Magma storage at Ocean Islands: insights from Cape Verde

Barker AK^{1,2}, Rydeblad E^{1,3}, Silva SMDM⁴

1 Mineralogy, Petrology, Tectonics, Department of Earth Sciences, Uppsala University, Villavägen 16, Uppsala SE-752 36, Sweden

2 Centre of Natural Hazards and Disaster Sciences (CNDS), Villavägen 16, Uppsala SE-752 36, Sweden

3 Department of Earth Science and Engineering, South Kensington Campus, Imperial College London, SW7 2AZ, UK

4 Faculty of Science and Technology, University of Cabo Verde, CP 279, Praia, Santiago, Cabo Verde

Abstract

The Cape Verde archipelago is a group of Ocean Islands in the Central Atlantic that forms two chains of islands trending Northwest and Southwest. Several of the islands are considered to be volcanically active, with frequent eruptions on Fogo. We examine the mineral chemistry and thermobarometry of the southern islands; Santiago, Fogo and Brava together with the Cadamosto Seamount. Our objective is to explore the magmatic storage system and implications for volcanic eruptions and associated hazards at Cape Verde. The volcanic rocks at Cape Verde are alkaline and dominantly mafic, whereas the island of Brava and the Cadamosto Seamount are unusually felsic. Clinopyroxene compositions range from 60 to 90 Mg# at Santiago and Fogo. In contrast, at Brava and the Cadamosto Seamount the clinopyroxene compositions are 5 to 75 Mg#. Mineral chemistry and zonation records fractional crystallization, recharge, aggregation of crystals, magma mixing and variations in thermal conditions of the magma at temperatures from 925 to 1250°C. Magma storage depths at Santiago, Fogo, Brava and the Cadamosto Seamount are between 12 and 40 km, forming deep sub-Moho magma storage zones. Transient magma storage in the crust is suggested by fluid inclusion re-equilibration and pre-eruption seismicity. A global compilation of magma storage at Ocean Islands suggests deep magma storage is a common feature and volcanic eruptions are often associated with rapid magma ascent through the crust. Shallow magma storage is more variable and likely reflects local variations in crustal structure, sediment supply and tectonics. Petrological constraints on the magma plumbing system at Cape Verde and elsewhere are vital to integrate with deformation models and seismicity in order to improve understanding and mitigation of the volcanic hazards.

Introduction

Ocean Islands occur in all of the world's oceans and are formed by intraplate volcanism. Several Ocean Islands are volcanically active with eruptions at Reunion, Canary Islands, Cape Verde and Hawaii between 2007 and 2018 (e.g. Di Muro et al., 2014; Longpré et al., 2014; Worsley, 2015; Liu et al., 2018). Many of the Ocean Islands are inhabited and have lucrative tourist industries, which put people directly at risk from the volcanic hazards associated with eruptions. Additionally, submarine and subaerial eruptions can also be hazardous for the shipping and aviation industries respectively (O'Mangain et al., 2007; Di Muro et al., 2014). Consequently, understanding the magma plumbing systems from the magma storage zones to the volcanic centers is imperative for assessing and monitoring the volcanic hazards. We present insights into the magma storage at Cape Verde, to explore the nature of the volcanically active parts of the archipelago which is motivated by the recent eruptions at Fogo (Ribeiro, 1960; Worsley et al., 2015).

The Cape Verde archipelago is an Ocean Island group built upon the Cape Verde Rise (Ramalho et al., 2010a,b). The older and now heavily eroded islands of Maio, Sal and Boa Vista are found in the East, from where two volcanic chains emerge. The northern chain consists of São Nicolau, Santa Luzia, São Vicente and Santa Antão. The islands of Santiago, Fogo, and Brava as well as the Cadamosto Seamount form the southern island chain (Figure 1). The island of Fogo displays the iconic Pico do Fogo, peaking at 2829 m above sea level, located inside the Bordiera, on the Chã das Caldeiras, where the historic eruption sites are found; the latest eruption was November 2014 to February 2015 (Silva et al., 2015; Fernandes & Faria, 2015; González et al., 2015; Worsley 2015; Richter et al., 2016; Jenkins et al., 2017). Prior to that nine eruptions occurred at Fogo between 1760 and 1995, with a total of 28 eruptions since settlement in the 15th century (Ribeiro, 1960; Nascimento, 2015). Brava and the Cadamosto Seamount have no historical record of volcanic activity. However, Holocene volcanic activity is documented at Brava and the Cadamosto Seamount records eruption ages of 21 ka (Madeira et al., 2010; Faria & Fonseca, 2014; Samrock et al., 2019). Additionally, seismicity extends from Fogo to the Cadamosto Seamount suggesting that Brava and the Cadamosto Seamount are also underlain by active magmatic systems (Heleno et al., 2006; Grevemeyer et al., 2010).

The volcanic hazards on Fogo are in the form of ash and lapilli associated with strombolian activity and lava flows as well as occasional sub-plinian to plinian eruptions (Torres et al., 1997; Eilese et al., 2015). The residents of Fogo are vulnerable to the direct volcanic hazards associated with an eruption, especially the approximately one thousand who live and work inside the Bordeira close to Pico do Fogo, followed by the 11,000 residents of the East coast of Fogo (Faria & Fonseca, 2014). The eruptions in 1995 and 2014/2015 led to large-scale evacuation as well as significant damage to property, agricultural land and buildings (Silva et al., 2015; Worsley 2015; Jenkins et al., 2017). Potential volcanic hazards at

Brava are connected to pyroclastic and phreatomagmatic eruptions and endanger the 6,000 inhabitants (Faria & Fonseca, 2014).

In addition, the onshore and offshore geological record of the Cape Verde archipelago shows evidence for flank collapses and landslide deposits (Masson et al., 2002, 2008; Paris et al., 2018; Barrett et al., 2020). The Bordeira cliffs of Fogo were potentially exposed by a huge landslide (Day et al., 1999; Foeken et al., 2009; Ramalho et al., 2015; Martínez-Moreno et al., 2018). Tsunami deposits have been used to constrain the associated flank collapse and landslide to approximately 73 ka (Ramalho et al., 2015). Furthermore tsunami deposits associated with local flank collapses are found, among other locations, on Santiago and Maio (Paris et al., 2011, 2018; Ramalho et al., 2015; Madeira et al., 2019). Therefore the Cape Verde Islands are associated with multiple natural hazards that may pose a threat to the population both locally and farther afield.

Monitoring of the volcanism is the responsibility of the National Institute for Metrology and Geophysics. In addition the University of Cape Verde has been conducting geochemical monitoring in partnership with national and international institutions (Fonseca et al., 2003; Barrancos et al., 2015; Pérez et al., 2015; Dionis et al., 2015a, b). Seismicity and deformation is monitored throughout Cape Verde with 15 seismometers and a geodetic network (Faria & Fonseca, 2014). The Fogo-Brava seismic network consists of seven seismometers on Fogo and two on Brava and Fogo hosts three geodetic stations. Additional monitoring targets volcanic gas emissions with semi-frequent measurements of CO₂ and SO₂ emissions and fumerole temperatures (Faria & Fonseca, 2014; Worsley 2015). The National Institute for Metrology and Geophysics work with the National Civil Protection service to provide risk communication (Faria & Fonseca, 2014).

The magmatic system can be investigated through minerals in the erupted products. The minerals capture the compositional signature of the host magma during growth, therefore they respond to variations in magma composition. Consequently minerals provide a chronological record of magma evolution (Davidson et al., 2007). On this basis mineral thermobarometry offers a petrological method to estimate the crystallization conditions of magma chambers, including the temperature and pressure of crystal growth (e.g. Putirka, 2008). Igneous thermobarometers have been widely applied in many settings such as volcanic arcs (e.g. Till, 2017; Lai et al., 2018) and to investigate continental magmatism (Putirka et al., 2003; Putirka & Condit, 2003). Clinopyroxene-melt thermobarometers have also been applied in Ocean Islands settings to explore tholeiitic to alkaline systems e.g. Hawaii, Cape Verde, Canary Islands (e.g. Putirka, 1997; Klügel et al., 2000, 2005, Nikogosian et al., 2002; Aulinas et al., 2010; Hildner et al., 2011, 2012; Longpré et al., 2014; Mata et al., 2017). Despite large uncertainties associated with

crystallization pressure estimates, thermobarometry offers several advantages that complement geophysical methods for determining magma chamber depth location (Magee et al., 2018). Firstly a magma pocket requires significant dimensions to be detected by seismic arrays, even at dense seismic station distribution (e.g. Gudmundsson, 2012). Secondly seismicity picks up differences in seismic velocities, and thereby requires magma pockets to be melt rich, if they are mush zones with a significant proportion of crystals then the melt:solid ratio may not be high enough to observe variations in seismic wave velocities (e.g. Gudmundsson, 2012). Additionally, studies of deformation often employ petrological data as inputs to estimate the shape, size and depth of the point source, to model the inflation of a magma chamber (e.g. Amelung & Day, 2002; González et al., 2015). Petrological techniques can thus provide crucial information on magma storage for integration with geophysical models (Magee et al., 2018). Whereas geophysical methods are not only useful for investigating the magma storage system but also have distinct advantages in real-time monitoring of magma movement prior to eruptions (Magee et al., 2018).

Furthermore, thermobarometry determines pressure and temperature of mineral growth and is often sensitive to crystals growing in small magma pockets (e.g. Geiger et al., 2016; Stock et al., 2018). In many cases these magma pockets are not detected until earthquakes trace magma ascent prior to an eruption. Several cases have shown that other methods confirm the models for magma plumbing systems developed from thermobarometry (e.g. Longpré et al., 2014; Barker et al., 2019). However thermobarometry is not without limitations, restricted by the minerals present as well as magma compositions and experimental conditions of calibrations (e.g. Putirka, 2008). For instance, although clinopyroxene crystallizes over wide temperature, pressure and compositional ranges the felsic magmas in many settings evolve beyond clinopyroxene crystallization (e.g. Weidendorfer et al., 2016). Therefore other minerals or features such as fluid inclusions and volatile solubility in glass are required to trace the more evolved parts of the magmatic system (e.g. Schwarz & Klügel 2004; Jeffery et al., 2016).

All volcanic activity is fed from the magma plumbing system, therefore it is important to understand the magma storage, magmatic processes and where they occur as well as the processes that lead to magma ascent (Cooper, 2017; Putirka, 2017; Sparks & Cashman, 2017). To address the connection between the magmatic system and volcanic hazards at Cape Verde we will discuss magmatic processes recorded by mineral chemistry and the magma plumbing system derived from thermobarometric modelling for the southern Cape Verde archipelago. Mineral chemistry from the islands of Santiago, Fogo, Brava and the Cadamosto seamount will be examined to provide insights into the magmatic processes. These examples have been selected based on availability of mineral chemistry and geophysical data as well as relevance from a volcanic hazard perspective. Mineral-melt equilibrium

thermobarometry for Santiago, Fogo, Brava and the Cadamosto Seamount will be integrated into a model for the magmatic plumbing systems in southern Cape Verde. Evidence for the shallower parts of the magma plumbing systems and the volcanic hazards at Fogo will be reviewed. Finally, the magma plumbing system for Cape Verde will be compared with other Ocean Islands globally.

The Cape Verde Archipelago

The Cape Verde Rise is situated 2000 km East of the Mid-Atlantic Ridge in the Central Atlantic, between 15 and 17°N and approximately 500 km West of Dakar, Senegal (Figure 1). The Cape Verde Rise forms a gigantic bathymetric swell reaching circa 2 km above the surrounding ocean crust and covers an area of more than 0.3 million km² (Crough, 1982; McNutt, 1988; Jørgensen & Holm, 2002; Lodge and Helffrich, 2006; Holm et al., 2008; Masson et al., 2008; Brown et al., 2009). The Cape Verde islands and seamounts are located in the Southwest of the Cape Verde Rise (Figure 1; Rona, 1971; Egloff, 1972; Dash et al., 1976). The islands are arranged in a semi-circle opening to the West, with older heavily eroded islands occurring in the East and two subparallel chains of islands in the North and South, which extend westwards (Figure 1; Gerlach et al., 1988).

The northern and southern island chains are subparallel to the fracture zones found to the West of the Cape Verde archipelago (Figure 3; Jacobi and Hayes, 1982). Therefore fracture zones may influence the orientation of the island chains (Klerkx et al., 1974; Le Pichon and Fox, 1971; Doucelance et al., 2003). Evidence for fracture zone offsets is documented by magnetic lineations and gravity anomalies (Figure 3; Hayes and Rabinowitz, 1975; Dash et al., 1976; Williams et al. 1990; Verhoef et al. 1991; Roest et al. 1992; Ali et al., 2003). Additionally brittle deformation through normal, reverse and transverse faults has formed horst and graben structures on the Cape Verde Rise (Figure 3b; Efimov & Skolotnev, 2006; Skolotnev et al., 2007; 2009).

The Cretaceous ocean crust, formed at the Mid Atlantic Ridge, is overlain by a several kilometer thick sequence of hemipelagic sediments (Figure 2; Dash et al. 1976; Courtney & White, 1986; Williams et al., 1990; Ali et al., 2003; Müller et al., 2008; Pim et al., 2008). The oceanic lithosphere is approximately 85 km thick, confirmed by elevated seismic velocities extending to ca. 80 km and a Lower Velocity Zone that appears at 90±6 km depth (Cazenave et al., 1988; Lodge and Helffrich, 2006). Although the ocean crust beneath Cape Verde has normal thicknesses of 7 to 8 km, the islands are underlain by thickened crust (Pim et al., 2008; Wilson et al., 2010, 2013). The Moho beneath individual islands decreases westwards from >20 km beneath Maio, to 17 to 18 km beneath Santiago and to 12 to 14

km beneath Fogo (Lodge and Helffrich, 2006; Pim et al., 2008; Wilson et al., 2010, 2013). The sedimentary stratigraphy observed at Deep Sea Drilling Project (DSDP) Site 368 is composed of marl and nannofossil ooze, underlain by clay, claystone and minor chert associated with turbidites, plus black shale interbedded with dikes below (Lancelot et al., 1978; Faugères et al., 1989).

The Cape Verde islands display core complexes composed of intrusive carbonatite to alkaline assemblages as well as submarine pillow basalts (Serralheiro, 1970; De Paepe et al., 1974; Serralheiro, 1976; Stillman et al., 1982; Gerlach et al., 1988; Davies et al., 1989; Madeira et al., 2010; Ramalho et al., 2010b). The extrusives are dominantly mafic as exemplified by nephelinites and basanites at Santiago and Fogo (Gerlach et al., 1988; Davies et al., 1989; Barker et al., 2009; Hildner et al., 2011, 2012). However several volcanic centers such as the island of Brava and the Cadamosto Seamount exhibit largely felsic assemblages (Assunção et al. 1965; Machado et al. 1968; Madeira et al., 2008, 2010; Barker et al., 2012; Weidendorfer et al., 2016; Samrock et al., 2019). Cape Verde is one of the few Ocean Island groups hosting carbonatites, which are widespread on Brava with minor exposures on Fogo, Santiago and São Vicente (e.g., Assunção et al., 1965; Silva et al., 1981; Turbeville et al., 1987; Hoernle et al., 2002; Jørgensen and Holm, 2002; Mata et al., 2010; Mourão et al., 2010).

Geochronology of the Cape Verde archipelago

The Cape Verde islands exhibit a general age progression from the early Miocene, for the oldest exposures, which decrease westwards to the presently volcanically active parts of the archipelago (Figure 4; Bernard-Griffiths et al. 1975; Stillman et al., 1982; Mitchell et al. 1983; Torres et al. 2002; Plesner et al. 2003; Jørgensen and Holm 2002; Duprat et al., 2007; Holm et al. 2008; Foeken et al. 2009; Dyhr and Holm 2010; Madeira et al. 2010; Samrock et al., 2019). The older islands in the East are highly eroded displaying relatively flat and low topography in contrast to the shield volcanoes that occur further West. The earliest age of the eastern islands is recorded by submarine pillow basalts from the island of Sal at 26 Ma (Torres et al., 2002). The volcano-magmatic activity continues until approximately 1 Ma at Sal (Figure 4; Mitchell et al., 1983; Torres et al., 1997; Holm et al., 2008; Dyhr & Holm, 2010). The northern islands show volcanic activity from approximately 8 to 10 Ma to between 100,000 and 300,000 years ago at São Nicolau to Santa Antão (Figure 4; Plesner et al. 2003; Duprat et al., 2007; Holm et al. 2008). Seismic activity beneath Santa Antão suggests this region continues to be volcanically active (Faria & Fonseca, 2014).

There is evidence that volcanism in the southern islands commenced at Santiago approximately 11 Ma ago (Figure 4; Stillman et al., 1982; Mitchell et al., 1983; Torres et al., 2002; Ali et al, 2003). The majority of the volcano-magmatic activity on the southern islands emerges from 5.5 Ma at Santiago, 4.5 Ma at Fogo and 2.3 Ma at Brava (Holm et al., 2008; Foeken et al. 2009; Madeira et al., 2010; Ramalho et

al., 2010). The most recent volcanic eruptions from Santiago are dated to 740,000 years ago and 300,000 years ago at Brava (Holm et al., 2008; Madeira et al., 2010). The islands typically show multiple stages of volcanism interrupted by periods of quiescence (Figure 4; Holm et al., 2008; Foeken et al., 2009; Madeira et al., 2010; Ramalho et al., 2010a, 2010b, 2010c). Volcanic activity at the Cadamosto Seamount appears to occur in the last 20,000 to 100,000 years (Samrock et al., 2019). The island of Fogo displays historical volcanism with around 30 eruptions since discovery and settlement in 1460 including recent eruptions in 1995 and 2014 to 2015 (Ribeiro, 1960; Da Silva et al., 1999; Hildner et al., 2011, 2012; Silva et al., 2015; Fernandes & Faria, 2015; Worsley 2015). Additionally Fogo, Brava and the Cadamosto Seamount display ongoing seismic activity (Heleno et al., 2006; Grevenmeyer et al., 2010; Faria & Fonseca, 2014).

