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Elevated fossil coral deposits in the Hawaiian Islands: A measure of island uplift in the Quaternary

Jones, Anthony Talbot, Ph.D.

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U·M·I
300 N. Zeeb Rd.
Ann Arbor, MI 48106

ELEVATED FOSSIL CORAL DEPOSITS IN THE HAWAIIAN ISLANDS:

A MEASURE OF ISLAND UPLIFT IN THE QUATERNARY

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By

Anthony T. Jones

Dissertation Committee:

Richard W. Grigg, Chairperson
Fred T. Mackenzie
Alexander Malahoff
Gary McMurtry
Johanna Resig

This dissertation is dedicated to Patricia E. O'Hagan and my family.

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ABSTRACT

The origin of emerged marine fossils in the Hawaiian Islands is investigated. Three alternative hypotheses tested are: (1) The Hawaiian Islands are stable and sea level fluctuations during the Late Pleistocene are responsible for the emerged marine terraces; (2) A set of giant waves swept up to 326 m on Lanai and neighboring islands depositing marine fossils 105 ka; (3) The islands have undergone uplift due to lithospheric flexure and therefore, the elevated marine terraces record interglacial periods. Fossil corals from the islands of Oahu, Molokai, and Lanai are dated by the electron spin resonance method. On Oahu, a double high stand for the last interglacial period is interpreted from ESR dates and stratigraphy; one at ~130 ka and one at ~118 ka. The Waialae reef at 14 m elevation is dated at 200 ± 20 ka. The 30-m Kaena stand is dated at 490 ± 40 ka. Based on the elevation and age of the deposits compared with geological records from elsewhere, Oahu is interpreted as having been uplifted 30 m in the last 500 ka at a mean rate of 0.06 mm yr^{-1}.

Three geomorphic terraces are identified on the south coast of Molokai. The lower two terraces at ~2 m and 10.5-12 m are assigned an age of 3.4 ka and 125 ka, respectively. The highest terrace at 20-30 m is dated at 300 ± 30 ka. The mean uplift rate there is calculated at 0.10 mm yr^{-1}.

On Lanai, corals collected from elevations of 12-74 m

correlate with the interglacial at approximately 200 ka and are interpreted as evidence for uplift during the last 300 ka.

The overall pattern of increasing elevation with age of the deposits is interpreted as uplift. The proposed mechanism for uplift is lithospheric flexure. The loading of the Hawaiian hot spot, centered under the island of Hawaii, causes local downward deformation of the lithosphere. Compensating upward lithospheric flexure occurs over a radius of approximately 200 km to 400 km, resulting in uplift of the islands of Lanai, Molokai, and Oahu that lie within this zone. This model is consistent with several independent data sets for flexure including geometry of the Hawaiian moat and arch, gravity anomalies, and seismic profiles. Therefore, the "stable island" and "giant wave" hypotheses are no longer tenable.

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INTRODUCTION

Sea level is the major reference surface on the earth separating oceanic, atmospheric, and terrestrial environments. During the last two million years of earth history, sea level has oscillated vertically by about 100 m on time scales of approximately 100,000 years (Kukla, 1977). The increase in the volume of the ocean and, therefore, the rise in sea level has been directly related to melting of the major continental ice sheets and linked directly to global climatic conditions (Shackleton and Opdyke, 1976). Pleistocene climatic cycles are attributed to orbitally controlled variations in solar insolation - the Milankovitch hypothesis (Hays et al., 1976). According to this hypothesis, the accumulation of ice sheets responds to changes in the solar input from minor changes in the geometry of the earth's orbit and axis of rotation (Imbrie and Imbrie, 1979). Present day sea level is within 10 m of the maximum height of sea level during the last 0.5 million years (Bender et al., 1979; Pirazzoli et al., 1991, 1993).

Marine carbonate sediments containing shallow-water marine fossils have been identified by Harold Stearns at unusually high elevations (up to 326 m) on several of the southern Hawaiian Islands (Stearns, 1935a, 1935b, 1961, 1974a, 1978a). Given that sea level has not been more than 10 m above its present level, these deposits, if associated with interglacials, contradict the current view of continuous subsidence after formation of the Hawaiian Islands (Moore, 1987). As reviewed by Moore (1987), subsidence in the Hawaiian Islands is attributed to lithospheric loading of the Pacific plate by the building of hot spot volcanoes, and thermal cooling of the reheated plate. The interplay of these processes results in rapid subsidence for the island of Hawaii (2.6 mm yr^{-1}) with decreasing rates of subsidence for the islands between Hawaii and Oahu (Moore, 1987).

Two competing hypotheses have been proposed to explain the marine fossil deposits and their anomalous elevations. Early workers, most notably Harold Stearns (1935a), suggested that the elevations of the various deposits were correlated between islands and that each deposit represented a specific sea level highstand. Thus, Stearns envisioned each island as a stable edifice recording the fluctuating global or eustatic sea level. This purely eustatic hypothesis does not agree with Pleistocene sea level studies from either stable oceanic islands such as Bermuda (Harmon et al., 1983, Hearty et al., 1992) or tectonically uplifted reef tracts such as Huon Peninsula in Papua New Guinea (Bloom et al., 1974) or Barbados (Radtke and Grün, 1990).

Moore and Moore (1984) proposed that the pattern of emerged marine fossil deposits on islands southeast of Oahu represented sedimentary traces of a giant wave or set of waves that swept up the coasts of Lanai and neighboring islands. A basic tenet of this interpretation is that the islands southeast of Oahu are subsiding at too high of a rate for shoreline features from previous interglacials to be preserved above present sea level. This giant wave hypothesis predicts that exposed corals on the southern islands date at circa 105 ka - the date of the giant wave event (Moore and Moore, 1988).

Lithospheric flexure was first identified around the island of Hawaii by Walcott (1970a). Loading at the hot spot[1] produces a deformation or warping in the elastic lithosphere. The Hawaiian observations have since been widely cited as an example of lithospheric flexure (see Turcotte and Schubert, 1982). However, the possiblity of flexure as the cause of uplift in the main Hawaiian Islands northwest of Hawaii has not been seriously considered.

In the last decade, the development of new dating techniques applicable to fossil corals, such as electron spin resonance (ESR), allows testing of these competing hypotheses by establishing the chronology of various uplifted deposits. ESR was selected for dating corals because the technique spans the range of ages, from the present to 1 Ma (Ikeya, 1984). This is the estimated age of the oldest fossil coral deposits in the main Hawaiian Islands. This dissertation is focused upon the development of a chronology of emerged coral-bearing deposits in the region from Hawaii to Kauai (Figure 1) and the

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[1]The term hot spot is used throughout the dissertation to describe the phenomena of mid-oceanic plate magmatic discharge that forms line island-seamount chains and is not intented to imply an origin or mechanism for the melting anomaly.

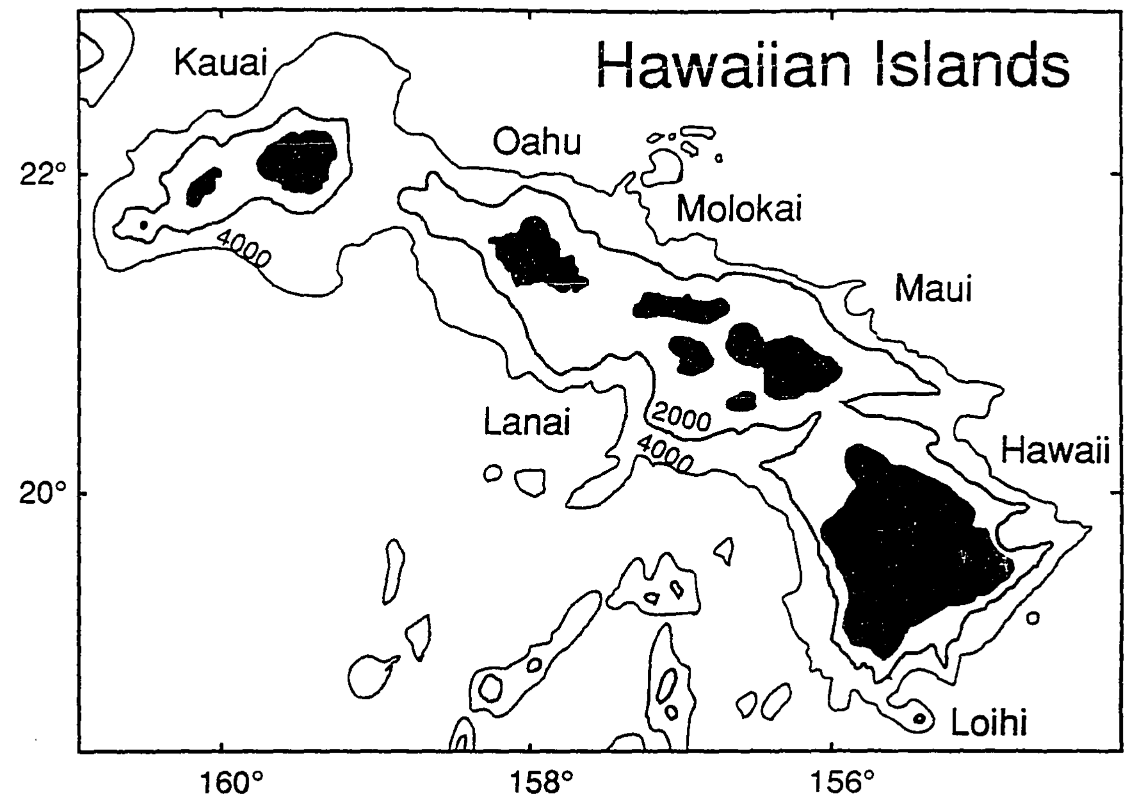


Figure 1. Map of the Hawaiian Islands showing the geographical distribution of the main islands. Bathymetry contoured at 2000 m.

possible causal mechanisms for their origin. The overall objectives of the study are: (1) to use ESR-dated corals to establish the vertical history of each island, thereby testing the "giant wave" hypothesis, (2) to examine the role of lithospheric flexure in the observed pattern of fossil coral deposits, and (3) to develop a model to account for the subsidence and uplift history of the islands.

The dissertation is organized into several sections as follows. In the first section, the subsidence history of each island along the Hawaiian Ridge is described. In the second section, the giant wave hypothesis is critically evaluated. In the third section, published dates on marine fossils from the Hawaiian Islands are reviewed. A chapter on ESR dating method as applied to corals is also presented. Results from field and dating studies of the islands of Oahu, Molokai, Lanai, and Kauai follow. The last section is a discussion of the distribution patterns found on these four islands. Lithospheric flexure is suggested as the most likely mechanism to explain the origin of emerged marine terraces in the high Hawaiian Islands.

CHAPTER II

SUBSIDENCE HISTORY

"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" (Moore, 1987).

Introduction

The construction of a volcanic island over a hot spot places a tremendous load on the underlying lithosphere. The result is an isostatic adjustment that depresses the lithosphere leading to partial sinking of the growing island. Most of the Hawaiian Islands have subsided 2-4 km since breaking the ocean surface, with the greater part of their subsidence occurring within the first 1 Ma of each island's existence (Moore, 1987). Thus, at a horizontal plate movement rate of 10 cm yr^{-1} to the northwest (Jackson et al., 1980), maximum subsidence occurs within 100 km of the hot spot.

Recently, Moore (1987) reviewed the evidence for subsidence along the Hawaiian Ridge. He used a wide variety of observations to infer subsidence rates, including tide-gauge measurements, seismic refraction, studies of deeply submerged reef terraces, cores of marine carbonates, the sulfur content of cored volcanics, and drowned cultural artifacts. Estimates for the rate of subsidence of the Island of Hawaii range from 1.8 mm yr^{-1} to greater than 3 mm yr^{-1} (Table 1); with the most recent and best supported estimate being 2.6 mm yr^{-1} (Ludwig et al., 1991).

Table 1.
Subsidence Rates for the Hawaiian Islands.

Feature and Location	Rate (mm yr^{-1})	Time Period (yr)	Reference
<u>Submerged Archaeological Features</u>			
Honoaunau, Hawaii	3	2×10^2	Apple & Macdonald, 1966
<u>Tide Records</u>			
Hilo, Hawaii	2.4	38	Moore, 1987
Kahului, Maui	0.3	38	Moore, 1987
<u>Submerged Carbonate Platforms</u>			
N.W. Hawaii	2.6	4.5×10^5	Ludwig et al., 1991
W. Hawaii	2.5	5×10^5	Moore & Campbell, 1987
Kealakekua, Hawaii	1.8-3+	1.3×10^4	Moore & Fornari, 1984
Kohala, Hawaii	2	5×10^5	Campbell, 1984
<u>Chemical Signatures in Lavas</u>			
Sulfur content	2.4	1.5×10^5	Moore & Thomas, 1988
<u>Height History of Haleakala Laser Station</u>			
Haleakala, Maui	0 ± 2	10	Smith et al., 1988

By examining submerged cultural artifacts carved in pahoehoe lava and old photographs of shorelines, Apple and Macdonald (1966) calculated a relative sea-level rise for Hawaii of 3 mm yr^{-1} for the last two centuries. They attributed this "sinking" primarily to isostatic adjustment of the island because of lithospheric loading. Recently, petroglyphs have been discovered in the intertidal zone at several locations on the Island of Hawaii (P.F. Kwiatkowski, pers. comm., 1992) indicating relative sea-level rise within the last few thousand years.

Tide-Gauge Records

Moore (1970) examined tide records from Hilo, Hawaii; Kahului, Maui; and Honolulu, Oahu for the years 1946 to 1967. The eustatic component was removed by comparing records from Oahu with tide-gauge records from the West Coast of North America. The rates of subsidence for Hawaii and Maui were calculated relative to Oahu which was assumed to be relatively stable. The results were subsidence rates for Hawaii of 4.1 mm yr^{-1} and for Maui of 1.7 mm yr^{-1} (Moore, 1970). Berg and Sutton (1977) also evaluated tide-gauge records from the Hawaiian Islands and concluded that there was "a long-term downward flexure of the Hawaiian Islands chain at its eastern end of about 4 mm yr^{-1} with respect to Honolulu." The authors carefully noted, however, that segments of the record provide opposite interpretations and further, that short-term variations on the order of tens of centimeters should be expected because plate tectonic motion is a discontinuous process. Recently, Moore (1987) extended his analysis by adding 17 years of data to his previous study. This analysis reports lower values for Hawaii and Maui than his earlier efforts, e.g. 2.4 and 0.3 mm yr^{-1}, respectively (see Table 1).

Updated tide-station records from the Hawaiian Islands are presented in Figure 2. Notwithstanding the high degree of variance for the record from Nawiliwili, Kauai (Table 2), the

Table 2.
Tide-gauge records from Hawaiian stations showing trend of rise in mean sea level.

Station	Years From	To	Records	Trend (mm yr^{-1})	Std. err
Hilo	1927	1986	45	3.46	± 0.22
Kahului	1947	1986	35	1.82	± 0.22
Honolulu	1892	1986	95	1.58	± 0.13
Nawiliwili	1955	1986	32	1.90	± 0.31

trend in mean sea level shows maximum subsidence at Hilo associated with crustal loading at the hot spot and a minimum at Honolulu. The record at Honolulu may be influenced by Oahu's position near or on the flexural arch (a region of uplift associated with lithospheric deformation) along the Hawaiian Ridge axis. The mean trend for Maui, Oahu and Kauai are close to values for historic global eustatic rise in sea level. It should be emphasized that previous accounts, especially by Moore (1970), interpret the island of Oahu as

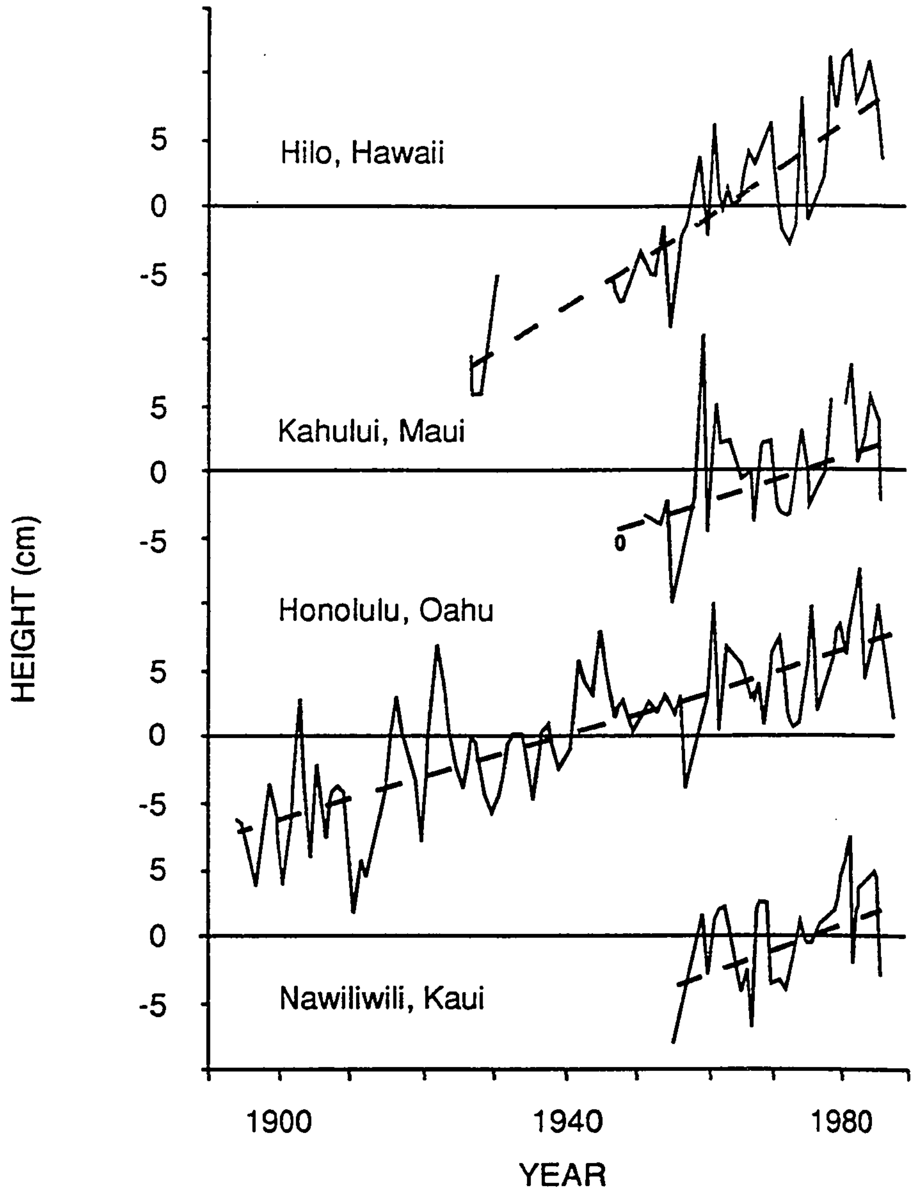


Figure 2. Tide-gauge data from Hawaii, Maui, Oahu, and Kauai.

stable and the island of Kauai as rising. Recent global sea-level rise estimated from tide gauge records range from 1.1 (± 0.1) mm yr^{-1} to 2.4 (± 0.9) mm yr^{-1} with more recent calculations around 1.8 mm yr^{-1} and higher (Table 3). Therefore, it is difficult to extract a relative sea level change for Maui, Oahu and Kauai over this period.

Table 3.
Estimates of present-day global mean sea level rise (See Fletcher (1992) for review of the problem).

Sea Level Rate (mm yr^{-1})	Reference
1.1	Gutenberg, 1941
1.5 ± 0.3	Hicks, 1978
3.0	Emery, 1980
1.1 ± 0.1	Gornitz et al., 1982
1.7	Aubrey and Emery, 1983
1.7	Barnett, 1983
1.2 ± 0.3	Gornitz and Lebedoff, 1987
1.15	Barnett, 1988
1.75 ± 0.13	Trupin and Wahr, 1990
2.4 ± 0.9	Peltier and Tushingham, 1989
1.8 ± 0.12	Douglas, 1991

Submerged Carbonate Platforms

Several authors have used deeply submerged marine terraces around Hawaii to estimate island subsidence rates. Moore and Fornari (1984), assuming a uniform subsidence rate for Hawaii, inferred rates of subsidence for deep terraces along the western portion of Hawaii. They based their study on ^{14}C dating of reef samples from the shallowest terrace at 150 m depth. Campbell (1984) calculated a subsidence rate of 2 mm yr^{-1} based on the depth of the 1000 m reef terrace and the youngest age of the subaerial portion of the Kohala volcano. By matching terrace depths with the time of glacial maxima from oxygen isotope records of Shackleton and Opdyke (1976), Campbell estimated the rates of subsidence for west Hawaii (2.43 mm yr^{-1}) and west Lanai (1.91 mm yr^{-1})(Campbell, 1986). Szabo and Moore (1986) determined the age of reef samples from the -360 m terrace off northwest Hawaii and calculated a subsidence rate for northern Hawaii (2.7 mm yr^{-1}). Recently, Ludwig and others (1991) dated a series of submerged reef terraces off northwest Hawaii. The six reef terraces, ranging in age from 17 to 475 ka, suggest that Hawaii has subsided approximately 2.6 mm yr^{-1} for the past 475 ka.

Because of the potential for differential subsidence along the Hawaiian Ridge, Moore (1987) cautioned against age correlation of submerged terraces based only on depth. In fact, some terraces tilt toward the region of greatest volcanic loading. This may be true of the terraces offshore of West Maui, designated the "H" and "K" terrace (Moore and Campbell, 1987; Moore, 1987). Although the actual ages of the tilted terraces have not been determined, Moore (1987) set limits for the terrace ages based on subsidence rates and the age of the volcanic pedestal on which they developed.

Drilling and Dredging

Moore and Thomas (1988) analyzed the sulfur content in drill holes on Kilauea's East Rift Zone. They classified sulfur-poor degassed lava as having erupted subaerially, and sulfur-rich lava erupting in submarine environments. From a sulfur profile in the rock core, Moore and Thomas (1988) interpreted submarine-subaerial transitions and determined a subsidence rate of 2.4 mm yr^{-1} over the last 200 ka.

Reef carbonates, 1280 m below sea level, were recovered from a scientific observation hole drilled in the geothermal field along the Puna Ridge of Kilauea (D. Thomas, pers. comm., 1990). An _in situ_ limestone section, approximately 180 m thick, was composed of coralline algae, corals, shallow-marine foraminifera, mollusk shells, and sponge spicules (J. Resig, written comm., 1991). The thickness of the marine section indicates that more than one marine highstand must have been required to build this deposit of shallow-water carbonates. The subsurface geological evidence implies that Kilauea subsided at least 1200 m probably within the last 0.5 Ma (assuming a subsidence rate of 2.4 mm yr^{-1}). The reef developed on an antecedent foundation of the volcanic predecessor to Mauna Loa and was subsequently buried by Kilauea lavas. Down faulting of the carbonate section has been proposed to partially explain the location of the marine unit (D. Thomas, pers. comm. 1992).

Discussion

In summarizing the available information on subsidence along the Hawaiian Ridge, Moore (1987) concluded that subsidence was fairly rapid for the initial 1 Ma of volcanic shield building. As the volcanoes are carried off the hot spot by the northwestward movement of the Pacific Plate, the rate of subsidence declines rapidly towards Oahu. The most recent estimate for subsidence shows that the island of Hawaii sinks at a rate of a few millimeters per year (2.6 mm yr^{-1}). The island of Maui is intermediate to Hawaii and Oahu. In the past Oahu has been assumed to have been stable for at least 120 ka (Veeh, 1966; Ku et al., 1974; Stearns, 1978a; Moore, 1987). The Moore model for differential but constant subsidence is evaluated in this thesis by dating fossil reef terraces along the Hawaiian Ridge.

Subsidence rates based on various sources of data are plotted as a function of distance along the Hawaiian Ridge in Figure 3. The plot displays a subsidence-rate curve for islands along the Hawaiian Ridge from Hawaii to Kauai. Tide-gauge records from four stations (D. Cox, pers. comm. 1991) provide short-term measures of subsidence when corrected for a global sea-level rise of 1.8 mm yr^{-1} (Douglas, 1991, see Table 3).

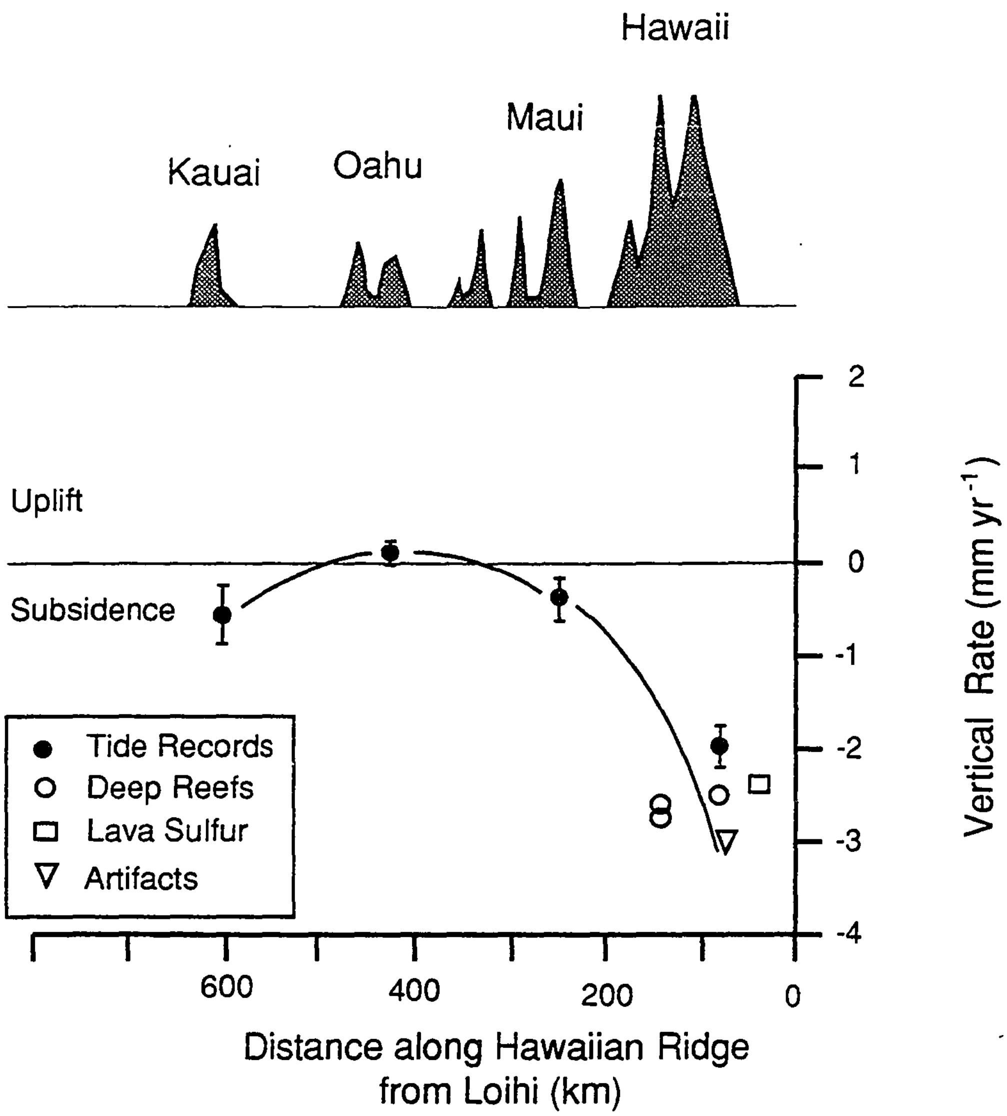


Figure 3. Subsidence rate curve for islands along the Hawaiian Ridge. Data sources given in text.

The resulting subsidence-rate curve for the southeast Hawaiian Islands shows rapid subsidence near the Hawaiian hot spot. A sharp decline in subsidence rates is observed at about 200 km from Loihi, the present site of Maui. A small amount of uplift is suggested at Oahu possibly corresponding to the position of the lithospheric bulge.

Thermal contraction of the oceanic lithosphere (Sclater et al., 1971; Parsons and Sclater, 1977) yields a subsidence rate of approximately 0.03 mm yr^{-1} for the first 5 Ma. This analysis treats the Hawaiian hotspot as if reheated to that of a 25 Myr-old lithosphere. Subsidence of islands and atolls north of Kauai are probably governed principally by cooling of the plate (Dietrick and Crough, 1978; Crough, 1978) with subsidence rates of 0.03 mm yr^{-1} for Midway Atoll (Ladd et al., 1967) at the northern end of the Hawaiian chain and 0.04 mm yr^{-1} for Kōkō Guyot in the southern Emperor Seamounts (Davies et al., 1972). Lithospheric thinning under the hotspot (Dietrick and Crough, 1978) has been suggested as mechanims for subsidence of the Hawaiian Islands. However, neither of these mechanisms, cooling plate or thinning plate, account for the observed emergence of volcanic islands within the Hawaiian Islands. Others have suggested mid-plate tectonics (Hšu and Schlanger, 1968, Menard, 1973), lowering of eustatic sea level (Stearns, 1974a, 1978a) or both (Schlanger and Douglas, 1974). For the south Pacific atolls, McNutt and Menard (1978) demonstrated that local tectonic effects caused by adjacent loading volcanic centers can cause flexure to result in uplift of nearby atolls up to 70 m. The primary mechanism controlling the vertical motion of the islands in the southeastern Hawaiian Ridge appears to be lithospheric flexure.

CHAPTER III

GIANT WAVE HYPOTHESIS

Stearns (1938) identified fossiliferous limestones on Lanai and interpreted the deposits as several ancient marine strandlines. Earlier, Wentworth (1925a) had mentioned coral fragments 46 m above sea level on Lanai, but concluded that sea level never stood more than 3 to 5 m above the present level. Stearns (1938) assigned names to two unique strandlines, the higher one being the "Mahana" stand at 365 m, and the lower one, the "Manele" stand, at 170 m. The Mahana stand was described as "several vein-like fillings of fossiliferous marine limestone in crevices in basalt" (Stearns, 1938, p. 618). The outcrops, exposed by a cattle path, were "only a quarter to half an inch wide and 2 to 3 feet long" and contained coral and shell fragments. Molluscan fossils identified from fragments collected by Stearns are listed in Table 4. The location of Stearns type locality for the Mahana stand is shown in Figure 4. Three attempts to relocate this type section have been unsuccessful.

The Manele stand was described from several locations on Lanai including the southwestern section of Kaluakapo Crater; in Kawaiu Canyon, where the deposit was reportedly 45 m thick; and two unnamed gulches that drain into Huawai and Poopoo Bays along the eastern side of Lanai (Figure 4).

Combining extensive field work throughout the Hawaiian Islands, Stearns in 1978 speculated on the chronological

Table 4.
Molluscan taxa identified from the +326 m Mahana site on Lanai (from Stearns, 1938).

Gastropoda:	
	Modulus tectum Gmelin
	Triforis sp.
	Strombus hellii Rousseau
Pelecypoda:	
	Pinctada sp.

sequence of the marine shorelines and terraces in Hawaii that he had described (Figure 5). By 1978, only 5 of the 35 named terraces were assigned dates based on limited radiometric dating.

In 1984, James G. Moore and George W. Moore published an article in *Science* in which they proposed that a giant wave or set of waves was (were) responsible for depositing marine fossils at unusually high elevations on Lanai and on nearby islands (Figure 6, Moore and Moore, 1984). A major premise of their argument was that the islands southeast of Oahu were subsiding too rapidly to leave a rock record of previous interglacials above present sea level. The authors base their argument on inter-pretation of tide gauge measurements and ^{14}C dated submerged reef terraces.

Tide gauge histories are available from several stations: Hilo, Hawaii; Kahului, Maui; Honolulu, Oahu; Nawiliwili, Kauai; and Point Allen, Kauai (Cox, pers. comm. 1991), but

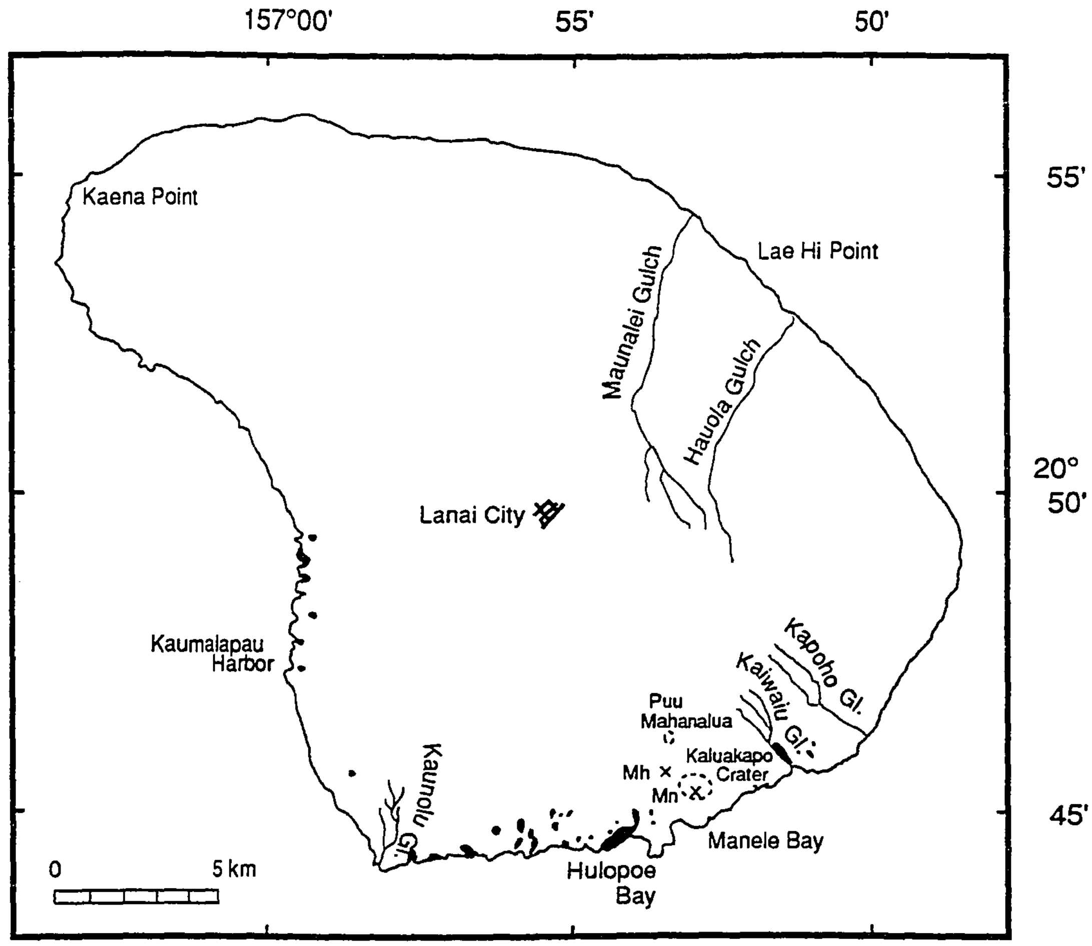


Figure 4. Map of Lanai showing type locality of Mahana (Mh) and Manele stand (Mn) and location of Manele marine deposits (from Stearns, 1938; 1940).

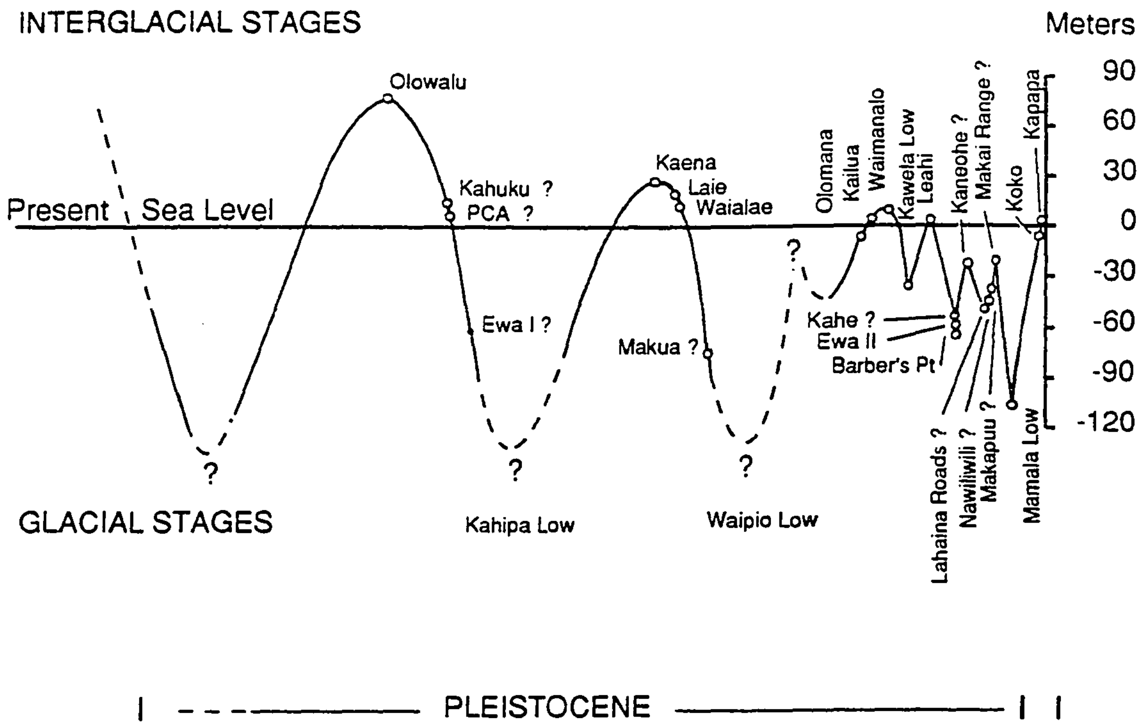


Figure 5. Sea level curve for Hawaiian shorelines as speculated by Stearns (1974a, 1978a).

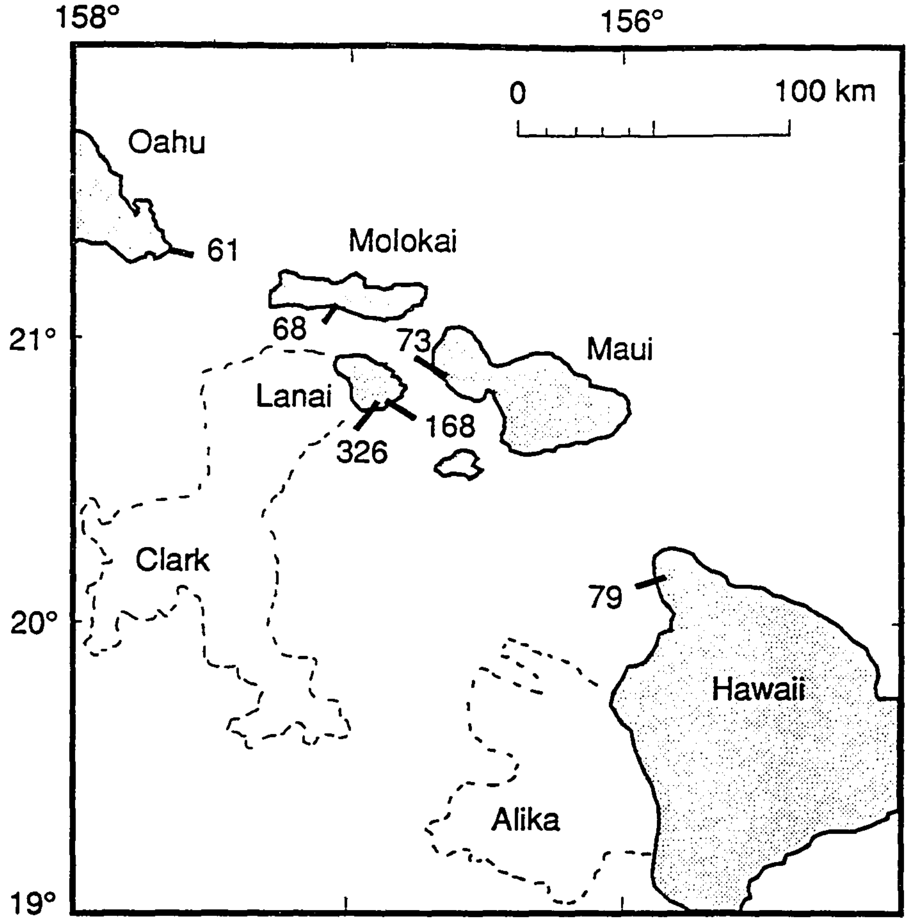


Figure 6. Map of southeastern Hawaiian Islands showing maximum height, in meters, of limestone clasts in the 'Hulopoe Gravel' on Lanai and tentatively correlated high-elevation limestone-bearing deposits on other islands (after Moore and Moore, 1984). Major submarine debris avalanches identified as possible source regions for the giant wave or waves are outlined by the dashed lines.

there are no records from either Lanai or Molokai. Extrapolating from one island to a nearby island may be misleading given differential subsidence/uplift along the Hawaiian Ridge.

The deeply submerged marine terraces that have been dated are restricted to the Island of Hawaii and imply rapid subsidence for that island (as do the tide gauge measurements at Hilo). Campbell (1986) and Moore and Campbell (1987) attempted to extend the age-depth relationship from the flight of terraces off northwest Hawaii to a series of submerged terraces off northwest Lanai. This must be questioned because the terraces have not been dated and there is no a priori reason to infer a subsidence rate for Lanai.

Moore and Moore (1984) recognized limestone-bearing gravel beds on Lanai, designating them the 'Hulopoe Gravel', and outlined several of their characteristics as follows: (1) 95 percent of the gravel was basalt with 5 percent "conspicuous white limestone boulders", (2) marine fossils occur as clasts and were not in growth position, and (3) cements, previously attributed to marine encrustation, extend into the bedrock and their lithology suggests a "calcite-cemented soil horizon".

Moore and Moore (1984) used high-elevation marine deposits identified by Stearns on adjacent islands to collaborate the magnitude of the giant wave event. The authors referenced a 65 m conglomerate bed on Molokai above Kaunakakai, large blocks of limestone talus at 61 m on Oahu at

Diamond Head, the 'Olowalu' stand at 73 m on Maui, and a site near Mahukona on Hawaii at 79 m in support of their hypothesis. They suggested from field observations that the Maui and Hawaii deposits "closely resemble" the 'Hulopoe Gravel' on Lanai. Stearns conceded in a letter to J. Moore that the elevation on Oahu of 61 m was probably in error due to problems with the altimeter. Stearns estimated that a more likely elevation for the coral-reef clast at Diamond Head is about 20 m. However, in his second edition of The Geology of the State of Hawaii, Stearns (1985) tended to agree with the giant wave hypothesis and suggested a coastal slide on the south west coast of Lanai caused a giant surge of waves.

Age of the Wave Event

Moore and Moore (1988) dated three coral clasts from the Hulopoe Gravel by uranium series. The coral clasts were from Kawaiu Gulch and Kaluakapo Crater at elevations above 100 m and dated at 101, 108 and 134 ka (Table 5). From these data,

Table 5.
Uranium-series ages of coral clasts in Hulopoe Gravel, Lanai (Moore and Moore, 1988).

Sample locality	Elevation (m)	Calcite percent	Age (ka)
Kawaiu Gulch	115	<3	108 ± 5
Kawaiu Gulch	120	<3	101 ± 4
Kaluakapo Crater	155	5	134 ± 7

Moore and Moore (1988) concluded that the time of the giant waves which swept up on the Island of Lanai was approximately 105 ka. The high elevation deposits on the other islands tentatively correlated to this event have never been dated (see Figure 6). The arguments used by Moore and Moore (1988) to establish the timing of the giant wave are open to debate. The most serious problem is that the age of the corals in the Hulopoe gravel give the time when the coral lived, not when the coral was transported to it's present locality. The three coral ages from Lanai are not significantly different from the last interglacial age from Oahu (Ku et al., 1974) at the 95% confidence level (t-test) (Figure 7). The interpretation of the coral ages for the Hulopoe gravel should correctly be that the Hulopoe gravel is no older than the youngest coral (101 ka) and that the Hulopoe gravel could have been deposited at any time since then. One way to verify the age of the alleged event would be to use exposure dating techniques applied to the basalt boulders in the Hulopoe Gravel (e.g. ^{3}He, Kutz et al., 1990; ^{36}Cl, Phillips et al., 1990).

Possible Source of the Wave Event

Tsunami impulse-generated wave trains can result from earthquakes, explosive volcanism, submarine landslide, and bolide-water impacts. Giant waves from a massive submarine landslide along the Hawaiian Ridge have been proposed as the mechanism for depositing the marine fossils at high elevations in the southeastern Hawaiian islands. Moore and Moore (1984)

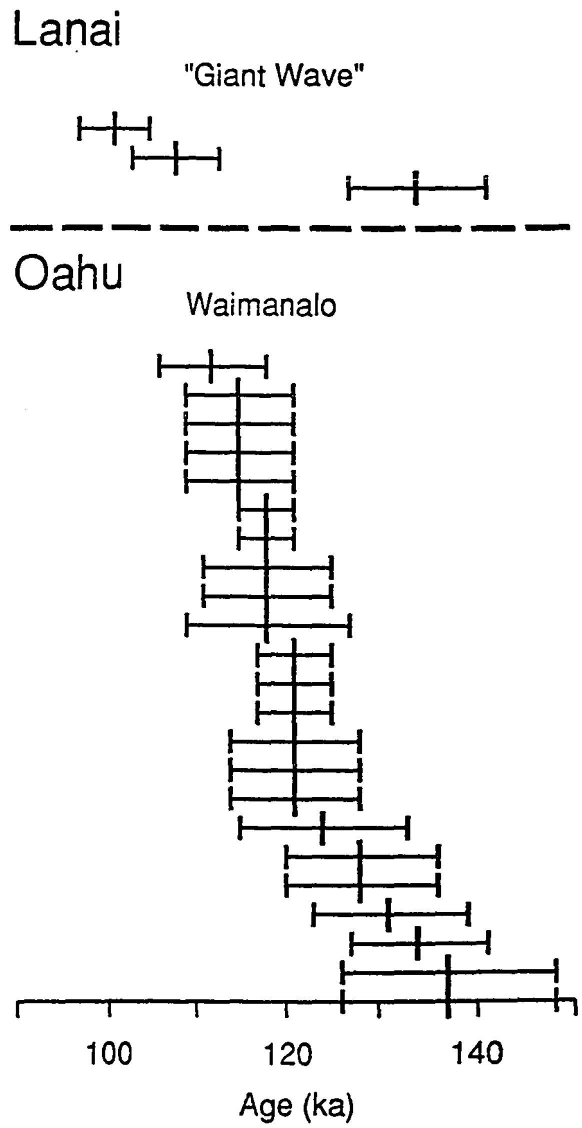


Figure 7. Display of coral dates from 'Hulopoe Gravel' on Lanai (Moore and Moore, 1988) and the Waimanalo limestone from Oahu (Ku et al., 1974).

initially identified a submarine slide off Lanai as a source region for generating the wave which washed up 326 m on the southern coast of Lanai (see Figure 6). They discounted seismic sea waves generated from an earthquake or an extraterrestrial bolide impact. Later, Moore and others (1989) suggested that the giant submarine Alika debris avalanche off Hawaii, identified from a GLORIA survey of the Hawaiian Islands, was the source for the giant waves.

Young and Bryant (1992) speculate that the Lanai giant wave destroyed last interglacial-aged (125 ka) sand barriers along the southern coast of New South Wales, Australia. They report scouring up to 15 m on coastal abrasion ramps from the northern side of Tura Head. The timing of the erosional features is correlated to the Lanai giant wave; although this interpretation is highly tenuous. A very active seismic region just north of the New South Wales coastline, the subducting margin of the Samoa-Tonga-Kemadec arc (Soloviev, 1970), is a much more likely candidate for the source of the New South Wales tsunamis. A point source tsunami generated in Hawaiian waters after traveling over 14,000 km to New South Wales would be greatly reduced in amplitude. Assuming the energy of the wave would dissipate at a rate proportional to the radius, $r^{-5/6}$ (Van Dorn, 1961, as cited in Tucholke, 1992), a 326 m wave would approach New South Wales with an amplitude of 0.1 m. This analysis, however, must be considered first order and approximate because it does not consider the complex bathymetry of the southwestern Pacific. Nevertheless, the magnitude of the decrease in wave amplitude casts serious doubt on Young and Bryant's hypo-thesis.

Soil Erosion

As further evidence to support the giant wave hypothesis, Moore and Moore (1988) state that "about 2 m of soil and weathered basalt were removed" by the wave in a section more than 2 km wide. They contrast the 2-3 m deep soil at higher elevations on Lanai with the apparent absence of soil on the southern slope of Lanai, reasoning that the stripped soil and weathered rock resulted from the same catastrophic event that deposited the 'Hulopoe Gravel'. Erosion of this soil on the south coast of Lanai occurs up to elevations of 365 m as described by Stearns (1938). The most likely agent for soil removal is wind.

