

Prodigious Submarine Landslides on the Hawaiian Ridge

J. G. MOORE,¹ D. A. CLAGUE,¹ R. T. HOLCOMB,² P. W. LIPMAN,³
W. R. NORMARK,¹ AND M. E. TORRESAN¹

The extensive area covered by major submarine mass wasting deposits on or near the Hawaiian Ridge has been delimited by systematic mapping of the Hawaiian exclusive economic zone using the side-looking sonar system GLORIA. These surveys show that slumps and debris avalanche deposits are exposed over about 100,000 km² of the ridge and adjacent seafloor from Kauai to Hawaii, covering an area more than 5 times the land area of the islands. Some of the individual debris avalanches are more than 200 km long and about 5000 km³ in volume, ranking them among the largest on Earth. The slope failures that produce these deposits begin early in the history of individual volcanoes when they are small submarine seamounts, culminate near the end of subaerial shield building, and apparently continue long after dormancy. Consequently, landslide debris is an important element in the internal structure of the volcanoes. The dynamic behavior of the volcanoes can be modulated by slope failure, and the structural features of the landslides are related to elements of the volcanoes including rift zones and fault systems. The landslides are of two general types, slumps and debris avalanches. The slumps are slow moving, wide (up to 110 km), and thick (about 10 km) with transverse blocky ridges and steep toes. The debris avalanches are fast moving, long (up to 230 km) compared to width, and thinner (0.05–2 km); they commonly have a well-defined amphitheater at their head and hummocky terrain in the lower part. Oceanic disturbance caused by rapid emplacement of debris avalanches may have produced high-level wave deposits (such as the 365-m elevation Hulopoe Gravel on Lanai) that are found on several islands. Most present-day submarine canyons were originally carved subaerially in the upper parts of debris avalanches. Subaerial canyon cutting was apparently promoted by the recently steepened and stripped slopes of the landslide amphitheaters.

INTRODUCTION

New perspectives on the extent and character of submarine slope failures have been obtained with the systematic mapping of the Hawaiian exclusive economic zone (EEZ) using the side-looking sonar system GLORIA (Geologic Long-Range Inclined Asdic). In the fall of 1988, GLORIA surveys extending 200 nautical miles from land were completed in a zone from the southeast end of the Hawaiian Ridge adjacent to the island of Hawaii northwest to beyond the island of Kauai. In this region which includes the principal Hawaiian islands, 17 well-defined major landslides were mapped (Figures 1 and 2 and Table 1), and remnants of several others partly covered by younger deposits also were found. This report documents the characteristics of these landslides, examines their morphologic variations, and attempts to relate them to the processes that built and modified the volcanic systems of the Hawaiian Ridge.

The study of submarine mass-wasting deposits composed of massive or poorly bedded volcanic rocks presents difficult problems not encountered in the better understood deposits composed of sedimentary materials. Slumps, block slides, debris flows, and other such processes are widely recognized in nearly all types of continental margin settings (see reviews by Moore [1978], Embley [1982], Kenyon [1987], Coleman and Prior [1988], Lee [1989]). These deposits are common in areas of high sedimentation rates and include thick accumulations of sediment with slope-parallel bedding. Features produced by sediment failure are well described on modern

delta slopes, where the relatively shallow water allows high-resolution surveys using side-looking sonar, reflection profiling, borehole drilling, and measurements of geotechnical properties [Coleman and Garrison, 1977; Prior and Coleman, 1982, 1984; Prior *et al.*, 1984a, b]. In slope regions with lower sediment accumulation rates, earthquakes are commonly the favored triggering mechanism for ground failure over areas of a few hundred square meters to as much as 80,000 km² [e.g., Field *et al.*, 1982; Normark, 1974; Kenyon, 1987] (see also review articles cited earlier).

The larger mass wasting deposits from continental margins extend well into deep water and, including the debris flows associated with many submarine slope failures, may reach several hundred kilometers in length [Jacobi, 1976; Bugge *et al.*, 1988]. Because much of the deposits from these larger failures lie in deep water (>2000 m), less is known about their internal structure and composition as compared to continental shelf structures (see discussion by Normark and Gutmacher [1988]). The Storegga submarine slide on the Norwegian continental margin exceeds 5000 km³ in volume and is one of the largest mass failures [Bugge *et al.*, 1988]. It apparently represents three episodes of failure, the second of which may be associated with a tsunami deposit along the east Scottish coast [Dawson *et al.*, 1988].

Slope failures on sedimented continental margins are comparatively easy to document as compared to those on volcanic slopes. The small-scale failures in bedded sediment appear on high-resolution reflection profiles because the missing sediment section is obvious, and the mass wasting deposits show as lens- or wedge-shaped units with irregular, hummocky surfaces and little or no coherent internal structure. Seismic reflection profiles can discern these deposits even when buried by later sediment, and both the scar and base of sediment mass-wasting deposits can be clearly recognized.

In contrast, mass failure on the submarine slopes of oceanic volcanoes must be recognized primarily on the basis

¹U.S. Geological Survey, Menlo Park, California.

²U.S. Geological Survey, University of Washington, Seattle.

³U.S. Geological Survey, Denver, Colorado.

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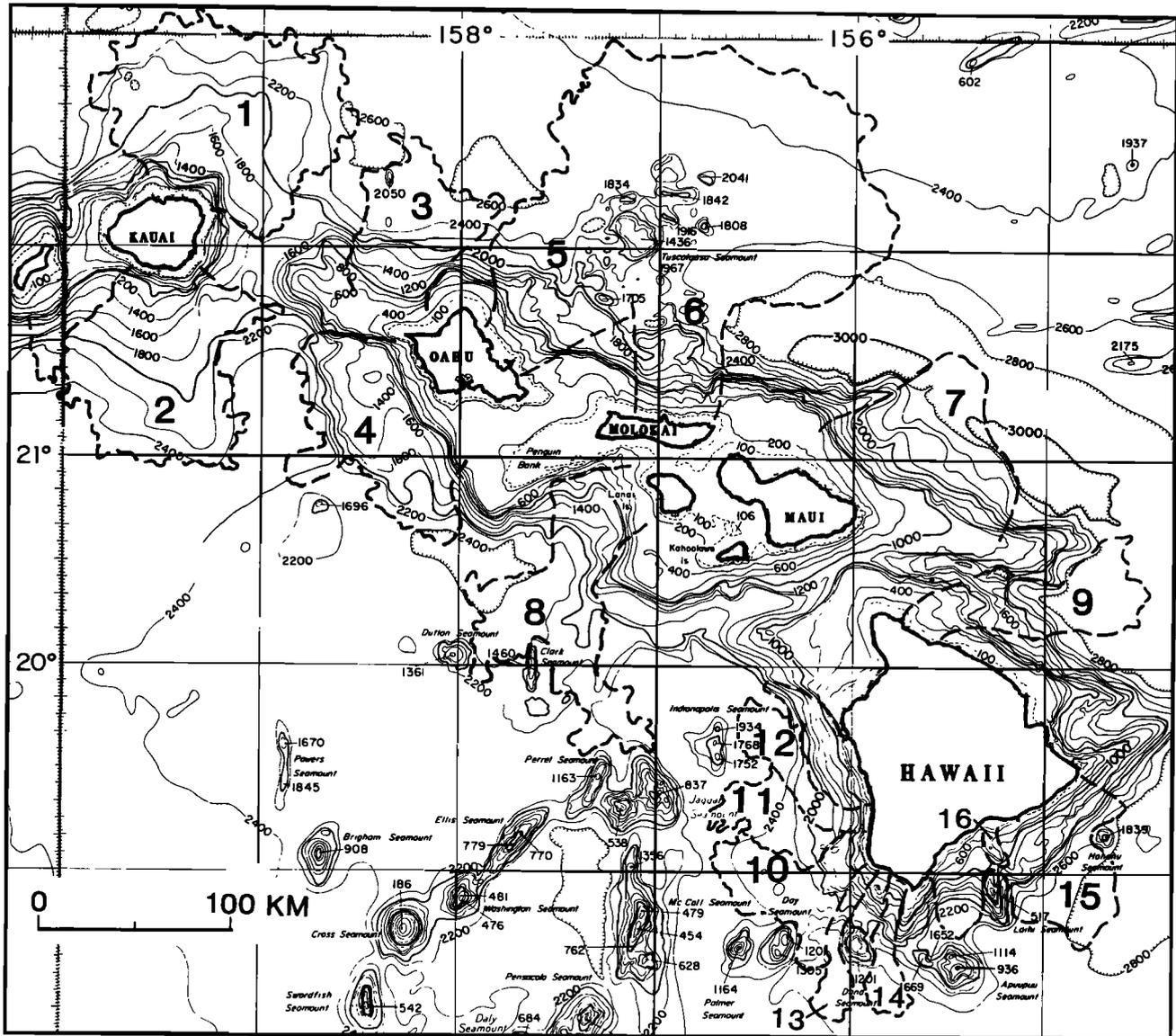


Fig. 1. Bathymetric map of the southeastern Hawaiian Ridge showing major mass wasting features (heavy dashed lines) identified by number in text and Table 1. Contour interval 200 fathoms (366 m). Base from *U.S. Naval Oceanographic Office* [1973].

of surface features as the basaltic rock and rubble are nearly opaque acoustically. Standard seismic reflection profiling systems generally do not provide subsurface information on the internal structure of slide components or of the slopes where they originated. With relatively slow accumulation of primarily pelagic sediment, the Hawaiian Ridge slope adjacent to the islands has insufficient sediment cover to provide much assistance in defining landslide features.

PREVIOUS WORK ON HAWAIIAN SLOPE FAILURES

The role of major landsliding in shaping Hawaiian volcanoes has long been a subject of debate. An early controversy concerned the relative importance of faulting versus marine erosion in producing high cliffs along the coasts of some islands. The high cliffs of northeast Oahu (Nuuanu Pali), as well as those of east Niihau, the Napali coast of Kauai, and northern Molokai were interpreted by *Dana* [1890, p. 290] as resulting from "profound fracturing of the mountain dome" and "downdrop" of the seaward part beneath the sea.

Dana's faulting model was rejected by *Hitchcock* [1900], who believed that erosion alone was sufficient to form the cliffs without the additional complexity of faulting. *Wentworth* [1927], comparing estimates of marine and fluvial erosion rates, concluded that marine erosion alone was insufficient to produce some of the larger sea cliffs (e.g., those of north Molokai) and that they were probably produced by faulting.

The idea of flank faulting eventually led to the concept that the faults bound the headwalls of giant landslides. *Stearns and Clark* [1930] and *Stearns and Macdonald* [1946] suggested that the 50-km-long Hilina fault system on the south flank of Kilauea volcano owes its origin to, and is at the head of, a major landslide on the steep seaward slope. *Macdonald* [1956] later discounted landsliding, suggesting that large-volume intrusion beneath the volcano center forcibly uplifted the summit causing faulting on the volcano flank.

Moore and Krivoy [1964] revived the landslide concept for Kilauea's south flank and suggested that not only is the

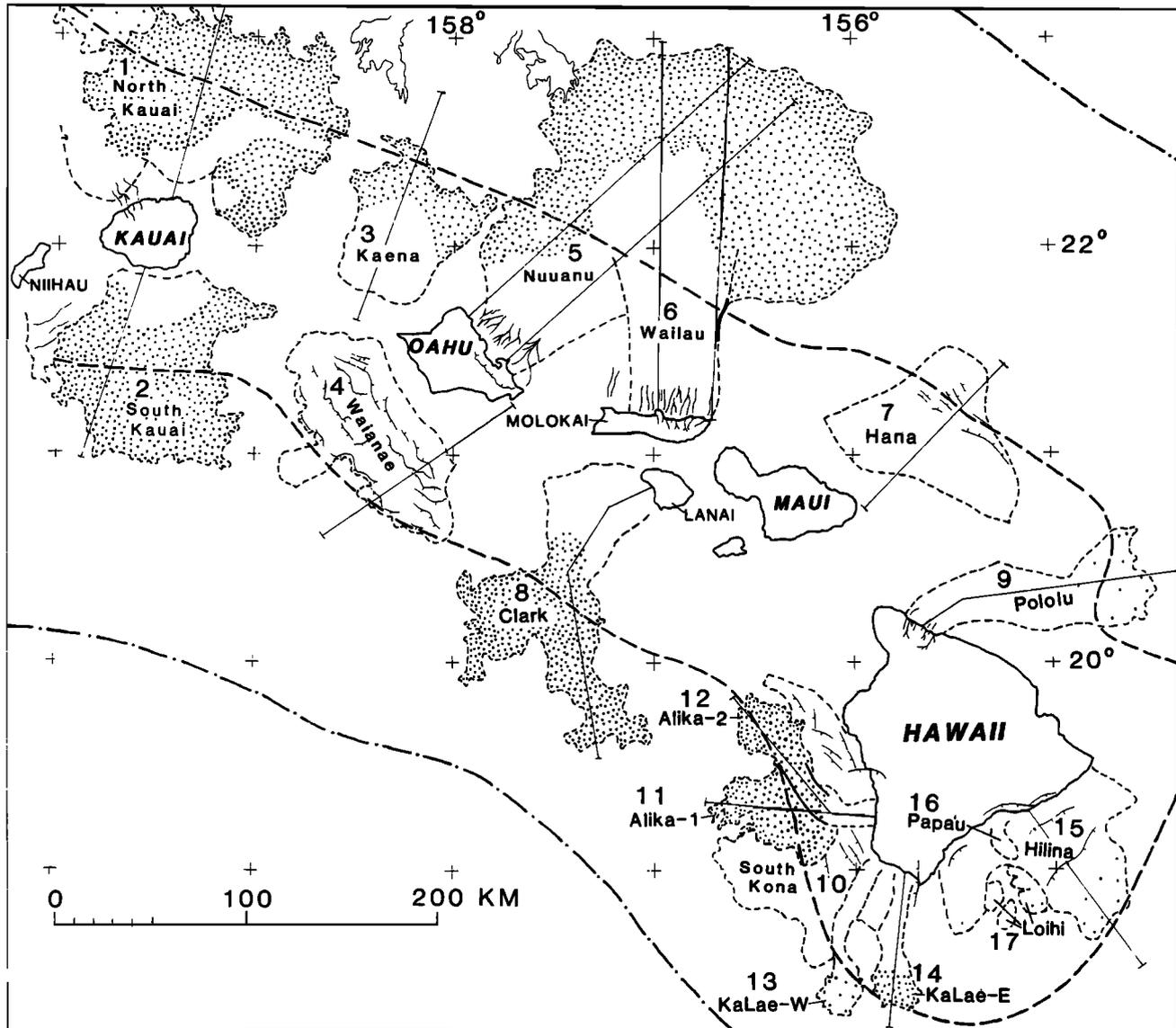


Fig. 2. Map of southeastern Hawaiian Ridge showing major slides bounded by dashed lines identified by number in text and Table 1; compare with Figure 1. Dotted area, hummocky ground (widely spaced where subdued); hachured lines, scarp; thin, downslope-directed lines, submarine canyons and their subaerial counterparts; heavy dashed line, axis of the Hawaiian Deep; dash-dotted line, crest of the Hawaiian Arch.

