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"UNDERSTANDING BASALTIC LAVA FLOW MORPHOLOGIES AND STRUCTURES FOR HAZARD ASSESSMENT"

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ABSTRACT

Lava flow surface morphologies are like pages of a book. If we are able to read the writing of that book, we can understand its content, and learn, act, and react accordingly. In the same way, if we understand lava surface morphology, recognise how it formed and the hazard it poses while flowing, we can adopt actions to protect from lava flow invasion our villages, infrastructures and local population. The surface of lava is a function of intrinsic and extrinsic qualities, and their combination results in different shapes, sizes, and complexities, as well as in different hazards. Initial sheet flows spreading at high speed have great potential for devastating land, as happened in Hawaii in May-August 2018 [Neal et al., 2018]. However, their destructive potential significantly decreases with time and distance from the vent. Conversely, lava oozing from the distal exit of lava tubes moves slowly but allows the tubes to expand, increasing gradually and slowly the potential hazard for invasion of more remote lands. In this paper, I present an overview of diverse lava flow surfaces, morphologies and structures in a framework of their generating eruptive parameters, in order to suggest preliminary but prompt hazard evaluations that could be applied during the initial phases of effusive volcanic crises at basaltic volcanoes worldwide.

1. INTRODUCTION

The final morphology of a solidified lava flow is the result of complex interactions between the lava and the environment in which it is emplaced. Lava flows display several shapes, sizes and surface morphologies, mainly determined by the physical and chemical properties of the lava, its temperature, rate of effusion, crystal and gas content, but also by the local conditions such as gravitational field strength and topography [Peterson and Tilling, 1980; Kilburn and Guest, 1993; Harris et al., 2017a, 2017b]. The main parameter influencing the morphology of a lava flow is its non-Newtonian rheology, which causes the lava to rest on a slope, although unconfined by topography, as soon as the supply ceases [Hulme, 1974]. In turn, lava rheology depends on its composition, temperature, crystal and gas contents [Giordano and Russell, 2018]. The second most important parameter influencing lava flow morphology is the rate of effusion [Walker, 1971], which defines the maximum length that a channelized lava flow can attain [Walker, 1973].

Morphologies of solidified basaltic lava flows can be broadly divided into two categories: (i) aa or (ii) pahoehoe, whereas the term "block lava" is generally used for thick brecciated lavas that are typically more silicic than basalt [Harris et al., 2017a, 2017b], and thus is not considered here. Aa is the type of lava that in solidified form is characterized by a rough, jagged, spiny and generally clinkery surface (Figure 1A). Conversely, pahoehoe lava displays a smooth, billowy, or ropy surface and normally comprises interconnected multiple flow lobes (Figure 1B). Harris et al. [2017b] offer a comprehensive review of all terms used to define lava flow morphology, describing also the many different varieties of pahoehoe and aa recognised in volcanology. For the aim of this paper, which is essentially to characterize lava flows in order to reveal their hazard, I consider only the two basic types, aa and pahoehoe. It is here worth noting that aa lava flows generally advance faster than pahoehoe, and therefore pose a greater hazard [Kauahikaua and Tilling, 2014]. Thus, a greater attention will be devoted to the emplacement of the faster and more hazardous aa lava flows.

Conversely, aa flows have vesicles that tend to be irregularly shaped, as a result of the deformation caused by movement during the final stages of solidification [Peterson and Tilling, 1980]. Even though aa lava tends to be more viscous, molten lava of approximately the same initial viscosity may form either pahoehoe or aa, with the transition from one to the second determined by a balance between viscosity and motion. Thus, in addition to the effect of increasing viscosity, lava tends to change into aa also when subjected to flow turbulence and internal shearing, such as during vigorous fountaining, pouring down steep slopes or prolonged flowage for great distances [Peterson and Tilling, 1980].

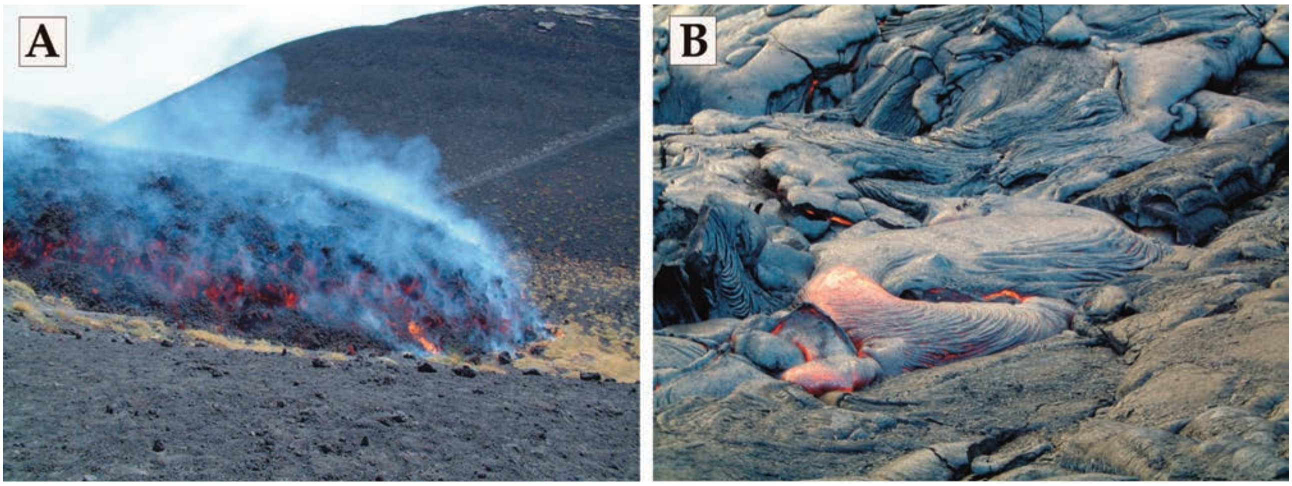

FIGURE 1. (A) Expanding aa lava flow front at Mount Etna (Italy), 18 July 2001. The width of the image is about 15 m. Photo by INGV. (B) Expanding pahoehoe flow lobes at Kilauea volcano (Hawaii). The width of the image is about 3 m.

