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Shallow-Level Migmatization of Gabbros in a Metamorphic Contact Aureole, Fuerteventura Basal Complex, Canary Islands

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Migmatization of gabbroic rocks at 2–3 kbar has occurred in the metamorphic contact aureole of a mafic pluton in the Fuerteventura Basal Complex (Canary Islands). Migmatites are characterized by a dense network of closely spaced millimetre-wide leucocratic veins with perfectly preserved igneous textures. They are all relatively enriched in Al, Na, P, Sr, Ba, Nb, Y and the rare earth elements compared with the unaffected country rock beyond the aureole. Migmatization under such low-pressure conditions was possible because of the unusual tectonic and magmatic context in which it occurred. Multiple basic intrusions associated with extrusive volcanic activity created high heat flow in a small area. Alkaline and metasomatized rocks present in the country rock of the intruding pluton were leached by high-temperature fluids during contact metamorphism. These enriched fluids then favoured partial melting of the host gabbroic rocks, and contaminated both the leucosomes and melanosomes. A transpressive tectonic setting at the time of intrusion created shearing along the contact between the intrusion and its host rock. This shearing enhanced circulation of the fluids and allowed segregation of the neo-formed melts from their restite by opening tension veins into which the melts migrated. Depending on the relative timing of melt segregation and recrystallization, leucosomes range in composition from a 40–60% mixture of clinopyroxene (± amphibole) and plagioclase to almost pure feldspathic veins. Comparable occurrences of gabbros migmatized at low pressure are expected only at a small scale in localized areas of high heat flow in the presence of fluids, such as in mid-ocean ridges or ocean-islands.

KEY WORDS: *crustal anatexis; gabbros; migmatites; partial melting*

INTRODUCTION

Three main parameters are intimately involved in the genesis of migmatites: temperature, a_{H_2O} and deformation [for reviews see Brown *et al.* (1995) and papers therein, and Wolde *et al.* (1996)]. Favourable conditions are commonly obtained in the lower crust under amphibolite or granulite facies conditions. Temperature increase is often the result of crustal thickening or basaltic underplating. Dehydration melting at low a_{H_2O} leads to fusion and the increase in volume allows segregation of the leucocratic melts. Ultimately, deformation is the motor of melt extraction.

Anatexis of basic material at low pressure is poorly documented. Flagler & Spray (1991) mentioned anatectic plagiogranites in amphibolitized gabbros from oceanic fragments within high-temperature shear zones, near a spreading centre in the Shetland Islands. Bédard (1991) and Mevel (1988) also reported occurrences of anatexis in oceanic crust or related units.

The *in situ* anatexis of gabbroic rocks in an ocean-island setting has, to our knowledge, never been described. In Fuerteventura, spectacular migmatites are exposed within a gabbroic complex emplaced in the shallow oceanic crust and its sediments. Migmatites crop out in a 200-m-wide aureole at the contact between two gabbroic intrusions. Deformation ceased soon after melt segregation and 'magmatic' textures are therefore preserved in the aureole and, in particular, in the tension veins. The purpose of this paper is to examine in detail the Fuerteventura migmatites in their geological context and

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to make a first attempt to define the pressure and temperature conditions necessary for their formation. We propose that these conditions result from a specific combination of magmatic events and tectonic setting (Hobson *et al.*, 1997). This could provide a model for the understanding of anatexis and migmatization of gabbroic rocks elsewhere.

GEOLOGICAL SETTING

Fuerteventura is the second largest island of the Canarian volcanic archipelago and the closest to the African coast (Fig. 1). Two main geological units have been described (Fúster *et al.*, 1968; Stillman *et al.*, 1975; Féraud *et al.*, 1985; Le Bas *et al.*, 1986; Stillman, 1987; Coello *et al.*, 1992; Ancochea *et al.*, 1996): (1) Tertiary lavas and pyroclastic flows and Plio-Quaternary lavas and aeolian sands which cover most of the island, and (2) a Basal Complex (Stillman *et al.*, 1975) comprising an uplifted series of sedimentary and submarine volcanic and volcaniclastic rocks ranging from Lower Jurassic (Steiner *et al.*, 1998) to Tertiary, five Tertiary alkaline plutons that intruded the sedimentary rocks, and a sheeted dyke complex that intruded both the plutonic and sedimentary formations. The main outcrop of the Basal Complex, the Betancuria Massif (Fúster *et al.*, 1968; Gastesi Bascuñana, 1969), is situated in the central western part of the island (Fig. 1).

The plutonic history of Fuerteventura appears short [Cantagrel *et al.* (1993) and references therein], as only ~5 my separate the second from the last of the five nested intrusions defined in the Betancuria Massif by Le Bas *et al.* (1986). There is no geological reason to believe that the first intrusion should be much older than the others. These intrusions are (in chronological order): (1) the Tierra Mala pluton (*sensu stricto*), (2) the Ajuy–Solapa carbonatites and ijolites (25–19 Ma) intrusive in Tierra Mala, (3) the PX1 and (4) PX2 plutons, and (5) the Vega de Rio de Palma (21–18 Ma) and Toto ring dykes.

The sheeted dyke complex was not emplaced as a single event but can be related to each of the intrusions and is probably associated with volcanic activity (Javoy *et al.*, 1986; Le Bas *et al.*, 1986; Ancochea *et al.*, 1996). The plutons could therefore represent cooled magma reservoirs that fed the Miocene volcanoes of Fuerteventura (Stillman, 1987; Ancochea *et al.*, 1996). Emplacement of the plutons at high levels in the crust is evidenced by the fact that they intruded the overturned ocean-floor sediments and by the isotopic composition of the meteoric fluids that metamorphosed them (Javoy *et al.*, 1986; Le Bas *et al.*, 1986). This hydrothermal metamorphism, which is thought to be due to the circulation of meteoric fluids along the dykes (Javoy *et al.*, 1986), reached the greenschist facies and affected all the Basal Complex. Intensity of metamorphism is linked to the density of the dykes.

