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Evolution of an active lava flow field using a multitemporal LIDAR acquisition

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[1] Application of light detection and ranging (LIDAR) technology in volcanology has developed rapidly over the past few years, being extremely useful for the generation of high-spatial-resolution digital elevation models and for mapping eruption products. However, LIDAR can also be used to yield detailed information about the dynamics of lava movement, emplacement processes occuring across an active lava flow field, and the volumes involved. Here we present the results of a multitemporal airborne LIDAR survey flown to acquire data for an active flow field separated by time intervals ranging from 15 min to 25 h. Overflights were carried out over 2 d during the 2006 eruption of Mt. Etna, Italy, coincident with lava emission from three ephemeral vent zones to feed lava flow in six channels. In total 53 LIDAR images were collected, allowing us to track the volumetric evolution of the entire flow field with temporal resolutions as low as ~ 15 min and at a spatial resolution of <1 m. This, together with accurate correction for systematic errors, finely tuned DEM-to-DEM coregistration and an accurate residual error assessment, permitted the quantification of the volumetric changes occuring across the flow field. We record a characteristic flow emplacement mode, whereby flow front advance and channel construction is fed by a series of volume pulses from the master vent. Volume pulses have a characteristic morphology represented by a wave that moves down the channel modifying existing channel-levee constructs across the proximal-medial zone and building new ones in the distal zone. Our high-resolution multitemporal LIDAR-derived DEMs allow calculation of the time-averaged discharge rates associated with such a pulsed flow emplacement regime, with errors under 1% for daily averaged values.

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1. Introduction

[2] Light detection and ranging (LIDAR) technology has been extensively used to produce high-spatial-resolution digital elevation models (DEMs) on Earth and other planets [e.g., *Smith et al.*, 2001]. LIDAR is an active system that transmits very short light pulses to the ground. These are then reflected or scattered back to the instrument. A photodiode detects the returning pulses and records the travel time of the light from the scanner to the ground and back again. The travel time is used to calculate the distance between the instrument and the ground. Combining the range measurements with the direction of pulse emission (determined by an inertial navigation system and a scan mirror angle encoder) and the position of the emitter

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(determined by a differential global positioning system), it is possible to reconstruct extremely accurate coordinates (with submeter precision) for all points sampled across the surveyed surface [e.g., *Baltsavias*, 1999; *Wehr and Lohr*, 1999; *Wagner et al.*, 2006]. LIDAR can be tripod or aircraft mounted. Airborne LIDAR surveys permit generation of high-accuracy DEMs for large areas, allowing detailed and comprehensive maps of all surface features within the image.

[3] Airborne LIDAR technology has already been extensively applied in volcanology, where accurate morphometric and volumetric measurement of surface features are crucial for understanding the dynamics of lava flow and dome emission [e.g., *Queija et al.*, 2005; *Ventura and Vilardo*, 2007; *Favalli et al.*, 2009a]. Several lava flow orientated studies have been conducted by analysing a single, highspatial resolution, LIDAR-derived DEM. *Mazzarini et al.* [2005] presented a detailed morphometric analysis of an active lava channel at Mt. Etna (Italy). *Harris et al.* [2007a] used these data to model the thermorheological conditions likely associated with the observed channel-fed unit, with

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Favalli et al. [2009b] using LIDAR data to map the distal flow segment of Etna's 2001 lava flow. Likewise, *Ventura and Vilardo* [2007] used airborne LIDAR data to map the surface morphology of Vesuvius' 1944 flow and to model the emplacement dynamics. *Bisson et al.* [2009] also used LIDAR to evaluate the risk of lava invasion on Etna's east flank, with *Marsella et al.* [2009] using a LIDAR-derived DEM of Stromboli to assess lava volumes erupted during the 2007 eruption.

[4] The increasing availability of LIDAR-derived DEMs has also resulted in many studies aimed at quantifying morphostructural and volumetric surface changes in volcanic areas, some using time series of DEMs. For example, Davila et al. [2007] used LIDAR, Advanced Spaceborne Thermal Emission and Relection Radiometer, and Landsat data to identify morphological changes in the drainage system, and map lahar emplacement, at Volcán de Colima (Mexico). Csatho et al. [2008] used LIDAR to provide the first high-precision topographic map of an active crater, applying data for Erebus volcano (Antarctica), and Fornaciai et al. [2010a] used LIDAR data to map the morphology of Stromboli volcano (Italy). Neri et al. [2008] used a time series of LIDAR data to map the morphostructural changes across Etna's summit area during the past two decades, with Tarquini and Favalli [2010] quantifying the consequences of the same changes on lava flow hazard maps. Favalli et al. [2009a, 2009c] and Fornaciai et al. [2010b] also used LIDAR time series to investigate the morphology of the scoria cones on Etna's flanks, as well as to estimate volumes of tephra and lava emplaced across, and eroded from, Etna's summit area during 2005–2007.

[5] To date, LIDAR-based studies of volcanic processes have considered DEM time series with time intervals of the order of years. However, airborne LIDAR data are usually collected in multiple strips during a single survey. Each strip is acquired by flying at a constant velocity along a straight path. The surveys are flown so that they have overlapping areas between adjacent strips. These areas of overlap are acquired at two different times, usually separated by a few minutes. In this way, DEMs of dynamic features, such as lava flows, can be generated with a temporal resolution of a few minutes. Favalli et al. [2009a] began to explore this capability by using LIDAR data for a channel-fed lava flow active on Etna during 2004. By comparing the DEMs derived from the region of overlap, some insight into the temporal evolution of the lava flow field in the areas of overlap could be obtained. However, the active lava flow was captured in only three of the nine NNE-SSW strips acquired during the overflight, with significant overlap occurring in only two strips [Favalli et al., 2009a]. Based on this experience, a new LIDAR survey was flown at Etna in 2006, during another lava-producing eruption. Over 2 d, 53 overlapping strips were acquired over the active lava flow field. Repeated LIDAR overflights along the same flight path allowed generation of multiple DEMs at time intervals ranging from a few minutes to 25 h, with vertical and horizontal resolutions of less than 1 m. This, through subtracting the DEMs obtained before and after lava flow emplacement, allows precise volumetric measurements of the emplaced units [e.g., Stevens et al., 1997, 1999].

[6] Here we show how a time series of LIDAR-derived DEMs allow the emplacement dynamics of a complex active

lava flow field to be quantitatively investigated. We focus on a data sequence collected during the morning of 18 November 2006, when 10 fully overlapping strips of LIDAR data allowed us to examine a 2 h period of activity at time intervals of about 10 min. Our results show how multitemporal LIDAR data acquired for active lava flows at a high temporal resolution represent a major step in the study and quantification of morphological changes occurring at an active lava flow field resulting from channel-contained flow, channel overflow, flow pulses advancing down the channels, and the advance of flow fronts.

