SUBSIDENCE OF THE HAWAIIAN RIDGE

By James G. Moore

ABSTRACT

The great load placed on the ocean floor by the Hawaiian volcanoes, combined with the removal of magma from depth to build the volcanoes, has caused the lithosphere to sag and the Hawaiian Ridge to subside. Constraints on the amount and age of subsidence have been derived from tide gauge measurements, geophysical measurements, and studies of submarine canyons, submerged carbonate reefs, drill holes, and dredge hauls. Most of the volcanoes have subsided 2–4 km since reaching the sea surface, and the bases of the volcanoes have subsided 5–8 km. The bulk of subsidence is complete about 1 m.y. after initiation of volcanism on the sea floor, and about one-half of the upbuilding of the volcanoes is reduced by subsidence. A substantial but unknown part of volcanic growth results from intrusion of dikes and sills into the volcanic edifice, and hence the subsidence of different parts of the volcanoes varies.

Major sea-level terraces form while the volcanoes are actively growing above sea level. As the site of volcanism and subsidence migrates toward the southeast, these terraces become tilted toward the point of active volcanic loading. This tilting makes it difficult to correlate terraces from volcano to volcano along the Hawaiian Ridge by use of depth alone.

The young volcanoes at the southeast end of the Hawaiian Ridge have not yet completed their growth and subsidence. Hence volcanic ages derived from them provide estimates of volcanic propagation rates that are too high. Likewise, volumes measured for the volcanoes only above the level of the ocean floor indicate eruption rates for the older, subsided volcanoes that are too low. These factors raise doubts about inferred increases in volcanic propagation rates and eruption rates during the last few million years.

Volcanic centers on the Hawaiian Ridge are commonly aligned along two or more curved loci roughly parallel to the ridge and approximately 40 km apart. These loci may mark the edges of downbending of the two sides of the lithospheric subsidence trough, where fracturing conducts magma to the surface. The curved map trends along the loci may be the result of interference in these subsidence-induced fractures by major pre-Hawaiian seamounts rafted into the area of active volcanism.

Some of the intense seismic activity concentrated at the southeastern end of the Hawaiian Ridge may result from bending and dislocation of the volcanic pile as its southeastern part subsides more rapidly than the part adjacent on the northwest.

INTRODUCTION

The Hawaiian Ridge is formed by the most vigorous volcanic center on Earth and consequently is one of the youngest, highest, and steepest topographic features on the planet. Northwestward movement of the Pacific plate relative to the mantle hot spot that produces the volcanic center carries the volcanic products away at a rate of about 10 cm/yr, thus forming the ridge. The combination of the weight of erupted material and the removal of material from depth to feed the volcanoes causes the lithosphere to flex downward and the volcanic islands to subside. Subsidence is also partly caused by cooling and contraction of the lithosphere as it moves away from the mantle hot spot.

The subsiding ridge has depressed the adjacent ocean floor more than one hundred kilometers away from the ridge axis. The volcanic products of the ridge fill the inner part of this depression, but the outer part remains a moat-like depression called the Hawaiian Deep. The axis of this deep parallels the ridge on each side and forms an arc around the southeast end of the ridge with a radius of about 140 km (fig. 2.1). The center of curvature falls on the island of Hawaii between the summits of Mauna Loa and Mauna Kea Volcanoes.

Outside of the Hawaiian Deep is a broad low upbowing of the ocean floor called the Hawaiian Arch. The arch probably formed mainly because mantle material squeezed out from beneath the sinking ridge and buoyed up the fairly rigid bend beyond the deep, but also perhaps partly through heating and uplift of mantle rocks as they encountered the Hawaiian hot spot.

The Hawaiian Ridge extends from its active southeast end at Kilauea, Mauna Loa, and Loihi Volcanoes nearly 3,500 km northwest across the central Pacific to the bend where the Emperor Seamounts begin. From this bend the seamount chain continues north for another 2,500 km to its end near the Aleutian Trench. The ages of volcanoes increase systematically from Kilauea toward the northwest; the age is about 43 Ma at the Hawaiian-Emperor bend and about 70 Ma at the north end of the Emperor Seamounts (Clague and Dalrymple, chapter 1). Across this same span of distance the volcano peaks become systematically lower in elevation (fig. 2.2): the highest at the young southeast end are more than 4 km above sea level; a large group located 1,000–3,000 km northwest are close to sea level; and the northern Emperor Seamounts are commonly about 3 km deep. Although the shape and height of the volcanoes have been modified by erosion and by growth of carbonate reefs, these elevation differences result mainly from subsidence.

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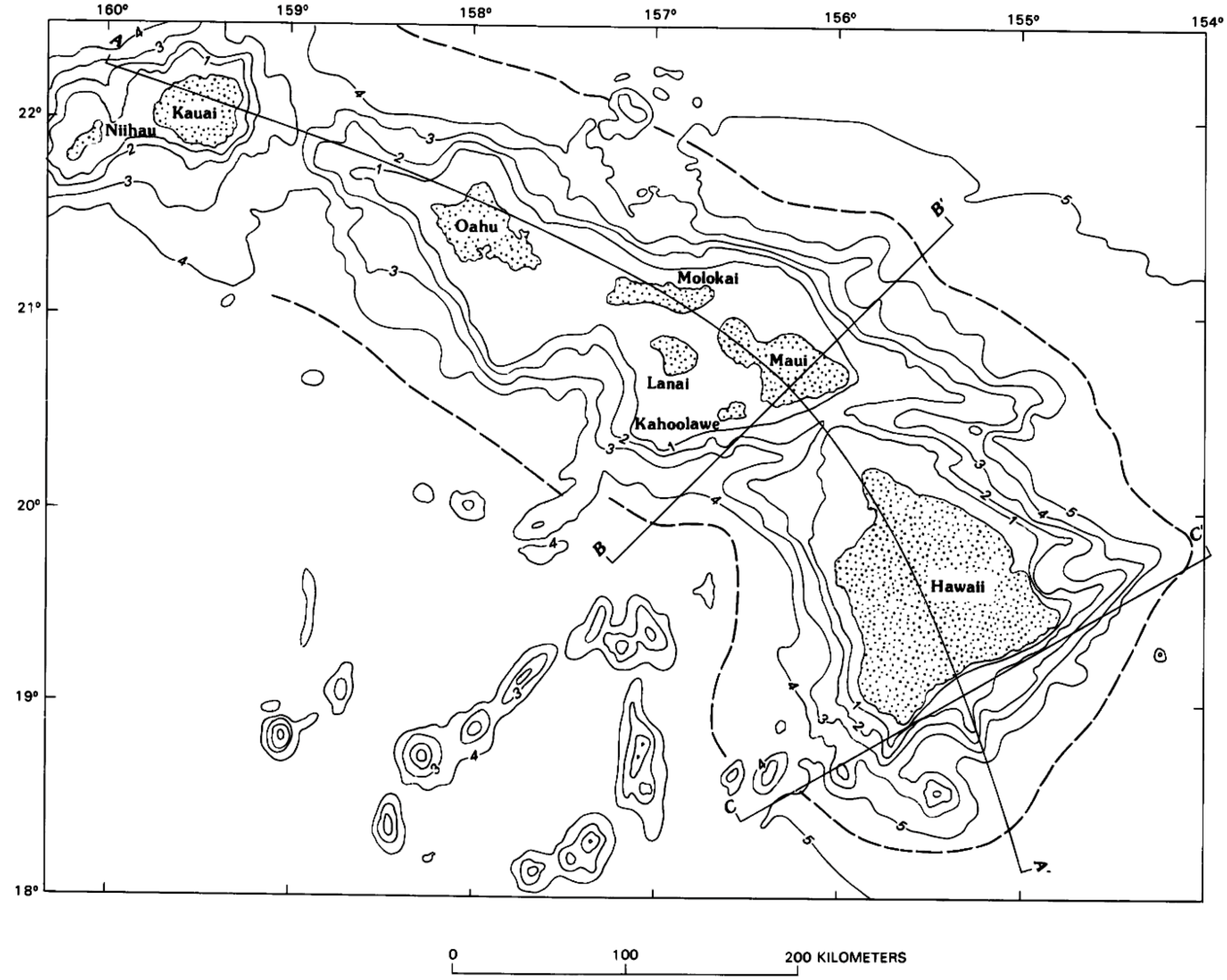

FIGURE 2.1.—Map of southeastern part of the Hawaiian Ridge showing axis of the Hawaiian Deep (dashed line) and section lines of figure 2.6. Depth contours in kilometers.

colleagues who have guided me through the intricacies and excitement of Hawaiian geology. The earliest and most effective of these were Donald Richter, Jerry Eaton, and Chester Wentworth.

TIDE GAUGE MEASUREMENTS

An analysis of tide-gauge data indicated that Honolulu on Oahu is approximately stable, because sea level at Honolulu is rising at a rate similar to that reported by many other stations that apparently reflect the worldwide eustatic rise (Moore, 1970). A reanalysis of Honolulu tidal data for the period 1935–83 indicates that sea level is rising there at a rate of 1.2 mm/yr. The close comparison of this rate with a recent estimate of global sea level increase of 1.5 mm/yr (Barnett, 1983) supports the notion that Honolulu is practically stable.

Some of the scatter inherent in tidal observations can be eliminated by comparing tide levels at other Hawaiian stations with those at the apparently stable Honolulu. Factors such as eustatic sea-level rise, barometric pressure, and sea temperature affect both Honolulu and other neighboring stations in the same way and hence are eliminated by using the differences in annual mean water levels for Honolulu and the neighboring stations. Such an analysis of measurements through 1967 indicated that sea level was rising at Hilo, Hawaii, at 4.1 mm/yr and at Kahului, Maui, at 1.7 mm/yr *relative to the assumed stable Honolulu* (Moore, 1970). Hence Hawaii and Maui were presumed to be undergoing absolute subsidence at those rates.

