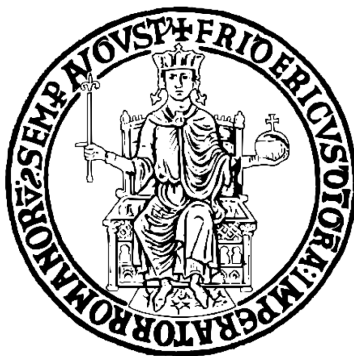


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Lava flow emplacement and lava tube morphologies and mechanisms of formation at Mt. Vesuvius, Italy

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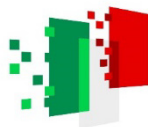
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Abstract

Volcanic eruptions have an important impact on the environment, on urbanised or rural areas. Large eruptions can have a worldwide impact. Effusive eruptions are usually harmless to human lives but can cause great damage to infrastructures, cultivated lands and houses as lava flows burn and bury anything they encounter. The distance travelled by lava flows is controlled by many parameters, such as eruptive rate and rheology of the lava. The formation and development of lava tubes, highly efficient lava transport structures, within a flow can greatly increase its distance of emplacement.

Somma-Vesuvius is famous for its explosive activity, but the last eruptive cycle from 1631 to 1944 was a long period of open-conduit conditions with effusive to mildly explosive eruptions. Historical documentation describes the presence of lava tubes at Vesuvius, opening the possibility of finding lava tubes on the slopes of the volcano. Vesuvius and its surroundings are highly inhabited and present a high risk of destruction by effusive eruptions. Past events are reminders of such risk. It is thus crucial to better understand the emplacement of lava flows and the formation and development of lava tubes, especially at volcanoes producing silica-undersaturated potassium-rich lavas due to the absence of studies in such context.

The work carried out during this Ph.D. aimed at understanding lava flow emplacement, and lava tube formation and development in the lava flow fields present at Vesuvius. A detailed morphometry analysis of the largest lava tube formed in the 1858 flow field was conducted using a combination of Terrestrial Laser Scanner and airborne imagery. Lavas forming the tube were characterised by petrological, geochemical, and textural analyses performed by Scanning Electron Microscope, Electron Microprobe Analysis and synchrotron x-ray microtomography. The stability of the lava tube structure was assessed by Acoustic Emissions tests with a novel signal processing method combining Empirical Mode Decomposition (EMD) and Hilbert-Huang Transform (HHT).

The analyses show that the 1858 lava flow field reached its maximum extent quickly and then expanded both laterally and vertically. The size of the flow field appears to be influenced by the topography and the amount of crystals in the lava. Our findings suggest that the lava tube studied was formed by inflation of a lava flow breakout. The tube remained active for a long period and experienced at least twelve drainage and refill cycles. The thermo-redox conditions inside the tube were unstable and did not display a consistent trend during the activity of the tube. Microlitic crystallisation within the lava tube was enhanced at the sides of

the conduit due to thermal conditions and shear rates compared to the centre of the tube. The structure of the lava tube is fractured and contains loosened masses in some areas, but the overall structure remains stable.

Moreover, a geotinerary and virtual tour along the 1858 lava flow field was designed to showcase five peculiar volcanic landforms, three lava tubes, an ephemeral vent and a tumulus, that allows people to evidence effusive activity and discover volcanic features formed during lava flow emplacement at Somma-Vesuvius. A website was developed to host the virtual tour, which comprises 360° images and an explicative panel for each site. The website aims to sensibilibize people either not used to hiking or with physical impediment about volcanic landforms and their scientific relevance. The pertinence of the geotinerary has been assessed by a Strengths, Weaknesses, Opportunities and Threats (SWOT) analysis. The geotinerary was conceptualized for visitors to discover volcanoes and volcanic activity and may contribute to divert tourist traffic away from the volcanic cone, while raising awareness on volcanic processes and volcanic hazards.

Keywords: synchrotron x-ray microtomography; 1858 lava flow field; thermo-redox conditions; effusive activity; volcanic hazards; volcano tourism

Sintesi

Le eruzioni vulcaniche hanno un grande impatto sull'ambiente, sulle aree urbane e su quelle rurali. Le grandi eruzioni possono avere conseguenze a livello mondiale. Le eruzioni effusive sono generalmente meno pericolose per la vita umana, ma possono causare gravi danni alle infrastrutture, ai terreni agricoli e alle abitazioni, poiché le colate laviche incendiano e seppelliscono tutto ciò che incontrano sul loro cammino.

La distanza percorsa da una colata lavica dipende da diversi parametri, come il tasso eruttivo e la reologia della lava. La formazione e lo sviluppo di tubi di lava, strutture efficienti per il trasporto della lava all'interno di una colata, possono aumentare notevolmente la distanza di propagazione della colata.

Il Somma-Vesuvio è famoso per la sua attività esplosiva, ma l'ultimo ciclo eruttivo, dal 1631 al 1944, è stato caratterizzato da un lungo periodo di condotto aperto con eruzioni effusive e moderatamente esplosive. Documentazioni storiche descrivono la presenza di tubi di lava sul Vesuvio durante la formazione dei campi lavici, il che apre la possibilità di poter trovare e studiare tali strutture sui fianchi del vulcano. Il Vesuvio e le aree circostanti sono densamente abitate e soggette a un elevato rischio di distruzione in caso di eruzioni effusive: gli eventi passati ci ricordano chiaramente questo pericolo. Pertanto, è cruciale comprendere meglio lo sviluppo delle colate laviche e la formazione dei tubi di lava, soprattutto nei vulcani che emettono lave a basso contenuto di silice e ricche di potassio perché questo tipo di fenomeni sono stati spesso ignorati.

Questo lavoro presenta un'indagine multidisciplinare sullo sviluppo delle colate laviche e sulla formazione dei tubi di lava generati dalla colata lavica del 1858 sul Vesuvio. È stata condotta un'analisi dettagliata della morfologia del tubo più grande della colata utilizzando una combinazione di scansione laser terrestre (Terrestrial Laser Scanner) e immagini aeree. Sono state effettuate analisi petrologiche, geochimiche e tessiturali delle lave che costituiscono il tubo mediante microscopio elettronico a scansione (SEM), microsonda elettronica (EMPA) e microtomografia a raggi X da sincrotrone (X-ray μ CT). La stabilità della struttura del tubo è stata valutata attraverso test di emissioni acustiche, applicando un innovativo metodo di elaborazione del segnale che combina la Decomposizione Modale Empirica (EMD) e la trasformata di Hilbert-Huang (HHT).

Le analisi mostrano che la colata lavica del 1858 raggiunse rapidamente la sua massima lunghezza, per poi espandersi lateralmente e verticalmente. La lunghezza della colata è stata influenzata dalla topografia e dalla quantità di cristalli presenti nella lava. I nostri risultati suggeriscono che il tubo principale da noi trovato si sia formato in seguito al rigonfiamento di una colata lavica secondaria. Il tubo rimase attivo per un lungo periodo e ha subito almeno 12 cicli di riempimento e svuotamento. Le condizioni termiche e di ossidoriduzione all'interno del tubo sono apparse instabili e non seguono tendenze regolari nel tempo; le condizioni termiche e di ossidazione hanno avuto un significativo effetto sulla cristallizzazione all'interno del tubo, visto che la cristallizzazione appare diversa dal resto della colata, in particolare lungo le pareti del tubo stesso. La struttura del tubo è fratturata e contiene, in alcune zone, masse instabili, ma nel complesso la struttura risulta stabile.

Per rendere fruibile ai turisti, è stato progettato un geoitinerario e un tour virtuale lungo il campo lavico del 1858; questo per mostrare cinque particolari morfologie vulcaniche: tre tubi di lava, una bocca effimera e un tumulo. Questi siti consentono di osservare l'attività effusiva passata e di scoprire le caratteristiche vulcanologiche delle colate laviche del Somma-Vesuvio. È stato inoltre creato un sito web per ospitare un tour virtuale dell'area, che comprende immagini a 360° e un pannello esplicativo per ogni punto di interesse.

Il sito web ha lo scopo di sensibilizzare le persone che non sono abituate alle escursioni o che presentano limitazioni fisiche, permettendo loro di conoscere le morfologie vulcaniche e la loro rilevanza scientifica. La pertinenza del geotour è stata valutata mediante un'analisi SWOT che valuta i punti di forza (Strengths), debolezze (Weaknesses), opportunità (Opportunities) e minacce (Threats). Il geotour è stato ideato per consentire ai visitatori di scoprire i vulcani e l'attività vulcanica, contribuendo al tempo stesso a sensibilizzare le persone sui processi e sui rischi vulcanici.

Parole chiave: microtomografia a raggi X da sincrotrone; campo di colata lavica del 1858; condizioni termiche e redox; attività effusiva; rischi vulcanici; turismo vulcanico.

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GENERAL INTRODUCTION

Since the dawn of time, volcanoes have posed significant threats to nearby communities. Volcanic dangers are diverse from direct hazards, caused by the eruption itself, such as pyroclastic flows, lava flows, and ash falls to indirect hazards, induced by the eruption, like lahars, flank collapses, tsunamis, pollution and famine (Thierry *et al.*, 2008; Sandri *et al.*, 2014; Weir *et al.*, 2024; Garcia *et al.*, 2025). Whilst generally being unharmed for human lives (Harris, 2015), lava flows destroy everything they encounter, inflicting severe socioeconomic damages (Pesaresi *et al.*, 2008; Coltelli *et al.*, 2012; Harris, 2015; Cubellis *et al.*, 2016; Brown *et al.*, 2017; Del Negro *et al.*, 2019; Meredith *et al.*, 2022, 2023). Historical examples of cities destroyed by lava flows abound: the 1669 lava flow at Mount Etna reached the city of Catania, the 1944 lava flow at Vesuvius destroyed the villages of San Sebastiano and Massa di Somma, and the 1973 Eldfell eruption in Iceland forced the evacuation of the island (Chester *et al.*, 2007; Branca *et al.*, 2013; Del Negro *et al.*, 2019). More recently, eruptions at La Palma in 2021 (Carracedo *et al.*, 2022a) and Nyiragongo in 2002 and 2021 (Tedesco *et al.*, 2007; Boudoire *et al.*, 2022) destroyed inhabited areas. These past and recent events testify to the risk and impact of lava flows on human activities. The distance a lava flow travels is largely determined by its transport mechanisms, with lava tubes playing a critical role in extending flow lengths (Peterson *et al.*, 1994; Keszthelyi, 1995).

The first documented description of a lava tube likely dates to the 18th century, when Eggert Olafsen and Bjarni Povelsen, during their Danish-sponsored expedition to Iceland from 1752 to 1757, explored a cavern called Surtur by locals (now called Surtshellir), coming from *Surtr*, a fire giant in Nordic mythology, supposed to devastate the world during the battle of Ragnarök. In their book describing Iceland, first published in 1772 (Ólafsen and Povelsen, 1805), they noted: *‘This varnish, as well as the stalactites just mentioned, affords a certain proof of the operation of subterraneous fires, and that the lava, in a state of fusion, has passed, like a rivulet, through this channel, while it began to cool on the sides and top of the cavern’*. They describe a cave extending 1.6 kilometres with a height of 10 metres and a width of 16 metres inside. Surtshellir was formed at the end of the 9th century (Smith *et al.*, 2021). It remained the longest known lava tube until the end of the 19th century.

The first account of an active lava tube was published in a letter to the Missionary Herald in 1844 by Titus Coan who observed a ‘subterranean canal’ during the 1843 eruption in Hawaii: *‘The lava on which we were treading gave indubitable evidence of powerful igneous action below, as it was hot and full of seams, from which smoke and gas were escaping. But we soon had ocular demonstration of what was the state beneath us; for in passing along we came to an*

opening in the super incumbent stratum, of two yards long and ten wide, through which we looked, and at the depth of fifty feet, we saw a vast tunnel or subterranean canal, lined with smooth vitrified matter, and forming the channel of a river of fire, which swept down the steep side of the mountain with amazing velocity' (Coan, 1844). Later, E.D. Baldwin described similar structures during the 1880-81 Mauna Loa eruption, noting: *'openings in the top of the flow that exposed the tunnel of red-hot flowing lava beneath'* (Baldwin, 1953).

Despite these early discoveries, systematic research on lava tubes only began in the 20th century. In February 1921, Thomas Jaggar observed and described, at Kilauea's Halema'uma'u pit crater, how *'the lava shot down the slope like a sluiceway; then it closed itself in and the river formed a tunnel of crust'* (Jaggar, 1921). Hawaiian volcanoes, with their frequent effusive activity, became natural laboratories for studying active lava tubes; thus, many observations were made in Hawaii and largely contributed to a better understanding of the formation, development and activity of lava tubes (Wentworth and Macdonald, 1953; Greeley, 1971a, 1971c, 1972, 1987; Peterson *et al.*, 1994; Kauahikaua *et al.*, 1998). Concurrently, inactive lava tubes were explored and surveyed in Victoria and Queensland, Australia (Ollier and Brown, 1965; Atkinson *et al.*, 1975; Webb *et al.*, 1982), in Hawaii, United States (Greeley, 1987; Allred and Allred, 1997, 1998; Kempe and Ketz-Kempe, 2016), and in Sicily, Italy (Guest *et al.*, 1980; Calvari and Pinkerton, 1998, 1999) revealing morphologies, internal features (stalactites, wall linings, ...), and proposing formation mechanisms. For decades, lava tubes were thought to form exclusively in basaltic volcanic centres and within pāhoehoe lava flows. However, Calvari and Pinkerton (1998) demonstrated their presence in 'a'ā lava flows at Mount Etna. This thesis extends this compositional range further, documenting lava tubes in silica-undersaturated alkali-rich lavas, a previously unrecognised context.

Somma-Vesuvius, one of the world's most tragically famous volcanoes, has been surrounded by human activity since at least 3000 years when the Greeks founded Parthenope, reestablished as Neapolis ('New city') in 470 BC, and became one of the largest cities of the Great Greece. While the Greeks never witnessed Somma-Vesuvius in an eruptive state, the Romans have seen, experienced and lived with volcanic activity at Somma-Vesuvius after the renewed activity of the volcano with the 79 AD eruption that destroyed the cities of Pompeii and Herculaneum (Fig. 1), during which the first description of an explosive eruption was made by Pliny the Younger.



Figure 1. Ancient mural of Bacchus and Somma-Vesuvius before the 79 AD eruption, House of the Centenary, Pompeii.

Though renowned for its explosive activity, Somma-Vesuvius has also undergone prolonged effusive phases, most recently from the 17th to the 20th century. It is during that eruptive cycle, in the middle of the 19th century, that lava tubes were first mentioned at Vesuvius when Luigi Palmieri, director of the Vesuvius Observatory, described an unusual phenomenon during the 1857 eruption: *‘At this time, while the lava was flowing on the slope of the cone, no point could be observed on the summit, so that I walked on the still hot petrified lava under which the lava was flowing down from the summit of the mountain’* (Palmieri, 1859). A few other testimonies of lava tubes were written later. A potential lava tube was drawn by John Phillips in a book published in 1869 (Fig. 2; Phillips, 1869).

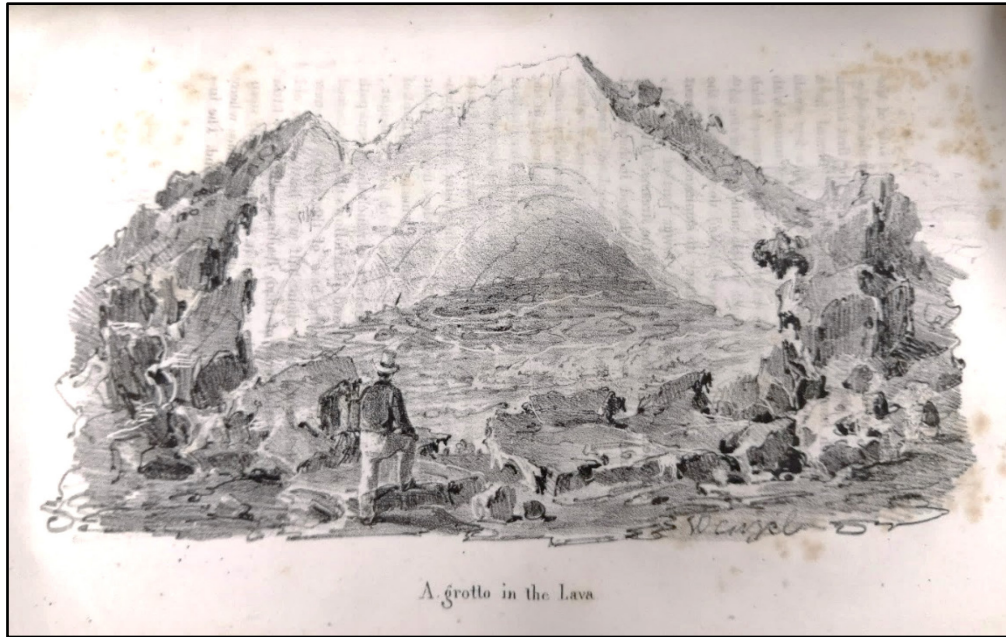


Figure 2. Drawing of a lava cave at Mount Vesuvius (Phillips, 1869)

Much later, a detailed description of a lava tube found in the 1858 lava flow field was made by Alessandro Malladra (1917) and published in the Bulletin of the Society of Naturalists of Naples (Fig. 3). Until 2024, Malladra's work remained the sole study of lava tubes at Vesuvius.

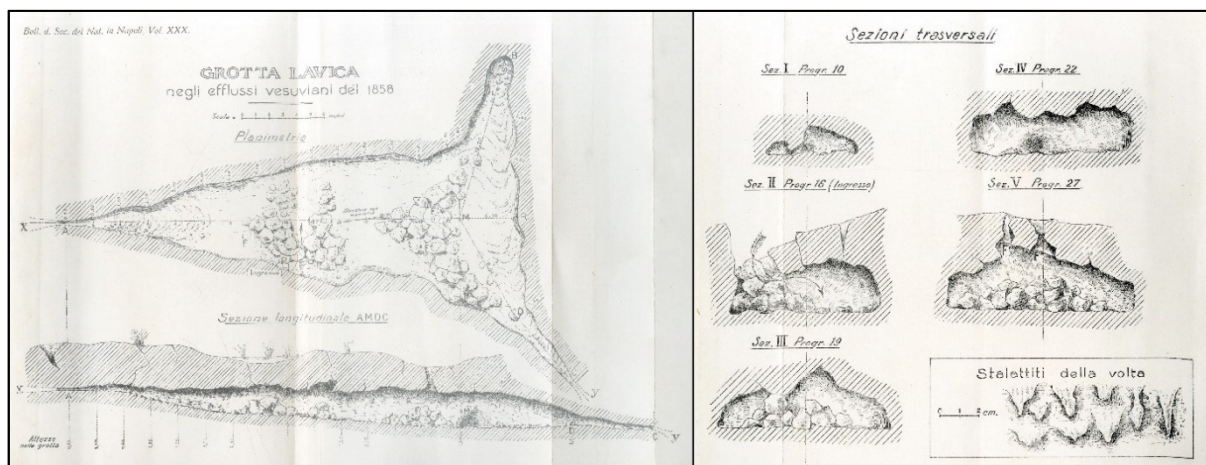


Figure 3. Drawn sections of a lava tube formed in the 1858 lava flow field at Mount Vesuvius by Malladra (1917)

This thesis aims to advance our understanding of lava tube formation and development through three primary objectives: (1) reconstructing lava flow emplacement dynamics, (2) characterising lava tube morphometry, and (3) analysing the petrological, geochemical and structural properties of the lavas forming the studied lava flows and lava tubes.

The main motivation for this study is to improve hazard mitigation during effusive eruptions in Italy and worldwide. Lava tube formation within a flow can increase its hazard. This is particularly important for densely inhabited volcanic centres like Somma-Vesuvius and

Mount Etna. Fatalities have happened in recent history at both edifices, notably the 1944 eruption at Somma-Vesuvius that emplaced a lava flow that destroyed the villages of San Sebastiano and Massa di Somma (Chester *et al.*, 2007), and the 1669 eruption of Mount Etna that emplaced a lava flow that reached 17 kilometres and destroyed part of the city of Catania (Branca *et al.*, 2013). Understanding the formation and development of lava tubes can help develop enhanced numerical models for hazard mapping and near-real-time lava flow forecasting. This work seeks to improve the preparation and response to future eruptive crises. While the thesis has local implications, it aims to give constraints on lava flow emplacement and lava tube formation at a global scale.

The second motivation is the renewed scientific interest in lava tubes for planetary exploration purposes after the discovery of extensive lava flow fields on Venus, Mars and the Moon. Scientists preparing exploration missions on other extraterrestrial bodies believe lava tubes are interesting structures to be used as efficient and cost-effective shelters against cosmic rays and asteroid impacts (Hörz, 1985; Angelis *et al.*, 2002; Haruyama *et al.*, 2012). Our findings may inform the detection and characterisation of extraterrestrial lava tubes, aiding in the preparation of planetary exploration missions.

During this Ph.D., I focused on the effusive activity of Somma-Vesuvius, where I rediscovered the presence of lava tubes by finding and re-analysing the lava tube described in the work of Malladra (1917), and by discovering new tubes. The methodology integrates a multidisciplinary approach, including: (1) extensive fieldwork to discover lava tubes in the 1858 lava flow field, (2) a detailed morphometric analysis of the largest lava tube to characterise its structure and dimensions, (3) petrological, geochemical, and textural analysis of tube-forming lavas to characterise their compositional and physical properties, (4) a stability assessment by acoustic emission testing to evaluate the mechanical integrity of the structure, and (5) the development of a geoitinerary that highlights five interesting volcanological structures, comprising three lava tubes, an ephemeral vent and a tumulus. The geoitinerary offers both a great tool to disseminate scientific knowledge and to raise awareness on volcanic hazards, specifically about lava flow hazard at Somma-Vesuvius.

CHAPTER 1

Geological context of the study and state-of-the-art

The first chapter of the thesis is dedicated to the description of the geological context and of the state-of-the-art of lava flow emplacement and lava tube formation. The first section introduces the volcanism of Italy. The following section presents the volcanic area of interest of this thesis, Somma-Vesuvius. The third section presents the state-of-the-art of lava flow emplacement. The fourth presents the state-of-the-art of lava tube formation. The last section details the presence of lava tubes on other planetary bodies and their importance for future exploration missions.

1.1. Volcanism in Southern Italy

Europe, and more specifically the Mediterranean area, is affected by complex geodynamic processes controlled by the convergence of the Eurasian and African plates since the Cretaceous (Fig. 4; Doglioni *et al.*, 1999; Mantovani *et al.*, 2002; Carminati *et al.*, 2012), which caused the closing of the Tethys Sea. The still ongoing collision of the two tectonic plates, and implicated microplates, compressed and piled up sedimentary rocks resulting in the formation of two mountain chains, the Alpine (120 Ma to 0 Ma) and Apennine (13 Ma to 0 Ma; Carminati *et al.*, 2010) orogenic belts, and in the opening of two oceanic basins, the Ligurian-Provencal sea (33 Ma to 14 Ma) and the Tyrrhenian sea (13 Ma to 0 Ma; Carminati *et al.*, 2010). Resulting from such complex geodynamic activity, a widespread magmatic activity developed.



Figure 4. Simplified tectonic map of western Europe showing the plate boundaries, collisional zones, and directions of extension and tectonic transport. MHSZ—mid-Hungarian shear zone; MP—Moesian platform; RM—Rhodope massif; IASZ—Izmir-Ankara-Erzincan suture zone. Modified from Dilek (2006).

Italy is scattered by magmatism. Volcanoes formed primarily along the Tyrrhenian coast of the Italian peninsula and within the Tyrrhenian Sea, where some edifices are fully or partially submerged (Fig. 5). The earliest volcanic activity began in Sardinia and the Ligurian Sea during the Miocene (12 Ma), before migrating to Tuscany (8.5 Ma), the Tyrrhenian Sea basin, and finally to central and southern Italy (< 1 Ma). While most of the volcanic activity ceased tens of thousands of years ago, several volcanic centres remained persistently active, such as Mount Etna in Sicily and Stromboli in the Aeolian Islands, while others, including Somma-Vesuvius and Campi Flegrei in the Campanian region, are currently in a state of quiescence.

The magmatic rocks found across Italy nearly span the entire compositional spectrum of the Total Alkali Silica (TAS) classification (Le Bas *et al.*, 1986; Le Maitre *et al.*, 2002), ranging from subalkaline to alkaline, and from mafic to silicic. This diversity not only reflects the mantle-derived origin of the magmas, highlighting the significant heterogeneity of mantle sources beneath the region, but also results from magmatic processes such as fractional crystallisation and crustal contamination.

Geographically, geodynamically and geochemically, Italy's magmatism can be divided into eleven distinct provinces (Fig. 5, Table 1; Peccerillo, 2017). These provinces are further categorised based on their association with orogenic processes (linked to subduction) or anorogenic processes (unrelated to subduction). Today, the only active subduction zone in the region is the Aeolian subduction, where the eastern Mediterranean oceanic lithosphere subducts beneath northeastern Sicily and southwestern Calabria. This subduction is responsible for the volcanism of the Aeolian arc (e.g. Alicudi, Filicudi, Panarea, Salina, Vulcano, Stromboli, Lipari). In contrast, volcanoes such as Mount Etna are not directly related to subduction processes.

The Campania magmatic province encompasses Somma-Vesuvius, Campi Flegrei and the Islands of Procida, Vivara, and Ischia. This volcanic region is situated at the intersection of the Apennine fold-thrust belt and the Tyrrhenian back-arc basin, where two main faulting systems oriented NE-SW and NW-SE converge (Peccerillo, 2017). The magmatic province is characterised by mafic to felsic alkaline potassic magmas, with geochemical evidence suggesting that melts are generated from a hybrid source, incorporating both Ocean Island Basalt (OIB)-like and subduction-related components (D'Antonio *et al.*, 2013; Peccerillo, 2017; Piochi *et al.*, 2021).



Figure 5. Magmatic provinces in Italy as grouped based on major, trace element and isotopic characteristics. Modified from Peccerillo (2017).

Table 1. List of the Italian magmatic provinces with the main volcanic centres, their physical and geochemical characteristics. From Peccerillo and Frezzotti, 2015.

Magmatic province	Main magmatic centres and ages (in Ma)	Volcanology	Petrology-geochemistry
<i>Orogenic magmatism</i>			
Tuscany (8.5–0.3 Ma)	Acid intrusions: Elba (8.5–6.3), Montecristo (7.1), Giglio (5), Campiglia–Gavorrano (5–4.3). Various hidden intrusions at Larderello–Amiata (<5) Acid lavas: San Vincenzo (4.5), Roccastrada (2.5), Amiata (0.3), Cimini (1.3–1.1), Tolfa–Cerite–Manziana (4.2–1.8) Mafic centres: Capraia (7.7–4.8), Elba (5.8), Orciatice and Montecatini (4.1), Radicofani (1.3), Cimini (0.9), Torre Alfina (0.8). The Sisco lamproitic dyke (14) in Corsica is often considered to belong to the Tuscany Province	Felsic rocks: granitoid rocks, aplites, pegmatites. Monogenetic lava flows and domes, and polygenetic cones (Amiata). A few ignimbrites (Cimini, Cerite); no pyroclastic rocks in other centres. Mafic rocks: small extrusive and subvolcanic bodies. Stratovolcano at Capraia. Few or no pyroclastic products	Mafic magmas: calc-alkaline, shoshonitic and ultrapotassic (lamproites) rocks, nearly saturated in silica, with high MgO. LILE abundances and $^{87}\text{Sr}/^{86}\text{Sr}$ (c. 0.708–0.717) increase from calc-alkaline to lamproitic compositions Felsic magmas: metaluminous to peraluminous rocks originated by crustal anatexis (acid rocks) plus mixing between these and various types of mafic magmas and crustal melts. $^{87}\text{Sr}/^{86}\text{Sr} \sim 0.712\text{--}0.723$, with highest values in rocks with no evidence of mixing with mafic magmas (e.g. Roccastrada and some San Vincenzo rhyolites)
Intra-Apennine Province (0.6–0.3 Ma)	San Venanzo (0.26), Cupaello (0.64), Polino (0.25), Acquasparta, Oricola (0.53), etc.	Monogenetic pyroclastic centres and rare lavas (San Venanzo and Cupaello)	Ultrapotassic melilititic (kamaufugites) composition. Carbonate-rich pyroclastic rocks of uncertain origin. Similar trace element patterns and radiogenic isotope composition ($^{87}\text{Sr}/^{86}\text{Sr} \sim 0.710\text{--}0.711$) to mafic rocks of the Roman Province
Roman Province (Latium) (0.8–0.006 Ma)	Vulsini (0.6–0.13), Vico (0.4–0.09), Sabatini (0.8–0.09), Alban Hills (0.6–0.006)	Large multicentre volcanic complexes and stratovolcanoes with polygenetic calderas formed prevalently by pyroclastic rocks and minor lavas	Roman-type potassic (trachybasalt to trachyte) and ultrapotassic (leucite tephrite to phonolite) rocks with high enrichments in LILE (K, Rb, Cs, Th, LREE, etc.) and depleted in HFSE (Ta, Nb, Zr). High $^{87}\text{Sr}/^{86}\text{Sr} \sim 0.709\text{--}0.711$

Ernici (or Mid Latin Valley)–Roccamonfina (0.7–0.1 Ma)	Ernici: Pofi, Ceccano, Patrica, etc. (0.6–0.25), Roccamonfina (0.6–0.05)	Monogenetic pyroclastic and lava centres (Ernici or Mid Latin Valley), and a stratovolcano with a central caldera and intra-caldera domes (Roccamonfina)	Silica-undersaturated Roman-type leucite-bearing ultrapotassic rocks, and younger Campania-type saturated to slightly undersaturated calc-alkaline to potassic rocks. The latter show lower LILE and radiogenic Sr than ultrapotassic rocks
Campania (including Ponza and nearby islands) (4 Ma–present)	Somma–Vesuvius (0.03–AD 1944), Campi Flegrei (0.3–AD 1538), Ischia (0.15–AD 1302), Procida (0.05–0.01), Ponza (4–1), Palmarola (1.5), Santo Stefano (1.2–0.6), Ventotene (0.9–0.1). Buried calc-alkaline volcano (2 Ma)	Stratovolcanoes and multicentre complexes mostly formed by pyroclastic rocks with minor lavas. Rhyolitic lavas and pyroclastic rocks at Ponza and Palmarola. Calc-alkaline rocks found by deep borehole drillings or as rare xenoliths	Shoshonitic–potassic (trachybasalt to trachyte) and ultrapotassic (leucite tephrite to phonolite) rocks, less enriched in LILE and radiogenic Sr ($^{87}\text{Sr}/^{86}\text{Sr} \sim 0.705\text{--}0.708$) than equivalent Roman rocks, but similar to Stromboli potassic rocks
Vulture (0.7–0.1 Ma)	Vulture, Melfi	Stratovolcano with a summit caldera, intracaldera maars and some parasitic centres formed by dominant pyroclastic rocks and lavas	Mildly to strongly silica-undersaturated Na–K-rich (variable K/Na) melilitites, tephrites, phonolites, foidites with hauyne. Late carbonatites. Intermediate OIB- and arc-type trace element signatures
Aeolian Arc (0.27 Ma–present)	Alicudi (0.1–0.03), Filicudi (0.25–0.03), Salina (0.25–0.016), Vulcano (0.12–AD 1888), Lipari (0.27–AD 1220), Panarea (0.15–0.008), Stromboli (0.2–present)	Large stratovolcanoes with some calderas, formed by alternating pyroclastic rocks and lavas. Active volcanism in the central–eastern sector of the arc	Dominant calc-alkaline (basalt–andesite–rhyolite) compositions in the western–central arc. Dominant shoshonitic and potassic rocks at Vulcano and Stromboli. More primitive radiogenic isotope compositions in the western–central islands. Silicic rocks are concentrated in the central island of the archipelago (Lipari, Vulcano)
Southern Tyrrhenian sea floor (12 Ma–present)	Cornacya (12), Anchise, Sisifo, Enarete, Eolo, Lametini, Alcione, Palinuro, Marsili Sites 650 and 651	Seamounts of various ages, shapes and volumes. Marsili is a 60 km NE–SW elongated ridge, rising c. 3000 m above the actively spreading Marsili basin	Shoshonitic to calc-alkaline and arc-tholeiitic compositions

Anorogenic magmatism

Sicily (7 Ma–present)	Etna (0.5–present), Iblei (7–1.5), Ustica (0.75–0.13), Pantelleria (0.3–0.005), Linosa (1–0.5), Sicily Channel seamounts (Miocene to AD 1891), ODP	Stratovolcanoes, diatremes, small plateaux, cinder cones, tuff cones and tuff rings, etc. Active volcanism at Etna and in the Sicily Channel (AD 1831 and 1891)	Tholeiitic to Na-alkaline rocks (basanite, hawaiite, trachyte) at Etna, Iblei, Linosa. Na-transitional basalts to peralkaline trachyte and rhyolite at Pantelleria. Trace element and isotopic signatures typical of FOZO–OIB-type rocks
Sardinia (12–0.1 Ma)	Isola del Toro (12), Capo Ferrato (c. 5), Montiferro (4–2), Orosei–Dorgali (4–2), Monte Arci (3.7–2.3), Logudoro (3–0.1)	Stratovolcanoes, basaltic plateaux and monogenetic centres	Mafic to silicic subalkaline, transitional to Na-alkaline rocks, sometimes with a K-affinity. Variable radiogenic isotope compositions, mostly with EM-1 characteristics, are unique in Europe
Southern Tyrrhenian Sea floor (7–0.1 Ma)	Magnaghi (3), Marsili, Vavilov, Aceste, Prometeo, ODP Sites 654 and 655, DSDP Site 373	Seamounts and lava fields	MORB-type tholeiite, Na-transitional and alkaline basalt to trachyte compositions

1.2. Somma-Vesuvius

1.2.1. Structure of the edifice and eruptive history

The Somma-Vesuvius volcanic complex is situated in southern Italy, at the southern extremity of the Campanian plain, within the Gulf of Naples. It lies approximately 15 kilometres east of the metropolitan city of Naples (Fig. 6).

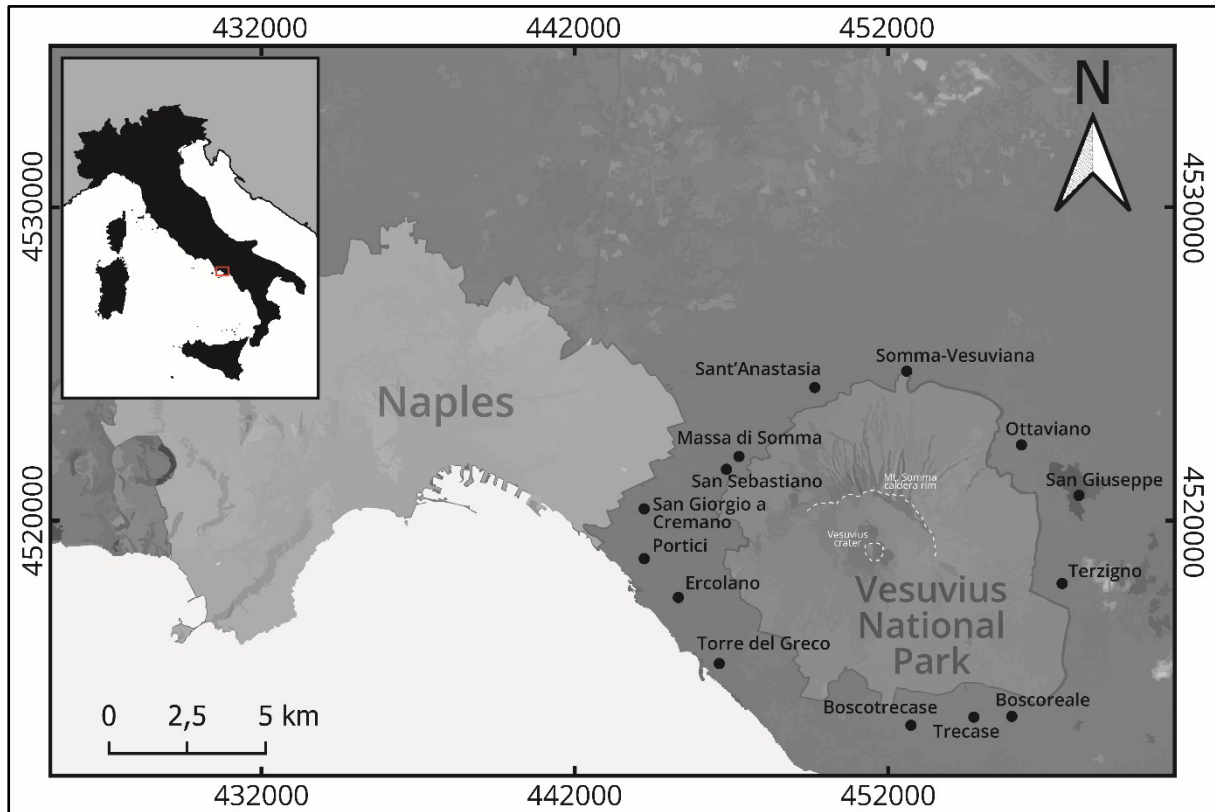


Figure 6. Location of Somma-Vesuvius in southern Italy, close to the city of Naples.

Somma-Vesuvius is a composite stratovolcano comprising two edifices, Monte Somma, an older, partially collapsed structure, and Vesuvius, a younger cone that rises 1281 metres above sea level (Fig. 7). The volcanic complex is characterised by an asymmetric composite caldera, formed through repeated collapses of Monte Somma following highly explosive eruptions that drained the magmatic reservoir (Fig. 6 and 7; Cioni *et al.*, 1999; Arrighi *et al.*, 2001; Cioni *et al.*, 2008). The northern rim of the caldera remains significantly higher than its southern and western counterparts, reflecting the structural asymmetry of the system. Vesuvius grew within the caldera after the Pompeii eruption in 79 AD (Cella *et al.*, 2007; Cioni *et al.*, 2008). Somma-Vesuvius is part of the Roman magmatic province. The most ancient activity produced slightly silica-saturated products (K-trachyte to K-latitude). The composition evolved through time to silica undersaturated potassium-rich products (phonolite-phonotephrite-tephrite). The volcanic

edifice is underlain and thus rests on a limestone basement at a depth of 2 kilometres (D'Argenio *et al.*, 1973; Ippolito *et al.*, 1975; Brocchini *et al.*, 2001).

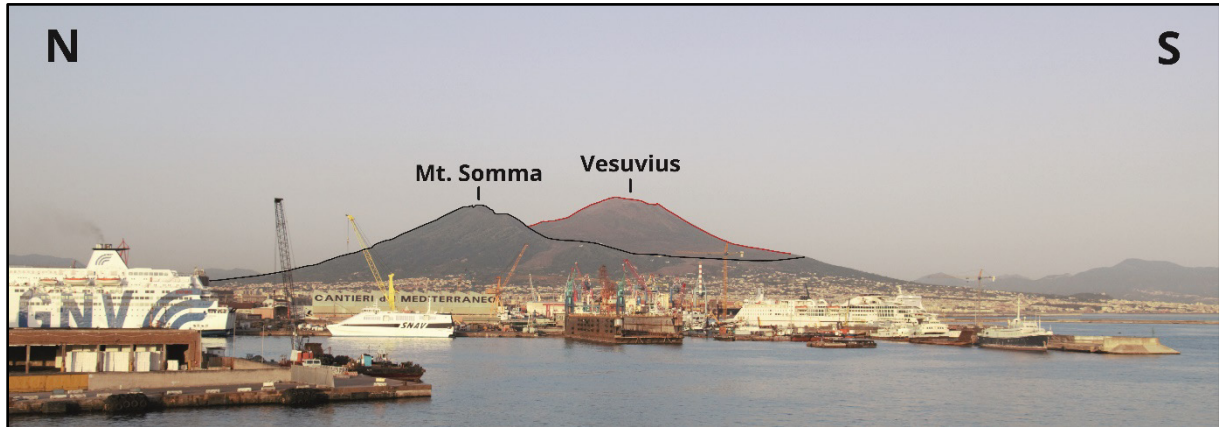


Figure 7. Photograph of the western side of Somma-Vesuvius showing both edifices.

Volcanic activity at Somma-Vesuvius initiated around 400 ka BP, as evidenced by lava flows identified in the Trecase well (Brocchini *et al.*, 2001). However, the eruptive record prior to 40 ka BP is not well documented, while significantly more of the activity is known after 25 ka BP, (Santacroce, 1987; Rolandi *et al.*, 1998; Santacroce *et al.*, 2008). Over the past 25 ky, the volcano has displayed an eruptive behaviour composed of repose periods, long periods of open-conduit conditions (Hawaiian and Strombolian eruptions) and periods of closed-conduit (sub-Plinian and Plinian eruptions). Up to 18 ka BP, Somma-Vesuvius had a mostly effusive activity that formed the edifice of Monte Somma by thin lava flow piling intercalated by spatter and cinder deposits. Then the activity of the volcano changed to mostly explosive (Santacroce and Sbrana, 2003; Santacroce *et al.*, 2008). At least four major Plinian eruptions produced by Somma-Vesuvius were identified, starting with the Pomici di Base eruption 22,000 yr BP, and followed by the Mercato eruption 8,800 yr BP, the Avellino eruption 3,800 yr BP, and the Pompeii eruption in 79 AD (Cioni *et al.*, 2008; Santacroce *et al.*, 2008). The sporadic Plinian eruptions were intercalated by sub-Plinian eruptions, notably the Greenish eruption 16,000 yr BP, the Pollena eruption in 472 AD and the eruption of 1631. Plinian eruptions dramatically reshaped the volcano's morphology. Deposits from the eruptions that followed the 79 AD Plinian eruption filled the caldera and buried the western and southern caldera rims. The highly explosive eruptions were followed by repose periods or long periods of open-conduit conditions during which Somma-Vesuvius produced effusive to mildly explosive eruptions. From all the known eruptions of Somma-Vesuvius, a positive correlation exists between the duration of the

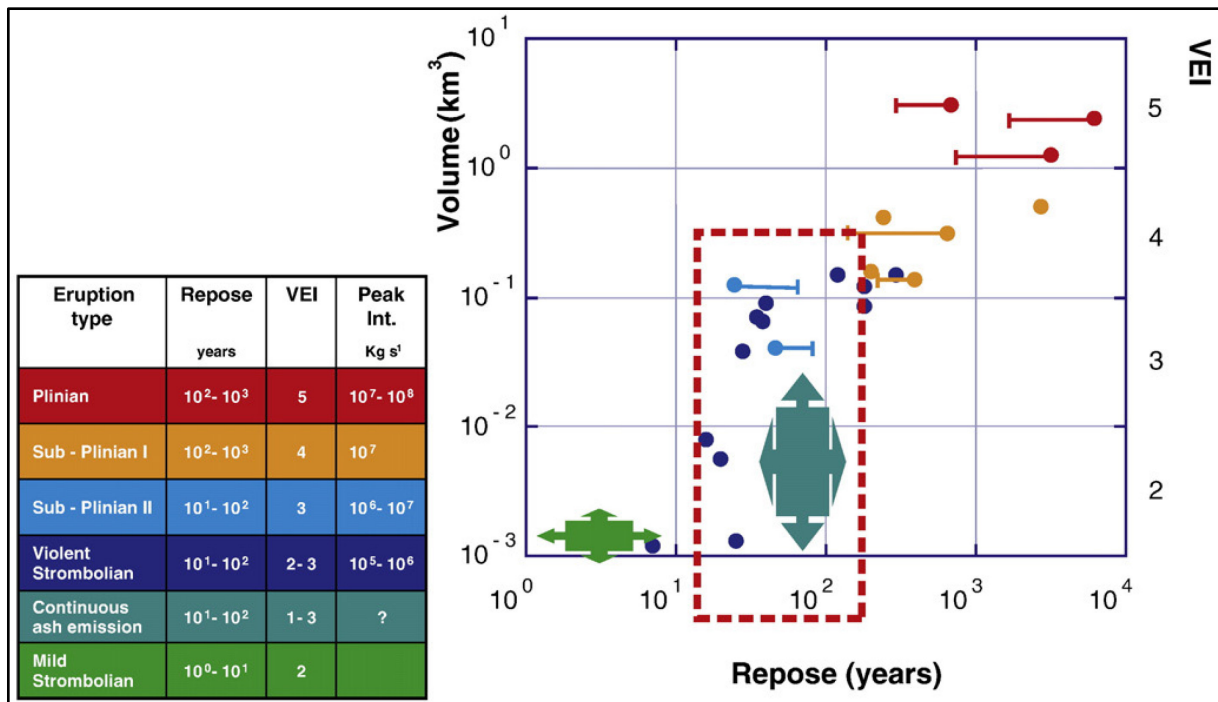


Figure 8. Apparent repose time in function of the magnitude of the eruption. The main physical parameters associated to each eruption type are shown in the inset. Figure from Cioni et al. (2008)

repose periods and the magnitude of the subsequent eruption (Fig. 8; Santacroce, 1983; Cioni *et al.*, 2008). Mafic lavas typically erupted during periods of open-conduit conditions, while more felsic, differentiated lavas were associated with highly explosive eruptions, following prolonged quiescence.

The last eruptive cycle of Vesuvius began after the sub-Plinian eruption of 1631 and continued as a prolonged period of open conduit condition, consisting mainly of summit and lateral effusive eruptions intercalated by semi-persistent, mild explosive activity (Strombolian and Hawaiian type; Carta *et al.*, 1981; Arrighi *et al.*, 2001). Arno *et al.* (1987) identified eighteen cycles of activity within the last eruptive cycle (1631-1944), each culminating in a ‘final eruption’, a more intense explosive event, and separated by short repose periods, up to seven years. Over one hundred lava flows of ‘a’ā and pāhoehoe types were emplaced during this eruptive cycle. Most of the lava flows and lava flow fields emplaced during this eruptive cycle have been covered by younger flows due to the morphology of the caldera. These lavas are predominantly phonotephritic in composition, with minor basanites and trachy-basalts (Fig. 9; Belkin *et al.*, 1993).

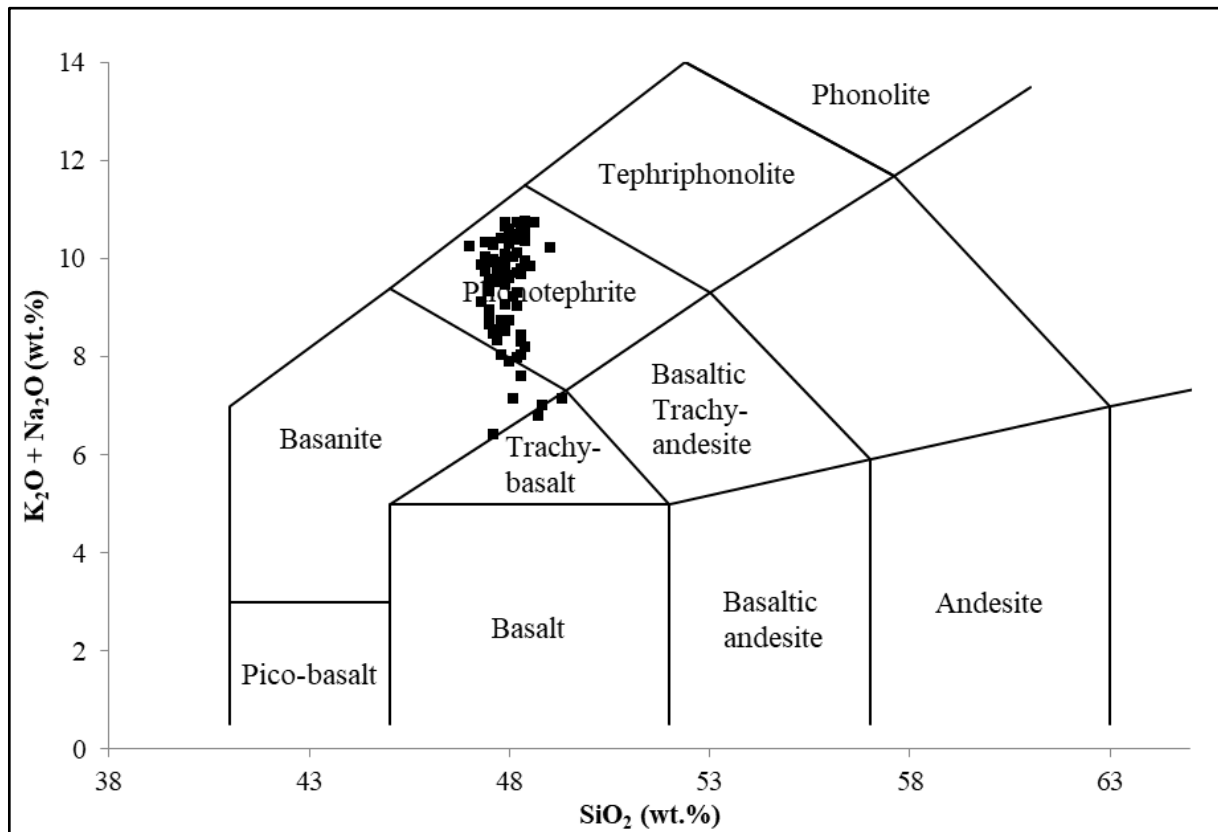


Figure 9. Representative analysis of AD 1631-1944 lavas from Mount Vesuvius. Data from Belkin *et al.* (1993)

This eruptive cycle concluded with an effusive-explosive eruption that occurred in March 1944 during World War II. The eruption started on 18 March 1944 with effusive activity until 21 March. Fountaining activity happened from late afternoon of 21 March to 22 March. Sustained explosive activity followed from 22 March to 23 March. The eruptive activity declined from 23 March with periodic Vulcanian eruptions and ended on 30 March 1944. The villages of San Sebastiano and Massa were destroyed by an ‘a’ā lava flow (Fig. 10; Chester *et*

al., 2007; Cubellis *et al.*, 2016). After the 1944 eruption, the volcano entered a phase of quiescence that continues to date.



Figure 10. Photograph of the 1944 'a'a lava flow destroying the town of San Sebastiano al Vesuvio on the 22nd of March. From Pesce and Rolandi (1994).

The proximity of Somma-Vesuvius to major cities (Naples, Rome) of southern Europe made it one of the most documented volcanoes in history. The famous description of the AD 79 Pompeii eruption by Pliny the Younger, who witnessed the eruption, stands as one of the oldest detailed accounts of a volcanic event. The 1631-1944 eruptive cycle, occurring during the post-Renaissance era, was particularly well-documented, as naturalists, aristocrats, and artists travelled through Europe to see, study, map, describe, or paint volcanoes and volcanic activity in southern Italy (Hamilton, 1772; Mercalli, 1884; Johnston-Lavis *et al.*, 1891; Gemmellaro, 1819; Maravigna, 1819; Gemmellaro, 1844; Mercalli, 1881, 1888, 1889; Pignonati, 1767; Guarini *et al.*, 1855; Phillips, 1869). Vesuvius became a central subject of the Grand Tour, inspiring countless paintings and works of art. In addition, the very first volcanological observatory, Osservatorio Vesuviano, was founded in 1841 under the Bourbon government at an altitude of 608 metres on the western slopes of Vesuvius. Equipped with both physical and

meteorological instruments, its operation started in 1848, leading to thorough descriptions of the following volcanic eruptions and landforms. The unparalleled amount of historical documentation about Vesuvius and its eruptive activity provides invaluable insights for reconstructing and understanding volcanic processes and interpreting the observed volcanic deposits.

1.2.2. Magma source and architecture of the plumbing system

Reconstructing the architecture of a volcanic plumbing system remains a complex challenge. At Somma-Vesuvius, decades of multidisciplinary research, combining geophysical and geochemical methods, have provided critical insights into magma storage, ascent dynamics, and crustal interactions.

A shallow magmatic reservoir (Fig. 11), located at depths of 3 to 5 kilometres, has been consistently identified through multiple geophysical techniques, including active seismic tomography (Zollo *et al.*, 1998; Auger *et al.*, 2001; Del Pezzo *et al.*, 2006), teleseismic studies (De Gori *et al.*, 2001), magnetotelluric surveys (Di Maio *et al.*, 1998; Patella and Mauriello, 1999; Pommier *et al.*, 2010), and integrated tomographies. These findings are corroborated by geochemical analyses of melt inclusions, which similarly indicate the presence of a shallow magma storage zone at a depth of 3 to 5 kilometres (Barberi *et al.*, 1981; Belkin *et al.*, 1985; Cortini *et al.*, 1985; Belkin and De Vivo, 1993; Fulignati *et al.*, 2000; Lima *et al.*, 2003). Additionally, shallower crystallisation has been documented up to 1 kilometre depth (Dolfi and Trigila, 1978; Marianelli *et al.*, 1999). This shallow magmatic reservoir between 3 and 5 kilometres depth is believed to have a small volume, with estimations significantly varying from 2-2.5 km³ (Barberi *et al.*, 1981) to 0.12-0.15 km³ (Morgan *et al.*, 2004).

Beyond the shallow reservoir, larger crustal reservoirs have been inferred at greater depths. Gravimetric (Berrino *et al.*, 1998) and seismic tomography studies (Zollo *et al.*, 1996; Auger *et al.*, 2001) initially proposed the existence of a deep crustal reservoir with a volume of 200 to 800 km³, located at 8 to 10 kilometres depth. Subsequent research has refined this model, identifying two crystallisation stages at depths of 8 to 12 kilometres (Pommier *et al.*, 2010) and 17 to 22 kilometres (Fig. 11; Dolfi and Trigila, 1978; Belkin *et al.*, 1985; Zollo *et al.*, 1996; Di Maio *et al.*, 1998; Marianelli *et al.*, 1999; Lima *et al.*, 2003). Moreover, magma storage is believed to extend down to the Moho discontinuity, detected at 30 kilometres depth (Corrado and Rapolla, 1981; Ferrucci *et al.*, 1989; De Natale *et al.*, 2001; Chiarabba *et al.*, 2005).

The magmas erupted at Somma-Vesuvius are derived from 5 to 10% partial melting of a heterogeneous MORB-type mantle source (Paone, 2006). This source is further affected by slab-contaminated mantle with fluids and sediments (Carminati *et al.*, 1998; Lucente *et al.*, 1999; De Gori *et al.*, 2001; De Natale *et al.*, 2001; Paone, 2004, 2006) and by crustal contamination from the Hercynian basement and carbonate deposits (Paone, 2006).

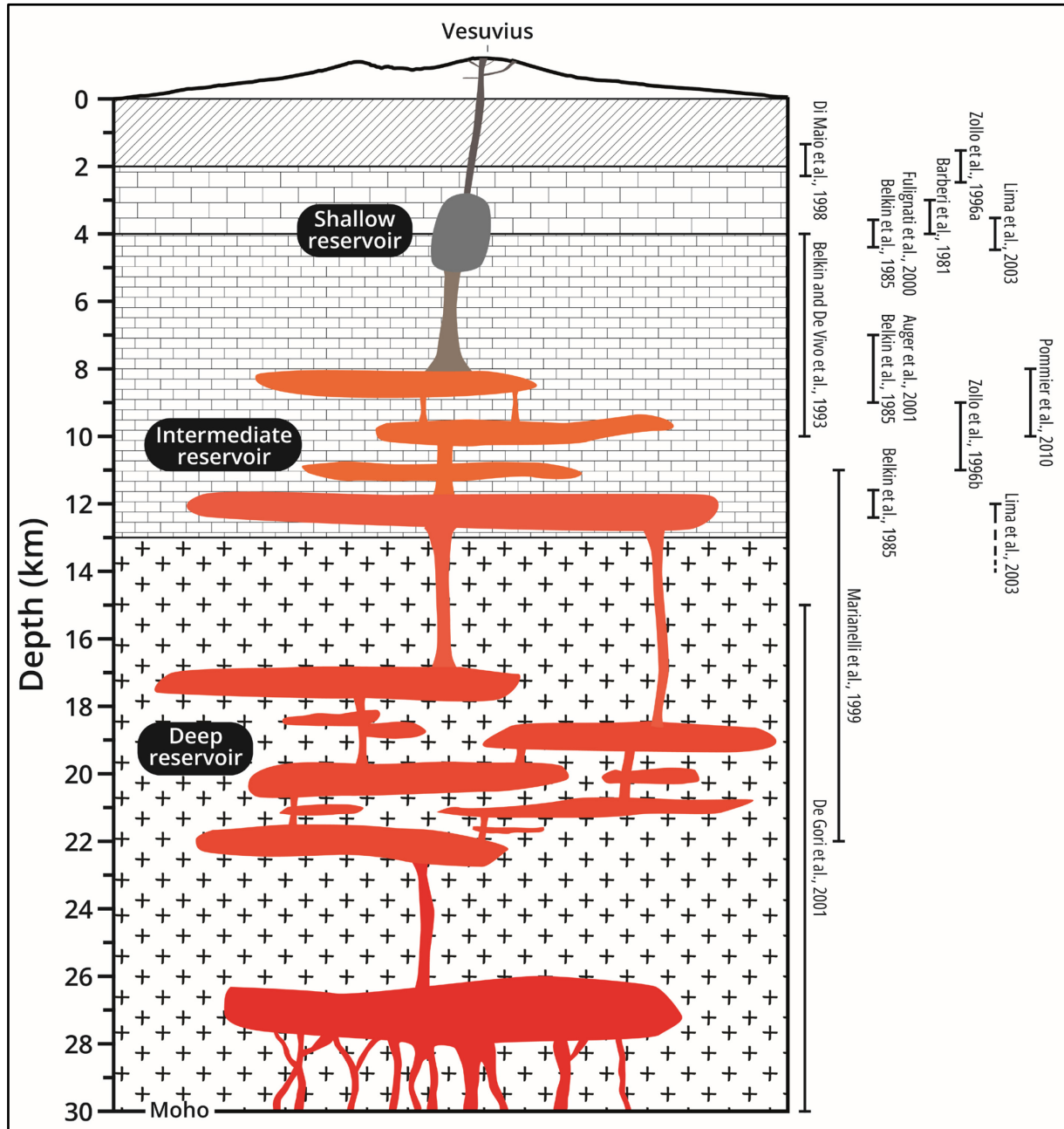


Figure 11. Representation of the plumbing system present under Somma-Vesuvius based on geophysical and geochemical studies.

1.3. Lava flow and emplacement mechanisms

Volcanoes are distributed around the world, both in oceanic and continental settings, and produce a wide range of deposits that extend from proximal to distal regions relative to the eruptive centre. The spatial distribution of these deposits is largely determined by the eruptive style. During effusive to mildly explosive eruptions, lava emerges from eruptive fissures and spreads onto the volcano's flanks, forming lava flows. The emplacement of lava flows is a gravity-driven process, strongly influenced by the underlying topography, specifically slope and roughness (Fig. 15a). Lava flows cover areas near volcanic centres but also most of the Earth's oceanic floor. Lava is a complex mixture of melt, crystals, and gases. Its composition varies from ultramafic to felsic. Erupted lava typically contains crystals that formed within the magmatic reservoir and during ascent through the eruptive conduit. These crystals, known as phenocrysts or microphenocrysts, can constitute less than 1 vol.% in some Kīlauea lavas to over 40 vol.% in lavas erupted at Vesuvius (Macdonald, 1949; Trigila and De Benedetti, 1993). Gas, a critical component, is released during magma ascent and lava emplacement. Bubbles nucleate on nanolite crystals, grow, deform, and coalesce, resulting in the formation of circular to elongated vesicles commonly observed in lava flows.

Macdonald (1953) classified lava flows into three primary types based on surface morphology and internal structure: (1) Pāhoehoe, *characterized by a smooth, billowy, or ropy surface, and spheroidal vesicles.* (Fig. 12a) (2) 'A'ā *'characterized by a fragmental and spinose surface, and irregularly shaped vesicles'* (Fig. 12b). And (3) Block lava *'differs from 'a'ā in greater regularity of the shapes of the fragments in the breccia phase, and less spinose surfaces* (Fig. 12c). *Beneath the fragmental tops of 'a'ā and block lava flows is a nearly continuous non-fragmental layer'.*



Figure 12. Main types of lava. a) Pāhoehoe flow at Kīlauea, Hawaii (USGS); b) 'a'ā flow at Kīlauea, Hawaii (USGS); and c) block lava at Lassen Volcanic Park (Frank Schulenburg, CC BY 4.0).

This classification was further refined by identifying sub-types based on surface morphology, vesicularity, type of breccia, clast size and morphology, and flow field structure (Fig. 13; Harris *et al.*, 2016).

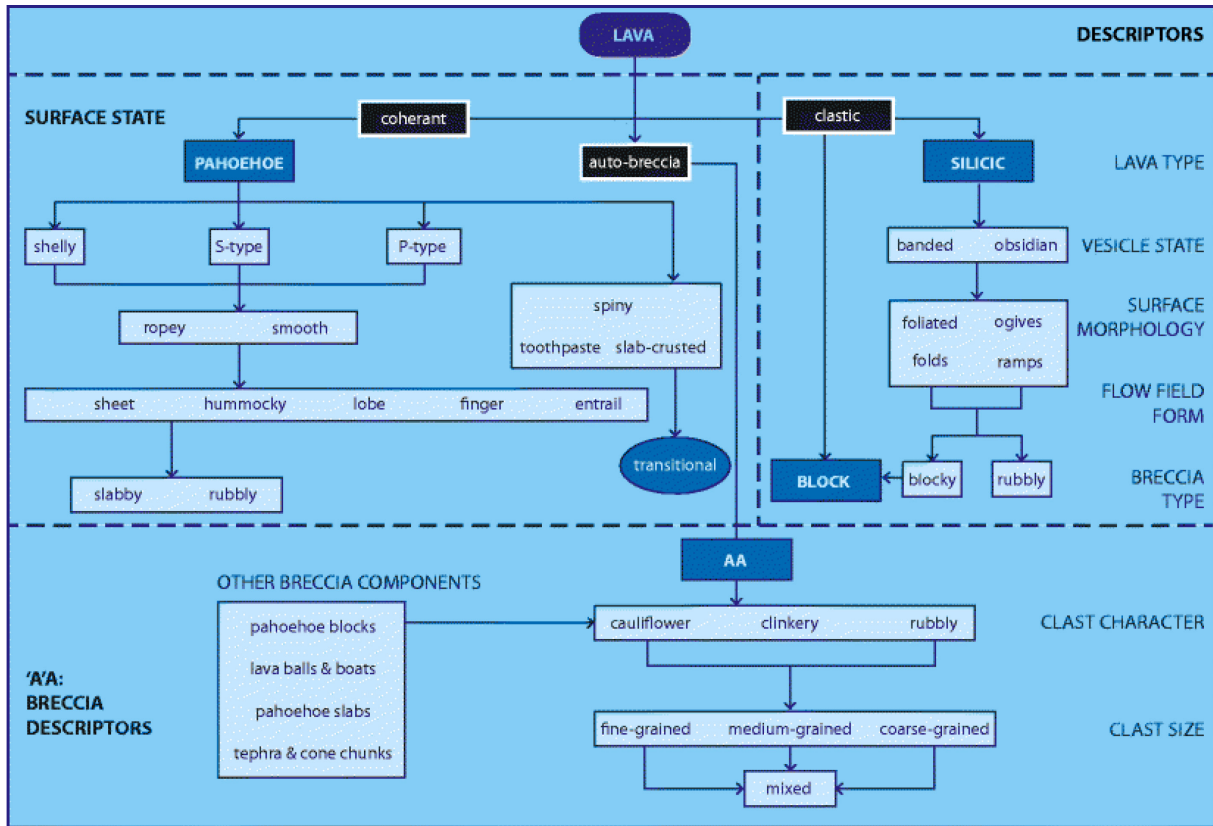


Figure 13. Classification of, and description terminology for, sub-types of pāhoehoe, 'ā, and block lava from Harris *et al.* (2016)

Pāhoehoe lava flows comprise hundreds to thousands of individual lava lobes, or 'toes', ranging from tens of centimetres to tens of metres in size. Each pāhoehoe lobe has a fluid core with a smooth, glassy, coherent surface that thickens over time. Lobe size is proportional to the lava supply rate. The core is typically well vesiculated. Pāhoehoe lava flows advance through 'toeing', when molten lava is supplied to an emplaced lobe with a thin cooled crust, the lobe inflates until its crust ruptures, breaking the base of the lobe, and forming a new lobe (Fig. 12a). On gentle slopes, pāhoehoe lobes have lateral velocities close to their advancing velocities, leading to lobe coalescence. This process can form a continuous liquid lava core with a single crust called sheet flow. Pāhoehoe sheet flows are typically a few tens of centimetres thick but tend to inflate to a thickness of a few metres.

'A'ā lava flows consist of a fluid, coherent core sandwiched between auto-brecciated basal and surface crusts. During emplacement, the core deforms and advances over the basal surface breccias that continuously fall from the surface crust to the base at the flow front in a conveyor belt motion (Fig. 12b). Solidified 'a'ā flows exhibit a solid core encased by two

breccia layers. ‘A‘ā breccias are sub-rounded dense clasts with rough surfaces formed due to shearing that tears apart the lava flow surface. ‘A‘ā flows are typically 1 to 20 metres thick.

Block lava flows resemble ‘a‘ā flows structurally, with a basal and surface breccia and a coherent core. However, block lava flows are significantly thicker than ‘a‘ā flows, being tens to hundreds of meters thick. Block lava flows are dense, crystal-rich and contain very few to no vesicles. Their surface morphology can be blocky, with angular clasts, or rubbly, with sub-rounded clasts, with sizes up to 5 to 10 metres (Fig. 12c).

The thermo-rheological structure of a pāhoehoe or ‘a‘ā lava flow comprises five distinct layers (Fig. 14; Hon *et al.*, 1994):

- The **surface crust**: a low-temperature zone cooled rapidly by radiation, with its base defined by the 800 °C isotherm.
- The **upper visco-elastic layer**: a partially molten zone between the 800 °C isotherm, at the top, and the 1070 °C isotherm, at the base. The layer thickness increases during emplacement.
- The **core**: a high-temperature, low viscosity zone. The core thickness decreases with time unless the flow is inflating.
- The **lower visco-elastic layer**: a second visco-elastic layer between the core and basal crust. Its thickness will increase more slowly due to the insulation provided by the overlying layer.
- The **basal crust**: a solid zone cooled rapidly at the contact with the underlying surface.

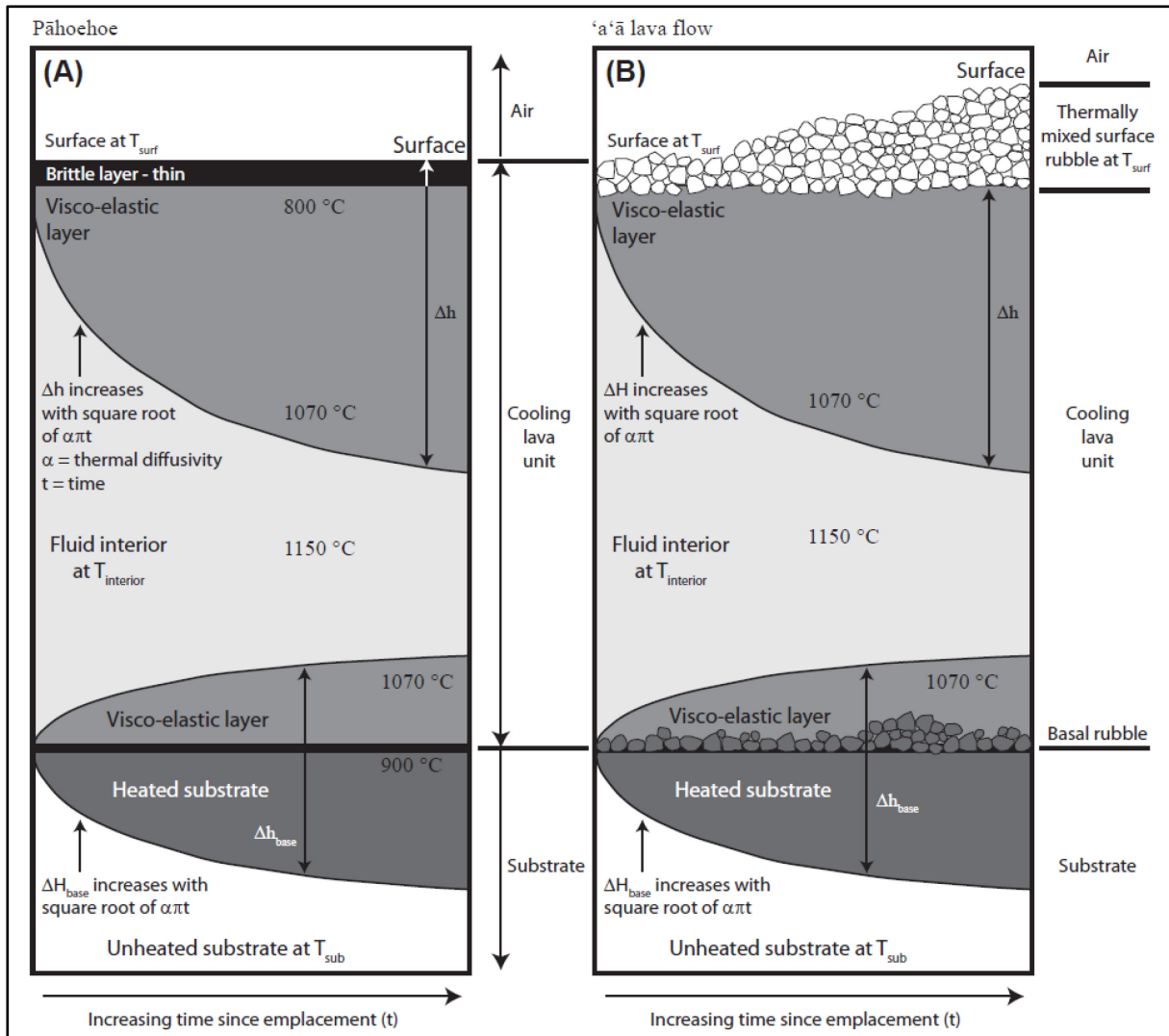


Figure 14. Thermo-rheological structure of (a) pāhoehoe lava flows and (b) 'a'ā lava flows. Figure from Harris and Rowland (2015).

The behaviour of lava flows during emplacement is governed by their rheological properties: viscosity, vesicularity, and crystallinity. Viscosity is primarily controlled by chemical composition, particularly the degree of polymerisation of the melt, which is controlled by network-forming oxides such as SiO_2 , TiO_2 , Al_2O_3 and Fe_2O_3 , which increase viscosity (Mysen and Virgo, 1994; Giordano *et al.*, 2008). Evolved lavas, rich in silica, tend to have higher viscosities. Silicate melts behave as Bingham fluids (Fig. 15b); thus, to be able to flow, a certain amount of shear stress, characterised by the yield strength of the fluid, needs to be reached. Basaltic melts typically have viscosities of $10^2 - 10^5$ Pa s when erupted (Walker, 1967; Shaw, 1969; Calvari *et al.*, 1994; Pinkerton and Norton, 1995; Harris *et al.*, 2005), while evolved melts can reach $10^7 - 10^{10}$ Pa s (Fink, 1980; Manley, 1996; Navarro-Ochoa *et al.*, 2002; Harris *et al.*, 2004). Lava rheology is also influenced by the amount and shape of crystals and vesicles present in the melt. Crystals increase apparent viscosity, whereas vesicles can either

increase or decrease it depending on their shape and connectivity (Sparks and Pinkerton, 1978; Harris *et al.*, 2019). Cooling during emplacement further enhances crystallisation and reduces vesicularity.

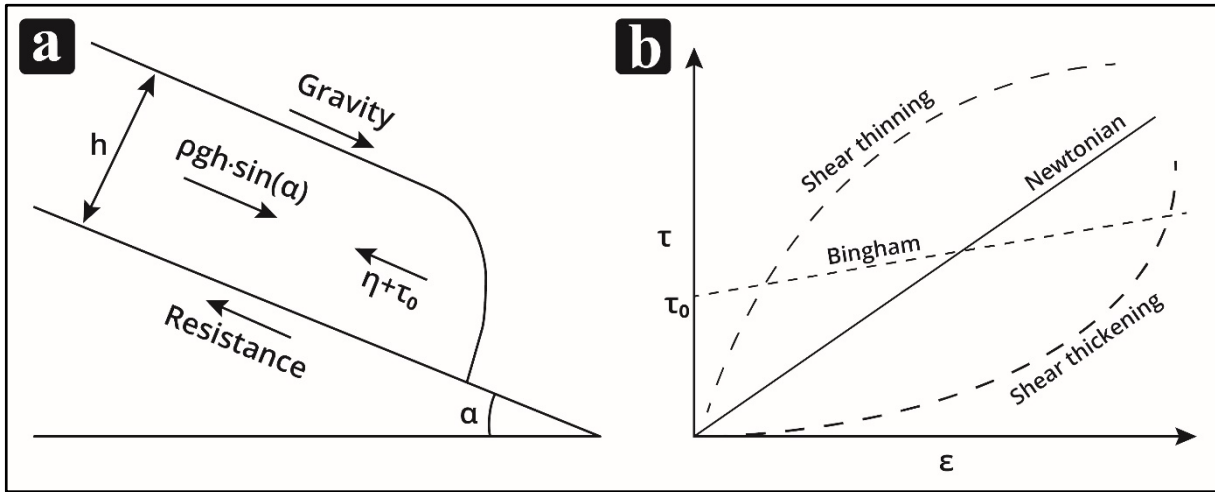


Figure 15. a) Emplacement parameters affecting lava flow emplacement. b) rheological behaviours of liquid

Lava flow emplacement is also controlled by eruptive characteristics, such as eruptive rate and eruption duration, pre-eruptive topography (slope and roughness) and cooling conditions (Walker *et al.*, 1973; Pinkerton and Wilson, 1994). These factors collectively determine the final shape and size of lava flows (Walker *et al.*, 1973; Malin, 1980; Pieri and Baloga, 1986; Pinkerton and Wilson, 1994; Gregg and Fink, 2000). Lava flows and lava flow fields typically extend from hundreds of meters to tens of kilometres on the Earth's surface.