2. Effusive Activity at Etna and the 2006 Eruption

[7] Mt. Etna (Figure 1), located on the east coast of Sicily (Italy), has a basal diameter of about 40 km and is the highest volcano in Europe with an elevation of 3329 m [*Neri et al.*, 2008]. Between 2000 and 2006, there were five periods of eruptive activity involving two flank eruptions in July–August 2001 [*Behncke and Neri*, 2003] and 2002–2003 [*Andronico et al.*, 2005], as well as three periods of sustanined effusive activity from fractures extending from the SE crater (SEC) during January–July 2001 [*Lautze et al.*, 2004], 2004–2005 [*Burton et al.*, 2005], and 2006 [*Neri et al.*, 2006; *Behncke et al.*, 2008, 2009]. Effusive activity tends to be channel and tube fed, forming extensive compound lava flow fields predominantly of type 'a'ā as described, for example, by *Kilburn and Guest* [1993] and *Calvari and Pinkerton* [1998].

[8] Etna's 2006 eruption began late in the evening of 14 July and continued intermittently for 5 months, with details being given in Neri et al. [2006] and Behncke et al. [2008, 2009]. The first phase of the eruption lasted 10 d and was fed by a short fissure on the lower east flank of the SEC cone. The second phase began from the summit vent of the SEC on 31 August and produced intermittent overflows over the next 2 weeks, before pauses in the activity marked a transition to an episodic style of eruptive behavior. Between early October and the middle of December, about 20 paroxysmal eruptive episodes produced intense Strombolian explosions, pulsating lava fountains, tephra emission, and lava flows from multiple vents on and near the SEC cone. These episodes were accompanied by persistent lava effusion from a vent at 2800 m elevation on the upper east flank of Etna, about 1 km from the SEC. This third phase began on 12 October and ended on 14 December, with minor lava effusion also occuring between late October and late November from further vents that opened between the 3050 and 3150 m elevations to the SW of the SEC. Large fluctuations in effusion rate from the 2800 m vent were correlated with the paroxysmal episodes, and often a conspicuous increase in lava effusion and the vigor of spattering preceded the onset of a new paroxysm by several hours.

[9] The 17–18 November 2006 LIDAR survey occurred during the third phase of activity and fell in an interval between two major paroxysms which occurred on 16 and 19 November. This interparoxysmal interval was characterized by low rates of lava effusion from the 2800 m vent. By the time of the overflight, the lava flow field fed during the previous \sim 4 months of effusive activity extended \sim 4 km down the steep W slope of the Valle del Bove (Figure 1)



Figure 1. (a) Lava flow fields of Mt Etna's 2006 eruption at the time of the LIDAR survey (17–18 November 2006). Yellow area marks the southwestern lava flow field which was not active at the time of the survey; orange area marks the active 2006 eastern lava flow field. (b) Coverage of the strips acquired during the 2006 LIDAR survey: each strip is represented by a different color. The white outline marks the lava flow fields given in Figure 1a.

/ork ^a

Strip Name	Number of Points	Average Intensity	Day of Acquisition	Local Time	$\Delta t (s)^{b}$
113	2,224,098	9.5	17	10:04	
213	2,306,325	9.5	18	08:31	80,765
223	2,350,751	9.5	18	08:46	919
233	2,171,182	9.6	18	09:03	1004
243	2,245,384	9.3	18	09:18	894
253	2,318,460	9.7	18	09:34	975
263	2,529,783	9.2	18	09:49	914
273	2,602,888	9.2	18	10:04	1015
283	2,562,350	9.4	18	10:21	900
293	2,463,741	9.7	18	10:49	1674
303	2,589,118	9.4	18	11:04	914

^aStrips were acquired on the 17 and 18 November 2006. Strips acquired in the second day are separated by intervals of 10 to 18 min. All strips almost perfectly overlap and have similar point densities and average returned intensity values.

^bTime difference, in seconds, between two successive strips; e.g., from 10:04 to 08:31 (LT) is 80,765 s.

north of M. Centenari and comprised numerous overlapping lobes that showed pronounced flow channels. At the time of the LIDAR surveys, several of these lobes were active.

3. LIDAR Survey: Experimental Setup and Data Description

[10] In 2004, an airborne LIDAR survey was performed on an active lava flow at Etna, as described in *Mazzarini et al.* [2005]. This survey was originally planned to capture a complete high-spatial resolution three-dimensional map of an active lava flow. In *Favalli et al.* [2009a] two overlapping strips of the 2004 survey, acquired a few minutes apart, were analyzed and used to generate two DEMs showing the time evolution of a short portion of the active lava flow. Despite the fact that only a small portion of the 2004 lava flow was imaged by only two strips, *Favalli et al.* [2009a] highlighted the great potential of multiple LIDAR data acquisitions at active lava flows over short time intervals for providing a detailed quantification of all morphological changes.

[11] The 2004 experience opened the way for this study in which a LIDAR survey was planned to image the 2006 lava flow at a high temporal resolution (~15 min). The 2006 LIDAR survey was performed during the 17 and 18 November 2006 eruption using an Optech airborne laser terrain mapper (ALTM) 3033 laser altimeter (http://optech. on.ca). These data have nominal accuracies that are dependent on the flight elevation above the terrain, decreasing with elevation. In our case, while the flight elevation was about 4500 m at sea level (asl), the active lava field extended between 2900 and 1850 m asl elevations, so the instrumental horizontal and vertical accuracies were in the ranges of 0.8-1.35 m and 0.25-0.35 m, respectively. A detailed discussion of systematic errors associated with this instrument, together with a rigorous algorithm for their correction, can be found in Favalli et al. [2009a].

[12] The 2006 lava flow was recorded in 53 strips, five of which imaged the western (inactive) portion of the flow field with a NE-SW strip orientation, and 48 of which imaged the active lava flows moving into the upper Valle del Bove with an E-W orientation (Figure 1). Strips were collected at different times and separated by variable time intervals ranging from a few minutes to around 1 d. Two of the NE-SW oriented strips were acquired on the first day of acquisition, with the other three being collected on the

second day. They cover an area of 13 km² and include the SEC and the 2006 lava flow field emplaced on the southwest flank of Etna (Figure 1). This lava flow field was not active at the time of the survey, but very minor volumes (on the order of 10^4 – 10^5 m³) of lava were added to it during the eruptive episodes of 19, 21, and 24 November.

