference (ID-TIMS value/measured ratio)

^d corrected for mass bias by normalising to GJ-1 (ID-TIMS/measured ratio; c. 0.5% per amu).

^e error correlation defined as $\text{err}^{^{206}\text{Pb}/^{238}\text{U}}/\text{err}^{^{207}\text{Pb}/^{235}\text{U}}$.

* Analysis in monazite