# Unravelling Temporal Geochemical Changes in the Miocene Mogan and Fataga group ignimbrite succession on Gran Canaria, Canary Islands, Spain

Peter A. Nicholls

Gran Canaria hosts a long-lived Miocene volcanic caldera system, the Tejeda Caldera, which produced an extensive succession of silicic ignimbrites and lavas (the Miocene Mogan and Fataga formations, > 1400 km$^3$) between 14 and 12 Ma. The ignimbrite succession exhibits temporal geochemical changes in radiogenic isotopes that were first examined by Cousens et al (1990). These changes were originally interpreted to be caused by changing mantle source compositions that fed the Tejeda volcanic system.

Recent work on large silicic caldera systems like Yellowstone employs a record of oxygen isotopes in the erupted products and postulates the progressive assimilation of crustal or older volcanic material into the successive magmas that fed the eminently large scale silicic eruptions (Bindeman et al 2001). This new theory has put forward the idea that crustal assimilation may be more important in such long-lived caldera systems than previously recognised.

The transition from the dominantly trachytic to rhyolitic Mogan group (14 – 13.2 Ma) to the dominantly trachytic to phonolitic Fataga group ignimbrites and lavas (ca 13.2 – 8 Ma) was sampled for oxygen isotopes to test whether temporal changes in the Mogan and Fataga group magmas were caused by crustal assimilation processes or indeed by temporal changes in mantle source composition (e.g. Cousens et al 1990).

Oxygen isotopes show gradually decreasing values up section between 14 and 12 Ma, which confirms that considerable changes occurred in the magmatic supply system through time. Progressively lower $\delta^{18}O$ values upsection are consistent with a low $\delta^{18}O$ component being added during the magma differentiation but we note that the Strontium isotope ratios decrease at around the same time, while Pb isotopes show correlated excursions too. These observations appear to point towards a mantle source change from the Mogan to Fataga group volcanism, the latter with the lower average $\delta^{18}O$ signature. This scenario is more plausible than crustal contamination as we would then expect an increase in strontium isotope ratios rather than an overall decrease. Crustal contamination is nevertheless detected to have played a role during the evolution of the Mogan group, where several units record additional crustal involvement. The overall oxygen-strontium isotope correlations however suggests a change in mantle source and requires change from a mixed source involving EM1, DM and HIMU to a higher proportion of the HIMU component towards the end of Miocene activity. This change was likely a function of the gradual waning of the DM and EM1-like components in the mantle supply column towards end of the Miocene volcanic episode on Gran Canaria.

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**Abstract**

Gran Canaria hosts a long-lived Miocene volcanic caldera system, the Tejeda Caldera, which produced an extensive succession of silicic ignimbrites and lavas (the Miocene Mogan and Fataga formations, > 1400 $km^3$) between 14 and 12 Ma. The ignimbrite succession exhibits temporal geochemical changes in radiogenic isotopes that were first examined by Cousens et al (1990). These changes were originally interpreted to be caused by changing mantle source compositions that fed the Tejeda volcanic system.

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The transition from the dominantly trachytic to rhyolitic Mogan group (14 – 13.2 Ma) to the dominantly trachytic to phonolitic Fataga group ignimbrites and lavas (ca 13.2 – 8 Ma) was sampled for oxygen isotopes to test whether temporal changes in the Mogan and Fataga group magmas were caused by crustal assimilation processes or indeed by temporal changes in mantle source composition (e.g. Cousens et al 1990).

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This change was likely a function of the gradual waning of the DM and EM1-like components in the mantle supply column towards end of the Miocene volcanic episode on Gran Canaria.

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## 1. Introduction

Gran Canaria hosts a long lived Miocene volcanic system that comprises shield basaltic lavas at the base that is overlain by an extensive succession of Miocene silicic ignimbrites and lavas (e.g. Schmincke 1969a,b, Crisp and Spera 1987, Clark and Spera 1990, Cousens et al 1990, Freundt and Schmincke 1992, 1995, Sumita and Schmincke 1998, Hansteen and Troll 2003, Troll and Schmincke 2002). The Mogan group was erupted between 14 to 13 Ma (Bogaard et al 1998) and forms the lowest part of the Miocene silicic succession on Gran Canaria. It ranges in composition from trachytes to rhyolites. The Fataga group has a trachytic to phonolitic composition and dates from ca. 12.9 to 9.2 Ma (Bogaard et al 1998). These two formations have a combined volume of ≥ 1400 $km^3$ (Sumita and Schmincke 1998), thus making Gran Canaria the ocean island with the largest volume of felsic eruptive deposits worldwide.

Temporal geochemical variations throughout this succession have been investigated previously (e.g. Crisp and Spera 1989, Cousens et al 1990, Hoernle et al 1991, Hoernle 1998) but as yet, no systematic oxygen isotope study is available for these rocks and consequently the question as to the reason for these variations remains (e.g. Hoernle 1998). Here we test, using oxygen isotopes in conjunction with published Sr-Nd-Pb isotope data, for the potential causes of the observed variations and three possible outcomes are considered: (i) no change is observed in the oxygen isotopes, meaning that crustal contamination has a negligible role and that the oxygen isotope composition of the mantle source(s) has not changed through time, (ii) oxygen ratios become higher upsection, indicating an increasing role of crustal contamination from e.g. hydrothermally altered tuffs and intrusions (Donoghue et al 2008, 2010), and (iii) oxygen isotope ratios become lower upsection, indicating mantle source changes in respect to oxygen isotopes, which would then confirm the older proposal of source variations that was based on a systematic change of Sr and Pb isotope ratios.

## 2. Geological Background

*2.1 Origin of Magmatism in the Canary Islands*

The Canary archipelago lies in the eastern central Atlantic, ~200 km off the coast of Morocco, and is made up of seven volcanic islands of different ages that are likely related to a hotspot (Fig 1) (e.g. Anguita and Hernan 1975, Schmincke 1976, 1982, Hoernle et al 1991, Hoernle and Schmin ke 1993, Hoernle et al 1995, Thirlwall 1997, Carrecedo et al 1998, Guren o et al 2001, 2006, 2010). The islands themselves have three main classes of volcanic rocks: (i) Mafic lavas ranging from subalkaline basalts to silica undersaturated nephelinites, (ii) felsic rocks which range from trachytes to varieties of rhyolites and phonolites, and (iii) plutonic rocks which include layered gabbro to alkali gabbro and

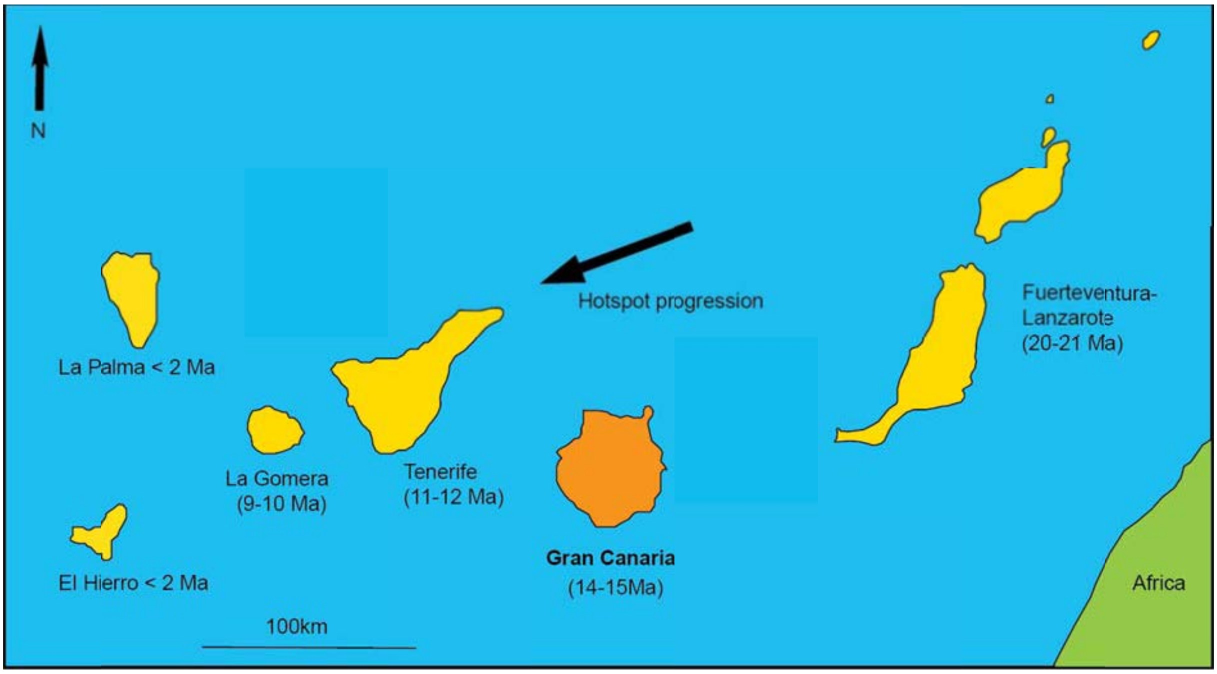


Figure 1 – Map of the Canary archipelago showing location of Gran Canaria and relative ages of islands (after Carrecedo et al 2001)

syenite intrusions (e.g. Schmincke 1976, Chayes 1977, Crisp and Spera 1987, Cousens et al 1990, Arana et al 1994, Ablay et al 1998, Carrecedo et al 2001, Freundt-Malecha et al 2001, Bryan et al 2002,Troll et al 2002, Troll and Schmincke 2002, Wiesmaier et al 2013) . In respect to volume, the most widespread magma type is the mafic lavas, while the highly evolved magmas make up only a small percentage of the island's volume and intermediate magma types are rarely present (e.g. Schmincke 1976, Freundt-Malecha et al 2001). A possible reason for the absence of widespread intermediate rocks is that they crystallise at depth and tend not to be erupted (Freundt-Malecha et al 2001).