[13] The 48 E-W oriented strips of the active lava flow field overlapped each other for about two thirds of their width. The strips cover an area of 28 km² and included the entire 2006 eastern lava flow field, including the flows which were active during the survey, as well as the summit craters and most of the Valle del Bove. Eighteen strips were acquired during the first day and 30 on the second day. To acquire the E-W strips, the airplane flew over the active lava flow field repeatedly during the 2 d of acquisition, almost always following the same three parallel flight lines: a northern one, a southern one, and a central one (Figure 1). Work presented here is based on data from 11 E-W strips of the active flow field, the details for which are given in Table 1. We used only 11 strips acquired along the central flight path, because they cover the entire active lava field (strips obtained from the lateral flight paths cover only a part of the lava flow field). We also had to discard most of the strips acquired during the first day because they were largely affected by gas emission and so lacked good data [Mazzarini et al., 2007].

[14] Of the many factors that affect the accuracy of LIDAR-derived DEMs, the point spacing or point density (i.e., LIDAR spatial data resolution) is one of the most important. There are numerous factors affecting the actual distribution of LIDAR pulse returns. These include instrument and survey characteristics, reflectance of the terrain, and environmental conditions. The terrain and environmental conditions across a volcanic area are particularly critical for the acquisition of LIDAR data. For example, the topography over most volcanic edifices means that the distance between the sensor and the ground will vary along the aircraft flight path, as will the morphology of the terrain. In addition, different volcanic surfaces will have very different optical and textural characteristics (e.g., lava flows of different ages and morphologies, ash, tephra, and vegetation). At active systems, the line of sight may also be contaminated by the presence of volcanic plumes. All of these factors make it almost impossible to obtain a uniform point density over volcanic areas [Fornaciai et al., 2010a]. Figure 2a summarizes the point density distribution for the 2006 LIDAR



Figure 2. (a) Map of the average point density normalized for the sensor-terrain distance. The average is for all the strips listed in Table 1. For descriptions of point density zones 1 to 6, see text. (b) Normalized intensity map of strip 303. Note: There is a strong correlation between intensity of the backscattered signal and the point density.

data set. The point density is dependent on the acquisition geometry: The smaller the distance between sensor and target (terrain), the narrower the acquired strip; thus we have the same number of points over a smaller area, so the point density is higher. The average point density (number of points per square meter) is calculated for all central strips and normalized for the sensor-terrain distance. We find that the average point density increases as the lava surface becomes younger. In the case presented here, lava flows older than a few years have point densities $\leq 0.10 \text{ pts/m}^2$ (Zone 1 in Figure 2a), with the 2004 lava having a point density of between 0.10 and 0.25 pts/m² (Zone 2 in Figure 2a) and lava that is 1–2 months old (Zone 3 in Figure 2a) having 0.15–0.40 pts/m². Lava that is a few days old has 0.25–0.40 pts/m² (Zone 4 in Figure 2a), but lava about one day old has point

densities of 0.50–0.60 pts/m² (Zone 5 in Figure 2a), and active lava has 0.50-1.20 pts/m² (Zone 6 in Figure 2a).

[15] The LIDAR data not only contain quantitative topographic information (x, y, and z) for investigated surfaces but also provide data regarding the reflectance characteristics of the Earth's surface in the near infrared (NIR) portion of the spectrum. The emitted laser pulse interacts with the surface, generating backscatter, and the received signal is recorded as a function of time. The return peak amplitude, or energy of each received echo, is commonly called intensity (*I*) and is considered proportional to surface reflectance [*Höfle and Pfeifer*, 2007]. LIDAR intensities are also inversely proportional to the squared distance between the instrument and the target. For this reason, in this work, LIDAR intensities were normalized to a standard distance of 1000 m by scaling all intensities by a factor of $(d/1000)^2$, where "d" is the slant range in meters [Mazzarini et al., 2007]. A map of the LIDAR normalized intensities is given here for the last strip of the 2006 survey (Figure 2b). This shows that zones of high reflectance correlate with zones of high point density which, in turn, are associated with recent and active lava flows (cf. Figures 2a and 2b). The LIDAR spatial resolution and intensity values are strongly related, because, for a fixed acquisition geometry and environmental conditions, the spatial resolution will depend only on surface reflectance [Höfle and Pfeifer 2007]. Figure 2 shows this: Both intensity and point density decrease from the active or most recent lava flow to older lava [e.g., Mazzarini et al., 2007].

4. LIDAR-Derived DEMs and Coregistration

[16] The high spatial and temporal resolution of the 2006 LIDAR data set allows generation of an accurate time sequence of DEMs. To quantify the submeter topographic changes and to make volume flux measurements using multitemporal DEMs, all the DEMs must be matched in order to minimize the DEM difference in areas not affected by natural changes. Coregistration was achieved before deriving the DEMs, that is, by directly correcting the LIDAR data points following a procedure similar to that described by Favalli et al. [2009a]. Registration between different strips was achieved by selecting a number of tie points evenly distributed across areas around and inside (e.g., on large kipukas) lava flows that were not modified by the flows active during the investigated time period. For each tie point, using a method based on triangular irregular networks to locally reconstruct the surfaces, mismatches between surfaces were calculated in each of the three directions: x, y, and z (Figure 3). Using one strip as a reference or master image (in this case strip 213, the first strip collected on the second day), the other slave strips were coregistered to it using a rubber sheeting method: A mesh of triangles was generated from the control points using a Delaunay triangulation, and linear transformations were then used to coregister the different datasets on a triangleby-triangle basis.

[17] The coregistration procedure produced a significant error reduction in DEM difference images for regions not affected by lava emplacement (Figure 3). By way of example, Figures 3a and 3c show the DEM difference between two strips (strip 303 and reference strip 213) before and after coregistration. The uncorrected DEM difference (Figure 3a) shows high systematic errors of up to 2 m distributed over a great portion of the strip. These mismatches completely disappear in Figure 3c where the DEM difference is calculated using geometrically corrected input data. The only remaining differences in elevation between the two strips are now due to height changes resulting from emplacement of new lava between the two acquisitions and hence are coincident with the active lava flows. There are small errors across a few small regions at the edge of the surveyed region with low data point density.

[18] Reduction in residual errors, after correction, was assessed by comparing the corrected and uncorrected DEMs in areas outside the region affected by the active lavas. Coregistration reduced the RMS vertical errors from 0.26 to 0.15 m. Figure 3d shows the initial asymmetric distribution of the vertical displacement between the two strips in the raw data, indicating the presence of systematic error. After correction, this distribution shrinks to a Gaussian distribution centered on $\Delta z = 0$, due to the removal of the main systematic errors.