The pattern of soil erosion on Kahoolawe was also used as an indication of the direction of the approaching wave train from the south west (Moore and Moore, 1988, Figure 8). The present eroded and wind swept appearance of Kahoolawe devoid of vegetation, contrasts with the early descriptions of the landscape as a savannah of grasslands, shrubs and trees (see Spriggs, 1991). The introduction of sheep on Kahoolawe in the late 1800's combined with periods of drought led to extensive soil erosion (Spriggs, 1991). By the time Harold Stearns visited the island and mapped the geology in the late 1930's, the island had lost a substantial amount of soil. The recent

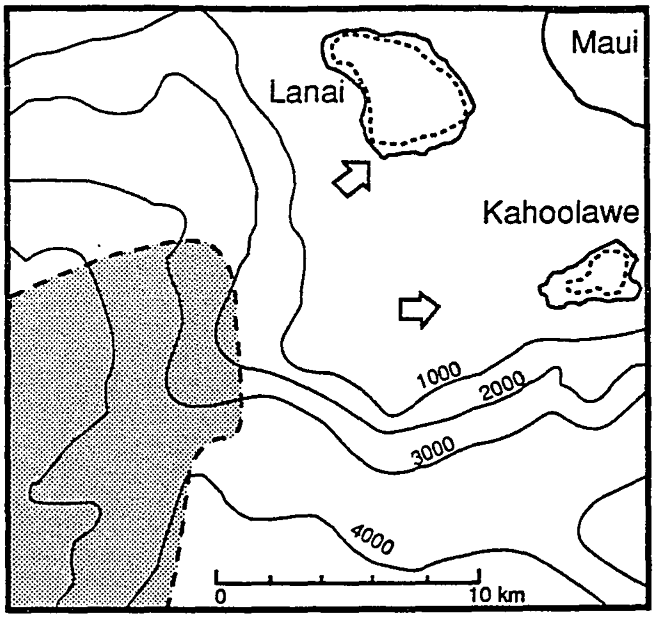


Figure 8. Geographic relationship of submarine slide and areas of soil stripping on Lanai and Kahoolawe. Stippling indicates area of submarine slide. Dashed lines enclose upper areas of unstripped soil (from Moore and Moore, 1988).

studies of landscape change on Kahoolawe (Table 6, Spriggs, 1991) support Stearns' contention that the massive erosion and

Table 6.
Changes to Kahoolawe Landscape (from Spriggs, 1991).

Time period (A.D.)	Human Activity	Environmental Effect
900-1600	Fishing	Minor
1600-c.1778	Agriculture	Dry forest vegetation converted to open savannah
1778-1830	Population decline, Goats introduced	Unknown
1830-1859	Penal colony	
1859-1909	Ranching sheep, goats, cattle	Destruction of vegetation, Massive erosion of topsoil

landscape degradation were due to overgrazing by cattle, sheep and goats (Stearns, 1940). The stripping of soil as evidence for a giant wave or it's direction is not supported since other erosional agents, such as deforestation, wind and rain, could be responsible for removal of soil if an extensive soil had developed on the slope of the islands.

Early suggestions that the marine fossils were carried uphill by natives (Wentworth, 1925a) or blown uphill by wind (Macdonald and Abbott, 1970, p. 208) were refuted by Stearns (1978a). This lead Moore and Moore (1984) to conclude that "all the proposed ancient high-level shorelines ... can be correlated with a single event" (Moore and Moore, 1984, p. 1314) - the 105 ka giant wave.

In order to test the giant wave hypothesis, part of this dissertation was designed to date marine fossils from high-elevations on Molokai and Lanai. If the corals were a result of a single event, then ages should correspond to the 105 ka age assigned to the giant wave event by Moore and Moore (1988).

CHAPTER IV

REVIEW OF CHRONOLOGY OF MARINE FOSSIL DEPOSITS

In this section, the published dates of marine deposits in the Hawaiian Islands are reviewed in order to elucidate patterns in Quaternary sea level and trends in island subsidence/uplift history. Prior to radiometric dating, shoreline ages were estimated by indirect means including stratigraphic position and known field relation of shorelines to volcanics such as the Honolulu Volcanics Group (e.g. Stearns, 1966; Ward, 1973). Several methods for determining the age of marine deposits have been used in the Hawaiian Islands. For each date, the sample location, type of material dated, dating method, and elevation are tabulated in Appendix 1.

The environmental settings of marine fossils in the Hawaiian Islands are varied, and include subaerial, submarine, and subsurface environments. Deposits in the subaerial realm include *in situ* fossil reefs, beach rock, dunes, high-elevation fossiliferous marine conglomerate, marine limestone bombs from phreatic eruptions, and storm and tsunami deposits. Further subaerial marine materials are associated with anthropological sites including temples known as heiaus and native Hawaiian fishing shrines (D. Schideler, pers. comm. 1992). Emerged marine fossil deposits are generally best preserved on the leeward side of each island where rainfall and erosion are minimal, leading to preservation of the marine fossils. A notable exception is windward Oahu, where marine deposits are widespread and well documented (Stearns, 1978a). Subsurface limestone deposits have also been identified particularly in several deep cores on Oahu (Stearns and Chamberlain, 1967; Resig, 1969; Lum and Stearns, 1970).

The dated materials are not evenly distributed between environmental settings nor islands (Figure 9). About half of the dated marine deposits in the Hawaiian Islands are from emerged reefs on Oahu. The second largest set of dated material is from the intensely studied shallow-water submerged reef at Hanauma Bay on Oahu with a third set from deeply submerged reefs off the Island of Hawaii. Eighty percent of the dated material is from Oahu. Little attention has been given to studying the other islands with regard to dating of their marine fossil-bearing deposits.

^{14}C Dating

The geochronology of the late Quaternary (< 40 ka), and particularly the Holocene (<10 ka), has been defined primarily on the basis of ^{14}C dated fossils. This method was applied to date marine fossils from several terraces in the Hawaiian Islands (Rubin and Berthold, 1961). Early results were ambiguous: most samples were older than the upper limit of ^{14}C dating (i.e. Shepard, 1961). At least 85 radiocarbon determinations have been reported from the Hawaiian Islands with many of these representing only a minimum age. Although the upper limit of dating with this method has increased with

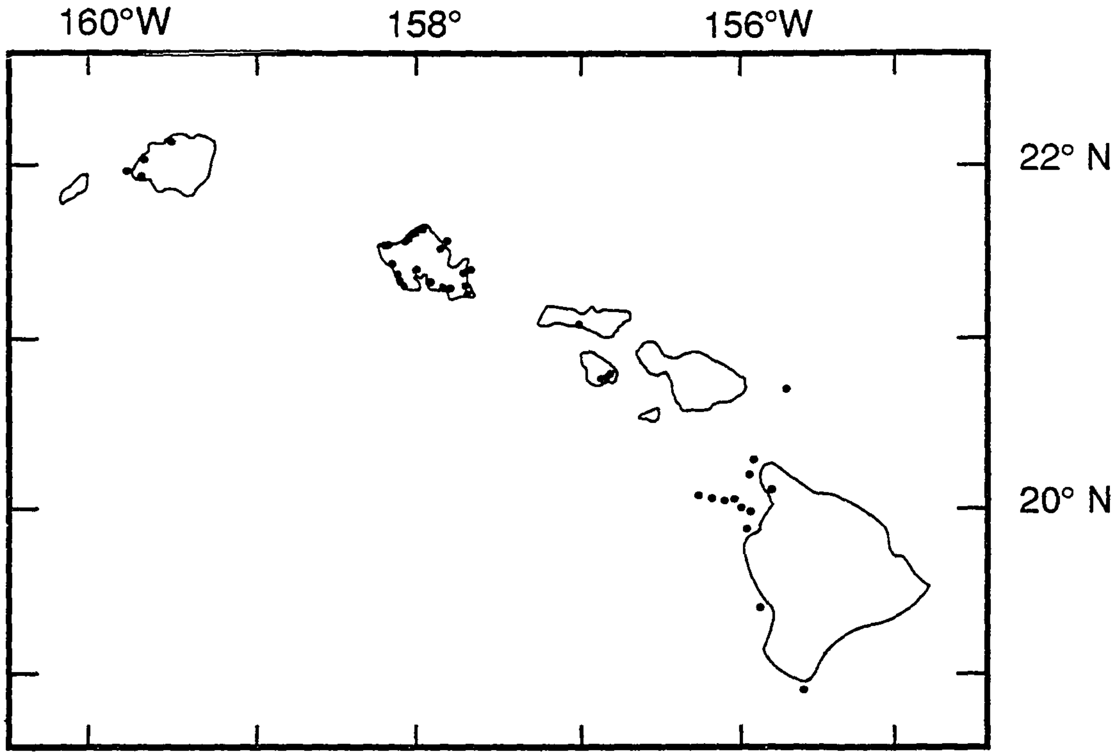


Figure 9. Distribution of dated marine fossils in the Hawaiian Islands.

improvements in technology (i.e., accelerator mass spectrometry can extend the dating limit beyond 50 ka), many Hawaiian samples are considerably older than the limitation of the ^{14}C method.

Radiocarbon dates have been reported from four islands: Oahu (Easton and Olson, 1976; Stearns, 1977), Molokai (Weisler, 1989), Kauai (Hubbs et al., 1965; Inman and Veeh, 1966), and Hawaii (Moore and Fornari, 1984; Moore et al., 1990a). The materials analyzed include coralline algae; mollusk shells, gastropods (Cypraea tigris, Venus reticulata, Conus sp.) and bivalves (oyster shells); various species of coral (Porites, Pocillopora, Cyphastrea sp.); and various geological materials such as limestone and coral sand.

On Oahu, the most intensive dating study was the ^{14}C profiling done at Hanauma Bay. Easton and Olson (1976) drilled ten cores across a coral-algae reef transect and reported more than 60 radiocarbon dates. They concluded that sea level has not been significantly above present level since 7 ka. Their work was widely cited as evidence for no high stand in Hawaii during the Holocene.

On Molokai, Weisler (1989) reported a mid-Holocene specimen of Porites cored from 400 m inland near the Kakahai'a fishpond. Unfortunately, the elevation of this sample, dated at 3.37 $\pm$ 0.07 ka, was described just as "near sea level". Weisler correctly noted that the sample age and position imply an open exchange with the sea at this site. Because sea level had to be above the coral for it to grow, this observation further suggests the possibility of a relative mid-Holocene highstand.

Inman and Veeh (1966) dated coral samples from a -18 m (10-fathom) terrace off northeastern Kauai with ^{14}C (8.37 ± 0.25 ka; Hubbs et al., 1965) and ^{230}Th (8.0 ± 1.0 ka; Veeh, 1966). Finally, nine submerged samples from Hawaii have been radiocarbon dated. The existence of these submerged reefs has been attributed to the combined interaction of rising sea level and rapid subsidence (Moore and Fornari, 1984; Moore and Campbell, 1987; Moore et al., 1990a). Moore and Fornari (1984) sampled the -150 m reef terrace off Kealakekua Bay, Hawaii with the submersible Makali'i. Three ^{14}C ages, averaging 13.12 ± 0.425 ka, were tentatively correlated with a wide shelf offshore of Oahu at -105 m (Ruhe et al., 1965), the so-called Kahipa-Mamala shelf of Stearns (1978a). Later, Moore et al. (1990a) collected corals from a submerged reef at depths of -155 to -305 m off Ka Lae (South Point) with the Pisces V. The average age was 10.8 ± 1.25 ka. Two samples collected from west Hualalai Volcano in 175-200 m water depth had an age of 11.52 ± 0.72 ka. In comparing ^{14}C and $^{230}Th/^{234}U$ dates, the ^{14}C dates tend to be about 3 ka younger (Moore et al., 1990a). The authors accepted the more accurate $^{230}Th/^{234}U$ dates for the age of the terrace. Most of the submersible samples collected by Moore et al. (1990a) were much deeper than the -150 m terrace. However, these samples were interpreted to be debris from the upslope reef.

To summarize the ^{14}C data set from Hawaii, a compilation for the last 10 ka is given in Figure 10. The data show that sea level rose from about -18 m to the present over a span of about 4,000 years. This rise may have been episodic. The -18 m datum is from a constructional terrace and the set of data from about -9 m indicate a second stillstand about 7 ka lasted approximately 1,000 years. But until further work on the nature of the rising sea level is undertaken, the implication of an episodic sea-level rise is speculative. The ^{14}C research demonstrates two attributes of the last deglaciation: first, that sea level has risen from about -18 m at about 8 ka to its present level and second, that mid-Holocene sea level was probably higher than present.

Uranium Series Dating

Corals take up uranium from sea water but exclude thorium during precipitation of their calcareous skeletons. ^{230}Th is a daughter product of the radioactive decay of ^{234}U. The ^{230}Th will accumulate in the coral skeleton over time if the skeletal system remains closed. Corals can therefore be dated by utilizing the disequilibrium in the uranium-series (Broecker et al., 1968). Two methods are presently available to determine the $^{230}Th/^{234}U$ ratio in corals: alpha spectrometry and mass spectrometry.

Alpha Spectrometry

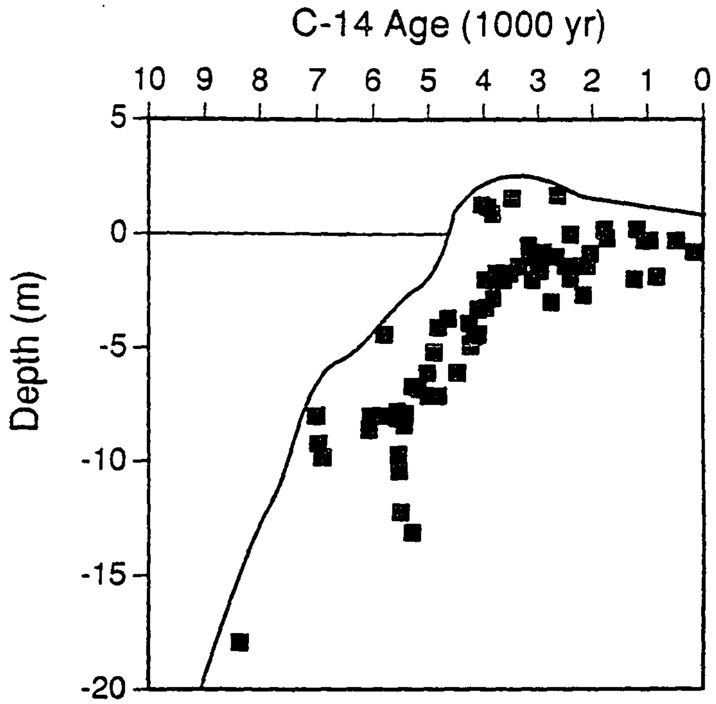


Figure 10. Sea level curve based on ^{14}C dates compiled from the Hawaiian Islands.

Alpha-spectrometry $^{230}Th/^{234}U$ dating of corals has established the age of the prominent Waimanalo stand at 122 ± 7 ka (Veeh, 1966; Ku et al., 1974). On Oahu, Ku and co-workers (1974) dated twenty-three samples of the Waimanalo stand with ages ranging from 112-137 ka. They concluded that the age of the Waimanalo stand is 122 ± 7 ka and represents a single eustatic sea level +7.6 m above present sea level. Based on similar elevations and age at sites elsewhere, Ku et al. (1974) considered Oahu to be stable for at least the last 125 ka, with some workers accepting Oahu as stable for as long as 500,000 yrs (Stearns, 1974a). Stearns (1976) and Chappell and Veeh (1978) noticed that the Oahu samples comprise two groupings: (1) samples of Waimanalo limestone and cemented conglomerate with an average age of 133 ka, and (2) samples from deposits that lie unconformably on the Waimanalo limestone separated from the older group by the Diamond Head Tuff. The later samples had on average an age of 119 ka. Thus, there were two fundamentally different geochronologic and lithostratigraphic interpretations for these data. This problem can now be resolved with the improved precision of mass spectrometric U-series analysis (Edwards et al., 1987).

Alpha-spectrometry has been used at two other locations in the Hawaiian Islands by Szabo and Moore (1986) and Moore and Moore (1988). One site is the submerged terraces off the Island of Hawaii. This series of terraces offers evidence for the timing of the termination of continental glacial events and the associated Pacific and global eustatic rise in sea level and accompanying changes in paleoclimate and paleoceanography during late Pleistocene lowstands. Szabo and Moore (1986) dated the -350 m reef terrace off northwest Hawaii from submersible-collected coralline algae. Szabo and Moore (1986) relate their $^{230}Th/^{234}U$ age of 120 ± 5 ka to the penultimate major glacial period (oxygen isotope stage 6), even though the interval for stage 6 is usually considered to be earlier: from 128 to 186 ka (Imbrie et al., 1984). From this one dated terrace, deeper terraces were correlated to the deep-sea oxygen isotope curve (adapted from Shackleton and Opdyke, 1973). Szabo and Moore predicted the ages of deeper members of the series off northwest Hawaii. Terraces at -700, -950, -1150, and -1300 m were related to the lowstands of the sea represented by oxygen isotope stages 8, 10, 12, and 14, which ended at 250, 350, 440 and 500 ka, respectively.

The other locale where α-spectrometry was used to determine the age of corals was the high-elevation corals in the 'Hulopoe Gravel' of Lanai. Moore and Moore (1988) report only three dates from coral in the 'Hulopoe gravel' above 100 m elevation on Lanai. The samples range in age from 101 to 134 ka. An age of 105 ka is used for the age of the "giant wave" event. Lipman and colleagues (1988) have linked the giant wave with the Alika debris avalanche off west Hawaii, thereby, dating the landslide. It should be noted that the wave event could not have occurred earlier than the date of the youngest coral (101 $\pm$ 4 ka).

Unfortunately, α-spectrometry allows dating of corals only from the last interglacial with relative confidence. Dating of previous interglacials, although theoretically possible, is generally not reported because of poor yields or very large uncertainties in the counting statistics.

Mass Spectrometry

The development of thermal ionization mass spectrometry (TIMS) to measure Th and U isotopes requires a significantly smaller sample and yields greater precision than α-spectrometry (Edwards et al., 1987). TIMS also allows coral samples older than 250 ka to be dated by using $^{234}U/^{238}U$ analysis. TIMS has been used in Hawaii to develop a chronology for the drowning of the terraces off Hawaii (Ludwig et al., 1991) and off Maui (Moore et al., 1990b).

Coral recovered in dredge hauls from the uppermost of a series of four subparallel arcuate belts on the Haleakala Ridge off southeast Maui was dated by a single mass spectrometric determination of $^{234}U/^{238}U$ (Moore et al., 1990b). These reefs appear to have been tilted about 20 m/km southward as a result of post-depositional volcanic loading of Hawaii. Further field and chronological study of the tilted reefs are needed to establish their horizontal extent and chronology.

A series of six deeply submerged reef terraces off northwest Hawaii were dated by mass spectrometric $^{234}U/^{238}U$

analysis (Ludwig et al., 1991). The results indicate (1) a relatively uniform rate of subsidence for the island of Hawaii at 2.6 mm yr^{-1} over the past 475,000 years and (2) drowning of the reefs coincided with intervals of rapid sea level rise - usually at the end of glacial periods (Figure 11). At the termination of the last glacial maximum, 21 ka, sea level was about -107 $\pm$ 5 m as estimated from the shallowest drowned terrace when corrected for subsidence of Hawaii. This lowstand can be verified by sampling and dating the -105 m terrace off Oahu (Ruhe et al., 1965). It is interesting to note that no reef in the series of deeply submerged terraces is associated with glacial stage 4, a relatively short interval of approximately 11,000 years. This may be due to lack of sufficient time to develop an extensive reef system on a subsiding platform and has bearing on the present state of reef development in the Hawaiian Islands because sea level has been at or near its present level for only 8,000 years. As pointed out by Ludwig and others (1991), the ages from the four deeper terraces were more difficult to correlate to glacial terminations because of scatter in the data set. Given this variance, the likely range of depths for the past five major glacial periods is between -70 to -150 m (Figure 11B). This problem can be resolved with better sampling that carefully selects in situ reef crest samples and with more replication. It should be noted that a linear extrapolation for the rate of subsidence assumes that the hot spot has been

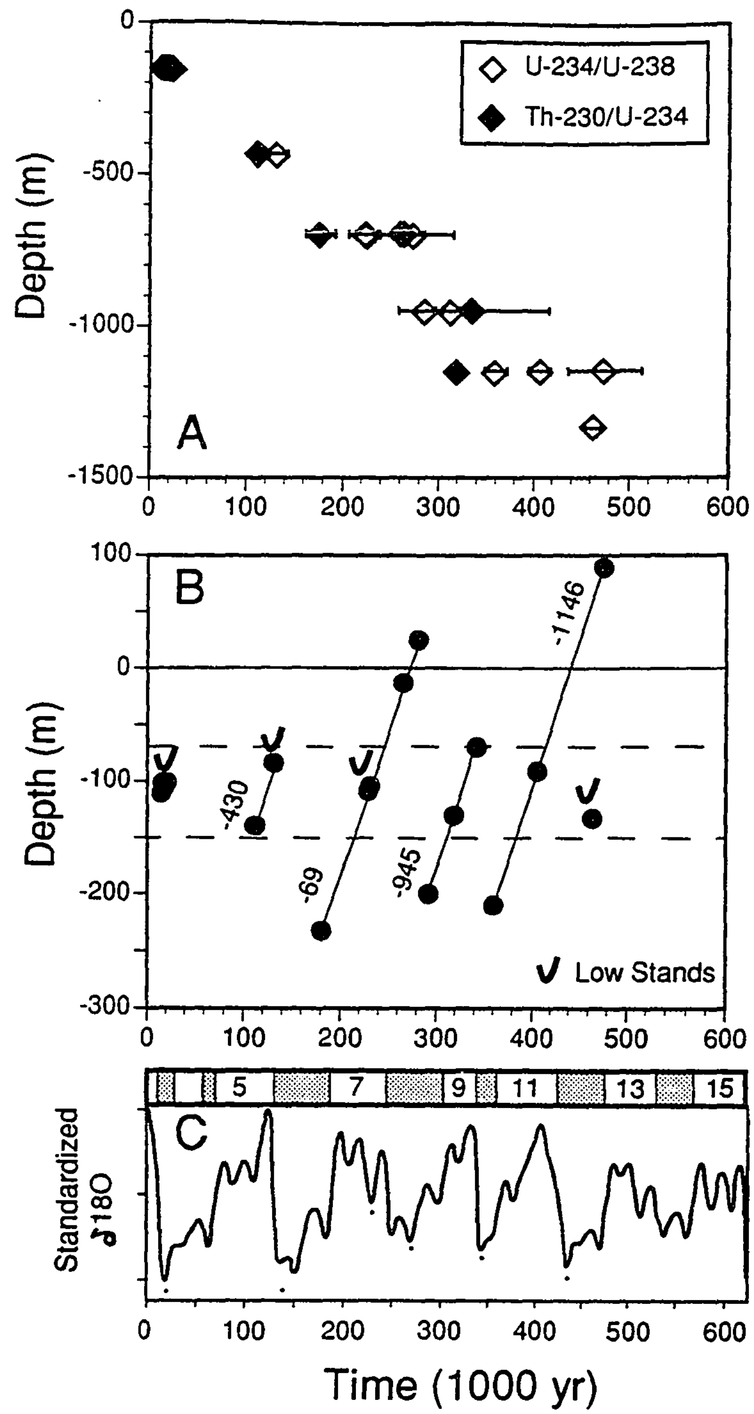


Figure 11. Age-depth plots for (A) the six terraces off northwest Hawaii (Ludwig et al., 1991) and (B) subsidence corrected depths of lowstands with (C) display of oxygen isotope curve showing the history of sea level over the last 600,000 years and stage boundaries (after Imbrie et al., 1984).

behaving in a regular manner for the last 500 ka, in terms of volume of lava and loading of the plate. It also assumes sea level fall was on the same order for each glacial period. Further efforts at establishing the chronology of the deeper terraces are needed.

$^{231}Pa/^{235}U$ Dating

A method based on using protactinium ($^{231}Pa/^{235}U$) to date corals was developed by Ku (1968) and has been applied as an independent verification of the α-$^{230}Th/^{234}U$ dates. Fifteen dates derived from this method have been reported from the Waimanalo Formation on Oahu (Ku et al., 1974). The $^{231}Pa/^{235}U$ ages generally support the $^{230}Th/^{234}U$ ages of the last interglacial period at 122 $\pm$ 7 ka on Oahu (Ku et al., 1974; Figure 12).

To summarize the findings from the U-series dating, three principal outcomes of the research are (1) the establishment of the age of the last interglacial period on Oahu, (2) the establishment of a chronology for the series of drowned reefs off Hawaii, and (3) the dating of the alleged "giant wave" event on Lanai.

ESR Dating

A relatively new technique, ESR dating of corals, was introduced in the early 1980's (Ikeya and Ohmura, 1983). Only one study has used this technique to date marine fossils from the Hawaiian Islands. Brückner and Radtke (1989) dated twenty-one samples using ESR from emerged marine deposits on

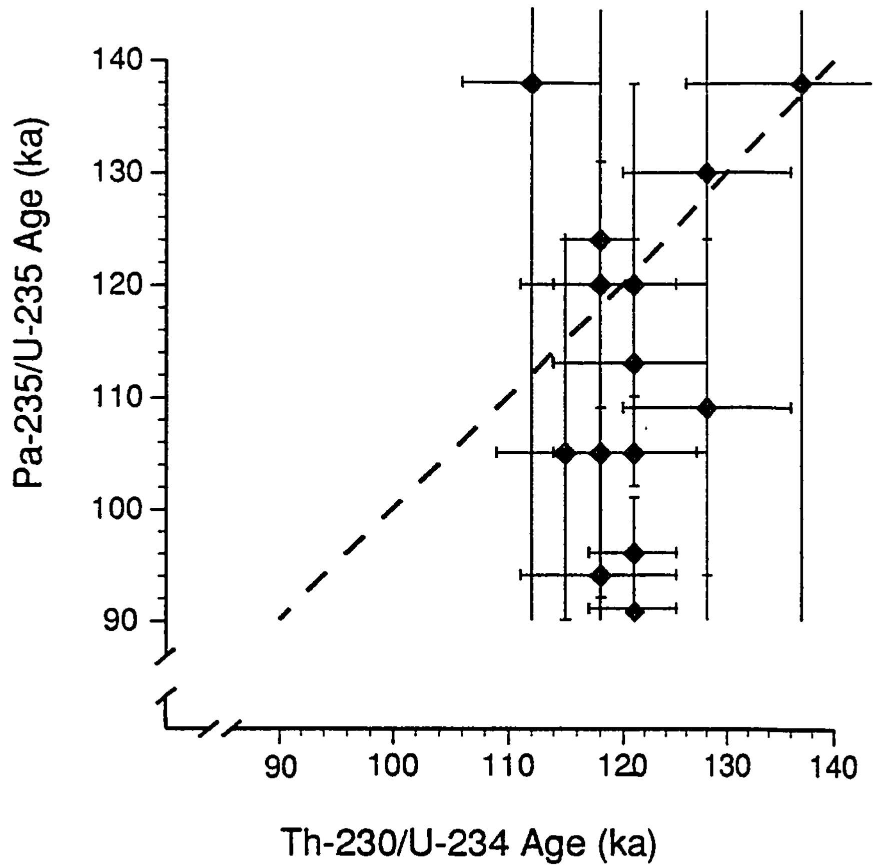


Figure 12. Plot of the $^{231}Pa/^{235}U$ dates versus $^{230}Th/^{234}U$ dates for samples from Waimanalo limestone (after Ku et al., 1974).

Oahu. The authors concluded that there were at least four fossil reef tracts or beach deposits. The age of the highest deposit at +30 m located at Kaena Point was found to be 450-500 ka. A set of sites from +17 to +25 m could not be dated because of alteration of the original aragonitic skeleton of the coral material. An elevated beach near Kahe Point at +15 m was dated at 250 ka. The last interglacial represented by the Waimanalo stand at +7.6 m was dated at 125 ka. A younger tract between -0.5 to -1 m was identified only as Holocene in age.

Brückner and Radtke (1989) performed nine α-$^{230}Th/^{234}U$ age analysis. Of these, three were from the same samples with ESR dates. Two of the ESR dates fell within the standard deviation of the α-$^{230}Th/^{234}U$ age, one was outside the range at $\pm$ 1σ (Table 7).

Table 7.
Comparison of ESR and $^{230}Th/^{234}U$ ages from Oahu (after Brückner and Radkte, 1989).

Sample No.	Th/U (ka) ± Std. Dev.	ESR† (ka) ± Std. Dev.
Nana-1	$137 ^{+20}_{-17}$	116 ± 17
NH-2	$230 ^{+20}_{-20}$	256 ± 38
Kahe-11c1	$210 ^{+10}_{-5}$	223 ± 33

† estimated error 15% (see Radtke and Grün, 1988).

Other Methods

Potassium-argon (K-Ar) dating of an intrusive basaltic dike that penetrated the Kaena Limestone placed a minimum age of 410 ± 40 ka on the Kaena marine deposit (Stearns and Dalrymple, 1978). This younger limit was in agreement with independent determinations, including $^{230}Th/^{234}U$ date of >200 ka, $^{234}U/^{238}U$ date of 600 ± 100 ka (Veeh in Stearns, 1973), and ESR dates of 482 ± 54 ka (Brückner and Radtke, 1989).

A method of dating marine notches by dating tuffs that have not been eroded was suggested by Stearns (1974a), but has not been pursued.

Discussion

Quaternary Sea Level

From this review of the chronology of the marine deposits in the Hawaiian Islands, a few general conclusions can be drawn. The last interglacial period is well represented on Oahu by the Waimanalo stand of Stearns (1978a). The age of this limestone deposit was established at 122 ± 7 ka by Ku and others (1974) and later workers (Easton and Ku, 1981; Brückner and Radtke, 1989). The widely quoted level for the Waimanalo shore at +7.6 m is the average of two wave-cut notches at +8.2 m and +6.7 m. Coral material, however, found at an elevation of +11-12 m has been dated at 125 ± 3 ka from Mokapu Point (Muhs and Szabo, 1991) and from Kahe Point (Ku et al., 1974). These data cast doubt that Oahu has been stable, given the evidence for a 5-6 m highstand during the last interglacial elsewhere in the world.

The Holocene transgression of the sea has been reasonably documented by the compiled ^{14}C data set for the Hawaiian Islands (Figure 9). However, the maximum sea level fall during the last glacial has not been well established in Hawaii. The expected depth for the last glacial maximum is around -110 m to -130 m (Kennett, 1982).

The age data from deposits older than 150 ka are limited. Samples from sites on Oahu that have been dated older than 150 ka are all above 15 m.

Trends in Uplift/Subsidence

Given the restricted data set, the trend in uplift or subsidence can only be accurately quantified for two of the islands, Oahu and Hawaii. The elevation of dated emerged deposits on Oahu suggests that the island has experienced 30 m of uplift during the late Quaternary (Figure 13). This conclusion is primarily based on comparison of sea level records from other areas. For example, the generally-accepted stable platform of Bermuda has been extensively studied in terms of sea level records during the Late Pleistocene (<500 ka). During the last four interglacials, sea level in Bermuda has been within 5 m of the present level (Table 8).

As for the island of Hawaii, the drowning of reefs during the termination of continental glaciation and the preservation of these reefs as a flight of terraces offshore northwest

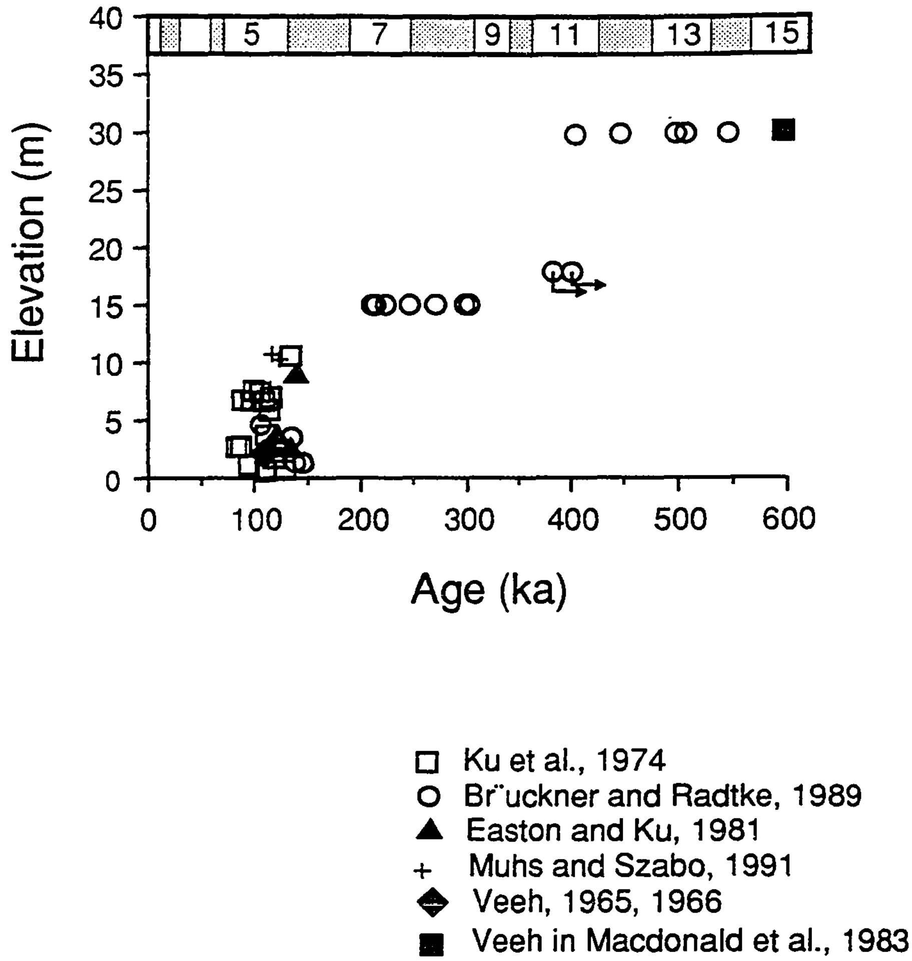


Figure 13. Compilation of dated marine fossils from Oahu.

Table 8.
Composite stratigraphy of Bermuda with maximum elevation of highstands, uranium-series, ESR and amino acid racemization (AAR) ages and correlations with isotopic stages.

Fm.[1]	Max. Elev. (m)	U-age[2]	AAR[3]	ESR[4]	Stage[5]
Southampton	≤+1[A]	85±6		102	5a
Rocky Bay	5±1[B]	120 125±4			5e
Belmont	+2.5[A]	204±11	190±20 265±30	188	7
Upper Town Hill	+5±3[C]		350±40 325±20	373	9
Lower Town Hill	+5±3[C]		475±35 430±15	490	11
"Gov. Quarry"	+22[A]		>700		
Walsingham			>880	709	

[1] From Vacher et al. (1989) with unmapped unit provisionally identified as "Gov Quarry" for sample from Government Quarry referred to as an unnamed formation in Hearty et al., (1992).
[2] Harmon et al. (1983)
[3] Hearty et al. (1992)
[4] Hearty and Mitterer (1990)
[5] Oxygen isotope stages

Elevational data from the following sources:
[A] Hearty et al., (1992)
[B] Harmon et al. (1983)
[C] Vacher pers. comm. (1992)

Hawaii is convincing evidence for subsidence of this island of approximately 1300 m over the last 475 ka (Figure 11).

The questions raised by review of the existing chronological data set are the rational for investigating the following unanswered problems:

- Was sea level at its present height during the mid-Holocene?
- Has the island of Oahu remained stable during the Late Pleistocene?
- Are the described marine deposits on Molokai and Lanai related to the alleged 105 ka giant Lanai wave?
- For the island of Lanai, are there any indications of emerged paleoshores (uplifted)?

CHAPTER V

ELECTRON SPIN RESONANCE (ESR) DATING OF CORAL

Introduction

The basic principles of electron spin resonance (ESR) spectroscopy were developed over fifty years ago (Gorter, 1936a, 1936b; Gorter and Kronig, 1936). In the late 1960's, Zeller and coworkers (Zeller et al., 1967; Zeller, 1968) suggested ESR as a means of dating, similar to thermoluminescence (TL) dating. Ikeya (1975) successfully dated the first carbonate samples, by measuring a radiation sensitive ESR signal in cave-precipitated calcite. Later, Ikeya and coworker applied this technique to date Pleistocene corals from uplifted reef tracts in the Ryukyu Islands (Ikeya and Ohmura, 1983). ESR has "developed considerably in the last decade as a dating technique for material in the geosciences and paleoanthropology" (Ikeya, 1988, p. 92).

In addition to its application to corals, ESR dating has been successfully developed for a variety of materials, including calcite in cave deposits (e.g., Ikeya, 1975; Hennig et al., 1981), mollusks (e.g., Ikeya and Ohmura, 1981; Radtke et al., 1981; Linke et al, 1985; Katzenberger and Grün, 1985; Radtke, 1985; Radtke et al., 1985), foraminifera (e.g., Sato, 1981, 1982; Siegele and Mangini, 1985), spring-deposited travertines (e.g., Hennig and Hours, 1982; Radtke et al., 1986; Schwarcz et al., 1988), hydroxyapatites in tooth enamel of mammals (Grün et al., 1987), quartz from faults (e.g., Ikeya et al., 1982, 1983; Buhay et al., 1988), and various minerals in sediments and volcanic ash (Imai and Shimokawa, 1985; Shimokawa and Imai, 1987; Toyoda and Ikeya, 1991). Current efforts are continuing to develop potential ESR dating for a broad spectrum of materials, such as bone (Sales et al., 1985; Oduwole and Sales, 1991; compare Grün and Schwarcz, 1987), animal skins (Miki, 1985), blood (Miki et al., 1988), fish scales (Blackwell et al., 1992), quartz sand from archaeological hearths (Blackwell et al., 1992) and barite from deep-sea hydrothermal chimneys (Jones, unpubl. data). ESR dating has been frequently applied in paleoanthropology for dating mammal teeth (Schwarcz and Zymela, 1985; Grün et al., 1987; Wieser et al., 1988; Zymela et al., 1988), burnt flint (Porat and Schwarcz, 1991), and speleothems (Ikeya and Poulianos, 1979; Hennig et al., 1981; 1982; Poulianous 1982; Xirotiris et al., 1982; Grün, 1985) at important cave sites inhabited by early hominoids.

In studying fossil coral deposits from the Hawaiian Islands, the ESR dating method was selected to estimate ages for several emerged deposits. ESR was chosen for several reasons:

(1) The estimated ages of the deposits (5-500 ka) were within the applicable range for this technique ($\leq$1 Ma), so one technique could span the estimated range for the sample dates.

(2) Relatively small amounts (<1 g) of unrecrystallized coral are required for ESR dating.

(3) The procedure is nondestructive, enabling other chemical analyses to be performed on the same samples.

(4) The technique has been successfully applied to marine deposits on Oahu (Brückner and Radtke, 1989) and elsewhere (Ikeya and Ohmura, 1983; Koba et al., 1985; Skinner, 1985b, 1988; Ikeya, 1988; Radtke et al., 1988; Ikeda et al., 1991; Pirazzoli et al., 1991; Gray et al., 1992).

This chapter focuses on the ESR dating method as applied to corals from the Hawaiian Islands. Experimental studies were carried out to determine aliquot size, sample pretreatment, post-irradiation treatment and examine spectrometric protocol.

Principles of the Method

The principles and practice of dating geological materials using ESR spectroscopy are now well established (Ikeya, 1978; Hennig and Grün, 1983; Ikeya, 1988; Grün, 1989a, 1989b). The dating method relies on the fact that unpaired electrons accumulate at crystal lattice defects as a result of natural irradiation in solids. The density of unpaired electrons trapped in such defects is measured directly and chemically non-destructively with the ESR spectrometer, therefore, measuring the cummulative effect of irradiation.

The ESR signal intensity increases through time as a function of the average radiation flux or dose rate. An ESR age is determined from the relationship:

$$AD = \int_0^T \dot{D}(t)\, dt \qquad (5.1)$$

where the accumulated dose, abbreviated AD, is the total dose acquired by the sample since its formation and the annual dose rate, $\dot{D}$, is generated from radioactive elements in the sample and surrounding sediment plus cosmic rays.

Experimentally, the total or accumulated dose (AD) is extrapolated by artificial additive irradiation. The ESR age is obtained from the AD by assessing the average annual dose rate.

In corals, the dominant radioactive components are uranium and its daughter products. Corals incorporate uranium from seawater as they build their calcium carbonate skeleton. Upon death, the skeleton acts as a passive dosimeter recording the cummulative natural environmental radiation at the sample site. Allowances for cosmic radiation effects are incorporated in computing the annual dose rate.

Dating Limits

The upper limit for ESR coral dating is approximately 1 Ma (Ikeya, 1984) and samples older than 800 ka have been reported (Radtke et al., 1988; Ikeda et al., 1991). Improved ESR measurement techniques have extended the applicable range of this technique beyond those of ^{14}C (<70 ka) and $^{230}Th/^{234}U$ (<350 ka) to bridge an important gap in Quaternary chronological studies between 500 ka and 1 Ma (Skinner, 1985a). For corals, the measurable minimum ESR age, approximately 500 years, depends on the U content in the coral and the radiation dose received by the coral (Ikeya and Ohmura, 1983) as well as instrument limitation. At ambient temperatures, the stored charges are stable on timescales up to 10^7 years (Grün, 1989a). In practice, the dating of corals less than 500 ka is readily determined.

Comparison with Other Dating Techniques

A comparison between ESR and other dating techniques is found in nearly all papers in which ESR is used to date coral. Generally, good agreement has been reported for ESR dates and those obtained by ^{14}C dating of individual coral heads (Ikeya, 1983; Ikeya and Ohmura, 1983; Radtke and Grün, 1988; Peng et al., 1989), $^{230}Th/^{234}U$ dating (α-spectrometry) (Ikeya, 1983a; Ikeya and Ohmura, 1983; Skinner, 1985b, 1988; Radtke and Grün, 1988; Radtke et al., 1988; Peng et al., 1989; Brückner and Radtke, 1989; Pirazzoli et al., 1991; Gray et al., 1992) and He/U dating (Radtke and Grün, 1988; Skinner, 1988; Figure 14). Holocene corals were compared with ESR and ^{14}C in a few studies. The correlation between the methods is remarkable considering that investigators used different functions (linear, exponential) to extrapolate an accumulated dose (AD) (Table 9). However, in studies of Pleistocene reefs, there is

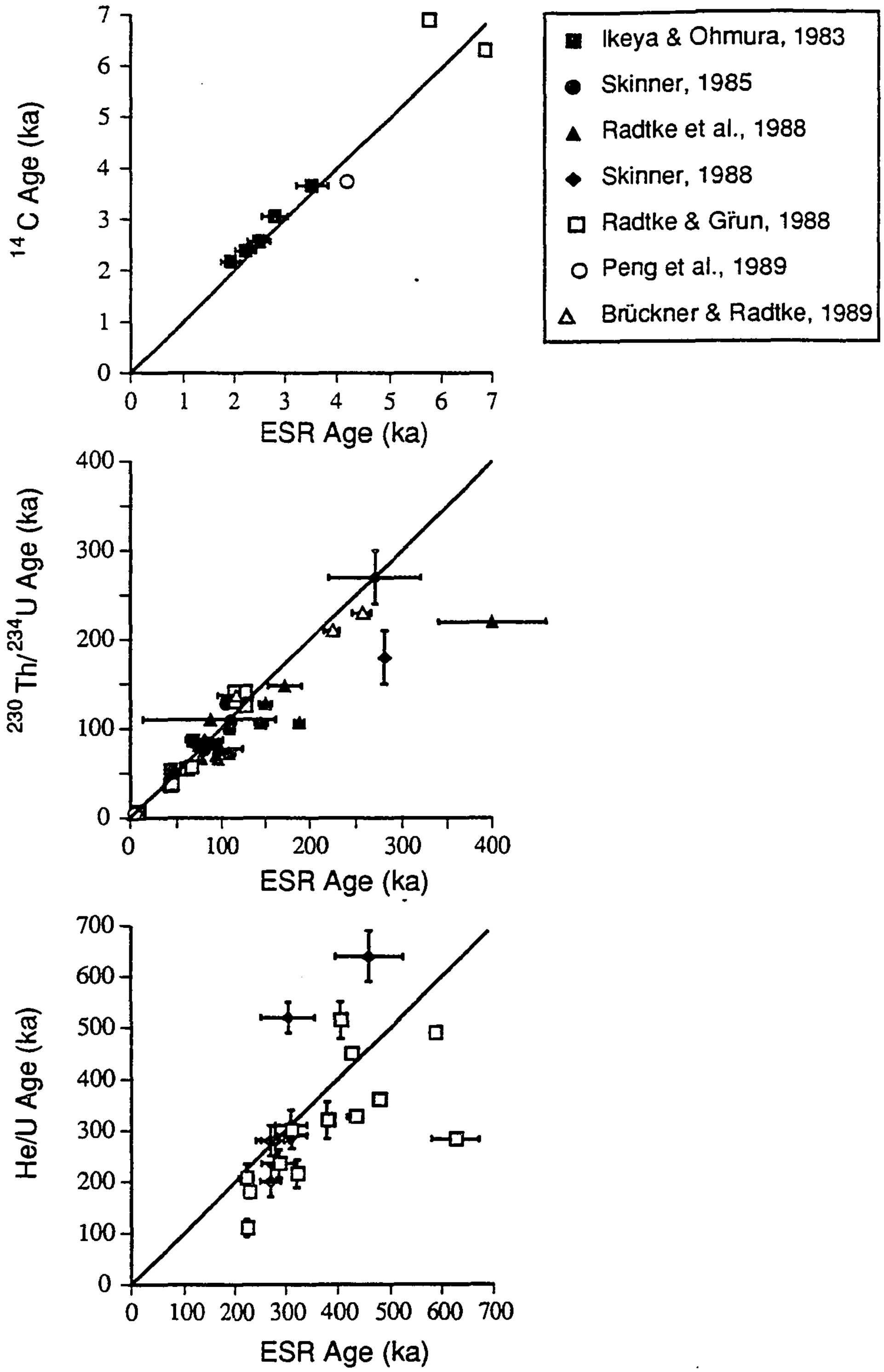


Figure 14. Comparison of ESR dates with other techniques. The 1:1 line is plotted for reference.

Table 9.
Correlation of ESR coral data with other independent methods.

Method	n	r^2
^{14}C	8	0.921
$^{230}Th/^{234}U$	46	0.743
He/U	20	0.320
He/U [less outlier]	19	0.614

a tendency for the ESR dates to be older than the $^{230}Th/^{234}U$ dates. Nevertheless, the agreement is very good for samples less than 200 ka. It should be kept in mind that the upper limit of $^{230}Th/^{234}U$ is generally around 250 ka. In addition, in studies comparing ESR with He/U, most ESR dates are older than the He/U dates. This result supports previous suspicions of the sensitivity of He/U dating to escape of He from the sample - thus violating the assumption of a closed system (Bender et al., 1979). Although the number of samples older than 400 ka is very limited, removal of the outlying points, substantially improves the correlation coefficient between the two methods. Both studies comparing ESR with He/U dates use published He/U results of Bender et al. (1979) from the uplifted reef tracts on Barbados. Radtke and Grün (1990) concluded that the "Barbados" model of sea level has larger uncertainties than previously reported caused by variation in local uplift rate and geochronological problems. It should be emphasized that the He/U method is an experimental dating technique with serious problems involving the migration of He and that the method has not been used for geochronological studies.

Besides corals from Oahu (Brückner and Radtke, 1989), ESR has been used to date corals in Japan (Ikeya, 1983; Ikeya and Ohmura, 1983; Koba et al., 1985, 1987), the South China Sea (Peng et al., 1989), Haiti (Skinner, 1985b), Barbados (Skinner, 1988; Radtke et al., 1988), Sumba Island, Indonesia (Pirazzoli et al., 1991), and the Cook Islands (Gray et al., 1992). Within the last decade, ESR coral dating has laid a foundation for extending the dating technique past the 300-350 ka limit commonly set by α-spectrometry $^{230}Th/^{234}U$ dating. Thermal ionization mass spectrometry (TIMS) $^{230}Th/^{234}U$ dating of coral yields maximum dates of about 500 ka (Edwards et al., 1986). Although recent developments in TIMS allow for high precision in dating corals, TIMS is expensive, destructive, and difficult to master, whereas ESR is nondestructive, easier to learn and faster.

<u>Instrumentation</u>

The ESR spectrometer detects paramagnetic centers generated by unpaired electrons held in electron traps or "holes". In an ESR spectrometer, the microwave frequency is held constant, while the magnetic field is varied linearly resulting in an ESR spectrum. In order to increase the signal-to-noise ratio, the ESR absorption line is transformed into its first derivative then plotted on an X-Y recorder (ESR intensity I) against the magnetic field. The ESR spectra reported in this work were recorded at room temperature with a Brücker ER 100D spectrometer with a microwave bridge ER-040-X. Instrumental settings used for all measurements are listed in Table 10.

Table 10.
Instrumental settings for routine measurement of coral ESR spectra.

Settings:	
Microwave Frequency	9.76 GHz
Mid-range	3480 G
Scan Range	50 G
Microwave Power	20 Db (2 mW)
Field Modulation	0.8 G_{pp}

The spectra were plotted on paper and the intensity measured directly. Selected spectra were recorded on a computer using an EPR Data Acquisition System (EPRware version 2.2.2; Morse, 1987) that can record, manipulate, and plot ESR spectra. A comparison of results obtained using EPRware with those obtained with the standard chart recorder suggests that no difference occurs in the spectra. Whereas greater signal processing can be undertaken when spectra have been stored digitally, recording the region of interest on a paper chart proved more efficient, because it reduced the time needed to record the full spectrum from 2.5 minutes to less than 30 seconds.

Coral ESR Spectra

In modern corals, the ESR signal intensity is zero. Upon irradiation, three signals are generated in coral at g-values of 2.0057, 2.0031, and 2.0007 (Figure 15). The signal at g-value 2.0007 is used in all dating studies. The g-value or Lande factor is a spectrometric splitting factor with free electrons having a g-value near 2.000. For further discussion of the physical basis of ESR spectrometry see Grün (1989a, appendice B and C). Other carbonate materials, including mollusks, foraminifera, and calcite speleothems, also exhibit similar signals at g-value 2.0007 (Grün, 1989a, 1989b). The identification of the paramagnetic centers specified in coral dating studies was originally suspected by Ikeya and Ohmura (1983) to be associated with the CO_3^{-3} radical. Other investigators, however, are not convinced as to its identification (Yokoyama et al., 1988, Barabas et al., 1989, Debuyst et al., 1991). For geochronological applications, however, the characteristics of the selected signal is more important than accurate identification of the center. In any event, the dating signal is not affected by exposure to light or by the procedures used in preparing the sample, nor does the signal fade after irradiation (Grün, 1989a, 1989b). No extraneous interference in the range of g-values from 2.020 to 1.9976 was reported during signal evaluation by the "plateau test" (Miki and Ikeya, 1985).

