The source of Proterozoic anorthosite and rapakivi granite magmatism: evidence from combined in situ Hf–O isotopes of zircon in the Ahvenisto complex, southeastern Finland

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Abstract: The isotope compositions of massif-type anorthosites in Proterozoic anorthosite–mangerite–charnockite–granite (AMCG) complexes are commonly dominated by crustal values. Olivine-bearing anorthositic rocks in several AMCG suites have, however, been shown to display juvenile character, suggesting that variably depleted mantle reservoirs were involved in their genesis. A coupled in situ zircon Hf–O isotope dataset from the 1.64 Ga Ahvenisto AMCG complex in the 1.54–1.65 Ga Fennoscandian rapakivi granite–massif-type anorthosite province reveals correlated juvenile isotope signals (δ18Ozzn = 5.4–6.6‰; initial εHf = −1.1 to +3.4) in the most primitive gabbroic rock type of the suite suggesting a depleted mantle origin for the anorthositic rocks. This signal is not as prominent in the more evolved co-magmatic anorthositic rocks (δ18Ozzn = 6.3–7.8‰; initial εHf = −2.1 to +0.5), most probably owing to contamination of the mantle-derived primary magma by crustal material. A rapakivi granite associated with the anorthositic rocks has different isotope composition (δ18Ozzn = 7.4–8.6‰; initial εHf = −0.8 to +2.0) that points to a crustal source.

Supplementary material: Full analytical results and details and SEM images of the analysed grains with spatial information on both methods are available at: http://www.geolsoc.org.uk/SUP18780.

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The origin and sources of the Proterozoic massif-type anorthositic rocks and related ferroan A-type (rapakivi) granites in the so-called anorthosite–mangerite–charnockite–granite (AMCG) suites found in nearly all continental areas still pose an unsettled petrological problem. Most notably, the role and nature of the mantle component purportedly involved in their petrogenesis have eluded petrologists for decades. The classical two-source model (mantle for the anorthositic members and crust for the felsic rocks; e.g. Anderson 1983; Haapala & Rämö 1990; Emslie et al. 1994; Frost et al. 1999; Scoates & Mitchell 2000; McLelland et al. 2010) has been challenged by two competing single-source hypotheses that advocate either the mantle (Turner et al. 1992; Frost et al. 2002) or the continental crust (Duchesne & Wilmart 1997; Duchesne et al. 1999; Schiellerup et al. 2000) as the only source for the rock types in these suites. In this paper, we provide a spatially coupled in situ isotope dataset of zircon oxygen and Hf analyses from the 1.64 Ga Ahvenisto rapakivi granite–massif-type anorthosite complex in southeastern Finland to evaluate the validity of the source hypotheses regarding both the anorthositic and granitic rocks in the 1.53–1.65 Ga Fennoscandian rapakivi suite. These data allow us to correlate a previously reported radiogenic Hf isotope signal in the anorthositic rocks (Heinonen et al. 2010a) with mantle-like δ18O values and, furthermore, to assess the importance of contamination by crust in AMCG petrogenesis.

Theoretical background

In part, the long-lived controversy over the sources involved in AMCG magmatism stems from the isotopic similarity of the postulated source components. In many cases, radiogenic isotope systems (namely Sm–Nd) in the upper mantle and contemporary juvenile crustal source reservoirs have not developed detectable compositional contrast during the relatively short time elapsed since crust formation in any one AMCG suite. Accumulating isotope data, however, suggest that contamination of mantle-derived magmas by lower crust is an important process in AMCG petrogenesis (see Bybee et al. 2014, and references therein).

Most studies of isotopic tracers utilized thus far have employed the bulk isotope dilution-thermal ionization mass spectrometry (ID-TIMS) technique, which, although superbly precise in terms of analytical accuracy, offers limited spatial resolution to register petrogenetic information potentially stored in heterogeneous populations of minerals. In situ Hf–O isotope studies of zircon provide additional information to resolve these issues (e.g. Kemp et al. 2007; Heinonen et al. 2010a). The spatial resolution of the secondary ionization mass spectrometry (SIMS) and laser ablation multicollector–inductively coupled plasma–mass spectrometry (LAM-ICP-MS) techniques is able to recover detailed petrogenetic information recorded by the mineral zircon during magmatic evolution. In addition, the shorter half-life of 176Lu compared with 147Sm results in stronger contrast in the initial 176Hf/177Hf values compared with the 143Nd/144Nd values in potential source-reservoirs. Stable oxygen isotopes complement the Hf isotope data as enrichment in 16O (higher δ18O) in zircon provides time-independent information of supracrustal influence in magmas (Valley 2003).