5. Volume Calculation and Errors

[19] DEM difference grids can be used to map surface changes and to calculate the volumes emplaced. The volume emplaced between the two times of DEM acquistion (V) can be calculated from the following [see, for example, *Coltelli et al.*, 2007]:

$$V = \sum_{i} \Delta x^2 \Delta z_i \tag{1}$$

in which Δx is the grid step and Δz_i is the height variation within grid cell *i*, that is, the height difference experienced by the grid cell at the location *i*. These values are then summed for all cells inside the area across which we want to calculate the volume changes.

[20] We find that the total volume emplaced during the LIDAR survey was 568,112 m³ (covering an area of 285,000 m²). This was emplaced over a period of 25 h to give a time averaged discharge rate of 6.31 m^3 /s over the entire period. Volumes emplaced and discharge rates over other time steps within our sampling period are given in Table 2.

[21] The standard variance propagation law, when applied to Equation (1), implies that the error estimation on the volume (σ_V) has the form [e.g., *Coltelli et al.*, 2007]:

$$\sigma_{V} = \sqrt{\sum_{i} \left[\left(\frac{\partial V}{\partial \Delta z_{i}} \right)^{2} \sigma_{\Delta z}^{2} + \left(\frac{\partial V}{\partial \Delta x} \right)^{2} \sigma_{\Delta x}^{2} \right]}, \qquad (2)$$

where $\sigma_{\Delta x}$ and $\sigma_{\Delta z}$ are the planimetric and vertical accuracies. However equation (2) has two major flaws. First, according to the definition of the errors associated with the grid cells, there is no error on the horizontal location of the cell *i*: The vertical error is the only measurable or perceivable error in the DEMs. This error may be partially attributable to horizontal errors existing in the DEM are the vertical errors [*United States Geological Survey*, 1998]. For this reason, the term depending on $\sigma_{\Delta x}$ must be dropped from equation (2). Second, equation (2) is valid only when the Δz_i values are uncorrelated. This is not normally the case when dealing with DEMs, where variations in Δz_i are, spatially, strongly correlated (Figure 4a).

[22] In general, the error on the volume is linearly dependent on the standard deviation on the height variations $(\sigma_{\Delta z})$. This can be calculated from regions where the volume has not changed (i.e., control region A_E). In our case $\sigma_{\Delta z}$ is 0.153 m, with the control region being located around our region of interest, having the same density of points as the region of interest and covering an area of 307,000 m². An upper bound on the error for the volume estimate is given by assigning each pixel the maximum possible error, giving:

$$\operatorname{Err}_{V,\operatorname{high}} = A\sigma_{\Delta z}.$$
 (3)

Figure 3. Coregistration of strip 303 to the 213 master strip. (a) The 303–213 DEM difference map before the coregistration: systematic errors of up to 2 m are evident as orange zones (see c for key). (b) The tie point distribution used for co-registration. Arrows represent the planimetric displacement calculated at each tie point location. The background map shows the lava thickness change during the entire survey period: note that tie points are located outside the area of lava flow activity. (c) The 303–213 DEM difference after the coregistration. Errors are highly reduced and now the only deviations in elevation between the two strips are due to the movement of active lava. Some small errors also remain in areas with low data point density at the edge of the surveyed region. (d) Distribution of the DEM differences between strips 303 and 213 calculated outside the area of active lava in the raw and corrected data. The asymmetric and dispersed distribution apparent in the raw data collapses into a Gaussian distribution tightly centered around 0 for the corrected data.

A lower bound on the error estimate is obtained by applying the equation for the standard deviation associated with the variance propagation for uncorrelated errors, i.e.:

$$\operatorname{Err}_{V,\operatorname{low}} = \sqrt{\sum_{i} \left(\frac{\partial V}{\partial \Delta z_{i}}\right)^{2} \sigma_{\Delta z}^{2}} = A \frac{\sigma_{\Delta z}}{\sqrt{N}}, \quad (4)$$

where N is the total number of grid cells in the sum of equation (1). For the upper bound, where all errors are assumed to be correlated among them, the ratio $\text{Err}_{V,\text{high}}/A$ is $\sigma_{\Delta z}$. For the lower bound, all errors are uncorrelated and the same ratio is a function of the number of pixels and scales as $\sigma_{\Delta z}/\sqrt{N}$. In the case of the total volume

Table 2. Total Emplaced Volumes and Time-Averaged Discharge Rates for All Channel-Fed Lava Flow Units Active Across the Eastern Lava Field (Figure 1a)^a

Strips	Time Range	Reference Figure	Δt	$\Delta t(s)$	Vol (m ³)	Error _{Vol} (m ³)	TADR (m ³ /s)	Error _{TADR} (m ³ /s)
303-113	10:04 (17/11) to 11:04 (18/11)	Figure 5c	24h 59' 34"	89,974	568,110	2690	6.31	0.03
213-113	10:04 (17/11) to 08:31 (18/11)	None	22h 26' 05"	80,765	515,160	2700	6.38	0.04
303-213	08:31 (18/11) to 11:04 (18/11)	Figure 5b	2h 33' 29"	9,209	52,960	2360	5.75	0.28
223-213	08:31 (18/11) To 08:46 (18/11)	Figure 5a	15' 19"	919	4,040	2420	4.40	2.80

^aDischarge rates are averaged over a range of time periods from 15 min to 25 h. Errors on volumes are calculated following equation (5) and errors on TADR following equation (6).

Figure 4. (a) Example of a DEM difference map showing the characteristic distribution of the errors. Errors have a correlation length of 1.87 m, forming patches with a typical dimension ~14 m². (b) Plot of the average correlation coefficient between pixels as a function of their relative distance. (c) Plot of $\sigma_V/(\Delta x^2 \sigma_{\Delta z})$ as a function of the number of pixels for various DEM pairs. The limiting cases N and \sqrt{N} are also shown (see text). This is the error as a function of the number of pixels within the area over which volume is calculated.

emplaced during the LIDAR survey we obtain an upper bound for the error ($\text{Err}_{V,\text{high}}$) of 43,750 m³ and a lower bound ($\text{Err}_{V,\text{low}}$) of 82 m³.

[23] In reality, errors are neither fully correlated nor totally uncorrelated. For DEMs, errors are spatially correlated: In our case errors have an average correlation length of 1.87 m (Figure 4). This means that, on average, errors patches have a typical dimension of 14 grid cells (Figure 4a). The error on the volume can be found using the generalized (as opposed to the standard) variance propagation formula:

$$\sigma_V^2 = \Delta x^4 \left(\sum_i \sigma_{\Delta z}^2 + \sum_i \sum_{j \neq i} \text{COV}(\Delta z_i, \Delta z_j) \right)$$
$$= \Delta x^4 \sigma_{\Delta z}^2 \sum_{i,j} \rho_{ij}, \tag{5}$$

where $\text{COV}(\Delta z_i, \Delta z_j)$ is the covariance between the height variations at grid cell *i* and at grid cell *j* and $\rho_{ij} = \text{COV}(\Delta z_i, \Delta z_i)/\sigma_{\Delta z}^2$ is the corresponding correlation coefficient.

