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Oxygen isotope composition of xenoliths from the oceanic crust and volcanic edifice beneath Gran Canaria (Canary Islands): consequences for crustal contamination of ascending magmas

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Abstract

Xenolith samples of marine terrigenous sediments and altered Jurassic MORB from Gran Canaria (Canary Islands) represent samples of sub-island oceanic crust. These samples are postulated to define end-members for crustal contamination of basaltic and felsic ocean island magmas. The meta-igneous rocks show great heterogeneity in oxygen isotope compositions (δ^{18}O 3.3–8.6 ‰), broadly correlating with their stratigraphic position in the oceanic crust. Gabbros interpreted as fragments of oceanic crust layer 3 have δ^{18}O values of 3.3–5.1 ‰, which is lower than MORB (5.7–6.0 ‰). Layer 2 lavas and dykes show a broader range of δ^{18}O of 4.1–8.6 ‰. Therefore, high-temperature metamorphism seems to have been the dominant process in layer 3, while both high- and low-temperature alteration have variably affected layer 2 rocks. Siliciclastic sediments have high δ^{18}O values (14.1–16.4 ‰), indicating diagenesis and low-temperature interaction with seawater. The oxygen isotope stratigraphy of the crust beneath Gran Canaria is typical for old oceanic crust and resembles that in ophiolites. The lithologic boundary between older oceanic crust and the igneous core complex at 8–10 km depth—as postulated from geophysical data—probably coincides with a main magma stagnation level. There, the Miocene shield phase magmas interacted with preexisting oceanic crust. We suggest that the range in δ^{18}O values (5.2–6.8 ‰) [Chem. Geol. 135 (1997) 233] found for shield basalts on Gran Canaria, and those in some Miocene felsic units (6.0–8.5 ‰), are best explained by assimilation of various amounts and combinations of oceanic and island crustal rocks and do not necessarily reflect mantle source characteristics.

Keywords: Oxygen isotopes; Oceanic crust; Xenoliths; Sediments; Gran Canaria

1. Introduction

Petrologic research has focussed for many years on the apparent heterogeneity of mantle source domains, assuming that ocean-island basalt (OIB) magmas truly record the composition of the underlying mantle. Several recent studies have shown, however, that variations in isotopic and trace element patterns of OIB magmas can to some extent be explained by crustal contamination (e.g., Wolff and Palacz, 1989; Eiler et al., 1996; Thirlwall et al., 1997; Bohrson and Reid, 1997; O'Hara, 1998; Harris et al., 2000; Wolff

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et al., 2000). Possible contaminants include the island's edifice and the underlying oceanic crust. Rocks of the edifice and oceanic crust can display a high degree of chemical heterogeneity due to diverse source compositions and a variable hydrothermal overprint. Pre-island oceanic crust and an island's core complex therefore need to be well characterized in order to truly account for their impact on ocean island petrogenesis.

For this reason, we studied the oxygen isotope composition of variably metamorphosed Jurassic MORB-type gabbro and basalt xenoliths (Schmincke et al., 1998), fresh Jurassic to Tertiary sedimentary xenoliths (Hoernle, 1998) and thermally overprinted plutonic xenoliths from the volcanic edifice at Gran Canaria. The results help to constrain the oxygen isotope composition of old Atlantic crust and the plutonic core of the island and, thus, assist evaluation of their role as end-members for contamination processes.

2. Geologic setting

The Canary Islands form a chain of seven large volcanic islands located off the NW African continental shelf (Fig. 1). All islands are underlain by oceanic crust as indicated by tholeiitic MORB gabbro xenoliths occurring on Lanzarote, Gran Canaria and La Palma (Schmincke et al., 1998; Hoernle, 1998). The age of the igneous crust is bracketed by paleomagnetic anomalies S1 (175 Ma) between the easternmost islands and the African coast and M25 (155 Ma) between La Palma and El Hierro, the westernmost and youngest islands (Roeser, 1982). The Canary Islands are characterized by multiple volcanic phases and a large chemical diversity as exemplified by the stratigraphy of Gran Canaria (e.g., Schmincke, 1982; Hoernle and Schmincke, 1993; Cousens et al., 1990). Briefly, the volcanism on Gran Canaria can be divided into three phases: (1) The Miocene phase (15–8.5 Ma) comprises the shield series of transitional to

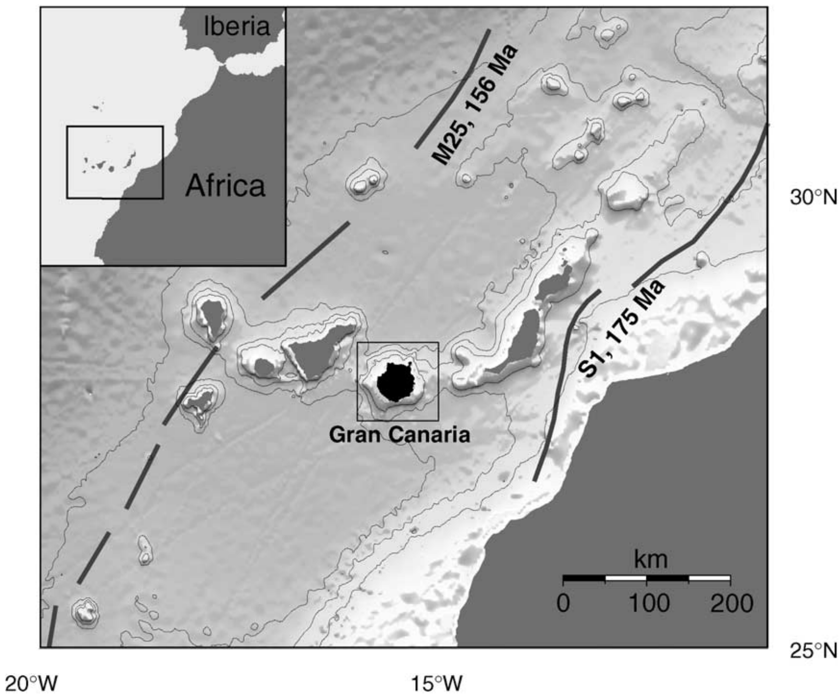

Fig. 1. Map of the northwest African margin and the Canary Archipelago. The Canary Islands are located between the magnetic anomalies S1 (175 Ma) and M25 (156 Ma) (Roeser, 1982). The bathymetric contour interval is 1000 m.

mildly alkalic basalts to picrites (ca. 15–13.8 Ma), and >1000 km^3 of subalkaline to peralkaline trachytes, rhyolites and phonolites (13.8–8.5 Ma) (Fig. 2). (2) The Pliocene phase (5.0–3.0 Ma) commenced with eruption of nephelinites followed by alkali basaltic to trachytic and basanitic to phonolitic series. (3) Nephelinites, melilite nephelinites and especially basanites dominate the late Quaternary to Holocene volcanism.

