THE HAWAIIAN-EMPEROR VOLCANIC CHAIN
Part I
Geologic Evolution

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ABSTRACT

The Hawaiian-Emperor volcanic chain stretches nearly 6,000 km across the North Pacific Ocean and consists of at least 107 individual volcanoes with a total volume of about 1 million km³. The chain is age progressive with still-active volcanoes at the southeast end and 80–75-Ma volcanoes at the northwest end. The bend between the Hawaiian and Emperor Chains reflects a major change in Pacific plate motion at 43.1 ± 1.4 Ma and probably was caused by collision of the Indian subcontinent into Eurasia and the resulting reorganization of oceanic spreading centers and initiation of subduction zones in the western Pacific. The volcanoes of the chain were erupted onto the floor of the Pacific Ocean without regard for the age or preexisting structure of the ocean crust.

Hawaiian volcanoes erupt lava of distinct chemical compositions during four major stages in their evolution and growth. The earliest stage is a submarine alkalic preshield stage, which is followed by the tholeiitic shield stage. The shield stage probably accounts for >95 percent of the volume of each volcano. The shield stage is followed by an alkalic postshield stage during which a thin cap of alkalic basalt and associated differentiated lava covers the tholeiitic shield. After several million years of erosion, alkalic rejuvenated-stage lava erupts from isolated vents. An individual volcano may become extinct before the sequence is complete. The alkalic preshield stage is only known from recent study of Loihi Seamount. Lava from later eruptive stages has been identified from numerous submerged volcanoes located west of the principal Hawaiian Islands.

Volcanic propagation rates along the chain are 9.2 ± 0.3 cm/yr for the Hawaiian Chain and 7.2 ± 1.1 cm/yr for the Emperor Chain. A best fit through all the age data for both chains gives 8.6 ± 0.2 cm/yr. Alkalic rejuvenated-stage lava erupts on an older shield during the formation of a new large shield volcano 190 ± 30 km to the east. The duration of the quiescent period preceding eruption of rejuvenated-stage lava decreases systematically from 2.5 m.y. on Niihau to <0.4 m.y. at Haleakala, reflecting an increase in the rate of volcanic propagation during the last few million years. Rejuvenated-stage lava is generated during the rapid change from subsidence to uplift as the volcanoes override a flexural arch created by loading the new shield volcano on the ocean lithosphere.

Paleomagnetic data indicate that the Hawaiian hot spot has remained fixed during the last 40 m.y., but prior to that time the hot spot was apparently located at a more northerly latitude. The most reliable data suggest about 7° of southward movement of the hot spot between 65 and 40 Ma.

The numerous hypotheses to explain the mechanism of the hot spot fall into four types: propagating fracture hypotheses, thermal or chemical convection hypotheses, shear melting hypotheses, and heat injection hypotheses. A successful hypothesis must explain the propagation of volcanism along the chain, the near-fixity of the hot spot, the chemistry and timing of the eruptions from individual volcanoes, and the detailed geometry of volcanism. None of the geophysical hypotheses proposed to date are fully satisfactory. However, the existence of the Hawaiian swell suggests that hot spots are indeed hot. In addition, both geophysical and geochemical hypotheses suggest that primitive undegassed mantle material ascends beneath Hawaii. Petrologic models suggest that this primitive material reacts with the ocean lithosphere to produce the compositional range of Hawaiian lava.

INTRODUCTION

The Hawaiian Islands; the seamounts, banks, and islands of the Hawaiian Ridge; and the chain of Emperor Seamounts form an array of shield volcanoes that stretches nearly 6,000 km across the north Pacific Ocean (fig. 1.1). This unique geologic feature consists of more than 107 individual volcanoes with a combined volume slightly greater than 1 million km³ (Bargar and Jackson, 1974). The chain is age progressive with still-active volcanoes at the southeast end whereas those at the northwest end have ages of about 75–80 Ma. The volcanic ridge is surrounded by a symmetrical depression, the Hawaiian Deep, as much as 0.7 km deeper than the adjacent ocean floor (Hamilton, 1957). The Hawaiian Deep is in turn surrounded by the broad Hawaiian Arch.

At the southeast end of the chain lie the eight principal Hawaiian Islands. Place names for the islands and seamounts in the chain are shown in figure 1.1 (see also table 1.2). The Island of Hawaii includes the active volcanoes of Mauna Loa, which erupted in 1984, and Kilauea, which erupted in 1986. Loihi Seamount, located about 30 km off the southeast coast of Hawaii, is also active and considered to be an embryonic Hawaiian volcano (Malahoff, chapter 6; Moore and others, 1979, 1982). Hualalai Volcano on Hawaii and Haleakala Volcano on Maui have erupted in historical times. Between Niihau and Kure Island only a few of the volcanoes rise above the sea as small volcanic islets and coral atolls. Beyond Kure the volcanoes are entirely submerged beneath the sea. Approximately 3,450 km northwest of Kilauea, the Hawaiian Chain bends sharply to the north and becomes the Emperor Seamounts, which continue northward another 2,300 km.

It is now clear that this remarkable feature was formed during the past 70 m.y. or so as the Pacific lithospheric plate moved north and then west relative to a melting anomaly, called the Hawaiian hot spot, located in the asthenosphere. According to this hot-spot hypothesis, a trail of volcanoes was formed and left on the ocean

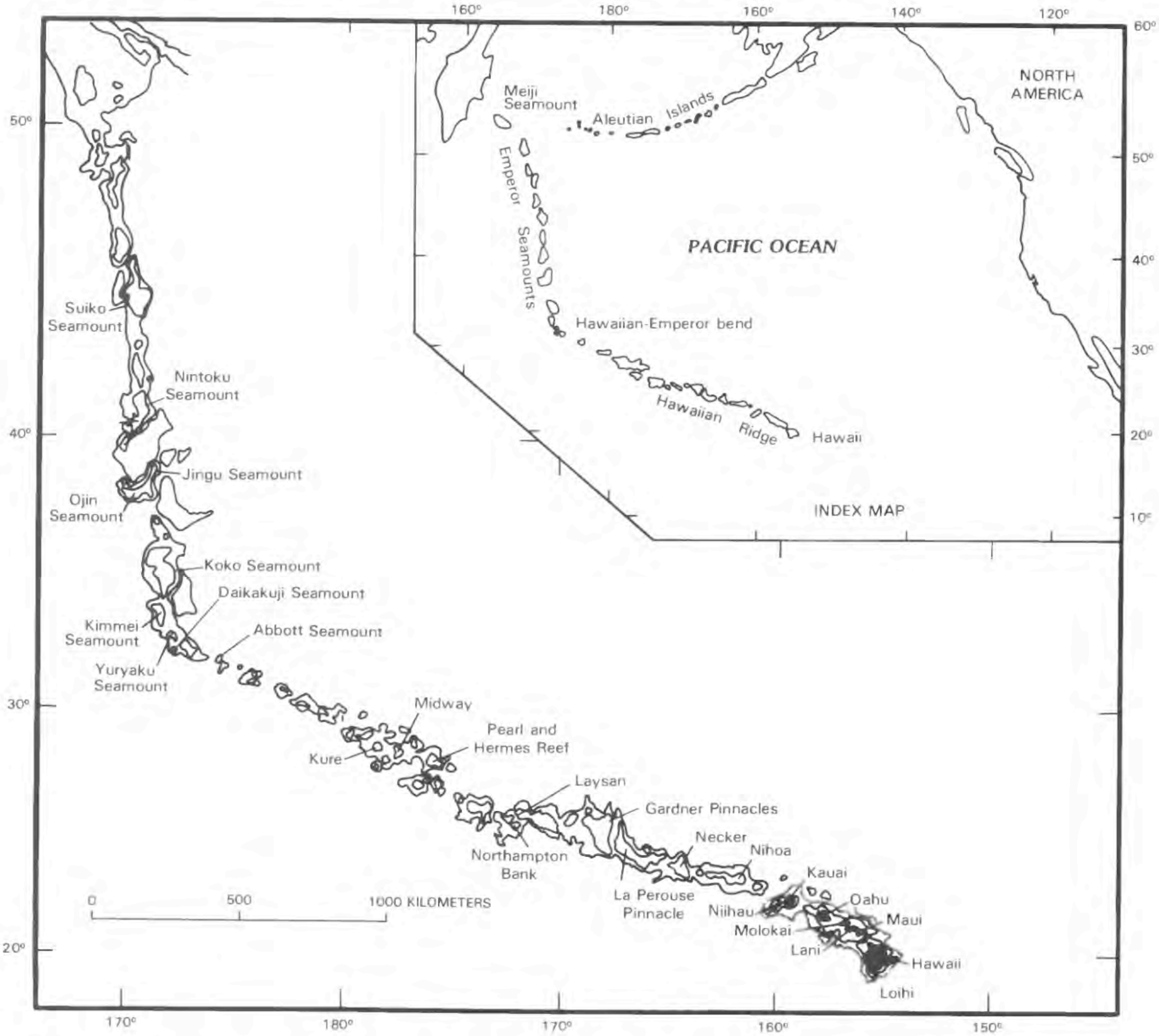

FIGURE 1.1.—Bathymetry of Hawaiian-Emperor volcanic chain modified from Chase and others (1970). Contours at 1-km and 2-km depths shown in area of the chain only. Inset shows location of chain (outlined by 2-km depth contour) in central North Pacific.

floor as each volcano was progressively cut off from its source of lava and a new volcano was born behind it.

Wilson (1963a, c) was the first to propose that the Hawaiian Islands and other parallel volcanic chains in the Pacific were formed by movement of the sea floor over sources of lava in the asthenosphere. Although the Emperor Chain was recognized as a northward continuation of the Hawaiian Chain by Bezrukov and Udintsev (1955) shortly after the Emperor Seamounts were first described by Tayama (1952) and Dietz (1954), Wilson confined his hypothesis to the volcanoes of the Hawaiian Islands and the Hawaiian Ridge. Christofferson (1968), who also coined the term "hot spot," extended Wilson's idea to include the Emperor Seamounts and suggested that the Hawaiian-Emperor bend represents a major change in the direction of sea-floor spreading from northward to westward. Morgan (1972a, b) proposed that the Hawaiian and other hot spots are thermal plumes of material rising from the deep mantle and that the worldwide system of hot spots constitutes a reference frame that is fixed relative to Earth's spin axis.

Although experimental testing of the various hypotheses proposed to explain hot spots has so far proven unproductive, the hot-spot hypothesis has several important corollaries that can and have been tested to varying degrees. Foremost among these is that the volcanoes should become progressively older to the west and north as a function of distance from the hot spot. This progressive aging should be measurable with radiometric methods and also should be evident in the degree of erosion, subsidence, and geological evolution of the volcanoes along the chain. A second important corollary is that the latitude of formation of the volcanoes, as recorded in the magnetization of their lava flows, should reflect the present latitude of the hot spot rather than the present latitude of the volcanoes. Third, because the active mechanism is beneath the lithosphere, the Hawaiian-Emperor Chain should show no relation to the structure of the sea floor. Finally, the volcanic rocks of the volcanoes should be similar in both chemistry and sequence of eruption for each volcano along the chain, or should change in a systematic and coherent way.

In this paper we describe the Hawaiian-Emperor volcanic chain and those individual volcanoes that have been sampled and studied. We review the evidence that indicates that all of the corollaries mentioned above are true and, therefore, that the hot-spot hypothesis is a viable explanation of the origin of the chain. We will also describe the various hypotheses that have been proposed to explain the hot-spot mechanism and discuss their strengths and weaknesses.

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STRUCTURE AND AGE OF THE UNDERLYING CRUST

The volcanoes of the Hawaiian-Emperor Chain were formed by eruption of lava onto the floor of the Pacific Ocean without regard for the age or preexisting structure of the ocean crust, or for the presence of preexisting volcanoes. The precise age of the ocean crust beneath much of the chain is poorly known because of the paucity of magnetic anomalies in the area (fig. 1.2). The Hawaiian Islands and Ridge east of about Midway Island lie on crust older than anomaly 34 but younger than anomaly M0. In a general way, both the Hawaiian seamounts and the underlying crust increase in age to the west so that the age of the crust beneath each volcano at the time it was built was between 80 and 90 m.y. (fig. 1.3). Volcanoes between Midway and the Hawaiian-Emperor bend and in the Emperor Seamounts south of Jingu Seamount are all built on crust whose age is between that of anomalies M0 and M3. Because the seamounts increase in age to the northwest but the underlying crust is roughly the same age, the age of the crust when the overlying volcano was built decreases systematically from about 80 m.y. at the bend to about 55 m.y. at Jingu Seamount. North of Jingu Seamount the age of the crust is not known, but plate reconstructions imply decreasing crustal ages to the north (Scientific Party DSDP 55, 1978; Byrne, 1979).

Northward from Jingu Seamount, we estimate that the crustal age at Suiko was roughly 40 m.y. and at the northernmost seamount, Meiji, was <20 m.y. when those volcanoes formed. If this extrapolation is extended beyond Meiji Seamount to hypothetical seamounts we presume existed once but which have been subducted or accreted in the Kuril Trench, we conclude that the Hawaiian hot spot was located, and perhaps originated, beneath the Kula-Pacific spreading axis at about 100–90 Ma.

Preexisting structures in and on the underlying crust appear to have had little or no influence on the formation of the Hawaiian-Emperor Chain (fig. 1.2). Several fracture zones, including the Mendocino, Murray, and Molokai, cross the chain, but none appears to have greatly affected the orientation of the chain, the rate of propagation of volcanism, or the volume of eruptive products. Likewise, the chain has overridden at least one Late Cretaceous seamount, again without obviously affecting the orientation, rate of propagation of volcanism, or the volumes of eruptive products (Clague and Dalrymple, 1975).

ERUPTIVE SEQUENCE

Hawaiian volcanoes erupt lava of distinct chemical compositions during four different stages in their evolution and growth (table 1.1). The three later stages are well studied and documented (Stearns, 1940a, b, c; Macdonald and Katsura, 1964; Macdonald, 1968), but the first (preshield) stage, which includes the early phase of the submarine history of the volcano, has only been examined recently (Moore and others, 1979, 1982).

The tholeiitic eruptive stage includes a long period of submarine eruption that forms a volcanic edifice with steep slopes and a subaerial eruptive phase that forms the shield-shaped volcano (Peterson and Moore, chapter 7). In this paper we refer to this entire stage of tholeiitic volcanism as the shield stage. In the shield stage, tholeiitic basalt flows construct the main volcanic edifice in the relatively short span of perhaps 1 m.y. or less (Jackson and others, 1972). Wright and others (1979) independently propose 200,000 yr as the duration of tholeiitic volcanism. Most of the mass of an individual volcano (95–98 percent) is formed from these voluminous eruptions. The shield stage usually includes caldera collapse and eruption of caldera-filling tholeiitic basalt. During the next stage, the alkalic postshield stage, alkalic basalt also may fill the caldera and form a thin cap of alkalic basalt and associated differentiated lava that covers the main shield. This alkalic lava accounts for less than 1 percent of the total volume of the volcano. After as much as a few million years of volcanic quiescence and erosion, very small amounts of SiO_2-poor lava may erupt from isolated vents; this stage is commonly called the posterosional stage; we refer to it here as the alkalic rejuvenated stage. An individual volcano may become extinct before this eruptive cycle is complete, but the general sequence is typical of the well-studied Hawaiian volcanoes (table 1.2). Some of these ideas are more than half a century old. Cross (1915) recognized that each of the Hawaiian volcanoes built a shield of lava, comparable to Kilauea flows, during a period of frequent voluminous eruptions. He also noted that this period was followed by a period of erosion and declining activity that produced cinder cones

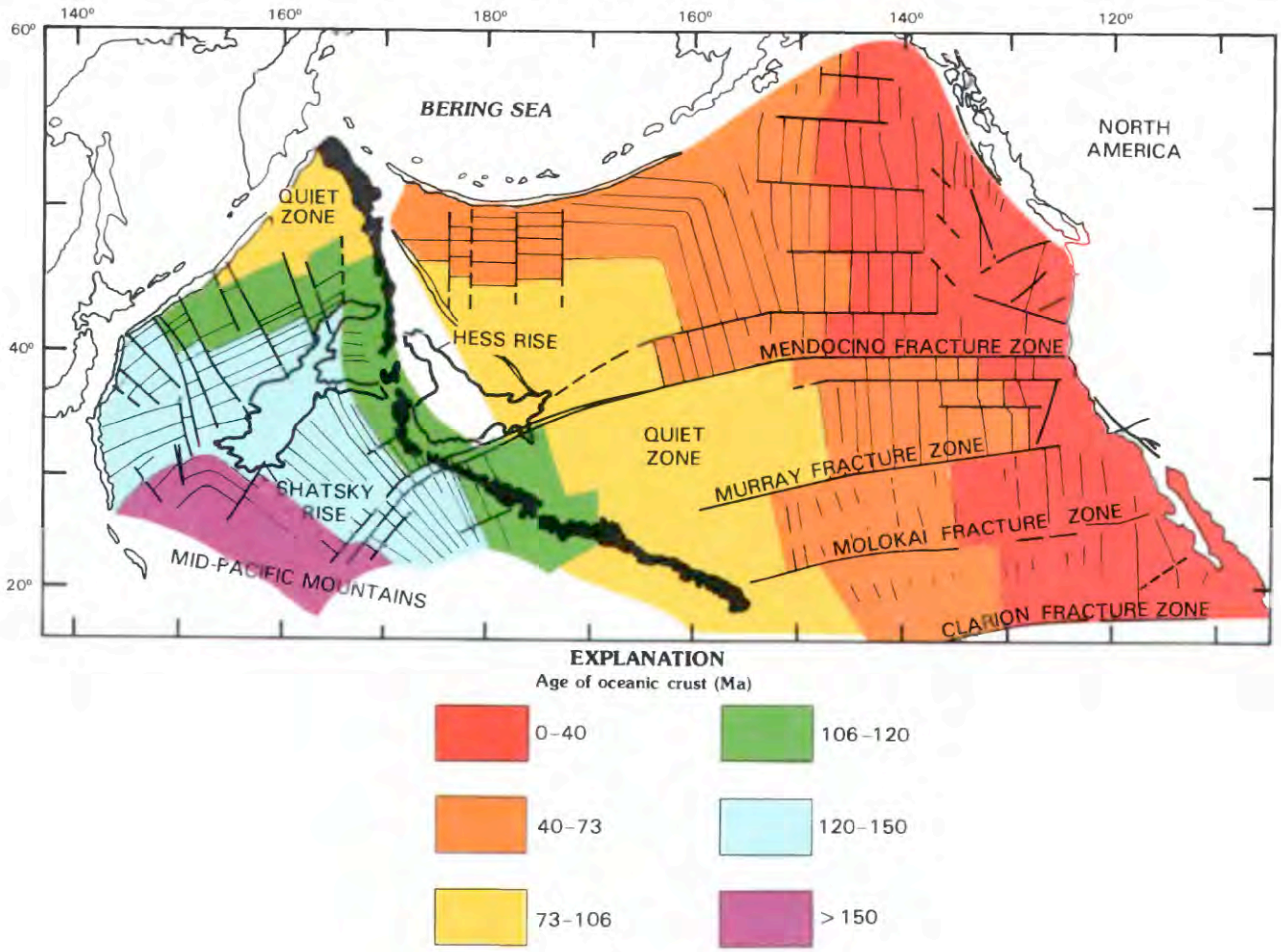

FIGURE 1.2.—Magnetic anomalies and structure of ocean floor crust in North Pacific modified from Hilde and others (1976). Hawaiian-Emperor Chain (black) crosscuts preexisting fracture zones and Mesozoic magnetic-anomaly sequence.

and small flows of lava richer in SiO_2 and FeO (these are the hawaiite and mugearite that characterize the alkalic postshield stage). S. Powers (1920) noted that eruptive centers of nepheline basalt on Kauai, Oahu, Molokai, and Maui were active long after the main volcano became quiet; he appears to have been the first to associate nepheline basalt with late-stage eruptions following an erosional hiatus.

New insight into the preshield stage has come from recent studies of Loihi Seamount, a small submarine volcano located about 30 km off the southeast coast of Hawaii. Its location, small size, seismic activity, and fresh, glassy lava all indicate that Loihi is an active volcano and the youngest in the Hawaiian-Emperor Chain. Some of the older lava samples recovered from Loihi Seamount are alkalic basalt and basanite, whereas the youngest lava samples recovered are tholeiitic and transitional basalt. This observation led Moore and others (1982) to conclude that Loihi Seamount, and perhaps all Hawaiian volcanoes, initially erupt alkalic basalt. Later, the bulk of the shield is built of tholeiitic basalt, but during declining activity the magma compositions revert to alkalic basalt. The alkalic preshield stage, like the alkalic postshield stage, produces only small volumes of lava, probably totaling less than a few percent of the volcano.

We have omitted the main caldera-collapse stage of Stearns (1966) from the eruption sequence because it can occur either during the shield stage or near the beginning of the alkalic postshield stage. The lava erupted may therefore be tholeiitic or alkalic basalt, or of both types.

GEOLOGY OF THE HAWAIIAN ISLANDS

Descriptions of volcanoes and their eruptions were made by nearly all the earliest visitors to the Hawaiian Islands. Descriptions of particular note are those of William Ellis (1823), George, Lord Byron (1826), Joseph Goodrich (1826, 1834), and Titus Coan

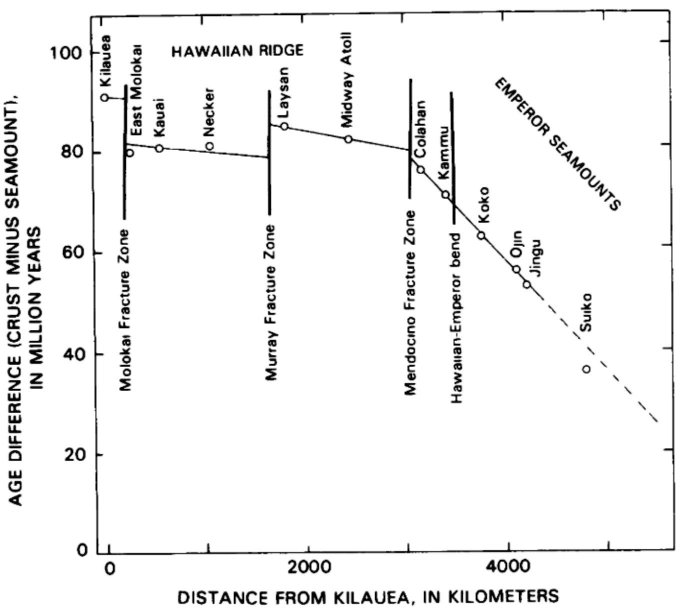

FIGURE 1.3.—Age of oceanic crust when overlying volcano formed, as a function of distance from Kilauea, for selected volcanoes in Hawaiian-Emperor Chain. Note offsets at fracture zones. Along Hawaiian Ridge both crust and volcanoes increase in age to west so crustal age when volcanoes formed is roughly constant. On the other hand, Emperor Seamounts increase in age to north but crust decreases in age; thus age of crust when seamounts formed decreases from roughly 75 Ma at the bend to less than 40 Ma at Suiko Seamount.

TABLE 1.1.—*Hawaiian eruptive products*

Eruptive stage	Rock types	Eruption rate	Volume (percent)
Rejuvenated --	alkalic basalt basanite nephelinite nepheline melilitite	Very low	<1
Postshield ---	alkalic basalt transitional basalt ankaramite hawaiite mugearite benmoreite trachyte phonolite	Low	~1
Shield -------	tholeiitic basalt olivine tholeiitic basalt picritic tholeiitic basalt icelandite (rare) rhyodacite (rare) [1]alkalic basalt (?)	High	95-98
Preshield ----	basanite alkalic basalt transitional basalt [1]tholeiitic basalt (?)	Low	~3

[1]Wright and Helz (chapter 23) suggest that the shield stage may include rare intercalated alkalic basalt and that the preshield stage includes tholeiitic basalt. We suspect that tholeiitic and alkalic basalt occur intercalated during the transitions from preshield to shield stage and from shield stage to postshield stage but that during the main shield stage only tholeiitic lava is erupted.

(1840 and other letters until 1882). Many of their letters describing the volcanoes were published in the American Journal of Science. Goodrich, in particular, provided detailed descriptions of the volcanoes on the Island of Hawaii. None of the earliest descriptions however, included information about the mineralogy or petrology of the lava.

The United States Exploring Expedition visited Hawaii in 1840–41. The commander of the expedition published a narrative (Wilkes, 1845) containing descriptions of caldera activity at Kilauea and new maps of both Kilauea and Mauna Loa calderas. James D. Dana, the geologist of the expedition, published a detailed report on the geology of the areas visited by the expedition (Dana, 1849). This report contains descriptions of lava flows including their mineralogy and flow morphology, in addition to numerous other observations on the active and inactive volcanoes that make up the islands. Later reports by Dutton (1884), J.D. Dana (1887, 1888, 1889), Green (1887), and Brigham (1909) added details on eruptions and expanded the geologic observations to other islands.

E.S. Dana (1877), Cross (1904), and Hitchcock (1911) presented detailed petrographic descriptions of lava from the islands. Daly (1911) and Cross (1915) described the mineralogy and petrology of Hawaiian lava flows at the time the Hawaiian Volcano Observatory was established, and Jagger (1917) described activity in Halemaumau lava lake. The paper by Cross (1915) is a milestone because it added detailed descriptions and chemical analyses of rocks from Hawaiian volcanoes other than Kilauea and Mauna Loa.

More detailed petrographic descriptions of lava from the islands were published by S. Powers (1920). Soon afterward, papers appeared by Washington (1923a, b, c) and Washington and Keyes (1926, 1928) with detailed accounts of the geology and petrology, new high-quality chemical analyses of lava from Hawaii and Maui, and a classification of Hawaiian volcanic rocks. Palmer (1927, 1936) added geologic descriptions and petrography of lava from Kaula and Lehua Islands, both of which are tuff cones of the alkalic rejuvenated stage. Lehua Island is just one of several rejuvenated-stage vents associated with Niihau, whereas Kaula Island sits atop a completely submerged shield.

These early, mainly descriptive and reconnaissance studies were superseded by detailed mapping of the islands beginning in the 1930's. H.T. Stearns and his coworkers, in a remarkable series of bulletins published by the Hawaii Division of Hydrography, published geologic maps and descriptions of Oahu (Stearns and Vaksvik, 1935; Stearns, 1939, 1940b), Lanai and Kahoolawe (Stearns, 1940c), Maui (Stearns and Macdonald, 1942), Hawaii (Stearns and Macdonald, 1946), Niihau (Stearns, 1947), and Molokai (Stearns and Macdonald, 1947). The bulletin on Kauai by Macdonald and others (1960) completed the monumental mapping job begun by Stearns; though Stearns did not coauthor the report, he did much of the mapping and is an author of the map. These maps and bulletins provide the geologic framework for all

TABLE 1.2.—*Eruptive stages represented on volcanoes of the Hawaiian-Emperor Chain*

[Volcano numbers from Bargar and Jackson (1974); numbers 0 and 65A added for consistency. Presence of stages: M, major unit; R, rare or of small volume; X, present but extent unknown; A, known to be absent; -, unknown. (T), transitional lava probably erupted during late shield stage or caldera-collapse phase of that stage. For volcanoes from Kilauea through Necker, data from detailed mapping and sampling; for remaining volcanoes, primarily from dredge and drill samples]

Volcano		Eruptive Stages			
Number	Name	Preshield (alkalic)	Shield (tholeiitic)	Postshield (alkalic)	Rejuvenated (alkalic)
	Hawaiian Islands				
0	Loihi	M	M	-	-
1	Kilauea	-	M	A	A
2	Mauna Loa	-	M	A	A
3	Mauna Kea	-	M	M	A
4	Hualalai	-	M	M	A
5	Kohala	-	M	M	A
6	East Maui	-	M	M	M
7	Kahoolawe	-	M	R	R
8	West Maui	-	M	R	R
9	Lanai	-	M	A	A
10	East Molokai	-	M	M	R
11	West Molokai	-	M	R	A
12	Koolau	-	M	A	M
13	Waianae	-	M	M	R
14	Kauai	-	M	R	M
15	Niihau	-	M	R	M
15A	Kaula	-	X	X	X
	Northwestern Hawaiian Islands and Hawaiian Ridge				
17	Nihoa	-	M	-	-
19	(Unnamed Seamount)	-	X(T)	-	-
20	(Unnamed Seamount)	-	X	-	X
21	(Unnamed Seamount)	-	X	-	-
23	Necker	-	M	X	X
26	La Perouse Pinnacles	-	X	-	-
28	Brooks Bank	-	X(T)	X	-
29	St. Rogatien Bank	-	-	X	-
30	Gardner Pinnacles	-	X	X	-
36	Laysan	-	-	X	-
37	Northampton Bank	-	X	-	-
39	Pioneer Bank	-	X	-	-
50	Pearl and Hermes Reef	-	-	X	-
51	Ladd Bank	-	-	-	X
52	Midway	-	M	X	-
53	Nero Bank	-	X	-	-
57	(Unnamed Seamount)	-	-	X	-
63	(Unnamed Seamount)	-	-	-	X
65	Colahan	-	X	-	X
65A	Abbott	-	X(T)	-	-
	Emperor Seamounts				
67	Daikakuji	-	X	-	-
69	Yuryaku	-	X	X	-
72	Kimmei	-	-	X	-
74	Koko (southern)	-	X	M	-
76	Koko (northwest)	-	X	-	-
81	Ojin	-	X	X	-
83	Jingu	-	-	X	-
86	Nintoku	-	-	X	-
90	Suiko (southern)	-	-	X	-
91	Suiko (central)	-	M	X	-
108	Meiji	-	M	-	-

subsequent studies of the islands and can also be used to put many of the earlier descriptions into a broader geological context. A number of derivative publications include summaries of the geology of the islands by Stearns (1946, 1966), an overview of the petrography of lava from the islands by Macdonald (1949), and a summary of the geology of the Hawaiian Islands by Macdonald and others (1983). The brief geologic summaries in appendix 1.1 have largely been extracted from the above publications. Additional unpublished observations by ourselves are included for Hualalai, East and West Molokai, Koolau, Kauai, and Niihau.

The maps of Stearns and coworkers separate rejuvenated-stage lava from earlier lava, but do not subdivide shield and postshield lava on the basis of chemical composition. The eruptive stages that are known to occur in each of the volcanoes of the Hawaiian Islands are summarized in table 1.2. Evidence for the alkalic preshield stage exists only at Loihi Seamount. If this stage is present in all Hawaiian volcanoes, it is completely buried by later, shield-stage tholeiitic lava. The tholeiitic shield stage is known to form the major portion of the subaerial and, we assume, the submarine part of each volcano. On the main islands, only Hualalai Volcano and Kaula Island do not have subaerial exposures of tholeiitic lava. Alkalic lava of the alkalic postshield stage occurs relatively late in the eruptive sequence and has not yet developed on Loihi Seamount or Kilauea and Mauna Loa Volcanoes. To the northwest of there it occurs on all volcanoes except Lanai and Koolau, although the volumes present on Kauai, Niihau, Kahoolawe, and West Molokai are small. Some volcanoes have predominantly mugearite, whereas others have predominantly hawaiite; these are called Kohala type and Haleakala type, respectively, by Macdonald and Katsura (1962). Wright and Clague (in press) propose two additional types: a Hualalai type with a bimodal trachyte-alkalic basalt lava distribution and a Koolau type with little or no alkalic postshield lava present.

Hawaiian volcanoes commonly have summit calderas and elongate curved rift zones from which much of the lava issues. Summit calderas exist on Loihi Seamount (Malahoff, chapter 6; Malahoff and others, 1982), Kilauea, and Mauna Loa. Each of these calderas is connected to two prominent rift zones. Not all Hawaiian volcanoes, however, had a summit caldera. West Molokai Volcano, in particular, shows no evidence of ever having had a caldera. Flat-lying lava ponded inside a caldera is not exposed on Hualalai, Mauna Kea, Kohala, or Niihau, but former calderas are inferred at those volcanoes from geophysical data (see Macdonald and others, 1983).

The formation and structure of the rift zones have been examined in an elegant paper by Fiske and Jackson (1972), who concluded that the orientation of the rift zones reflects local gravitational stresses within the volcanoes. Isolated shields such as Kauai and West Molokai had nearly symmetrical stress fields represented by generally radial dikes and thus have only poorly defined rift zones. The rift zones of these isolated volcanoes tend to align parallel to the orientation of the chain, suggesting the influence of a more regional stress field that also controls the orientation of the chain. In contrast, the rift zones of the other volcanoes tend to be aligned parallel to the flanks of the preexisting shields against which they abut.

GEOLOGY OF THE HAWAIIAN RIDGE

The Northwestern Hawaiian Islands were the focus of all geologic investigations along the Hawaiian Ridge west of Kauai until oceanographic techniques were applied to the area in the 1950's. Geological descriptions of the leeward islands include those of S. Powers (1920) for Nihoa and Necker Islands, and Washington and Keyes (1926) and Palmer (1927) for Nihoa, Necker, Gardner Pinnacles, and French Frigates Shoal (LaPerouse Pinnacles). These reports cite earlier sketchy descriptions. Macdonald (1949) reexamined Palmer's samples and added more detailed petrography. The petrology of the basaltic basement of Midway Atoll is described from two drill cores by Macdonald (1969) and Dalrymple and others (1974, 1977), whereas the geology of the site is detailed by Ladd and others (1967, 1969). Paleomagnetic data on flows and dikes from Nihoa and Necker Islands are given by Doell (1972), whereas similar data from the Midway drill core are given by Gromme and Vine (1972).

Marine geologic investigations of the Hawaiian Ridge began with Hamilton's pioneering work in 1957. Much subsequent work has focused on the structure of the oceanic crust in the vicinity of the chain, but few cruises have actually been conducted that dealt mainly with the geology of the Hawaiian Ridge. In the early 1970's, Scripps Institution of Oceanography and the Hawaii Institute of Geophysics conducted cruises to the Hawaiian Ridge. Samples collected by these cruises are described in Clague (1974a, 1974b) and Garcia, Grooms, and Naughton (in press). Subsequent cruises to the area by the Hawaii Institute of Geophysics and the U.S. Geological Survey are cited in appendix 1.1.

Lava samples recovered from the Hawaiian Ridge and Emperor Seamounts are more difficult to assign to volcanic stages because they are recovered by dredging and drilling or are collected from small islets and the field relations are usually unknown or only poorly known. Table 1.2 summarizes the available data from the Hawaiian Ridge. Based on the sequence and volumes of lava in the Hawaiian Islands, we have assumed that tholeiitic basalt always represents the shield stage and that strongly alkalic, SiO_2-poor lava represents the alkalic rejuvenated stage. Differentiated alkalic lava has been assigned to the alkalic postshield stage. Some alkalic basalt occurrences could be assigned to either the alkalic postshield or rejuvenated stages; they have been assigned on the basis of trace-element signatures and mineral chemistry using criteria outlined in a later section of this paper. No lava samples have been assigned to the alkalic preshield stage because we assume that the small volumes of such early lava have been buried by the later voluminous tholeiitic lava of the shield stage.

The samples recovered by dredging are probably not representative of the lava forming the bulk of the individual seamounts, but instead represent the youngest lava types erupted on the volcanoes. This natural sampling bias should result in an overrepresentation of alkalic lava from both the postshield and rejuvenated stages. In addition, selection of recovered samples for further study introduces another bias because the freshest samples are commonly alkalic lava, particularly hawaiite, mugearite, and trachyte. With these biases in mind, it is still possible to note general trends along the entire chain.

Tholeiitic basalt and picritic tholeiitic basalt, similar to those of the shield stage of subaerial Hawaiian volcanoes, have been recovered from 11 seamounts, banks, and islands in the Hawaiian Ridge west of Kauai and Niihau (table 1.2; appendix 1.1). The abundance of tholeiitic basalt from the Hawaiian Ridge implies that these volcanoes are genetically related to the Hawaiian Islands and that the general sequence of Hawaiian volcanism, in which tholeiitic basalt forms a major portion of each volcano, has occurred along the entire Hawaiian Chain.

GEOLOGY OF THE EMPEROR SEAMOUNTS

Little was known of the geology of the Emperor Seamounts until quite recently. The chain was recognized as the continuation of the Hawaiian Ridge by Bezrukov and Udintsev (1955), but not until 1968 were the first samples recovered from Suiko Seamount (Ozima and others, 1970). These samples are dominantly, if not completely, ice-rafted detritus. Subsequent studies included a cruise to the southern part of the chain by Scripps Institution of Oceanography in 1971 (ARIES Leg VII; Davies and others, 1971, 1972), Deep Sea Drilling Project (DSDP) Site 192 on Meiji Seamount (Creager and Scholl, 1973), DSDP Sites 308 and 309 on Koko Seamount (Larson and others, 1975), a cruise by the Hawaii Institute of Geophysics (Dalrymple and Garcia, 1980), a cruise by the U.S. Geological Survey in 1976 that surveyed the sites for Leg 55 of DSDP (Dalrymple and others, 1980a), and Leg 55 DSDP Sites 430, 431, 432, and 433 in the central part of the chain (Jackson and others, 1980). The Scripps Institution of Oceanography cruise ARIES VII in 1971 and the Leg 55 DSDP cruise in 1977 were particularly successful, and most of our knowledge of the Emperor Seamounts is derived from these two cruises.

The petrology of lava samples recovered by these two cruises is described in Clague (1974a) and Kirkpatrick and others (1980), respectively. A detailed seismic interpretation of the carbonate caps of many of the seamounts is given by Greene and others (1980), and overviews of the results of DSDP Leg 55 are given by Jackson and others (1980) and Clague (1981).

Table 1.2 summarizes the available data on eruptive stages represented by samples from the Emperor Seamounts, and details are given in appendix 1.1 for individual volcanoes. We have assumed that tholeiitic basalt represents the shield stage and that alkalic lava postdates the tholeiitic shield stage; only at Ojin and Suiko Seamounts does drilling show that the alkalic lava overlies the tholeiitic flows.

Tholeiitic basalt and picritic tholeiitic basalt similar to those of the shield stage of subaerial Hawaiian volcanoes have been recovered by drilling and dredging from six volcanic edifices in the Emperor Seamounts. The abundance of tholeiitic lava from the Emperor Seamounts is strong evidence that these volcanoes are genetically related to the Hawaiian Islands and Hawaiian Ridge. Likewise, the general eruptive model for the Hawaiian Islands is apparently applicable to the Emperor Seamounts.

Alkalic postshield-stage lava has been recovered by dredging and drilling from nine seamounts in the chain. In general these samples are alkalic basalt, hawaiite, mugearite, and trachyte similar to lava erupted in the Hawaiian Islands, but lava from Koko Seamount includes anorthoclase trachyte and phonolite that are interpreted to have erupted during the alkalic postshield stage (Clague, 1974a). Lava of the rejuvenated alkalic stage has not been identified from any of the Emperor Seamounts.

SUBSIDENCE OF THE VOLCANOES

Charles Darwin (1837, 1842) was the first to suggest that coral atolls might grow on subsiding platforms and that drowned atolls and certain deeply submerged banks with level tops could be explained by subsidence. Hess (1946) recognized that flat-topped submarine peaks, which he named guyots, were drowned islands. He thought that they were volcanic, bare of sediments and coral, and had been planed off by erosion at sea level. He attributed their depth to rising sea level caused by sediment deposition in the oceans. Menard and Dietz (1951) agreed with Hess that submergence was primarily due to a rise in sea level, but they thought that local subsidence might also play a role. Hamilton (1956), in his classic study of the Mid-Pacific Mountains, which included a program of dredging and coring, concluded that those (and other) guyots were formerly basaltic islands that had been wave and stream eroded and on which coral reefs subsequently grew. Their eventual submergence, he thought, was primarily caused by regional subsidence of the sea floor. It is now known that Darwin and Hamilton were basically correct about the steps leading to the formation of guyots, and about the predominant role of subsidence in the process.

The Hawaiian-Emperor volcanic chain is an excellent example of the gradual transformation of volcanic islands to guyots. From southeast to northwest there is a continuous progression from the active volcanoes such as Mauna Loa and Kilauea through the eroded remnants of Niihau, Nihoa, and Necker, through growing atolls like French Frigates Shoal and Midway Islands, to deeply submerged guyots like Ojin and Suiko. The progression can be observed not only along the chain but within the stratigraphy of individual seamounts. Drilling, dredging, and seismic observations have shown conclusively that the atolls and guyots of the chain are capped by carbonate deposits that overlie subaerial lava flows (see, for example, Ladd and others, 1967; Davies and others, 1971, 1972; Greene and others, 1980; Jackson and others, 1980).

The subsidence of Hawaiian volcanoes with time results from thermal aging of the lithosphere and isostatic response to local loading. The depth of the sea floor (and of volcanoes sitting upon it) increases away from spreading ridges because the lithosphere cools, thickens, and subsides as it moves away from the source of heat beneath the ridge (Parsons and Sclater, 1977; Schroeder, 1984). Detrick and Crough (1978) pointed out that the subsidence of many islands and seamounts, including those along the Hawaiian-Emperor Chain, was far in excess of that which could be accounted for by this normal lithospheric aging or by lithospheric loading. They proposed that the lithosphere is thermally reset locally as it passes over a hot spot and that the excess subsidence is largely a consequence of renewed lithospheric aging.

The Hawaiian-Emperor Chain rests on crust of Cretaceous age (circa 120–80 Ma) for which the depth should be about 5.5–5.9 km. The depth near Hawaii, however, is less than 4.5 km (fig. 1.4). The depth increases along the chain to about 5.3 km near the Hawaiian-Emperor bend in a manner consistent with the thermal resetting hypothesis. Thus, the subsidence of Hawaiian volcanoes as they move away from the hot spot is in part a function of their distance from the hot spot, that is, of the reset thermal age of the lithosphere beneath the chain. The volcanoes are passively riding away from the Hawaiian hot spot on cooling and thickening lithosphere that is subsiding at about 0.02 mm/yr.

Superimposed on the effect of crustal aging is subsidence caused by the immense and rapid loading of the lithosphere by the growing volcanoes (Moore, chapter 2). This effect is local, but while the volcano is active the rate of subsidence caused by loading may exceed that from lithospheric aging by more than two orders of magnitude. Moore (1970) found, from a study of tide-gage records in the Hawaiian Islands and on the west coast of North America, that Hilo on the Island of Hawaii has been subsiding at an absolute rate of 4.8 mm/yr since 1946. Recent data on drowned coral reefs near Kealakekua Bay indicate an absolute subsidence rate for the western side of Hawaii of 1.8 to 3+ mm/yr averaged over the past 300,000 yr and also indicate that the rate may have accelerated during that time (Moore and Fornari, 1984). Moore's (1970) tide-gage data also show that absolute subsidence decreases systematically away from the Island of Hawaii, with rates for Maui and Oahu of 1.7 mm/yr and 0 mm/yr, respectively. Some of this decrease in subsidence may be due to compensating uplift as the volcanoes are carried over the Hawaiian Arch, but an analysis of gravity data indicates that there is no appreciable viscous reaction to the seamount loads over time (Watts, 1978). Thus, it is probable that the volcanoes are isostatically compensated within a few million years of their birth, and that thermal aging of the lithosphere is the major cause of subsidence along the chain.

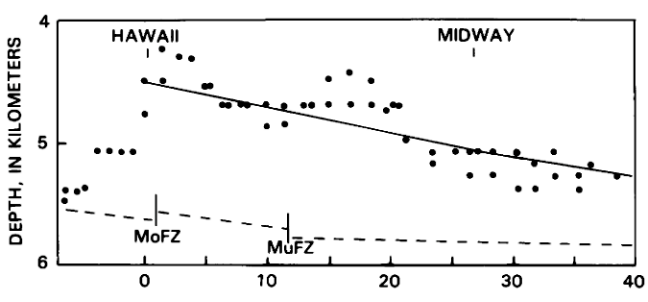

FIGURE 1.4.—Minimum depth to sea-floor swell as a function of time since reheating (or age of volcanoes along chain). Dashed line is predicted depth for normal aging of lithosphere away from spreading ridge. Solid line is predicted depth for thermally reset lithosphere 45 km thick. MoFZ and MuFZ, Molokai and Murray Fracture Zones, respectively. Modified from Detrick and Crough (1978) and Crough (1983).

GEOCHRONOLOGY AND PROPAGATION OF VOLCANISM

EARLY WORK: LEGENDS AND DEGREE OF EROSION

According to Hawaiian legend, the goddess Pele first inhabited Kauai, but then moved southeastward island by island to Kilauea Volcano, where she now resides (Bryan, 1915). The reasoning behind this legend is unknown, but it was probably based in large part on the relative appearance of age of the various volcanoes. Many centuries after this legend originated, J.D. Dana (1849) rendered the first scientific opinion confirming the general age progression implied by the legend.

Dana was not only the first geologist to conclude that the order of extinction of Hawaiian volcanoes was approximately from northwest to southeast, he also recognized that the Hawaiian Chain included the islets, atolls, and banks that stretch for some distance to the northwest of Kauai. Dana saw no reason to think that the volcanoes of the chain did not originate simultaneously: "No facts can be pointed to, which render it even probable that Hawaii is of more recent origin than Kauai" (Dana, 1849, p. 280). Their relative degree of erosion, however, provided ample evidence to indicate their order of extinction: "From Kauai to Mount Loa all may thus have simultaneously commenced their ejections, and have continued in operation during the same epoch till one after another became extinct. Now, the only burning summits out of the thirteen which were once in action from Niihau to Hawaii, are those of Loa and Hualalai: we might say farther that these are all out of a number unknown, which stretched along for fifteen hundred miles, the length of the whole range. This appears to be a correct view of the Hawaiian Islands" (Dana, 1849, p. 280). Subsequent workers agreed with Dana on the general order of extinction (for example, Brigham, 1868; Dutton, 1884; Hillebrand, 1888; Hitchcock, 1911; Cross, 1915; Martin and Pierce, 1915; Wentworth, 1927; Hinds, 1931; Stearns, 1946), although the sequences they proposed invariably differed in detail (table 1.3). Of these various workers only Stearns (1946), who studied the Hawaiian Islands in more detail than any of his predecessors, had the sequence exactly correct as judged by present data.

The idea that the volcanoes of the Hawaiian Chain originated simultaneously and only became extinct progressively seems to have persisted until a few decades ago. Stearns (1946), for example, mentions the lack of evidence to indicate when any of the Hawaiian volcanoes began but shows all of the main shields except Hualalai and Kilauea erupting simultaneously at the end of the Pliocene (Stearns, 1946, p. 97, fig. 25). Two exceptions were Cross (1904) and Wentworth (1927), who thought that the degree of erosion was probably a function of when the volcanoes emerged above the sea as well as of the elapsed time since they ceased to erupt. Cross (1904, p. 518) states: "It appears to me plausible to assume that the earliest eruptions occurred at or near the western limit of this zone (the more than 1000 mile expanse of the island chain), and that in a general way at least, the centers of activity have developed successively farther and farther to the east or southeast, until now the only active loci of eruption are those of Mauna Loa and Kilauea on the island of

TABLE 1.3.—*Early estimates of the order of extinction of the principal Hawaiian volcanoes*

[Criteria used are given beneath each source; volcanoes listed in proposed order of extinction, oldest at top]

Dana (1849) Erosion	Brigham (1868) Erosion	Dana (1888) Erosion	Hillebrand (1888) Floral diversity	Wentworth (1927) Erosion	Hinds (1931) Erosion	Stearns (1946) Erosion and stratigraphy
Kauai	West Kauai, Niihau	Kauai	West Oahu, Kauai	Koolau	Waianae	Kauai
Waianae	Waianae	Waianae	Molokai, East Oahu	Kauai	Koolau	Waianae
West Maui	East Kauai	West Maui	Kohala, West Maui	East Molokai	Niihau	Koolau
Koolau	West Molokai	Kohala	Mauna Kea	West Maui	Kauai	West Molokai
Mauna Kea	West Maui	Koolau	East Maui	Mauna Kea	West Molokai	East Molokai
East Maui	Kohala	East Maui	Hualalai	Waianae	East Molokai	West Maui, Lanai
Mauna Loa	Koolau	Mauna Kea	Mauna Loa, Kilauea	East Maui	Lanai	Kahoolawe
	East Molokai	Hualalai		Lanai	West Maui, Kohala	East Maui
	Mauna Kea	Mauna Loa, Kilauea		Niihau, West Molokai	Kahoolawe	Kohala
	Lanai, Kahoolawe			Kahoolawe	East Maui	Mauna Kea
	East Maui			Kohala	Mauna Kea	Haulalai, Mauna Loa, Kilauea
	Hualalai			Hualalai, Mauna Loa, Kilauea	Hualalai	
	Mauna Loa, Kilauea				Mauna Loa, Kilauea	

Hawaii." He specifically noted the difference between his hypothesis and that of Dana.

Estimates of the geologic ages of the Hawaiian volcanoes varied considerably among those early workers willing to hazard a guess on the basis of the meager data then available. Dana (1849) thought it likely that the eruptions commenced as early as early Carboniferous or Silurian time; this estimate was based on the concept that the Earth had cooled from a molten globe producing fissuring and volcanism, the apparent lack of post-Silurian volcanism in the interior of the North American continent, and the presumption that the oceans would cool after the continents. Cross (1904) speculated that the western part of the leeward islands formed in the early part of the Tertiary. Wentworth (1925, 1927) attempted to quantify erosion rates for several of the islands and estimated the extinction ages of some of the volcanoes as follows:

Lanai	0.15 Ma
Kohala	0.22 Ma
Koolau	1.00 Ma
Kauai	2.09 Ma

On the basis of physiographic evidence, Wentworth doubted that any part of the Hawaiian group emerged above sea level before late Tertiary time. Hinds (1931), like Cross (1904), recognized that the atolls and banks of the leeward islands were the remnants of once-larger volcanoes: "The landscapes of the leeward group—the volcanic stacks, the reef limestone and calcareous sand islands rising from submarine platforms, and submerged platforms from which no islands rise, represent the final stages in the destruction of a volcanic archipelago. Such a fate awaits the windward islands unless they be rejuvenated by volcanic or diastrophic forces" (Hinds, 1931, p. 196). He recognized that the amount of erosion and subsidence required to reduce a mammoth Hawaiian volcano to a coral atoll was probably considerable and concluded: "The complete or nearly complete destruction of the Leeward islands suggests that volcanism ceased there well back in the Tertiary, hence the mountains must have risen above the ocean long before, perhaps even in Mesozoic time" (Hinds, 1931, p. 205). On the basis of geomorphic considerations, Stearns (1946) thought that the volcanoes of the main Hawaiian Islands rose above sea level in the Tertiary.

RADIOMETRIC AND FOSSIL AGES

The first radiometric ages for Hawaiian volcanoes were determined by McDougall (1963), who measured ages of 2.8 to 3.6 Ma for his Middle and Upper Waianae Series on Oahu (all K-Ar ages have been converted to the new constants; Steiger and Jaeger, 1977). He also reported an age of 8.6 Ma for what he called the Mauna Kuwale Trachyte of the Lower Waianae Series(?), an age that later proved to be incorrect, probably because of excess argon in the biotite analyzed (Funkhouser and others, 1968). In subsequent studies McDougall and Tarling (1963) and (primarily) McDougall (1964) reported K-Ar ages of lava from 7 of the principal Hawaiian volcanoes and concluded that the ages of the shield stages were approximately:

Kauai	5.8–3.9 Ma
Waianae	3.5–2.8 Ma
Koolau	2.6–2.3 Ma
West Molokai	1.8 Ma
East Molokai	1.5–1.3 Ma
West Maui	1.3–1.15 Ma
East Maui	0.8 Ma
Hawaii (all 5 volcanoes)	<1 Ma

McDougall thus confirmed Stearns' extinction sequence and also suggested that the main shield stage of a Hawaiian volcano essentially was complete before the next volcano rose above the sea.

Since the pioneering work of McDougall, many additional radiometric ages have been determined for the volcanoes of the main islands, and the dating has been extended to the volcanoes of the leeward islands, the western Hawaiian Ridge, and the Emperor Seamounts. In total, there are now reasonably precise radiometric age data for 35 of the volcanoes in the Hawaiian-Emperor Chain (see appendix 1.1). Radiometric ages of two volcanoes on the Hawaiian Ridge are not included in appendix 1.1 because it is probable that the samples are not from Hawaiian volcanoes (Clague and Dalrymple, 1975). These include a minimum age of 71 ± 5 Ma for altered basalt from Wentworth Seamount, 80 km northwest of Midway, and an age of 77.6 ± 1.7 Ma for a sample of rhyolite (probably an erratic, see appendix 1.1) dredged from the northern slope of Necker Island.

