A SHORT REVIEW OF PYRODUCTS (LAVA TUBES)

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Abstract: Lava tubes and caves, when accessible for man, are very common in volcanic environments and poorly known to non-specialists. This short overview presents the distribution and forms of lava tubes, their speleothems and mineralogy and modes of formation. Studies of lava caves outside of Earth currently are a topic of great interest, as they may be potential locations for some life forms and future bases in space exploration. Basic features of lava tubes are illustrated with reference to the longest lava cave in the world, Kazumura Cave, in Hawaii.

Key words: caves, speleothems, extraterrestrial, volcanic environment, Kazumura Cave.

INTRODUCTION

In describing the processes of formation and the forms of lava caves, researchers are in an exceptionally favourable position. The formation of lava tubes, especially at an early stage, has been well described from contemporary volcanoes, for example in Hawaii (Fig. 1) and on Mount Etna. Lava tubes, defined by Halliday (2004) as a "roofed conduit of flowing lava, either active, drained, or plugged", are common volcanic features with tunnel-like appearances. They are natural conduits, through which lava travels beneath the surface of a lava flow. When the supply of lava stops at the end of an eruption or the lava is diverted elsewhere, lava in the tube system typically drains downslope and leaves partially empty conduits beneath the ground. However, almost all the accessible lava caves have undergone moderate to substantial downward erosion. Thus, the lava conduit may leave an accessible cave without draining (S. Kempe, pers. comm., 2020). Recent overviews of problems, related to the description, genesis, nomenclature, etc. of lava caves, were presented by Halliday (2002) and Kempe (2012b, 2019).

Lava tubes (pyroducts) form in volcanic flows and are found in many volcanic regions of the world, for example, in the USA (Kilauea Volcano in Hawaii; Kauahikaua *et al.*, 1998), Italy (Mount Etna; Calvari and Pinkerton, 1998), Australia (Undara Volcano; Atkinson *et al.*, 1975), Iceland, the Canary Islands, Azores, India, Vietnam, Korea, Japan, Galapagos Islands, Easter Island, Mexico, Kenya, Saudi Arabia and Jordan (Sauro *et al.*, 2020 and references therein). Lava tubes are well known to people, who live in the

areas where they are common, but poorly known in Europe, except perhaps in Iceland and Italy. They are the subject of volcanospeleologic research, a term coined by William R. Halliday (Kempe, 2012a). Various terms are used for the primary forms: lava tubes, lava pipes, lava tunnels, or pyroducts. A crust forms on the surface of the cooling lava above the conduit. The prime difference between "channel" and "tube" is that the crust on the channel is discontinuous and moving in downstream direction, while the crust over the tube is stationary (e.g., Keszthelyi *et al.*, 1999). An open lava channel can evolve into a lava tube and it is not always easy to distinguish between a lava tube and a lava channel (e.g., Duraswaimi *et al.*, 2004). Accordingly, some researchers have used the term "canals/pipes" (see Sen *et al.*, 2012 for discussion). Although the term "lava tube" is used most often nowadays, it is worth noting the development of these terms in the geological literature, as summarized in table 6.2 of Lockwood and Hazlett (2010). These authors, also Kempe (2012b, 2019), strongly advise the use of the term "pyroduct" – "any internal lava conduit in a flow, irrespective of shape and size, regardless of whether it contains molten lava during eruptive activity or is preserved as an elongate cave after eruptive activity ends and molten rock drains away". There are two reasons for this. One is that the older term should take precedence. The term "pyroduct" was coined already in 1844 by the Reverend Titus Coan, after seeing a pyroduct in action on Mauna Loa, in March 1843 (Coan, 1844 *fide* Lockwood and Hazlett,
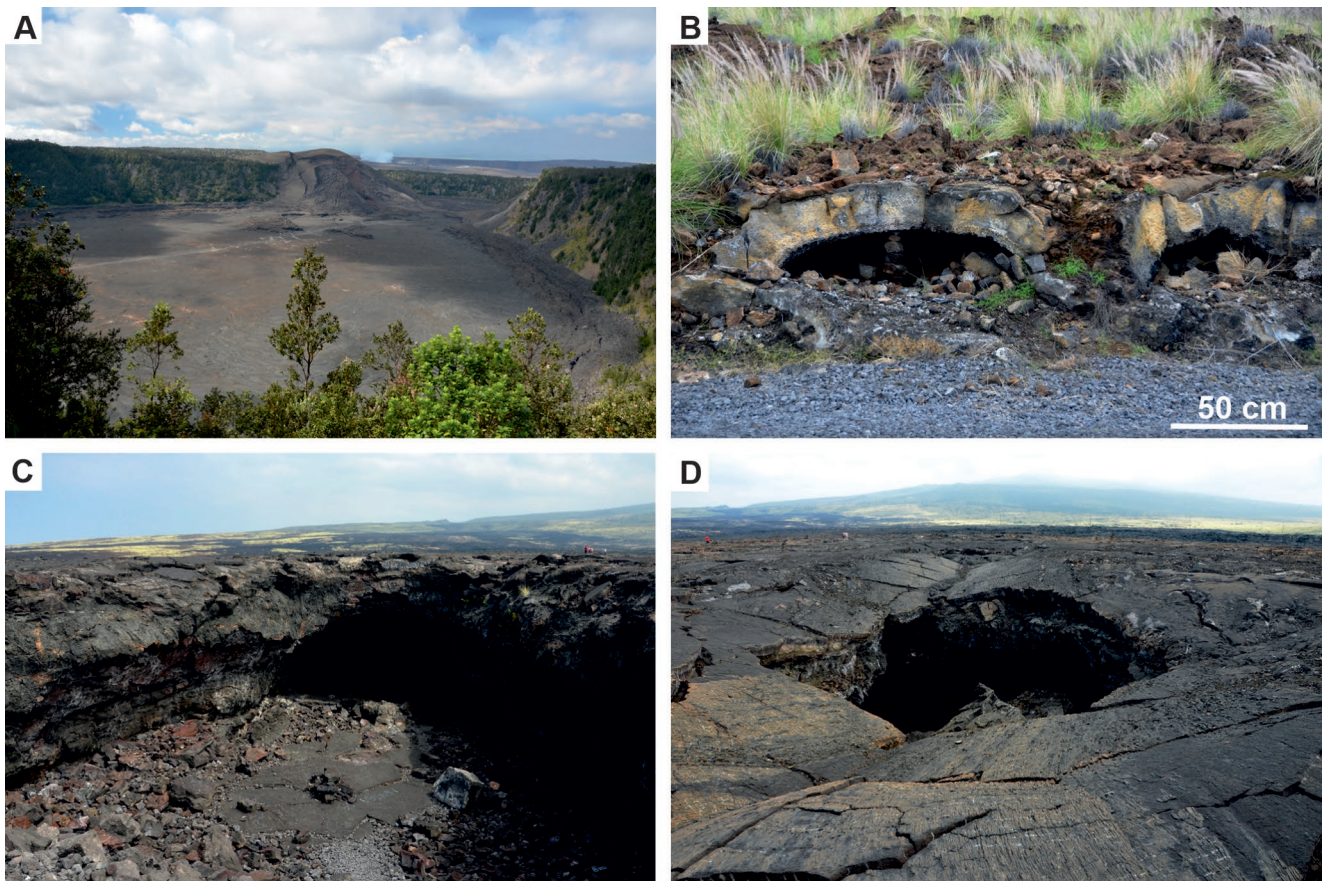
Fig. 1. Surface expressions of volcano and lava tubes. **A.** Kilauea Iki crater and fumes of the Halema ʻuma ʻu crater (active vent) of the Kilauea caldera in the backround. **B.** Upwarped lava sheets detached from their sheet below owing to lateral pressure. **C.** Large collapse giving horizontal access to cave passage. **D.** Collapse (puka) above large cave.

2010). The term "lava tube" was used later, for the first time by Tom Jaggar in 1919 (Jaggar, 1919 *fide* Lockwood and Hazlett, 2010). This term creates the impression of a tubular passage that can carry lava under pressure, which typically is not the case. Most passages are rectangular and the roof of a pyroduct does not have the tensile strength to sustain the pressure on flowing lava, apart from its own weight. Thus, the term "pyroduct", as defined above, seems to be more appropriate and should be used preferentially. In saying this, the present author cannot negate the terminology that is rooted in the extensive literature on the subject and will use different terms interchangeably herein.

Some authors (e.g., Gradziński and Jach, 2001) distinguish between a lava tube and a lava cave, the latter being accessible to humans. Some researchers describe lava caves as "pseudokarst" but this term does not seem to be appropriate (see discussion in Eberhard and Sharples, 2013).

Lava caves are quite common in the volcanic environments but poorly known to the public, including many speleologists. They pose a challenge to the standard thinking of a geologist because, unlike karst caves, they form in a geological instant, from a few weeks to a few years. In this short overview the author would like to show why lava caves are of great potential importance, describing: 1) the environment and modes of formation; and 2) the differences and similarities between lava and karst caves.

Kazumura Cave (Big Island, Hawaii), the longest lava cave in the world, is presented here as providing excellent and representative examples of various lava features.

GENERAL CHARACTERISTICS OF LAVA TUBES

Distribution and forms

Volcanic caves are a broad group, which includes the caves created by volcanic explosion, lava effusion and various volcanic processes. Bella and Gaál (2013, and references therein) distinguished roofed lava channels, lava tube caves, drainage tubes of active rift zones, lava rise caves, lava ridge caves, lava tumulus caves, boulder caves in lava flows, volcanic crater shafts, volcanic exhalation caves, chimneys of spatter cones and hornitos, lava blister caves, polygenetic spatter cone caves in carbonatite volcanoes, volcanic pyrogenetic tree moulds, and volcanic eruptive fissure caves. By far the most common are the first two types. Most of the pyroducts that actively transmit lava in a subterranean setting formed syngenetically with the laterally flowing lava: they are described in detail below. Lava tubes and channels occur also in seafloor volcanic terrains of the East Pacific rise and in seamounts in the eastern Pacific. They can form horizontal intracrustal pathways for the circulation

of hydrothermal fluids (Fornari, 1986). Secondary caves in lava flows often are formed along tectonic fissures as a result of the collapse of the ceilings of igneous chamber fragments and by erosion (the shorelines of the seas and lakes or the banks of rivers; Kempe, 2012a, and references therein). It is worth noting, however, that distinguishing the origins of some caves is not easy. For example, one cave (Kukaʻau Cave) in the Hamakua volcanites on Mouna Kea, was formed as a result of erosion (Kempe and Werner, 2003), while the other (Pʻaʻauhau), was established originally as a lava tube, but was later modified as a result of erosion by a stream (Kempe *et al.*, 2003).

