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Seismic and Acoustic Signatures of Surficial Mass Movements at Volcanoes

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Abstract

Surficial mass movements, such as debris avalanches, rock falls, lahars, pyroclastic flows, and outburst floods, are a dominant hazard at many volcanoes worldwide. Understanding these processes, cataloging their spatio-temporal occurrence, and detecting, tracking, and characterizing these events would advance the science of volcano monitoring and help mitigate hazards. Seismic and acoustic methods show promise for achieving these objectives: many surficial mass movements generate observable seismic and acoustic signals, and many volcanoes are already monitored. Significant progress has been made toward understanding, modeling, and extracting quantitative information from seismic and infrasonic signals generated by surficial mass movements. However, much work remains. In this paper, we review the state of the art of the topic, covering a range of scales and event types from individual rock falls to sector collapses. We consider a full variety of volcanic settings, from submarine to subaerial, shield volcano to stratovolcano. Finally, we discuss future directions toward operational seismo-acoustic monitoring of surficial mass movements at volcanoes.

Keywords: volcano seismology, infrasound, landslides, mass movements, avalanches, lahars

Highlights

- Surficial mass movements are common in volcanic areas and generate signals that are picked up by seismic and acoustic monitoring arrays
- Our understanding of the relation of these signals to characteristics of the mass movement is limited but improving
- We review the literature on the study of mass movements at volcanoes using seismic and acoustic monitoring
- We discuss future research directions and steps toward operational monitoring

1. Introduction

Volcanoes are fundamentally unstable features. Magmatic processes deform the ground, extrude new domes, emplace dikes, deposit thick layers of unconsolidated tephra, and leave behind steep-walled craters. Such actions can affect slope stability even after a volcano is no longer active. Rivers and glaciers incise deeply into already weak layers and can undercut slopes. Water seeping into the subsurface weathers and weakens the rock layers. Hot fluids and gases transform strong igneous rocks to weak clay (López and Williams, 1993; McGuire, 1996; Vallance and Scott, 1997; Reid et al., 2001). As a consequence, surface mass movements are a common occurrence in volcanic settings that can occur at any time, not just during eruptions. Though there are other types of mass movement at volcanoes (e.g., ejection of mass from volcanic vents, explosions, movements of magma and fluids in the subsurface), here we use the term to refer exclusively to gravitationally driven movements of surface material.

The hazards that mass movements pose to human populations are significant, in large part because they can be large enough and travel fast enough (tens to hundreds of kilometers per hour) to reach distant populated areas downstream quickly (seconds to hours). Large volcanic debris flows (lahars) have travelled >100 km (e.g., Vallance and Scott, 1997; Worni et al., 2012). Tsunamis generated by events at coastal and submarine volcanoes can inundate distant coastal areas (e.g., Moore and Moore, 1988; Tinti et al., 2006), and landslide dams can block rivers and may then fail catastrophically, causing rapid flooding downstream (e.g., Costa and Schuster, 1988; Bovis and Jakob, 2000). Tanguy et al. (1998) estimated that 48% of the documented volcano-related fatalities from 1783 to 1998 (>100,000 total) were directly caused by pyroclastic density currents, debris avalanches, and mudflows, and another 17% by volcanogenic tsunamis, which are often related to gravitational movements. A tragic recent case occurred in 1985 when more than 23,000 people were killed by lahars during the eruption of Nevado del Ruiz, Colombia (Pierson et al., 1990), but examples stretch back in human history. Victims of pyroclastic density currents were found at the archaeological site of Pompeii (Luongo et al., 2003), and oral traditions of indigenous people living around the volcanoes of the Pacific Northwest describe “rivers of liquid rock... killing all growing things and forcing the Indians to move far away” (Cashman and Cronin, 2008, pg. 414, originally from Clark, 1953).

Modern volcano monitoring that incorporates lessons from the past can lead to successful evacuations that save human lives, such as during the 2007–2008 eruptions of Nevado del Huila in Colombia where authorities evacuated cities that were later overrun by lahars (Pulgarin et al., 2011; USGS Volcano Disaster Assistance Program, 2011). However, economic losses can still be substantial. For example, pyroclastic density currents and lahars buried Plymouth, the capital of the island of Montserrat, during the Soufrière Hills eruption of 1996–1997 (Cole et al., 1998). While most residents had been evacuated, this led to the complete abandonment of the city and lasting economic problems for the island. Mass movements frequently occur during volcanic unrest. Since eruptions are commonly accompanied by precursory activity, preemptive evacuations can be made if the expected hazard is well known. However, some event types, such as large landslides, lahars, and outburst floods, can also happen independently of volcanic unrest, potentially without warning, making them particularly hazardous to nearby communities.

While large catastrophic events are most notable, it is important to also monitor and study smaller events for two reasons: (1) smaller events can still be locally hazardous for people working or recreating around the volcano; and (2) smaller events are more frequent, and therefore often provide the only data available for studying mass-movement processes and for understanding the types of signals to expect from larger events. Understanding and monitoring spatiotemporal patterns of activity can inform current and future hazard assessments and can be used to understand the larger and rarer events.

Many types of mass movements generate characteristic signals that can be detected by seismic, acoustic, and submarine instrumentation (e.g., hydrophones). These signals contain information about an event's size, location, and propagation dynamics. However, detecting, classifying, and extracting quantitative information from the seismic and acoustic (or seismo-acoustic) signals these events generate remains an outstanding challenge in this relatively young research field. This is primarily because monitoring networks are not typically designed to record mass movements; many, if not most, events reported in the literature were recorded serendipitously on networks designed for other purposes. Nevertheless, significant progress toward understanding and modeling seismo-acoustic signals from such events has been made in recent years. The purpose of this review is to outline the state of the art on the topic in the context of volcanic environments.

In our review, we consider seismo-acoustic studies spanning several decades. Although our intent is to focus on physical processes and their associated signal features, it is inescapable that the available observations for a given time period are influenced by the available technology. Seismic and infrasonic instrumentation, hardware, and acquisition methodologies have evolved and improved dramatically over the past several decades (e.g., Agnew and Lee, 2002; Bormann, 2012). Notable developments include the increasing availability of broadband seismic instrumentation in the 1990s, the transition from analog to digital over the past decade or so, and the increasing availability of disk storage which enabled, for example, the transition from recording primarily event-triggered waveform data focused on earthquake signals to recording continuous digital waveform data in the 1990s. Present trends include continual improvements in equipment portability, speed of deployment, field ruggedness, power consumption, and cost, which in turn are allowing for deployments of increasing numbers of high-quality sensors. Similarly, major improvements in infrasound technology have occurred since the opening for signature of the Comprehensive nuclear Test-Ban Treaty in 1996. While it is not practical in this review paper to state the instrumentation used in each particular study, the reader should keep in mind that technological improvement has influenced the type, quality, and quantity of available data over time. We highlight the instrumentation used by a particular study whenever necessary to provide context for observations and the utilized data analysis methods.

This review is broken into five sections. In Section 2, we outline the different types of mass movements that generate seismo-acoustic waves at volcanoes and describe their typical signal characteristics. In Sections 3 and 4, we provide the theoretical background for seismo-acoustic wave generation and propagation as related to surface mass movements. In Section 5, we summarize detection, location, and characterization techniques. Finally, in Section 6, we discuss what is needed to move this area of active research toward more operational applications.

2. Event types and signal characteristics

Mass movements occur in all types of volcanic settings, including shield volcanoes, cinder cones, and submarine volcanoes, but the archetypal producers are the steep, layered cones of stratovolcanoes. Traditional classification schemes typically classify events by style of motion (e.g., fall, slide, topple, flow), material type (e.g., rock, debris, earth, ice), velocity, and moisture

content (Varnes, 1978; Cruden and Varnes, 1996; Hung et al., 2014). However, only a subset of events described by these taxonomies are both common in volcanic environments and occur with sufficient volume and speed to generate observable seismic or acoustic waves. Therefore, we focus on summarizing that subset here in terms of their seismic and acoustic manifestations. Examples of these event types are shown in Figure 1 and summarized in Table 1.

2.1 General signal characteristics

The range of seismic and acoustic signal characteristics from different event types is represented in Figure 2 and Figure 3, respectively. Most seismic signals at intermediate and high frequencies ($> \sim 0.1$ Hz) have emergent waveforms (sometimes described as “cigar” or “spindle” –shaped) that lack clear phase arrivals (P and S waves) and have peak amplitudes typically towards the midpoint of the signal (e.g., Figure 2A) (Tilling et al., 1975; Norris, 1994; Levy et al., 2015). Such signals imply a source that builds up relatively gradually and extends over time, as opposed to the sudden brittle slip of faults in the case of earthquakes. One proposed explanation for the emergent character is that surficial mass movements frequently grow in momentum and size, so it can take time before sufficient material is moving fast enough for the signal to exceed the noise level at a given recording station (Surinach et al., 2000; Havens et al., 2014). Another contributing factor is that seismic waveforms are scattered in the process of travelling through the shallow subsurface, which is exceptionally heterogeneous at many volcanoes (Weaver and Malone, 1976; Weaver and Malone, 1978).

Longer period seismic signals (less than ~ 0.1 Hz) are typically observed only for exceptionally large ($> 10^7$ m³) events. When they are observed, long-period signals commonly arrive before the high-frequency signals have built up in amplitude (e.g., Allstadt, 2013a; Hibert et al., 2014a). Long-period signals can be observed thousands of kilometers away from the source, and waveforms are simpler and coherent over large distances.

Atmospheric acoustic signals are also widely observed in relation to surficial mass movements. Acoustic signals are as varied as the mass movement processes themselves, with a complex variety of waveform and frequency characteristics reported (Figure 3). Analysis is commonly restricted to frequencies lower than the threshold for human hearing (~ 20 Hz), a range that is referred to as “infrasound.” Observed infrasound signals range from transient signals associated

with “airblast” (sudden air displacement, e.g., by rock fall; Zimmer et al., 2012), gas blowout, and impacts of objects with the ground (e.g., Green and Neuberg, 2005; Oshima and Maekawa, 2001), to more sustained tremor-like signals associated with avalanches, debris flows, and pyroclastic flows (Yamasato, 1997; Johnson and Palma, 2015; Marchetti et al., 2015; Schimmel and Hübl, 2015). A challenge is that the emergent and sustained infrasound waveforms characteristic of surficial mass movement events may superficially resemble wind noise (Walker and Hedlin, 2010). Thus, it is advantageous to collect infrasound data using an array, which can isolate coherent infrasound signals from incoherent wind noise (Matoza et al., 2007; see also Section 5.2.1).

Durations of mass movements and, therefore, the duration of their corresponding seismo-acoustic signals, can vary drastically from seconds to hours depending on the style and details of the event sequence. Scattering, travel-time differences between different wave types, and surface wave dispersion can elongate the signal beyond the actual duration of the event. Commonly, this elongation is a fraction of the total duration of the source process, especially for events with long source durations. Though the duration of the source process is sometimes assumed to be approximately equal to the duration of the signal (e.g., Hibert et al., 2011), due to the emergent character of the waveforms, estimating the start and end times of an event precisely using a seismic or acoustic waveform can be problematic. For example, if the noise level is high on a given station, the signal may not emerge from this noise until several seconds after it would have if the station were in a quieter setting. Furthermore, if two or more components of a progressive failure event occur close enough in time, they may appear as one event on seismic and acoustic records. This lengthens the perceived duration of the event and complicates the interpretation (Norris, 1994).

2.2 Event types

While many mass movements share the general characteristics described above, there are also differences that reflect different styles of movement. In this section, we summarize the different types of volcanic mass movements that generate seismic and acoustic signals along with noteworthy distinguishing signal characteristics. Basic information such as typical velocities and common triggers are outlined in Table 1, while relevant details about event dynamics and the seismo-acoustic signal characteristics are discussed below. Note that precise quantitative and

mutually exclusive descriptions of the expected signal from different event types do not necessarily exist; we therefore limit our review to cases presented in the literature. Additionally, while many studies of specific events report frequency ranges observed, the frequency content observed depends strongly on where the seismic station is relative to the event and the characteristics of the subsurface that are specific to the study location. Precisely defined “typical” frequency ranges are not necessarily useful when applied out of context from where they were determined. Therefore, we focus on relative differences in frequency content and general trends rather than exact numbers.

2.2.1 Debris, rock, and ice avalanches

Debris, rock, and ice avalanches (Figure 1A) are highly mobile and extremely rapid flows of fragmented rock, ice, and/or debris (Hungr et al., 2014). The first part of the name describes the material composing the original source mass (Cruden and Varnes, 1996; Hungr et al., 2014). The term “debris avalanche” is sometimes used (including in this publication) as a catchall term at volcanoes, where “debris” is defined as a mix of dominantly coarse material. Debris avalanches frequently shift between several types of movement (e.g., fall to slide to flow) over the course of an event, so capturing this variety in a simple classification can prove futile. While the bulk of the avalanche material is rarely saturated, high velocities and long runout distances may be promoted by the entrainment and liquefaction of saturated material at the base of the flow (Hutchinson and Bhandari, 1971).

Due to high speeds and potentially large volumes, debris avalanches are among the most seismogenic types of mass movements. Close-in recordings (Figure 2A-3) contain the most complete information if recorded on instruments with enough fidelity to not clip at high amplitudes. The high-frequency (>1 Hz) energy generated by such events is commonly high enough in amplitude that, despite rapid attenuation at these frequencies, the resultant seismic signal is often observable at distances of several hundred kilometers. At long periods (tens to hundreds of seconds) the signal can sometimes be observed for thousands of kilometers (Allstadt et al., 2017). Though reports in the scientific literature are few, debris avalanches can also generate strong infrasound signals visible for hundreds of kilometers (e.g., Pankow et al., 2014) (Figure 3A-3).

2.2.2 Edifice collapses

The largest mass movements that occur at volcanoes, commonly referred to as “edifice collapses,” are not necessarily distinct from debris avalanches, except for their exceptional size. Edifice collapses can be subdivided into sector collapses and flank collapses. Sector collapses involve a large portion of the volcanic edifice including the summit, and they usually occur during volcanic unrest (Scott et al., 2001). Some examples include the 2.3 km³ sector collapse that removed the summit of Mount St. Helens in 1980 (Figure 1B) (Voight et al., 1981) and the 3.8 km³ sector collapse that removed the summit of Mount Rainier, Washington, ~5600 years ago (Vallance and Scott, 1997). Flank collapses also involve a significant portion of the edifice, but exclude the summit (Scott et al., 2001). Flank collapses from basaltic shield volcanoes may be orders of magnitude larger than sector collapses on stratovolcanoes. Many past offshore flank collapses from the Hawaiian volcanoes involved hundreds to thousands of cubic kilometers of material (Normark et al., 1993).

Edifice collapses are rare, so few events have been seismically recorded. When they are recorded, they appear on seismic stations hundreds to thousands of kilometers away. Two of the earliest papers on landslide seismic signals were on edifice collapses at Mount St. Helens and Kilauea, Hawaii, respectively (Kanamori and Given, 1982; Eissler and Kanamori, 1987). Both were studied from teleseismic distances and generated strong long-period surface waves such as we see for many large debris avalanches.