In addition to the radiometric age data, there are paleontologic ages for several of the Hawaiian-Emperor volcanoes based on material recovered by dredging and drilling programs. In general, these ages postdate volcanic activity and are consistent with the radiometric data. From southeast to northwest they include (1) an age of 28–31 Ma for late Oligocene nannofossils in volcanogenic sediments at DSDP Site 311 on the archipelagic sediment apron of an unnamed seamount (no. 58 of Bargar and Jackson, 1974) 240 km northwest of Midway (Bukry, 1975); (2) an age of 15–32 Ma (East Indies Tertiary stage Te) for larger foraminifers (Cole, 1969) and smaller foraminifers (Todd and Low, 1970) in reef limestone above basalt in a drill hole at Midway Atoll; (3) an age of 39–41 Ma for dredged late Eocene larger foraminifers from Kammu Seamount (Sachs, quoted in Clague and Jarrard, 1973); (4) an age of 50.5 ± 3.5 Ma for early Eocene coccoliths in volcanogenic sediments cored at DSDP Site 308 atop Koko Seamount (Bukry, 1975); (5) an age of 57–59 Ma for late Paleocene calcareous nannofossils (Takayama, 1980) and pelagic foraminifers (Hagn and others, 1980) in sediments above basalt at DSDP Site 430 on Ojin Seamount; (6) late Paleocene planktonic foraminifers and probable early Eocene benthic foraminifers in sediments above basalt at DSDP Site 432 on Nintoku Seamount (Butt, 1980); (7) an age of 59–61 Ma for middle Paleocene calcareous nannofossils in sediments above basalt at DSDP Site 433 on Suiko Seamount (Takayama, 1980); and (8) an age of 70–73 Ma for lower Maestrichtian nannofossils from sediments above basalt at DSDP Site 192 on Meiji Seamount at the northern end of the Emperor Seamounts (Worsley, 1973). None of these fossil ages is in conflict with the radiometric data.

On the other hand, Menard and others (1962) describe Miocene corals and pelagic foraminifers dredged from a submarine terrace 10 km southwest of Oahu. The authors note the difficulty in assigning an age to these samples and state that the "planktonic foraminifera *Globigerinoides quadralobates* [=*G. trilobus auct.*] *plexus* suggest a lower limit of early Miocene. The upper age limit is less definitive" (Menard and others, 1962, p. 896). Present nomenclature would identify these samples as *Globigerinoides triloba*, which ranges in age from early Miocene to Pleistocene (Kenneth and Srinivasan, 1983). Menard and others (1962) cite as additional evidence of the Miocene age of the sample the 60 percent of extinct coral species in the sample. We conclude that none of these criteria unequivocally supports a Miocene age and no conflict exists between the ages of the reef and that of the underlying volcanic basement of 1.8–2.7 Ma.

Two samples of Eocene terrigenous sediment recovered 250 km east of Hawaii and 100 km south of Kauai (Schreiber, 1969) are also anomalous. These samples were probably derived from volcanoes that predate the Hawaiian Chain, or they may have been reworked from sediment on the sea floor during formation of the Hawaiian volcanoes.

The available radiometric data are summarized in table 1.4 and plotted in figure 1.5 as a function of distance measured from Kilauea Volcano along the Hawaiian-Emperor trend. Because some of the volcanoes are unnamed and some seamounts and islands consist of more than one major volcanic edifice, each dated volcanic center is identified in the table with the number assigned to it by Bargar and Jackson (1974). The exceptions are Abbott Seamount, a small volcano between Colahan and Kammu Seamounts, and Kaula Island, which were not previously numbered and to which we have assigned numbers 65A and 15A, respectively.

As can be seen from figure 1.5, the age data confirm the general age progression along the chain as first suggested by Dana (1849) and required by the hot-spot hypothesis of Wilson (1963a), and they show that the progression is continuous from Kilauea at least to Suiko Seamount, more than half way up the Emperor Seamounts Chain and nearly 5,000 km from the active volcanoes of Mauna Loa and Kilauea. The data also substantiate the hypothesis that the Emperor Seamounts are a continuation of the Hawaiian Chain, as proposed by Christofferson (1968) and Morgan (1972a, b).

RATES OF VOLCANIC PROPAGATION

In order to determine accurately the rate of volcanic propagation along the Hawaiian-Emperor Chain, we would like to know the time that each tholeiitic shield volcano first erupted onto the sea floor, but such data clearly are not obtainable. What is available for the dated volcanoes is one or more radiometric age on lava flows erupted during one or more stage of volcanic activity (see appendix 1.1). In order to calculate propagation rates, therefore, it is necessary to adopt some consistent strategy for selecting the numerical age used to represent the age of each dated volcano. Different authors have approached this problem in different ways. McDougall (1971) used the youngest age of tholeiitic basalt as representing the time of cessation of volcanism for each dated volcano in the principal Hawaiian Islands. In contrast, Jackson and others (1972) and Dalrymple and others (1980b, 1981) used the oldest age for tholeiitic volcanism as the best available approximation of the age of the volcanoes. McDougall (1979) and McDougall and Duncan (1980) adopted yet another approach and used the average age of tholeiitic shield volcanism. For table 1.4, we have chosen the oldest reliable ages for tholeiitic volcanism available, but the choice of which ages to use is probably not critical when considering the data for the chain as a whole. The reason is that the existing data on the rate of formation of Hawaiian volcanoes indicate that the tholeiitic shields are probably built up from the sea floor in as little as 0.5–1.5 m.y. (see summary in Jackson and others, 1972). This amount of time is within the analytical uncertainty of the K-Ar ages at about 20 Ma, or less than one-third of the way along the dated

TABLE 1.4.—*Summary of K-Ar geochronology along the Hawaiian-Emperor volcanic chain*

[Volcano number and distance from Bargar and Jackson (1974) and K.E. Bargar (written commun., 1978). Best K-Ar age is oldest reliable age of tholeiitic basalt, where available; all data converted to new constants $\lambda\varepsilon + \lambda\varepsilon' = 0.581 \times 10^{-10}$/yr, $\lambda\beta = 4.962 \times 10^{-10}$/yr, $^{40}K/K = 1.167 \times 10^{-4}$ mol/mol]

Volcano Number	Volcano Name	Distance from Kilauea along trend of chain (km)	Best K-Ar age (Ma)	Data source	Remarks
1	Kilauea	0	0-0.4	--	Historical tholeiitic eruptions
3	Mauna Kea	54	0.375±0.05	1	Samples from tholeiitic shield (Hamakua Volcanics)
5	Kohala	100	0.43±0.02	2	Samples from tholeiitic shield (Pololu Basalt)
6	Haleakala	182	0.75±0.04	3	Samples from tholeiitic shield (Honomanu Basalt)
7	Kahoolawe	185	>1.03±0.18	3	Samples from alkalic postshield stage (upper part of Kanapou Volcanics)
8	West Maui	221	1.32±0.04	4	Samples from tholeiitic shield (Wailuku Basalt)
9	Lanai	226	1.28±0.04	5	Samples from tholeiitic shield (Lanai Basalt)
10	East Molokai	256	1.76±0.07	3	Samples from tholeiitic shield (lower member of East Molokai Volcanics)
11	West Molokai	280	1.90±0.06	3	Samples from tholeiitic shield (lower part of West Molokai Volcanics)
12	Koolau	339	2.6±0.1	4,6	Samples from tholeiitic shield (Koolau Basalt)
13	Waianae	374	3.7±0.1	6	Samples from tholeiitic shield (lower member of Waianae Volcanics)
14	Kauai	519	5.1±0.20	7	Sample from tholeiitic shield (Napali Member of Waimea Canyon Basalt)
15	Niihau	565	4.89±0.11	8	Samples from tholeiitic shield (Paniau Basalt)
15A	Kaula	600	4.0±0.2	21	Phonolite from postshield stage (?)
17	Nihoa	780	7.2±0.3	9	Samples from tholeiitic shield
20	Unnamed	913	9.2±0.8	20	Dredged samples of alkalic basalt
23	Necker	1,058	10.3±0.4	9	Samples from tholeiitic shield
26	La Perouse Pinnacles	1,209	12.0±0.4	9	Samples from tholeiitic shield
27	Brooks Bank	1,256	13.0±0.6	20	Dredged samples of hawaiite and alkalic basalt
30	Gardner Pinnacles	1,435	12.3±1.0	20	Dredged samples of alkalic and tholeiitic basalt
36	Laysan	1,818	19.9±0.3	10	Dredged samples of hawaiite and mugearite
37	Northampton Bank	1,841	26.6±2.7	10	Dredged samples of tholeiitic basalt
50	Pearl and Hermes Reef	2,281	20.6±0.5	11	Dredged samples of phonolite, hawaiite, and alkalic basalt
52	Midway	2,432	27.7±0.6	12	Samples of mugearite and hawaiite from conglomerate overlying tholeiitic basalt in drill hole
57	Unnamed	2,600	28.0±0.4	11	Dredged samples of alkalic basalt
63	Unnamed	2,825	27.4±0.5	11	Dredged samples of alkalic basalt
65	Colahan	3,128	38.6±0.3	13	Dredged samples of alkalic basalt
65A	Abbott	3,280	38.7±0.9	13	Dredged samples of tholeiitic (?) basalt
67	Daikakuji	3,493	42.4±2.3	14	Dredged samples of alkalic basalt
69	Yuryaku	3,520	43.4±1.6	11	Dredged samples of alkalic basalt
72	Kimmei	3,668	39.9±1.2	14	Dredged samples of alkalic basalt
74	Koko (southern)	3,758	48.1±0.8	14,15	Dredged samples of alkalic basalt, trachyte, and phonolite
81	Ojin	4,102	55.2±0.7	16	Samples of hawaiite and tholeiitic basalt from DSDP Site 430
83	Jingu	4,175	55.4±0.9	17	Dredged samples of hawaiite and mugearite
86	Nintoku	4,452	56.2±0.6	16	Samples of alkalic basalt from DSDP Site 432.
90	Suiko (southern)	4,794	59.6±0.6	18,19	Single dredged sample of mugearite
91	Suiko (central)	4,860	64.7±1.1	16	Samples of alkalic and tholeiitic basalt from DSDP Site 433

Data sources:

1. Porter and others (1977)
2. McDougall and Swanson (1972)
3. Naughton and others (1980)
4. McDougall (1964)
5. Bonhommet and others (1977)
6. Doell and Dalrymple (1973)
7. McDougall (1979)
8. G.B. Dalrymple (unpub. data, 1982)
9. Dalrymple and others (1974)
10. Dalrymple and others (1981)
11. Clague and others (1975)
12. Dalrymple and others (1977)
13. Duncan and Clague (1984)
14. Dalrymple and Clague (1976)
15. Clague and Dalrymple (1973)
16. Dalrymple and others (1980a)
17. Dalrymple and Garcia (1980)
18. Saito and Ozima (1975)
19. Saito and Ozima (1977)
20. Garcia and others (1986b)
21. Garcia and others (1986a)

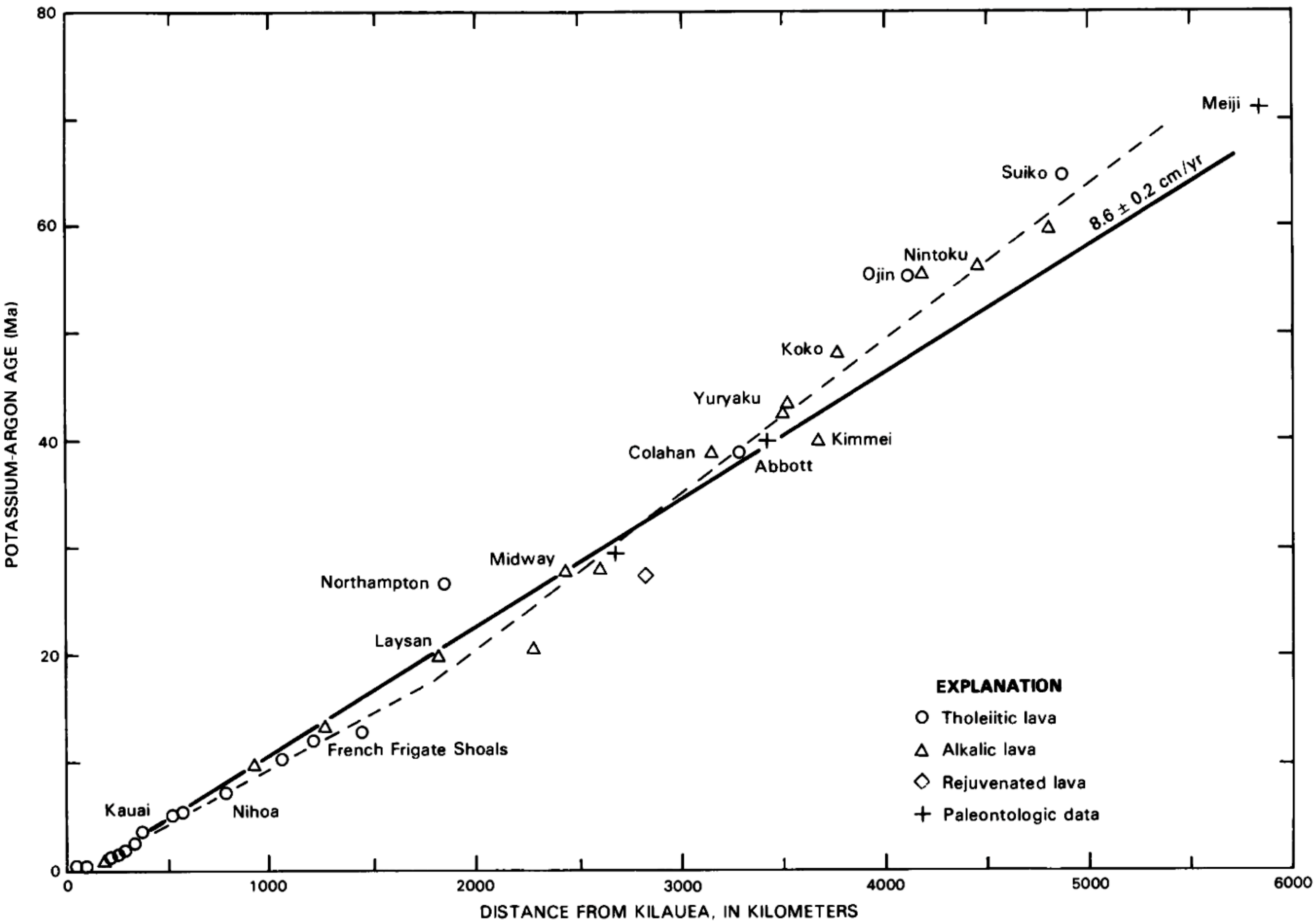

FIGURE 1.5.—Age of volcanoes in Hawaiian-Emperor Chain as a function of distance from Kilauea. Solid line is least-squares cubic fit (York 2) from table 1.5 and represents average rate of propagation of volcanism of 8.6 ± 0.2 cm/yr. Dashed line is two-segment fit using data from Kilauea to Gardner and Laysan to Suiko (table 1.5). Radiometric data from table 1.4, paleontologic data discussed in text.

part of the chain. The question of which ages to use is in any case moot for most of the volcanoes west of Kauai because so few suitable samples have been recovered that there is rarely a choice to make.

A majority of the age data from islands and seamounts west of French Frigates Shoal were obtained on alkalic rocks rather than on tholeiitic basalt. This is because the alkalic rocks, being younger than the tholeiitic basalt, are more likely to be recovered by dredging and drilling and are more resistant to submarine alteration than tholeiitic basalt. This bias toward ages of alkalic lava also probably makes very little difference because the difference between the ages of the postshield alkalic and youngest tholeiitic rocks is only a few hundred thousand years in the Hawaiian Islands (McDougall, 1964, 1969; Funkhouser and others, 1968; McDougall and Swanson, 1972; Doell and Dalrymple, 1973) and, presumably, in the other volcanoes of the Hawaiian-Emperor Chain. For one unnamed volcano on the western Hawaiian Ridge (63 in appendix 1.1 and table 1.4) the only data available are from lava erupted during the alkalic rejuvenated stage. In the principal Hawaiian Islands, lava erupted during the rejuvenated stage may postdate the tholeiitic shield and alkalic postshield stages by more than 4 m.y. (McDougall, 1964; G.B. Dalrymple, unpublished data, 1985), so the main shield of volcano 63 may be several million years older than indicated in table 1.4.

Previously, age-distance data along the Hawaiian-Emperor Chain have been regressed using a simple linear regression of age (dependent variable) on distance (independent variable), either unconstrained (for example, McDougall, 1971, 1979; Jackson and others, 1972; McDougall and Duncan, 1980) or forced through the origin (for example, Dalrymple and others, 1980b, 1981). The resulting volcanic propagation rates for the Hawaiian segment of the chain have ranged from as little as 6 cm/yr (Jackson and others, 1975) to as much as 15 cm/yr (McDougall, 1971; Jackson and others, 1972), although most recent estimates have been between 8 and 10 cm/yr (McDougall, 1979; McDougall and Duncan, 1980;

Dalrymple and others, 1981). Simple linear regression models have the disadvantages that they presume no error in distance and they do not take into account the experimental errors of individual determinations.

We have treated the data in table 1.4 using a two-error cubic fit (York 2), which allows for errors in both age and distance and weights the data accordingly (York, 1969). Errors for the age determinations are straightforward and are either provided in the original references or have been estimated by us from the array of data available on an individual volcano. Jackson and others (1975) estimated the cumulative errors in distance to be about 1.5 km at Kilauea to as much as 20 km near the western end of the Hawaiian Chain. We have interpolated and extrapolated these values to find errors for the distances in table 1.4. The results of both the York 2 regressions and the two simpler regression models for various segments of the Hawaiian-Emperor Chain are given in table 1.5. For the entire chain, the average rate of volcanic propagation is 8.6 ± 0.2 cm/yr with an intersection (that is, theoretical zero time) 89 km northwest of Kilauea using the York 2 regression. The simple regression models yield similar, though slightly lower, values of propagation rate.

Rates of propagation for the Hawaiian segment of the chain, that is, Kilauea through Abbott, have been calculated using both the maximum and minimum ages of tholeiitic volcanism. The results do not vary with model and range from 8.6 to 9.2 cm/yr. For comparison, we have included comparable calculations using the average ages of McDougall (1979). The resulting rates are somewhat higher than the rates calculated from either the maximum or

TABLE 1.5.—*Rates of propagation of volcanism along segments of the Hawaiian-Emperor Chain for several linear-regression models*

[Rates are in centimeters per year. Intercept, in kilometers, and correlation coefficient, r, given in parentheses where relevant. Simple regression is of age on distance, unweighted data; York 2 fit is two-error cubic, weighted data]

Chain segment	Data source	Simple regression		York 2 fit
		Unconstructed	Forced through origin	
Hawaiian-Emperor ---	table 1.4	7.8±0.2 (175, 0.992)	8.2±0.1	8.6±0.2 (89)
Hawaiian -----------	table 1.4 (maximum ages)	8.6±0.3 (102, 0.985)	9.1±0.2	9.2±0.3 (80)
	McDougall, 1979 (average ages)	9.4±0.3 (91, 0.994)	9.9±0.2	11.3±0.1 (3)
	appendix 1.1 (minimum ages)	8.6±0.3 (119, 0.986)	9.1±0.2	9.1±0.3 (97)
Emperor ------------	table 1.4	6.5±0.8	7.9±0.2	7.2±1.1
Kilauea to Gardner -	table 1.4	9.9±0.3 (57, 0.992)	10.6±0.3	9.6±0.4 (73)
Laysan to Suiko ----	table 1.4	6.9±0.4 (0.971)	8.1±0.2	6.8±0.3
Gardner to Waianae -	table 1.4	9.5±0.4 (54, 0.993)	10.1±0.2	10.1±0.8 (9)

minimum data, but the difference is largely a consequence of differences in the data sets, the ones in table 1.4 and appendix 1.1 being more current.

Rates calculated for the Emperor Chain, that is, Daikakuji through Suiko, are markedly lower than for the Hawaiian Chain, ranging from 6.5 to 7.2 cm/yr when not forced through the origin. Separate rates for these two major segments of the chain are only meaningful, however, if there was a rate change at the time of formation of the bend. This hypothesis can be tested by using the linear equations found from the York 2 regressions to predict the age of the bend, which we estimate to be 3,451 km from Kilauea at the position of volcano 68. The predicted ages for the bend are 36.7 Ma and and 43.0 Ma for the Hawaiian and Emperor segments, respectively. The Hawaiian prediction, which is similar to the value of 37.8 Ma found by McDougall (1979), differs significantly from the measured bend age of 43.1 ± 1.4 Ma as determined from the ages of Daikakuji and Yuryaku Seamounts. This suggests that if there was a significant change in volcanic propagation rate it did not occur at bend time, but some time after, a conclusion also reached by Epp (1978).

For some time, it has been apparent to us that a change in rate near or before the time of formation of Midway is consistent with the available data (Dalrymple and others, 1980b). For example, the fits of the data for the chain segments Kilauea-Gardner and Laysan-Suiko are slightly better than the fits for the Hawaiian and Emperor segments (table 1.5; fig. 1.5). The two former lines intersect near Gardner Pinnacles at an age of about 18 Ma. Epp (1978) concluded that a rate change occurred around 20–25 Ma. We have tried various ways to determine the most likely time for a change in the rate of volcanic propagation, including correlation with eruption volumes along the chain (see section below on eruption rates) and age-predictive models for the central parts of the chain, but we are not convinced that the results are meaningful. We can only conclude that the data imply, but do not require, a change of rate sometime after the formation of the Hawaiian-Emperor bend and before or near the time of formation of Laysan Volcano.

In addition to the possibility of a major change in the volcanic propagation rate, as discussed above, there are also indications of short-term departures from linearity. Short-term changes in the volcanic propagation rate were first proposed by Jackson and others (1972) to explain the apparent acceleration of propagation during the past 5 m.y. or so. They did not suggest that short-term variations in propagation rate reflected variations in relative motion of hot spot and plate. Shaw (1973) and Walcott (1976) proposed thermal feedback mechanisms to account for such variations without varying the relative rate of motion between the hot spot and the Pacific plate (see section below on models). Nonlinear models have been disputed by McDougall (1979) and McDougall and Duncan (1980), who argue that linear regressions fit the Hawaiian data so well that no other model needs to be considered.

It seems obvious to us from the geometry alone, however, that the volcanic propagation rates must be nonlinear in detail. If this were not so, then either the volcanism would have formed a ridge rather than individual volcanoes, or the volcanoes in the chain would be spaced in proportion to their ages along a single line. Neither is the case; the volcanoes are irregularly spaced within a band some 200–300 km wide, indicating clearly that volcanic propagation is irregular.

Although some of the irregularities in the age-distance data no doubt reflect dating errors and differences in the stage of volcanism sampled, some of the deviations appear to be larger than can reasonably be attributed to these causes. For example, the ages of Laysan and Northampton Bank should differ by only about 0.3 m.y. rather than the 6.7 m.y. indicated by their measured ages. A similar discrepancy occurs in the ages of volcanoes near the bend (table 1.4; fig. 1.5). There are also volcanoes in the chain that appear to have been active simultaneously even though they were separated by distances of hundreds of kilometers. Examples include Laysan, and Pearl and Hermes, as well as Midway and Northampton. Indeed, Mauna Loa, Kilauea, and Loihi are currently active, erupting tholeiitic basalt, and are separated by more than 80 km.

The primary reason that Jackson and others (1972) suggested short-term nonlinearities in propagation rates was the pronounced curvature in the age-distance data from the volcanoes of the principal Hawaiian Islands. When plotted as a function of distance from Kilauea, the ages for these volcanoes clearly indicate an acceleration of volcanic propagation over the past 5 m.y. (Jackson and others, 1972). This curvature is also one reason that virtually all regressions intersect the distance axis northwest of Kilauea (table 1.5) and predict a negative age for that volcano. McDougall (1979) has argued that the curvature is caused by a bias toward young ages for the less eroded volcanoes, but this cannot be so. Even though Kohala Mountain is relatively uneroded, it is deeply incised on the windward side by several canyons whose floors are near sea level, and it is unlikely that further erosion will expose lava significantly older than is now exposed. In addition, the rapid subsidence of Hawaii (Moore, 1970) may carry the oldest subaerial lava flows below sea level before they can be exposed by erosion. Similar arguments can be made for West Maui, Lanai, Kahoolawe, East Molokai, and Koolau Volcanoes, where lava deep within the subaerial part of the tholeiitic shield has been exposed by marine or stream erosion or by faulting.

We have plotted the known age range for tholeiitic shield, alkalic postshield, and alkalic rejuvenated-stage volcanism for the principal Hawaiian volcanoes in figure 1.6, from which the acceleration of volcanic propagation over the past 3–5 m.y. is evident. It is also clear from this figure that the curvature in the age-distance data is not a function of which eruption stage, tholeiitic shield or alkalic postshield, is chosen to represent the age of the volcanoes. Furthermore, a bias toward younger ages for the less eroded volcanoes, taken to include Kilauea through Haleakala, cannot produce the curvature because older ages for these volcanoes would exaggerate, not lessen, the apparent acceleration. Thus, the acceleration of volcanic propagation in the principal islands, as proposed by Jackson and others (1972), appears to us to be real.

While the overall rate of propagation of volcanism along the chain (or at least major segments of it) may be linear and reflect the relative motion between the Pacific plate and the Hawaiian hot spot, there also appears to be ample justification for retaining nonlinear

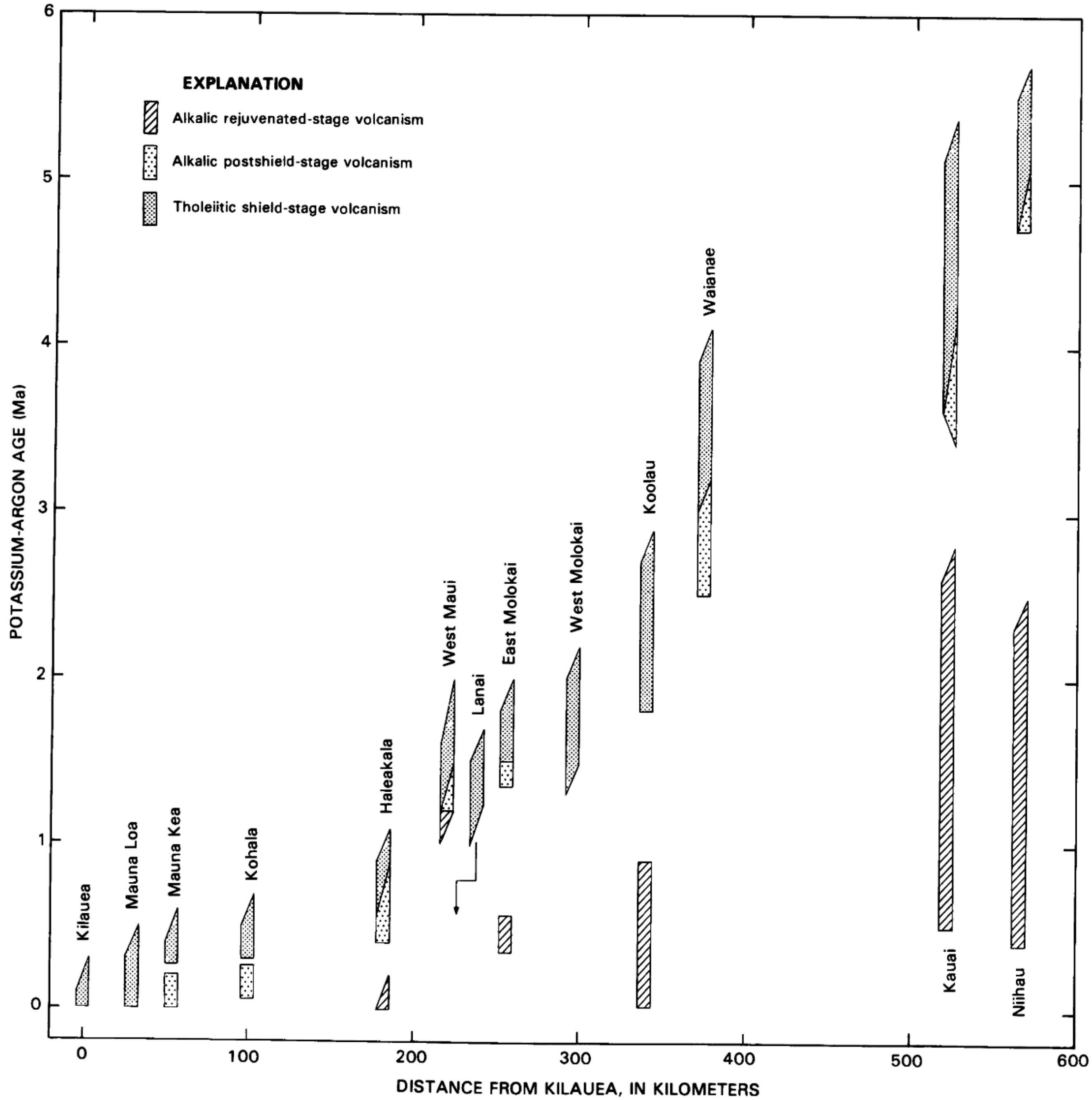

FIGURE 1.6.—Known durations of tholeiitic shield, alkalic postshield, and alkalic rejuvenated-stage volcanism for dated volcanoes of principal Hawaiian Islands. Angled lines indicate overlapping or uncertain ages or overlapping volcanism. Data from sources discussed in appendix 1.1. Data for Niihau and for Kauai (Koloa Volcanics) are from G.B. Dalrymple (unpublished data, 1985).

propagation on a small scale as a working hypothesis. It is unlikely that the cause of this nonlinear propagation, if real, will be known until more is learned about the hot-spot mechanism.

ERUPTION RATES ALONG THE CHAIN

The bathymetry of the chain as a whole is not well known, particularly for the western Hawaiian Ridge, and the 1970 charts for the North Pacific (Chase and others, 1970) and their 1973 derivative (Chase and others, 1973) are probably still the best published sources available. An updated bathymetric chart for the Emperor Seamounts (Clague and others, 1980) was based on the data used by Chase and others (1970) and additional geophysical profiles collected between 1970 and 1979; recently published bathymetry for much of the central part of the Emperor Seamounts (Smoot, 1982) is based on previously classified Navy multibeam data. The gross structure of the Emperor Seamounts is little changed in the later charts, but the shapes and locations of some individual volcanoes changed dramatically (fig. 1.7).

Bargar and Jackson (1974) compiled volume data along the chain and identified individual volcanic centers and their rift systems using the bathymetry of Chase and others (1970). From the more accurate multibeam data it is clear that many of the volcanic centers and rift zones identified by Bargar and Jackson are incorrect in detail. Because the number and general sizes of the volcanoes change little on the later charts, we have used Bargar and Jackson's volume estimates rather than engage in the laborious process of calculating new ones from the newer data. We suspect that the volumes based on multibeam data would vary relatively little from those of Bargar and Jackson.

The cumulative volume of the volcanoes is plotted in figure 1.8 against distance from Kilauea beginning at Tenchi Seamount, 500 km north of Suiko. It is clear that the volume of eruptive products per unit distance along the chain has not been constant over the past 70 million years.

We have calculated dV/dx, where V is volume and x is distance, for segments of the chain, which are summarized in table 1.6. Also listed are dx/dt, where t is time, calculated from the age relations along the chain, and the derived quantity dV/dt for segments along the chain. These calculations clearly show that the volumes erupted per unit distance along the chain and per unit time increase from the Emperor Seamounts to the Hawaiian Ridge and to the Hawaiian Islands. The present-day eruption rate for Kilauea alone, when compared to eruption rates along the Hawaiian Ridge and Emperor Seamounts, demonstrates that the Hawaiian hot spot is presently producing large volumes of lava at the greatest eruption rates in its known history. The average eruption rate from Hualalai to Kilauea is 5 times that for the islands as a whole and nearly 22 times the rate for the entire chain. The only section of the chain where volumes do not increase toward the present is the westernmost section of the Hawaiian Ridge, which formed immediately following the change in plate motion recorded as the Hawaiian-Emperor bend. This change in plate motion was followed by a virtual cessation of volcanic activity that lasted for nearly 10 m.y.

PETROLOGY OF THE HAWAIIAN-EMPEROR VOLCANIC CHAIN

EARLY WORK: LAVA SERIES AND DIFFERENTIATES

Early observers of Hawaiian eruptions rather uniformly agreed that the lava originated "in the bowels of the earth." As time progressed this view was expanded upon, but it was not until the 1950's and 1960's that the lava source and the processes generating the various lava types were discussed in detail. S. Powers (1920) proposed that nepheline basalt and trachyte were formed by differentiation of basaltic magma because of their occurrence late in the eruptive sequence. He wrote (S. Powers, 1920, p. 280): "Each volcano has arisen at an intersection in a fracture system in the earth's crust, has been fed from the same primal source, and has finally lost connection with that source. When this takes place differentiation may proceed in the magma chambers of large volcanoes and the extreme products of Hawaiian volcanism, nepheline basalt and trachyte, may appear either at the close of the main volcanism or in a later phase after extensive erosion."

This viewpoint, that there was a single primary Hawaiian magma from which the varieties of lava evolved by means of differentiation, was popular well into the 1960's. H. Powers (1935) expressed a similar view, although he showed that fractional crystallization alone could not explain the differentiation of Hawaiian basalt. Macdonald (1949) proposed that the sole primary Hawaiian lava was olivine basalt, although his calculated average included both alkalic and tholeiitic olivine basalt analyses (which he did not distinguish at that time). He also proposed that andesine andesite (hawaiite), oligoclase andesite (mugearite), and trachyte were successive differentiates from an olivine basalt parental lava. This view is now known to be incorrect because Macdonald's calculated olivine basalt was basically a tholeiitic basalt in composition. He further inferred that picritic basalt of the oceanite type (here termed picritic tholeiitic basalt) formed by the accumulation of olivine and that ankaramite was not an oceanite that simply accumulated clinopyroxene. This last idea is correct, ankaramite being alkalic in composition whereas picritic basalt of the oceanite type is tholeiitic. In order to differentiate ankaramite and nepheline basalt from the parental olivine basalt, Macdonald (1949) proposed that limestone assimilation and selective remelting (wall-rock assimilation) were important processes. He correctly inferred that the dunite xenoliths so common in alkalic lava from the postshield and rejuvenated stages formed by accumulation of olivine followed by recrystallization.

Tilley (1950) recognized that the bulk of the primitive shields was made of tholeiitic basalt and that alkalic rocks erupted only during the declining stages of activity. He proposed that alkalic olivine basalt was derived from tholeiitic basalt by crystal fractionation. H. Powers (1935) had, however, earlier noted that primitive lava was silica saturated, whereas the late differentiated lava was silica undersaturated; he argued that these lavas could therefore not be simply related to one another by crystal fractionation.

New concepts important to understanding the origin of Hawaiian lava were introduced by H. Powers (1955). He clearly established that the abundant rocks termed olivine basalt in the shields are silica saturated and noted that they are compositionally distinct for individual volcanoes. He proposed the concept of magma batches to account for the subtle differences between the lava of different shields. He also reiterated that olivine basalt erupted during the declining stages of activity is silica undersaturated. His discussion of fractionation trends for tholeiitic and alkalic basalt is nearly identical to present-day views. He further recognized that earthquakes associated with volcanic activity gave a minimum depth of magma generation which he took to be 48–56 km (30–35 miles). His estimates of the source rocks that could be melted to produce basalt and of the causes of melting provided a framework for experimental research for many years. In particular, he noted that basalt could be generated at depth by wholesale melting of rocks of basaltic composition or by partial melting of peridotite. The models of melting he considered all assumed that it was caused by an increase in temperature. He dismissed exothermic nuclear processes

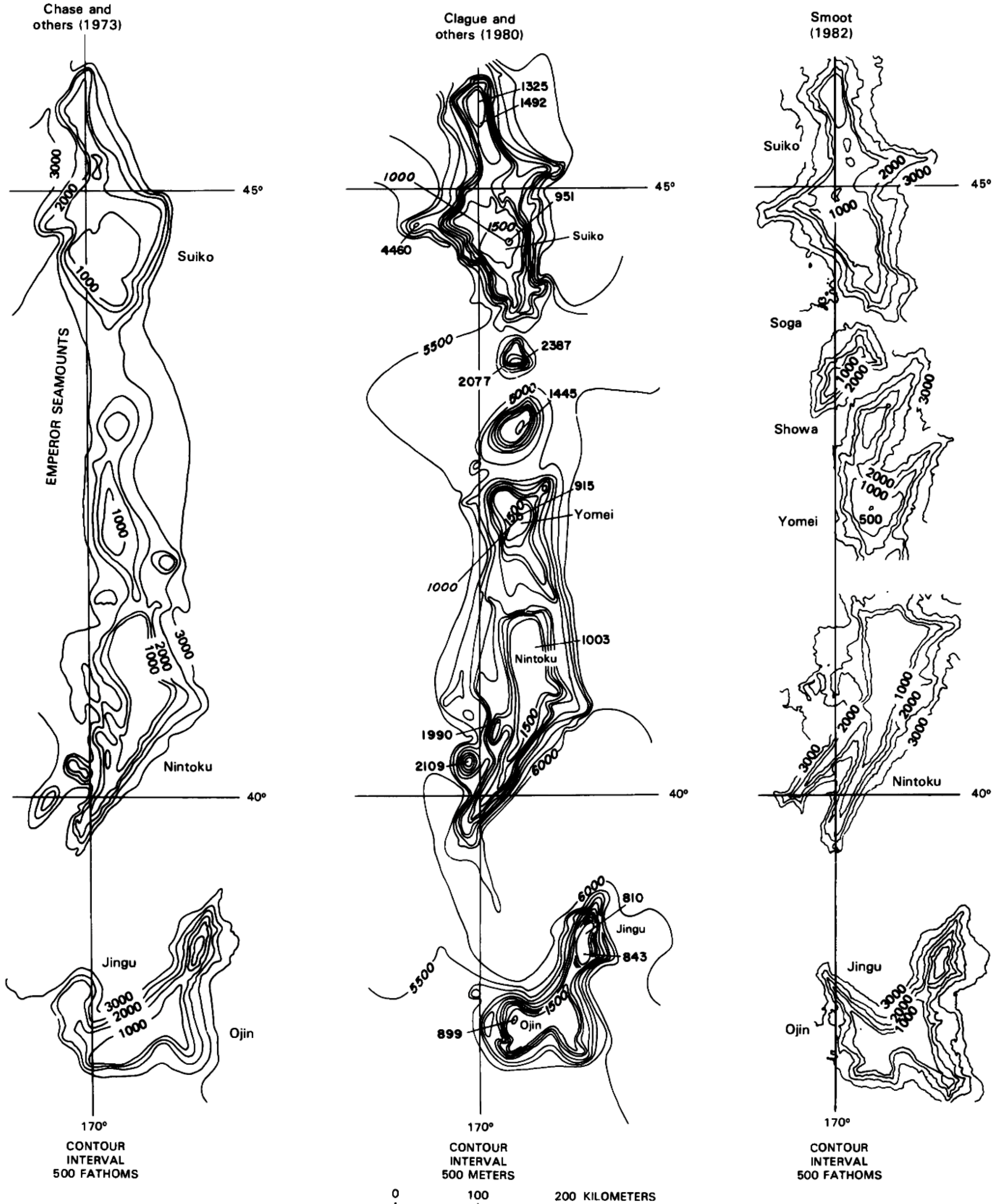

FIGURE 1.7.—Comparison of bathymetry of central Emperor Seamounts from Chase and others (1973; left), Clague and others (1980b; center) and Smoot (1982; right). General size and shape of seamounts were fairly well mapped by bathymetric sounding (left and center), but multibeam bathymetry (right) adds wealth of detail. Contour intervals are 500 fathoms for left and right figures and 500 m for center figure.

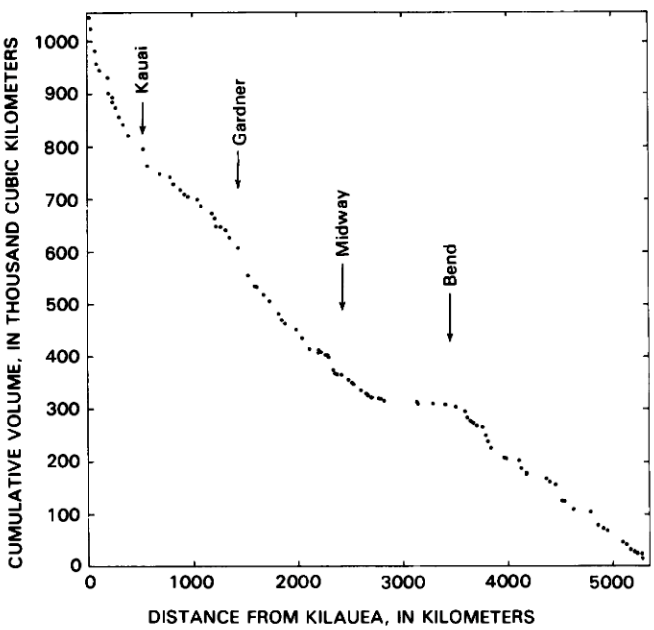

FIGURE 1.8.—Cumulative volcanic volume along Hawaiian-Emperor Chain plotted as a function of distance from Kilauea (along trend of chain). Average volume increment rate (dV/dx) for Emperor Chain is 0.16×10^3 km³/km. Just east (left) of the bend there is a segment of very low volcanic productivity, in which only 0.02×10^3 km³/km was erupted. Remainder of submarine portion of the Hawaiian Chain has an average dV/dx of 0.20×10^3 km³/km. In Hawaiian Islands section from Kauai to Haleakala can be fit by 0.40×10^3 km³/km and that from Haleakala to Kilauea by 1.1×10^3 km³/km.

in the crust because Hawaiian lava is not enriched in U, Th, or K. He emphasized convection from hotter regions deep in the mantle and friction produced by dynamic processes (he proposed tides). His convection model is very similar to later plume models and his friction model to at least a part of later thermal-feedback models, both of which are discussed more thoroughly later in this paper.

Kuno and others (1957) clearly demonstrated that closed-system crystal fractionation of tholeiitic basalt led to generation of granophyre in the differentiation of thick lava bodies. They related picritic tholeiitic basalt, olivine tholeiitic basalt, tholeiitic basalt, and granophyre as one differentiation sequence and alkali olivine basalt, picrite basalt of ankaramite type (here called ankaramite), hawaiite, mugearite, and trachyte as another differentiation sequence. They presented trace-element data, which they used to evaluate the proposed fractionation trends. Most importantly, they dismissed any fractionation relation between tholeiitic and alkalic basalt and proposed that they originated independently "through melting of the earth's material under different sets of physical conditions" (Kuno and others, 1957, p. 214). They argued that both lava types were generated by partial melting of peridotite, but that alkalic basalt was derived at greater depth. Their discussion implies that the source rocks were homogeneous and that only the physical conditions of melting varied. In addition, the alkali-silica diagram, widely used to distinguish between tholeiitic and alkalic lava, was first presented in this paper.

Eaton and Murata (1960) published a detailed study locating earthquake epicenters as deep as 60 km beneath Kilauea using basically the same data cited by Powers (1955). These observations establish a minimum depth of magma generation for tholeiitic shield lava of greater than 60 km, although they still allow for selective melting (wall-rock reaction) at shallower depths (Macdonald, 1968).

Macdonald and Katsura (1962, 1964) established the tholeiitic character of the early lava on Kohala and West Maui, where the known late-stage lava is alkalic. They also demonstrated that the subtle variations in composition of the tholeiitic basalt from different volcanoes were unrelated to the type of alkalic cap that followed (Kohala or Haleakala type). In contrast to the interpretations of Tilley (1950), H. Powers (1955), and Kuno and others (1957), Macdonald and Katsura (1964) proposed that alkalic basalt formed by differentiation of tholeiitic basalt. They cited as evidence the interbedded sequence of tholeiitic and alkalic flows, including some of transitional composition, that occur in Haleakala-

TABLE 1.6.—*Eruptive rates along the Hawaiian-Emperor Chain*

[Volume/distance data from figure 1.8; propagation rate data from table 1.5, except for Kilauea to Hualalai; recent data for Kilauea from Dzurisin and others (1984), based on combined eruption-intrusion rate]

Segment	Volume/distance, dV/dx (10^3 km³/km)	Propagation rate, dx/dt (km/m.y.)	Eruption rate dV/dt (10^3 km³/m.y.)
Kilauea (1956-1983)	---	---	86
Kilauea to Hualalai	1,500	250	290
Hualalai to Waianae	400	101	40
Hawaiian Islands (0-5.5 Ma)	---	---	56
Waianae to Gardner Pinnacles	190	101	19
Gardner Pinnacles to volcano 57	200	68	14
Volcano 57 to Hawaiian-Emperor bend	20	68	1
Emperor Seamounts	160	72	12
Average for entire chain	---	---	13

type volcanoes between the tholeiitic shield lava and the later alkalic postshield-stage lava and proposed that volatile transfer might be an important differentiation process. They discussed thoroughly the fractionation sequences of both alkalic and tholeiitic lava and related the Mauna Kuwale rhyodacite to tholeiitic lava of the Waianae Range.

The years 1964–68 produced new insights from a variety of studies. The first isotopic data from Hawaiian lava (Lessing and Catanzaro, 1964; Hamilton, 1965; Powell and DeLong, 1966; Tatsumoto, 1966) clearly demonstrated that the source rocks for Hawaiian lava were heterogeneous. Experimental studies at high pressure and temperature (for example, O'Hara, 1965; Green and Ringwood, 1967) added new data bearing on the mineralogy of potential source rocks and the physical conditions of melting. At the same time, trace-element data began to be used to evaluate Hawaiian petrogenetic processes (Schilling and Winchester, 1966; Gast, 1968). Studies of Hawaiian xenoliths (White, 1966) added to the abundant new data being used to evaluate the petrogenesis of Hawaiian lava and the nature of the source rocks. This period marks a transition from relatively qualitative models of petrogenesis to modern quantitive modeling and testing of basalt petrogenesis. To a great degree present day petrogenic models are based on these same types of data and similar quantitative modeling techniques. However, because modern isotopic and trace-element data are more accurate and precise, the models proposed are more refined and complex.

Macdonald (1968) followed Green and Ringwood (1967) and proposed that the tholeiitic, alkalic, and nephelinitic lava types were derived from a single parent magma of olivine tholeiitic basalt composition. He proposed that the compositional variations reflect the depth at which fractional crystallization occurred with tholeiitic basalt fractionated at shallow depths, alkalic basalt at moderate depths, and nephelinitic lava at depths of several tens of kilometers. He further argued that the primary magma is olivine tholeiitic basalt rather than tholeiitic basalt because most lava has lost olivine, which accumulated within the magma chambers to form the high-density masses discovered by gravity surveys (summary and references in Jackson and others, 1972). Macdonald (1968) noted that the highest temperature lava from the Kilauea Iki eruption contained 27–30 percent olivine and proposed that this closely approximated primary magma. Wright (1973) calculated the bulk composition of the same eruption and proposed it as a representative parental (but not necessarily primary) composition for Kilauea tholeiitic basalt. Macdonald (1968) presented quantitative models showing that ankaramite is alkalic basalt plus olivine and clinopyroxene and that hawaiite is alkalic basalt minus olivine, clinopyroxene, plagioclase, and magnetite. This type of mass-balance approach was later refined by Wright (1971) and Wright and Fiske (1971) to demonstrate the roles of fractionation and hybridization in generating basalt at Kilauea and Mauna Loa. The debate about whether primary tholeiitic magma is olivine rich or olivine poor continues today (see Wright and Helz, chapter 23; Wright, 1984; Budahn and Schmitt, 1984). Extreme compositions are liquids with 20 percent MgO (Wright, 1984) and average tholeiitic basalt with 9 percent MgO (Powers, 1955).

Macdonald (1968) also calculated average compositions of Hawaiian lava from the different eruptive stages. His average compositions, recalculated on a dry-reduced normalized basis, are presented in table 1.7. These averages clearly show that lava of the tholeiitic shield stage is silica saturated and that of alkalic postshield and rejuvenated stages is silica undersaturated. The presence of normative hypersthene in the mugearite, benmoreite, and trachyte, which are derived from undersaturated alkalic basalt, reflects fractionation of Fe-Ti oxides, which enriches the residual melt in silica. Rejuvenated-stage alkalic basalt contains greater than 5 percent normative nepheline, and average nephelinite and nepheline melilitite contain normative leucite. Alkalic basalt of the preshield and rejuvenated stages are similar in composition. The preshield Loihi Seamount averages are calculated from Moore and others (1982), Frey and Clague (1983), and D.A. Clague (unpub. data, 1985).

XENOLITH DISTRIBUTIONS

In a detailed analysis of the xenolith populations in Hawaiian lava, Jackson (1968) subdivided the xenoliths into dikes and sills, cumulates, and metamorphic rocks. His breakdown of the relative abundances and types of xenoliths is given in table 1.8, modified to include xenoliths found in Loihi Seamount alkalic preshield lava. Jackson noted that only those xenoliths with metamorphic textures could represent either mantle source rocks or mantle residua left after partial melting. The dikes, sills, and some, but not all, of the cumulate rocks are cognate or from shallow depths. The cumulate xenoliths from Hualalai Volcano appear to come from a variety of sources, including cumulates formed as part of oceanic crustal layer 3, cumulates of tholeiitic Hawaiian shield lava, and cumulates of lava from the alkalic postshield stage (D.A. Clague, unpub. data, 1985). The single cumulate xenolith from Loihi Seamount presumably represents a cumulate of ocean crustal layer 3.

Jackson and Wright (1971) proposed that the abundant dunite xenoliths in the Honolulu Volcanics represent residue left after melting the mantle to form Koolau shield tholeiite, an interpretation with which we disagree. Jackson and Wright (1971) inferred that the garnet lherzolite and lherzolite found only in alkalic lava from the rejuvenated stage were potential mantle source rocks. The difference in xenolith populations for the three alkalic eruptive stages is striking since only the lava from the preshield stage and rejuvenated stage contain xenoliths that formed at depths greater than about 20 km. Although it would be useful if these xenolith populations reflected the mantle through which the lava ascends, it seems more likely that they reflect the development of shallow magma storage reservoirs, which act as hydraulic filters and remove xenoliths carried up from greater depths in much the same way as lakes remove sediment from rivers.

Lava stored in a shallow magma reservoir, either within a few kilometers of the surface or at the base of the oceanic crust, lose any xenoliths they may have acquired during ascent and from this point can only entrain xenoliths that occur at shallower levels in the volcanic system. However, lava of the preshield and rejuvenated stages erupts in small volumes at infrequent intervals and probably no shallow magma storage reservoirs exist. During the shield stage, tholeiitic lava erupts in large volumes at frequent intervals from a

TABLE 1.7.—*Average compositions and norms of major Hawaiian lava types*

[All figures in weight percent; -- not present. Normative components calculated with original FeO/Fe_2O_3 ratios. See Macdonald (1968) for remainder of norms for all but the alkalic preshield stage. Data for alkalic preshield stage from Frey and Clague (1983) and D.A. Clague (unpub. data, 1985)]

Stage	Alkalic preshield				Tholeiitic shield		Alkalic postshield						Alkalic rejuvenated			
Lava type	Tholeiite and olivine tholeiitic basalt	Transitional basalt	Alkalic basalt	Basanite	Picritic tholeiitic basalt	Tholeiitic and olivine tholeiitic basalt	Ankaramite	Alkalic basalt	Hawaiite	Mugearite	Benmoreite	Trachyte	Alkalic basalt	Basanite	Nephelinite	Melilitite
Chemical composition																
SiO_2	48.4	48.3	45.6	43.5	46.7	50.0	44.6	45.9	48.6	52.1	58.3	62.8	45.2	44.6	40.6	37.8
Al_2O_3	12.2	13.6	11.6	11.1	8.5	14.1	12.2	14.9	16.1	17.1	18.0	18.3	12.8	12.8	11.6	11.2
FeO*	12.0	12.3	12.4	13.6	12.1	11.3	12.6	13.0	12.2	10.0	7.5	4.5	12.4	12.5	13.3	14.6
MgO	11.2	7.5	13.5	12.2	20.9	8.5	13.1	7.9	4.9	3.3	1.6	.4	11.5	11.3	12.4	13.0
CaO	11.0	11.4	10.4	11.4	7.4	10.4	11.6	10.6	8.1	6.2	3.6	1.2	11.5	10.7	13.1	14.1
Na_2O	2.1	2.7	2.5	3.2	1.6	2.2	1.9	3.0	4.3	5.5	6.0	7.5	2.7	3.6	3.9	4.2
K_2O	.4	.7	.8	1.3	.3	.4	.7	1.0	1.5	2.1	2.9	4.3	.9	1.0	1.2	1.0
TiO_2	2.3	3.1	2.7	3.1	2.0	2.5	2.7	3.0	3.4	2.4	1.2	.5	2.3	2.6	2.9	2.9
P_2O_5	.2	.3	.3	.5	.2	.3	.3	.4	.7	1.1	.7	.2	.5	.5	.9	1.1
MnO	.2	.2	.2	.2	.2	.2	.2	.2	.2	0.2	.2	.2	.2	.2	.2	.1
Normative composition																
Q	--	--	--	--	--	2.2	--	--	--	--	--	--	--	--	--	--
Ne	--	--	2.7	11.8	--	--	2.6	2.6	0.3	--	--	--	6.0	10.5	17.3	18.7
Lc	--	--	--	--	--	--	--	--	--	--	--	--	--	--	5.7	4.8
Hy	14.3	6.3	--	--	15.9	21.5	--	--	--	2.7	4.0	.4	--	--	--	--
Ol	10.5	5.9	23.1	17.7	29.5	--	22.2	13.2	6.7	2.1	--	--	19.3	17.9	14.9	20.2

shallow magma storage reservoir and perhaps a deeper staging zone (see review by Decker, chapter 42). During the alkalic postshield stage, lava erupts in small volumes at infrequent intervals, though in larger volumes and at more frequent intervals than during the alkalic preshield or rejuvenated stages. During this stage, lava apparently resides in reservoirs below the base of the oceanic crust (Clague and others, 1981) for time periods sufficient for the dense peridotite xenoliths to settle out. Tholeiitic lava at Kilauea passes through two such filters, one an intermediate-depth (20–30 km) staging area, the second a well-defined and complex shallow reservoir system 3–7 km beneath the surface. After passing through these filters, the lava can only incorporate as xenoliths the wall rocks occurring at depths shallower than the shallowest reservoir (dikes, sills, and olivine cumulates). Alkalic lava of the postshield stage contains abundant xenoliths of dunite, which have CO_2 inclusions that were trapped at depths of at least 15 km, and cumulate xenoliths of rocks from oceanic crustal layer 3 (Roedder, 1965; D.A. Clague, unpub. data, 1985). These observations suggest that in the postshield stage any shallow magma chamber of the shield stage no longer exists but an intermediate staging area at 20–30 km, similar to that beneath Kilauea, acts as an effective filter that removes any lherzolite or garnet peridotite xenoliths. Lava may fractionate in this zone and, upon movement to the surface, entrain xenoliths of ocean crust rocks and cumulates formed in earlier volcanic stages at shallow depth. The presence of xenoliths that originate at great depth in alkalic lava of both the preshield and rejuvenated stages implies that neither shallow nor intermediate staging area acts as an effective filter in these stages. The near-primary character of the host lava also indicates that the lava was not stored at shallow depths but rather moved from its source region to the surface in short time periods (Clague and Frey, 1982).