Almost all caves are formed in alkaline and tholeiitic basalt; among the few exceptions are caves formed in phonolites and carbonatites (McFarlane *et al.*, 2004; Kempe, 2013 and references therein). Two main types of lava host the majority of lava caves and channels: aʻa (rough, brecciated basalt lava) and pāhoehoe (Fig. 2; smooth, glassy basalt lava; Hawaiian nomenclature; Harris *et al.*, 2017). Channels are commonly found in both aʻa and pāhoehoe flows, whereas
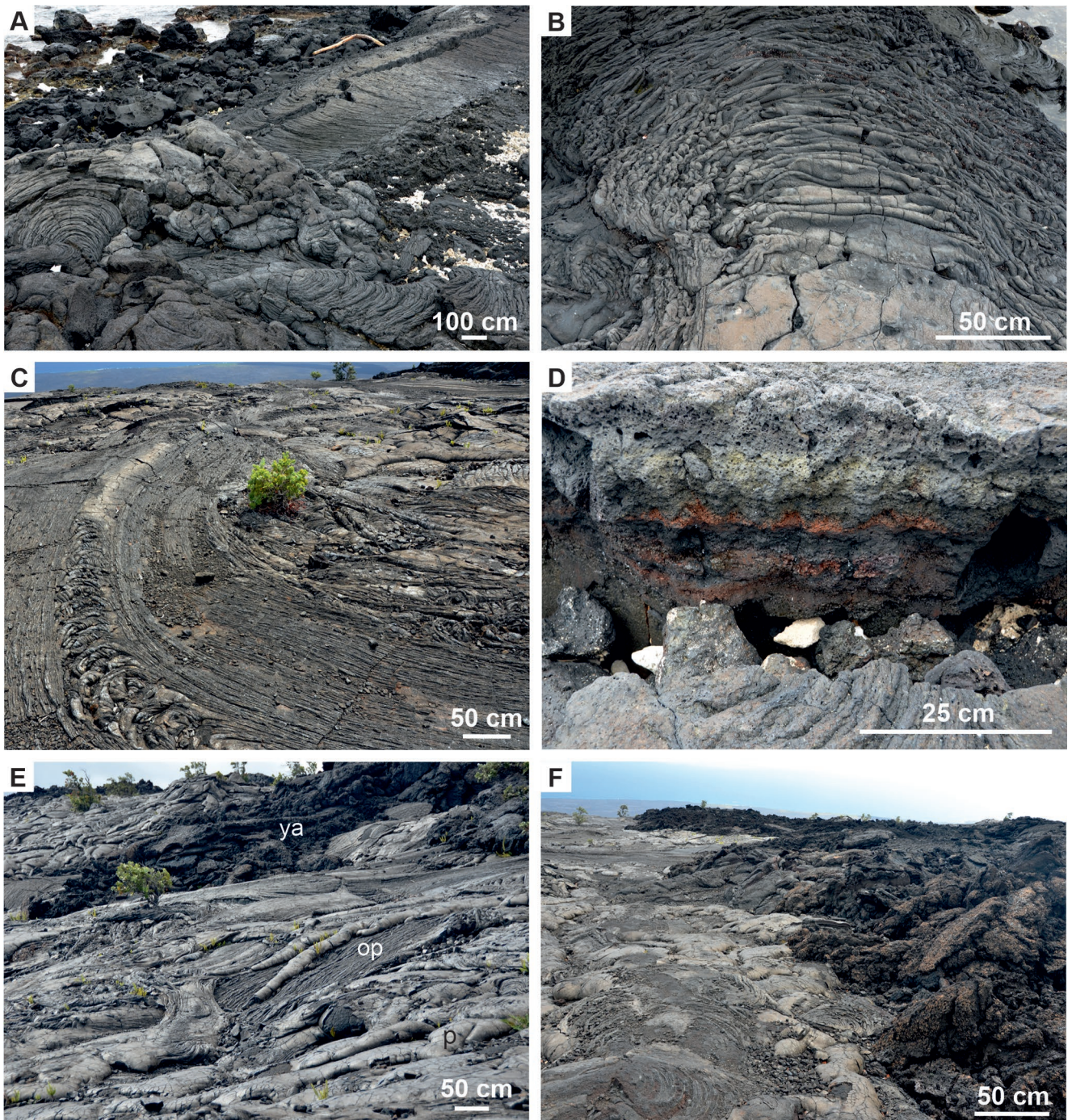

Fig. 2. Various types of Hawaiian basalt lavas (Big Island). **A.** Pāhoehoe lava on the coast. Note a pressure-ridge in the backround with possible cave underneath. **B.** Undulating surface of the frozen tongue of ropy pāhoehoe lava. **C.** Pāhoehoe lava flow. **D.** Cross-section of a pāhoehoe flow exposed in a sea cliff. Different colours of the individual flow units resemble those observed in lava tubes. **E.** Sequence of flows: older pāhoehoe (op) – pāhoehoe (p) – younger ʻa (ya). **F.** Aʻa flow (darker) over older pāhoehoe flow (shiny).

tubes are much more common in pāhoehoe flows (e.g., Hon *et al.*, 1994; Kauahikaua *et al.*, 2003). However, more and more evidence is emerging that the lava tubes in a ʾa flows are not as rare as previously was thought (e.g., Coombs *et al.*, 1995; Calvari and Pinkerton, 1999; Wantim *et al.*, 2013). A ʾa lava is found on the floors of lava caves as a result of terminal cooling that increases the viscosity of the lava, pulling the lava apart to form clinker (S. Kempe, pers. comm., 2020). There are several other rarer lava tubes. Discrete lava tubes form in hummocky flows, in which lava pathways have many dead ends, so that the lava cannot flow (Wise, 2014). Special "littoral" tubes associated with pillows, rarely drained and mostly plugged, form when the lava flow reaches the sea (Peterson *et al.*, 1994). Lava tumuli can be hollow inside (e.g., in Kilauea caldera; Halliday, 1998). Pedersen *et al.* (2017) described tumuli hosting small caves and chambers, associated both with a ʾa and pāhoehoe inflationary tubes, which resulted from lava being injected from below. Deep "inflated-entrenched" lava tubes, some being very large, were considered to be formed through inflation along deep inception horizons, following previous lava flow boundaries. There, after inflation, the conduit is enlarged by downward, thermic erosion and breakdown phenomena (see Kempe, 2019). Channel curvature affects the surface morphology and dynamics of the flow and can stimulate the formation of a lava tube (Valerio *et al.*, 2011).

Lava tubes/pyroducts may be divided into single-trunked, double (or multiple)-trunked, and superimposed-trunked systems (Kempe, 2009). Lava channels and tubes often form distinctly complex networks with anastomosing and braided reaches. They often intersect or join in different lava flows (Kempe, 1999). Examples of various systems can be found in Kempe (2013). Such systems can greatly influence the development of the coastline in volcanic areas (e.g., Kauahikaua *et al.*, 1993; Umino *et al.*, 2006; Ramalho *et al.*, 2013).

The formation of lava tubes is a relatively fast process. Their growth is hampered by the rate of lava cooling and local topography. It also is influenced by changes of internal pressurization, associated with the inflation of a lava flow (Glaze and Baloga, 2015). The population density of the tubes is inversely proportional to their diameter (Coombs *et al.*, 1995). Sustained flow forms small tubes over periods of hours to days (e.g., Peterson *et al.*, 1994; Byrnes and Crown, 2001), whereas large tubes form over weeks to months (e.g., Harris *et al.*, 1997; Calvari and Pinkerton, 1998). Kempe *et al.* (2018) pointed out that, contrary to popular opinion, lava caves are not created at the end of the eruption "when the tube runs empty", but prolonged activity creates a gas space above the down-cutting lava river.

Most of the intact lava tubes are restricted to lavas that are less than a few million years old. Among that oldest are the San Antonio Mountain Cave (SAM) in New Mexico (nearly 4 Ma, Rogers *et al.*, 2000) and lava caves of the Al-Shaam plateau in Jordan (about 7 Ma; Kempe and Al-Malabeh, 2005). Tubes in older volcanic environments typically collapse, forming rills, sinuous ridges and channels, and/or are filled with sediments.

Lava tubes start typically at the shield volcano that issued the lava. For example, tubes, as much as 20 m high

and 10–25 m wide, were observed within a kilometre of the Pu'u 'O'o-Kupaianah eruption of Kilauea (Kauahikaua *et al.*, 1998) or at a distance of 600 m from a vent at the Mount Cameroon Volcano (Wantim *et al.*, 2013). The distance and the length are dependent on many factors, such as morphology, volume, channel width, viscosity of lava, etc. The lengths of lava caves vary from a few metres up to tens of kilometres and their width and height vary from about 0.2–0.5 m up to 30 m (Bunnell, 2008), but it should be noted that locally the dimensions of lava caves do not correspond to primary pyroducts, owing to, for example, erosion. The depth below the surface ranges from a few centimetres to a depth of a few tens of metres, typically less than 10–20 m (the total depth of lava caves is described as the difference between maximum and minimum altitudes above sea level). In most cases, tubes are sub-parallel to the surface itself.

Various factors are important in the internal development of the pyroducts, such as the joining of the smaller ducts, expansion during inflation, and thermal erosion of the substrate (Coombs *et al.*, 1995). During lava flow, tubes can be filled completely or only partly. Typically, the lava level in the conduit decreases, as a result of either reduced lava supply, or enlargement of the corridor cross-section due to erosion, with constant lava flow (e.g., Kauahikaua *et al.*, 1998). The shape of the tube can be very complex and varied (see for example Calvari and Pinkerton, 1999). They can continue to evolve in dimensions and shape, from an elliptical cross-section to arched, round, oval, or alternatively keyhole shaped, owing to the formation and failure of blockages, changes in effusion rate, thermal and mechanical erosion, or slope changes (e.g., Kauahikaua *et al.*, 1998; Kerr, 2001; Dragoni and Santini, 2007; Diniega *et al.*, 2013). Bends in a lava flow are often observed, and these can strongly affect the flow dynamics and the formation of lava tubes (Greeley, 1971; Peterson *et al.*, 1994). Common lava features in caves are lava stalactites, stalagmites, helictites, columns, flowstone, coralloids, grooves (flow lines), shelves, ceiling cusps, linings, lava falls, dams, levees, gutters and benches (e.g., Larson, 1993). The tube walls exhibit a single layer of lava lining at some places, whereas elsewhere it shows additional layer (e.g., Duraswaimi *et al.*, 2004; McHale, 2013; Fig. 3). Subsequent linings of the lava tube can have
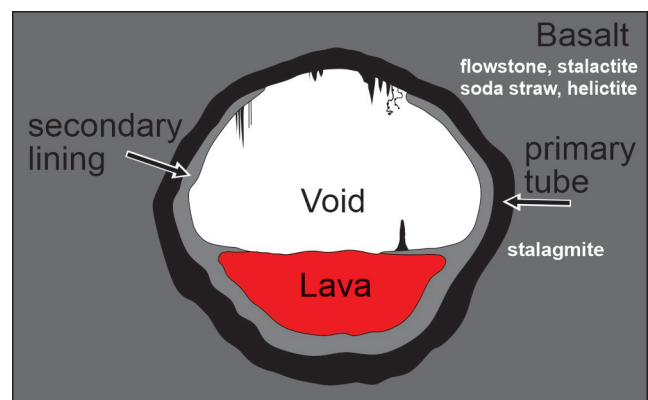

Fig. 3. Simplified cross-section through a lava tube, with lining, levee and most common types of lava speleothems (after Onac and Forti, 2011, modified).

different colours and compositions, depending on the primary composition and/or oxidation processes in the lava tube. Lava also can change its composition, owing to melting and the incorporation of the bedrock.