Lava is transported from the vent to the flow front via lava channels or lava tubes. Lava channels are lava streams confined by levees, which are stationary banks of solidified lava formed by rapid margin cooling and higher drag. Levees start to form early during flow emplacement and are present throughout most of the flow length. Levees are classified into four types (Sparks *et al.*, 1976):

- **Initial levees** are the first to form. They determine the width of the channel, but it can be modified by a change of slope or effusive rate. Other types of levees are usually built on initial levees.
- **Accretionary levees** develop near the eruptive vent through the accumulation and welding of ductile material on the top and sides of the levees due to the stable and slow rise of the lava level.
- **Rubble levees** consist of an accumulation of unwelded 'a'ā clinkers formed in well-developed areas of an 'a'ā lava flow.

- **Overflow levees** result from repeated draping of overflowing lava on the existing levees, widening and heightening the levees. Overflow levees are thought to occur due to fluctuating effusion rates or due to rising lava level by back pressure.

Lava tubes are ‘roofed conduits within which molten lava flows’ (Fig. 16; Kauahikaua *et al.*, 1998) or ‘conduits beneath the surface of solidified lava through which molten lava flows’ (Greeley, 1987). The mechanisms of formation and the structures that formed within lava tubes are detailed in the next section (Section 1.4).

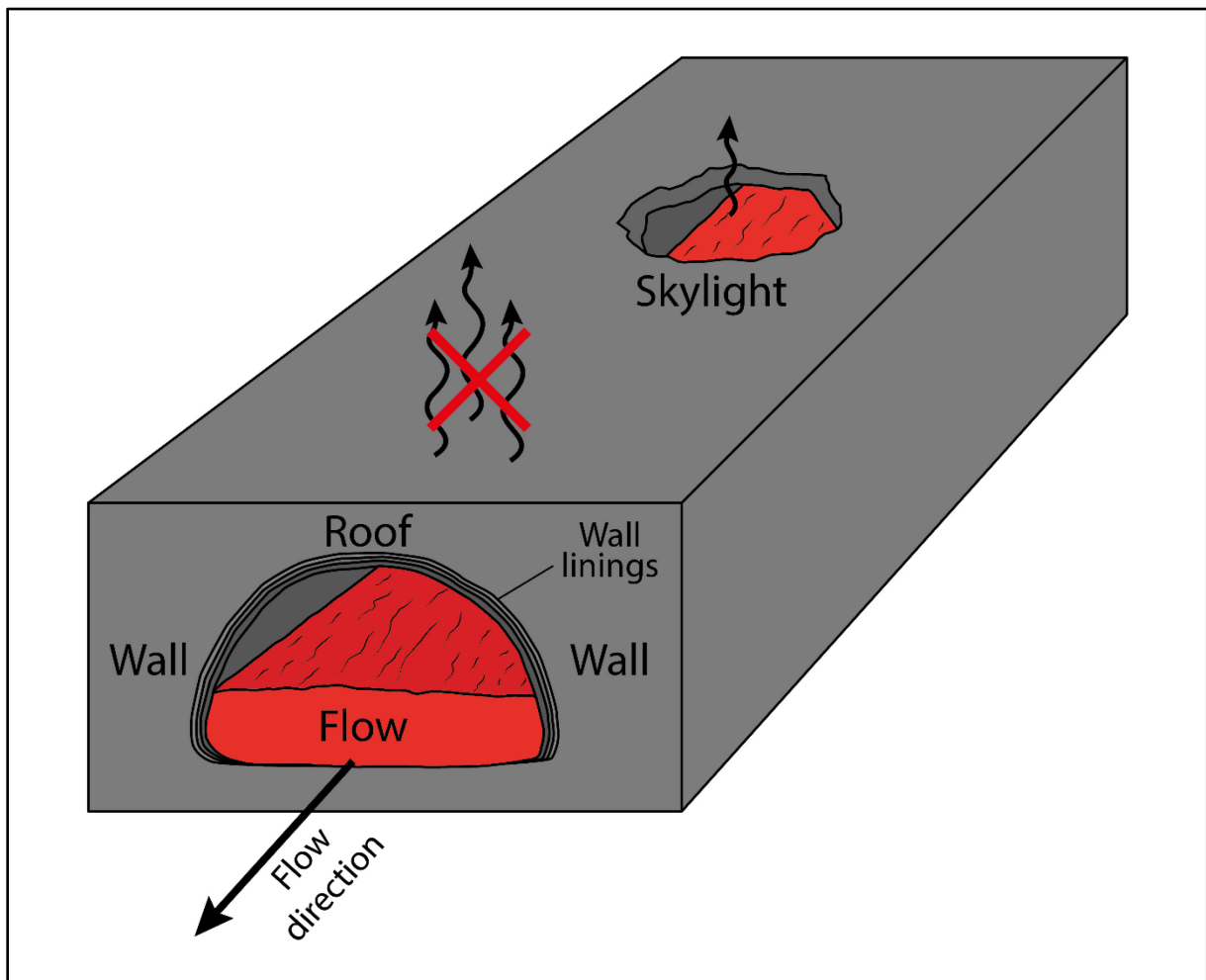


Figure 16. Schematic of a lava tube. The presence of a roof made of solidified lava inhibits atmospheric cooling, highly decreasing cooling rates. Each passage of a flow inside a lava tube leaves a layer of lava on the walls and ceiling.

Inflation processes during flow emplacement can generate different structures at the surface of a lava flow:

Tumuli are ‘positive topographic features that form by a mechanism of crustal uplift on lava flow fields’ (Duncan *et al.*, 2004). Tumuli can range from a metre to tens of metres high. They are believed to be associated with underlying lava tubes (Calvari *et al.*, 1994; Calvari and Pinkerton, 1998). Three types of tumulus have been described by Duncan *et al.* (2004): (1)

focal, formed over the main eruptive vents, (2) satellite, produced by important inflation of lava flows, and (3) distributary, formed by up-arching of the crust over a tube with only limited and short-lived lava breakouts.

Lava rises and **lava rise pits** are uplifted lava plateaus bordered by a monoclinical convex outward-sloping scarp. They are usually a few metres high and tens of metres wide. The surface of lava rises has a smooth pāhoehoe texture that sometimes presents ropy structures. Shallow pit-like structures can occur on some of the lava rises. These pits can have broad flat bottoms resulting from surface collapse after lava drained below the crust. Lava rises when being sufficiently inflated can develop a medial crack, becoming analogous to satellite tumuli.

Lava domes are semi-spherical positive topographic features formed by the accumulation of lava. They can reach tens of metres in height and up to one hundred metres in width.

1.4. Lava tube formation, development, and internal features

The length of a lava flow is significantly enhanced by the formation of lava transport structures, particularly lava tubes, which are the most efficient structures in transporting lava away from the eruptive vent. These structures are predominantly observed in basaltic volcanic systems, such as those in Hawaii (USA) and Mount Etna (Italy). The presence of a static roof, formed by the solidification of surface lava, drastically reduces atmospheric cooling of the molten lava flowing beneath. As a result, heat transfer in a lava tube is limited to conduction through the walls and ceiling, leading to an exceptionally low cooling rate, approximately 1 °C per kilometre (Keszthelyi, 1995; Sakimoto and Zuber, 1998; Witter and Harris, 2007). This thermal efficiency contrasts with the cooling rates in open channels, which range from 5 to 7 °C per kilometre (Crisp *et al.*, 1994; Cashman *et al.*, 1999; Soule *et al.*, 2004; Riker *et al.*, 2009; Robert *et al.*, 2014; Rhéty *et al.*, 2017). Consequently, lava tubes enable lava to travel far greater distances from the eruptive fissure compared to open channels. Lava tubes exhibit a range of lengths from tens of meters (Calvari *et al.*, 2024; Lemaire *et al.*, 2024) to tens of kilometres, such as the Kazumura lava tube in Hawaii, which extends over 57 kilometres (Allred and Allred, 1997) or the Undara tube system in Australia, which extends over 100 kilometres (Ollier and Brown, 1965).

Three primary mechanisms for lava tube formation have been described in the literature (Greeley, 1971a, 1971c, 1972, 1987; Peterson *et al.*, 1994; Kauahikaua *et al.*, 1998; Calvari and Pinkerton, 1998, 1999):

(1) by **roofing over a lava channel**. This mechanism is subdivided into three processes, depending on the rheology of the lava and on the eruptive rate (Greeley, 1987; Peterson *et al.*, 1994):

- Roofing over a channel: when the effusion rate is stable over time, it allows cooling of the lava surface, forming a fine crust that extends from the channel levees across the lava surface, joining in the middle of the channel and gradually thickening to form a solid roof (Fig. 17a). Fluctuations in lava flow level can disrupt this crust, carrying fragments downstream.
- Floating slabs accumulation: In constricted or sinuous channels, crustal pieces aggregate and merge to form a continuous roof (Fig. 17b).
- Accretion to levees: variations in the eruptive rate cause lava overflows over the channel levees. Repeated overflows build up the levees inward toward the centre of the channel, eventually merging to form a roof after a sufficient period of time (Fig. 17c).

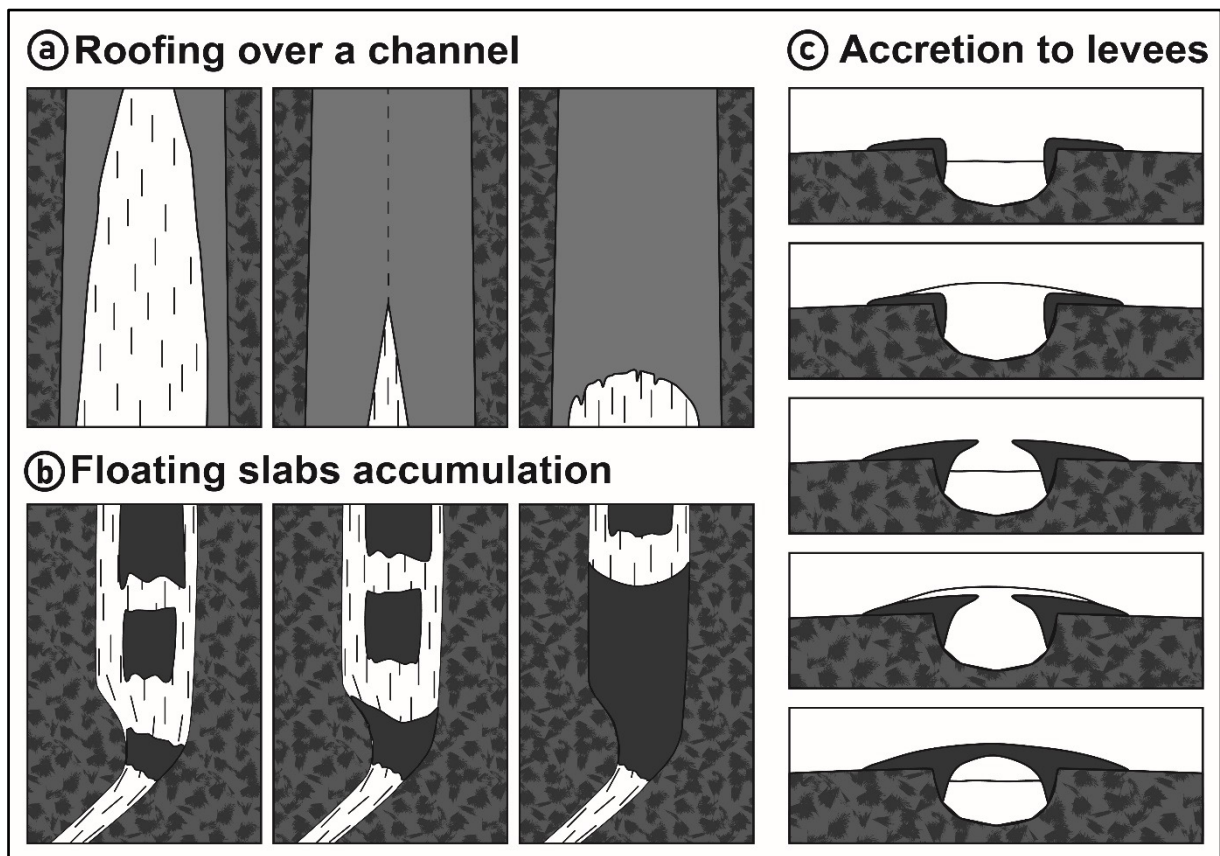


Figure 17. Sub-types of lava tube formation by roofing over a lava channel. a) Roofing over a lava channel; b) floating slabs accumulation; c) accretion to levees.

(2) by **pāhoehoe lobe extension**. Lava tubes can also form through the alignment and coalescence of pāhoehoe lobes (Fig. 18a). As lobes are emplaced in the flow direction, they create a primary lava pathway that may evolve into a lava tube (Wentworth and Macdonald, 1953).

(3) by **lava flow inflation** (Fig. 18b). Inflation occurs when the lava flow front cools rapidly, forcing the flow to thicken vertically. In ‘a‘ā lava flows, this process is driven by rapid cooling at the flow front (Calvari and Pinkerton, 1998). In pāhoehoe lava flows, inflation can result from insufficient slope to sustain flow, leading to the formation of sheet flows that thicken over time (Hon *et al.*, 1994; Lemaire *et al.*, 2024).

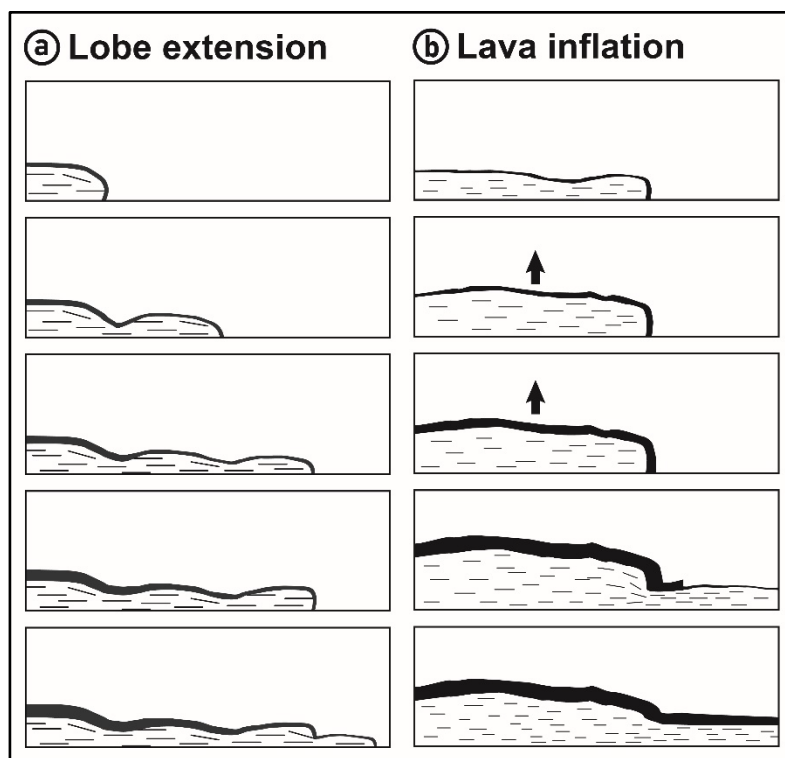


Figure 18. Lava tube formation mechanisms by (a) lobe extension and (b) lava inflation.

Lava tubes predominantly form on low slopes ($<10^\circ$; Keszthelyi, 1995; Kauahikaua *et al.*, 1998), although their formation has been observed on slopes up to 30° (Calvari and Pinkerton, 1998).

Lava tubes are classified into two main morphological types: simple and compound (Fig. 19). Simple lava tubes consist of a single conduit. Their cross-sectional shape depends on the formation mechanism. When formed by roofing over a lava channel, a simple lava tube has a ‘triangular’ section (Fig. 19a), while when formed by inflation of a lava flow, a simple tube has a more rounded section (Fig. 19b). Compound lava tubes comprise two or more tubes that coalesce. Both types can be symmetric or asymmetric. For simple lava tubes, the degree of

asymmetry depends on the flow direction. A straight lava tube will be symmetric, while when a tube changes direction, it develops an asymmetry (Fig. 19c and d). This asymmetry is either due to a more important deposition of lava on the slower-moving inner side of a turn (Fig. 19c) or due to a difference of charge supported by the tube roof (Fig. 19d) from accumulation of lava flows on one side, opening of ephemeral vents on the walls of the tube, or secondary collapse events. For compound lava tubes, when tube coalescence is vertical, they have symmetric transverse sections (Fig. 19e), while when the overlying lava tube is not straight above the lower tube during coalescence, their sections are asymmetric (Fig. 19f). Lava tubes can also be single or braided. Single lava tubes have a unique conduit, while braided ones have lateral branching of at least two conduits.

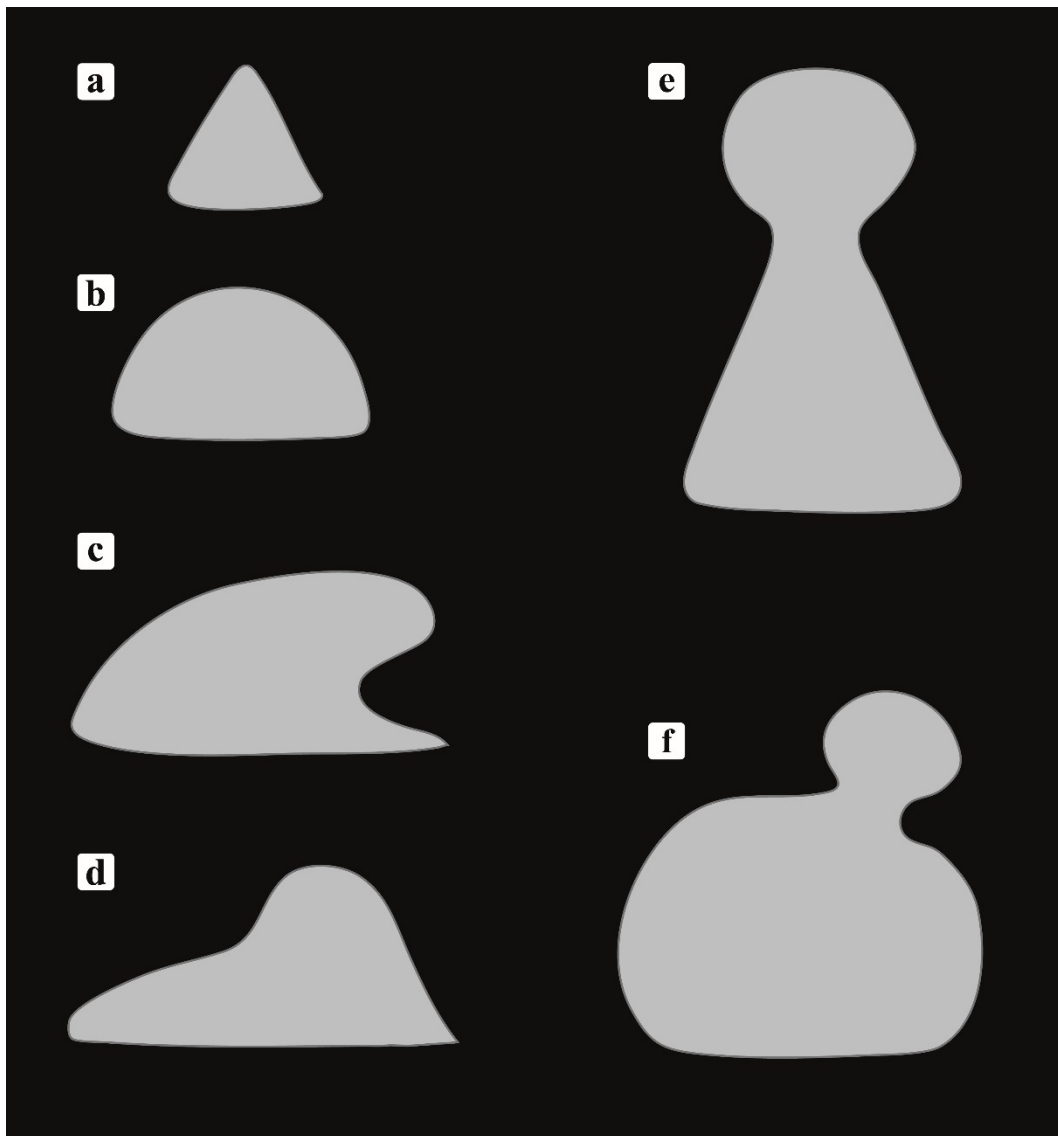


Figure 19. Lava tubes morphologies. Simple (a-d) and compound (e-f). transverse sections of lava tubes. (a-b) Simple and symmetric sections; (c) simple, asymmetric section due to bending of the tube; (d) simple asymmetric section due to roof loading; (e) symmetric, compound tube section with typical key-shape, indicating coaxial capture; (f) asymmetric compound tube section. Figure modified from Calvari and Pinkerton (1999).

Lava tubes host a variety of internal features, developed during or after their activity, that reflect their thermo-rheological and lava transport processes taking place inside (Larson, 1991; Calvari and Pinkerton, 1999):

- **Lava stalactites** are solidified masses of lava that hang from a ceiling or overhanging surface. Lava stalactites form with plastic lava, while some harden unchanged, others can be distorted externally by gas currents or internally by vesiculation before solidifying. Solidified stalactites can be coated by later flows. Many types of lava stalactites have been described based on their shapes:
 - *Botryoid* lava stalactites are a cluster of short, branching lava helictites that resemble an aggregate of grapes.
 - *Bulbous* lava stalactites are significantly thicker at the bottom than at the top.
 - *Dip-layered* stalactites are composed of highly vesicular, concentric layers due to repeated coating by fluid lava.
 - *Elephant's foot* stalactites have their growth terminated on an obstruction later removed.
 - *Pipe stem* stalactites are partially deformed tubular lava stalactites that have an oval or pinched transverse cross-section.
 - *Shark tooth* stalactites are blade-like stalactites usually as long as their width. They occur as individuals or can be joined along a vertical centreline. Shark tooth stalactites are typically smooth and thinly glazed externally, uniformly vesiculated internally.
 - *Soda straw* stalactites are predominantly straight tubular stalactites.
 - *Spatter* stalactites are lumpy, irregular, highly vesiculated, concentrically layered lava stalactites accumulated from spatter. Common inside spatter cones.
 - *Teat* stalactites are shaped like an inverted cone, having a generally circular transverse cross-section and a drip-sized tip.
 - *Tubular* lava stalactites are slightly and uniformly tapered. Their diameter averages about 0.7 centimetres, and usually slightly decreases toward the tip. Tubular stalactites range from millimetres to over a meter. The tip can be hemispherical or open, but the interior is usually a succession of elongated vesicles and thin partitions. Tubular stalactites are usually marked with shallow annular grooves caused by the incremental growth.

- **Lava stalagmites** are vertically oriented accretions of lava droplets. They are associated with a ceiling or wall structure which drips (a low point, stalactite, etc.), but usually the accretion surface moves or is inundated, causing disparities between stalactites and stalagmites.
- **Lava helictites** are cylindrical, even tubular extrusions unaffected by gravity. Lava helictites form on linings, other helictites or stalactites due to gas pressure. They range in diameter from five millimetres up to a centimetre.
- **Wall linings** are layers of hardened lava on the walls and ceiling of a lava tube. They form by deposition of a layer of lava during successive passage of flows inside, either partially or completely filling the lava tube. Each layer corresponds to a cycle of infilling and drainage of the lava tube, thus preserving a record of the tube's activity. The thickness of a wall lining ranges from a few millimetres to several metres and is not necessarily uniform. The thickness depends on the duration and rheology of the lava flow. Wall linings are fragile and tend to collapse from the ceiling or detach from the walls. In certain conditions, wall linings can detach from the wall and roll down on themselves, forming rolls on the sides of the lava tube. This happens when a lava tube is drained rapidly while the lining is still plastic, contrarily to the outer linings that are already rigid and stable.
- **Lateral benches** are banks along the sides of a lava tube formed by lateral solidification of lava on the walls due to a maintained stable lava level inside a lava tube. When multiple lateral benches are observed, it indicates several stages of magma supply rates.
- **Shelves** are elongated, overhanging crusts, remnants of crusts, or accumulations of linings attached to a wall.
- **Pillars** are bodies of lava linking the floor to the ceiling of a tube that divides it for a short distance.
- **Cracks** can be longitudinal or transverse. Longitudinal cracks in the roof of lava tubes are due to deformation caused by the weight of the lava tube roof or by both the weight of the roof and an overlying flow. Cracks can develop in the middle of the roof or at the sides. High deformation can produce whale-back structures. Intense deformation of a longitudinal crack can produce pillars.
- **Grooves** are formed when cooled lava at the surface of a flow is transported and mark the plastic walls of the lava tube. Grooves can be horizontal, inclined or vertical, and record flow direction. Vertical grooves indicate a sudden drop of the lava level during which solid lava at the margins marked the plastic wall lining.

- **Gas blisters** are remnants of gas bubbles that were accumulated between wall linings. When gas blisters form, it suggests that previous wall linings are still degassing. While most of the time, gas blisters are observed as unoxidized rounded areas on the wall linings, some of them indicate that the gas bubble was still surrounded by molten lava when the lining detached, leaving thin glassy lava.
- **Pāhoehoe squeeze-outs** are injected lava volumes in between rubbles, welding the clasts together.
- **Lava falls** are precipitous drops in the floor of a lava tube over which lava flowed.

1.5. Lava tubes on other planetary bodies

Since the 1970s, structures resembling terrestrial lava tubes have been identified on other planetary bodies (Greeley, 1971b; Hurwitz *et al.*, 2013). On the surface of the Moon, elongated structures are observed. Some of these structures are linear and correspond to large-scale faults, while others have a sinuous shape, called sinuous rilles, and are interpreted as collapsed lava tubes. On the surface of Mars, aligned pit craters are observed, known as sinuous collapse chains are also hypothesised to be lava tubes. These structures are distinguished from impact craters by the absence of elevated crater rims or ejecta deposits, suggesting their formation through the collapse of subsurface cavities (Leone, 2014; Zhao *et al.*, 2017; Sauro *et al.*, 2020). Many of these collapses and chains of collapses occur in lava flows or within volcanic centres, further supporting their volcanic origin (Fig. 20a, b). The presence of lava tubes on both Mars and on the Moon is confirmed by the observation of skylights, openings in the roof of lava tubes that reveal subsurface voids (Fig. 20c; Haruyama *et al.*, 2012; Robinson *et al.*, 2012; Chappaz *et al.*, 2017). However, the precise morphology of these extraterrestrial lava tubes remains poorly characterised due to the lack of in-situ exploration missions, limiting current analysis to satellite image analysis and surface observations. Estimates indicate that Martian and lunar lava tubes may be 80 times and 300-700 times more voluminous, respectively, than their terrestrial counterparts (Sauro *et al.*, 2020). This size discrepancy is attributed to lower gravitational acceleration affecting the planetary bodies and higher flow velocities (Murase and McBirney, 1970). Furthermore, extraterrestrial lava tubes seem to extend over longer distances than terrestrial lava tubes, ranging from tens of kilometres to hundreds of kilometres (Hurwitz *et al.*, 2013; Chappaz *et al.*, 2017; Sauro *et al.*, 2020).

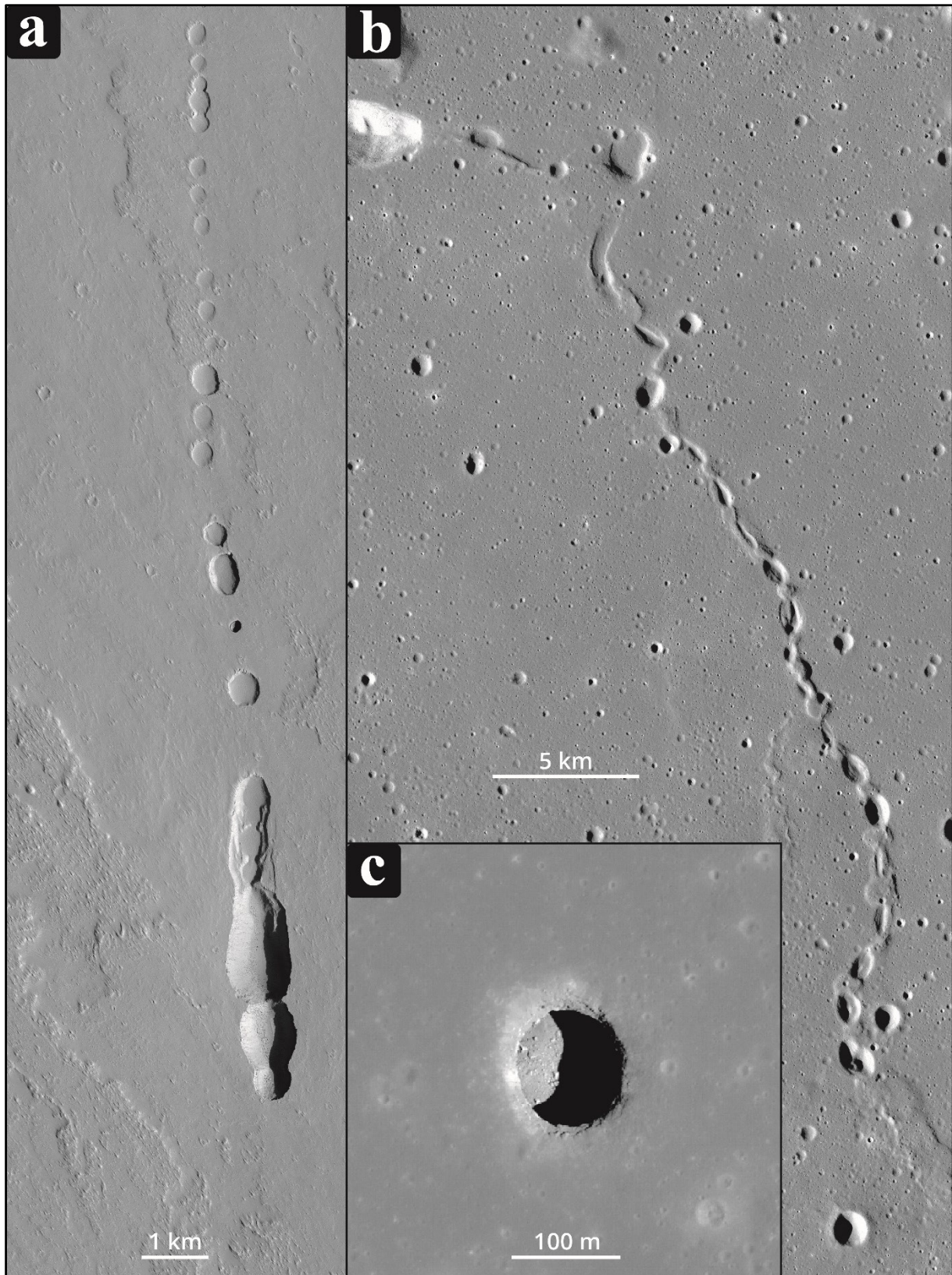


Figure 20. Examples of skylight, collapse chains and sinuous rilles on Mars and the Moon. a) Pit crater chain in the southern area of Arsia Mons on Mars captured by the HiRISE on MRO (NASA/JPL-Caltech/UA Arizona). b) Sinuous rille near Gruithuisen crater on the Moon captured by the LROC (NASA/GFSC/ASU). c) Mare Tranquillitatis pit crater on the Moon in a high-sun view captured by LROC. The image of the sub-circular skylight reveals collapsed rocks on a smooth interior floor (NASA/Goddard/ASU).

The discovery of lava tubes on other planetary bodies holds significant promises for future space exploration. These structures could serve as long-term natural shelters to host human settlements, offering protection from galactic cosmic rays, solar particle events, and meteorite impacts, particularly critical on the Moon, which lacks a protective atmosphere (Hörz, 1985; Angelis *et al.*, 2002; Haruyama *et al.*, 2012; Blair *et al.*, 2017). Additionally, lava tubes are thermally stable environments, with internal temperatures hovering around -20 °C, compared to extreme surface fluctuations of 300 °C (Hörz, 1985; Coombs and Hawke, 1992). Given their larger dimensions relative to terrestrial lava tubes, lunar and Martian lava tubes could accommodate permanent bases, provided their structural integrity is confirmed (Theinat *et al.*, 2020). Utilising these natural structures would reduce the need for energy and construction materials, while also offering direct access to subsurface resources, such as volatiles or water (Blamont, 2014; Williams *et al.*, 2010). Several projects are exploring the design of habitats within lunar lava tubes or the adaptation of these structures into safe, habitable environments (Fig. 21; Daga *et al.*, 1992; Blamont, 2014; Grandl, 2017; Mari *et al.*, 2021; Martin and Benaroya, 2023; Zhang *et al.*, 2024). Beyond their potential for human habitation, lava tubes may also advance the search for extraterrestrial life, as they could have preserved evidence of past and present life on Mars (Boston *et al.*, 2001; Lévêillé and Datta, 2010; Michalski *et al.*, 2018).

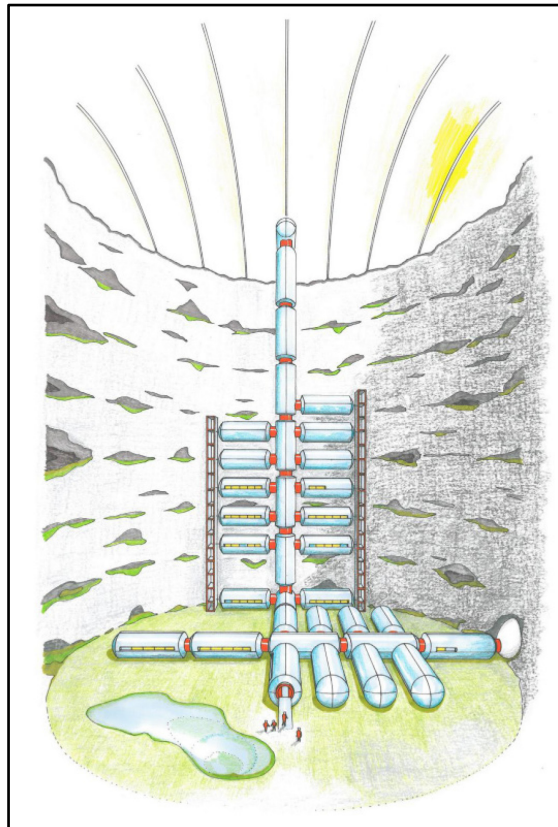


Figure 21. Conceptual lunar habitat for 100 inhabitants inside Mare Tranquillitatis Hole (From Grandl, 2017)

CHAPTER 2

Methodologies and techniques

This chapter is dedicated to describing all the methodologies used during the Ph.D. to characterise the studied lava flows and their associated lava tubes. The first section details and explains the working principles of the analytical techniques applied to the petrological, chemical, and textural characterisation of the lava samples. The second section relates the models used to characterise the thermo-redox conditions. The third section shows the different techniques used to numerically reconstruct the morphology of the studied lava tubes and explains the workflow from the data acquisition to the final 3D model. The fourth section explains the method using Acoustic Emission testing to assess the stability of the tube's structure and the novel processing method of the signals. The last section describes the applied methodology to design a geotinerary and a virtual tour for geoheritage and conservation purposes.

2.1. Petrological, chemical, and textural analyses

2.1.1. Sampling methodology

Several sampling campaigns were performed at Vesuvius between November 2023 and November 2024 to fully investigate the 1858 lava flow field and the largest lava tube rediscovered within the flow field. Samples were collected on the surface along the 1858 lava flow field, close to the eruptive vent, and close to the flow front. The presence of an abandoned quarry close to the flow front allowed us to collect samples from the base and core of the 1858 lava flow field. To characterise the studied lava tube, samples were collected on its surface and in its interior. Inside the lava tube, samples were collected within the walls to get a sequence of wall linings either by breaking or drilling.

2.1.2. Sample preparation

The collected samples were cut into smaller pieces. Fresh and unaltered pieces were used to prepare thin sections (Section 2.1.3) and epoxy mounts for Electron Microprobe Analysis (EMPA; Section 2.1.6). The thin sections were used for optical analysis with an optical microscope (Section 2.1.3) and Scanning Electron Microscope (SEM) imaging (Section 2.1.5). Part of each sample was crushed and pulverised into a fine powder to prepare pressed powder pellets for X-ray fluorescence (XRF) analysis (Section 2.1.4). Samples were also prepared for synchrotron X-ray microtomography (Section 2.1.7) as rectangular cuboids that have a base of 4 by 4 millimetres and an unconstrained height (commonly around 10 millimetres).

2.1.3. Thin sections

Thin sections of the collected samples were made to optically investigate the texture of the sampled lavas and characterise the mineral phases present in the assemblage. Thin sections (28 × 48 mm) were realised with a sample thickness of 30 µm. An OPTIKA B-600 POL polarising microscope was used for the characterisation of the samples, alongside a Zeiss Axiocam 105 color optical camera with a resolution of 5 MP to capture images of the thin sections.

2.1.4. X-ray fluorescence (XRF)

X-ray fluorescence was used to measure the bulk rock compositions of major and trace elements of the collected samples. The samples were ground into powders with a grain size less than 75 µm using a Retsch PM-100 planetary mill and agate grinding balls. To prepare pressed powder pellets, we used 4 grams of lava powder mixed with a millimetre of Polyvinyl alcohol for each sample. The mixtures were dried for three hours at 90 °C and reduced to powder again.

Powders were placed in aluminium capsules onto which we applied a pressure of 20 tons for 20 seconds.

X-ray fluorescence analyses are performed by bombarding the pressed powder pellets with high-energy X-rays. The samples react by emitting secondary (or fluorescent) X-rays with characteristic energies depending on the elements that compose the sample. The wavelengths of the secondary X-ray are diffracted on a monochromator and detected by a gas counter. XRF analyses were performed at the Department of Earth, Resource and Environment Sciences (DiSTAR) of the University of Naples, Italy, using an Axios Panalytical sequential wavelength dispersion X-ray fluorescence spectrometer (WDS) equipped with six analyser crystals, three primary collimators, and two detectors (flow counter and scintillator), operating at different kV and mA settings for each analysed element. Analytical uncertainties are in the order of 1–2% for major elements and 5-10% for trace elements.

2.1.5. Scanning Electron Microscope (SEM)

A Scanning Electron Microscope works by bombarding the sample with electrons while scanning the surface of the sample. The surface of the sample emits particles or rays. The detectors can get secondary or backscattered electrons, for imagery, or X photons, for chemical analysis. During this Ph.D., the Scanning Electron Microscope was used to observe at high magnification the collected samples. Images obtained from backscattered electrons (BSE) show the chemical contrasts present within the analysed sample. The backscattered electrons are the incident electrons that interacted with the atoms of the sample and were then reemitted towards the detector. The energy of backscattered electrons depends on the elements encountered in the ‘deep’ zone they interacted with. Elements with a higher atomic number reemit the incident electrons more strongly than elements with a lower atomic number, meaning that the heavier an element is, the brighter the image will be.

SEM analyses were performed at the Department of Earth, Resource and Environment Sciences (DiSTAR) of the University of Naples, Italy, with a Zeiss Merlin VP Compact microscope. Analyses were made on carbon-coated thin sections and epoxy mounts. Analyses were made with a 15 kV tension.

2.1.6. Electron microprobe analysis (EMPA)

The microprobe is a non-destructive in-situ analytical instrument that was used here to measure major and minor element concentrations of crystals within the samples. Electron microprobe analysis is based on wavelength dispersion spectrometry (WDS) of X-rays emitted

by the sample. The polished surface of samples, placed in a void chamber, is bombarded by a beam of electrons. The bombardment excites atoms in the selected area. The atoms emit X-rays when returning to their normal energy level. The emitted X-rays have wavelengths that are characteristic of the elements forming the sample and intensities that depend on element concentration. The emitted X-rays are selected depending on their wavelength. A monochromatic crystal with a variable orientation is used to diffract the wavelengths. The intensity of the wavelengths is detected by a gas counter. A quantitative value of each element concentration can be obtained by comparing the obtained intensities with those of a standard. It allows precise measurements of glass and crystal compositions.

The analyses were performed at the Laboratoire Magmas et Volcans (LMV) of the Université Clermont Auvergne (UCA), Clermont-Ferrand, France, with a Cameca SXFiveTactis microprobe equipped with a LaB6 source. The microprobe was calibrated using natural and synthetic standards (Table 2).

Table 2. Standards used to calibrate the microprobe.

Element	Mineral	Standard	Standard origin
Si, Ca	Cpx	Wollastonite	Natural
Fe	Ol, Opx, Cpx, Gt	Fayalite	Synthetic
Mg	Ol, Opx, Cpx, Gt	Forsterite	Synthetic
Cr	Ol, Opx, Cpx, Gt	Cr ₂ O ₃	Synthetic
Al	Ol, Opx, Cpx	Orthoclase	Natural
Mn, Ti	Ol, Opx, Cpx, Gt	MnTiO ₃	Synthetic
Ni	Ol, Opx, Cpx, Gt	NiO	Synthetic
K	Ol, Opx, Cpx, Gt	Orthoclase	Natural
Na	Ol, Opx, Cpx, Gt	Albite	Natural

The analysed samples were epoxy-mounted lava pieces with a polished surface and carbon-coated. To measure crystal compositions, a focused beam with a 15 kV acceleration voltage and a 15 nA beam current was set as the operating condition. Ten elements (Si, Ti, Al, Cr, Fe, Mn, Mg, Ca, Na, K) were chosen to be analysed. For each sample, at least ten analyses

of each phase (Leucite, Plagioclase, Clinopyroxene and Fe-Ti oxides) were performed. Analyses are given as oxides in wt.%. Analyses with a total sum of oxides below 96 wt.% or above 104 wt.% were not considered. The error is expressed as the standard deviation at 3σ of an element i :

$$SD_{3\sigma} = 3 \left(\frac{C_s}{\lambda_s} \right) \cdot \sqrt{\sigma_{\lambda_s}^2 + \left(\frac{\lambda_s}{\lambda_{sd}} \right)^2 \sigma_{\lambda_{sd}}^2} \quad (1)$$

with:

$$\lambda_s = \frac{\sum_1^{i_{max}} \left(\frac{P_i}{t_{P,i}} - \frac{B_i}{t_{B,i}} \right) / I_j}{i_{max}} \quad (2)$$

$$\sigma_{\lambda_s}^2 = \sqrt{\frac{\sum_1^{i_{max}} (P_i/t_{P,i}/I_j)/i_{max}}{\sum_1^{i_{max}} t_{P,i} \cdot I_i} + \frac{\sum_1^{i_{max}} (B_i/t_{B,i}/I_j)/i_{max}}{\sum_1^{i_{max}} t_{B,i} \cdot I_i}} \quad (3)$$

$$\sigma_{\lambda_{sd}}^2 = \max (SD_{sd,measured}^2, SD_{sd,theoretical}^2) \quad (4)$$

where C_s is the element concentration. λ_s is the element intensity (peak maximum corrected from the background; in cps/nA), and $\sigma_{\lambda_s}^2$ is the variance of the emission of the sample. λ_{sd} is the element intensity (peak background in cps/nA) and $\sigma_{\lambda_{sd}}^2$ is the variance of the emission of the standard. P and B are the peak and background counts (cps) of element i . t is the total counting time (s) on the peak (P) or the background (B). I_i is the current from the Faraday cup. i_{max} is the total number of measurements of element i (combining peak and background).

2.1.7. Synchrotron X-ray Microtomography

X-ray microtomography or micro-CT is a radiographic imaging technique used to reconstruct in 3D the internal structure of any material up to a spatial resolution better than 1 micrometre. X-ray microtomography is a non-destructive technique allowing many scans to be made of the same sample under different conditions. None or very simple sample preparation is necessary. X-ray microtomography originates from Computerised Axial Tomography (CAT or CT) scans, used for medical imaging for decades, but a rapid adaptation of the technique in other fields took place, including materials science and geology.

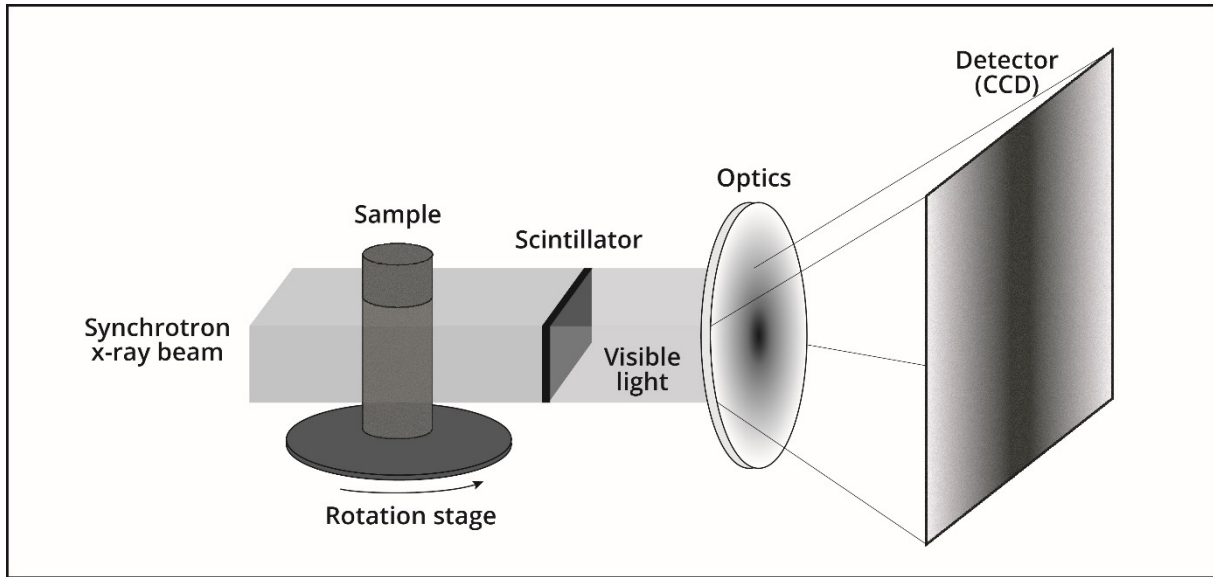


Figure 22. Experimental set-up for synchrotron x-ray microtomography. Modified from Burlion et al. (2006).

The technique consists of acquiring a series of 2D X-ray absorption images recorded as the object is rotated along a single axis (Fig. 22). This series of images is used to reconstruct a three-dimensional digital image where each voxel (volume element or 3D pixel) represents the X-ray absorption in this volume using mathematical principles of tomography. The relationship between X-ray absorption and material density is used to infer the 3D internal structure from the images. The resulting 3D reconstructed samples are typically displayed as a series of 2D “axial slices” (Fig. 23).

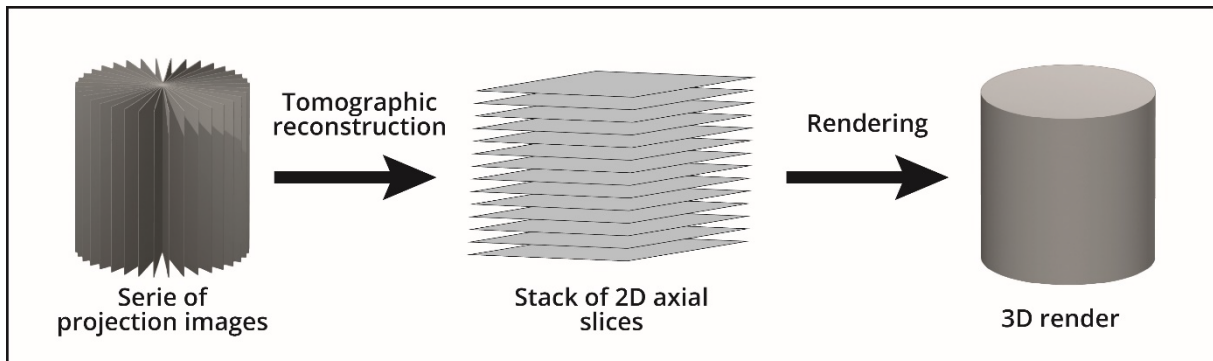


Figure 23. Processing steps from raw images acquired by synchrotron X-ray microtomography.

Spatial resolution obtained using conventional CT systems is limited by the geometry of the X-ray beam and by the characteristics of the detector. For a point source, fan beam or cone beam, the spot size of the X-ray source plays an important role. The smaller the spot size, the smaller the penumbral blurring, which will help produce a more accurate projected image. A larger spot size means that photons hitting a particular pixel can be traced back through different ray paths through the specimen, adding significant noise to the tomographic reconstruction. For this reason, commercial CT systems usually employ micro-focused X-ray

sources with spot sizes of a few micrometres or smaller. Commonly, one-megapixel detectors are used, but some instruments integrate detectors with over ten megapixels.

A significant advance in X-ray microtomographic imaging was the use of synchrotron radiation, greatly enhancing the performed imaging. Synchrotron radiation results from the bending of a high-energy electron beam by a magnetic field stored in a ring. The emitted light is many orders of magnitude greater in brightness than that emitted by laboratory X-ray sources. The high flux allows resolution of very subtle variations in absorptivity and thus allows imaging of detailed internal structures. Synchrotron radiation also allows X-ray beam collimation, simplifying tomographic reconstruction algorithms and specification of the X-ray energy to a narrow energy band. Using a monochromatic X-ray beam improves the accuracy of the reconstructed tomographic images by eliminating the issue of energy dependence on X-ray absorption. In order to obtain high-quality images using synchrotron X-ray sources, relatively small samples, typically 5–10 mm, should be analysed.

Analyses of the samples collected at Vesuvius were performed at the SYRMEP (SYnchrotron Radiation for MEdical Physics) beamline of the Elettra synchrotron in Basovizza, Italy (Fig. 24). Measurements were performed in propagation-based phase contrast mode using a monochromatic, near-parallel X-ray beam set to a mean energy of 23 keV. Lava samples of 4 by 4 millimetres and varying heights were mounted on a rotating head at a distance of 100 millimetres from the detector. 1800 radiographic images have been acquired for each sample with a 0.01° step over a total range of 180°. Each radiograph had an exposure time of 2 seconds, leading to a scan duration on the order of 1 h for each sample. A 0.9 µm pixel size of the detector has been fixed.

The data processing workflow (Fig. 25) starts by reconstructing the 2D axial slices of the samples with an anisotropic voxel size of 0.9 µm using a custom software called Syrmep Tomo Project 4.0. The axial slices were imported into the software ImageJ to be cropped and converted into 8-bit raw images. Images with eventual ring artefacts were removed during this step. The axial slices were imported into the software Avizo, in which an edge-preserving filter was applied to the images to reduce image noise before segmentation. The filtered axial slices were then imported back into ImageJ to segment each mineral phase. Vesicles and crystals were segmented by applying the 3D Otsu automatic thresholding algorithm (Otsu, 1979) with manual refinement following visual inspection of the samples using Avizo to ensure the selection of thin pore-melt films. Voids (vesicles and fractures) and phenocrysts from all phases were separated and subtracted from initial volumes to obtain comparable volumes (Fig. 48).

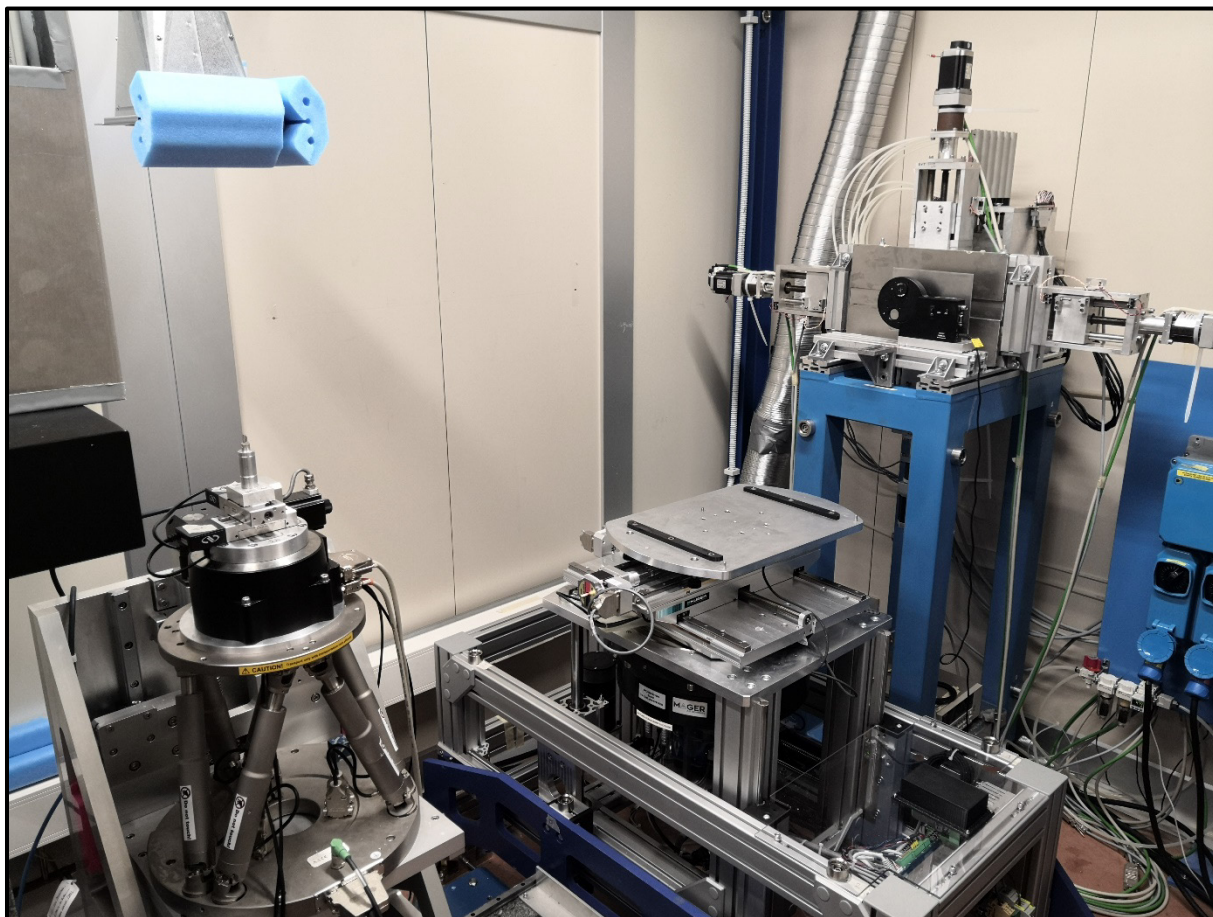


Figure 24. Photograph of the experimental set-up at the synchrotron of Trieste.

The 3D reconstructed samples were then analysed to extract the number, size, and volume of crystals in Avizo. Crystals were separated by a combination of watershed, distance transform, and numerical reconstruction algorithms. Crystals were then counted and measured to calculate the crystal size distribution (CSD) of each phase. Rendered 3D visuals of the raw and processed volumes were produced using Avizo and VGStudio 2022.

Synchrotron 3D microtomography workflow

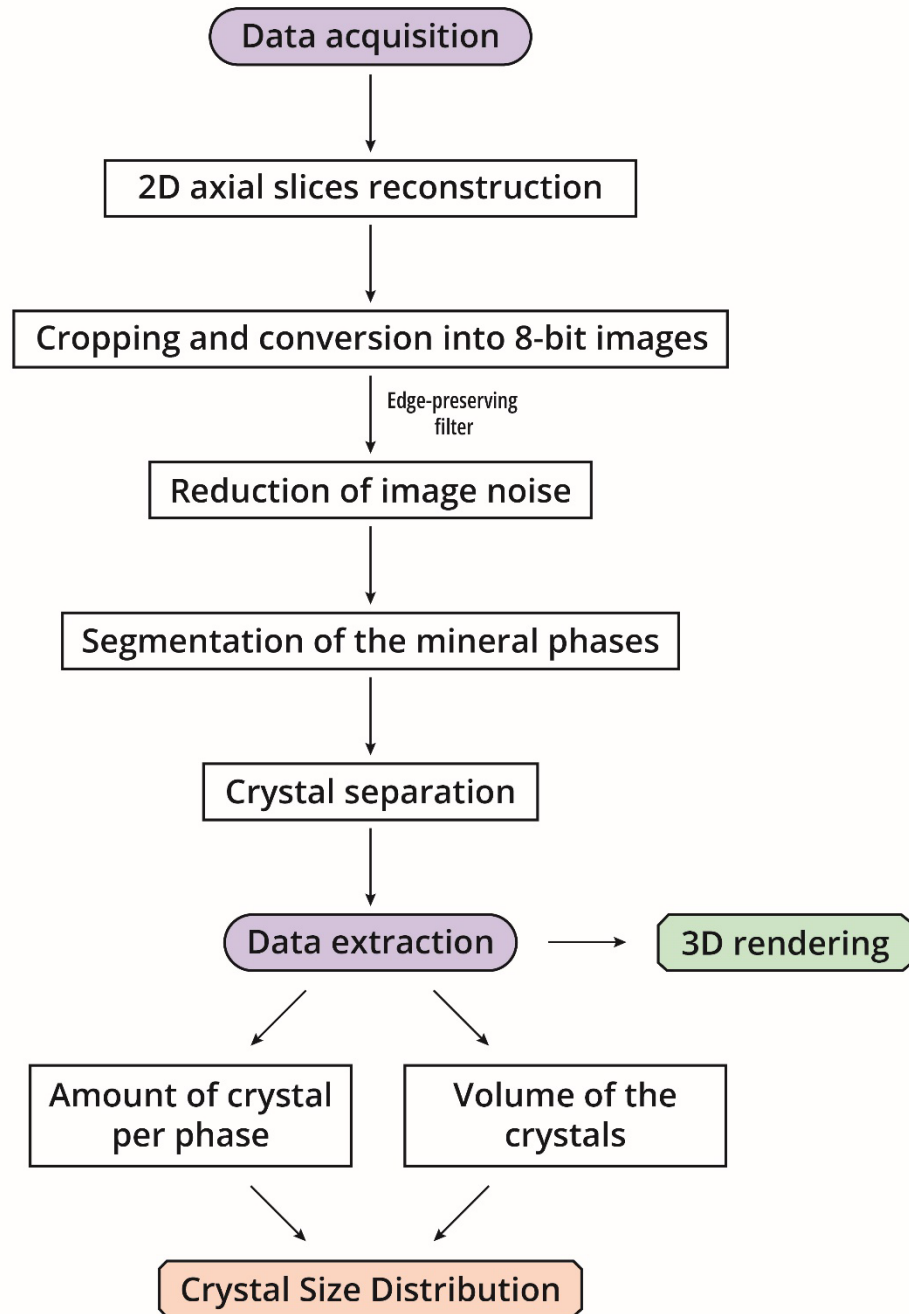


Figure 25. Synchrotron X-ray microtomography data processing workflow

2.2. Thermo-rheological properties and redox conditions

2.2.1. Thermometry

Geothermometers are tools used to estimate the temperature of crystallisation of a mineral in a rock. Mineral compositions vary as a function of temperature and pressure. Geothermometers are based on the chemical equilibrium between minerals and mineral-liquid. Since minerals at equilibrium are dependent on temperature and pressure, it is possible to estimate the temperature of crystallisation by measuring the compositions of coexisting minerals.

In this study, crystallisation temperature (i.e. the temperature of formation of crystals) was estimated using three geothermometers from Putirka (2008):

(1) A clinopyroxene-liquid thermometer (equation 33 in Putirka, 2008):

$$\frac{10^4}{T(K)} = 7.53 - 0.14 \ln \left(\frac{X_{Jd}^{cpx} X_{CaO}^{liq} X_{Fm}^{liq}}{X_{DiHd}^{cpx} X_{Na}^{liq} X_{Al}^{liq}} \right) + 0.07(H_2O^{liq}) - 14.9(X_{CaO}^{liq} X_{SiO_2}^{liq}) - 0.08 \ln(X_{TiO_2}^{liq}) - 3.62(X_{NaO_{0.5}}^{liq} + X_{KO_{0.5}}^{liq}) - 1.1(Mg\#^{liq}) - 0.18 \ln(X_{EnFs}^{cpx}) - 0.027P(kbar) \quad (5)$$

Global calibration of the thermometer yields an estimated error of ± 42 °C in hydrous conditions.

(2) A plagioclase-liquid thermometer (equation 24a in Putirka, 2008):

$$\frac{10^4}{T(K)} = 6.4706 + 0.3128 \ln \left(\frac{X_{An}^{pl}}{X_{CaO}^{liq} (X_{AlO_{1.5}}^{liq})^2 (X_{SiO_2}^{liq})^2} \right) - 8.103(X_{SiO_2}^{liq}) + 4.872(X_{KO_{0.5}}^{liq}) + 1.5346(X_{Ab}^{pl})^2 + 8.661(X_{SiO_2}^{liq})^2 - 3.341 \times 10^{-2}(P(kbar)) + 0.18047(H_2O^{liq}) \quad (6)$$

The thermometer has an estimated error of ± 36 °C.

(3) An olivine-liquid thermometer (equation 22 in Putirka, 2008):

$$T(^{\circ}C) = (15294.6 + 1318.8P(GPa) + 2.4834[P(GPa)]^2) / (8.048 + 2.8352 \ln D_{Mg}^{ol/liq} + 2.097 \ln[1.5(C_{NM}^L)] + 2.575 \ln[3(C_{SiO_2}^{liq})] - 1.41NF + 0.222H_2O^{liq} + 0.5P(GPa)) \quad (7)$$

The estimated error of the thermometer is ± 29 °C when water is present.

The analysed samples are fully crystallised; thus, we assume that crystal cores are in equilibrium with a liquid having the average bulk rock composition (Table 6).

2.2.2. Oxygen fugacity

Oxygen fugacity (fO_2) is the measure of the chemical activity of oxygen in a system, accounting for non-ideal interactions unlike partial pressure. Oxygen fugacity controls the oxidation state of multivalent elements, such as iron, sulphur and copper, and thus for a magma, influences mineral stability, magma viscosity and volatile speciation. Some mineral phases record the fO_2 of the lava during crystallisation, making the estimation of the redox state of a lava possible by measuring crystal compositions. Redox buffers are used to compare oxygen fugacity between different systems, the most common being Quartz-Fayalite-Magnetite (QFM), Hematite-Magnetite (HtMt), Nickel-Nickel-Oxide (NNO) and Iron-Wüstite (IW).

Here, major element compositions of Fe-Ti oxides present in the mineral assemblage were measured to determine oxygen fugacity. The method of Berry and O'Neill (2021) was used, which links the ferric/ferrous iron ratios of Fe-Ti oxides to the fO_2 while also taking into account the chemical composition of the lava. Oxygen fugacity is calculated as $\Delta \log fO_2$ relative to the Quartz-Fayalite-Magnetite buffer (QFM). The method can be used for lavas with a silica content of less than 60 wt.%. The average bulk rock composition of the analysed lavas was used. Berry and O'Neill (2021) defined the following empirical equation (Equation 8.2 in their work):

$$\Delta QFM = \frac{\log_{10} \frac{Fe^{3+}}{Fe^{2+}} + 1.35 - 0.023[CaO] - 0.034[Na_2O] - 0.044[K_2O] + 0.018[P_2O_5]}{0.25} \quad (8)$$

where [CaO], [Na₂O], [K₂O] and [P₂O₅] are the concentrations of the oxide components in weight per cent. The concentrations of ferrous iron (Fe²⁺) and ferric iron (Fe³⁺) were estimated as follows:

$$Fe^{2+} = \frac{[Fe^{2+}] + 4[Ti] - 2[Mg] - 2[Mn]}{3} \quad (9)$$

$$Fe^{3+} = \frac{2 \times ([Fe^{2+}] - 2[Ti] + [Mg] + [Mn])}{3} \quad (10)$$

where [Fe²⁺], [Ti], [Mg] and [Mn] are the number of cations contributing to the oxide formula.

The calculated proportions of total iron are used in equation 8 as the ferric/ferrous iron ratio

$$\frac{Fe^{3+}}{Fe^{2+}}$$

2.2.3. Thermodynamic modelling

Thermodynamic modelling consists of predicting the state of a magma using thermodynamic principles. One of these principles was introduced in 1873 by J.W. Gibbs as ‘available energy’, now called Gibbs free energy (Gibbs, 1873). It can be expressed as:

$$G = H - TS \quad (11)$$

where H is the enthalpy of the system, S is the entropy of the system, and T is the temperature of the system. If a small change occurs in a system kept at constant temperature and pressure, the resulting free energy change is given by:

$$\Delta G = \Delta H - T\Delta S \quad (12)$$

At chemical equilibrium, the Gibbs free energy change (ΔG) for a reaction is zero, and the system is at its minimum possible energy state. This state is defined by the equation:

$$\Delta G^\circ = -RT \ln K \quad (13)$$

where ΔG° is the standard Gibbs free energy change, R is the ideal gas constant, T is the temperature (in K), and K is the equilibrium constant. A negative standard Gibbs free energy change leads to a large K (products favoured), while a positive standard Gibbs free energy change leads to a small K (reactants favoured).

Calculation of the Gibbs free energy of solid, liquid and gaseous phases allows to determine which phases are stable for a given composition, temperature, pressure, and oxygen fugacity. Thermodynamic models are calibrated on experimental data.

The software MELTS (Ghiorso and Sack, 1995), specifically Rhyolite-MELT version 1.2 (Gualda *et al.*, 2012; Ghiorso and Gualda, 2015), developed to model chemical mass transfer in magmatic systems, was used to constrain the crystallisation processes that took place within the lava tube. This software allows to (1) model equilibrium or fractional crystallisation of specified bulk compositions, (2) model crystallisation paths under constrained temperature, oxygen fugacity, total enthalpy, total entropy, total volume or any combination of these constraints, and (3) model assimilation or magma mixing in evolving magmatic systems. The input composition corresponds to the average bulk rock composition measured by XRF analysis (Table 6), combined with the estimated water content of 0.50 wt.% from Belkin *et al.* (1993). The crystallisation sequence was modelled at atmospheric pressure (1 bar) from 1300 °C down to the solidus temperature at 970 °C in 10 °C steps. Simulations were performed under three

different oxidation conditions, represented by oxygen fugacity buffers: Hematite-Magnetite (HtMt) and Quartz-Fayalite-Magnetite (QFM +2 and QFM +3). Outputs of the model are crystallisation paths showing liquidus temperatures, phase proportions and compositions.

2.2.4. Viscosity estimations

Viscosity is a fundamental rheological property that measures a fluid's internal resistance to flow under an applied stress. For magmatic systems, viscosity is a critical parameter controlling eruption dynamics. Specifically for lava flow emplacement, viscosity is used to determine flow rate, internal forces, structure formation and ultimately emplacement distance (Llewellyn and Manga, 2005; Melnik and Sparks, 2005; Rowland *et al.*, 2005; Caricchi *et al.*, 2007; Gonnermann and Manga, 2007; Massol and Jaupart, 2009; Applegarth *et al.*, 2010). The rheology of any fluid consists of the relation of shear stress (τ) with shear strain-rate ($\dot{\gamma}$), usually plotted as flow curves (Fig. 15b). The simplest rheology is called Newtonian, it is defined by the equation: $\tau = \mu\dot{\gamma}$ and is plotted as a straight line passing by the origin with a slope of μ , its viscosity. Many fluids have non-Newtonian rheology, plotting as non-linear flow curves (Fig. 15b). As it is not possible to define a single strain-rate-independent viscosity for such flow curves, the rheology is described in terms of an apparent viscosity η , the ratio of stress to strain-rate ($\eta = \tau/\dot{\gamma}$), varying with strain-rate.

Natural magmas are complex mixtures of crystals and gas suspended in a silicate melt (Harris, 2015). The rheology of silicate melts has been well-studied across a wide range of compositions and conditions (Hess and Dingwell, 1996; Giordano *et al.*, 2006; Neuville, 2006; Poe *et al.*, 2006; Vetere *et al.*, 2006; Hui and Zhang, 2007; Giordano *et al.*, 2008, 2009; Vetere *et al.*, 2008; Whittington *et al.*, 2009), and can be treated as Newtonian liquids. Viscosity depends mainly on (1) chemical composition, particularly SiO_2 , which forms polymerised networks in the melt, increasing resistance to flow (Mysen and Virgo, 1994; Giordano *et al.*, 2008), (2) temperature, decreasing viscosity exponentially with temperature, and (3) multiphase suspensions of crystals and vesicles. Crystals are rigid, incompressible particles which increase viscosity due to particle-particle interactions, while vesicles are deformable, compressible and inviscid particles which can either reduce or increase viscosity (Harris *et al.*, 2019). To estimate lava viscosity, it is necessary to calculate the viscosity of the liquid and add the effect of crystals and vesicles.

In this work, lava is considered as a two-phase suspension, meaning only crystals are suspended in a viscous melt, as the laws describing three-phase suspensions, including vesicles,

are not well constrained. The apparent viscosity (η_{app}) of the lava is calculated as follow:

$$\eta_{app} = \eta_{liq} \times \eta_r \quad (14)$$

where η_{liq} is the viscosity of the liquid phase and η_r accounts for the effect of suspended crystals and vesicles. The viscosity of the liquid phase η_{liq} was determined following the GRD model (Giordano *et al.*, 2008). This model is widely used as it considers a wide range of compositions and integrates the non-Arrhenian temperature dependence of silicate melt viscosity modelled by the Vogel-Fulcher-Tammann (VFT) equation (Vogel, 1921; Fulcher, 1925; Tammann and Hesse, 1926). Giordano *et al.* (2008) write the VFT equation as follows:

$$\log \eta = A + \frac{B}{T(K)-C} \quad (15)$$

where A, B, and C are adjustable parameters. A is the value of $\log \eta$ (Pa s) at infinite temperature, representing the high temperature limit to silicate melt viscosity ($\log \eta_\infty$). A is assumed to be an unknown constant; thus, the effect of composition on viscosity is ascribed to B and C. These fitting parameters are calculated based on the melt composition obtained every 10 °C step from MELTS simulations (Section 2.2.3). Here, the effect of suspended particles (η_r) only considers crystals. The effect of crystals on lava viscosity was incorporated to the liquid viscosity using the computed volume fraction of crystals. To estimate the effect of crystals on the viscosity, three rheological models were applied.

1. The Shaw model (Shaw, 1969), based on the Einstein-Roscoe equation, which considers all crystals to be spherical (assuming a theoretical maximum packing fraction of 0.74):

$$\eta_r = (1 - 1.35\phi)^{-2.5} \quad (16)$$

where ϕ is the volume fraction of crystals in the lava.

2. The Krieger and Dougherty method (Krieger and Dougherty, 1959), also based on the Einstein-Roscoe equation but takes into account the shape of crystals:

$$\eta_r = 1 - \left(\frac{\phi}{\phi_m} \right)^{B\phi_m} \quad (17)$$

where ϕ_m is the maximum packing fraction (Mueller *et al.*, 2010).

We approximated with this model the effect of large rounded particles (with a maximum packing fraction is 0.641) and fine elongated particles (with a maximum packing fraction is 0.454) by calculating the product of the effect for both crystal shapes (Farris, 1968):

$$\eta_r = \eta_r(\phi_l) \times \eta_r(\phi_f) \quad (18)$$

3. The model of Costa (Costa, 2005; Costa *et al.*, 2009) based on a semi-empirical non-Newtonian relationship that takes into account the shape of crystals and the effect of shear rate:

$$\eta_r = \frac{1 + \frac{\phi}{\phi_m}^\delta}{[1-F]^B \phi_m} \quad (19)$$

where F is calculated as follows:

$$F = (1 - \varepsilon) \operatorname{erf} \left[\frac{\sqrt{\pi}}{2(1-\varepsilon)} \frac{\phi}{\phi_*} \left(1 + \frac{\phi^\gamma}{\phi_*^\gamma} \right) \right] \quad (20)$$

in which the error function (*erf*) expresses a rheological transition below and above the maximum packing fraction. The fitting parameters for crystal types A and D from Cimarelli *et al.* (2011) were used. Both the shape and the amount of crystals of the two types of particles can be introduced in the model using the following equations:

$$\phi_{*lf} = \phi_{*f} x^{1.34} + \phi_{*l} (1 - x)^{1.34} \quad (21)$$

$$\gamma_{*lf} = \gamma_{*f} x + \gamma_{*l} (1 - x) \quad (22)$$

$$\varepsilon_{*lf} = \varepsilon_{*f} x + \varepsilon_{*l} (1 - x) \quad (23)$$

where x is the fraction of finer to coarser particles.

Relative viscosity was calculated with the Costa model at shear rates of 10^{-4} and 1 s^{-1} .

Table 3. Fittings parameters from Cimarelli *et al.* (2011) were used to estimate the relative viscosity using the Costa model.

	A		D	
Strain rate (s^{-1})	10^{-4}	1	10^{-4}	1
ϕ	0.62	0.67	0.24	0.28
γ	1.52	1.6	0.1	0.11
ξ	5000	10000	30000	21000

In both the Krieger and Dougherty and Costa models, a bi-modal mixture of crystals was assumed. Leucite, Fe-Ti oxides and olivine were treated as large rounded particles, while plagioclase and clinopyroxene were modelled as smaller elongated particles.

2.3. Detailed morphological reconstruction

Different high-end instruments were used together to numerically reconstruct the morphology of the studied lava tubes. Both ground and airborne surveys were carried out to fully capture their geometry, allowing a detailed investigation of the lava tube morphometries.

2.3.1. Terrestrial Laser Scanner (TLS) and ground photogrammetry

The interior of the investigated lava tube was captured using a terrestrial laser scanner (TLS). A Riegl VZ-400 scanner mounted on a tripod was used. It is a time-of-flight (ToF) scanner that measures the distance to a specific point by calculating the time it takes for a light signal to travel from the scanner to the object and back (Fig. 26). The laser scanner emits a pulsed light signal, here a high laser pulse with a repetition rate of up to 1.2 MHz and measures the time it takes for the signal to return to the sensor, allowing precise distance calculations. The scanner has an angular measurement resolution of 0.0005° . Several scans from different points of view were necessary to minimise shadow areas produced by the complex morphology of the tube (collapses, tube extensions). Reflectors of varied sizes were placed inside and outside the lava tube to obtain common points of reference between the TLS scans for easier merging with scans produced by other techniques for precise alignment. Some key parts of the lava tube that link the interior and exterior of the site, such as skylights and large fractures in the roof, were photographed with a GoPro 10 camera to perform photogrammetry work. These surveys allow a better connection between the TLS scans and the airborne scans.