Issues inherent to bulk-rock isotopic analysis have also hindered the study of the origin of the 1.54–1.65 Ga Finnish rapakivi suite (see Rämö & Haapala 2005). Several conventional isotopic studies (Sm–Nd, Rb–Sr, Pb–Pb, Lu–Hf) have produced an array of overlapping compositions that point to a (lower) crustal source for the rapakivi granites and a major crustal influence in
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the associated, potentially mantle-derived, anorthositic rocks (e.g. Patchett et al. 1981; Rämö 1991; Neymark et al. 1994). A recent in situ zircon Lu–Hf isotope (LAM-ICP-MS) and trace element (LAQ-ICP-MS; laser ablation quadrupole–inductively coupled plasma–mass spectrometry) study (Heinonen et al. 2010a), however, provides further evidence that the Finnish rapakivi granites were derived from a Palaeoproterozoic crustal source with a relatively homogeneous Hf isotope composition. Furthermore, in contrast to the rapakivi granites, a considerably greater within-sample variation in the initial Hf isotope compositions of zircon from the associated anorthositic and monzodioritic rocks (εHf from c. 0 to c. +9 at 1640 Ma; Heinonen et al. 2010a) points to thorough contamination by crust of a depleted, possibly even asthenospheric, mantle (DM) source. In spite of strong isotopic overprinting by crust, some of the zircon crystals in the most primitive rock types of the anorthosite suite might still preserve the original isotope composition of the primary mantle source. Other recent Hf isotope studies also imply strong crustal influence on the parental magmas of massif-type anorthositic rocks in AMCG complexes in general (e.g. Morisset et al. 2006; Bickford et al. 2010; Heinonen et al. 2012).

Laser fluorination oxygen isotope studies on AMCG suite rocks (e.g. Valley et al. 1994; Peck & Valley 2000; Elliott et al. 2005; Peck et al. 2010) have provided similar implications to the Hf results. They show that the geochemically most primitive (commonly olivine-bearing) gabbrö-anorthositic rocks have more mantle-like δ18O signatures than the coeval anorthositic granitic rocks, which in turn have overlapping compositions that imply a stronger supracrustal influence. For example, in the Grenville and Nain AMCG suites (Peck & Valley 2000; Peck et al. 2010), contrasting oxygen isotope signatures in anorthositic rocks have been attributed to contamination of mantle-derived magmas by material from varying crustal sources (Peck et al. 2010). Results of previous bulk zircon oxygen isotope studies (Elliott et al. 2005) tentatively imply a similar relationship between the rock types of the Finnish AMCG suite.

Geological background

The Finnish rapakivi granites and associated anorthositic rocks form four major granitic batholiths and several smaller batholiths and stocks (Rämö & Haapala 2005; Fig. 1b). These rocks were intruded into the Palaeoproterozoic (1.8–1.9 Ga) Svecofennian crust of southern Finland in two stages, at 1.65–1.62 and 1.59–1.54 Ga. The c. 1.64 Ga Ahvenisto complex (Alviola et al. 1999; Heinonen et al. 2010b; Fig. 1c) is the most prominent locality of Proterozoic massif-type anorthosite in southern Finland (Rämö & Haapala 2005; Heinonen et al. 2010a,b). At the current level of exposure, the Ahvenisto complex (Fig. 1c) covers c. 350 km² and comprises a multi-phase rapakivi granite batholith (70% of areal extent) and associated gabbro–anorthositic rocks (anorthosites, leucogabbronorites, leuconorites, and leucotroctolites) in a horseshoe-shaped zone around the granite batholith and consist primarily of leucogabbroanorthosite, of which c. 10% is locally olivine-bearing. True anorthosite (plagioclase >90 vol.%) is found only as autolithic inclusions (up to 200 m across) in the leucogabbroanorthosite (Heinonen et al. 2010b). Monzodioritic rocks are present mainly as discontinuous ring-dikes in the eastern and southeastern parts of the complex. The country rocks of the complex are Palaeoproterozoic (c. 1.89 Ga) mica schists, granodiorites, garnet- and cordierite-bearing late-tectonic leucogranites, and
Table 1. Background information on the studied samples

| Sample number | Rock type: | A118 | A1271 | A1306 | A1360 | A1933 |
|---------------|------------|------|-------|-------|-------|-------|
|               | Sample     | lgbbrnor | iltroct | qmonzod | br-hbl gr | an     |