However, precise determination of sea level trends requires a long period of measurement. The 16 years of record for Kahului and the 21 years for Hilo utilized for the previous estimates were

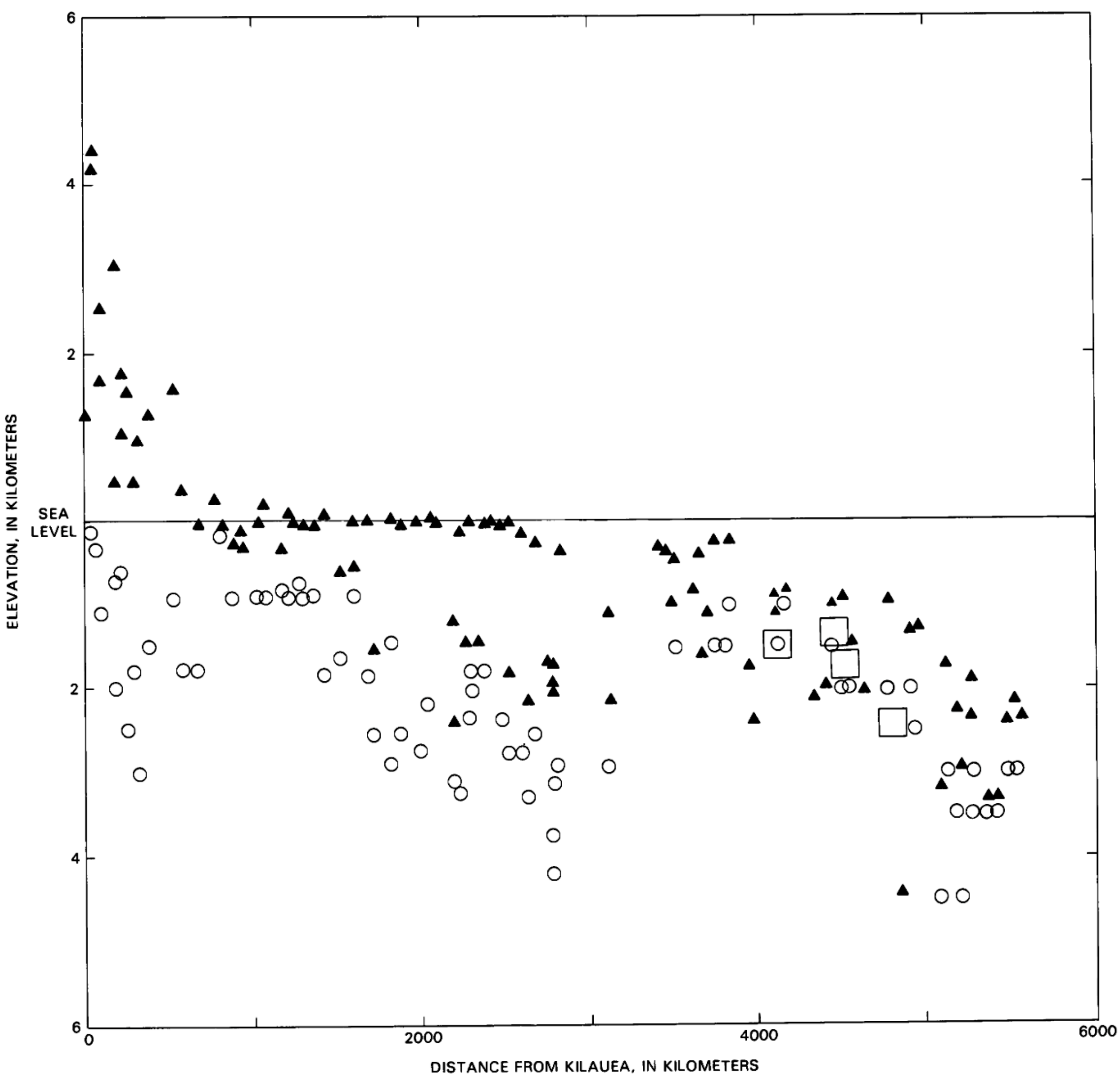

FIGURE 2.2.—Elevation and depth of volcanic features of the Hawaiian Ridge-Emperor Seamounts area. Triangles, tops of individual volcanoes; circles, depths of prominent terraces believed to have formed at sea level; squares, depth of penetration of drill holes on seamounts that encountered only subaerially erupted lava.

minimal. A reanalysis using the 17 additional years of measurements since 1967 provide a better data base for estimation of longer term trends in subsidence rates (fig. 2.3).

The primary measurement used is the annual mean of sea level at the various stations as reported by the U.S. National Ocean Survey. This annual mean is the stand of sea level relative to a fixed datum measured every hour and averaged for one year. The trends of differences of the annual mean of Hilo minus Honolulu and of Kahului minus Honolulu have been determined through 1983. Linear regression fits of this data indicate that Hilo is subsiding relative to Honolulu (assumed stable) at 2.4 mm/yr and Kahului at 0.3 mm/yr. The reason for the differences between the older estimates and these revised estimates is clear from the plot of annual means of sea level (fig. 2.3). The trend through 1967 was steeper than the longer trend through 1984. This change affects both the Kahului and Hilo stations, but is not a worldwide effect because the Honolulu means have been subtracted from the plotted data. The reason for this change in slope is unknown; it may reflect regional

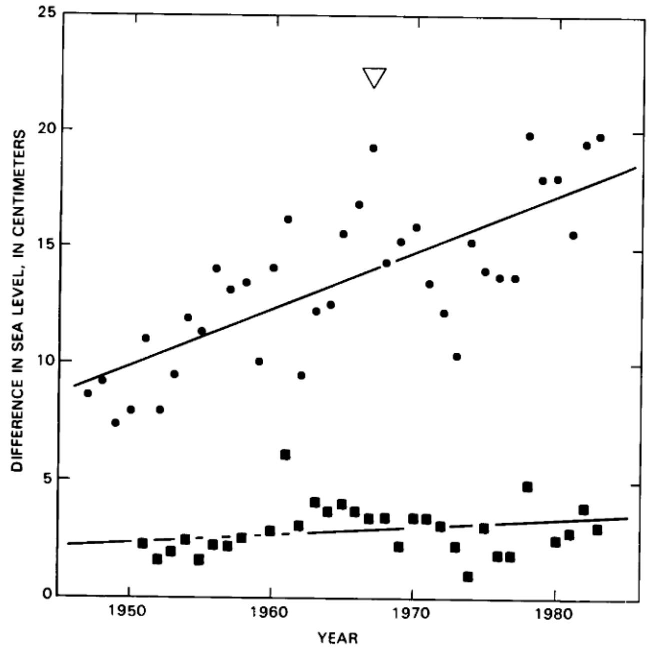

FIGURE 2.3.—Differences in annual means of sea level through 1983. Dots, means at Hilo, Hawaii, minus means at Honolulu, Oahu; squares, means at Kahului, Maui, minus means at Honolulu, Oahu. Open triangle, limit of data (1967) for analysis of Moore (1970).

variations in atmospheric or marine conditions on sea level or a real pulsation in subsidence rates.

In evaluating the subsidence of coastal stations such as Kahului and Hilo, the location of these stations relative to the deep subsiding trough underlying the Hawaiian Ridge must be considered (fig. 2.4). Subsidence varies between different locations on the island of Hawaii and probably reaches a maximum near the island center, where the base of the crust is depressed about three times as much as it is at Hilo on the coast. The surface subsidence at the island center may not actually be three times greater than it is at Hilo, however, because other processes taking place between the surface and the base of the crust, such as intrusion and variations in magma storage within shallow magma chambers, may affect surface subsidence.

DRILLING AND DREDGING

Many holes have been drilled on the Hawaiian Islands principally for the purpose of developing fresh water supplies. Most of these holes penetrate only a short distance below sea level, but a few deep research holes have been particularly valuable in providing information on the history of subsidence as well as on variations in sea level caused by other factors.

A hole drilled on the Ewa coastal plain, Oahu, reached a depth of 337 m and bottomed in red clinker and weathered vesicular subaerial basalt of the Koolau Volcano. The hole records a long period of subsidence and deposition of shallow marine and lagoonal muds, marsh deposits (lignite), and coral reefs, with intermittent periods of erosion and weathering (Stearns and Chamberlain, 1967).

Two holes were drilled on Midway Atoll in 1965. The deeper one penetrated limestone reef material until it encountered weathered basalt, overlain by basaltic conglomerate, at a depth of 384 m. The hole then went through a series of basaltic lava flows to the bottom at 504 m below sea level. The high vesicularity of these lavas (fig. 2.5) indicates that all were erupted subaerially.

A geothermal test well was drilled at an elevation of 182 m on the east rift zone of Kilauea Volcano in early 1976. This hole was 1,962 m deep and bottomed 1,780 m below sea level. The drill cuttings reveal the first appearance of pillow-like glass at 348 m below sea level, and the apparent transition zone between subaerially erupted and subaqueously erupted basalt is placed between 275 and 382 m below sea level (Stone, 1977). We will assume the change to be at 348 m. The age of the lavas encountered in the hole is not known, but it is evident that since the time the drill-hole region was at sea level, 182 + 348 = 530 m has been added to the upper surface of Kilauea Volcano. If the sea-level stand were the same then as now, the volcano has subsided 348 m. However, sea level may have been as much as 100 m below its present stand, in which case the volcano may have subsided only 248 m. One-half to two-thirds of the volcanic buildup of this part of the volcano has therefore been lost by subsidence. This ratio is similar to that for the southeastern Hawaiian Ridge as a whole. It now stands 5–6 km above the old ocean floor, which is depressed 5–8 km below the ridge axis (fig. 2.6). Hence about one-half of the buildup of the ridge has been lost by subsidence.

During Leg 55 of the Deep Sea Drilling Project (DSDP), holes were drilled on the tops of four seamounts of the Emperor chain (Jackson, Koisumi, and others, 1980). All of the core material recovered from these holes indicates that the lava flows were erupted subaerially and that the sediments are of shallow-water origin. Evidence for the subaerial nature of the lava flows is the presence of coarsely vesicular flow tops, red oxidized flow tops and bases, interbedded red tropical soils, and absence of pillow lavas; the sediments above the basalts contain shallow reef and littoral organisms and structures.

Because the water depth was great where these DSDP holes were started, they provide information on subsidence that is deeper than holes drilled on land (fig. 2.2). Hole 430 on Ojin Seamount (lat 37°59′ N.) reached 1,566 m below sea level and bottomed in material of subaerial origin. Hole 432 on Nintoku Seamount (lat 41°20′ N.) reached 1,327 m below sea level in lava flows with red oxidized tops interbedded with red tropical soil. Hole 433 on Suiko Seamount (lat 44°30′ N.) reached 2,422 m below sea level and penetrated a considerable thickness of reef limestone overlying subaerially erupted basalt. The Emperor DSDP holes document minimum subsidence of nearly 2.5 km after the volcanoes reached sea level and produced subaerially erupted lava.