This analysis leads us back to Jackson's (1968) conclusion that only the xenoliths with metamorphic textures could possibly be mantle source rocks or residua. We conclude that only the lherzolite and garnet peridotite xenoliths represent mantle rocks from below the magma storage zone that appears to have existed beneath Hawaiian shield volcanoes at depths of 20–30 km. The dunite and wehrlite xenoliths are either deformed cumulates formed during

TABLE 1.8.—*Distribution of Hawaiian xenolith types*

[Data from Jackson (1968) except the alkalic preshield stage which is from D.A. Clague (unpub. data, 1985)]

Eruptive stage	Lava type	Xenolith type (percent): Dike rocks and vein fillings	Cumulates	Metamorphic rocks
Alkalic rejuvenated	alkalic basalt basanite nephelinite nepheline melilitite	(<1)	(1) olivine cumulates dominant	(99) dunite ≅ wehrlite ≅ lherzolite> harzburgite, garnet peridotite locally
Alkalic postshield	alkalic basalt	(3) veins dominant	(35) olivine cumulates> pyroxene cumulates	(62) dunite>> wehrlite
	ankaramite	(14) dikes and sills dominant	(53) pyroxene cumulates> olivine cumulates	(33) dunite>> wehrlite
	hawaiite mugeartie trachyte	(14) dikes and sills dominant	(57) pyroxene cumulates> olivine cumulates	(29) dunite>> wehrlite
Tholeiitic shield	tholeiitic basalt	(75) dikes and sills dominant	(25) olivine cumulates dominant	none
Alkalic preshield (Loihi)	alkalic basalt basanite	none	(1) olivine cumulates	(99) dunite>> lherzolite

earlier stages in the volcano's growth or cumulates formed during formation of oceanic crust (see Sen, 1983, 1985; Kurz and others, 1983; Sen and Presnall, 1985).

The remaining xenoliths of spinel lherzolite, rare harzburgite, and rare garnet peridotite that occur in alkalic lava of the rejuvenated stage and even more rarely in alkalic lava of the preshield stage therefore are the only xenoliths of deeper mantle material. Spinel lherzolite xenoliths have many characteristics that imply a close genetic relationship to midocean-ridge basalt; however, both the Sr-isotopic and rare-earth data indicate that these xenoliths have been enriched by mixing between residua left after formation of midocean-ridge basalt and an enriched magma or vapor (Frey 1980, 1984; Wright, 1984; Frey and Roden, in press). These xenoliths probably represent depleted oceanic lithosphere modified by processes related to Hawaiian magmatism.

The final group of xenoliths consists of pyroxenite, websterite, and garnet-bearing pyroxenite and websterite. These occur in only a few vents of the alkalic rejuvenated stage in the Honolulu Volcanics (Jackson and Wright, 1970) and on Kaula Island (Garcia, Frey, and Grooms, in press). These rocks occur both as separate xenoliths and as layers in xenoliths. Some of these xenoliths have been called garnet lherzolite (Jackson and Wright, 1970), but they are not merely a higher pressure assemblage of spinel lherzolite because their bulk compositions are distinct (Jackson and Wright, 1970; Sen, 1983). The iron-rich olivine in all the xenoliths of this group led Sen (1983) to argue that they represent neither source rocks nor residua related to Hawaiian lava. Frey (1980, 1984) has argued that they may represent crystal accumulates from alkaline Hawaiian magma. We conclude that none of the xenoliths found in Hawaiian lava represent mantle source rocks or residua related to Hawaiian volcanism. They do, however, provide insight into the conduit systems through which much of this lava passed.

Jackson and Wright (1970) demonstrated that xenoliths in the Honolulu Volcanics were compositionally zoned in a geographic sense with respect to the Koolau caldera; abundant dunite near the caldera grades into lherzolite and finally garnet-bearing websterite and pyroxenite away from the caldera. Jackson and Wright (1970) combined these observations with experimental petrologic and geophysical data to construct a cross section through the mantle and crust beneath Oahu (fig. 1.9). Their cross section emphasizes the mineralogic and compositional heterogeneity of the mantle beneath Hawaiian volcanoes. However, the origins of many of the rock types are now thought to be different from those proposed by Jackson and Wright (1970). A more recent model by Sen (1983) shows plagioclase lherzolite beneath the oceanic crust to a depth of about 30 km (defined by the limit of plagioclase stability), spinel lherzolite from 30 to nearly 50 km, and garnet lherzolite below about 50 km. The zone beneath the volcanoes includes cumulate dunite to depths

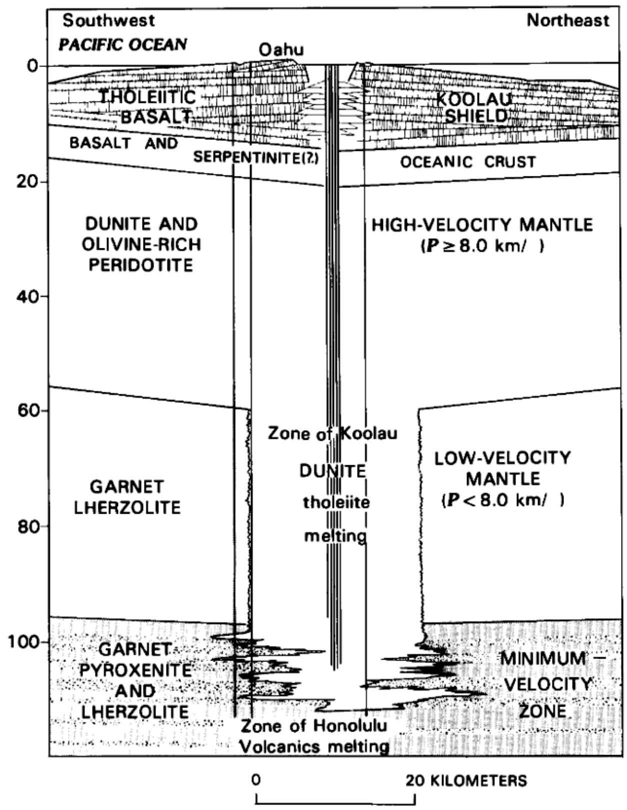

FIGURE 1.9.—Cross section beneath Oahu from Jackson and Wright (1970). Dunite zone inferred to be mantle residue left behind from partial melting that forms Koolau tholeiitic shield lava. Configuration of rock types and their mode of origin are far different from those shown in figure 1.10.

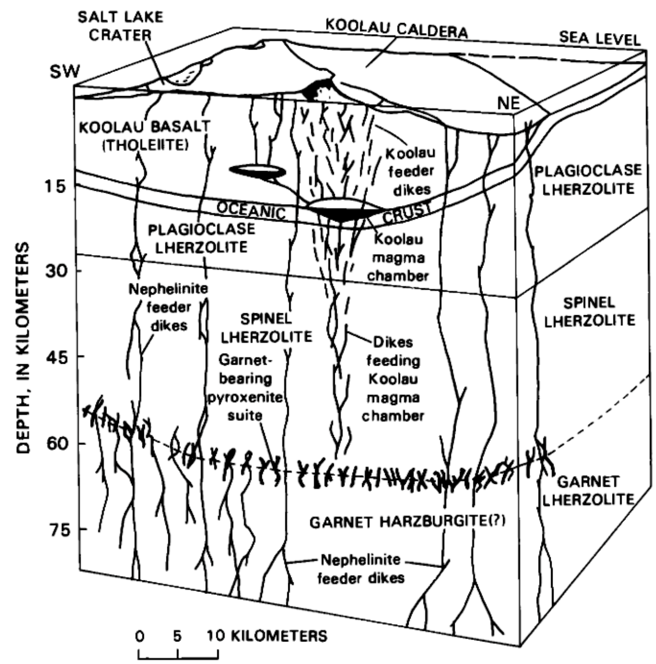

FIGURE 1.10.—Cross section beneath Oahu from Sen (1983) showing configuration of various mantle source and residual rocks brought to surface as xenoliths of abundant dunite, lherzolite, and garnet peridotite by rejuvenated-stage Honolulu Volcanics.

of about 15 km (fig. 1.10). The areal distribution of xenoliths observed by Jackson and Wright (1970) reflects passage of the Honolulu Volcanics lava through the zone of dunite cumulates.

PETROLOGY OF LAVAS ALONG THE VOLCANIC CHAIN

In the middle to late 1970's new studies added data to the already complex data array on Hawaiian volcanoes. Studies on lava recovered from the older submarine portions of the chain (Clague, 1974; Dalrymple and others, 1974, 1977, 1981; Clague and others, 1975; Dalrymple and Clague, 1976; Kirkpatrick and others, 1980; Clague and Frey, 1980; Lanphere and others, 1980; Dalrymple and Garcia, 1980; Garcia, Grooms, and Naughton, in press) clearly demonstrate that the volcanoes of the entire Hawaiian-Emperor volcanic chain erupted tholeiitic basalt and picritic tholeiitic basalt similar to those of the shield stage in the Hawaiian Islands. In addition, alkalic lava similar to that erupted during the postshield stage in the Hawaiian Islands, including hawaiite, mugearite and trachyte, is commonly recovered from the older volcanoes. In the drill holes on Ojin and Suiko, tholeiitic lava occurs below alkalic lava, as in the Hawaiian Islands. Some samples are chemically and mineralogically similar to alkalic rejuvenated-stage lava from the Hawaiian Islands (table 1.2). The identification of which dredged or drilled lava samples erupted during which eruptive stage relies on comparison of the major-element compositions to those of the various Hawaiian lava types (table 1.7) in conjunction with trace-element ratios (Clague and Beeson, 1980; Clague and others, 1980c; Frey and Clague, 1983) and mineral compositions (Keil and others, 1972; Fodor and others, 1975; Clague and others, 1980a). In particular, we have found that the composition of groundmass pyroxene and the K/Ba and P_2O_5/Zr ratios seem to separate alkalic basalt of the postshield and rejuvenated stages. Rejuvenated-stage alkalic basalt has lower K/Ba and higher P_2O_5/Zr ratios and pyroxene with more calcic compositions and higher concentrations of Na, Ti, and Al than postshield-stage alkalic basalt. Lava samples recovered from the volcanoes west of the principal Hawaiian Islands are discussed in appendix 1.1 and summarized in table 1.9.

Several conclusions may be drawn from these studies of samples from along the Hawaiian-Emperor Chain. The first conclusion is that the Hawaiian hot spot has produced very similar lava types in the same eruptive sequence for at least the last 65 m.y. (table 1.10). Samples from the same eruptive stage are similar to one another in both major- and trace-element compositions, including rare-earth elements (Clague and Frey, 1980; Frey and Roden, in press). Isotopic studies indicate, however, that small systematic changes occur over time (Lanphere and others, 1980; Unruh and

TABLE 1.9.—*Rock types and inferred volcanic stages represented along the Hawaiian-Emperor Chain*

[X, present; --, not present or not known; (T), transitional; volcano numbers from Bargar and Jackson (1974)]

Volcano		Shield stage			Alkalic postshield stage						Alkalic rejuvenated stage				
Number	Name	Tholeiitic basalt	Picritic tholeiitic basalt	Rhyodacite	Alkalic basalt	Ankaramite	Hawaiite	Mugearite	Trachyte	Phonolite	Alkalic basalt	Basanite	Nephelinite	Melilitite	Tephrite
1	Kilauea	X	X	--	--	--	--	--	--	--	--	--	--	--	--
2	Mauna Loa	X	X	--	--	--	--	--	--	--	--	--	--	--	--
4	Hualalai	X	X	--	X	X	--	--	X	--	--	--	--	--	--
3	Mauna Kea	X	X	--	X	X	X	--	--	--	--	--	--	--	--
5	Kohala	X	--	--	--	--	--	X	X	--	--	--	--	--	--
6	Haleakala	X	--	--	X	X	X	X	--	--	X	X	--	--	--
7	Kahoolowe	X	X	--	--	X	--	--	--	--	X?	--	--	--	--
8	West Maui	X	X	--	X	X	--	X	X	--	--	X	--	--	--
10	East Molokai	X	X	--	X	X	X	X	X	--	X	X	--	--	--
9	Lanai	X	X	--	--	--	--	--	--	--	--	--	--	--	--
11	West Molokai	X	X	--	X	--	X	--	--	--	--	--	--	--	--
12	Koolau	X	X	--	--	--	--	--	--	--	X	X	X	X	--
13	Waianae	X	X	X	--	--	X	--	--	--	--	--	--	--	--
14	Kauai	X	X	--	--	--	X	--	--	--	X	X	X	X	--
15	Niihau	X	X	--	X	--	--	--	--	--	X	--	--	--	--
15A	Kaula	X	--	--	--	--	--	--	--	X	--	X	--	--	--
17	Nihoa	X	X	--	--	--	--	--	--	--	--	--	--	--	--
19	Unnamed Seamount	--	X	--	--	--	--	--	--	--	--	--	--	--	--
20	Unnamed Seamount	X	--	--	--	--	--	--	--	--	--	X	--	--	--
21	Unnamed Seamount	--	X	--	--	--	--	--	--	--	--	--	--	--	--
23	Necker	--	X	X?	--	--	X	--	--	--	--	--	--	--	X?
26	La Perouse Pinnacles	--	X	--	--	--	--	--	--	--	--	--	--	--	--
30	Gardner Pinnacles	--	X	--	X	--	--	--	--	--	--	--	--	--	--
38	Brooks Bank	X	--	--	--	--	X	--	--	--	--	--	--	--	--
29	St. Rogatein Bank	--	--	--	--	--	X	--	--	--	--	--	--	--	--
36	Laysan	--	--	--	--	--	X	X	--	--	--	--	--	--	--
37	Northampton Bank	X	X	--	--	--	--	--	--	--	--	--	--	--	--
39	Pioneer Bank	X	--	--	--	--	--	--	--	--	--	--	--	--	--
50	Pearl and Hermes Reef	--	--	--	X	--	X	--	--	X	--	--	--	--	--
51	Ladd Bank	--	--	--	--	--	--	--	--	--	--	--	X	--	--
52	Midway Island	X	--	--	--	--	X	X	--	--	--	--	--	--	--
53	Nero Bank	--	X	--	--	--	--	--	--	--	--	--	--	--	--
57	Unnamed Seamount	--	--	--	X	--	--	--	--	--	--	--	--	--	--
63	Unnamed Seamount	--	--	--	--	--	--	--	--	--	--	--	--	--	[1]X
65	Colahan Seamount	(T)	--	--	--	--	--	--	--	--	--	--	--	--	[1]X
65A	Abbott Seamount	(T)	--	--	--	--	--	--	--	--	--	--	--	--	--
67	Daikakiyi Seamount	X	--	--	X	--	--	--	--	--	--	--	--	--	--
69	Yuryaku Seamount	--	--	--	X	--	--	--	--	--	--	--	--	--	--
72	Kimmei Seamount	--	--	--	X	--	--	--	--	--	--	--	--	--	--
74	Koko Seamount (southeast)	X	--	--	X	--	X	X	X	X	--	--	--	--	--
76	Koko Seamount (northwest)	X	--	--	--	--	--	--	--	--	--	--	--	--	--
81	Ojin Seamount	X	--	--	--	--	X	--	--	--	--	--	--	--	--
89	Jingu Seamount	--	--	--	--	--	X	X	--	--	--	--	--	--	--
86	Nintoku Seamount	--	--	--	X	--	--	--	--	--	--	--	--	--	--
90	Suiko Seamount (southern)	--	--	--	--	--	--	X?	--	--	--	--	--	--	--
91	Suiko Seamount (central)	X	X	--	X	--	--	--	--	--	--	--	--	--	--
108	Meiji Seamount	X	--	--	--	--	--	--	--	--	--	--	--	--	--

[1]Dredges from unnamed seamount (63) and Colahan Seamount recovered ankaramite, tephrite, and amphibole-bearing hawaiite that are probably rejuvenated stage lava.

TABLE 1.10.—*Composition of tholeiitic basalt from volcanoes of the Hawaiian-Emperor Chain*

[All figures in weight percent, dry reduced normalized average analyses; olivine added or subtracted so that Mg/(Mg+0.85 Fe)=0.70; --, not analyzed; P_2O_5 value in parentheses is high due to marine phosphatization]

Volcano	Kilauea	Mauna Loa	Mauna Kea	Hualalai	Kohala	East Maui	West Maui	Lanai	East Molokai	West Molokai	Koolau	Waianae (upper)	Waianae (lower)
SiO_2 --	48.9	50.5	46.4	49.4	48.2	49.7	47.3	49.0	46.4	49.2	51.5	46.9	47.9
Al_2O_3 -	12.1	12.2	12.8	11.8	13.5	13.3	12.6	12.8	12.9	11.8	13.2	13.4	14.1
FeO ---	11.4	11.0	12.3	11.9	11.8	10.9	12.2	11.0	12.4	11.9	10.4	11.9	11.2
MgO ---	12.7	12.3	13.6	13.4	13.1	12.1	13.7	12.3	13.8	13.2	11.7	13.2	12.5
CaO ---	9.7	9.2	9.9	9.2	9.4	9.6	9.8	8.8	9.7	8.9	8.3	9.1	9.6
Na_2O --	1.99	1.98	1.82	1.84	1.62	1.76	1.66	2.13	1.89	2.42	2.47	2.07	1.86
K_2O ---	.44	.36	.28	.28	.10	.33	.23	.11	.20	.18	.27	.53	.26
TiO_2 --	2.34	1.85	2.40	1.75	1.93	2.01	2.14	1.66	2.22	2.08	1.68	2.46	2.20
P_2O_5 --	.22	.21	.23	.16	.20	.13	.20	.18	.25	--	.22	.35	.25
MnO ---	.17	.16	.17	.15	.18	.18	.17	.14	.18	.19	.15	.15	.17
Volcano	Kauai	Niihau	Nihoa	Unnamed (20)	Unnamed (21)	Necker Island	La Perouse Pinnacles	Gardner Pinnacles	Northampton Bank	Pioneer	Daikakiji	Suiko high TiO_2	Suiko low TiO_2
SiO_2 --	48.4	48.4	47.4	46.9	47.1	47.15	47.5	47.6	48.7	48.2	49.7	47.6	47.9
Al_2O_3 -	12.5	12.1	11.5	11.2	11.3	11.1	11.7	12.5	11.4	11.2	11.5	12.7	13.0
FeO ---	11.9	12.1	12.4	12.6	12.4	12.8	11.8	11.3	11.7	11.9	11.5	12.3	11.9
MgO ---	13.3	13.6	13.8	14.0	13.8	14.1	13.1	12.6	13.0	13.3	12.9	13.7	13.2
CaO ---	9.2	8.1	9.4	9.6	9.2	9.8	10.7	10.3	10.1	9.8	9.05	8.7	9.3
Na_2O --	1.93	2.28	1.77	2.06	2.25	1.59	1.82	2.06	2.00	2.20	2.08	2.26	2.26
K_2O ---	.30	.49	.26	.24	.64	.38	.39	.16	.32	.67	.43	.28	.14
TiO_2 --	2.06	2.49	2.59	2.97	2.72	2.47	2.44	1.98	2.26	2.36	2.32	2.10	1.86
P_2O_5 --	.23	.31	(.70)	.28	.35	.36	.34	.26	.23	.17	.36	.22	.16
MnO ---	.17	.16	.15	.15	.15	.17	.20	.16	.16	.14	.16	.18	.18

others, 1983). The lower $^{87}Sr/^{86}Sr$ ratios and higher $^{143}Nd/^{144}Nd$ ratios of lava from the central Emperor Seamounts compared to those of lava from the Hawaiian Islands and Ridge imply that tholeiitic lava erupted 65 m.y. ago was derived from a more depleted source than that erupting today. Lanphere and others (1980) correlated this observation with the data shown in figure 1.3 to suggest that the chemistry of Hawaiian tholeiitic lava has varied as a function of the age and thickness of the oceanic lithosphere beneath each volcano when it was constructed. The correlation suggests that the oceanic lithosphere forms at least part of the source material for Hawaiian tholeiitic magma, or that the magma partially reequilibrates with the oceanic lithosphere. Wright (1984) proposes that Hawaiian magma originates from oceanic lithosphere converted to asthenosphere.

STRATIGRAPHIC STUDIES IN THE HAWAIIAN ISLANDS

Studies of stratigraphically controlled samples (Beeson, 1976; Clague and Beeson, 1980; Chen and Frey, 1983; Clague and others, 1983; Feigenson, 1984; Lanphere and Frey, 1985) have shown that major-element, trace-element, and isotopic ratios change systematically as a function of time at some Hawaiian volcanoes. These observations are not universal (Stille and others, 1983; Frey and others, 1984) in as much as Waianae and Mauna Kea erupted isotopically similar lava during the tholeiitic shield and alkalic postshield stages. Chen and Frey (1983) observed systematic stratigraphic trends in $^{87}Sr/^{86}Sr$, Rb/Sr, $^{143}Nd/^{144}Nd$, and Sm/Nd ratios in samples from East Maui Volcano. Their data indicate that the tholeiitic lava had higher $^{87}Sr/^{86}Sr$ and Sm/Nd ratios and lower ^{143}/Nd ^{144}Nd and Rb/Sr ratios than the later alkalic lava from the postshield and rejuvenated stages. They proposed a complex mixing model to explain the apparent paradox of having more radiogenic isotopic ratios combined with more depleted trace-element ratios in the same rocks. Their model proposes two sources, a primitive mantle-plume source and a depleted oceanic-lithosphere source, which can mix before melting or can produce partial melts which then mix. They argue that small amounts of small-percentage melts from the oceanic lithosphere (midocean-ridge source) are mixed with enriched mantle or with melts derived from enriched mantle. This model is similar to earlier selective-melting models qualitatively proposed by Green and Ringwood (1967) and Macdonald (1968). It would also be possible to mix small-percentage melts of enriched mantle with the oceanic-lithosphere source (basically an enrichment model) to create the source rocks for Hawaiian magma (Clague and others, 1983; Chen and Frey, 1985), although the measured and calculated compositions do not match as closely as in the model of Chen and Frey (1983). All these models predict a range of source compositions from which the lava is generated.

Other studies emphasize the bulk composition of the source rocks and the processes and physical conditions of the melting process. For example, Clague and Frey (1982), from a detailed trace-element analysis of the rejuvenated stage Honolulu Volcanics on Oahu, concluded that lava ranging from nepheline melilitite to alkalic basalt was generated by 2–11 percent partial melting of a homogeneous garnet (<10 percent) lherzolite source that was carbon bearing. The source had been recently enriched and had a chondrite-normalized La/Yb ratio of 4.4. During melting, phlogopite, amphibole, and a Ti-rich phase (oxide?) remained in the residua, but apatite was completely melted. This model can be combined with the model of Chen and Frey (1983) to generate the source composition indicated for the Honolulu Volcanics (see Roden and others, 1984). The homogeneous composition of the mantle source for the Honolulu Volcanics implies that the recent enrichment event affects a large volume of depleted mantle from which the lava is then generated by partial melting. This is not the same process espoused in the model preferred by Chen and Frey (1983) in which enriched mantle or partial melts of enriched mantle mix with partial melts of depleted mantle. Note that all these models consider only two mixing end members, whereas the isotopic data clearly indicate that at least three distinct source compositions are required (Tatsumoto, 1978; Staudigel and others, 1984). Feigenson (1984) proposed three-end-member mixing models for Kauai lava but did not identify the trace-element signatures of the source components.

The generation of large volumes of tholeiitic lava has been the focus of recent studies by Wright (1984) and Budahn and Schmitt (1984), who used different approaches and reached dramatically different conclusions. Wright (1984) used mass-balance considerations to calculate the components and abundance of material that must be added to depleted lithosphere to generate Hawaiian tholeiitic basalt by large percentages of partial melting (35–42 percent melting). His models did not attempt to calculate the variations in source composition for tholeiitic basalt from the different volcanoes (Leeman and others, 1977, 1980; Basaltic Volcanism Study Project, 1981). Budahn and Schmitt (1984) used inverse procedures to estimate the variations in source composition required to generate the tholeiitic basalt from a number of Hawaiian volcanoes. Their estimated sources had 74–86 percent olivine plus orthopyroxene, 11–21 percent clinopyroxene, and 3–5 percent garnet. All the calculated sources had slightly enriched light-rare-earth-element contents, and low heavy-rare-earth abundances (0.9 to 1.6 times chondrites). They calculated the partial melting at 2–10 percent for these sources. Budahn and Schmitt (1984) did not address the processes that led to creation of these different source compositions, nor did they consider the volumes of mantle source regions required, or the constraints on the production of partial melts provided by Kilauea's magma supply and eruption processes. Wright's (1984) model follows from consideration of these additional constraints. The large difference between the models of Wright (1984) and Budahn and Schmitt (1984) emphasizes the uncertainties concerning the compositions and processes that create the source rocks and the lava of the Hawaiian Islands.

PETROLOGIC STUDIES OF LOIHI SEAMOUNT

Studies of Loihi Seamount have provided new insight in magma genesis in the Hawaiian Islands. Trace-element and isotopic studies demonstrate that the source rocks beneath a single volcano are heterogeneous and require at least three mantle components (Frey and Clague, 1983; Lanphere, 1983; Staudigel and others, 1984). Perhaps more important is the observation of very high $^3He/^4He$ ratios, which imply a primitive undegassed source of volatiles (Kaneoka, chapter 27; Kurz and others, 1983; Rison and Craig, 1983; Kaneoka and others, 1983). The ratio of $^3He/^4He$ is inversely related to the volume of the volcanoes on the Island of Hawaii (Kaneoka, chapter 27; Kurz and others, 1983), suggesting that at smaller volcanoes lava is generated from sources that are largely primitive and not degassed. Another observation is that Hawaiian volcanoes initially erupt alkalic lava generated from heterogeneous source compositions by rather small percentages of partial melting (Moore and others, 1982; Frey and Clague, 1983). The evolutionary sequence at a Hawaiian volcano is therefore from small-volume, infrequent eruptions of small-percentage melts to large-volume, frequent eruptions of large-percentage melts, and then back to small-volume, infrequent eruptions of small-percentage melts (Wise, 1982).

PETROLOGIC OVERVIEW

In summary, the petrology of lava from along the Hawaiian-Emperor Chain indicates that at least three source materials are involved in the generation of Hawaiian lava; one of these sources is apparently the depleted ocean lithosphere, whereas another is relatively primitive undegassed mantle. The third component is less well defined. Since multiple sources are required, mixing of these sources or of melts generated from these sources must occur. The compositions of lava along the chain apparently are related to the age (thickness) of the underlying oceanic lithosphere; the volcanoes formed on younger and thinner oceanic lithosphere were generated from a source with a larger component of the depleted ocean lithosphere. Detailed overviews of the petrology of Hawaiian tholeiitic lava and Hawaiian alkalic lava are presented in Wright and Helz (chapter 23) and Clague (in press), respectively.

Volcano volume reflects the degree of melting of the tholeiitic basalt and probably also the size and frequency of intrusion of magma batches. The inverse correlation of volcano volume with $^3He/^4He$ ratio in tholeiitic basalt indicates that smaller percentage melts are derived from sources with more of the primitive component and larger percentage melts are derived from sources with less of the primitive component. Volcano volumes and compositions of the shield tholeiitic basalt are also related, less-enriched tholeiitic basalt forms larger shields and more-enriched tholeiitic basalt forms smaller shields (Clague and Frey 1979), although this correlation is imperfect. Models developed in the future should address the problem of characterizing the isotopic and trace-element compositions of the three mantle components and address the timing and processes of mixing of these sources. The source volumes inferred from different melting models must be considered. Wright's (1984) model requires only modest source volumes, whereas the model of Budahn and Schmitt (1984) requires partial melting of a mantle zone 100 km thick and 100 km wide to generate the volcanoes of the principal Hawaiian Islands. Such enormous inferred volumes of mantle source rock pose numerous problems for models advocating small-percentage melting to generate the tholeiitic shields.

A separate problem is the cause of the alkalic rejuvenated stage. Jackson and Wright (1970) used tide-gage data from Moore (1970) to suggest that generation of the rejuvenated-stage Honolulu Volcanics might be caused by uplift as Oahu has passed over the

Hawaiian Arch. They argued that the Hawaiian Arch, an isostatic response to volcanic loading on the oceanic crust, follows the progression of active volcanic centers by several hundred kilometers and several million years. Clague and others (1982) showed that the duration of the quiescent period preceding eruption of the rejuvenated-stage lava has decreased systematically from nearly 2.5 m.y. on Niihau to <0.4 m.y. at Haleakala (see fig. 1.6). They suggested that a new mechanism should be sought to explain the age data. We have reexamined the data and conclude that they are consistent with the model proposed by Jackson and Wright (1970) because the rate of volcanic migration is increasing. The rejuvenated stage follows the formation of the shield not by a constant time but by a constant distance. The rejuvenated-stage Koloa Volcanics on Kauai and Kiekie Basalt on Niihau began erupting during formation of the Koolau shield located 180–225 km to the southeast. Likewise, the Honolulu Volcanics on the Koolau Range of Oahu began erupting during formation of the East Maui shield located 160 km to the southeast. The rejuvenated-stage Kalaupapa Volcanics on East Molokai erupted during formation of the Mauna Kea shield located 200 km to the southeast. Finally, the rejuvenated-stage Hana Volcanics on East Maui began erupting during formation of the Mauna Loa shield, located 160 km to the southeast. In each case, the lava of the rejuvenated stage began erupting during formation of a large shield 190 ± 30 km to the southeast. The Hawaiian Arch is about 250 km from the center of the volcanic ridge, but only 210 km to the east-southeast of Hawaii (Walcott, 1970). It is therefore likely that a factor in magma generation during the rejuvenated stage is the rapid change from subsidence to uplift as the volcanoes override the flexural arch created by formation of large shields. To the northwest of the Hawaiian Islands the rates of volcanic propagation were slower and more constant; we predict that lava of the rejuvenated stage will postdate the shield stage by 2–3 m.y. We also suggest that the apparent paucity of lava from the rejuvenated stage to the northwest of the Hawaiian Islands may reflect the absence of large volcanic edifices capable of flexing the lithosphere sufficiently. Likewise, the absence of any lava samples of the rejuvenated stage from the Emperor Seamounts may result from the rather wide spacing between volcanic edifices: by the time the next younger volcano formed, the previously constructed volcano was already beyond the arch. The fact that the Emperor volcanoes were constructed on young, thin lithosphere would amplify this effect because the distance from the load to the flexural arch decreases as the lithosphere becomes less rigid.

FIXITY OF THE HAWAIIAN HOT SPOT

Wilson's original hypothesis for the origin of the Hawaiian and other island chains by passage of the crust over a source of lava in the mantle (Wilson, 1963a, b, c) did not require that the hot spot be fixed, only that it have some motion relative to the crust above it. Morgan (1972a, b), on the other hand, specified that a worldwide system of thermal plumes (hot spots) was fixed in the mantle and that the relative movement between them was small or negligible. Several workers (for example, Minster and others, 1974; Gordon and Cape, 1981; Morgan, 1981) have shown from relative plate motions and paleomagnetic and other data that Morgan's hypothesis of relative hot-spot fixity is basically correct, but that the fixity of the hot-spot frame of reference with respect to the spin axis, particularly in early Cenozoic and Late Cretaceous times, is not established.

PALEOMAGNETIC TESTS OF HOT-SPOT FIXITY

Age data along the chain have shown that there has been more or less continuous relative motion between the Hawaiian hot spot and the Pacific plate, thereby proving the kinematic aspect of the hot-spot hypothesis, but these data have little or no bearing on the question of hot-spot fixity. The lava flows that form volcanoes of the Hawaiian-Emperor Chain, however, contain a nearly continuous magnetic record of the latitude of the Hawaiian hot spot for the entire Cenozoic and the latest Cretaceous. Although only a small fraction of this magnetic record has been read, there are now sufficient data to provide a partial test of the fixity hypothesis for the Hawaiian hot spot. Paleomagnetic data from volcanoes along the chain show that the Hawaiian hot spot (and thus presumably the worldwide hot-spot frame) has been, to a first approximation, fixed with respect to the spin axis since the time of formation of the Hawaiian-Emperor bend. The limited data indicate, however, that there was motion between the hot spot and the spin axis, that is, true polar wander, before that time.

The paleolatitudes of several Hawaiian-Emperor volcanoes, as determined from paleomagnetic studies on individual rock samples and from shipboard magnetic surveys, are given in table 1.11 and plotted in figure 1.11 as a function of volcano age. In general, the data indicate that the Hawaiian-Emperor volcanoes formed not at their present latitudes but at a latitude near the present latitude of Hawaii. Thus, the latitude of the Hawaiian hot spot has been approximately fixed throughout the Cenozoic. The data are not of uniform quality, however, and some care must be exercised in their interpretation.

The paleomagnetic data have been discussed and evaluated by Jackson and others (1980), Kono (1980), and Sager (1984), who point out that the paleomagnetic sampling of Meiji, Nintoku, Ojin, Midway, Nihoa, and the Island of Hawaii involved a small number of lava flows, making it doubtful that the secular variation is adequately averaged out. The errors for the Ojin and Nintoku sites reflect this uncertainty, but there is reason to suspect that the errors assigned to the paleolatitudes of Midway, Meiji, and Nihoa are too small. This is because of the unusually low dispersions and the likelihood of serial correlation of some of the flows, which further decreases the number of independent measurements from the sites.

The paleolatitude of $17.5° \pm 5°$, determined for Suiko by Kodama and others (1978) from magnetic survey data, is suspect for several reasons. First, the magnetic anomaly over Suiko is complex, resulting in a low statistical test of fit ($R = 1.1$) for the inversion. Second, it is likely that Suiko is constructed from several coalesced volcanoes (Bargar and Jackson, 1974), possibly of different ages, and the necessary assumption of uniform magnetization is probably invalid for this seamount. In addition, the paleolatitude is inconsistent with that obtained from the paleomagnetic study of

TABLE 1.11.—*Paleolatitudes determined for volcanoes of the Hawaiian-Emperor Chain*

[From paleomagnetic measurements on lava flows and from shipboard magnetic surveys (SM). Data from compilations by Kono (1980) and Sager (1984); the more reliable data (Sager, 1984) underlined; uncertainties are the values of α_{95}]

Volcano Number	Volcano Name	Number of flows determined	Present latitude (degrees north)	Paleolatitude (degrees north)
1,2,4,	Kilauea, Mauna Loa Hualalai (historical)	17	19.5	19.6±1.4
1,2,3	Kilauea, Mauna Loa Mauna Kea(^{14}C dated)	8	19.5	17.7±10.7
12	Koolau	33	21.4	16.8±3.6
13	Waianae	55	21.5	15.7±3.3
14	Kauai (Makaweli Member, Waimea Canyon Basalt)	25	22.0	15.6±3.1
14	Kauai (Napali Member, Waimea Canyon Basalt)	46	22.1	14.9±3.1
17	Nihoa	14	23.1	21.0±6.6
52	Midway	13	28.2	15.4±5.4
65A	Abbott	(SM)	31.8	17.5±2.4
81	Ojin	6	38.0	17.6±13.2
86	Nintoku	4	41.3	36.0±24.6
91	Suiko	(SM)	44.8	16.7±5
91	Suiko	65	44.8	27.1±3.5
108	Meiji	6	53.0	19.2±4.1

Suiko (Kono, 1980), which is the best study of its kind for any of the volcanoes in the chain.

Sager (1984) included in his compilation two additional determinations from the principal Hawaiian Islands that we have chosen to omit from table 1.11. These include a group of 129 flows from the Islands of Hawaii and Niihau, and a second group of 19 flows from Kauai. Both groups include flows from the rejuvenated stage that were erupted several million years after the hot spot had moved (relatively) southeastward to form new tholeiitic shields. Although both of these determinations were included by Sager in his list of more reliable paleolatitudes, they are so close to the present position of the hot spot that their elimination has no significant impact on the conclusions drawn from the data.

Taken at face value, the more reliable paleolatitude data (fig. 1.11) indicate that the Hawaiian hot spot may have been a few degrees south of its present position during the late Cenozoic, near its present position when Abbott Seamount formed at about 39 Ma, and 7° north of its present position at Suiko time, 65 Ma. Analyses of paleoequator (Sager, 1984) and worldwide paleomagnetic data (Livermore and others, 1983), however, show that there has been little or no motion of the spin axis relative to the worldwide hot-spot frame during the past 40 m.y. or so. From this comparison of independent data Sager (1984) concluded that the apparent southward displacement of the Hawaiian hot spot shown by the data from younger volcanoes in the chain (fig. 1.11) reflected changes in the magnetic field rather than relative movement between the hot spot and the spin axis.

The apparent displacement indicated by the Suiko data, however, is probably real. The Suiko paleolatitude is based on analysis of a large number of flows (table 1.11) recovered by coring over an interval of 550 m (Kono, 1980). Even when certain flows thought to represent a very short time interval are grouped, there are still a minimum of 40 independent data. There are also 12 places in the cores where the inclination changes by more than 15°, which indicates that at least 13 secular variation cycles have been sampled, making it likely that secular variation has been adequately averaged out. Other paleomagnetic stability indices indicate that 27.1°±3.5° is a highly reliable measure of the latitude of formation of Suiko Seamount (Kono, 1980).

BIOFACIES AND TEMPERATURE DATA FROM THE HAWAIIAN-EMPEROR CHAIN

Although northward displacement of the Hawaiian hot spot relative to the spin axis is only indicated by the single paleolatitude from Suiko, it is supported by a variety of additional data. Analysis of Pacific deep-sea sediment cores, for example, shows that the paleoequator was 10°–16° farther north than at present between about 75 Ma and 65 Ma (Sager, 1984).

Biofacies data from DSDP Leg 55 drilling in the Emperor Seamounts provide semiquantitative substantiation of the Suiko paleolatitude. The bioclastic sediment on Suiko, Nintoku, and Ojin Seamounts consists primarily of coralline algae and bryozoans with ostracodes, foraminifers, and assorted shell fragments typical of a shallow-water, high-energy environment (Jackson and others, 1980). Only a single coral was found in the Suiko material, and none was recovered from either Ojin or Nintoku, indicating that corals were not significant contributors to the carbonate buildups.

Schlanger and Konishi (1975) have pointed out that carbonate buildups in the Pacific can be divided into bryozoan-algal and coral-algal facies, the distribution of which depends largely on water temperature and solar insolation and thus is, to a large degree, a function of latitude. They observe that in the modern Pacific, the coral-algal facies dominates at latitudes less than about 20°, whereas the bryozoan-algal facies is predominant above about 30° latitude. They locate the boundary between these facies at about 25° latitude, but emphasize that the transition is gradual. In the central Pacific, the annual surface-water temperature at 25° latitude is about 22 °C (Muromtsev, 1958), which is usually considered the minimum for active coral-algal reef growth (Vaughn and Wells, 1943; Heckel, 1974). The optimum temperature for vigorous reef growth is 25–29 °C. Thus, the existence of carbonate buildups of the bryozoan-algal facies atop Ojin, Nintoku, and Suiko Seamounts indicates that the reefs atop the volcanoes formed in water temperatures less than about 22 °C.

Using the oxygen-isotope temperature data of Savin and others (1975) for the North Pacific, Greene and others (1978) reconstructed the approximate latitude variation through time for the 20 °C and 22 °C isotherms (fig. 1.12). They showed that if the Hawaiian hot spot were fixed, then Suiko Seamount would have formed in water warm enough to have developed active coral-algal reefs. Following the analysis of Greene and others (1978), Jackson and others (1980) showed that the Paleocene water temperature at the latitude determined by the paleomagnetic data from Suiko was appropriate for the bryozoan-algal carbonates that occur immediately above the basalt.

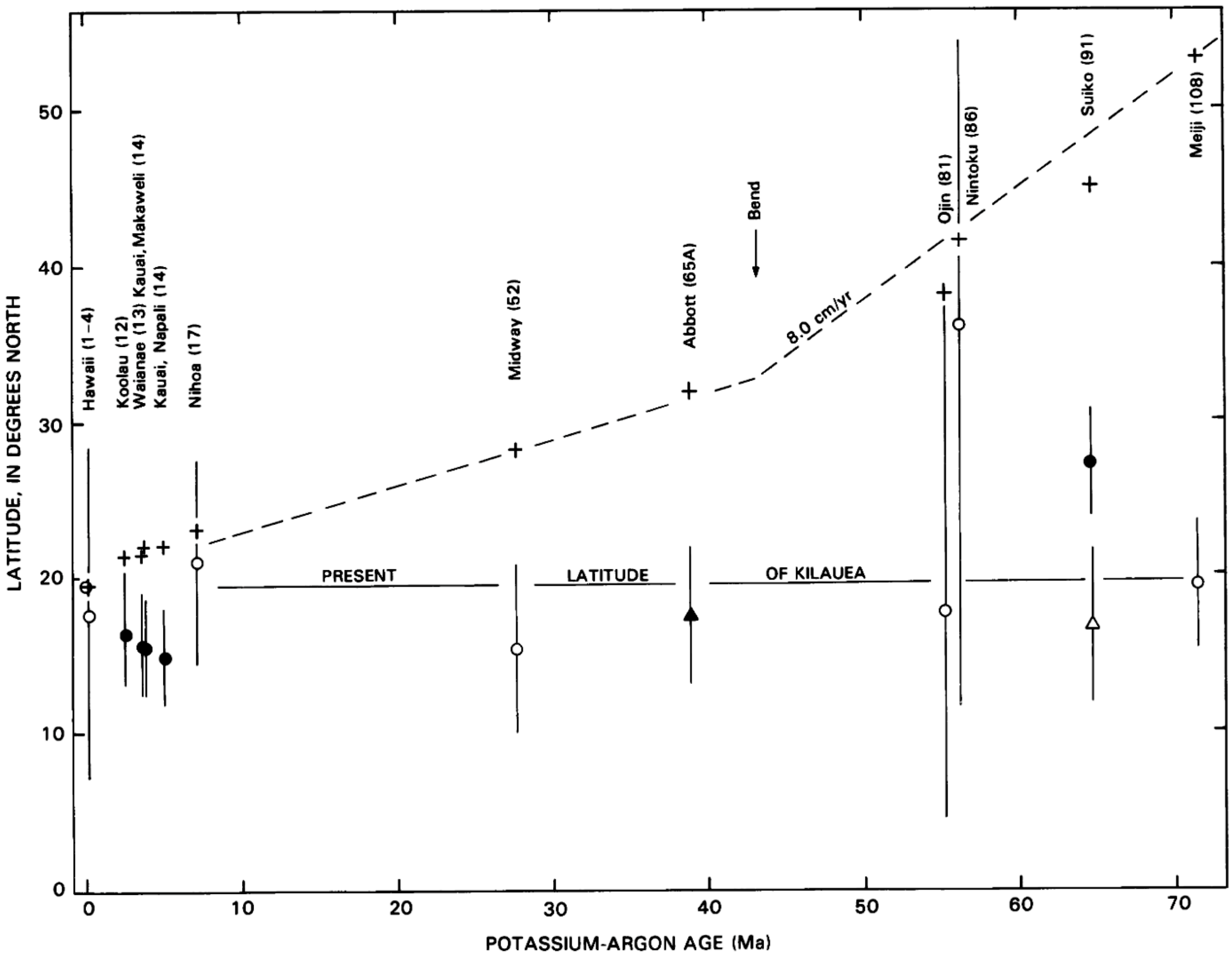

FIGURE 1.11.—Paleolatitude plotted against age for volcanoes along Hawaiian-Emperor Chain. Crosses, present latitude; dots and circles, paleolatitudes determined from paleomagnetic data; triangles, paleolatitudes determined from shipboard magnetic surveys. More reliable data indicated by solid symbols. Error bars show α_{95}. Dashed reference line is backtracked position of hot spot relative to Pacific plate assuming a constant velocity of 8 cm/yr. Paleomagnetic data from table 1.11, age data from table 1.4 and appendix 1.1.

The paucity of coral material on seamounts in the central Emperor Chain is in contrast to Koko and the seamounts on the bend, where corals are more common but still less abundant than in a region of vigorous coral reef growth (Davies and others, 1971, 1972; Matter and Gardner, 1975). Oxygen-isotope temperatures of carbonate diagenesis for Suiko, Nintoku, Ojin, and Kammu Seamounts (McKenzie and others, 1980) show a gradual warming from Suiko to Koko, at least in part caused by southward migration of the hot spot (Jackson and others, 1980). Thus, the biofacies and paleotemperature data from Leg 55 are consistent with the paleomagnetic data, indicating a latitude of 27° for the hot spot at Suiko time. The data are also consistent with Sager's (1984) suggestion that the hot spot had reached its approximate present latitude by the time Abbott Seamount formed just after formation of the Hawaiian-Emperor bend; the slightly cooler temperatures indicated by the carbonate facies and temperature data from the bend seamounts are probably related to the sudden drop in ocean temperature in the late Eocene rather than to a more northerly hot spot.

Thus, the paleomagnetic data from Suiko, the biofacies and temperature data from the central and southern Emperor Seamounts, and the Pacific paleoequator data all indicate southward migration of the Hawaiian hot spot in the early Tertiary and Late Cretaceous. This conclusion is consistent with previous findings, based on analysis of worldwide paleomagnetic data in the hot-spot frame of reference, of about 10° of southward movement of the hot-spot frame relative to the spin axis (that is, true polar wander)

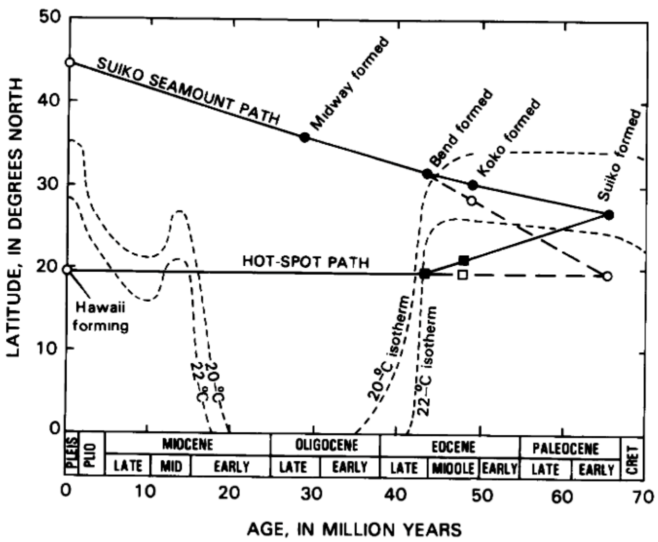

FIGURE 1.12.—Approximate position of 20 °C and 22 °C surface-water isotherms in north-central Pacific during Cenozoic, modified from Greene and others (1978), based on data of Savin and others (1975). Dots, paleolatitudes of Suiko Seamount assuming that Suiko formed at 27° N. and that hot spot has been fixed since time of bend formation. Circles, paleolatitudes for Suiko assuming a fixed hot spot for past 65 m.y. Squares, positions of formation of Koko Seamount and bend under same assumptions. Backtracking was about an Emperor pole at lat 17° N., long 107° W. and a Hawaiian pole at lat 69° N., long 68° W. (Clague and Jarrard, 1973). The 22 °C isotherm is approximate boundary between coral-algal (warmer) and bryozoan-algal (colder) facies of Schlanger and Konishi (1975).

during the latest Cretaceous and earliest Tertiary (for example, Gordon and Cape, 1981; Jurdy, 1981, 1983; Morgan, 1981; Gordon, 1982).

The paleolatitude of 17.5°±4.4° N. found for Abbott Seamount puts the Hawaiian hot spot at about its present latitude by 40 Ma, which is consistent with the conclusion of Livermore and others (1983) that true polar wander did not occur during the past 35 m.y. The paleoequator analysis of Sager (1984) suggests that there was no true polar wander after formation of the bend, that is, after 43 Ma. This requires approximately 7.6° of southward latitudinal motion of the hot spot between 65 Ma and 43 Ma and 5° of northward motion of the Pacific plate in order to satisfy the relative motion of about 0.65° latitude per million years indicated by the age-distance data.

COMPARISON TO OTHER PACIFIC LINEAR ISLAND CHAINS

Another way of evaluating the movement of hot spots is to compare the orientation and age progression quantitatively along volcanic chains formed during the same time period on the same plate. Several studies (Clague and Jarrard, 1973; Jackson, 1976; Jarrard and Clague, 1977; Epp, 1978; McDougall and Duncan, 1980; Turner and others, 1980; and Duncan and Clague, 1985) have attempted such evaluations for the Pacific plate since Morgan (1971) first proposed the technique. Most of the linear volcanic chains in the Pacific basin are oriented roughly west-northwest and apparently formed sequentially over nearly stationary hot spots during the last 43 m.y. as the Pacific plate rotated clockwise about a pole located near lat 69° N., long 68° W. (Clague and Jarrard, 1973). Another group of linear chains exhibit roughly north-trending orientations and apparently formed by the same mechanism between at least 80 Ma and 43 Ma as the Pacific plate rotated clockwise about a pole located near lat 17° N., long 107° W. (Clague and Jarrard, 1973).

The hot spots that formed the Hawaiian, Austral-Cook, Society, Marquesas, Caroline, Pitcairn-Gambier, Samoan, and Islas Revilla Gigedos island chains and the Pratt-Welker and Cobb-Eickelberg seamount chains moved very little with respect to one another (Duncan and Clague, 1985). The most convincing evidence that hot spots move with respect to one another comes from the orientation of the Marquesas Islands, which is discordant by about 25° with that predicted, implying motion of the Marquesas hot spot to the northeast with respect to the hot-spot reference frame at several centimeters per year (Jarrard and Clague, 1977) during the last 5 m.y. The rates of volcanic migration along the chains younger than 43 Ma fit a pole of rotation at lat 68° N., long 75° W. and an angular rotation rate of 0.95°±0.02°/m.y. (Duncan and Clague, 1985).

CAUSE OF THE HAWAIIAN-EMPEROR BEND

An especially knotty problem over the past decade has been the relationship between Pacific sea-floor spreading, worldwide plate motion, and the Hawaiian-Emperor bend. Since there is now firm evidence that the motion of the hot-spot frame was small during the early Cenozoic and has been negligible since then, the 120° angle in the Hawaiian-Emperor bend must represent a major (circa 60°) change in the absolute motion of the Pacific plate. Since the motions of individual plates are not independent, we would expect such a significant change to be part of a worldwide reorganization of both absolute and relative plate motions. Various authors have suggested that the bend might correlate with circum-Pacific tectonic events (Jackson and others, 1972; Clague and Jarrard, 1973; Moore, 1984), that it may be caused by the collision of India and Eurasia (Dalrymple and Clague, 1976), or that it may be the result of new subduction zones along the southwestern margin of the Pacific plate (Gordon and others, 1978). However, completely satisfactory correlations have not been achieved.

A major feature of the northeast Pacific magnetic-anomaly pattern is the major change in the trend of the magnetic anomalies, that is, the magnetic bight, between anomalies 24 and 21. Reconstruction of the Pacific plate shows that this change in the anomaly pattern is the result of a change in spreading about the Pacific-Kula-Farallon triple junction, in particular the cessation of spreading on the Kula Ridge (Scientific Party DSDP 55, 1978; Byrne, 1979). This occurred perhaps as early as anomaly 24 but no later than the time of anomaly 21, which is approximately the time of the major change in spreading direction between Greenland and Europe (Vogt and Avery, 1974) and shortly before an apparent increase in the frequency of geomagnetic reversals (Jacobs, 1984). The change in anomaly orientation can also be correlated with numerous events associated with worldwide reorganization of plate motions (Rona and Richardson, 1978).