| Geophysical compositions from Heinonen et al. (2010b), except for A1271, for which average of five olivine-bearing gabbro rocks as reported by Alviola et al. (1999) was used, and A1933, for which equivalent sample MH87a of Heinonen et al. (2010b) was used. U-Pb TIMS age data (on zircon and baddeleyite) from Alviola et al. (2010) and Heinonen et al. (2010b). Laser fluorination oxygen isotope values from Elliott et al. (2005). Hf isotope values from Heinonen et al. (2010a) and Nd isotope values from Heinonen et al. (2010b). an, anorthosite; br-hbl gr, biotite–hornblende granite; lgbbrnor, leucogabbro-norite; iltroct, leucotroctolite; qmonzod, quartz-monzodiorite.

The Nd–Sr–Pb isotopic compositions and elemental geochemistry, the monzodiorites and the anorthositic rocks of the Alvenisto complex are deemed to share a common source (Heinonen et al. 2010b). On the basis of Nd–Sr–Pb isotopic compositions and elemental geochemistry, the monzodiorites and the anorthositic rocks of the Ahvenisto complex are well characterized in the previous studies. All the major rock types of the Ahvenisto complex were sampled for combined Hf–O analysis. These comprise rapakivi granite (A1360), leucogabbro-norite (A118), leucotroctolite (A1271), anorthosite (A1933), and quartz monzodiorite (A1306), and have been well characterized in the previous studies (Table 1).

Zircon separated from the five samples was analysed for oxygen isotopes by SIMS and for Hf isotopes by LAM-ICP-MS. Clear, non-metamict and inclusion-free zircon grains for in situ analysis were picked from pure, non-magnetic zircon fractions produced by standard mineral separation methods (Wilfley table, heavy liquid and magnetic separation). The grains were mounted in epoxy together with a standard in the central 12 mm area of the 25 mm mount, sectioned approximately in half, polished to surface, and Au-coated for SIMS and Hf isotope analyses. SEM imaging was performed prior to coating to avoid targeting unsuitable crystals. Save for magmatic zonation in some of the crystals, no internal structures or inherited domains were recognized. For the LAM-ICP-MS Hf analyses, the mounts were polished after the O isotope analyses to remove the Au coating and possible minor damage caused by the primary ion beam.

In situ data acquired for the isotope systems were spatially coupled by targetting the same portions of the analysed crystals wherever possible. We analysed the grains for O isotopes from 86 spots (16–20 per sample; sample volume some hundreds of µm³) and Hf isotopes from 46 spots (6–12 per sample; sample volume c. 50000 µm³). The resulting spatially coupled dataset, with both oxygen and Hf isotope data, comprises 42 analysis spots with 6–12 spots per sample.

**SIMS measurement of oxygen isotope values**

The oxygen isotope values in zircon were measured with a CAMECA IMS1280 multicollector ion microprobe at the NordSIM Laboratory, Department of Geosciences, Swedish Museum of Natural History in Stockholm. The analysis was performed with a c. 2 nA Cs⁺ primary ion beam together with a normal incidence, low-energy, electron gun for charge compensation, medium field magnification (c. 80×) and two Faraday detectors (channels L2 and H2) at a common mass resolution of c. 2500. Measurements were performed in pre-programmed chain-analysis mode with automatic field aperture and entrance slit centring on the 16O signal. The magnetic field was locked using nuclear magnetic resonance regulation for the entire analytical session. Each data-acquisition run comprised a 20 µm × 20 µm pre-sputter to remove the Au layer, followed by the centring steps and 64 s of data integration performed using a non-rastered, c. 10 µm spot. Field aperture centring values were found to be well within those for which no bias has been observed during tests on standard mounts (Whitehouse & Nemchin 2009). A total of 86 unknowns was measured during two analysis sessions. In the measurement chain, every set of four unknowns was bracketed by two analyses of the Geostandards 91500 zircon. A δ18O value of +9.86‰ (Wiedenbeck et al. 2004) (SMOW) was assumed for the 91500 zircon in data normalization, and small linear-drift corrections were applied to each session. External reproducibility of...
Masses 172–179 were measured simultaneously in Faraday collectors. Ablation was conducted in helium, at the following conditions: beam diameter 55 µm (aperture imaging mode), pulse frequency 5 Hz, beam fluence c. 1 J cm⁻², and static ablation. Each ablation was preceded by a 30 s measurement of the background signal. The total Hf signal obtained was in the range 1.5–3.0 V. Under these conditions, 120–150 s of ablation was required to obtain an internal precision of ±0.000020 (1SE).