Dredging of the submarine flanks of the Hawaiian volcanoes can also provide information on the amount of subsidence. Four dredge hauls from the submarine east rift zone of Mauna Kea Volcano recovered lavas that are much more vesicular than similar

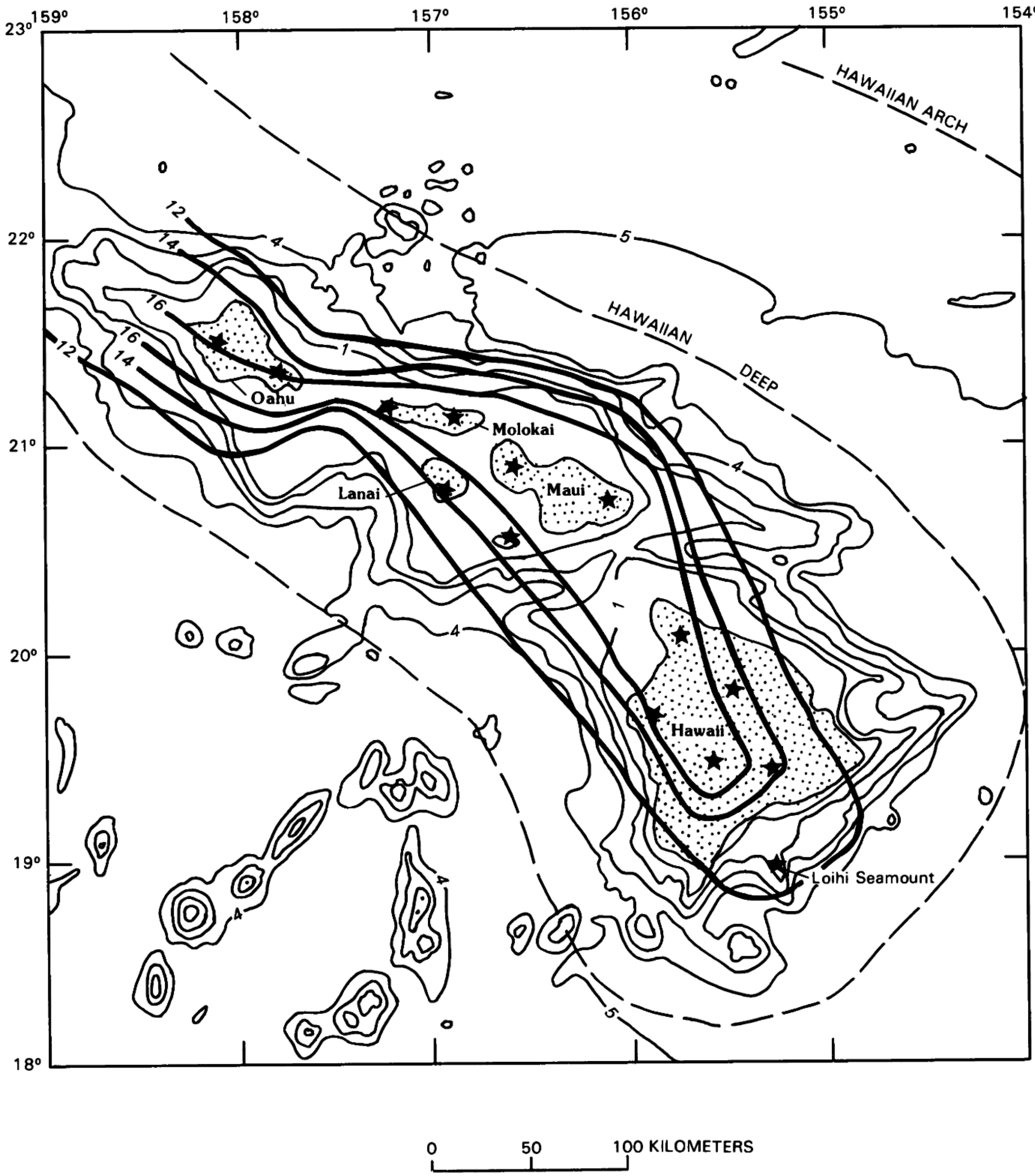

FIGURE 2.4.—Structure contours (heavy lines), in kilometers below sea level on the base of the crust, as determined by seismic refraction for the southeastern part of the Hawaiian Ridge. Data from Shor (1960), Furumoto and others (1973), and Zucca and others (1982). Depth contours in kilometers. Stars, volcanic centers.

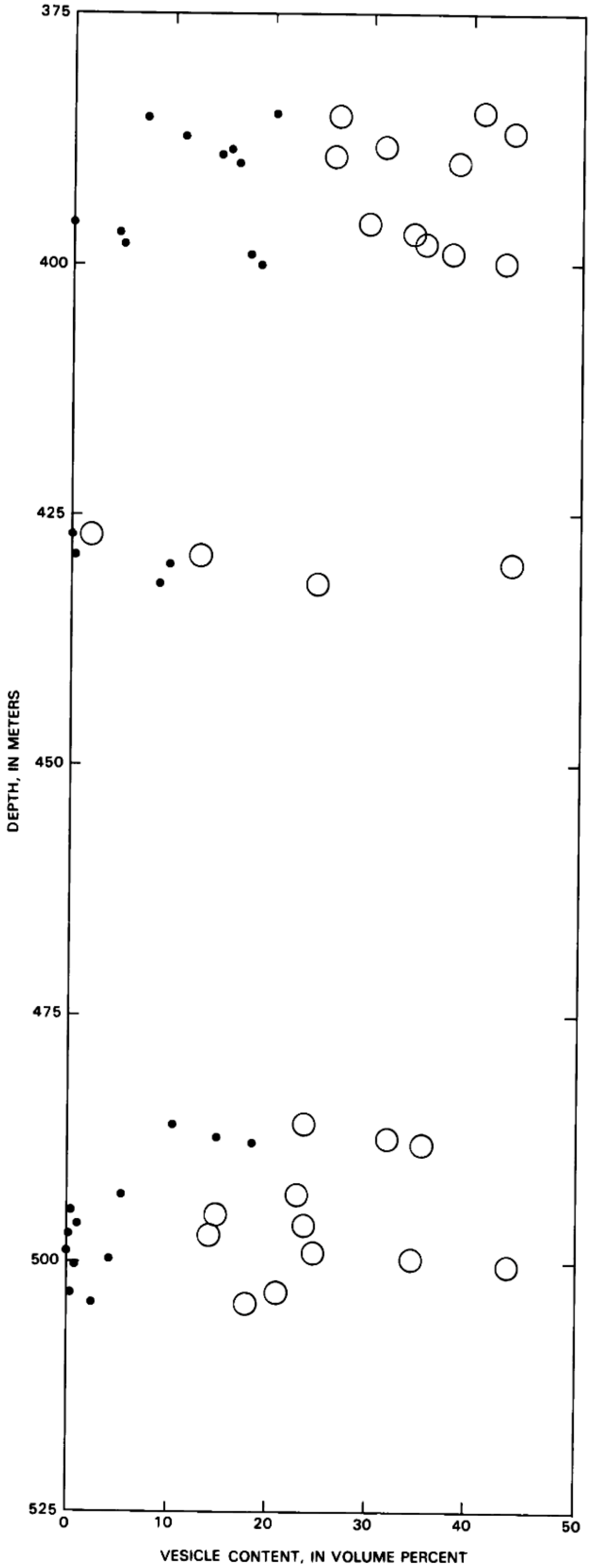

FIGURE 2.5.—Vesicularity of lava samples from 1965 drillhole on Midway Island. Open circles, volume percent of all vesicles; dots, volume percent of unfilled, open vesicles.

lavas collected from the same depth on the currently active rift zones of Kilauea and Mauna Loa Volcanoes. Comparison of vesicularity indicates that Mauna Kea has subsided about 1,000 m since extrusion of these lavas (Moore and Peck, 1965).

On the northwest rift of Hualalai Volcano, sulfur-poor degassed lavas have been dredged from depths down to 1,750 m. The low sulfur content of these lavas indicates that they were erupted under subaerial conditions and therefore that the volcano has subsided more than 1.7 km since their eruption (D.A. Clague, written commun., 1985).

ZONE OF ONGOING SUBSIDENCE

The southeast end of the Hawaiian Ridge, including the Island of Hawaii, is currently undergoing the most rapid subsidence. Tide gauge measurements indicate that Hilo has subsided at a rate of 2.4 mm/yr during the last 36 years. An analysis of the depth of reefs off the northwest coast in comparison with eustatic sea-level curves indicates that subsidence there has ranged from 1.8 to more than 3 mm/yr and has increased over the last 0.3 m.y. (Moore and Fornari, 1984). A study of submerged Hawaiian artifacts carved in lava flows indicates that Honaunau on the Kona coast of Hawaii is submerging at about 3 mm/yr (Apple and Macdonald, 1966). Photographs of the Kealakekua Bay region of west Hawaii taken in 1929 show a distinct sea-level algae line which is now 18.5 cm lower than the present-day line of the same character at the same place, indicating a submergence rate of 3.4 mm/yr (Moore and Fornari, 1984). Assuming that the current rise of sea level is 1.5 mm/yr, then these submergence rates translate to subsidence rates of 1.5 mm/yr for Honaunau and 1.9 mm/yr for Kealakekua Bay.

A growing body of data indicates that in addition to the general long-term subsidence, parts of the ridge are subject to landsliding and other forms of rapid subsidence. Such processes are an effective means of moving material to a lower elevation and of reducing the slope of the Hawaiian Ridge. During the November 1975 magnitude 7.2 earthquake, a 25-km-long zone along the southeast coast of Hawaii subsided as much as 3.5 m (Tilling and others, 1976). Such processes can be an efficient way of accommodating the rapid and dramatic subsidence that must be occurring on the southeast end of the ridge (fig. 2.6).