The primary agent of melt generation beneath the Canaries is likely a mantle plume (e.g. Anguita and Hernan 1975, Schmincke 1976, 1982, Hoernle et al 1991, Hoernle and Schmincke 1993, Thirlwall 1997, Carrecedo et al 1998, Gurenko et al 2001, 2006, 2010, Montelli et al 2004). However, the

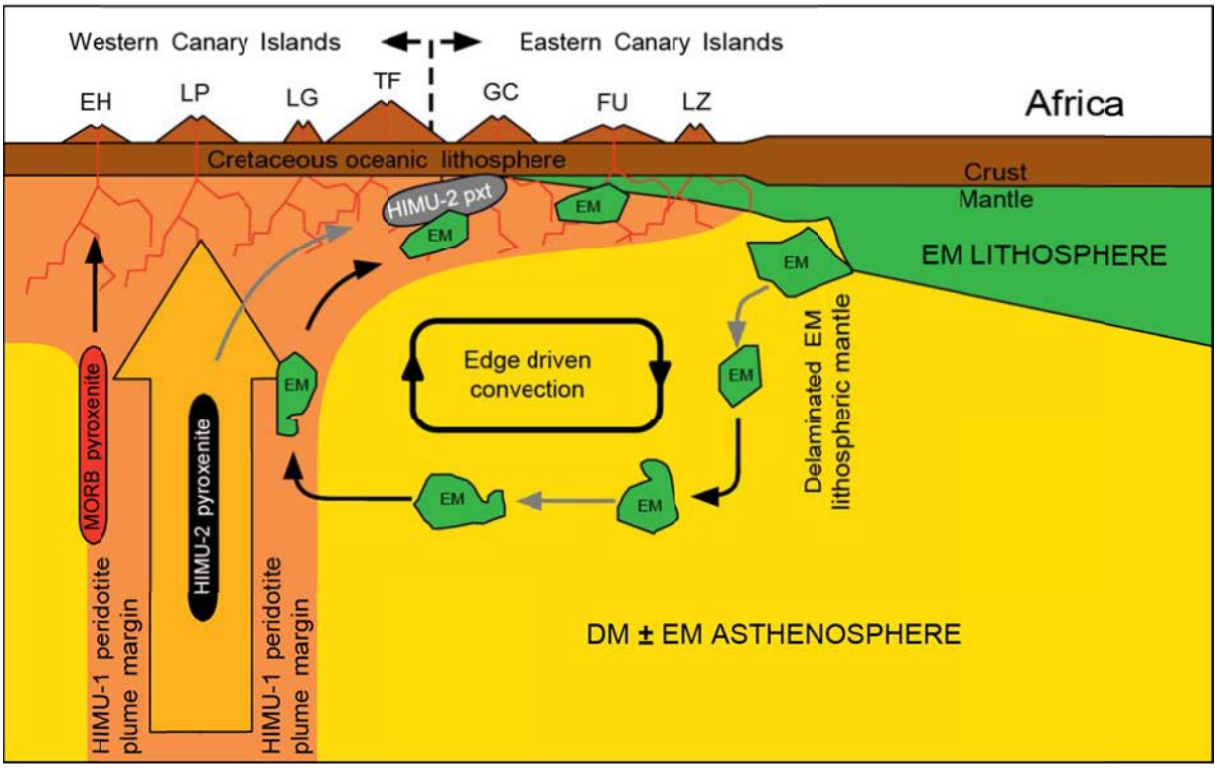


Figure 2 – Diagram of possible antle components feeding Canarian volcanism. Redrawn from Gurenko et al 2010.

Canary mantle plume has been suggested to entrain and sample different parts of the lithosphere and asthenosphere over the course of its evolution and a series of studi s have identified a range mantle domains that likely contributed to the Canary magmatism (e.g. Hoernle et al. 1991, Hoernle et al 1995, Thirlwall et al 1997, 2000, Widom et al 1999 Abratis et al 2002, De eny et al 2004, Geldmacher et al 2005, Gurenko et al 2006, 2009, 2010, Deegan et al 2012). Three distinct mantle source components have most recently been identified and are discussed in detail in e.g. Gurenko et al (2010) and Deegan et al 2012 (Fig 2). The first is a derivative of the HIMU component and is believed to be generated by the recycling of formerly subducted oceanic crust (e.g. Zindler and Hart 1986, Harmon and Hoefs 1995, Stracke et al 2003, Gurenko et al 2006, 2009, 2010, Deegan et al 2012). There is probably a HIMU-type pyroxenite component that originates in the centre of the plume and a HIMU-type peridotite component which originates from the margins of the plume (Gurenko et al 2010). In addition to the HIMU-type components, there is also a depleted MORB-type pyroxenite component that is derived from the upper mantle (DM) and entrained within the plume (e.g. Deegan et al 2012). Finally there is an EM-type peridotite component that likely comes from recycled continental lithosphere adjacent to the African continent (e.g. Gurenko et al 2010) and which is still considered influential on Gran Canaria, but seems to not reach as far as Tenerife (Gurenko et al 2009, 2010; Deegan et al 2012).

From the identification of these mantle components models of the plume beneath the Canary Islands have been constructed (e.g. Hoernle and Schmincke 1993, Gurenko et 2010). A HIMU-type plume entrains EM1 and DM components during ascent where entrainment of EM components can be accomplished in a number of ways. Thermal erosion of the lithosphere beneath the African continent or wholesale detachment of the lithospheric mantle could have distributed subcontinental lithosphere via edge-driven convection during the break up of Pangea (Hoernle 1991, 1995, 1998, Gurenko et al 2009, 2010).

The Canarian magma compositions may not be exclusively down to the mantle processes, however, and higher level contamination by crustal components may also play a part. Studies employing radiogenic and stable isotopes have revealed contamination of Canarian basaltic and evolved magmas by oceanic sediments and altered oceanic crust and/or by self-recycling of the volcanic edifices (e.g. Thirlwall et al 1997, Wolf et al 2000 Gurenko et al 2001, Hansteen and Troll 2003; Troll and Schmincke 2002, Deegan et al 2012, Wiesmaier et al 2013).

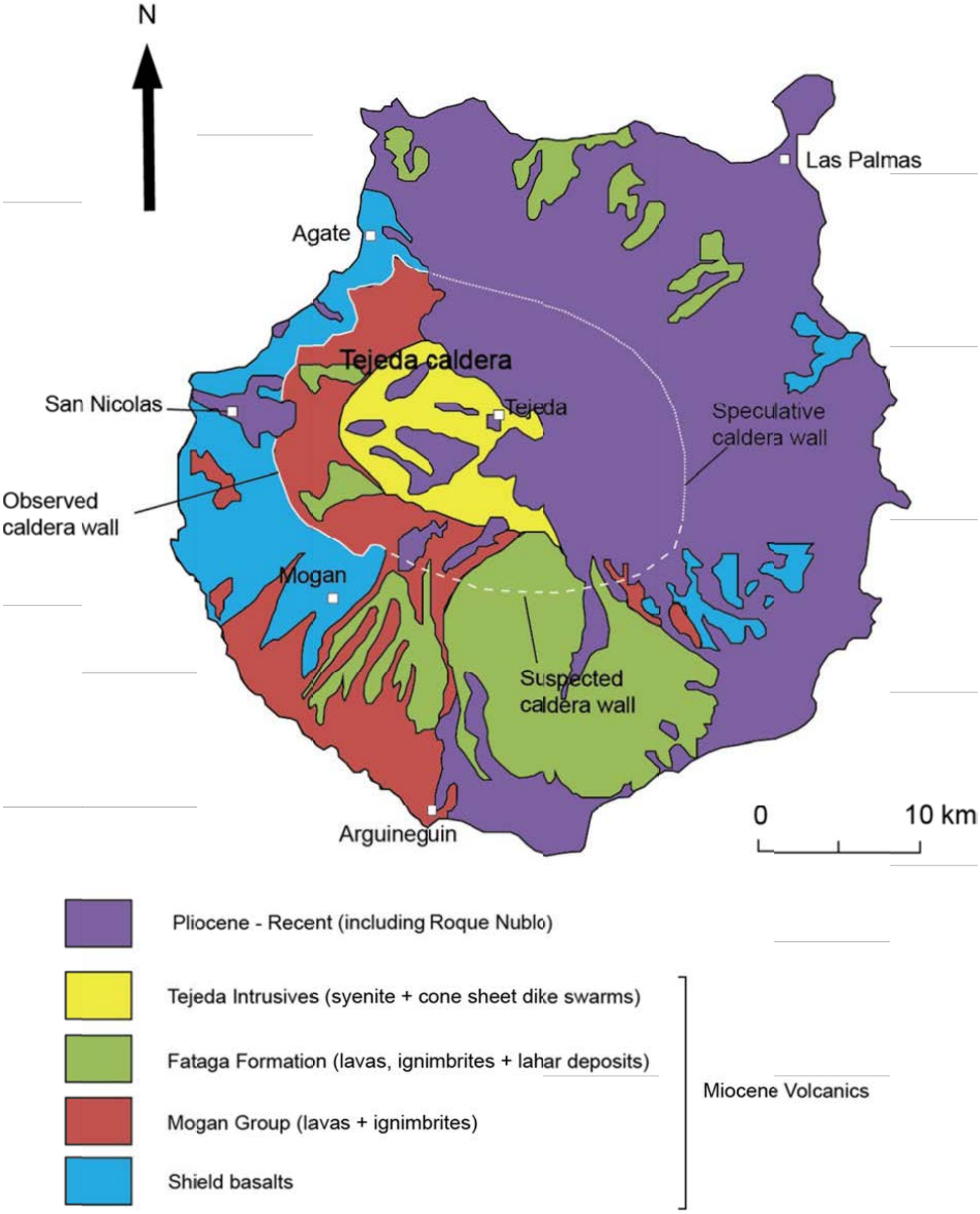


Figure 3 – Map of Gran Canaria volcanics redrawn from Cousens et al (1990) and Schmincke and Sumita (2001)

*2.2. Island evolution and the geology of Gran Canaria*

The islands of the Canary archipelago are seen to go through a broad cycle of volcanic stages: the first is the "Shield stage" with the effusive eruption of picritic, ankaramitic, and basanitic lavas that starts off as a seamount dominated by pillow lavas, before moving onto a sub-aerial stage in which copious lavas are erupted either from a central shield or along a broad rift (Fig 1) (e.g. Thirwall et al 1997, Carrecedo et al 2001, Walter et al 2005, Longpre et al 2008). The islands of La Palma and El Hierro are currently observed to be in this stage (Carrecedo et al 2001). The shield stage is often punctuated by large sector collapses which have created broad submarine debris aprons all around the islands (e.g. Ancochea et al 1994, Carracedo 1994, Ablay and Hurliman 2000, Boulesteix et al 2013). The next stage, the "Erosional stage", is when the shield is completed and eruptive frequency declines. This stage may be characterized by large-scale explosive eruptions that produce ignimbrites associated with caldera collapse, like the silicic ignimbrites associated with the Las Canadas caldera on Tenerife (Byran et al 1998, Anochea et al 1999, Edgar et al 2002, Brown et al 2003, Brown and Branney 2004) or in fact the Mogan and Fataga groups on Gran Canaria (e.g. Schmincke 1969b, Crisp and Spera 1987, Clark and Spera 1990, Cousens et al 1990, Freundt and Schmincke 1992, 1995, Sumita and Schmincke 1998, Hansteen and Troll 2003, Troll and Schmincke 2002, Troll et al 2003). The "Post-erosional" stage that follows is frequently characterized by the construction of a large strato volcano such as Pico Teide on Tenerife or voluminous fissure eruptions such as those recorded on Lanzarote (Carracedo et al 1992, Anochea et al 1994, Ablay and Marti 2000, Carracedo et al 2007, Carracedo et al 2011, Wiesmaier et al 2013). On Gran Canaria, this stage is represented by the Pliocene Roque Nublo deposits that are derived from a now largely lost volcano in the centre of the island (Brey and Schmincke 1980, Cacho-Garcia 1994, Hoernle et al 1991, Perez-Torrado et al 1995). The final stage is is the "rejuvenation stage" that is expressed in the form of alkaline, small-volume volcanism and is represented on Gran Canaria by a series of young cinder cones and maars of dominantly basanite and nephelinte compositions (e.g. Rodriguez Gonzales et al 2009). These island evolutionary stages are often separated by long hiatuses which can last several millions of years (Carrecedo et al 2001, Anochea et al 2006, Cousens et al 1990, Crisp and Spera 1989).

