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ORIGINAL PAPER

# Low-pressure, water-assisted anatexis of basic dykes in a contact metamorphic aureole, Fuerteventura (Canary Islands): oxygen isotope evidence for a meteoric fluid origin

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**Abstract** Migmatites produced by low-pressure anatexis of basic dykes are found in a contact metamorphic aureole around a pyroxenite–gabbro intrusion (PX2), on Fuerteventura. Dykes outside and inside the aureole record interaction with meteoric water, with low or negative  $\delta^{18}\text{O}$  whole-rock values (+0.2 to  $-3.4\text{‰}$ ), decreasing towards the contact. Recrystallised plagioclase, diopside, biotite and oxides, from within the aureole, show a similar evolution with lowest  $\delta^{18}\text{O}$  values ( $-2.8$ ,  $-4.2$ ,  $-4.4$  and  $-7.6\text{‰}$ , respectively) in the migmatite zone, close to the intrusion. Relict clinopyroxene phenocrysts preserved in all dykes, retain typically magmatic  $\delta^{18}\text{O}$  values up to the anatectic zone, where the values are lower and more heterogeneous. Low  $\delta^{18}\text{O}$  values, decreasing towards the intrusion, can be ascribed to the advection of meteoric water during magma emplacement, with increasing fluid/rock ratios (higher dyke intensities towards the intrusion acting as fluid-pathways) and higher temperatures promoting increasing exchange during recrystallisation.

**Keywords** Basic dykes · Contact aureole · Meteoric water · Oxygen isotopes · Partial melting

## Introduction

Partial melting of basic igneous rocks in low-pressure contact metamorphic environments is a rarely documented phenomenon, as geological settings combining a high heat flow and shallow crustal levels, are uncommon. This process is, however, likely to occur at mid-ocean ridges and in oceanic plateaus (LIPs; Large Igneous Provinces). Flagler and Spray (1991) document anatectic plagiogranites in amphibolitised gabbros from oceanic fragments within high temperature shear zones, near a spreading centre in the Shetland Islands. Bédard (1991) and Mevel (1988) also report occurrences of anatexis in oceanic crust or related units. However, the in situ anatexis of basic rocks in an ocean island setting has, to our knowledge, only been described in Fuerteventura Basal Complex (Hobson et al. 1998, 2000).

The Basal Complex of Fuerteventura, Canary Islands, represents the hypabyssal root zone of an ocean island volcano, exposure of which can be attributed to uplift and subsequent erosion of the overlying formations. This complex records a long magmatic activity characterised by the intrusion of numerous magma batches as small plutons, dykes, dyke swarms, and ring-dyke complexes of alkali-gabbros, pyroxenites, syenites and carbonatites (Cantagrel et al. 1993; Demény et al. 1998; Hoernle et al. 2002). The high heat flow generated by the successive magma pulses feeding the overlying subaerial volcanic centres induced metasomatic and metamorphic processes, culminating with the migmatization of basic dykes around two layered pyroxenite–gabbro intrusions called PX1 and PX2 (Stillman 1987).

The PX2 migmatized dykes are characterised by a dense network of closely spaced millimetre-wide leucocratic segregations with perfectly preserved, comb-like, igneous textures. Mineralogically these migmatites consist of

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diopside, andesine plagioclase, Ti-rich biotite, ilmenite, magnetite and minor apatite. They are dominated by plagioclase in the leucosome and diopside in the melanosome. They contain about 25% of leucosomes, considered to represent the original amount of melt present.

Pressure conditions during anatexis are difficult to assess, as no suitable mineral associations are available. An upper bound can be fixed at ca. 4 kb, which corresponds to the present depth of the Moho under the Fuerteventura-Lanzarote ridge (about 11 km; Banda et al. 1981), but as some of the Basal Complex intrusions were emplaced into overturned ocean-floor sediments, the depth of intrusion was probably much lower than this, roughly estimated at 3–6 km, i.e., ca. 1–2 kb, approaching the maximum depth of most known meteoric-hydrothermal systems (e.g., Nabelek 1991).

Temperature estimates for the shallow level partial melting event, around the PX1 intrusion, are in the order of 820–890°C, based on orthopyroxene–clinopyroxene and amphibole–plagioclase thermometry (Muñoz and Sagredo 1994). However, low pressure (2 kb) partial melting experiments on ferrogabbros from the PX1 country rock performed by Koepke et al. (2003), require temperatures up to 1,000°C in order to produce 25% melt (Fig. 3a, Koepke et al. 2003). These experiments also highlight the importance of water in the process of partial melting: at 1,000°C, 2 kb, and the addition of 0.3 wt% water, only a few percent of melt was produced. The 25% melt observed in the PX2 migmatized dykes would require the addition of approximately 2.7 wt% water. Hence, a large amount of fluid is necessary to produce such a high degree of partial melting. The source of the fluids could have been meteoric, oceanic, or magmatic, either in circulation at the time of anatexis or released by hydrous lithologies upon heating from the PX2 intrusion.

Oxygen isotopes are a powerful tool for deciphering fluid histories, i.e., sources and flow paths, in contact metamorphic environments, provided  $\delta^{18}\text{O}$  values of the protoliths and the fluid are sufficiently different (e.g., Valley 1986; Roselle et al. 1999). Ocean island basalts have fairly heterogeneous  $\delta^{18}\text{O}$  values ranging from 4.6 to 7.5‰ (Hoefs 1997), although most are around 5.5‰ (Harmon and Hoefs 1995). In the Canary Islands the relatively large range has been attributed to recycling of ancient hydrothermally altered oceanic crust (causing  $^{18}\text{O}$  depletion in the rock and lowering of the assumed primitive mantle  $\delta^{18}\text{O}$  values of 5.5–6.0‰) and crustal contamination (causing enrichment) (Thirwall et al. 1997). The tracing of magmatic fluid flow by means of oxygen isotopes can be an extremely subtle affair in a system where the intrusion and its country rock are both basic and igneous. However, the influence of sea or meteoric water, with their low or negative  $\delta^{18}\text{O}$  values, respectively, could be more easily tracked.