A regional, i.e. western Central Atlantic and Northwest African, NNE–SSW transtensive stress regime was active at the time of emplacement of most of the plutons. This regime would correspond to the extensional stress field suggested by Stillman *et al.* (1975), Féraud *et al.* (1985) and Stillman (1987), to the NE–SW 'African' fractures of Araña & Ortiz (1991), and to the shear zones of Fernandez *et al.* (1997). These NNE–SSW transtensive movements could have been a response to the N–S Alpine compression illustrated by the folding of the Rif and High Atlas chains in north-western Africa (e.g. Féraud *et al.*, 1985; Araña & Ortiz, 1991). This extensional regime is responsible for the orientation of the Fuerteventura–Lanzarote volcanic ridge, the more or less oblong shape of the intrusions (Fig. 1), the orientation of the dykes, the internal geometry of some of the intrusions and the lack of deformation in the country rocks of the plutons (Fúster *et al.*, 1968; Stillman *et al.*, 1975; Féraud *et al.*, 1985; Le Bas *et al.*, 1986). The end of the extensional stress regime is marked by the intrusion of the Vega de Rio de Palma and Toto ring dykes (Stillman *et al.*, 1975; Féraud *et al.*, 1985; Le Bas *et al.*, 1986; Stillman, 1987) and the beginning of a centred and more differentiated volcanism. PX1 is the only intrusion which seems to have been emplaced in a compressional regime, deforming and migmatizing its host rock. Emplacement must have taken place at a time when magmatic activity and crustal dilation did not coincide (Féraud *et al.*, 1985; Stillman, 1987). Like Stillman (1987), we think this period of crustal extension influenced the geometry of the magmatic activity in the eastern Canary Islands but that it did not trigger it.

THE CONTACT AUREOLE

Field observations

PX1 is a banded intrusion formed of NNE–SSW alternating horizons of alkaline gabbros and pyroxenites (Hobson *et al.*, 1995) (Fig. 1). It intruded a composite host formation which consists of Tierra Mala gabbros, syenites, basic dykes, and pyroxenites all locally fenitized by the Ajuy–Solapa carbonatites and ijolites. This host formation will hereafter be called the Tierra Mala Massif (TM). Therefore, migmatization as a result of emplacement of PX1 affected rocks of different composition, grain size and structure in a zone up to 200 m wide that follows the contact with TM (Fig. 1). This zone is characterized by (1) migmatites which are formed of a dense network of leucocratic veins and fractures crosscutting a darker matrix (Fig. 2), and (2) abundant, 1–2 m large leucocratic bodies that locally brecciate the various

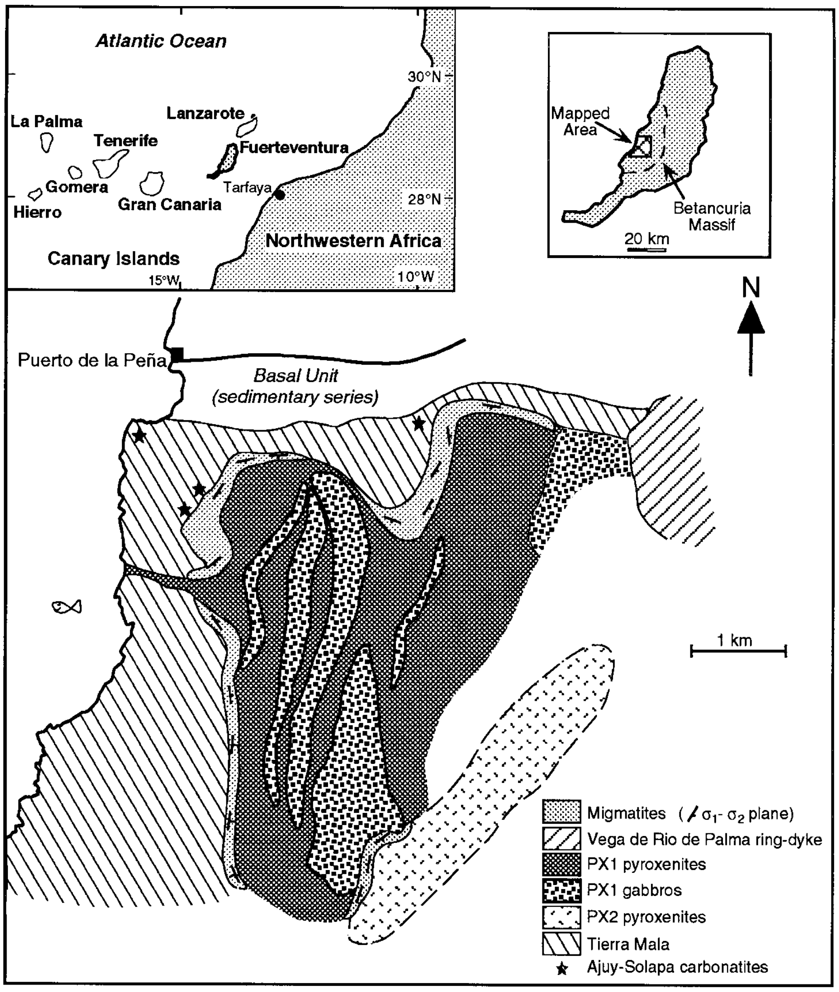

Fig. 1. Geological map of the central western part of the Betancuria Massif (Fuerteventura Basal Complex). Tierra Mala includes gabbros, pyroxenites and syenites, as well as rocks fenitized by the Ajuy–Solapa carbonatites. The Toto ring-dyke is slightly to the south of PX2, out of the field shown here. The sheeted dyke complex is not shown for clarity as it intrudes all other formations.

lithologies in the aureole. The transition from the migmatites to the undeformed country rock occurs over a few metres. Thermal metamorphism, on the other hand, is recorded up to 1 km beyond this limit (Stillman, 1987).

Partial melting and the subsequent segregation of the melts into veins were not homogeneous throughout the migmatized zone and were apparently not related to the distance from the contact. Partial melting was controlled by rock composition, i.e. rocks with seemingly no plagioclase (e.g. pyroxenites) did not melt. Segregation of the melt was controlled by deformation; there is a transition from undeformed rocks with particular patchy textures in which the leucosome forms *in situ* pods, to deformed rocks in which the melt segregated into veins. (Throughout this paper, the terms leucosome and melanosome are used purely descriptively referring respectively to the light and dark parts of the migmatite.) In the same way, pyroxenite and gabbro relics can be found in intensely veined zones. These observations imply that deformation did not trigger partial melting but only segregated pre-existing melts by creating pressure gradients in the migmatites.