Hilina fault system related to gravitational sliding but also that the east rift zone of the volcano forms the upper margin of a large seaward moving slump. An analysis of geodetic and seismic evidence led Swanson *et al.* [1976] to support this notion of a mobile south flank of Kilauea and to add the concept that the south flank moves in response to forceful intrusion of magma into the rift zones. They assumed that the volcano flank is mobile downward to a depth of 5–8 km and suggest that the base of the mobile zone is not defined by a single plane or sole but that seaward displacement gradually decreases downward [see also Crossen and Endo, 1982]. Lipman *et al.* [1985] assessed the deformation associated with the 1975 earthquake ($M = 7.2$) on Kilauea's south flank and concluded that a deep-seated gravitational slump, partly related to forcible intrusion of magma into the rift zone at its head, could explain the geologic and geophysical observations. They suggested that the sole of the creeping mass was

the pre-Hawaiian seafloor and hence that the slide was as much as 10 km thick. Episodic movement of this landslide has been incorporated into a model for the long-term eruptive behavior of Kilauea [Holcomb, 1987].

Irregular blocky topographic welts extending downslope from giant amphitheaters on the submarine north flanks of Molokai and Oahu were interpreted as representing giant landslides by Moore [1964]. Macdonald and Abbott [1970, p. 365], however, questioned the prodigious size of these features, compared with previously known landslides, and proposed that the blocky seamounts are in situ volcanoes. The results of three separate oceanographic investigations by Mathewson [1970], Langford and Brill [1972], and Andrews and Bainbridge [1972, p. 112] caused each of them to reject the landslide hypothesis, with the latter two studies favoring the in situ volcano concept. Later measurements of the magnetic stratigraphy of East Molokai volcano led

TABLE 1. Hawaiian Submarine Slides

No.	Name	Location	Area, km ²	Length, ^a km	Width, ^b km	Type ^c
1	North Kauai	North Kauai	14,000	140	100	D (0.6–0.9)
2	South Kauai	South Kauai	6,800	100	50	D (0.6–1.2)
3	Kaena	NE Oahu	3,900	80	45	D
4	Waianaean	SW Oahu	6,100	50	80	S (1.0)
5	Nuuuanu	NE Oahu	23,000	235	35	D (0.8–1)
6	Wailau	North Molokai	13,000?	<195	40	D (0.7)
7	Hana	NE Maui	4,900	85	110	S
8	Clark	SW Lanai	6,100	150	30	D (0.5–1)
9	Pololu	North Hawaii	3,500	130	20	D
10	South Kona	West Mauna Loa	4,600	80	80	S
11	Alika-1	West Mauna Loa	2,300	88	15	D (0.9–1)
12	Alika-2	West Mauna Loa	1,700	95	15	D (2–5)
13	Ka Lae, west	South Hawaii	850	85	10	D
14	Ka Lae, east	South Hawaii	950	75	10	D (2)
15	Hilina	South Hawaii	5,200	40	100	S
16	Papa'u	South Hawaii	200	20	6	SF
17	Loihi	South Hawaii	500	15	10–30	L
Total			97,600			

Located by number in Figure 2.

^aLength of Waianaean and South Kilauea landslides omits indistinct irregular topography beyond steep toe.

^bWidth at head of landslide.

^cD, Debris avalanche; (number) is average number of hummocks appearing in GLORIA images per square kilometer; S, Slump; SF, sand rubble flow; L, three unclassified landslides.

Holcomb [1985] to support the notion that the north half of the caldera is missing and probably was carried beneath the sea by landsliding.

Bathymetric mapping of Papa'u seamount on the submarine south flank of Kilauea volcano led Moore and Peck [1965] to propose that this feature is a relatively young landslide, but Macdonald and Abbot [1970, p. 317] favored its origin as a shield volcano built on a rift zone with the same trend as that of Hualalai volcano. Later ocean floor photography, submersible diving, and analysis of dredged samples led Fornari *et al.* [1979] to conclude that Papa'u seamount is in fact a relatively recent landslide lobe with a volume of about 40 km³.

Irregular bathymetry suggesting submarine mass wasting on the west flank of Mauna Loa was investigated from the U.S. Geological Survey vessel *S. P. Lee* in 1976 and 1978. This work showed that a large region (including the south Kona slump and the Alika slide complex) between the previously enigmatic, subaerially exposed Kealakekua and Kahuku faults was involved in seaward sliding that reached 80 km offshore [Normark *et al.*, 1979]. Analysis of bathymetry and newly acquired Sea MARC II side scan sonar images of small areas led Fornari and Campbell [1987] to emphasize the importance of gravitational slumping and debris flows on the flanks of the Hawaiian Ridge.

METHODS

The GLORIA system, which generates sonographic images of the seafloor as much as 22.5 km to each side of the ship track, can resolve features as small as a few hundred meters across [Somers *et al.*, 1978]. In addition to GLORIA, 3.5-kHz echo-sounding and air gun seismic reflection profiling were employed during the survey of the seafloor extending 200 nautical miles (370 km) off the island shores. This regional mapping of the U.S. EEZ around Hawaii was

conducted aboard the British vessel *M/V Farnella* by the U.S. Geological Survey and the U. K. Institute of Oceanographic Sciences. The systematic and uniform nature of the GLORIA regional surveys has been particularly valuable in a comparative study of this type. The surface features and other sonic characteristics can be directly compared between different parts of the same landslide or from widely spaced slides of different age.

Early utility of the GLORIA system for investigating submarine landsliding in the Hawaiian region came in 1986 when sonar images collected off the southwest flank of Mauna Loa outlined a major region of submarine slope failure. These surveys delineated the Alika and Ka Lae debris avalanches, covering an area of about 5000 km² [Lipman *et al.*, 1988].

DESCRIPTION OF SLIDES

Along the 700 km reach of the Hawaiian Ridge from Kauai to Hawaii, well-exposed slope failures involve roughly one-half of the flanks of the ridge (Figures 1 and 2). Most of these slides have been known only since the recent completion of the GLORIA surveys, and our understanding of them is based largely on this data base of restricted scale. Finer blocky facies, and related turbidite deposits, doubtless extend beyond the slide margins as shown on Figure 2. We hope that this preliminary reconnaissance study will help guide more detailed surveys and sampling programs on specific slides. Brief descriptions of the individual landslides, from northwest to southeast along the ridge, are given below (see also Table 1).

The large Hawaiian slope failures can generally be divided into two structural types based on the classification of Varnes [1978], but the term landslide is retained as a general term for all slope failures including these submarine features. The first type (example, Hilina slump) includes the broad

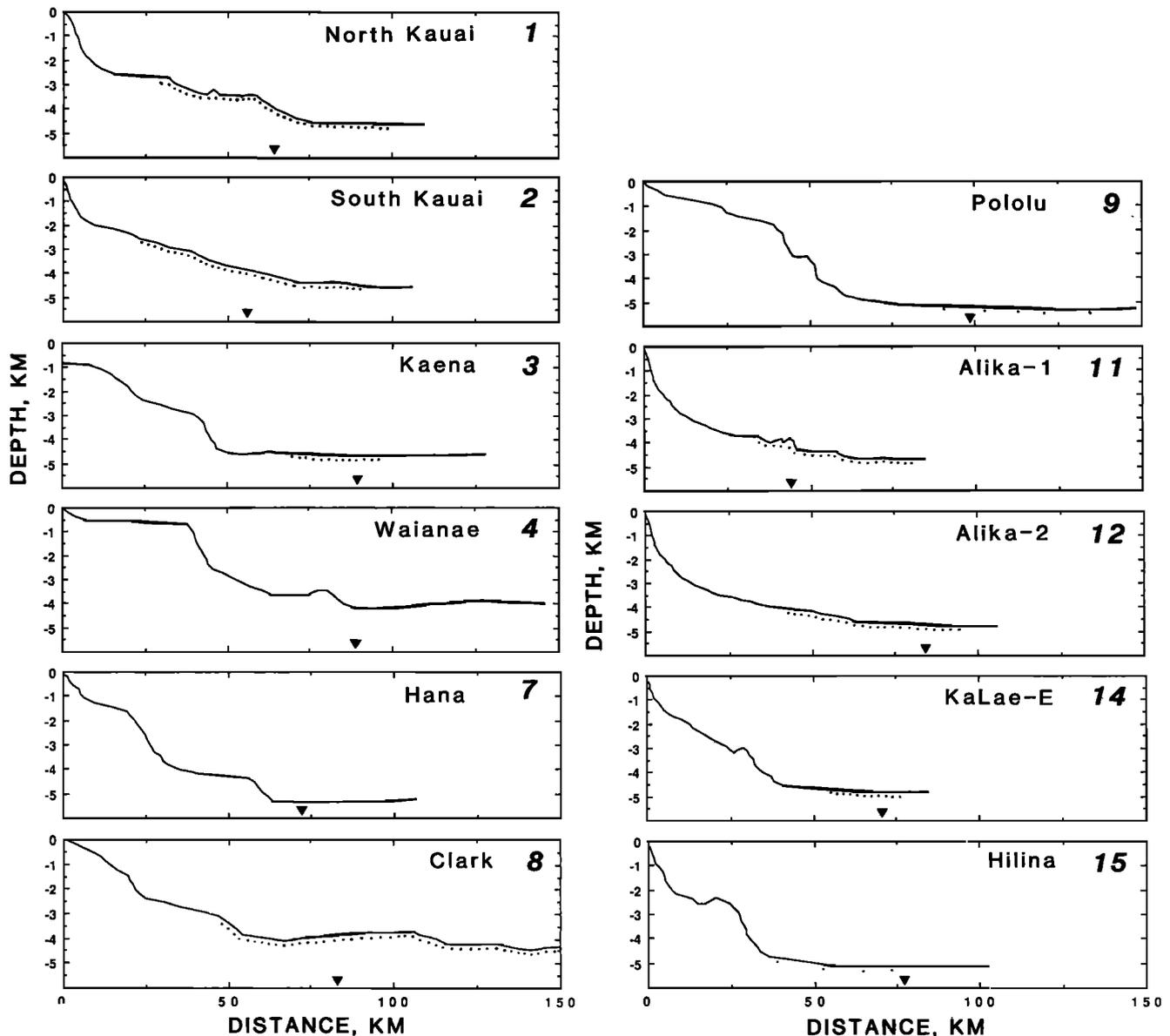


Fig. 3. Longitudinal profiles of landslides as numbered in Table 1 and shown in Figures 1 and 2. Hummocky terrain is indicated by dots (widely spaced where faintly visible), and lower end of dots is known lower limit of debris avalanche. Projected trace of Hawaiian Deep axis shown by triangle. Vertical exaggeration is 7.5.

and steeper failures that have a steep scarp at their toe, are cut by transverse faults into a few large blocks, and commonly lack a well-developed amphitheater at their head. The second type (example, Nuuanu debris avalanche) includes the narrow and more far-reaching landslides that show well-developed amphitheaters in their upper parts, are broken into large blocks in their midsections, and terminate in an apron of distinctively hummocky terrain. The first type are termed slumps and appear to have moved slowly and intermittently. The second type are called debris avalanches and commonly represent a single episode of rapid failure.

Slide 1: North Kauai Debris Avalanche

The volcanoes on Kauai and Niihau completed their shield-building stage about 5 Ma and are the oldest volca-

noes in the study area [Clague and Dalrymple, 1987]. Consequently, the landslide features on their slopes are the most degraded owing to subsequent island subsidence, erosion, and sedimentation. GLORIA images show that a landslide complex covers 14,000 km² along the north and northeast sides of Kauai (Figures 1 and 2), and much of this terrain is characterized by acoustic returns suggestive of hummocky or blocky material. The margins of continuously hummocky terrain extend as far as 115 km from the island, and areas of more isolated blocks and hummocks appear 20–40 km farther seaward particularly on the northwest side. The irregular edge of the hummocky terrain apparently results from accumulation of overlying sediment. A broad ridge, possibly constructed of large landslide blocks, extends 120 km northeast from the north side of the island to beyond

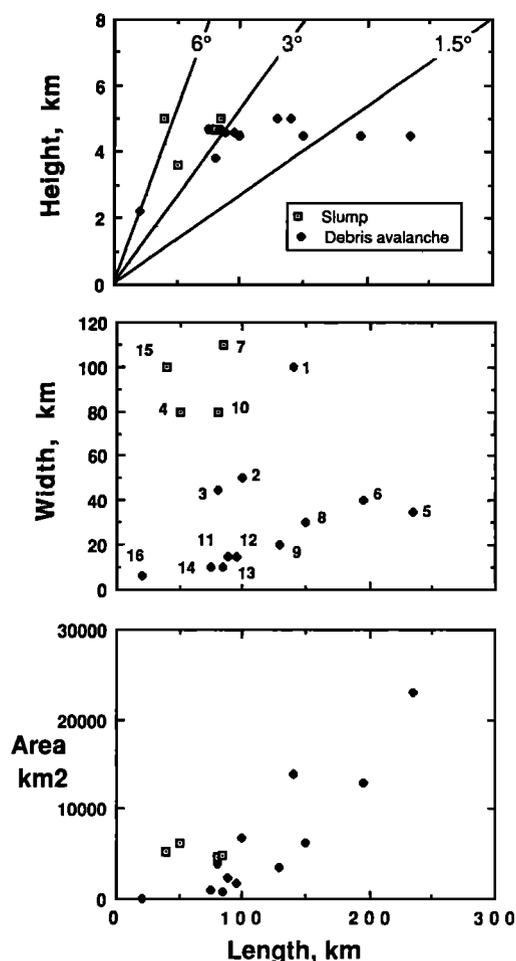


Fig. 4. Length versus height, width, and area for large Hawaiian slumps and debris avalanches identified by number in text and Table 1. Height is measured from toe to sea level; lines show overall gradient in degrees. Width of the head of the landslide and area are measured below sea level.

the axis of the Hawaiian Deep (Figures 1 and 3). Hummocky terrain is not recognized on much of the southeast side of this ridge, perhaps because of deep burial by turbidites moving down the volcano flank and ponding behind this ridge.

