How lava flows: New insights from applications of lidar technologies to lava flow studies

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ABSTRACT

Mafic lava flows are common; for this reason, they have long been a focus of volcanological studies. However, field studies of both older and active flows have been hampered by difficulties in field access; active flows are hot, whereas older flows have rough and jagged surfaces that are difficult to traverse. As a result, morphometric studies of lava flows have generally lagged behind theoretical studies of flow behavior. The advent of laser scanning (LS) (i.e., lidar, light detection and ranging) technologies, both airborne mapping (ALSM) and terrestrial (TLS), is promoting detailed studies of lava flows by generating data suitable for production of high-resolution digital elevation models (DEMs). These data are revolutionizing both the visual and quantitative analysis of lava flows. First and foremost, this technology allows accurate mapping of flow boundaries, particularly in vegetated areas where bare earth imaging dramatically improves mapping capabilities. Detailed imaging of flow surfaces permits mapping and measurement of flow components, such as channels, surface folds, cracks, blocks, and surface roughness. Differencing of preruptive and posteruptive DEMs allows analysis of flow thickness variations, which can be related to the dynamics of lava emplacement. Multitemporal imaging of active flows provides information not only on the rates and locations of individual flow lobes, but also measurement of pulsed lava transport. Together these new measurement capabilities can be used to test proposed models of channel formation, lava tube formation, rates of flow advance, and flow conditions within lava channels; they also provide new ways to assess the hazard and risk posed by lava flow inundation. Early published studies illustrate the potential of applying lidar to volcanic terrain; it is clear, however, that the availability of high-resolution digital topography is poised to revolutionize the study of mafic lava flows.

INTRODUCTION

Mafic lava flows are a persistent and widespread form of volcanic activity that, while rarely fatal, pose a common threat to communities around the world. Central to lava flow hazard assessment is the construction of probabilistic flow hazard maps and development of tools for real-time prediction of flow paths, flow advance rates, and final flow lengths. Construction of lava flow hazard maps requires accurate information on the areal distribution and temporal history of effusive activity, in addition to probable vent locations. Accurate prediction of flow paths and advance rates requires not only rapid assessment of eruption conditions (especially eruption rate) but also improved models of lava flow emplacement. Together these dual goals of lava flow hazard mapping prior to eruptive activity, and predictive modeling in response to eruption initiation, demand accurate documentation of preexisting topography, past flow volumes and areal coverage, syneruptive variations in lava flux, and an improved understanding of the controls on flow advance.

All of these needs are met by new laser altimetry (lidar, light detection and ranging) technologies that are radically changing the ways in which we view and study volcanic landscapes that are continually resurfaced by lava flows. The development of lidar using both airborne (ALSM, airborne laser scanner mapping) and ground-based (TLS) laser scanners has dramatically improved the resolution of digital topography available for lava flow research. This technology not only allows collection of better data on the spatial distribution and volumes of older lava flows, but also provides views of active flows that offer new insights into the processes that formed them. Here we first review the history, goals, and strategies of lava flow research and describe the capabilities (and limitations) of both ALS and TLS imaging of volcanic landscapes before providing an overview of recent applications of lidar technology to lava flow studies. We then discuss the potential of lidar to improve our understanding of the hazards and dynamics of mafic lava flows.

BACKGROUND

Lava flow studies have been critical to volcanology as a science, as illustrated by early founding of volcano observatories at Vesuvius, Italy (in 1841, at a time when it was having frequent effusive eruptions) and Kilimaea, Hawaii (in 1912). The frequency of mafic lava flows in these locations, and at other locations around the world (such Etna, Italy; Piton de la Fournaise, Reunion; and Iceland), and the potential for renewed volcanism near major population centers (such as Auckland, New Zealand, and Mexico City, Mexico) pose unique hazards to surrounding populations. A compilation of eruptive volumes (magnitude) and eruption rates (intensity; Fig. 1) shows that, except in unusual cases (e.g., Tazieff, 1977), historic lava flows have rarely exceeded mass eruption rates of 10⁶ kg/s (an intensity index of 9; Pyle, 2000); this contrasts with eruption rates of 10³–10⁹ kg/s (intensity index 10–12) for most explosive eruptions. Figure 1 also shows that lava flows can persist from weeks to years, and thus pose a long-lived hazard. However, the duration and frequency of effusive activity also provide unique opportunities for volcanologists to study active volcanic processes. These dual motivations, i.e., hazard and opportunity, work together to make location of frequent lava flow activity the focal point for testing and applying new innovations in technology. Here we frame the primary questions related to lava flows by reviewing both the types of information required for hazard assessment and, briefly, some of the technological
innovations that have contributed to lava flow studies. We return to lava flow hazards at the end of this review, to illustrate ways in which lidar technology, in particular, is shaping new methods of volcanic hazard and risk assessment.

### Lava Flow Hazards: Questions

From a hazards perspective, the questions related to lava flows are simple. To construct hazard maps, we need to know which regions lava flows are most likely to cover, and how frequently this will occur. Hazard maps are generated either by analysis of past eruption history (e.g., Heliker and Wright, 1992; Kauahikaua et al., 1995; Trusdell, 1995) or by simulations that include both probable vent location (e.g., Wade et al., 1994) and the historic range of eruption conditions (e.g., Crisci et al., 2010). When an eruption is in progress, hazard managers want to know where the lava flow will go, how fast it will travel, and how long the eruption is likely to last. Lava flows are confined by topography, thus flow paths can be broadly predicted from digital elevation models (DEMs) once the vent location is known (e.g., Kauahikaua et al., 1995). Rates of lava flow advance depend on the volumetric effusion rate and proximity to the vent (Kauahikaua et al., 2003; Soule et al., 2004). Flow length is also strongly dependent on eruption rate (Walker, 1973), although the slope traversed (Kilburn, 2004), steadiness of effusion from the vent (Guest et al., 1987), extent of topographic confinement (Soule et al., 2004), initial lava temperature (Riker et al., 2009), emplacement conditions (open channel or insulating tube; Cashman et al., 1999; Griffiths et al., 2003), and planform flow geometry also affect flow length. To date, models of flow coverage and dynamics have been limited, in part, by the resolution of both preeruptive and posteruptive digital topography. Here we show ways in which high-resolution lidar-generated DEMs can improve our understanding of how lava flows resurface the landscape.

