VOLCANO INSTABILITY AND LATERAL COLLAPSE

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ABSTRACT

Active volcanoes are dynamically evolving structures, the life-cycles of which are punctuated by episodes of flank instability and lateral failure. Such behaviour is now recognised as ubiquitous and lateral collapses are estimated to have occurred at least four times a century over the past 500 years. In the Andes, three quarters of the large volcanic edifices have experienced collapse, while over a hundred debris avalanche deposits around the Japanese Quaternary volcanoes testify to repeated episodes of instability and collapse over the last million years.

A range of both internal (endogenetic) and external (exogenetic) factors contribute to the destabilisation of a volcanic edifice. The former include gravitational spreading, the development of steep slopes and a mechanically unsound structure, surface loading by erupted materials, and oversteepening or incremental lateral displacement due to magma intrusion. External factors include basement uplift or subsidence, fault activity, climatic effects, and changing sea levels. The triggering of lateral collapse at a destabilised volcano typically occurs in response to a short-lived dynamic event such as the intrusion of a body of fresh magma (magmagenic), or to volcanic or tectonic earthquakes (seismogenic). In the former case, collapse may be due to either mechanical push or to pore-pressure changes in ground water, while in the latter increased ground accelerations are the probable cause.

Volcanic edifices may become unstable and experience structural failure at any scale. This ranges from relatively small rock falls involving a few hundred to a few thousand cubic kilometres of debris, to the giant ocean-island megaslides, such as those identified around the Hawaiian and Canary archipelagoes, that may incorporate up to 5000 cubic kilometres of material. While the lowest volume collapses probably occur at one active volcano or another every few weeks, the largest events have frequencies of tens to hundreds of thousands of years. Such major lateral collapses are primarily confined to large, long-lived polygenetic volcanoes, located on either continental (eg Etna, Mount Rainier, Colima) or oceanic (eg Mauna Loa, Kilauea, Piton de la Fournaise) crust.

Edifice instability may develop rapidly in response to a discrete event, such as the emplacement of the Mount St. Helens cryptodome in 1980. Alternatively, the volcano may become destabilised incrementally over many thousands or tens of thousands of years due, for example, to rift-related lateral displacement associated with persistent dyke intrusion.

The hazard implications of volcano lateral collapse are significant at both the local and regional scales. In addition to the production of debris avalanches that may exceed 100km in length, the scale of tsunami generated by the largest collapses of ocean island volcanoes are sufficient to cause ocean-wide destruction. In the latter context, candidate volcanoes for future major collapse and tsunami generation include the Cumbre Vieja (La Palma, Canary Islands) and Fogo (Cape Verde Islands).

1 • INTRODUCTION

The spectacular landslide that triggered the climactic eruption of Mount St. Helens during May 1980 (Lipman & Mullineaux, 1981) focused attention on the instability of volcanic edifices, and their tendency to experience lateral collapse. Such behaviour is now recognised as being commonplace, and is recorded both in the geological record and also at many recently active volcanoes (Siebert 1984; Ui 1983). Inokuchi (1988) reports, for example, over a hundred debris avalanches produced by lateral collapse around Quaternary volcanoes in Japan, while Francis (1994) notes that three quarters of large volcanoes in the Andes have experienced lateral failure. The potential hazard posed by such behaviour is significant, with at least four such collapses occurring per century for the last five hundred years (Siebert 1992). Lateral collapses are not confined solely to subaerial volcanoes on Earth. Improved submarine imaging methods have revealed giant debris avalanches and slumps around the flanks of ocean island volcanoes such as those of Hawaii (Fornari & Campbell 1987; Moore et al. 1994; Garcia 1996), Reunion Island (Lenat et al. 1989; Labazuy 1996), Stromboli (Kokelaar & Romagnoli, 1995), Augustine Island (Beget & Kienle 1992), and the Canary Islands (Holcomb & Searle, 1991; Carracedo 1996; Weaver et al. 1994). Similarly, imagery gathered using the Viking and Magellan spacecraft have revealed evidence for lateral collapses of volcanoes on both Mars (Crumpler et al. 1996) and Venus (Bulmer & Guest 1996).

2 • THE DEVELOPMENT OF STRUCTURAL INSTABILITY IN VOLCANIC EDIFICES

Volcano instability may be defined as ‘the condition within which a volcanic edifice has been destabilised to a degree sufficient to increase the likelihood of the structural failure of all or part of the edifice' (McGuire 1996, p1). Growing volcanoes may become unstable and experience failure at a range of scales, from relatively minor rock falls consisting of a few hundred to a few thousand cubic metres of debris to the giant megaslides around ocean-island volcanoes that may involve volumes in excess of $1000km^3$. The lowest volume collapses probably occur on some volcano or another with frequencies of the order of weeks or months, while the largest events have return periods of tens to hundreds of thousands of years.

A whole spectrum of factors may act, independently or in concert, to increase the instability of a volcanic edifice, and some volcanoes demonstrate a much greater potential for instability development and lateral failure than others. Large-scale structural instability is invariably confined to major polygenetic volcanoes, located either on continental (e.g. Etna in Sicily, Rainier in the Cascade Range, and Colima in Mexico) or oceanic (e.g. Mauna Loa and Kilauea on Hawaii, and Piton de la Fournaise on Reunion Island) crust.