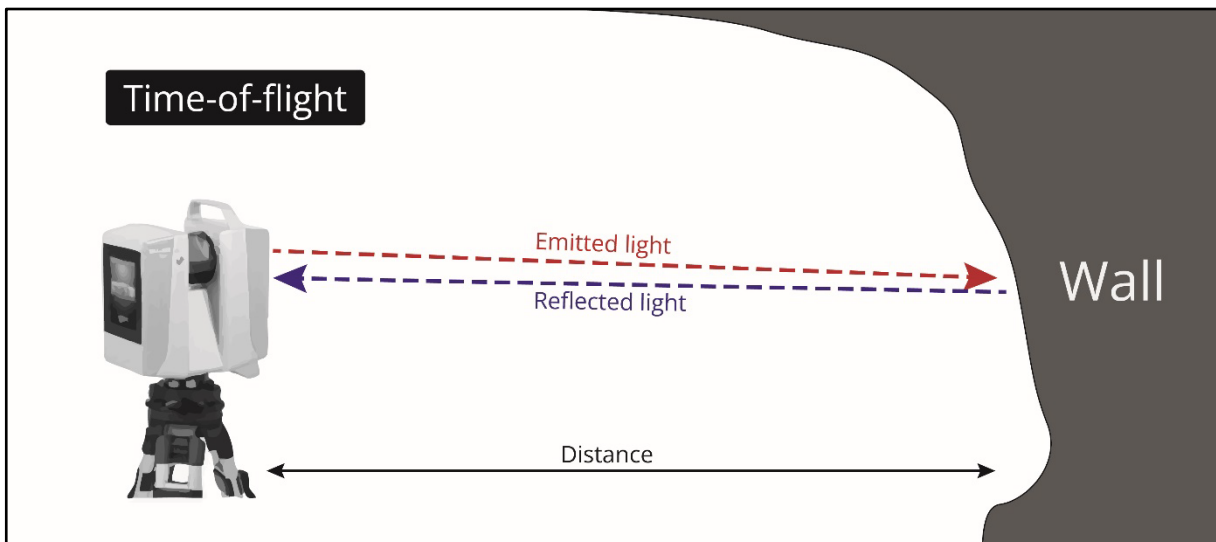


Figure 26. Physical process behind time-of-flight laser scanning.

The outputs of the scans are point clouds. For each lava tube, point clouds are merged in a specific software, RiSCAN Pro, to build a complete point cloud of the tube interior that can be transformed into a mesh.

2.3.2. Airborne LiDAR and photogrammetry

A Matrice 300 RTK UAV (Unmanned Aerial Vehicle) from DJI (Fig. 27) was used to carry out airborne surveys to scan the overlying surface of the lava tube. This UAV can be



Figure 27. In-flight DJI Matrice 300 RTK UAV equipped with the Zenmuse P1.

equipped with different sensors that are stabilised on three axes thanks to the Zenmuse stabiliser. Two sensors were used during this Ph.D.: (1) the Zenmuse P1, an optical camera that has a full frame CMOS sensor of 45 MP accompanied by a 35 mm lens opening from f/2.8 to f/16 and with a FOV of 63.5°, and (2) the Zenmuse L1, a LiDAR (Light Detection And Ranging) sensor that captures point clouds of the investigated areas with a resolution of 3 centimetres at 100 metres altitude. For optimal surface scanning and later georeferencing, an accurate localisation system needs to be employed. The Matrice 300 is equipped with a really accurate positioning technology called RTK for Real Time Kinematic, which is a satellite-based positioning technique that uses multiple systems (GPS, GLONASS, Galileo) and a local base that corrects the positioning data in real time. RTK positioning has a precision of within centimetres.

Aerial photogrammetry uses optical images taken from different points of view to reconstruct an area in 3D via a technique called photogrammetry (Figure 28). This technique produces 3D data from 2D photographs, allowing precise measurements in a three-dimensional space. Photogrammetry uses the Structure-from-motion (SfM) algorithm to determine the local motion signals of similar points used to estimate the 3D structure. Aerial photographs are captured at the same altitude using the follow-the-terrain feature of DJI and with an overlap of at least 70%. These 2D images are imported into the software Agisoft Metashape Professional

to perform the photogrammetry treatment and produce 3D data files (point clouds, meshes, DEM, orthophotos).

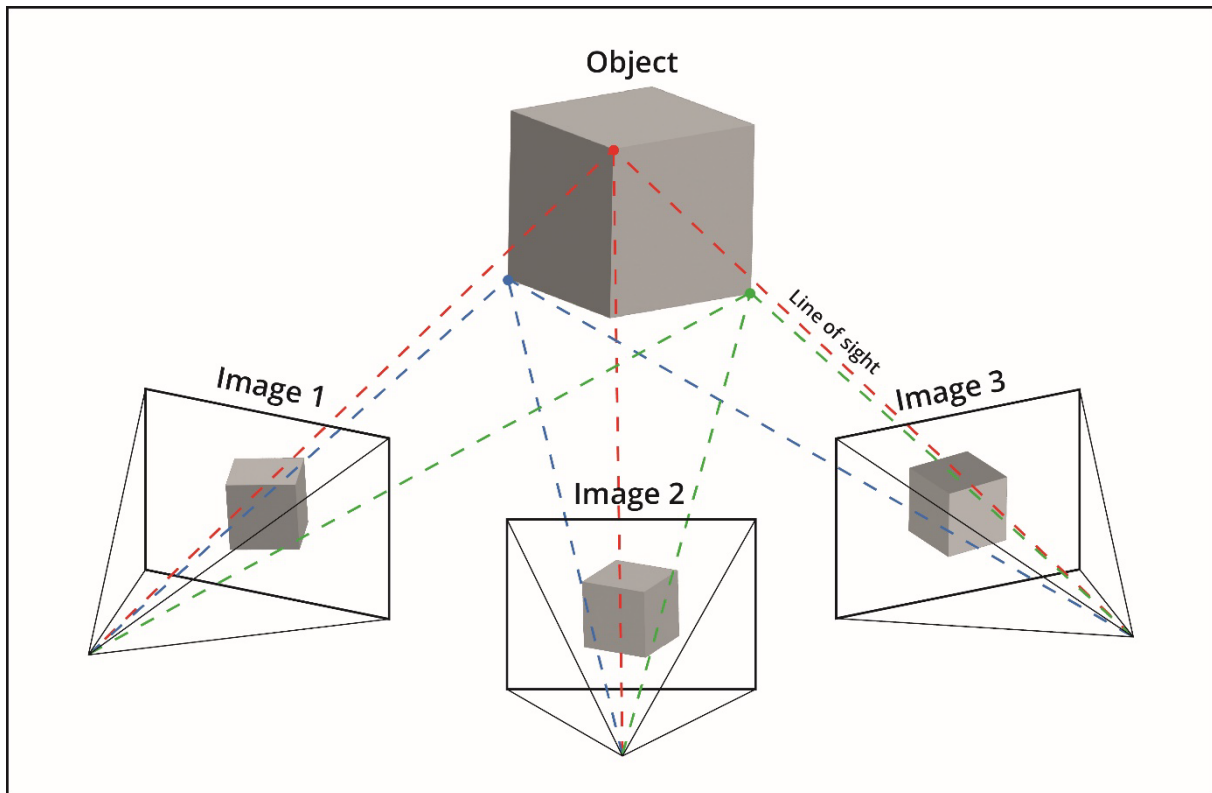


Figure 28. Principle of photogrammetry.

Airborne LiDAR is a technique that determines the range between the sensor and a surface by illuminating the surface with a laser and measuring the time needed for the reflected light to travel back to the sensor (Fig. 29). The distance sensor/surface is calculated using the time for the laser light to travel to the surface and back to the sensor. LiDAR produces point cloud data. These point clouds were treated using the DJI Terra software.

Markers were placed on and in the lava tube to facilitate the alignment of scans produced by the different methods.

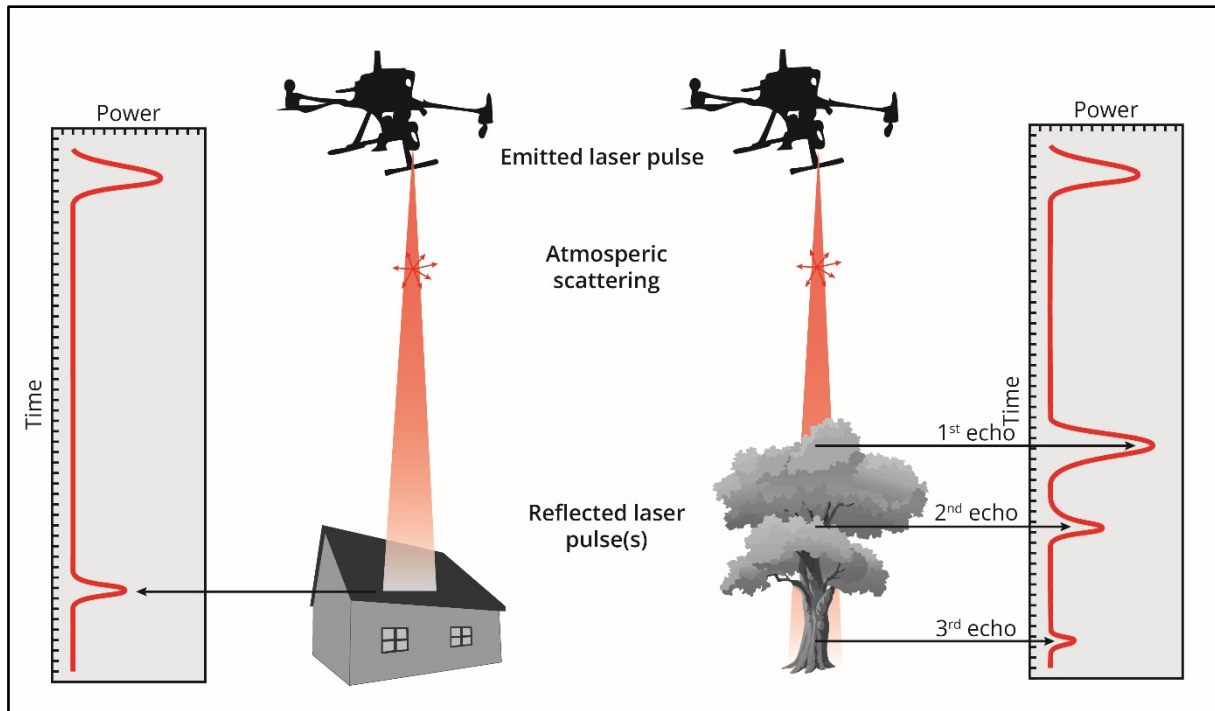


Figure 29. Principle of LiDAR acquisition.

2.3.3. Complete 3D model and morphometry analysis

All the point clouds produced by the different acquisition techniques were imported into the software CloudCompare. Alignment of the scans using the position of common reflectors and recognisable features in the overall geometry of the lava tube is made. The aligned point clouds are then merged together. The complete point cloud was transformed into a mesh and exported to the software Rhino7 to be analysed. The 3D reconstructed lava tubes were cut into multiple longitudinal, lateral, and vertical sections. Length, width, thickness, and roof thickness measurements were performed every metre along the sections. The overall internal slope of the lava tubes was measured following flow lines.

2.4. Stability assessment

2.4.1. Acoustic Emission (AE) testing

To assess the stability of lava tubes, we used Acoustic Emission (AE) testing, a non-destructive testing (NDT) method. AE testing is a passive NDT method as it captures acoustic signals produced within the material. Acoustic Emission (AE) testing consists of capturing transient elastic waves produced by a sudden redistribution of stress in a material when a structure is affected by an external stimulus such as a change in pressure, load, or temperature (Fig. 30). The technique is used during the loading of a structure and thus has the advantage of

being performed during operation. Sources of AE vary from large, like earthquakes and rock bursts, to small, like the formation and growth of cracks, slip, and dislocation movements. Detecting and analysing AE signals gives information on the origin and importance of a discontinuity in a material. Acoustic Emission testing has many industrial applications, like assessing structural integrity, detecting flaws, testing for leaks, or monitoring weld quality, but is also used extensively in research. AE testing captures dynamic processes in a material, highlighting active features such as crack growth. It is thus possible to determine if a crack is developing or is stable if enough loading is affecting the structure. It gives AE testing the ability to immediately give information on the risk of failure and the stability of a structure.

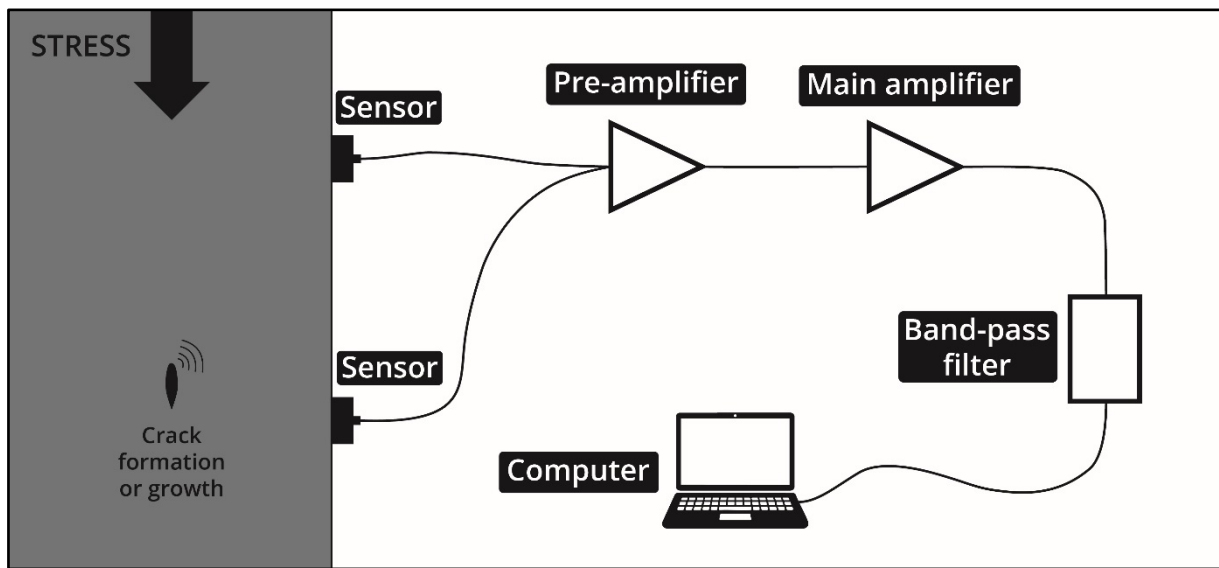


Figure 30. Acoustic emission testing principle and common acquisition system.

An AE testing system (Fig. 30) usually consists of acoustic sensors that detect and respond to dynamic motion caused by acoustic emissions. Transducers convert the dynamic motions at the surface of the material into electric signals. The transducer element in an AE sensor is usually a piezoelectric crystal, commonly made from a ceramic. The selection of transducers is based on operating frequency, sensitivity, and environmental characteristics. Transducers are either resonant or broadband. Typically, AE equipment responds to movement in a frequency range of 30 kHz to 1 MHz. AE signals are weak and need to be magnified by a pre-amplifier that boosts the voltage. The pre-amplifier is commonly placed close to the transducer to minimise interference. Noises in the signals are eliminated by a band-pass filter that removes low frequencies, which are commonly background noises, and high frequencies. Enhanced and filtered signals can be further enhanced by a main amplifier and then are recorded on a computer. An AE testing system records parameters such as count, hit, event, rise time, duration, peak amplitude, energy, root mean square (RMS) voltage, frequency spectrum, and arrival time difference.

2.4.2. Instrumentation

The literature shows that acoustic emission signals from cavernous structures, or signals from the damage progressions in materials collected from cavernous structures, are characterised by low amplitude and low frequency. Therefore, to acquire acoustic emission signals in lava tubes, we assume that narrowband sensors with high sensitivity are suitable. Due to the high sensitivity required, we exclude the use of contactless sensors. Possible instabilities in the structure of the lava tube also inhibit the use of multi-crystal piezoelectric sensors due to their low shock resistance.

The ISR15 sensor was selected based on the constraints mentioned previously and on the availability of commercial instrumentation. The ISR15 is a narrowband sensor operating between 50 kHz and 200 kHz with a peak sensitivity of $69 \text{ dB ref } V/\left(\frac{m}{s}\right)$. It is an intrinsically safe sensor, with a piezoelectric crystal protected by a thick ceramic wear plate and completely enclosed with an epoxy coating. The surface area of the transducing element is approximately $1.6 \times 10^3 \text{ mm}^2$. The dimensions of the sensor do not have any adverse effects on the signal acquisition because the source of the acoustic emission signals generally acquired in cavernous structures is relatively smaller than the wavelength of interest. The sensor can operate without losing sensitivity at temperatures between -45°C and 125°C and has a shock limit of 1000 g, making it potentially suitable for acquiring acoustic emission signals inside lava tubes.

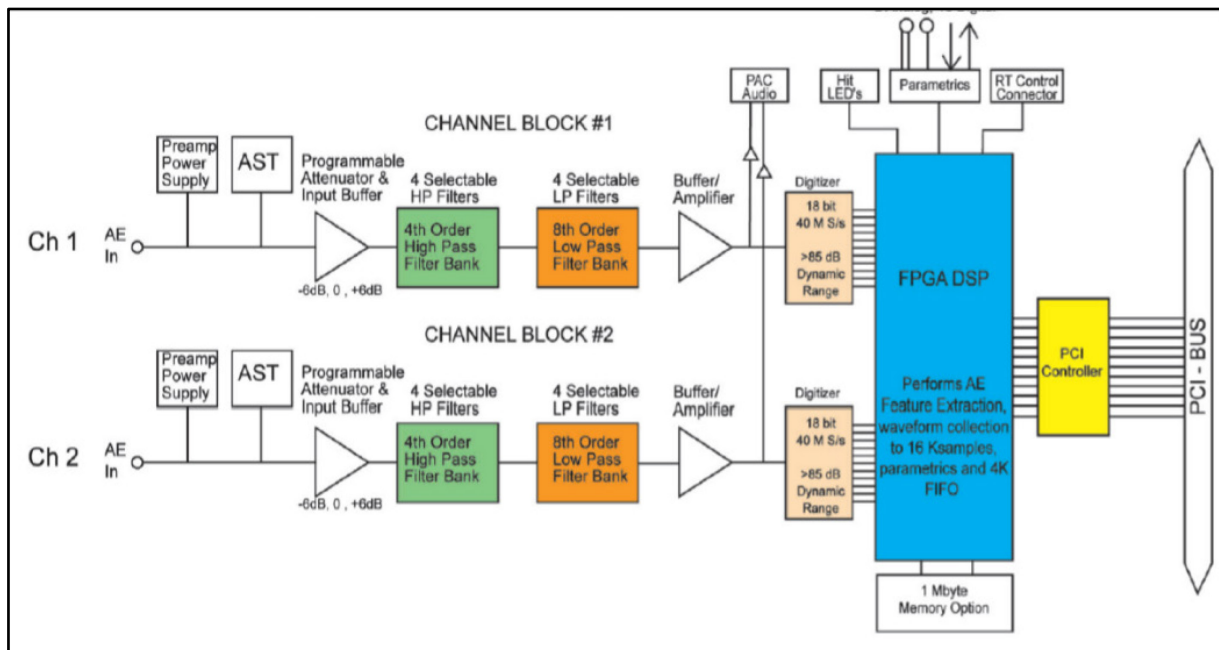


Figure 31. Technical sketch of the PCI-2 acquisition card from the manufacturer.

The PAC PCI-2 data acquisition system from MISTRAS (Fig. 31) was chosen for the on-site AE tests. The setup is composed of a four-channel acquisition system, in which the sensors are powered by a 2/4/6 integral preamplifier. A two-meter BNC coaxial cable is used to power the preamplifier and to register the acquired AE signals. The acquired AE waveforms are sampled at a rate of 2 MHz with a sample length of 1024.

The AE tests start by placing the sensors in position for 500 seconds, acquiring AE signals continuously. If no signals are registered during the first 100 seconds period, then the tests are interrupted and repeated. The AE signals are pre-processed before analysis. Events with counts (i.e., the total number of instances of the signal amplitude crossing the detection threshold of 26 dB) less than 3 are discarded as noise. The signal waveforms are then filtered using the built-in analogue and digital filters of the data acquisition system. The filtered signals are sampled at 2 MHz and stored for analysis.

2.4.3. Data processing

AE signals are typically analysed in the Time Frequency (TF) domain, allowing for representation of the signals over both time and frequency. AE signals are commonly analysed using short-time Fourier transform (STFT), continuous wavelet transforms (CWT) and discrete wavelet transforms (DWT), or wavelet-free transforms such as empirical mode decomposition (EMD; Mousavi *et al.*, 2016; Ebrahimian *et al.*, 2019; Zhang *et al.*, 2021; Dong *et al.*, 2024). STFT is limited by a fixed resolution but is particularly useful to represent continuous and quasi-stationary signal segments (Maia *et al.*, 2024). While CWT and DWT allow customised TF resolution using different wavelet functions, the selection of wavelets is not trivial. Therefore, in this work, a novel approach based on EMD and on the Hilbert-Huang transform (HHT) is used. The EMD method decomposes the whole data set into several components that satisfy Intrinsic Mode Functions (IMF). EMD is an adaptive method that generates its own basis functions considering the properties of the signals, allowing meaningful decomposition of signals that change over time. The IMFs, ordered from highest to lowest frequency, are used with the Hilbert-Huang Transform (HHT) to create a Hilbert-Huang Spectrum, providing insights into the instantaneous frequency and amplitude of the signal (Huang and Shen, 2014).

We developed a novel procedure to process the AE signals recorded inside lava tubes. The AE signals in the time domain are decomposed into different IMFs and residuals using the EMD method. Then, by comparing each IMF with the original signal using permutation entropy (PE), the IMFs with significant signal content are selected. PE is the measure of complexity in the time series signal and is used in prior studies to select the IMFs (Si *et al.*, 2014; Mohd Jaafar *et al.*, 2019; Koppolu and Chemmangat, 2023). It is hypothesised that the IMFs with low PE contain significant signal content, as they are characterised by low complexity. In this context, the PE of the raw signal and decomposed IMFs are compared. Finally, the IMFs sharing a similar PE to the raw signal with a significance level of 95% are selected. The selected IMFs are summed up to reconstruct the time series signal. This allows the selection of significant signal content without the necessity of windowing or selecting wavelets, which are essential in STFT, CWT, and DWT. The resulting signal has an improved signal-to-noise ratio compared to the original signal. The frequency spectrum of the denoised AE signals is then analysed using the fast Fourier transform (FFT) method. This reveals the information about the central frequency or the principal frequency bands of the signal. The TF analysis is carried out using the HHT method. In HHT, the instantaneous frequency, instantaneous amplitude, and their temporal information are extracted while working within the predefined frequency limits. For the sample signal, the HHT is applied within the frequency limits of 50 kHz and 200 kHz. This novel approach uses PE-based EMD and frequency-localised HHT to extract the principal spectral contents in the TF domain of the AE signal. In addition, this method is less likely to be susceptible to variations in the signal-to-noise ratio and thus can be used to analyse the AE signals acquired from high noise environments such as cavernous structures.

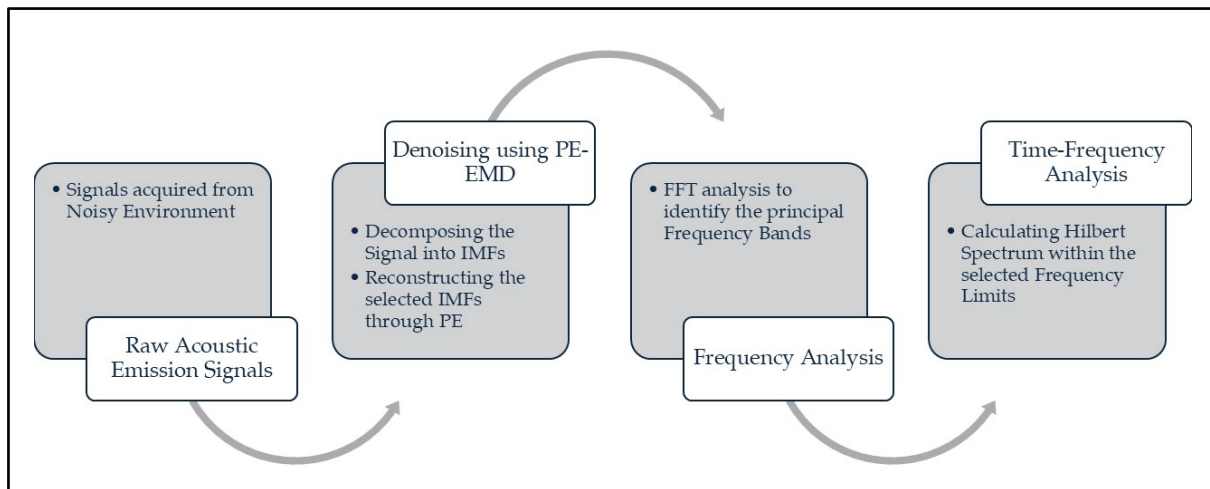


Figure 32. Workflow of the novel processing method applied in this work.

2.5. Geoheritage, conservation and scientific dissemination

The creation of a geoitinerary and virtual tour, its geoheritage relevance and its dissemination involves fieldwork, 3D scanning, 360° photography, explicative panels and SWOT analysis. This multidisciplinary approach aims to give visitors a full understanding of effusive activity at Vesuvius and the associated hazards.

2.5.1. Fieldwork

Extensive fieldwork was carried out on the 1858 lava flow field, first to find and identify peculiar and preserved surface features. Once found, the sites were measured and analysed. The GPS positions of the sites were taken using a Garmin handheld GPS. Photographs of the sites were taken using a Canon 70D camera equipped with an 18-135 mm lens. We captured 360° HDR photos with a 72 MP resolution at each site using an Insta360 X4 camera. One of the sites of the geoitinerary (Site 5) was scanned in 3D as detailed previously (section 2.3).

2.5.2. Image processing

Raw photographs were edited, corrected for optical distortion and exported as JPG files in the software Adobe Photoshop Lightroom. The raw HDR 360° photos captured as INSP files were edited using the constructor's software Insta360 Studio. The dewarped field of view was applied to the photos and they were exported as 360° JPG photos.

2.5.3. Panel creation

Panels describing the formation and growth of the selected volcanic landforms to help visitors understand the mechanisms that drive the creation of such features during lava flow field emplacement were developed. Each panel shows photos and 360° previews of the site, gives a definition of the observed structure and presents the mechanism that allowed the formation of the site. On each panel, a QR code allows the visitor to access the site's webpage (Section 2.5.4). These panels should be placed next to the sites, as their use as a communication strategy is common in geotourism research (Melelli, 2019).

2.5.4. Geoitinerary and virtual tour development

The geoitinerary was designed to link the sites efficiently and easily for the visitors to follow. The GPS position of each site was imported into the software QGIS. We also used the GPS tracking of each fieldwork, which recorded the paths we followed to help design the itinerary along the 1858 lava flow field. The geoitinerary was drawn in QGIS. A map was exported that indicates the site's location, the geoitinerary to follow and parking options.

Access to geoitinerary still requires travelling to the location and necessitates being able to hike in uneasy environments. Thus, we created a virtual tour allowing people with physical impediments to discover the volcanic landforms of the 1858 lava flow field. We designed a website that showcases the geoitinerary and sites. Each site has its own subpages that comprise its explicative panel and immersive 360° photos (<https://sites.google.com/view/vesuvius-1858lavaflow/>). The 360° photos are hosted on Momento360 and integrated on the website in 360° compatible viewers.

2.5.5. Geoheritage relevance

To assess the geotourism potential of the proposed geoitinerary and provide initiatives for their efficient and effective use (Carrión-Mero *et al.*, 2020; Kubalíková and Kirchner, 2016; Valente *et al.*, 2021), we carried out the Strengths, Weaknesses, Opportunities and Threats (SWOT). SWOT analyses are used to precise an objective and identify internal and external factors favourable or unfavourable to the project. Strengths are characteristics of the project giving it an advantage compared to others, while Weaknesses are the characteristics that give a disadvantage. Opportunities are elements of the environment that can be used to the advantage of the project, while Threats are elements that can cause problems to the project.

CHAPTER 3

Lava flow field development and lava tube formation during the 1858–1861 eruption of Vesuvius (Italy), unravelling by historical documentation, lidar data and 3D mapping

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Abstract

Somma-Vesuvius is well known for its powerful Plinian explosive eruptions, however, during the last eruptive cycle (1631-1944), persistent activity took place on the stratovolcano as mild and violent Strombolian, and effusive eruptions, forming more than one hundred lava flow fields. An important mechanism of lava transport within lava flow fields is the formation and development of lava tubes. The presence of lava tubes in a flow field can greatly increase their distance of emplacement. Observations of lava tubes at Vesuvius have been documented in historical records and speleological reports, but no modern scientific studies are available. This work focuses on lava tubes formed in the compound lava flow field of the long-lived 1858 eruption (from 27 May 1858 to 12 April 1861) that was fed by seven eruptive fissures. The temporal and spatial evolution of the 1858 lava flow field was reconstructed using historical documentation. The exposed lava flow field surface was analysed using a 1-m resolution lidar Digital Surface Model (DSM). Surveys to fully digitise the interior and the overlying surface of the largest lava tube found in the 1858 lava flow field were conducted using a terrestrial laser scanner, optical cameras, and an Unmanned Aerial Vehicle (UAV). The accurate 3D model obtained was used to precisely quantify the inner dimensions and to better constrain the morphologies of the lava tube. Observed internal features were described and used to gain information on the formation and activity of the lava tube. Our data allowed us to understand that the described lava tube formed as an inflated lava flow inside which lava flowed through during an extended period ultimately draining out completely at the end of the eruption. Understanding how lava flow fields develop and how lava tubes form on Vesuvius is crucial to re-evaluate the last effusive activity of the volcano and its impact on hazard assessment.

Keywords: lava tunnel, tumulus, inflation, stratovolcano, 3D digitization, effusion

3.1. Introduction

Somma-Vesuvius is one of the first and most studied volcanoes in the world. Most of the attention of the scientific community was on its explosive activity, especially on the famous AD 79 Plinian eruption. However, the last eruptive cycle of Vesuvius, that started in 1631 with a sub-Plinian eruption and stopped in 1944, was predominantly effusive (Carta *et al.*, 1981; Arno *et al.*, 1987; Arrighi *et al.*, 2001; Cioni *et al.*, 2008; Scandone *et al.*, 2008). For this eruptive activity many historical descriptions, chronicles of the time, drawings, paintings, and maps are available. Modern studies on the effusive activity of Vesuvius are limited in comparison to the ones on its explosive behaviour. Apart from the works of Belkin *et al.* (1993), Trigila and De Benedetti (1993) and Villemant *et al.* (1993) on the petrology and mineralogy of the lavas of Vesuvius and the study of Ventura and Vilardo (2008) on the 1944 lava flows, the emplacement mechanisms of lava flows on Vesuvius and their associated surface features were not fully investigated. During more than three hundred years, over one hundred lava flows were emplaced on the slopes of Vesuvius. Despite the low hazard linked to effusive eruptions due to the usually slow progression of lava flows, they can still pose a danger for populations living on the flanks of volcanoes (Duncan *et al.*, 1996; Chester *et al.*, 2007). In the past, fatalities caused by lava flows have occurred (Harris, 2015; Brown *et al.*, 2017). The main problem caused by lava flows for the population is the destruction of immovable properties like houses, businesses, agricultural lands, and critical facilities such as schools, hospitals, roads, and administrative buildings (Coltelli *et al.*, 2012; Jenkins *et al.*, 2017; Del Negro *et al.*, 2019; Carracedo *et al.*, 2022b; Meredith *et al.*, 2022, 2024). Lava flows can also cause secondary hazards such as wildfires or air pollution (Ainsworth and Boone Kauffman, 2009; Vasconez *et al.*, 2018; Filippi *et al.*, 2021; Carracedo *et al.*, 2022b; Hernández *et al.*, 2022).

An important mechanism of lava transport in lava flows is by the development of lava tubes (Swanson, 1973; Greeley, 1987). Lava tubes are defined as ‘roofed conduit of flowing lava, either active, drained or plugged’ (Kauahikaua *et al.*, 1998; Halliday, 2004), and in this study, we apply this term independently of their shape or size. The presence of lava tubes can significantly increase lava flow length due to the insulation of the hot molten lava by an overlying cooled crust (Swanson, 1973; Peterson *et al.*, 1994; Keszthelyi, 1995; Witter and Harris, 2007). Lava tubes on basaltic volcanoes are thoroughly described and their formation has been observed in active flows (Greeley, 1971a, 1971c, 1987; Peterson *et al.*, 1994; Calvari and Pinkerton, 1998; Kauahikaua *et al.*, 1998) or theorized through the study of cooled lava tubes (Atkinson *et al.*, 1975; Calvari and Pinkerton, 1999). Five formation mechanisms of lava

tubes by channel roofing or inflation have been theorized: (1) the rooted crust mechanism; (2) the accretional levees mechanism; (3) the slabs aggregation mechanism; (4) the pahoehoe lobe extension mechanism; and (5) the inflation lava mechanism (Greeley, 1971a, 1971c, 1987; Hon *et al.*, 1994; Peterson *et al.*, 1994; Calvari and Pinkerton, 1998; Kauahikaua *et al.*, 1998; Sauro *et al.*, 2020).

The flanks of Vesuvius are intensely urbanized with more than 300 000 inhabitants (Pesaresi *et al.*, 2008; Harris *et al.*, 2016) presently living in the path of lava flows emplaced from 1669 to 1944. On Vesuvius, lava flows from the last eruptive cycle primarily affected the western and southern flanks, occasionally reaching the coastline, more than seven kilometres away from the summit cone. Urbanization around Vesuvius continues to expand, especially in municipalities along the coastline, which are currently the most densely populated (Petrosino *et al.*, 2004). Several rural dwellings are located at the footslopes of Vesuvius, and many houses or residential complexes have been built around the half of last century along the main roads which from the towns of Ercolano and Torre del Greco reach the crater. Consequently, in the event of an effusive eruption, the risk of lava flow destroying buildings is very high. Although lava tubes are commonly observed in basaltic lava flows (Ollier and Brown, 1965; Greeley, 1987; Peterson *et al.*, 1994; Kauahikaua *et al.*, 1998), some have been observed in phonotephritic lava flows (Belkin *et al.*, 1993) of Vesuvius (Palmieri, 1859; Phillips, 1869). Malladra (1917) describes a lava tube in the 1858 lava flow that emplaced on the western slopes of Vesuvius. This lava tube was visited in 2006 by speleologists (Damiano and Del Vecchio, 2006), but no scientific investigations were carried out either on the 1858 lava flow field of Vesuvius nor on its lava tubes.

In this study, we focus our interest on the 1858 lava flow field of Vesuvius as it was produced by a long duration eruption, shows a large variety of surface features, and is up-to-now the only flow field with preserved and accessible lava tubes at Vesuvius. We present a reconstruction of the temporal and spatial evolution of the 1858 lava flow field and an analysis of its surface features. We describe the morphology and internal features of the lava tubes found in this lava flow field and analyse in detail the morphology of the largest lava tube using high-end technology. Finally, based on all the observations, we propose an interpretation of the mechanism that allowed the formation and development of the largest lava tube.

3.2. Geological setting

3.2.1 Somma-Vesuvius

The Somma-Vesuvius complex is located in the southern part of Italy, to the East of Naples in the southernmost part of the Campanian plain (Fig. 33a). The Somma-Vesuvius is a composite stratovolcano formed by (a) a pre-existing edifice, Monte Somma, that has collapsed repeatedly in consequence of highly explosive eruptions to form a composite caldera (Cioni *et al.*, 1999; Arrighi *et al.*, 2001; Cioni *et al.*, 2008), and (b) a younger cone, Vesuvius (1281 m

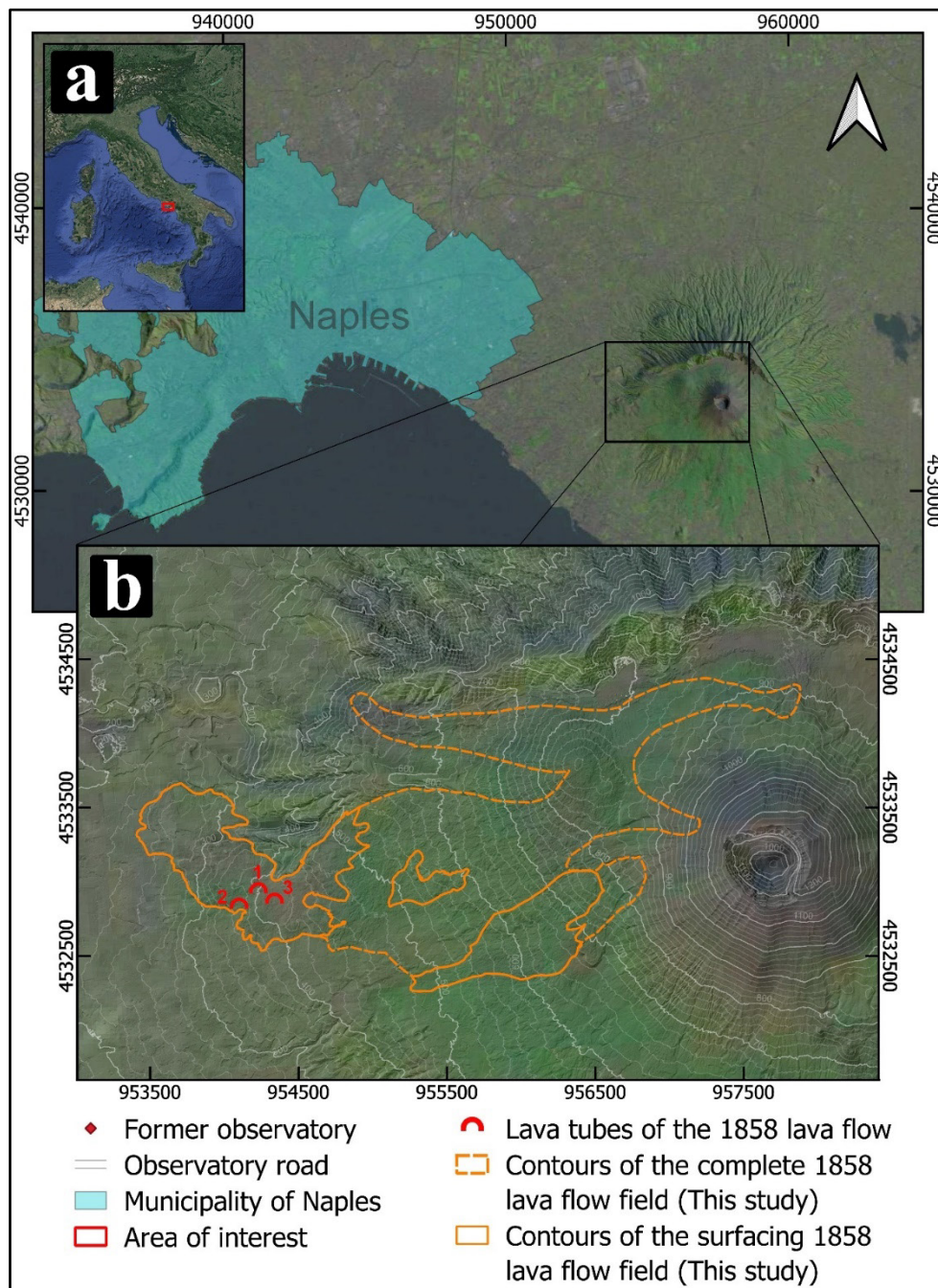


Figure 33. a) location of Vesuvius on the west coast of southern Italy, east of the city of Naples; b) location of the 1858 lava flow and its lava tubes on the west flank of Vesuvius.

a.s.l.) that grew after the AD 79 Plinian eruption (Cella *et al.*, 2007; Cioni *et al.*, 2008). Four major Plinian eruptions were identified: Pomici di Base (22,000 yr BP), Mercato (8,800 yr BP), Avellino (3,800 yr BP) and Pompeii (79 AD) (Santacroce *et al.*, 2008). The last eruptive cycle of Vesuvius started after the sub-Plinian eruption of 1631 and continued as a prolonged period of open conduit condition with mainly effusive eruptions intercalated by Strombolian and Hawaiian eruptions (Carta *et al.*, 1981; Arrighi *et al.*, 2001). Eighteen cycles of activity separated by short repose periods, maximum seven years long, were characterized by Arno *et al.* (1987). This eruptive cycle ended with the 1944 eruption that destroyed the village of San Sebastiano (Chester *et al.*, 2007; Cubellis *et al.*, 2016). Following this event the volcano entered in a phase of quiescence that continues to date. The lava flows emplaced during the 1631-1944 cycle have both 'a'ā and pāhoehoe surfaces. Lavas erupted on Vesuvius have in majority phonotephritic compositions, with few basanites and trachy-basalts (Belkin *et al.*, 1993). Lavas show a range of silica content from 47.00 to 49.30 wt% with a mean of 48.00 wt% and a high alkali content ranging from 6.42 to 11.87 wt% with a mean of 9.60 wt% (Belkin *et al.*, 1993). Lavas are porphyritic with an average of 40% phenocrysts mostly leucites and clinopyroxenes (Trigila and De Benedetti, 1993).

3.2.2 The 1858 eruption

Like many eruptive events on Vesuvius, the 1858 eruption was observed by the inhabitants of the Naples Bay but also by tourists and scientists coming from Western Europe. Several books were published on the volcanoes of Southern Italy (Hamilton, 1772; Mercalli, 1884; Johnston-Lavis *et al.*, 1891) describing the eruptions of Etna (Gemmellaro, 1819, 1844), Stromboli (Mercalli, 1881, 1888), Vulcano (Mercalli, 1888, 1889) and Vesuvius (Pigonati, 1767; Guarini *et al.*, 1855; Phillips, 1869). Geological maps depicting the extent of lava flows were made (Marsigli, 1832; Roth, 1857; Le Hon, 1866; "Monte Vesuvio," 1860s (unknown precise publication date)). Many paintings and drawings were produced by artists or scientists capturing eruptions at a certain time or shedding light on strange phenomenon such as lava tubes (Phillips, 1869). Luigi Palmieri, director of the Vesuvius Observatory at the time, described in detail the course of the 1858 eruption up to July 1859 in the annals of the Observatory (Palmieri, 1859) and resume the 1858 eruption in a report to the Accademia Pontaniana (Palmieri *et al.*, 1862). The 1858 eruption started on the 27 of May and lasted nearly three years to end the 12 April 1861 (Palmieri *et al.*, 1862; Scandone *et al.*, 1993). During this eruption, seven eruptive fissures opened, and a compound lava flow field emplaced on the western flank of Vesuvius (Fig. 33b). The lava flow field reached a maximum of five kilometres

from the vents and inundated two deep valleys. Two samples of the 1858 lavas were analysed using X-ray fluorescence (XRF) by Belkin *et al.* (1993). Lavas are phonotephritic with a mean SiO₂ content of 47.45 wt% and a mean alkali content of 9.90 wt%. Lavas of the 1858 eruption are strongly porphyritic with an average crystal content of 40% (Trigila and De Benedetti, 1993). The 1858 lava flow has a pāhoehoe surface with a variety of surface features. These inflated surface features form by local overpressures that cause crustal uplift creating a few meters high reliefs (Walker, 1991; Rossi and Gudmundsson, 1996; Duncan *et al.*, 2004; Anderson *et al.*, 2012). The lava tubes we describe in this study are found in the shallowest portions of the flow field. Other and previous lava tubes might have been covered by the 1858 last lavas or by more recent lava flows (1895-1899, 1944), thereby obscuring them from observation. Up to now, these lava tubes are unique features on Vesuvius because they are the only ones discovered, preserved and accessible.

3.3. Materials and methods

3.3.1. Historical and contemporary documents

Historical documents were used to reconstruct the temporal and spatial evolution of the 1858 eruption, the 1858 lava flow field contours and to obtain information about lava tubes. We carried out a thorough literature review and found eight geological and topographical maps from 1832 to 2016 (Marsigli, 1832; Roth, 1857; Le Hon, 1866; “Monte Vesuvio,” 1860s (unknown precise publication date); Istituto Geografico Militare, 1908a, 1908b; Santacroce and Sbrana, 2003; Paolillo *et al.*, 2016). The two maps from 1832 (Marsigli, 1832) and 1857 (Roth, 1857) show the topography or the lava flows prior to the 1858 lava flow. The two maps from the end of the 1860s (Le Hon, 1866; “Monte Vesuvio,” 1860s (unknown precise publication date)) show the 1858 lava flow before being largely covered by the 1895-99 lava flow. The four other maps were published in 1908 (Istituto Geografico Militare, 1908a), 1908 (Istituto Geografico Militare, 1908b), 2003 (Santacroce and Sbrana, 2003) and 2016 (Paolillo *et al.*, 2016).

Moreover, we analysed paintings of the 1858 eruption (Fig. 34a, b) to better reconstruct its development through time. We also used a precise description of the evolution of the 1858 eruption by the director of the Vesuvius Observatory, Luigi Palmieri, reported in the 1859 Annals of the Vesuvius Observatory. In the literature, an important paper from Malladra was

published in 1918 where the author described in detail the lava tube formed by the 1858 lava flow of Vesuvius.

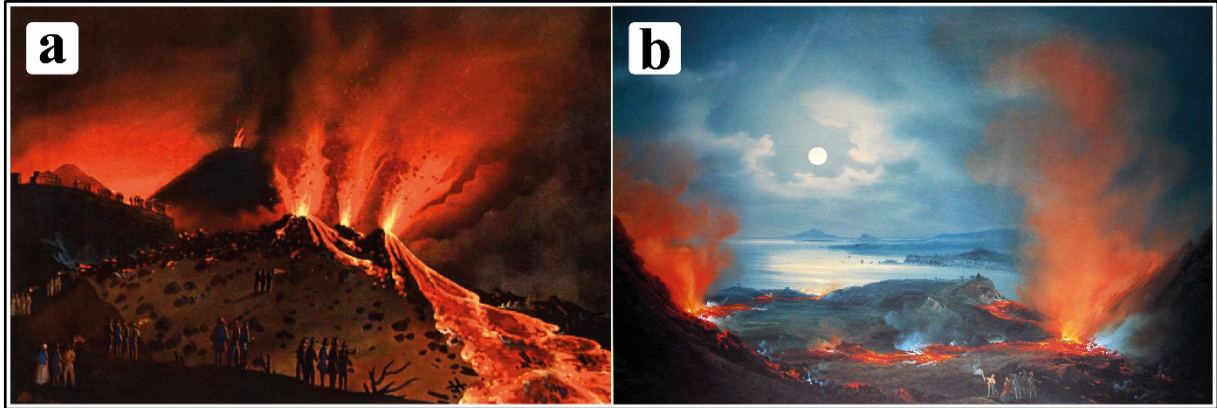


Figure 34. Paintings of the 1858 eruption. a) painting with a point of view from the observatory by an unknown artist, private collection; b) Painting *Chiaro di Luna durante l'eruzione del 1858* by Giovan Di Battista Gatti representing the 1858 eruption during a full moon, 48 hours after its beginning, from the north side of the Gran cono.

3.3.2. Topography

Two different elevation datasets were used to describe lava flow morphology, lava flow surface and lava tubes surfaces: a Digital Terrain Model (DTM) with a 1-meter resolution from the municipality of Naples (www.cittametropolitana.na.it/pianificazione_territoriale/) and a DTM covering the entire Italian territory with a 10-meters resolution from the Tinitaly project (Tarquini *et al.*, 2023) of the Istituto Nazionale di Geofisica e Vulcanologia (INGV). From the 1-meter DTM we extracted three topographic parameters: slope, relative relief, and aspect (Ventura and Vilardo, 2008). Slope and aspect were calculated using the Horn formulas (Horn, 1981). Relative relief was calculated as the difference between minimum and maximum altitude within a five pixels radius to highlight fine scale features and within a hundred pixels radius to identify large scale features. We produced precise Digital Elevation Models (DEM) of the overlying surface of the lava tube in order to describe precisely the morphology of the surface over the lava tube. The proceedings to scan this surface are described in the next section.

3.3.3. 3D scanning of the lava tube

The lava tube interior was captured using a time-of-flight terrestrial laser scanner (TLS), the Riegl VZ-400, which has an angular measurement resolution of 0.0005° (Fig. 35a). The data was acquired during field work in March 2023. Nineteen scans from different points of view were necessary to obtain a satisfactory result by minimizing shadow areas produced by the morphology of the tube (collapses, tube extensions). Reflectors of varied sizes were used for common points of reference between the scans. Twelve out of the nineteen scans made during the survey (Fig. 35b) were used to build the 3D model in RiSCAN Pro. Additionally,

photogrammetry work was performed in the northern and southeastern points of the tube using a Gopro 10 camera to complete the 3D scan obtained by the TLS.

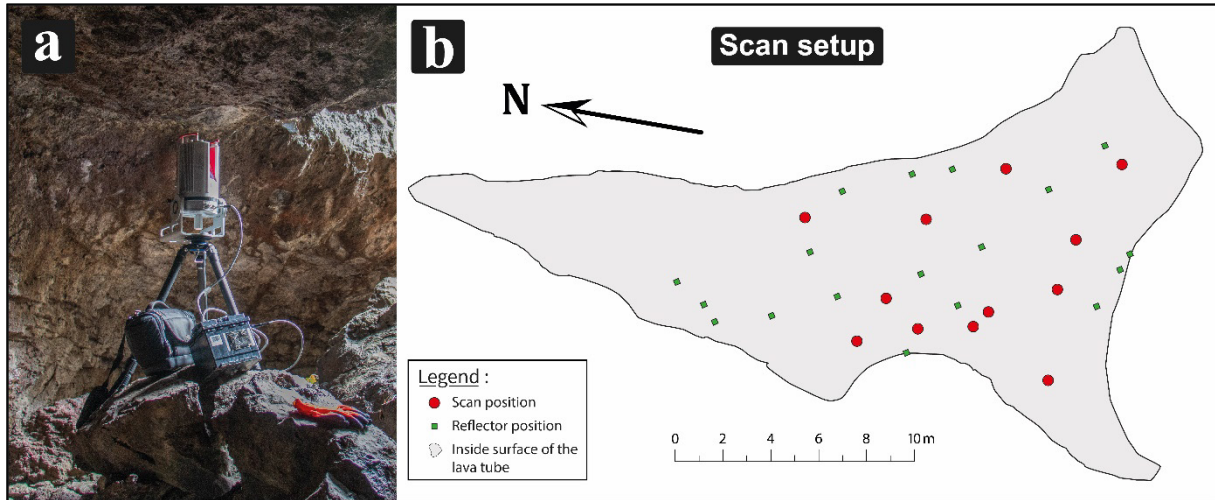


Figure 35. 3D scan setting. a) Photograph of the Riegl VZ-400 terrestrial laser scanner during the lava tube survey; b) Scan and reflector positions into the lava tube.

The overlying surface of the lava tube was scanned using a DJI Matrice 300 UAV. Markers were placed on and in the lava tube to facilitate scan alignment. Three flights were performed in July 2023. (1) A global photogrammetric survey of a large area around the lava tube was carried out using the UAV equipped with a Zenmuse P1 optical camera. (2) A low altitude flight over the lava tube using the same optical camera, and (3) a survey using the UAV equipped with a Zenmuse L1 LiDAR (Light Detection And Ranging) sensor were done. The optical data was processed in Agisoft Metashape Professional to obtain point clouds, meshes and DEMs, while the LiDAR data was treated using the DJI Terra software.

Another survey of key parts of the lava tube, the skylight and a large fracture in the roof, was conducted with a Gopro 10 camera to perform photogrammetry work. This survey allows the connection between the scanned georeferenced surface and the scanned interior. These images were transformed into point clouds and meshes with textures using Agisoft Metashape. The 3D scan of the interior of the lava tube was aligned with the georeferenced DTM surface in CloudCompare to form a complete 3D representation.

3.4. Results

3.4.1. The 1858 lava flow field

3.4.1.1. The 1858 eruption and the emplacement of the lava flows

The historical documents that describe the 1858 eruption were used to reconstruct its temporal and spatial evolution (Fig. 36). The complete contour of the 1858 lava flow field and the position of the vents were extrapolated from old geological maps (Le Hon, 1866; “Monte Vesuvio,” 1860s (unknown precise publication date); Istituto Geografico Militare, 1908a, 1908b; Santacroce and Sbrana, 2003; Paolillo *et al.*, 2016). From the descriptions of the 1858 eruption, we extracted the position and formation time of the different vents, but also the position and direction of the flow fronts and the size and speed of the lava flows at different dates.

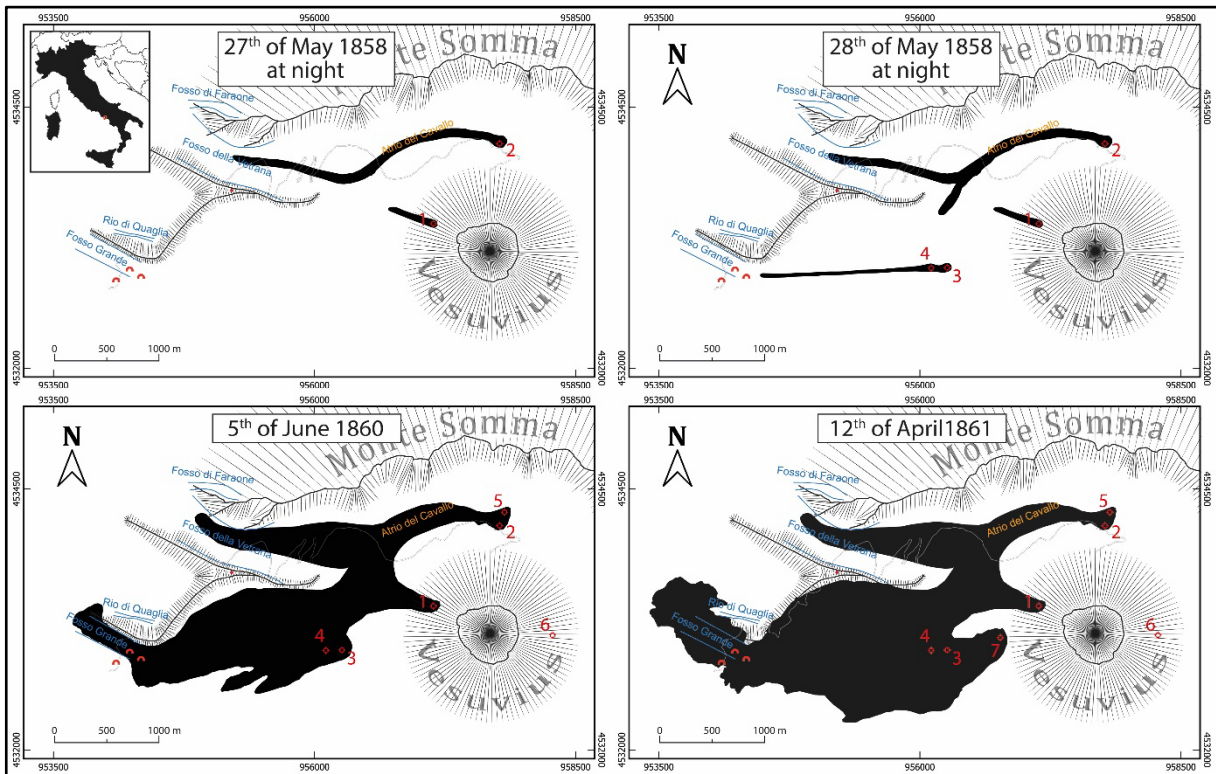


Figure 36. Sketches of the temporal and spatial evolution of the 1858 lava flow field reconstructed using historical documents, paintings and geological maps. Fissures or vents are red circles. Lava flow is solid black. Blue lines are the boundaries of the main actual and past valleys. Red diamond is the position of the former Observatory.

In the 1859 Annals of the Vesuvius Observatory, Palmieri described the 1858 eruption by saying that it started with the opening of a fissure in the west side of the Gran Cono (the cone that grew within the Monte Somma caldera, now called Vesuvius) on the 27 of May 1858 at 04:17 a.m. He observed the opening of a second fissure on the same day (precise time of opening is not mentioned by Palmieri) in the Atrio del Cavallo and at the end of the 27 of May, he saw that the lava erupting from this vent was starting to flow into the Fosso della Vetrana

(Fig. 36). Palmieri calculated that, on the 27 of May, the lava had a velocity of 2 ms^{-1} . The next day, he stated that a third fissure opened in the Piano delle Ginestre at 04:30 a.m. (Fig. 36). Soon, he observed a fourth vent opening next to the third and abundant lava flowing out of it (precise time of opening is not mentioned by Palmieri). This painting by Giovan Battista Gatti called “Chiaro di luna durante l’eruzione del 1858” represents the 1858 eruption during a full Moon seen from the north side of the Gran Cono (Fig. 34b). We determined the date on which the artist painted this representation using a mean lunar cycle of 29 days, 12 hours, 44 minutes, and 3 seconds (29.53058885 days). The painting was probably depicted during the night of Friday 28 of May 1858, about 48 hours after the beginning of the eruption. The overall extension of the flows at that time can be determined using this painting. Palmieri described the development of the eruption noting that the lava was then flowing into a ~100-m-deep valley called Fosso Grande. He described the opening of a fifth fissure next to the second one and the opening of a sixth fissure in the eastern part of the Gran Cono from which a moderate amount of lava flowed out.

Table 4. Chronology of events for the 1858 eruption of Vesuvius.

Date	Time	Event
27 May 1858	4:17 a.m.	Opening of vent 1 on the western side of Vesuvius
27 May 1858	Not mentioned	Opening of vent 2 in the Atrio del Cavallo
28 May 1858	4:30 a.m.	Opening of vent 3 in the Piano delle Ginestre
28 May 1858	Not mentioned	Opening of vent 4 next to fissure 3. Opening of vent 5 next to vent 2. Opening of vent 6 on the eastern side of Vesuvius
Not mentioned	Not mentioned	Closure of vent 6 after low emission of lava
15 June 1858	Not mentioned	Closure of vent 1,2 and 3. The eruption seemed to have come to an end but vent 4 and 5 still emit.
30 June 1858	Not mentioned	Closure of vent 5 after a month of activity.
March 1860	Not mentioned	Closure of vent 4 after nearly two years of activity, eight days later, opening of vent 7.
12 April 1861	Not mentioned	Closure of vent 7 after a year of activity, end of the eruption after 1051 days of activity.

According to Palmieri, most of the vents stopped emitting lava after a few days to a few weeks. He noticed that vents 4 and 5 were the only ones active on the 16 of June 1858 and that fissure five stopped emitting on the 30 of June 1858. Palmieri wrote in a report to the Accademia Pontaniana that only fissure 4 in the Piano delle Ginestre kept emitting lava until the end of March 1860 (Palmieri *et al.*, 1862). Eight days after the closure of vent 4, a seventh fissure opened 600 meters higher and emitted lava until the 12 of April 1861 (Palmieri *et al.*, 1862; Scandone *et al.*, 1993).

3.4.1.2. Flow margins of the 1858 lava flow

Six historical and recent geological maps from the 1860s to 2016 were used to extrapolate the flow margins of the 1858 lava flow field. When georeferenced into a GIS (Fig.

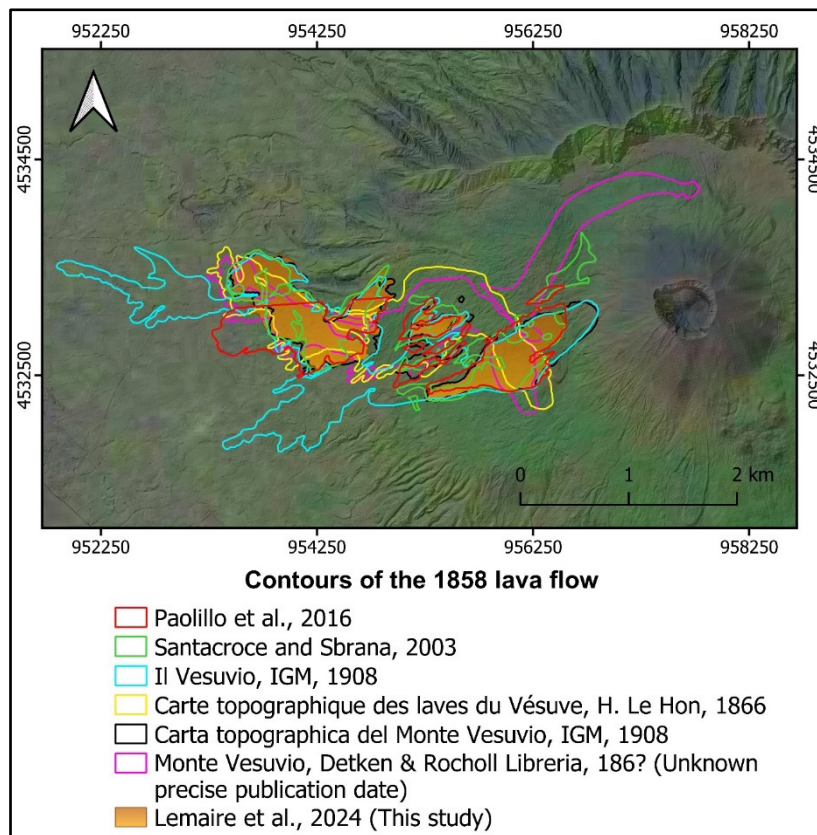


Figure 37. Representation of the flow margins of the 1858 lava flow field extracted from six georeferenced historical and recent geological maps (coloured lines) and from the results of this study (solid orange).

37), discrepancies between the covered areas appear. Some discrepancies are related to the deposition of new lava flows after 1858, some are dependent on the choices of the authors, and some are due to errors during the georeferencing of the oldest geological maps. In an attempt to overcome the addressed discrepancies amongst geological maps, new margins were redefined, one of which represents the outline of the exposed 1858 lava flow field and the other represents the outline of the complete lava flow field. Both were defined based on the geological maps, on the 1-meter and 10-meters resolution DEMs and on detailed field work (Fig. 37).

Using these new flow margins, we measured that the area of the complete flow field is 3.9 million square meters and that the exposed area is 1.3 million square meters which implies that 66% of the 1858 lava flow field is covered by other flows. Centre lines were determined using the methodology of Kereszturi *et al.* (2016). A minimum flow field length of 3.069 kilometres and a maximum of 5.207 kilometres were measured. We measured the width of the lava flow field every hundred meters along the centre lines, the mean width being 0.607 kilometres with a minimum width of 0.116 kilometres and a maximum width of 1.496 kilometres. Volume and thickness of the 1858 lava flow field are not measurable but will be estimated in the following sections.

3.4.1.3. Topographic parameters and surface features

Using the 1-meter DEM of Vesuvius, a surface analysis of the 1858 lava flow field was carried out. Three topographic parameters were used to describe the still exposed part of the 1858 lava flow field (Fig. 38): slope, relative relief, and aspect (Ventura and Vilardo, 2008). The slope of the surface is comprised between 0.01° and 84.41° with an average of 15.87° . Three breaks in slope are visible in the western part of the lava flow field. The relative relief calculated using a five meters radius shows a large number of very fine-scale features with a mean difference of altitude of 0.17 meters. These correspond to small tumuli or tilted slabs of lava. On the other hand, the relative relief calculated using a one hundred meters radius shows tens of large-scale features. These correspond to large tumuli a few meters high or dome-shaped features up to sixty meters in diameter. These large surface features are mostly present in the

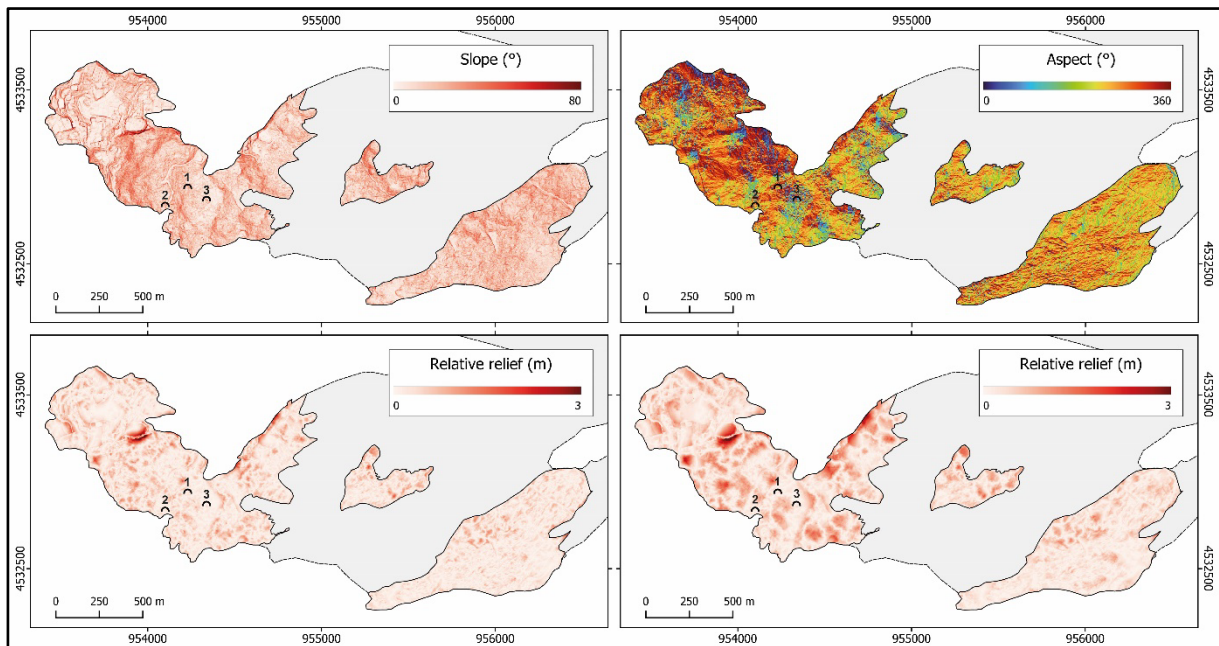


Figure 38. Surface parameters for the exposed areas (uncovered by more recent lava flows) of the 1858 lava flow field. Top left, slope. Top right, aspect. Down left, relative relief with a radius of 5 meters. Down right, relative relief with a radius of 100 meter.

western distal part of the lava flow field, while the eastern proximal part has a flatter surface (Fig. 38). Similarly, the aspect shows these large-scale features but also the overall south to north south direction of the lava flow. We can observe more surface features in the distal part of the lava flow field, but the description of the lava flow field surface is made difficult by covering of more recent lava flows and by the presence of dense vegetation.

3.4.1.4. Estimation of the emitted volume and eruption rates

Using the historical description of the 1858 eruption (Palmieri, 1859), we know the opening and closing times of the eruptive vents (Fig. 39). Vents 1 and 2 opened the first day followed by vents 3 and 4 the second day and by vents 5 and 6 the fourth day. Most of the vents stopped emitting lava after a few days to a few weeks. Vents 1, 2 and 3 stopped erupting on the 15 of June 1858, twenty days after the beginning of the eruption. Vent 5 stopped emitting lava on the 30 of June 1858, after thirty-two days of activity. The closing of vent 6 was not described but probably occurred a few days after its opening as Palmieri wrote that a moderate amount of lava was emitted from this vent. Only vent 4 kept emitting lava until March 1860. Vent 7 opened eight days after the closure of vent 4 and was the only vent to emit until the end of the eruption.

Palmieri estimated an erupted volume of 120 million cubic meters of lava in July 1859, with 36 million cubic meters emitted from vent 4. This value is inevitably underestimated because the eruption stopped more than eight months later. If we consider the time of emission of this volume from the 27 of May 1858 to the 1 of July 1859 (400 days) and use the eruption rate equation:

$$Q = \frac{V(t)}{t} \quad (24)$$

then the mean eruption rate during this period is $3.47 \text{ m}^3\text{s}^{-1}$ (Fig. 39). We calculated the eruption rate at vent 4 which is $1.04 \text{ m}^3\text{s}^{-1}$ for the same period. Using the average eruption rate of the whole flow field from day 1 to day 400 and the eruption rate of vent 4 for the rest of the eruption, we estimated the total volume of the lava flow field at the end of the eruption (after 1051 days of eruption) to be 178 million cubic meters. The mean 1858 flow field thickness is then equal to 45.85 meters. The thickness of the flow field can seem high compared to other flows at Vesuvius like the 1944 lava flow that has a mean thickness between 6 to 17 meters (Ventura and Vilardo, 2008). This difference could be explained by the pre-emplacement topography, consisting in deep valleys cut within pyroclastic deposits, by the compound morphology of the

flow field and by the long duration of the eruption. Based on the calculated volume and duration of the eruption we estimate the mean output rate to be $1.97 \text{ m}^3\text{s}^{-1}$.

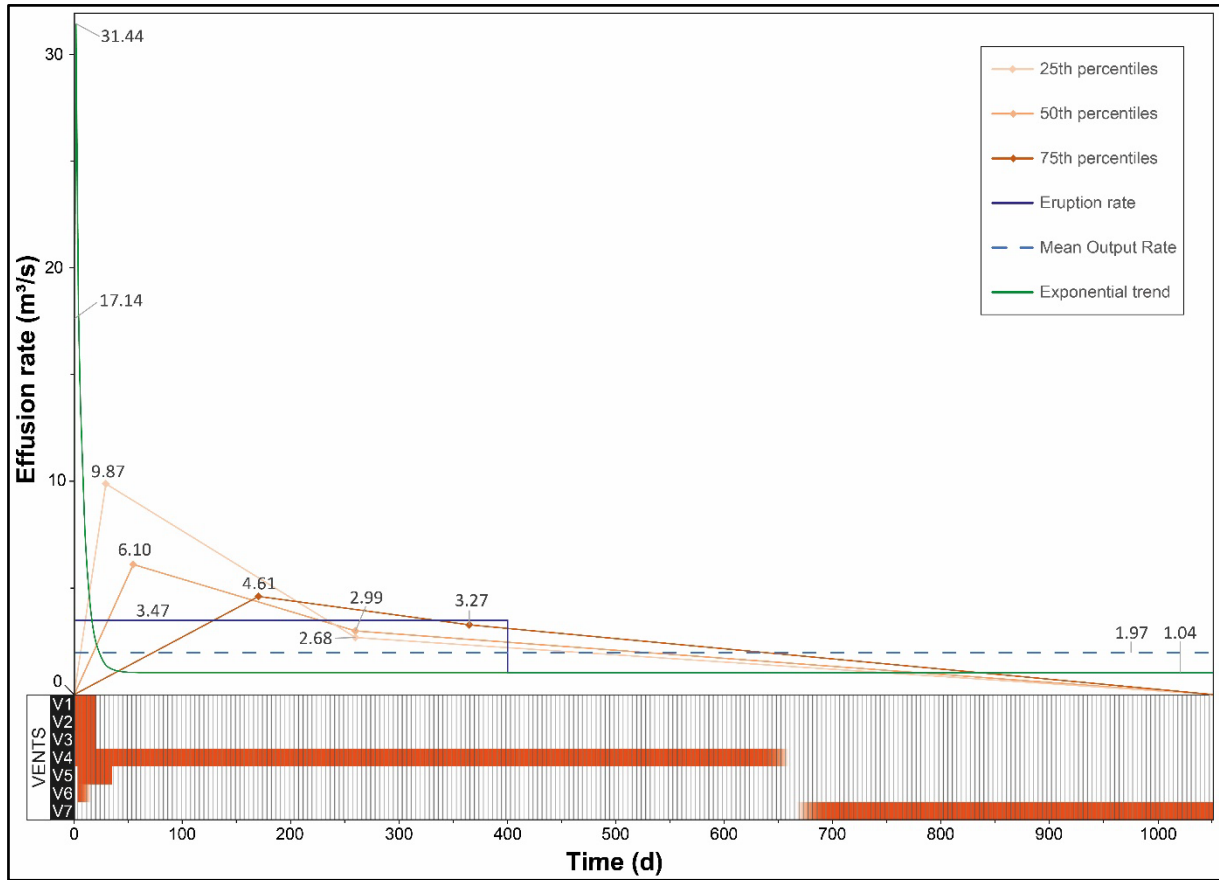


Figure 39. Opening and closing time of the eruptive vents of the 1858 eruption and estimated eruption rates. Eruption rate estimated based on the volumes of Palmieri (solid blue line), mean output rate of the 1858 eruption (dashed blue line), characteristic effusion rates for the 1858 eruption based on the characteristic curves from Zuccarello *et al.*, 2022 (solid orange lines) and exponential effusion rate trend based on the estimated volume and mean output rate of the 1858 eruption (solid green line).

To have an idea of the emplacement dynamics, it is necessary to define effusion rate trends, namely both the trends based on eruption duration and a trend based on initial effusion rates, estimated volume and mean output rate. We used the characteristic effusion rate curves identified on Etna by Zuccarello *et al.* (2022; Fig. 39) as for lava flow morphology, erupted volumes, and emplacement on steep slopes, eruptions at Etna are similar to those that occurred during the 1631-1944 cycle of Vesuvius, thus making the work of Zuccarello *et al.* (2022) a good proxy to estimate eruption rates. There are three curves corresponding to the 25th, 50th and 75th percentiles of the occurrence of effusion peaks. These curves have classic waxing and waning phases. Applied to the 1858 eruption, these characteristic curves yield instantaneous effusion rates attaining maximum effusion peaks of 4.6 to $9.9 \text{ m}^3\text{s}^{-1}$ after 29 to 170 days of eruption and diminishing to 2.7 and $3.3 \text{ m}^3\text{s}^{-1}$ (respectively) after 260 to 365 days.

