Earth-Science Reviews 57 (2002) 1–35

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Slope failures on the flanks of the western Canary Islands

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Received 7 June 2000; accepted 17 May 2001

Abstract

Landslides have been a key process in the evolution of the western Canary Islands. The younger and more volcanically active Canary Islands, El Hierro, La Palma and Tenerife, show the clearest evidence of recent landslide activity. The evidence includes landslide scars on the island flanks, debris deposits on the lower island slopes, and volcaniclastic turbidites on the floor of the adjacent ocean basins. At least 14 large landslides have occurred on the flanks of the El Hierro, La Palma and Tenerife, the majority of these in the last 1 million years, with the youngest, on the northwest flank of El Hierro, as recent as 15 thousand years in age. Older landslides undoubtedly occurred, but are difficult to quantify because the evidence is buried beneath younger volcanic rocks and sediments. Landslides on the Canary Island flanks can be categorised as debris avalanches, slumps or debris flows. Debris avalanches are long runout catastrophic failures which typically affect only the superficial part of the island volcanic sequence, up to a maximum thickness of 1 to 2 km. They are the commonest type of landslide mapped. In contrast, slumps move short distances and are deep-rooted landslides which may affect the entire thickness of the volcanic edifice. Debris flows are defined as landslides which primarily affect the sedimentary cover of the submarine island flanks. Some landslides are complex events involving more than one of the above end-member processes.

Individual debris avalanches have volumes in the range of 50–500 km^3, cover several thousand km^2 of seafloor, and have runout distances of up to 130 km from source. Overall, debris avalanche deposits account for about 10% of the total volcanic edifices of the small, relatively young islands of El Hierro and La Palma. Some parameters, such as deposit volumes and landslide ages, are difficult to quantify. The key characteristics of debris avalanches include a relatively narrow headwall and chute above 3000 m water depth on the island flanks, broadening into a depositional lobe below 3000 m. Debris avalanche deposits have a typically blocky morphology, with individual blocks up to a kilometre or more in diameter. However, considerable variation exists between different avalanche deposits. At one extreme, the El Golfo debris avalanche on El Hierro has few large blocks scattered randomly across the avalanche surface. At the other, Icod on the north flank of Tenerife has much more numerous but smaller blocks over most of its surface, with a few very large blocks confined to the margins of the deposit. Icod also exhibits flow structures (longitudinal shears and pressure ridges) that are absent in El Golfo. The primary controls on the block structure and distribution are inferred to be related to the nature of the landslide material and to flow processes. Observations in experimental debris flows show that the differences between the El Golfo

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PII: S0012-8252(01)00069-1

and Icod landslide deposits are probably controlled by the greater proportion of fine grained material in the Icod landslide. This, in turn, relates to the nature of the failed volcanic rocks, which are almost entirely basalt on El Hierro but include a much greater proportion of pyroclastic deposits on Tenerife.

Landslide occurrence appears to be primarily controlled by the locations of volcanic rift zones on the islands, with landslides propagating perpendicular to the rift orientation. However, this does not explain the uneven distribution of landslides on some islands which seems to indicate that unstable flanks are a 'weakness' that can be carried forward during island development. This may occur because certain island flanks are steeper, extend to greater water depths or are less buttressed by the surrounding topography, and because volcanic production following a landslide my be concentrated in the landslide scar, thus focussing subsequent landslide potential in this area. Landslides are primarily a result of volcanic construction to a point where the mass of volcanic products fails under its own weight. Although the actual triggering factors are poorly understood, they may include or be influenced by dyke intrusion, pore pressure changes related to intrusion, seismicity or sealevel/climate changes. A possible relationship between caldera collapse and landsliding on Tenerife is not, in our interpretation, supported by the available evidence.

Keywords: Canary Islands; Landslides; Debris avalanches; Slumps; Debris flows

1. Introduction

It is now firmly established that large-scale landsliding is a key processes in the evolution of oceanic islands. Detailed studies of landslides have been carried out around the Hawaiian Islands (Lipman et al., 1988; Moore et al., 1989, 1994), Reunion (Cochonat et al., 1990; Labazuy, 1996; Ollier et al., 1998) and the Canary Islands (Holcomb and Searle, 1991; Krastel et al., 2001; Masson, 1996; Masson et al., 1998; Teide Group, 1997; Urgeles et al., 1997, 1999; Watts and Masson, 1995). Some of the clearest evidence for landsliding, in the form of large fields of blocky landslide deposits, has been reported offshore. Landslide deposits can be transported several hundred kilometres and cover many hundreds of km^2 of seafloor on the submarine island flanks. Individual landslides can involve up to a few thousand km^3 of material, but more typically are a few hundred km^3 in volume. Onshore, landslide headwalls are typically expressed as arcuate embayments and steep cliffs (Cantagrel et al., 1999; Navarro and Coello, 1989; Ollier et al., 1998; Ridley, 1971).

In the Canary Islands, the relatively recent discovery of landslide deposits offshore (Holcomb and Searle, 1991; Masson, 1996; Watts and Masson,

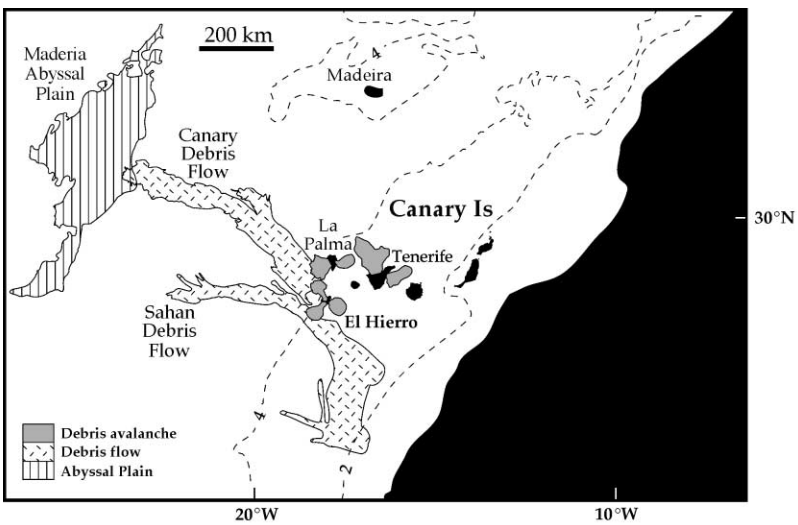

Fig. 1. Map showing location of the Canary Islands and areas where landslide deposits have been mapped.

1995) confirms earlier controversial interpretations based on the onshore geology (Bravo, 1962; Navarro and Coello, 1989). Prior to the study presented here, all the offshore studies have concentrated on areas of island flank downslope of suspected subaerial landslide scars, in particular the Orotava, Icod and Guimar valleys on Tenerife, the Taburiente Caldera/Cumbre Nueva Arc on La Palma, and the El Golfo embayment on El Hierro (Holcomb and Searle, 1991; Masson, 1996; Masson et al., 1998; Teide Group, 1997; Urgeles et al., 1997, 1999; Watts and Masson, 1995). Here we present the results of a more comprehensive study of flank collapse processes on Tenerife, La Palmas and El Hierro. The paper is partly a review and summary of previously published data, but also draws on a considerable volume of new

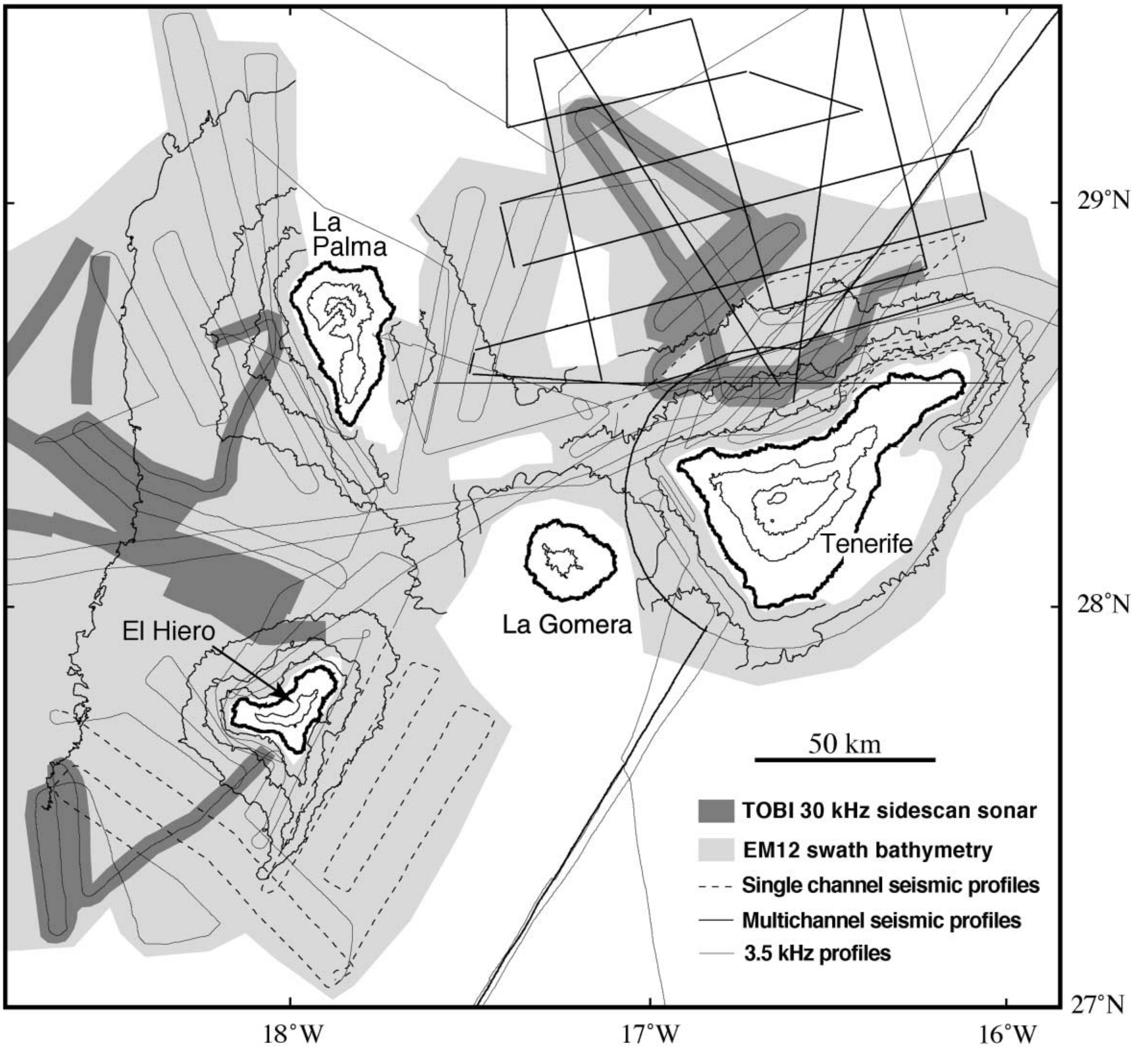

Fig. 2. Summary of data coverage around the islands of El Hierro, La Palma and Tenerife. EM12 swath bathymetry data was collected on RRS Charles Darwin cruises 82 (1993) and 108 (1997) and on the 'Crescent-94' cruise of the BIO Hesperides (1994). TOBI sidescan sonar data was collected on RRS Discovery cruise 205 (1993) and RRS Charles Darwin cruise 108. Multichannel seismic reflection profiles were collected on RRS Charles Darwin cruises 82 and on a cruise of the Dutch research vessel Zirfaea (1994), and single channel data were collected on RRS Charles Darwin cruise 108, RRS Discovery cruise 188 (1989) and the 'Crescent-94' cruise of the BIO Hesperides. All cruises collected 3.5 kHz profiles.

material. Much of the discussion, particularly the section on flow processes, is based on a new comparison between landslides on the different islands.

A Simrad EM12 multibeam system was used to map the submarine morphology and backscatter characteristics of large areas of island flank. These data clearly distinguish between unfailed island slopes and those affected by landsliding processes. High resolution, deep-towed, sidescan sonar data, acquired with the TOBI 30 kHz system, was used to examine the surface structure of landslide deposits in greater detail, to gain a better understanding of landsliding processes. Our results show that landsliding on the flanks of the islands is more widespread than previously supposed, and that landslide processes are both variable and complex. At least 14 individual landslides have been identified.

1.1. Study area and data collection

The Canary Islands are a group of seven volcanic islands in the eastern Atlantic Ocean off the northwest African margin (Fig. 1). There is evidence for a general decrease in the age of the islands from east

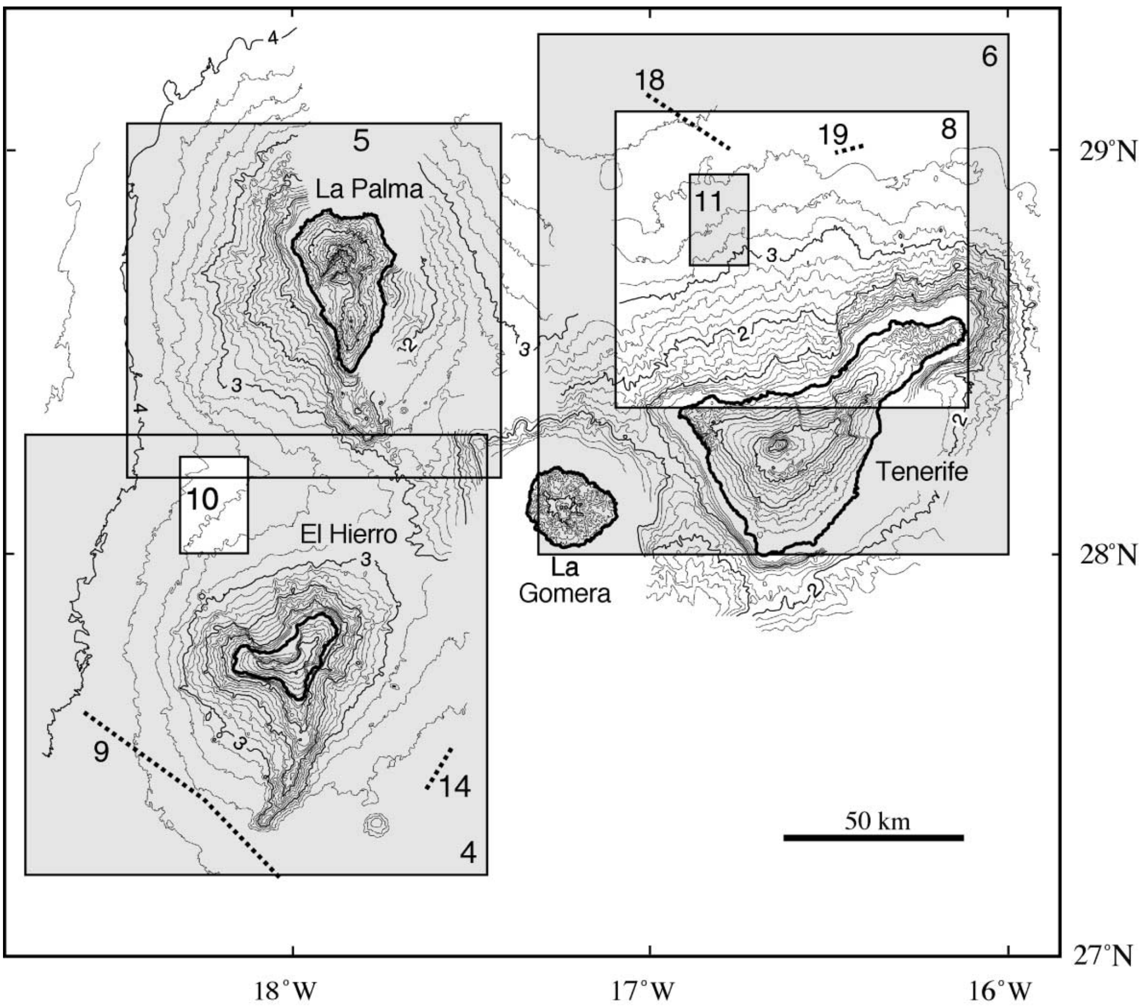

Fig. 3. Summary of swath bathymetry coverage around the islands of El Hierro, La Palma and Tenerife. Contour interval 100 m, with thousand metre contours shown by heavier lines. Land topography from maps published by the Spanish Geographical Survey. Figure locations shown by numbered boxes and heavy dashed lines.

to west, suggesting a hotspot origin for the island chain, although volcanic activity has occurred within historic times on all islands apart from La Gomera (Carracedo et al., 1998). El Hierro and La Palma are the most westerly and youngest of the Canary Islands, and with Tenerife appear to have been the most active, in terms of both volcanic and landslide activity, in the recent past (Urgeles et al., 1997). The most recent large landslide in the Canaries was probably the El Golfo failure on El Hierro, which occurred at about 15 ka (Masson et al., 1996).

The study area covers much of the submarine flanks of Tenerife, La Palma and El Hierro (Figs. 1–3). Complete seafloor coverage of the north flank of Tenerife, the west flank of La Palma and all around El Hierro up to water depths of around 4000 m was obtained using an EM12 multibeam system (Figs. 2 and 3). Less comprehensive surveys were carried out around the remainder of Tenerife and the eastern flank of La Palma. TOBI 30 kHz sidescan sonar images were obtained north of Tenerife, south of El Hierro and west of both La Palma and El Hierro (Fig. 2). 3.5 kHz profile data were recorded along all survey tracks. Seismic profiles consist of 12-channel sleeve-gun data collected north of Tenerife and 4-channel airgun data collected north of Tenerife, west of La Palma and both southeast and southwest of El Hierro. A complete list of the cruises from which data was used is given in the caption to Fig. 2.

