Lateral collapse and tsunamigenic potential of marine volcanoes

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Abstract: The predominantly constructive life cycles of large, long-lived, stratovolcanoes and basaltic shields are punctuated by transitory episodes during which large volumes of material are divested from the flanks. Such shedding typically takes place catastrophically in the form of a lateral collapse, generating a debris avalanche and leaving a scar that may attain caldera dimensions. Collapse may follow instability development arising from a single, discrete, event, such as a crypto-dome intrusion, or may be the end-product of progressive destabilization over a long period of time. Lateral collapses may also occur at persistent slumps, which may have been active over periods as long as 10^4–10^5 a prior to catastrophic failure. Collapse velocities may exceed 40 m s^{-1}, leading to completion of the process within a few hundreds of seconds. Collapse volumes span several orders of magnitude, ranging from less than 1 km^3 to more than 10 km^3 at many continental and subduction-zone volcanoes, to 1000 km^3 or more at the great basalt shields of Hawaii. Lateral collapse may be accompanied by a wide range of associated hazards, including atmospheric shock wave, pyroclastic flows and surges, extensive tephra fall, and secondary lahars. For volcanoes in the marine environment, potentially destructive and lethal tsunamis can be added to the inventory. Here, the potential for a lateral collapse at an ocean island volcano to generate a 'mega-tsunami' (more than 100 m high at source and destructive at oceanic distances) is discussed and evaluated.

The life cycles of large, long-lived volcanoes are primarily constructive. Periodically, however, edifice growth is punctuated by episodes of instability, leading to structural failure and collapse of the flanks. Such lateral collapse typically generates a debris avalanche and leaves behind a scar that can be of caldera dimensions, hence the discussion of the process in this volume. In this context, it is noteworthy that the 'type' caldera – Caldera Taburiente (Fig. 1) on the Canary Island of La Palma – was formed by lateral collapse and subsequent erosion, rather than by vertical subsidence along ring faults. Volcano lateral collapses are now known to be ubiquitous, and are estimated by Siebert (1992) to have happened around four times a century over the past 500 years. The phenomenon may prove to be even more commonplace, with Belousov (1994) drawing attention, for example, to three major, twentieth century lateral collapses in the Kurile–Kamchatka region of Russia alone. Increased interest in volcanic edifice destabilization and collapse was generated by the climactic Mount St Helens eruption of 18th May, 1980, which was triggered by the catastrophic failure of the north flank (Lipman & Mullineaux 1981). Subsequent surveys (e.g. Ui 1983; Siebert 1984) highlighted numerous examples of lateral collapse in the geological record. In particular, Inokuchi (1988) described over a hundred debris avalanches from lateral collapses at Quaternary volcanoes in Japan, while Francis (1994) observed that three-quarters of large volcanoes in the Andes have experienced lateral failure. Lateral collapse is no respecter of edifice size or form, and can occur at individual cinder cones and lava domes, large strato-volcanoes, and major basaltic shields; it also occurs at volcanoes of all compositions. Collapse volumes bridge several orders of magnitude, ranging from less than 1 km^3 to $\geq 10^3$ km^3 at many continental and subduction-zone volcanoes, to ≥ 1000 km^3 at the great basalt shields of Hawaii.

As demonstrated by the 1980 eruption of Mount St Helens, lateral collapse is a major hazard (Siebert *et al.* 1987), which has taken more than 23 000 lives since the middle of the seventeenth century (Table 1). The main product is usually a debris avalanche. If the collapse triggers an explosive eruption, as at Mount St Helens, this may be accompanied by an atmospheric shock wave, pyroclastic flows and surges, secondary lahars and widespread tephra-fall. Whether associated with an eruption or not, lateral collapse in the marine environment has the potential to generate large tsunamis. An estimated five per cent of all recorded tsunamis are attributed to volcanic activity and more than one per cent of these are ascribed to the entry of landslides into the ocean (Smith & Shepherd 1996). Most recently, in 2002, two small landslides (total volume *c.* 5.6×10^6 km^3) within the prehistoric Sciara del Fuoco lateral collapse scar on the island volcano of Stromboli, generated a

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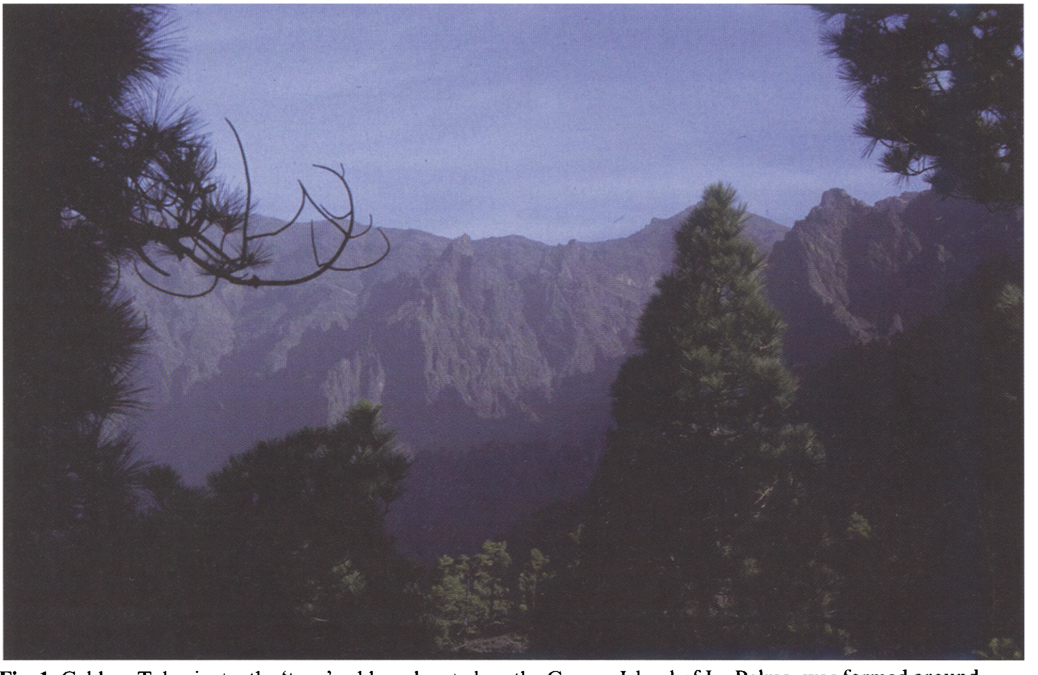

Fig. 1. Caldera Taburiente, the 'type' caldera, located on the Canary Island of La Palma, was formed around 500 000 years BP by lateral collapse, and has subsequently been modified by erosion.

Table 1. *Historical lateral collapse events that have resulted in loss of life (data from Siebert & Simkin 2002)*

Volcano	Year	Deaths	Notes
Komaga–Take (Japan)	1640	700	Deaths due to tsunami
Oshima–Oshima (Japan)	1741	1 475	Earthquake-triggered; deaths due to tsunami
Papandayan (Indonesia)	1772	*c.* 3 000	Forty villages obliterated
Unzen (Japan)	1792	14 528	Deaths due to tsunami
Bandai-san (Japan)	1883	*c.* 300	Several villages buried
Ritter Island (Papua New Guinea)	1888	*c.* 3 000	Deaths due to tsunami
Mount St Helens (USA)	1980	57	First detailed observations of lateral collapse event
Casita (Nicaragua)	1998	*c.* 1 500	Flank collapse triggered debris flow

tsunami that reached 10 m a.s.l. along the east coast of the island, causing substantial damage to coastal communities (Bonaccorso *et al.* 2003).