Oxygen and hydrogen isotope studies on gabbros and dykes, all metamorphosed under static greenschist facies

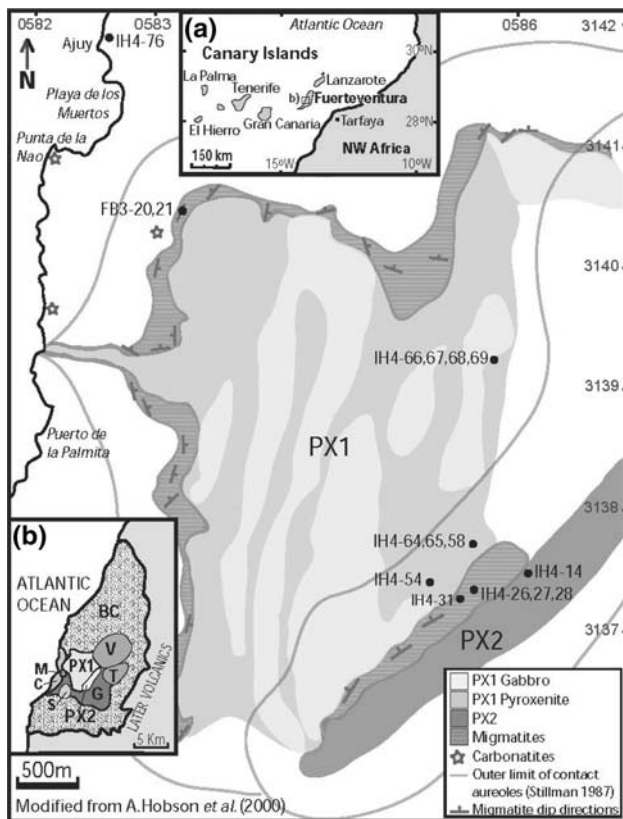
conditions, from the Basal Complex around Pajara, were done by Javoy et al. (1986). They obtained mean whole-rock  $\delta^{18}\text{O}$  values of  $+1.5 \pm 0.3\text{‰}$  for dykes and  $+4.0 \pm 0.6\text{‰}$  for plutonics, arguing that isotopically light meteoric water must have been present at the time of dyke emplacement. This assumption would require high altitude precipitation in order to further fractionate the composition of predicted Tertiary meteoric waters from the palaeolatitudes of the Canary Islands, thus suggesting the former existence of volcanic edifices as high as ca. 2,500–4,000 m in elevation (Javoy et al. 1986).

The aim of this study is to investigate the nature and origin of the fluids that assisted the low-pressure anatexis of basic dykes in the aureole of the PX2 intrusion, using oxygen isotope values measured for whole-rocks and mineral-separates from within and beyond the aureole.

### Geological setting

Fuerteventura is one of the easternmost islands of the Canary archipelago (Fig. 1a), a chain of Early Miocene to recent volcanic islands that lie on oceanic crust ranging in age from ca. 150 to 175 Ma (Schmincke et al. 1982), about 100 km off the north-western coast of Morocco. Two major lithologic units have been distinguished on the island (Fig. 1b): the Basal Complex (BC), and the unconformably overlying Subaerial Volcanic Series of flat-lying flows of Neogene and Quaternary ages.

The Basal Complex (e.g., Fúster et al. 1968; Stillman et al. 1975; Stillman 1987; Le Bas et al. 1986) consists of a thick Mesozoic sedimentary sequence overlain by Tertiary volcanic sedimentary rocks of Oligocene age, intruded by a very dense, early Miocene (e.g., Coello et al. 1992; Cantagrel et al. 1993), NNE–SSW-trending sheeted dyke complex, which formed in association with the emplacement of a series of nested alkaline plutons. The ‘Tierra Mala’ pluton [a series of alkali pyroxenites, hornblendites, gabbros and syenites (*G*, *M* and *S* in Fig. 1b)], followed by the ‘Ajuy-Solapa carbonatite complex’ [a series of apatite pyroxenites, ijolites, syenites, carbonatites and fenites (*C* in Fig. 1b)] (Cantagrel et al. 1993; Le Bas et al. 1986; Balogh et al. 1999), were intruded by the PX1 vertically layered, alkali-gabbro and pyroxenite intrusion (Fig. 1), causing the metamorphism and partial melting of dykes in its aureole. The second layered gabbro–pyroxenite intrusion, PX2, was emplaced along the SE margin of PX1 and caused the partial melting of dykes intrusive to PX1. The youngest intrusion in the Basal Complex, with a mean age of  $20.6 \pm 1.7$  Ma (proposed by Le Bas et al. 1986), is the Vega de Rio de Palmas ring complex [a syenitic ring dyke with a central alkali gabbro stock (*V* and *T* in Fig. 1b)].



**Fig. 1** Geological map of layered gabbro–pyroxenite intrusions PX1 and PX2 and their aureoles (modified from Hobson et al. 2000), showing sample localities and principal migmatite orientations. NNE to NE trending dykes are ubiquitous, but not represented on this map for clarity. *Inset a* Canary Island archipelago showing location of *inset b* on Fuerteventura. *b* Basal Complex of the Betancuria Massif (generalised outlines only) in reverse order of intrusion: *V* Vega de Rio de Palmas, alkali gabbro–syenite ring complex (youngest); *T* Toto, alkali gabbro–syenite ring complex; *M* Tierra Mala ‘Vega-type’ syenite; *PX2* latest gabbro–pyroxenite pluton; *PX1* late gabbro–pyroxenite pluton; *C* carbonatite complex; *G* early gabbro–pyroxenite (=Tierra mala s.l.); *BC* bedded host rock to dyke swarm and plutons (oldest) (Le Bas et al. 1986)

Basic dykes traverse the whole Basal Complex, sometimes occupying more than 70% of the outcrops. Dykes post-PX1 but pre-PX2, with sub-vertical dips towards the NW, i.e., sub-parallel to PX2, are hydrothermally metamorphosed (static greenschist facies), and increasingly hornfelsed and ultimately migmatized, with proximity to the PX2 intrusion.

## Analytical techniques

### Whole-rock major and trace element analyses

Samples were crushed and subsequently reduced to a very fine powder by grinding in a tungsten carbide ring mill. Major and trace elements of whole-rock samples were

analysed by X-ray fluorescence spectrometry and LA-ICP-MS, at the University of Lausanne. Major elements (Si, Ti, Al, Fe, Mn, Mg, Ca, Na, K, P, Cr and Ni) were measured on fused lithium borate glass disks using a Philips PW2400 X-ray fluorescence spectrometer. Trace elements were measured from pressed powder pellets on the same XRF spectrometer (Nb, Zr, Y, Sr, U, Rb, Th, Pb, Ga, Zn, Ni, Cr, V, Ce, Ba and La), and by LA-ICP-MS on glass disks (Be, Sc, Ti, V, Cr, Ni, Cu, Zn, Y, Zr, Nb, Cs, La, Ce, Pr, Nd, Sm, Eu, Gd, Tb, Dy, Ho, Er, Tm, Yb, Lu, Hf, Ta, Pb, Th and U). Tests were made to assess the amount of trace element contamination, such as Ta and Nb, from the tungsten carbide mill. The samples analysed in this study are relatively rich in these elements, therefore contamination is considered negligible.