2. Data processing and interpretation techniques

The EM12 swath mapping system collects both bathymetric and seafloor backscatter data. Bathymetric data was acquired using Simrad's Mermaid system and processed using the Neptune software. Gridded bathymetric data were combined with topographic data obtained from geographical maps published by the Spanish Geographical Survey. A final grid of bathymetry and topography was constructed

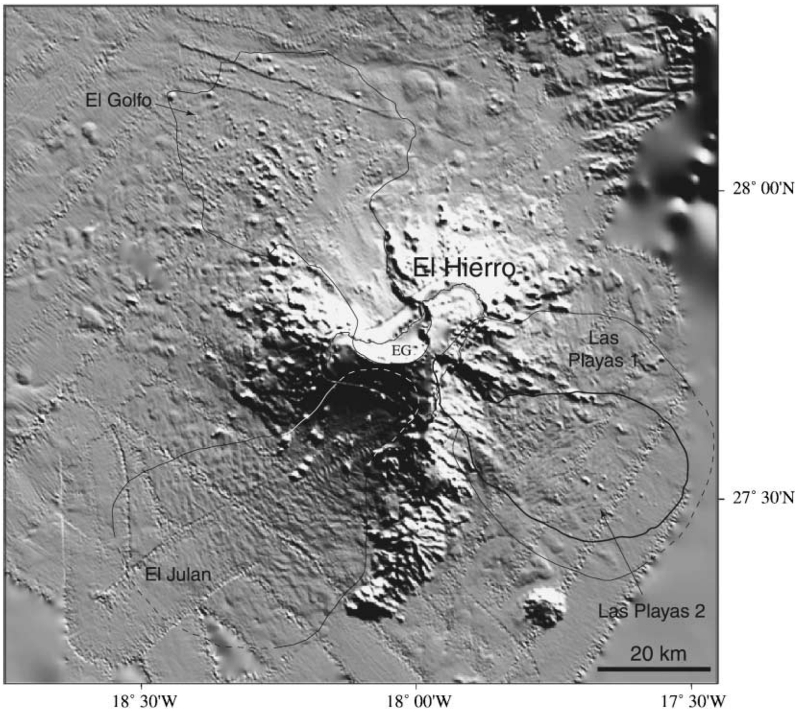

Fig. 4. Shaded relief image of El Hierro viewed from above. Shading reflects the intensity of the horizontal gradient in the direction of illumination (045°). Areas affected by landslides are outlined. Note strong topographic and textural contrasts between landslides and constructional volcanic areas. EG = El Golfo. For location see Fig. 3.

at 0.1×0.1 min intervals using GMT software (Wessel and Smith, 1991). The gridded data was used to produce contour maps and shaded relief maps (Figs. 3–6). TOBI 30 kHz data and EM12 backscatter data were processed using PRISM and ERDAS software for display both in simple map form and in a variety of combinations with the bathymetry data.

2.1. Recognition of landslides

The interpretation of a combination of EM12 bathymetry and backscatter data is the key to the identification of island flank areas which have been affected by landsliding. Although the scars left by the more recent failures (e.g., El Golfo embayment on El Hierro; Figs. 3 and 4) are immediately obvious on simple displays of bathymetric data, the topographic expression of older failure scars may be reduced by infilling by later lava flows (mainly onshore) or sediments (mainly offshore) and by the degradation of marginal scarps and other areas of rough topography generated by the landslide. For example, the Icod Valley on Tenerife is largely filled with late-stage volcanic products of Teide; the Anaga landslide scar offshore north Tenerife is masked by up to 100 m of sedimentary cover; the walls of the Orotava Valley on Tenerife are heavily gullied. In these older landslide areas, a variety of interpretation techniques (or a combination of techniques) may need to be applied to the bathymetry and backscatter data to discriminate between slopes subject to landsliding and adjacent more stable slopes.

For example, as we have previously demonstrated (Gee, 1999; Watts and Masson, 1995), landslides on the flanks of the western Canary Islands radically change the topographic profile of the areas of island flank where they occur (Fig. 7). This change in profile can still be recognised even when the superficial traces of the landslide have been buried by later events. Landslide valleys typically have smooth slope profiles with mean slope gradients that decrease

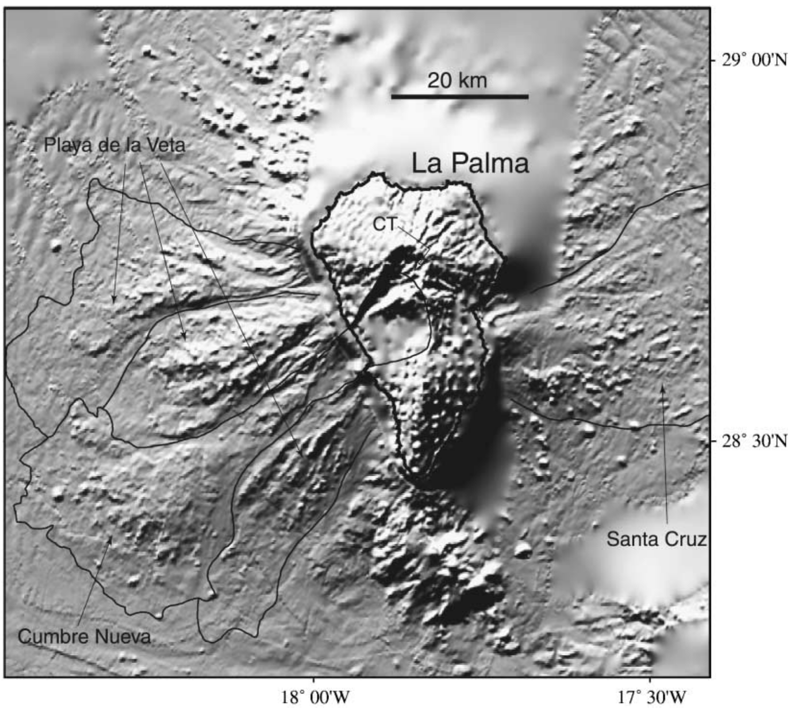

Fig. 5. Shaded relief image of La Palma viewed from above. Shading reflects the intensity of the horizontal gradient in the direction of illumination (340°). Areas affected by landslides are outlined. Note strong topographic and textural contrasts between landslides and constructional volcanic areas. CT = Caldera de Taburiente. For location see Fig. 3.

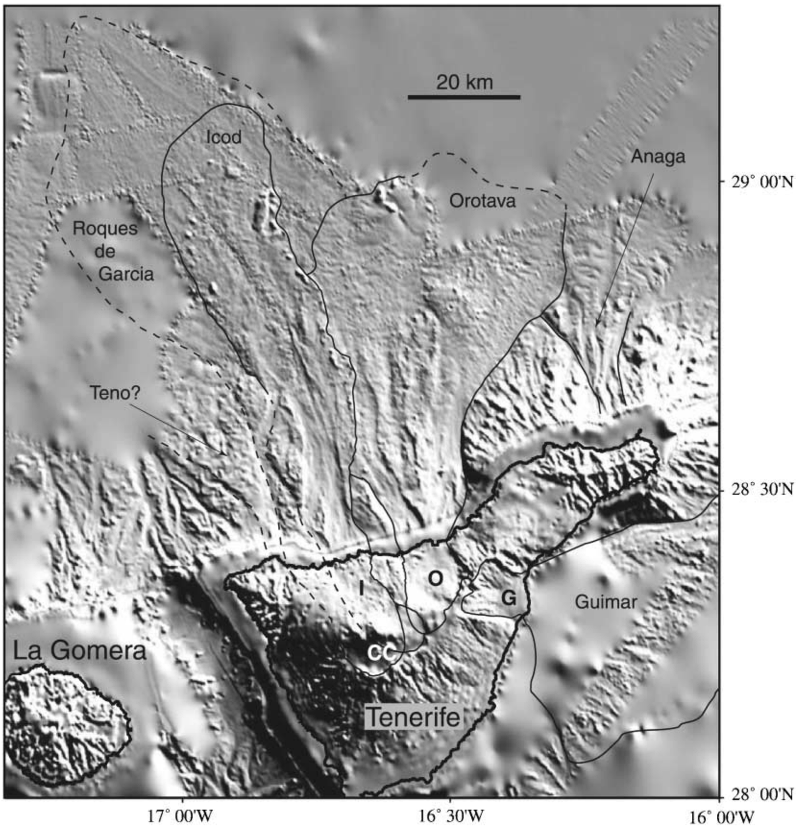

Fig. 6. Shaded relief image of Tenerife viewed from above. Shading reflects the intensity of the horizontal gradient in the direction of illumination (45°). Areas affected by landslides are outlined. Boundary of Guimar debris avalanche is from Krastel et al. (2001). Note strong topographic and textural contrasts between landslides and constructional volcanic areas. CC = Canadas Caldera, I = Icod Valley, O = Orotava Valley, G = Guimar Valley. For location see Fig. 3.

from about 10° on the upper slope to 5° at 3000 m (Fig. 7). The related landslide deposits usually occur in water depths greater than about 3000 m and have slope angles in the range 1–2°. Landslide deposits can also generate a distinct bathymetric bulge, typically 100–300 m high, at about the 3000 m bathymetric contour (e.g., off North Tenerife and western La Palma, Fig. 3). Flanks which appear not to have been affected by landsliding are more irregular and much steeper, with a wider range of measured slope gradients (typically 12° to > 30°) on the upper slope, which decrease very rapidly downslope at water depths > 3000 m (Fig. 7). The rugged terrain adjacent to landslide regions is particularly clearly seen on shaded relief images of island slopes (Figs. 4–6).

Variations in backscatter can also be used to distinguish between smooth sedimented seafloor and the rough blocky seafloor that typically characterises landslide deposits. Backscatter data derived from the 13 kHz Simrad EM12 echosounder is a particularly powerful tool in this respect (Watts and Masson, 1998), allowing us to recognise thin sheets of debris which have little bathymetric expression. Since backscatter levels are closely related to the degree of sediment cover overlying a landslide deposit, they can also be used to give a qualitative assessment of the relative ages of deposits, particularly where overlapping debris deposits occur (Fig. 8).

One problem in the definition of landslide boundaries is that different datasets may indicate different positions for a particular landslide boundary. For example, TOBI sidescan data often suggests that blocky landslide debris extends over a wider area than backscatter data derived from the EM12 system.

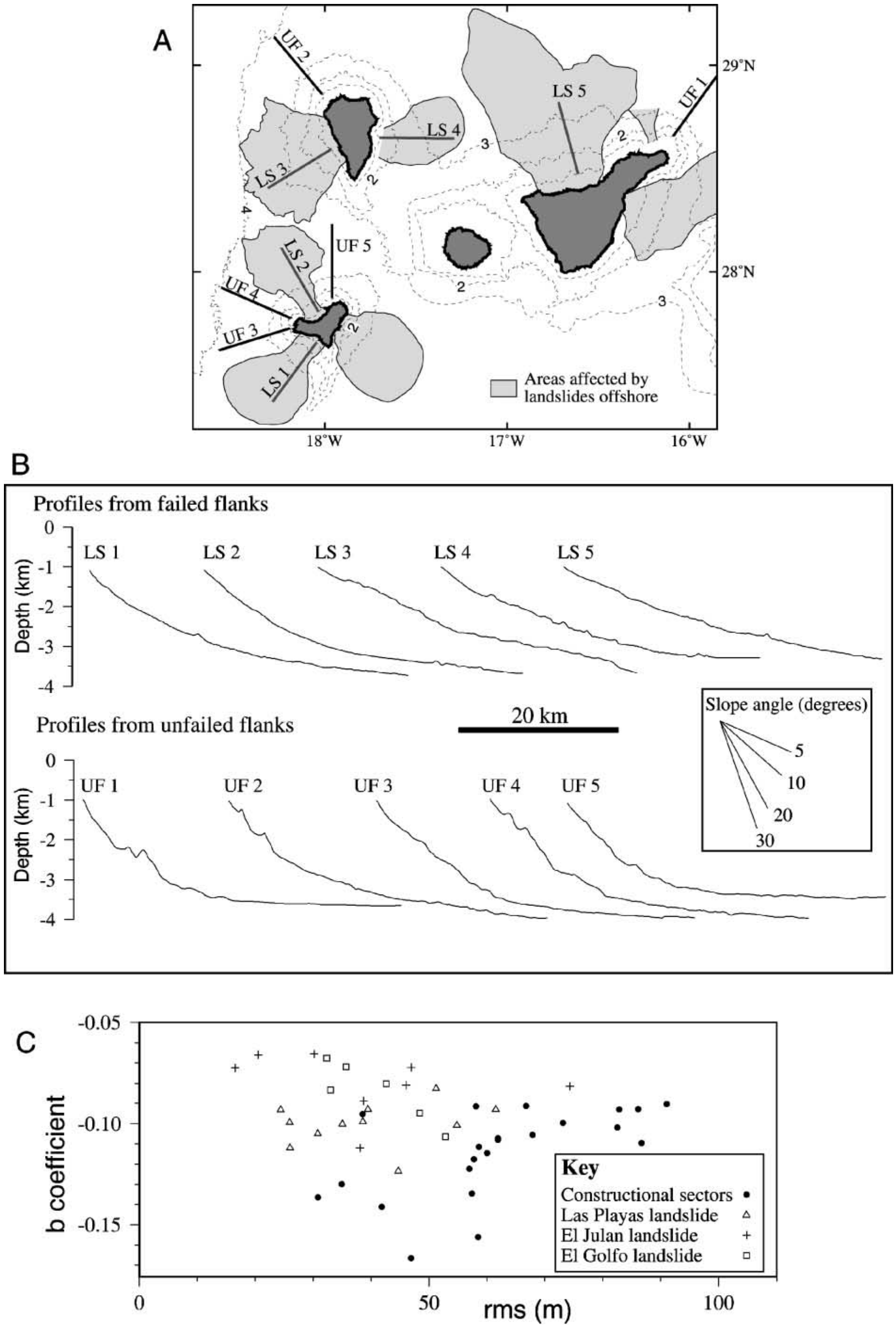

Fig. 7. Summary diagram showing slope gradients of failed and unfailed slopes. (A) Location of selected profiles from stable flank areas and from flanks affected by landsliding. (B) Comparative profiles from landslides and from stable flank areas. Profiles begin at a depth of 1000 m on the submarine island flanks because swath bathymetry data does not generally extend further landward. All profiles have a general exponential form, but those from landslides have overall lower gradients, are much less concave upwards, and are smoother than profiles from stable flanks. (C) Plot of the exponential coefficient (b) of an exponential curve fitted to each profile, against the rms residual of the profile relative to the exponential curve, for 48 radiating profiles around the island of El Hierro. The smoother landslide profiles have smaller negative exponential coefficients and smaller rms residuals relative to profiles from unfailed slopes.

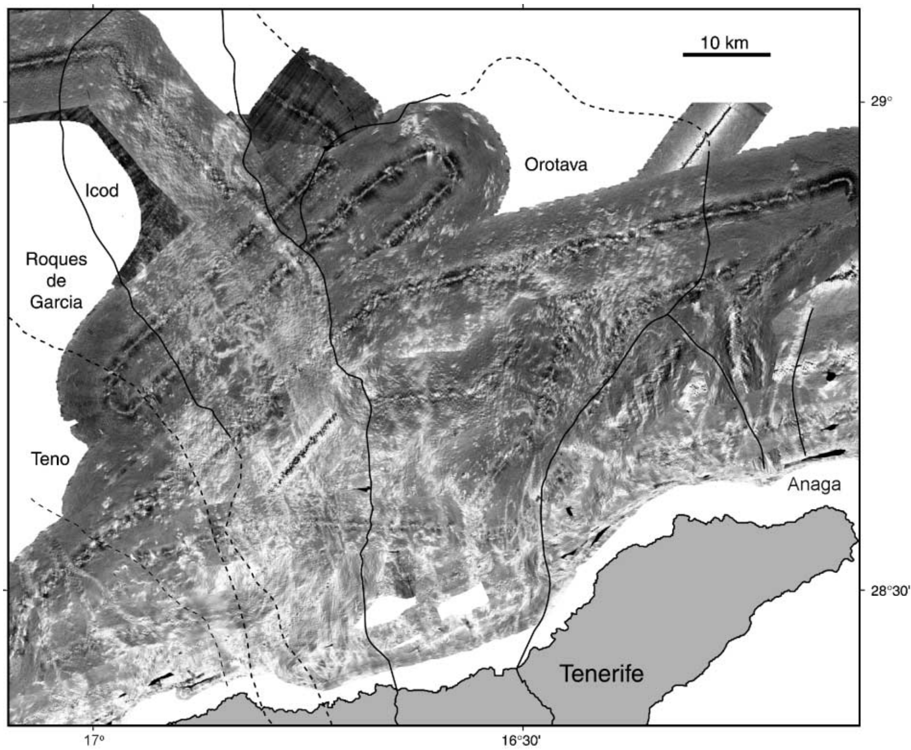

Fig. 8. Example of EM12 backscatter contrast in area of overlapping debris deposits on the north flank of Tenerife. Debris avalanche deposits (outlined) show distinctive speckled patterns which are most obvious on young landslides with little sediment cover (e.g. Icod). Older debris avalanches (e.g. Orotava) show patterns transitional between the speckled debris avalanche pattern and the smooth backscatter characteristic of sediment covered seafloor, reflecting avalanche deposits partially buried by later sedimentation. Stable island flanks show irregular backscatter patterns, within which some canyons and gullies can be recognised. For location see Fig. 3.

This appears to result primarily from the lower resolution of the EM12 system, with the result that small blocks are more difficult to recognise on EM12 backscatter data. Similarly, areas of chaotic debris facies seen on seismic profiles (e.g., Fig. 9) are often difficult to reconcile with backscatter data, sometimes because very thin debris sheets are poorly resolved by seismic profiling, at other times because partially buried debris deposits may have limited surface expression.

Debris flows made up of island flank sediments are most easily recognised on 3.5 kHz profiles, typically taking the form of lens-shaped bodies of characteristically acoustically transparent material (Gee et al., 1999; Masson et al., 1993). Flow patterns associated with debris flows are also frequently seen on sidescan sonar and EM12 backscatter data.

TOBI 30 kHz sidescan sonar data have been acquired over several landslide deposits and are used primarily to observe the detailed structure of these deposits (Figs. 10 and 11). Flow-deposit morphology and structure gives information on flow types and processes, and the assessment of sediment cover gives some qualitative information on the relative ages of flows.

2.2. Landslide processes / types

Landslides on the flanks of the Canary Islands are classified as ‘debris avalanches’, ‘slumps’, or ‘debris flows’. Debris avalanches and slumps (Moore et al., 1989) are used to describe landslides which cut into the volcanic and intrusive rocks of the island. Debris flows affect only the sedimentary cover of the submarine island slopes.