[24] We have calculated the average correlation coefficient as a function of the distance, R, between two grid cells inside our control region, $A_{\rm E}$. Using the correlation coefficients we can calculate the error using equation (5). The plot of the $\chi = \frac{1}{\sqrt{2}}$

average values for the quantity $\sigma_V / (\Delta x^2 \sigma_A)$

$$\sigma_{\Delta z} = \left(\sum_{i \ i} \rho_{ij}\right)$$
 is

given in Figure 4c as a function of the number of pixels over which the volume is calculated. As explained above, the limiting cases are N (when all grid cells have errors which are completely correlated) and \sqrt{N} (when all grid cells have uncorrelated errors). Using the generalized variance propagation equation, the error on the total volume emplaced during the LIDAR survey (568,100 m³) is only 2700 m³, that is, less than 0.5%.

6. Morphological Evolution of a Channel-Fed Lava Flow Field

[25] The 17–18 November 2006 LIDAR survey had been preceded by a LIDAR survey on 29 and 30 September 2005. Favalli et al. [2009a] give a description of the 2005 LIDAR survey and correct the systematic errors in the initial data, achieving horizontal and vertical RMS errors for the corrected data of 0.48 and 0.16 m, respectively. Topography from this first survey provides an accurate and up-to-date surface, onto which the 2006 flow units were emplaced. The difference between the 2006 and the 2005 LIDAR-derived DEMs show that the lava flow field emplaced by the time of the 2006 LIDAR survey (Figure 1) has thicknesses up to over 10 m. Repeated surveys during 2006 also allowed us to describe and quantify the topographical changes due to the emplacement and extension of channel-fed lava flow units over a variety of time scales. Here, we analyze this evolution over three time scales: ~ 15 min, ~ 2.5 h, and ~ 1 d using the DEM difference between the strips 223-213, 303-213, and 303–113, respectively (Table 2, Figure 5).

[26] Figure 5a is the DEM difference map (strips 223–213) showing the morphological changes that occurred over a 15 min period, between 08:31 and 08:46 local time (LT) on 18 November 2006. In this image we can identify six active channels. We see that the flow of lava in each channel is highly unsteady: All the active channels contain an undulating surface (areas of increased elevation separated by areas of decreased elevation, bounded by the channel levees). This is consistent with a number of small pulses moving down the channel, with the thickness differences implying that series of lava bulges have advanced in the time between the two images.

Figure 5. Lava thickness changes at the flow field over three different time scales: (a) \sim 15 minutes, between 08:31 and 08:46 LT on the 18 November 2006; (b) \sim 2.5 h, between 08:31 and 11:04 LT on the 18 November 2006; and (c) \sim 1 d, between 10:04 LT on the 17 November 2006 and 11:04 LT on the 18 November 2006. Ephemeral vent zones marked 1, 2, and 3 located the main feeding points, with the six active channels being labelled accordingly (1.1 through 3.2). Point A in (b) and (c) marks the position of the active front in channel 1.2.

[27] Figure 5b shows the DEM difference map between strips 303 and 213 and highlights the morphological changes that occurred over a period of 2 h and 33 min, between 08:31 and 11:04 LT on 18 November 2006. The map again

reveals six active channels, as well as a number of channel overflows and smaller secondary flows. Lava is supplied by three ephemeral vent zones (vent systems 1 to 3 in Figure 5b). We term these ephemeral vent zones because they were not coincident with the original effusive vent but instead had formed at the end of a braided tube system that had developed during the preceeding weeks (a similar situation was apparent for the SE crater channel system considered by *Bailey et al.* [2006]). Ephemeral is used to stress that the location at which moving, active lava becomes visible at the surface can change in time as tube systems develop (e.g., *Calvari et al.*, 1994; *Calvari and Pinkerton*, 1998).

[28] Upslope from these three ephemeral vent zones, the DEM difference map reveals no surface changes. Vent system 1 fed two separate channel-fed flows extending up to 2 km from the vent (flows 1.1 and 1.2, Figure 5b), plus a short (~200 m long) flow extending east from the vent. This small flow was moving parallel to the master channel that fed flows 1.1 and 1.2. Channel 1.1 originates from the left side of channel 1.2, at a distance of ~230 m from vent system 1. This is probably not a simple bifurcation of master channel 1.2, but instead 1.1 looks like it is fed by a tube that emerges from beneath 1.2 (Figure 5b). The path of the northern channel (1.1) was influenced by the ~ 3 m high levees of a preexisting channel immediately to the north, with channel 1.1 following the base of this levee for most of its course. The well-formed channel section of 1.1 extends 1200 m to feed a 20 m long zone of distal, dispersed flow. Channel 1.2 is somewhat longer (2030 m) and also feeds a 90 m long zone of distal, dispersed flow.

[29] However, the flow front of 1.2 is now static, with the active portion of the flow retreating up the main channel (point A in Figure 5b). On the first day the active flow front was located 1810 m from the vent (point A in Figure 5c), and on the second day 1560 m, giving a retreat of 250 m in 22.4 h. The flow front of unit 1.1 is advancing slowly (only 5 m/h), with a number of pulses again being apparent in both channels. The uppermost pulse is the longest (~200 m long) and is at roughly the same location in both channels, extending between downflow locations of 270 m and 500 m in channel 1.1 and between 250 m and 410 m in 1.2. Pulses close to the ephemeral vents are evident in all the six active channels, with all pulses being at similar position. Further down the channels, seven shorter (≤ 40 m long) pulses are apparent in channel 1.1, and eight in 1.2. Typically each pulse forms a thickening of the active lava flow within the channel by 1.5-3.3 m, and are separated by sections along which flow levels are much lower.

[30] Vent system 2 feeds two active channels (2.1 and 2.2), which have some small overflows within 400 m of the vent (Figure 5b). The overflows typically follow the levee base for downchannel distances of 30 to 90 m. While channel 2.1 is 910 m long, channel 2.2 is 1090 m long. Both channels feed short (40 m long in both cases) lengths of dispersed flow. While the advance rate of flow front 2.1 is again very slow (only 3 m/h), that of 2.2 is much faster (advancing at an average velocity of 90 m/h). Frontal advance of flow 2.2 is described in detail in the next section. In the distal section of channel 2.1, at least three small pulses are recognizable, with no pulses being visible in channel 2.2. Vent system 3 feeds two channel-fed flows. The main flow (3.1) comprises a 1340 m long channel feeding a 70 m long section of dispersed flow. This channel contains a series of small pulses in its proximal section, plus three major pulses in its medial/distal section. The flow front is advancing at an average velocity of 20 m/h. System 3 also

feeds a second much shorter (570 m long) channelized flow (3.2 in Figure 5b). This, for almost 100 m, runs in close contact with flow 2.2.

