Volcanic and hydrothermal processes in submarine calderas: the Kulo Lasi example (SW Pacific)

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Abstract :

The study area is located at the transition between the northern end of the Tonga Trench and the North Fiji fracture zone, where tectonic movements are reputed to be the fastest in the world. To the southeast of Futuna Island, a broad area of volcanism occurs within a region characterized by a change in the tectonic fabric between a NE-SW oriented volcanic graben and the N-S oriented Alofi ridge. In 2010, the active volcano Kulo Lasi, which represents the most recent volcanic episode in the Futuna area, was discovered in the center of this extensive volcanic zone. Kulo Lasi is a 20 km diameter shield volcano that rises 400 m above the seafloor. It is composed of basaltic to trachy-andesitic lava with no obvious geochemical affinity with the Tonga subduction that occurs 500 km to the east. The central caldera is 5 km in diameter and 300 m deep and is located at a water depth of 1500 m. Diving operations with the submersible Nautile and high-resolution AUV mapping, have revealed the presence of numerous active and inactive hydrothermal fields on the floor and the walls of the caldera. Four tectono-volcanic stages can be distinguished at Kulo Lasi caldera. In stage 1, the shield volcano is built. Annular reverse faults develop at the summit and control circulation of water/rock-dominated hydrothermal fluids and high-temperature alteration of rocks along the nascent normal faults. Mixing of hydrothermal fluids with seawater is favored along normal superficial faults, leading to the formation of low-temperature Fe/Mn mineralization at the summit of the volcano. During stage 2, the caldera collapse, gradually revealing outcrops of the altered and mineralized zones formed during Stage 1. As the magma chamber cools and collapses, less heat is available. As a result, medium to low-temperature (<100 °C) Fe/Si deposits form on the floor of the caldera. In stage 3, refilling and ascent of the magma chamber at depth promote the uplift of a central resurgent dome devoid of recent lava emission. The depth of the magma chamber under the dome is estimated to be 1.6 km. Annular dykes feeding the new lava flows are controlled by the deep reverse faults at the outer rim of the caldera floor. The eruptive events are accompanied by short-lived emission of magmatic fluids, rich in SO₂, materialized by the presence of native sulfur depositions on the surface of the most recent lava flows. During stage 4 (present day), black smoker sulfide chimneys, controlled by the deep reverse faults, form at the surface of the most recent lava flows at the outer part of the caldera floor. The fluids emitted result from the mixing of a deep fluid of the water/rock reaction type and of shallow seawater superheated in contact with the hot dykes. Hydrogen is generated during reaction of seawater with the hot dikes and H₂ is enriched in the mixed fluids. Active (43°C) siliceous chimneys at the base of the caldera walls and low-temperature vents (6°C) perched on

the caldera walls suggests a decrease in the venting temperature at a distance from the floor of the caldera. In chimneys, the sphalerite/pyrrhotite/isocubanite/barite paragenesis, probably linked to the high H₂ concentrations in the fluids, indicate unusual reducing conditions for a back-arc. The sulfide mineralization also has unusual trace element concentrations. The concentration in Pb and Ba appears characteristic of back-arc environments while the enrichment in Co, Se, and Sn is more common to mineralization associated with basalt in mature back-arcs and associated with ultramafic rocks on mid ocean ridges.

Graphical abstract

Highlights

► A new major active submarine caldera in a back-arc environment. ► Different types of hydrothermal deposits in relation to the different stages of eruptive events and collapse of the caldera. ► Venting different types of fluids. ► Unusual mineralogy and chemistry for back-arc mineralisation.

Keywords : Subaqueous volcanism, hydrothermal activity, Kulo Lasi caldera, sulfides, SW Pacific

1 Introduction

1.1 Hydrothermal activity in back-arc environments

Since the late 1980s, arcs and back-arc basins have attracted the attention of economic geologists because the vast majority of ancient Volcanogenic Massive Sulfides (VMS) deposits mined on land are believed to have formed in this type of environment. For example, several major hydrothermal deposits are known to occur along the Kermadec volcanic arc (e.g., (Baker et al. , 2005, Baker et al. , 2012, Berkenbosch et al. , 2012, de Ronde et al. , 2001, Leybourne et al. , 2012a), in the Lau Basin (Baker et al. , 2010, Baker et al. , 2006, Bertine and Keene, 1975, Evans et al. , 2017, Fouquet et al. , 1991, Fouquet et al. , 1993, Hawkins et al. , 1986., Herzig et al. , 1998, Kim et al. , 2009, Lisitsyn et al. , 1992, Lupton et al. , 2009, Massoth et al. , 2007, Von Stackelberg, 1988)), in the North Fiji Basin (Auzende et al. , 1995, Bendel et al. , 1993, Halbach et al. , 2001, Ishibashi et al. , 1994), in

the Manus Basin (e.g., (Auzende et al. , 1996, Bach et al. , 2007, Binns and Scott, 1993, Both et al. , 1986, Gamo et al. , 1997, Gemmell et al. , 1999, Gena et al. , 2001, Kim et al. , 2004, Lisitsyn et al. , 1993, Moss and Scott, 2001, Petersen et al. , 2005, Taylor et al. , 1986), in the Mariana basin (e.g., (Baker et al. , 2008, Baker et al. , 2005, de Ronde et al. , 2014a, Hein et al. , 2014, Ishibashi et al. , 2006, Resing et al. , 2007) and in the Okinawa basin (Glasby, 2008, Halbach et al. , 1989, Halbach and Pracejus, 1993, Hou et al. , 2005, Iizasa et al. , 1992, Kimura et al. , 1988, Monecke et al. , 2014, Noguchi et al. , 2007). Exploration over the last 30 years has shown that intense hydrothermal activity occurs along the back-arc ridges and on a high proportion of submarine volcanoes of island arcs (de Ronde et al. , 2003, de Ronde and Stucker, 2015). The volcanic and tectonic instability of back-arc basins and the wide variability of the associated volcanic rocks generate extremely varied hydrothermal fields with respect to fluid composition and mineralization (de Ronde et al. , 2015, de Ronde and Stucker, 2015, Resing et al. , 2011). Back-arc basins can be classified into different types (Monecke et al. , 2014). Young basins can develop within continental crust (e.g., Okinawa Basin, Woodlark Basin) or island arc crust (e.g., Lau Basin, Manus Basin) while the most evolved back-arc basins (e.g., North Fiji Basin and central Manus) are dominated by tholeiitic basalt and the corresponding mineralizing processes here are very similar to those that occur on mid-ocean ridges (MOR) (Bendel et al. , 1993).

1.2 Caldera formation and fossil mineral deposits

Caldera volcanoes represent an important ore-forming environment for a range of deposit types, including porphyry copper, epithermal deposits, polymetallic veins, and volcanogenic massive sulfide (VMS) deposits (Elston, 1994, Mueller et al. , 2009, Rytuba, 1994, Stix et al. , 2003). Large ancient VMS deposits are found in paleo back-arc environments and show a close spatial and temporal relationship with silicic submarine calderas (Ohmoto, 1978, 1996, Ohmoto et al. , 1983). For example, the Kuroko VMS deposits are associated with clusters of subaqueous calderas (Ohmoto, 1978, 1996, Ohmoto et al. , 1983).

Calderas are formed by the collapse of the top of the volcano after eruption of a large quantity of magma from a subsurface reservoir. They can also form following the withdrawal of magma beneath the volcano. The collapse is triggered by large-volume eruptions and sustained effusive volcanism that partially drain high-level magma chambers (Acocella, 2007, Kennedy and Stix, 2003, Lipman, 1997, Mueller et al. , 2009). The subsurface magma provides the heat and magmatic volatiles that drive the overlying hydrothermal systems. Permeability associated with the caldera faults focuses the hydrothermal circulation (Eddy et al. , 1998). The size of the calderas can vary from 2 to 100 km in diameter (Lipman, 2000).

Calderas display near-circular geometry, with ellipsoidal forms consistent with multiple eruptive events and regional tectonics (Mueller et al. , 2009). In recent years, monitoring of caldera collapse (Geshi et al. , 2002), analog experiments (Acocella et al. , 2000, Geyer et al. , 2006, Kennedy et al. , 2004, Roche et al. , 2000) and numerical modeling (Gudmundsson, 1998) have considerably increased our understanding of caldera formation. For example, field and experimental works (Branney, 1995, Kennedy and Stix, 2003, Kennedy et al. , 2004, Kennedy et al. , 2008, Lavallee et al. , 2004, Roche et al. , 2001, Roche et al. , 2000, Stix et al. , 2003) have shown that the principal caldera ring faults, which form as a result of magma-chamber evacuation, dip vertically or outwards. These latter are reverse faults forming a ring with a comparatively smaller diameter relative to the topographic rim of the caldera. A second set of normal faults, associated with late-stage peripheral extension, dip inwards and are located on the margins of the caldera. These two sets of faults provide an ideal fracture system for fluid circulation and ore formation (Stix et al. , 2003).

1.3 Hydrothermal activity in modern deep submarine calderas

Sulfide deposits in modern submarine calderas are documented in the deep ocean in about 29 hydrothermal fields (Table 1). Eleven sites are located in back-arc environments, sixteen are related to arc volcanoes, and only two are known on MORs. However, on fast-spreading ridges, the processes involved in the formation of the axial summit graben are similar to those involved in caldera formation as demonstrated by field observations, depth of the magma chamber, and analog models (Lagabriele and Cormier, 1999, Lagabriele et al. , 2001). In the modern ocean, calderas form at relatively shallow water depths compared to typical hydrothermal fields on mid-ocean ridges, and only two have been identified at a depth greater than 2000 m. In back-arc settings, the dominant lavas are basalt and andesite, while island arcs lavas are more felsic (Table 1).

Among these calderas, only a few have been studied in any detail using a combination of high resolution mapping and/or diving operations. Major detailed studies are along the Tonga-Kermadec arc and on the Brothers caldera (Baker et al. , 2012, Berkenbosch et al. , 2012, de Ronde et al. , 2005, de Ronde et al. , 2011, Embley et al. , 2012, Gruen et al. , 2014, Tontini et al. , 2012, Wright and Gamble, 1999), and Volcano 19 (Stoffers et al. , 2006); and along the Izu-Bonin arc: Suiyo seamount (Marumo et al. , 2008), Myiojin Knoll (Honsho et al. , 2016) and Bayonnaise knoll (Honsho et al. , 2015). In contrast with hydrothermal systems on MORs, the involvement of magmatic fluids rich in metals has been demonstrated in several back-arc basins sites (Berkenbosch et al. , 2012, Berkenbosch et al. , 2015, Butterfield et al. , 2011, de Ronde et al. , 2001, de Ronde et al. , 2015, de Ronde et

al. , 2011, de Ronde and Stucker, 2015, Herzig et al. , 1998, Resing et al. , 2007, Yang and Scott, 1996).

1.4 Objectives

The Alaufi cruise (geophysics and seabed mapping), conducted west of Futuna in 2000, revealed the existence of an active volcanic ridge and recent volcanic areas to the south and east of Futuna Island (Pelletier et al. , 2001). The Kulo Lasi volcano was discovered in 2010 during regional mapping to explore the French Exclusive Economic Zone (EEZ) (Fouquet et al. , 2015). Vertical and tow-yo hydrocasts have revealed intense hydrothermal activity (Konn et al. , 2016) in the Kulo Lasi caldera. In this publication, we present results obtained on the Kulo Lasi volcano in 2010 (Futuna 1 cruise) and 2012 (Futuna 3 cruise). Local high-resolution bathymetry and seafloor backscatter images were completed in the caldera using an Autonomous Underwater Vehicle (AUV). Seven dives (six in 2010 and one in 2012), using the manned submersible Nautilie, were then performed to explore and sample volcanic rocks, hydrothermal products, fluids and animals within the caldera. Several active and inactive hydrothermal sites were discovered (Fouquet et al. , 2015). During the dives, preliminary observations of biota associated with the active vents was also made. We present detailed geological data based on high-resolution maps and on seafloor observations using the manned submersible Nautilie. We also discuss compositions of fluids and of hydrothermal deposits in relation to the evolution of the volcano. Our data provides insights into the factors controlling the location, type, and timing of hydrothermal discharge in submarine calderas. The young age of this volcano, combined with the availability of high-resolution data make it a suitable locality at which to investigate the volcano-tectonic controls and temporal succession of different types of hydrothermal mineralization in back-arc calderas with effusive volcanic activity.

2 Materials and methods

2.1 Seabed mapping, AUV, and Nautilie submersible

The regional bathymetric maps and seafloor backscatter images were mainly acquired at 8 knots with the ship-borne Kongsberg EM122 multibeam echosounder (12 kHz) installed on the R/V L'Atalante. Data were processed with Caribes and Sonarscope/3DViewer softwares developed at IFREMER. The resolution of the maps reaches 25 and 10 m for the bathymetry and seafloor backscatter, respectively. Some surveys conducted at lower speed (5 knots) in the Kulo Lasi area led to even better resolution, with 10 m and 5 m for the bathymetry and seafloor backscatter resolutions, respectively.

The high-resolution bathymetric map of the central Kulo Lasi caldera was constructed using a Kongsberg/Simrad EM2000 multibeam echosounder (200 kHz 1.5 x 3.5°) (CNRS Geoazur equipment) integrated in an AUV (AsterX, IFREMER) that was navigated at 70 m above the seafloor at 3 knots with survey lines spaced by 200 m. Acoustic data were processed to provide 2 and 0.5 m resolution for the bathymetry and seafloor backscatter maps, respectively. In addition to the bathymetric multibeam survey, a magnetometer (INSU/IPG /Paris) and a CTD were installed in the AUV during the mapping operations. Details on the AsterX AUV, and the navigation and the EM2000 multibeam payload can be found in (Dupre et al. , 2008). Six AUV dives of 8 hours each were carried out to cover the Kulo Lasi caldera area (35 km²) corresponding to a 225 km long track.

2.2 Fluid samples analyses

In order to collect duplicate samples of hot fluids from different vents, titanium samplers of 750 mL capacity were deployed in pairs and manipulated by the arm of the Nautilie submersible. The samplers were specially conditioned to avoid outgassing and gas leakage during recovery and on-board gas extraction. The temperature of the sampled fluids was measured by using autonomous sensors (Micrel-NKE sensors) placed on the sampler; pH and H₂S concentrations were measured immediately after recovery of the samples and total gas were then extracted following the procedure as described in (Charlou and Donval, 1993). Extracted gases were stored in special gas-tight flasks of 100 or 200 mL capacity, then separated by gas-chromatography (Agilent GC 7890A, Agilent Technologies) and quantitatively analyzed by triple detection using mass (MS 5975C, Agilent technologies), flame ionization and thermal conductivity detectors. Preliminary concentrations of major gases (CO₂, H₂, CH₄, and N₂) were obtained on board using a portable microsensor Technology Instruments Inc. chromatograph coupled to the gas extractor. Aliquots of gas were stored in vacuum-tight tubes (Labco, Ltd.) and sent for carbon isotope studies (Isolab b. v., Netherlands). After gas extraction, hydrothermal solutions were transferred either to acid-cleaned polyethylene bottles for major and minor elements analysis or to precombusted glass bottles for extraction of organic compounds. Major elements including anions (Cl, Br, and SO₄²⁻) and cations (Na, Mg, K, and Ca) were measured by ion chromatography (Dionex Ion Chromatograph System 2000) after appropriate dilution. Fe, Mn, Cu, Zn, Sr, Li, and Rb were measured by flame atomic absorption spectrometry using standard additions (Aanalyst, Perkin Elmer Inc.). Aliquots for silica determination were immediately diluted 100- to 200-fold and analyzed on board using the silicomolybdate automatic colorimetric method. In addition, dissolved organic compounds were recovered from fluids on board using stir-bar sorptive

extraction and then qualitatively analyzed by gas chromatography – mass spectrometry on a 6890 – 5973N GC-MS from Agilent (Konn et al. , 2009, Konn et al. , 2018).