The geology of Santiago

The island of Santiago is composed of an intrusive complex, followed by submarine and subaerial volcanics (Figures 5 & 6; Bebiano, 1932; Serralheiro et al., 1976; Ramalho, 2011). The Complexo Antigo consists of alkaline silicate intrusives and carbonatites. The submarine stage is represented by the Flamengos Formation and exhibits pillow lavas embedded in hyaloclastite with eruption ages of 4.0 to 5.5 Ma (Figures 4, 5 & 6; Serralheiro, 1976; Holm et al., 2008; Barker et al., 2009; Ramalho et al., 2020c; Ramalho, 2011). The shield stage is formed by subaerial lavas of the Pico da Antonía Formation between 1.5 and 3.2 Ma (Figures 4, 5 & 6; Serralheiro, 1976; Holm et al., 2008; Barker et al., 2010; Ramalho et al., 2020c). Occasional carbonatite lava flows belong to the Pico da Antonía Formation (Jørgensen and Holm, 2002). The Assomada and Monte das Vacas formations erupted post-erosional lavas and scoria cones between 0.7 and 1.1 Ma (Figures 4 & 6; Serralheiro, 1976; Holm et al., 2008; Barker et al., 2010; Ramalho et al., 2020c). Additionally, a sliver of the Jurassic ocean crust has been uplifted and emplaced at Baía de Angra, plus uplifted sediments are found at Tarrafal (Serralheiro, 1976; Gerlach et al., 1988; Ogg et al., 1995; Ramalho, 2011).

The geology of Fogo

The geology of Fogo can be divided into four units. The stratigraphy begins with the submarine stage composed of carbonatites and alkaline basalts from approximately 4.5 Ma (Figure 4; Day et al., 1999; Madeira et al., 2010). The emergence of Fogo as an island is recorded by the subaerial lavas of the Monte Barro Group (Day et al., 1999). Subsequently a large stratovolcano known as Monte Amarelo developed during the Quaternary, reaching up to 3 km above sea level (Figure 7; Day et al., 1999). A giant landslide

occurred at approximately 73 ka, leaving the Bordiera cliffs as a remnant of the caldera wall or flank collapse scarp (Figure 1; Elsworth & Day, 1999; Masson et al., 2008; Foeken et al., 2009; Paris et al., 2011, 2018; Ramalho et al., 2015; Madeira et al., 2019). The associated structure is filled with volcanic materials to form a plateau known as the Chã das Caldeiras, which hosts the majority of the volcanic activity from 62 ka to present (Figures 4 & 7; Ribeiro, 1960; Torres et al., 1997; Day et al., 1999; Foeken et al., 2009). The spectacular Pico do Fogo, rising 2829 m above sea level formed by Hawaiian and Strombolian activity that produced basanite to tephrite lapilli and lava (Figures 5 & 7; Torres et al., 1997). Historic eruptions emanated from the base of Pico do Fogo, the eastern flank of Pico do Fogo or from vents on the Chã das Caldeiras plain. The eruptions have typical durations of 2 months, and the last eruption was 23rd November 2014 to 7th February 2015 (Figure 5; Silva et al., 2015; Fernandes & Faria, 2015; Worsley 2015). The eight previous eruptions occurred in 1785, 1799, 1816, 1847, 1852, 1857, 1951 and 1995 (Figure 4; Ribeiro, 1960; Torres et al., 1997; Day et al., 1999).

The geology of Brava

Brava is smaller than Fogo and Santiago with an area of 64 km², located 18 km from Fogo on the Brava-Fogo platform, which is only 1400 mbsl (Figure 1; Da Silva et al., 1999; Madeira et al., 2010). The geology was first investigated by Bebiano (1932) and later by Machado et al. (1968). Recent work by Madeira et al. (2010) has divided the stratigraphy into three units. The submarine volcanism consists of hyaloclastite and pillow basalt breccias and flows that radiate from the centre of the island (Figure 8; Madeira et al., 2010). They are intruded by N-S and E-W oriented dikes. Subsequently the intrusive complex was emplaced in the submarine volcanics. The alkaline-carbonatite intrusives are represented by pyroxenites, ijolite-melteigites-urites and nepheline syenites plus carbonatites (Figure 8; Madeira et al., 2010; Mourão et al., 2010). Post-erosional volcanism is dominantly phonolitic and ranges from ignimbrites and tephra deposits due to phreatomagmatic eruptions, to phonolite lava flows and domes (Figures 5 & 8). Mafic volcanism is rare in the post-erosional sections, however carbonatites in the form of ignimbrites and lava flows are not uncommon (Figure 8; Madeira et al., 2010). Seismicity, particularly NE and SE of Brava associated with volcanic cones on the Brava-Fogo platform, indicate that Brava is magmatically active. Seismic swarms at depths of 2 to 10 km, along with seismic tremor suggest volcano-tectonic events related to submarine intrusions (Faria & Fonseca, 2014).

The Cadamosto Seamount

The Cadamosto Seamount is a large circular submarine volcano, with a diameter of approximately 15 km, and a height of 3 km, reaching water depths of up to 1380 mbsl (Figure 1; Hansteen et al., 2014). The seamount has three main craters in the summit area, where pillow and sheet flows have been observed (Figure 5). The samples are mainly phononephelinites to phonolites, with occasional basaltic samples (Barker et al., 2012; Hansteen et al., 2014; Samrock et al., 2019). New ^{40}Ar - ^{39}Ar geochronology gives eruption ages of 20 to 97 ka confirming the young nature of the seamount (Samrock et al., 2019). Additionally the Cadamosto Seamount is seismically active with notable earthquakes of magnitudes 4.9 and 4.3 recorded in 1998 and 2004 respectively (Grevemeyer et al., 2010). The seismicity is interpreted to be formed by brittle rock failure resulting from submarine intrusion (Grevemeyer et al., 2010).

Volcanic eruptions and hazards at Fogo

The natural hazards faced by the population of Fogo are weather related, such as droughts that lead to famine, as well as flash floods and related landslides that destroy roads and buildings. Less common and therefore considered less of a risk are volcanic eruptions and associated earthquakes. The volcanic eruptions pose hazard to the communities living in the Chã das Caldeiras area (964 people) plus potentially the towns on the East Coast of Fogo (11,000 people). Lava flows are the greatest hazard, however volcanic gases and tephra also contribute to the volcanic hazards (Texier-Teixeira et al., 2014; Nascimento, 2015; Silva et al., 2015, 2017; Richter et al., 2016).

There have been 28 eruptions on Fogo since discovery of the island by Portuguese sailors in 1460 (Ribeiro, 1960). The most recent eruption commenced on 22nd November 2014 and continued for 77 days, finally ceasing on 7th February 2015. The eruption consisted of a 6 km high eruption cloud causing tephra deposits in eight villages and towns, as well as other islands such as Brava to the West and São Vicente to the North (Nascimento, 2015). The lava flows rapidly cut off the road access to Chã das Caldeiras and went on to destroy the villages of Portela and Bandeira including 75% of the buildings in the Chã das Caldeiras area and 25% of the agricultural land and water storage facilities (Jenkins et al., 2017).

The volcano monitoring activities carried out by the Institute of National Meteorology and Geophysics and the University of Cape Verde are essential to volcanic crisis management. Regular automated monitoring covers seismicity and deformation of the volcano (Fonseca et al., 2003; Faria & Fonesca, 2014; Dionis et al., 2015a, b; Fernandes & Faria, 2015). During the eruption in 1995, volcanic tremor and seismic swarms occurred between 1 and 5 km depths (Da Silva et al., 1999). Approximately seven weeks prior to the eruption in 2014-2015, earthquakes were detected originating at depths of >15 km (Fernandes & Faria, 2015). Additional geochemical monitoring of the volcanic gases is manual and therefore

sampling is conducted during specific campaigns and does not provide a continuous record of volcano degassing (Dionis et al., 2015a,b; Pérez et al., 2015). For instance, air quality was monitored during the 2014-2015 eruption by the deployment of an instrument at one station in São Filipe on the West coast of Fogo (Nascimento, 2015).

Magmatic processes

We use mineral chemistry from Cape Verde to explore the magmatic processes recorded by clinopyroxene (Davidson et al., 2007). Advantages of focusing on clinopyroxene include crystallization over a wide temperature range and during a large extent of magmatic differentiation. The volcanic rocks from Santiago and Fogo classify as basanite, tephrite, alkali basalt, nephelinite and melanepehlinite, with minor tephriphonolite occurring at Fogo. Whereas those from Brava and the Cadamosto Seamount are dominantly phonolite and phonotephrite combined with the presence of intrusive syenite at Brava (Figure 9; Barker et al., 2009, 2012; Hildner et al., 2011, 2012; Weidendorfer et al., 2016; Mata et al., 2017; Samrock et al., 2019).

Clinopyroxene from Santiago and Fogo classifies as diopside-augite, whereas Brava and the Cadamosto Seamount host both diopside-augite and green aegirine-augite (Barker et al., 2009, 2012; Hildner et al., 2011, 2012; Weidendorfer et al., 2016; Rydeblad, 2018). Clinopyroxene from Santiago and Fogo display compositions of dominantly 60 to 90 Mg# and 60 to 85 Mg# respectively (Figure 10; $\text{Mg\# mol\%} = (\text{Mg}/\text{Mg} + \text{Fe}^{\text{total}})$). The samples from Fogo also show minor abundances of clinopyroxene at 10 to 20 Mg# and 45 to 60 Mg#. Analysed clinopyroxene from Brava are dominantly aegirine-augite that display a much wider range of compositions from 5 to 65 Mg# (Figure 10). The nephelinites and ijolites host clinopyroxene with the highest frequency between 35 and 55 Mg#, whereas the syenites host clinopyroxene with 10 to 35 Mg# (Figure 10). The seven diopside-augite analyses have 45 to 65 Mg#. Clinopyroxene from the Cadamosto Seamount shows a skewed distribution with a range of 40 to 75 Mg# and mode at 40 to 55 Mg# (Figure 10). When the diopside-augite and aegirine-augite are distinguished, they range from 41 to 74 Mg# and 42 to 72 Mg# respectively (Figure 10). High $\text{Fe}^{3+}/\text{Fe}^{\text{total}}$ observed for volcanic rocks from Cape Verde indicates crystallization under oxidizing conditions with high oxygen fugacity (Holm et al., 2006; Duprat et al., 2007; Mata et al., 2017). This may increase the Mg# of the clinopyroxene and would likely have a more significant effect as the iron content increases, thereby having the greatest influence on the clinopyroxene compositions from Brava and the Cadamosto Seamount (Freise et al., 2003). Additionally, the presence of huyne and high sulphur contents in the Brava and Cadamosto Seamount samples would further contribute to higher Mg# contents in

clinopyroxene (Andújar et al., 2008; Barker et al., 2012; Weidendorfer et al., 2016). The clinopyroxene compositions reflect the overall composition of the volcanic centers with a restricted range of mafic volcanic rocks found at Santiago and Fogo with high clinopyroxene Mg#, therefore differentiation has played a limited role in these locations. At Brava and the Cadamosto Seamount the volcanic centers become more felsic with wider compositional ranges and lower clinopyroxene Mg#. Consequently, the clinopyroxene compositions exhibit wider ranges or skewed distributions recording the influence of extensive fractional crystallization.

More detailed examination of the mineral chemistry shows that clinopyroxene from Santiago and Fogo display increasing Al_2O_3 , TiO_2 and low but increasing Na_2O and MnO with decreasing Mg# (Figure 11). In contrast, the clinopyroxene crystals from Brava and the Cadamosto Seamount form a different trend at low and decreasing Al_2O_3 , TiO_2 with decreasing Mg#. The clinopyroxene crystals from Brava exhibit high Na_2O of 2.1 to 6.2 wt% compared to 0.3 to 1.3 wt% Na_2O in the clinopyroxene from Santiago and Fogo. At the Cadamosto Seamount Na_2O contents increase from diopside-augite with 0.8 to 2.7 wt% to the aegirine-augite population with Na_2O of 2.7 to 4.5 wt% (Figure 11). The MnO contents increase from 0.2 to 1.3 wt% in clinopyroxene with Mg# between 35 and 75 including all crystals from the Cadamosto Seamount (Figure 11). Clinopyroxene crystals with $\text{Mg}\# \leq 35$ from Brava show a decrease in MnO contents from 1.0 to 0.1 wt% as Mg# decreases (Figure 11).

The majority of the clinopyroxene crystals from Santiago and Fogo display strong correlations, with only a relatively small proportion of clinopyroxene from Fogo plotting at lower Mg# for given Al_2O_3 and TiO_2 contents (Figure 11). This suggests dominantly equilibrium growth during fractional crystallization of the magma and phenocryst formation (Mollo et al., 2013, 2018; Welsch et al., 2016). The extensive differentiation trends in the clinopyroxene chemistry and the liquid lines of descent observed in the whole rock geochemistry (Barker et al., 2009; Mata et al., 2017), suggest that equilibrium crystallization occurred during melt evolution (e.g. Mollo et al., 2018). Thus, much of the crystal population is in disequilibrium with the final host melt, and this disequilibrium is not homogenized by diffusion or mixing of magmas. Consequently, the erupted products host antecrysts crystallized under different conditions compared to the final melt compositions (e.g. Davidson et al., 2007; Barker et al., 2009; Mollo et al., 2013, 2018; Welsch et al., 2016). At Brava and the Cadamosto Seamount the clinopyroxene populations are more scattered illustrating greater compositional diversity. The equilibrium mineral-melt pairs indicate that the diopside-augite population grew in equilibrium with low Na_2O melts (<8 wt% Na_2O), whereas the host rocks have much higher Na_2O contents (8 to 14 wt% Na_2O ; Barker et al., 2012; Weidendorfer et al., 2016). Therefore despite their similar compositions, the diopside-augite and aegirine-augite populations were likely aggregated from different melts (Figure 11).

Whole rock geochemistry for Santiago, Fogo and Brava confirms significant fractional crystallization. At Santiago the full range of compositions indicate a total of 24% fractional crystallization of predominantly clinopyroxene (Barker et al., 2009). The Fogo lavas record significant fractional crystallization to produce the basanites and tephrites, which requires crystallization of up to 21% clinopyroxene and minor olivine, Fe-Ti-oxides and apatite (Magnusson, 2016; Mata et al., 2017). Likewise, the magmatic system at Brava shows 15% olivine and 35% clinopyroxene crystallization, with subordinate Fe-Ti-oxides and apatite, which explains the ijolite and nephelinite compositions (Weidendorfer et al., 2016). Furthermore, a total of 90% fractional crystallization is required to produce the entire range of whole rock compositions (Weidendorfer et al., 2016). The lavas at Santiago and a few of those from Fogo display high MgO contents implying a role for aggregation of crystals (Barker et al., 2009; Magnusson, 2016; Mata et al., 2017). These accumulated crystals must have grown in very similar magmas following the liquid line of descent for clinopyroxene crystallization shown by their Mg# and Al₂O₃ content (Figure 11).

Now we consider zoning patterns in clinopyroxene crystals and the processes they reveal. Petrographically at Santiago we frequently observe patchy cores that are rounded by resorption. Apparent in this example the resorbed, patchy core passes into an area of concentric zonation, where oscillatory zoning emphasizes the different growth rates for different sectors of the crystal (Figure 12). The associated chemistry for the zoned clinopyroxene crystal from Santiago exhibits a plateau over the patchy core with only small variations. The outer oscillatory zonation displays trends of fractional crystallization with decreasing Mg# and MnO content simultaneous with increasing TiO₂ content (Figure 12; Streck, 2008). Additionally, at Santiago the oscillatory zonation records recharge events with increasing Mg# and MnO associated with a decrease in TiO₂ content. The behaviour of Mg# and MnO are also observed to be decoupled, which indicates the occurrence of magma mixing. The selected clinopyroxene crystal from Fogo also reveals a patchy core followed by two outer zones. The fractures highlight differences in resorption of the core. The zone surrounding the core shows oscillatory zoning and then passes into a relatively homogeneous rim (Figure 12). The patchy core shows small variations in Mg#, whereas the oscillatory zone exhibits large variations in Mg# and the rim shows homogeneously high Mg# (Figure 12). The variations in TiO₂ and MnO contents are not consistent with the variations in Mg#, therefore much of the variation cannot be attributed to fractional crystallization or recharge events. Instead the zonation records magma chamber dynamics associated with temperature and compositional differences causing convection and mixing (Figure 12; Streck, 2008; Rydeblad, 2018). The clinopyroxene crystal from the Cadmosto Seamount shows a resorbed core surrounded two concentric outer zones (Figure 12). The concentric zones are homogeneous and incongruent with curved boundaries. In the associated

chemical variations, we observe a decrease in both Mg# and TiO₂ in the outer zones of the crystal, reflecting differentiation in this felsic system (Figure 11, 12). The variation in MnO content in the outer zones, despite the decrease in Mg# suggests that MnO behaves differently in felsic alkali magmas than their mafic counterparts (Streck, 2008). However the offset in composition of Mg# and TiO₂ in the two outer zones and the change in MnO also points towards magma mixing or variations in thermal conditions of the magma (>32 µm; Figure 12). The small variations in the inner part of the crystal (<27 µm), reflect variations in magma chamber compositional and thermal conditions. The zonation and textures of clinopyroxene crystals from southern Cape Verde indicate that the host magmas were not homogeneous in composition or temperature. This is traced by significant differences in behaviour between the major cation species in the mineral structure such as MgO, FeO and CaO and the trace elements TiO₂, MnO and Al₂O₃. The latter record the chemical and thermal disequilibrium of the magma chamber (Mollo et al., 2013, 2018; Welsch et al., 2016).