2.2.3 Rock falls

Rock falls (Figure 1D) are masses of rock of any size that move mainly by free-fall, bouncing, and rolling rather than by shear displacement (Varnes, 1978). If sourced in unconsolidated deposits, they are called “debris falls.” A rock fall signal, even if resulting from the collapse of only one block, can be complex, potentially including signals from the initial failure, breakup and falling of rocks, and multiple impacts, sliding, and airblasts along the rock fall path and when it reaches the bottom of the slope (Zimmer and Sitar, 2015). If the source mass starts as a coherent block and remains relatively intact as it falls, the signals can have onsets nearly as impulsive as earthquakes, in some cases resulting in observable *P*- and *S*-wave arrivals on seismic records (Dammeyer et al., 2011). Rock falls are common at volcanoes because volcanoes

are commonly composed of inhomogeneous mixtures of relatively strong materials such as lava flows and weaker materials such as ash layers. Stronger layers can protect weaker layers, maintaining the steep, unstable slopes and cliffs where rock falls are typically concentrated. At volcanoes where freezing temperatures and ice are common, rock falls are more likely when temperatures are above freezing. At Mount St. Helens, Washington, for example, rock fall from the crater walls is much more frequent during the day, when they are triggered by melting of ice in the wall of the crater. Winter conditions stabilize the walls, largely suppressing the occurrence of rock falls (Mills, 1991). Example seismic characteristics from a larger-than-normal rock fall off the dome of Mount St. Helens in 2006 are shown in Figures 2A-4 and 2C (Moran et al., 2008).

Infrasound signals of rock falls (Figure 3A-2, C) vary in complexity and may include separable signals from different parts of the failure. The signal from displaced air is generally lower in frequency (<1 Hz to ~ 0.1 Hz), whereas signals from impacts tend to be higher frequency (>1 Hz to tens of Hertz) (Moran et al., 2008; Johnson and Ronan, 2015; Zimmer et al., 2012). Airblast and individual impact signals result in relatively simple pulses, while more complex signals may result from breakup of the rock mass and transition to sustained flow, multiple impacts, or compressed air pockets as the rock fall progresses (Moran et al., 2008). It is noteworthy that the infrasound waveform and spectral signatures of rockfalls observed to date (e.g., Figure 3) are distinct from those of volcanic degassing and explosion signals (e.g., Fee and Matoza, 2013; Matoza et al., in press), indicating that infrasound can help to discriminate between these processes.

2.2.4 Pyroclastic density currents

Pyroclastic density currents (PDC), also called pyroclastic flows (Figure 1E), are rapid flows of hot material fluidized by escaping and entrained gasses. Since PDCs occur mainly during eruptions, isolating the seismic and acoustic signals related to a PDC from other signals (e.g., volcano-tectonic or low-frequency earthquake swarms, tremor, rock falls) can be a challenge without additional information about the sequence of events (e.g., Yamasato, 1997, who used video, or Breard et al., 2016, using analog experiments), though this is also true of other event types when they occur during eruptions.

The seismology of PDCs has been extensively documented at Soufrière Hills volcano, Montserrat (Calder et al., 2002; Jolly et al., 2002; Luckett et al., 2002; De Angelis et al., 2007) and Unzen (Uhira et al., 1994). Similar to many other types of mass movements, PDCs generate signals with emergent arrivals and typical durations of just a few minutes (Zobin et al., 2009). Frequency spectra are broad, with energy generally concentrated within the 3-20 Hz band (Jolly et al., 2002) and if observed from similar distances, typically peaking at lower frequencies than lahar seismic signals (Zobin et al., 2009, 3-4 Hz vs. 6-8 Hz). Many PDC signals have an initial low-frequency ($< \sim 2$ Hz) phase that has been interpreted differently at different volcanoes. For example, at Soufrière Hills, the low-frequency phase is attributed to gas escape and resonance of the conduit that forces instability of the dome (Luckett et al., 2002). On the other hand, at Unzen, the low-frequency signal is explained by the detachment and impact of part of the dome on the ground below (Uhira et al., 1994). Zobin et al. (2009) compared PDC's generated by lava dome collapses with those generated by the collapse of the eruptive column at Volcán de Colima, Mexico, and found that there was no difference in the spectral content of the two, and the only difference being the initiating part of the signal: the lava dome collapse PDC started with an emergent signal related to the actual collapse, the eruption column collapse PDC started with a large-amplitude explosion. The explosion signal starts with longer-period pulses of energy that the authors attribute to magma movement followed by an explosion in the conduit (Figure 4).

PDCs are also well-documented sources of infrasound (e.g., Oshima and Maekawa, 2001; Delle Donne et al., 2014) that consist of long-duration, broadband signals. In a pioneering study using analog infrasound data, Yamasato (1997) observed variations in the period of observed waveforms from pyroclastic flows at Mount Unzen, Japan. These observed period changes were interpreted as a Doppler shift associated with motion of the PDC, which could consequently be used to estimate the flow velocity. Based on preliminary field observations on the intensity and size of PDCs at Soufrière Hills volcano, Ripepe et al. (2010) speculated that the frequency of the dominant spectral peak may scale inversely to the size of the flow and reflect propagation dynamics and grain composition of the pyroclastic flow.

2.2.5 Lahars and outburst floods

Lahar is an Indonesian term for flows of mud, debris and water coming from volcanoes (Figure 1F). Lahars are a primary volcano hazard because their high mobility (Iverson, 1997) allows

them to reach populations distant from volcanoes with little or no warning. While many triggers are related to eruptive activity, lahars with a long reach may also be triggered in the absence of volcanic activity by landslides or heavy rain events (Table 1). The recognition of this hazard has driven the development of lahar warning systems at several volcanoes that use seismic and acoustic signals to warn downstream populations in time for evacuation (Section 5.1.2.2).

The term lahar is used only in volcanic settings and has been adopted worldwide by the volcanology community as a general term that incorporates debris flows and hyperconcentrated flows (Smith and Fritz, 1989). Though debris flows and hyperconcentrated flows also occur in non-volcanic settings, lahars can be much larger, faster and more extensive than their non-volcanic counterparts and can be triggered by events and conditions unique to volcanoes. Debris flows in any setting often start as saturated landslides or watery streamflows that gain momentum and incorporate solids as they flow downhill (Iverson et al. 1997). However, since volcanoes are rich sources of easily-scavenged debris, ponded and interstitial fluids, and clays from hydrothermally-altered volcanic materials (Vallance, 2005), the initiating landslides can be larger and watery flows can bulk up by a greater factor than in non-volcanic settings (e.g., a factor of ~ 2.5 at Ruapehu, New Zealand, in 2007 (Procter et al., 2010), and ~ 4.5 at Nevado del Ruiz, Colombia, in 1985 (Pierson et al., 1990)).

When describing the seismic and acoustic wavefields generated by lahars, it is useful to specify whether recordings are “near-field”, where the instrumentation is located tens of meters from the flow channel and the signal is dominated by energy coming from the part of the flow passing right by the station, or “far-field”, where stations are located kilometers from the flow and the signal may be dominated by energy coming from many parts of the flow simultaneously. The observing distance is important for any mass movement event, but near-field recordings of other event types are uncommon either because they are too destructive or because their path is not easily anticipated. Lahars occur in channels and run out for long distances so instrumentation may be placed right next to the channel where flows are expected.

Since lahars encompass a range of flow types (debris flows, hyperconcentrated flows) and many lahar triggers are also seismogenic, the seismic and acoustic signal characteristics can vary widely from event to event in both the near-field and the far-field. Far-field signal characteristics may vary at different locations for a single event because the behavior of a single flow can vary

substantially along its path, and due to their long runout distances, source to station distances can change significantly. In the absence of an impulsive trigger (e.g., landslide or PDC), far-field seismic signals generated by flows in all settings typically have a very emergent onset, with the signal growing in amplitude sometimes over the course of tens of minutes as the flow bulks and gains momentum. Signals can last for up to several hours before fading (Figure 2A-1, 2A-2, 3A-1).

Small flows ($< \sim 10^5 \text{ m}^3$) may only be observable at high frequencies ($> \sim 1 \text{ Hz}$) at seismic stations located within a few kilometers of the event, while large flows ($> 10^6 \text{ m}^3$) generate observable seismic waves over a wider frequency band at greater distances (Allstadt et al., 2017) (Figure 2B). In the near-field, flows with a higher concentration of solids have generally been observed to generate more lower frequency ($< 50 \text{ Hz}$) seismic energy than wetter hyperconcentrated flows ($> 50 \text{ Hz}$, Doyle et al., 2010; Cole et al., 2009). Though this is also true when observed from the far-field, the exact frequency ranges observed depend strongly on distance from the source and vary from study to study. Infrasound signals of lahars and debris flows (Figure 3A-1) are similar to seismic signals in that they have long durations and can have broadband frequency content. Chou et al. (2007) observed that flows containing larger clasts generated higher frequency infrasonic energy than muddier flows. As flows progress down channels, they commonly develop surges where localized higher velocities and coarser grained material are concentrated, forming a dominant boulder-rich surge at the front (Zanuttigh and Lamberti, 2007). These higher-energy surges produce higher amplitude signals than the intersurge flow (e.g., Kean et al., 2015; Doyle et al., 2010).

Rain on fresh ash often mobilizes into small or intermediate-sized lahars (Jones et al., 2015; Kumagai et al., 2009; Lavigne et al., 2000; Doyle et al., 2010). Signals of lahars of this type would typically have a very emergent onset. Rainfall-induced lahars can persist for years after the ash has been deposited (e.g., Marciál et al., 1996; Tuñgol 2002), and therefore can appear unrelated to volcanic activity. The high repeatability of this type of lahar can provide datasets needed to allow for fine tuning and calibrating of near-field lahar monitoring systems and the thresholds used for alarms (e.g., Lavigne et al., 2000; Tuñgol 2002, Jones et al., 2015). However, such calibrations, which are often in the form of relations between discharge and seismic amplitude, are usually empirical and valid only at the site where they were derived. Locations

with high event repeatability can also provide a good natural laboratory for studying lahars and lahar seismicity in general because dozens of lahars may occur in the same drainage and complementary instrumentation (rain gauges, flow height sensors, cameras) can be collocated with near-field seismic instrumentation in anticipation of future flows. For example, Doyle et al. (2010) found a power-law relation between near-field peak seismic amplitude and the cross-sectional area of the flow (wetted area) and was able to make inferences about particle concentration based on relative frequency content for a lahar at Semeru volcano, East Java based on independent bucket-sampler sediment measurements. Doyle et al. (2010) found that there was less high frequency energy and lower amplitudes overall for dilute flows than concentrated flows, but echoing the findings of Cole et al. (2009), they noted that this relation was not direct because dense, laminar style flows were quieter and less rich in high frequencies as well. They attribute this to differences between lower-frequency frictional and higher-frequency collisional flow styles.

Far-field recordings are more difficult to relate to other quantitative data even if observations of many lahars are available, but general trends can still be identified. For example, Zobin et al. (2009) compared numerous rain-fall induced lahar signals to those of PDC's traveling down the same ravines and recorded at the same station, located ~1 km from the closest part of the flows at Volcán de Colima, Mexico, and found that the peak frequencies of the lahars (6-8 Hz) were consistently higher than those of PDC's (3-4 Hz).

Lahars can also be triggered by mass movements discussed elsewhere in this paper (PDC's, individual or coalescing landslides) that have their own distinct seismic and acoustic signals. In the case that the triggering event grades directly into the lahar, the lahar signal may emerge and build up to a peak while the signal of the trigger fades (e.g. Figure 2A-1, eruption and subsequent lahar at Mount Redoubt), and the separation of the two sources in time may not always be easily discernable. Spectrograms, which show the evolution of frequency content in time like that shown in Figure 2A-1, can help detect shifts in frequency content related to changes in dominant flow type that may not be apparent in the seismogram alone. In other cases, lahar mobilization can be delayed, allowing for easier signal discrimination. For example, dewatering and liquefaction of the 1980 Mt St Helens debris avalanche deposits generated large lahars that began hours after the eruption (Janda 1981, Fairchild 1987).

Sudden releases of large amounts of water (outburst floods, Figure 1C, Table 1), also generate seismo-acoustic signals and can evolve into lahars (e.g., Walder and Driedger, 1994; de la Fuente et al., 2016). Glacial outburst floods (jökulhlaups) from subglacial or periglacial reservoirs can be linked to meteorological events (heavy rains, summer melting) that add water to the glacial bed at non-erupting glacier-clad volcanoes, and to increased heat production at erupting volcanoes (Walder et al., 1995; Vogfjörð et al., 2005). Lahar-initiating floods can also come from breaches of subaerial lakes or water-filled craters (e.g., Procter et al., 2010; Cole et al., 2009; Waythomas et al., 2013; Buurman et al., 2013; Badrudin, 1994), and less commonly, the expulsion of groundwater and/or hydrothermal water (Newhall et al., 2001; Pulgarin et al., 2015). Outburst floods are on a continuum with debris flows, so they share many of the same general seismic and acoustic characteristics (e.g., Schimmel and Hübl, 2016). Schimmel and Hübl (2016) suggest that, in general, infrasonic signals from outbursts may be higher in frequency than those from debris flows. Outburst floods tend to be more seismogenic than precipitation-induced floods because they start suddenly, are more concentrated in time, and can have much larger discharge rates. An example signal from an outburst flood and subsequent lahar (unrelated to eruptive activity) at Mount Shasta, California, is shown in Figure 2A-2 (de la Fuente et al., 2016).

2.2.6 Mass movements at submarine volcanoes

Many of the previously described event types can also occur beneath the ocean surface (Figure 1H). Mass movements are a major contributor to the morphologic evolution of submarine volcanoes. Bathymetric data reveal features associated with slides and slumps, including landslide headwalls and chutes, as well as debris on the surrounding seafloor (Moore et al., 1989; Mitchell et al., 2002; Wright et al., 2008). In the case of several active submarine volcanoes where repeat surveys were conducted, slides have been identified in bathymetric difference mapping (Wright et al., 2008; Chadwick et al., 2008, 2012; Embley et al., 2014; Haney et al., 2014). Active landsliding in the submarine environment generates broadband acoustic signals that may be identified on hydrophones or seismometers, even at great distances from the source (Caplan-Auerbach et al., 2001; Okal, 2003; Wright et al., 2008; Chadwick et al., 2012). Sliding generates a characteristic broadband (0-200 Hz) signal that may last for tens of seconds to tens of minutes (Caplan-Auerbach et al., 2001; Caplan-Auerbach et al., 2014). Hydroacoustic signals may also couple into the ground and be recorded by land-based seismometers, as occurred for the

sector collapse at Monowai volcano, New Zealand (Wright et al., 2008). Acoustic signals generated by submarine landslides on Kilauea typically begin with a 1-20 Hz “rumble” that Caplan-Auerbach et al. (2001) interpreted as tumbling large blocks. The “rumble” signals were subsequently replaced by a broadband “hiss” that may reflect the sliding of fragmental material down the volcano’s flank.

Because they are remote and underwater, mass movements at submarine volcanoes can be difficult to identify or study. However, such events can sometimes be identified in the hydroacoustic record because they generate strong interference patterns caused by wavefront reflections in the water column. The simultaneous arrival of direct waves with waves reflected off the ocean surface results in some wavelengths interfering constructively, while others interfere in a destructive manner. Integer multiples of these wavelengths behave similarly, resulting in a characteristic interference pattern called a Lloyd’s Mirror (Carey 2009) that is dependent on source-receiver geometry. Changes in the interference pattern therefore represent a moving source, most likely indicative of mass wasting (Caplan-Auerbach et al., 2014).

2.2.7 Ice and Snow

Many volcanoes are glaciated and receive large amounts of snow each winter. Glacier motion and snow avalanches also generate seismic and acoustic waves that are recorded by volcano monitoring networks. While not strictly of volcanic origin, we briefly review seismic characteristics of these mass movements because their signals can be difficult to distinguish from other mass movements more directly related to volcanic hazard. They also provide insights into other types of events. Glacier seismicity is an extensive subdiscipline that is reviewed in detail by Aster and Winberry (2017), so we restrict our discussion to aspects relevant to volcanoes and mass movements.