We defined an exponential curve to represent a waning phase after the effusion peak using the equation from Wadge (1981; Fig. 39):

$$Q_t = Q_0 e^{-\lambda t} \quad (25)$$

in which Q_0 is the initial effusion rate and λ a constant. Firstly, we calculated the effusion rate at each vent after 15.5 and 39.5 hours from the beginning of the eruption using the equation of Wadge (1978) that is based on the length of the lava flow:

$$Q = 4.8L \quad (26)$$

This equation yields a total effusion rate of $17.14 \text{ m}^3\text{s}^{-1}$ at 15.5 hours and $31.44 \text{ m}^3\text{s}^{-1}$ at 39.5 hours. We modified equation (25) as:

$$Q_t = Q_p e^{-\lambda t} + c \quad (27)$$

where Q_p is the effusion rate at its peak, and c a constant. Constrained by the volume of the lava flow field and the mean output rate of the eruption, equation 27 becomes:

$$Q_t = 31.44 e^{-(0.154t)} + 1.04 \quad (28)$$

3.4.2. Lava tubes in the 1858 lava flow field

3.4.2.1. Description of the lava tubes

We found three lava tubes in the distal part of the 1858 lava flow field (Fig. 33b), each showing a different size and morphology. The first one is a small lava tube, labelled 1 in Figure

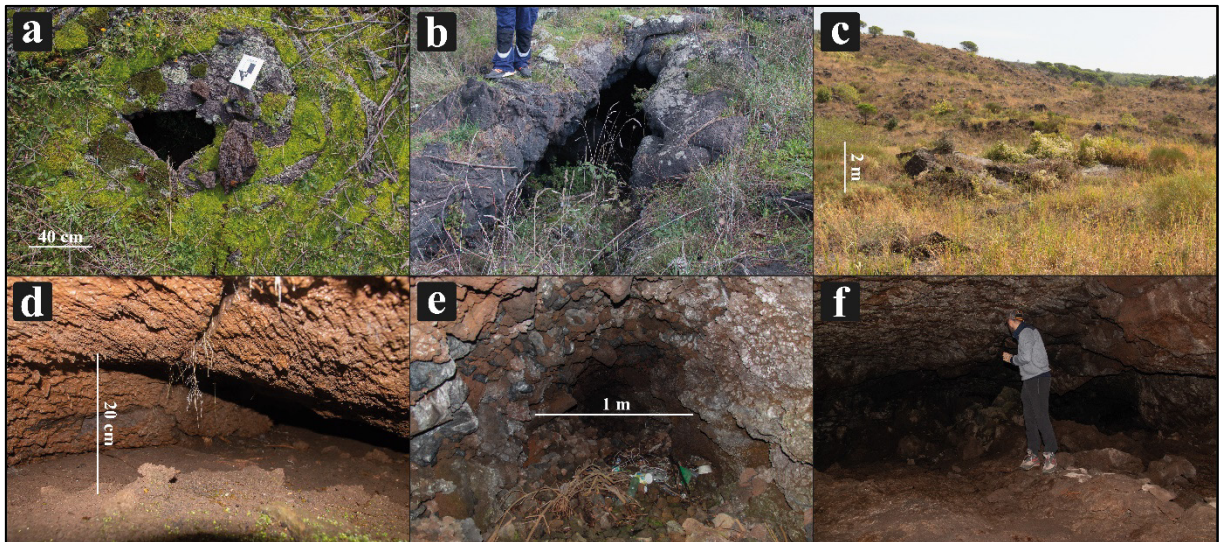


Figure 40. Photographs of the three lava tubes found in the 1858 lava flow field. a) Exterior surface of tube 1; b) Exterior of tube 2; c) Exterior surface of tube 3; d) Interior of tube 1; e) Interior of tube 2; f) Interior of tube 3.

33, located next to a tumulus, and formed in one of its pahoehoe lava outpourings (Fig. 40a). Its roof is formed by two thin layers of ropy pāhoehoe lava and is a few centimetres thick. This small lava tube has a twenty centimetres internal diameter (Fig. 40d). The second, labelled 2 in figure 33, is located in a steep part of the 1858 lava flow field. It is an early-stage lava tube that was probably active for a few hours to a few days. It is five meters long and nearly two meters high (Fig. 40b). Inside we observed 'a'ā rubbles fringed by squeezed out toothpaste pāhoehoe (Fig. 40e). The third one, labelled 3 in figure 33 (Fig. 40c, f), is the largest found in the 1858 lava flow field and possibly the largest ever found at Vesuvius. It is located in a very flat area of the flow field. Its morphology is described in detail in the next section.

3.4.2.2. Morphology of the largest lava tube

On the different Digital Elevation Models showing the 1858 lava flow field, the largest lava tube appears as a triangularly shaped relief one to two meters high oriented north south (Fig. 41). A skylight, by which it is possible to enter the lava tube, appears as a small depression. The high-resolution DEM of the lava tube surface produced by UAV photogrammetry shows the slabs of ropy pāhoehoe lava forming the roof of the lava tube and the fractures separating them. The exterior of the lava tube has a strong morphological resemblance to a tumulus.

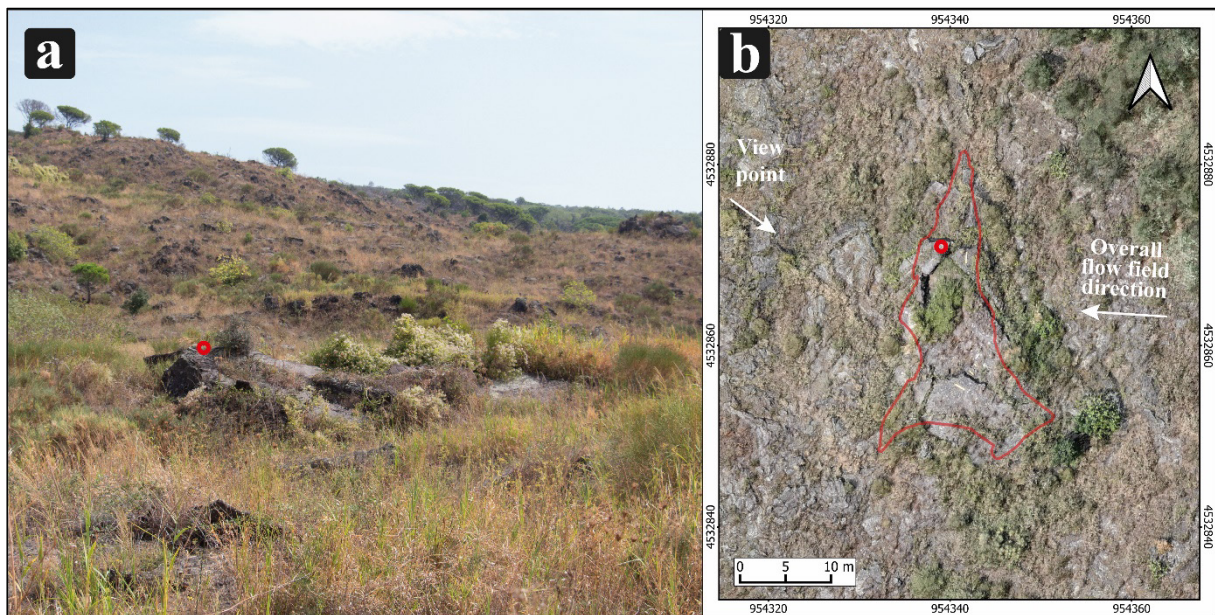


Figure 41. Lava tube surface. a) Photograph of the lava tube surface from the north west with a reference point (red circle); b) colorized DEM with the contour of the lava tube (solid red line) and a reference point (red circle).

Inside, the lava tube (Fig. 42) also has a triangular shape with a length of 30.15 meters (A-R) in the north south direction and a width that varies from 1.20 meters at the northern point (A) to a maximum of 17.41 meters in its southern part (S-B). We measured a mean inside height of 1.40 meters with a maximum height of 2.47 meters in the southern part of the lava tube (Fig. 42). Flow direction in the lava tube was from north to south. The lava entered by a twenty-five

centimetres wide vent and exited the lava tube to the southwest by a small passage at point C (Fig. 42). A section of 34.46 meters along that flow path (A-M-C) showed an average slope of 4.43° . The slope inside the tube from B to S is 2.80° . We defined a plane by taking three points at the tips of the tube (A, B and C) and calculated an overall slope of 5.05° to the southwest. We measured the internal floor surface of the lava tube being 222.6 m^2 and the actual volume of the lava tube being $328 \text{ m}^3 \pm 41 \text{ m}^3$. An area of the roof collapsed forming a skylight, from where it is possible to enter the lava tube. In addition, several blocks fell from the roof covering some sections of the floor of the lava tube. This is especially true under the east west fracture at point M (Fig. 42) where a large pile of rock debris is present. For every vertical section, we measured the roof thickness at distances of one meter. The mean roof thickness is 2.40 meters for the A-C section, 2.15 meters for the A-R section, 2.73 meters for the S-B section and 2.19 for the E-T section (Fig. 42). The overall roof thickness is 2.37 meters with a thinner area of 0.74 meters and a thicker area of 4.29 meters.

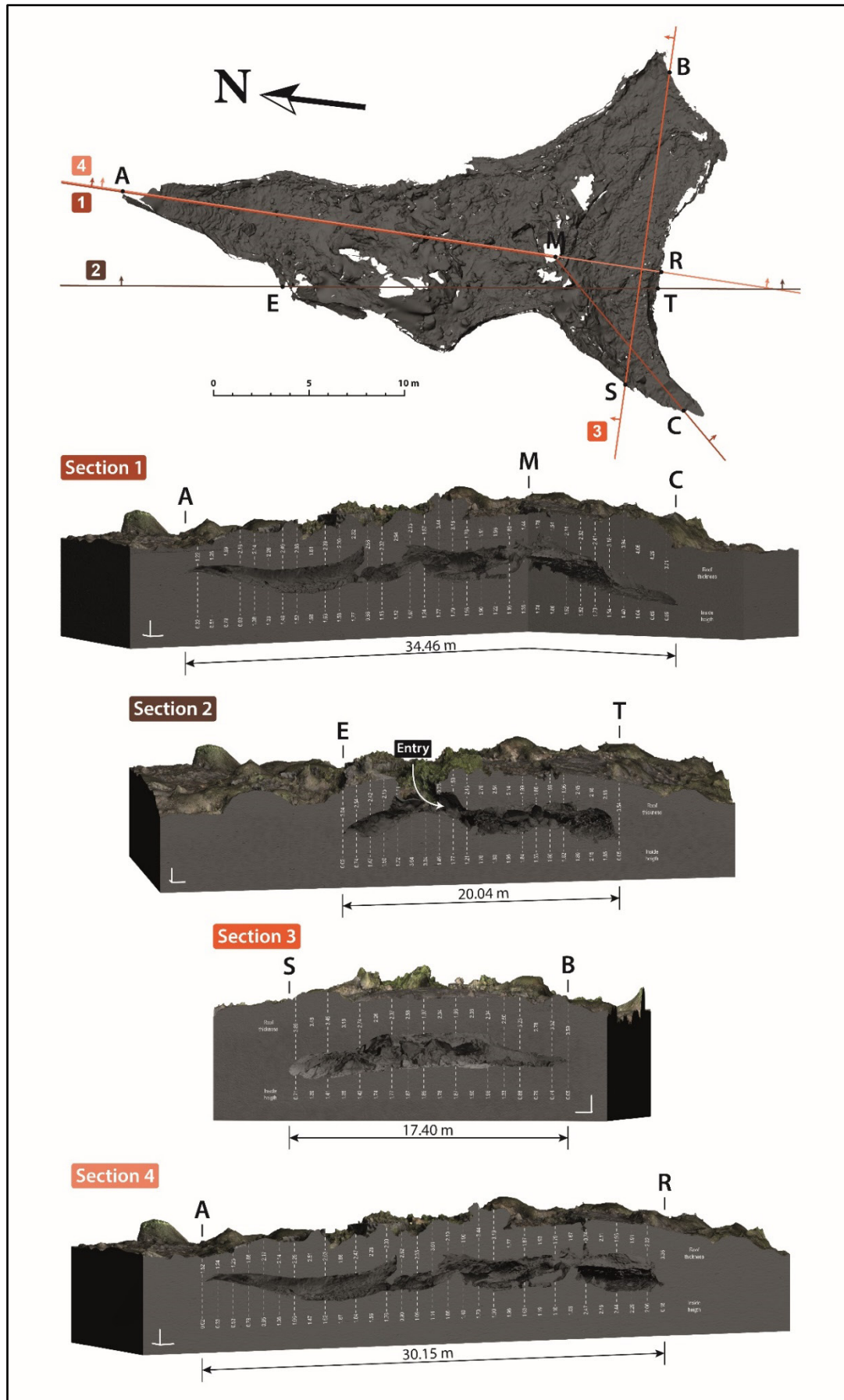


Figure 42. Four axonometric 3D sections and cross-sections of the lava tube with measured roof thickness and inside height values every meter.

Taking into account the resources available at the time, Malladra masterfully described the overall morphology of the lava tube in comparison with our high-resolution model (Fig. 43). However, few differences appear when we compare the same sections as Malladra drew and using the same points of reference. One of the main differences is at the southern end of the lava tube where the eastern extension is narrower in the drawing of Malladra. On his sections, he measured 33 meters for the A-R distance, which is 3 meters longer than our measurement. Malladra measured 42.80 meters for the flow path A-M-C, which is more than 8 meters longer than what we measured. B-C length is 24 meters on his drawings, more than 6 meters longer than the length we measured. Malladra measured a difference of altitude between point A and point C of 3 meters, corresponding to a slope of 4.01° , which is very close to the 4.43° that we measured on the 3D model. These differences show the importance of new morphology analyses using high-end technologies in order to obtain extremely detailed measurements.

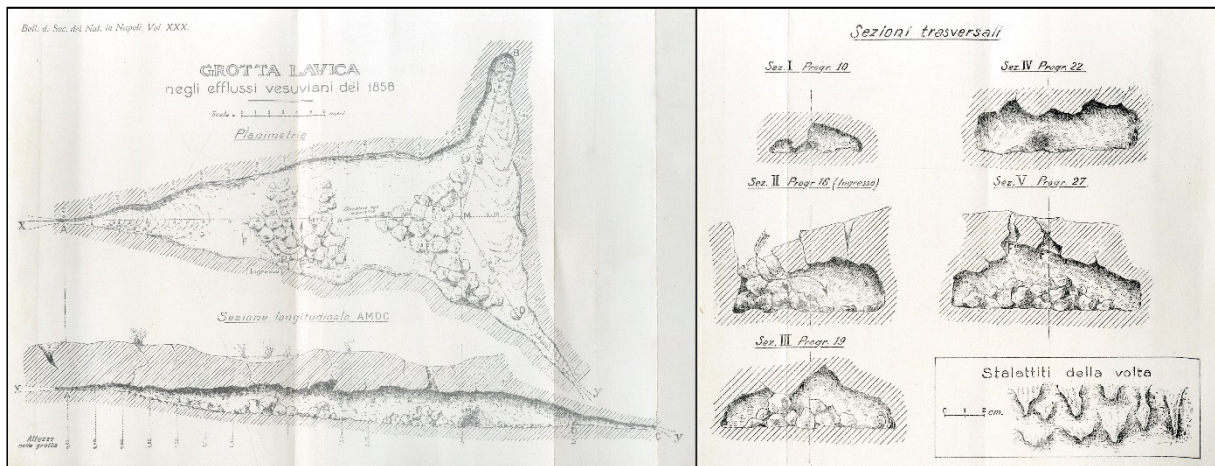


Figure 43. Sections and cross-sections of the lava tube from Malladra (1918).

3.4.2.3. Internal features

Inside the lava tube we observed several features showing evidence of prolonged lava flowage (Fig. 44). Where parts of the wall collapsed, we observed that the lava tube is lined with at least twelve, millimetres to centimetres thick, layers of lava (Fig. 44a). The thickness of those linings depends partly on the rheology of the lava that filled the tube, and the number of linings depends on the number of times the lava tube filled and drained (Calvari and Pinkerton, 1999). The last lining is five to seven centimetres thick indicating that a higher viscosity lava flowed into the lava tube before it ultimately emptied completely. In certain areas, this wall lining detached from the wall and rolled down on itself forming rolls on the sides of the lava tube (Fig. 44b). It indicates that the lava tube emptied rapidly while this lining was still plastic, in contrast with the outer linings that were already rigid and stable. In other areas, we can

observe vertical grooves in the wall lining (Fig. 44c). This indicates a vertical drop of solid lava that marked the plastic wall lining when the flow suddenly dropped inside the lava tube flowing at a lower level. Gas blisters are visible on the walls of the lava tube where the last lining detached or collapsed (Fig. 44d, e). Their diameters are comprised between 4 and 10 centimetres. Gas blisters are remnants of gas bubbles that were present in between linings. Some of them indicate that the gas bubble was still surrounded by molten lava when the lining detached leaving thin glassy lava. The ceiling of the lava tube has large bulbous in shape features on its surface (Fig. 44f). On the roof at the northern and southwestern points of the tube (A and C, Fig. 42), small stalactites with a maximum length of two centimetres are present (Fig. 44g). These stalactites have smooth surfaces and are covered by a thin light brown glassy film. They have a cone-like shape and form along small ridges aligned in the flow direction. This type of stalactite is believed to be formed by remelting of the lava by hot gases accumulating at the roof of the tube (Peterson *et al.*, 1994; Kauahikaua *et al.*, 1998). On the walls of the lava tube, we also observed pull-apart stalactites where the lining detached. Those stalactites form by the stretching of still plastic lava in between two linings that confers them a spiky morphology (Calvari and Pinkerton, 1999; Corsaro *et al.*, 2005). Their length is less than half a centimetre. Another type of stalactite can be observed on the walls of the lava tube as thin horizontal tubular features with a length up to half a centimetre (Fig. 44h). From a conical base, a thin tube goes out and at the tip of this kind of stalactite is usually present a small spherical piece of lava. Where a part of the wall collapsed, we can observe the coating of a portion of a block with fluid lava (Fig. 44i). The fissure that detached this block was formed before the lava tube emptied. The last lava that entered the lava tube via the entry point (Point A), flowed and solidified as a ropy pāhoehoe flow (Fig. 44j). It emplaced 6.60 meters inside the lava tube and has a maximum width of 1.40 meters. It is confined by 0.20 meters wide levees.

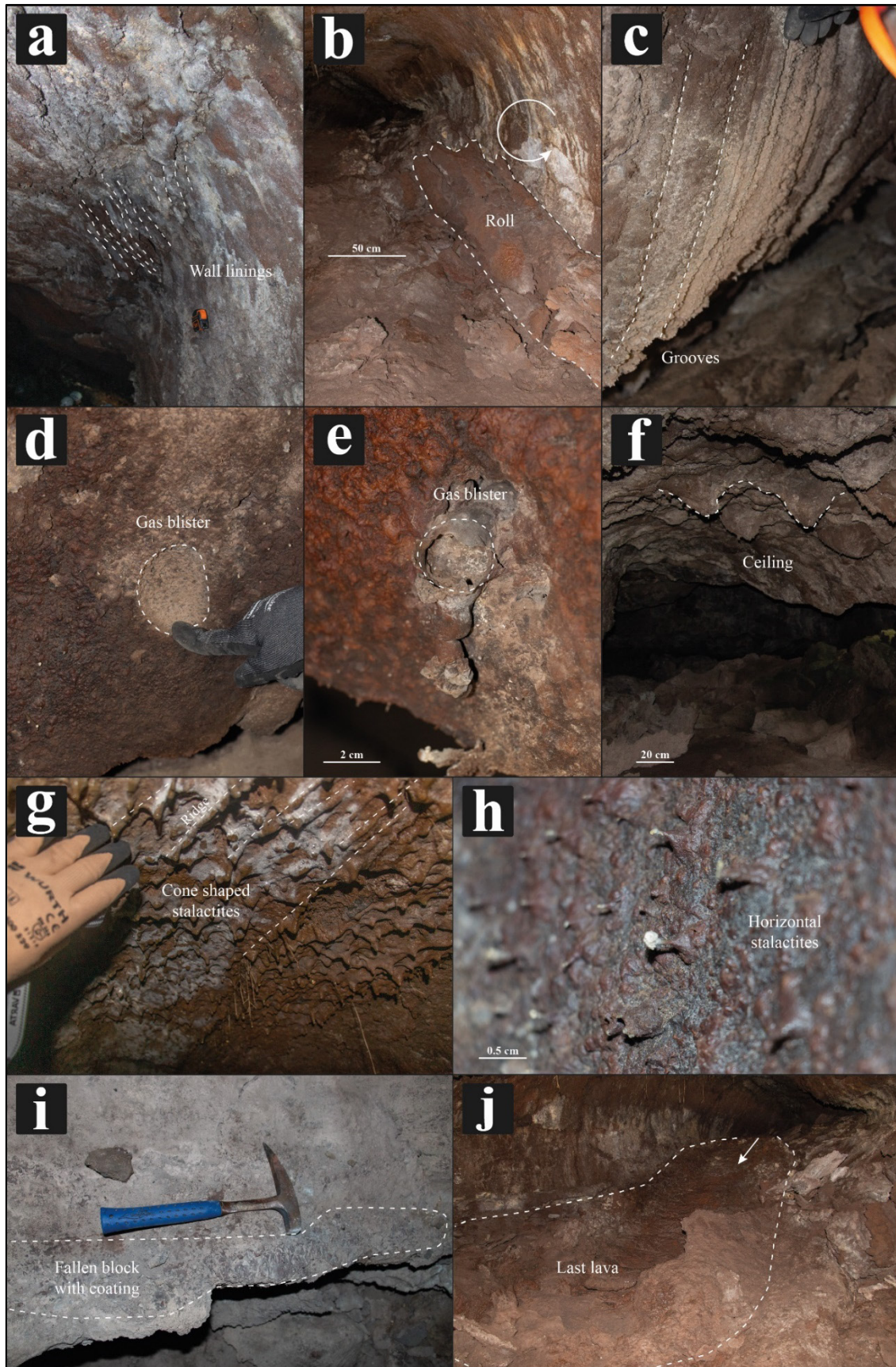


Figure 44. Photographs of the observed internal features. a) Wall linings; b) Rolled down wall lining; c) Vertical grooves; d) Imprint of a gas blister; e) Gas blister; f) Bulbous in shape features on the lava tube ceiling; g) Cone shaped stalactites; h) Horizontal tubular stalactites; i) Lava coating on a fallen block; j) Last lava entering the lava tube.

3.5. Discussion

This study gives insights into lava flow emplacement and morphologies of lava tubes on Vesuvius. We conducted a spatial and temporal reconstruction of the 1858 lava flow field emplacement on Vesuvius, a detailed morphological and surficial analysis of the 1858 lava flow field and a complete morphological analysis of its associated lava tubes.

The 1858 eruption was a fountaining event with seven eruptive vents that formed a compound lava flow field that inundated deep valleys on the western flank of Vesuvius. The low eruptive rates and long duration of the 1858 eruption, as well as the confined emplacement topography contributed to the compound morphology of the lava flow field (Anderson *et al.*, 2005). Most of the 1858 lava flow field shows a pāhoehoe surface with features that are evidences of inflation within the last flow-unit where the molten lava uplifted its overlying crust. These inflation structures are tumuli that inform us about local overpressures and flow accumulations that happened during lava flow emplacement. The western part of the 1858 lava flow, furthest from the vents, was much more affected by these processes.

After one hundred years, we reconsidered the presence of lava tubes in the 1858 lava flow field of Vesuvius and scientifically studied one of them using modern technologies. The 1858 lava flow has a phonotephritic composition, demonstrating the possibility of lava tube formation and development in volcanic systems that erupt more alkaline and highly porphyritic lavas such as Vesuvius.

We presented a detailed morphological analysis of the largest lava tube and its overlying surface using a complete 3D model of the area. In term of shape and size, the studied lava tube is much smaller and has a different morphology than lava tubes in basaltic settings such as Hawaii, Canary Islands, Iceland or Etna where lava tubes are hundreds to thousands of meters long, tens of meters wide, and shaped like meandrous rivers (Ollier and Brown, 1965; Greeley, 1971a, 1971c; Hróarsson and Jónsson, 1991; Peterson *et al.*, 1994; Kauahikaua *et al.*, 1998; Calvari and Pinkerton, 1999; Tomasi *et al.*, 2022; Calvari *et al.*, 2024). On Vesuvius, this lava tube is relatively small in size, measuring only 30 meters in length, and has a triangular shape typical of a pāhoehoe breakout. A more reasonable comparison can be made with lava tubes formed at Mount Etna. One lava tube formed by pāhoehoe lavas at Mt. Etna is the 105 meters long Mt. Arcimis lava tube (Calvari *et al.*, 2024). While being small in size like the studied tube at Vesuvius, the Mt. Arcimis lava tube has a near tubular braided morphology. The morphology of the lava tube at Vesuvius shows signs of lava inflation processes, not as the succession and

connection of pāhoehoe lobes (Crown and Baloga, 1999) but as a small breakout that inflated (Byrnes and Crown, 2001; Patrick *et al.*, 2017). The surface of the tube is composed of uplifted slabs of pāhoehoe lava. Its surface morphology strongly resembles the morphology of a satellite tumulus (Duncan *et al.*, 2004), unlike the Mt. Arcimis lava tube overlaid by lava channels, inflated flows and tumuli (Calvari *et al.*, 2024). These characteristics interrogate us on the way the lava tube at Vesuvius formed and developed.

Inside the lava tube, we observed internal features showing that lava flowed within the lava tube from north to south. Stalactites formed by remelting of the ceiling by hot gases indicate a prolonged lava flowage into the tube. The multiple layers of wall lining also indicate that the lava filled and drained the tube several times (Calvari and Pinkerton, 1999), possibly feeding secondary flows and vents. In addition, wall lining thickness increases from the outer to the innermost portions of the lava tube, testifying an increase in the viscosity and thus a change in the rheological behaviour of the lava flowing inside the lava tube. The last wall lining shows that the lava tube drained suddenly while the wall lining was still plastic allowing it to detach from the wall and roll down on itself to form rolls on the lava tube floor and small spiky horizontal stalactites on the wall. These internal features showed the evolution of the conditions inside the lava tube, its formation and development.

An interpretation of the mechanism that formed the lava tube is in Figure 45. (1) Initially, a small lava flow of a few tens of meters emplaced. The outer part cooled rapidly and formed a crust, while the internal part stayed hot. The emplaced lava flow was soon under compressive forces as its front and sides cooled down and solidified. (2) The lava flow inflated and the cooled surface crust started to lift. Fractures appeared in the cooled crust forming slabs of pāhoehoe lava. (3) A sustained lava supply allowed the support and consolidation of the fractured roof. (4) The lava tube emptied partially or completely coating the walls and ceiling with lava. (5) The supply of lava increased, and the lava tube was filled again by lava. This process of emptying and refilling of the lava tube took place multiple times. Every time it emplaced a new layer of lava on the walls and ceiling forming numerous wall linings. Throughout time the rheological properties of the lava changed, increasing its viscosity. (6) After a long-lived activity, the supply of lava stopped, and the lava drained out of the lava tube rapidly, leaving a pāhoehoe flow surface and an empty cavity.

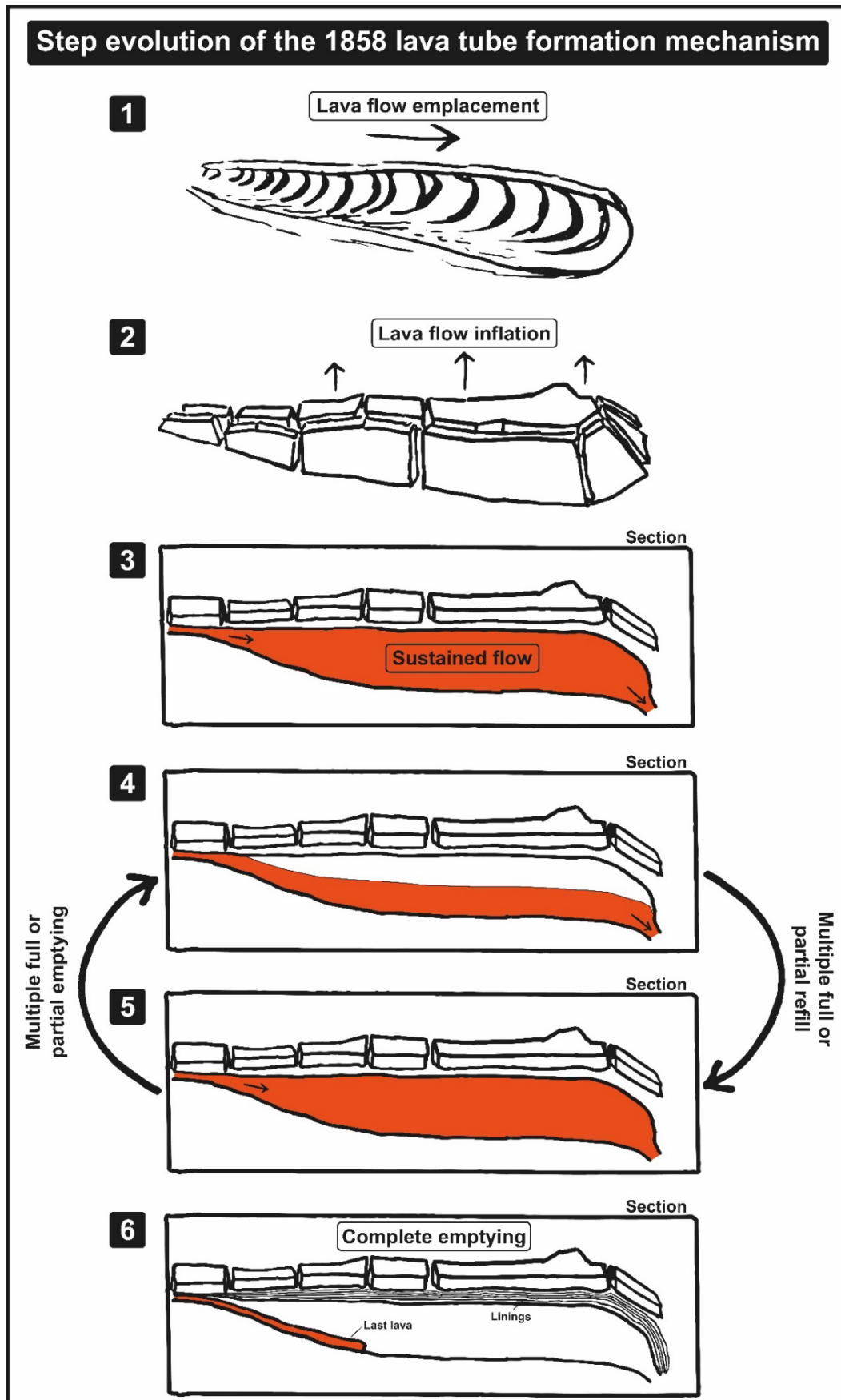


Figure 45. Step evolution model of the lava tube formation mechanism proposed for tube 3 based on the lava flow inflation process.

The study of lava tubes is essential to comprehend how lava is transported in lava flows as it affects their distance of emplacement. We described a lava tube that was able to transport lava to a limited distance. However, it must be stressed that this tube is located in the shallowest part of the lava flow field and transported lava for a time long enough to inflate, fill and drain several times. Thus, it could be inferred that larger tubes were present within the lava flow field at a deeper level, feeding the distal margins of the flow field. In addition, it is here worth noting that the volcanic and anthropogenic context in which this lava tube is situated can be even distressed by a limited reach. The highly inhabited flanks of Vesuvius are easily affected in case of lava flow emplacement and the effect of a lava tube as we described in this study can have a significant impact in the reach of damageable areas. Knowing the conditions of lava tube formation and their possible ways of development is fundamental to evaluate and respond adequately to the hazard during or even prior volcanic crises, especially in volcanic areas where constructed and agricultural lands are very close to the eruptive activity.

3.6. Conclusion

On Vesuvius, the effusive processes are not well studied, given that most of the scientific literature concentrated on the past explosive activity. Our study is one of the first in several decades to focus on the effusive activity and its processes at Vesuvius. This study focuses on the morphological description of the 1858 lava flow field and on the morphologies and mechanisms of formation of lava tubes in the same lava flow field at Vesuvius.

The reconstruction of the development of the 1858 lava flow field at various stages was inferred from historical documents. This reconstruction shows the dynamics of lava flow emplacement which helped understand the formation and development of a lava tube in that lava flow.

Through the digitization and description of this lava tube we showed the presence of lava tubes in volcanic systems in which effusive phases were fed by alkali-rich magmas, different from the typical basaltic magmas. The largest lava tube found in the 1858 lava flow field was re-examined after more than 100 years with the aid of new technologies. The complete 3D scan allowed the study of the lava tube morphology. The presence of internal features such as multiple inner coatings and lava stalactites is evidence that lava flowed into the cavity and therefore that this cavity could be described as a lava tube. Internal features are also important to know and understand the thermal and rheological history of the lava flowing inside the lava

tube. The shape in plan view of this lava tube is explained by its origin, namely the emplacement of a pāhoehoe breakout eventually drained. The formation of this lava tube is mainly explained by the process of lava flow inflation (Hon *et al.*, 1994).

In the frame of hazard assessment at Vesuvius, this study is the first step into getting knowledge about the presence and mechanism of formation of lava tubes in the lava flows of Vesuvius. The possible formation of lava tubes, even of limited size, has to be taken into account when forecasting the reach limit of lava flows as it can quickly affect populated and built-up areas.

CHAPTER 4

Crystallisation and thermo-redox conditions governing lava tube formation and development during the 1858 eruption at Vesuvius

(Submitted to Bulletin of Volcanology)

Abstract

Lava flows pose a significant volcanic hazard for communities living near volcanoes. Although usually not life-threatening, lava flows can cause substantial damage to infrastructure, leading to severe economic and social consequences. The distance a lava flow can travel is often extended by the formation of efficient transport structures, such as lava tubes. Tubes are well-insulated, limiting heat loss to the atmosphere and allowing molten lava to remain at high temperatures over long distances. Understanding the physical and chemical conditions during lava tube formation and development is essential – not only for comprehending lava flow dynamics but also for improving hazard assessment.

Here, we combine synchrotron X-ray computed microtomography, petrological analysis and thermodynamic modelling to investigate the conditions that controlled crystallisation and rheological evolution of the lava flow and lava tubes by examining natural samples collected along the 1858 lava flow field and inside a lava tube at Vesuvius. Our results indicate that variations in groundmass crystallisation were driven by differences in cooling conditions, ranging from 1120 ± 20 °C to 970 °C, during both the emplacement of the lava flow field and development of the lava tube.

Crystallisation kinetics are mostly affected by the degree of undercooling and the shear rate affecting the lava during emplacement. The morphology and groundmass crystallisation of the 1858 lava flow field suggest an emplacement dynamic that consists of piling up of small lava flows, explaining the short emplacement distance of the flow field. The variation in crystallinity within the groundmass of the wall linings indicates that the thermo-redox conditions inside the lava tube were not stable during its activity. Shear rate and cooling by conduction control the formation of wall linings. The evolution of the cooling conditions inside the lava tube affects its structure and the lava flowing toward the flow front. Constraining these variations within lava tubes will help improve models that simulate lava tubes or lava flows, the creation of hazard maps, and response strategies during an eruptive crisis. Constraining the structure of a lava tube and determining its evolution throughout the tube activity helps improve stability models used to assess the risk of collapse, both for terrestrial and extraterrestrial volcanoes.

Keywords: lava flow field; volcanic hazard; flow emplacement; cooling; oxygen fugacity

4.1. Introduction

Predicting eruptive scenarios at high-risk volcanic systems remains a significant challenge, particularly for Somma-Vesuvius due to its proximity to densely populated urban areas. The stratovolcano is known for its powerful explosive eruptions, particularly for the 79 CE Plinian eruption that destroyed the cities of Pompeii and Herculaneum (Santacroce *et al.*, 2008), but also for its last eruption in 1944 (Chester *et al.*, 2007; Cubellis *et al.*, 2016). These explosive eruptions are the most hazardous and threaten nearly a million inhabitants living within the red zone (Pesaresi *et al.*, 2008; Harris *et al.*, 2016; ISTAT, 2023; Italian Civil Protection, 2024), but the volcano also produced several effusive eruptions in the past, and especially during the last eruptive cycle (1631-1944), throughout which more than one hundred phonotephritic lava flows were emplaced (Carta *et al.*, 1981; Arno *et al.*, 1987; Belkin *et al.*, 1993; Arrighi *et al.*, 2001; Cioni *et al.*, 2008; Santacroce *et al.*, 2008). In case of an eruptive crisis, these effusive eruptions may affect inhabited areas, important facilities (e.g. schools, medical centres, administrative buildings) and major transport infrastructures (roads, railways) (Coltelli *et al.*, 2012; Harris, 2015; Brown *et al.*, 2017; Jenkins *et al.*, 2017; Del Negro *et al.*, 2019; Carracedo *et al.*, 2022b; Meredith *et al.*, 2022, 2024). While the 1858 lava flow did limited damage, a similar eruption nowadays would certainly reach and destroy several buildings. In the case of a flow reaching the coast, the socio-economic impact would be high. Assessing the hazard posed by future eruptions producing lava flows requires a detailed understanding of lava properties (composition, erupted temperature, viscosity) and transport mechanisms (lava channels, lava tubes). Identifying lava tubes and understanding the conditions necessary for their formation are critical to improve hazard assessment and update hazard maps at active volcanoes (Felpeto *et al.*, 2001; Behncke *et al.*, 2005). This knowledge is also essential for more effective management of effusive crises, and for mitigating potential social and economic impacts (Pesaresi *et al.*, 2008; Coltelli *et al.*, 2012; Harris, 2015; Cubellis *et al.*, 2016; Brown *et al.*, 2017; Del Negro *et al.*, 2019; Meredith *et al.*, 2022, 2023).

Lava tubes are defined as roofed conduits within which molten lava flows and are very efficient in transporting lava within lava flow fields (Greeley, 1971a, 1971c, 1987; Peterson *et al.*, 1994; Kauahikaua *et al.*, 1998; Calvari and Pinkerton, 1999; Halliday, 2004). Unlike in lava channels, where atmospheric radiation and convection are the main heat exchange mechanisms (Harris *et al.*, 2005), inside a lava tube, the molten lava is covered by a solid crust, inhibiting atmospheric cooling and thus leading to much lower cooling rates (Keszthelyi, 1995; Sakimoto and Zuber, 1998; Witter and Harris, 2007). The lava flowing inside a lava tube remains molten

for a longer time, with the ability to emplace further away from the eruptive vents (Atkinson *et al.*, 1975; Keszthelyi and Pieri, 1993; Allred and Allred, 1997).

Lava tubes are widespread on Earth's surface and have also been identified on extraterrestrial bodies such as Mars or the Moon (Greeley, 1971b; Sauro *et al.*, 2020; Lèveillé and Datta, 2010; Qiu and Ding, 2023). Lava tubes typically form in low viscosity lavas such as basalts (Greeley, 1971a, 1971c, 1987; Peterson *et al.*, 1994; Kauahikaua *et al.*, 1998; Calvari and Pinkerton, 1998), and their formation is principally controlled by the composition and viscosity of the erupted lava, the eruptive rate and the pre-emplacement topography (Greeley, 1971a, 1971c, 1987; Peterson *et al.*, 1994; Kauahikaua *et al.*, 1998; Calvari and Pinkerton, 1998). Notable examples are found in Hawaii, such as the Kazumura cave formed at Kilauea (Greeley, 1971a, 1971c, 1987; Peterson *et al.*, 1994; Allred and Allred, 1997; Kauahikaua *et al.*, 1998; Kempe and Ketz-Kempe, 2016) or in the Canary Islands, such as La Corona tube system formed by the homonymous volcano on the Island of Lanzarote (Tomasi *et al.*, 2022). Lava tubes have recently been discovered within effusive products produced during the most recent eruptive cycle of Vesuvius (Lemaire *et al.*, 2024). The effusive activity of Vesuvius has received limited attention from the scientific community, with an absence of studies specifically focused on the presence, formation and development of lava tubes at the volcano.

The mechanisms leading to the formation of lava tubes have been described through observations of active tubes during eruptions, as well as inferred from studies of inactive tubes (Greeley, 1971a, 1971c, 1987; Peterson *et al.*, 1994; Kauahikaua *et al.*, 1998; Calvari and Pinkerton, 1999). Currently, tube formation processes are primarily characterised from a physical perspective. Among the key physical parameters, the rheology of the flowing lava plays a fundamental role in the initiation and development of lava tubes. The rheological evolution of a melt is influenced by its initial composition, the degree of degassing and crystallisation, and the shear rate (Burkhard, 2001; Di Fiore *et al.*, 2022, 2023a, 2023b). All these factors, except the initial composition are, in turn, dependent on the degree of undercooling which is governed by local cooling conditions. Therefore, it is fundamental to investigate the chemistry and microstructural features of tube-forming lavas to improve our understanding of the formation, development, and activity of lava tubes.

Building on the physical characterisation of lava tubes in the 1858 lava flow field of Vesuvius presented by Lemaire *et al.* (2024), this study investigates the petrological, geochemical and textural properties of the lavas that formed the flow field and its largest lava tube. The bulk chemical composition of the lava appears homogeneous throughout the 1858

lava flow field. However, variations in the matrix textures reflect changes and localised differences in the cooling conditions. The studied lava tube formed by inflation of a lava breakout, during which slabs of lava were uplifted, and numerous wall linings formed inside throughout the activity of the tube (Lemaire *et al.*, 2024). By analysing samples from different wall lining layers inside the lava tube, key parameters such as temperature, viscosity, and crystallinity were determined to retrieve the crystallisation kinetics of the lava during various stages of the tube formation and development. Cooling conditions including cooling rate and oxygen fugacity were estimated by examining crystals, specifically microlites, within the sampled lavas. Analysis of the variations in lava textures and physical properties can provide insights into the processes leading to lava tube formation and their periods of activity. Here, we present a detailed study of the petrology, geochemistry and texture of the lava which formed lava tubes, combining XRF, EMPA and micro-CT analyses. Our results provide crucial insights into the formation mechanisms of lava tubes, with the aim of assessing the factors that favour or inhibit their development and their efficiency as lava transport structures.

4.2. The 1858 eruption and lava flow field

Somma-Vesuvius is a volcanic complex located in southern Italy, in the Campania region, fifteen kilometres to the east of Naples (Fig. 46a). Of the numerous lava flows produced during the last eruptive cycle (1631-1944), one of them, the 1858 lava flow field, formed lava tubes (Fig. 46b). The 1858 eruption began on 27 May 1858 and ended on 12 April 1861, producing around 178 million cubic meters of erupted lava (Belkin *et al.*, 1993; Lemaire *et al.*, 2024). Seven fissures opened asynchronously during the eruption, leading to the formation of a compound lava flow field on the western flank of Vesuvius (Lemaire *et al.*, 2024). The surface of the 1858 lava flow field is mainly pāhoehoe, with the presence of lobes, lava ropes, and numerous inflation structures such as tumuli are observed. Recent work identified three lava tubes within this flow field, each exhibiting a distinct morphology which suggests different formation mechanisms (Lemaire *et al.*, 2024). This study focuses on the geochemical and textural analysis of the lavas forming the 1858 lava flow field, with particular attention to its largest lava tube.

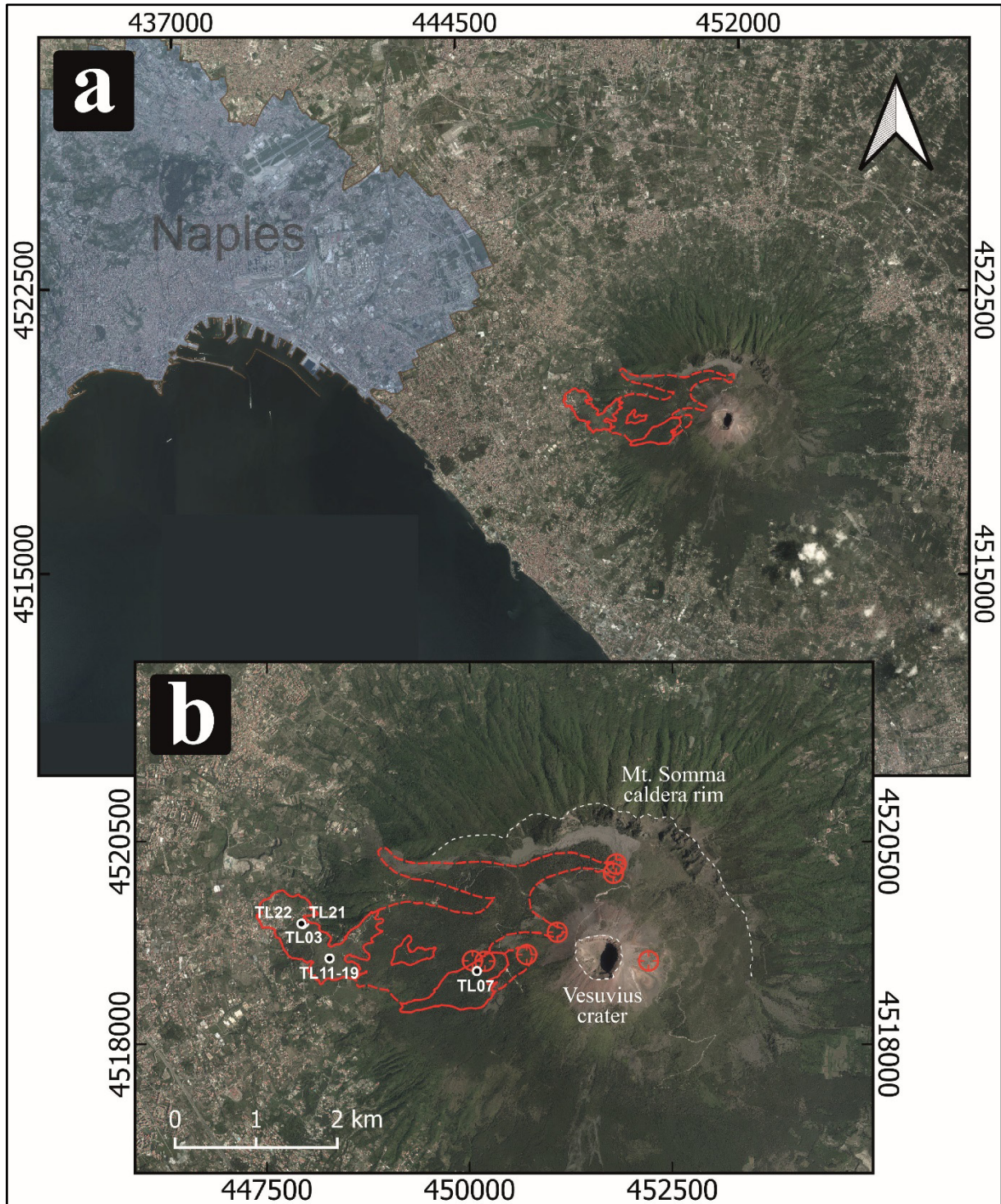


Figure 46. a) Location of Vesuvius and b) position of the 1858 lava flow field on its western flank and sampling locations (Red dashed line shows the extent of the flow field and the red full line shows the uncovered area of the flow field).

4.3. Methods

4.3.1. Sampling

Twelve samples of the 1858 lava flow field were collected to provide a complete overview of the flow field and its largest tube. Four samples were taken along the flow to characterise the spatial variability of the flow field, one close to the vent (TL07), three close to the front specifically from the basal part (TL22), the middle part (TL21) and the surface (TL03; Fig. 46b). Six samples were collected to characterise the largest lava tube identified in the flow field, one on the surface of the lava tube (TL19) and five at successive depths within the tube walls (TL11, TL12, TL13, TL14, TL16; Fig. 47). Both sample TL11 and TL13 derive from the last lining (Fig. 47d), easily recognisable as it is not always properly stuck to the wall.

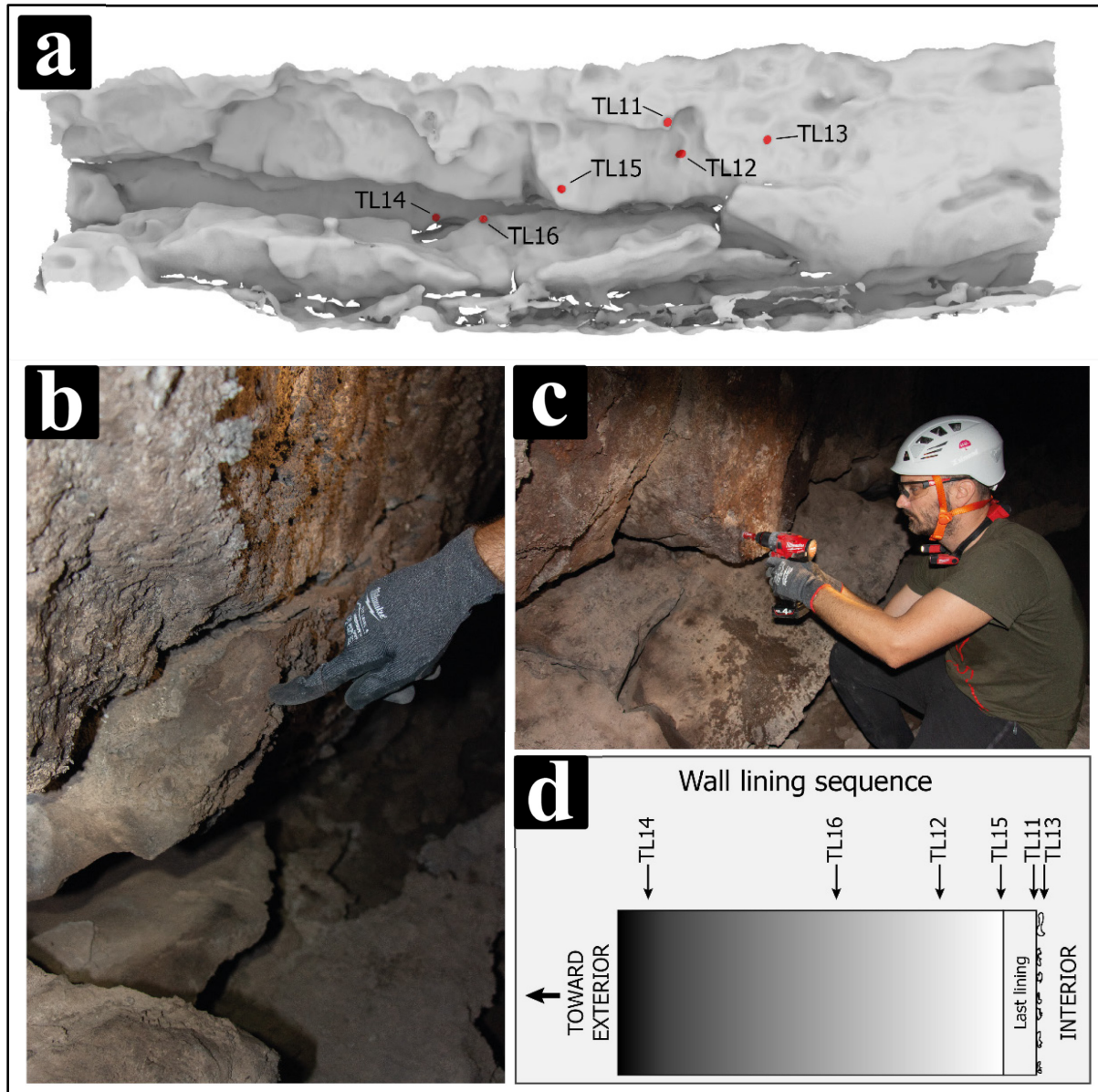


Figure 47. Sampling locations inside the largest lava tube found in the 1858 lava flow field.

4.3.2. X-ray Fluorescence (XRF)

The bulk rock compositions of the collected samples were characterised by X-ray Fluorescence (XRF). Samples were ground into powders with a grain size less than 75 μm using a Retsch PM-100 planetary mill and agate grinding balls. To prepare pressed powder pellets, we used 4 grams of lava powder mixed with a millimetre of Polyvinyl alcohol for each sample. The mixtures were dried and reduced to powder again. Powders were placed in aluminium capsules onto which we applied a pressure of 20 tons for 20 seconds.

XRF analyses were performed at the Department of Earth, Resource and Environment Sciences (DiSTAR) of the University of Naples, Italy, using an Axios Panalytical sequential wavelength dispersion x-ray fluorescence spectrometer (WDS) equipped with six analyser crystals, three primary collimators, and two detectors (flow counter and scintillator), operating at different kV and mA settings for each analysed element. Analytical uncertainties are in the order of 1–2% for major elements and 5–10% for trace elements.

4.3.3. Scanning Electron Microscope (SEM)

Backscattered electron (BSE) images were acquired using the Zeiss Merlin VP Compact electron microscope at the Department of Earth, Environment and Resource Sciences (DiSTAR), University Federico II, Naples, Italy. Four standard thin sections and six one-inch epoxy mounts were analysed. The images were used to compare groundmass textures and to quantify phenocryst volume contents using the software ImageJ (Abramoff *et al.*, 2004).

4.3.4. Electron microprobe analysis (EMPA)

Crystal and glass compositions were measured at the Laboratoire Magmas et Volcans (LMV) in Clermont-Ferrand, France, using a Cameca SxFiveTactis electron microprobe. All the samples were epoxy mounted. Crystal analyses were carried out using a focused beam with a 15 kV acceleration voltage and a 15 nA beam current to analyse ten elements (Si, Ti, Al, Cr, Fe, Mn, Mg, Ca, Na, K).

4.3.5. Synchrotron X-ray computed microtomography (SR μ CT)

The X-ray computed microtomography analyses were performed at the SYRMEP beamline of the Elettra synchrotron in Basovizza, Italy. The measurements were performed in propagation-based phase contrast mode (Cloetens *et al.*, 1996; Arzilli *et al.*, 2016) using a monochromatic, near-parallel X-ray beam set to a mean energy of 23 keV. Samples were mounted on a rotating head at a distance of 100 millimetres from the detector. 1800 radiographic images were acquired for each sample with a 0.01° step over a total range of 180°. Each

radiograph had an exposure time of 2 seconds, leading to a scan duration on the order of 1 h for each sample. A 0.9 μm pixel size of the detector has been fixed.

The custom software SYRMEP Tomo Project 4.0 (Brun *et al.*, 2017) was used for slice reconstruction, correcting ring artefacts (Rivers, 1998). A phase-retrieval algorithm (Paganin *et al.*, 2002) was applied to projections before the slice reconstruction using a δ/β parameter, which represents the ratio between the real and imaginary parts of the refractive index ($n = 1 - \delta - i\beta$), set to 50 (Arzilli *et al.*, 2016). This allowed enhancement of the contrast between different minerals and glass. The axial slices were imported into the software ImageJ to be cropped and converted into 8-bit raw images. Images with eventual ring artefacts were removed during this step. Edge-preserving filters were applied to reduce the image noise before segmentation using Avizo software. Vesicles and crystals were segmented by applying the 3D Otsu automatic thresholding algorithm (Otsu, 1979) with manual refinement following visual inspection of the samples using Avizo to ensure the selection of thin pore-melt films. Voids (vesicles and fractures) and phenocrysts from all phases were segmented from the initial volumes (Fig. 48).

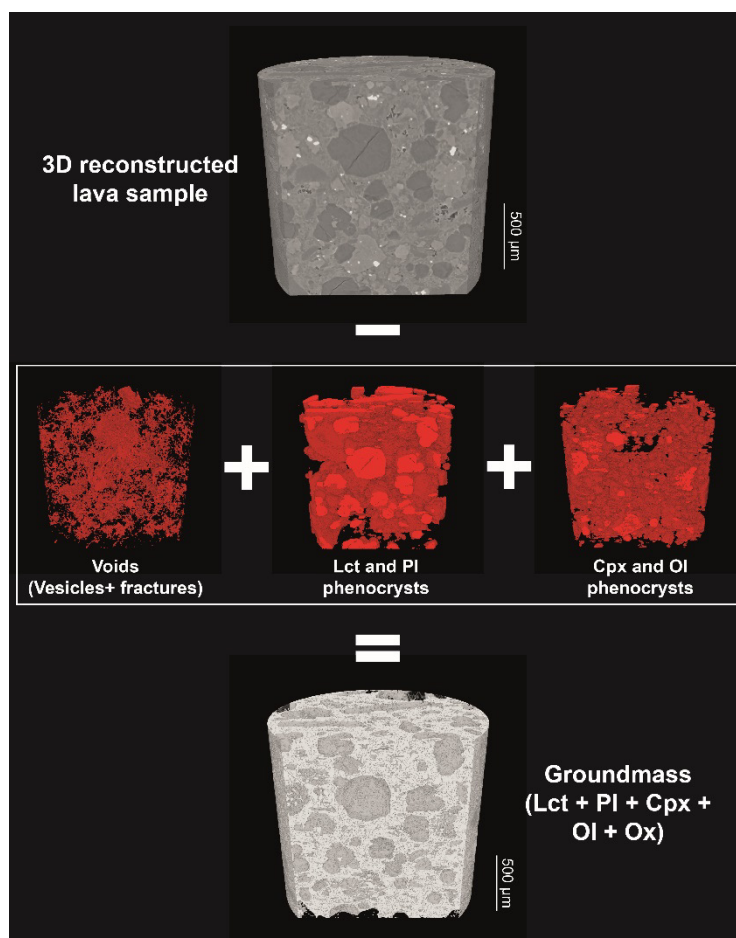


Figure 48. Methodology applied to the 3D reconstructed samples. Subtraction from the complete volume of the voids and phenocrysts volumes to conserve only the crystallised groundmass.

The 3D reconstructed samples were then analysed to extract the number, size, and volume of crystals in Avizo software. Crystals were separated by a combination of watershed, distance transform and numerical reconstruction algorithms. Crystals were then counted and measured to calculate the crystal size distribution (CSD) of each phase. Volume renderings were produced using Avizo and VGStudio 2022.

4.3.6. Geothermometry

Geothermometers were used to estimate the temperature of crystallisation. We applied the clinopyroxene-liquid (equation 33), plagioclase-liquid (equation 24a) and olivine-liquid (equation 22) thermometers from Putirka (2008). As the analysed samples are fully crystallised, we assume that the crystal cores are in equilibrium with a liquid having the average bulk rock composition (Table 6).

4.3.7. Oxygen fugacity

Oxygen fugacity (fO_2) was calculated relative to the Quartz-Fayalite-Magnetite buffer (QFM) following the method from Berry and O'Neill (Equation 8.2; 2021) using the ferric/ferrous iron ratios of Fe-Ti oxides and the average bulk rock composition:

$$\Delta QFM = \frac{\log_{10} \frac{Fe^{3+}}{Fe^{2+}} + 1.35 - 0.023[CaO] - 0.034[Na_2O] - 0.044[K_2O] + 0.018[P_2O_5]}{0.25} \quad (8)$$

where [CaO], [Na₂O], [K₂O] and [P₂O₅] are the concentrations of the oxide components in weight percent. The concentrations of ferrous iron (Fe²⁺) and ferric iron (Fe³⁺) were estimated as follows:

$$Fe^{2+} = \frac{[Fe^{2+}] + 4[Ti] - 2[Mg] - 2[Mn]}{3} \quad (9)$$

$$Fe^{3+} = \frac{2 \times ([Fe^{2+}] - 2[Ti] + [Mg] + [Mn])}{3} \quad (10)$$

where [Fe²⁺], [Ti], [Mg] and [Mn] are the number of cations contributing to the oxide formula.

The calculated proportions of total iron are used in equation 8 as the ferric/ferrous iron ratio $\frac{Fe^{3+}}{Fe^{2+}}$.

4.3.8. Thermodynamic calculations

The thermodynamic calculator MELTS (Ghiorso and Sack, 1995), specifically Rhyolite-MELTS version 1.2 (Gualda *et al.*, 2012; Ghiorso and Gualda, 2015), was used to constrain the crystallisation processes that took place within the lava tube. The input composition

corresponds to the average bulk rock composition measured by XRF, combined with the estimated water content from Belkin *et al.* (1993) of 0.50 wt.%. The complete crystallisation sequence from above the estimated liquidus was modelled at atmospheric pressure (1 bar) from 1300 °C down to the solidus temperature at 970 °C in 10 °C steps. Simulations were performed under three different oxidation conditions, represented by oxygen fugacity buffers: Hematite-Magnetite (HtMt) and Quartz-Fayalite-Magnetite (QFM +2 and QFM +3).

4.3.9. Modelled viscosity

The apparent viscosity (η_{app}) of the lava is calculated as follow:

$$\eta_{app} = \eta_{liq} \times \eta_r \quad (14)$$

where η_{liq} is the viscosity of the liquid phase and η_r accounts for the effect of suspended crystals and vesicles. The viscosity of the liquid phase η_{liq} was determined following the model from Giordano *et al.* (2008) based on the calculated melt composition obtained at every 10 °C step from Rhyolite-MELTS simulations. The relative increase in viscosity attributed to crystallisation was estimated using the calculated crystal volume fraction. To estimate the effect of crystals on the viscosity, three rheological models were applied.

The Shaw model (Shaw, 1969), based on the Einstein-Roscoe equation, which considers all crystals to be spherical:

$$\eta_r = (1 - 1.35\phi)^{-2.5} \quad (16)$$

where ϕ is the volume fraction of crystals in the lava.

The Krieger and Dougherty method (Krieger and Dougherty, 1959) also based on the Einstein-Roscoe equation but taking into account the crystal shape:

$$\eta_r = 1 - \left(\frac{\phi}{\phi_m}\right)^{B\phi_m} \quad (17)$$

where ϕ_m is the maximum packing fraction (Mueller *et al.*, 2010).

We approximated with this model the effect of large rounded particles (with a maximum packing fraction is 0.64) and fine elongated particles (with a maximum packing fraction is 0.454) by calculating the product of the effect for both crystal shapes (Farris, 1968):

$$\eta_r = \eta_r(\phi_l) \times \eta_r(\phi_f) \quad (18)$$

The model of Costa (Costa, 2005; Costa *et al.*, 2009) based on a semi-empirical non-Newtonian relationship which takes into account crystal shape and the effect of shear rate:

$$\eta_r = \frac{1 + \frac{\phi}{\phi_m}^\delta}{[1-F]^B \phi_m} \quad (19)$$

where F is calculated as follows:

$$F = (1 - \varepsilon) \operatorname{erf} \left[\frac{\sqrt{\pi}}{2(1-\varepsilon)} \frac{\phi}{\phi_*} \left(1 + \frac{\phi^\gamma}{\phi_*^\gamma} \right) \right] \quad (20)$$

in which the error function (*erf*) expresses a rheological transition below and above the maximum packing fraction. We used the fitting parameters for crystal types A and D from (Cimarelli *et al.*, 2011). Both the shape and fraction of crystals of the two types of particles can be introduced in the model using the following equations:

$$\phi_{*lf} = \phi_{*f} x^{1.34} + \phi_{*l} (1 - x) \quad (21)$$

$$\gamma_{*lf} = \gamma_{*f} x + \gamma_{*l} (1 - x) \quad (22)$$

$$\varepsilon_{*lf} = \varepsilon_{*f} x + \varepsilon_{*l} (1 - x) \quad (23)$$

where x is the fraction of finer to coarser particles.

Relative viscosity was calculated with the Costa model at shear rates of 10^{-4} and 1 s^{-1} .

Table 5. Fittings parameters from Cimarelli *et al.* (2011) were used to estimate the relative viscosity using the Costa model.

	A		D	
Strain rate (s^{-1})	10^{-4}	1	10^{-4}	1
ϕ	0.62	0.67	0.24	0.28
γ	1.52	1.6	0.1	0.11
ξ	5000	10000	30000	21000

In both the Krieger and Dougherty and Costa models, a bi-modal mixture of crystals was assumed. Leucite, Fe-Ti oxides and olivine were treated as large rounded particles, while plagioclase and clinopyroxene were modelled as smaller elongated particles.

4.4. Results

4.4.1. Bulk rock compositions, petrography and mineral chemistry

Whole-rock analyses, together with geochemical data from previous studies (Belkin *et al.*, 1993) indicate that the 1858 lavas fall within the phonotephrite field of the Total Alkali Silica (TAS) diagram (Le Bas *et al.*, 1986; Le Maitre *et al.*, 2002) with SiO₂ content ranging from 47 to 48 wt.% and total alkali content exceeding 9 wt.%. According to the classification proposed by Joron *et al.* (1987) for Somma-Vesuvius rocks, the 1858 lavas plot in the leucitic tephrit-phonolites field, characterised by an MgO content of 3.69 wt.% and a sum of normative feldspathoids of 38.78 wt.%. The total FeO is around 9 wt.%. The water content, estimated to be around 0.55 wt.% by Belkin *et al.* (1993), is consistent with the absence of hydrous mineral phases in the assemblage, indicating a low volatile content in the eruptive products.

Table 6. Bulk rock composition of the 1858 lava from Belkin *et al.* (1993), Joron *et al.* (1987), and this study obtained by XRF analysis.

	Belkin <i>et al.</i> , 1993		Joron <i>et al.</i> , 1987		This study	
Sample	V80	V82	X	TL07	TL21	MBRC*
SiO ₂	47.6	47.3	47.63	47.07	46.61	46.84
TiO ₂	1.08	1.07	0.98	1.13	1.18	1.16
Al ₂ O ₃	17.6	17.6	19.3	17.59	17.85	17.72
Fe ₂ O ₃	2.08	2.08	1.82	8.87	9.34	9.11
FeO	6.2	6.1	5.87			
MnO	0.15	0.16	0.15	0.17	0.18	0.18
MgO	4.2	3.98	3.91	3.79	3.59	3.69
CaO	9.2	8.67	8.04	9.37	9.62	9.50
Na ₂ O	2.65	2.63	3.05	2.45	3.26	2.86
K ₂ O	7.27	7.25	7.78	8.11	6.63	7.37
H ₂ O ⁺	0.34	0.31				
H ₂ O ⁻	0.11	0.2				
P ₂ O ₅	0.99	0.97	0.8	0.78	0.91	0.85
Total	99.9	98.53	99.99	99.33	99.17	99.25

*Mean bulk rock composition calculated as the mean of the TL07 and TL21 sample compositions.

The petrography of the 1858 lavas was investigated through thin sections analysis and BSE imaging. Crystals size measurements were carried out following the method of Zellmer (2021), where we considered four classes: phenocrysts ($>500\ \mu\text{m}$), microphenocrysts ($>30\ \mu\text{m}$ and $<500\ \mu\text{m}$), microlites ($\geq 1\ \mu\text{m}$ and $<30\ \mu\text{m}$) and nanolites ($\geq 30\ \text{nm}$ but $<1\ \mu\text{m}$). The maximum value between the length and width of the crystals was taken. The lavas have a glomeroporphyritic texture with a phenocryst assemblage of leucite (Lct), plagioclase (Pl) and clinopyroxene (Cpx). Most samples are holocrystalline with a glass-free microlitic groundmass. However, one sample (TL19) contains a glassy groundmass with evidence of incipient microlite crystallisation. The phenocrysts are typically euhedral, reaching sizes up to 3 mm. Leucites can be found as glomerophyres. Phenocrysts, crystals larger than $500\ \mu\text{m}$, represent, on average, 40 to 45 vol.% of the rock as estimated by microtomography and SEM across 10 samples (Table 7; Appendix 2).

Table 7. Volume fractions of phenocrysts determined by X-ray micro-CT and backscattered imaging.

Method	3D μCT			SEM			
Sample	lct + pl	cpx + ol	Vtot (%)	lct	pl	cpx	Vtot (%)
TL03	39.24	3.80	43.04	33.48	5.47	6.66	44.80
				19.62	5.71	18.66	
				31.19	3.91	5.17	
				34.84	4.04	8.42	
TL07	30.43	12.59	43.02	25.97	5.84	5.87	41.75
				20.07	7.12	4.76	
				27.40	8.95	5.91	
TL11	37.35	1.74	39.09	31.51	11.85	8.92	37.11
				26.83	12.06	11.34	
				33.68	11.99	9.38	
TL12	67.05	2.60	69.65	21.84	11.67	4.70	52.53
TL13	27.25	13.31	40.55	21.61	17.50	10.24	38.21
TL14	58.95	5.60	64.55	22.65	17.32	10.56	49.35
				14.03	17.48	10.43	
				24.01	11.91	11.68	
TL16	39.51	2.16	41.67	26.22	3.19	0.00	46.69
				26.41	6.19	6.68	
				15.07	10.07	8.72	
TL19	33.78	1.55	35.34	15.99	6.70	10.26	34.35
				22.09	7.74	2.41	
				25.35	5.93	10.14	
TL21	26.23	5.52	31.75	34.22	7.47	4.62	33.01
				17.68	7.92	11.63	
				24.86	9.04	8.14	
TL22	41.39	5.10	46.49	17.68	7.92	11.63	41.65
Average			45.52	24.86	9.04	8.14	41.94

Clinopyroxenes are present across the entire crystal size spectrum, from phenocrysts ($>500\ \mu\text{m}$) to microlites ($\geq 1\ \mu\text{m}$ and $<30\ \mu\text{m}$). Most clinopyroxene phenocrysts are oscillatory zoned. Clinopyroxene crystals of the 1858 lavas plot within the diopside field (Fig. 49a; Appendix 1). A positive correlation is observed between titanium and aluminium content.

Plagioclase crystals are present across all size classes from phenocrysts to microlites. Plagioclase phenocrysts plot within the bytownite (An_{70-80}) and labradorite (An_{60-70}) fields, while smaller crystals, microphenocrysts ($>30\ \mu\text{m}$ and $<500\ \mu\text{m}$) and microlites, also extend toward more sodic composition in the Andesine field (An_{30-50} ; Fig. 49b; Appendix 1).

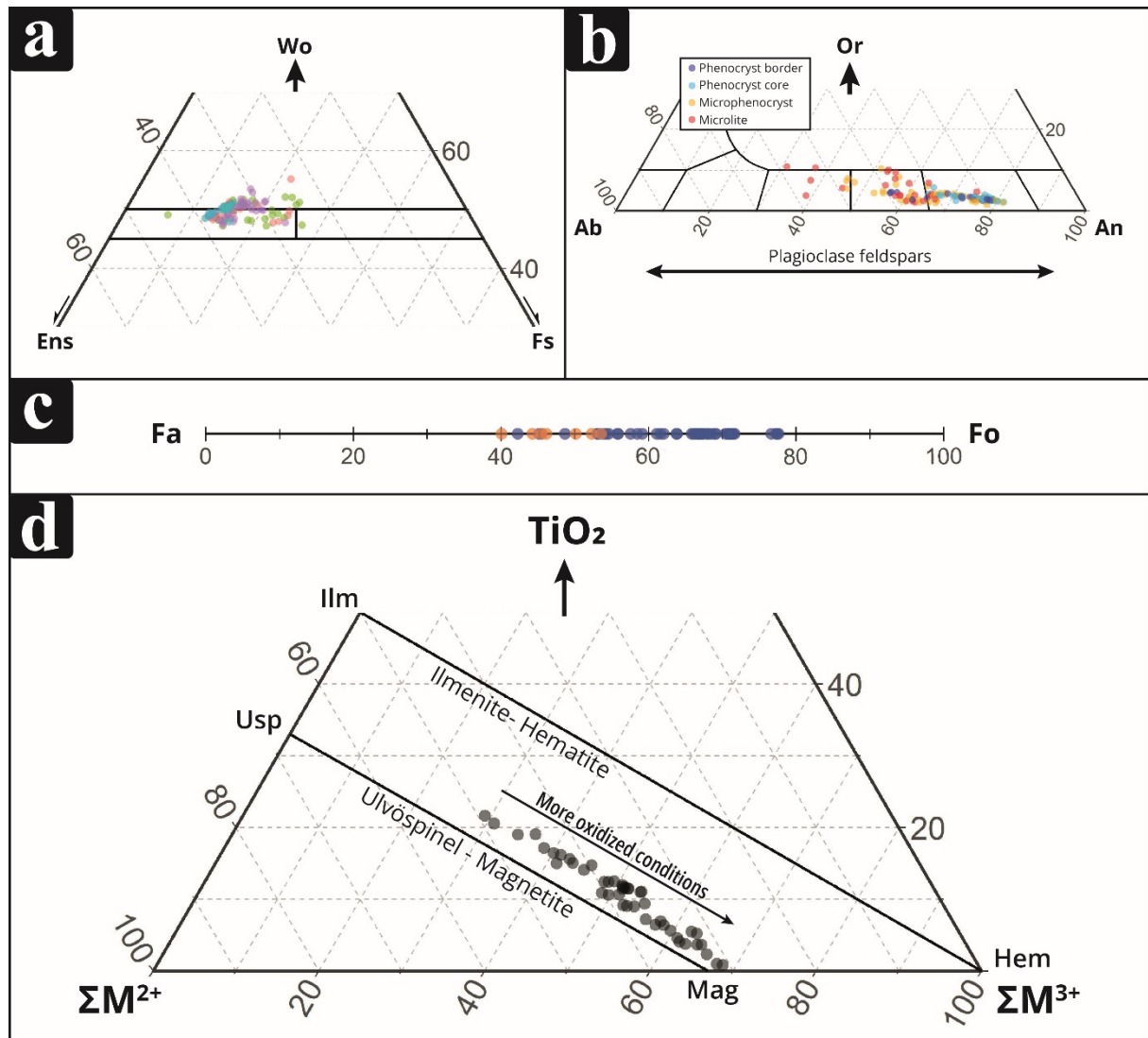


Figure 49. Mineral compositions. a) Clinopyroxene compositions plotted in a Ens-Wo-Fs ternary diagram. b) Plagioclase compositions plotted in an Ab-Or-An ternary diagram. c) Olivine compositions plotted on the Fa-Fo solid solution. d) Fe-Ti oxide compositions plotted in a ΣM^{2+} vs. ΣM^{3+} vs. ΣM^{4+} ternary diagram.