Most Hawaiian lavas initially erupt as pahoehoe, and may change to aa downstream as they flow away from the vent [Peterson and Tilling, 1980; Lipman and Banks, 1987; Cashman et al., 1999]. This transition is caused by cooling and by increase in viscosity and yield strength during flowage [Kilburn, 1981]. At Etna, Stromboli and Fogo volcanoes instead, aa lavas are the most common surface morphology among early lava flows, with pahoehoe becoming more widespread later in an eruption when effusion rate declines [Calvari et al., 1994, 2005, 2018; Calvari and Pinkerton, 1998; Lodato et al., 2007]. Although pahoehoe and aa can be part of the same lava flow, they differ for temperature, viscosity, and vesicles shapes, with most active pahoehoe flows being less viscous and erupted at higher temperature than aa flows, and having vesicles normally as regular spheroids [Peterson and Tilling, 1980].

Pahoehoe and aa lavas have different modalities of emplacement. Pahoehoe starts as small lobes inflating to form wide and thick sheet flows several times greater than initial flow lobes [Figure 1B; Hon et al., 1994; Keszthelyi and Denlinger, 1996; Harris et al., 2007b]. Conversely, aa lavas move as a caterpillar, producing an accumulation of clinkers in the front zone overrun by fluid lava, thus forming a sort of sandwich with clinkers above and below and a fluid compact lava in between (Figure 1A). Aa lavas move by pulses and display significant variations of the magma flux, with pulses of increased flux in the front region considered as the distal result of more frequent flux changes in the vent region [Bailey et al., 2006; James et al., 2007; Favalli et al., 2010]. Pahoehoe flows also expand intermittently, with alternating phases of lobes inflation and frontal expansion [Hon et al., 1994]. As with pahoehoe, aa lava

Lava type	Volcano	Eruption	Peak TADR (m^3s^{-1})	Flow front velocity, reported	Flow front velocity m s^{-1}	Reference
Aa sheet flow	O Shima (Japan)	1951		50 km/h	13.889	Mason and Foster, 1953
Aa sheet flow	Etna (Italy)	1991-93		2 km/day	0.023	Calvari et al., 1994
Channelized aa lava flows	Piton de la Fournaise, (La Réunion)	Dec 2010	38		7.30	Soldati et al., 2018
Channelized aa lava flows	Holuhraun (Iceland)	2014-15	350	1.13 km/day	0.131	Pedersen et al., 2017
Channelized aa flows	Mauna Loa (Hawaii)	1984	806			Lipman and Banks, 1987
Channelized aa flows	Etna (Italy)	May 2001	0.7	0.29 m/s	0.29	Bailey et al., 2006
Channelized aa flows	Etna (Italy)	1981	640		1.67 0.56	Guest et al., 1987 Coltelli et al., 2012
Channelized aa flows	Etna (Italy)	1983	50		0.02	Guest et al., 1987
Aa flow front	Mauna Loa (Hawaii)	1950	1,179-1,769	5.8 miles/h	2.59	Finch and Macdonald, 1953
Aa flow front	Etna (Italy)	2006		5 m/h	5.00	Favalli et al., 2010
Pahoehoe sheet flow	Kilauea (Hawaii)	2014-15		400-500 m/day	0.006	Poland et al., 2016; Patrick et al., 2017
Pahoehoe sheet flow	Kilauea (Hawaii)	23 Jan 1988	1.1	150 m/1.25 h	0.029	Hon et al., 1994
Pahoehoe flow	Kilauea (Hawaii)	1960	58-116	30 km/h	8.333	Macdonald 1962
Pahoehoe	Laki (Iceland)	1783-84	8,700	15-17 km/day	0.17-0.20	Thordarson and Self, 1993; Guilbaud et al., 2005

TABLE 1. Peak values of lava flows emplacement rates (TADR = time-averaged discharge rate) as a function of their surface morphology.

flows are prone to inflate, forming extensive and complex lava tubes, although this process is more difficult to detect than in pahoehoe flows [Calvari and Pinkerton, 1998; James et al., 2009]. At Etna volcano, pulses during the expansion of aa lava flows [Lautze et al., 2004; Bailey et al., 2006; James et al., 2007, 2010, 2012; Favalli et al., 2010] generate characteristic surface morphologies, influence volume distribution around the lava flow field, and construct the distal, medial and proximal channel segments [Favalli et al., 2010].

The largest historic basaltic lava flow field is probably that formed in Iceland during the 1783-1784 Laki eruption, when 14.7 km^3 of lava erupted at initial instantaneous effusion rate (IER) of up to 8.7 × 10^3 m^3 s^-

[1], forming in 8 months ~600 km^2 of a mostly pahoehoe lava flow field [Thordarson and Self, 1993; Guilbaud et al., 2005]. Lava flow speed during the initial phases of that eruption was up to 15-17 km/day [Table 1; Thordarson and Self, 1993]. The lava flow field initially emplaced as thin pahoehoe lobes that gradually coalesced into larger sheet lobes [Guilbaud et al., 2005], an expansion similar to most pahoehoe lava flow fields observed today on Kilauea [Hon et al., 1994; Kauahikaua et al., 1998; Patrick et al., 2017].