2.3 Rock sample analyses

Samples collected by dredges or during the dives were characterized and prepared directly on-board. Rocks were cut and then dried in an air oven at 50°C. A representative portion of each sample was ground to a fine powder using an agate mortar. Mineralogical and quantitative chemical data were acquired using X-ray techniques (X-ray diffraction and X-ray fluorescence) at the Laboratoire de Géochimie et Métallogénie at IFREMER institute (Brest, France). X-ray diffraction (XRD) analyses were conducted with a BRUKER AXS D8 Advance and a BRUKER D2 PHASER diffractometer. Samples were top loaded into 2.5-cm-diameter circular cavity holders, and analyses were run between 5° and 70° 2 θ , in steps of 0.01° 2 θ at 1s/step (monochromatic Cu K α radiation, 40 kV, 30mA). Minerals were identified using Diffrac Suite EVA software. X-ray fluorescence analyses were conducted with a wavelength dispersive X-ray fluorescence spectrometer (WD-XRF; BRUKER AXS S8 TIGER) on fusion beads (for major elements) or compressed pellets (for trace elements). After data acquisition, the measured net peak intensities, corrected for inter-element effects, were converted into concentrations using calibration curves generated from analysis of certified geochemical standard powders (measured under identical analytical conditions). Calibrations were established using a set of certified materials obtained mainly from the Canadian Certified reference materials Project (CCRMP) (for example, CCU-1, CZN-1, Fer-1, PTC-1 or PTM-1) Geological Survey of Japan (GSJ)(JP-1) and the Centre de Recherches Pétrochimiques et Géochimiques in France (CRPG)(BE-N). For sulfur isotopes, sampling was done in two ways: (1) micro-drilling of polished sections, and (2) separation by hand under a binocular microscope. The collected minerals were digested in inverted aqua regia and then purified by elution through a cationic resin to remove any interfering matrix. The solution was analyzed by a Neptune MC-ICP-MS at (PSO- UBO/IFREME-Brest).

3 Geological setting and morphology of the Kulo Lasi volcano

3.1 Regional setting and morphology of the volcano

The Futuna Islands (Futuna and Alofi), are located south of the ancient subduction zone of Vityaz, in the junction area between the Lau back-arc basin and the North Fiji back-arc basin

at the eastern end of the North-Fiji transform fault (Figure 1). The study area is located about 200 km to the west of the northern tip of the Tonga trench. The fastest plate velocity recorded on earth (18 to 24 cm per year) (Bevis et al. , 1995) is located in this area. The sinistral North-Fiji transform zone connects the northern end of the Tonga trench to the spreading axis of the North Fiji back-arc basin and passes through the northern edge of the Fijian platform (Pelletier et al. , 1998, Pelletier et al. , 2001). The islands of Futuna and Alofi are located along this fault and are regularly subject to both superficial and local earthquakes of high magnitude (Pelletier et al. , 2001).

The Kulo Lasi volcano is situated about 100 km SE of Futuna Island. North of the volcano, the main faults are oriented NE-SW, while in the south, the regional tectonic structure fabrics are progressively folded to a nearly N-S orientation (Figure 2). The volcano is active and is superimposed on a highly tectonized, but young volcanic basement (Figure 2). Given the absence of sediments in the backscatter images, the basement appears to have formed during several episodes of extremely recent volcanic events. The older, partly sedimented area, south of Futuna Island and at the west of the volcano, is host to two groups of faults; the oldest of which are oriented NNE-SSW and the younger are N-S. Thus, the non-tectonized Kulo Lasi volcano represents the youngest known volcano in the investigated area south Futuna volcanic domain.

The Kulo Lasi volcano is about 20 km in diameter and 360 m high and has a relatively flat summit located at a depth of 1200 m. It is composed of a succession of non-tectonized lava flows (Figure 3). Close to the summit are three flat-topped and circular volcanic cones of about 2 km in diameter and 150 m high. Two of them have a central deep crater and one has a small mound in the center. This type of volcanic cone is common in the Futuna area and has been described in many places along MOR and near Hawaii. They form as submarine lava ponds continuously overflow, forming a circular structure with a flat horizontal top and steep pillowed margins (Clague et al. , 2000). Factors that facilitate the construction of these structures include: effusive eruption, low volatile content, moderate to high confining pressure, long lived steady eruption, moderate effusion rates and relatively flat slope (Clague et al. , 2000).

The caldera is sub-circular, with an outer diameter of 5.6 x 4.5 km and contains a 1.5 km in diameter central resurgent dome. The internal diameter of the caldera is 4.2 x 3 km. Between the resurgent dome and the caldera walls, the flat bottom of the caldera is covered by recent lavas. The inner walls are composed of tilted old lava flows, talus, huge boulders, and angular blocks. The central resurgent dome consists of tectonized lava due to the uplift of the

old caldera floor. The recent lava flows at the bottom of the caldera are formed of pillow, lobate, and ropy lavas that cover scoriaceous lava or pillow lava. These different lavas are all rich in fresh glass. Even though the floor of the caldera is extremely young, as determined by combining high-resolution mapping and dives observations, it is possible to identify several generations of flows, most of which appear to have fresh surfaces. The absence or paucity of sediment and weathering attests to very recent volcanic activity in the NW and S sectors of the caldera. Lava flows in the SW and central parts appear to be older and sometimes show light sediment coverage. The SE margin of the caldera floor is covered by very recent flows. The Kulo Lasi volcano is volcanically and hydrothermally active today, as is demonstrated by the numerous active hydrothermal sites and hydrothermal plumes discovered inside the caldera (Konn et al. , 2016).

3.2 Morphology of the caldera based on high-resolution bathymetry

The bathymetry (Figures 3 and 4) reveals four types of terrains: (1) The caldera walls, which are almost vertical and up to 300 m high; (2) the outer part of the caldera floor, formed of old collapsed volcanic lava flows that are tilted towards the center of the caldera and marked by numerous normal faults oriented parallel to the main caldera walls; (3) the resurgent central dome which has been affected by a NE-SW oriented tectonic fabric. (4) Fresh lava flows on the floor of the caldera. The total volume of lava emitted by the Kulo Lasi (27 km^3) before the formation of the caldera is estimated considering the average diameter of the volcano (17 km) and its height (360 m), before the collapse, estimated from the slopes.

3.2.1 Caldera walls

The inner wall of the caldera is asymmetric with the western side composed dominantly of tilted old lavas and a steep and narrow eastern side composed of abundant volcanic scree (Figure 4).

The northern wall, traversed during dive #1, consists of recent lava scree, often sorted and lacking sedimentary cover. The slope contains a number of steps marked by normal faults before reaching the caldera floor. The base of the wall is vertical and has been flooded by very recent lobate lava flows that infill the caldera floor. The surface of this lava is typically a shiny glass. One of the dives along the base of the western side of the caldera (dive #4) mapped only talus and scree generated by the collapse of several lava flows. The collapse formation here is about one kilometer wide. When presented in 3D, the high-resolution bathymetry clearly shows a series of old lava flows that are tilted towards the center of the caldera and are reminiscent of stack of pancakes (Figure 4 left part of the image). The

eastern side of the caldera shows a vertical wall with abundant scree that often covers the entire wall section from the top to the caldera floor. At the base of the wall, on the caldera floor, older pillow lava ridges are buried under recent lava flows. Between these ridges and the base of the caldera wall, a long depression has been filled by recent lava flows that cover the older scree. On this side of the caldera, numerous hyaloclastites are observed near the recent flows. In the depression, a network of large N-S oriented cracks reveals sections of the prismatic flows and indicates extension at the base of the caldera wall. Some of these cracks are lined with active and/or inactive sulfide chimneys about one meter in height. The active chimneys are discharging a clear fluid. Several yellow colored hydrothermal sediment fields are observed around the chimneys. During dive #5, a complete section along the southern caldera wall showed old tectonized lava and scree at the base, an extensive alteration and mineralized zone exposed on the caldera wall on the upper part of the slope and massive prismatic lava flows at the top of the caldera wall.

3.2.2 Central resurgent dome

The resurgent dome consists of old pillow lava flows and tectonized and still forming debris on the slopes. There is no evidence of hydrothermal activity at the top or the base of the dome. The top of the resurgent dome is tectonized and made with volcanic boulders while the slopes show relatively old lavas flows partially covered by scree. In the south, the resurgent dome strikes towards the caldera wall and is again made up of ancient and tectonized basalts. These older lavas are mostly pillows with a few lobate flows often covered by a film of sediment. Some small areas (1 to 2 m²) of ocher-colored hydrothermal sediments are observed locally and does not seem to be fault controlled. The southern half of the dome appears to be dissected by a series of E-W to SE-NE oriented faults that run parallel to the regional faults observed on the old basement to the north (Figures 3 and 4).

3.2.3 Recent lava flows

Most of the area between the caldera walls and the resurgent dome is occupied by recent lavas flows (i.e. pillow lavas, lobate flows, ropy and scoriaceous lava flows). To the north, draped lavas form a recent flow a few meters high on the top of the recent lobate flows (Dive #1). In the SE part of the caldera, nearer to the walls, the lava flows also appear to be very young; for example, the glassy edges of the pillow lava are very fresh and sediments almost nonexistent. Based on the lava typology, and of their overlapping relationships, five recent eruptive episodes are distinguished in the SE quarter of the caldera (Figure 5). Episode 1 forms pillow lava flows with light sediment coverage at the deepest part (> 1500 m) of the caldera floor. Episode 2 occurs as scoriaceous lava flows and sheet flows, indicating very fluid lava. Episode 3 is made up of two large sheet flows of SSW-NNW orientations and

issued from the southern caldera wall. The surface of these flows is very flat (1490m of water depth) and becomes rougher towards the NE, away from the emission area. Episode 4, between 1490 and 1470m of water depth, consists of a continuous front of pillow lava lining the south wall of the caldera. These lavas, extremely recent, are devoid of sediments and overlap the lava flows of previous episodes. The last episode, episode 5, results in narrow flows (20 to 50 m wide for 200 to 400 m long) parallel and close to (< 100m) the south wall of the caldera and surmounting the lava of episode 4.

4 Composition of the volcanic rocks

4.1 Petrography

All of the samples recovered by dredges and/or during submersible dives are very fresh. Most of the samples have glassy textures and contain abundant vesicles. A few samples have doleritic textures and contain plagioclase, clinopyroxene, olivine, and oxides. The remainder has a homogeneous mineralogy. Phenocrysts, when present, are of plagioclase and olivine and more rarely, clinopyroxene. The same minerals comprise the groundmass and are disseminated in the glassy matrix. In many samples, the plagioclase and olivine crystals show evidence for quenched crystallization, with dendritic shapes for the olivines and box shapes for the plagioclases.

4.2 Geochemistry

Major and trace element concentrations were measured in 49 samples. The samples were recovered by dredging or from submersible dives performed along the caldera walls, inside the caldera, on the resurgent dome and on two satellite volcanoes located on either side of the caldera. The samples can be divided into two groups. Group 1 comprises rocks that were sampled along the caldera walls and old tectonized lava flows. These represent the oldest part of the volcano and were emplaced before caldera formation. Group 2 comprises samples retrieved from the floor of the caldera and the two satellite volcanoes at the top of the caldera. Recent lava flows, from which the samples were collected by submersible, overlap the scree at the base of the caldera walls, thus these lavas are clearly younger and erupted after the caldera collapse.

The Group 1 samples show the largest variations in major elements, with 48.6 to 61.5 wt% SiO₂ and corresponding MgO from 8.4 to 2.1 wt%. The Group 2 lavas show a more restricted range of compositions: 49.2 to 51.9 wt% SiO₂ and 3.9 to 7.7 wt% MgO. In the Total Alkali Silica (TAS) diagram (Figure 6 and table 2) all samples plot in the subalkaline field, close to

the boundary between low-K and medium-K volcanic rocks, and range from basalt to andesite in composition.

The major element compositions (Figures 6 and 7) suggest that the two groups of samples have different evolution trends. The Group 1 samples form a standard fractional crystallization trend, indicating removal of plagioclase, clinopyroxene, and olivine from the onset of fractionation with Fe-Ti oxides from 52 wt% silica. These minerals correspond to those observed in thin section. The Group 2 lavas, however, form a different trend in these diagrams, with lower Al, Ca and Mg contents but higher Fe and Ti contents than the Group 1 lavas.

Rare earth element patterns (Figure 8) show a broad range of compositions for Group 1 lavas, but with parallel patterns and a deeper europium negative anomaly for the most differentiated samples. These patterns are compatible with evolution via fractional crystallization. As expected, Group 2 lavas have a more restricted range of REE compositions, although their patterns are parallel to those of Group 1.

5 Hydrothermal mineralization

Six different types of mineralization (figures 9 and 10) were observed and sampled during the dives: 1) Sulfide impregnation in highly altered volcanic rocks; 2) Fe/Mn low-temperature precipitates; 3) low-temperature Fe/Si hydrothermal sediments; 4) high-temperature sulfide/sulfate chimneys; 5) active siliceous chimneys; and 6) iron oxides. In addition, also, areas with low-temperature venting without mineral precipitates were observed in many places along the western and southern caldera walls. Observations made during the dives were used to confirm the different acoustic facies identified and delineated in the seafloor backscatter images. These maps show the presence of several extensive hydrothermal sediment formations on the floor of the caldera and zones of hydrothermal alteration on the caldera walls (Figures 11 and 12). No large VMS formation was found. However, the large number and diversity of hydrothermal mineralization allows us discuss the evolution of the different types of mineralization in relation to the caldera geometry and the formation history.

5.1 Alteration and stockwork mineralization along the caldera wall

Several zones of intense hydrothermal alteration including disseminated sulfides and local stockwork mineralization were found along the caldera walls. The alteration is extensive and is clearly visible in several places in the high-resolution seafloor backscatter images (Figure

11). One dredge and two submersible dives were carried out in this locality. One of the largest zone is located near the top of the southern wall of the caldera (Figure 11). A traverse made during dive #5 revealed scree composed of white blocks of altered lava at the base, overlain by a steep slope containing larger ($> 1\text{ m}$) white blocks of altered lava at depth of between 1340 and 1260 m. The top of the caldera wall (1200 m) comprises a vertical cliff composed of prismatic lava flows. The alteration zone consists of discolored and bleached lavas impregnated with sulfide that have left ocher residues on the rock surfaces. Some larger blocks exhibits a network of oxidized sulfide veins, indicative of an earlier stockwork zone that has been exposed after the collapse of the caldera along normal faults. A second extensive alteration zone of this type was visited in the NW sector of the caldera, near to the base of the caldera wall and close to the margins of recent lava flows (Figure 13). Two large areas ($100 \times 40 \text{ m}$ and $100 \text{ m} \times 80 \text{ m}$) of altered gray basalt with disseminated pyrite were sampled.

5.2 Low temperature Fe/Mn and Si/Fe mineralization

An inactive hydrothermal site located along the northern upper rim of the caldera (Figure 9) consists of black manganese oxides covering orange iron oxides. This hydrothermal mineralization, which is up to several tens of cm thick, formed on a series of basalt flows and, covers an area several hundred square meters in size. No chimneys were observed in the area.

Inactive accumulations of silica and iron oxides are relatively abundant on the caldera floor at Kulo Lasi (Figures 9 and 12), occurring near the caldera walls throughout the southwestern, northwestern, and northeastern parts. These areas were seen as low backscatter intensity patches in images acquired during the AUV survey and correspond to older low-temperature hydrothermal sites. Several zones of the same type were visited during dives # 3, 4 and 5. These appears to form brown-ocher irregular patches of hydrothermal sediments varying from a few tens of meters to $10\,000 \text{ m}^2$ in size. The contact between the hydrothermal sediments and the lava flows is distinct. None of these sites were seen to be active. Most of the formations are located on the outer parts of the caldera floor and a short distance from the base of the walls. Very often, white pelagic sediments almost entirely cover the earlier very friable hydrothermal formations. These have black surfaces covered with manganese oxides and a core of ocher-colored material and are rarely more than a meter thick. Since they lie both on tilted old lava flows (NE part of the caldera) and on undisturbed recent lava flows (SW part of the caldera floor), most of the hydrothermal sediment formations, even not

active, are considered to be relatively young. They also occur preferentially at the outer part of the caldera.

5.3 Light orange chimneys near the base of the caldera wall

A field of siliceous chimneys was seen along the northern edge of the caldera, less than 2 km east of the black smoker field. Here, the northern wall is marked by older lava scree. The chimneys are composed of a light-orange porous material and measure up to 5 m in height. The maximum temperature of the transparent fluid is 43°C (Figure 10C). The chimneys were not sampled, but from their texture, color, the absence of dark color in the broken parts, and the temperature of venting they appear to be similar to the classical silica chimneys observed at many hydrothermal sites in the world's oceans.