Large volcanic edifices built on continental crust are particularly prone to lateral collapse, although not on a scale comparable with the largest failures of their oceanic counterparts. Continental edifices are characteristically stratovolcanoes with internally-weak structures resulting from a high pyroclastic content and hydrothermal alteration. The potential for instability development and lateral failure is compounded by steep slopes and often high precipitation rates that may contribute to changes in edifice pore-water pressures.

Despite the low slope angles and homogeneous structure, instability and lateral collapse are frequently induced in large, basaltic shield volcanoes such as those of Hawaii. The major contributing factor towards instability development is rifting associated with persistent dyke emplacement, while failure may be triggered by earthquakes, changes in pore-water pressure, or environment factors such as large, rapid variations in sea level. Instability may also develop due to the spreading of such basaltic shields on weak horizons of oceanic sediment (Nakamura 1980) or in response to seaward-creeping bodies of olivine cumulate (Clague & Denlinger 1994).

The tendency towards the general development of instability at large, long-lived, polygenetic volcanoes is enhanced by continuous changes in morphology. These may be either endogenetic – resulting from the intrusion of fresh magma – or exogenetic – due to the addition of material at the surface, and characteristically lead to oversteepening or overloading at the surface. Subsidence or uplift of the sub-volcanic basement may also contribute towards instability, as may the growth of a volcanic edifice on a sloping or weak (e.g. clay-rich) substrate. The latter is particularly effective in leading to the detachment of the down-slope sectors of so-located volcanoes (e.g. Etna), and increasing the potential for catastrophic lateral failure.

Destabilisation of a volcanic edifice may take thousands or tens of thousands of years, or it may be achieved in only a few months. Rapid-onset destabilisation was well demonstrated by emplacement of the new cryptodome only a few months before the failure of the north flank of Mount St. Helens in May 1980 (Christiansen & Peterson, 1981). Contrastingly, slow, progressive destabilisation is characteristically incremental, and often results from the cumulative effects of many small events. These may take the form of successive eruptions leading to the overloading of a weak slope by the addition of numerous lava flows, or – as at Kilauea (Swanson et al. 1976) and Etna (McGuire et al. 1990; McGuire et al. 1997) – may result from progressive lateral displacements associated with persistent dyke emplacement along long-lived, intra-volcanic rift systems.

Provided that similar conditions are maintained, volcano lateral collapse related to progressive destabilisation may be cyclic. At Augustine volcano (Alaska) (Beget & Kienle 1992), collapse leading to debris avalanche formation has a return time of only 150 – 200 years. At Colima (Mexico), recurrent collapse occur every few thousand years (Komarowski et al. 1994), while the giant flank collapses of the Hawaiian volcanoes have frequencies of 25 – 100 ka (Lipman et al. 1988).

3 • TRIGGERING EDIFICE FAILURE

As for the development of edifice instability, structural failure may itself occur over a range of time scales, although it most often takes the form – as at Mount St. Helens - of a near-instantaneous, catastrophic event. In some situations, however, deformation following lateral failure may take place more slowly with little in the way of rapid mass movement. Etna is a prime example for this type of behaviour, where a large sector of the eastern flank has become detached from the remainder of the edifice and has been sliding seawards – probably for tens of thousands of years (e.g. Kieffer 1985; Borgia 1992; McGuire & Saunders 1993). Such behaviour does not require a discrete, instantaneous failure event to initiate movement, nor does any high velocity transportation of a large mass of material need to be involved.

Both phenomena do, however, occur during formation of debris. Once a block of volcanic terrain has become destabilised it becomes susceptible to failure in response to one or more of a number of 'internal' or 'external' triggers. The former are dominantly related to the extrusion or intrusion of fresh magma that have the potential to trigger failure through gravity loading, mechanical push, or temperature-related changes in pore pressure (Elsworth & Voight 1996; Day 1996). Other internal triggers include ground accelerations associated with strong volcanogenic earthquakes, and displacements associated with long-term edifice spreading. Many external triggers reflect the dynamic geological settings in which volcanoes are commonly located, and include basement fault movement and related tectonic seismicity. Environmental factors may also be important, with both precipitation (on a short time-scale) and changing sea levels (on a longer time-scale) having the potential to initiate structural failure.

The role of magma intrusion is typically paramount in triggering structural failure occurring during eruption, which may occur in response to a critical, destabilising change in edifice morphology, as in the Mount St. Helens case, or may result from other less obvious effects. In particular, Voight & Elsworth (1992) and Elsworth & Voight (1996) have drawn attention to the potential role of dykes in triggering structural failure as a consequence of their raising pore pressures by means of mechanical and/or thermal straining.

There is, in fact, increasing evidence that water plays an important role in the destabilisation and mechanical failure of volcanic edifices, and in determining the manner in which the failed mass behaves. Not only does hydrothermal alteration often play a major part in increasing susceptibility to failure (Siebert et al., 1987), but the common association of lateral collapse events with phreatic explosions (e.g. at Bandai-san, Japan, in 1888) supports a significant involvement by hydrothermal pore fluids in the failure process. Day (1996) proposes that this results from a reduction in rock strength in response to the generation of pore pressures which are a large fraction of, or even higher, than contemporary confining pressures. As mentioned in the previous section, this situation may arise due to the heating effects associated with the intrusion of fresh bodies of magma (Voight & Elsworth, 1992; Elsworth & Voight 1996). Additionally, Day proposes that similar conditions may result from intrusion degassing, the discharge of pressurised fluids from depth via clastic dykes, or by faulting-associated deformation and pore collapse.