The early magnetic time scales of Heirtzler and others (1968) and LaBreque and others (1977) put anomaly 21 at about 54–53 Ma and 52–51 Ma (corrected for new K decay and abundance constants), respectively, which implies a lag of at least 10 m.y. between the reorganization of Pacific magnetic anomalies and the formation of the Hawaiian-Emperor bend at 43.1 ± 1.4 Ma (Dalrymple and Clague, 1976). More recent time scales, however, have narrowed this somewhat awkward gap. Ness and others (1980) put anomaly 21 at about 49–48 Ma, Lowrie and Alvarez (1981) at about 48.5–47.5 Ma, and Butler and Coney (1981) at about 47–46 Ma. As suggested by Butler and Coney, a lag of 3–4 m.y. is close enough to suggest a causal relationship between the relative motion change represented by the magnetic bight and the absolute change represented by the Hawaiian-Emperor bend.

Gordon and others (1978) suggested that the change in direction of the Pacific plate at ~43 Ma was caused by the development of new trenches along the southwestern boundary of the plate. These new trenches, which replaced an earlier set of ridges and transform faults, were the result of rifting of Australia from Antarctica and the accompanying convergence of the Australia-Indian and Pacific plates. Gordon and others (1978) suggest that some time would have elapsed before the subducting plate would have been long enough and dense enough to exert sufficient torque on the Pacific plate to change its direction of motion. This could explain the lag between the timing of reorientation of the magnetic anomalies, which record the change in relative plate motion, and the age of the Hawaiian-Emperor bend, which records the change in absolute plate motion. The duration of the lag time would depend on the rate of plate convergence. As noted by Gordon and others (1978) a lag time of perhaps as much as 10 m.y. might be explained if convergence were sufficiently slow. Their mechanism is more plausible, however, if the lag can be shortened to a few million years, as now seems likely.

HAWAIIAN HOT-SPOT MODELS

Although there is now little doubt that the Hawaiian-Emperor Chain owes its origin to a hot spot that has been approximately fixed with respect to the Earth's spin axis throughout the Cenozoic, there is scant information concerning the exact mechanism involved. Even the term "hot spot" may be misleading, for excess heat is not necessarily involved. Alternatively, it could be the result of pressure release in a mantle source area (Green, 1971; McDougall, 1971; Jackson and others, 1972).

A successful hypothesis for the Hawaiian hot-spot mechanism must explain the propagation of volcanism along the chain, the near-fixity of the hot spot, the chemistry and timing of the eruptions from individual volcanoes, and the detailed geometry of volcanism, including volcano spacing and departures from absolute linearity. Over the past decade or so several mechanisms have been advanced to explain how a linear chain of volcanoes might be progressively erupted onto the sea floor, but most are highly generalized and suffer from lack of detail. Few of the hypotheses address all of the kinematic and petrological issues, and none seems to be amenable to experimental test. Nonetheless, they are interesting speculations on solutions to an extremely difficult problem.

All of the proposed mechanisms can be grouped into four basic types:

1. Propagating fracture driven by lithospheric stresses.
2. Thermally or chemically driven convection.
3. Melting caused by shear between the lithosphere and the asthenosphere.
4. Mechanical injection of heat into the lithosphere.

PROPAGATING-FRACTURE HYPOTHESES

Dana (1849) was the first to associate the Hawaiian volcanic chain with crustal fracturing. He proposed that the Hawaiian and other volcanic chains in the Pacific were each emplaced along a series of short echelon fractures (or "rents") that were widest at the southeast end where volcanism was the most prolonged. He considered these fractures to be part of a worldwide system reflecting tension in the crust resulting from cooling of the Earth from an initially molten state. S. Powers (1917) agreed that the eruptions occurred through a superficial set of echelon fractures following the trend of the chain, but he attributed the trend to some deeper seated lines of weakness. Chubb (1934) thought that the Hawaiian swell represented the surface manifestation of a broad anticline trending in the direction of the chain and produced by compression oriented north-northeast and south-southwest. He proposed that the Hawaiian volcanoes erupted along strike faults and dip faults atop and aligned with the anticline.

Betz and Hess (1942) found no evidence of vertical displacement along fault scarps but thought that the chain might be the manifestation of a great transcurrent strike-slip fault resulting from crustal shortening within the Pacific basin caused by Tertiary volcanism along the margins of the basin. In view of the Earth's sphericity, the straightness of the chain indicated to them that the fault plane was essentially vertical. Considering the strength and thickness of the ocean crust, they thought that an anticline the dimensions of the Hawaiian swell was unlikely and proposed instead that the swell represented a thick lava pile related to the presumed fault zone. Dietz and Menard (1953) thought this idea improbable because of the enormous volume of lava that would be required to produce the swell.

Other early authors subscribed to the idea that the Hawaiian Chain developed atop a propagating fracture (for example, Stearns, 1946; Eaton and Murata, 1960; Jackson and Wright, 1970), but they were vague or noncommital as to the cause of the rupture.

Most recent authors who have advanced propagating-fracture hypotheses have attempted to relate the cause of the fracture to either local or regional stress fields within the Pacific plate. Green (1971) suggested that divergent flow vectors caused by the movement of the plate over an imperfect sphere, that is, an uneven upper mantle surface, caused local tension and intermittent failure of the lithosphere. The fracturing would allow rapid upwelling and partial melting of material from the low-velocity zone. One problem with this hypothesis is the means by which the irregularities on the asthenosphere are maintained, but Menard (1973) suggested that such persistent asthenospheric "bumps" might be caused by a rising thermal plume in the mantle.

McDougall (1971), following the ideas of Green (1971), proposed that the physical feature that subjected the plate to local tension might be either a thermal high or an incipient upwelling caused by a local concentration of heat-producing radioisotopes. According to McDougall's model, fracturing results in the diapiric rise of peridotitic material from the asthenosphere into the lithosphere (fig. 1.13*A*, *B*), where partial melting then generates tholeiitic magma. Movement of the plate and counterflow of the asthenosphere eventually decapitate the diapir, but replacement of material from deeper levels of the asthenosphere perpetuates the high and a new diapir is created (fig. 1.13*C*, *D*). Noting that the rate of propagation of volcanism along the Hawaiian Chain is slightly more than twice the half-spreading rate of the East Pacific Rise, McDougall concluded that there must be counterflow of material in the asthenosphere in a zone of thickness comparable to that of the lithosphere. Jackson and others (1980) showed that it was not possible to reconcile equal-but-opposite hot-spot motion with the paleolatitude of Suiko Seamount if the counterflow had persisted throughout the history of the Hawaiian-Emperor Chain. Hot-spot countermovement until the time of the bend followed by latitudinal stability from then to the present is, however, kinematically permissible.

Another mechanism for producing a local stress field and lithospheric rupture was proposed by Walcott (1976), who related the stress to the volcanic load on the lithosphere. He suggested that large volcanoes will produce lithospheric stresses during growth that may be large enough to cause disruption of the plate. If the plate is under a normal state of horizontal compression, then the failure of the lithosphere will occur preferentially parallel to the direction of the compressive stress. The direction of rupture will remain linear as long as the ambient stress direction remains constant, and the rate of propagation will depend on the speed of formation of the load. Noting the rapidity with which Hawaiian volcanoes form, Walcott concluded that the propagation of volcanism along the Hawaiian chain must be limited by the availability of magma source material. Thus, the mechanism would be self-perpetuating and self-regulating. Although this mechanism will result in a line of volcanoes, it does not explain the observed age progression nor does it account for hot-spot fixity, but it might be locally important and explain the detailed distribution of volcanoes within the Hawaiian Chain (Walcott, 1976).

Expanding on the original idea of Dana (1849), Jackson and others (1972) observed that the individual volcanic centers of the Hawaiian-Emperor Chain appear to lie on short, sigmoidal, overlapping loci that are echelon in a clockwise sense in the Hawaiian Chain and in a counterclockwise sense in the Emperor Chain (fig. 1.14), although the latter was based on inadequate bathymetric data. They proposed that the pattern of loci may be caused by extensional strain resulting from tension within the Pacific plate, but they did not speculate on the ultimate cause of the stresses. Jackson and Shaw (1975) developed this idea more fully and extended it to other chains in the Pacific. They argued that linear hot-spot chains track and record the states of stress in the Pacific plate as a function of time, and that the stress was reflected in the detailed geometry of volcanoes within a chain, that is, in the orientation of the volcanic loci, which represent the injection of magma along lines perpendicular to least principal stress directions. On the basis of their analysis of the Hawaiian-Emperor, Pratt-Welker, Tuamotu, and

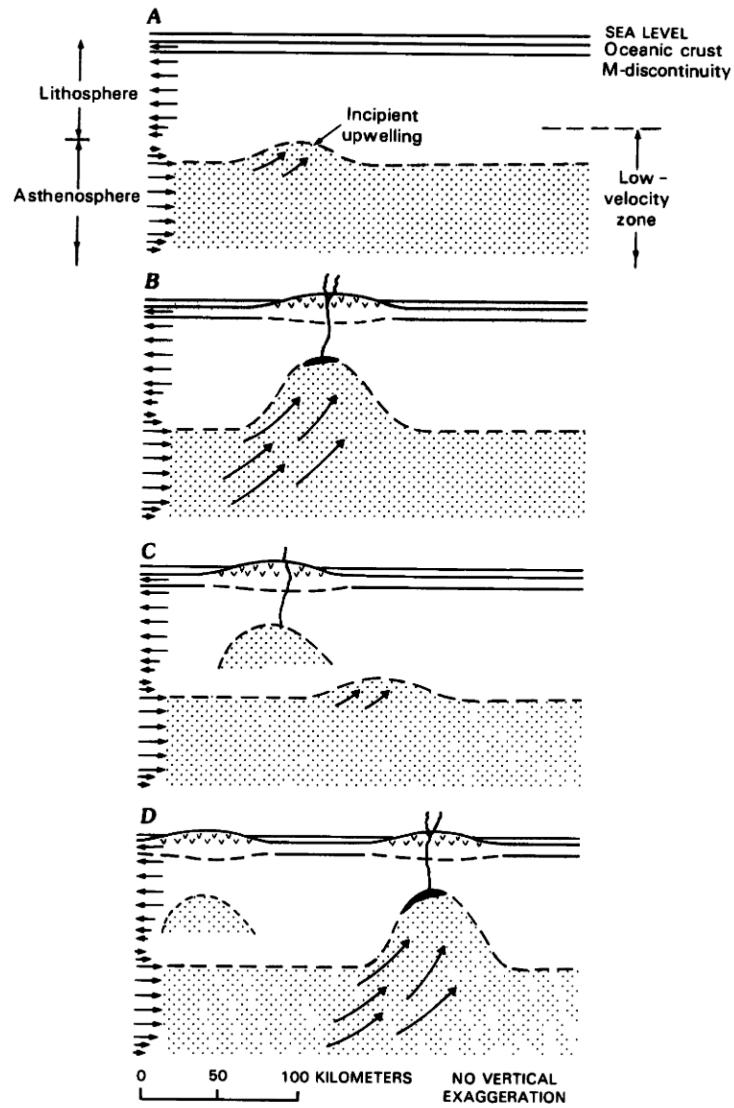

FIGURE 1.13.—Schematic diagram of McDougall's (1971) propagating-fracture hypothesis. Propagating tensional fracture allows diapiric upwelling from asthenosphere (***A***) and partial melting (***B***). Relative motion between lithosphere and asthenosphere (arrow along left margin) eventually decapitates diapir (***C***) and cycle begins at a new position (***D***). The shaded zone represents peridotitic material that is the source rock of the lava. From McDougall.

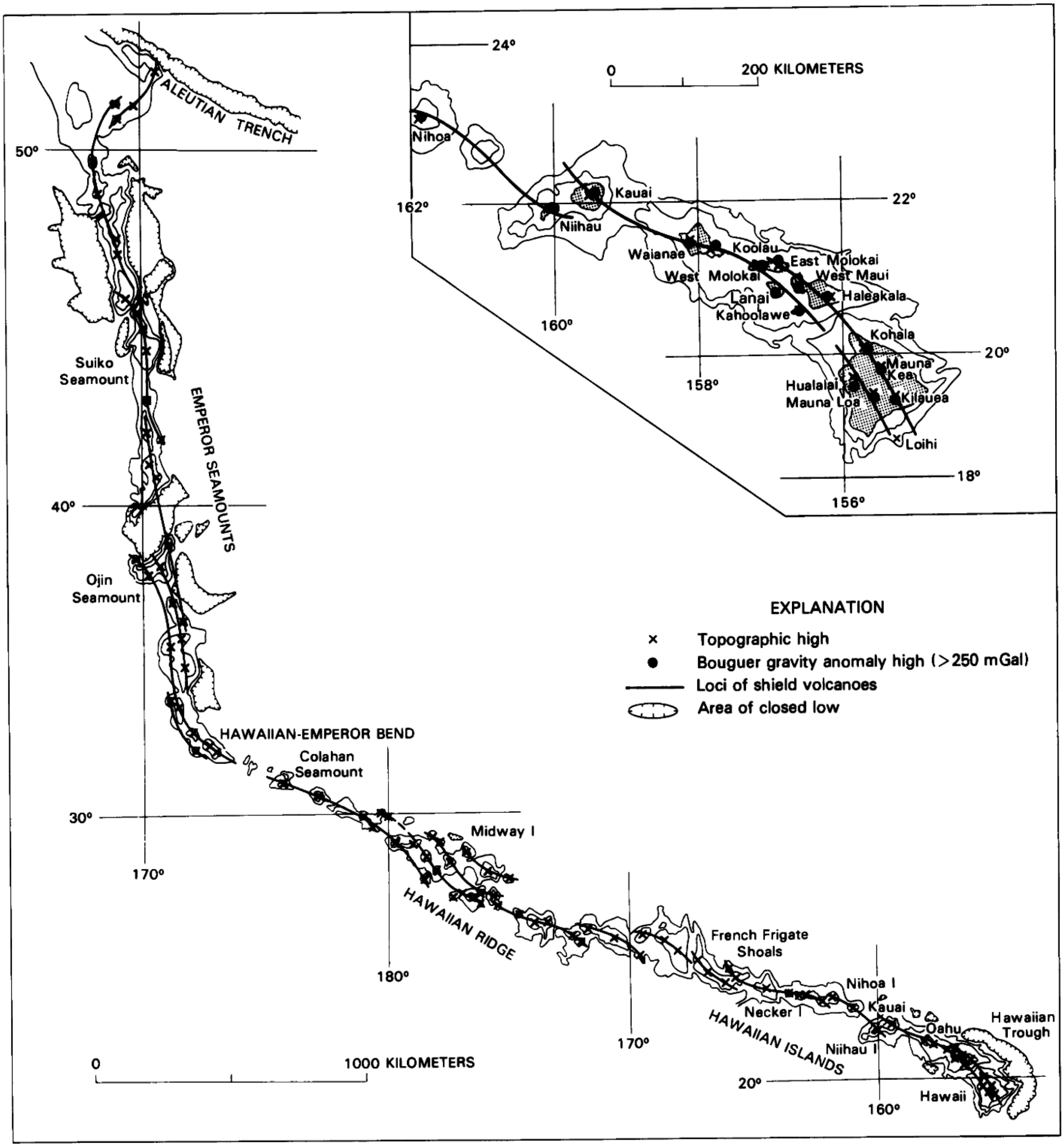

FIGURE 1.14.—Loci of shield volcanoes in Hawaiian-Emperor Chain according to Jackson and others (1972). Inset shows detailed relation between topographic highs, Bouguer gravity anomaly highs, and loci for principal Hawaiian Islands. Contour interval is 1 km; hachures indicate area of closed low.

Austral-Ellice-Gilbert-Marshall Chains, they concluded that the stress orientations since the time of formation of the Hawaiian-Emperor bend were caused by a right-lateral rotational couple acting within the plane of the Pacific plate. This couple resulted in the minimum principal stress oriented in a northeast-southwest direction. Before the time of the bend, the rotational couple was left-lateral and the minimum stress was oriented north-northwest and south-southeast. The curvature in the volcanic loci, they proposed, reflects episodic swings of the minimum-stress directions that averaged about 12 m.y. per episode and were perhaps a consequence of episodic changes in the force vectors at plate boundaries. Jackson and Shaw (1975) were uncertain about the exact causes of the stress field within the Pacific plate, but noted that possible contributors included convergence and divergence at plate boundaries, varying convection rates in the asthenosphere, and volume changes within the plate resulting from changing pressure and temperature.

On the basis of an analysis of volcano spacing and the relation of volcanic chains to preexisting plate structures, Vogt (1974) suggested that the factors that controlled the path of hot-spot chains are not clearly of one origin, but included simple shear, reactivated sea-floor-spreading structures, and local stresses. He concluded that the sigmoidal loci (fractures) postulated by Jackson and others (1972) for the Hawaiian Chain had no counterpart in other chains, although Jackson and Shaw (1975) claimed to have found a similar pattern on other chains in the Pacific.

Solomon and Sleep (1974) preferred the propagating-fracture hypothesis, in part because it avoided the necessity of an abnormal and unknown source of heat in the asthenosphere. They emphasized that the stresses in the Pacific plate can be explained entirely in terms of the forces acting on plate boundaries, and that such mechanisms have the attractive feature of being amenable to numerical treatment. They proposed that the continued motion of the plate with respect to the boundary force field and to secondary convection cells in the asthenosphere might cause a linear propagating fracture as new parts of the plate moved into zones of tension, and that such a tensional fracture would permit the passive upwelling of volcanic material from below. This model offers no explanation why Hawaii is located where it is. In addition, passive mantle upwelling seems inadequate to produce the enormous volume of lava that comprises the Hawaiian Islands.

Turcotte and Oxburgh (1973, 1976, 1978) also subscribed to propagating tensional fractures as a possible cause of linear midplate volcanism. They noted that although brittle failure may occur at the surface of a plate, plastic failure is more likely at depth where lithostatic pressure is large compared with the yield stress. Theoretically, plastic and brittle failure will occur at angles of 35° and 45°, respectively, to the direction of tension (fig. 1.15*A*). Possible causes of tension include thermal stresses in the cooling and thickening plate as it moves outward from the spreading ridge and membrane stresses caused by the movement of plates on the surface of the nonspherical Earth (fig. 1.15*B*, *C*). Turcotte and Oxburgh note that the angle between the Hawaiian Chain and the direction of sea-floor spreading, as deduced from magnetic anomalies and fracture zones, is 34°, in good agreement with the predicted value. They also note that the angle between the trend of the chain and the

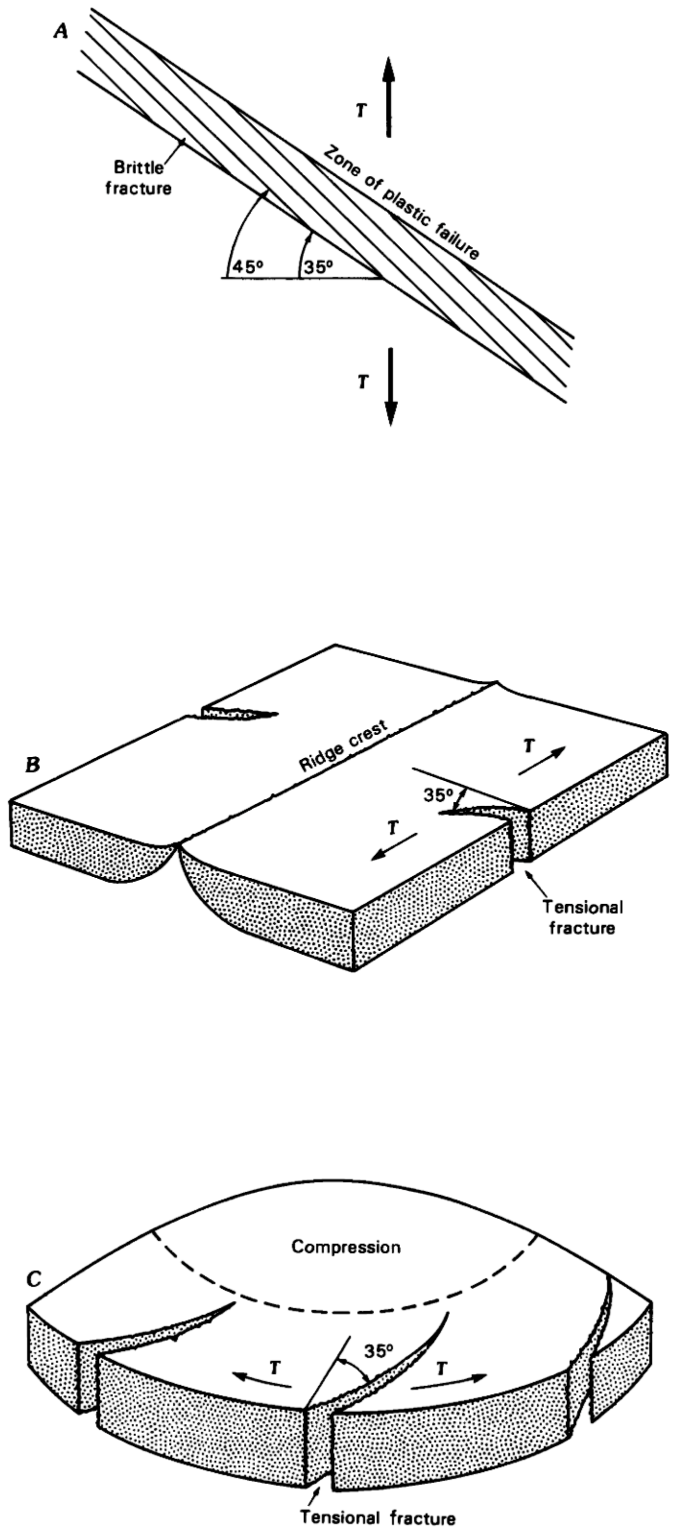

FIGURE 1.15.—Tensional stresses as a possible cause of propagating fractures in lithospheric plates. ***A***, Orientation of plastic and brittle failure in thin plate under tension (*T*) according to Turcotte and Oxburgh (1973, 1976, 1978). ***B***, Fracture in lithospheric plate from tensional stress caused by cooling and thickening of plate away from spreading ridge. ***C***, Fracture caused by membrane stresses in northward-moving plate on oblate Earth.

loci of Jackson and others (1972) is approximately correct for brittle fracture. The Cook-Austral, Tuamoto-Pitcairn, and Kodiak-Bowie Chains also lie at angles of between 31° and 42° to spreading directions, but the angle made by the Marquesas is 60°, which is much larger than that predicted by Turcotte and Oxburgh (1978).

Both plastic and brittle failure, as proposed by Turcotte and Oxburgh, provide a means of propagating a fracture as a function of plate motion and might account for some degree of hot-spot fixity. The thermal mechanism relies on cooling and thickening of the plate as a function of time and distance from the spreading ridge. Once started, the fracture will propagate from a point that remains at a fixed crustal age from the ridge. As these authors point out, fractures due to membrane stresses would be most likely in middle latitudes because the change in the radius of curvature of the Earth is a maximum at a latitude of about 45°. For a plate in the northern hemisphere moving northward, the fracture would propagate southward from a point that remains latitudinally fixed. This mechanism does not, however, account for the great variety of latitudes of active Pacific hot spots, the parallelism of Pacific volcanic chains, or the Hawaiian-Emperor bend (Solomon and Sleep, 1974).

Handschumacher (1973) advanced three fracture-related explanations for the Emperor Seamount Chain, but he did not extend them to include the Hawaiian Chain. Two of the mechanisms, extrusion along a strike-slip fault and interaction between a stable part of the Pacific plate on the west and a spreading ridge on the east, have since been disproved by the age progression (younger southward) of the Emperor volcanoes. The third mechanism, secondary activity along a zone of weakness between eastern and western parts of the plate, invokes preexisting structural control, but, like all propagating fracture hypotheses, does not provide any insight into the lava-producing mechanism.

THERMAL AND CHEMICAL CONVECTION HYPOTHESES

Numerous authors have associated the Hawaiian Chain with thermally driven convection in the asthenosphere. Among the earliest were Dietz and Menard (1953) and Menard (1955), who hypothesized that the Hawaiian Arch or swell occurred over the intersection of two upwelling and diverging convection cells. This would put the lithosphere under tension and produce fracturing as the volcanic load increased, providing a reasonable explanation for the geometry and form of the Hawaiian Arch, Ridge, and Deep. It does not, however, account for the constant rate of propagation of volcanism along the chain, although in 1955 this was poorly known. Although he did not discuss the Hawaiian Chain, Wilson (1962) showed it to be coincident with an early Tertiary ridge which he suggested formed by diverging convection cells.

Wilson (1963a, b, c, d) was the first to suggest a thermal convection mechanism that specifically addressed the age progression in the Hawaiian and eight other parallel chains in the Pacific. He speculated that the source of lava resided in the stagnant, or at least more slowly moving, region of a mantle convection cell (fig. 1.16). Spreading of the sea floor above this fixed source would result in an age-progressive chain of volcanoes. Wilson (1963a) tentatively put the source at a depth of about 200 km, below the low-velocity zone, but did not speculate on the ultimate cause of the lava source.

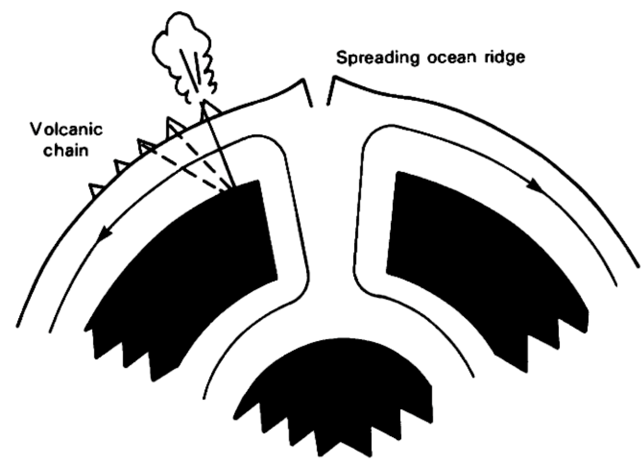

FIGURE 1.16.—Wilson's proposed possible origin of Hawaiian Chain. If lava is generated in stable core of mantle convection cell and surface is carried along by plate motion, then one source can give rise to a chain of successively extinct volcanoes. Modified from Wilson (1963), by permission of the National Research Council of Canada.

The hypothesis that has undoubtedly received the most attention since Wilson's is that of Morgan (1971, 1972a, b) who proposed that the Hawaiian and other Pacific hot spots were narrow thermal zones of upwelling, which he termed "plumes," that originate deep within the Earth's mantle, possibly near the core. They arise because of thermal instabilities (excess heat), which cause upward convection of hot plumes of mantle rock in much the same way that thermal instabilities in the atmosphere cause thunderhead clouds. According to this hypothesis, the plumes are of relatively low viscosity, about 150 km in diameter, and convect upward at a rate of about 2 m/yr. In addition to providing lava for volcanic chains, plumes are considered by Morgan to be a driving force of plate tectonics, to be capable of rifting continents, and to occur on midocean ridges as well as in the middle of plates. Morgan identified about 20 hot spots, but subsequent authors have tended to be more generous (for example, Burke and Wilson, 1976; Crough, 1983).

One aspect of Morgan's hypothesis that has proven extremely important to the study of plate tectonics, whether or not hot spots are actually plumes, is the concept that hot spots are fixed relative to one another and to the Earth's spin axis. As we discussed earlier, hot-spot fixity appears to be generally true for long periods of geologic time, and thus hot spots provide a stable reference frame for studies of absolute plate motions.

Morgan (1972a, 1972b) observed that most hot spots were characterized by a positive gravity anomaly and a topographic high, both of which, he said, are symptomatic of rising thermal currents in the mantle. Morgan calculated that as few as 20 plumes could bring up from depth an estimated 500 km^3/yr of mantle material and half of the total heat flow from the Earth.

Wilson (1973) endorsed the plume hypothesis and likened plumes to other natural diapiric mechanisms such as salt domes, thunderheads, and volcanic pipes. Menard (1973) noted that the Hawaiian, Austral-Cook (Macdonald Seamount), and Gulf of Alaska hot spots all lie on the updrift side of asthenospheric bumps and concluded that equally persistent rising plumes were required to sustain the asthenospheric relief at sites not associated with hot spots. Strong (1974) noted that the compositions of Kilauea and Mauna Loa lava were not the same, concluded that the Hawaiian plume was probably not the direct source of lava, and questioned whether Morgan's plumes were necessarily zones of mass transport. Alternatively, he suggested they might be zones of high thermal conductivity or concentrated diffusion.

Morgan (1972b) proposed four tests of the plume hypothesis, including seismic detection, prediction of plate motions from plume dynamics, evaluation of the necessity of plumes for heat transport from the deep mantle, and correlation of changes in Cenozoic and Cretaceous spreading patterns with the disappearance or emergence of new hot spots. Of the four, only the seismic test had any real potential for yielding a conclusive answer. Davies and Sheppard (1972), Kanasewich and others (1972, 1973), and Kanasewich and Gutowski (1975) analyzed seismic rays passing beneath the Hawaiian Islands from earthquakes in the southwest Pacific. They concluded that there is a zone of abnormally high velocities near the core-mantle boundary beneath Hawaii and that the seismic data are generally consistent with Morgan's plume hypothesis, although there were no data indicating an extension of the velocity anomaly upward through the upper mantle. The interpretation of the seismic data was questioned by Wright (1975) and Green (1975), who concluded that the observed travel time anomalies were most likely the result of upper mantle inhomogeneities beneath the seismic detector arrays in western North America. From a study of teleseismic arrivals from 55 earthquakes recorded at 21 stations on Hawaii, however, Ellsworth and others (1975) found evidence of lower than average velocities at depths of 30–50 km beneath the island. Whether this anomaly extends into the asthenosphere is unknown. Thus, the seismic evidence for a thermal plume beneath Hawaii appears to be, at best, inconclusive.

One difficulty with the plume hypothesis is that narrowly confined convection is unstable in fluids with high Prandtl numbers (kinematic viscosity divided by thermal diffusivity) such as mantle material (Turcotte and Oxburgh, 1978). Narrow plumes might be sustained, however, if confined to the upper mantle and heated from below by a lower mantle source (Turcotte and Oxburgh, 1978). Another problem is that the amount of partial melting that would result from the adiabatic decompression of mantle material rising from the core-mantle boundary is much too high to result in Hawaiian basalt (Turcotte and Oxburgh, 1978). This objection might not apply if mantle plumes are a source of heat for melting of the lower lithosphere or the uppermost asthenosphere rather than a direct source of magma.

An alternative to thermal plumes, proposed by Anderson (1975), is that the plumes are relict compositional conduits. According to Anderson's hypothesis, the Earth accreted inhomogeneously and in the sequence in which compounds would condense from a cooling nebula. Thus, the primitive deep mantle was a material enriched in Ca, Al, Ti, and the refractory trace elements, including U and Th. This material, being less dense than the overlying layers, rose as chemical plumes through buoyancy early in Earth's history and partially melted to yield anorthosites. Present-day hot spots occur above the mantle residua of this partial melting. These plumes provide heat to the base of the lithosphere because they are enriched in heat-producing elements, principally U and Th, and so constitute what could be called radioactive hot spots. Chemical plumes might explain both asthenospheric bumps and also the episodic nature of volcanism. Anderson proposed that the rapid withdrawal of heat by magma could periodically outstrip heat production and therefore temporarily halt magma generation. However, one would think that the chemical inhomogeneities should be seismically detectable.

Richter (1973) and Richter and Parsons (1975) have suggested that the Hawaiian-Emperor Chain and other linear chains might be a consequence of the nonlinear interaction of two different scales of mantle convection, one involving sea-floor spreading and the return flow necessary to conserve mass, and the other a Rayleigh-Benard convection reaching to depths of about 650 km. This latter convection forms rolls whose axes initially are aligned perpendicular to the spreading direction. In time, however, the latitudinal rolls give way to longitudinal rolls with axes parallel to the direction of plate motion (fig. 1.17). The time for the transition to occur depends on the spreading velocity, but may be as short as about 20 m.y. for a fast (about 10 cm/yr) plate like the Pacific plate. Longitudinal rolls will generate alternating bands of tension and compression in the overlying plate. Linear volcanic chains might form along the zones of tension and, either because of modulation of convection amplitude along the roll or because of the fracture properties of the plate, could propagate opposite to the direction of spreading. A feature of this mechanism is that ages out of order can occur. In addition, an age gap of some tens of millions of years could occur near the bend in the Hawaiian-Emperor Chain because of the time required for a new set of longitudinal rolls to be established

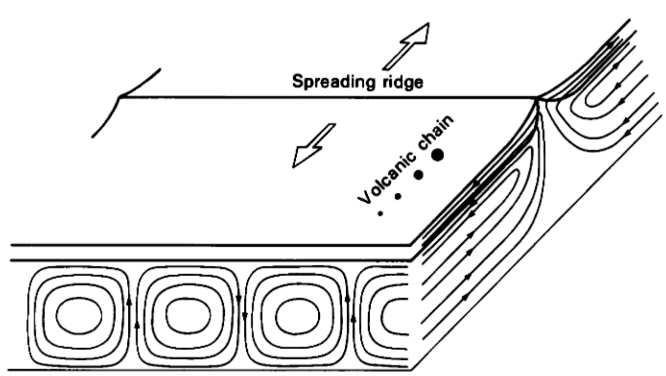

FIGURE 1.17.—Schematic diagram of large-scale asthenospheric flow related to sea-floor spreading and superimposed small-scale longitudinal rolls. Volcanic chain might occur in zone of tension between diverging rolls and would propagate opposite to direction of spreading. From Richter and Parsons (1975).

following a change in spreading direction. The age data for the Hawaiian-Emperor Chain (fig. 1.5), however, show that the propagation is continuous around the bend, although seamounts are sparse on the westernmost Hawaiian Ridge. Another feature of the longitudinal roll model is that parallel volcanic chains should be spaced at some multiple of twice the depth of the convecting layer, which is in accord with the geometry of the major Pacific chains for a convecting depth of about 600 km (Richter, 1973).

SHEAR-MELTING HYPOTHESES

Shear melting with thermal feedback to regulate the propagation rate was proposed by Shaw (1973) to explain the nonlinear time-distance-volume relations along the Hawaiian Chain noted by Jackson and others (1972) and Swanson (1972) (fig. 1.18). According to his hypothesis, the hot spot is the result of a delicately balanced thermomechanical process that derives energy from plate motion and is regulated by a feedback process inherent in the physical properties of the rocks involved. In principle, the idea is quite simple and is based on the observation that a viscous medium will rise in temperature when sheared. Shear occurs within a finite zone between the lower lithosphere and the upper asthenosphere because of their relative motion. As shear proceeds the temperature rises and the viscosity decreases within the shear zone. This allows an increase in the rate of shearing, which in turn produces a further increase in temperature. The increasing temperature eventually results in partial melting and the formation of magma, which rises to the surface to form the volcanoes. The magma carries off excess heat, the temperature decreases rapidly, viscosity increases, and melting stops temporarily as a new cycle is initiated. Each cycle lasts a few million years and is characterized by accelerating propagation of volcanism and eruption volume followed by a sudden halt.

A means of localizing shear melting and fixing the resulting hot spot relative to the mantle was advanced by Shaw and Jackson (1973). They proposed that once partial melting began the residua sinks, forming a type of gravitational anchor that reaches down into the mantle, perhaps to the core-mantle boundary (fig. 1.19). The downwelling anchor not only forms a geographic pinning point for the hot spot but also results in the inflow of fresh mantle material beneath the hot spot, which thus is not limited by supply. There is strong evidence, however, that the depleted residua from partial melting of the most likely parent rocks are less dense than the parent material and would not sink (see, for example, O'Hara, 1975; Boyd and McCallister, 1976; Jordan, 1979). Thus, unless the source of Hawaiian basalt is something quite unusual, the formation of a gravitational anchor seems unlikely, and the shear-melting hypothesis suffers from the lack of both a starting mechanism and a means of localization.

HEAT-INJECTION HYPOTHESIS

It has long been known that the Hawaiian hot spot, among others, is associated with a broad topographic anomaly on the ocean floor, the Hawaiian swell, which has been attributed to some sort of thermal anomaly for more than three decades (see, for example, Dietz and Menard, 1953; Menard, 1955). Only recently, however, has it become clear that the swell may be the result of thermal resetting and thinning of the aging and thickening lithosphere. Detrick and Crough (1978) observed that long-term rates of subsidence of volcanoes in the Pacific are higher than can be accounted for by the subsidence that accompanies the cooling and thickening of the lithosphere as it moves away from the spreading ridge (Parsons and Sclater, 1977; Schroeder, 1984). They proposed that the excess subsidence is the result of thermal resetting of the lithosphere as the aging plate rides over the hot spot. The resetting is accompanied by lithospheric thinning and a rise in the elevation of the sea floor. The rapid subsequent subsidence then represents a gradual return of these shallow areas to normal depths, that is, depths commensurate with the age of the sea floor (see also Crough, 1979, 1983; Epp, 1984). The hypothesis of thermal resetting is supported by anomalously high heat flow along the

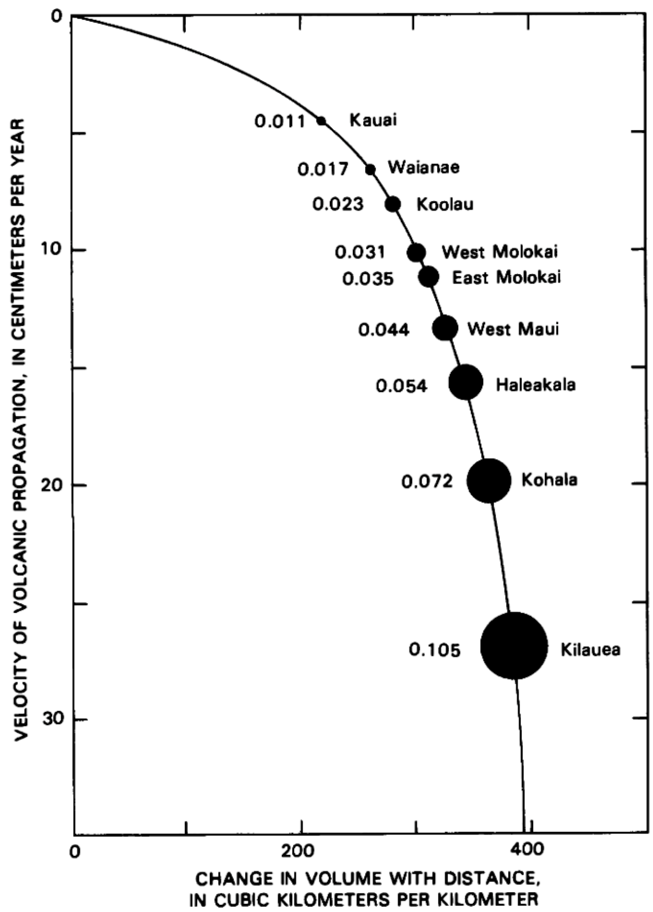

FIGURE 1.18.—Change in volume of lava with respect to unit distance versus change in distance with respect to time (velocity of volcanic propagation) for principal Hawaiian Islands. Diameter of circles approximately proportional to apparent eruption rates, which are also given (in cubic kilometers per year) next to circles. From Shaw (1973).

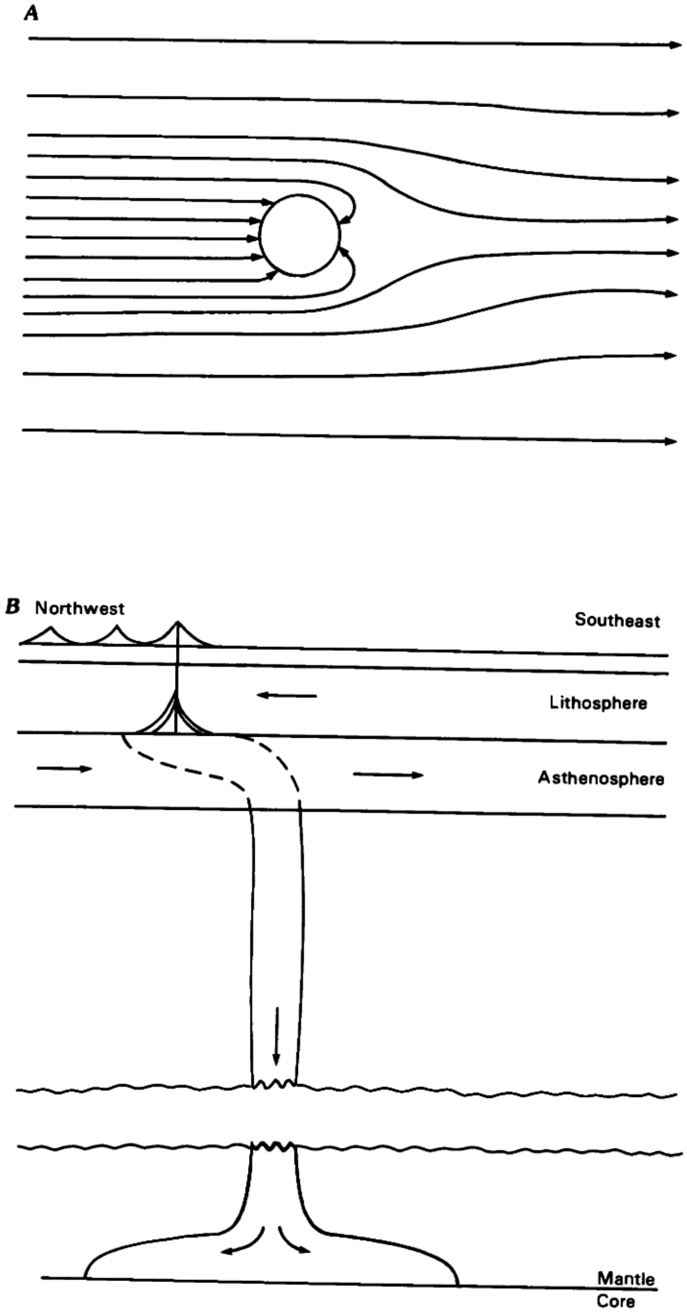

FIGURE 1.19.—Schematic views of possible downwelling of dense residua from tholeiitic melting. *A*, Plan view showing hypothetical flow lines in asthenosphere along horizontal plane taken at time near culmination of melting episode. *B*, Vertical section showing proposed gravitational anchor. From Shaw and Jackson (1973).

Hawaiian Ridge (Detrick and others, 1981). The concept of lithospheric thinning over hot spots is substantiated by the flexural data, which indicate that the lithosphere over hot spots is much thinner than that of comparable age flexed at subduction zones (McNutt, 1984).

Detrick and Crough (1978) recognized that the major problem with their thermal model for the Hawaiian swell is that it requires extremely rapid heating of the lithosphere; a heat flux more than 40 times normal is indicated, if the heating is entirely by conduction. This is because the kinematics of plate motion relative to the hot spot requires the swell to rise in only a few million years, whereas it would take about 100 m.y. at twice the normal heat flux to raise the swell. This problem, however, may not be as serious as it once seemed. More recent modeling by Nakiboglu and Lambeck (1985) demonstrates the sensitivity of these calculations to the lower boundary condition. They argue that most of the Hawaiian swell can be produced by thermal conduction, but a small dynamic component may also be required to support the swell. Another potential solution to this problem, proposed by McNutt (1984), is lithospheric delamination, a process invoked by Bird (1979) to explain volcanism in continental interiors. According to this hypothesis, a strip of the lower lithosphere separates from the upper lithosphere and descends into the asthenosphere. This produces the sudden rise in temperature at the base of the remaining lithosphere required to produce the swell without invoking an unreasonable heat flux. The lateral resistance of the descending strip might also provide the necessary stability of the hot spot with respect to the mantle. It is unclear how delamination might begin, but once started theory suggests that it can propagate at plate velocities (Bird and Baumgardner, 1981).

One problem with delamination is that it requires the sinking into the asthenosphere of the lower lithosphere, which is thought to be one component of the source of ocean-island basalt. For Hawaii the proposed depth of delamination, that is, the thickness of the lithosphere over the hot spot, is slightly less than 30 km (McNutt, 1984), a depth considered to be well above the source region of Hawaiian basalt. It is also clear from pressure-temperature relations that the descending slab would not melt (and if it did the residua would rise rather than sink). Therefore, Hawaiian basalt would have to be generated from the material of the upper asthenosphere, although at lower lithosphere depths, and the lithosphere-asthenosphere boundary would be a purely mechanical one (that is, with no compositional differences across the boundary).

In summary, geophysical models for the Hawaiian hot spot tend to be highly generalized and difficult if not impossible to test. None has yet been advanced that satisfactorily explains all of the geometric, kinematic, physical, and chemical observations from the Hawaiian-Emperor Chain. Although many intriguing and clever ideas have been advanced, the hot-spot mechanism is still somewhat mysterious. Detrick and Crough's (1978) idea that the Hawaiian swell is caused by thermal resetting of the aging ocean crust implies that hot spots are indeed hot. In addition, the possibility that the swell is dynamically supported (Detrick and Crough, 1978) implies that material wells up beneath the lithosphere. Petrologic studies indicate that Hawaiian lava is generated from mantle sources consisting of at least 3 geochemical components; one of these is a primitive undegassed component. The cause of the Hawaiian hot spot is still unknown, but present hypotheses are consistent with Morgan's plume hypothesis in which hot primitive mantle material ascends beneath the ocean lithosphere below Hawaii and reacts with the lithosphere to produce the compositional range of Hawaiian lava.

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APPENDIX 1.1

NOTES ON THE GEOLOGY AND GEOCHRONOLOGY OF INDIVIDUAL VOLCANOES IN THE HAWAIIAN-EMPEROR CHAIN

Following are brief descriptions and comments on the geology and geochronology of all volcanoes in the Hawaiian-Emperor Chain that have been mapped or sampled. References to sources of data are included in the reference list of the main text of the paper. The radiometric ages are conventional K-Ar age determinations unless otherwise indicated; errors are the estimated standard deviations of precision. Volcanoes are discussed from southeast to northwest, that is, from youngest to oldest, using the numbering system (in parentheses) of Bargar and Jackson (1974). Stratigraphic nomenclature used here is from Langenheim and Clague (chapter 1, part II).

Loihi Seamount (no number).—Loihi Seamount is inferred to be the youngest Hawaiian volcano on the basis of its location and seismic activity, the presence of fresh glassy pillow lava, and the occurrence of hydrothermal discharge into the water column. Loihi is located on the southeast flank of Kilauea on the Island of Hawaii and is 30 km offshore and 60 km south-southeast from the summit of Mauna Loa. The seamount rises to 950 m below sea level and has a distinct north-south orientation, delineating two rift zones that extend from the roughly 2.8-km by 3.7-km summit caldera (Malahoff and others, 1982). The volcano is active seismically; persistent swarms of shallow earthquakes that are probably caused by volcanic intrusion or submarine eruptions (Klein, 1982) occur periodically near the summit (Klein and Koyanagi, 1979). Moore and others (1982) describe fresh glassy pillow lava recovered from Loihi Seamount that includes basanite, alkalic basalt, basalt transitional between alkalic and tholeiitic basalt, and tholeiitic basalt. They also demonstrated that the alkalic lava is generally older than the tholeiitic lava and suggested that Loihi Seamount, and presumably all Hawaiian volcanoes, have a stage of alkalic volcanism before the tholeiitic shield stage that is characterized by infrequent, small-volume eruptions of alkalic lava. Further evidence that Loihi Seamount is an active volcano is provided by the discovery of a plume of hydrothermal methane and helium in the water column above Loihi (Horibe and others, 1983; Kim and Craig, 1983) and the presence of water-temperature anomalies recorded by the ANGUS camera sled (Malahoff and others, 1982).

Kilauea Volcano (1).—The lava of Kilauea, currently active and historically the most active volcano in Hawaii, can be divided into the older Hilina Basalt and the younger Puna Basalt, which are separated by the Pahala Ash. The exposed lava consists of tholeiitic basalt and picritic tholeiitic basalt that issued from the 3-km by 5-km summit caldera and the two rift zones. The rift zones extend to the east and the southwest, with the east rift zone extending nearly 50 km from the summit caldera to Cape Kumikahi at the northeast corner of the island and for at least an additional 90 km beneath the sea.

Mauna Loa Volcano (2).—The lava of Mauna Loa is divided into the Ninole Basalt (oldest), the Kahuku Basalt, and the Kau Basalt (youngest). The Kahuku and Kau are separated by the Pahala Ash and are thought to be coeval with the Hilina Basalt and Puna Basalt, respectively, of Kilauea. The exposed lava is all tholeiitic basalt and picritic tholeiitic basalt that issued from the 2.5-km by 4-km summit caldera, named Mokuaweoweo, and two rift zones. The rift zones extend southwest and east-northeast. The southeastern and the southwestern slopes of Mauna Loa are steepened by downfaulting along the Kaoiki and Kealakakua fault systems. Mauna Loa last erupted in 1984. Possible remnants of two earlier shield volcanoes are exposed in the Ninole Hills (Ninole Basalt) and in the vicinity of Kulani. Neither of these earlier shield volcanoes is well delineated and both may be merely deeper parts of the Mauna Loa shield uplifted along normal faults in a manner analogous to that in the Hilina fault system near Puu Kapukapu. With the exception of numerous ^{14}C ages, the sole published age data for Mauna Loa were obtained by Evernden and others (1964) on two samples from the Ninole Basalt. One sample contained negligible radiogenic ^{40}Ar. The other sample had a radiogenic ^{40}Ar content of 2.5 percent and a calculated age of 0.54 Ma. No uncertainty was given for the age, but from the quality of the data we estimate a standard deviation of approximately 0.4 m.y. It seems unlikely that the Ninole Basalt is more than a few hundred thousand years old.

Mauna Kea Volcano (3).—Mauna Kea Volcano last erupted some 3,600 years ago (Porter and others, 1977). The volcano has a well exposed section of shield lava capped by postshield alkalic lava. The rocks are divided into the older Hamakua Volcanics and the younger Laupahoehoe Volcanics. The Hamakua Volcanics consists of shield-stage tholeiitic basalt, olivine tholeiitic basalt, and picritic tholeiitic basalt, and overlying postshield-stage alkalic basalt, ankaramite, and hawaiite. There is no clear boundary between the shield and postshield lavas; tholeiitic and alkalic lavas are intercalated near the boundary (Frey and others, 1984). Most of the surface of Mauna Kea is blanketed by the younger Laupahoehoe Volcanics which is mostly hawaiite, with much less ankaramite and alkalic basalt (West and Garcia, 1982). The Hamakua Volcanics is exposed only in deep erosional canyons.

The rift zones of Mauna Kea are not well defined, but cinder cones are roughly aligned in westerly and southerly directions from the summit. A nearly buried east rift zone is still clearly delineated by a submarine ridge extending nearly 40 km to sea. It is uncertain if a summit caldera existed, but the crude arcuate alignment of some cinder cones, coupled with a large gravity high just south of the summit, indicate that a former caldera may be buried beneath the Laupahoehoe Volcanics.

Porter and others (1977) obtained K-Ar ages on three samples from the Hamakua Volcanics. The two younger flows gave ages of 0.27 ± 0.04 Ma, whereas the older flow gave an age of 0.375 ± 0.050 Ma. K-Ar and ^{14}C ages for flows from the Laupahoehoe Volcanics range from about 0.19 Ma to 4.5 ka. Funkhouser and others (1968) reported a K-Ar age of 0.6 ± 0.3 Ma for a single sample of hawaiite from the Laupahoehoe Volcanics.

Hualalai Volcano (4).—Hualalai Volcano was last active in 1800-01 when two major and several smaller flows issued from a series of vents on the northwest rift zone. With the exception of a large trachyte cone and flow at Puu Waawaa (Waawaa Trachyte Member), the entire subaerial surface of the volcano consists of alkalic basalt flows of the alkalic postshield stage. All subaerial rocks of Hualalai are called the Hualalai Volcanics. A detailed study of the volcano (Moore and others, chapter 20) shows that nearly all the lava is alkalic basalt, with only a few flows that are gradational to hawaiite. Some ^{14}C ages of charcoal from beneath many of these flows demonstrate the youth of the alkalic basalt surface. The structure of the volcano is poorly known, although the northwest and south rift zones are well defined. It is unknown if a summit caldera existed in the past. Recent studies on the submarine northwest rift zone recovered tholeiitic basalt and picritic tholeiitic basalt (Clague, 1982), indicating that Hualalai, like all other Hawaiian volcanoes, had a tholeiitic shield stage. The submarine portion of the rift is overlain by terrace deposits that are inferred to be as old as 120 ka (Moore and Fornari, 1984), which indicates that the tholeiitic shield stage had ended by this time.

Funkhouser and others (1968) reported an age of 0.4±0.3 Ma for the Waawaa Trachyte Member of the Hualalai Volcanics; recent K-Ar results indicate the trachyte is about 105 ka (G.B. Dalrymple, unpub. data, 1985).