### Technological Advances in Lava Flow Mapping: Short History

Studies of lava flows have advanced in tandem with technological advances. For much of human history, accounts of lava flow behavior and impact were based on close observation, and ranged in form from the rich oral traditions of the Hawaiians (e.g., Cronin and Cashman, 2007; Swanson, 2008) to the detailed drawings and careful descriptions of seventeenth and eighteenth century activity at Italian volcanoes Etna and Vesuvius (e.g., Scarth, 2009). The early nineteenth century saw the addition of both a theoretical framework for Earth history (deep time) and technological advances in drafting and mapping techniques that propelled both the Scrope and Poulett (1825) study of the volcanoes of central France, and Lyell's (1830) compilation of observations of volcanic activity at Etna and Vesuvius (e.g., Rudwick, 2008). By the late nineteenth century, systematic observations of frequent lava flows from Mauna Loa volcano, Hawaii, provided a coherent observational framework for the development of models of lava flow behavior (e.g., Dana, 1891). Airplanes proved the first major technological advance to lava flow studies, when aerial overviews (first used in the 1919 eruption of Mauna Loa) gave volcanologists new perspectives on the plan-view complexity of individual lava flows. Starting in the 1980s, routine helicopter support allowed volcanologists to create detailed maps of flow advance (e.g., Wolfe et al., 1988) to be supplemented with repeat measurements of flow velocity and effusion rates from stations located along persistent lava channels (e.g., Lipman and Banks, 1987; Calvari et al., 1994) and robust lava tube systems (e.g., Kauahikaua et al., 1996).

The past few decades have seen an explosion of new tools that have been applied to lava flow studies. The wide availability of global positioning systems (GPS) provided detailed maps of flow extent, topographic maps gave way to DEMs, and satellites carried instruments that allowed remote sensing of effusive activity. Of the satellite-based imaging capabilities, the most thoroughly utilized for lava flow studies have been satellite systems that detect short wavelength infrared signals such as GOES (geostationary satellites), ATSR (along track scanning radiometer), AVHRR (advanced very high resolution radiometer), and MODIS (moderate resolution imaging spectroradiometer); for reviews, see Oppenheimer, 1998; Wright et al., 2004). These satellites provide images with low spatial (1-4 km/pixel) but high temporal resolution, and therefore can be used for whole flow field monitoring. Although flow widths are commonly smaller than individual pixels, the intensity of thermal emissions records the fractional pixel areas occupied by active flows, allowing flow area to be determined and converted to effusion rate, given sufficient calibration data (e.g., Pieri and Baloga, 1986; R. Wright et al., 2001). Higher resolution thermal imaging data can be obtained using airborne (e.g., Realmuto et al., 1992) and hand-held (e.g., Harris et al., 2005; Ball and Pinkerton, 2006) thermal imaging cameras that provide high spatial resolution, but usually at the expense of both spatial and temporal coverage.

Satellite-based radar images have the advantages of both seeing through cloud cover and having higher resolution than satellite-based thermal imaging techniques. Radar-generated DEMs (e.g., from SRTM, Shuttle Radar Topography Mission; http://www2.jpl.nasa.gov/srtm/) typically have a horizontal resolution of 90 m (although 30 m resolution is possible), which limits applications of DEM analysis to changes on the scale of the volcanic edifice (e.g., Wright et al., 2006; Hugel et al., 2008). Radar correlation imaging (e.g., using SIR-C, Space Shuttle Imaging Radar), in contrast, provides image resolution sufficient for monitoring individual lava flows (e.g., Zebker et al., 1996; Deittrich et al., 2012), as well as postemplacement flow volumes (e.g., Stevens et al., 1997; Lu et al., 2003) and cooling-induced subsidence (Stevens et al., 2001). Airborne interferometric radar, such as TOPSAR (topographic synthetic aperture radar; e.g., Zebker et al., 1992), provides sufficient resolution (1-2 m vertical) to measure the thickness of individual lava flows (e.g., Evans et al., 1992), although Mourgues-Mark and Garbeil (2005) recommended combining TOPSAR with lidar data to obtain sufficiently high resolution DEMs for lava flow modeling.

Ground-based radar has a range of a few kilometers and can measure topographic changes that exceed the instrument resolution of ~5 m; to date, ground-based radar techniques have been used primarily for monitoring changes in slow-moving blocky flows (Macfarlane et al., 2006) and lava domes (Wadge et al., 2008).

### Lidar (Light Detection and Ranging)

By the end of the twentieth century, this technological explosion included implementation of laser scanning (lidar) imaging, which was
fueled not only by instrument development but also by critical improvements in the computational infrastructure required to collect and analyze large data sets (including the National Science Foundation–sponsored National Center for Airborne Laser Mapping). Airborne laser swath mapping (ALSM) scans the ground surface using laser pulses centered on the flight path of the plane; the scan direction is perpendicular to the flight path and the swath width is a function of distance from the ground. Conversion of the laser pulse returns to x-y-z points generates a point cloud that can be filtered and gridded to produce a high-resolution DEM. Details of ALSM can be found in Shan and Toth (2008); recent overviews include Jaboyedoff et al. (2010), Glennie et al. (2013), and Roering et al. (2013). It is important that, depending on the filtering, the DEM can show either bare earth elevation or the uppermost (first return) surface, which can be used to map either urban or forest structure (e.g., Bisson et al., 2009; Deligne et al., 2012).

Most ALSM systems operate at near-infrared frequencies with significant absorption in water so that data cannot be collected through clouds; areas with persistent cloud cover are therefore problematic and require either additional survey time or pairing with radar imaging techniques to ensure complete coverage. The resolution of swath mapping using lidar systems is strongly dependent on the accuracy of the aircraft GPS control and inertial measurement unit, the elevation of individual flight lines, and the extent and density of ground cover (vegetation). Of these, location accuracy is most important, particularly in referencing side-by-side flight swaths (e.g., Favalli et al., 2009a); systematic relative swath errors can be minimized by flying some flight lines perpendicular to the main flight line direction (Latypov, 2002) and by applying iterative closest point matching algorithms to overlapping swaths (Kumari et al., 2011). ALS lidar DEMs used in studies of young (unvegetated or sparsely vegetated) lava flows are typically gridded at <1 m, a substantial improvement over 30 or 10 m DEMs commonly used in current lava flow hazard simulations.