At basaltic volcanoes like Etna and Stromboli (Italy), Kilauea (Hawaii), Piton de la Fournaise (La Réunion), Fogo (Cape Verde), or Holuhraun (Iceland), the general shape of a complex lava flow field is defined by a few arterial lava flows generally displaying aa texture, with its outline modified by secondary lava flows normally having a pahoehoe surface [Guest et al., 1987; Kilburn and Lopes, 1988, 1991; Calvari et al., 2005, 2018; Rhéty et al., 2017; Pedersen et al., 2017]. When lava tubes develop within complex lava flows, their hidden path is revealed by the distribution of skylights, ephemeral vents or breakouts, tumuli, shatter rings, or pressure ridges [Guest et al., 1980; Mattox et al., 1993; Calvari and Pinkerton, 1998; Kauahikaua et al., 1998, 2003; Calvari et al., 1994, 2018]. The formation of lava tubes significantly increases the hazard posed by an expanding lava flow because their insulating effect allows the flow to spread further down slope. In the following sections, I will give a brief introduction of the terms used to describe the effusion rate when calculated on different time spans, and then describe the modalities of emplacement of lava flows and flow fields, their structures and speed of formation considering several examples worldwide, in order to assess their potential hazard and suggest how to organize risk mitigation.

2. TADR, ER, IER

Harris et al. [2007a; 2017a] present a review of effusion rate definitions (Figure 2), here summarised to have clear in mind the time scale and size of the phenomena that I am considering. Following Harris et al. [2007a, 2017a], I use instantaneous effusion rate (IER) for the volume flux of erupted lava that is feeding a flow at any particular point in time; time-averaged discharge rate (TADR) for the volume fluxes over a given time period (e.g., monthly, weekly, daily); eruption rate (ER) for the total volume of lava emplaced since the beginning of the eruption divided by the time since the eruption began; and mean output rate (MOR) for the final volume of erupted lava divided by the total duration of the eruption, which can be obtained only once that the eruption is over (Figure 2). Dense Rock Equivalent (DRE) standard is desirable when comparing available data because it is independent from vesicle content, but past data and those collected from ground measurements rather than from satellite often fail to mention if the volume is expressed as bulk (including vesicles) or DRE (excluding vesicles). This is the case of most of the data in Table 1.

At the beginning of many basaltic fissure eruptions, the discharge rate increases rapidly to a maximum value, then gradually drops to lower values as the eruption proceeds and the feeder dike empties [Wadge, 1981]. This behaviour has been observed and documented at several basaltic effusive eruptions and on different volcanoes, such as Etna and Stromboli (Italy), Fogo (Cape Verde), Holhuraun (Iceland), Piton de la Fournaise (La Réunion), Kilauea (Hawaii), [Hon et al., 1994; Calvari et al., 2005, 2018; Harris et al., 2005a, 2011, 2012; Lodato et al., 2007; Spampinato et al., 2008a, 2008b; Ganci et al., 2012a, 2013, 2018; Cappello et al., 2016; Coppola et al., 2017; Patrick et al., 2017; Pedersen et al., 2017]. The general trend of declining discharge rate is overlapped by a local variability giving rise to pulses of the order of hours along the lava flow channel [Lautze et al., 2004; Bailey et al., 2005; Harris et al., 2005b], that re-

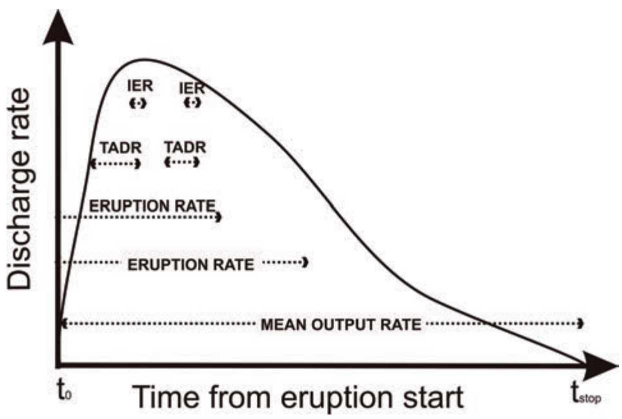

FIGURE 2. Scheme modified after Harris et al. [2007a], showing a theoretical discharge rate curve and the temporal scale at which each of the measurement definitions apply. IER = Instantaneous effusion rate; TADR = time-averaged discharge rate (weekly, daily); ER = eruption rate; MOR = mean output rate, t_0 = eruption start time; t_{stop} = eruption stop time.

flect more frequent pulses occurring at the main vent [James et al., 2007]. High effusion rate has been related to a greater crystal content of the lava (up to 70%), whereas it apparently has no effect on lava porosity, which is uniformly decreasing with distance from the vent [Soldati et al., 2018]. An important parameter affecting lava flow emplacement speed is the roughness of the topographic surface on which the flow is spreading, with greater roughness increasing flow brecciation and favouring the change from pahoehoe to aa [Rumpf et al., 2018].

3. LAVA FLOWS AND LAVA FLOW FIELDS

Many lava flows are divisible into smaller lava bodies, or flow units, each of which has a top which cooled and solidified before another flow unit superposed on it [Walker, 1971]. A lava flow field or compound lava flow is a lava body divisible into flow units (Figure 3) produced by a single vent or fissure during an uninterrupted period of effusion [Walker, 1971; Kilburn, 1996]. Superposed flow units in a compound flow are separated by an interval of time ranging from minutes to more than a year [Walker, 1971].