The Kalaupapa Peninsula on the north coast of Molokai is the exposed part of a small shield volcano that initially erupted below sea level, at the base of the giant northern sea cliff of the island. The vent in the center of the present peninsula erupted lava that covers a total area of about 20 km^2, half of which is now above sea level. Radiometric dating indicates that this activity occurred at 0.34–0.57 Ma (Clague and others, 1982). Available bathymetry (Mathewson, 1970) demonstrates that the submarine part of the shield has a markedly steeper slope below a depth of about 50 m (fig. 2.7). This slope break apparently marks the level of the sea at the time of volcanism (Mark and Moore, chapter 3). Therefore sea level is presently 50 m above its level during Kalaupapa volcanism. Sea level now stands relatively high compared to much of the Pleistocene when it probably averaged about 50 m below the present level. If we assume that sea level was at its present level

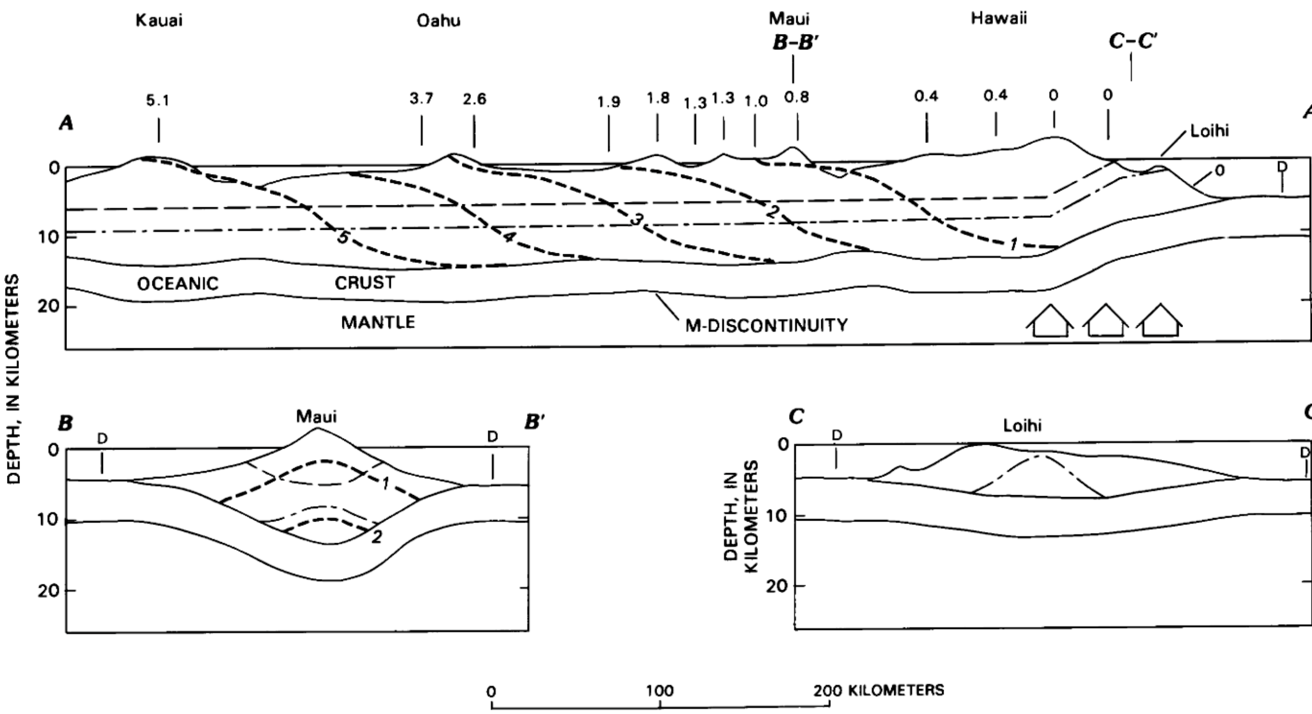

FIGURE 2.6.—Diagrammatic sections of the Hawaiian Ridge based primarily on seismic refraction and gravity measurements (Zucca and others, 1982). Location of sections shown in figure 2.1: $A-A'$, curved longitudinal crustal section along the length of the ridge; $B-B'$, transverse section through Haleakala Volcano on Maui; $C-C'$, transverse section through Loihi Volcano south of Hawaii. Radiometric ages of volcanic rock series in Ma (Clague and Dalrymple, chapter 1) are shown above section $A-A'$. Heavy dashed lines, isochrons of volcanic products in Ma (note that the zero isochron is the present southeastern slope of the ridge); dashed line, boundary between subaerially and subaqueously erupted lava; dash-dot line, boundary between tholeiitic basalt above and early alkalic basalt below. D, axis of Hawaiian Deep. Vertical exaggeration, ×4. Magma chambers and volcanic conduits are omitted for clarity, but position of mantle hot-spot area shown by arrows in $A-A'$.

during growth of the Kalaupapa shield, then subsidence occurred at the low rate of about 0.1 mm/yr. If sea level was below its present level, as it was during much of the Pleistocene, then subsidence would be even less. The Kalaupapa evidence, therefore, indicates that subsidence of central Molokai has been 0.1 mm/yr or less during the last one-half million years.

The presence of coral reefs above sea level provides constraints on the timing and amount of subsidence of the Hawaiian Ridge, and this evidence consequently is critical to an understanding of the subsidence history. Recent studies have shown that coral deposits, previously believed to have been reefs uplifted hundreds of meters, were actually laid down by a giant wave that swept up on land more than 300 m high about 115,000 years ago (Moore and Moore, 1984; Moore and others, 1984). This evidence removes the need for the volcanoes to have risen tectonically hundreds of meters, as previously proposed (Stearns, 1978). Instead, rapid subsidence during volcano growth, and continuing perhaps 1 m.y. after growth, is consistent with most evidence.

Oahu is the southeasternmost island on which extensive coral reefs, clearly in growth position, occur above sea level. These reefs include the Waimanalo reef at 7.5 m above sea level dated at 125 ka, the Laie reef at 21.5 m, and the Kaena reef at 29–30 m dated at about 650 ka (Stearns, 1978). These reefs presumably grew and developed when sea level was higher than today owing to glacioeustatic causes, and they testify to the stability of Oahu and the lack of significant present-day subsidence. However, the presence of important terraces below sea level on Oahu, especially those below 150 m (fig. 2.8), which is commonly considered to be the lowest glacioeustatic stand of the sea, indicates that Oahu, like the other islands, subsided before the growth of reefs that now occur above sea level.

Available evidence therefore indicates that contemporary subsidence at rates of a few millimeters per year can be documented at several sites on the island of Hawaii, including Hilo and the west (Kona) coast. Subsidence on Maui, if occurring, is an order of magnitude less, and Molokai and Oahu have been virtually stable for a long period.

SUBMARINE TERRACES

A large number of submarine benches, terraces, and flat-topped ridges or seamounts appear on large- and small-scale bathymetric maps of parts of the Hawaiian Ridge. Many are rather subtle features, only visible on detailed bathymetric charts. Several

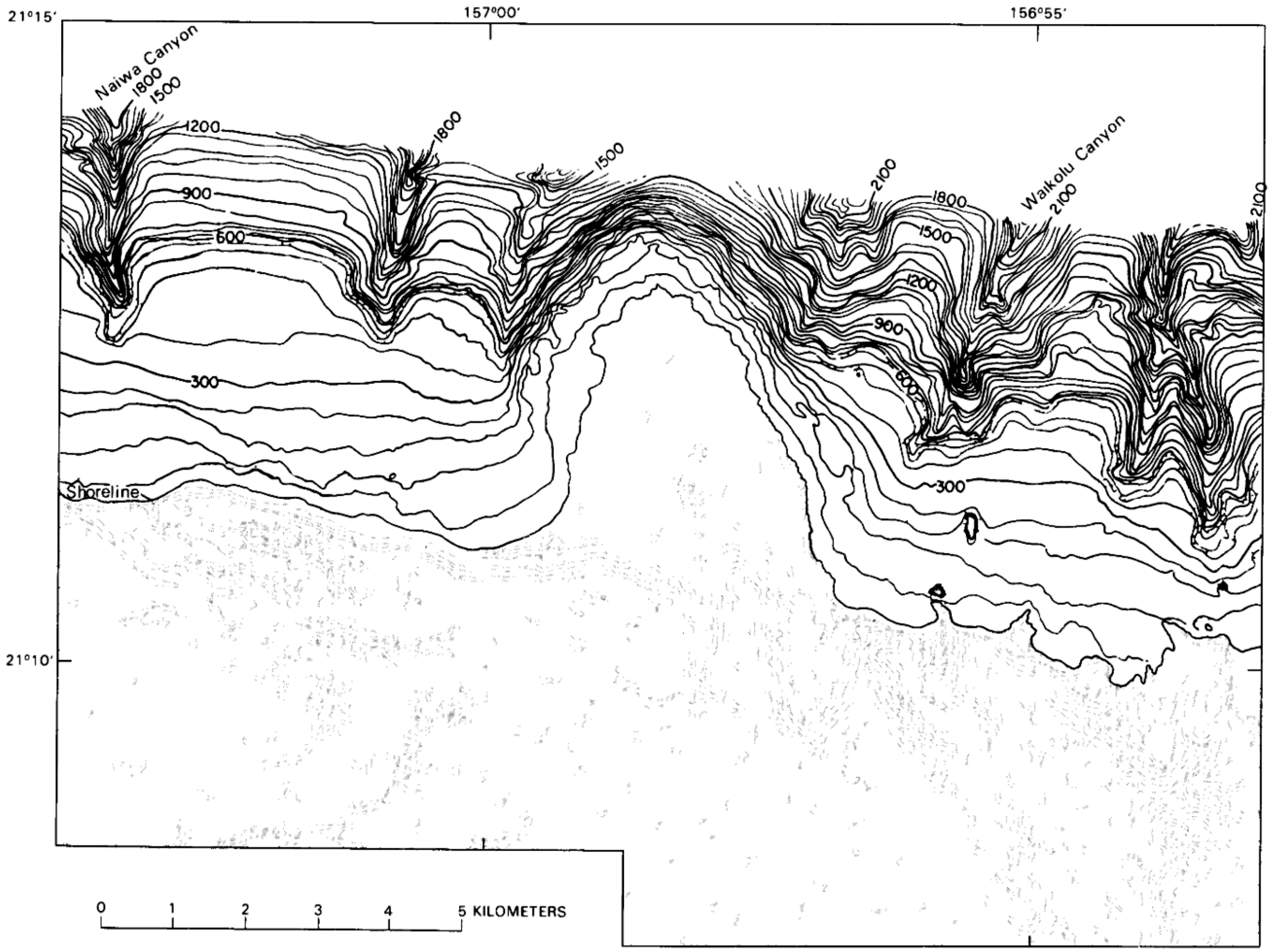

FIGURE 2.7.—Map of the Kalaupapa Peninsula on northern Molokai. Bathymetry after Mathewson (1970). Elevations and depths in feet; bathymetric contour interval, 60 ft.

of these that have been investigated by diving, dredging, and photography proved to be submerged coral reefs. A study of detailed nearshore bathymetric charts indicates that submarine terraces are common off several of the chief Hawaiian Islands at depths of 100–150 m. Curiously, deeper terraces in the depth range 200–500 m seem to be present only off the Island of Hawaii (fig. 2.8*A*). A prominent terrace at about 150-m depth occurs off west Hawaii on Kohala, Mauna Kea, and Hualalai Volcanoes, but younger lavas from Mauna Loa Volcano flood it south of lat 19°15′ N. Submersible dives reveal that it is a well-developed coral reef. Radiocarbon ages of the reef limestone show that the reef drowned 13,250 years ago (Moore and Fornari, 1984).