Once structural failure has been initiated, water may also have a role to play in determining the behaviour of the collapsing mass, and in particular whether the event is aborted or proceeds to form a debris avalanche. The islands of La Palma and El Hierro (Carracedo 1996) in the Canary Islands, for example, provide excellent evidence for major, rift-related, flank failure, but also two examples of lateral collapse events which appear to have been aborted during the early stages of sliding. On the western side of the Cumbre Vieja ridge at La Palma, eruptive activity during 1949 was accompanied by the opening of a concave-downslope fracture system some 4km in length. The vertical displacement on the fracture amounts to only a few metres, and testifies to a seaward sliding event which was aborted immediately following initial failure.

On the neighbouring island of El Hierro, a similar event appears to have taken place, forming the San Andres fault system, although here over about 300m of vertical displacement occurred prior to cessation of deformation. The faults of this aborted collapse are well exposed and are characterised by dry fault breccias and ultra-cataclasites, rather than the extremely fluidised gouge muds and mud-rich breccias encountered in exhumed collapses elsewhere in the Canary islands (Day 1996). It is possible that this lack of pressurised fluids on the faults, and a consequent lack of significant slip weakening or of brecciation of the collapsing slump blocks by gouge dykes may have caused fault movement to cease without debris avalanche generation. This evidence from the Canaries permits the possibility of reinterpreting large flank slumps, such as those of Kilauea, as aborted collapse events which might have generated major debris avalanches if conditions had been more appropriate to the generation of high pore pressures, either in the fault zones or within the slump blocks as a whole, and thus to sustained slumping and slump-block disaggregation.

4 • POST-FAILURE MASS TRANSPORT AND EMPLACEMENT

Lateral failure of volcanic edifices inevitably involves the downslope, gravity-driven, mass-transfer of material from the source to an area of deposition. In cases of relatively minor rock-falls and slumps, the distances involved may amount to only a few tens or hundreds of metres, whereas at the other end of the scale, gigantic debris avalanches may be transported to distances of several hundred kilometres (Stoopes & Sheridan, 1992; Moore et al., 1994). Table 1 below summarizes the more important parameters of some of the longer, larger volume, volcanogenic debris avalanches.

TABLE 1.

Volcano	Deposit	Volume (km^3)	Runout (km)
Nevado di Colima		22-33	120
Socompa		17	35
Volcán de Colima		6-12	43
Shasta		26	50
Popocatapetl		28	33
Chimborazo	Riobamba	8.1	35
Mawenzi		7.1	60
Akagi	Nashikizawa	4	19
Galunggung		2.9	25
Mount St. Helens	1980	2.5	24
Fuji	Gotenba	1.8	24
Shiveluch	1964	1.5	12
Bandai-san	1888	1.5	11
Egmont	Pungarehu	0.35	31
Unzen	1792	0.34	6.5
Asakusa	Migisawa	00.4	6.5

Volumes and runout distances of selected subaerial volcanic debris avalanches. Data from Hayashi & Self (1992), Stoopes & Sheridan (1992), Wadge et al. 1995)

Depending upon a number of factors, including the nature of the failed material and the underlying terrain, and the precise failure mechanism, the transported mass may travel in a largely coherent manner or may become totally disrupted. Transport velocities may vary enormously from over 100m/sec where failure has been catastrophic to as little as 1-2cm/y where displacement of the detached mass involves creep-like behaviour. Structural failure may be confined solely to the volcano, or may involve the underlying basement, with consequences, in the latter case, for the composition of the deposits formed.

Inevitably, the downslope removal of material during the failure process leaves voids in the source area. Where large-scale, catastrophic failure has take place, these commonly take the form of near parallel-sided amphitheatre-like depressions which open downslope (Siebert, 1984) and which are typically surrounded, or partly bounded, by steep walls which may be in excess of a kilometre in height, although more complex forms also occur. Less spectacular events of an effectively instantaneous nature may leave small collapse scars, for example along the margins of actively growing domes or unstable caldera rims. Where failure-related transport is a slower, longer-term process, the area between the mobile, detached block and the remainder of the edifice is typically marked by incipient or open fractures or by active fault systems. Although the more impressive morphological features associated with catastrophic failure may remain extant and easily recognisable for a considerable period of time, those associated with less-rapidly operating mechanisms may be difficult to discern, and the recognition and nature of the displacement may require the use of ground deformation monitoring techniques.

Because catastrophic failure events are commonly associated with eruptive or intrusive activity, the deposits they produce often reveal a wide range of lithological and sedimentological characteristics (Ui, 1989). The generation of 'dry' avalanches (Ui, 1983), in the absence of accompanying volcanic activity, typically produces deposits made up of more or less disrupted volcanic material with or without a basement contribution. Where fresh magma is involved, either in triggering the collapse or due to post-failure unroofing and decompression of a shallow reservoir, juvenile material may be present in the form of chilled lava fragments which may show evidence of a plastic nature during transport. Where structural failure triggers a major decompressive eruption, as during the May 1980 event at Mount St. Helens, base-surge and pyroclastic flow formation closely follows the generation of the debris avalanche, and deposits of these magma-rich events may be mixed with or overlie the more lithic-dominated avalanche material.