### Electron microprobe

Microprobe analyses were made at the University of Lausanne with a CAMECA SX50. The accelerating voltage was 15 kV for a beam current of 10–20 nA, depending on the sample analyzed. The beam was rastered over a surface of  $10 \times 8$  microns for feldspars and  $3 \times 2$  microns for hydrated minerals. Natural silicates were used as standards.

### LA-ICP-MS

Laser ablation measurements were made with a 193 nm Lambda Physik<sup>®</sup> Excimer laser (Geolas 200M system) coupled to a Perkin-Elmer 6100 DRC ICP-MS, at the University of Lausanne, Switzerland. Laser settings were 27 kV with a 10 Hz repetition rate, yielding a fluence of about  $12 \text{ J/cm}^2$  on the ablation site. Helium was used as carrier gas (1.1 l/min) and NIST612 glass was used as the external standard, and Ca and Al as internal standards (on the basis of electron microprobe measurements on the ablation pit site). BCR2 basaltic glass was regularly used as a monitor to check for reproducibility and accuracy of the system. Results were always within  $\pm 10\%$  of the values reported by Witt-Eickschen et al. (2003), while Rb, Cs, Y and especially Cr were sometimes off the  $\pm 10\%$  range of the USGS recommended values for BCR2. Wherever possible, several spots were analyzed for each phase to document the trace element variability within individual crystals and between crystals of the same thin section. Data reduction was done using LAMTRACE<sup>®</sup>, a spreadsheet developed by S. E. Jackson (Macquarie University, Sydney).

### Oxygen isotope analyses

Selected whole-rock powders and hand-picked mineral separates of clinopyroxene, plagioclase, biotite, oxides and amphiboles, were analysed. The samples were selected to

cover the range of rock types found within each zone of the aureole, including samples from the hydrothermally metamorphosed zone, the PX2 intrusion itself and also a dyke unaffected by any metamorphism for comparison. The sample localities are shown in Fig. 1.

The oxygen isotope composition ( $^{16}\text{O}$ ,  $^{18}\text{O}$ ) of the samples were measured at the University of Lausanne, using a method similar to that described by Sharp (1990) and Rumble and Hoering (1994) and which is described in more detail in Kasemann et al. (2001). Between 0.5 and 2.5 mg of sample was loaded onto a small Pt-sample holder and pumped out to a vacuum of about  $10^{-6}$  mbar. After prefluorination of the sample chamber overnight (or only 30 min for whole-rock powders or feldspars), the samples were heated with a  $\text{CO}_2$ -laser in 50 mbar of pure  $\text{F}_2$ . Excess  $\text{F}_2$  is separated from the  $\text{O}_2$  produced by conversion to  $\text{Cl}_2$  using  $\text{KCl}$  held at  $150^\circ\text{C}$ . The extracted  $\text{O}_2$  is collected on a molecular sieve (13 $\times$ ) and subsequently expanded into the inlet of a Finnigan MAT 253 isotope ratio mass spectrometer. Oxygen isotope compositions are given in the standard  $\delta$ -notation, expressed relative to VSMOW in permil (‰). Replicate oxygen isotope analyses of the standards used [UWG-2 garnet ( $n = 2$ ); Valley et al. 1995, and LS1 quartz ( $n = 2$ ) an in-house standard normalised to NBS-28] for any one set of samples generally have an average precision of  $\pm 0.2\text{‰}$  for  $\delta^{18}\text{O}$ . The accuracy of  $\delta^{18}\text{O}$  values is commonly better than  $0.2\text{‰}$  compared to accepted  $\delta^{18}\text{O}$  values for UWG-2 of  $5.8\text{‰}$ .

### Field relationships and petrography

PX2 is a vertically layered intrusion, located just west of Pájara on the SE fringe of the larger PX1 intrusion (Fig. 1). It is approximately 600–700 m wide and 3.5 km long (Hobson 2000), elongated NE–SW. It consists of alternating layers of alkali gabbro, wehrlite and pyroxenite, although it is largely dominated by pyroxenite with variable amounts of plagioclase and olivine. Layering is on a centimetric to metric scale, and is sub-parallel to the length of the intrusion, dipping  $65\text{--}85^\circ$  NW–NNW. PX1, the older alkali gabbro–pyroxenite, layered intrusion, forms the country rock to PX2 on its well-exposed, north-western side, but does not show abundant evidence of metamorphism of partial melting in the field.

The intensity of basanitic and clinopyroxene-phyric, ankaramitic dykes, in PX1, increases towards the PX2 contact (from 10 to 80%). It is the nature of these dykes, in the vicinity of PX2, that clearly shows the extreme effects of thermal metamorphism, to the point of anatexis. With increasing proximity to the PX2 intrusion an evolution from unaltered ankaramitic to hydrothermally altered (greenschist facies) dykes is observed outside the aureole,

and from hornfelsed to migmatised dykes within the aureole.

### Migmatised dykes

Dykes found within the first 250 m of the PX2 intrusion (samples IH4–26, 27 and 31; Fig. 1) all resemble ‘migmatites’ (Fig. 2). They are characterised by a dense network of closely spaced, millimetre-wide, leucocratic segregations (Fig. 2a), but differ from migmatites produced in regional metamorphic terrains by the lack of deformation textures in their melanosome and leucosome. Predominant leucosome orientations are sub-parallel to PX2 and their respective dyke margins, although conjugate leucosomes are also observed (Fig. 2a).

Leucosomes, thought to be representative of a ‘melt’, are composed of 70% andesine plagioclase ( $\text{An}_{32\text{--}36}$ ), 24% diopside, 4% oxides and 2% biotite. No post-solidus deformation appears to have taken place, as plagioclase and diopside form comb-like, igneous textures, elongated perpendicular to leucosome walls (Fig. 2c). The finer grained melanosome, is a fairly homogeneous lithology with a granoblastic, polygonal texture, and well-developed triple junctions (Fig. 3c and inset). Mineralogically it is similar to the leucosome, but modes are different due to the lower abundance of plagioclase. We interpret this melanosome as a completely recrystallised, more or less restitic mineral assemblage from which melt segregated.