Olivine crystals are present as microlites and microphenocrysts. Olivine cores display a wide compositional range from Fo₅₅ to Fo₇₈, while their rims have a narrower compositional range from Fo₄₀ to Fo₅₅, indicating that olivine crystals were in disequilibrium with the

remaining liquid (Fig. 49c; Appendix 1). Apatite is present as an accessory mineral in all samples.

Oxides are present as nanolites, microlites and also as microphenocrysts. Fe-Ti oxides occur in concentrations ranging from 0.36 to 4.84 wt.%, and lie on the Ulvöspinel-Magnetite solid solution line when plotted in a ΣM^{2+} vs. ΣM^{3+} vs. ΣM^{4+} ternary diagram (Fig. 49d; Appendix 1). These oxides are iron-rich, with a titanium content varying from trace amounts to 20 wt.%. A decrease in TiO_2 is associated with an increase in ferric iron (Fe^{3+}), suggesting progressively more oxidising conditions of the crystallisation. Similar to clinopyroxene crystals, the titanium content increases with increasing aluminium.

All samples are fully crystallised; however, there are variations in the groundmass texture. Along the 1858 lava flow field (Fig. 50a), the lava collected at the surface near the eruptive vent (TL07) has a groundmass composed of skeletal to euhedral plagioclase microlites and clinopyroxene nanolites (≥ 30 nm but < 1 μm). In contrast, the lava sampled at the surface close to the flow front (TL03) has larger euhedral microlitic crystals of plagioclase and clinopyroxene interspersed in a very fine crystallised matrix. Lava samples collected from a quarry at the base (TL22) and centre (TL21) of the lava flow show a homogenous distribution of subhedral to euhedral microlites of plagioclase, clinopyroxene, olivine and Fe-Ti oxides.

Samples collected at the surface and inside the lava tube also show variations in groundmass textures (Fig. 50b). The sample from the surface of the lava tube (TL19) has a very finely crystallised groundmass with skeletal plagioclase microlites (< 30 μm) and clinopyroxene nanolites, indicating high cooling rates. Additional samples were collected from within the lava tube wall to represent a sequence of wall linings. These show microcrystalline groundmasses containing clinopyroxene, plagioclase and oxide microphenocrysts (30-100 μm) and microlites (Fig. 50b). Variations in microlite size across these samples suggest different cooling rates. The samples can be ranked by decreasing microlite size as follows: TL14, TL16, TL12, TL11 and TL13. Sample TL13 contain abundant microlites and rapidly crystallised oxides having dendritic structures. The differing proportions and development of mineral phases indicate that these lavas went through different cooling paths under varying redox conditions.

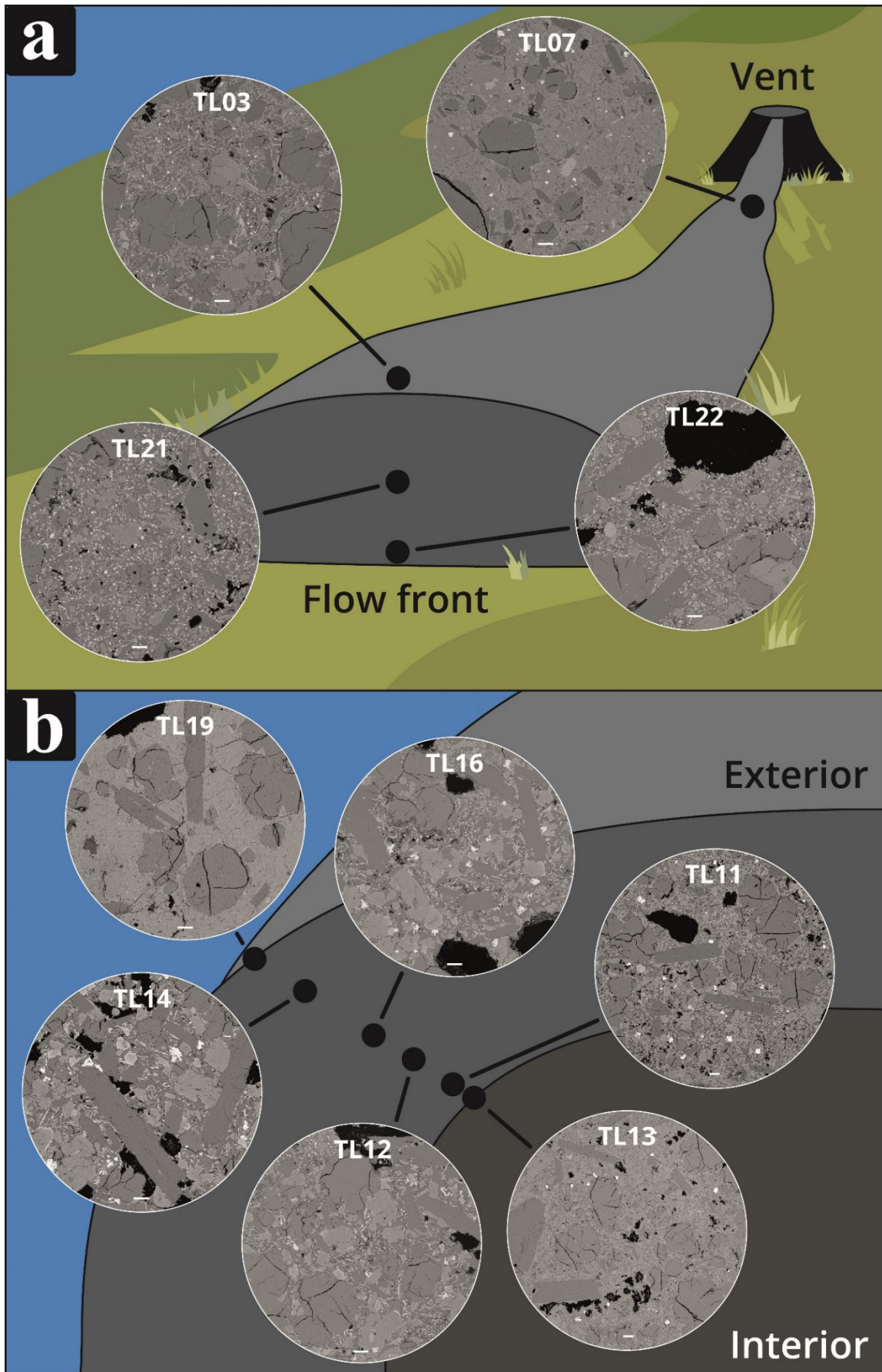


Figure 50. BSE images of the samples show the variations in textures. a) Along the lava flow field. b) Inside the lava tube.

4.4.2. Crystal morphologies and crystal size distributions

Initial 3D reconstructed samples had a volume of 3.91 mm^3 (3.34 mm^3 for TL12). The analysis was carried out on a vesicle and phenocryst-free basis (Fig. 48); thus, the analysed volumes were reduced to between 0.80 to 2.39 mm^3 . Crystal size distributions (CSD) were performed exclusively on the microlitic and dendritic groundmass of the samples to investigate crystallisation that happened during lava flow emplacement. Synchrotron X-ray μ CT provides high-resolution reconstructions, allowing for precise 3D measures. But in this case, due to the low X-ray absorption contrast between chemically similar mineral phases, it was not always possible to separate each phase individually, limiting the volume estimations of each phase. Instead, three composite phase groups were distinguished: (1) leucite and plagioclase, (2) clinopyroxene and olivine and (3) oxides (Fig. 51 and 52).

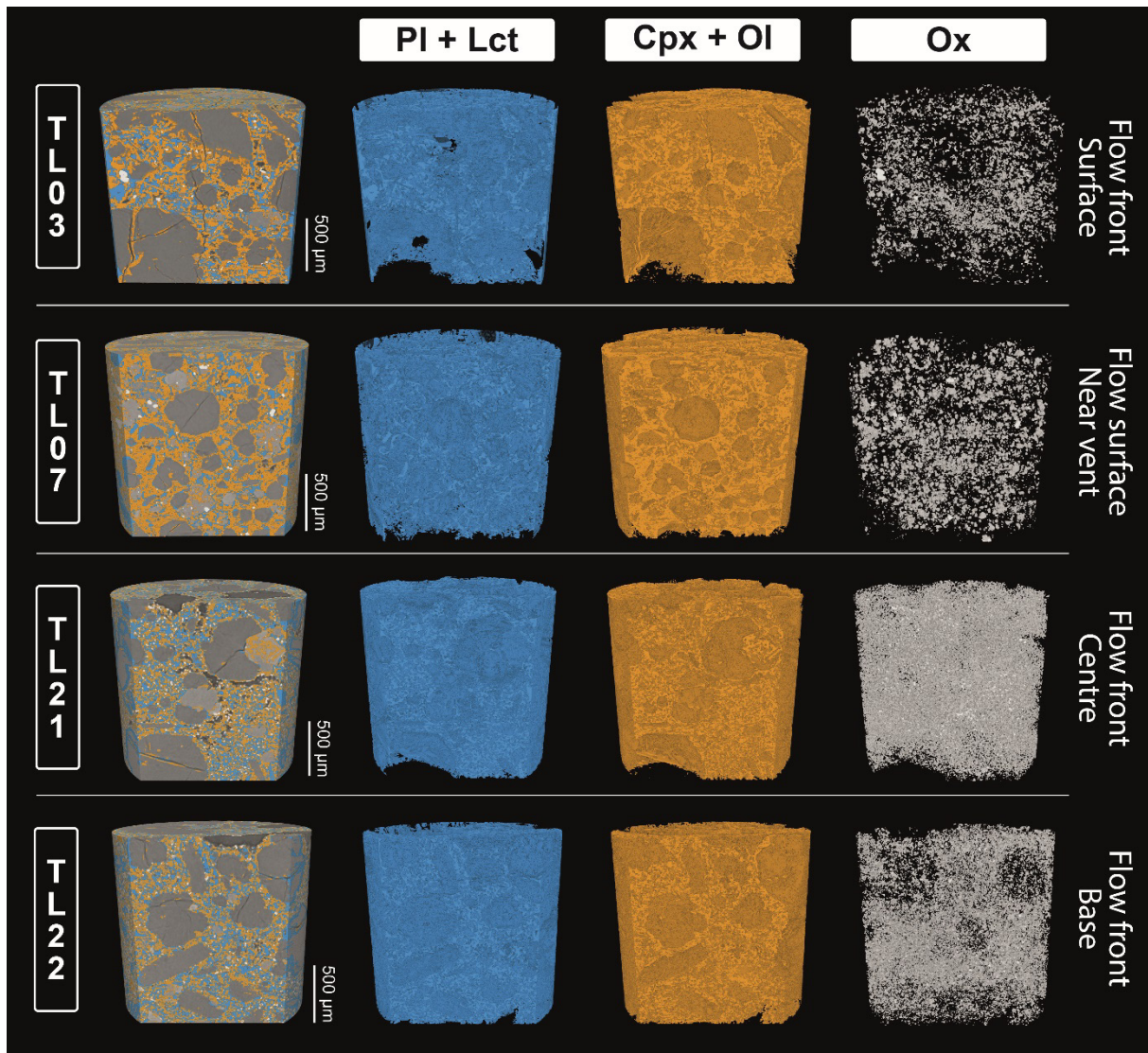


Figure 51. Samples collected along the 1858 lava flow field reconstructed in 3D and separated phases.

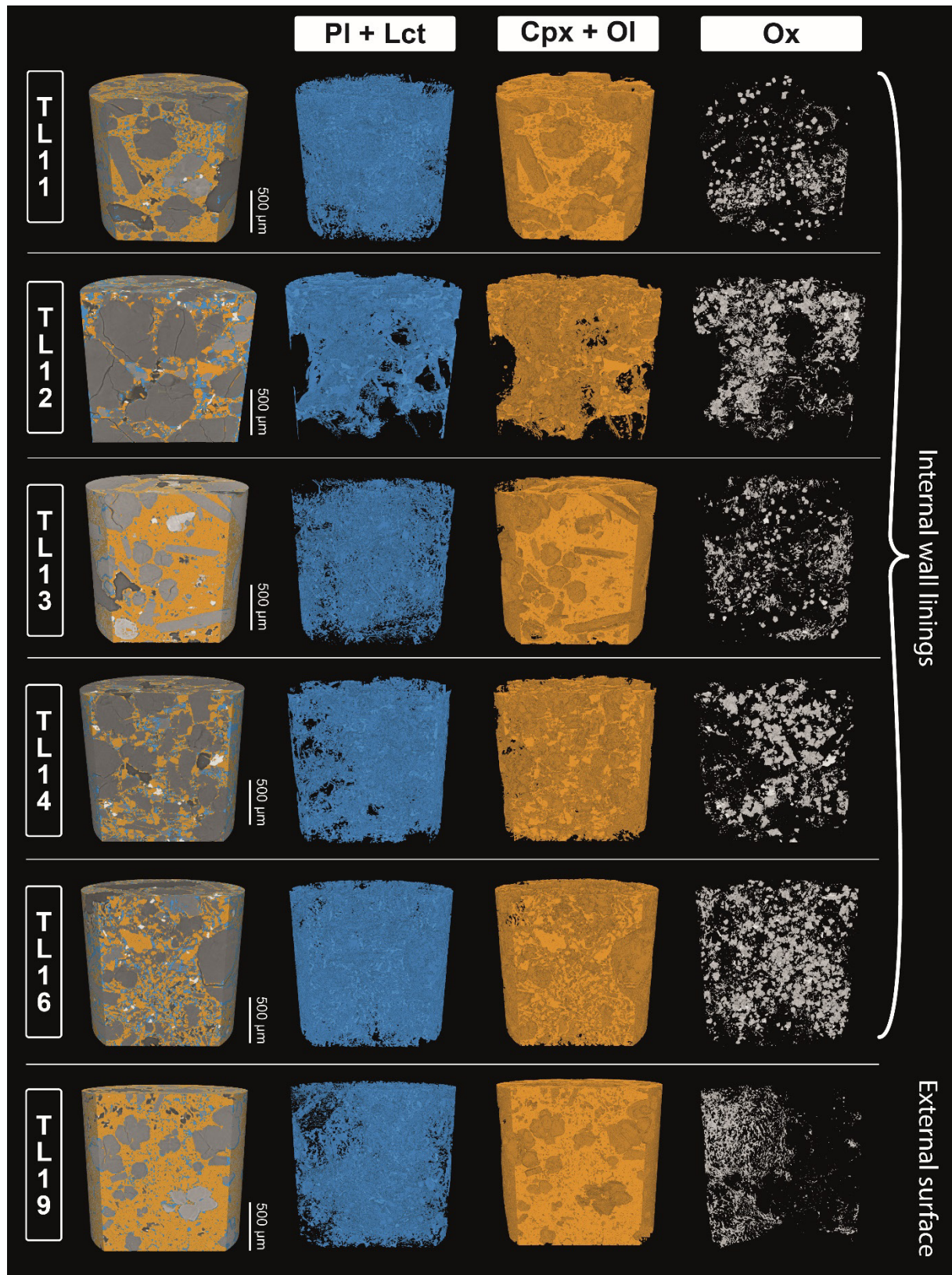


Figure 52. Samples collected inside the lava tube reconstructed in 3D and separated phases.

The volume fractions of microlites within the groundmass were computed for each sample (Fig. 53; Appendix 3). Inside the lava tube, volume fractions plotted in function of their distance from the innermost lining, show variations (Fig. 53). The volume of plagioclase microlites (solely plagioclase because leucite is not present as microlites) at the surface of the tube is 11 %, it reaches 30 % at 0.2 metres from the innermost lining, decrease to 12 % at the penultimate lining and drops below 1 % in the last wall lining. The volume fractions of oxides follow a similar trend, ranging from 0.3 to 3.6 %. Volume fractions of clinopyroxene and olivine microlites follow an opposite trend, with 80 % at the surface of the tube, decreasing to 52 % at 0.2 metres from the innermost lining, increasing to 77 % at the penultimate lining and reaching 85 % at the last lining.

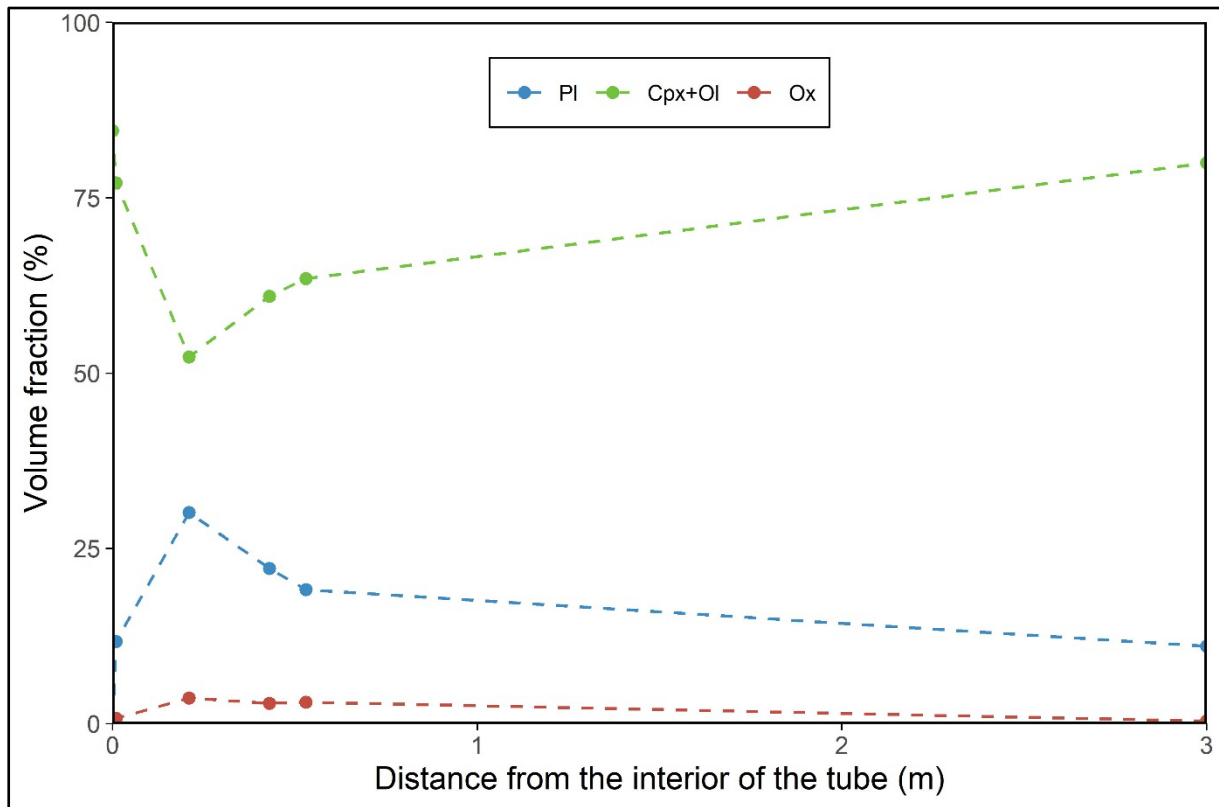


Figure 53. Volume fractions of microlites present in the groundmass.

All crystal size distributions show a decrease in the crystal number densities (Nd) with increasing crystal volume (Fig. 54; Appendices 4, 5, and 6). At the scale of the lava flow (Fig. 54a, 54b and 54c), distributions have similar trends for pl + lct and cpx + ol, with crystal number densities for crystals with a volume between 200 and 300 μm^3 , ranging from 4.3×10^3 to $6.6 \times 10^3 \text{ mm}^{-3}$ and from 1.1×10^4 to $2.2 \times 10^4 \text{ mm}^{-3}$ (Fig. 54a and 54b), respectively. Instead, distributions are broader for oxides, showing two trends (Fig. 54c). One trend with a high Nd of small oxide crystals, 1.1×10^4 to $1.3 \times 10^4 \text{ mm}^{-3}$, and lower Nd of large oxide crystals ($> 9000 \mu\text{m}^3$), 33 to 53 mm^{-3} , that represent samples from the bottom (TL22) and core (TL21) of the flow close to the flow front, while the other trend is showing a lower Nd of small oxide ($1.6 \times 10^3 \text{ mm}^{-3}$), and the presence of larger oxides with volumes up to $40000 \mu\text{m}^3$, which represents samples from the surface of the flow (TL03 and TL07).

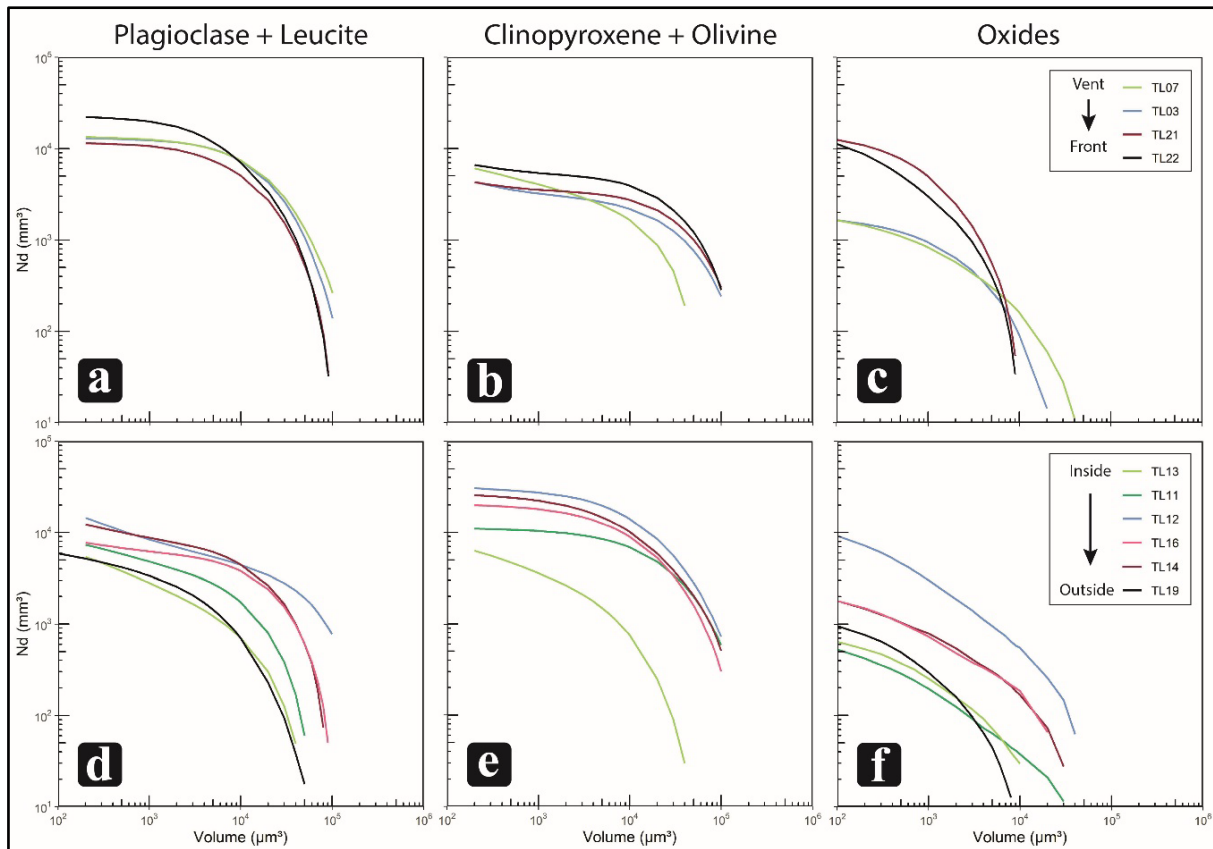


Figure 54. Crystal size distributions per separated phases for the lava flow samples (a, b, c) and for the lava tube samples (d, e, f).

Distributions of the samples collected inside the lava tube show more dispersed trends (Fig. 54d, 54e and 54f). For pl + lct crystals, distributions have crystal number densities for small volume crystals, ranging from 5.2×10^3 to $1.4 \times 10^4 \text{ mm}^{-3}$ (Fig. 54d). The main variation lies in the size and number of larger crystals. For cpx + ol crystals, three distinct trends are observed (Fig. 54e). The first trend includes three of the linings (TL12, TL14, TL16), which

display the highest crystal number densities, for small volume crystals, ranging from 2.0×10^4 to $3.0 \times 10^4 \text{ mm}^{-3}$. The second trend corresponds to TL11, which shows a similar distribution to the first trend but with a lower Nd of $1.1 \times 10^4 \text{ mm}^{-3}$ for small crystals. The third trend is represented by rapidly cooled lava, specifically the ultimate lining (TL13) and likely the surface of the tube (TL19; cpx + ol separation was not possible due to the very small crystal size). These show the lowest Nd value of $6.3 \times 10^3 \text{ mm}^{-3}$ and contain fewer large crystals (maximum $50000 \mu\text{m}^3$). For oxides, distributions have similar patterns but span different orders of magnitude (Fig. 54f). Among small crystals (between 100 and $200 \mu\text{m}^3$), TL12 shows the highest Nd at $9.2 \times 10^3 \text{ mm}^{-3}$. Two of the inside linings, TL12 and TL14, have nearly identical distributions starting at $1.8 \times 10^3 \text{ mm}^{-3}$. TL11, TL13 and TL19 show the lowest Nd values, ranging from 5.3×10^2 to $9.6 \times 10^2 \text{ mm}^{-3}$ while following similar trends.

4.4.3. Geothermometry

Leucite, clinopyroxene and plagioclase are the phases that had sufficient time to crystallise, grow and equilibrate with the melt to form phenocrysts. We use the compositions of clinopyroxene and plagioclase phenocrysts along with olivine compositions to estimate their crystallisation temperature.

One of the first phases to crystallise from the melt is clinopyroxene. Ten compositions of clinopyroxene in equilibrium ($K_D(\text{Fe-Mg})^{\text{cpx-liq}} = 0.25 \pm 0.01$) with the ‘liquid’ (mean bulk rock composition) were used in equation 33 from Putirka (2008), yielding a mean temperature of 1124°C with a propagated error of $\pm 46^\circ\text{C}$ (Fig. 55).

Seventy-five compositions of plagioclase in equilibrium ($K_D(\text{An-Ab})^{\text{pl-liq}} = 0.27 \pm 0.11$) with the ‘liquid’ (bulk rock composition) were used in equation 24a from Putirka (2008), yielding a mean temperature of 1115°C with a propagated error of $\pm 36^\circ\text{C}$ (Fig. 55).

The temperature of the onset of olivine crystallisation was estimated using the olivine-liquid thermometer described in Putirka (2008) as equation 22. We assume that olivine cores are in equilibrium with a liquid that has the mean bulk rock composition (Table 6). Twelve olivine compositions yield an average temperature of 1122°C with a propagated error of $\pm 45^\circ\text{C}$ (Fig. 55).

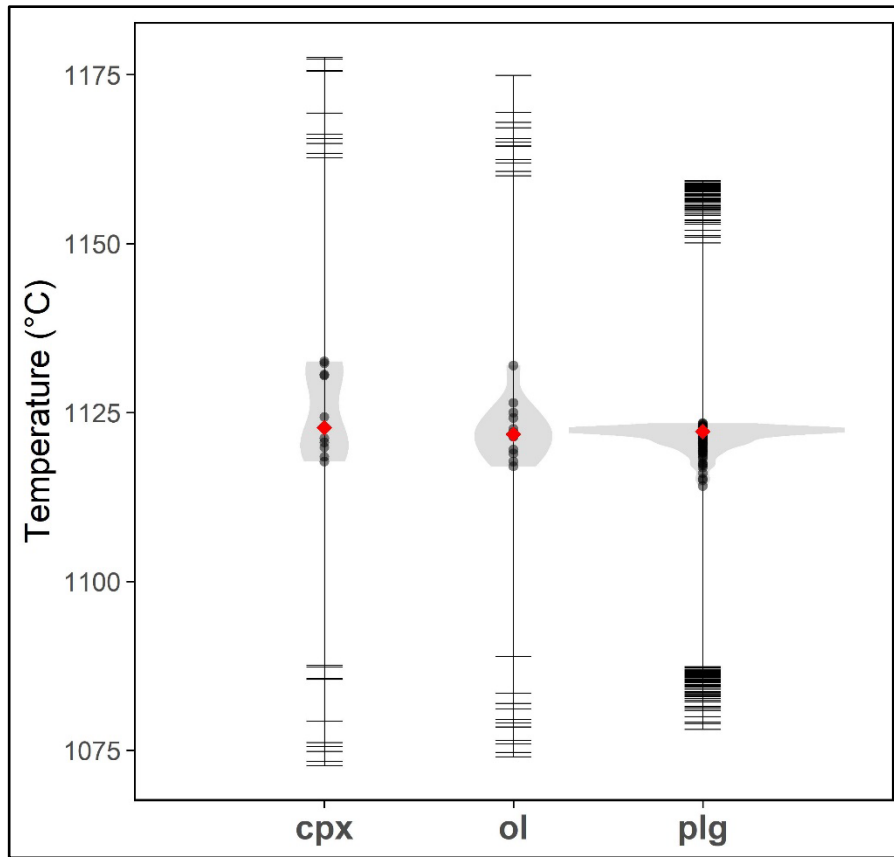


Figure 55. Eruption temperature estimates using cpx-liquid, ol-liquid, and pl-liquid thermometers.

4.4.4. Oxygen fugacity (fO_2)

The previously observed variation in the ferric/ferrous iron ratio in Fe-Ti oxides shows a change in the oxidation state of iron in the lava. Oxygen fugacity is influenced by melt composition, temperature, and the oxidation state of iron. In our case, melt composition remains constant, while only temperature and iron oxidation state vary. Determining the oxygen fugacity under which the lava crystallised allows the degree of oxidation during crystallisation to be constrained.

All the samples cooled under oxidising conditions (Fig. 56). The $\log(fO_2)$ values of the sample collected at the base of the lava flow (TL22) relative to the QFM (quartz–fayalite–magnetite) buffer are QFM +2.5, while the samples collected inside the lava tube (TL13, TL16, TL19) yield even more oxidising conditions between QFM +3.5 and QFM +4 (Fig. 56).

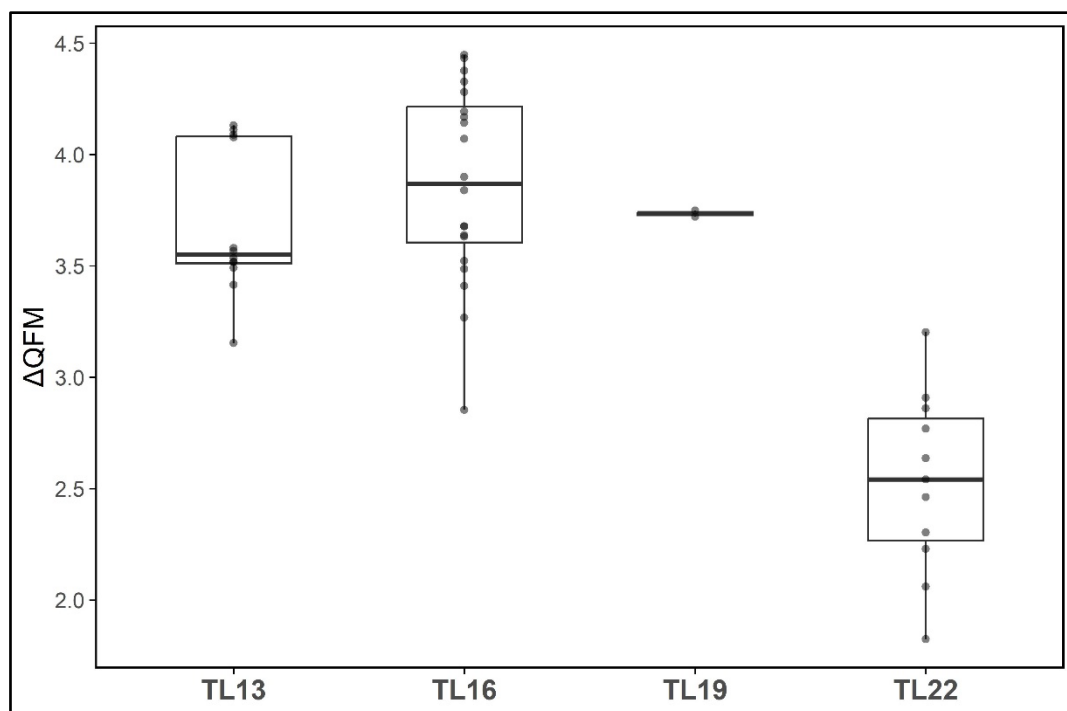


Figure 56. Oxygen fugacity (fO_2) estimations.

4.4.5. Thermodynamic calculations

Textural analysis and thermodynamic phase modelling indicate that phenocryst crystallisation occurred at depth, whilst crystallisation of microlites occurred during lava flow emplacement and lava tube development. Calculations were performed using Rhyolite-MELTS v1.2, using the mean bulk rock composition as an input (Table 6). The results predicted by Rhyolite-MELTS were combined with the experimental results obtained by Di Fiore *et al.* (2023b), which investigated crystallisation for a similar phonotephritic composition. We observe a relatively small difference between the liquidus temperature of clinopyroxene ($\approx 30^\circ\text{C}$) predicted by Rhyolite-MELTS at ambient pressure and that observed by Di Fiore *et al.* (2023b). While we observe a larger difference between the liquidus temperature of leucite ($\approx 40^\circ\text{C}$) predicted by Rhyolite-MELTS and that experimentally constrained by Di Fiore *et al.* (2023b). Agreement in the plagioclase liquidus ($<1150^\circ\text{C}$) is observed between modelling and experimental results. The combination of these results allowed us to constrain a phase equilibria diagram at ambient pressure for the phonotephritic composition and the evolution of crystal fraction and mineral assemblage as a function of temperature and oxygen fugacity (Fig. 57).

The three thermodynamic calculations performed at different fO_2 conditions show similar volume fractions for each of the phases which comprise the 1858 lava (Fig. 57; Appendices 7, 8, and 9). Variation in oxidation conditions has a minor influence on the

crystallisation of phonotephritic lava but still affects the initial crystallisation temperature of leucite, which occurs at 1260 °C, 1210 °C and 1180 °C at QFM +2, QFM +3 and HtMt, respectively, and Fe-Ti oxides >1300 °C. Clinopyroxene starts crystallising between 1175-1185 °C and plagioclase at 1130 °C.

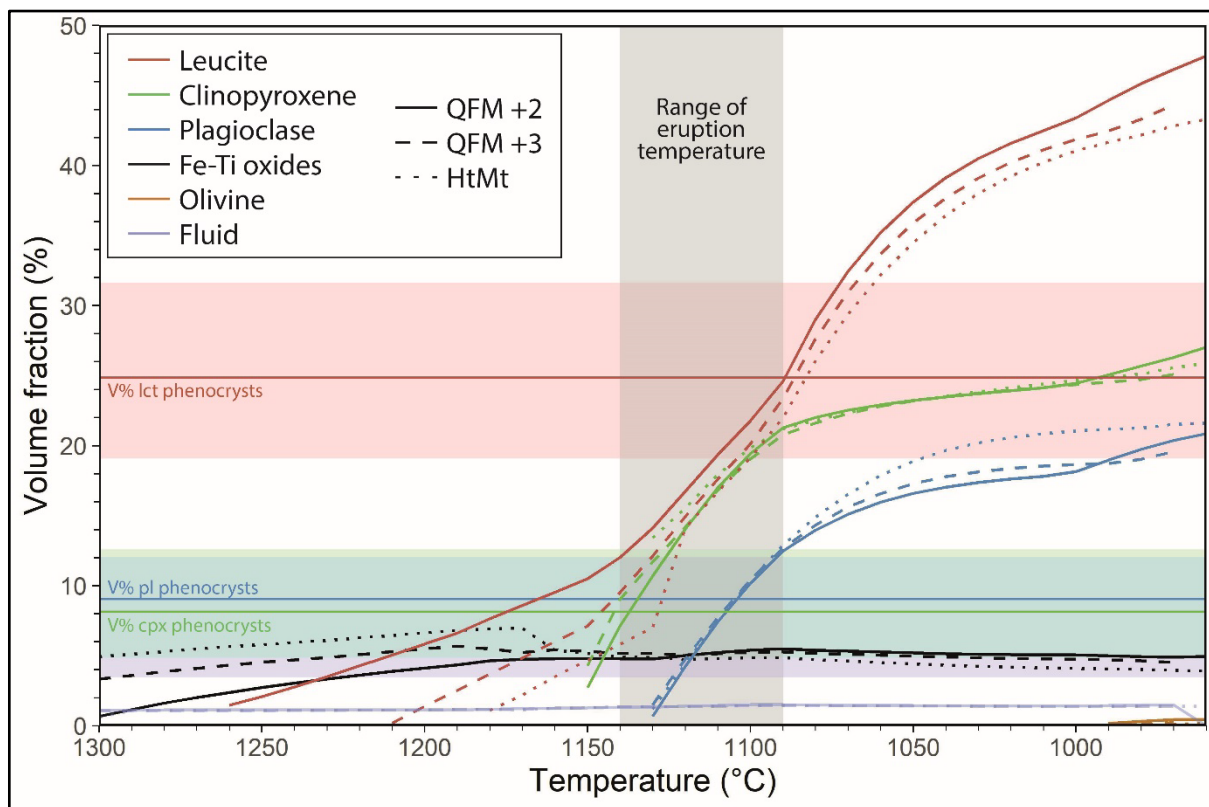


Figure 57. Thermodynamic calculation as a proxy for the cooling of a phonotephritic melt represented by the volume fraction of each phase at three fO_2 conditions with decreasing temperature. Horizontal lines represent the mean amount of phenocrysts measured in the sample set. Coloured areas correspond to the standard deviation. The grey area is the estimated range of temperature at which the lava flows inside the lava tube.

To validate the estimated crystallisation temperature obtained by thermometry, the phenocryst volume content of each phase was measured (Table 7) and compared with the crystallisation paths from thermodynamic calculations. The temperatures at the intersection of both curves correspond to the temperature of crystallisation of the phenocrysts. Mean phenocrysts volume content for leucite, plagioclase and clinopyroxene are 24.86, 9.04 and 8.14 V.% respectively, corresponding to temperatures of 1086, 1104.5 and 1142 °C. Therefore, the estimated temperature at which phenocrysts formed at depth is 1111 ± 25 °C. The lava flowing inside the lava tube must be slightly cooler than the crystallisation temperature, depending on ascent rate and heat loss.

4.4.6. Viscosity modelling

The apparent viscosity (η_{app}) of the melt was estimated as the viscosity of the liquid phase (η_{liq}) multiplied by the effect of crystals in suspension (η_r). The effect of vesicles was not considered in this calculation. The viscosity of the liquid calculated using the model of Giordano *et al.* (2008) based on compositions computed by Rhyolite-MELTS, ranges from 35 to 171 Pa s at 1300 °C to $1.15 - 1.48 \times 10^4$ Pa s at 970 °C, the solidus temperature (Fig. 58).

Subsequently, the effect of crystals was added to the viscosity of the liquid using three different models. Common viscosity models are generally not suitable for highly crystalline lavas and tend to overestimate the effective viscosity. As a first approach, the Shaw method was used (Shaw, 1969), assuming all crystals to be spherical in shape. According to this model, the rheological cut-off, the point at which the viscosity becomes so high that the lava ceases to flow, is reached at approximately 1090 °C.

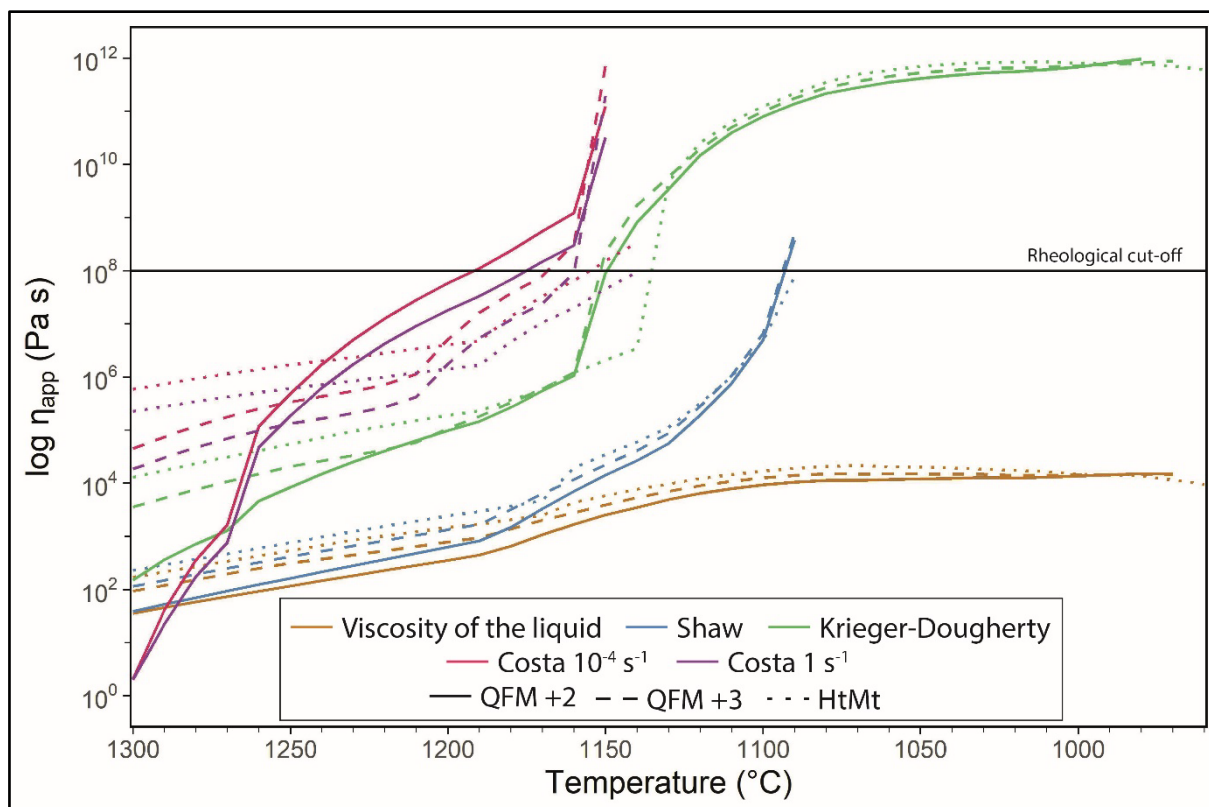


Figure 58. Viscosity estimates at different fO_2 conditions. The viscosity of the liquid is represented in yellow. Apparent viscosity is represented in blue.

For the Krieger-Dougherty and Costa models, a bi-modal crystal size distribution was used with large spherical particles representing leucite, olivine and Fe-Ti oxides phenocrysts and needles-like fine particles representing plagioclase and clinopyroxene microlites. Both models rapidly overestimate the lava viscosity with a rheological cut-off reached above 1150 °C.

4.5. Discussion

4.5.1. Erupted lava and lava flow emplacement

Our results show that the 1858 eruption at Vesuvius erupted phenocrysts-rich ($\approx 40\%$) phonotephritic magma, in agreement with previous studies on the effusive products of Vesuvius (Belkin *et al.*, 1993). Phenocrysts are commonly believed to form in the magma reservoir, while smaller crystals form during magma ascent in the conduit and during lava flow emplacement. The growth of crystals increases lava viscosity and changes its behaviour to non-Newtonian (Mueller *et al.*, 2010), changing the way lava flows and emplaces (Soldati *et al.*, 2016). Here, differences of crystallisation are observed from 3D μ CT data in the groundmasses of our samples collected along the 1858 lava flow field. The changes in number, size, and volume of the microlitic crystals are due to variations in local cooling conditions. Samples from the surface of the flow have a microcrystalline groundmass texture, indicating a high degree of undercooling, consistent with air-cooled lava. The samples collected at the base and centre near the lava flow front have a different groundmass texture with a more homogeneous crystal size, indicating slower cooling. The cooling rate was lower due to piling up of single lava flows, keeping previously emplaced flows at a high temperature for a prolonged period.

The high phenocryst content in the lava increases its viscosity, reducing its ability to flow. This parameter, alongside the low slope of a large part of the emplacement surface and the low effusion rate, explains the short distance of emplacement (≈ 3 kilometres; Lemaire *et al.*, 2024) of the 1858 lava flow field.

Observing the crystallisation of lava can be useful to decipher its ability to flow. At Vesuvius, some flows reached the sea, emplacing over 7 kilometres, while others, like the 1858 lava flow field, were much shorter. The effects of crystallisation on lava flow emplacement need to be constrained to better simulate lava flows and improve hazard maps, especially at highly inhabited volcanic centres like Vesuvius, where cities are at risk.

4.5.2. Lava tube formation and activity

The lava at the surface of the lava tube is very finely crystallised with skeletal plagioclase microlites, indicating that it was quenched. Compared to experiments, the surface lava seems to have cooled around $180\text{ }^{\circ}\text{C/h}$ (Vetere *et al.*, 2015). Instead, lavas successively flowing inside the tube are more crystalline, suggesting a longer time cooling. The cooling rates inside the lava tube seem to be comprised between 1 and $60\text{ }^{\circ}\text{C/h}$ (Vetere *et al.*, 2015). This observation is in agreement with the proposed model of formation by inflation of a lava flow

outbreak for this lava tube (Lemaire *et al.*, 2024), where the crust of the outbreak flow cools rapidly, which forces the flow to inflate and later form a tube. Lava tubes are known to be highly efficient lava transport structures with important insulation of the molten lava flowing inside, inhibiting atmospheric cooling. This is the case here, inside the studied lava tube, where lava was flowing between 1115 ± 25 °C and 1090 °C with a viscosity ranging from 10^4 to 10^8 Pa s.

Our results show that the interior of the lava tube was a more oxidising environment than the lava flow. While the variation of the redox conditions can affect the crystallisation of a lava (Kolzenburg *et al.*, 2018), the performed thermodynamic calculations show that, for a phonotephritic melt emplacing on the Earth's surface, it does not have a major influence on the liquidus temperatures and the amount of crystals. It seems unnecessary to take oxygen fugacity into account when modelling the flow of silica-undersaturated alkaline potassic melts as crystallisation seems to be mainly driven by the degree of undercooling and shear rate (Vona and Romano, 2013; Di Fiore *et al.*, 2023b).

Our results show that lava flowing inside a lava tube is insulated and maintained at high temperature. Still, the interior of a lava tube is subject to variations in temperature and oxygen fugacity during its activity. These changes impact the crystallisation kinetics of the lava flowing inside the tube towards the flow front. Constraining the thermal and rheological state of lava within lava tubes is crucial to build numerical models that simulate flow within a tube or lava flows with lava tube formation.

When lava flowed inside the lava tube, crystallisation was favoured at the contact of the walls and sometimes ceiling, where the heat gradient was higher due to heat conduction through the tube's structure and where lava is affected by stronger shear rates. Crystallisation is initiated at higher temperatures when lava is sheared, up to 20-30 °C higher at a shear rate of 1 s^{-1} (Di Fiore *et al.*, 2023b). Higher shear rate promotes crystallisation, increasing its viscosity and reaching its rheological cut-off at higher temperatures. While shear rate tends to increase lava viscosity by accelerating crystallisation, it can also decrease its viscosity due to disaggregation of leucite clusters (Campagnola *et al.*, 2016). The heat gradient due to conduction and shear rate are the main parameters controlling the growth of wall linings during the passage of lava inside the lava tube, increasing the viscosity and immobilising lava at the flow boundaries. The sampled wall linings have different groundmass textures that indicate a variation in local cooling conditions.

Better understanding lava tube formation and development is crucial to improve preparedness and hazard assessment. The 1858 lava flow field has a limited extent, even if it formed lava tubes, due to the limited size of the structures. However, at volcanoes like Vesuvius, small tubes can help lava flows reach inhabited areas due to their proximity to the eruptive centre. Our constraints on lava tubes internal conditions, namely temperature and viscosity, can be implemented in flow models to improve lava flow modelling by considering lava tube formation. Therefore, preparation of lava flow hazard maps will be improved and help accurately plan the response in case of effusive events.

The stability of a lava tube through time depends on the way its structure is formed. Here, a thick crust cooled, and then successive layers of lava were attached to it under variable conditions. If not well attached, the layers can detach and weaken the integrity of the lava tube's structure. Our results can be used in stability models to constrain the structure of lava tubes on Earth and on other planetary bodies. For geoheritage and geotourism, determining the stability of lava tubes is required to ensure safe exploration by the public. Our work is also important in the domain of space exploration. Being able to precisely model the stability of a lava tube structure is particularly important to prepare missions aimed at developing habitats in lava tubes on Mars or the Moon for future explorations.

4.6. Conclusion

Our study is the first to provide petrological, geochemical and textural analyses of tube-forming lavas. The 1858 lava flow field at Vesuvius is a unique deposit to study the emplacement of phonotephritic lava rich in phenocrysts and the formation of lava tubes in alkaline lavas. Our analyses show the effect of the degree of undercooling and oxidation conditions on crystallisation kinetics. The difference in crystallisation influenced the emplacement dynamics of the 1858 lava flow field and the formation and development of the lava tube. Several lava flows were piled up to a maximum height of over on hundred metres, limiting the length of the flow field to 3 kilometres. The lava tube formed when a small breakout surface solidified rapidly by atmospheric cooling, forcing the flow to uplift. Here, lava tube formation is controlled by cooling rate and effusive rate. Inside the lava tube, several linings solidified on the walls and ceiling under temperatures ranging from 1115 ± 25 °C to 970 °C and with an oxygen fugacity ranging from QFM +3 to QFM +4.5. Lava tube activity is controlled by the effusion rate, which changes the level of lava flowing inside the tube. When lava flows

inside the lava tube, crystallisation depends on the degree of undercooling, ranging here from high (60 °C/h) to low (1 °C/h), and the amount of shear rate, increasing lava viscosity up to two orders of magnitude. This controls not only the thermo-rheological properties of lava arriving at the flow front and thus the ability of the flow to advance, but also the crystallisation of lava on the walls and ceiling of the lava tube that impacts its structure, forming millimetre to centimetre thick layers.

CHAPTER 5

**Progress in Acoustic Emission testing: a new tool aimed at
assessing the stability of lava tubes**

Abstract

This research work investigates the stability of a lava tube cave using Acoustic Emission (AE) tests. Lava tubes are preferred pathways within fluid lava flows, forming cavernous structures after drainage. Once drained, lava tube pathways become lava tube caves. The complex nature of their inner and outer layers, and the unidentified structural stability limits the application of conventional AE instrumentation. In this work, AE instrumentation required for on-site AE tests of lava tube caves is selected through a systematic investigation. In addition, advanced signal processing techniques are developed to analyse the AE signals typical of cavernous structures. On-site AE tests are carried out inside a lava tube cave formed in the 1858 lava flow field at Vesuvius. Six layouts were deployed in different areas inside the lava tube cave and correspond to the inner linings. The structural stability of the linings in different portions of the lava tube cave is investigated using the time-frequency (TF) analysis of the AE signals. Our results show the possible presence of loosened or fractured masses in the vicinity of two layouts, namely the left and right linings of the lava tube cave. However, detailed analysis in the TF domain reveals that the microstructural movements in the layouts due to the presence of loosened mass do not contribute to any severe damage progression, indicating that the lava tube cave structure of our case study is stable. This technique can be effectively and rapidly applied to verify the stability and safety of lava tube caves before being opened to the access of people for touristic or other purposes.

Keywords: lava tube, Vesuvius volcano, lava tube stability, acoustic emission, Hilbert-Huang transform, empirical mode decomposition

5.1. Introduction

The study of drained lava tubes, namely cavities or empty spaces which formed and developed within lava flows or lava flow fields, is facing renewed interest due to the possibility of using these volcanic structures as human shelters in the exploration of extraterrestrial rocky planets. Many of the inner planets of our Solar System comprise volcanic rocks with extensive lava flow fields, where the possibility of finding cavities deep enough to protect future inhabitants from cosmic rays, temperature variability and meteorite fallout is widely recognised (Greeley, 1971a; Sauro *et al.*, 2020; L  veill   and Datta, 2010; Qiu and Ding, 2023). As soon as they form, lava tube caves are affected by the development of internal fracturing, with vertical and transversal fractures due to contraction by cooling (Calvari and Pinkerton, 1999; Calvari *et al.*, 2024). Overlap of multiple lava overflows along the same lava channel can also increase the thickness of the lava pile above a lava tube cave, causing instability and deformation of the roof that, if still plastic, may bend downward and even collapse, obstructing the cavity (Calvari and Pinkerton, 1998). After cooling, the stability of lava tube caves can be affected by several processes, such as weather erosion, root penetration along pre-existing fractures, temperature change and formation of ice in the cracks, overload, etc.

On Earth, the increasing development of geo-tourism and correlated facilities for exploring lava tube caves as extraterrestrial analogues is causing a growing interest in lava tube studies. This renewed interest brings the need to accurately evaluate the safety and stability of lava tube cave structures, especially if they are to be used as shelters for people.

Several non-destructive techniques are available to evaluate the stability of cavernous structures. However, most of them are limited by their scalability or complex instrumentation (Dwivedi *et al.*, 2018; Chanteraud *et al.*, 2021; Nsengiyumva *et al.*, 2021). Acoustic Emission (AE) testing is a passive non-destructive evaluation technique, which is predominantly used in recent times to quantitatively evaluate the stability and predict damage progressions such as crack formation, nucleation of the existing cracks, or presence of loosened masses in the structure.

Acoustic emission phenomena are caused by the propagation of elastic waves, which are associated with a sudden inelastic deformation in materials or structures. In AE testing, these elastic waves are acquired using contact-based piezoelectric sensors or non-contact sensors. In cavernous structures, the AE is associated with microscale deformations in large structural masses, abrupt movement of existing fractures, or generation of new ones (Aoki *et al.*, 1989,

1991; Khoshouei and Bagherpour, 2019; Plenkers *et al.*, 2021). The AE testing is based on acquiring elastic waves and analysing them in the time, frequency, or time-frequency (TF) domains (Pollock, 2018; Grosse *et al.*, 2021). The primary objective of the AE testing is to trace the elastic waves back to their source, thereby using it as a diagnostic and prediction tool to evaluate the stability of the structure.

Due to the complex nature of cavernous volcanic structures and their intricate internal and external layers, overlying irregular surface and often remote access, the conventional instrumentation is not applicable. Therefore, acquiring AE signals is challenging. In cavernous structures or inaccessible regions, the AE signals are generally acquired through waveguides coupled with piezoelectric transducers (Dixon *et al.*, 2003). However, the AE signal acquisition through waveguides is susceptible to large attenuation, especially when a proper acoustic coupling between the waveguide and the structure is not established (Hamstad, 2006; Ono, 2017).

The aim of this study is to develop a suitable AE testing instrumentation to carry out on-site tests in the selected lava tube cave and to investigate its structural stability. First, preliminary experiments are carried out to model the suitable AE instrumentation. Second, a novel signal processing technique is developed and tested to analyse the AE signals in the TF domain. Finally, on-site AE tests are carried out in the lava tube cave. The acquired AE signals are analysed, and the stability of the cave is discussed. In this framework, this paper presents a new survey of a lava tube developed into a lava flow field emplaced during the 1858 eruption of Vesuvius and can be applied to other lava tube caves at this and other volcanoes. A list of all acronyms used in this paper can be found in Appendix 10.

5.2. Methods

5.2.1. Instrumentation for on-site measurement

5.2.1.1. Auto-Sensor-Test for sensor selection

Prior studies report that the AE signals from cavernous structures, or the signals from the damage progressions in materials collected from cavernous structures, are characterised by low amplitude and low frequency (Khoshouei and Bagherpour, 2019; Plenkers *et al.*, 2021). Therefore, it is reasonable to assume that narrowband sensors with high sensitivity are suitable to acquire AE signals in the lava tube cave. The high sensitivity requirement precludes the use of contactless sensors. The possible instability of the lava tube also inhibits the use of multi-

crystal piezoelectric sensors, as these have low shock resistance. Based on the aforementioned rationales and on the availability of commercial piezoelectric transducers, the ISR15 was selected. The ISR15 is a narrowband sensor operating between 50 kHz and 200 kHz with a peak sensitivity of 69 dB re 1 V/(m/s). It is an intrinsically safe sensor, with a piezoelectric crystal protected by a thick ceramic wear plate and completely enclosed with an epoxy coating. The surface area of the transducing element is approximately $1.6 \times 10^3 \text{ mm}^2$. The dimensions of the sensor do not have any adverse effects on the signal acquisition because the source of the AE signals generally acquired in cavernous structures is relatively smaller than the wavelength of interest. The sensor can operate without losing sensitivity at temperatures between -45 °C and 125 °C and has a shock limit of 1000 g, making it potentially suitable for acquiring AE signals from lava tube caves.

To analyse their applicability, a preliminary experiment is carried out in the laboratory setup using samples collected from different interior linings of the lava tube. Eight samples with irregular geometry and surface porosity were collected from the lava tube. The coupling of the sensor to the surface of the specimens and its acquisition capability of the stress wave (acoustic wave) signals are analysed using the “Auto-Sensor-Test” (AST) mode incorporated in the PAC PCI-II data acquisition system (Finkel *et al.*, 2000; Barile *et al.*, 2019). In this approach, the sensor is coupled to the specimen surface and is spiked with a 24 V burst voltage to generate stress waves. The reflected component of these stress waves from the specimen surface is acquired by the sensor. The coupling is analysed by the peak amplitude, signal power, counts, duration, and risetime. Provided there is a good coupling between the sensor and the specimen surface, the different AE descriptors return similar values with low variance.

Figure 59 shows the eight samples collected inside the studied lava tube cave. The surface porosity of the samples is measured by adopting the k-means image clustering approach (Arthur and Vassilvitskii, 2006). This approach is commonly used in several image processing techniques, and the procedures are detailed by Angelin *et al.* (2017) and Venkataraman *et al.* (2007).

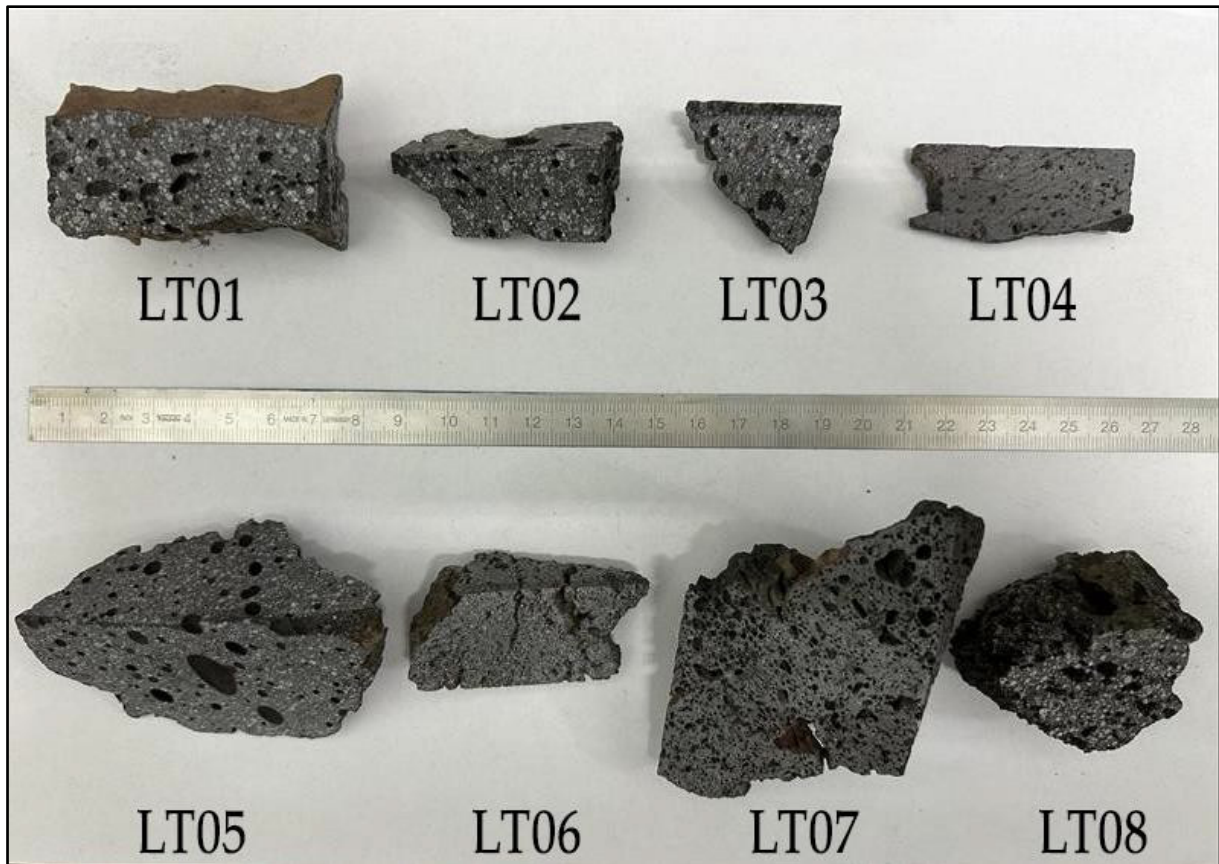


Figure 59. Test samples collected along the 1858 lava flow field at Vesuvius.

The surface micrographs of the specimens are acquired using an optical microscope and have an area of 225 mm^2 (see Fig. 60a). The acquired images are then segmented into a predefined number of clusters using an unsupervised k-means clustering technique (see Fig. 60b). The clusters containing the porous regions are then identified and extracted (Fig. 60c). The surface area of the porous region is calculated by binarizing the image (Fig. 60d) and subsequently, the surface porosity percentage is calculated. Figure 60 presents the sequence of operations followed to measure the surface porosity. The image processing technique is further explained in a previous study (Barile *et al.*, 2025).

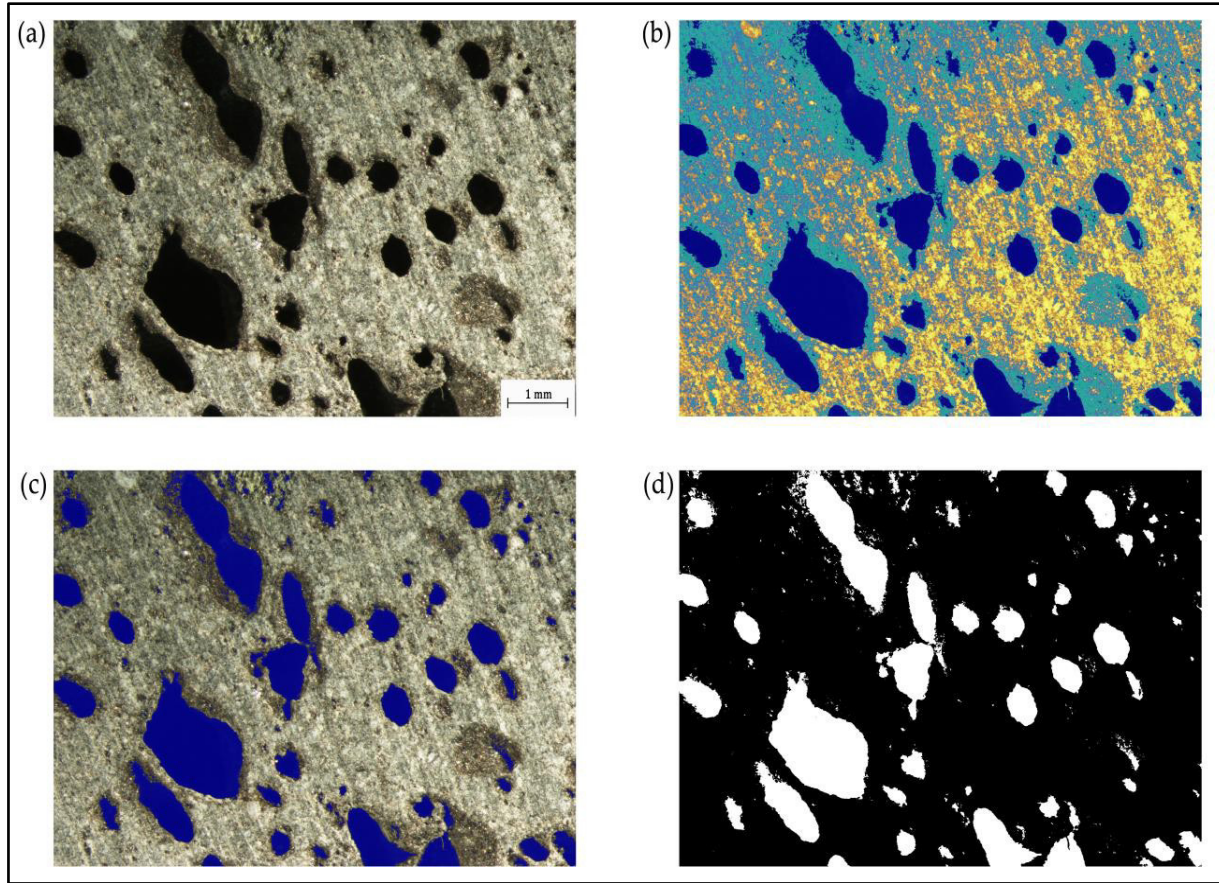


Figure 60. (a) Microscopic image of the lava tube specimen; (b) Image segmented into 5 clusters; (c) Extracted cluster of interest containing porous regions; (d) Converting the image into black and white to measure the surface area of the porous regions.

The porosity percentages of all the lava tube specimens are reported in Figure 61a, and their disparities are presented in the form of a boxplot in Figure 61b. Despite the variance in porosity, it can be noted that the maximum surface porosity of the lava tube specimens is 111.15 mm², which is sufficiently smaller than the transducing surface area of the ISR15 sensor. Nevertheless, the influence of porosity percentage on the sensor coupling and AE signal acquisition is analysed by propagating stress waves using the AST mode.

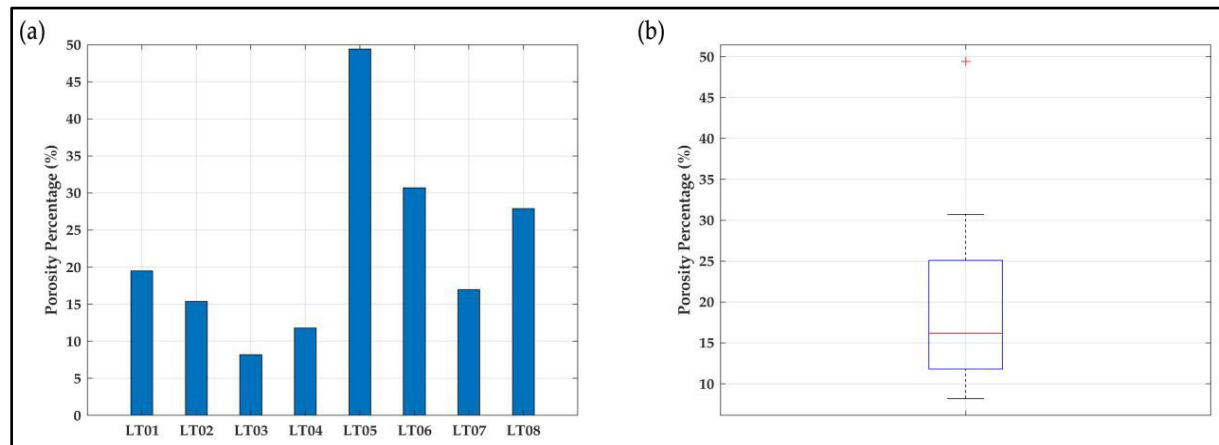


Figure 61. (a) Porosity percentage of the lava tube specimens calculated using the k-means image segmentation method; (b) Boxplot showing the distribution of the porosity percentage.

The different stress wave (acoustic emission) descriptors reflected from the surface of the lava tube specimens during the AST tests are analysed and reported in Table 8.

Table 8. Descriptors of the reflected acoustic emission signal from each specimen.

Specimen Name	Peak Amplitude <i>dB</i>	Signal power <i>a.u.</i>	Counts	Duration <i>μs</i>	Risetime <i>μs</i>
LT01	91	0.1365	26	51.20	20.00
LT02	90	0.1314	26	51.80	20.60
LT03	91	0.1417	24	50.00	18.80
LT04	91	0.1271	24	42.60	18.80
LT05	90	0.1510	25	43.20	19.80
LT06	91	0.1404	24	42.40	18.80
LT07	91	0.1117	25	43.00	19.40
LT08	91	0.1561	25	57.60	19.00
<i>Mean</i>	<i>90.75</i>	<i>0.1370</i>	<i>24.8750</i>	<i>47.7250</i>	<i>19.4000</i>
<i>Std</i>	<i>0.4629</i>	<i>0.01396</i>	<i>0.8345</i>	<i>5.7171</i>	<i>0.6761</i>
<i>Variance</i>	<i>0.2143</i>	<i>0.0002</i>	<i>0.6964</i>	<i>32.6850</i>	<i>0.4571</i>

As observed in Table 8, there is no disparity between the AE descriptors of the different specimens. In fact, the standard deviation and the variance of the descriptors are quite small, indicating that there are no statistically significant differences between the descriptors acquired from the eight specimens. Essentially, the reflected AE signals acquired from the AST test are not unaffected by the disparity in the surface porosity between the specimens. This confirms the good acoustic coupling between the sensor and the specimens. This is because the transducing surface of the sensor is sufficiently larger than the surface porosity, thus enabling proper acquisition of the AE signals. Besides, the sensor comprises a single piezoelectric crystal, thus requiring minimal contact with the transducing surface. Based on this analysis, it can be determined that the ISR15 sensor is suitable for acquiring on-site AE signals in lava tube caves.

5.2.1.2. Hammer impact tests to define acquisition parameters

Since the cavernous structures generate AE signals with low frequency, a piezoelectric sensor operating in the low frequency band with high sensitivity has been selected. However, this increases the probability of acquiring low-frequency noise, signals with a low signal-to-noise ratio, and reverberations during onsite tests. Consequently, it is imperative to define the optimal acquisition parameters to avert these issues. To this end, preliminary AE data acquisition tests are carried out in the laboratory premises. Two ISR15 sensors are mounted on a stable concrete wall and are connected to the data acquisition system through preamplifiers.

Impact events are then artificially created by using a wooden hammer away from the sensor. The hammer impacts generate stress waves in the walls, which propagate as acoustic emission signals and are acquired by the sensors. The tests are repeated for different parameter ranges, as presented in Table 9.

Table 9. Parameters evaluated for acoustic emission signal acquisition.

Parameters	Minimum	Maximum
Acquisition Threshold	16 dB	45 dB
Preamplifier	20 dB	60 dB
Lowpass Filter	100 kHz	200 kHz
Highpass Filter	200 kHz	600 kHz
Peak Definition Time (PDT) Hit	20 μs	1000 μs
Definition Time (HDT)	100 μs	2000 μs
Hit Lockout Time (HLT)	300 μs	5000 μs

The objective of the test is to record one AE event per hammer impact (Dvořák and Topolář, 2021). This ensures that unnecessary noise and reverberations are not registered by the sensor. Through a trial-and-error process, the optimal parameters to acquire AE signals are determined. These parameters were found to be as follows: threshold – 26 dB; preamplifier – 40 dB; bandpass filters – 100 kHz/400 kHz; PDT – 100 μs ; HDT – 600 μs ; and HLT – 1000 μs . As the ISR15 sensor is fully encapsulated in epoxy coating, it offers radio frequency and electromagnetic immunity, thus permitting the low-threshold signal acquisition. The selection of other parameters is also influenced by the large volume of the structure. Nevertheless, with these defined parameters, each hammer impact event is registered as one acoustic event. Figure 62 shows the frequency spectrum of some sample signals acquired during the hammer impact tests, which are analysed using FFT.

Since the frequency content of the signals is localised within a narrow frequency band due to the nature of the source and the operating frequency of the sensor, frequency analysis alone is insufficient. This work proposes a novel time-frequency (TF) analysis, which will be explained in the following sections.

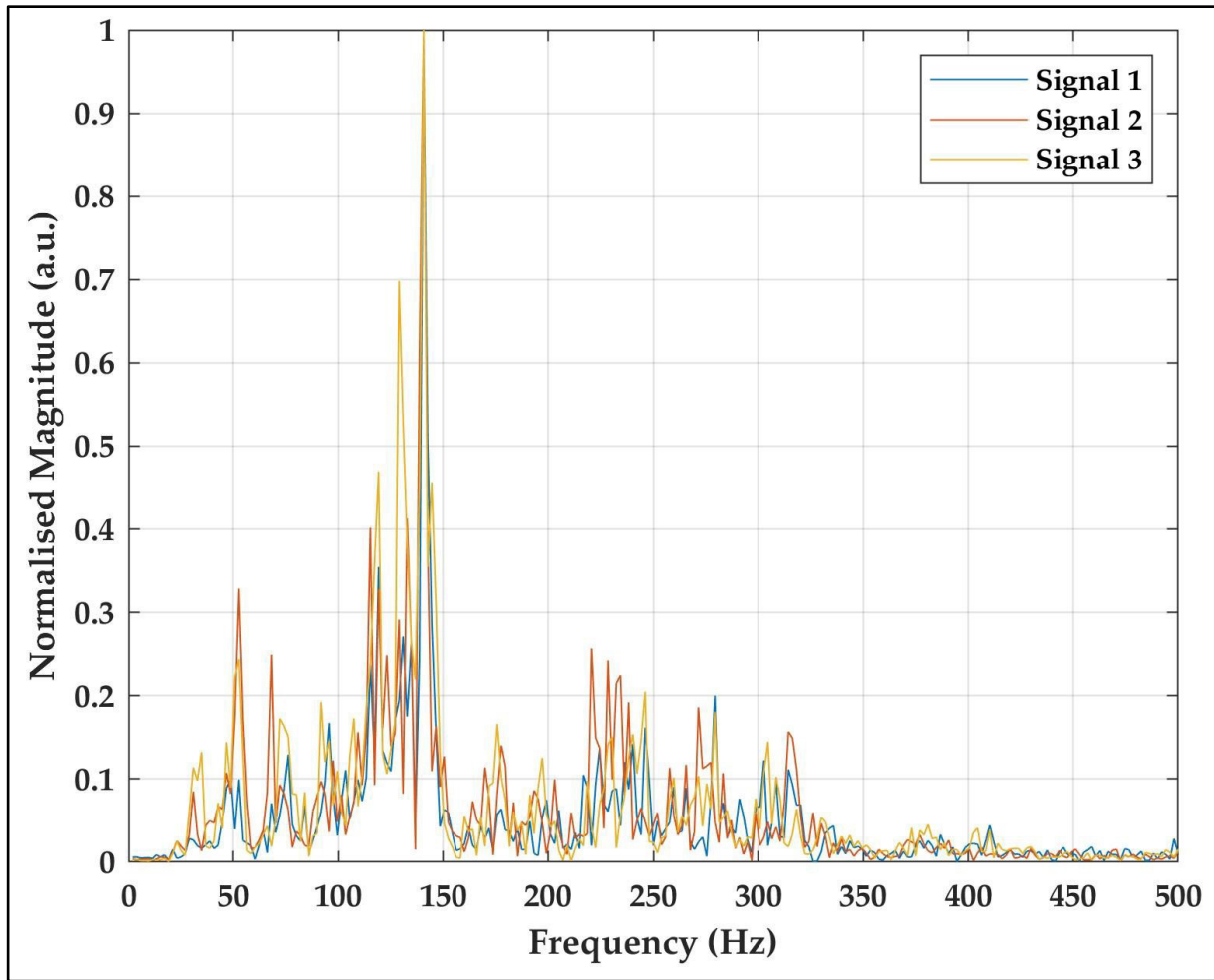


Figure 62. Acoustic emission signals acquired from the hammer test using the optimal parameters.