Research by Williams *et al.* (2003) in Cave Basalt (Mount St. Helens) has shown that the oldest outer lava linings in direct contact with dacite substrates are contaminated with substrate material, while younger, internal lava linings are uncontaminated. Contraction cracks, observed in the roof, are occasionally filled with sediments (Waichel *et al.*, 2013).

In active volcanic fields, tubes typically do not leave much surface expression, but occasionally tubes are marked by features, such as hornitos, elongate tumuli, skylights and break-outs (Fig. 1; Hon *et al.*, 1994; Calveri and Pinkerton, 1998; Kauahikaua *et al.*, 2003). Various holes/skylights are especially common above lava tubes and often used in exploration for them. They are called "pukas" in Hawai'i or "jameos" in the Canary archipelagos and usually form sinuous chains (Sauro *et al.*, 2020). Skylights usually result from collapse of the roof, which can be due to gravitational effects or flow overpressure (e.g., Cushing *et al.*, 2015; Sauro *et al.*, 2020) or through incomplete crusting over a lava channel developing into a lava tube. Kempe (2012a) distinguished two types of collapse forms: 1) a "hot puka", when the lava flow is still active and the rubble is carried away; and 2) a "cold puka", formed after the flow terminated and breakdown remains. Hot pukas cool the lava inside the tube and allow observations of the hot lava flowing under the surface. The temperature difference between the core of the lava stream and its surface can range from 29 to 144 °C (Witter and Harris, 2007). These primary skylights are very important in the formation of lava flows, changing air flow (cooling) or redirecting a lava flow. For example, Kempe (2012a) observed the collapse of the primary ceiling (hot puka), which allowed the inflow of cold air and consequently the solidification of the lava stream surface and the formation of a secondary ceiling. When lava is ejected through an opening in the lava crust, conical structures, hornitos, form. Lava can also invade a cave through cold pukas, forming a variety of large, vertical flow features (Kempe, 2013), for example, lava falls and lava curtains that fall into pre-existing pukas. Occasionally a 'a flows over older flows (see Fig. 2E, F) and enters pre-existing tubes through skylights or collapsed lava tubes (Shervais *et al.*, 2005), locally blocking them. Secondary skylights can form later, owing to tectonics or erosion.

The significance of lava tubes in volcanic flows is difficult to overestimate. Shield volcanoes owe their shape to the fact that low-viscosity, high-temperature lavas form internal tunnels, in which the lava can be transported for tens of kilometres (Kempe, 2009). The specific morphology of shield basalt volcanoes is largely due to the presence of numerous lava tubes that evenly and over long distances distribute the discharged lava (e.g., Peterson *et al.*, 1994). Within many lava flows in Hawaii, for example, lava tubes regularly facilitate transport of basaltic lava over distances of 10–20 km with temperature drops of ≤1 °C/km (Kauahikaua *et al.*, 2003). Lava tubes are important in the development

of lava flow areas, both pāhoehoe (e.g., Peterson and Tilling, 1980), and a 'a (e.g., Calvari and Pinkerton, 1999), enabling the formation of longer and wider lava flows (e.g., Cashman *et al.*, 1998; Al-Malabeh and Kempe, 2012). They facilitate the swelling/inflation of pāhoehoe flows (Hon *et al.*, 1994; Pasquarè *et al.*, 2008). Lava flows not only use previously formed lava tubes but can also destroy them.

Thermal and mechanical erosion

One of the important mechanisms for the formation of lava channels and pyroducts is thermal and/or mechanical erosion by flowing lava (e.g., Kauahikaua *et al.*, 1998). It involves some combination of thermal melting and assimilation and/or mechanical plucking and entrainment of underlying substrates by hot flowing lava (Hulme, 1973; Huppert *et al.*, 1984; Williams *et al.*, 1998). In the case of loose rubble, for example of a 'a lava, simple, mechanical erosion can be a major force (Kempe, 2012b). The distinction between thermal and mechanical erosion is difficult and often problematic, especially that often both processes are very effective and responsible for the final result. Gallant *et al.* (2020) documented extreme thermo-mechanical erosion by a small volume of lava. Downcutting by a basaltic-andesitic lava flow on the volcano Motombo (Nicaragua) was a hundred times the rate reported for thermal erosion in lava flow fields.

Generally, thermal erosion is more effective: in consolidated, non-volcanic substrates, having a lower melting temperature than lava, low mechanical strength, higher volatile content and lower conductivity; at lower slopes; at lower gravity; with a prolonged period of flow and higher turbulence (Hulme, 1982; Greeley *et al.*, 1998; Kerr, 2009; Hurwitz *et al.*, 2010). The efficiency of thermal erosion depends also on the physical properties of magma, which depend to a major extent on magma composition (Whittington *et al.*, 2020 and references therein). Thermal erosion channels in lavas have been identified, for example, in Hawaiian basaltic flows (Kauahikaua *et al.*, 1998), in low-viscosity and low-temperature carbonatite lava (Kerr, 2001), and in Archaean komatiite lava flows (Williams *et al.*, 1999). Coombs *et al.* (1995) calculated the thermal erosion rate of lava flows and tubes on Kilauea at 5.4 cm per day for 74 days.

Mechanical erosion is more effective in unconsolidated substrates (e.g., regolith; Hurwitz *et al.*, 2010). Siewert and Ferlito (2008) developed a model for mechanical erosion that explains the main field observations. Williams *et al.* (2003) found a greater abundance of xenoliths and xenocrysts relative to xenomelts in the Cave Basalt (Mt. St. Helens, USA) and suggested that mechanical erosion rather than thermal erosion was the dominant, erosional process.

It seems reasonable to use the term "mechanical/thermal erosion" in all questionable cases. It is also important to bear in mind the fact that in the processes of channel and tube formation, erosion is not the only possibility. For example, caves in a carbonatite volcano in Tanzania were formed by thermal erosion and the aqueous dissolution of spatter cones (McFarlane *et al.*, 2004).

Temperature, viscosity and lava flow rate in an active lava tube

Lava tubes can be excellent insulators of a lava flow. Isolating the lava in the tube reduces its rate of cooling and thus increases its range of flow (e.g., Keszthelyi, 1995; Keszthelyi and Self, 1998). The maximum surface temperature of the lava runoff in a Kilauea tube was 1,138 °C, and 1,020 °C at the outflow from the tube (Pinkerton *et al.*, 2002). The transport through lava tubes over 100 km in length of lava in inflated flows permits the molten lava core to reach the flow front, with a cooling rate of less than 0.5 °C/km (e.g., Keszthelyi, 1995; Sakimoto and Zuber, 1998; Riker *et al.*, 2009). The effective insulation provided by the roof means that tube-fed flow has the potential to extend tens to hundreds of kilometres before the core cools by 200 °C, in spite of low (1–4 m³/s) effusion rates (Harris, 2006). In a master tube with a thick roof, the cooling rate is the lowest among all types of lava flow. Length of lava flow, assuming that composition, isolation and morphology are constant, will rise with the effusion rate (Harris and Rowland, 2009). The heat loss of the magma in the lava tube is different than that in the surface runoff. The studies of Pinkerton *et al.* (2002) of Kilauea flows and tubes showed that tubed flows have a different surface thermal profile compared to those of active channels in subaerial flows. Tubed flows have margins that are cooler than the centres (see also Flynn and Mouginis-Mark, 1994); this reflects the increased importance of conductive heat loss through the tube walls, compared to radiant heat loss from the surface. Detection of thermal anomalies can help to locate the positions of active lava tubes, Temperature distributions on pāhoehoe flow fields revealed temperature anomalies of up to 150 °C above active tubes and tumuli (Pinkerton *et al.*, 2002).

The rate of lava flow in a tube and its width and height can be determined, among other properties, by measuring its flow between two skylights (Tilling and Peterson, 1993). The velocities of lava flows measured within channels and tubes are typically 1–3 m/s but may approach as much as 10 m/s (e.g., Hon *et al.*, 1994; Kauahikaua *et al.*, 2003; Harris *et al.*, 2007). Changes in velocity and viscosity of lava flows lead to the formation of different lava types. Belousov and Belousova (2018) studied different types of lava flows of the 2012–2013 eruption of the Tolbachik volcano. They showed significant differences in the propagation velocities of flows and the viscosity of a ʻa and pāhoehoe (velocity 2 to 25 mm/s and 0.5 to 6 mm/s and viscosities 1.3×10^5 to 3.3×10^7 Pa·s and 5×10^3 to 5×10^4 Pa·s, respectively). The parental lava was identical for both lava types. The viscosity of lava runoff generally increases with time, as the cooling effect becomes important. Diniega *et al.* (2013) studied the influence of viscosity on lava flow dynamics and created a model that presented a plausible explanation for why channels and tubes are common features of basaltic flows. Modelling of the lava flow in the tubes shows that pressure is the determining factor for very slight slopes, while for higher slopes, gravity becomes the determining factor (Sakimoto *et al.*, 1997).

Lava tube mapping methods and lava temperature assessment

Most lava tube mapping to date on Earth has been done directly by humans. There is a growing interest in use of sophisticated instrumental, mainly geophysical, methods. Location of the lava flows, channels and tubes has been studied by means of several techniques, e.g., geoelectromagnetic (Bozzo *et al.,* 1994), changes in the magnetic field (Budetta and Del Negro, 1995), radar (GPR; Miyamoto *et al.*, 2005), laser scanning (TLS – 3D models; Nelson *et al.*, 2011), very low frequency electromagnetic induction (VLF; Kauahikaua *et al.*, 1990), Forward Looking InfraRed (FLIR) thermal camera (Spampinato *et al.*, 2008), multispectral infrared images, tracing a 10–15 °C temperature anomaly on the surface, directly above the tube (TIMS; Realmuto *et al.*, 1992), and estimation of lava flow temperatures, using Landsat night-time images (Nádudvari *et al.*, 2020).