A volcanic 'dry' avalanche deposit has been defined (p135) by Ui (1983) as a 'volcaniclastic deposit formed as a result of large-scale sector collapse of a volcanic cone associated with some form of volcanic activity'. This is, however, an unnecessarily restrictive definition as it implies that deposits formed by similar mass-removal events at volcanoes which are not associated with volcanic activity are somehow different. A more appropriate definition may be that proposed by McGuire (1995), which defines volcanic 'dry' avalanche deposits as ' having been formed by the large-scale collapse of a volcanic edifice, or part thereof, in the absence of significant amounts of water'.

In many cases, however, the avalanches formed during catastrophic edifice failure are far from 'dry'. Available water from saturated volcaniclastic sequences, from snow and ice fields, or from surface water bodies commonly becomes entrained into the avalanching mass causing a progressive transition from dry debris avalanche to debris flow or debris-laden flood (broadly termed lahars when generated in volcanic environments). Large volumes of surface water are not, however, essential to transform a debris avalanche into a debris flow. According to Fairchild (1987), the water source for the North Fork lahars at Mount St. Helens was finely comminuted ice incorporated within the avalanche deposit during collapse. Lahar formation here is attributed to liquefaction of the water saturated debris avalanche due to harmonic tremor associated with the post-collapse eruption.

5 • THE ROLE OF PERSISTENT RIFTING

Siebert (1984), in his global review of the occurrence of volcanogenic debris avalanches, drew attention to the fact that structural collapse was more common at edifices characterised by the existence of parallel dyke swarms. This link, which supports the role of persistent, dyke-induced rifting as one of the major causes of edifice destabilisation and failure, is clearly illustrated by the orientations of the long axes of flank collapse scars relative to adjacent dyke zones. At Stromboli (Tibaldi 1996) and La Palma (Canary Islands) (Carracedo 1996) for example, these are oriented normal to a single dominant dyke zone, whereas at Etna (McGuire & Pullen, 1989; McGuire et al., 1993), Piton de la Fournaise (Réunion Island) (Duffield et al., 1982), and El Hierro (Canary Islands) (Carracedo 1996), sector collapses bisect the angles formed by two intersecting zones of persistent dyking. Where repeated dyke emplacement along a preferential path has taken place over a long period of time (eg 10^4y or more), accumulated erupted and intruded products typically result in the growth of a pronounced topographic ridge overlying the rift zone. Due to oversteepening and loading effects, this structure tends to become less stable over time, and more susceptible to failure in response to single destabilizing events such as an earthquake (volcanogenic or otherwise), or dyking event

Due to the manner in which dyke orientations at shallow depths are controlled by a gravitational stress regime which is a reflection of the edifice morphology (McGuire & Pullen, 1989), the existence of large sector collapse structures, which occurred in response to dyke-induced rifting, may in turn control the disposition of post-collapse zones of persistent rifting, and therefore the locations and directions of subsequent collapses. This behaviour is illustrated by the orientation of post-lateral collapse dykes at both Etna (McGuire et al., 1990, 1991) and Stromboli (Tibaldi 1996).

In addition to essentially vertical dykes, subhorizontal sheet intrusions (sills) have also been suggested to have a destabilising role. In particular, Adushkin et al. (1995) and Delemen (1995) have proposed that the emplacement of such a magma body at the unstable Klyuchevskoi volcano in Kamchatka may be sufficient to trigger future edifice failure resulting in a debris avalanche with a volume of 4-8km^3.

6 • LATERAL EDIFICE GROWTH

Recent papers have highlighted a number of different ways in which volcanic edifices can become destabilised and experience failure during lateral edifice growth. Two contrasting mechanisms involve (i) relatively deep gravitational spreading along basal thrusts, due to their increasing mass, of volcanic structures such as those constituting the Hawaiian, and the Concepción and Maderas volcanoes in Nicaragua, and (ii) shallow gravitational sliding of sectors of volcanoes due to oversteepening, peripheral erosion, basement slope or tilting, or a combination of these and other factors. Both mechanisms lead to lateral edifice growth which may contribute towards greater edifice instability. The manner in which the terms gravitational spreading and gravitational sliding are often used synonymously when applied to volcanic edifices, illustrates that the differences between the two mechanisms have not been clearly defined. Although, as the terminology indicates, gravity has a major role to play in both types of behaviour, this role is not identical and does not lead to the formation of the same phenomena.

McGuire (1996) proposes that the term volcanic spreading should not be used in a genetic manner, and should be confined simply to describing the phenomenon of lateral edifice enlargement. Mechanisms by which spreading is accomplished can then be summarised as gravitational sliding, gravitational thrusting, or edifice collapse. Each mechanism may furthermore be described as being radial, where the entire circumferance of the edifice is involved, or sector, where spreading involves only part of the edifice. Using this classification, the form of spreading proposed for Kilauea would be described as sector thrusting, and that for Etna, as sector sliding. Other forms of spreading include radial thrusting at Mombacho (Nicaragua), radial collapse at Augustine Island (Alaska), and sector collapse at numerous volcanoes.