The vein orientations define the stress field which prevailed during the intrusion. Large (2–3 mm wide) veins form conjugate sets (Fig. 2). Smaller sigmoid veinlets are in the acute bisector plane of the conjugate veins (Fig. 2) and follow the contact between PX1 and TM. These smaller veins were opened parallel to the direction of maximum compression (σ_1), in the near-vertical σ_1–σ_2 plane (σ_2 defined by the intersection of the conjugate veins). This is the same stress field as observed in a dyke system, and we suggest that PX1 was emplaced by way of multiple 'dyke-like' intrusions, which would account for its banded structure (Fig. 1). Deformation within the

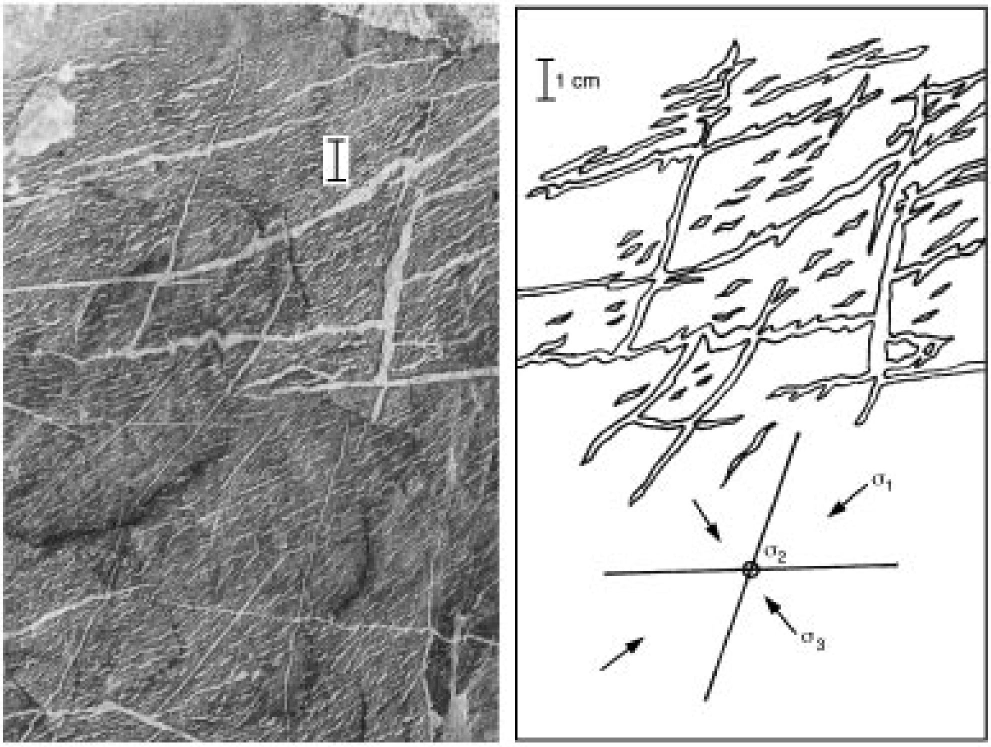

Fig. 2. Veining in a Fuerteventura migmatite. The small sigmoid veins are parallel to σ_1 and the intersection of the larger conjugate veins defines σ_2.

aureole can be directly related to the emplacement of PX1 and not to a regional event, as the σ_1–σ_2 plane (i.e. the plane containing the two main compression directions) follows the PX1–TM contact.

Textures and parageneses

The TM gabbros are homogeneous and coarse grained. Typical parageneses include augite, olivine, brown hornblende, biotite, plagioclase (An_{40-60}), apatite, titanite, ilmenite, magnetite and epidote. Ijolites and syenites contain aegirine, melanite, perovskite, alkali feldspar, plagioclase (An_{30-50}), nepheline, apatite, titanite, calcite, epidote, sodalite, ilmenite and magnetite. Magmatic textures are partly preserved in rocks which experienced partial melting without deformation. In some mafic dykes, partial melts concentrated in the interstices of a still recognizable framework of partly resorbed elongate clinopyroxenes. These leucocratic pods consist of an assemblage of neo-formed crystals of plagioclase, small grains of augite and clusters of long needles of apatite (up to 0·5 mm). Biotite is a late phase either in the leucocratic pods or as a replacement product of the clinopyroxene.

With increasing deformation, small tension veins opened and a migmatitic texture developed as the melts segregated into veins. The leucosomes consist of 60–85% plagioclase containing patches of Na-, K- and Ba-rich feldspars with some neo-formed augite and minor magnetite, ilmenite and apatite. Crystals are euhedral, large (millimetre-sized) and oriented perpendicular or oblique to the vein walls, forming a comb-like texture (Fig. 3). Augite often grows preferentially along the vein walls as xenomorphic polygonal grains, differing from the melanosome clinopyroxene in their lower inclusion content. There is hardly any evidence of solid-state deformation apart from some undulose extinction of plagioclase grains. The inter-mineral contacts are rectilinear and do not display sub-grains. Deformation must have ended before crystallization of vein material was complete.

The inter-vein dark matrix of the migmatites has a mosaic texture with triple-point contacts (Fig. 3). In some samples, this mosaic has a preferred orientation given by the alignment of elongate crystals. Grain-size is usually very small (<100 μm) except for a few 1–2 mm relic porphyroclasts of clinopyroxene and olivine. In some cases, crystals in the matrix are oriented parallel to those in the veins, indicating crystallization under stress. Typical parageneses include augite and plagioclase (An_{20-30}) with some titanite, magnetite, ilmenite and apatite. Proportion of plagioclase is variable; it is,