Several geophysical methods were used beyond the Earth. For example: on the Moon (NASA's GRAIL mission – Chappaz *et al.*, 2014) and on Mars (images of THEMIS, MOC and HiRISE – Giacomini *et al.*, 2009; VNIR and TIR – Crown and Ramsey, 2017; Helmholz resonance – Williams *et al.*, 2017). Various technologies for robotic expeditions that will explore skylights, lava tubes and caves on other planets and moons have been proposed (Antol, 2005; Whittaker, 2011; Kalita *et al.*, 2018).

Formation of pyroducts (lava tubes)

The formation of pyroducts may be a very complex and multi-stage process, involving several lava flows (see e.g., Bauer *et al.*, 2013). The first, genetic observations of actively forming lava tubes by "overcrusting" of an open channel were made by Peterson and Swanson (1974) during long-lasting effusive eruptions at the Kīlauea volcano, on the Island of Hawaii. Excellent reviews of various formation processes can be found in Kempe (2012a, 2019) and more recently in Sauro *et al.* (2020).

Several modes of formation of lava tubes (pyroducts) exist (Fig. 4). Two main ways (Kempe *et al.*, 2010) are: 1) "Inflationary" (Fig. 4A) – The lava flows grow at their distal tips, where hot lava quickly covers the ground in thin sheets. The next advance makes this sheet swell before the formation of the next distal surface sheet and later downward erosion. This process can be repeated. 2) "Crusting over of channels" by closure by slab jam and closure by lateral shelf growth (Fig. 4C left and right, respectively). The tubes that form by roofing over the channels tend to have relatively thin roofs (roof thickness << tube diameter), whereas the tubes that form within inflated sheets tend to have thicker roofs (Keszthelyi *et al.*, 1999). Most pyroducts form by the first process at the tip of the lava flow by a repeated process of advance and inflation (Kempe, 2013). Those interested in the differences between "inflationary" *versus* "crusted-over roofs of pyroducts" are directed to Kempe *et al.* (2010).

Kempe (2012a) distinguished also caves, formed by coalescence of small ducts and consecutive downward erosion (Fig. 4B). During long-lasting eruptions, several small tubes can merge focusing the flow along one main path. The focused flow has stronger thermal erosion potential and can enlarge and entrench the conduit (Kempe, 2019). Bauer *et al.* (2013) described the Kahuenaha Nui Cave (Hawaii), which formed in four different lava flows. First, the trunk passage formed by eroding an underlying a'a rubble layer. Then, a stack of several superimposed pāhoehoe flows with small ducts combined into one flow, eroding the main trunk underneath.

Lava tubes, formed by shallow inflation, usually are characterized by a surficial bulge (due to inflation) and by an original horizontal elliptical cross-section that can be entrenched by thermal erosion (Sauro *et al.*, 2020). The development of tunnels that carried lava to the distal fronts due to the reduction in effusion rates may generate localized inflation phenomena throughout the lava flow (Bernardi *et al.*, 2019).

Crusting over the channel can happen through different mechanisms, operating separately or in combinations , depending on the flow rate, turbulence and channel geometry (Dragoni *et al.,* 1995; Sauro *et al.*, 2020): 1) the growth of solidified, rooted crusts from lava stream banks; 2) overflows and spatters, accreted to form shelves and levees that progressively grow, forming a roof across the stream; and 3) plates and lithoclasts of solidified lava floating downstream, welding together and forming a blocky roof. Overcrusted tubes are usually limited to the width and depth of the
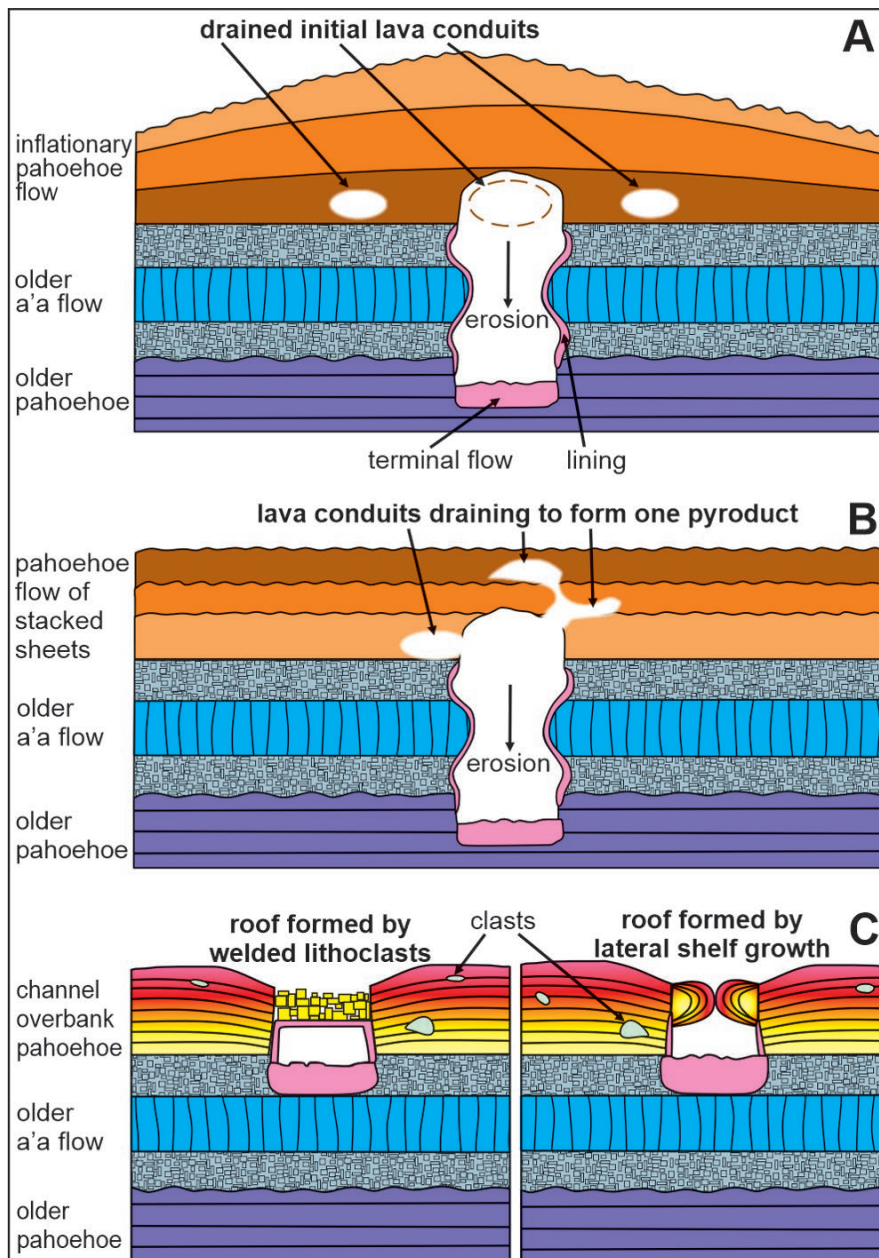

Fig. 4. Different cave formation modes within the pyrodct. **A.** Inflation of the lava flow front and later downward erosion. **B.** Coalescence of small ducts and consecutive downward erosion. **C.** Crusting-over of channels by floating lithoclasts, welded together (left) and by lateral shelf accretion and consecutive closure (right; after Kempe, 2012b, modified).

channels, where they originally formed (Sauro *et al.*, 2020). Various rheological models have attempted to explain crust formation in lava flows and lava tube formation (Dragoni *et al.*, 1995; Cashman *et al.*, 2006; Valerio *et al.*, 2011; Filippucci *et al.*, 2011). Near the eruption site, when the lava surface temperature is high, the crust is thin and fragmented. With cooling, the fragments thicken and merge into a continuous shell (Dragoni *et al.*, 1995). The amount of crust coverage is mainly controlled by the channel curvature and width, narrow channels have a greater coverage than wide ones (Valerio *et al.*, 2011).

Speleothems and mineralogy

Mineral speleothems in lava caves and karst caves show many similarities (Kempe, 2013; see also comparison in Hill and Forti, 1997). However, many of the minerals of the former are known only from lava caves (e.g., Etna Caves; Forti, 2000).

There are both primary forms, consisting of the minerals that make up tholeiitic basalt (rock speleothems or lavacicles), and secondary forms (mineral speleothems, *sensu* Kempe, 2013; e.g., calcium carbonates, gypsum, opal, mirabilite, tenardite, and vanadate phases; Forti, 2005; Hon *et al.*, 2009; Guimbretièr *et al.*, 2014). Kempe (2013, table 1) elegantly compared the morphology of mineral and rock speleothems. These two types result from different processes. Rock speleothems form in a very short time through the flow of molten material , whereas mineral speleothems form over long time scales through minute additions from solution (Kempe, 2013).

Among lava (rock) speleothems are tubular stalactites, stalagmites, columns, soda straws, and corraloids, flowstones, helictites, barnacle-like stretched lava, runner, runner channels, and lava blisters or squeeze-ups. Lava stalagmites can exceed 3 m in height and a lava column in a Korean lava tube cave is 7.6 m high (Hong, 1992).

According to Allred and Allred (1998), interstitial fluid is expelled from lava as a result of retrograde boiling. Stalactites are the most common rock speleothems. Kempe (2013) distinguished several classes, differing in mode of formation: 1) an extrusion of small amounts of melt from the roof at regular distances; 2) a quick lowering of the lava level, drawing down residual melt sheets from the ceiling; 3) the extruded residual melt first forms drops and erratic extrusions and finally forms a cylinder, drawn down because of the weight of the "pigtail" at the end; 4) a repeated dipping of the pulsating lava flow; 5) spattering (Bosted and Bosted, 2009); and 6) the cascading of lava down through ceiling holes. Rock stalagmites also can be divided into several morphological classes (Kempe, 2013): 1) agglutinated, small and similar-sized drops of lava and broken fragments of cylindrical stalactites; 2) larger spatter; and 3) fused, larger amounts of lava, cascading down from upper passages or through pukas from the surface.

Cave minerals in volcanic environments and their modes of formation are described in the excellent overview by Forti (2005). Lava cave minerals constitute up to 40% of the secondary chemical deposits found in all the caves of the world and 35% of them are restricted to volcanic environments.

Processes, mechanisms, temperature ranges and related chemical deposits are summarized in table 1 of Forti (2005). Two mechanisms are exclusive to volcanic environments. In the early stages, processes are practically controlled by the temperature of the cave atmosphere. When the lava walls cool, the chemical composition and the morphology of speleothems change (Forti, 2005). A list of minerals restricted to volcanic environments can be found in table 2 of Forti (2005).