[31] The DEM difference map for strips 303–113 is given in Figure 5c and shows the morphological and volumetric changes that took place over a ~25 h period between 10:04 LT on 17 November (strip 113) and 11:04 LT on 18 November (strip 303). It shows the construction of a compound channel-fed flow field, fed by six channels. Many of the channels follow each others levees in a generally down hill direction (modified by the existence of preexisting levee structures) to form a flow field of coalesced and overlapping levees and overflow units. While strongly positive volume gains in the medial to distal section of the flow field show this to be the main zone of emplacement and construction, the proximal sections are zones of transport in stable channels which are experiencing lower degrees of construction/deposition. Construction in the proximal sections tends to result from overflow to add volume to the levees. In constrast, deposition in the medial-distal sections also occurs along the channel behind advancing pulses, as well as at, and just behind, active flow fronts where new levees are being created.

[32] The series of panels in Figure 5 shows how LIDAR time series can be used to execute a morphological analysis of an active lava flow field at different time scales, allowing complex flow field emplacement phenomena to be unravelled. We note that from Figure 5c alone it is impossible to understand the succession of flow unit emplacement event. However, using the full time series as given in Figures 5a and 5b, the series of events and associated emplacement dynamics that led to the construction of the final compound flow field given in Figure 5c can be recreated.

6.1. Time-Averaged Discharge Rates

[33] Table 2 collates the total lava volumes emplaced over each of the time intervals separating the five DEMs, with the volume errors being calculated using equation (5). Table 2 also reports the time-averaged discharge rates (TADR), with the relative errors, for each interval. Errors in the derived TADR (ϕ) are calculated using the standard propagation formula for a ratio between two quantities ($\phi = V/T$, in which V is volume and T is time):

$$\operatorname{Err}_{\varphi} = \frac{V}{T} \left(\frac{\operatorname{Err}_{V}}{V} + \frac{\operatorname{Err}_{T}}{T} \right). \tag{6}$$

The active flow field was 2150 m long in the flight path direction. The flight time above the field was, on average, 34 s (corresponding to an average flight velocity of 63.2 m/s = 227 km/h). The times reported in Tables 1 and 2 refer to the instants when the airplane was over the center of the scene containing the active flow field. The error with which any point is imaged is therefore ± 17 s, which implies that the error in the time intervals (Err_T in equation 6) is ± 34 s. The absolute volume errors are dependent on the area over which the volume changes are evaluated, and in our case all areas are very similar so that the absolute error is, essentially, fixed. This means that large volumes have a lower percentage error than small volumes. The errors on the time averaged discharge rates through equation (6). Hence

time-averaged discharge rates (TADR) calculated for time intervals of around 1 d have extremely low percentage errors (less than 1%), thanks to the accurate strip to strip coregistration. TADRs calculated for a time interval of 2.5 h have a higher percentage error (~5%). Finally, errors on TADRs for time intervals of 15 min are affected by very high errors (over 60% in our case). Our results show that a bulk volume of ~0.6 × 10⁶ m³ was emplaced over a 25 h time period to give a TADR of 6.31 ± 0.03 m³/s for that period. TADRs given in Table 2 suggest that TADR may have been slightly lower (5.75 ± 0.28 m³/s) during the last 2.5 h.

[34] In the same way that we calculate the TADR for the whole field, we can calculate the volume rate though any section along a given channel as follows: We calculate the volume difference from that section down to the front of the flow and we divide it by the time interval. Our measurements give bulk volume changes, so if the degree of lava vesicularity changes, for example, between the proximal and distal parts of the flow or between the front and the tail of a pulse, then the volume change will not be a direct measure of actual lava mass flux: It will include the variations in the bulk volume due to vesicularity changes.

6.2. Temporal Dynamics of Pulsed Flow Emplacement

[35] Our data set allows the quantification of the temporal evolution of an advancing lava flow fed by a channel experiencing a variable supply rate, as well as analysis of topographic influences on emplacement. We focus on the distal portions of the two southernmost flow channels: 3.1 and 2.2. These were the fastest advancing flows and thus show the most evident topographic changes over the sampled time interval.

6.2.1. Dynamics and Volume of Pulses in Channel 3.1 [36] Figure 6 details the distal portion of channel 3.1 (see Figure 5b for location) showing, step by step, the passage of three rapidly advancing pulses down the channel between 08:31 and 11:04 LT on 18 November. Over 2.5 h (8291 s) the flow front advances ~20 m at an average rate of ~8.7 m/h. Behind the flow front a second pulse advances ~41 m at an average rate of ~18 m/h. A third, much more complex, pulse travels 60 m in about 1 h (3713 s) at an average rate of ~60 m/h. As already discussed, behind this pulse we appear to track a series of smaller surges.

[37] Profiles marked by black lines on Figure 6f, locate the sections for which we calculate the TADR for three time steps: 08:31-08:46 (Figure 6g), 10:06-10:21 (Figure 6h), and 08:31-11:04 (Figure 6i). The 08:31 to 08:46 LT time step (Figure 6g) shows the presence of four TADR maxima. The first two maxima relate to the inflated flow front and lowermost pulse and reveal maximum volumetric flow rates of about 0.5 m³/s during pulses, separated by periods when the TADR declines to <0.5 m³/s. The uppermost pulse comprises two closely spaced maxima (separated by a distance of 100 m). This pulse is transporting a large amount of lava at peak rates of about 1.5 and 2.5 m³/s. During subsequent time steps (Figure 6h), the amplitude of the TADR oscillations marking the flow front and lowermost pulse have decreased noticeably but are still visible. The uppermost pulse now displays a single maximum at 2.5 m³/s and is rapidly advancing. In Figure 6i we display the TADR averaged over the full 2.5 h period. We again see the three pulses, although they are now somewhat smoothed due to

the longer time averaging. The front and median pulses remain small with peak rates of less than 0.5 m³/s, while the third pulse is the largest with a peak rate of ~ 2 m³/s and a length of at least 150 m. In Figure 6i we also calculate the total volume added per unit length of the channel over the full 2.5 h long period. The two most advanced pulses are carrying/emplacing 75 m³ of lava per m, while the third is carrying about 125 m³ per m. Small fluctuations of ± 25 m³ per m are also apparent within the third pulse.