The island of Gran Canaria has gone through all of these “stages” of volcanic activity, commencing at 14 -15 Ma ago and continuing into the Holocene (Fig 2) (e.g. MacDougall and Schmincke 1976, Hoernle et 1991, Thirlwall et al 1997 Bogaard et al 1998, Carracedo et al 1998, Hansteen and Troll 2002). The youngest eruptions on Gran Canaria are dated at 1500 – 3000 years BP (Rodriguez-Gonzalez et al 2009).

*2.2.1 The Miocene Stratigraphy of Gran Canaria*

Gran Canaria contains one of the broadest suite of volcanic rocks in the Canaries ranging from mafic nephelinites to rhyolites (e.g. Schmincke 1969a,b, Crisp and Spera 1989, Cousens et al 1990 Hoernle et al 1993a, Thirlwall et al 1997, Sumita and Schmincke 1998, Gurenko et al 2001, Troll and Schmincke 2002, Troll et al 2002 Hansteen and Troll 2003, Rodriguez-Gonzalez et al 2009). The Miocene period offers the greatest range of differentiated rocks and is characterized by an initial mafic phase which constructed the shield volcano, before commencement of a felsic phase that led to the formation of the Tejeda caldera and the deposition of two extensive pyroclastic groups, the Mogan and Fataga groups (e.g. Schmincke 1969a,b, Crisp and Spera 1989, Cousens et al 1990, Hoernle et al 1993a , Thirlwall et al 1997, Gurenko and Schmincke 1998, Sumita and Schmincke 1998, Freundt et al 2001, Gurenko et al 2001, Sumner and Branney 2001, Troll and Schmincke 2002, Troll et al 2002, Hansteen and Troll 2003, Troll et al 2003, Sumner and Wolff 2003, Jutzeler et al 2009).

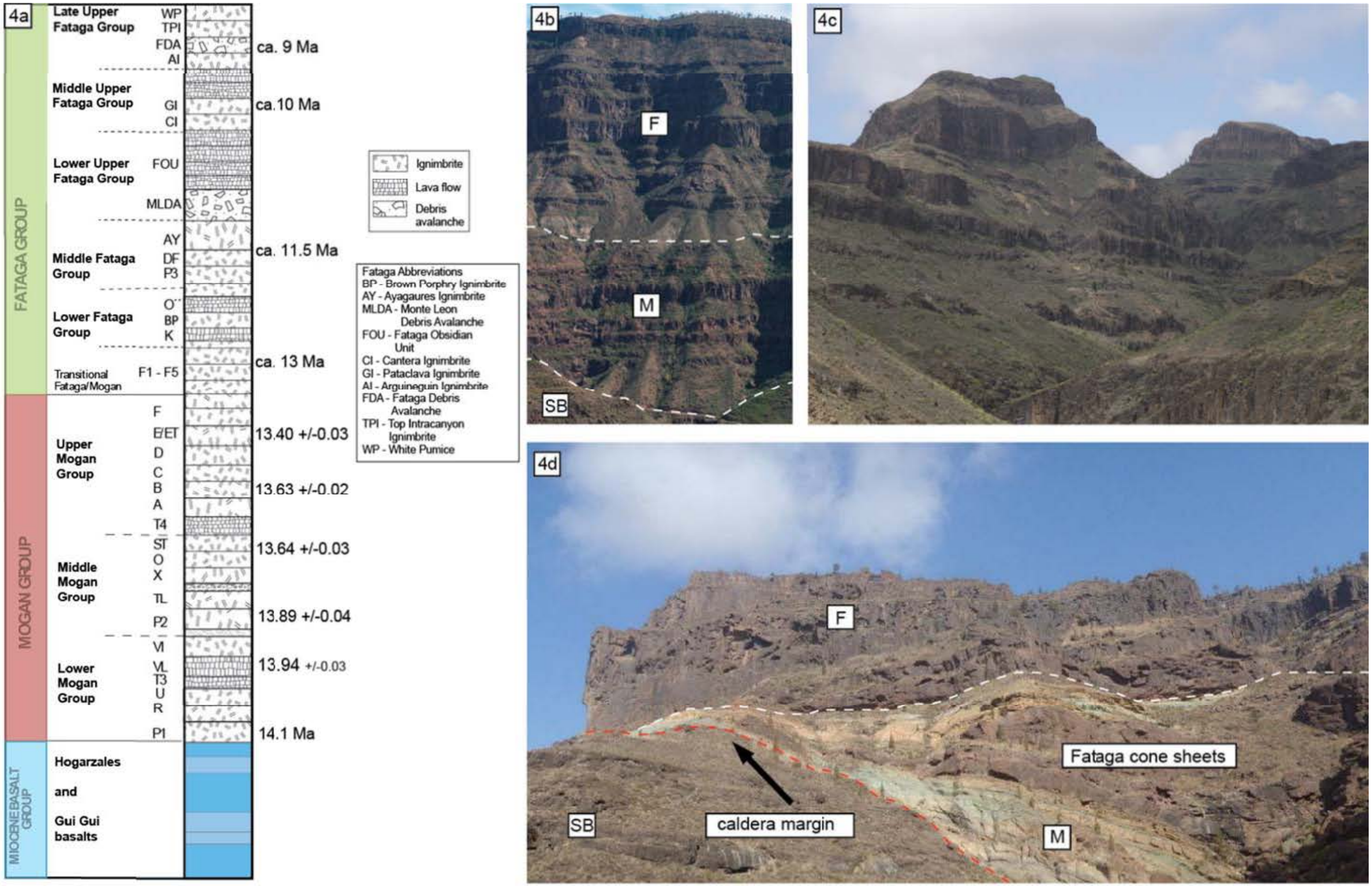


Figure 4 - a: Detailed stratigraphy of main Miocene units, redrawn and adapted from Sumita and Schmincke 1998. b Outcrop of Miocene units in Barranco de Arguineguin (SB = shield basalts, M = Mogan Group, F = Fataga). c: Fataga Group exposure at caldera margin near St. Lucia. d: Monte Horno formation at Los Azulejos, showing margin of Tejeda caldera with ponded Mogan ignimbrites against the caldera featuring later hydrothermal alteration which has affected the Mogan but not the overlying Fataga units.

*Shield basalts*

Initial activity was characterized by the construction of a seamount that eventually broke through the ocean surface to begin the construction of a shield volcano (Thirlwall et al 1997, Gurenko 2001, 2009, 2010). Tholeiitic and picritic basalts of the Guigi and Hogarzales formations were erupted and they together form about 80% of the subaerial edifice (Thirlwall et al 1997). Lava effusion at this time was fairly prodigious with at least 1000km$^3$ of lavas erupted occurring between 14.5 and 14 Ma. (MacDougall and Schmincke 1976, Thirlwall et al 1997, Gurenko and Schmincke 2000, Gurenko et al 2001, 2004, 2006, 2010, Herr et al 2002) . The evolution of the early basaltic Gran Canaria volcano was punctuated by periodic gravitational collapses which generated collapse scars and submarine debris fans (e.g. Funck and Schmincke 1998).

*Felsic caldera-forming Phase (Post-shield)*

Following these early mafic lava effusions, Gran Canaria entered a new stage at around 14 Ma, which was characterized by the formation of the 20 km diameter Tejeda caldera and the deposition of over 1400 km$^3$ of evolved felsic ignimbrites and lavas that represent the largest volume of silicic rocks on any ocean island worldwide (Sumita and Schmincke 1998). The eruptions during this period would have been highly explosive with the generation of pyroclastic density currents and plinian columns that would have covered virtually the entire island and would have resulted in the destruction of all life on the island on repeated occasions (e.g. Leat and Schmincke 1993, Freundt and Schmincke 1995, Kobberger and Schmincke 1999, Sumner and Branney 2001). The initial phase of this felsic activity lasted from ca. 14.1 to 13.4 Ma and is represented by the tracyhitic to rhyolitic Mogan group and is associated with the formation of the main caldera escarpment of the Tejeda caldera. The later phase (ca.13 – 9 Ma) is represented by the trachytic to phonolitic Fataga group that also erupted from within the former caldera enscarpment (Bogaard et al 1998, Jutzeler et al 2009, Donoghue et al 2010).

*The Mogan group*

During Mogan times (c.a. 14 to 13.2 Ma), pyroclastic density currents deposited 15 to 20 trachytic, comenditic, and pantelleritic ignimbrites. Rare mafic to intermediate flows also occur (Fig 4a, b). These ignimbrites are noted for their strongly rheomorphic and lava-like lithofacies, which indicates high temperature low viscosity pyroclastic “boiling over” type activity (Leat and Schmincke 1993, Kobberger and Schmincke 1999, Sumner and Branney 2001). The main ignimbrite units of the Mogan group are summed up in the stratigraphic column in Figure 4a. The Mogan group is in total over 400 metres thick and is divided into Lower, Middle and Upper domains (e.g. Schmincke 1969a,b, Crisp and Spera 1987, Clark and Spera 1990, Cousens et al 1990, Sumita and Schmincke 1998, Freundt et al 2001, Troll and Schmincke 2002). Moreover, the Mogan group is characterized by compositionally zoned ignimbrites (Crisp and Spera 1987, Cousens 1990, 1992, Sumita and Schmincke 1998) with usually a rhyolitic base grading into a trachytic top (Cousens et al 1990, Leat and Schmincke 1993, Kobberger and Schmincke 1999, Sumita and Schmincke 1998, Troll and Schminke 2002, Troll et al 2003).

The first ignimbrite of the Mogan group, ignimbrite P1, offers a good example of chemical zoning. P1 consists of a crystal-poor rhyolite at the base which grades upwards into sodic trachyandesites to sodium poor trachyandesites. Finally the ignimbrite is topped by two units of basalt which constitutes locally over 50% of the deposit (Freundt and Schmincke 1992, 1995a,b).

Lava flows are sporadically present in the Mogan group with the plagioclase-clinopyroxene mugearite lava of T3 separating ignimbrite units U and VI. Further up section in the Upper Mogan group, a Hawaiite lava flow can also be observed, it is named T4 and underlies ignimbrite A (Schmincke

1969b, Crisp and Spera 1987, Cousens et al 1990, Sumita and Schmincke 1998, Freundt and Schminke 1976).

The formation of the Tejeda caldera occurred concurrently with the deposition of the Mogan group (Freundt and Schmincke 1995, Sumita and Schmincke 1998, Troll et al 2002). Initial caldera subsidence likely took place at the time of the eruption of ignimbrite P1, with consequent eruptions triggering further episodes of caldera unrest and collapse (Freundt and Schmincke 1995, Sumita and Schmincke 1998, Troll et al 2002). The main caldera escarpment is visible as an uncomformity where Mogan age ignimbrites pinch out against older shield basalts, such as that seen at e.g. Monte Horno (Fig 4d) (Schmincke and Swanson 1966, Schmincke 1967). The caldera margin is surrounded with concentric peripheral faults that display periods of successive reactivation, indicating that caldera collapse happened in an incremental and probably piecemeal fashion (Troll et al 2002). Total subsidence of the caldera is estimated in the region of 1 km (Schmincke 1976).