Landslides from ocean-island volcanoes (Whelan & Kelletat 2003; Keating & McGuire 2000, 2004) are among the biggest catastrophic mass movements on the planet. Around 70 major landslides have been identified around the Hawaiian Island archipelago (Fig. 2), the largest having volumes in excess of 5000 km^3 and lengths of over 200 km (e.g. Moore *et al.* 1994). Such volcanic landslides are now proving to be widespread in the marine environment and have been identified around other island groups, including the Canary Islands (Fig. 3), Cape Verde Islands and Galapagos Islands, and adjacent to individual island volcanoes including Stromboli (central Mediterranean), Piton des Neiges and Piton de la Fournaise (Réunion Island, Indian Ocean), Tristan de Cunha (South Atlantic), Augustine Island (Alaska), and Ritter Island (Papua New Guinea) (Holcomb & Searle

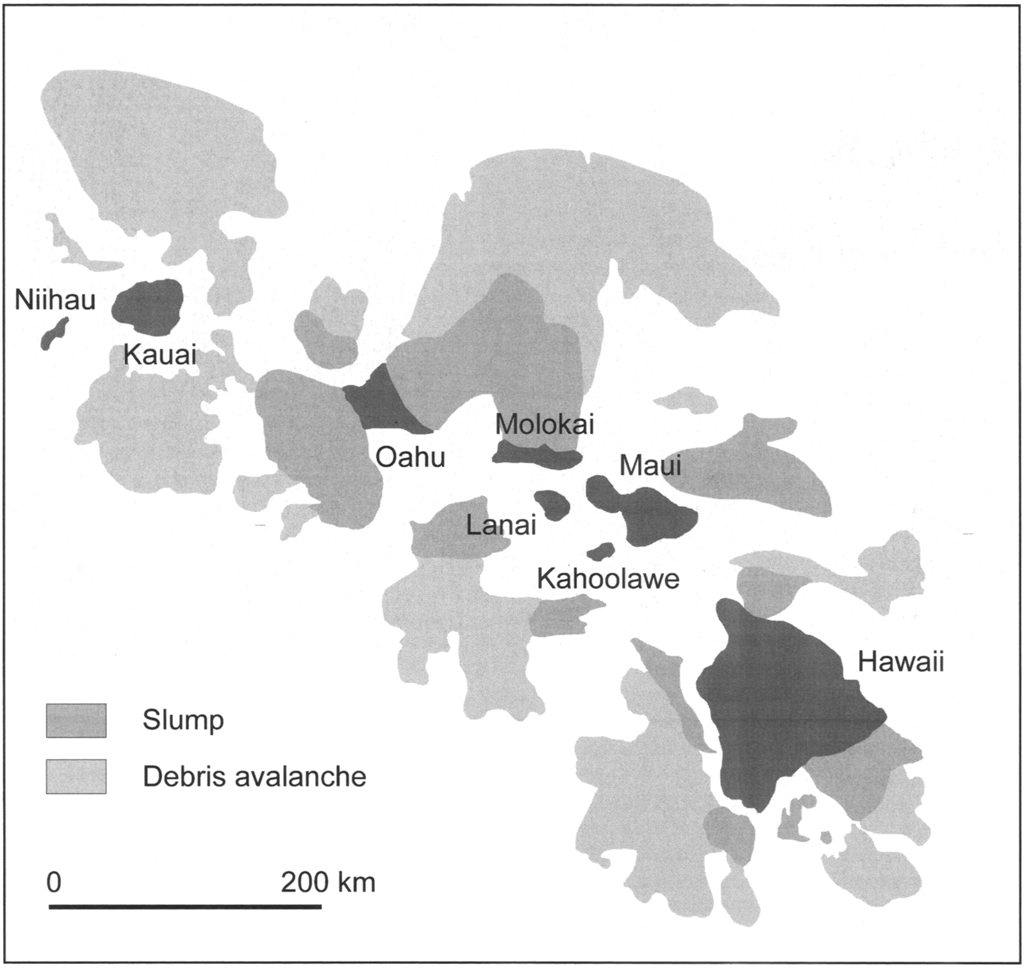

Fig. 2. Distribution of lateral-collapse related-slumps and debris avalanches around the eastern Hawaiian Islands. Modified from Whelan & Kelletat (2003).

1991; McGuire 1996; Keating & McGuire 2000, 2004).

Due to the disposition of tectonic plates and the frequent congruity of plate margins and coastlines, many of the world's active volcanoes either form islands or are coastally located; indeed, 57% of currently active (*c.* 600) volcanoes occupy such locations (McGuire *et al.* 1997). Assuming that the *c.* 1500 Holocene volcanoes (Siebert & Simkin 2002) are regarded as active, therefore, over 850 volcanoes can be considered as potential tsunami sources. Abundant evidence exists of lethal tsunamis triggered at volcanoes in recent human history (Table 2), and many of these have been attributed to landslides arising from flank collapse. Similarly, the geological record supplies evidence of giant tsunamis triggered by major lateral collapse (e.g. McMurtry *et al.* 2004; Pérez-Torrado *et al.* 2006).

Recent attention has focused on the potential for very large lateral collapses at island volcanoes to trigger so-called 'mega-tsunamis'. No strict definition exists for the term 'mega-tsunami', which is essentially a media-driven descriptor. McGuire (2006), however, has proposed an arbitrary definition, reserving the term for waves that are ≥100 m in height at source, and which remain destructive at oceanic distances. While the work of McMurtry *et al* (2004) and Pérez-Torrado *et al.* (2006) supports waves well in excess of 100 m close to source, incontrovertible evidence is currently lacking – either from observations of historical tsunamis, or, in the geological record – for such waves remaining

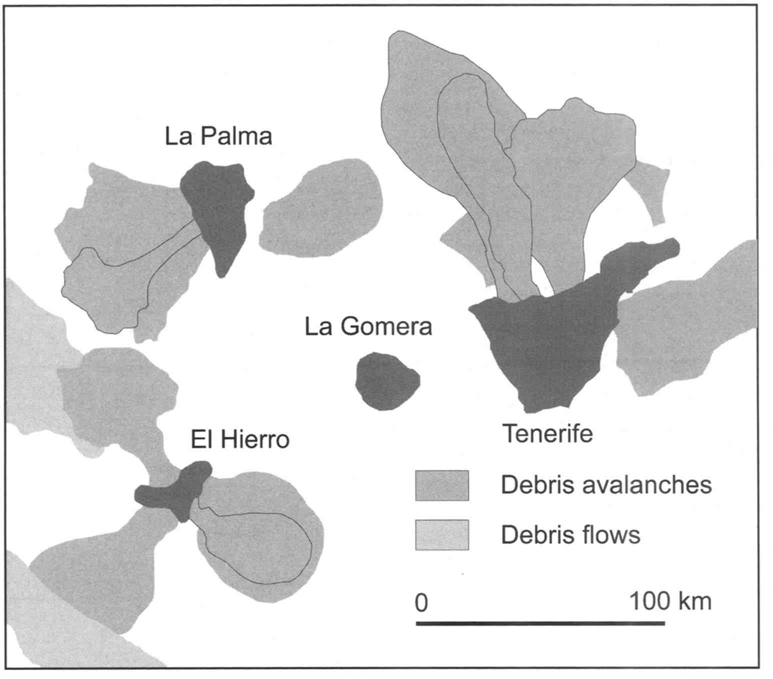

Fig. 3. Distribution of debris-avalanche and debris-flow deposits adjacent to the western Canary Islands. Modified from Whelan & Kelletat (2003).