Standard reference zircon crystals 91500, Temora-2, Mud Tank and GJ-1 were run as unknowns at frequent intervals. Data obtained over a 2 year period indicate accuracy within ±0.000020; the observed external reproducibility on Temora-2 of ±0.000058 (2SD) gives a conservative estimate of an uncertainty of ±2 epsilon units.

Isotope ratios were calculated using the Nu Plasma time-resolved analysis software. The raw data were corrected for mass discrimination using an exponential law; the mass-discrimination factor for Hf was determined assuming 179 Hf/177 Hf = 0.7325; the observed 2SE uncertainty of the Hf mass-discrimination factor was better than 0.5%, which was propagated through to the error of the final result. The interference on mass 176 from 176 Yb was determined from the interference-free isotope 172 Yb, using 176 Yb/172 Yb = 0.58715. The Yb mass-discrimination factor was calculated from the 174 Yb/172 Yb value using 174 Yb/172 Yb = 1.46182, after removal of a minor interference on mass 174 from 174 Hf, following the protocol described by Elburg et al. (2013).

The value 176 Lu/177 Lu = 0.02669 was used for correction of Lu interference on mass 176 (DeBievre & Taylor 1993). On the basis of this revised correction, Temora-2 gives a present-day 176 Hf/177 Hf value of 0.282679 ± 51 (2SD, n = 322), which is comparable with the published analytical data on a solution, 0.282686 ± 8 (Woodhead & Hergt 2005). The low-REE Mud Tank reference zircon gave 176 Hf/177 Hf = 0.282510 ± 43 (2SD, n = 351), which is indistinguishable from the solution analysis at 0.282507 ± 6 from previous studies (Woodhead & Hergt 2005). The in-house high-Yb (176 Yb/177 Hf = 0.26 ± 26) LV-11 standard gave 176 Hf/177 Hf = 0.282824 ± 80 (2SD, n = 254), comparable with the solution value at 0.282837 ± 28 (Heinonen et al. 2010a). A value for the decay constant of 176 Lu of 1.867 × 10⁻¹¹ a⁻¹ was used in all calculations (Scherer et al. 2001; Söderlund et al. 2004). For the calculation of ε Hf values, present-day chondritic 176 Hf/177 Hf = 0.282785 and 176 Lu/177 Hf = 0.0336 values (Bouvier et al. 2008) were used. A depleted mantle model (Griffin et al. 2000) was used, producing a present-day 176 Hf/177 Hf value (0.28325, ε Hf = +16.4), similar to that of average mid-ocean ridge basalt over 4.56 Ga, from chondritic initial hafnium at 176 Lu/177 Hf DM = 0.0388 modified to the above-mentioned λ 176 Lu and chondritic composition.

Results
Zircon δ¹⁸O values
The leucogabbro (A118), anorthosite (A1933), and quartz monzodiorite (A1306) have higher weighted average δ¹⁸O zrn values (6.9–7.2‰) and more limited within-sample variation (A118 6.3–7.2‰; A1933 6.5–7.8‰; A1306 6.6–7.6‰) than the leucotroctolite (A1271 δ¹⁸O zrn = 5.9‰, 5.4–7.0‰) and the leucogabbro (A118, quartz monzodiorite (A1306), and rapakivi granite (A1360) have comparable δ¹⁸O zrn SIMS values to those determined earlier by laser fluorination (Fig. 2; Elliott et al. 2005).

Lu–Hf isotopes
Weighted mean initial ε Hf isotope values of the five samples analysed follow the previously determined pattern for the Finnish rapakivi suite rock types but the absolute values and single
analyses seem to plot at the lower end of the previously determined within-sample ranges (Heinonen et al. 2010a; Figs 3 and 4). The new results from the anorthositic rocks and the quartz monzodiorite have chondritic to slightly positive average εHf (at 1640 Ma) values (0.0 in the quartz monzodiorite to +1.4 in the leucotroctolite). The rapakivi granite has a somewhat lower negative mean εHf value (−0.6 at 1640 Ma). Contrary to earlier measurements (Heinonen et al. 2010a), the analyses from the anorthositic rocks in this study do not produce a broader within-sample compositional variation than in the granite but do still range to higher absolute values (> +3 εHf). The initial εHf values calculated for the granite A1360 extend to as low a value as −2.1. All the samples have initial εHf ranges that overlap significantly with the evolution path of the Svecofennian crust (Fig. 3).