*The Fataga group*

Following transitional units F1 to F5, the explosive activity on Gran Canaria continued with the deposition of the trachytic to phonolitic Fataga group (between 12.4 and 8.3 Ma) and marks the final period of extrusive Miocene volcanic activity on Gran Canaria (Crisp and Spera 1989, Clark and Spera 1990, Cousens et al 1990, Bogaard et al 1998, Sumita and Schmincke 1998). The trachyphonolitic Fataga group is interbedded with debris avalanche and lahar deposits (Fig 4c) (Crisp and Spera 1987, Cousens et al 1990, Sumita and Schmincke 1998) and unlike the Mogan group, the Fataga group is comparatively poorly studied. It is still unknown as to how many ignimbrites make up the group exactly, with current estimations between 8 and 12 (Fig 4b). The eruptive volume of the Fataga group is estimated to comprise c.a. 500 $km^3$ (Crisp and Spera 1987, Cousens et al 1990, 1992, Sumita and Schmincke 1998) which together with the volume of the Mogan group adds up to $\geq$ 1400 $km^3$.

From 12.3 to 7.3 Ma, intrusive activity appears prominent on the island with the emplacement of trachytic to phonolitic cone sheet dykes and syenite intrusions centred inside the Tejeda caldera. This intrusive activity is in part concurrent with the Fataga group, however, it does appear to continue after the Fataga eruptions have ceased (Schmincke 1967, Schirnick et al 1999). Concurrent with the Fataga magma chamber extensive hydrothermal alteration took place of both the Mogan ignimbrites and intrusions (Donoghue et al 2008, 2010).

*Pliocene Roque Nublo group*

Following the Fataga events a hiatus of over four million years is recorded, after which volcanic activity resumed on the island with the deposition of the Pliocene Roque Nublo group. The Roque Nublo volcanics take the form of alkaline lavas and predominantly non-welded ignimbrites, which display some features of phreatomagmatism possibly related to ground water interaction (Perez-Torrado et al 1995). During this time a large stratovolcano was built up within the Miocene caldera and this volcano subsequently decayed in a series of collapse events in a series of collapse events that generated the Roque Nublo debris avalanche deposits (Brey and Schmincke 1980, Cacho-Garcia 1994, Hoernle et al 1991). The close of the Roque Nublo stage of activity is marked by the extrusion of lava domes in the crescent left by these collapse events (Cacho-Garcia 1994).

*Recent Activity*

The volcanic system on Gran Canaria remains active, albeit at a much lower level than during previous eras. A suite of monogenetic cinder cones and associated lava flows erupted during the Holocene with the focus of activity predominantly in the Northeast of the island. The types of lavas include basanites, alkali basalts and tephrites (e.g. Hoernle et al 1991, Rodriguez-Gonzalez et al 2009). The most recent eruption on Gran Canaria took place ~1,500 yrs BP, represented by for example the Jinamar volcano and Montanion Negro cinder cone (Rodriguez-Gonzalez et al 2009).

## 3. Methods

### *3.1. Introduction to oxygen isotopes*

Isotopes of oxygen provide a powerful tool to elucidate the petrogenesis of igneous rocks and are widely used to chart geochemical changes in magmas erupted in many different settings (e.g. Muehlenbachs et al 1974, Taylor 1987, Woodhead et al 1987, 1993 Bindeman and Valley 2000, 2001, Eiler et al 1995, 1996a,1996b, 1997, Harmon and Hoefs 1995, Harris et al 2000, Bindeman and Valley 2000, 2001). Unlike radiogenic isotopes (such as strontium and lead) oxygen isotopes fractionate during crystal fractionation, e.g. as the melt evolves from a mafic to silicic composition. The full difference is for basalt to rhyolite ≤ 1‰ (Bindeman 2008) and if the magma fractionated in a closed system without external influences, then basalt lavas will have $\delta^{18}O$ values of between 5 and 6 ‰, and rhyolites will show values of between 6 and 7 ‰ (Ingerson 1953, Silverman 1957, Taylor 1980 Bindeman and Valley 2000). If magmas undergo open system differentiation instead then the isotope budget will be upset and should be distinct from mantle-like values and from the range of mantle-derived fractionation products. The value for a rhyolite would then be either much higher than 6 to7‰, if contaminated by e.g. high $\delta^{18}O$ sediments, or it could much be lower if contaminated with low $\delta^{18}O$ volcanic components (e.g. Taylor 1980, Harmon and Hoefs 1995, Bindeman and Valley 2000, 2001, Troll and Schmincke 2002, Troll and Hansteen 2003, Bindeman et al 2011, Deegan et al 2012).

In primitive mantle-derived rocks, in turn, oxygen isotope variations does also occur and may reflect deeper level changes in the mantle, and thus mantle heterogenity. As the mantle beneath ocean islands, including the Canary Islands, is presumed to be isotopically heterogeneous, changes in $\delta^{18}O$ may reflect the sampling of different mantle components by ascending plume materials through time (e.g. Harmon and Hoefs 1995, Lassiter and Hauri 1998, Kempton et al 2000, Farnetani et al 2001, Skovgaard et al 2001, Gurenko et al 2010). It is therefore instructive to compare oxygen isotope data of igneous rocks with results from radiogenic isotope analysis (e.g. lead, neodymium and strontium) in order to obtain a comprehensive picture of potential sources involved in the Miocene explosive volcanism of Gran Canaria (e.g. Zindler and Hart 1986, Harmon and Hoefs 1995, Smith et al 1996, Stracke et al 2003, Bhattacharya et al 2013, Deegan et al 2012, Wiesmaier et al 2013).

### *3.2 Analytical methods*

Samples were collected from Mogan and Fataga group ignimbrites and were then washed and cut to remove altered and weathered surfaces. Samples were then crushed in a jaw crusher to sand-sized fragments. Crystals were hand-picked using a binocular microscope and only pristine feldspar and biotite was selected. After picking, the mineral seperates were either powdered in an agate ball mill or were analysed as wholegrain crystal seperates.

The majority of oxygen isotope ratios were determined at Oregon State University using the hand - picked minerals. The $\delta^{18}O$ ratios for these samples were measured by laser fluorination, where samples are heated with an infrared (9.6 µm $CO_2$) laser in the presence of $BrF_5$ to release the oxygen (Bindeman et al 2008, Borisova et al 2013). Oxygen was then changed to $CO_2$ gas using hot graphite before being analysed with IRMS in dual inlet mode. The unknown oxygen isotopes were then analysed with garnet standards, which have a precision better than 0.1‰ (Bindeman et al 2008).

Several feldspar and biotite powders were prepared and analysed at the University of Cape Town (UCT), employing a conventional silicate line for oxygen separation and following the strategy outlined in Vennemann and Smith (1990), Fagereng et al (2008) and Harris and Vogeli (2010). The samples were reacted at 10k Pa with $BrF_5$ and the purified oxygen released was captured in a 5 Å molecular sieve inside a glass bottle. The oxygen isotope ratios were then measured directly from the gas on a Finnegan DeltaXP mass spectrometer using an internal reference standard.

The results were then reported using standard oxygen isotope notation i.e. $\delta = (R_{sample}/R_{standard} - 1) * 1000$, where $R = {}^{18}O/{}^{16}O$, relative to SMOW (Standard Mean Ocean Water) see e.g. Taylor (1980) for further details.

*3.3 Normalisation of Oxygen isotope values*

When elucidating the nature of oxygen isotope values in the Mogan and Fataga group and how they originated (i.e. closed vs open system behaviour) it is necessary to recalculate the oxygen isotopes firstly to the value of magma and then back to the primitive magmatic value prior to fractionation. The oxygen isotope compositions of the host magmas can be calculated from phenocrysts which permits an assessment of magmatic variability. For feldspar this is $\Delta$ fsp – melt = 0.2 and for pyroxene $\Delta$ px – melt = - 0.3 (e.g. Kyser et al 1981; Harris et al 2005). This melt correction is applied to all mineral oxygen isotope values in this study and yields an approximation of the original melt value.

The oxygen isotope values of the ignimbrites and lavas of the Mogan and Fataga groups used were all from feldspars. In addition $\delta^{18}O$ values for the shield stage basalts from Thirlwall et al 1997 were also employed for comparison, and those values come from pyroxenes.

Once melt values have been calculated, the oxygen isotopes must be corrected for fractional crystallization to approximate the original value of the primitive magma. Generally speaking, a magma experiences an approximate 0.2 per mil increase in $\delta^{18}O$ over 5% wt. increase in $SiO_2$. For the Gran Canaria magmas values of basalt from the Mogan group and Shield stages were averaged to a primitive silica value of 47.6 % to approximate the $\delta^{18}O$ of the original primitive magmas. This was then inserted into the following equation:

$$(X-47.63)/5*0.2$$

where X = the evolved silica composition of a particular sample. The equation yields the amount (per mil) of isotope increase between the sample's melt values and a primitive parental liquid prior to fractionation. To gain the pristine magma composition, the derived factor (X) is then subtracted from the $\delta^{18}O$ melt value of a selected sample (c.f. Taylor 1986, Bindeman 2008).

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Although these normalised values represent the values before fractional crystallization, they still will reflect changes caused by e.g. crustal contamination. Contamination processes may be detected through non-systematic variations of the normalized values when compared with the accepted mantle range. Alternatively changes in mantle supply should yield systematic and gradual variations of the normalized values through time.

## 4 Results

### *4.1 General trends of Gran Canaria oxygen isotopes*

To elucidate on the general trends of the oxygen isotopes for the Miocene and Pliocene igneous rocks, the values were plotted together with hydrothermally altered tuffs and intrusive rocks from Gran Canaria (Donoghue et al 2008, 2010) to give a fuller spectrum of oxygen isotopes values available from Gran Canaria. These values were plotted against $SiO_2$ weight percent for each of the different units (Fig 5).

At the time of the eruption of the shield lavas and their olivine mantle xenoliths, oxygen isotope values were 5 to 6 ‰ with a few higher outliers (Thirlwall et al 1997). The $SiO_2$ content increased as the Mogan group was erupted and so did the oxygen isotopes, falling in the range of 6 to 7 ‰ with some lower values present (Fig 5) (Cousens et al 1992, Freundt and Schmincke 1995, Hansteen and Troll 2003, Troll et al 2003, Troll and Schmincke 2000). However, the younger Fataga group oxygen isotope values are seen to take a noted downwards shift, moving from 5.8 ‰ to 5.2‰, with a greater similarity to the shield basalt values than to the underlying Mogan group (Fig 5).