Table 2. *Historical volcanogenic tsunamis that resulted in significant loss of life (data from Siebert & Simkin 2002)*

Volcano	Year	Cause	Deaths	Notes
Komaga–Take (Japan)	1640	Landslide	700	
Santorini (Greece)	1650	Eruption	50	
Long Island (Papua New Guinea)	1660	Eruption	*c.* 2 000	Tsunamis and pyroclastic flows
Gamkandra (Indonesia)	1673	Eruption	Many	
Oshima–Oshima (Japan)	1741	Landslide	1 475	
Unzen (Japan)	1792	Landslide	14 528	
Tambora (Indonesia)	1815	Eruption	Many	10 000 killed by direct effects of eruption
Ruang (Indonesia)	1871	Landslide	400	Collapse of lava dome
Krakatau (Indonesia)	1883	Eruption	36 417	Most killed by tsunamis
Ritter Island (Papua New Guinea)	1888	Landslide	*c.* 3 000	Waves 12–15 m high
Ta'al (Philippines)	1965	Eruption	>200	Most drowned due to boats capsizing
Iliwerung (Indonesia)	1979	Landslide	539	Waves 9 m high

large enough to be destructive at oceanic distances. This capability is the focus of intense debate, particularly with respect to a possible lateral collapse of the Cumbre Vieja volcano on the island of La Palma in the Canary Islands (e.g. Mader 2001; Ward & Day 2001, 2003; Pararas-Carayannis 2002; Gisler *et al.* 2005; Masson *et al.* 2006). Given its major hazard and risk implications, the potential for volcano lateral collapse to generate waves that are destructive at oceanic distances forms the core of this review paper.

Development of instability

Active volcanoes are dynamically evolving structures, the growth and development of which are typically punctuated by episodes of edifice instability, structural failure, and ultimately collapse (McGuire 1996). Volcano instability has been defined by McGuire (1996, p. 1) as 'the condition within which a volcanic edifice has been destabilized to a degree sufficient to increase the likelihood of the structural failure of all or part of the edifice'. This condition is the product of the manifold factors that together have a tendency to destabilize volcanoes and make them more prone to lateral collapse than other elevated landforms. In particular, volcanoes are morphologically dynamic, undergoing continuous changes in form in response to both magma intrusion and extrusion. Destabilizing morphological changes related to magma intrusion may arise due to: (1) episodic replenishment and tapping of an established magma reservoir leading to cycles of edifice-wide tumescence and contraction; (2) discrete emplacement of a new magma body of irregular form (e.g. a crypto-dome); or (3) repeated injection of minor intrusions, in particular dykes, along preferential trends (rift zones). All three mechanisms may lead to eventual edifice destabilization and lateral collapse, although the length of the preparation phase prior to collapse will depend upon the nature of the intrusive behaviour. Weeks to months may be sufficient time for a single, large intrusive event – such as the 1980 Mount St Helens crypto-dome – to trigger collapse (Voight *et al.* 1981). In contrast, thousands or tens of thousands of years are likely to be required before instability due to progressive dyke-induced rifting, such as that characterizing the flanks of the Hawaiian volcanoes, is translated into lateral collapse. Although less prevalent, instability and collapse may also result from magma extrusion. Potential mechanisms include the progressive loading of a steep slope by the products of successive eruptions, especially lava flows (e.g. Murray & Voight 1996), or the growth and over-steepening of an active lava dome (e.g. Begét & Kienle 1992).

While magma is often a key element, many other factors may contribute to the destabilization and lateral collapse of a volcanic edifice. Of particular note are the typically elevated levels of both volcanogenic and tectonic seismicity, environmental factors such as precipitation (Day *et al.* 2000) and sea-level change (McGuire *et al.* 1997), morphology, internal composition, and strength of the edifice (e.g. Reid *et al.* 2001; Thomas *et al.* 2004*a*; Cecchi *et al.* 2005), and the behaviour, form or structure of the underlying basement (e.g. Wadge *et al.* 1995; Belousov *et al.* 2005)

Although this paper focuses on large-volume slope failure, it is important to note that growing volcanoes may become destabilized and experience subsequent collapse at any scale, ranging from minor rock-falls with volumes in the order of a few hundred to a few thousand cubic metres, to the giant 'Hawaiian-type' mega-slides involving $\geq 1000\ km^3$. Low-volume collapses occur many times every year, while the largest events occur at intervals of 10^4–10^5 a. Often the conditions leading to instability and collapse are persistent throughout the history of the volcano, leading to cycles of growth and collapse on characteristic time-scales. At Augustine volcano (Alaska), for example, Begét & Kienle (1992) have recognized at least 11 landslide events over the past 2 ka, recurring every 150–200 years. These they relate to episodes of reconstructive lava-dome growth, each terminated by collapse ($< 1\ km^3$) due to over-steepening. At Colima (Mexico) the cycle of edifice destabilization is longer, culminating in large-scale collapse ($\leq 12\ km^3$) and landslide formation every few thousand years (Komorowski *et al.* 1994), while at Hawaii a complete collapse-to-collapse cycle, characterized by volumes of up to $5\,000\ km^3$, has a duration of between 25 and 100 ka (Lipman *et al.* 1988).

Large-volume lateral collapse

Despite the ubiquity of the phenomenon, some volcanoes clearly have a greater proclivity for large-scale lateral collapse than others. While not correlated with slope angle – the largest landslides occurring around the margins of gently sloping oceanic shield volcanoes – such behaviour is generally confined to major volcanic structures, particularly those of a polygenetic nature (i.e. constructed over a long period of time from several centres of activity). There appears to be an upper limit to the size of single landsliding events, which tend to have volumes of 20% or less of the volume of the failed edifice (Holcomb & Searle 1991). This upper limit on landslide volume may be explained if the maximum width of the collapse is limited by conjugate shear planes, the angle of which is usually smaller than the 90° expected under ideal conditions.

McGuire *et al.* (2002) note that the maximum possible lateral collapse occurs when the entire thickness of a volcano collapses in a given sector. Approximating a volcanic edifice to a cone, therefore, the maximum ideal collapse volume is 25% of the volume of the whole cone (90/360 = 0.25). The observed 20% maximum volume may thus reflect secondary departures from the ideal case (e.g. maximum angle of collapse sector is less than 90°, or the maximum collapse thickness is less than the height of the edifice). Macroscopic stress distributions within cones may thus define an upper limit to collapse volumes. The wholesale structural instability that can generate such large collapses may develop at both continental and subduction-zone strato-volcanoes, such as Etna (Sicily), Rainier (Cascade Range, USA), Colima (Mexico), and Fuji (Japan), and at marine volcanoes including Mauna Loa and Kilauea (Hawaii), Piton de la Fournaise (Réunion Island, Indian Ocean), and La Palma and El Hierro (Canary Islands). The fact that continental and marine volcanoes are characterized by very different morphologies, structures and magmatic plumbing systems serves further to demonstrate that the development of instability and lateral collapse is not restricted to a specific set of environmental conditions or of volcano composition.

Polygenetic edifices in continental and subduction zone environments are typically strato-volcanoes composed of interleaved horizons of lava and fragmental pyroclastic materials, which together build a mechanically unsound structure that may be further weakened by hydrothermal alteration. The potential for increased instability and structural failure is compounded by the steep (typically $\geq 20°$) slopes characteristic of such volcanoes, and the high precipitation rates that often accompany their elevation and which may contribute to destabilizing changes in edifice pore pressures. Counter-intuitively, the largest lateral collapses are associated with those volcanic edifices that have the most subdued topographic expression. Large basaltic shield-volcanoes, such as those making up the Hawaiian Islands, have slope angles as low as 3°, and are mechanically stronger than their continental strato-cone counterparts, being composed almost entirely of superimposed lava flows. Clearly, the development of instability at such edifices, and subsequent flank collapse, reflects a different set of conditions to that which pertains at steeper volcanoes. In particular gravity-induced edifice spreading, augmented by a growing hot, viscous core (Borgia 1994) or dense outward-creeping masses of olivine-rich cumulates (Clague & Denlinger 1994), may play an important role in edifice destabilization. This may be enhanced by progressive rifting associated with repeated dyke emplacement over a long period of time, local seismicity, changes in edifice pore pressures, and environmental factors such as large, rapid changes in sea-level.