5.4 Active low-temperature diffusion with no precipitates

In the southwestern sector of the caldera (Figure 9), three small (< 50 m) active hydrothermal sites are currently discharging transparent fluids. In the largest area, a depression is filled with ocher-colored hydrothermal sediment. Close to this site, shimmering and transparent fluids are being released along a fault scarp (1418 m water depth) at a temperature below 6°C. The biota here is very dense and varied and comprises various fixed and mobile organisms including white and pink shrimps, crabs, large clusters of siboglinidae worms (Figure 10D) fish mussels and galathea crabs. Two other active vent areas of this type, at water depths of 1372 m and 1411 m respectively, occur further west along the slope (Figure 9).

5.5 Active black smoker chimneys on the caldera floor

5.5.1 Location

Chimneys (either active or inactive) were observed in four of the six dives (Figures 10), with all sulfides-rich chimneys located on the caldera floor in the north-western and eastern sectors of the caldera (Figures 9 and 13). They systematically occur on the most recent lava flows that have been stained by iron oxides. Iron oxides also accumulate as sediments in the depressions. Although small in size (up to 1-2 m high), the chimneys are locally quite numerous and very active. However, they are all located near the base of the caldera wall, and no high-temperature activity was seen in the central zone of the caldera, or on the walls themselves. The average distance between the chimneys and the wall is 145 m (min. 80 m, max. 240 m). The main active areas in the NW of the caldera (Figure 13) strike S-SW/N-NW for approximately 0.5 km and forms a relatively narrow (< 50 m) alignment near the base of the western caldera wall (Figure 13). The water depths of the high-temperature active

chimneys are relatively constant, between 1476 and 1478 m in the west and at 1462 m depth in the east. Marine organisms are abundant near many chimneys and in areas of diffuse venting. However, only mobile animals such as shrimps, galathea crabs, and fish were seen. No sessile animals were observed, suggesting, like the small size of the chimneys, that hydrothermal activity is extremely recent in this area. Two types of smokers are in the NW area (Figure 13): black smokers with temperatures up to 345°C and clear smokers with temperature of up to 332°C. Anhydrite is the dominant mineral in both types of chimneys. The second active area with sulfide chimneys is on the eastern side of the caldera where chimneys located near a recent crack expel a transparent fluid. The temperature was not measured on these chimneys.

5.5.2 Fluid composition

Three smokers on the north-western of the caldera floor were sampled at a water depth of 1475m (Figure 13). Two of the smokers were short narrow chimneys and the other had a beehive structure. Composition of collected fluid is shown on table 3 (Konn et al. , 2018). The fluid sampled from the beehive-form smoker was found to be significantly mixed with seawater (Mg: 49.8 mM, T°: 134°C and pH: 6.07). Black and translucent fluids with similar temperatures (342 and 343°C) were observed at the different smokers. The clear fluids have a chlorinity (497 mM) lower than seawater, while the black fluids have a chlorinity of up to 735 mM and higher than seawater. Silica concentrations range from 8.4 mM in fluids with low chlorinity, to 14.6 mM in fluids with high chlorinity. The high chlorinity fluids are enriched in metals (maximum: Fe: 9884 µM, Mn: 1416 µM, Cu: 20µM, and Zn: 175 µM) compared to the low chlorinity fluids (maximum: Fe: 4796 µM, Mn: 862 µM, Cu:14µM, and Zn: 45 µM. K (16.6 to 29 mM), Li (149 to 249 µM) and Rb (26 to 46 µM) are highly enriched compared to seawater. Na concentrations (388 to 612 mM) are principally linked to the chlorinity. Ca concentrations (10.8 to 11.6 mM) are similar to that of seawater). Sr (3.4 to 9.4 µM) is extremely depleted compared to seawater.

One specific characteristic is the high concentration of Mg (24.6 and 26.5 mM) and SO_4^{2-} (8.8 and 6.2 mM) in the fluids despite the very low pH levels (2.24 and 2.36). CH_4 is nearly absent (<0.01 mM) whereas the H_2 concentration is variable at a very high level in all samples (1 to 19.8 mM). H_2S is up to 5 mM in the low chlorinity fluid, and 0.4 mM in the high chlorinity one. The concentration in CO_2 (5.1 mM) is low compared to magmatic dominated fluids (e.g. up to 206 mM in magmatic dominated fluids at Brother seamount (de Ronde et al. , 2011). Due to different solubility, CO_2 and sulfur gases (SO_2 and H_2S) exsolved at different pressures and may be decoupled (de Ronde and Stucker, 2015). Fluids from the Kulo Lasi volcano are clearly enriched in organic carbon (TOC = 1.78-3.40 mg/L compared to seawater ~0.5 mg/L

(Bennett et al. , 2011, Lang et al. , 2006, Sharp et al. , 2002). Both aliphatic and aromatic hydrocarbons, as well as fatty acids, were present. The fluids are also enriched in longer chain alkanes (mainly C10 and $\geq$ C16), aromatic hydrocarbons (mainly naphthalene, fluorene, and anthracene mainly) and linear fatty acids (C12 and C16).

6 Composition of the hydrothermal mineralization

6.1 Mineral and chemical compositions of chimneys

The mineralogy of 16 samples was studied using X-Ray diffraction (major minerals) and reflected light microscopy (major and trace minerals and paragenetic studies). The sulfate chimneys are mainly composed of anhydrite with disseminated pyrrhotite (Figures 14, 15 and Table 4) and rare chalcopyrite and sphalerite. Anhydrite represents almost 100% of the sample in some cases. Some samples contain significant amounts of pyrrhotite in the form of monoclinic rods. Euhedral pyrite was found to be more common near the exterior of the chimneys (Figure 15-E). Pyrrhotite blades are commonly corroded and partly replaced by pyrite (Figure 15-H).

Most of the inactive chimneys are dominated by sulfide (Table 4). At the core of the chimneys pyrrhotite (70-80%) is dominant and pyrite is relatively rare, (Table 4 and figure 15) while sphalerite is dominant (70%) near the exterior margins of the chimneys. The sphalerite content gradually increases from the core towards the external part of the chimney where it sometimes forms a massive layer (Figure 14-D, E, G). Chalcopyrite and isocubanite are less abundant and are localized towards the center of the chimneys (Figure 14-H and 15-B, E). Pyrite occurs in the outer margin together with sulfates (barite $\pm$ anhydrite) and iron oxides. Pyrrhotite occurs as blades or hexagons in sizes ranging from a few microns to hundreds of microns (Figure 15-A, C, D). These blades are often mantled by overgrowths of isocubanite and/or sphalerite and sometimes contain inclusions of these minerals. Some sphalerite inclusions form concentric structures. The sphalerite is generally in the form of ovoid aggregates (Figure 14-E, G) with massive to semi-massive or collomorph textures and occurs in the inferred low-temperature outer part of the chimney. Isocubanite appears as anhedral crystals within the massive sphalerite, ranging in size from a few microns to tens of microns. The isocubanite has generally crystallized around pyrrhotite or sphalerite crystals

but can also occur in isolated forms in association with pyrrhotite, or line the hydrothermal conduits. Chalcopyrite frequently occurs as exsolutions in isocubanite (Figure 14-H) and commonly replaces sphalerite. One of the chimneys in the eastern sector of the caldera is dominated by pyrite. However, the presence of pyrrhotite ghost aggregates (Figure 15-F) is indicative of an early hydrothermal stage dominated by pyrrhotite. In other chimneys, pyrite associated with sphalerite is sub-automorphic and replaces pyrrhotite (Figure 15-D). Galena is often observed within sphalerite (Figure 14-B, D, E) and is mainly sub-automorphic, or xenomorphic, but sometimes present a skeletal habit. The galena is much more abundant in colloform sphalerite in the outer parts of the chimneys and gradually disappears towards the copper-sulfide-rich inner parts. Barite and opal are the most frequent gangue mineral phases. When present, opal is distributed homogeneously as a thin and late precipitate around sulfides and locally may replace pyrrhotite or sphalerite. Opal is considered to be a late phase since its precipitation occurs at relatively low temperatures (i.e., ~ 50-100°C) (Scott, 1997). Barite preferentially develops in the outer chimney zones in association with sphalerite (Figure 14-D) and pyrite.

Chemical compositions of various hydrothermal precipitates were determined by XRF and are presented in table 4. For individual chimneys, several analyses were carried out, each analysis corresponding to either the complete chimney or to a particular zone within the chimney (e.g., a zone rich in pyrrhotite or sphalerite). Analyses can be divided into three groups: (1) sulfate-rich active chimneys dominated by anhydrite. (2) Zn-rich, representing areas enriched in sphalerite (Zn > 10 wt%) and (3) Fe-rich dominated by pyrrhotite (Fe > 40 wt%). The sulfate-sulfide chimneys show a wide range of Zn concentrations (from 0.2 to 47.7 wt%, average ~10 wt%) and exhibit relatively low Cu contents (0.2 - 2 wt%: Average ≈ 1 wt%). Cobalt (up to 2664 ppm), selenium (up to 867 ppm), tin (up to 694 ppm), and gallium (up to 734 ppm) are, in turn, characterized by high values. Concentrations of Pb (up to 1.5 wt%), Ba, (up to 5 wt%) As (up to 1724) and Ge (up to 324 ppm) are relatively high. Average compositions of sulfide mineralization in other tectonic environments in the modern seafloor are presented in table 5.

6.2 Mineral and Chemical compositions of hydrothermally altered lavas

Eleven altered volcanic samples were analyzed to define the main mineralogical and chemical characteristics of the alteration zones. The white color of the leucocratic rocks (Figure 16) indicates strong alteration of the mafic minerals. Among the different alteration minerals present, chlorite, nontronite, and pyrite were identified by XRD in all of the samples.

The presence of plagioclase and clinopyroxene in the altered samples reflects the mineralogy of the basaltic protolith. Pyritization is relatively high and occurs as millimeter-sized grains of pyrite that may represent over 10% of the sample (Figure 16-C). Petrographic studies confirmed the presence of large quantities of subhedral to euhedral pyrites and showed the presence of chalcopyrite, either as isolated grains or as inclusions in pyrite (Figure 16-D, E). Rare pyrrhotite inclusions were also observed in pyrite (Figure 16-F).

The chemical composition of the hydrothermally altered basalt reveals several stages of alterations. Silica has been leached from most samples, and SiO_2 concentrations are generally lower than 50% (31 to 46 %). Only one sample contained up to 55% SiO_2 ; this sample also has the highest quartz concentration and is probably silicified. Total sulfur is high (up to 9.4 %) and reflects the presence of disseminated sulfides. Compared to fresh basalt (7.7 % MgO), all samples, except one, exhibit Mg enrichment (up to 14 % MgO). Potassium (as low as 0.04 %) has been leached from most samples. Ca (average 0.5 % CaO) was completely leached in the alteration zone at the base of the caldera wall (fresh basalt: 12.3% CaO). In the alteration zone near the top of the caldera wall, the Ca is also leached (average 7 % CaO). Only one sample, in which calcite was identified is enriched in calcium. Trace element analysis shows an intense leaching of Rb in contrast to enrichments in V and Ni. Barium up to 370 ppm is slightly enriched in some samples. However, no barite was detected by XRD on the altered samples.

6.3 Mineral and chemical compositions of Si/Fe hydrothermal sediments

The Fe/Si metalliferous sediments are mainly composed of iron oxides, clay minerals (nontronite) and amorphous silica. The poorly crystalline character of some of these mineral phases (ex. opal, goethite, nontronite) makes them difficult to identify using XRD. Amorphous silica likely represents the major component of the sediments.

The results of XRF analysis of these hydrothermal sediments are given in Table 6. Data for the major elements indicate high concentrations of SiO_2 (26.2 to 86.6 wt%) and Fe_2O_3 (4.6. to 53.6 wt%), relatively high concentrations of Na_2O (1.8 to 3.4 wt%) and MgO (0.4 wt%) but relatively low concentrations of manganese and aluminum (e.g., MnO <1.4 wt%; Al_2O_3 <0.6 wt%, respectively). This suggests that nontronite is very low in aluminum and manganese oxides (i.e., vernardite, birnessite, or todorokite) represent minor components. These mineralization are characteristic of low-temperature (<100°C) hydrothermal precipitates.

6.4 Sulfur Isotopes

Fourteen sulfur isotope analyzes were performed on mineral fractions of pyrrhotite, chalcopyrite, sphalerite, pyrite and anhydrite from four inactive and one active chimney. One anhydrite exhibits a high positive isotopic signature of $\delta^{34}\text{S} = 21.11\text{‰}$ from seawater sulfate (20.9 ‰). The data obtained for the sulfides are consistent ranging from 1.3 to 3.5 ‰ with an average of 2.3 ‰. Two sphalerite samples and one pyrrhotite sample from the FU-PL06-05 chimney yielded the highest $\delta^{34}\text{S}$ values ($> 3\text{‰}$). The high-temperature minerals, such as chalcopyrite, and pyrrhotite have lower isotope signatures ($< 3\text{‰}$). The lowest values (i.e., 1.25 ‰) were obtained on pyrrhotite sampled by micro drilling at the core of the chimneys. Three pyrite fractions from samples of three different altered volcanic rocks were also separated for sulfur isotope analysis. The $\delta^{34}\text{S}$ values are higher than those observed for the sulfides from the chimneys and range between 5 and 9 ‰. The $\delta^{34}\text{S}$ of the native sulfur is 3.12 ‰.

7 Discussion

7.1 Regional geological setting and implications of lava composition for the formation and evolution of Kulo Lasi volcano.

At Kulo Lasi, the two groups of magma appear to have followed a different crystallization path and did not reach the same degree of differentiation. As the CaO, MgO and TiO_2 contents of the two groups overlap, it is possible that the two magmas had the same starting composition and were derived from the same source. According to the REE patterns, the two series are from a similar magma source but with different evolutionary trends within the magma. Group 2 samples clearly represent a new batch of magma ascending through the Kulo Lasi volcano after the caldera collapse, but with a shorter residence time in the volcano and a different crystallization sequence. Group 2 samples are younger from field observations, we also show that they are less differentiated (see major element diagrams) and they possibly have the same parental magma (see major element diagrams and parallel REE pattern, Figures 6, 7 and 8).

The trace element compositions of lavas sampled from large volcanoes in the Lau and Tonga-Kermadec arcs and back-arcs are consistent with their geodynamical setting. They show substantial enrichments in fluid mobile elements (e.g., LILE large-Ion Lithophile Elements, such as Rb, Ba, K, and Sr) and low concentrations of fluid immobile high field strength elements (e.g., HFSE such as Zr, Hf, Nb, and Ta) (Hawkesworth et al. , 1993,

Tatsumi and Eggins, 1995). These geochemical characteristics are recognized in arc lavas, and to a lesser extent, in back-arc volcanics. In contrast, the samples from Kulo Lasi volcano are not LILE-enriched and do not show HFSE negative anomalies and thus, they do not exhibit the normal geochemical characteristics of back-arc volcanism. In the very complex tectonic setting of the Futuna area, this indicates that the magmatic activity in the Kulo Lasi region is not directly related to subduction.

Several large volcanoes with calderas have been described in the Manus, Lau and Tonga-Kermadec arcs and back-arcs in the SW Pacific area where the Futuna area is also located (see Table 1). Most back-arc calderas are dominated by basaltic or slightly differentiated rocks, while, felsic volcanic rocks and associated pyroclastic products are common in arcs. The felsic volcanoes have steep external slopes, with a large caldera at the summit and are associated with large deposits of silica-rich pyroclastic products (Barker et al. , 2013, Graham et al. , 2008, Wright et al. , 2006). Kulo Lasi volcano is very different from the se large felsic volcanoes. Its very flat shape is more like that of a shield volcano formed mostly from the extrusion and accumulation of mafic lavas. This is also consistent with the fact that the most differentiated rocks sampled so far on Kulo Lasi have only intermediate compositions with no more than 61.5 wt% silica. No ignimbrite or pumices were sampled by the dredges or observed during the dives. Thus, we suggest that the caldera probably formed through incremental collapse due to frequent and massive outpouring of mafic to intermediate magmas from the summit or flanks of the volcano. The absence of highly differentiated lavas might be explained by the regular input of new basaltic lava to the shallow magma chamber. This is compatible with the intense volcanic activity of the whole SE Futuna area as suggested by the lack of sediments in seafloor backscatter images (Figure 2) in this area and by direct observation during several of the submersible dives. Compared to the surrounding areas, Kulo Lasi volcano is non tectonized. Furthermore, several dives in the recent volcanic area 20 km north of Kulo Lasi revealed that the most recent lavas erupted on the caldera floor. Thus, Kulo Lasi can be considered the focus of the most recent volcanic activity at the intersection between the regional NE-SW and N-S fault systems (Figures 2 and 3).