Terrestrial laser scanning can provide complementary three-dimensional (3D) data to airborne laser swath mapping (e.g., Petrie and Toth, 2008a, 2008b, 2008c). Depending on the range to the target, pulse rate of the laser, and reflectance of the surface, collecting TLS data at subcentimeter point spacing with subcentimeter accuracy is fairly routine. Commercially available TLS systems have a wide range of maximum target distances (from a few meters to >1.5 km) and precisions are typically reported as subcentimeter, although precision depends not only on target reflectivity but also on the range to the target, because beam spreading can degrade precision at large ranges. The benefits of TLS over ALS are its ability to resolve smaller length scale features (decimeter to centimeter), its ability to image in locations that are inaccessible to ALS (e.g., vertical faces and/or outcrops), lower cost of operation, greater portability, and potential for rapid deployment. As with ground-based radar, the primary limitations of TLS relative to ALS are its much more restrictive spatial scale and the difficulty in achieving high-incidence angles to the imaging target. The latter limitation means that acquiring complete coverage of a surface with TLS may require collecting multiple scans from different vantage points to fill in line-of-sight shadows. In such cases, scans are registered relative to each other using surface matching algorithms (e.g., Besl and McKay, 1992) where sufficient overlap between scans is available, or by georeferencing individual scans to a common geodetic reference frame. In both cases, it is advantageous to use a network of widely distributed surveying targets that can be imaged from multiple vantage points and used to coregister individual scans to a coherent geodetic reference frame (e.g., Buckley et al., 2008).

Looking to the future, opportunities for innovative digital terrain model (DTM) analysis will only continue to grow as new lidar acquisition and analysis techniques are developed. For example, whereas many commercial lidar systems measure only the first and last return of each laser pulse, new systems record the entire waveform of the energy pulse that is backscattered from the reflecting surface. Although currently used primarily for forest research, full waveform lidar offers the possibility of obtaining more information on both the geometry and reflectance of illuminated surfaces (e.g., Bretar et al., 2008; Mallet and Bretar, 2009). In addition, airborne lasers are now being developed for measurement of bathymetry; water-imaging systems use a short-wavelength (blue-green) laser capable of penetrating the water-air interface (C.W. Wright et al., 2001; Kinzel et al., 2007; McKean et al., 2008; Glennie et al., 2013). Although restricted to water depths of <10 m, this technique has the potential to image flows erupting in shallow water, or, after emplacement, the shallow subaqueous extent of lava (e.g., lava dams and lava deltas), which is often ignored in flow volume measurements.

It is important to discuss some of the challenges involved with lidar work. First, obtaining airborne lidar data is expensive. However, a single data set can often be used for a range of applications, and therefore can form the focal point of collaborative projects. In addition, with time the general availability of lidar data will certainly increase, particularly through data sharing portals such as the National Science Foundation’s OpenTopography (http://www.opentopography.org). Another challenge is the huge numbers of individual x-y-z values (often in excess of 10^3) in point clouds from typical ALSM or TLS surveys; such large data sets can present significant computational challenges. This data management challenge is exacerbated for the enormous data sets generated from new-generation full waveform lidar systems (Bretar et al., 2008). However, for most users, working with segments of larger data sets, or DTMs, and choosing point spacing appropriate for the features of interest, can mitigate these challenges. In addition, data analysis tools are widely available (e.g., through OpenTopography), and many students now come to the geosciences with geographic information system (GIS) skills.

**Classification of Main Applications of Lidar to Lava Flow Studies**

The application of lidar-generated DEMs to lava flow studies is a very recent phenomenon; to our knowledge, the first lidar-based lava flow publications appeared in 2005. In subsequent years the use of lidar has slowly increased as more research groups obtain access to these data. Here we illustrate many important questions in lava flow research that can be addressed using high-resolution digital topography. For the purpose of this overview, we classify applications of lidar-derived high-resolution topography (both ALS and TLS) as follows: (1) mapping solidified flows; (2) monitoring active flows; (3) modeling lava flow emplacement; and (4) hazard mapping and risk assessment. This classification allows us to review the primary areas of lava flow studies from the perspective of questions that can be addressed using very high resolution digital topographic data. We focus primarily on subaerial mafic lava flows because (1) most applications of lidar to lava flow studies to date have been for mafic flows, (2) the frequency of mafic eruptions at places like Hawaii and Mount Etna both provide the opportunity for testing new models and new technologies, and (3) frequent resurfacing of these regions requires frequent remapping of the volcanic edifice to obtain up-to-date base maps for hazard assessment. However, we also include examples from intermediate composition lava flows, and from locations other than Hawaii and Etna, to illustrate the full range of lidar-based techniques that can be applied to lava flow problems.

**MAPPING SOLIDIFIED LAVA FLOWS**

Challenges to accurate lava flow mapping are both technological and intrinsic to the nature of lava flow resurfacing processes. Intrinsic
problems include the decrease in the spatial and the temporal accuracy of mapped lava flows with increasing flow age and distance from the vent. Because lava flows typically erupt from localized vent areas, proximal parts of older flows are covered by younger flows. For this reason, accurate planimetric maps of lava flow coverage in near-vent regions are available only for the most recent or historic flows. Although some sense of the frequency of lava flow coverage can be obtained from studies of drill cores (e.g., Katz and Cashman, 2003; Stolper et al., 2009), assessments of expected flow coverage must rely on (1) past rates of surface coverage (e.g., Behnke et al., 2005; Wright et al., 1992); (2) simple probability models of surface coverage through time (Kauahikaua et al., 1995); (3) mapping of lava sheds (i.e., drainage patterns that will confine flow movement; Kauahikaua, 2007); or (4) Monte Carlo simulations of future flow paths (e.g., Favalli et al., 2009b; Crisci et al., 2010). There are several mapping challenges, however, that can be addressed by lidar, including both improved mapping of older flows, and quantitative mapping of younger flows and flow features.