The chemical composition of secondary minerals found in lava caves is diverse. A number of sulphate, chloride and fluoride minerals are formed, both in and around active lava tubes. Hon *et al.* (2009) gave an excellent description of these primary forms. Gases emitted from the lava in the tubes react with the surrounding atmosphere and tube walls and form both sublimates and precipitates. The spatial and temperature zoning of these forms is interesting. Inside the glowing tubes at temperatures >1000 °C, subhedral anhydrite, glauberite and magnesioferrite are formed. Cu-bearing Na_2SO_4 and $KNaSO_4$ cover the walls directly outside the glow zone. Their lower temperature polymorphs, tenardite and afthitalite, are formed at temperatures <300 °C in subsurface fumaroles. At temperatures of 100 to 400 °C, anhydrite and native sulphur are formed in the vacuoles. Surface and near-surface sulphatars form a mineral complex of acid hydrated Mg-Fe-Al-Na sulphates (locally fluorides are also present). When the cooling lava comes into contact with acid gases, gypsum, ralstonite, opal and iron chlorides and hydroxides are formed. The drying of the lava tube leads to the collapse of the transformation zones and the formation of precipitates at 30–60 °C, dominated by Na-Mg-K sulphates. After the tube cools down to the temperature of the surrounding environment, these highly soluble sulphates are removed by meteoric waters and gypsum is formed in their place.

Minerals in the same lava tube can differ, depending on the location. In basaltic caves at the Craters of the Moon National Monument (USA) McHenry (2009) found that most mineral coatings on the walls and ceilings were calcite or silica-dominated, whereas mounds of Na-sulphate and Na-carbonate precipitates were present on the floors. The morphology, the occurrence and the numeric density of several lava speleothems depend on which section of the lava tube they formed in and the morphological diversity of the different sections of the tube (Gadanyi, 2010).

Comparison of the chemical composition of the lava stalactites (frozen lava drops) and the roof of the Etna lava tube showed their high degree of similarity, with simultaneous large differences in the composition and texture of plagioclases (Lanzafame and Ferlito, 2014). The floatation of megacrysts (plagioclase laths) towards the top of a lava tube was observed in the Snake River Plain Basalts, USA (Shervais *et al.*, 2005) and in the Khedrai Dam lava channel (Deccan Volcanic Province; Sen *et al.*, 2012).

Processes of oxidation are common and often very intensive during the first stages of lava tube formation. Oxidation marks (oxidized olivine phenocrysts, ferruginous clinopyroxene and large amounts of hematite) are visible, both in the rocks around the lava tubes and in the tubes themselves (Atkinson *et al.*, 1975).

Secondary minerals result from various chemical and physical processes inside a lava tube. Many of them are similar to those in karst caves, but some are not. Large speleothems (up to 3 m long) with complex mineralogy (mainly sodium carbonate) have been described from a cave in carbonatites as resulting from the interaction of endogenous condensates and meteoric waters (McFarlane *et al.*, 2004). Chemtob and Rossman (2014) described opaque surface coatings, composed of amorphous silica and Fe-Ti oxides. These coatings were the product of interaction of the basaltic surface with volcanically derived, acidic fluids. Secondary minerals form within the host rock and may seal the primary porosity of the lava (Kempe *et al.*, 2003). These fragile forms can be easily destroyed, by natural and human forces. Guimbretière *et al.* (2014) studied secondary minerals (from thenardite to vanadates phases) in the lava tubes from a locally hot, recent volcanic flow of the Piton de la Fournaise volcano (Réunion) and found that environmental conditions of the lava tube evolved very quickly during three months and all the speleothems disappeared.

A number of secondary speleothem types (e.g., moonmilk and vermiculations) are formed through the action of microorganisms or processes related to organic remains. Soft moonmilk-type carbonate deposits were found in Cueva del Viento (Tenerife, the Canary Islands; Gradziński and Jach, 2001) and recently were studied extensively in the Show Caves on the Galapagos Islands (Ecuador; Miller *et al.*, 2020).

The presence of a high silica content in the basaltic walls and/or sediments of volcanic caves is especially important. The dispersed remains of filamentous microorganisms were found on the coral-like formations in the Chilean Ana Heva lava cave, composed of amorphous material, containing magnesium and silica (Miller *et al.*, 2014). Kashima *et al.* (1987) described speleothems in the caves in Japan, consisting mainly of the skeletons of diatoms, alternating with layers of clay and detrital material, cemented by silica. De los Rios *et al.* (2011), examining the ochreous speleothems in a lava tube on the island of Terceira (Azores), found bacterial structures in them and suggested that bacteria, through their metabolic activity, not only precipitate ferrihydrite, but may also, through nucleation, contribute to passive precipitation of minerals. Secondary mineral deposits in the lava caves are related to the biogenic mineralization of bones and guano deposits. Forti *et al.* (2004) described 19 such minerals from lava in Saudi Arabia. Three of them are extremely rare organic compounds strictly, related to the combustion of guano.

In some lava caves, ice occurs the whole year round (e.g., Mauna Loa Icecave, Hawaii – see Kempe and Ketz-Kempe, 1979, or Bat Cave, New Mexico – see Crumpler and Aubele, 2001; Parmenter, 2018). A specific shape of lava cave (a collapsed entrance in the upper section and deeper, descending, sinuous passages) and a high-altitude location (not necessary) may form a cold air trap (see Perşoiu and Onac, 2019). These caves may contain so-called cryogenic minerals. In two ice caves on the Mauna Loa volcano (Hawaii), Teehera *et al.* (2018) found multi-phase deposits consisting mainly of secondary amorphous silica, cryptocrystalline calcite, and gypsum.

Water and lava caves

Lava caves are generally dry, but in some cases the presence of internal sediments evidences secondary water flow (Gradziński and Jach, 2001, and references therein). On Rapa Nui, Paulo (2009) observed that the silty infilling of the lava caves is easily removed by waves, resulting in the exposure of lava corridors, mostly at the coast; the recent ones are in the wave range and the older ones high on the cliffs. In 2004, Edmonds and Gerlach (2006) observed that Kilauea lava was emerging from tubes on the lava delta and flowing into the sea as several continuous streams of lava. Such flows generally flow into the sea passively but explosive activity occurs, when the sea flows into underground lava tubes (Mattox and Mangan, 1997). Some lava cave systems, such as Lanzarote, include both dry and submerged passages (e.g., Wilkens *et al.*, 2009). In the case of the Corona cave, it is assumed that it formed under subaerial conditions and was later flooded during subsequent post-glacial sea level rise (Wilkens *et al.*, 2009, after Carracedo *et al.*, 2003). It is noteworthy that diving or snorkelling in water-flooded lava caves is becoming popular, like geothermal snorkelling in the Leitharendi lava cave on Iceland.

Lava tubes may be important in groundwater transport and storage, forming a complex of fissured and conduit aquifers within lava flows (for a comparison with karst aquifers see Kiernan and Middleton, 2005).

Habitats

Living creatures of all sizes can be found in the majority of caves, although only some of them live in lava caves. Lava caves are used as shelters for various bigger mammals, for example in Saudi Arabia, hyenas, wolves and foxes can be found in lava caves (Forti *et al.*, 2004). Rodents, birds, and bats are common in entrances, skylights, and passages. Carpets of moss develop near entrances and below skylights in the lava tube caves. Together with thin roots from trees, growing above the passages and often extending into the cave, they are an excellent environment to host cave-adapted communities of troglobites (spiders, insects, crayfish, salamanders, and fish), fungi, and microbes. Spiders are especially common and some of them are without eyes (Gertsch, 1973). Over 100 invertebrate species have evolved in the Hawaiian lava tubes, through a reduction in or loss of characteristics (e.g., Bousfield and Howarth, 1976). Chapman (1985) studied Hawaiian lava caves and found that most are inhabited periodically, but some of them, with a favourable microclimate, may become the main place of residence for certain organisms. The main energy sources in Hawaiian lava tubes are plant roots, especially Ohia-lehua, slimes deposited by organically rich, percolating ground water and accidentals, which are those animals that blunder in and die (Howarth, 1978). Different organisms are common at different locations. For example, diplurans (hexapods) are present in many lava tubes of the northwestern United States but are poorly represented in volcanic caves elsewhere in the world (Ferguson, 1992). In the submerged parts of lava caves, a variety of aquatic fauna may develop, for example, remipede crustaceans and polychaete worms (Wilkens *et al.*, 2009).

Volcanic caves are filled with colourful microbial mats on the walls and ceilings (e.g., Garcia *et al.*, 2009; Moya *et al.*, 2009). Spores, coccoid, diatoms, and filamentous cells, many with hair-like or knobby extensions, were some of the microbial structures observed in biofilms called "lava wall slime". Two types are common, snoottites (jelly-like icicles hanging from the ceiling and walls of a cave) and biovermicultions (various patterns of dots, lines, or networks, resembling dendrites or hieroglyphs on the walls of a cave). Microbiological studies of lava caves are becoming especially important in the context of detection of life in the subsurface of extraterrestrial bodies (Northup *et al.*, 2011).

Microbial communities in lava caves range from hard to soft and from mineral deposits to the microbial mats that line cave walls. Multi-coloured microbial mats and inorganic secondary mineral deposits host a wide variety of microorganisms (Northup *et al.*, 2011). On the other hand, the white and yellow microbial mats of the Azores and Hawaii do not show much morphological differentiation (Hathaway *et al.*, 2014). Variations in local deposit parameters probably govern the composition of microorganisms in recent volcanic deposits (Gomez-Alvarez *et al.*, 2007). Diatoms, including new species, are quite common in lava caves near their entrances, in areas illuminated by natural light (Rushforth *et al.*, 1984; Lowe *et al.*, 2013). It is interesting to note that *Navicula thurstonensis* was found in the Thurston Lava Tube (Hawaii), also in artificially illuminated sections (Rushforth *et al.*, 1984)

Lava caves can have a potential as sources of novel microbial species and bioactive compounds, especially because Actinobacteria (and Proteobacteria) usually predominate in a lava cave environment (Cheeptham *et al.*, 2013; Riquelme *et al.*, 2015; Lavoie *et al.*, 2017). Interestingly, unique microbial diversity, distinct from other environments, including cave environments, has been found in Hawaiian ice caves (Teehera *et al.*, 2018). Mineralized microbialites are found not only deep in the darkness, but also grow on the ceilings and walls in the photic zone of several open caves in Hawaiian basalts, where fresh water seeps out of the rock (Léveillé *et al.*, 2000).