7.2 Caldera formation and fluid pathways

The structure and shape of collapse and resurgence are often controlled by preexisting discontinuities (Acocella et al. , 2004). The emplacement of the Kulo Lasi volcano is regionally controlled by the crossing of the N25 directions visible at the south and west of the volcano with the N70 direction largely expressed in the north (Figure 2). These two regional

fault systems are also seen within the caldera. Part of the western collapsed area is controlled by N25 faults. The N70 directions are better expressed at the top and the west side of the resurgent dome. The two types of faults also control the southern extension of the resurgent dome. Thus, the structure and shape of collapse and resurgence are partly controlled by preexisting regional discontinuities and probably explain the slight E-W elongation of the caldera.

Collapse experiments show that the outward dipping reverses ring faults generate the first depression and are followed by an outer concentric depression formed by the inwards dipping normal ring faults (Acocella et al. , 2000). Reverse faults propagate upwards from the magma chamber and then normal superficial faults initiate from tension fractures propagating downwards from the surface (Burchardt and Walter, 2010). When these structures intersect, the resulting higher permeability is favorable for hydrothermal circulation. Cold seawater is drawn down through the normal faults and returns as hot metal-rich fluids along the reverse faults, favoring VMS formation at the base of the caldera walls (de Ronde et al. , 2011, Embley et al. , 2012, Mueller et al. , 2009). In general, the largest massive sulfide deposits preferentially form at the base of the caldera walls. The heat from the magma chamber drives high-temperature convective fluids. These fluids may deposit mineralization where the main fault intersects the seafloor, or where there are permeable pyroclastic rocks in the area of maximum subsidence (Stix et al. , 2003). Massive sulfide formation is favored during two stages: (1) during and immediately after caldera collapse and (2) during post-caldera resurgence (de Ronde et al. , 2014b).

Theoretical models also indicate that reverse ring faults controlling caldera subsidence will form at the margins of the magma chamber. Consequently, the area covered by the resulting caldera depression will be of the same order as the surface projection of the underlying magma reservoir (Marti et al. , 2008). Calderas develop by increasing their depth/diameter ratio and evolve from a depression controlled by the collapse along the deep reverse ring faults (= structural diameter - D_s) to an enlargement of the diameter controlled by erosion along the wall (= tectonic diameter - D_t) (Geshi et al. , 2012). Mature calderas are characterized by $S_s/D_s > 0.07$ (S_s = structural depth of the caldera) (Geshi et al. , 2012). At Kulo Lasi, the position of the deep reverse faults is materialized by the last eruptive episode rooted on annular dikes close to the base of the walls. Based on the position of these faults, the structural diameter (D_s) of the caldera is 3.7 km, thus for a structural depth (S_s) of 300 m, the S_s/D_s ratio of 0.08 indicates a mature stage for the caldera evolution.

Numeric and analog models show that mafic calderas are more favorable to the injection of dikes in circumferential fissures (Corbi et al. , 2016). At Kulo Lasi, traces of the annular reverse faults are given by the narrow and most recent lava flows that occur on the caldera floor close to the base of the caldera wall (Figures 5 and 13). Here, the most recent flows form annular structures parallel to the caldera wall, and we interpret these as having formed in response to recent diking along reverse faults focusing the lavas towards the seafloor. As the black smokers are aligned along these annular structures, it can be deduced that the high-temperature hydrothermal discharge is also most likely controlled by the normal reverse faults. No recent lava flows are emitted at the top or from the base of the resurgent dome. The eruptive fissures are the surface expression of dike emplacement controlled by the state of stress in the volcano (Chadwick and Howard, 1991). This probably explains the situation observed at Kulo Lasi and is in agreement with results of analog experiments (Scandone and Acocella, 2007) where low aspect ratio (<1.6) form a coherent collapse ensuing connection from the chamber to the surface and the creation of continuous ring conduit feeding annular vents. Unlike our observation at Kulo Lasi, active calderas are seldom associated with circumferential eruptive fissures. However, eroded magmatic complexes reveal widespread circumferential dikes (Corbi et al. , 2016). In addition, at Kulo Lasi, no hydrothermal fields are seen at the base of the resurgent dome or on the central part of the caldera floor, indicating that the focus of lava, and thus of heat, at the periphery of the caldera controls the active hydrothermal discharge (Figure 13). All this suggests that at Kulo Lasi, recent diking, recent lava flows, and hydrothermal activity are controlled by the deep reverse faults. In fact, dike intrusion along the deep ring faults depressurize the reservoir inducing subsidence of the above roof and opening of the faults (Acocella, 2007, Gudmundsson, 2016). This is the case at the Kulo Lasi where the deepest part of the caldera is located in the external part of the caldera floor (Figure 5). The only large cracks observed are also located at the foot of the east wall of the caldera. In addition, topographical observations show that the floor of the caldera has locally subsided near the wall after the emission of the first two generations of recent flows. For example, near the south wall, from where they are emitted, some flows are 5 to 10 m lower than their distal effusion zone towards the center of the caldera (see flow 3 in figure 5). All these observations testify to a control of the recent volcanic and hydrothermal activities by recent dykes injected in the reverse faults. In addition, this geometry will facilitate the descent of seawater at the outer part of the caldera.

Similar control of the hydrothermal system by injection of a recent dike at the border of the caldera has also been proposed for the Brothers volcano (de Ronde et al. , 2011). However, at the Brothers caldera, there is no morphologically well-defined continuous ring fault wall

(Embley et al. , 2012). On the Clark volcano, during the caldera stage hydrothermal activity was focused along the caldera ring faults (de Ronde et al. , 2014b).

7.3 Depth of the reaction zone

The geometry of the collapsed structure depends strongly on the size and depth of the magma chamber (Cole et al. , 2005, Kusumoto and Takemura, 2005, Roche et al. , 2000). According to end-member mechanisms proposed for caldera collapse, the Kulo Lasi has the characteristics of a piston model (Cole et al. , 2005, Lipman, 1997, Roche et al. , 2000). This model has a roof aspect ratio R (thickness /width of the magma chamber roof) of < 1.4 (Roche and Druitt, 2001). The depth of the magma chamber can be estimated from the ratio of the area tectonized by normal faults on the periphery of the caldera (annular extension zone) to the total surface of the caldera (Roche et al. , 2000). For the Kulo Lasi, the extension zone covers 10.04 km^2 (~50%) of the total surface of the caldera (20.2 km^2). According to the model proposed by Roche (Roche et al. , 2000), 50 % of extensional zone corresponds to a roof aspect ratio above the magma chamber (thickness/width) of 0.25. Considering the maximum width of the caldera (5.3 km) the depth of the magma chamber before the caldera collapse is thus estimated to be at about 1.3 km.

The resurgent dome is very well defined and surrounded by relatively flat floor. Its slopes are made of fragmented pillow lava. The absence of sediments on the talus indicates that the resurgent dome is a very recent structure. Experimental results show that there is a close correlation between the final shapes of the uplifted area and those of the intrusions (Galland, 2012). The asymmetry of the uprising depends directly on the asymmetry of the intrusion. In addition, the uplifted zones are directly located above the shallow part of the intrusion (Galland, 2012) and the deformation pattern directly reflects the 3D shape of the underlying intrusion. Thus, at Kulo Lasi the position of the resurgent dome in the NW quarter of the caldera floor indicates the shallowest part of the new magma chamber. This is also coherent with the position of the high-temperature vents in the NW part of the caldera. No extrusion of recent lava is observed on the resurgent dome and there is no graben at its top. Only preexisting N25 and N70 faults are reactivated. This can be interpreted as the rise of a rigid block, as described by Acocella for both analog models and field observations (Acocella et al. , 2001). This morphology corresponds to an aspect ratio (depth of the chamber/width of the resurgent block) close to 1 (Acocella et al. , 2001). The diameter of the resurgent dome is 1.6 km, thus for an aspect ratio of 1, the depth of the new magma chamber should be close to 1.6 km under the resurgent dome. This indicates that the top of the new magma chamber may be slightly deeper than the magma chamber existing prior to the collapse of the caldera.

This also promotes the rheological model recently proposed by Galetto (Galetto et al. , 2017) and where the formation of a caldera creates, at the level of the chamber, a transition zone of residual magma. In this model, the viscosity contrast between the residual magma chamber and the newly injected magma underneath provides a rheological barrier that favors the trapping and accumulation of new magma at depth and hinders the propagation upwards of the newly injected magma. This process allows the formation of a resurgent dome and of a large magma chamber (Galetto et al. , 2017). In this configuration, diking and eruptions occur preferentially at the periphery of the system along the deep reverse faults.

Even if andesitic lava are found in the Kulo Lasi caldera, no explosive products are observed. Geochemical and mineral variability in the composition of lava may be used to probe the conditions of magma storage and processes that drive magma evolution of caldera-producing magmatic systems. Geochemical and petrological studies are underway to discuss this issue for the Kulo Lasi volcano and will be published in a specific paper.

7.4 Type of early alteration

The Mg enrichment in the altered volcanic rocks probably reflects a loss of magnesium from seawater in the descending part of a convective cell. This magnesium is most likely taken up by chlorite (Mottl, 1983). Mineral assemblages in Desmos and the nearby PacManus hydrothermal systems of the Manus Basin, are interpreted to have formed during advanced argillic alteration by acid-sulfate fluids, and typically contain Na- and K-bearing alunite, pyrophyllite, silica phases, anhydrite and pyrite, but no Mg-bearing minerals (Gena et al. , 2001, Lackschewitz et al. , 2004, Paulick and Bach, 2006, Seewald et al. , 2015). The same assemblage has been described for the NW Rota-1 volcano in the Mariana arc and for the Brothers volcano (de Ronde et al. , 2005). At Kulo Lasi volcano, the presence of chalcopyrite typically reflects high-temperature hydrothermal alteration ($> 300^{\circ}\text{C}$), while the pyrrhotite could indicate somewhat more reducing conditions during the early stages of alteration. No specific minerals, indicative of magmatic fluid and acid-sulfate alteration, such as alunite or pyrophyllite, were identified in our samples. Thus, during the first hydrothermal stage at Kulo Lasi, we should consider a classical water/rock type hydrothermal system, with high-temperature fluids and deposition of pyrite and chalcopyrite as disseminated and stockwork mineralization during the fluid ascent. This is similar to the water/rock-dominated system discussed by de Ronde and Stucker (de Ronde and Stucker, 2015) for arc and back-arc volcanoes.

At Kulo Lasi volcano, the alteration outcrops are found on the caldera walls, with significant alteration and stockwork mineralization near the top of the wall. This indicates that this alteration occurred prior to, or during, the early stages of the caldera formation. The occurrence of extensive alteration also favors a model of a progressive collapse of the caldera rather than a single rapid event as described by Geshi (Geshi et al. , 2002). At Brothers volcano, the distribution of alteration zones and extinct and active chimneys has also been suggested to infer 'progressive' migration of the hydrothermal system as the caldera was formed (de Ronde et al. , 2005, Embley et al. , 2012). A study of Clark volcano has suggested that hydrothermal activity associated with caldera formation can then be superseded by the building of a cone inside the caldera (de Ronde et al. , 2014b).

7.5 Fluid composition

7.5.1 Unusual composition of the fluid

In modern submarine hydrothermal systems, two types of fluids may be present: evolved seawater generated by water/rock-dominated convective systems, and magmatic fluids. During water/rock-dominated hydrothermal circulation, Mg and SO_4^{2-} from seawater are completely trapped in the altered volcanic rocks, and a pure hydrothermal fluid contains no Mg or SO_4^{2-} (Seewald et al. , 2015, Von Damm et al. , 1985). Pure hydrothermal end-members can, therefore, be calculated by extrapolation of the hydrothermal composition to Mg = 0 (Von Damm et al. , 1985). The high temperature fluids (>300°C) from the Kulo Lasi (Table 3) have unusually high concentrations of Mg (24.6 to 34.9 mM) and SO_4^{2-} (6.2 to 12.0 mM). However, the low pH (2.24 to 3.32) and the high temperature (338-343°C) of these fluids indicate that significant mixing with cold seawater at subsurface or during sampling is rather unlikely (Konn et al. , 2018). Mixing with cold seawater would not maintain the temperature of 343°C and the low pH while increasing the concentration of Mg from 0 to ~50% of the seawater concentration.

7.5.2 Evidence for phase separation

At their sampled seafloor conditions the Kulo Lasi vent fluids (P=147 bars and T=343°C) are located on the two-phase curve for seawater with low CO_2 concentration (Bischoff and Rosenbauer, 1984, Monecke et al. , 2014) and within the field of subcritical boiling. Thus, variations in chlorinity, from 34% higher to 9% lower than that of the seawater, can be explained by a process of phase separation (Bischoff and Rosenbauer, 1987). An adiabatic decompression of supercritical seawater (Bischoff and Pitzer, 1989, Foustoukos and Seyfried, 2007a) would cause it to separate, at about 320-350 bars and 415-420°C (Konn et al. , 2018) into two phases having the NaCl percentages observed at Kulo Lasi.

7.5.3 Magmatic fluid contribution?

A possibility to explain high Mg and SO_4^{2-} concentrations at Kulo Lasi is the contribution of magmatic fluids. Magmatic-hydrothermal systems are thought to make up around 70% of hydrothermal systems related to arc volcanoes and are also present in back-arcs (de Ronde and Stucker, 2015). In arc and back-arc environments, degassing of SO_2 reflects a fundamental difference in the speciation of sulfur in magma. In general, sulfur extracted from volcanic rocks occurs as H_2S in water/rock-dominated fluids, whereas magmatic SO_2 strongly partitions into aqueous fluids exsolved from magma (Scaillet and Pichavant, 2003). Magmatically-derived SO_2 undergoes disproportionation in the upflow zone to produce reduced and oxidized sulfur species upon cooling (Butterfield et al. , 2011, de Ronde et al. , 2015, de Ronde et al. , 2005, de Ronde et al. , 2011, Embley et al. , 2006, Gamo et al. , 1997, Herzig et al. , 1998, Resing et al. , 2007, Seewald et al. , 2015) according to the reactions:

This process contributes to a lowering of pH and a rise in SO_4^{2-} and S^0 . Elemental sulfur is the most commonly deposited phase in magmatic-hydrothermal systems and can form lakes of molten sulfur (de Ronde et al. , 2015, de Ronde and Stucker, 2015). These acid-sulfate fluids represent a different style of hydrothermal activity to black smoker fluids at oceanic spreading centers where water/rock-dominated fluid venting is dominant. In particular, the magmatic fluids are more acidic and contain substantially higher aqueous SO_4^{2-} and Mg. They also show similarities to the fluids responsible for high sulfidation epithermal deposits in subaerial settings (Hedenquist and Lowenstern, 1994, Seewald et al. , 2015). For example, at the Desmos and North Su sites in the Manus back-arc basin, where magmatic-hydrothermal fluids are venting, concentrations of Mg and SO_4^{2-} are high (Gamo et al. , 1997, Gena et al. , 2006, Seewald et al. , 2015) despite the very low pH (pH ~ 2), indicating that magmatic SO_2 has mixed with the fluid. The input of magmatic volatiles has also been demonstrated at the NW Rota-1 and Brothers arc volcanoes (Butterfield et al. , 2011, de Ronde et al. , 2011) and in the PacManus field (Yang and Scott, 1996, 2002). High magnesium concentrations in magmatic fluids make it difficult to regress the data through Mg = 0 to calculate the end-member hydrothermal fluid concentrations, as is done with water/rock systems. In this situation, sulfate has poor correlation with magnesium, and with pH due to the disproportionation reaction (equation 2) (Butterfield et al. , 2011, de Ronde et

al. , 2011, de Ronde and Stucker, 2015, McMurtry et al. , 1993, Seewald et al. , 2015, Stucker et al. , 2017). This is not the case for the Kulo Lasi fluids where the positive correlation, similar to that observed in water-rock systems, between Mg and SO_4^{2-} ($R^2 = 0.97$) and between pH and SO_4^{2-} ($R^2 = 0.95$) rather suggests mixing with seawater. Si and Fe/Mn are on a mixing line ($R^2 = 0.80$) with seawater, indicating that the silica concentrations correspond to a dilution of hydrothermal fluid with seawater. The low CO_2 concentration in the fluid implies completely degassed lava, which has lost the magmatic volatiles (CO_2 , SO_2) during the eruptive phase. Thus, low CO_2 and high H_2S probably also indicate a low contribution of magmatic fluids because the magmatic degassing is characterized by strong enrichment in CO_2 (Butterfield et al. , 2011, de Ronde et al. , 2011, de Ronde and Stucker, 2015, Lupton et al. , 2006, Lupton et al. , 2008, Seewald et al. , 2015). At Kulo Lasi, H_2S concentrations (<5 mM) are within the range of hydrothermal fluids from water-rock-dominated systems on mid-Ocean Ridges and CH_4 (<0.001 to 0.006 mM) concentrations are relatively low. Therefore, we have no evidence of magmatic fluid contribution in the collected fluids.