Collapses from walls of ice (seracs) and ice avalanches share seismic and acoustic characteristics with rock falls and avalanches, respectively. Seismo-acoustic signals from glacial outbursts sometimes contain discrete, impulsive signals related to glacier movement that occur in conjunction with outburst flood mobilization. Two glacial outbursts at Mount Spurr volcano, Alaska, in 1993 and 2013 were accompanied by discrete seismic events thought to be related to the settling of glacial ice in response to the release of subglacial water (Nye et al., 1995; Walter

et al., 2013). Outburst floods flowing under glaciers can be harmonic due to the resonance of cracks and conduits (Winberry et al., 2009). This can make subglacial floods difficult to distinguish from volcanic tremor, an important consideration in settings such as Iceland where tremor from both volcanoes and outburst floods can occur in similar locations below the ice sheets (e.g., Eibl et al., 2015).

Signals from glacier basal slip are especially problematic for volcano monitoring, as such signals can be indistinguishable in character from low-frequency volcanic earthquakes (e.g., Thelen et al., 2013). Repeating sequences of glacier basal-slip have been observed at several glacier-clad volcanoes and are difficult to distinguish from repeating volcanic earthquake sequences that sometimes precede or accompany eruptions (Weaver and Malone, 1979; Jónsdóttir et al., 2009; Allstadt and Malone, 2014). Repetitive glacier basal slip is likely a similar phenomenon to precursory quakes that have been observed accelerating in recurrence prior to multiple ice avalanches on Iliamna volcano, Alaska (Caplan-Auerbach and Huggel, 2007)(Figure 5). Such repeating precursory sequences are likely due to stick-slip motion at the ice-rock interface and therefore their signals are akin to small earthquakes (e.g., Thelen et al., 2013). In some cases, however, the stick-slip events precede glacial failure, for which the seismic signal is like that of other mass wasting events (Caplan-Auerbach and Huggel, 2007). The presence of ice may promote this type of behavior, but it is not required, since similar behavior has been observed prior to some landslides that did not directly involve ice (Yamada et al., 2016, Poli et al., 2017; Schöpa et al., 2018).

Snow avalanches (Figure 1G) also occur frequently on steep volcanoes in climates that experience significant snowfall and generate seismic and infrasonic signals. However, their signals tend to be weak, seismically visible only within a few kilometers (e.g., Allstadt et al., 2017) so they are less problematic to volcano monitoring than glacier seismicity. In contrast, infrasound waves from large snow avalanches have been observed hundreds of kilometers from the source (Bedard, 1989), though this is likely only under ideal propagation and recording conditions. The literature on infrasound and seismic studies of snow avalanches is potentially the most extensive of any mass movement type discussed here (e.g., Kishimura and Izumi, 1997; Sabot et al., 1998; Suriñach et al., 2000; Biescas et al., 2003; Comey and Mendenhall, 2004; Kogelnig et al., 2011; Havens et al., 2014). This is likely driven by the demand for avalanche

monitoring for risk management in mountainous areas, as well as the fact that snow avalanches are easier to study. They can be triggered artificially, the location of avalanche-prone slopes can be easily identified, and avalanches can reoccur many times over the same flow path. These factors allow for more experimental control in seismic and acoustic studies and provide a good opportunity to gain insight applicable to the less-easily studied surface mass movements if differences in material type and physical process can be taken into account. The seismic signals of snow avalanches are similar to those from other surface flows in the high-frequency range (>1 Hz), with emergent onsets and long durations (Kishimura and Izumi, 1997; Suriñach et al., 2000). The seismic signal is thought to be generated primarily by interactions of the flow with the terrain and obstacles (Sabot et al., 1998; Suriñach et al., 2000). In contrast, infrasound signals are thought to be primarily generated by the turbulent motion of the snow and air at the avalanche front (Naugolnykh and Bedard, 2002). An example of a snow-avalanche infrasound signal is shown in Figure 3A-4 (Havens et al., 2014).

2.3 Recurrence and time clustering

In general, larger mass movements are less common than smaller mass movements (McGuire, 1996). Regional landslide inventory mapping has revealed power-law scaling between landslide size and the frequency of occurrence of medium and large landslides (areas $> \sim 1000 \text{ m}^2$) (Hovius et al., 1997; Guzzetti et al., 2002; Malamud et al., 2004). This suggests that if all mass movements in a region could be perfectly detected and characterized using seismo-acoustic methods, power-law scaling akin to the Gutenberg-Richter frequency-magnitude relation for earthquakes (Stein and Wyss, 2002) might be expected. McGuire (1996) illustrated this concept for volcanic mass movements schematically (Figure 6). The actual frequency of occurrence of a given type and size of mass movement in a given region depends heavily on susceptibility (steepness, slope strength, vegetation) and the availability and occurrence of triggers (storms, earthquakes, volcanic eruptions).

Similar to earthquakes, mass movements commonly occur in temporal clusters. Due to their precise time resolution, seismo-acoustic data are well suited to studying these temporal patterns. Such studies are important for providing insight into the temporal relation of smaller and larger landslides, which can in turn help with hazard assessments. For example, after analyzing four debris avalanches at Mount Rainier volcano, Washington, and finding that each was composed

of multiple signals minutes to hours apart, Norris (1994) suggested that progressive failure was actually the norm for Mount Rainier and that areas experiencing an uptick in rock fall activity should be avoided for some time afterwards.

The clustering of mass movements in space and time may in fact be the norm in many settings. Rock falls in particular occur in clusters, especially during periods of volcanic unrest. Both infrasound and seismic methods have been used to count the occurrence and relative size of rock falls (e.g., Caplan-Auerbach et al., 2004; Hibert et al., 2011; Johnson and Ronan, 2015). Clustering can reflect the influence of changes in stress due to various volcanic processes. For example, intense rock fall activity at Makaopuhi crater at Kilauea, Hawaii, in 1972 correlated with eruptive activity and was inferred to be due to changes in stress patterns in the crater wall (Tilling et al., 1975). Another example is the shift from earthquake to rock fall seismic signals that marked the start of the dome-building eruptions at Mount St. Helens in 1981-1986 (Malone et al., 1983). Clustering may also occur because one event unbuttresses the slope above, as observed in sand models (Acocella, 2005). There is evidence that this phenomenon also occurs in volcanic systems: in clusters of submarine landslides at West Mata submarine volcano, source locations were found to move gradually closer to the ocean surface, suggesting progressive failure up the flank (Caplan-Auerbach et al., 2014; Drobiarz, 2017).

A common triggering factor can also cause a cluster of events. For example, multiple landslides can be triggered by one precipitation event (e.g., Helmstetter and Garambois, 2010; Hibert et al., 2011) or earthquake (e.g., Dai et al., 2017).

Some cases of time clustering are similar to foreshock-mainshock-aftershock sequences, where one or more smaller mass movements precedes and follows the largest collapse (e.g., Allstadt, 2013a; Coe et al., 2016; Gualtieri and Ekstrom, 2018). Smaller collapses preceding the main event may be a side effect of accelerating deformation as the larger mass begins to mobilize. Alternatively, the smaller event(s) could trigger the larger event if the stress perturbation is sufficiently destabilizing. Smaller collapses that decay in number over time can follow a large event, akin to earthquake aftershocks (e.g., Allstadt et al., 2017; Moore et al., 2017), which may reflect the adjustment of the slope to the new stress configuration.

3. Seismic Theory

Seismic waves are elastic waves traveling through the Earth. A given observed seismic signal can be simplified as a convolution of the source signal $s(t)$, the effects of earth's structure on the signal along its path $g(t)$, and the response of the seismometer $r(t)$ (Stein and Wysession, 2003, pg. 379):

$$\mathbf{u}(t) = \mathbf{s}(t) * \mathbf{g}(t) * \mathbf{r}(t)$$

Equation 1

Typically, we are most interested in using the seismic signal recorded at a given station $\mathbf{u}(t)$ to obtain the source signal $\mathbf{s}(t)$, which we want to then relate to characteristics of the source process. This requires deconvolving both the effects of earth's structure $\mathbf{g}(t)$ and the response of the seismometer $\mathbf{r}(t)$ from $\mathbf{u}(t)$. Correcting for the response of the seismometer $\mathbf{r}(t)$ is the easiest step, as procedures for removing instrument response are well-established and seismic data is often archived by professional data centers with detailed instrument response information (e.g., IRIS Data Management Center, <https://ds.iris.edu>). For a review of instrument correction, we refer the reader to Haney et al. (2012). Isolating the effects of the earth's structure $\mathbf{g}(t)$, on the other hand, is exceptionally challenging in volcanic regions and we will discuss this in detail later in this section. Here, we will focus on the source process itself because even if we could easily obtain $\mathbf{s}(t)$, we still need to know how it relates to event dynamics if we want to use it to understand, study, or characterize the mass movement. This requires an understanding of how surface mass movements actually generate seismic waves.

3.1 Source theory

Seismic waves generated by mass movements are the result of spatially and temporally variable tractions applied to the surface of the earth by the moving material. In order to be seismically observable, the source process needs to occur over timescales that are in the seismic frequency band, which is technically less than about an hour (Stein and Wysession, 2003), but practically much shorter, less than about a minute, due to noise and instrumentation limits. For this reason, a slowly deforming landslide is generally not seismogenic regardless of size, though effects of the slow deformation such as brittle cracking, stick-slip sliding, temporal changes in seismic velocity, or small rock falls may be seismically observed (Spillmann et al., 2007; Helmstetter and Garambois, 2010; Mainsant et al., 2012; Yamada et al., 2016). For the same reason, it is not static stresses that are important so much as rapid changes in basal tractions due to spatial and temporal changes in velocity and flow thickness (e.g., Sabot et al., 1998; Suriñach et al., 2000; Favreau et al., 2010), essentially, unsteady flow. Even in the simplest cases, the source process of a surface mass movement in reality consists of processes occurring over many scales simultaneously over an extended and changing

footprint. This range of scales is, in part, responsible for the broadband frequency content often observed for mass movement signals (e.g., Figure 2B). To understand this, consider if, in

Equation 1, we hold $g(t)$ and $r(t)$ constant and vary only the source signal $s(t)$, a longer duration source signal will result in a lowering of the frequency content of the signal. A shorter duration source will have the opposite effect. Therefore, a flow in which there are momentum exchanges related to the flowing material encountering bumps and curves in many different locations over many different length scales (and thus durations) at the same time would yield a diverse set of simultaneous source signals and thus result in a similarly broad range of frequency content in the resulting signal. That signal would, in theory, be dependent on changes in local flow velocities, which in turn depend on details of the flow dynamics and topography of the sliding path.

Practically speaking, this concept is useful to explain lower frequency (< 1 Hz) seismic observations but explaining the higher frequency energy observed in the context of this or other theoretical models has been more enigmatic because higher frequencies attenuate rapidly and are more sensitive to small-scale heterogeneities and signals are often complex (e.g., Hibert et al., 2017a; Levy et al., 2015). In studies where the source process of the higher frequency energy has been investigated, the signal is often treated as a reflection of a stochastic process and correlations between amplitudes and modeled bulk flow characteristics are sought. Schneider et al. (2010) found that the 1-20 Hz signal correlated best with frictional work rate, though momentum and total kinetic energy also showed strong correlations. Hibert et al. (2017) found that a high correlation between time series of landslide momentum and envelopes of 3-10 Hz energy, Levy et al. (2015) found that observed seismic energy > 1 Hz correlated best with the simulated forces applied to the ground by the flow. Though all three studies only considered large, rapid landslides.

It is helpful, especially when considering the source process of the highest frequency energy in a flow, to abandon the continuum approach where instead of considering continuous tractions, consider the forces exerted by the individual particle interactions with the flow bed. The forces generated by individual particle interactions are generally treated as many random “impacts” with distributions of fluctuating velocity, particle size, and timing between impacts. The distribution used should reflect the type of flow that is being modeled, for example particles

accelerated by gravity, such as a saltating particle on a downward trajectory or impact of a rock fall block, would be modeled differently than impacts of agitated particles at the base of a flow that deviate from the mean flow velocity. Theoretical models of this type, that consider only impacts of individual particles and the higher frequency wavefield, have been proposed for (e.g., Tsai et al., 2012) and debris flows (Kean et al., 2015, Lai et al., 2018). Iverson (1997) observed that the fluctuations of normal stress measured at the base of a debris flow were much higher when measured on a force plate of about the same length scale as individual particles than when measured on larger length scales (Figure 7). This indicates that the bouncing, rolling, and sliding of individual particles may be important to seismogenesis on smaller scales (i.e., higher frequencies), even if the forces average out over the larger spatial scales typically considered in a continuum approach.

As with the momentum exchanges in the continuum discussed earlier, individual impacts also radiate energy at frequencies that are inversely proportional to the duration of the impact. The duration of the impact is longer for a larger mass at a slower speed, meaning a larger slower impact or collision would radiate lower frequencies than a smaller higher-velocity impact (e.g., Tsai et al., 2012; Farin et al., 2015). The total radiated energy, on the other hand, is higher for both a larger mass and larger impact velocity (Farin et al., 2015). Therefore, consistent with intuition, an impact from a block in a rock fall with a significant free-fall component would radiate more seismic energy than an individual saltating particle of the same size at the base of a flood, if all else is equal, because the impact velocity would be higher. Similarly, a flow with larger particles would generate lower frequencies than a similar flow composed of smaller particles, consistent with observations (e.g., Burtin et al., 2009) and simulations (e.g., Feng et al., 2017). Though Tsai et al. (2012) and Lai et al. (2018) argue that the duration of the impact can be assumed to be elastic and instantaneous for the frequency ranges typically of interest, and therefore that there may not be a direct dependence between frequency and particle size, though they consider only impacts on bedrock, not a common substrate for many types of flows in volcanic settings.

The grain-size distribution can evolve throughout the flow duration and vary spatially within parts of the flow due to grain-size segregation, deposition and entrainment, fracturing of the source volume, and other details of the event dynamics. These changes can be observed

seismically in the form of, for example, changing frequency content and amplitudes over time (e.g., Suriñach et al., 2005; Cole et al., 2009; Zobin et al., 2009). However, the source is typically moving and also spatially elongated. Therefore, changes in frequency related to changes in the source process must be carefully distinguished from changes in frequency related to attenuation and path effects (Section 3.4). This is because source-to-station distances may change significantly over the course of an event and seismic energy may be coming from multiple locations simultaneously.

In movements that involve significant proportions of fluid (e.g., hyperconcentrated flows, outburst floods), turbulent flow and breaking waves can also generate seismic noise in the higher frequency range (~1-100 Hz) but lower in frequency than the signal from the individual impacts of bedload (Schmandt et al., 2013; Gimbert et al., 2014). Coherent turbulence structures and gravity waves are thought to be capable of generating lower frequency energy (Gimbert et al., 2014). Fluids may still play a role in seismogenesis in non-fluid motion; for example, variations in basal stress due to pore fluid fluctuations in a debris flow (Iverson and LaHusen, 1989) could generate seismic waves. However, the importance of this fluctuation relative to other potential sources of seismic energy has not been explored, to our knowledge.

Clearly, the problem of understanding the seismic wavefield source process can become complex quite fast when considering all scales of behavior in realistic mass movements. However, there is a special case where a simple solution exists: if the mass movement can be approximated as a rigid block where displacements between the block and the earth are the same everywhere and the material above this surface does not deform, then at the long-wavelength (long-period) limit, the wavefield generated by surface tractions between the earth and the sliding block is identical to a single body force vector $\vec{f}(t)$ applied at the event centroid that varies with time. The force is equal but opposite in direction to the mass of the moving material m multiplied by the acceleration vector $\vec{a}(t)$ of the entire mass:

$$\vec{f}(t) = -m\vec{a}(t)$$

Equation 2

(Kawakatsu, 1989).