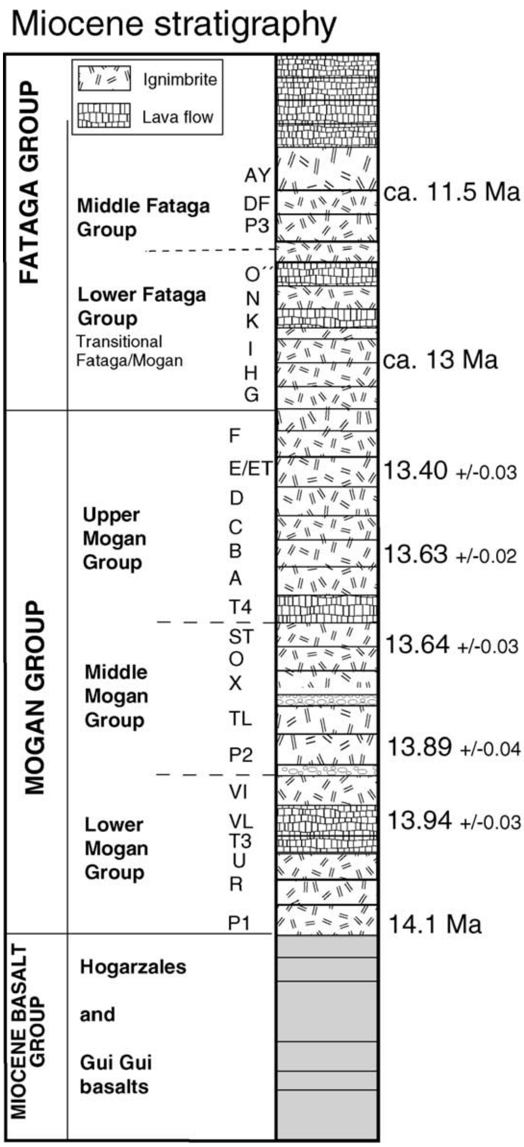

Fig. 2. Simplified stratigraphic section of Miocene eruptive episodes on Gran Canaria (modified after Sumita and Schmincke, 1998).

3. Petrography

3.1. MORB-type xenoliths

MORB-type xenoliths occur in Miocene fanglomerates overlying the shield basalts on Gran Canaria in Barranco de Balos, and are interpreted to represent fragments of layer 3 (gabbros) and layer 2 (lavas and dykes) of the oceanic crust beneath Gran Canaria (Schmincke et al., 1998; Hoernle, 1998) (Fig. 3). The rocks comprise a wide spectrum of weakly to strongly metamorphosed greenschist facies metabasalts, metadiorites and metagabbros, but contain some relatively fresh clinopyroxene ± orthopyroxene-bearing gabbros. Some clasts apparently represent fragments of pillows or dykes, as is evident from their fine-grained margins. We selected typical rocks from the following three textural groups for further analysis:

(a) Nearly aphyric and very fine grained lavas and dykes (grain sizes < 1 mm) (Fig. 3c,d).
(b) Medium- to fine-grained (average grain sizes about 1–2 mm) and commonly phyric rocks texturally dominated by abundant plagioclase, probably representing slowly cooled lava flows.
(c) Cumulate textured gabbros with grain sizes ≥ 2 mm (Fig. 3a,b). Common varieties include plagioclase + clinopyroxene ± orthopyroxene accumulate, and clinopyroxene ± orthopyroxene-rich plagioclase heteroaccumulate in which the ≤ 2 cm large pyroxenes are plagicolase-poikilitic. Clinopyroxene and orthopyroxene cores contain exsolved orthopyroxene and clinopyroxene, respectively.

Primary igneous minerals are plagioclase (An 60–80, Or 0.01–0.6), augitic clinopyroxene (En 44–54, Fs 8–16, Wo 38.5–41) ± orthopyroxene and Fe–Ti oxides (Schmincke et al., 1998). The primary mineralogy has been modified to a variable extent at upper greenschist facies conditions, as reflected in secondary actinolite or ferroactinolite, Fe–Mg chlorite, epidote, sphene, ilmenite ± hematite, ± Ab-rich plagioclase (An 5–15), ± muscovite, ± biotite. Several

