Absence of hydrothermal oxygen isotope variations in host rocks supports magmatic origin of the giant Grängesberg iron oxide–apatite (IOA) deposit, Central Sweden

Franz Weis1 · Valentin R. Troll1 · Erik Jonsson1,2 · Karin Högååhl1 · Chris Harris3 · Weian Sun1,4 · Katarina P. Nilsson5 · Börje Dahrén1

Received: 7 December 2020 / Accepted: 6 October 2021 / Published online: 13 November 2021
© The Author(s) 2021

Abstract
The origin of Kiruna-type iron oxide–apatite ores is controversial, and debate presently centres on a ‘magmatic’ versus a ‘hydrothermal’ mode of formation. To complement recent investigations on the Grängesberg iron oxide–apatite ore deposit in the northwestern part of the Palaeoproterozoic Bergslagen ore province in central Sweden, we investigated the oxygen isotope composition of the host rocks of this large iron oxide–apatite ore body. As the metavolcanic and metagranitoid country rocks around the Grängesberg ore body either pre-date or are coeval with ore formation, they would be expected to record an extensive isotopic imprint if the ore body had formed by large-scale hydrothermal processes involving an externally sourced fluid. A direct magmatic formation process, in turn, would have produced localized alteration only, concentrated on the immediate vicinity of the ore body. Here, we test these two hypotheses by assessing the oxygen isotope variations in the host rocks around the main Grängesberg iron oxide–apatite ore body. We analysed oxygen isotopes in quartz from metavolcanic (n = 17) and metagranitoid host rocks (n = 14) from the vicinity of the ore body, and up to 2 km distance along and across the strike of the ore body. Remarkably, we find no significant variation in δ18O values with distance from the ore body, or any deviations in country rock δ18O from common magmatic and/or regional values. Only two samples show shifts to values more negative than the common magmatic range, indicating highly localized hydrothermal overprint only. As a large-scale, low-temperature hydrothermal origin of the ore body through voluminous fluid percolation would be expected to have left a distinct imprint on the oxygen isotope values of the country rocks, our results are more consistent with an ortho-magmatic origin for the Grängesberg iron oxide–apatite ore.

Keywords Kiruna type · Iron oxide–apatite ore · Hydrothermal alteration · Host rock · Oxygen isotopes · Grängesberg · Bergslagen · Sweden

Introduction and approach
The genesis of ‘Kiruna-type’ iron oxide–apatite ores is the focus of a long-standing debate (e.g. Geijer 1910, Parak 1975, Weidner 1982, Naslund 1983, Nyström and Henriquez 1994, Barton and Johnson 1996, Sillitoe and Burrows 2002). Genetic models range from a direct magmatic (ortho-magmatic) mode of origin from Fe–P-rich magmas, via formation from high-temperature magmatic fluids by direct crystallization or liquid immiscibility (≥ 800 °C) (cf. Nyström et al. 2008; Jonsson et al. 2013; Weis et al. 2013; Knipping et al. 2015; Bilenker et al. 2016; Tornos et al. 2016, 2017; Hou et al. 2017, 2018; Sun et al. 2019; Troll et al. 2019; Sarlès et al. 2020), to a hydrothermal mode of formation involving transport and precipitation, including replacement-type
be used to identify channelized or pervasive movement of hydrothermal fluids due to a localized or pervasive overprint in the form of, e.g. a zonal or “bullseye” pattern around the ore body, or a more asymmetric pattern if a certain sector or domain was particularly strongly affected (e.g. Taylor 1971, 1974; Nabelek 1987; Beaty et al. 1988; Singleton and Criss 2002; Skelton et al. 2007). In these cases, systematic isotopic gradients have been identified in country rocks over distances ranging from a few hundred metres to several kilometres from the investigated ore bodies or intrusions.

Here, we use quartz grains in addition to whole rock samples, since quartz is usually relatively robust and often resistant to secondary low-temperature alteration processes (cf. Taylor 1968; Larson and Taylor 1986; Dilles et al. 1992; Budd et al. 2017). However, quartz may change its isotope composition by, e.g. infilling of microfractures within individual crystals during hydrothermal alteration (e.g. King et al. 1997). For example, hydrothermal activity around the Kidd Creek VMS deposit affected host rock quartz in the immediate surroundings (~350–400 m) of the ore body, leading to a 5‰ increase from the originally magmatic δ18O values of the quartz. However, a fluid-mediated overprint on the δ18O composition related to later intrusions or regional metamorphism cannot be entirely ruled out (Hyslop et al. 2008), but such processes seem not to have had a significant impact on the isotope compositions of the Grängesberg ore itself (Jonsson et al. 2013). In contrast, hydrothermally precipitated quartz and amorphous forms of silica tend to display comparatively high oxygen isotope values (e.g. >15‰, Götze et al. 2001; Kleine et al. 2018), allowing for a clear distinction of magmatic versus hydrothermally derived types of quartz. Assuming that all sampled country rocks to the ore were present at the time of ore formation and that tectono-thermal activity post-dating ore formation had not fully overprinted the O isotope composition, we can now resolve if traceable isotope compositional gradients typical for hydrothermal alteration exist in the surroundings to the Grängesberg ore body (cf. Taylor 1971, 1974; Dilles et al. 1992; King et al. 1997; Donoghue et al. 2008).