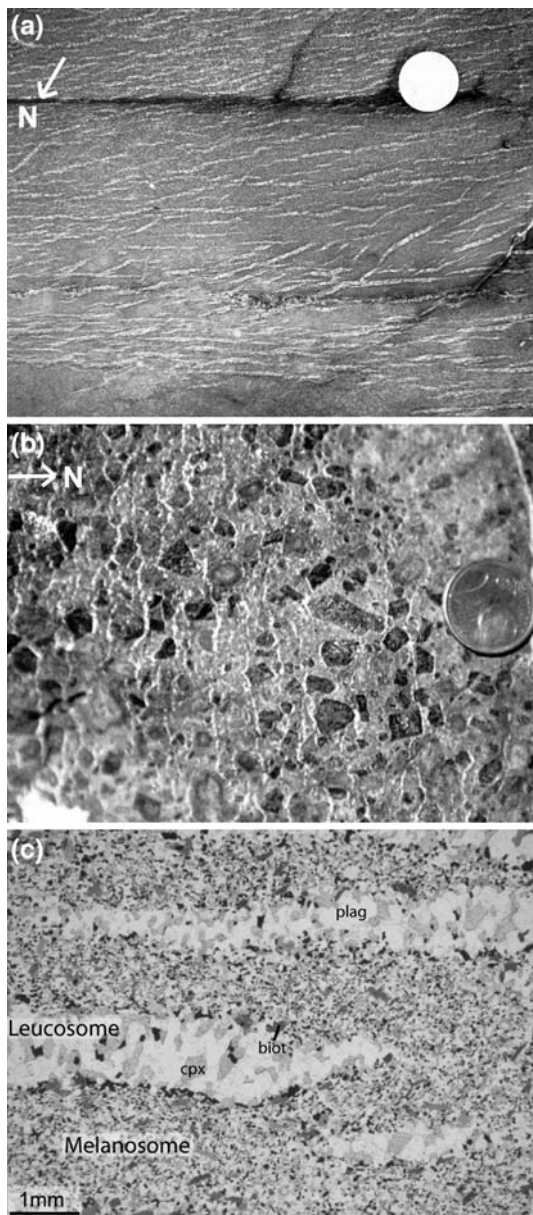
Dyke centres preserve large (2–10 mm), relict clinopyroxene phenocrysts, around which leucosomes are distributed (Fig. 2b). They appear to be the only igneous remnants that escaped complete recrystallisation. They do however contain abundant ilmenite exsolution lamellae (Fig. 3f), a feature possibly related to thermal metamorphic processes. Neocrystallised diopside appears to have grown epitaxially along the rim of such relict phenocrysts (Fig. 3f).

No significant differences are noted between leucosome and melanosome major element mineral compositions, however, relict clinopyroxene phenocrysts have slightly higher Ti and Al contents than the small neocrystallised diopside from the matrix (eTable 1a; Holloway and Bussy 2007).

### Hornfelsed dykes

Beyond 250 m from the PX2 intrusion no more evidence for partial melting is found. Matrix textures largely resemble those found in the migmatite melanosomes, but with a smaller grain-size (Fig. 3b). They consist of small, recrystallised diopside, oligoclase-andesine plagioclase ( $\text{An}_{23\text{--}46}$ ; a slightly larger compositional range than plagioclase in the migmatised dykes, probably due to a lower





**Fig. 2** Field photographs of **a** PX2 aureole migmatite (coin for scale, diameter = 21 mm), and **b** central phytic part of migmatized dyke from PX1 aureole (coin for scale, diameter = 19 mm). **c** Photomicrograph of a migmatite showing well-developed leucosomes and melanosomes. Minerals include plagioclase (*plag*), diopside (*cpx*), biotite (*biot*) and oxides (*ilmenite* and *magnetite*). Non-deformed, igneous textures—with crystals aligned perpendicular to vein walls—are preserved in these leucosomes

degree of equilibration), fairly abundant biotite (that forms pseudomorphs or haloes around relict phenocrysts), ilmenite, magnetite, kaersutite, and larger, relict minerals. Few elongate, labradorite plagioclases, resembling igneous remnants are preserved. In these hornfelsed dykes, the large, relict clinopyroxene phenocrysts, only contain ilmenite exsolution lamellae on outer parts of crystals (Fig. 3e, eTable 1).

Such characteristics are found up to approximately 700 m from PX2, defining the *hornfels zone* (samples IH4–54, 64, 65; Fig. 1) and delimiting the metamorphic contact aureole.

#### Hydrothermally metamorphosed (Greenschist facies) dykes

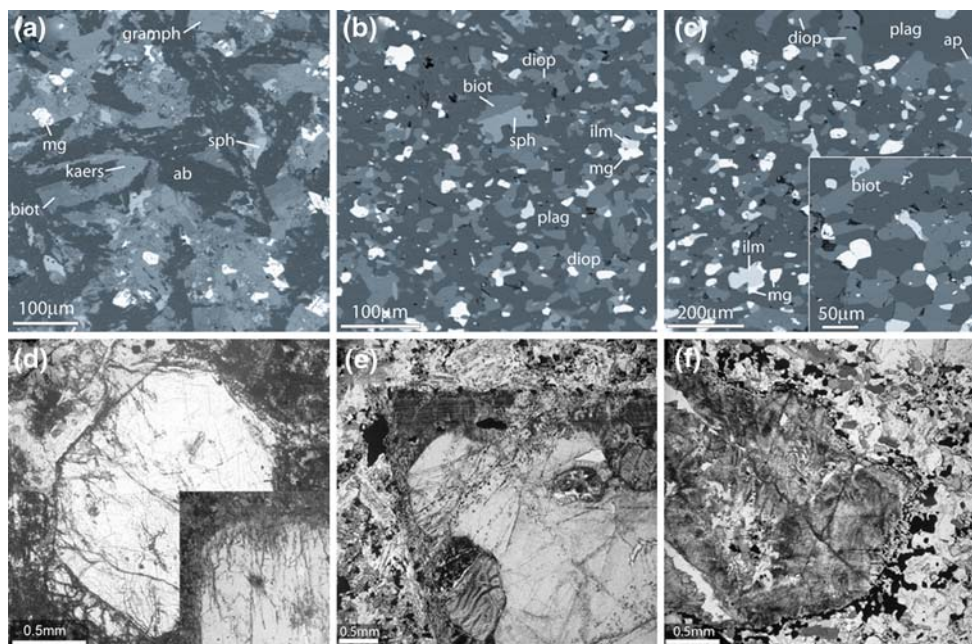
Outside the contact aureole, beyond 700 m from PX2 but still within the PX1 intrusion, the intensity of dykes diminishes rapidly, and they appear increasingly hydrothermally altered to the greenschist facies (samples IH4–66, 67, 68; Fig. 1). Texturally there is no longer any sign of contact metamorphic recrystallisation, as igneous intersertal/intergranular, seriate textures are preserved (Fig. 3a). Nevertheless, these dykes appear to have undergone a pervasive, hydrothermal event, as they are dominated by secondary hydrous minerals, e.g., edenite. They also contain albitic plagioclase ( $An_{0-48}$ ), often sericitised or altered to zeolites, kaersutite, magnetite, ilmenite, sphene and rare scapolite and calcite. The large, relict igneous phenocrysts of clinopyroxene, that are unaltered by any metamorphism (Fig. 3d), do not contain any ilmenite exsolution lamellae, but sometimes have Ti-rich patches or rims. However, some are partially or entirely replaced by edenite (Fig. 3d, inset). Numerous phenocryst pseudomorphs (after clinopyroxene or olivine) contain a mixture of fibrous, hydrous, green to white minerals such as edenite, tremolite and actinolite.