However, at Kulo Lasi, a magmatic episode is suggested by widespread occurrence, near the active vents, of patches of pure native sulfur (Figure 10G) at the surface of the most recent lava flows. This indicates that the reaction 2 may have been occurring on the seafloor in the past. Native sulfur is absent from the active chimneys dominated by sulfides and anhydrite. In fact, no activity is associated with sulfur deposits. Thus, even if both types of deposits grow on the surface of the same recent lava flows, it appears that the native sulfur deposits formed prior to the sulfide chimneys during an early episode of magmatic fluid venting. Indeed, mineralogical composition of the chimneys is not compatible with high-sulfidation fluids. At Kulo Lasi, FeS content in high temperature sphalerite from the core of active chimneys (average FeS: 29.5 mol%) and in inactive chimneys (9.6% FeS on the NW field and 16.2% in the Eastern field) point to a low sulfidation state (Evans et al. , 2017) with pyrrhotite and isocubanite as the stable sulfide phases. The occurrence of pyrrhotite implies relatively reducing conditions that are not compatible with the presence of SO_4^{2-} of magmatic origin in the fluid. The high H_2 concentrations recorded in all collected fluids (1 to 19.8 mM) could explain the precipitation of low- $f\text{O}_2$ - $f\text{S}_2$ mineral assemblages. In recent study, (Kawasumi and Chiba, 2017) conclude that H_2 concentrations strongly control the $f\text{S}_2$ of hydrothermal fluids and that dissolved H_2 concentrations of over 1 mm/kg result in the low $f\text{O}_2$ and $f\text{S}_2$ necessary for pyrrhotite to be stable. The replacement of pyrrhotite by pyrite in some chimneys (Figure 15H) might also reflect rapid changes in the H_2S and/or H_2 concentration in the fluid. In addition, minerals characteristic of high sulfidation magmatic fluids, such as alunite and enargite, are not found at Kulo Lasi.

In general, magmatic-hydrothermal systems contain very little Li (de Ronde and Stucker, 2015) ($<100 \mu\text{M}$) while the water/rock-dominated fluids have high Li concentrations ($> 500 \mu\text{M}$). At Kulo Lasi, the concentrations of mobile elements such as Li (266 to $519 \mu\text{M}$ in the end-member fluid) and Rb (44 to $95 \mu\text{M}$ in the end-member fluid) also point to water/rock-dominated fluids and indicate interaction with fresh volcanic rocks under conditions with low water/rock ratios (Seewald et al. , 2015, Seyfried, 1987, Von Damm, 1995). The slightly positive sulfur isotope signature (between 1.3 and 3.5 ‰) measured in sulfides from the chimneys is typical of many oceanic sulfide mineralization and suggests no contribution of magmatic SO_2 ($\delta^{34}\text{S} < 0 \text{ ‰}$).

Thus, the combination of low CO_2 , the presence of H_2S in the fluid, the positive sulfur isotopic values in sulfides, the high H_2 concentration in the fluid and the low-sulfidation characteristics of mineral assemblages are not compatible with a major contribution of magmatic fluid at the time of sampling the fluids.

7.5.4 Rapid evolution of the system

The low CH_4 concentration in the fluids collected in 2010 cannot account for the elevated concentration of CH_4 in the plume existing above the caldera at the same time. The strong difference in the CH_4/Mn ratios between the plume (0.7 – 4.5) (Konn et al. , 2016) and the sampled fluids (0.001 – 0.01) (Konn et al. , 2018) also indicate that the methane plume in 2010 is not generated by the hot fluids venting in 2010. In addition, the elevated CH_4 levels in the plume had vanished in 2011. This is most likely caused by a recent eruptive event (Konn et al. , 2016) followed by the new venting of fluids with low CH_4 concentration. The observation of very small chimneys, without fixed animals, growing on top of the most recent lava flows also indicate the beginning of a new hydrothermal episode after a major eruptive event. A rapid decrease in the CH_4 output after the eruptive event was also observed in the Endeavour segment (Lilley et al. , 2003). We thus consider that a CH_4 -poor plume, generated by the new hydrothermal episode following the eruptive event, replaces the CH_4 -rich plume, accumulated prior to the volcanic eruption. This rapid change suggests that the dike episode feeding the most recent lava flows occurred less than one year before sampling of the fluids in 2010. Unfortunately, due to bad weather conditions, time series sampling of the fluid was not possible in 2012.

7.5.5 High concentration of hydrogen

The high concentrations of hydrogen both in low salinity fluids (1.14 and 19.8 mM of H_2) and in high salinity fluids (1.04 and 6.18 mM of H_2) is unusual. Hydrogen is generally lower than

0.5 mM in most oceanic hydrothermal fields generated by water/rock reactions. In deep-sea volcanic environments, the highest H₂ enrichment (8 to 26 mM) in hydrothermal fluids is in ultramafic environments where hydrogen is generated during serpentinization (Charlou et al. , 2010). As there is no evidence for shallow mantle outcrops at Kulo Lasi, this process cannot be considered here.

A potential source of H₂ is a primary magmatic gas released through magma degassing. However, lavas in arc environments are more oxidizing than at mid-ocean ridges and point to low H₂ released from the magma (Holloway and Blank, 1994). For this reason, fluid collected from submarine magmatic-hydrothermal systems, such as the Brother Volcano, have a low H₂ concentration (maximum 67.7 µM) (de Ronde et al. , 2011). In the same way, plumes generated by acid-sulfate fluids in arc environments at Monowai volcano (Leybourne et al. , 2012b) and Rota-1 volcano (Resing et al. , 2007) are not enriched in H₂ compared to plumes related to eruptive events.

A possible explanation for the high levels of H₂ at Kulo Lasi is the reaction of seawater with hot lava according to the reaction (Holloway, 2004):

Indeed, hydrogen can be temporarily concentrated in fluids during eruptive events and episodes of diking on ocean ridges. H₂ is produced by the oxidation of Fe(II) to Fe(III) during high-temperature reactions of seawater with lava. The oxygen necessary for this oxidation is derived from the seawater (Berndt et al. , 1996, Christie et al. , 1986, McCollom and Seewald, 2007). Experimental studies have shown that fluids that have reacted with orthopyroxene achieved the highest dissolved H₂ concentration (Allen and Seyfried, 2003).

At Kilauea, large quantities of H₂ (up to 2 µM) are generated when lava flows interact with seawater (Sansone and Resing, 1995, Sansone et al. , 1991). At 9°46'N on the East Pacific Rise (EPR), hydrogen concentrations in the fluid were up to 40.2 mM two weeks after the emplacement of a new dyke (Lilley et al. , 2003) and returned to normal low concentrations (<0.5 mM) less than a year after the dike injection. Similar observations were made in the Endeavour segment of the Juan de Fuca Ridge, where hydrogen concentration in the fluid was up to 1.31 mM four months after a diking event and returned to normal values one year later (Lilley et al. , 2003). At 18°26' S on the EPR high concentration of H₂ (1.3 mM) are reported in fluids venting on extremely recent lobate flows (Charlou et al. , 1996). In back-arc

basins, H_2 is also enriched in fluids generated during eruptive events at W-Mata (16.6 mM H_2) and NW Rota-1 (2.2 mM H_2).

Immediately following eruptive events, high H_2 concentrations are also reported in the plumes from Gorda Ridge (47 nM) (Kelley et al. , 1998), Loihi volcano (20 nM) (Sansone et al. , 1996) (Sansone and Resing, 1995) and axial volcano (183 nM) (McLaughlin-West et al. , 1999). High H_2 concentration is also found in the plumes sampled during the eruptive events at West Mata (up to 15,9 nM) (Baumberger et al. , 2014) and on the NELSC (9,1 nM) (Baker et al. , 2011). Thus, H_2 detection in plumes is now considered a clear evidence of extremely recent or ongoing eruption.

Similarly, at Kulo Lasi, based on field observation and fluid chemistry, we suggest that high H_2 concentration in the fluid is generated when seawater is heated in contact with the still-hot dykes recently injected at the periphery of the caldera. In addition, the fluids associated with the eruptive events at 9-10°N on the EPR (Von Damm, 2000) have a low pH (between 2 and 3) and condensed gases produced during the interaction between molten rocks and seawater in Hawaii also have low pH (1.1 to 2.1) (Resing and Sansone, 1999, 2002). The low pH may also be explained by subsurface deposition of Cu and Fe sulfides. Petrographic and geochemical data for the sulfide deposits clearly supports formation from reduced fluids (pyrrhotite), and the presence of significant silica, low Cu, and abundant Zn suggests moderate to high fluid pH (Evans et al. , 2017). Loss of sulfate and calcium from subsurface deposition of anhydrite can also occur.

As a result, the fluid is hot, enriched in H_2 , and may have a low pH. Thus, fluids from the Kulo Lasi may be considered as the mixing of a deep hydrothermal fluid of the water/rock type with seawater heated on contact with the hot dykes. This process can explain high temperature and high concentrations of SO_4^{2-} and Mg in the mixed fluid. As the contact of cold seawater with the dike at a shallow level is probably of short duration, the composition of the seawater is preserved (Table 3). Considering the mixing of heated seawater (Mg =53.2 mM) with an evolved water/rock type fluid (Mg = 0) we can calculate the maximum percentage of mixing and the end-member compositions for the high temperature collected fluid (Tab 4).

7.5.6 Silica and Fe/Mn geothermometer and geobarometer

As there is no Fe and Mn in the seawater, the Fe/Mn ratio is preserved during subsurface mixing with the heated seawater. Using the Fe/Mn geothermometer (Pester et al. , 2011), temperatures in the reaction zone are estimated to be 415°C and 414°C for the low salinity

fluids and 426°C and 422°C for the high salinity fluids (Table 3). Using these maximum temperatures and the measured temperature in the sampled fluid (considered a mixture of an end-member of water/rock type fluid with various amounts of heated seawater) we have calculated the minimum temperature of the heated seawater. This temperature is close to 250°C both for the high and low salinity fluids (Table 3). This probably indicates that both types of fluids have mixed with the same heated seawater. If we consider some cooling during the ascent of the deep fluid, the end-member temperature used for our calculation may be lower, and the temperature of the heated seawater could be closer to 300°C. The Fe/Mn temperatures can also be used in conjunction with dissolved Si and Cl to constrain possible subseafloor reaction conditions in the pressure-temperature space (Bischoff and Rosenbauer, 1985, Fontaine et al. , 2009, Foustoukos and Seyfried, 2007b, Pester et al. , 2012, Pester et al. , 2011).

The silica isopleths can be used as a pressure proxy when they coincide with the associated temperature calculated using the Fe/Mn ratio (Seewald and Seyfried, 1990). For low salinity fluids, assuming quartz-fluid equilibrium, we can estimate the Si concentration and pressure condition for the formation of end-member fluids using the diagram of chlorinity and silica iso-concentration lines in the P-T space proposed by Fontaine (Fontaine et al. , 2009) (Figure 17) and the Cl-SiO₂ diagram of Foustoukos (Foustoukos and Seyfried, 2007b). Considering the temperature of the reaction zone estimated by the Fe/Mn geothermometer (415°C for the low salinity fluid) and the chlorinity of the end-member (average 0.464 mM for the low salinity fluid) a pressure of 315 bars and silica concentration of about 8.5 mM are estimated for the end-member fluid in the reaction zone (Figure 17). Using the percentage of mixing with seawater calculated with Mg (Table 3), the calculated Si concentration in the end-member fluid implies that the heated seawater is enriched in silica at a concentration estimated to be around 8.1 mM. Indeed, during the lava seawater reactions, the fluids may be enriched in silica (Resing and Sansone, 2002).

Thus, we can propose that the Kulo Lasi fluids have equilibrated at a temperature between 415°C (low salinity) and 425°C (high salinity fluid) and at a pressure of about 315 bars. This indicates equilibrium of Si, Fe, and Mn at about 1650 m below the seafloor. For condensed vapor phases that rise more rapidly than brines, these conditions are considered close to those at the reaction zone near the upper limit of the magma chamber (Fontaine et al. , 2009). These values are consistent with the depth of the new magma chamber under the resurgent dome, estimated to be at 1600 m below the seafloor, using the roof aspect ratio (see discussion above).

In summary, our results suggest a rapidly evolving system, with an early episode of magmatic fluid enriched in SO_2 and characterized by sulfur deposits at the surface of the flows just after the last eruption. This episode is quickly followed by the emission of fluids, resulting from the mixing of deep fluids of water/rock reactions type with heated seawater and without evidence of magmatic contribution. The rapid change in composition suggests a young, immature, and unstable hydrothermal system related to the recent volcanic episode. This instability has been observed in other areas of recent magmatic events on the EPR (Lilley et al. , 2003) and in the back-arc and arcs (Baumberger et al. , 2014, Butterfield et al. , 2011). Considering the rapid decrease in the hydrogen concentration of fluids on the EPR and Juan de Fuca Ridge after an eruptive event, the high concentrations of H_2 in Kulo Lasi fluids probably reflect an eruptive episode that occurred only a few months before the sampling of the fluids. This is coherent with the evolution of the CH_4 concentration in the plume discussed earlier in this paper.

7.6 Atypical composition of hydrothermal mineralization for a back-arc setting

The mineralization at Kulo Lasi appears to be relatively atypical compared to known sulfide mineralization in arc and back-arc basin environments, where the dominant iron sulfide is pyrite, and the dominant copper sulfide is chalcopyrite. This is the case for the Brothers volcano caldera (Berkenbosch et al. , 2012, de Ronde et al. , 2005, de Ronde et al. , 2011), the Sunrise site at the Myojin Knoll (Fiske et al. , 2001, Glasby et al. , 2008), Palinuro (Petersen et al. , 2014), the volcano 19 caldera (Stoffers et al. , 2006) and the East Diamante (Hein et al. , 2014). Pyrite is also the dominant iron sulfide in the Okinawa basin (Halbach and Pracejus, 1993), Lau Basin (Fouquet et al. , 1993, Herzig et al. , 1998) and Manus Basins (Binns and Scott, 1993).

In general, mineralization in back-arc basins sites are typically dominated by pyrite, which is indicative of relatively oxidizing hydrothermal fluids. The occurrence of pyrrhotite/isocubanite as the dominant sulfides at Kulo Lasi is rarely reported in a back-arc environments. Pyrrhotite has previously been found in trace amounts in chimneys on Brothers volcano (Berkenbosch et al. , 2012). Pyrrhotite and isocubanite (cubic cubanite) were recently described in chimneys from the Kilo Moana vent field in the Lau Basin (Evans et al. , 2017, Tivey et al. , 2012). Pyrrhotite and isocubanite are indicative of more reducing conditions than pyrite and chalcopyrite. These conditions are probably explained by the high H_2 concentration in the fluids (Table 3). The presence of significant amounts of galena and barite are typical of arc and back-arc mineralization. While the dominant sulfides are pyrrhotite and isocubanite,

replacement of pyrrhotite by pyrite is observed (Figure 13H). This indicates that the more reduced phases may be replaced by much more oxidized phases. This is compatible with the SO_4^{2-} enrichment in the fluids. As this is occurring in young chimneys, we suggest that there has been a rapid change in the fluid chemistry, from highly reducing fluids to the more oxidizing fluid due to instabilities in mixing between the reduced hydrothermal fluid and heated seawater. This can also be due to rapid change in the H_2 content of the fluid. The instability of the Kulo Lasi vents was also seen in the plume studies, which showed dramatic changes over a one-year period (Konn et al. , 2016).