The observation that maximum collapse volume increases as a volcano's mean slope angle becomes more gentle is consistent with the interplay of gravity (driving collapse) and rock strength (resisting collapse). Following the argument in McGuire *et al.* (2002), for shear failure to occur, the mean gravitational stress over the future failure surface will be in the order of $\rho g h \sin\alpha$, where h is the maximum depth to the shear surface, ρ is the mean density of overlying rock, and α is the mean angle of slope of the ground. (Strictly, this relation holds only for failure as a slab with the failure surface parallel to the ground; however, to an order of magnitude, it will also hold for other common failure geometries, such as a wedge-shaped failure, provided that the length of the failure surface, in the direction of collapse, is much greater than h.) For geometrically similar failure patterns, h will also be approximately equal to $CV^{1/3}$, where V is the collapse volume and C is a geometric constant. For slope failure under shear, the driving stress must equal the resisting strength, S, from which it follows that $\rho g C V^{1/3} \sin\alpha \propto S$.

Hence, for negligible variations in bulk density among volcanoes, the collapse volume is approximately proportional to $(S/\sin\alpha)^3$, yielding an inverse relation between volume and slope angle. Since typical values of α on volcanoes range from about 3° to 20°, the variation in collapse volume between gentle and steep volcanoes is expected to be about 250–300 ($\approx (\sin 20/\sin 3)^3$). The corresponding ratio observed for maximum collapse volumes (from *c.* 10^3 km^3 to *c.* 1 km^3) is on the order of 5 000–10 000, some 20 to 30 times more than expected from changes in slope alone. Thus the bulk strength of volcanoes must also vary by as much as a factor of three, with larger strengths being associated with shallower slopes. Although such a relation has to be verified in detail, it is consistent with the observation above, that, compared with steep-sided edifices, volcanoes with shallower slopes are expected to contain fewer major horizons of weak pyroclastic material and, thus, to have a greater bulk strength.

In practice, determining the bulk strength of an edifice is problematic because it is a function not only of the mechanical properties of constituent rock, but also of a volcano's structure. Thomas *et al.* (2004*a*) describe a volcanic edifice, at its most basic level, as being 'made up from discrete layers of different rock types cut through on a range of scales by pervasive and penetrative discontinuities including faults, fractures and contact surfaces.' Assigning a meaningful overall strength to such a body is clearly difficult, and laboratory-determined rock properties only provide a part of the picture. Schultz (1995), for example, determined that the rock-mass compressive strength of basalt is up to 80 per cent weaker than its intact compressive strength as measured under laboratory conditions. Applied to volcanic rocks from Snowdonia (North Wales and Tenerife (Canary Islands), established geotechnical engineering methodologies for estimating strength criteria and other rock-mass properties, Thomas *et al.* (2004*a*) show that edifice rock-mass compressive strength may be up to 96% lower than that of intact rock value based on laboratory tests. Indeed, Thomas *et al.* (2004*a*) conclude that, in general, volcanic edifices can be very weak, with cohesive strengths less than 1 MPa and rock-mass angles of friction ranging from 28 to *c.* 38°. They also show that, for crystalline volcanic material, the rock-mass strength is largely independent of either composition or age. The application of rock-mass strength determination to specific volcanoes is rare, but has been attempted for Stromboli by Apuani *et al.* (2005) in light of the 2002 flank collapse and tsunami. Using geotechnical methodologies comparable to those of Thomas *et al.* (2004*a*), they determine cohesive strengths ranging from 0.6–1.4 MPa for pyroclastic units to 1.5 to 3.9 MPa for lava flows, and friction angles ranging, respectively from 15 to 23° and from 31–43°.

Bulk strengths might be smaller than for intact rock. To propagate a new failure surface, however, it may be necessary to break intact rock between existing failure planes. Local stresses, therefore, may still need to overcome the strength of intact rock, and this requirement might constrain rates of deformation before edifice collapse.

The bulk strength of a volcanic edifice, or a part thereof, can be significantly reduced by hydrothermal alteration, and a number of large collapses have involved weak, clay-rich, hydrothermally altered rock (e.g. Siebert *et al.* 1987; Lopez & Williams 1993; Day 1996; van Wyk de Vries *et al.* 2000; Reid *et al.* 2001; Cecchi *et al.* 2005). From a hazard point of view, the contribution of such material in the collapse process is important, as its ability to retain pore-water provides a ready means for transforming the resulting debris avalanche into debris flows with longer potential run-outs (e.g. Iverson *et al.* 1997). This is the situation at Mount Rainier volcano (Washington State, US), which has generated over 55 Holocene debris flows that have travelled for distances of over 70 km. Reid *et al.* (2001) attribute this behaviour to the preferential collapse of areas of voluminous, weak, hydrothermally altered rock on the upper flanks of the edifice. They also show that areas of relatively unaltered rock have remained stable over the period.

On a larger scale, Cecchi *et al.* (2005) describe how hydrothermal alteration may weaken the entire core of a volcanic edifice, making it prone, initially, to slow spreading of the flanks and, later, to catastrophic collapse. One volcano that demonstrates this behaviour, Casitas in Nicaragua, suffered flank collapse during torrential rains that accompanied the passage of Hurricane Mitch in 1998 (Van Wyk de Vries *et al.* 2000; Kerle 2002), triggering a debris flow that took an estimated 1 500 lives.

Triggering lateral collapse

Volcano lateral collapse may only occur once the edifice or a part thereof has become sufficiently destabilized and, most critically, when sufficient strain has accumulated along the future slide plane. McGuire *et al.* (2002) propose that collapse triggering is related to particular processes or events that reduce the strength of the edifice to values much lower than normal, or to transient events such as intrusions or major earthquakes, that accelerate ground movement. Researchers have proposed a wide variety of such processes, of which some are extensions of processes known to operate in non-volcanic terrains, whereas others are unique to volcanoes. Table 3, from McGuire *et al.* (2002) divides these processes into overstressing processes (in which catastrophic failure occurs as a result of an increase in the stresses operating on the volcano) and weakening processes (in which failure results from a reduction in the mechanical strength of the volcano). It is noteworthy that many of the processes listed involve the effects of pore-fluid pressurization by a variety of mechanisms, some of which are similar to the liquefaction processes encountered in unconsolidated or partly consolidated sediments. Others involve the specific effects of fluid pressure upon fractures in more competent rock, and thermal expansion arising from magma intrusion which, in confined systems, induces thermal pressurization. On the basis of physical and numerical modelling, Thomas *et al.* (2004*b*) have determined that while pore-fluid pressurization alone is unlikely to cause collapse, it provides a mechanism for driving the edifice close to the

Table 3. *Processes proposed as triggers of volcano lateral collapse (after McGuire* et al. *2002)*

Process	Over-stressing (OS) or material weakening (W)	Limitations upon operation	Proposed examples	Reference(s)
Over-steepening by endogenous dome growth or crypto-dome emplacement	OS	Small-volume collapses only	Bezymianny (Russia) 1956; Mount St Helens (US) 1980	Voight *et al.* (1983); Paul *et al.* (1987)
Excess magma pressure from crypto-dome or dyke intrusion	OS		Etna (Italy)	McGuire *et al.* (1990)
Over-steepening by tilting of volcano above fault scarp	OS	Volcanoes above active fault systems	Socompa (Chile); Kamchatka (Russia)	Francis & Self (1987); Wadge *et al.* (1995); Belousov *et al.* (2005)
Over-steepening by erosion of buttressing flank	OS	Rapid erosion concentrated on lower flanks of volcano	Stromboli (Italy)	Tibaldi (1996)
Direct ground acceleration associated with earthquake	OS		Mount St. Helens 1980 (US)	
Static failure following alteration of edifice to materials (e.g clays) with abnormally low coefficients of friction	W	Volcanoes with intense hydrothermal systems only	Bandai-san (Japan) 1888; Casita (Nicaragua); Mount Rainier (US)	Siebert *et al.* (1987); van Wyk de Vries *et al.* (2000); Reid *et al.* (2001); Cecchi *et al.* (2005)
Mechanical compression of confined pore fluids by elastic deformation on intrusion emplacement, causing elevated pore-fluid pressures	W	Porous, fluid-saturated volcanoes with low permeability		Elsworth & Voight (1995)
Mechanical compression of confined pore fluids through anelastic deformation, particularly in shear or fault zones, causing elevated pore-fluid pressures	W	Porous, fluid-saturated volcanoes. Most likely when localized brittle or semi-brittle deformation style is dominant	Roque Nublo volcano (Gran Canaria, Spain)	Day (1996)
Co-seismic collapse of pore spaces causing pressurization of pore fluids (liquefaction *sensu stricto*)	W	Porous, fluid-saturated volcanoes		