The Grängesberg iron oxide–apatite deposit

Grängesberg is located in the northwestern part of the Palaeoproterozoic Bergslagen ore province in south-central Sweden (Allen et al. 1996, 2008; Stephens et al. 2009; Stephens and Jansson 2020). The Grängesberg mine itself was the largest iron ore mine in southern and central Sweden, and was in operation until 1989. It was closed due to a combination of low metal prices at the time and political reasons, despite considerable ore reserves remaining. Efforts towards a re-opening of the mine are currently underway. The Grängesberg iron oxide–apatite ore body itself is dominated by
massive magnetite ore with subordinate haematite. Both oxides occur, however, also as veins and disseminations in the immediate surroundings of the main ore body (Fig. 1; Magnusson 1970). In the massive magnetite ore, there are characteristic bands of fine-grained fluorapatite, with variable amounts of silicate minerals such as amphiboles, biotite, chlorite as well as diverse REE minerals (Jonsson et al. 2013). The main part of the deposit, the so-called ‘Export Field’, consists of approximately 80% magnetite and 20% haematite in terms of oxide ore minerals (e.g. Jonsson et al. 2013). The main mineralization occurs as massive lenses or “logs” that dip between 50° and 75° towards the south-east and can be traced for more than 900 m at the surface, where its width ranges between 50 and 100 m (Johansson 1910; Geijer and Magnusson 1944).

The Grängesberg iron ore body is hosted in a suite of metavolcanic rocks with basaltic to rhyolitic composition that belongs to the ca. 1.91–1.87 Ga volcano-sedimentary succession of Bergslagen. This was previously known as the “leptite-hälleflinta formation” (e.g. Allen et al. 1996). The majority of metavolcanic rocks in Bergslagen are rhyolitic (and to a lesser extent dacitic) and have been affected by regional K- and/or Na-alteration (Allen et al. 2008; Stephens et al. 2009; Stephens and Jansson 2020, and references therein). The immediate host rocks to the Grängesberg ore are dominated by metavolcanic to meta-subvolcanic rocks of intermediate to basic compositions (Jonsson et al. 2013). Moreover, the ore body as well as the host rocks in the Export Field (Fig. 1) are transected by dykes and sills of mafic to more felsic compositions, as well as by a set of significantly younger granitic pegmatite dykes (Geijer and Magnusson 1944; Jonsson et al. 2011, 2013). Notably, haematite-dominated ore occurs mostly in the structural footwall, and in the vicinity of crosscutting pegmatite dykes. Although most of the haematite in the Grängesberg ore body is believed to be of magmatic origin, some of it, especially in the form of veins and disseminations, most likely is a product of later alteration processes (Jonsson et al. 2013). Alteration zones within the host rocks next to, and within the

![Fig. 1](image-url)
ore body, comprise disseminated and discrete phyllosilicate (biotite, chlorite) and amphibole-rich assemblages (so-called sköl assemblages) and these also contain variable amounts of accessory iron oxides and fluorapatite (Frietsch 1982; Jonsson et al. 2011). Besides these assemblages, there are also traces of alkali alteration (K-feldspar, albite) present in the Grängesberg area. This, however, is a regional feature throughout the whole Bergslagen region, not confined to the Grängesberg area (Allen et al. 2008; Stephens et al. 2009; Stephens and Jansson 2020), and hence not directly associated with the ore-forming process here. In fact, country rocks in the vicinity of the ore body preserve a predominantly igneous chemical character as has previously been noted by Persson et al. (2013), e.g. on the basis of their K₂O and Na₂O relationships (see below). Moreover, the massive magnetite ores seem also unaffected by this alkali alteration or by the regional metamorphism. The hydrothermal alteration assemblages present are, in turn, always spatially related with the ore and are absent beyond the main ore zone and the direct vicinity of the ore body (Jonsson et al. 2013; Persson et al. 2013), implying a genetic relationship between the two and the absence of a pronounced mineralogical aureole to the Grängesberg ore body.

The host rocks and the Grängesberg iron oxide–apatite ores have been affected by at least two phases of ductile deformation during the ca. 1.9–1.8 Ga Svecokarelian orogeny (e.g. Stephens et al. 2009; Högdahl et al. 2013). If a larger-scale hydrothermal “aureole” was generated during ore formation, i.e. prior to deformation and metamorphism, the O isotope pattern around the deposit might be expected to have been affected by these hydrothermal processes (cf. Taylor 1971, 1974; Beaty et al. 1988; Dilles et al. 1992; Ayers et al. 2006). Thus, some pattern of variation in isotopic values around the deposit would be expected, for instance in the form of a folded pattern or as a seemingly random mixture of high and low values due to, e.g. channelized flow and later deformation, despite later overprinting processes. As the main host rock in the hanging wall east of the ore body, a gneissic granitoid with a strong stretching lineation, is possibly younger than the ore itself (Jonsson et al. 2011; Högdahl et al. 2013), a recorded alteration pattern would primarily be expected in the foot wall to the west of the Export Field, whereas the east of the Export Field would be more representative of regional oxygen isotope signatures.

**Methods**

Samples (n=23) of the host rocks to the Grängesberg ore body have been collected from outcrops in the area around the iron oxide–apatite deposit in a suite of traverses up to 2 km away from the main ore body (Figs. 2, 3, 4; Table 1). From these samples, we analysed whole rock (n=17) and quartz separates (n=14) for their oxygen isotope compositions (Table 1).