#### Ankaramitic dykes

An ankaramitic dyke from outside the PX2 and PX1 contact aureoles, that does not appear to have been influenced by the hydrous metamorphic event (sample IH4–76; Fig. 1, top left near Ajuy), was sampled as representative of an unaltered igneous rock. It is dominated by phenocrysts of titanite (eTable 1a), with few olivines. The matrix consists of microcrysts of titanite, magnetite and olivine surrounded by small but often poikilitic plagioclase enveloping even smaller crystals of the aforementioned minerals, including fine apatite needles.

#### Geochemistry

The PX2 pyroxenite and all the lithologies in and around its aureole, are mafic to ultramafic, and diopside, olivine, and nepheline normative. They are all metaluminous, alkaline, and relatively  $TiO_2$ ,  $FeO_{total}$  and  $MgO$  rich (3–5 wt%, 13–20 wt% and 5–10 wt%, respectively). All dykes are basaltic to basaltic in composition, and generally have comparable whole-rock, major-element compositions.



**Fig. 3** Back-scattered electron (BSE) images of dyke matrices **a** greenschist facies **b** hornfels **c** migmatite, showing increasing recrystallisation towards the intrusion (migmatite zone). Note: preservation of igneous intergranular-interstitial textures in the greenschist facies dyke, a noticeably smaller grain-size in the hornfels with similar equigranular textures but a larger grain-size than in the

migmatite zone. Photomicrographs of clinopyroxene phenocrysts from **d** greenschist facies **e** hornfels and **f** migmatite dyke centres, showing increasing ilmenite exsolution towards the intrusion (migmatite zone). *ab* albite, *sph* sphene, *biot* biotite, *kaers* kaersutite, *mg* magnetite, *gr amph* green amphibole, *diop* diopside, *ilm* ilmenite, *plag* plagioclase, *ap* apatite

Only the greenschist dykes outside the aureole differ in their higher volatile contents (up to 3 wt%). In terms of trace elements all the dykes retain their OIB signatures with a strong enrichment in the most incompatible elements of up to 200 times the primitive mantle. Smaller scale variations of dyke compositions within and outside the aureole are not pronounced and do not follow any trend towards/away from the intrusion (Holloway and Bussy 2007).

### Oxygen isotope compositions of rocks and minerals

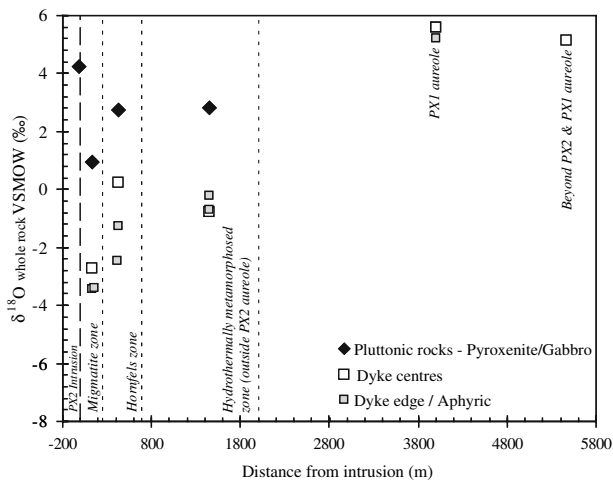
The similarity of the whole-rock geochemistry of dykes from the migmatite, hornfels, and greenschist zone, allow oxygen isotope compositions to be compared directly without having to correct for fractionations caused by original magmatic differences, or transport of material into or out of the dykes during metamorphism.

Low pressure anatexis of basic dykes in the contact metamorphic aureole of the PX2 pyroxenitic intrusion, was attained by the successive intrusion of numerous magma batches, producing a continuously high heat flow, along with the inferred presence of fluids. In order to gain insight into the nature and origin of these fluids, oxygen isotope compositions of whole-rock powders and mineral

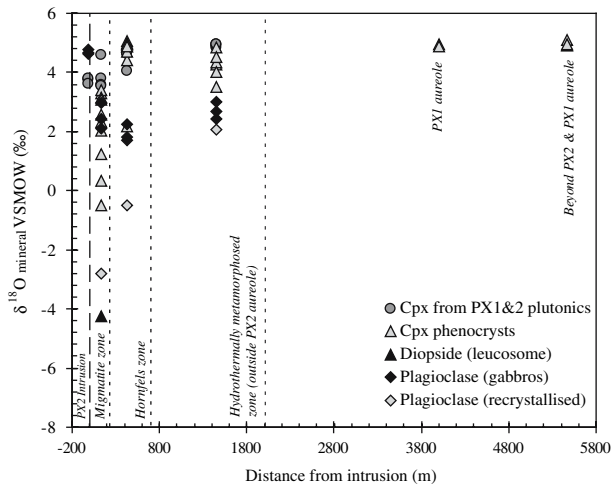
separates of clinopyroxene, feldspar, biotite, amphibole, and oxides from in and around the PX2 aureole were measured. Results (given in eTable 2) provide unequivocal evidence for interaction of migmatite, hornfelsed, and greenschist facies dykes just outside the aureole, with meteoric fluids.

### Whole-rocks

All dykes in and around the PX2 aureole, in the migmatite, hornfels and greenschist zone, have low or even negative whole-rock  $\delta^{18}\text{O}$  values (+0.2 to  $-3.4\text{‰}$ ), decreasing towards the intrusion (Fig. 4), with lowest values in the anatectic zone. Furthermore, the aphyric dykes or dyke edges have lower values than the phyric central parts of dykes (Fig. 4). The PX1 gabbro, forming the country rock to the PX2 intrusion, also has fairly low  $\delta^{18}\text{O}$  values (+1.0 to  $+2.8\text{‰}$ ) decreasing within the aureole towards the intrusion, although this trend is much less pronounced than that shown by the dykes. The PX2 intrusion itself has a  $\delta^{18}\text{O}$  value of  $+4.2\text{‰}$ , approaching a value more typical for basic, mantle-derived melts. Migmatite dykes from the PX1 aureole and an ankaramitic dyke from beyond both aureoles, unaffected by any meteoric-hydrothermal events, have the highest  $\delta^{18}\text{O}$  values ( $+5.1$  to  $+5.6\text{‰}$ ), likely representing original magmatic values.