Each debris avalanche is believed to be the result of a single catastrophic failure, rapidly emplaced and producing a field of blocky rock debris spread over a

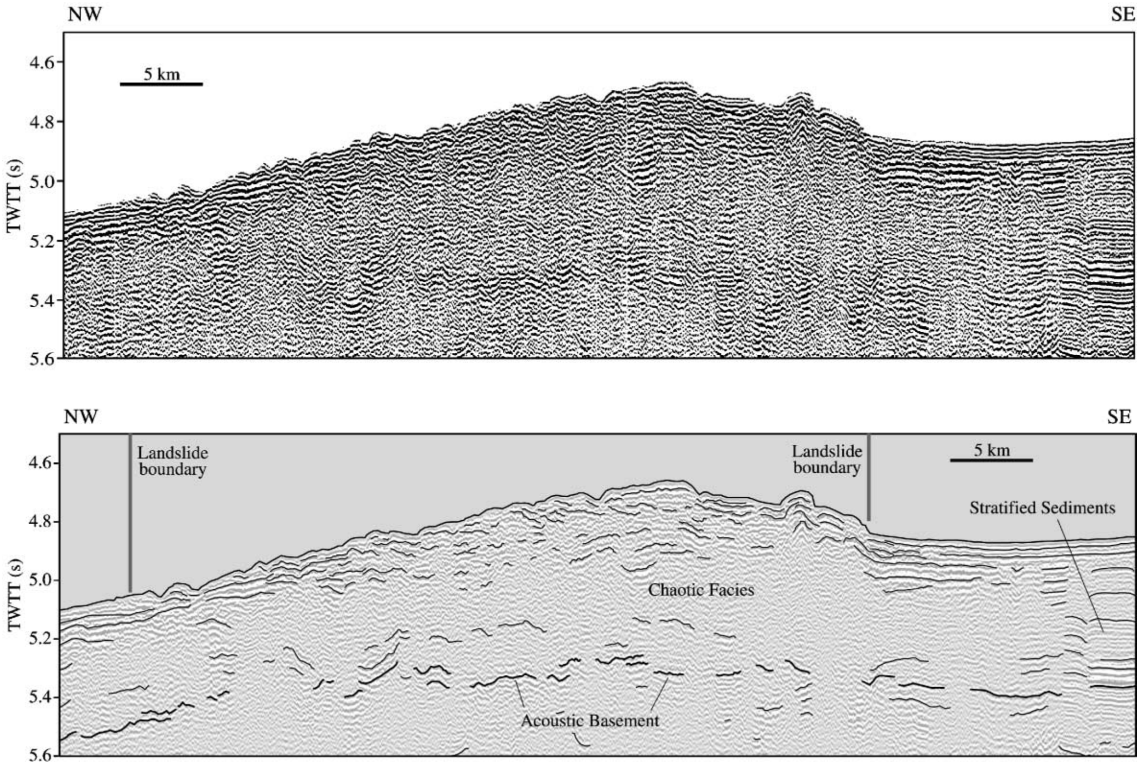

Fig. 9. Seismic profile across the El Julan landslide on the southwest flank of El Hierro. Note the chaotic seismic facies underlying the topographic bulge of the landslide, contrasting with the more strongly layered facies at the southeastern end of the profile. For location see Fig. 3.

wide area. The area affected by a debris avalanche tends to be relatively elongate, with the length (downslope) greater than the width. Although it is difficult to generalise about the thickness of the failed section, debris avalanches are usually associated with relatively superficial landslides, affecting sections a few hundred metres to perhaps 1 or 2 km thick (Moore et al., 1989). Around the Canary Islands, the resultant debris avalanche deposits are typically a few tens to a few hundred metres thick (Masson, 1996; Urgeles et al., 1997, 1999; Watts and Masson, 1995).

In comparison to debris avalanches, slumps can affect a much greater thickness within the volcanic island sequence (up to 10 km thick on the Hawaiian Ridge) and tend to be wide relative to their length (Moore et al., 1989). Slumps are thought to be slow-moving events, involving creep over an extended period of time. They involve a coherent mass of material, which typically deforms during slumping to produce a series of transverse ridges, scarps and folds on the surface of the slump (Moore et al., 1989). Possible slump-like activity has been reported from the southwestern and southeastern flanks of El Hierro in the Canary Islands (Gee, 1999). However, the lack of seismic data to image the deeper flank structures of the Canary Islands make recognition difficult.

Debris flow is used to describe landslides which primarily affect the sedimentary cover of the submarine slopes of the Canary Islands. It is recognised that the larger sedimentary landslides (e.g., the Canary and Saharan debris flows) probably include elements of rotational slumping and sliding, in addition to true debris flow (Masson et al., 1996). Some debris flows, for example the Canary Debris Flow on the western flank of La Palma and El Hierro (Fig. 12), can be related to debris avalanche events (Mas-

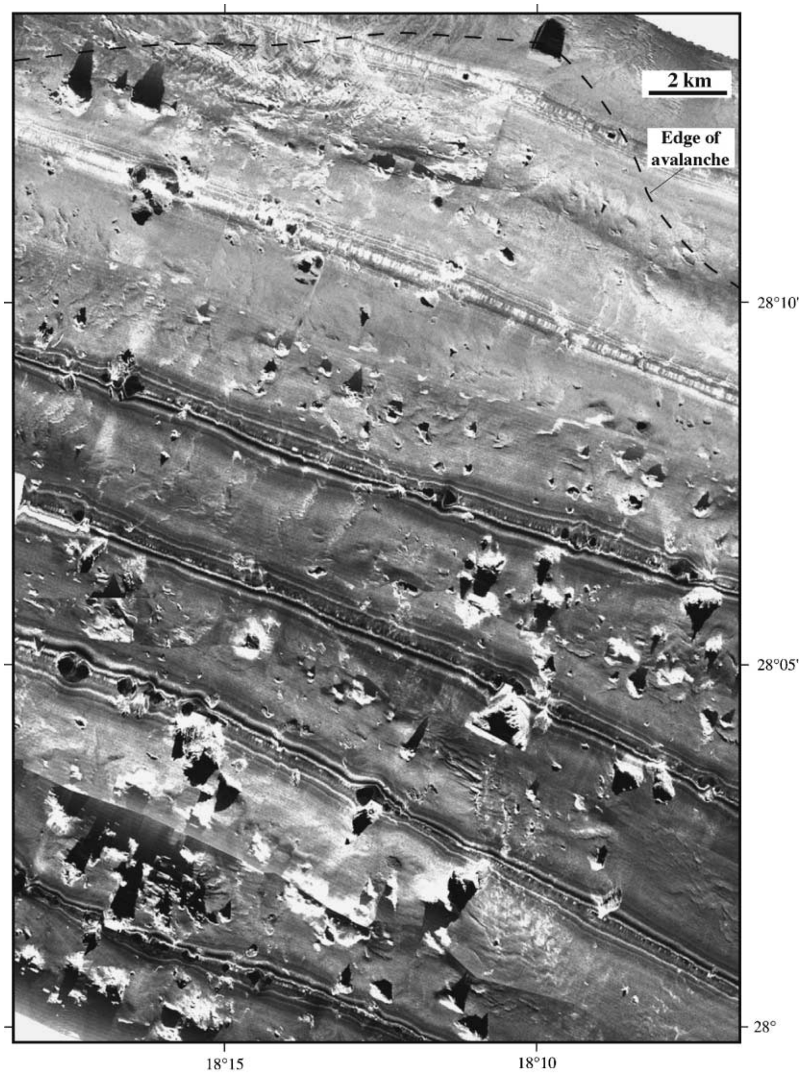

Fig. 10. TOBI 30 kHz sidescan sonar image of part of the El Golfo debris avalanche deposit northwest of El Hierro. High backscatter = light tones. Avalanche blocks are randomly scattered on the deposit surface and individual blocks are up to 1.2 km in diameter. Note the contrast in block size and distribution between the El Golfo and Icod avalanche deposits (see Fig. 11). For location see Fig. 3.

son, 1996; Masson et al., 1998; Urgeles et al., 1997). In contrast, the Saharan Debris Flow (Embley, 1976) originated on the Northwest African continental margin to the south of the Canaries, independent of any island flank landslide, but caused significant substrate erosion as it flowed over the lower flanks of Hierro (Gee et al., 1999).

Areas affected by landsliding may be complex, consisting of multiple overlapping debris avalanche events (Watts and Masson, 1995, 1998) or single events which evolve downslope from debris avalanche to debris flow (Masson, 1996; Masson et al., 1998; Urgeles et al., 1997).

3. Description of landslides

The following is a brief summary of landslides and landslide deposits that we have identified around

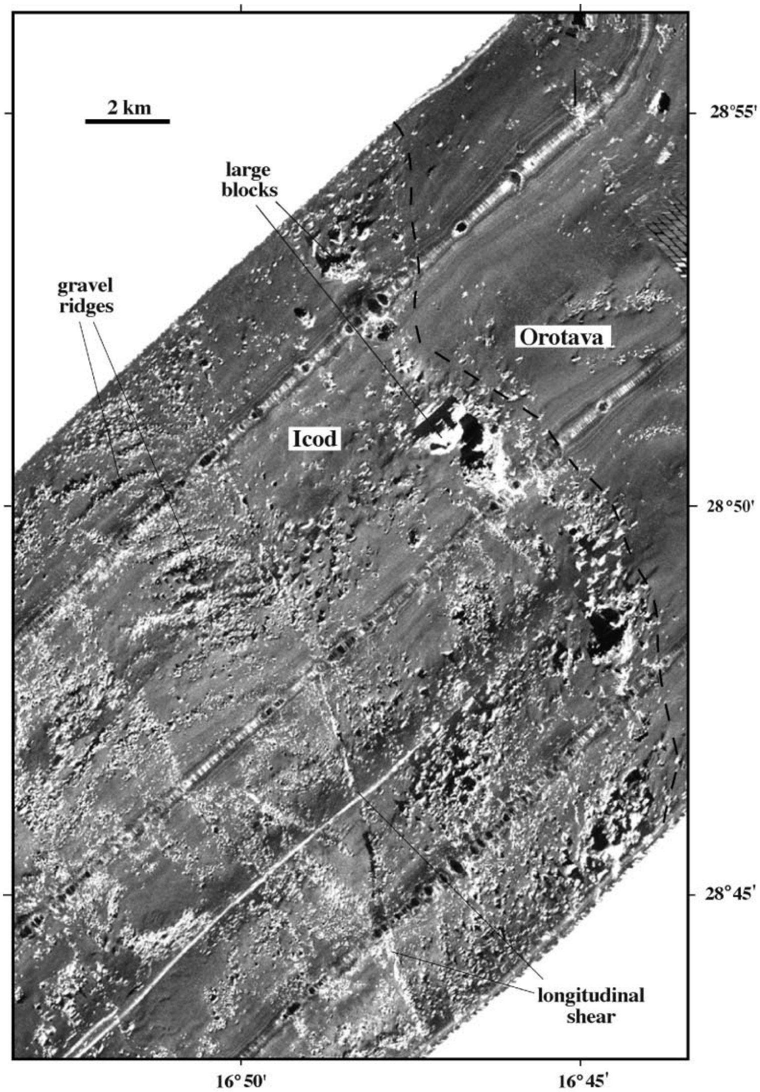

Fig. 11. TOBI 30 kHz sidescan sonar image of part of the Icod and Orotava debris avalanche deposits north of Tenerife. High backscatter = light tones. Avalanche blocks in the Icod deposit are more numerous and, in general, smaller than blocks in the El Golfo avalanche deposit (see Fig. 10). Large blocks are confined to the margins of the Icod deposit. Flow structures such as gravel ridges and longitudinal shears, absent in the El Golfo deposit, are also seen. The smoother surface of the Orotava avalanche deposit is due to a covering of sediment up to 20 m thick (see Fig. 19). For location see Fig. 3.

the western Canary Islands, grouped according to the island affected. For ease of reference, informal names are given to those landslides not already named (Fig. 12). Brief descriptions of landslide source areas and deposits are given, with comments on those aspects of their structure and morphology which give an insight into the processes involved in their formation. Landslide statistics are summarised in Table 1.

3.1. El Hierro

The subaerial structure of El Hierro has been interpreted in terms of three rift arms arranged at 120°, along which pronounced ridges have developed (Carracedo, 1996). The flanks between these three ridges are characterised by three large embayments which extend offshore and are interpreted as being of

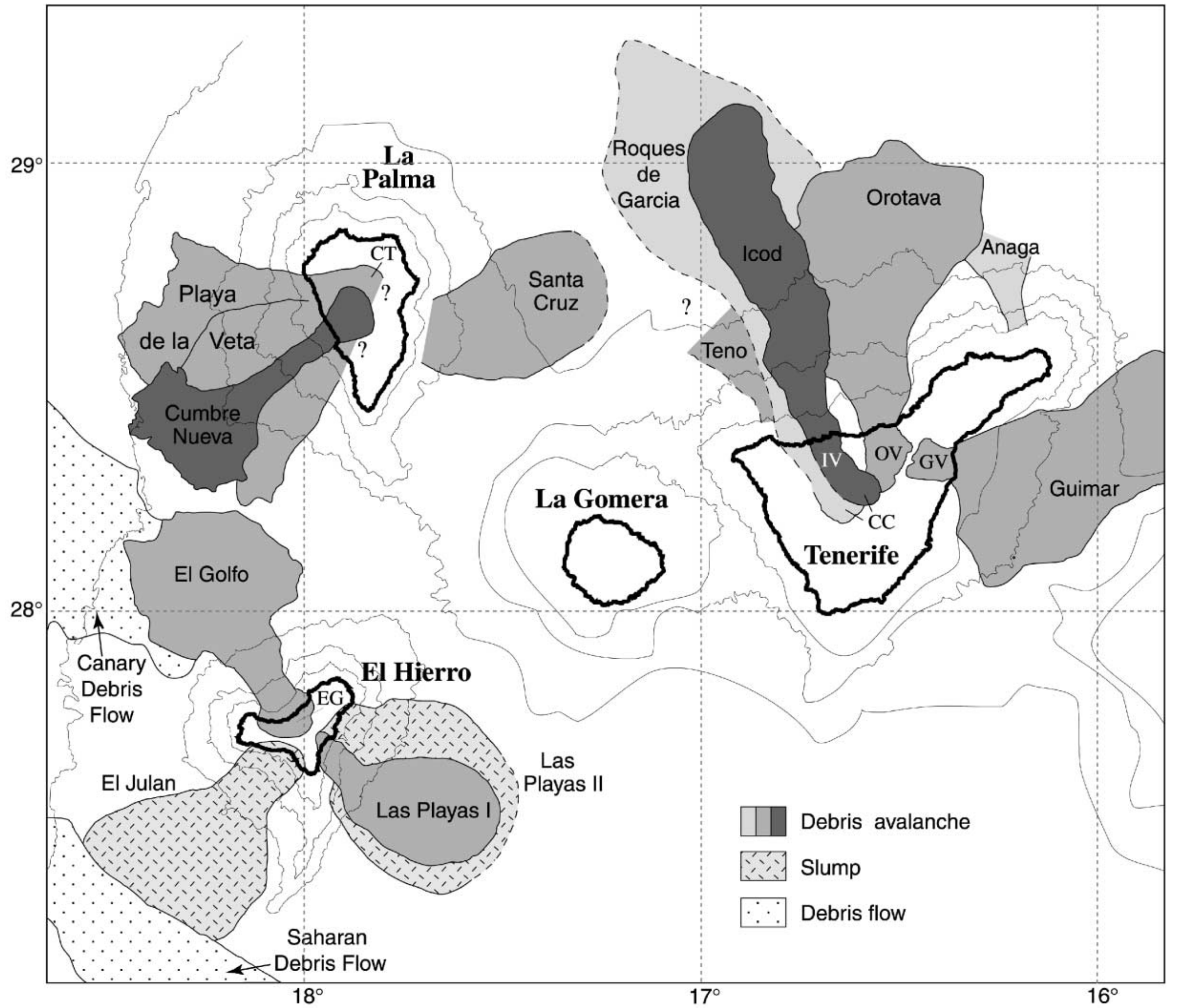

Fig. 12. Summary of mapped landslides on the flanks of El Hierro, La Palma and Tenerife. Variable grey shading of debris avalanche areas serves only to distinguish the different deposits. EG = El Golfo, IV = Icod Valley, OV = Orotava Valley, GV = Guimar Valley, CC = Canadas Caldera, CT = Caldera de Taburiente.

landslide origin (see Discussion). Downslope of these embayments, major debris avalanche deposits have been recognised. The scar of a further landslide, Tinor, has been mapped onshore on the older part of the island, but is not recognised offshore (Carracedo et al., 1997a). In addition, two major sedimentary debris flows have also influenced island flank development.

3.1.1. El Golfo Debris Avalanche

The El Golfo Debris Avalanche, on the northwest flank of El Hierro (Figs. 4, 10 and 13), is the most recent, most clearly defined and best described of the Canary Island flank failures (Table 1) (Masson, 1996; Masson et al., 1998; Urgeles et al., 1997). The avalanche scar is the clearly defined El Golfo embayment on the island of El Hierro (Fig. 4) with a headwall scarp in excess of 1000 m high. Downslope from El Golfo, the proximal erosional area of the El Golfo Avalanche consists of a smooth chute bounded by lateral scarps. The scarps are up to 600 m high, but decrease in height downslope, disappearing between 3000 and 3200 m water depth. Evidence from the offshore island flank, in the form of a turbidite

Table 1
Landslide parameters
In addition to data from this study, data has been included from the following sources: El Golfo (Masson, 1996; Masson et al., 1998); Las Playas (Day, personal communication, 1998); El Julan (Holcomb and Searle, 1991); Canary (Masson et al., 1996, 1998); Saharan (Embley, 1982; Gee et al., 1999); Cumbre Nueva and Playa de la Veta (Urgeles et al., 1999) (Day, personal communication, 1998); Orotava (Cantagrel et al., 1999; Marti, 1998; Watts and Masson, 1995); Icod, Roques de Garcia (Cantagrel et al., 1999); Guimar (Ancochea et al., 1990; Cantagrel et al., 1999; Krastel et al., 2001)

Island	Landslide	Type	Area (km^2)	Volume (km^3)	Height (m)	Runout length (km)	h/l	$A/V^{2/3}$	Age (ka)
El Hierro	El Golfo	Debris Avalanche	1500	150–180	5000	65	0.076	53	13–17
	Las Playas 1	? Slump	1700	?	?4000	?50			176–545
	Las Playas II	Debris Avalanche	950	< 50	4500	50	0.090	70	145–176
	El Julan	? Slump or Debris Avalanche	1800	? 130	4600	60	0.077		> 160
	Canary	Debris Flow	40,000	400	1450	600	0.0024	740	13–17
	Saharan	Debris Flow	48,000	1100	3200	700	0.0036	450	60
La Palma	Cumbre Nueva	Debris Avalanche	780	95	6000	80	0.075	37	125–536
	Playa de la Veta	Debris Avalanche complex	2000	? 650	6000	80	0.075	27	?800–1000
	Santa Cruz	Debris Avalanche	? 1000	?	?3500	50	? 0.070		? > 900
Tenerife	Icod	Debris Avalanche/ Flow	1700	? 150	6800	105	0.065	60	150–170
	Orotava	Debris Avalanche	2100	? 500	6600	90	0.073	33	540–690
	Roques de Garcia	? Debris Avalanche	? 4500	? 500	7000	130	0.054	71	? > 600
	Anaga	? Debris Avalanche	? > 400	?	> 3500	?	?		≫ 1000
	Guimar	Debris Avalanche	1600	120	> 4000	> 50	?	66	780–840

and debris flow linked to the debris avalanche, indicates an age of about 15 ka (Masson, 1996). Onshore evidence suggests a much greater age, in the range 100–130 ka; this has been reconciled with the younger age deduced offshore by postulating a two-phase failure (Carracedo et al., 1997a). However, there is no offshore evidence on the El Golfo flank for a failure older than 15 ka.