Magma storage in southern Cape Verde

The volcanics at Cape Verde are highly alkaline, in contrast most thermobarometers for igneous minerals are calibrated for subalkaline tholeiitic and calc-alkaline systems (e.g. Nimis, 1995; Putirka, 2008, Neave & Putirka, 2017). The clinopyroxene thermobarometers that are appropriate for alkaline systems often target mafic and ultramafic compositions (Ashchepkov et al., 2011), and are thereby not suited to the range of differentiated compositions observed at Cape Verde. The only available clinopyroxene-melt thermobarometer that has been developed for evolved alkaline systems is calibrated for relatively high K/Na (Masotta et al., 2013). Unfortunately, the lavas from Cape Verde are highly enriched in Na, giving a low K/Na ratio and thus this formulation is not ideal for clinopyroxene from Cape Verde (Barker et al., 2012, 2019). The only thermobarometer that has been calibrated for a range of alkaline compositions, providing a method for intercomparison between all samples, is the clinopyroxene-melt equilibrium thermobarometer of Putirka (Putirka et al., 1996, 2003; Putirka 2008). The Putirka clinopyroxene-melt thermobarometer has been widely applied in many alkaline settings (e.g. Klügel et al., 2000, 2005, Nikogosian et al., 2002). It has also been shown to be consistent with independent depth estimates such as seismic tomography, seismic reflection profiles or earthquakes patterns (e.g. Longpré et al., 2014; Barker et al., 2019). Unfortunately the data used by Putirka (2008) for calibration of evolved alkali compositions is sparse and therefore also not ideal (Putirka et al., 1996, 2003; Putirka 2008). Hence, we compare the results from the models of Masotta with those of Putirka (Putirka et al., 1996, 2003; Putirka 2008; Masotta et al., 2013). Therefore, we employ models A and B for pressure and

temperature respectively from Putirka et al. (2003) for consistency with Hildner et al. (2011, 2012) and additionally for the evolved compositions equations P-alk and T-alk from Masotta et al. (2013).

The clinopyroxene-melt thermobarometer is based upon jadeite (NaAl) and diopside-hedenbergite (CaMgFe) exchange between clinopyroxene and melt for determination of temperature. The corresponding pressure is estimated from the distribution of the jadeite component (NaAl) between clinopyroxene and the equilibrium melt (Putirka, 2008). Employment of the clinopyroxene-melt equilibrium thermobarometer thus requires demonstration of equilibrium between the clinopyroxene and melt. There are two commonly used approaches to test for equilibrium. Firstly, testing for equilibrium from Fe-Mg partitioning utilizing a Rhodes diagram, with $K_d(\text{Fe-Mg})$ of 0.275 ± 0.067 (Putirka, 2008; Barker et al., 2009, 2012). A second approach is to employ the variations between clinopyroxene components as observed from the mineral chemistry and calculated from the melt composition (Putirka, 2008; Mollo et al., 2013). Typically, equilibrium is considered to be within ± 0.06 DiHd, ± 0.07 EnFs and ± 0.02 Jd of unity (Putirka, 1999). Mollo et al. (2013) proposed an additional multicomponent equilibrium test, expressed as the difference between the DiHd component and tested on the alkaline compositions characteristic of Mt. Etna. The multicomponent equilibrium test derives from a comparison between the clinopyroxene components calculated from the equilibrium with the melt and directly from the clinopyroxene composition.

Ideal mineral-melt pairs are considered to be analysis of groundmass glass and proximal crystal rims. However it is rare to have much glass in lavas from Cape Verde, which tend to display a microcrystalline groundmass (Barker et al., 2009, 2012; Hildner et al., 2011, 2012). Glassy melt inclusions are typically modified during cooling of volcanic rocks and are therefore not suitable (Baker, 2008). Glass or groundmass separates would provide equilibrium melt compositions for crystal rims, representing the final growth from the melt. Consequently, whole rock compositions that include minerals and groundmass are more likely to represent core compositions (Barker et al., 2009). Employing the whole rock composition of a sample as nominal melts for testing clinopyroxene-melt equilibrium is therefore not only suitable but also necessary to understand the crystallization conditions of the entire crystal population. Equilibrium between clinopyroxene and melt can be further optimized, by subtracting the compositions of the modal mineral content from the whole rock composition (Barker et al., 2009). Another approach to find an optimal equilibrium match between clinopyroxene and melt is to select equilibrium melt compositions from a related suite of volcanic rocks (Barker et al., 2015; Neave & Putirka, 2017). The minimum ΔDiHd for data from Santiago, Fogo, Brava and the Cadamosto Seamount occurs at a temperature difference of 20°C (Figure 13). The ΔDiHd ranges from +0.02 to +0.15, meanwhile the temperature decreases to +0.4°C at ΔDiHd of +0.03 and increases up to 75°C as the ΔDiHd

increases to a maximum of +0.15. The multicomponent equilibrium appears to be equally good for evolved compositions such as at the Cadamosto Seamount as for the more mafic systems, for example Fogo.

Evaluation of the oxygen fugacity for the most recent eruptions at Fogo suggests fO_2 of +0.9 to +2.3 log ΔNNO with Fe^{3+}/Fe^{total} of 0.21 (Mata et al., 2017). These relatively oxidized conditions are confirmed by ultramafic xenoliths with fO_2 of up to +2.2 log ΔQFM (Ryabchikov et al., 1995). Other approaches have determined the Fe^{3+}/Fe^{total} to 0.19 to 0.26 at Santo Antão and São Nicolau (Holm et al., 2006; Duprat et al., 2007). Cape Verde like many Ocean Island Basalts is apparently relatively oxidized compare to Mid Ocean Ridge Basalts. Herzberg and Asimow (2008) calculate primary magma compositions for Ocean Island Basalts globally and find a range in Fe^{3+}/Fe^{total} of 0.15 to 0.25. Their findings are consistent with estimates of oxygen fugacity and Fe^{3+}/Fe^{total} at Cape Verde that represent relatively oxidized conditions. The influence of oxidation on the clinopyroxene compositions may impact the equilibrium especially that based on Mg# and clinopyroxene components that contain iron (Freise et al., 2003; Andújar et al., 2008). It is possible to vary Fe^{3+}/Fe^{total} in the thermobarometric modeling, and therefore we have adopted a Fe^{3+}/Fe^{total} of 0.21. However, the potential implications on the resulting crystallization conditions are unclear and the applicability of the calibration data to such oxidized conditions is also uncertain.

Conversion of pressure to depth has been performed with a crustal density of 2800 kg/m³ for the mafic ocean crust (Barker et al., 2009, 2012; Hildner et al., 2011, 2012). Uncertainties associated with the experimental calibration data are $\pm 33^\circ C$ and ± 0.17 GPa (SEE; Putirka, 2008), which equates to depths of ± 6 km. Hence, much of the range in pressure estimates described below are within error of the method (Figure 14).

Now turning to southern Cape Verde, the equilibrium clinopyroxene crystals from Santiago display compositions of 75 to 90 Mg#. These clinopyroxene crystals belong to alkaline basalts, basanites, tephrites and melaneophelinites and are in equilibrium, $\Delta DiHd$ of 0.01 to 0.08, with compositions in the range of 52 to 65 Mg# (Figure 9, 13; Barker et al., 2009). The corresponding crystallization conditions are temperatures of 1165 to 1255°C and pressures of 0.40 to 1.15 GPa (Figure 14; Barker et al., 2009).

From Fogo the clinopyroxene mineral chemistry shows the highest frequency compositions between 60 and 85 Mg# (Figure 10). The host rocks are tephrite, basanite, foidite and tephriphonolite and equilibrium, $\Delta DiHd$ of <0.12 and corresponding melt compositions for the main clinopyroxene population range between 30 and 54 Mg# (Figures 9 & 13; Hildner et al., 2011, 2012; Rydeblad, 2018). The resulting crystallization conditions are 1010 to 1255°C and 0.45 to 1.35 GPa, which compare well

with existing data of 1010 to 1150°C and 0.42 to 0.90 GPa for pre-eruptive magma storage conditions (Hildner et al., 2011, 2012; Mata et al., 2017). Mata et al. (2017) further divide the crystallization conditions for the 2014-2015 eruption at Fogo by host magma into 1045 to 1065°C and 0.56 to 0.78 GPa for phonotephrites and 1100 to 1145°C at 0.70 to 0.90 GPa for tephrites. This study extends the range of crystallization conditions to higher temperatures and pressures, thereby also characterizing the deeper part of the magma storage system (Figure 14; Hildner et al., 2011, 2012; Mata et al., 2017).

There are only a few analyses of diopside-augite from Brava that are hosted by nephelinite with Mg# 47 to 62 (Weidendorfer et al., 2016). They show equilibrium, ΔDiHd of 0.02 to 0.13, with low Na melt compositions of Mg# 20 to 32 (Figures 10 & 13). The Putirka et al. (2003) models estimate crystallization at 1170 to 1300°C and 0.2 to 1.0 GPa (Figure 14). In comparison the Masotta et al. (2013) model indicates crystallization at lower temperatures of 930 to 980°C and 0.29 to 0.67 GPa and shows less scatter than the estimate from Putirka et al. (2003).

Diopside-augite from the Cadamosto Seamount have 40 to 75 Mg# (Figure 10; Barker et al., 2012). The corresponding whole rocks are phonolite to phononephelinite and equilibrium melt compositions are 29 to 37 Mg# with ΔDiHd of 0.01 to 0.16 (Figures 9 & 13; Barker et al., 2012). The Putirka et al. (2003) model gives crystallization conditions of 970 to 1025°C and 0.45 to 1.35 GPa, whereas the Masotta et al. (2013) model indicates 925 to 980°C and 0.26 to 0.70 GPa (Figure 14). Hence the results from the Masotta et al (2013) model overlap with those estimated by the model of Putirka et al. (2003) and extend to lower temperature and pressure. These evolved magmas from Brava and the Cadamosto Seamount crystallized at lower temperatures compared to the clinopyroxene from Santiago and Fogo (Figure 14; Barker et al., 2012).

The clinopyroxene crystals feature significant zonation through which we track the magma chamber dynamics for samples from Fogo. Rapid decreases in clinopyroxene Mg# occur under several scenarios, sometimes with no discernable temperature or pressure changes, suggesting mixing processes (Figure 15a). A second scenario observed in the zonation is a rapid decrease in clinopyroxene Mg# with a decrease in temperature on the order of 30°C and simultaneous decrease in pressure suggesting convection and ascent within the magma reservoir (Figure 15b). In contrast, rapid increase in clinopyroxene Mg# occurs with large variations in temperature (Figure 15b). This implies that the thermal and compositional gradients are not perfectly contemporaneous and that thermal and chemical mixing occur at different rates. Whereas there are also examples of gradual increase in clinopyroxene Mg# corresponding with slight increases in temperature and pressure (Figure 15c). In some places the outer

rims record a decrease in clinopyroxene Mg#, with simultaneous decreases in temperature and pressure (Figure 15d), indicating magma evolution during ascent within the magma reservoir.

Integrating the thermobarometry from the different locations, we find that crystallization was in the range of 12 to 40 km (Figure 14 & 16). Therefore, the magma storage beneath southern Cape Verde occurs below the Moho in the oceanic lithospheric mantle. At Santiago and Fogo the highest frequency of crystallization is at 30 to 40 km (Figure 16; Barker et al., 2009; Hildner et al., 2011, 2012; Rydeblad, 2018). We also observe in the samples from Fogo that the rims crystallize with highest frequency between 20 and 30 km and extend to lower pressures approaching the Moho (Figure 14). This suggests that the more mafic magmas begin to crystallize clinopyroxene deep in the system and move upwards towards the Moho as crystallization continues and zonation and final rim compositions progressively form (Figure 15d). Evolution of magma storage depths with differentiation is consistent with the increase in Al_2O_3 contents up to 13 wt% as MgO content decreases signaling magmatic differentiation and associated with lower pressures of crystallization (Figure 11; Marianelli et al. 1999; Morgan et al., 2004). Hildner et al. (2012) proposed that the pre-eruptive magma storage conditions become shallower with time in the historic eruptions. However, no trends with time appear when we investigate the crystallization conditions of the whole range of clinopyroxene compositions. The absence of systematic temporal variations is confirmed by clinopyroxene from the 2014-2015 eruption recording pressures of 0.56 to 0.9 GPa, which are deeper than for the 1995 and 1951 eruptions (Mata et al., 2017).

The dominance of the deep magma storage zone indicates that the magmatic processes of differentiation, recharge, mixing and potentially convection recorded by the crystal populations take place in this deep magma storage zone (Figures 12, 14 & 15). The accumulation of antecrysts and aggregation of crystal populations is likely to also occur in this deep magma storage zone before magmas ascend and erupt. Thus, the resorbed kaersutite crystals frequently found in lavas from Fogo probably crystallized in the oceanic lithospheric mantle, consistent with amphibole thermobarometry yielding pressures of approximately 0.6 GPa (Mata et al., 2017). Subsequently they would have been incorporated into magmas with starkly different compositions leading to their break down (Hildner et al., 2011, 2012).

Controls on the depth of magma storage

The location of the magma storage is controlled by the crustal structure beneath the South of the Cape Verde archipelago. The islands and seamounts stand several kilometers above the seafloor at approximately 4000 mbsl and beneath them is 1 to 2 km of sediments (Figure 2; Lancelot et al., 1978;

Pim et al., 2008). The sediments overlie relatively old 130-135 Ma Central Atlantic ocean crust which extends down to the Moho at depths of 12 to 18 km, that decrease from East to West (Ali et al., 2003; Lodge & Helffrich, 2006; Pim et al., 2008). Hence the deep mantle plume source generates magmas which ascend in to the oceanic lithospheric mantle where they stagnant below the mantle-crust boundary (Figure 16; Barker et al., 2009).

The density of the ocean islands and seamounts, as well as the ocean crust extending down to the crust mantle boundary at 12 to 18 km beneath Cape Verde is estimated to be 2.75 to 2.80 g/cm³ (Ali et al., 2003; Lodge & Helffrich, 2006; Pim et al., 2008). The sediments would be expected to provide a low density layer between the mafic ocean crust and the Cape Verde islands and seamounts (Pim et al., 2008). The expected upper mantle density is 3.1 g/cm³ (Tenzer et al., 2013). The density contrast between the upper mantle and the oceanic crust evidently controls the depth and pressures of magma storage (Figure 14 & 16). The basanite, tephrite and melanephelinite at Fogo and Santiago have calculated densities of 2.65 to 2.80 g/cm³ (Bottinga & Weill, 1970; Bottinga et al., 1982). This range is similar to the mafic ocean crust and extends to slightly lower densities, therefore the mafic magmas would be expected to pool at the Moho or slowly ascend. However the actual density of magma is a function of both the melt density and the crystal cargo. Many of the erupted basanitic lavas have crystal contents of 5 to 20% and even more in ankaramite samples (Barker et al., 2009). Such crystal cargoes of dominantly clinopyroxene and olivine would increase the density of the magma, leading to a decrease in buoyancy and stagnation of the magma below the crust-mantle boundary.

Common magma storage depths, irrespective of magma composition and age are likely associated with significant magma intrusion, solidification and formation of crystal mush zones. This would serve to underplate the oceanic crust with mafic materials, building a magma storage zone of lower density than the surrounding upper mantle lithologies (Klügel et al., 2015). Evidence for this comes from Fogo lavas that host several cumulate nodules, containing olivine, clinopyroxene, kaersutite, spinel and phlogopite (Hildner et al., 2012). The spinel, kaersutite and phlogopite are consistent with crystallization in the oceanic lithospheric mantle. Therefore, these nodules are likely sampling cumulates formed in the deep magma storage zone. Furthermore, crustal thickening associated with underplating and potentially shallower intrusions contributes to localized uplift on individual islands, such as observed on Santiago and Brava (Madeira et al., 2010; Ramalho et al., 2010a, 2010b).

The phonolites and syenites at Brava and the Cadamosto Seamount have calculated densities of 2.35 to 2.48 g/cm³ and even the clinopyroxene-equilibrium melts show densities of 2.45 to 2.48 g/cm³, consistent with recent experimental constraints (Bottinga & Weill, 1970; Bottinga et al., 1982; Seifert et

al., 2013). Such densities for phonolitic magma carrying relatively low density minerals should promote magma buoyancy and ascent through the ocean crust. Although the viscosity of magma would also change during differentiation and influence the magma dynamics. The resulting increase in viscosity of the magma plus the entrained crystal population potentially inhibit magma ascent. Crystal contents of 40 to 50% cause the magma to behave as a rigid crystal network, trapping it in the magma storage zone (Cooper, 2017). The crystal content ranges from 10 to 50% for the Cadamosto Seamount, thus the crystal assemblages along with the magma evolution controls the viscosity and thereby promote continued magma storage in the upper oceanic lithospheric mantle (Barker et al., 2012).

The equilibrium clinopyroxene compositions at Brava and the Cadamosto Seamount are 45 to 75 Mg# with equilibrium melt densities of 2.45 to 2.48 g/cm³ (Figure 10). However, it is likely that the host magma carrying the existing crystal cargo, would have continued to evolve with a corresponding density change to 2.35 g/cm³. Such a change in density provides a larger density contrast between the mafic ocean crust and the most evolved phonolitic compositions that may have promoted magma ascent.