Table 3. *Continued*

Heating of confined pore fluids by intrusions or magmatic gases, causing elevated pore-fluid pressures by thermal expansion effect	W	Porous, fluid-saturated volcanoes with low permeability		Elsworth & Voight (1995); Day (1996).
Influx of highly pressurized fluids expelled from greater depths, causing elevated pore-fluid pressures	W	Porous, fluid-saturated volcanoes	Unzen (Japan) 1792	Day 1996
Mechanical strain-weakening in localized faults and shear zones	W	Localized, brittle or semi-brittle deformation style		
Co-seismic strain weakening in fault rocks (acoustic fluidization mechanism)	W	Localized, brittle or semi-brittle deformation style		Melosh (1996)

critical state in which the factor of safety (F_s) is reduced to below one. Reid (2004) proposed that lateral collapses not associated with eruptions (e.g. Bandai-san, Japan, in 1888) may be triggered as a result of the deep intrusion of magma generating temporarily elevated pore-fluid pressures that propagate upwards into the edifice. Given appropriate rock hydraulic conditions, Reid demonstrated through numerical modelling that these pressures have the potential to destabilize the core of the edifice, leading to massive deep-seated collapse. For more detailed discussions of lateral collapse triggering mechanisms, the reader is referred to the reviews in McGuire *et al.* (2002) and McGuire (2003).

Lateral collapse in the marine environment

Volcano lateral collapse is ubiquitous in the marine environment, occurring in all tectonic settings, including island and continental arcs and 'hot-spot' related archipelagos and individual islands (Fig. 4). Collapses may be purely sub-aerial, may be confined wholly to the submarine flanks, or may involve material both from above and below sea-level. The most voluminous lateral collapses are recorded at large, long-lived, ocean-island volcanoes, where – as a consequence – the potential for future collapse provides the highest risk of generating major tsunamis. A comprehensive review of volcano instability and collapse within such environments, and the associated tsunami threat, is addressed in Keating & McGuire (2000), and evaluation of links between climate change and ocean island collapse in Keating & McGuire (2004).

The Hawaiian Islands typify those geological features associated with ocean island volcanoes that have histories of lateral collapse. They are surrounded by submarine aprons of allochthonous debris emplaced by sliding and slumping (e.g. Moore *et al.* 1994) (Fig. 2), which may grade into 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.* 1994), Tristan de Cunha (Holcomb & Searle 1991), the Galapagos Islands (Chadwick *et al.* 1992), the Canary Islands (Holcomb & Searle 1991; Weaver *et al.* 1994; Carracedo 1996; Wynn & Masson 2003; Masson *et al.* 2006) (Fig. 3), Stromboli and Alicudi

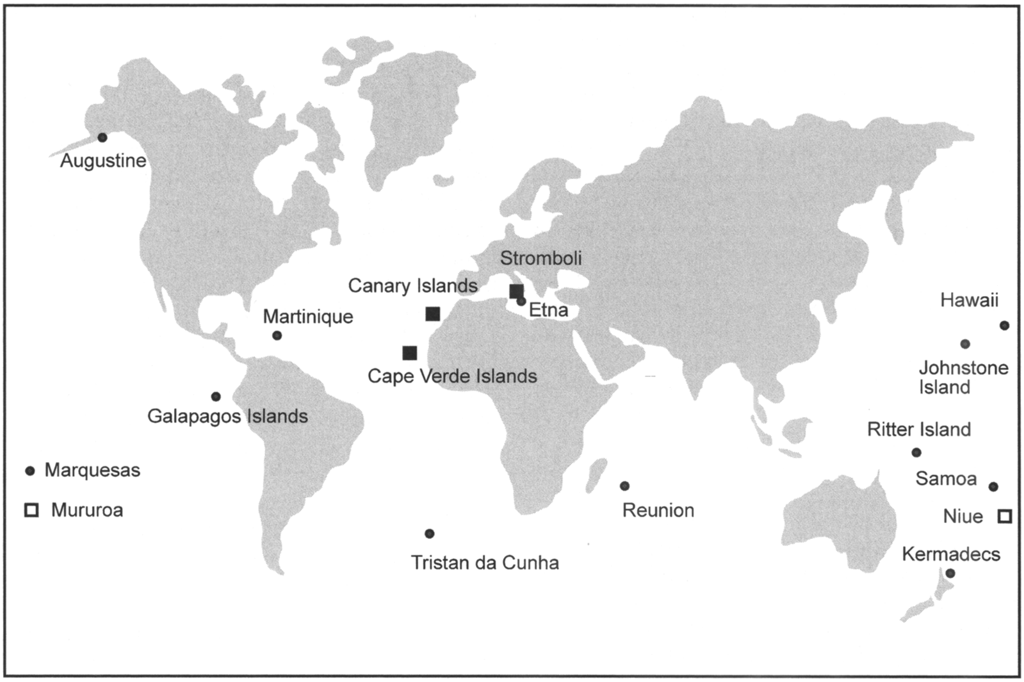

Fig. 4. Volcanic islands, archipelagos, and coastal volcanoes that display evidence of past collapse (*filled circles*), incipient instability (*open squares*), and past collapse together with incipient instability (*filled squares*).

(Aeolian Islands) (Kokelaar & Romagnoli 1995), and at Augustine Island (Alaska) (Begét & Kienle 1992).

The common occurrence of aprons of destabilized material around or adjacent to 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 like Hawaii, where younger centres (such as Kilauea) are buttressed on the landward side by older edifices (e.g. 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 down-faulting, or of the episodic production of debris avalanches, or a combination of all three. The relatively unstable nature of the seaward-facing flanks of a marine volcano is further enforced by the dynamic nature of the land–sea contact. Not only does marine erosion provide a constant destabilizing agent, but large changes in global sea-levels of up to 130 m, occurring over periods as short as 18 000 years, with catastrophic rises recently identified of 11.5 m in $<160 \pm 50$a (Blanchon & Shaw 1995), offer a means of modifying internal stress regimes and water pore-pressures in favour of edifice destabilization (McGuire *et al.* 1997).

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. Sixty-eight landslides with lengths in excess of 20 km have been identified. In a comprehensive review of landslide generation along the Hawaiian Ridge, Moore *et al.* (1994) highlight a number of characteristic features that are likely to be applicable to the destabilization 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 (Fig. 2) 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 100 km) and thick (up to 10 km), the latter are long (up to 230 km) and relatively thin (0.5–2 km). 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 is 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 2 km across; and evidence for uphill transport on the Hawaiian Arch submarine ridge, implying very 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 destabilization and collapse in constraining the morphological and structural evolution of marine volcanoes. Any chosen point in the life cycle of such an edifice represents a 'snapshot' 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 restabilize 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) reports turbidite currents related to Hawaiian landslides, which travelled over 1000 km and flowed over sea-bed obstructions 500 m high. Similarly, Weaver *et al.* (1994) present evidence for turbidity currents over 600 km in length that appear to be related to slope failure on the flanks of the westernmost Canary Islands of La Palma and El Hierro around 18 ka BP (Fig. 3). More recently, Wynn & Masson (2003) interpret stacked turbidite subunits in the region in terms of prehistoric lateral collapses on the islands of El Hierro and Tenerife. Furthermore, they propose that each member of the stacks is representative of a single landslide occurring within a multi-stage lateral collapse event. This interpretation minimizes the tsunami threat from such events, through favouring the formation of multiple small collapses rather than a single major landslide. This critical issue is addressed further below.