From a geochemical point of view, the Kulo Lasi mineralisation has relatively low levels of copper (0.8 wt%), perhaps due to subsurface precipitates, and is relatively rich in zinc (≈ 6 wt%). Lead (≈ 0.3 wt%) and barium (1%) are similar to other back-arc basin (BAB) mineralization in the SW Pacific (Table 5). In the Cu-Zn-Pb triangular diagram (Fouquet et al. , 1993) (Figure 18), the Kulo Lasi mineralization plots in the field of back-arc basins mineralization associated with basaltic and felsic host rocks. This contrast with mature back-arc with MORB-type lavas (i.e; North Fiji basin) for which the mineralization, depleted in Pb, plots in the field of mid ocean ridges mineralisation.

High concentrations of certain trace elements appear to be directly related to the presence of particular mineral phases. For example, the positive correlation between Pb and Zn is consistent with petrographic observations of inclusions of galena in sphalerite. The abundance of sphalerite also controls the enrichment in Cd, Sn, and Ag. Similarly, the positive correlations between cobalt, selenium, and iron indicate an affinity between these elements and pyrrhotite. Variations in the gallium and germanium contents are more complicated to interpret though high levels appear to correspond to sphalerite-rich samples.

Regarding the minor elements, Ga (> 100 ppm) is unusually high. Compared to back-arc mineralization, the Kulo Lasi sulfides are enriched in Co (≈ 0.15 wt%), Sn (168 ppm) and especially in Se (> 350 ppm) (Table 5). Cobalt and selenium enrichments are commonly associated with MORs mineralization. The highest known average concentrations in Co (5086 ppm), in Sn (347 ppm) and in Se (772 ppm) are in mineralization found on slow spreading ridges in ultramafic rocks environments (Evrard et al. , 2015, Fouquet et al. , 2010). Like the Kulo Lasi mineralization, these ultramafic mineralization are characterized by an abundance of isocubanite and pyrrhotite and reducing conditions that may favor the concentration of Co and Se. As expected for a pyrrhotite-dominant mineralization, gold is not enriched (Hannington and Scott, 1989).

Thus, the Kulo Lasi mineralization has chemical affinities with both back-arc and ocean-ridge mineralization. All of these geochemical features indicate a complex signature that reflects characteristics of both immature felsic back-arc basins (i.e., Lau basin) and mature basaltic back-arc basins (i.e., North Fiji basin). In the North Fiji Basin, for example, Se and Co are also enriched (Table 5). This is also coherent with the volcanic rock composition, which indicates that the magmatic activity in the Kulo Lasi region is not directly related to subduction. Thus, the Kulo Lasi mineralization is atypical and differs from mineralization associated with arc and back-arcs located closer to the subduction zone.

7.7 Timing and control of the different types of mineralization in large basaltic submarine calderas based on Kulo Lasi volcano

The large amount and diversity of hydrothermal mineralization at Kulo Lasi allows for discussion of the evolution of different types of mineralization in relation to caldera formation, history, and geometry. Based on high-resolution mapping, field observations, vent fluid compositions, and types of hydrothermal mineralization, we propose a four-stage model for evolution of the caldera and formation of mineralization in deep-sea basaltic/andesitic calderas. This model is also based on recent advances in the understanding of the formation of aerial and submarine calderas.

7.7.1 Stage 1: Early stage of caldera formation

During **Stage1**, the volcano is built by a series of lava flows that pour from the summit (Figure 19). The basement of the volcano is made up of recent tectonized lava. Early Fe/Mn mineralization associated with low-temperature diffuse venting precipitate at the top of the volcano. At the end of this construction stage, the magma chamber empties and its summit starts to collapse. This collapse is not a catastrophic event, as observed in large silicic volcanoes (Geshi et al. , 2002), but is instead a slow process that take place over a long period. Cracks and ring fissures, precursors to the formation of the caldera, appear in the summit area. When the deep reverse faults are initialized, the resulting increased permeability allows access to hot hydrothermal fluids from deep within the system. The deep hot fluids can rise to the seafloor along reverse faults and associated fissures that formed prior to the major collapse of the caldera. Evidence for this episode is provided by external normal faults that have exposed outcrops of highly altered zones impregnated with sulfides and associated with stockwork mineralization. The presence of Cu sulfides suggests flows of high-temperature fluids towards the summit of the caldera along the initial faults, while the

absence of alunite suggests a lack of significant acid-sulfate-like magmatic fluids. The positive value of the sulfur isotopes also indicates a lack of magmatic fluids, which would typically lead to negative values (de Ronde et al. , 2005, de Ronde et al. , 2011, Herzig et al. , 1998). In addition, the high values of δS_{34} (6 to 9 per mil) in disseminated pyrites in the alteration zones indicate a seawater contribution to the subsurface hydrothermal fluid. High-temperature sulfide chimneys should be able to form at the summit of the volcano during this stage, although this has not yet been observed as the dives so far conducted have all been focused on the inside of the caldera. A possibility is that these chimneys were destroyed during the collapse of the caldera. In summary, this stage is characterized by a classic water/rock-dominated convective hydrothermal system, mixed with cold seawater, with no evidence for significant contributions from magmatic fluids.

7.7.2 Stage 2: Caldera formation

Stage 2 corresponds to the collapse of the roof of the magma chamber (Figure 20). The formation of the caldera starts with collapse along the internal reverse faults, and then along the normal external faults, forming the morphology now visible on the seafloor. Lava flows are tilted toward the center of the caldera, and talus and scree piles form at the base of the walls. The collapse movements along the outer normal faults result in exhumation of the altered zones and stockwork formed during Stage 1. Freshly broken rocks and new external faults are highly permeable and facilitate the downward penetration of cold seawater. This, combined with the deepening, emptying, and then collapse and cooling of the magma chamber leads to the hydrothermal circulation of low-temperature fluids particularly at the end of the caldera formation. At Kulo Lasi, some areas of silica-rich metalliferous sediments covered by pelagic sediments, located on the caldera floor near the base of the walls, formed during this low-temperature episode.

7.7.3 Stage 3: New volcanic episode

This stage begins by refilling of the magma chamber causing its top to swell (Figure 21). The visible consequence is a rising of the caldera floor and the formation of a central resurgent dome from uplift and deformation of the old lava. Several successive lava flows are issued from the base of the caldera walls at the level of the reverse faults emplacement. Scree on the lower slopes of the caldera is consumed by the recent lava flows. Sedimentary formations of low temperature hydrothermal silica and iron oxides are locally precipitated on the early flows at the outer margins of the caldera floor as part of this new volcanic episode. It appears that the final hydrothermal episode of Stage 2 and the earliest hydrothermal episode of Stage 3 are characterized by low temperature hydrothermal emissions on the caldera floor. At Kulo Lasi, none of these sites is active as indicated by the cover of pelagic

sediments. These hydrothermal mineralization are thought to have been formed by shallow hydrothermal circulation in the highly permeable zones along the outer normal faults. The last lava flows, controlled by the deep reverse fault, probably erupted less than one year ago. At the surface of these recent flows and close to the reverse faults, numerous small deposits of native sulfur are formed by the discharge of magmatic-dominated fluids just after the last eruptive event. This episode was probably very short, less than a few months after the eruptive event.

7.7.4 Stage 4: High temperature black smoker activity

Hydrothermal fluids, resulting from water-rock reaction in a deep reaction zone, discharges on the seafloor. Seawater heated ($>250^{\circ}\text{C}$) by the still hot dikes mixes with the deep hydrothermal fluid near the surface (Figure 22). Reactions of fluids with the hot dike also result in hydrogen enrichment. On the western side of Kulo Lasi caldera, the alignment of active vents on narrow lava flows is interpreted as a control of both, lava diking and fluid discharge, by the reverse faults. Consequently, black smokers and sulfide deposits occur preferentially on the caldera floor near the base of the walls. This geometry is demonstrated in analog models, and high-temperature discharge has also been documented for other submarine caldera volcanoes near the intersection of the caldera walls and floor (e.g., (Baker et al. , 1999, de Ronde et al. , 2011)). On the eastern side of the caldera, weakly active sulfide chimneys are located near large cracks that indicate tectonic extension, facilitating the discharge, near the base of the caldera walls. No chimneys are observed on new flows in the center of the caldera. Low temperature silicic chimneys are active at the base of the caldera walls and diffuse very low temperature venting occurs on the caldera walls. Both types probably result from mixing of hydrothermal fluids with various amounts of cold seawater. Positive sulfur isotopes values and dominant pyrrhotite and isocubanite indicate low sulfidation fluids of water-rock type without magmatic fluid contribution during this stage. Late replacement of pyrrhotite by pyrite and rapid variation in the plume chemistry indicate an unstable hydrothermal system.

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Figure captions

Figure 1: Regional location of the Wallis and Futuna volcanic areas. The white box shows the position of figure 2. E-W black line = North Fiji transform fault. Location of spreading centers and faults are from Pelletier 1996 and 2001 (Pelletier and Auzende, 1996 , Pelletier et al. , 2001).

Figure 2: Bathymetric (left) and seafloor backscatter amplitude (right) maps of the Kulo Lasi volcanic area with 25 m and 10 m resolution, respectively. Light gray = old areas covered with sediments. Dark gray = recent volcanic zones. Kulo Lasi volcano is the most recent volcanic structure in the area. See white box in Figure 1 for regional setting.

Figure 3: 3D bathymetric view of Kulo Lasi volcano with a 10 m resolution derived from the shipborne multibeam surveys at low speed (looking north). To the north of the volcano is a highly tectonized area probably similar to the older area underlying the volcano. The Kulo Lasi volcano itself is relatively non-tectonized compared to the surrounding volcanic basement, except an old tectonized volcanic construction on its NW flank. Note the three flat-topped satellite cones near the summit of the volcano. Water depth: red color = 1200 m to 1500 m; green color = 1500 m to 1800 m; blue-color 1800 m to 2300 m.

Figure 4: High-resolution (1 m) AUV-derived 3D bathymetric map of the Kulo Lasi caldera with no vertical exaggeration. Note the collapsed and tilted old lava flows on the left side. The resurgent dome in the center is composed of old tectonized pillow lava flows. The relatively flat caldera floor is made up of several generations of 1-2 m thick recent flows. The most recent lava flows are clearly visible. Note the abundance of scree along the eastern caldera wall. View looking north from the top of the southern wall of the caldera. Dashed white line shows the position of figure 5.

Figure 5: High-resolution (1 m pixel grid) bathymetric AUV-derived map showing recent lava flows in the southeast of the caldera. See position on figure 4. White numbers refer to five successive eruptive episodes: Episode 1- Pillow lava with light sediment coverage (> 1500 m of water depth). Episode 2- (Purple contours and arrows), scoriaceous lava flows, and sheet flows. Episode 3- (Blue contours and arrows) sheet flows issued from the southern caldera wall. Episode 4- (Yellow contours and arrows), fresh pillow lava without sediment coverage (1490 to 1470m of water depth). Episode 5- (Green dashed lines and white arrows).

Discontinuous and narrow lava flows parallel and close to (~100 m) the base of the caldera walls. These probably represent the trace of dikes rooted in the reverse faults. For episodes, 2, 3 and 4 the color arrows give the direction of the flow.

Figure 6: TAS (Total Alkali-versus-Silica) diagram (Le Bas et al. , 1986). The red circles represent samples of the older lava taken from the caldera walls, and the blue circles represent samples taken from the most recent lava flows. The dashed line approximately separates alkaline volcanic rocks from their subalkaline counterparts.

Figure 7: Selected major elements (in wt%) plotted against total silica content for the volcanic rocks of the Kulo Lasi. Same symbols as in Figure 6. The red circles represent samples of the older lava taken from the caldera walls, and the blue circles represent samples taken from the most recent lava flows.

Figure 8: Chondrite-normalized rare earth element patterns of Group 1 (Old lava - red lines) and Group 2 (Young lava - blue lines). Chondrite composition from Anders and Grevesse (Anders and Grevesse, 1989). These patterns are very similar and likely derive from the same source.

Figure 9: Locations of the different types of mineralization in the Kulo Lasi caldera superimposed in the high-resolution bathymetric map derived from AUV surveys (1 m pixel grid). The different diving tracks are also indicated.

Figure 10: Different types of hydrothermal mineralization observed in the Kulo Lasi caldera (see figure 9). A: Fe-Mn mineralization at the top of the north caldera wall. B: Fe-Si hydrothermal sediments on the SW caldera floor. C: Active silica chimneys (43°C) near the base of the caldera wall. D: Tubeworms and crabs in an area of diffuse venting on the southern caldera wall. E: Inactive sulfide chimney near a large crack in a prismatic lava flow at the base of the SE caldera wall. F and H: Active black smokers (up to 343°C) and iron oxide staining on the surface of a recent lava flow. G: Native sulfur (white) at the surface of a recent lava flow. H: Surface of a recent lava flow. Photos taken from Nautilie, each view cover a surface of about 5 to 20 m².

Figure 11: Detailed AUV-derived bathymetry (left) and seafloor backscatter amplitude maps (right) of the southern wall of the Kulo Lasi caldera, with a resolution of 2 m and 0.5 m, respectively. Red and yellow lines = dive tracks. Altered zones near the top of the caldera

wall are outlined in black. The upper section of the caldera is composed of prismatic lava flows. (See figure 9 for location).

Figure 12: High resolution AUV-derived seafloor backscatter map (0.5 m pixel grid). Low-temperature Si-Fe hydrothermal sediments (light gray patches indicated by arrows) on the caldera floor. Several of the patches were visited during the dives. See figure 9 for location.

Figure 13: Location of mineralization and high temperature hydrothermal activity in the NW sector of the caldera. Left: Bathymetric map and main hydrothermal fields. Right: Geological map showing the different types of lava flows and areas of intensive hydrothermal alteration near the base of the caldera wall. Contours based on dive observations, high-resolution bathymetry and interpretation of seafloor backscatter Images acquired with the AUV Aster^X. Positions and samples numbers of fluids and chimneys collected during the dives is shown. See figure 9 for location.

Figure 14: Micro-photographic plate - reflected light. A: massive pyrrhotite – the red box corresponds to photo B. B: Overgrowth of isocubanite and sphalerite mantling pyrrhotite. C: Inclusions of pyrite and relics of a former aggregate of sphalerite in pyrrhotite. D: Aggregates of sphalerite with galena inclusions. E: Aggregate of sphalerite with two successive textures: a collomform texture at the center with inclusions of galena (G) and a crystallized texture edge associated with inclusions of isocubanite and late galena - the red dotted line marks the boundary between the two textural forms. F and G: "Inclusions" of isocubanite and chalcopryrite in sphalerite. In Photograph G the chalcopryrite is aligned along the crystallization boundary. H. Pyrrhotite, sphalerite, and isocubanite with chalcopryrite exsolutions.

Figure 15: Micro-photographic plate -reflected light. A, B, and C: Association of sphalerite, isocubanite, pyrrhotite, and pyrite. B and C: Late opal mantles other minerals and may gradually replace sphalerite. D: Replacement of pyrrhotite by a pyrite-sphalerite assemblage. E: Chimney dominated by pyrite and marcasite. The center is enriched in chalcopryrite. F: Ghosts of early pyrrhotite crystals in a pyritic chimney. G. Pyrrhotite that has been strongly altered and partially replaced by pyrite. H: Active chimney. Pyrrhotite blades partially replaced by pyrite.

Figure 16: A-B: Macroscopic samples of altered basalts with disseminated sulfides. Inset: images = polished sections enlarged in C, D, E and F. C-F: Micro-photographs taken under reflected light (showing (C) pyrites dispersed in a strongly altered basaltic matrix, and (D)

isolated chalcopyrite grains and (E) inclusion of chalcopyrite in pyrite. The presence of chalcopyrite suggests hydrothermal alteration at high-temperature ($> 300^{\circ}\text{C}$). F: Pyrrhotite inclusions in pyrite.

