Susceptibility of mid-ocean ridge volcanic islands and seamounts to large-scale landsliding

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[1] With a view to assessing the incidence of large-scale landsliding, a morphologic database was created for volcanic islands and seamounts on young oceanic lithosphere. The database included 44 mid-ocean ridge seamounts, Jasper seamount and the islands of Ascension, Bouvet, Guadalupe, and several of the Galapagos and Azores Islands, supplemented with published reports from a further five volcanic edifices. The data reveal that major landslides are common on edifices taller than 2500 m but are rare in shorter edifices, implying a threshold of instability at around 2500 m. A number of causes of this threshold are discussed. For example, many structures taller than 2500 m are, or were originally, volcanic islands, and therefore their flanks probably include extensive weak hyaloclastite built up from lava-sea interactions around coasts. Compaction of hyaloclastites in larger edifices lead to regions of low permeability, which may help to explain the more deeply seated slope failures. It is intriguing that the threshold also coincides with the edifice height at which volcanic ridges become observable, a stage at which dike intrusion is probably common because slope oversteepening or excess pore pressure associated with dike intrusions have been proposed elsewhere as landslide triggers. The current indications of landsliding reveal no observable relation to rainfall, as landslides occur equally in wet and dry climates, and no relation to tectonic setting, as there are relatively few major landslides in the seismically active Azores group.

INDEX TERMS: 3035 Marine Geology and Geophysics: Mid-ocean ridge processes; 3022 Marine Geology and Geophysics: Marine sediments—processes and transport; 8499 Volcanology: General or miscellaneous;

KEYWORDS: landslides, volcanic debris avalanche, slope stability, multibeam sonar, side-scan sonar


1. Introduction

[2] Volcanic islands, like many terrestrial volcanoes, can become unstable and fail causing giant landslides. Surveys of their submerged flanks with seafloor imaging sonars have frequently revealed landslides of a range of sizes [Holcomb and Searle, 1991; Lenat et al., 1989; Masson et al., 2002; Moore et al., 1989]. In this paper, landslides are classified as suggested by Moore et al. [1989], following Varnes [1978], as debris avalanches (fast moving granular rock flows) and slumps (slowly or intermittently moving bodies of rock or debris). Figure 1 shows the morphology of an example debris avalanche headwall, eroded chute and deposit of El Hierro of the Canary Islands [Masson et al., 2002], and Figure 2 shows its corresponding slope profiles described in more detail later. The origin of the instability is the process of volcanic growth which creates steep, unstable slopes by adding erupted material to the tops of slopes and ground movements, and slope oversteepening associated with intrusions. In addition, a wide range of internal and external factors have been proposed to cause or contribute to failure, which have recently been reviewed by Keating and McGuire [2000]. Unfortunately, quantitatively resolving the relative importance of these factors is difficult with proper stability modeling, because representative rock samples are rarely available for geotechnical stability analysis [Hurlimann et al., 1999] owing to their inaccessibility at depths of potential failure surfaces and the heterogeneous nature of volcanic materials. The conditions leading up to or contributing to prehistoric landslides (rainfall, water table, seismicity, volcanic activity, etc.) are also poorly known.

[3] An alternative approach is to study volcanic islands from different tectonic and environmental settings, where contrasting landsliding character or incidence between areas might provide clues to the reasons for failure. Mitchell et al. [2002], for example, compared the Canary and Hawaiian Islands and noted that, whereas Hawai’i shows evidence for several deep-seated slumps in addition to debris avalanches, the Canaries show almost exclusively evidence for debris avalanches, with only minor evidence for slumping. This difference might ultimately reflect the greater volcanic building rate of Hawai’i, for example, if slump systems are driven by plastic flow in hot olivine cumulates within the island [Carracedo, 1999; Clague and Denlinger, 1994]. Other differences include a greater block size within
Hawaiian landslide deposits [Mitchell et al., 2002] and a greater frequency of failure in the Canaries, possibly due to longer periods of volcanism on each island and more abundant pyroclastic material there [Krastel et al., 2001]. These observations suggest that different landsliding character and incidence could emerge from studies of seamounts and islands of different sizes, tectonic and climatic settings, and that such information could provide clues to the underlying causes of instability.

Whereas small seamounts on mid-ocean ridges are relatively simple truncated cones [Batiza and Vanko, 1983], larger seamounts, guyots, and volcanic islands are complex stellate features, with multiple volcanic centers, protruding ridges due to growth along volcanic rift zones and embayments due to landslides [Vogt and Smoot, 1984]. Studying the transition between these two end-members is difficult because it typically occurs when the edifice is 2000–4000 m tall and submerged [Mitchell, 2001]. Owing to the shallow regional topography at mid-ocean ridges, however, many volcanic edifices within the 2000–4000 m height range rise above sea level and are more accessible. The geology and age of volcanic islands is also commonly better known than for seamounts, so studies of islands may shed light on the processes affecting this change in gross structure.

The mid-ocean ridge volcanic islands and seamounts have not been previously studied systematically for major landslides, although various authors have noted their presence on individual islands as related below. The following paper describes a compilation of data and information available from the areas in Figure 3 and from the sources given in Appendix A (which also describes data quality and processing). The data are shown in Figures 4–13 as

**Figure 1.** The bathymetry of El Hierro Island of the Canaries, showing a deep embayment and associated debris field on its north side created by the El Golfo debris avalanche, and further embayments to the southwest and southeast caused by older collapse events [Gee et al., 2001; Holcomb and Searle, 1991; Masson et al., 2002; Urgeles et al., 1998]. Data from the two outlined areas were used to generate Figure 2. (Data courtesy of D.G. Masson and A.B. Watts.) See color version of this figure at back of this issue.
2. Criteria for Identifying Landslides

Landslides were identified based on known characteristics of subaerial landslides [Varnes, 1978] and volcanic landslides in particular [Siebert, 1984; Ui et al., 1986]. Much larger landslides around volcanic ocean islands also show similar features to the smaller landslides [Moore et al., 1989]. Because common landslide morphologies are observed at extremes of scale, a common set of identification criteria can be used for the intermediate-sized mid-ocean ridge edifices. The features that Varnes [1978] describes for subaerial debris avalanches (fast moving landslides of disaggregated material) are well illustrated by the El Golfo embayment and associated submarine deposit in Figure 1 [Masson, 1996]. As marked on the figure, the avalanche has left a steep head scarp which encompasses the El Golfo valley, and an elongate chute which is topographically depressed compared to the surrounding terrain, with inward facing scarps along its margins. Subaerial debris avalanche structures can have V-shaped head scarps in plan view [Varnes, 1978], whereas those in ocean islands are typically U-shaped (Figure 1). Other features can include longitudinal flow structures and transverse pressure ridges, which are also observed in ocean island debris avalanche structures [Watts and Masson, 1995; Lipman et al., 1988]. In volcanic island debris avalanche structures, the chute is commonly smoother than the surrounding terrain if the surrounding terrain has a rugged volcanic morphology. A debris lobe or debris field containing the avalanche deposit can form a hummocky terrain of transported blocks (such an area is just visible in the top of Figure 1). Such debris lobes typically have a speckled pattern in long-range side-scan sonar records owing to the backscattering contrast between blocks in the avalanche and their surrounding sediment [Holcomb and Searle, 1991].