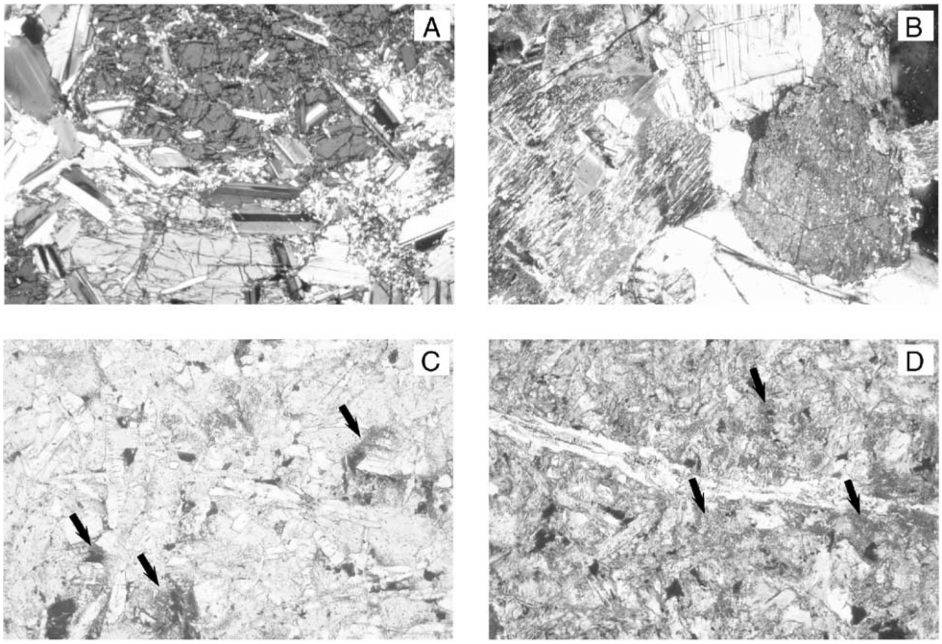

Fig. 3. Photomicrographs of igneous xenoliths from the Jurassic ocean crust. (A) Noritic gabbro HATB914 showing optically fresh plagioclase and incipient chlorite formation (fine-grained aggregates) at clinopyroxene (top) and orthopyroxene (bottom) rims. Crossed polarisers; field of view ca. 1 mm. (B) Incipient internal chloritization along exsolution lamellae in clinopyroxene (grain to the left). Gabbro HATB9125, crossed polarisers; field of view ca. 1 mm. (C) Metamorphosed lava HAT9110 containing plagioclase, almost completely chloritized clinopyroxene, Fe-actinolite, aggregates of microcrystalline sphene (arrows) and minor opaque oxides. Field of view ca. 0.5 mm. (D) Strongly metamorphosed lava 303905 containing plagioclase, chlorite, epidote, Fe-actinolite, aggregates of microcrystalline sphene (arrows) and minor opaque oxides. The sample is penetrated by epidote-filled fractures. Field of view ca. 0.5 mm.

samples contain iron sulphide. The contents of ilmenite ± hematite approach 1 vol.% in some cases (e.g., sample B9117). Sphene, which is finely disseminated in the fine-grained samples, may exceed 1–2 vol.% (e.g., samples B913 and B9117). Most fine-grained metabasalts (e.g., samples 303905 and B912; Fig. 3d) consist almost entirely of this greenschist facies assemblage. The more altered samples are locally penetrated by fractures up to about 0.5 mm wide, filled with epidote, sphene, ± biotite, ± chlorite. However, many of the coarser grained or phyric samples contain igneous pyroxene and plagioclase with optically fresh cores (e.g., samples B914 and B9117; Fig. 3a,b).

3.2. Sedimentary xenoliths

The sedimentary xenoliths, which range in size from 3 to about 12 cm, occur in several Quaternary basanitic fallout deposits (tephra cones) on Gran Canaria, and are thought to represent terrigeneous material transported across the African continental shelf. The rocks are subrounded to rounded due to melting and assimilation in the host magma, and some samples have glassy crusts (Fig. 4). The sedimentary xenoliths are porous and have a foamy appearance due to vesicle contents of 20–50% and glass contents between 10% and 35%. Some samples are partly disaggregated. Dominating pre-melting textures are partly interlocking and tightly quartz-cemented quartz and feldspar grains typically ranging in size between 100 and 300 μm. Quartz grains are commonly surrounded by "dusty zones" of quartz cement, typical for deeply buried siliceous sediments that experienced diagenesis. Due to the high glass contents, the xenoliths contain two types of textural domains: (a) a patchwork of tightly cemented quartz and feldspar grains, and (b) glass-dominated vesicular interstices. Glass in the xenoliths comprises a dark (in this case, basanitic) type intruding along cracks from the host