TOBI sidescan sonar images of the debris avalanche deposit show a typical blocky seafloor,

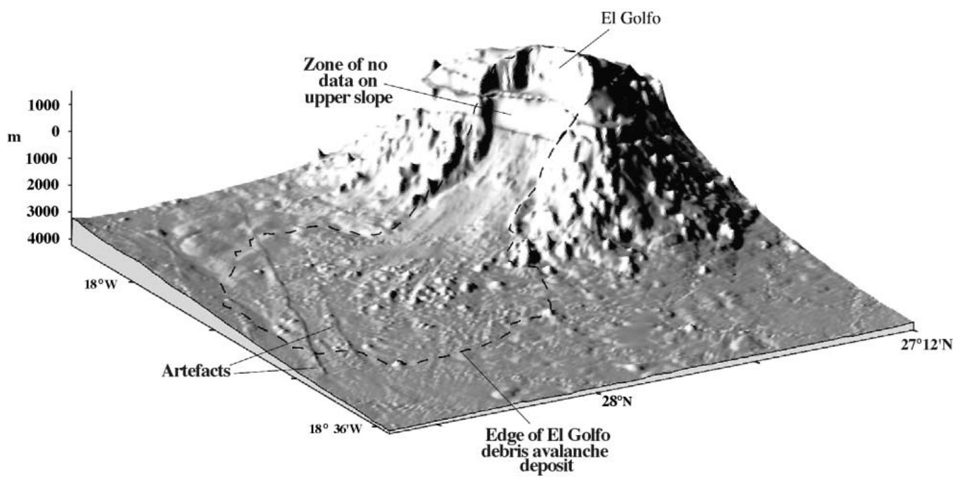

Fig. 13. 3D image of the El Golfo landslide scar and deposit on the northwest flank of El Hierro, viewed from the west.

with angular blocks up to 1.2 km across (Fig. 10) (Masson, 1996; Masson et al., 1998). Blocks can be up to 300 m high. The random distribution and orientation of blocks on the avalanche surface is a characteristic feature. Features indicative of flow, such as flow parallel lineations or shears, pressure ridges or alignment of block orientations, are not seen.

3.1.2. Las Playas Debris Avalanche (or avalanche / slump complex)

The Las Playas Debris Avalanche is a newly discovered feature on the southeast flank of El Hierro (Figs. 4 and 12; Table 1). The central part of this feature consists of a blocky debris avalanche deposit defined on the basis of EM12 backscatter data (Las Playas II, see Table 1). This debris avalanche appears to be superimposed on a wider area of deformed strata seen on seismic data (Las Playas I, see Table 1) (Gee, 1999).

The collapse scar associated with the central blocky debris avalanche is characterised by a narrow embayment <10 km wide which extends from the subaerial part of the flank to around 2500 m water depth (Figs. 3 and 4). The embayment is flanked by lateral scarps up to 500 m high (Fig. 3). Beyond 2500 m, EM12 backscatter data and 3.5 kHz profiles define a slightly elongate lobe of blocky avalanche debris with a typical block relief of a few tens of metres (Fig. 14). However, this lobe has no discernible positive bathymetric expression as can be observed, for example, over debris deposits west of La Palma (Fig. 3). Seismic profiles show a chaotic sequence ranging in thickness from a few tens of milliseconds distally to a maximum approaching 200 ms in the centre of the deposit, overlying a more transparent sequence. An average thickness of 100 m was used in estimating the volume of the Las Playas deposit (Table 1).

Onshore, the subaerial part of the Las Playas embayment was mapped as a landslide scar by Fuster et al. (1993). However, Day et al. (1997) re-interpreted the embayment as the result of preferential erosion along a series of faults orientated perpendicular to the coast. He considered the faults to be strike–slip faults marking the southern end of an aborted giant flank collapse which affected a broader region of the eastern flank of El Hierro. However, Day (personal communication, 1998) has now re-examined the area and agrees that the Las Playas area has been affected by landsliding. Based on the ages of volcanic sequences cut by and post-dating the Las Playas embayment, he dates the failure (i.e., Las Playas II) as between 145 and 176 ka.

Seismic profiles indicate a zone of deformation considerably broader than that recognised as related to the debris avalanche. Landward of this broad deformation zone (Las Playas I, Fig. 12), the subaerial part of the Las Playas flank is associated with a series of seaward-dipping, normal faults parallel with the coastline (Day et al., 1997). These faults, with a downthrow up to 300 m towards the coast, were interpreted by Day et al. as evidence for an aborted giant flank collapse which occurred between 545 and 261–176 ka. Our evidence, however, indicates that the faults are the headwall of the broad deformation zone, and that this may be a slump-type failure analogous to those seen off Hawaii (Moore et al., 1989). Failure of parts of slumps, giving rise to debris avalanches, as is suggested by the spatial relationship between Las Playas I and II, is also seen on the Hawaiian Ridge (Moore et al., 1989, 1994).

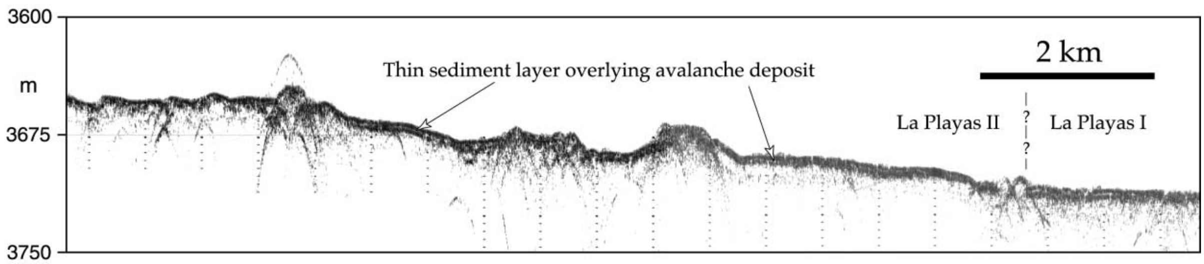

Fig. 14. 3.5-kHz profile showing upstanding blocks and hyperbolic echos in the area of the interpreted Las Playas debris avalanche deposit. Both the Las Playas I and II deposits appear to be buried beneath a sediment layer approximately 5 m thick.

3.1.3. El Julan Debris Avalanche (or slump?)

The El Julan Debris Avalanche was first interpreted on the basis of a characteristic speckle pattern on GLORIA long-range sidescan sonar data (Holcomb and Searle, 1991). EM12 bathymetry and backscatter, seismic, 3.5 kHz and TOBI sidescan sonar data collected on CD108 show a chaotic landslide deposit up to 300 m thick within the submarine El Julan embayment (Gee et al., 2001).

High-resolution TOBI sidescan sonar data collected from the El Julan slope show a stepped sediment-covered slope with only a few landslide blocks. The stepped seafloor is remarkably similar to that downslope from the El Golfo Debris Avalanche and could thus be interpreted as a partially failed slope cut by shallow rotational faults (Masson et al., 1998). Most of the avalanche blocks are angular features typical of landslide deposits elsewhere in the islands. However, a few of the larger ‘blocks’, some in excess of 400 m high, have a regular conical shape and an even, high backscatter. These are interpreted as post-landslide volcanic edifices, possibly the submarine equivalent of the small volcanic cones which are abundant on the present-day surface of El Hierro (Gee et al., 2001). Although the TOBI data suggest that landslide structures are exposed at the seabed, 3.5 kHz profiles show a regular undulating seafloor with a distinct surficial sediment layer between 10 and 12 m thick, particularly on the lower part of the island slope, indicating that El Julan is a relatively old structure. Dates obtained from the lavas filling the onshore part of the El Julan embayment indicate only that any landsliding must be older than 158 ± 4 ka (Carracedo et al., 1997a; Day et al., 1997).

The El Julan landslide poses a number of questions relating to its age and process of emplacement. Firstly, while the dating of lavas onshore and sediment drape seen on profiles offshore suggests an age > 160 ka, faulting of the surficial sediments observed on sidescan sonar images appears fresh, suggesting much more recent movement. One possibility is that the faults are not a consequence of the original landslide emplacement, but are later structures superimposed on the landslide deposit. Erosion of the toe of the El Julan deposit by the Saharan debris flow at 60 ka, or seismic activity related to the El Golfo landslide at about 15 ka, are both possible triggers for re-activation of the El Julan landslide deposit. Secondly, the rarity of blocks on the landslide surface, coupled with the observation of faults within the deposit, suggests that much of the deposit may not be highly disaggregated, but remained as a relatively coherent mass. This suggests a slump deposit or complex slump/debris avalanche deposit, rather than a fully developed debris avalanche. A slump, with relatively limited displacement in the headwall region, might also explain why later volcanism has been able to completely bury that headwall. Elsewhere, on La Palma and Tenerife, most avalanche headwalls leave some visible remnant or topographic expression, even after several hundred thousand years or later volcanism and erosion.

3.1.4. Canary Debris Flow

The Canary Debris Flow affected a wide area of the island slopes to the west of El Hierro and La Palma (Masson et al., 1992). Several studies indicate that it consists predominantly of fragmented sedimentary material, including pelagic sediments, turbidites and volcaniclastic material from the island flanks (Masson et al., 1992, 1997, 1998; Simm et al., 1991). The flow formed a thin (average 10 m thick) deposit over an area of around 40,000 km^2, on a slope $\ll 1°$, suggesting a highly mobile flow. The deposit is a classic debris flow mixture of clasts and matrix. Clasts range from centimetre-sized to large slabs up to 300 m across (Masson et al., 1997).

The Canary Debris Flow is directly related to the El Golfo Debris Avalanche and is interpreted to have been triggered by loading of the slope sediments by the avalanche (Masson et al., 1998). A turbidite emplaced simultaneously with the debris flow (Fig. 15) was probably derived primarily from the avalanche, although the precise source area and trigger of the turbidite cannot be ascertained. Emplacement age for the Canary Debris Flow is about 15 ka (Masson, 1996).

3.1.5. Saharan Debris Flow

The Saharan Debris Flow originated at about 24°N on the Northwest African continental margin to the south of the Canary Islands and flowed just to the southwest of El Hierro (Embley, 1976). Its main influence on Canary Island landsliding has been to rework and possibly bury part of the El Julan land-

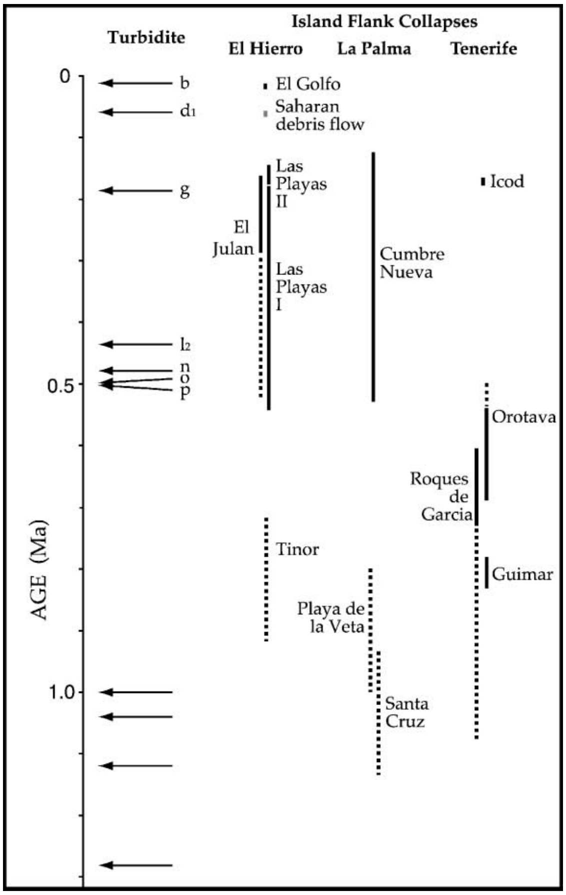

Fig. 15. Summary of the ages of volcaniclastic turbidites emplaced on the Madeira Abyssal Plain within the last 1.3 Ma (left) compared with the ages of landslides derived by dating onshore volcanic sequences (right). Onshore dates shown by solid lines are well constrained age ranges; dotted lines show greater uncertainty. Some dates for landslides older than 1.3 Ma have been determined onshore Tenerife, but are too poorly constrained to be plotted. The record of volcaniclastic turbidites on the abyssal plain extends back to 17 Ma, although such turbidites became much more common after 7 Ma (see text for further discussion).

slide deposit (Gee et al., 1999). TOBI sidescan sonar data collected on CD108 clearly shows erosion of El Julan landslide material in some areas and onlap of Saharan debris flow material onto the El Julan slope in other areas. Calculations based on the area and possible thickness of the volcaniclastic component of the Saharan Debris Flow (Gee et al., 1999) indicate that 40 km^3 or more of volcaniclastic Canary Island slope sediments must have been removed by and incorporated into the Saharan Debris Flow. Much of this is likely to have come from reworking of El Julan landslide deposit. It is possible that erosion of the toe of the El Julan landslide may have caused failure within the landslide deposit upslope. This might account for the apparent freshness of the shallow rotational faults which cut the landslide deposit on the lower slope. The Saharan Debris Flow has an age of about 60 ka, based on nannofossil dating of cores which penetrated the feather edge of the deposit (Gee et al., 1999).

3.2. La Palma

A large complex of debris avalanche deposits, covering an area of about 2000 km^2, forms a distinct lobate topographic bulge on the seafloor on the western flank of La Palma (Figs. 3, 5 and 16) (Urgeles et al., 1999). Four distinct debris avalanche lobes can be distinguished, representing at least two and probably as many as four landslide events. The youngest Cumbre Nueva landslide can be clearly distinguished from older deposits, grouped under the name Playa de la Veta landslide complex. A single landslide deposit, Santa Cruz, is also recognised on the eastern flank of La Palma.

3.2.1. Cumbre Nueva Debris Avalanche

The Cumbre Nueva Debris Avalanche, covering an area of around 780 km^2 and with a volume of about 95 km^3, is the youngest debris avalanche on the west flank of La Palma (Urgeles et al., 1999). The avalanche deposit forms a clear topographic bulge on the island flank between 2500 and 4000 m water depth (Figs. 3 and 12). The failure scar associated with this landslide extends onshore into the valleys bounded by the Caldera de Taburiente and the Cumbre Nueva Ridge, although the scar has been much degraded by later fluvial erosion (Carracedo et al., 1999). The blocky nature of the debris avalanche deposit is confirmed by TOBI sidescan sonar data (Urgeles et al., 1999). Evidence for the age of the Cumbre Nueva Avalanche is limited. The youngest rocks in the headwall, approximately 530 ka in age, give the maximum age; the oldest overlying lavas, 125 ka in age, give the minimum age. The record of volcaniclastic turbidites in the Madeira Abyssal Plain (MAP), which would be expected to contain evidence of any event on the western La Palma flank, suggests an age of between 420 and 500 ka. This inference can be drawn because the MAP record contains no younger volcaniclastic turbidites unattributed to other sources (Fig. 15) (Masson, 1996).

3.2.2. Playa de la Veta Debris Avalanche Complex

On the west flank of La Palma, debris avalanche deposits older than those of the Cumbre Nueva landslide are grouped under the name Playa de la Veta Debris Avalanche Complex (Urgeles et al., 1999). Three distinct debris lobes can be mapped between about 1000 and 3000–4000 m water depth, each forming a topographic bulge with rough surface topography (Figs. 5 and 12). Seismic profiles show a consistent chaotic signature within the interpreted debris lobes (Urgeles et al., 1999). The boundaries between the lobes are marked by incised channels which post-date avalanche emplacement, with channel locations apparently controlled by the topographic lows between debris lobes (Fig. 16). Although the age relationship between lobes is unclear, it seems likely that each represents a distinct avalanche event (Urgeles et al., 1999). In total, the Playa de la Veta Debris Avalanche complex may contain as much as 650 km^3 of material. The age of emplacement is poorly constrained, but is probably in the range 800 ka to 1 Ma (Urgeles et al., 1999).

3.2.3. Santa Cruz Debris Avalanche

A limited reconnaissance survey of the east flank of La Palma was carried out using the EM12 swath system. An area of blocky terrain, identified from EM12 backscatter data, is interpreted as a single lobe of debris avalanche deposits, here referred to as the Santa Cruz Debris Avalanche. The area of debris avalanche deposit is in the order of 1000 km^2. The debris avalanche headwall appears to lie in the vicinity of the coastal embayment around Santa Cruz on the east side of the island. However, there is little morphological evidence for a collapse scar either on land or on the upper island slope. This would suggest that the failure scar has been filled by later volcanism associated with the formation of the Cumbre Nueva and Taburiente volcanoes, suggesting an age in the order of 1 Ma (Carracedo et al., 1997b).