Kohala Volcano (5).—Kohala Volcano is composed of the older Pololu Basalt and the younger Hawi Volcanics. The Pololu Basalt is a succession of thin flows of tholeiitic basalt, olivine tholeiitic basalt, and picritic tholeiitic basalt except near the top of the section, where alkalic basalt occurs. Most of the Pololu Basalt represents the shield stage, whereas the uppermost, alkalic part represents the alkalic postshield stage. The younger Hawi Volcanics formed during the alkalic postshield stage and is separated from the Pololu Basalt by an erosional unconformity. Most of the Hawi lava is mugearite, but benmoreite and hawaiite are also present. Cinder cones of the Hawi Volcanics align along two rift zones trending northwest and southeast. Arcuate faults near the summit indicate that a caldera probably formed during eruption of Pololu lava, but Hawi lava has entirely buried it. A gravity high located southeast of the summit may correspond to the approximate location of this inferred caldera. Hawi lava is absent from an 11-km section on the northwest side of the volcano. Flows are deflected from this zone by a series of fault scarps in the summit region that bounds a northwest-trending graben 10 km long and 2–5 km wide. Lava erupted inside the graben, filled it, flowed to the northwest and southeast ends, and overflowed most of the southwest rim. Feigenson and others (1983) made a detailed petrologic study of stratigraphically controlled lava samples from the Pololu Volcanics, and Lanphere and Frey (1985) made a similar study of samples from the Pololu Basalt and Hawi Volcanics.

Evernden and others (1964) obtained an age of 0.43 Ma for one sample of tholeiitic basalt from the Pololu Basalt. Dalrymple (1971) obtained scattered age results that averaged 0.7±0.15 Ma on five samples from the Pololu Basalt. The best data for lava of the Pololu is from McDougall and Swanson (1972), who dated nine flows ranging in age from 0.459±0.028 to 0.304±0.091 Ma. The average of the three oldest flows is 0.43±0.02 Ma. Ages on 12 samples from the Hawi Volcanics (McDougall, 1969; McDougall and Swanson, 1972) range from 0.261±0.005 to 0.061±0.001 Ma.

East Maui Volcano (6).—East Maui Volcano is the youngest volcano in the Hawaiian Islands having rejuvenated-stage volcanics. The volcano was last active in about 1790 (Oostdam, 1965). The oldest unit is the Honomanu Basalt, a series of tholeiitic basalt, olivine tholeiitic basalt, and picritic tholeiitic basalt flows that represents the shield stage. Above sea level these flows are nearly completely buried by those of the overlying Kula Volcanics and Hana Volcanics. The Kula Volcanics is composed predominantly of hawaiite with some ankaramite and alkalic basalt and represents the alkalic postshield stage. There is little evidence of extensive erosion between the shield and postshield lavas. The Hana Volcanics is composed mostly of the same rock types as the Kula Volcanics, but it erupted after an erosional period; it represents an alkalic rejuvenated stage of volcanism. Three rift zones are delineated by the location of vents for the Kula and Hana Volcanics. The east and southwest rifts are characterized by vents of both these units, whereas the north rift has only Kula Volcanics vents. The Hana Volcanics is unique among Hawaiian rejuvenated stage volcanic rocks in that its vents are aligned along the preexisting rift zones, the duration of the erosional period is rather short (<0.4 m.y.), and ankaramite and hawaiite are present. Chen and Frey (1983, 1985) present a detailed geochemical study of lava from all three eruptive stages.

Naughton and others (1980) reported ages for 7 samples of the Honomanu Basalt from three localities. The individual sample ages range from 0.54 to 0.91 Ma. Probably the best age for the Honomanu Basalt is the mean of 0.75±0.04 Ma of the four measurements on samples from the so-called crater of Haleakala. McDougall (1964) dated two samples of the Kula Volcanics at 0.46 and 0.86 Ma, whereas the mean of four samples from the Kula Volcanics dated by Naughton and others (1980) is 0.41±0.09 Ma. No ages have been determined for the Hana Volcanics.

Kahoolawe (7).—The volcanic rocks of Kahoolawe have not been subdivided. The only formation, the Kanapou Volcanics, includes tholeiitic basalt and olivine tholeiitic basalt of the shield stage, tholeiitic and alkalic basalt of the caldera-filling phase, and alkalic basalt and hawaiite of the alkalic postshield stage. Five vents along the seacliff in Kanapou Bay erupted alkalic basalt following an extended period of volcanic quiescence; these vents presumably represent an alkalic rejuvenated stage. The volcano was built by eruptions along a prominent west-southwest rift zone and two less pronounced rifts trending east and north. Most of the vents have been removed by erosion, but remnants of about six vents remain. The caldera of 5 km diameter lies at the eastern end of the island and has been breached by the sea.

Naughton and others (1980) dated two samples collected by H.S. Palmer in 1925 from the upper (alkalic) part of the Kanapou Volcanics. The mean of the two measurements is 1.03±0.18 Ma.

West Maui Volcano (8).—The volcanic rocks of West Maui are divided into the Wailuku Basalt, Honolua Volcanics, and Lahaina Volcanics. The Wailuku Basalt consists of tholeiitic basalt, olivine tholeiitic basalt, and picritic tholeiitic basalt of the shield stage and of alkalic basalt of a caldera-filling phase. The Honolua Volcanics, which represents an alkalic postshield stage, consists of a thin discontinuous cap of mugearite with some trachyte and hawaiite. The Lahaina Volcanics followed a long period of erosion and consists of the cones and flows of four small eruptions of basanite and olivine-rich basanite. The Lahaina Volcanics represents the alkalic rejuvenated stage. The volcano has ill-defined rift zones delineated by dike swarms trending northeast and south and by vents of the Honolua Volcanics trending north and south from the small central caldera. The caldera-filling lava is severely altered by late gases; erosion has preferentially removed these altered rocks in Iao Valley.

McDougall (1964) dated three samples of the Wailuku Basalt, and the ages all fall within the narrow range of 1.30–1.33 Ma with a mean of 1.32±0.04 Ma. Naughton and others (1980) obtained ages of 1.58 and 1.97 Ma on two samples from the Wailuku, but the precision of the measurements is poor. The samples dated by Naughton and his colleagues may be from an older part of the shield than those dated by McDougall (Naughton and others, 1980). McDougall's results for four samples of the Honolua Volcanics range only from 1.18 to 1.20 Ma, whereas Naughton and others (1980) measured an age of 1.50±0.13 Ma for a single sample of the Honolua Volcanics. Naughton and others (1980) dated one sample from the Lahaina Volcanics at 1.30±0.10 Ma, an age which appears to be too old on stratigraphic grounds.

Lanai (9).—Only tholeiitic lava was erupted during the shield stage and caldera-collapse phase of Lanai Volcano; no later alkalic lava is known. The Lanai Basalt consists of tholeiitic basalt, olivine tholeiitic basalt, and picritic tholeiitic basalt that erupted from the northwest, southwest, and southeast rift zones and from the summit caldera underlying the present-day Palawai Basin. An extensive dike swarm marking the southwest rift crops out along the Kaholo Pali. Most of the dikes are nearly vertical and are about 30 cm thick.

Bonhommet and others (1977) measured ages for six samples of the Lanai Basalt that were collected on the southern part of the Lanai shield. The data fit a K-Ar isochron indicating an age for the Lanai Basalt of 1.28±0.04 Ma. Naughton and others (1980) obtained ages of 0.71±1.27 to 0.86±0.55 Ma for three samples from the northeastern part of the island. They speculate that the northeastern part of the shield may be somewhat younger than the southern part, but the mean of their three ages (0.81±0.66 Ma) is not significantly different from the more precise isochron age of Bonhommet and others (1977) at the 95-percent level of confidence.

East Molokai Volcano (10).—The lava of East Molokai is subdivided informally into an upper and a lower member of the East Molokai Volcanics. The lower member consists of tholeiitic basalt, olivine tholeiitic basalt, and picritic tholeiitic basalt characteristic of the shield stage, but alkalic basalt of the postshield stage occurs in the upper part of the lower member. Beeson (1976) and Clague and Beeson (1980) have shown that tholeiitic and alkalic basalt are intercalated in the upper part of the lower unit. The upper member consists predominantly of mugearite, with smaller amounts of hawaiite and trachyte and represents the alkalic postshield stage. Following an extended erosional period, during which the 1,200-m cliffs on the north side of the island formed, alkalic basalt and basanite of the rejuvenated stage Kalaupapa Volcanics erupted to form the Kalaupapa Peninsula and Mokuhooniki Island, a tuff cone located offshore of the east end of the island. East Molokai Volcano was built by eruptions along eastward- and west-northwestward-extending rift zones and from a summit caldera about 2.5 km by 7 km in size. The caldera-filling lava flows are similar to those of the lower member except that they are horizontal, more massive, and more extensively altered.

McDougall (1964) dated two samples from the lower member of the East Molokai Volcanics. Both ages agree at 1.52 Ma. Naughton and others (1980) obtained an age of 1.76 ± 0.07 Ma based on several analyses of a single sample. McDougall also dated three samples of the upper member and obtained ages of 1.35 to 1.49 Ma. A basalt sample from the Kalaupapa Peninsula was dated at 1.24 ± 0.16 Ma by Naughton and others (1980), but this appears to be too old in view of the younger ages (0.35 ± 0.03 to 0.57 ± 0.02) obtained for three samples of the Kalaupapa Volcanics by Clague and others (1982).

West Molokai Volcano (11).—The volcanic rocks of West Molokai belong to the West Molokai Volcanics. Most of the exposed lava is tholeiitic basalt of the shield stage. This lava erupted from an east-northeast rift zone that crosses the summit area. A less pronounced rift trends toward the northwest. There is no evidence of a summit caldera. Alkalic lava of the postshield stage, predominately hawaiite with subordinate alkalic basalt, erupted from a series of cinder and spatter cones located mainly on the northwest rift zone. No rejuvenated-stage lava is known. A shoal area extends beneath the sea to the west-southwest nearly 65 km and includes Penguin Bank. This bank probably represents a separate volcanic center, but it has not been studied and little is known of its history.

McDougall (1964) dated one sample from the summit of the West Molokai shield at 1.89 Ma. From the sample locality, we conclude that it is from the tholeiitic shield (lower part of West Molokai Volcanics). Naughton and others (1980) dated six samples of the lower part of the West Molokai Volcanics from three localities. The three localities gave mean ages of 1.84 ± 0.07, 1.90 ± 0.06, and 1.52 ± 0.06 Ma.

Koolau Volcano (12).—Koolau Volcano on Oahu is composed of tholeiitic basalt, olivine tholeiitic basalt, and rather rare picritic tholeiitic basalt that erupted from a long rift system oriented northwest-southeast. These shield-building flows make up the Koolau Basalt. Lava that ponded in the 16-km by 13-km caldera is also tholeiitic and is called the Kailua Member of Koolau Basalt. The rift zones are identified by an extensive dike complex consisting of hundreds of nearly vertical dikes that average 60–90 cm thick. A few of the youngest flows of the Koolau Basalt appear to be transitional between tholeiitic and alkalic basalt. The rejuvenated-stage Honolulu Volcanics erupted from about 36 groups of vents following a long period of volcanic quiescence and erosion. The lava is strongly alkalic and ranges in composition from alkalic basalt, basanite, and nephelinite, to melilitite. The vents from which these rocks erupted show no relationship to the preexisting rift zones or caldera-bounding faults. Many of the vents formed by violent hydromagmatic eruptions that formed tuff cones commonly containing accidental blocks of Koolau Basalt and coral limestone. Clague and Frey (1982) presented a detailed trace-element geochemical study of the lava and summarized the geology of the Honolulu Volcanics.

K-Ar ages have been reported for a large number of samples from the Koolau Basalt by McDougall (1964), Funkhouser and others (1968), McDougall and Ur-Rahman (1972), and Doell and Dalrymple (1973), who also summarized and evaluated the ages. The best ages for the Koolau Basalt range from 1.8 to 2.7 Ma. Ages for lava of the rejuvenated-stage Honolulu Volcanics range from about 0.03 to 0.9 Ma (Funkhouser and others, 1968; Gramlich and others, 1971; Stearns and Dalrymple, 1978; Lanphere and Dalrymple, 1980).

Waianae Volcano (13).—Waianae Volcano is divided into the older Waianae Volcanics and the younger Kolekole Volcanics. The Waianae Volcanics is subdivided into the Lualualei, Kamaileunu, and Palehua Members. The Lualualei Member consists of tholeiitic basalt, olivine tholeiitic basalt, and picritic tholeiitic basalt of the shield stage. The Kamaileunu Member consists of rocks that accumulated inside the 14-km-wide caldera and is also composed mostly of tholeiitic lava, although alkalic rocks are present near the top. The Kamaileunu Member also includes the only occurrence of icelandite and rhyodacite (Mauna Kuwale-Rhyodacite Flow) in the Hawaiian Islands. The Palehua Member consists mainly of hawaiite with rather rare alkalic basalt flows; it represents the alkalic postshield stage. The Kolekole Volcanics represents the group of young cones near the southwest end of the island and a single flow of alkalic lava erupted in Kolekole Pass. The tholeiitic shield lava erupted from three rift zones trending northwest, south-southeast, and northeast. There is no unconformity between tholeiitic and alkalic lava within the caldera; the boundary is transitional and may be similar to the one on East Molokai. K-Ar ages have been determined for a large number of samples from the Waianae Volcanics by McDougall (1964), Funkhouser and others (1968), McDougall and Ur-Rahman (1972), and Doell and Dalrymple (1973), who also summarized and evaluated all of the data. Ages from the Lualualei and Kamaileunu Members range from about 3.0 to 3.9 Ma. Ages from the Palehua Member range from about 2.5 to 3.2 Ma.

Kauai (14).—The Island of Kauai consists of a single large shield volcano with a summit caldera 16–19 km across. The Waimea Canyon Basalt has been divided into four members, but all consist of tholeiitic basalt, olivine tholeiitic basalt, and abundant picritic tholeiitic basalt. The Napali Member represents the shield stage, whereas the Olokele and Makaweli Members represent the caldera-filling phase, having filled the summit caldera and a 6-km-wide graben on the south flank, respectively. Two other calderas formed on the flanks of the Kauai shield volcano: the Lihue depression, 11–16 km across, was apparently not filled by tholeiitic lava; the Haupu caldera, roughly 3 km across was filled with thick ponded flows called the Haupu Member of the Waimea Canyon Basalt. These are the only flank calderas known in the Hawaiian Islands. Near the top of the Olokele and Makaweli Members, a single flow of hawaiite rests on a soil 30–60 cm thick. This single flow apparently represents the alkalic postshield stage on Kauai. Unlike most Hawaiian volcanoes, Kauai has no well-defined rift zones; dikes radiate from the summit caldera in all directions, although they are more concentrated in the northeast and west-southwest directions.

Following a long period of volcanic quiescence and deep erosion, the alkalic rejuvenated stage Koloa Volcanics erupted from at least 40 vents concentrated on the south and east flanks of the shield. The lava ranges from alkalic basalt, basanite, and nephelinite, to melilitite. The abundant vents located along the southeast coast erupted almost entirely alkalic basalt.

McDougall (1964) reported ages for three samples from the Napali Member ranging from 3.63 to 5.77 Ma. Evernden and others (1964) obtained an age of 3.43 Ma for a single sample from the Napali. In a more recent study, McDougall (1979) reported K-Ar ages ranging from 3.81 ± 0.06 to 5.14 ± 0.20 Ma for 16 samples of the Napali collected from three localities. He concluded that some of the variation was probably due to differential Ar loss, that the Napali Member was erupted between about 5.1 and 4.3 Ma, and that the Napali lava in Waipio Valley was erupted over a short time interval at about 5.1 ± 0.2 Ma.

Ages of 4 samples from the Makaweli Member range from 3.60 to 4.15 Ma (McDougall, 1964). Only three samples from the Koloa Volcanics (rejuvenated stage) have been dated; two samples have ages of 0.62 and 1.21 Ma (Evernden and others, 1964) and another an age of 1.46 Ma (McDougall, 1964).

Niihau (15).—The Island of Niihau consists of a deeply eroded shield volcano mantled by lava of the alkalic rejuvenated stage on the north, west, and south sides. The Paniau Basalt consists of tholeiitic basalt and olivine tholeiitic basalt of the shield stage and the remnants of a single alkalic postshield stage vent at Kaeo. Several dikes exposed near the eastern coastline are also of alkalic basalt and presumably fed vents that have been completely removed by erosion. A magnificent dike swarm is exposed in the eastern seacliff (Dalrymple and others, 1973, fig. 5); these dikes trend southwest and represent a rift zone. The summit of the volcano was northeast of the present-day island, and the eastern side of the volcano has been removed by erosion or downfaulting. The period of volcanic quiescence and marine erosion that removed the eastern side of the shield was followed by eruption of the alkalic rejuvenated-stage lava of the Kiekie Basalt, which is entirely alkalic basalt. Lehua Island off the north shore is a breached tuff cone of the Kiekie Basalt.

Ages for Niihau have not been published, but data for 11 tholeiitic flows and dikes of the Paniau Basalt fit a K-Ar isochron with an age of 4.89 ± 0.11 Ma (G.B. Dalrymple, unpub. data, 1983).

Kaula Island (15A).—Kaula Island is a small tuff cone on a large submarine edifice. The edifice almost surely represents a separate shield volcano related to the Hawaiian Islands, but it has not been sampled. The tuff cone is probably a vent of the alkalic rejuvenated stage. Garcia, Frey and Grooms (in press) described accidental blocks of tholeiitic basalt, basanite, phonolite, and ultramafic xenoliths that occur in the tuff.

Garcia, Grooms, and Naughton (in press) obtained K-Ar ages for two phonolite blocks and a basanite from the tuff; they determined ages of 4.00 ± 0.09 and 4.22 ± 0.25 Ma for phonolite samples, 3.98 ± 0.70 Ma for a biotite separated from the phonolite, and 1.8 ± 0.2 Ma for the basanite.

On the basis of the composition and age of the phonolite, we propose that the phonolite is from the alkalic postshield stage, the basanite blocks are from rejuvenated-stage flows underlying the tuff cone (Garcia, Frey, and Grooms, in press), and tholeiitic basalt represents the shield stage.

Nihoa Island (17).—Nihoa Island is a remnant of a large tholeiitic shield with flows dipping 5°–10° to the southwest. All the flows exposed on Nihoa are of tholeiitic basalt, and they range from aphyric to porphyritic in texture (Dalrymple and others, 1974).

Funkhouser and others (1968) obtained an age of 7.5 ± 0.4 Ma for a single sample from Nihoa Island. Dalrymple and others (1974) reported a best weighted mean age of 7.2 ± 0.3 Ma for six samples of tholeiitic basalt from the island.

Unnamed Seamount (19).—A single sample dredged by the Hawaii Institute of Geophysics from this seamount has been analyzed (Garcia, Frey, and Naughton, in press). The samples in dredge 9-11 are moderately altered picritic tholeiitic to transitional basalt with about 30 percent olivine phenocrysts. These probably erupted during the late shield stage or the caldera-filling phase. The seamount is undated.

Unnamed Seamount (20).—On the second leg of the Hawaii Institute of Geophysics cruise 72-07-02, dredge 51 recovered tholeiitic basalt with 20 percent phenocrysts of augite, plagioclase, and olivine, and aphyric alkalic lava of basanite composition (Clague, 1974a; Garcia, Frey, and Naughton, in press). These rocks probably erupted during the shield and the alkalic rejuvenated stages, respectively. Garcia, Frey, and Naughton (in press) obtained a weighted mean age of 9.6 ± 0.8 Ma for four analyses of two of the dredged alkalic basalt samples.

Unnamed Seamount (21).—On the second leg of the Hawaii Institute of Geophysics cruise 72-07-02, dredge 49 recovered tholeiitic to transitional basalt containing 20 percent phenocrysts of olivine and augite (Clague, 1974a). This lava probably erupted during the shield stage. The volcano is undated.

Necker Island (23).—Samples have been collected from Necker Island (Dalrymple and others, 1974), and others have been dredged from the submarine flanks of the volcano during Hawaii Institute of Geophysics cruises 72-07-02 (second leg, dredge 48) and KK84-08-06-01 (Clague, 1974a; Campbell and others, 1984). The subaerial samples are mostly picritic tholeiitic basalt collected from flows dipping 5°–10° to the north-northwest. Palmer (1927) described two dikes that are alkalic lava. One is highly altered, but on the basis of chemical analysis appears to be a nephelinite; the other is described as a hawaiite. These two dikes probably fed vents during the alkalic rejuvenated stage and alkalic postshield stage, respectively. The single lava sample dredged in 1972 is a rhyolite porphyry (Clague and Dalrymple, 1975). This rock type is unknown from elsewhere in the Hawaiian-Emperor Chain, and we suspect that it is either an ice-rafted erratic or a piece of ship's ballast. The 1984 dredges have not yet been analyzed but contain calcareous sediment, volcaniclastic breccia, basalt, and hyaloclastite.

Funkhouser and others (1968) reported an age of 11.3 ± 0.6 Ma for a single sample of subaerial basalt. Dalrymple and others (1974) dated two samples of tholeiitic basalt from the island; they gave a mean age of 10.3 ± 0.4 Ma. The rhyolite porphyry has a Cretaceous age (Clague and Dalrymple, 1975).

La Perouse Pinnacle (French Frigates Shoal) (26).—La Perouse Pinnacle and an even smaller adjacent rock are the only subaerial exposures of volcanic rock within French Frigates Shoal, a coral atoll consisting of 15 or 16 small sand islets. La Perouse Pinnacle is a stack of lava flows that dip 1°–2° to the northwest. The subaerial flows are picritic tholeiitic basalt (Dalrymple and others, 1974) that probably erupted during the shield stage. Four dated samples have a mean age of 12.0 ± 0.4 Ma.

Brooks Bank (28).—Three samples have been analyzed from dredge 41 of the second leg of the Hawaii Institute of Geophysics cruise 72-07-02 (Clague, 1974a; Garcia, Frey, and Naughton, in press). Two of these samples are hawaiite, probably from the same flow, and the third sample is an olivine basalt transitional between tholeiitic and alkalic basalt. The hawaiite probably erupted during the alkalic postshield stage and the transitional basalt during either the late shield stage or the caldera-collapse phase. The hawaiite and alkalic basalt have a mean age of 13.0 ± 0.06 Ma.

St. Rogatien Bank (29).—A single sample has been analyzed from dredge 44 of the second leg of the Hawaii Institute of Geophysics cruise 72-07-02 (Clague, 1974a). The sample is an aphyric hawaiite that probably erupted during the alkalic postshield stage; it has not been dated.

Gardner Pinnacles (30).—The two rocks that constitute Gardner Pinnacles are the westernmost subaerial exposures of volcanic rock in the Hawaiian Chain. The alkalic basalt flows that make up the rocks dip 15° to the west and are cut by several east-trending dikes (Dalrymple and others, 1974). Dredged samples of geochemically similar, though less differentiated, alkalic basalt were recovered in dredge 37 from the second leg of the Hawaii Institute of Geophysics cruise 72-07-02 (Clague, 1974; Garcia, Frey, and Naughton, in press). A later dredge on the flank of Gardner Pinnacles (HIG dredge 6-7; see Garcia, Frey, and Naughton, in press) recovered largely unaltered picritic tholeiitic basalt. We infer that the picritic tholeiitic basalt erupted during the shield stage and the alkalic basalt flows during the postshield stage. Additional samples have recently been recovered during Hawaii Institute of Geophysics cruise KK84-04-28-05 from a number of dredge stations on Gardner Pinnacles, but these have yet to be analyzed (Campbell and others, 1984).

Samples from the island were too altered for dating (Dalrymple and others, 1974), but Garcia, Frey, and Naughton (in press) obtained a weighted mean age of 12.3 ± 1.0 Ma for two dredged samples of alkalic basalt and one of tholeiitic basalt.

Laysan Island (36).—A single dredge during U.S. Geological Survey cruise LEE8-76-NP recovered a variety of hawaiite and mugearite pebbles (Dalrymple and others, 1981) that probably erupted during the alkalic postshield stage. Conventional K-Ar and ^{40}Ar-^{39}Ar measurements on five of the samples fall within the range 18.8–21.4 Ma, and ^{40}Ar-^{39}Ar incremental heating experiments on three samples gave a mean age of 19.9 ± 0.3 Ma.

Northampton Bank (37).—A Hawaii Institute of Geophysics cruise sampled the south side of Northampton Bank and recovered coral-reef debris, picritic tholeiitic basalt, and olivine tholeiitic basalt that probably erupted during the shield stage. Dalrymple and others (1981) reported conventional K-Ar and ^{40}Ar-^{39}Ar age data for three dredged samples of tholeiitic basalt. Only one of the samples gave a ^{40}Ar-^{39}Ar age spectrum plateau. The inferred age for that sample is 26.6 ± 2.7 Ma.

Pioneer Bank (39).—On the second leg of the Hawaii Institute of Geophysics cruise 72-07-02, dredge 25 recovered pillow breccia of olivine tholeiitic basalt (Clague, 1974a) that probably erupted during the shield stage. The volcano is undated.

Pearl and Hermes Reef (50).—On the second leg of the Hawaii Institute of Geophysics cruise 72-07-02, dredge 24 recovered round clasts of alkalic basalt, hawaiite, and nepheline phonolite (Clague and others, 1975) that probably erupted during the alkalic postshield stage. It is possible that the phonolite sample erupted during an alkalic rejuvenated stage, although other phonolite samples from Koko Seamount in the Emperor Seamounts are all interpreted to have erupted during the alkalic postshield stage (Clague, 1974a). The weighted mean age of phonolite, hawaiite, and alkalic basalt is 20.6 ± 0.5 Ma.

Ladd Bank (51).—On the second leg of the Hawaii Institute of Geophysics cruise 72-07-02, dredge 23 recovered a single fresh clast of ankaramite vitrophyre that is compositionally similar to a basanite or nephelinite (Clague, 1974a). This sample probably erupted during an alkalic rejuvenated stage; it is undated.

Midway Island (52).—In 1965, two holes were drilled through the reef on Midway and into flows of tholeiitic basalt (Ladd and others, 1967). Analyses of the tholeiitic flows are presented in Dalrymple and others (1974) and of hawaiite and mugearite cobbles from a conglomerate overlying the flows in Dalrymple and others (1977). We infer that the tholeiitic flows erupted during the shield stage and the hawaiite and mugearite during an alkalic postshield stage.

Dalrymple and others (1974) reported ages for four samples of tholeiitic basalt from the reef drill hole at Midway. The ages ranged from 10.8 to 18.2 Ma. In the later study, Dalrymple and others (1977) reported an age for Midway of 27.7 ± 0.6 Ma based on conventional K-Ar and ^{40}Ar-^{39}Ar analyses of two unaltered samples of hawaiite and mugearite from the conglomerate. Incremental heating experiments showed that the conventional K-Ar ages obtained earlier for the tholeiitic basalt samples do not represent crystallization ages.

Nero Bank (53).—Scripps Institution of Oceanography cruise TASADAY III recovered a vitrophyre of picritic tholeiitic basalt that probably erupted during the shield stage (Clague, 1974a); it is undated.

Unnamed Seamount (57).—On the second leg of the Hawaii Institute of Geophysics cruise 72-07-02, dredge 20 recovered several samples of open-textured alkalic basalt that probably erupted during an alkalic postshield stage (Clague and others, 1975). Three samples of the basalt have concordant K-Ar ages with a mean of 28.0 ± 0.04 Ma.

Unnamed Seamount (58).—DSDP Site 311, located 240 km west of Midway Island, recovered volcanogenic deposits from the archipelagic apron of this volcano (Larson and others, 1975) that yielded a nannoplankton age of 31–28 Ma (Bukry, 1975).

Unnamed Seamount (63).—Scripps Institution of Oceanography cruise TASADAY III recovered a wide range of alkalic lava types including ankaramite, analcime tephrite, amphibole-bearing tephrite, and amphibole-bearing hawaiite (Clague, 1974a, 1974b; Clague and others, 1975). These strongly alkalic rocks probably originated by crystal fractionation from alkalic rejuvenated-stage basanitic parental magma. Clague and others (1975) obtained concordant K-Ar results from three samples of the alkalic lava; the mean age is 27.4 ± 0.05 Ma.

Hancock Seamount (64).—Hawaii Institute of Geophysics cruise KK84-04-28-05 recently recovered samples in a number of dredges from Hancock Seamount; these samples have not been analyzed or dated (Campbell and others, 1984).

Colahan Seamount (65).—On U.S. Geological Survey cruise L8-82-NP, dredge 4 recovered samples of transitional basalt, tephrite, and amphibole-bearing hawaiite (Duncan and Clague, 1984; D.A. Clague, unpub. data, 1983) that probably erupted during an alkalic rejuvenated stage. Analysis of these samples is still in progress; therefore the identification of eruptive stage is less certain than for the other seamounts. Duncan and Clague (1984) have reported ^{40}Ar-^{39}Ar total-fusion ages of 37.5 ± 0.3 and 39.8 ± 0.2 Ma for two alkalic basalt samples.

Abbott Seamount (65A).—On U.S. Geological Survey cruise L8-82-NP, dredges 2 and 3 recovered samples of transitional to alkalic basalt that probably erupted during the late shield stage or caldera-collapse phase (Duncan and Clague, 1984). Analysis of these samples is still in progress. Duncan and Clague (1984) reported ^{40}Ar-^{39}Ar total-fusion ages of 40.4 ± 0.5 and 36.3 ± 0.3 Ma for two of the samples.

Kammu Seamount (66).—On Scripps Institution of Oceanography cruise AIRES VII, dredge 54 recovered abundant carbonate reef debris but no volcanic rocks. N. Sachs (quoted in Clague and Jarrard, 1973) identified *Spiroclypeus variablis* Tan., a large foraminifer of late Eocene age.

Daikakuji Seamount (67).—On Scripps Institution of Oceanography cruise AIRES VII, dredge 55 recovered a range of lava samples including hypersthene-bearing tholeiitic basalt, basalt transitional between tholeiitic and alkalic basalt, and alkalic basalt (Clague, 1974a; Dalrymple and Clague, 1976). Microprobe analyses of glass rinds on some of these samples are in agreement with the published analyses on altered whole-rock samples. The tholeiitic basalt is interpreted to have erupted during the shield stage, the transitional basalt during the late shield stage or caldera-filling phase, and the alkalic basalt during the alkalic postshield stage.

Dalrymple and Clague (1976) made conventional K-Ar and ^{40}Ar-^{39}Ar age determinations on tholeiitic and alkalic basalt and on plagioclase separates. On the basis of ^{40}Ar-^{39}Ar incremental-heating results from the alkalic basalt and ^{40}Ar-^{39}Ar total-fusion analyses of the plagioclases samples, they concluded that the best age for the seamount was 42.4 ± 2.3 Ma.

Yuryaku Seamount (69).—On Scripps Institution of Oceanography cruise AIRES VII, dredge 53 recovered several fairly fresh pebbles of alkalic basalt (Clague, 1974a; Clague and others, 1975; Dalrymple and Clague, 1976). These samples probably erupted during the alkalic postshield stage.

Clague and others (1975) determined an age of 43.4 ± 1.6 Ma for Yuryaku on the basis of ^{40}Ar-^{39}Ar incremental-heating experiments on two dredged samples of alkalic basalt.

Kimmei Seamount (72).—On Scripps Institution of Oceanography cruise AIRES VII, dredges 51 and 52 recovered several samples of alkalic basalt that have been analyzed (Clague, 1974a; Dalrymple and Clague, 1976). Two of these samples are rather severely phosphatized, but all three probably erupted during the alkalic postshield stage.

Dalrymple and Clague concluded that the best age for Kimmei was 39.9 ± 1.2 Ma from ^{40}Ar-^{39}Ar incremental-heating experiments on three dredged samples of alkalic basalt.

Koko Seamount, southeast part (74).—On Scripps Institution of Oceanography cruise AIRES VII, dredge 43 recovered a large collection of rounded volcanic beach cobbles and abundant coral fragments. The volcanic cobbles include tholeiitic basalt, alkalic basalt, hawaiite, mugearite, trachyte, and phonolite (Clague, 1974a). The tholeiitic basalt probably erupted during the shield stage and the entire suite of related alkalic lava types probably erupted during the alkalic postshield stage. DSDP Leg 32 drilled two shallow holes on Koko Seamount, but neither reached volcanic basement (Larson and others, 1975). The structure and seismic stratigraphy of the seamount are described by Davies and others (1972) and Greene and others (1980).

Clague and Dalrymple (1973) obtained conventional K-Ar and ^{40}Ar-^{39}Ar total fusion data on seven dredged samples of sanidine trachyte, alkalic basalt, and phonolite. Krummenacher (cited in Clague and Jarrard, 1973) obtained K-Ar ages of sanidine from two trachyte samples. The data are concordant and have a mean of 48.1 ± 0.8 Ma (Dalrymple and Clague, 1976).

Koko Seamount, northwest flank (76).—On Scripps Institution of Oceanography cruise AIRES VII, dredge 44 recovered pillow fragments of differentiated tholeiitic basalt from the northwest flank of Koko Seamount (Clague and Dalrymple, 1972; Clague 1974a). This lava probably erupted from a rift zone during the shield stage.

Ojin Seamount (81).—DSDP Leg 55 drilled site 430 through a lagoonal sediment pond near the center of Ojin Seamount (Jackson and others, 1980). Five lava flows were penetrated, including four flows of aphyric to sparsely porphyritic hawaiite and an underlying flow of tholeiitic basalt (Kirkpatrick and others, 1980). The overlying sediment consists of shallow-water carbonate reef or bank deposits. The flows were clearly erupted subaerially: a red soil zone was recovered between two of them. The four hawaiite flows were apparently erupted rather rapidly, because their paleomagnetic inclinations are very similar (Kono, 1980). The lowermost tholeiitic flow probably erupted during the shield stage, whereas the hawaiite flows probably erupted during an alkalic postshield stage.

Dalrymple and others (1980) obtained an age of 55.2 ± 0.7 Ma for Ojin on the basis of ^{40}Ar-^{39}Ar incremental-heating results from two samples of hawaiite and one sample of tholeiitic basalt recovered during drilling of DSDP site 430.

Jingu Seamount (83).—A Hawaii Institute of Geophysics cruise in July 1977 recovered several fresh samples and abundant moderately altered samples of hawaiite and mugearite (Dalrymple and Garcia, 1980) that probably erupted during an alkalic postshield stage.

Dalrymple and Garcia (1980) reported an age of 55.4±0.9 Ma for Jingu based on ^{40}Ar-^{39}Ar incremental-heating experiments on three of these dredged samples of hawaiite and mugearite.

Nintoku Seamount (86).—DSDP Leg 55 drilled site 432 into a lagoonal sediment pond on the top of Nintoku Seamount (Jackson and others, 1980). Samples of three lava flows were recovered from beneath sandstone, conglomerate, and a thin red clay horizon. The flows are all alkalic lava. The top two flows are identical feldspar-porphyritic alkalic basalt, and the bottom flow is transitional between alkalic basalt and hawaiite. All three flows probably erupted during the alkalic postshield stage. As on Ojin Seamount, these flows were clearly erupted subaerially.

Dalrymple and others (1980) obtained ^{40}Ar-^{39}Ar data from two samples recovered during drilling of DSDP site 432. Only one of the samples gave easily interpretable results, and that one indicated an age of 56.2±0.6 Ma.

Yomei Seamount (88).—DSDP Leg 55 drilled two holes at site 431 on a faulted terrace (Jackson and others, 1980). Neither hole reached volcanic basement. The upper 7.5 m consisted of fragments of manganese-oxide crust, authigenic silicates, phosphate, ice-rafted pebbles, and calcareous sand of Quaternary age. The lower 9.5 m consisted of authigenic silicates, manganese-oxide crust fragments, altered basalt clasts, and calcareous sand of middle Eocene age.

Suiko Seamount, southern part (90).—Saito and Ozima (1975, 1977) obtained a ^{40}Ar-^{39}Ar incremental-heating isochron age of 59.6±0.6 Ma for a single sample of mugearite dredged from the southern part of Suiko. The reliability of this age has been questioned, however, on the basis of (1) selection of the sample from a variety of ice-rafted material dredged from Suiko and (2) the unorthodox and potentially misleading treatment of the ^{40}Ar-^{39}Ar data (Dalrymple and others, 1980). Three conventional K-Ar determinations ranging from 22 Ma to 43 Ma on samples from the same dredged material (Ozima and others, 1970) are unreliable because of severe sample alteration. The sample of mugearite could represent lava of an alkalic postshield stage; however, the presence of abundant ice-rafted material (Ozima and others, 1970) creates obvious difficulties in identifying an indigenous sample from among the erratics.

Suiko Seamount, central part (91).—DSDP Leg 55 drilled a deep reentry hole (433C) in a lagoonal sediment pond (Jackson and others, 1980) on top of Suiko Seamount. The hole penetrated 550.5 m, the lower 387.5 m entirely in basalt. Samples of more than 100 flows or flow lobes were recovered, of which the upper three flow units are alkalic basalt and the remainder are tholeiitic basalt and picritic tholeiitic basalt. The three alkalic flows probably erupted during a postshield stage, whereas the thick sequence of tholeiitic lava represents the shield stage.

Dalrymple and others (1980) determined an age of 64.7±1.1 Ma for two samples of alkalic and tholeiitic basalt recovered during drilling of DSDP site 433C. The data were obtained by ^{40}Ar-^{39}Ar incremental heating.

Tenji Seamount (98).—A single dredge was obtained from Tenji Seamount by the U.S. Coast Guard Cutter *Glacier* in September 1971 (Bargar and others, 1975). The small group of rocks recovered included samples of basalt, crystal tuff, volcaniclastic sandstone, mudstone, graywacke, and a manganese nodule. Some of the lava samples could be derived from the seamount, but the rest are clearly glacial erratics. None of the samples was dated because of the uncertainty of their origin.

Meiji Seamount (108).—DSDP Leg 19 drilled site 192 on top of Meiji Seamount. A thickness of 13 m of pillow basalt with glassy margins was recovered; the rocks are highly altered, but interpretation of the immobile trace elements suggests that they are tholeiitic basalt erupted during the shield stage (Dalrymple and others, 1980b).

The only radiometric data available for Meiji is a minimum age of 61.9±5.0 Ma for highly altered basalt recovered during drilling of DSDP site 192 (Dalrymple and others, 1980b). This age is considerably less than the 70–68 Ma for overlying sediments based on nannoflora (Worsley, 1973).

THE HAWAIIAN-EMPEROR VOLCANIC CHAIN
Part II
Stratigraphic Framework of Volcanic Rocks of the Hawaiian Islands

By Virginia A.M. Langenheim *and* David A. Clague

ABSTRACT

Stratigraphy is an important tool for understanding the geologic history of the volcanoes of the Hawaiian Islands, providing a framework for much information from other geologic and related fields. Three major eruptive stages in a Hawaiian volcano's life—shield stage (tholeiitic), postshield stage (alkalic), and rejuvenated stage (alkalic)—have generally provided a basis for dividing the volcanic rocks into stratigraphic units. Such units are basic to stratigraphy, and suitable nomenclature for them helps promote unambiguous scientific communication regarding the spatial and temporal relations of rocks. The stratigraphic nomenclature of the Hawaiian Islands is herein reviewed and updated to reflect current scientific needs and to be consistent with the most recent (1983) North American Stratigraphic Code.

The major divisions of volcanic rocks on each island formerly called "Volcanic Series" are all considered to be of formational rank and renamed accordingly. Their names reflect either a predominant commonly accepted lithologic type (such as "Basalt") or the variety of volcanic lithologies in the unit (those units are called "Volcanics"). Only those subdivisions of the major units that are currently considered to be useful as formally named units of member or lesser rank are retained; others are considered to be informal. Principal and other reference localities are designated for those well-established units for which a type locality was not previously specified.

We give in tabular form a brief summary of each stratigraphic unit, including its lithology, occurrence, thickness, type and reference localities, stratigraphic relations, age, and any stratigraphic changes made herein.

INTRODUCTION

Stratigraphy is an important tool for understanding the geologic history of the Hawaiian Islands, providing a framework into which much of the scientific information contributed by other geologic and related fields can be fit in an organized manner.

The purpose of this paper is to present a brief summary of the volcanic stratigraphy of the islands, and, because the discrimination of stratigraphic units is basic to stratigraphy, to review the stratigraphic nomenclature for these rocks and update it as far as possible to reflect current scientific needs and to conform to the most recent North American Stratigraphic Code (North American Commission on Stratigraphic Nomenclature, 1983). In order to present a better view of the stratigraphy of the volcanic rocks of the islands, we begin with a short discussion of the geologic setting.

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GEOLOGIC SETTING

The Hawaiian Islands consist of a chain of volcanoes that stretches about 2,700 km (1,700 mi) across the northern Pacific Ocean in a northwesterly direction from the Island of Hawaii to Kure Island (fig. 1.20). The principal (so-called Windward) Hawaiian Islands of Hawaii, Maui, Kahoolawe, Lanai, Molokai, Oahu, Kauai, Niihau, and Kaula lie at the southeastern end of the chain. All of these islands are formed by large volcanoes, though Kaula is only a small crescent-shaped erosional remnant of a tuff cone, presumably resting on a large submerged volcano. Some of the islands are formed by a single volcano, others by two or more coalesced volcanoes. The subaerial part of these volcanoes, which constitutes only a small fraction of the total mass of each volcano, is typically shield shaped. In older volcanoes, this shield shape is largely modified by erosion. The rocks of the southeastern islands are almost entirely of volcanic origin; only minor amounts of sedimentary rocks occur. Northwest of the Windward Islands lie the so-called Leeward Hawaiian Islands (fig. 1.20), which consist of small volcanic islets and atolls; only few of the volcanoes there rise above the sea.

The age of the volcanoes increases progressively from the southeast end, where the volcanoes are still active, to the northwest end, where the volcanoes are about 30 Ma. Most of the volcanoes have been extinct for millions of years. The only historical eruptions have been at East Maui (Haleakala) Volcano on the Island of Maui, and Hualalai, Mauna Loa, and Kilauea Volcanoes on the Island of Hawaii; Mauna Loa and Kilauea are frequently active.

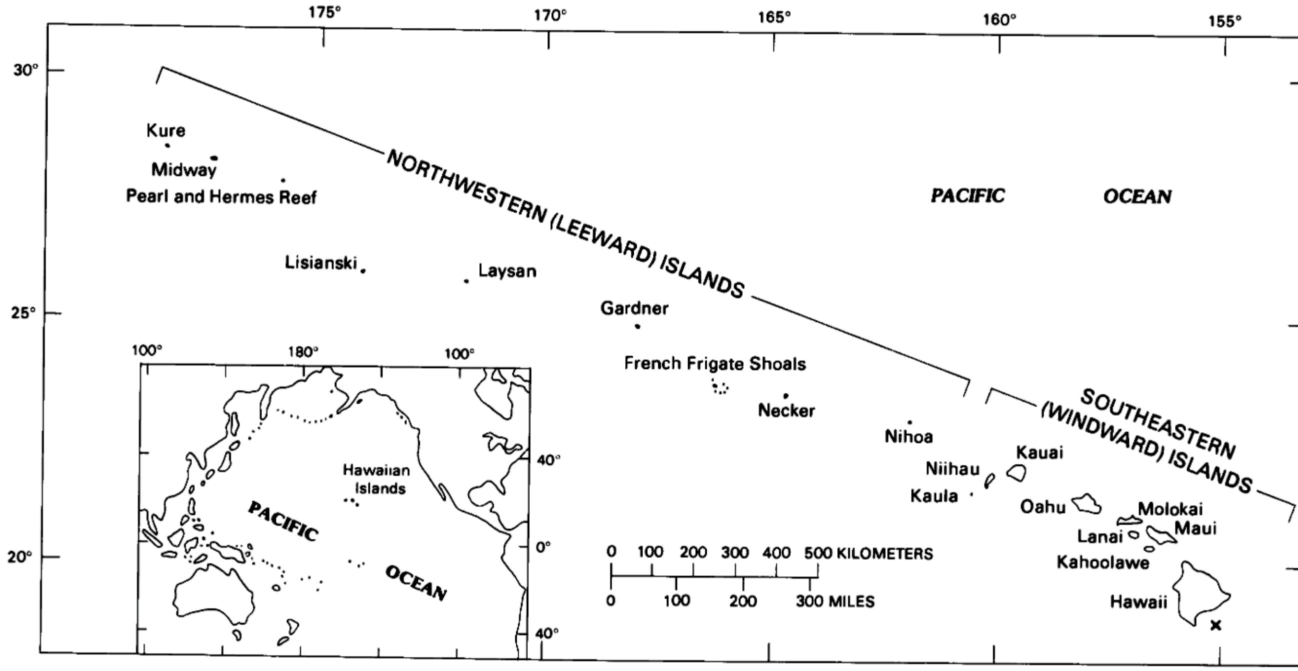

FIGURE 1.20.—Index map of Hawaiian Islands. x, Loihi Seamount.

Hawaiian volcanoes go through four major eruptive stages—preshield, shield, postshield, and rejuvenated stages—in their evolution and growth, and each stage is represented by rocks of distinct chemical and mineralogical composition. The stages used in this report are from Clague and Dalrymple (chapter 1, part I); see also Peterson and Moore (chapter 7) for discussion of alternative nomenclature. All but the preshield stage are well studied and documented (Stearns, 1940b; Macdonald and Katsura, 1964; Macdonald, 1968). Although an individual volcano may become extinct before all the stages are complete, the general sequence of stages is typical of well studied Hawaiian volcanoes.

Very little is known about the preshield stage, which includes the earliest phase of submarine activity and, in the one known example (Loihi Seamount, fig. 1.20), apparently consists of alkalic basalt and basanite (Moore and others, 1982). The main volcanic edifice (more than 95 percent of the total volume of the volcano) is constructed, in perhaps a million years or less, by voluminous eruptions of silica-rich tholeiitic basalt (and rare rhyodacite and icelandite) during the shield stage. The shield stage, as used here, includes the submarine eruption of tholeiitic basalt, which precedes the subaerial shield-forming eruptions. During the shield stage or at the beginning of the next stage, the postshield stage, a caldera may form and be filled with tholeiitic and (or) alkalic basalt. This process of caldera formation and filling was previously referred to as the caldera-filling or caldera-collapse stage, but is here considered to be a phase of the shield stage or postshield stage. During the postshield stage, a relatively thin cap of alkalic basalt and associated differentiated lava (ankaramite, hawaiite, mugearite, benmoreite, and trachyte) covers the main shield. This alkalic lava makes up less than 1 percent of the total volume of the volcano. Later, after a relatively long period of volcanic quiescence and erosion, a very small amount of silica-poor lava (alkalic basalt, basanite, nephelinite, and nepheline melilitite; rare ankaramite and hawaiite known on East Maui Volcano only) erupts from isolated vents; this stage is referred to here as the rejuvenated stage.

During the shield and postshield stages, eruptions are not confined to the summit area of a volcano, but also occur along extensive zones of fissures, called rift zones, that extend down the flanks of the volcano. These zones are usually marked at the surface by collapse craters, cinder and spatter cones, and grabens, and below the surface by large numbers of dikes. During the rejuvenated stage, however, eruptions occur at vents unassociated with the preexisting rift zones, with the exception of rejuvenated-stage lava on East Maui Volcano.

The shield stage is represented by lava that is dark colored and forms relatively long and thin flows. Most of these flows have sparse to abundant olivine phenocrysts, and some have abundant large phenocrysts of plagioclase. Pyroclastic deposits are minor components during the shield stage.

The postshield stage is represented by lava that is commonly lighter colored than shield-stage lava and contains pyroxene, olivine, and plagioclase phenocrysts. This lava forms shorter and thicker flows. Pyroclastic deposits are more abundant during the postshield stage. There is little evidence of erosion between the eruption of shield-stage lava and postshield-stage lava.

The lava that ponds in a caldera during the shield stage or at the beginning of the postshield stage is usually massive and may show columnar structure.

Rejuvenated-stage lava is also dark colored, but usually forms thick flows with few or no phenocrysts; pyroclastic deposits are common during this stage.

The shield, postshield, and rejuvenated stages have generally formed the basis for the major stratigraphic divisions or subdivisions of the rocks of the Hawaiian volcanoes (fig. 1.21).

PREVIOUS STUDIES

The geology of the Hawaiian Islands has been studied for more than a hundred years (see Clague and Dalrymple, chapter 1, part I, and Peterson and Moore, chapter 7, for summaries of geologic investigations). Early geologic investigations concentrated on the main southeastern islands and were mainly reconnaissance and general descriptive studies. These early studies were superseded by detailed mapping of the eight major southeastern islands beginning in the 1930's and continuing for nearly 30 years. H.T. Stearns and his coworkers, in a series of bulletins published by the Hawaii Division of Hydrography, produced detailed geologic maps (at scale of 1:62,500, except Hawaii at 1:125,000) and descriptions of the Islands of Oahu (Stearns and Vaksvik, 1935; Stearns, 1939, 1940a), Lanai and Kahoolawe (Stearns, 1940b), Maui (Stearns and Macdonald, 1942), Hawaii (Stearns and Macdonald, 1946), Niihau (Stearns, 1947b), Molokai (Stearns and Macdonald, 1947), and Kauai (Macdonald and others, 1960; Stearns was not a coauthor of this report, but did much of the geologic mapping).

These bulletins have provided the basic stratigraphic framework for subsequent petrologic, mineralogic, geochemical, and geophysical investigations of the islands. Only a few stratigraphic studies have been done since publication of the bulletins, and those studies and detailed geologic mapping have been mainly on the Island of Hawaii.

STRATIGRAPHIC NOMENCLATURE

The classification and naming of stratigraphic units (used here as synonomous with lithostratigraphic units), although to some extent arbitrary and artificial, helps promote concise and unambiguous scientific communication as to the spatial and temporal relations of rocks. Stratigraphic units can be formally or informally named; both are useful in stratigraphic work. Formally named units, however, are named and defined in accordance with procedures outlined in the Code, and any major changes made to them, such as boundary or rank changes, need to be justified.

In the Hawaiian Islands, formal stratigraphic names have been applied only to rock units on the eight main southeastern Hawaiian Islands (Hawaii, Maui, Kahoolawe, Lanai, Molokai, Oahu, Kauai, and Niihau), and these names are almost entirely restricted to volcanic rocks and the sedimentary rocks closely associated with them. Other sedimentary units, such as beach and reef deposits, have been formally named, but they are not discussed here.

Stearns and his coworkers divided the rocks of each volcano into one or more major units that they formally named as "Volcanic Series" (see remarks column in table 1.12). These "Series" consisted of rocks resulting from a succession of extensive eruptions. Some of the "Series" consisted entirely of volcanic rocks, whereas others consisted of volcanic rocks, related intrusions, and their weathering products. Some of the "Series" were not subdivided, whereas others were subdivided either into formations or into members. Although the use of "Volcanic Series" as part of a formal name conformed to the "stratigraphic code" in use at the time the unit was named (Ashley and others, 1933), it does not conform to later codes, including the present (1983) code (American Commission on Stratigraphic Nomenclature, 1961, 1970; North American Commission on Stratigraphic Nomenclature, 1983), which restricts the use of the term "Series" to chronostratigraphic units.

The above inconsistencies, plus recent petrologic, mineralogical, and geochemical studies, have necessitated changes in the stratigraphic nomenclature for the volcanic rocks of the Hawaiian Islands[1] because (1) the lithic or descriptive terms of the names do not conform to the current stratigraphic code or do not reflect modern petrologic classification; (2) the ranks of the units are unclear or inappropriate; and (3) formalization of some names is not now considered to be necessary or useful.

The basic guidelines used in this report for updating the stratigraphic usage in the Hawaiian Islands are those recommended by the 1983 North American Stratigraphic Code (North American Commission of Stratigraphic Nomenclature, 1983). Any changes to the stratigraphic nomenclature made in this report, such as revision (change of rank or lower or upper boundary), redefinition (change of name), and abandonment of formally named stratigraphic units, have been approached as follows:

(1) Formal names are retained only for units that serve a useful purpose and require the stability of nomenclature that formalization affords. As stratigraphic units of any rank become established by repeated demonstration of their usefulness, those formal names that have not been used for a few decades or are currently not thought to serve a useful purpose are abandoned as formal names, but the geographic term of their name may be used informally to identify particular flows (and their associated cones) or beds.

(2) Major stratigraphic units on each volcano are all considered to be of formational rank, and are all formally named units. The ranks of stratigraphic units are important in that they give some concept of the scale of the units in relation to other units. Some of these units could have been considered to be units of group rank, but this does not seem necessary. The volcanic products of one volcano (except for some ash deposits) do not occur on other volcanoes, although they may overlap or interfinger at their boundaries where two or more volcanoes coalesce. Therefore, the units of a volcano can be considered to be essentially confined to one large "mountain," and group rank does not seem appropriate. This approach would also give some consistency to the stratigraphic nomenclature of all the islands.

(3) Formally named subdivisions of the major stratigraphic units are those units of member or lesser rank that are distinctive and (or) extensive. Many of the major units of the Hawaiian volcanoes are very often difficult to subdivide because of rapid lateral changes and lack of key beds.

[1]G.A. Macdonald had planned to revise the stratigraphic nomenclature of the Hawaiian Islands (written commun. to R.W. Kopf, 1976, 1977) before his untimely death in 1978.