Although the load of a volcanic edifice may play a part in the initiation of gravitational sliding, it is the form of the underlying substrate which provides the appropriate conditions. While the load-driven spreading mechanism is initiated on a flat substrate, which thereafter becomes downwarped, gravity-controlled sliding requires the development of an asymmetrical sloping surface beneath the edifice, which may, as at Etna, result from differential uplift beneath the volcanic pile. As with the thrusting mechanism, the sliding of large sectors of volcanic edifices along a sloping basement typically takes place at centimetric annual rates, but displacements occur along a décollement (or series thereof) which slopes downwards. Further similarities with gravitational thrusting involve the requirement for a ductile sub-volcanic horizon, the development of tensional conditions in the upper levels of the edifice and the potential for generating peripheral compressional features. In gravitational sliding, the long axis of the mobile sector of the volcano is typically oriented parallel to the slope of the underlying substrate, and is separated from the remainder of the edifice by faults or fault zones characterised by significant strike-slip components.

While convincing cases for the operation of gravitational thrusting have been made for the Hawaiian volcanoes, and a number of other smaller the mechanism has also been proposed, less credibly, to attempt to explain the unstable and mobile nature of the eastern flank of the Etna volcano (Borgia et al., 1992; Borgia, 1994). Here, however, a combination of generally shallow and the common occurrence of creep-related surface, argue strongly for gravitational sliding of the eastern sector of the edifice over a clay-smeared substrate which is downfaulted seaward and continues to be uplifted at an average annual rate of 0.8 to 1.4 mm/y (Stewart et al., 1993; Firth et al. 1996).

7 • INSTABILITY AND FAILURE AT ISLAND AND COASTAL VOLCANOES

Many of the largest landslides resulting from instability and structural failure are located adjacent to the margins of island and coastal volcanoes. All the Hawaiian volcanoes, for example, are surrounded by submarine aprons of allochthonous volcanic material emplaced by sliding or slumping (Moore et al., 1989; 1994), which may grade into volcanic turbidites (Garcia 1996). Similar deposits have been recognised around many marine volcanoes, using techniques such as sea-beam bathymetry and high-resolution side-scan sonar imaging, including Piton des Neiges and Piton de la Fournaise (Réunion Island) (Rousset et al., 1987; Lenat et al., 1989; Labazuy 1996), Piton du Carbet (Martinique) (Semet & Boudon, 1994), the Marquesas volcanoes (Barsczus et al., 1992; Filmer et al., 1992), Tristan de Cunha (Holcomb & Searle, 1991), the Galapagos Islands (Chadwick et al., 1992), the Canary Islands (Holcomb & Searle, 1991; Weaver et al., 1994), Stromboli and Alicudi (Aeolian Islands) (Romagnoli & Tibaldi, 1994), and at Augustine Island (Alaska) (Begét & Kienle, 1992). The sizes of deposits are enormously variable although Holcomb & Searle (1991) report that many single landslides affecting oceanic volcanoes may have been sufficiently large as to involve the transport of up to 20% of the edifice volume. Some of the Hawaiian landslides have volumes greater than 5000km^3 and lengths in excess of 200km making them the largest such structures recorded on Earth (Moore et al., 1992).

The common occurrence of aprons of destabilised material around marine volcanoes is to be expected for a number of reasons. Most significantly, the seaward-facing flank of any volcano located at the land-sea interface is inevitably the least buttressed. This applies both to coastal volcanoes such as Etna, where the topography becomes increasingly elevated inland, and to island volcanoes such as Hawaii where younger centres (such as Kilauea) are buttressed on the landward side by older edifices (eg Mauna Loa). The morphological asymmetry resulting from this effect leads to the preferential release of accumulated intra-edifice stresses, due for example to surface-overloading or to repeated dyke-emplacement, in a seaward direction. This stress release may take the form of the slow displacement of large sectors of the edifice in the form of giant slumps, of co-seismic downfaulting, or of the episodic production of debris avalanches, or a combination of all three. The relatively unstable nature of the seaward-facing flanks of any volcano is further enforced by the dynamic nature of the land-sea contact. Not only does marine erosion provide a constant destabilising agent, but large changes in global sea levels of up to 130m, occurring over periods as short as 18,000 years, with catastrophic rises recently identified of 11.5m in $<160 \pm 50$y (Blanchon & Shaw, 1995), offer the means of modifying internal stress regimes and water pore pressures in favour of edifice destabilisation.

The Hawaiian Ridge, extending from near Midway Island to Hawaii, provides by far the most impressive evidence for volcano instability and collapse in the marine environment. Following a cooperative submarine survey by the United States Geological Survey and the UK Institute of Oceanographic Sciences, using the GLORIA side-scan sonar system, sixty- eight landslides with lengths in excess of 20km have been identified. In a comprehensive review of landslide generation along the Hawaiian Ridge, Moore et al. (1994) highlight a number of characteristic features which are likely to be generally applicable to the destabilisation and collapse of marine volcanoes. Although occurring throughout the lifetimes of the volcanoes, the largest landslides occurred when the centres were young and unstable, were close to their maximum size, and when seismic activity was at a high level. The authors differentiate slumps from debris avalanches and report the existence of intermediate forms. Slumping and avalanching can therefore be viewed as end-members of a continuous sequence of emplacement mechanisms which is probably applicable to large-scale mass-wasting processes at all volcanoes located in marine environments. The two mechanisms are not mutually exclusive, with debris avalanches often forming from the disaggregation of oversteepened or overpressured slumps, or from injection of pore fluids or gouge muds from the lower regions of slumps into the upper layers, which may then disintegrate.