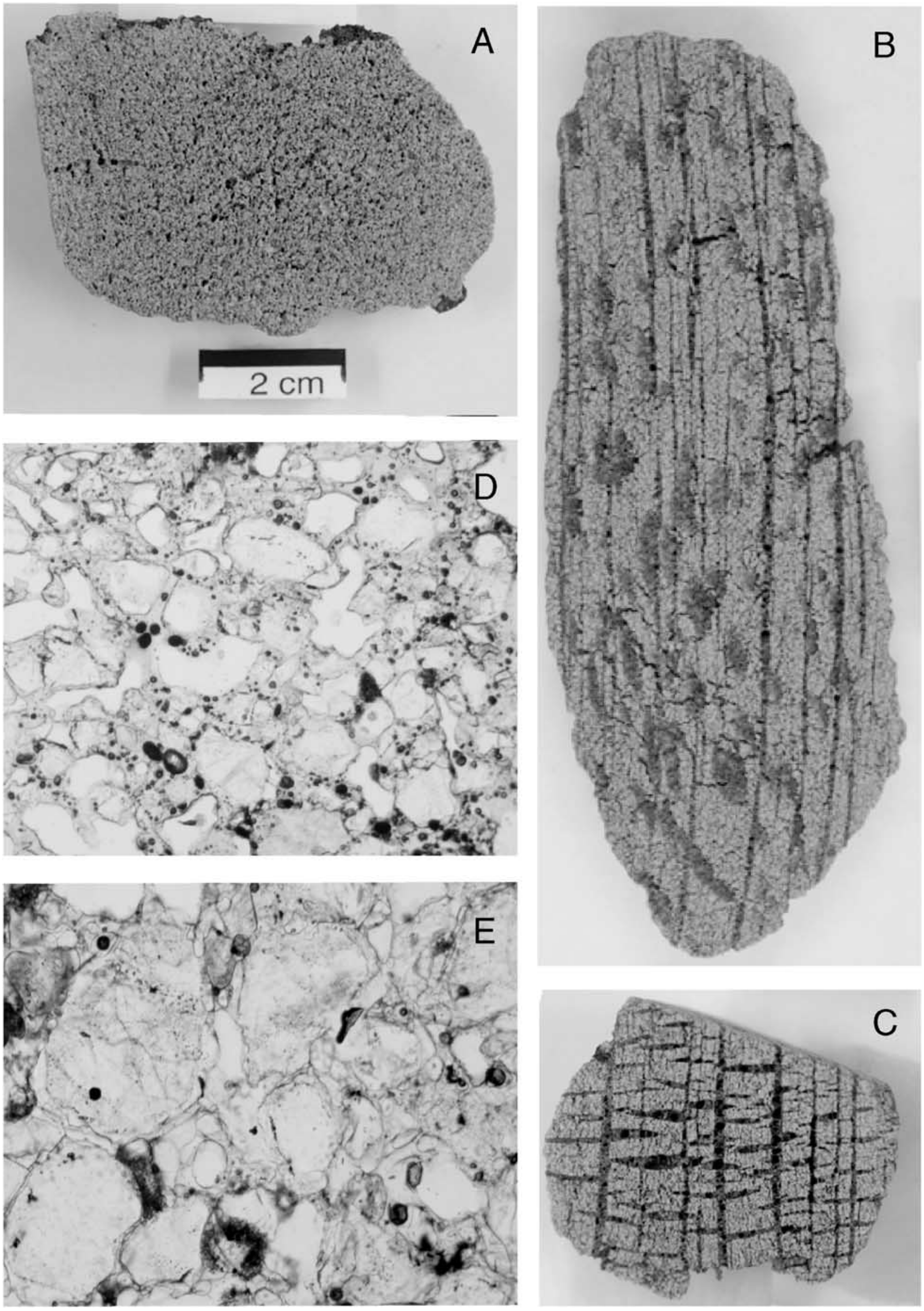

Fig. 4. Hand specimen photos and photomicrographs of sedimentary xenoliths. A to C show sawn surfaces of rounded xenoliths containing numerous mm-sized to microscopic vesicles and dark glassy patches. Scale bar in A applies to all hand specimens. (A) The comparatively homogeneous sample HAT917B. (B, C) Two perpendicular sections through sample HAT917C. Dark layers and patches contain highly vesicular basaltic glass. (D) Photomicrograph showing a mixture of irregular and rounded quartz and less abundant feldspar crystals, about 20 vol.% vesicles and 20 vol.% nearly colourless interstitial glass (sample HAT917B, field of view ca. 2 mm). (E) Photomicrograph showing rounded quartz grains with diagenetic overgrowths, surrounded by nearly colourless interstitial glass (sample HAT917D, field of view ca. 1 mm).

Table 1
Compositions of siliciclastic sedimentary xenoliths from Gran Canaria

	HAT917B	HAT917C	HAT917D	HAT917E	HAT91249
wt.%					
SiO_2	87.2	86.7	87.3	87.1	85.6
TiO_2	0.3	0.4	0.2	0.4	0.3
AL_2O_3	4.5	4.9	2.9	4.8	5.1
Fe_2O_3	2.0	2.5	1.8	1.7	2.5
MnO	0.05	0.06	0.05	0.02	0.09
MgO	1.4	1.3	1.8	0.8	0.9
CaO	2.0	1.2	2.7	0.7	3.0
Na_2O	0.8	0.8	0.5	1.1	0.9
K_2O	0.7	0.9	0.2	1.1	0.9
P_2O_5	0.05	0.06	0.04	0.05	0.05
H_2O	1.6	1.5	1.0	1.1	0.7
CO_2	0.11	0.07	0.17	0.03	0.02
Sum	100.74	100.38	98.76	99.04	100.00

magma into xenolith interiors, and a nearly colourless (felsic) type associated with partial melting of the sedimentary xenolith minerals. To our knowledge, this is the first description of such sedimentary xenoliths from Atlantic islands. Bulk-rock analyses of xenolith interiors are given in Table 1.

4. Analytical methods

4.1. XRF

Central pieces of the xenoliths were cut and selected for whole-rock analyses, avoiding cracks and rims affected by surface weathering. Samples for whole-rock analyses were crushed and powdered in an agate mortar and an agate ball mill. Samples were dried at 110 °C prior to analyses. Major and trace elements were determined by XRF on fused beads using an automated Philips PW1480 spectrometer at GEOMAR. All analyses were performed with a Rh tube; calibration was performed using international geological reference samples BHVO-1 (Govindaraju, 1994), JA-2, JB-2, JB-3 and JR-1 (www.iaea.or.at/programmes/nahunet/e4/nmrm/material/gsjjr1.htm). Standard analyses are given by Abratis et al. (2002). H_2O and CO_2 were analysed upon ignition of powders at 1200 °C using a Rosemount CWA 5003 infrared photometer. Full descriptions of analytical method is available from the authors upon request.

4.2. Oxygen isotopes

Whole-rock samples and feldspar separates were hand-picked under high-magnification stereomicroscopes, and only carefully selected chips and crystals were used for analyses. Mineral separates were cleaned in a 10% ultrapure HF-solution for 5–10 min. Oxygen isotopes were analysed at the University of Capetown by C. Harris using a conventional vacuum extraction line employing ClF_3 (cf. Borth-

Table 2
$\delta^{18}O$ of oceanic crust xenoliths, core complex xenoliths and Miocene volcanics

Sample no.	Rock type	SiO_2	$\delta^{18}O$
Layer 3			
HAT B 9111	altered microgabbro	49.2	3.37 (wr)
HAT B 9125	fresh gabbro	49.0	4.38 (wr)
HAT B 914	fresh noritegabbro	51.2	5.1 (wr)
HAT B 912	altered dolerite	50.7	3.25 (wr)
Layer 2			
303905	very altered basalt	48.8	4.84 (wr)
HAT B 9110	altered basalt	49.2	8.62 (wr)
HAT B 9117	altered basalt	49.8	4.11 (wr)
HAT B 913	altered basalt	49.1	7.68 (wr)
B 9117.2	altered basalt	49.8	4.19 (wr)
Layer 1			
HAT 91249	quartz-rich sediment	85.6	14.51 (wr)
HAT 917 B	quartz-rich sediment	87.2	15.48 (wr)
HAT 917 C	quartz-rich sediment	86.7	14.37 (wr)
HAT 917 E	quartz-rich sediment	87.1	16.4 (wr)
HAT 917 D	quartz-rich sediment	87.3	14.14 (wr)
Core complex fragments			
HAT 919 C1	thermally altered trachyte	59.3	6.47 (wr)
HAT 919 C3	altered lava xenolith	61.9	4.33 (wr)
A-Pl-1-fsp	syenite xenolith in ign A	59.6	6.81 (fsp)
A-Pl-11-fsp	syenite xenolith in ign A	58.3	6.91 (fsp)
A-Pl-16-fsp	syenite xenolith in ign A	58.3	6.81 (fsp)
Mogan Group fsp separates			
T3-PR-fsp	T3 muagerite/hawaiite	58.0	6.67 (fsp)
X-Rb-Btau	ign X rhyolite	68.6	6.34 (fsp)
T4-PR	T4 alkali basalt	43.6	6.78 (fsp)
A-F6-fsp	ign A-trachyte fiamme	67.0	6.78 (fsp)
D-TF-1-fsp	ign D-trachyte fiamme	65.5	6.74 (fsp)
E-14m-F1	ign E-trachyte fiamme	63.6	6.74 (fsp)
F-I-F1-Bto	ign F-felsic fiamme	65.4	6.84 (fsp)
F-TF7-fsp	ign F-trachyte fiamme	64.7	6.48 (fsp)

ign = ignimbrite; wr = whole rock; fsp = feldspar separate.

wick and Harmon, 1982; Harris et al., 2000). The analytical error is estimated to be $< \pm 0.15$‰. Repeated analyses of in-house standards overlapped well within the given analytical error. All values are given relative to standard mean ocean water (V-SMOW).