Insights into the shallower magmatic system

Evidence of magmatic processes occurring shallower in the crust at Cape Verde is limited. At the Cadamosto Seamount, the majority of the diopside-augite crystallization occurs below the Moho in the deep magma storage zone (Figure 16). However the magma differentiation and aegirine-augite crystallization is unaccounted for by this crystallization model. Furthermore, at the Cadamosto Seamount the $\delta^{18}\text{O}$ for clinopyroxene (diopside-augite) have mantle derived values of +5.3‰, whereas the feldspathoids show $\delta^{18}\text{O}$ of +6.3 to +7.1‰, elevated over the mantle equilibrium value of +6.25‰ ($\delta^{18}\text{O}$ SMOW; Zhao and Zheng 2003; Barker et al., 2012). Additionally the sulfur isotopes for whole rock powders gave $\delta^{34}\text{S}$ of +4.7 to +5.9‰, highly enriched compared to mantle values with $\delta^{34}\text{S}$ of $+0.8 \pm 0.2\%$ (Sakai et al. 1984; Barker et al., 2012). Barker et al. (2012) interpreted these feldspathoid and whole rock isotope signatures to be caused by assimilation of a few percent of oceanic sediments and anhydrite. Hence the magmas associated with the Cadamosto Seamount interact with the sediment crustal layer, which is likely located at 4 to 6 km below sea level (Figures 2 & 16; Lancelot et al., 1978; Pim et al., 2008). This evidence suggests that less differentiated magmas ascend from the deep magma storage zone and evolve to phonolitic compositions through assimilation during storage in the upper crust. Seismicity in the vicinity of the Cadamosto Seamount also records crustal magmatic activity at depths of 0.5 to 15 km, which is likely associated with shallow hydrothermal processes and slightly deeper volcano-tectonic events such as magma intrusion in the ocean crust (Figure 16; Grevenmeyer et al., 2010). The

volcano-tectonics events provide evidence for crustal magma storage, consistent with assimilation of sedimentary components by the magmas.

Brava hosts a carbonatite-silicate intrusive complex, where the silicate rock types are represented by ijolite, nephelinite and nepheline syenite (Weidendorfer et al., 2016). Extreme fractional crystallization takes place, up to 90%, which is reflected in the large range of clinopyroxene compositions of 5 to 65 Mg# (Figure 10). The presence of clinopyroxene from 3 to 90% fractional crystallization is consistent with the wide range in clinopyroxene compositions (Figure 10). Significant crystallization may have occurred in the deep magma storage zone (<30%; Weidendorfer et al., 2016), in the presence of olivine, perovskite and magnetite as illustrated by the crystallization of clinopyroxene with Mg# 47 to 62 at and below the Moho in the nephelinites from Brava (Figure 13). Given the challenges of mobilizing magmas containing more than 40 to 50% crystals (Cooper, 2017), deep crystallization is likely followed by magma ascent and intrusion into the upper crust beneath the volcanic edifice, where *in situ* differentiation and crystallization continue. Thereby crystallizing the majority of the aegirine-augite and other phases of the syenite in the ocean crust. Persistent and relatively shallow magma storage is required to provide the heat source for the active geothermal system on Brava that contains a mixture of magmatic and biogenic gases (Dionis et al., 2015a). Additionally, Brava is seismically active recording volcano-tectonic events at depths of 2 to 10 km in the ocean crust and volcanic edifice (Figure 16; Faria & Fonseca, 2014). The seismicity is consistent with continued magma differentiation in the volcanic edifice.

There is evidence for temporary magma storage at Fogo, associated with magma ascent and eruption. Olivine and clinopyroxene hosted fluid inclusions from the eruptions in 1995 and 1951 re-equilibrated at depths of 8 to 13 km (Figure 16; Hildner et al., 2011; 2012). Hildner et al. (2011, 2012) report relatively rapid re-equilibration during stagnation at these lower crustal depths over timescales of hours to days. Hence eruption related temporary magma pooling occurs in the lower ocean crust. Seismicity associated with the 1995 eruption at Fogo locates depths of magma storage between 1 and 5 km in the volcanic edifice and oceanic crust (Figure 16; Da Silva et al., 1999). Additionally, the surface deformation signals from the 1995 eruption have been explained by a 2 km deep magma source, fed from a conduit that extends to depths of more than 16 km (Figure 16; Amelung and Day, 2002). Precursory seismicity to the 2014 eruption commenced in early October before the eruption on 23rd November 2014 (Fernandes & Faria, 2015). The initial earthquakes were recorded at depths exceeding 15 km, associated with the sub-Moho magma storage zone. On 22nd November the seismicity was dominated by high frequency and magnitude long-period earthquakes at depths of 2 to 4 km (Fernandes & Faria, 2015; Jenkins et al., 2017). The 2014/2015 surface deformation is consistent with a dike propagating vertically from sub-Moho depths and arriving below Pico do Fogo where the magma flowed laterally towards the

Southwest before erupting (González et al., 2015). Therefore, the transient eruption related magma pathways can be traced from the sub-Moho magma storage zone, through the lower ocean crust to the magma pockets in the volcanic edifice feeding the magma to the eruption fissure.

Long-term magma storage in the crust is suggested by shallow seismicity and the sustained geothermal system. Fogo hosts a geothermal system with typical temperatures of 90-100°C recorded at degassing fumaroles and values up to 130 to 190°C reported in the months leading up to the 2014 eruption (Risby & Sandback 2014; Dionis et al., 2015b). The active degassing is composed of a mixture between magmatic and biogenic-atmospheric gases (Dionis et al., 2015b). Shallow seismicity at depths of 300 to 400 m below the Chã das Caldeiras plateau are confirmed by drilling to correspond to the geothermal system (Faria & Fonseca, 2014). This requires a relatively shallow long-term heat source to sustain the geothermal activity, although it has not yet been detected by geophysical methods (Caranova & Silva, 2012). The heat source may be associated with the volcano-tectonic earthquakes at depths of up to 7 km, this would imply that these earthquakes are caused by magma intrusion (Faria & Fonseca, 2014).

This information of magma storage depths and magma ascent timescales provides vital information to scientists at the Institute of National Meteorology and Geophysics and the University of Cape Verde attempting to interpret seismicity, deformation and volcanic gas geochemistry. Thereby contributing to the decision making and communication with the National Civil Protection of Cape Verde in times of impending volcanic crisis (Faria & Fonesca, 2014; Dionis et al., 2015a, b). Additional information on the volcanic hazards offered by Earth science comes in the form of volcanic hazard maps. Richter et al. (2016) have examined the probable pathways of lava flows following the 2014-2015 eruption. Their results suggest that the locations of Portela and Bandaeira continue to be a risk of future lava flows.

Ocean Islands globally

The southern chain of the Cape Verde hotspot therefore exhibits a predominantly deep magma storage zone at depths of between 12 and 40 km traced by clinopyroxene thermobarometry, seismicity and deformation (Figure 17). Magma ascent through the ocean crust is documented by fluid inclusions and seismicity. Additionally, shallow crustal magma storage is found by seismicity and deformation in the upper 5 km of the oceanic crust and volcanic edifice. Prior to the eruption in 2014/2015 seismicity traced the magma ascent from below the Moho, through the crust to the upper crust in the months leading up to the eruption (Fernandes & Faria, 2015; González et al., 2015; Jenkins et al., 2017).

A comparison with the Canary Islands, suggests that the clinopyroxene equilibrium thermobarometry and seismicity also display similarly deep magma storage relative to a Moho depth of 13 to 15 km (Figure 17; Ranero et al., 1995). Gran Canaria records magma storage at 15 to 36 km below the Moho with a subsidiary zone at 3 to 4 km (Table 1; Aulinas et al., 2010). The island of Tenerife also shows deep magma storage at 20 to 45 km (Longpré et al., 2008). Estimates for magma storage at La Palma also range from 12 to 45 km placing the magma storage within the oceanic lithospheric mantle (Klügel et al., 2000; 2005; Nikogosian et al., 2002; Galipp et al., 2006; Barker et al., 2015). Magma storage for the 2011 eruption at El Hierro has been reported to occur at depths of 17 to 24 km consistent with seismic activity (Longpré et al., 2014 and references therein). Fluid inclusions and deformation suggest magma storage in the lower crust (Longpré et al., 2014). Additionally investigation of the deformation places a magma storage zone in the upper crust (Figure 17). In the 2011 eruption offshore el Hierro the seismicity migrated upwards and laterally tracking the movement of magma supplied from sub-Moho magma storage through the crust to eruption at the seafloor (González et al., 2013).

At Madeira, magma storage from clinopyroxene equilibrium thermobarometry has been shown to occur at depths of 15 to 35 km and therefore deep relative to Moho depths of 14 to 15 km (Figure 17; Schwarz & Klügel 2004; Klügel and Klein, 2006). Furthermore, fluid inclusions record temporary stagnation in the lower and upper crust (Schwarz & Klügel 2004; Klügel and Klein, 2006).

The Azores also record deep crystallization of clinopyroxene, olivine and plagioclase as well as entrapment of primary fluid inclusions (Figure 17; Renzulli & Santi 2000; Beier et al., 2006; Dias et al., 2007; Madureira et al., 2011; Zanon et al., 2013; Zanon & Pimentel 2015). In the lower crust magma storage is recorded by continued crystallization of clinopyroxene, along with fluid inclusion equilibration and the appearance of seismicity (Renzulli & Santi 2000; Beier et al., 2006; Dias et al., 2007; Silva et al 2012; Madureira et al., 2011; Zanon et al., 2013; Zanon & Pimentel 2015). A shallow crustal magma pocket is picked out by seismicity, water solubility and plagioclase-olivine experimental phase petrology (Renzulli & Santi 2000; Silva et al 2012; Jeffery et al., 2016).

Clinopyroxene crystallization in basanites traces magma storage at Tristan da Cunha to the oceanic mantle lithosphere and for more evolved trachyandesite to the middle of the ocean crust (Figure 17; Weit et al., 2017). Earthquakes were felt on Tristan da Cunha commencing several months prior to the eruption in 1961 and again associated with the nearby submarine eruption in 2004 (O'Mangain et al., 2007). These earthquakes showed high signal to noise ratios on the nearest seismometers, thus the earthquakes were deemed to have shallow epicenters and the seismicity was employed to locate the eruption to 40 to 50 km South of Tristan da Cunha (O'Mangain et al., 2007).

Information for the magma storage system at Ascension is derived from melt inclusions, placing stagnation in the lower crust (Figure 17; Chamberlain et al., 2016). Additionally, seismic data suggests volcano-tectonic events some of which occur in the shallow crust of the Ascension rift zone (Hanson et al., 1996).

In the Pacific Ocean, the two hotspots with records of magma plumbing systematics are Hawaii and the Galapagos. Investigations at Hawaii record clinopyroxene crystallization in the oceanic lithosphere and through to the lower crust (Figure 17; Putirka et al. 1996; Putirka 1997; Chatterjee et al., 2005; Hammer et al., 2016). Whereas seismicity occurs in the upper crust beneath the central volcanoes and extends into the middle crust at rift zones (Poland et al., 2015). Wolf island in the Galapagos shows clinopyroxene crystallization in the lower crust (Stock et al., 2018). Lower crustal magma storage is confirmed by deformation, which places magma storage at 6 to 9 km depth as well as in the volcanic edifice at approximately 1 km (Stock et al., 2018).

Reunion hotspot in the Indian Ocean hosts Piton de la Fournaise volcano, which frequently displays volcanic activity, including an eruption in 2007. Magma storage at the Moho and below is traced by seismicity, as well as melt inclusions and volatile solubility which extend to much greater depths (Figure 17; Bureau et al., 1998; Peltier et al. 2009; Di Muro et al., 2014). Mid-crustal magma stagnation is indicated by melt inclusions and seismicity (Bureau et al., 1998; Peltier et al. 2009), whereas shallow magma storage below the volcanic edifice is shown by fluid inclusions, volatile solubility and seismicity above a deformation center (Figure 17; Famin et al., 2009; Peltier et al. 2009; Di Muro et al., 2014; Fontaine et al., 2014). Notably for the 2007 eruption at Piton de la Fournaise volcano, the seismicity started shallow and penetrated downwards with time, as the eruption was fed from progressively deeper (Massin et al., 2011).

Studies of the magma plumbing system associated with the volcanic islands of the Kerguelen archipelago are scarce. Partly due to their inaccessibility and relatively evolved character often hosting phonolites (e.g. Quilty & Wheller, 2000; Freise et al., 2003). However several islands seem to be volcanically active, as observed by passing vessels on long Ocean voyages (Quilty & Wheller, 2000). Volcanic eruptions have been reported from Heard Island in 1910, 1948 to 1954, 1985 to 1987 and 1996, whilst a submarine eruption likely occurred in 1992. An eruption was reported from McDonald Island in 1997 and satellite images show evidence for an eruption in 2001 (Quilty & Wheller, 2000; Wunderman, 2003; Stephenson et al., 2005). The only constraints on the magma storage suggest mid-crustal magma evolution of phonolitic magmas based on experimental petrology (Figure 17; Freise et al., 2003). Given the recent history of eruptions in the Kerguelen archipelago and potential hazard for shipping, we

recommend detailed investigation into the magma storage system. Sample material exists from several of the islands that can be used for petrological investigation (Barling et al., 1994; Freise et al., 2003) and InSAR would provide useful insights into the shallow magma storage system without the need for an expedition or deployment of equipment.

This comparison shows that deep magma storage is common to Ocean Islands, however the shallow crustal magma storage systems vary considerably and are therefore likely controlled by local factors opposed to the tectonic setting (Figure 17). Magma storage in the crust may be controlled by density differences such as the transition between the ocean crust and sediments, the presence of which will vary depending on the supply of sedimentary materials to the Ocean Islands. Sedimentary layers provide a low density zone between the mafic ocean crust and the base of the volcano, hence creating an opportunity for magmas to stall. Evidence for magma-sediment interaction is observed in lavas from Cape Verde and seismicity, plus intrusions at this level likely contribute to local uplift at Santiago and Brava (Da Silva et al., 1999; Ramalho et al., 2010a, 2010b; Barker et al., 2012; Fernandes & Faria, 2015). Magmas stalling deeper in ocean crust may be influenced by the gabbro-dike or lava-dike transitions, where porosity and permeability change. Additionally at these contacts, the ocean crust changes from massive gabbro to vertically aligned dikes and then to horizontal lava flows and random orientations of pillow lavas in the volcanic pile. Such changes potentially modifying the relationship between the magma and crust, leading to readjustment of magma flow. The characteristics of magmatic conduits will also influence magma ascent, with high permeability associated with fractures or faults aiding direct magma ascent, whereas sealing of previous pathways may enforce stalling of magma. Some Ocean Islands, such as Tenerife, Canary Islands have well developed rift zones, which may promote magma ascent (Carracedo et al., 2007). The regional fracture zones that potentially control the alignment of volcanic islands in Cape Verde likely provide effective pathways for magma ascent (Klerkx et al., 1974; Le Pichon and Fox, 1971; Jacobi and Hayes, 1982). Furthermore, magmatic properties may also influence magma dynamics on ascent through the crust, with magma buoyancy affected by density contrasts with the wall rock as well as viscosity associated with the crystal cargo and magma composition (Bottinga & Weill, 1970; Bottinga et al., 1982; Seifert et al., 2013; Cooper, 2017). This may explain storage of mafic magmas in the lower crust, as the magma density may have changed just enough for magma to ascend above the Moho and then stall in the lower crust where a similar density is encountered.

Conclusions and recommendations

Globally at Ocean Islands, we observe that clinopyroxene crystallization tends to occur deep in the magmatic system; in the oceanic lithosphere and sometimes up into the lower crust. Therefore, clinopyroxene captures the magmatic processes and magma storage that occur deep in these relatively mafic volcanic islands. Recent, well monitored eruptions of Fogo, Cape Verde (2014-2015) and el Hierro, Canary Islands (2011) have been fed from deep magma storage zones and seismicity traced the ascent of magma through the crust to the eruption site (González et al., 2013, 2015). In contrast the eruption at Piton de la Fournaise, Reunion (2007) appears to have started with fracturing of the shallow magma storage zone followed by magma ascent from progressively deeper to feed the ongoing eruption (Massin et al., 2011).

Therefore, knowledge of the deep magma storage system from clinopyroxene crystallization is appropriate for volcanoes where eruptions are fed from deep magma chambers, allowing monitoring teams to interpret enhanced deep seismicity. However for others such as Piton de la Fournaise it is essential to understand the shallow magma storage system. The presence of such long-lived shallow magmatic systems may influence the eruption style, therefore the presence or absence of protracted shallow magma storage may be a useful indicator of likely eruptive behaviour. To investigate the temporary as well as long-term shallow magma storage zones a combination of methods is recommended. Petrological methods that lend themselves to investigating the shallower magmatic system are water solubility, experimental studies, glassy melt inclusions and to some extent fluid inclusions depending on the timescales available for re-equilibration (Renzulli & Santi 2000; Freise et al., 2003; Famin et al., 2009; Zanon et al., 2013; Di Muro et al., 2014; Jeffery et al., 2016). Ideally seismicity and deformation data are used both for developing models of the magma storage system as well as real-time monitoring (Magee et al., 2018).