Effusive volcanic eruptions form either single lava flow units or compound lava flow fields, depending on the eruption duration: short-lived effusive pulses form single lava flows, also defined as volume-limited lava flows [Guest et al., 1987], whereas long-lasting eruptions promote the emplacement of complex lava flow fields [Kilburn and Lopes 1988, 1991; Calvari and Pinkerton, 1998; Calvari et al., 2005, 2010, 2018; Patrick et al., 2017; Pedersen et al., 2017]. A flow may generate major new lava streams by bifurcation, breaching and overflow of vents, channels, or tubes. Bifurcation is controlled by topography and is a random process, breaching and overflows are cooling controlled and more systematic [Kilburn and Lopes, 1988]. Thus, the growth pattern of a lava flow field may follow comparable cooling trends beneath the randomizing effect of bifurcation [Kilburn and Lopes, 1988]. A simple evolutionary sequence starts with lava flow field widening until an equilibrium width is reached, followed by flow lengthening until inhibited by its crust [Kilburn and Lopes, 1988]. With further crustal growth, flow thickening starts at the front and migrates upstream, leading to maintained thickening, breaching or overflow [Kilburn and Lopes, 1988; Kilburn and Guest, 1993]. The development of lava tubes may significantly modify this general model, promoting longer flows than the chan-

FIGURE 3. Aerial view of the active complex lava flow field during Stromboli's 2002-03 flank eruption, showing several vents (red) feeding several flow units displaying both pahoehoe and aa surfaces. Photo by INGV.

nel-fed equivalent [Calvari and Pinkerton, 1998; Kauahikaua et al., 1998].

Walker [1971] observed that whether a flow is compound or simple seems to be determined primarily by the rate of extrusion of lava at the surface, with a low rate favouring the formation of compound flows, and a high rate producing simple flows. In addition, the rate of extrusion also influences the morphology of lava flows, with aa surfaces forming in conditions of high IER, and pahoehoe flows forming at low IER [Griffiths and Fink, 1992a, 1992b; Lyman et al., 2005; Cashman et al., 2006].

The hazards posed by a single lava flow unit and by a compound lava flow field are extremely different as are their sizes. A compound lava flow field usually has a larger surface area than a single flow unit. In addition, compound lava flow fields often comprise lava tubes, which increase even more the possibility of lava to travel longer distances [Calvari and Pinkerton, 1998, 1999; Kauahikaua et al., 1998]. Long lasting eruptions fed by lava tubes develop on the flow field surface several structures like tumuli, pressure ridges, inflation clefts, ephemeral vents, skylights, toothpaste flows, squeeze-outs [Walker, 1991]. All these features suggest low emplacement rate, and conditions of a slowly expanding lava flow field with low hazard.

4. SHEET FLOWS

Sheet flows are characterised by flat surfaces and are common during the initial phases of an eruption, when IER is high and the flow spreads laterally from the vent inundating the surrounding topography [Figures 4A and B, Figure 5; Ballard et al., 1979; Lipman and Banks, 1987; Kilburn and Guest, 1993; Hon et al., 1994; Calvari et al., 2005]. At this stage, lava channel development is inhibited by the high IER, that can reach values as high as 180 $m^3 s^{-1}$ [1989 Etna eruption; Bertagnini et al., 1990], 280 $m^3 s^{-1}$ [2002-03 Stromboli eruption; Calvari et al., 2005], 350 $m^3 s^{-1}$ [2014-15 Holuhraun eruption; Pedersen et al., 2017], or even 980 $m^3 s^{-1}$ [20 August 2011 Etna paroxysm; Behncke et al., 2014]. A flow front consisting of sheet lava is often diagnostic of high advance rate, and it is possible to judge whether a flow was in a fast-advance regime based simply on a quick look at the flow front morphology [Patrick et al., 2017]. Sheet flow front advance rates of up to 2 km/day (0.023 $m s^{-1}$; Table 1) have been measured during the 1991-1993 eruption at Mt Etna on aa lava flows [Calvari et al., 1994]; up to 500 m/day (0.006 $m s^{-1}$; Table 1) during the 2014-2015 Pahoa effusive crisis at Kilauea on pahoehoe flows [Poland et al., 2016]; and of 50 km/h (13.889 $m s^{-1}$; Table 1) during the initial phases of the 1951 O Shima