A second major reef terrace occurs at about 400-m depth off northwest Hawaii (Campbell, 1984) and is almost certainly correlative with the terrace at that depth off the northeast coast of Mauna Kea Volcano (figs. 2.1, 2.8). Submersible dives in 1985 off northwest Hawaii recovered reef limestone samples which yielded preliminary uranium-series ages of about 115 ka (B. Szabo, written commun., 1985).

These shallow reef terraces apparently deepen and become separated toward the southeast, forming an imbricate pattern tilted southeast (fig. 2.8). The terraces, which were presumably formed horizontally at sea level, have since been depressed more toward the southeast where subsequent subsidence has been greater. Toward the northwest, where Pleistocene subsidence occurred at only a very low rate, terraces of many different ages formed atop one another at the lower limit of the eustatically shifting sea, 100–150 m below present sea level.

Small-scale bathymetric charts of the Hawaiian Ridge show a series of large and prominent benches and terraces bounded below by a steep slope (fig. 2.9). The shapes of these abrupt slope changes indicate that they formed at the subaerial-subaqueous lava transition zone during island growth by grading of lava flows to sea level (Mark and Moore, chapter 3). A compilation of the depths of such

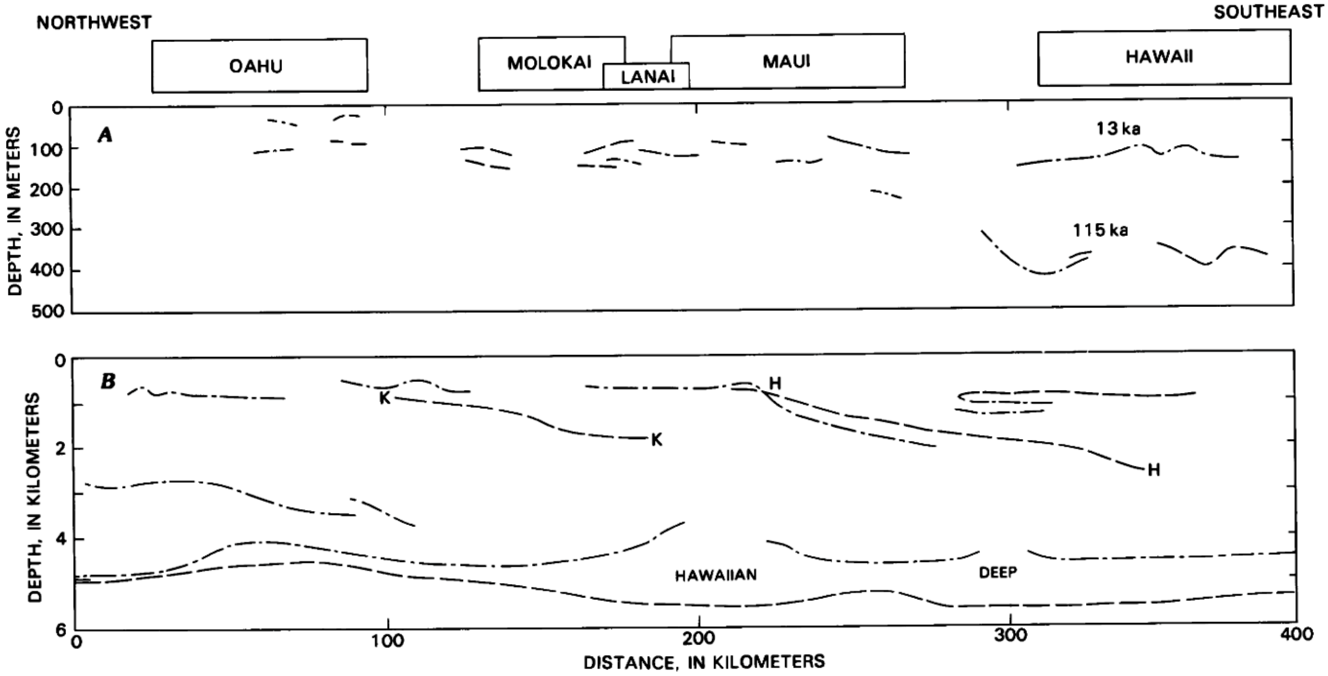

FIGURE 2.8.—Depths of terraces and axes of Hawaiian Deep projected onto a vertical plane striking N. 45° W. ***A***, Terraces in upper 500 m, including two dated terraces off west Hawaii. ***B***, All major terraces above Hawaiian Deep, including the tilted H and K terraces shown in figure 2.9. Dashed line, features on northeast side of ridge; dash-dot line, features on southwest side of ridge.

benches on the entire Hawaiian-Emperor chain indicates that they range from 1 to 4 km but are commonly 2–3 km (fig. 2.2). These depths, coupled with drill-hole evidence, suggest that the Hawaiian volcanoes have generally subsided 2–3 km after they had grown above sea level and could be marked by such sea-level features.

One of the more prominent benches surmounts the east Maui Ridge (called the Haleakala Ridge by Wilde and others, 1980). This terrace, here called the H terrace, is tilted down to the southeast and can be traced more than 150 km from north of west Maui at about 400-m depth to north of the Island of Hawaii at 2,000-m depth (figs. 2.8, 2.9). The terrace slopes about 1 km per 100 km, or about one-half degree, but it is believed to have been an original horizontal feature that has since tilted down toward the Hawaiian hot spot. A second such tilted terrace off north Molokai projects down beneath the H terrace and consequently is believed to be older (figs. 2.8, 2.9). This second feature, called the K terrace, can be traced nearly 100 km from about 900-m depth northwest of Molokai to 1,800-m depth northeast of Molokai. The K terrace is clearly younger than the proposed giant submarine landslide north of Molokai (Moore, 1964), but older than the Kalaupapa Volcano of north Molokai (0.34–0.57 Ma). The Molokai submarine canyons (see section "Submarine Canyons") could have been cut subaerially when the K terrace was about at sea level.

The ages of these tilted terraces are unknown, but they can be crudely approximated by comparing their depths with those of terraces of known ages in the same general area, assuming generally uniform rates of subsidence that are more rapid in the center than at the edge of the ridge. The H terrace, which was probably formed during the period when the present surface of Haleakala Volcano was developing, is about five times as deep as the 400-m terrace of Hawaii dated at 115 ka (fig. 2.8). Considering the lateral position of the H terrace, its age is probably 0.5–1.0 Ma. The K terrace, about twice as deep (and old) as the H terrace, probably formed when volcanism was active on Molokai at 1.8–1.9 Ma (Clague and Dalrymple, chapter 1). When the K terrace formed, the unsubsided Molokai volcanoes were more than 3 km above the sea, a height that may have led to gravitational instability and landsliding of the island's north flank.

SUBMARINE CANYONS

Submarine canyons have been identified off Kauai, Oahu, Molokai (Shepard and Dill, 1966), and Hawaii (this report), and doubtless many more will be found as improved bathymetric data become available. The Kauai submarine canyons head about 3.5 km offshore on the northwest coast and can be traced to depths of more than 2,200 m. The canyons are steep walled; Honopu Canyon has walls about 750 m high at an axial depth of 1,500 m.

Canyons off the northeast coast of Oahu head 3–4 km offshore and can be traced downward to depths exceeding 2,000 m. These canyons have been extensively filled by sediment in their upper parts, and they also passed sediment to fan deposits near the Hawaiian

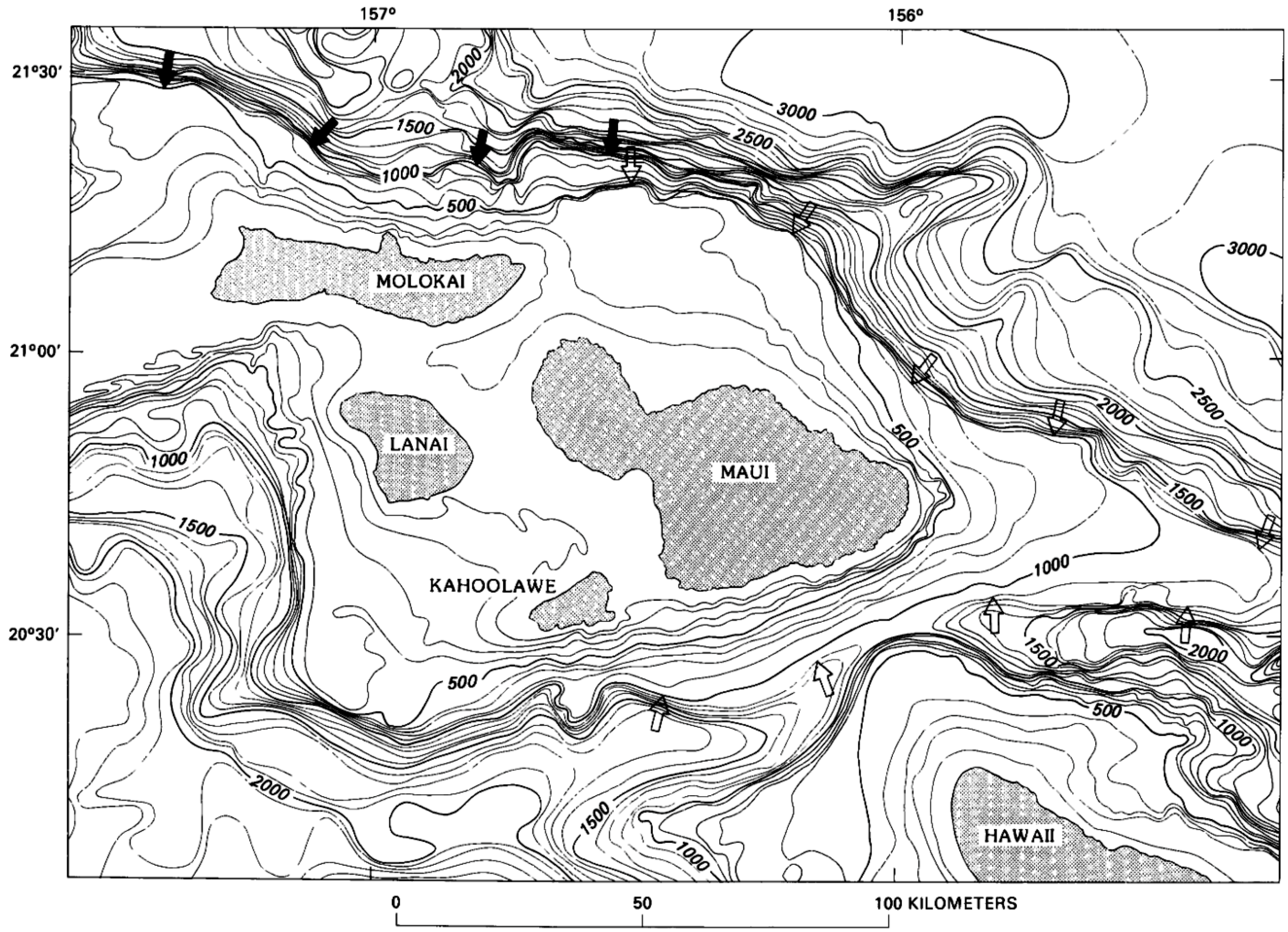

FIGURE 2.9.—Major tilted submarine terraces on southeastern Hawaiian Ridge. Open arrows, H terrace; solid arrows, K terrace. Contour interval, 100 fathoms (600 ft).