Investigation of timescales of magmatic processes that provide indications of typical timescales for magma ascent from a magma storage zone to eruption are also important. Such indications may be assessed on a case by case basis from the extent of re-equilibration of fluid inclusions (Hildner et al., 2011, 2012). Further information about magmatic timescales that may give insights into the magma ascent processes involved in volcanic eruptions may be gleaned by exploring diffusion between compositional zones within crystals. An example of olivine diffusion chronometry from Fogo, Cape Verde suggests rapid ascent of magmas and corresponding eruption on the order of 3 to 12 hours (Hildner et al., 2011). We recommend integration of all types of petrological data to provide indications of magmatic processes and their location and timescales with geophysical methods such as seismicity to

track magma storage and transport as well as InSAR and geodetics to reveal deformation patterns and assess source location and geometry (Magee et al., 2018).

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References:

- Ali, M.Y., Watts, A.B. and Hill, I., 2003. Structure of Mesozoic oceanic crust in the vicinity of the Cape Verde Islands from seismic reflection profiles. *Marine Geophysical Researches*, 24(3-4): 329-343.
- Ali, M.Y., and Watts, A.B., (2003). A seismic reflection profile study of lithospheric flexure in the vicinity of the Cape Verde Islands. *Journal of Geophysical Research* **108(B5)**: 2239-2263.
- Amelung, F., & Day, S. (2002). InSAR observations of the 1995 Fogo, Cape Verde, eruption: Implications for the effects of collapse events upon island volcanoes. *Geophysical Research Letters*, 29(12), 47-1.
- Andújar, J., Costa, F., Martí, J., Wolff, J.A. and Carroll, M.R., (2008). Experimental constraints on pre-eruptive conditions of phonolitic magma from the caldera-forming El Abrigo eruption, Tenerife (Canary Islands). *Chemical Geology*, 257(3-4), pp.173-191.
- Ashchepkov, I.V.; André, L.; Downes, H.; Belyatsky, B.A. (2011) Pyroxenites and megacrysts from Vitim picrite-basalts (Russia): Polybaric fractionation of rising melts in the mantle? *J. Asian Earth Sci.* 42, 14–37.
- Assunção CFT, Machado F, Gomes RAD (1965) On the occurrence of carbonatites in the Cape Verde Islands. *Boletim da Sociedade Geológica de Portugal* 16:179–188
- Aulinas M, Gimeno D, Fernandez-Turiel JL, Pérez-Torrado FJ, Rodriguez- González A, Gasperini D (2010) The Plio-Quaternary magmatic feeding system beneath Gran Canaria (Canary Islands, Spain): constraints from thermobarometric studies. *Journal of the Geological Society* 167:785-801.
- Baker, D.R. (2008) The fidelity of melt inclusions as records of melt composition. *Contrib. Mineral. Petrol.* 156, 377–395.

- 831 Barker, A.K., Nilsson, D., Hansteen, T., (2019). Unravelling the crustal architecture of Cape Verde; the xenolith
832 record. Submitted to Special Issue: Mineral Textural and Compositional Variations as a Tool for Understanding
833 Magmatic Processes editor: Silvio Mollo Minerals, 9(2), 90.
- 834 Barker, A.K., Troll, V.R., Carracedo, J.C., Nicholls, P., (2015). The magma plumbing system for the 1971 Tengüfa
835 eruption, La Palma, Canary Islands. *Contrib Mineral Petrol* 170(5) doi: 10.1007/s00410-015-1207-7
- 836 Barker, A.K., Troll, V.R., Ellam, R.M., Hansteen, T.H., Harris, C., Stillman, C.J., Andersson, A., (2012), Magmatic
837 evolution of the Cadamosto Seamount, Cape Verde: Beyond the spatial extent of EM1. *Contributions to Mineralogy*
838 and *Petrology* 163, 949-965. doi: 10.1007/s00410-011-0708-2.
- 839 Barker, A.K., Holm, P.M., Peate, D.W., & Baker, J.A., (2010), A five million year record of compositional
840 variations in mantle sources to magmatism on Santiago, southern Cape Verde archipelago. *Contributions to*
841 *Mineralogy and Petrology* doi: 10.1007/s00410-009-0470-x
- 842 Barker, A., Holm, P., Peate, D. & Baker, J., (2009). A 5 million year record of compositional variations in mantle
843 sources to magmatism on Santiago, southern Cape Verde archipelago. *Mineralogy & Petrology*, 160: 33-154.
- 844 Barling, J., Goldstein, S. L., & Nucholls, I. A. (1994). Geochemistry of Heard Island (southern Indian Ocean):
845 characterization of an enriched mantle component and implications for enrichment of the sub-Indian Ocean mantle.
846 *Journal of Petrology*, 35(4), 1017-1053.
- 847 Barrancos, J., Dionis, S., Quevedo, R., Fernandes, P., Rodríguez, F., Pérez, N., Silva, S., Cardoso, N., Hernández,
848 P., Melián, G., Padrón, E., Padilla, G., Asensio-Ramos, M., Calvo, D., Semedo, H. & Alfama, V., (2015). Sulphur
849 dioxide (SO₂) emissions during the 2014-15 Fogo eruption, Cape Verde. *Geophysical Research Abstracts* Vol. 17,
850 2015. EGU General Assembly 2015.
- 851 Bebiano, J., (1932). A geologia do Arquipélago de Cabo Verde, *Comunicações dos Serviços Geológicos de*
852 *Portugal*, (18), 167–187.
- 853 Beier, C., Haase, K.M. and Hansteen, T.H., (2006). Magma evolution of the Sete Cidades volcano, São Miguel,
854 Azores. *Journal of Petrology*, 47(7), pp.1375-1411.
- 855
- 856 Bernard-Griffiths, J., Cantagrel, J.-M., Alves, C., Mendes, F., Serralheiro, A., & Macedo, J., (1975).
857 Geochronologie: Données radiométriques potassium-argon sur quelques formations magmatiques des îles de
858 l’archipel du Cap Vert, *C. R. Seances Acad. Sci. Ser. D.*, 280, 2429–2432.
- 859 Bottinga, Y., & Weill, D. F. (1970). Densities of liquid silicate systems calculated from partial molar volumes of
860 oxide components. *American Journal of Science*, 269(2), 169-182.
- 861 Bottinga, Y., Weill, D., & Richet, P. (1982). Density calculations for silicate liquids. I. Revised method for
862 aluminosilicate compositions. *Geochimica et Cosmochimica Acta*, 46(6), 909-919.
- 863
- 864

- 865 Brown, M. C., Singer, B. S., Knudsen, M. F., Jicha, B. R., Finnes, E., & Feinberg, J. M. (2009). No evidence for
866 brunhes age excursions, santo antão, cape verde. *Earth and Planetary Science Letters*, 287(1-2), 100-115.
- 867 Bureau, H., Pineau, F., Meß trich, N., Semet, M. P. & Javoy, M. (1998). A melt and fluid inclusion study of the gas
868 phase at Piton de la Fournaise volcano (Reunion Island). *Chemical Geology* 147, 115-130.
- 869 Caranova, R., & Silva, R. P. (2012) Geothermal resource assessment in volcanic islands-Fogo Island, Cape Verde.
- 870 Carracedo, J-C., Perez-Torrado, F., Rodriguez- Gonzalez, A., Paris, R., Troll, V.R., Barker, A.K. (2015) Volcanic
871 and structural evolution of Pico do Fogo, Cape Verde. *Geology Today* 31 (4), 146-152. 10.1111/gto.12101
- 872 Carracedo, J.C., Rodríguez Badiola, E., Guillou, H., Paterne, M., Scaillet, S., Pérez Torrado, F.J., Paris, R., Fra-
873 Paleo, U., Hansen, A., (2007). Eruptive and structural history of Teide volcano and rift zones of Tenerife, Canary
874 Islands. *Geological Society of America Bulletin* 19, 1027–1051. <http://dx.doi.org/10.1130/B26087.1>.
- 875 Cazenave, A., Dominh, K., Rabinowicz, M., & Ceuleneer, G. (1988). Geoid and depth anomalies over ocean swells
876 and troughs: evidence of an increasing trend of the geoid to depth ratio with age of plate. *Journal of Geophysical*
877 *Research: Solid Earth*, 93(B7), 8064-8077.
- 878 Chamberlain, K.J., Barclay, J., Preece, K., Brown, R.J. and Davidson, J.P., (2016). Origin and evolution of silicic
879 magmas at ocean islands: Perspectives from a zoned fall deposit on Ascension Island, South Atlantic. *Journal of*
880 *Volcanology and Geothermal Research*, 327, pp.349-360.
- 881 Chatterjee, N., Bhattacharji, S. and Fein, C., (2005). Depth of alkalic magma reservoirs below Kolekole cinder cone,
882 Southwest rift zone, East Maui, Hawaii. *Journal of Volcanology and Geothermal Research*, 145(1-2), pp.1-22.
- 883 Cooper, K. M. (2017). What does a magma reservoir look like? The “crystal's-eye” view. *Elements*, 13(1), 23-28.
- 884 Courtney, R. C. & White, R.S. (1986). Anomalous heatflow and geoid across the Cape Verde Rise: Evidence for
885 dynamic support from a thermal plume in the mantle. *Geophysical Journal of the Royal Astronomical Society*. **87**:
886 815-867.
- 887 Crough, S. (1982). Geoid height anomalies over the Cape Verde Rise. *Marine Geophysical Researches*, 5, 263–271.
- 888 Da Silva, S.H., Day, S.J. and Fonseca, J.F.B.D., (1999). Fogo Volcano, Cape Verde Islands: seismicity-derived
889 constraints on the mechanism of the 1995 eruption. *Journal of volcanology and geothermal research*, 94(1-4),
890 pp.219-231.
- 891 Dash, P., Ball, M.M., King G.A., Butler, L.W. and Rona, P.A. (1976). Geophysical investigation of the Cape Verde
892 Archipelago. *Journal of Geophysical Research* **81(29)**: 5249-5259.
- 893 Davidson, J. P., Morgan, D. J., & Charlier, B. L. (2007). Isotopic microsampling of magmatic rocks. *Elements*, 3(4),
894 253-259.

- 895 Davies, G., Norry, M., Gerlach, D. & Cliff, R., (1989). A combined chemical and Pb.Sr.Nd isotope study of the
896 Azores and Cape Verde hot spots, the geodynamic implications. In: Saunders, A.D. & Norry, M.J. (eds) *Magmatism*
897 *in the Ocean Basins*. Geological Society, Special Publications, London 42, pp. 231-255.
- 898 Day, S., Heleno, S. & Fonseca, J., (1999). A past giant lateral collapse and present day flank instability of Fogo,
899 Cape Verde Islands. *Journal of Volcanology and Geothermal Research*, 94: pp. 191-218.
- 900 De Paepe, P.D., Klerkx, J., Hertogen, J. & Plinke, P., (1974). Oceanic tholeiites on Cape Verde Islands -
901 petrochemical and geochemical evidence. *Earth and Planetary Science Letters*, 22(4): 347-354.
- 902 Di Muro, A., Métrich, N., Vergani, D., Rosi, M., Armienti, P., Fougereux, T., Deloule, E., Arienzo, I., & Civetta, L.
903 (2014). The shallow plumbing system of Piton de la Fournaise Volcano (La Reunion Island, Indian Ocean) revealed
904 by the major 2007 caldera-forming eruption. *Journal of Petrology*, 55(7), 1287-1315.
- 905
906 Dias, N.A., Matias, L., Lourenço, N., Madeira, J., Carrilho, F. and Gaspar, J.L., (2007). Crustal seismic velocity
907 structure near Faial and Pico Islands (AZORES), from local earthquake tomography. *Tectonophysics*, 445(3-4),
908 pp.301-317.
- 909
910 Dionis, D., Melián, G., Rodríguez, R., Hernández, P., Padrón, E., Pérez, N., Barrancos, B., Padilla, G., Sumino, H.,
911 Fernandes, P., Bandomo, Z., Victória, S., Pereira, J. & Semedo, H., (2015b). Diffuse volcanic gas emission and
912 thermal energy release from the summit crater of Pico do Fogo, Cape Verde. *Bulletin of Volcanology*. Bull
913 *Volcanol.* 77:10.
- 914 Dionis, S., Pérez, N., Hernández, P., Melián, G., Rodríguez, F., Padrón, E., Sumino, H., Barrancos, J., Padilla, G.,
915 Fernandes, P., Bandomo, Z., Silva, S., Pereira, J. & Semedo, H. (2015a) Diffuse CO₂ degassing and volcanic activity
916 at Cape Verde islands, West Africa. *Earth, Planets and Space* 67:48.
- 917 Doucelance, R., Escrig, S., Moreira, M., Gariépy, C. and Kurz, M. D., (2003). Pb-Sr-He isotope and trace element
918 geochemistry of the Cape Verde Archipelago. *Geochimica et Cosmochimica Acta*, 67(19), 3717-3733.
- 919 Duprat, H., Friis, J., Holm, P., Grandvuiet, T., & Sørensen, R., (2007). The volcanic and geochemical development
920 of São Nicolau, Cape Verde Islands: Constraints from field and ⁴⁰Ar/³⁹Ar evidence, *Journal of Volcanology and*
921 *Geothermal Research*, 162(1-2), 1 – 19.
- 922 Dyhr CT, Holm PM (2010) A volcanological and geochemical investigation of Boa Vista, Cape Verde Islands;
923 ⁴⁰Ar/³⁹Ar geochronology and field constraints. *J Volcanol Geotherm Res* 189:19–32
- 924 Efimov, V. N., & Skolotnev, S. G. (2006). New data on the structure of sedimentary cover at the Cape Verde
925 Seamount and the Cape Verde Basin (central Atlantic) based on continuous seismic profiling data. In *Doklady earth*
926 *sciences* (Vol. 407, No. 2, pp. 376-380). MAIK Nauka/Interperiodica.
- 927 Egloff, J. (1972). Morphology of ocean basin seaward of northwest Africa: Canary Islands to Monrovia, Liberia.
928 *AAPG Bulletin*, 56(4), 694-706.

- 929 Eisele, S., Reißig, S., Freundt, A., Kutterolf, S., Nürnberg, D., Wang, K. L., & Kwasnitschka, T. (2015). Pleistocene
930 to Holocene offshore tephrostratigraphy of highly explosive eruptions from the southwestern Cape Verde
931 Archipelago. *Marine Geology*, 369, 233-250.
- 932 Elsworth, D., & Day, S. J. (1999). Flank collapse triggered by intrusion: the Canarian and Cape Verde
933 Archipelagoes. *Journal of Volcanology and Geothermal Research*, 94(1-4), 323-340.
- 934 Famin, V., Welsch, B., Okumura, S., Bachèlery, P., & Nakashima, S. (2009). Three differentiation stages of a single
935 magma at Piton de la Fournaise volcano (Reunion hot spot). *Geochemistry, Geophysics, Geosystems*, 10(1).
936
- 937 Faria, B. & Fonseca, J.F.B.D. (2014). Investigating volcanic hazard in Cape Verde Islands through geophysical
938 monitoring: network description and first results. *Natural Hazards and Earth System Sciences*, v.14, pp.485–494.
- 939 Faria, B. (2010) Monitorização Geofísica do Vulcão do Fogo e Níveis de Alerta, Ph.D thesis, Universidade Técnica
940 de Lisboa, (in Portuguese).
- 941 Faugères, J., Legigan, P., Maillet, N., and Latouche, C., (1989), Pelagic, turbiditic, and contouritic sequential
942 deposits on the Cape Verde Plateau (Leg 108, Site 659, Northwest Africa): Sediment record during Neogene time:
943 Proceedings of the Ocean Drilling Program, Scientific Results, v. 108, p. 311–327.
- 944 Fernandes, R., & Faria, B. (2015). FOGO-2014: monitoring the Fogo 2014 eruption, Cape Verde. In EGU General
945 Assembly Conference Abstracts Vol. 17 id.12709
- 946 Foeken, J., Day, S., & Stuart, F., (2009). Cosmogenic ³He exposure dating of the Quaternary basalts from Fogo,
947 Cape Verdes: Implications for rift zone and magmatic reorganisation, *Quaternary Geochronology*, 4(1), 37–49.
- 948 Fonseca, J. F., Faria, B. V., Lima, N. P., Heleno, S. I., Lazaro, C., d'Oreye, N. F., Ferreira, A.M.G, Barros, I.J.M.,
949 Santos, P., Bandomo, Z., Day, S. J., Osorio, J.P., Baio, M., Matis, J.L.G (2003). Multiparameter monitoring of Fogo
950 Island, Cape Verde, for volcanic risk mitigation. *Journal of volcanology and geothermal research*, 125(1-2), 39-56.
- 951 Fontaine, F. R., Roult, G., Michon, L., Barruol, G., & Muro, A. D. (2014). The 2007 eruptions and caldera collapse
952 of the Piton de la Fournaise volcano (La Réunion Island) from tilt analysis at a single very broadband seismic
953 station. *Geophysical Research Letters*, 41(8), 2803-2811.
954
- 955 Freise, M., Holtz, F., Koepke, J., Scoates, J. and Leyrit, H., (2003). Experimental constraints on the storage
956 conditions of phonolites from the Kerguelen Archipelago. *Contributions to Mineralogy and Petrology*, 145(6),
957 pp.659-672.
- 958 Galipp K, Klügel A, Hansteen TH (2006) Changing depths of magma fractionation and stagnation during the
959 evolution of an oceanic island volcano: La Palma (Canary Islands). *Journal Volcanol Geotherm Res* 155:285–306
- 960 Geiger, H., Barker, A.K., Troll, V.R., (2016). Locating the depth of magma supply for volcanic eruptions, insights
961 from Mt. Cameroon. *Sci. Rep.* 6, 33629; doi: [10.1038/srep33629](https://doi.org/10.1038/srep33629)
- 962 Geissler, W.H., Jokat, W., Jegen, M., Baba, K., (2016). Thickness of the oceanic crust and the mantle transition zone
963 in the vicinity of the Tristan da Cunha hot spot estimated from ocean-bottom and ocean-island seismometer receiver
964 functions. *Tectophysics this volume*.
- 965 Gerlach, D.C., Cliff, R.A., Davies, G.R., Norry, M. and Hodgson, N., (1988). Magma sources of the Cape-Verdes
966 Archipelago - isotopic and trace-element constraints. *Geochimica Et Cosmochimica Acta*, 52(12): 2979-2992.