3.3. Tenerife

Tenerife is the highest and, in terms of surface area, the second largest of the Canary Islands. Debris avalanche deposits were first recognised offshore Tenerife by Watts and Masson (1995), who noted that these deposits covered much of the submarine northern flank of the island (Figs. 6, 12 and 17). They also recognised that these deposits were the product of several episodes of avalanching, although

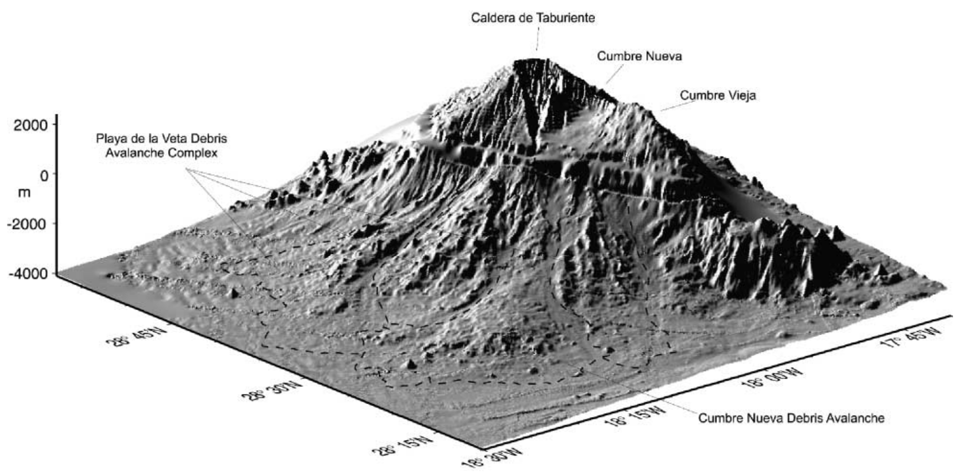

Fig. 16. 3D image of the western flank of La Palma showing the complex topographic bulge made up of multiple debris avalanche deposits. Dashed lines mark the various debris lobes, separated by channels which are post avalanche features which follow the topographic lows between lobes. View is from the southwest.

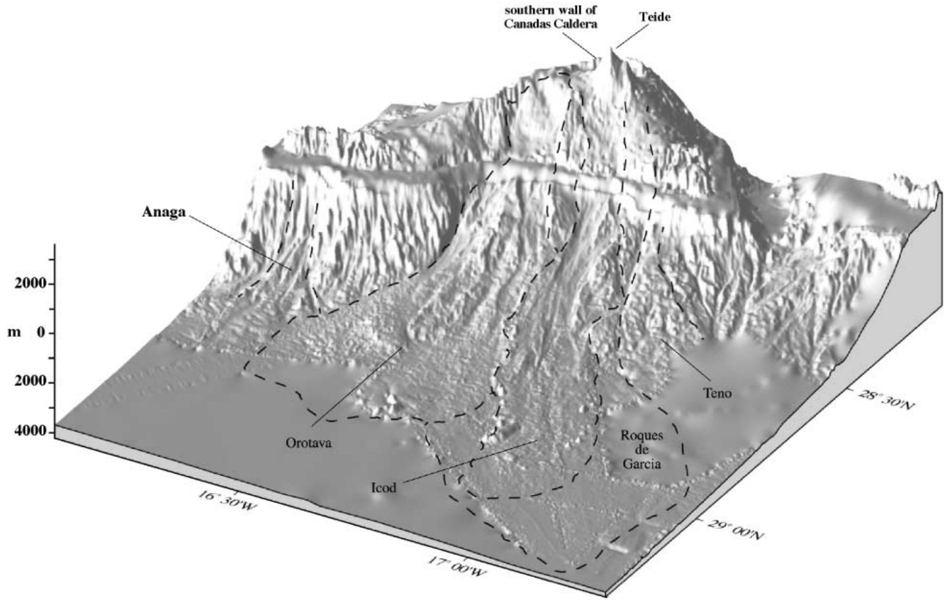

Fig. 17. 3D image of the north flank of Tenerife viewed from the northwest. Note the strong morphological contrast between the stable slope in the east and the area affected by multiple landslides in the west. The southern wall of the Canadas Caldera, to the south of Teide, is interpreted as the headwall of the Icod and probably Roques de Garcia landslides (see text for further discussion).

they were unable to separate the various events with any degree of confidence. Watts and Masson (1998) further developed the concept of multiple avalanche episodes, recognising at least four avalanche events (Figs. 6, 12 and 17). These have produced a large area of debris avalanche deposits with a combined volume estimated at 1000 km^3 (Teide Group, 1997; Watts and Masson, 1995, 1998). A fifth debris avalanche deposit is also recognised on the south-eastern flank of Tenerife, downslope from the Guimar valley (Fig. 12) (Krastel et al., 2001). In addition, at least two older landslide breccias (at 2–3 Ma and about 6 Ma) have been recognised from onshore rock exposures of limited extent (Cantagrel et al., 1999). The deposits from these landslides have not been mapped offshore.

3.3.1. *Icod debris avalanche / flow*

The Icod debris avalanche (or debris flow, see Discussion) is the youngest on the north flank of Tenerife, with an estimated age of 170 ka (Cantagrel et al., 1999; Watts and Masson, 1995). The area affected by the avalanche is up to 20 km wide and 105 km long (Fig. 12). Constraints on the thickness (and thus volume) of the avalanche deposit are poor. Steep margins to the deposit, up to 45 m high (Watts and Masson, 2001), give a minimum thickness, but a lack of penetration on high-resolution seismic profiles prevents assessment of its thickness over most of its area. TOBI 30 kHz sidescan sonar data show a high backscatter, blocky seafloor in the area of the avalanche deposit. Blocks are much smaller and far more numerous than seen on comparable images from the El Golfo Debris Avalanche deposit, except at the margins of the Icod deposit which are marked by a 'halo' of very large blocks up to 1.5 km across (Fig. 11). Flow structures, in the form of flow-parallel shear structures, aligned blocks and flow-perpendicular ridges, are clearly visible on the sidescan images (Watts and Masson, 2001). On 3.5 kHz profiles, no discernible sediment cover can be seen draping the debris deposit (Fig. 18). At water depths shallower than about 3 km, the deposit can be traced upslope into a linear depression about 10 km wide and up to 400 m deep, with a relatively flat bottom and steep-sided margins (Watts and Masson, 2001). This "chute-like" structure extends into the subaerial Icod valley and appears to be erosional in origin. The

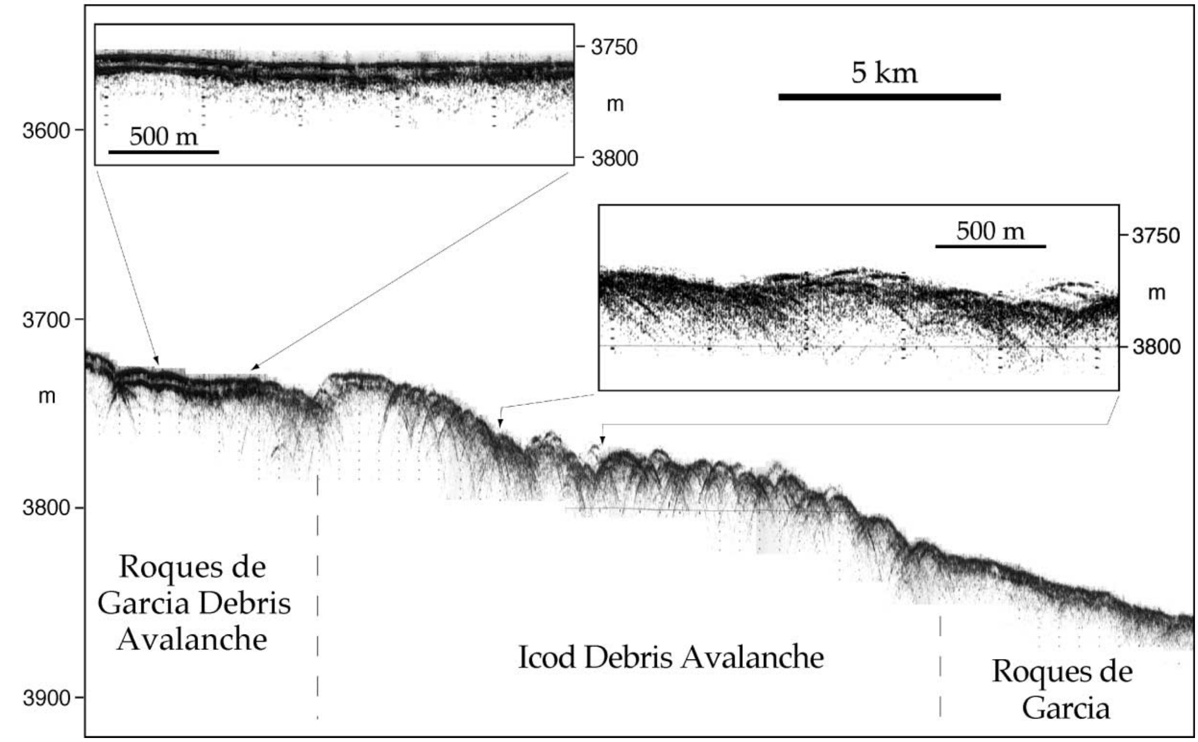

Fig. 18. 3.5-kHz profile crossing part of the Icod and Roques de Garcia debris avalanche deposits. The Roques de Garcia avalanche deposit is overlain by a distinct sediment layer which is absent over the Icod deposit. For location see Fig. 3.

head of the Icod Valley appears to extend into the Canadas Caldera, but the relationship between the valley and the caldera is largely obscured by the volcanic products of the younger strato-volcano of Teide Pico Viejo. The origin of the caldera is the focus of much current debate, with arguments for its creation by vertical collapse related to magmatic activity, a combination of vertical and lateral collapse, or solely by lateral collapse due to large-scale landsliding (see Discussion) (Ancochea et al., 1998, 1999; Cantagrel et al., 1999; Marti et al., 1997; Navarro and Coello, 1989; Ridley, 1971; Watts and Masson, 1995, 1998, 2001).

3.3.2. *Orotava Debris Avalanche*

New offshore data collected during Charles Darwin cruise 108, integrated with the most recent work onshore (Cantagrel et al., 1999), suggests that the Orotava Debris Avalanche can be mapped as a distinct debris lobe extending due north from the Orotava Valley (Figs. 6 and 12). The Orotava Debris Avalanche has a typical blocky character on sidescan sonar and EM12 backscatter data (Watts and Masson, 1998, 2001) (Figs. 8 and 10). The apparent density of blocks and the backscatter level are both lower than for the adjacent Icod Debris Avalanche (Figs. 8 and 10), probably reflecting the greater sediment cover on, and hence age of, the Orotava avalanche. High resolution seismic profiles crossing the distal part of the Orotava Debris deposit show that this sediment cover is thicker than previously recognised, reaching 20 m or so in thickness (Fig. 19). At a large scale, the surface of the avalanche is characterised by longitudinal ridges and valleys with a peak to trough amplitude of 300 to 400 m and a wavelength of 5 to 8 km (Watts and Masson, 1995). These are interpreted as ridges of debris separated by channels or chutes which were the main debris transport pathways. Sidescan sonar data gives the impression that the ridges are characterised by larger and

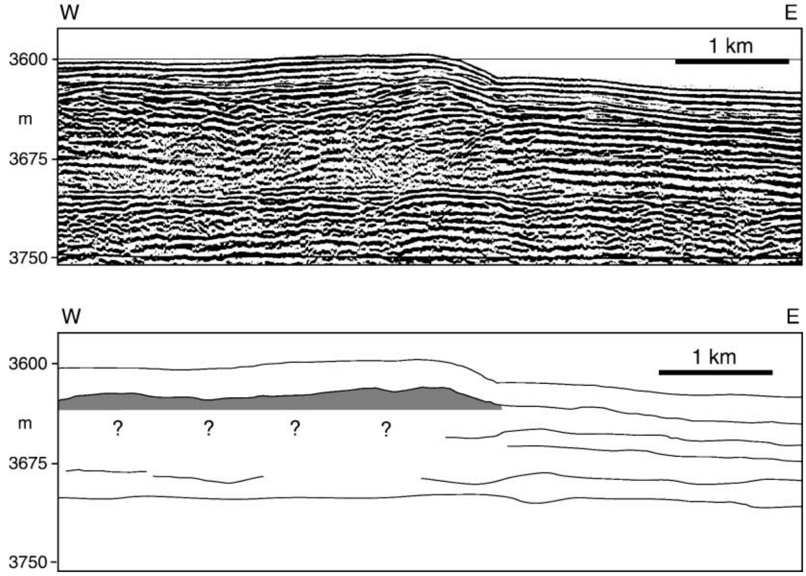

Fig. 19. High resolution multichannel seismic profile (and interpretation) across the distal part of the Orotava debris avalanche deposit. This profile indicates a layer of post-avalanche sediment up to 20 m thick.

more numerous blocks compared to the valleys. However, this may be due to greater post emplacement sedimentation in the valleys.

The Orotava Debris Avalanche can be traced upslope into the Orotava Valley on Tenerife (Figs. 6 and 12). Morphologically, the subaerial Orotava Valley, with its flat floor flanked by steep scarps (Palacios, 1994; Ridley, 1971), is the most distinct landslide-related valley on Tenerife. The age of formation of the Orotava Valley is estimated at between 540 and 690 ka, based on K/Ar dating of basaltic lavas in the upper part of the landslide scarp (Cantagrel et al., 1999). An age in the range 500–700 ka is compatible with the 20 m or so of sediment cover seen draping the debris deposit offshore (Fig. 19). Orotava is believed to have been the product of the lateral collapse of the Canadas volcano which preceded the Teide Pico Viejo complex. It is one of the largest landslides to have occurred in the western Canary Islands. Taken together, the volume of debris avalanche deposits on the central part of the north flank of Tenerife (from the Icod, Orotava, Roques de Garcia and possible other older avalanches) has been estimated at more than 1000 km^3 (Teide Group, 1997; Watts and Masson, 1995). Probably $< 50\%$ of this volume can be ascribed to the Orotava avalanche, although the existing data does not allow us to estimate volumes for the individual debris avalanche deposits with any confidence. In particular, uncertainty in the estimate of the thickness of the Orotava deposit leaves a considerable error margin in the volume estimate.

3.3.3. Roques de Garcia and Teno Debris Avalanches

To the west of the Icod Debris Avalanche, Watts and Masson (1998) described a poorly defined, apparently older area of landslide deposits, which they referred to as the Teno Debris Avalanche. Although data collected during Charles Darwin cruise 108 has allowed us to better define this avalanche area (Figs. 6, 12 and 17), the new data also confirms the difficulty in mapping landslide boundaries within this topographically complex area which was probably affected by at least two distinct landslides prior to the Icod landslide. The more recent of these can be traced onshore to the west of the Icod Valley, strongly suggesting a correlation with the Roques de Garcia debris avalanche inferred onshore (Cantagrel et al., 1999). Thus we now refer to this avalanche as the Roques de Garcia debris avalanche. ‘Teno debris avalanche' is now used to describe possible older avalanche deposits, recognised on the upper part of the submarine slope west of the area affected by the Roques de Garcia landslide.

The character of the Roques de Garcia avalanche deposit, when seen on sidescan sonar, EM12 or 3.5 kHz profile data, is similar to that of the Orotava avalanche deposit (see previous section). In the most northern part of the debris deposit, away from the areas affected by the Orotava and Icod events, sidescan sonar and EM12 backscatter data show numerous, widely scattered small blocks on a background of smooth sedimented seafloor. Seismic profiles show that this smooth seafloor is the result of between 10 and 25 m of sediment overlying the avalanche deposit.

The age of the inferred Roques de Garcia avalanche is not well constrained onshore, although it is probably >600 ka and possibly >1 Ma in age (Cantagrel et al., 1999). An age slightly greater than that of the Orotava avalanche is compatible with the slightly thicker post-emplacement sediment cover seen offshore. The volume of the Roques de Garcia avalanche deposit cannot be determined, but given the large area it covers, it is likely to be of the same order as that of the Orotava deposit. The area of the Teno massif, in the extreme northwest of Tenerife, appears to have a landslide history extending back to between 5.0 and 6.0 Ma (Cantagrel et al., 1999). However, the limited extent of the evidence for landsliding, both in the form of landslide breccias onshore and slope morphology offshore, allows little of the detail of this history to be deduced.

3.3.4. Anaga Debris Avalanche

The Anaga landslide scar is recognised primarily on the basis of its topographic and morphological expression (Watts and Masson, 1998) (Figs. 6 and 17). It forms a distinct embayment which extends from the shelf edge to about 3000 m water depth. The seafloor within this embayment is characterised by a series of large-scale gullies and ridges, with a wavelength of up to 5 km and an amplitude of a few tens of metres (Fig. 6). A possible associated debris deposit is indicate by a topographic bulge in the region of the 3200 m contour (Fig. 3).

Anaga appears to be one of the older landslides to have affected Tenerife. Its western boundary is onlapped by the Orotava Debris Avalanche, indicating it must at least be older than about 600 ka. Watts and Masson (1998) report up to 100 m of post-landslide drape covering this landslide, indicating that Anaga may be $\gg 1$ Ma in age. The age of the landslide is also reflected by the presence of a 5-km-wide wave-cut shelf around the Anaga massif, which has been formed since landsliding.

3.3.5. Guimar Debris Avalanche

On the southeastern flank of Tenerife, a broad area of hummocky topography, corresponding to an area of distinctive speckled pattern on EM12 backscatter images, is interpreted as a debris avalanche extending east from the subaerial Guimar Valley (Krastel et al., 2001; Teide Group, 1997) (Figs. 6 and 12). This landslide is unusual because it is the only example described from the Canary Islands to have occurred on a buttressed flank, the buttress being the island of Gran Canaria. This island appears to have deflected the debris avalanche slightly toward the northeast.

The subaerial valley of Guimar is around 10 km wide with a relatively flat base and well-defined flanking scarps between 300 and 600 m high. In its upper part, the Guimar Valley forms two diverging sub-valleys separated by a low ridge. Both sub-valleys extend onto the dorsal ridge, a narrow ridge separating the Orotava and Guimar Valleys. Based on the age of infilling lavas, the valley is estimated at around 0.78 to 0.83 Ma in age (Ancochea et al., 1990; Cantagrel et al., 1999).

4. Discussion

4.1. Flow and deposit statistics

4.1.1. Rates and volumes of landslide erosion

An understanding of the effect of large-scale landsliding on volcanic island evolution requires a comparison between the rate of island construction by volcanic processes and the rate of material removal by landsliding. The rate of erosion by landslides critically depends on our ability to produce accurate estimates of landslide ages and volumes. The latter can be estimated either from the volume of the landslide scar or the volume of the landslide deposit (Masson, 1996; Urgeles et al., 1997, 1999; Watts and Masson, 1995). Ideally, volumes should be calculated for both scar and deposit, thus reducing the amount of error.