Discussion

Mantle source of the anorthositic rocks

The differing oxygen isotope signatures in the Grenvillian anorthosites were suggested by Peck et al. (2010) to result from contamination of mantle-derived magmas by materials from varying crustal sources. Similar oxygen isotope patterns have also been reported for the Finnish rapakivi suite by Elliott et al. (2005). Their laser-fluorination study included three samples of bulk zircon from the Ahvenisto complex. Elliott et al. (2005) observed that the leucogabbronorite and the quartz monzodiorite had significantly lower δ18O values (A118 and A1306; δ18OZrn = 7.04–7.15‰) than the rapakivi granite (A1360; δ18OZrn = 8.4‰). This they interpreted to reflect difference in the sources of primary magmas (mantle for the leucogabbronorite and the quartz monzodiorite and crust for the granite). These observations were confirmed by our new in situ measurements and, furthermore, an even more juvenile δ18O signature akin to the olivine-bearing anorthositic rocks in the Grenville (Peck et al. 2010) was detected in the Ahvenisto leucotroctolite (A1271; δ18OZrn = 5.4–7.0‰). Similar magmatic processes thus may have affected both suites. The leucotroctolite is also the only anorthositic sample that displays mantle-like δ18OZrn values within the external uncertainty of the SIMS method (±0.6‰, 2SD). The other anorthositic rocks and the monzodiorite have surprisingly homogeneous and uniform δ18O compositions that could be interpreted to reflect a homogeneous crustal source but for the reasons stated below we prefer a model in which a mantle-derived magma is contaminated by the crust, as in the Grenville Province (see Peck et al. 2010).

Previous analyses on the leucotroctolite A1271 (Heinonen et al. 2010a) show a rather large range in the zircon Hf isotope compositions (initial εHf from 0.0 to +7.9; n = 53; Fig. 4), typical of the anorthositic rocks of the rapakivi suite of southern Finland. The previously determined initial Hf isotope values for the leucogabbronorite (A118) and monzodiorite (A1306) samples (Heinonen et al. 2010a) do not range to such high values as for the leucotroctolite, but generally the variation pattern was interpreted as a depleted (asthenospheric) mantle source signal. The variation in the initial Hf isotope values in the coupled Hf–O isotope dataset (n = 8) from the leucotroctolite is more limited (initial εHf from −1.1 to +3.4) and borders on being meaningful within the external reproducibility of the method but, nevertheless, shows a tentative correlation with the radiogenic Hf and low δ18OZrn values (5.4–6.6‰; Fig. 4). Also, the nominally lower εHf values in the leucogabbronorite and monzodiorite samples seem to be correlated with overall higher and more constant δ18OZrn values compared with the leucotroctolite (Fig. 4). We interpret the more limited within-sample ranges in the present dataset to reflect the relatively smaller sample sizes and note that they represent expected samples from the positively skewed (sample S = 0.88; n = 61 for A1271) total Hf composition distributions. Thus, we argue that the observed trend could conform to magmatic evolution of a mantle-derived magma via a combined assimilation–fractional crystallization (AFC) process. This possibility is explored by EC–AFC (energy-constrained assimilation–fractional crystallization; Bohrson & Spera 2001; Spera & Bohrson 2001; Table 2) models plotted in Figure 4 in relation to the analysed data.