5. Results

$\delta^{18}O$ values range from 3.3% to 5.1‰ for the gabbros, from 4.1% to 8.6‰ for the lavas and dykes, and from 14.1% to 16.4‰ for the siliciclastic sediments (Tables 1 and 2; Fig. 5). The gabbros have $\delta^{18}O$ values lower than fresh MORB ($\delta^{18}O$ 5.7–6.0‰; Eiler et al., 1996), suggesting high-temperature alteration and/or metamorphism as the dominant process in layer 3, lowering the $\delta^{18}O$ values due to oxygen exchange with heated seawater (e.g., Muehlenbachs, 1986). The lavas and dykes have $\delta^{18}O$ values both higher and lower than MORB, indicating both high and low-temperature alteration as important processes in layer 2. This is likely to reflect the high-temperature regime associated with dykes and their emplacement, and additional near-surface low-temperature alteration of lavas that enriches ^{18}O in the rocks upon weathering and oxygen exchange with cold seawater (cf. Alt et al., 1986). Siliciclastic sediment samples, in turn, are all above 14‰, indicating low-temperature oxygen exchange with seawater. The terrigenous sediments in this study, however, have $\delta^{18}O$ values somewhat lower than those typical for marine sediments (about 26‰; Muehlenbachs, 1998). The sedimentary xenoliths clearly interacted with the host lavas and contain up to 35 vol.% of interstitial mafic glass. The mafic glass could not be entirely separated from the sedimentary matrix, suggesting that the pre-emplacement $\delta^{18}O$ values were probably as high as those reported by, e.g., Muehlenbachs (1998). There is nevertheless a broad positive correlation between $\delta^{18}O$ and $^{87}Sr/^{86}Sr$ ratios for all investigated oceanic crust samples, the latter ranging from 0.70286 to 0.70524 for layer 2 and 3 rocks, and from 0.70929 to 0.72361 for the sediments (Hoernle, 1998) (Table 2), confirming the important role of alteration processes.

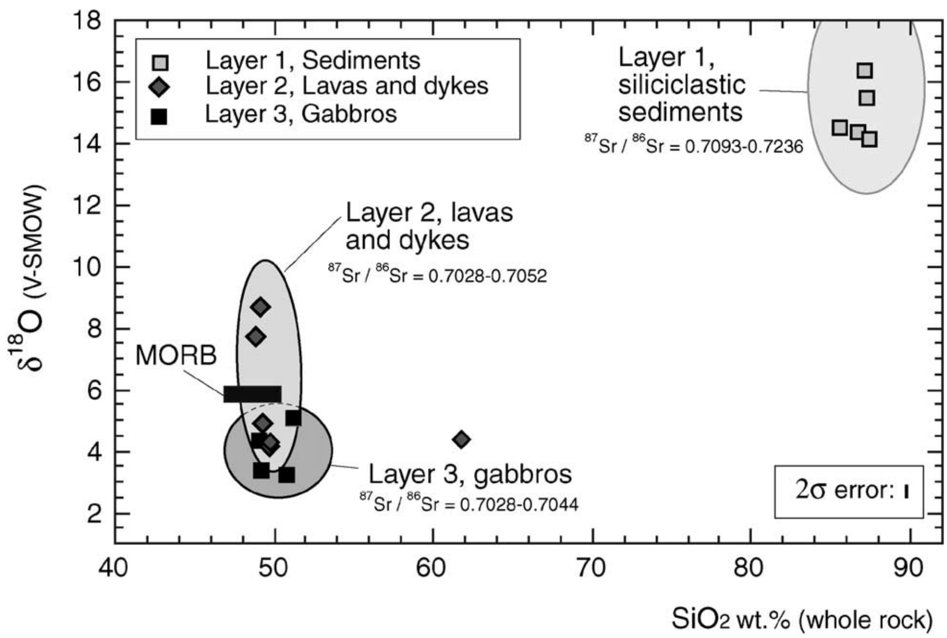

Fig. 5. Whole rock SiO_2 versus $\delta^{18}O$ for ocean crust xenoliths. Oceanic layer 3 is characterized by low SiO_2 and low $\delta^{18}O$, whereas layer 2 shows a similar spread in SiO_2 but ranges in $\delta^{18}O$ up to 8.6‰. Layer 1 sediments have the highest SiO_2 and $\delta^{18}O$ values. This trend broadly correlates with the Sr isotope data of Hoernle (1998).

Carbonate-rich sediments from the Canary Islands have even higher $\delta^{18}O$ values reaching 23–27‰ (Thirlwall et al., 1997; K. Hoernle, personal communication), thus representing the most extreme oxygen isotope end-member for the east Atlantic.

Fragments of the Gran Canaria core complex, such as syenite xenoliths from ignimbrite "A" and thermally altered felsic lava xenoliths in Quaternary basanite fallout deposits, have $\delta^{18}O$ values of 4.3–7.5‰ (Table 2). In comparison, feldspar separates from Mogan group lavas and ignimbrites have $\delta^{18}O$ values of 6.0–7.3‰, while hydrothermally altered caldera-margin tuffs have values up to 16‰ (Troll and Schmincke, 2002).