Mapping Flows and Flow Units

When combined with historic accounts, existing flow maps, and high-resolution aerial photos, lidar data can be used to generate very accurate maps of both flow boundaries and flow surface morphology. An illustration of a lidar-generated DEM and associated interpretive geologic map is shown in Figure 2 (Pyle and Elliott, 2006). Figure 2A illustrates the problem of cloud cover (seen as the blurry center part of the DEM, where a 15-m-resolution conventional DEM has been used to fill a cloud-generated data gap). However, Figure 2A also illustrates the exquisite topographic detail that lidar provides of complex lava flow surface morphologies. This lidar image shows a succession of historic dacitic lava flows on Nea Kameni Island in the center of the Santorini caldera, Greece. The accompanying interpretive map (Fig. 2B) shows the age of separate flows as well as the locations of vents (domes and fissures), individual flow lobes, and flow surface features (levees and surface folds). These individual features can also be mapped separately, which is particularly useful where solidified flow features can be linked directly to observed emplacement processes (e.g., Favalli et al., 2010a).

The interpretive maps of Nea Kameni and Etne rely not only on a wealth of observational data, but also on the youth (and consequent lack of vegetation) of the constituent lava flows. An important advantage of lidar technology, however, is its unique contribution to flow mapping in older vegetated terrains. In particular, raw point cloud data can be filtered to obtain bare earth DEMs of lava flow surfaces in heavily forested areas (e.g., Hofton et al., 2006). This capability of lidar alone is revolutionizing mapping of older lava flows. As an example, Figure 3 shows aerial imagery, a 10 m DEM, a bare earth lidar hillshade, and a lava flow map from the upper McKenzie River in the central Oregon Cascades. The lava flow map was created from a lidar-generated DEM by mapping flow boundaries and variations in flow surface textures. Field checking of flow boundaries, combined with selective sampling and chemical analysis, allows individual flows to be distinguished. Lidar mapping can also be combined with high-precision age dating to determine the volcanic history of a region. For example, Crow et al. (2008) refined the history of lava flow inundation of, and removal from, the Grand Canyon by mapping remnant terraces that mark canyon-filling lava flows. Correlation of individual terraces through vertical elevation and age provided a complete record of both the frequency and extent of canyon volcanism.

Flow mapping and relative age determination can be accomplished with lidar data alone by using either lidar-based intensity data or roughness analysis of lidar DEMs. An example of intensity mapping is provided in Figure 4, which shows flow-specific intensities for a region of Mount Etna. For a given target distance, the intensity of the lidar return depends on surface roughness and/or texture, in turn a function of original flow emplacement conditions (e.g., Peterson and Tilling, 1980; Rowland and Walker, 1990; Soule and Cashman, 2005) and subsequent surface weathering. At Mount Etna, the lidar signal intensity for a given flow surface type first decreases sharply because of flow cooling, and then increases gradually with increasing lava flow age as the flow surface weathers and becomes vegetated. Where revegetation is slow (e.g., arid climates), lava flow surfaces can alter by infilling with airborne dust (loess; e.g., Vaughan et al., 2011). Surface infilling will affect the intensity and the surface roughness, which can be measured using the variability in slope and aspect in local patches of a DEM, as measured by vectors constructed to each DEM cell (McKean and Roering, 2004). Local variability of vector orientation is measured using eigenvalue ratios that record the extent and nature of clustering of the vector orientations. We show an example of this type of analysis in Figure 5, which compares the surface textures of 1 km x 1 km DEM patches from a 3-k.y.-old lava flow from the upper McKenzie River (Fig. 4) with a much older pahoehoe lava flow (the 60-k.y.-old West Crater flow) from the Owyhee River, Oregon (Brossy et al., 2008). In Figure 5, red (strong clustering) corresponds to smooth topography, while blue (strong scattering) corresponds to rough topography. The contrast between these two flow surfaces reflects primarily the difference in age and consequent extensive infilling of the older West Crater flow surface with loess.

Measuring Erupted Volume

Another important parameter derived from mapping lava flows is the total flow volume (e.g., Fig. 1). Accurate measurements of lava flow volumes are surprisingly difficult to make. A combination of digital orthophotos, GIS software, and/or GPS mapping of flow boundaries generally provides excellent planimetric maps of flow surface coverage. Conversion of area to volume, however, relies on good estimates of flow thickness. In the field, flow thickness is typically estimated by averaging measurements of levee and/or flow front thickness. However, Coltelli et al. (2007) showed that the use of average flow thicknesses measured in this way may cause large errors in volume measurements (25% or greater). For example, using the average thickness of the flow margin will underestimate flow volumes where flows have filled topographic depressions (e.g., Lu et al., 2003) or undergone extensive syneruptive levee construction (e.g., Sparks et al., 1976). These errors can be reduced using accurate data on both preeruption and posteruption topography. Unfortunately, preeruption high-resolution topography is rarely available and flow volumes are estimated either by differentiating high-resolution lidar relative to existing (lower resolution) DEMs (e.g., Ventura and Vilardo, 2008; Favalli et al., 2010a) or, for prehistoric flows, using lidar data to compute thickness along the flow length (e.g., Deardorff and Cashman, 2012).

Where the preexisting topography is of sufficient resolution, comparison of preemplacement and postemplacement topography can provide important insight into flow emplacement processes. For example, Figure 6 shows thickness variations over a small segment of a lava flow erupted from Mauna Loa volcano in 1984. Here the preeruption DEM was constructed from high-resolution pre-1984 stereo-aerial photos referenced to the lidar image outside the boundaries of the 1984 flow (e.g., Corsini et al., 2009). This method provides submeter vertical accuracy, such that differentiating preeruptive and posteruptive DEMs yields a detailed imaged of flow thickness variations. In this channel segment of the 1984 lava flow, both the channel and levees are >6 m thick and much thicker than the flow margins, thus illustrating the problem
with using flow margin thicknesses to estimate flow volumes. Thickening of the flow interior relative to the flow margin is also evident in the distal segment of the 2001 Etna lava flow, where Favalli et al. (2010a) show that the early (rapidly emplaced) flow margins are less than half the thickness of the later, more slowly emplaced, channel fill.