Following a similar approach as that used to show that the equivalent source mechanism to slip on a fault is a double couple, Fukao (1995) uses Gauss' theorem to show that the surface tractions due to slip on the failure plane produce the same seismic wavefield as if a body force was applied simultaneously to every particle in the source volume equal to the particle's density times the acceleration of the block, which when integrated over the volume at the long wavelength limit equates to Equation 2. This is an important simplification because it is much easier to model the seismic wavefield generated by a body force applied at a single point at long periods (long wavelengths), than to model the wavefield generated by temporally varying surface tractions. Long wavelengths attenuate slowly and are insensitive to subsurface heterogeneities so simple 1D earth models can be used. Because it is possible to model the long-period (~ 0.1 - 0.001 Hz) seismic wavefield from a single body force without requiring detailed information about the subsurface or topography, long-period seismic waves generated by a mass movement that can be well-approximated as a rigid sliding block can be inverted to obtain the single force vector time series that best fits the observed seismic data. Then, if the mass is known or can be estimated, the trajectory can be obtained. Alternatively, if the trajectory is known, the mass can be obtained (Equation 2).

Though, there are few, if any, mass movements that actually fully meet the criteria required, many studies have found the single-force approximation to be a reasonable model for many large ($\gg 1$ million m^3), rapid landslides with relatively simple failure sequences when studied using wavelengths much longer than the size of the event (e.g., Kanamori and Given, 1982; Kawakatsu, 1989; Brodsky et al., 2003; Ekstrom and Stark, 2013). In reality, most masses deform significantly, consist of multiple collapses closely spaced in time, and entrain and deposit sediment during the event. While the single-force approximation can still be useful, these factors complicate the interpretation (Allstadt, 2013b; Moretti et al., 2015; Coe et al., 2016). Single-force inversions can be used to roughly estimate mass and trajectory for simple landslides (Ekström and Stark, 2013; Yamada et al., 2013; Hibert et al., 2014a; Schöpa et al., 2018), to understand the sequence of events (e.g., Allstadt, 2013b; Iverson et al., 2015; Moore et al., 2017), to extract quantitative information such as velocity and basal friction (Brodsky et al., 2003; Moretti et al., 2012; Yamada et al., 2013), and to constrain numerical landslide models (Favreau et al., 2010; Moretti et al., 2015; Yamada et al., 2016).

Unfortunately, while this single-force approximation should technically apply to a mass movement of any size that meets or approximates the assumptions of the model, long-period seismic waves are rarely observed above the noise level for moderate and smaller-sized events. There are, at most, a handful of events worldwide in a given year that generate strong long-period seismic waves and can be studied this way (e.g., Ekstrom and Stark, 2013). The extension of the single-force mechanism to smaller events is tenuous (e.g., Dammeier et al., 2015). This may be because smaller masses result in lower force amplitudes and thus lower seismic amplitudes, but also because smaller events may have shorter durations and thus the waves generated by the bulk acceleration would have shorter periods that may overlap with the two strongest peaks in earth's ambient noise field which occur at periods of 4-8 sec and 10-16 sec (McNamara and Buland, 2004). Shorter period waves also attenuate much more rapidly and are more sensitive to subsurface heterogeneities so simple 1D earth models are no longer sufficient.

Many of the events studied using the single-force approach occurred in volcanic settings (e.g., Kanamori and Given, 1982; Kawakatsu, 1989; Brodsky et al., 2003; Moretti et al., 2012; Allstadt, 2013b), likely because volcanoes are a common source of large, rapid mass movements. The first was a study of the seismic signals from the Mount St. Helens 1980 sector collapse and debris avalanche (Kanamori and Given, 1982), where the single-force mechanism was first proposed. The largest recorded example of a single-force landslide source was a massive ($\sim 10^3 - 10^4 \text{ km}^3$) tsunamigenic submarine slump off the flank of Kilauea volcano, Hawaii (Eissler and Kanamori, 1987), though this source designation has never been definitively settled (Ma et al., 1999). A complicating factor of the use of the single-force approximation in volcanic settings is that large mass movements that occur as a result of eruptive activity can be concurrent with other seismic sources (explosions, very long period earthquakes). Some of these other sources can also be represented seismically using single-force mechanisms and produce seismic energy in similar frequency bands (e.g., Nishimura, 1998; Chouet et al., 2003; Haney et al., 2013). These similarities can make it difficult to separate the contributions from different sources.

3.2 Seismic Efficiency

In addition to the source process described above, another factor that controls the seismic signal generated at the source, $s(t)$, is how much of the total energy released is converted into seismic waves. Seismic efficiency is the ratio between the total radiated seismic energy and the change in

potential energy of the event. Unlike the elastic strain energy released by earthquakes, mass movements release gravitational potential energy (Fukao, 1995). Only a fraction of the potential energy loss is converted into seismic energy, the rest is dissipated by other means such as frictional heating, plastic deformation, or fragmentation. Seismic efficiency ratios that have been estimated for surface mass movements range widely from 10^{-6} to 10^{-3} (Berrocal et al., 1978; Deparis et al., 2008; Hibert et al. 2011; Levy et al., 2015) with a few outliers (e.g., Vilajosana et al., 2008, ~ 0.2). Earthquakes, in comparison, typically have a higher seismic efficiency, where the ratio of radiated seismic energy to stress drop is generally around 0.06 (McGarr, 1999). This suggests significant differences in energy partitioning between the processes. The wide range in estimated ratios for mass movements may be partially due to methodological and instrumentation differences (e.g., Levy et al., 2015) and uncertainties and assumptions made about wave types, attenuation, and velocity structure in the computation of radiated seismic energy. Radiated energy measurements for earthquakes can vary by a factor of 2 or more because of these uncertainties (e.g., Shearer, 2009, pg. 275).

Another reason for the wide range of seismic efficiency values reported in the literature is that the efficiency can depend heavily on the style of movement and the surface over which the flow is traveling. Tilling et al. (1975) noted that a rock fall moving over a gentle slope onto mushy cooling lava in Hawaii generated no observable signal, while similarly sized rock falls with primarily vertical trajectories falling onto solid ground had very clear signals on the same station. Similarly, Hibert et al. (2011) found that rock falls with a free-fall component seemed to be more seismically efficient than granular flows. Numerous authors have noted that flows and impacts onto looser soils generally have lower seismic efficiencies than those on harder surfaces such as bedrock. For example, Cole et al. (2009) compared signals from several lahars with different characteristics passing the same site and found that a flow initially was much quieter seismically than the very similar previous flow until the second flow eroded down through the deposits of the first, at which point it became more energetic. They concluded that the loosely compacted deposits were damping the signal. Similarly, Kean et al. (2015) found that seismic amplitudes of debris flows over a bedrock channel were an order of magnitude higher than for similarly sized flows in the same channel when it was coated by a relatively thin layer of loose deposits, an observation the authors attribute to the dissipation of impact energy into friction due to inelastic collisions with multiple points of contact when particles impacted deformable, loose deposits.

3.3 Seismic Wavefield

In this section, we discuss the characteristics of the seismic wavefield that is generated at the source and how it is partitioned between different wave types. For context, we first briefly review the three main types of seismic waves (for detailed background the reader is referred to Shearer, 2009). Within an elastic continuum, the fastest seismic waves generated by the source and the first to arrive at a given station are *P* waves. *P* waves are compressional—the ground motion is in the same direction as wave propagation. *S* waves, also called shear or transverse waves, arrive next, typically with higher amplitudes and ground motion that is perpendicular to the direction of wave propagation. *S* waves can be further subdivided into horizontally (*SH*) and vertically (*SV*) polarized waves. The first arrival of each wave type is referred to as a direct phase arrival. Surface waves (e.g., Rayleigh and Love waves) arrive after *S* waves, typically with even higher amplitudes. Unlike the previous two types, which are body waves that travel through the Earth, surface waves propagate along the Earth's surface. Rayleigh waves have particle motion in the vertical plane; Love waves, like *S* waves, have transverse motion. In layered media, surface waves are dispersive; lower frequencies typically travel faster than higher frequencies.

Theoretically, the partitioning of the radiated seismic energy into different wave types depends on the orientation of the generating forces. When the seismogenic source process consists primarily of vertical forces, Rayleigh waves are dominant (e.g., Tsai et al., 2012). When there is a horizontal component, Love waves will be important as well, and the proportions will depend on the ratios of the horizontal to vertical force components (e.g., Gualtieri and Ekström, 2016). This is demonstrated quantitatively in Figure 8, which shows the dependence of seismic wave type on orientation of a seismic force source at the surface of an earth model of constant velocity (half-space). These quantities are derived by integrating waves over a hemisphere (body waves) or a cylinder (surface waves) at large distances (far-field) from the source based on the original solutions for vertical and horizontal forces in seminal papers by Miller and Pursey (1955) and Cherry (1962). Those solutions can be considered spatially-integrated, time-harmonic versions of what is known as Lamb's problem, which considers the seismic response of a point force at the free surface of an elastic halfspace (Lamb, 1904; Aki and Richards, 2002; Gualtieri and Ekström, 2016). The ratio between Rayleigh and *SH*-polarized waves trades off as the force vector becomes more horizontal, and at an angle of approximately 37 degrees the two wave types are

generated in equal proportion. P and SV waves show less variation in terms of radiated energy as a function of the force orientation and account for less than one-third of the energy at all angles. Though it is sometimes assumed that surface waves dominate for surface events, Figure 8 demonstrates that body waves can contribute significantly to the generated wavefield, most markedly when force angles are shallow, though path effects (Section 3.4) may change the ratios by the time they are recorded. Note that, since the halfspace is homogeneous, Love waves are not addressed by the existing theory (Cherry, 1962) and are not shown in Figure 8. For a realistic layering, most of the SH-polarized energy would likely propagate as Love waves instead of SH body waves.

3.4 Attenuation and Path Effects

So far in this section we have considered all of the factors that control the wavefield generated at the source, $\mathbf{s}(t)$. Now we consider how this signal changes as it travels from the source to the seismic station, the effects of the earth's structure, $\mathbf{g}(t)$. The energy partitioning between different wave types (Section 3.3) may change substantially from what was emitted at the source by the time the wavefield is recorded on seismometers due to geometrical spreading, scattering, and attenuation that occurs along the path between source and seismic station. The simplest is geometrical spreading, which is simply the conservation of energy across a wavefront. Assuming a homogeneous medium, in the case of body waves, the wavefront is spherical, for surface waves it is cylindrical. As a result, energy per unit wavelength decays at a rate of $1/r^2$ for body waves and $1/r$ for surface waves (e.g., Stein and Wysession, 2003). This implies that for distal observations, regardless of the initial partitioning, the energy recorded may be composed dominantly of surface waves. However, this difference may be offset by differences in intrinsic attenuation along the paths each type of wave takes. Body waves can take refractive or reflective paths through deeper less-attenuating layers, whereas surface waves have to propagate through the highly attenuating near-surface. This has a particularly strong effect on higher frequency surface waves (smaller wavelengths) that are sensitive to the shallowest layers where attenuation is higher.

While most seismic signals from surface mass movements lack clearly identifiable phase arrivals due to their extended and complex source processes, many authors have found that the signals observed tend to be dominated by surface waves based on particle motion analysis (e.g.,

Vilajosana et al., 2007; Vilajosana et al., 2008; Uhrhammer, 2009; Dammeier et al., 2011; Gualtieri and Ekstrom, 2016), slow wave propagation speeds (e.g., Tilling et al. 1975), and the fact that shallow sources, such as those from surface mass movements, produce waves that graze and interact with the free surface and encourage the generation of surface waves (Neuberg and Pointer, 2000).

In addition to geometrical spreading, volcanoes, with their heterogeneous, layered structure and potentially large contrasts in seismic wave velocities, can have dramatic path effects that significantly affect the observed waveform. Path effects describe scattering and attenuation that occurs along the path between source and seismic station. In volcanic settings, seismic signals are commonly classified into event types based on their frequency and/or duration. Path effects may significantly distort the recorded signal, complicating the interpretation. In the most extreme cases, two sources with completely different initial signal characteristics may be so altered by the filtering effect of the volcano that they look quite similar by the time they reach a seismometer (Weaver and Malone, 1976; Weaver and Malone, 1979).

Intrinsic attenuation arises from anelastic properties of the rock through which seismic waves travel, whereas scattering attenuation arises from interactions with small-scale heterogeneities (Aki and Richards, 2002). While both intrinsic and scattering attenuation are frequency dependent, scattering attenuation is typically more dependent on frequency than intrinsic attenuation (Mayeda et al., 1992). Topographic interaction and interactions of waves with layer boundaries, which result in the reflection or transmission of a particular wave, also further alter seismic waveforms (Neuberg and Pointer, 2000).

Practically, intrinsic and scattering attenuation filter out higher frequencies and scattering attenuation increases the duration of the signal. The intrinsic quality factor (Q) quantifies attenuation as the inverse of the fractional energy loss per wavelength (Shearer, 2009). Intrinsic Q is frequency dependent, inversely proportional to damping, and is dependent largely on the composition of the rock. In addition, Q tends to increase with depth, especially at volcanoes (e.g., Mayeda et al., 1992; Tusa et al., 2004). A couple of end-member examples include a highly-attenuating stratovolcano such as Mount St. Helens, WA, with a Q of ~ 130 at 10 Hz (Haskov et al., 1989) and a less-attenuating basaltic shield volcano such as Kilauea, Hawaii, with a Q of ~ 430 at 9 Hz (Mayeda et al., 1992).

4. Acoustic Theory

Atmospheric acoustic waves are volumetric longitudinal waves and can be thought of as *P* waves traveling through the fluid air (*S* waves cannot propagate in a fluid because a fluid cannot support shear). Shallow subsurface and subaerial volcanic processes generate a wide variety of atmospheric acoustic signals, including explosive eruptions, shallow degassing, surface flow, and mass movements. Acoustic methods provide a natural complement to seismic monitoring of surficial mass movements at volcanoes.

4.1 Source theory

The theory developed for audible acoustic frequencies is generally valid for infrasound (frequencies less than 20 Hz) down to the acoustic cutoff frequency, which is about 3.3 mHz in the lower atmosphere (Evers and Haak, 2010). It is commonly useful and appropriate to use a linear equivalent source representation for infrasound (for further information on equivalent source theory in acoustics, see Lighthill, 1962; Morse and Ingard, 1968; Lighthill, 1978; Pierce, 1989). Outside a theoretical surface bounding an arbitrary source of any degree of complexity, the source can be represented as a sum of equivalent sources provided that the source region is compact (i.e., small compared to the wavelength divided by 2π). Seismic source theory commonly uses a representation in terms of a moment-tensor of force couples or a single-force vector. In contrast to the single-force mechanism from seismic source theory, in acoustics, the monopole is the most elementary linear equivalent source and radiates sound equally in all directions (a monopole is also called a simple source or point source). For the monopole, acoustic pressure is related to the rate of change of the mass outflow rate at the source. Higher-order equivalent sources (dipoles and quadrupoles) can be mathematically constructed from combinations of monopoles (Lighthill, 1962; Morse and Ingard, 1968; Lighthill, 1978; Pierce, 1989). The relationship between acoustic and seismic equivalent sources was recently investigated by Haney et al. (2018).