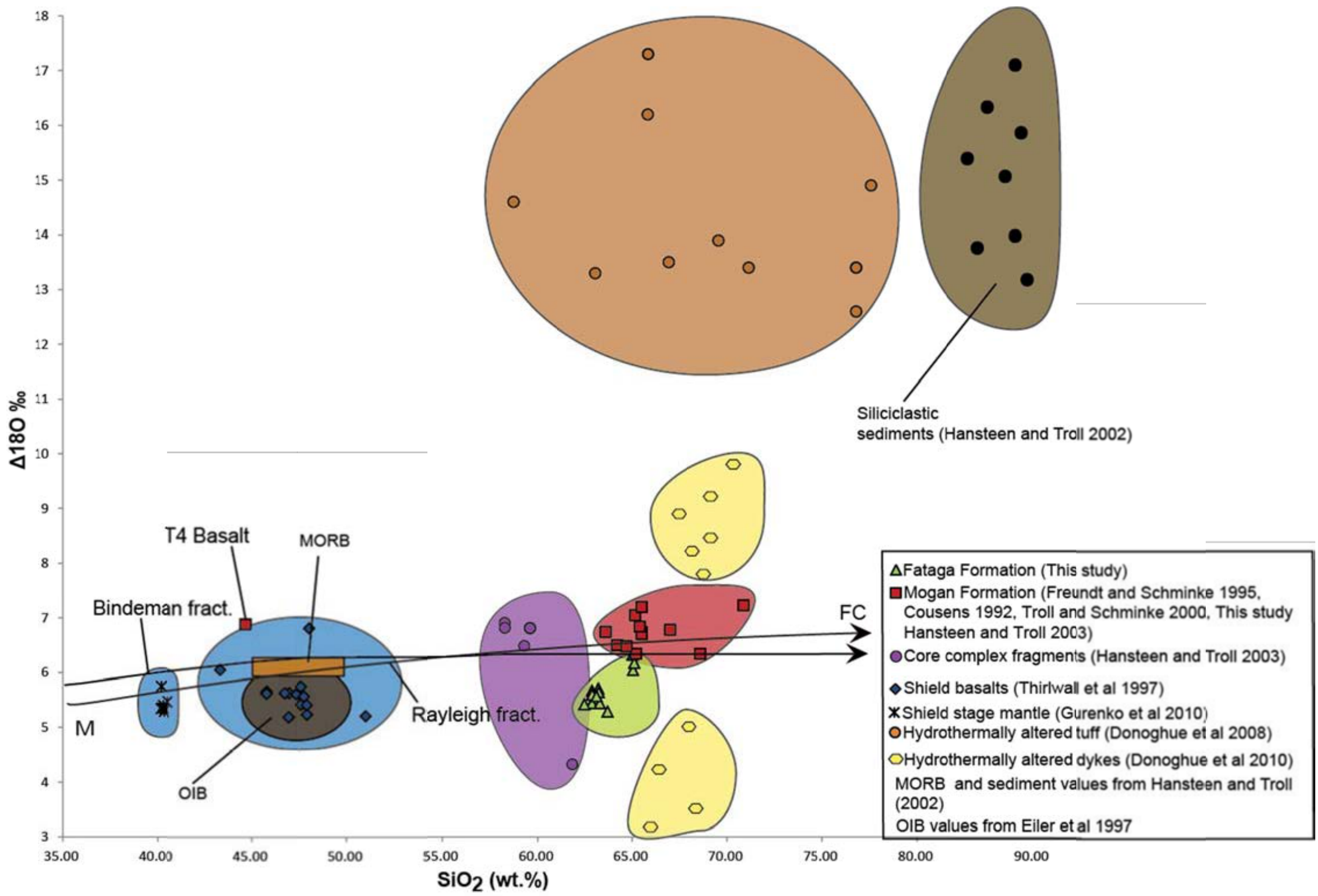



Figure 5 – Gran Canaria oxygen isotopes plotted against silica values.

Two closed system fractional crystallization curves have also been added to the diagram (Taylor and Shephard 1986, Bindeman et al 2008) to show the magmatic progression in oxygen isotopes from an original mantle source starting component. Interestingly, very few points in any of the three main units lie on these lines. Values plot either above (in the case of the Mogan group) or below (in the case of Shield basalts and the Fataga group) these reference curves (Fig 5).

To provide further reference, the common oxygen isotopes values for MORB and OIB were added (Eiler et al 1997, Hansteen and Troll 2003). None of the Gran Canaria oxygen isotopes fall directly within the MORB field (Fig 5b), but notably the values for the shield basalts fall within the OIB domain. Values for marine silciclastic sediments on Gran Canaria were also added (Hansteen and Troll 2003) and it can be seen that there is a trend towards these sediment values for both the Mogan and the hydrothermally altered tuffs and intrusions (Fig 5), but this appears to discontinue with the start of Fataga volcanism.

Following the oxygen vs silica plots, oxygen isotope values for the Mogan, Fataga and Roque Nublo stages were tabulated and then corrected to magma compositions when appropriate (eg. from mineral samples – see methods for exact procedure). Following this step, the data were corrected for fractional crystallization to derive the approximate parental magma composition for the sample. This data is summed up in Figure 5 and 6 and Tables 1.1, 1.2 and 1.3. In addition, oxygen isotopes data are compared with available Sr and Pb isotopes (Fig 6). For the sake of space, values that overlap from individual ignimbrite values have been omitted from the plots.

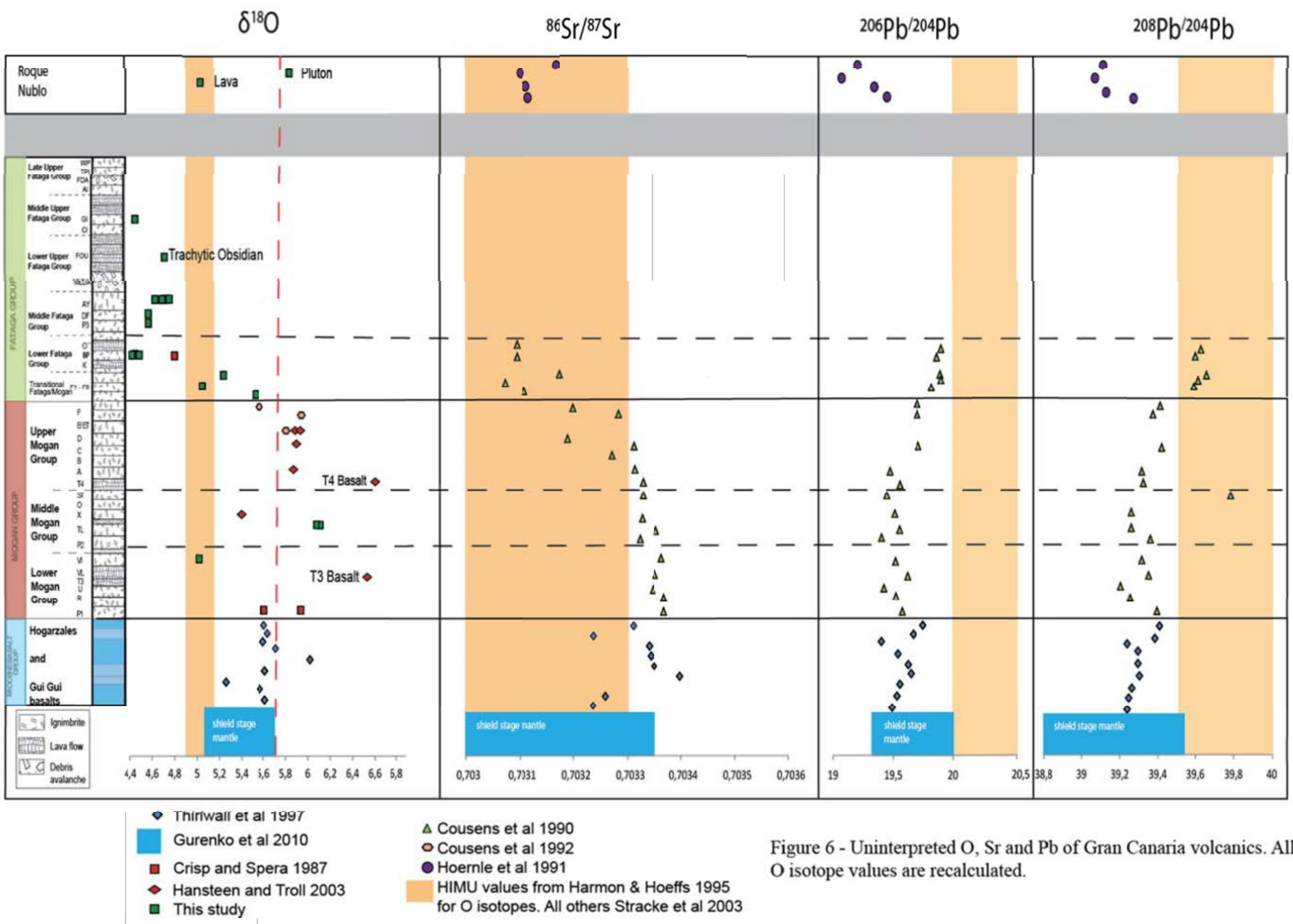



Figure 6 - Uninterpreted O, Sr and Pb of Gran Canaria volcanics. All O isotope values are recalculated.

**Table 1.1 Melt and fractional crystallization corrected oxygen isotopes of the Mogan group**

| Strat. Unit | Eruptive Unit | | Analysed d18O ‰ | SiO% | F.C. | d18O Parental magma | Data Source |
|---|---|---|---|---|---|---|---|
| Lower Mogan | Ignimbrite P | P1a | 5 | 65.2 | 0.7 | 4.3 | Crisp and Spera 1987 |
| | | P1b | 6.1 | 65.2 | 0.7 | 5.4 | Crisp and Spera 1987 |
| | | P1c | 6.6 | 65.2 | 0.7 | 5.9 | Crisp and Spera 1987 |
| | Basalt | T3 basalt | 6.4 | 58 | N/A | N/A | Hansteen and Troll 2003 |
| | Ignimbrite VI | VI a | 5.6 | 70 | 0.8 | 4.7 | **This study*** |
| | | VI b | 5.5 | 70 | 0.8 | 4.7 | **This study*** |
| Middle Mogan | Ignimbrite TL | TL Vit MCA | 6.8 | 66.8 | 0.7 | 6 | **This study**** |
| | | TL Vit MCA2 | 7.0 | 70.5 | 0.9 | 6.1 | **This study**** |
| | Ignimbrite X | X | 6.1 | 68.6 | 0.8 | 5.3 | Hansteen and Troll 2003 |
| | | X | 6.0 | 68.5 | 0.8 | 5.2 | Troll et al 2003 |
| Upper Mogan | Alkali Basalt | T4 basalt | 6.5 | 43.6 | N/A | N/A | Hansteen and Troll 2003 |
| | Ignimbrite A | A | 6.5 | 67 | 0.7 | 5.8 | Hansteen and Troll 2003 |
| | | PAT-MCA | 6.1 | 69.9 | 0.8 | 5.2 | Troll and Schmincke 2002 |
| | | PAT-Agu | 6.2 | 67.9 | 0.8 | 5.4 | Troll and Schmincke 2002 |
| | | A-II-Mca | 6.3 | 67.5 | 0.7 | 5.5 | Troll and Schmincke 2002 |
| | | A-F-28 | 6.3 | 69.4 | 0.8 | 5.4 | Troll and Schmincke 2002 |
| | | A-F-32 | 6.4 | 68.6 | 0.8 | 5.6 | Troll and Schmincke 2002 |
| | | A-O-Mca | 6.6 | 69.2 | 0.8 | 5.7 | Troll and Schmincke 2002 |
| | | A-F-36 | 6.7 | 69.6 | 0.8 | 5.8 | Troll and Schmincke 2002 |
| | | A-F-27 | 7.0 | 69.7 | 0.8 | 6.1 | Troll and Schmincke 2002 |
| | | A-F1-BTTS | 6.9 | 70.1 | 0.9 | 6 | Troll and Schmincke 2002 |
| | | A-F2-Bto | 7.0 | 70.5 | 0.9 | 6.1 | Troll and Schmincke 2002 |
| | | A-F11-Bto | 7.1 | 70.6 | 0.9 | 6.2 | Troll and Schmincke 2002 |
| | | A-F8 | 7.0 | 67.5 | 0.7 | 6.2 | Troll and Schmincke 2002 |
| | | A-F3 | 6.9 | 66.1 | 0.7 | 6.1 | Troll and Schmincke 2002 |
| | | A-F6 | 6.6 | 66.9 | 0.7 | 5.8 | Troll and Schmincke 2002 |
| | | A-F38 | 6.4 | 65.8 | 0.7 | 5.7 | Troll and Schmincke 2002 |
| | | A-PI-1-PR | 6.9 | 59.5 | 0.4 | 6.4 | Troll and Schmincke 2002 |
| | | A-PI-12-PR | 6.9 | 66.9 | 0.7 | 6.1 | Troll and Schmincke 2002 |
| | | A-PI-11-SP | 7.3 | 58.2 | 0.4 | 6.9 | Troll and Schmincke 2002 |
| | Ignimbrite D | D | 6.5 | 65.5 | 0.7 | 5.8 | Hansteen and Troll 2003 |
| | Ignimbrite E | E | 6.5 | 65.5 | 0.7 | 5.7 | Hansteen and Troll 2003 |
| | | E | 6.5 | 63.6 | 0.6 | 5.9 | Cousens et al 1992 |
| | Ignimbrite ET | ET | 6.3 | 64.2 | 0.6 | 5.6 | Cousens et al 1992 |
| | Ignimbrite F | F | 6.64 | 65.4 | 0.7 | 5.9 | Cousens et al 1992 |
| | | F | 6.28 | 64.7 | 0.6 | 5.5 | Cousens et al 1992 |