In summary, the use of lidar-generated data to make high-resolution bare earth DEMs is rapidly enhancing our ability to map flow outlines and/or areas of old and young flows alike, to make detailed maps of flow surfaces, and to track the evolution of those surfaces through time. This ability will not only improve flow hazard maps, particularly in regions of older (and partially vegetated) lava flows, but will also allow new types of analysis, such as investigations of the ways in which lava flows are revegetated (Deligne et al., 2012). Moreover, as time goes on, coupled preeruptive and posteruptive lidar surveys will provide increasingly accurate measurements of flow thickness distributions and flow volumes produced by specific eruptions.

**MONITORING ACTIVE FLOWS**

Instantaneous lava flux is probably the most important parameter controlling both the rate of lava flow advance (e.g., Kauahikaua et al., 2003)
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Figure 3. Comparative maps of part of ~3000-yr-old mafic lava flow field in the upper McKenzie River basin, Oregon. (A) Google Earth image showing Clear Lake (dammed by the flow) and variable vegetation cover on the flows. (B) 10 m digital elevation model (DEM) hillshade. (C) Lidar (light detection and ranging)-generated DEM of the same region; note the detail provided of the flow surface. (D) Geologic map of individual flows constructed using the lidar data and associated mapping, sampling, and geochemical analysis.

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Figure 4. Lidar (light detection and ranging) intensity values superimposed on shaded relief image of a portion of Etna volcano, Italy. Data are normalized to a standard aircraft elevation of 1000 m. Image illustrates the variation in intensity data with flow age and surface texture (from Mazzarini et al., 2007, copyright 2007 American Geophysical Union. Reproduced by permission of American Geophysical Union.).
and the distance a flow will travel (e.g., Walker, 1973). Accurate time-averaged volumetric effusion rates can be determined from measured flow volumes and known eruption durations (e.g., Hofton et al., 2006; Ventura and Vilardo, 2008). Reconstructing temporal variations in flux, however, requires more than simple preeruption and posteruption topographic data. For this reason, acquisition of image time series of active flows has been a recent priority in lava flow studies. Here we briefly review approaches that do not rely on lidar before discussing the contributions of lidar to this important research topic.

Ground-based multitemporal studies of active lava flows have focused on changes at the flow front, a region that is relatively accessible and safe for study. For example, a study of a blocky lava flow from Arenal volcano, Costa Rica, using ground-based millimeter wave radar (AVTIS, All-Weather Volcano Topography Imaging Sensor; Macfarlane et al., 2006) allowed images to be obtained from a (safe) distance of 3 km. Recorded changes in both vertical elevation profiles and flux over an 8-day observation period were used to measure the rates of lava flow advance and to estimate the volumetric flux. Similar measurements have been employed in monitoring the growth of a lava dome at Soufrière Hills volcano, Montserrat (Wadge et al., 2008), where the combination of long viewing distance and insensitivity to the persistent cloud cover has proved particularly valuable. Photogrammetric time series obtained during recent eruptions of Mount Etna (Coltelli et al., 2007; James et al., 2007, 2010) have also captured detailed information on short-term variations in lava flux to the flow front. These studies have documented order-of-magnitude changes in volumetric flux at the flow front caused by pulses of lava that travel down the channel at rates of 0.16–0.33 m/s (James et al., 2007, 2010). The lidar data show simultaneous movement of these pulses through multiple parallel channels, providing evidence that the flux changes originated near the vent and not by flow variations through individual channel reaches. The multitemporal lidar images also document levee overflows when the capacity of the channel is exceeded by the temporary increase in lava volume caused by a pulse. This process of pulse-driven overflows may explain the excessive levee thicknesses at the large channel bend in the Mauna Loa lava channel (shown in Fig. 6).

Another example of the application of multitemporal lidar imaging to active lava flows is a study of flow emplacement on the steep slopes of Sciara del Fuoco, Stromboli volcano, Italy.
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Figure 6. Preeruption and post-eruption eruption differencing of high-resolution digital elevation models (DEMs) for a region of the 1984 Mauna Loa lava flow. Preeruption DEM was created from high-resolution aerial photos (e.g., Corsini et al., 2009); posteruption DEM is generated from lidar (light detection and ranging) data. Off-flow error is <1 m horizontal and ~0.15 m vertical. Channel construction is particularly evident at the location of a major breakout and/or flow bifurcation (Flow1B of Lipman and Banks, 1987).

Figure 7. Volumetric changes across the distal portion of Etna lava channel measured by digital elevation model differencing. (A) At 919 s. (B) At 2817 s. (C) At 4706 s. (D) At 6621 s. (E) At 9290 s. (F) Total change over 2.5 h. Color scale denotes relative elevation differences (flow thickness): red is a relative increase, blue is a relative decrease. Black lines locate cross sections shown in G–I as variations in time-averaged discharge rate (TADR) along the channel, plotted as a function of distance from the flow front position in F. G–I: Measurements in time steps. (G) 08:46–08:31. (H) 10:21–10:06. (I) 08:31–11:04. Red line in I gives the total volume emplaced per unit length (L) (from Favalli et al., 2010b, copyright 2010 American Geophysical Union. Reproduced by permission of American Geophysical Union.).

dV/dx—change in measured flow volume per unit length along the channel.
that is ultimately delivered to the flow front. They also demonstrate the necessity of repeat surveys, particularly near the beginning of an eruption when flux tends to be highest and conditions change rapidly.

MODELING LAVA FLOW EMLACEMENT

The studies described here demonstrate the use of lidar data to create high-resolution DEMs before, during, and after lava flow activity, thereby providing critical information on the spatial and temporal evolution of lava flow fields. Repeat surveys of an active lava flow (Favalli et al., 2010b) further show the potential of using lidar for imaging short-term variations in the movement of lava through complex channel systems. These studies raise new questions about channel development and the behavior of channelized lava flows, many of which can be addressed by detailed analysis of lidar-generated DEMs.