Debris avalanches in subaerial volcanoes can excavate the peaks of the volcanoes to leave amphitheaters open to one side and associated deposits [Siebert, 1984; Ui et al., 1986]. Sonar mapping around the submarine flanks of volcanic islands has revealed similar features. Examples include those of the tall volcanic ocean islands of Hawai‘i [Moore et al., 1989; Lipman et al., 1988], the Canary Islands [Masson et al., 2002], and Reunion [Lenat et al., 1989]. Although there are complications (e.g., differing block size [Mitchell et al., 2002]), debris avalanche structures appear similar in these different areas. The island arc volcanoes of Stromboli [Kokelaar and Romagnoli, 1995], the Lesser Antilles [Deplus et al., 2001], and Oshima-Oshima, Japan [Satake and Kato, 2001], show subaerial or submarine U-shaped embayments, topographically depressed elongate chutes and debris fields that are similar to those of the ocean islands. Debris avalanche chutes in the Canary Islands have nearly perfectly exponential shapes in profile [Gee et al., 2001] in contrast to the Hawaiian landslides which apparently do not always show this form [Moore et al., 1989].

Subaerial slumps can possess some of the above features, for example, movement can create an upper depressed region (zone of depletion) bounded by a main or head scarp and a corresponding lower elevated region (zone of accumulation [Varnes, 1978]). However, slumps around Hawai‘i are broader along slope than debris avalanche structures and generally lack a characteristic chute [Moore et al., 1989, 1994; Moore and Chadwick, 1995; Smith et al., 2002]. In volcanic islands, the slump headwall is commonly obscured by later volcanism. The Hilina slump system of Kilauea, recently reviewed by Lipman et al. [2002], is the classic example of a large-scale volcanic

![Figure 2](attachment:image.png)

**Figure 2.** (a) The graphs show the average slope for the El Golfo debris avalanche structure (dotted lines) and for the west unfailed volcanic sector (solid lines) of El Hierro Island for the sectors located in Figure 1. The different lines show the effect of filtering the original data over different spatial scales up to 3000 m. They reveal that the avalanche and volcanic regions are still discriminated in terms of average slope despite the filtering. Hence the reduced slopes of avalanche deposits should be interpretable from sparse sounding data if the avalanches and volcanic regions are similar to those of El Hierro. (b) Corresponding variability of slope (represented by slope standard deviation) is readily decreased by filtering such that the avalanche and volcanic slopes become difficult to discriminate if the data are filtered over >2 km. (Differences in slope roughness should be discernible in single-beam sounding data if the survey lines are spaced less than 2 km apart.)

contoured bathymetry in color in the online version of this manuscript and gray in the print version. For some Azores Islands, sonar and radar image data were also available and are shown alongside the bathymetry. The bathymetry data of the larger edifices are mostly derived from single-beam soundings and are not of such a high quality as the Hawaiian, Canary, and Reunion data. Furthermore, the procedure used here aims to locate primarily the largest landslides, i.e., those that have modified the gross size distribution of landslides. Nevertheless, the results suggest a threshold in major landslide incidence in edifices taller than 2500 m, which could imply an important functional control on edifice stability.
slump and shows topographic evidence for landsliding in the form of normal faults stepping seaward and trending parallel to the coast [Lipman et al., 2002] but these are less prominent than the El Golfo head scarp (Figure 1). The Hilina slump has been suggested to involve movement of the entire flank of the edifice over a decollement close to the edifice’s contact with its underlying oceanic crust [Dieterich, 1988], driven by pressure in magma bodies [Swanson et al., 1976] or cumulates within the volcano [Clague and Denlinger, 1994]. Modeling of recent Global Positioning System (GPS) data indicates movement also occurs higher within the edifice [Cervelli et al., 2002].

Submarine data are important for identifying the thrust faulting of slump toes, in particular the morphology of blind thrusts in sonar data [Smith et al., 1999, 2002] and deformation in seismic reflection images [Morgan et al., 2000].

To illustrate the effect of data spatial resolution, Figures 2a and 2b show the average slope magnitude and standard deviation of slopes, respectively, for the El Golfo debris avalanche (dotted lines) and the west volcanic flank (continuous lines) of El Hierro plotted versus depth. These were calculated from the data in Figure 1 by the method in the work of Mitchell et al. [2002]. The different lines show the effect on slopes after filtering the bathymetry with the filter widths shown. The west area has a rugged morphology of volcanic cones and lava terraces, which is reflected in a high standard deviation. The graphs suggest that volcanic geomorphology should be observable even in low-resolution maps, because the data preserve a higher standard deviation than the El Golfo chute even when filtered over >1 km. With progressively coarser filtering, the two flank types also retain their distinctively different average slopes, although slope magnitude is unfortunately not very helpful because other features of low slope can occur and cross sections of small subaerial nonvolcanic debris avalanche structures [Siebert, 1984] can show no reduction in gradient.

Various types of data were interpreted to provide the landslide indicators in Table 1, taking account of the varied data resolution. The principal indicators were (1) headwall embayment of the submarine flank or island shelf, (2) embayment of the subaerial edifice and/or arcuate normal faults, (3) contrasting surface morphology (a smooth chute missing a typical rugged volcanic morphology of cones and terraces), (4) a topographic debris lobe, and (5) a speckle pattern in side-scan sonar data. The data included published reports, the bathymetry, slope magnitudes, and remote-sensing spaceborne imaging radar-C (SIR-C) and side-scan sonar imagery if available. The indicators are more suitable for identifying debris avalanche structures than slumps (also because the data are inadequate to resolve slump thrust faulting); nevertheless, some possible slumps are noted based

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**Figure 3.** Locations of the volcanic islands described here. Plate boundaries (solid lines) and 30 Ma seafloor isochrons (dotted lines) are from Muller et al. [1997].
on observed headwall faults and topography. Furthermore, not all the above criteria may be met for any given debris avalanche feature. For example, some known deposits lack an obvious chute (e.g., La Palma [Urgeles et al., 1999]), a subaerial embayment can be infilled by later volcanism or modified by erosion, and a hummocky debris field can be obscured by sediment. Embayments can be confused with caldera collapse structures. Coastal erosion creates steep sea cliffs around volcanic islands, so the presence of steep scarps near sea level is not useful if they are the only landslide indicators. In the case of small seamounts, their margins are steep talus slopes [Fornari et al., 1984], so a strong contrast in surface morphology or roughness (indicator 3) is not anticipated. The existence of a topographic debris lobe depends on the topography existing before the avalanche. From all these complications, the final results in Table 1 are therefore biased in part by nature, tending to remove subaerial evidence, etc., and by data quality, tending to lead to landslides being identified where both subaerial and submarine indications exist. Because of the latter bias, the data set only represents the larger landslides that have had a major effect on landslide morphology.