The upper submarine slopes off north Kauai are scalloped by three amphitheaterlike indentations, each of which probably represents the head of a major slope failure (Figures 1 and 2). The compound nature of the landslide is further suggested by the greater width of the deposits, when compared with others of similar length (Figure 4). The northeastern scallop fed the smallest landslide that moved northeast. The central scallop is the headwall of a major landslide which moved north and is characterized by the most distinctive hummocky terrain. The northwestern scallop is apparently the amphitheater of a third landslide, represented by hummocky terrain extending at least 140 km north-northwest from the island.

This northwestern amphitheater (offshore from the Napali coast area) is scored by several submarine canyons that extend from about 200 to 2300 m depth and can be projected upslope to major subaerial canyons [Shepard and Dill, 1966]. The submarine canyons begin about 3 km offshore and are best developed above a 900-m-deep bench about 8 km

offshore. The horizontal nature of this postslide bench indicates that it is a sea level feature carried down by regional subsidence. Hence slope failure took place when the island stood at least 900 m above its present level and was about twice as large. Therefore subaerial parts of the island were involved in the failures. Likewise, much of the canyon cutting was subaerial, but extension of the canyons below the 900-m-deep bench may have resulted from submarine erosion processes.

Slide 2: South Kauai Debris Avalanche

South of the island of Kauai is a broad area of blocky and hummocky terrain, about 80 km wide, that extends 110 km offshore and covers an area of 6800 km² (Figures 1, 2, and 5). This deposit was apparently derived from a series of debris avalanches originating from both the south flank of Kauai volcano and the east flank of Niihau volcano on the west. The eastern part of the deposit is apparently overlain by lavas belonging to a younger volcanic field between Oahu and Kauai. The west part of the deposit is augmented by slope failures from the steep southeast flank of Niihau. Linear sediment courses, one heading offshore from the Makaweli Graben, trend down the south flank of Kauai to the upper limit of imaged avalanche deposits (Figure 5). The axis of the landslide is marked by a broad longitudinal ridge that crosses the axis of the Hawaiian Deep and forms a low welt separating two basins.

Block size as seen on GLORIA images varies from pixel dimensions (50 × 125 m) to masses more than 5 km on a side. Most recognizable blocks are 500 m to 2 km in size, and commonly stand 50–100 m above the general surface as shown in 3.5-kHz echo records. The average maximum sediment thickness on top of the blocks based on 3.5-kHz echo records is about 25 m, assuming that this postfailure, acoustically transparent material has the same sound velocity as water. This is the greatest sediment thickness seen on slide debris reviewed in this study suggesting that this slide is among the oldest.

Slide 3: Kaena Debris Avalanche

The Kaena debris avalanche heads in a broad amphitheater 40 km wide on the north side of the Kaena Ridge, the submarine ridge extending northwest from Oahu toward Kauai (Figure 1). The landslide moved north more than 90 km into the axis of the Hawaiian Deep (Figure 2) where it produced a low swell between two deeper basins. The upper part of the slope failure deposit, below the 1-km-high amphitheater headwall, contains tilted fault blocks whose long axes are normal to the direction of movement. One such block is 12 km long and 5 km wide and stands more than 300 m above its base [Brocher and ten Brink, 1987, Figure 8].

Beyond the blocks, the hummocky nature of the deposit is subdued and poorly resolved in the GLORIA images. The individual hummocks are generally widely spaced, suggesting that the landslide is older, and more deeply sedimented, than the neighboring Nuuanu debris avalanche to the east. The feature shows some characteristics of a slump; it is comparatively wide relative to its length (Figure 4) and meets the deep seafloor at a steep zone more than 1 km high (Figure 3). The slope failure may have originated as a block

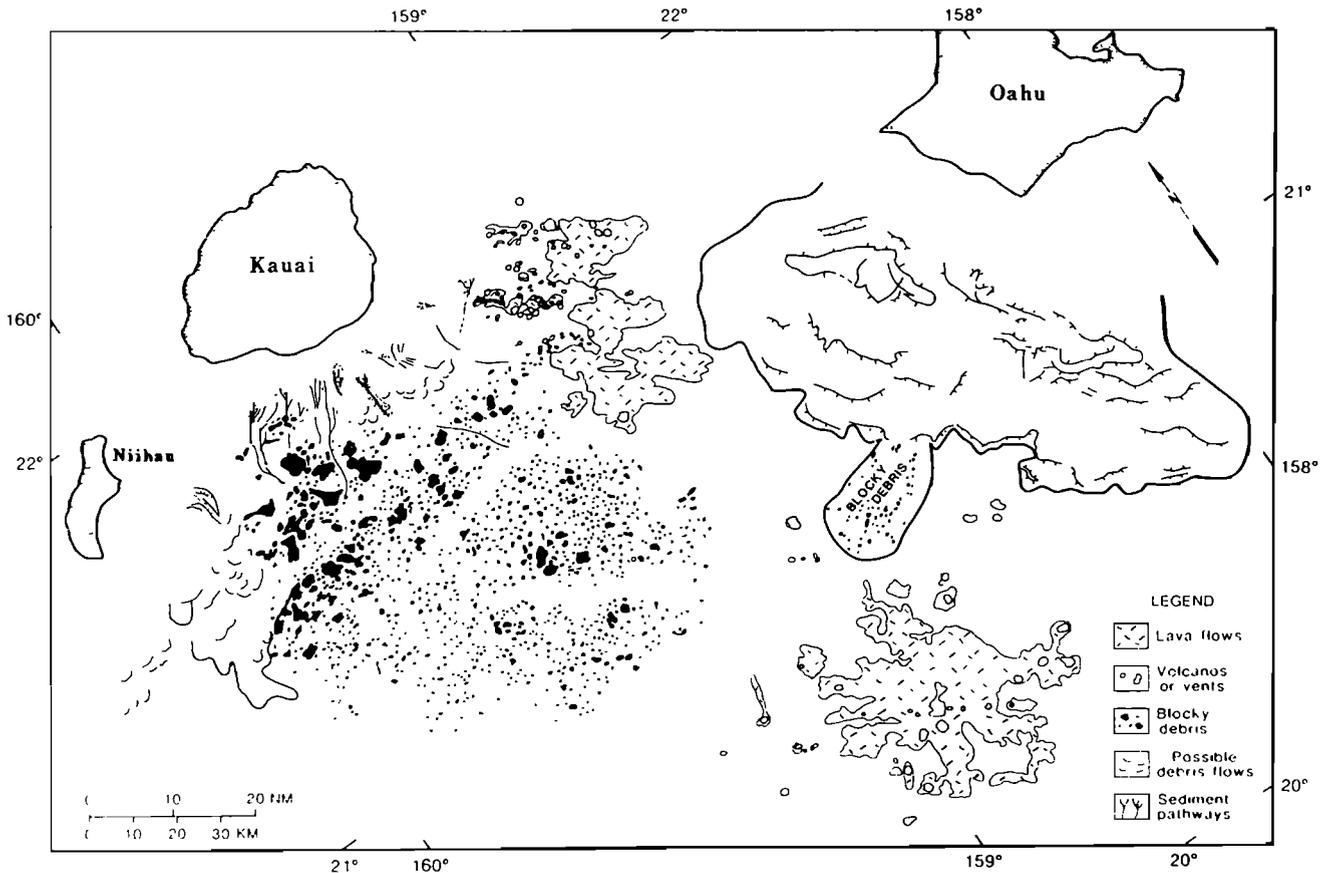


Fig. 5. Tracing from GLORIA image showing contrasting features of the South Kauai debris avalanche and the Waianae slump.

slump which partly disintegrated to form the hummocky debris avalanche apron.

Slide 4: Waianae Slump

The Waianae slump heads in a 130-km-wide zone extending from the Kaena Ridge northwest of Oahu to a point west of Penguin Bank (Figures 1, 2, and 5). This feature was originally called the West Oahu Giant Landslide [Hussong *et al.*, 1987]. However, to avoid confusion with other Oahu mass wasting features, we call it the Waianae slump, because of its location primarily on the flank of Waianae volcano which makes up western Oahu. However, the southeastern part of the slide apparently belongs to the presumably younger eroded and submerged volcano forming Penguin Bank, and the northwestern part may lie on a third volcano forming the Kaena ridge. The great width of the head of this slump (Figure 4), together with the fact that it apparently is rooted on more than one volcano, suggests that multiple slide events occurred. The surface features seen in GLORIA images, however, do not provide a basis for subdividing the slump.

The slump extends about 90 km offshore and covers an area of 6100 km². Large transverse scarps or ridges tens of kilometers long and spaced 5–20 km apart divide the slump mass into several large fault-controlled blocks (Figure 5). The slide generally lacks hummocky topography except in one lobe at its central, distal part. Most of the slide toe is steep and meets the adjacent seafloor abruptly (Figure 3).

The steep toe, large width–length ratio, and general lack of a hummocky surface indicates that it moved as a few large blocks rather than as a mass of broken fragments. However, the steep toe apparently failed near the center producing the smaller debris avalanche which moved 15 km farther out from the main slump mass.

Slide 5: Nuuanu Debris Avalanche

The Nuuanu debris avalanche covers 23,000 km² (the largest landslide mapped), on the northeast flank of Koolau volcano, the shield comprising the eastern half of the island of Oahu (Figures 1, 2, 6, and 7). The debris avalanche is about 230 km long as measured from its headwall at Nuuanu Pali on Oahu to its toe half way up the southwest flank of the Hawaiian Arch. In crossing the Hawaiian Deep, the debris avalanche (and the Wailau sister avalanche to the east) produced a welt across the deep bounding two depressions, the Kauai Deep on the west, and the Maui Deep on the east (Figure 1) [Wilde *et al.*, 1980].

After crossing the axis of the Hawaiian Deep, the avalanche moved uphill a distance of about 140 km from a present depth of at least 4600 m in the deep to its terminus at about 4300 m on the arch (Figure 6). Hence the vertical upslope transport is greater than 300 m. This does not consider the thickness of the slide in the deep which will increase the upslope distance, nor does it consider the extent of post-avalanching ridge subsidence, downwarp of the moat, and uplift of the arch.

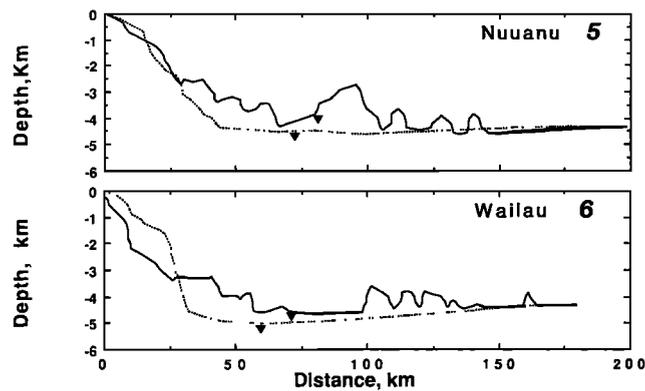


Fig. 6. Longitudinal profiles from Nuuanu and Wailau debris avalanches. Dotted profiles are on northwest margin of Nuuanu and east margin of Wailau landslides. Projected trace of Hawaiian Deep axis shown by triangles. See Figure 2 for location. Vertical exaggeration is 10.

It is important to note, however, that the deep and arch continue around the southeastern end of the Hawaiian Ridge ringing the youngest volcanoes of the chain on the island of Hawaii. Because the depth interval between the deep axis and the arch crest off southeast Hawaii is also several hundred meters, a similar slope probably existed off northeast Oahu, at the time of slope failure, and the Nuuanu avalanche must have traveled up that slope. The momentum necessary for the debris to ride up such a slope requires that it moved rapidly. A fast moving submarine landslide of this size probably produced strong bottom-scouring currents as well as large surface waves.

Two gravity cores taken about 130 km north of the north limit of the blocky facies imaged with GLORIA (at 24°04.9'N; 157°04.7'W and 23°42.1'N; 156°40.7'W) near the crest of the Hawaiian Arch bottomed in consolidated silty material that may be distal fine-grained sediment deposited by the Nuuanu debris avalanche. These turbidite-like deposits overlie the voluminous flood basalts that cover this part of the Hawaiian Arch [Shaw and Moore, 1988]. The sediments are apparently thin (less than a few meters), however, because the GLORIA sonic pulses readily penetrate them and reveal the subjacent lava flow surfaces.

A giant amphitheater above the debris avalanche deposits is defined by bathymetric contours deflected as much as 10 km shoreward from the projected original shape of the volcanic shield (Figure 1). The apparent thickness of material removed from this amphitheater is 700–1100 m. Below the amphitheater is a weltlike mass studded with giant blocks that are tens of kilometers in maximum dimension and rise 0.5–1.8 km above the regional slope (Figures 1, 6, and 7). These blocks commonly have tilted flat summits; they apparently originated in the amphitheater and have moved 50 km or more to their present position. The largest of these blocks, Tuscaloosa seamount, is 30 km long and 17 km wide and has a broad flat summit about 1.8 km above its base [Langford and Brill, 1972]. The tabular aspect of such large, yet thin, blocks requires that they slid to their present position.

The distal 50 km of the debris avalanche deposit is well-defined on the GLORIA images which show scattered blocks ranging up to about 1 km in size (Figure 7). The

average maximum thickness of sediment on these blocks (as determined from 3.5-kHz echo records) is 9 m.

Longitudinal bathymetric profiles that show the size and height of the large blocks in the central part of the Nuuanu debris avalanche led Moore [1964] to postulate that the bases of the large blocks probably extend somewhat farther below the general surface than they rise above it. This model suggests a thickness of about 2 km for the landslide. Insights on debris avalanche thickness are also obtained by comparison of longitudinal profiles down the axis of the avalanche with parallel profiles on the side (Figure 6). These profiles suggest an average minimum thickness of the central part of the avalanche below the amphitheater of about 400 m and an average of perhaps 200 m over the entire landslide area.