Tsunamigenic potential

There is increasing evidence in the geological and geomorphological records of lateral collapses at ocean-island volcanoes generating large tsunamis. Due to an often greater vertical drop and to the high velocities achieved, the tsunami-producing potential of a large body of debris *entering* the sea is much greater than that of a similar-sized submarine landslide, and even small sub-aerial volcanic landslides can locally generate locally highly destructive waves if they enter a large body of water. In 1792 at Mount Unzen (Japan), for example, a landslide with a volume of only *c.* 0.33×10^9 m^3 – which was not connected with an eruption – triggered a tsunami that caused more than 14 500 deaths. In 1888, several hundred deaths are thought to have resulted from the lateral collapse of the Ritter Island volcano (Papua New Guinea), which generated tsunamis with wave run-up heights as great as 12–15 m (Johnson 1987; Ward & Day 2003). Ritter Island is the largest known historical lateral collapse at a volcano, involving around 5 km^3 of material. Most recently, in 2002, the two small landslides (total volume *c.* 5.6×10^6 km^3) triggered within the prehistoric Sciara del Fuoco lateral collapse scar on the island volcano of Stromboli, generated a tsunami with a 10 m run-up that caused substantial damage along the east coast of the island (Bonaccorso *et al.* 2003).

Tsunamis associated with *giant* collapses at oceanic-island volcanoes, however, can have run-up heights an order of magnitude greater. For example, a wave train associated with collapse of part of Mauna Loa (Hawaii) – the so-called Alika 2 Slide – around 120 000 a BP has been implicated in the emplacement of coral and other debris to an altitude of over 400 m above contemporary sea-level and more than 6 km inland on the neighbouring island of Kohala (McMurtry *et al.* 2004). Giant waves generated by ancient collapses in the Hawaiian Islands may have been of Pacific-wide extent. Young & Bryant (1992) explained signs of catastrophic wave erosion up to 15 m above current sea-level along the New South Wales coast of Australia – 14 000 km distant – in terms of impact by tsunamis associated with a major collapse in the archipelago around 1.05×10^5 a BP. These phenomena have also, however, been interpreted in terms of a tsunami generated by a comet or asteroid impact in the ocean.

Putative giant-tsunami deposits have been identified at increasing numbers of locations. On Gran Canaria in the Canary Islands, for example, deposits consisting of volcanic clasts and broken marine shell and coral debris (Fig. 5) have been recognized at elevations of up to 188 m above present sea-level (Pérez-Torrado *et al.* 2006), while similarly elevated deposits occur on the neighbouring island of Fuerteventura (S. J. Day. pers. comm.). Pérez-Torrado *et al.* (2006) propose that the Gran Canaria deposits were emplaced by tsunamis caused by flank failure of another island in the archipelago, and tentatively suggest the 30 km^3 Gûimar lateral collapse (age $< 8.3 \times 10^5$ a) on neighbouring Tenerife as a possible source. Large coral boulders weighing up to 2000 tonnes on the Rangiroa reef (French Polynesia) have been linked by Talandier & Bourrouilh-le-Jan (1988) with giant tsunamis formed by the early nineteenth-century collapse of the Fatu Hiva volcano (Marquesas Islands, SE Pacific). Coral boulders, of up to 10^3 m^3, scattered along the northeast coast of the Bahamian island of Eleuthera, together with associated geomorphological features (Hearty 1997; Hearty *et al.* 1998), may provide evidence of giant tsunamis from a major collapse at the Canary Island of El Hierro.

The La Palma case

With respect to future lateral collapses at ocean islands, considerable attention has been focused on the Cumbre Vieja volcano (La Palma, Canary Islands) (Fig. 6) and its tsunamigenic potential in relation to expected failure of its western flank.

Fig. 5. Deposits of rounded cobbles and shell debris attached to the walls of the Agaete Valley on the northwest coast of Gran Canaria (Canary Islands) at heights ranging from 41 to 188 m above current sea-level. These are interpreted as tsunami deposits by Pérez-Torrado *et al.* (2006), and tentatively linked to the prehistoric Gûimar lateral collapse on the neighbouring island of Tenerife.

Fig. 6. Aerial view of the western flank of the Cumbre Vieja volcano, La Palma (Canary Islands).

This interest arises from events that took place during the volcano's penultimate eruption in 1949 (Bonelli Rubio 1950), when – in association with intense seismicity – part of the flank dropped 4 m (maximum) along a 3 km west-facing system of normal faults that opened along the crest of the volcano (Fig. 7). The faults are interpreted by Day *et al.* (1999*a*) as the first surface ruptures produced by a developing detachment fault or fault zone beneath the western flank of the edifice. The ability of past collapses in the Hawaiian and Canarian archipelagos to generate tsunamis in excess of 150 m high (McMurtry *et al.* 2004; Pérez-Torrado *et al.* 2006), along with evidence from the Ritter Island slide (Ward & Day 2003), suggest that flank failure of Cumbre Vieja may occur catastrophically. A rudimentary geodetic monitoring campaign undertaken between 1994 and 1997 (Moss *et al.* 1999) further suggests, rather inconclusively, that the western flank may currently be moving at around 1 cm a^{-1}. While forecasts of the timing of translation of this creep into catastrophic failure are entirely unconstrained, this is most likely to occur during a future eruption when elevated levels of seismic shaking, the pressure of intruded magma and additional impetus provided by magma-heated groundwater, will provide optimum conditions for collapse. This can only happen, however, once sufficient strain has accumulated along the future slide plane. The Ritter Island collapse involved transport velocities of 45 m s^{-1} and perhaps as high as 80 m s^{-1} (Ward & Day 2003), and Ward & Day (2001) proposed an entry velocity as high as 100 m s^{-1} for a future major collapse of the Cumbre Vieja. There continues, however, to be serious disagreement about the nature of a future collapse of the Cumbre Vieja and its tsunamigenic potential (e.g. Masson *et al.* 2006). Areas of dispute centre on the volume of material available for collapse, the slide velocity of the collapsing mass, the extent to which the collapse is retrogressive rather than a single slide, and the ability of any resulting tsunamis to retain sufficient energy to be destructive along coastlines around the North Atlantic Basin.

Fig. 7. Part of a 3-km-long west-facing system of normal faults that opened along the crest of the Cumbre Vieja volcano during the 1949 eruption. Maximum vertical displacement is 4 m. West is to the left of the picture.

Discussion concerning the volume of material available concentrates, in particular, on how this is constrained by the thickness of the slide. Urgules *et al.* (1997) and Hürliman *et al.* (2004), based on pre-collapse reconstructions of, respectively, the El Golfo (on El Hierro) and Orotava (on Tenerife) collapses, arrive at lower maximum thicknesses of 1–1.5 km. Assuming the maximum length (25 km) and width (15 km) dimensions used by Ward & Day (2001) in their worst-case model, this would result in a maximum available volume for a future Cumbre Vieja west flank collapse of less than 300 km^3. These assumptions are made on the basis, however, that in each case the summit of the failed volcano coincides with the line of the collapse head-scarp. As demonstrated by Mount St. Helens in 1980, volcano lateral collapses typically retrogress significantly into the opposing flank. For both the El Golfo and Orotava collapses, therefore, the original summits are likely to have been considerably higher than the altitude of the collapse scar rim, resulting in a greater depth between the summit and the basal slide plane. Furthermore, reconstructions of pre-collapse edifices on Fuerteventura (Stillman 1999) and Fogo (Cape Verde Islands) (Day *et al.* 1999*b*) and of the Cumbre Nueva on La Palma itself (Day *et al.* 1999*a*), all suggest initial thicknesses of the collapsing mass of 2–3 km beneath the summit. From this, Ward & Day (2001) derive a worst-case collapse with a volume of *c.* 500 km^3. An intact lateral collapse block with dimensions (22×11 km) comparable to those used by Ward & Day (2001) in their

Cumbre Vieja tsunami model, has been identified by Acosta *et al.* (2003) adjacent to the west coast of the Canary Island of Fuerteventura (Fig. 8).