Source theory for infrasound from mass movements is currently not well understood. Moran et al. (2008) applied a monopole source model to an unusual 50 s very-long-period (VLP) infrasound signal associated with a large rock fall at Mount St. Helens (Figure 3c). However, microbarometers with the response necessary to observe the VLP signal were only installed at

one array ~13.4 km from the source, effectively representing a point-sample of the wavefield at a single azimuth. Thus, it was not possible to observe any directivity in the radiation pattern of the rock fall event that may have been present. For example, it is possible that a piston-like push of air away from the nose of a directed mass movement may be better modeled by an acoustic dipole, which is equivalent to a force acting on the air (e.g., Russell et al., 1999; Haney et al., 2018). A common example of a dipole source in audible acoustics is a loudspeaker set in a flat panel (e.g., Russell et al., 1999). Motion of the loudspeaker cone produces a directional (dipolar) radiation pattern. Future field studies with more sensors at different azimuths from the source are needed to assess directivity. In addition, higher frequency source components, as commonly indicated by the broadband records of lahars or pyroclastic flows, likely represent much more complex sources for which a single equivalent source assumption is too simplistic. For example, in the case of debris flows and pyroclastic flows, the elongated source may best be modeled as an extended distribution of summed higher-frequency monopoles. Furthermore, for some event types, turbulence may be a significant source of infrasound, but this has not been quantitatively addressed as far as we are aware. Turbulence is thought to be a major source of infrasound in volcanic eruptions involving jet flow (Matoza et al., 2013, 2009). Further work is required to quantify and assess the role of turbulence as an infrasound source for flows.

4.2 Wave types and propagation

A review of infrasound wave propagation in volcanic environments is provided by Fee and Matoza (2013). To first order, infrasound propagation in the atmosphere is controlled by vertical gradients in temperature and horizontal winds. Infrasound propagation at local distances from volcanoes (<10 km) is primarily influenced by the atmospheric boundary layer, including microscale interactions with volcanic topography (e.g., Fee and Garcés, 2007; Marcillo and Johnson, 2010; Matoza et al., 2009; Johnson et al., 2012; Lacanna et al., 2014; Kim et al., 2018), which can diffract, reflect, and even effectively block the signal. Topographic interactions can control whether or not the infrasound signal from a surface event is detectable at a given near-source location (Kogelnig et al., 2014; Johnson and Palma, 2015). For larger events recorded at longer ranges, infrasound propagation takes place in waveguides formed by the vertical temperature and wind structure of the atmosphere. The majority of long-range infrasound propagation occurs in the anisotropic waveguide formed by stratospheric wind jets at altitudes

between 35 and 55 km (Le Pichon et al., 2009). An additional approximately isotropic waveguide is formed by refraction at altitudes between 90 and 120 km from the strong temperature gradient in the thermosphere. In addition, long-range directed ducting in the troposphere is also possible—usually as a result of strong winds. Absorption is relatively low for atmospheric infrasound, approximately 5×10^{-5} dB/km at 1 Hz (Sutherland and Bass, 2004). Acoustic waves generated by most mass movements are relatively low power and therefore only recordable at local distances from volcanoes; thus, boundary layer variability and tropospheric ducting are the most likely propagation scenarios (e.g., Fee and Garces, 2007; Kim et al., 2018). However, in the case of major volcanic landslides such as the 1980 Mount St. Helens debris avalanche, regional (hundreds of kilometers) to global (thousands of kilometers) infrasound propagation is relevant (Reed, 1987; Delclos et al., 1990).

5. Detection, Location, and Characterization

5.1 Detection

5.1.1 Detectability

The ability to detect seismic waves from a given surface mass movement in a given frequency band is dependent on the type and size of the event (Section 3), the proximity and bandwidth of instrumentation, and noise levels. Empirically based estimates of seismic detection limits can be extracted for different frequency bands from catalogs of surface mass movements. Existing catalogs mostly include landslides (Dammeier et al. 2011; Chen et al., 2013; Allstadt et al., 2017). Domains of detectability, undetectability, and conditional detectability can be derived from such catalogs based on the farthest detections at high frequencies (>1 Hz) of different event types of different volumes. We have done this for different event types in Figure 9 using the catalog and 1-5 Hz detection information compiled by Allstadt et al., (2017). We outlined areas of conditional detectability that spans the range of inter-event-type variability by fitting a line to the datapoints and taking the slope of that line to create an upper and lower limit encompassing all detections and assume that all events below that line are detectable and all above are likely not detectable. This exercise shows that there is substantial variability in distance to furthest detection, even when grouped by event type, but in general, a fall is seismically detectable at a much greater distance than a debris flow composed of the same amount of material. Rock, ice,

and debris avalanches fall in between the two. Figure 9 only addresses one frequency band. The smallest volume for which seismic waves with periods of tens to hundreds of seconds are observable seems to lie around 10^5 to 10^6 m³ based on data reported by Dammeier et al. (2015) and Allstadt et al., (2017).

In addition to topographic effects and temporal variability in the atmosphere described in Section 4.2, the detectability of infrasound signals from mass movements also depends heavily on noise levels, as well as atmospheric propagation conditions. Wind-noise reduction and persistent ambient infrasound (clutter) discrimination represent major challenges for infrasound detection. Wind is the dominant noise source in the frequency band 0.01–5 Hz (Walker and Hedlin, 2010). At a given infrasound station, wind variations can account for 4 orders of magnitude difference in the background noise at a given frequency (Hedlin et al., 2002; Bowman et al., 2005; Brown et al., 2014). Wind noise varies with time of day, season, and geographic location (Bowman et al., 2005). In addition to the use of arrays, wind noise can be mitigated via site selection (e.g., locating the array in a wind-sheltered site such as a forest or area of dense vegetation) and/or by attaching a wind-noise reduction system to the infrasound sensor. Persistent sources of infrasound (e.g., oceanic microbaroms, surf infrasound, or anthropogenic activity such as airplanes, gas flares, wind-farms, or urban noise) can create false alarms and potentially interfere with or obscure signals of interest. Thus, site surveys that statistically quantify wind noise and ambient coherent infrasound arrivals are an important component of establishing an infrasound monitoring system (e.g., Matoza et al., 2007, 2013).

5.1.2 Detection methods

5.1.2.1 *Traditional earthquake techniques*

Surface mass movement detection by seismometers and/or geophones is common in volcanic environments throughout the world. Automated detection algorithms for earthquake monitoring typically use a short-term average to long-term average ratio (STA/LTA) (Allen, 1982). This is also the most straightforward means for seismically detecting mass movements, since STA/LTA is often already implemented on existing networks. STA/LTA detection is best for impulsive arrivals that typically are associated with local earthquakes, but also can work to detect mass-movement signals if they are not too emergent. However, STA/LTA algorithms can also

generate many false triggers on single stations. This can be mitigated by using STA/LTA triggers on multiple seismic stations.

Once a transient is identified, additional parameters must be considered in order to properly distinguish mass movements from other events (e.g., earthquakes, cultural noise). This can be done manually by analysts who are familiar with different signal types. Automated methods are less common. The challenge in a volcanic environment is that seismic records are often cluttered with emergent and elongated signals from many different types of sources, many too small to be recorded on more than a few stations, making it difficult to uniquely classify events.

Partsinevelos et al. (2016) used qualities of the STA/LTA detections such as pulse duration, multi-station detection, and relative arrivals between stations to successfully distinguish between rock fall onto a roadway and passing cars and walking people. STA/LTA ratios can be tuned for the expected signals and will vary by setting and event type. For example, Fuchs et al., (2018) found an STA window of 5 s and LTA window of 120 s with a trigger ratio of 4 worked well for detecting rock falls and rock slides using stations tens to 100's of km from the source on a regional network. Partsinevelos et al. (2016), used much shorter windows optimized to detect small rock falls from sensors tens of meters away, with an STA window of 0.002 s, and an LTA window of 0.4 s and a trigger ratio of 2. Dietze et al. (2017) found that STA window of 0.5 s, and an LTA of 90 s worked best to detect rock falls within ~1 km of the sensors.

Although the STA/LTA time windows and ratios implemented on existing monitoring networks are not typically optimized for mass movement signal detection, such events are still frequently detected and sometimes appear in earthquake catalogs. However, the inclusion of surface mass movements in earthquake catalogs is inconsistently handled and far from comprehensive. This is partly because such events are difficult to confirm or locate without independent verification, which is practical only for the largest events. It is also because source parameters of surface phenomena do not fit well in typical earthquake catalogs. For landslides and debris flows, we are interested in quantities such as total volume or discharge, runout distance and drop height, which we cannot yet easily obtain from their signals alone, while for earthquakes we are interested in quantities such as magnitude and epicentral location. Surface events are rarely recorded on enough stations to obtain a location, and those that do are difficult to locate with typical phase-picking methods because of the emergent character of the signals, as discussed in Section 5.2.

Equivalent earthquake magnitudes are sometimes computed for mass movement signals using amplitude or duration-based methods (e.g., Pankow et al., 2014). However, due to the difference in the physical process, such magnitudes do not hold the same physical meaning as for earthquake magnitudes (e.g., Lin et al., 2015) nor do they necessarily scale directly with event size. Even so, amplitude or duration-based magnitudes can be used to differentiate earthquake signals from mass movement signals. For example, Manconi et al. (2016) found that the ratio of local magnitude (M_L) to duration magnitude (M_D) could be used to discriminate regionally-recorded rock slide signals from earthquakes, where the ratio is about equal to 1 for earthquakes, and typically less than 0.8 for rockslides, reflecting that rock slide signals often have longer durations.

5.1.2.2 *Lahar detection with acoustic flow monitors*

Few seismically-based mass movement detection systems have ever truly been operationalized. The exception is lahar detection. The acoustic flow monitor (AFM) was developed in 1989 at the U.S. Geological Survey (USGS) Cascades Volcano Observatory (Hadley and LaHusen, 1995) as a low-cost lahar detection device. An AFM detects lahars using an exploration-class geophone (e.g., Sercel HF-10) whose frequency response of approximately 10-300 Hz acts as a high-pass filter to reduce the effect of eruption-related tremor or distant seismic sources on the AFM signal amplitude. AFMs are usually installed along the stream valley as part of a lahar-warning system. To conserve power and telemetry bandwidth, the AFM performs onboard processing and transmits the instantaneous peak ground velocity, averaged over just a fraction of a second, at intervals ranging from 2 to 10 minutes in three frequency bands (full band 10-300 Hz, low band 10-100 Hz and high band 100-300 Hz). A built-in threshold-duration event detector sends an alarm code if the chosen threshold-duration is exceeded. Relative amplitudes of the high- and low-frequency bands may help to distinguish watery flows from more viscous flows, with watery flows dominating energy in the higher frequency band (Hadley and LaHusen, 1995; Marcial et al., 1996).

An AFM threshold-duration alarm can be triggered by nearby vehicular traffic, people moving near the sensor, or wind and heavy storm noise. To reduce false alarms in a lahar detection system, multiple AFMs are usually installed at different points along the monitored drainage, typically within a few tens to hundreds of meters from the channel. The differences in AFM

alarm times between sites can be used to estimate flow velocity and confirm detection. Rain gages and tripwires have been added to AFMs to provide independent checks of the AFM seismic data. The tripwire allows for estimation of cross-sectional area of a lahar flow, which may be correlated to volume and planimetric area and therefore to the expected downstream extent of a lahar. This independent corroboration of lahar size is necessary because no general relation between AFM amplitude and discharge rate has yet been published, though some researchers have correlated AFM amplitudes and discharge rates for lahars within limited size ranges at specific AFM sites (Lavigne et al., 2000; Andrade et al. 2006b; Almeida, 2016; Andrade and Almeida, 2017).

AFM systems have been deployed for lahar detection at volcanoes around the world. They have been most effective in detecting small repetitive lahars, especially those heralded by eruptions or rainfall and used in conjunction with other monitoring techniques that detect conditions necessary for lahar formation. Table 2 summarizes volcanoes where AFM systems have been installed and how they have performed. AFM system and data from 1992 lahars at Pinatubo reported by Tungol and Regalado (1996) are shown in Figure 10.

Despite their success, AFMs provide only very simple, limited data and are becoming outdated as technology and methods improve. Some other considerations driving this shift away from AFMs include: 1) the requirement that the low-gain AFMs be installed close to the drainage—a factor that requires installing and maintaining AFM systems on every drainage for which lahar monitoring is desired. The cost of maintaining AFM systems on multiple drainages commonly results in fewer hazardous drainages being monitored, or AFM systems installed further downstream where several drainages have coalesced, reducing warning times; and 2) due to the infrequent nature of lahars at most sites and the inapplicability of AFM data to any other volcano-monitoring task, installation of new AFM networks and operation and maintenance of existing ones has often been given a low priority.

5.2 Event Location

5.2.1 Phase picking methods

Emergent arrivals make phase-picking location methods problematic because pick times are controlled by station noise levels and it can be impossible to cleanly identify first body-wave

arrival times. However, several studies have used phase picks for emergent arrivals based on the kurtosis, which is the fourth moment of the amplitude distribution of the windowed signal (Baillard et al., 2014). Hibert et al. (2014b) found that the kurtosis provided relatively accurate *P*-wave arrival times for locating rock falls on a local scale at Piton de la Fournaise volcano. Fuchs et al. (2018) successfully extended the method to obtain first arrival times of both rock falls and rock slides on a regional seismic network. With these pick times, the authors located 14 events with a mean deviation of 8 km from the known locations. Even when first arrivals are not clear upon manual inspection, kurtosis-based algorithms can detect a change in the statistical distribution of the time series when the signal changes from normally distributed noise to non-normally distributed signal plus noise—but only if the signal amplitude is sufficiently high for the change to be detectable. Kurtosis-based picking allows for the use of traditional phase-pick-based earthquake location methods. However, it is also prone to the same limitations (e.g., velocity model inaccuracies, poor network geometry) and, to our knowledge, has not been tested on a wide range of event types.

5.2.2 Array techniques

While traditional earthquake location and detection techniques may not often work well for mass movements, several techniques can be used to detect and locate events using seismic and acoustic arrays. Any arrangement of instruments can be called an array. However, in this section, we use the term array to refer to a collection of multiple sensors (array elements) where the sensor spacing is chosen with the dual goals that a target incident wave is coherent across the array and that noise is incoherent between individual array elements. Array techniques are useful because they can improve signal-to-noise ratios. However, the primary benefit in the context of mass movements is that the parameters of coherent incident waves can be estimated (e.g., back azimuth and apparent velocity across the array) using techniques such as beamforming, where waveform cross-correlation or stacking is used to find minor time shifts between similar waveforms arriving on different stations. This allows moving sources to be tracked if the array-to-source azimuth changes sufficiently; if a specific drainage or travel path can be assumed, then a moving source can potentially be located. With multiple arrays, triangulation between the arrays can be used to identify source locations in different time windows. Depending on the array

response, arrays can also be used to identify and separate multiple coincident sources (e.g., volcanic tremor occurring at the same time as a lahar).

Array design takes into account a range of factors, such as the number of array elements available (cost and logistics), the detection frequencies of interest (larger aperture arrays are optimal for lower frequency sources and vice versa), the expected wavelengths of those frequencies, (which depends on subsurface structure directly below the array), and site conditions. Site conditions include wind noise (e.g., Walker and Hedlin, 2010) and repetitive coherent ambient signals, such as microbaroms from distant ocean storms or cultural noise from local sources (e.g., Matoza et al., 2013). For an overview of arrays and array design, the reader is referred to Rost and Thomas (2002) for seismic arrays and to Christie and Campus (2010) for infrasound arrays.