A detailed seismic reflection and refraction line crosses the western part of the debris avalanche about 100 km north of the east cape of Oahu near the axis of the Hawaiian Deep (line ESP-9 of Watts *et al.* [1985] and Brocher and ten Brink [1987]). This 100-km-long east-west line extends from the eastern part of the Kaena debris avalanche deposit to the western part of the Nuuanu deposit. Analysis of the seismic returns indicates that the total moat fill is about 2.6 km thick, but a higher-velocity layer (0.9 km thick) is sandwiched between two lower-velocity layers (a 1.5-km layer below and a 0.2-km layer above). We regard the higher-velocity layer as landslide material, which overlies and is overlain by pre-failure and post-failure sediment (probably moat-filling turbidite material); if so, the average thickness of slide material here is 900 m. Of the foregoing thickness limits, we favor the conservative estimate of 200 m over the entire area of the slope failure (23,000 km²), which yields a volume estimate of about 5000 km³ for the debris avalanche.

In the region where the Nuuanu debris avalanche and the neighboring Wailau avalanche on the east meet, the major elongate blocks are more nearly normal to the long axis of the Nuuanu avalanche. Moore [1964] suggested that the Nuuanu slide was younger than the Wailau slide, because the long axes of landslide blocks are more commonly perpendicular to the movement direction of the Nuuanu landslide. However, GLORIA images (Figure 7) reveal a field of hummocky terrain typical of distal debris avalanche facies east of the upper Wailau debris avalanche. This material is probably not a part of the Wailau avalanche because of its location off to the side and outside of a levee on the right (east) side of the Wailau debris avalanche; hence it must be part of the great distal debris fan of the Nuuanu debris avalanche. If the Wailau avalanche had been emplaced first, finer material from the Nuuanu avalanche could not likely have crossed it. For this reason we now consider the Wailau debris avalanche to be younger than the Nuuanu debris avalanche.

The Wailau landslide apparently impinged on the blocky deposit of the Nuuanu debris avalanche, but did not change the orientation of the massive blocks of the older feature. The younger relative age inferred for the Wailau debris avalanche is also in agreement with the relative ages of the two volcanoes that spawned these slope failures. The Nuuanu debris avalanche deposits appear to overlie older lava of the North Arch volcanic field and in turn are overlain by younger lava of the same volcanic field (Figures 2 and 7). Therefore age constraints for the North Arch lavas will better bracket the ages of these slope failures.

Several submarine canyons incise the headwall area of the



Fig. 7. Mosaic of GLORIA sonographs showing the Nuuanu and Wailau debris avalanches. Light shades designate regions of greater acoustic backscatter. The apparent elongation of hummocks parallel to ship tracks is an artifact caused by yaw of the towed vehicle.

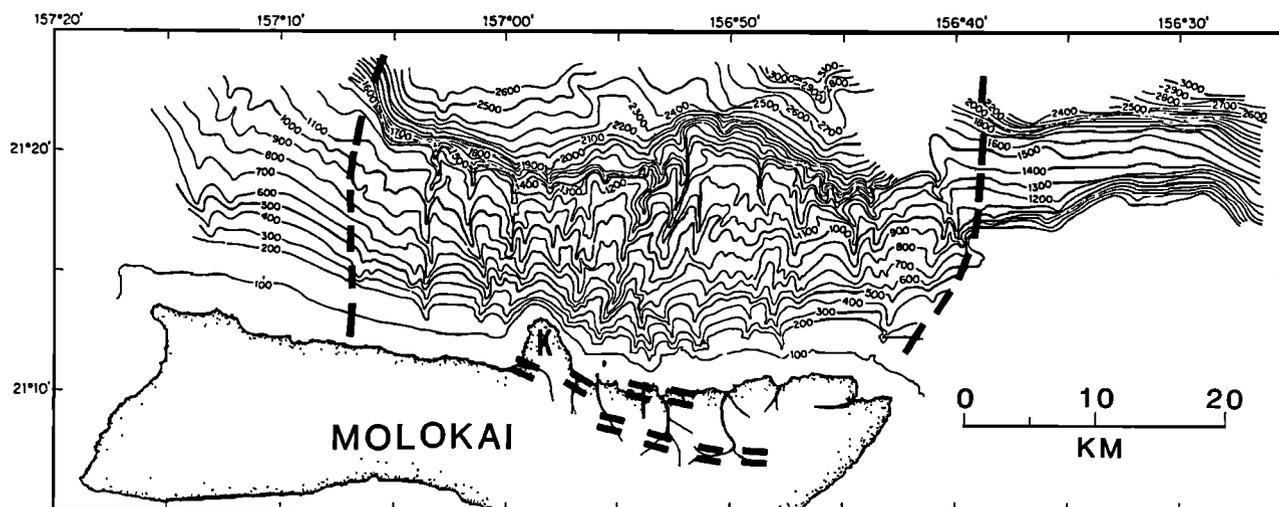


Fig. 8. Bathymetry north of Molokai from A. Malahoff and C. D. Hollister (written communication, 1987); contour interval 100 m. Margins of the Wailau debris avalanche shown by heavy dashed lines and trend of dike swarms is shown by double dashed lines [from *Stearns and Macdonald*, 1947]. Note that major canyons above sea level (shown by curved lines) are commonly aligned with submarine canyons. The postslide Kalaupapa volcano (K) grew in the slide amphitheater after subsidence was complete and partly filled the upper part of adjacent, submerged canyons.

Nuuanu debris avalanche (Figure 2). These submarine canyons feed sediment down the slope where it has accumulated to produce a 500-m-thick deposit behind and between the blocky seamounts which are a part of the upper landslide deposit. The fact that the northwest canyons are clearly associated with land canyons led *Andrews and Bainbridge* [1972] and *Coulbourn et al.* [1974] to conclude that the canyons were carved as subaerial canyons, and subsequently drowned during subsidence of the Hawaiian Ridge. This subsidence has been as much as 1800 m because the canyon morphology is well-developed to that depth [*Andrews and Bainbridge*, 1972]. Because the canyons are cut in the upper part of the slide scar, we infer that the ridge stood at least 1800 m higher at the time of failure.

The 40-km-long, northwest trending dike swarm of Koolau volcano [*Walker*, 1987] at the head scarp of the debris avalanche is aligned normal to the downslope direction of movement of the avalanche. Although the dikes generally tend to dip outward on the two sides of the swarm, most dip northeast at the southeastern part of the swarm. This dip was perhaps controlled by gravitational failure and removal of support from that side of the volcano.

Slide 6: Wailau Debris Avalanche

The Wailau debris avalanche, named for a major canyon on the north side of Molokai Island, has removed the north part of East Molokai Volcano and left a well-defined amphitheater 40 km wide at its head (Figures 1 and 7). The debris avalanche apparently overrode the south side of the older Nuuanu debris avalanche, and the extent of its distal facies and its area is unknown because of our inability to distinguish debris from the two features with available data. A known fact, however, is that the largest block (Tuscaloosa Seamount, Figure 1) in this crossover zone is not a part of the Wailau debris avalanche but belongs to the Nuuanu avalanche, as based on the composition of lava dredged from it (*M. S. Pringle*, written communication, 1988).

A regional terrace that abruptly steepens below 1300 m

water depth provides some constraints on the age of the slide (Figure 8). This terrace extends completely across the headwall amphitheater of the Wailau landslide, and caps a scarp about 1 km high [*Shepard and Dill*, 1966]. The nearly horizontal nature of this bench indicates that it is a sea level feature that postdates the debris avalanche. Hence the failure occurred when the volcano stood at least 1300 m higher than at present, or nearly 3 km above sea level. The terrace was either built by (1) formation of a lava delta during the end of shield-building activity of East Molokai volcano, or (2) growth of a particularly vigorous carbonate reef during a period of stable shoreline conditions. The reef hypothesis is supported by the appearance on the terrace of mounds in echo records that are interpreted as reefs [*Mathewson*, 1970].

The 1300-m-deep terrace is about 800 m deeper than the tilted H terrace which extends from east Molokai to east Maui [*Moore*, 1987] and is about 500 m deep east of the Wailau landslide headwall. This H terrace can be traced south of east Maui, where it is about 2 km deep. A comparison of the depth of this terrace with those of dated reefs west of northern Hawaii, which are currently subsiding at about 2.5 mm/yr, suggests that the H terrace was formed about 850 ka [*Moore and Campbell*, 1987]. If the Molokai subsidence rate was also 2.5 mm/yr during the period between formation of the 1300-m Wailau terrace and the H terrace, then the estimated age of the Wailau terrace is 320 thousand years older than the H terrace or about 1170 ka, and the landslide somewhat older. The landslide could not have been much older than this terrace, however, because postfailure volcano growth has been quite minimal. The estimated age of the Wailau debris avalanche is 1400 ± 200 ka, which is in general agreement with K-Ar ages of lavas that mark the end of the shield stage of the East Molokai volcano about 1500 ka [*Clague and Dalrymple*, 1987].

The landslide scar is largely unfilled except that the small Kalaupapa volcano has grown within the upper part of it (Figure 8). This small volcano (which has been dated by

K-Ar at 570–340 ka [Clague *et al.*, 1982]) grew after completion of the 1300-m subsidence of the Hawaiian Ridge that followed landsliding. The lack of post-Kalaupapa subsidence is indicated by the fact that a notable slope change on the volcano flank that presumably formed at sea level is now only slightly below present sea level [Moore, 1987].

The present position of the great northern sea cliff of East Molokai is not entirely of landslide origin. Those parts of the cliff that flank the sea seem to stand, on average, about 500 m south of the part which is behind (and protected by) the Kalaupapa volcano. A similar recession (and heightening) of the sea cliff due to marine erosion must have taken place during the approximately one million years after the debris avalanche occurred and before growth of the Kalaupapa volcano.

Several major submarine canyons occur in the Wailau debris avalanche amphitheater and most of them head offshore from major canyons above sea level on East Molokai volcano (Figure 8) [Shepard and Dill, 1966; Mathewson, 1970]. Some of the canyons cross the Wailau 1300-m terrace and extend several hundred meters below it. The canyons were chiefly carved by subaerial stream erosion before the volcano subsided about 1300 m to its present depth. Subsequently, some submarine erosion extended the canyons below the terrace level, supporting the notion that the Wailau terrace is composed of softer sediment and coral, rather than harder lava.

If the Wailau terrace indeed is composed of sediment or coral, then the question is raised as to why sea level remained for such a long period at this level during terrace construction? A possible answer is that the ongoing isostatic subsidence of the volcano caused by the addition of volcanic material to the ridge was temporarily interrupted and reversed by the rapid and massive transfer of material from the crest to the flank of the ridge by mass wasting. Perhaps this mass transfer caused the ridge to stabilize for a period before resuming its subsidence. Because the debris avalanche volume is roughly 1000 km^3 (assuming a thickness of 75 m over an area of $13,000 \text{ km}^2$), it represents 10 ka of Hawaiian hotspot volcanic production at the current rate of $0.1 \text{ km}^3/\text{yr}$.

Slide 7: Hana Slump

The Hana slump is centered north of the east cape of Maui on the northeast flank of Haleakala volcano (Figures 1 and 2). The 4900 km^2 slump separates the two deepest segments of the Hawaiian Deep, the Maui Deep on the west and the Hawaiian Trough on the east [Wilde *et al.*, 1980]. The northeast directed feature shows no hummocky surface on GLORIA images. Instead, it is characterized by subdued transverse ridges, probably reflecting deeply sedimented block slide scarps, and a steep toe area (Figure 3). At the head of the slump, two broad amphitheaters separated by a low ridge suggest that two slope failures may have merged to produce this feature.

The western amphitheater is flanked on the north by a northeast trending ridge that may represent an older rift zone ridge. South of this ridge, the slump surface is defined by repetitive northwest trending, regularly spaced linear ridges 2–8 km long and 1–4 km wide. The slump surface, below the eastern amphitheater, is marked by larger northwest trending blocky ridges up to 25 km long and 8 km wide. This

southeastern segment of the slide is apparently more sediment covered than the northwestern segment and is regarded as the older feature.

Slide 8: Clark Debris Avalanche

The 6100 km^2 Clark debris avalanche can be traced from west of Lanai south for 150 km (Figures 1 and 2); its lower part divides into two lobes which flowed on either side of Clark seamount, a 1500-m-high volcanic ridge (apparently of Cretaceous age) situated beyond the axis of the Hawaiian Deep. The landslide heads in the steep southwest facing amphitheater west of the island of Lanai and is flanked on the north by Penguin Bank and on the south by the west plunging ridge south of Lanai. The head of the debris avalanche is apparently marked by the major northwest trending system of normal faults which transect the island of Lanai [Stearns, 1940]. These faults, which are dominantly downthrown toward the southwest, probably mark the upper limit of the region of mass failure.

The debris avalanche separates into two branches 80 km from the source region. The eastern branch, which moved almost directly south, extends an additional 80 km to terminate on the lower slopes of Perret and Jagger seamounts. The western branch extends 60 km southwest and ends after flowing between Dutton and Clark seamounts. These two branches may actually be two independent debris avalanches with the eastern branch originating from the north wall of the amphitheater on the south flank of Penguin Bank and the western branch originating to the east on the west flank of Lanai volcano.

Most of the surface of the debris avalanche in water deeper than 3.5 km is characterized by a speckled pattern on GLORIA images, indicating hummocky terrain with individual blocks generally less than a few kilometers in size. Maximum sediment thickness on these blocks averages 9 m, suggesting that the slide may be roughly the age of the Nuuanu slide which appears to be mantled with about the same amount of sediment.

The steep scarp west of Lanai at the upper part of the feature is surmounted at a depth of 1200 m by gently sloping terrain that is the subsided subaerial surface of Lanai volcano [Moore and Campbell, 1987, Figure 2]. On this gentle surface is a steplike series of five coral reefs. This general surface apparently was constructed by subaerial, post-avalanche volcanic activity, after which reefs successively grew as the region subsided below sea level. The lowest and oldest reef, at a depth of 1000 m, has been estimated to be about 650 ka [Moore and Campbell, 1987], which is therefore a younger limit on the age of the Clark slope failure. This observation requires a revision in the assumption that the wave-deposited gravels on Lanai (dated at 105 ka) and similar deposits on adjacent islands were deposited by a giant wave caused by a landslide (here named the Clark debris avalanche) on the ridge flank southwest of Lanai [Moore and Moore, 1988]. Age relations indicate, rather, that the giant waves probably were produced by movement of the Alika debris avalanche west of the island of Hawaii [Lipman *et al.*, 1988].