Measuring Lava Channels

Prior to the advent of lidar technology, there were few systematic studies of lava channel geometry. Particularly challenging are measurements of active channel depth, which is commonly estimated by measuring the maximum (assumed) neutral buoyancy height of rafted lava fragments (Lipman and Banks, 1987) or, rarely, by probing lava streams within lava tubes (Kauahikaua et al., 1998). Postemplacement channel measurements employ either simple depth:width aspect ratios (e.g., Kauahikaua et al., 2002) or time-consuming (and thus limited) differential GPS surveys of complete channel sections (e.g., Zimbelman et al., 2008). The paucity of accurate data on channel cross sections is unfortunate given the demonstrated importance of lava channels as the primary conduits of lava transport. As a result, thermomechanical models of flow emplacement typically assume either that channels maintain a constant depth (e.g., Harris and Rowland, 2001) or a constant width (e.g., Kilburn, 1996) from the vent to distal end of the flow; these assumptions are clearly oversimplifications for lava flow fields that host multiple active channels, and in which short-term changes in lava flux cause frequent channel overflows.

Extraction of channel cross sections from lidar-generated high-resolution DEMs allows analysis of along-flow channel geometry. Model predictions suggest that flow (and channel) geometry should reflect both the material properties of the lava and the volumetric flux (e.g., Hulme, 1974; Lipman and Banks, 1987; Kilburn, 1996; Harris and Rowland, 2001). A simple comparison of channel cross sections from Hawaii (1984 Mauna Loa flow; Fig. 8 [left]) and central Oregon (basaltic andesite Collier Cone lava flow; Deardorff and Cashman, 2012; Fig. 8 [right]) confirms that the (more viscous) basaltic andesite lava channel is both wider and deeper than the basaltic lava channel on Mauna Loa. The two flows have roughly similar volumes (0.12 km³ for the Collier Cone flow, as compared to 0.22 km³ for the Mauna Loa flow); in addition, the Collier Cone flow traversed steeper
average slopes (≤10° compared to ≤5°) and was probably erupted at significantly lower effusion rates (Deardorff, 2011) than the Mauna Loa flow. As both of these factors should decrease, not increase, the channel dimensions, it appears that the lava rheology was the most important factor in determining channel geometry.

If more detail is required, ALS imaging can be supplemented with TLS imaging of individual channel reaches. We find that although channel dimensions from ALS are consistent with TLS measurements, ALS-derived channel depths are ~10% less, and widths ~20% more, than TLS-derived measurements of the same location (Fig. 9). Moreover, although gridding the ALS data at 0.5 m provides a better fit than the 1 m gridding interval, the ALS model surface shows a U-shaped channel, whereas the TLS resolves a rectangular channel, which is more faithful to the actual channel shape (and has important implications for models of flow through lava channels). This difference results from both wider point spacing in ALS relative to TLS data as well as larger grid spacing in DEM surface models that average elevations over small-scale features such as steep channel walls. Accurately capturing bank geometry is also a common problem with ALS in fluvial studies (e.g., McKean et al., 2009).

TLS imaging is also able to resolve small-scale flow features that are not visible in ALS gridded data. As an example, a TLS data set from another portion of the Mauna Loa 1984 channel system imaged “armadillo” (Naranjo et al., 1992) structures lining the channel floor (Fig. 10). These structures form during the late stages of flow emplacement due to shear failure of partially solid, but ductile lava, and were observed to be the dominant mode of lava transport during the final phase of emplacement of the 1988–1990 Lonquimay basaltic-andesite flow (Naranjo et al., 1992). The Mauna Loa structures are 50–75 cm in length and 5–15 cm in height, comparable to those described for Lonquimay (~1–5 m in length, tens of centimeters in height). Their morphology is reminiscent of low-angle detachment faults with progressively rotated fault planes, steep breakaways, and fault-parallel corrugations (Cann et al., 1997; Tucholke et al., 1998). Both the scale and the geometry of these structures provide information on rheological changes in late-erupted magma that may provide clues to conditions that cause lava to stop flowing.

**Modeling Lava Channels**

To fully understand lava flow emplacement, we need to understand how and why lava flows construct channels. Channels develop in flows of non-Newtonian fluids when fluid yield strength limits lateral spreading (Johnson, 1970; Hulme, 1974). Although this simple model has been widely applied to lava flows (e.g., Fink and Zimbelman, 1986; Moore, 1987), levees in active lava flows clearly form by multiple mechanisms, many of which involve cooling and solidification of the flow surface (Sparks et al., 1976; Lipman and Banks, 1987). To explore the effect of cooling on flow channelization, Kerr et al. (2006) developed a model in which flow width (w) is determined by competition between cross-slope flow spreading (controlled by the physical properties of the magma, the volumetric flow rate, Q, and the slope) and cooling and crust formation:

\[
w = 2 \sqrt{\frac{(g\Delta \rho)^2 \sigma^4 \kappa^2 \sin^2 \theta \cos^8 \theta}{\kappa^2 \sin^4 \theta}} \]

where g is gravity, \(\Delta \rho\) is the density contrast between the magma and surrounding medium (air), \(\mu\) is magma viscosity, \(\sigma\) is crust strength, \(\kappa\) is thermal diffusivity, and \(\theta\) is slope.

A superb lidar-generated data set for a 2004 lava flow from Mount Etna (Mazzarini et al., 2005; Fig. 11A) provides along-flow data on flow width, channel width, and slope that can be used to illustrate the application of the Kerr et al. (2006) model. Figure 11B shows the excellent correspondence between calculated and observed channel width for a flux of 0.6 m³/s, the best-fit effusion rate determined by analysis of the root mean square error (RMSE, calculated using the difference between predicted and measured channel width). At this flux, the RMSE is 2.7 m, ~15% of the average channel width.

**Figure 9.** Topographic profiles across a rectangular-shaped channel in the Mauna Loa 1984 flow by TLS (terrestrial laser scanning) and ALSM (airborne laser scanner mapping) at the same location. TLS data are gridded at 0.1 m (blue), and ALSM data are gridded at 1 m (black) and 0.5 m (red). The rectangular shape is resolved by the TLS data, but 0.5 m and 1 m ALSM data suggest a U-shaped channel.

**Figure 10.** Imaging and measurements of “armadillo” (Naranjo et al., 1992) structures on the floor of a lava channel in the Mauna Loa 1984 flow. (A) Point cloud viewed from above, with colors reflecting intensity of returned laser pulses. (B) TLS (terrestrial laser scanning) point cloud gridded at 2 cm with illumination from the west. Corrugations parallel to the direction of flow (left to right) are well resolved. Red line shows profile transect illustrated in C. (C) Profile across the armadillo structures showing the curved shear surface and steep breakaway present on most individual structures.
width along the flow. A flux of 0.6 m/s is slightly less than the minimum estimated total volumetric flow rate of 1 m/s (Mazzarini et al., 2005). The difference between the channel-derived flow rate and minimum (1 m/s) and volume-averaged (2.2 m/s) flow rates demonstrates that the channel transported only a fraction of the total lava volume. Evidence of flow outside of the well-defined channels can be found in the highly irregular flow margins, channel spillovers, and flow bifurcations around topographic obstructions (Fig. 11A).