6. Discussion

6.1. Oxygen isotope stratigraphy of the oceanic crust

The mafic xenoliths in this study were previously interpreted as oceanic crust fragments, on the basis of characteristic MORB compositions and petrography, and correlate with the various layers of the oceanic crust, i.e., gabbros (layer 3), lavas and dykes (layer 2) and sediments (layer 1) (Schmincke et al., 1998; Hoernle, 1998). The $\delta^{18}O$ values of the igneous and meta-igneous samples reflect their inferred positions within this crustal profile (Fig. 6). Our results compare well with the oxygen isotope stratigraphy obtained from ophiolite sequences in, e.g., Chile, Oman and Macquarie Island (Stern et al., 1976; Gregory and Taylor, 1981; Cocker et al., 1982; Muehlenbachs, 1986). Oceanic crust samples from various DPSP and ODP cruises have oxygen isotope compositions (Muehlenbachs, 1986; Alt et al., 1986) overlapping with our data, suggesting that our oceanic layer 2 and 3 rocks represent typical igneous oceanic crust. Interestingly, the contrast in $\delta^{18}O$ between the gabbros and lavas is also characteristic for the oceanic crust close to mid-oceanic ridges. The oxygen isotope compositions of rocks from the oceanic crust in general seem to be controlled by their stratigraphic position, and only to a smaller extent by the local thermal regime. A detailed study designed to average the composition of oceanic layer 3, which was recovered from DSDP hole 735B (southwest Indian Ridge), yielded an average $\delta^{18}O$ of 4.4 ± 1.0‰ (Muehlenbachs, 1998), virtually identical to our values. The inferred stratigraphy of the Jurassic oceanic crust

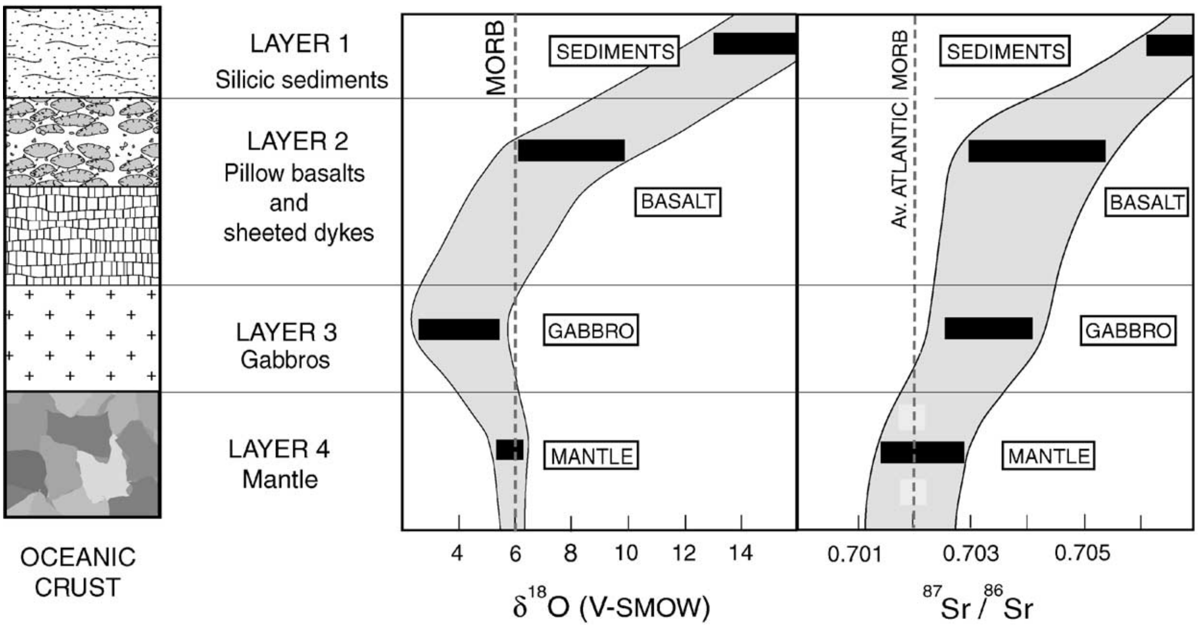

Fig. 6. Schematic column of the oceanic crust beneath Gran Canaria. To the right are the compositional ranges in $\delta^{18}O$ and $^{87}Sr/^{86}Sr$ as reconstructed from xenoliths. Sr isotope data are from Hoernle (1998).

beneath Gran Canaria thus closely resembles that of both younger and older oceanic crust worldwide.

6.2. Crustal contribution to ocean island basalts

Barometry using magmatic fluid inclusions in olivine from Miocene subaerial shield stage lavas and submarine hyaloclastites (ODP-Leg 157) revealed two main stagnation levels within the crust beneath Gran Canaria (Hansteen et al., 1998; Gurenko et al., 1998). Most inclusions reflect pressures of about 300 MPa (10 km), which is within the lower crust, while a second group of inclusions correspond to a pressure of about 110 MPa (3.5 km). The latter group probably represents a high-level reservoir and/or a level of neutral buoyancy. Seismic data from the periphery of Gran Canaria indicate a lithologic transition from Mesozoic oceanic crust to Mesozoic sedimentary rocks at about 10 km depth (Ye et al., 1999; Krastel and Schmincke, 2002), thus coinciding with the lower crustal fractionation level for the shield basalts.

Considering the oxygen isotope compositions of xenoliths from the oceanic crust, from the plutonic core complex of the island and from erupted mafic to felsic lavas, we present a model for crustal contamination beneath Gran Canaria (Fig. 7).

Assuming the observed isotopic variations to occur on various scales within the pre-island crust, our data can be used to explain the broad $\delta^{18}O$ range of 5.2–6.8‰ found for the Miocene shield basalts on Gran Canaria (e.g., Thirlwall et al., 1997). Values lower than MORB (5.7–6.0‰) may be the result of assimilation of metamorphosed and altered low $\delta^{18}O$ oceanic crust. In the SiO_2–$\delta^{18}O$ diagram (Fig. 7), the shield basalts appear to be influenced dominantly by oceanic layer 2 and 3 rocks, but some sediment influence for the high $\delta^{18}O$ sample cannot be excluded. Extensive partial melting of the metamorphosed oceanic crust is indicated by a pronounced deviation of shield basalt $\delta^{18}O$ values toward the lower spectrum of layer 2 and layer 3 samples. Independently of whether we use a hypothetical