- 967 González, P. J., Bagnardi, M., Hooper, A. J., Larsen, Y., Marinkovic, P., Samsonov, S. V., & Wright, T. J. (2015).
968 The 2014–2015 eruption of Fogo volcano: Geodetic modeling of Sentinel- 1 TOPS interferometry. *Geophysical*
969 *research letters*, 42(21), 9239-9246.
- 970 González, P.J., Samsonov, S.V., Pepe, S., Tiampo, K.F., Tizzani, P., Casu, F., Fernández, J., Camacho, A.G. and
971 Sansosti, E., (2013). Magma storage and migration associated with the 2011–2012 El Hierro eruption: implications
972 for crustal magmatic systems at oceanic island volcanoes. *Journal of Geophysical Research: Solid Earth*, 118(8),
973 pp.4361-4377.
- 974 Grevemeyer I, Helffrich G, Faria B, Booth-Rea G, Schnabel M, Weinrebe RW (2010) Seismic activity at Cadamosto
975 seamount near Fogo Island, Cape Verde—formation of a new ocean island? *Geophys J Int* 180:552–558.
976 doi:10.1111/j.1365-246X. 2009.04440.x
- 977 Gudmundsson, A. (2012). Magma chambers: Formation, local stresses, excess pressures, and compartments. *Journal*
978 *of Volcanology and Geothermal Research*, 237, 19-41.
- 979 Hammer, J., Jacob, S., Welsch, B., Hellebrand, E. and Sinton, J., (2016). Clinopyroxene in postshield Haleakala
980 ankaramite: 1. Efficacy of thermobarometry. *Contributions to Mineralogy and Petrology*, 171(1), p.7.
- 981 Hanson, J. A., Given, H. K., & Berger, J. (1996). Earthquake activity near Ascension Island, South Atlantic Ocean,
982 as seen by a combined seismic/hydrophone array. *Geothermics*, 25(4-5), 507-519.
- 983
- 984 Hansteen, T., Kwasnitschka, T., Klügel, A., (2014). Cape Verde Seamounts - Cruise No. M80/3: December 29, 2009
985 - February 1, 2010 - Dakar (Senegal) - Las Palmas de Gran Canaria (Spain). DFG-Senatskommission für
986 Ozeanographie, Bremen, 1–42. doi:10.2312/cr_m80_3, 42 pp.
- 987 Hayes, D. E. and Rabinowitz, P. D., (1975), Mesozoic Magnetic Lineations and the Magnetic Quiet Zone off
988 Northwest Africa, *Earth Planet. Sci. Lett.* **28**, 105–115.
- 989 Heleno, S.I.N., Faria, B.V.E., Bandomo, Z. & Fonseca, J.F.B.D., (2006). Observations of high-frequency harmonic
990 tremor in Fogo, Cape Verde Islands, *J. Volc. Geotherm. Res.*, **158**, 361–379.
- 991 Herzberg, C.; Asimow, P. D. (2008) Petrology of some oceanic island basalts: PRIMELT2. XLS software for
992 primary magma calculation. *Geochemistry, Geophysics, Geosystems*, 9(9).
- 993 Hildner, E.; Klügel, A.; Hansteen, T. (2012) Barometry of lavas from the 1951 eruption of Fogo, Cape Verde
994 Islands: Implications for historic and prehistoric magma plumbing systems. *Journal of Volcanology and Geothermal*
995 *Research* 217-218 73–90.
- 996 Hildner, E.; Klügel, A.; Hauff F. (2011) Magma storage and ascent during the 1995 eruption of Fogo, Cape Verde
997 archipelago. *Contributions to Mineralogy and Petrology*, 162, 751-772, doi:10.1007/s00410-011-0623-6.
- 998 Hoernle K. Tilton G, Le Bas MJ, Duggen S, Garbe-Schönberg D (2002) Geochemistry of oceanic carbonatites
999 compared with continental carbonatites: mantle recycling of oceanic crustal carbonate. *Contributions to Mineralogy*
1000 *and Petrology*, 142(5), 520-542.
- 1001 Holm, P. M., Grandvuinet, T., Friis, J., Wilson, J. R., Barker, A. K. and Plesner, S., (2008). An ⁴⁰Ar/³⁹Ar study of
1002 the Cape Verde hot spot: Temporal evolution in a semistationary plate environment. *Journal of Geophysical*
1003 *Research: Solid Earth* (1978 – 2012), 113 (B8).

- 1004 Holm, P.M.; Wilson, J.R.; Christensen, B.P.; Hansen, L.; Hansen, S.L.; Hein, K.H.; Mortensen, A.K.; Pedersen, R.;
1005 Plesner, S.; Runge, M.K. (2006) Sampling the Cape Verde mantle plume: Evolution of melt compositions on Santo
1006 Antão, Cape Verde Islands. *Journal of Petrology*, 47, 145-189.
- 1007 Jacobi, R. D., & Hayes, D. E. (1982). Bathymetry, microphysiography and reflectivity characteristics of the West
1008 African margin between Sierra Leone and Mauritania. In *Geology of the Northwest African continental margin* (pp.
1009 182-212). Springer, Berlin, Heidelberg.
- 1010 Jeffery, A. J., Gertisser, R., O'Driscoll, B., Pacheco, J. M., Whitley, S., Pimentel, A., & Self, S. (2016). Temporal
1011 evolution of a post-caldera, mildly peralkaline magmatic system: Furnas volcano, São Miguel, Azores.
1012 *Contributions to Mineralogy and Petrology*, 171(5), 42.
- 1013 Jenkins, S. F., Day, S. J., Faria, B. V. E., & Fonseca, J. F. B. D. (2017). Damage from lava flows: insights from the
1014 2014–2015 eruption of Fogo, Cape Verde. *Journal of Applied Volcanology*, 6(1), 6.
- 1015 Jørgensen, J. & Holm, P. (2002). Temporal variation and carbonatite contamination in primitive ocean island
1016 volcanic from São Vicente. Cape Verde Islands. *Chemical Geology* (192), pp. 249–267.
- 1017 Klerkx, J., Deutsch, S., and De Paep, P., (1974). Rubidium, strontium content and strontium isotopic composition
1018 of strongly alkalic basaltic rocks from the Cape Verde Islands. *Contributions to Mineralogy and Petrology* 45: 107-
1019 118.
- 1020 Klingelhöfer, F., Minshull, T.A., Blackman, D.K., Harben, P. and Childers, V., (2001). Crustal structure of
1021 Ascension Island from wide-angle seismic data: implications for the formation of near-ridge volcanic islands. *Earth*
1022 *and Planetary Science Letters*, 190(1-2), pp.41-56.
- 1023 Klitgord, K., & Schouten, H. (1987). Plate kinematics of the Central Atlantic: in Tucholke, BE, and Vogt, PR, eds.
1024 *The Geology of North America: the Western Atlantic Region: Geol. Soc, America, Decade of North American*
1025 *Geology*.
- 1026 Klügel, A., Klein, F., (2006). Complex magma storage and ascent at embryonic subma- rine volcanoes from
1027 Madeira Archipelago. *Geology* 34, 337–340.
- 1028 Klügel, A., Longpré, M. A., García-Cañada, L., & Stix, J. (2015). Deep intrusions, lateral magma transport and
1029 related uplift at ocean island volcanoes. *Earth and Planetary Science Letters*, 431, 140-149.
- 1030 Klügel A, Galipp K, Hansteen TH (2005) Magma storage and underplating beneath Cumbre Vieja volcano, La
1031 Palma (Canary Islands). *Earth Planet Sci Lett* 236:211–226
- 1032 Klügel A, Hoernle KA, Schmincke HU, White JDL (2000) The chemically zoned 1949 eruption on La Palma
1033 (Canary Islands): petrologic evolution and magma supply dynamics of a rift- zone eruption, *J Geophys Res*
1034 105(B3):5997–6016
- 1035 Kokfelt, T. F., Holm, P.M., Hawkesworth, C.J., Peate, D.W. (1998). A lithospheric mantle source for the Cape
1036 Verde Island magmatism: Trace element and isotopic evidence from the Island of Fogo. *Mineralogical Magazine*
1037 62A, 801-802.
- 1038 Lai, Z., Zhao, G., Han, Z., Huang, B., Li, M., Tian, L., Bu, X. (2018). The magma plumbing system in the Mariana
1039 Trough back-arc basin at 18° N. *Journal of Marine Systems*, 180, 132-139.
- 1040 Lancelot, Y., Seibold, E., Cepek, P., Dean, W., Ereemeev, V., Gardner, J., Jansa, L., Johnson, D., Krashennnikov,
1041 V., Pflaumann, U., Rankin, J. G., Trabant, P. and Bukry, D., (1978). Site 368: Cape Verde Rise, in Initial Rep. Deep
1042 Sea Drill. Proj., vol. 41, pp. 233–326, eds Lancelot, Y., Seibold, E., Cepek, P., Dean, W., Ereemeev, V., Gardner, J.,
1043 Jansa, L., Johnson, D., Krashenin- Nikov, V., Pflaumann, U., Rankin, J. G., Trabant, P. and Bukry, D., Washington
1044 (U.S. Government Printing Office).

- 1045 Le Maitre RW, Streckeisen A, Zanettin B, Le Bas MJ, Bonin B, Bate- man P, Bellieni G, Dudek A, Efremova A,
1046 Keller J, Woolley AR (2002) Igneous rocks. A classification and glossary of terms, Recommendations of the IUGS
1047 subcommission on the systematics of igneous rocks
- 1048 Le Pichon, X.L., and Fox, P.J., (1971). Marginal offsets, fracture zones and the early opening of the North Atlantic.
1049 *Journal of Geophysical Research* **75**: 6294-6307.
- 1050 Leva, C., Rümper, G., Link, F. and Wölbern, I., (2019). Seismological evidence for subcrustal magmatic injection
1051 beneath Fogo volcano, Cape Verde hotspot. EarthArXiv 10.31223/osf.io/f6zdp
- 1052 Liu, C., Lay, T., & Xiong, X. (2018). Rupture in the 4 May 2018 MW 6.9 earthquake seaward of the Kilauea east
1053 rift zone fissure eruption in Hawaii. *Geophysical Research Letters*, 45(18), 9508-9515.
- 1054 Lodge, A. and Helffrich, G., (2006). Depleted swell root beneath the Cape Verde Islands. *Geology*, 34(6): 449-452.
- 1055 Longpré, M. A., Klügel, A., Diehl, A., & Stix, J. (2014). Mixing in mantle magma reservoirs prior to and during the
1056 2011–2012 eruption at El Hierro, Canary Islands. *Geology*, 42(4), 315-318.
- 1057 Longpré, M. A., Troll, V. R., & Hansteen, T. H. (2008). Upper mantle magma storage and transport under a
1058 Canarian shield- volcano, Teno, Tenerife (Spain). *Journal of Geophysical Research: Solid Earth*, 113(B8).
- 1059 Machado F, Azeredo Leme J, Monjardino J, Seita MF (1968) Carta geológica de Cabo Verde, notícia explicativa da
1060 Ilha Brava e dos Ilhéus Secos. Garcia de Orta 16:123–130
- 1061 Madeira, J., Ramalho, R. S., Hoffmann, D. L., Mata, J., & Moreira, M. (2019) A geological record of multiple
1062 Pleistocene tsunami inundations in an oceanic island: The case of Maio, Cape Verde. *Sedimentology*.
1063 <https://doi.org/10.1111/sed.12612>
- 1064 Madeira J, Mata J, Mourão C, da Silveira AB, Martins S, Ramalho R, Hoffmn DL (2010) Volcano-stratigraphy and
1065 structural evolution of Brava Island (Cape Verde) based on 40Ar/39Ar, U-Th and field constraints. *J Volcanol*
1066 *Geotherm Res* 196:219–235
- 1067 Madeira, J., Brum da Silveira, A., Mata, J., Mourão, C., & Martins, S. (2008). The role of mass movements on the
1068 geomorphologic evolution of island volcanoes: examples from Fogo and Brava in the Cape Verde archipelago.
1069 *Comun. Geol*, 95, 93-106.
1070
- 1071 Madureira, P., Mata, J., Mattielli, N., Queiroz, G., and Silva, P. (2011) Mantle source heterogeneity, magma
1072 generation and magmatic evolution at Terceira Island (Azores archipelago): Constraints from elemental and isotopic
1073 (Sr, Nd, Hf, and Pb) data. *Lithos*, 126, 402–418.
- 1074 Madureira, P., Mata, J., & Moreira, M. (2008). The Ne isotopic signature of Terceira lavas (Azores): evidence for
1075 Ne recycling?. *e-Terra*, 5(11).
- 1076 Magee, C., Stevenson, C. T., Ebmeier, S. K., Keir, D., Hammond, J. O., Gottsmann, J. H., O’Driscoll, B. (2018).
1077 Magma plumbing systems: a geophysical perspective. *Journal of Petrology*, 59(6), 1217-1251.
- 1078 Magnusson, E., (2016). Temporal evolution of historic mafic lavas from Fogo, Cape Verde. Examensarbete vid
1079 Institutionen för geovetenskap, ISSN 1650-6553 ; 365. urn:nbn:se:uu:diva-296417
- 1080 Marianelli, P., Metrich, N., Sbrana, A. (1999) Shallow and deep reservoirs involved in magma supply of the 1944
1081 eruption of Vesuvius. *Bulletin of Volcanology*, 61, 48-63

- 1082 Martínez-Moreno, F. J., Santos, F. M., Madeira, J., Pous, J., Bernardo, I., Soares, A., Esteves, M., Adão, F., Ribeiro,
1083 J., Mata, J., & da Silveira, A. B. (2018). Investigating collapse structures in oceanic islands using magnetotelluric
1084 surveys: The case of Fogo Island in Cape Verde. *Journal of Volcanology and Geothermal Research*, 357, 152-162.
- 1085 Masotta, M.; Mollo, S.; Freda, C.; Gaeta, M.; Moore, G. (2013) Clinopyroxene–liquid thermometers and barometers
1086 specific to alkaline differentiated magmas. *Contributions to Mineralogy and Petrology*, 166(6), 1545-1561.
- 1087 Massin, F., Ferrazzini, V., Bachèlery, P., Nercessian, A., Duputel, Z., & Staudacher, T. (2011). Structures and
1088 evolution of the plumbing system of Piton de la Fournaise volcano inferred from clustering of 2007 eruptive cycle
1089 seismicity. *Journal of Volcanology and Geothermal Research*, 202(1-2), 96-106.
- 1090 Masson, D.G., Le Bas, T.P., Grevemeyer, I. & Weinrebe, W., (2008). Flank collapse and large-scale landsliding in
1091 the Cape Verde Islands, off West Africa, *Geochem. Geophys. Geosyst.*, **9**, Q07015, doi:10.1029/2008GC001983.
- 1092 Mata, J., Martins, S., Mattielli, N., Madeira, J., Faria, B., Ramalho, R. S., Silva, P., Moreira, M., Caldeira, R.,
1093 Moreira, M., Rodrigues, J., & Martins, L. (2017). The 2014–15 eruption and the short-term geochemical evolution
1094 of the Fogo volcano (Cape Verde): Evidence for small-scale mantle heterogeneity. *Lithos*, 288, 91-107.
- 1095 Mata, J., Moreira, M., Doucelance, R., Ader, M. and Silva, L. C., (2010). Noble gas and carbon isotopic signatures
1096 of Cape Verde oceanic carbonatites: implications for carbon provenance. *Earth and Planetary Science Letters*,
1097 291(1), 70-83.
- 1098 McBride, J. H., White, R. S., Henstock, T. J., & Hobbs, R. W. (1994). Complex structure along a Mesozoic sea-floor
1099 spreading ridge: BIRPS deep seismic reflection, Cape Verde abyssal plain. *Geophysical Journal International*,
1100 119(2), 453-478.
- 1101 McNutt, M. K., (1988) Thermal and mechanical properties of the Cape Verde Rise, *J. Geophys. Res.*, 93, 2784–
1102 2794.
- 1103 Mitchell, J. G., M. J. LeBas, J. Zielonka, and H. Furnes (1983), On dating the magmatism of Maio, Cape Verde
1104 Islands, *Earth Planet. Sci. Lett.*, 64, 61–76.
- 1105 Mollo, S., Blundy, J., Scarlato, P., De Cristofaro, S. P., Tecchiato, V., Di Stefano, F., Vetere, F., Holtz, F., &
1106 Bachmann, O. (2018). An integrated PT-H₂O-lattice strain model to quantify the role of clinopyroxene fractionation
1107 on REE+ Y and HFSE patterns of mafic alkaline magmas: Application to eruptions at Mt. Etna. *Earth-Science*
1108 *Reviews*, 185, 32-56.
- 1109 Mollo, S., Putirka, K., Misiti, V., Soligo, M., & Scarlato, P. (2013). A new test for equilibrium based on
1110 clinopyroxene–melt pairs: clues on the solidification temperatures of Etnean alkaline melts at post-eruptive
1111 conditions. *Chemical Geology*, 352, 92-100.
- 1112 Morgan, D. J., Blake, S., Rogers, N. W., DeVivo, B., Rolandi, G., Macdonald, R., & Hawkesworth, C. J. (2004).
1113 Timescales of crystal residence and magma chamber volume from modelling of diffusin profiles in phenocrysts:
1114 Vesuvius 1944. *Earth and Planetary Science Letters* **222**, 933-946, doi:10.1016/j.epsl.2004.03.030
- 1115 Mourão, C., Mata, J., Doucelance, J., Madeira, J., Brum da Silveira, A., Silva, L. & Moreira, M. (2010). Quaternary
1116 extrusive calciocarbonatite volcanism on Brava Island (Cape Verde). *Journal of African Earth Sciences*, 55, 59–74.
- 1117 Müller, R., Sdrolias, M., & Roest, W., (2008). Age, spreading rates and spreading symmetry of the world's ocean
1118 crust, *Geochemistry Geophysics Geosystems*, 9(Q04006).
- 1119 Nascimento A. (2015) Post disaster needs assessment: Fogo volcanic eruption 2014-2015.
1120 <https://www.gfdrr.org/en/cape-verde-post-disaster-needs-assessment-volcano-eruption-2014-2015>