In relation to slide velocity, there is considerable evidence – both observational and in the geological record – for catastrophic failure and high transport velocities being the norm for volcano lateral collapse. In May 1980, failure of the north flank of Mount St Helens (Washington State, USA) occurred in less than a minute, on a slide plane of *c.* 10° (comparable with slide-plane slopes on the submarine flanks of the Canary Islands), resulting in a landslide achieving velocities in excess of 80 m s^{-1} (Lipman & Mullineaux 1981). Similarly, large, non-volcanic, sub-aerial landslides (*sturzstroms*), such as that responsible for the 1963 Vajont (NE Italy) reservoir disaster (Kilburn & Petley 2003), are also catastrophic phenomena. Ward & Day (2003) estimate that bulk average peak velocities during the 1888 Ritter Island collapse reached *c.* 45 m s^{-1} after an initial drop of the centre of mass of just 700 m, and that velocities as high as 80 m s^{-1} may have been achieved. Scaling up in terms of the fall height and other parameters, Ward & Day (2003) propose that the Ritter Island observations and modelling support velocities in excess of 100 m s^{-1} during a future collapse of the Cumbre Vieja. Such velocities are reasonably consistent with, for example, the transport of the 5 000 km^3 Nuuanu landslide (Oahu, Hawaii), which must have been travelling at more than 80 m s^{-1} in order to have sufficient energy to climb 350 m up the far slope of the Hawaiian Arch submarine ridge (Ward 2001). In the Canary Island archipelago itself, Day *et al.* (1997) also hold up the occurrence of pseudotachylite (friction-melted rock) on the 300 m slip surface of an aborted prehistoric landslide on the island of El Hierro, as evidence of exceptionally high transport velocities. Most conclusively, the tsunami deposits recently identified in Hawaii and on Gran Canaria, at elevations in excess of 150 m (McMurtry *et al.* 2004; Pérez-Torrado *et al.* 2006), must have required gigantic waves for their emplacement, waves that can only have resulted from very high entry velocities.

Notwithstanding such evidence for the achievement of exceptional velocities, Wynn & Masson (2003) and Masson *et al.* (2006) argue

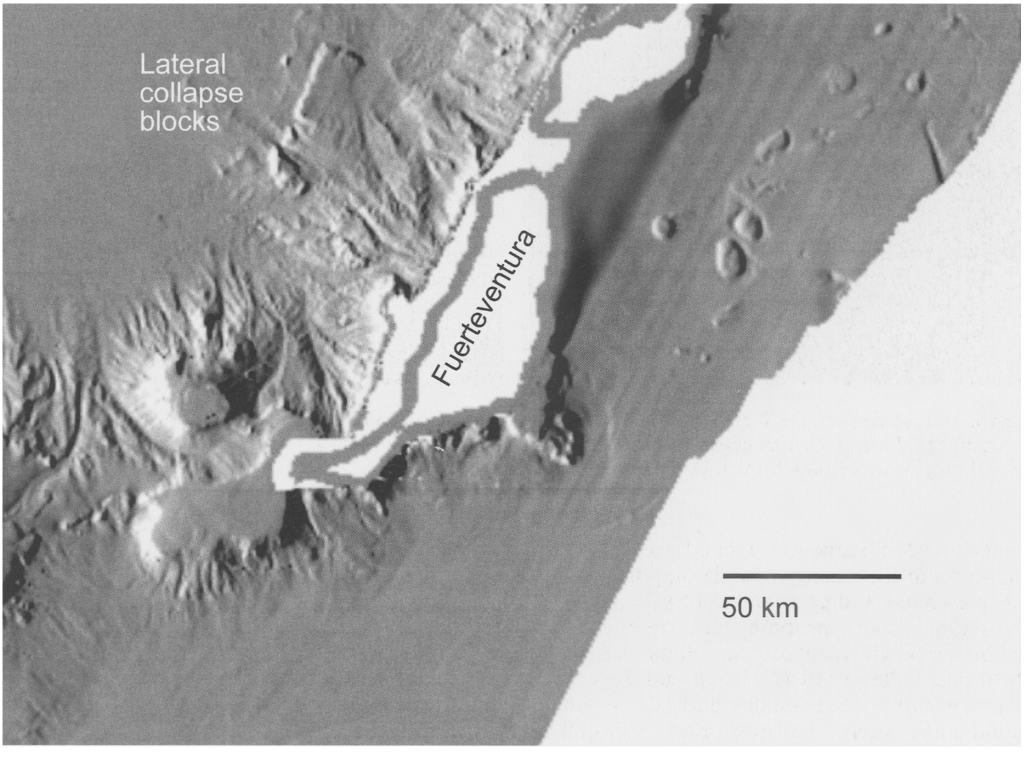

Fig 8. Gigantic blocks, one 22 × 11 km in size, emplaced to the west of the eastern Canary Island of Fuerteventura as a result of the lateral collapse of the Northern and Central Volcanic Complexes in Miocene to Pliocene times. Image: courtesy of Juan Acosta (Acosta *et al.* 2003).