Two cases with different mantle sources, an enriched sub-continental lithosphere mantle (SCLM) and an asthenospheric depleted mantle (DM), both contaminated with similar material, are considered as plausible alternatives. In both cases, a viable contamination model requires a relatively enriched (lower 176Hf/177Hf) contaminant with higher δ18O values and higher abundance (concentration, C) of Hf (Cmelt/Ccontaminant ≤ 0.1 for the DM case and 0.1–0.5 for the SCLM case) than those of the magma. For lower crustal Hf abundances of c. 2–4 ppm (e.g. Rudnick & Gao 2004) a relatively high grade of partial melting of either mantle source is required (DM c. 0.1–0.2 ppm Hf (e.g. Workman & Hart 2005) and somewhat higher Hf concentration for the enriched SCLM source) to produce viable Cmelt/C ratios. Svecofennian crust is assumed to be a suitable contaminant and is approximated in the models by the highest whole-rock
oxygen isotope compositions reported for Palaeoproterozoic rocks in southern Finland (Hoefs & Epstein 1969) and Hf isotope data for Svecofennian rocks by Andersen et al. (2009) and Kurhila et al. (2010). Oxygen isotope compositions for both mantle reservoirs are assumed to conform to the mantle zircon value of Valley et al. (1998; δ18 O = 5.3 ± 0.3‰), and the Hf isotope evolution at 1640 Ma is calculated according to Griffin et al. (2000) for DM and Andersen et al. (2009) for Svecofennian SCLM (or Svecofennian lithosphere mantle; SFLM). Based on the generally incompatible behaviour of Hf in the major mineral phases involved over the plausible compositional range of the magma and the contaminant and allowing for the possible influence of oxide phases in the mafic magma and the likely, but in terms of mass balance small, effect caused by zircon in the crustal contaminant, we have adopted a uniform Dz value of 0.2 for all EC-AFC calculations. If above-mentioned reasonable estimates of Cw/Cr ratios (Table 2) are used, an evolution of a magma derived from either of the mantle reservoirs could produce the isotope compositions observed in the zircon (Fig. 4). For the coupled dataset the viable models with an SCLM source allow for more variation in the Cw/Cr ratio and also otherwise seem to produce a better fit than the one for a DM source, but a magma derived from an SCLM source would not be able to produce the more radiogenic Hf isotope values of the uncorrelated dataset (Fig. 4) reported by Heinonen et al. (2010a). An estimate of the amount of crustal material incorporated into the magma during assimilation (M*, mass of assimilant normalized to the original mass of magma) is similar for both source alternatives. Less than 10% input of crustal material is required to produce the observed range in zircon of the leucotroctolite and less than 20% would account for the other anorthositic rocks. This proportion seems to be a reasonable mass-balance estimate considering the boundaries set by major element geochemistry of the anorthositic rocks. It should, however, be noted that the limited resolution of the Hf isotope dataset means that the majority of the constraints derived from the EC-AFC model reflect mostly variation in the O isotope values.

As the heat required to melt the assimilant comes mostly from crystallization of the parental magma, the thermal properties of the EC-AFC model are capable of producing an estimate of the degree of crystallization at the onset of assimilation. For the standard lower crustal case (Bohrson & Spera 2001) adjusted for a Proterozoic geotherm at assumed near-solidus equilibration temperature (Teq = 975 °C), the mass of cumulates (M*) relative to the original mass of magma produced before any contamination had occurred in the system was near 75%. The actual proportion of the original magma exposed to contamination by crustal material thus may have already been relatively evolved at the onset of assimilation. This possible late onset of contamination, therefore, provides a mechanism to explain how zircon crystallizing at a fairly late stage and in a relatively narrow time window during the magmatic evolution of a mafic magma would still be capable of registering mantle isotope signatures from the original source.

The source of the rapakivi granite

The initial Hf isotope compositions of the Finnish rapakivi granites are surprisingly homogeneous (2SD < 2.7 ε units; Heinonen et al. 2010a) compared with other granitic suites that have been studied utilizing the in situ method (e.g. Andersen et al. 2002, 2004, 2007). They also overlap significantly with the evolution of the Palaeoproterozoic Svecofennian crust (Fig. 3). The granites were thus derived either from a crustal source with a homogeneous Hf isotope composition (Heinonen et al. 2010a) or from a heterogeneous source with the isotope compositions homogenized during subsequent magmatic evolution at high temperatures.

Combining the zircon Hf and oxygen isotope data collected from the rapakivi granite A1360 sheds additional light on the source question. The slightly higher δ18 O values (7.4–8.6‰) and lower initial 176 Hf/177 Hf ratios (εHf from –2.1 to +0.5) of the Ahvenisto rapakivi granite compared with those of the anorthositic rocks and the quartz monzodiorite (δ18 O from 6.3 to 7.8‰ and εHf from –1.8 to +3.4, Fig. 4) suggest a separate origin for the different rock types. The mass-balance requirement set by the EC-AFC models provides some additional constraints to rapakivi petrogenesis, as incorporation of c. 30% of crustal material (Fig. 4) into a
mantle-derived anorthosite parental magma, which is most probably not capable of generating a granitic magma. It should, however, be noted that the thermal conditions in the magmatic system are not well constrained and, therefore, we have adopted the thermal EC-AFC parameters for the standard lower crustal case justified by Bohrson & Spera (2001) adjusted for a higher Proterozoic geotherm (initial assimilant temperature of 650 °C). Other thermal parameters have no qualitative effect on the mass-balance estimates but the extent of country rock interaction and therefore also the $M/M_m$ ratio provided by the EC-AFC models are more sensitive to the equilibration temperature parameter $T_{eq}$. We have adopted an intermediate near-solidus value of 975 °C for $T_{eq}$, which is similar to the value used by Bohrson & Spera (2001) and allows for moderate interaction between the magma and wall rock. Deviation from this conservative $T_{eq}$ value is in our opinion not warranted, but would in an extreme case be capable of precluding interaction of the modelled reservoirs altogether or producing somewhat higher estimates of $M/M_m$ ratio. The system, however, reaches full crystallization before attaining significantly higher $M/M_m$ values, which in our opinion suggests that crystallization of anorthositic rocks and the rapakivi granite from a common parental magma is rather unlikely.