*Analysed at University of Oregon **Analysed at University of Cape Town

**Table 1.2 Melt and fractional crystallization corrected Oxygen isotopes of the Fataga group**

| Strat. Unit | Eruptive Unit | | Analysed d18O ‰ | SiO% | F.C. | d18O Parental Magma | Data Source |
|---|---|---|---|---|---|---|---|
| **Transitional Mogan/ Fataga** | | F III BTO | 6.1 | 65.0 | 0.6 | 5.4 | **This study*** |
| | | F I BTO | 5.8 | 65.0 | 0.6 | 5.1 | **This study*** |
| | | F F5 BTO | 5.9 | 65.1 | 0.6 | 5.2 | **This study*** |
| **Lower Fataga** | Ignimbrite G | G | 5.5 | 62.8 | 0.6 | 4.9 | Crisp and Spera 1987 |
| | | BPB-FSP | 5.1 | 62.8 | 0.6 | 4.5 | **This study*** |
| | | BPM-FSP 1.45 | 5.1 | 62.8 | 0.6 | 4.5 | **This study*** |
| | | BP Top FSP 1.59 | 5.2 | 62.8 | 0.6 | 4.5 | **This study*** |
| **Middle Fataga** | Ignimbrite P | P3 V | 5.2 | 63.3 | 0.6 | 4.6 | **This study*** |
| | | P3 I | 5.2 | 62.4 | 0.5 | 4.6 | **This study*** |
| | Ignimbrite DF | DF IV | 5.2 | 62.8 | 0.6 | 4.6 | **This study*** |
| | Ayaguras Ignimbrite | AY BAY 11 | 5.5 | 63.2 | 0.6 | 4.8 | **This study*** |
| | | AY II BTO | 5.4 | 62.8 | 0.6 | 4.8 | **This study*** |
| | | AY VBTO | 5.4 | 63.2 | 0.6 | 4.8 | **This study*** |
| | | AY 10 BTO | 5.4 | 62.8 | 0.6 | 4.7 | **This study*** |
| | | AY VI BTO | 5.3 | 63.1 | 0.6 | 4.7 | **This study*** |
| **Lower upper Fataga** | Trachyte Obsidian | F-Ob-ML-3 | 5 | 56 | 0.3 | 4.7 | **This study**** |
| **Middle upper Fataga** | Ignimbrite GI | FGI-BAG2 | 5 | 63.7 | 0.6 | 4.4 | **This study*** |

*Analysed at University of Oregon **Analysed at University of Cape Town

**Table 1.2 Melt and fractional crystallization corrected oxygen isotopes of the Roque Nublo group**

| Strat Unit | Eruptive unit | Analysed d18O‰ | SiO% | F.C. | d18O parental magma | Source |
|---|---|---|---|---|---|---|
| Lava | RNL-1 | 5.8 | 60.5 | 0.5 | 5.0 | **This study*** |
| Pluton | RN PLUTO | 6.1 | 49 | 0.05 | 5.8 | **This study*** |
| Ignimbrite | RQNB | 5.8 | 51.3 | 0.1 | 5.6 | **This study*** |

*Analysed at University of Oregon

*Parent magma isotope composition for the Mogan and Fataga groups*

The majority of corrected isotope values for the Mogan group samples fall in the range 4.3 to 6.9 ‰ (Table 1.1, Fig 6). The lowest value is observed in ignimbrite P1 of the Lower Mogan group which falls at 4.3 ‰ (Crisp and Spera 1987) while the highest are at 6.9 ‰ in Ignimbrite A. The Lowerr Mogan has values in the range of 4 ‰ to 6.4‰ (Crisp and Spera 1987, Hansteen and Troll 2003). Into the Middle Mogan, the values of parental magmas fall in the range of 5.2 ‰ to 6 ‰ (Troll and Schmincke 2000, Hansteen and Troll 2003, Troll et al 2003). There is an increase in values in ignimbrite A (start of the upper Mogan group) from 5.8 ‰ to 6.9 ‰ (Hansteen and Troll 2003, Troll and Schmincke 2000, Troll et al 2003), which starts a trend of rising parental melt values before a slight lowering towards the beginning of the transition stage between the Mogan and the Fataga is again observed (Fig 6).

Corrected oxygen isotopes values for the Fataga group show a range from 4.4 ‰ to 5.4 ‰, with an overall decrease from the lower to the upper Fataga units (Table 1.2, Fig 6). Overall, the oxygen isotope values for the parental magmas of the Fataga group show a decreasing trend upwards and the values are noted to be lower than those for the underlying Mogan group.

*Parental magmas of the Roque Nublo group*

Oxygen isotope melt values for the Roque Nublo range from 5 ‰ to 5.8 ‰ (Table 1.3 Fig 6), only three values have been collected, however as the Roque Nublo was not the focus of this study. Although useful, they cannot reveal any particular trends at this stage.

*4.2 Temporal evolution of Miocene to Pliocene parental magmas*

To fully characterize the parental melts of each of the successive volcanic stages on Gran Canaria, the corrected oxygen isotopes will be used in conjunction with the available radiogenic isotopes (Fig 6).

It appears that the shield stage melts have fairly stable oxygen and strontium isotopes that are slightly higher than corresponding values for local mantle (Thirlwall et al 1997, Gurenko et al 2010) and Sr isotopes can be seen to increase upsection through the shield basalts (Thirlwall et al 1997).

The Mogan samples show higher oxygen isotope values for the parental magma throughout the Lower, middle and upper groups when compared with the shield stage (Freundt and Schmincke 1995, Hansteen and Troll 2003, Troll et al 2003, Troll and Schmincke 2000). Strontium isotopes remain uniform throughout much of the lower and middle groups before a pronounced decrease in the upper Mogan group (Fig 6) (Cousens et al 1990, Troll and Schmincke 2002). Lead Isotopes are observed to have the corresponding trends with a rather uniform signature through the lower and middle Mogan groups but a slight increase in the upper Mogan group samples for both isotopes (Cousens et al 1990).

At the transition from the Mogan to the Fataga groups, oxygen isotopes are observed to notably decrease in value and this trend continues into the lower Fataga group and into the middle and upper Fataga groups (Fig 6). At this point, the strontium isotope curve also shows a shift to lower values, similar to the oxygen isotopes (Fig 6) (Cousens et al 1990). For the lead isotope values, an increase is seen at this point in time and the two lead isotopes behave in a synchronous fashion (Fig 6).

A comparison with the values of HIMU mantle for ocean island settings (e.g. Harmon and Hoefs 1995, Stracke et al 2003), reveals that oxygen isotopes for the shield basalts and Mogan group plot at higher values than what is expected for pure HIMU or plume-type oxygen isotope component (Harmon and Hoefs 1995). In contrast, the Fataga group falls towards the HIMU field (Fig 6), while the Roque Nublo group yields values that once again plot above that of pure HIMU. A similar trend is seen for strontium isotopes, where shield basalts and Mogan group values are initially higher than pure HIMU or OIB mantle (Stracke et al 2003), before trending towards a HIMU average value in the Fataga period (Fig 6). Lead isotope values are observed to be below those for HIMU in the shield basalts and the Mogan group before showing an increase towards HIMU-OIB-type mantle values in the Fataga group. By the time the Roque Nublo group is deposited, lead isotope values once again fall below HIMU-OIB-type mantle compositions, indicating mixed mantle sources at play (Fig 6). Notably, a progressive change is seen in all isotope systems at the boundary of the Mogan to the Fataga groups, but no such change is observed between the shield basalts and the Mogan group (Fig 6). A compositional change is again inferred for the Pliocene Roque Nublo group that erupted after a considerable hiatus of ~3 to 4 Ma.

## 5. DISCUSSION

### *5.1 Origins of magmas feeding Gran Canaria volcanism*

Following correction for crystal-magma fractionation and fractional crystallization processes, two key processes can be inferred to control the isotopic characteristics of Miocene Gran Canarian magmas. These are changes in the type of mantle supplying the magmatic activity and crustal modifications derived from the melting and assimilation of ocean crust or island edifice material. To assess the degree to which each of these processes is relevant for magma generation and evolution it is necessary to interpret the stratigraphic isotope data in regard to changes through time (Figs 7 and 8).