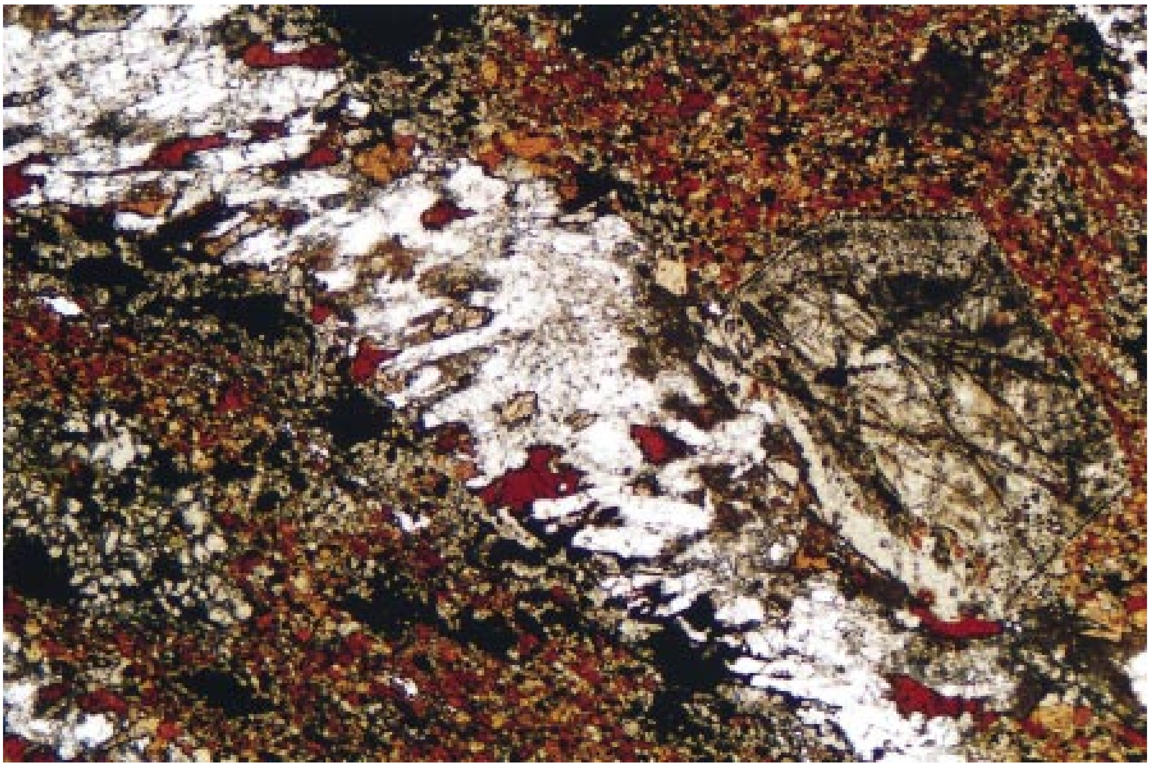

Fig. 3. Photomicrograph of a migmatite. Plagioclase forms a comb-like texture in the vein with some clinopyroxene, kaersutite and interstitial Fe–Ti oxides. The melanosome is mainly constituted of clinopyroxene, kaersutite, and Fe–Ti oxides. A relic magmatic clinopyroxene (1·5 mm long), with opaque exsolution lamellae, survived recrystallization, whereas another one (to the left of the vein) has been replaced by an aggregate of polygonal grains.

however, always absent along vein contacts where mm-wide mafic selvages appear. The latter are interpreted as restites from which all the plagioclase has been extracted.

Relic porphyroclasts include olivine and augite, which both show signs of disequilibrium. The olivine is surrounded by a reaction rim of coronitic augite, kaersutite and biotite, whereas the augite contains numerous exsolution needles of ilmenite (Fig. 3). Locally, the migmatites are hydrated, and kaersutite and biotite occur in the restite as well as in the veins, together with variable proportions of clinopyroxene.

The larger leucocratic bodies found in the aureole are syenitic unlike the migmatitic veins. Locally, small apophyses of these syenites penetrate the migmatites and join into the general vein trend. The syenites must have been at least partly molten during deformation and contributed to the leucocratic material (see the section on geochemistry). This material is therefore not entirely autochthonous (i.e. originating from the rock itself). The syenites also stayed ductile longer than the migmatites, as the latter are sometimes dismantled by the syenites.

MINERAL CHEMISTRY

In a given sample, clinopyroxene and plagioclase have slightly different compositions in the veins and the restite (Table 1 and Fig. 4), as a result of local equilibrium (re)crystallization. More important variations exist between minerals from samples of different whole-rock composition (Table 1).

Clinopyroxene

Clinopyroxenes in the migmatites are Ti-rich augites (TiO_2 = 1·6–2·76 wt %). They are zoned with Ti-, Al- and Na-richer cores, possibly reflecting higher crystallization temperatures (Adam & Green, 1994) than for the rims. Clinopyroxenes from the restite are richer in Ca, Mg and Fe^{3+} than those in the veins, which in turn contain more Al, Ti, Fe^{2+} and Na (Fig. 4).

Relic porphyroclasts are augites with sector zoning, containing abundant ilmenite exsolution lamellae (Fig. 3). They have diopside-rich cores evolving to Ti-augite rims. They contain more Ca, Al and Cr than the recrystallized clinopyroxene (Fig. 4). Their composition is similar to that of the clinopyroxene from the TM gabbros.

Table 1: Representative microprobe analyses from minerals of the PX1 and Tierra Mala intrusions

Representative plagioclase													
Migmatites (16-94-1)	SiO_2	Al_2O_3	Fe_2O_3*	MgO	CaO	Na_2O	K_2O	SrO	BaO	Total	An	Ab	Or
in melanosome	61·90	22·87	0·54	0·01	4·02	8·68	0·52	0·68	0·47	99·68	19·75	77·23	3·03
in leucosome	61·35	22·96	0·28	0·00	4·39	8·42	0·79	0·66	0·47	99·32	21·32	74·13	4·55
	61·11	22·32	0·27	0·00	3·47	7·29	2·88	0·70	0·97	99·02	17·29	65·68	17·04
Tierra Mala gabbro (AH53)	54·75	26·72	0·33	0·03	9·57	5·70	0·34	0·51	0·14	98·09	47·16	50·82	2·01

Representative clinopyroxene												
Migmatites		SiO_2	TiO_2	Al_2O_3	Cr_2O_3	Fe_2O_3	FeO	MnO	MgO	CaO	Na_2O	Total
15-94 (in melanosome)	(a) rim	50·22	1·41	3·38	0·03	3·42	3·71	0·32	14·09	22·66	0·68	99·91
	(b) core	50·29	1·54	3·22	0·05	3·32	3·97	0·33	14·30	22·39	0·66	100·07
15-94 (in leucosome)	(a) rim	51·03	1·18	2·61	0·03	3·20	4·17	0·33	14·47	22·52	0·63	100·18
	(b) core	49·91	1·63	3·56	0·05	3·03	4·43	0·27	13·68	22·31	0·75	99·62
58-94 (in melanosome)	(a) rim	50·68	1·45	3·22	0·02	3·23	3·78	0·30	14·47	22·50	0·68	100·32
	(b) core	49·41	2·11	4·01	0·01	3·55	3·88	0·25	13·92	22·16	0·79	100·10
58-94 (in leucosome)	(a) rim	49·79	2·02	3·82	0·02	3·13	3·95	0·25	14·01	22·17	0·80	99·97
	(b) core	49·54	2·15	3·98	0·01	3·15	4·20	0·24	13·83	21·99	0·84	99·94
58-94 (relic porphyroclast)	(a) rim	50·18	1·65	3·66	0·04	2·39	4·54	0·24	14·12	22·22	0·66	99·69
	(b) core	48·99	2·26	4·68	0·16	3·38	3·67	0·20	13·68	22·42	0·78	100·22
Tierra Mala gabbro (AH53)		48·80	2·11	4·37	0·04	4·14	3·17	0·20	14·00	22·30	0·73	99·86

Representative kaersutite													
Migmatites (16-94-1)	SiO_2	TiO_2	Al_2O_3	Cr_2O_3	FeO*	MnO	MgO	CaO	Na_2O	K_2O	F	O = F	Total
	39·91	6·37	11·46	0·00	12·95	0·22	11·32	11·58	2·70	1·61	0·24	0·10	98·46

Representative phlogopite												
Migmatites (15-94)	SiO_2	TiO_2	Al_2O_3	FeO*	MnO	MgO	BaO	Na_2O	K_2O	F	O = F	Total
	33·81	8·15	14·71	9·29	0·12	16·31	5·28	0·5	7·93	0·82	0·34	96·92

All oxides in wt %. FeO* and Fe_2O_3* are total iron measured as FeO and Fe_2O_3, respectively. An, Ab, and Or are anorthite, albite and orthoclase contents of plagioclase.