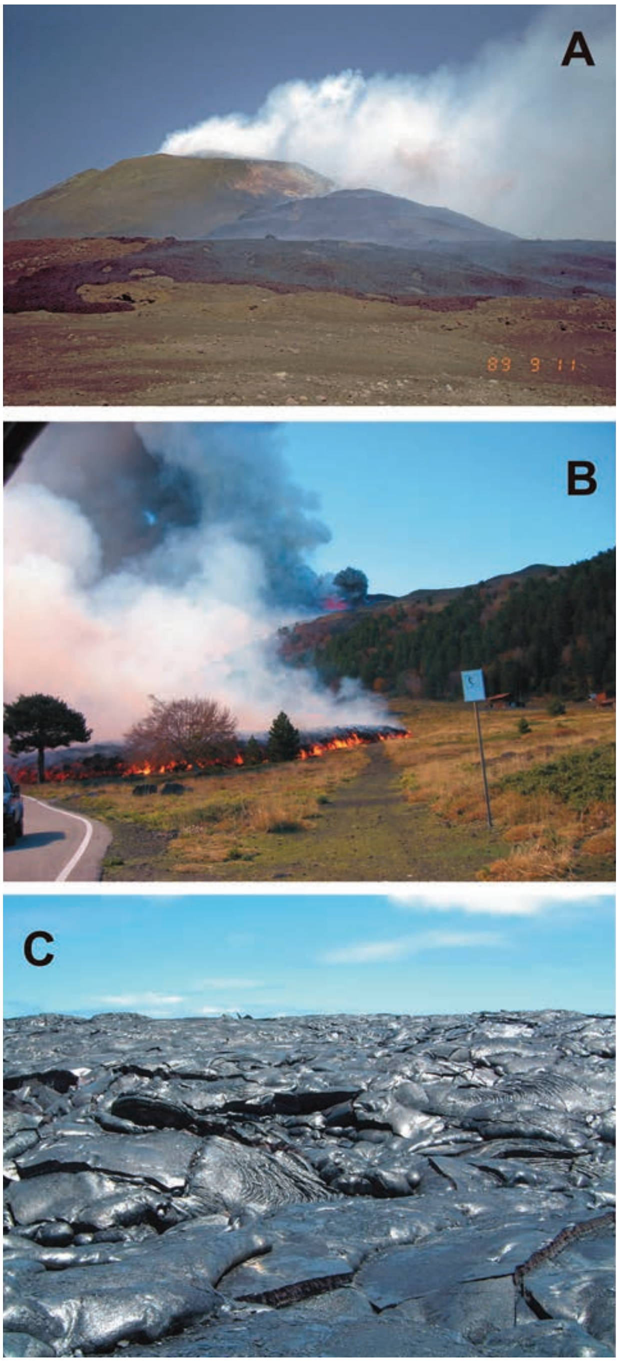

FIGURE 4. (A) Example of a sheet flow with aa surface overflowing the rim of the South-East Crater during the initial phases of Etna's 1989 eruption. (B) Aa sheet flow spreading during the 2002-03 Etna's flank eruption (foreground), while the eruptive fissure was producing lava fountains and ash plume (background). Photo by INGV. (C) Pahoehoe inflated sheet flow, Kilauea, where individual flow lobes can still be distinguished.

FIGURE 5. Cinder cone formed along the 2001 eruptive fissure, south flank of Mt. Etna, in 2001, with initial volume-limited sheet flows (at the two sides of the channel), and in the middle the main lava channel bounded by multiple levées. Photo by INGV.

eruption [Japan; Mason and Foster, 1953]. Sheet flows normally do not contain lava channels or lava tubes because they appear to be conditioned by the combination of a flat topography and a high output rate [Lipman and Banks, 1987; Hon et al., 1994]. However, long-lived pahoehoe sheet flows comprising multiple inflated lobes (Figure 4C) propagate downslope as a series of interconnected lobes that eventually can develop lava tubes [Hon et al., 1994]. Sheet flows normally last 1-2 days during the initial phases of an eruption [Lipman and Banks, 1987], and tend to develop a convex upward upper surface when slowly moving, and a nearly flat upper surface when rapidly advancing [Lipman and Banks, 1987]. The hazard posed by sheet flow expansion is very high because the spread fast, but it is normally limited to the few first days of an eruption. Inflated pahoehoe sheet flows instead advance more slowly, and can be distinguished by their fast counterpart because they display a different top surface, where initial individual flow lobes can still be distinguished (Figure 4C).

5. LEVÉES

When a lava flow spreads along a surface, cooling results in the development of regions of stationary liquid at its margins, called levées [Figure 6A; Hulme, 1974]. Sparks et al. [1976] described four different types of levées: initial, accretionary, rubble and overflow. Initial levées are formed in active lava flows by the stagnation and cooling of lava at the margins of an initial flow unit [Figure 5; Hulme, 1974; Sparks et al., 1976], and are characterized by a broad zone of marginal clinker bounding the central flowing plug. Effectively these are the rubbly flanks of the aa flow left behind after the initial flow passed through [Lipman and Banks, 1987]. Narrowing of the actively flowing central zone results in inward growth of the initial levées, an effect observed in the morphologies of Etna's proximal channel levées [Figure 5; Bailey et al., 2006]. Initial levées determine channel width, and their sizes depend on slope [Sparks et al., 1976]. Accretionary levées are slowly built up by smearing of the hot, ductile clinkers onto the sides and tops of the levées, welding together to form a solid levée [Sparks et al., 1976]. Formation of aa clinker at the flow margins by shearing and milling results in piles of clinker being gradually piled up at the margins of the flow zone [Naranjo et al., 1992]. This process is responsible for rubble levées [Sparks et al., 1976] that overlie the inner edge of the initial levée. Rubble levées are laid down by the initial flow front passage [Sparks et al., 1976] and may later become overprinted with overflow levées [e.g., Lipman and Banks, 1987; Naranjo et al., 1992; Bailey et al., 2006]. Overflow or accretionary levées are emplaced during periods of increased effusion rate, when flux exceeds channel capacity, causing the channel to overflow for a short period of time [Kilburn and Guest, 1993; Bailey et al., 2006]. An additional type of levée is called "swollen levées", and occurs when channel lava laterally intrudes the levée causing swelling, brecciation and local lava extrusion [Kilburn and Guest, 1993]. Similarly, "seeps" of viscous spiny or toothpaste pahoehoe have been observed as extruded from shield flanks or from perched lava channels [Patrick and Orr, 2012]. Active levées become static levées when they become strong enough to withstand pressure from the channelled lava, or when lateral pressure on the margins is removed by draining of the channel, and a static levée may become active if material accumulating in the channel increases the pressure exerted on the levées [Guest et al., 1987]. During waning flow, levées become accreted to the inner walls of the channel [Naranjo et al., 1992], and nested levées may form [Lipman and Banks, 1987]. Continued narrowing of the central flow zone can leave a series of abandoned rubble levées that attest to sustained inward levée growth. An additional type of levées has been observed at Stromboli, and called "excavated debris levées" [Figure 6B; Calvari et al., 2005] because the lava spreading on the steep Sciara del Fuoco slope has excavated downward and pushed laterally the cold and loose debris, thus improving channel formation. These levées can be distinguished from the other types by their lower temperature [Calvari et al., 2005].