Whole rocks were analysed with a conventional vacuum extraction line at the University of Cape Town, South Africa. About 10 mg of sample was dried overnight at 50 °C and then loaded into Ni tubes. The samples were then degassed at 200 °C for 2 h under vacuum, then reacted with ClF₃ for 4–6 h at 550 °C. The liberated oxygen was passed over a hot carbon rod and converted to CO₂, which was then collected in glass tubes. The in-house standard MQ was run in duplicate with each batch of eight samples and used to convert the raw data to the SMOW scale; MQ has a δ¹⁸O value of 10.1‰ calibrated against NBS-28, assuming a value of 9.64‰ (Coplen et al. 1983). Quartz was analysed by the laser fluorination method described by Harris and Vogeli (2010), whereby they were reacted with 10 kPa of Br₂ and the purified O₂ was collected onto a 5 Å molecular sieve in a glass storage bottle. Oxygen isotope analysis was carried out using a Finnigan DeltaXP dual inlet gas source mass spectrometer. The in-house Monastery garnet (MON GT) standard was analysed in duplicate with each batch of ten samples. The MON GT was calibrated against UWG-2 garnet standard from Valley et al. (1995) and has a δ¹⁸O= 5.38‰ (Harris and Vogeli 2010), which assumes δ¹⁸O = 5.80‰ for the UWG-2. All raw data were normalized to a MON GT value of 5.38‰. All oxygen data were recorded in the usual δ notation relative to SMOW where δ¹⁸O = (Rsample/Rstandard − 1) × 1000, and R = the measured ratio ¹⁸O/¹⁶O. Based on the long-term duplication of both in-house standards, the δ¹⁸O values have 2σ errors of 0.2‰. Detailed information on sample preparation is given in Harris et al. (2015).

**Results**

The country rock samples comprise intermediate- to high-silica metavolcanic and metaplutonic rocks and thin section images of representative rock samples are provided in Fig. 3. All samples and analytical data are reported in Table 1. The assemblages are not particularly altered for most parts and show predominantly igneous mineralogical and chemical signatures (Figs. 2, 3, 4), although some metamorphic overprint and regional alkali alteration are detected in some samples. With respect to oxygen isotope systematics, the Grängesberg metavolcanic and metaplutonic host rocks record δ¹⁸O values exclusively between +5.8 and +8.6‰, which essentially overlaps with the range of δ¹⁸O values for ‘normal’ intermediate arc magmas (cf. Taylor 1968; Bindemann 2008; Budd et al. 2017, Fig. 5; Table 2). Quartz from meta-granites and meta-dacites in the vicinity around the ore body exhibits δ¹⁸O values from +6.0 to +10.3‰, with the majority...
showing values around +8 and +9‰ (n = 9), which is also within the range of igneous quartz reported in the literature and considerably below the values typical for hydrothermal quartz (cf. Taylor 1968; Larson and Taylor 1986; Blatt 1987; Chiba et al. 1989; Ghent and Valley 1998; Götze et al. 2001; Bindeman 2008; Jourdan et al. 2009; Budd et al. 2017; Kleine et al. 2018; Fig. 4). Equilibrium magma compositions have been calculated from the δ18O values in quartz by applying the appropriate fractionation factors between quartz and andesite, dacite and rhyolite melts at the corresponding temperatures (1000 °C, 900 °C and 800 °C); (Δquartz-andesite = +0.91‰, Δquartz-dacite = +0.84‰, Δquartz-rhyolite = +0.77‰, Zhao and Zheng 2003). Grängesberg host rock quartz samples with δ18O ≥ +6.7‰ (n = 11) yield equilibrium magma isotope compositions between δ18O = +6.1 and +9.5‰ (Table 2), with the majority of calculated magma isotope values falling between +7.0 and +8.5‰ (n = 11). This range corresponds to the measured whole rock data from the area (Table 1) and to values characteristic for intermediate- to high-silica igneous rocks in general (cf. Bindeman 2008). The calculation of equilibrium magmas for the spectrum of relevant rock types (magma compositions) that occur in the Grängesberg area.

Fig. 2 Field photographs of major rock types in the Grängesberg area (see also Table 1). a Coarse porphyric, granitic gneiss with mafic enclaves. b Fine and even-grained felsic metavolcanic rock with a pronounced east-dipping foliation. c Feldspar–porphyric, phyllosilicate altered felsic metavolcanic rock. d Very fine-grained banded and laminated felsic volcanic rock with skarn layers (green and brown). e Plagioclase–porphyric, massive, felsic volcanic rock that is associated with amphibolites. f Mafic (metabasic) amphibolite with plagioclase porphyroclasts.
shows that little variation of magmatic values is in fact expected in the region, strengthening the coherence of our result of a magmatic origin of the majority of analysed quartz grains. A small number of quartz grains, however, have $\delta^{18}O \leq +6.7\%$ ($n = 3$) and are not consistent with magmatic values. This may be because the quartz in these samples was disturbed by later infilling of microfractures during medium- to high-temperature (> 400 °C) hydrothermal processes or that the separated quartz contained vein or amygdale quartz that had formed at relatively high temperatures from localized hydrothermal processes (cf. King et al. 1997; Zheng et al. 1999; Donoghue et al. 2010; Kleine et al. 2018).