To provide a rough measure of reliability, the presence of the above landslide indicators was noted for each edifice and given in Table 1 in three levels (yes, no, or moderate evidence, or numerically 0, 1, or 0.5, respectively). The confidence level in Table 1 was then calculated by adding these scores. Given the above considerations, however, this confidence level is not fully quantitative and only a general guide to observational quality. Furthermore, the scheme does not address the confidence in observations of where major landslides are absent. A further iteration of this scheme, once more comprehensive multibeam sonar and other data become available, will therefore need to address confidence in both presence and absence of landslides.

3. Island and Seamount Geology and Morphological Observations

Islands and seamounts from the Pacific are described first and then those in the Atlantic and Indian Oceans. In addition, where evidence concerning landslides has been given in publications but the data were not freely available (and hence figures are not provided herein), these edifices are also described. In each map, the large arrow locates the identified landslide embayment. Small open triangles locate the cross sections shown with bold lines in Figure 14, and solid triangles locate the adjacent reference sections. The map insets show the data coverage.

3.1. Small Mid-Ocean Ridge Seamounts

A multibeam sonar data set of 44 mid-ocean ridge seamounts was examined. It comprised data from 8°S on the East Pacific Rise [Cochran et al., 1993] and from the Juan de Fuca Ridge [Hammond, 1997]. Edifices ranged in height from 200 to 1700 m. The edifices showed a range of morphologies from sharp to flat topped [Smith, 1988]. The round, flat-topped morphology has been attributed either to overtopping caldera [Clague et al., 2000a] or to circular eruptions such as from ring dikes [Batiza et al., 1989; Simkin, 1972] or cone sheets [Mitchell, 2001], and hence their eruptive geometry may differ from the larger edifices described later.

Only two seamounts showed evidence of slope failures large enough to embay the seamount flanks and/or summit (Figure 4). The flank topographic gradients are not strongly reduced for the avalanche chutes in Figure 14 as might be expected for small volume landslides. These two seamounts are 1300 and 1400 m tall. Circular embayments were also observed in other seamounts but their circular morphologies are more readily attributed to caldera collapse [Hammond, 1997]. Therefore although deep-tow side-scan sonar images commonly show gullying of the steep flanks of seamounts [Fornari et al., 1984] probably caused by small slope failures and other superficial erosional processes [Tucholke et al., 1997], larger failures sufficient to indent seamount flanks are rare. Considering the spatial resolution of multibeam data for these water depths (100–150 m), the lack of landslides is not an artifact of resolution and indicates that mid-ocean ridge volcanoes are genuinely more stable than the other edifices described below.

3.2. Jasper Seamount

Jasper seamount (Figure 5) lies off the California coast on 24 Ma seafloor, slightly older than Chron C6C.
It grew with a main tholeiitic-transitional shield-building phase dated (Ar-Ar) at 10.0–10.3 Ma, continuing with two later phases, the last a summit alkalic phase as late as 4.1–4.8 Ma [Pringle et al., 1991]. On the basis of magnetic polarity, Gee et al. [1991] suggested that the shield phase may have occurred at 11.5–10.5 Ma on seafloor originally of 13 Ma. The seamount is >3500 m tall, rising from a base at around 4100 m to a summit at 527 m depth. The petrology of samples suggests that summit eruptions were explosive and, assuming a subsidence history similar to Fieberling Guyot, could have occurred at 200–300 m depth [Gee et al., 1991].

The bathymetry (Figure 5) reveals, around the base of the edifice, a series of flat-topped features that are lobate in plan view and are most likely solidified submarine lava lobes or lava lakes such as observed around Hawai’i [Clague et al., 2000b; Moore and Chadwick, 1995]. Above them, the flanks are mostly 15°–25°, but rugged with slopes varying from <5° to >30°. Although no clear headwall is visible, a likely slope failure occurs along the north slope of the edifice around 30°30’N, 122°45’W. This area is smooth with slopes of 15°–20° declining to <15° with depth. A possible debris lobe occurs at 30°37’N, 122°44’W. A small downslope channel along the west side of the avalanche chute marks its boundary with a prominent ridge (marked “Terraced morphology” in Figure 5), which was interpreted as a volcanic rift zone based on morphology [Gee et al., 1991] and high seismic velocities [Hammer et al., 1994].

3.3. Guadalupe Island

Guadalupe Island, or Isla Guadalupe (Figure 6) lies on a mid-ocean ridge that ceased spreading at ~12 Ma [Lonsdale, 1991]. The island has been historically active, whereas K-Ar dating has yielded ages of 3.7, 5.4, and 7 Ma for the older formations [Batiza et al., 1979; Engel and Engel, 1970]. The geology of the island comprises two lava shield volcanoes covered with younger rift series lavas [Batiza, 1977], with the youngest of the shields in the north.

Despite low resolution, the bathymetry data (Figure 6), reveal a prominent embayment of the northeast submarine flank (29°08’N, 118°15’W) which mimics an embayment of the coastline. Downslope from the
embayment, seaward deflection of the contours around 3000 m depth suggests a debris avalanche depositional lobe. A horseshoe-shaped semicircle of cinder cones and faults and fissures (white dashed line in Figure 6) occurs around the top of an embayment in the subaerial topography [Batiza, 1977]. This feature is most likely caused by eastward collapse of the northern shield volcano. The missing eastern half of the subaerial embayment implies that failure involved the caldera rim of the volcano.

3.4. Henderson Seamount

Henderson Seamount is 3400 m high, has been dated (Ar-Ar) as 7.8 ± 0.8 Ma, and occurs on 22 Ma seafloor [Honda et al., 1987]. The bathymetry [Taylor et al., 1980]
shows an arcuate embayment in the east side of the seamount’s summit and a broad lobe on the seamount east flank. Although Taylor et al. [1980] were unsure whether this feature was caused by a landslide or a trapdoor-type caldera collapse, the morphology strongly resembles the debris avalanche lobes on the west side of La Palma [Urgeles et al., 1999], and it is most likely the result of slope instability.

3.5. **Galapagos Islands**

The southwesterly Galapagos Islands are considered here (Figure 7), because these young volcanoes have grown...
against the steep edge of the Galapagos platform, providing potential instability. The volcanoes of Cerro Azul and Fernandina have been historically active. Dating their ages of inception is difficult because the lack of erosion associated with their dry climate has prevented exposure of deeper units, but Naumann and Geist [2000] predicted that both islands are <500 ka. Two K-Ar dates of lavas dredged from the escarpment immediately south of Isabela island are 0.1 and 0.2 Ma [Sinton et al., 1996].