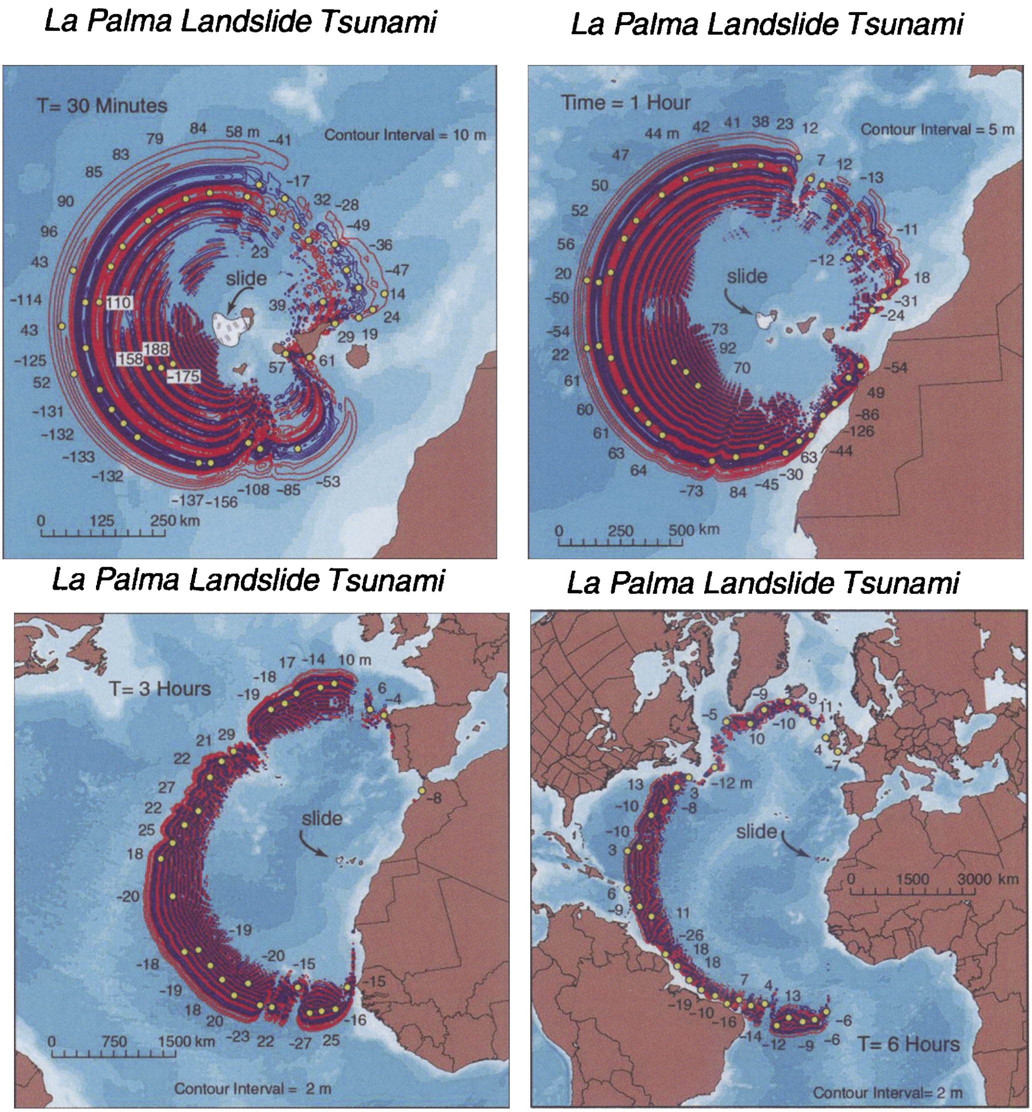

Fig. 9. Time-slices from the Ward & Day (2001) Cumbre Vieja tsunami model, showing the location of the wave train 30 min, 1 h and 6 h after collapse. Positive numbers = wave-crest heights (m); negative numbers = wave-trough heights (m). From Ward & Day (2001).

that turbidite sequences related to prehistoric collapses in the Canary Islands support a model of piecemeal rather than wholesale landslide formation. They propose that each member within stacked turbidite sub-units associated with past collapses at El Hierro and Tenerife is representative of a single landslide within a multi-stage lateral collapse. Each sub-unit is internally graded, but there is no intervening pelagic sediment between sub-units, which would indicate an extended period of time between the emplacement of successive units. Clearly, the smaller volumes that would be involved in such an eventuality would minimize the associated tsunami threat. There are, however, a number of problems with the proposition that each turbidite sub-unit represents a phase of collapse. In particular, several mechanisms exist by which multiple turbidity currents – or multiple pulses within a single current – may be generated by a single submarine landslide. Studies of the submarine landslide deposits generated by the 1888 Ritter Island collapse, for example, reveal the formation of large numbers of discrete debris-flow lobes, each with the potential to transform into a turbidity current (Simon Day, pers. comm.). Similar behaviour emerges from a granular flow computer model for a future collapse of Cumbre Vieja. The simulations show distal elements of the landslide forming discrete lobes, which follow distinct paths that overlap and interact at distances in excess of 200 km from the source (Steve Ward, pers. comm.). Such a situation would seem highly appropriate to producing a stack of turbidite units, closely associated in time and therefore without intervening pelagic sediment.

The major area of dispute concerns the ability of any collapse-driven tsunamis to retain sufficient energy to be destructive at oceanic distances. Ward & Day (2001) modelled the consequences for a worst-case scenario (a 500-km^3 slide block moving at 100 m s^{-1}) (Fig. 9), predicting the formation of an initial dome of water 900 m in height that subsides to a series of waves hundreds of metres high. For collapse scenarios involving a range of slide blocks from 150–500 km^3, Ward & Day (2001) predict a wave train that transits the entire Atlantic Ocean with wave heights along the coast of the Americas ranging from 3–8 m (for a 1.5×10^2 km^3 slide) to 10–25 m (for a 5×10^2 km^3 slide). Mader (2001), however, proposes that wave heights along the east coast of the US would be less than 3 m and lower for smaller collapse volumes, while Paras-Carayannis (2002) predicts that they will be as low as 1 m.

The opposing viewpoints centre on the efficiency with which large trans-oceanic tsunamis can be generated by what is essentially a point-source (rather than the linear source typical of the biggest seismogenic tsunamis). Sub-aerial and submarine landslides generate waves that have characteristically short wavelengths and short periods (the time taken for a tsunami wave to complete a full cycle). These can be extremely destructive locally, but their damage potential at distances of thousands of kilometres is questioned. Based upon calculations for the tsunami potential of a Cumbre Vieja collapse, Gisler *et al.* (2005) used the SAGE hydrocode to show that both the periods and wavelengths of the waves produced are short compared to tele-tsunamis observed in the Indian and Pacific Oceans. While recognizing that a future lateral collapse could be expected to generate waves capable of being highly destructive within the Canary Island archipelago itself, and along the coasts of Morocco, Spain and Portugal, Gisler *et al.* (2005) suggest that the absence of long-period–long-wavelength waves will probably not permit the propagation of destructive waves to the east coast of North America. It may be, however, that Gisler *et al.* (2005) cannot rule out transoceanic mega-tsunamis until they have propagated the waves for more than the few hundred kilometres used in their simulations (Simon Day, pers. comm.).

Focusing on the only major tsunamigenic collapse for which data are available, Ward & Day (2003) highlight the formation and importance of short- (1–5 minutes) period tsunamis during the Ritter Island collapse, and their ability to transmit damaging and potentially lethal energy to distances of several hundred kilometres. They argue that a scaled-up Ritter Island collapse – in terms of fall height and other parameters *would* be capable of generating a transoceanic mega-tsunami (Fig. 10). Support

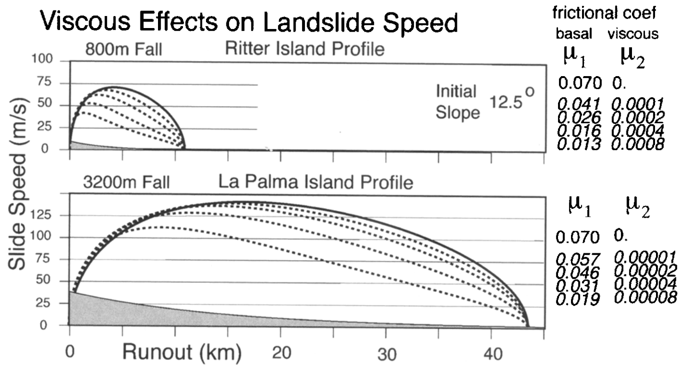

Fig. 10. Velocity trajectories of Ritter Island (top) and La Palma (bottom) landslides with ($\mu_2 > 0$, dashed lines) and without ($\mu_2 = 0$, solid line) viscous dissipation. Uppermost dashed lines represent the most plausible μ_2 values. The grey area is the slope profile. Ward & Day (2003) propose that giant landslides, such as that threatened by the instability of the Cumbre Vieja's west flank, should run at speeds exceeding 100 m s^{-1}, even in the presence of viscous dissipation. From Ward & Day (2003).

for the persistence of short-period waves across oceanic distances comes from observations of tsunamis generated by nuclear tests at Bikini Atoll in the late 1950s and early 1960s. Van Dorn (1961) reports the measurement of 3–9-minute waves (compared to 10 min in the Ward & Day 2001 Cumbre Vieja model) out to a distance of 2 700 km. The wave decay rate described by Van Dorn is also in agreement with that proposed by Ward & Day (2001).

Conclusions

Lateral collapse at marine volcanoes is commonplace in the geological record, and is certain to occur again. Locally, such collapses have the potential to trigger major tsunamis with run-ups in excess of 150 m, and with enormous destructive and lethal capability. The capacity for large, lateral collapses to generate tsunamis with sufficient energy to be destructive at oceanic distances remains to be incontrovertibly established. Nevertheless, it is a threat that must be considered. In addition to possible future tsunami generation from collapse of La Palma's Cumbre Vieja volcano, Day *et al.* (1999*b*) also highlight the development of instability at Fogo volcano in the Cape Verde Islands, which may prove to be a future tsunami source. Elsewhere, Stromboli maintains the potential for further tsunamigenic behaviour, while Begét & Kienle (1992) expect another collapse-related tsunami at Augustine volcano (Alaska), comparable to an event in 1883 that generated 20-m waves.

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