6.2.2. Dynamics and Volume of Pulses in Channel 2.2

[38] Figure 7 details the advance of the 2.2 lava flow front (see Figure 5b for location) over the same 2.5 h period. Over this period the flow front advanced ~203 m, reaching 2095 m asl, at an average advance rate of ~88 m/h. Figure 7 shows a single, large, and sustained pulse at the flow front: It is at least 100 m long with a TADR of between 2.5 and 3.5 m³/and is carrying between 75 and 125 m³ per unit length. Between 10:06 and 10:21, advance accelerates and causes the flow front pulse to extend more rapidly and increase in length to ~200 m. Distribution of the volume over a greater length causes, by conservation of mass, the local TADR to decline to 2.5 m^3/s (Figure 7h). The 2.5 h time-averaged plots (between 08:31 and 11:04 LT, Figure 7i) also show a long, single pulse comprising the active flow front. Just prior to, and during, the acceleration (Figures 7c and 7d), we note the formation of a small overflow just behind the leading (flow front) pulse. This is due to lava that spills out of the channel due to high lava levels and a local depression in the topography at this point, allowing a breach. This overflow gets left behind as an overflow levee once the pulse moves away to cut the overflow supply. Removal of this volume from the pulse further explains the decline in local TADR down the channel of this point: the volume being lost to (overflow) levee construction.

6.2.3. Flow-to-Channel Evolution During Passage of a Pulse-Fed Flow Front

[39] In Figure 8a we chart the temporal evolution of the flow cross section during the arrival and passage of the lava flow front of channel 2.2. The flow front itself is marked by a high-volume pulse comprising 3100 m³ of lava in a 40 m long and 20 m wide pulse. The flow front pulse is moving down a steep (18.5°) slope, with the flow front having a velocity of 0.026 m/s (1.5 m/min). Behind this we see the characteristic waning tail that defines a pulse (Figures 8c–8f). The flow front pulse itself has a TADR of ~3 m³/s and is followed by an almost steady TADR of ~1.5 m³/s (Figure 7g).

[40] The flow front pulse is following a V-shaped ravine in the initial terrain between the parallel levees of two, partially superimposed, older channels (black profile in Figure 8a). The ravine has lateral slopes of 22° and 14° on the right and left sides (relative to the flow direction), respectively. These are extremely effective in guiding the path of the current flow. Passage of the flow front through our reference station (A-A', Figure 8a) causes the TADR to rise from 0 to 3 m³/s in about 9 min. When the volume rate through the cross section reaches its maximum (3 m³/s) so too does the flow thickness (7.6 m) and total cross section area (116 m²). At this point, the profile is like a smooth, flattopped dome, characteristic of a zone of dispersed flow. The flow width at this point (25 m) remains constant after passage of the flow front, but with time the flow thickness

(c) 4706 s, (d) 6621 s, and (e) 9290 s. For the color scale, see the caption for Figure 5. (f) Total volumetric change over the entire time 2.5 h period (08:31 to 11:04). Black lines locate cross sections where time averaged discharge rates were calculated, as given in plots (g)-(i). (g)-(i) Variation in TADR down channel: values are plotted as a function of distance Figure 6. Volumetric changes across the distal portion of lava channel 3.1. DEM differences at (a) 919 s, (b) 2817 s, from the flow front position in the final image (i.e., at 11:04). Variation is given over two time steps: (g) 08:46-08:31 and (h) 10:21-10:06, as well as for the entire period, (i) 08:31 to 11:04. Red line in Figure 7i gives the total volume emplaced per unit length. Error bars are not reported for this measurement because errors are too small to plot.

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Figure 8. Evolution of the distal portion of the lava channel 2.2. Flow front position and volume distribution at (a) 08:46–08:31, (b) 09:03–08:46, (c) 09:18–09:03, and (d) 10:06–09:49. For color scale, see legend in Figure 5. Profile A-A' marks the location of cross-channel profile given in Figure 8e, profile B-B' marks the location of downchannel profile given in Figure 8f. (e) Temporal evolution of cross-channel profile, with preexisting surface given in black (no vertical exaggeration). (f) Temporal evolution of downchannel profile (no vertical exaggeration). Transect down B-B' at 08:46 is given by black line, and at 08:31 using dashed line.

begins to decline across the center of the flow, reaching a minimum of ~5.5 m (total cross section area = 84 m^2) about 30 min after the lava flow front had reached the cross-section location. At this point, levees have begun to form and a channel has become established, as is apparent from the profile (blue profile, Figure 8e). Fifteen minutes later the flow thickness increases to 6.8 m (total cross

section area of 103 m²; green profile of Figure 8e) and the TADR to \sim 1.5 m³/s. The flow thickness and TADR then remained roughly constant for the following hour, that is, up to the end of the survey.

[41] The channel-forming stage (blue profile, Figure 8e) reveals a lava channel with a width of 13 m (as compared to the total flow width of 25 m). The initial levee marking the

left bank had a width of 4.5 m and was only 0.3 m higher than the average level of the lava flowing inside the channel. The initial levee marking the right bank was 4.5 m wide and 1.3 m higher than the average level of the lava in the channel. During the following phase of "steady" flow, the level of the flowing lava reaches the height of the right (higher) levee and is 1 m higher than the initial right levee. This behavior is characteristic of the passage of a pulse. In this case passage of a flow front pulse is apparent from the series of longitudinal profiles given in Figure 8f, as well as the image sequence of Figures 8a to 8d. This shows that (1) the front of the pulse is abrupt and steep, behind which there is a zone of (2) high level, high volume flux flow, followed by (3) a zone of lower flow levels and volume fluxes and, finally, (4) a zone of recovery to flow levels typical of interpulse flow.

7. Discussion

[42] The main aim of this work has been to propose and describe a new methodology that can be applied to airborne LIDAR data to allow morphological analysis of active lava flows, permitting precise calculations of volume and timeaveraged discharge rate. The method is based on an analysis of topographic data collected during a series of airborne LIDAR overflights. Such LIDAR time series allows the movement and emplacement of erupted lava volumes to be tracked and quantified. The method is based on two steps, which can be adapted also to high temporal resolution terrestrial LIDAR data acquisitions which require further geometric treatment to take into account the oblique view [e.g., James et al., 2009]. The first step involves creation of a multitemporal LIDAR data set separated by short time intervals (~15 min) for an active lava flow at a spatial resolution of ~1 m. The second step involves application of accurate geometric correction and minimization of the errors to allow precise volume and TADR calculations.