- 1121 Neave, D. A., & Putirka, K. D. (2017). A new clinopyroxene-liquid barometer, and implications for magma storage
1122 pressures under Icelandic rift zones. *American Mineralogist*, 102(4), 777-794.
- 1123 Nikogosian IK, Elliott T, Touret JLR (2002) Melt evolution beneath thick lithosphere: a magmatic inclusion study
1124 of La Palma, Canary Islands. *Chemical Geology* 183: 169-193.
- 1125 Nimis, P. (1995). A clinopyroxene geobarometer for basaltic systems based on crystal-structure modelling.
1126 *Contributions to Mineralogy and Petrology* **121**: 115-125.
- 1127 O'Mongain, A., Ottemoller, L., Baptie, B., Galloway, D., & Booth, D. (2007). Seismic activity associated with a
1128 probable submarine eruption near Tristan da Cunha, July 2004–July 2006. *Seismological Research Letters*, 78(3),
1129 375-382.
- 1130 Ogg, J.G., (1995). Magnetic polarity time scale of the Phanerozoic, in *Global Earth Physics: A handbook of physical*
1131 *constants. AGU Ref. Shelf, vol. 1*, edited by T.J. Ahrens, p 240-270. Washington D.C.
- 1132 Paris, R., Giachetti, T., Chevalier, J., Guillou, H. and Frank, N., (2011). Tsunami deposits in Santiago Island (Cape
1133 Verde archipelago) as possible evidence of a massive flank failure of Fogos volcano. *Sedimentary Geology*, 239(3-
1134 4), pp.129-145.
- 1135 Paris, R., Ramalho, R. S., Madeira, J., Ávila, S., May, S. M., Rixhon, G., Engel, M., Brückner, H., Herzog, M.,
1136 Schukraft, G., Pérez-Torrado, F. J., Rodrigues-González, A., Carracedo, J.C. & Giachetti, T. (2018). Mega-tsunami
1137 conglomerates and flank collapses of ocean island volcanoes. *Marine Geology*, 395, 168-187.
- 1138 Peltier, A., Bachèlery, P. and Staudacher, T., (2009). Magma transport and storage at Piton de La Fournaise (La
1139 Réunion) between 1972 and 2007: A review of geophysical and geochemical data. *Journal of Volcanology and*
1140 *Geothermal Research*, 184(1-2), pp.93-108.
- 1141 Pérez, M., Dionis, S., Fernandes, P., Barrancos, J., Rodríguez, F., Bandomo, Z., Hernández, P., Melián, G., Silva,
1142 S., Padilla, G., Padrón, E., Cabral, J., Calvo, D., Asensio-Ramos, M., Pereira, J., Gonçalves, A., Barros, I., &
1143 Semedo, H., (2015). Precursory signals of the 2014-15 Fogo eruption (Cape Verde) detected by surface CO₂
1144 emission and heat flow observations. *Geophysical Research Abstracts* Vol. 17, 2015, EGU General Assembly 2015.
- 1145 Pim, J., Peirce, C., Watts, A. B., Grevemeyer, I. and Krabbenhöft, A., (2008). Crustal structure and origin of the
1146 Cape Verde Rise. *Earth and Planetary Science Letters*, 272(1), 422-428.
- 1147 Plesner, S., Holm, P. M., & Wilson, J. R. (2003). 40Ar–39Ar geochronology of Santo Antão, Cape Verde Islands.
1148 *Journal of Volcanology and Geothermal Research*, 120(1-2), 103-121.
- 1149 Poland, M.P., Miklius, A. and Montgomery-Brown, E.K., (2015). Magma supply, storage, and transport at shield-
1150 stage. *Charact. Hawaii. Volcanoes*, 179(1801), pp.179-234.
- 1151 Putirka, K. D. (2017). Down the crater: where magmas are stored and why they erupt. *Elements*, 13(1), 11-16.
- 1152 Putirka, K. (2008) Thermometers and barometers for volcanic systems. *Reviews in Mineralogy and Geochemistry*,
1153 69, 61-120.
- 1154 Putirka K, Mikaelian H, Ryerson FJ, Shaw H (2003) New clinopyroxene-liquid thermobarometers for mafic,
1155 evolved and volatile-bearing lava compositions, with applications to lavas from Tibet and the Snake River Plain, ID.
1156 *Am Mineral* 88:1542–1554
- 1157 Putirka K (1999) Clinopyroxene + liquid equilibria to 100 kbar and 2450 K. *Contrib Mineral Petrol* 135: 151-163

- 1158 Putirka K (1997) Magma transport at Hawaii: Inferences based on igneous thermobarometry. *Geology* 25: 69-72
- 1159 Putirka K, Johnson M, Kinzler R, Longhi J, Walker D (1996) Thermobarometry of mafic igneous rocks based on
1160 clinopyroxene-liquid equilibria, 0-30 kbar. *Contrib Mineral Petrol* 123: 92-108
- 1161 Putirka K, Condit C (2003) A cross section of a magma conduit system at the margins of the Colorado Plateau.
1162 *Geology* 31:701-704
- 1163 Quilty, P. G., & Wheller, G. E. (2000). Heard Island and the McDonald Islands: a window into the Kerguelen
1164 Plateau. In *Papers and Proceedings of the Royal Society of Tasmania* (Vol. 133, No. 2, pp. 1-12).
- 1165 Ranero CR, Torne M, Banda E (1995) Gravity and multichannel seismic reflection constraints on the lithospheric
1166 structure of the Canary Swell. *Mar Geophys Res* 17:519–534
- 1167 Ramalho, R. S., Winckler, G., Madeira, J., Helffrich, G. R., Hipólito, A., Quartau, R., Adena, K. & Schaefer, J. M.
1168 (2015). Hazard potential of volcanic flank collapses raised by new megatsunami evidence. *Science advances*, 1(9),
1169 e1500456.
- 1170 Ramalho, R. A. dos Santos (2011). Vertical Movements of Ocean Island Volcanoes: Insights from a Stationary
1171 Plate. In *Building the Cape Verde Islands* (pp. 177-197). Springer, Berlin, Heidelberg.
- 1172 Ramalho, R. S., Helffrich, G., Cosca, M., Vance, D., Hoffmann, D., & Schmidt, D. N. (2010c). Vertical movements
1173 of ocean island volcanoes: insights from a stationary plate environment. *Marine Geology*, 275(1-4), 84-95.
- 1174 Ramalho, R., Helffrich, G., Schmidt, D. & Vance, D. (2010b). Tracers of uplift and subsidence in the Cape Verde
1175 Archipelago. *J. Geol. Soc. Lond.* 167, pp. 519-538.
- 1176 Ramalho, R., Helffrich, G., Cosca, M., Vance, D., Hoffmann, D. & Schmidt, D. (2010a). Episodic swell growth
1177 inferred from variable uplift of the Cape Verde hotspot islands. *Nature Geoscience*, Vol. 3, pp. 774-777.
- 1178 Renzulli, A. and Santi, P., (2000). Two-stage fractionation history of the alkali basalt-trachyte series of Sete Cidades
1179 volcano (São Miguel Island, Azores). *European Journal of Mineralogy*, 12(2), pp.469-494.
- 1180 Ribeiro, O. (1960). A ilha do Fogo e as suas Erupções. Memórias Série Geográfica I. Lisboa: Junta de Investigações
1181 do Ultramar.
- 1182 Richter, N., Favalli, M., de Zeeuw-van Dalssen, E., Fornaciai, A., Fernandes, R. M. D. S., Pérez, N. M., Levy, J.,
1183 Silva, S.V. & Walter, T. R. (2016). Lava flow hazard at Fogo Volcano, Cabo Verde, before and after the 2014-2015
1184 eruption. *Natural Hazards & Earth System Sciences*, 16(8).
- 1185 Risby, O. & Sandback, B. (2014) Mapping report Fogo, Cape Verde. Uppsala University.
- 1186 Roest, W. R., Danobeitia, J. J., Verhoef, J., & Collette, B. J. (1992). Magnetic anomalies in the Canary Basin and the
1187 Mesozoic evolution of the central North Atlantic. *Marine Geophysical Researches*, 14(1), 1-24.
- 1188 Rona, P. A. (1971). Bathymetry off central northwest Africa. In *Deep Sea Research and Oceanographic Abstracts*
1189 Vol. 18, No. 3, pp. 321-327. Elsevier.
- 1190 Ryabchikov, J. D.; Ntaflos, T.; Kurat, G.; Kogarko, L. N. (1995) Glass-bearing xenoliths from Cape Verde:
1191 evidence for a hot rising mantle jet. *Mineralogy and Petrology*, 55(4), 217-237.

- 1192 Ryan, W.B.; Carbotte, S.M.; Coplan, J.O.; O'Hara, S.; Melkonian, A.; Arko, R.; Weissel, R.A.; Ferrini, V.;
1193 Goodwillie, A.; Nitsche, F.; Bonczkowski, J. (2009) Global multi- resolution topography synthesis. *Geochemistry,*
1194 *Geophysics, Geosystems*, 10, 3.
- 1195 Rydeblad, E. (2018). Mineral Constraints on the Source Lithologies at Fogo, Cape Verde. Unpublished MSc thesis,
1196 Uppsala University. ISSN 1650-6553 ; 433. [urn:nbn:se:uu:diva-354077](https://nbn-resolving.org/urn:nbn:se:uu:diva-354077)
- 1197 Sakai H, Des Marias DJ, Ueda A, Moore JG (1984) Concentrations and isotope ratios of carbon, nitrogen and sulfur
1198 in ocean-floor basalts. *Geochim Cosmochim Acta* 48:2433–2441
- 1199 Samrock L.K., Wartho, J-A., Hansteen, T.H. (2019). ⁴⁰Ar-³⁹Ar dating of the active phonolitic Cadamosto Seamount
1200 (Cape Verde): Indications for periodic volcanic activity and a possible relationship to global sea level changes.
1201 *Lithos*
- 1202 Sandwell, D. T., & Smith, W. H. (1997). Marine gravity anomaly from Geosat and ERS 1 satellite altimetry. *Journal*
1203 *of Geophysical Research: Solid Earth*, 102(B5), 10039-10054.
- 1204 Schwarz, S. & Klügel, A. (2004). Melt extraction pathways and stagnation depths beneath the Maderia and Desertas
1205 rift zones (NE Atlantic) inferred from barometric studies. *Contributions to Mineralogy and Petrology* **147**: 228-240.
- 1206 Serralheiro, A. (1976). A geologia da Ihla de Santiago (Cabo Verde). *Boletim do museu e laboratório mineralógico*
1207 *e geológico da Faculdade de Ciências* 14: 157-369.
- 1208 Serralheiro, A., (1970). Geologia da Ilha de Maio (Cabo Verde). Junta de Investigações do Ultramar.
- 1209 Seifert, R., Malfait, W. J., Petitgirard, S., & Sanchez-Valle, C. (2013). Density of phonolitic magmas and time scales
1210 of crystal fractionation in magma chambers. *Earth and Planetary Science Letters*, 381, 12-20.
- 1211 Silva, L., Le Bas, M. & Robertson, A. (1981). An oceanic carbonatite volcano on Santiago, Cape Verde Islands.
1212 *Nature*, 294, 644–645.
- 1213 Silva, R., Havskov, J., Bean, C. and Wallenstein, N., (2012). Seismic swarms, fault plane solutions, and stress
1214 tensors for São Miguel Island central region (Azores). *Journal of Seismology*, 16(3), pp.389-407.
- 1215 Silva, S.; Calvari, S.; Hernandez, P.; Pérez, N.; Ganci, G.; Alfama, V.; Barrancos, J.; Cabral, J.; Cardoso, N.; Dionis,
1216 S.; Fernandes, P.; Melian, G.; Pereira, J.; Semedo, H.; Padilla, G. & Rodriguez, F., (2017). Tracking the hidden
1217 growth of a lava flow field: the 2014-15 eruption of Fogo volcano (Cape Verde). *Geophysical Research Abstracts*
1218 Vol. 19, EGU 2017-14514, 201. EGU General Assembly 2017.
- 1219 Silva, S., Cardoso, N., Alfama, A., Cabral, J., Semedo, H., Pérez, N., Dionis, S., Hernández, P., Barrancos, J.,
1220 Melián, G., Pereira, J. & Rodríguez, R., (2015). Chronology of the 2014 volcanic eruption on the island of Fogo,
1221 Cape Verde. *Geophysical Research Abstracts* Vol. 17, 2015 EGU General Assembly 2015.
- 1222 Skolotnev, S. G., Turko, N. N., Sokolov, S. Y., Peyve, A. A., Tsukanov, N. V., Kolodyazhnyi, S. Y., Eskin, A. E.
1223 (2007). New data on the geological structure of the junction of the Cape Verde rise, Cape Verde abyssal basin, and
1224 bathymetrists seamounts (Central atlantic ocean). In *Doklady Earth Sciences* (Vol. 416, No. 1, pp. 1037-1041).
1225 MAIK Nauka/Interperiodica.
- 1226 Skolotnev, S. G., Kolodyazhny, S. Y., Tsukanov, N. V., Chamov, N. P., & Sokolov, S. Y. (2009). Neotectonic
1227 morphotructures in the junction zone of the Cape Verde Rise and Cape Verde Abyssal Plain, Central Atlantic.
1228 *Geotectonics*, 43(1), 51-66.

- 1229 Sparks, R. S. J., & Cashman, K. V. (2017). Dynamic magma systems: Implications for forecasting volcanic activity.
1230 *Elements*, 13(1), 35-40.
- 1231 Stephenson, J., Budd, G. M., Manning, J., & Hansbro, P. (2005). Major eruption-induced changes to the McDonald
1232 Islands, southern Indian Ocean. *Antarctic Science*, 17(2), 259-266.
- 1233 Stillman, C., Furnes, H., Le Bas, M., Robertson, A. & Zielonka, J. (1982). The geological history of Maio, Cape
1234 Verde Islands. *Journal of the Geological Society*, Vol. 139, n°3, pp. 347-361. Londres.
- 1235 Stock, M. J., Bagnardi, M., Neave, D. A., MacLennan, J., Bernard, B., Buisman, I., Gleeson, L.M. & Geist, D.
1236 (2018). Integrated petrological and geophysical constraints on magma system architecture in the western Galápagos
1237 Archipelago: insights from Wolf volcano. *Geochemistry, Geophysics, Geosystems*, 19(12), 4722-4743.
- 1238 Streck, M. J. (2008). Mineral textures and zoning as evidence for open system processes. *Reviews in Mineralogy*
1239 *and Geochemistry*, 69(1), 595-622.
- 1240 Tenzer, R., Bagherbandi, M., & Vajda, P. (2013). Global model of the upper mantle lateral density structure based
1241 on combining seismic and isostatic models. *Geosciences Journal*, 17(1), 65-73.
- 1242 Texier-Teixeira, P., Chouraqui, F., Perrillat-Collomb, A., Lavigne, F., Cadag, J. R., & Grancher, D. (2014).
1243 Reducing volcanic risk on Fogo Volcano, Cape Verde, through a participatory approach: which outcome?. *Natural*
1244 *Hazards and Earth System Sciences*, 14(9), 2347-2358.
- 1245 Till, C. B. (2017). A review and update of mantle thermobarometry for primitive arc magmas. *American*
1246 *Mineralogist*, 102(5), 931-947.
- 1247 Torres, P. C., Madeira, J., Silva, L. C., Brum da Silveira, A., Serralheiro, A. and Mota Gomes, A., (1997). Carta
1248 geológica das erupções históricas da Ilha do Fogo: revisão e actualização. In “A erupção vulcânica de 1995 na ilha
1249 do Fogo, Cabo Verde”; Edição do Instituto de Investigação Científica Tropical e Ministério da Ciência e
1250 Tecnologia: 119-132.
- 1251 Torres, P. C., Madeira, J., Silva, L. C., da Silveira, A. B., Serralheiro, A., & Gomes, A. M. (1998). Carta Geológica
1252 das Erupções Históricas da Ilha do Fogo (Cabo Verde): revisão e actualização. *Comunicações do Instituto*
1253 *Geológico e Mineiro*, 84, A193-196.
- 1254 Torres, P., Silva, L., Serralheiro, A., Tassinari, C., & Munhá, J., (2002). Enquadramento geocronológico pelo
1255 método K/Ar das principais sequências vulcano-estratigráficas da Ilha do Sal - Cabo Verde, Garcia de Orta, Serviços
1256 Geológicos, 18(1-2), 9–13.
- 1257 Turbeville, B. N., Wolff, J. A., & Le Bas, M. J. (1987). An oceanic nephelinite–phonolite–carbonatite association,
1258 Brava, Cape Verde Islands. *Eos*, 68, 1522.
- 1259 Vales, D., Dias, N.A., Rio, I., Matias, L., Silveira, G., Madeira, J., Weber, M., Carrilho, F. and Haberland, C.,
1260 (2014). Intraplate seismicity across the Cape Verde swell: A contribution from a temporary seismic
1261 network. *Tectonophysics*, 636, pp.325-337.
- 1262 Verhoef, J., Collette, B.J., Dañobeitia, J.J., Roeser, H.A., Roest, W.R., (1991). Magnetic anomalies off West-Africa
1263 (20–38° N). *Marine Geophysical Research* 13 (2), 81–103.
- 1264 Weidendorfer, D., Schmidt, M. W., & Mattsson, H. B. (2016). Fractional crystallization of Si-undersaturated
1265 alkaline magmas leading to unmixing of carbonatites on Brava Island (Cape Verde) and a general model of
1266 carbonatite genesis in alkaline magma suites. *Contributions to Mineralogy and Petrology*, 171(5), 43.