Fig. 3 Photomicrographs of major rock types in the vicinity of the Grängesberg iron oxide–apatite ore deposit. a Representative granitic gneiss. b Biotite-bearing, foliated fine-grained felsic metavolcanic rock. c Phyllosilicate-rich felsic metavolcanic rock, in this example with muscovite and phlogopite. d Layered and laminated felsic metavolcanic rock with alternating fine-grained and very fine-grained layers. e, f Plagioclase–porphyric metavolcanic rock associated with amphibolite and larger plagioclase phenocrysts. a–e are in crossed polarised light, f is in plane-polarised light. The scale bar is 1000 µm.
Discussion

The δ18O values for pristine intermediate to felsic volcanic rocks, such as arc-derived andesites and dacites, are typically between +6.0 and ~ +8.0‰ and are higher than primitive magmas derived from the depleted upper mantle (5.7 ± 0.3‰), due to a combination of fractional crystallization and crustal contamination (Hoefs 1997; Eiler 2001; Bindeman 2008). During hydrothermal alteration of volcanic rocks, circulating water will lead to isotopic exchange with the rock (e.g. Taylor 1974). Since the infiltrating water in most hydrothermal systems will have had lower δ18O values than the rock, water–rock interaction will lower the δ18O in the rock, provided that the fractionation factor between rock and water is low enough (cf. Taylor 1971, 1974; Hoefs 1997; Faure and Mensing 2005; Donoghue et al. 2010; Kleine et al. 2018). In contrast, an increase in δ18O of the rocks above magmatic values is usually caused by fluid rock interaction at low temperatures (< 300 °C), most often through the breakdown of original minerals and precipitation of secondary minerals (e.g. transformation of feldspar to clay minerals) that tend to have high δ18O, thus causing

![Fig. 4 Alkali oxide variation diagram (K2O + Na2O vs 100 × K2O/(K2O + Na2O); Hughes 1973) for country rock samples to the Grängesberg ore body as determined by Persson et al. (2013). The country rocks show predominantly unaltered igneous character, although some samples are affected by contact metamorphic processes and some others show the effects of regional alkali alteration (see text for details).](image)

| Table 1 | Country rock sample set Grängesberg iron oxide–apatite deposit, Central Sweden |
|---------|--------------------------------------------------------------------------------|
| Sample  | Description                                                                 |
| VRT-GMD29 | Reddish-grey metagranite, foliated                                           |
| VRT-GMD31 | Greyish-pink metagranite, foliated                                           |
| VRT-GMD40 | Greyish metagranite, foliated                                                |
| VRT-GMD45 | Greyish-pink biotite-rich metagranite, foliated                              |
| VRT-GMD48 | Grey biotite-rich granitic gneiss, foliated                                  |
| VRT-GMD49 | Dark-grey granitic gneiss, foliated                                          |
| VRT-GMD52 | Grey-pinkish metagranite, foliated                                           |
| VRT-GMD53 | Pale pink-coloured metavolcanic rock, foliated                              |
| VRT-GMD56 | Grey-pinkish granite, isotropic                                              |
| KPN-090007A | Grey-reddish intermediary metavolcanic rock, foliated                      |
| KPN-090026-4 | Grey-pink felsic metavolcanic rock, foliated                              |
| KPN-090033A | Grey-pinkish intermediary metavolcanic rock, foliated                      |
| KPN-090042A | Grey-pink felsic metavolcanic rock, foliated                               |
| KHO-09003 | Reddish-grey felsic metavolcanic rock with rounded magnetite, foliated     |
| KHO-09005 | White Na-altered felsic metavolcanic rock with magnetite disseminations, foliated |
| KHO-090010 | Grey felsic metavolcanic rock, foliated                                       |
| KHO-090011A | Grey fine-graind felsic metavolcanic rock, foliated                        |
| KHO-090012 | Pale grey, fine-grained felsic metavolcanic rock, foliated                 |
| KHO-090013 | Grey biotite-rich intermediary metavolcanic rock, foliated                 |
| KHO-090013G | Greyish-green amphibolite, metagneous, foliated                            |
| KHO-090017 | Pink fine-grained felsic metavolcanic rock, foliated                        |
| KHO-09127b | Grey felsic metavolcanic rock, foliated                                       |
| KHO-09129A | Fine-grained granitoid metavolcanic rock                                    |

Table 1 | Coordinates RT90 | Quartz δ18O (‰) | 2σ | Whole rock δ18O (‰) | 2σ |
|---------|------------------|-----------------|----|---------------------|----|
| VRT-GMD29 | 1,454,958, 6,662,272 | 7.9 ± 0.2 – – |  – | – |
| VRT-GMD31 | 1,456,585, 6,661,782 | 6.7 ± 0.2 – – |  – | – |
| VRT-GMD40 | 1,456,305, 6,661,990 | 9.3 ± 0.2 – – |  – | – |
| VRT-GMD45 | 1,454,534, 6,661,130 | 8.3 ± 0.2 – – |  – | – |
| VRT-GMD48 | 1,454,386, 6,660,546 | 8.4 ± 0.2 7.9 ± 0.2 |  – | – |
| VRT-GMD49 | 1,455,395, 6,661,477 | 9.2 ± 0.2 7.8 ± 0.2 |  – | – |
| VRT-GMD52 | 1,456,209, 6,662,922 | 8.7 ± 0.2 6.9 ± 0.2 |  – | – |
| VRT-GMD53 | 1,454,816, 6,662,790 | – 8.4 ± 0.2 |  – | – |
| VRT-GMD56 | 1,455,284, 6,661,802 | – 7.4 ± 0.2 |  – | – |
| KPN-090007A | 1,454,596, 6,663,236 | – 5.8 ± 0.2 |  – | – |
| KPN-090026-4 | 1,453,806, 6,662,229 | 9.1 ± 0.2 6.8 ± 0.2 |  – | – |
| KPN-090033A | 1,453,794, 6,662,943 | 10.3 ± 0.2 8.6 ± 0.2 |  – | – |
| KPN-090042A | 1,453,884, 6,663,419 | 6 ± 0.2 7.5 ± 0.2 |  – | – |
| KHO-09003 | 1,454,400, 6,663,551 | – 7.7 ± 0.2 |  – | – |
| KHO-09005 | 1,454,562, 6,663,500 | 7 ± 0.2 5.8 ± 0.2 |  – | – |
| KHO-090010 | 1,454,196, 6,661,292 | – 6.8 ± 0.2 |  – | – |
| KHO-090011A | 1,454,420, 6,661,821 | – 6.8 ± 0.2 |  – | – |
| KHO-090012 | 1,454,571, 6,661,728 | 6.7 ± 0.2 7.0 ± 0.2 |  – | – |
| KHO-090013 | 1,454,467, 6,661,972 | – 7.7 ± 0.2 |  – | – |
| KHO-090013G | 1,454,473, 6,661,971 | – 6.0 ± 0.2 |  – | – |
| KHO-090017 | 1,453,588, 6,660,757 | 8.8 ± 0.2 – – |  – | – |
| KHO-09127b | 1,454,632, 6,662,214 | – 8.3 ± 0.2 |  – | – |
| KHO-09129A | 1,453,535, 6,662,113 | 8.6 ± 0.2 – – |  – | – |
Fig. 5  