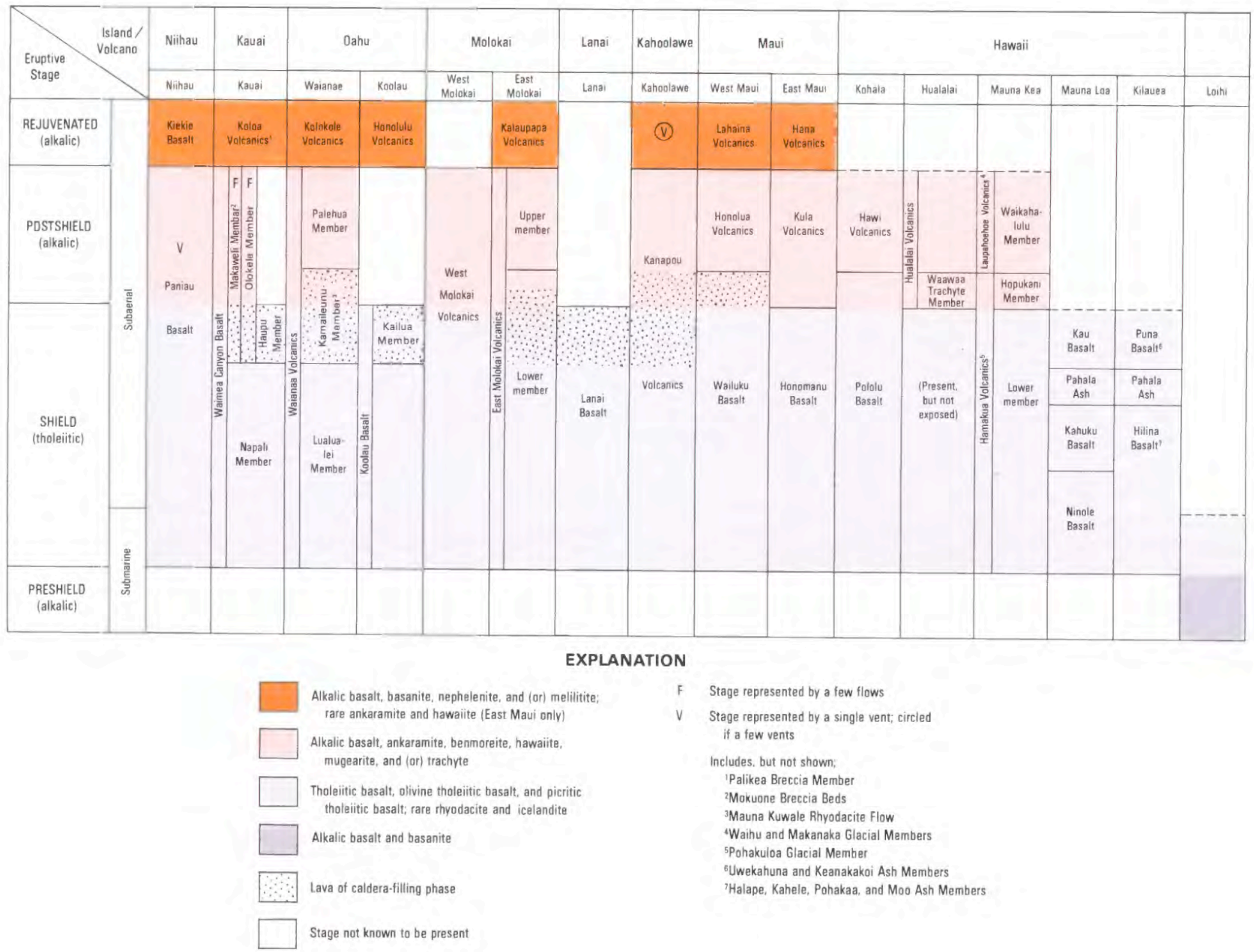

FIGURE 1.21.—Chart showing relations of eruptive stages to volcanoes and stratigraphic units, southeastern Hawaiian Islands, arranged approximately from northwest (left) to southeast (right) (see fig. 1.20 for location). No age correlation between units on different volcanoes implied. Dashed lines indicate end of stage not implied. Base of oldest unit on each volcano (except Loihi) assumed to be at the base of the shield stage.

(4) Formal names of the major stratigraphic units consist of a unique geographic term followed by a simple and generally accepted lithic term, such as "Basalt," or by the term "Volcanics." The initial letter of each term is capitalized. "Basalt" is used for units that consist entirely or almost entirely of basalt, whether tholeiitic or alkalic. An example is the Puna Basalt. The general term "Volcanics" (equivalent to the term "Formation") is used for units that do not have a predominant rock type or are composed of rock types such as hawaiite, mugearite, trachyte and basanite. Examples are the Lahaina Volcanics, Hamakua Volcanics, and Honolulu Volcanics (table 1.12).

The term "Formation" was not used because it does not convey that the unit is composed of volcanic rocks. The term "Volcanic Complex," which indicates a diverse assemblage of extrusive volcanic rocks, related intrusions, and their weathering products, could have been used for some units. However, we tentatively have opted for the term "Volcanics," because it is a shorter term, has precedence of use in the Hawaiian Islands, and does not introduce a new term until the need is confirmed by future investigations.

In general, the term "Basalt" is used for units that represent all or part of the shield stage, although a minor part of the postshield stage may also be represented. The term "Volcanics" is used for units that represent all or part of the postshield stage and for units that represent both shield stage and postshield stage. All units that represent the rejuvenated stage, with the exception of the Kiekie Basalt of Niihau Volcano, are termed "Volcanics."

(5) Formal names of subdivisions of the major units consist of a unique geographic term followed by the appropriate rank term—"Member," "Flow(s)," or "Bed(s)." An intervening lithic or descriptive term may be used in member names and is used in bed and flow names. The initial letter of each term is capitalized. Examples include the Napali Member, Kahele Ash Member, Makanaka Glacial Member, Mokuone Breccia Beds, and Mauna Kuwale Rhyodacite Flow.

Members, flows, and beds can also be informally named. Informal members are designated solely by their lithology or by their stratigraphic position, and none of the initial letters of their names is capitalized. Examples are ash member and lower member. The terms "historic member" and "prehistoric member" are not used in this report because age should play no part in differentiating lithostratigraphic units. Informal flows and beds are similarly designated, such as lower flow, upper flows, ash bed, and breccia beds, but may also combine a geographic term with the unit term or terms; only the initial letter of the geographic term is capitalized. Examples are the Kona ash beds, Kilea flow, and Makapipi flows. With one exception (see Waianae Volcano), all named flows in the Hawaiian Islands are considered to be informal in this report.

(6) A principal reference locality is herein designated for some well established units for which a type locality was never specified. For units that previously had more than one type locality specified, the one here considered to be the more accessible is retained as the type, and the other localities are redesignated as reference localities.

A designated stratotype (type section or type locality) is essential in the definition of a formal stratigraphic unit because it serves as the standard for the unit and constitutes the basis for its recognition. It should therefore be representative of the concept of the unit. Because of the nature of Hawaiian volcanic activity, the stratotype for many of the Hawaiian volcanic units is not truly representative of those units, and reference sections or localities become invaluable in illustrating the lithologic diversity within a unit or the stratigraphic relations with other units. We have not attempted to specify additional reference sections or localities in this report, but we hope this will be done in future stratigraphic studies.

STRATIGRAPHIC SUMMARY

The stratigraphy of the eight main Hawaiian Islands is briefly discussed below by island and volcano from southeast to northwest.

A summary of the formally named stratigraphic units used in this report is given in table 1.12, including a brief description of the lithology, occurrence, thickness, type and reference localities, stratigraphic relations, and age of each. The descriptions of the units were largely taken from the sources cited in the remarks column. Informal units are listed if they were previously formally named. Type localities are used throughout the table, even though some of the units have specified type sections.

Almost all of the isotopic ages shown in the table are K-Ar ages. Numerous radiocarbon ages have been determined for the youngest volcanoes but, with a few exceptions, are not shown. For a summary of radiocarbon ages for the Island of Hawaii, see Rubin and others (chapter 10). The relative ages assigned to the stratigraphic units are based on the Decade of North American Geology time scale (Palmer, 1983), which has the following epoch boundaries: Miocene-Pliocene, 5.3 Ma; Pliocene-Pleistocene, 1.6 Ma; and Pleistocene-Holocene, 10 ka. A correlation diagram of the major stratigraphic units is presented as figure 1.22.

Stratigraphic changes that are made in this report, using the guidelines of the Code and the approach discussed previously, are indicated in the remarks column and discussed below. Former names that have been applied to the units are also indicated in the remarks column. We have not used glottal stops in the geographic part of lithostratigraphic-unit names, although some authors have used them in other publications.

HAWAII

The Island of Hawaii, the largest of the Hawaiian Islands, consists of five coalesced volcanoes: Kilauea, Mauna Loa, Mauna Kea, Hualalai, and Kohala (fig. 1.23). Mauna Kea is the highest of these volcanoes; Mauna Loa is the largest by volume. Little erosion has occurred on these volcanoes except on the northeastern sides of Kohala and Mauna Kea.

KILAUEA VOLCANO

Kilauea, the youngest volcano of the island and still very active, consists entirely of shield-stage tholeiitic lava that issued from the summit caldera and the east and southwest rift zones. The rocks of the volcano are divided into the Hilina Basalt (older) and the

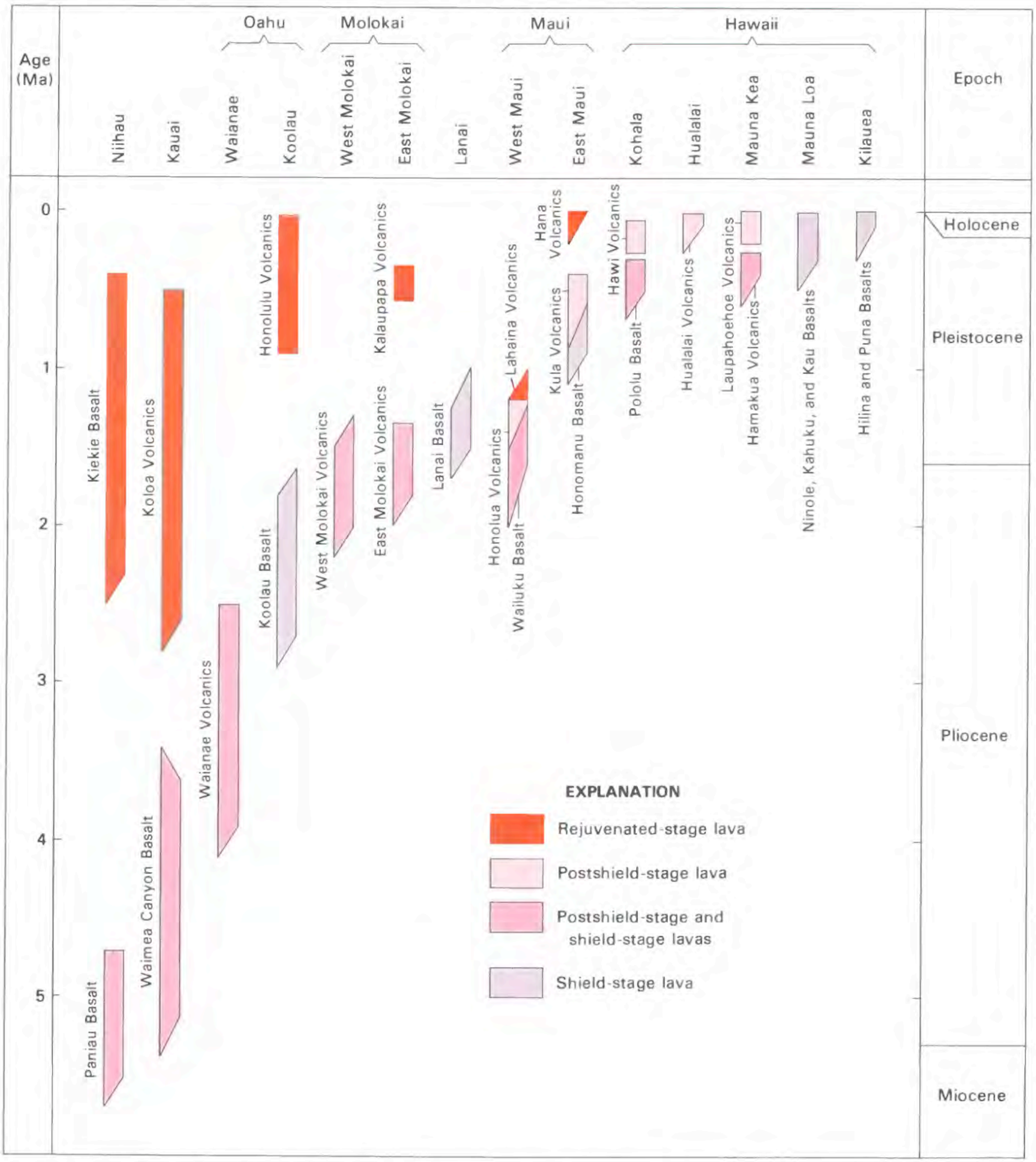

FIGURE 1.22.—Correlation diagram of isotopically dated major volcanic stratigraphic units of southeastern Hawaiian volcanoes. Modified from Clague and Dalrymple (chapter 1, part I, fig. 1.6). Diagonal lines indicate overlapping or uncertain ages. Epoch boundaries from Palmer (1983).

Puna Basalt (younger), generally separated from one another by the Pahala Ash (see subsection "Mauna Loa Volcano"; see also Easton, chapter 11). The Hilina Basalt is exposed only in fault scarps located along the south flank of the volcano. The Hilina Basalt includes the Halape (oldest), Kahele, Pohakaa, and Moo (youngest) Ash Members (Easton, chapter 11). The Puna Basalt covers almost the entire surface of Kilauea and essentially consists of all post-Pahala lava (see discussion below). The Puna includes the prehistoric Uwekahuna (older) and historical Keanakakoi (younger) Ash Members.

The ages of the Hilina and Puna Basalts are not well known. The Hilina Basalt is probably older than the approximately 31-ka

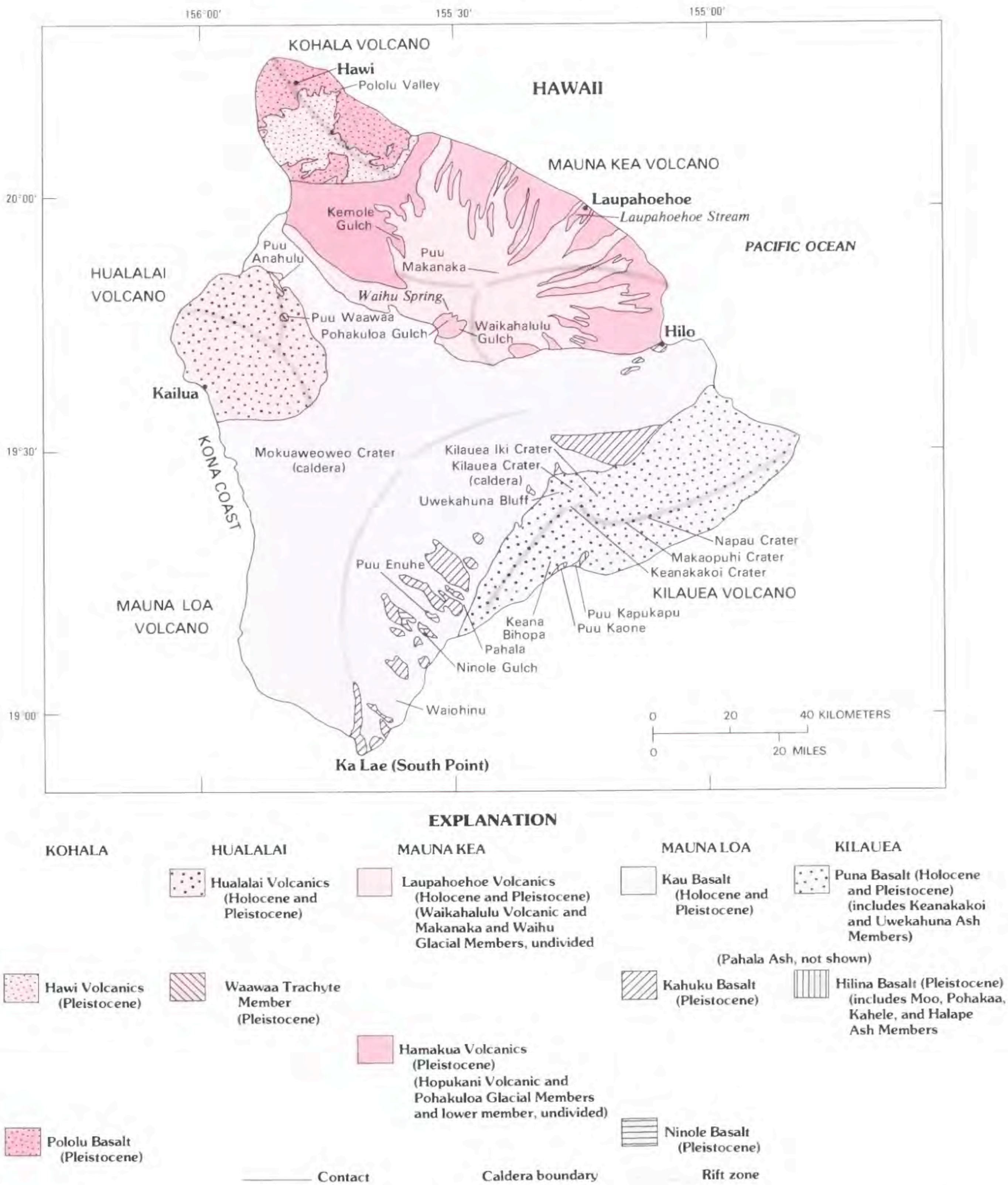

FIGURE 1.23.—Generalized geologic map of the Island of Hawaii, showing localities mentioned in text. Modified from Macdonald and others (1983), Easton (chapter 11), and Porter (1979a, 1979b). Rift zones from Fiske and Jackson (1972). Caldera boundary of Mauna Kea Volcano is inferred to be buried beneath younger lava (see Porter, 1972).

radiocarbon age obtained from the base of the Pahala Ash on Mauna Loa, and the oldest exposed Hilina flows are estimated to be about 100 ka (Easton, chapter 11). Most of the Puna Basalt was erupted during the last 10,000 years, but a Puna flow intercalated with the Pahala Ash at Puu Kaone was radiocarbon dated at about 22 ka (see Easton, chapter 11). The Uwekahuna Ash Member of the Puna Basalt has radiocarbon ages ranging from 2.17 ka to 1.04 ka; however, the age of the eruption or eruptions is uncertain. The ash could have been deposited about 2.1 ka (Casadevall and Dzurisin, chapter 13; Lockwood and Rubin, 1986) or about 1.5 ka (Holcomb, chapter 12).

MAUNA LOA VOLCANO

The lava of Mauna Loa, like that of Kilauea, is all shield-stage tholeiitic lava that has mostly issued from the summit caldera and southwest and northeast rift zones.

The oldest exposed rocks belong to the Ninole Basalt, which forms a series of steep-sided hills on the southeast flank of the volcano. The Ninole, which was originally named the Ninole Basalt by Stearns (1926), was later called the Ninole Volcanic Series by Stearns and Macdonald (1946). Recently, Lipman (1980) used the name Ninole Volcanics. As the Ninole is all tholeiitic basalt, the term used by Stearns (1926) is more informative and is used here. The Ninole was thought to represent remnants of an earlier shield volcano mostly buried by lavas from Mauna Loa (Stearns and Macdonald, 1946), but recent studies indicate that it may merely be deeper parts of the Mauna Loa shield, no more than a few hundred thousand years old, that have been uplifted along normal faults (Lipman, 1980).

Unconformably overlying the Ninole are the Kahuku Basalt (older) and Kau Basalt (younger) (table 1.12), which are separated by the Pahala Ash. Lipman (1980) recently used the terms Kahuku Volcanics and Kau Volcanics, but as these units are composed entirely of basalt, they are here renamed the Kahuku Basalt and Kau Basalt, respectively. The Kahuku Basalt crops out on the east and south sides of the volcano. The Kau Basalt covers most of the surface of Mauna Loa and consists of all post-Pahala lava.

The Pahala Ash is a distinctive yellowish vitric ash that has been largely altered by weathering to a reddish brown mixture of clay minerals and hydrated oxides. The thickness of the unit varies considerably, but it is locally more than 15 m thick (Easton, chapter 11).

These ash deposits were originally the uppermost part of what Noble and Clark (in Washington, 1923) called the Pahala Series, which also included the underlying flows. Stone (1926) used the term Pre-Kilauea Series for the uppermost ash and underlying flows, and restricted the name Pahala to the uppermost ash deposits, calling them the Pahala Ash. He applied the name Pahala Ash to ash deposits of Mauna Loa and Kilauea, but not to those of Mauna Kea. Stearns and Clark (1930) used the name Pahala Basalt for both the uppermost ash deposits and the underlying flows. The flows of their Pahala Basalt are herein called the Kahuku Basalt on Mauna Loa and the Hilina Basalt on Kilauea. Wentworth (1938) used the term Pahala Tuff in the same general sense as the Pahala Ash of Stone (1926), but also used the term Waiau Formation (recently abandoned by Porter, 1973) for equivalent ash on Mauna Kea and the term Glenwood Tuff for equivalent ash on parts of the east slope of Mauna Loa. Stearns and Macdonald (1946) and Davis and Macdonald (in Avias and others, 1956), however, considered the Pahala Ash to be a persistent ash formation that was derived from several sources. They believed the Pahala occurred on all the volcanoes, separating their major units on Kilauea (Hilina and Puna Volcanic Series), Mauna Loa (Kahuku and Kau Volcanic Series), and Mauna Kea (Hamakua and Laupahoehoe Volcanic Series) and capping the older unit (Pololu Volcanic Series) of Kohala and a member (Waawaa Volcanics) of the Hualalai Volcanic Series of Hualalai.

Current mapping on Mauna Kea, Hualalai, and Kohala indicates that the extensive surficial ash deposits on these volcanoes were locally derived and are genetically distinct from the Pahala Ash on Kilauea and Mauna Loa (E.W. Wolfe, oral commun., 1986). Therefore, we are geographically limiting the Pahala to the occurrences on Kilauea and Mauna Loa.

The age of the Ninole Basalt is considered to be no more than a few hundred thousand years. The single published K-Ar age of 0.54 Ma has an estimated uncertainty of 0.4 m.y. (see Clague and Dalrymple, chapter 1, part I). The Kahuku Basalt and the Kau Basalt are approximately coeval with the Hilina Basalt and Puna Basalt, respectively, of Kilauea. The Pahala Ash on Mauna Loa and Kilauea is considered by Easton (chapter 11) to be entirely of Pleistocene age (between about 30 ka and 10 ka).

MAUNA KEA VOLCANO

Mauna Kea last erupted about 3.6 ka (Porter, 1979a). The volcano passed through the shield stage into the postshield stage and produced a cap of differentiated lava that almost completely buried the original subaerial shield. It is uncertain if a summit caldera existed; however, a small caldera has been inferred to lie buried beneath the younger lava (Porter, 1972). Rift zones are less pronounced than on Kilauea and Mauna Loa, but westerly, southerly, and easterly rifts are suggested by alignments of cinder cones. The lower northeastern slope of the volcano has gulches cut into it, whereas the other slopes are generally little affected by erosion. The upper slopes of the volcano were glaciated during the Pleistocene.

The rocks of the volcano were divided into the Hamakua Volcanic Series and the overlying Laupahoehoe Volcanic Series by Stearns and Macdonald (1946). The Hamakua represents the shield stage and part of the postshield stage; the Laupahoehoe represents the rest of the postshield stage. These units, considered to be of formational rank by Stearns and Macdonald (1946), were subdivided by them into informal members.

Recently, Porter (1979a, 1979b; see also Porter, 1973, 1974; Porter and others, 1977) formally redefined the Hamakua and Laupahoehoe to include glacial deposits, raised them to group rank, and subdivided both units into formally named volcanic and glacial formations. They subdivided the upper part of the Hamakua Group into the Hopukani Formation (volcanic) and the Pohakuloa Formation (glacial), but Porter (1979a, 1979b) called the lower part of the group the "lower member of Stearns and Macdonald (1946)." Porter and others (1977; see also Porter, 1973, 1974) originally subdivided their Laupahoehoe Group into three volcanic formations—Liloe (oldest), Hanaipoe, and Waikahalulu (youngest) Formations—and two interstratified glacial formations—Waihu and Makanaka Formations. The Makanaka included the Kemole Member, a volcanic unit. This usage was superseded by a threefold subdivision of the Laupahoehoe Group (Porter, 1979a, b), in which all of the volcanic rocks were assigned to the Waikahalulu Formation; the names Liloe, Hanaipoe, and Kemole were abandoned as formal names of lithostratigraphic units and were formally applied to chronostratigraphic units—the Liloean, Hanaipoean, and Kemolean Stages.

Because the use of the Hamakua and Laupahoehoe as units of group rank is not consistent with the approach used here for the rest of the volcanoes of the Hawaiian Islands, these units are here reduced to formation rank and renamed the Hamakua Volcanics and Laupahoehoe Volcanics. Their subdivisions are reduced to member rank and renamed to reflect the overall nature of their lithologies (see table 1.12).

The Hamakua is considered to be of Pleistocene age and the Laupahoehoe of Pleistocene and Holocene age on the basis of K-Ar and radiocarbon dating (Porter, 1979a) (table 1.12).

HUALALAI VOLCANO

Hualalai Volcano last erupted in 1800–1801, when several flows issued from the well-defined northwest rift zone. Less well defined rift zones trending north and southeast are marked by numerous cinder and spatter cones. It is not known whether the volcano had a summit caldera. Shield-stage tholeiitic lava is not exposed on Hualalai Volcano, but tholeiitic rocks are known to occur in the subsurface and in the submarine part of the volcano (Moore and others, chapter 20; Clague, 1982).

The entire subaerial surface of the volcano consists of postshield-stage alkalic basalt, with minor hawaiite and trachyte, named the Hualalai Volcanic Series by Stearns and Macdonald (1946). They included within the Hualalai a trachyte cone and flow, which they called the Waawaa Volcanics. Though they gave the Waawaa what is now considered to be a formational rank name, they clearly considered it to be a member of the Hualalai (Stearns and Macdonald, 1946, p. 143), and it is so considered here (table 1.12).

Stearns and Macdonald (1946) did not apply a name to the ash deposits that mantle the slopes of the volcano, although Wentworth (1938) had earlier called these deposits the Kona Tuff Formation. Later, Davis and Macdonald (in Avias and others, 1956) presumably included the Kona Tuff Formation within the Hualalai, effectively giving the Hualalai group status. R.B. Moore (oral commun., 1986) suggests that this unit should never have been formally named; it is here abandoned as a formal name and the unit is informally called the Kona ash beds. Thus, the Hualalai is here reduced to formational rank and renamed the Hualalai Volcanics.

The Hualalai Volcanics is largely of Holocene age, but the oldest flows are of Pleistocene age (Moore and others, chapter 20). The Waawaa Trachyte Member has a published K-Ar age of 0.4 ± 0.3 Ma (Funkhouser and others, 1968), but recent unpublished K-Ar determinations by G.B. Dalrymple (oral commun., 1986) indicate the trachyte is about 0.105 Ma.

KOHALA VOLCANO

Kohala Volcano is an oval volcano built up around northwest and southeast rift zones. It is deeply dissected on its northeast side. Arcuate faults near the summit of the volcano suggest that a caldera formed during the shield stage but was later buried by the younger lava in the postshield stage.

The volcanic rocks of Kohala Volcano were originally divided by Stearns and Macdonald (1946) into the Pololu Volcanic Series (older), composed of shield-stage tholeiitic basalt with caldera-filling postshield-stage alkalic basalt near the top, and the Hawi Volcanic Series (younger), consisting of differentiated alkalic lava of the postshield stage (table 1.12). Neither of these units has been subdivided, and they are here renamed the Pololu Basalt and Hawi Volcanics, respectively, to reflect their lithologies. The Pololu Basalt and Hawi Volcanics are of Pleistocene age on the basis of K-Ar determinations (McDougall, 1969; McDougall and Swanson, 1972) (table 1.12).

MAUI

The Island of Maui (fig. 1.24), the second largest of the Hawaiian Islands, consists of two large coalesced volcanoes, East Maui (or Haleakala) Volcano and West Maui Volcano, connected to one another by an isthmus formed when lava of East Maui banked against the already existing West Maui Volcano. The Maui volcanoes are more dissected than the volcanoes of the Island of Hawaii.

EAST MAUI (OR HALEAKALA) VOLCANO

East Maui Volcano last erupted about 200 years ago and has a large summit crater called Haleakala Crater, which is primarily of erosional origin (Macdonald and others, 1983). East Maui is the youngest Hawaiian volcano to have rejuvenated-stage lava.

The rocks of the volcano were originally divided by Stearns and Macdonald (1942) into the Honomanu Volcanic Series (oldest), Kula Volcanic Series, and Hana Volcanic Series (youngest), representing the shield, postshield, and rejuvenated stages, respectively (table 1.12). The Kula eruptions took place along southwest, east, and northwest rift zones. The Hana eruptions are unique among Hawaiian rejuvenated-stage eruptions because their vents are aligned along preexisting rift zones (southwest and northwest rift zones), the erosional period preceding these eruptions was rather short (<0.4 m.y.), and ankaramite and hawaiite are present.

The Honomanu Volcanic Series is almost completely buried by later lava and is only exposed in the seacliffs along part of the north coast. The Honomanu Volcanic Series was not subdivided by Stearns and Macdonald (1942), and it was more recently called the Honomanu Formation in the Haleakala Crater area by Macdonald

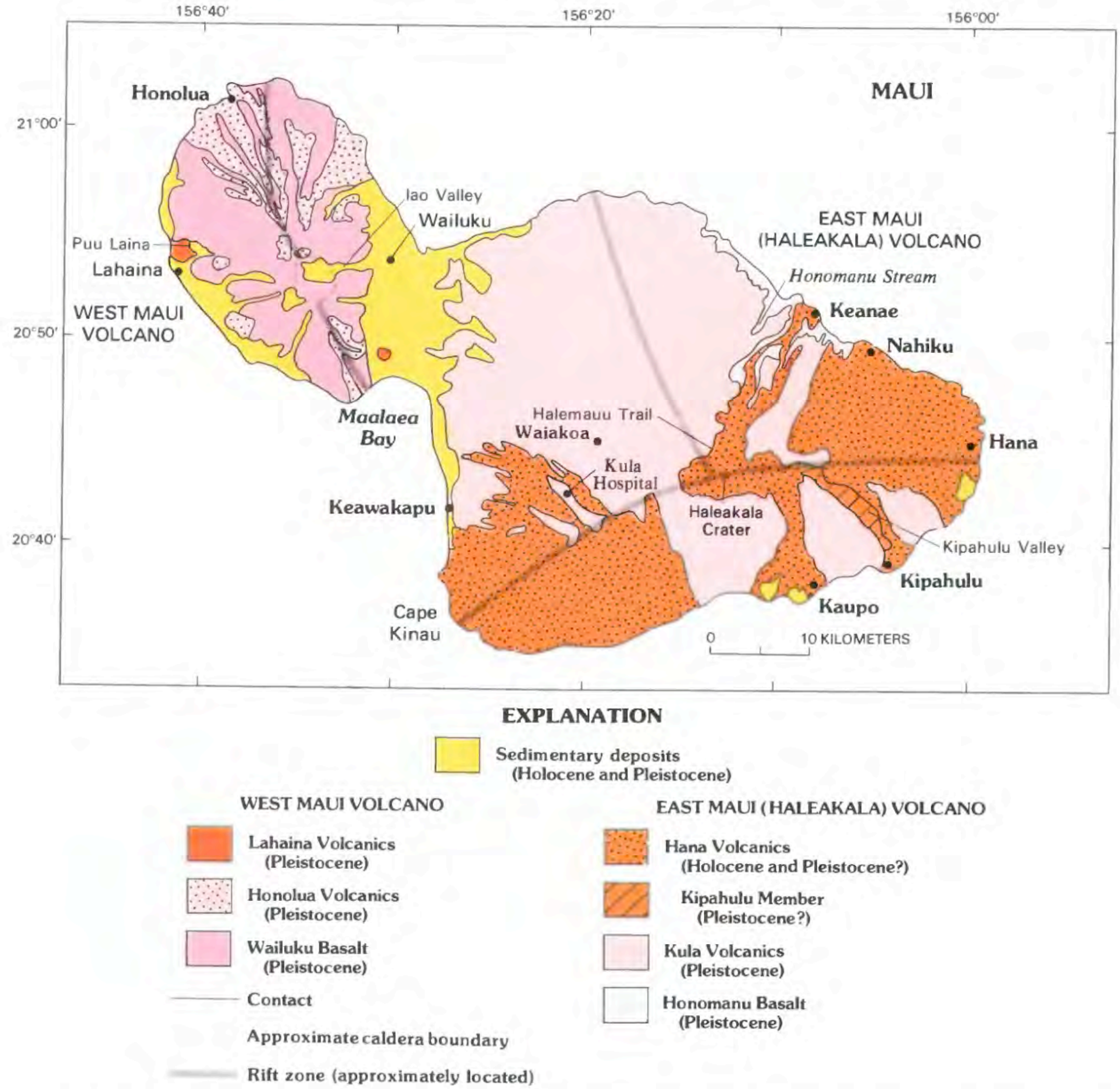

FIGURE 1.24.—Generalized geologic map of Maui, showing localities mentioned in text. Modified from Macdonald and others (1983). Rift zones from Fiske and Jackson (1972). West Maui Volcano caldera boundary from Macdonald and others (1983, fig. 20.4).

(1978). As it consists entirely of tholeiitic basalt, it is here renamed everywhere the Honomanu Basalt.

The Kula Volcanic Series of Stearns and Macdonald (1942), which was later called the Kula Formation by Macdonald (1978), is here renamed the Kula Volcanics to reflect its varied lithology. Rocks in the lower part of the south wall of Haleakala Crater that Stearns and Macdonald (1942) had originally assigned to the Honomanu were excluded from the Honomanu and named the Kumuiliahi Formation by Macdonald (1978). As these rocks appear more properly to belong to the Kula (Macdonald and others, 1983, p. 391), they are here tentatively included in the Kula, and the name Kumuiliahi Formation is not used.

The Kula was named by Stearns (1942) for the settlement of Kula (now called Waiakoa) along the roads leading to the Kula Sanatorium or Hospital on the west slope of the volcano, but he did not designate a type locality. Macdonald and Davis (in Avias and others, 1956), however, did specify the type locality as "Kula, a district on the west slope of East Maui Mountain." This "locality" is here considered to be along Highway 37 near Waiakoa (Kula Post Office), about 7 km northeast of Kula Hospital. Other accessible sections of the Kula, according to Stearns, are along the Kaupo-Kipahulu road near Kipahulu and along Halemauu Trail in the Haleakala Crater area; these are here considered to be reference localities.

Stearns and Macdonald (1942) subdivided the Hana Volcanic Series into many local formally named units in the Keanae and Nahiku areas along the northeast coast. It was not clear whether they intended these units to be of member or formation rank (see Stearns and Macdonald, 1942, p. 94, 95), but they applied formational-rank names to these units. Macdonald and Davis (in Avias and others, 1956) considered them to be formations. In Kipahulu Valley in the southeastern part of the island, however, Stearns and Macdonald (1942) included only the Kipahulu Member within the Hana. The Kipahulu was later raised in rank by Macdonald and Davis (in Avias and others, 1956) to Kipahulu Formation, though in later publications by Macdonald and Abbott (1970), Macdonald and others (1983), and Stearns (1985) the term Kipahulu Member is used. Thus the Hana had previously been considered to be of group rank. Macdonald (1978) did reduce the rank of the Hana in the Haleakala Crater area, using the name Hana Formation, but he did not address the status of the formational rank units in the Keanae and Nahiku areas.

Therefore, the following changes are here made to the Hana: The Hana is reduced to formational rank everywhere and renamed the Hana Volcanics. Its subdivisions in the Keanae and Nahiku areas, being mainly single flows of very limited extent, are abandoned as formally named units and are used informally (table 1.12). The Kipahulu is retained as a member of the Hana because it is a fairly extensive unit and is separated from earlier Hana flows by an erosional unconformity. The historical lava flow of about 1790 in the Cape Kinau area (Oostdam, 1965) is considered to be part of the Hana, although it was not included in the Hana by Stearns and Macdonald (1942), and they gave "1750?" as the date of the flow. Finally, to correct an error in Keroher and others (1966, p. 1976), the Hana does not include the Kaupo Mudflow.

The Pleistocene age of the Honomanu Basalt and Kula Volcanics is well documented by K-Ar determinations (Naughton and others, 1980) (table 1.12). No isotopic ages have been determined for the Hana Volcanics, but it is here considered to be Pleistocene(?) and Holocene. The Kipahulu Member is probably Pleistocene, judging by its relations to older units.

WEST MAUI VOLCANO

West Maui Volcano is incised by deep valleys and is considered to be extinct. Lava was erupted from a small central caldera and from the north and southeast rift zones.

The rocks of West Maui Volcano were divided by Stearns and Macdonald (1942) into three major units—Wailuku Volcanic Series (oldest), Honolua Volcanic Series, and Lahaina Volcanic Series (youngest). These units represent the shield stage and postshield caldera-filling phase, the postshield stage, and the rejuvenated stage, respectively.

Stearns and Macdonald did not subdivide the Wailuku and Honolua, which are here renamed the Wailuku Basalt and Honolua Volcanics, respectively. They did, however, include two formational-rank units, the Kilea Volcanics and the Laina Volcanics, within their Lahaina Volcanic Series. The Lahaina was recently reduced to formational rank and renamed the Lahaina Volcanics by Clague and others (1982), but they did not address the status of the Kilea and Laina. The Kilea Volcanics and Laina Volcanics are two small flow units with associated cinder cones; because they are of such limited extent, they are here abandoned as formally named units and their names used informally (table 1.12). The name Lahaina Volcanics of Clague and others (1982) is retained here.

Isotopic age determinations of the Wailuku Basalt and Honolua Volcanics (McDougall, 1964; Naughton and others, 1980) (table 1.12) indicate a Pleistocene age for both units. The Lahaina Volcanics is less well dated, but it is also considered to be Pleistocene. The single K-Ar age of 1.30 ± 0.10 Ma (Naughton and others, 1980) is considered to be too old on stratigraphic grounds.

KAHOOLAWE AND LANAI

Each of these islands consists of a single shield volcano with a summit caldera, and each has been little dissected.

KAHOOLAWE VOLCANO

The lava of Kahoolawe Volcano was erupted along a prominent southwest rift zone (fig. 1.25). The caldera was almost completely buried beneath a cap of later lava.

The rocks that form essentially all of the Island of Kahoolawe, the smallest of the major islands, were called the Kanapou Volcanic Series by Stearns (1946). The Kanapou, which is not subdivided and is here renamed the Kanapou Volcanics, represents the shield stage, a caldera-filling phase of both the shield and postshield stages, and the postshield stage. The small rejuvenated-stage vents that occur in the sea cliffs on the west side of Kanapou Bay were not considered by Stearns (1946) to be part of the Kanapou, and they are not so considered here.

The alkalic part of the Kanapou Volcanics has been dated at about 1 Ma (Naughton and others, 1980); the tholeiitic part is undated, but is presumed here to be Pleistocene also. The rejuvenated-stage vents are not isotopically dated. They were considered to be of Holocene age by Macdonald and others (1983), but this age is probably too young.

LANAI VOLCANO

The Island of Lanai was built by eruptions from the summit and along northwest, southwest, and southeast rift zones (fig. 1.26). The caldera was mostly filled by lava flows, but its remnant is now covered by alluvium.

The volcanic rocks of Lanai represent the shield stage, including the caldera-filling phase, and they were called the Lanai Volcanic Series by Stearns (1946). Wentworth (1925) had originally applied the name Lanai Basalt to the lava flows of Lanai Volcano and the name Manele Basalt to the small crater remnant

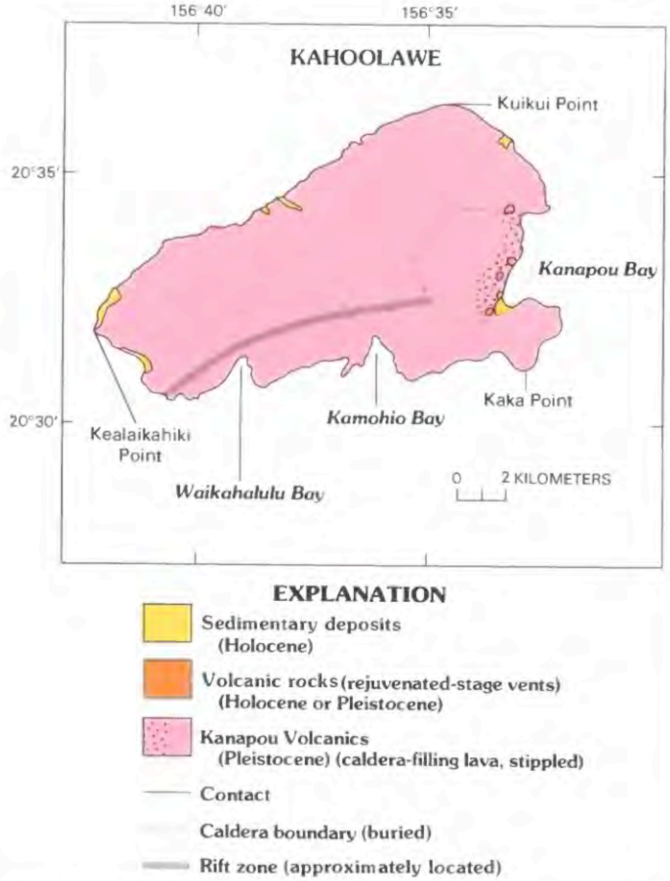

FIGURE 1.25.—Generalized geologic map of Kahoolawe, showing localities mentioned in text. Modified from Macdonald and others (1983). Rift zone from Fiske and Jackson (1972).

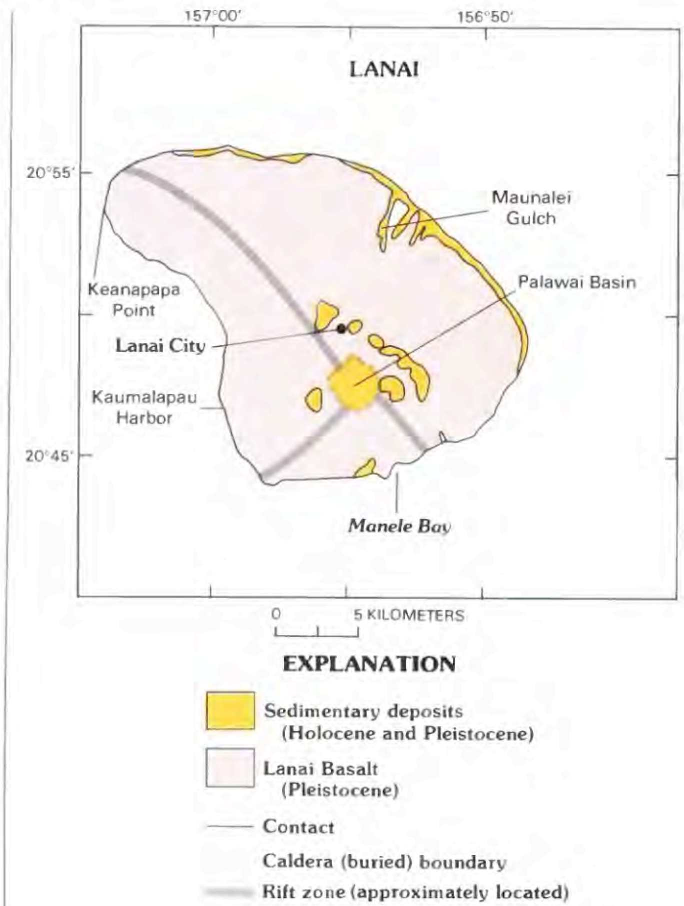

FIGURE 1.26.—Generalized geologic map of Lanai, showing localities mentioned in text. Modified from Macdonald and others (1983). Rift zones from Fiske and Jackson (1972). Caldera boundary inferred to be largely coincident with Palawai Basin.

that forms the headland southwest of Manele Bay. Stearns (1946) redefined the Lanai to include not only the lava flows but also other associated rocks (pyroclastic and intrusive rocks) in a unit he called the Lanai Volcanic Series; he did not include Wentworth's terms Lanai and Manele Basalts as part of the Lanai Volcanic Series. However, Macdonald and Davis (in Avias and others, 1956) included (improperly) the Lanai Basalt, which included Manele Basalt, in the Lanai Volcanic Series, seemingly giving group status to the later unit.

The Lanai Volcanic Series of Stearns (1946) is here reduced to formational rank and renamed the Lanai Basalt. Though there is a chance of confusion with the Lanai Basalt of Wentworth, which consisted only of the lava flows, it would be more misleading to call the unit "Lanai Volcanics" because it consists totally of shield-stage tholeiitic basalt.

The term Manele Basalt is here abandoned as a formal name because it has essentially the same lithology and same age as the Lanai Basalt, as used here, and is of extremely limited extent and the term has been applied to a former high stand of the sea.

The Lanai Basalt is of Pleistocene age based on a K-Ar isochron age of 1.28 ± 0.4 Ma (Bonhommet and others, 1977) (table 1.12).

MOLOKAI

The Island of Molokai is another volcanic doublet, made up of two coalesced volcanoes—East Molokai and West Molokai Volcanoes (fig. 1.27). East Molokai Volcano had a summit caldera; there is no evidence that West Molokai Volcano had one. Both volcanoes are deeply dissected along their northern coasts.

EAST MOLOKAI VOLCANO

East Molokai Volcano was built principally by eruptions from the summit caldera and along east and northwest rift zones.

The volcanic rocks of East Molokai Volcano were divided by Stearns (1946, 1947a) into the East Molokai Volcanic Series (older) and the Kalaupapa Basalt (younger).

The East Molokai Volcanic Series, here renamed the East Molokai Volcanics, was subdivided by Stearns into two informal members—lower and upper members. The lower member represents the shield stage and part of the postshield stage; both stages include a caldera-filling phase. The upper member represents the rest of the postshield stage.

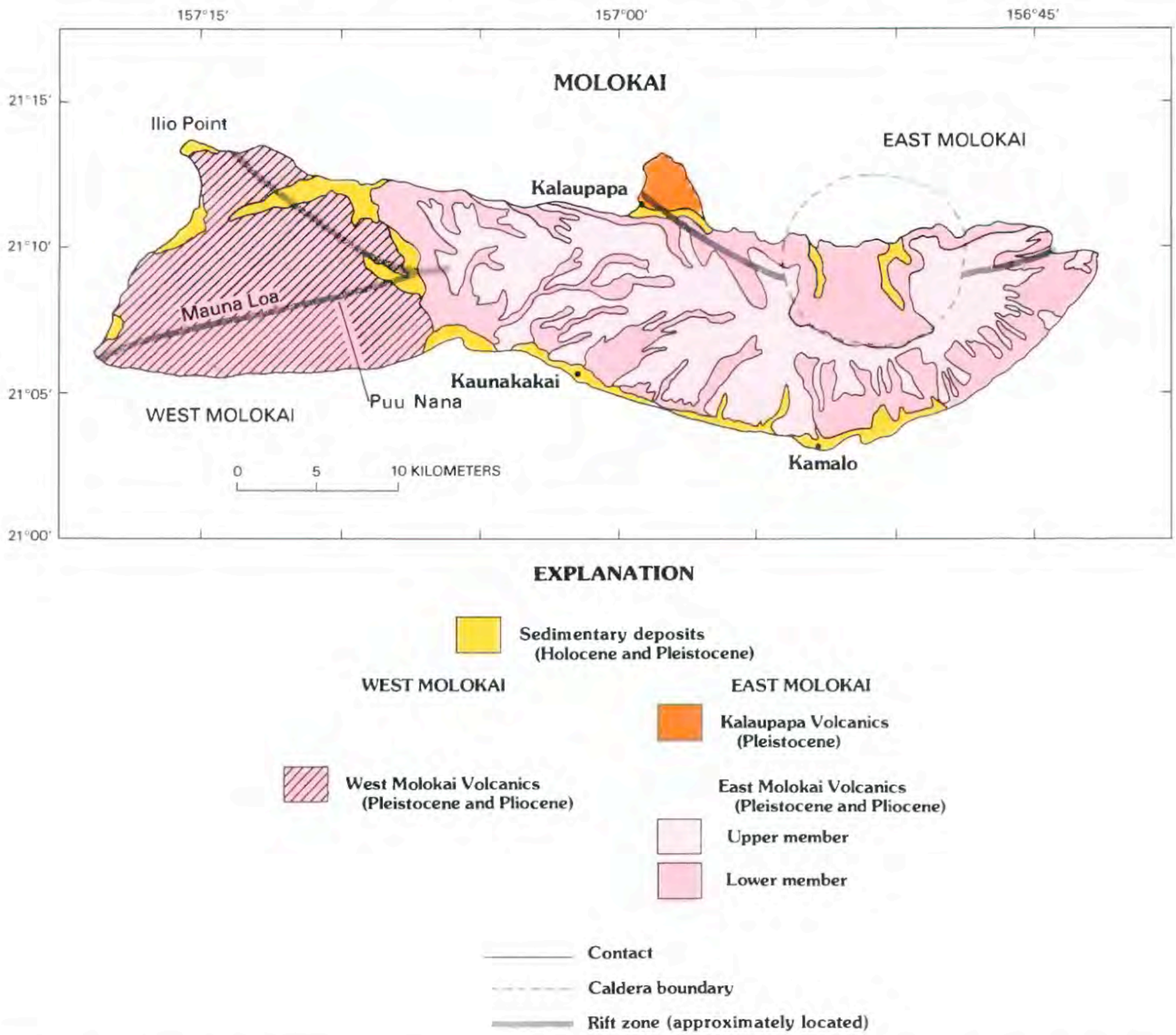

FIGURE 1.27.—Generalized geologic map of Molokai, showing localities mentioned in text. Modified from Macdonald and others (1983). Rift zones from Fiske and Jackson (1972). East Molokai Volcano caldera boundary from Holcomb (1985).

The Kalaupapa Basalt, which consists of alkalic basalt and basanite, is here renamed the Kalaupapa Volcanics to reflect the range of compositions present. The Kalaupapa has been proposed as a separate shield (see Macdonald and others, 1983; Holcomb, 1985), but it is here considered to represent a rejuvenated-stage vent associated with East Molokai Volcano on the basis of its age and chemistry (see Clague and others, 1982).

The East Molokai Volcanics is largely Pleistocene, but its lowermost flows are Pliocene on the basis of K-Ar age determinations on its lower and upper members (McDougall, 1964; Naughton and others, 1980) (table 1.12). The Kalaupapa Volcanics has a Pleistocene age based on K-Ar determinations of 0.57 ± 0.02 and 0.35 ± 0.03 Ma (Clague and others, 1982).

WEST MOLOKAI VOLCANO

West Molokai Volcano was built by eruptions principally along a northeast rift zone that crosses the summit area (Puu Nana) and along a northwest rift zone. There is no evidence of a summit caldera.

All of the volcanic rocks of West Molokai Volcano were called the West Molokai Volcanic Series by Stearns (1946, 1947a). The West Molokai Volcanic Series, which was not subdivided by Stearns, is here renamed the West Molokai Volcanics to reflect its varied rock types. The West Molokai Volcanics represents the shield and postshield eruptive stages.