5.2.1.3. Data acquisition setup

We chose to use the PAC PCI-II data acquisition system for the on-site AE tests. The setup comprises a four-channel acquisition system, in which the sensors can be powered by a 2/4/6 integral preamplifier. A 2-meter BNC (Bayonet Neil-Concelman) coaxial cable is used to power the preamplifier and to register the acquired AE signals. The acquired AE waveforms are sampled at a rate of 2 MHz with a sample length of 1024.

5.2.2. Novel signal processing technique for Time-Frequency analysis

AE signals are typically analysed in the TF domain using short-time Fourier transform (STFT), continuous wavelet transforms (CWT), and discrete wavelet transforms (DWT), or wavelet-free transforms such as empirical mode decomposition (EMD) (Mousavi *et al.*, 2016; Ebrahimian *et al.*, 2019; Zhang *et al.*, 2021; Dong *et al.*, 2023). STFT is limited by a fixed resolution but is particularly useful to represent continuous and quasi-stationary signal segments (Maia *et al.*, 2024). While CWT and DWT allow customised TF resolution using different wavelet functions, the selection of wavelets is not trivial. Therefore, in this work, a

novel approach based on EMD and on the Hilbert-Huang transform (HHT) is used. The traditional procedures of EMD and HHT are extensively discussed in the literature (Huang and Shen, 2014; Zhang *et al.*, 2016; Yuan *et al.*, 2018; Quinn *et al.*, 2021). Therefore, this section only details the adopted novelty.

The novel procedure employed is as follows. The AE signals in the time domain are decomposed into different intrinsic mode functions (IMFs) and residuals using EMD. Figure 63 shows a sample AE signal acquired from the lava tube cave, which is decomposed into 7 IMFs (5 out of 7 IMFs are presented in Figure 63). Then, by comparing each IMF with the original signal using permutation entropy (PE), the IMFs with significant signal content are selected. PE is the measure of complexity in the time series signal and is used in prior studies to select the IMFs (Si *et al.*, 2014; Mohd Jaafar *et al.*, 2019; Koppolu and Chemmangat, 2023). It is hypothesised that the IMFs with low PE contain significant signal content, as they are characterised by low complexity. In this context, the PE of the raw signal and decomposed IMFs are compared. Finally, the IMFs sharing similar PE of the raw signal with a significance level of 95% is selected.

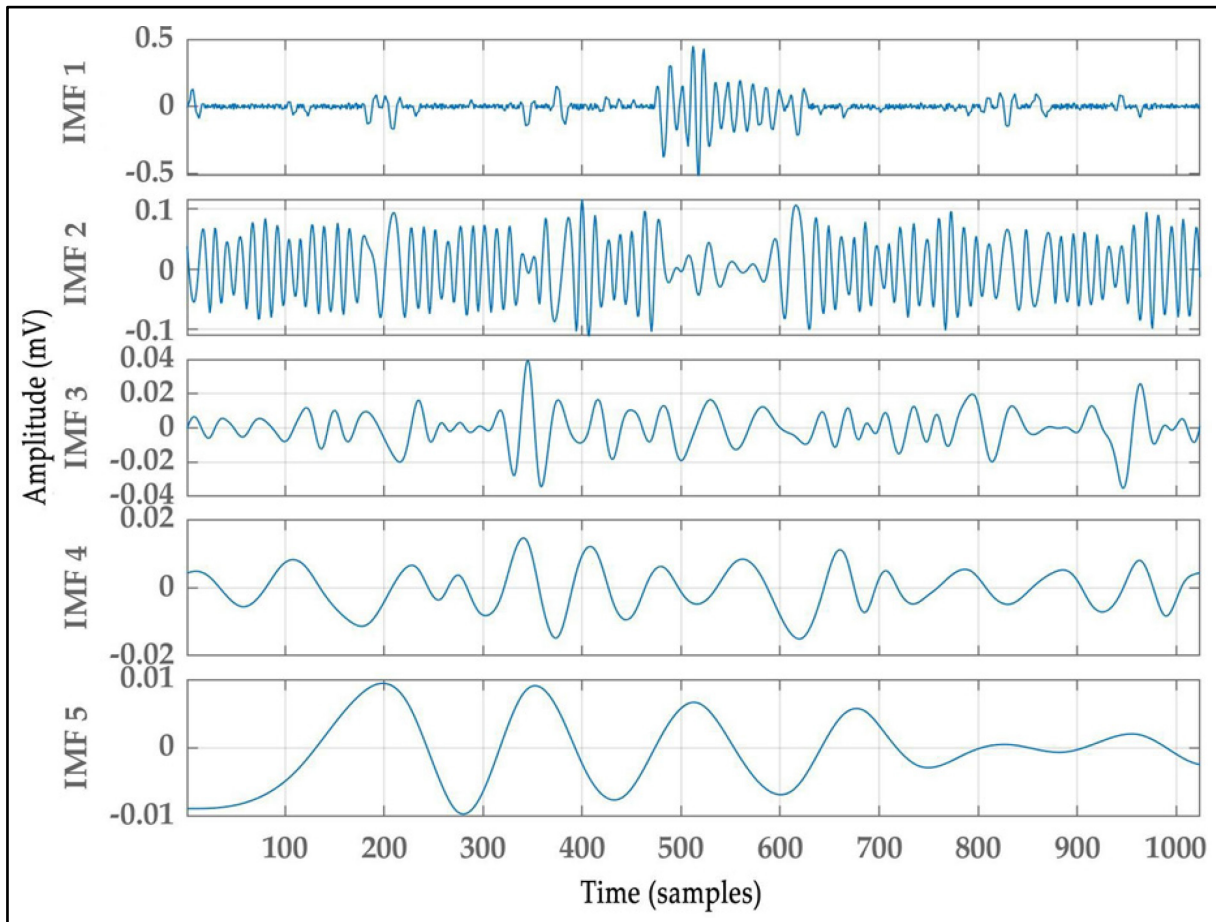


Figure 63. IMFs of the decomposed acoustic emission signal.

The selected IMFs are summed up to reconstruct the time series signal. This allows the selection of significant signal content without the necessity of windowing or selecting wavelet, which are essential in STFT, CWT, and DWT. The resulting signal has an improved signal-to-noise ratio compared to the original signal. Figure 64 shows the raw and denoised signal.

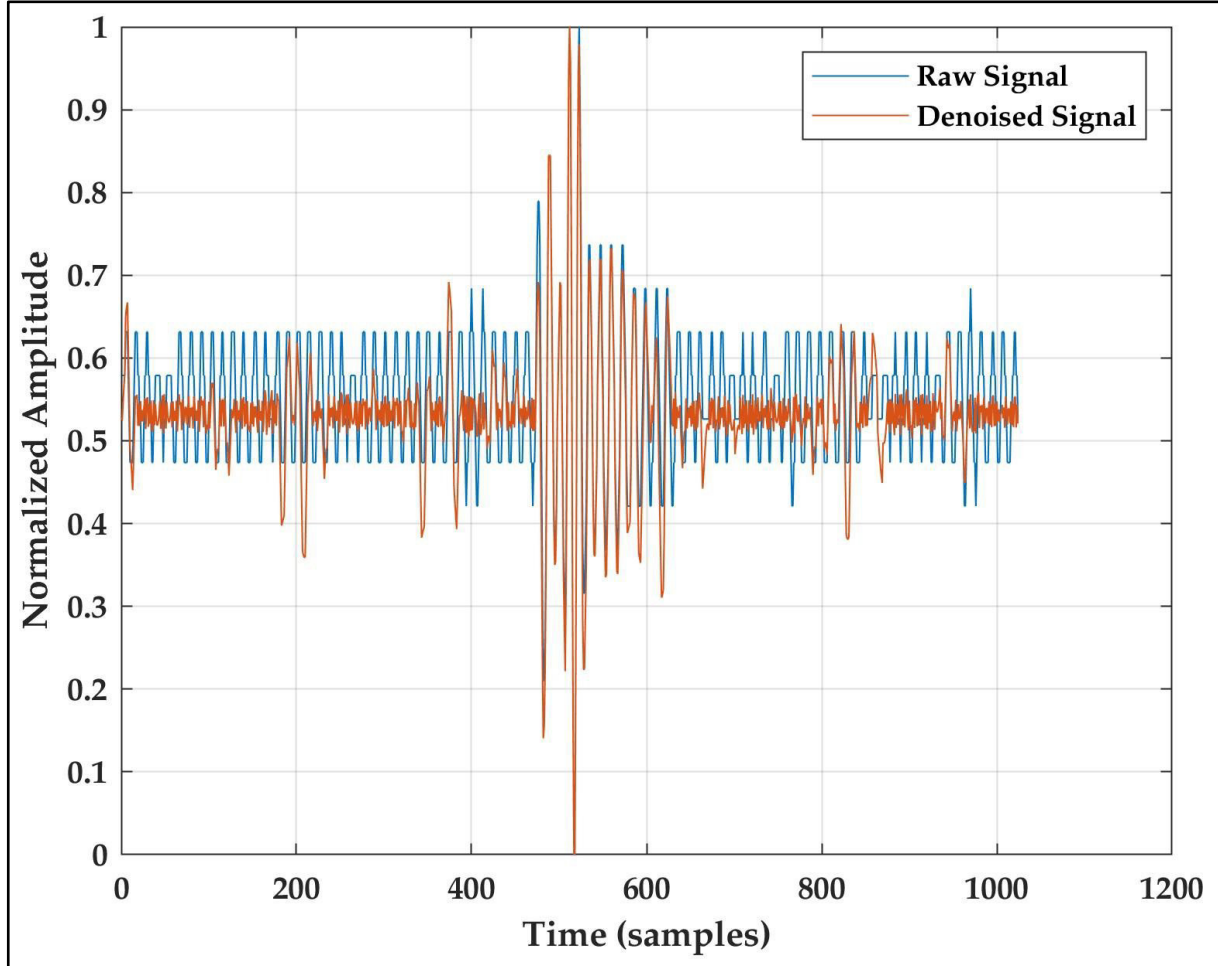


Figure 64. Raw acoustic emission signal and its denoised form in Time Domain.

The frequency spectrum of the denoised AE signals is then analysed using the fast Fourier transform (FFT) method. This reveals the information about the central frequency or the principal frequency bands of the signal. Figure 65 presents the FFTs of the raw and denoised sample signal. In the example signal, it can be observed that most of the frequency content of the signal is localised between 50 kHz and 200 kHz. Therefore, the TF analysis is carried out strictly within this frequency limit. This improves the resolution of the signal content in the TF domain.

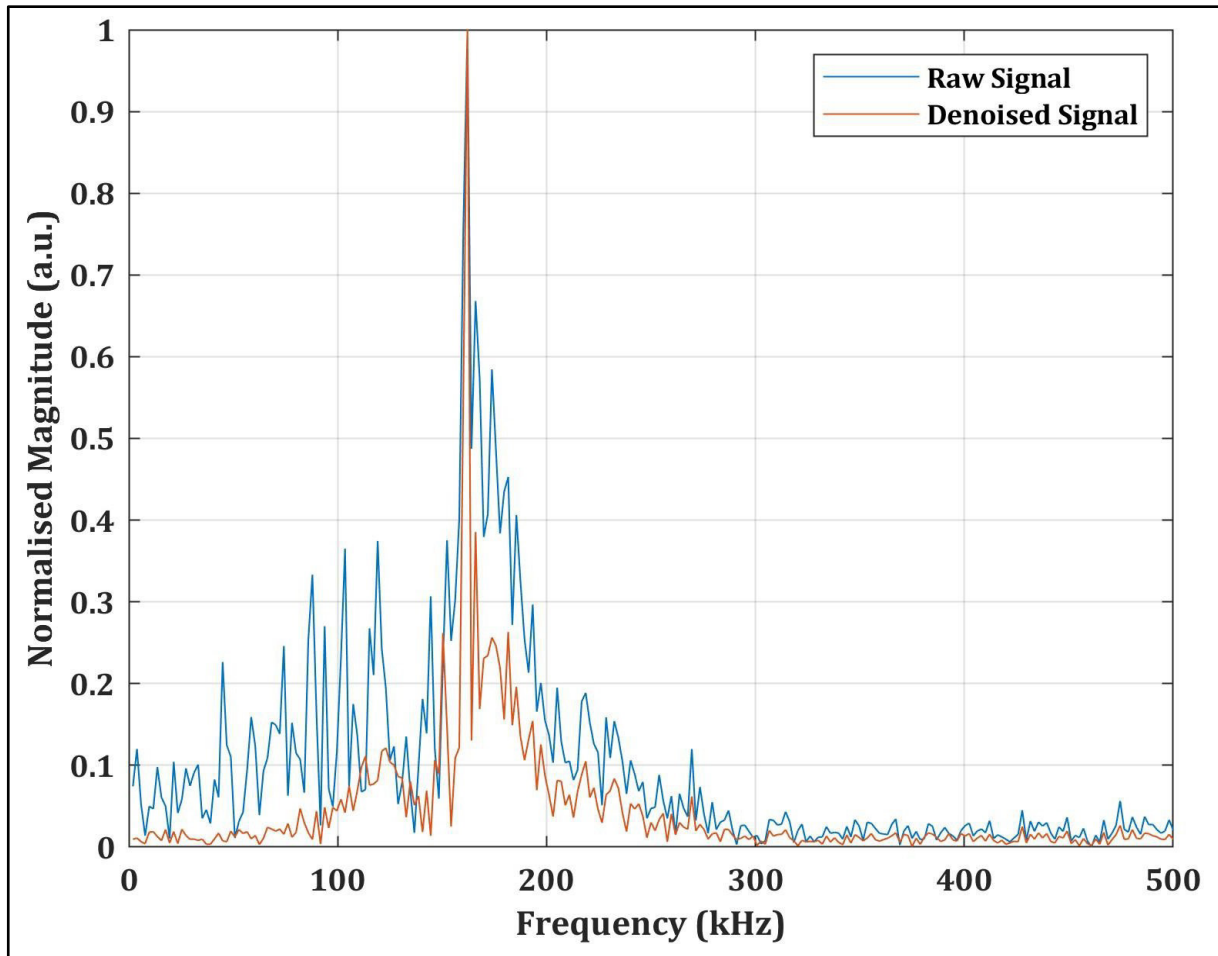


Figure 65. FFT results of the raw and denoised AE signal.

The TF analysis is carried out using the HHT method. As mentioned earlier, the documentation for HHT is well-established in the literature (Huang and Shen, 2014). In HHT, the instantaneous frequency, instantaneous amplitude, and their temporal information are extracted while working within the predefined frequency limits. For the sample signal, the HHT is applied within the frequency limits of 50 kHz and 200 kHz. The results are presented in Figure 66.

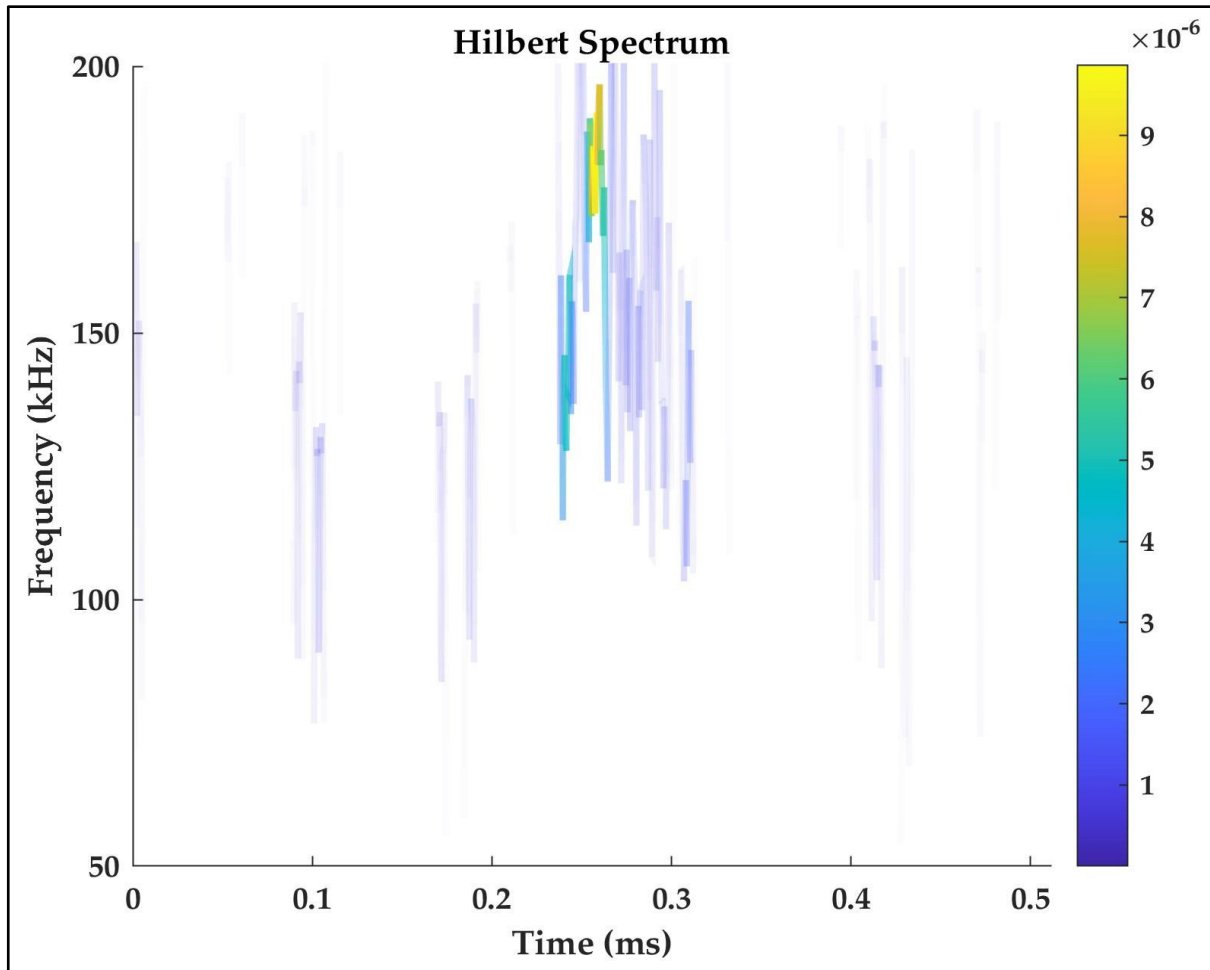


Figure 66. HHT of the sample signal showing the temporal changes in the instantaneous frequency and instantaneous amplitude (colour bar).

This novel approach uses PE-based EMD and frequency-localised HHT to extract the principal spectral contents in the TF domain of the AE signal. In addition, this method is less likely to be susceptible to variations in the signal-to-noise ratio and thus can be used to analyse the AE signals acquired from high noise environments such as cavernous structures. A schematic of the proposed approach is presented in Figure 67.

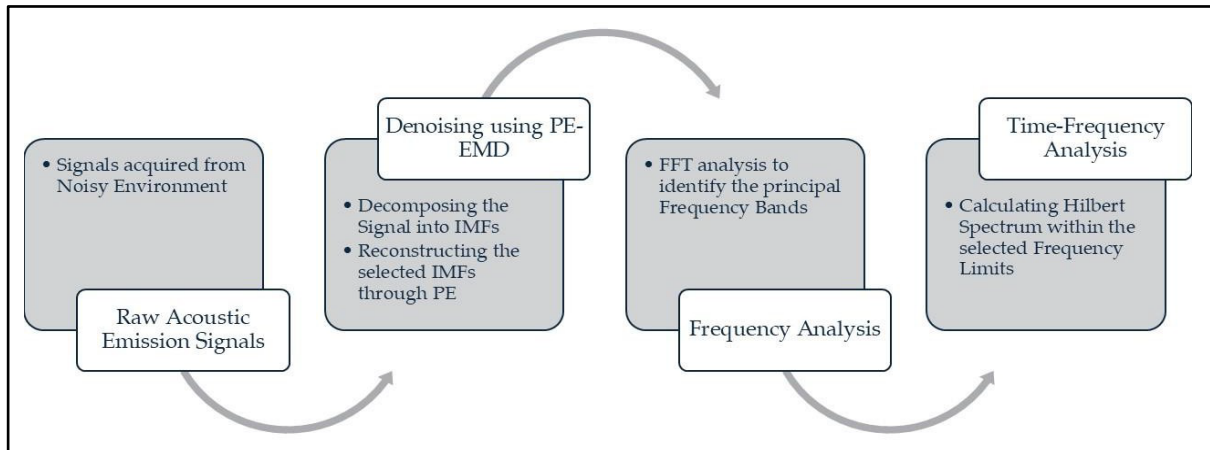


Figure 67. Schematic of the proposed signal processing technique for Time-Frequency analysis.

5.3. On-site tests in a lava tube at Vesuvius

The technique described here was applied to a lava tube formed during the 1858 eruption at Somma-Vesuvius. During its last eruptive cycle from 1631 to 1944, Vesuvius showed a mostly effusive activity intercalated by mildly explosive eruptions. Numerous lava flows were emplaced during this eruptive cycle on the western and southern flanks of the volcano. The 1858 eruption lasted nearly three years, from 27 May 1858 to 12 April 1861 and is one of the longest eruptions taking place during the 1631-1944 eruptive cycle. It erupted 178 million cubic

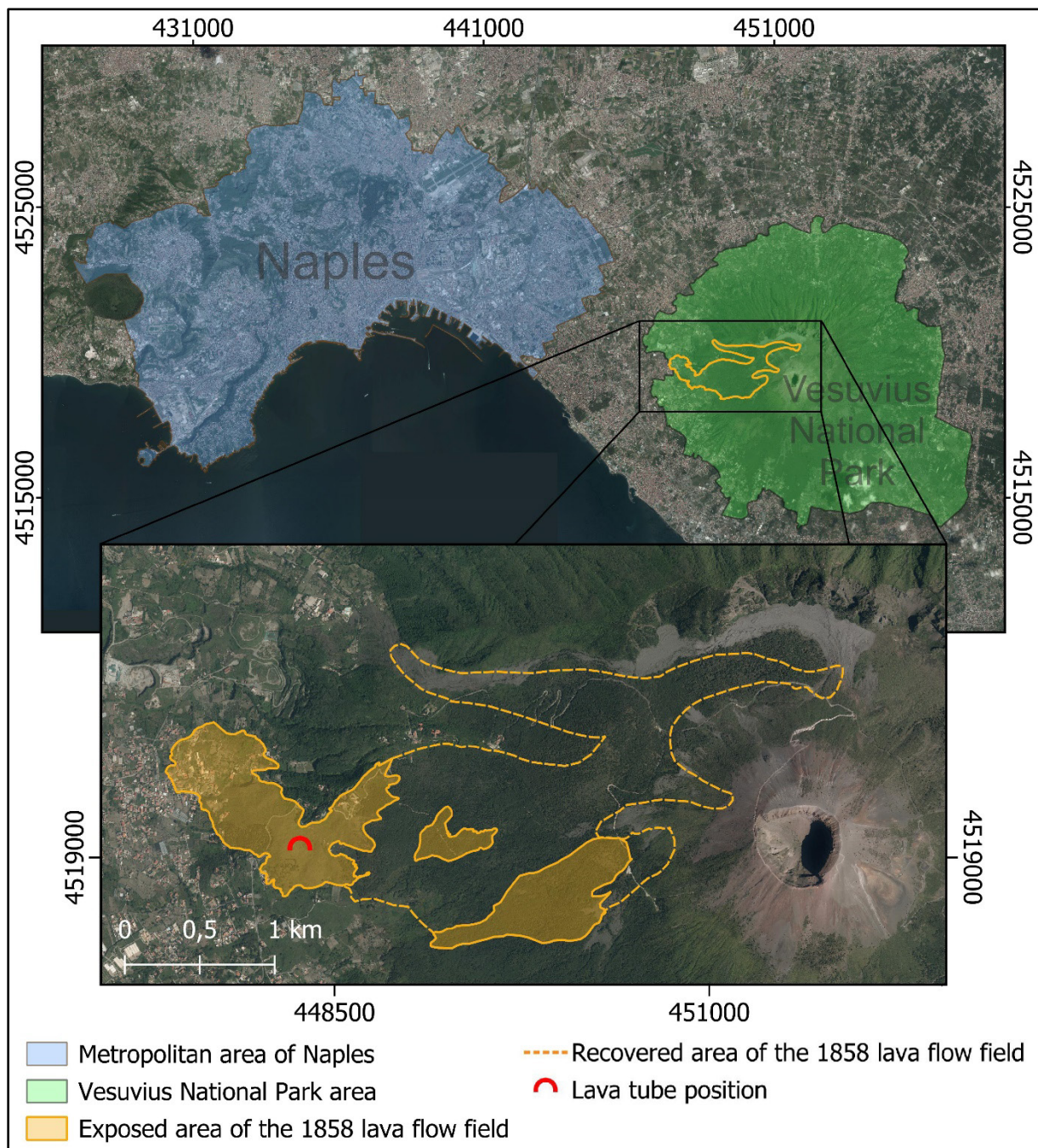


Figure 68. Inundation area of the 1858 lava flow field and position of the studied lava tube within the flow field.

metres of lava, emplacing a three kilometres long lava flow field on the western flank of the volcano (Fig. 68; Lemaire *et al.*, 2024). The lava flow filled two deep valleys that were incised into the AD 79 eruption deposits. The lava erupted during the 1858 eruption is phonotephritic in composition and has a large number of phenocrysts, around 40%, mostly leucite but also clinopyroxene and plagioclase. Several lava tubes were formed in the 1858 lava flow field, and the largest lava tube found in the lava flow field was first described by Malladra (1917). More than a century later, this lava tube was rediscovered and scanned in 3D by Lemaire *et al.* (2024). This lava tube was formed by the inflation of a small pāhoehoe lava flow breakout. Its activity ended when the tube drained completely, leaving an empty cavity. It is 30 metres long and reaches a width of up to 17 metres. The mean thickness of the roof is about 2.5 metres. Several internal features are observed inside the tube, some of which can affect the stability of the tube structure, such as wall linings and gas blisters.

5.3.1. Experimental setup

Based on the analysis described in the previous sections, the instrumentation and the acquisition parameters were defined (Table 10). The AE tests are carried out by installing two ISR15 sensors in different inner linings of the lava tube. A thin layer of silicone grease is applied between the sensor and the lining surface to minimise the acquisition of reverberations and improve the acoustic coupling. AE signals above the detection threshold are amplified using an external preamplifier and stored as events using the PAC PCI-II data acquisition system. The signal waveforms are sampled at 2 MHz over 1024 samples.

Table 10. Acoustic emission instrumentation and acquisition parameters for onsite tests.

Instrumentation		General Description
Sensor	ISR15	Intrinsically safe resonant frequency sensor Operating frequency – 50 – 200 kHz Resonant Frequency – 150 kHz
Preamplifier	2/4/6 Preamplifier	Voltage Preamplification by 40 dB
Acquisition Parameters		
	Detection Threshold	26 dB
	Analogue Filters	100 kHz / 1 MHz
	Digital Filters	100 kHz / 400 kHz
	Pre-trigger	256 (50% of the Signal Length)
	Sample Rate	2 MHz
	Signal Length	1024 samples

The experiments were carried out in five different areas inside the lava tube cave shown in Figure 69. Details about the layouts are presented in Table 11 and displayed in Figure 69.

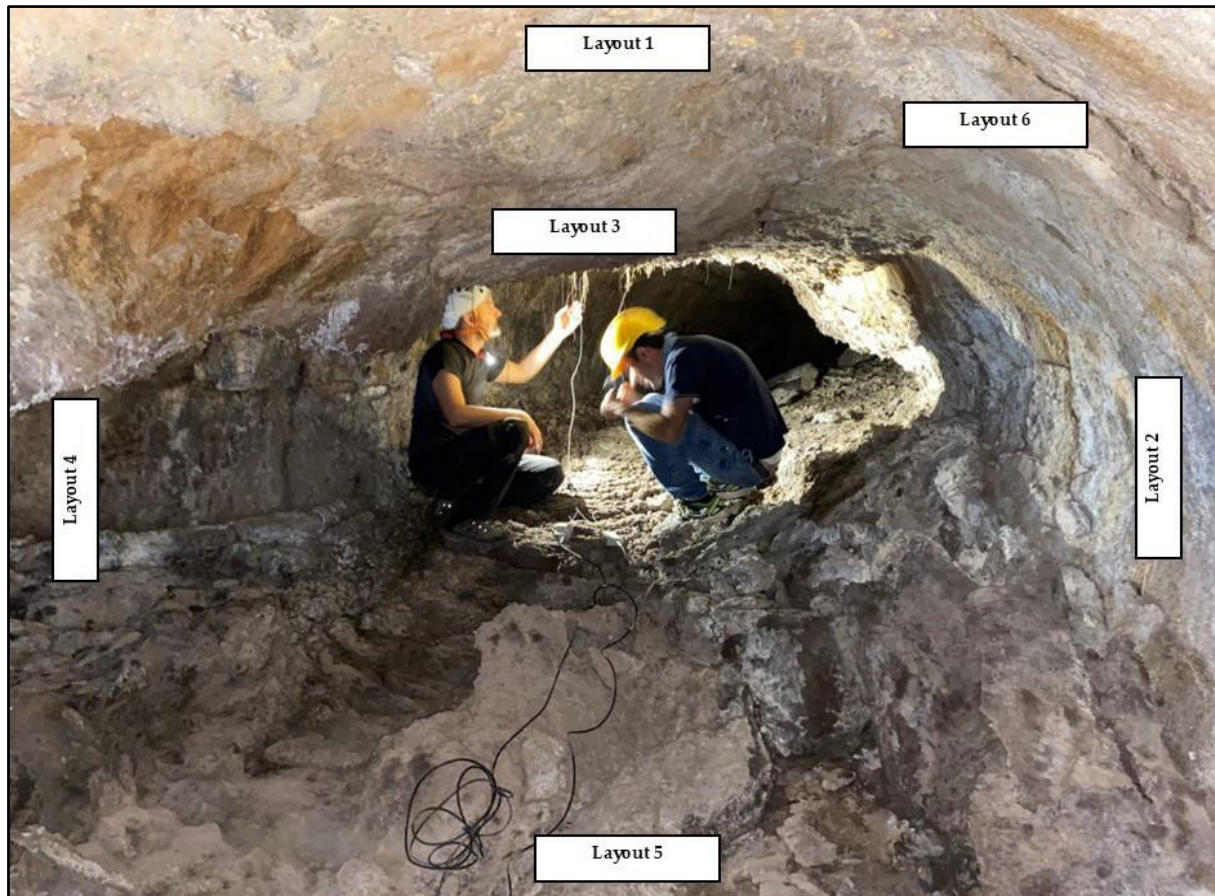


Figure 69. Lava tube at Vesuvius, with numbers indicating the different areas analysed for stability and listed as layouts in Table 11.

In each area, the AE tests are carried out by placing the sensors approximately 300 mm apart. The separating distance between the sensors is only approximated due to the unstable structure of the lava tube linings, which did not permit the precise positioning of the sensors at the desired locations. As minimal contact is sufficient for the ISR15 sensor (see Section 5.2.1.1), AE signals are acquired from the structures despite the unevenness.

Table 11. Areas of deployment of the layouts inside the lava tube.

Layout Number	Description
Layout 1	Stalactites
Layout 2	Left Lining Structure
Layout 3	Central Lining
Layout 4	Right Lining Structure
Layout 5	Floor
Layout 6	Roof

5.3.2. Signal acquisition and preprocessing

The sensors are placed in position for 500 s, and the AE signals are acquired continuously. If no signals are registered during the first 100 s period, then the tests are interrupted and repeated. After acquiring the AE signals, they are pre-processed prior to analysis. Following the recommendations of AE testing, events with counts less than 3 are discarded as noise. Counts are the total number of instances of the signal amplitude crossing the detection threshold of 26 dB. The signal waveforms are then filtered using the built-in analogue and digital filters of the data acquisition system. The filtered signals are sampled at 2 MHz and stored for analysis.

From Layout 3 (Central Lining) and Layout 5 (Floor), no AE signals are recorded. All the acquired signals from the other layouts are analysed in the time, frequency, and TF domains. The results are presented and discussed in the subsequent sections.

5.4. Results and discussion

5.4.1. Time Domain Analysis

All the acquired AE signals are analysed in the time domain using traditional descriptors. In this context, the peak amplitude and counts of the AE signals are analysed. Peak amplitude is defined as the largest signal amplitude measured within 100 μ s PDT and is

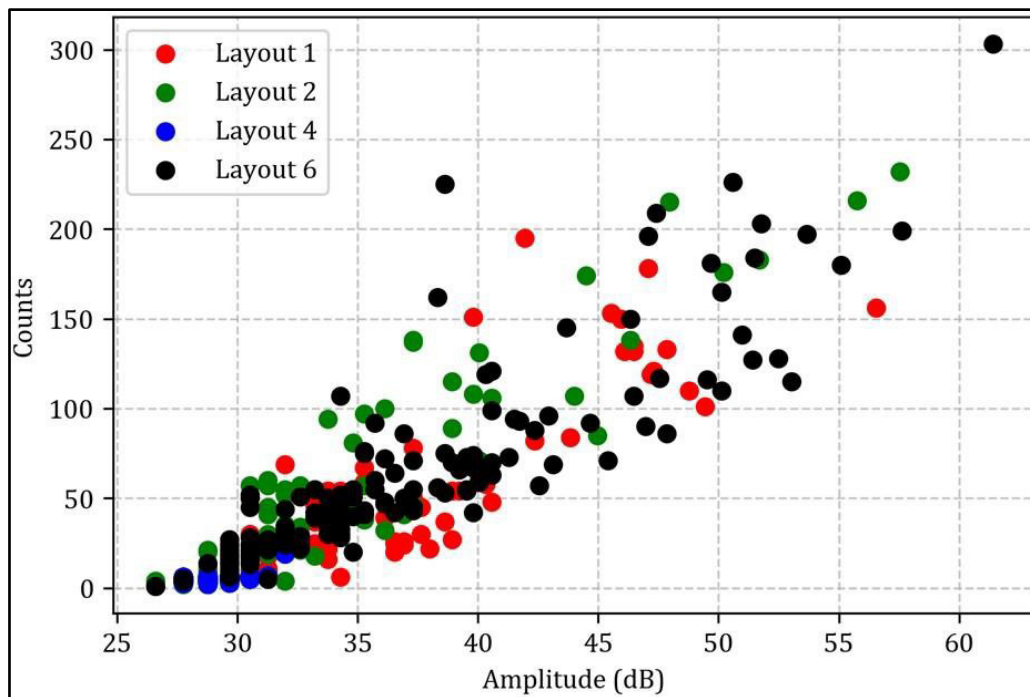


Figure 70. Amplitude-count distribution of acoustic emission signals acquired from the different layouts.

expressed in dB. Figure 70 shows a comparison of these two AE descriptors acquired from all the layouts.

The absence of AE signals registered in Layouts 3 and 5 (Central lining and floor) indicates their good stability. This could also suggest the absence of loosened mass, microstructural motions, or moving bodies in their vicinity.

From the amplitude-count distribution of the AE signals from the other layouts, the following observations are made. About 98.46% of AE signals have peak amplitude less than 35 dB and counts less than 70. Notably, all the signals acquired from Layout 4 fall within this range. During the acquisition period, the total number of signals acquired from Layouts 1 and 6 is merely 81 and 166, respectively, compared to 5328 and 3105 signals from Layouts 2 and 4. However, the average counts recorded by the signals differ considerably. These results are summarised in Table 12.

Table 12. Time Domain analysis of acoustic emission signals acquired from different layouts.

Layouts	Total Number of Signals	Signals with Amplitude >35 dB	Average Counts per Signal
Layout 1	81	37	50.57
Layout 2	5328	24	5.85
Layout 4	3105	0	4.81
Layout 6	166	73	57.77

The general behaviour of AE signals in terms of their amplitude and counts has been observed in other cavernous structures in the literature (Aoki *et al.*, 1990; Hata *et al.*, 1998). Previous studies also reported that the continuous acquisition of AE signals from a cavernous structure indicates the presence of loosened zones or loosened masses around that structure (Aoki *et al.*, 1990; Hata *et al.*, 1998; Fan *et al.*, 2011). However, the very low average count of the AE signals must be taken into consideration. These signals cannot be simply dismissed as noise. Preliminary tests reported in Section 5.2.1.2 indicate that the instrumentation is calibrated for stable structures and thus acquiring noise is unlikely. AE signals with low counts are often associated with friction between microstructural elements, which generally do not contribute to any catastrophic events in materials or structures. Therefore, it is safe to assume that these signals do not indicate an accumulation of damage or failure in the structure.

It should be noted that the layouts with a stable structure generally do not generate any AE signals (Layouts 3 and 5 – Central lining and floor), or a very low number of AE signals (Layouts 1 and 6 – stalactites and roof). Since the layouts 2 and 4 correspond to the side walls

of the lava tubes, the continuous generation of AE signals with very low average counts was expected, as they are possibly bounded by loosened masses created by the lava flow. Nevertheless, it is important to analyse the AE signals in their frequency and TF domains before drawing conclusions.

5.4.2. Frequency Domain analysis

To explicitly identify the differences between the AE signals acquired from different layouts and differentiate their sources, the signals are analysed in the frequency domain using the FFT method. FFT results for a sample signal taken from each layout are presented in Figure 71. The signals are selected based on the overall amplitude-count distribution of their group. AE signals which share similar amplitude-count features within a group are selected for analysis.

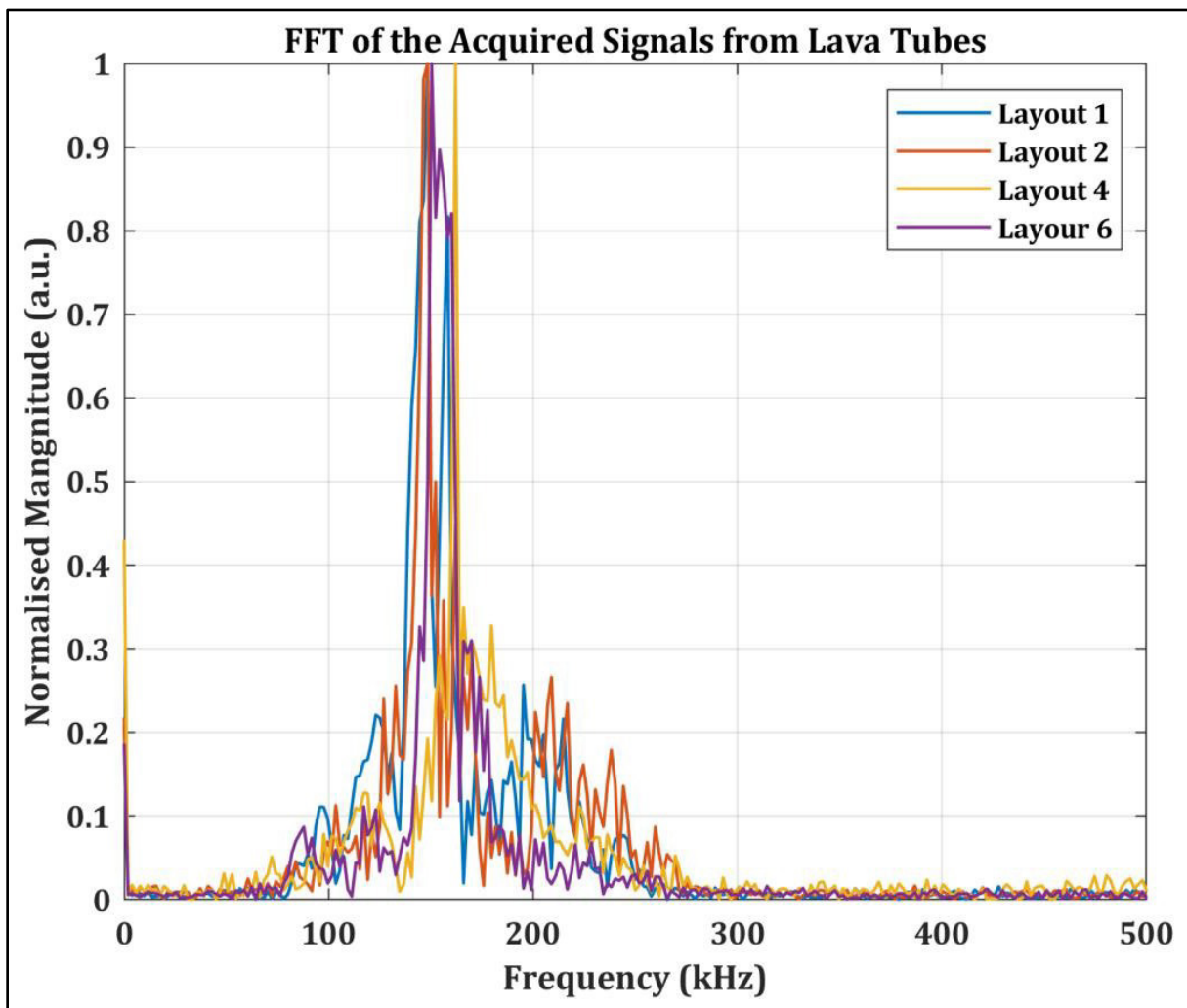


Figure 71. Fast Fourier Transform results of acoustic emission signals acquired from the different layouts.

No explicit differences can be observed in the frequency characteristics of these signals. All the signals have their frequencies centred between 50 kHz and 160 kHz, while the peak frequencies of the signals from all layouts are between 150 kHz and 160 kHz. Despite providing the overall frequency distribution of the AE signals, it is unable to differentiate between the signals from different sources. Therefore, taking the FFT results as a basis, the central frequencies are adopted for the TF analysis method developed and reported in Section 5.2.2. The TF domain results are presented and discussed in the next section.

5.4.3. Time-Frequency Domain analysis

The signal processing technique is essential for this case study, as the signals acquired from the different layouts have very low frequencies, and the sensor operates within a very narrow frequency band. The TF results in the Hilbert spectrum shed more light on the necessity of this procedure. The Hilbert spectrums of the representative AE signals of each layout (see Fig. 71) are presented in Figure 72.

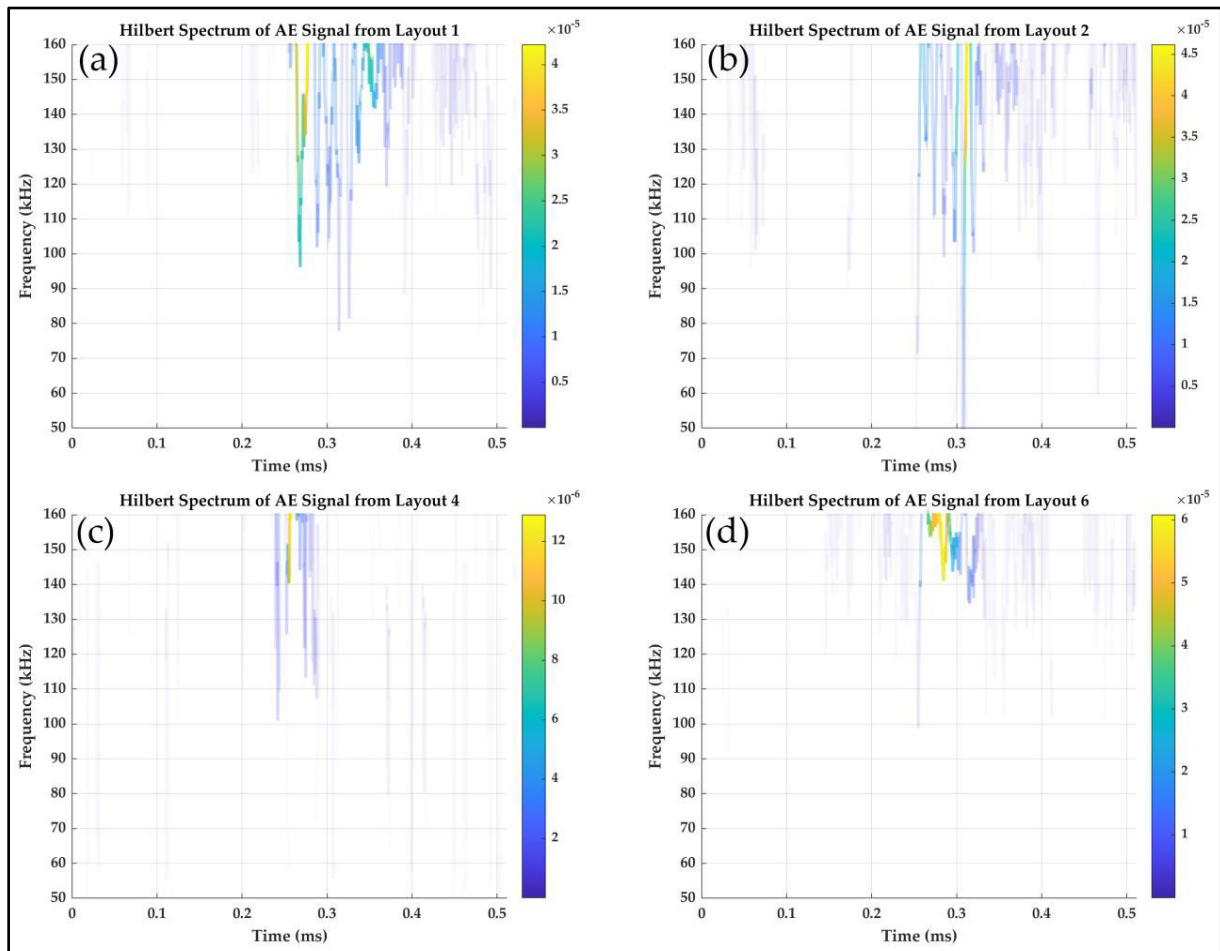


Figure 72. Hilbert Spectrum of the acoustic emission signals acquired from (a) layout 1; (b) layout 2; (c) layout 4; (d) layout 6.

The Hilbert spectrum reveals critical differences between the signals acquired from the different layouts. The frequencies of the AE signals acquired from layouts 1 (Fig. 72a) and 2 (Fig. 72b) are distributed longer in the time domain compared to the signals acquired from Layouts 4 (Fig. 72c) and 6 (Fig. 72d).

Furthermore, the frequency distributions in the time domain also show significant differences. The signal from layout 1 has frequencies ranging from 95 kHz to 160 kHz and is distributed in the time domain between 0.25 ms and 0.4 ms. While the signal from layout 2 also has the frequencies distributed in the same time domain, the frequencies are between 100 kHz and 160 kHz. At 0.31 ms, however, a wider frequency distribution is observed. The signal from layout 4 has much narrower frequencies distributed over a very short period. The signal from layout 5 exhibits highly localised TF behaviour with the frequencies between 140 kHz and 160 kHz distributed over a period of 0.25 ms to 0.325 ms.

As seen in Figure 72, the Hilbert spectrum analysis can highlight the differences between the AE signals acquired from different layouts, indicating that these signals originate from different sources. Nevertheless, it is imperative to analyse whether all the AE signals from each layout share the same TF characteristics. Analysing each signal as spectrograms to find the similarities is time and cost intensive. To simplify this process, statistical tools are employed.

5.4.4. Similarity analysis using statistical tools

Two statistical analyses are carried out to find the similarities between the AE signals from the same source using two different methods. The two methods used for similarity analysis are the Bhattacharyya Coefficient and the Kullback-Leibler (KL) divergence. These are statistical distance parameters which are used in image processing applications to determine the similarity between the probability distributions in a dataset. In this case, these distance metrics are used in two ways: (1) to find the similarity between the AE signals in their time domain and (2) to find the similarity between their instantaneous frequencies of the Hilbert spectra.

Both methods are popular statistical tools, and detailed explanations can be found in the literature (Hall, 1987; Djouadi *et al.*, 1990; Theodoridis and Koutroumbas, 2003). BC values range from 0 to 1, where 0 indicates high dissimilarity and 1 indicates high similarity. Conversely, a lower KL value indicates high similarity, while the larger values indicate dissimilarity.

The BC and the KL values of each signal in the layout are compared with a reference signal both in the time and frequency domains. The signals presented in Figures 71 and 72 are taken as the reference signal in each layout. The average values in each layout are reported in Table 13. In Table 13, the BC_t and KL_t refer to the similarity indices calculated in the time domain signal, while BC_f and KL_f refer to the similarity between the instantaneous frequencies of the signals, as calculated using the HHT.

Table 13. Bhattacharyya Coefficient and KL Divergence of the AE signals compared in Time and Frequency Domains.

Layout	Time Domain		Frequency Domain	
	BC_t	KL_t	BC_f	KL_f
Layout 1	0.74	2.77	0.62	5.14
Layout 2	0.76	3.66	0.73	4.40
Layout 4	0.80	5.05	0.61	6.17
Layout 6	0.70	3.46	0.70	4.60

The results presented in Table 13 show that the AE signals in each group show high similarity. Despite the large number of signals acquired in layouts 2 and 4, the similarity indices suggest that the signals may have originated from a very similar source. To validate this similarity, AE signals with different similarity indices from all datasets are analysed. Here, only signals from layout 4 are presented and discussed in the next section.

5.4.5. Similarity in the Time-Frequency Domain

The AE signals from layout 4 that share very close BC_t values are selected for analysis. The AE signals with different Bhattacharyya coefficients are analysed in their TF domain using the method adopted in Section 5.2.2 (Fig. 73).

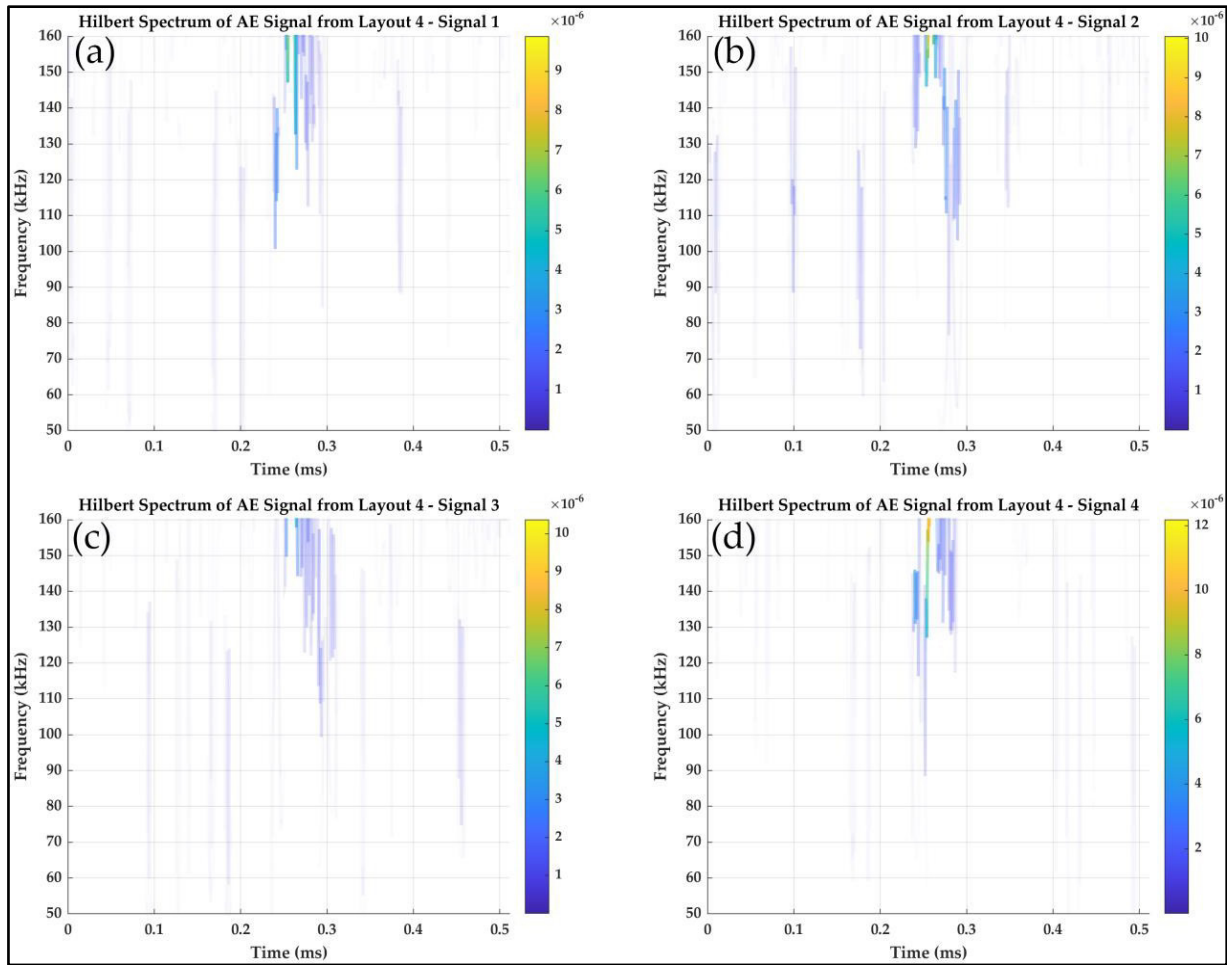


Figure 73. Acoustic emission signals from layout 4 with Bhattacharyya Coefficients (in Time Domain) (a) 0.8094; (b) 0.8089; (c) 0.7989; (d) 0.8062.

All the AE signals from layout 4 (Fig. 73) exhibit some similarities in both their frequency and time distributions. This indicates the high probability that all the signals are generated from a similar source. Using this example, this conclusion can be drawn for each layout, namely that there is a high probability that, in each layout, the signals are generated from a similar source.

5.5. Concluding remarks

The AE signals are acquired from six different areas in a lava tube cave at Vesuvius. No signals with amplitudes above the detection threshold were acquired in layouts 3 and 6 (central lining and roof), indicating high stability in these two structures. Very few AE signals are acquired from layouts 1 and 5 (stalactites and floor), which also indicates their fairly high stability, but the presence of signals produced by external sources. While a significantly larger number of AE signals are acquired from the two remaining layouts, layout 2 and 4 (left and

right and horizontal linings), which represent the side walls of the lava tube. This suggests the presence of loose masses or materials near the walls, which are responsible for these signals. However, the time domain analysis reveals that these signals have a very low amplitude and count. Typically, AE signals with such low amplitudes and counts are related to insignificant microstructural movements in the structure that do not affect its stability. These signals could potentially be generated by friction between the smaller structural elements present in the vicinity of these side walls. While the frequency analysis could not discriminate the differences between the AE signals generated from the different layouts, the time-frequency analysis using HHT explicitly showed the differences. The similarity indices verified that all the signals from each layout may have originated from a similar source. Our results show that all six tested layouts are generally stable with no potential loosened mass around them, which could contribute to the instability. These results also explain why the analysed section of the lava tube is still visitable since its formation and has not collapsed.

5.6. Future scope and development

This work has introduced a new methodology for investigating lava tube stability; this method was never applied before to the study of lava tubes, and our results highlight how successful this methodology was in assessing lava tube stability. The protocol used in this work should be extended to other volcanic areas, in particular those where tourism is especially developed in lava tube exploration.

Furthermore, in the future, it is going to be important to investigate lava tube stability systematically, comparing narrower and wider regions of the lava tube in order to better analyse their structural stability. Moreover, sensors could be installed over a prolonged period to determine whether there is any tempestuous period of AE signals present in the different layouts. Finally, if a significant variation in the AE signals generated by each layout (i.e., internal morphological variation) is measured, machine learning methods could be applied to identify and classify the signal differences.

CHAPTER 6

Discovering the 1858 lava flow field at Somma-Vesuvius (Italy) through geotinerary, virtual tour and 3D reconstruction of volcanic landforms

(Published in Geoheritage)

Abstract

Geotourism in volcanic areas attracts millions of visitors each year travelling to see and explore volcanic areas. One of the world's most famous and visited volcanic areas is Somma-Vesuvius. Here, tourism is mainly concentrated around the central crater while the rest of the volcano, part of Vesuvius National Park, is largely unvisited. The volcano's worldwide fame is primarily linked to its explosive activity, most notably the famous eruption that destroyed Pompeii and Herculaneum in 79 CE. Yet Somma-Vesuvius underwent long periods of effusive activity, such as the 1631-1944 eruptive cycle. On the volcano, there are examples of archetypal lava flows, including the 1858 flow field in which lava tubes formed. We propose a geoitinerary through the 1858 lava flow field that includes five sites, three lava tubes, an ephemeral vent and a tumulus, that allows people to evidence effusive activity and discover volcanic features formed during such activity at Somma-Vesuvius. A website was developed to host a virtual tour, which includes 360° images and an explicative panel for each site. The largest lava tube was 3D scanned to reconstruct its geometry, and to conserve a virtual copy of this fragile structure. Both the website and 3D scan aim to sensitize people either not used to hiking or with physical impediment about volcanic landforms and their scientific relevance. The pertinence of the geoitinerary has been assessed by a SWOT analysis. The geoitinerary may contribute to divert tourist traffic away from the volcanic cone, while raising awareness on volcanic processes.

Keywords: lava tube; 3D scan; geotourism; volcano tourism; geomorphology; geoconservation

6.1. Introduction

6.1.1 Geotourism in volcanic regions

Active volcanic regions are of high scientific interest due to their dynamic and geologically transformative nature. The eruption of molten lava, the seismic activity that often precedes and accompanies volcanic events, and the rapid formation of new landforms highlight the dynamic processes occurring within the Earth's lithosphere. These phenomena, including the creation of craters, lava flows, pyroclastic deposits, and the alteration of surrounding ecosystems, offer direct insight into the mechanisms of plate tectonics, magma movement, and landform evolution (Németh *et al.*, 2017; Dóniz-Páez *et al.*, 2020; Pérez-Umaña *et al.*, 2020). Volcanic eruptions can reshape landscapes on a geologically short timescale, emphasising the Earth's ever-changing nature. These phenomena attract people who may travel to appreciate volcanic activity, thus contributing to geotourism development. The concept of geotourism significantly increased its diffusion in scientific papers in the last decades (Hose, 1995, 1996; Miccadei *et al.*, 2011; Dowling and Newsome, 2018; Migoñ *et al.*, 2018; Ólafsdóttir and Tverijonaite, 2018; Sisto *et al.*, 2020; Santangelo and Valente, 2020; Valente *et al.*, 2020). The definition of geotourism was first proposed by Hose (1995, 1996) as 'the provision of interpretive and service facilities to enable tourists to acquire knowledge and understanding of the geology and geomorphology of a site beyond the level of a mere aesthetic appreciation'. This definition has been refined by Dowling and Newsome (2005), who introduced the concept of scale by including both small geological/palaeontological outcrops and large landforms as geotourism attractions. The latter include volcanic areas whose geotourism potential has been assessed in many areas of the world (Erfurt-Cooper, 2011, 2022; Erfurt-Cooper *et al.*, 2015; Alessio and De Lucia, 2017; Megerle, 2020). Volcanoes like Kilauea in Hawaii (US), Piton de la Fournaise on the island of La Reunion (France), Etna in Sicily or Stromboli in the Aeolian islands (Italy) and more recently the volcanoes of the Reykjanes peninsula (Iceland) are attracting many visitors each year due to their eruptive activities (Langridge and Michaud, 2023). For example, the eight national parks of Hawaii attracted over 4.7 million visitors in 2020 (<https://www.nps.gov/state/hi/index.htm>).

Visiting dormant volcanoes, as currently is Vesuvius, or extinct volcanic regions provides a valuable opportunity to understand many of the processes shaping our planet in greater depth. These areas have preserved geological features that reflect the Earth's volcanic history, offering tangible means to study past eruptions, patterns of volcanic activity, and long-term impacts on both the environment and human populations (Németh and Moufti, 2017;

Zangmo *et al.*, 2017; Planagumà and Martí, 2018; Quesada-Román *et al.*, 2020; Baadi *et al.*, 2023). Dormant volcanoes, while not currently active, remain vital for studying eruption recurrence, the long-term stability of volcanic structures, and potential future hazards.

From a naturalistic perspective, many volcanic regions support diverse ecosystems that have adapted to the unique conditions created by past volcanic activity (Sklenář *et al.*, 2010; Fazlutdinova *et al.*, 2021; Berrios *et al.*, 2022). Volcanic soils, rich in minerals, often foster lush vegetation, while animal populations may thrive in these nutrient-dense environments. In contrast, some volcanic landscapes appear barren, offering an opportunity to observe the gradual process of ecological succession as life returns to previously disturbed areas (Kienle *et al.*, 2022). These regions can also hold significant cultural and historical value, as human civilisations have long interacted with volcanic landscapes, adapting to, or even leveraging volcanic activity for their livelihoods (Beltrán-Yanes *et al.*, 2020; Zangmo Tefogoum *et al.*, 2022). In addition, barren volcanic land can offer unique opportunities to test technologies and operational scenarios, providing inputs for future lunar volcanic cave explorations during spatial missions (Sauro *et al.*, 2024).

The value of visiting dormant or extinct volcanoes extends beyond the aesthetic and ecological appeal, and plays an important role in hazard perception and risk management (Petrosino *et al.*, 2019). Exposure to volcanic landscapes enhances awareness of the potential risks posed by future eruptions or secondary hazards such as landslides or mudflows, even in currently dormant volcanoes. Studying the past behaviour of volcanoes, identifying precursor signals such as gas emissions or ground deformation, and understanding historical eruption patterns can inform early warning systems and emergency response strategies. This knowledge is critical for improving risk mitigation measures in areas vulnerable to volcanic hazards.

The interest in geotourism in volcanic areas is growing with many recent works in various volcanic centres around the world (Vereb *et al.*, 2020; Ertekin *et al.*, 2021; Sánchez-Cortez *et al.*, 2023; Vegas *et al.*, 2023; Yudiantoro *et al.*, 2024; Mülâyim *et al.*, 2025). Multiple volcanic areas were listed in the UNESCO World Heritage List, like Jeju Island in 2010 (Jeon *et al.*, 2024), Mount Fuji in 2013 (Chakraborty and Jones, 2018), Mount Etna in 2013 or the Chaîne des Puys monogenetic volcanic field in 2018 (Merle *et al.*, 2023). Several other volcanic areas are within geoparks included in the Global Geopark Network (GGN), like El Hierro island in Spain (Casillas Ruiz *et al.*, 2023), the Colca y volcanes de Andagua in Peru (Galaś *et al.*, 2018), and the Tungurahua volcano in Ecuador (Soria *et al.*, 2020) or will be considered for inclusion as Nevado del Ruiz in Colombia (Cifuentes-Correa *et al.*, 2023). Geotourism in

volcanic areas, particularly through guided geotours, offers an opportunity to deepen public understanding of the processes that shape the Earth. These tours can integrate educational content about volcano formation and eruption mechanisms, while also offering insights into the long-term environmental and societal impacts of volcanic activity (Armiero *et al.*, 2011; Alberico *et al.*, 2023). Through scientific curiosity or educational outreach, these tours enrich the visitor's understanding of the interconnectedness between geological forces, natural landscapes, and human societies. In this context, geotours serve not only as a mean of exploring the awe-inspiring beauty of volcanic landscapes but also as a platform to raise awareness about the dynamic and potentially hazardous nature of Earth's volcanic activity.

On the volcanoes of southern Italy, geotourism dates back to the 18th century when the noble young men from other European countries came to witness volcanic eruptions or to discover volcanic rocks and deposits, as stops of a journey known as the “Grand Tour”. While being well developed, recent works still aim at preserving and discovering Italian volcanoes (Pasquaré Mariotto *et al.*, 2022; Platania *et al.*, 2022; Alberico *et al.*, 2023; Di Vito *et al.*, 2023). In Sicily, guided excursions on Mount Etna are offered to visitors and the presence of expert guides is mandatory past a certain altitude fixed by local authorities. In the Aeolian Islands, guided excursions to observe the eruptions of Stromboli and free access to the crater of Vulcano to observe fumaroles are possible but limited by local authorities to certain elevations (below 400 m above sea level at Stromboli) and localities (Vulcano) (<https://cme.ingv.it/statodi-attivita-dei-vulcani-eoliani/archivio-documenti-protezione-civile/ordinanze-sindacali-comune-di-lipari>). In the Campanian region, a national park was created in 1997 to protect the nature of the Somma-Vesuvius stratovolcano. Here, the majority of tourists come to access the summit cone while the rest of the volcano is unvisited. With a maximum capacity of about one million visitors each year, Vesuvius is one of the most visited volcanoes in the world (Art. 2 of the Protocol for the tourist use of Path n. 5 of the Vesuvius National Park (“Crater of Vesuvius”; 620 000 visitors in 2024, <https://www.parconazionalelvesuvio.it/nel-2024-nuovo-record-di-visitatori-sul-gran-cono-del-vesuvio/>).

As previously stated, many landforms develop during volcanic eruptions, explosive or effusive. Among these are lava tubes (less commonly named lava tunnel or pyroduct by specialists in scientific fields other than volcanology) which form during effusive events (Greeley, 1971a, 1971c, 1987; Peterson *et al.*, 1994; Kauahikaua *et al.*, 1998; Calvari and Pinkerton, 1998, 1999). Lava tubes are a peculiar, but very infrequent geological feature at Vesuvius. The recent discovery and analysis of lava tubes on Vesuvius (which were formed

during the 1858 eruption; Lemaire *et al.*, 2024) interrogates us on the opportunity to evaluate the geotourism potential of these landforms and to relocate tourist traffic in areas of the National Park different from the central crater. As the 1858 lava flow surface features are unknown landforms at Vesuvius, most visitors of the Vesuvius National Park are not aware of their presence and ignore how these kinds of volcanic landforms develop. In this paper, we propose a geoitinerary through the 1858 lava tube field at Somma-Vesuvius by integrating field work with a virtual tour and a 3D reconstruction of the largest lava tube geometry.

6.1.2 Lava tubes as part of volcanic geoheritage

Lava tubes are preferred pathways of molten lava under a static solid crust (Greeley, 1987). The formation of a cooled roof over the flowing molten lava stops the convective cooling of the lava by the ambient air and allows it to remain hot for a longer time. Measured and theoretical cooling in lava tubes show a ≈ 1 °C/km cooling rate of the lava (Keszthelyi, 1995; Sakimoto and Zuber, 1998; Witter and Harris, 2007) compared to the 5 to 7 °C/km cooling rate in open lava channels (Crisp *et al.*, 1994; Cashman *et al.*, 1999; Soule *et al.*, 2004; Riker *et al.*, 2009; Robert *et al.*, 2014; Rhéty *et al.*, 2017). The lava flowing inside lava tubes is then able to travel longer distances from the vents than the lava flowing on the ground or within open channels. Master lava tubes are hundreds of metres to tens of kilometres long. Among the longest lava tubes on Earth are the Kazumura lava tube, located on the slopes of Kilauea volcano in Hawaii with a length of 65 kilometres (Allred and Allred, 1997), and the ≈ 100 kilometres long Undara lava tube in Australia, even visible from satellites (Atkinson *et al.*, 1975; Stephenson *et al.*, 1998).

Three main lava tube formation mechanisms have been described in the literature. A lava tube can form by roofing over a lava channel in several ways, depending on the rheology of the lava and on the effusion rate (Greeley, 1987; Peterson *et al.*, 1994). When the effusion rate is stable over time, a crust can extend from the channel levees on the lava surface, joining in the middle of the channel and thickening to form a solid roof. When the lava flow rate is fluctuating, the surface crustal slabs can break, detaching from the levees or from the lava surface and floating downflow. If they encounter a constriction or a turn, the crustal lava slabs can pile up, aggregate and merge to form a roof. When the effusion rate varies, overflows of lava over the levees make them grow towards the middle of the channel until they merge. Another way a lava tube can form is by inflation of a pāhoehoe sheet flow (Hon *et al.*, 1994) or of an ‘a‘ā lava flow front (Calvari and Pinkerton, 1998). And lastly, lava tubes can form by pāhoehoe lobe budding (Peterson *et al.*, 1994; Kempe, 2019). Pāhoehoe lava flows advance by

the repeating process of lava lobe creation from the front or sides of previous lobes. In some cases, the lava lobes align in the flow direction, allowing the formation of a primary lava pathway that can develop into a lava tube (Wentworth and Macdonald, 1953).

Lava tubes are commonly found in basaltic volcanic areas all over the world, either on volcanic islands such as Hawaii, Iceland, Canary Islands, the Azores, Galapagos, Sicily, Reunion Island, Jeju island in South Korea, Comores or Mauritius (Hernandez *et al.*, 1991; Hróarsson and Jónsson, 1991; Nunes and Braga, 1991; Calvari and Pinkerton, 1999; Kempe and Ketz-Kempe, 2016; Claudino-Sales, 2019; Tomasi *et al.*, 2022) or in intracontinental settings such as Australia, Saudi Arabia, Jordan or China (Atkinson *et al.*, 1975; Moufti *et al.*, 2013; Wang *et al.*, 2014). Lava tubes in volcanic areas such as Somma-Vesuvius are less common and were unknown up to recently (Lemaire *et al.*, 2024).

Geotourism focused on lava tubes is developed in the most famous volcanic places where lava tubes can be found (King, 2010; Galindo *et al.*, 2019; Dóniz-Páez and Pérez, 2023). On Hawaii island, it is possible to visit and explore lava tubes either self-guided or through guided tours provided by private companies or by the US National Park Service (<https://www.nps.gov/state/hi/index.htm>). Since a few years, Iceland has become an attractive destination for volcano enthusiasts due to the recurrent eruptions (Parks *et al.*, 2023). Private companies have developed tourism here in several lava tubes. The exploration of at least some portions of lava tubes is then possible for anyone, either self-guided in adapted parts of the lava tube or under the supervision of a guide to explore the entirety of the lava tube. In the Canary Islands, on the island of Lanzarote, a lava tube has been transformed by Cesar Manrique, a local artist, into an artistic, cultural and touristic centre (<https://cactlanzarote.com/en/centre/jameos-del-agua-guide/>). A restaurant, a natural pool, an artificial pool, and a natural auditorium can be found inside part of a large lava tube called La Corona at a skylight called ‘Jameos del agua’. Other lava tubes are proposed for visit in less renowned areas, but located in natural parks that are UNESCO World Heritage Sites, such as Jeju volcanic island (South Korea), La Réunion Island off the coast of Madagascar (France), or Mount Etna in Sicily (Italy) (<https://whc.unesco.org/en/list/>). However, on many of the less famous volcanic areas, the exploration of lava tubes is limited to experts or persons accompanied by local speleologists.

6.2. Geological setting

Vesuvius is one of the volcanoes of southern Italy (Fig. 74a), situated in the Campanian region, 15 kilometres to the east of Naples (Fig. 74b). Vesuvius is part of the Somma-Vesuvius complex, which is a stratovolcano that went through multiple collapses and regrowth phases (Cioni *et al.*, 1999). The first volcanic activity of the stratovolcano started after the Campania Ignimbrite eruption of the Phlegraean Fields (ca. 40 ka; Giaccio *et al.*, 2017) by the emplacement of a repetitive series of thin lava flows from a central vent and from cinder cones mostly developed as parasitic and eccentric vents (Macdonald *et al.*, 2016). This mainly effusive activity changed to mainly explosive 22 ka.