The possibility of a source isotopically like the mantle cannot, however, be ruled out only on the basis of isotopic compositions of zircon. The overlap observed in the Hf–O isotope values could result from, for example, remelting of earlier products of crystallization of mantle-derived magmas at lower crustal levels, as implied by the ‘tholeiitic’ model (Frost & Frost 1997) for some of the North American rapakivi-type granites. In this scenario, the small differences between the rock types could be ascribed to secondary (further crustal) contamination of the granitic magma. However, isotopic fractionation between zircon and melt complicates the interpretation of the oxygen isotope data and may account for some of the observed $\delta^{18}O_{melt}$ patterns.

**Effects of oxygen isotope fractionation on $\delta^{18}O_{melt}$**

Isotopic fractionation may lead to crystallization of (in terms of $\delta^{18}O$) heterogeneous populations of zircon from a magma that has initially been homogeneous with respect to oxygen isotope composition, or, conversely, crystallization of zircon with similar oxygen isotope compositions from magmas of different isotopic signatures. Fractionation of oxygen isotopes in magmas is controlled by the crystallization temperature and compositions of the melt and crystallizing assemblage, which means that oxygen isotope compositions of the melt ($\delta^{18}O_{melt}$) can be approximated from $\delta^{18}O_{melt}$ values on the basis of whole-rock geochemistry (e.g. Valley et al. 1994; Bindeman 2008; Lackey et al. 2008).

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Table 2. EC-AFC model details

| Thermal parameters | Isotopic parameters |
|--------------------|---------------------|
| $T_{eq}$ magma     | $Hf$ isotopes       |
| $T_{ini}$ magma    | $C_\delta/ C_{bulk}$|
| $T_{ini}$ assimilant| $D_m$               |
| $T_{liq}$ assimilant| $176$Hf/$177$Hf magma | 0.281853 |
| $T_{liq}$ solidus   | $18$Hf/$177$Hf assimilant | 0.281700 |
| $T_{liq}$          |                      |
|                     |                     |
| $C_{magma}$        | $\delta^{18}O$ magma (%) | 5.3   |
| $C_{assimilant}$   | $\delta^{18}O$ assimilant (%) | 10.2  |
| $H_{ini}$          | $C_a/ C_s$          | 0.94  |
| $H_{liq}$          |                      |
|                     |                     |

Isotope evolution of the magmas is calculated according to the EC-AFC model of Spera & Bohrson (2001) for a ‘standard’ lower crustal case at near-solidus equilibration temperature ($T_{eq}$=975°C) and adjusted for Proterozoic geotherm ($T_{ini}$ assimilant=650°C). Oxygen isotope values for SCLM and DM are mantle zircon values from Valley et al. (1998). Hf isotope value for the Svecofennian crustal assimilant is the maximum whole-rock value reported for Palaeoproterozoic rocks of southern Finland (Hoefs et al. 1994; Bindeman 2008; Lackey et al. 2008). Svecofennian Hf isotope composition is averaged from the Svecofennian crustal evolution of Andersen & Epstein 1969), DM Hf isotope value is calculated according to Griffin et al. (2009), Svecofennian Hf isotope composition is averaged from the Svecofennian crustal assimilant is the maximum whole-rock value reported for Palaeoproterozoic rocks of southern Finland (Hoefs et al. 1994; Bindeman 2008; Lackey et al. 2008). Oxygen isotope value for the Svecofennian crustal assimilant is the maximum whole-rock value reported for Palaeoproterozoic rocks of southern Finland (Hoefs et al. 1994; Bindeman 2008; Lackey et al. 2008).