Man and mineral resources in lava caves

Lava caves have been used by man for many functions. In Hawaii, lava caves (usually those that could be entered through local ceiling collapses), especially in the near portion of the opening, were used as temporary or permanent shelters. Longer caves were places of refuge during wars (Sinoto, 1992). Kempe *et al.* (1993, 2009) noted various kinds of fortifications in some caves. Radiocarbon dating of charcoal supports the opinion that lava caves were used by the first inhabitants of the Hawaiian Islands (Allred *et al.*, 1999; Kempe *et al.*, 2009). Lava caves were also used as burial or religious sites (La Plante, 1992; Sinoto, 1992) and some important cultural materials still exist in many of them. Extensive studies of Polish speleologists and archaeologists in the lava caves on Easter Island (Rapa Nui) proved that use of these caves by humans was very diversified and was changing with time (Sobczyk, 2009). They became established as ceremonial objects; tombs; comfortable

night shelters; during inter-clan conflicts, they were shelters giving protection against enemies, occasionally with camouflaged entrances. They served as natural water reservoirs and the sites of arable land (*manavai*), where even today fruit trees grow.

Caves are common places of occurrence of bat guano. Such deposits occur mostly in limestone karst caves, but lava caves are not excluded, for example in Saudi Arabia (Forti *et al.*, 2004) and Australia. There is commercial mining of guano as fertilizer in Kenya (Simmons, 1998) and the USA (Crumpler and Aubele, 2001; Parmenter, 2018).

Environmental pollution and the development of tourism endanger lava cave systems, which are extremely sensitive to external influences. Halliday (2003) pointed out that surface pollutants that flow into caves or are deliberately dumped there threaten drinking water supplies, their ecosystems and cultural artifacts. Discharges into lava caves may put potable water reservoirs at risk of contamination. Where the lava pipes have collapsed, small reservoirs of stagnant water are formed, being available to local people (Kiernan *et al.*, 2003).

Lava tubes beyond the Earth

Lava caves also are widely distributed in volcanic fields on planetary surfaces beyond the Earth (Fig. 5). The formation of Hadley Rille on Moon (Apollo 15 mission) through a lava channel/tube mechanism was proposed already in 1988 by Spudis *et al.* (1988); for the Moon see also e.g., Greeley (1971) and Coombs and Hawke (1992). The famous Labyrinthus Noctis-Valles Marineris system on Mars was proposed by Leone (2014) as a network of lava tubes. Sinuous collapse chains and skylights in lunar and Martian volcanic regions often have been interpreted as collapsed lava tubes (Fig. 5B; e.g., Keszthelyi *et al.*, 2008; Sauro *et al.*, 2020). Lava tubes have been proposed to occur also on
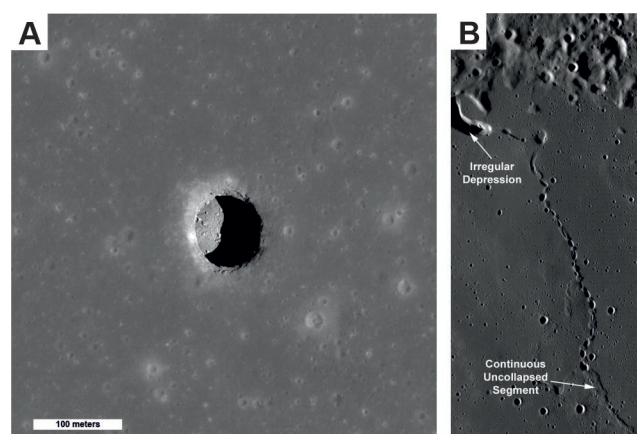
Fig. 5. Lunar lava tubes formations. **A.** Sinuous chain of collapse pits transitioning into a continuous uncollapsed segment of a lunar lava tube (copied from: http://www.nasa.gov/ mission_ pages/LRO/ multimedia/lroimages/ lroc-20110217-chain.html; author – NASA/GSFC/Arizona State University). **B.** A 100 m deep Lunar pit crater (Mare Tranquillitatis) – possible access to a lava tube (copied from: http://photojournal.jpl.nasa.gov/ catalog/ PIA13518; author – NASA/GSFC/Arizona State University).

other planets and some natural satellites: Venus (Byrnes and Crown, 2002), Mercury, Titan (Bleacher *et al.*, 2015), and Io (Crown *et al.*, 1992; Keszthelyi *et al.*, 2001). Their existence is indicated mainly by images taken of vertical holes "skylights" (Fig. 4A) by the LRO and SELENE spacecraft and by gravity data from the GRAIL Mission (http://www.nasa. gov/; Haruyama *et al.*, 2009; Modiriasari *et al.*, 2018). More than 300 of these potential cave entrances have been identified on the Moon (Wagner and Robinson, 2019) and more than 1,000 on Mars (Cushing, 2019). Interestingly, holes have been discovered on the surface of Mars by American middle school students on the basis of NASA's HiRISE photos (http://asunews.asu.edu/20100617_skylight). It should be noted that not all collapse structures, "skylights" or crater pits must be related to lava tubes. Visual similarities can lead us astray. Kempe (2017) showed deep hypogene sinkholes in Arabia, which are strikingly similar to the holes on the Arsia Mons volcano, on Mars. He suggested that the latter could be associated with permafrost collapse sinks. Therefore, simple visual interpretations should be assessed with caution. Extraterrestrial lava tubes have enormous dimensions, being 1 to 3 orders of magnitude more voluminous than the terrestrial analogues (Sauro *et al.*, 2020). The GRAIL data indicated that lava tubes can be 1–2 km wide and several hundred kilometres long (Modiriasari *et al.*, 2018). Zhao *et al.* (2017) used CTX (Context Camera) and HiRISE (High Resolution Imaging Science Experiment) images and DTM (digital terrain model) derived from them to identify the geomorphology of sinuous ridges with a lava-tube origin in Tharsis (Mars). Their lengths are estimated as varying between ~14 and ~740 km and most of them occur on slopes <0.3°. GRAIL observations and modelling show that lava tubes even 1 km wide are likely to exist and remain stable (Theinat *et al.*, 2020). Sam *et al.* (2020) studied unmanned aerial vehicle (UAV)-derived images of the Icelandic volcanic-aeolian environment and fissure volcanoes and found a number of small caves and openings. They suggest that similar openings could lead to vast subterranean hollow spaces on Mars. S. Kempe (pers. comm., 2020) contests this suggestion, considering the polar shifts of Mars and the associated dust covers of former ice caps. Dust should have clogged any of these over billions of years.

Some researchers regard lava tubes as important means of lava transportation beyond the Earth. They could have facilitated magma transportation over the relatively low topographic slopes and be important in the formation of long lava flows. They are formed at low lava viscosity, relatively low local flow rate, and sustained magma supply during a long period (Schinella *et al.*, 2011; Zhao *et al.*, 2017 and references therein). The formation processes of extraterrestrial lava tubes may have varied and could have been complex. Sauro *et al.* (2020) suggested that inflation and overcrusting processes, similar to those on the Earth, were active on Mars, while deep inflation and thermal entrenchment were the predominant mechanisms of emplacement on the Moon. Branching channel networks on Mars could have resulted from initial flow thickening, followed by the partial drainage of preferred lava pathways (Bleacher *et al.*, 2015).

Subsurface life on extraterrestrial bodies can be evaluated on the basis of analogous subsurface environments on Earth. Growing interest in the exploration of extraterrestrial planets and moons is justified by various expectations: 1) understanding of the formation of the Earth and other planets; 2) building future human bases, sheltered against hazards, such as meteorite impacts, temperature fluctuation, and seismic activity (Haruyama *et al.*, 2009; Perkins, 2020). A reasonable overburden (greater than 6–8 m) will reduce the radiation, due to high-energy particles of cosmic rays (GCR) down to an "Earth-normal" background (Hong *et al.*, 2014; Turner and Kunkel, 2017); 3) potential sites for the search for extraterrestrial organisms and finding biogenic/organic compounds (Léveillé and Datta, 2010). Lava tubes may contain groundwater or water ice deposits and provide habitable environments, both past and present (Grin *et al.*, 1998; Williams *et al.*, 2010; Schulze-Makuch *et al.*, 2015).

KAZUMURA CAVE, BIG ISLAND, HAWAII THE LONGEST LAVA CAVE IN THE WORLD

Kazumura Cave, located approximately 20 km south of Hilo (Puna District), on the Big Island (Hawaii; Fig. 6), is the longest lava cave in the world (65.5 km; Shick, 2012; Gulden, 2021). Kazumura Cave is located on the northeastern slope of Kilauea, a currently active volcano, stretching almost to the sea from the caldera. The host rocks are tholeiitic basalt of the Ailʻaʻau lava flows, which spread from the 1.5-km-long Kilauea Iki Crater (Fig. 6B), situated east of Kilauea Caldera (Holcomb, 1987). This eruption is considered the longest eruption of Kilauea in memory, with lava covering an area of about 430 km². Studies of Ailʻaʻau basalt showed that the temperature drop in lava over a 39-km section was only 4 °C. In contrast, the temperature in the cave rises gradually from Kilauea (15 °C) to the coast (22 °C; Allred and Allred, 1997). Greeley (1987) estimated that 58 % of the lava flows on Kilauea are likely to be fed by lava tubes. Within the Ailʻaʻau Lava Field, on both sides of Kazumura Cave, there are several other large lava tubes and their systems, constituting one primordial system (Halliday, 1994). The Kazumura, Keʻala and Ainahou caves, all on Kilauea, are good examples of single-trunked systems (Allred *et al.*, 1997; Kempe, 2002). Kazumura Cave also is an example of the intersection of different lava flows. Lava of the Kazumura Flow intruded into the upper and lower ends of the previously formed Keala tube in the Ailʻaʻau Flow Field (Kempe, 1999).

The cave descends down the volcano from an elevation of 1,130 m near the summit to 28 m at its lower end, giving a depth of 1,102 m for the lava cave. The depth below the ground surface does not exceed 20 m (Allred and Allred, 1997). The average slope of the cave is 1.90–1.75° with the upper sections of the cave being steeper and the lower ones less steep (Halliday, 1994; Allred and Allred, 1997). The formation of the cave and its host rocks is estimated as taking place 350–500 years ago (Holcomb, 1987).