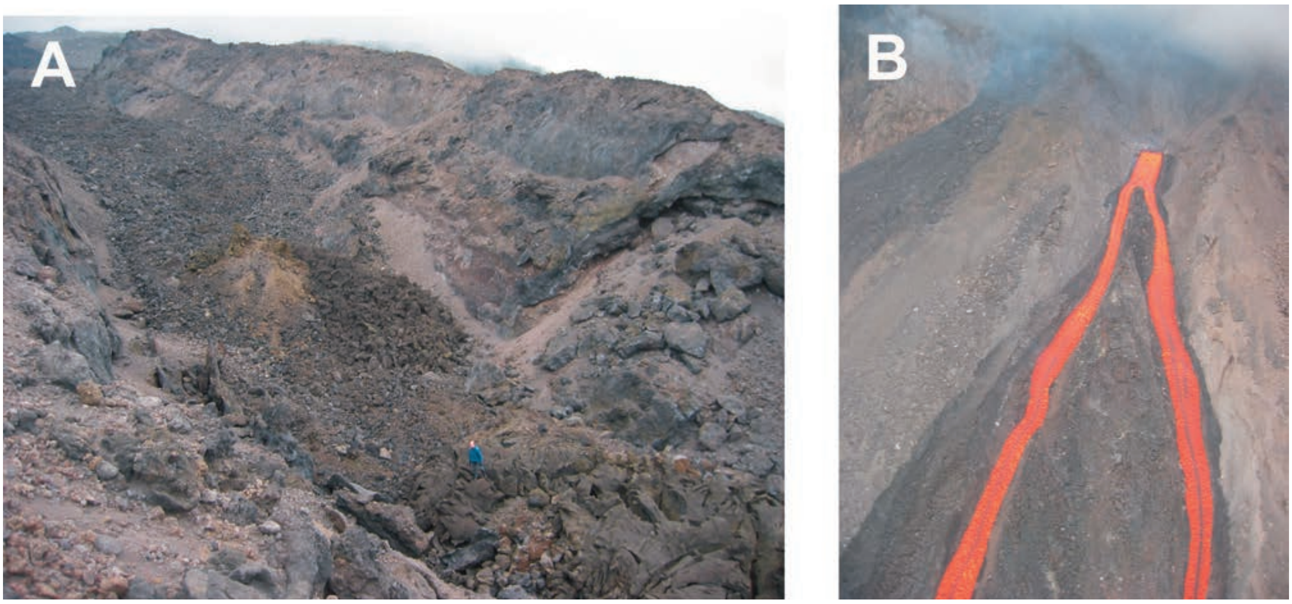

FIGURE 6. (A) Lava channel, about 30 m wide and 20 m deep, formed in aa flow during the 2001 Etna's flank eruption, bounded by high levées. Note the person with blue jacket in the middle of the channel for scale. (B) Active lava channel during the 2007 Stromboli's flank eruption, with "excavated debris levées" (light brown) around the vent and uppermost channel. The width of the vent is about 6 m.

6. LAVA CHANNELS

Lava flows are non-Newtonian or Bingham fluids, and as soon as a lava flow stabilises, it develops a channel zone. Thus the channel is a zone of flowing lava contained between static levées [Sparks et al., 1976; Lipman and Banks 1987; Kilburn and Guest 1993; Soldati et al., 2018]. Channel formation results from an increase in flow surface velocity in the axial zone of a flow, with the velocity gradient between the rapidly moving central area and the stagnating margins becoming greater with time and eventually abrupt, and is marked by an initial channel generally much wider than in the stable channel zone [Lipman and Banks, 1987]. This process is normally quite fast [at Mauna Loa in 1984; Lipman and Banks, 1987], and lava may start spreading within channels just a few hours [at Mt. Etna in 1991; Calvari et al., 1994; Calvari and Pinkerton 1998] or 24 h after the eruption start [at Holuhraun in 2014; Pedersen et al., 2017]. Lipman and Banks [1987] have classically described the architecture of channel-fed lava flow systems, recognising four zones down a channel-fed lava flow: (1) a proximal, stable channel zone where levées are well-developed and mostly static (Figure 6); (2) a medial, transitional channel zone, where the flow is channelized and bounded by incipient rubble levées still deformable; (3) a distal dispersed flow zone, where there are no levées and the lava is moving across its entire width; and (4) the flow front or flow toe, where the system is still spreading and expanding. The channel propagates downslope along the transitional and dispersed flow zones, where there are no levées present and the flow core is covered by breccia all the way to the flow front [Soldati et al., 2018]. This change confined a progressively greater volume of faster moving lava to a smaller channel cross section in an upstream direction, until the surface of the central area became more incandescent and finally consisted dominantly of fluid lava [Lipman and Banks, 1987]. Lava channels develop as a consequence of cooling, which forms as soon as a crust develops, thick enough to prevent lateral flow [Hulme, 1974]. Following Hulme [1974], the time needed to form this crust is proportional to the fourth power of the flow depth, and thus time increases rapidly with increasing depth of flow.

Pre-existing topography affects the form of the channel network in different ways, depending if we are dealing with aa or pahoehoe flows. At Mauna Loa, Dietterich and Cashman [2014] found that steeper slopes correspond to higher braiding indices on pahoehoe flows, whereas Soldati et al. [2018] at Piton de la Fournaise described structures of the aa channel network varying with distance from vent rather than with time, with single channels forming on steeper slopes and braided channels on shallower slopes. Soldati et al. [2018] observed that changes in the architecture of the aa flow are reversible because the flow can switch back and forth between single and braided channel configurations multiple times during its emplacement. In addition, Soldati et al. [2018] recognise the importance of flow rate on lava flow structures, with lava flows characterized by low effusion rates more prone to braiding compared to high effusion rate flows on any given slope. Channel networks govern the distribution of lava supply within a flow, because changes in the channel topography can dramatically alter the effective volumetric flux in any one branch, which affects both flow length and advance rate [Dietterich and Cashman, 2014]. Specifically, branching will slow and shorten flows, while merging can accelerate and lengthen them [Dietterich and Cashman, 2014], with important effects on lava flow hazard evaluation. Consideration of channel networks is thus important for predicting lava flow behaviour and possibly mitigating flow hazards with diversion barriers.