Deep (Andrews and Bainbridge, 1972). Ponds of such sediments occur between the blocky seamounts northeast of Oahu downslope from the canyons and indicate that these seamounts (variously interpreted as volcanoes or landslide blocks) predate canyon cutting.

The upper parts of 12 submarine canyons have been mapped in detail by Mathewson (1970) off the north coast of Molokai (fig. 2.7). Several of these canyons are directly offshore from major subaerial canyons and they head 1–2 km from shore. It is likely that the submarine canyons that flank the Kalaupapa Peninsula were cut before the peninsula was built at about 0.4 Ma (fig. 2.7).

A recent compilation of unpublished NOAA charts reveals three submarine canyons off windward Kohala Volcano (fig. 2.10). These canyons head less than 1 km from shore and can be traced downslope only to the limit of available bathymetry at a depth of 440 m. These submarine canyons are 40–80 m deep and hence are much less deeply incised than the 600-m-deep land canyons. This contrast can be attributed to the fact that the tops of the submarine canyons, as well as the upper part of the terrain between the canyons, has been planed off by erosional recession of the 400-m-high seacliff during subsidence. The canyons are offshore from major and minor subaerial canyons cut in Kohala Volcano. However, no submarine canyon lies offshore from Waipio Valley, the most southern Kohala canyon. This absence probably results from filling of the submarine continuation of Waipio Valley by lava from the younger, adjacent Mauna Kea Volcano (Stearns, 1966, p. 125).

Detailed bathymetric mapping off western Hawaii to locate anchor sites for an ocean thermal-energy conversion test platform has revealed a system of submarine canyons that generally head at depths greater than 1,000 m (Campbell and Erlandson, 1979). The western limit of this detailed mapping does not show the lower limit of these canyons, but several descend to depths greater than 1,600 m.

All who have investigated the Hawaiian submarine canyons note that they commonly align with present-day subaerial valleys but generally show little or no bathymetric expression across the nearshore zone down to depths of about 100 m or more. Seismic-reflection profiling off windward Oahu (Coulbourn and others, 1974) shows that submarine troughs, filled and concealed by young sediments nearshore, connect the subaerial valleys with the visible submarine canyons. This fact provides support for a subaerial origin of at least some submarine canyons.

Hamilton (1957) and Macdonald and Abbott (1970) proposed erosion below sea level by submarine processes as the major cause of the Molokai submarine canyons, because the canyons extend deeper than any previously postulated estimate of submergence. On the other hand, other workers support subaerial erosion for the carving of all the canyons. Stearns (1966) believes submarine erosion simply could not cut such canyons in hard basalt; Mathewson (1970) points to the continuity of subaerial-submarine stream profiles to make the case for subaerial canyon cutting and later subsidence. Comparison of the depths of the lower ends of canyons with depths of major terraces believed to be sea-level features, as well as the evidence from drillholes (fig. 2.2), leaves little doubt that subsidence has been adequate to permit all of the canyons to have been cut by subaerial processes, although post-subsidence modification of the canyons by submarine processes is likely.

CRUSTAL STRUCTURE

Measurements of gravity and determinations of crustal structure by seismic refraction and reflection experiments have provided important constraints on the structure of the Hawaiian Ridge. This work shows that the depth to the base of the crust as marked by the M-discontinuity increases from about 10 km far from the ridge to as much as 20 km beneath the ridge. A 100-km-long seismic-refraction profile shot in 1978 offshore from the west coast of Hawaii indicates that the base of the crust has been depressed about 8 km beneath Mauna Loa Volcano (Zucca and others, 1982; Hill and Zucca, chapter 37). The base of the crust, normally 10–11 km below sea level away from the ridge, has been depressed to a depth of 13–14 km on both the west and south coasts of the island and down to 18.5 km beneath Mauna Loa (fig. 2.6). Seismic-refraction data farther northwest indicate that the M-discontinuity south of Oahu is about 20 km below sea level (Furumoto and Wollard, 1965; Watts and others, 1985).

The inferred lowering of the base of the crust beneath the ridge may be caused in part by injection of relatively low-density basaltic dikes and sills into the mantle beneath the original crustal layer. Such basaltic material would reduce the density and seismic velocity of the the upper mantle, convert it to a crustlike material, and cause a downward migration of the M-discontinuity. An interpretation based on seismic, gravity, and geoid data suggests that the lower 4 km of the depressed oceanic crust south of Oahu is composed of such a dense intrusive complex (Watts and others, 1985). This work also indicates that the old ocean floor on top of the oceanic crust has been warped down 5–7 km beneath the ridge to depths as great as 12 km below sea level.

Seismic refraction and gravity measurements have also shown that a significant mass fraction of the Mauna Loa and Kilauea volcanic edifices above the oceanic crust is composed of intrusive cores beneath the summit region and rift zones (Hill and Zucca, chapter 37). Intrusion of such material will add weight to the crust and cause subsidence of the volcano base, but will not cause the same degree of subsidence of the top of the volcanoes as it would if erupted subaerially as lava flows.

Depression of the top of the oceanic crust is 5–7 km near Oahu, and depression of the base of the oceanic crust is about 8 km beneath Mauna Loa, as revealed by seismic experiments. Despite the possibility that some of the apparent depression of the base of the crust is caused by intrusive activity, subsidence of the base of the volcanoes is probably 5–7 km and may be greater.

The major depression of the crust caused by the weight of the growing volcanoes atop the Hawaiian hot spot forms a giant trough or canoe-like crustal warp (fig. 2.4). This southeast-trending trough underlies the Hawaiian Ridge and terminates about 50 km southeast of the Island of Hawaii near the area where the Hawaiian Deep curves around the end of the ridge (fig. 2.1). The propagation of the Hawaiian Ridge and its underlying trough onto new virgin crust toward the southeast can be compared to the movement of a massive caterpillar tread over muddy ground (fig. 2.6). The sharp downflexing of the crust and lithosphere at the head of the caterpillar is probably the cause of some of the earthquakes concentrated around the volcanic front (fig. 2.11).

The edges of the deepest part of the downbowed crust are roughly coincident with the two or more loci, commonly about 40 km apart, that connect individual volcanic centers (fig. 2.4). Sharp bending of the crust and mantle on the two sides of the base of this trough may cause fractures that tap magmas from the hot spot and are related to the formation and progagation of these volcano loci. The changes in trend of these loci along the ridge (Shaw and others, 1980) are perhaps caused by the presence of major pre-Hawaiian volcanic seamounts that were rafted over the hot-spot area. Such seamounts and their underlying subsidence troughs could affect the downflexing of the lithosphere caused by the new volcanic load and perturb the positions of the fractures that conduct magma from the hot spot to the surface to feed the new volcanoes.

The prevolcanic sea floor underlying the Hawaiian Ridge is, like the underlying base of the crust, warped down into a trough. This trough, which is filled with material erupted from the volcanoes, extends beneath the level of the old ocean floor about as far down as the largest volcanoes presently rise above that level. The cross-sectional area of this trough beneath the level of the old ocean floor is about 750–1,000 km^2, equal to or somewhat greater than the cross-sectional area of the volcanic ridge above the old ocean floor. If the rate of plate motion is assumed to be 10 cm/yr, 0.075–0.1 km^3 of volcanic material subsides beneath the level of the old ocean floor per year. This volume is similar to or slightly less than the current measured output of Kilauea Volcano of 0.11 km^3/yr (Swanson, 1972). Hence the bulk of the material erupted from the Hawaiian hot spot is not only deeply buried, but has subsided below the general level of the surface on which it was originally deposited (the old ocean floor). Models considering the origin and structure of the ridge, and of the magmas that formed it, need take this fact into account.