Moore et al. (1994) draw attention to both the different characteristics and the alternative mechanisms responsible for the emplacement of the Hawaiian slumps and debris avalanches. While the former are typically both wide (sometimes over 100km) and thick (up to 10km), the latter are long (up to 230km) and relatively thin (0.5 - 2km). The authors explain that the slumps are deeply rooted in the edifice and may be bounded by rift zones on their landward side, and by the edifice-substrate interface at their base. Movement is typically slow and creep-like, although evidence for major co-seismic displacements are recorded on the active Hilina slump of Kilauea (Lipman et al., 1985). In contrast, the debris avalanche features reported by Moore et al. (1994) include well-defined amphitheatres in their source regions, hummocky terrain with megablocks up to 2km across, and evidence for uphill transport on the Hawaiian Arch submarine ridge, implying high emplacement velocities. From the range of features described from the Hawaiian Ridge, it becomes apparent that the term 'landslide', which has common usage in describing the products of seaward mass-movement at coastal and island volcanoes, is not wholly appropriate. Strictly-speaking this term is best confined to describing the rapidly emplaced debris avalanche deposits, rather than the relatively slow-moving slump blocks.

The results of the extensive submarine surveys conducted around the Hawaiian volcanoes have highlighted the important role of large-scale edifice destabilisation and collapse in constraining the morphological and structural evolution of marine volcanoes. Any chosen point in the lifecycle of such an edifice represents a 'snap-shot' of a continuing conflict between constructive forces represented by endogenous and exogenous growth during, respectively, intrusive and extrusive activity, and destructive influences dominated by mass-wasting due to slumping, avalanching, and other erosive mechanisms. As suggested by Fornari & Campbell (1987), these latter phenomena may actually act in concert in the long term to restabilise the edifice by widening its base.

Many collapse scars on island and coastal volcanoes, such as the Grand Brûlé on Piton de la Fournaise, the Valle del Bove on Etna, and the Sciara del Fuoca on Stromboli were formed during the Holocene, and are only visible due to their youth. Older structures are, however, rapidly buried or covered, and may only be recognizable by unexplained variations in edifice morphology. At Mauna Loa, for example, anomalously steep slopes developed along the entire length of the west flank have been tentatively interpreted (Moore et al., 1994) in terms of young lava flows filling a sequence of older collapse amphitheatres. Similar anomalously steep slopes on the flanks of the Cumbre Vieja ridge on La Palma (Canaries), and other marine volcanoes may conceal structures generated by older collapse events, evidence for which may only be found in the submarine record.

The transport of volcanically-derived debris into the marine environment due to catastrophic landsliding may be enhanced by the triggering of large-scale turbidite formation at the distal ends of the avalanches. Garcia (1996), report turbidite currents related to Hawaiian landslides, which travelled over 1000km and flowed over sea-bed obstructions 500m high. Similarly, Weaver et al. (1994), present evidence for turbidity currents over 600km in length which appear to be related to slope failure on the flanks of the westernmost Canary Islands of La Palma and Hierro around 18 ka BP.

Further submarine surveys, using increasingly advanced imaging techniques, can be expected, over the next decade, to improve our knowledge of the morphologies and structures of volcanic landslides and associated deposits emplaced in the submarine environment. The numbers of landslides recognised are also certain to increase, as imagery is obtained for as yet relatively poorly studied volcanic island chains such as the Canaries and the Azores. Already, however, there is a sufficient body of data to indicate that repeated flank collapse is a ubiquitous occurrence in the normal lifecycle of marine volcanoes.

8 • HAZARD IMPLICATIONS OF VOLCANO INSTABILITY AND LATERAL COLLAPSE

As shown by collapse of the northern flank of Mount St. Helens in 1980, the consequences of edifice failure can be both dramatic and catastrophically destructive. Failure to forecast such an event and/or to initiate the appropriate mitigation procedure - which can only involve evacuation of the entire population of the area at risk - will result in major loss of life. With velocities in excess of 100m/sec (360km/h), and momentum sufficient to mount topographic barriers hundreds of metres high, no man-made structures can survive the impact of a major debris avalanche. The area affected is also likely to be large; the relatively small-scale debris avalanche at Mount St. Helens travelled over 24km from its source, but this is minimal compared with the 120km long late-Pleistocene debris avalanche at Nevado di Colima (Columbia) (Stoopes & Sheriden, 1992), which covers an area of around 2200km^2. A volcanic landslide on this scale has not been observed during historic times, and the problems involved in forecasting and mitigating such an event, particularly in a densely populated area, would be enormous. Nevertheless, such a scenario will eventually occur and provision must be made for mitigating its effects.

Edifice failure at Mount St. Helens was particularly important, from the hazard mitigation point of view, because it demonstrated clearly that volcanic landslides occurring in the presence of a shallow magma body can generate a whole spectrum of destructive phenomena in addition to the landslide itself, including lateral blast, lahars, pyroclastic flows, and extensive ash-fall. Furthermore, the speed with which these phenomena were unleashed following landslide initiation indicated that no mitigation procedures could be implemented once the sequence of events had started. The lesson being that such measures must be in place well in advance of the expected event.