**From Lava Channels to Lava Tubes**

Most of the discussion here has focused on construction of, and flow through, well-defined lava channels. However, basaltic lava transport can also occur through lava tubes, a common feature of lava flows and a thermally efficient form of lava transport (e.g., Kesztthelyi, 1995; Helz et al., 1995, 2003). Experimental models show that channel versus tube flow behavior is a direct consequence of the physical conditions of flow emplacement (Griffiths et al., 2003). Specifically, the extent to which a stable surface crust can be maintained on an advancing flow is determined by the relative time scales of surface cooling ($t_c$) and flow advection ($t_h$, where $u$ is flow velocity and $h$ is a characteristic length scale for advection, typically the flow thickness). This balance can be combined into a single parameter $\Psi = u/t_h$ (Griffiths, 2000). Also important is the Rayleigh number of the flow: $Ra = g\beta(\Delta T)h^3/\kappa\nu$, where $\beta$, $\kappa$, and $\nu$ are the thermal expansion coefficient, thermal diffusivity, and kinematic viscosity of the fluid lava, respectively, and $\Delta T$ is the temperature difference between the flow interior and flow surface.

The applicability of this model to natural lava flows has been tested in only a few cases. Soule et al. (2005) predicted $\Psi$ using modeled flow velocities for specific channel geometries derived from high-resolution bathymetry. Locations where a mobile crust was predicted correlated with the presence of wide autobrecciated bands at the channel margins, a morphologic expression of open channel flow (Fig. 12A). Similarly, Ventura and Vílardo (2008) applied the model to the 1944 Somma-Vesuvius lava flow, where they modeled flow velocities from lidar-derived flow thickness and estimates of lava rheology. They found a similar correlation between $\Psi$ and flow morphology, with mobile crust regions characterized by autobrecciated flow surfaces and stable crust regions by smooth sheets (Fig. 12B). Additional experiments (Cashman et al., 2006) provide a framework for extending this approach to a variety of flow regimes and channel geometries that could be easily exploited with lidar-derived DEMs to predict conditions of lava tube formation.

Once formed, lava tubes can transport lava over great distances. Observations of active tubes show that (1) tube systems are often complex, with more than one branch active at a given time, (2) individual tube segments evolve in morphology with time, particularly through downcutting by a combination of mechanical and thermal erosion, and (3) older tubes can be reoccupied during renewed eruptive activity (e.g., Greeley et al., 1998; Kauahikaua et al., 1998, 2003; Williams et al., 2004). Despite their importance, however, there are only a limited number of studies of either the physical characteristics of lava tubes or their mode of formation (e.g., Peterson et al., 1994; Calvari and Pinkerton, 1998; Greeley, 1987; Dragoni et al., 1995; Kauahikaua et al., 1998; Kerr, 2001). These studies have recognized a diversity of tube morphologies that appear related to the dynamics of flow through the tubes, their mechanisms of formation, the extent of thermal erosion, and environmental conditions such as local slope. However, the almost complete absence of morphometric data on lava tube shapes along their lengths makes development and testing of hypotheses for tube formation and evolution difficult.
Figure 12. Relationship between flow Rayleigh number (Ra) and the ratio between advective and thermal time scales (Ψ) for flows with difference surface morphologies imaged using high-resolution topographic data. A) Submarine channelized lava flows erupted along the East Pacific Rise (following Soule et al., 2005). B) Channelized lava flow erupted from Vesuvius (Italy) in 1944; here intermediate (IS) and distal (DS) flow segments show contrasting (and unexpected) behavior with a transition from tube flow to open channel flow downslope because of an abrupt increase in slope (modified following Ventura and Vilardo, 2008).

The portability of TLS systems make them excellent tools for detailed surveying of the structure and morphology of lava tubes; in particular, TLS systems are able to resolve the cross-sectional tube geometries as well as internal features such as lava falls and lava highstands. In the example shown in Figure 13, a 30 m section of the Thurston lava tube on Kilauea volcano was scanned by an Optech Iliris 36D TLS. Two scans, one tilted up and the other down, were collected from the same scanner location. Tube-normal cross sections every 4 m illustrate the length scale (10–12 m) over which the tube cross section changes from a symmetric roughly circular shape to an asymmetric shape. The cyan dashed line in Figure 13 is at the same elevation in each cross section and shows that the tube volume increases through downdropping of the floor, possibly due to capture of the flow by a deeper tube or by thermal and/or mechanical erosion of the tube floor (e.g., Kauahikaua et al., 1998; Kerr, 2001). The inset at the lower left illustrates TLS imaging of lava highstands. The ability to obtain accurate 3D images of lava tubes thus presents the opportunity to study lava tube structure and origin in completely new ways, and certainly represents a unique opportunity for advancing our understanding of tube development in both terrestrial and planetary environments.

Morphological Measurements of Lava Flow Surfaces

Prominent morphological features of lava flows include not only lava channels and levees, but also surface folds, tumuli, fractures, blocks, and lobate flow fronts (see Fig. 2). These features record the dynamics of flow emplacement, resulting from competition between the fluid processes driving the flow (controlled primarily by lava flux and rheology) and the restraining presence of a growing (and brittle) crust (e.g., Griffiths, 2000; Applegarth et al., 2010). Of these surface features, prominent surface folds are common on lava flows of all compositions. They range in amplitude and wavelength from centimeters (in Hawaiian pahoehoe) to tens or hundreds of meters (in obsidian flows). In theory, the geometry of flow surface folds can be used to constrain the thickness and viscosity of the folding layer (from the fold wavelength and the compressional stress (from the fold amplitude; e.g., Fink and Fletcher, 1978; Gregg et al., 1998); for this reason, surface folds have been used to estimate lava rheology, particularly in remote (or planetary) environments (Theilig and Greeley, 1986; Gregg et al., 1998; Warner and Gregg, 2003). In practice, although fold wavelengths can be measured with reasonable accuracy from high-resolution orthophotos, measurement of fold amplitude requires high-resolution topographic data.