7.1. Method and Precision

[43] The subtraction of high spatial resolution sequences of DEMs allows movement and advance of an active lava volume to be tracked and quantified. First, data have to be acquired at a suitable time interval, which for an advancing lava flow should be a few minutes to tens of minutes. Accurate DEM coregistration and systematic error removal then allows significant error reduction, giving RMS vertical discrepancies between different DEMs of just 15 cm. This is well below the nominal vertical accuracy of the instrument, which is between 25 and 35 cm for our flight elevations above the terrain (1600 to 2600 m). RMS vertical discrepancies between DEM pairs are usually directly used to calculate the errors on DEM-derived volumes under the wrong assumption that pixel errors are either totally correlated (thus overestimating the real error on the volume) or totally uncorrelated (thus underestimating the real error on the volume) in the region within which the volume is calculated. In this work we instead calculate the average correlation coefficient between the pixels in the two DEMs as a function of the distance between the two pixels. We use this to calculate accurate errors on DEM-derived volumes by using the generalized variance propagation formula.

[44] Volume calculations and relative errors are summarized in Table 2 and are as low as 0.5%. Error propagation is then used to calculate the errors on the derived TADRs, which can be calculated for any section down the channel. This allows spatial variation in TADR to be examined down channel, as well as through time, using the full image time series, at time scales ranging from 10 min through 2.5 h to 1 d (Table 2). Compared with most other methods, which typically have an error of ~50% [Harris et al., 2007b], calculated daily TADRs are very accurate, with percentage errors under 1%. This makes multiple LIDAR acquisitions an ideal tool for the calibration of other methods for the estimation of daily TADRs. In this regard we note that time averaged discharge rates "consider volume fluxes averaged over a given time period" which "is typically obtained by measuring the volume emplaced over a known interval, and dividing by the duration to give volume flux over that interval" [Harris et al., 2007b]. Typically this can be obtained from a post- (or syn-) eruption DEM, assuming a preemplacement surface, calculating the volume difference, and dividing by the time period over which that volume was emplaced [e.g., Stevens et al., 1997; Coltelli et al., 2007; Favalli et al., 2009b]. TADRs are more typically obtained from satellite thermal data [Wright et al., 2001; Harris and Baloga, 2009]. Our advantage is that the vertical and horizontal precision of the derived DEMs, as well as their temporal frequency and exact knowledge of the acquisition time, allows for accurate estimation of the volume difference over extremely well-constrained time periods.

7.2. Pulsed Flow Dynamics

[45] The availability of high-spatial-resolution, multitemporal, LIDAR-derived DEMs for an evolving channelfed, compound lava flow field allows the detailed study of all the complex dynamics of flow field emplacement. We here focus on spatial and temporal fluctuations in flow rate, flow front advance, and levee formation. Specifically we consider the pulsed nature of the volume flux down the channel, and the effect this has on flow dynamics, channel construction, and channel morphology.

[46] Recently interest has focused on the short-time period oscillations in volume flux that most persistently fed channel and tube systems appear to experience [e.g., Bailey et al., 2006; James et al., 2007, 2010; Harris et al., 2009]. Peterson et al. [1994] observed lava flow in tube systems active during the 1969-1974 eruption of Mauna Ulu (Kilauea, Hawaii) and noted that "as the rate of lava supply varied, the level of the stream and the rate of flow in the tube fluctuated accordingly." Such oscillations in volume flux cause variation in flow velocity and level and have also been noted in channel-fed systems at Mauna Loa [Lipman and Banks, 1987], Kilauea [Harris and Baloga, 2009], and Etna [Bailey et al., 2006]. During Etna's 1983 eruption, Frazzetta and Romano [1984] observed "almost continous oscillations ranging from 15% to 20% of the estimated effusion rate" in the main channel over a time period of several hours. Such oscillations can overwhelm the channel to build characteristic overflow levees [Sparks et al., 1976; Guest et al., 1987; Lipman and Banks, 1987], adding to the height and volume of the initial levee, as well as modifying the shape and morphology of the channel unit [Bailey et al., 2006; Harris and Baloga, 2009]. While pulses may reflect changes in the bulk volume of magma arriving at the eruptive vent and feeding the channel [*Bailey et al.*, 2006; *James et al.*, 2010], surges can also result from instabilities at blockages forming in the channel [*Guest et al.*, 1987; *Lipman and Banks*, 1987; *Bailey et al.*, 2006], as well as flow instabilities developing in a channel [*James et al.*, 2007].

[47] Here we are able to quantify the pulsing process, as well as its effect on flow morphology and flow front advance. We find the following:

[48] 1. The existence of volume pulses at similar positions in all six channels, fed by three different vent zones, indicates that pulses recorded here were due to a variation in volume flux in the master feeding system common to all six channels. We thus suggest that pulses were due to changes in the bulk volume flux of magma arriving at the master vent. They thus likely reflect a time variation in the supply rate of gas and magma to the shallow system, as has been proposed as a likely mechanism for pulse generation at Etna by *Bailey et al.* [2006] and *James et al.* [2010] and also to explain variations in magma levels and thermal emission observed during persistent explosive activity at, for example, Stromboli [*Ripepe et al.*, 2002, 2005].

[49] 2. The pulse has a characteristic form, consistent with that described by *Bailey et al.* [2006], of a steep flow front and a long waning tail. Pulses are often preceeded, and followed, by anomalously low flow levels. Passage of the pulse typically involves an increase in the volume flux by up to a factor of 5, resulting in a coincident increase in flow level. This frequently ovewhelms the channel to supply overflow that construct overflow levees and feed new secondary flows that move down and then along (parallel to) the levee base of parent channel.

[50] 3. Arrival of a pulse at the flow front causes rapid advance and formation of a characeristic flow front bulge, dome, or slug of lava in the zone of dispersed flow, behind which the stable channel rapidly develops.

[51] Pulses are a temporally and spatially common feature within channel-fed flows at Etna and generate characteristic surface morphologies. They also influence the volume distribution around the flow field, as well as the construction of distal, medial, and proximal channel segments. In our case, while construction of distal-medial segments only occurred during pulse passage (by formation of overflow units), arrival of the pulse at the flow font accelerated the construction and extension process across the distal sector.

8. Conclusion

[52] High-temporal-resolution time series of LIDAR data, especially when acquired from a synoptic perspective (as is possible from the airborne vantage point), allows precise quantification of flow-field-wide dynamics, volume fluxes, and emplacement conditions, as well as their spatial and temporal variations. Our results not only point to the potential of such data sets in allowing major advances in understanding lava flow dynamics and emplacement processes but also in understanding the complex interactions controlling the final dimensions, form, and morphology of a lava flow field. Methods and analyses such as those presented here will thus likely be fundamental in improving our ability to model and predict lava flow emplacement, with direct consequences for lava flow hazard assessment, as well as analysis of remotely sensed data for extraterrestrial flow fields.

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