Assessing the hazard posed by structural instability at a particular volcano should ideally adopt a two-pronged approach based upon mapping and surveillance. The former provides information on the nature, extent, and frequency of past collapse events, and identifies areas likely to fail at some point in the future, while the latter concentrates on detecting and monitoring the onset and development of instability using electronic distance measurement (EDM) and related techniques. Because edifice failure and debris avalanche formation is often an episodic event which constitutes part of the normal life-cycle of a volcano, considerable information about the behaviour of a future landslide can be gleaned from examining older deposits associated with similar events. Determining the extent of such deposits is particularly important in hazard zonation map preparation, as it provides an estimate of the area which might be affected by a future debris avalanche. The size-estimates and shapes of hazard zones determined on this basis must always, however, be regarded only as a guide to future eruptive behaviour. A volcano may not, for any one of a number of reasons, precisely replicate past activity, and destructive phenomena may affect larger or different parts of the surrounding terrain during a future collapse event. Furthermore, not all the effects associated with edifice failure may be sufficiently well-preserved in the geological record. At Mount St. Helens, for example, the hazard zonation maps of Crandell & Mullineaux (1978) proved to be highly accurate for all phenomena except the lateral blast which accompanied the May 18th landslide. This extended three times further, and covered an area up to fifteen times greater than predicted (Miller et al., 1981), probably reflecting the low preservation potential of the products of earlier, similar events. In assessing the potential for future instability and failure, detailed structural mapping also has an important role to play, particularly in identifying features which may provide clues as to the nature and locations of future collapse. These may include creeping/episodic faults, peripheral thrust faults and folds, old collapses, active rift zones, and zones of alteration. Recognizing changes with time of fault geometries or dyke-zone orientations may also be important in terms of reflecting stress regime modifications prior to edifice destabilisation.

Debris avalanche production in active volcanic terrains may be accompanied by other destructive phenomena. A particular threat lies in the transition, during transport, from debris avalanche to debris flow (lahar), due to the melting of snow and ice fields or to the entrainment of surface water during the collapse event. Such behaviour effectively increases the length of the runout, compared to dry avalanches, thereby enlarging the extent of damage and destruction. Debris avalanches may also provide conditions favouring lahar formation by damming water catchments and forming new lakes which may drain catastrophically (Costa & Shuster, 1988). Additionally, lahars and mud-laden floods may be generated due to dewatering of the avalanche material (Janda et al., 1981), or may form by means of remobilization caused by heavy rainfall. In these ways, a major collapse event may provide a sufficient source of debris to feed numerous precipitation-related lahar/flood events over a period of years or decades. This is an extended-hazard problem which has become familiar to the inhabitants of towns in the vicinity of the Pinatubo volcano (Philippines) since the eruption in June 1991 (Pierson, 1992). Although here the source of the lahars is a thick mantle of pyroclastic flow, rather than debris avalanche, material, a similar scenario could be expected to follow emplacement of a major volcanic landslide in an area of very high rainfall.

The Nevado del Ruiz (Columbia) catastrophe in 1985 (Herd et al., 1986; Voight, 1990), in which close to 25,000 lives were lost, demonstrated graphically the devastating potential of lahars. Ironically, considering this was one of the worst volcanic disasters of the century, the detrimental effects of these phenomena are some of the easiest to mitigate. This is largely due to their being more topographically constrained than debris avalanches, tending to follow river valleys and pond in areas of low relief. It should be relatively easy, therefore, given the funding and political will, to effectively mitigate the lahar problem at any particular volcano. This may be accomplished by means of (i) a judicious construction policy which avoids susceptible areas, (ii) an effective monitoring system, based on alarmed trip-wires and seismometers, which is able to provide sufficient time for evacuation of the area, and (iii) having in place an effective plan for rapid evacuation. Damage to property can be minimised by the upstream installation of a sequence of sediment dams and baffles designed to reduce the sediment and boulder load thereby reducing the destructive power of the lahar.

The preference for destabilization and collapse on the seaward-facing flanks of volcanic edifices has already been discussed in terms of the buttressing effect of adjacent terrain on the landward side. From the hazard point of view, this effect has major implications, favouring as it does the formation of volcanogenic tsunamis. Some 5% of all tsunami are estimated to have been formed by volcanic activity, and at least one fifth of these result from volcanic landslides (Smith & Shepherd 1996). One of the most recent occurred at Harimkotan (Severgina) volcano (Kurile Islands) during 1933, when a small (0.5km^3) debris avalanche entered the sea and generated a 9m high tsunami (Belousov, 1994). As revealed by the Mount Unzen (Japan) collapse in 1792, even small landslides can generate highly destructive waves if they enter a large body of water. At Unzen, a collapsing volume of only about 0.34km^3 (Hayashi & Self, 1992), which was not connected with volcanic activity, entered Ariake Bay and triggered a tsunami which caused 14,500 deaths. Similar 'cold' collapses are probably relatively common events, particularly on steep-sided volcanoes in the marine environment. Kick 'em Jenny volcano in the Lesser Antilles, for example (Smith & Shepherd 1996), must be viewed as a volcanogenic tsunamis source. This submarine volcano has erupted 12 times in the last 53 years and possesses steep flanks bearing signs of previous collapse events. The steep-sided Stromboli volcano in the Aeolian Islands (Tibaldi 1996; Kokelaar & Romagnoli 1995) might also be considered a prime candidate for flank collapse sufficient to generate a tsunamis capable of affecting the coastal regions of northern Sicily and western Calabria. Begét & Kienle (1992) also highlight the tsunami risk at Mount St. Augustine volcano (Alaska). They report the formation of a 20m high tsunami due to the emplacement of a small debris avalanche into the Cook Inlet during the 1883 eruption, and expect a similar event to occur sometime during the next century. Some of the largest volcanogenic tsunamis were undoubtedly associated with emplacement of the Hawaiian Island debris avalanches. A wave-train associated with a collapse on the flanks of Lanai volcano around 105 ka BP appears not only to have reached an elevation of around 375m on the island itself (Moore & Moore, 1984), but also to have crossed the Pacific and impinged energetically upon the coast of New South Wales in Australia. Here, Young & Bryant (1992) report the results of catastrophic wave erosion related to the tsunami, at heights of at least 15m above present sea level. Both forecasting the onset of such a collapse and mitigating its effects are beyond current capabilities, and likely to remain so for some time. More recently, tsunami deposits located around 100m above current sea level have been reported from both Gran Canaria and Lanzarote (S. J. Day, pers.comm.), providing further evidence of major lateral failure of the Canary island volcanoes during the Quaternary.