Feldspars

Plagioclase (An_{40-60}) is the only feldspar in the TM gabbros. In the migmatites, An_{20-30} oligoclase is found in both leucosomes and melanosomes alongside some primary K-feldspar. Pervasive replacement phenomena led to the development of alkali-rich patches and veinlets along grain boundaries and within individual plagioclase crystals (Fig. 5). A wide range in Ba, Na and K contents is observed. Within-grain filament-like patches range from $Ab_{80}An_{15}Or_5$ to $Ab_{20}An_0Or_{80}$ with up to 2 mol % celsian (Ce). Larger patches and grain boundary stringlets are Ce-richer K-feldspar ($Or_{81}Ce_4Ab_{15}$) with up to 6000 ppm Sr, whereas late fracture fillings and small spots have up to 25 mol % Ce. Pure albite vermiform patches are also present. The grain boundary-preferred location of these replacement features points to a pervasive late- to post-magmatic circulation of alkaline fluids.

Amphibole and biotite

These minerals are always associated and mainly found in the restite. The short prismatic amphiboles are kaersutites (Leake *et al.*, 1997) of homogeneous composition with Si < 6·25 p.f.u. They contain fine needles of opaque exsolution

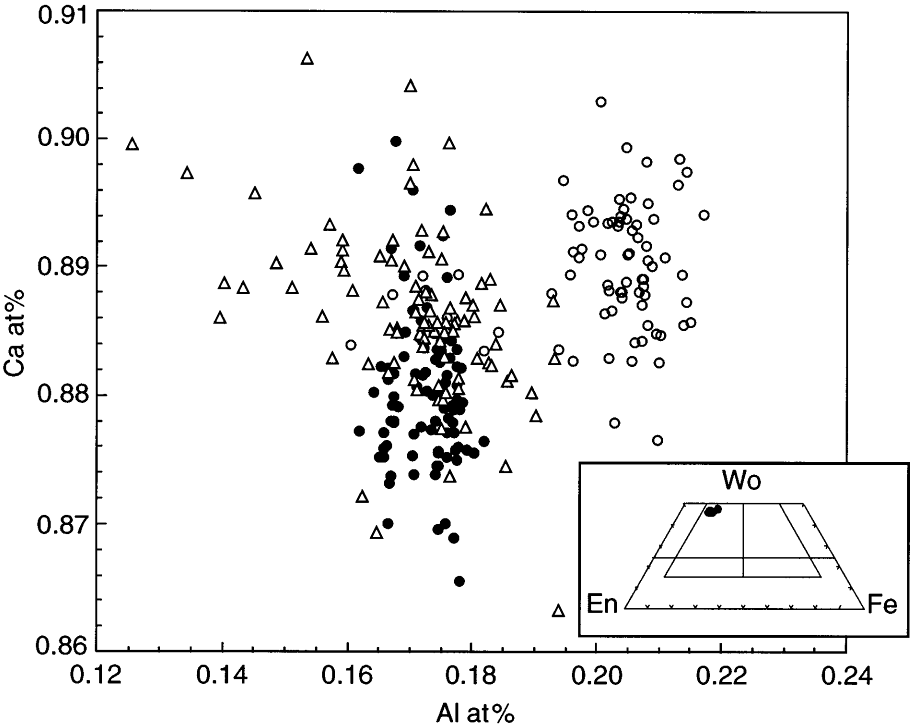

Fig. 4. Microprobe analyses of the Ca and Al content of different clinopyroxenes in the leucosome and melanosome of a single sample. A magmatic relic from the same sample was also analysed. ○, relic clinopyroxene; ●, clinopyroxene in leucosome; △, clinopyroxene in melanosome. The relic contains more Al than the neo-formed grains. Clinopyroxene is slightly more calcic in the melanosome than in the leucosome. All clinopyroxenes analysed are augites (inset).

lamellae. Biotite is a Ba-rich (~4 wt % BaO) phlogopite with up to 1 wt % F.

GEOCHEMISTRY

Migmatites are nepheline-normative, ultrabasic to basic rocks with *mg*-numbers between 29 and 37. Their composition range is comparable with that of the TM gabbros (*mg*-number 25–53). They are, however, enriched in Al, Na, P, Sr, Ba, Nb, Y and the rare earth elements (REE), and depleted in Rb, Ni and Cr (Table 2 and Fig. 6a–c). The *mg*-number of the gabbros covers a wider range (25–53). Major element composition of the leucosome and melanosome (Table 2) is controlled by their respective abundance in plagioclase and clinopyroxene. Trace elements were redistributed during partial melting according to their affinities for specific mineral phases, through solid–solid and solid–liquid reactions. Melting of the plagioclase produced an Al-, Sr- and Ba-rich melt. Alumina, preferentially incorporated by the melt phase, segregated into the veins. This Al segregation and a lower temperature explain the lower Al content of the recrystallized clinopyroxene compared with the magmatic one (Fig. 4). Cr and Ni from the clinopyroxene were incorporated in the recrystallizing Fe and Ti oxides and remained in the restite (Table 2). P and the REE, which are clearly enriched in the melanosome (Table 2 and Fig. 6a and d), are concentrated in fine apatite needles included in recrystallized clinopyroxene, residual plagioclase and hydrous phases when present.

Migmatites are all enriched in Al compared with their protolith, the TM gabbros (Table 2). Therefore, the recrystallization of Al-poor sodic plagioclase (Table 1) and pyroxene (Fig. 4 and Table 1) cannot satisfy the mass balance of Al in the migmatites without an input of an Al-rich (i.e. feldspathic) phase. Such a phase could also supply Na, Ba and Sr. It probably consisted of silicate melts and fluids rich in dissolved elements, percolating during and after migmatization. Melts derived from the anatexis of syenites are locally observed in the field. Circulation of alkaline fluids is evidenced by the composition and distribution of alkali- and Ba-rich feldspars (Fig. 5), and the Ba and F content of the phlogopite in the migmatites.