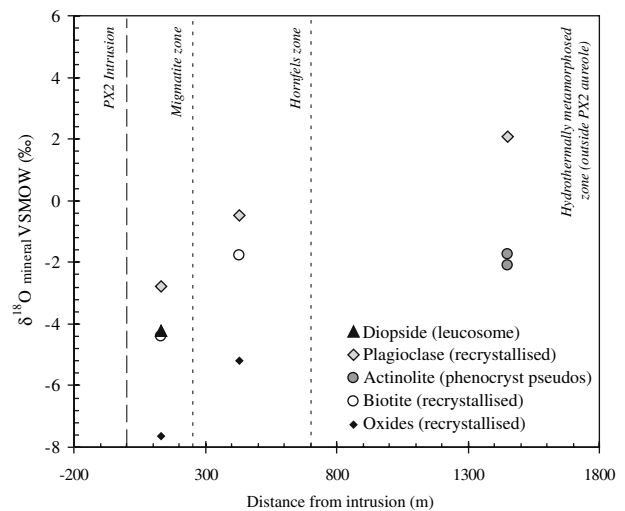
**Fig. 4** Whole-rock  $\delta^{18}\text{O}$  values for representative samples from the PX2 intrusion and its contact aureole, plotted as a function of distance from the intrusion



**Fig. 5**  $\delta^{18}\text{O}$  values of clinopyroxene and plagioclase; (a) from plutonic rocks (PX1 host gabbro and PX2 pyroxenite), (b) neo-crystallised in dykes; and (c) relict clinopyroxene phenocrysts. Plotted as a function of distance from the PX2 intrusion

**Mineral separates**

PX2 pyroxenite mineral separates only deviate slightly from the whole-rock composition, clinopyroxene with lower (about +3.7‰) and plagioclase with higher (about +4.7‰) values (Fig. 5). The large igneous clinopyroxenes from the PX1 gabbros and relict phenocrysts in the dykes from within the aureole, that escaped recrystallisation during contact metamorphism, have higher  $\delta^{18}\text{O}$  values than their respective whole-rock compositions (compare Figs. 4, 5). However, there is a large heterogeneity between and within single relict clinopyroxene phenocrysts in the aureole (see eTable 2), with the lowest  $\delta^{18}\text{O}$  values



**Fig. 6**  $\delta^{18}\text{O}$  values of neo- or re-crystallised minerals; diopside, plagioclase, actinolite, biotite and oxides, found in migmatitised, hornfelsed and greenschist dykes, plotted as a function of distance from the PX2 intrusion

measured from the migmatite zone (−0.5‰) and highest from the hornfels zone (+5.1‰).

Plagioclase from the PX1 gabbro has lower  $\delta^{18}\text{O}$  values (+1.7 to +3‰) than the PX2-pyroxenite plagioclase (about +4.7‰), but not as low as the recrystallised plagioclase from the dykes of the aureole (−2.8 to −0.5‰; Fig. 5), that decreases with proximity to the intrusion.

All neo- and recrystallised minerals from dykes within and just outside the aureole, in the hydrothermally metamorphosed zone, have low  $\delta^{18}\text{O}$  values, with a pronounced decrease towards the intrusion (Fig. 6). Plagioclase, diopside, biotite and oxides (magnetite and ilmenite) from the migmatites, yield consistent  $\delta^{18}\text{O}$  values of −2.8, −4.2, −4.4 and −7.6‰, respectively, in terms of their relative order of decreasing content of  $^{18}\text{O}$ . Similar values were obtained for minerals separated from the leucosomes and melanosomes. However, in the hornfelsed dykes, plagioclase, biotite and oxides yield consistently higher values of −0.5, −1.8 and −5.2‰, respectively (diopside was unfortunately too fine-grained for separation).

In summary,  $\delta^{18}\text{O}$  values of dykes, in and around the PX2 aureole, that pre-date PX2 but cut PX1, are lower than those of normal, mantle-derived mafic rocks, and decrease towards the PX2 intrusion. Negative  $\delta^{18}\text{O}$  values can only be obtained by interaction with meteoric fluids and/or melting of material that has previously been depleted in  $^{18}\text{O}$  through exchange with meteoric water. Higher or even normal mantle-derived magmatic values for some relict clinopyroxene phenocrysts supports an exchange with low  $\delta^{18}\text{O}$  meteoric water during emplacement though, with minerals that are recrystallising being open to this exchange. Therefore, any decrease in  $\delta^{18}\text{O}$  observed in relict

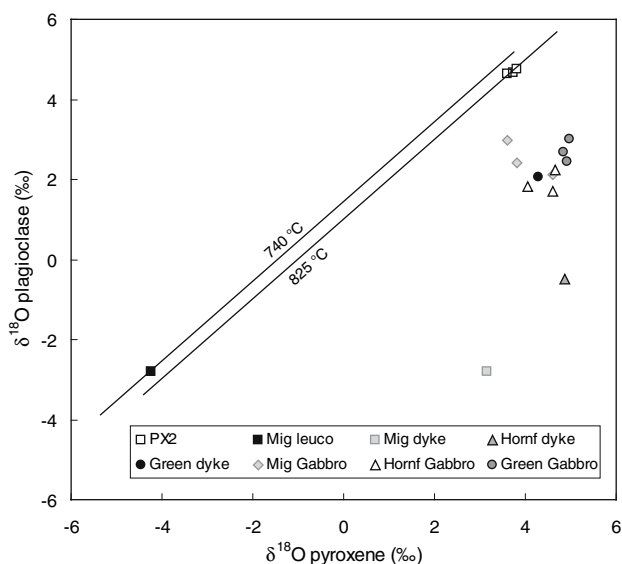


phenocrysts may be related to diffusive exchange or exchange along small zones of alteration/recrystallisation within the grains but not visible in thin section. However, smaller, recrystallised minerals from dykes within the aureole and outside the aureole in the hydrothermal zone, all have negative  $\delta^{18}\text{O}$  values, appearing to have recrystallised in the presence of and in equilibrium with a relatively  $^{18}\text{O}$ -poor meteoric fluid phase.

Diffusion distances were calculated for clinopyroxene (diopside), using the following equations:  $X = \sqrt{Dt}$  and  $D = D_0 e^{(-E_a/RT)}$  (after Valley 2001), where the pre-exponential diffusion coefficient  $D_0 = 1.5 \times 10^{-6} \text{ cm}^2/\text{s}$ , the activation energy  $E_a = 226 \text{ kJ/mol}$  (Farver 1989), and the gas constant,  $R = 8.3143 \text{ J}^\circ\text{Kmol}$ . An estimation of the maximum time ( $t$ ), during which the temperatures remained just below melting temperatures ( $T = 1,000^\circ\text{C}$  or  $1,323 \text{ K}$ , Holloway and Bussy 2007; Koepke et al. 2003) but above closure temperatures for diffusion, was made in order to calculate the maximum diffusion distance ( $X$ ) possible for the clinopyroxene phenocrysts. In 1,000,000 years (a cautious overestimate) the diffusion distance would not have exceeded  $2 \times 10^{-3} \mu\text{m}$ . Oxygen isotope exchange by diffusion does not, therefore, appear to be a plausible mechanism for the depletion of  $^{18}\text{O}$  in the 2–10 mm clinopyroxene phenocrysts of the PX2 aureole.