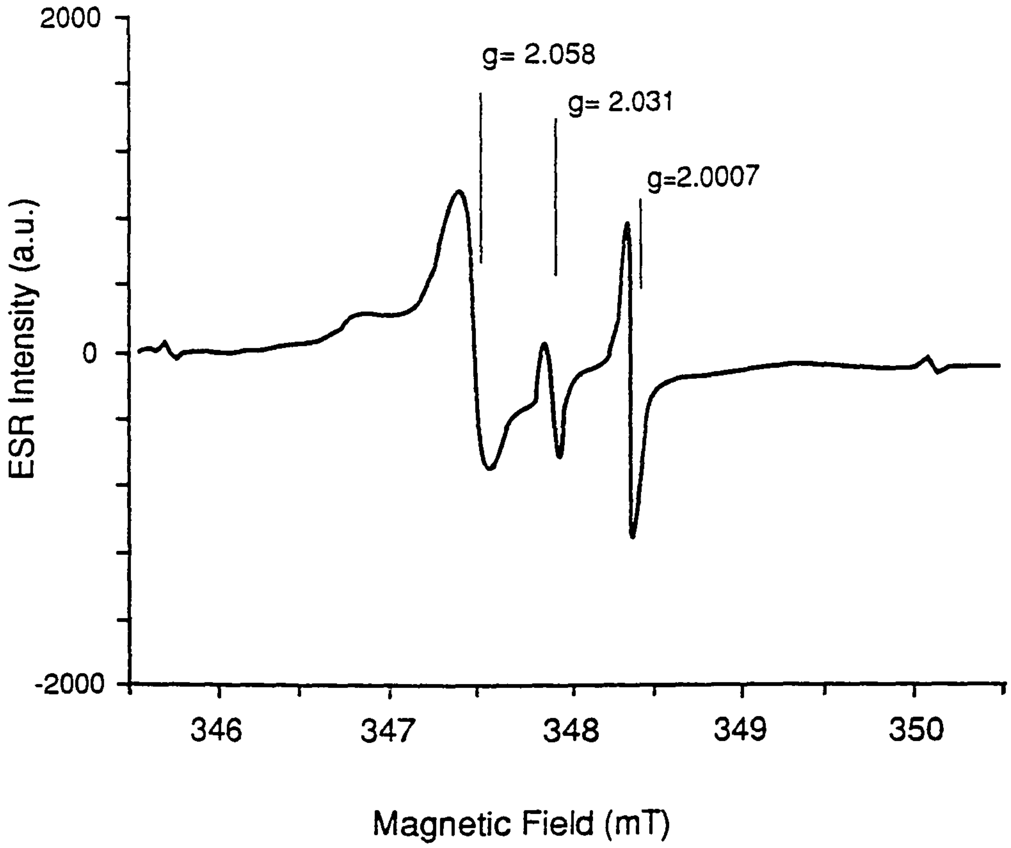


Figure 15. Coral ESR spectra showing the dating signal at g-value 2.0007.

Sample Preparation

Protocols for preparing coral samples for ESR analysis are similar to those reported by Grün (1989a, 1989b), with the exception that significantly smaller aliquot sizes were used in this study (see below; Figure 16). Selected corals were cut with a rock saw into slabs, which were visually examined for alteration. Regions with obvious recrystallization textures were avoided. Outer surfaces were excluded to eliminate externally material that has been exposed to α radiation (Aitken, 1985). Samples were crushed in a mortar using a gentle hand-powdering procedure. The fraction less than 425 μm was dry sieved and the larger fraction returned for further crushing. Based on experiments with Hawaiian coral, the accumulated dose (AD) is independent of grain size. The size fraction, 425-250 μm, is preferred simply because it is easier to handle and to transfer into quartz glass tubes for ESR spectrometry. The fraction between 425 and 250 μm was used for ESR analysis while the fraction between 250 and 125 μm was used for neutron activation analysis (NAA). The <125 μm fraction was analyzed by X-ray diffraction (XRD) to determine the aragonite:calcite ratio (Lowenstam, 1954).

Coral samples were divided into 13 to 15 aliquots for γ irradiation (^{60}Co) at the McMaster Nuclear Reactor. After irradiation, samples were annealed for 3 days at 90°C to remove unstable signal interference generated by the artifical γ irradation. Approximately 1 g of each coral sample was

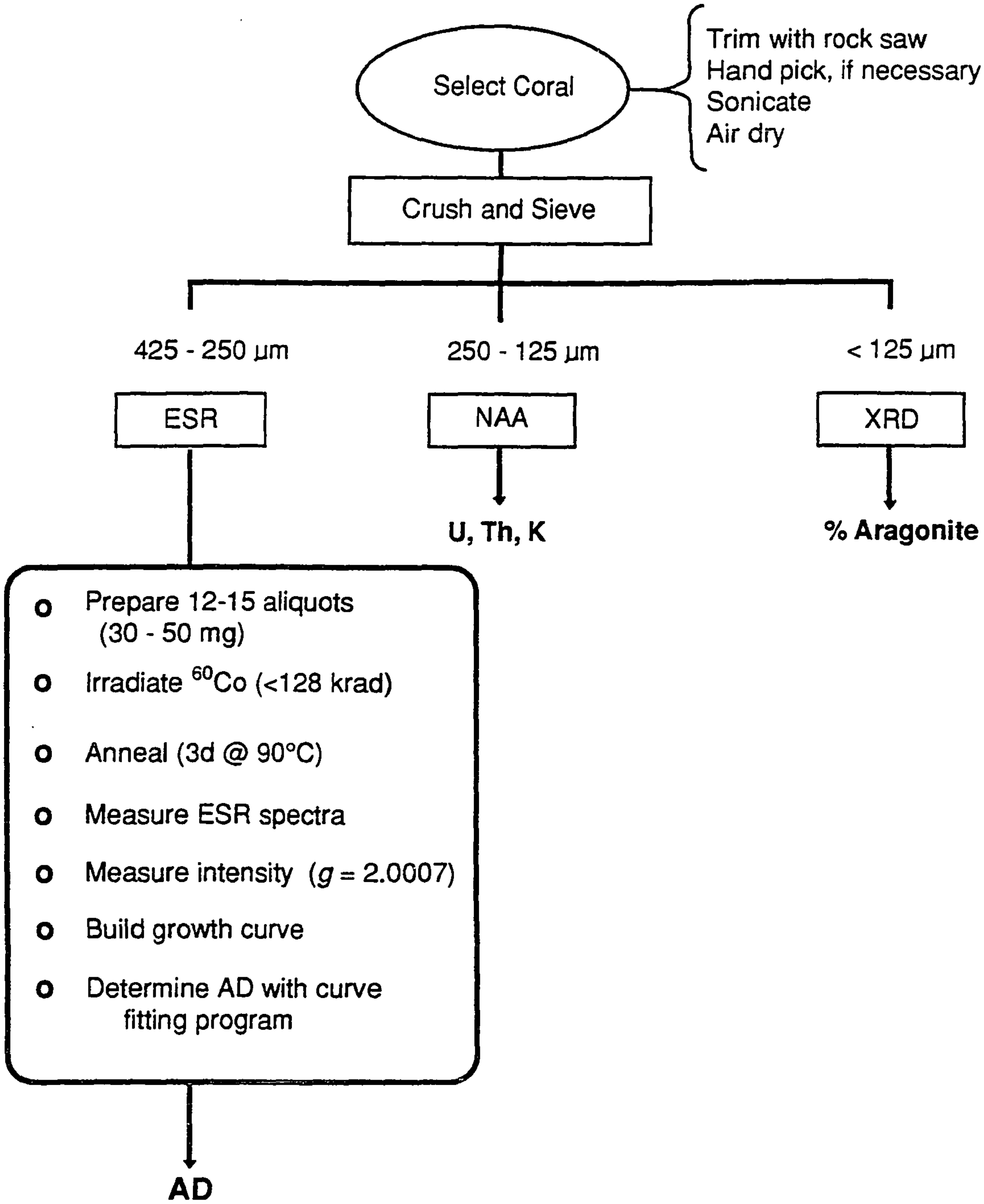


Figure 16. Scheme for preparation of coral samples for ESR dating.

analyzed using delayed neutron counting NAA for ^{235}U, and by instrumental NAA for ^{238}U, ^{232}Th, ^{42}K, plus several other elements.

Aliquot Weight

Aliquots of 30 to 50 mg were used for ESR dating. This decision is based on my study that compared the measured ESR dating signal intensity (g-value 2.0007) of fossil coral material with increasing aliquot size (Figure 17). The ESR intensity is proportional to aliquot weight in the range of 20 to 100 mg. Using a 2 mm ID quartz glass tube, the spectrometer cavity becomes filled by a 110 mg aliquot, so that the signal intensity saturates above this weight. To enhance reproduciblity, the quartz tube is marked so that its height and orientation are precisely aligned for each packed sample. Previous studies reported aliquot sizes ranging from 200 to 400 mg (Table 11).

Pre-treatment: Etching

Coral pre-treatment by etching with dilute acid had been reported (Skinner, 1985a, 1988; Peng et al., 1989; Ikeda et al., 1991), but was not evaluated prior to this study. To test this procedure, the ESR signal was measured in 30 mg aliquots of corals etched in weak acetic acid (0.001N) for 10 to 60 minutes and overnight. The ESR intensity at g-value 2.0007 did not differ with etching, nor did the etching improve the signal-to-noise ratio. The computed accumulated doses (AD) for the etched and unetched coral were within the

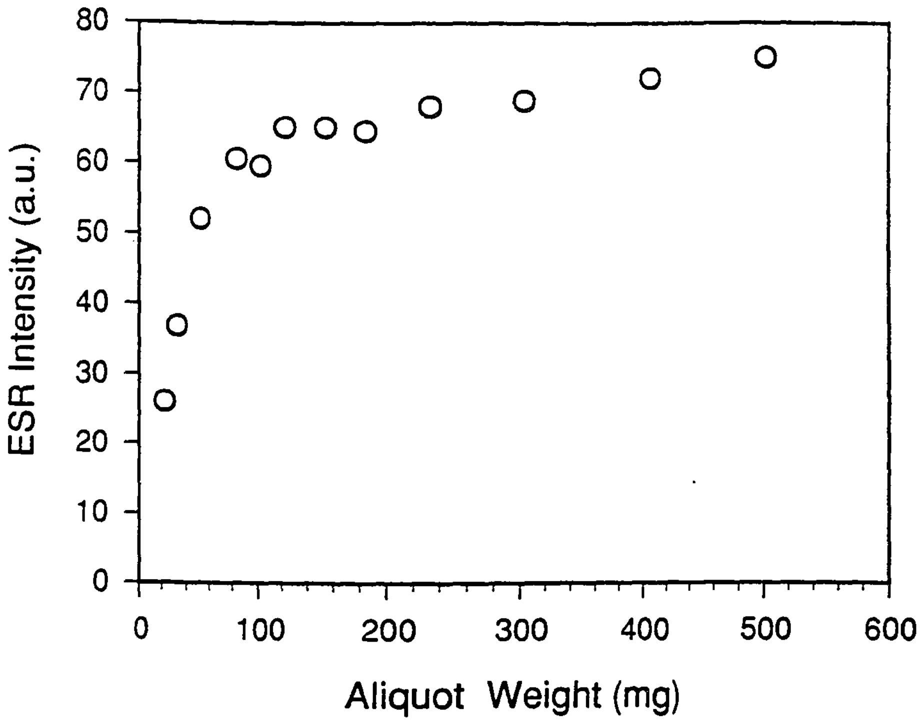


Figure 17. ESR dating signal response to aliquot weight.

Table 11.
Comparison of laboratory procedures for ESR coral dating.

Aliquot		Grain	Treatment	Reference
No.	Size (mg)	Size (μm)	Etching	
2	200	100-1000	--	Ikeya and Ohmura, 1983
7	400	250-500	--	Koba et al., 1985
--	--	100-400	dilute acid	Skinner, 1985b
9	--	--	--	Radtke et al., 1988
6-7	--	100-325	1% Acetic	Skinner, 1988
7-9	--	100-200	--	Radtke and Grün, 1988
--	300	125-180	5% HCl	Peng et al., 1989
12	100-200	100-200	none	Barbabas (written comm., 1990)
14	--	250-500	HCl	Ikeda et al., 1991
13-15	30-50	250-425	none	This study

range of error as determined by a statistical procedure which removes each point and recalculates the curve (Grün and Macdonald, 1989; Figure 18). Acid etching as a means for "eliminating surface effects" (Skinner, 1985b) is therefore considered unnecessary in ESR coral dating.

Annealing Experiments

Two post-irradiation treatments for corals have been recently suggested in an interlaboratory comparison, storage at room temperature for 4 weeks or at 60-90°C for 3-5 days (M. Barbabas, written comm. "Intercomparison Project on ESR Dating", 1990). I compared samples stored at room temperature (25°C) for 14 days with those stored at 90°C for 3 days. There were no significant differences in computed AD. Variation was greater within replicated treatments than between treatments at 25°C and 90°C. At temperatures above 200°C, the dating signal at *g*-value 2.0007 is altered (Jones et al., 1992). Annealing at 90°C for 3 days was adopted for the post-irradiation treatment.

Determination of Accumulated Dose (AD)

The accumulated dose is generated by all radioactive sources that penetrate the coral and is evaluated empirically by the additive dose method (Grün et al., 1987). Twelve to fifteen aliquots are artificially irradiated at geometrically progressive intervals over the range from 1 krad to 128 krad. The SI unit for radiation dose is the Gray (Gy), however,

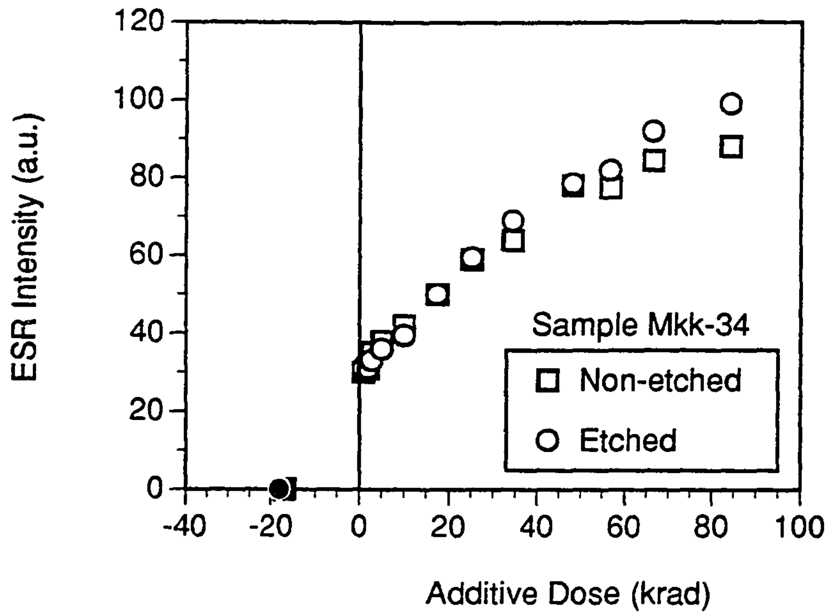


Figure 18. ESR signal intensity versus additive dose in etched and unetched corals.

to avoid confusion with the geological usage of giga years (Gy) radiation doses are reported in rads (100 rads = 0.1 krad = 1 Gy). A growth curve is constructed from the response of the ESR signal intensity (I) measured at g-value 2.0007 to the artificial γ irradiation. The AD is defined by extrapolating the growth curve to a zero ESR signal intensity (see Figure 19). Because the number of trap sites is fixed, the probability of filling traps decreases as more traps are filled. Thus, the growth curve approaches an asymptote, the maximum intensity (I_{MAX}), where all traps are filled. The relation between ESR intensity and laboratory-applied radiation dose, as suggested by Grün and Macdonald (1989), is expressed as:

$$I = I_{MAX} \left(1 - e^{(-a (D - D_E))} \right) \qquad (5.2)$$

where I = ESR intensity,

a = constant,

D = radiation dose,

D_E = equivalent γ-dose also known as the accumulated dose (AD), and

I_{MAX} = ESR intensity at saturation.

The exponential saturation function fits the growth curve data better than a linear function which was used in pioneering coral dating studies by Ikeya and Ohmura (1983). A curve fitting program, FitBas (by R. Grün), utilizing an iterative

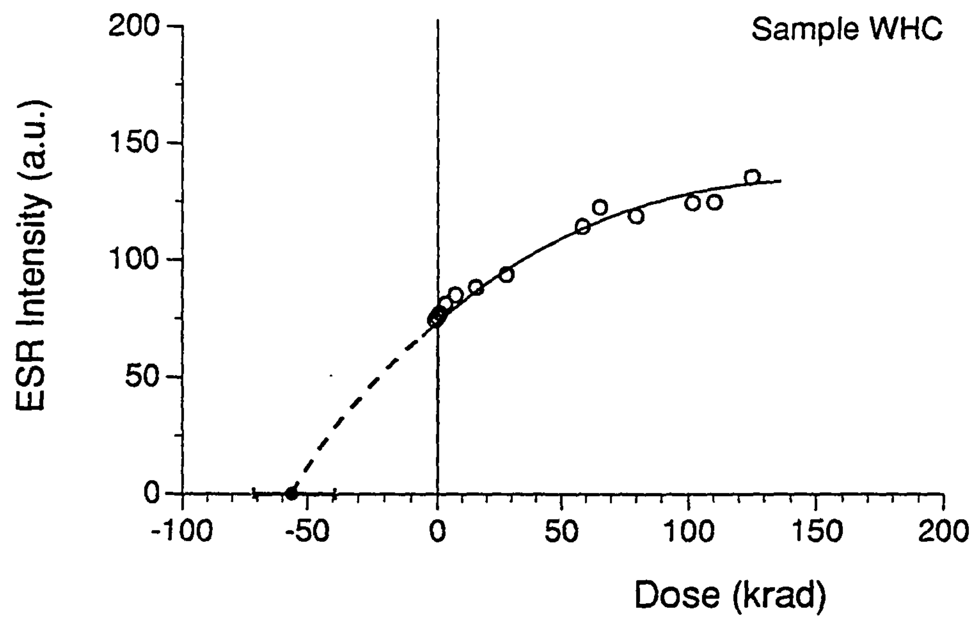


Figure 19. Additive dose method for determining AD.

exponential fit, is used to compute the AD. As a measure of the uncertainty in the AD, the associated error was calculated by a jackknifing procedure (Grün and Macdonald, 1989) which may overestimate the error in the AD (Porat and Schwarcz, 1991). Repeated determinations of AD in corals yield an uncertainty of 10-15% for samples 80 ka to 300 ka (Radtke et al., 1988).

Annual Dose Rate ($\dot{D}$) Evaluation

Corals may experience radiation from three sources, cosmic rays; internally generated radiation from U incorporated into the coral skeleton; and external radiation from the surrounding sediment matrix. The cosmic radiation is dependent on altitude, latitude and attenuation of overlying rocks and sediment and is generally small relative to the internal radiation field, averaging less than 3.0 rad/a (Prescott and Stephen, 1982; Prescott and Hutton, 1988; Grün, 1989a, 1989b). For ESR age calculations reported in this study, the cosmic radiation dose rate is assumed to be 2.5 rad/a based on studies of the total dose rate on Oahu (Brückner and Radtke, 1989). This value is comparable with the cosmic dose rate from other studies, including a value of 2.8 rad/a for sea level at 20°N (Aitken, 1985).

In calculating the ages, the external radiation dose rate is assumed to be similar to the internal radiation dose rate because of the geometry of the corals and the composition of the enclosing matrix. This assumption of an infinite medium proved successful in previous studies that dated corals from Barbados (Skinner, 1985b, 1988; Radtke et al., 1988), New Guinea (Radtke and Grün, 1988), and Oahu (Brückner and Radtke, 1989). In these studies, ESR dates agreed with $^{230}Th/^{234}U$ and He/U dates.

The internal radiation dose was determined by measuring the radioactive element concentration, i.e., U, Th, and K. Uranium concentration typically ranged from 1.8 to 3.2 ppm, which agrees with reported ranges for other fossil corals (Veeh and Turekian, 1968; Livingston and Thompson, 1971; Cherry and Shannon, 1974; Veeh and Burnett, 1982; Radtke et al., 1988; Skinner, 1988; Hamelin et al., 1991). Thorium was at or near the detection limit for the analysis (<0.01 ppm). Potassium values were negligible.

Coral ESR Age

Ages were calculated with the DATA V program (by Grün), which considers U-series disequilibria in the sample. Table 12 gives the values of the parameters used in calculating the coral ESR age. For corals, radon loss does not normally occur (Skinner, 1986).

α/γ Effectiveness Ratio (k)

Because the AD is determined with a γ source, a correction for equivalence of α radiation is necessary. The α/γ efficiency ($\underline{k}$) for corals was estimated to be 0.06 ± 0.01, based on empirical studies by Radtke et al. (1988). Within the coral, only internally generated α particles contribute to

Table 12.
Parameters used for determining ESR ages in coral.

Parameter	Reference
No Rn loss	Skinner, 1986
$\underline{k}$ = 0.06 ± 0.01	Radtke et al., 1988
$(^{234}U/^{238}U)_0 = 1.14$	Moore et al., 1990b
Th, K negligible	This study

the annual dose, because the outer layer (20-30 μm thick), which is dosed by the environment, was removed during sample preparation. The pioneering studies of Ikeya and Ohmura (1983) proposed $\underline{k}$ values near 0.015-0.020 based on comparing ESR with ^{14}C ages for Holocene corals, whereas Koba et al. (1987) preferred a value of 0.055-0.0815 based on comparison with $^{230}Th/^{234}U$ ages from Late Pleistocene corals. A recent study empirically determined the α-efficiency in corals to be 0.06 ± 0.01 (Radtke et al., 1988), using an artificial ^{241}Am α source. Relying on ^{241}Am as an artificial α source may be questionable, because Bradley and Chong (1991) have demonstrated γ emission from ^{241}Am. Lyons and Brennan (1991) preferred using a low-energy research accelerator to determine the α/γ effectiveness for calcite speleothems and recommend a value of 0.05 for $\underline{k}$. They indicated no dependence of $\underline{k}$ on AD, ESR age, or U content. For the ESR age calculations reported in this study, $\underline{k}$ was assumed to be 0.06 ± 0.01. The sensitivity of this parameter is addressed below.

Initial $^{234}U/^{238}U$ Ratio

Modern corals accurately record the $^{234}U/^{238}U$ ratio of seawater in which they live without appreciable fractionation of uranium from calcium during skeletogeny. The activity ratio of $^{234}U/^{238}U$ for modern Hawaiian corals is 1.147 $\pm$ 0.011 (Moore et al., 1990b), similar to the $^{234}U/^{238}U$ ratio in open ocean environments of 1.144 $\pm$ 0.004 (Chen et al., 1986). Recently there has been some debate about whether corals do indeed always have the $^{234}U/^{238}U$ ratio of seawater (Ku et al., 1990). The $^{234}U/^{238}U$ ratio for modern Hawaiian corals are similar to the $^{234}U/^{238}U$ ratio for modern corals from elsewhere (Edwards, 1988; Edwards et al., 1988; Bard et al., 1990; Hamelin et al., 1991; Table 13). However, corals from the last interglacial period appear to have slightly elevated initial $^{234}U/^{238}U$ ratios when compared with modern seawater values (i.e. 1.16, Hamelin et al., 1991), and corals older than 150 ka could have initial $^{234}U/^{238}U$ ratios approaching 1.2 (Bard et al., 1991). However, alteration of the original aragonite and secondary precipitation of calcite from pore waters may be responsible for the observed higher initial ratio in older samples. For the ESR age calculations, a stable initial $^{234}U/^{238}U$ ratio, $(^{234}U/^{238}U)_0$, of 1.14 $\pm$ 0.01 was selected.

Parameter Sensitivity

Over the range of accumulated doses experienced in this study, the sensitivity of the age calculations to variations

Table 13.
High precision mass spectrometric measurement of the ratio of $^{234}U/^{238}U$ in modern corals.

Location Species	$(^{234}U/^{238}U)_0$	Reference
BARBADOS		
Acropora palmata	1.146 ± 0.005	Edwards, 1988
Acropora palmata	1.148 ± 0.002	Hamelin et al., 1991
WINDWARD ISLANDS		
Acropora palmata	1.147 ± 0.004	Edwards, 1988
GALAPAGOS		
Pavona clavus	1.148 ± 0.007	Hamelin et al., 1991
VANUATU		
Goniastria retiformis	1.147 ± 0.003	Edwards et al., 1988
Platygyra sinensis	1.149 ± 0.006	Edwards et al., 1988
Favia sp.	1.148 ± 0.004	Edwards et al., 1988
HAWAII		
Porites sp.	1.147 ± 0.011	Moore et al., 1990b

in key parameters, the α/γ efficiency parameter k, and the initial $^{234}U/^{238}U$ ratio, needs examination. By changing the α/γ efficiency k from 0.06 to 0.05, the age increases by less than 5% (Table 14). The response of varying k on computed ESR ages over the range of AD's reported in this study is plotted in Figure 20. As mentioned above, values for k reported for corals range from 0.015 to 0.0815 with empirically determined values near 0.06 ± 0.01 (Radtke et al., 1988; Grün et al. 1992). The older the sample the greater the degree of response to changes in k and therefore the greater the effect on the computed age. The relative effectiveness of the different types of radiation (α, β, γ) in generating ESR signal intensity requires further evaluation. Whereas the utilization of a low energy accelerator as a source for mono-energetic α particles is a significant step toward resolving this problem in stalactites (Lyons, 1988), experiments with corals using this approach have not yet been performed. The selection of the relative α/γ effectiveness value near 0.06 is based on an extremely limited data set.

The $(^{234}U/^{238}U)_0$ ratio influences the annual dose rate. However, the response of calculated ESR age to varying $(^{234}U/^{238}U)_0$ is not very sensitive over the range of anticipated initial ratios (Figure 21). If $(^{234}U/^{238}U)_0$ were significantly higher, for example 1.2 at 200 ka, and 1.7 at 450 ka, as speculated by Bard et al. (1991), then the computed ESR coral ages would be only slightly younger than those reported in

Table 14.
ESR ages: effect of α-efficiency and initial $^{234}U/^{238}U$.

AD (Gy)	α-efficiency (k)	$(^{234}U/^{238}U)_0$	ESR Age (ka)
100	0.05	1.14	120.7
		1.15	120.3
	0.06	1.14	116.3
		1.15	116.0
200	0.05	1.14	208.1
		1.15	207.4
	0.06	1.14	199.9
		1.15	199.3
400	0.05	1.14	367.0
		1.15	366.6
	0.06	1.14	351.9
		1.15	350.8

this study (Table 13). Therefore, the computed ESR ages are reasonably insensitive to changes in either k or the $(^{234}U/^{238}U)_0$ ratio.

Recommendations for Sampling

Sampling coral for ESR dating requires a few precautions. Corals representative of the geological feature of interest should be preferably in growth position. Where possible, samples from sites exposed to direct sunlight should be avoided, because elevated rock surface temperatures (up to 80°C) may significantly affect the rate of retrapping of unpaired electrons (Grün, 1989a, 1989b). In the field, corals should be examined for any diagenetic alteration. By breaking open a hand specimen, a visual assessment of the amount of

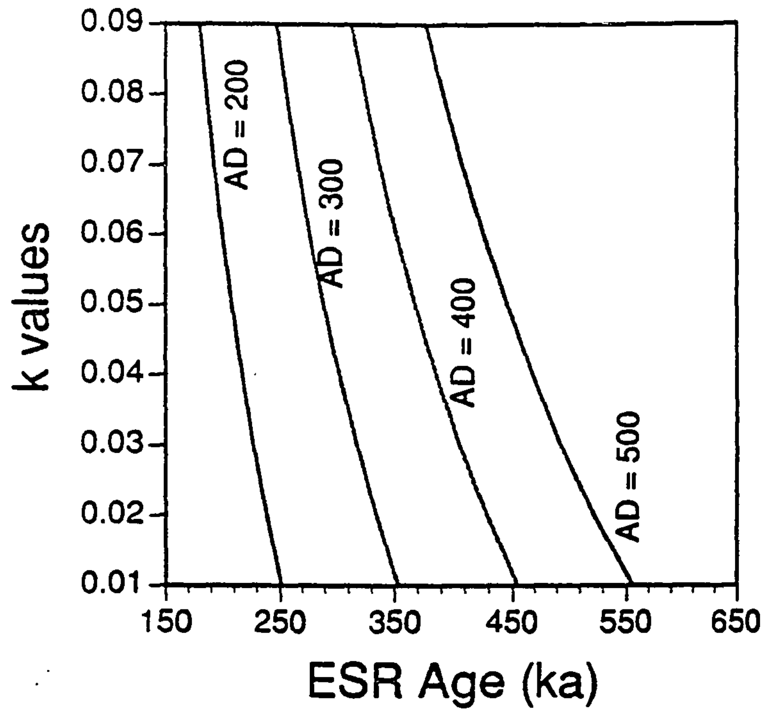


Figure 20. The effect of α/γ efficiency on ESR coral age.

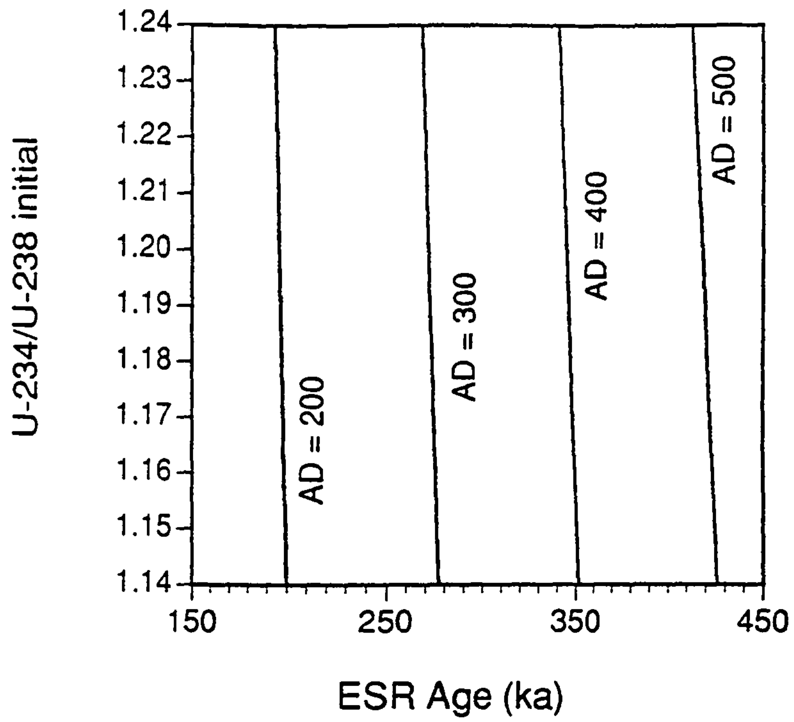


Figure 21. The effect of initial $^{234}U/^{238}U$ on ESR coral age.

alteration can be made by examining the exposed surface for calcite cleavage planes which glisten in sunlight. Aragonite-rich samples can later be carefully screened in the laboratory for datable material. The original internal architecture should also be apparent in hand samples, with the characteristic porous structure and intricate design normally seen in the individual calices plainly visible. Any discoloration or chalky surfaces should be avoided.

Although it is preferable to measure the external γ activity and cosmic dose rate _in situ_, it is rarely accomplished. This is due to several factors, including cost of a portable gamma spectrometer or cost of revisiting the site to retrieve the TL dosimeters used to measure the cosmic dose component. Laboratory analysis of radioisotopes in the sample and surrounding matrix is currently the accepted alternative to extensive field measurement of the radiation field. Therefore, if the coral is in a reef structure, the sediment, or matrix surrounding the coral should also be collected for NAA analysis. However, if the corals are near a regolith, γ dose rates should be measured _in situ_. Detailed notes on stratigraphy and elevation should be recorded and interpreted to provide stratigraphic control on the reliability of ESR results. As outlined above, each sample should be analyzed by XRD to avoid samples with calcite concentrations greater than about 5 percent, because aragonite to calcite inversion causes underestimation of AD (Radtke et al., 1988) and is often accompanied by U migration (Kaufman, 1986).

Summary

The development of ESR dating provides a simple and relatively quick means for recording an age-dependent signal. The ESR dating signal is derived from the natural radiation that the coral has received since its time of formation. The accumulated dose (AD) is determined by the additive dose method. An ESR age is computed from the AD by evaluating the rate of annual radiation (D). In the use of ESR dating of Hawaiian fossil corals, several factors related to sample processing were investigated, including etching and post-irradiation annealing. For sample preparation, an aliquot size of 30-50 mg and a grain size between 425-250 μm are recommended. No chemical pretreatment may be required, although post-irradiation annealing at 90°C for 3 days is suggested. The post-irradiation annealing is considered adequate, but further clarification of signal response is needed. With regard to the annual dose rate, various parameters should be evaluated, including the α/γ efficiency, secondary mineralization, diagenetic alteration, uranium accumulation, and disequilibria in the U-decay series.

ESR dating applied to corals has been successful in dating the late Pleistocene interglacial periods.

CHAPTER VI

UPLIFTED PLEISTOCENE REEFS ON OAHU

Introduction

Studies on the chronology of marine fossil deposits in the Hawaiian Islands have concentrated on raised reefs on Oahu, especially the Waimanalo Formation. Several high stands, which precede the last interglacial, have been described by Stearns (1935a, 1974a, 1978a) (Table 15). Studies on Oahu historically have been used in evaluating eustatic sea levels, especially of the last interglacial. Veeh (1966) was the first to date the Waimanalo formation and correlate this sea level stand with similar high stands elsewhere in the world, demonstrating a eustatic sea level rise of +2 to +9 m during the last interglacial ca. 120 ka. Ku et al. (1974) documented the age of the Waimanalo Formation at 122 $\pm$ 7 ka. The elevation for the Waimanalo stand as specified by Ku et al. was +7.6 m taken from the mean of two wave-cut notches. The conclusion reached by Ku et al. (1974), based on the elevations of the last interglacial at other tectonically stable locations, was that Oahu has been tectonically stable for at least 122 ka. Since 1974, several workers, most notably Easton and Ku (1981) and Brückner and Radtke (1989), have dated fossil deposits on Oahu, but the conclusion that Oahu has been a stable platform must remain open.

In 1973, Ward attempted to correlate the Australian Pleistocene shorelines in Gippsland with shorelines from Oahu.

Table 15.
Elevation and estimated age of emerged shorelines described from Oahu (after Stearns, 1978a).

Hawaiian Shoreline Name	Approximate Elevation (m)	Estimated Age (ka)
Present	0	
Kapapa	1.5	4
Leahi	0.6	115
Waimanalo	7.5	125
Kailua	3.6	
Waialae	12	
Laie	21.5	
Kaena	29-30	650
PCA	7.5	
Kahuku Point	17	

For the Oahu shorelines, Ward (1973) relied on the known relation of the former shorelines to the post-erosional basalts and the available K-Ar dates for the Honolulu Volcanics (Gramlich et al., 1971). His postulated ages for several emerged shorelines are summarized in Table 16. In comparing the sequence of elevated shorelines on Oahu relative to shorelines from Gippsland, Australia; South Carolina; and Mangaia, Cook Islands; Ward (1973) concluded that Oahu was rising at a rate of 0.016 m kyr^{-1} contrary to the later findings of Ku et al. (1974). Stearns (1974b) refuted Ward's conclusions stating that "Oahu is not rising now and did not rise at any time in the Pleistocene" (Stearns, 1974b, p. 1189).

Table 16.
Estimated age of Pleistocene shorelines based on relation of shorelines with post-erosional basalts on Oahu (from Ward, 1973).

Hawaiian Terrace Name	Suggested Age (ka)
Leahi I	40-120
Waimanalo	120 ± 20
Waipio	270-320
Kaena	400-450
Olowalu	<2,150

One objective of the present study was to test competing hypotheses regarding the stability of Oahu (Figure 24). The chronology of marine deposits developed from new data obtained in this study at previously-unstudied exposures and published chronological studies are used to refine estimates for the age of elevated Pleistocene reef deposits on Oahu and re-evaluate the tectonic stability of the island. Previous studies, primarily by Ku et al. (1974), Easton and Ku (1981) and Brückner and Radtke (1989), provide useful data to test these hypotheses. In total, more than 100 dates of marine fossils from deposits on Oahu have been reported (Appendix 1).

New Data

Several new exposures of elevated reef deposits not previously studied were investigated. These sites include exposures along a drainage canal at Barber's Point (BPC), a temporary trench at West Beach (WB), and along a road cut at the Waianae Comprehensive Health Center (WHC), as well as a limestone outcrop near Sandy Beach known locally as Lanai

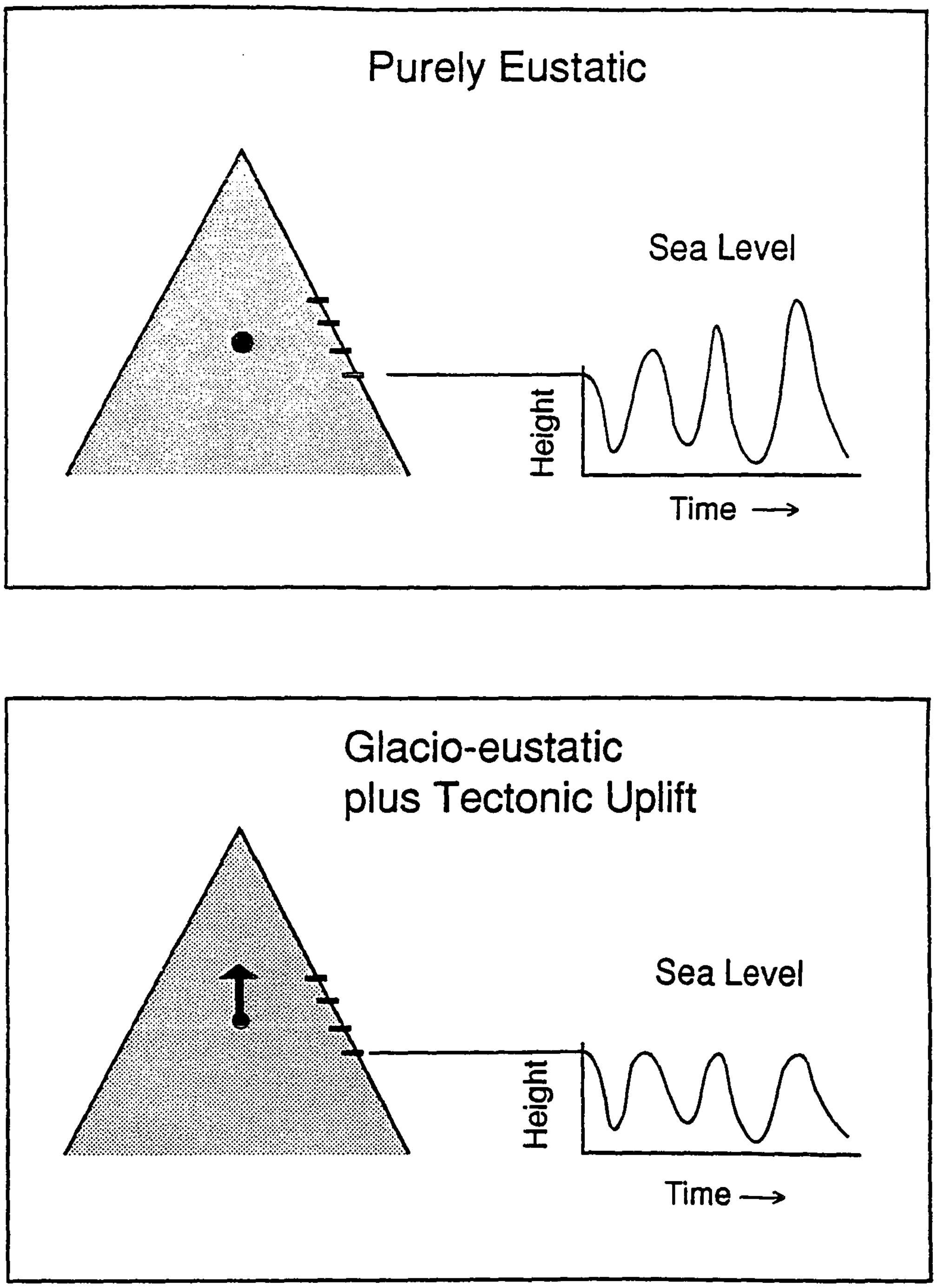


Figure 22. Competing models for Oahu emergent relic shorelines.

Lookout (LL). The type locality of the +30 m stand at Kaena Point (Stearns, 1935a) and a lower elevation (+4 m) stand at Kaena Point were also reexamined (Figure 23). All coral specimens sampled at these sites were preserved in their growth positions. All of the coral samples are >98% aragonite in composition, as determined by X-ray diffraction, indicating no significant chemical alteration. It is assumed that the Porites and Platygyra sampled, lived in water depths of 1-10 m based on the distribution of the modern genus. In the Hawaiian Islands, the genus Porites is a common member of the modern reef community and the principle reef builder (Maragos, 1977). In contrast, the genus Platygyra is not a member of the modern assemblage of corals in Hawaii, but is commonly found on upper reef slopes, back reef margins and on reef flats throughout the Indo-west Pacific (Veron, 1986).

Barber's Point

At Barber's Point, a drainage canal dug in 1972 normal to the coast, exposes a well preserved reef section approximately +6 m above sea level (Figure 24). Corals collected at the Barber's Point canal (BPC) were dated by ESR. The stratigraphy, depositional and diagenetic history of these fossil deposits are described in detail elsewhere (Sherman, 1992).

West Beach (Ko Olina)

During construction of a golf course at "West Beach" (Ko Olina), a trench was dug exposing a section of coral reef

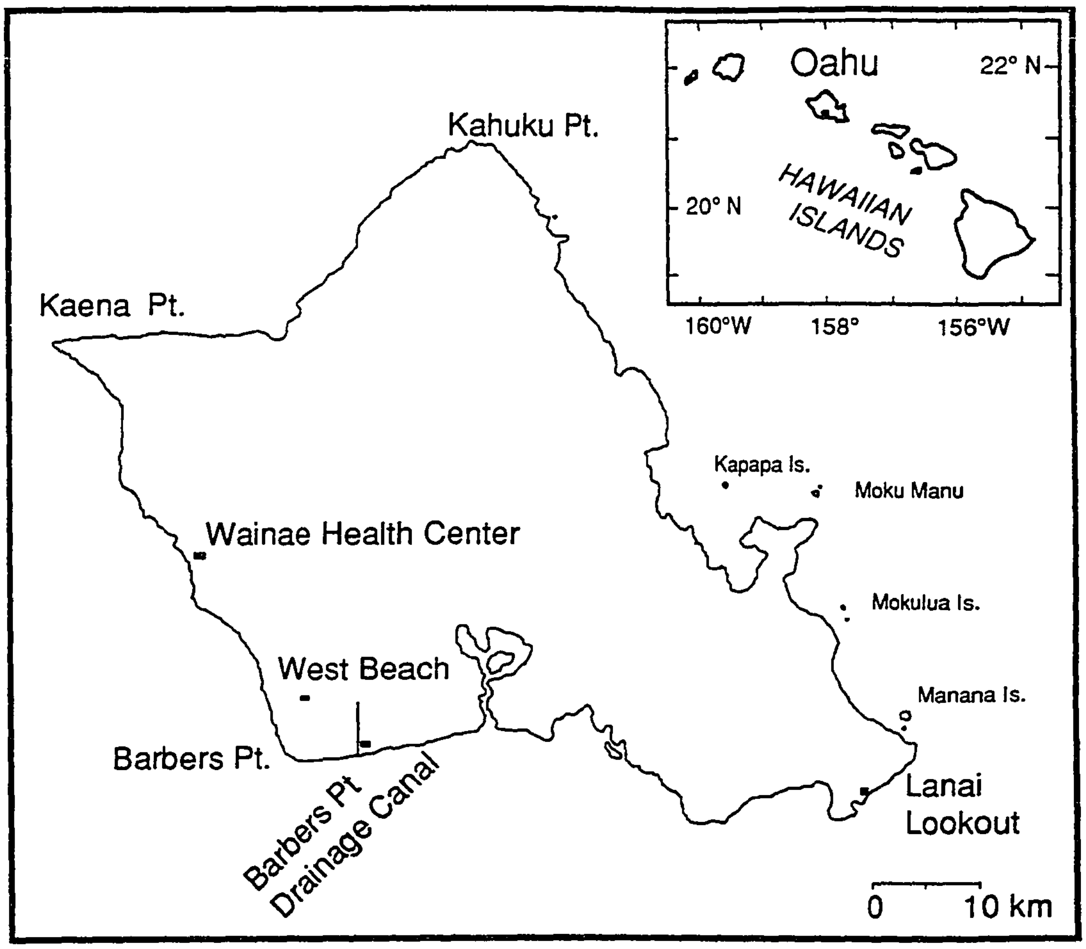


Figure 23. Index map of Oahu showing locations of study sites.

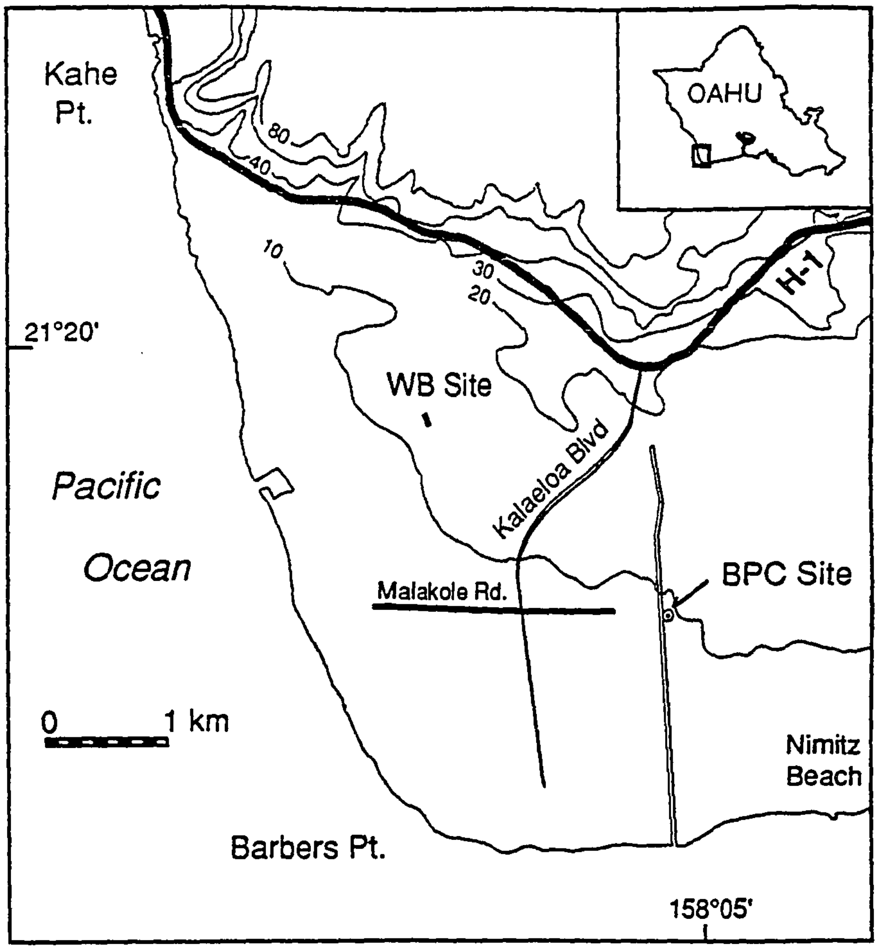


Figure 24. Location of Barber Point drainage canal (BPC) and West Beach sites.

(Figure 25). Corals were collected in situ at an elevation of approximately +14 m and dated using the ESR method. The nature of the bedrock is unknown because it was not exposed. This site has been re-landscaped and covered since sampling in 1989.

Waianae Health Center

Three large (1 m diameter) dome-shaped corals (*Platygyra* sp.) situated in growth position behind the Waianae Health Center (WHC) at an elevation of +26.5 m were sampled and dated with ESR (Figure 26). This site has been protected from extensive diagenesis by its microtopography. Immediately north of the WHC site around the Puu Haleakala ridge, in Lualualei Valley, highly altered limestone has been quarried. The quarry walls are very fossiliferous, but the material is strongly recrystallized with most of the macrofossils now represented by molds of corals and mollusks.

Lanai Lookout

An inlier of limestone is exposed at the tuff bench below the Lanai Lookout Scenic Point parking area along the Kalanianaole Highway. The limestone is mantled by ash erupted from the Hanauma Bay - Koko Fissure as part of a post-erosional, rejuvenation stage of volcanism in eastern Oahu (Honolulu Volcanics)(Moberly and Walker, 1987). This outcrop is described by Moberly and Walker (1987; Figure 27). The highest elevation of outcropping limestone is +3 m which is the base of the contact between the limestone and a paleosol.