The migmatites' high content in P and high field strength elements (HFSE; Nb, Y and REE) cannot be

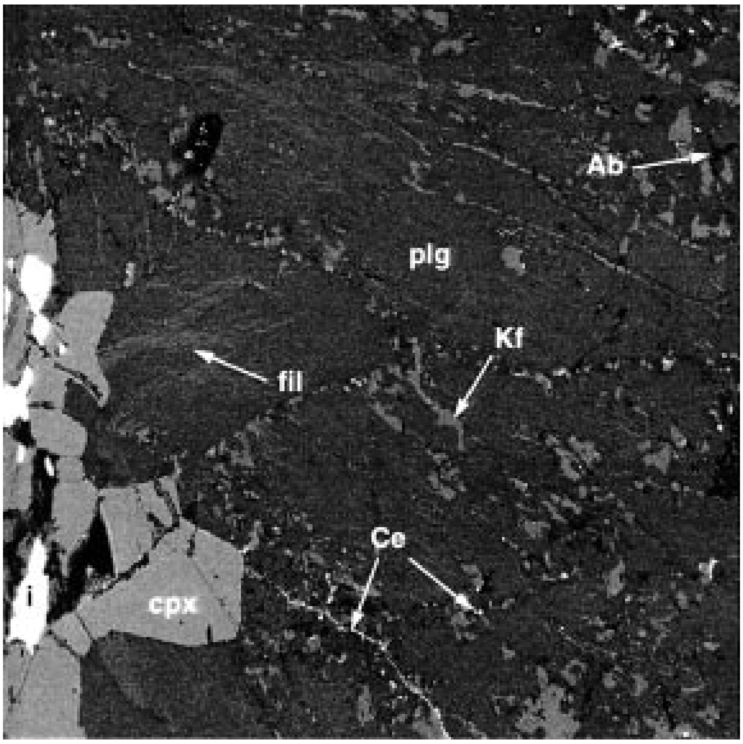

Fig. 5. Back-scattered electron image of a leucosome at the contact of a melanosome. Length of picture is 600 μm. cpx and i are clinopyroxene and ilmenite in the melanosome; plg is plagioclase; Ab (darker grey) is a small replacement 'spot' of vermiform albite in a plagioclase; Kf (medium grey) is alkali feldspar of varying composition (see text) forming replacement patches in all plagioclase grains; Ce (white) is celsian-rich alkali feldspar found both as fine stringlets along grain boundaries and as small replacement spots and late fracture fillings; fil, filament-like K-feldspar replacement patches in plagioclase grains. The primary K-feldspar grains mentioned in the text are not visible here.

explained by a relative enrichment as a result of loss of an anatectic melt, nor by percolation of simple alkaline fluids. However, Cl-, F- and/or S-rich fluids are known to be able to dissolve P- and REE-rich minerals such as apatite (Ayers & Watson, 1991) and to efficiently transport these elements (Gieré & Williams, 1992). Fluid circulation, enhanced by shearing along the PX1–TM contact, also best explains the homogeneous enrichment of the migmatites throughout the aureole.

Potential sources for P and HFSE are the fenitized TM rocks (Fig. 6b and Table 2) and the syenites for Al and alkali elements. The fluids leaching these source rocks could be: (1) late magmatic fluids related to the Ajuy–Solapa carbonatite complex; (2) meteoric water heated by PX1; (3) fluids released by the country rock during thermal metamorphism.

P–T ESTIMATIONS

Common cation-exchange geothermobarometers could not be applied because of the absence of key phases such as quartz, orthopyroxene or garnet. Fe–Ti oxides are entirely re-equilibrated at low temperature and, similarly, preliminary results from stable isotope thermometry on the basis of oxygen isotope fractionation between clinopyroxene and plagioclase yielded temperatures of <600°C. These values probably reflect the temperature of alkaline fluids which circulated in the aureole.

The available experimental work on partial melting mostly refers to quartz-normative or quartz-bearing mafic rocks and phase diagrams cannot be strictly applied to mafic alkaline nepheline-normative rocks. We can, however, somewhat constrain the *P–T* conditions. Concerning the pressure, the depth to the Moho under the Fuerteventura–Lanzarote ridge is ~11 km (Banda *et al.*, 1981). Also, Fuerteventura probably once hosted a volcano ~3 km high, according to oxygen and hydrogen isotopes (Javoy *et al.*, 1986). Therefore, the pressure at which the plutons were emplaced could not have exceeded 4 kbar, and was probably less (maybe only 2 or 3 kbar), as the intrusions were emplaced in the overturned ocean-floor sediments. Such pressures are in agreement with the absence of garnet.

The presence of hornblende and biotite in most samples implies that sufficient water was available in the melts.

Table 2: Representative whole-rock analyses of the different lithologies (PX1, Tierra Mala and migmatites)

Formation:	PX1	Tierra Mala						Migmatites				
Rock:	g.	g.	g.	g.	sy.	fen.g.	fen.g.	mig.	mig.	mig.	leuco.	mela.
Sample no.:	22-94	30-95	12-95	59-95	16-95	22-92	10-92	37-94	01-95	22-96	22-96	22-96
SiO_2	44·10	45·21	40·10	42·22	59·79	39·26	33·75	46·65	39·80	45·79	61·37	42·89
TiO_2	3·00	2·73	5·08	5·51	0·43	2·07	5·16	3·55	5·53	3·58	0·08	4·80
Al_2O_3	14·70	7·86	10·12	14·81	19·14	17·59	10·00	15·84	14·25	14·80	22·98	14·07
Fe_2O_3	9·27	6·31	8·16	7·29	2·80	1·75	9·95	7·69	9·46	7·27		
FeO	3·26	7·09	6·90	6·94	1·22	2·84	4·42	3·35	5·10	4·19	0·20	12·82
MnO	0·13	0·17	0·21	0·17	0·07	0·19	0·26	0·19	0·21	0·22	0·01	0·24
MgO	7·91	14·49	9·12	5·09	0·40	4·39	9·51	4·94	6·21	5·40	0·01	7·61
CaO	14·47	11·37	14·71	11·20	1·96	12·63	17·41	8·76	13·25	10·32	3·93	10·06
Na_2O	1·94	1·64	2·25	3·22	6·78	6·18	1·00	4·59	2·35	4·26	8·19	4·03
K_2O	0·10	0·87	0·63	0·84	4·51	0·61	1·58	1·39	0·39	1·24	1·93	1·34
P_2O_5	0·12	0·23	0·67	0·52	0·05	0·12	2·94	1·09	1·53	1·05	0·01	0·91
H_2O	0·22	0·48	0·99	0·58	1·03	6·48	2·69	1·00	1·05	0·94		
CO_2	0·09	0·18	0·14	0·13	0·21	3·64	0·10	0·04	0·05	0·40		
Total	99·41	98·63	99·08	98·52	98·39	97·75	98·77	99·09	99·17	99·45	98·72	98·77
mg-no.	41	53	39	27	10	50	42	32	31	33	6	37
Nb	<5	20	59	49	<5	203	91	124	74	64		
Zr	60	195	349	301	3388	101	532	378	212	309		
Sr	1046	298	913	890	1867	6224	1461	1295	1665	814	2687	815
Rb	6	20	14	21	71	20	58	14	5	18		
Ga	22	6	13	17	43	13	22	25	23	26		
Zn	88	98	122	101	47	77	156	105	115	112		
Cu	95	115	35	60	66	37	156	116	209	163		
Ni	169	479	98	77	<2	6	4	36	24	105		
Cr	230	1250	215	28	<3	20	29	42	14	198		
V	409	322	604	504	113	97	330	268	470	422		
Ba	101	191	350	370	1224	444	807	870	356	550	1576	947
Th	<1	2·1	2	<1	<1	26	41	5·8	<1	20·4	25·6	11·3
U	<2	0·7	0·6	<2	8·3	24	9	1·3	<2	6·6	9·5	1·9
S	278	355	89	86	2·8	673	1335	272	105	147		
Hf	2	2	6	3	56	14	11	7	7	8		
Y	14	18	36	30·1	15	38	58	34	33	39	29	53
La	7·8	21·4	39·8	86	18·9	176	112	84·6	27	127	71	152
Ce	21·3	49·1	99·9	41	28	323	1335	164	202	240	129	289
Pr	3·4	6·6	14·4		2·7			19·5		28·4	15·2	34·8
Nd	17·6	27·3	62·8	370	9·2	156	11	72·2	81	101	57·3	127
Sm	4·9	6·1	13·5		1·7			12·5		16·8	10·6	22
Eu	1·92	2	4·39		1·84			4·26		5·42	4·35	5·55
Gd	4·5	5·7	12·1		1·8			11		14·4	8	17
Tb	0·6	0·9	1·7		0·3			1·5		1·9	1·2	2·3
Dy	3·4	4·1	7·6		1·8			6·3		8·3	5·7	10·2
Ho	0·58	0·76	1·36		0·47			1·23		1·6	1·09	1·84
Er	1·4	1·7	3		1·4			2·8		3·5	2·4	4·2
Tm	0·2	0·2	0·3		0·3			0·3		0·5	0·3	0·6
Yb	0·9	1·5	2		2			2·1		3	2·1	3·5
Lu	0·12	0·18	0·3		0·34			0·29		0·4	0·3	0·52