The age of the West Molokai Volcanics is considered to be

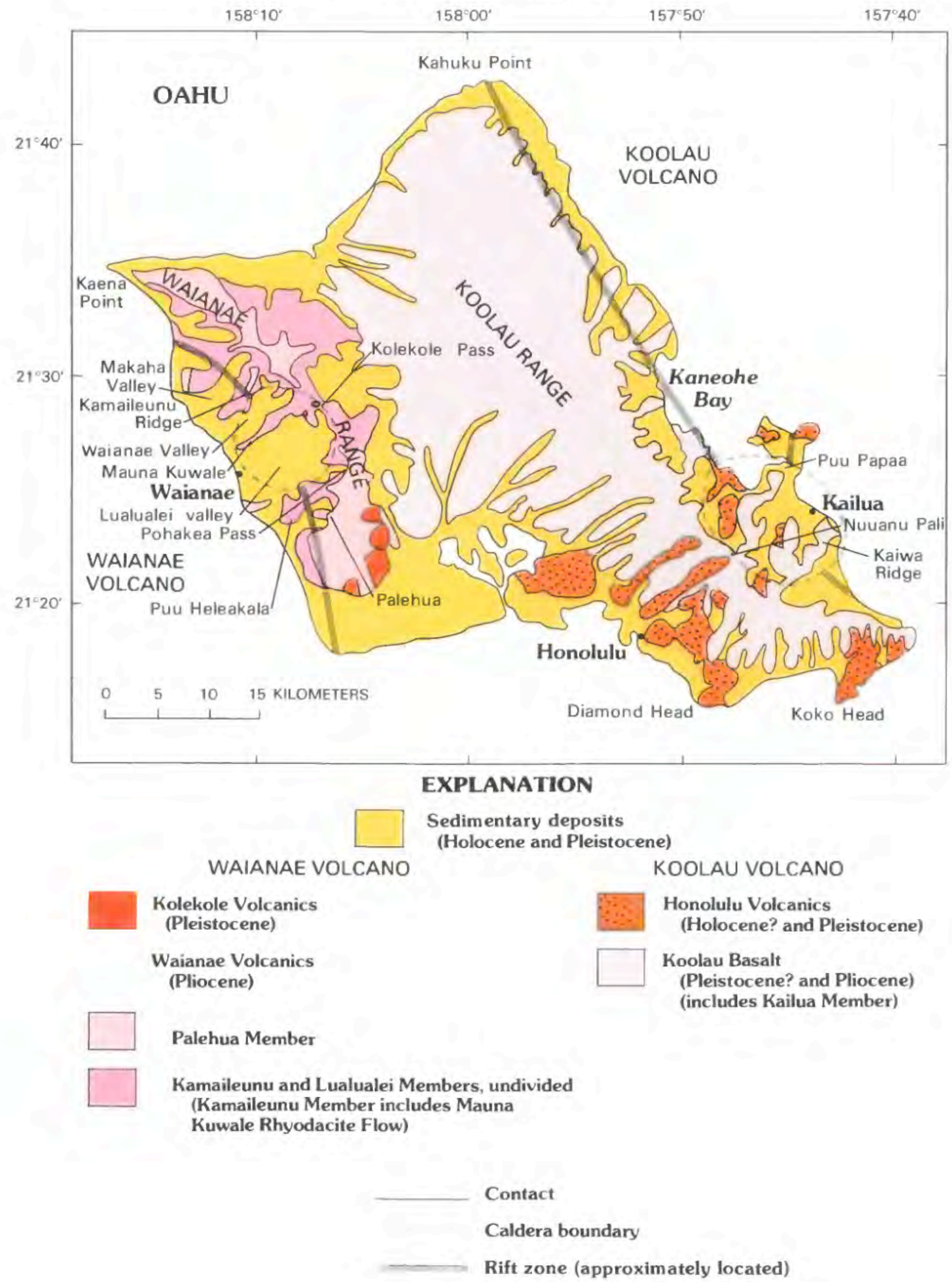

FIGURE 1.28.—Generalized geologic map of Oahu, showing localities mentioned in text. Modified from Macdonald and others (1983) and Sinton (in press). Rift zones from Fiske and Jackson (1972). Koolau Volcano caldera boundary from Clague and Frey (1982); Waianae Volcano caldera boundary from Sinton (in press).

Pliocene and Pleistocene on the basis of K-Ar determinations by Naughton and others (1980) (table 1.12).

OAHU

This island is made up of the highly dissected remnants of two shield volcanoes—Koolau Volcano (east) and Waianae Volcano (west) (fig. 1.28)—that have lost their original shield outline and are now two northwest-trending ridges shaped mainly by erosion. Both volcanoes had summit calderas.

KOOLAU VOLCANO

The lava of Koolau Volcano was principally erupted from the caldera and along the northwest and southeast rift zones. A major dike complex occurs in the rift zones (Walker, chapter 41). The vents for the youngest lava (rejuvenated stage) show no relationship to the preexisting rift zones.

The rocks of this volcano were originally divided by Stearns (1935, 1939) into three major volcanic units—the Kailua Volcanic Series (oldest), the Koolau Volcanic Series, and the Honolulu Volcanic Series (youngest). The Koolau and the Honolulu represent the shield stage and rejuvenated stage, respectively. Stearns originally believed the Kailua Volcanic Series represented an older lava series that was under or close by the summit caldera of Koolau Volcano, but that it was not part of Koolau Volcano. He later (Stearns, 1940a) recognized that the Kailua was part of the Koolau Volcano, representing the caldera complex. Macdonald and Davis (in Avias and others, 1956) included (improperly) the Kailua Volcanic Series in the Koolau Volcanic Series. Although the ranks they intended for these units are not clear, it is here presumed that the Kailua Volcanic Series was of formational rank.

More recently the Koolau has been considered to be of formational rank. It was called the Koolau Formation by Wentworth (1951) and the Koolau Volcanics by Lanphere and Dalrymple (1979) (table 1.12). It is here renamed the Koolau Basalt because it consists entirely of shield-stage tholeiitic basalt. Although Wentworth (1926, fig. 15) also used the term "Koolau basalt" in the explanation of a page-size geologic map of a small area of Oahu, he did not describe the unit and there does not seem much chance of confusion with the Koolau Basalt as used here. The Kailua Volcanic Series is here reduced in rank and renamed the Kailua Member of the Koolau Basalt.

The Honolulu Volcanic Series was defined by Stearns (1935) to consist of rejuvenated-stage lava that erupted from more than 35 vents on the southern slopes of the volcano. These lavas are strongly alkalic and form a variety of cones (tuff, cinder, spatter), many with associated lava flows. The eruptions presumably did not come in rapid succession, but occurred over a long period of time. Their general sequence (Macdonald and others, 1983) (table 1.12) has been based largely on the relative degree of weathering and erosion and on relations to stands of the sea because few superpositional relations are known and isotopic ages are considered unreliable.

The Honolulu Volcanic Series was of group rank as defined by Stearns (1935, 1939) because it was made up of numerous formally named units of formational rank (table 1.12). Hay and Iijima (1968) and Clague and others (1982) also considered the Honolulu to be of group rank. Lanphere and Dalrymple (1979) and Clague and Frey (1982) used the formational-rank term Honolulu Volcanics, but did not address the status of its formally named subdivisions. The Honolulu is here considered to be of formational rank, and the term Honolulu Volcanics is retained to reflect its range of lithologic compositions.

The formally named eruptive units that make up the Honolulu could have been reduced in rank and retained as formally named members, but, because most are single flows of very limited extent, this seems unnecessary; they are here all considered to be informal units (table 1.12).

The age of the Koolau Basalt is Pliocene based on K-Ar ages ranging from 2.7 to 1.8 Ma (Doell and Dalrymple, 1973); however, a Pleistocene age cannot be ruled out for the youngest flows (table 1.12). The Kailua Member has not been isotopically dated, but is probably Pleistocene.

The K-Ar ages reported for the Honolulu Volcanics range from about 0.9 to about 0.03 Ma, but there are large differences (sometimes a factor of ten) in the ages reported by different investigators for the same eruptive units (see Clague and Dalrymple, chapter 1, part I; Macdonald and others, 1983). The Honolulu is certainly in part Pleistocene on the basis of published K-Ar ages, but because of the unreliability of some of the ages, a Holocene age cannot be ruled out for the youngest flows.

WAIANAE VOLCANO

Waianae Volcano was built by eruptions from the summit caldera and along the principal northwest and southeast rift zones.

The rocks of the volcano are divided into the Waianae Volcanics, representing the shield and postshield stages (both including a caldera- filling phase), and the Kolekole Volcanics, representing the rejuvenated stage (Sinton, in press). The Waianae was originally named the Waianae Volcanic Series and subdivided into three informal members (lower, middle, and upper) by Stearns (1935). He described these members but did not map them separately (Stearns, 1939). The Waianae has recently been subdivided by Sinton (in press) into three formally named members—the Lualualei (oldest), Kamaileunu, and Palehua (youngest) Members. These members are, for the most part, equivalent to Stearns' lower, middle, and upper members, respectively (table 1.12; see Sinton, in press). The Kamaileunu Member includes the Mauna Kuwale Rhyodacite Flow and several icelandite flows. These are the only known occurrences of rhyodacite and icelandite in the Hawaiian Islands. The rhyodacite flow was given formal status by Sinton (in press), and because it is of such distinctive lithology it is also considered to be formal here.

A Pliocene age for the Waianae Volcanics is inferred from K-Ar determinations ranging from 3.9 to 2.5 Ma (Doell and Dalrymple, 1973) (table 1.12). Funkhouser and others (1968) reported an average K-Ar age of about 2.4 Ma for the Mauna Kuwale Rhyodacite Flow, but this age is inconsistent with their reported ages for the overlying lava flow (4.3 Ma) and dikes (about 3 Ma) cutting the rhyodacite. According to Sinton (in press), the Mauna Kuwale probably has a minimum age of 3.2 Ma. The Kolekole Volcanics has not been isotopically dated but is tentatively considered by Sinton (in press) to be Pleistocene.

KAUAI

KAUAI VOLCANO

The Island of Kauai (fig. 1.29) consists of a single deeply eroded shield volcano with a summit caldera 15-20 km across, the largest in the Hawaiian Islands, and at least two flank calderas, the only ones known in the islands. Lava erupted not only in the calderas, but also from northwest and southeast rift zones. After a period of erosion, rejuvenated-stage lava erupted from about 40

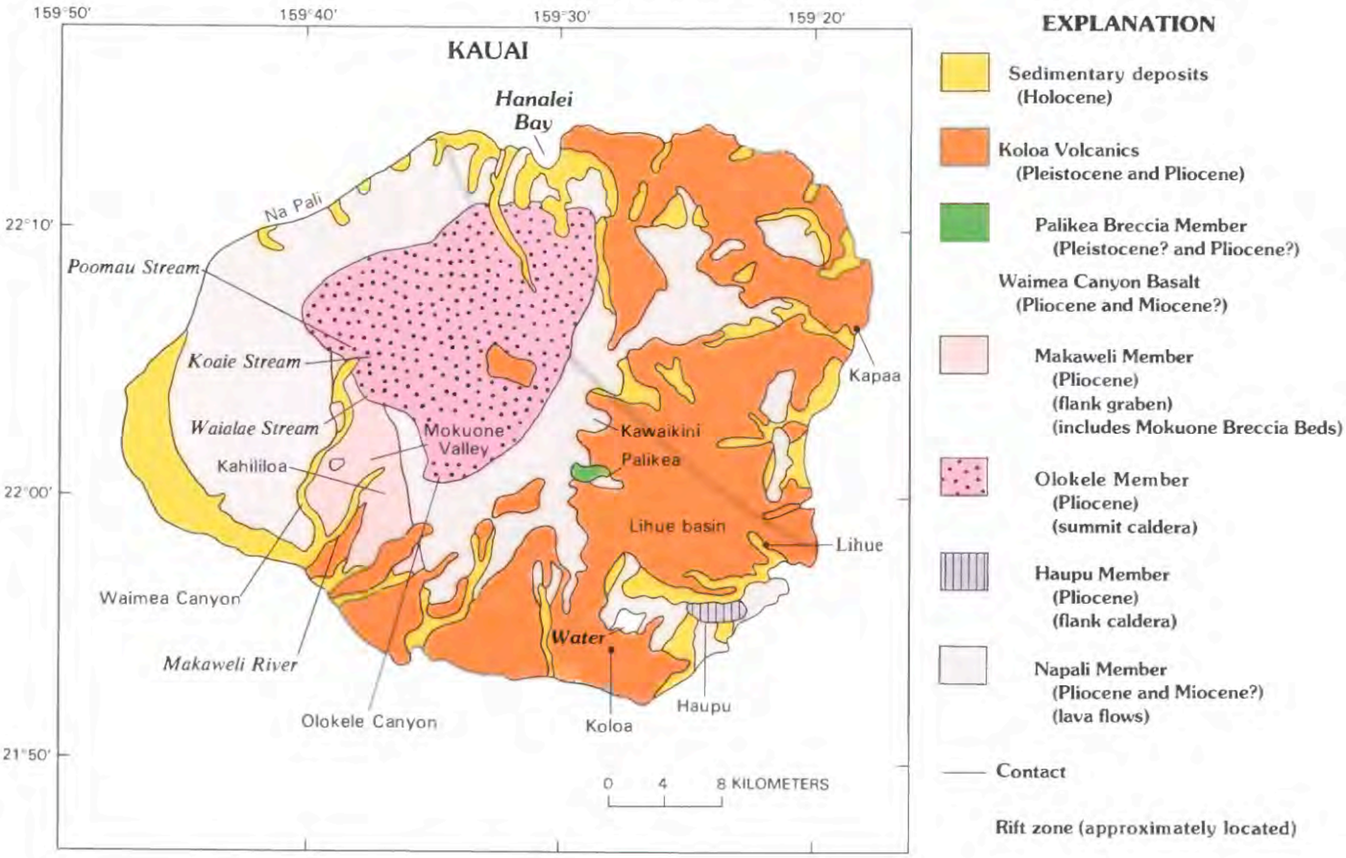

FIGURE 1.29.—Generalized geologic map of Kauai, showing localities mentioned in text. Modified from Macdonald and others (1983). Rift zones from Fiske and Jackson (1972). Summit caldera and southeast flank caldera are represented by Olokele Member and Haupu Member, respectively. Another flank caldera is represented by Lihue basin, a subcircular basin 10-15 km wide, which is almost bisected by southeast rift zone. There is no evidence that shield-stage or postshield-stage lava poured into this caldera, but it was later filled with rejuvenated-stage lava (Koloa Volcanics).

vents scattered over the eastern two-thirds of the island; these vents show no relation to the older rift zones.

The rocks of Kauai Volcano were originally divided by Stearns (1946) into the Waimea Volcanic Series (older), later renamed the Waimea Canyon Volcanic Series (Macdonald, 1949), and the Koloa Volcanic Series (younger). The Waimea Canyon represents the shield stage including the caldera-filling phase. Postshield-stage hawaiite is rare and occurs only at the top of the Waimea Canyon (fig. 1.21, table 1.12). The Koloa represents the rejuvenated stage.

The Waimea Canyon Volcanic Series was subdivided by Macdonald and others (1954, 1960) into four formally named formations—the Napali, Haupu, Olokele, and Makaweli Formations. The Makaweli Formation as defined by Macdonald and others (1960) included associated sedimentary deposits, which were called the Mokuone Member.

In order to have a nomenclature consistent with the other volcanoes of the Hawaiian Islands, the following changes are here made to the Waimea Canyon Volcanic Series: The Waimea Canyon is reduced to formational rank and renamed the Waimea Canyon Basalt to reflect its predominant lithology. Its subdivisions are reduced to member rank and called the Napali, Haupu, Olokele, and Makaweli Members. The sedimentary unit included within the Makaweli is reduced in rank and renamed the Mokuone Breccia Beds (table 1.12).

Although the Napali, Haupu, Olokele, and Makaweli Members consist of essentially the same rock types, they are distinguishable and extensive units. The Napali Member consists of thin-bedded flank flows, whereas the Olokele and Makaweli Members consist of massive or thick-bedded flows. The Olokele is separated from the Makaweli and from the Napali by faults. The Haupu, which presumably represents a small caldera on the southeast flank of the volcano, also consists of thick-bedded flows, but it is isolated from the Olokele and Makaweli Members and is surrounded by older thin-bedded flows of the Napali Member.

The Napali, Olokele, and Makaweli Members had more than one type locality specified by Davis and Macdonald (in Avias and others, 1956) and Macdonald and others (1960). One of the localities is here retained as the type locality for each unit, and the other localities are redesignated as reference localities (table 1.12.).

The Koloa Volcanic Series, as defined by Stearns (1946) and Macdonald and others (1960), included within it a sedimentary unit that they named the Palikea Formation. The Palikea Formation consists largely of breccia underlying and interbedded with Koloa lava flows. The breccia, which grades laterally in some places into stream-laid conglomerate, is primarily made up of fragments of the Waimea Canyon Basalt and represents the rapid shedding of debris from the steep slopes of the volcano before and during Koloa time. Clague and others (1982) reduced the Koloa Volcanic Series to formational rank and renamed it the Koloa Volcanics, but they did not discuss the status of the Palikea Formation. The Palikea is here reduced in rank and renamed the Palikea Breccia Member of the Koloa Volcanics.

The Waimea Canyon Basalt is largely of Pliocene age, but the oldest flows may be Miocene, as indicated by K-Ar ages ranging from about 5.1 to 3.6 Ma (McDougall, 1964; 1979). The age relations between the caldera-filling-phase members are not known, but the Olokele Member is considered to be older than the Makaweli Member, and the Haupu Member is thought to be coeval with the Olokele.

The Koloa Volcanics is largely Pleistocene, an age indicated by all the published K-Ar ages, but unpublished K-Ar ages of 2.59 and 2.01 Ma (G.B. Dalrymple, oral commun., 1986) suggest that the oldest flows are Pliocene. The age of the Palikea Breccia Member is not known, but it is probably Pliocene and Pleistocene on the basis of its stratigraphic relations to Koloa lava flows.

NIIHAU

NIIHAU VOLCANO

The Island of Niihau (fig. 1.30) is the deeply eroded remnant of a shield volcano. The central highland area, which consists of shield-stage tholeiitic lava, is fringed by a low coastal platform, which consists of rejuvenated-stage lava. The shield-stage lava was erupted from a southwest rift zone; the summit of the shield (and former caldera?) was presumably northeast of the present island.

The rocks of the volcano were divided by Stearns (1946, 1947b) into two major units—the Paniau Volcanic Series (older) and the Kiekie Volcanic Series (younger). The Paniau almost entirely represents the shield stage; a single postshield-stage vent that occurs at Kaeo and a couple postshield alkalic basalt dikes along the east coast of the island are also considered in this report to be part of the Paniau Basalt. These units are here renamed the Paniau Basalt and Kiekie Basalt, respectively. Although all of the rejuvenated-stage units on the other islands are called "Volcanics," the name Kiekie Volcanics is not used here because the Kiekie consists entirely of alkalic basalt.

No isotopic ages for the Island of Niihau have been published. However, unpublished K-Ar data for 11 flows of the Paniau indicate a Miocene and Pliocene age (G.B. Dalrymple, oral commun., 1986). The Kiekie Basalt was considered to be of Pleistocene age by Stearns (1947b) on the basis of its relations to Pleistocene shorelines, but unpublished K-Ar ages (G.B. Dalrymple, oral commun., 1986) indicate a Pliocene and Pleistocene age.

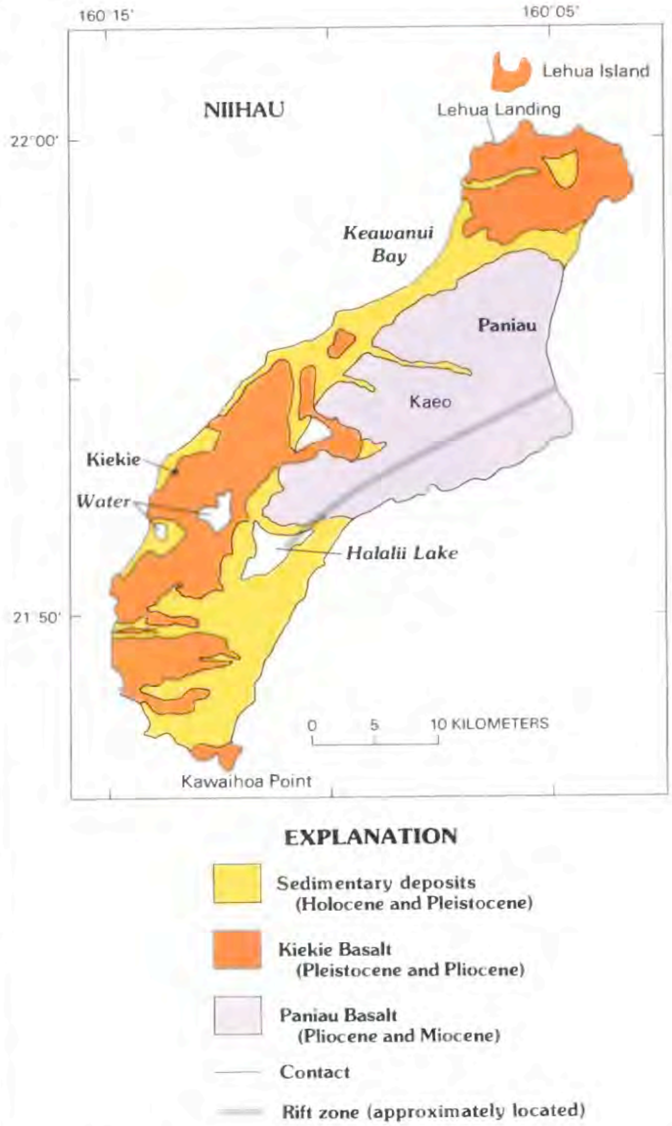

FIGURE 1.30.—Generalized geologic map of Niihau, showing localities mentioned in text. Modified from Macdonald and others (1983). Rift zone approximately follows dike swarm mapped by Stearns (1947b).

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TABLE 1.12.—*Summary of stratigraphic units for main southeastern Hawaiian Islands*

[Volcanic rocks and closely associated intrusive rocks and sedimentary deposits only. Units on each volcano are listed from youngest to oldest; see text for further explanation. All isotopic ages (in parentheses) and K-Ar ages (in millions of years), recalculated from new decay constants, unless otherwise noted; see summary by Clague and Dalrymple (chapter 1, part I). The term "partly" (in parentheses) preceding a former name indicates that the unit as then defined is equivalent to only part of the unit as used here; the term "part" (in parentheses) following a former name indicates that the unit as used here is equivalent to only part of the unit as then defined]

Unit (this report)	Description	Age	Remarks
		HAWAII	
		Kilauea Volcano	
Puna Basalt	Shield-stage lava of tholeiitic basalt and picritic tholeiitic basalt. Historic and prehistoric lava flows and minor intercalated pyroclastic deposits. Exposed over most of surface of Kilauea. Maximum exposed thickness: > 100 m. Type locality: Uwekahuna Bluff at Kilauea Crater (caldera). Reference localities: at Napau, Makaopuhi, and Kilauea Iki Craters, and near Keana Bihopa (see Easton, chapter 11). Overlies and is locally intercalated with Pahala Ash; interfingers with Kau Basalt of Mauna Loa Volcano. Includes Keanakakoi Ash Member and Uwekahuna Ash Member:	Holocene and Pleistocene (^{14}C, 1.13-22.6 ka) (see text)	Puna Basalt (Easton, chapter 11). Formerly: Kamehame Basalt (part) (Stearns, 1926, 1930); Puna Volcanic Series (Stearns and Macdonald, 1946); Puna Formation (Easton and Garcia, 1980).
Keanakakoi Ash Member	Lithic and vitric ash. Pyroclastic surge and minor airfall deposits. Mainly represents explosive eruption of A.D. 1790, but has slightly older reticulite pumice at base and post-1790 but pre-1823 reticulite pumice at top. Occurs around Kilauea Crater (caldera). Maximum exposed thickness: ~11 m. Type locality: just southwest of Keanakakoi Crater. Is interbedded with historic Puna lava flows.	Holocene	Keanakakoi Ash Member (Easton, chapter 11). Formerly: Keanakakoi Formation (part) (Wentworth, 1938; Powers, 1948); "aa" [erroneously for ash?] member (part) of Puna Volcanic Series (Stearns and Macdonald 1946, p. 108); Keanakakoi Formation (part) of Puna Volcanic Series (Davis and Macdonald in Avias and others, 1956); Keanākakoi Member (part) of Puna Formation (Easton and Garcia, 1980); (partly) Keanakakoi Formation (Malin and others, 1983).
Uwekahuna Ash Member	Mostly ash. Mainly pyroclastic surge deposits Exposed near base of Kilauea Crater (caldera) and on southeast flank of Mauna Loa. Maximum exposed thickness: ~5 m. Type locality: Uwekahuna Bluff. Interbedded with prehistoric Puna lava flows.	Holocene (^{14}C, see text)	Uwekahuna Ash Member (Easton, chapter 11). Formerly: Uwekahuna Ash of Kilauea Series (Stone, 1926); Uwekahuna Formation (Wentworth, 1938; Powers, 1948); Uwekahuna Tuff of Puna Volcanic Series (Stearns and Macdonald, 1946, p. 194); Uwekahuna Ash of Puna Volcanic Series (Macdonald, 1949, p. 65, 67; Davis and Macdonald in Avias and others, 1956); Uwekāhuna Member of Puna Formation (Easton and Garcia, 1980).
Pahala Ash	Vitric ash, largely palagonitized. Occurs on Mauna Loa and Kilauea Volcanoes (see text). Maximum exposed thickness: ~15 m. No type locality designated; named for exposures near town of Pahala on south slope of Mauna Loa. Principal reference locality: Moolelo, just southwest of Keana Bihopa; reference localities: Puu Kapukapu and Puu Kaone, all on Kilauea Volcano (see Easton, chapter 11). Overlies Hilina Basalt on Kilauea and Kahuku Basalt on Mauna Loa.	Pleistocene	Geographically restricted (see text). Pahala Ash (Easton, chapter 11). Formerly: Pahala Series (part) (Noble and Clark in Washington, 1923); Pahāla Ash of Pre-Kilauea Series (Stone, 1926); Pahala Basalt or Formation (part) (Stearns, 1930); Pahala Tuff (Wentworth, 1938); Pahala Ash (part) (Stearns and Macdonald, 1946; Davis and Macdonald in Avias and others, 1956); Pahāla Formation (Easton and Garcia, 1980).

TABLE 1.12.—*Summary of stratigraphic units for main southeastern Hawaiian Islands*—Continued

Unit (this report)	Description	Age	Remarks
	HAWAII--Continued		
	Kilauea Volcano--Continued		
Hilina Basalt	Shield-stage lava of tholeiitic basalt, olivine tholeiitic basalt, and picritic tholeiitic basalt. Predominantly flows with associated pyroclastic deposits. Exposed in Hilina fault system escarpments along south flank of volcano. Maximum exposed thickness: ~300 m. Type locality: just south of Keana Bihopa. Reference localities: Keana Bihopa and Puu Kapukapu. Base not exposed. Includes Moo, Pohakaa, Kahele, and Halepe Ash Members, all intercalated with Hilina lava flows.	Pleistocene (see text)	Hilina Basalt (Easton, chapter 11). Formerly: Pahala Basalt (part) (Stearns and Clark, 1930); Hilina Volcanic Series (Stearns and Macdonald, 1946); Hilina Formation (Easton and Garcia, 1980; Decker and Christiansen, 1984).
Moo Ash Member	Mostly yellow-brown ash and palagonite. Maximum exposed thickness: ~3 m. Type locality: Keana Bihopa. Reference locality: Puu Kapukapu.	Pleistocene	Moo Ash Member of Hilina Basalt (Easton, chapter 11). Formerly: Moo Member of Hilina Formation (Easton and Garcia, 1980).
Pohakaa Ash Member	Thickest of ash members. Yellow- to reddish-brown-weathering palagonite, vitric ash, and soil. Exposed thickness: 1-4 m. Type locality: Pohakaa Arroyo, about 3 km southwest of Keana Bihopa. Reference locality: Puu Kapukapu.	Pleistocene	Pohakaa Ash Member of Hilina Basalt (Easton, chapter 11). Formerly: Pohakaa Member of Hilina Formation (Easton and Garcia, 1980).
Kahele Ash Member	Crudely bedded red clay with palagonite. Exposed thickness: 10-125 cm. Type locality: Pohakaa Arroyo, about 3 km southwest of Keana Bihopa. Reference locality: Puu Kapukapu.	Pleistocene	Kahele Ash Member of Hilina Basalt (Easton, chapter 11). Formerly: Kahele Member of Hilina Formation (Easton and Garcia, 1980).
Halape Ash Member	Palagonite and poorly bedded clay. Exposed thickness: 10-50 cm. Type locality: Keana Bihopa. Reference locality: Puu Kapukapu.	Pleistocene	Halape Ash Member of Hilina Basalt (Easton, chapter 11). Formerly: Halape Member of Hilina Formation (Easton and Garcia, 1980).
	Mauna Loa Volcano		
Kau Basalt	Shield-stage lava of tholeiitic basalt, olivine tholeiitic basalt, and picritic tholeiitic basalt. Historic and prehistoric lava flows and minor intercalated pyroclastic deposits. Exposed over most of surface of Mauna Loa. Maximum exposed thickness: > 185 m. Type locality: west wall of Mokuaweoweo Crater (caldera). Overlies Pahala Ash and Kahuku Basalt; interfingers with Puna Basalt of Kilauea Volcano.	Holocene and Pleistocene	Renamed. Formerly: Kamehame Basalt (part) (Stearns, 1926, 1930); Kau Volcanic Series (Stearns and Macdonald, 1946); Kau Formation (Porter, 1971, 1974); Kau Volcanics (Lipman, 1980).
Pahala Ash	See description under Kilauea.	Pleistocene	See "Remarks" under Kilauea; see also text under Mauna Loa.
Kahuku Basalt	Shield-stage lava of tholeiitic basalt, olivine tholeiitic basalt, and picritic tholeiitic basalt. Lava flows and minor intercalated pyroclastic deposits. Exposed on east and south sides of Mauna Loa. Maximum exposed thickness: 180 m. Type locality: Kahuku Pali, a fault escarpment running north of Ka Lae. Overlies (fault or unconformity) Ninole Basalt.	Pleistocene	Renamed. Formerly: lower member of Pahala Basalt (part) (Stearns 1930); Kahuku Volcanic Series (Stearns and Macdonald, 1946); Kahuku Volcanics (Lipman, 1980).
Ninole Basalt	Shield-stage lava of tholeiitic basalt, olivine tholeiitic basalt, and picritic tholeiitic basalt. Lava flows with pyroclastic deposit near top. Exposed north of Waiohinu on southeast slope of Mauna Loa. Maximum exposed thickness: ~600 m. Type locality: walls of valley at Puu Enuhe, near Ninole Gulch. Base not exposed.	Pleistocene (0.54±0.4) (see text)	Ninole Basalt (Stearns, 1926, 1930) (see text). Formerly: (partly) Ninole Tuff (Wentworth, 1938); Ninole Volcanic Series (Stearns and Macdonald, 1946); Ninole Volcanics (Lipman, 1980).

TABLE 1.12.—*Summary of stratigraphic units for main southeastern Hawaiian Islands*—Continued

Unit (this report)	Description	Age	Remarks
	HAWAII--Continued		
	Mauna Kea Volcano		
Laupahoehoe Volcanics	Postshield-stage alkalic lava flows and associated volcanic and sedimentary deposits. Occurs on most of surface of Mauna Kea. Typical exposed thickness: 50-100 m. Type locality: near town of Laupahoehoe. Reference localities: in Waikahalulu and Pohakuloa Gulches. Conformably overlies Hamakua Volcanics. Divided into Waikahalulu Volcanic Member, and Makanaka and Waihu Glacial Members:	Holocene and Pleistocene	Reduced in rank and renamed (see text). Formerly: (partly) Laupahoehoe Volcanic Series (Macdonald, 1945; Stearns and Macdonald, 1946); Laupahoehoe Series (Porter, 1971); Laupahoehoe Group (Porter, 1973, 1974, 1979a, 1979b; Porter and others, 1977).
Waikahalulu Volcanic Member	Postshield-stage lava predominantly of hawaiite, but some alkalic basalt and ankaramite. Flows and associated pyroclastic deposits. Occurs on upper and lower slopes of volcano. Typical exposed thickness: 50-100 m. Type locality: in Waikahalulu Gulch. Reference locality: in Pohakuloa Gulch. Conformably, and locally unconformably, overlies Hamakua Volcanics. Interstratified with Makanaka and Waihu Glacial Members.	Holocene and Pleistocene (^{14}C and K-Ar, 0.0045-0.19)	Reduced in rank and renamed (see text). Formerly: (partly) Laupahoehoe Volcanic Series (Stearns and Macdonald, 1946); (partly) Waikahalulu Formation of Laupahoehoe Group (Porter, 1974, Porter and others, 1977); Waikahalulu Formation of Laupahoehoe Group (Porter, 1979a, 1979b).
Makanaka Glacial Member	Gravel and diamicton. Exposed on upper slopes of volcano. Exposed thickness: > 10 m. Type locality: Puu Makanaka. Reference localities: in Waikahalulu and Kemole Gulches. Intercalated with Waikahalulu lavas. Younger than Waihu Glacial Member.	Pleistocene	Reduced in rank and renamed (see text). Formerly: Makanaka Drift (Wentworth and Powers, 1941; see also Stearns, 1945); Makanaka Formation (part) of Laupahoehoe Group (Porter, 1974; Porter and others, 1977); Makanaka Formation of Laupahoehoe Group (Porter, 1979a, 1979b).
Waihu Glacial Member	Mainly conglomerate and diamictite. Exposed mainly on upper slopes below summit of Mauna Kea. Exposed thickness: generally ≤ 30 m. Type locality: near Waihu Spring. Reference localities: in Pohakuloa Gulch and near Waihu Spring.	Pleistocene	Reduced in rank and renamed (see text). Formerly: Waihu Drift (Wentworth and Powers, 1941); Waihu Fanglomerate (Stearns, 1945); Waihu Formation of Laupahoehoe Group (Porter and others, 1977; Porter, 1974, 1979a, 1979b).
Hamakua Volcanics	Divided into three members: upper two formally-named members represent postshield-stage lava flows and pyroclastic cones and associated sedimentary (glacial) deposits; informal lower member represents shield-stage lava flows. Type locality: in south wall of Laupahoehoe Stream gulch. Reference locality: Pohakuloa Gulch.	Pleistocene	Reduced in rank and renamed (see text). Formerly: (partly) Hamakua Volcanic Series (Stearns and Macdonald, 1946); Hamakua Group (Porter, 1973, 1974, 1979a, 1979b; Porter and others, 1977) (see text).
Pohakuloa Glacial Member	Gravel and diamicton. Exposed mainly in Pohakuloa and Waikahalulu Gulches. Maximum exposed thickness: ~40 m. Type locality: west wall of upper Pohakuloa Gulch. Reference locality: in Waikahalulu Gulch. Locally overlies upper flows of Hopukani Volcanic Member.	Pleistocene	Reduced in rank and renamed (see text). Formerly: Pohakuloa Drift (Wentworth and Powers, 1941); Pohakuloa Formation of Hamakua Group (Porter, 1974, 1979a, 1979b; Porter and others, 1977)
Hopukani Volcanic Member	Postshield-stage lava of alkalic basalt, ankaramite, and hawaiite. Flows and cinder cones. Exposed mainly in gulches on lower windward slopes of Mauna Kea and in saddle between Mauna Kea and Kohala Volcanoes. Exposed thickness: ~25 m. Type locality: in south wall of Laupahoehoe Stream gulch. Reference localities: in Pohakuloa and Waikahalulu Gulches. Overlies lower member.	Pleistocene (0.27±0.04, 0.375±0.050)	Reduced in rank and renamed (see text). Formerly: upper member of Hamakua Volcanic Series (Stearns and Macdonald, 1946); Hopukani Formation of Hamakua Group (Porter, 1974, 1979a, 1979b; Porter and others, 1977).

TABLE 1.12.—*Summary of stratigraphic units for main southeastern Hawaiian Islands*—Continued

Unit (this report)	Description	Age	Remarks
	HAWAII--Continued		
	Mauna Kea Volcano--Continued		
lower member	Shield-stage lava of tholeiitic basalt, olivine tholeiitic basalt, and picritic tholeiitic basalt. Exposed in sea cliffs and gulches along northeast coast of Mauna Kea (so-called Hamakua coast) and near mouths of valleys between Hilo and Laupahoehoe. Exposed thickness: probably ~150 m. Base not exposed.	Pleistocene	Formerly: lower member of Hamakua Volcanic Series (Stearns and Macdonald, 1946, p. 154, fig. 31) (see text).
	Hualalai Volcano		
Hualalai Volcanics	Postshield-stage lava of alkalic basalt and rare hawaiite. Historic and prehistoric lava flows and associated pyroclastic deposits. Covers entire surface of Hualalai Volcano. Maximum exposed thickness: ~150 m. Type locality: around town of Kailua. Interfingers with Kau Basalt of Mauna Loa Volcano. Base not exposed. Includes Kona ash beds (informal) and Waawaa Trachyte Member (formal) (see text).	Holocene and Pleistocene	Reduced in rank and renamed (see text). Formerly: Hualalai Volcanic Series (Stearns and Macdonald, 1946).
	Kona ash beds		Abandoned as formally named unit (see text). Formerly: Kona Tuff Formation (Wentworth, 1938); unnamed unit of Hualalai Volcanic Series (Stearns and Macdonald, 1946); Kona Tuff Formation of Hualalai Volcanic Series (Davis and Macdonald in Avias and others, 1956).
Waawaa Trachyte Member	Trachyte cone and flow. Occurs in Puu Waawaa-Puu Anahulu area. Exposed thickness: ~275 m. Type locality: Puu Waawaa. Base not exposed.	Pleistocene (0.105, see text)	Renamed. Formerly Waawaa Volcanics [Member] of Hualalai Volcanic Series (Stearns and Macdonald, 1946, p. 143) (see text).
	Kohala Volcano		
Hawi Volcanics	Postshield-stage lava of mostly mugearite and hawaiite with some trachyte and benmoreite. Lava flows and associated pyroclastic deposits. Occurs in summit area and parts of slopes of Kohala Volcano. Exposed thickness of flows: ~30 m. Type locality: in Kumakua Gulch, about 1 km east of Hawi. Underlies Quaternary surficial deposits. Conformably, locally unconformably, overlies Pololu Basalt. Basalt.	Pleistocene (0.061±0.001, 0.261±0.005)	Renamed. Formerly: Hawi Volcanic Series (Stearns and Macdonald, 1946).
Pololu Basalt	Shield-stage lava of tholeiitic basalt, olivine tholeiitic basalt, and picritic tholeiitic basalt; postshield-stage caldera-filling alkalic basalt near top. Thin lava flows and associated pyroclastic deposits. Occurs on upper and lower slopes of Kohala. Maximum exposed thickness: ~900 m. Type locality: on northwest side of Pololu Valley. Base not exposed.	Pleistocene (0.304±0.091-0.459±0.028)	Renamed. Formerly: Pololu Volcanic Series (Stearns and Macdonald, 1946).

TABLE 1.12.—*Summary of stratigraphic units for main southeastern Hawaiian Islands*—Continued

Unit (this report)	Description	Age	Remarks
	MAUI		
	East Maui (Haleakala) Volcano		
Hana Volcanics	Rejuvenated-stage lava of alkalic basalt and basanite; rare ankaramite and hawaiite. Lava flows and associated intrusive rocks and pyroclastic and sedimentary deposits. Exposed in summit region and on eastern and southwestern slopes. Exposed thickness: > 300 m. Type locality: village of Hana. Unconformably overlies Kula Volcanics. Includes Kipahulu Member in Kipahulu Valley. Also includes historic flow of A.D. 1790 in Cape Kinau area and numerous informal flows in Nahiku and Keanae areas (see text) that are listed below in sequence (youngest to oldest) from Stearns and Macdonald (1942):	Holocene and Pleistocene(?)	Reduced in rank, revised to include historic (A.D. 1790) flow, and renamed (see text). Formerly: Hana Volcanic Series (Stearns, 1942); Hana Formation (Macdonald, 1978); Hana Group (Clague and others, 1982).
			Abandoned as formally named units (see text). Formerly all in Hana Volcanic Series (see Macdonald, 1942; Macdonald and Davis <u>in</u> Avias and others, 1956):
	Nahiku area:		
	Hanawi flow		Hanawi Basaltic Andesite
	Paakea flow		Paakea Basalt
	Kuhiwa flow		Kuhiwa Basaltic Andesite
	Mossman flow		Mossman Picritic Basalt
	Kapaula flow		Kapaula Basaltic Andesite
	Makaino flow		Makaino Basaltic Andesite
	Waiaaka flow		Waiaaka Basaltic Andesite
	Makapipi flows		Makapipi Basalt(s)
	Big Falls flows		Big Falls Picritic Basalt(s)
			Formerly all in Hana Volcanic Series (see Stearns, 1942):
	Keanae area:		
	Keanae flow		Keanae Basalt
	Waiokamilo flow		Waiokamilo Basalt
	Piinaau flow		Piinaau Basalt
	Ohia flow		Ohia Basalt
	Wailuanui flow		Wailuanui Basalt
	Pauwalu flow		Pauwalu Basalt
Kipahulu Member	Late alkalic rejuvenated-stage lava of alkalic basalt and basanite; rare ankaramite and hawaiite. Lava flows. Exposed thickness: > 400 m. Type locality: Kipahulu Valley. Unconformably overlies Hana Volcanics (older flows), Kula Volcanics, and Honomanu Basalt.	Pleistocene(?)	See text. Formerly: Kipahulu Member of Hana Volcanic Series (Stearns, 1942); Kipahulu Formation of Hana Volcanic Series (Macdonald and Davis <u>in</u> Avias and others, 1956); Kipahulu Member of Hana Volcanic Series (Macdonald and Abbott, 1970; Macdonald and others, 1983; Stearns, 1985).
Kula Volcanics	Postshield-stage lava of hawaiite with some ankaramite and alkalic basalt. Lava flows with associated intrusive rocks and pyroclastic and sedimentary deposits. Exposed in summit region and on all slopes. Maximum exposed thickness: ~600 m. Type locality: near Waiakoa, on west slope of volcano; reference localities: near Kipahulu and along Halemauu Trail (see text). Overlies Honomanu Basalt.	Pleistocene (0.46-0.86, 0.41±0.09)	Renamed. Formerly: Kula Volcanic Series (Stearns, 1942); (partly) Kula Formation (Macdonald, 1978). Includes rocks formerly called Kumuiliahi Formation by Macdonald (1978) (see text).
Honomanu Basalt	Shield-stage lava of tholeiitic basalt, olivine tholeiitic basalt, and picritic tholeiitic basalt. Lava flows and associated intrusive rocks and rare pyroclastic deposits. Exposed in summit region and on north, northeast, and southwest slopes. Maximum exposed thickness: ~250 m. Type locality: in Honomanu Stream valley. Base not exposed.	Pleistocene (0.75±0.04)	Renamed. Formerly: Honomanu Volcanic Series (Stearns, 1942); (partly) Honomanu Formation (Macdonald, 1978).

TABLE 1.12.—*Summary of stratigraphic units for main southeastern Hawaiian Islands*—Continued

Unit (this report)	Description	Age	Remarks
	MAUI--Continued		
	West Maui Volcano		
Lahaina Volcanics	Rejuvenated-stage lava of basanite and picritic basanite. Lava flows and associated pyroclastic deposits. Exposed in small areas on west and southeast sides of West Maui. Exposed thickness of flows: 3-20 m; cones, ~45 m. No type locality designated. Principal reference locality[1]: Puu Laina. Unconformably overlies Honolua Volcanics. Locally includes two informal flows:	Pleistocene (1.30±0.10, see text)	Lahaina Volcanics (Clague and others, 1982) (see text). Formerly: Lahaina Volcanic Series of Stearns (1942); Lahaina Group (Naughton and others, 1980).
			Abandoned as formally named units. Formerly of Lahaina Volcanic Series (Stearns, 1942):
	Laina flows and cinder cone		Laina Volcanics
	Kilea flow and cinder cone		Kilea Volcanics
Honolua Volcanics	Postshield-stage lava of mugearite, trachyte, and hawaiite. Lava flows and associated domes dikes, and pyroclastic deposits. Caps ridges on all flanks of volcano. Maximum exposed thickness: ~300 m. Type locality: village of Honolua. Overlies Wailuku Volcanics.	Pleistocene (1.18-1.20, 1.50±0.13)	Renamed. Formerly: Honolua Volcanic Series (Stearns, 1942).
Wailuku Basalt	Shield-stage lava of tholeiitic basalt, olivine tholeiitic basalt, and picritic tholeiitic basalt, and postshield-stage caldera-filling lava of alkalic basalt. Lava flows with associated intrusive rocks and pyroclastic and sedimentary deposits. Widely exposed on West Maui Volcano. Maximum exposed thickness: > 1,500 m. Type locality: south wall of Iao Valley. Base not exposed.	Pleistocene (1.32±0.04, 1.58-1.97)	Renamed. Formerly: Waikulu Volcanic Series (Stearns, 1942).
	KAHOOLAWE		
	Kahoolawe Volcano		
Kanapou Volcanics	Postshield-stage lava of alkalic basalt and hawaiite; postshield-stage caldera-filling lava of alkalic basalt; shield-stage caldera-filling lava of tholeiitic basalt; shield-stage lava of tholeiitic basalt and olivine tholeiitic basalt. Lava flows and associated pyroclastic deposits and intrusive rocks. Forms essentially all of island. Maximum exposed thickness: ~450 m. Type locality: cliffs of Kanapou Bay. Base not exposed.	Pleistocene (upper alkalic part, 1.03±0.18)	Renamed. Formerly: Kanapou Volcanic Series (Stearns, 1946; Macdonald and Davis in Avias and others, 1956).
	LANAI		
	Lanai Volcano		
Lanai Basalt	Shield-stage (including caldera-filling phase) lava of tholeiitic basalt, olivine tholeiitic basalt, and picritic tholeiitic basalt. Lava flows and associated pyroclastic deposits and intrusive rocks. Forms almost all of island. Maximum exposed thickness: > 1,000 m. No type locality designated. Principal reference locality[1]: Maunalei Gulch on northeast side of island (J.G. Moore, oral commun., 1986). Base not exposed.	Pleistocene (1.28±0.4)	Reduced in rank and renamed (see text). Formerly: (partly) Lanai Basalt and Manele Basalt (Wentworth, 1925); Lanai Volcanic Series (Stearns, 1946); Lanai Volcanic Series (which [improperly] included Lanai Basalt, and Lanai Basalt [improperly] included Manele Basalt) Macdonald and Davis in Avias and others, 1956).

TABLE 1.12.—*Summary of stratigraphic units for main southeastern Hawaiian Islands*—Continued

Unit (this report)	Description	Age	Remarks
	LANAI--Continued		
	Lanai Volcano--Continued		
	(Manele crater remnant)		Abandoned as formally named unit (see text). Formerly: Manele Basalt (Wentworth, 1925); Manele Basalt of Lanai Basalt of Lanai Volcanic Series (Macdonald and Davis in Avias and others, 1956).
	MOLOKAI		
	East Molokai Volcano		
Kalaupapa Volcanics	Rejuvenated-stage lava of alkalic basalt and basanite. Lava flows and associated cone. Makes up Kalaupapa peninsula. Maximum exposed thickness: ~125 m. Type locality: Kalaupapa peninsula. Unconformably overlies East Molokai Volcanics.	Pleistocene (0.35±0.03, 0.57±0.02)	Renamed. Formerly: Kalaupapa Basalt (Stearns, 1946, 1947a; Clague and Frey, 1982).
East Molokai Volcanics	Divided into two informal members. Upper member: postshield-stage lava of mugearite, with lesser amounts of hawaiite and trachyte. Lava flows and associated pyroclastic deposits and intrusive rocks. Lower member: shield-stage (including caldera-filling phase) lava of tholeiitic basalt, olivine tholeiitic basalt, and picritic tholeiitic basalt, and postshield-stage (including caldera-filling phase) alkalic basalt. Lava flows and associated intrusive rocks and pyroclastic and sedimentary deposits. Makes up almost all of East Molokai. Maximum exposed thickness: > 1,500 m. Type locality: cliff south of Kalaupapa peninsula. Locally overlies West Molokai Volcanics of West Molokai Volcano. Base not exposed.	Pleistocene and Pliocene (1.35-1.49, upper member; 1.52 and 1.76±0.07, lower member)	Renamed. Formerly: East Molokai Volcanic Series (Stearns, 1946, 1947a).
	West Molokai Volcano		
West Molokai Volcanics	Postshield-stage lava of hawaiite and alkalic basalt; shield-stage lava of tholeiitic basalt. Lava flows and associated intrusive rocks and pyroclastic deposits. Forms most of West Molokai. Maximum exposed thickness: > 400 m. Type locality: West Molokai mountain (Mauna Loa). Locally overlapped by East Molokai Volcanics. Base not exposed.	Pleistocene and Pliocene 1.52±0.6, 1.89, 1.84±0.7, 1.90±0.06)	Renamed. Formerly: West Molokai Volcanic Series (Stearns, 1946; 1947a)
	OAHU		
	Koolau Volcano		
Honolulu Volcanics	Rejuvenated-stage lava, ranging from alkalic basalt, basanite, and nephelinite to melilitite. Exposed on southwest and northeast flanks of Koolau Range. No type locality designated; named for exposures in city and county of Honolulu. Honolulu Volcanics consists of an assemblage of local informally named lava flows, cinder, spatter, and tuff cones, and ash deposits (see text and below; sequence listed below from Macdonald and others, 1983). Unconformably overlies Koolau Basalt.	Holocene(?) and Pleistocene (0.03-0.9) (see text)	Honolulu Volcanics (Clague and Frey, 1982) [also Lanphere and Dalrymple, 1979]. Formerly: Honolulu Volcanic Series (Stearns, 1935, 1939); Honolulu Series (Winchell, 1947); Honolulu Group (Hay and Iijima, 1968; Clague and others, 1982).

TABLE 1.12.—*Summary of stratigraphic units for main southeastern Hawaiian Islands*—Continued

Unit (this report)	Description	Age	Remarks
	OAHU--Continued		
	Koolau Volcano--Continued		
			Abandoned as formally named units see text). Formerly all in Honolulu Volcanic Series (Stearns, 1935, 1939, 1940a; see also Avias and others, 1956, for other former names):
	Hawaiiloa flow (and associated cone)		Hawaiiloa Volcanics
	Pali Kilo flow (and associated cone?)		
	Pyramid Rock flow		Pyramid Rock Basalt
	Moku Manu tuff cone		Moku Manu Volcanics
	Ulupau tuff cone (and flow?)		Ulupau Tuff (also Ulupau Head Tuff); Mokapu Basalt (part)
	Mokolea flow		Mokolea [misspelled Mokulea by Stearns (1935)] Basalt; Mokapu Basalt (part)
	Kalihi flow (and associated cone)		Kalihi Volcanics
	Haiku flow (and associated cone)		Haiku Volcanics
	Rocky Hill flow (and associated cones)		Rocky Hill Volcanics (Basalt)
	Manoa cinder cone (and lava flow?)		
	Aliamanu flow (and associated cone)		Aliamanu Tuff (Basalt)
	Kaneohe flow (and associated cones)		Kaneohe Volcanics
	Luakaha flow (and associated cone)		Nuuanu Volcanics (Basalt) (part)
	Makuku flow (and associated cone)		Nuuanu Volcanics (Basalt) (part)
	Pali flow (and associated cone)		Pali Volcanics
	Makawao tuff and breccia deposits		Makawao Breccia
	Kaau flows (and associated tuff and mudflow deposits)		Kaau Volcanics (Basalt, Tuff)
	Mauumae flow (and associated cone)		Mauumae Volcanics
	Salt Lake tuff cone		Salt Lake Tuff
	Makalapa tuff cone		Makalapa Tuff
	Ainoni flow (and associated cone)		Ainoni Volcanics
	Castle flow (and associated cone)		Castle Volcanics
	Maunawili flow (and associated cone)		Maunawili Volcanics
	Training School flow (and associated cone)		Training School Volcanics
	Diamond Head tuff cone		Diamond Head Tuff [also Diamond Head Black Ash; Black Point Ash]
	Kaimuki flows		Kaimuki Volcanics (Basalt)
	Black Point flow		Black Point Basalt
	Kamanaiki flows		Kamanaiki Basalt
	Punchbowl flows (and associated cone)		Punchbowl Volcanics
	Manana tuff cone		Manana Tuff
	Koko Crater tuff cone		Koko Volcanics (part)
	Kahauloa flows (and associated cone)		
	Hanauma tuff cone		Koko Volcanics (part)
	Koko Head flow (and associated cone)		Koko Volcanics (part)
	Kalama flow (and associated cone)		Kalama Volcanics
	Kaohikaipu flow (and associated cone)		Kaohikaipu Volcanics
	Kaupo flow (and associated conelet)		Kaupo Basalt
	Round Top cinder and ash deposits		
	Sugarloaf flow (and associated cinder cone and ash deposits)		Sugar Loaf Basalt
	Tantalus flow (and associated cinder cone and ash deposits)		Tantalus Basalt
Koolau Basalt	Shield-stage lava of tholeiitic basalt, olivine tholeiitic basalt, and rare picritic tholeiitic basalt; near top, rocks transitional between tholeiitic and alkalic basalt. Lava flows (typically thin-bedded) with associated intrusive rocks (dikes) and minor pyroclastic and sedimentary deposits. Widely exposed in Koolau Range. Maximum exposed thickness of flows: probably > 1,000 m. No type locality designated. Principal reference locality[1]: Nuuanu Pali. Locally unconformably overlies Waianae Volcanics of Waianae Volcano. Base not exposed. Includes Kailua Member:	Pleistocene(?) and Pliocene (1.8-2.7)	Renamed (see text). Formerly: Koolau Basalt (Wentworth, 1926); Koolau Volcanic Series (Stearns, 1935); Koolau Series (Wentworth and Jones, 1940; Winchell, 1947); Koolau Basalt Series (Wentworth and Winchell, 1947); Koolau Formation (Wentworth, 1951); Koolau Volcanics (Lanphere and Dalrymple, 1979).