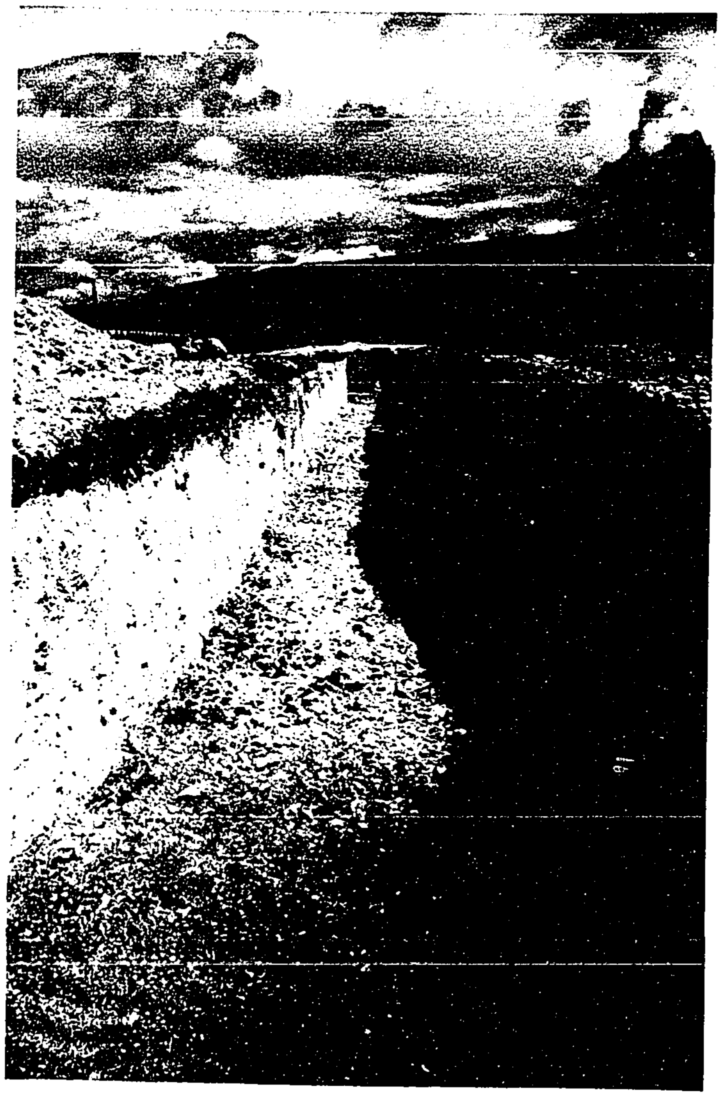


Figure 25. Photograph of West Beach sample location looking north with the HECO electric power plant stack at Kahe Point in the background.

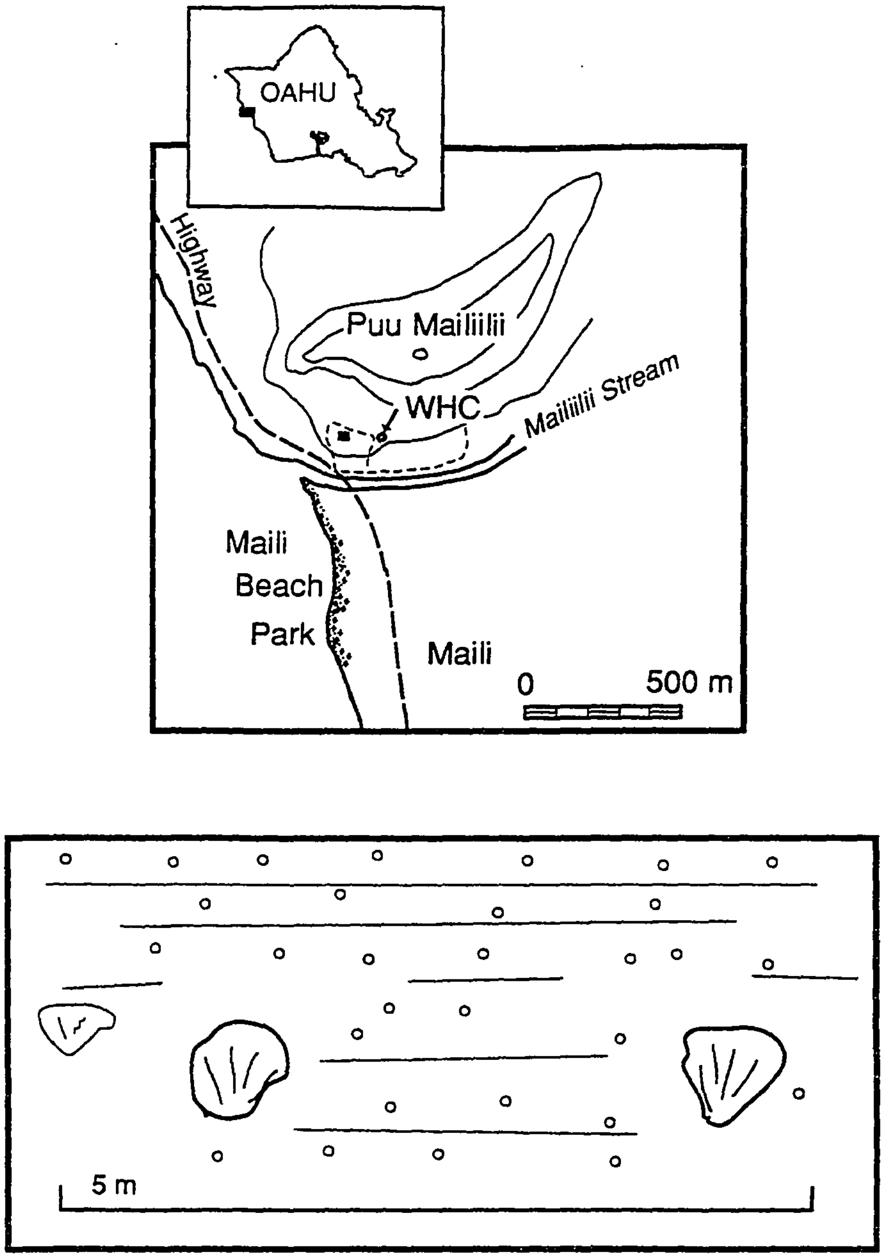


Figure 26. Sketch of exposure of <u>Platygyra</u> at the Waianae Health Center.

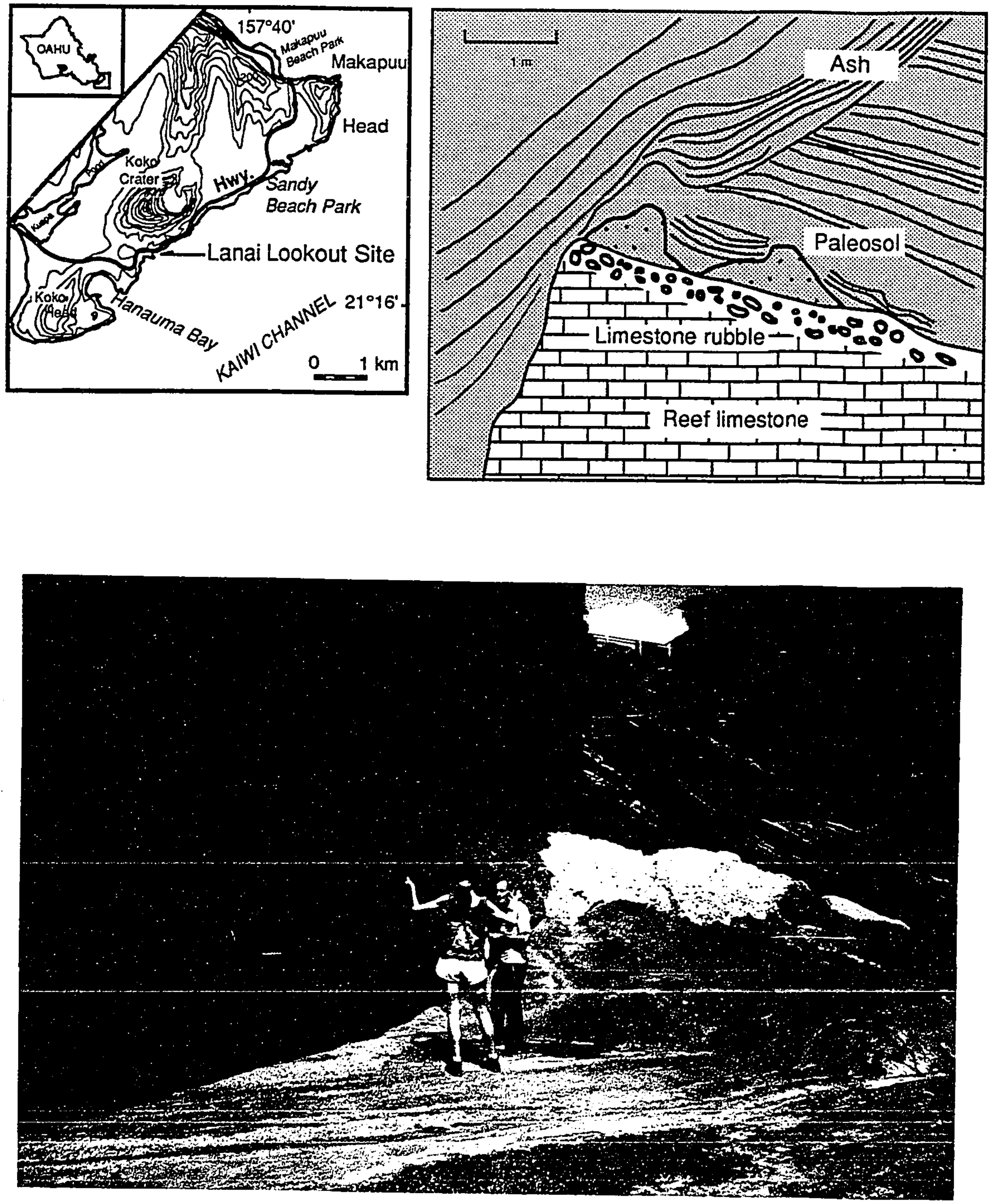


Figure 27. Photograph of Lanai Lookout outcrop showing limestone capped with paleosol.

The limestone outcrop extends below sea level. Prior to the addition of volcanic material from the pyroclastic eruptions along the Koko Head - Koko Crater fissure, the limestone reef capped with a paleosol stood as an offshore islet. The original Oahu shoreline was located approximately 2 km north of the study site and now is buried by the ash flows (Stearns, 1939; Moberly and Walker, 1987).

A red paleosol, 79 cm thick and extending horizontally for 1.3 m, mantles the well lithified limestone unit. A fossil coral, _Platygyra_ spp., was sampled in growth position at +2 m elevation. The upper surface of the limestone unit has a scattering of _Porites_ fragments. _Porites compressa_ dominates this coral assemblage.

Kaena Point

Corals were collected at the type locality of the Kaena stand as described by H.T. Stearns (1935a). The elevation of the site is approximately 26.5 m. Stearns refers to the Kaena stand or the "+30 m stand" based on the uppermost reefal deposit of limestone at Kaena Point. This stand can be traced for at least 2 km along the southern portion of Kaena Point towards Makaha.

A carbonate unit located at a lower elevation (~4 m) at Kaena Point was also sampled. Coral (_Porites_) from a coral algal framestone (KP3-2) at approximately +3 m elevation was dated by ESR and mass spectrometry U-series. Sherman (1992) has described the lithofacies and interpreted the depositional environment at this locality. From the coral morphology in the framestone and the macro-scale form of the framestone, Sherman (1992) suggested a near-shore high energy environment of deposition. James (1989) has reported on the molluscan assemblage of this unit and concluded that all the species examined were shallow water species (intertidal to shallow subtidal) indicating a shallow water depositional environment.

Results

The results of ESR analysis and calculated ESR ages of late Pleistocene coral samples from Oahu are presented in Table 17. The uranium and thorium isotope results from corals are shown in Table 18 with the calculated U-series age. The mass spectrometry analysis was preformed at McMaster University.

"Last Interglacial Reef" Deposits

The ESR ages of four samples indicate deposition of these reefs during the last interglacial when sea level reached a maximum of 6 m above the present (Land et al., 1967; Matthews, 1973; Bloom et al., 1974). The age of the lower elevation (<5 m) Kaena Point coral framestone unit (KP3-2, MS-$^{230}Th/^{234}U$: 129.4 $\pm$ 1.3 ka; ESR: 139 $\pm$ 12 ka) agree with dates for the Waimanalo Formation (115 - 137 ka, Ku et al., 1974) within limits of experimental error. The three dates from BPC, however, are generally older than the age of the last interglacial period (120-135 ka). An indurated coarse grainstone comprising slabs (0.5 m thick, 1-2 m in length)

Table 17.
The Accumulated Dose (AD) and ESR dates of corals from Oahu.

Sample	Elevation (m)	AD (krad)	ESR Age (ka)
BPC-10	5	10.9 ± 6.8	122 ± 8
KP3-2	3	14.4 ± 1.3	139 ± 14
BPC-2A	5	14.8 ± 1.0	142 ± 20
BPC-9	5	16.2 ± 2.6	152 ± 25
WB	14	19.6‡	196 ± 29
LL-3	3	40.0 ± 10.0	312 ± 79
WHC	26.5	55.7 ± 16.0	468 ± 136
KAE-3	26.5	72.5‡	547 ± 82
BPC-6	2	79.2 ± 12.9	562 ± 96

† Isotope stages follow Imbrie et al. (1984).
‡ Analysis by U. Radtke, Dusseldorf University (written comm., 1989) assumes a 15% error, see Radtke et al. (1988).

form a shingle pavement separating the upper coral bafflestone sample (BPC-10, ESR: 122 ± 8 ka) from the lower coral framestone samples (BPC-2A, MS-$^{230}Th/^{234}U$: 176.1 ± 11 ka; ESR: 142 ± 20 ka; BPC-9, MS-$^{230}Th/^{234}U$: 146.2 ± 9.5 ka; ESR: 152 ± 25 ka). The grainstone slabs are interpreted as a regressive beach facies indicating a minor regressive phase between the two units (Sherman, 1992).

For mass spectrometric analysis of coral samples, Hamelin et al. (1991) have proposed that samples with calculated initial $^{234}U/^{238}U \geq 1.17$ have been diagenetically altered. This caveat suggests that only the MS-$^{230}Th/^{234}U$ date of sample KP3-2 is reliable. The remaining analytical runs must be considered

Table 18.
Isotopic composition (U, Th) and age for corals from Oahu.

Sample	$^{230}Th/^{234}U$	$^{234}U/^{238}U$	$(^{234}U/^{238}U)_{init}$	Age (ka)
BPC-2A	0.8348 ± 0.0016	1.1189 ± 0.0011	1.2003 ± 0.0024	183.9 ± 1.2
	0.8034 ± 0.0029	1.1127 ± 0.0012	1.1815 ± 0.0028	168.2 ± 1.8
BPC-9	0.7354 ± 0.0014	1.1111 ± 0.0015	1.1649 ± 0.0025	139.5 ± 1.0
	0.7699 ± 0.0158	1.1146 ± 0.0011	1.1771 ± 0.0049	152.9 ± 7.0
KP3-2	0.7090 ± 0.0035	1.0993 ± 0.0012	1.1437 ± 0.0022	130.3 ± 1.0
	0.7033 ± 0.0010	1.0974 ± 0.0012	1.1402 ± 0.0019	128.5 ± 1.0
WHC	1.0636 ± 0.0035	1.0462 ± 0.0012		>240

Notes: Reported errors quoted are 1 standard deviation.

suspect with possible initial ^{230}Th incorporated as detrital material during skeletogeny. Although the ^{230}Th age may represent an older estimate of their age of formation (Table 18).

The lithofacies at Barber's Point Canal suggest two highstands at least 5 meters above the present level (Sherman, 1992). This interpretation is supported by the ESR dates (Figure 28). The presence of two highstands during the last interglacial (oxygen isotope substage 5e) agrees with the re-interpretation of Ku et al.'s data by C.E. Stearns (1976) and Chappell and Veeh (1978) who have suggested a double transgression with a highstand at 133 ka and a later event at 118 ka (see Moore, 1982).

Double Nips of the Waimanalo Stand

Harold Stearns (1935a) first described two wave-cut notches or nips in eolianite at +8.2 and +6.7 m to mark the paleoelevation of the Waimanalo stand (Figure 29). Later, Ku and others (1974) used an average value of +7.6 m for the elevation of the Waimanalo stand. The field relations and ESR dates for the BPC are consistent with the interpretation of the ages of the two nips indicating two stands of the sea during the last interglacial period. From the difference in elevation of the two notches, I speculate that the height difference between the two sea level events was approximately 1.5 m.

The idea of a pair of highstands during the last

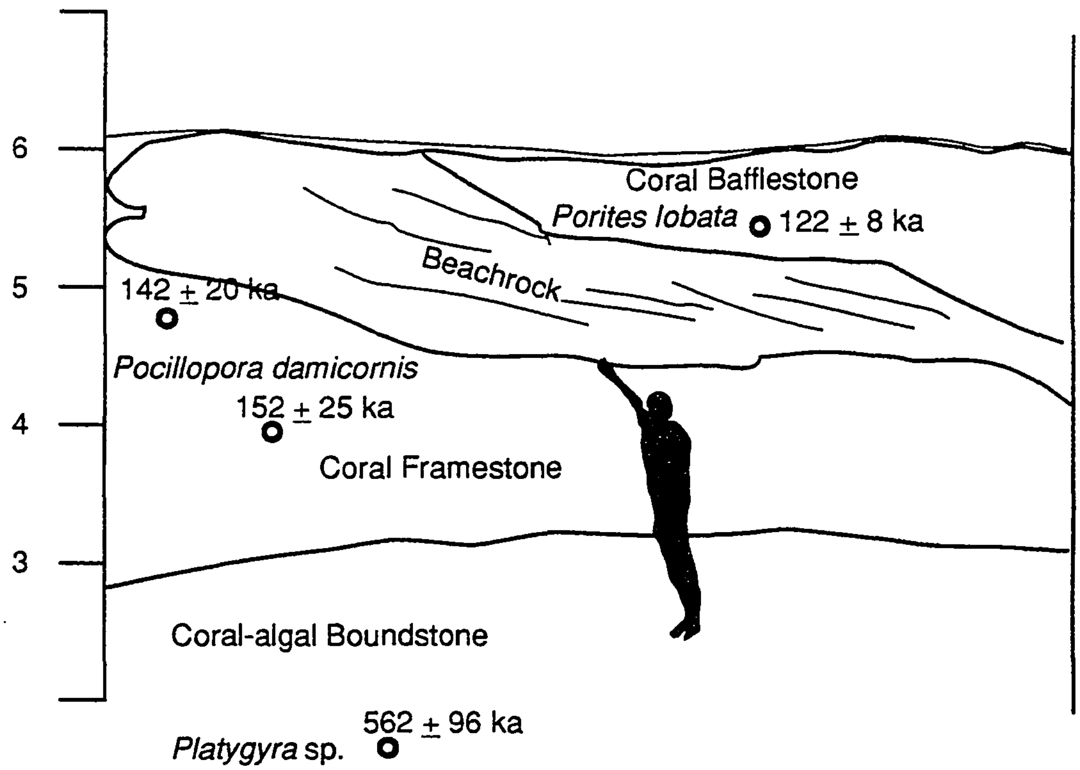
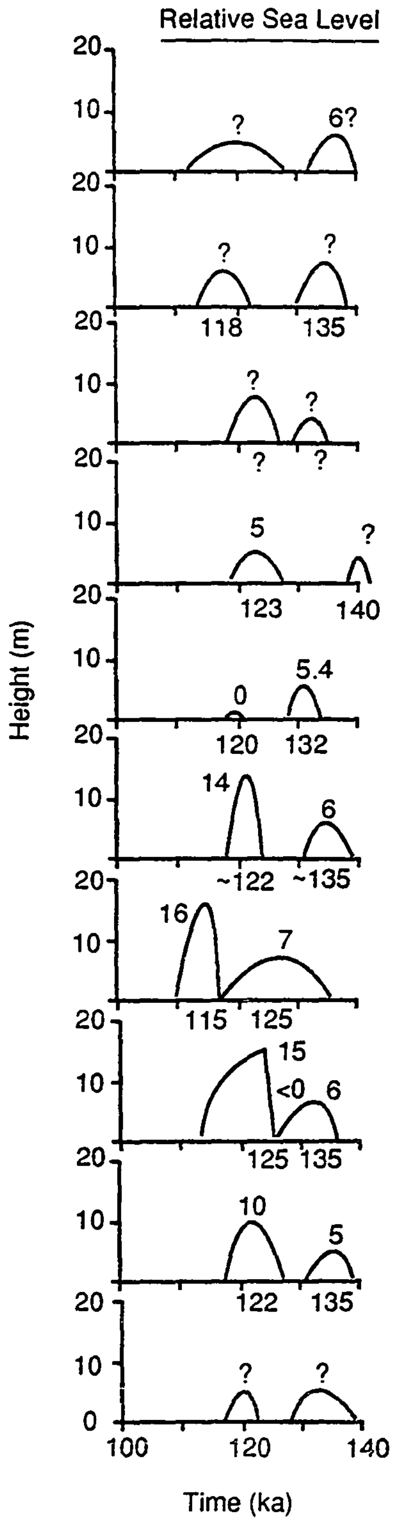


Figure 28. Stratigraphy of the Barber's Point Canal with ESR dates (adapted from Sherman, 1992).

Figure 29. Photograph near Waimanalo of paired marine notches in eolianite (from Stearns, 1935a).

interglacial period (stage 5e) has been increasing recognized at widely scattered localities (Hollins, 1977; Hollins and Hearty, 1990; Sirkin et al., 1990; cf. Kaufman, 1986; see Table 19). However, it has also been strongly disputed whether the last interglacial period comprises a single event (Harmon et al, 1983, CLIMAP, 1984) or two separate events (Chappell and Veeh, 1978; Moore, 1982). The interpretation of the uplifted reef tracts from Barbados have been central to this debate. A double transgressive sequence was not originally recognized from this locality (Broecker et al., 1968; Mesolella et al., 1969; Matthews, 1973; Bender et al., 1979), whereas a double transgression was described from the uplifted reef tracts on the Huon Peninsula in Papua New Guinea (Bloom et al., 1974; Chappell, 1983). Recently, Johnson (1991) proposed a possible double highstand on Barbados based on uplift rates and precisely-measured relative elevation of sea level features. In contrast to the uplifted coasts of Barbados, recent mass spectrometric U-series dating studies of corals from tectonically stable, San Salvador Island, Bahamas, have been interpreted as a single event for the duration of this period (Chen et al., 1991). These new data, which suggest that the last interglacial sea level preceded the calculated Milankovitch maximum insolation by almost 5 kyr and continued 8 kyr after the maximum insolation, seriously question the paired highstands and the relationship of sea level to the astronomical theory of climate change.

Table 19. Double transgressions during the last interglacial period (Oxygen isotope substage 5e).



Locality	Reference
Fradis Minoris Sardinia	Ulzegas and Hearty, 1986
Hergla Tunisia	Miller et al., 1986
Campo de Tiro Mallorca	Hearty, 1987
La Paz Baja California	Sirkin et al., 1990
Barbados	Johnson, 1991, 1992
Myrtle Beach South Carolina	Hollin and Hearty, 1990
Thames Estuary England	Hollin, 1977
Global	Hollin, 1982 as modified in Hollin and Hearty, 1990
Mediterranean	Hearty, 1986 as modified in Hollin and Hearty, 1990
Atauro	Chappell and Veeh, 1978

A compilation of published dates from Oahu for the last interglacial period indicates that two stages of sea level are recorded during the period 140-105 ka (Figure 30, data from Appendix 1). Further studies directed toward this question are needed and several exposures on Oahu of the Waimanalo Formation would provide excellent study sites.

The current assumption, based on elevations of reefs on tectonically-stable islands, is that sea level reached a maximum of 6 m during the last interglacial (e.g. Matthews, 1973; Bloom et al., 1974). The elevation proposed for the highest event on Oahu during the last interglacial is 8.2 m which suggests uplift of ca. 2.2 m in 125 kyr. If this difference is due to uplift, then it represents a net rate of 0.018 m kyr^{-1} (0.018 mm yr^{-1}) over this time period. If, however, the 12 m elevation proves to be the highest elevation during the last interglacial, then a faster uplift rate would be required (0.048 m kyr^{-1}).

Stage 7 Interglacial

The Stage 7 interglacial is defined for this discussion as the interglacial that occurred between 186 and 245 ka. It is referred to as Stage 7 from the deep-sea oxygen isotope nomenclature. The West Beach reef is assigned an age of 196 $\pm$ 29 ka based on ESR dating. This reef, situated at approximately 14 m above sea level, can be correlated chronologically with the isotope Stage 7 interglacial.

On the eastern end of Oahu, Easton (1963) has described

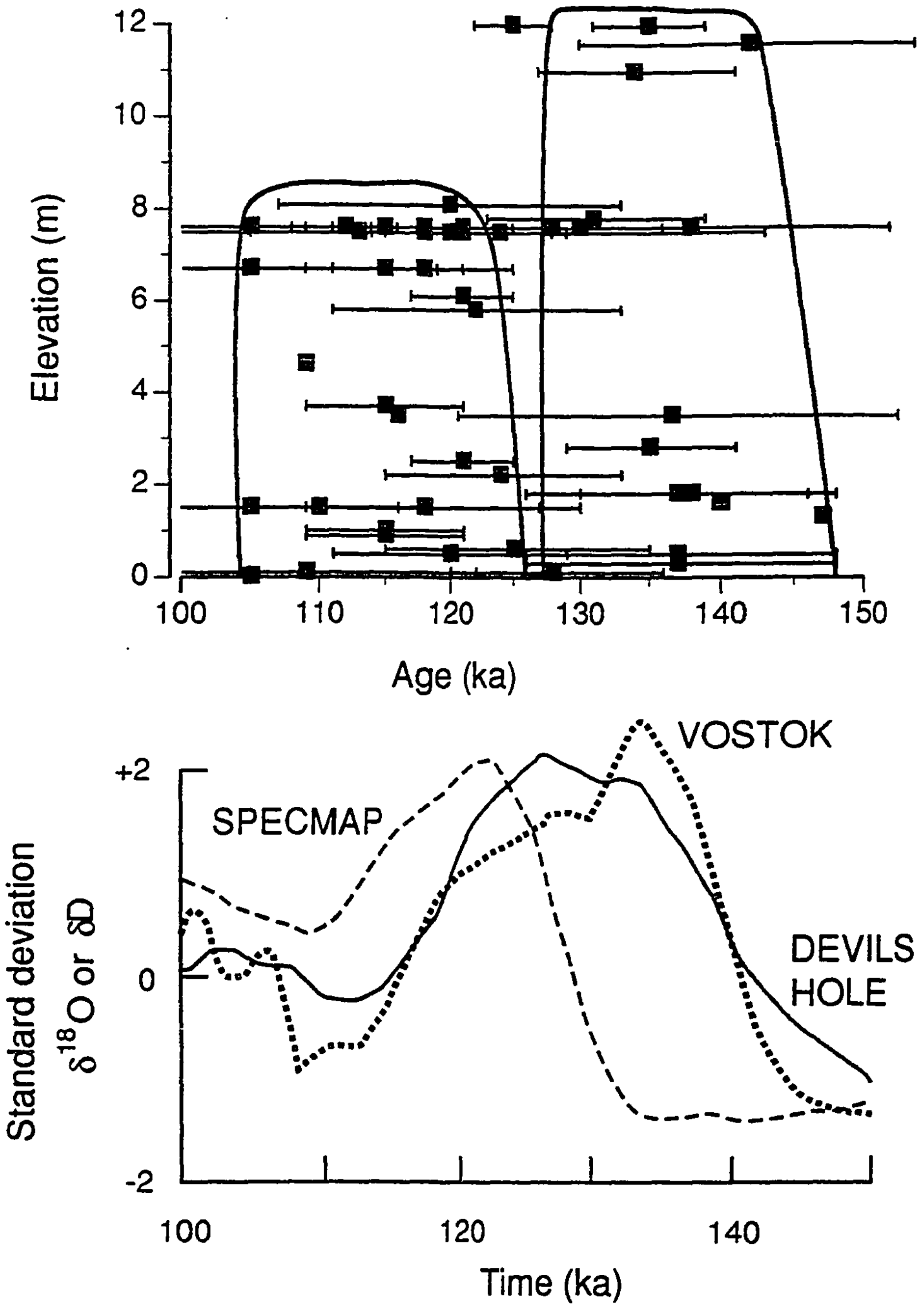


Figure 30. Compiled record of elevation of dated marine limestones on Oahu for the period 100-150 ka shown with three climatic curves: $\delta^{18}O$ continental record from Devils Hole, Nevada (Winograd et al., 1992), oceanic $\delta^{18}O$ curve from compiled deep-sea cores (SPECMAP, Williams et al., 1988) and δD from ice core at Vostok (Jouzel et al., 1987).

a wave-cut bench with a veneer of marine fossils uncovered during excavation of Hawaii Kai on Manua O Ahi Ridge. This geomorphological feature, which includes the presence of sea caves, is evidence for the existence of a ca. 12 m shoreline representing the Waialae shoreline. Based on the ESR date from the reef at West Beach and the descriptions and elevation on Oahu of a +12 m or "40-ft" shoreline of Easton (1963), the Waialae shoreline of Wentworth (1926) is herein assigned an age of 196 ± 29 ka. Further study and dating of this shoreline are necessary to confirm its geographical extent and chronology. Application of amino acid racemization dating to mollusc materials collected by Easton (1963) from the Waialae shoreline would be one such avenue toward establishing the chronology of this shoreline.

During the Stage 7 interglacial, sea level has been estimated to -12 to +2.5 m elsewhere in the world (Table 20). The chronology of emerged corals and submerged speleothems from stable platforms such as Bermuda, suggest that sea level during the stage 7 interglacial was between about present level and -6 m from about 195 - 210 ka (Harmon et al., 1983). Hearty et al. (1992) subsequently have proposed that the maximum elevation of sea level on Bermuda during this interglacial was 2.5 m above present sea level. Slightly different sea level histories are recorded on tectonically uplifted coasts where using a simple assumption

Table 20.
Estimated maximum height of past three major interglacial highstands from stable and uplifted coastal terraces.

		Oxygen Isotope Stage		
		5	7	9
Stable	BERMUDA†			
	Terrace:	Rocky Bay	Belmont	Upper Town Hill
	Age (ka)	124 ± 4	204 ± 11	330 ± 20
	Elev. (m)	5 ± 1	2.5 ± 1	5 ± 3
Uplifted	BARBADOS‡			
	Terrace:	Rendezvous Hill	Kingsland	Rowans
	Age (ka)	125	229 ± 37	310 ± 8
	Elev. (m)	(+54 ± 3)	(+79 ± 3)	(+110 ± 5)
	Paleo-elev.* (m)	6	0 ± 10	0 ± 10
	SUMBA IS., INDONESIA§			
	Terrace:	I_2	II_2	III_1
	Age (ka)	123 ± 15	190 ± 65	348 ± 42
	Elev. (m)	(+12 ± 7)	(+51 ± 6)	(+145 ± 10)
	Paleo-elev.§ (m)	0 ± 5	-12 ± 12	0 ± 5

† Harmon et al., 1983; Hearty et al., 1992; see Table 7 for details.
‡ Bender et al., 1979; Radtke and Grün, 1990.
* Assume 6 m at 125 ka.
§ Uplift rate of 0.5 mm/a; Pirazzoli et al., 1991.

of constant uplift, the present elevation of interglacial highstands can be adjusted for vertical uplift. This procedure enables reconstruction of paleosea level elevations at the time of the interglacials. Late Pleistocene sea level reconstructed from uplifted reef tracts on Papua New Guinea indicates that sea level was close to the present level during the Stage 7 interglacial around 212 ka ago (Chappell, 1983). On Barbados, a Stage 7 interglacial reef tract has been dated at 229 $\pm$ 37 ka and the maximum elevation of sea level during that period is estimated to have been at 0 $\pm$ 10 m (Bender et al., 1979). Similarily on Suma Island, Pirazzoli et al (1991, 1992) have back calculated paleo-elevations of -12 $\pm$ 12 m for sea level during the Stage 7 interglacial 190 $\pm$ 65 ka.

From each of these selected studies, the maximum height of Stage 7 is always less than the maximum height of Stage 5e. The presence of a Stage 7 reef, the West Beach reef at 14 m, elevated above the Stage 5e Waimanalo stand is convincing evidence for continued uplift of Oahu between 200 and 125 ka. The rate of uplift can be estimated from the above studies to be approximately 0.06 $\pm$ 0.01 mm yr^{-1} for Oahu (Table 21).

The ages of the WB reef can be correlated to a warm stage, Stage 7.1, of the deep-sea oxygen isotope record (Imbrie et al., 1984). Stage 7.1 has an age of 193 ka as inferred from a composite oxygen isotope record that has been orbitally tuned according to Milankovitch theory (Martinson et al., 1987). In conclusion, the estimated ages for this

Table 21.
Net uplift rate for elevated reefs on Oahu for period 200 ka to 125 ka.

Locality[†]	Highstand Age (ka)	Max. Elev. of Sea Level (m)	Uplift Rate for Oahu[‡] (m/ka)
Stable Coast:			
Bermuda	204 ± 11	2.5 ± 1	0.06 ± 0.01
Tectonically Uplifted Coasts:			
Barbados	229 ± 37	0 ± 10	0.07 ± 0.05
Suma Is. Indonesia	190 ± 65	-12 ± 12	0.13 ± 0.06
Papua New Guinea	212	0	0.07

† See text for references to localities.
‡ Assumes a 196 ka elevated shoreline at 14 m.

highstand is constrained to 200 ± 10 ka.

Earlier Interglacials

The Lanai Lookout limestone (LL) is estimated to be 312 ± 79 ka and can be correlated to isotope stage 9. The ESR date indicate that when the Koko Rift erupted (ca. 30 ka, Gramlich et al., 1971; Lanphere and Dalrymple, 1980), this reef had been fossilized and that the fragments of limestone ejected during the phreatic explosions and preserved as bombs in the ash layers along the Koko fissure, do not date the phreatic eruption but the age of the fossil reef that was intruded. The elevation of this limestone at 3-4 m points to the fact, recognized by Easton and Ku (1981), that assigning an age to a shoreline deposit based solely on elevation can be inappropriate. Reef limestone ejecta in the tuffs on Manana Island, off the windward coast of Oahu, display a similar diagenetic state as the LL reef limestone and the carbonate ejecta observed around Hanauma Bay and Makapuu Head.

The species of Platygyra from the Waianae Health Center was ESR dated at 468 ± 136 ka. The MS-$^{230}Th/^{234}U$ age of this sample is >240 ka. A set of ESR dates from the +30 m Kaena Point stand, including sample KAE-3, range in age from 406 to 547 ka with a mean of 482 ka (Brückner and Radtke, 1989). The average age (n=5) is close to the 490 ka age for isotope stage 13 based on oxygen isotope curves given by Prell et al. (1986). The ESR ages of the highest stands on Oahu (+30 m Kaena stand) which includes the WHC reef, at an elevation of 26 m, are correlated to isotope stage 13. Based on studies of uplifted reef tracts from Barbados and Indonesia which reconstruct paleo-sea level over the last 700 ka, sea level highstands were within 10 m of present sea level during this entire period (Bender et al., 1979; Pirazzoli et al., 1991). The age and elevation of the 28-30 m reefs at Kaena and WHC imply that Oahu has undergone uplift for the past 500,000 years. The Kahuku reef, as described by Stearns (1935a), underlies the Kaena stand, and I propose that the Kahuku stand represents an earlier shoreline, possibly isotope stage 15.

Past workers have interpreted marine benches and near-shore deposits on Oahu as primarily eustatic in origin (Easton and Ku, 1981). Harold Stearns developed a highly speculative sea level curve for the Hawaiian Islands based purely on this hypothesis (see Figure 5). It is interesting to note that the original sequence of relative ages (but not sea level) given by Stearns in 1935 and the subsequent incorporation of limited radiometric dating (Stearns 1978a) match with the recent data reported here. However, the interpretation of Oahu as a tectonically stable platform is inconsistent with sea level records from stable platforms (Bermuda: see Table 8) and tectonically elevated islands (Barbados: Bender et al., 1979; New Guinea: Chappell and Shackleton, 1986; Suma, Indonesia: Pirazzoli et al., 1991).

If we remove the Olowalu stand from consideration for which there is little evidence on Oahu (Macdonald et al, 1983, p. 421), the remaining sequence of emerged marine deposits on Oahu is as follows: Waimanalo (8.2 m), Waialae (14 m), Laie (21 m), Kaena (29 m) and Kahuku (17 m); a sequence in good agreement with recent chronology based on absolute dates (Figure 31). Unaltered corals for dating from the Laie and Kahuku stands have not been found. Based on all available dates, I have assigned isotope stages to these deposits as follows: Waimanalo is stage 5e, Waialae is stage 7, Laie is stage 9/11, Kaena is stage 13, and Kahuku is stage 15 (see Figure 31).

Ward's correlation and speculation on Oahu shorelines based on K-Ar dates of Honolulu basalts are more or less

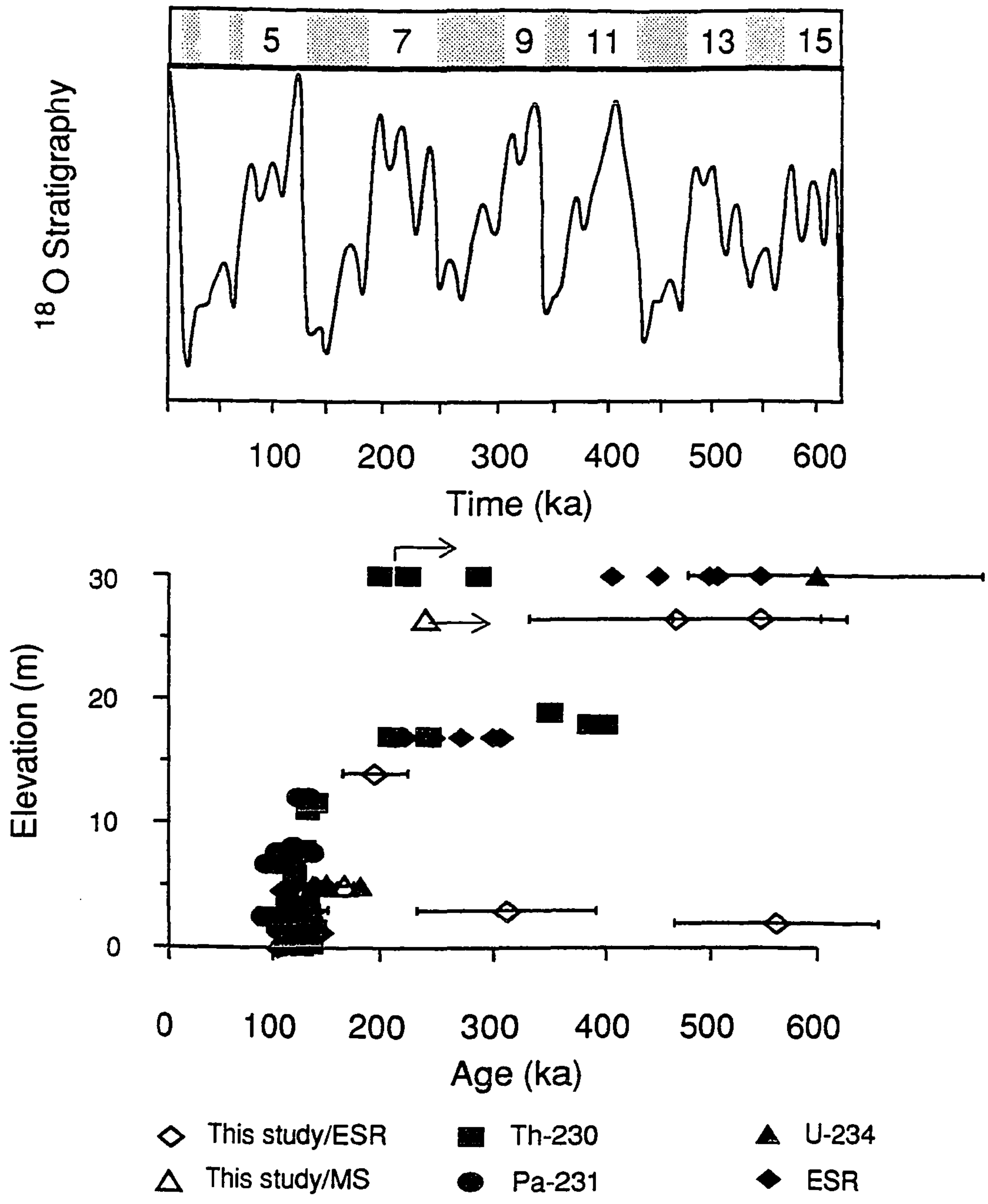


Figure 31. Chronological data from several emerged reef terraces on Oahu with oxygen isotope record from Imbrie et al., 1984.

consistent with the present observations (Ward, 1973). Although, Ward did not recognize the Waialae shoreline and had a younger postulated age for the Kaena stand. The calculated overall rate of uplift is probably closer to 0.06 m kyr^{-1} than to Ward's original estimate of 0.016 m kyr^{-1} (Table 22).

Although Oahu has been considered to be a stable platform in many of the classic studies of the last interglacial, the evidence provided here strongly suggests that within the last 0.5 Myr, the island has been uplifted at an average rate of

Table 22.
Rates of Uplift of Pleistocene reefs on Oahu.

Period (ka)	Paleo-sea Level (m)	Present Elevation (m)	Uplift Rate (m/ka)
0-125	6	8.2	0.02
125-200	0	14	0.06
200-500	0	30	0.06

about 0.06 m kyr^{-1}. This uplift pattern may explain the presence of elevated limestones on other islands farther northwest in the chain such as the +30 m reef described by Hinds (1930) from Niihau and the coastal Mana plain of Kauai (Macdonald et al., 1983).

Flexure as Uplift Mechanism

A hypothesized mechanism to account for vertical uplift of Oahu is the deformation of the lithosphere as a consequence of the applied load at the hot spot. Watts and ten Brink (1989) in conducting a study of the crustal structure of the Hawaiian Islands state that "beyond the area of subsidence there is a broad flexural bulge which reached its maximum uplift on Oahu". They further point out that "the passage of Oahu over Hawaii's flexural bulge has caused uplift". In order to demonstrate a general agreement with the hypothesis, the amplitude and wavelength of the flexure are predicted from theory and the available data on topography, gravity and seismic reflection surveys. If the uplift is associated with lithospheric flexure, then using simple models of flexure we can compare the magnitude of the bulge amplitude and the wavelength of the deformed surface with the observed elevation of elevated reefs and the distance from the hot spot. Two models have been used to describe the isostatic effect of applied loads to an elastic plate overlying a weak fluid substratum. The continuous plate model, first proposed by Vening Meinesz (1941) and Gunn (1943), considers the lithosphere as a continuous elastic plate. The second model, proposed by Walcott (1970a), considers a fractured lithosphere beneath the loading volcano, the so-called broken plate model (Figure 32).

Using these two-dimensional elastic plate models, where the load is applied as a point load at $x = 0$, as shown in Figure 32, the height of the forebulge can be determined (Turcotte and Schubert, 1982). For a continuous plate, the

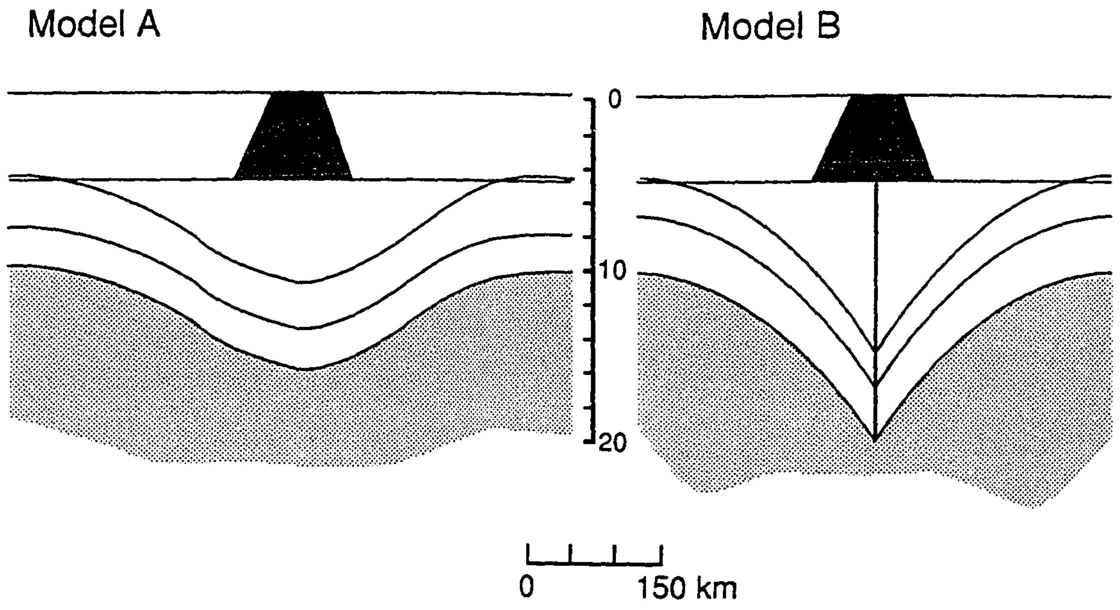


Figure 32. Diagrammetic cross-section of two simple models for flexure of the lithosphere due to the load of a seamount or island: (A) continuous Plate model, and (B) broken Plate model (redrawn from Watts and Cochran, 1978).

deflection surface, w, is given by

$$w = w_0 \, e^{-\alpha x} (\cos \alpha x + \sin \alpha x) \qquad (7.1)$$

where w_0 is the maximum amplitude of the deflection at x = 0 and α is the flexural wave number (defined below)(modified from Turcotte and Schubert, 1982). The height of the forebulge, w_b, is obtained by evaluating the above equation when the slope is zero such that

$$w_b = -w_0 \, e^{-\pi} = -0.0432 \, w_0 \qquad (7.2)$$

Therefore, the amplitude of the forebulge is approximately 4% of the maximum depressed lithosphere at the point source.

For a broken plate, the deflection surface as a function of w_0 is given by

$$w = w_0 \, e^{-\alpha x} \cos \alpha x \qquad (7.3)$$

The amplitude of the forebulge can be expressed as

$$w_b = -w_0 \, e^{-3\pi/4} \cos \frac{3\pi}{4} = -0.0670 \, w_0 \qquad (7.4)$$

Notice that the amplitude of the forebulge for the broken plate model is approximately 7% of the maximum deformation at the loading point (Turcotte and Schubert, 1982). To evaluate the maximum depression of the lithosphere, we can examine the drowned coral reefs off northwest Hawaii that imply subsidence of 1400 m during the last 475 kyr (Ludwig et al., 1991). Thus, a bulge height of 56 - 98 m would be predicted from these calculations. The predicted uplift is greater than the observed uplift of ~30 m for the elevated reefs on Oahu. The results of three-dimensional modeling, however, would dampen considerable the calculated two-dimensional uplift of 56-98 m (P. Wessel, pers. comm. 1992).

Using simple two-dimensional elastic plate models, the distance to the maximum bulge height from the loading point can be calculated knowing the flexural wave number (α). The flexural wave number is defined as follows:

$$\alpha = \left(\frac{(\rho_m - \rho_i)\, g}{4\, D} \right)^{1/4} \qquad (7.5)$$

where ρ_m and ρ_i are the densities of the underlying fluid mantle ($\approx$3.4 g cm^{-3}) and infilling material ($\approx$2.8 g cm^{-3}), respectively, g is acceleration due to gravity (980 cm s^{-2}) and D is the flexural rigidity (units dyne cm). The flexural wave number has dimensions of reciprocal length. Two cases will be discussed, the continuous plate model and the broken plate model. For a continuous plate model (A), it can be shown that the distance from the center of the load to the maximum positive deformation or bulge can be expressed as a function of the flexural number is:

$$x_{b_A} = \frac{\pi}{\alpha} \qquad (7.6)$$

For the broken plate model (B), a similar relations exist to describe the distance from the center of the load to the bulge and can be expressed as:

$$x_{b_B} = \frac{3\ \pi}{4\ \alpha} \qquad (7.7)$$

Therefore, the distance to the bulge for a broken plate is closer to the loading (Further discussion of the full development of these derivations is outlined in Watts and Cochran, 1974). Walcott (1970a) has suggested a flexural wave number of 1/125 km. Using this figure the resulting distance to the bulge would be 390 km for X_{bA} and 290 km for X_{bB}. Oahu is approximately 340 km from Kilauea, the active Hawaiian hot spot. This distance is midway between the bulge maxima for a continuous plate and a broken plate model.

An alternative, but less likely, hypothesis for the observed uplift is isostatic adjustment owing to large-scale erosion or catastrophic landslides. Recent GLORIA images off the north coast of Oahu suggest an enormous landslide (Moore et al., 1989). Preliminary two-dimensional model calculations indicate an isostatic rebound on the order of +30 m for a seamount the size of a typical Hawaiian volcano loosing 8% of its' mass by a submarine slide (J. Smith, pers. comm., 1992).

However, the likely age of the Nuuanu debris avalanche off north Oahu is older than 1.4 ± 0.2 Ma (Moore et al., 1989), an age that greatly precedes the uplift described above.

In summary, from the new data reported here, Oahu appears to have undergone uplift for at least 500,000 years at a rate of ~0.06 mm yr^{-1} for ~400 ka and 0.02 mm yr^{-1} for the last 100 ka. The history of the island's vertical motion can be divided into a series of stages. The sequence of stages is envisioned as follows: once the shield building phase is complete, the island subsides. As the island migrates away from the hot spot, the rate of subsidence is reduced. Subsidence is well documented in cores from the deep boreholes on Oahu such as on the Ewa plain and at Waimanalo (Stearns and Chamberlain, 1967, Resig, 1969, Lum and Stearns, 1970). The flexing of the lithosphere caused by loading at the hot spot results in islands between 270 km and 350 km from the hot spot to undergo uplift reaching a peak rate at the node of the bulge and then decreasing to the northwest. The uplift leads to a series of raised reef tracts while the island passes over the bulge. The islands then return to a subsidence mode controlled largely by lithospheric cooling after a distance of about 500 km from the hot spot is reached.