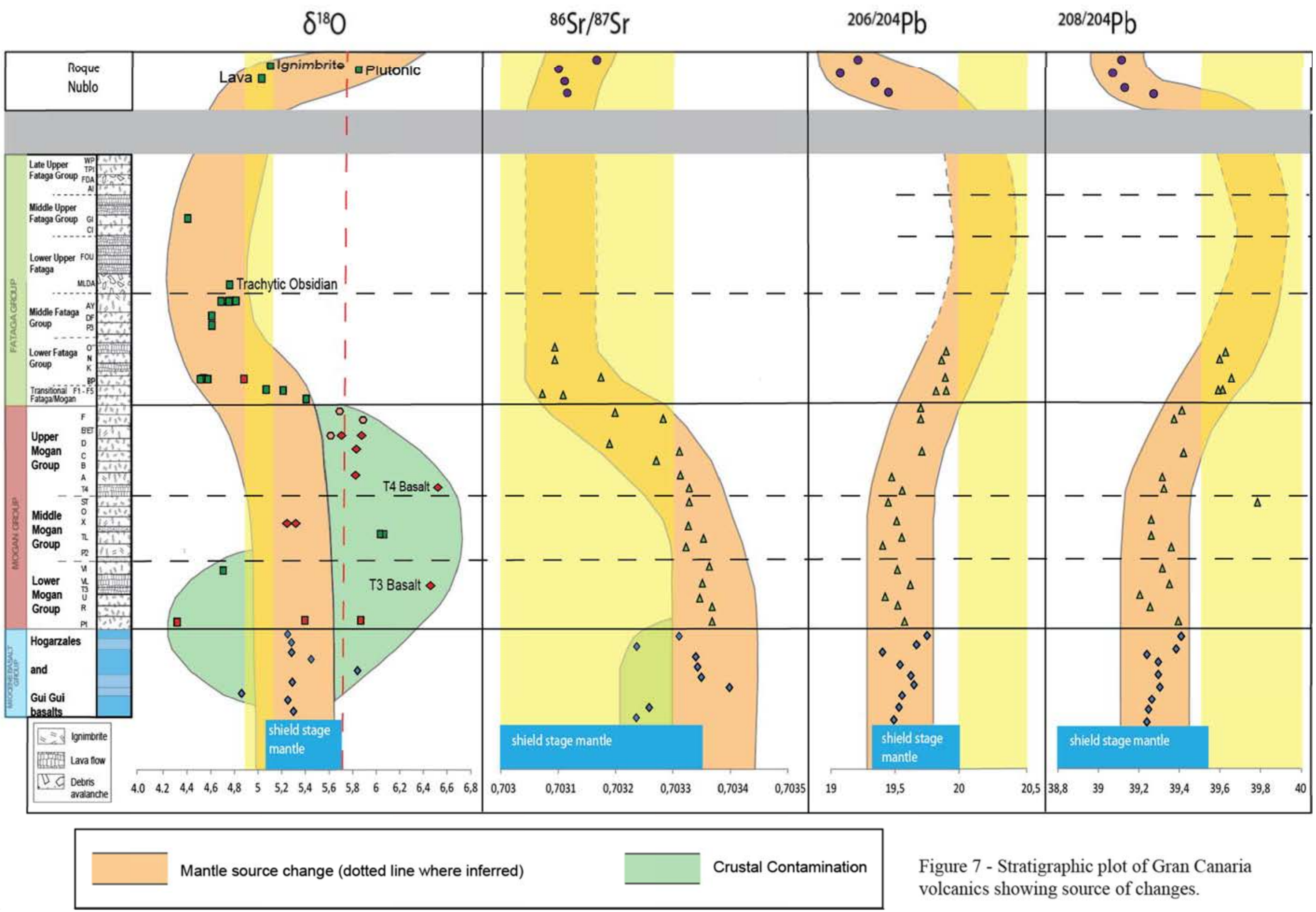



Figure 7 - Stratigraphic plot of Gran Canaria volcanics showing source of changes.

*5.1.1 Mantle changes*

*Shield basalts*

The origin of the shield basalt magmas has been discussed in previous studies (eg Thirlwall et al 1997, Gurenko 2010) and a brief summary is given here. In figures 6 and 7 a, a uniform range of values through time is observed for the shield stage and its probable mantle source. In figure 8a, the shield stage magmas are seen to lie between a HIMU and a DM component (+ a possible EM component), implying that during this period mantle sources were likely a mixture of at least a HIMU and DM component.

*The Mogan group*

In Fig 5 and 7 it is apparent that the oxygen isotope values of the Mogan group likely evolved from mantle melts that were compositionally similar to that of the shield basalts. To start with, the Mogan group is apparently dominated by the DM component, but evolves towards a more HIMU-rich input with time, which is consistent with available strontium and lead isotopes (Fig 7) (Cousens et al 1990).

*The Fataga group*

In Fig 7 the Fataga group shows influence of mantle components with a different oxygen isotope signature to that of the preceding Mogan group and shield basalts. As to the nature of this source, it can be seen in Fig 7 that the oxygen, strontium and lead isotopes all move towards a more HIMU-like component (c.f. Cousens et al 1990). This shift is also observed in figure 8 where the Fataga group values fall closer towards the HIMU field than the Mogan group.

Starting at the transition of the Mogan to the Fataga groups, the DM component (and a possible EM component) appears to wane and the HIMU component came to dominate the parental magmas which fed the majority of the Fataga group eruptions. This shift indicates the appearance of the main part of the HIMU heterogeneity according to Cousens et al (1990) and Hoernle (1998).

*The Roque Nublo*

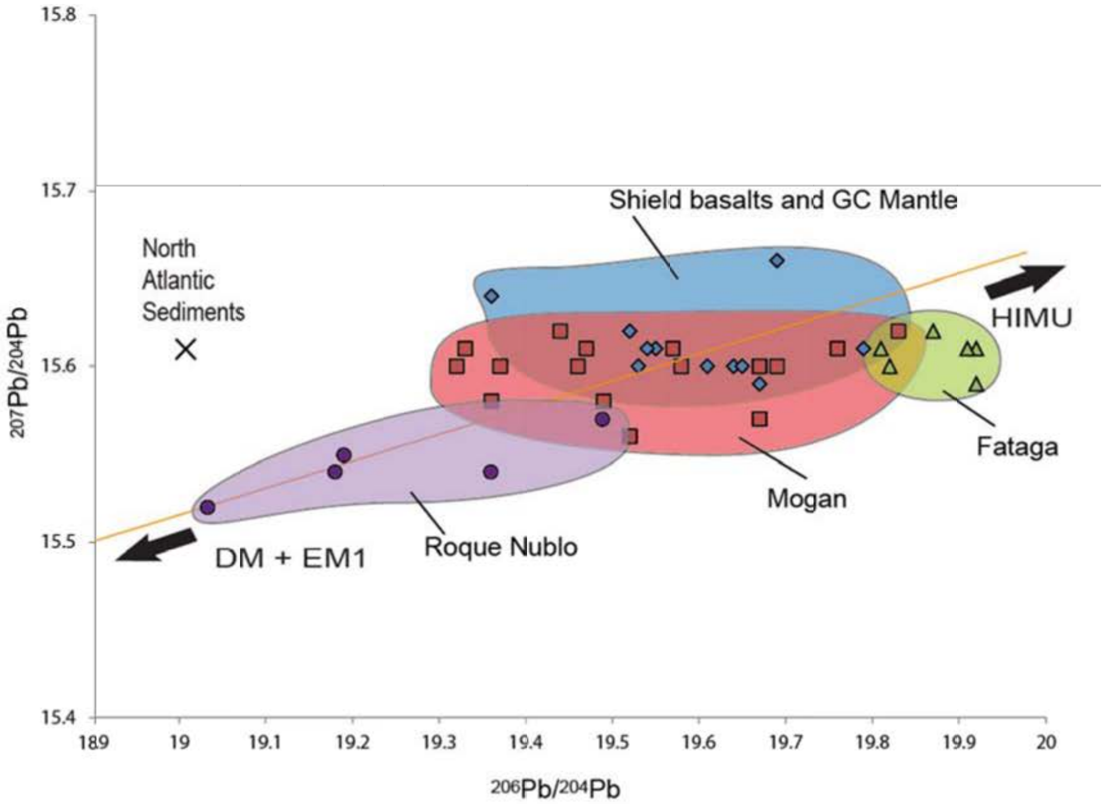
By the time the Roque Nublo eruptions commenced, another change in the mantle source component had taken place. As can be seen in figure 7, the oxygen and lead isotopes have moved away from the HIMU influence, indicating that a different component is fuelling this younger magma than the mixture that gave rise to the Fataga eruption phases. The lead plot (Fig 8), also shows that the Roque Nublo magmas are moving away from the earlier HIMU dominance towards one which is more dominated by a DM (and likely also EM) component (Hoernle et al 1991).

### 5.1.2 *Crustal modifications/additions*

A useful way to assess the levels of crustal additions is by the employment of oxygen isotope values in addition to the existing radiogenic isotope data available (Fig 6). Crustal additions should appear as semi-random and as anomalous excursions from overall trends in figure 7, unlike mantle changes, which should be gradual and long-lived. This notion is due to the fact that crustal input may be variable in individual eruptions and will thus differ between events.

*Shield basalts*

Although most oxygen isotope values for the shield basalts are consistent with the mantle range, there are a few samples which are not. These either exhibit values higher or lower than those expected for the mantle (Fig 7). In figure 8 it appears that some shield stage magmas are displaced towards values



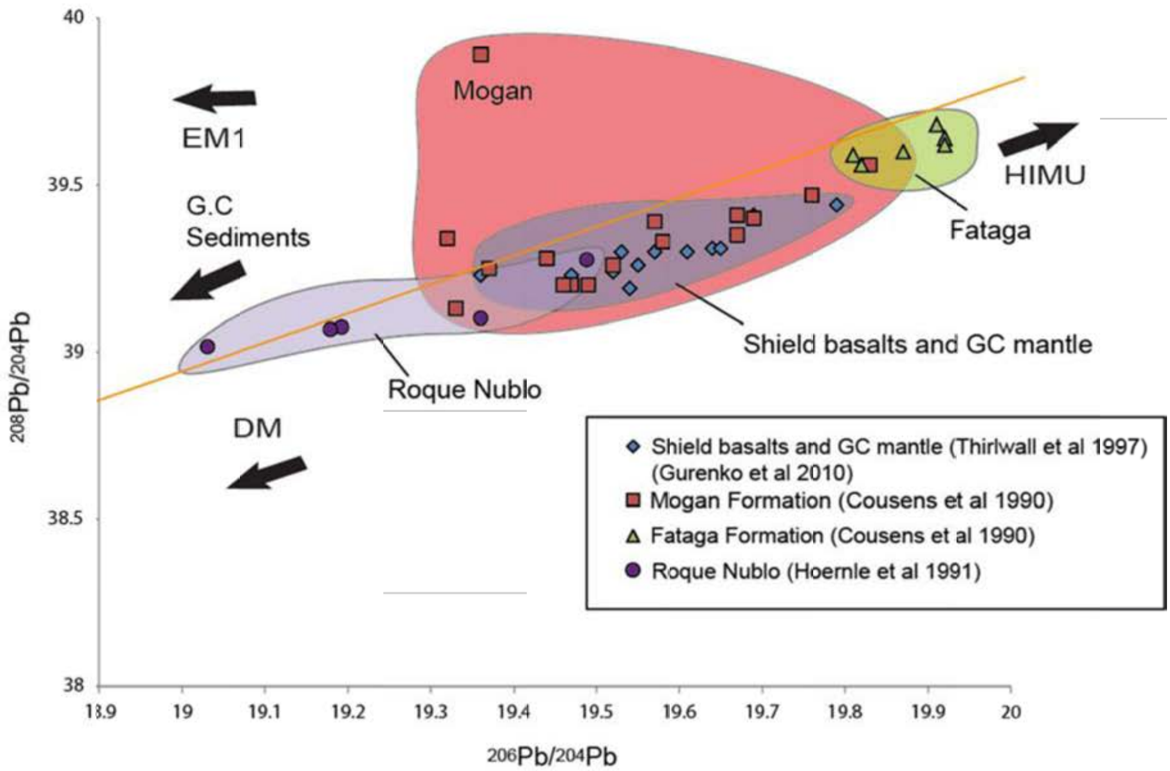




Figure 8 – Lead isotopes values  or Gran Canaria volcanics.

of local Canary and regional Atlantic sediments (e.g. Thirlwall et al 1997, Hansteen and Troll 2003). As has been highlighted by Thirlwall et al (1997), some of the Shield stage lavas have been contaminated by another component, a low $\delta^{18}O$ one, which Hansteen and Troll (2003) suggest to be dominantly altered ocean crust.