The best source of information on Kazumura Cave, its morphology and development is the paper by Allred and Allred (1997). A large amount of affordable information
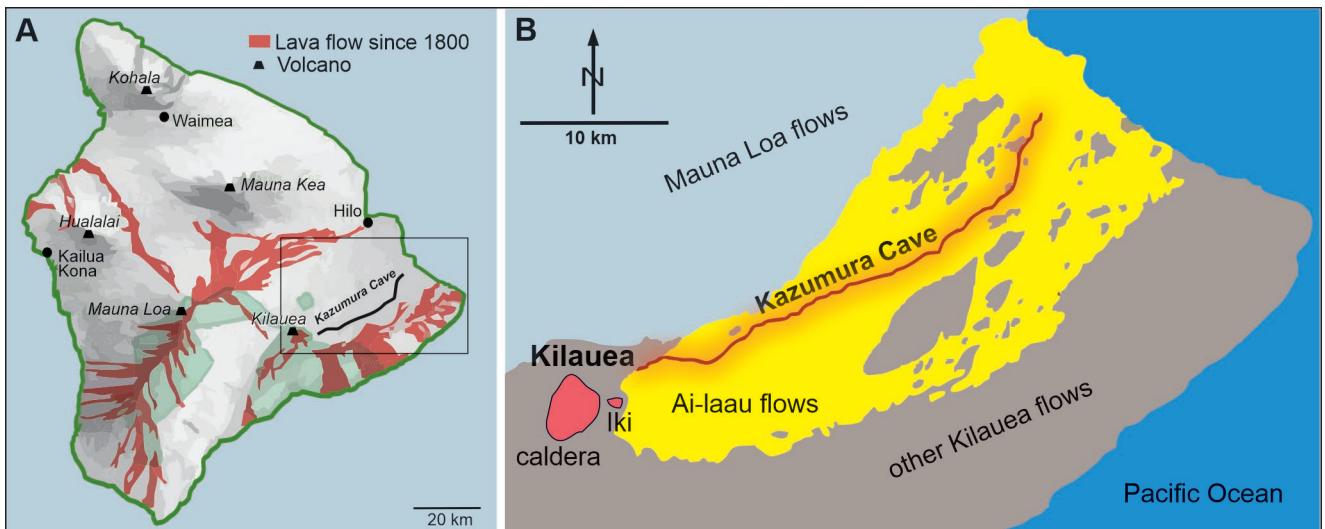
Fig. 6.   Kazumura Cave. **A.** Big Island (Hawaii) with its volcanoes and main towns. **B.** Location of Kazumura Cave (frame in A).

on lava caves, based on extensive knowledge of Kazumura Cave, also can be found in the book by Shick (2012), the guide to this cave.

The Kazumura Cave Atlas lists 101 entrances, all on private property. Most of the openings (maximum height about 3 m and width about 8 m; Fig. 7) leading to subterranean passages result from the collapse of the roof. The corridors have a maximum size of 21 m wide and 18 m high (Allred and Allred, 1997). However, their cross-sectional areas are different in different areas of the cave (Halliday, 1994). They also vary at different levels. In upper levels, they are likely to be wider and exhibit more uneven floors and walls than that of the lowest and last active level (Allred and Allred, 1997). Much of Kazumura Cave began as braided networks, which evolved into a master tube. Thermal erosion increased with the turbulence, caused by the steeper slopes. Reinsulation of the lava stream created multi-level development in spacious, downcut passages, especially below the entrances (Allred and Allred, 1997). The floors of passages usually have a ropy appearance. A ʻa lava, sometimes observed on the floor of the passages, results from the

cooling of the terminal flow, which increased the viscosity of the lava and led to the disruption of the flowing lava to form clinker (S. Kempe, pers. comm., 2020). Rafted lava blocks in the frozen residual flow are observed locally.

Many structures, common in world-wide lava caves, can be studied in Kazumura Cave, such as: lava falls (Fig. 8A), lava plunge pools (Fig. 9B), meanders, loops, multi-level passages, windows, stacked and offset balconies, bridges, sealed-off windows forming cupolas, injections, backcutting, eddy current (Fig. 9A), collapses, lava explosions, floaters, and others (Allred and Allred, 1997; Shick, 2012). In many places, accretionary linings (glazing) formed, owing to fluctuating lava levels, and are detached partly or have broken loose and allow detailed observations. Wall surfaces are especially in cul-de-sacs, very rough and uneven, popcorn-like, possibly owing to degassing (Allred and Allred, 1997). Interesting features of Kazumura Cave are black lava flows, which intruded the cave through several entrances. On the basis of the glassy skin and unmatching cracks, Allred and Allred (1997) suggested that these flows entered the cave after it had cooled.
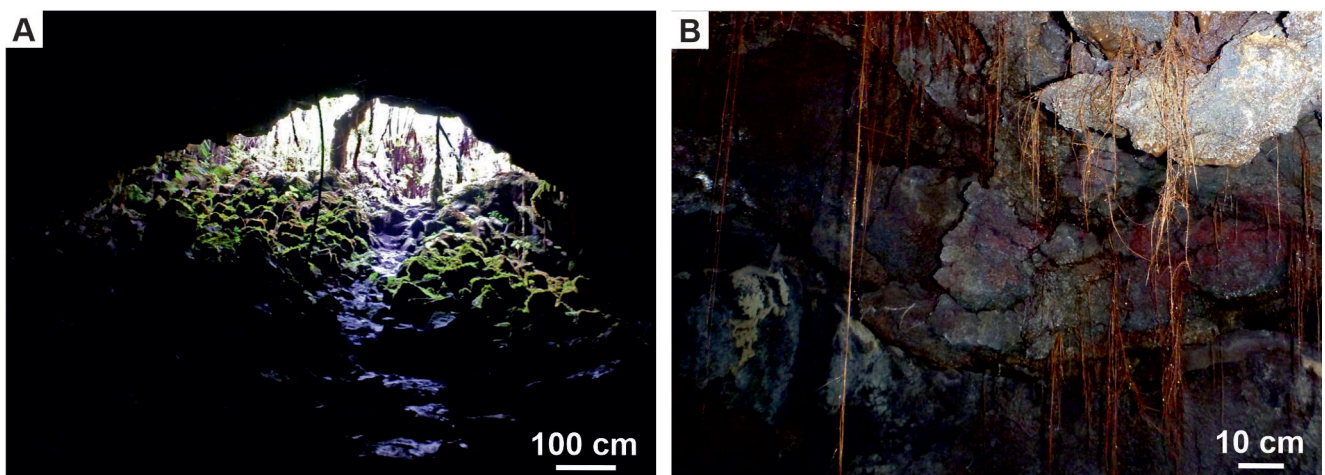

Fig. 7.   Kazumura cave. **A.** One of the exits from the cave. **B.** Tree roots in the cave, indicating short distance from the surface.

Various types of braiding are observed in different parts of the cave. Numerous undercuts and steep walls are observed locally, especially in the sharp bends of the corridors (Allred and Allred, 1997). Overlying tubes and lavafalls indicate that the cave was formed in several stages. Lavafalls are especially important for the development of Kazumura Cave. They are numerous, especially in the upper sections (Fig. 8A; Halliday, 1994). 41 falls and cascades (height from 0.9 to 13.7 m) are listed in table 3 of Allred and Allred (1997). Mature lava falls may develop undercutting and secondary widening, which resulted in cross-sections of caves that are larger than original channels (Kempe, 1997). After the migration of falls multiple levels and other accretion may form downstream (Allred and Allred, 1997).

Different types of lava speleothems (lavacicles), both on roofs and walls, can be observed: tubular and conical lava stalactites (Fig. 10B), triangular blades (knobs; Fig. 10C), soda straws (Fig. 10A), small lava stalagmites formed by fused individual drops of lava, and lava helictites (Fig. 10D), stringy lava, stretched lava, corraloids, and Pele's hair (Pele is Hawaiian *goddess* of volcanoes and fire). Secondary mineral speleothems are not common.

Variations in colour observed on the surface of the rocks in Kazumura Cave are spectacular, but finding an explanation for them is not easy. Differences in composition, oxidation, microbes and cooling rate, all can be responsible. Typical tholeiitic basalt lava is black, but with a high pyroxene or olivine content becomes green. Successive lava linings and runners (Fig. 9C) also often have different colours, indicating early changes in chemical, and perhaps mineralogical, composition, Various shades of red are usually related to oxidation. The iron in volcanic glass is oxidized to fine hematite in the outside air, while the lava is still hot. In the presence of water vapour, also goethite and limonite are formed (Kempe, 2012a). Red lavas are often found around skylights or other openings that bring fresh air into the tube. Microbial mats (Fig. 9D) and vermiculations also reveal various colours. White or yellowish crusts are composed of secondary minerals.
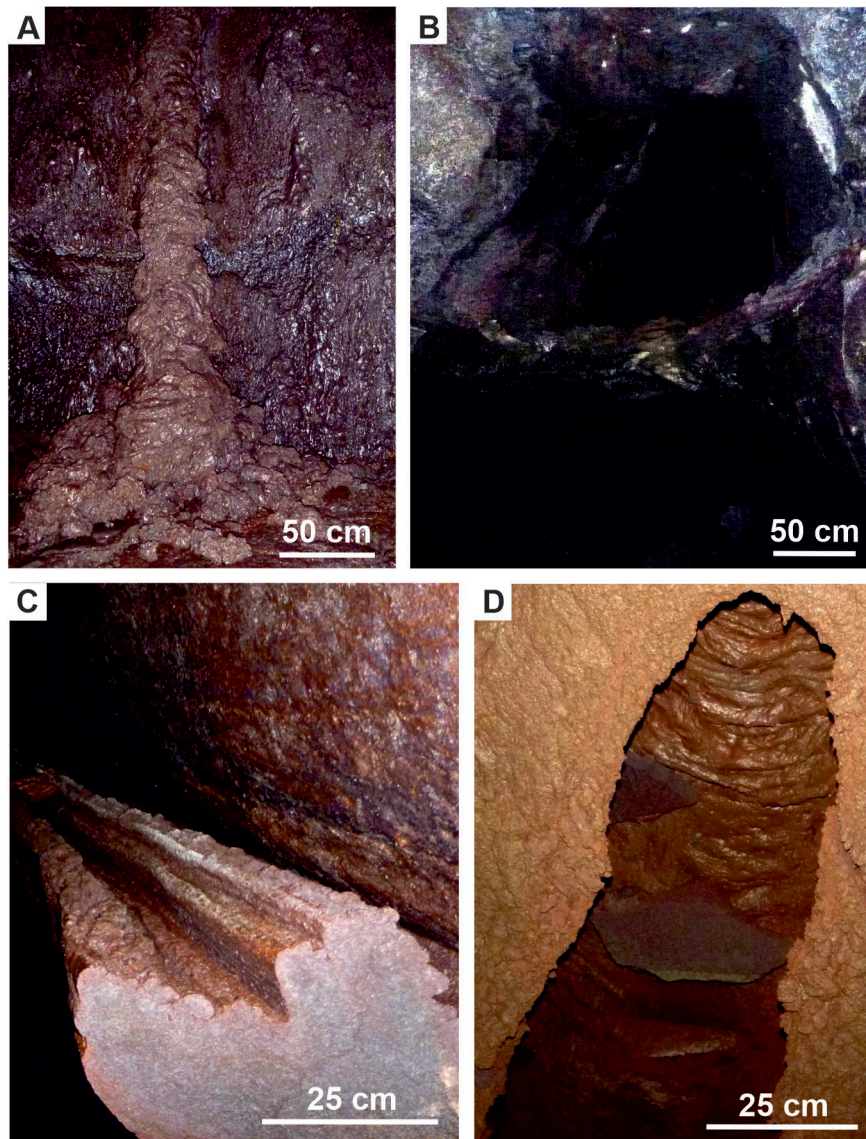