TABLE 1.12.—*Summary of stratigraphic units for main southeastern Hawaiian Islands*—Continued

Unit (this report)	Description	Age	Remarks
	OAHU--Continued		
	Koolau Volcano--Continued		
Kailua Member	Shield-stage caldera-filling tholeiitic basalt and olivine tholeiitic basalt. Thick-bedded lava flows and dikes cutting them. Exposed near town of Kailua. Exposed thickness: ~500 m. No type locality specified. Principal reference locality[1]: Kaiwa Ridge. Reference locality[1]: east side of Puu Papaa. In fault(?) contact with Koolau extracaldera flows. Base not exposed.	Pleistocene(?)	Reduced in rank (see text). Formerly: Kailua Volcanic Series (Stearns, 1935); Kailua Volcanic Series of Koolau Volcanic Series [improperly] (Macdonald and Davis *in* Avias and others, 1956).
	Waianae Volcano		
Kolekole Volcanics	Rejuvenated-stage alkalic lava. Flows and cones. Occurs near Kolekole Pass and at south end of Waianae Range. Thickness at type locality: ~2 m. Type locality: Kolekole Pass. Unconformably overlies flows of Waianae Volcanics and alluvium.	Pleistocene	Kolekole Volcanics (Sinton, in press). Formerly: (partly) Kolekole Volcanics (Stearns, 1946; Macdonald and Davis *in* Avias and others, 1956).
Waianae Volcanics	Postshield-stage lava of hawaiite with rare alkalic basalt and mugearite; postshield-stage caldera-filling lava of alkalic basalt; shield-stage caldera-filling lava of tholeiitic basalt; shield-stage lava of tholeiitic basalt, olivine tholeiitic basalt, and picritic tholeiitic basalt. Flows and associated intrusive, pyroclastic, and sedimentary rocks. Exposed over most of Waianae Range. Base not exposed. Divided into Palehua (youngest), Kamaileunu, and Lualualei (oldest) Members.	Pliocene	Waianae Volcanics (Sinton, in press) (see text). Formerly: Waianae Volcanic Series (part) (Stearns, 1935); Waianae Volcanic Series (Stearns, 1946).
Palehua Member	Postshield-stage lava of hawaiite with minor alkalic basalt and mugearite. Flows and associated pyroclastic deposits. Occurs at higher elevations throughout Waianae Range. Maximum exposed thickness: > 300 m. Type locality: near Palehua. Conformably, locally unconformably, overlies Kamaileunu Member.	Pliocene (2.5-3.2)	Palehua Member of Waianae Volcanics (Sinton, in press). Formerly: upper member (part) of Waianae Volcanic Series (Stearns, 1935); upper member of Waianae Volcanic Series (Stearns, 1946).
Kamaileunu Member	Postshield-stage caldera-filling lava of alkalic basalt, hawaiite, and rare ankaramite, grading down into shield-stage caldera-filling lava of tholeiitic basalt and rare icelandite. Thick flows and associated rocks. Well exposed in ridges bounding Waianae, Lualualei, and Makaha Valleys. Maximum exposed thickness: > 600 m. Type locality: Kamaileunu Ridge. Reference localities: near Puu Heleakala and Pohakea Pass. Conformably, locally unconformably, overlies Lualualei Member. Includes formally named Mauna Kuwale Rhyodacite Flow:	Pliocene (3.0-3.5)	Kamaileunu Member of Waianae Volcanics (Sinton, in press). Formerly: lower member (part) and middle member (part) of Waianae Volcanic Series (Stearns, 1935).
Mauna Kuwale Rhyodacite Flow	Rhyodacite. Exposed on Mauna Kuwale and Kauaopuu ridges. Exposed thickness: ~135 m. Type locality: Mauna Kuwale. Underlies hawaiite and overlies icelandite of Kamaileunu Member.	Pliocene (see text)	Mauna Kuwale Rhyodacite Flow of Kamaileunu Member (Sinton, in press). Formerly: part of lower member of Waianae Volcanic Series (Stearns, 1935); part of upper member of Waianae Volcanic Series (Stearns, 1940a); Mauna Kuwale Trachyte (McDougall, 1963, 1964); Mauna Kuwale Rhyodacite of upper Waianae Volcanic Series (Funkhouser and others, 1968).
Lualualei Member	Shield-stage lava of tholeiitic basalt, olivine tholeiitic basalt, and picritic tholeiitic basalt. Thin-bedded flows and associated rocks. Occurs mainly in Lualualei Valley. Exposed thickness: > 450 m. Type locality: Puu Heleakala. Base not exposed.	Pliocene (3.0-3.9)	Lualualei Member of Waianae Volcanics (Sinton, in press). Formerly: lower member (part) and middle member (part) of Waianae Volcanic Series (Stearns, 1935).

TABLE 1.12.—*Summary of stratigraphic units for main southeastern Hawaiian Islands*—Continued

Unit (this report)	Description	Age	Remarks
	KAUAI		
	Kauai Volcano		
Koloa Volcanics	Rejuvenated-stage lava of alkalic basalt, basanite, nephelinite, and melilitite. Lava flows and associated pyroclastic deposits. Widely exposed over eastern two-thirds of island. Maximum exposed thickness: ~650 m. Type locality: vicinity of town of Koloa. Unconformably overlies Waimea Canyon Basalt. Locally includes Palikea Breccia Member, which underlies and is intercalated with Koloa lava flows.	Pleistocene and Pliocene (0.62, 1.21, 1.46, 2.01, 2.59) (see text)	Koloa Volcanics (Clague and others, 1982) (see text). Formerly: Koloa Series (Hinds, 1930); Koloa Volcanic Series (Stearns, 1946; Macdonald, 1949; Macdonald and others, 1954, 1960; Macdonald and Davis in Avias and others, 1956).
Palikea Breccia Member	Breccia and lesser conglomerate. Exposed in narrow bands over eastern two-thirds of island. Maximum exposed thickness: ~215 m. Type locality: Palikea ridge, about 5 km southeast of Kawaikini peak.	Pleistocene(?) and Pliocene(?)	Reduced in rank and renamed (see text). Formerly: Palikea Formation of Koloa Volcanic Series Macdonald and others, 1954, 1960; Davis and Macdonald in Avias and others, 1956).
Waimea Canyon Basalt	Shield-stage lava of tholeiitic basalt, olivine tholeiitic basalt, and picritic tholeiitic basalt; rare postshield-stage hawaiite near top. Lava flows and minor pyroclastic deposits. Widely exposed. Total exposed thickness: ~800 m. Type locality: walls of Waimea Canyon. Divided into Makaweli, Olokele, Haupu, and Napali Members.	Pliocene and Miocene(?) (3.6-5.1)	Reduced in rank and renamed (see text). Formerly: Kauai Lavas (Hinds, 1930); Waimea Volcanic Series (Stearns, 1946); Waimea Canyon Volcanic Series (Macdonald, 1949; Macdonald and others, 1954, 1960; Davis and Macdonald in Avias and others, 1956; Stearns, 1967).
Makaweli Member	Shield-stage caldera-filling lava of tholeiitic basalt, olivine tholeiitic basalt, and picritic tholeiitic basalt; postshield-stage hawaiite flow near top of unit. Represents south flank graben. Predominantly thick-bedded flows. Exposed in southwestern part of island. Maximum exposed thickness: ~450 m. Type locality[1]: West wall of Makaweli River canyon. Reference locality[1]: east wall of lower Waimea Canyon. Includes Mokuone Breccia Beds at base of and intercalated with flows. In fault contact with Olokele and Napali Members.	Pliocene (3.60-4.15)	Reduced in rank (see text). Formerly: Makaweli Formation of Waimea Canyon Volcanic Series (Macdonald and others, 1954, 1960; Davis and Macdonald in Avias and others, 1956).
Mokuone Breccia Beds	Breccia with lesser conglomerate. Exposed in narrow bands in southwestern part of island. Maximum exposed thickness: ~300 m. Type locality: walls of Mokuone Valley, west of Kahililoa.	Pliocene	Reduced in rank and renamed (see text). Formerly: Mokuone Formation (Macdonald and others, 1954); Mokuone Member of Makaweli Formation (Davis and Macdonald in Avias and others, 1956; Macdonald and others, 1960).
Olokele Member	Shield-stage caldera-filling lava of tholeiitic basalt, olivine tholeiitic basalt, and picritic tholeiitic basalt; postshield-stage hawaiite flow near top. Represents large summit caldera. Thick-bedded flows and associated pyroclastic deposits. Exposed in central part of island. Maximum exposed thickness: ~800 m. Type locality[1]: walls of upper Olokele Canyon. Reference localities[1]: walls of Poomaa, Koaie, and Waialae Streams. In fault contact with Makaweli and Napali Members.	Pliocene	Reduced in rank (see text). Formerly: Olokele Formation of Waimea Canyon Volcanic Series (Macdonald and others, 1954, 1960; Davis and Macdonald in Avias and others, 1956).
Haupu Member	Shield-stage caldera-filling lava of tholeiitic basalt, olivine tholeiitic basalt, and picritic tholeiitic basalt. Represents southeast flank caldera. Thick-bedded flows exposed on Haupu ridge. Maximum exposed thickness: ~500 m. Type locality: south side of Haupu ridge. Separated from Napali Member by buried caldera faults(?).	Pliocene	Reduced in rank (see text) Formerly: Haupu Volcanic Series (part) (Stearns, 1946); Haupu Formation of Waimea Canyon Volcanic Series (Macdonald and others, 1954, 1960).

TABLE 1.12.—*Summary of stratigraphic units for main southeastern Hawaiian Islands*—Continued

Unit (this report)	Description	Age	Remarks
	KAUAI--Continued		
	Kauai Volcano--Continued		
Napali Member	Shield-stage lava of tholeiitic basalt, olivine tholeiitic basalt, and picritic tholeiitic basalt. Mostly thin-bedded flows and associated pyroclastic deposits. Widely exposed. Maximum exposed thickness: ~800 m. Type locality : west wall of Waimea Canyon. Reference locality : northwest coast of island (Napali coast). Base not exposed.	Pliocene and Miocene(?) (4.3-5.1)	Reduced in rank (see text). Formerly: Napali Formation of Waimea Canyon Volcanic Series (Macdonald and others, 1954, 1960).
	NIIHAU		
	Niihau Volcano		
Kiekie Basalt	Rejuvenated-stage lava of alkalic basalt. Lava flows and pyroclastic deposits. Forms low coastal plain. Maximum exposed thickness: ~90 m. Type locality: town of Kiekie. Unconformably overlies Paniau Basalt.	Pleistocene and Pliocene (see text)	Renamed. Formerly: Kiekie Volcanic Series (Stearns, 1946, 1947b); Kiekie Volcanics (Clague and others, 1982).
Paniau Basalt	Shield-stage lava of tholeiitic basalt, olivine tholeiitic basalt, and rare postshield-stage alkalic basalt. Lava flows and associated intrusive rocks, minor ash. Single post-caldera stage alkalic vent at Kaeo. Comprises central highlands of island. Maximum exposed thickness: ~350 m. Type locality: east side of Paniau hill. Base not exposed.	Pliocene and Miocene (see text)	Renamed. Formerly: Paniau Volcanic Series (Stearns, 1946, 1947b).

[1]Here designated.

Addendum

Some units on the Islands of Hawaii, Maui, and Oahu listed in table 1.12 and discussed in the text were also formerly called "Formations" as follows (listed in alphabetical order): Hana Formation (East Maui Volcano, Maui); Hawi Formation (Kohala Volcano, Hawaii); Honolua Formation (West Maui Volcano, Maui); Honolulu Formation (Koolau Volcano, Oahu); Honomanu Formation (East Maui Volcano, Maui); Hualalai Formation (Hualalai Volcano, Hawaii); Kahuku Formation (Mauna Loa Volcano, Hawaii); Kau Formation (Mauna Loa Volcano, Hawaii); Kolekole Formation (Waianae Volcano, Oahu); Koolau Formation (Koolau Volcano, Oahu); Kula Formation (East Maui Volcano, Maui); Lahaina Formation (West Maui Volcano, Maui); Ninole Formation (Mauna Loa Volcano, Hawaii); Pololu Formation (Kohala Volcano, Hawaii); Waawaa Formation (Hualalai Volcano, Hawaii); Waianae Formation (Waianae Volcano, Oahu); and Wailuku Formation (West Maui Volcano, Maui). See Easton, R.M., and Gaiswinkler-Easton, M., 1983, A guide to the geology of the Hawaiian Islands--Hawaii, Maui, and Oahu: Joint Annual Meeting of the Geological Association of Canada, Mineralogical Association of Canada, and Canadian Geophysical Union, Field Guidebook, v. 2, p. 1-91.

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.

ACKNOWLEDGMENTS

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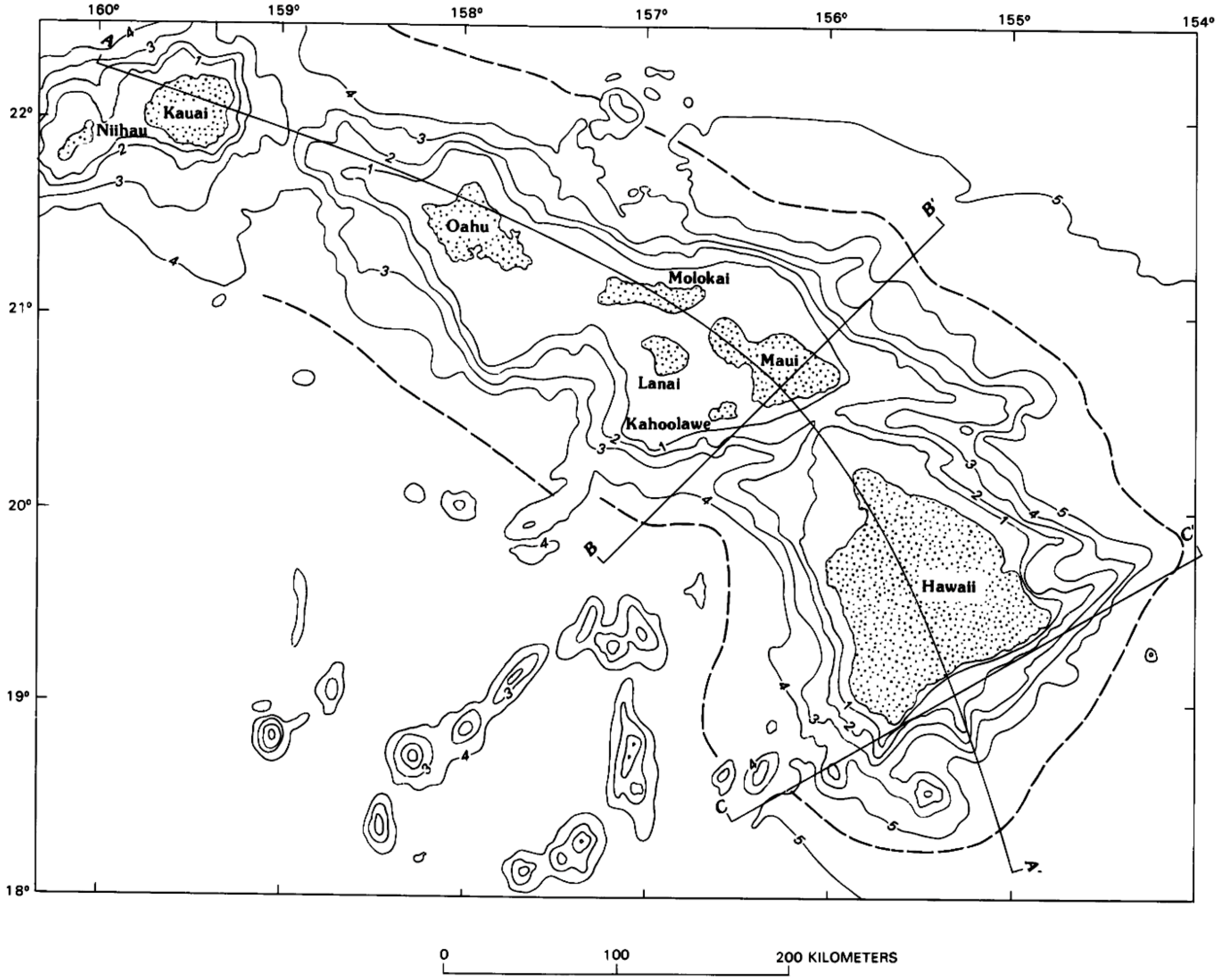

FIGURE 2.1.—Map of southeastern part of the Hawaiian Ridge showing axis of the Hawaiian Deep (dashed line) and section lines of figure 2.6. Depth contours in kilometers.

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

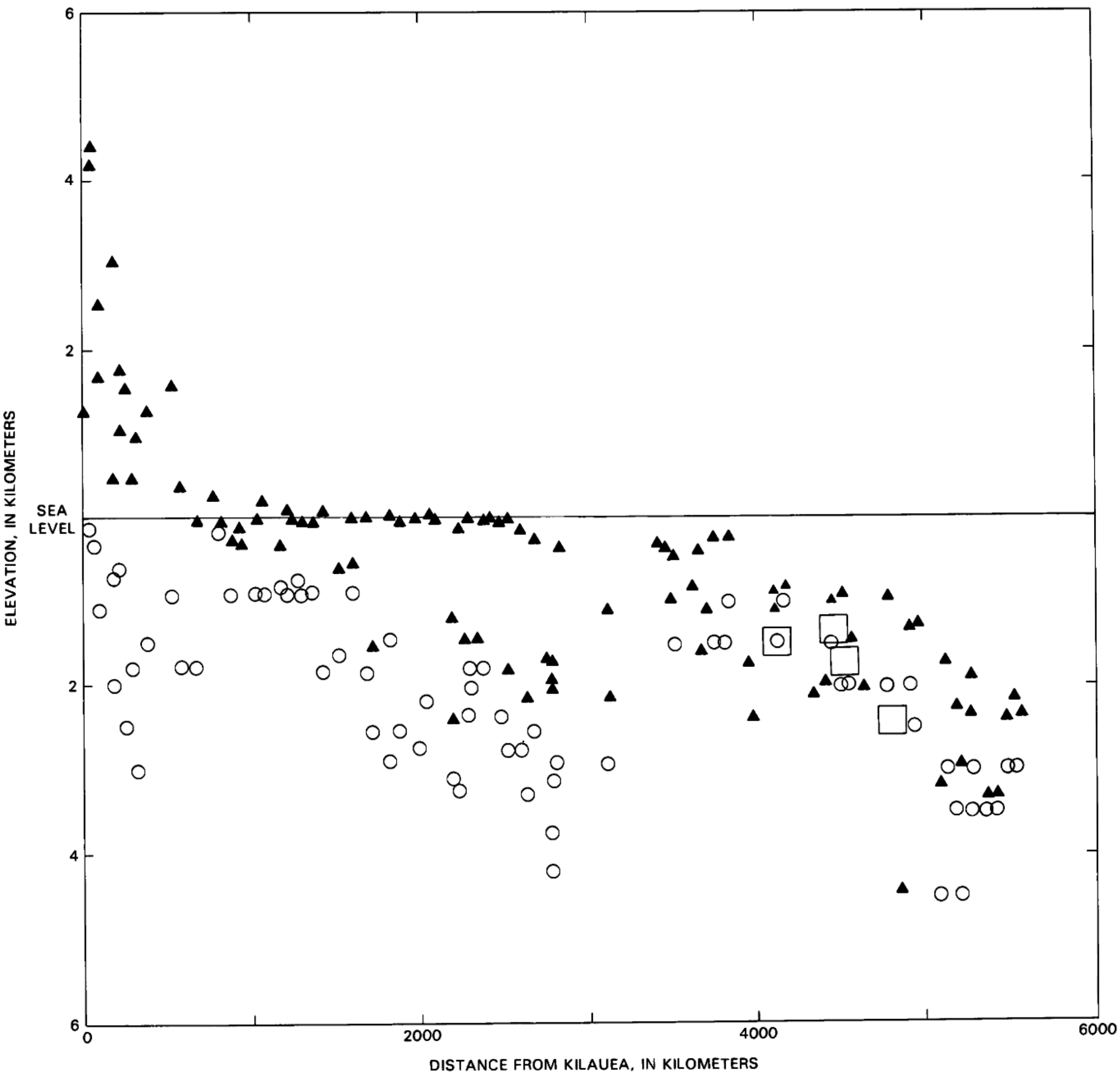

FIGURE 2.2.—Elevation and depth of volcanic features of the Hawaiian Ridge-Emperor Seamounts area. Triangles, tops of individual volcanoes; circles, depths of prominent terraces believed to have formed at sea level; squares, depth of penetration of drill holes on seamounts that encountered only subaerially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

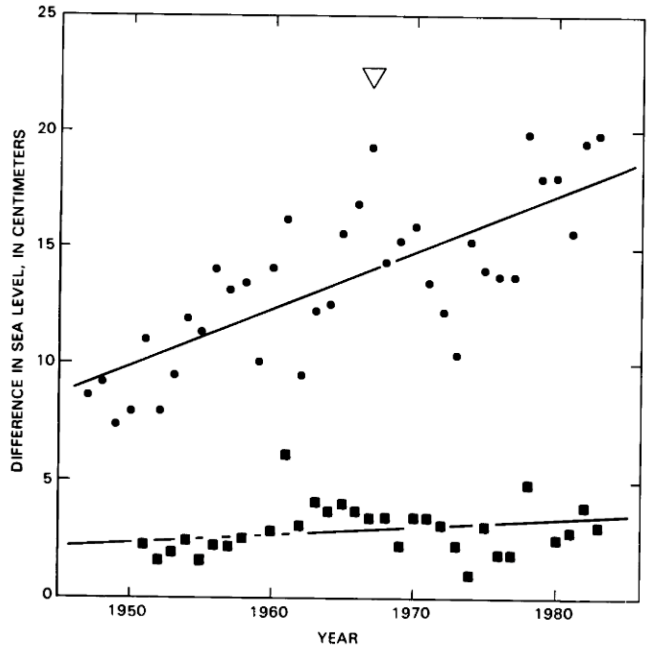

FIGURE 2.3.—Differences in annual means of sea level through 1983. Dots, means at Hilo, Hawaii, minus means at Honolulu, Oahu; squares, means at Kahului, Maui, minus means at Honolulu, Oahu. Open triangle, limit of data (1967) for analysis of Moore (1970).

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

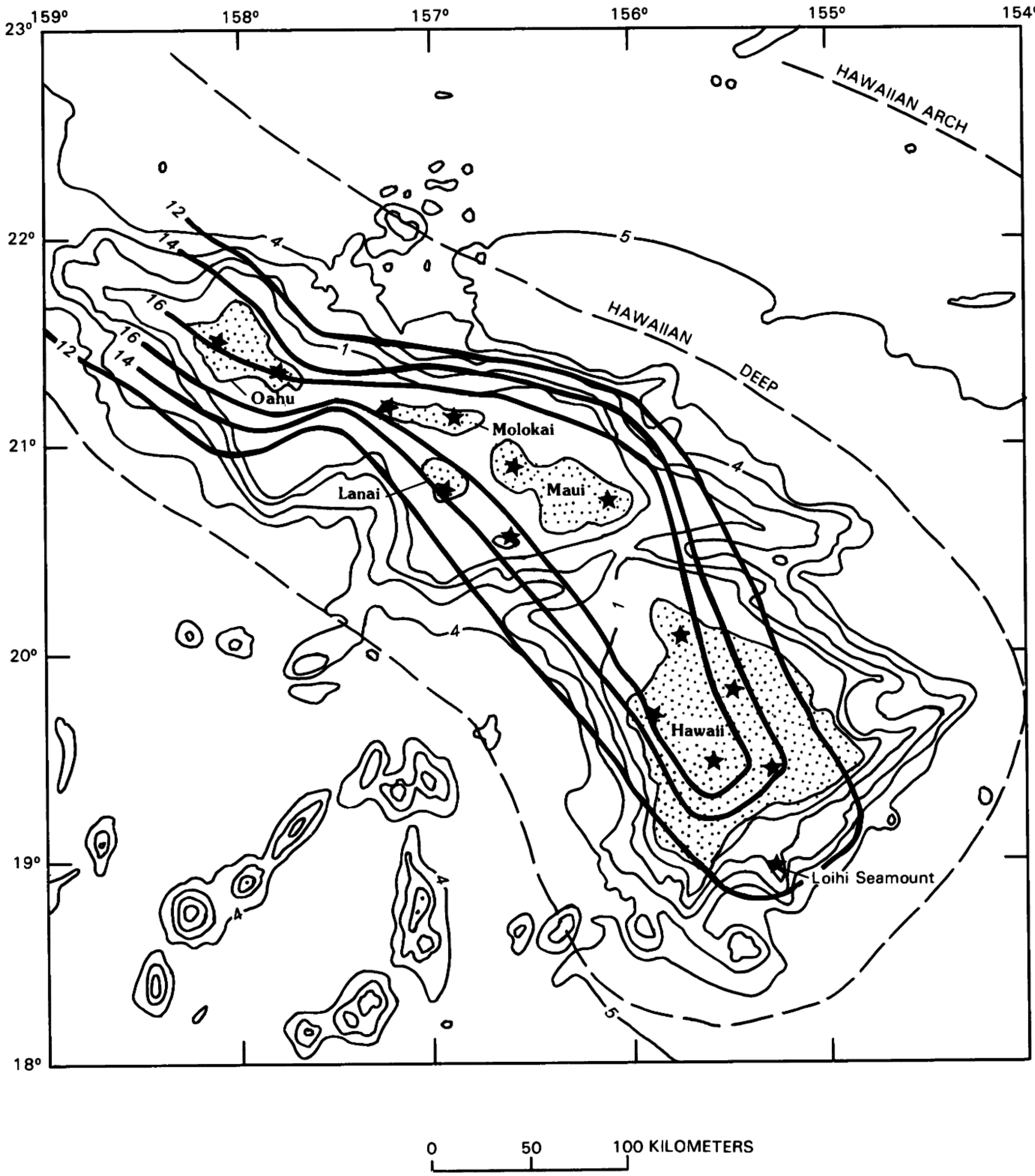

FIGURE 2.4.—Structure contours (heavy lines), in kilometers below sea level on the base of the crust, as determined by seismic refraction for the southeastern part of the Hawaiian Ridge. Data from Shor (1960), Furumoto and others (1973), and Zucca and others (1982). Depth contours in kilometers. Stars,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

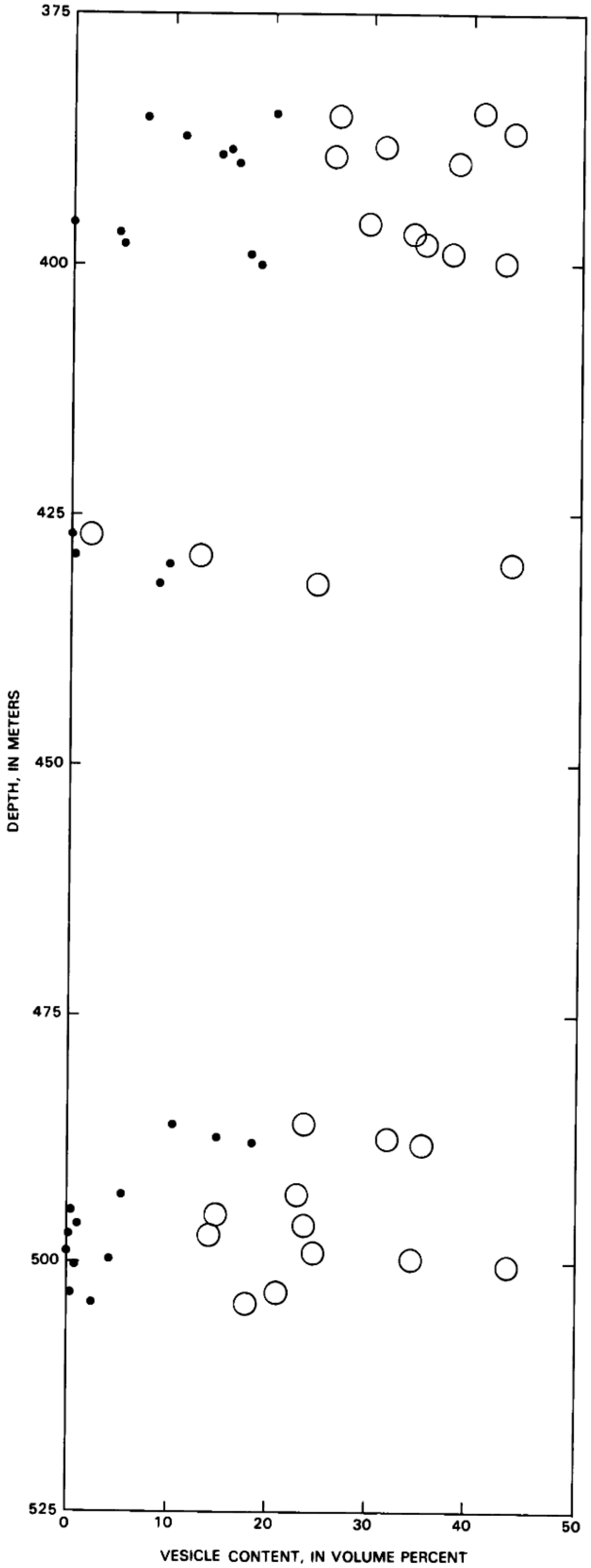

FIGURE 2.5.—Vesicularity of lava samples from 1965 drillhole on Midway Island. Open circles, volume percent of all vesicles; dots, volume percent of unfilled, open vesicles.

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

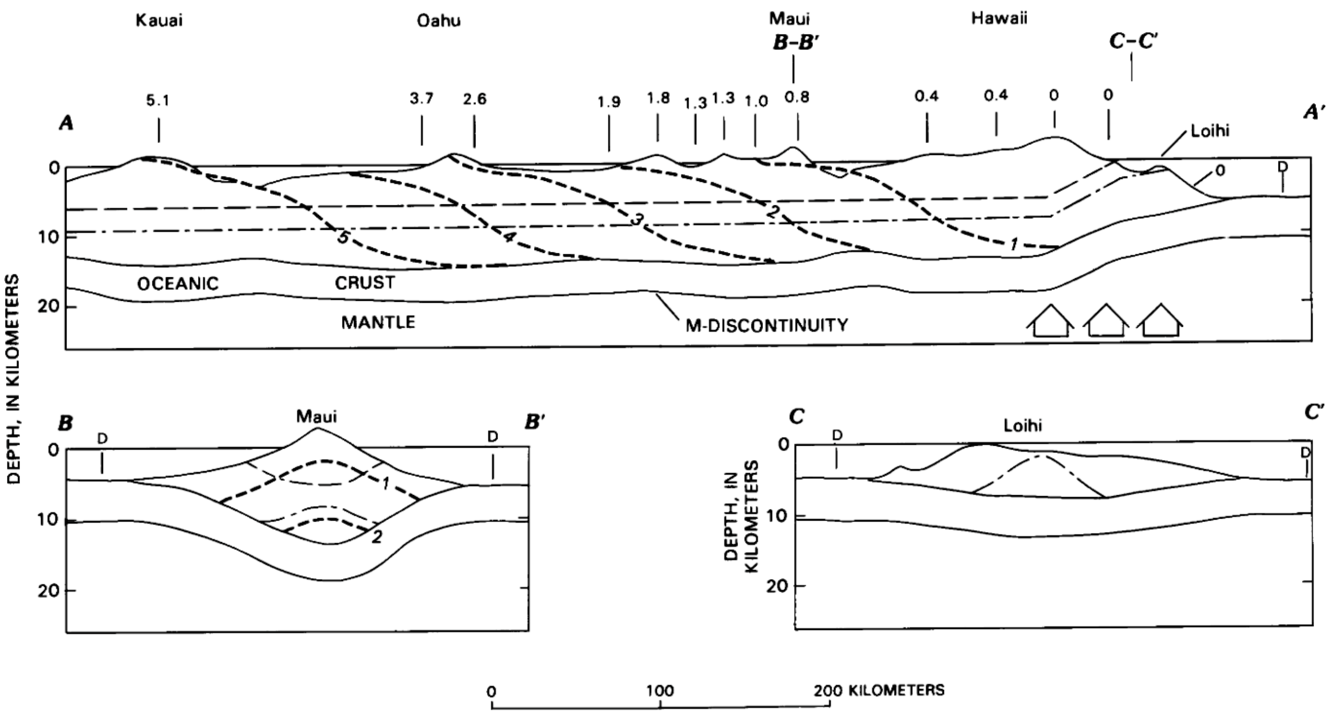

FIGURE 2.6.—Diagrammatic sections of the Hawaiian Ridge based primarily on seismic refraction and gravity measurements (Zucca and others, 1982). Location of sections shown in figure 2.1: $A-A'$, curved longitudinal crustal section along the length of the ridge; $B-B'$, transverse section through Haleakala 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

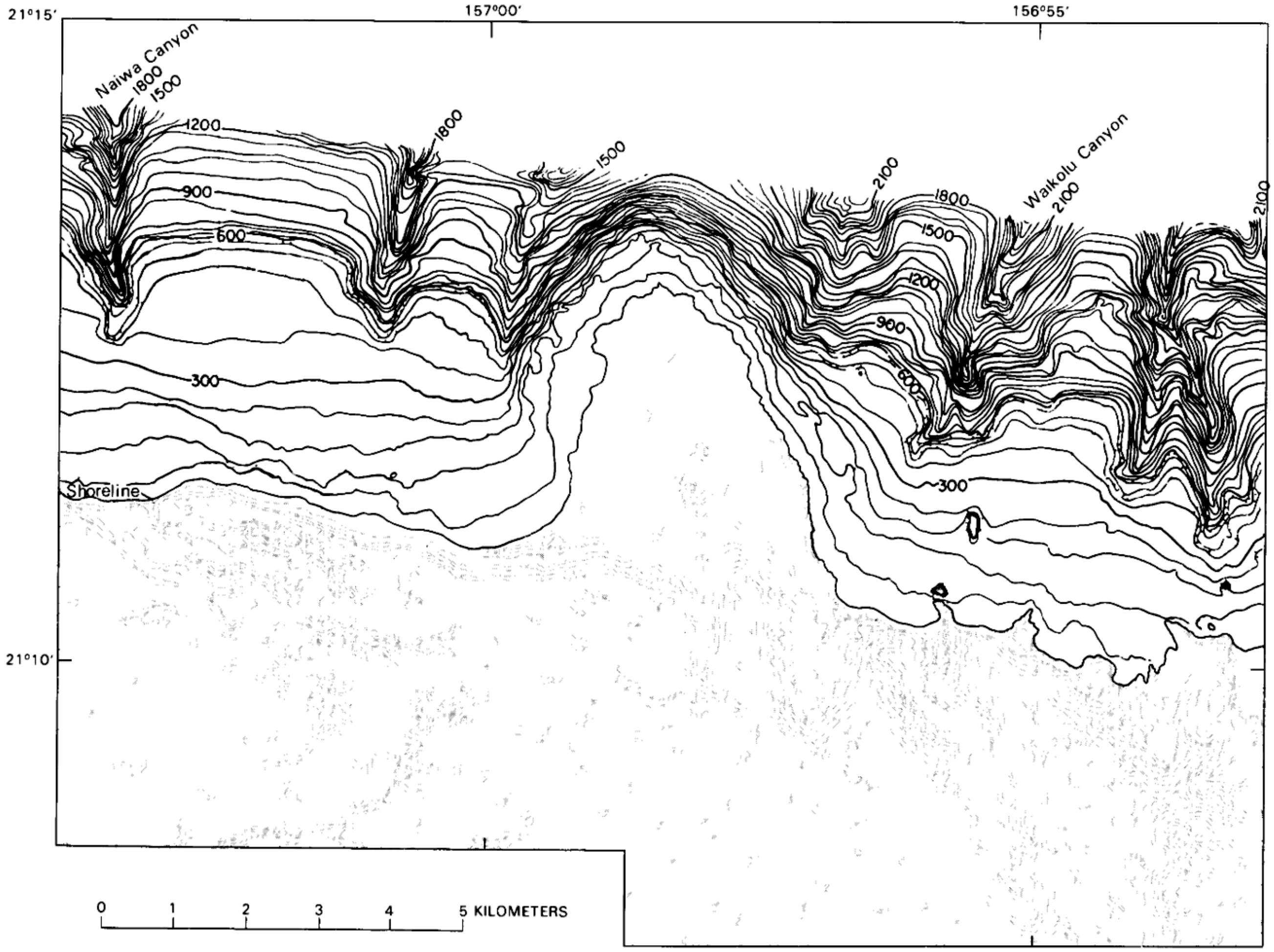

FIGURE 2.7.—Map of the Kalaupapa Peninsula on northern Molokai. Bathymetry after Mathewson (1970). Elevations and depths in feet; bathymetric contour interval, 60 ft.

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

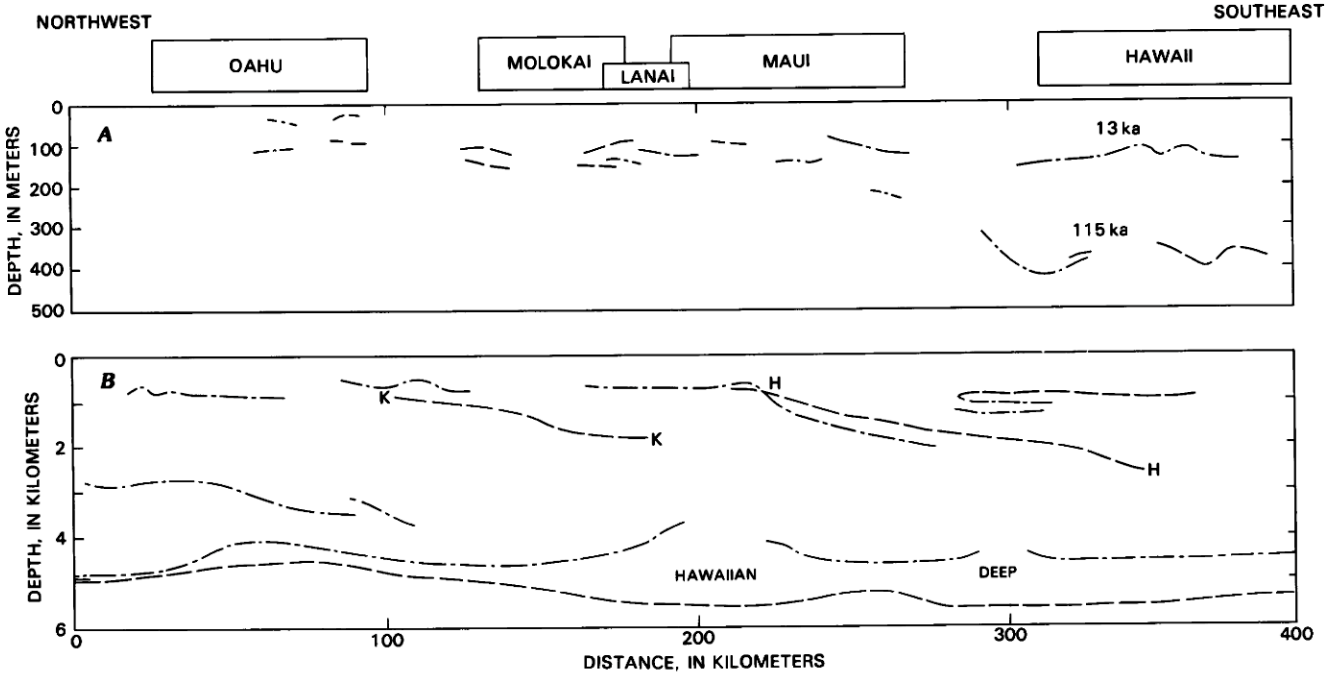

FIGURE 2.8.—Depths of terraces and axes of Hawaiian Deep projected onto a vertical plane striking N. 45° W. ***A***, Terraces in upper 500 m, including two dated terraces off west Hawaii. ***B***, All major terraces above Hawaiian Deep, including the tilted H and K terraces shown in figure 2.9. Dashed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

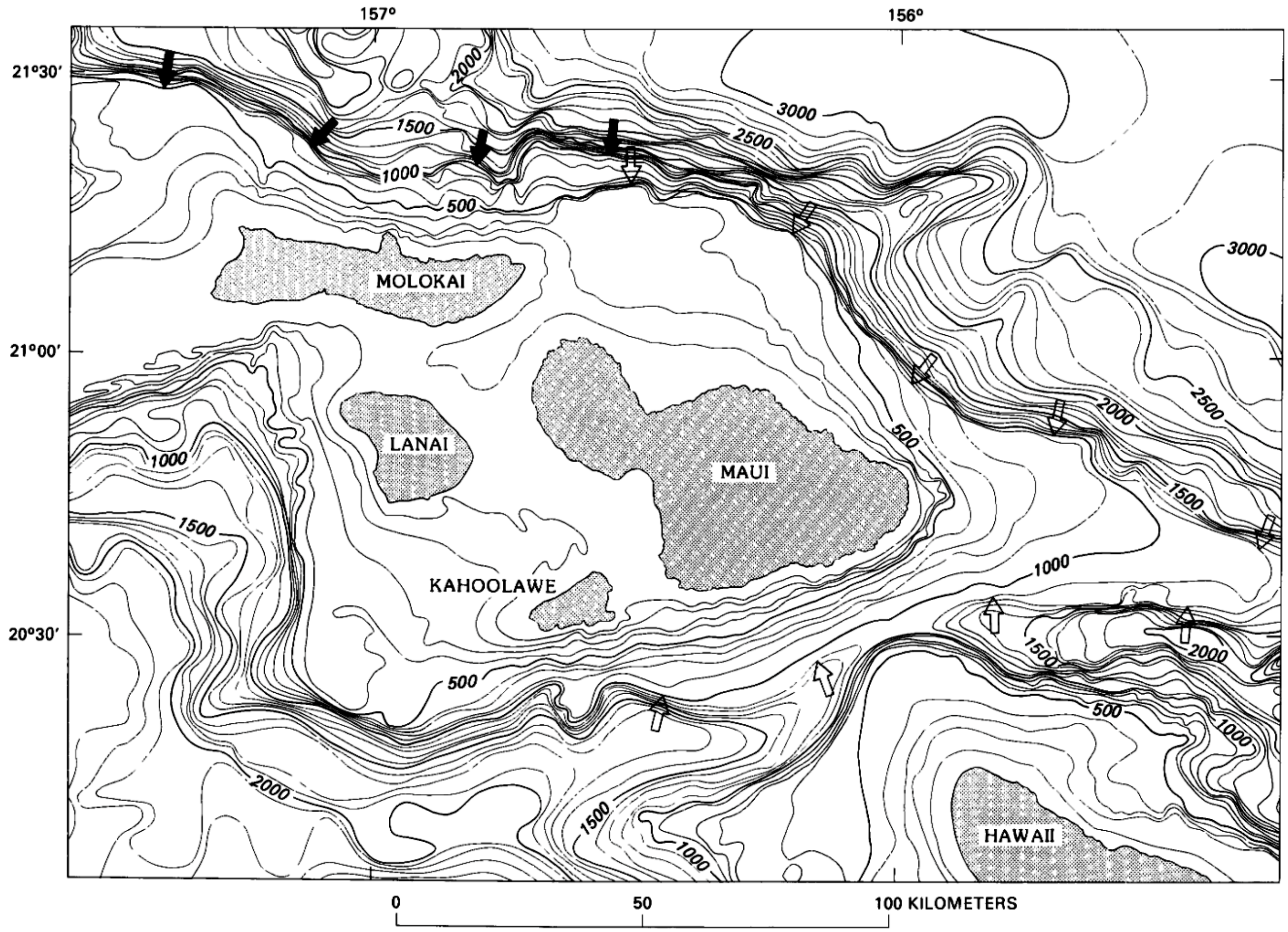

FIGURE 2.9.—Major tilted submarine terraces on southeastern Hawaiian Ridge. Open arrows, H terrace; solid arrows, K terrace. Contour interval, 100 fathoms (600 ft).

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.

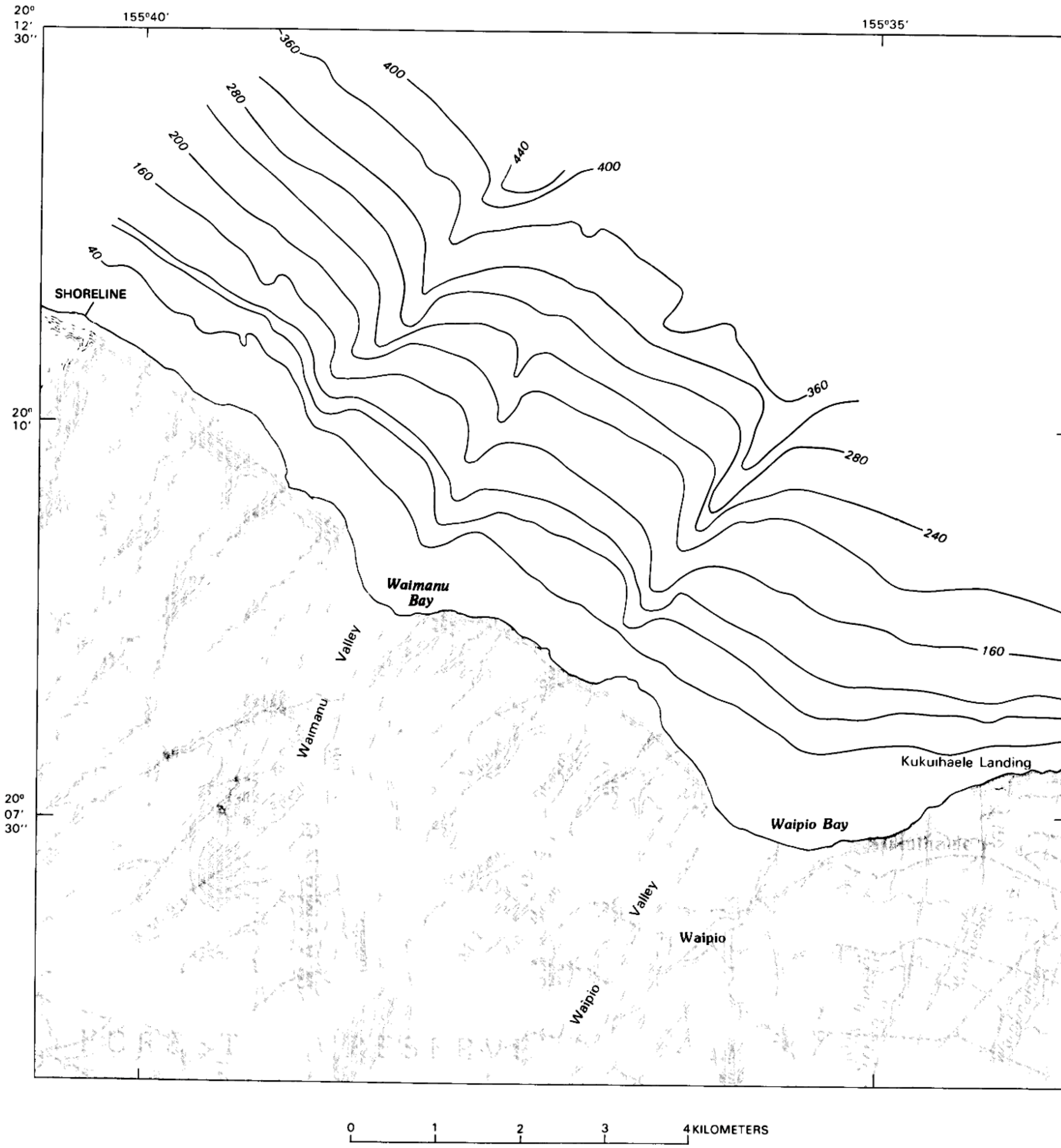

FIGURE 2.10.—Map of northeastern Kohala showing subaerial and submarine canyons. Bathymetric contour interval, 40 m.

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

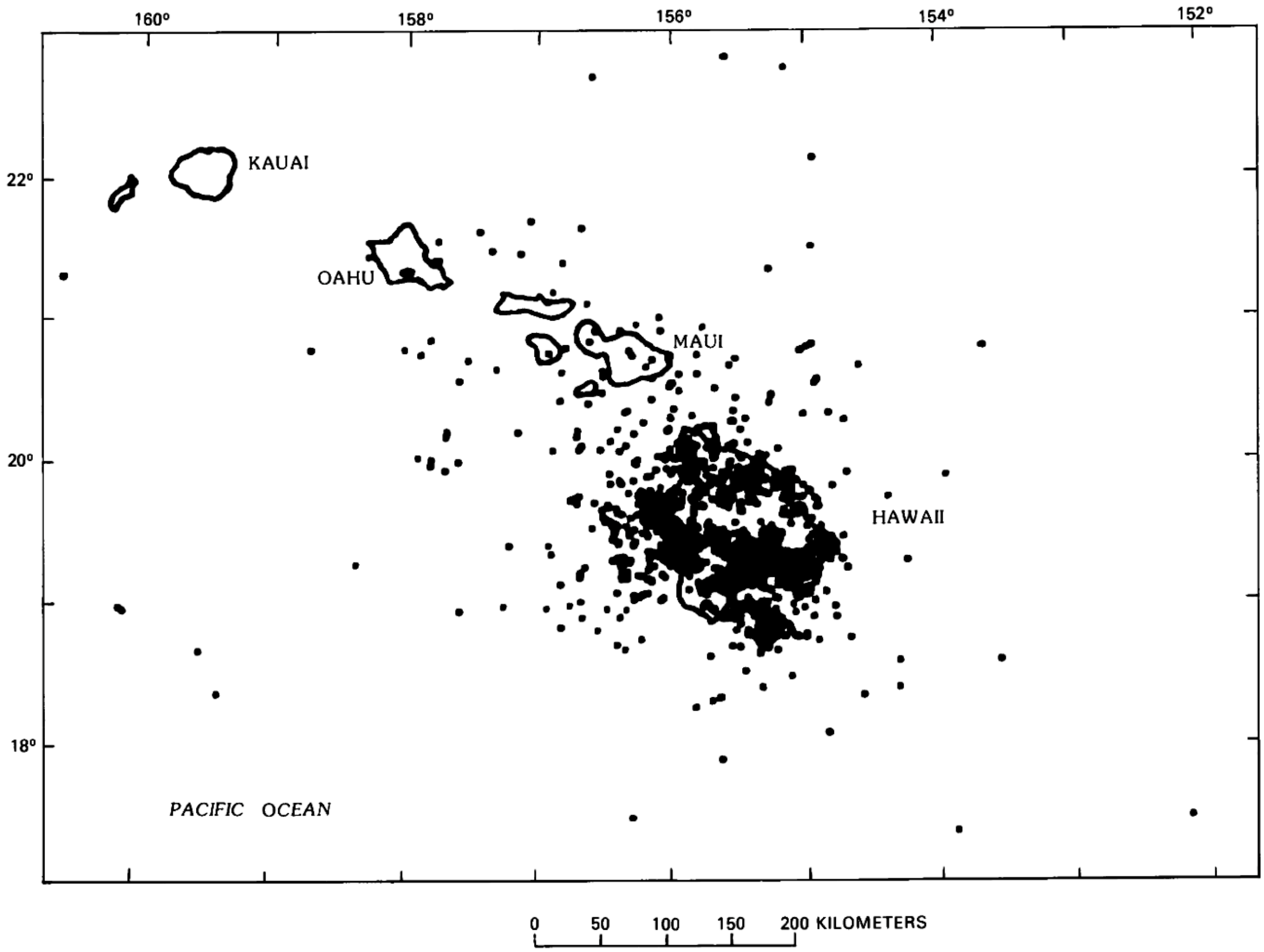

FIGURE 2.11.—Epicentral locations of earthquakes with magnitude 3 or larger in the Hawaiian region during 1968–83 (F.W. Klein, written commun., 1985).

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

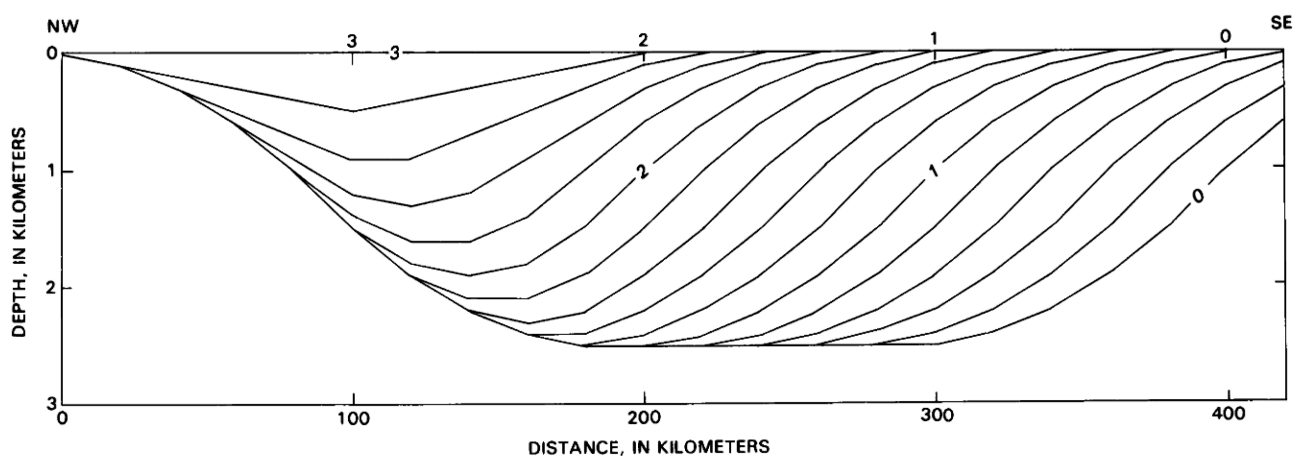

FIGURE 2.12.—Vertical section passing through center of migrating cone-like subsidence zone (100-km radius) that migrates southeast at 10 cm/yr; position of the center of the cone at specific times (Ma) is shown at top. The center of the cone subsides 2.5 mm/yr and subsidence and migration have occurred 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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