Slide 9: Pololu Debris Avalanche

Kohala volcano, which forms the northern peninsula of the island of Hawaii, has undergone a major slope failure

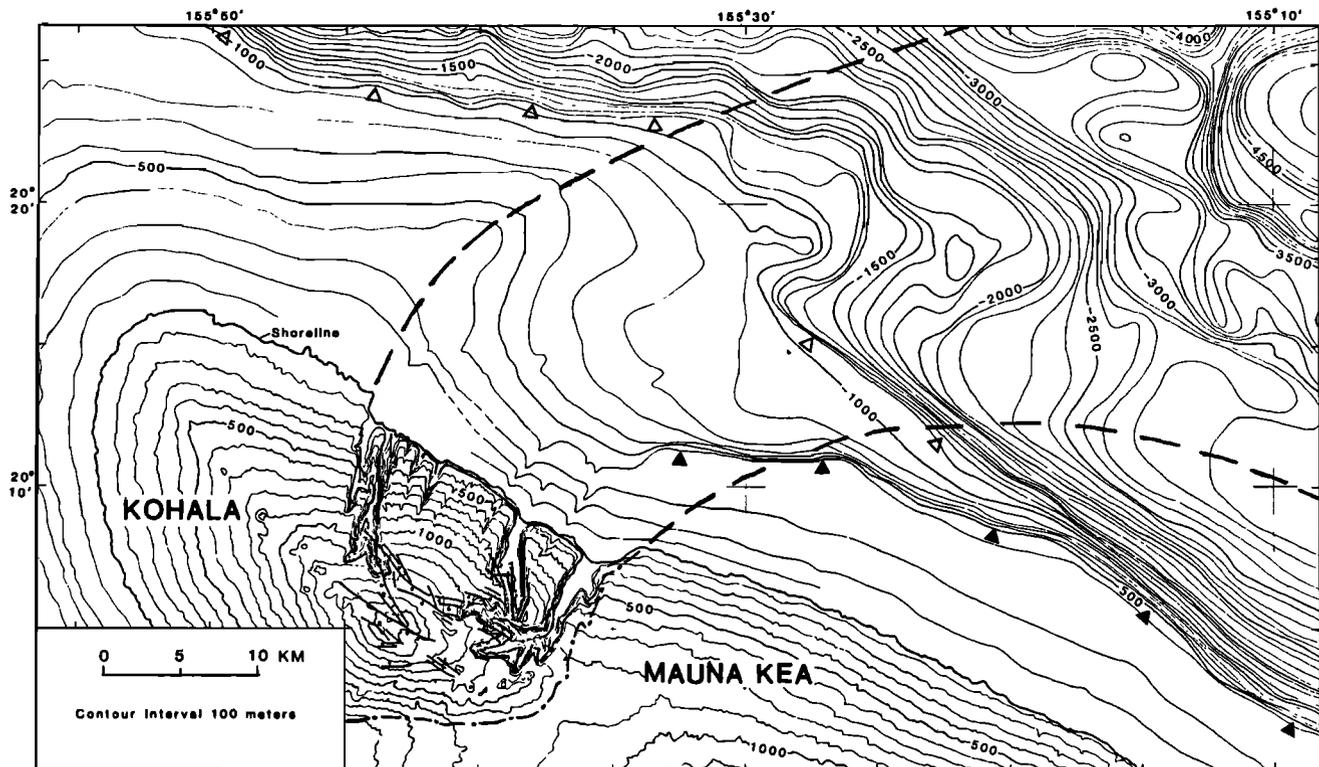


Fig. 9. Bathymetric-topographic map of the upper part of the Pololu debris avalanche (margins shown by dashed line). Also shown are the edge of Mauna Kea lava flows lapping onto Kohala volcano (dash-dotted line), the Kohala shelf break (open triangles), and the Mauna Kea shelf break (solid triangles). Bathymetry south of 20°22' contoured from *National Ocean Survey* [1981] and north of 20°22' from *Campbell* [1987], and land contours from USGS 1:100,000 1980 topographic map. Contour interval above and below sea level is 100 m.

that covers 3500 km² of its east flank (Figures 1, 2, and 9). This feature is called the Pololu debris avalanche after Pololu Valley, a major subaerial canyon in the northwest part of the headwall area of the feature. The landslide amphitheater extends from above sea level down to a depth of about 900 m and produces a marked reentrant of the shoreline 20 km long and extending 2 km inland (Figure 9). The submarine amphitheater is 10–20 km wide and its central depth is 150–400 m below the regional slope, causing the bathymetric contours to be displaced shoreward about 5 km.

The upper part of the amphitheater is incised by four submarine canyons [Moore, 1987] that can be traced on available bathymetric charts to depths somewhat below 400 m. These canyons are directly on line with prominent subaerial canyons on Kohala volcano. It is notable that the submarine canyons occur only within the landslide reentrant and the large subaerial canyons occur only upslope from the reentrant. Submarine canyons are not known to occur elsewhere on this volcano.

Downslope from the amphitheater an irregular bulge in the regional contours defines the main landslide welt. The debris avalanche turns east where it enters the trough between the steep flanks of Kohala and the east ridge of Haleakala on the north. The west flank of the landslide has dammed and isolated an unfilled depression, 13 km long, 6 km wide, and about 800 m deep, along the axis of this trough (Figure 1). This major trough, within which the depression occurs, must be an important sediment pathway, but surprisingly, the closed depression has not yet filled with sediment.

The lower part of the debris avalanche extends 60 km

beyond the mouth of the Kohala–East Haleakala trough and produced a broad fan in the Hawaiian Deep. The landslide can be traced a total of 130 km from the shoreline to its toe by means of the subtle hummocky character shown on GLORIA images.

The age of the landslide can be estimated from its relation to the abrupt change in slope of Kohala volcano at about 1000 m depth (Figure 9). This slope change or terrace is believed to have formed at the point in the evolution of the volcano when shield building ceased, that is, when copious lava flows no longer crossed and remade the shoreline. Subsidence rates and model ages of coral reefs indicate that this slope break was formed (at sea level) about 370 ka [Moore and Campbell, 1987]. The debris avalanche appears to be slightly older than the formation of this terrace because the terrace is partly constructed across the landslide surface, particularly on its eastern side (Figure 9). The terrace apparently formed after slope failure when subaerial lava flows erupted from southeastern Kohala (in a region now covered by Mauna Kea lava flows) crossed the subaerial part of the landslide and extended the shoreline. The limited coverage of the landslide topography by these lava flows indicates that mass failure took place not long before 370 ka and that the volcano stood 1000 m higher (or about 2.7 km above sea level) at the time of failure.

Kohala volcano is mantled by the postshield Hawi Volcanics which unconformably overlie the older shield stage Pololu Basalt. K-Ar ages determined on the Hawi Volcanics are 261 ± 4 ka to 61 ± 1 ka, and on the Pololu Basalt, 459 ± 28 ka to 304 ± 91 ka [McDougall and Swanson, 1972]. The

Pololu debris avalanche apparently occurred during the late shield stage during eruption of the Pololu Basalt; Pololu Valley itself, carved in the landslide scar, is partly filled by a late Hawi lava flow. Hence the age estimated from the morphology and relations of the 1000-m terrace is in accord with available radiometric ages of Kohala lavas.

The end of shield building is considerably younger on Mauna Kea volcano than on Kohala. A submerged terrace 7–15 km wide occurs for a distance of 70 km along the northeast flank of Mauna Kea volcano. The depth of the shelf break is 375 ± 15 m which corresponds with oxygen isotope stage 6 ending at 128 ka [Moore and Campbell, 1987]. This terrace (in addition to the older 1000-m-deep Kohala terrace) has also developed across the southern part of the Pololu landslide amphitheater for a distance of about 5 km (Figure 9). This terrace overlap apparently results from lavas from the growing Mauna Kea shield spilling across the subsided, slide-scarred east flank of Kohala. These Mauna Kea lavas also apparently filled and covered the submarine extension of Waipio valley, the prominent easternmost Kohala canyon.

A major northwest trending system of normal faults about 13 km long near the summit of Kohala mountain has been interpreted by Stearns and Macdonald [1946, p. 42] as “relict faults of a former caldera,” yet the system is more grabenlike than circular in plan. The fault zone is directly upslope from the offshore landslide and the shoreline reentrant, and most individual faults trend at right angles to the apparent direction of slope failure. Offset of the mapped faults is variable, but generally those on the southwest side of the swarm are downthrown toward the northeast, and those on the northeast side are downthrown toward the southwest. The resulting structural depression is regarded as a pull-apart graben developed at the head of the landslide.

Slide 10: South Kona Slump

A broad zone of blocky and irregular submarine terrain covering 4600 km² occurs along the west flank of Mauna Loa volcano extending 90 km south from the Kailua area to the south cape of the island (Figures 1 and 2) [see also Fornari and Campbell, 1987]. Subaerial indications of massive gravitational instability along this coast are the west-side-down displacements on two major normal faults, the Kealakekua fault toward the north margin of this zone and the Kahuku fault at the south side [Normark et al., 1979]. Several arcuate sets of buried subaerial faults (identified by anomalously steep slopes) occur in the 50-km coastal segment between these major faults and suggest repeated slump events.

The age of movement on the Kealakekua fault is bracketed by the fact that it cuts a lava sequence containing a soil dated by radiocarbon at 31 ka and in turn is overlain by a 13 ka unfaulted, 150-m-deep coral reef [Lipman et al., 1988]. However, the relationship of movement of the Kealakekua fault to the movement history of the south Kona slump is not obvious. Several offshore volcanic vents extending from near the shoreline to depths of 1000 m are apparently associated with the Kealakekua fault. The shallow vent was associated with a submarine eruption in 1877 and the two deeper vents, which erupted small fresh pillowed flows of similar chemistry, are assumed to be of the same age [Moore et al., 1985]. Presumably, dislocations at the north margin of this giant slump extend deep enough to have tapped the magma conduit system of Mauna Loa volcano.

The offshore slump terrain is marked by benches and scarps [Lipman et al., 1988] and shows a particularly steep regional slope near the base of the volcano flank [Mark and Moore, 1987]. This broad, oversteepened slump has undergone younger local collapse in four places to produce overriding debris avalanches that have moved out tens of kilometers beyond the base of the slump to cross the axis of the Hawaiian Deep. These include the Alika debris avalanche (phases 1 and 2) in the center of the slump and the west and east Ka Lae debris avalanches near the southern margin of the slump.

Slides 11 and 12: Alika Debris Avalanches

The Alika debris avalanches are probably among Hawaii's youngest [Lipman et al., 1988]. These two partly overlapping debris avalanches, designated phases 1 and 2, cover 2300 and 1700 km², respectively (Figures 1 and 2). They moved nearly 100 km from their source near the present shoreline to depths of 4800 m in the Hawaiian Deep. The estimated thickness of the debris avalanches is 50–200 m based on the size of the blocks and the height of the levees, yielding an estimated volume for both failures together of 200–800 km³.

Much of the debris avalanche deposit, except for that on the steeper slopes above about 4 km depth, is marked by closely spaced hummocky mounds. The individual mounds, which are generally equant in plan, are commonly up to a few hundred meters across and 25–75 m high, but some are several kilometers across and several hundred meters high. Over much of this terrain, the hummocks are closely spaced and similar in size and height, suggesting that they were all produced by the same processes in the same episode. Ocean floor photographs show thin (<0.5 m) sediment on rocky surfaces of the mounds, and no apparent sediment cover is detected by examination of 3.5-kHz echo records. Down-slope from the zone of closely spaced hummocks is a fringing zone from 5 to 20 km wide of isolated blocks or mounds. Where this zone on the phase 2 Alika debris avalanche was examined with a high-resolution side scan sonar system, the smallest resolvable mounds are about 50 m in diameter and the larger are 300–400 m across. Toward the distal margin of this zone the resolvable mounds are increasingly separated, commonly by 0.5–1 km. The outer zone of isolated blocks probably resulted from the removal of the muddy matrix of the distal part of the debris slide. Turbidity flows formed from segregated matrix probably carried finer-grained material away leaving the isolated blocks [Lipman et al., 1988].

The hummocky topography and marginal levees of the Alika landslides resemble the 1980 Mount St. Helens subaerial debris avalanche and suggest that a similar high emplacement velocity prevailed for the Alika features. The water disturbance produced by rapid avalanching of rock volumes of this size could produce ocean waves of large size. The general youth and location of the Alika phase 2 debris avalanche makes it a reasonable source for the extensive 105 ka wave deposits that have been documented on Lanai and nearby islands up to heights of 326 m above sea level [Lipman et al., 1988; Moore and Moore, 1984, 1988].