Fig. 8. Various features of Kazumura Cave. **A.** Lavafall. **B.** Two levels passages. **C.** Broken bench (levee) with two gutters. **D**. Exfoliations of thin lava linings.
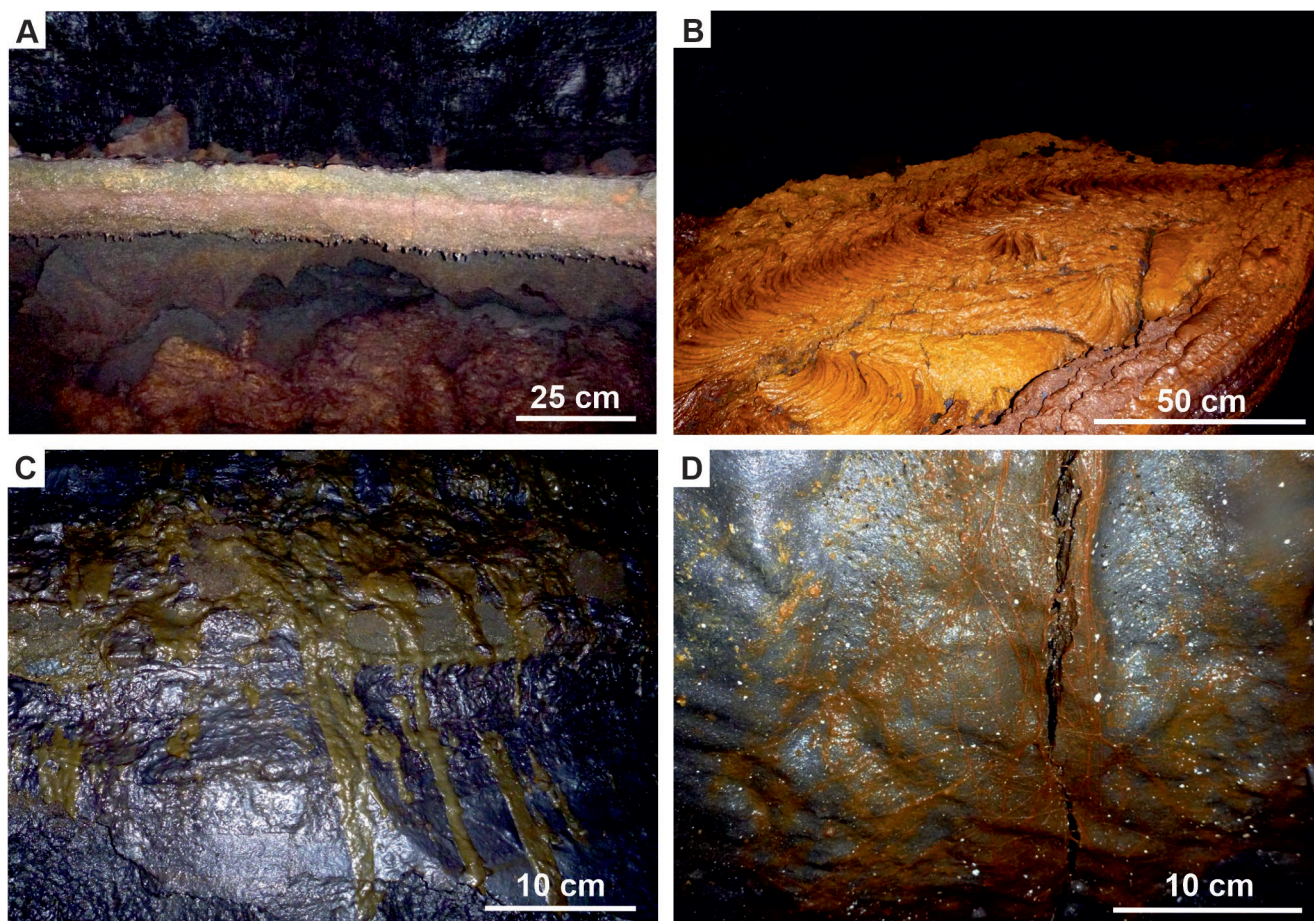
Fig. 9. Various features of Kazumura Cave. **A.** Rim (former crust) of the plunged lava pool. **B.** Frozen boiling lava texture. Note a miniature „volcano" in the middle. **C.** Lava runners (dribblets) on the wall. **D.** Microbial red-brown slime with well-visible filaments on the surface of the basalt wall.

Kazumura Cave is inhabited periodically by various species (Chapman, 1985). Native cavernicolous animals are predominantly arthropods. Ten troglobitic and seven native troglophilic (facultative cavernicoles) species where found (Howarth, 1978). Multi-coloured moulds or fungus layers occur on walls and ceilings, commonly along paths of the frequent contraction cracks (Allred and Allred, 1997).

People have used caves, including Kazumura Cave, since prehistoric times, especially its last 9-km stretch near the ocean (Allred and Allred, 1997). Many of the entrances and collapses and the locally thin roof allow pollution in many parts of the cave, both by natural processes at the surface and by anthropogenic influences. Extreme vandalism and destructive impacts have been observed over the years (Allred and Allred, 1997); fortunately, human awareness is growing and the situation is slowly improving.

Some parts of the cave are accessible with specialist equipment and a local guide, even by non-speleologists. It goes without saying that visitors to lava caves in advance should be made familiar with the information on appropriate behaviour in these caves and their preservation (see Shick, 2012). In 1994, Halliday asked several questions about the speleogenesis of Kazumura Cave. Many of them still are not answered and should provide a challenge for the future.

CONCLUSIONS

Lava tubes (pyroducts) form in volcanic flows and are found in many volcanic regions of the world. Their distribution and forms are described and their significance for the extention of volcanic flows is denoted. Thermal and mechanical erosion are important factors in their formation. Different ways of pyroducts formation are discussed, together with important parameters like temperature, viscosity, and lava flow rate in active lava tubes. Primary lava speleothems are different from those in karst caves and built of different minerals. Methods of lava tube mapping and lava temperature assessment in the Earth and beyond are presented.

Habitats of lava caves, from bacterial to human, and their mineral resources are shortly discussed. Lava caves may be potential locations for life forms and future bases for space exploration.

Basic features of lava tubes are illustrated with reference to the longest lava cave in the world, Kazumura Cave (Hawaii).
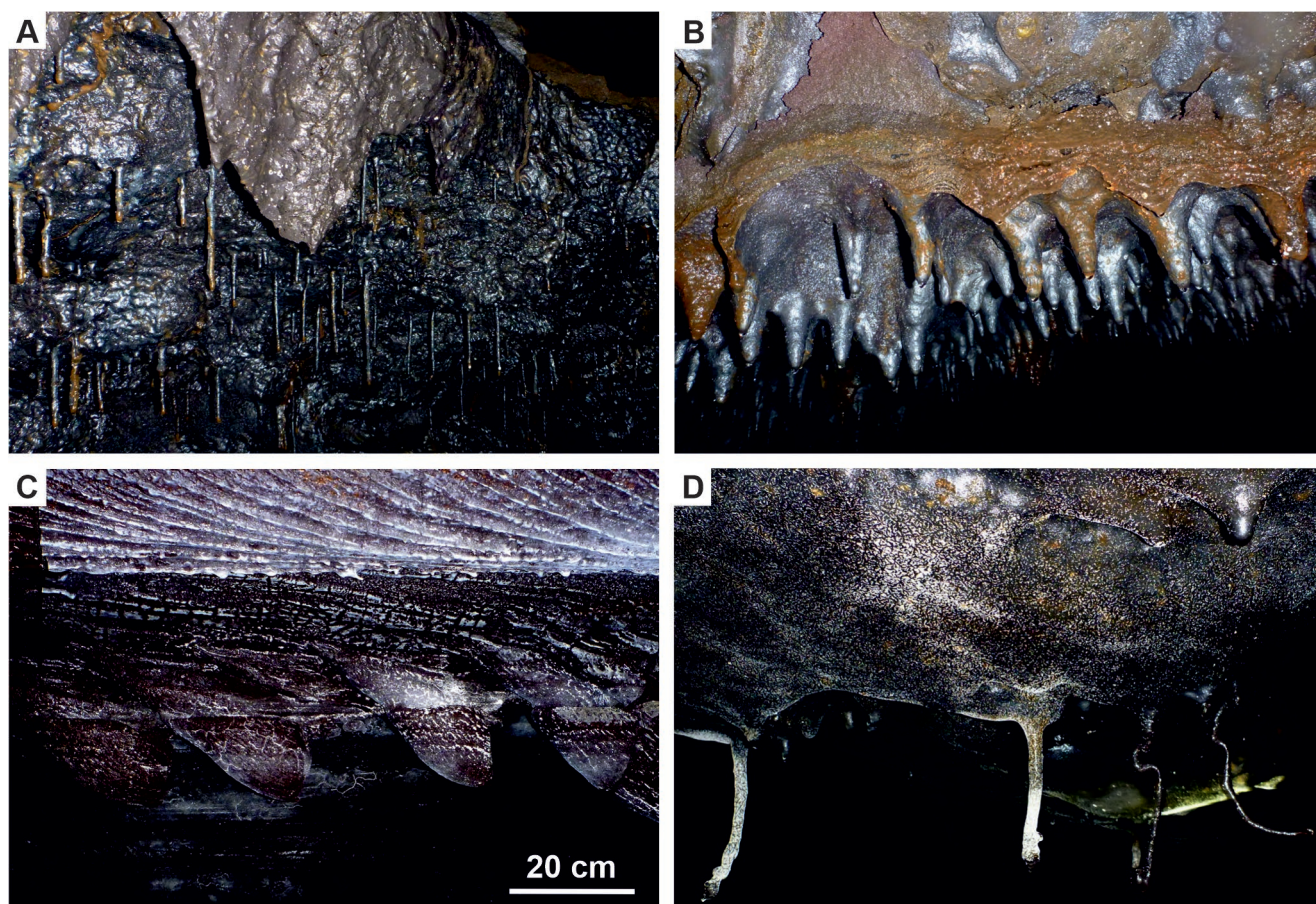
Fig. 10. Lavacicles in Kazumura Cave. **A.** Sharktooth (foreground) and tubular stalactites (width about 1–2 cm) and soda straws (width about 0.5 cm; background). **B.** Conical stalactites (length about 5–10 cm). **C.** Accreted knobs (fins/blades), resulted probably from eddy currents. **D.** Helictites (width about 0.5–1 cm).

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