Figure 17: Chlorinity (in molal, blue lines) and silica (in mmolal, red lines) iso-concentration lines in P-T space (redrawn after Fontaine et al. 2009). The intersection of iso-Cl and iso-Si defines a single P-T equilibration point. The black circle indicates the estimated equilibration point for vapor-like fluids of the Kulo Lasi field. The position of the Kulo Lasi fluids is determined by the means of temperatures calculated using the Fe/Mn geothermometer (415°C) (Pester et al. , 2011) and the chlorinity of the end-member low salinity fluid (average 0.464 mM). This position suggests that, for mineral-fluid equilibrium, the fluid had to reach a pressure of 415 bar. Under these conditions, the silica concentration would be 8.5 mM. These conditions are very close to the two-phase boundary and to the critical point of seawater and suggest a reaction zone at a depth of about 1650m below the seafloor.

Figure 18: Ternary diagrams showing the composition of the sulfide chimneys from Kulo Lasi., The Pb-Zn-Cu diagram (left) highlights an affinity of Kulo Lasi with mineralization in back-arc basins,(Fouquet et al. , 1993). The Co-Cu-Se diagram (right) highlights an affinity with deposits located in the Mid-Ocean ridges (Fouquet et al. , 2010). (See table 5 for detailed compositions and references).

Figure 19: Stage 1 - End of the main volcanic episode and initiation of collapse. A: recent lava flows forming the shield volcano. B: fractured volcanic basement. M: magma chamber. Initiation of faulting at the top of the volcano: Red dashed line = reverse faults; Dark blue dashed lines = normal faults. Colored numbers: 1- red arrows: high-temperature hydrothermal convective cells. 2- blue arrows: mixing with cold seawater. Black numbers: 1- high-temperature sulfide chimneys (red triangles) rooted on altered lava and stockwork mineralization (red crosshatching). 2- low-temperature Fe/Mn hydrothermal mineralization (brown patches).

Figure 20: Stage 2 – End of the collapse of the magma chamber and caldera formation. A: lava flows. B: fractured volcanic basement. M: magma chamber. Caldera faults: red line = reverse faults; dark blue lines: normal faults. Colored numbers: 1- blue arrows: Low-temperature hydrothermal convective cells. Black numbers = hydrothermal precipitates: 1- outcrop of hydrothermal alteration and stockwork mineralization (red crosshatching). 2- low-temperature Fe/Mn hydrothermal mineralization (brown patches). 3- low-temperature Si/Fe

mineralization at the base of the caldera walls (orange patches. Talus and scree are not represented.

Figure 21: Stage 3: New magmatic episode and magmatic fluid venting. A: old lava flows. B: fractured volcanic basement. M: magma chamber. R: resurgent dome. C: recent lava flows and direction of flows (black arrows). Caldera faults: red dotted line = reverse faults; Dark blue lines = normal faults. Colored numbers: 1- yellow arrows: magmatic fluid. 2- red arrows: high-temperature hydrothermal convective cells and fluids generated by deep water-rock reaction. 3-blue arrows, cold seawater. Black numbers = hydrothermal precipitates: 1- outcrop of hydrothermal alteration and stockwork mineralization (red crosshatching). 2- low-temperature Fe/Mn hydrothermal mineralization (brown patches). 3- low-temperature Si/Fe mineralization at the base of the caldera walls (orange patches). 4- native sulfur deposits generated by the discharge of magmatic fluids (yellows dots). Talus and scree are not represented.

Figure 22: Stage 4: New hydrothermal episode and high-temperature venting. A: old lava flows. B: fractured volcanic basement. M: magma chamber. R: resurgent dome. C: recent lava flows. Caldera faults: red dotted line = reverse faults; dark blue lines = normal faults. Colored numbers: 1- red arrows: high-temperature hydrothermal fluid generated by deep water-rock reaction. 2- white arrows: seawater heated ($>250^{\circ}\text{C}$) at shallow levels in contact with the hot recent dyke controlled by the reverse faults (red line). This heated seawater is mixed with the hydrothermal fluid. 3- Blue arrow, cold seawater. Black numbers = hydrothermal precipitates: 1- stockwork mineralization (red crosshatching). 2- low-temperature Fe/Mn deposits (brown patches). 3- Low-temperature Si/Fe mineralization (orange patches). 4- native sulfur deposits generated by the discharge of magmatic fluids (yellow dots). 5- recent (< 1 year) black smokers sulfidic chimneys (red triangles). 6- low-temperature silica chimney near the base of the caldera wall (yellow triangles). 7- Diffuse low-temperature venting without precipitates. Talus and scree are not represented.

Table captions

Table 1: Known deep sea calderas associated with hydrothermal sulfides deposits: 1:(Honsho et al. , 2015, Iizasa et al. , 2004); 2:(Gamo et al. , 1996, Seewald et al. , 2015); 3: (Berkenbosch et al. , 2012, Kim et al. , 2009, Lupton et al. , 2015); 4: (Falloo et al. , 1992, Lisitsyn et al. , 1992); 5: (Ligi et al. , 2014); 6: This paper; 7,8,9,10: (Lupton et al. , 2012); 11:

(Anderson et al. , 2015); 12: (Iizasa et al. , 1999, Tanahashi et al. , 2008); 13: (Iizasa et al. , 1992); 14: (Tsunogai et al. , 1994); 15: (Urabe et al. , 1987); 16: (Gamo et al. , 2004); 17: (Hein et al. , 2014); 18: (Stern et al. , 2008); 19: (Brounce et al. , 2016); 20: (Lupton et al. , 2008); 21: (Stuben et al. , 1992) 22-23: (Stoffers et al. , 2006); 24: (de Ronde et al. , 2005, de Ronde et al. , 2011); 25: (de Ronde et al. , 2001, Wright et al. , 2003); 26: (Leybourne et al. , 2012b); 27: (Boschen et al. , 2013); 28: (CASM, 1985, Hannington and Scott, 1988); 29: (Haymon et al. , 2008).

Table 2: Chemical compositions of representative lava samples collected from the Kulo Lasi volcano. Old lava are in red and young lava in blue in figs 6, 7 and 8.

Table 3: fluid compositions (Konn et al. , 2018). Samples TiG3 and TiD3 correspond to the chimney sample FU-PL-06-02 and samples TiG1 and TiD1 correspond to the chimney sample FU-PL-06-03. See location of samples on figure 13. End-members are calculated in this paper, considering mixing between deep hydrothermal fluid generated by water-rock reaction and shallow seawater heated by the recent dike. The temperature estimated using the Fe/Mn geothermometer (Pester et al. , 2011) is indicated. See text for details.

Table 4: Results of XRF and XRD analyses of sulfides and sulfates chimneys. See Fig. 13 for correspondence between chimneys and fluid samples. Zn-rich and Fe-rich samples are from inactive chimneys. For the location: NW = North West part of the caldera. E = Eastern part of the caldera. The lower part of the table shows the mineral proportions determined from X-ray diffraction data and calculated using the TOPAS software. Estimating the proportion of amorphous phases (opal, iron oxyhydroxydes) is not possible; values proposed are therefore semi quantitative.

Table 5: Average composition of sulfide mineralization in typical oceanic environments. Source of data: (1) = (Fouquet et al. , 1996); (2, 3, 4, 14) = Fouquet, unpublished data; (5 to 12) = (Fouquet et al. , 2010); (13, 16, 17, 19, 20, 23, 24, 25, 26, 27, 28, 29) = (Hannington et al. , 2005.) and references therein; (15) = (Bendel et al. , 1993); (18) = this study. (21, 22) = (Fouquet et al. , 1993). MORB = Mid-Ocean Ridge Basalt, E-MORB = Enriched MORB; BABB = back-arc basin basalt; IAB = Island-arc tholeiite.

Table 6: Chemical composition of Si-Fe hydrothermal sediments

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Figures

Figure 1 – in colors

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Figure 2 – in colors

Figure 3 – in colors

Figure 4 –in black and white

Figure 5 – in colors

Figure 6 – in colors

Figure 7 – in colors

Figure 8 – in colors

Figure 9 – in colors

Figure 10 – in colors

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Figure 11 – in colors

Figure 12 – in black and white

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Figure 13 – in colors

Figure 14 – in colors

Figure 15 – in colors

Figure 16 – in black and white

Figure 17 – in black and white

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Figure 18 – in black and white

Figure 19 – in colors

Figure 20 – in colors

Figure 21 – in colors

Figure 22 – in colors

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Tables

Table 1

Name	Geodynamic Setting	Volcanism	Diameter Caldera (km)	Mineralisation	Temperature °C	Depth m
Back-arc basins						
1 Bayonnaise	Izu Bonin	Andesite-dacite	0.5x0.7	Sulfides	-	800-700
2 Desmos	Manus	Basalt-andesite	1.5x2	Sulfides/sulfur	120	1810
3 Niuatahi (Caldera O)	Lau	Dacite	15	Sulfides/sulfur	-	1280
4 CLSC A3	Lau	Basalt-andesite	?	Sulfides	-	2500
5 Palinuro	Tyrrhenian	Basalt-andesite	11x7	sulfides	-	1000-650
6 Kulo Lasi	Futuna	Basalt-trachyandesite	5.6x4.5	Sulfide/silica/sulfur	345	1200-1500
7 Southern caldera	NWLSC	Basalt	1.5x6	plume studies	-	2160
8 Central caldera	NWLSC	Basalt	2.8x7	plume studies	-	2150
9 Lobster caldera	Rochambeau Rift	Basalt	2x2.8	plume studies	-	1500
10 Dugong	North Lau Basin	Basalt	5	plume studies	-	1170
11 Nifonea	New Hebrides	Basalt	5x8	Sulfides	HT°	1875-2080
Arc volcanoes						
12 Myojin-Sunrise	Izu Bonin	Andesite-dacite-rhyolites	6x7	Sulfides	278	1360-1220
13 Myojinsho	Izu Bonin	Andesite-dacite-rhyolites	5x6	Sulfides	-	1110-850
14 Suyio	Izu Bonin	Andesite-dacite	1.7	Sulfides	317	1380-1320
15 Kaikata	Izu Bonin	Andesite	3	Sulfides/manganese	22	930-460
16 Toto	Mariana	?	13	?	170	2960-2300
17 East Diamante	Mariana	Dacite	5x10	Barite/sulfides	260	367-406
18 W Rota	Mariana	Andesite-rhyolite-basalt	6x10	feeder zone	-	1500
19 Fina Nagu Cone 5	Mariana	Basalt-basaltic andesite	?	Chimneys	?	2000
20 Nikko	Mariana	Basalt-dacite	0.36	Sulfur	215	410-490
21 Esmeralda	Mariana	Basalt-basaltic andesite	4	Sulfides/sulfur	78	300
22 Volcano 19	Tonga	Basalt-basaltic andesite	1.9	Sulfides /barite	265	970-385
23 Volcano 1	Tonga	Basaltic andesite-andesite	7x4.5	Sulfure/Fe-oxides	152	65
24 Brothers	Kermadec	Basalt-andesite	3x3.5	Sulfides	302	1800-1150
25 Healy	Kermadec	Basalt-felsic	2x2.5	plume studies	-	1370-1140
26 Monowai	Kermadec	Basalt-basaltic andesite	8	Sulfides/sulfates	-	1040
27 Kemp	Sandwich	?	8x7	White smokers	40	800-1600
Mid-ocean ridges						
28 Axial seamount	Juan de Fuca	E-MORB	3x8	sulfides	348	1540
29 Los Huellos	Galapagos	MORB	4x2	sulfides	-	1050-1530

Table 1

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Table2

Sample	FU-DR-33-01gl	FU-DR-33-03	FU-DR-38-01	FU-DR-36-02	FU-PL-03-06
	Old lava from the caldera			Young lava from surface and caldera floor	
	Basaltic trachy- andesite	Trachy-andesite	Basalt	Basalt	Basalt
%					
SiO ₂	53.6	60.1	48.6	49.2	51.3
Al ₂ O ₃	13.9	13.4	15.4	14.4	13.8
Fe ₂ O ₃	12.0	8.9	10.8	11.6	15.0
MgO	3.7	2.1	8.4	7.6	3.9
CaO	7.4	4.7	12.6	12.3	8.1
Na ₂ O	4.7	5.4	2.6	2.7	4.1
K ₂ O	0.7	1.3	0.2	0.2	0.7
TiO ₂	1.9	1.2	1.1	1.2	2.4
P ₂ O ₅	0.4	0.3	0.1	0.1	0.4
MnO	0.2	0.2	0.2	0.2	0.2
SO ₄	0.2	0.1	0.0	0.1	0.2
LOI	0.2	1.1	-0.3	-0.3	-0.1
total	99.0	99.0	99.8	99.4	100.0
ppm					
V	258	61	255	285	340
Co	40	17	44	45	42
Ni	177	10	79	62	19
Cu	92	38	170	171	89
Zn	87	112	78	75	130
Li	11	13	5	5	7
Rb	16	27	2	3	11
Sr	129	95	118	115	143
Y	77	98	27	29	62
Zr	293	474	58	69	197
Nb	18	26	4	4	12
Cs	0.2	0.3	0.0	0.0	0.2
Ba	138	216	39	33	113
La	16.6	22.0	3.4	3.7	11.9
Ce	41.6	56.8	8.8	9.9	30.2
Pr	6.1	7.9	1.4	1.6	4.5
Nd	29.3	36.8	7.5	8.3	22.3
Sm	8.6	10.5	2.6	2.9	6.9
Eu	2.5	2.6	1.0	1.1	2.3
Gd	10.7	12.7	3.7	4.0	8.8
Tb	1.9	2.2	0.7	0.7	1.5
Dy	12.4	15.1	4.5	4.8	10.2
Ho	2.6	3.3	1.0	1.0	2.2
Er	7.9	9.9	2.9	3.0	6.4
Yb	7.6	9.8	2.7	2.8	6.0
Lu	1.1	1.5	0.4	0.4	0.9
Hf	7.3	11.3	1.8	2.0	5.1
Ta	1.1	1.5	0.2	0.2	0.8
Pb	1.4	2.0	1.1	0.6	1.3
Th	2.2	3.4	0.4	0.4	1.4
U	0.7	1.2	0.1	0.1	0.5

Table 2

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Table 3

Zone	unit	Sea Water	N1	N1	N2	N2	N3	N3
Sample name		FU-PL-05-TiG2	FU-PL-06-TiG4	FU-PL-06-TiD4	FU-PL-06-TiG3	FU-PL-06-TiD3	FU-PL-06-TiD1	FU-PL-06-TiG1
Type			Beehive	Beehive	Chimney	Chimney	Chimney	Chimney
Color of the fluid			Black	Black	Clear	Clear	Black	Black
Depth	m	1150	1475	1475	1475	1475	1475	1475
Mesured concentrations								
T chimney	°C		332	332	331	331	345	345
pH		7.8	6.07	5.58	2.24	2.37	2.36	3.32
CH ₄	mM	0	-	0.001	0.006		0.004	0.001
H ₂	mM	0	-	1.14	19.83	-	6.18	1.04
H ₂ S	mM	0	0.46	1.66	5.05	3.81	0.39	-
CO ₂	mM	2.3	2.6	1.3	5.1	-	3.0	1.0
Cl	mM	551	516	485	497	517	735	689
NaCl	wt%	3.2	3.0	2.8	2.9	3.0	4.3	4.0
Mg	mM	53.2	49.8	45.7	24.6	28.6	26.5	34.9
Si	mM	0.1	1.0	2.1	8.2	8.4	14.6	10.8
Fe	μM	0	123	78	4796	4283	9884	6845
Mn	μM	0	32	76	862	788	1416	1064
Cu	μM	0	1.7	1.3	14.2	4.2	2.5	20.1
Zn	μM	0	3.1	1.5	44.8	41.5	175.1	77.2
SO ₄ ²⁻	mM	29	27	23.9	8.8	10.7	6.2	12
Na	mM	457	448	406	388	405	612	565
K	mM	9.8	10.6	9.5	18.5	16.6	29.5	23.7
Ca	mM	10.6	10.5	10.2	11.6	10.8	10.9	10.8
Li	μM	28	54	55	156	149	249	197
Rb	μM	4.4	5.3	6.1	25.8	24.2	46.4	35.6
Sr	μM	93	84.2	74	7.3	9.4	3.4	9.1
Br	μM	833	822	994	738	770	1135	1051
Fe/Mn		-	3.8	1.0	5.6	5.4	7.0	6.4
Br/Cl			1.59	2.05	1.48	1.49	1.54	1.53
Calculated end-members								
T° maximum measured	°C		134	136	342	338	343	343
T° estimated using Fe/Mn	°C		397	333	415	414	426	422
T° calculated for heated	°C		-	-	257	250	247	251
Cl	mM		-	-	451	477	918	952
Si	mM		-	-	15.2	18.0	29.0	31.2
Fe	μM		-	-	8921	9262	19694	19899
Mn	μM		-	-	1603	1704	2821	3093
Cu	μM		-	-	26	9	5	58
Zn	μM		-	-	83	90	349	224
Na	mM		-	-	329	345	766	771
K	mM		-	-	26	25	49	50
Ca	mM		-	-	12	11	11	11
Li	μM		-	-	266	290	468	519
Rb	μM		-	-	44	47	88	95
Sr	μM		-	-	-66	-88	-86	-151
Br	μM		-	-	656	697	1435	1467