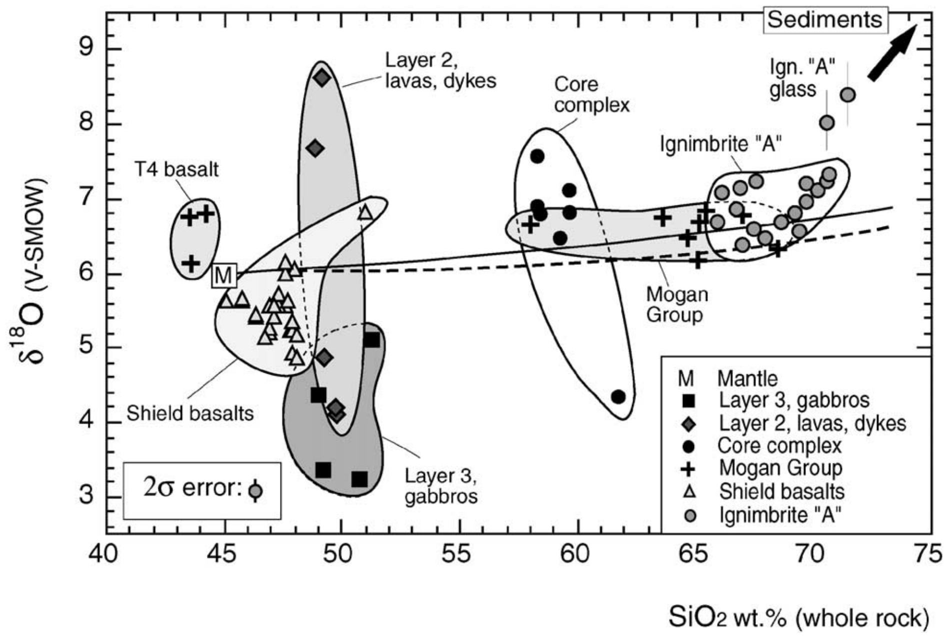

Fig. 7. Fractionation and assimilation trends for mafic and felsic volcanics from Gran Canaria depicted in a SiO_2–$\delta^{18}O$ diagram. Shield basalt data are from Thirlwall et al. (1997), and data from ignimbrite "A" and additional plutonic fragments are from Troll and Schmincke (in press). Note that ignimbrite "A" glass data were obtained by SIMS. Miocene felsic volcanics show generally a good correlation with closed system fractionation (solid line; Taylor and Sheppard, 1986), and compare well with other oceanic rock suites interpreted to have evolved as closed systems (dashed line from Ascension Island; Harris, 1983). Deviations from the fractionation trends are best explained by contamination of the magmas by various amounts of sediments and/or altered igneous rocks from the pre-island and/or island crust. See text for discussion.

mantle composition or a more evolved shield basalt composition, about 20–25% of bulk-rock assimilation of the lowest $\delta^{18}O$ gabbros (3.25‰) has to be assumed. Although the actual melting and assimilation processes within layers 2 and 3 are not well constrained for Gran Canaria, we find selective melting that preferentially affects rock volumes with high contents of fusible hydrous minerals (e.g., chlorite and actinolite) to be more plausible. Values as low as about 2.5‰ are common for hydrous medium-grade metamorphic minerals (cf. Alt et al., 1986; Muehlenbachs, 1998) and are also found in uplifted oceanic complexes in other Canary Islands (Javoy et al., 1986). Assuming preferential melting and assimilation of hydrous phases with very low $\delta^{18}O$ (~ 2–3‰), contamination of shield basalts would be on the order of 5–10% for the lowest $\delta^{18}O$ data of Thirlwall et al. (1997). These results implies that the low $\delta^{18}O$ values observed in some of the shield basalt samples may in fact reflect contamination processes rather than a low $\delta^{18}O$ mantle component.

The highest $\delta^{18}O$ values among the shield basalts can, in turn, be explained by assimilation of several percent of passive margin sediments, in accordance with the model of Thirlwall et al. (1997). We suggest that the most probable contaminant corresponds to the siliciclastic sediments investigated in this study. Similar sediments also occur as xenoliths in the submarine Hijo de Tenerife volcano ($\delta^{18}O$ 15.8–16.4‰; Hansteen, unpublished data), a 600-m-high cone in the channel between Gran Canaria and Tenerife (Schmincke and Graf, 2000), testifying to their regional occurrence.

Additional evidence for crustal contamination of the Gran Canaria shield basalts has been provided by variable sulfur ($\delta^{34}S$ − 1.0‰ to 8.5‰) and oxygen ($\delta^{18}O$ 5.0‰ to 7.8‰) isotope compositions of melt inclusions in clinopyroxene (Gurenko et al., 2001). Their preferred model is a less than 10% contamination of the magmas by hydrothermally altered basaltic crust and marine sediment, very much in accordance with our data.

6.3. Crustal contamination in Miocene felsic magmas from Gran Canaria

The subalkaline to peralkaline felsic volcanics that followed the shield stage on Gran Canaria (13.8–9 Ma) have $\delta^{18}O$ in feldspars ($\delta^{18}O_{fsp}$) ranging from 6‰ to 7.3‰ (cf. Cousens et al., 1990; Freundt and Schmincke, 1995; Troll and Schmincke, in press). Miocene felsic volcanics show a reasonably good correlation with closed-system equilibrium fractionation, and with oceanic rock suites that are interpreted to represent closed-system evolutionary trends (e.g., Ascension: Taylor and Sheppard, 1986; Harris, 1983; Sheppard and Harris, 1985) (Fig. 7). Although many of the feldspar data plot within, or close to, the range expected for fractional crystallisation, some highly evolved rhyolite end-members of the felsic ignimbrite succession show $\delta^{18}O_{fsp}$ values up to 7.3‰ and $\delta^{18}O_{glass}$ values of up to 8.5‰, not compatible with closed-system evolution (Troll and Schmincke, in press). In particular, $\delta^{18}O$ values of feldspar and glass from ignimbrite "A" rhyolite of the Upper Mogan Group (13.63 Ma) show a trend towards the sediment composition of the oceanic crust. Also, whole-rock samples of T4 alkali basalt (Fig. 2) are slightly elevated in $\delta^{18}O$ with respect to an assumed mantle composition. The simplest explanation for the spread in Mogan Group data is contamination with terrigenous sediment and rocks of oceanic layer 2 and/or the island's plutonic core. The diversion in $\delta^{18}O$ in ignimbrite "A" feldspar and glass is, for example, accompanied by a depletion of incompatible elements such as Zr and Nb and elevated $^{87}Sr/^{86}Sr$ (Troll, 2001), implying the input of a component that was high in $\delta^{18}O$ and $^{87}Sr/^{86}Sr$ but low in Zr and Nb. Oceanic sediments fulfill all these criteria and Troll and Schmincke (in press) concluded that an input of 8–12% of an oceanic sediment component is able to account for the elevated $\delta^{18}O$ values and the depletion in Zr and Nb in, e.g., the feldspars of the most evolved rhyolite end-member of this ignimbrite.