**Fig. 5.** (a) $\Delta^{18}$O$_{melt-zrn}$ values calculated from normative whole-rock compositions of the studied rocks (Table 1) over the temperature range of 700–1200 °C. Furthest to the right are $\Delta^{18}$O values calculated on the basis of the SiO$_2$ content of the whole-rocks (SiO$_2$ calc.; Lackey et al. 2008). (b) Melt $\delta^{18}O$ values calculated with corresponding $\Delta^{18}$O$_{melt-zrn}$ and average measured $\delta^{18}O_{melt}$ values (Table 3).
Fractionation of oxygen isotopes between the melt and crystallizing zircon in the rocks studied was estimated by two methods (Table 3). We first calculated the total Δ$^{18}$O values for normative mineral compositions over a magmatic temperature range from Δ$^{18}$O$_{min-zrn}$ values in the literature (Valley 2003). The second method was to use a simple formula relating the SiO$_2$ composition of the rocks to their magmatic Δ$^{18}$O$_{melt-zrn}$ value (Lackey et al. 2008), including an estimate of change in the fractionation temperature. Values used and obtained in these calculations are given in Tables 1 and 3 and the results are illustrated in Figure 5. Reliance on the geochemical composition of cumulate rocks renders both of these methods potentially suspect in terms of absolute values, but both indicate a similar pattern. Regardless of the method used, the rapakivi granite A1360 exhibits the highest Δ$^{18}$O$_{melt-zrn}$ values and the leucotroctolite A1271 shows only negligible fractionation. To compare single zircon grains and illustrate the possible effects of fractionation, δ$^{18}$O$_{melt}$ compositions calculated from the temperature-independent SiO$_2$-based Δ$^{18}$O$_{melt-zrn}$ values (Lackey et al. 2008) are shown in Figure 5.

These results suggest that it would be possible to obtain the δ$^{18}$O values observed in the zircon from samples A118, A1306 and A1933 from a magma homogeneous in oxygen isotopes (Figs 5 and 6). However, calculations on samples A1360 and A1271 suggest marked differences in parental δ$^{18}$O$_{melt}$ compositions. Because of the stronger fractionation effect (higher Δ$^{18}$O$_{melt-zrn}$ values) of the granite A1360, it seems very likely that the granitic parental magma had a higher δ$^{18}$O value than the melt that was parental to the anorthositic rocks. Furthermore, the oxygen isotope values calculated for the granitic parental magma overlap with the modelled crustal composition (Fig. 6). Conversely, the low Δ$^{18}$O$_{melt-zrn}$ value derived for the leucotroctolite A1271 suggests relatively lower δ$^{18}$O values for the parental melt.

The reader should also note that unknown fractionation temperatures of the magmas introduce further uncertainties in comparisons of δ$^{18}$O$_{melt}$ values. However, it seems clear that, regardless of absolute values, consideration of these fractionation effects amplifies the isotopic differences between the rock types and, furthermore, the conclusions we have drawn from the δ$^{18}$O$_{zrn}$ values.

**Conclusions**

Combined Hf–O isotope data on zircon from the Ahvenisto rapakivi granite–anorthosite complex provide evidence about the mantle source of the anorthositic rocks in the c. 1.54–1.65 Ga Finnish rapakivi granite suite. Based on the coupled Hf–O dataset of this study, an SCLM-type enriched mantle source would seem to be the most likely candidate for the source of the anorthositic rocks but the previously measured more radiogenic Hf isotope values (Heinonen et al. 2010a) can be explained only by derivation from a more depleted mantle source (DM). The bulk of the anorthositic
rocks in the complex do not show the mantle signal, most probably because of contamination of the mantle-derived magma by crust. Tentative EC-AFC modelling suggests that the observed Hf and O isotope values could have been generated by an AFC process in which a mantle-derived magma was contaminated by Palaeoproterozoic Svecofennian crust. Therefore, the asthenospheric DMM is favoured as the source of the primary magmas of the anorthositic rocks of the Ahvenisto complex. The Ahvenisto rapakivi granite does not show direct isotopic evidence of a mantle signature. From this, considered together with oxygen isotope fractionation effects, it is concluded that the granite was derived from a crustal source, separate from the source of the anorthositic rocks. Thus, petrogenetic models that invoke a single source, either the crust (e.g. Duschenes et al. 1999) or the mantle (e.g. Frost et al. 2002), for the parental magmas of all AMCG rock types cannot be applied to the Ahvenisto complex. Combined Hf–O zircon analysis offers a possible test for mantle-source considerations also in other AMCG complexes. For example, the anorthositic rocks in the Grenville and Nain Provinces display similar oxygen (Peck et al. 2010) and Hf (Bickford et al. 2010) isotope characteristics to those in the Ahvenisto complex. A closer examination using a coupled SIMS–LAM-ICP-MS analysis of zircon in olivine-bearing rocks (where available) in those terranes would be interesting.

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