Historically, the field of earthquake seismology has rarely deployed sensors in small-aperture arrays. Local and regional seismic networks are typically arranged in distributed, wide-aperture formations intended to maximize spatial coverage since earthquakes can be located effectively using *P*- and *S*-wave arrival times. Another reason is that seismic waves travel faster than sound waves in the atmosphere and therefore seismic arrays must have a wider aperture than infrasound arrays for similar frequency bands of interest. Wider apertures translate into more complex logistical challenges. The detection of mass movements using coherence on small-aperture arrays has only recently been reported by Burtin et al. (2009), Lacroix and Helmstetter (2011), and Lacroix et al. (2012). However, seismic arrays are capable of detecting moving sources on a local scale (Almendros et al., 2002) and thus we expect the use of coherence on seismic arrays for studying mass movements to increase in the future. One important consideration when using array techniques on seismic signals generated by mass movements is that, as discussed in Section 3.3, the energy may be dominated by surface waves. The advantage of this is that apparent velocities passing over the array are equal to the actual velocities so the expected surface waves can be easily identified and separated from body waves. The disadvantage is that surface waves are usually dispersive, meaning velocity is frequency-dependent. Lower frequencies generally travel faster than higher frequencies and phases at different frequencies may separate enough as the wavefield passes over the array that the signal becomes decorrelated if the frequency band used is too broad or corresponds with a steep part of the dispersion curve. Therefore, it is

important to characterize the subsurface below the potential array location and its dispersion characteristics when designing the array.

Infrasound array techniques, on the other hand, have more commonly been applied in volcanic settings for tracking moving and flowing distributed sources (e.g., Yamasato, 1997; Scott et al., 2007; Olivieri et al., 2011; Havens et al., 2014; Johnson and Palma, 2015; Marchetti et al., 2015; Schimmel et al., 2017). A single infrasound array can provide directional information (backazimuth), giving information on relative locations and velocities of moving sources with respect to the array (e.g., Matoza et al., 2007). Back projection of the observed infrasonic backazimuth onto assumed propagation paths based on topography models provides an estimate of the instantaneous velocity of the moving front (e.g., Ripepe et al., 2009; Marchetti et al., 2015; Johnson and Palma, 2015). Multiple infrasound arrays can be combined to locate sources using backazimuth cross-bearings (e.g., Matoza et al., 2017) or semblance methods (e.g., Ripepe and Marchetti, 2002). Networks composed either of single infrasound sensors or arrays of sensors provide information on relative arrival time and/or frequency content variations for improved tracking of moving sources. For example, the path and speed of pyroclastic flow fronts has been tracked using arrival times and Doppler shifts at a microphone network (Yamasato, 1997) and with array processing (Ripepe et al., 2009). Delle Donne et al. (2014) used infrasound and seismic signals from pyroclastic flows at Soufrière Hills in combination with thermal camera imagery to track multiple fronts. Similar source tracking using infrasound arrays has been applied to snow avalanches (e.g., Havens et al., 2014; Marchetti et al., 2015) and lahars (Johnson and Palma, 2015). In the latter case, a lahar at Villarica was recorded serendipitously for >2 hours on a three-element infrasound array (VIH) deployed 9.4 km from the summit and 4 km from lahar's closest approach. Johnson and Palma (2015) were able to track an initial lahar flow pulse down a drainage at an average speed of 38 m/s, until it evolved into a stationary tremor signal associated with the flow passing through a topographic notch (Figure 11). One potential downside to infrasound arrays for monitoring purposes is that topographic blocking and noise can obscure the signal. For example, of the 10 infrasound arrays deployed around Villarica by Johnson and Palma (2015), only one recorded the lahar well, which the authors attribute to topographic blocking.

5.2.3 Envelope cross correlation

Envelope cross-correlation (ECC), also sometimes referred to as signal migration, is a method similar to those described in the previous section but that does not require a small-aperture array (e.g., Obara, 2002; Burtin et al., 2009). The method is similar to beamforming in that the seismic traces, in this case the envelopes of the waveforms, from several stations are time shifted to find the set of offsets that give the maximum correlation (or maximum stacked energy). Rather than searching over a grid of slownesses and azimuths, the search is performed over a grid of source positions. This produces a probability density function (or correlation function) of the likely location in space. The downside is that searching over a grid requires computing travel times, which requires knowledge or assumptions about the velocity structure of the entire area searched. Although this method can be used with any seismic array, as for any location method, the geometry of the array strongly controls how effectively events can be located.

Though perhaps best known for its use to locate non-volcanic tremor (Obara, 2002; Wech and Creager, 2008), ECC is suited to any tremor-like source that lacks clear phase arrivals but has coherent variations in amplitude across a group of stations. The technique has been used to locate debris flows (Burtin et al., 2009), landslides (Chen et al., 2013; Chao et al., 2016), rockfalls (Dietze et al., 2017), and river bedload during floods (Chao et al., 2015). The accuracy of locations is strongly dependent on the network geometry and the quality of the velocity model used to estimate travel times for each grid point (Burtin et al., 2009; Chen et al., 2013). For example, Chen et al. (2013) were able to locate 12 landslides to within an average accuracy of 2 km using a seismic network spanning all of Taiwan, but this degree of accuracy required the availability of a regional seismic velocity model.

5.2.4 Amplitude location methods

Amplitude source location (ASL) is another technique used to locate seismic events for which traditional phase-pick based location methods are not possible. Since seismic amplitude is a function of distance to the source, relative differences in amplitudes measured by a network located near to and surrounding the seismic signal source can be used to locate the source.

The ASL technique, like ECC, uses a grid search over possible source locations, typically with grid spacing of a few hundred meters (Ogiso and Yomogida, 2015). For each grid point, time

window, and source amplitude, the theoretical seismic amplitudes at the location of each station are calculated given the seismic velocity model under the assumption of isotropic scattering. The process is repeated for each grid point. The output is a map of residuals for where the best source location is the grid point for which the residuals between the modeled amplitudes and actual amplitudes are lowest.

The signal recorded at each station is typically bandpass filtered to accentuate the frequency that best characterizes the event; in the case of small lahars, Kumagai et al. (2010) found that 5-12 Hz worked well, possibly because the more pronounced scattering of these higher frequencies reduces or removes the effects of the source radiation pattern on relative amplitudes (Takemura et al., 2009). Both body-wave and surface-wave spreading factors have been used successfully in the ASL method. Kumagai et al. (2010) tried both and found only small differences in locations for small debris flows. ASL can be implemented using existing seismic networks, though, as usual, the location precision depends strongly on network geometry. Varying the inputs such as the attenuation quality factor (Q), frequency bands, wave velocities, site amplification factors, and time-window length can change the resulting location, but not to a first-order importance (Battaglia and Aki, 2003; Kumagai et al., 2010; Ogiso and Yomogida, 2015). Site amplification corrections seem to have the strongest effect on location results.

For short duration, compact sources such as small rock falls that do not travel farther than is resolvable by the chosen grid, a single time window containing the entire signal may be used. For longer duration sources such as lahars that can travel many kilometers, several time windows may be necessary to track the flow over time. An example of an ASL-estimated location over time for a small lahar flowing down the north flank of Cotopaxi, Ecuador, is shown in Figure 12 (Kumagai et al., 2009). The figure also gives an indication of the resolution that could be expected for a network of similar sparsity. The location is on the correct flank of the mountain, but in some cases, is more clearly resolved to a specific drainage (Figure 12a) than others (Figure 12b).

ASL has been used for the location and tracking of lahars and debris flows (Kumagai et al., 2015, 2010, 2009; Ogiso and Yomogida, 2015; Walter et al., 2017), rock falls (Battaglia and Aki, 2002), and pyroclastic flows (Yamasato, 1997; Jolly et al., 2002). Similar to most of the techniques described here, ASL has only been used as a post-event research and proof-of-

concept tool. It also has promise as a potential real-time debris flow detector/locator with the added benefit of being able to also locate other event types such as rock falls and tremor. This would be a valuable contribution to hazard mitigation, since locations could help determine in real time which drainages are experiencing lahars and which are not. However, ASL has not yet been tested for automation for mass movement detection to our knowledge.

5.2.5 Long-period detection and location

As described in Section 3.1, very large and energetic surface mass movements generate long-period (tens to hundreds of seconds) surface waves that can be observed for thousands of kilometers (e.g., Kanamori and Given, 1982). Some researchers have developed techniques to automate detection and location using existing seismic networks, though none have focused solely on volcanic environments. One of the earliest results stemming from semi-automated detection of long-period signals was a study by Lin et al. (2010), who used 20 - 50 second surface waves to detect 52 landslide and submarine slumps during a typhoon in Taiwan in 2009. While detections were made by manually examining the seismic records during the storm, to locate the events, they performed double-couple and single-force inversions using pre-computed Greens functions over a grid to find the best fit. This method of searching over a grid using pre-computed Greens functions for the best fitting solution is called GRiD MT (Kawakatsu, 2003) and was originally developed for application to earthquakes.

A similar but slightly more automated approach used Global Centroid Moment Tensor (CMT) inversions. The Global CMT project (<http://www.globalcmt.org/>) continuously searches over a global grid for potential seismic sources using 35-150 sec surface waves recorded by the Global Seismic Network (Ekstrom, 2006). Most detected sources are earthquakes, but large landslides and glacial calving events also can be detected (Ekstrom and Stark, 2013). Landslides can be located within 20 to 100 km of the actual source. Potential landslide detections are not automatically reported currently, to our knowledge, though a real-time algorithm has been proposed by Lin et al. (2015).

A major issue with the application of such methods to volcanic settings is that the long wavelengths result in a location uncertainty that can be much larger than the total footprint of a given volcano. Another issue is differentiating landslide sources from other sources that generate

long-period surface waves (e.g., teleseismic earthquakes, explosions). One potential approach would be to use the ratio of the long-period energy to short-period energy, as originally suggested by Weichert et al. (1994) who observed that, compared to earthquakes and explosions, landslides had much higher surface-wave magnitudes than body-wave magnitudes (Figure 13). Chao et al. (2016) extends this idea further and proposes a hybrid approach that combines long-period detection and force inversion with more precise locations using ECC on the high-frequency waveforms. However, their study was based on a set of known events, so whether the approach could be implemented effectively in real-time remains to be seen.

5.3 Characterization

One of the benefits of including seismic and acoustic methods in the toolkit for studying mass movements is that they contain temporal information about the event. However, characterizing event type and size from seismo-acoustic signals alone is challenging. The problem is not as simple as defining a single value, such as magnitude for earthquakes; volumes can change significantly over time as an event entrains and deposits material along its path. The dominant flow regime can also evolve, for example from debris avalanche to a debris flow as ice melts and/or water is incorporated. Furthermore, the change in potential energy, which relates to the seismic energy released, is dependent on the fall height. The fall height, in turn, depends largely on the path, the initial mass, and the rheology of the flow—factors that can vary widely from event to event.

5.3.1 Trajectory and mass from long period inversions

Currently, deterministic methods to estimate event size are limited to large, rapid landslides and are based on single-force histories obtained from inverting long-period (tens to hundreds of seconds) waves (Section 3.1). The premise is that, according to Equation 2, the force felt by the Earth is equal but opposite in direction to the landslide mass times its acceleration. Therefore, if the trajectory or mass are known or can be estimated (assuming the mass is constant), then the other variable can be estimated using the force history. Since there are two unknowns, researchers usually try different masses until the trajectory is approximately correct based on what they know about the landslide in order to obtain estimates of both (e.g., Ekström and Stark, 2013; Chao et al., 2016). This allows for the estimation of the velocity, momentum, and location

of the center of mass at each moment in time, though it is important to note that these values for the center of the mass may be considerably different from the flow front. This approach also requires some non-seismic information about the landslide, which may not always be available, so Ekström and Stark (2013) proposed an empirically derived simple scaling relation that relates peak force ($F_{\max}$) and landslide mass (m):

$$m = 0.54F_{\max}.$$

Equation 3

When plugged into Equation 2, Equation 3 essentially states that the peak acceleration is centered around $\sim 2 \text{ m/s}^2$. In reality the peak acceleration varies widely and is highly dependent on the specifics of the event, as is apparent given the significant scatter of the data used to derive the relation. For this reason, and because the uncertainties of the absolute force amplitudes can be significant (Moretti et al., 2015), Equation 3 realistically provides an order of magnitude estimate of mass. Furthermore, obtaining the trajectory requires double integration; therefore, any noise in the force history can significantly affect the trajectory. The trajectory method may not always provide useful results, for example when events fail in multiple time-delayed subevents or where the total mass moving at a given point in time changes significantly due to entrainment and deposition.

For more complex events for which a trajectory cannot be obtained by methods described above, the timing of peak forces from single-force time histories can be interpreted and tied to locations of corresponding features along the path, such as sharp curves or large barriers, to get an estimate of speed. This approach has been shown to give velocities comparable to field-based techniques such as superelevation (Allstadt et al., 2013a; Coe et al., 2016). However, it is important to note that such field-based analytical models can also have large errors (Iverson et al., 2016).

5.3.2 Propagation Velocity

For the majority of events that do not generate observable long-period surface waves and thus cannot be characterized using the approach of the previous section, there are other means of estimating propagation velocity. Array techniques can be used to track moving sources and estimate their velocities (Section 5.2.2 and 5.2.4). Simpler still is to divide the runout length by the duration of the seismic or acoustic signal to estimate the velocity under the assumption that

the signal represents one distinct failure pulse that generated observable seismic waves until it reached its final runout distance. This calculation cannot be done until the event has been identified and its length estimated, but it may provide a reasonable rough estimate for some events. However, it can be problematic if stations are far from the source because the perceived duration at a given seismic station is dependent on the local noise level and distance from the source (Dammeier et al., 2011). Duration is also sensitive to the method used to estimate start and end times (e.g. Dammeier et al., 2011; Levy et al., 2015). Furthermore, many signals have elongated durations, not because they ran out farther or were slower but because they reflect multiple overlapping sequential events (e.g., Norris, 1994).

Constraining landslide speed in the submarine domain is important for hazard analysis, as submarine volcanoes are prevalent, and landslides are a potential tsunami source. A method for estimating speed for submarine events was applied to landslides on West Mata volcano. Acoustic waves traveling on different paths from the source through the water column generated an interference pattern at hydrophones moored within the ocean low-velocity zone, the Lloyd's Mirror effect (Carey 2009). The frequency content of the interference pattern depends on source-receiver geometry, so any changes in the spectral content require that either the source or receiver be moving. Caplan-Auerbach et al. (2014) used this interference pattern to both identify active landsliding on West Mata and to estimate propagation velocities of 10-25 m/s. Further analysis by Drobiarz (2017), however, using a modified source location and data from multiple hydrophones suggests lower speeds, of 4-10 m/s.

5.3.3 Empirical methods

Numerous authors have tried to extract relationships between known event characteristics and seismic parameters. Success is generally limited to cases where most of the variables that affect the signal (Section 3) are held constant. For example, Norris, (1994) observed a linear relation between volume and amplitude from data on known rock fall and avalanche events but found that the relation was only valid for simple block failures (single collapse) of similar material over similar slope profiles over similar terrain types recorded on the same seismic station. In fact, most of the events Norris (1994) analyzed could not be easily characterized, since they were composed of multiple overlapping collapses and had much more complex waveforms. Similarly, Suriñach et al. (2000) found that even for two similarly sized snow avalanches down the same

path, if the flow type (dry vs. dense) was different or the slope surface was fresh rather than old snow, the seismic energy dissipation rates were different. This is not to say amplitude holds no information; as Mills (1991) points out, amplitude and/or duration can still be used to differentiate events with volumes that are orders of magnitude apart.

When so many variables affect the signal, a single simple relation between a seismically or acoustically measurable parameter and the physical characteristics of the mass movement such as mass or volume is not necessarily obtainable. In spite of this, there are a number of studies where authors have successfully developed empirical relationships between seismic parameter(s) and event parameters(s) (e.g., Helmstetter and Garambois, 2010; Dammeier et al., 2011; Chen et al., 2013; Hibert et al., 2014b). These relations are successful because they reduce the number of unknowns by using datasets derived from a single setting, one event type, and/or a specific seismic network. However, this generally also makes the methods site- or event-specific. Applying the same methods elsewhere would require tuning the model to a new location where datasets needed for tuning may not be available. Data fits for more generalized relations often show scatter of more than an order of magnitude (e.g., Ekstrom and Stark, 2013; Manconi et al., 2016). Despite the lack of transportable empirical relationships that can be used on any network for any event type and size, the fact that researchers have had success in estimating event characteristics under some conditions indicates that such a goal may in the end be possible, if effective means for controlling site-to-site variations can be developed.