Slides 13 and 14: Ka Lae Debris Avalanches

Two south directed debris avalanches originated on the steep terrain of the South Kona slump west of Ka Lae (the

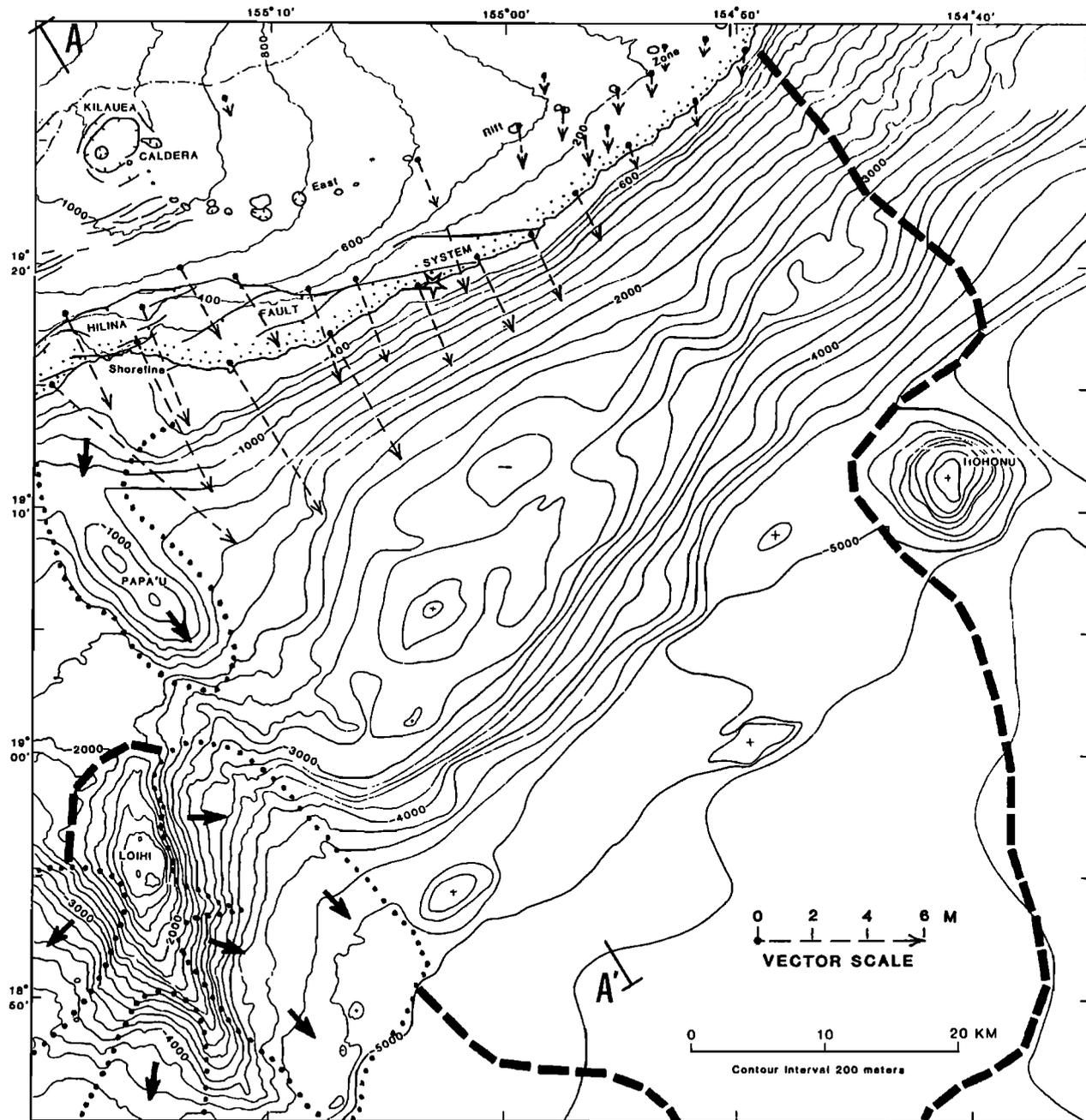


Fig. 10. Bathymetric-topographic map of the Hilina slump area. Bathymetry from *National Ocean Survey* [1981, 1982] and land contours from USGS 1:100,000 topographic map. Contour interval is 200 m. Heavy dashed line is margin of Hilina slump, and dotted lines bound smaller overriding Papa'u and Loihi landslides with movement direction shown by short arrows. Dash-dotted line is boundary between Kilauea volcano and Mauna Loa volcano to north. The epicenter of the magnitude 7.2 (1975) earthquake is shown by star, and dashed arrows are geodetically measured vectors showing horizontal movement associated with the earthquake [after *Lipman et al.*, 1985]. A-A' indicates location of section of Figure 11.

southmost point of the island of Hawaii) and moved down the slope on either side of Dana seamount to the Hawaiian Deep (Figures 1 and 2). The western Ka Lae debris avalanche (850 km²) extends 85 km south between Day and Dana seamounts and spreads out in the Hawaiian Deep as a broad field of faintly hummocky terrain. The east Ka Lae debris avalanche (950 km²) moved 75 km down a narrow chute between Dana seamount on the west and the steep west slope of the south rift zone ridge of Mauna Loa volcano on the east. The excellent preservation of the hummocky

field at its terminus suggests that this may be one of the youngest debris avalanches imaged by GLORIA.

Slide 15: Hilina Slump

The Hilina slump is an active landslide that involves most of the south flank of Kilauea volcano (Figures 1, 2, and 10). The slump covers about 5200 km², an area more than 3 times that of the volcano above sea level. Its location on an active volcano atop the Hawaiian hotspot places it at the locus of

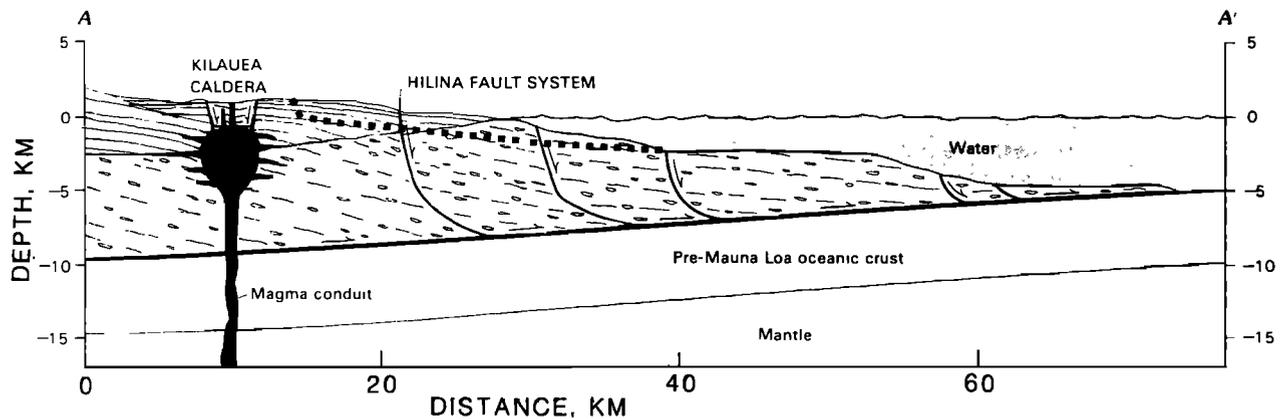


Fig. 11. Section through Hilina slump with no vertical exaggeration modified from Lipman *et al.* [1985]. Location of section shown in Figure 10. Dashed pattern forming base of volcano is subaerially erupted clastic rocks including phreatic explosion ash, littoral cone ash, flow foot breccias, minor pillowed flows, and pillow breccias, landslide debris, and turbidity flow deposits. Most of this material was probably erupted from Mauna Loa except for the uppermost few kilometers. Lined pattern forming upper part of volcano is subaerially erupted tholeiitic basalt, predominantly thin (2–10 m thick) pahoehoe and aa lava flows, but including ash layers. Lava from Mauna Loa predominates over that from Kilauea in north part of section. Dotted line indicates possible sole of future debris avalanche favored by steep regional slope.

sustained magmatic, seismic, and hydrothermal activity, all of which probably influence slope failure. This creeping zone is about 100 km wide; it extends west approximately to the south rift zone ridge of Mauna Loa and east as far as the east cape of the island. The slump is bounded at its upslope margin on the northwest by the southwest rift zone of Kilauea volcano and possibly the Kaoiki fault system (which separates Kilauea volcano from Mauna Loa volcano) and on the northeast by the east rift zone of Kilauea. The Hilina fault system is a major system of normal faults (downthrown to the south) parallel to the south coast and 5–10 km south of the north boundary of the sliding zone that apparently defines a major dislocation in the slump (Figures 10 and 11).

This vast, actively sliding slope is moving around the buttress formed by Loihi volcano and shows an anomalously steep slope where it meets the flatter seafloor both west and east of Loihi [Mark and Moore, 1987]. The southeast flank of the slide impinges on Hohonu seamount, an apparent Cretaceous volcano from which weathered basalt has been recovered [Moore and Fiske, 1969]. Marking the middle of the slump northeast of Loihi is a conspicuous midslope bench 2600–2800 m deep, 15 km wide, and 60 km long (Figure 10). The general backtilted nature of this bench, as shown by a closed depression on its shoreward side, suggests that it represents the top of a giant rotational slump block (Figure 11). Alternatively, features such as these “bulges” [Mark and Moore, 1987] may be giant folds, a notion suggested by the aseismic aspect of the middle and lower part of the Hilina slump [Crossen and Endo, 1982]. South of the steep slump toe are three small elongate seamounts (3–8 km long) which apparently represent the tops of buried slide blocks. The steep toe of the main slump zone is about 35-km offshore, but faintly hummocky terrain continues another 40 km into the axis of the Hawaiian Deep. This subdued, irregular backscatter on GLORIA images probably represents material deposited by early debris avalanches, now largely covered by finer volcanoclastic sediment.

Coincident with the 1975 magnitude 7.2 earthquake (Figure 10), much of the south coast for a distance of 60 km

subsided up to 3.5 m and moved seaward several meters; moreover a 30-km segment of the Hilina fault system was reactivated, and a small summit eruption broke out [Tilling *et al.*, 1976]. Geodetic measurements show that subaerial vertical and horizontal ground deformation related to the quake was as much as 3.5 m down and 8 m southeast, respectively (Figure 10), perhaps greater below sea level. Similar episodes have occurred previously, as in 1823 and 1868. If the 1975 displacements are representative, the long-term deformation of the south flank of Kilauea must be dominated by episodic large-scale gravitational slumping [Tilling *et al.*, 1976; Lipman *et al.*, 1985] with a seaward movement of the order of 0.1 m/yr. Analysis of the Love and Rayleigh seismic waves from the 1975 earthquake led Eissler and Kanamori [1987] to conclude that the bulk of the seismic radiation was produced by large-scale slumping of a part of the south flank of Kilauea.

The main hypocenter of the 1975 shock, and most aftershocks, were about 10 km deep, situated in the vicinity of the prevolcanic seafloor (Figure 11), showing that the entire volcanic mass down to the top of the Cretaceous crust was probably involved in block slumping [Crossen and Endo, 1982; Lipman *et al.*, 1985; Holcomb, 1987]. Much of the lateral movement may have been accommodated within poorly consolidated submarine volcanic rubble, material which earlier in its history had already experienced downslope sliding, and also within the pelagic sediment accumulated on the Cretaceous seafloor before Kilauea (and Mauna Loa) began growing.

Although the measured horizontal displacements are directed uniformly seaward toward the unbuttressed south flank of the volcano, the pattern of vertical and horizontal movements, on a smaller scale, reflects the summit caldera and rift zones and apparently is partly related to the movement of magma. The slumping results not only from the gravitational instability of the volcano flank but apparently also from forceful injection of magma into the summit reservoir and rift zones [Swanson *et al.*, 1976; Lipman *et al.*, 1985].

Slide 16: Papa'u Sand-Rubble Flow

The Papa'u landslide [Fornari *et al.*, 1979] is a 200 km² lobate debris tongue directly south of Kilauea caldera on top of the Hilina slump (Figures 1, 2, and 10). It is 19 km long, 6 km wide, and up to 1 km thick with a volume of about 40 km³ that has moved downslope from near the shoreline to a depth of 1850 m. Dredge hauls, remote camera photographs, and submersible observations indicate that it is largely composed of unconsolidated angular glassy basalt sand and blocks up to about 1 m in size. Because of the small size of the constituent fragments and the regular lobate shape, the feature has been termed a sand-rubble flow [Fornari *et al.*, 1979].

The landslide was apparently fed from a vast oversteepened, nearshore bank of fragmental basaltic material that was generated as subaerial lava flows from the summit region of Kilauea poured into the sea, quenched, and shattered. The landslide appears to have been emplaced during a single flow event as indicated by its simple lobate shape and the fact that multiple small debris flows would have filled in the saddle landward of the high point of the flow lobe. Estimates based on the present lava production of Kilauea volcano suggest that the landslide was emplaced several thousand years ago [Fornari *et al.*, 1979].

Slide 17: Loihi Landslides

Loihi is an active submarine volcano, growing against the south flank of Kilauea [Moore *et al.*, 1982]. The volcano is elongate north-south and has a generally flat summit about 1 km deep indented by several pit craters (Figures 1, 2, and 10). The site of the volcano at the extreme southeastern end of the Hawaiian Ridge in close proximity to the Hawaiian hotspot suggests that it will develop into the next major shield. The Hilina slump on the south flank of Kilauea is moving south against Loihi and divides around it on its course down to abyssal depths in the Hawaiian Deep.

Detailed bathymetric mapping utilizing multibeam sonar systems has revealed several amphitheatres on the flanks of Loihi (the "armchair indentations" of Malahoff [1987]; see also Fornari *et al.* [1988]), which are interpreted as landslide scars. Below these amphitheatres a zone of irregular, hummocky topography apparently representing landslide debris is evident, but these surveys did not extend far enough downslope to map and characterize the lower course of these landslides. The GLORIA images of Loihi are of limited value because of the small size of the failures and their location near steep slopes in relatively shallow water. These landslides cover more than three-quarters of the volcanic cone above 2500 m depth and the slide headwalls are extraordinarily steep, up to 40° in the first vertical kilometer (Figure 10).

STRUCTURE AND NATURE OF LANDSLIDES

The two general types of large Hawaiian landslides, slumps and debris avalanches [Varnes, 1978], range from the more coherent to the more disaggregated forms respectively. No doubt a whole spectrum of intermediate and compound landslides exists between these end-members.

The slumps are particularly steep and thick (up to 10 km) and are wide relative to their length (Figure 4). The overall gradient is generally greater than 3°, and the width/length

ratio exceeds one. The slumps maintain physical continuity, that is, they apparently never broke loose and moved far from their source. Several lack well-defined amphitheatres, probably because they were continually fed at their head by addition of volcanic material during the period of episodic movement. The slumps are commonly marked by transverse ridges and scarps that are generally bounded by faults but perhaps also in some places represent bulges or folds. The toe of the slumps commonly forms a distinct scarp where it meets the preslide seafloor.

Taking the Hilina slump as a model, these slope failures apparently developed during the active stage of shield building, when rift zones were well established, and owe their movement not only to the gravitational instability of the growing and widening shield but also to the lateral hydraulic jacking resulting from injection of magma into the rift zones. The slumps move intermittently at a rate controlled by the accumulation of lava flows on their upper reaches, as well as by magma intrusion in the rift zones which mark the boundary between the slumps and the more stable upper part of the volcano. Seismic shaking contributes to the development of these slope failures and is also a direct result of their movement.

Earthquake focal depths for the 1975 magnitude 7.2 Kilauea earthquake indicate that the base of the Hilina slump is close to the 10-km-deep, pre-Hawaiian seafloor surface. Because the top of the oceanic crust dips under the volcano at about 3° [Hill and Zucca, 1987], the sole of the slump most likely slopes upward at this angle, if indeed it is controlled by the pre-Hawaiian ocean floor. Upslope movement on this surface can only occur if the overall top of the slump dips seaward more steeply. Measurements of the average downslope angles of all the slope failures show that generally the slumps exceed 3°, whereas the debris avalanches have smaller slope angles (Figure 4).