Some assumptions usually have to be made in the estimate of landslide deposit volume. For example, estimates of the deposit volume are often based on its topographic anomaly, since the base of such deposits is rarely imaged by seismic profiling techniques. Little is known about the porosity of landslide deposits, so any increase in volume caused by fragmentation during landslide emplacement is ignored. Loss of material from the landslide mass may occur due to its entrainment into turbidity currents or debris flows, while material may be added by erosion of the seafloor over which a landslide travels. However, these changes are usually impossible to estimate and have to be ignored, with the assumption that they are small compared to overall landslide volume.

Estimates of landslide volume independently based on the volume of its scar and deposit usually give comparable results when applied to relatively recent landslides, where the landslide scar has been little modified by erosion or later volcanic activity and the deposit can be identified as a distinct topographic bulge with clearly definable boundaries. For the El Golfo avalanche on El Hierro, the volume of the avalanche scar can be estimated at about 180 km^3 (Fig. 20). This compares favourably with a volume of about 150 km^3 calculated for the deposit (Urgeles et al., 1997), with the difference probably accounted for by avalanche material incorporated into the Canary debris flow or transported downslope by the turbidity current associated with avalanche emplacement (Masson, 1996). The correspondence between the two volume estimates gives clear confidence in the calculation methods. The reliable volume estimate, in turn, gives confidence that El Golfo resulted from a simple, single flank failure. Suggestions that the most recent El Golfo landslide occurred within the El Golfo embayment, but affected only the offshore segment (Carracedo et al., 1997a) are not supported by the volume assessment (Fig. 20), nor the offshore mapping evidence (Masson et al., 1996). Similarly, suggestions that El Golfo might have been location of a 2 km + high volcano (Carracedo et al., 1997a) are not supported by the volume of debris mapped offshore.

When the landslide scar has been heavily modified by later volcanism and/or erosion, as in the case of the Caldera de Taburiente on La Palma (Carracedo et al., 1999), landslide volume can only be estimated from the volume of the deposit. For a single deposit on any island flank, a reliable estimate, based on the volume of the topographic anomaly associated with that deposit, appears to be possible (Urgeles et al., 1999). Where multiple overlapping landslide deposits occur, such as north of Tenerife or west of La Palma, it is usually possible only to estimate overall debris volumes (Urgeles et al., 1999). In the case of multiple deposits covering large areas, the assumptions relating to the geometry of the pre-landslide island flank become correspondingly large, and volume estimates correspondingly less reliable. For example, where landslides cover a broad sector of island flank, such as the Playa de la Veta complex on the west side of La Palma, assuming a flat seafloor may lead to an overestimate of landslide volumes because of the natural curvature of the island flank.

In some cases, such as north of Tenerife and west of La Palma, the volume of landslide deposits offshore is substantially greater than the volume of the landslide scars onshore (Cantagrel et al., 1999; Urgeles et al., 1999; Watts and Masson, 1995). In the case of the north flank of Tenerife, this apparent discrepancy can be explained, on the basis of onshore geological evidence, by repeated cycles of volcanic construction and failure of the same general area of flank (Cantagrel et al., 1999). A similar situation is seen on the island of Reunion and can also be explained by repeated volcanic buildup and failure cycles.

When compared to the volume of the individual volcanic islands, debris avalanche deposits around El Hierro and La Palma account for about 10% of the total volcanic edifice. For La Palma, the volume of the edifice (above the present day seafloor) is about 6500 km^3 (Urgeles et al., 1999) and the volume of known debris avalanches is about 600–800 km^3 (Table 1). Corresponding figures for El Hierro are 5500 km^3 and 400–500 km^3 (Gee, 1999). Individual landslides may remove as much as 25% of the

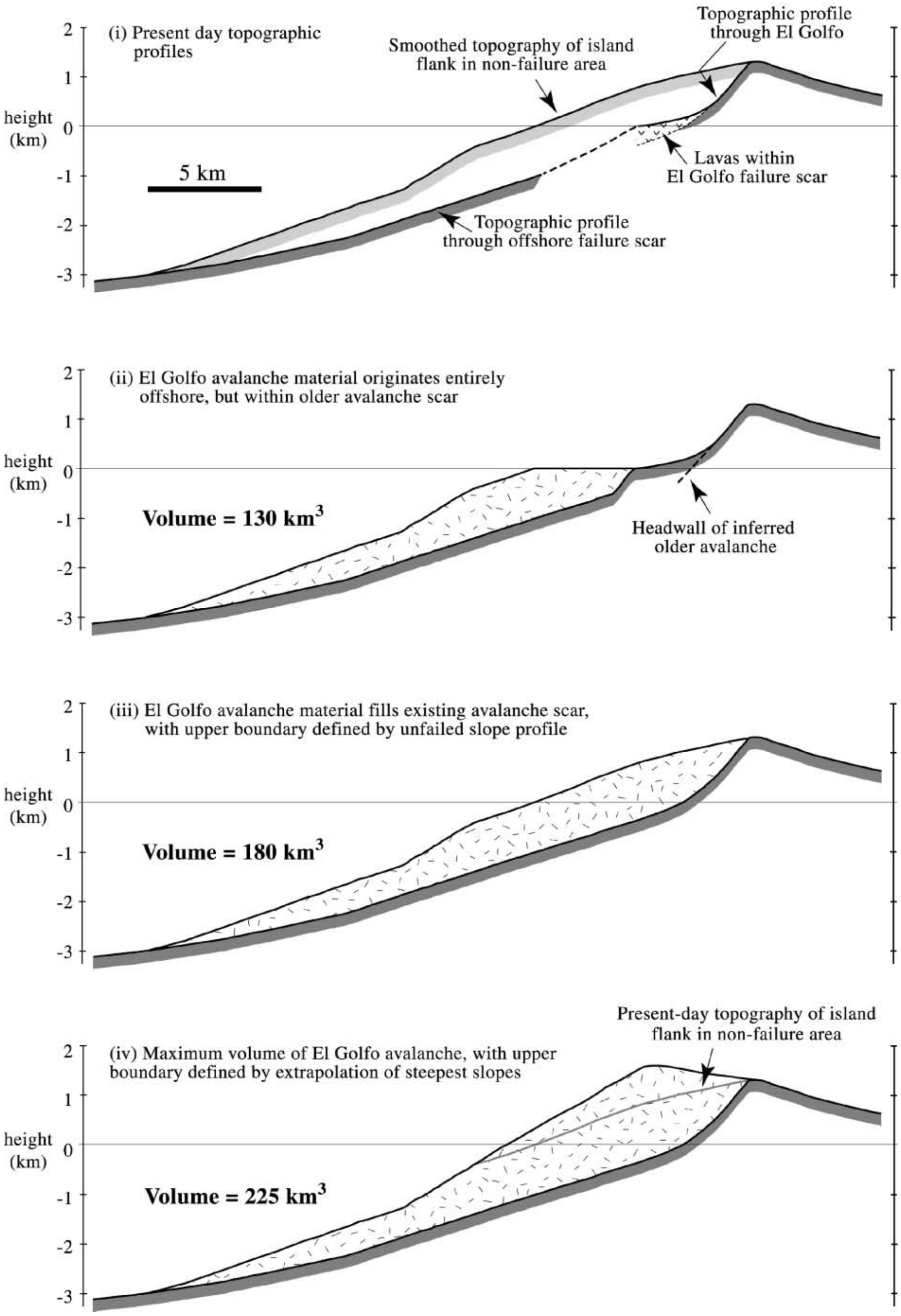

Fig. 20. Series of reconstructed cross-sections through the El Golfo landslide scar showing the uncertainty in calculating landslide volumes on the basis of the scar. The upper cross-section shows present day profiles, heavily smoothed to remove the effects of local features, such as volcanic cones or gullies. The lower three profiles show calculations of the possible landslide volume, based on variable assumptions detailed on the figure.

subaerial part of an individual island. This is important in understanding the hazard potential of such landslides. For example, the El Golfo landslide removed some 50 km^3 of material from an island of perhaps 200 km^3 in volume. Almost 50% of the seafloor within 60 km of the shoreline of El Hierro is covered with landslide debris (Gee, 1999).

4.1.2. Relationship between volume, runout distance, and height

The relationship between the volume of a landslide (V), its runout distance (l) and the height drop between the headwall and the landslide toe (h), usually expressed as h/l plotted against V, is commonly used as a measure of the relative efficiency of a landslide (Hampton et al., 1996; Scheidegger, 1973). Even allowing for differences in volumes, landslides on the flanks of the Canary Islands clearly fall into two categories, with sedimentary debris flows (Canary and Saharan flows) having a h/l ratio more than an order of magnitude lower than any of the debris avalanches on the island flanks (Table 1). Canary Island debris avalanches plot in the same field as Hawaiian debris avalanches (Hampton et al., 1996). The Icod 'debris avalanche', which we argue below exhibits many of the characteristics of a debris flow, has an h/l ratio toward the lower end of the distribution for 'debris avalanches', but is barely distinguished from other Canary Island 'debris avalanches' by this measure.

Dade and Huppert (1998) showed that the runout of rockfalls was related to their potential energy and to the area over which they spread. A relationship between area (A) and volume $(V)^{2/3}$ was determined. Canary Island debris avalanches fall on the trends predicted by this analysis, suggesting typical rockfall behaviour. Canary Island sedimentary debris flows, in contrast, spread over much greater areas relative to their volume (see Table 1). The spread of values of the $A/V^{2/3}$ ratio seen for the various Canary Island debris avalanches is not considered significant, given the uncertainties in estimating both A and V. The former, in particular, should strictly be calculated as the area covered by the landslide deposit, rather than the total area affected by landsliding, i.e., scar plus deposit, as it is only possible to estimate here.

4.2. Character and distribution of landslides

4.2.1. Canarian landslides

Most Canarian landslides take the form of debris avalanches (Table 1). The typical Canary Island debris avalanche has an overall lobate shape and extends from a subaerial headwall scarp to a blocky debris deposit in 3000–4000 m water depth. Most debris avalanche scars have a relatively narrow headwall region and 'chute' above 3000 m water depth, where erosion and downslope transport dominate, and a broader deposition lobe below 3000 m water depth (Fig. 12). Sidescan sonar images show that although debris avalanche deposits have a typical blocky structure, block size and the degree of fragmentation of the avalanche material are variable (Figs. 10 and 11). At one extreme, El Golfo has large blocks scattered randomly across the avalanche surface. At the other, Icod has much more numerous but smaller blocks over most of its surface, with a few large blocks confined to the margins of the deposit. The primary controls on the block structure and distribution are likely to be the nature of the landslide material (e.g., volcanic or intrusive rocks, pyroclastic deposits or volcaniclastic sediments) and the flow processes (e.g., flow or avalanche, speed of emplacement). This is further discussed below.

Only two landslides in the western Canaries might tentatively be identified as slumps (as defined by Moore et al., 1989; Table 1). These are the Las Playas I and possibly El Julan landslides on the flanks of the island of El Hierro (Fig. 12). Evidence from Hawaii indicates that slumps affect a considerable thickness of or even the whole volcanic edifice, and that movement on the decollement at the base of the slump occurs within sediments between the oceanic crust and the base of the volcanic edifice (Moore et al., 1994).

One possible explanation for the more common occurrence of deep-seated slumps around Hawaiian Islands relative to the Canary Islands is that Hawaii, unlike the Canary Islands, is associated with a large amplitude topographic swell. The Hawaiian plume also has greater melt production rates than the Canary plume (e.g., White, 1993). The combined effect of these processes is that Hawaii may have been maintained at a greater elevation for longer times than the Canary Islands, thus increasing the gravitational potential which drives the slumping process. It might also be speculated that slumps would be more common on young volcanic islands, such as El Hierro, where rapid growth of the volcanic edifice imposes a rapidly changing load on the underlying poorly consolidated sediments. For islands with longer histories, such as Tenerife, consolidation of the underlying sediments under loading and a tendency toward a balance between volcanic growth and erosive processes may reduce the likelihood of slumping. In addition to the landslides identified offshore, faulting associated with the 1949 volcanic eruption on La Palma could be a superficial indicator of deeper slump-type deformation of the young, southern volcanic edifice on that island (Carracedo et al., 1999). Fault downthrow to the west of a "few metres" was observed on a 4-km-long fault system (Carracedo, 1996; McGuire, 1996). The magnitude of this movement is similar to that observed during the 1975 event on the Hilnia Slump in Hawaii, although fault movement on Hawaii occurred over a much greater lateral extent (Moore et al., 1989). However, no evidence of slumping has been observed on the offshore flank of La Palma, and all of the landslides affecting the older northern part of the island are debris avalanches (Fig. 12).

Large-scale sediment failures leading to debris flows do not appear to be common on the flanks of the Canary Islands, with only one documented example, the Canary debris flow, having its source on the island slopes. Relationships between debris flows and other landslide types are further discussed below.

4.2.2. Controls on landslide locations in the western Canary Islands

A model linking volcanic rift zones and landsliding in the Canary Islands was proposed by Carracedo (1994,1996) and Carracedo et al. (1998). This model is based on the observation that volcanism at many oceanic islands and seamounts is concentrated on a series of volcanic rift zones radiating from the centre of the volcanic edifice (Fiske and Jackson, 1972; Vogt and Smoot, 1984). The resulting stellate edifice geometry may be further enhanced by landsliding between the rift arms, with landslides commonly propagating perpendicular to the rift direction (Moore et al., 1989). Three-armed rifts, spaced at 120°, seem to be the naturally preferred (least horizontal stress?) situation, as in the case of El Hierro and Tenerife (Carracedo, 1994,1996; Carracedo et al., 1998). However, single rifts (or systems where one arm of a multiple rift system becomes dominant), leading to elongate volcanic edifices, are also seen, as in southern La Palma (Carracedo et al., 1999). In the case of La Palma, the single rift may result from buttressing of the young southern rift by the older northern part of the island.

Although the proposed link between rifts and landslide location can explain the overall distribution of landslides and stable island flanks, it does not entirely explain the uneven distribution of landslides around some islands. On both La Palma and Tenerife, multiple landslides have occurred on one particular flank, while other areas have remained more stable over long periods (Fig. 12) (Krastel et al., 2001; Urgeles et al., 1999; Watts and Masson, 1998). This suggests that once established, an unstable flank becomes a weakness which can be carried forward in the development of the island. Several factors may contribute to this phenomenon. Firstly, failures may be most easily initiated on island flanks which have the greatest gravitational potential for movement, i.e., those which are steepest and/or extend to greater water depths and are least buttressed by the surrounding topography. This might result in extension on island rift zones being asymmetrically directed toward the least buttressed flank, increasing the possibility of this flank becoming detached and failing. This would appear to be a possible scenario in the cases of La Palma and Tenerife. Secondly, by locally decreasing the overburden on any underlying magmatic system, the removal of part of an island by landsliding may focus later magmatism to the region of the landslide scar. Preferential accumulation of new volcanic products within the scar may then increase the probability of future instability in that area. This may be a factor in the case of Tenerife, where multiple failures appear to have originated from approximately the same source area (Cantagrel et al., 1999). Thirdly, landslide breccias within old landslide scars may form weak layers, prone to failure when loaded by new volcanic products (Ancochea et al., 1999; Cantagrel et al., 1999; Labazuy, 1996; Watts and Masson, 2001). Finally, in the case of Tenerife in particular, the concentration of rainfall on the north side of the island may increase pore water availability at shallow depths within the volcanic edifice, with the possibility that this may increase the risk of landsliding.

4.2.3. Comparisons with other areas

Large-scale landsliding is recognised on many oceanic island volcanoes (Holcomb and Searle, 1991). Blocky debris avalanche deposits, similar to those on the Canary Island flanks, are widespread on the flanks of the Hawaiian Ridge (Moore et al., 1989, 1994) and off Reunion (Labazuy, 1996; Ollier et al., 1998). Other volcanic islands, such as Stromboli and Etna in the Mediterranean Sea, exhibit clear morphological evidence of landslide scars onshore (Rust and Neri, 1996; Tibaldi, 1996).

A close relationship between volcanic rift zones and landslides, similar to that proposed for the Canaries, is seen in the Hawaiian Islands, where it is generally observed that landslides (both slumps and debris avalanches) move perpendicular to rift zones (Moore et al., 1989). Proposed controlling mechanisms imposed by the rift zone include severing of the updip attachment of a landslide mass due to dyke injection and the creation of excess pore pressures in the landslide mass due to magma injection (Elsworth and Voight, 1995). However, a circular rift/landslide relationship can also be envisaged, because extension on rift zones generated by gravitational movement may also facilitate magma injection (Moore et al., 1989).

The eastern flank of Piton de la Fournaise volcano on Reunion Island appears to have suffered repeated failures similar to those on the north flank of Tenerife (Labazuy, 1996). Here, an offshore landslide deposit some 550 km^3 in volume, much of which consists of subaerially erupted basalts, is associated with an onshore landslide scar almost an order of magnitude smaller on the island, a situation similar to that on Tenerife. Three landslides appear to have occurred in the last 150 ka, indicating a landslide system more active than that on Tenerife.

4.3. Failure processes on the Canary Island flanks

4.3.1. Triggering mechanisms

Although the general relationship between rift zones/dyke injection and landsliding is largely accepted, little is known about the mechanisms which trigger individual large-scale volcano flank landslides. At one end of the spectrum, large slump-type landslides, driven by gravitational instability, may creep continuously and require no specific triggers to initiate movement (McGuire, 1996; Rasa et al., 1996). However, as evidenced by the 1975 movement on the Hilnia Slump on Hawaii, even large slumps are capable of essentially instantaneous movement (Moore et al., 1989). The 1975 slump movement was accompanied by both a large earthquake and a small volcanic eruption. However, as noted by Moore et al. (1989), it is not clear whether slump movement caused these associated phenomena, or dyke injection linked to the volcanic eruption or the earthquake was the original trigger of slump movement.