A subaerial embayment of the southwest flank of Cerro Azul has been interpreted as a landslide scar [Naumann and Geist, 2000]. The adjacent island shelf is also very narrow below the subaerial scar (00°58′S, 91°28′W) and a broad embayment of the deeper island flank occurs in the bathymetry (Figure 7). Slopes in the embayment are steeper in the upper flank (>15° at depths shallower than 2000 m) than the west promontory (slopes generally 5°–15°), whereas slopes in the embayment decline to less than 5° at 3000 m depth, significantly gentler than in the adjacent promontory. The embayment is significantly less rugged than the adjacent promontory. This feature is interpreted as caused by one or more slope failures of Cerro Azul. Profiles in Figure 14 reveal an abruptly steepened upper submarine slope along both sections (Figures 14a and 14b) in Figure 7, which could indicate lava deltas grown over the upper landslide embayments.

Figure 8. (a) Bathymetry as in Figure 6 and (b) imagery from long-range side-scan sonar and SIR-C of São Jorge Island of the Azores. The side-scan sonar data were collected with the vessel’s track running east-west along the north margin of the map and with the sonar ensonifying seafloor toward the southeast as shown (hence black area is acoustically blocked by the island). See color version of this figure at back of this issue.
[22] A similar embayment of the western submarine flank of Fernandina Island occurs at 00°20′S, 91°42′W, forming a smooth surface where continuous multibeam sonar data exist, compared with the northwest promontory of the island, which has lobate structures in the corresponding slope map and is a submarine ridge typical of volcanic rift zones [Vogt and Smoot, 1984].

[23] The extreme northwest point of Isabela Island (Cape Berkely) was interpreted by McBirney and Williams [1969] as the remains of a volcano caldera with its western two thirds downfaulted into the sea. Although the bathymetry coverage is poor, the steep nearshore bathymetry revealed in hydrographic soundings implies that the erosion could have occurred by a failure of the edifice to the west. More recently collected multibeam and side-scan sonar data [Fornari et al., 2001] confirm these interpretations of landslides around Cerro Azul, Fernandina, and at Cape Berkely.

3.6. Easter Island

[24] Easter Island consists of three volcanoes dated (K-Ar) 0.1–0.7 Ma [Miki et al., 1998] and <0.7 Ma [Kaneoka and Katsui, 1985]. A long-range side-scan and multibeam sonar survey around the north and west of the island [Hagen et al., 1990] shows only a volcanic morphology. Rounded structures of varied backscatter in side-scan sonar images suggest volcanic cones, craters, lava flows, and possible lava terraces but no observable debris avalanche structures.

3.7. São Jorge Island

[25] São Jorge Island (Figure 8) is a remarkably linear feature; the island’s morphology is probably controlled by...
major faults of the Azores deformation zone [Lourenco et al., 1998]. The western half of the island is volcanically active and young, including historical eruptions [Weston, 1964], whereas the eastern half has yielded K-Ar dates of 0.11–0.55 Ma [Feraud et al., 1980]. This difference is revealed in the subaerial topography, with a weakly eroded volcanic ridge on the western half of the island, marked by cinder cones and a ridge-perpendicular drainage system, whereas the eastern half contains deep erosional amphitheaters and more mature dendritic drainage systems. The western half of the island also has a narrower shelf (up to 2 km) compared with the eastern half (3–4 km wide).

The bathymetry shows that the island flanks are everywhere steep, mostly >15°, and 20°–30° for 50% of the flanks. Whereas irregularities in the submarine slope are consistent with minor slope failures, there is no clear evidence for major landslide embayments of the shelf break or slope. Furthermore, the side-scan sonar data show a monotonous basin north of the island and no evidence for debris avalanche deposits. The radar image data show small amphitheater-shaped valleys around the southeast third of the island (east of 28°00'W) presumably partly created by small-scale hillslope landsliding, but the data otherwise show no major embayments or arcuate faults which could be attributed to major landslides.

3.8. Flores and Corvo

The westerly Azores Islands of Flores and Corvo (Figure 9) lie on ~10 Ma oceanic crust west of the Mid-Atlantic Ridge [Freire Luis, 1996]. Flores has been volcanically active since at least 0.50–0.62 Ma [Feraud et al., 1980] or 2 Ma [Azevedo and Portugal Ferreira, 1999] based on K-Ar dating. The subaerial edifice of Flores is morphologically complex due to the combined effects of volcanism and erosion, with a complex coastline reflecting coastal erosion and recent uplift [Azevedo and Portugal Ferreira, 1999].

[28] The bathymetry shows a steep, rugged submarine slope. Although the presence of minor slope failures is not precluded by the available data, the shelf breaks of the two islands (200–300 m on Flores and 100–200 m on Corvo) are rounded in plan view with no significant deep embayments. A number of small elevated features with closed contours are observed around the submarine apron of the island. These do not lie downslope of embayments of the submarine slope, so they are interpreted as more likely volcanic cones than landslide blocks.

3.9. Southwest Terceira Island

The west flank of Terceira Island (Figure 10) has grown against a >1000-m-deep basin of the Terceira Rift, creating a steep, potentially unstable slope. The large westerly strato-volcano on the island is Santa Barbara. The island has been historically active [Self, 1976; Weston, 1964], with recent eruptions on the submarine ridge running northwest from Santa Barbara [Freire Luis et al., 1999]. A sample from within the volcano’s caldera has been dated (K-Ar) at less than 0.029 Ma [Feraud et al., 1980]. No landslides or likely slump headwall faults are mentioned in reports from subaerial field mapping [Self, 1976; Zbyszewski et al., 1971] and no evidence for them exists in the subaerial topography (Figure 10a).

[30] The bathymetry (Figure 10a) shows that the island has a narrow shelf to the southwest, only 1 km wide, and a shelf break that parallels the coastline without any major landslide embayments. The island bathymetry steepens upward to slopes >25° and locally 30° immediately below the shelf break. The side-scan sonar data (Figure 10b) show no evidence for major landslides, although some linear features running down the submarine slope are probably small ero-
Figure 11. (a) Bathymetry as in Figure 6 and (b) imagery from long-range side-scan sonar and SIR-C of Pico Island of the Azores. The side-scan sonar data were collected with the vessel’s track running east-west along the south margin of the map and with the sonar ensonifying seafloor toward the north. High backscatter is shown as white. See color version of this figure at back of this issue.
sional gullies not resolved in the corresponding bathymetry. The southwest flank of Terceira is interpreted as a volcanic constructional slope which has been affected by only minor superficial slope failures and erosion.

3.10. Pico Island

[31] Pico Island (Figure 11) comprises the tall central volcano, Pico (altitude 2351 m above sea level (asl)), forming the main edifice in the west, the smaller Topo volcano (at 38°25′N, 28°13′W) and a volcanic ridge whose axis is well defined by the line of cinder cones visible in the radar image in Figure 11b. K-Ar dates [Feraud et al., 1980] from the older north side of the easterly end of the island and one south of the Topo volcano are quite young, the latter taken near 38°23′N, 28°13′W found to be <0.037 Ma. The island has been historically active [Weston, 1964]. High Bouguer gravity anomalies centered on Pico and Topo volcanoes suggest intrusive complexes [Ridley et al., 1974].