The debris avalanches generally are longer and not as thick as the block slumps. The amphitheater at their head marks the source area from which the lower avalanche material was derived. The debris avalanches typically include a long middle and distal train of hummocky debris. Blocks up to tens of kilometers in size may be present immediately below the amphitheater, and blocks 1 km in size are common in the distal part of the deposit. The deposits are of variable thickness along their course, but the greatest average thickness in the blocky and hummocky sectors is probably 0.4–2 km, and the overall average thickness may be as little as 50 m in the smaller avalanches. Little debris may remain in the upper part of the amphitheater, and the distal part is thin and perhaps transitional to turbidity flows. Some of the debris avalanches may develop by partial or complete failure of the volcano slope already oversteepened by a slump and therefore provide a mechanism to reduce the overall gradient of the slump and carry material farther out beyond the base of the volcano. Unlike the slumps, most debris avalanches probably move at high velocity in a single episode, as shown by the uniform mound size and spacing in the distal hummocky facies and the fact that some avalanches have moved uphill several hundred meters.

LANDSLIDING AND VOLCANO DEVELOPMENT

Major landsliding occurs throughout the growth of Hawaiian volcanoes and apparently continues at a reduced rate

long after dormancy. More than half the surface area of Loihi, the youngest volcano in the chain, is extensively modified by landsliding even though its summit is still 1 km below sea level and the volcano is perhaps a hundredth the size of a mature shield.

Some of the largest landslides occur near the end of shield building, when the volcanoes stand 2–4 km above sea level and are erupting and extending their shorelines most vigorously. Mass wasting probably occurs, however, intermittently throughout most of the shield-building stage with the slope failure deposits being incorporated into the edifice as it grows so that little surface evidence remains of their existence. Fragmental landslide debris, associated with normal faults and slide slip planes, are believed to be important elements in the layered structure of these volcanoes from top to bottom.

Some of the debris avalanches (Wailau and Pololu) are characterized by prominent submarine benches 1 km or more deep which represent a postfailure shoreline (Figures 7 and 8). Such benches indicate that significant extension of the shoreline by accumulation of volcanic and/or coralline material occurred after movement. Additionally, they indicate that after failure, major regional subsidence occurred, a process that typically occurs only during, and shortly after the end of, shield building [Moore and Campbell, 1987]. The apparent hiatus in subsidence, which allowed considerable shoreline extension, may result from the rapid removal, by landsliding, of a substantial part of the volcanic load which causes the subsidence in the first place.

The subaerial structural features of the volcanoes commonly show an apparent relationship to the major slope failures of the submarine flanks of the volcanoes. Most of the major failures have moved in directions normal to the rift zones of the volcanoes on which they occur. The upper limit of most of the southward movement of the Hilina slump, for example, is bounded by the east and southwest rift zones of Kilauea volcano, and horizontal displacement vectors are subperpendicular to the rift zones [Lipman et al., 1985]. Moreover, some volcanic activity is related to landslide activity as shown by the 1877 submarine eruption in Kealakekua Bay that occurred near the north margin of the South Kona slump, the small Kalaupapa volcano growing within the amphitheater of the Wailau debris avalanche, and the 1975 Kilauea summit eruption apparently triggered by movement of the Hilina slump.

The Nuanu, Wailau, and Pololu debris avalanches and the Waianae, Hilina, and South Kona slumps (Figures 1 and 2) all head at, and have moved perpendicular to, the primary rift zones of their host volcanoes. Other slope failures, such as the Kaena debris avalanche and the Hana slump, head on the sides of long, shallow submarine ridges which apparently represent submerged rift zones. This rift zone/gravity failure relationship is part of the cause-effect paradox of the gravitational instability of these volcanoes. Rift zones form along the tensional zones at the head of landslides, but also landsliding is induced by the lateral shoving caused by magma injection into the rift zones.

Continued mass wasting long after volcanic activity has ceased is suggested by the modification of submarine slopes on the older volcanoes of the Hawaiian-Emperor Ridge toward the northwest. Detailed multibeam sonar bathymetry shows that the form of some volcanoes is modified by extensive mass wasting that has produced an angular, almost

skeletal, shape with sharp ridges radiating from the center (the "starfish" shape of Vogt and Smoot [1984]). Submerged 43 Ma volcanoes of this character [Smoot, 1985] near the Hawaiian-Emperor bend (about 3500 km northwest from Hawaii) probably owe their shape to dense dike rock of the rift zones exposed by downslope movement and removal of less resistant, weathered lava flow material between the rift zone dike swarms. Between the ridges deep amphitheatres above and fanlike debris welts below apparently mark the site of landslide events.

The widespread development of landslides of diverse type along the submarine slopes of the Hawaiian Ridge suggests that analogous slides should be present on other oceanic volcanoes constructed in deep water. Comparable bathymetric and sonar data exist only for a few areas. Isle de la Reunion, in the Indian Ocean, closely resembles the volcanic island of Hawaii. The presence of large-scale submarine landslides on the east flank of Piton de la Fournaise, the active volcano on La Reunion, was inferred by Duffield et al. [1982], in part by analogy with the structural features of Kilauea volcano in Hawaii. Recent Sea Beam bathymetric mapping, high-resolution sonar imaging (SLAR system; 0.3-m resolution), seafloor photography, and dredging along the submarine east flank of Piton de la Fournaise has confirmed the widespread presence of submarine landslide material [Lenat et al., 1989]. Hummocky terrain, similar to the Hawaiian debris slide deposits, extends at least 40 km offshore to 2500 m depth. Photographed and dredged material consists entirely of subaerially erupted, fragmented lava. The full dimensions of the large slumped submarine terrain on the east flank of La Reunion have not yet been delimited. Additionally, the regional bathymetry for the island shows steep submarine slopes along older parts of the island (Piton des Neiges) that grade down into a broad gently sloping apron on the lower slopes and suggest that mass failure is widespread around the entire island.

RELATION OF LANDSLIDES TO SUBMARINE CANYONS

The known submarine canyons offshore from the Hawaiian islands occur on the upper part of the major debris avalanches, and typically, these canyons extend shoreward to connect with major subaerial canyons. The subaerial counterparts of the submarine canyons include some of the largest canyons of Hawaii (Figure 2) including those of northwest Kauai, northeast Molokai (Figure 8), and northeast Hawaii (Figure 9). However, connection between subaerial and subsea canyons is everywhere obscured by a shallow submarine shelf, a few kilometers wide adjacent to the coast, in which no canyons occur. Seismic reflection profiling has shown that buried canyons, filled by sediment, are present beneath the shelf and provide connections between the subaerial and submarine canyons [Coulbourn et al., 1974].

Submarine canyons occur on the upper parts of the North Kauai, Nuanu, Wailau, and Pololu debris avalanches. These canyons are developed mostly above the shoreline terraces, produced shortly after slope failure occurred, which are 800, 900, 1300, and 1000 m deep on these four avalanches, respectively. Hence the submarine canyons were subaerially carved to about their present depth after landsliding. Subaerial canyon cutting in the hard basaltic rock was promoted by the oversteepened, and recently

stripped, slopes of the upper amphitheatres of the debris avalanches, especially in the amphitheatres formed on the windward (northeast) volcano slopes having high rainfall. Subsequent regional subsidence carried all but the upper parts of the canyons below sea level and permitted them to be alluviated in a coastal strip a few kilometers wide. In deeper water, away from the sediment source generated by subaerial canyon cutting and coastal erosion, the soft sediment filling the upper parts of the submarine canyons was cleared by submarine currents and turbidity flows, and some submarine erosion extended the canyons below the shoreline that existed initially after sliding.

RELATION OF LANDSLIDES TO THE HAWAIIAN DEEP

The Hawaiian Deep or moat is a bathymetric trough which flanks both sides of the Hawaiian Ridge about 100 km from the ridge center and loops around its southeast end. The axis of the deep is represented by a series of closed basins; the deepest seafloor in the Hawaiian region, exceeding 5600 m, occurs in one such basin north of Maui. The deep is produced by the downwarp of the oceanic crust by the weight of the volcanic ridge which pushes down and outward to produce a broad upbowing beyond the deep called the Hawaiian Arch. The distance between the arch crests on both sides of the ridge is about 500 km (Figure 2).

Virtually all of the debris avalanches extend outward across the axis of the deep and hence have apparently moved uphill onto the flank of the arch (Figures 2 and 3). The long-term effect of major mass wasting on the flanks of the Hawaiian Ridge is to transport volcanic material outward from eruptive vents and thus widen the ridge and force the axis of the deep to migrate outward. The continued subsidence of the ridge counters this effect, narrows the ridge and causes the axis of the deep to migrate inward. Hence the present configuration of the ridge and deep are a balance between these two processes.

The expected cycle of mass wasting during the life history of an individual volcano is for the largest and most far traveled debris avalanches to occur during the final stages of the shield building cycle when volcanic production peaks, and when the volcanoes stand highest above sea level. Before and after that time smaller landslides that travel shorter distances are expected to occur. Hence we infer that an onlap-offlap stratigraphic sequence of landslide and related deposits should occur on the inner flank of the Hawaiian Arch. The landslide deposits that lap furthest up the arch flank should be closest in age to the end of shield building of the host volcano.

Where the landslides cross and partly fill the axis of the deep, they produce low ridges that compartmentalize the deep into a chain of separate basins (Figures 1 and 12). The differences in depth between the tops of these landslide welts and the bottoms of the basins range from 0.4 to 1.5 km. These depth differences provide a conservative estimate of the thickness of the landslide welts where they have ponded in the deep (Figure 12).

CONCLUSIONS

A primary result of long range side looking sonar (GLORIA) surveys of the Hawaiian exclusive economic zone was to document the enormous scale of submarine slope failure

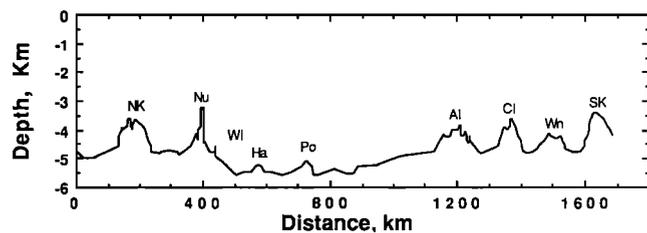


Fig. 12. Bathymetric profile along axis of Hawaiian Deep proceeding clockwise from the north side of Kauai, around the island of Hawaii, and northwest to the south side of Kauai (see Figure 2 for location). Humps produced by distal parts of landslides are North Kauai (NK), Nuuanu (Nu), Wailau (WI), Hana (Ha), Pololu (Po), Alika (Al), Clark (Cl), Waianae (Wn), and South Kauai (SK).

along the Hawaiian Ridge. Slumps and debris avalanches were found to cover nearly 100,000 km² of the ridge and adjacent seafloor from Kauai to Hawaii, an area more than 5 times the combined land area of the islands. Some of the individual debris avalanches are more than 200 km long and about 5000 km³ in volume, ranking them among the largest on Earth. These slope failures begin early in the history of individual volcanoes when they are small submarine seamounts, culminate near the end of subaerial shield building, and continue long after dormancy. Hence mass-wasting deposits and structures are no doubt an important element in the internal makeup of the volcanoes.

Most of the landslides flowed into, and locally across, the axis of the Hawaiian Deep which flanks the Hawaiian Ridge. Therefore mass wasting is the primary mechanism for filling the deep and widening the ridge. The eruptive behavior of the volcanoes can be modulated by landsliding, and the subaerial geometry of the slope failures is related to the structural features of the volcanoes including rift zones and fault systems. The landslides are of two general types, slumps and debris avalanches. The slumps are slow moving, wide (up to 110 km), and thick (about 10 km) with transverse blocky ridges and steep toes. The debris avalanches are fast moving, long (up to 230 km) compared to width, and thinner (0.05-2 km); they commonly have well-defined amphitheatres at their head and hummocky terrain in the lower part. Water disturbances caused by the rapid movement of the debris avalanches probably produced the high waves which have left deposits on Lanai and adjacent islands. Most submarine canyons were carved subaerially in the upper amphitheatres of debris avalanches. Subaerial canyon cutting in the hard basaltic rock was promoted by the oversteepened, and recently stripped, nature of these amphitheatres.