At the other end of the spectrum, debris avalanches are instantaneous failures. Small debris avalanches are frequently associated with volcanic eruptions, such as the 1980 eruption of Mount St. Helens in the United States. In this type of debris avalanche, failure usually results from oversteepening due to rapid edifice growth over a relatively short period; seismicity related to the eruption may also play a role in defining the precise moment of failure (see review in McGuire, 1996). In the case of oceanic islands, however, large-scale debris avalanches are infrequent events which punctuate long periods of apparently stable volcanic growth. In the western Canary Islands, the return period for landslides on each individual island is a few hundred thousand years (Fig. 15). There is no particular evidence that failures are related to periods of unusual volcanic activity, although we recognise the limited resolution of the island volcanic records in this context. Thus landsliding appears to a response to long-term volcanic build-up to a point where the load on the volcanic edifices exceeds that which it can support. When this critical point is reached, a relatively minor trigger may be all that is required to initiate a landslide. Thus events that have occurred commonly throughout construction of the (stable) volcanic edifice, such as dyke injection or seismicity, may be critical in triggering failure. Increases in pore pressure related to magma intrusion, which can destabilise a previously stable edifice without any change in the morphology of that edifice, may be crucial (Day, 1996; Elsworth and Voight, 1995).

Climate change and, more particularly, related fluctuations in sealevel may also affect the stress regime within an oceanic volcanic island, as suggested by links between sealevel and the intensity of explosive volcanism in the Mediterranean Sea (McGuire et al., 1997). Such stress changes could also affect flank stability. More speculatively, for the Canaries, climate change may be linked to changes in rainfall, in turn affecting groundwater level within the volcanic edifices. Heating of groundwater by magma intrusion may contribute to pore pressure increases and hydrothermal alteration, leading to weakening of the edifice structure in the manner suggested by Day (1996). Accurately dated Canary Island landslides are too few in number to allow a statistical evaluation of any possible link with climate. Dating of volcaniclastic turbidites (thought to be generated by landsliding on the island flanks) in the Madeira Abyssal Plain sequence to the west of the Canaries is also inconclusive (Fig. 15). Seven such turbidites are recognised within the part of the sequence dated using the oxygen isotope age scale, directly correlated with climate. Four of these coincide with isotope stage boundaries, indicating periods of climate change, while two occurred during sealevel highstand and one during sealevel lowstand.

4.3.2. Flow processes

Landslides on the Canary Island flanks involve a range of processes including slumping, debris flow and debris avalanche. It is important to recognise that these are not distinct and separate processes, but part of a process continuum (e.g., Mulder and Cochonat, 1996). Most of the data discussed here gives information on the surface morphology of landslide deposits and is most useful in understanding the emplacement of the more superficial landsliding processes, such as debris avalanches and debris flows, which are dominated by laminar flow processes. We have little data on slump emplacement and have assumed a ‘Hawaiian’ model for these landslides (Moore et al., 1989, 1994).

Debris (or rock) avalanches are defined as large-scale catastrophic events where the avalanche material becomes highly disaggregated and energy is dissipated through the avalanche mass by collisions between clasts (Mulder and Cochonat, 1996). Debris avalanches are the dominant landslide type in the western Canary Islands (Table 1, Fig. 12). Most originate from a headwall scarp with a distinct amphitheatre shape, are restricted to a narrow ‘chute’ as they cross the upper part of the island flank, and spread abruptly into a broad depositional lobe beyond the distinct decrease in slope gradient which occurs at the base of the volcanic edifice (Fig. 12). High-resolution sidescan sonar images of a typical debris avalanche deposit, El Golfo on the island of El Hierro, show the characteristic features to be a random distribution of clasts up to 1 km or more in diameter and a general absence of structures indicative of flow, such as longitudinal shears or pressure ridges (see Description of landslides).

The characteristic shape of the debris avalanche deposit beyond the break of slope at the edifice base and the lack of flow structures on the surface of the deposit strongly support the idea that the energy of the debris avalanche is dissipated by collisions between clasts. As a result of this energy transmission mechanism, the avalanche spreads laterally in all directions as soon as it escapes the confines of the chute on the upper island slope, giving the characteristic broad lobate deposit on the lower slope. Spreading of the debris avalanche beyond the confines of the chute, thus increasing its surface area and lowering its thickness, may also contribute to the loss of excess pore pressures within the moving avalanche.

The Icod debris landslide deposit differs from all other blocky landslide deposits on the Canary Island flanks in being much more elongate than typical avalanche deposits, because of the occurrence of flow structures within the deposit, and because of the halo of coarse blocks which mark its perimeter (Figs. 11 and 12; Watts and Masson, 2001). Almost all of the characteristics of the Icod flow (see Description of landslides) can be reproduced in large-scale coarse-grained debris flow experiments, giving a strong insight into the flow processes which controlled Icod emplacement (Iverson, 1997; Major, 1997; Major and Iverson, 1999). Such experimental flows typically produce elongate debris lobes with gravel ridges on the surface of the deposit (Major, 1997). They also produce flows with steep margins, particularly in the distal reaches of the flow, often with a concentration of coarser particles at the margin (Major and Iverson, 1999). Measurements of pore pressure in the interior of flows during their emplacement indicate that high pore pressures, "nearly sufficient to cause liquefaction" persist through all phases of the flow, from mobilisation to deposition. However, these high pore pressures are absent from the flow margins (Major and Iverson, 1999). Thus the steep flow margins result from 'strength' produced by friction between grains and between the flow and the underlying bed. High friction at the flow margin causes deposition of marginal ridges or levees, effectively damming the flow and halting its forward progress, even though high pore pressures continue to exist in the flow interior. Both experimental and natural subaerial flows typically develop a series of surges; successive surges deposit by pushing aside and forward into material already deposited. Because high pore pressures persist in the deposit interior after it comes to rest, giving a very weak material, successive surges are able to deform deposits from earlier surges, producing flow structures such as gravel ridges and longitudinal shears. Surging flows also produce a single amalgamated deposit, indistinguishable from a single en masse debris deposit.

Examination of the shape of debris deposits around the western Canary Islands suggests that El Golfo and Icod form the end members of a series of landslide shapes (Fig. 12), probably representing a series of flow processes ranging from debris avalanche (more lobate) to coarse-grained debris flow (more elongate), respectively. Although most of the landslides originating on La Palma and El Hierro are similar in shape to El Golfo, landslides originating from Tenerife exhibit the whole range of shapes. We would suggest that the difference in landslide type primarily results from differences in the rock types involved in the landslides. La Palma and El Hierro consist almost entirely of volcanic extrusives and intrusives of basaltic composition, with only minor pyroclastic intervals (Carracedo et al., 1997a, 1999). In contrast, pyroclastic deposits form a significant proportion of the Canadas and younger volcanic sequences preserved on Tenerife (Bryan et al., 1998; Cantagrel et al., 1999). Our images suggest that landslides dominated by basaltic rocks give rise to a coarse-grained breccia (Fig. 10); this results in a transport process dominated by grain to grain collision, i.e., a debris avalanche mechanism. Pyroclastic material is much more friable, giving a finer grained breccia (Fig. 11) and probably a significant proportion of fine-grained matrix; this results in a debris flow mechanism, with the flow supported by elevated pore pressure. However, the large blocks at the margins of the Icod flow may be evidence that it was initiated as a debris avalanche, converting to a debris flow as disaggregation took place during transport. The large blocks might have survived either because they are composed of less friable basaltic material, or because they were carried as rafts by the flow. Such rafted blocks typically accumulate at flow margins (Iverson, 1997; Johnson, 1970; Major and Iverson, 1999).

Experimental studies of coarse-grained debris flows confirm that the grain-size of the debris is fundamental in determining flow characteristics. In particular, the addition of even small amounts ($<$ 10%) of fine-grained material greatly increases the runout of coarse-grained flows, because "the fines help sustain high pore pressures that reduce frictional resistance", thus increasing runout distance (Iverson, 1997). In the experimental studies, lobate, relatively short-runout flow deposits, similar in plan view to the El Golfo landslide deposit, are rapidly deposited immediately in front of the flume channel mouth by partially saturated flows. These are considered to be analogous to Canarian debris avalanches, which cannot sustain high pore pressures because of their coarse grain-size and corresponding high permeability. In contrast, saturated experimental flows, particularly those containing some fine-grained material, form elongate bodies with relatively long runout. These are considered to be analogous to the Icod debris flow, which is considered to have been supported by excess pore pressure which was maintained over an extended period during flow emplacement, allowing the typical debris flow characteristics to develop. This would indicate that the landslide contained a significant proportion of fine-grained matrix.

4.3.3. Relationships between slumps, debris avalanches, debris flows and turbidity currents

In the case of the El Golfo landslide, there is considerable evidence to suggest that a single landslide event on the flank of El Hierro gave rise to three discrete flow phases: debris avalanche, a sedimentary debris flow and a turbidity current (Masson et al., 1996, 1998; Urgeles et al., 1997). Both the debris flow deposit and the turbidite contain a complex sedimentary assemblage consisting of mixed volcaniclastics and pelagic slope sediments (Masson et al., 1997; Rothwell et al., 1992). The sedimentary debris flow appears to have been triggered by loading of the sediments on the lower slopes of the island (Masson, 1996; Masson et al., 1998; Roberts and Cramp, 1996). The origin of the turbidity current is less certain, but Masson (1998) suggested that it may have been sourced directly from the debris avalanche, rather than from the debris flow. A similar relationship between the Icod debris avalanche and a volcaniclastic turbidite seen both in the Madeira Abyssal Plain to the west of the Canaries (Rothwell et al., 1992) and the Agadir Basin to the north (Wynn et al., in press), is suggested by the similarity in age between avalanche deposit and turbidite (Fig. 15), although in this case we have no direct evidence linking the two deposits. However, it does support a general relationship between island flank landslides and turbidites in distal basins (Masson, 1996).

In contrast, there is no evidence that debris avalanches generally trigger sedimentary debris flows, since the El Golfo avalanche/Canary debris flow is the only known example of this relationship. The large age difference the El Julan slump/avalanche and the Saharan debris flow (Fig. 15) indicates that despite their partial overlap (Fig. 12), there is no direct relationship between them. However, the presence of volcaniclastic debris from the El Julan landslide in the sub-seafloor succession overridden by the Saharan debris flow was clearly a major factor controlling the exceptional runout of the Saharan flow (Gee et al., 1999).

The turbidite record of the Madeira Abyssal Plain shows a history of volcaniclastic turbidite emplacement extending back to 17 Ma, but with a large increase in the number of such turbidites at about 7 Ma, possibly corresponding to the first turbidites derived from the growing island of Tenerife (Weaver et al., 1998). On average, one volcaniclastic turbidite has reached the abyssal plain approximately every 100 ka during the last 7 Ma, although the distribution of turbidites in time is irregular (Fig. 15). If each turbidite corresponds to a landslide on the islands, then the total of 80 volcaniclastic flows recorded at ODP site 951 in the last 7 Ma represents the minimum number of landslides to have affected the Canaries during that period.

4.3.4. Landslide versus volcanic formation for Canarian 'caldera'

The contributions of vertical and lateral collapse to the genesis of the caldera-like depressions on the western Canary Islands has been extensively discussed in the recent scientific literature. Evidence from El Hierro and La Palma overwhelmingly favours a landslide origin for the depressions and embayments on these islands, in that there is no evidence for the vertical tectonics or the large volumes of pyroclastic material normally associated with caldera-forming eruptions (Carracedo et al., 1997a,b, 1999). In addition, there is abundant evidence for landslide deposits offshore (Figs. 4, 5, 9, 10, 13 and 16) (Masson, 1996; Masson et al., 1998; Urgeles et al., 1997, 1999).

On Tenerife, most of the debate has centred on the origin of the Canadas Caldera. Building on the early work of Navarro and Coello (1989), and with the support of several studies which show an extensive field of landslide debris offshore (Teide Group, 1997; Watts and Masson, 1995, 1998, 2001), several recent papers have proposed that the Canadas Caldera is primarily a complex landslide scar, created by several episodes of lateral collapse (Ancochea et al., 1998, 1999; Cantagrel et al., 1999; Carracedo, 1994; Watts and Masson, 1998, 2001). According to these authors, volcanic activity is mainly responsible for the construction of (unstable) volcanic edifices, which are then destroyed by lateral collapse, i.e., landsliding, rather than by vertical collapse related to caldera formation. The contrasting view is that the Canadas Caldera formed during a "complex sequence of vertical collapse events" (Marti et al., 1997) associated with explosive volcanic activity (Booth, 1973; Bryan et al., 1998; Marti et al., 1990, 1994, 1997; Ridley, 1971). In this scenario, lateral collapse of the north wall of the caldera occurred only after it was weakened by the vertical collapse.

Our interpretation of the Canadas Caldera is that its structure, as observed at the present day, is entirely the product of repetitive large-scale landslides. The southern wall of the Canadas Caldera is the headwall of this landslide complex. True caldera-forming volcanic eruptions, if they occurred, have had a relatively minor influence on the overall development of the caldera structure, and all evidence for such eruptions has been removed from the area of the caldera by landsliding (Ancochea et al., 1999). The principal lines of evidence supporting our interpretation can be summarised as follows:

(1) The absence of a northern caldera wall and the lack of surface or subsurface geological evidence that this hypothetical structure ever existed (Carracedo, 1994; Navarro and Coello, 1989).

(2) The huge volumes of landslide debris mapped offshore require a source area encompassing the whole area of the caldera, rather than an area limited to its northern wall (Watts and Masson, 1995, 1998).

(3) Cross-sections through the north flank of Tenerife are entirely consistent with a failure surface which can be traced beneath the caldera floor to its southern rim (Fig. 21) (Cantagrel et al., 1999; Navarro and Coello, 1989; Watts and Masson, 2001).

(4) Pyroclastic deposits on the southern flank of Tenerife are not unequivocal evidence of a caldera-forming eruption, as the presence of "discordances and palaeosoils" appears to suggest a long period of eruption (Carracedo, 1994). Where deposits created by explosive volcanic eruptions coincide in time with landslide episodes, for example at 170 ka (Ancochea et al., 1999), this may also be interpreted as due to the release of pressure on a magma chamber by the landslide.

4.4. Geohazards

Potential geohazards in the Canary Islands include volcanic activity, seismicity, landslides and tsunamis. Although the history of volcanic activity and associated seismicity is outside the scope of this paper, all the islands except La Gomera have seen volcanic activity in historic times (Carracedo et al., 1998),

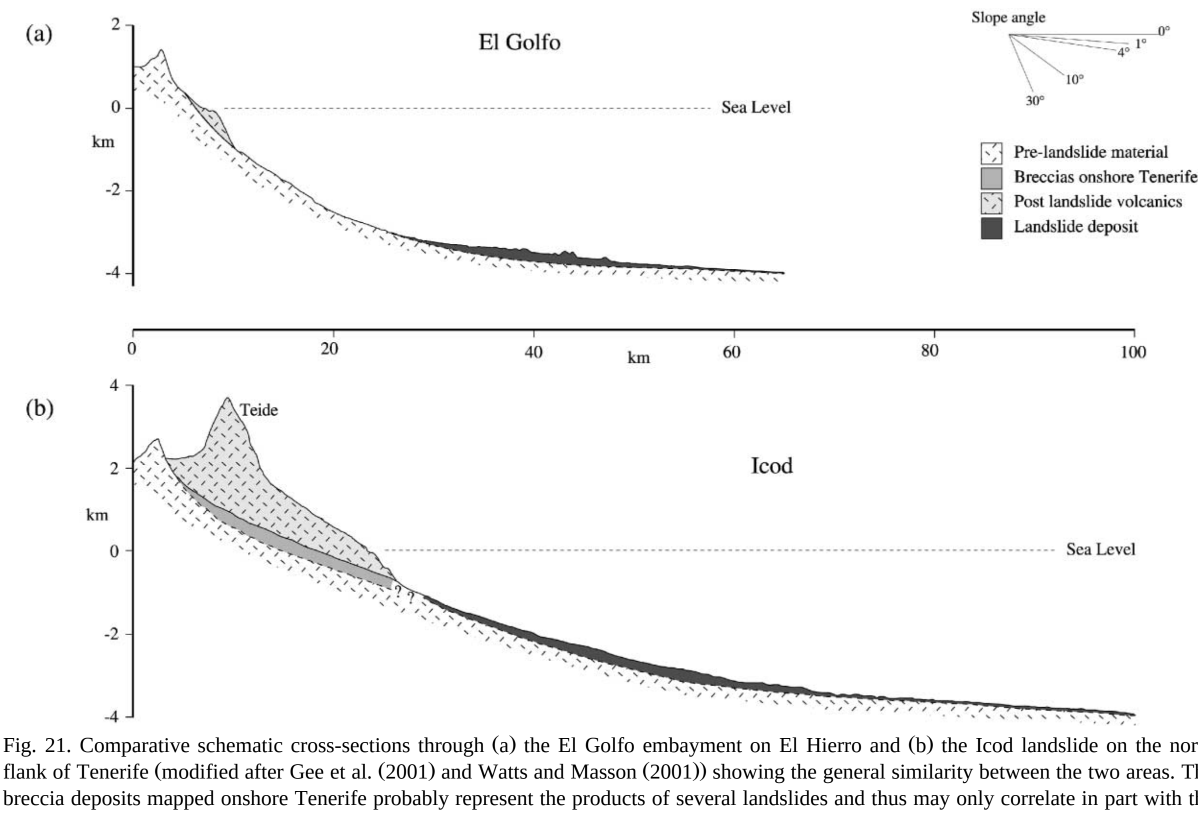

Fig. 21. Comparative schematic cross-sections through (a) the El Golfo embayment on El Hierro and (b) the Icod landslide on the north flank of Tenerife (modified after Gee et al. (2001) and Watts and Masson (2001)) showing the general similarity between the two areas. The breccia deposits mapped onshore Tenerife probably represent the products of several landslides and thus may only correlate in part with the landslide deposit offshore (Watts and Masson, 2001). The lesser slope curvature of the Icod profile relative to El Golfo appears to be typical of older island slopes which have experienced multiple landslide episodes (see Fig. 7).

and it is clear that hazards associated with such activity pose the greatest risks for the inhabited areas of the islands. The hazard, and more particularly the risk, posed by giant landslides is more difficult to evaluate. Clearly, a landslide which removes up to 25% of the subaerial volume of an island, such as the El Golfo landslide on El Hierro, would have a catastrophic impact on the island itself. However, the risk from such landslides is relatively low, because they occur on average only once in 100,000 years over the whole island chain, and only once every 300,000 years for each individual island. Landslides occur almost exclusively in the early, “shield-building” stage of island growth, when volcanic production rates are highest (Carracedo et al., 1998, 1999). Thus risks from landslides primarily affect El Hierro, La Palma and Tenerife. Geologically young landslides are not recognised on any of the older, “post-shield-stage” islands, such as Gran Canaria, although this island had a lengthy history of landsliding between 15 and 3.5 Ma (Krastel et al., 2001).