[32] One possible landslide on the south side of Pico volcano at 38°26′N, 28°25′W, originally identified by Woodhall [1974], forms an arcuate embayment in the subaerial topography. The radar image reveals gullies that are similar to those on Hawaiian subaerial landslide scars [Moore et al., 1989]. Detailed field mapping [Nunes, 1999], however, has revealed that the avalanche deposits overlie undisrupted coastal lava and volcanic stratigraphy is coherent across the gullies, so the slope mass wasting may instead be caused by oversteepening such as caused by a fault [Nunes, 1999]. Furthermore, a linear target in the side-scan sonar imagery southwest of Pico volcano along the coastline (Figure 2) suggests a probable fault escarpment. This feature is coherent and inconsistent with a large-scale debris avalanche structure or slump of the volcano’s flank.

[33] Faults observed on Topo volcano (marked on Figure 11b) form an arc as though caused by movement to the southeast. A speckle pattern occurs in the side-scan...
sonar imagery beneath it in Figure 11b. Cruz and Silva [2001] described this feature as a giant landslide. The island shelf is relatively narrow immediately below the Topo landslide. The landslide is difficult to classify as its movement has not wholly evacuated the headwall region (see profile in Figure 14), nevertheless, the speckle pattern in side-scan sonar is typical of debris avalanches. Other, more ambiguous candidates for landslides include the embayment at 38°30’N, 28°18’W (its linear headwall, however, parallels other Azores tectonic features [Lour- enco et al., 1998]) and the embayment of the island slope at 38°28’N, 28°11’W.

3.11. Jan Mayen Island

[34] Jan Mayen Island lies at the northern end of the Jan Mayen Ridge where it abuts the active Jan Mayen transform valley. The dominant strato-volcano in the north of the island (Beerenberg, altitude 2277 m asl) has grown into the transform valley. Dating of rock samples suggests that Beerenberg is less than 0.5 Ma [Saemundsson, 1986]. A volcanic rift zone extends to the northeast corner of the island [Fitch, 1964] and a submarine ridge extends for a further 20 km offshore (UK Hydrographic Office, Navigation Chart 4100, 1997). A fissure eruption occurred on the east side in 1970, trending from 600 m asl to the coast along an arc subparallel to the east coastline and open to the southeast [Imsland, 1986; Maaloe et al., 1986]. The extension toward the southeast and coast-parallel orientation of fissures is analogous to the configuration of the larger-scale Hilina faults of Kilauea and may represent slumping of Beerenberg’s east flank.

3.12. Ascension Island

[35] Ascension Island (Figure 12) lies 110 km west of the Mid-Atlantic Ridge on 7 Ma seafloor [Minshull and Brozena, 1997; van Andel et al., 1973]. K-Ar ages of lavas from the island range from 0.65 to 1.5 Ma [Harris et al., 1982; Nielson and Sibbett, 1996] although there are also basaltic flows that have been described as “clearly very young, perhaps only a few hundred years old” [Atkins et al., 1964]. The most recent volcanic growth has occurred on the eastern side of the island represented by the peak in Figure 12 [Nielson and Sibbett, 1996]. The older western half of the edifice is eroded, forming a broad shelf which coincides with high gravity anomalies and seismic velocities, probably representing an older intrusive complex [Klingelhofer et al., 2001; Minshull and Brozena, 1997]. Klingelhofer et al. [2001] further proposed that the island may have begun growing much earlier than the K-Ar data suggest, perhaps initiating as a volcano within or adjacent to the Mid-Atlantic Ridge median valley.

[36] The lack of data on the western side prevents detailed interpretation, but the multibeam data at 3000 m depth show no clear evidence for landslides. The younger east side is

Figure 13. Bathymetry of the north side of Bouvet Island as Figure 6 (region above sea level is shown black). See color version of this figure at back of this issue.
<table>
<thead>
<tr>
<th>Name</th>
<th>Lat, b/C</th>
<th>Long, b/C</th>
<th>Altitude, c/m</th>
<th>Edifice Height, d/m</th>
<th>Precipitation, e/mm/yr</th>
<th>Historical Activity</th>
<th>Min/Max Ages, f/Ma</th>
<th>Embayment Submarine</th>
<th>Subaerial</th>
<th>Debris Lobe</th>
<th>Debris Field</th>
<th>Morphological Contrast</th>
<th>Gradient Contrast</th>
<th>Feature Width, g/km</th>
<th>Feature Altitude, h/m</th>
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<td>0/1.5</td>
<td>Y</td>
<td>M</td>
<td>...</td>
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<td>0/-</td>
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<td>...</td>
<td>...</td>
<td>...</td>
<td>8.4/5.0</td>
<td>900</td>
<td>2.5</td>
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</tbody>
</table>

aY, yes; N, no; M, moderate evidence; dots, data inadequate or unavailable.
bLat, latitude; long, longitude; approximate location only.
cSummit altitude of appropriate volcano (islands) or water depth (seamounts).
dEdifice height above the adjacent ocean floor.
epAnnual precipitation recorded at nearest rain gauge station [National Climate Data Center, 1969].
fActive historically or in the Holocene based on publications, Smithsonian Institution Global Volcanism Program and Weston [1964].
gReported minimum/maximum radiometric ages (zero minima where active in Holocene).
hEmbayment of shelf edge or island/seamount slope.
iEmbayment or faults above sea level or in seamount summit.
jInterpreted debris lobe typical of debris avalanche deposits.
kSpeckle pattern in side-scan sonar image data.
lLocal contrast in topographic roughness or missing volcanic morphology on island or seamount flank.
mLocal contrast in gradient between proposed chute and adjacent terrain within island or seamount upper flank.

\[\text{Width of embayment in edifice submarine flank unless stated in footnote explanations q–v.}\]


\[\text{Highest evidence for the landslide (most speculative locations indicated by question marks).}\]

\[\text{CL, confidence level; nominal confidence level given by the sum of scores in indicators h–l with scores assigned 0, 0.5, and 1 for N, M, and Y, respectively.}\]

\[\text{Westerly chutes/wider embayment.}\]

\[\text{Embayment of submarine slope/shelf.}\]

\[\text{SW arcuate sea cliffs and subaerial fault [Chevallier, 1987].}\]

\[\text{Onshore eruptive fissures/embayment in hydrographic chart.}\]

\[\text{Subaerial Santa Rosa valley [Chevallier, 1986].}\]

\[\text{Slope chute/subaerial embayment.}\]
Figure 14. Profile sections of the landslides (bold lines) and adjacent terrain (fine lines) located by the open and solid triangles, respectively, in Figures 4–13. The vertical exaggeration is 3:1, and the horizontal line in each case represents current sea level. The arrows mark the interpreted headwalls or an uppermost steep region, which may represent an overprint by coastal volcanism or lava delta (e.g., Fernandina, Cerro Azul (b), and Guadalupe).
steep, with slopes steeper than 10° on average and locally up to 30°. The rugged morphology suggests it is mostly volcanic. However, at around 7°59’S, 14°19’W and at 7°55’S, 14°19’W, the island shelf and shelf break are missing in the sounding data, and the flanks are embayed. The former embayment also occurs adjacent to a linear section of steep cliffs along the coastline (Figure 12). These are interpreted to be small landslides that have removed two sections of the island’s shelf. A further large embayment occurs on the south side at 8°00’S, 14°23’W but the data coverage is too poor to assess this feature.