7. LAVA TUBES

A lava tube is a roofed conduit through which molten lava travels away from its vent [Figure 7; Kauahikaua et al., 1998]. Lava tubes normally form during long-lasting eruptions in both pahoehoe and aa lava flow fields [Kilburn and Guest, 1993; Calvari and Pinkerton, 1998]. At Etna, Calvari and Pinkerton [1998] described a minimum time of 4 days in order to form a tube sector within channelled aa lava flows, and considered a steady magma discharge rate as the main requirement for tube formation. In turn, the discharge rate defines the size of the tube, with longer and wider tubes being formed by higher discharge rates. Calvari and Pinkerton [1998] describe tube formations at Etna within aa lava flows in conditions of discharge rate spanning three orders of magnitude. By contrary, lava tubes within pahoehoe flows described in Hawaii apparently form only in a restricted and low range of discharge rate [1 to 5 m^3 s^{-1}, Peterson et al., 1994]. At Nyiragongo during the 2002 flank eruption lava tubes transporting lava within lake Kivu apparently formed along the arterial flow in less than a week [Komorowski et al., 2003], although it appears from the description that lava surface cooling was favoured by the contact with the lake water. At Kilauea, Hon et al. [1994] describe the process of lava tube formation within pahoehoe sheet flows in absence of lava channels and as a gradual concentration of the hot flow interior, with the flow margins cooling within hours after emplacement by edge effect, developing preferential pathways along which lava tubes will gradually develop. Lava tubes within pahoehoe sheet flows begin to form after 2-4 weeks of their emplacement, but information about this process are scant because normally, being formed on low topographic gradient, they

FIGURE 7 (A) Skylight above an aa master lava tube formed during the 2014-15 eruption of Fogo volcano, Cape Verde. Note the smooth pahoehoe slabs bounding the topographic surface and inner skylight walls resulting from late tube drainage. (B) Skylight above an active lava tube emplaced within pahoehoe inflated sheet flows, Kilauea volcano, Hawaii. Note the gas clouds from skylights that reveal the tube path.

do not drain and therefore are difficult to observe [Hon et al., 1994]. Lava tubes formed within pahoehoe sheet flows have been observed to develop with time an equidimensional section, whereas immature tubes (less than 1 week old) have widths more than one order of magnitude greater than their height [Hon et al., 1994; Kauahikaua et al., 1998].

Four processes have been recognised by which lava channels evolve into tubes: (i) inward growth of channel crust, rooted to flow margins; (ii) repeated overflow and accretion causing levées to arch and seal over channels; (iii) jamming of crustal fragments on the channel surface; (iv) and wholesale attachment of a complete channel crust to the bounding levées [Peterson and Swanson 1974; Greeley, 1987; Kilburn and Guest, 1993; Calvari and Pinkerton 1998; Kauahikaua et al., 1998].

Broad, flat pahoehoe sheet flows evolve into elongated tumuli with an axial crack as the flanks of the original flow were progressively buried by breakouts [Kauahikaua et al., 1998]. Sometimes, the tubes began to thermally erode the floor (downcutting), frequently observed through skylights, with rates of 10 cm/day [Kauahikaua et al., 1998]. This process increases the insulation of tubes making them increasingly deeper and more difficult to detect. Lava streams normally occupy only a small fraction of the tube interior, and the stream has a free surface that is primarily gravity-driven [Kauahikaua et al., 1998].

The presence of lava tubes within an active lava flow field is very important for hazard assessment because their insulation can allow lava to flow farther from the vent. In fact, a cooling of just ~1°C/km has been estimated for the lava flowing inside a lava tube [Calvari et al., 1994; Keszthelyi 1995; Kauahikaua et al., 1998; Heltz et al., 2003]. Tubes allow lava to travel for longer distances than when flowing within channels, threatening areas very distant from the active vent. The effect of lava tubes is thus similar to the downslope displacement of the effusive vent. This is the reason why the most devastating lava flow fields have involved the formation of extensive lava tube networks [Calvari and Pinkerton, 1998; Kauahikaua et al., 1998; Crisci et al., 2003; Branca et al., 2013; Patrick et al., 2017; Pedersen et al., 2017; Calvari et al., 2018; Soldati et al., 2018].

8. HAZARD IMPLICATIONS

One of the more challenging aspects for predicting lava flow dynamics is that the rheological properties of lava flows evolve during eruption and emplacement as a consequence of cooling, degassing and crystallization. This produces strongly heterogeneous flow conditions, with lava textures and morphologies that evolve both in space and time [Kolzenburg et al., 2017]. The evolution of the lava's rheology determines, for instance, whether lava advances as sheet- or channel-like flows, and also dictates its surface morphology, which in turn has great influence on heat loss from the flow [Kolzenburg et al., 2017]. In addition to changes in rheology, lava surface morphology changes also as a result of climaxing or declining discharge rate. Thus, as an eruption proceeds, we normally observe a first phase displaying the formation of sheet flows when the discharge rate is high and the flow is still free to expand laterally invading the topography that surrounds the effusive vent. This phase normally lasts from one to a few days. This stage corresponds to a great hazard for the areas surrounding the vent, especially because lava can reach high speeds (Table 1), but the hazard is limited in time.

A second stage corresponds to the formation of levées and one or more well-defined channels, confining the lava within well-established paths and driving it more efficiently to longer distances from the vent. This stage can last from weeks to years, and the greatest hazard is located at the frontal zone, where the flow has not yet established its path and is free to expand laterally. However, sudden changes in the supply rate, or upstream blockages formation and removal, can result in sudden and even large channel overflows that may invade the area around the channel. Overflows are normally short-lived, but their damage can be potentially devastating.

The hazard posed by the expanding frontal zone, being it increasingly far away from the vent, is also increasingly lower due to its decreasing speed of expansion, a result of cooling and crust formation. This stage normally offers enough time to evacuate buildings or infrastructures avoiding loss of life. With time, and if magma supply is steady enough, lava tubes may develop within the lava flow field [Calvari and Pinkerton, 1998; Kauahikaua et al., 1998]. The insulating effect of lava tubes allow lava to travel longer distances from the main vent, increasing the possibility of inflation or endogenous growth of the lava flow field [Hon et al., 1994; Calvari and Pinkerton, 1998], and resulting in increased hazard for the distal portion of the lava flow field, where breakouts can suddenly open giving rise to fast spreading secondary flows [Calvari et al., 1994, 2018]. In addition, an underestimated hazard derives from the reactivation of lava flow fronts due to the overlapping of two or more lava flow units when the lower/earlier has not yet solidified, thus allowing a combination of the molten core of the two overlapped flows [Applegarth et al., 2010].