Summary

ESR dates and MS-U-series dates from new limestone exposures and re-interpretation of previous research are presented which support a tectonic origin for the elevated coastal plains of Oahu. Dates for several elevated marine reefs on Oahu are interpreted as the result of a combination of vertical uplift of Oahu during the past 500 kyr and eustatic sea level fluctuations over this period. The last interglacial appears to be represented by two high stands. The difference in paleoelevation between these two events is speculated to be ~1.5 m. A shoreline at 14 m is assigned an age of 200 ± 20 ka. The highest stand, the 30-m Kaena stand, is assigned an age of 490 ± 40 ka. Sometime prior to 490 ka (pre-Kaena), the island's vertical motion switched modes from subsidence to uplift. The magnitude of the observed uplift is one half to a third of the predicted height determined from simple geophysical models for lithospheric flexure (Turcotte and Schubert, 1982). The distance of Oahu from the hot spot is intermediate between the calculated distance of two models with lithospheric flexure, the continuous plate model and the broken plate model. The data and interpretations are consistent with the hypothesis of uplift associated with lithospheric flexure as opposed to isostasy caused by undersea landslides or tsunami deposits. Therefore, the classical view of Oahu as a stable platform for studies of Quaternary sea level is no longer tenable.

CHAPTER VII

ESR DATING OF PLEISTOCENE CORALS FROM MOLOKAI, HAWAII

> "At Molokai, an island a few miles north-west of Maui, Mr. B. Munn, teacher for the Mission, assured me that he had seen masses of coral apparently in their original position, imbedded in calcareous rocks, one hundred and even one hundred and fifty feet above sea level. I suspect, however, that here is some error, either of calculation or observation, having seen nothing on any of the other islands to warrant the belief in such an elevation as this would indicate. Still from testimony of all the missionaries, there can be no question of the fact that there are really in Molokai raised coral beaches of height at least equal to those of Oahu and Kauai" (Couthouy, 1844, p. 150).

Introduction

Marine fossils deposited at anomalously high elevations on the island of Molokai were first reported by Couthouy in 1844. Harold Stearns (1935a) interpreted the deposits as high stands of the sea resulting from purely glacio-eustatic fluctuations during the Pleistocene. He correlated the elevated deposits on Molokai with deposits from other Hawaiian Islands based strictly on their height above sea level. Later, Stearns (1978a) proposed a chronology for these and other shorelines and terraces in the Hawaiian Islands (Figure 5). Moore and Moore (1984) tentatively correlated the high-elevation marine deposits on Molokai with deposits on Lanai and attributed them to a giant wave event, dated 105 ka (Moore and Moore, 1988). However, past attempts to date Pleistocene samples from Molokai have met with problems caused by diagenetic alteration (T.L. Ku, pers. comm., 1991). The objective of this study was to apply a recently developed technique, electron spin resonance (ESR) dating, to date diagenetically unaltered corals from several marine limestones on Molokai.

Geologic Setting

The island of Molokai is located between the islands of Maui and Oahu in the Hawaiian Island chain, 260 km northwest of the presently active hot spot Kilauea volcano (Figure 33). Molokai is composed of two coalescing composite volcanoes; the Mauna Loa or West Molokai feature is 1.9 Ma and the younger East Molokai volcano is 1.75 Ma (Naughton et al., 1980; Table 23). A post-erosional addition of a small shield volcano (leading to the formation of a peninsula at Kalaupapa) was formed 340-570 ka (Clague et al., 1982). This was about one million years after the conclusion of the principal shield-building activity on East Molokai. A detailed account of the geology of the island is provided in a monograph by Stearns and Macdonald (1947) along with a geological map of the island. The map delineates patches of "calcareous marine deposits, consolidated reef limestone and breccia and poorly consolidated calcareous sandstone" (Stearns and Macdonald, 1947, plate 1, legend). A paleontological study of Molokai marine material was completed by Ostergaard (1939), who, at the request of Harold Stearns, examined mollusks from nine

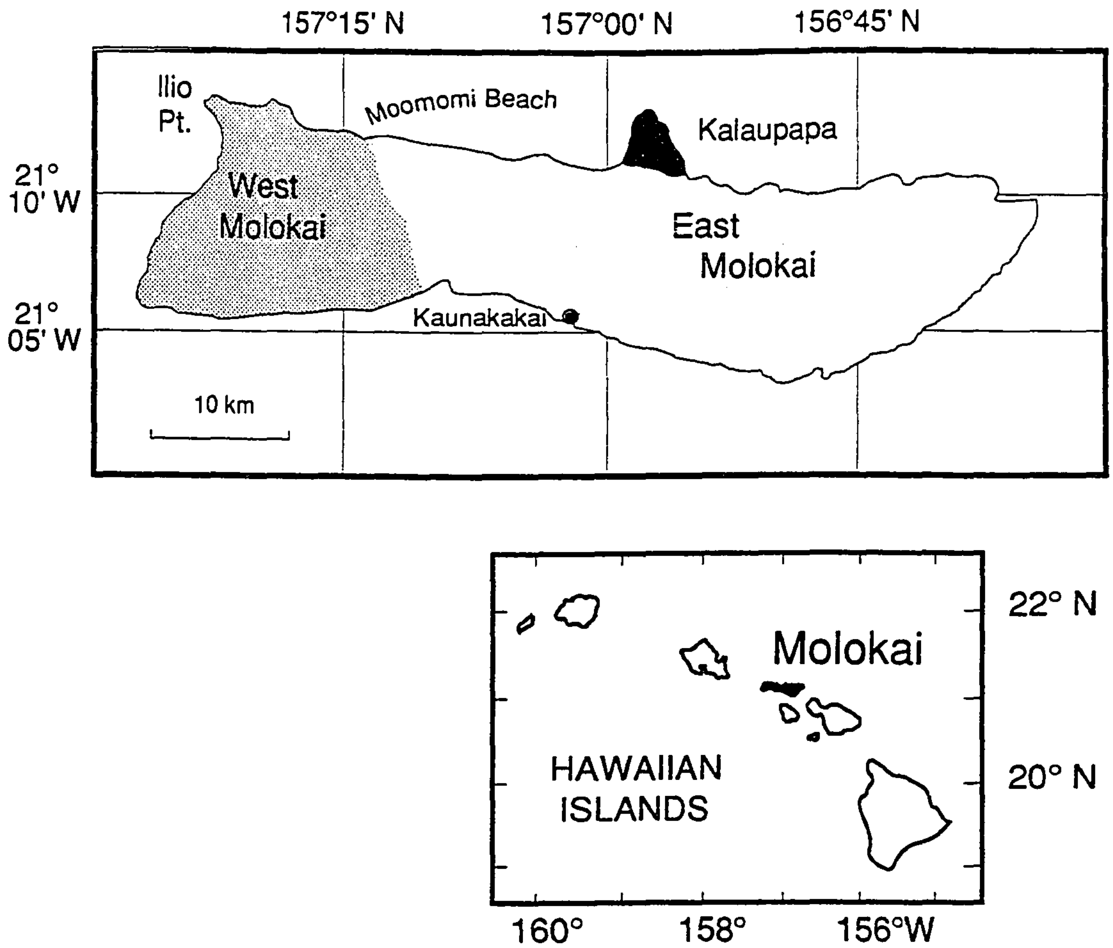


Figure 33. Map of the Island of Molokai in the Hawaiian Archipelago showing the three principal volcanic features West Molokai, East Molokai and Kalaupapa.

Table 23.
Estimates of age of volcanoes on Molokai, Hawaii.

Volcano	Age (Ma)	Reference
West Molokai	1.89	McDougall, 1964
	1.52-1.84	Naughton et al., 1980
East Molokai	1.51-1.53	McDougall, 1964
	1.75	Naughton et al., 1980
Kalaupapa	0.344-0.570	Clague et al., 1982

horizons ranging in elevation from 3 to 59 m. Based on the fairly homogeneous assemblage of mollusk species between the horizons, Ostergaard concluded that the horizons represented a rapid change in relative sea level with the highest formation laid down first. The other units were deposited progressively as the sea retreated. He suggested a late Pleistocene age for the deposits (Ostergaard, 1939).

Distribution of Emerged Marine Fossils

Marine carbonate deposits are exposed on the southern slope of East Molokai (Figure 34). Lindgren reported that "small amounts of coral rock, indicating a former higher water level, are found all along the southern coast" reaching an elevation of +7.6 m with one fossil coral reef extending to +40 m near Pu'u Maninikolo (Lindgren, 1903, p. 15). Stearns (1935a) reported coral fragments imbedded in limestone conglomerate in a gulch east of Pu'u Maninikolo at +58 m. Stearns and Macdonald (1947) correlated terraces and marine deposits on East Molokai at an elevation of approximately

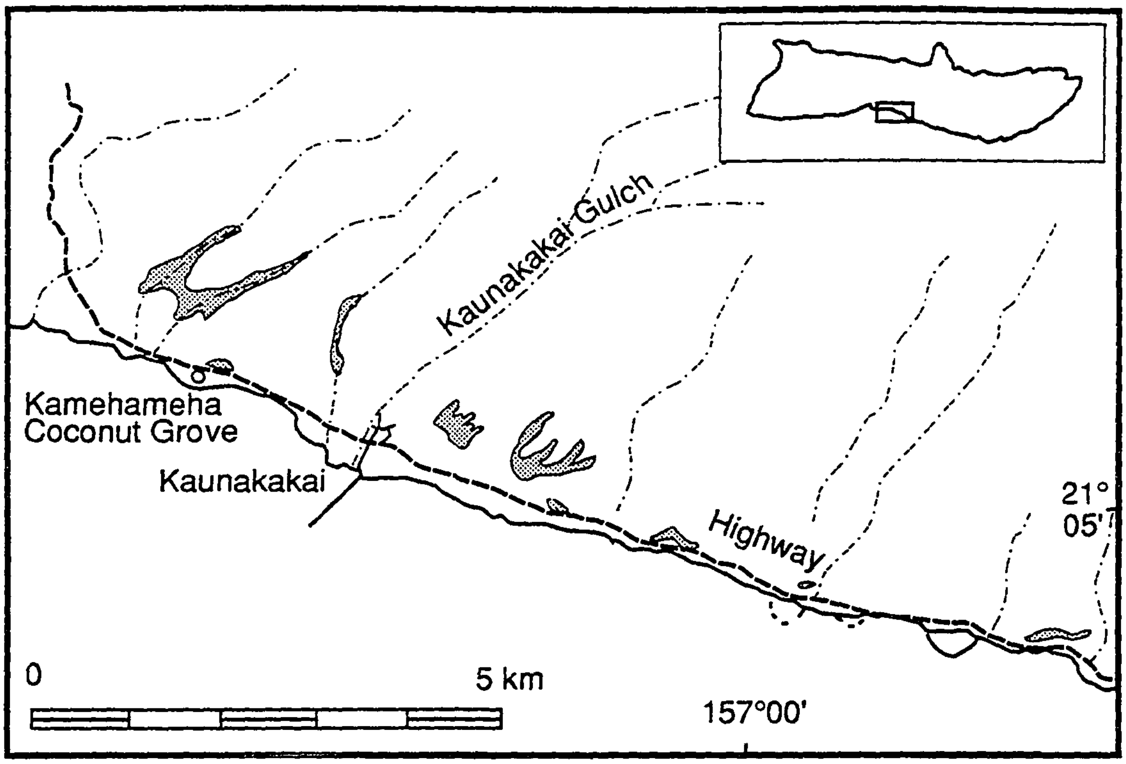


Figure 34. Stearn's mapped distribution (shaded) of emerged marine fossiliferous conglomerate along the southern coast of Molokai.

+7.6 m with the Waimanalo stand on Oahu (Table 24). Stearns and Macdonald (1947) also correlated wave-cut platforms and marine fossiliferous conglomerate at an elevation of approximately +30 m on Molokai with the +30 m Kaena stand on Oahu. Finally, a fossiliferous marine conglomerate at an elevation of about +76 m was correlated by Stearns and Macdonald (1947) with the Olowalu stand on Maui. This

Table 24.
Correlation of Molokai shorelines based strictly on comparison of inter-island elevations (Stearns, 1935a, 1978a).

Elevation (m)	Stand	Estimated Age (ka)
+ 7.6	Waimanalo, Oahu	125
+ 30	Kaena, Oahu	650
+ 58 ?	Olowalu, Maui	ca. 1,150

fossiliferous marine limestone crops out in a series of gulches west of Kaunakakai. However, the larger and steeper gulches, Manawainui Gulch and Kaunakakai Gulch, lack limestones (Figure 35). The highest specific mention of coral fragments was from a site 1.6 km west of Kaunakakai Gulch at an elevation of about +85 m (Stearns and Macdonald, 1947, p. 14).

Lum (1972) reported the occurrence of marine fossiliferous conglomerate and wave-cut caves at +63 m in a gully 0.8 km east of Manawainui Gulch and interpreted these features as evidence of a relic shoreline. Later, Stearns

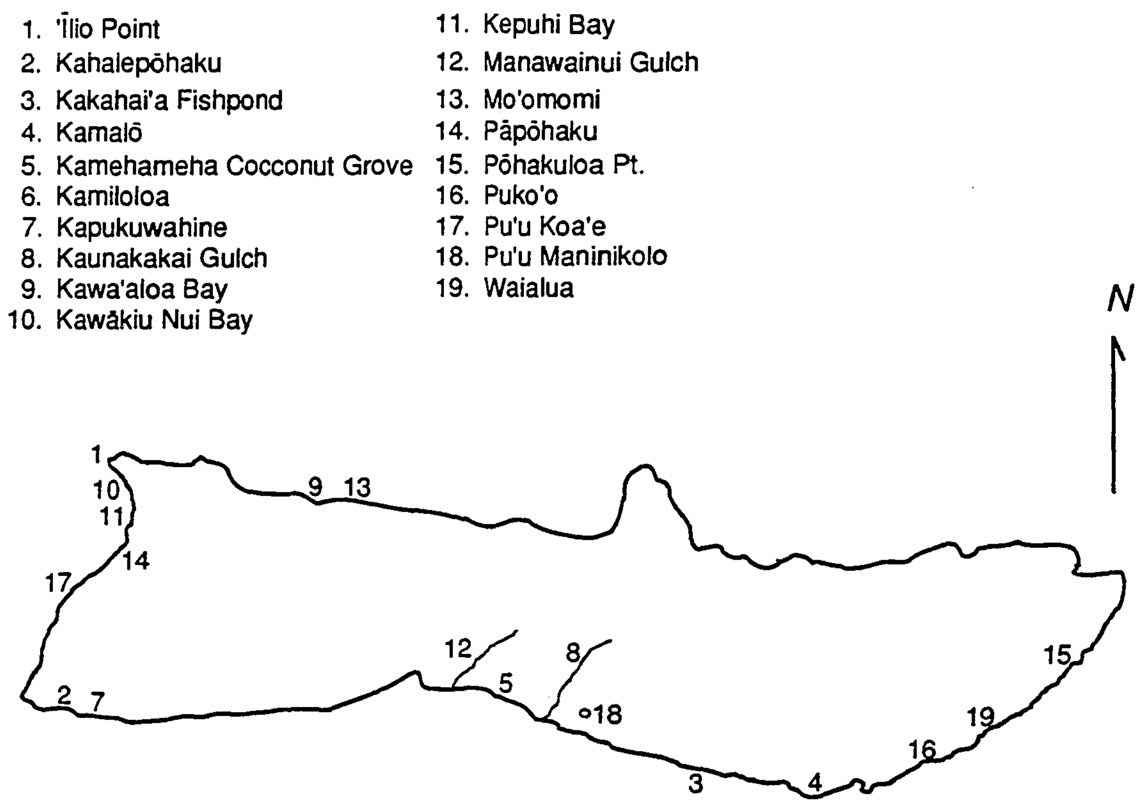


Figure 35. Index map of localities on Molokai mentioned in text.

(1978a) presented a geologic section of this exposure and interpreted the wave-cut caves as fluvial features. Despite this interpretation of these features as erosional, Stearns concluded that the sea had been to this level in the past. Coral samples collected by Stearns from this site were too altered for U-series dating (Stearns, 1978a). This +63 m shoreline was later incorporated into Stearns' road guide to points of geological interest, although the site is farther from the road than indicated by his guidebook (Stearns, 1978b).

Sample Localities

Sample localities containing marine fossils were identified using Stearns' map (Stearns, 1947) and other published literature. Marine fossils, especially scleractinian corals, were collected in the field. Site locations are given in Figure 35. Site elevations were determined with an altimeter ($\pm$ 1 m)(American Paulin System, Micro Surveying Altimeter, Model M-1), and agree with elevations based on USGS topographic sheets (1:25,000 scale; USGS, 1983) and previous reported values (Stearns, 1935a; Stearns and Macdonald, 1947; Ostergaard, 1939; Lum, 1972; Stearns, 1978a).

Results

Description of Terraces

Three geomorphic terraces were identified in the field. The geographical distribution of the terraces identified in aerial photographs and verified in the field is mapped in Figure 36. A topographic profile, along a leveled transect normal to the coast near Kamehameha Coconut Grove, is presented in Figure 37.

A low altitude terrace at 1.8 m, herein named Terrace I, is best developed as a coastal plain west of Kaunakakai and extends along major sections of the southern coast of Molokai including Kamalō, Pulo'o and Waialua. Several geomorphic features are related to Terrace I. These include: wave-abraded notches carved into basalt and eolianite at approximately 2-3 m elevation, coastal dunes (both lithified and unlithified), emerged wave-abrasion platforms with characteristic erosional features such as sea stacks, sea caves, emerged intertidal bioerosion structures and exposed beachrock (Table 25, Figures 38, 39). A sketch of the wave abrasion platform at Pōhakaloa Point is shown in Figure 40.

A second terrace (II) at 10.5 m elevation is apparent as a break in slope in the topographic profile at Kamehameha Coconut Grove (Figure 37). This second terrace is fairly narrow and extends from Kamehameha Coconut Grove to immediately west of the cinder cone at Pu'u Maninikolo. Evidence for terracing can also be seen east of Kaunakakai at Kamiloloa Heights where the road traverses the terraces. The existence of a second terrace is further supported by a planation surface truncating the bedding plane of eolianite at Pu'u Koa'e on the west shore of Molokai (Figure 41) and

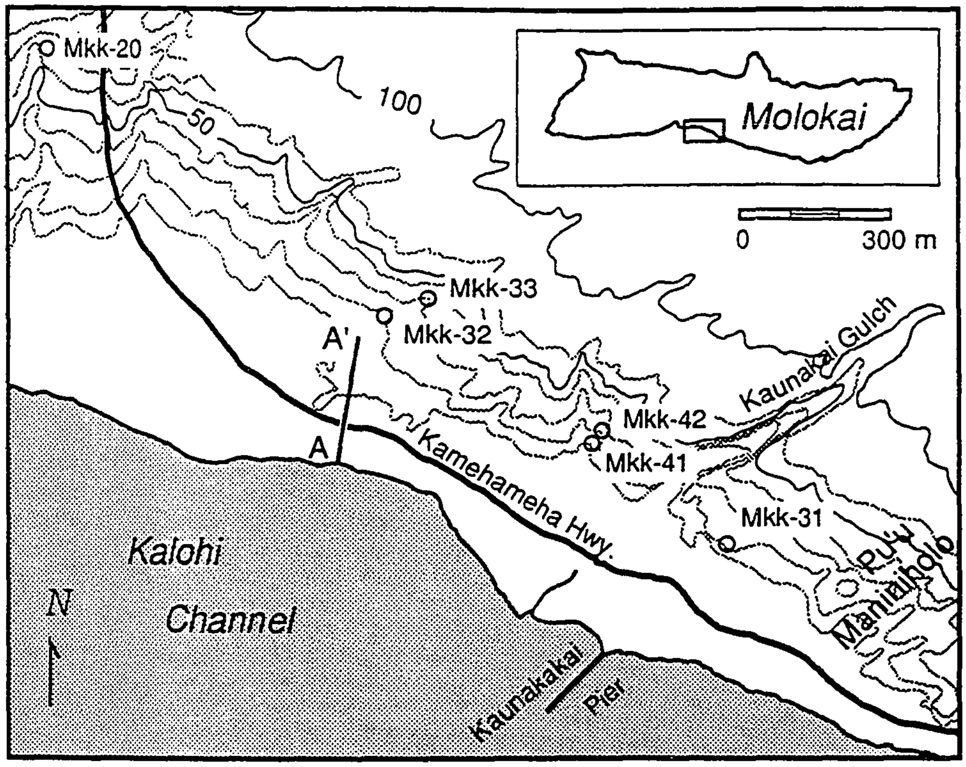


Figure 36. Geographic distribution of marine terraces on the southern shore of Molokai. Topographic profile, A-A', near Kamehameha Coconut Grove, is presented in Figure 37.

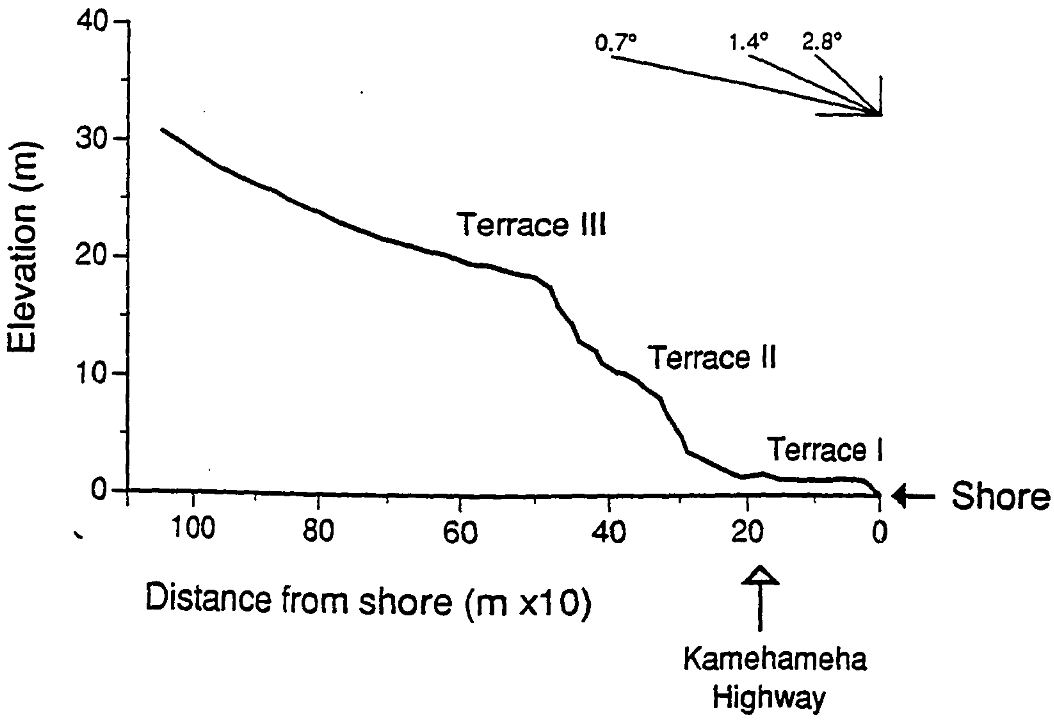


Figure 37. Topographic profile near Kamehameha Coconut Grove, Molokai.

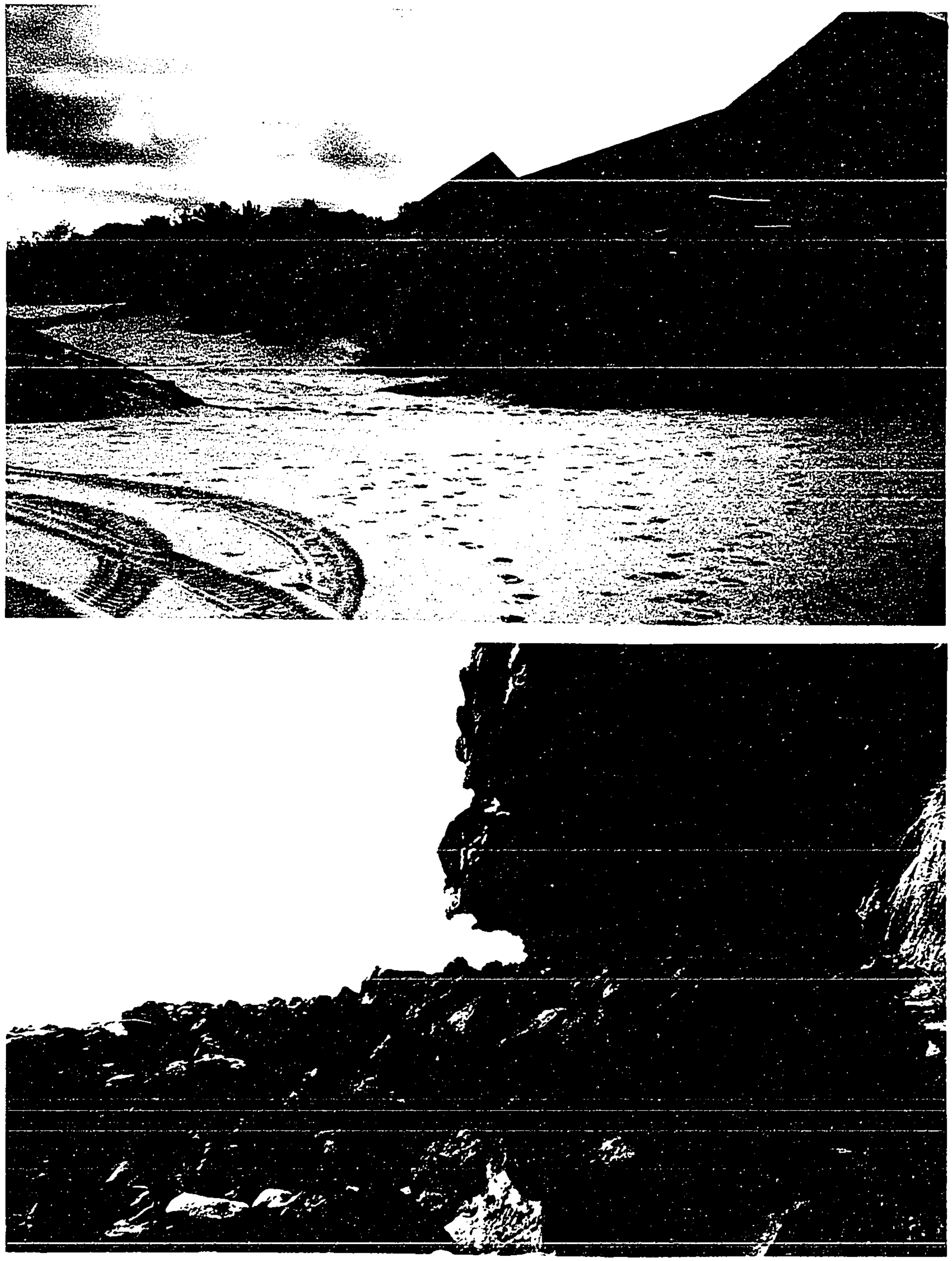


Figure 38. A. A marine notch in eolianite at Pāpōhaku Beach, Molokai. Elevation of notch is 3 m. B. A marine notch in basalt along the southwest coast of Molokai near Kapukuwahine. Elevation is ≈3 m.

Table 25.
Localities with supportive geomorphic evidence for Terrace I highstand on Molokai.

Geomorphic Feature	Locality
Coastal plain	Kamehameha Coconut Grove Kamalō Puloʻo Wailalua
Coastal dunes	Pāpōhaku Beach Moʻomomi Beach Kamehameha Coconut Grove Kamiloloa
Notches:	
Basalt	Kapukuwahine Kahalepōhaku N. Kawākiu Nui Bay Puʻu Koaʻe Moʻomomi Beach
Eolianite	Pāpōhaku Beach
Wave-abrasion platforms	Kahalepōhaku Pōhakuloa Pt.
Emerged sea stacks	Kapukuwahine Pōhakuloa Pt.
Emerged sea caves	Kawakiunui Bay
Emergent beachrock	Kepuhi Bay

notches in basalt observed at Pōhakuloa Point and on headlands at Kapukuwahine Beach and Kawākiu Nui Bay. Terrace II may represent a highstand during the last interglacial period. However, the elevation is higher than +6 m, as expected from other studies of the last interglacial on stable platforms (see Table 20).

A third terrace above 20 m is shown in Figure 37. The terrace has a slope of 1° significantly less than the typical Hawaiian Shield volcano slopes of 3-6° for Mauna Kea and Mauna Loa (Mark and Moore, 1987). A coral-bearing boulder field

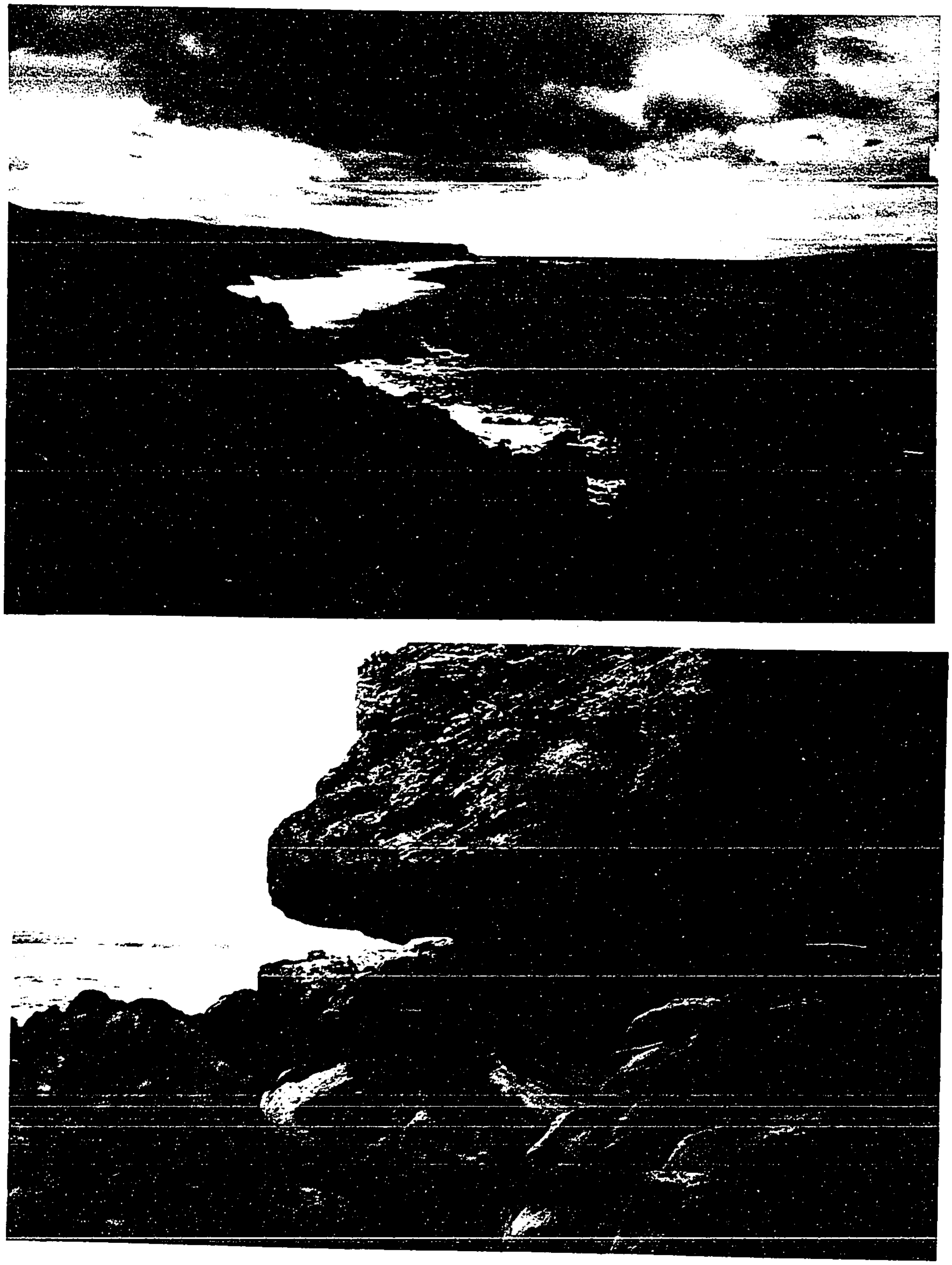


Figure 39. A. Emergent wave-abraded platform with sea stacks along the southwest coast of Molokai near Kahalepōhaku. B. A marine notch cutting eolianite at Pu'u Koa'e exposing basement lava flows.

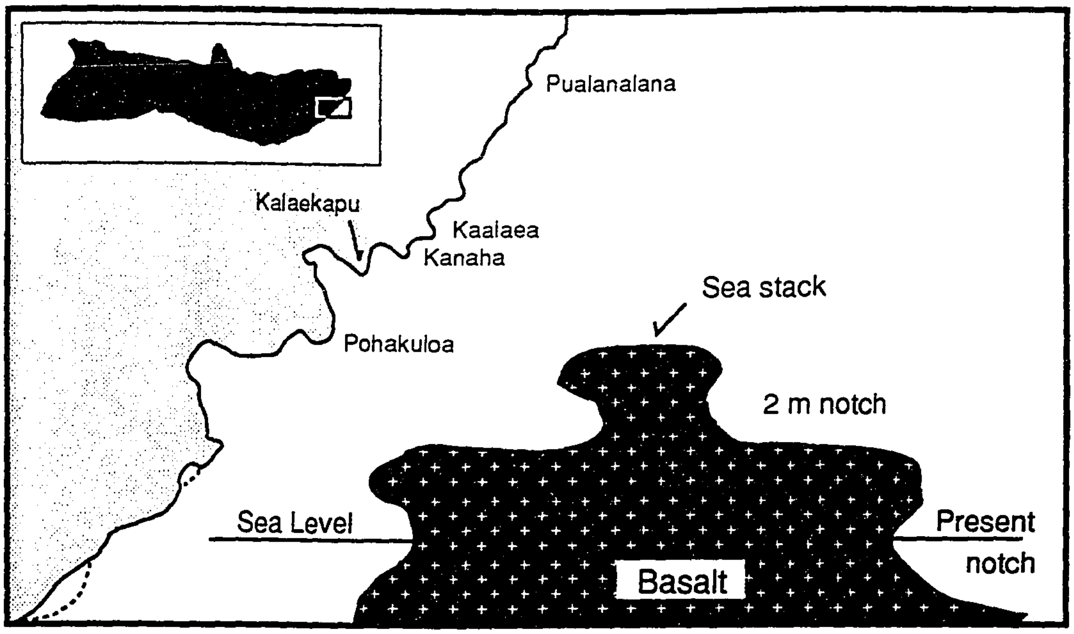


Figure 40. Sketch of wave-abrasion platform at Pōhakuloa Point, Molokai with several geomorphic features related to Terrace I.

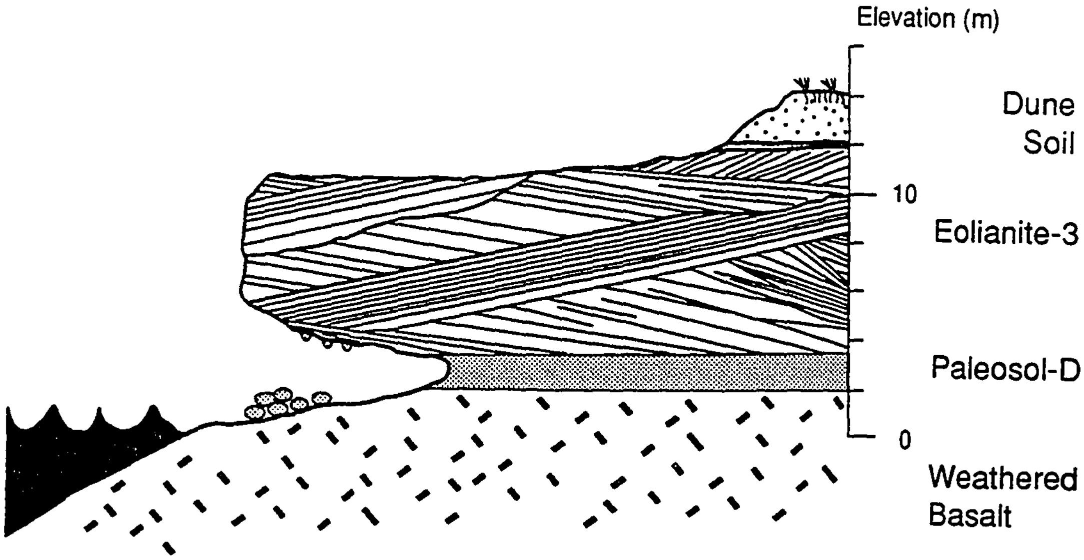


Figure 41. Sketch of cross section at Pu'u Koa'e, Molokai indicating two sea level events, one at 3 m and one at 10.5 m.

lies upon this terraces from which corals were sampled. Fossil corals from the upper terrace were carefully collected and then dated by ESR dating method.

Table 26.
Results of ESR dating of corals from Molokai.

Elev. (m)	Sample No.	U (ppm)	AD (krad)	ESR Age (ka)
13	Mkk-41	2.08	13.5 ± 1.0	163 ± 14
14	Mkk-31	2.65	28.8 ± 1.4	258 ± 17
	Mkk-31a	2.89	27.7 ± 4.4	237 ± 40
	Mkk-31b	2.32	27.8 ± 3.0	272 ± 32
	Mkk-31c	2.67	27.1 ± 2.1	245 ± 21
	Mkk-tw	2.56	27.4 ± 0.7	253 ± 12
15	Mkk-32a	2.28	22.5 ± 1.5	232 ± 18
	Mkk-32b	2.41	24.6 ± 1.5	241 ± 18
18	Mkk-421	2.46	17.0 ± 2.1	177 ± 22
30	Mkk-33b	2.38	30.6 ± 3.0	290 ± 31
58	Mkk-20	2.68	35.3 ± 2.5	303 ± 25

ESR Dating

The ESR results are listed in Table 26. Neutron activation analysis for uranium concentration in the samples showed an average value of 2.49 ± 0.23 ppm. This value is consistent with published values for fossil corals (<200 ka; Veeh and Burnett, 1982) and does not differ substantially from U content of modern corals (Swart and Hubbard, 1982; Radtke et al., 1988; Hamelin et al., 1991)(Figure 42). Note that sample Mkk-41 has a low U concentration. Recalculation of this

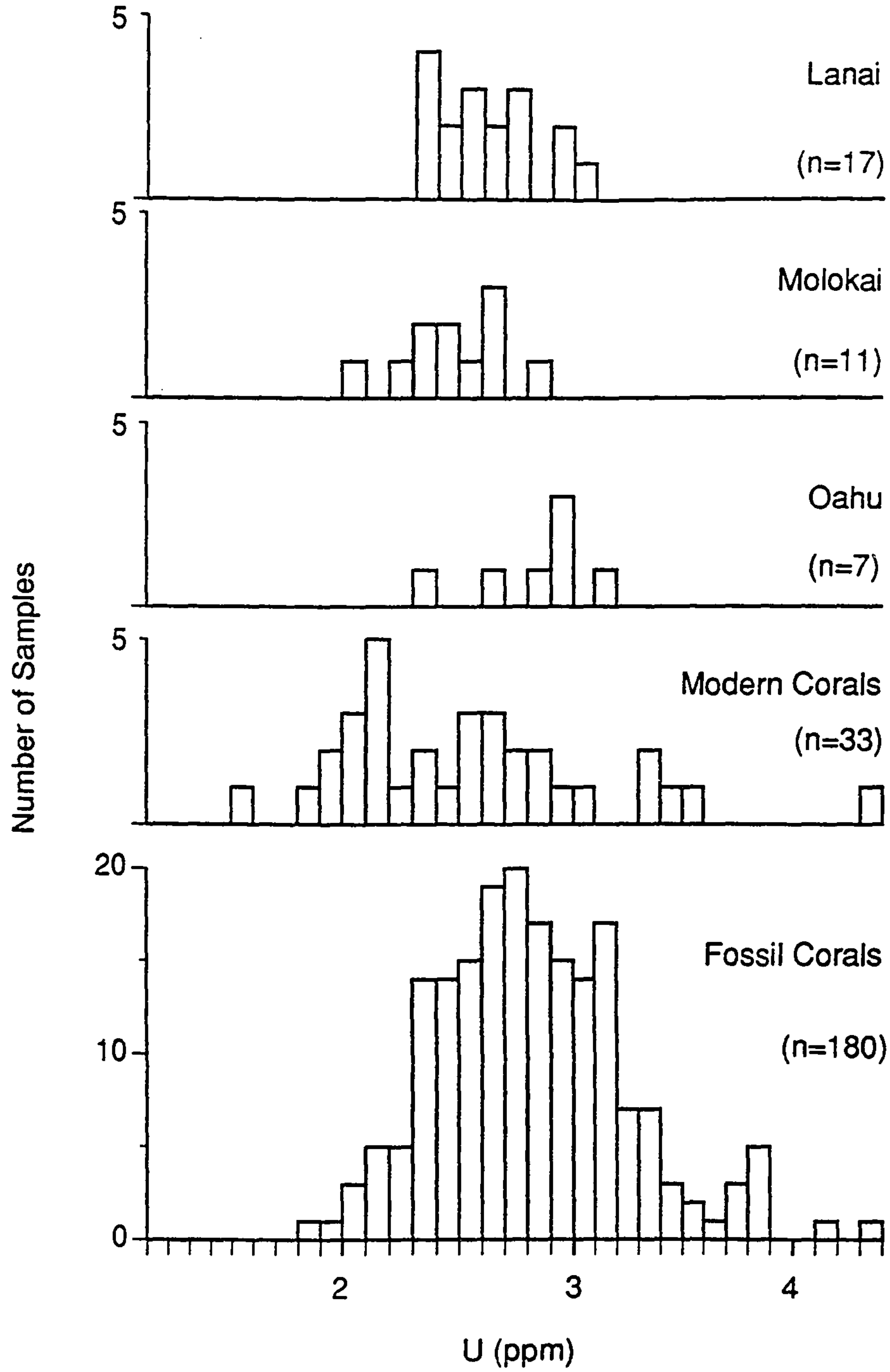


Figure 42. Histogram of U content in fossil corals from Lanai, Molokai and Oahu compared with modern and fossil corals from Veeh and Burnett (1982).

sample using the average U concentration of Molokai samples gives an ESR age of 146 ± 14 ka.

The ages associated with the marine terraces can be related to warm stages of the deep-sea oxygen isotope record (Imbrie et al., 1984) as illustrated in Figure 43. Samples collected from 13 to 18 m elevation appear to correlate with the Stage 7 interglacial, while the two highest samples may correlate with Stage 9.

The ages must be regarded with some caution, because they represent only a few analyses. However, when splits of more than one sample were analyzed, there was generally good reproducibility of results. For example, sample Mkk-31 ages average 253 ± 15 ka, well within the estimated 10-15% uncertainty for ESR coral dating (Radtke and Grün, 1988). Until the ESR method is further refined, the method can not be used to resolve substages of the major interglacials.

Interpretation and Discussion

Terrace I: Holocene

The numerous geomorphic features related to Terrace I listed above (see Table 25) are supported by ^{14}C dates of coral (*Porites*) obtained from drill holes at Kakahai'a fishpond (3370 ± 70 yr B.P.; Weisler, 1989) and at Kamehameha Coconut Grove (3290 ± 70 yr B.P.; C. Fletcher, unpubl. data). Additionally, a low-elevation eolianite inland from Kamehameha Coconut Grove was dated as mid-Holocene (5460 ± 80 yr B.P.; C. Fletcher, unpubl. data). As noted by Weisler, this shoreline

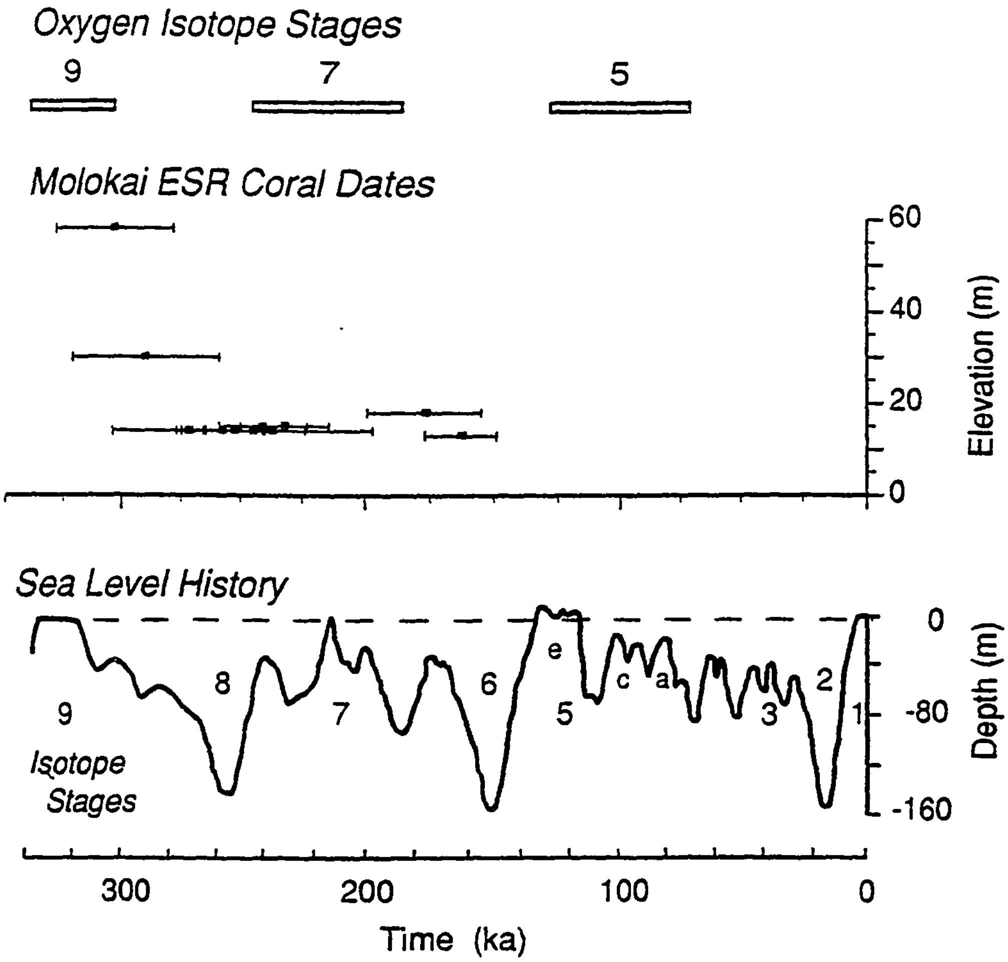


Figure 43. Correlation of coral ESR age from Molokai with oxygen isotope stages (after Imbrie et al., 1984) and sea level history (Chappell and Shackleton, 1986).

corresponds to the 1.5 m Kapapa Stand on Oahu which has recently been dated at ~3,400 yr B.P. (Fletcher and Jones, 1992). Consequently, the very broad low-elevation coastal terrace located at an elevation of ~2 m along the southern coast of Molokai is proposed to represents a mid-Holocene relative highstand.

Terrace II: Last Interglacial

Extensive accumulations of marine carbonates attributable to the last interglacial, such as the well-documented Waimanalo Formation on Oahu (Ku et al., 1974), have not been found on Molokai. However, the elevation of Terrace II at Kamehameha Coconut Grove and the 10.5 wave-cut terrace at Pu'u Koa'e suggest a highstand at this level. Although erosional surfaces are not readily dated, the most plausible scenario is that the 10.5 m terrace represents the last interglacial when eustatic sea level was about 6 m above present (Harmon et al., 1983; Radtke and Grün, 1990; Pirazzoli et al., 1991; Hearty et al., 1992). If so, then this field evidence suggests an uplift of 4.5 m for Molokai since the last interglacial. The uplift rate can be calculated as:

$$\frac{10.5\ m - 6\ m}{125\ kyr} = 0.04\ mm\ yr^{-1}$$

A possible mechanism for this uplift is discussed below.

Terrace III

ESR dates of Terrace III corals at 30 m elevation suggest an age of approximately 290 $\pm$ 30 ka, which corresponds to

Stage 9, an interglacial highstand. The uniform nature of this terrace surface, its consistently flat slope, and the sharp change in slope at both the upslope and downslope edge strongly suggest an origin as an erosional surface constructed during a prolonged period of exposure to wave abrasion. If the uplift rate for Molokai during the last 300 ka were similar to that computed for the last 125 ka (0.04 m/ka), then the Stage 7 shoreline with a paleo-elevation of ≈0 m (Radtke, 1987) would have been uplifted to +3 m by the time of the transgression of the last interglacial to +6 m. This transgression would have overtaken the raised Stage 7 shoreline.

A conglomerate unit of coral clasts and subangular basalt boulders overlies Terrace III. Along natural exposures in gullies on the southern slope of East Molokai, the conglomerate unit shows no internal features typical of a reef deposit laid down at sea level, such as in situ corals in a reef framework, secondary encrustation by coralline algae, or evidence of biological or physical erosional processes and marine cementation. Based on the ESR age of the corals from this conglomerate, this unit may not have been deposited by the 105 ka Lanai giant wave as argued by Moore and Moore (1988), but instead may have been deposited by an earlier wave event occurring sometime during early part of Stage 7 (186 - 245 ka). The earlier event may have had an estimated runup of height of 10-20 m, based on the estimate of paleo-sea level at about present level. Further study of this terrace and its deposits are needed before more definitive conclusions can be reached. In Table 27, Terraces I, II and III roughly correlate with deep-sea oxygen isotope Stages 1, 5 and 9.

Table 27.
Correlation of Molokai terraces with oxygen isotope stages†.

Stage 1 (<12 ka)	Stage 5 (71-128 ka)	Stage 9 (303-339 ka)
Terrace I	Terrace II	Terrace III

† Age brackets based on Imbrie et al. (1984).

Uplift Mechanism

The uplift of Molokai may occur as a result of flexure caused by volcanic accretion on the island of Hawaii. The magnitude of uplift needed to account for anomalous paleo-shoreline elevations is consistent with models of bulge heights as determined by geophysical calculations for lithospheric flexure (Turcotte and Schubert, 1982). Using a two-dimensional elastic plate model, the maximum positive deformation (maximum bulge height) above sea level is approximately 4 % of the maximum deflection for a continuous plate model, and almost 7 % of the maximum deflection for a broken plate model (P. Weissel, pers. comm. 1992). Drowned coral reefs in the Alenuihaha Channel suggest that the island of Hawaii has subsided at least 1 km since 340 ka (Jones, unpubl. data). Thus, 40-70 m of total uplift at the maximum height of the bulge could have been produced in the last 340 ka. This amount of uplift agrees approximately with the age and elevation of paleo-shorelines on Molokai.