a Distribution of surface samples used in this study around the Grangesberg ore body (the red ellipse) and their respective oxygen isotope composition. The area presented in Fig. 1b is indicated by the black box.  
b The positions of two traverses (T1 in E–W and T2 in N–S orientation) are indicated by yellow bars across the Grangesberg area, comprising the samples that are shown in Fig. 4. Notably, no major variation beyond the regular magmatic values is apparent (Source of base map: Google Earth)
the opposite effect to medium to high-temperature hydrothermal overprinting (cf. Criss and Taylor 1986; Rhodes and Oreskes 1999; Faure and Mensing 2005; Donoghue et al. 2008, 2010; Berg et al. 2018). Therefore, if a hydrothermal origin (e.g. 300–500 °C) for the Grängesberg ore body would apply, as suggested for other ore deposits including some iron oxide–apatite ores (e.g. Taylor 1974; Parak 1975; Smith et al. 2013; Dare et al. 2014; Westhues et al. 2016, 2017), a widespread, systematic and characteristic low δ18O alteration pattern in the host rocks in the vicinity of the ore would be expected as a result (cf. Taylor 1974; Beaty et al. 1988; Singleton and Criss 2002). The magnitude of these effects would depend on the δ18O value of the input surface water, and the water/rock ratio, but the effect would be concentric variation of δ18O around the focus of hydrothermal activity, especially since the palaeoenvironment is interpreted as shallow marine back-arc-setting, an at least in part open system with fluid influx of marine and/or meteoric isotopic compositions would be envisaged. Tornos et al. (2016) find that hydrothermal alteration at the El Laco volcano in Chile involved fluids at temperatures around 900 °C, which would suggest a dominantly magmatic fluid (δ18O usually 5–10‰; Hoefs 1997). Even if such a fluid is considered, hydrothermal fluids circulating in the host rock volume around the Grängesberg ore would experience distal changes in relation to the host rocks, due to progressive cooling of the magmatic fluids away from the ore body and consequent changes in fractionation behaviour. Moreover, progressive interaction between the percolating fluids and the host rocks would lead to a changing fluid composition and hence a notable isotope variation away from the ore body, thus, temperature changes and progressive fluid–rock interaction ought to leave a detectable variation in the aureole surrounding a large hydrothermal ore deposit (e.g. Taylor 1974; Nabelek 1987; Berg et al. 2018). In contrast, low-temperature hydrothermal fluids that circulate through various host rocks (e.g. emanating from a medium temperature hydrothermal ore deposit) may be representing the distal effects of a high-temperature hydrothermal system, and will lead to an increase in whole rock δ18O of the host rocks.