Major elements in wt % and trace elements in ppm. Major and trace elements analysed by XRF except for the leucosome and melanosome, which were analysed by electron microprobe (focused beam rastered over an area of 100 mm × 80 mm). Leucosome and melanosome whole-rock analyses are means of up to 50 points in profiles in the leucosomes and melanosomes. REE, Y, U and Th analysed by ICP-MS except for Y, La, Ce and Nd for samples 10-92, 22-92, 37-94 and 1-95, which are by XRF. *mg*-number = 100 × MgO/(MgO + FeO*), with FeO* calculated using FeO = $0·9Fe_2O_3$. FeO in the leucosome and melanosome measured as total Fe. Analyses preceded by < are below the detection limit. When no data are shown there have been no analyses. g., gabbro; sy., syenite; fen.g., fenitized gabbro; mig., migmatite; leuco., leucosome; mela., melanosome.

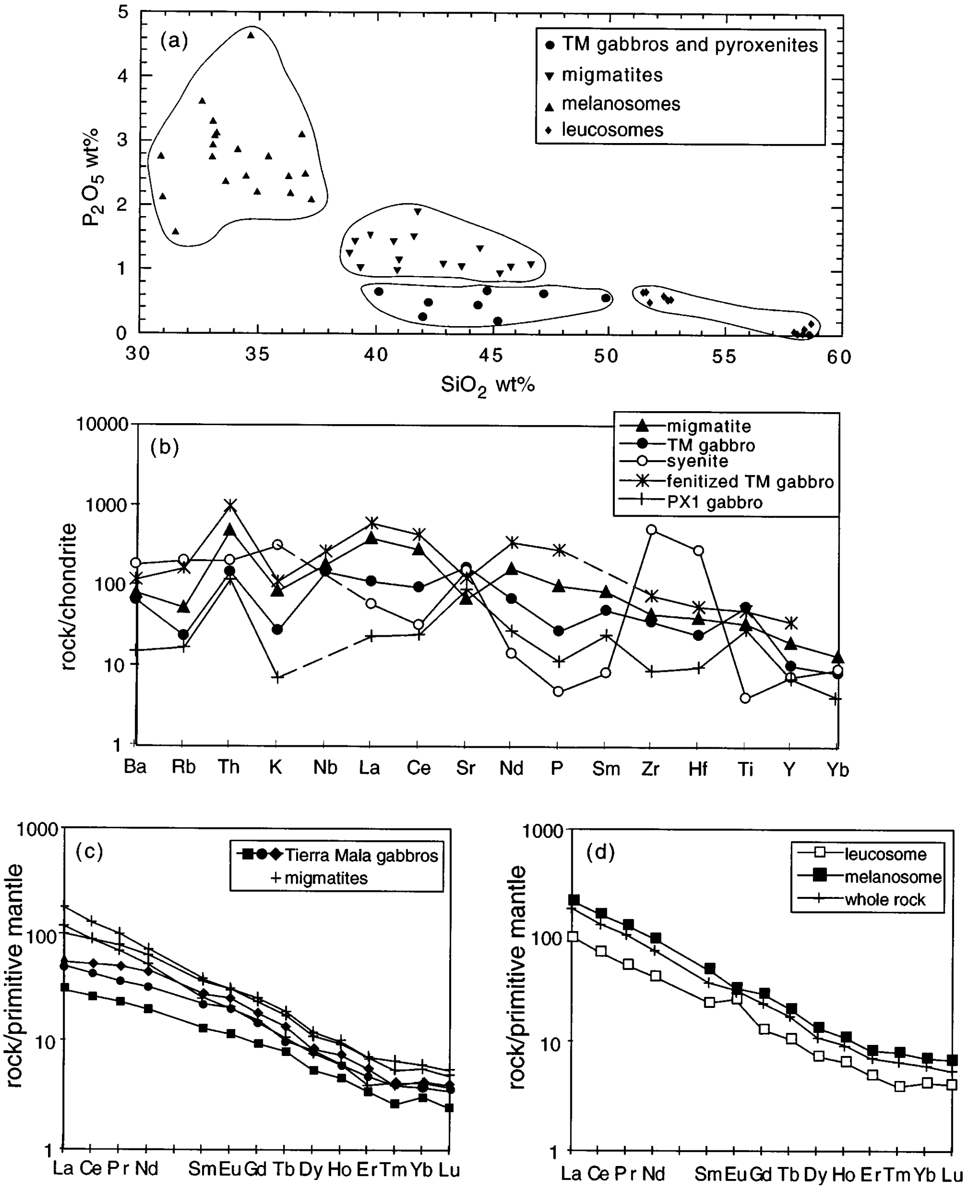

Fig. 6. (a) SiO_2 (wt %) vs P_2O_5 (wt %) variation diagram of the Tierra Mala gabbros and pyroxenites, the migmatites, and the leucosomes and melanosomes. Gabbros, pyroxenites and migmatites are XRF analyses. Whole-rock leucosome and melanosome analyses as in Table 1. (b) Trace element abundances of a PX1 gabbro and rocks from its contact aureole. Abundances normalized to the values of Thompson (1982). (c) REE abundances in migmatites and Tierra Mala gabbros and pyroxenites. Abundances normalized to the primitive mantle of McDonough *et al.* (1991). (d) REE abundances in a migmatite and its leucosome and melanosome. Abundances normalized to the primitive mantle of McDonough *et al.* (1991).