- 1267 Weit, A., Trumbull, R.B., Keiding, J.K., Geissler, W.H., Gibson, S.A. and Veksler, I.V., (2017). The magmatic
1268 system beneath the Tristan da Cunha Island: Insights from thermobarometry, melting models and
1269 geophysics. *Tectonophysics*, 716, pp.64-76.
- 1270 Welsch, B., Hammer, J., Baronnet, A., Jacob, S., Hellebrand, E., & Sinton, J. (2016). Clinopyroxene in postshield
1271 Haleakala ankaramite: 2. Texture, compositional zoning and supersaturation in the magma. *Contributions to*
1272 *Mineralogy and Petrology*, 171(1), 6.
- 1273 White, R. S. (1984). Atlantic oceanic crust: seismic structure of a slow-spreading ridge. *Geological Society, London,*
1274 *Special Publications*, 13(1), 101-111.
- 1275 Williams, C., Hill, I., Young, R. & White, R. (1990). Fracture zones across the Cape Verde Rise, NE Atlantic.
1276 *Journal of Geological Society of London*, 1
- 1277 Wilson, D., Peirce, C., Watts, A., Grevemeyer, I., Krabbenhoft, A., (2010). Uplift at litho- spheric swells-I: seismic
1278 and gravity constraints on the crust and uppermost mantle structure of the Cape Verde mid-plate swell. *Geophysical*
1279 *Journal International* 182, 531–550.
- 1280 Wilson, D., Peirce, C., Watts, A., Grevemeyer, I., (2013). Uplift at lithospheric swells-II: is the Cape Verde mid-
1281 plate swell supported by a lithosphere of varying mechanical strength? *Geophysical Journal International* 193, 798–
1282 819.
- 1283 Worsley, P. (2015). Physical geology of the Fogo volcano (Cape Verde Islands) and its 2014–2015 eruption.
1284 *Geology Today*, 31(4), 153-159.
- 1285 Wunderman, R. (2003). Summary of recent volcanic activity. *Bulletin of Volcanology*, 65(6), 458-459.
1286
- 1287 Zanon, V., Kueppers, U., Pacheco, J.M. and Cruz, I., (2013). Volcanism from fissure zones and the Caldeira central
1288 volcano of Faial Island, Azores archipelago: geochemical processes in multiple feeding systems. *Geological*
1289 *Magazine*, 150(3), pp.536-555.
- 1290
- 1291 Zanon, V. and Pimentel, A., (2015). Spatio-temporal constraints on magma storage and ascent conditions in a
1292 transtensional tectonic setting: the case of the Terceira Island (Azores). *American Mineralogist*, 100(4), pp.795-805.
1293
- 1294 Zhao Z-F, Zheng Y-F (2003) Calculation of oxygen isotope fractionation in magmatic rocks. *Chem Geol* 193:59–80

Figure captions:

Figure 1. Bathymetric and topographic map of the Cape Verde Archipelago as well as the Cadamosto Seamount, and islands of Brava, Fogo and Santiago. Note the existence of a “northern chain” composed of the islands of Santo Antão, São Vicente, Santa Luzia and islets, and São Nicolau, and a “southern chain” composed of the islands Santiago, Fogo and Brava and the Cadamosto Seamount. Relatively deep water separates most islands. Sources: [Ryan et al., 2009](#); www.geomapapp.org; [Hansteen et al., 2014](#).

Figure 2. Compiled stratigraphy of DSDP Site 368, the Cape Verde Rise and the Cape Verde islands. Several kilometer thick sequences of sediment are found above the ocean crust throughout the region ([Dash et al. 1976](#); [Lancelot et al., 1978](#); [Pim et al., 2008](#)). The Moho is thickened beneath the islands and occurs at depths of 12 to 18 km, becoming deeper towards the East ([Dash et al. 1976](#); [Lancelot et al., 1978](#); [Ali & Watts, 2003](#); [Lodge & Helffrich, 2006](#); [Pim et al., 2008](#)).

Figure 3 (a) Map showing fractures zones, magnetic lineations, gravity and magnetism of the Cape Verde Rise. Black lines show fracture zones as mapped by [Williams et al. \(1990\)](#). Magnetic lineations of M4–M25 are plotted after [Klitgord and Schouten \(1986\)](#). The satellite-derived free-air gravity anomalies are shown ([Sandwell and Smith, 1997](#)). DSDP Site 368 is labeled. Magnetic quiet zone boundary from [Hayes and Rabinowitz \(1975\)](#). (b) Schematic cross-section NNW to SSE across the Cape Verde Rise following the yellow line ([Stillman et al., 1982](#); [Ryan et al., 2009](#); www.geomapapp.org).

Figure 4. Age distribution of exposed volcanic and intrusive products across the Cape Verde Archipelago. Black bars mark $^{40}\text{Ar}/^{39}\text{Ar}$ geochronology and He exposure dating whereas grey bars show K/Ar geochronology and other geological inferences. Modified from [Holm et al. \(2008\)](#) and based on the geochronological data of [Bernard-Griffiths et al. \(1975\)](#); [Mitchell et al. \(1983\)](#); [Torres et al. \(2002\)](#); [Plesner et al. \(2003\)](#); [Jørgensen and Holm \(2002\)](#); [Duprat et al., \(2007\)](#); [Holm et al. \(2008\)](#); [Foeken et al. \(2009\)](#); [Dyhr and Holm \(2010\)](#); [Madeira et al. \(2010\)](#); [Ramalho et al. \(2010c\)](#); [Samrock et al. \(2019\)](#). Note the tentative age progression of the oldest exposed rocks in East to the youngest in the West.

Figure 5. Photographs from the islands of Santiago, Fogo and Brava and the Cadamosto Seamount. a) Hyaloclastite from the submarine Flamengos Formation overlain by subaerial lavas of the Pico de Antónia Formation, Santiago. b) The Assomada Formation, Santiago, c) Pillow lavas in hyaloclastite from the Flamengos Formation overlain by a beach conglomerate followed by lavas of the Pico de Antónia Formation, Santiago, d) the iconic Pico do Fogo standing on the Chã das Caldeiras plateau,

Fogo, e) Bordeira cliffs that surround the western part of the Chã das Caldeiras, Fogo, f) the volcanic eruption at Fogo in November 2014, g) pillow lavas and hyaloclastites intruded by dike swarms, Brava h) sequence of ignimbrites, Brava, i) phonolitic lava near the summit of the Cadamosto Seamount (M80/3-033ROV), and j) steep cliffs of a crater at the Cadamosto Seamount (M80/3-033ROV).

Figure 6. Geological and topographic map of Santiago after [Serralheiro, 1976](#); [Ryan et al., 2009](#); www.geomapapp.org. The Flamengos Formation is exposed along river valleys and in highly eroded areas. The Pico da Antonia Formation is wide spread and represents the shield building stage of volcanism. An erosional phase was followed by lavas of the Assomada Formation that flowed into valleys and the Monte das Vacas scoria cones.

Figure 7. Geological and topographic map of Fogo after [Torres et al., 1998](#); [Ryan et al., 2009](#); www.geomapapp.org; [Carracedo et al., 2015](#). The shield volcano is composed of the Monte Amarelo Group volcanism and a large landslide created the Bordeira cliffs delineating the Cha das Caldeiras plateau that hosts the Pico do Fogo and recent volcanic eruptions.

Figure 8. Geological and topographic map of Brava after [Madeira et al., 2010](#); [Ryan et al., 2009](#); www.geomapapp.org. The island core complex is composed of alkaline and carbonatite intrusive rocks, whereas the volcanic rocks are mostly nephelinite to phonolite.

Figure 9. Compositional classification of the volcanic rocks from Santiago, Fogo, Brava and the Cadamosto Seamount after [LeMaitre et al. 2002](#). Santiago and Fogo host (mela)nephelinite, basanite and tephrite to basalt and even tephriphonolite at Fogo. In contrast Brava and the Cadamosto Seamount host dominantly phonolite and syenite. Data are unnormalized. Data sources: [Kokfelt, 1998](#); [Barker et al., 2009, 2012](#); [Hildner et al., 2011, 2012](#); [Magnusson, 2016](#); [Weidendorfer et al., 2016](#).

Figure 10. Clinopyroxene Mg# histograms for Santiago, Fogo, Brava and the Cadamosto Seamount. All locations host diopside-augite, whereas Brava and the Cadamosto Seamount display highest abundances of aegirine-augite. Data sources: [Barker et al., 2009, 2012](#); [Hildner et al., 2011, 2012](#); [Weidendorfer et al., 2016](#); [Rydeblad, 2018](#).

Figure 11. Clinopyroxene mineral chemistry for Santiago, Fogo, Brava and the Cadamosto Seamount. The clinopyroxene crystals from Santiago and Fogo display a distinct trend from the clinopyroxene found

1359 *at Brava and the Cadamosto Seamount. Data sources: [Barker et al., 2009, 2012](#); [Hildner et al., 2011,](#)*
 1360 *[2012](#); [Weidendorfer et al., 2016](#); [Rydeblad, 2018](#).*

1361 *Figure 12. Zonation in clinopyroxene from Santiago, Fogo and the Cadamosto Seamount, shown by Mg#*
 1362 *(mol%), TiO₂ and MnO content. Santiago Coastal group of the Flamengos Formation (121-306; [Barker](#)*
 1363 *[et al., 2009](#)); Fogo eruption 1847 (CVF07 pyroxene 11; [Rydeblad, 2018](#)); Cadamosto Seamount (D885*
 1364 *J1; [Barker et al., 2012](#)). Data sources: [Barker et al., 2009, 2012](#); [Hildner et al., 2011, 2012](#);*
 1365 *[Weidendorfer et al., 2016](#); [Rydeblad, 2018](#).*

1366 *Figure 13. Multicomponent equilibrium ΔDiHd versus temperature for diopside-augite from Santiago,*
 1367 *Fogo, Brava and the Cadamosto Seamount calculated after [Mollo et al. \(2013\)](#).*

1368 *Figure 14. Thermobarometry for Santiago, Fogo, Brava and the Cadamosto Seamount. Data sources:*
 1369 *[Barker et al., 2009, 2012](#); [Hildner et al., 2011, 2012](#); [Rydeblad, 2018](#). The Moho is from [Lodge &](#)*
 1370 *[Helffrich 2006](#) and [Pim et al. 2008](#).*

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1372 *Figure 15 Crystallization conditions; temperature and pressure, with composition for zoned*
 1373 *clinopyroxene from Fogo; a) Prehistoric eruption CVF05 pyroxene 8, b) Eruption in 1785 CVF08*
 1374 *pyroxene 15, c) Eruption in 1799 CVF09 pyroxene 5 and d) Eruption in 1847 CVF06 pyroxene 7.*

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1376 *Figure 16. Crystallisation depth and model for the magma plumbing system beneath Santiago, Fogo,*
 1377 *Brava and the Cadamosto Seamount. The shallow magma storage system is illustrated by fluid inclusion*
 1378 *barometry as well as seismicity, deformation and the active hydrothermal systems. Data sources: [Da](#)*
 1379 *[Silva et al., 1999](#); [Amelung & Day, 2002](#); [Barker et al., 2009, 2012](#); [Ryan et al., 2009](#);*
 1380 *www.geomapapp.org; [Heleno et al., 2006](#); [Grevemeyer et al., 2010](#); [Hildner et al., 2011, 2012](#); [Faria &](#)*
 1381 *[Fonseca, 2014](#); [Rydeblad, 2018](#). The Moho is from [Lodge & Helffrich 2006](#) and [Pim et al. 2008](#).*

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1383 *Figure 17. Schematic of the magma storage systems for Ocean Islands globally including results from*
 1384 *clinopyroxene-melt thermobarometry, fluid inclusions, melt inclusions, experimental petrology*
 1385 *deformation and seismicity. Cape Verde; [Da Silva et al 1999](#); [Amelung & Day 2002](#); [Lodge & Helffrich,](#)*
 1386 *[2006](#); [Pim et al., 2008](#); [Barker et al., 2009, 2012](#); [Grevemeyer et al., 2010](#); [Hildner et al., 2011, 2012](#);*
 1387 *[Faria & Fonseca 2014](#); [Fernandes & Faria, 2015](#); [Vales et al., 2015](#); [Jenkins et al., 2017](#); [Leva et al.,](#)*

1388 2019; Mata et al., 2017; this study. *Canary Islands*; Klügel et al., 2000; 2005; Galipp et al., 2006;
 1389 Longpré et al., 2008, 2014; Aulinas et al., 2010; Barker et al., 2015; González et al., 2013. *Madeira*;
 1390 Schwarz & Klügel 2004; Klügel & Klein, 2006. *The Azores*; Renzulli & Santi 2000; Beier et al., 2006;
 1391 Dias et al., 2007; Silva et al 2012; Jeffery et al., 2016; Madureira et al., 2008; Zanon et al., 2013; Zanon
 1392 & Pimentel 2015. *Tristan da Cunha*; Geissler et al., 2016; Weit et al., 2017. *Ascension*; Klinghofer et al.,
 1393 2001; Hanson et al., 1996; Chamberlain et al., 2016. *Hawaii*; Putirka et al. 1996, Putirka 1997;
 1394 Chatterjee et al., 2005; Poland et al 2015; Hammer et al., 2016. *Galapagos*; Stock et al., 2018. *Reunion*;
 1395 Bureau et al., 1998; Famin et al., 2009; Peltier et al. 2009; Di Muro et al., 2014; Fontaine et al., 2014.
 1396 *Kerguelen*; Freise et al., 2003.

DSDP site 368

Cape Verde Rise

Cape Verde Islands

Legend

Atlantic Ocean

Cape Verde	Canary Islands	Madeira	Azores	Tristan da Cunha	Ascension
1 to 5 km seismicity, deformation	4.5 km deformation	2 to 4 km fluid inclusions	2 to 4 km water solubility, fractionation, seismicity	earthquakes prior to eruptions	1.5 to 2 km seismicity
8 to 13 km fluid inclusions, seismicity	10 to 16 km fluid inclusions, deformation	8 to 10 km fluid inclusions	>8 km fluid inclusions, cpx-melt, seismicity	6 to 10 km cpx-melt	8.5 km melt inclusions
12 to 18 km Moho	14 km Moho	14 to 15 km Moho	14 km Moho	19 km Moho	12 to 13 km Moho
13 to 46 km cpx- melt, seismicity, deformation	12 to 45 km cpx- melt, seismicity	15 to 35 km melt	<30 km cpx- melt, fsp-ol melt, fluid inclusions	24 to 36 km cpx-melt	
da Silva et al 1999; Amelung & Day 2002; Lodge & Helffrich, 2006; Pim et al., 2008; Barker et al., 2009, 2012; Grevemeyer et al., 2010; Hildner et al., 2011, 2012; Faria & Fonseca 2014; Fernandez & Faria, 2015; Vales et al., 2015; Jenkins et al., 2017; Leva et al., 2019; Mata et al., 2017; this study.	Klügel et al., 2000; 2005; Galipp et al., 2006; Longpre et al., 2008, 2014; Aulinas et al., 2010; Barker et al., 2015; Gonzalez et al., 2013.	Schwarz et al., 2004; Klügel & Klein, 2006	Renzulli & Santi 2000; Beier et al., 2006; Dias et al., 2007; Silva et al 2012; Jeffery et al., 2016; Madureira et al., 2008; Zanon et al., 2013; Zanon & Pimentel 2015	Geissler et al., 2016; Weit et al., 2017	Klinghofer et al., 2001; Hanson et al., 1996; Chamberlin et al., 2016

<u>Pacific Ocean</u>		<u>Indian Ocean</u>	
Hawaii	Galapagos	Reunion	Kerguelen
2 to 4 km (<10 km) seismicity	1 km deformation	0.5 to 2.5 km water solubility, melt inclusions, seismicity	
9 to 18 km cpx-melt	8 to 13 km cpx-melt, deformation	7 to 14 km water solubility, melt inclusions, seismicity	6 to 8 km experimental
18 to 19 km Moho	13 km Moho	14 to 15 km Moho	
19 to 46 km cpx-melt		15 to 24 km water solubility, melt inclusions, seismicity	
Putirka 1996, 1997; Chatterjee et al., 2005; Poland et al 2015; Hammer et al., 2016	Stock et al., 2018	Bureau et al., 1998; Famin et al., 2009; Peltier et al. 2009; Di Muro et al., 2014; Fontaine et al., 2014	Freise et al., 2003