| Sample     | Quartz δ18O (‰) | Whole rock δ18O (‰) | 2σ | Re-calculated andesite δ18O (‰) | Re-calculated dacite δ18O (‰) | Re-calculated rhyolite δ18O (‰) |
|------------|-----------------|----------------------|----|-------------------------------|-------------------------------|-------------------------------|
| VRT-GMD29  | 7.9             | –                    | ±0.2 | 7.0                          | 7.1                           | 7.1                           |
| VRT-GMD31  | 6.7             | –                    | ±0.2 | 5.8                          | 5.8                           | 5.9                           |
| VRT-GMD40  | 9.3             | –                    | ±0.2 | 8.4                          | 8.4                           | 8.5                           |
| VRT-GMD45  | 8.3             | –                    | ±0.2 | 7.4                          | 7.4                           | 7.5                           |
| VRT-GMD48  | 8.4             | 7.9                  | ±0.2 | 7.5                          | 7.6                           | 7.6                           |
| VRT-GMD49  | 9.2             | 7.8                  | ±0.2 | 8.3                          | 8.4                           | 8.5                           |
| VRT-GMD52  | 8.7             | 6.9                  | ±0.2 | 7.8                          | 7.9                           | 8.0                           |
| VRT-GMD53  | –               | 8.4                  | ±0.2 | –                            | –                             | –                             |
| VRT-GMD56  | –               | 7.4                  | ±0.2 | –                            | –                             | –                             |
| KPN-090007 | –               | 5.8                  | ±0.2 | –                            | –                             | –                             |
| KPN-090026-4 | 9.1           | 6.8                  | ±0.2 | 8.2                          | 8.3                           | 8.3                           |
| KPN-090033A | 10.3           | 8.6                  | ±0.2 | 9.4                          | 9.5                           | 9.5                           |
| KPN-090042A | 6.0             | 7.5                  | ±0.2 | 5.1                          | 5.2                           | 5.2                           |
| KHO-09003  | –               | 7.7                  | ±0.2 | –                            | –                             | –                             |
| KHO-09005  | 7.0             | 5.8                  | ±0.2 | 6.1                          | 6.2                           | 6.2                           |
| KHO-090010 | –               | 6.8                  | ±0.2 | –                            | –                             | –                             |
| KHO-090011A | –              | 6.8                  | ±0.2 | –                            | –                             | –                             |
| KHO-090012 | 6.7             | 7.0                  | ±0.2 | 5.8                          | 5.9                           | 5.9                           |
| KHO-090013 | –               | 7.7                  | ±0.2 | –                            | –                             | –                             |
| KHO-090013G | –              | 6.0                  | ±0.2 | –                            | –                             | –                             |
| KHO-090017 | 8.8             | –                    | ±0.2 | 7.9                          | 8.0                           | 8.0                           |
| KHO-09127b | –               | 8.3                  | ±0.2 | –                            | –                             | –                             |
| KHO-09129A | 8.6             | –                    | ±0.2 | 7.6                          | 7.7                           | 7.8                           |

Fractionation factors for equilibrium calculations were taken from Zhao and Zheng (2003)

\[\frac{1000 \ln \alpha_{(mt-andesite)}}{\delta^{18}O} = +0.91 \%\]
\[\frac{1000 \ln \alpha_{(mt-dacite)}}{\delta^{18}O} = +0.84 \%\]
\[\frac{1000 \ln \alpha_{(mt-rhyolite)}}{\delta^{18}O} = +0.77 \%\]
due to a reversal of fractionation factors (e.g. Taylor 1974; Donoghue et al. 2008; Berg et al. 2018). In such a case, an increase in \( \delta^{18}O \) values at distance from the ore body would be expected if fluid circulation was an important process. In the Grängesberg case, we consider host rocks in the tectono-stratigraphic hanging wall to the east of the Grängesberg ore body to be possibly younger than the ore and thus to possibly reflect a pristine original composition or regional pattern unrelated to the ore formation, while a potential offset pattern would be expected to the west of the Export Field in case hydrothermal transport and deposition of metals was the main ore-forming process. There are, however, no distinct difference between the western and eastern part in the \( \delta^{18}O \) distribution around the ore body and the \( \delta^{18}O \) values do not show any systematic patterns for either the whole rock or quartz data with regard to geographical distribution (Figs. 4, 5). Indeed, there is very little variation beyond a regular range of essentially magmatic values (+6.0 to +8.0‰; Fig. 5) and, furthermore, there is no real variation from this scheme in whole rock \( \delta^{18}O \) data presented by Jonsson et al. (2013) for host rocks sampled from drill cores (from 570 and 650 m depth) through the Grängesberg ore zone. These drill core samples also range from +5 to +9‰, with the majority of values between +6 to +7‰, which are thus consistent with the predominantly magmatic isotope composition derived from our surface survey. Furthermore, we note that calculated equilibrium magma compositions for ore magnetite from Grängesberg also yield calculated equilibrium magma \( \delta^{18}O \) values of between +6 and +8‰, show, moreover, magmatic Fe isotope values (Jonsson et al. 2013; Troll et al. 2019), and thus align with the magmatic range of the host rock data. These values are also consistent with the theoretical equilibrium magma compositions for our quartz samples, which show calculated \( \delta^{18}O \) magma values between +6.1 and +9.5‰, even when applying fractionation factors for the whole spectrum of intermediate to felsic magmas available in the entire Grängesberg region (Fig. 5, Table 2). The results are thus consistent with a dominantly magmatic mode of origin for the country rocks and the Grängesberg iron ore deposit.

The quartz phenocrysts \((n = 14)\) in the host rocks to the Grängesberg ore have an average \( \delta^{18}O \) value of +8.2‰, a value representative of relatively normal igneous quartz (e.g. Larson and Taylor 1986; Blatt 1987; Ghent and Valley 1998; Bindeman 2008; Budd et al. 2017). Even though quartz is often seen as being relatively resistant to hydrothermal alteration, as it exchanges oxygen isotopes very slowly with hydrothermal fluids, some potential alteration can be seen in the Grängesberg samples. The quartz samples which exhibit \( \delta^{18}O \) values below 7‰ \((n = 3)\) were most likely influenced by localized high-temperature fluid–rock interaction (cf. Donoghue et al. 2010; Kleine et al. 2018) that would be expected for a high-temperature magmatic setting, or possibly by fluid circulation during regional metamorphism. Our whole rock \( \delta^{18}O \) data are almost exclusively within the range of common igneous rocks (Fig. 5) and only two whole rock samples from the northern part of the Grängesberg area, one intermediate and one felsic metavolcanic rock, exhibit a \( \delta^{18}O \) value of 5.8‰, and are thus at the lower end of the typical range for intermediate magmas of +6 to 8‰ (Fig. 6).