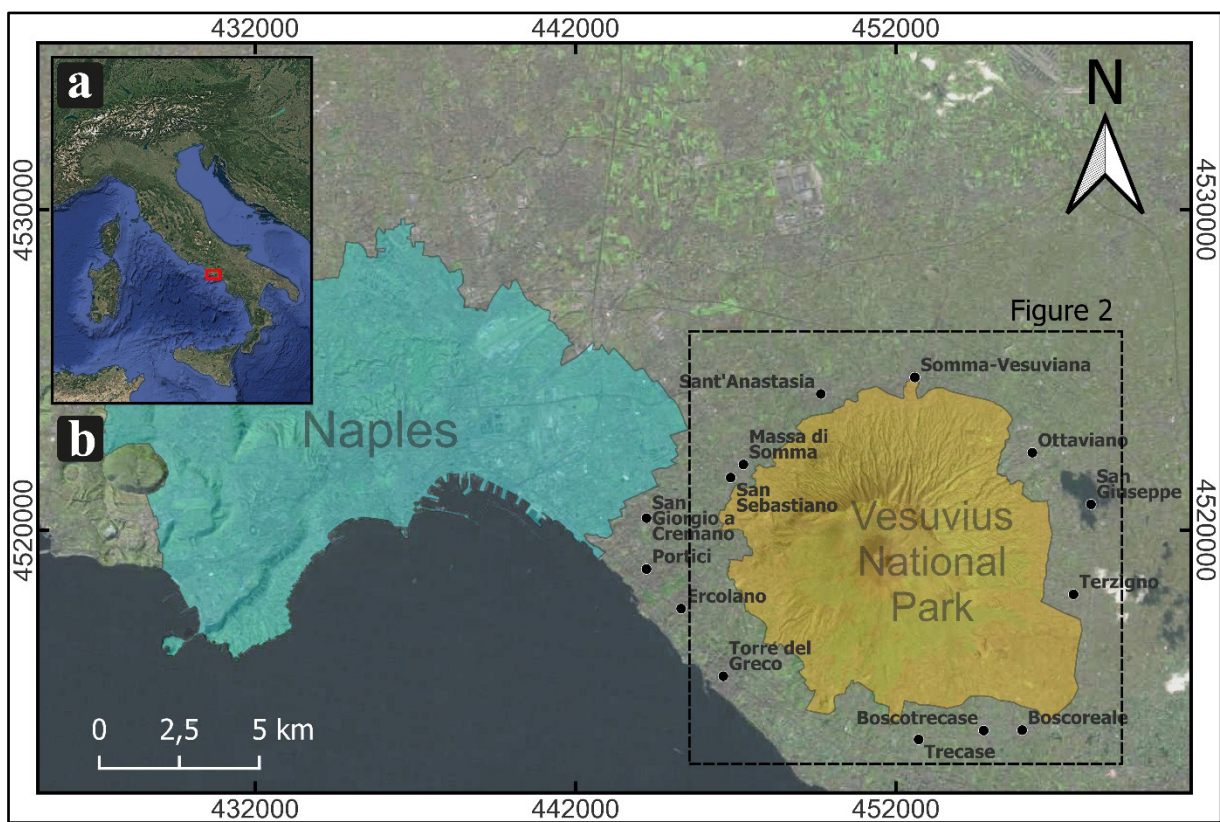


Figure 74. Localisation of the Vesuvius National Park 15 kilometres east of the city of Naples in Southern Italy, and main towns located in the surrounding area (black dots).

The volcano produced four major Plinian eruptions: the Pomici di Base (22 ka), the Mercato pumice (ca. 9 ka), the Avellino pumice (ca. 3.9 ka) and the Pompeii pumice (79 CE) (Santacroce *et al.*, 2008). Poorly known minor explosive events occurred in between these major eruptions. In particular, between the Avellino and Pompeii eruptions, at least six explosive events occurred (Rolandi *et al.*, 1998), named AP1-AP6 eruptions (Andronico and Cioni, 2002). Monte Somma is the older edifice that collapsed multiple times due to extensive drainage of the magma chamber, in concomitance with the explosive eruptions (Cioni *et al.*, 1999). It resulted in the formation of a sub-circular caldera, and Vesuvius is the new cone that

grew within the Monte Somma caldera after the 79 CE Pompeii eruption during periods of open-conduit activity (Arrighi *et al.*, 2001; Cella *et al.*, 2007; Cioni *et al.*, 2008). The last open-conduit period initiated with a sub-Plinian eruption in 1631. During this phase, lateral and summit effusive and semi-persistent mild explosive eruptions took place (Carta *et al.*, 1981; Arrighi *et al.*, 2001). More than one hundred lava flows were emplaced on the slopes of Vesuvius during this phase (Fig. 75), which ended with the 1944 eruption (Chester *et al.*, 2007; Cubellis *et al.*, 2016).

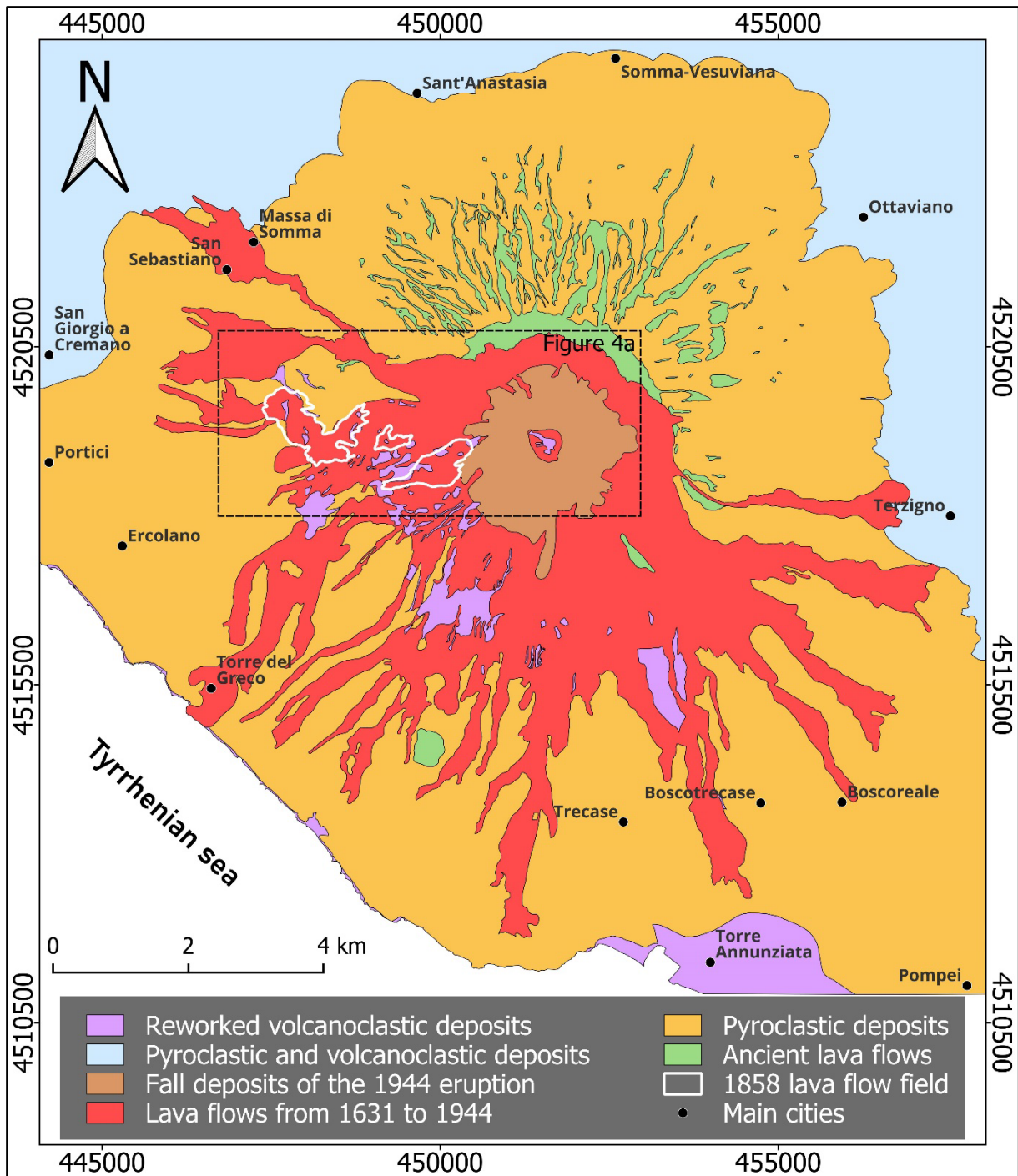


Figure 75. Simplified geological map of Somma-Vesuvius modified from Santacroce and Sbrana (2003).

The surroundings of Somma-Vesuvius have been highly inhabited for a long time. The eruptions of the 1631-1944 period were observed and described in historical documentation. In a few historical documents, descriptions of lava tubes can be found. One lava tube is drawn in the book titled ‘Vesuvius’ by Phillips (1869). Palmieri describes the 1858 eruption in the Annals of the Vesuvius Observatory in 1859: “Thus the curious phenomenon of flowing lavas was seen without being able to know the mouth from which they came. [...] When the outer part of the lava has hardened and the inner part is still liquid, it often happens that it flows downwards as if through a tunnel of impetrated matter, as if it were a vaulted aqueduct worked by art. As the lava disappears, it no longer fills the entire capacity of that tunnel, and often through some fissure made in it one can see the fire flowing below” (Palmieri, 1859). Another description of lava tubes is found in the Annals of the Vesuvius Observatory of 1870, where Palmieri wrote about the 1867 eruption saying: “This appearance arose from the fact that the lavas on the Vesuvian cone had formed a covered channel or tunnel of slag within which invisible slag descended from the summit to the base of the aforesaid cone, and when, because of the periodicity mentioned above, the tunnel could not accommodate a new flood, its walls broke in one or more places and streams of fire could be seen appearing on the cone” (Palmieri, 1870). Lava tubes were observed during eruptions but the only description of an inactive lava tube at Vesuvius was done by Malladra (1917), who described a lava tube preserved within the 1858 lava flow field. The 1858 eruption is well described. Nowadays, numerous surface features can be observed and are of interest to understand lava flow field emplacement (see Lemaire *et al.*, 2024).

6.3. Methods

In this work, we used for the first time at Vesuvius a multidisciplinary approach combining fieldwork, 3D scanning, 360° photography, explicative panels, and SWOT analysis, aimed at creating and providing to visitors an easy and comprehensive itinerary for them to discover volcanic activity. As a result, we acquired data to create a virtual tour of the study area. Virtual tours allow people to visit and get knowledge of a site from their laptop or mobile devices anywhere in the world. The role of virtual tours in geotourism has been recently diffused in the scientific literature (Martínez-Graña *et al.*, 2013, 2023; Filocamo *et al.*, 2020; Fassoulas *et al.*, 2022; Papadopoulou *et al.*, 2024; Williams *et al.*, 2025), even in the volcanic landscapes of Iceland (Pasquaré Mariotto *et al.*, 2022). Each method used in this study is described below.

Our study began with field surveys that were extensively carried out to identify surface features on the 1858 lava flow field of Vesuvius. As a result, we discovered and selected five landforms to be included in the geotinerary, which are three lava tubes, a tumulus, and an ephemeral vent that formed during the 1858 eruption.

We developed panels that describe the formation and growth of the selected volcanic landforms to help visitors understand the mechanisms that drive the creation of such features during lava flow field emplacement. These panels should be placed next to the five selected sites, as their use as a communication strategy is common in geotourism research (Melelli, 2019).

For the accessibility of the selected features, we captured 360° photos using an Insta360 X4 camera. The photos were edited and exported within the Insta360 Studio software. The 360° photos are hosted on Memento360 and implemented on a website alongside the explanatory panels, made on Google sites, where anyone can explore the virtual geotinerary, allowing an immersion into the sites (<https://sites.google.com/view/vesuvius-1858lavaflow/>).

Furthermore, a 3D reconstruction of the interior and surface of the largest lava tube found in the 1858 lava flow field using high-end technologies was carried out by Lemaire *et al.* (2024) and used in this work to preserve the morphology of the tube as a digital twin through time. The interior of the lava tube was scanned using a Riegl VZ-400 time-of-flight terrestrial laser scanner mounted on a tripod. Nineteen scans in different positions within the tube were acquired to avoid shadows due to the complex geometry of the inside. Twelve out of the nineteen scans were used to create a mesh in the RiSCAN Pro software. The overlying surface of the lava tube was scanned using a DJI Matrice 300 RTK UAV equipped with a stabilised Zenmuse P1 optical camera. Aerial images were transformed into a surface Digital Elevation Model (DEM) and a mesh in Agisoft Metashape. To correctly assemble both scans, we first positioned markers in different areas of the lava tube during the survey and then we carried out photogrammetry work using a GoPro 10 camera in two key areas to produce two meshes that show both the interior and exterior of the lava tube. The interior and surface scans were assembled in the software CloudCompare.

Finally, we carried out the Strength, Weaknesses, Opportunity and Threats (SWOT) analysis to test the geotourism potential of the 1858 lava flow field and to propose initiatives for its efficient and effective use (Carrión-Mero *et al.*, 2020; Kubalíková and Kirchner, 2016; Valente *et al.*, 2021).

6.4. Results

6.4.1 The lava flow field geotinerary

The 1858 lava flow field is located on the western flank of Vesuvius at an elevation ranging from 200 to 800 metres above sea level (Fig. 76). It was emplaced during the 1858-1861 eruption that began on 27 May 1858 and ended on 12 April 1861, lasting almost three years. The maximum flow length is around 5.2 kilometres, and the mean flow width is around 0.6 kilometres. The emplaced lava flow field covered around 3.9 million square metres but the exposed area, uncovered by more recent lava flows, is just 1.3 million square metres, one third of the original area. The seven vents that opened during the eruption produced an estimated volume of 178 million cubic metres of lava (Lemaire *et al.*, 2024), and are now covered by more recent flows.

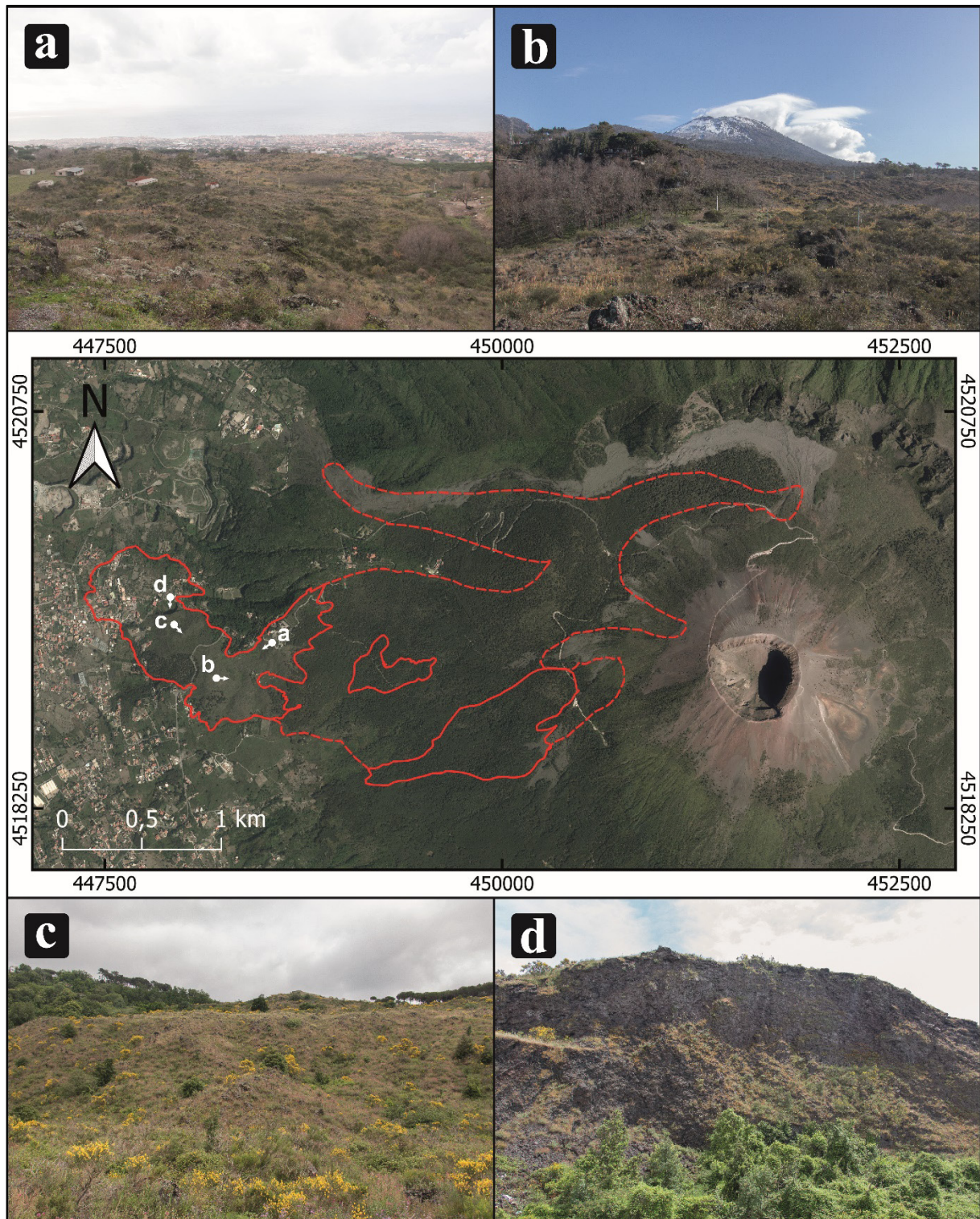


Figure 76. Orthophoto of Vesuvius with the margins of the 1858 lava flow field highlighted in red (continuous line: exposed, dashed line: covered by more recent lava flows) and photos of the flow field. a) Photo of the lava flow field looking toward the southwest, the city of Torre del Greco is visible along the seaside. b) Photo of the lava flow field looking toward the Vesuvius crater taken close to site 5. c) Photo of the flow field taken close to the flow front. d) Photo of a quarry wall located close to the flow front showing the piling up of small flows for a total thickness of about 100 m.

Within the 1858 lava flow field, five sites are described and proposed for on-site visit along a geotinerary (Fig. 77) and online through a virtual geotinerary composed of 360° photos. For each site, a descriptive panel was designed to give scientific information about the site and explain the mechanism behind its formation. The geotinerary is 500 metres long and is mainly addressed to people used to hiking, as walking on a lava flow field may be difficult. People not used to hike may appreciate the geotinerary through the virtual tour (Section 5.2.4). The best periods to enjoy the geotinerary are autumn and winter, whereas vegetation blooming in spring and summer may reduce visibility. The description of each stop and the related explanatory panel are reported below. The presence of panels next to each site is essential to the comprehension and dissemination of the characteristics and scientific information we aim to disseminate, and it is our intention to collaborate with the local park authorities to realise them.

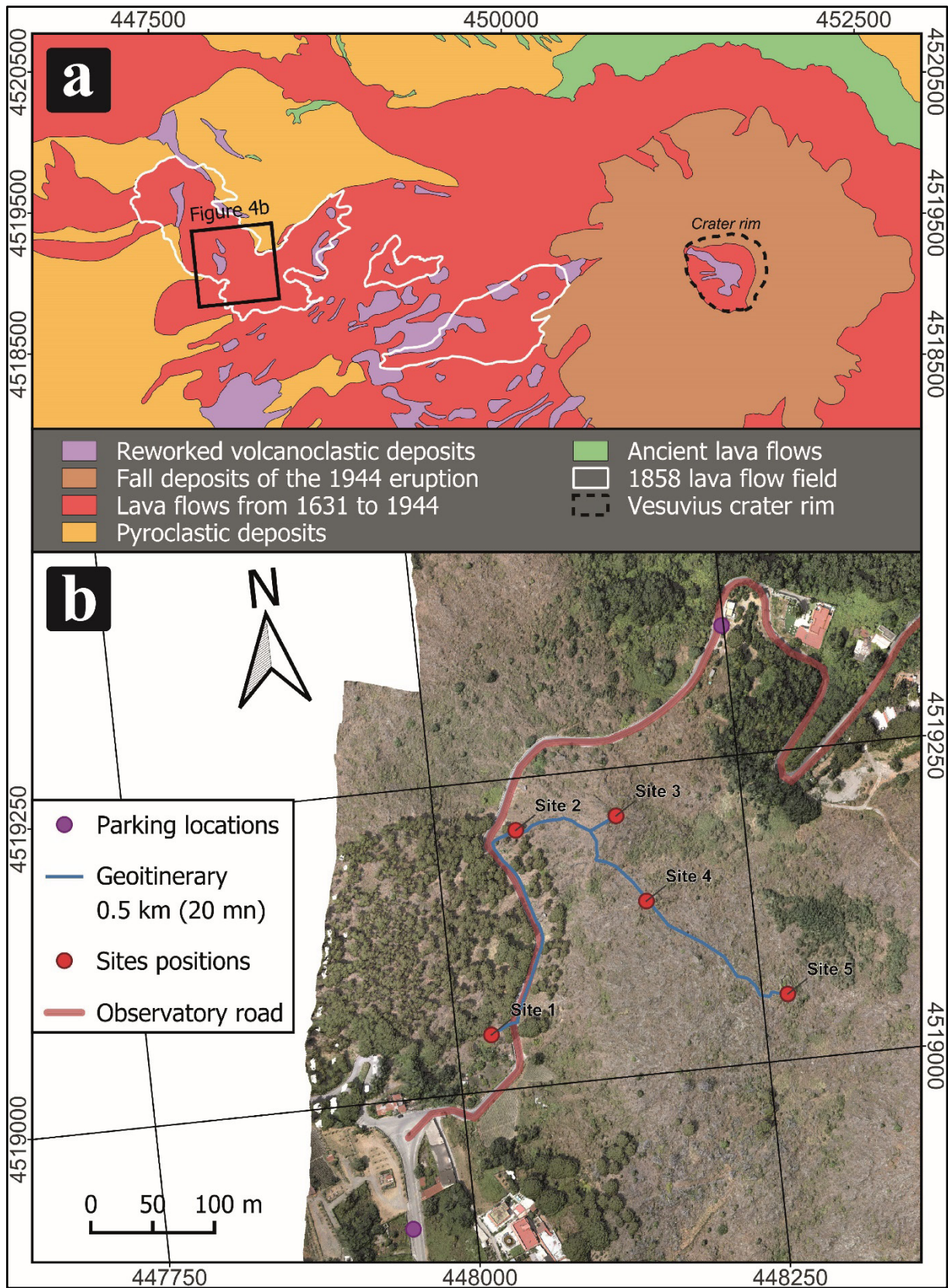


Figure 77. a) Detail of the geological map of Figure 2 with location of the area (black box) where the geoitinerary is located; b) Proposed geoitinerary (blue line) featuring five sites (red dots) along the 1858 lava flow field of Vesuvius.

Site 1: Early-stage lava tube

Site 1 corresponds to a lava tube (Fig. 78) located a few metres next to the Observatory Road. It is 5 metres long, 1 metre wide and about 2 metres high. The outer rim (Fig. 78) is surrounded by overflows with pāhoehoe textures. Inside, in the deepest part of the tube, it is visible the ephemeral vent that originated the tube and probably formed at the margins of a former and larger lava flow located upslope. On the lateral walls there are several lava flows overlapped, showing that this tube was probably the vertical merging of the voids formed within several lava flow units. In between the stacked flows, there are pāhoehoe squeeze-outs fingering into the ‘a‘ā rubbles (Fig. 78), suggesting that the tube was active for some time (probably days), allowing the pāhoehoe lobes to fill the voids left over in between the ‘a‘ā rubbles comprising the outer crust of the lava flow that originated the tube. These lava fingers were probably attempts of the fluid lava inside the tube to open ephemeral vents at the sides of the tube when the pressure inside the tube was growing due to the accumulation of lava.

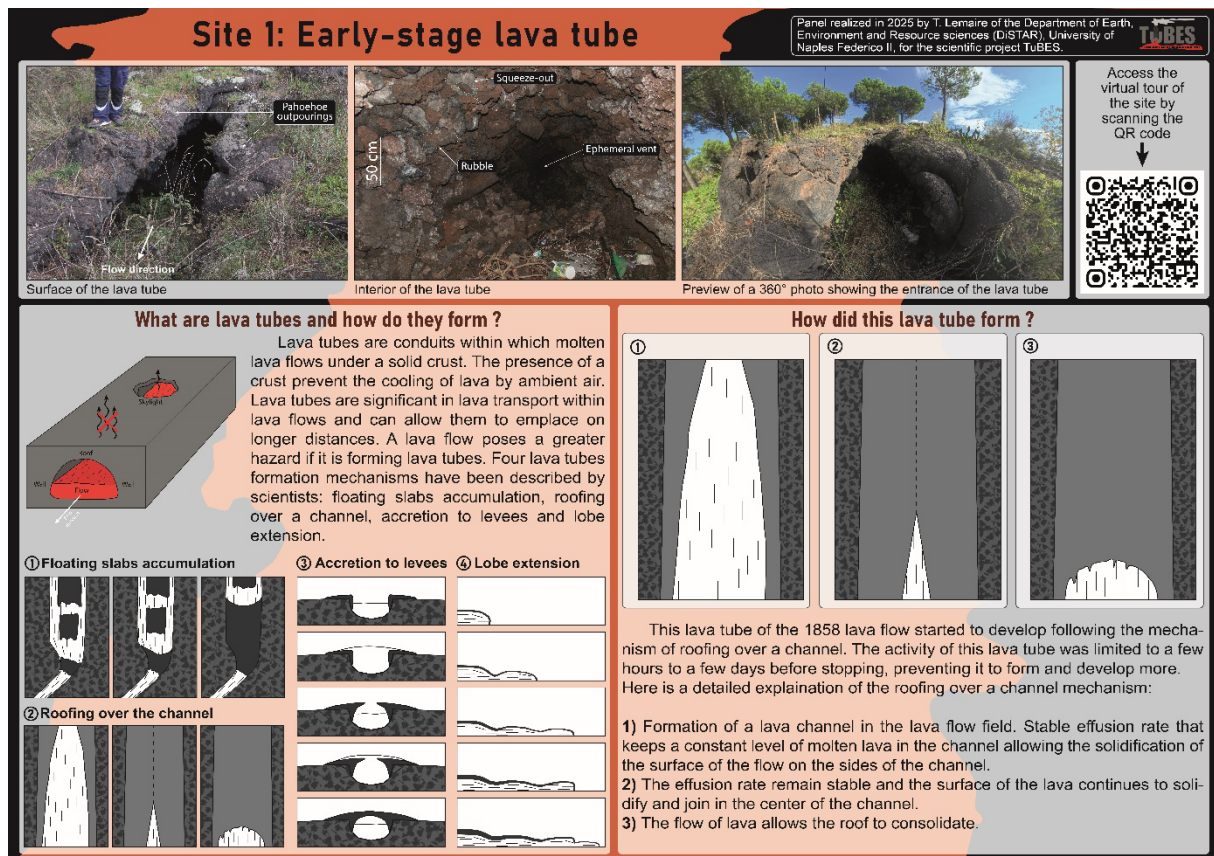


Figure 78. Explicative panel presenting site 1, detailing the general processes of lava tube formation, and showing the model for the formation of the observed lava tube.

Site 2: Ephemeral vent

This feature corresponds to the breakage of the crust at a lava flow front from which molten lava outpoured forming a new flow, usually short in duration and/or length. It is commonly associated with lava tubes, as it causes their drainage and thus the formation of a cavity into the drained flow upslope. Here, the lava outpoured from the former flow margins and formed a 2 metres wide lava channel with pronounced lateral levees (Fig. 79). The lateral levees are 20 to 30 centimetres higher than the flow surface, suggesting a partial drainage after its initial emplacement. The location of this ephemeral vent suggests that the outpoured lava flow originated from Site 3 (Fig. 80).

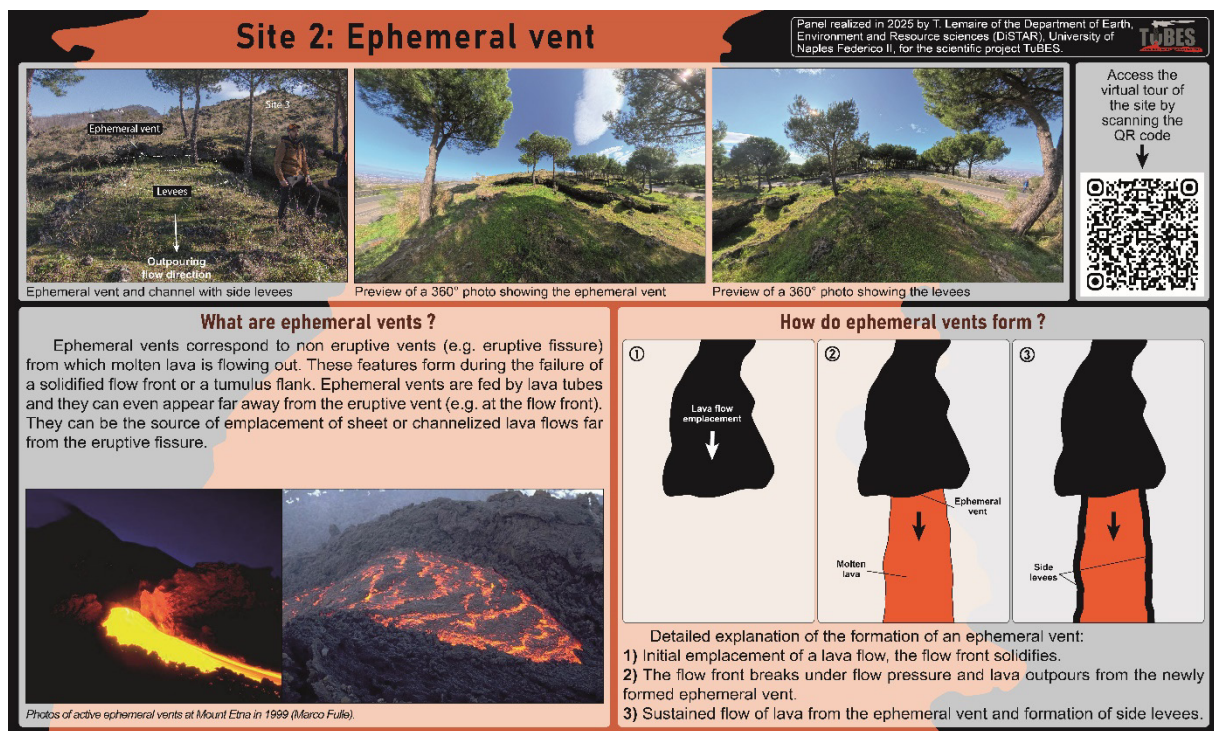


Figure 79. Explicative panel presenting site 2, detailing the process of formation of ephemeral vents, such as the one observed here.

Site 3: Tumulus

Tumuli are positive topographic features that form by a mechanism of crustal uplift on lava flow fields (Duncan *et al.*, 2004). A tumulus forms normally on a shallow sloping ground and at a break in slope, where the magmastatic pressure within the lava flow led to localised inflation processes that took place during lava flow emplacement (Duncan *et al.*, 2004; Anderson *et al.*, 2012). It is a sign of lava accumulation below a stable crust and of flow inflation. The tumulus surface feature here described is 10 to 20 metres high and has a 160 metres diameter (Fig. 80). This tumulus shows numerous breakouts or ephemeral vents from its sides and top forming small channels and even small tubes on its flanks. A tumulus indicates

that magmastatic pressure within the lava flow led to localised inflation processes that took place during lava flow emplacement (Duncan *et al.*, 2004; Anderson *et al.*, 2012), often located in places along the flow path where the ground slope suddenly decreases or increases. Tumuli are often indicative of the presence of lava tubes underneath because they emplace along the lava tube paths where the ground surface presents a break in slope, allowing flow drainage and thus the formation of a cavity (Mattox *et al.*, 1993; Calvari and Pinkerton, 1999).

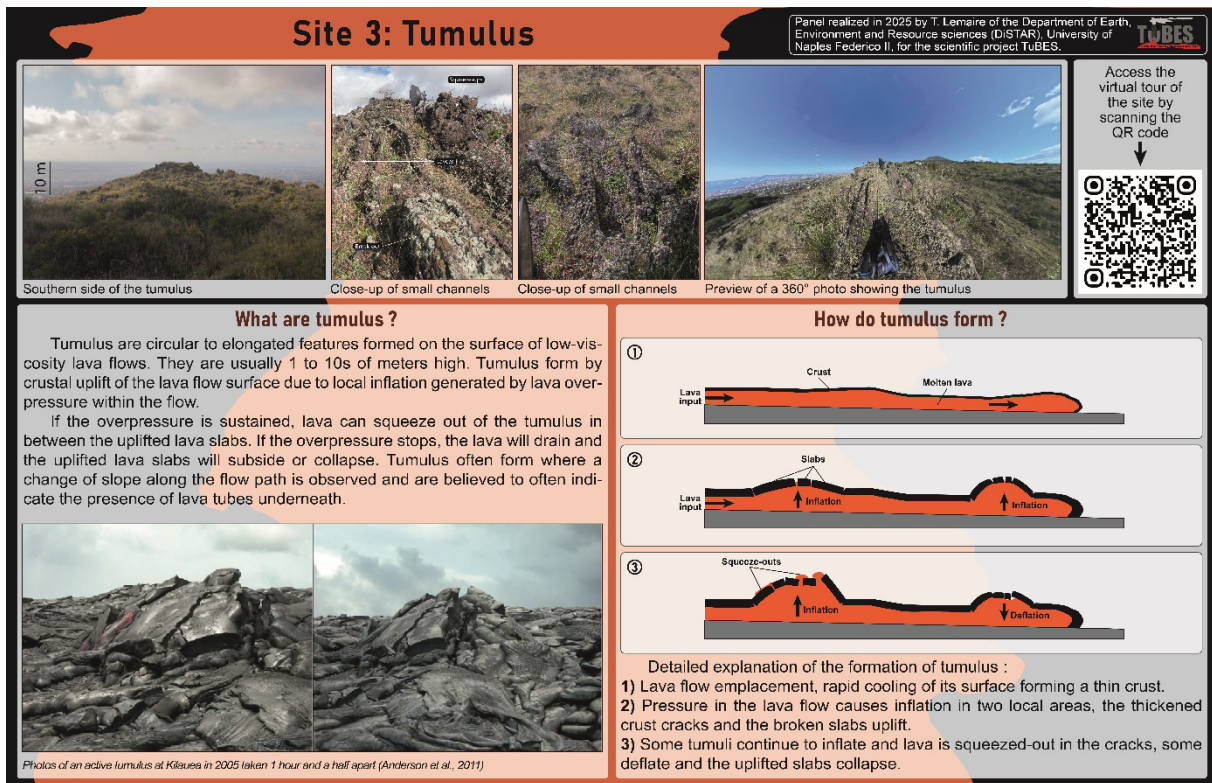


Figure 80. Explicative panel presenting site 3, a tumulus, and detailing the mechanism of formation of such structure.

Site 4: Breakout with a small lava tube

Site 4 is a small lava tube located next to the base of a small tumulus and originated from a side breakout. The tube is 20 centimetres wide conduit and with a thin broken crust about 1-2 centimetres thick allowing the view of the inside (Fig. 81). The roof is formed by two overlapped ropy pāhoehoe sheets of lava that broke in one place forming a triangular skylight (Fig. 81a). Inside it, we can observe smooth rounded and reddish parallel stalactites on the ceiling elongated along the flow direction, and a concave roof indicating flow inflation and later flow drainage. Some transversal fractures contain roots growing into the cavity (Fig. 81b), a common feature in cases of a thin tube crust increasing tube instability and erosion. The floor is mostly horizontal and with pāhoehoe textures.

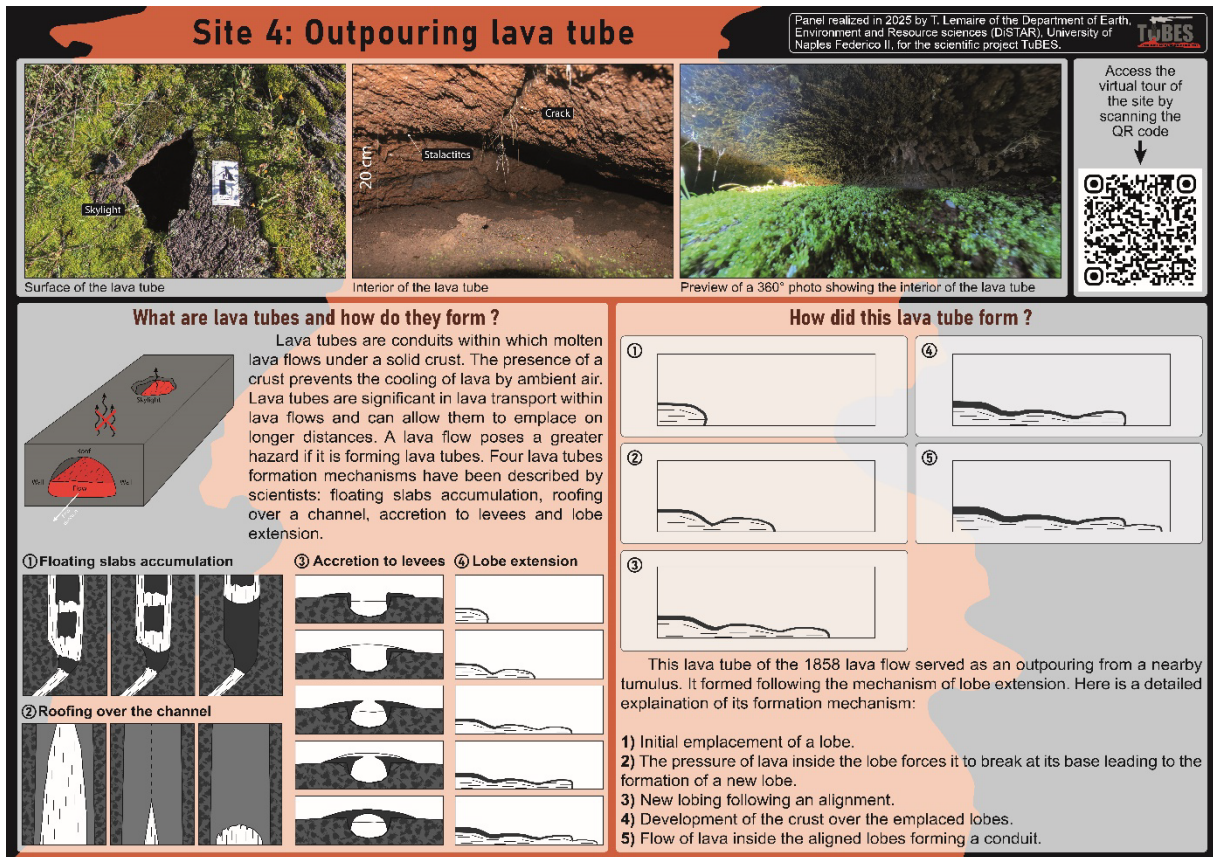


Figure 81. Explicative panel presenting site 4, detailing the general processes of lava tube formation, and showing the model for the formation of the lava tube observed at Site 4.

Site 5: The largest lava tube

Located in a very flat area, this lava tube is the largest one found in the 1858 lava flow field (Fig. 82). The outer surface is characterised by an inflated pāhoehoe flow showing longitudinal cracks and one skylight allowing access into the tube. The tube outer surface rises ≈ 1.5 meters above ground, and is surrounded by vegetation, whose blooming in spring and summer may prevent the recognition of the site. The tube entrance is through the skylight that allows easy access through several blocks collapsed from the ceiling. The roof is 1 to 3 metres thick and the lava tube is about 30 metres long, 1.5 to 17 metres wide and the mean height inside is 1.4 metres. Inside the tube there are lava stalactites, several inner linings, local collapses of the roof, and the floor is made of pāhoehoe surface texture. Flow direction in the lava tube was from north to south. The lava entered from a small twenty-five centimetres wide vent and exited the lava tube to the southwest by a small passage.

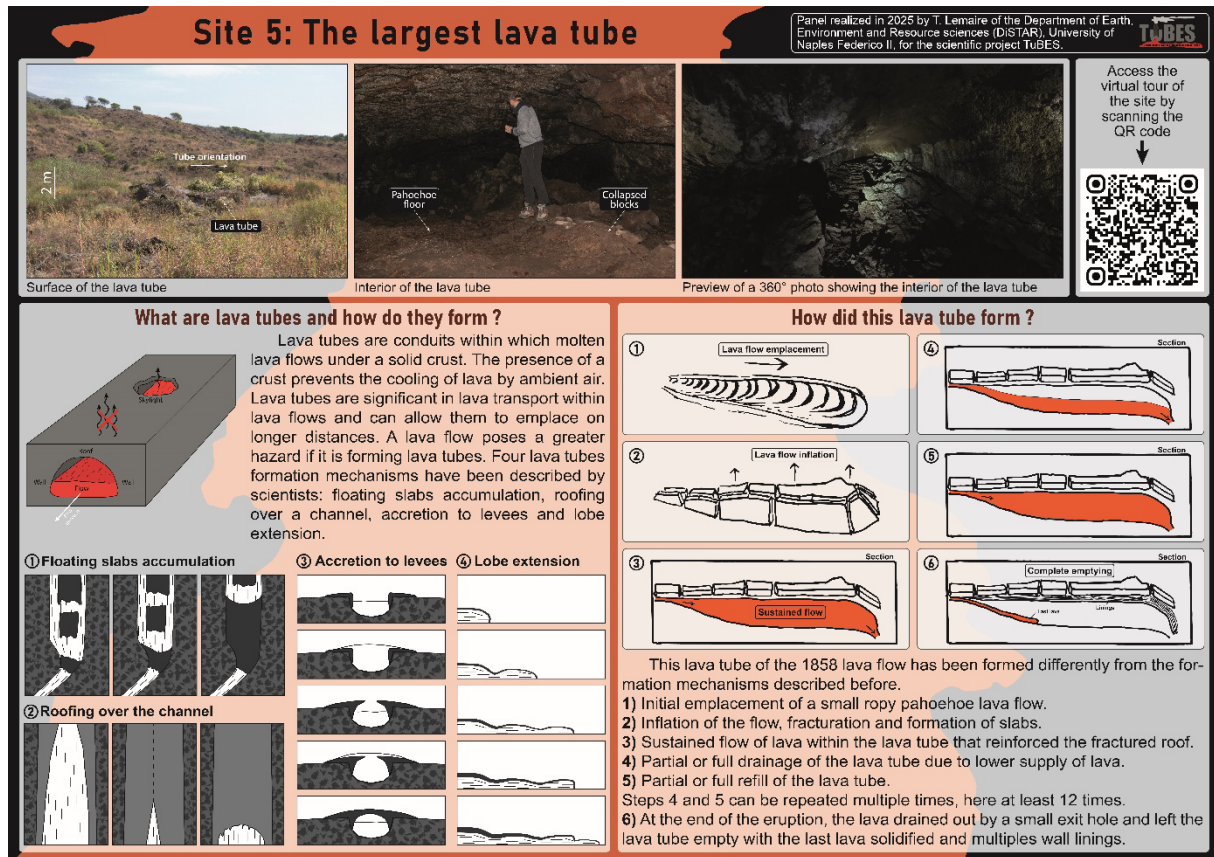


Figure 82. Explicative panel presenting site 5, detailing the general processes of lava tube formation, and showing the model for the formation of the lava tube found at Site 5.

6.4.2 Virtual tour

The 360° images collected in the field, as well as the panels and explanations of each site, are published on a dedicated website accessible online via the following link: <https://sites.google.com/view/vesuvius-1858lavaflow/>. The ‘Geoitinerary’ page introduces this study and the proposed geoitinerary through a descriptive text and an online interactive map (Fig. 83a). The ‘Terminology’ page defines the volcanological terms used in the descriptions of the sites (Fig. 83b). The ‘Sites’ page is redirecting to each of the five site’s sub-pages that are reachable by clicking on the Explore button of each site photograph (Fig. 83c). The five sub-pages show a description, the associated 360° photos and the explanatory panel of each site. These 360° photos are interactive and allow visitors to navigate and virtually explore each site (Fig. 83d).



Figure 83. Screenshots of the website. a) Geoitinerary page with an introduction to the geoitinerary and an interactive map b) Terminology page on which volcanological terms are defined c) Sites page that redirects to the five subpages of the sites d) Subpage of Site 1 with a description of the site, the interactive 360° photos and an explanatory panel.

6.4.3 3D digitisation of the largest lava tube

Lemaire *et al.* (2024) have investigated the largest lava tube (Site 5) found in the 1858 lava flow field. The field surveys to 3D scan this lava tube allowed the construction of a complete 3D model in which the internal scan is aligned with the georeferenced overlying surface. Using the 3D model, Lemaire *et al.* (2024) described in detail the morphology of the lava tube (Fig. 84). The lava tube is 30.15 metres long and 1.20 to 17.41 metres wide. The mean height inside is 1.40 metres while the maximum height is 2.47 metres in the southern part of the lava tube. The 3D digital copy is incredibly helpful to characterise precisely its morphology but is also an excellent capture of its state at a moment in time, allowing to explore it even after possible collapses, as well as to preserve the record of this unique volcanic feature at Vesuvius.

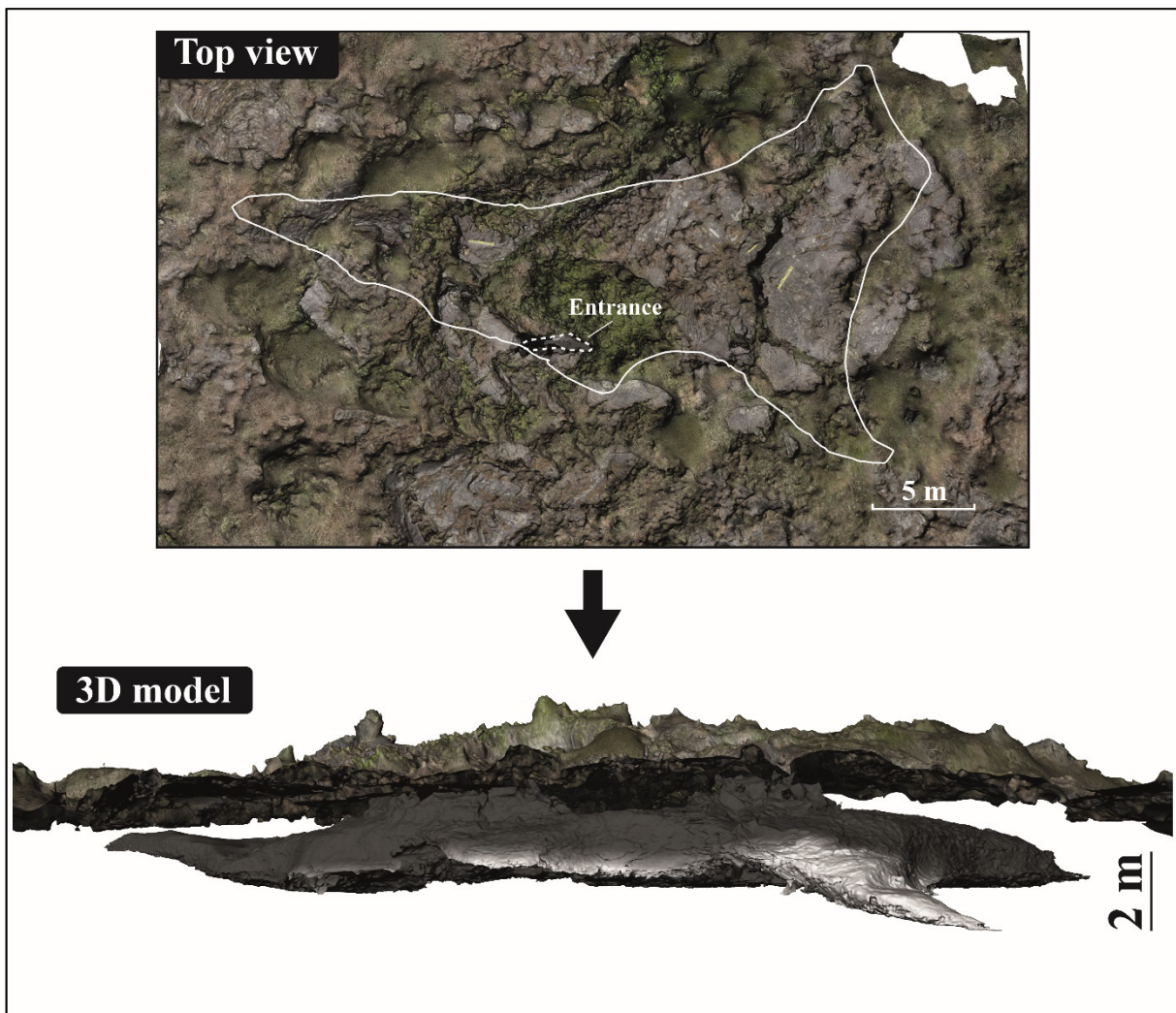


Figure 84. 3D model of the largest lava tube found in the 1858 lava flow field (Site 5) at Vesuvius. It is composed of the internal scan of the lava tube and the georeferenced scan of its overlying surface. At the top, a detailed view of the overlying lava tube surface with the contours of the skylight (dashed white line) and the underlying tube (white line).

6.5. Discussion

6.5.1 Mechanisms of lava tube formation in the 1858 lava flow field

Lava tubes are formed by three main mechanisms: roofing over a lava channel, pāhoehoe lobe extension or lava flow inflation (Greeley, 1971a, 1987; Peterson *et al.*, 1994; Kauahikaua *et al.*, 1998; Calvari and Pinkerton, 1998, 1999). In the 1858 lava flow field, the lava tubes (sites 1, 4 and 5) have different sizes and morphologies indicating that they were subject to various mechanisms of formation. Site 1 formed by roofing over an ‘a‘ā lava channel that produced several lateral pāhoehoe overflows increasing the thickness of the lateral levees. The ‘a‘ā nature of the initial flow is testified by the stacked rubbles alternated to massive flows, each about 0.5 metres thick, observed on the lateral inner walls of the tube (Fig. 78). This is an early-stage lava tube and given its small extent was probably active for a few hours to a few days, not being able to develop more due to its location at the margins of the lava flow field, thus receiving only a limited amount of lava supply for a short time. Site 4 formed by inflation of a very small pāhoehoe lava flow and served as a draining path for the lava from a nearby tumulus (Site 3). Site 5 also formed by inflation of a pāhoehoe lava flow at a larger scale. This mechanism of formation was described as a new interpretative model by Lemaire *et al.* (2024) and detailed here in the related explanatory panel (Fig. 82). The process involves at first the emplacement of a small ropy pahoehoe lava flow. Its margins cooled down and solidified while the supply of lava continued to feed the flow leading to its inflation. Flow inflation in turn causes fracturing of the surface crust and the formation of large slabs of solidified crust. A sustained and steady flow of lava allowed the formation, growth and consolidation of the crust. The lava tube partially or completely drained due to the opening of ephemeral vents from its margins, then by a new supply of lava was partially or completely refilled. This change of lava level inside the tube happened at least twelve times throughout the activity of the tube, as suggested by the presence of twelve wall linings. Each lining, in fact, reveals that the tube was full of lava and drained to allow this thin film of lava to attach and coat the inner tube walls. When the lava supply stopped, the lava tube drained completely, leaving the empty cavity we observe nowadays.

It is worth noting that the young age of the lava tubes here described allows a vision not just of the flow surface and of the inner tube shape, but also of the inner features such as stalactites and inner coatings described above, normally not preserved within much older lava tubes such as those of Australia (Undara tube, Atkinson *et al.*, 1975), Lanzarote (La Corona

tube, Tomasi *et al.*, 2022), or of the Kingdom of Saudi Arabia (Umm Jirsan, Németh and Moufti, 2017) to give just a few examples.

6.5.2 The 1858 lava tubes as possible geotourism attractions

The geoitinerary is a major opportunity for the scientific community to reach out to the public and disseminate scientific knowledge about volcanoes and their associated hazards, but also about lava flows and their surface features. Filling the gap in knowledge that the population can have about lava flow field occurrence is essential, especially for the local inhabitants living next or even on the flanks of an active stratovolcano. Knowing how a volcano works will help people understand the hazards associated that they may face in the future. Since the last eruptive activity at Vesuvius was mostly effusive, the hazard due to effusive activity exists, but is, at the moment, largely unknown to local inhabitants. This is particularly important because the risk posed by lava flows at Somma-Vesuvius is increased by the proximity of inhabited areas to the eruptive centre, with the first buildings located just 3 kilometres from the Vesuvius crater (Chester *et al.*, 2007). This is a characteristic that Vesuvius shares with other Italian volcanoes like Mount Etna, and other volcanic areas at Hawaii, Iceland and Chile. Cities all around the coast of Hawaii Island are located a few kilometres from possible vents opening on the flanks of Mauna Loa, Kilauea, Mauna Kea, Hualalai and Kohala volcanoes, sometimes being less than a kilometre (Kauahikaua and Tilling, 2014). The village of Vik in Iceland is exposed to volcanic risk due to the activity of the Katla volcano (Jóhannesdóttir and Gísladóttir, 2010), and the 2021 Geldingardalur eruption directly impacted the Reykjanes Peninsula, the town of Grindavik and the South Coast Road (Sigtryggisdóttir *et al.*, 2025). In the southern Andes of Chile, human reoccupation of rural settlements affected by volcanic events has been testified in the Carran-Los Venados and Puyehue-Cordón Caulle volcanic systems (Vergara-Pinto and Marín, 2023). Noteworthy, the formation of lava tubes could increase the distance that lava flows can travel, further increasing the areas potentially at risk (Calvari and Pinkerton, 1998; Kauahikaua *et al.*, 1998). The use of questionnaires and interviews with citizens living near or within a volcanic area is a common practice that helps understand their risk perception related to volcanic events, and the local visitors of the geoitinerary could be profitably involved in such a programme. This method has been tested in Indonesia (Lavigne *et al.*, 2008), Iceland (Jóhannesdóttir and Gísladóttir, 2010) and Vesuvius (Avvisati *et al.*, 2019).

The geoitinerary is visitable on-site by people used to hike because of an uneven path, but the presence of a wooden path in the future would help expand the possibility of visiting the sites. In addition, the proposed virtual tour allows everyone to access the geoitinerary. The

free and easy access to an online virtual tour regrouping 360° photos of each site, allows the dissemination of knowledge, the observation and explanation of these volcanic features to a broader audience and to persons unable to reach the sites (e.g. persons with disabilities, elders). It also gives the possibility to see, explore and understand the sites without the possible risks of going to the field (e.g. collapses) or the difficulty of finding the sites due to vegetation overgrowth. The complete 3D digitisation of Site 5, the largest lava tube found in the 1858 lava flow field, is crucial because it provides a record of this fragile structure's heritage. This record stands as a numerical copy of the site that can be shown, explored, and studied in any place at any time. It is one of the very few 3D scans of lava tubes produced in the world (e.g. Tomasi *et al.*, 2022).

In terms of tourist management, the presence of the 1858 lava flow field and its associated features in a lower part of the stratovolcano could help redistribute the flow of tourists within the National Park, which is presently focused on the summit crater. In this sense, it is close to the entrance of the "Lava River" path, which leads to the 1944 lava flow and close to the historical site of the Vesuvius Observatory. The three visits can easily be combined in a single itinerary to be fitted along the whole of a spring or summer day.

As introduced at the beginning of this study, in other volcanic areas, local communities are taking advantage of lava tubes as touristic attractions. Even if the morphological scale is different between lava tubes at Vesuvius and Hawaii, the comparison in terms of geotouristic attractions and management is still interesting. In Hawaii, the National Park Service made itineraries to and within lava tubes. The Thurston lava tube present at Kilauea is adapted and lit up during the day for visitors to freely access and explore (<https://www.nps.gov/places/nahuku.htm>). Other lava tubes are only accessible by booking tours like the Kazumura cave (http://kazumuracave.com/#Tour_Information), the Pua Po'o lava tube (<https://www.fhvn.org/institute/wild-caves-exploration/>) or the Kula Kai caverns (<https://www.kulakaicaverns.com/>). Even if Hawaii is internationally renowned for its volcanoes, lava tubes are important geotouristic sites for the natural parks of Hawaii. At Vesuvius, lava flows and lava structures could attract new visitors on the volcano but also to the city of Naples and its surroundings (e.g. Ercolano, Torre del Greco).

6.5.3 Quality of the sites and visitor targets

A site that could be part of the geoheritage and worthy of promotion and development must have some characteristics that can be summarised as scientific, educational, additional and aesthetic features (Różycka and Migoń, 2018). The scientific value of all the sites is unquestionable, due to the uniqueness of the Vesuvius lava tubes and the possibility of bringing together, on a reduced scale, all the structures that characterise this peculiar feature of lava fields. The sites are located in a lava flow field that can be easily observed, which is not a frequent characteristic of Vesuvius lava flow fields. The educational value derives directly from the scientific one, since the sites represent a sort of open-air classroom where the characteristics of low-viscosity lava flows can be demonstrated in a volcanic area where effusive products are, in general, poorly represented. The added value lies in the link between the lava tube and the history of observation and monitoring of this worldwide known volcano. The chronicles of the time (see section 6.2) tell of the pioneering descriptions of the phenomenon, which had never been observed on the volcano (Palmieri, 1870), and of the first speleological investigations by Malladra (1917), who described the largest lava tube found in the 1858 lava flow field for the first time. These chronicles remind the visitor of the progress made by scientific research in the last 150 years, but also of the importance of direct observation of volcanic phenomena, guaranteed here since the second half of the 19th century by the presence of the first volcanological observatory in the world. From a naturalistic point of view, the vital role of pioneer plant species such as the lichen (*Stereocaulon Vesuvianum*) in the creation of ecosystems can be demonstrated. This lichen is currently in the process of colonising barren lava flows, breaking down rocks and contributing to the formation of soil. The aesthetic value of the whole area is exceptionally high, as it offers a panoramic view of the Gulf of Naples, with the Sorrento peninsula and the island of Capri to the south, and Posillipo and the Camaldoli hills to the north-east (see the panoramic view from site 3 on the website: <https://sites.google.com/view/vesuvius-1858lavaflow/sites/site-3-tumulus>). This breathtaking landscape contains all the natural beauty that attracted the ancient Greek and Roman civilisations to settle in this active volcanic area. Unfortunately, the site's accessibility to the general public is limited by the steep ascent and very uneven paths. In addition, there are currently no dedicated car parks. Working with the local authorities would certainly address these problems, as there is a state-owned area just outside the entrance to the lava flow field, where there is also a small building used as a tool shed. This area could be used profitably for bus and car parking, and the building could be restored to provide a visitor point with equipment suitable for accessing the website and projecting videos of the sites.

The possible visitors who can profitably enjoy the sites can be divided into three main target groups. The first group could include people who are unfamiliar with the scientific context of volcanic areas, but who enjoy walking in the open air and beautiful panoramas, and who can increase their curiosity about geology thanks to a visit to unique sites such as the lava tubes. Meanwhile, this geo-tour can help make them aware of the hazards to which the Vesuvius area is exposed, which in the long term could increase the preparedness of the local population at risk. The second group of people for whom the visit could be fruitful are the students from secondary schools, who can join the visit of the sites with the visit of the Vesuvian Observatory Museum, in the framework of school programmes dedicated to the diffusion of volcanological themes among young people. The final aim of these activities could be improving the level of resilience of the area, by increasing the correct knowledge of the volcano as a source of hazard, but also as a resource for the area. Finally, geo-experts would visit the sites to learn more about the various manifestations of effusive activity at Somma-Vesuvius and the characteristics of lava tubes formed in tephriphonolitic lava flows.

6.5.4 SWOT analysis

The assessment of each site has highlighted the strengths and opportunities of this volcanic landforms geotinerary along with the weaknesses and threats associated. These aspects are recapitulated using a SWOT analysis for the complete geotinerary (Table 14).

Table 14. Results of the SWOT (Strengths, Weaknesses, Opportunities, and Threats) analysis.

Strengths	Weaknesses
1. Area internationally renowned 2. Discovery of a poorly known area of Vesuvius and its National Park 3. One area of the National Park already attracts many tourists (the summit cone, called Gran Cono) 4. High didactic value to explain the formation of volcanological landforms during lava flow emplacement 5. The site is easily accessible by car or public transport 6. Opportunity to observe the bay of Naples from a nice point of view	1. The geoitinerary is at an initial stage and deserves to be further developed 2. The geoitinerary is a personal initiative by the Authors and lacks, up to now, of an intense collaboration with local administration 3. The full development of the geoitinerary needs funding and a management policy 4. Local people and local administration have not yet fully understood the high touristic and didactic potential of the area 5. Within the National Park, the tourist traffic is limited to the summit cone 6. Lack of parking spaces near the geoitinerary 7. The geoitinerary can be very hard to follow during spring and summer due to the intense blooming of the vegetation, therefore maintenance is necessary
Opportunities	Threats
1. The geoitinerary could be integrated into a wider tour because of its closeness to the main touristic attraction of the National Park, the city of Naples and the archaeological sites of the Vesuvian area (i.e., Pompeii, Herculaneum, Oplontis) 2. Local accommodation facilities can easily host a large amount of tourists 3. The geoitinerary can be explored rapidly (less than 1 hour) 4. The geoitinerary can be combined easily with other sites of utmost volcanic interest in the vicinity	1. Lack of financial resource to develop the geoitinerary 2. Low interest in local authorities for the development of geotourism

6.5.5 Strengths and weaknesses of the applied methods

Our employment of a multimethodological approach involving fieldwork, immersive photography and terrestrial laser scanning to document the distinctive Vesuvius lava tubes for the first time, ultimately prompts us to evaluate the strengths and limitations of the chosen methodology.

Fieldwork is necessary in order to locate interesting and important geological sites. It is also crucial to describe volcanological structures in detail, and to design and test a geotourism itinerary that links all the sites together. However, fieldwork can be challenging when the terrain is rough or heavily vegetated. In some cases, it can be dangerous, particularly in underground environments where there is a high risk of collapse. Capturing immersive photos of the sites using a 360° camera allows the visitor to navigate and observe the volcanological structure from

anywhere. The instrumentation is small, cheap, mainstream and easily manipulated. Treatment of the data is simple in the constructor's software. The only limitation of this method resides in hosting and publishing the 360° images on compatible servers/platforms and online viewers to make them available to online visitors.

Digitising the lava tube using a Terrestrial Laser Scanner provided a high-resolution 3D model that captures the morphometry and state of this site at a moment in time. 3D scanning allows the conservation of a digital twin of the site to be re-examined or shown throughout time, even if the original site erodes and/or collapses. However, this method is more complex to apply and requires large, expensive instrumentation operated by experts. The treatment of the data can be tedious, and the constructor's software is complex to use.

6.6. Conclusive remarks

The Somma-Vesuvius volcanic complex boasts incredible geological features largely neglected by tourists. Although it is visited by large numbers of tourists, access to the Vesuvius National Park is limited to the top of the Vesuvius crater. To communicate and share knowledge about volcanoes, their products and their hazards, we believe that it is necessary to expand the visitor experience in the National Park to include peculiar geosites. In this study, we focused our attention on the 1858 lava flow field due to the large variety of surface features that are beautiful volcanological examples of landforms formed during recent effusive eruptions and easily accessible by visitors. We proposed an on-site geoitinerary and an online virtual tour regrouping five interesting volcanological sites. We assessed and highlighted the geoheritage potential of the geoitinerary for local tourism, scientific dissemination, conservation and enhancement of volcanic risk perception.

GENERAL CONCLUSION

The main objective of this thesis was to investigate lava flow emplacement dynamics and lava tube formation and development. The effusive processes at Somma-Vesuvius lacked attention due to the concentration of most scientific literature on its past explosive activity. This thesis is one of the first works in several decades focusing on the effusive activity and its processes at Vesuvius and the first showing the presence of lava tubes at volcanoes that erupted silica-undersaturated alkali-rich lavas. Our multidisciplinary study was carried out as follow: (1) analysis of the 1858 lava flow field emplacement dynamics and morphometry analysis of the largest lava tube, (2) petrological, geochemical and textural analysis of tube-forming lavas, (3) stability assessment of the lava tube structure and (4) development of a geotinerary and virtual tour.

The reconstruction of the 1858 lava flow field development at various stages was inferred from historical documents. This reconstruction shows the dynamics of lava flow emplacement. The lava flow field rapidly reached its maximum extent and then grew both laterally and vertically. Several lava flows were piled up to a maximum height of over one hundred metres, limiting the length of the flow field to 3 kilometres. Our findings suggest that the limited extent of the flow field is a consequence of both the pre-eruptive topography and the amount of phenocrysts in the lava. Reconstructing the lava flow field emplacement dynamics helps understand the formation and development of lava transport structures.

The largest lava tube formed in the 1858 lava flow field was re-examined after more than 100 years using modern technologies. The digitisation produced a complete 3D model used to carry out a detailed morphological study of the lava tube. The shape in plan view of this lava tube is explained by its origin, namely the emplacement of a pāhoehoe breakout that eventually drained. The formation of this lava tube is mainly explained by the process of lava flow inflation. The presence of internal features such as multiple inner coatings and lava stalactites is evidence that lava flowed into the cavity and, therefore, that this cavity could be described as a lava tube. Internal features are also important to know and understand the thermal and rheological history of the lava flowing inside the lava tube. We provided a petrological, geochemical and textural analysis of the lavas forming the tube. Our analyses show the effect of the degree of undercooling and redox conditions on crystallisation kinetics. The crystallisation variations influenced lava tube formation and development. The lava tube formed when a small breakout surface solidified rapidly by atmospheric cooling, forcing the flow to uplift. Here, lava tube formation is controlled by cooling rate and effusive rate. Inside the lava tube, several linings solidified on the walls and ceiling under temperatures ranging

from 1115 ± 25 °C to 970 °C and with an oxygen fugacity ranging from QFM +3 to QFM +4.5. Lava tube activity is controlled by the effusion rate, which changes the level of lava flowing inside the tube. When lava flows inside the lava tube, crystallisation depends on the degree of undercooling, ranging here from high (60 °C/h) to low (1 °C/h), and the amount of shear rate, increasing lava viscosity up to two orders of magnitude. This controls not only the thermo-rheological properties of lava arriving at the flow front and thus the ability of the flow to advance, but also the crystallisation of lava on the walls and ceiling of the lava tube that impacts its structure, forming millimetres to centimetres thick layers.

In the frame of hazard assessment at Vesuvius, this study is the first step in getting knowledge about the presence and mechanism of lava tube formation in the lava flows of Vesuvius. The possible formation of lava tubes, even of limited size, has to be taken into account when forecasting the reach limit of lava flows, as it can quickly affect populated and built-up areas. Vesuvius is densely populated, and lava flow hazard is a main concern.

The acoustic emission testing with AE signals acquired in six different areas inside the largest lava tube of the 1858 lava flow field suggests a high stability of the lava tube structure. An important number of AE signals were acquired at the sides of the cavity, on the wall linings, indicating the presence of loosened masses or materials. The time domain analysis revealed the very low amplitude and count of these signals, commonly related to insignificant microstructural movements in the structure that do not affect the general stability. Unlike the frequency analysis, the time-frequency analysis using a combination of Empirical Mode Decomposition and Hilbert-Huang Transform explicitly showed the differences between the AE signals generated from the different layouts. The similarity indices verified that all the signals acquired inside the lava tube may have originated from a similar source.

This thesis investigated incredible structures, lava tubes, at Somma-Vesuvius. As the large numbers of tourists visiting the Vesuvius National Park are limited to the top of the Vesuvius crater, such volcanic landforms present lower on the slopes of the volcano are largely neglected. The 1858 lava flow field formed a multitude of interesting and preserved volcanic landforms. Thus, to expand the visitor experience at Somma-Vesuvius, we developed a geotinerary and virtual tour along the 1858 lava flow field that highlights five peculiar volcanic landforms, comprising three lava tubes, an ephemeral vent and a tumulus. The aim is to allow visitors to discover volcanic landforms, disseminate scientific knowledge and raise awareness on volcanoes and lava flow hazards.

PERSPECTIVES

Our results suggest that the work carried out during this study needs to be continued to further understand lava tube formation and development. The petrology, geochemistry, and texture of tube-forming lavas in volcanoes producing lavas of different compositions must be characterised to constrain the influence of lava composition and crystal content on the formation and development of lava tubes. It would be interesting to go further into the investigation of lava tubes internal conditions. Constraining the evolution of the temperature, oxygen fugacity, and viscosity could be performed by analysing internal features or by measuring gas composition and temperature during active flow.

The AE testing analysis could be extended by analysing other areas of the lava tube, especially narrower regions, to analyse their structural stability. Furthermore, sensors could be installed over a prolonged period of time to determine whether there are any tempestuous periods of AE signals present in the different layouts. Machine learning methods could be applied to identify and classify the signal differences if a significant variation in the AE signals generated by each layout is measured. AE testing of lava tubes accessible to visitors must be conducted to attest the stability of the structure and ensure safe access.

Further work must be carried out in the detection and mapping of lava tubes, both in old and active lava flows. It would be interesting to calibrate the detection of lava tubes by geophysical methods, especially by airborne magnetic surveys, using the morphometry analysis carried out by 3D scanning and airborne imagery. Airborne magnetic surveys would allow the detection and mapping of unknown and inaccessible lava tubes over large and perilous areas. Such technology will also be used to detect lava tubes on other planetary bodies to detect extraterrestrial lava tubes during exploration missions. Low interval monitoring of active lava flow emplacement and lava tube formation could be carried out using airborne thermal imagery during and after eruption. Thermal imagery reveals lava transport pathways to precisely map and track through time the evolution of lava tube systems.

It would be interesting to implement the lava tubes 3D models into Augmented Reality (AR) or Virtual Reality (VR) environments to allow a full immersion and exploration of these underground features for detailed morphological analysis, and even for scientific dissemination and public outreach.

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Appendices

Appendix 1. Crystal compositions, raw data. Compositions are expressed in wt.%.