Each country with frequently erupting volcanoes has developed proper systems to face volcanic crises, depending on the most common lava flow morphologies, speeds and features. At Kilauea volcano, where an effusive eruption is going on since 1983, the scientists of the Hawaiian Volcano Observatory (HVO) in charge of volcano monitoring and hazard assessment produce lava flow maps including potential flow paths based on topographic steepest-descent calculations [e.g., Kauahikaua and Tilling, 2014; Poland et al., 2016; Neal et al., 2018]. This is enough to forecast the possible expansion of lava, given that pahoehoe lava flows can be very fast, change rapidly directions following minor topography changes, and cause a reversal topography due to inflation [Hon et al., 1994].

By contrast on Etna, where frequent effusive eruptions normally occur every few years [Harris et al., 2011, 2012; Bonaccorso and Calvari, 2013], satellite measurements of IER [Ganci et al., 2011, 2012b] are crucial during the initial phases of an eruption [Bonaccorso et al., 2015]. These are used at first to estimate the maximum distance that a single flow unit can travel [Calvari and Pinkerton, 1998; Wright et al., 2001; Bonaccorso et al., 2015], and in the meanwhile the same parameters are used to run lava flow simulations using cellular automata models, that provide a more accurate and reliable forecast of lava flow expansion [Crisci et al., 2003; Vicari et al., 2011; Ganci et al., 2012b; Del Negro et al., 2013] as well as a probabilistic modelling of future eruptions [Cappello et al. 2011, 2012, 2013]. However, the discovery that lava tubes develop and grow even within aa lava flow fields typical of Etna [Calvari and Pinkerton, 1998; 1999] has revealed an important hazard that must be considered when dealing with long-lived effusive eruptions having a steady supply [Calvari et al., 1994; Calvari and Pinkerton, 1998; Solana et al., 2017; Calvari et al., 2018]. Reliable tube growth simulations are not available at the moment, and all we can do when lava tube presence has been detected within growing lava flow fields is to run lava flow simulation moving the effusive vents where new breakouts open at the end of the tube path [Cappello et al., 2016; Neal et al., 2018].

9. CONCLUDING REMARKS

The analysis of lava flow morphology can help evaluate the possible hazard posed by active lava flows and lava flow fields, because each lava type forms in different phases of an eruption and in different conditions of discharge rate, lava rheology and topography. The initial stages of effusive eruptions normally involve the highest rates of effusion [Wadge 1981; Harris et al., 2011, 2012], rapidly declining to a lower value and steady state when the feeder dike stabilizes and drain. Thus, the initial phases of effusive eruptions at highest flow rate normally produce sheet flows, where a thin sheet of lava spreads laterally at high speed (Table 1) around the vent [Pedersen et al., 2017]. Although very hazardous, they normally last few hours to days, and are followed by aa lava flows spreading at still high discharge rates. These soon form channels, whose levées protect the villages and infrastructures located at the two sides, but at this stage it is still possible that overflows from the main channel may invade properties and land [Neal et al., 2018]. When discharge rate declines, earlier aa flows may be covered by slower pahoehoe flows. If we can measure the IER during an effusive eruption, we can apply the simple formula proposed by Walker [1971] that relates IER and maximum flow length, in order to estimate a priori the maximum distance that a single flow unit can reach from its vent, as has been done on Etna using Walker's formula appropriately modified for Etna's lava [Calvari and Pinkerton, 1998; Wright et al., 2001; Bonaccorso et al., 2015; Solana et al., 2017]. If instead the first flows forming at high rates are pahoehoe, then we can apply the expertise of HVO and produce a map with lines of maximum steepness of the ground, in a way to forecast the path followed by pahoehoe lobes and inflated pahoehoe sheet flows [Kauahikaua and Tilling, 2014; Poland et al., 2016; Neal et al., 2018]. If the eruption proceeds, and a complex lava flow field forms, it is possible to run appropriate models simulating lava flow emplacement [Crisci et al., 2003; Cappello et al., 2011, 2016; Ganci et al., 2011, 2012b; Vicari et al., 2011]. However, we must take into account that complex and long-lived lava flow fields may promote the formation of hidden lava tubes, which allow the lava to expand with very little heat loss and spread further downslope than when travelling free on the surface or within channels [Calvari and Pinkerton, 1998; Kauahikaua et al., 1998; Calvari et al., 2018]. Because no reliable models exist at the moment for simulating the effect of lava tubes, the only possibility to evaluate lava flow invasion in these cases is to apply existing lava flow models to any new major breakouts opening at the edge of the lava flow field [e.g., Cappello et al., 2016]. Identification of inflated zones of the lava flow field and of potential ephemeral vent sites in its distal regions is critical for hazard assessment because they allow flows to lengthen significantly over their calculated cooling-limited lengths [Pinkerton and Sparks 1976; Calvari and Pinkerton, 1998]. In all conditions, the possibility to extract useful lava flow parameters from satellite images may help when ground measurements are impossible or dangerous [Ganci et al., 2011, 2012a, 2012b, 2013, 2018; Cappello et al., 2016].

The experience gained at well monitored volcanoes can be applied to those cases where a monitoring system is not well developed or is lacking, and the analysis of lava flow morphology can help evaluating the hazard when and where there is no possibility to obtain reliable lava flow mapping and effusion rate measurements from ground or remote sensing techniques. This was the case for example of the 2014-15 eruption at Fogo volcano (Cape Verde), or the 2002 Nyiragongo eruption [Komorowski et al., 2003; Cappello et al., 2016; Calvari et al., 2018].

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