Molokai is approximately 250 km from Kilauea, the active Hawaiian hot spot. Because the distance between the crest of the Hawaiian Arch and the Hawaiian Ridge is approximately 250 km (Dietz and Menard, 1953; Walcott, 1970), Molokai is on or near the maximum deformation of the lithosphere bulge. Thus, the marine terraces on Molokai are at a distance from the hot spot where maximum uplift is expected, and the magnitude of the observed uplift agrees with geophysical arguments for bulge heights based on the observed subsidence of Hawaii.

An alternative hypothesis for the observed uplift is isostatic adjustment owing to large-scale erosion or catastrophic landslides. Recent GLORIA images off the north coast of Molokai suggest enormous landslides have occurred during the development history of Molokai (Moore et al., 1989). However, the likely age of the Wailau debris avalanche off the north slope of Molokai is 1.4 $\pm$ 0.2 Ma (Moore et al., 1989), an age that greatly precedes the timing of the uplift based on paleo-shoreline evidence.

The interpretation of Molokai uplifting during the last 300 ka contradicts the prevailing view that the island of Molokai is subsiding. Moore (1987) has argued for a subsidence rate of $\leq$0.1 mm yr^{-1} for central Molokai during the last 0.5 My. He assumes that the slope change on the Kalaupapa Peninsula, now 50 m below sea level, represents the level of the sea at the end of volcanism. From the age of the Kalaupapa volcano, Moore calculated a subsidence rate for the island of Molokai.

A timetable of events is presented in Table 28 and is illustrated in Figure 44. The age of the Kalaupapa Volcanics

Table 28.
A timetable for volcanic and paleo-sea level events.

Events	Molokai Features	Time Period* (ka)
	VOLCANISM:	
Shield-Building	West Molokai	1,520-1,840
	East Molokai	1,500-1,530
Rejuvenation	Kalaupapa	340-570
	INITIATION OF UPLIFT	
	PALEO-SEA LEVEL EVENTS:	
Stage 9 Interglacial	Terrace III	303-339
Last Interglacial	Terrace II	120-132
Mid-Holocene	Terrace I	3.4-4.4

* Time periods are estimated as follows: for the Holocene highstand based on radiocarbon data from Kauai (see Chapter VI); for the last interglacial period based on Chen et al. (1991) and for Stage 7 and 9 interglacial periods based on Imbrie et al. (1984).

(340-560 ka) suggests that the secondary eruptive phase on Molokai had ended prior to the uplift observed on the southern shore. The establishment of an approximate chronology provides an opportunity for future research on geological

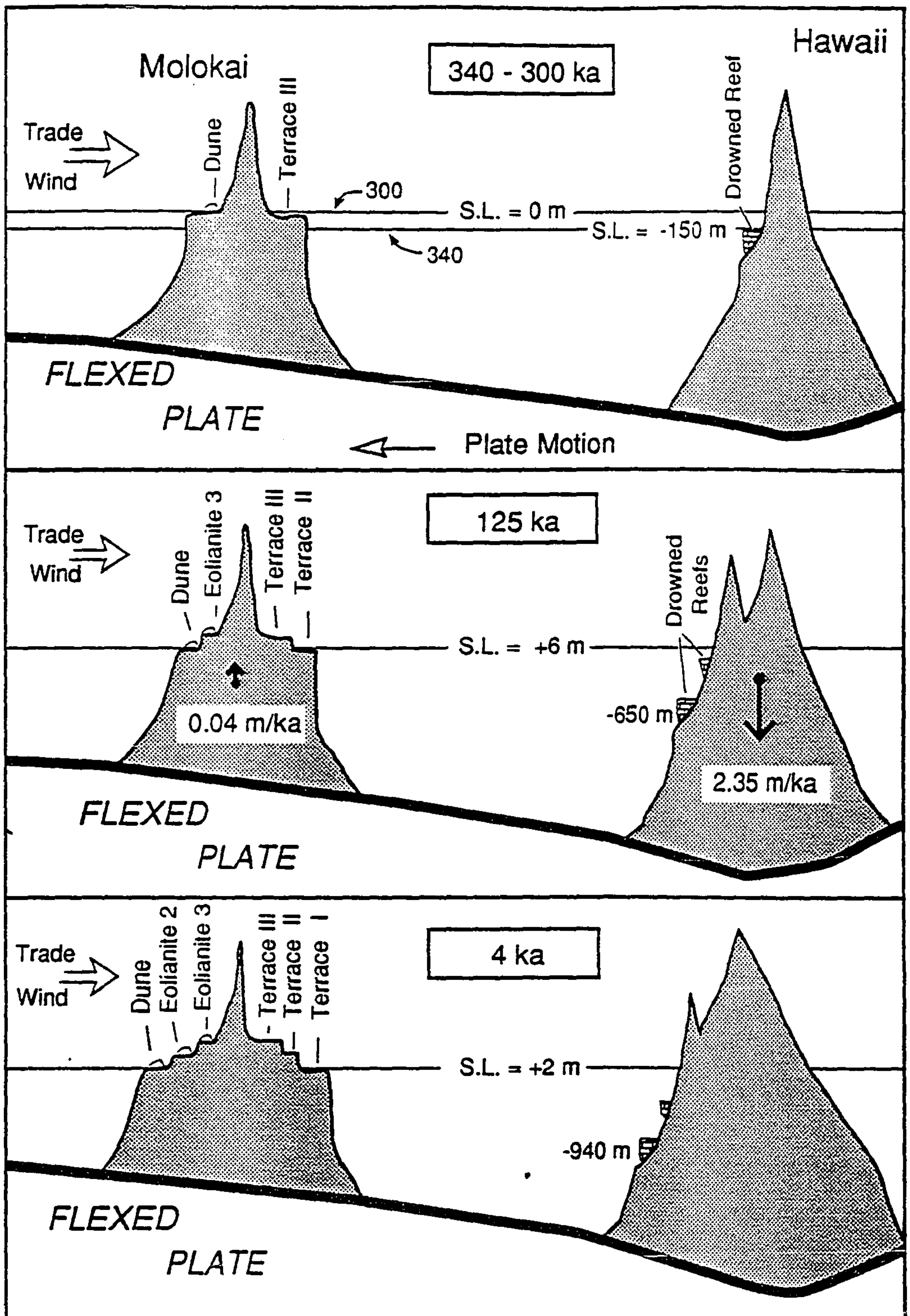


Figure 44. Schematic representation of paléo-sea level events on Molokai relative to hot spot volcanic loading on the island of Hawaii. Drawing not to scale.

processes and features on Molokai. One such feature is the eolian dune fields on the northeast coast.

Relation of Terraces to Eolian Dune Fields

The orientation of the lithified dunes in Hawaii and the region's predominately northeasterly trade winds have been incorrectly used as a basis for paleoclimatic interpretation during glacial periods (see Gavenda, 1992). Several authors have stated that Hawaiian dunes developed during glacial low stands of the sea (Stearns, 1940, 1947; Stearns and Macdonald, 1942, 1947; Macdonald et al., 1960, 1983). The uplifted marine terraces on the south coast of Molokai challenges this idea and clearly furnish evidence important for the interpretation of the dunes on Molokai.

Two prominent calcareous dune fields in the vicinity of Moʻomomi Beach and ʻIlio Point are located on the northwestern coast of Molokai. The dunes near Moʻomomi Beach extend for at least 5 km, parallel to the direction of the predominant northeast wind. The dune field is composed of active and fossil dunes.

A second dune field at ʻIlio Point has been significant in the development of paleo-shorelines on Molokai. The 2.5 km^2 lithified calcareous dune at ʻIlio Point is notched at 1.5 and 7.5 m elevations (Stearns and Macdonald, 1947) and formed prior to the last interglacial high stand. In Volcanoes in the Sea, Macdonald and co-authors suggest that the dunes at Ilio Point were "formed from sand blown inland from beaches exposed to wind at a time of lower sea level" (Macdonald et al., 1983, p. 412).

In order to build up carbonate dunes, a proximate source of material is required. Carbonate production by shallow-water marine organisms during episodes of high sea level, such as the present active dune system at Mo'omomi Beach, contribute skeletal debris for dune deposition. A lowering of sea level is therefore may not be required to produce calcareous sand for dune formation. Indeed, such a lowering could lead to weathering and cementation of potential nearshore sources of calcareous sand (Land et al., 1967). Also during glacial periods when the sea level was on the order of 100 m lower than present, the source region would actually be farther from the present location of the dune field and significantly stronger trade winds would be required to transport dune sand to their present locations several meters above present sea level. Coastal dunes do not generally migrate such distances (Vacher, 1973).

In the Mo'omomi dune field, three eolianite units can be recognized along the western headlands of Kawa'ahoa Bay where each is bracketed by a paleosol. A geological map of the coast at Kawa'ahoa Bay is presented in Figure 45. The eolianite units become progressively more lithified.

The Mo'omomi carbonate eolianites are composed of skeletal fragments of shallow-water marine organisms, predominately foraminifera, pelecypods, gastropods and coral

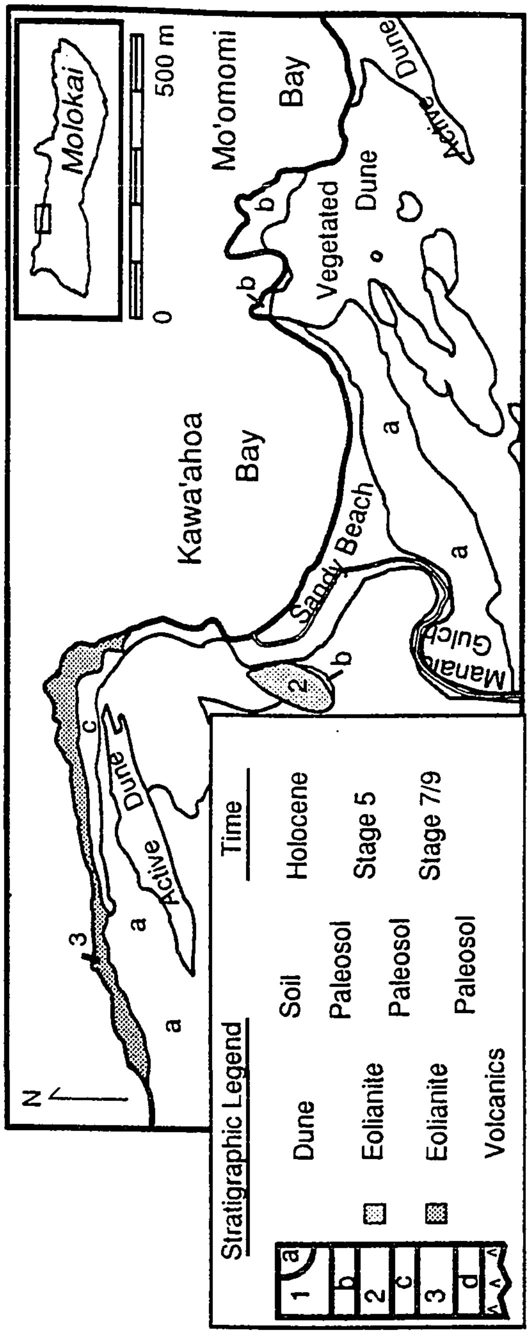


Figure 45. Geological map of dune field near Mo'omomi Beach, Molokai.

fragments: similar in composition to the present beach at Kawa'ahoa Bay and in general to modern Hawaiian beaches (Moberly et al., 1965) with the exception that the modern beach at Kawa'ahoa Bay has a small but significant component of volcanics. Volcanic products are absent from eolianite units 2 and 3.

Because there is evidence for three paleo-shorelines on the south slope of West Molokai, the connection between sand production and dune formation suggests that at least 3 sets of dunes are present in the Mo'omomi dune field. The active unconsolidated dune field is composed of recent and Holocene age sands, while the lithified dunes, units 2 and 3, may represent two of the past interglacial periods (Stages 5, and 7/9). The age relationship of the dunes could be readily confirmed by dating the dune fields. The chronology combined with detailed measurements of the dune geometry would indicate the magnitude and direction of trade winds in the northern central Pacific during interglacial periods. Such an approach has been used by Mackenzie (1964) for Bermuda Pleistocene coastal eolianites. It is proposed that the eolianites on Molokai's northwest shore were formed during periods of high sea level and that the paleosols developed during low sea level (Bretz, 1960; Land et al., 1967; Vacher, 1973).

Summary

Marine terraces in coastal southern Molokai form extensive peneplaned terraces at 20-30, 10.5-12, and 2-3 m above present sea level. Terraces on Molokai at elevations of <3 m and 10.5 m are inferred to be 4 and 125 ka, respectively. These terraces ages can be correlated to interglacial stages derived from oxygen isotope stratigraphy of deep-sea sediments. The formation and elevation of the marine terraces are as a result of both sea-level glacial eustatic fluctuations and uplift of the island during the last 125 kyr. Lithospheric flexure is proposed as the mechanism for uplift. Molokai, approximately 250 km from Kilauea, is near the maximum deformation of the lithospheric bulge (Walcott, 1970). Simple calculations of the amount of uplift indicate 40-70 m of uplift are anticipated as the lithosphere flexes due to volcanic loading of the island of Hawaii.

Dune fields of the northwestern coast of Molokai are most likely related to high sea level events and not to lower sea levels as has been previously thought. Three eolianite units are recognized from Mo'omomi Beach area, and it is postulated that the three units formed during interglacial periods.

CHAPTER VIII

ESR DATING OF RAISED PLEISTOCENE CORALS FROM LANAI, HAWAII

> "In a talus mass at a place northeast of Manele and about 150 feet above sea level shells and coral fragments were found to be so abundant that a natural origin seemed reasonable. Subsequent search in other gulches failed to reveal similar evidence and I have concluded that the deposit must be in part of artificial origin. It is impossible to believe that the sea has stood more than 10 to 15 feet above its present level at any time since Lanai was formed. Had it done so, it seems certain that there would be clear indications at more than one place and of more than one sort." (Wentworth, 1925, p. 33).

Introduction

Marine fossils deposited at peculiarly high elevations on the island of Lanai were first reported by Wentworth in 1925. Harold Stearns (1938) described several other elevated sites with marine fossils on Lanai. He interpreted the deposits as high stands of the sea as a result of purely glacio-eustatic fluctuations during the Pleistocene and correlated the elevated deposits on Lanai with deposits from other Hawaiian Islands based strictly on interisland comparison of elevation (Table 29). Stearns later speculated on the chronology of these and other shorelines and terraces in the Hawaiian Islands (Stearns, 1974A, 1978a; Figure 5).

Moore and Moore (1984) proposed that the Lanai high-elevation fossiliferous marine conglomerates were sedimentary

Table 29.
Ancient Shorelines on Lanai (Stearns, 1938, 1978a).

Shoreline	Elevation (m)
Present	0
Kapapa	2
Waimanalo	7.6
Kaena	30
Olowalu	76
Manele	170
Kaluakapo	190
Mahana	365

traces of a giant wave generated by a massive submarine landslide that occurred 105 ka (Moore and Moore, 1988). One objective of this study was to apply a recently developed technique, ESR dating, to distinguish between these two competing hypotheses. A further objective was to test a third hypothesis that the marine deposits on Lanai are due to uplift of the island combined with eustatic high sea level stands during interglacials.

Geologic Setting

The island of Lanai is located 15 km west of Maui approximately midway between Molokai and Kahoolawe in the Hawaiian Island chain, 220 km northwest of the active hot spot volcano, Kilauea (Figure 46). Lanai is composed of a single volcano which rises more than 6,000 m from the ocean floor and abuts part of a larger complex of five volcanoes which form the islands of Maui, Molokai and Kahoolawe. The

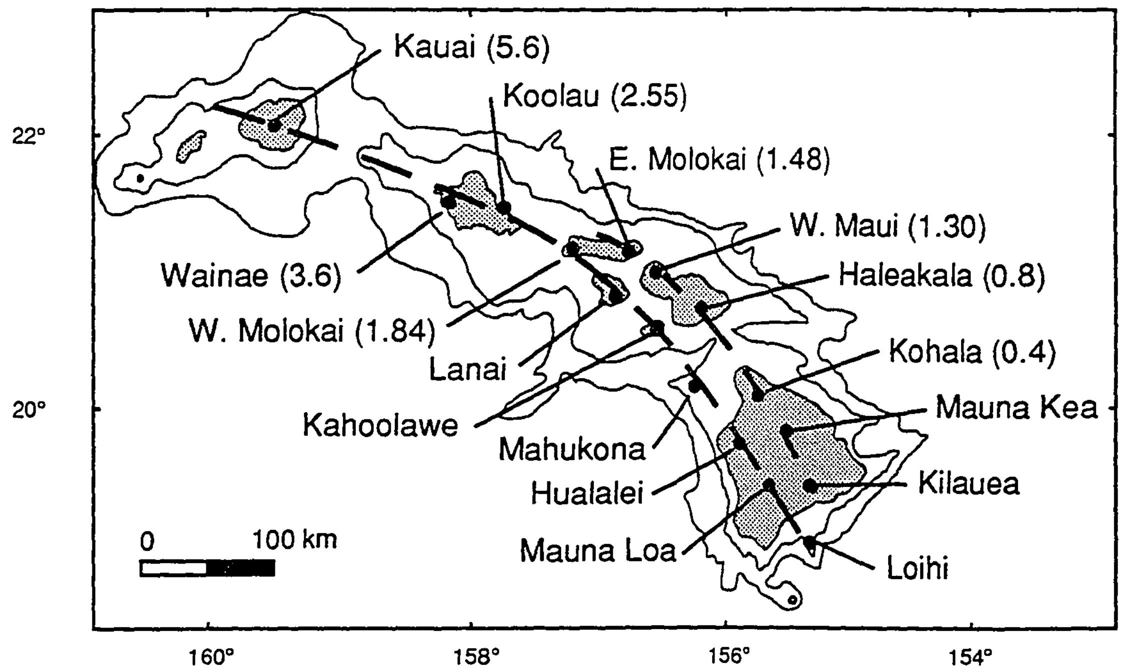


Figure 46. Map of southern Hawaiian Islands showing location of Lanai and paired sequence of volcanoes that form the islands. Black dots are individual volcanic centers. Dashed lines are the tracks of volcanic loci of Jackson et al. (1972) as modified by Garcia et al. (1990).

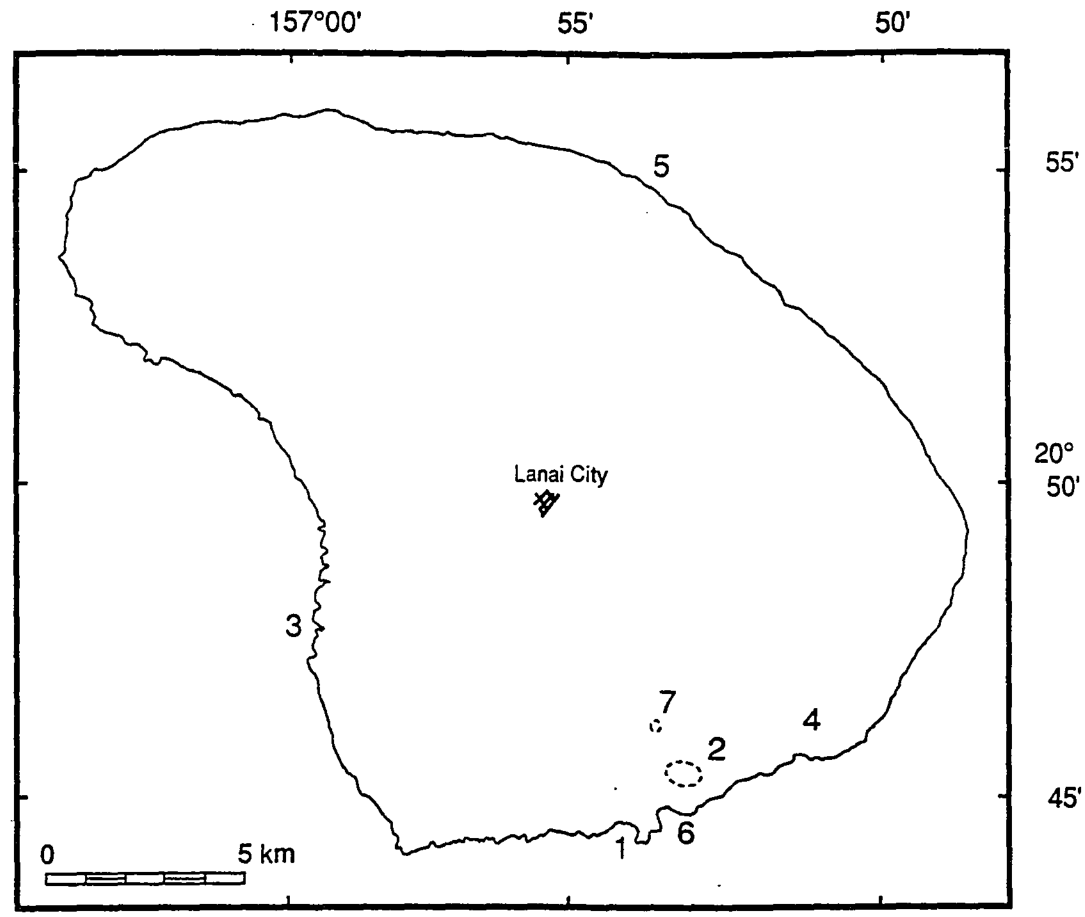
island's subaerial basalts have been dated at 1.25 ± 0.04 Ma (Bonhommet et al., 1977). A detailed account of the geology of the island of Lanai is provided in a monograph by Stearns (1940). A geological map of Lanai by Stearns (1940) delineates patches of marine conglomerate primarily on the southern coast, but also patches of marine deposits on the west coast in selected gullies north of Kaumalapau Harbor and in Kawaiu Gulch (see Figure 47 for index map to locations).

Sample Locality

Sites identified as containing marine fossils from Stearns (1938) and Moore and Moore (1984, 1988) were investigated and sampled (Table 30). Marine fossils especially scleractinian corals were collected in the field. The locations of collection sites are given in Figure 47 and 48. The elevation of the sites was determined with an altimeter (American Paulin System, Micro Surveying Altimeter, Model M-1) and agree with topographical maps (1:25,000 scale USGS, 1984). The highest reported coral deposit at +326 m (Stearns, 1938) could not be relocated. It was concluded after field investigation of this locality that the "deposit" does not exist because of the lack of collaborating geomorphic or sedimentological features representative of coastal landforms.

Description of Samples

Natural exposure of marine fossils located in gullies



1. Hulopoe Bay
2. Kaluakapo Crater
3. Kaumalapu Harbor
4. Kawaiu Gulch
5. Lae Hi Point
6. Manele Bay
7. Puu Mananalua

Figure 47. Index map to locations on Lanai mentioned in text.

near Manele Bay were sampled. Starting from sea level and moving up a section of well-lithified limestone is exposed along various sections of Lanai's coast including the eastern shore of Hulopoe Bay. Mollusk shells are exposed along the sea cliff immediately west of Hulopoe Bay to an elevation of 12 m. Within the carbonate matrix are coral fragments, generally smaller than 10 cm. Subrounded basalt cobbles and boulders are intermixed with the carbonates. Sample (L-10) is from this locality. Moving up, sample L-21 was taken from under a ledge on the west side of the gulch that enters into Kapihua Bay. This is believed to be the type locality of the Hulope Gravel as described in Moore and Moore (1988). It is about 200 m from the shore and a rock cairn marks the site. Further up this same gulch samples L-23 and L-24 were collected in conglomerate that was clast supported with subangular basalt boulders. Analysis of other coral clasts collected in this area showed extensive alteration with the majority of the skeletal aragonite converted to calcite. The corals were also well worn and no evidence for reef structures such as framework were apparent. However, further up this gulch a large in situ brain coral (Platygyra sp.) was found at an elevation of ~30 m (L-95, L-95+). This coral sample was oriented in growth position in a block of limestone that had been exposed by the erosion of the gully. No marine fossils were found above 74 m elevation in this gulch.

Below a newly constructed tennis court at the Manele Bay resort, an exposure of 4 m section of marine mudstone is present. The marine unit is about 41-45 m above sea level and overlies weathered basalt and a paleosol caps the unit as well as hill debris (Figure 52). A single coral clast was removed from the lower portion of this unit (sample t.c.).

In Kawaiu Gulch on the eastern side of Lanai, a 6 m thick deposit at about 53 m was sampled including <u>Pocilliopora</u> <u>meandrina</u>. A remnant boulder beach with characteristic berm was noted in this gulch. Such a structure would require prolonged exposure to sea level and not result from run-up of a tsunami. Even higher, sample L-32 was collected at about 170 m elevation from Kaluakapo Crater in the eastern stream bed within the bowl.

A reconnaisance of the west coast of Lanai found several localities with many marine fossils exposed (Figure 53).

Results

ESR Coral Dates

The ESR results are listed in Table 30. Neutron activation analysis for uranium concentration in the Lanai samples showed an average value of 2.60 ± 0.22 (n = 17), compared with the average for Molokai of 2.49 ± 0.23 ppm (n = 11) and for Oahu of 2.83 ± 0.27 (n = 7). These values are consistent with published values for fossil corals (Veeh and

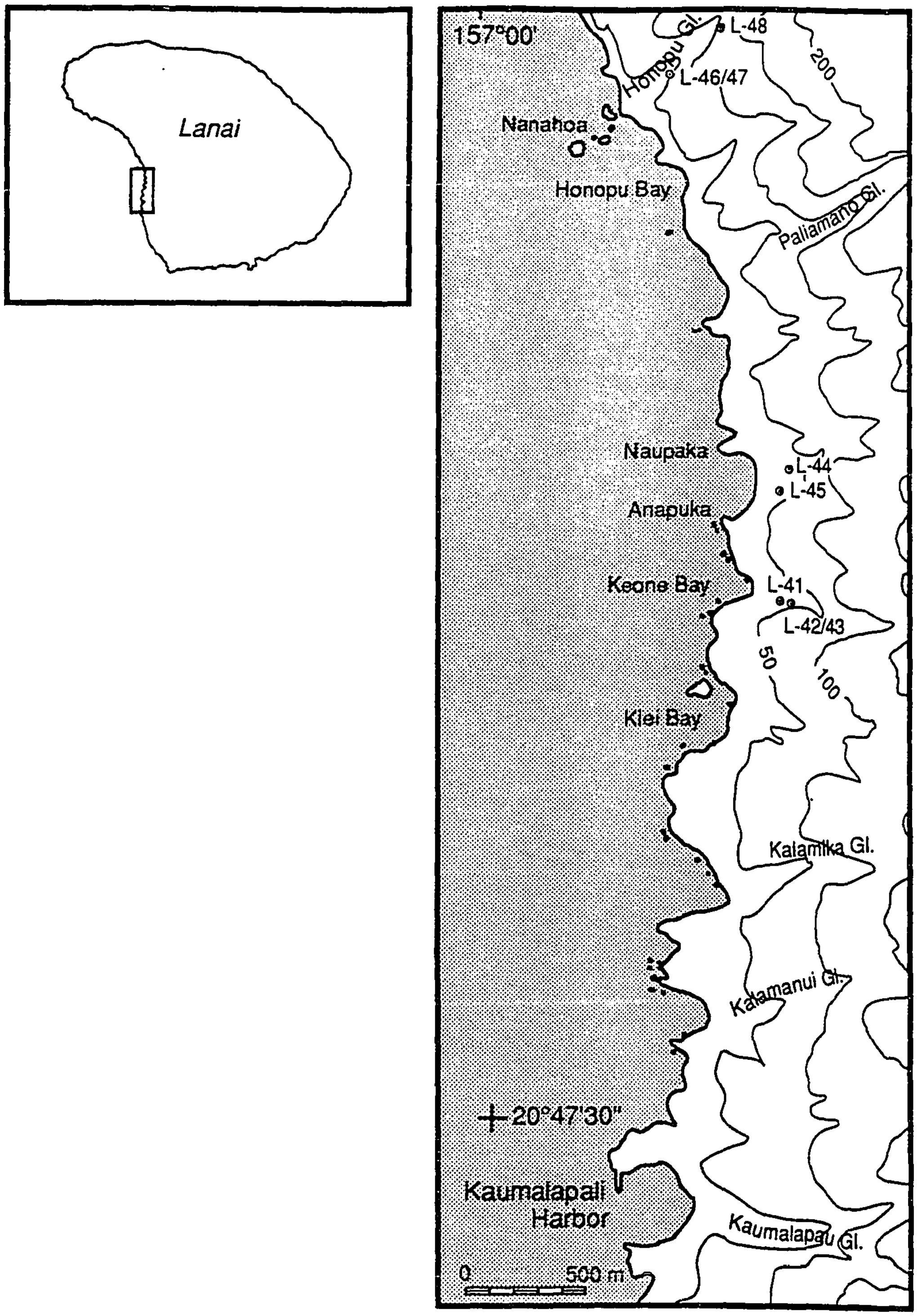


Figure 48. Locations of several sites on the west coast of Lanai with fossil coral outcrops.

Table 30.
Results of ESR coral dating from Lanai, Hawaii.

Elev (m)	Sample No.	U (ppm)	AD (krad)	ESR Age (ka)
12	L-10	2.94	24.12 ± 4.03	210 ± 36
16	L-21S	3.06	21.21 ± 4.47	185 ± 40
16	L-21	3.06	23.62 ± 1.46	202 ± 15
23	L-41B	2.31	18.92 ± 2.03	200 ± 23
23	L-41A	2.38	22.14 ± 1.52	223 ± 18
27	L-42AX	2.68	21.76 ± 2.06	205 ± 21
27	L-42AR	2.59	20.19 ± 1.44	197 ± 16
27	L-42AS	2.38	21.10 ± 1.19	215 ± 15
27	L-42B	2.36	22.06 ± 3.79	224 ± 39
29	L-95	2.77	21.8 ± 1.60	201 ± 17
29	L-95+	2.77	17.98 ± 1.42	173 ± 15
29	L-95	2.56	17.19 ± 4.78	181 ± 49
40	L-47	2.51	2.39 ± 0.69	37 ± 11
44	L-t.c.	2.91	14.31 ± 1.34	140 ± 14
47	L-23B	2.42	23.94 ± 3.17	235 ± 33
47	L-23A	2.54	24.59 ± 1.96	233 ± 21
52	L-26	2.49	22.14 ± 1.74	217 ± 19
74	L-24S	2.74	20.81 ± 3.91	195 ± 38
74	L-24R	2.74	24.57 ± 2.22	223 ± 22
170	L-32A	2.74	13.15 ± 1.24	137 ± 14
170	L-32B	2.60	9.84 ± 1.0	112 ± 12

Burnett, 1982) and do not differ substantially from U content of modern corals (Swart and Hubbard, 1982; Radtke et al., 1988; Hamelin et al., 1991)(Figure 42). Note that for three samples (L-21, L-24, and L-95) a single U analysis was preformed for the subsamples.

The ESR coral ages can be associated with warm stages of the deep-sea oxygen isotope record (Imbrie et al., 1984) as illustrated in Figure 49. Most samples collected from 12 to 74 m elevation appear to correlate with the penultimate interglacial (oxygen isotope stage 7), while two of the highest samples appear to correlate with stage 5.

Interpretation and Discussion

The lowest coral sample from 6 m elevation as previously reported by Veeh in Stearns was dated 110 ± 10 ka (1973). The single age from a coral clast at the "tennis court" site was dated at 140 ± 14 ka. However, before assigning a definitive age to this exposure, more dates are needed. Nevertheless, the mudstone in this deposit indicates a fairly low energy environment of deposition, suggesting that this exposure represents an interglacial high stand.

By far the largest subset of samples dated were on the order of about 200 ka, ranging from 173 ± 15 to 235 ± 33 ka with a mean of 209 ± 16 ka. These samples are correlated to oxygen isotope Stage 7.

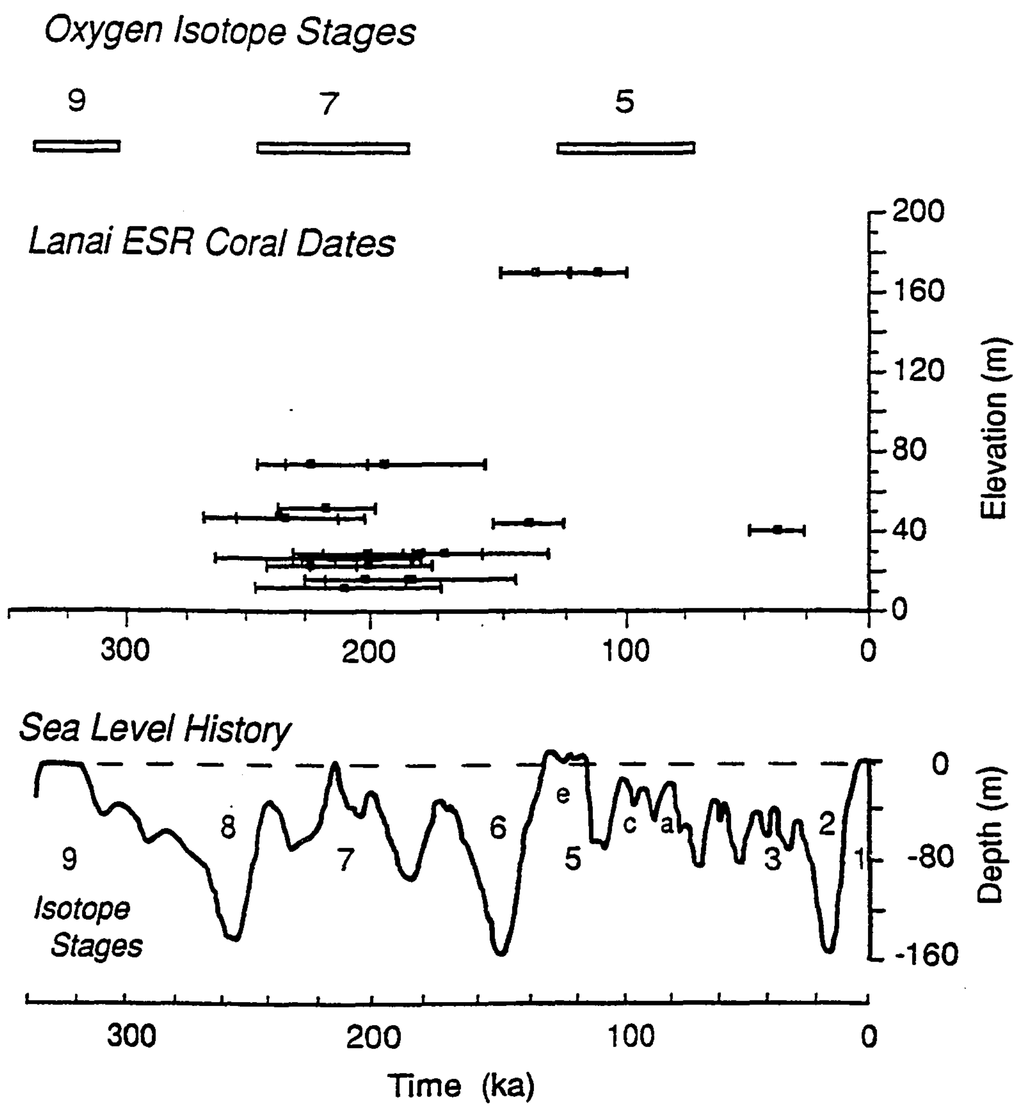


Figure 49. Correlation of coral ESR age from Lanai with oxygen isotope stages (after Imbrie et al., 1984) and sea level history (Chappell and Shackleton, 1986).

Hulopoe Gravel

Moore and Moore (1988) describe a sedimentary formation which includes low boulder ridges (<2 m in height) exposed on the surface near the south coast of Lanai as the Hulopoe Gravel. Moore and Moore argue that the presence of these boulder ridges indicate that a train of giant waves struck the southern coast of Lanai. They describe the bedforms as consisting of "branching dunelike gravel ridges, generally 1 m high with a spacing of 10 m, that are roughly transverse to the slope of the land".

It is interesting to consider that Stearns did concede to the idea of the giant wave. In the second edition of his book, Geology of the State of Hawaii he writes, "The Moores have described a giant wave that hit Lanai from a subsea landslide south of Lanai. The author believes, instead, that the wave was caused by a huge slice of the southwest coast of Lanai sliding into the sea and forming the cliff shown in Figure 9-8. The wave reached about 1,200 feet on Lanai, which is the height of the Mahana shoreline listed in Table 2 as having been due to a still stand of the sea, as are the Kaluakapo shorelines at 625 feet, the Manele at 560 feet, and the two unnamed at 375 ± and 325 ± feet. These shorelines were caused by the landslide as the great wave continued to surge. The Olowalu shoreline at 250± feet on Maui probably represents the height the wave reached on Maui, and the marine deposit on Kohala Mountain on Hawaii at 175 feet probably represents the height of the wave on that island" (Stearns, 1985, p. 245).

An alternative explanation to a giant wave creating the basalt boulder ridges, is that the ridges were artificially created by early Hawaiians as part of a dry-land agricultural complex. Similar dryland agricultural complexes of boulder terraces and mounds are well documented on the leeward side of several islands (e.g. Hawaii: Pearson, 1969; Tuggle and Griffin, 1973; Kirch, 1985; Maui: Kirch, 1985; Molokai: Weisler and Kirch, 1985; Weisler, 1989; Oahu: Hommon, 1969 in Kirch, 1985; Kauai: Sinoto, 1975; Hammatt et al., 1978; and Niihoa: T. Hunt, pers. comm. 1992). Moore and Moore recognized native Hawaiian structures within a section of the gravel bedform that they mapped (Moore and Moore, 1988, their Fig. 4), but dimissed their significance. A plan view of their sketch is reproduced as Figure 50. As with many dryland cultivation sites, field shelters were typically constructed within the complex (Kirch, 1985). The u-shaped "aboriginal house walls" identified by Moore and Moore are consistent with the area being used for dryland cultivation (Weisler, written comm, 1992).

One way to demonstrate that the boulder ridges represent a relict dryland agricultural terrace is through the use of exposure dating such as with cosmogenically-produced stable isotope ^{3}He (Kurz et al., 1990) or with ^{36}Cl (Phillips et al., 1986) which has proven successful in identifying glacial

Figure 50. Plan view of 'Hulopoe Gravel' of Lanai from Moore and Moore (1988).

moraines on Mauna Loa. Exposure dating with ^{3}He has been applied to young Hawaiian lava flows on Hualalai and Mauna Loa by Kurz et al. (1990) [For a review of surface exposure dating see Dorn and Phillips (1991)].

A second way to test the hypothesis that the boulder fields are man-made is to identify characteristics directly related to construction of agricultural terraces such as high organic content of the soil, abundance of charcoal flakes from human-induced burning as a agricultural practice. The dry stone masonry structures should have several attributes such as size, materials, construction technique, configuration, orientation and spatial arrangement which can distinguish them from deposits of a catastrophic wave.

It should be noted that one sample (LNY-32) collected in a swale east of Puu Mahanalua at 170 m elevation was ESR dated at 124 $\pm$ 17 ka. This site is several meters higher than the one dated by Moore and Moore (1988) from Kaluakapo Crater. Further field investigation at this site is necessary before the giant wave hypothesis can be evaluated. Excavation should be done to determine the underlying sedimentary structure. At this time, the field relations of the Kaluakapo Crater and the dated deposit at Kawaiu Gulch can not be determined.

Wentworth (1925) noted that the wide-spread coral fragments and shells over the central plateau of Lanai was associated with Hawaiian stone artifacts as he stated "clearly of human distribution". He also cautioned against misinterpretation of calcium carbonate filling joints as an indicator of past sea levels.

Archaeological Evidence

Recent archaeological investigations (Kaschko and Athens, 1987; Hammatt et al., 1990; Stride and Hammatt, 1990) on Lanai have uncovered greater use of the south coast of Lanai than previously documented (Emory, 1924). These recent surveys have identified adze quarries, shelters, and other archaeological structures in the area mapped by Moore and Moore as "giant wave" deposit. The archaeological teams were familiar with the report by Moore and Moore (1984) of a giant wave deposit in the Hulopoe Bay area and specifically stated that "a conscious effort was made during the field survey to distinguish between naturally-occurring marine materials and possible culturally-transported items" (Kaschko and Athens, 1987, p. 3).

Kaschko and Athens (1987) conducted an archaeological survey of the Hulopoe Bay and Manele Bay coastal areas. They report numerous sites with archaeological features. Three of which had culturally-transported marine materials above 85 m in elevation. One such site, a collapsed oval cairn (3 m x 2.6 m x 0.5 m) at 87 m elevation, had two water-worn coral clasts and shells of Cypraea. The structure was constructed with basalt cobbles and boulders. At a second site, a walled structure (4.3 m x 3.8 m x 0.8 m), weathered discoidal coral artifacts were exposed on the surface. Nearby were shells of

Cypraea, Cellana, and Drupa in sparsely-spaced midden remains.

A far more startling discovery was a Hawaiian foot trail leading from near Manele Bay up the ridge crest toward the inland plateau (Kaschko and Athens, 1987). Along this route, 15 features were identified and described (Kaschko and Athens, 1987; Appendix 2). The highest feature identified was at an elevation of about 183 m. Presumably the trail route extended on up the ridge crest toward the upland plateau. If so, then the prehistoric trail would intersect with the type locality for the 326-m Manele shoreline (Stearns, 1940). Stearns had originally described this deposit as

> "several vein-like fillings of fossiliferous marine limestone in crevices in basalt...These outcrops are only a quarter to half an inch wide and 2 to 3 feet long. They contain, however, distinctly recognizable coralline algae and one gastropod. Some fragments of coral are discernible." (Stearns, 1938, p. 618).

It is quite conceivable that Stearns had identified and described a part of the trail marking cairn. Note that the species described by Stearns (1938, see Table 4) at the Manele type locality are all "edible" mollusks: mollusks that general attain a sufficient body size that the meat is worth the effort to collect and prepare for eating. Similar species are described from several of the archaeological structures along the trail leading from Manele Bay up the ridge crest (see

Appendix 2). Native Hawaiians, who have occupied the Hawaiian Islands since 500 A.D., are known to have used materials from the sea for a variety of purposes including ceremonial, religious, marking of special sites, namely fishing (ko'a) and birding shrines, and "liming" of the island's acidic soil (David Shideler, pers. comm. 1992, see also Kirch, 1985).

An archaeological survey conducted in the coastal region bounded on the east by Kapihua Bay and on the west by Huawai Bay, identified agricultural terraces which had been "modified by Hawaiians for planting with rock clearing and stacking on the natural retaining alignments" (Hammatt et al., 1990). The report further cites locations of basaltic dike intrusions that were quarried for raw materials to form adzes and other stone tools. Hammatt et al. (1990) document archaeologically significance sites in a section of the southern Lanai coast that Moore and Moore (1988) had mapped as Hulopoe Gravel (Figure 51). In a related study, Stride and Hammatt (1990) surveyed a parcel of land upslope from the Hammatt et al. study. Stride and Hammatt (1990) describe a single site comprising a C-shaped structure with an adjoining circular enclosure. A piece of coral was visible on the surface of the C-shaped structure at an elevation of approximately 160 m.

The conclusion drawn from recent archaeological reconnaisance of the south coast of Lanai in the proximity of Hulopoe Bay is that the boulder ridges in the area are not relict depoists of a catastrophic wave but rather are

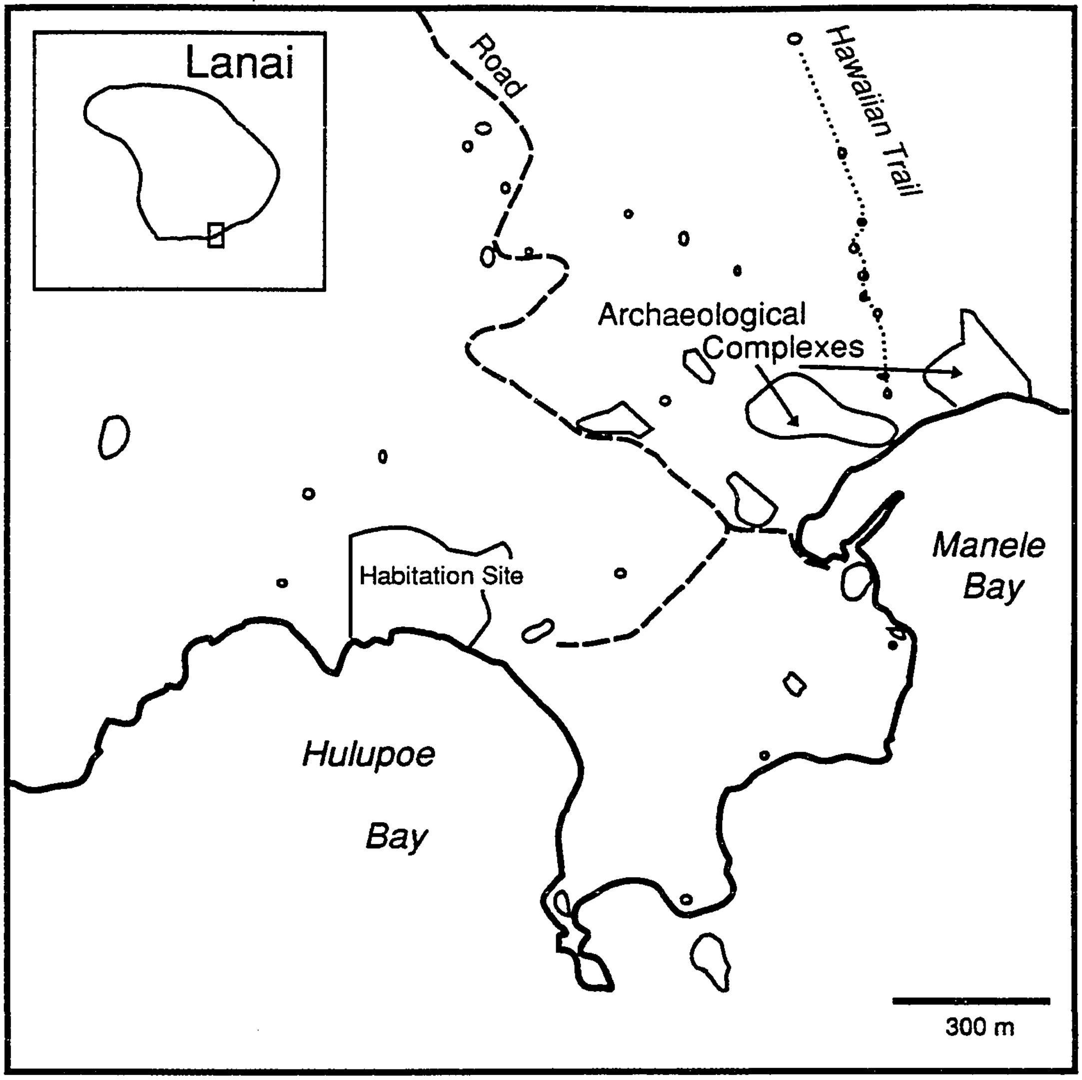


Figure 51. Location of archaeologically significant features identified on the southern coast of Lanai. Compiled from Kaschko and Athens (1987), Hammatt et al. (1990) and Stride and Hammatt (1990).

naturally occurring materials built by native Hawaiians (basalt cobbles, boulders, marine skeletal and reef material). in construction of habitation sites and in agricultural terraces.

It should also be noted that along the southern coast of Lanai, the only area with access to the sea from the shore are the sandy beaches of Hulopoe and Manele Bays.

SUMMARY

Moore and Moore (1984, 1988) re-examined sites originally described by Harold Stearns as ancient shorelines. They hypothesized that a giant wave caused by a massive submarine avalanche, located off Hawaii (Moore et al., 1989), swept the southern coast of Lanai. They dated this event at 105 ka based on $^{230}Th/^{234}U$ age of three corals from two localities.

The pattern of emerged marine fossils on Lanai can be explained without invoking a cataclysmic event. Abundant archaeological evidence suggests that the boulder ridges identified as Hulopoe Gravel by Moore and Moore (1988) are in fact, archaeological features known as dryland agricultural terraces.Based on ESR dating of coral from several sites on Lanai, the age and elevation of fossil corals is interpreted to be evidence for uplift of the island during the last 300 kyr. This uplift can be readily explained by the flexing of the lithosphere under Lanai caused by the loading of hot spot volcanos Mauna Kea and Mauna Loa.

CHAPTER IX

HOLOCENE CORAL REEF ON KAUAI

Introduction

Sea-level fluctuations during the Pleistocene have profoundly influenced the evolution of coastal systems. The volcanic islands of Hawaii are no exception. During the last 20 kyr, sea level has risen from 100-130 m (Kennett, 1982), flooding former coasts, valleys, and plains. Whether sea level was actually higher than present during this period has been a concern of geologists for some time (Fairbridge, 1961; Milliman and Emery, 1968, and others). Recent studies in the south Pacific have documented a higher than present sea level during the last 6 ka (Pirazzoli and Montaggioni, 1988) and compilation of global Holocene sea-level curves present a complex nature of Holocene sea-level change (Pirazzoli, 1991).

In this chapter, new data are presented for a mid Holocene paleo-shoreline on Kauai, providing convincing evidence for a sea level high stand about 1.8 m above present in the north central Pacific. This is the first unequivocal evidence for a Holocene sea level high stand in the Hawaiian Islands.

Geologic Setting

Kauai is the sixth major island in the Hawaiian Archipelago which is a linear chain of volcanic islands and coral atolls that stretch between Hawaii in the southeast to Kure in the northwest. Built over a stationary hot spot

(Wilson, 1963; Morgan, 1972), the islands are carried northwestwardly on the Pacific plate, displacing progressively older islands to the northwest. These islands are located in the subtropics where vertical coral reef growth ranges from about 14 mm yr^{-1} in the southern part of the chain to about 1 mm yr^{-1} at Kure Atoll (Grigg, 1982).