5.3.4 Machine learning

Machine learning, or artificial intelligence (AI), has recently emerged as a method for classifying seismic event types including surface mass movements. It is advantageous in part because there is no need for a physical or geometric model linking the event types, and AI can perform well in noisy environments with temporally evolving data. This technique has been especially popular at volcanoes, where automated classification of diverse event types can play a role in volcano hazard assessment. Similar to empirical methods, machine learning requires training datasets. Most techniques start by taking a set of events with known classifications and dividing them into at least two subsets—a training and a testing dataset. Next, waveforms are parameterized into many variables in the time-domain, frequency-domain, and polarity- or particle-motion domain. The model is trained to recognize patterns in the parameters that are presented using the training

dataset. The resulting model is then assessed against the testing dataset that was not used to train the model. Depending on the results of this assessment, the model parametrization may be updated or tuned so a third evaluation or validation dataset that was not used for training or tuning may be used to evaluate the final model (Luongo et al., 2004). In general, the choice of parameters strongly changes the results of the classification, as does the size of the training set. Accumulating the required datasets for surface mass movements may be problematic because such events generate a wide range of signals, some event types are rare, seismic networks may not have been deployed long enough to capture sufficient data, and event type designation is typically dependent on expert opinion rather than independent verification. For this reason, the technique has only been used for common mass movement signals, primarily for rock falls (e.g., Ohrnberger, 2001; Langer et al., 2003; Esposito, 2006; Hibert et al., 2017b; Maggi et al., 2017).

Several different AI approaches have been used in seismic waveform classification. Artificial neural networks (ANNs) attempt to emulate the way that neurons in the brain operate by weighting various inputs and “firing” if a threshold is exceeded, passing the information to the next unit (e.g., Falsaperla et al., 1996). ANNs have been successful in classifying rock falls at Soufrière Hills (Langer et al., 2003, 2006) and Stromboli (Esposito, 2006). ANNs tend to over-fit training datasets, leading to poor or variable performance on the testing dataset (Luongo et al., 2004; Masotti et al., 2006). The problem can be exacerbated when additional parameters that do not distinguish the different event types are included. Optimal configuration of the different neurons and layers can require extensive data exploration.

The random forest algorithm (RF; Breiman, 2001) addresses some of these shortcomings and has been used to classify rock falls at Piton de la Fournaise (Hibert et al., 2017b; Maggi et al., 2017) and the Super-Sauze landslide in France (Provost et al., 2017). RF uses randomized decision trees to vote on the best classification. The algorithm does not tend to overfit the data, even with a large number of parameters, and minimizes the influence of parameters that are not useful for classification. However, RF still requires an extensive testing and evaluation dataset.

Some other techniques that have been used to classify rock fall include Hidden Markov Models (Ohrnberger, 2001; Hammer et al., 2012), Fuzzy Logic (Hibert et al., 2013), and Support Vector Machines (Curilem et al., 2014). Other methods have also been applied for seismic classification

at volcanoes (e.g., Cortés et al., 2014; Unglert et al., 2016). However, a detailed overview of all existing machine learning methods is beyond the scope of this review.

6. Discussion

It is clear that significant progress is being made in understanding the seismo-acoustic signatures of mass movements at volcanoes; however, a number of outstanding challenges remain. In this section, we discuss some of these challenges and how they may be addressed. What is generally needed are: 1) more and better data, 2) better techniques for analyzing the data, and 3) better and more quantitative models to explain the signals. The following subsections focus on each of these three areas. We discuss current research as well as what could be done to further accelerate scientific progress, ultimately leading to improvements in volcano hazard quantification and mitigation.

6.1 Seeking better datasets

To first order, seismic and acoustic signals from many different types of surface mass movements share many of the same general characteristics; however, higher order details (e.g., amplitude, timing, duration, and frequency content) are controlled by a range of physical variables. The underlying variables are of interest to researchers and include velocity, mass, material type, water content, flow domain, and the sequence of events and subevents. The radiated wavefield may contain information that is otherwise difficult or impossible to observe directly. The controlling variables can vary widely between events and can also vary spatially and temporally within a single event, making it difficult to untangle the effects of one from another in an observed signal without additional information. Clearly, there is a need to improve our theoretical understanding and ability to quantify or isolate various contributions to the wavefield so that seismic and acoustic methods can be better used to help advance our understanding of mass movements. This will require us to collect better datasets recorded on networks designed explicitly for this purpose and design laboratory and field-scale experiments with this objective in mind.

Existing monitoring networks that are optimized for earthquake monitoring are often already useful for surface-event monitoring. Several groups have published catalogs tying

serendipitously recorded mass movements with independent information about the events (e.g., Charbonnier and Gertisser, 2008; Dammeier et al., 2011; Chen et al., 2013; Ekström and Stark, 2013; Allstadt et al., 2017; IRIS DMC, 2017). However, there are weaknesses in relying on only preexisting datasets to understand the connection between variables of interest and the radiated wavefield: 1) the events are characterized with differing levels of uncertainty and completeness, in some cases with no observations except for the seismo-acoustic signals themselves; and 2) they rely on existing monitoring network configurations. To improve basic theoretical understanding, it is important to perform experiments designed specifically to address scientific questions related to seismo-acoustic signals generated by mass movements. For example, researchers have started to study debris flows using seismic and/or acoustic waves at the heavily instrumented natural debris flow laboratories of Chalk Cliffs, Colorado, USA (e.g., Kean et al., 2016) and Illgraben catchment in Switzerland (e.g., Burtin et al., 2014; Walter et al., 2017; Kogelnig et al., 2014). Quantitative understanding of seismic efficiency and the effects of particle concentration on near-field lahar seismic signals has come out of instrumenting channels experiencing multiple lahars (e.g., Cole et al., 2009; Doyle et al., 2010). Similarly, seismic and acoustic experiments at full-scale experimental test sites for avalanches, Ryggfjonn, Norway and Vallée de la Sionne, Switzerland (e.g., Vilajosana et al., 2007; Biescas et al., 2003) have yielded insights about snow avalanche seismicity and acoustic wave generation that are relevant to other types of flows. In addition, some well-monitored volcanoes experience upticks in activity and have acted as de-facto natural laboratories for mass movement studies, such as Piton de la Fournaise (e.g., Hibert et al., 2014b) and Volcán de Colima (Zobin et al., 2009). The limitation of natural laboratories is that in many cases, these sites lack the independent means of verification and control (e.g., direct observations of what occurred, where, when, and how) needed to truly progress our collective theoretical understanding. Increased and more routine video monitoring of volcanic slopes would help to capture more mass movement events, which would help to calibrate seismo-acoustic detection systems; however, traditional video monitoring requires clear visibility and is thus limited during inclement weather (cloud or fog cover).

On the other end of the spectrum, there are controlled laboratory experiments focused on seismo-acoustic wavefields generated by mass movements (e.g., Farin et al., 2015) where there is a great deal of control on parameters such as material type, volume, and initial conditions. Insights can be made, but the trade-off is that the experiments may not always be very realistic (e.g., perfectly

spherical particles, flat surfaces, continuous flows), so scaling-up and applying such lessons directly to natural events is not always straightforward. In between the natural laboratory and the actual laboratory are large-scale semi-controlled experiments, such those at the USGS debris flow flume (Iverson et al., 2010) where the material type and volume can be systematically varied and factors such as velocity, flow depth, and basal stresses can be independently measured. Though only recently have experiments there focused on the seismic and acoustic wavefield generated by flows (e.g., Allstadt et al., 2016; Iverson et al., 2017).

Progress should continue to be made through natural and controlled laboratory studies, and could be accelerated through publicly available high-quality datasets accessible for use in aggregated studies. One possible location for sharing such data is the recently created Exotic Seismic Events Catalog (<http://ds.iris.edu/ds/products/esec/>, IRIS DMC, 2017), which was designed to store diverse data types related to non-earthquake events such as mass movements alongside seismic and acoustic data.

A community natural laboratory experiment could also serve to accelerate progress and provide high-quality datasets. Such an experiment should be designed specifically with scientific objectives related to seismo-acoustic studies of mass movements. For example, dense “large-N” seismo-acoustic networks could be deployed at some well-chosen representative volcanoes where mass movements are common. Such unusually dense deployments could provide valuable insights into the limits, requirements, and ideal designs of more limited deployments. In addition, they could be paired with, for example, video, photos, ground-based radar interferometry, repeat lidar scanning, or other means of providing independent constraints on the mass movement events and processes.

6.2 Generalizing processing techniques

As described throughout this review, many authors have proposed and implemented various detection, location, and characterization techniques with success. However, many examples in the literature are case studies implemented for just one event type or in just one locale where the number of unknowns can be reduced or considered constant (e.g., Helmstetter and Garambois, 2010; Chen et al., 2013; Hibert et al., 2014b). In addition, several studies had the benefit of a well-characterized prior event catalog, allowing for signal detection algorithms to be tuned to the

event characteristics expected based on previous observations. Each volcano and monitoring network is different, so transporting these methods to other locations and applying them to other event types requires that the sources of differences (e.g., different source processes, source to station geometries, topography, subsurface characteristics, propagation effects) be well-understood and quantified and that approaches are developed that can successfully account for these differences. Additionally, developing standardized practices for different techniques, such as through community, collaborative coding efforts (e.g., Dietze, 2018) can also help standardize methods and reduce barriers.

In order to truly understand the capabilities and limitations of each approach, the community needs to investigate and publish failures as well as successes. The community also needs to thoroughly investigate and openly communicate sources of uncertainties. One potential approach that could reduce barriers to the application of existing methods at new sites would be to perform sensitivity studies of promising methods to determine exactly what input parameters (attenuation, seismic velocities, material properties, window length) are most important to quantify in order to use that method on a new volcano. It then would be important to develop consistent methods for estimating or measuring these parameters at new sites or establish published ranges of typical values for different types of settings. The success of many of the characterization methods discussed in Section 5 depend on how well the velocity structure, attenuation, site effects, and path effects are known or can be approximated. Detailed velocity and attenuation models have not been published for many volcanoes and even if they are available, they can be cumbersome to use in the context of some of the characterization methods because they may require 3D waveform modeling. A simpler approach is to calibrate the methods against known active or passive source. For example, Jolly et al. (2014) and Walsh et al. (2016) calibrated the Amplitude Source Location (ASL, Section 5.2.4) method to locate and characterize a debris avalanche and lahar using active source signals generated by dropping large masses from a helicopter along a channel.

A driver of this research area is the desire for remote monitoring and warning systems. This requires the development and improvement of methods and algorithms and monitoring network design to go with it. For example, though drainage-specific lahar monitoring systems are already in place and have been effective (Table 2), there is a desire for more cost-effective methods that

allow for longer warning times and monitoring of multiple drainages from a single system. Joint small-aperture seismic and acoustic array systems (Sections 5.2.2-5.2.4) show promise for operational monitoring. A single array can detect changes in backazimuth, while multiple arrays can be used to infer source location; this aids with signal discrimination and may allow flow velocities to be estimated.

Another complicating factor to monitoring efforts related to processing techniques is that in volcanic settings, and most acutely during volcanic unrest, signals from surface mass movements may be concurrent with other seismic sources (explosions, very long period earthquakes, tremor, repeating earthquakes). Even in periods of quiescence, glaciers and fumaroles can generate signals that can be difficult to differentiate from mass movement signals. Wind noise represents a major challenge for the acoustic detection of mass movement if the instruments are not deployed in arrays. Wind noise can generate long-duration broadband waveforms that superficially resemble signals from some mass movement events. Joint seismic and acoustic monitoring are complementary and can aid significantly with the robust detection and discrimination of mass movement signals because the two wavefields contain different information; seismic waves reflect mainly momentum exchanges between the mass movement and the subsurface, whereas acoustic waves reflect interactions between the mass movement and the atmosphere. The character of the interactions at these two different interfaces is dependent on the source process. However, seismo-acoustic arrays have historically not commonly been deployed as part of monitoring networks; thus, only a limited number of data examples currently exist. This point further motivates the deployment of test arrays in sites where mass movements are known to frequently occur. Machine learning and spectral methods have also been shown to help differentiate event types in some settings (Sections 5). However, their effectiveness depends on the quality of the data available. Event detection methods often require training for a specific site and network configuration, requiring preexisting classified datasets that take time to acquire. In any case, the robustness of detection methods is likely to improve through deployments of larger numbers of well-placed sensors.

6.3 Modeling improvements

For some existing applications, natural or lab-generated datasets may not be sufficient. Impactful events are rare and are recorded on whatever network existed at the time. For the goal of

developing a new monitoring system targeted at rare but impactful events, it is not feasible to wait for the target event to occur to test the system and its configuration. The robust system must be developed before the target event occurs (e.g., Lockhart and Murray, 2004). Therefore, it would be helpful to be able to reliably simulate the expected seismo-acoustic signals from scenario events. Such scenario events would represent the suite of possibilities targeted for detection, tracking, and alarming. Such simulations could be used to design and refine the network configuration prior to any deployment. This would require joint numerical modeling of the expected surface mass movement events, their radiated wavefields, and the expected noise levels at the stations.

Several researchers have started to develop the tools needed for this approach, such as modeling mass movements and the corresponding wavefields for long-period seismic waves generated by large-scale accelerations (Favreau et al., 2010; Moretti et al., 2012, 2015). On the other end of the spectrum, a few models focus on the highest frequencies generated by individual impacts (e.g., Tsai et al., 2012; Farin et al., 2015; Lai et al., 2018). However, there has been little research on the part of the signals between these end members that are likely valuable to monitoring efforts, though a few studies have compared correlations between higher and intermediate period seismic energy and numerical landslide outputs such as frictional power, momentum, and total forces (Schneider et al., 2010; Levy et al., 2015; Hibert et al., 2017a). We are not aware of any studies that simulate the acoustic wavefield from mass movements, though rapid advances have been made in acoustic waveform modeling from arbitrary point sources (e.g., Matoza et al., 2009; Lacanna and Ripepe, 2013; Kim et al., 2015). To our knowledge, there are no theoretical models that directly relate parameters that can be simulated by depth-averaged numerical models (velocity, depth, basal tractions) to the higher frequency stochastic seismic or acoustic wavefield.

If there were stochastic theoretical models that related, for example, a time series of basal tractions to the expected high-frequency seismo-acoustic radiation, it might be possible to take an approach similar to one commonly used in the generation of broadband synthetic seismograms for earthquake scenarios (Frankel, 2009; Graves and Pitarka, 2010). At long periods, the mass movement process can be modeled deterministically because path topography can be measured and the stresses a flow imparts on the subsurface can be modeled (e.g., Moretti et al., 2012; Moretti et al., 2015). However, as smaller scales (and thus higher frequencies) are

approached and modeling each individual impact or interaction becomes unfeasible, stochastic models could be used in place of deterministic ones. However, such an ideal depends on the development of seismo-acoustic source models, representing basic research on the source physics of mass movements.

7. Conclusion

We have outlined the scientific and societal benefits of advances in the study of mass movements using seismic and acoustic methods. Numerous researchers have found success in detecting, characterizing, and studying mass movement events using seismic and acoustic methods. A variety of methodologies and processing techniques have been applied successfully or could be applied to the detection, tracking, and characterization of mass movements at volcanoes, where mass movement hazard can be acute.

However, much work remains for this research area to reach its full potential, and basic questions still abound. To date, successes have been limited mainly to site-specific settings where sufficient preexisting data and network coverage are available. Most methodologies have not yet reached the state where they can be used for operational monitoring. Additionally, there are scientific gaps in our collective theoretical understanding of and ability to model the seismo-acoustic wavefield and how it relates to event parameters, particularly at smaller spatial scales and higher frequencies.