Also, the melting point of plagioclase is significantly lowered in presence of water (e.g. Housh & Luhr, 1991), which probably facilitated partial melting of the gabbros. In some migmatites, plagioclase has been entirely melted. According to Beard & Lofgren (1991), the plagioclase-out curve for a water-saturated basalt is 1020°C at 2 kbar and 975°C at 3 kbar. Similar values were found by Holloway & Burnham (1972) with water pressure less than total pressure: plagioclase is 'out' at 1000°C at 2 kbar and at 960°C at 3 kbar. These values are in agreement with the temperature at which clinopyroxene is stable (Spear, 1981). Temperatures of 960–1010°C could be within the range reached by the migmatites during contact metamorphism. However, taking into account the higher albite content of plagioclase in the aureole compared with the plagioclase in experimental work, lower temperatures could have been more likely.

DISCUSSION

Partial melting of mafic rocks in various geological settings has been well documented within the last 15 years. Many cases relate to the migmatization of tholeiitic rocks metamorphosed under amphibolite facies conditions in orogenic belts (e.g. Sorensen, 1988; Sawyer, 1991; Williams *et al.*, 1995) or oceanic shear zones (e.g. Flagler & Spray, 1991). Common features of such migmatites are quartz-bearing leucosomes of tonalitic to trondhjemitic composition and garnet-bearing residues produced under relatively high confining pressures above 8 kbar and temperatures around 750°C. Fluids often play a major role in lowering solidus temperatures and as a means for mass transfer, especially for large ion lithophile elements (LILE). Vapour-absent melting has also been reported for amphibolites (Williams *et al.*, 1995) and mafic granulites (Tait & Harley, 1988). Deformation is always essential to melt segregation (Vigneresse *et al.*, 1991; Sawyer, 1994, 1996) and is often thought to facilitate fluid circulation if present, and, correlatively, partial melting.

Anatexis of gabbros in Fuerteventura took place under very different conditions in terms of initial rock composition and lithostatic pressure, and resulted from the unusual conjunction of several parameters. The shallow-level intrusion of a gabbro is not expected to supply sufficient heat to trigger partial melting in a country rock of similar composition. However, in Fuerteventura, the ocean-island tectonic context allowed the development of a long-lived basic magmatic system that was focused in a restricted area. This resulted in anomalously high heat flow, which probably kept the country rocks close to their solidus temperature. Relatively small additional heat input supplied by subsequent intrusions would then be sufficient to start the melting process. According to the parageneses in the migmatites, temperatures of 900–1000°C could have been reached for pressures of 2–3 kbar. However, this unusual thermal regime would probably not have been sufficient to induce a 50 vol. % melting of nepheline-normative gabbros with complete dissolution of the plagioclase phase. High-temperature alkaline and probably halogen-rich fluids, present in the contact aureole at the beginning of PX1 intrusion, were most probably a decisive factor in lowering the overall solidus temperature of the host rocks.

Although shearing did not directly trigger anatexis, it is definitely responsible for melt segregation and was essential to enhanced fluid circulation as evidenced by the pervasive chemical transformations within the whole aureole. Shearing within the Fuerteventura nested intrusions was not expected, as the whole region was in an overall extensional–transtensional setting and experienced up to 30 km crustal distention (Le Bas *et al.*, 1986). However, in the case of PX1, periods of crustal dilation and magma emplacement did not coincide and the intrusion took place in a compressive–transpressive environment. This resulted in intense shearing all along the PX1–TM contact and the subsequent opening of abundant tension veins in the migmatites. The few areas of the aureole which were shielded from deformation did not develop veins, and partial melts remained as *in situ* pods in the rock.

The volumetrically dominant small tension veins (parallel to the maximum compression direction σ_1) usually consist of leucosomes with up to 40 vol. % ferromagnesian minerals, as is expected from experimental phase diagrams. The wider conjugate veins are generally much more feldspathic, with as little as 5 vol % Fe–Mg minerals. These veins are not considered to reflect the initial composition of partial melts, but that of fractionated liquids with a contribution from TM syenitic melts. The unusually high albite normative content of the leucosomes compared with regular gabbroic liquids is thought to influence the crystallization path. Neo-crystallization of clinopyroxene from the partial melt is expected to start before liquid segregation, which would increase the molar proportion of feldspar in the residual liquid. After segregation of the latter, clinopyroxene saturation and crystallization along vein walls could further fractionate the plagioclase component in the vein.

At first sight, the spectacular comb-like texture of the leucosome crystals would suggest that deformation stopped just after melt segregation and before crystallization. In fact, recent experimental work on analogue material (norcamphor) by Rosenberg & Handy (1997, and personal communication) showed that partial melting under continuous oriented compressional stress led to the opening of small veinlets parallel to σ_1, into which melt segregated and started to crystallize at right angles to the vein walls (i.e. parallel to σ_3), as observed in the migmatites. Consequently, these melts might well have started to crystallize during shearing, but the latter ended before complete solidification, as plagioclase crystals do not show any sign of subsolidus or fragile deformation.

The very particular circumstances which led to the low-pressure partial melting of nepheline-normative gabbros make it an exceptional phenomenon, which is not expected to occur frequently nor at a large scale. Similar conditions could be imagined in other high heat flow contexts such as in mid-ocean ridges experiencing intra-oceanic deformation in the presence of fluids or, possibly, in some Large Igneous Provinces (LIPs) such as oceanic basaltic plateaux, in which very high production rates of basalts might supply enough heat to melt rocks of similar composition.

CONCLUSION

Partial melting of gabbroic rocks in the contact aureole of the PX1 pluton (Fuerteventura Basal Complex) occurred at low pressure (2–3 kbar), high temperature (1000°C) and in the presence of an alkaline fluid phase. These conditions result from the combination of magmatic (nested intrusions, hot-spot context, metasomatism) and tectonic factors (regional shear stress). The Fuerteventura migmatites provide a rare opportunity to address fundamental mechanisms such as: (1) the formation of anatectic melts in an oceanic environment; (2) the nature and origin of fluids and their impact on selective element enrichment in migmatites; (3) melt segregation mechanisms and crystallization processes in tensional fractures in a transpressive stress field. The fusion mechanisms of basic material at low pressure illustrated here may potentially exist in other areas, where the above-cited conditions are satisfied, but the preservation of magmatic recrystallization textures as observed in Fuerteventura seems to be exceptional.

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