$\delta$ - $\delta$  diagrams (Fig. 7) may illustrate whether rocks have undergone hydrothermal alteration events after first formation of the minerals or whether they have reached and preserved isotopic equilibrium at the time of crystallisation. Plagioclase, being more susceptible to low temperature sub-solidus alteration (e.g., Criss and Taylor 1986; Criss et al. 1987), may exchange with water down to lower closure temperatures compared to pyroxene, and hence have lower  $\delta^{18}\text{O}$  values if exchanging with meteoric water (e.g., Taylor and Forester 1979). Hence, high  $\Delta^{18}\text{O}$  px – plag values, such as all points lying below the equilibrium isotopic fractionation lines on Fig. 7, represent mineral-pairs out of isotopic equilibrium, where the plagioclase was open to exchange with meteoric fluids. The disequilibrium between mineral pairs in the PX1 gabbros (*Mig*, *Hornf* and *Green Gabbros* in Fig. 7) and the greenschist facies dyke (*Green dyke*) thus provide evidence of sub-solidus hydrothermal alteration. However, the accentuated fractionations observed between the migmatite and hornfels relict, clinopyroxene phenocrysts and their matrix plagioclase (*Mig* and *Hornf dyke*, Fig. 7) is probably related to increased exchange between the plagioclase and the fluid during and after recrystallisation.

Pyroxene and plagioclase from the PX2 pyroxenite and the recrystallised migmatite matrix (leucosome and melanosome) (*PX2* and *Mig leuco* in Fig. 7) appear to have crystallised under equilibrium magmatic conditions as they have small fractionations that also correspond to approximate



**Fig. 7** A plot of  $\delta^{18}\text{O}$  plagioclase versus  $\delta^{18}\text{O}$  pyroxene for dykes and plutonics in the PX2 aureole, including the PX2 intrusion itself. Equilibrium isotopic fractionation lines are shown: (a)  $740^\circ\text{C}$ ; a fractionation of +1.44, for a plagioclase composition of  $\text{An}_{32}$ , and (b)  $825^\circ\text{C}$ ; a fractionation +0.93–0.97, for an average plagioclase compositions of  $\text{An}_{63}$ . *Mig leuco* minerals from migmatite leucosome (IH4–27); *Mig dyke* relict cpx phenocrysts from migmatite and matrix (leucosome/melanosome) plagioclase (IH4–26), *Hornf dyke* relict cpx phenocrysts from hornfelsed dyke and matrix plagioclase (IH4–65), *Green dyke* relict cpx phenocrysts from greenschist facies dyke and matrix plagioclase (IH4–68), *Mig Gabbro* gabbro from migmatized zone (IH4–28), *Hornf Gabbro* gabbro from hornfelsed zone (IH4–58), *Green Gabbro* gabbro from greenschist facies zone (IH4–69)

closure temperatures for pyroxene in intrusive gabbroic rocks. Nevertheless, the migmatite is obviously in equilibrium with a low  $\delta^{18}\text{O}$  meteoric fluid, whereas PX2 appears to have only slightly lower than normal magmatic values. Temperatures implied by such fractionations (calculated using the equation:  $\delta^{18}\text{O}_{\text{plag}} - \delta^{18}\text{O}_{\text{cpx}} = A \times 10^6/T^2$ , where  $T$  is temperature, and  $A$  coefficients for plagioclase–clinopyroxene fractionation factors (migmatite;  $\text{An}_{32} - \text{Di} = 1.47$ ; PX2 pyroxenite;  $\text{An}_{63} - \text{Di} = 1.15$ ) taken from Table 5 in Chacko et al. 2001) are in the order of  $740$  and  $825^\circ\text{C}$ , for the migmatite and PX2 pyroxenite, respectively. These temperatures may well represent cooling temperatures and/or closure temperatures in a more or less binary pyroxene–plagioclase system, where the plagioclase closes to isotopic exchange once the pyroxene has closed. The different temperatures recorded may be due to the smaller grain-size of the migmatites compared to the pyroxenite.

## Discussion

Estimated temperatures necessary for the partial melting of the basic igneous dykes in the PX2 contact metamorphic



aureole at pressures of 1–2 kb are in the order of 1,000°C, with the assumption that fluids were present to decrease their solidus temperatures. Experiments by Koepke et al. (2003, Fig. 3b), on *Tierra Mala* ferrogabbros (from the PX1 country-rock), show that the proportion of melt produced is not only sensitive to temperature and pressure, but is also highly dependant on the bulk water content: at 2 kb and 1,000°C, 1 wt% water content produced 10% melt, whereas 3 wt% water content produced 35% melt. The unaltered ankaramitic dyke (IH4–76, Holloway and Bussy 2007) far away from the PX2 aureole, contains a mere 0.7 wt% H<sub>2</sub>O, not nearly enough to produce the 25% melt thought to be represented by leucosomes in the migmatized dykes close to the intrusion. However, the hydrothermally metamorphosed basanitic dykes found just beyond the aureole, the hypothesised migmatite protoliths, contain up to 3 wt% H<sub>2</sub>O, sufficient to account for the melt proportions at attainable temperatures.

Circulation of high altitude meteoric water through a volcanic edifice at temperatures of 300–500°C, i.e., static greenschist facies conditions, may well have caused the hydrous recrystallisation of all dykes in the vicinity of the edifice, thereby decreasing their  $\delta^{18}\text{O}$  values. Fluid convection driven by an intrusion would have enhanced the circulation of meteoric fluids and also oxygen exchange (Valley 1986). Low  $\delta^{18}\text{O}$  values of all rocks showing signs of subsolidus hydrothermal alteration, such as the greenschist facies dykes found just outside the PX2 thermal aureole, can therefore be attributed to meteoric-hydrothermal fluid circulation. Many studies on oxygen isotope variations in ophiolites (e.g., Gregory and Taylor 1981; Muehlenbachs 1986; Eiler 2001; Stakes and Taylor 2003), focussing on low and high temperature hydrothermal alteration processes, have described changes in whole-rock  $\delta^{18}\text{O}$  values caused by seawater interaction with oceanic crust. However, the changes noted do not result in depletions in  $^{18}\text{O}$  in the order of those observed in the present system.