3.13. Gough Island

[37] Gough Island lies on ~38 Ma oceanic crust, and contains lavas dated (K-Ar) 0.13–1.0 Ma [Le Roex, 1985], 0.12-2.5 Ma [Maud et al., 1988], and (Ar-Ar) <1.0 Ma [O’Connor and le Roex, 1992], the latter assumed to be the most reliable. The island’s submarine flank to the southwest has a 25-km-wide concave-seaward embayment which suggests one or more major landslides. Chevallier [1987] interpreted a series of concave-southwest arcuate slope breaks and cliffs, one of which encompasses the island’s 910 m peak, as evidence for slumping.

3.14. Tristan Da Cunha Island

[38] Tristan da Cunha Island lies on 9 Ma seafloor [Holcomb and Searle, 1991] and includes rocks dated (Ar-Ar) at less than 1 Ma [O’Connor and le Roex, 1992] and (K-Ar) 0.01–0.21 Ma [McDougall and Ollier, 1982], with historical activity also reported. The island has a radial distribution of eruptive vents and dikes exposed by erosion [Chevallier and Verwoerd, 1987]. Holcomb and Searle [1991] recognized a speckle pattern in long-range side-scan sonar records below a steep concave-seaward cliff on the northwest side of the island, most likely marking the headwall of a major landslide. The cliff, however, does not affect much of the subaerial part of the edifice (affecting only lower subaerial slopes).

3.15. Bouvet Island

[39] Bouvet Island or Bouvetoya (Figure 13) lies only 45 km southwest of the Southwest Indian Ridge axis on 4–5 Ma oceanic crust [Sclater et al., 1976]. The age of the island is poorly defined. There is evidence for recent activity [Baker, 1967; Baker and Tomblin, 1964], but a 3–4 km wide shelf to the north (Figure 13) suggests that it has completed its main growth phase. Insland et al. [1977] suggested that the island has evolved beyond a caldera collapse stage based on the summit morphology, although summit details are largely obscured by ice.

[40] The northeast submarine flank of the edifice (54°22’S, 3°35’E) forms a broad promontory with varied slopes ranging from less than 5° to more 25° (“Terraced promontory” in Figure 13). The corresponding slope map shows terraces resembling those in submarine lava sequences around Hawaii [Moore and Chadwick, 1995]. The submarine slope immediately to the west has more uniform slopes, up to 10°–15° at depths shallower than 1000 m, declining to less than 5° below 1000 m, and is embayed relative to the promontory to the east. Contours on its surface are crenellated revealing a series of downslope canyons, similar to those observed on some Hawaiian landslide scars [Moore et al., 1989].

[41] The north side of the island is interpreted to have a volcanic constructional sector to the northeast and an embayment from one or more slope failures across an area at least 13 km wide on the north slope. The contours show a continuous decline in relief from the embayment down to the 2 km contour at the north edge of the map (“Sediment pathway” in Figure 13) and the bathymetry is generally smooth with isolated abyssal hills protruding. This smooth morphology could represent sediments shed from the island and its submarine slope, or it could represent widespread deposits from slope failures of Bouvet’s north flank.

3.16. Marion Island

[42] Marion reaches >1200 m altitude asl, has been historically active, and dates (K-Ar) extend to 450 ± 10 ka [McDougall et al., 2001]. Soundings show that the island’s west and south slopes drop sharply by at least 1000–2000 m b.s.l. [Chevallier, 1986]. A 2-km-wide embayment above sea level on the south side marks a collapse structure that is similar morphologically to that of the Grand Brulé slide of Piton de la Fournaise, Reunion [Duffield et al., 1982]. Chevallier [1986] also interpreted more superficial slump headwalls from the subaerial topography.

4. Discussion

[43] The landslide observations are summarized in Table 1 along with other data. In Figure 15a, the width of the landslide is plotted versus edifice height (with width normalized to height). Because the data interpretation focused on major landslides capable of modifying the edifice’s gross morphology, many smaller landslides will not have been detected. Nevertheless, the figure shows an apparent threshold in major landslide abundance: whereas small mid-ocean ridge seamounts only rarely show landslides (2 out of 44 seamounts), landslides become more common for edifices taller than 2500 m. A second observation is summarized in Figure 16, in which the highest observed evidence for landsliding is plotted as a black-to-white transition against the vertical extent of the edifice. Most of the interpreted landslides lie near sea level, whereas others lie near the peak of the island or seamount. The distribution partly represents bias in the identification method (slides being more readily identified where there are indications in data both above and below sea level). Nevertheless, they suggest different types of failure which are discussed below.

4.1. Correlation With Appearance of Volcanic Rift Zones

[44] Elsworth and Voight [1995, 1996] suggested that some affects associated with dike intrusion can potentially explain triggering of some landslides, as dikes cause excess pore pressures associated with the magma and groundwater heating and other mechanical effects. It is therefore interesting that the landslide threshold in Figure 15a correlates with the first appearance of volcanic ridges (interpreted as rift zones) in multibeam bathymetry data sets of ocean basin seamounts [Vogt and Smoot, 1984] shown in Figure 15c. Furthermore, Figure 15b shows a measure of the non-
The circularity of seamount morphology [Mitchell, 2001]. This parameter was calculated by extracting the bathymetry contour at half the height of the seamounts and calculating the moment of inertia of the shape formed by that contour. Circular seamounts have values of 0.1 and increasingly irregular seamounts, such as those with large volcanic ridges, satellite cones, or landslide embayments, have higher values [Mitchell, 2001].

The graphs in Figure 15 therefore suggest a possible connection between landsliding and growth of volcanic rift zones and morphological change, for which landslide triggering by intruding dikes in rift zones [Elsworth and Voight, 1995] could provide a connection. This is difficult to prove without detailed fieldwork and knowledge of the chronology of landsliding and dike intrusion. Some of the landslides identified here occur away from any evidence of volcanic rift zones or in edifices without rift zones (e.g., the mid-ocean ridge seamounts, Bouvet, Cerro Azul, Tristan da Cunha, and Ascension). Nevertheless, the landslides of Fernandina, Jasper, and Topo do occur adjacent to volcanic rift zones and therefore a connection will be worth exploring further to explain some of these edifice failures.