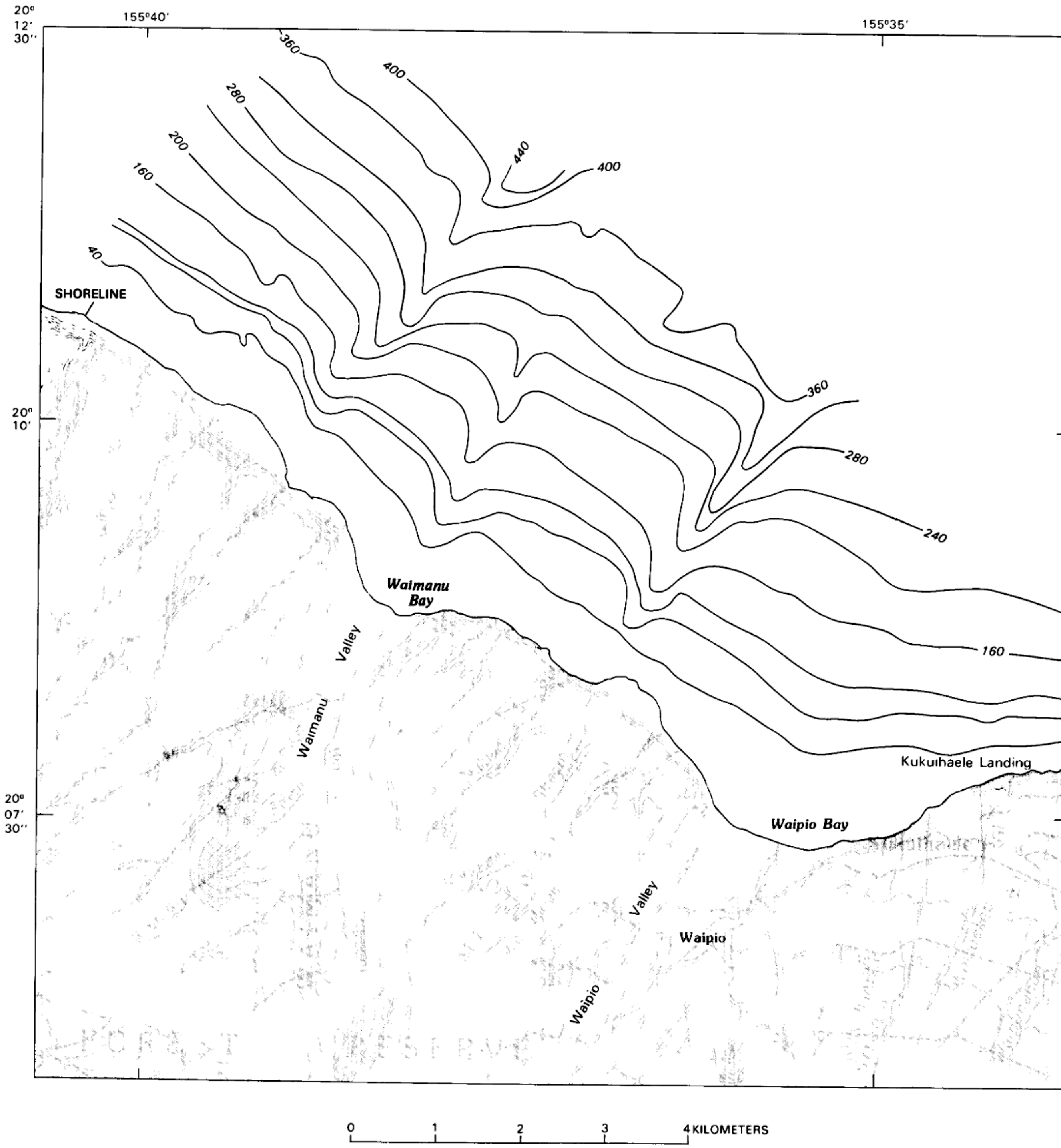

FIGURE 2.10.—Map of northeastern Kohala showing subaerial and submarine canyons. Bathymetric contour interval, 40 m.

The axis of the trough beneath the ridge has undergone most of its subsidence within a short distance (and hence time) from the unsubsided oceanic crust to the southeast. The base of the crust (and presumably also the old ocean floor) has subsided 7.5 km in the 100 km from the beginning of downwarping to the region beneath Mauna Loa (fig. 2.6). This is equivalent to a subsidence of 7.5 mm/yr if we assume that the volcanic system propagates laterally at a rate of 10 cm/yr relative to the crust.

The axis of the Hawaiian Deep becomes deeper northwest of the hot spot because of continued subsidence of the ridge and crust in

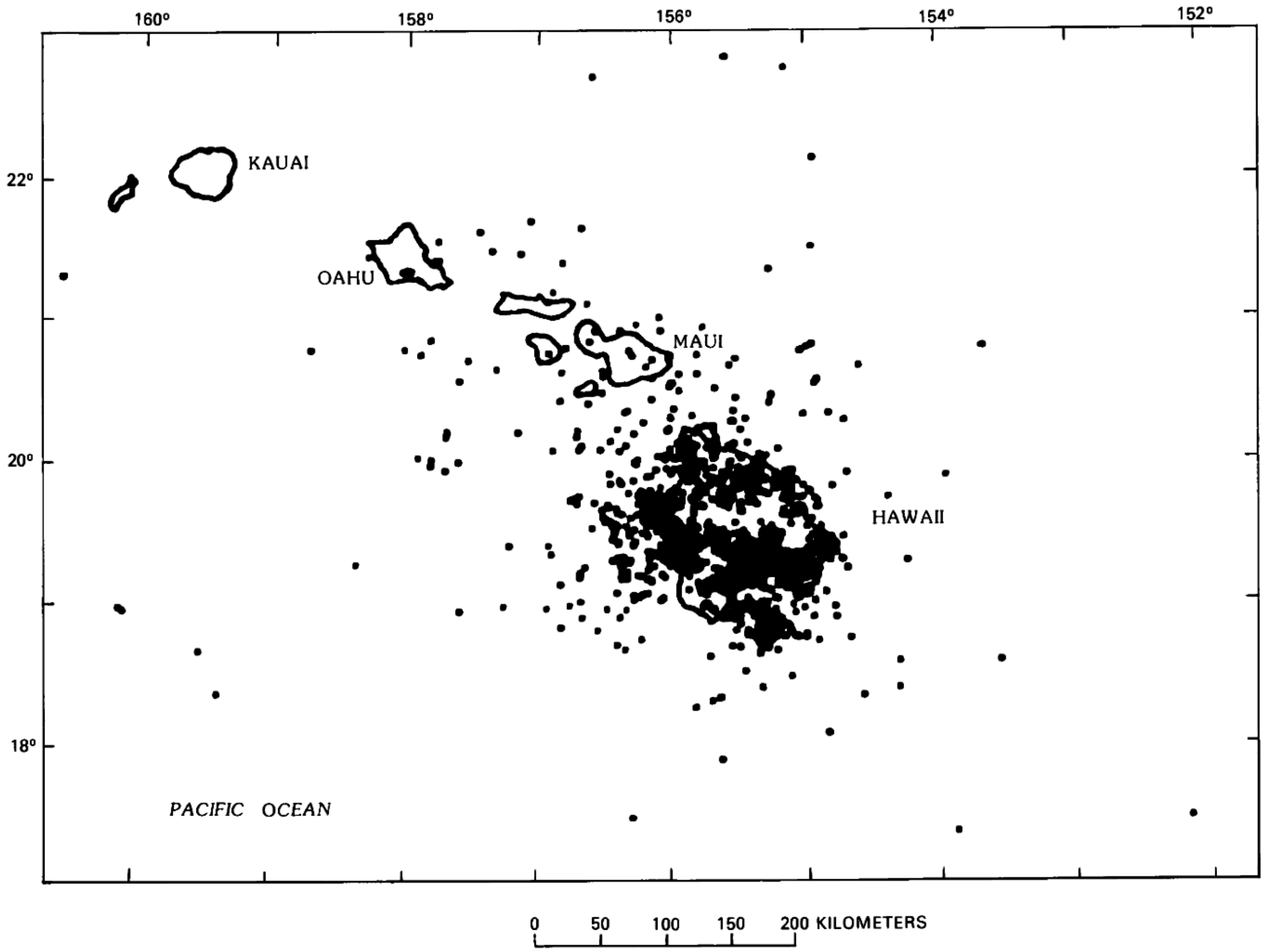

FIGURE 2.11.—Epicentral locations of earthquakes with magnitude 3 or larger in the Hawaiian region during 1968–83 (F.W. Klein, written commun., 1985).

response to the weight of erupted products. In addition, the lithosphere may have been reheated, and hence elevated adjacent to the hot spot to partly create the Hawaiian Arch (Von Herzen and others, 1982). Cooling and subsidence of the lithosphere northwest of the hot spot adds to the depression of the trough axis. Seismic refraction experiments indicating that the depth to the base of the crust south of Oahu is 20–23 km (Furumoto and Woollard, 1965) suggest the possibility of minor continued subsidence from Mauna Loa to Oahu.

CHARACTER OF SUBSIDENCE

The extent of the depression of the M-discontinuity from unaffected crust southeast of the ridge, through Kilauea, and northwestward to the older volcanoes provides information on the history and character of subsidence (fig. 2.6). Assuming a volcanic propagation rate of 10 cm/yr, each 100 km represents 1 m.y.

The bulk of subsidence occurred in about 1 m.y. The center of the base of the volcanic ridge subsides about 8 km, and the part of the volcano that first broke the sea surface subsides about 5 km. However, the sides of a volcano undergo considerably smaller amounts of subsidence than the center, producing a trough-like shape to the boundary between subaerially and submarine erupted basalt (fig. 2.6, section *B-B′*).

The southeast end of the ridge is downshearing against that part to the northwest because ongoing subsidence is greater to the southeast (fig. 2.6). This southeasterly dislocation probably occurs at all scales in southeast Hawaii and no doubt produces some of the seismic activity (fig. 2.11). Moreover, it must be an important process in shaping the geometry of volcanic conduits, magma chambers, and rift zones. Major fault systems such as the Kaoiki, which separates Mauna Loa from Kilauea, and the Hilina, on the south flank of Kilauea, are probably related to this process.

The general pattern of subsidence permits an estimate of the volume proportions of certain divisions of the volcanoes. A section through the embryonic part of the ridge at Loihi Volcano (fig. 2.6) indicates a cross-sectional area of the ridge here of about 880 km^2 of basalt, all subaqueously erupted or intrusive. Petrologic studies (Moore and others, 1982) indicate that Loihi has recently undergone a transition from eruption of early alkalic lava to eruption of tholeiitic lava, and the section suggests that about 30 percent of the ridge here is composed of the early alkalic material. This estimate does not consider the possibility that much of the basal part of the volcano is composed of intrusive rock, possibly of a different composition. At this early stage already about 40 percent of the volume of the volcano has subsided beneath the level of the prevolcanic ocean floor.

In contrast, Haleakala Volcano on Maui has virtually completed its volcanic cycle, except for lingering small-scale eruption of late alkalic lavas. The cross-sectional area of the volcanic ridge here is about 1,450 km^2, of which about 16 percent is the early alkalic lava and only a few percent is late alkalic lava. Presently only 4 percent of the volcanic pile is above sea level, but 25 percent of the volume of the ridge here was erupted above sea level and built shield volcanoes largely of tholeiitic basalt. The other 75 percent of the volcanic ridge, although also composed principally of tholeiitic basalt, was erupted beneath the sea and is composed largely of pillowed lavas that predate the growth of the subaerial shield volcanoes and have steeper slopes (Moore and Fiske, 1969). About 53 percent of the volume of volcanic materials here has subsided beneath the level of the prevolcanic ocean floor.

The primary constraint on the measured rate of volcanic propagation along the Hawaiian Ridge is the body of radiometric ages of rocks collected from the volcanoes (Clague and Dalrymple, chapter 1). It is evident that virtually all of these samples are from near the top of the volcanoes. Wherever possible, tholeiitic lavas rather than the later alkalic lavas were collected and dated, so the ages more nearly define the age of the major shield-building stage of volcano growth.