Kauai is approximately 5 Myr old and has well developed beaches, large nearshore sand deposits, extensively eroded valleys and large river systems. Kauai is the wettest island in the Hawaiian chain, receiving over 10,000 mm of precipitation at Waialeale annually (Ronck, 1984). The large river systems on Kauai, such as the Hanalei River, occasionally flood the valley plain as evidenced by extensive alluvial deposits. Hanalei River on the north shore of Kauai feeds into an almost semicircular bay, which opens to the north. The Hanalei Bay is well protected by two headlands: Puu Poa Point to the east and Makanoa Point to the west (Figure 52).

The Holocene Record

The Holocene record from the Hawaiian Islands has been described primarily from Oahu. Easton and Olson (1976) drilled the coral-algal reef in Hanauma Bay, Oahu. They attempted to reconstruct Holocene sea level from reef growth and concluded from radiocarbon chronology that sea level had not risen above its present level on Oahu since 7 ka.

In addition, Bryan (1989) surveyed the low surf benches

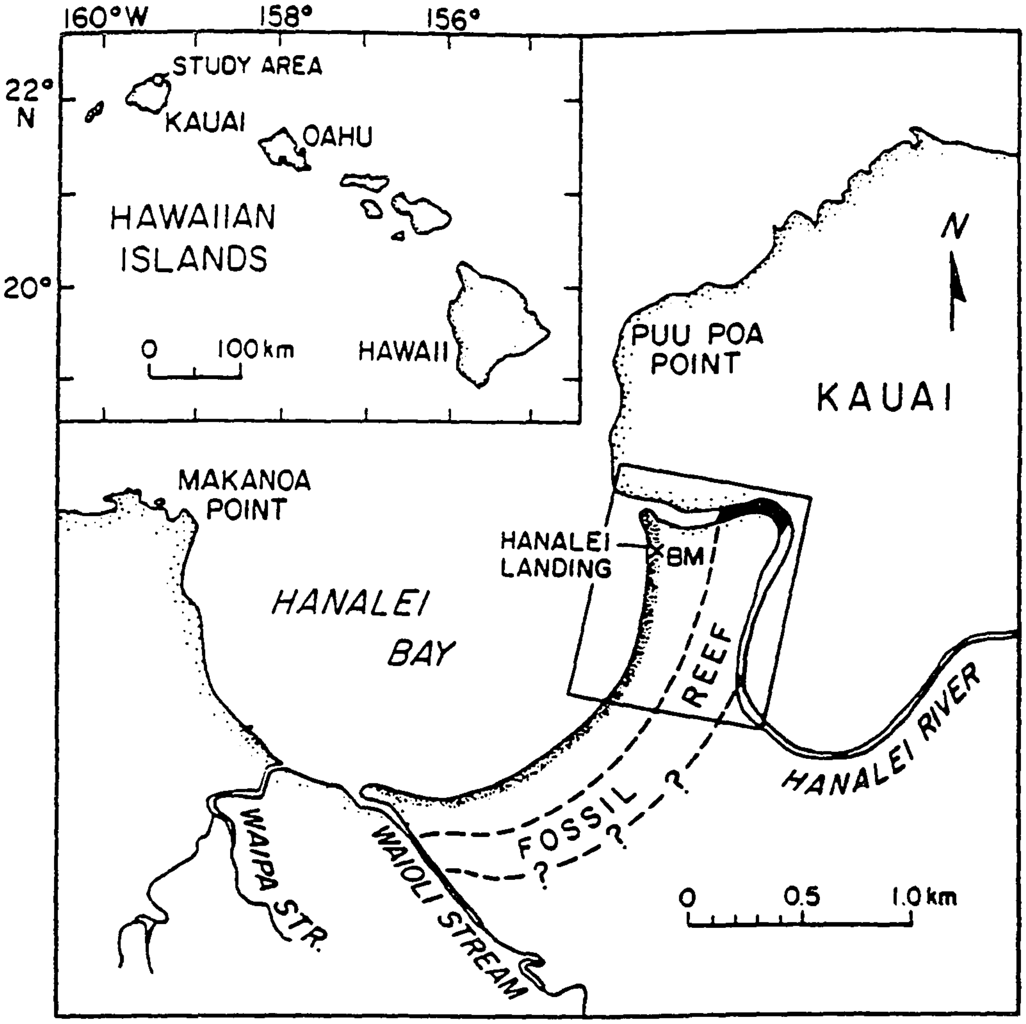


Figure 52. Location of Holocene reef in Hanalei River estuary, north Kauai. Extent of fossil reef is shown. Box indicates coverage of Figure 53.

on Oahu and concluded that the wave-cut benches do not provide evidence for sea level significantly above its present level during the Holocene.

Harold Stearns (1977), however, argued for a Holocene high stand, his so-called "Kapapa stand" (Stearns, 1935a), based on a dated coral (3.482 $\pm$ 0.160 ka) from an outcrop of beach rock in Hanauma Bay ranging in elevation from +1.5 to +2.1 m above present sea level. Stearns, noting the elevation of the conspicuous wave-cut terraces and notches around Oahu at 2 m, correlated the elevation of the Kapapa stand with the age and elevation of benches elsewhere in the Pacific. However, Easton (1977) suggested that the coral sample dated by Stearns had been thrown up by storm waves from the lower reef that dated about 3 ka. Furthermore, Easton emphasized the difficulties inherent in correlating elevations of terraces and notches across ocean basins. Hence, the existence of a Holocene high stand in the Hawaiian Islands has been in question for more than fifteen years.

Methods

Field surveys of the Hanalei River estuary and the Waioli Stream were undertaken to map the geographic extent of the fossil reef discovered during a visual census in the Hanalei River estuary. Multiple elevation measurements from a mean sea level datum were determined using a digital altimeter with a precision of $\pm$0.3 m as specified by the manufacturer. Altimeter elevations were calibrated to a nearby US Coast and

Geodetic Survey tidal benchmark at the Hanalei Landing. Mean tidal range (MLW-MHW) in the Hanalei Bay is 0.4 m (National Ocean Survey, 1989).

^{14}C Dating

Corals (Porites compressa) preserved in growth position were sampled for ^{14}C dating (Figure 53): the outer surface was scraped to remove encrusted estuarine bivalves, calcareous serpulid tubes, and filamentous algae. Subsequent X-ray diffraction analysis revealed that all samples were composed of >98 percent aragonite. Five or more grams of the carbonate material were etched in weak acid to remove the outer layer and were radiocarbon dated by Beta Analytical Laboratories (Miami, Florida) using benzene synthesis. Ages are based on a ^{14}C half-life of 5568 years and are reported as conventional ^{14}C years before 1950 A.D. The errors given in Table 31 represent one standard deviation. Measurement of $\delta^{13}C$ and the isotopically corrected ages ("^{13}C adjusted ages") are also provided in Table 15. Corrections for isotopic fractionation of ca. +400 yrs was compensated by the "carbon age" of the oceanic reservoir (ca. -400 yrs)(see Stuiver and Pollach, 1977, Stuiver et al., 1986). In the following discussion, conventional ^{14}C age will be used to facilitate direct comparison of results from Easton and Olson (1976).

Results

Approximately 500 m^2 of fossil reef have been uncovered in a meander of the channel in the mouth of the Hanalei River;

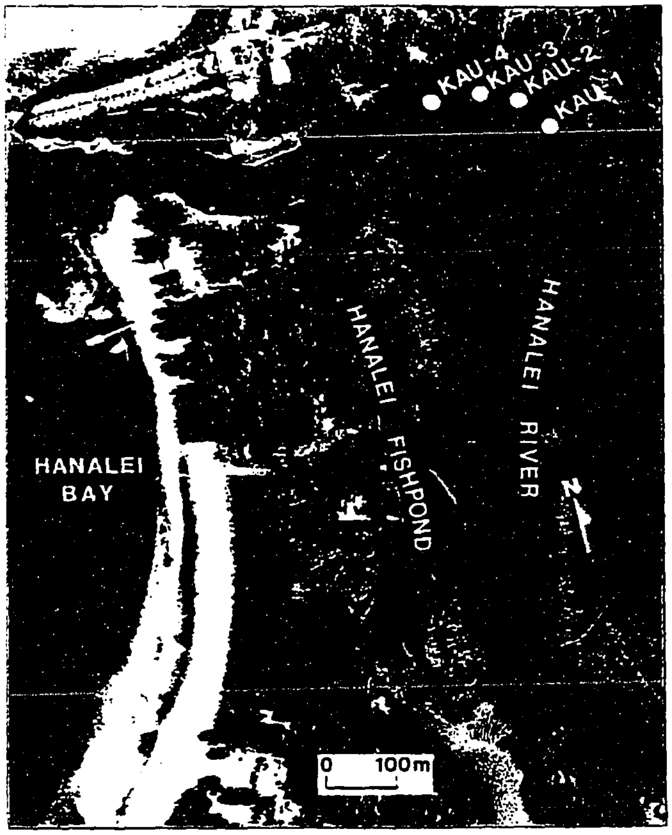


Figure 53. Sample sites for ^{14}C dating in Hanalei River estuary. Aerial photograph from U.S. Corps of Engineers (1983).

a smaller reef section, about 700 m upstream, is also exposed. The reef corals are typically in growth position. Inspection of pot-holes and recesses in the fossil reef reveal the framework is at least 1.5 m thick. The dominant coral in the Hanalei River exposure is Porites compressa, which forms a nearly monospecific stand of upright finger-like branching colonies, some greater than 1 m in diameter. Maragos (1977) documents Porites compressa growing in wave-protected areas of Hawaii, where it is spatially competitive and the dominant species. Furthermore, this species accounts for a large portion of the Waimanalo reef material found on Oahu (Pollock, 1928). The solitary coral, Fungia scutaria, was also collected at the Hanalei River exposures. It is a common member of the modern reef community which is generally restricted to the upper 3 m of water depth on the reef flat. Waioli Stream, approximately 2 km southwest of the Hanalei River estuary, also contains an exposure of about 150 m^2 of reef flat. Pocilliopora meandrina and Fungia scutaria were collected from this location. In modern Hawaiian waters, Pocilliopora meandrina is a dominant coral growing on hard substratum in wave exposed environment commonly between 1 to 15 m water depth (Maragos, 1977; R. Grigg, pers. comm., 1991). No reef material was observed during field work in a third stream discharging in Hanalei Bay, the sediment-filled Waipa Stream.

The results of the radiocarbon dating are reported in

Table 31.
Radiocarbon dates of *in situ* corals from Hanalei River estuary, Kauai.

Lab. no.	Field no.	$\delta^{13}C$	^{13}C Adjusted Age	Conventional ^{14}C Age B.P.
Beta-41951	Kau-1	-2.3	4570 ± 70	4200 ± 70
Beta-41952	Kau-2	-2.7	4040 ± 80	3670 ± 80
Beta-41953	Kau-3	-0.7	3630 ± 80	3230 ± 80
Beta-41954	Kau-4	-0.4	4470 ± 80	4070 ± 80

Table 31. The coral samples range in age from 4.2 to 3.23 ka. The reef framework is found at a maximum elevation of +1.8 ± 0.3 m in the Hanalei River.

Discussion

In situ reef corals provide a useful estimate of the minimum elevation of sea level in tectonically inactive areas and are readily dated (e.g., Hopley, 1986). Their stenohaline requirements ensure that coral samples represent growth in normal marine saline waters (ca. 35 ‰). The presence of *Porites compressa*, *Pocilliopora meandrina* and *Fungia scutaria* in the Hanalei River exposures implies an ancient shallow-water reef environment, possibly analogous to present day Kaneohe Bay reef flat on the windward side of Oahu, where *Porites compressa* comprises more than 80% of the coral cover (Maragos, 1977). In Kaneohe Bay, some *Porites compressa* colonies are exposed at extreme low tides. Even though it is not restricted to depths near the surface, *Porites compressa* often forms monospecific reefs in shallow areas tens to hundreds of meters from the present shoreline. However, the presence of corals in growth position does not give a definable relationship to the paleo-sea level except for a minimum elevation. The lack of coral species with a restricted depth range in Hawaii limits the ability to interpret the maximum height of sea level during the Holocene.

Previously reported dates of the absolute ages of fossil corals from Kauai are few: twenty-five years ago, a 10 fathom deep (-18 m) reef terrace off northeast Kauai (Kaheko Reef) was radiocarbon dated at 8.37 $\pm$ 0.25 ka (Inman and Veeh, 1966) and the age was confirmed by a U-series date of 8 $\pm$ 1 ka (Veeh, 1966).

On Kauai, Matsumoto et al. (1988) report ^{14}C dates of the pelecypod Tellina (3.85-4.04 ka) from a coastal plain near Kapaa. The samples were from a marine unit 0.7-1.2 m above mean sea level. Overlying the marine sediment was a non-marine stratum of fine sand with terrestrial gastropods (Leptachatina spp. and Carelia spp.). A date of 2.65 $\pm$ 0.17 ka was determined for Leptachatina above +1.3 m. Matsumoto et al. suggest a possible mid-Holocene high stand, although marine materials could have been deposited by a storm and do not necessary represent sea level indicators such as the in situ reef corals.

The ^{14}C dated corals from the Hanalei River exposure indicate that relative sea level stood at least + 1.8 m above its present level on Kauai at 4.2 - 3.2 ka. Sea level was probably at least 1 m higher to account for the development of the reef. Since 4 ka, relative sea-level has apparently dropped to its present level. Whether this retreat was smooth or episodic and punctuated with minor regressions cannot be established from the fossil reef data. However, from the spread of coral ages, one can speculate that the reef development in the Hanalei River exposure represents a stillstand of at least several hundred years. A similar Holocene high stand has been documented for islands in Micronesia (Tracey and Ladd, 1975; Buddemeier et al., 1975), the South Pacific at French Polynesia (Pirazzoli and Montaggioni, 1988) and the Cook Islands (Woodroffe et al., 1990a), and the central Indian Ocean (Woodroffe et al., 1990b). A complete atlas of Holocene sea level curves has been compiled by Pirazzoli (1991).

Preservation of the Hanalei River fossil reef probably resulted from burial by fluvial sediment or periodic flooding and delta construction with progradation that buried the exposed fossil reef. As the shoreline shifted seaward, the reef structure remained buried. These reefs would then have been protected from mechanical erosion. Finally, during recent times, the Hanalei River removed the overburden at the seaward bend exposing the reef structure.

Implications

The Holocene high sea-level stand on Kauai agrees in amplitude and timing with a high-resolution global model of the last deglaciation (Figure 54). This recently developed model (ICE-3G of Tushingham and Peltier, 1991) considers three components; solid earth, ice, and ocean, and preserves the equal potential gravitational field and accounts for glacial isostatic adjustment. The model was constrained by ^{14}C-controlled relative sea-level curves from ice-covered sites and was tested against geologically controlled glacial retreat isochrons, oxygen isotope data from deep-sea sedimentary cores, and the elevation of coral terraces. The model was later validated with relative sea-level curves from ice-free sites (Tushingham and Peltier, 1992).

This geophysical model predicts a Holocene sea-level rise in the central Pacific of 2 m at 5 ka (Tushingham, 1989). Data from Easton and Olson (1976) generally fall below the predicted sea-level curve, with the exception of material at about -10 m. Easton and Olson's data from Hanauma Bay indicate no high stand; but samples from above sea level were not collected. Deposits from a high stand may not have been well preserved in Hanauma Bay. The fringing reef in this semi-protected crater would not be preserved by a prograding delta, as in the case of the Hanalei River, but instead, may have been planed off by wave action and bioerosion. In an earlier study, Easton (1973) published a composite cross section with a wave-cut bench at +1.5 m (Figure 55). A still stand from about 7.1 to 6.1 ka at around -9 m is indicated in the radiocarbon profiling of Hanauma Reef. Easton and Olson

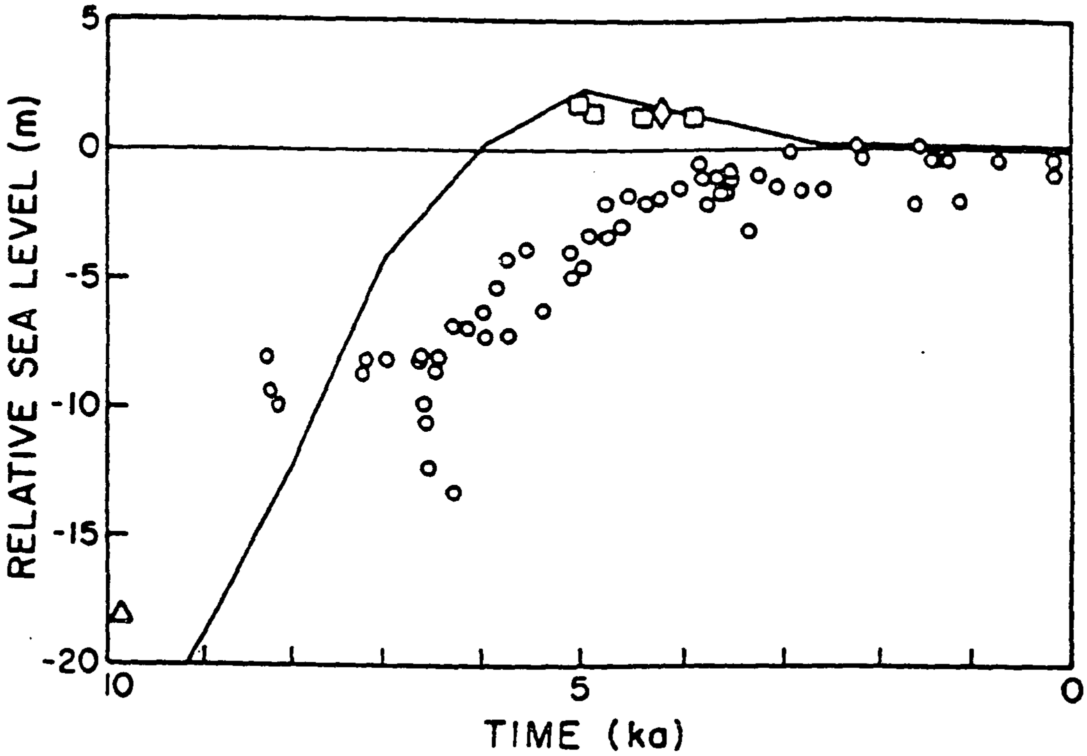


Figure 54. Comparison of the observed data from Oahu (circle: Easton and Olson, 1976; diamond: Stearns, 1977) and Kauai (triangle: Inman and Veeh, 1966; squares: this study) with the predicted relative sea level history curve for Oahu using the ICE-3G melting chronology (Tushingham and Peltier, 1991).

(1976) originally interpreted this still stand as possible evidence for uplift of Oahu since they were unable to find records of such a still stand at other locations. Recent studies of submergent shorelines of the southwest Pacific suggest an episodic transgression from the last glacial period (Carter et al., 1986). Among the still stands listed by Carter et al. is one at -9 m that occurred around 7 ka.

Montaggioni (1988) re-evaluated Easton and Olson's data. He considered the reconstruction of the Holocene sea level changes as propounded by Easton and Olson to be a minimum curve with a paleo-sea level 2-15 m above their estimated minimum curve. The corals studied by Easton and Olson are not restricted to a narrow depth zone. Stearns' (1977) date falls near the predicted curve, although the dated coral fragment may have been transported from the reef as mentioned above.

The ICE-3G model, however, does not consider any vertical adjustments owning to local crustal loading at the central Pacific hot spot. Lithospheric flexure resulting from crustal loading should be confined to a radius of 400 km around the hot spot (Watt and ten Brink, 1989). Oahu and Kauai are 350 and 500 km from the hot spot, respectively. Therefore, Kauai should be beyond the primary influence of this flexure and, over the time frame of the Holocene, considered stable.

The age and elevation of the Kauai fossil reef reported here provide the first undisputable evidence for a mid-Holocene high stand in the Hawaiian Islands. The growing

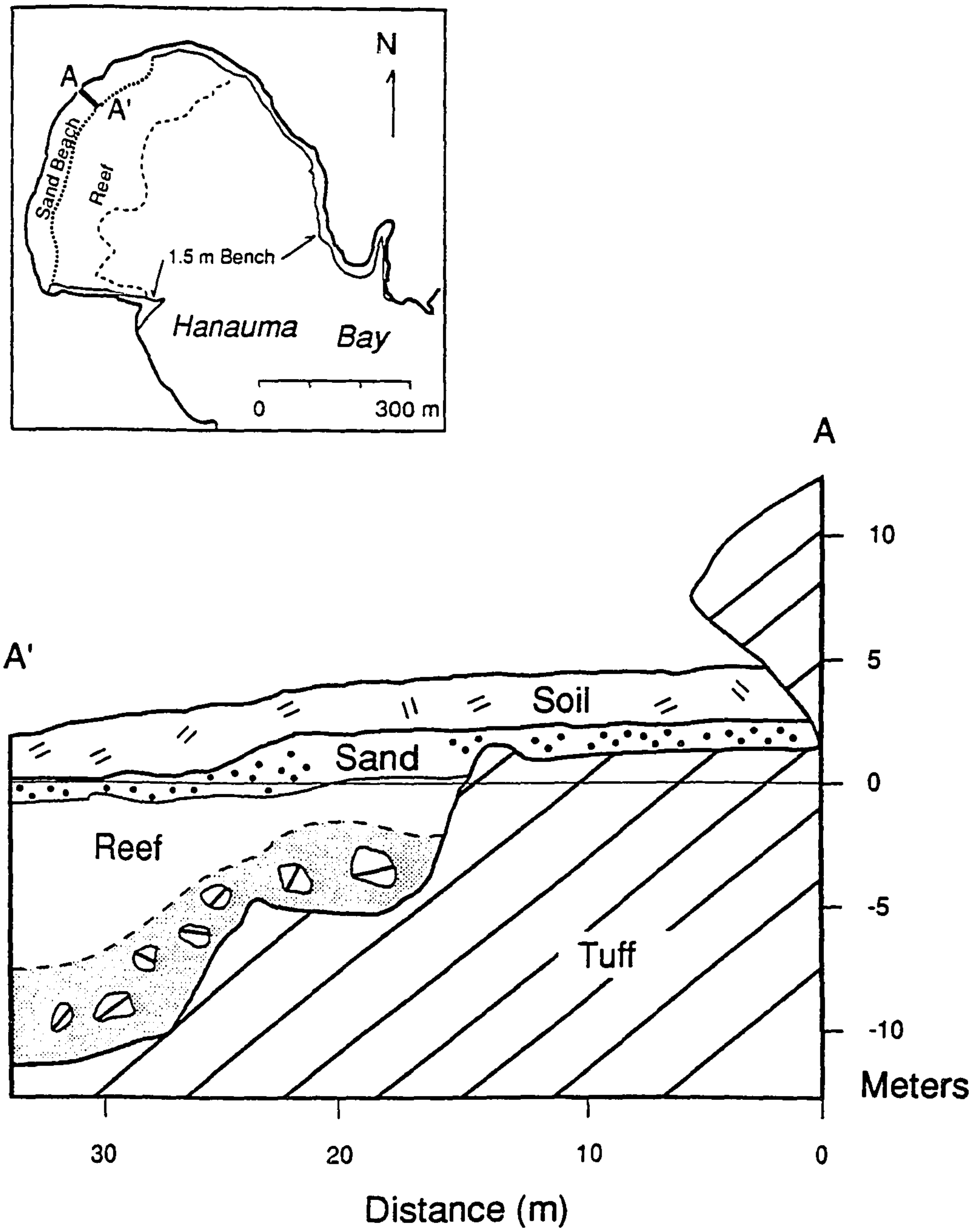


Figure 55. Composite cross section of Hanauma Reef. Note the notch in the tuff. Redrawn from Easton (1973).

consensus of reports on a widespread Holocene high stand draws attention to the meager data set available for evaluation of Holocene sea level and the potential for rapid change in the global climate system.

To summarize, a shallow-water coral reef discovered in Hanalei River on Kauai at +1.8 m above sea level was dated by ^{14}C to be 4.2 - 3.2 ka. Recent geophysical modeling of hotspot flexure indicate that at least during the Holocene, Kauai can be considered stable. The Hanalei River reef suggests a higher than present sea level existed in the central Pacific during the mid-Holocene. These results agree with recent high-resolution geophysical models of Holocene deglaciation and evidence at other Pacific Island oceanic sites.

The reefs in the Hanalei River should prove to be an excellent location for extraction of a complete record of the Holocene transgression using modern drilling techniques and new high-precision methods of dating corals. The scope of the current study excluded any extensive drilling, however, this study does suggests that a critical re-examination of sites on Oahu, such as the Kapapa stand of Stearns (1978a), is warranted.

CHAPTER X

DISCUSSION OF RESULTS

Introduction

The previous chapters have described evidence for uplift of the islands of Oahu, Molokai and Lanai. This chapter will focus on a discussion of a proposed mechanism to account for the uplift of these three islands. Two possible explanations for the data will be considered in detail. They are (1) uplift associated with lithospheric flexure due to loading at the Hawaiian hot spot and (2) isostatic compensation from subaerial erosional loss and submarine mass wasting (landslides).

Uplift Pattern

To summarize the previous three chapters, field evidence and ESR dating from the islands of Oahu, Molokai, and Lanai suggest that each island underwent uplift during the Late Pleistocene. The rates of uplift for the islands vary and are summarized in Table 32. For these calculated rates, it is assumed that the paleo-sea level height for Stage 5e was +6 m, for Stage 7 was +2.5 m, and that the paleo-sea level height for the other interglacials discussed (Stage 9, 11, 13) was about 0 m. Figure 56 illustrates the correlations between the elevation of the raised reefs on Oahu, elevated terraces on Molokai and Lanai and marine climatic record (SPECMAP, Williams et al., 1988).

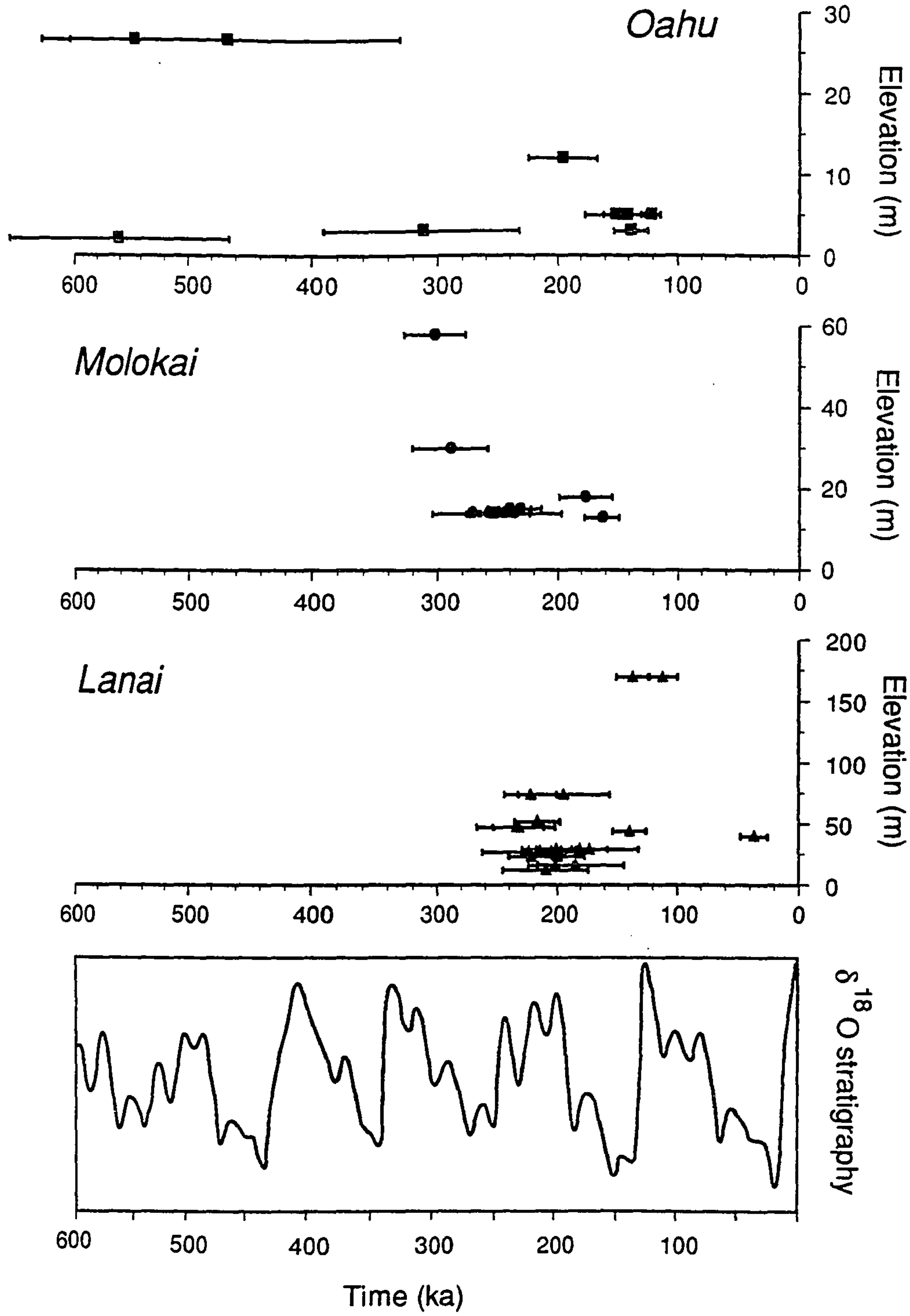


Figure 56. Plot of present elevation and age of emerged marine deposits from Oahu, Molokai, and Lanai with oxygen isotope records from the deep ocean (Williams et al., 1988).

Table 32.
Estimated Uplift Rates for Oahu, Molokai, and Lanai.

Island	Time Period (ka)	Uplift Rate ($mm\ yr^{-1}$)
Oahu	0-125	0.02
	125-200	0.06
	200-500	0.06
Molokai	0-125	0.04
	0-300	0.10
Lanai	0-200	0.34

Lithospheric Flexure

Lithospheric flexure will be the first mechanism examined to account for the uplift. Because the lithosphere can be modeled as an elastic plate, loads applied to the lithosphere such as volcanic islands and seamounts exert a force on the lithosphere which responds by deformation. Part of this deformation is a proximate trough and a distal arch or lithospheric bulge at a radial distance of a few hundred kilometers. The distance depends on the elastic properties of the lithosphere (rigidity, effective elastic thickness of the lithosphere) and the amount of loading (density structure of the load and degree of infilling of the moat). The basic theory of bending elastic materials is provided by Nadai (1963).

The proximity of the three uplifted Hawaiian islands to the active hot spot volcanism at Kilauea and the magnitude of the uplift imply, but do not require, that the uplift is directly related to the compensating flexural bulge of the lithosphere. Several observations support flexure of the lithosphere surrounding the island of Hawaii including topography, gravity measurements, seismic data (Watt and ten Brink, 1989), and moat seismic stratigraphy (Rees et al., 1993).

Early workers such as Vening Meinesz and Gunn analyzed gravity data from the Hawaiian Islands. This allowed Vening Meinesz (1934, 1941) to recognize that the volcanic pile was regionally compensated. Gunn (1943) using some simple assumptions applied a two-dimensional beam analysis to determine the wavelength of flexure. Watt and Cochran (1974) determined the deformation of the lithosphere along the entire Hawaiian-Emperor chain using gravity anomaly and topography profiles. They concluded that the amplitude and wavelength of the free-air gravity anomaly profile can be explained by a simple model of lithospheric flexure.

The overall topography of the Hawaiian Islands lends support to the flexure interpretation. The prominent Hawaiian Ridge is superimposed on a broad elongate, low rise, called the Hawaiian Swell, which is about 1200 km across (Dietrick and Crough, 1978). At the base of the Hawaiian Ridge is a depression or trough called the Hawaiian Deep (Dietz and Menard, 1953; Hamilton, 1957). The Deep is best developed along the east side of the island of Hawaii. A large arcuate peripheral arch, the Hawaiian Arch was recognized in early submarine bathymetric profiles by Dietz and Menard (1953). Dietz and Menard (1953) related the offshore bathymetry to the structure of the islands by recognizing that the Deep was a depression caused by the superposed load and that the Arch resulted from a compensating "elastic bulge". Thus, the geomorphology of the Hawaiian Ridge and surrounding bathymetry strongly support flexure of the lithospheric plate.

Walcott (1970a) first quantified the hypothesis that the Hawaiian Deep and Arch were produced by elastic flexure of the lithosphere under a load. His best fit model to explain the wavelength and amplitude of the arch surrounding the island of Hawaii was a broken elastic plate model (Figure 57). Several basic observations on the geometry of the moat and arch complex fit with Walcott's broken plate model. These include the distance from the ridge to the crest of the Arch (~ 250 km), the amplitude of the Arch, on the order of few hundred meters (~200 m), and the width of the Arch (~ 500 km). Although there is difficulty in assessing the width and the amplitude of the Arch, nevertheless, the basic conclusion is that the secondary features of the Hawaiian Arch and Deep are caused by loading of the lithosphere.

Only a segment of the Hawaiian Ridge will be arched by the flexed plate resulting in the uplift of overlying islands. Using the broken plate model and a representative value for the distance from the point load to the maximum height of the bulge, x_b = 250 km, we can determine the overall length for

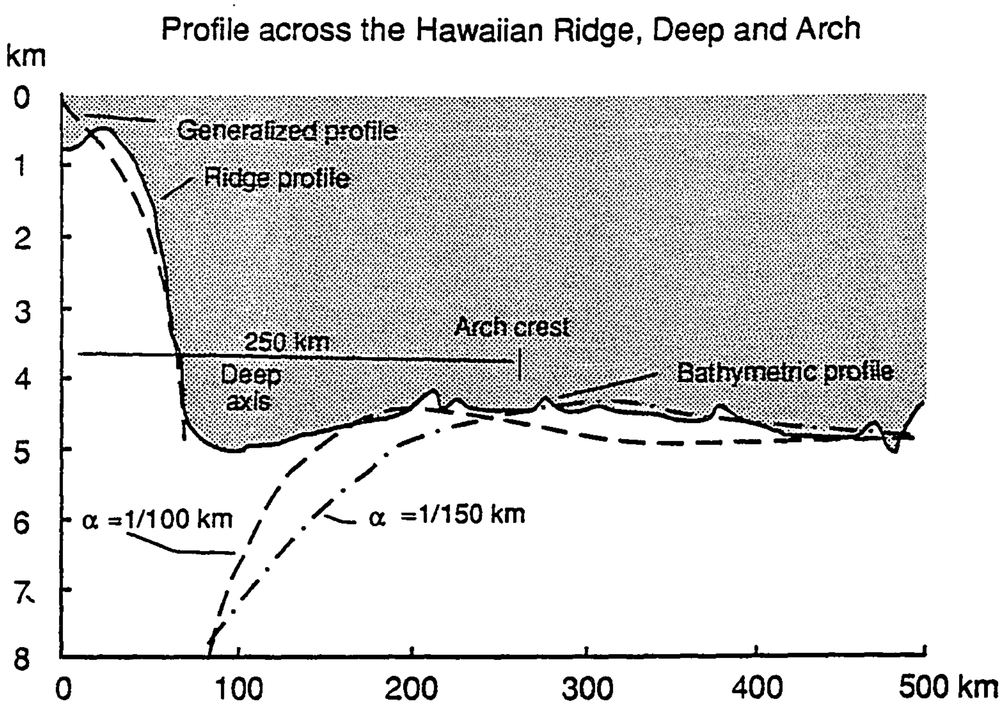


Figure 57. Bathymetric profile off southeast Hawaii with calculated profiles (dashed lines) for a broken elastic plate. The best overall fit with the arch is for a flexural parameter of approximately 1/125 km, however the 1/100 km curve provides a good fit between 150 and 225 km away from the hot spot (redrawn from Walcott, 1970a).

the arched segment. The first crossing, x_0 is expressed as

$$x_0 = 0.75\frac{\pi}{\alpha}$$

where the flexural wave number, α, for the Hawaiian Archipelago is approximately 1/80 km. The second crossing x_1, can be approximated by

$$x_1 = 1.55 \times \frac{\pi}{\alpha}$$

and the segment, therefore, spans approximately 200 km. Assuming that the hot spot loading has been fairly constant over the Quaternary, we can estimate the amount of time an island is within the zone of arched lithosphere. With a horizontal plate movement of approximately 10 cm yr^{-1} (= 10 km 100 ka^{-1}) (Jackson et al., 1980), islands are subject to uplift for about 2 Ma.

Estimation of maximum uplift from flexural model

Using simple models of flexure, Lambeck (1981a) calculated the distance from the load center to the bulge and the maximum height of the bulge for values of flexural rigidity (D) ranging from 2 to 8 x 10^{29} dyne cm and varying degrees of moat infilling (β). Table 33 presents the maximum elevation of the arch and the distance from the load to the arch for a load with a density of 2.5 g cm^{-3}. As is readily apparent from Table 33, as the degree of infilling increases,

Table 33.
Distance to Arch from center of load and elevation of arch for a flexural model with various flexure rigidity (D) and degree of infilling (β).†

D Flexural Rigidity ($x10^{29}$ dyne cm)	β Degree of Infilling	Radius to Arch (km)	Maximum Elevation (m)
2	0	195	220
	0.5	205	180
	1	225	50
5	0	225	245
	0.5	235	220
	1	260	45
8	0	245	280
	0.5	260	220
	1	285	40

† From Lambeck (1981a).

the radius of the arch increases and the elevation of the arch decreases, and, as the flexural rigidity increases, the radius to the arch increases. For the Hawaiian Islands, the degree of infilling can be assumed to be quite low because predominately pelagic sediment (50-100 m thick) overlies volcanic basement off the island of Hawaii (Rees et al., 1993) and the topographic expression of flexure is revealed in the offshore bathymetry. The radius from Mauna Loa to the volcanic centers for the principal Hawaiian Islands and their volcanos are listed in Table 34. The distance from the hot spot to Lanai, Molokai and Oahu are within the segment expected to undergo uplift.

Table 34.
Approximate distance from hot spot to Principal Hawaiian Islands.

Island	Volcano	Distance (km)
Maui	Haleakala	182
	West	221
Lanai		226
Molokai	East	256
	West	280
Oahu	Koolau	339
	Waianae	374
Kauai		519

* From Clague and Dalrymple (1989).

The elastic plate theory assumes that once the plate is loaded the flexure is instantaneous. This is probably not the case. A response time on the order of 10^3-10^4 yr is considered realistic for the mantle to react. This time is derived from the response of the continental lithosphere to the unloading of the massive ice sheets during deglaciation (Walcott, 1970c), but oceanic plates are less well characterized in this regard. The system is also probably far more complex than the simple models imply. The Hawaiian Ridge acts as a line load on the plate in addition to the hot spot volcano acting as a point load. Individual pedestals vary in volume and the volcanic production rates have undoubtedly changed through time. Also the influence of the Molokai Fracture dissecting the ridge axis separates lithosphere of different ages and therefore, its thickness, rigidity and rheology. Whether the flexural properties of both sides on the fracture zone are adequately represented in the model studies has not been addressed. Hence, the combined effect of a line load - the Hawaiian Ridge - and a point source loading interact to yield the observed trough parallel to the ridge and around the southern end of the island of Hawaii. Simple two dimensional models can not account for the above mentioned complicating factors. However, two dimensional models do provide an approximation of the physical properties and the expected resulting vertical movement of islands near hot spots.

Isostasy

Isostatic adjustment as a result of subaerial erosion or mass wasting is an alternative to lithospheric flexure as a mechanism to explain the uplift of the islands of Oahu, Molokai, and Lanai. In order to examine the effect of isostasy related to erosion, it is first necessary to discuss erosion rates for the Hawaiian Islands and then proceed to calculate, based on isostasy, the amount of uplift that it might produce.

Rates of Denudation

Present-day rates of erosion for various types of rocks can be estimated from dissolved chemical species in river runoff. Moberly (1963) estimated the total denudation rate for the Kaneohe Bay watershed based on the removal rate of dissolved calcium by rivers. Later, Li (1988), using an extensive water resource database, examined the chemical constituents of Hawaiian rivers and groundwater to estimate the chemical and physical denudation for the islands of Hawaii, Oahu, and Kauai (Table 35). His rates agree with

Table 35.
Denudation Rates for Hawaiian Islands.

Island	Denudation Rates ($mm\ yr^{-1}$)	Reference
Hawaii	0.04 - 0.05	Li, 1988
Oahu	0.13	Moberly, 1963
	0.04 - 0.16	Li, 1988
Kauai	$\leq$ 0.19	Li, 1988

rates determined from Ca ion leaching rates for each island. A general trend of increasing chemical denudation with increasing age of the island was observed and, coincidently, lower mean island elevation. Given the overriding influence of the trade winds on orographic precipitation, islands in the lee of high islands, such as Lanai, may have drastically different chemical denudation rates than those islands exposed to fairly constant trade winds. Differences between windward and leeward sides of the island were also noted by Li (1988). Clearly, physical or mechanical denudation of an island is a complex problem and difficult to quantify. For Oahu estimates vary between 32 - 128 km^3 of material removed by erosion during the last 500 ky. Below, the present-day denudation rates are used to calculate the thickness of the removed surfaces.

Isostatic Compensation

Can the uplift observed for the islands of Oahu, Molokai and Lanai be explained by isostatic compensation due strictly to erosion. Referring to Figure 58, let us first assume that the island is in isostatic equilibrium. From the conditions of isostasy, we know that

$$\rho_1 h_1 + \rho_2 h_2 = \rho_3 D_1$$

$$\rho_2 h_2 = \rho_3 D_2$$

where ρ_1 is the density of the surface material, ρ_2 is the density of the volcanic structure and ρ_3 is the density of the underlying mantle, while h_1 is the thickness of the eroded section and D_1 and D_2 are the distance from the base of the compensated volcano to a reference plane such as the ocean floor. By subtracting the above two equations, we obtain

$$\rho_1 h_1 = \rho_3 (D_1 - D_2)$$

or

$$(D_1 - D_2) = h_1\left(\frac{\rho_1}{\rho_3}\right)$$

where $(D_1 - D_2)$ is simply the amount of uplift and is directly related to the depth of the eroded section and the difference in density between the eroded material and the underlying mantle. The thickness of the eroded section, h_1, can be

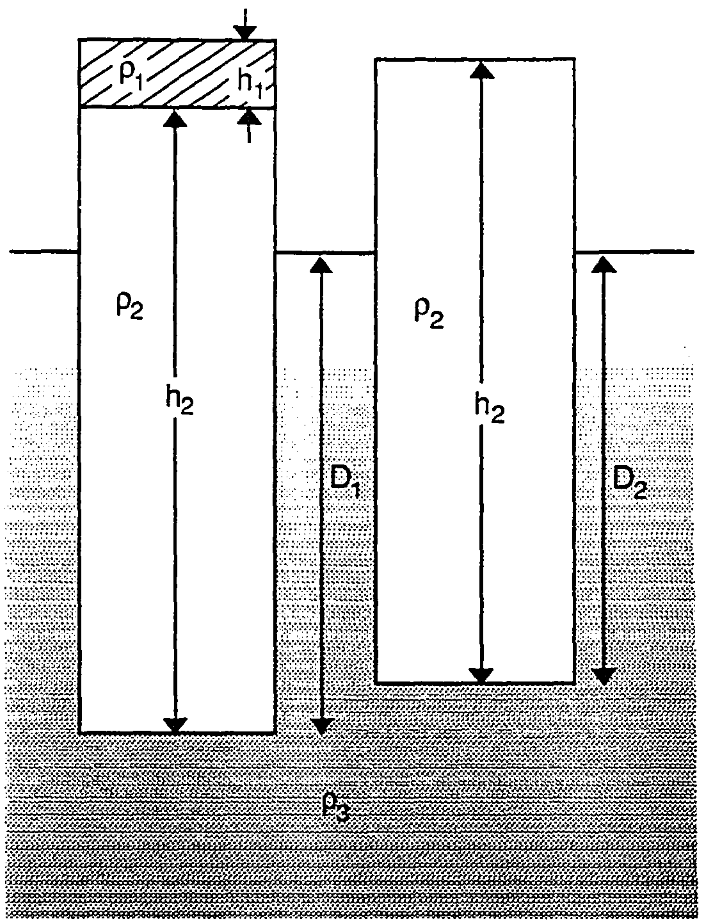


Figure 58. Isostatic equilibrium of an island (a) before erosion and (b) after erosion.

estimated from the denudation rates mentioned above. The range of denudation rates for Oahu allow for a 20 to 80 m to be removed during the last 0.5 My. The density of the Hawaiian Islands is approximately 2.6 g cm^3 while the density of the underlying mantle is 3.0 g cm^3 (Strange et al., 1965). Given the range of 20 - 80 m, an uplift of 17 - 69 m can be determined from the above relation for Oahu for the last 0.5 My.

Although the amount of uplift associated with erosion of an island is significant over 500 ky compared to the observed uplift of Pleistocene reefs, regional isostatic compensation due to erosion is probably a minor factor in contributing to uplift for the southeastern Hawaiian Islands for the following reasons. The majority of the eroded sediment is not lost from the system, as assumed in the simple analysis above, but rather is retained by sedimentation in the nearshore or in nearby offshore areas. The local reef systems around the Hawaiian Islands capture sediment removed by stream erosion. The above calculations do not consider the increase in island mass due to biological construction of coral reefs (ρ_{reef} = 1.45 g cm^3) by calcium carbonate secreting organisms.

Implications

Evolution of Hawaiian Islands

The uplifted reefs and marine deposits described in earlier chapters have bearing on the vertical history of the islands and the evolutionary succession in island formation.

As the Hawaiian Islands progressively age, erosion and subsidence eventually transform them into low-lying coral islands or atolls. This evolution can typically be divided into six stages: a genesis stage of shield-building, a phase of rapid subsidence followed by differential subsidence as the island moves away from the active hot spot, an emergent stage resulting from lithospheric flexure and two final stages of slow subsidence for atoll development and guyot construction before a given island succumbs to subduction in the Aleutian Trench. A schematic representation of the six stages of evolutionary history for the Hawaiian islands is given in Figure 59.

1. Island Genesis

All of the Hawaiian Islands are active submarine volcanoes which have erupted over the Hawaiian hot spot. Some have developed on the flanks of pre-existing volcanoes such as the present situation for the submarine volcano of Loihi (Malahoff, 1987; Moore and Clague, 1992). A typical seamount in the chain requires about 300 ky (Moore and Clague, 1992) to reach sea level where it then becomes an island. The areal growth rate for an island the size of Hawaii, is about 0.02 $km^2\ yr^{-1}$ over its first 600 ky of history (Moore and Clague, 1992). During this time, it will subside 2-4 km with the bulk of the subsidence occurring within the first 1 My (Moore, 1987). From seismic refraction profiles, the structure under Mauna Loa indicates that the maximum vertical depression of

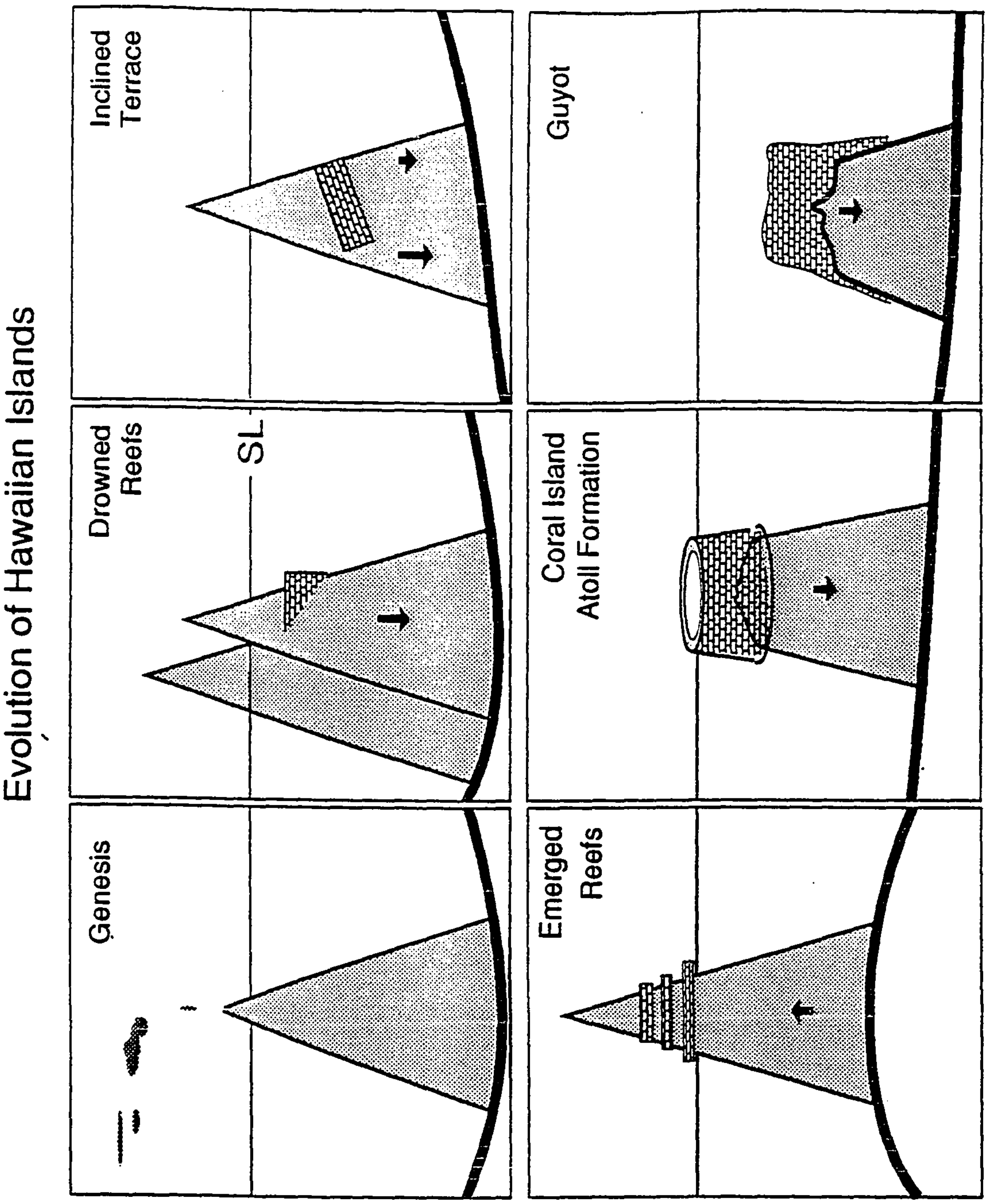


Figure 59. Schematic representation of the evolutionary stages in the history of the Hawaiian Islands.

the base of the crust has been about 9 km (Zucca et al. 1982).