Table 3

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Table 4																
Type		Sulfate-rich					Zn-rich sulfides						Fe-rich sulfides			
Location		NW	NW	NW	NW	NW	NW	E	E	E	NW	E	E	NW	NW	E
Sample n°		FU-PL01-03	FU-PL01-04	FU-PL06-03	FU-PL04-02	FU-PL06-02	FU-PL01-06-ext	FU-PL03-05a	FU-PL03-02b	FU-PL03-02c	FU-PL01-06-int	FU-PL03-04	FU-PL03-05b	FU-PL01-07a-haut	FU-PL01-07b-bas	FU-PL03-03
%																
S		26.18	25.57	26.20	25.44	26.38	32.56	32.01	33.12	33.85	37.09	39.92	40.17	39.35	33.32	46.49
Ca		28.69	27.89	27.64	27.62	19.73	0.38	0.34	<0.3	<0.3	<0.3	0.31	<0.3	<0.3	<0.3	1.47
Fe		3.64	1.48	1.25	1.31	13.76	9.90	11.10	35.50	36.95	46.94	39.50	58.76	55.07	44.72	42.32
Cu		0.11	<0.1	<0.1	<0.1	0.27	1.04	0.29	0.77	0.83	2.02	1.53	1.70	1.88	1.01	1.27
Zn		<0.2	<0.2	<0.2	<0.2	0.21	47.70	43.47	13.20	11.84	8.67	7.56	2.78	2.63	1.50	1.11
Pb							1.46	1.53	0.47	0.33	0.35	0.29				
Ba		<0.2	<0.2	<0.2	<0.2	<0.2	5.04	2.91	0.81	<0.2	1.75	0.28	<0.2	<0.2	3.88	0.73
Al2O3		0.70	1.46	1.56	1.11	1.79	0.57	0.67	0.66	0.67	0.63	0.70	0.59	0.59	0.79	0.62
SiO2		0.32	<0.3	<0.3	<0.3	5.76	2.15	6.53	8.91	16.32	3.15	10.94	<0.3	2.22	12.72	2.23
MgO		<0.3	<0.3	0.30	<0.3	0.99	<0.3	0.31	<0.3	<0.3	<0.3	<0.3	<0.3	<0.3	<0.3	1.04
P2O5		0.03	0.04	0.05	0.04	0.06	0.04	0.06	0.04	0.03	0.03	0.07	0.03	0.03	0.03	0.07
K2O		<0.01	0.01	0.02	<0.01	0.05	0.04	0.09	0.04	0.02	0.02	0.03	0.02	0.01	0.03	<0.01
TiO2		<0.01	<0.01	<0.01	<0.01	0.18	0.03	0.02	0.04	0.01	0.01	<0.01	0.01	0.01	0.02	<0.01
MnO		0.11	<0.01	<0.01	<0.01	0.03	0.10	0.35	0.05	0.02	0.09	0.04	0.03	0.02	0.02	0.04
ppm																
Mo			17	<10	17	23	64	62	37	32	36	63	22	56	154	65
Sr		1800	2851	1717	1688	1943	1121	560	188	28	409	73	11	9	1130	502
Pb		35	12	15	27	51	14892	17042	3949	2495	2902	2121	392	501	767	640
As		10	6	<5	<5	240	780	1724	532	551	920	1161	558	1525	126	267
Se			10	<5	<5	256	33	<5	207	132	501	489	812	483	867	370
Ga			<5	<5	<5	7	45	734	157	145	15	107	49	20	<5	<5
Ge			<10	<10	<10	<10	160	324	59	38	14	26	<10	<10	<10	<10
U			<5	<5	<5	<5	<5	<5	<5	6	<5	13	6	6	<5	9
Co		<10	<10	80	52	938	754	1311	1075	749	2254	2471	2649	2664	1703	1077
Ni		13	<10	<10	<10	32	157	128	26	22	25	82	<10	22	35	77
Ag			<20	<20	<20	<20	298	232	68	62	43	26	<20	<20	<20	<20
Cd			<15	<15	<15	<15	2482	1829	635	576	455	367	151	178	84	60
In			<10	<10	<10	<10	47	30	12	11	26	13	18	10	48	18
Sn			<20	<20	<20	<20	694	652	311	348	156	240	106	68	42	<20
Sb			<20	<20	<20	<20	77	215	93	120	37	86	22	34	<20	<20
Au			0.3	0.4	0.3	0.6	<2	<2	<2	<2	<2	<2	<2	<2	0.9	0.7
W			<0.3	<0.3	<0.3	<0.3							>0.8		>0.7	>0.4
Pt			<0.2	<0.2	<0.2	<0.2	<0.2	<0.2	<0.2	<0.2	<0.2	<0.2	<0.2	<0.2	<0.2	<0.2
Bi			<1	<1	1.9	2.6	<1	<1	<1	<1	<1	3.3	<1	4.9	10.6	<1
Mineral composition																
Pyrrhotite						18	8		47	40	51	14	84	70	61	7
Sphalerite/wurtzite							69	67	28	27	16	20	7	7	3	2
Pyrite			1	<1	<1	3	1	6	6	9	2	43		1	4	50
Chalcopyrite							6	8	7	8	15	10	8	14	7	13
Isocubanite							<1	0	<1	<1	3	<1	<1	4	1	<1
Marcasite																21
Barite							14	8	4		6	2	0	2	13	4
Anhydrite			99	99	99	74	<1									
Halite								6								1
Nontronite							<1				4					
Opal			<0.3	<0.3	<0.3	6	2	7	9	16	3	11	<0.3	2	13	2

Table 4

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Table 5

Name	Host Rocks	Regional setting	n	depth (m)	Cu	Zn	Pb	Fe	S	SiO2	Ba	Ag	Au	Cd	As	Sb	Ni	Co	Se	Mo	In	Sn	references	
								(wt %)																
Fast-spreading mid-ocean ridges																								
13°N	MORB	East Pacific Rise	31	2650	4.56	15.26	0.09	25.48	36.54	2.21	0.00	88	0.56	436	210	3	2	408	96	48	<5	<10	1	
17°30S	MORB	East Pacific Rise	28	2573	2.65	11.99	0.05	32.76	41.57	4.14	0.00	44	0.10	302	211	3	1	218	263	63	4	3	2	
18°30S	MORB	East Pacific Rise	18	2640	10.19	8.54	0.04	31.28	46.89	2.05	0.03	55	0.32	273	141	15	24	1321	243	197	9	8	3	
21°30S	MORB	East Pacific Rise	26	2820	9.90	4.10	0.06	37.50	22.99	4.04	0.01	62	0.42	220	31	1	219	1185	-	90	-	-	4	
Slow-spreading mid-ocean ridges																								
Menez Gwen	E-MORB	Mid Atlantic Ridge	26	820	1.75	1.66	0.05	7.12	14.25	25.99	20.77	27	0.15	100	114	30	22	18	63	55	-	-	5	
Lucky Strike	E-MORB	Mid Atlantic Ridge	154	1650	5.37	4.16	0.05	21.05	30.97	12.38	10.85	74	0.30	195	334	28	10	89	110	79	1	1	6	
TAG	MORB	Mid Atlantic Ridge	66	3670	8.34	7.63	0.04	25.13	26.56	8.89	0.01	81	1.40	203	98	15	68	95	30	84		2	7	
Snake Pit	MORB	Mid Atlantic Ridge	71	3500	8.26	6.10	0.05	36.95	36.03	3.00		80	1.79	242	312	27	1	135	135	60	4	25	8	
Krasnov	MORB	Mid Atlantic Ridge	16	3700	2.21	0.14	0.01	38.96	42.90	11.21	0.27	18	1.10	19	66	20	15	628	23	88	10	23	9	
Rainbow	Ultramafic rocks +- gabbros	Mid Atlantic Ridge	116	2400	12.43	14.99	0.03	28.56	32.75	0.71	0.29	188	5.10	391	214	34	490	5086	186	29	10	138	10	
Achadze-1	Ultramafic rocks +- gabbros	Mid Atlantic Ridge	49	4100	14.21	14.11	0.03	32.81	31.74	1.30	0.05	79	6.30	251	231	29	973	2882	200	31	11	347	11	
logatchev-1	Ultramafic rocks +- gabbros	Mid Atlantic Ridge	18	3000	28.03	3.82	0.02	28.60	32.44	2.83	0.08	45	6.50	94	65	26	92	654	772	23	14	249	12	
Back-arc basins																								
Central Manus	MORB, minor BABB	Central Manus basin	152	2500	1.00	17.60	0.60	11.60	24.50	24.20	8.70	514	-	500	425	-	19	24	-	-	<20	-	13	
Vienne Wood	MORB, minor BABB	Central Manus basin	18	2500	1.04	21.07	0.21	6.46	26.19	13.40	2.20	114	3.00	929	156	29	<20	<10	7	82	<5	<10	14	
North Fiji	MORB, E-MORB, minor BABB	North fiji basin	26	1950	7.16	6.38	0.05	30.05	36.67	16.24	0.70	145	0.83	180	175	18	4	156	162	270	<5	<10	15	
Central Mariana trough 18°N	MORB, minor BABB	Marianne trough	11	3675	1.20	10.00	7.40	2.40	17.80	1.20	33.30	185	0.78	465	125	190	-	2	10	5	-	-	16	
Kings Triple Junction	BABB, MORB, low K andesite	Northern Lau basin	78	2000	5.60	28.20	0.10	8.70	-	14.40	-	17	-	555	380	-	5	27	-	-	53	-	17	
Kulo Lasi	MORB, trachy-andesite	Futuna volcanic area	16	1500	0.80	8.97	0.29	26.71	32.21	6.36	1.00	46	<1	433	531	44	41	1121	265	42	15	168	18	
Pacmanus	BABB, MORB, low K andesite, dacite	Eastern Manus basin	111	1650	6.40	24.80	1.40	12.40	24.70	10.80	6.40	290	13.49	960	4600	670	7	7	10	97	57	9	19	
SuSu	BABB, MORB, low K andesite, dacite	Eastern Manus basin	19	1460	14.20	1.10	0.20	20.80	29.70	14.10	5.30	97	14.52	92	7890	385	-	118	39	470	-	-	20	
Vai Lili	BABB, MORB, low K andesite, dacite	Southern Lau basin	24	1710	7.52	20.65	0.21	10.15	28.13	4.41	9.02	113	1.21	566	1213	29	9	1	1	24	73	6	21	
White Church	BABB, MORB, low K andesite, dacite	Southern Lau basin	21	1966	3.80	11.09	0.38	7.06	18.83	21.62	23.98	109	2.26	318	430	85	2	2	2	53	1	0	22	
Northern Valu Fa Ridge	BABB, MORB, low K andesite, dacite	Southern Lau basin	63	1850	3.20	12.10	0.40	6.80	18.10	21.30	21.80	86	2.68	300	450	78	22	3	<5	36	11	8	23	
Central Valu Fa Ridge	BABB, MORB, low K andesite, dacite	Southern Lau basin	87	1750	5.50	27.60	0.30	7.40	27.40	3.80	10.80	120	3.21	855	1900	76	13	3	<5	16	125	19	24	
Southern Valu Fa Ridge	BABB, MORB, low K andesite, dacite	Southern Lau basin	25	1850	1.70	11.60	0.70	32.30	38.90	1.80	3.30	405	1.43	350	3815	21	28	23	35	18	1	11	25	
Arc																								
Myojin knoll	IAB, low K andesite, dacite, Rhyolite	Izu-Bonin arc	37	1300	5.50	21.90	2.30	10.50	-	8.70	6.70	1215	20	1860	5045	3565	-	-	-	-	-	-	26	
Suiyo Seamount	IAB, low K andesite, dacite	Izu-Bonin arc	13	1300	12.80	17.50	0.70	15.00	-	-	<0.1	160	27.73	885	1170	585	4	19	1	120	-	15	27	
Brothers	IAB, minor BABB, med. K andesite	Kermadec arc	9	1600	2.80	3.30	0.20	7.40	-	-	31.90	-	-	105	2155	-	66	39	-	-	-	-	28	
Intracontinental back-arc																								
Jade site - Izena cauldron	Basalt, K andesite, dacite, rhyolite, sediments	Central Okinawa trough	40	1400	3.30	20.20	11.80	6.20	18.10	10.10	7.20	2305	3.13	820	17500	6710	-	-	4205	-	-	-	29	

Source of data: (1) = Fouquet et al. 1996; (2, 3, 4, 14) = Fouquet, unpublished data; (5 to 12) = Fouquet et al., 2010; (13, 16, 17, 19, 20, 23, 24, 25, 26, 27, 28, 29) = Hannington et al., 2004, 2005 and references therein; (15) = Bendel et al. 1993; (18) = this study. (21, 22) = Fouquet et al., 1993. MORB = mid-ocean ridge basalt, E-MORB = enriched MORB; BABB = back-arc basin basalt; IAB = Island-arc tholeite.

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Table 6

Sample	FU-PL02-07	FU-PL02-08	FU-PL02-09	FU-PL02-11
%				
SiO ₂	86.53	43.80	42.83	26.26
Al ₂ O ₃	0.30	0.36	0.49	0.53
Fe ₂ O ₃	4.61	34.50	36.16	52.55
MnO	0.02	1.40	1.34	0.81
CaO	0.29	1.16	1.33	1.41
MgO	0.48	2.92	2.11	1.36
K ₂ O	0.15	0.52	0.47	0.35
Na ₂ O	1.88	3.40	3.13	2.86
TiO ₂	0.02	0.02	0.04	0.04
P ₂ O ₅	0.12	0.87	1.08	1.00
SO ₄	0.15	0.17	0.14	0.18
LOI	6.01	10.70	10.45	11.27
Total	100.56	99.82	99.57	98.62
ppm				
V	10	171	106	211
Co	3	19	13	50
Ni	2	30	30	27
Cu	10	36	48	82
Zn	8	70	73	71
As	11	118	108	93
Rb	4	9	8	6
Sr	38	264	270	300
Y	<3	5	7	3
Zr	<10	<10	<10	<10
Nb	1	1	1	1
Ba	19	127	51	88
Ce	<10	<10	<10	<10
La	<6	6	10	<6
Sc	<3	<3	<3	<3
Mo	6	46	44	49
Pb	9	10	11	12
Sn	5	<3	-	-
Sb	6	<2	-	-
U	4	<2	3	-

Table 6

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Figure 22

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3D Diagram of the current stage of formation of the Kulo Lasi caldera (Stage 4). A: old lava flows. B: fractured volcanic basement. M: magma chamber. R: resurgent dome. C: recent lava flows. Caldera faults: red dotted line = reverse faults; dark blue lines = normal faults. Colored numbers: 1- red arrows: high-temperature hydrothermal fluid generated by deep water-rock reaction. 2- white arrows: seawater heated (>250°C) at shallow levels in contact with the hot recent dyke controlled by the reverse faults (red line). 3- Blue arrow, cold seawater. Black numbers = hydrothermal precipitates: 1- stockwork mineralization (red crosshatching). 2- low-temperature Fe/Mn deposits (brown patches). 3- Low-temperature Si/Fe mineralization (orange patches). 4- native sulfur deposits generated by the discharge of magmatic fluids (yellow dots). 5- recent (< 1 year) black smokers sulfidic chimneys (red triangles). 6- low-temperature silica chimney near the base of the caldera wall (yellow triangles). 7- Diffuse low-temperature venting without precipitates.