Surface folds can be characterized using lidar-generated DEMs, either as line transects taken perpendicular to the fold axes (e.g., Figs. 14 and 15A) or by generating spectrograms via Fourier analysis of digital surface data (Fig. 15B). The appropriate DEM accuracy depends on the scale of folding; for example, surface folds preserved within the Mauna Loa 1984 flow channel, with fold wavelengths of 5–10 m and amplitudes of ~0.5 m, can be resolved by ALS data (Fig. 14A). In contrast,ropy folds generated on pahoehoe flows from the coastal plain of Kilauea volcano are not visible with ALS, but can be resolved with TLS (Fig. 14B). Here fold wavelengths of 2–3 cm and 5–10 cm are similar to the first- and second-generation folds observed by Fink and Fletcher (1978). A third generation of folds that give the pahoehoe ropes a braided appearance (Fig. 14C) cannot be resolved even by the TLS data.

Lidar-derived transects can be analyzed by Fourier analysis to track changes in wavelength along the flow. The data must first be interpolated to unit spacing and detrended to remove the influence of slope. Pyle and Elliott (2006) also suggested using a cosine taper to improve data quality; their analysis of surface folding along one of the Nea Kameni lava flows (shown in their Fig. 3; see Fig. 15B herein) shows that the dominant wavelength of the surface folds increases from ~20 m near the vent to ~40 m at the flow front (a distance of 2 km), suggesting gradual thickening of the folded viscous layer. Qualitatively, the decrease in wavelength and amplitude from the more dacitic Kameni lava flows to the basaltic but viscous Mauna Loa 1984 lava flow, to the low-viscosity Kilauea coastal plain lava flow is consistent with viscosity-based models of fold generation; however, a more comprehensive survey of these features, now easily afforded by ALS and TLS data, is necessary to develop quantitative models for fold formation.

LIDAR APPLICATIONS TO HAZARD AND RISK ASSESSMENT

The hazard posed by lava flows to exposed communities is often assessed by modeling the probability of lava flow invasion into affected...
areas. The traditional approach to hazard modeling has been to map hazard zones on the basis of past activity (e.g., Wright et al., 1992), an approach that assumes that future activity will follow, statistically, the same patterns. An alternative approach is to run simulations of future activity and use the results of those simulations to map the susceptibility of different areas to lava flow invasion (e.g., Felpeto et al., 2001; Rowland et al., 2005; Favalli et al., 2009b, 2009c; Crisci et al., 2010). Critical for these models are DEMs that are (1) updated sufficiently frequently to account for topographic changes caused by ongoing eruptive activity (both lava flows and scoria cones, e.g., Neri et al., 2008; Tarquini and Favalli, 2010), and (2) of sufficiently high resolution to characterize expected activity. Most simulations are run on 10 m DEMs, even if the digital topographic data were originally collected at a higher resolution (e.g., Tarquini and Favalli, 2010); this gridding is used for numerical convenience and, until recently, because of limited high-resolution topographic data. The growing availability of very high resolution lidar-generated topography,
How lava flows

Figure 15. Surface folding analysis of a dacitic lava flow, Nea Kameni, Greece. (A) Transect perpendicular to surface folds showing multiple characteristic fold wavelengths. (B) Spectral analysis showing changes in dominant flow wavelength (indicated by spectral power, with red being high) from proximal (window 11) to distal (window 1) parts of the flow (from Pyle and Elliott, 2006).

We have reviewed the application of ALS and TLS data to mapping and morphologic interpretations of lava flows. It is important that collection of digital x-y-z point cloud data allows not only (1) extraction of bare earth topography in vegetated terrains, but also (2) detailed analysis of surface features using spectrogram, Fourier transform, and vector analysis techniques, (3) extraction and measurement of morphologic features such as lava channels and bounding levees, and (4) quantitative data required for risk assessment. The extraordinarily high resolution of TLS data makes TLS particularly useful for detailed geomorphic studies. As the intensity of the lidar return from the imaged surface is sensitive to surface texture (as well as imaging distance), flows of different ages and emplacement style can be easily distinguished using normalized intensity data. Multitemporal topographic data for a single lava flow provide detailed records of flow thickness distributions and interactions with topography (if collected before and after flow emplacement) as well as important information on conditions of flow advance (if collected during emplacement). Together these new capabilities allow testing of existing models of channel formation and flow advance and improved capabilities to assess both the hazard and risk of lava flow inundation; we expect that continued acquisition of lidar data will also promote new ways of thinking about lava flows and flow fields. Additionally, it seems clear that new analysis techniques and interpretive frameworks developed for ALS and TLS data on terrestrial lava flows will be rapidly transferred to interpretation of lava flows in submarine (e.g., Soule et al., 2005) and planetary (e.g., Glaze et al., 2003, 2009; Hiesinger et al., 2007) environments.

CONCLUSIONS

Figure 16. Analysis of data shown visually in Figure 16. (A) Percentage of each hazard area (H1–H4) occupied by the urban environment, forest, or bare earth. (B) Percentage of each land type (urban, forest) in each hazard area (following Bisson et al., 2009).

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However, raises the question of what scale of topography is optimal for modeling flows of different rheologies.

Modeling the risk of lava flow invasion requires combining inundation hazard probabilities with maps of land use and infrastructure. Land use classification is often undertaken using a combination of remote sensing (usually satellite imaging) and field checking (e.g., Harris et al., 2011); classification is typically presented in the form of the percentage of land area covered by a certain crop or population density. However, a recent study of the risk of lava flow invasion into the town of Zafferana Etnea (on the slopes of Mount Etna) illustrates ways in which lidar-generated bare earth DEMs can be combined with first return (canopy) DEMs to measure the volumes (rather than simply area) of urban and forested areas, thereby helping to quantify the risks associated with lava flow hazards (Bisson et al., 2009). These data can be analyzed by the percentage of land use within each hazard zone (Fig. 16A) or by percentage of each hazard zone in urban versus forest environments (Fig. 16B).