One of these two outlier samples is rich in magnetite disseminations, which will lower the \( \delta^{18}O \), while the other appears to record the effect of localized alteration with a
mineral replacement texture visible on inspection. These rare alteration phenomena seen in the wider array of host rocks to the Grängesberg ore body contrast the hydrous alteration assemblages, such as the biotite, chloride or amphibole-rich assemblages that are present within the Grängesberg ore body and its direct contact zone, and which are by now interpreted to be directly associated with the emplacement of the ore body (Jonsson et al. 2013; Troll et al. 2019). In summary, apart from rare and localized exceptions, our data do not exhibit any major offset of δ18O values from the regional average in the wider range of host rocks around the main iron oxide–apatite ore zone at Grängesberg, which implies that a large-scale hydrothermal circulation of external fluid into the host rocks of the Grängesberg ore body was likely not taking place during the ore-forming event. Therefore, the magmatic δ18O values recorded in the host rocks are more consistent with an igneous rather than a hydrothermal origin of the main ore body and are in line with the magmatic δ18O and δ56Fe values from massive magnetite ore samples from drill core material in the main Grängesberg ore zone (Jonsson et al. 2013; Troll et al. 2019).

**Conclusions**

Quartz and whole rock isotope data from host rocks to the Grängesberg iron oxide–apatite ore do not show any systematic offset relative to normal magmatic values, and no distinct isotopic variation is observed on either side of the main ore zone (the The Export field) at Grängesberg. In particular the footwall rocks of the ores that are older or synchronous with ore formation do not seem to have been affected by any intense hydrothermal overprinting, as would be expected if this very large iron oxide ore body had formed from dominantly hydrothermal processes. A long-time, steady-state hydrothermal system capable of forming this type of large-scale deposit, such as sub-seafloor replacements or seafloor exhalations, for example, would generate a spatially widespread offset of primary isotope values, which is not the case here. Lower than ‘normal’ magmatic δ18O values are observed in a very small number of country rock samples only, and when present, they are from locations relatively close to the ore body. The absence of a large-scale and marked δ18O “hydrothermal aureole” to the Grängesberg ore, coupled with the broadly magmatic stable isotope values of the host rocks and the complementary magmatic data from the centre of the Grängesberg ore body, shows that there is a) no significant variation across the area and b) that there is a broadly similar magmatic signal preserved in the ore as well as in the country rocks. The new δ18O country rock data from around the Grängesberg ore thus corroborate the interpretation of a mainly ortho-magmatic origin of the Kiruna-type iron oxide–apatite ore deposit at Grängesberg.

**Acknowledgements** We thank Fayrooza Rawoot for help with the O isotope analyses at Cape Town University. We are, moreover, grateful to Alasdair Skelton, Ulf Bertil Andersson, Taofa Zhou, Noel White and Fernando Tornos for inspiring discussions. Finally, we are grateful to the Geological Survey of Sweden (SGU), the Swedish Research Council (VR), and the National Research Foundation of South Africa (NRSA) for financial support of our work.

**Funding** Open access funding provided by Uppsala University.

**Open Access** This article is licensed under a Creative Commons Attribution 4.0 International License, which permits use, sharing, adaptation, distribution and reproduction in any medium or format, as long as you give appropriate credit to the original author(s) and the source, provide a link to the Creative Commons licence, and indicate if changes were made. The images or other third party material in this article are included in the article’s Creative Commons licence, unless indicated otherwise in a credit line to the material. If material is not included in the article’s Creative Commons licence and your intended use is not permitted by statutory regulation or exceeds the permitted use, you will need to obtain permission directly from the copyright holder. To view a copy of this licence, visit http://creativecommons.org/licenses/by/4.0/.

**References**

Allen RL, Lundström I, Ripa M, Simeonov A, Christofferson H (1996) Facies analysis of a 1.9 Ga, continental margin, back-arc, felsic caldera province with diverse Zn–Pb–Ag–(Cu–Au) sulfide and Fe oxide deposits, Bergslagen Region, Sweden. Econ Geol 91:979–1006

Allen R, Ripa M, Jansson N (2008) Palaeoproterozoic volcanic-and limestone-hosted Zn–Pb–Ag–(Cu–Au) massive sulphide deposits and Fe oxide deposits in Bergslagen, Sweden. IGC Excursion Nr. 12:16–21

Ayers JC, Loflin M, Miller CR, Barton MD, Coath CD (2006) In situ oxygen isotope analysis of monazite as a monitor of fluid infiltration during contact metamorphism: Birch Creek Pluton aureole, White Mountains, eastern California. Geology 34:653–656

Barton MD, Johnson DA (1996) Evaporitic-source model for igneous-related Fe oxide–(REE-Cu-Au-U) mineralization. Geology 24:259–262

Beaty DW, Taylor HP, Coad PR (1988) An oxygen isotope study of the Kidd Creek, Ontario, volcanogenic massive sulfide deposit: evidence for a high 18O ore fluid. Econ Geol 83:1–17

Berg SE, Troll VR, Harris C, Deegan FM, Riishuus MS, Burchardt S, Krumholz M (2018) Exceptionally high whole-rock δ18O values in intra-caldera rhyolites from Northeast Iceland. Min Mag 82:1147–1168

Bilenker LD, Simon AC, Reich M, Lundstrom CC, Gajos N, Bindeman I, Barra F, Munizaga R (2016) Fe–O stable isotope pairs elucidate a high-temperature origin of Chilean iron oxide–apatite deposits. Geochim Cosmochim Acta 177:94–104