Indeed two main stagnation levels for the shield basalts have been identified; one at 10 km and another at 3.5 km (Hansteen et al 1998, Gurenko et al 1998). At 10 km, the crustal material changes from altered (low $\delta^{18}$ O) Jurassic ocean crust to (high $\delta^{18}$ O) Mezosoic sediments (Hoernle 1998, Ye et al 1999, Krastel and Schmincke 2002, Hansteen and Troll 2003). It is therefore probable that the mixture of low and high oxygen isotope values reflects assimilation of altered ocean crust and later oceanic sediments (Fig 7).

*The Mogan group*

In the Mogan group, many of the oxygen isotope values are elevated relative to the projected mantle components (Fig 7). These elevations are particularly noted in T3 and T4 basalts (Figs 5a, 7) which are considerably higher than the preceding shield basalts. We note that within the stratigraphic plot, the data by Cousens (1990) for Sr and Pb isotopes (Fig 7) do not show these excursions, although several units have previously been shown to be affected by crustal additions (e.g. T4 basalts, Ignimbrite A; Troll and Schmincke 2002, Troll and Hansteen 2003).

Based on our oxygen evidence, it can be suggested that several of the magmas feeding the Mogan eruptions interacted with a high $\delta^{18}O$ component. The most likely candidates for the high $\delta^{18}O$ input are Jurassic sediments or recycled altered igneous components at higher levels within the island edifice (Hansteen and Troll 2003). During the time of the Mogan group eruptions, the main depth of the magma chambers was higher than during the shield stage, with magmas ponding at a depth of 5 to

7 km (Crisp and Spera 1987, Cousens et al 1990, Freundt and Schmincke 1995, Freundt-Malecha et al 2001, Troll and Schmincke 2002, Hansteen and Troll 2003). If magmas were pooling at a shallow level then it is extremely likely that assimilation of the volcanic edifice itself, in a similar manner to that seen at Yellowstone (Bindeman 2001) might have occurred to a degree. Geophysical and petrological data confirms the existence of a mafic to felsic intrusive complex at the centre of the island (Crisp and Spera 1987, Freundt and Schmincke 1995, Ye et al 1999, Freundt-Malecha et al 2001, Krastel and Schmincke 2002, Troll and Schmincke 2002, Hansteen and Troll 2003). Alternatively, Hansteen and Troll (2003) considered sediment to be present too at this depth due to intrusive uplift (c.f. Robertson and Stillman 1979).

*The Fataga group*

The Fataga group does not show any of the anomalous excursions towards higher $\delta^{18}O$ as seen in the preceding Mogan group. Although the oxygen isotopes do plot lower than MORB values (Fig 5a Fig 7), the trends seen in the oxygen isotopes are mimicked by a drop in the Strontium isotopes and an increase in Pb isotopes and remarkably, changes in values for all isotopes happened at the same time (Fig 7). Notably the strontium isotope changes show the Fataga group samples change towards less radiogenic values. In combination with decreasing $\delta^{18}O$ values this change does most likely not reflect regular crustal additions, but a systematic change in source supply (Fig 8b).

This realization implies that by the time of the Fataga group eruptions, crustal input was minimal or non-existent and did not affect the isotope signatures significantly. Indeed, Cousens et al (1990) argued that the magma chamber system associated with the Fataga group could have become insulated from the surrounding rocks, thus reducing crustal input with time (see also Kerr et al 1998). Curiously, however, the Fataga group represents a decline of Miocene activity at Gran Canaria and it is thus not clear how the proposed HIMU input can be explained at this point in time. Most probably, the DM (and EM) components were progressively used up by mantle melting, leaving a HIMU enriched residue only as the mantle source for the Fataga volcanism.

*Roque Nublo*

It is difficult to ascertain the role of crustal contamination in the Roque Nublo magmas with the limited oxygen isotope data available. However, the plutonic xenolith might have undergone some degree of crustal contamination (Fig 7) as it has an unusually elevated value.

**6. Charting temporal geochemical changes in the Miocene volcanics of Gran Canaria.**

The diagrams in figure 9 detail the geochemical evolution of Gran Canaria via both changing mantle components and crustal additions. At the beginning of the Miocene, a large blob of HIMU-like material rose through the asthenosphere (plume head) and entrained DM (+EM) mantle material from the surrounding lithosphere during this stage (Fig 9A). Decompression melting then generated melts which were temporarily stored in the upper mantle. After traversing the Moho, magmas then got in contact with layer 2 and layer 1 of the ocean crust and assimilation of small portions of crustal materials did cause shifts in oxygen isotope signatures of some of the shield stage magmas, thus explaining deviations from mantle values (Fig 9 A).

Following the shield stage, the geochemical nature of the blob(s) remained broadly similar (Fig 9B) although the location of magma storage changed with the Mogan group reservoirs situated higher up and within the volcanic edifice itself (Fig 9 B). This change resulted in the assimilation of pre-existing (likely hydrothermally altered) intrusive materials and thus generated the contaminated melts of the Mogan group. Towards the close of Mogan group activity, however, a broad change indicates a

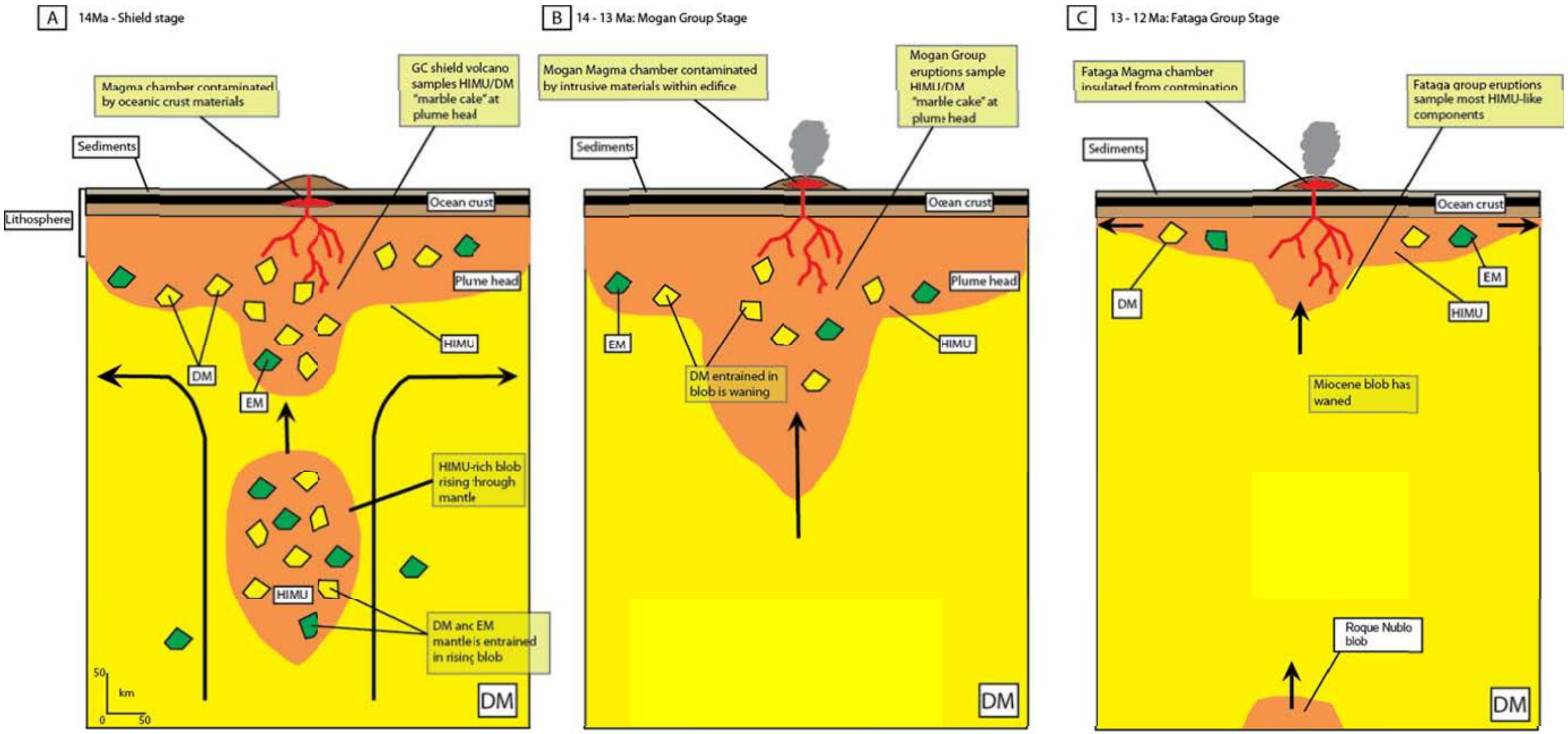



Fig 9 - Cartoons showing various crustal and mantle inputs during A: the Shield Stage, B: The Mogan Group Stage and C: The Fataga Stage waning of the DM (and EM) component in the source. During the shield stage and during Mogan times, the supplying plume likely had a “marble cake” appearance, containing different pockets of HIMU, DM and some EM-derived materials (Fig 9 A, B).

By the time of the eruption of the Fataga group, the feeder system and magma chambers had become largely insulated from crustal influences.The change in $\delta^{18}$ O recorded in the Fataga group is closely correlated with a change in radiogenic isotopes (Strontium and Lead), which leads to the conclusion that the waning of the DM component in the mantle blob, rather than any significant crustal contamination (Fig 9 C). Either the waning blob simply entrained less DM, or the DM component was at this point largely melted out thus causing a dominantly HIMU-like isotope signature. The dying nature of the blob is confirmed by the fact that although the Fataga goup is the same volume as the preceding Mogan, it took a much longer period to be erupted (c.a. 4 Ma as opposed $\leq$ 1 Ma for the Mogan group) and is thus a likely reflection of the volcanic system gradually shutting down prior to the post-Miocene hiatus.

This long hiatus of 4Ma, between the Fataga and the Roque Nublo groups, further indicates the decline of the volcanic system and the waning of the Miocene blob. By the time of the Pliocene recommencement of volcanism, a new blob had arrived as available O, Sr and Pb suggest a different source component mixture than in Fataga group. Upon impact with the lithosphere the new Pliocene blob spread out and triggered the rejuvenation of volcanic systems on Gran Canaria leading up to the Pliocene Roque Nublo volcanic episode.

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