4.2. Location Near Sea Level

Although the distribution of landslide heights (Figure 16) partly reflects bias in the interpretation, many of these particular major landslides occur near sea level. In Figure 17, three general locations of landslides (arrows a–c) are plotted against a cross section of Ascension Island. Whereas a few of the landslides involved high-level failure
(location “a”), e.g., Topo and Guadalupe, most of the others occurred in the lower half of the subaerial edifice or not far below sea level (the failures of Henderson and Jasper may also have occurred when they were less submerged).

One potential explanation could lie in the development of extensive hyaloclastite deposits, whose weak geotechnical properties could be an important cause of some volcanic island landslides [Duffield et al., 1982]. Hyaloclastite is produced in two phases; as a volcano grows toward sea level, lower pressure explosive eruptions are expected to create abundant hyaloclastite [Staudigel and Schmincke, 1984] and, once grown above sea level, interactions of coastal lava with sea swell can create slopes dominated by fragmented volcanic debris and hyaloclastite [Moore and Krivoy, 1964; Moore et al., 1973; Moore and Chadwick, 1995; Fornari et al., 1979]. Depending on geometrically how they grow and how rapidly they grow relative to subsidence, extensive hyaloclastite bodies probably exist in mid-ocean ridge volcanoes that have grown near or above sea level. Their geometry is unknown but coastal lava is likely to create extensive hyaloclastite around island flanks, and its location could explain the shallow debris avalanche failures of types “b” and “c” of Figure 17.

Hyaloclastites could also explain the higher-level landslides and slumps. From the geometry in Figure 17 (also see the Guadalupe and Topo sections in Figure 14), these type “a” landslides probably failed more deeply within the edifice. A geothermal exploration well on Ascension [Nielson and Stiger, 1996] provides one possible explanation because an extensive section of compacted and impermeable hyaloclastite was observed deep within the edifice, which was found to produce an extremely slow pressure recovery following drilling. The well is summarized in Figure 17, with hyaloclastite intervals shown in black. Compaction is facilitated by high smectite contents such that compacted hyaloclastite is unable to sustain open fractures [Nielson and Stiger, 1996]. If such bodies are widespread in these volcanic islands, buildup of pore pressures, such as associated with magmatic intrusions, may provide an explanation for some type “a” failures.

4.3. Other Factors

The ground accelerations associated with passage of seismic waves are widely considered to be important triggers of landslides [e.g., Hampton et al., 1996]. Seismicity is widespread among the Azores Islands, with 11 earthquakes of magnitude ($m_s$) > 5 during the period 1963–1984 and two with $m_s$ > 7 [Buform et al., 1988], from which we might expect many $m_s$ > 6 earthquakes over historical or recent geological times. However, the Azores Islands show relatively little evidence for giant landslides, and generally the data studied here suggest no evidence that seismicity is an important factor.

Rainfall is another important trigger of subaerial hillslope landslides and might be expected to affect volcanic islands. Indeed, a movement on the H dinner slump system recorded with GPS measurements [Cervelli et al., 2002] was correlated with an extreme downpour. There is, however, little obvious correlation of landslide incidence with rainfall in this study as they occur equally in low- and high-latitude environments and in areas of both high and low rainfall. Figure 18a shows the landslides confidence levels

![Figure 18a](image-url)

Figure 18. (a) Landslides of Figure 15 versus altitude of the island peak above sea level for islands within 40°C of the equator (tropical and subtropical). The Ascension and Fernandina landslides are largely submarine and not relevant to the issue of rainfall addressed by this figure so they are shown reclassified as no subaerial landslide observed. The islands of Corvo, Flores, and Gough were excluded because they are eroded and their present altitudes are not representative of their altitudes at the time of failure (Gough) or growth (Corvo and Flores). (b) Landslides versus precipitation at nearby meteorological stations with altitudes within 100 m of sea level where data were available [National Climate Data Center, 1969]. Precipitation in the vicinity of landslides of the Azores and Galapagos is represented by measurements at Horta (Faial Island) and on Seymour Island, respectively. Annotation represents SJ, São Jorge; T, Terceira; To, Topo volcano (Pico); A, Ascension; Gu, Guadalupe; E, Easter Island; JM, Jan Mayen; F, Fernandina (Galapagos); CA, Cerro Azul (Galapagos); and Tr, Tristan da Cunha.
of Figure 15a plotted versus altitude above sea level of the edifice peak. Altitude is used as a proxy for excess rainfall caused by condensation of air rising over topography, based on the observation that shallow islands are locally drier than tall islands [Whittaker, 1998] and precipitation commonly increases with altitude on individual islands [Peterson, 1972]. Precipitation measured near sea level, however, varies greatly between oceanic island groups, so the failure classes are also plotted in Figure 18b against nearby meteorological station readings [National Climate Data Center, 1969]. Whereas there is a weak change in landslide incidence with altitude (Figure 18a), there are few islands of low altitude to confirm the trend and the change could partly reflect the trend with edifice height (Figure 15a). Figure 18b shows that islands with landslides in their low subaerial slopes occur in both dry (Cerro Azul) and wet (Pico) climates.

The apparent lack of correlation with rainfall probably reflects the high permeability of the subaerial part of volcanic islands because the water table commonly lies near sea level regardless of climatic setting. Where the water table topography has been measured, its gradient toward the coast is typically only $10^{-4}$ to $10^{-2}$ [Cruz and Silva, 2001; Ecker, 1976; Peterson, 1972; Violette et al., 1997]. Above the water table, the vadose zone contains compartments of groundwater trapped by impermeable dikes or at contacts between permeable and impermeable formations [Ecker, 1976]. Since islands in regions of both high and low rainfall have small water table gradients (e.g., $10^{-4}$ for Pico [Cruz and Silva, 2001] an island with comparatively high precipitation), it is difficult to envisage how climatic changes in precipitation could affect edifice large-scale stability [McGuire, 1996]. For extreme rainfall events to cause landsliding [e.g., Cerrelli et al., 2002], a possibility is that the runoff temporarily fills columns of standing water in vadose zone cavities leading to localized high pore pressures sufficient to trigger failure. Such possibilities require further study of volcanic island hydrology.

5. Conclusions

The landslide identification based on these data sets was hampered by varied data quality and by data of mostly lower resolution than other studies. Designing a quantitative classification scheme is complicated also by natural processes, as volcanism and erosion can obscure evidence for landslides and the results are biased toward the larger landslides that have strongly modified the edifices. Nevertheless, an important result of this study is that landslides are uncommon in edifices shorter than 2500 m, but are common when they are taller than 2500 m.

Landslide triggering by dikes [Elsworth and Voight, 1995] could provide a partial explanation for this threshold, as the change in landslide incidence coincides with the appearance of volcanic rift zones in seamounts where dike intrusion is expected to have been more common. Indeed, some landslides, such as that of Fernandina, occur adjacent to volcanic rift zones. A second possible explanation is that many edifices of 2500 m or taller lie (or previously lay) near sea level and therefore shallow marine eruptions and coastal lava-sea interaction will have created extensive hyaloclastite deposits. This may explain why many of the major landslides lie near present sea level as failure could have initiated in shallow hyaloclastite. Observations of extremely low permeability in compacted hyaloclastite within a geothermal well on Ascension Island hint at an origin for the more deeply seated landslides, as compacted hyaloclastite may retain high pore pressures associated with volcanic intrusions or extreme rainfall events.