The Cumbre Vieja volcanic ridge, occupying the southern half of La Palma, has been identified as the most rapidly growing volcano in the Canaries (Carracedo et al., 1999). This area has been the site of intense volcanic activity in the last 20,000 years and almost half its surface area is covered with lava < 7000 years old. It has been suggested that a fault system which developed near the crest of the ridge during an eruption in 1949 is a sign of stresses that will eventually lead to collapse of the ridge and the generation of a westward-directed landslide (Carracedo, 1996; Day et al., 1997). However, we have no means of predicting the timing of any future collapse, or even of being certain that such a collapse will occur at all.

A tsunami generated by a landslide on the Canary Islands could have a widespread impact, both on the adjacent islands and on a wider geographic scale. No evidence for tsunamis associated with known landslides has been discovered to date, although this may be due mainly to the steep rocky nature of the island topography which does not provide suitable geological environments for the preservation of tsunami deposits. In the Hawaiian Islands, landslide-related tsunami with run-ups exceeding 300 m on adjacent islands have been recognised (Moore and Moore, 1984), although it should be recognised that Hawaiian landslides can be up to an order of magnitude larger than those discovered around the Canaries (Moore and Moore, 1984; Moore et al., 1989). Modeling of tsunamis based on the size and character of known landslides in the Canaries is an extremely uncertain science, because we know little of the key parameters which would influence tsunami size. These include the initial failure mechanism (e.g., single catastrophic failure or multiple retrogressive failure) and the rate of movement in the early stages of landsliding, when the subaerial part of the landslide is entering the ocean.

5. Conclusions

Landslides are an important process in the evolution of the western Canary Islands of El Hierro, La Palma and Tenerife. Our main conclusions are as follows:

(1) At least 14 landslides can be recognised, with most of the recognised landslides less than 1 Ma in age.

(2) Many young landslides can be identified from a combination of a landslide scar onshore and a field of blocky debris offshore. However, even where this primary evidence has been buried by later volcanism or sedimentation, the distinctive topographic profile of landslide areas is diagnostic.

(3) Debris avalanches are the most important landslide type in the western Canaries, although slumps and debris flows also occur. Compared to the Hawaiian Islands, the relative lack of slump activity in the Canaries probably relates to the lower topographic anomaly associated with the weaker Canarian mantle plume.

(4) Individual landslides have volumes in the range of 50–500 km^3, can cover several thousand km^2 of seafloor, and have runout distances of up to 130 km from source. Landslide volume can be difficult to estimate, since the base of landslide deposits is rarely imaged by seismic techniques.

(5) Debris avalanches have an overall lobate shape with a relatively narrow headwall and shute above 3000 m water depth and a broad lobate depositional lobe, typically composed of randomly distributed blocky debris, below 3000 m. However, a gradation to more elongate deposits showing evidence of more

structured flow processes is seen in some areas, particularly north of Tenerife. This probably relates to differences, between islands, in the volcanic materials involved in landsliding.

(6) Landslide locations are primarily controlled by the locations of volcanic rift zones on the islands, with landslides propagating perpendicular to the rifts. This does not, however, explain why some islands flanks have been the site of multiple landslides when others have been stable through long periods of island evolution. Buttressing by adjacent topography, focussing of volcanism in old landslide scars and formation of weak layers which can be re-exploited by later landslides may be important 'local' factors controlling landslide location.

(7) Landslides are a response to long-term volcanic build-up to a point where the load on the volcanic edifice exceeds that which it can support. Very little is known of the mechanisms which actually trigger failure, but dyke intrusion, pore pressure increases due to intrusion and seismicity may be important.

(8) Landsliding is shown to be the key process in forming large erosional valleys and 'caldera' on the Canary islands.

References

Ancochea, E. et al., 1990. Volcanic evolution of the island of Tenerife (Canary Islands) in the light of new K–Ar data. Journal of Volcanology and Geothermal Research 44, 231–249.

Ancochea, E., Cantagrel, J.M., Fuster, J.M., Huertas, M.J., Arnaud, N.O., 1998. Vertical and lateral collapses on Tenerife (Canary Islands): comment. Geology 26, 861–862.

Ancochea, E. et al., 1999. Evolution of the Canadas edifice and its implication for the origin of the Canadas Caldera (Tenerife, Canary Islands). Journal of Volcanology and Geothermal Research 88, 177–199.

Booth, B., 1973. The Granadilla pumice of southern Tenerife, Canary Islands. Proceedings of the Geologists Association 84, pp. 353–370.

Bravo, T., 1962. El circo de las Canadas y sus dependcias. Boletin de la Real Sociedad Espanola de Historia Natural 60, 93–108.

Bryan, S.E., Marti, J., Cas, R.A.F., 1998. Stratigraphy of the Bandas del Sur formation: an extracaldera record of Quaternary phonolitic explosive eruptions from the Las Canadas edifice, Tenerife (Canary Islands). Geological Magazine 135, 605–636.

Cantagrel, J.M., Arnaud, N.O., Ancochea, E., Fuster, J.M., Huertas, M.J., 1999. Repeated debris avalanches on Tenerife and genesis of Las Canadas caldera wall (Canary Islands). Geology 27, 739–742.

Carracedo, J.C., 1994. The Canary Islands: an example of structural control on the growth of large oceanic-island volcanoes. Journal of Volcanology and Geothermal Research 60, 225–241.

Carracedo, J.C., 1996. A simple model for the genesis of large gravitational landslide hazards in the Canary Islands. In: McGuire, W.J., Jones, A.P., Neuberg, J. (Eds.), Volcano Instability on the Earth and Other Planets. Geological Society Special Publication, Geological Society of London, London, pp. 125–135.

Carracedo, J.C., Day, S., Guillou, H., Torrado, F.J.P., 1997a. El Hierro Geological Excursion Handbook. Estacion Volcanologica de Canarias and the Universidad de Las Palmas, Tenerife / Gran Canaria, 43 pp.

Carracedo, J.C., Day, S.J., Guillou, H., Gravestock, P.J., 1997b. Geological Map of Cumbre Nueva Volcano, La Palma, Canary Islands. Estacion Volcanologica de Canarias, La Laguna, Tenerife, Canary Islands.

Carracedo, J.C. et al., 1998. Hotspot volcanism close to a passive margin: the Canary Islands. Geological Magazine 135, 591–604.

Carracedo, J.C., Day, S., Guillou, H., Gravestock, P., 1999. Later stage of volcanic evolution of La Palma, Canary Islands: rift evolution, giant landslides, and the genesis of the Caldera de Taburiente. Bulletin of the Geological Society of America 111, 755–768.

Cochonat, P. et al., 1990. Importance des dépôts gravitaires dans la mise en place d'un système volcano-sédimentaire sous-marin (Volcano de la Fournaise, Ile de la Réunion). Compte Rendu Academie des Sciences 311 (Series II), 679–686.

Dade, W.B., Huppert, H.E., 1998. Long-runout rockfalls. Geology 26, 803–806.

Day, S.J., 1996. Hydrothermal pore pressure and the stability of porous, permeable volcanoes. In: McGuire, W.J., Jones, A.P., Neuberg, J. (Eds.), Volcano Instability on the Earth and Other Planets. Geological Society Special Publication, Geological Society of London, London, pp. 77–93.

Day, S.J., Carracedo, J.C., Guillou, H., 1997. Age and geometry of an aborted rift collapse: the San Andreas fault system, El Hierro, Canary Islands. Geological Magazine 134, 523–537.

Elsworth, D., Voight, B., 1995. Dike intrusion as a trigger for large earthquakes and the failure of volcano flanks. Journal of Geophysical Research 100, 6005–6024.

Embley, R.W., 1976. New evidence for the occurrence of debris flow deposits in the deep sea. Geology 4, 371–374.

Embley, R.W., 1982. Anatomy of some Atlantic margin sediment slides and some comments on ages and mechanisms. In: Saxov, S., Neiuwenhuis, J.K. (Eds.), Marine Slides and Other Mass Movements. Plenum Press, New York, pp. 189–214.

Fiske, R.S., Jackson, E.D., 1972. Orientation and growth of volcanic rifts: the effect of regional structure and gravitational stresses. Proceedings of the Royal Society of London, Series A 329, pp. 299–326.

Fuster, J.M., Hernan, F., Cendrero, A., Coello, J., Cantagrel, J.M., Ancochea, E., Ibarrola, E., 1993. Geocronologia de la Isla de El Hierro (Islas Canarias). Boletin Real Sociedad Espanola Historia Natural (Geologia) 88, 85–97.

Gee, M.J.R., 1999. The catastrophic collapse of oceanic islands and the mechanics of long runout debris flows: examples from the NW African margin. PhD Thesis, University of Oxford, 243 pp.

Gee, M.J.R., Masson, D.G., Watts, A.B., Allen, P.A., 1999. The Saharan Debris Flow: an insight into the mechanics of long runout debris flows. Sedimentology 46, 317–335.

Gee, M.J.R., Watts, A.B., Masson, D.G., Mitchell, N.C., 2001. Landslides and the evolution of El Hierro in the Canary Islands. Marine Geol. 177, 271–293.

Hampton, M.A., Lee, H.J., Locat, J., 1996. Submarine landslides. Reviews of Geophysics 34, 33–59.

Holcomb, R.T., Searle, R.C., 1991. Large landslides from oceanic volcanoes. Marine Geotechnology 10, 19–32.

Iverson, R.M., 1997. The physics of debris flows. Reviews of Geophysics 35, 245–296.

Johnson, A.M., 1970. Physical Processes in Geology. Freeman, Cooper and Co., San Francisco, 577 pp.

Krastel, S., Schmincke, H.-U., Jacobs, C.L., Rihm, R., Le Bas, T.P., Alibés, B., 2001. Submarine landslides around the Canary Islands. Journal of Geophysical Research 106, 3977–3997.

Labazuy, P., 1996. Recurrent landslide events on the submarine flank of Piton de la Fournaise volcano (Reunion Island). In: McGuire, W.J., Jones, A.P., Neuberg, J. (Eds.), Volcano Instability on the Earth and Other Planets. Special Publication. Geological Society of London, London, pp. 295–306.

Lipman, P.W., Normark, W.R., Moore, J.G., Wilson, J.B., Gutmacher, C.E., 1988. The giant Alika debris slide, Mauna Loa, Hawaii. Journal of Geophysical Research 93, 4279–4299.

Major, J.J., 1997. Depositional processes in large-scale debris-flow experiments. Journal of Geology 105, 345–366.

Major, J.J., Iverson, R.M., 1999. Debris-flow deposition: effects of pore-fluid pressure and friction concentrated at flow margins. Geological Society of America Bulletin 111, 1424–1434.

Marti, J., 1998. Comment on A giant landslide on the north flank of Tenerife, Canary Islands, by A.B. Watts and D.G. Masson. Journal of Geophysical Research 103, 9945–9947.

Marti, J., Mitjavila, J., Arana, V., 1990. Stratigraphy and K–Ar ages of the Diego Hernandez wall and their significance on the La Canadas Caldera formation (Tenerife, Canary Islands). Terra Nova 2, 148–153.

Marti, J., Mitjavila, J., Arana, V., 1994. Stratigraphy, structure and geochronology of the La Canadas caldera (Tenerife, Canary Islands). Geological Magazine 131, 715–727.

Marti, J., Hurlimann, M., Ablay, G.A., Gudmundsson, A., 1997. Vertical and lateral collapses on Tenerife (Canary Islands) and other volcanic ocean islands. Geology 25, 879–882.

Masson, D.G., 1996. Catastrophic collapse of the flank of El Hierro about 15,000 years ago, and the history of large flank collapses in the Canary Islands. Geology 24, 231–234.

Masson, D.G., Kidd, R.B., Gardner, J.V., Huggett, Q.J., Weaver, P.P.E., 1992. Saharan continental rise: sediment facies distribution and sediment slides. In: Poag, C.W., de Graciansky, P.C. (Eds.), Geologic Evolution of Atlantic Continental Rises. Van Nostrand Reinhold, New York, pp. 327–343.

Masson, D.G., Huggett, Q.J., Brunsden, D., 1993. The surface texture of the Saharan Debris Flow deposit and some speculations on debris flow processes. Sedimentology 40, 583–598.

Masson, D.G., Kenyon, N.H., Weaver, P.P.E., 1996. Slides, debris flows and turbidity currents. In: Summerhayes, C.P., Thorpe, S.A. (Eds.), Oceanography, an Illustrated Guide. Manson Publishing, London, pp. 136–151.

Masson, D.G., van Neil, B., Weaver, P.P.E., 1997. Flow processes and sediment deformation in the Canary Debris Flow on the NW African Rise. Sedimentary Geology 110, 163–179.

Masson, D.G., Canals, M., Alonso, B., Urgeles, R., Hühnerbach, V., 1998. The Canary Debris Flow: source area morphology and failure mechanisms. Sedimentology 45, 411–432.

McGuire, W.J., 1996. Volcano instability: a review. In: McGuire, W.J., Jones, A.P., Neuberg, J. (Eds.), Volcano Instability on the Earth and Other Planets. Geological Society Special Publication, Geological Society of London, London, pp. 1–23.

McGuire, W.J. et al., 1997. Correlation between rate of sea-level change and frequency of explosive volcanism in the Mediterranean. Nature 389, 473–476.

Moore, J.G., Moore, G.W., 1984. Deposit from a giant wave on the island of Lanai, Hawaii. Science 226, 1312–1315.

Moore, J.G. et al., 1989. Prodigious submarine landslides on the Hawaiian Ridge. Journal of Geophysical Research 94, 17465–17484.

Moore, J.G., Normark, W.R., Holcomb, R.T., 1994. Giant Hawaiian landslides. Annual Review of Earth and Planetary Sciences 22, 119–144.

Mulder, T., Cochonat, P., 1996. Classification of offshore mass movements. Journal of Sedimentary Research 66, 43–57.

Navarro, J.M., Coello, J., 1989. Depressiones formadas por deslizamiento gravitacional en Tenerife, ESF meeting on Canarian Volcanism, Lanzarote, Canary Islands. European Science Foundation.

Ollier, G., Cochonat, P., Lenat, J.F., Labazuy, P., 1998. Deep-sea volcaniclastic sedimentary systems: an example from La Fournaise volcano, Reunion Island, Indian Ocean. Sedimentology 45, 293–330.

Palacios, D., 1994. The origin of certain wide valleys in the Canary Islands. Geomorphology 9, 1–18.

Rasa, R., Azzaro, R., Leonard, O., 1996. Aseismic creep on faults and instability at Mount Etna volcano, Sicily. In: McGuire, W.J., Jones, A.P., Neuberg, J. (Eds.), Volcano Instability on the Earth and Other Planets. Geological Society Special Publication, Geological Society of London, London, pp. 179–192.

Ridley, W.I., 1971. The origin of some collapse structures in the Canary Islands. Geological Magazine 108, 477–484.

Roberts, J.A., Cramp, A., 1996. Sediment stability on the western flanks of the Canary Islands. Marine Geology 134, 13–30.

Rothwell, R.G., Pearce, T.J., Weaver, P.P.E., 1992. Late Quaternary evolution of the Madeira Abyssal Plain, NE Atlantic. Basin Research 4, 103–131.

Rust, D., Neri, M., 1996. The boundaries of large-scale collapse on the flanks of Mount Etna, Sicily. In: McGuire, W.J., Jones, A.P., Neuberg, J. (Eds.), Volcano Instability on the Earth and Other Planets. Geological Society Special Publication, Geological Society of London, London, pp. 193–208.

Scheidegger, A.E., 1973. On the prediction and reach of catastrophic landslides. Rock Mechanics 5, 231–236.

Simm, R.W., Weaver, P.P.E., Kidd, R.W., Jones, E.J.W., 1991. Late Quaternary mass movement on the lower continental rise and abyssal plain off Western Sahara. Sedimentology 38, 27–40.

Teide Group, 1997. Morphometric interpretation of the northwest and southeast slopes of Tenerife, Canary Islands. Journal of Geophysical Research 102, 20325–20342.

Tibaldi, A., 1996. Mutual influence of dyking and collapses at Stromboli volcano, Italy. In: McGuire, W.J., Jones, A.P., Neuberg, J. (Eds.), Volcano Instability on the Earth and Other Planets. Special Publication. Geological Society of London, London, pp. 55–63.

Urgeles, R., Canals, M., Baraza, J., Alonso, B., Masson, D.G., 1997. The most recent megaslides on the Canary Islands: the El Golfo Debris Avalanche and the Canary Debris Flow, west El Hierro Island. Journal of Geophysical Research 102, 20305–20323.

Urgeles, R., Masson, D.G., Canals, M., Watts, A.B., Le Bas, T., 1999. Recurrent large-scale landsliding on the west flank of La Palma, Canary Islands. Journal of Geophysical Research 104, 25331–25348.

Vogt, P.R., Smoot, N.C., 1984. The Geisha Guyots: multibeam bathymetry and morphometric interpretation. Journal of Geophysical Research 89, 11085–11107.

Watts, A.B., Masson, D.G., 1995. A giant landslide on the north flank of Tenerife, Canary Islands. Journal of Geophysical Research 100, 24487–24498.

Watts, A.B., Masson, D.G., 1998. Reply to comment on "A giant landslide on the north flank of Tenerife, Canary Islands", by J. Marti. Journal of Geophysical Research 103, 9949–9952.

Watts, A.B., Masson, D.G, 2001. New sonar evidence for recent catastrophic collapses of the north flank of Tenerife, Canary Islands. Bulletin of Volcanology 63, 8–19.

Weaver, P.P.E. et al., 1998. Neogene turbidite sequence on the Madeira Abyssal Plain: basin filling and diagenesis in the deep ocean. In: Weaver, P.P.E., Schminke, H.U., Firth, J.V., Duffield, W. (Eds.), Proceeding of the Ocean Drilling Programme, Scientific Results. Ocean Drilling Programme, College Station, Texas, pp. 619–634.

Wessel, P., Smith, W.H.F., 1991. Free software helps map and display data. EOS, Transactions of the American Geophysical Union 72, 441–446.

White, R.S., 1993. Melt production rates in mantle plumes. Philosophical Transactions of the Royal Society of London, A 342, 137–153.

Wynn, R.B., Weaver, P.P.E., Stow, D.A.V., Masson, D.G., in press. Turbidite depositional architecture across three interconnected deep-water basins on the northwest African margin. Sedimentology.