The systematic  $\delta^{18}\text{O}$  decrease in dykes of the PX2 contact metamorphic aureole, towards higher metamorphic grades, is a trend observed in many other meteoric-hydrothermal systems around plutonic complexes. Such systems have been well reviewed by Criss and Taylor (1986, Table 1). One example of a gabbroic intrusion with  $^{18}\text{O}$  depleted volcanic wall rocks is the Tertiary Cuillin and Red Hills intrusive complexes, Isle of Skye, Scotland:  $\delta^{18}\text{O}$  values of the contact metamorphosed basaltic country rocks systematically decrease towards the plutonic complex, down to  $-7.1\text{‰}$  in the highest grade olivine zone (Forester and Taylor 1977). Explanations proposed for this radial inward depletion in  $^{18}\text{O}$ , i.e., the ‘bullseye’ pattern (Criss and Taylor 1986), essentially involve increasing temperatures toward the intrusive centres: Higher temper-

atures promote more rapid and extensive isotopic exchange causing the rocks to have  $\delta^{18}\text{O}$  values that approach those of the meteoric H<sub>2</sub>O (Forester and Taylor 1977). In addition, the fractionation factors between minerals and fluid decrease with increasing temperature. Ferry et al. (1987) also suggested that most of the hydrothermal alteration in the high-grade zone occurred at temperatures  $>1,000^\circ\text{C}$  while heated by the intrusions.

Large xenoliths of metabasalt found in the layered gabbros of the Kap Edvard Holm Intrusive Complex (East Greenland), closely resemble the hornfelsed dykes in the PX2 aureole, in terms of mineralogy, their fine-grained equigranular granoblastic textures, and their substantial depletion in  $^{18}\text{O}$  compared to fresh basalts (Brandriss et al. 1996). Oxygen isotope fractionation between coexisting plagioclase and pyroxene in the xenoliths reflects equilibration with low  $\delta^{18}\text{O}$  fluids at temperatures of approximately 1,000°C (Brandriss et al. 1992). Interpretations involve subsolidus meteoric-hydrothermal alteration of basaltic country rocks prior to stoping (Brandriss et al. 1995), followed by reheating to magmatic temperatures on assimilation, causing dehydration, partial melting and hornfelsic recrystallisation at temperatures  $>1,050^\circ\text{C}$  (Brandriss et al. 1996). This would have restored a pyroxene–plagioclase equilibrium with low  $\delta^{18}\text{O}$  values. A similar scenario could be envisaged for the dykes in the PX2 aureole (without stoping), whereby low temperature meteoric-hydrothermal alteration occurred prior to or during the initial stages of PX2 emplacement. Intrusion of the PX2 magmas heated the wall rocks sufficiently to cause partial to complete recrystallisation further away and partial melting closer to the intrusive centre. Changes in  $\delta^{18}\text{O}$  values are then closely coupled to the extent of recrystallisation and extent of fluid–mineral exchange during the recrystallisation.

Decreasing  $\delta^{18}\text{O}$  values towards the intrusion could also be partially attributed to the devolatilisation of the hydrothermally altered rocks with increasing temperature during contact metamorphism (Valley 1986): the fractionation factor of  $^{18}\text{O}/^{16}\text{O}$  between H<sub>2</sub>O and melt or minerals is larger than unity at high temperatures (above 600°C, e.g., Chacko et al. 2001) i.e.,  $^{18}\text{O}$  preferentially partitions into the aqueous fluid and the  $\delta^{18}\text{O}$  of melt and minerals after dehydration will be lower. Increasing temperatures with proximity to the intrusion, causing increasing dehydration of hydrothermally altered dykes, may thus also cause a decrease in  $\delta^{18}\text{O}$  values. However, the magnitude of this isotopic effect is controlled by the amount of water expelled (limited to 2–3 wt% for the greenschist dykes), and can not exceed 1‰ (e.g., Baumgartner and Valley 2001), i.e., a process much less effective in lowering the  $\delta^{18}\text{O}$  values than infiltration of low  $\delta^{18}\text{O}$  fluids.

Increasing fluid/rock ratios towards the PX2 intrusion could also account for the observed trends in  $\delta^{18}\text{O}$  values. Increasing dyke intensity towards PX2, probably reflects a higher fracture intensity towards the intrusion, along which fluid-flow may have focussed, preferentially altering the finer-grained dykes. Such channelised fluid flow would have created higher bulk system fluid/rock ratios towards the PX2 intrusion as dyke intensity is greater in proximity to the intrusion.  $\delta^{18}\text{O}$  whole-rock values would thus decrease heterogeneously towards the intrusion, with larger depletions in the dykes than their gabbroic hosts, as is observed in this study. Increasing fluid/rock ratios in addition to an increase in temperature can also propagate a radially inward fluid flow (e.g., Taylor 1971) of a more pervasive nature, thereby possibly increasing water/rock ratios towards the margin of the intrusion (in a relatively closed system). However, in order to have decreasing  $\delta^{18}\text{O}$  values with approach to the intrusive contact, it is imperative that the fluid has not been extensively buffered by the wall rocks with approach to the contact as this would only raise the  $^{18}\text{O}$  content of the fluid during increasing exchange. Hence, a fluid source close to the intrusive system and/or an increasing channel density for fluid transport are required (e.g., Conrad et al. 1997). Fluid flow paths would have been largely deflected around the impermeable PX2 magma sheets while hot, resulting in the preservation of relatively high  $\delta^{18}\text{O}$  values compared to wall rocks. Slight depletions compared to normal mantle values may be due to partial exchange with meteoric fluids upon cooling of the intrusion. Nevertheless, a lack of evidence for late, sub-solidus hydrothermal alteration in the high-grade rocks near the PX2 intrusive contact, and in the PX2 intrusion itself, preclude the possibility of a post-PX2 hydrothermal event.

## Conclusions

The low  $\delta^{18}\text{O}$  values of basic dykes in and around the PX2 contact metamorphic aureole imply that the fluids necessary for their low-pressure anatexis were of meteoric origin. Exchange of the rocks with meteoric-hydrothermal fluids may have started during initial stages or possibly even before the intrusion of PX2. The intrusion of PX2 may have created fluid convection cells, thus enhancing fluid infiltration and exchange with the  $^{18}\text{O}$  depleted fluids during recrystallisation of the country rock.

Decreasing  $\delta^{18}\text{O}$  values towards the PX2 intrusion could be attributed to increasing fluid/rock ratios and/or temperature. Increasing dyke intensity towards PX2 supports emplacement of PX2 into a structurally weakened and fractured zone that also offers an increase in possible fluid pathways and hence in overall fluid/rock ratio with approach to the intrusion. In parallel, increasing tempera-

ture towards PX2 facilitates exchange between advecting meteoric fluids and recrystallising minerals, thus similarly leading to a decrease in  $\delta^{18}\text{O}$  values towards the intrusion.

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