Nonetheless, seismo-acoustic methods show strong promise for the development of detection and early warning systems. While many of the topics discussed in this review are applicable also in non-volcanic settings, volcanoes require special attention. Volcanoes are a useful laboratory to make scientific advances because they are reliable producers of mass movements, where one is almost guaranteed to observe mass movements if one installs sensors and waits. More importantly, mass movements, and in particular lahars, are a significant hazard to populations living near and downstream from volcanoes. The societal benefit of operational seismo-acoustic monitoring is clear and attainable in areas of concentrated hazard.

In summary, the development and implementation of accurate and robust seismo-acoustic monitoring of surficial mass movements will require coordination across disciplines, but would

advance scientific understanding of the kinematics, dynamics, and spatio-temporal variability of these events. Seismo-acoustic monitoring of surficial mass movements shows promise for providing more accurate hazard assessments and early warning systems.

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Tables

Table 1 Summary of Mass Movement Event Types

Type (Figures)	Description	Typical velocities	Common triggers
Debris, rock, & ice avalanches (Figures 1A, 2A-3, 3A-3)	Highly mobile, and extremely rapid flows of fragmented rock, ice, and/or debris ⁽¹⁾	~20 – 90 m/s ⁽²⁾	• Destabilizing eruptive processes • Ground deformation • Snow or ice melt • Precipitation • Subsurface hydrological changes • Earthquakes • Long-term oversteepening • Overloading • Strength degradation due to hydrothermal alteration⁽³⁾
Edifice collapses (Figure 1B)	The largest mass movements that occur at volcanoes, subdivided into sector collapses, which involve the summit, and flank collapses, which do not ^(Error! Reference source not found.)	See debris avalanches	• Same as debris avalanches • Sector collapses usually occur only during periods of volcanic unrest⁽⁴⁾
Rock fall, debris fall (Figure 1D, 2A-4, 2A-5, 2C, 3A-2)	Masses of rock/debris of any size that move mainly by free-fall, bouncing, and rolling rather than by shear displacement ⁽⁵⁾	Can be very large due to free-fall component	• Same as Debris avalanches • Growth and inflation of active domes⁽⁶⁾ • Freeze/thaw
Lahars (Figure 1F, 2A-1, 2A-2, 3A-1)	“A rapidly-flowing, gravity-driven mixture of rock, debris and water (other than normal streamflow) from a volcano.” ⁽⁸⁾ Includes debris flows and hyperconcentrated flows ⁽⁹⁾	~5-40 m/s ⁽¹⁰⁾	• Mobilized from landslides that are saturated and/or entrain water⁽¹¹⁾ • Heavy precipitation on fresh ash and tephra⁽¹²⁾ • Outburst floods • Rapid melting of ice and snow from eruptive processes, such as ash or pyroclastic density currents mixing with and melting snow and ice⁽¹¹⁾
Outburst floods (Figure 1C, 2A-2)	Sudden releases of large amounts of water	Similar to debris flows	• Sudden releases of water from subglacial or periglacial lakes⁽¹³⁾ • Rapid melting of snow and ice due to eruptive processes⁽¹⁴⁾ • Breaches of man-made or natural dams – especially water-filled craters or calderas⁽¹⁵⁾ • Explosions of groundwater and/or from hydrothermal systems sometimes associated with volcanic activity⁽¹⁶⁾
Pyroclastic density currents (PDCs) (Figure 1E)	Rapid flows of hot material fluidized by escaping and entrained gasses ⁽¹⁷⁾	~8-170 m/s ⁽¹⁸⁾	• Collapse of eruptive ash plumes • Collapse of lava domes • Sector collapses • Retrogressive collapse of previous PDC deposits⁽¹⁷⁾

1. Hungr et al., 2014
2. Scott et al., 2001; Schneider et al., 2010; Favreau et al., 2010; Allstadt, 2013a
3. McGuire, 1996
4. Scott et al., 2001
5. Varnes, 1978
6. Calder et al., 2002
7. Havens et al., 2014
8. Smith and Fritz, 1989
9. Vallance, 2005
10. Johnson and Palma, 2015; Pierson, 1985; Scott et al., 2001
11. Iverson et al., 1997
12. Marciál et al, 1994, Tuñigol 2002
13. Björnsson, 2003
14. e.g., Waythomas et al., 2013
15. Waythomas et al., 1996, Regan et al., 2008; Massey et al., 2010
16. Newhall et al, 2001

17. Branney and Kokelaar, 2002
18. Charbonnier and Gertisser, 2008; Loughlin et al., 2002; Lube et al., 2007; Moore and Sisson, 1981

Table 2 Summary of Acoustic Flow Monitor Usage worldwide

Volcano	Location	Period of use	Description of performance	Reference(s)
Nevado del Huila	Colombia	2003-present	19 Feb 2007 (several million m ³) lahar not detected due to instrument power outage. 18 Apr 2007 (~50 million m ³) and 20 Nov 2008 (~300 million m ³) detected; Colombian Geological Survey successfully evacuated populations based on eruption signals and precursors prior to lahar passage.	Cardona et al., 2015; Pulgarin et al., 2015; Worni et al., 2012
Nevado del Ruiz	Colombia	1995-2016	No significant lahars occurred, AFMs recently replaced with standard seismometers	Cristian Lopez, SGC, personal information
Cotopaxi	Ecuador	2002-present	AFM System designed to detect large lahars (10's to 100's of millions of m ³), Small lahars have been observed but no large lahars have occurred since installation	Andrade et al., 2006a
El Reventador	Ecuador	2003-2004	Successful detections of frequent small lahars (<0.3 million m ³); flow rates estimated for calibrated sites	Almeida, 2016; Andrade et al., 2006a, 2006b; Andrade and Almeida, 2016
Tungurahua	Ecuador	1999-present	Successful detections of frequent small lahars (<0.3 million m ³); used to close roadways; flow rate estimates for calibrated individual stations	Almeida, 2016; Andrade et al., 2006a, 2006b; Andrade and Almeida, 2016
Mt Pelée	Martinique	1999-present	Successful detection of small lahars as part of an alarm system on the Revier du Precheur.	Rapport d'activite 2015 de L'Observatoire Volcanologique et Sismologique de Martinique
Merapi	Indonesia	1995-present	Successful detection of frequent lahars. Measured velocities between sites and estimated volumes for calibrated individual stations	Lavigne et al., 2000
Ruapehu	New Zealand	2002-present	Successful detection of 2007 lahar from tephra dam collapse (1.3-3.8 million m ³), one non-lahar trigger per month; Snow-rich eruption lahars in 2007 were detected but did not set off alarm due to apparent low energy transfer to bed	Keys and Green, 2008; Leonard et al., 2008
Mount Pinatubo	Philippines	1991-1992 and later	Lahar warning system including watch stations detected lahars after eruption; relative flow volume estimates	Marcial et al., 1996; Tungol and Regalado, 1996
Redoubt	United States (AK)	1990	Successful lahar detection in 1990; velocity estimates based on peak AFM values between stations	Brantley, 1990; Dorava and Meyer, 1994
Mount Rainier	United States (WA)	1998-present	Automated warning system operating on two drainages, no lahars large enough to test system	Lockhart and Murray, 2004

			have occurred	
Mount St. Helens	United States (WA)	1991-present	Successful detection of small debris flows.	Kurt Spicer, USGS, personal information

Figures

Figure 1 Examples of mass movements that generate seismic and acoustic signals A) Deposits of the 20 October 1997 debris avalanche, Mount Adams, WA, USA (photo: Richard Iverson, USGS, 1997) B) View of the crater left behind from the edifice collapse that occurred 18 May 1980 at Mount St. Helens, WA, USA, the deposits of the debris avalanche are shown in the foreground (photo: Tom Casadevall, USGS, 1980) C) August 1932 outburst flood from a glacier-dammed lake in the Copper River, AK, USA, shown sweeping away a Northwestern Railway bridge (photo: Alaska Dept. of Highways) D) Source area (cliff) and deposits from the

June 2011 sequence of rock falls onto Nisqually glacier at Mount Rainier, WA, USA (photo: Kate Allstadt, 2011) E) 7 August 1980 pyroclastic flow at Mount St. Helens, WA, USA (photo: Skamania County, WA, 1980) F) View east down Drift River Valley that was inundated by lahars caused by the 2009 eruption of Mount Redoubt, AK, USA (photo: R.G. McGimsey, AVO/USGS, 2009) G) Deposits from several snow avalanches on the Plains of Abraham at Mount St. Helens, WA, USA (photo: C. Dan Miller, USGS, 2012), H) Debris from a submarine mass movement composed of large blocks and coarse tephra on the flank of NW Rota-1 volcano (Jason frame grab from 14 April 2009 dive J2-408, from Schnur et al., 2017, used with permission).

Figure 2 Comparison of different types of seismogenic surface movements. A) Spectrogram and seismogram of 1) 24 March 2009 eruption of Mount Redoubt followed by lahar, starting at ~1900 sec (Lahar 4 from Waythomas et al., 2013), 2) Mud Creek outburst flood and debris flow at Mount Shasta, CA, in Sept. 2014 (de la Fuente et al., 2016), 3) Red glacier rock and ice avalanche at Iliamna volcano, May 2016 (Allstadt et al., 2017), 4) Rock fall off lava dome during 2006 Mount St. Helens eruption, May 2006 (Moran et al., 2008), and 5) Rock fall into lava lake at Kilauea, January 2016 (Orr et al., 2013). B) Power spectral density of each of the recordings at left with parts of the spectrum where the signal to noise ratio (S/N) compared to pre-event noise is greater than 1.5. The exception is the Mount Redoubt lahar, which shows a broadband station, RDW, which is closer to the source area, instead of DFR shown on A, which is close to the lahar runout area. C) Power spectral density of the 2006 Mount St. Helens rock fall on stations of increasing distance from the

source showing only parts of the spectrum where $S/N > 1.5$ relative to pre-event noise. Note that all records shown are corrected only for sensitivity to physical units of ground velocity, but corrections for differing frequency responses of different instruments have not been done.

Figure 3 Comparison of infrasound signals from different types of surface movements. A) Spectrogram and time series of 1) lahar during the 3/3/2015 eruption of Volcán Villarica, Chile (Johnson and Palma, 2015), 2) 2014 rock fall at Santiaguito, Guatemala (Johnson and Ronan, 2015), 3) rock and ice avalanche Red Glacier, Iliamna volcano, Alaska, May 2016, recorded on Transportable Array (TA) station O20K (Allstadt et al., 2017), 4) January 2012 snow avalanche in central Idaho (Havens et al., 2014). B) Power spectral density of each of the recordings at left, plus the 2006 rock fall at Mount St. Helens (Moran et al., 2008). C) Infrasound record of the 2006 Mount St. Helens rock fall showing the broadband record (top), high-pass filtered at 1 Hz (middle), and low-pass filtered at 0.1 Hz (bottom), modified from Moran et al. (2008). Note that waveforms in all panels were corrected for sensitivity to Pa, but corrections of differing frequency responses of different instruments have not been done.

Figure 4 Seismogram of a pyroclastic density current (PDC) at Volcán de Colima, Mexico in 2005 that was generated by the collapse of an eruptive column. Arrows with numbers indicate the following interpretations of the impulses by Zobin et al. (2009): 1) fragmented magma movement in the conduit, 2) explosion in the conduit, 3) rockfall from the collapsing eruptive column, 4) movement of rocks along the slope, 5) the pyroclastic flow. Figure from Zobin et al. (2009)

Figure 5 Time series (top) and spectrogram (bottom) of an ice avalanche at Iliamna volcano, Alaska on 10 September 2004. The failure plane was at the ice-rock interface and the event entrained rock. The seismic data show precursory repeating discrete stick-slip events that become more and more frequent until they merge into a continuous signal that culminates in the highest amplitude signal generated by the actual avalanche. This event involved a total volume of 4-6 million m^3 , dropped 1750 m and ran out 4950 m horizontally over the course of 110 seconds. Figure from Caplan-Auerbach and Huggel (2007).

Figure 6 Schematic illustration of the approximate volumes and relative recurrence intervals of various types of volcanic collapse events, modified from McGuire (1996).

Figure 7 Simultaneous measurements of basal normal stress during a debris flow experiment at the USGS debris flow flume on (A) a 500 cm² force plate and (B) a 1 cm² force plate, modified from Figure 11 in Iverson (1997).

Figure 8 Fraction of total radiated energy of different seismic wave types for a force source at the surface of a homogeneous halfspace as a function of orientation of the force vector. A force perpendicular to the surface radiates mostly Rayleigh waves whereas one oriented parallel radiates mostly SH-polarized waves. Poisson's ratio of the halfspace is equal to 0.25.

Figure 9 Most distant high-frequency detection (examined in 1-5 Hz band) for different event types and range of volumes showing A) rock, ice, and debris avalanches, B) rock and debris falls and C) debris flows and outburst floods. Horizontal error bars show volume uncertainties, vertical error bars show the distance to the next closest station from the one on which the farthest detection was made. Shaded regions show areas of conditional detectability for each event type, below this area is approximately where events are likely detectable, above this region indicates the distances at which each event is likely not detectable for a given volume. This figure was generated using the most distant detection in the 1-5 Hz frequency band compiled for a catalog of events by Allstadt et al. (2017). The areas of conditional detectability were defined by finding the slope of the line of best fit and shifting that line up and down so that the upper and lower lines encompassed all events in each category.

Figure 10 Acoustic flow monitor setup and data from 1992 lahar monitoring network at Mount Pinatubo showing (top) map of system and (bottom) data from the low frequency (10-100 Hz) band of the acoustic flow sensor (square on map) compared to the corresponding hydrograph at Mactan (MC on map) for August 29, 1992 lahars (from Tungol and Regalado, 1996)

Figure 11 a) Waveform (0.5-50 Hz) and spectrogram of acoustic signal generated by a 2015 lahar at Volcán Villarrica (Chile) recorded on array VIH (see map c), black line indicates centroid frequency. b) Location of the inferred lahar flow front along the flow profile (shown on map c). The inferred source location at each overlapping 5 sec time window is indicated with a dot. In panel b), the label “notch” indicates the distance at which the topographic notch is reached, which is inferred to have dominated the signal after the flow started passing through it, thus explaining why the location stabilizes after the lahar reaches ~11 km. Wf indicates the location of waterfalls along the path. Background colors indicate the signal to noise ratio of the array beam stack, dark red is 1 (best beam signal), dark blue corresponds to 0.45 (worse beam signal). Gray lines indicate location error due to timing uncertainty of 0.01 sec. Modified from Johnson and Palma (2015), figures 2 and 5.

Figure 12 Maps of normalized residuals for two different time windows computed by Kumagai et al., (2009) using the Amplitude Source Location method to track a small lahar down the northeast flank of Cotopaxi volcano with the star indicating the best location. The time windows are 5 seconds long. The time window of the lower map is 735 seconds after the upper map.

Figure 13 Comparison of surface wave magnitude to body wave magnitude for the Hope (H), Vajont (V), and Mantaro (M) landslides compared to nuclear explosions (solid circles) and earthquakes (open circles), from Weichert et al. (1994).

Highlights for “Seismic and Acoustic Signatures of Surficial Mass Movements at Volcanoes” by Allstadt et al.

- Surficial mass movements are common in volcanic areas and generate signals that are picked up by seismic and acoustic monitoring arrays
- Our understanding of the relation of these signals to characteristics of the mass movement is limited but improving
- We review the literature on the study of mass movements at volcanoes using seismic and acoustic monitoring
- We discuss future research directions and steps toward operational monitoring