Bindeman I (2008) Oxygen isotopes in mantle and crustal magmas as revealed by single crystal analysis. Rev Mineral Geochem 69:445–478

Blatt H (1987) Oxygen isotopes and the origin of quartz: perspective. J Sediment Res 57:373–377

Budd DA, Troll VR, Deegan FM, Jolis EM, Smith VC, Whitehouse MJ, Harris C, Greda C, Hilton DR, Halldorsson SA, Bindeman
host rocks to the Malmberget iron oxide–apatite deposit, Sweden. Precambir Res 342:1056–52
Sillitoe RH, Burrows DR (2002) New field evidence bearing on the origin of the El Laco magnetite deposit, northern Chile. Econ Geol 97:1101–1109
Singleton MJ, Criss RE (2002) Effects of normal faulting on fluid flow in an ore-producing hydrothermal system, Comstock Lode, Nevada. J Volcanol Geotherm Res 115:437–450
Skelton A, Hode Vuorinen J, Arghe F, Falllick A (2007) Fluid–rock interaction at a carbonatite-gneiss contact, Alnö, Sweden. Contrib Min Pet 154:75–90
Smith MP, Gleeson SA, Yardley BWD (2013) Hydrothermal fluid evolution and metal transport in the Kiruna District, Sweden: contrasting metal behaviour in aqueous and aqueous-carbonic brines. Geochim Cosmochim Acta. https://doi.org/10.1016/j.gca.2012.10.015
Stephens MB, Jansson N (2020) Paleoproterozoic (1.9–1.8 Ga) synorogenic magmatism, sedimentation and mineralization in the Bergslagen lithotectonic unit, Svecofennian orogen. In Sweden: Lithotectonic framework, tectonic evolution and mineral resources. Eds Stephens MB, Bergman Weihed j, Memoirs of the Geol Soc Lond 50:155–206
Stephens MB, Ripa M, Lundström I, Persson L, Bergman T, Ahl M, Wahlgren C-H, Persson P-O, Wickström L (2009) Synthesis of bedrock geology in the Bergslagen region, Fennoscandian Shield, south-central Sweden. Geol Surv Sweden Ser Ba 58:1–259
Sun W, Yuan F, Jowitt SM, Zhou T, Liu G, Li X, Wang F, Troll VR (2019) In situ LA-ICP-MS trace element analyses of magnetite: genetic implications for the Zhonggu orefield, Ningwu volcanic basin, Anhui Province, China. Min Depos 54:1243–1264
Taylor HP (1968) The oxygen isotope geochemistry of igneous rocks. Contrib Min Pet 19:1–17
Taylor HP (1971) Oxygen isotope evidence for large-scale interaction between meteoric ground waters and tertiary granodiorite intrusions, Western Cascade Range, Oregon. J Geophys Res 76:855–8774
Taylor HP (1974) The application of oxygen and hydrogen isotope studies to problems of hydrothermal alteration and ore deposits. Econ Geol 69:843–883
Tornos F, Velasco F, Hanchar JM (2016) Iron-rich melts, magmatic magnetite, and superheated hydrothermal systems: the El Laco deposit, Chile. Geology 44:427–430
Tornos F, Velasco F, Hanchar JM (2017) The magmatic to magmatic-hydrothermal evolution of the El Laco deposit (Chile) and its implications for the genesis of magnetite–apatite deposits. Econ Geol 112:1595–1628
Troll VR, Weis FA, Jonsson E, Andersson UB, Majidi SA, Högdahl K, Harris C, Millet M-A, Chinnasamy SS, Kooijman E, Nilsson KP (2019) Global Fe–O isotope correlation reveals magmatic origin of Kiruna-type apatite–iron-oxide ores. Nat Commun 10:1712
Valley JW, Kitchen N, Kohn MJ, Niendorf CR, Spicuzza MJ (1995) UWG-2, a garnet standard for oxygen isotope ratios: strategies for high precision and accuracy with laser heating. Geochim Cosmochim Acta 59:5223–5231
Weidner JR (1982) Iron-oxide magmas in the system Fe–C–O. Can Mineral 20:555–566
Weis F, Troll VR, Jonsson E, Högdahl K, Barker A, Harris C, Millet M-A, Nilsson KP (2013) Iron and oxygen isotope systematics of apatite–iron-oxide ores in central Sweden. In: Abstract volume, SGA biennial meeting 2013, Uppsala, Sweden, pp 1675–1678
Westhues A, Hanchar JM, Whitehouse MJ, Martinsson O (2016) New constraints on the timing of host-rock emplacement, hydrothermal alteration, and iron oxide–apatite mineralization in the Kiruna District, Norrbotten, Sweden. Econ Geol 111:1595–1618
Westhues A, Hanchar JM, Voisey CR, Whitehouse MJ, Rossman GR, Wirth R (2017) Tracing the fluid evolution of the Kiruna iron oxide apatite deposits using zircon, monazite, and whole rock trace elements and isotopic studies. Chem Geol 466:303–322
Zhao Z-F, Zheng Y-F (2003) Calculation of oxygen isotope fractionation in magmatic rocks. Chem Geol 193:59–80
Zheng Y-F, Fu B, Xiao Y-L, Li Y-L, Gong B (1999) Hydrogen and oxygen isotope evidence for fluid–rock interactions in the stages of pre- and post-UHP metamorphism in the Dabie Mountains. Lithos 46:677–693