Appendix A: Sources of Data and Processing

The Ascension Island bathymetry (Figure 12) was mostly created from two lines encircling the island at >2000 m depth with a Hydrosweep-DS multibeam sonar on the “FS Meteor” in 1998 [Devey et al., 1999]. Poor performance of the sonar necessitated removal of significant parts of the data during processing, but they nevertheless reveal broad volcanic features. Data on the island’s shelf, shelf break, and upper slope were obtained by manually digitizing sounding charts provided by the UK Hydrographic Office (providing the dense coverage nearshore and the east-west lines on the north side of the island shown in the inset to Figure 12). These were further supplemented with echo soundings from the National Geophysical Data Center (NGDC), Boulder, CO, and two short lines of SeaBeam multibeam data east of the island from a database at Lamont-Doherty Earth Observatory.

The Bouvet Island bathymetry (Figure 13) was mostly created from a series of 3.7-km-spaced east-west lines of the R/V Strakhov using a 15-beam Finnyards ECHOS 625 multibeam sonar [Bulychev et al., 1998; Ligi et al., 1999], supplemented with sounding data on the shelf (M. Ligi, personal communication, 1999). The Guadalupe Island bathymetry (Figure 6) was derived from single-beam sounding data, mostly collected by J. Hildebrand on the R/V Robert Sproul in 1991 and made available through the NGDC. Figure 7 shows the bathymetry around Isabela and Fernandina Islands of the Galapagos archipelago, derived from a compilation made available by W. Chadwick of Oregon State University, including soundings from research ships and soundings derived from hydrographic charts. The main details of the island slope morphology are revealed by data from a 16-beam SeaBeam multibeam sonar collected on the R/V Thomas Washington in 1990 by D. Christie and R. Duncan (heavy lines in Figure 7 inset). The bathymetry of Jasper seamount (Figure 5) was collected during a cruise of the Thomas Washington in 1983 by P. Lonsdale and made available by the NGDC.

The bathymetry data from the Azores Islands (Figures 8, 9, 10, and 11) were originally digitized from sounding charts of the Istituto Hidrografico Portugal by the Defense Mapping Agency (USA) and made publicly available by the NGDC. The original sounding data were collected with single-beam echo sounders in the 1950s and 1960s, and the observations in this paper were made accounting for the likely navigation accuracy of this age of data. The islands shown are Flores and Corvo (Figure 9), Terceira (Figure 10), Pico (Figure 11), and São Jorge (Figure 8). The topography above sea level was derived from topographic maps. The gray images also shown in Figures 8, 10, and 11 include GLORIA long-range side-scan sonar data originally collected by Searle [1980]. These data
were digitized from analogue magnetic tapes and processed using modern digital methods at the Southampton Oceanography Centre. The data have a resolution of 45 m perpendicular to the ship’s track and resolution parallel to track varying from 50–100 m near the sonar to ~1000 m farthest from the sonar [Mitchell and Somers, 1989; Somers et al., 1978]. The images are plotted with positive polarity such that high backscatter is white. Although these data have low resolution and can suffer from foreshortening in rugged terrain [Mitchell, 1991], they are useful for revealing gross morphology and general differences of seabed geology (bare rock is highly backscattering whereas sediment is low backscattering). The subaerial regions in Figures 8 and 11 are represented by 25-m-resolution SIR-C imagery obtained from the Jet Propulsion Laboratory, USA. The bathymetry data were interpolated onto fine grids using a surface-fitting algorithm [Smith and Wessel, 1990] with moderate tension factors to reduce oscillations in areas of sparse soundings.

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Figure 1. The bathymetry of El Hierro Island of the Canaries, showing a deep embayment and associated debris field on its north side created by the El Golfo debris avalanche, and further embayments to the southwest and southeast caused by older collapse events [Gee et al., 2001; Holcomb and Searle, 1991; Masson et al., 2002; Urgeles et al., 1998]. Data from the two outlined areas were used to generate Figure 2. (Data courtesy of D.G. Masson and A.B. Watts.)
Figure 4. Two examples of slope failure structures in mid-ocean ridge seamounts. Contours are plotted every 100 m with annotation in kilometers. Large arrows mark interpreted landslides. The triangles locate the profiles in Figure 14 (open symbols for the landslide profile and solid symbols for adjacent reference profiles). Some 44 such seamounts between 200 and 1700 m tall were examined but the data shown represent the only evidence for major flank failures found.
Figure 5. Bathymetry of Jasper seamount as Figure 4. The data were collected with a SeaBeam multibeam sonar on the Thomas Washington by P. Lonsdale in 1983 and made available by the NGDC. Gaps between data are plotted in midgray without contours.
Figure 6. Bathymetry of Guadalupe Island as Figure 12 (region above sea level is shown black). The symbols on the north half of the island represent (open circles) cinder cones, and (lines) faults and fissures according to Batiza [1977]. Large arrow marks interpreted landslide. The triangles locate the profiles in Figure 14 (open symbols for the landslide profile and solid symbols for adjacent reference profiles). Inset shows the data coverage.
Figure 7. Bathymetry of Isabela and Fernandina Islands of the Galapagos archipelago as Figure 6.
Figure 8. (a) Bathymetry as in Figure 6 and (b) imagery from long-range side-scan sonar and SIR-C of São Jorge Island of the Azores. The side-scan sonar data were collected with the vessel’s track running east-west along the north margin of the map and with the sonar ensonifying seafloor toward the southeast as shown (hence black area is acoustically blocked by the island).
Figure 9. Bathymetry of Flores (south) and Corvo (north) Islands of the Azores as Figure 6 (region above sea level is shown black).
Figure 10. (a) Bathymetry as in Figure 6 and (b) long-range side-scan sonar of the steep west flank of Terceira Island of the Azores. The side-scan sonar data were collected with the vessel’s track running from the SE corner of the map to 38°40′N, 27°30′W, with the sonar ensonifying the seafloor toward the north-northeast as indicated. High backscatter is shown as white. Areas above sea level shown in black.
Figure 11. (a) Bathymetry as in Figure 6 and (b) imagery from long-range side-scan sonar and SIR-C of Pico Island of the Azores. The side-scan sonar data were collected with the vessel’s track running east-west along the south margin of the map and with the sonar ensonifying seafloor toward the north. High backscatter is shown as white.
Figure 12. Bathymetry of Ascension Island as Figure 6. The data included bathymetry manually digitized from hydrographic sounding charts (adapted from Admiralty surveys by permission of the Controller of Her Majesty’s Stationary Office and the UK Hydrographic Office (http://www.ukho.gov.uk)).
Figure 13. Bathymetry of the north side of Bouvet Island as Figure 6 (region above sea level is shown black).