Sample	Area	Phase	SiO ₂	TiO ₂	Al ₂ O ₃	Cr ₂ O ₃	FeO	MnO	MgO	CaO	Na ₂ O	K ₂ O	Total
TL13	1	Pl	48.71	0.07	31.84	0.01	0.95	0.00	0.06	16.04	2.27	0.41	100.36
TL13	1	Pl	54.81	0.19	26.58	0.00	1.76	0.00	0.14	10.37	4.28	1.75	99.89
TL13	1	Pl	50.21	0.11	30.90	0.00	0.99	0.04	0.09	14.60	2.89	0.62	100.45
TL13	1	Cpx	47.27	1.91	7.12	0.00	8.73	0.16	11.85	23.29	0.35	0.02	100.68
TL13	1	Cpx	48.49	1.42	5.13	0.02	8.43	0.17	12.72	23.11	0.32	0.04	99.84
TL13	1	Ol	37.73	0.07	0.03	0.00	28.72	0.61	32.93	0.68	0.05	0.02	100.84
TL13	1	Cpx	48.57	1.49	5.35	0.04	8.49	0.12	12.69	22.97	0.35	0.06	100.12
TL13	1	Cpx	42.82	2.69	8.30	0.01	13.16	0.27	9.58	21.56	0.44	0.05	98.90
TL13	1	Cpx	39.86	3.45	11.35	0.00	13.47	0.30	8.17	21.60	0.58	0.22	99.00
TL13	1	Cpx	46.43	2.04	7.96	0.08	8.74	0.14	11.54	23.67	0.33	0.00	100.95
TL13	1	Cpx	45.99	1.88	8.16	0.01	9.35	0.13	11.48	23.20	0.35	0.02	100.57
TL13	1	Cpx	48.76	1.38	5.11	0.06	8.73	0.17	12.49	23.17	0.32	0.02	100.20
TL13	1	Cpx	46.53	1.94	7.56	0.06	9.55	0.14	11.50	22.74	0.38	0.05	100.46
TL13	1	Cpx	41.36	3.07	10.48	0.10	13.63	0.21	8.39	21.35	0.60	0.13	99.32
TL13	1	Cpx	45.22	2.25	8.69	0.03	8.87	0.19	11.13	23.18	0.37	0.04	99.97
TL13	1	Cpx	44.37	2.55	8.97	0.00	9.67	0.15	10.75	22.58	0.39	0.05	99.47
TL13	1	Pl	50.39	0.11	30.86	0.00	1.07	0.01	0.04	14.62	2.78	0.60	100.47
TL13	1	Pl	51.80	0.09	30.16	0.01	0.94	0.00	0.05	13.66	3.24	0.72	100.68
TL13	1	Pl	50.40	0.10	31.08	0.01	1.09	0.02	0.07	14.61	2.78	0.62	100.80
TL13	1	Pl	50.51	0.10	30.64	0.00	1.09	0.01	0.10	14.48	2.82	0.62	100.37
TL13	1	Pl	51.39	0.10	30.26	0.01	0.90	0.04	0.10	14.10	3.19	0.69	100.78
TL13	1	Pl	54.09	0.17	27.18	0.00	1.57	0.03	0.08	10.46	4.03	1.67	99.26
TL13	1	Pl	52.91	0.14	28.73	0.04	1.15	0.00	0.09	11.93	3.81	1.14	99.94
TL13	1	Pl	53.63	0.16	27.49	0.03	1.62	0.00	0.11	11.19	3.93	1.39	99.54
TL13	1	Pl	52.19	0.15	28.73	0.02	1.55	0.02	0.12	12.69	3.38	1.02	99.87
TL13	1	Ox	0.14	11.42	7.70	0.05	70.08	0.66	3.66	0.20	0.05	0.03	93.99
TL13	1	Ox	1.85	5.21	1.81	0.00	77.07	1.40	1.59	1.03	0.18	0.24	90.39
TL13	1	Cpx	46.16	1.28	17.64	0.01	7.57	0.08	4.53	14.98	2.56	1.39	96.22
TL13	1	Pl	49.00	0.05	31.77	0.00	0.91	0.04	0.05	15.48	2.28	0.50	100.08
TL13	1	Pl	49.75	0.07	31.49	0.00	0.83	0.00	0.12	15.14	2.53	0.49	100.43
TL13	2	Ox	0.19	11.38	8.04	0.17	70.40	0.44	4.23	0.07	0.03	0.02	94.98
TL13	2	Ox	0.27	11.42	7.99	0.09	70.33	0.54	3.84	0.22	0.04	0.04	94.78
TL13	2	Ox	0.20	10.23	6.71	0.01	65.63	0.49	2.82	4.10	0.00	0.04	90.24
TL13	2	Lct	55.59	0.08	23.26	0.07	0.30	0.00	0.01	0.00	0.49	20.98	100.77
TL13	2	Lct	55.39	0.10	23.03	0.00	0.43	0.02	0.00	0.00	0.43	21.10	100.49
TL13	2	Pl	48.51	0.05	32.34	0.01	0.65	0.03	0.07	16.12	2.07	0.43	100.27
TL13	2	Pl	48.20	0.06	31.86	0.00	0.69	0.00	0.11	15.98	2.09	0.41	99.40
TL13	2	Pl	48.68	0.06	31.83	0.04	0.94	0.02	0.07	15.69	2.24	0.47	100.05
TL13	2	Pl	48.53	0.07	31.92	0.00	0.97	0.02	0.09	15.33	2.42	0.52	99.88
TL13	2	Pl	54.06	0.14	27.36	0.01	1.33	0.05	0.09	10.44	4.08	1.61	99.17

Appendices

TL13	2	Pl	48.60	0.07	32.05	0.00	0.91	0.04	0.09	16.04	2.18	0.44	100.42
TL13	2	Pl	49.72	0.08	31.28	0.01	0.87	0.03	0.10	15.11	2.56	0.57	100.32
TL13	2	Cpx	49.54	1.24	5.17	0.00	8.14	0.25	13.25	23.44	0.33	0.00	101.36
TL13	2	Cpx	43.29	2.81	9.87	0.00	10.20	0.17	10.42	22.83	0.38	0.04	100.00
TL13	2	Cpx	47.27	1.86	6.43	0.00	9.16	0.13	11.95	22.92	0.33	0.02	100.06
TL13	2	Cpx	42.77	2.30	8.96	0.02	13.68	0.17	9.35	21.81	0.46	0.03	99.57
TL13	2	Cpx	43.54	2.92	9.28	0.10	10.97	0.18	9.82	22.86	0.42	0.04	100.13
TL13	2	Cpx	44.16	2.56	9.32	0.01	9.44	0.17	10.47	23.05	0.36	0.01	99.53
TL13	2	Ol	37.57	0.05	0.07	0.07	29.32	0.65	32.49	0.86	0.02	0.01	101.11
TL13	2	Ol	37.54	0.08	0.05	0.04	28.18	0.63	33.58	0.71	0.04	0.02	100.86
TL13	2	Cpx	49.08	1.34	5.02	0.02	8.78	0.24	12.78	22.95	0.30	0.02	100.53
TL13	2	Cpx	44.97	2.33	8.83	0.04	9.10	0.18	11.22	23.24	0.30	0.00	100.22
TL13	2	Cpx	44.84	2.32	8.65	0.03	8.75	0.13	11.21	23.32	0.32	0.00	99.58
TL13	2	Cpx	49.06	1.26	5.20	0.00	7.75	0.20	13.09	23.14	0.32	0.00	100.02
TL13	2	Cpx	48.41	1.39	5.18	0.05	8.31	0.23	12.85	23.08	0.31	0.01	99.82
TL13	2	Ap	0.48	0.01	0.01	0.00	0.42	0.03	0.27	53.51	0.06	0.01	54.80
TL13	2	Pl	49.70	0.12	30.93	0.00	0.98	0.00	0.07	14.33	2.92	0.60	99.63
TL13	2	Pl	50.73	0.10	30.34	0.02	0.82	0.01	0.08	14.43	2.96	0.70	100.21
TL13	2	Pl	50.97	0.11	30.32	0.00	0.88	0.01	0.11	13.98	3.15	0.74	100.27
TL13	2	Ox	2.71	5.86	2.04	0.06	75.79	1.70	1.84	1.07	0.12	0.24	91.43
TL13	2	Cpx	46.11	2.18	7.91	0.02	9.20	0.17	11.50	23.06	0.33	0.02	100.49
TL13	3	Ox	0.08	11.70	7.95	0.04	69.27	0.49	4.11	0.13	0.03	0.03	93.83
TL13	3	Ox	0.21	14.36	8.36	0.18	66.81	0.33	4.22	0.21	0.00	0.02	94.70
TL13	3	Ox	0.17	11.26	8.17	0.12	70.01	0.42	4.05	0.14	0.04	0.01	94.39
TL13	3	Ox	0.65	4.22	1.22	0.06	80.78	0.66	0.68	0.23	0.16	0.16	88.83
TL13	3	Ox	0.63	3.47	1.18	0.00	81.59	0.42	0.41	0.47	0.12	0.10	88.37
TL13	3	Ox	0.09	12.24	7.78	0.03	69.91	0.52	3.71	0.09	0.04	0.03	94.45
TL13	3	Ap	0.38	0.00	0.00	0.03	0.28	0.07	0.19	54.49	0.03	0.03	55.50
TL13	3	Incl.	45.86	1.15	17.12	0.05	9.10	0.26	4.42	14.01	3.61	3.57	99.15
TL13	3	Ap	0.38	0.00	0.01	0.00	0.25	0.02	0.26	54.83	0.05	0.01	55.82
TL13	3	Ol	38.32	0.05	0.06	0.00	25.90	0.57	35.19	0.59	0.02	0.01	100.72
TL13	3	Cpx	45.79	2.02	7.86	0.06	8.65	0.12	11.30	23.63	0.34	0.01	99.78
TL13	3	Cpx	49.05	1.27	5.01	0.00	8.38	0.19	13.08	23.08	0.33	0.02	100.43
TL13	3	Cpx	45.38	2.41	8.42	0.00	9.46	0.16	11.06	23.25	0.36	0.02	100.54
TL13	3	Cpx	46.24	1.99	7.79	0.06	8.56	0.08	11.56	23.13	0.32	0.00	99.72
TL13	3	Pl	49.68	0.05	31.54	0.02	0.69	0.00	0.09	15.43	2.50	0.58	100.57
TL13	3	Pl	53.31	0.17	27.29	0.00	1.65	0.01	0.12	10.86	3.91	1.57	98.88
TL13	3	Pl	49.94	0.09	31.05	0.00	0.93	0.00	0.13	15.07	2.77	0.59	100.58
TL16	1	Lct	56.03	0.08	23.75	0.09	0.42	0.00	0.00	0.01	0.68	20.72	101.77
TL16	1	Cpx	52.26	0.53	2.66	0.13	4.25	0.08	15.82	24.41	0.18	0.03	100.35
TL16	1	Cpx	46.90	1.90	7.95	0.03	8.65	0.15	11.79	23.44	0.35	0.00	101.16
TL16	1	Ol	37.68	0.03	0.04	0.01	27.92	0.58	34.06	0.70	0.03	0.01	101.05
TL16	1	Cpx	42.18	3.39	10.65	0.02	10.14	0.23	9.80	22.87	0.39	0.02	99.70
TL16	1	Ol	36.78	0.07	0.06	0.01	32.68	0.65	29.41	0.72	0.03	0.01	100.42

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TL16	1	Ap	2.36	0.00	0.03	0.00	1.80	0.08	1.07	52.26	0.03	0.03	57.67
TL16	1	Cpx	48.89	1.46	4.87	0.02	9.46	0.33	12.59	22.24	0.34	0.03	100.23
TL16	1	Cpx	49.21	1.35	4.75	0.05	9.41	0.16	12.58	22.99	0.33	0.04	100.87
TL16	1	Cpx	48.67	1.49	5.24	0.00	9.02	0.18	12.39	22.44	0.34	0.01	99.79
TL16	1	Cpx	48.10	1.86	5.40	0.00	11.10	0.30	10.92	22.15	0.37	0.02	100.22
TL16	1	Cpx	43.93	2.86	9.10	0.00	9.77	0.22	10.36	22.68	0.43	0.05	99.40
TL16	1	Cpx	41.54	4.69	9.75	0.00	13.48	0.31	7.52	21.65	0.64	0.04	99.63
TL16	1	Cpx	46.66	1.97	6.21	0.00	9.60	0.24	11.57	23.32	0.37	0.01	99.95
TL16	1	Ox	2.76	0.78	2.41	0.06	83.41	0.88	0.29	0.47	0.76	0.25	92.07
TL16	1	Ox	2.01	6.42	1.93	0.00	74.02	2.70	2.67	0.27	0.16	0.33	90.51
TL16	1	Ox	0.24	2.21	1.03	0.05	84.37	1.19	1.43	0.26	0.04	0.07	90.89
TL16	1	Ox	0.28	3.81	0.70	0.00	83.19	0.16	0.28	0.33	0.27	0.15	89.17
TL16	1	Ox	0.07	0.88	1.11	0.06	85.61	0.21	0.11	0.36	0.00	0.12	88.54
TL16	1	Ox	0.09	5.16	8.39	0.08	77.48	0.67	2.65	0.16	0.04	0.03	94.76
TL16	1	Ox	0.08	5.32	7.79	0.03	76.04	0.67	2.42	0.41	0.00	0.06	92.82
TL16	1	Ox	0.07	3.57	2.96	0.02	83.12	1.32	1.77	0.22	0.03	0.03	93.12
TL16	1	Ox	0.09	10.43	4.91	0.09	75.12	1.42	1.72	0.11	0.03	0.04	93.97
TL16	1	Pl	50.10	0.10	31.41	0.00	0.82	0.04	0.10	14.93	2.81	0.57	100.88
TL16	1	Pl	48.99	0.04	32.32	0.02	0.95	0.07	0.08	15.21	2.54	0.45	100.67
TL16	1	Pl	48.01	0.06	32.68	0.00	0.68	0.05	0.05	16.06	1.95	0.38	99.91
TL16	1	Pl	52.73	0.10	29.45	0.00	0.77	0.04	0.04	12.47	3.96	0.77	100.32
TL16	1	Pl	55.97	0.16	27.15	0.00	0.74	0.00	0.00	8.93	5.12	1.22	99.28
TL16	1	Pl	51.68	0.09	30.30	0.06	0.90	0.00	0.07	13.34	3.37	0.68	100.48
TL16	1	Pl	49.26	0.08	31.74	0.00	0.79	0.01	0.06	15.13	2.53	0.53	100.13
TL16	1	Pl	48.95	0.09	32.18	0.02	0.67	0.01	0.08	15.47	2.14	0.47	100.09
TL16	1	Pl	48.43	0.08	32.16	0.00	0.68	0.02	0.09	16.12	2.17	0.47	100.22
TL16	1	Pl	60.94	0.11	24.52	0.06	0.69	0.04	0.00	6.54	6.78	1.91	101.58
TL16	1	Pl	49.35	0.08	31.87	0.00	1.00	0.01	0.13	15.32	2.52	0.46	100.74
TL16	1	Pl	51.01	0.11	30.25	0.03	0.72	0.03	0.06	13.27	3.47	0.69	99.63
TL16	1	Pl	62.85	0.45	22.52	0.00	0.84	0.00	0.00	4.01	6.48	4.70	101.85
TL16	1	Pl	57.03	0.16	25.68	0.05	0.47	0.00	0.03	7.00	5.43	1.66	97.51
TL16	1	Pl	52.58	0.11	29.69	0.03	0.67	0.00	0.05	12.39	4.03	0.55	100.09
TL16	2	Pl	51.97	0.11	30.02	0.00	0.79	0.02	0.10	13.11	3.53	0.69	100.33
TL16	2	Pl	48.69	0.06	32.25	0.00	0.66	0.02	0.08	15.72	2.26	0.50	100.25
TL16	2	Pl	49.73	0.09	31.44	0.00	0.71	0.02	0.06	14.74	2.79	0.51	100.09
TL16	2	Pl	50.20	0.07	31.06	0.08	0.78	0.00	0.05	14.65	2.78	0.57	100.25
TL16	2	Pl	48.57	0.03	32.07	0.04	0.64	0.06	0.09	15.93	2.16	0.47	100.06
TL16	2	Pl	47.99	0.04	32.71	0.05	0.70	0.01	0.06	16.47	1.83	0.35	100.21
TL16	2	Pl	51.18	0.07	30.48	0.00	0.66	0.00	0.10	13.99	3.08	0.71	100.28
TL16	2	Pl	50.89	0.11	30.47	0.00	0.85	0.00	0.12	13.36	3.31	0.67	99.78
TL16	2	Lct	55.83	0.17	23.91	0.07	0.67	0.04	0.00	0.00	0.31	21.37	102.36
TL16	2	Pl	54.08	0.15	28.40	0.02	0.83	0.00	0.07	10.95	4.53	0.80	99.82
TL16	2	Pl	55.65	0.17	27.18	0.04	0.96	0.05	0.04	9.12	5.43	0.94	99.57
TL16	2	Pl	56.17	0.12	26.91	0.00	0.75	0.03	0.05	8.95	5.05	1.32	99.34

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TL16	2	Pl	55.32	0.17	27.47	0.00	0.71	0.01	0.01	9.47	5.06	1.17	99.40
TL16	2	Pl	58.10	0.14	25.89	0.04	0.47	0.00	0.01	7.53	6.04	1.26	99.47
TL16	2	Ox	0.09	9.11	7.66	0.08	72.81	0.84	2.33	0.18	0.00	0.02	93.11
TL16	2	Ox	0.09	12.23	8.10	0.10	71.24	0.56	2.25	0.13	0.02	0.07	94.78
TL16	2	Ox	0.10	8.77	4.22	0.03	77.92	0.88	1.07	0.13	0.00	0.07	93.18
TL16	2	Ox	1.24	10.03	1.88	0.02	75.99	1.46	0.88	0.37	0.26	0.16	92.29
TL16	2	Ox	0.16	8.77	1.23	0.00	79.07	1.32	0.78	0.16	0.00	0.05	91.54
TL16	2	Ox	0.50	8.67	1.82	0.00	77.85	1.49	1.06	0.22	0.08	0.12	91.81
TL16	2	Ox	0.14	6.10	2.15	0.00	79.68	1.94	0.14	0.17	0.00	0.11	90.42
TL16	2	Ox	0.08	3.53	4.70	0.05	80.93	0.73	1.68	0.64	0.02	0.03	92.39
TL16	2	Ol	36.35	0.05	0.06	0.00	32.25	0.69	29.60	0.77	0.06	0.03	99.87
TL16	2	Ol	34.47	0.12	0.06	0.00	42.77	1.66	20.00	0.83	0.07	0.03	100.00
TL16	2	Ol	36.42	0.06	0.04	0.07	33.43	0.74	29.46	0.73	0.03	0.02	101.01
TL16	2	Ol	37.37	0.04	0.05	0.04	27.47	0.56	34.31	0.66	0.01	0.02	100.53
TL16	2	Ol	35.40	0.08	0.04	0.00	39.13	0.99	24.06	0.76	0.04	0.07	100.57
TL16	2	Ap	0.34	0.01	0.01	0.00	0.25	0.00	0.24	54.79	0.00	0.02	55.67
TL16	2	Incl.	39.02	0.02	0.04	0.00	20.72	0.46	38.86	1.52	0.01	0.00	100.66
TL16	2	Pl	54.17	0.13	28.44	0.00	0.70	0.00	0.05	11.51	4.41	0.78	100.19
TL16	2	Pl	53.11	0.14	28.98	0.00	0.76	0.00	0.08	11.82	4.30	0.76	99.95
TL16	2	Pl	59.75	0.34	21.73	0.00	1.21	0.06	0.10	5.86	6.27	3.78	99.10
TL16	2	Lct	55.20	0.12	23.56	0.00	0.44	0.00	0.02	0.02	0.19	21.52	101.07
TL16	2	Cpx	45.27	2.35	8.87	0.04	8.83	0.12	11.08	23.69	0.29	0.00	100.54
TL16	2	Cpx	49.01	1.38	5.39	0.00	7.98	0.11	12.65	23.22	0.30	0.01	100.05
TL16	2	Cpx	48.23	1.59	5.62	0.04	9.47	0.13	12.21	22.82	0.37	0.04	100.52
TL16	2	Ap	0.66	0.02	0.04	0.00	0.47	0.07	0.27	54.11	0.05	0.02	55.70
TL16	2	Lct	55.64	0.10	23.76	0.00	0.50	0.02	0.01	0.00	0.27	21.52	101.82
TL16	2	Cpx	48.60	1.47	5.19	0.04	8.82	0.19	12.73	22.89	0.29	0.02	100.24
TL16	2	Ol	37.11	0.05	0.03	0.00	29.13	0.58	32.71	0.65	0.03	0.04	100.33
TL16	2	Cpx	44.31	2.83	9.25	0.01	9.80	0.23	10.43	23.09	0.42	0.02	100.39
TL16	2	Cpx	45.73	1.97	8.17	0.00	9.11	0.15	11.29	23.27	0.38	0.02	100.08
TL16	2	Cpx	44.83	2.58	7.87	0.02	12.20	0.36	9.13	22.12	0.54	0.11	99.76
TL16	2	Cpx	44.44	3.11	8.87	0.05	10.56	0.22	9.77	22.78	0.47	0.04	100.29
TL16	2	Cpx	46.20	2.23	6.05	0.00	12.70	0.31	9.93	21.22	0.46	0.37	99.46
TL16	2	Cpx	42.73	3.65	9.56	0.04	12.90	0.32	7.57	22.33	0.62	0.03	99.75
TL16	2	Cpx	41.17	5.03	9.89	0.02	13.55	0.33	7.30	21.71	0.60	0.09	99.69
TL16	3	Ol	37.58	0.03	0.07	0.03	28.34	0.63	33.80	0.64	0.03	0.01	101.15
TL16	3	Ol	37.50	0.03	0.06	0.03	28.67	0.61	33.23	0.66	0.02	0.00	100.80
TL16	3	Ol	36.92	0.07	0.06	0.00	31.36	0.73	31.10	0.62	0.07	0.01	100.93
TL16	3	Cpx	40.87	4.89	11.09	0.00	12.32	0.23	7.89	22.37	0.54	0.03	100.24
TL16	3	Cpx	46.43	2.39	5.98	0.03	12.42	0.25	9.95	22.04	0.41	0.02	99.90
TL16	3	Cpx	43.54	2.83	9.79	0.01	9.41	0.18	10.22	23.02	0.38	0.03	99.42
TL16	3	Cpx	45.16	2.84	6.60	0.03	13.12	0.33	9.10	22.23	0.54	0.03	99.99
TL16	3	Cpx	47.96	1.89	5.07	0.04	11.49	0.23	10.80	22.17	0.34	0.09	100.08
TL16	3	Cpx	44.63	3.00	7.49	0.00	12.67	0.26	9.33	21.96	0.48	0.02	99.86

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TL16	3	Cpx	46.92	1.82	7.73	0.01	8.57	0.19	11.90	23.25	0.33	0.01	100.74
TL16	3	Ol	38.41	0.03	0.04	0.03	26.04	0.52	35.42	0.58	0.02	0.00	101.09
TL16	3	Cpx	46.14	1.97	8.15	0.03	8.51	0.22	11.48	23.21	0.38	0.01	100.09
TL16	3	Ap	0.44	0.00	0.01	0.05	0.81	0.09	0.24	53.63	0.03	0.02	55.32
TL16	3	Cpx	45.63	3.08	6.17	0.00	13.63	0.35	8.77	21.57	0.54	0.07	99.82
TL16	3	Cpx	47.36	2.76	11.76	0.00	11.40	0.27	5.99	15.46	1.87	1.78	98.65
TL16	3	Ox	0.16	10.51	1.56	0.00	78.30	1.32	0.47	0.33	0.01	0.05	92.70
TL16	3	Ox	0.46	6.94	1.21	0.07	81.18	1.31	0.68	0.16	0.01	0.04	92.07
TL16	3	Ox	5.82	13.73	3.12	0.03	70.32	0.88	0.20	0.28	1.10	0.40	95.89
TL16	3	Pl	51.10	0.08	30.63	0.01	0.87	0.01	0.05	13.85	3.29	0.63	100.50
TL16	3	Pl	53.83	0.12	28.88	0.00	0.74	0.00	0.04	11.47	4.44	0.74	100.26
TL16	3	Pl	51.89	0.08	30.18	0.01	0.93	0.00	0.11	13.22	3.31	0.97	100.71
TL16	3	Pl	51.54	0.13	30.47	0.00	0.96	0.01	0.02	13.23	3.64	0.51	100.50
TL16	3	Pl	51.23	0.09	30.38	0.00	0.94	0.00	0.03	13.25	3.62	0.34	99.90
TL16	3	Pl	55.89	0.14	27.44	0.00	0.89	0.03	0.04	9.44	5.39	0.92	100.18
TL16	3	Pl	65.35	0.10	21.06	0.04	0.74	0.01	0.00	1.92	7.24	5.09	101.55
TL16	3	Pl	54.77	0.13	28.21	0.00	0.73	0.00	0.04	10.61	4.76	0.73	99.97
TL16	3	Pl	52.24	0.09	30.26	0.03	0.91	0.04	0.05	12.91	4.01	0.36	100.91
TL16	3	Pl	52.76	0.11	30.05	0.00	0.88	0.02	0.04	12.61	4.01	0.39	100.87
TL16	3	Pl	65.87	0.03	20.44	0.00	0.64	0.04	0.00	1.50	6.62	5.54	100.68
TL16	3	Pl	50.14	0.09	31.09	0.03	0.89	0.01	0.05	14.43	3.00	0.62	100.36
TL22	1	Pl	48.82	0.07	32.40	0.00	0.63	0.03	0.11	15.70	2.21	0.47	100.45
TL22	1	Pl	50.39	0.08	31.52	0.00	0.78	0.00	0.09	14.36	2.74	0.57	100.53
TL22	1	Pl	48.47	0.05	32.66	0.00	0.71	0.00	0.10	15.47	2.38	0.49	100.32
TL22	1	Pl	49.98	0.08	31.40	0.00	0.83	0.04	0.11	14.75	3.03	0.42	100.64
TL22	1	Pl	48.93	0.06	32.22	0.00	0.83	0.02	0.07	15.91	2.26	0.41	100.73
TL22	1	Pl	52.24	0.12	29.67	0.00	1.04	0.03	0.09	12.59	4.12	0.55	100.46
TL22	1	Pl	51.66	0.09	29.85	0.05	1.11	0.05	0.11	12.00	4.27	0.43	99.61
TL22	1	Pl	52.44	0.10	30.04	0.03	0.82	0.00	0.06	12.97	3.89	0.45	100.80
TL22	1	Pl	52.13	0.14	29.55	0.03	1.19	0.01	0.13	12.50	4.25	0.38	100.31
TL22	1	Ox	0.19	19.07	4.10	0.10	65.91	1.06	0.32	0.71	0.11	0.06	91.60
TL22	1	Cpx	42.29	3.38	9.97	0.02	11.33	0.15	9.67	22.30	0.42	0.03	99.56
TL22	1	Pl	55.22	0.08	27.49	0.00	0.74	0.00	0.05	8.60	7.10	0.69	99.98
TL22	1	Pl	53.11	0.07	28.96	0.06	0.67	0.00	0.05	11.97	4.24	0.63	99.75
TL22	1	Pl	55.47	0.07	24.90	0.03	0.50	0.00	0.02	0.07	0.42	21.05	102.54
TL22	1	Cpx	43.82	1.79	7.89	0.00	9.38	0.23	9.48	23.55	0.81	0.31	97.26
TL22	1	Ol	34.58	0.10	0.06	0.00	44.03	1.28	18.75	0.93	0.13	0.07	99.91
TL22	1	Pl	52.47	0.10	29.94	0.03	0.76	0.01	0.11	12.43	4.18	0.42	100.43
TL22	1	Ol	36.55	0.16	0.43	0.06	33.59	0.86	27.30	1.36	0.17	0.22	100.69
TL22	1	Ol	35.44	0.08	0.06	0.02	36.17	0.84	25.70	0.91	0.04	0.06	99.32
TL22	1	Ol	36.96	0.03	0.05	0.00	31.27	0.68	30.87	0.69	0.03	0.02	100.61
TL22	1	Ol	34.33	0.11	0.06	0.00	43.47	1.30	19.37	0.97	0.04	0.02	99.66
TL22	1	Ol	35.96	0.06	0.20	0.00	36.55	0.87	25.88	0.86	0.11	0.06	100.54
TL22	1	Ol	36.40	0.09	0.13	0.04	35.34	0.81	26.89	0.80	0.03	0.04	100.56

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TL22	1	Ol	34.53	0.17	0.05	0.00	43.36	1.14	20.32	0.79	0.07	0.05	100.50
TL22	1	Ox	0.09	12.12	7.59	0.08	72.19	0.45	1.49	0.10	0.00	0.13	94.22
TL22	1	Ox	0.14	15.57	8.77	0.20	67.00	0.79	1.00	0.17	0.01	0.01	93.65
TL22	1	Ox	0.70	16.86	8.93	0.04	57.60	0.64	2.18	5.15	0.08	0.07	92.24
TL22	1	Ox	1.09	20.43	5.42	0.17	65.30	0.85	0.35	0.48	0.31	0.07	94.48
TL22	1	Pl	48.78	1.42	4.69	0.00	9.20	0.22	12.69	22.97	0.32	0.01	100.30
TL22	1	Ap	0.39	0.01	0.03	0.02	0.27	0.00	0.22	54.57	0.01	0.03	55.54
TL22	1	Pl	43.01	3.03	9.25	0.08	10.74	0.17	9.91	23.16	0.49	0.04	99.87
TL22	1	Pl	48.01	1.40	4.72	0.00	9.24	0.20	12.29	23.03	0.35	0.06	99.28
TL22	1	Pl	44.32	2.60	8.66	0.03	9.59	0.16	10.65	22.91	0.43	0.03	99.39
TL22	1	Pl	47.80	1.69	5.44	0.00	9.44	0.17	12.56	22.40	0.32	0.02	99.86
TL22	1	Pl	47.71	1.83	5.73	0.00	10.12	0.20	11.85	22.86	0.32	0.02	100.66
TL22	1	Pl	42.94	3.67	10.45	0.02	9.88	0.14	10.12	22.82	0.38	0.03	100.46
TL22	1	Pl	45.15	2.62	9.03	0.04	10.30	0.12	9.97	22.20	0.66	0.03	100.12
TL22	1	Pl	44.98	2.12	8.32	0.00	10.74	0.20	10.26	22.63	0.48	0.15	99.89
TL22	1	Pl	41.48	3.16	10.37	0.00	11.46	0.21	9.49	21.97	0.36	0.06	98.55
TL22	1	Pl	46.08	2.04	8.05	0.05	9.78	0.28	10.28	22.50	0.56	0.87	100.48
TL22	1	Lct	55.85	0.10	23.80	0.08	0.33	0.02	0.00	0.01	0.81	20.80	101.79
TL22	1	Incl.	38.55	3.74	8.01	0.03	12.75	0.33	6.44	24.83	0.94	0.93	96.56
TL22	2	Ap	0.41	0.01	0.02	0.00	0.40	0.00	0.31	54.00	0.01	0.01	55.17
TL22	2	Ap	0.46	0.00	0.03	0.00	0.44	0.02	0.21	54.02	0.02	0.03	55.22
TL22	2	Ol	37.73	0.06	0.02	0.00	25.85	0.48	35.61	0.59	0.00	0.01	100.35
TL22	2	Ox	0.20	16.56	7.91	0.07	67.98	0.84	0.39	0.32	0.00	0.02	94.28
TL22	2	Ol	35.34	0.12	0.22	0.02	37.36	0.92	25.19	1.05	0.07	0.10	100.39
TL22	2	Ol	30.89	0.09	0.06	0.00	39.25	0.99	16.11	7.56	0.03	0.03	95.00
TL22	2	Ol	35.68	0.10	0.07	0.03	34.94	0.87	27.59	0.76	0.06	0.04	100.13
TL22	2	Ol	37.36	0.06	0.03	0.04	29.04	0.63	32.70	0.67	0.03	0.04	100.60
TL22	2	Ol	36.42	0.05	0.08	0.00	28.56	0.63	32.86	0.68	0.06	0.03	99.37
TL22	2	Ol	35.42	0.10	0.06	0.01	41.22	1.15	21.91	0.92	0.03	0.06	100.86
TL22	2	Cpx	49.81	1.23	4.74	0.02	7.72	0.17	13.20	23.39	0.29	0.01	100.58
TL22	2	Cpx	49.54	1.32	4.45	0.04	9.61	0.23	11.80	22.46	0.32	0.01	99.78
TL22	2	Cpx	47.72	1.80	5.77	0.00	10.00	0.24	11.75	22.18	0.35	0.03	99.84
TL22	2	Cpx	48.12	1.47	4.99	0.03	9.35	0.27	12.04	23.20	0.31	0.05	99.83
TL22	2	Cpx	39.64	2.69	10.38	0.02	14.39	0.15	9.10	21.82	0.50	0.10	98.79
TL22	2	Ox	0.16	17.90	6.12	0.06	66.28	0.78	0.42	0.23	0.01	0.11	92.07
TL22	2	Cpx	44.24	2.67	8.92	0.00	9.73	0.21	10.43	22.83	0.37	0.02	99.41
TL22	2	Lct	55.15	0.16	24.33	0.07	0.65	0.02	0.00	0.05	0.43	21.28	102.13
TL22	2	Cpx	44.46	2.96	8.80	0.01	9.81	0.20	10.55	23.01	0.37	0.03	100.20
TL22	2	Cpx	46.18	1.93	6.71	0.00	9.81	0.26	10.62	23.32	0.48	0.05	99.38
TL22	2	Cpx	47.60	1.68	5.32	0.04	9.25	0.19	11.84	22.76	0.29	0.06	99.00
TL22	2	Cpx	46.29	1.95	7.77	0.07	9.42	0.19	10.85	23.01	0.43	0.02	100.01
TL22	2	Cpx	43.27	2.01	7.53	0.00	10.00	0.26	9.78	24.00	0.48	0.06	97.40
TL22	2	Cpx	44.29	2.90	8.94	0.00	9.91	0.17	10.11	22.57	0.40	0.05	99.34
TL22	2	Pl	50.68	0.11	30.50	0.00	0.74	0.00	0.08	13.71	3.51	0.32	99.65

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TL22	2	Pl	52.53	0.07	29.56	0.01	0.71	0.04	0.07	12.46	4.03	0.50	99.99
TL22	2	Pl	49.44	0.06	32.01	0.00	0.87	0.04	0.08	15.24	2.42	0.42	100.57
TL22	2	Pl	48.52	0.03	32.23	0.00	0.88	0.02	0.12	16.03	2.31	0.26	100.39
TL22	2	Pl	51.69	0.08	29.93	0.00	0.86	0.00	0.06	12.98	3.87	0.35	99.81
TL22	2	Pl	50.27	0.07	30.62	0.03	1.03	0.03	0.07	13.90	3.49	0.41	99.92
TL22	2	Pl	49.18	0.07	31.98	0.02	0.81	0.00	0.07	15.73	2.33	0.47	100.65
TL22	2	Pl	48.01	0.05	32.72	0.00	0.75	0.05	0.15	16.33	1.98	0.40	100.44
TL22	2	Pl	49.01	0.06	31.65	0.02	0.86	0.00	0.10	15.00	2.57	0.57	99.84
TL22	2	Pl	48.23	0.03	32.34	0.01	0.67	0.07	0.13	15.89	2.15	0.40	99.92
TL22	2	Pl	51.49	0.09	30.03	0.02	0.80	0.00	0.06	13.31	3.54	0.62	99.97
TL22	2	Pl	53.26	0.14	28.14	0.03	1.13	0.00	0.19	11.19	4.08	1.33	99.49
TL22	2	Pl	53.40	0.13	28.82	0.05	0.80	0.00	0.07	11.93	4.25	0.65	100.10
TL22	2	Pl	55.61	0.13	23.58	0.00	0.60	0.00	0.00	0.06	0.21	21.30	101.49
TL22	2	Pl	52.62	0.07	29.22	0.00	0.59	0.00	0.02	11.81	4.37	0.63	99.33
TL22	2	Pl	52.62	0.08	29.90	0.00	0.74	0.02	0.07	12.53	3.87	0.65	100.49
TL22	2	Incl.	53.55	0.10	26.90	0.01	1.85	0.00	0.35	0.82	12.94	0.70	97.23
TL22	2	Lct	55.63	0.08	23.69	0.01	0.36	0.00	0.00	0.00	0.87	20.57	101.21
TL22	3	Ap	0.29	0.01	0.00	0.00	0.24	0.07	0.24	53.96	0.03	0.01	54.86
TL22	3	Ap	0.41	0.00	0.01	0.00	0.23	0.03	0.29	53.66	0.04	0.01	54.67
TL22	3	Ol	36.62	0.05	0.10	0.00	29.27	0.70	31.69	1.19	0.05	0.00	99.67
TL22	3	Ol	34.74	0.09	0.17	0.08	40.24	1.06	22.70	1.38	0.09	0.08	100.63
TL22	3	Ol	35.01	0.04	0.05	0.02	37.25	0.89	24.80	1.99	0.02	0.02	100.08
TL22	3	Ol	35.50	0.06	0.07	0.00	38.52	0.93	24.56	0.76	0.03	0.03	100.45
TL22	3	Ol	33.64	0.06	0.06	0.00	37.09	0.96	23.50	2.85	0.06	0.06	98.30
TL22	3	Ol	35.47	0.11	0.06	0.09	38.07	0.86	24.89	0.86	0.06	0.02	100.50
TL22	3	Ol	37.22	0.05	0.04	0.00	29.34	0.65	32.09	0.74	0.04	0.05	100.22
TL22	3	Ol	34.33	0.09	0.18	0.02	42.80	1.19	20.58	0.83	0.05	0.28	100.35
TL22	3	Incl.	60.43	0.12	24.82	0.00	0.57	0.01	0.00	0.16	9.86	4.36	100.33
TL22	3	Lct	56.23	0.09	23.57	0.00	0.43	0.00	0.04	0.00	0.83	20.55	101.74
TL22	3	Ox	0.12	15.94	8.53	0.04	68.56	0.92	0.20	0.23	0.01	0.06	94.61
TL22	3	Ox	0.13	14.80	7.65	0.11	70.46	0.62	1.46	0.06	0.02	0.09	95.39
TL22	3	Cpx	49.30	1.27	5.04	0.05	7.84	0.19	12.91	23.39	0.32	0.00	100.32
TL22	3	Cpx	45.10	2.29	8.84	0.01	9.06	0.11	11.37	23.22	0.34	0.06	100.40
TL22	3	Cpx	44.73	2.83	8.82	0.00	9.38	0.17	10.47	22.97	0.40	0.04	99.80
TL22	3	Cpx	48.24	1.57	5.27	0.00	9.01	0.20	12.56	22.70	0.30	0.01	99.86
TL22	3	Cpx	41.68	3.40	10.21	0.03	11.17	0.20	9.21	22.57	0.51	0.02	98.98
TL22	3	Cpx	48.29	1.52	5.25	0.04	9.44	0.19	12.62	22.73	0.31	0.05	100.43
TL22	3	Cpx	43.39	2.80	8.73	0.03	11.02	0.21	9.94	22.21	0.47	0.11	98.89
TL22	3	Cpx	48.75	1.35	4.98	0.10	8.32	0.25	12.93	23.32	0.33	0.03	100.38
TL22	3	Cpx	44.71	2.65	8.44	0.00	9.86	0.22	10.60	22.87	0.43	0.03	99.82
TL22	3	Cpx	45.80	2.14	7.48	0.05	9.40	0.17	10.58	23.66	0.34	0.02	99.65
TL22	3	Cpx	42.41	3.66	9.75	0.00	11.10	0.22	9.06	22.45	0.50	0.05	99.20
TL22	3	Cpx	46.44	2.08	6.91	0.05	9.38	0.20	11.36	22.87	0.36	0.09	99.74
TL22	3	Pl	48.39	0.07	31.77	0.00	0.84	0.04	0.28	16.19	2.35	0.56	100.49

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TL22	3	Pl	52.34	0.09	30.05	0.03	0.71	0.00	0.07	12.44	3.95	0.45	100.13
TL22	3	Pl	50.63	0.10	30.43	0.02	1.18	0.03	0.16	13.56	3.28	0.65	100.04
TL22	3	Pl	48.14	0.05	32.57	0.00	0.86	0.01	0.11	16.02	2.10	0.46	100.31
TL22	3	Pl	50.85	0.07	31.18	0.00	0.72	0.00	0.10	14.74	2.78	0.58	101.02
TL22	3	Pl	50.33	0.11	30.77	0.00	0.76	0.00	0.05	14.35	2.94	0.63	99.94
TL22	3	Pl	50.19	0.08	31.36	0.01	0.86	0.02	0.06	14.76	2.78	0.59	100.70
TL22	3	Pl	51.32	0.09	30.94	0.00	0.80	0.02	0.07	13.48	3.47	0.62	100.82
TL22	3	Pl	53.11	0.14	28.32	0.05	1.10	0.02	0.11	11.44	4.16	1.10	99.56
TL22	3	Pl	48.34	0.07	32.09	0.03	0.98	0.04	0.09	15.36	2.50	0.36	99.86
TL22	3	Pl	51.34	0.05	30.07	0.02	0.75	0.03	0.10	13.00	3.73	0.40	99.50
TL22	3	Pl	52.94	0.13	28.33	0.03	1.31	0.02	0.17	11.25	4.65	0.74	99.56
TL22	3	Pl	50.85	0.12	30.24	0.00	1.00	0.04	0.09	13.64	3.48	0.48	99.94
TL22	3	Pl	49.52	0.07	31.75	0.00	0.85	0.02	0.03	15.23	2.59	0.40	100.46
TL22	3	Ol	34.71	0.10	0.03	0.00	43.38	1.24	20.05	0.93	0.08	0.07	100.59
TL22	3	Ol	37.60	0.06	0.05	0.05	28.77	0.61	32.61	0.70	0.04	0.05	100.54
TL22	3	Cpx	44.41	2.87	9.19	0.00	9.61	0.12	10.47	23.18	0.37	0.09	100.32
TL22	4	Ol	38.51	0.02	0.02	0.00	25.71	0.58	35.85	0.63	0.02	0.00	101.34
TL22	4	Ol	32.99	0.05	0.05	0.00	33.81	0.80	21.88	5.29	0.10	0.16	95.14
TL22	4	Ol	37.63	0.01	0.05	0.01	25.70	0.53	35.48	0.59	0.00	0.01	100.02
TL22	4	Ol	38.16	0.02	0.08	0.00	26.26	0.51	35.11	0.57	0.02	0.04	100.76
TL22	4	Ol	37.81	0.02	0.05	0.00	26.99	0.63	33.89	0.62	0.03	0.00	100.04
TL22	4	Incl.	46.88	1.88	7.27	0.06	9.45	0.20	11.80	23.09	0.34	0.01	101.00
TL22	4	Ol	33.94	0.06	0.05	0.00	45.76	1.20	17.16	0.93	0.05	0.11	99.25
TL22	4	Emb.	48.78	1.42	5.83	0.03	8.56	0.22	13.01	23.34	0.31	0.02	101.52
TL22	4	Ol	38.08	0.05	0.04	0.00	25.68	0.46	35.52	0.63	0.06	0.02	100.54
TL22	4	Cpx	49.17	1.25	5.43	0.02	7.96	0.16	13.07	23.62	0.30	0.02	101.00
TL22	4	Incl.	53.20	0.53	23.26	0.04	4.31	0.02	1.74	1.99	9.60	1.56	96.25
TL22	4	Lct	55.65	0.12	23.13	0.00	0.28	0.03	0.02	0.00	0.82	20.89	100.95
TL22	4	Cpx	47.99	1.43	5.79	0.00	8.15	0.18	12.64	23.44	0.34	0.02	99.98
TL22	4	Pl	48.30	0.05	32.16	0.00	0.60	0.00	0.10	15.75	2.25	0.43	99.65
TL22	4	Pl	48.06	0.06	32.06	0.05	0.57	0.00	0.08	16.17	2.18	0.59	99.82
TL22	4	Pl	48.57	0.07	31.90	0.08	0.59	0.01	0.09	15.90	2.25	0.47	99.93
TL22	4	Pl	48.54	0.05	31.86	0.00	0.69	0.00	0.08	15.74	2.34	0.41	99.71
TL22	4	Cpx	49.01	1.29	5.31	0.08	7.84	0.12	13.29	23.45	0.30	0.01	100.69
TL22	4	Pl	48.10	0.08	32.36	0.00	0.69	0.01	0.05	15.73	2.20	0.47	99.68
TL22	4	Ox	0.23	13.85	8.66	0.17	70.93	0.69	0.78	0.26	0.04	0.05	95.65
TL22	4	Cpx	45.29	2.36	8.56	0.08	8.88	0.19	11.30	23.33	0.32	0.00	100.31
TL22	4	Cpx	44.08	2.70	9.55	0.07	9.41	0.14	10.64	22.70	0.36	0.02	99.68
TL22	4	Cpx	46.24	1.90	7.79	0.00	8.43	0.13	11.77	23.43	0.30	0.00	100.00
TL22	4	Pl	48.54	0.04	32.58	0.00	0.54	0.00	0.05	15.81	2.23	0.67	100.44
TL22	4	Pl	45.49	0.07	29.63	0.07	0.62	0.01	0.07	14.65	2.12	0.50	93.23
TL22	4	Cpx	45.22	1.97	8.06	0.05	8.72	0.17	11.35	23.35	0.33	0.01	99.23
TL22	4	Ox	0.17	14.99	8.70	0.00	67.01	0.72	1.50	0.43	0.05	0.03	93.61
TL22	4	Pl	46.92	0.04	32.18	0.02	0.73	0.00	0.09	16.57	1.88	0.34	98.77

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TL22	4	Cpx	46.06	1.72	7.10	0.04	8.33	0.15	12.10	23.39	0.35	0.02	99.25
TL22	4	Cpx	48.80	1.42	5.68	0.05	7.81	0.17	12.84	23.39	0.36	0.00	100.50
TL19	1	Pl	48.55	0.05	31.80	0.00	0.62	0.01	0.08	16.01	2.08	0.48	99.68
TL19	1	Ap	0.37	0.00	0.00	0.00	0.37	0.05	0.27	54.61	0.01	0.00	55.66
TL19	1	Ap	0.35	0.00	0.00	0.00	0.29	0.00	0.25	52.13	0.00	0.01	53.04
TL19	1	Lct	55.88	0.09	23.56	0.06	0.36	0.00	0.00	0.00	0.81	20.11	100.88
TL19	1	Pl	48.70	0.06	32.40	0.02	0.90	0.03	0.10	15.99	2.16	0.47	100.83
TL19	1	Pl	55.74	0.11	23.67	0.00	0.38	0.00	0.01	0.00	0.84	20.22	100.97
TL19	1	Pl	48.73	0.06	32.19	0.00	0.81	0.00	0.08	15.80	2.25	0.48	100.40
TL19	1	Pl	48.90	0.07	31.97	0.00	0.87	0.00	0.11	15.67	2.27	0.48	100.35
TL19	1	Pl	48.89	0.09	32.16	0.03	0.97	0.00	0.12	16.04	2.18	0.46	100.96
TL19	1	Pl	49.10	0.07	32.48	0.00	0.63	0.03	0.06	15.62	2.14	0.43	100.57
TL19	1	Pl	55.30	0.16	23.79	0.03	0.65	0.05	0.00	0.03	0.33	20.87	101.21
TL19	1	Pl	54.88	0.18	23.74	0.00	0.93	0.03	0.02	0.08	0.61	20.44	100.93
TL19	1	Pl	51.94	0.23	28.68	0.01	1.25	0.05	0.23	12.94	3.39	1.18	99.89
TL19	1	Pl	48.65	0.05	32.39	0.00	0.62	0.04	0.08	16.27	2.11	0.45	100.67
TL19	1	Pl	48.61	0.06	32.34	0.05	0.69	0.00	0.08	15.92	2.13	0.44	100.31
TL19	1	Pl	55.87	0.09	24.54	0.00	0.36	0.01	0.03	0.02	0.42	20.59	101.93
TL19	1	Pl	48.33	0.06	32.42	0.01	0.71	0.02	0.06	15.76	2.13	0.45	99.94
TL19	1	Lct	55.37	0.09	23.64	0.00	0.40	0.03	0.00	0.00	0.69	20.45	100.67
TL19	1	Ol	38.96	0.01	0.04	0.02	20.70	0.39	39.58	0.44	0.03	0.00	100.17
TL19	1	Ol	38.35	0.00	0.07	0.03	20.18	0.45	39.24	0.50	0.04	0.02	98.89
TL19	1	Cpx	46.36	2.02	7.98	0.04	8.59	0.11	11.85	23.13	0.34	0.02	100.44
TL19	1	Lct	55.87	0.09	23.91	0.04	0.35	0.00	0.00	0.00	0.87	20.12	101.25
TL19	1	Cpx	45.76	2.30	8.67	0.00	8.79	0.14	11.36	23.41	0.31	0.02	100.77
TL19	1	Pl	55.55	0.11	23.67	0.02	0.43	0.03	0.04	0.00	0.35	21.10	101.31
TL19	2	Pl	48.86	0.07	32.40	0.00	0.83	0.00	0.08	15.99	2.10	0.47	100.81
TL19	2	Pl	49.02	0.06	32.18	0.00	0.76	0.00	0.11	15.88	2.23	0.48	100.73
TL19	2	Pl	48.56	0.09	32.03	0.00	0.84	0.01	0.06	15.92	2.06	0.49	100.05
TL19	2	Pl	45.52	0.09	25.23	0.04	1.07	0.05	0.21	18.90	2.93	0.97	95.00
TL19	2	Ap	0.42	0.02	0.00	0.00	0.56	0.02	0.29	54.05	0.12	0.08	55.56
TL19	2	Cpx	46.15	2.04	7.99	0.03	8.68	0.12	11.77	23.00	0.29	0.02	100.09
TL19	2	Cpx	46.32	1.96	7.97	0.04	8.74	0.13	11.67	23.41	0.30	0.04	100.59
TL19	2	Ap	0.27	0.02	0.00	0.00	0.36	0.00	0.28	54.56	0.02	0.00	55.51
TL19	2	Ol	39.12	0.02	0.04	0.00	20.69	0.46	39.86	0.58	0.02	0.00	100.79
TL19	2	Ol	38.30	0.05	0.03	0.01	25.22	0.48	35.80	0.64	0.00	0.02	100.55
TL19	2	Cpx	49.34	1.29	5.54	0.05	7.59	0.11	12.96	23.49	0.34	0.01	100.72
TL19	2	Ol	39.18	0.01	0.03	0.00	21.59	0.38	39.71	0.47	0.01	0.00	101.38
TL19	3	Pl	48.77	0.07	32.32	0.04	0.94	0.00	0.11	16.00	2.21	0.49	100.94
TL19	3	Pl	49.17	0.08	32.04	0.00	0.94	0.04	0.07	15.66	2.23	0.50	100.73
TL19	3	Pl	48.78	0.06	32.37	0.00	0.77	0.06	0.11	15.80	2.33	0.46	100.75
TL19	3	Pl	49.48	0.06	32.09	0.00	0.97	0.00	0.08	15.66	2.31	0.52	101.16
TL19	3	Pl	56.12	0.12	23.61	0.00	0.74	0.03	0.00	0.07	0.82	20.34	101.84
TL19	3	Pl	48.96	0.06	32.04	0.03	0.79	0.03	0.09	15.68	2.27	0.48	100.43

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TL19	3	Pl	55.83	0.12	23.45	0.03	0.60	0.00	0.03	0.00	0.89	20.14	101.08
TL19	3	Cpx	49.58	1.26	5.45	0.00	7.88	0.20	13.13	23.31	0.31	0.04	101.17
TL19	3	Cpx	46.66	1.82	7.51	0.03	8.53	0.12	11.97	23.24	0.32	0.03	100.22
TL19	3	Ap	0.43	0.02	0.00	0.00	0.39	0.02	0.26	54.49	0.03	0.02	55.66
TL19	3	Cpx	45.75	2.18	8.40	0.00	8.58	0.17	11.44	23.31	0.33	0.04	100.20
TL19	3	Lct	56.01	0.10	24.71	0.06	0.42	0.04	0.00	0.00	0.73	19.90	101.97
TL19	4	Ox	0.21	10.82	9.04	0.06	68.37	0.33	5.30	0.23	0.02	0.06	94.43
TL19	4	Ap	0.50	0.08	0.13	0.00	0.96	0.03	0.30	54.03	0.04	0.04	56.11
TL19	4	Ap	0.40	0.00	0.01	0.04	0.41	0.02	0.24	54.82	0.04	0.04	56.02
TL19	4	Ol	38.31	0.04	0.03	0.01	25.49	0.53	35.26	0.73	0.01	0.01	100.43
TL19	4	Ol	38.38	0.02	0.04	0.02	25.47	0.56	35.80	0.62	0.02	0.01	100.94
TL19	4	Cpx	48.65	1.43	5.92	0.12	8.13	0.18	12.51	23.41	0.35	0.04	100.75
TL19	4	Cpx	48.96	1.31	5.34	0.06	7.96	0.13	12.93	23.41	0.29	0.01	100.41
TL19	4	Cpx	46.98	1.78	7.31	0.09	8.47	0.20	11.89	23.63	0.35	0.03	100.71
TL19	4	Cpx	45.11	2.31	8.91	0.00	9.03	0.18	11.14	23.61	0.31	0.01	100.63
TL19	4	Cpx	48.62	1.49	4.98	0.00	9.82	0.22	12.20	22.66	0.37	0.04	100.38
TL19	4	Pl	49.86	0.05	31.86	0.06	0.90	0.06	0.09	15.25	2.47	0.59	101.19
TL19	4	Lct	55.65	0.09	23.75	0.00	0.39	0.02	0.02	0.01	0.82	20.14	100.89
TL19	4	Pl	49.17	0.08	32.30	0.00	0.89	0.00	0.14	15.82	2.18	0.52	101.09
TL19	4	Pl	49.57	0.05	31.64	0.01	0.75	0.02	0.10	15.52	2.34	0.49	100.49
TL19	4	Pl	48.98	0.05	31.94	0.06	0.86	0.02	0.07	15.98	2.28	0.40	100.64
TL19	5	Cpx	45.33	2.28	8.83	0.04	8.77	0.09	11.30	23.02	0.31	0.01	99.99
TL19	5	Ap	0.50	0.01	0.01	0.03	0.49	0.04	0.30	54.41	0.00	0.02	55.82
TL19	5	Cpx	46.95	1.74	7.34	0.10	8.43	0.09	11.91	23.34	0.40	0.10	100.40
TL19	5	Ol	38.34	0.03	0.05	0.02	25.94	0.55	35.69	0.74	0.04	0.01	101.42
TL19	5	Cpx	46.24	1.73	7.99	0.00	8.74	0.20	11.71	23.16	0.36	0.01	100.15
TL19	5	Lct	55.33	0.11	23.43	0.01	0.55	0.03	0.06	0.13	0.08	21.43	101.16
TL19	5	Cpx	46.00	2.09	7.96	0.01	8.85	0.11	11.51	23.17	0.38	0.02	100.11
TL19	5	Cpx	49.25	1.21	4.99	0.06	7.54	0.16	13.36	23.05	0.37	0.18	100.17
TL19	5	Cpx	46.56	2.01	7.85	0.00	8.54	0.14	11.76	23.36	0.33	0.03	100.57
TL19	5	Cpx	46.07	2.18	8.43	0.03	8.93	0.15	11.49	23.12	0.35	0.01	100.76
TL19	5	Ol	38.22	0.03	0.03	0.00	26.22	0.52	35.80	0.68	0.02	0.00	101.54
TL19	5	Ox	0.18	10.69	8.83	0.13	68.51	0.46	4.86	0.57	0.07	0.02	94.30
TL19	5	Pl	48.84	0.06	32.26	0.00	0.68	0.03	0.11	16.14	2.28	0.46	100.87
TL19	5	Pl	48.78	0.06	32.18	0.02	0.63	0.02	0.10	15.37	2.24	0.47	99.88
TL19	5	Pl	48.94	0.05	32.18	0.02	0.78	0.00	0.11	15.84	2.19	0.47	100.59
TL19	5	Pl	49.22	0.06	32.24	0.07	0.70	0.05	0.08	15.73	2.30	0.48	100.93
TL19	5	Pl	48.73	0.06	32.35	0.00	0.78	0.03	0.10	15.75	2.14	0.45	100.40
TL19	5	Pl	48.63	0.07	32.66	0.01	0.86	0.00	0.09	16.24	2.03	0.44	101.03
TL19	5	Pl	48.79	0.06	32.17	0.00	0.73	0.03	0.06	15.60	2.22	0.46	100.12
TL19	5	Lct	55.65	0.12	24.52	0.00	0.52	0.03	0.04	0.15	0.51	20.67	102.20
TL19	5	Ap	0.59	0.01	0.04	0.00	0.55	0.00	0.24	54.21	0.02	0.14	55.7934
TL19	5	Pl	54.52	0.20	27.15	0.01	1.72	0.03	0.20	11.10	4.46	1.07	100.4634
TL19	5	Lct	55.81	0.12	24.55	0.06	0.51	0.04	0.02	0.06	0.71	20.12	102.0011

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TL19	5	Lct	55.65	0.05	23.73	0.01	0.36	0.00	0.02	0.02	0.99	19.77	100.6175
TL19	5	Lct	55.82	0.08	24.02	0.12	0.32	0.00	0.02	0.00	1.02	20.01	101.4066
TL19	5	Lct	55.83	0.10	24.06	0.00	0.37	0.00	0.01	0.01	0.90	20.18	101.4506
TL19	5	Lct	55.62	0.10	23.73	0.07	0.29	0.02	0.00	0.01	1.02	19.97	100.8146
TL19	5	Pl	48.69	0.08	32.62	0.00	0.74	0.00	0.07	16.08	2.19	0.46	100.925
TL19	5	Pl	48.88	0.07	32.65	0.00	0.61	0.00	0.08	15.77	2.20	0.46	100.7368
TL19	5	Pl	49.16	0.05	32.27	0.00	0.82	0.00	0.07	16.33	2.20	0.45	101.3434

Appendix 2. Volume of phenocrysts for each group of phases, volume of the voids and volume of analysis, all in mm³.

Samples	Width (px)	Height (px)	Nb slices	V _{tot}	V _{ph} Pl + Lct	V _{ph} Cpx + Ol	V _{tot} phenocrysts	V _{void}	V _a
TL03	1950	1950	1800	3.92	1.54	0.15	1.69	0.13	2.10
TL07	1950	1950	1800	3.92	1.19	0.49	1.69	0.06	2.18
TL11	1950	1950	1800	3.92	1.46	0.07	1.53	0.41	1.98
TL12	1800	1800	1800	3.34	2.24	0.09	2.33	0.21	0.80
TL13	1950	1950	1800	3.92	1.07	0.52	1.59	0.36	1.97
TL14	1950	1950	1800	3.92	2.31	0.22	2.53	0.28	1.11
TL16	1950	1950	1800	3.92	1.55	0.08	1.63	0.78	1.51
TL19	1950	1950	1800	3.92	1.32	0.06	1.38	0.75	1.78
TL21	1950	1950	1800	3.92	1.03	0.22	1.24	0.28	2.39
TL22	1950	1950	1800	3.92	1.62	0.20	1.82	0.26	1.84

Appendix 3. Measured and calculated volumes, expressed in mm³.

Samples	V _a	V _m Pl + Lct	V _m Cpx + Ol	V _m Ox	V% Pl + Lct	V% Cpx + Ol	V% Ox	Total
TL03	2.10	0.40	1.25	0.02	19.13	59.65	1.09	79.87
TL07	2.18	0.45	1.43	0.03	20.63	65.68	1.59	87.90
TL11	1.98	0.23	1.53	0.01	11.72	77.10	0.75	89.57
TL12	0.80	0.24	0.42	0.03	30.10	52.31	3.64	86.05
TL13	1.97	0.01	1.66	0.01	0.66	84.54	0.68	85.89
TL14	1.11	0.21	0.70	0.03	19.11	63.44	3.06	85.62
TL16	1.51	0.33	0.92	0.04	22.13	60.96	2.89	85.98
TL19	1.78	0.20	1.42	0.01	11.04	79.97	0.36	91.38
TL21	2.39	0.88	1.04	0.12	36.63	43.64	4.84	85.12
TL22	1.84	0.65	0.85	0.05	35.24	46.30	2.65	84.18

Appendix 4. Crystal count of plagioclase and leucite microlites per volume. Separation parameter set to 2 (1 for TL19).

Volume (μm^3)	TL03	TL07	TL11	TL12	TL13	TL14	TL16	TL19*	TL21	TL22
200	4264	6029	7346	14489	5401	12252	7730	5170	4266	6557
300	3909	5456	6655	12552	4676	11222	7289	4741	3995	6154
400	3693	5087	6181	11331	4171	10551	6993	4418	3849	5916
500	3544	4806	5834	10462	3799	10051	6778	4175	3749	5762
600	3442	4585	5557	9831	3516	9669	6621	3976	3676	5650
700	3363	4405	5324	9364	3291	9388	6499	3797	3624	5551
800	3299	4267	5125	8966	3102	9129	6391	3636	3582	5477
900	3250	4150	4969	8657	2939	8942	6300	3490	3548	5416
1000	3205	4041	4825	8388	2802	8775	6223	3373	3520	5364
2000	2962	3322	3922	6935	2033	7675	5728	2540	3372	5075
3000	2811	2911	3429	6198	1646	7036	5401	2056	3273	4889
4000	2695	2624	3049	5757	1403	6548	5115	1696	3190	4726
5000	2588	2389	2743	5424	1217	6133	4867	1432	3113	4575
6000	2497	2186	2483	5147	1068	5722	4616	1219	3033	4429
7000	2415	2028	2257	4937	959	5378	4397	1046	2962	4299
8000	2328	1891	2056	4756	858	5047	4188	912	2882	4177
9000	2248	1758	1877	4580	775	4734	3992	799	2803	4044
10000	2178	1645	1722	4426	706	4474	3816	705	2732	3904
20000	1632	873	795	3431	298	2610	2379	226	2111	2867
30000	1257	460	381	2786	126	1647	1528	93	1632	2122
40000	975	191	170	2308	49	1000	974	38	1269	1597
50000	760		60	1917		604	613	18	998	1215
60000	604			1607		369	381	9	786	921
70000	479			1324		185	230	7	615	693
80000	379			1092		74	127	4	479	524
90000	298			917			51	3	382	389
100000	241			777				1	302	283

Appendix 5. Crystal count of clinopyroxene and olivine microlites per volume. Separation parameter set to 2.

Volume (μm^3)	TL03	TL07	TL11	TL12	TL13	TL14	TL16	TL19	TL21	TL22
200	13005	13381	10976	30479	6315	25547	19904	10119	11463	22186
300	12870	13219	10875	29737	5589	24985	19567	10022	11327	21778
400	12778	13108	10787	29217	5078	24492	19315	9942	11225	21422
500	12700	13013	10713	28786	4691	24051	19059	9872	11129	21108
600	12625	12925	10646	28435	4389	23648	18810	9804	11042	20854
700	12555	12829	10573	28132	4145	23272	18601	9732	10955	20592
800	12486	12753	10506	27811	3931	22936	18394	9677	10872	20332
900	12415	12669	10442	27545	3745	22574	18186	9614	10784	20086
1000	12353	12589	10376	27280	3590	22264	17988	9556	10693	19832
2000	11729	11781	9814	24939	2629	19639	16246	9037	9722	17385
3000	11087	11075	9313	22957	2103	17668	14822	8624	8812	15162
4000	10457	10407	8882	21167	1763	16040	13638	8270	8025	13302
5000	9849	9801	8483	19643	1489	14673	12612	7986	7349	11708
6000	9268	9245	8122	18269	1283	13537	11712	7710	6762	10443
7000	8738	8733	7787	17053	1111	12512	10942	7457	6239	9371
8000	8242	8280	7464	15924	970	11625	10224	7217	5794	8442
9000	7779	7825	7169	14980	847	10856	9569	7004	5388	7659
10000	7345	7417	6887	14066	752	10181	8986	6801	5045	6994
20000	4250	4523	4790	8464	249	5907	5297	5182	2686	3314
30000	2599	2905	3470	5616	90	3877	3392	4062	1525	1782
40000	1635	1914	2592	3865	30	2738	2268	3225	889	1014
50000	1047	1304	1982	2795		2031	1591	2593	514	561
60000	678	919	1524	2093		1540	1124	2125	309	302
70000	445	654	1187	1542		1178	801	1731	182	158
80000	307	487	926	1207		908	591	1429	90	83
90000	206	359	736	928		681	428	1191	35	32
100000	139	263	585	721		506	302	995		
200000								199		
300000								37		

Appendix 6. Crystal count of oxide microlites per volume. Separation parameter set to 4.

Volume (μm^3)	TL03	TL07	TL11	TL12	TL13	TL14	TL16	TL19	TL21	TL22
100	1646	1646	527	9184	642	1797	1791	957	12518	11303
200	1496	1437	421	7195	532	1472	1497	762	10861	8691
300	1386	1292	356	6078	470	1272	1289	645	9599	7027
400	1299	1186	317	5258	415	1146	1141	556	8587	5926
500	1221	1099	286	4630	373	1044	1031	492	7767	5116
600	1154	1026	260	4220	340	964	946	431	7041	4490
700	1094	967	240	3795	317	903	873	388	6415	4002
800	1036	917	222	3509	289	861	819	351	5879	3602
900	989	875	207	3227	269	824	771	319	5390	3280
1000	942	829	195	3007	252	789	729	294	4962	3000
2000	636	572	124	1863	156	542	482	158	2484	1587
3000	467	437	91	1409	116	412	382	96	1443	967
4000	351	360	72	1121	91	345	332	64	895	620
5000	272	301	63	952	71	301	292	44	563	403
6000	220	260	55	827	58	266	261	29	361	260
7000	176	228	50	738	48	231	236	19	221	159
8000	143	202	45	643	39	210	214	13	123	86
9000	113	180	41	581	34	184	198	7	54	34
10000	90	160	38	551	30	167	185			
20000	14	60	21	255	10	72	65			
30000		28	11	146		28				
40000		11	5	62						

Appendix 7. Weight corrected and normalized proportion of phases calculated using MELTS from 1300 °C to 960°C at HtMt conditions.

T°C	liquid	olivine	clinopyroxene	feldspar	leucite	spinel	apatite	fluid
1300	94.48	0.00	0.00	0.00	0.00	4.94	0.00	0.58
1290	93.80	0.00	0.00	0.00	0.00	5.11	0.00	1.09
1280	93.62	0.00	0.00	0.00	0.00	5.30	0.00	1.09
1270	93.44	0.00	0.00	0.00	0.00	5.47	0.00	1.09
1260	93.27	0.00	0.00	0.00	0.00	5.64	0.00	1.09
1250	93.11	0.00	0.00	0.00	0.00	5.80	0.00	1.09
1240	92.95	0.00	0.00	0.00	0.00	5.95	0.00	1.09
1230	92.81	0.00	0.00	0.00	0.00	6.09	0.00	1.10
1220	92.61	0.00	0.00	0.00	0.00	6.28	0.00	1.11
1210	92.43	0.00	0.00	0.00	0.00	6.45	0.00	1.12
1200	92.24	0.00	0.00	0.00	0.00	6.63	0.00	1.13
1190	92.05	0.00	0.00	0.00	0.00	6.81	0.00	1.14
1180	90.86	0.00	0.00	0.00	1.10	6.90	0.00	1.14
1170	89.72	0.00	0.00	0.00	2.19	6.96	0.00	1.14
1160	88.45	0.00	0.00	0.00	3.98	6.13	0.00	1.44
1150	86.98	0.00	0.00	0.00	5.48	6.03	0.00	1.51
1140	85.47	0.00	0.00	0.00	6.99	5.94	0.00	1.60
1130	71.17	0.00	13.41	1.29	7.47	5.20	0.00	1.45
1120	58.55	0.00	15.54	4.66	14.84	4.97	0.00	1.43
1110	50.49	0.00	17.92	7.80	17.36	4.98	0.00	1.45
1100	43.25	0.00	19.85	10.71	19.73	5.00	0.00	1.46
1090	36.46	0.00	21.18	13.27	22.62	4.97	0.00	1.50
1080	30.01	0.00	21.80	15.24	26.65	4.83	0.00	1.47
1070	24.62	0.00	22.35	16.88	30.00	4.71	0.00	1.45
1060	20.19	0.00	22.80	18.20	32.77	4.59	0.00	1.46
1050	16.61	0.00	23.18	19.20	35.07	4.48	0.00	1.44
1040	13.73	0.00	23.54	19.95	36.98	4.39	0.00	1.42
1030	11.49	0.00	23.84	20.47	38.49	4.30	0.00	1.41
1020	9.72	0.00	24.12	20.83	39.70	4.23	0.00	1.40
1010	8.30	0.00	24.39	21.08	40.67	4.17	0.00	1.39
1000	7.15	0.00	24.66	21.25	41.44	4.13	0.00	1.38
990	6.25	0.00	24.89	21.36	42.03	4.08	0.00	1.40
980	5.48	0.00	25.13	21.43	42.51	4.05	0.00	1.40
970	3.94	0.11	25.56	21.74	43.25	3.99	0.00	1.40
960	2.96	0.17	25.90	21.83	43.78	3.95	0.00	1.41

Appendix 8. Weight corrected and normalised proportion of phases calculated using MELTS from 1300 °C to 960°C at QFM+2 conditions.

T°C	liquid	olivine	clinopyroxene	feldspar	leucite	spinel	apatite	fluid
1300	98.24	0.00	0.00	0.00	0.00	0.67	0.00	1.08
1290	97.72	0.00	0.00	0.00	0.00	1.16	0.00	1.12
1280	97.27	0.00	0.00	0.00	0.00	1.61	0.00	1.13
1270	96.84	0.00	0.00	0.00	0.00	2.02	0.00	1.14
1260	95.01	0.00	0.00	0.00	1.48	2.38	0.00	1.13
1250	94.06	0.00	0.00	0.00	2.09	2.72	0.00	1.13
1240	93.05	0.00	0.00	0.00	2.78	3.04	0.00	1.13
1230	92.01	0.00	0.00	0.00	3.52	3.34	0.00	1.14
1220	90.97	0.00	0.00	0.00	4.28	3.62	0.00	1.14
1210	89.93	0.00	0.00	0.00	5.05	3.88	0.00	1.14
1200	88.90	0.00	0.00	0.00	5.83	4.12	0.00	1.14
1190	87.89	0.00	0.00	0.00	6.62	4.35	0.00	1.14
1180	86.54	0.00	0.00	0.00	7.64	4.64	0.00	1.18
1170	84.84	0.00	0.00	0.00	8.94	4.94	0.00	1.28
1160	83.18	0.00	0.00	0.00	10.30	5.17	0.00	1.35
1150	79.29	0.00	2.74	0.00	11.39	5.21	0.00	1.38
1140	73.58	0.00	7.12	0.00	12.80	5.10	0.00	1.40
1130	67.35	0.00	10.68	0.71	14.85	5.02	0.00	1.39
1120	57.44	0.00	14.06	4.42	17.41	5.22	0.00	1.46
1110	48.65	0.00	16.99	7.62	19.87	5.38	0.00	1.49
1100	41.11	0.00	19.43	10.32	22.13	5.48	0.00	1.53
1090	34.39	0.00	21.28	12.54	24.74	5.52	0.00	1.53
1080	28.21	0.00	22.00	13.95	28.97	5.39	0.00	1.48
1070	23.43	0.00	22.52	15.03	32.28	5.30	0.00	1.45
1060	19.72	0.00	22.92	15.82	34.88	5.21	0.00	1.45
1050	16.86	0.00	23.22	16.40	36.95	5.14	0.00	1.43
1040	14.60	0.00	23.48	16.81	38.61	5.09	0.00	1.41
1030	12.83	0.00	23.71	17.12	39.90	5.04	0.00	1.40
1020	11.38	0.00	23.92	17.35	40.95	5.00	0.00	1.39
1010	10.21	0.00	24.13	17.52	41.78	4.98	0.00	1.38
1000	8.71	0.00	24.44	17.83	42.67	4.97	0.00	1.38
990	5.75	0.16	25.06	18.68	43.99	4.93	0.00	1.43
980	2.90	0.32	25.69	19.49	45.27	4.88	0.00	1.45
970	0.45	0.43	26.29	20.16	46.37	4.83	0.00	1.47
960	0.00	0.45	27.02	20.55	47.11	4.87	0.00	0.00

Appendix 9. Weight corrected and normalised proportion of phases calculated using MELTS from 1300 °C to 960°C at QFM+3 conditions.

T°C	liquid	olivine	clinopyroxene	feldspar	leucite	spinel	apatite	fluid
1300	95.60	0.00	0.00	0.00	0.00	3.33	0.00	1.07
1290	95.32	0.00	0.00	0.00	0.00	3.59	0.00	1.06
1280	95.06	0.00	0.00	0.00	0.00	3.86	0.00	1.06
1270	94.82	0.00	0.00	0.00	0.00	4.10	0.00	1.07
1260	94.58	0.00	0.00	0.00	0.00	4.33	0.00	1.07
1250	94.36	0.00	0.00	0.00	0.00	4.54	0.00	1.07
1240	94.10	0.00	0.00	0.00	0.00	4.71	0.00	1.09
1230	93.86	0.00	0.00	0.00	0.00	4.88	0.00	1.10
1220	93.63	0.00	0.00	0.00	0.00	5.08	0.00	1.11
1210	93.24	0.00	0.00	0.00	0.18	5.29	0.00	1.12
1200	91.90	0.00	0.00	0.00	1.38	5.49	0.00	1.13
1190	90.63	0.00	0.00	0.00	2.52	5.68	0.00	1.14
1180	89.15	0.00	0.00	0.00	3.82	5.49	0.00	1.14
1170	87.59	0.00	0.00	0.00	5.20	5.22	0.00	1.16
1160	85.96	0.00	0.00	0.00	6.69	6.17	0.00	1.44
1150	81.00	0.00	4.29	0.00	7.87	6.21	0.00	1.51
1140	74.80	0.00	9.03	0.00	10.36	6.22	0.00	1.60
1130	67.39	0.00	11.70	1.60	13.02	5.52	0.00	1.45
1120	57.58	0.00	14.19	5.01	15.89	5.31	0.00	1.43
1110	49.02	0.00	16.82	8.05	18.51	5.35	0.00	1.45
1100	41.47	0.00	19.05	10.75	20.96	5.41	0.00	1.49
1090	34.58	0.00	20.76	13.04	24.14	5.42	0.00	1.50
1080	28.27	0.00	21.61	14.66	28.35	5.31	0.00	1.47
1070	23.33	0.00	22.28	15.93	31.67	5.22	0.00	1.45
1060	19.46	0.00	22.80	16.88	34.32	5.13	0.00	1.46
1050	16.44	0.00	23.19	17.57	36.47	5.05	0.00	1.44
1040	14.07	0.00	23.50	18.05	38.20	4.98	0.00	1.42
1030	12.20	0.00	23.75	18.37	39.56	4.92	0.00	1.41
1020	10.72	0.00	23.97	18.59	40.64	4.86	0.00	1.40
1010	9.47	0.00	24.16	18.73	41.53	4.82	0.00	1.39
1000	8.49	0.00	24.36	18.84	42.22	4.79	0.00	1.38
990	7.67	0.00	24.53	18.89	42.80	4.75	0.00	1.37
980	5.73	0.09	24.72	19.19	43.70	4.67	0.00	1.40
970	3.01	0.23	25.09	19.77	44.89	4.57	0.00	1.40

Appendix 10. List of acronyms used in Chapter 5.

Acronym	Full form
AE	Acoustic Emission
AST	Auto-Sensor-Test
CWT	Continuous Wavelet Transform
DWT	Discrete Wavelet Transform
EMD	Empirical Mode Decomposition
FFT	Fast Fourier Transform
HDT	Hit Definition Time
HHT	Hilbert-Huang Transform
HLT	Hit Lockout Time
IMFs	Intrinsic Mode Functions
KL	Kullback-Leibler Divergence
PDT	Peak Definition Time
PE	Permutation Entropy
STFT	Short-Time Fourier Transform
TF	Time-Frequency



Abstract

Volcanic eruptions have an important impact on the environment, on urbanised or rural areas. Large eruptions can have a worldwide impact. Effusive eruptions are usually harmless to human lives but can cause great damage to infrastructures, cultivated lands and houses as lava flows burn and bury anything they encounter. The distance travelled by lava flows is controlled by many parameters, such as eruptive rate and rheology of the lava. The formation and development of lava tubes, highly efficient lava transport structures, within a flow can greatly increase its distance of emplacement.

Somma-Vesuvius is famous for its explosive activity, but the last eruptive cycle from 1631 to 1944 was a long period of open-conduit conditions with effusive to mildly explosive eruptions. Historical documentation describes the presence of lava tubes at Vesuvius, opening the possibility of finding lava tubes on the slopes of the volcano. Vesuvius and its surroundings are highly inhabited and present a high risk of destruction by effusive eruptions. Past events are reminders of such risk. It is thus crucial to better understand the emplacement of lava flows and the formation and development of lava tubes, especially at volcanoes producing silica-undersaturated potassium-rich lavas due to the absence of studies in such context.

The work carried out during this Ph.D. aimed at understanding lava flow emplacement, and lava tube formation and development in the lava flow fields present at Vesuvius. A detailed morphometry analysis of the largest lava tube formed in the 1858 flow field was conducted using a combination of Terrestrial Laser Scanner and airborne imagery. Lavas forming the tube were characterised by petrological, geochemical, and textural analyses performed by Scanning Electron Microscope, Electron Microprobe Analysis and synchrotron x-ray microtomography. The stability of the lava tube structure was assessed by Acoustic Emissions tests with a novel signal processing method combining Empirical Mode Decomposition (EMD) and Hilbert-Huang Transform (HHT).

The analyses show that the 1858 lava flow field reached its maximum extent quickly and then expanded both laterally and vertically. The size of the flow field appears to be influenced by the topography and the amount of crystals in the lava. Our findings suggest that the lava tube studied was formed by inflation of a lava flow breakout. The tube remained active for a long period and experienced at least twelve drainage and refill cycles. The thermo-redox conditions inside the tube were unstable and did not display a consistent trend during the activity of the tube. Microlitic crystallisation within the lava tube was enhanced at the sides of

the conduit due to thermal conditions and shear rates compared to the centre of the tube. The structure of the lava tube is fractured and contains loosened masses in some areas, but the overall structure remains stable.

Moreover, a geotinerary and virtual tour along the 1858 lava flow field was designed to showcase five peculiar volcanic landforms, three lava tubes, an ephemeral vent and a tumulus, that allows people to evidence effusive activity and discover volcanic features formed during lava flow emplacement at Somma-Vesuvius. A website was developed to host the virtual tour, which comprises 360° images and an explicative panel for each site. The website aims to sensibilibize people either not used to hiking or with physical impediment about volcanic landforms and their scientific relevance. The pertinence of the geotinerary has been assessed by a Strengths, Weaknesses, Opportunities and Threats (SWOT) analysis. The geotinerary was conceptualized for visitors to discover volcanoes and volcanic activity and may contribute to divert tourist traffic away from the volcanic cone, while raising awareness on volcanic processes and volcanic hazards.

Keywords: synchrotron x-ray microtomography; 1858 lava flow field; thermo-redox conditions; effusive activity; volcanic hazards; volcano tourism