Time-distance plots of the age data show an apparent acceleration of the volcanic propagation rate (as well as the volcanic eruption rate) in the last approximately one million years. Shaw and others (1980) find that the average rate of volcanic progression from the Hawaiian-Emperor bend to West Maui Volcano is 7.9 cm/yr, and the average rate from West Maui Volcano to Kilauea is 17.8 cm/yr. Clague and Dalrymple (chapter 1) support the notion of a relatively recent acceleration in the volcanic propagation rate and note that otherwise the curvature of the time-distance plot causes it to intersect the distance axis northwest of Kilauea and predict a negative age for that volcano.

However, the fact that the southeast end of the ridge has not undergone all of its subsidence is an important limitation to the use of unqualified time-distance plots to determine the propagation rate. The exposed and dated parts of the southeastern volcanoes will clearly submerge beneath the sea and be covered with younger lavas. When the first phase of rapid subsidence is complete for Kilauea, the present summit will be 1 km or more below sea level. Loihi Volcano (figs. 2.1 and 2.6), now 1 km below sea level, will have to grow more than 2 km up before it breaks the sea surface because more than one-half of its volcanic growth is offset by subsidence. In order for the ages of Kilauea and Mauna Loa to be comparable with those of the northwestern volcanoes, one must obtain samples of yet unerupted lavas that will mantle the presently exposed surface. Hence, if such unerupted future lava could be dated now it would have a negative age as suggested by Clague and Dalrymple. Until more is known about the volume and age of the submarine volcanic products, it seems premature to propose a dramatic increase in the rate of volcanic propagation in the last million years or so. Assuming that the rate has remained constant, then straight-line projection of age-distance data to Kilauea and on to Loihi suggests a negative (future) age for Loihi of 1–1.5 m.y. (Clague and Dalrymple, chapter 1). Since Loihi is already of finite age, the duration of the chief period of volcano growth is about 1.5–2 m.y. (fig. 2.6)

The proposed increase in the volume of volcanic products and hence eruption rates at the young end of the ridge (Clague and Dalrymple, chapter 1; Shaw and others, 1980) is also uncertain, because all of the previously measured volcanic volumes include only that part of the ridge above the level of the flat ocean floor surrounding the Hawaiian Ridge. The volume of volcanic material in the trough underlying the ridge has not been considered. Because the old ocean floor is fully depressed beneath the entire length of the ridge except at the extreme southeast end, it is not surprising that larger apparent volumes (and higher eruption rates) have been calculated for the unsubsided region.

The base of the volcanic material erupted between 1 Ma and the present is not depressed as much as that erupted between 2 and 1 or 3 and 2 Ma (fig. 2.6). The volume between the 1 Ma and zero isochrons is even further reduced if we consider the fact that the sides of the depressed zone loop around the end of the ridge (compare with fig. 2.4). This part of the ridge will subside many kilometers within the next one million years.

As the Hawaiian Ridge propagates southeasterly in the lithosphere by the northwesterly movement of the Pacific plate, the southeast end of the ridge and the surrounding ocean floor bend down toward the hot-spot area or toward the youngest volcano. Movement of the lithospheric plate away from the hot spot progressively removes a given island or volcano from the subsiding zone adjacent to the hot spot. At the present time Kahului, Maui, which is 180 km from Mauna Loa (near the hot-spot center), is not subsiding rapidly (fig. 2.3). and Molokai and Oahu appear to be stable. Hence the rapidly subsiding zone is largely in the vicinity of the Island of Hawaii, and it may be outlined by the concentration of recent earthquakes (fig. 2.11). Downwarping of the base of the crust begins about 100 km both southeast and southwest of Mauna Loa (Zucca and others, 1982). and available evidence suggests that the downwarping zone is roughly circular with a 100-km radius. The center of downwarping as defined by the curvature of the Hawaiian Deep (fig. 2.4) is about midway between the summits of Mauna Loa and Mauna Kea Volcanoes.

The manner of tilting of pre-existing terraces is clarified if we trace the history of a horizontal surface affected by this downwarping sag (fig. 2.12). Close to the center of the sag, assumed to be cone-like, the terrace will subside faster than on its edges where no subsidence occurs. As the terrace moves relatively northwest from the sag, progressively all subsidence will stop and the tilt will be "frozen in" on the terrace. If we assume that the sag moves 10 cm/yr southeast relative to the crust, and that subsidence is 2.5 mm/yr at the cone center, then an original horizontal surface will finally end tilted toward the sag at about 1.5 km/100 km (fig. 2.12). This tilt is similar to that of the H terrace off east Maui (fig. 2.8).

The depression of the M-discontinuity beneath the Hawaiian Ridge is a measure of subsidence of the volcanoes. The high proportion of the volcanic pile that is below the level of the prevolcanic ocean floor even at the youngest, southeasternmost volcanoes indicates that subsidence is very rapid. Much of the subsidence occurs while the volcanoes are growing. A large part of the total subsidence has occurred by the time the volcanoes reach sea level; Loihi Volcano, still 1 km below sea level, has already subsided more than 1 km (fig. 2.6) and will presumably subside an additional 6 km if it grows to the size of Mauna Loa. However, subsidence is much more rapid after the volcanic products are deposited above sea level because the buoyancy of water does not affect this material and its effective weight is 1 g/cm^3 greater. This fact is demonstrated by the slope of the depressed M-discontinuity at the volcanic front. This slope becomes abruptly steeper beneath the southeast shoreline of Hawaii (fig. 2.6). These slopes (assuming a volcanic propagation rate of 10 cm/yr) indicate subsidence rates of 4 mm/yr for the submarine part of the volcano, and 12 mm/yr for the subaerial part of the volcano. Comparison of these slopes at the base of the crust with volcanic profiles at the top of the crust (fig. 2.6) indicates that downbowing of the crust is an almost immediate response to volcanic loading.

Through the life of the growing volcano, new material deposited on top undergoes progressively less subsidence than the older material below it. Hence when the volcano achieves enough mass above sea level to be marked by sea-level terraces, these features will undergo only a small part of the total subsidence experienced by the older, lower parts of the volcano.

CONCLUSIONS

Subsidence of the Hawaiian Ridge can be constrained by tide-gauge measurements, geophysical measurements of crustal structure, and information gained from submarine canyons, submerged carbonate reefs, deep drilling, and dredging. An analysis of these data indicates the following:

1. The Hawaiian Ridge has undergone rapid and dramatic subsidence. Most volcanoes have subsided 2–4 km since reaching the sea surface, and the bases of the volcanoes have subsided 5–8 km.

2. Between one-half and two-thirds of the upbuilding of the volcanoes is offset by subsidence.

3. Major sea-level terraces formed during the chief periods of volcanic growth above sea level tend to tilt toward the point of active volcanic loading, that is toward the southeast. Hence depth correlation of terraces along the island chain may be misleading.

4. Because the young volcanoes at the southeast end of the Hawaiian Ridge have not completed their subsidence and growth, volcanic ages derived from them provide estimates of volcanic propagation rates that are too high. Likewise, volumes measured for the volcanoes only above the level of the ocean floor indicate eruption rates for the older, subsided volcanoes that are too low. These factors raise doubts about both the proposed increase in volcanic propagation rate and eruption rate during the last few million years.

5. The alignment of volcanoes along two or more curving loci about 40 km apart and parallel to the Hawaiian Ridge may result from sharp bending of the crust and mantle on the two sides of the base of the subsided trough underlying the ridge. Such bending may

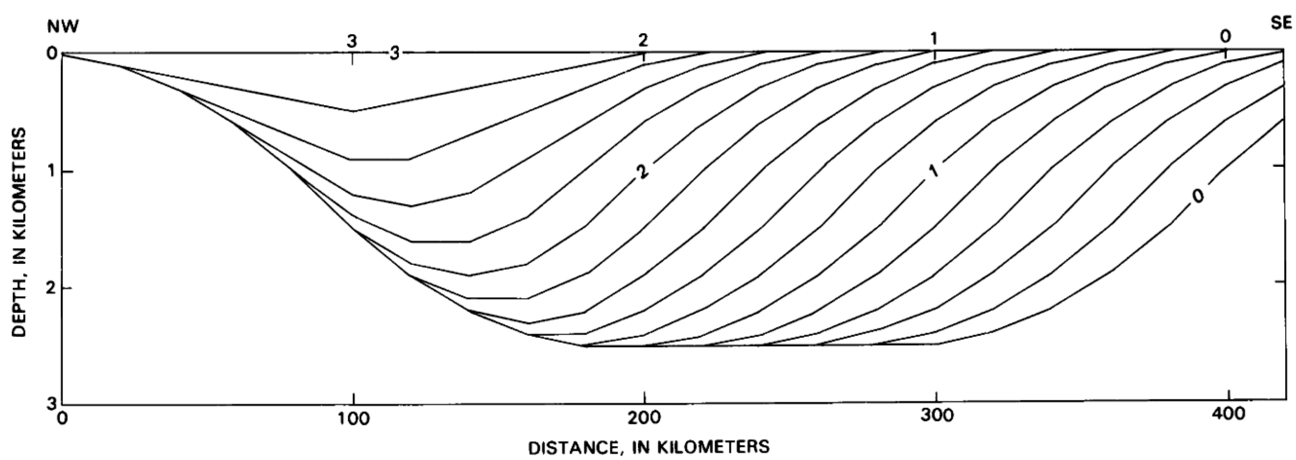

FIGURE 2.12.—Vertical section passing through center of migrating cone-like subsidence zone (100-km radius) that migrates southeast at 10 cm/yr; position of the center of the cone at specific times (Ma) is shown at top. The center of the cone subsides 2.5 mm/yr and subsidence and migration have occurred for 3 m.y. Profiles show successive shapes of a warped surface that was horizontal at 3 Ma; numbers give ages in Ma. Note the formation of a tilted terrace sloping about 1.5 km/100 km to the southeast that is left behind in the wake of the migrating subsidence zone. This model can apply to the deformation either of a sea-level terrace or of the ocean floor at the beginning of the propagation of the subsiding Hawaiian Ridge.

initiate fractures that tap magmas from the hot spot and feed the volcanoes. The changes in mapped trend of the loci may be caused by the rafting of pre-Hawaiian seamounts over the hot spot area, where they perturbed this downflexing of the lithosphere and the location of the conduit fractures.

6. An estimate of volumes based on patterns of subsidence indicate that as much as 16 percent of the volume of the ridge is early alkalic lava, about 25 percent was erupted subaerially, and about 53 percent has subsided beneath the level of the old ocean floor. These estimates may be reduced if a substantial volume of the volcanoes is composed of intrusive rather than extrusive rocks.

7. A part of the seismic activity that is centered at the southeastern end of the Ridge may result from bending and dislocation of the volcanic pile as the southeastern part subsides relative to that adjacent on the northwest.

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