9 • MONITORING VOLCANO INSTABILITY AND FORECASTING FAILURE

Successfully forecasting collapse in active volcanic terrains remains strongly dependent upon the monitoring of ground deformation and displacement using geodetic and related methods. Such techniques are particularly important in identifying sites of increasing destabilization on the flanks of newly reactivated volcanoes, and in monitoring the development of instability with a view to defining progressively smaller predictive 'windows' in order to attempt to forecast the timing of eventual structural failure. Defining such windows relies upon observing an increasing acceleration in the rate of deformation or displacement, with failure becoming increasingly likely as the rate of deformation becomes greater.

This technique proved particularly effective, when combined with data gathered from other monitoring methods, in forecasting dome-destroying eruptions at Mount St. Helens during the early 1980s (Swanson et al. 1983, 1985). Unfortunately, it was not so successful in predicting the major landslide event of May 1980. In this case, cumulative growth of the cryptodome-induced bulge demonstrated a constant displacement rate right up to the time when structural failure and landslide initiation was triggered by a magnitude 5 earthquake (McClelland et al. 1989). It is a matter for conjecture whether or not an acceleration in the growth rate of the bulge would have been observed had it not been prematurely detached from the northern flank by seismogenic ground accelerations. The situation does, however, illustrate an important point, which is that external events - in this case seismic activity - may accelerate the onset of instability caused by a separate, although in this case related, phenomena (eg magma intrusion), and may initiate failure sooner than might otherwise be expected.

Monitoring edifice instability has also been shown to be a useful tool in predicting future eruptive activity at Etna, where geodetic monitoring has revealed a relationship between the rate of downslope creep on the upper eastern flanks of the volcano and the timing of future eruptions. Murray & Voight (1996) propose that accelerating rates of downslope creep occur in response to increasing magma pressures. The creep behaviour eventually reduces the effective tensile strength of the rock to a level at which it is exceeded by magma pressure, thereby permitting eruption. On the basis of inverse-rate analysis of geodetic data accumulated during the 1980s, the authors also believe that this method permits good eruption predictions more than three months in advance.

A range of low- and high-tech ground deformation monitoring techniques are now in use for detecting and observing the surface displacements at active volcanoes which may signal increasing instability and presage forthcoming failure. At the low-tech end of the spectrum, simple steel tapes have been effectively used to measure displacements across basal thrust faults reflecting the progressive growth of the previously mentioned post-May 1980 lava dome at Mount St. Helens, contributing invaluable data towards successful forecasts of future episodes of dome destruction (Swanson et al. 1983, 1985). More commonly used, less risky, but more expensive techniques involve using either infra-red or laser electro-optical distance meters (EDMs), which are capable of monitoring horizontal distance changes of only a few centimetres over distances of up to tens of kilometres (eg McGuire et al., 1990; 1991; Iwatsubo & Swanson, 1992; Murray et al., 1995;). Displacement data gathered in this way may also usefully be supplemented by tiltmeters and precise-levelling surveys (Dzurisin, 1992; Murray et al., 1995; Toutain et al., 1995) to build a more comprehensive picture of the pattern of deformation associated with a destabilization event. Airborne (Garvin 1996), and space-based systems are also becoming more important, and the increasingly accessible and cost-effective global positioning system (GPS), in particular, can be viewed as an effective new tool in the armoury of geodetic hardware (Shimada et al., 1989; Nunnari & Puglisi, 1995). Radar interferometry techniques using successive synthetic aperture radar (SAR) images from satellite platforms have also been shown recently (Massonet et al., 1993) to have enormous potential in measuring centimetric surface displacements associated with co-seismic deformation fields around active faults. This technique is now being applied in volcanology (Massonet et al., 1995) and offers the possibility of monitoring very precisely, and at one go, changes in the overall surface deformation pattern at a chosen volcano. On the ground, better prediction of the timing of landslide initiation may come from the seismic monitoring of seismogenic faults which have the potential to act as slide surfaces, the application of acoustic emission techniques, which may permit detection of very small-scale micro-fracturing preceding and presaging major structural failure, and ground penetrating radar (GPR) which may be suitable for mapping fracture patterns in unstable areas.

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