6.4. Contamination levels

Depending on the geometry and size of the magmatic transport and storage systems and the residence times of magmas, crustal contamination can occur in both the lower and the upper crust (Fig. 8). For example, felsic ignimbrite magmas that erupted during post-shield volcanism at Gran Canaria are generally accepted to have formed under low pressures at depths of 5–7 km (Crisp and Spera, 1987; Cousens et al., 1990; Freundt and Schmincke, 1995; Freundt-

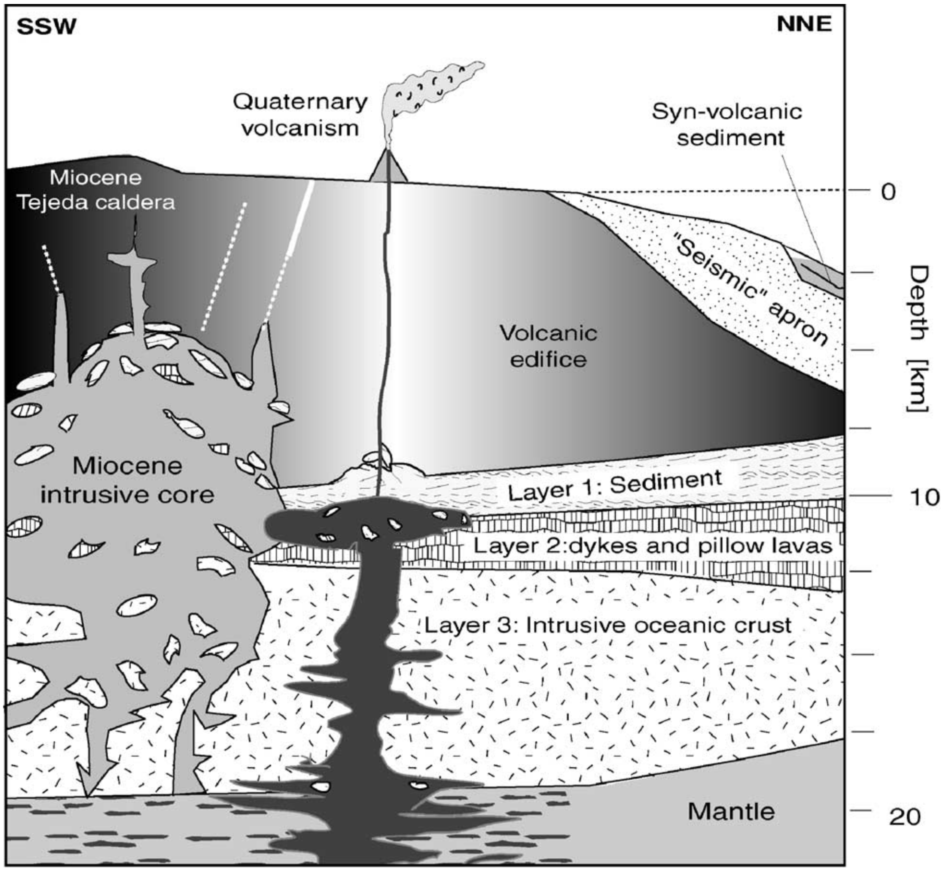

Fig. 8. Sketch depicting levels of possible crustal interaction between basaltic and felsic magmas and crustal rocks beneath Gran Canaria. Lithologic boundaries are after Ye et al. (1999) and Krastel and Schmincke (2002).

Malecha et al., 2001; Troll and Schmincke, in press). However, these rocks also show $\delta^{18}O$ evidence for sedimentary contamination with increasing degree of differentiation (Fig. 7), consistent with radiogenic isotope variations (Troll, 2001). We therefore invoke the availability of a sedimentary contaminant also at shallow level within the island's edifice. On the basis of the above-mentioned pressure determinations, we suggest that at least remnants of such sedimentary lithologies are present in the interior of the Gran Canaria edifice, possibly as uplifted chunks within the island itself. This reflects the possibility of an uplifted oceanic complex within the island's interior as observed on Lanzarote, La Palma and some of the Cape Verde islands (Staudigel and Schmincke, 1984; Stillman, 1985; Gerlach et al., 1988). However, alternative mechanisms such as giant phreatomagmatic explosions were also considered as a mechanism to distribute oceanic crustal fragments within the island's core (Schmincke et al., 1998).

Based on oxygen isotope measurements, high-level magma contamination has also been proposed, e.g., the Puu Oo eruption on Hawaii (Garcia et al., 1998). $\delta^{18}O$ values of 4.3–7.3‰ obtained for altered felsic lava and plutonic xenoliths from Gran Canaria fall within the range (−1.4‰ to +11‰) found for the uplifted plutonic complexes on the islands La Palma, Gomera and Fuerteventura (Canary Islands) (Javoy et al., 1986). Thus, high-level contamination appears very likely for the felsic magmas of Gran Canaria and additional high-level contamination for shield basalts can also not be excluded. Our preferred model is that the core complex beneath Gran Canaria is a mixture of layer 1, 2 and 3 rocks of the old oceanic crust and mafic to felsic intrusions of OIB-derived magmas (Fig. 8), which is supported by petrologic and geophysical data (Crisp and Spera, 1987; Freundt and Schmincke, 1995, Ye et al., 1999; Krastel and Schmicke, 2002; Freundt-Malecha et al., 2001; Troll and Schmincke, in press). The plutonic island core is thus the most probable region for crustal contamination of mafic through felsic magmas beneath Gran Canaria and most probably also in other Atlantic ocean islands along the passive margin of the African continent.

7. Conclusions

- Remnants of pre-island oceanic crust still exist beneath Gran Canaria and exhibit an oxygen isotope stratigraphy closely resembling those of both young and old oceanic crust and ophiolites.
- End-members for crustal contamination of mafic and felsic lavas at Gran Canaria comprise both terrigenous sediments and thermally altered oceanic crust, making it difficult to clearly identify a low $\delta^{18}O$ mantle reservoir.
- Contamination of magmas at Gran Canaria probably occurs within the plutonic core complex, which is likely to contain remnants of pre-island oceanic crust.

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