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REFERENCES

- Andrews, J. E., and C. Bainbridge, Submarine canyons off eastern Oahu, *Pac. Sci.*, 26, 108–113, 1972.
- Brocher, T. M., and U. S. ten Brink, Variations in oceanic layer 2 elastic velocities near Hawaii and their correlation to lithospheric flexure, *J. Geophys. Res.*, 92, 2647–2661, 1987.
- Bugge, T., R. H. Belderson, and N. H. Kenyon, The Storegga Slide, *Philos. Trans. R. Soc. London*, 325, 357–388, 1988.
- Campbell, J. F., Bathymetric atlas of the southeast Hawaiian Islands, *Tech. Rep. 87-1*, 21 pp., Hawaii Inst. of Geophys. Honolulu, 1987.
- Clague, D. A., and G. B. Dalrymple, The Hawaiian-Emperor volcanic chain, part I, *U.S. Geol. Surv. Prof. Pap.*, 1350, 5–54, 1987.
- Clague, D. A., D. Chen, R. Murnane, M. H. Beeson, M. A. Lanphere, G. B. Dalrymple, W. Friesen, and R. T. Holcomb, Age and petrology of the Kalaupapa Basalt, Molokai, Hawaii, *Pac. Sci.*, 36, 411–420, 1982.
- Coleman, J. M., and L. E. Garrison, Geological aspects of marine slope stability, northwestern Gulf of Mexico, *Mar. Geol.*, 2, 9–44, 1977.
- Coleman, J. M., and D. B. Prior, Mass wasting on continental margins, *Annu. Rev. Earth Planet Sci.*, 16, 101–119, 1988.
- Coulbourn, W. T., J. F. Campbell, and R. Moberly, Hawaiian submarine terraces, canyons, and Quaternary history evaluated by seismic-reflection profiling, *Mar. Geol.*, 17, 215–234, 1974.
- Crossen, R. S., and E. T. Endo, Focal mechanisms and locations of earthquakes in the vicinity of the 1975 Kalapana earthquake aftershock zone 1970–1979: Implications for tectonics of the south flank of Kilauea volcano, island of Hawaii, *Tectonics*, 1, 495–542, 1982.
- Damuth, J. E., and R. W. Embley, Mass-transport processes on Amazon Cone: western equatorial Atlantic, *Am. Assoc. Pet. Geol. Bull.*, 65, 629–643, 1981.
- Dana, J. D., *Characteristics of Volcanoes*, 399 pp., University Press, John Wilson, Cambridge, Mass., 1890.
- Dawson, A. G., D. Long, and D. E. Smith, The Storegga slides: Evidence from eastern Scotland for a possible tsunami, *Mar. Geol.*, 82, 271–276, 1988.
- Duffield, W. A., L. Stieltjes, and J. Varet, Huge landslide blocks in the growth of Piton de la Fournaise, La Reunion, and Kilauea Volcano, Hawaii, *J. Volcanol. Geotherm. Res.*, 12, 147–160, 1982.
- Eissler, H. K., and H. Kanamori, A single-force model for the 1975 Kalapana, Hawaii, earthquake, *J. Geophys. Res.*, 92, 4827–4836, 1987.
- Embley, R. W., Anatomy of some Atlantic margin sediment slides and some comments on ages and mechanisms, in *Marine Slides and Other Mass Movements*, edited by S. Saxov and J. K. Nieuwenhuis, pp. 189–213, Plenum, New York, 1982.
- Field, M. E., J. V. Gardner, A. E. Jennings, and B. D. Edwards, Earthquake-induced sediment failures on a 0.25° slope, Klamath River delta, California, *Geology*, 10, 542–546, 1982.
- Fornari, D. J., and J. F. Campbell, Submarine topography around the Hawaiian Islands, *U.S. Geol. Surv. Prof. Pap.*, 1350, 109–124, 1987.
- Fornari, D. J., J. G. Moore, and L. Calk, A large submarine sand-rubble flow on Kilauea volcano, Hawaii, *J. Volcanol. Geotherm. Res.*, 5, 239–256, 1979.
- Fornari, D. J., M. O. Garcia, R. C. Tyce, and D. G. Gallo, Morphology and structure of Loihi seamount based on seabeam sonar mapping, *J. Geophys. Res.*, 93, 15,227–15,238, 1988.
- Hill, D. P., and J. J. Zucca, Geophysical constraints on the structure of Kilauea and Mauna Loa volcanoes and some implications for seismomagmatic processes, *U.S. Geol. Surv. Prof. Pap.*, 1350, 903–917, 1987.
- Hitchcock, C. H., The geology of Oahu, *Geol. Soc. Am. Bull.*, 11, 23, 1900.
- Holcomb, R. T., Preliminary map showing products of eruptions, 1962–1974, from the upper east rift zone of Kilauea Volcano, Hawaii, *U.S. Geol. Surv. Misc. Invest. Map, MF-811*, 1976.
- Holcomb, R. T., The caldera of East Molokai Volcano, Hawaiian Islands, *Natl. Geogr. Soc. Res. Rep.*, 21, 81–87, 1985.
- Holcomb, R. T., Eruptive history and long-term behavior of Kilauea volcano, *U.S. Geol. Surv. Prof. Pap.*, 1350, 261–350, 1987.
- Hussong, D. M., J. F. Campbell, D. Hills, D. Peat, and J. Williams, Detailed mapping of the submarine geology of Oahu, Hawaii, using the SeaMARC/S system, *Eos Trans. AGU*, 68, 1336, 1987.
- Jacobi, R. E., Sediment slides on the northwestern continental margin of Africa, *Mar. Geol.*, 22, 157–173, 1976.
- Kenyon, N. H., Mass-wasting features on the continental slope of northwest Europe, *Mar. Geol.*, 74, 57–77, 1987.
- Langford, S. A., and R. C. Brill, Giant submarine landslides on the Hawaiian Ridge: A rebuttal, *Pac. Sci.*, 26, 254–258, 1972.
- Lee, H. J., Undersea landslides: Extent and significance in the Pacific Ocean, in *Landslides: Extent and Economic Significance, Proceedings of the 28th International Geological Congress Symposium on Landslides*, edited by E. E. Brabb, pp. 361–374, Balkema, Rotterdam, 1989.
- Lenat, J.-F., P. Vincent, and P. Bachelery, The off-shore continuation of an active basaltic volcano: Piton de la Fournaise (Reunion Island, Indian Ocean), *J. Volcanol. Geotherm. Res.*, 36, 1–36, 1989.
- Lipman, P. W., J. P. Lockwood, R. T. Okamura, D. A. Swanson, and K. M. Yamashita, Ground deformation associated with the 1975 magnitude-7.2 earthquake and resulting changes in activity of Kilauea volcano, Hawaii, *U.S. Geol. Surv. Prof. Pap.*, 1276, 45 pp., 1985.
- Lipman, P. W., W. R. Normark, J. G. Moore, J. B. Wilson, and C. E. Gutmacher, The giant submarine Alike debris slide, Mauna Loa, Hawaii, *J. Geophys. Res.*, 93, 4279–4299, 1988.
- Macdonald, G. A., The structure of Hawaiian volcanoes, *Verh. K. Ned. Geol. Mijnbouw. Genoot.*, 16, 274–295, 1956.
- Macdonald, G. A., and A. T. Abbott, *Volcanoes in the Sea: The Geology of Hawaii*, 441 pp., University of Hawaii Press, Honolulu, 1970.
- Macdonald, G. A., D. A. Davis, and D. C. Cox, Geology and ground-water resources of the island of Kauai, Hawaii, *Hawaii Div. Hydrogr. Bull.*, 13, 212 pp., 1960.
- Malahoff, A., Geology of the summit of Loihi submarine volcano, *U.S. Geol. Surv. Prof. Pap.*, 1350, 133–144, 1987.
- Mark, R. K., and J. G. Moore, Slopes of the Hawaiian Ridge, *U.S. Geol. Surv. Prof. Pap.*, 1350, 101–107, 1987.
- Mathewson, C. C., Submarine canyons and the shelf along the north coast of Molokai Island, Hawaiian Ridge, *Pac. Sci.*, 24, 235–244, 1970.
- McDougall, I., and D. A. Swanson, Potassium-argon ages of lavas from the Hawi and Pololu Volcanic Series, Kohala volcano, Hawaii, *Geol. Soc. Am. Bull.*, 83, 3731–3738, 1972.
- Moore, D. G., Submarine slides, in *Rockslides and Avalanches*, vol. 1, *Natural Phenomena, Dev. Geotech. Eng.*, vol. 14A, edited by B. Voight, pp. 563–604, Elsevier Scientific, New York, 1978.
- Moore, G. W., and J. G. Moore, Large-scale bedforms in boulder gravel produced by giant waves in Hawaii, *Spec. Pap. Geol. Soc. Am.*, 229, 101–110, 1988.
- Moore, J. G., Giant submarine landslides on the Hawaiian Ridge, *U.S. Geol. Surv. Prof. Pap.*, 501-D, D95–D98, 1964.
- Moore, J. G., Subsidence of the Hawaiian Ridge, *U.S. Geol. Surv. Prof. Pap.*, 1350, 85–100, 1987.
- Moore, J. G., and J. F. Campbell, Age of tilted reefs, Hawaii, *J. Geophys. Res.*, 92, 2641–2646, 1987.
- Moore, J. G., and R. S. Fiske, Volcanic substructure inferred from dredge samples and ocean-bottom photographs, Hawaii, *Geol. Soc. Am. Bull.*, 80, 1191–1202, 1969.
- Moore, J. G., and H. L. Krivoy, The 1962 flank eruption of Kilauea volcano and structure of the east rift zone, *J. Geophys. Res.*, 69, 2033–2045, 1964.
- Moore, J. G., and G. W. Moore, Deposit from a giant wave on the island of Lanai, Hawaii, *Science*, 226, 1312–1315, 1984.
- Moore, J. G., and D. L. Peck, Bathymetric, topographic, and structural map of the south-central flank of Kilauea volcano, Hawaii, *U.S. Geol. Surv. Map, I-456*, 1965.
- Moore, J. G., D. A. Clague, and W. R. Normark, Diverse basalt types from Loihi seamount, Hawaii, *Geology*, 10, 88–92, 1982.
- Moore, J. G., D. J. Fornari, and D. A. Clague, Basalts from the 1877 submarine eruption of Mauna Loa, Hawaii: New data on the variation of palagonitization rate with temperature, *U.S. Geol. Surv. Bull.*, 1663, 11, 1985.
- National Ocean Survey, Hydrographic survey sheets H-9855, H-9858 (south of Kalapana, southeast of Ka Lae) scale 1:80,000, Rockville, Md., 1981.

- National Ocean Survey, Hydrographic survey sheet H-9856 (Palma Pt. to Cape Kumukahi), scale 1:80,000, Rockville, Md., 1982.
- National Ocean Survey, Hydrographic survey sheet H-9974 (Off-shore Upolu Pt. to Laupahoehoe Pt.), scale 1:80,000, Rockville, Md., 1983.
- Normark, W. R., Ranger submarine slide, northern Sebastian Vizcaino Bay, Baja California, *Geol. Soc. Am. Bull.*, 85, 781-784, 1974.
- Normark, W. R., and C. H. Gutmacher, Sur submarine slide, Monterey Fan, central California, *Sedimentology*, 35, 629-647, 1988.
- Normark, W. R., P. W. Lipman, and J. G. Moore, Regional slump structures on the west flank of Mauna Loa volcano, Hawaii, paper presented at Hawaii Symposium on Intraplate Volcanism and Submarine Volcanism, Int. Assoc. of Volcanol. and Chem. of the Earth's Interior, Hilo, Hawaii, July 16-22, 1979.
- Prior, D. B., and J. M. Coleman, Active slides and flows in underconsolidated marine sediments on the slopes of the Mississippi delta, in *Marine Slides and Other Mass Movements*, edited by S. Saxov and J. K. Nieuwenhuis, vol. 6, pp. 21-49, Plenum, New York, 1982.
- Prior, D. B., and J. M. Coleman, Submarine slope instability, in *Slope Instability*, edited by D. Brunnsden and D. B. Prior, pp. 419-455, John Wiley, New York, 1984.
- Prior, D. B., J. M. Coleman, and E. H. Doyle, Antiquity of the continental slope, U.S. middle-Atlantic margin, *Science*, 223, 926-928, 1984a.
- Prior, D. B., B. D. Bornhold, and M. W. Johns, Depositional characteristics of a submarine debris flow, *J. Geol.*, 92, 707-727, 1984b.
- Shaw, H. R., and J. G. Moore, Magmatic heat and the El Nino cycle, *Eos Trans. AGU*, 69, 1553, 1564-1565, 1988.
- Shepard, F. P., and R. F. Dill, *Submarine Canyons and Other Sea Valleys*, 381 pp., Rand McNally, Chicago, 1966.
- Smoot, N. C., Guyot and seamount morphology and tectonics of the Hawaiian-Emperor elbow by multi-beam sonar, *Mar. Geol.*, 64, 203-215, 1985.
- Somers, M. L., R. M. Carson, J. A. Revie, R. H. Ege, V. J. Barrow, and A. G. Andrews, GLORIA II—An improved long range sidescan sonar, in *Proceedings of the IEE/IERE Subconference on Ocean Instruments and Communication: Oceanology International*, pp. 16-24, B.P.S. Publications, London, 1978.
- Stearns, H. T., Geology and ground-water resources of Lanai and Kahoolawe, Hawaii, *Hawaii Div. Hydrogr. Bull.*, 6, 177 pp., 1940.
- Stearns, H. T., and W. O. Clark, Geology and water resources of the Kau District, Hawaii, *U.S. Geol. Surv. Water Supply Pap.*, 616, 194 pp., 1930.
- Stearns, H. T., and G. A. Macdonald, Geology and ground-water resources of the island of Hawaii, *Hawaii Div. Hydrogr. Bull.*, 9, 363 pp., 1946.
- Stearns, H. T., and G. A. Macdonald, Geology and ground-water resources of the island of Molokai, Hawaii, *Hawaii Div. Hydrogr. Bull.*, 11, 113 pp., 1947.
- Swanson, D. A., W. A. Duffield, and R. S. Fiske, Displacement of the south flank of Kilauea volcano: The result of forceful intrusion of magma into the rift zones, *U.S. Geol. Surv. Prof. Pap.*, 963, 30 pp., 1976.
- Tilling, R. I., R. Y. Koyanagi, P. W. Lipman, J. P. Lockwood, J. G. Moore, and D. A. Swanson, Earthquakes and related catastrophic events, Island of Hawaii, November 29, 1975: A preliminary report, *U.S. Geol. Surv. Circ.*, 740, 33 pp., 1976.
- U.S. Naval Oceanographic Office, *Bathymetric Atlas of the North Pacific Ocean*, Washington, D. C., 1973.
- Varnes, D. J., Slope movement types and processes, in *Landslides, Analysis and Control, Spec. Rep. 176*, edited by R. L. Schuster and R. J. Krizek, pp. 11-33, National Academy of Sciences, Washington, D. C., 1978.
- Vogt, P. R., and N. C. Smoot, The Geisha Guyots: Multibeam bathymetry and morphometric interpretation, *J. Geophys. Res.*, 89, 11,085-11,107, 1984.
- Walker, G. P. L., The dike complex of Koolau volcano, Oahu: Internal structure of a Hawaiian rift zone, *U.S. Geol. Surv. Prof. Pap.*, 1350, 961-993, 1987.
- Watts, A. B., U. S. ten Brink, P. Buhl, and T. M. Brocher, A multichannel seismic study of lithospheric flexure across the Hawaiian-Emperor seamount chain, *Nature*, 315, 105-111, 1985.
- Wentworth, C. K., Estimates of marine and fluvial erosion in Hawaii, *J. Geol.*, 1, 117-133, 1927.
- Wilde, P., T. E. Chase, W. R. Normark, J. A. Thomas, and J. D. Young, Oceanographic data off the southern Hawaiian Islands, *LBL Publ. 359*, Lawrence Berkeley Lab., Univ. of Calif., Berkeley, 1980.
- Zucca, J. J., D. P. Hill, and R. L. Kovach, Crustal structure of Mauna Loa volcano, Hawaii, from seismic refraction and gravity data, *Bull. Seismol. Soc. Am.*, 72, 1535-1550, 1982.
- D. A. Clague, J. G. Moore, W. R. Normark, and M. E. Torresan, U.S. Geological Survey, 345 Middlefield Road, MS 999, Menlo Park, CA 94025.
- R. T. Holcomb, U.S. Geological Survey, School of Oceanography WB10, University of Washington, Seattle, WA 98195.
- P. W. Lipman, U.S. Geological Survey, Federal Center Box 25046, MS913, Denver, CO 80225.

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