2. Drowned Reefs

Submersible observations on the flanks of the Kohala and Haleakala volcanoes reveal a flight of deeply-submerged coral reefs (Jones, pers. observation). Campbell (1984) was the first to propose a model for drowning of coral reefs on rapidly subsiding Hawaiian volcanos (Figure 60). His observations of bathymetric benching combined with deep-sea bottom photographs of coral rubble led to the development of a model to explain the drowned reefs on the Kohala slope down to approximately 1000 m depth. As the island sinks from addition of new mass and sea level falls, the combined effect is a stable shoreline allowing formation of nearshore coral reefs. At the conclusion of continental glaciation, melt water raises sea level causing the reefs on a subsiding edifice to experience rapid relative sea level rise thereby drowning the reef. The stair-like sequence of drowned reefs off northwest Hawaii (Ludwig et al., 1991) represents the termination of glacial periods, as well as is a measure of the rate of sinking of the island.

3. Inclined Terraces

Several deeply submerged reef terraces identified off the island of Maui are inclined or tilted toward the center of volcanic loading on the island of Hawaii (Figure 61). Two of the prominent benches, the H- and K- terrace, extend for about 100 km along the slope of Maui and Molokai (Moore, 1989).

Reef-Drowning Model

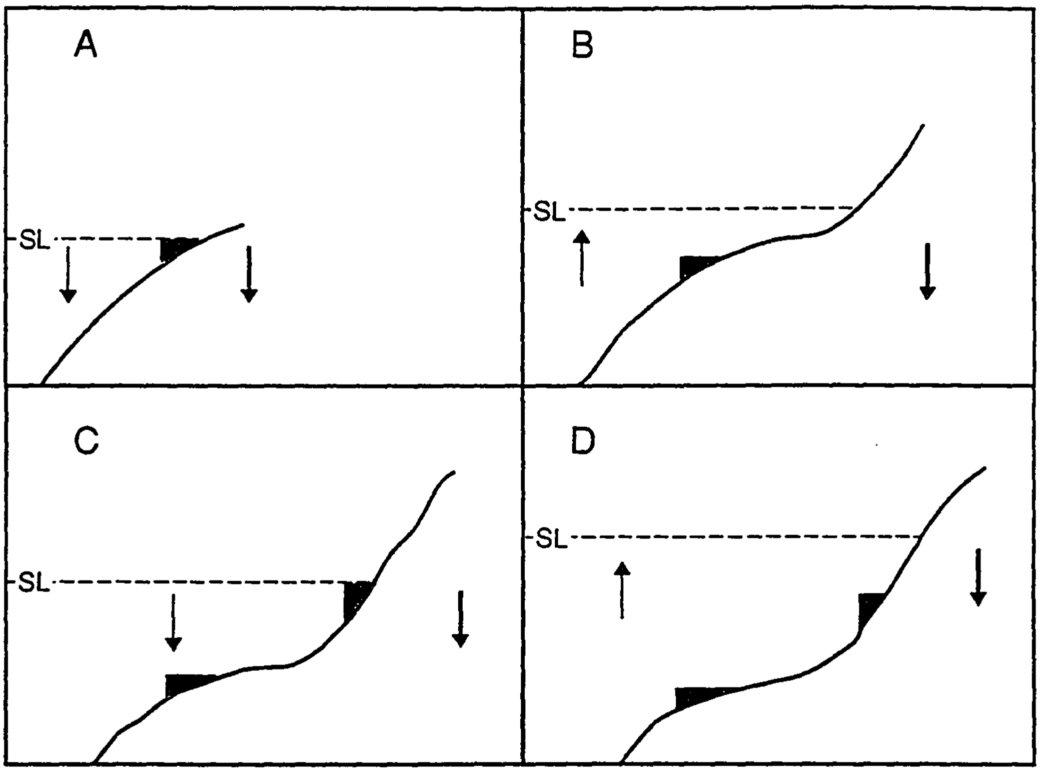


Figure 60. Campbell's model of reef drowning on a subsiding volcanic edifice (redrawn from Campbell, 1984).

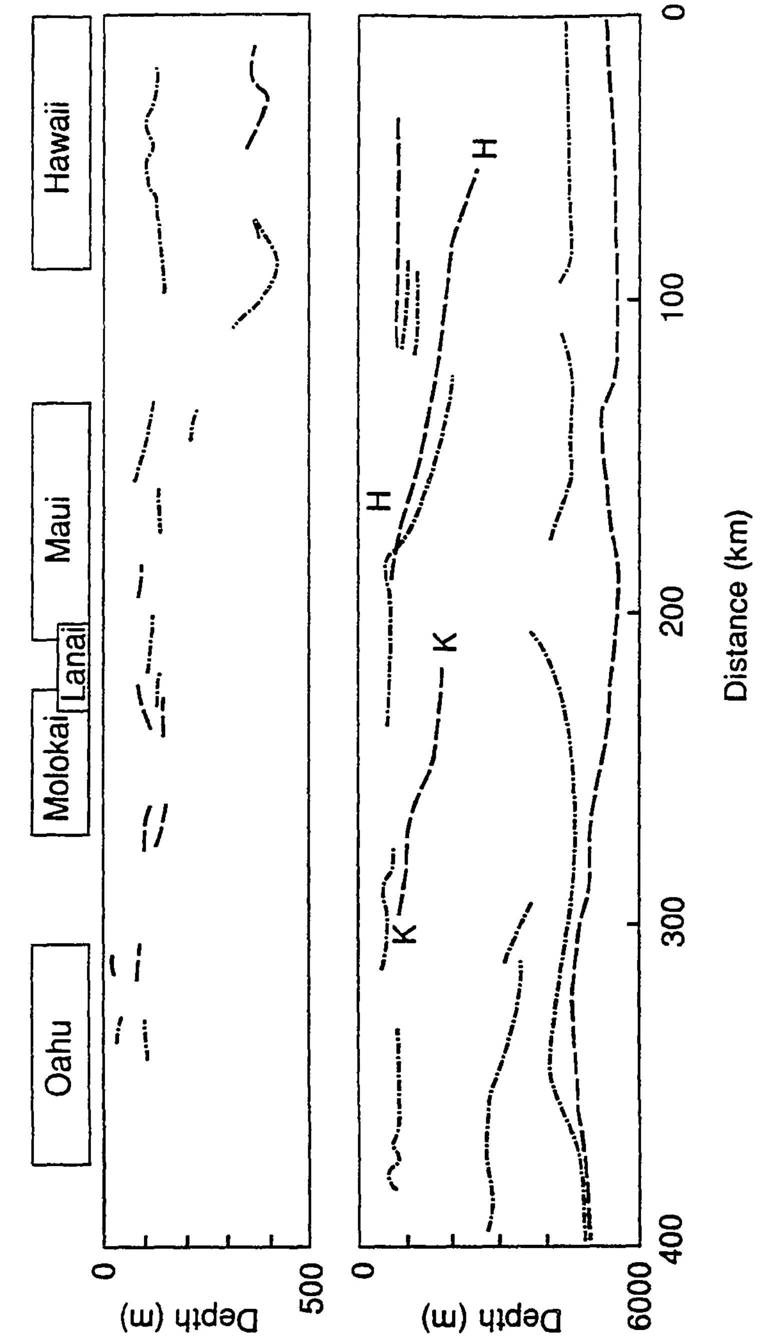


Figure 61. Depths of reef terraces projected onto a vertical plane running N 45° W. Upper panel water depth to 500 m. Lower panel water depth to 6000 m with lowermost profile of the depth of the Hawaiian Deep. Features shown for the north (dashed lines) and south (dash-dot lines) side of the ridge (modified from Moore, 1989).

These observations provide evidence for differential subsidence along a segment of the Hawaiian Ridge.

4. Emerged Reefs

In this thesis it has been shown that at a distance of a few hundred kilometers from the hot spot loading, lithosphere flexure can account for the uplift of islands causing interglacial coral reefs to be elevated. This phase may last 1-2 Ma depending on the timing and amount of loading and the spacing between islands. The building of the island of Hawaii within the last 0.5 My is the loading source responsible for the uplifted reefs on Oahu, Molokai and Lanai. Figure 62 illustrates a history of the areal growth of the island of Hawaii (Moore and Clague, 1992). Note the significant growth of Mauna Kea and Mauna Loa within the last 500 ky.

5. Coral Island/Atoll Formation

After an island moves off the lithospheric bulge, it gradual undergoes subsidence. Subsidence is related to thermal contraction of the plate causing the sea floor to generally deepen with age (Crough, 1978). This phase is the beginning of the classical development of coral islands as originally proposed by Darwin (1842). As high islands, fringed by coral reefs, subside, the encircling reef gradually grows up keeping pace with sea level and eventually develops into a barrier reef system, and finally into a coral island or atoll. In the Hawaiian Archipelago after approximately 12 Ma, islands develop into an atoll. Due to transport on the

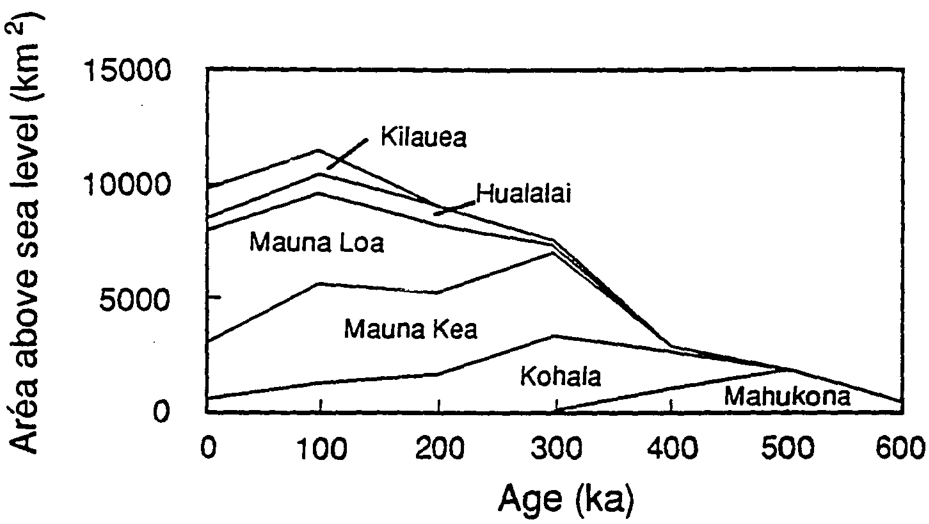


Figure 62. Growth history of the subaerial extent of the six volcanoes that comprise the island of Hawaii (redrawn from Moore and Clague, 1992).

Pacific Plate, they eventually reach a critical latitude where reef growth and accretion are balanced by subsidence and erosion. Beyond this threshold, dubbed the Darwin Point by Grigg (1982), the coral island drowns and enters its final stage of evolution.

6. Guyot

In the final stage in the Hawaiian Archipelago, drowned atolls become flat-topped planated seamounts or guyots (Hess, 1946). For the next 40 Ma or so, they will progressively subside as the plate as it heads toward eventual subduction in the Kamchatka Trench.

Chronology of Submerged Reef Terraces

A series of deeply submerged reef terraces off Lanai were identified from detail bathymetric maps of the southeastern Hawaiian Islands by Campbell (1986). These reefs probably drowned by a combination of rapid sea level rise at the termination of continental glaciation and the rapid subsidence of the island due to hot spot loading in a manner similar to the process that drowned the series of reef terraces on the flanks of the Kohala Volcano (Campbell, 1984; Ludwig et al., 1991). The shallowest Lanai terrace at -125 m most likely represents drowning because of rapid sea level rise at the termination of the last glacial maxima (~19 ka). The series of deeper terraces have been correlated by Campbell (1986) with the sequence of terminations of glacial maxima derived from deep-sea oxygen isotope stratigraphy (stage boundaries

2/1, 6/5, 8/7, 10/9, 12/11 and 16/15)(Figure 63). These transitions occurred at 13, 128, 251, 347, 440 and 592 ka, respectively (from Shackleton and Updyke, 1976 as cited in Campbell, 1986). However, these deeper terraces could have formed considerable earlier after the initial shield building of Lanai was complete around 1.25 Ma when the island of Lanai was closer to the hot spot and near the present location of the Kohala Volcano. Based on the ESR dates of the emerged reefs, the drowning of the reefs off Lanai probably occurred before 0.5 Ma, when Lanai was undergoing substantial subsidence possibly during the genesis of Haleakala.

If the major slope break off Lanai at 1350 m water depth marks the shoreline at the conclusion of the shield building stage as others have postulated (e.g. Moore and Campbell, 1987) and the age of Lanai is approximately 1.25 Ma (Bonhommet et al., 1977), then by using the oxygen isotope chronology developed by Williams et al (1988), the deepest reef could correlate with the end of stage 40. The next four shallower terraces correlate with the end of stage 38, 36, 34, and 32 (Figure 63, Williams et al., 1988). At this juncture however it is not possible to assign a definitive age to the five terraces deeper than 200 m.

Rejuvenated Volcanism

The flexing of the Hawaiian Islands may have initiated or coincided with secondary eruptions resulting in the prominent post-erosional vents, such as the landmarks of Diamond Head,

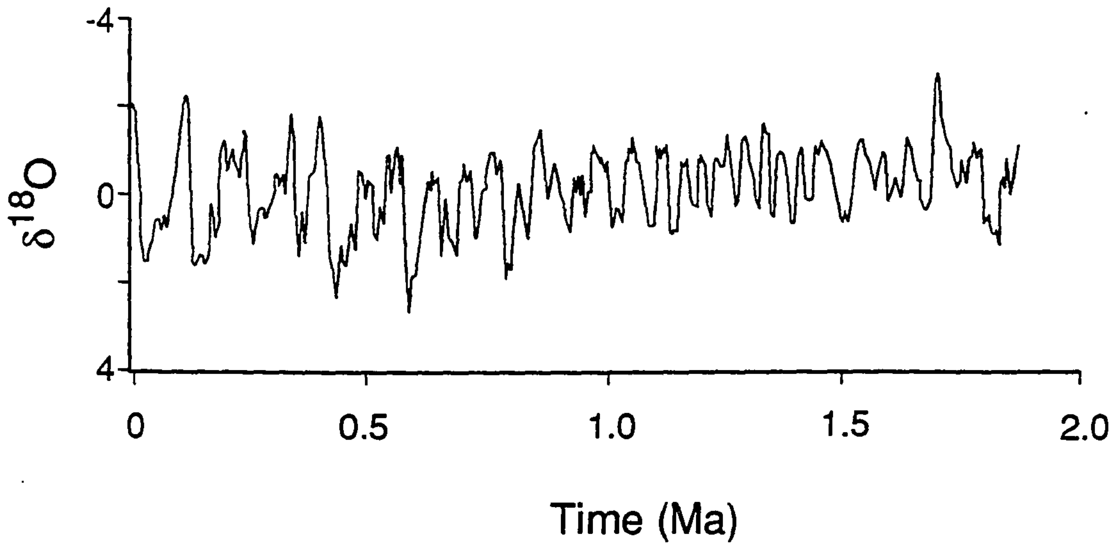


Figure 63. Oxygen isotope chronology from Williams et al. (1988).

Punchbowl, Hanauma Bay and Koko Crater. For example, after a substantial hiatus on the order of probably 1 Myr, a rejuvenation phase of volcanism produced 37 vents on eastern Oahu, termed the Honolulu Volcanics. Jackson and Wright (1970) first suggested that the generation of the Honolulu Volcanics might result from tensional stress introduced by the flexing of the lithosphere as the island rises over the Hawaiian Arch. Clague et al. (1982) show that from available data, the duration of the quiescent period decreases steadily from 2.5 Myr on Niihau to less than 0.4 Myr at Haleakala (Figure 64). They presume that because the volcanoes in the Hawaiian chain younger than the Koolau Volcano (<2.6 Ma), that is islands south of Oahu, are still subsiding "rapidly", a new mechanism is required to explain the occurrence of the Hawaiian secondary volcanics after the principle theolitic shield building stage had ended. However, Clague et al. (1982) did not offer a proposal. Later, Clague and Dalrymple (1989) reexamined the data and concluded that the data are consistent with the model proposed by Jackson and Wright (1970). Clague and Dalrymple (1989) suggest that

> "the alkalic 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 to 225 km to the east. Likewise, the Honolulu Volcanics

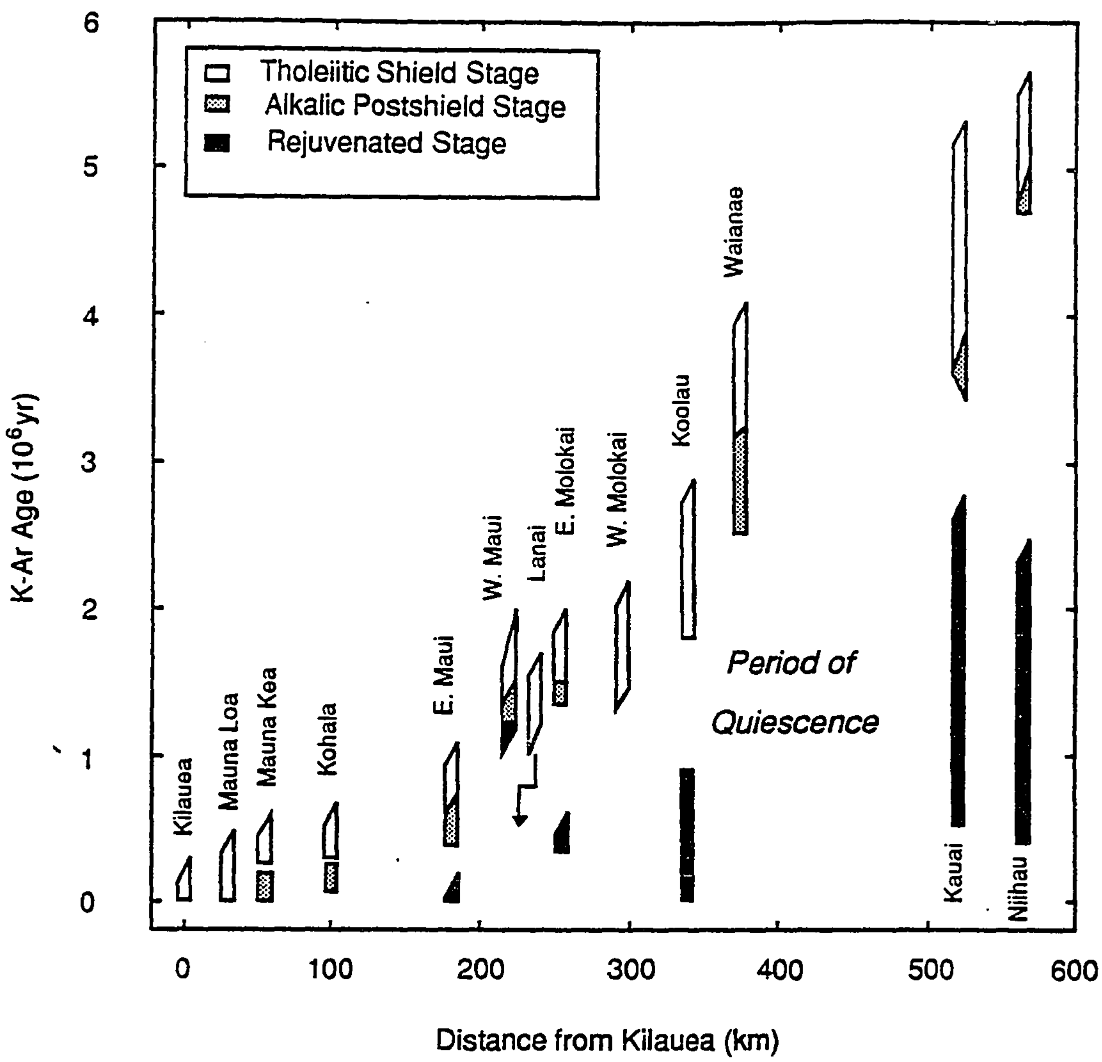


Figure 64. Duration of quiescence between various stages of volcanism - tholeiitic shield stage (open), alkalic postshield stage (shaded) and alkalic rejuvenated stage (dark) volcanism for dated volcanoes on the principle Hawaiian Islands. See Clague and Dalrymple (1987) for data sources.

on the Koolau Range of Oahu began erupting during formation of the East Maui shield located 160 km to the east. The rejuvenated stage Kalaupapa Volcanics on East Molokai erupted during formation of the Mauna Kea shield located 200 km to the east. Finally, the rejuvenated stage Hana Volcanics on East Maui began erupting during formation of the Mauna Loa shield located 160 km to the east. In each case, the rejuvenated stage lava began erupting during formation of a large shield 190 $\pm$ 30 km to the east" (Clague and Dalrymple, 1989,p. 202).

Furthermore, Clague and Dalrymple (1989) argue that northwest of the Hawaiian Islands, the rejuvenated stage postdated the shield-building stage by about 2-3 Ma because volcanic propagation rates were slower. They further speculate that the lack of huge volcanic edifices in the northern section of the Hawaiian chain indicates that the lithosphere was not sufficiently loaded for rejuvenation. This is reflected in the lack of rejuvenated stage lavas found in the northern part of the chain. Likewise, in the Emperor Seamounts, with relatively widely spaced seamounts, the building of the seamounts did not influence the neighboring seamount because of distances separating adjacent seamounts. Clague and Dalrymple (1989) also point out that the Emperor volcanoes were formed on younger and thinner lithosphere so the lithosphere was less rigid and, therefore, the distance from the load to the flexural arch would be shortened.

On Oahu, the age of several of the Honolulu Volcanic vents (31 ka - 1.0 Ma, Gramlich et al., 1971; Lanphere and Dalrymple, 1980) suggests that the secondary eruptive phase for this island broadly overlaps with the period of uplift as evident from the elevated reefs on Oahu described earlier in Chapter VII. However, only a quarter of the vents identified as post-erosional have been dated, and there is a lack of agreement between the two chronological studies on the sequence of venting and the age of the more recent vents. The alignment of the Honolulu Volcanics along fissures in a northeasterly trend is consistent with tensional release along faults in the Molokai fracture system (Walker, 1990). A similar pattern of northeast trending cones is observed in the Koloa Volcanics of Kauai, and young vents in the Waianae range on Oahu and on the Kalaupapa peninsula on Molokai (Walker, 1990). Until further refinement of the Honolulu Volcanics chronology, it is premature to speculate on interaction of volcanic rejuvenation and local uplift, but nevertheless, this could be an additional mechanism for localized uplift.

Implications for Other Hot Spot/Linear Chains

Several hot spot or linear chains of islands exist in the Pacific basin. McNutt and Menard (1978a) proposed a model of lithospheric flexure to explain uplifted atolls in the South Pacific (Table 36). They adopted an elastic plate model to

Table 36.
Uplifted Pacific Atolls attributed to lithospheric flexure.

Uplifted Atoll	Distance from Atoll to Loading Volcano (km)			Uplift (m)	
				Obs.	Pred.
Tuamoto Islands:	Mahetia	Tahiti			
Anaa	290	430		6	5.3
Niau	280	370		5	6.5
Kaukura	310	380		0	0.3
Makatea	250	250		70	71.4
Rangiroa	360	370		0	0
Tikehau	360	360		0	0.5
Matahiva	370	340		3.5	3.1
Cook Islands:	Raratonga	Aitutki	Manuae		
Atui	205	190	110	20	18
Mitiaro	245	230	140	27	27
Mauke	275	270	185	30	29
Mangaia	195	370	300	70	50

* Data from McNutt and Menard (1978a).

represent the lithosphere and calculated the wavelength and magnitude of flexure from several loading volcanoes. Volcanic size and distance between loading volcanoes and uplifted atolls was derived from hydrographic charts. The uplifted atolls were implied to represent a time-dependent horizontal surfaces. McNutt and Menard proposed that "the uplifted reefs at Oahu could be a product of loading by the volcanoes of Hawaii" (McNutt and Menard, 1978a: 1207).

Stearns (1979) had argued against the flexure model when it was first proposed by McNutt and Menard (1978a) to explain the elevated atolls or makateas in the South Pacific. He reasoned that because a trough runs parallel to the trend of the Hawaiian Islands and there is no bathymetric expression of the arch crossing the trench, that the "dimple theory" could not be used to explain the "well-documented marine fossiliferous beach conglomerates" from Lanai at elevations of 170, 190 and 324 m and thus the model did not operate for linear island chains. Stearns also questioned the lack of high emerged reefs on Kauai that should have resulted from the loading of the Waianae and Koolau volcanoes in the building of Oahu. He reiterated his conviction on the eustatic nature of the last interglacial Waimanalo stand on Oahu. McNutt and Menard (1979) replied that their model was not invoked to explain the emergent reefs on Lanai, but added that the absence of the arch crossing the Hawaiian trough does not necessarily imply that flexure is not important. Using a ratio of moat depth to arch amplitude, McNutt and Menard argued that the bathymetric expression would be difficult to trace even without sediment or volcanic debris infilling the moat. They also state that there is not enough water available to raise sea level globally to 170 m as Stearns had earlier hypothesized for an explanation of the high elevation

Lanai shorelines. McNutt and Menard in their calculations suggested that the nodal point of flexure from the loading of Hawaii was positioned at the southeast end of Oahu (McNutt and Menard, 1978b).

The study by McNutt and Menard (1978a) and follow-up study by Lambeck (1981a, b) relied heavily on makatea elevations picked from topographic maps with poor control. Spencer and co-workers (1987) provided high quality altitudinal data of selected geomorphic features from the Society Islands, north western Tuamotu Archipelago and southern Cook Islands to use in evaluating island uplift due to lithospheric flexure (Spencer et al. 1987).

The close match between the observed makatea elevations and the predicted elevations originally reported by McNutt and Menard (1978a) for a lithospheric flexure model may have been purely fortuitous. For the Cook Islands, McNutt and Menard (1978a) proposed that three volcanic centers, Aitutaki, Manuae, and Raratonga, loaded the plate and together were responsible for uplift of four neighboring atolls (Atui, Mitiaro, Mauke and Mangaia). For the Tuamotu Islands, two volcanic centers, Tahiti, and Mahetia, were considered to be responsible for loading the plate and uplifting the atolls of Matahiva, Tikehau, Makatea, Niau, and Anaa. The highest makatea is on Makatea which is reported to have uplift 70 m above sea level.

CHAPTER XI

CONCLUSIONS

The origin of elevated fossil coral deposits in the southeastern Hawaiian Islands has long been a subject of speculation and controversy. The first theory developed to explain the origin of elevated fossil coral deposits in Hawaii was Harold Stearn's view that the deposits represented relict high eustatic sea level stands throughout the Pleistocene. Another well-known theory is Moore and Moore's giant wave hypothesis which attributes the deposits to a tsunami. A major objective of this thesis is to test and critically evaluate these hypotheses by dating fossils from elevated deposits and reviewing and synthesizing the literature. A new model of lithospheric flexure is proposed to explain a pattern of differential uplift caused by loading of the Hawaiian hot spot.

A new dating technique, electron spin resonance (ESR), was used to date elevated coral deposits on the island of Lanai, Molokai, and Oahu. The basic pattern of increasing age of these deposits with increasing elevation on all three islands is interpreted as evidence for uplift. The mechanism invoked to explain this uplift is lithospheric flexure caused by crustal loading at the hot spot. For the past 300 kyr, the approximate rates of uplift calculated for Oahu, Molokai and Lanai are 0.06 m/ka, 0.10 m/ka, and 0.34 m/ka, respectively with the greatest uplift associated with the island of Lanai.

On Oahu, five new carbonate deposits were dated using ESR and mass spectrometric U-series methods. Lithofacies and ESR dating of last interglacial deposits at Barber's Point Canal suggest two high stands separated by a minor regressive period. Using the higher of these two terraces, the rate of uplift from 125 ka to present is calculated to be 0.048 m ka^{-1}. A reef from West Beach at an elevation of 14 m was dated at approximately 200 ka. The highest stand, the 30-m Kaena stand was dated at approximately 490 ka. Based on the Kaena deposit, uplift of Oahu from 500 ka to 125 ka is calculated to be 0.06 m ka^{-1}.

On Molokai, three geomorphic terraces are described and all appear to be related to high sea level stands. The youngest is a prominent fossil coral deposit at an elevation of +2 m exposed along the southern coast of Molokai. This terrace was radiocarbon dated at 3.31 $\pm$ 0.03 ka and correlates with the Kapapa stand on Oahu. A truncated eolianite and marine terrace at +10.5 m elevation is correlated to the last interglacial period (122-130 ka) when sea level worldwide was about 6 m above present. Calculation of uplift indicates a rate of 0.04 m ka^{-1} for the last 125 kyr. A third terrace at 30 m was tentatively dated at approximately 290 ka and is correlated to Stage 9 interglacial. The uplift rate for this terrace, assuming that the paleoelevation of Stage 9 was at today's sea level, is calculated to be 0.07 m ka^{-1}.

On Lanai, fossil corals from 12 - 74 m elevation from a variety of sites were ESR dated at approximately 207 $\pm$ 18 ka providing further evidence for the uplift of this island within the last 300 ka. Assuming a paleoelevation for Stage 7 sea level of 2 m above present level, a mean rate of uplift at 0.34 m ka^{-1} for the last 200 ka has been calculated. This is the highest uplift rate calculated for the three islands. The highest deposit on Lanai previously described from 326 m may be an archaeological artifact associated with an ancient Hawaiian trail leading from Manele Bay to the upland plateau. Boulder ridges previously ascribed to a giant wave bedform may also be archaeological in origin and may represent dryland agricultural terraces built by native Hawaiians.

Overall, the results of this study provide strong evidence in support of lithospheric flexure as the mechanism to explain uplift of the islands of Oahu, Molokai and Lanai. The distance from Mauna Loa to the areas of uplift on Oahu, Molokai and Lanai is consistent with results from two simple geophysical models for the wavelength of lithospheric flexure. The maximum height of the uplifted marine terraces approximately match the amplitude of the lithospheric bulge given by these two geophysical models. Other lines of evidence which support the lithospheric flexure hypothesis include (1) moat and arch geometry, (2) gravity anomalies, and (3) seismic studies of the crustal structure beneath the Hawaiian Ridge.

New data are also presented for a mid-Holocene paleo-shoreline on Kauai, providing unequivocal evidence for a Holocene sea level high stand in the Hawaiian Islands. A coral reef discovered in the Hanalei River on Kauai at 1.8 m above sea level was radiocarbon dated from 4.2 to 3.23 ka and suggests a higher than present sea level existed in the north central Pacific during the mid-Holocene. These results agree with a climatic maximum in the Holocene. They are also consistent with geological evidence at other oceanic sites in the Pacific, and recent geophysical numerical models of sea level rise during the Holocene deglaciation. The Hanalei reef suggest that Holocene sites on Oahu, especially the Kapapa stand, should be re-evaluated.

Harold Stearn's view that the emerged deposits represented purely eustatic sea level stands is inconsistent with records of paleo-sea level reconstructed from tectonically-elevated reef terraces from elsewhere (Papua New Guinea, Chappell, 1983; Barbados, Bender et al., 1979; Radtke and Grün, 1990; Suma Islands, Indonesia, Pirazzoli et al., 1991, 1993). Hence, the classical view of Oahu as a stable platform for eustatic sea level studies is not supported by results of this research.

The giant wave hypothesis is no longer tenable since there is lack of support for a single event and the sequence of ages and elevations indicate the islands have been uplifted.

Numerous other elevated fossil coral and dune deposits exist in the major Hawaiian Islands that have not been sampled or dated and considerable more research is needed before a more definitive model of lithospheric flexure can be developed.

Appendix 1:
DATES OF MARINE FOSSILS.
LISTED BY ISLAND WITH DATING METHOD AND SAMPLE ELEVATION.

Location	Material[1]	Method[2]	Elevation (m)	Age (ka)	Error (ka)	Source[3]
EMERGED DEPOSITS						
<u>Hawaii</u>						
Keawanui Bay	coral	T	6	110†	10	[23]
<u>Lanai</u>						
Kawaiu Gulch	coral clast	T	115	108	5	[13]
	coral clast	T	120	101	4	[13]
Kaluakapo Crater	coral clast	T	155	134	7	[13]
Manele?	coral?	T	171	>350		[16]
<u>Maui</u>						
No dated deposits						
<u>Molokai</u>						
Kawela	Porites	C		3.37	0.07	[24]
<u>Oahu</u>						
Kaena Point	coral	E	30	507		[1]
	coral	E	30	499		[1]
	coral	E	30	449		[1]
	coral	E	30	406		[1]
	coral	E	30	547		[1]
	coral	T	30	>225		[1]
	coral	T	30	>287		[1]
	shells	C	3.7-4.0	>50		[6,30]
Lualualei Valley	coral	U	30	600	100	[23]
	coral	T	30	>200		[23]
Kahe Point	coral	E	15-17	215		[1]
	coral	E	15-17	223		[1]
	coral	E	15-17	307		[1]
	coral	E	15-17	272		[1]
	coral	E	15-17	247		[1]
	coral	E	15-17	300		[1]
	coral	T	15-17	210	20	[1]
	coral	T	15-17	>243		[1]
	Porites	T	9.8-11.6	142	12	[3]
	Beach cong	T	15.8-18.9	>350		[9]
	shells	C	2.7-3.1	>40		[6,30]
Mokuauia Is.	Porites	T	0.5	120	9	[3]

Appendix 1. (Continued) DATES OF MARINE FOSSILS.

Fort Hase	coral	E	0.5	<10		[1]
	beachrock	E	0.5	<10		[1]
	P. meand.	E	1.3	147		[1]
Nu'upia Ekolu		E	1.6	140		[1]
Kaena State Park	Pocillopora	E	4.6	109		[1]
New Harbor	coral	E		112		[1]
	urchin sp	E		256		[1]
		T		230	20	[1]
Nanakuli		E	3.5	116		[1]
		T	3.5	136.6	16	[1]
	Leptastrea	T	3	120	30	[22]
Waimanalo	Porites	T	7.5	120		[8]
Honolulu Airport	Porites	T	2.8	135	6	[3]
Kahulu Pt.	shells	C	0.6	19.6	1.5	[18,19]
	shells	C	1.5	>50		[6,30]
	coral	C	1.5	>42		[6,30]
	snail	C	2.7	>50		[6,30]
	limpet	C	1.5	7.54	0.3	[6,30]
	Leptastrea	T	2	110	20	[22]
Mokapu	limestone	C	1.5	>37		[19]
	shells	C	3.6	>32		[19]
Upupau Head	Porites	T	4-5.8	122	11	[3]
	Pocillopora	T	1.5	118	9	[8]
	Pocillopora	P	1.5	105	11	[8]
Black Pt.	Porites	P	7.6	105	11	[8]
	Porites	P	7.6	105	15	[8]
	Porites	P	7.6	138	24	[8]
	Porites	P	6.7	94	10	[8]
	Porites	P	6.7	105	14	[8]
	Porites	P	7.6	130	22	[8]
	Porites	P	1.8	138	8	[8]
	Porites	T	2.2	124	9	[8]
	Porites	T	1.8	137	11	[8]
	Porites	T	7.6	128	8	[8]
	Porites	T	6.7	115	6	[8]
	Porites	T	6.7	118	7	[8]
	Porites	T	7.6	112		[8]
	Porites	T	7.6	121	7	[8]
	Porites	T	7.6	118	3	[8]
	Porites	T	7.6	115	6	[8]
	Porites	T	7.6	118	7	[8]
Kaikoo Place	cowry	E	0	105		[1]
		T	18	>385		[1]
		T	18	>400		[1]
	corals	T	0.6	125	10	[19]
Makai Range	shell	C	1.5	23.22	1.2	[19]
Leahi	coral	T	0.3	137	11	[19]
	coral	T	0.9	115	6	[19]
Hanauma Bay	corals	C	1.5	3.485	0.16	[19]
Popaia Is.	gastropod	C	1.5	39.1	1.5	[6,30]

Appendix 1. (Continued) DATES OF MARINE FOSSILS.

Popaia Is.	Leptastrea	T	1.5	110	20	[23]
Waipahu	oyst.	C	4.3	>38		[17]
	oyst.	C	4.3	>38		[17]
Waialua Bay	Cypraea	C	1.5	28.2	1.3	[5]
	Cypraea	C	1.5	24.14	0.8	[5,28,29]
	Cypraea	C	1.5	26.64	1.1	[5,29]
Waimea Bay	Cypraea	C	3.7	18.07	0.45	[5]
Waimea Bay	Cypraea	C	3.7	31.54	1.3	[5,28,29]
	Cypraea	C	3.7	31.84	1.0	[5,29]
	P. ret.	C	3.4-3.7	>40		[6,30]
Haleiwa	Leptastrea	T	1.5	140	30	[22]
	Leptastrea	U	1.5	140	50	[22]
Kawaihae St.	Porites	T	7.6	121	7	[8]
	Pocillopora	T	6.1	121	4	[8]
	Porites	P	8.1	120	13	[8]
Hawaii Loa Ridge	Porites	T	7.5	121	7	[8]
	Porites	T	7.5	118	3	[8]
	Porites	P	7.5	113	16	[8]
	Porites	P	7.5	124	19	[8]
Farrington Hwy	Pocillopora	T	3.7	115	6	[8]
KMC	Porites	T	7.8	131	8	[8]
	Pocillopora	T	11	134	7	[8]
RCA	Pocillopora	T	0.5	137	11	[8]
	Porites	T	1.0	115	6	[8]
Diamond Head	Porites	T	2.5	121	4	[8]
	Porites	T	2.5	96	9	[8]
	Porites	P	2.5	91	9	[8]
Diamond Head Rd.	Porites	T	0.1	128	8	[8]
	Porites	P	0.1	109	13	[8]
Pupukea Beach	C. tigris	C	2.1-2.4	>40		[6,30]
Puu O Hui Kai	Conus	C	3.5	18	0.6	[6,30]
Mokapu Pt	coral	T	11-12	135	4	[27]
	coral	T	11-12	125	3	[27]
<u>Kauai</u>						
Kauhou Valley	sandstone	C	0.9-1.8	15	0.6	[6]
Oomano Pt.	beachrock	C	1.0	1.6	0.16	[6]
Kelia	Tellina	C	1.1	3.93	0.17	[31]
	Tellina	C	0.8	3.85	0.22	[31]
	Leptachatina	C	1.6-1.3	2.65	0.17	[31]
	Tellina	C	1.2	4.04	0.18	[31]
SUBMERGED DEPOSITS						
<u>Hawaii</u>						
Ka Lae	coral	C	-305	10.73	0.07	[14]
	coral	C	-220	9.51	0.06	[14]
	coral	C	-155	10.46	0.06	[14]
	coral	C	-200	12.50	0.045	[14]
	coral	T	-160	13.90	0.3	[14]

Appendix 1. (Continued) DATES OF MARINE FOSSILS.

Hualalai	coral	C	-200	11.01	0.045	[14]
	coral	C	-175	12.03	0.08	[14]
Kealakekua Bay	limestone	C	-204	12.94	0.11	[12]
	limestone	C	-219	12.82	0.06	[12]
	limestone	C	-207	13.61	0.05	[12]
	coral	T	-207	15.80	0.5	[14]
	coral	T	-207	15.7	0.5	[14]
Northwest Hawaii	c.algae	T	-360	99	4	[21]
	c.algae	T	-360	120	5	[21]
	Porites	U	-150	17	5	[10]
	P. lobata	U	-150	19	5	[10]
Northwest Hawaii	Porites	T	-150	15.8	7	[10]
	P. lobata	T	-150	14	5	[10]
	P. lobata	U	-430	133	10	[10]
	P. lobata	T	-430	112	6	[10]
	P. lobata	U	-693	225	12	[10]
	Porites	U	-693	226	13	[10]
	P. lobata	U	-693	276	9	[10]
	P. lobata	T	-693	261	55	[10]
	Porites	T	-693	>260		[10]
	P. lobata	T	-693	177	15	[10]
	P. lobata	U	-945	314	10	[10]
	P. lobata	U	-945	287	10	[10]
	P. lobata	T	-945	337	79	[10]
	P. lobata	U	-1146	406	12	[10]
	P. lobata	U	-1146	360	12	[10]
	P. lobata	U	-1146	475	38	[10]
	P. lobata	T	-1146	>320		[10]
	P. lobata	U	-1336	463	8	[10]
<u>Lanai</u>						
No dated deposits						
<u>Maui</u>						
Haleakala Terrace	P. varians	U	-1555-1705	750	13	[15]
<u>Molokai</u>						
No dated deposits						
<u>Oahu</u>						
Hanauma Bay	coral	C	-2.0	3.64	1.2	[2]
	coral	C	-2.0	3.78	1.5	[2]
	coral	C	-5.0	4.24	1.5	[2]
	algae	C	-0.1	2.4	0.2	[4]
	Cyphastrea	C	-0.9	2.91	0.07	[4]
	algae	C	-2.1	3.98	0.13	[4]
	Porites	C	-2.1	3.64	0.12	[4]
	algae	C	-2.1	3.11	0.11	[4]
	Cyphastrea	C	-4.0	4.27	0.09	[4]
	Porites	C	-4.9	4.24	0.15	[4]

Appendix 1. (Continued) DATES OF MARINE FOSSILS.

Hanauma Bay	Porites	C	-6.8	5.3	0.11	[4]
	Pocillopora	C	-7.9	5.57	0.11	[4]
	Pocillopora	C	-8.5	5.44	0.13	[4]
	Porites	C	-0.3	1.74	0.07	[4]
	Pocillopora	C	-2.1	1.25	0.09	[4]
	beach rock	C	0.1	1.2	0.19	[4]
	Porites	C	0.1	1.79	0.11	[4]
	Porites	C	-1.8	3.78	0.15	[4]
	Porites	C	-1.8	3.53	0.08	[4]
	algae	C	-0.6	3.18	0.1	[4]
	algae	C	-1.5	2.51	0.09	[4]
	algae	C	-1.5	3.36	0.1	[4]
	algae	C	-3.3	3.96	0.1	[4]
	Porites	C	-4.2	4.82	0.12	[4]
	Porites	C	-8.2	5.57	0.11	[4]
	algae	C	-1.1	3.04	0.11	[4]
	algae	C	-1.1	2.91	0.08	[4]
	algae	C	-2.0	0.84	0.08	[4]
	Pocillopora	C	-2.9	3.84	0.12	[4]
	algae	C	-3.8	4.65	0.11	[4]
	algae	C	-5.3	4.9	0.15	[4]
	algae	C	-6.2	5.02	0.21	[4]
	algae	C	-6.9	5.18	0.14	[4]
	algae	C	-1.5	2.1	0.12	[4]
	algae	C	-1.5	2.27	0.1	[4]
	algae	C	-3.4	4.1	0.14	[4]
	Cyphastrea	C	-4.6	4.17	0.13	[4]
	Pocillopora	C	-8.1	7.01	0.12	[4]
	Porites	C	-9.3	6.97	0.1	[4]
	Porites	C	-9.9	6.9	0.17	[4]
	algae	C	-1.1	2.67	0.22	[4]
	algae	C	-1.1	3.17	0.12	[4]
	algae	C	-1.7	3.0	0.13	[4]
	algae	C	-1.7	2.96	0.1	[4]
	Cyphastrea	C	-7.2	5.0	0.12	[4]
	Cyphastrea	C	-7.2	4.82	0.1	[4]
	Porites	C	-8.1	5.6	0.11	[4]
	Porites	C	-8.1	5.87	0.16	[4]
	Porites	C	-8.1	6.06	0.13	[4]
	Porites	C	-8.1	6.05	0.15	[4]
	Porites	C	-8.1	5.57	0.22	[4]
	Porites	C	-8.7	6.08	0.14	[4]
	sand	C	-10.5	5.54	0.08	[4]
	algae	C	-0.4	0.49	0.18	[4]
	algae	C	-3.1	2.76	0.07	[4]
	algae	C	-6.2	4.49	0.2	[4]
	algae	C	-8.0	5.42	0.16	[4]
	Porites	C	-9.8	5.55	0.08	[4]
	Porites	C	-12.3	5.5	0.21	[4]
	Porites	C	-13.2	5.3	0.16	[4]

Appendix 1. (Continued) DATES OF MARINE FOSSILS.

Hanauma Bay	Porites	C	-0.4	0.48	0.1	[4]
	algae	C	-0.4	0.97	0.1	[4]
	algae	C	-0.4	1.05	0.1	[4]
	algae	C	-0.4	1.07	0.09	[4]
Kawainui Swamp		C	-1.0	7.05		[25]
	Eufenella	C	-2.80-3.05	2.17	0.2	[20]
	mud	C	-2.05-2.15	2.40	0.19	[20]
	peat	C	-0.90-1.1	0.17	0.12	[20]
Ukoa Pond	peat	C	-0.95-1.20	2.05	0.14	[31]
Punaoolapa	peat	C		30.8	1.1	[20]
Kahuku	peat	C	-1.05-1.30	5.97	0.2	[20]
Koko		C	-4.5	4.1		[2]
		C	-4.5	5.8		[4]
-500 m terrace	Leptastrea	T	-500	>200		[22]
Nu'upia Ekolu	P. comp.	E	-0.5	<10		[1]
Kahulu Pt.	shells	C	-0.3	21.6	0.08	[18,19]
<u>Kauai</u>						
Kaheko Reef	Porites	C	-18	8.37	0.25	[6,7]
Kaheko Reef	Porites	T	-18	8	1	[22]
SUBTERRANEAN DEPOSITS						
<u>Oahu</u>						
Ewa Core I	coral (?)	T	-15.7	140	40	[26]
Ewa Core I	coral (?)	U	-272	>700		[26]

Notes:
† 15% calcite see Stearns (1973).
[1]Materials are abbreviated as follows: c.algae is coralline algae; P. lobata is <u>Porites lobata</u>;. P. varians is <u>Pavona varians</u>; P. comp. is <u>Porites compressa</u>; Oyst. is oyster shell. P. meand. is <u>Porites meandrina</u>; Urchin sp is urchin spine; Beach cong. is beach conglomerate; C. tigris is <u>Conus tigris</u>; and P. ret is <u>Periglypta reticulata</u>.

[2]Methods are abbreviated as follows: C = radiocarbon; T = $^{230}Th/^{234}U$; U = $^{234}U/^{238}U$; P = $^{231}Pa/^{235}U$; and E = electron spin resonance.

[3]References are (1) Brückner and Radtke, 1989; (2) Easton, 1973; (3) Easton and Ku, 1981; (4) Easton and Olson, 1976; (5) Hubbs et al., 1962; (6) Hubbs et al., 1965; (7) Inman and Veeh, 1966; (8) Ku et al., 1974; (9) T.L. Ku cited in Stearns, 1978a; (10) Ludwig et al., 1991; (11) E. Matsumoto cited in Brückner and Radtke, 1989; (12) Moore and Fornari, 1984; (13) Moore and Moore, 1988; (14) Moore et al., 1990a; (15) Moore et al., 1990b; (16) J.K. Osmond cited in Stearns, 1966 (p. 23); (17) Rubin and Berthold, 1961; (18) Stearns, 1972; (19) Stearns, 1974; (20) Matsumoto and Kayanne, 1986; (21) Szabo and Moore, 1986; (22) Veeh, 1965, 1966; (23) H.H. Veeh cited in Macdonald et al., 1983 (p. 208); (24) Weisler, 1989; (25) J.C. Kraft in Brückner and Radtke, 1989; (26) Veeh in Hammond, 1970; (27) Muhs and Szabo, 1991; (28) Shepard, 1961;

Appendix 1. (Continued) DATES OF MARINE FOSSILS.

References (continued) (29) Shepard, 1963; (30) Shepard and Curray, 1967; and (31) Matsumoto and Kayanne, 1988.

Appendix 2.
Archaeological Features Associated with Hawaiian Trail Leading from Manele Bay Inland along Ridge Crest (adapted from Kaschko and Athens, 1987).

Site No.-Feature	Description
1501	A light surface midden scatter (approximately 30 by 25 m) is situated on the high rocky slope of a ridge line overlooking Manele Bay at an elevation of about 183 m. The site area contains low bedrock outcrops and exposed boulders, is subject to some slope wash erosion, and has a low cover of unidentified grasses with klu (Acacia farnesiana). Visible on the surface is a very light scatter of shells including Cellana, Conus, Cypraea, Thais, and coral fragments. The site datum marker is located about 5 m upslope from a large round boulder. This feature is apparently associated with the prehistoric trail route extending inland up this general ridge line from Site-1524.
1524	An inland-seaward, ridge top trail route with associated shelter structures and other remains (approximately 50 by 495 m in area) includes 14 features and spans a range in elevation from about 26 to 122 m. It is situated on the top and front slope of a prominent ridge line, whose surface is a rock-strewn moderate slope with exposed bedrock boulders and outcrops. The vegetation in the site area varies somewhat but generally consists of a moderate cover of kiawe (Prosopis pallida), klu (Acacia farnesiana), koa haole (Leucaena leucocephala), Hawaiian cotton (Gossypium sandvicense), and dry grasses. Sparse or light shell midden remains are scattered intermittently on the surface along the route of the trail depression. The site datum marker was positioned in the east side of the Feature A trail depression, adjacent to the location of Feature B. Associated with temporary use shelter structures for the most part, this prehistoric trail route probably provided access from the coast to inland agricultural areas.

1524-A The eroded, inland-seaward trail (roughly 1 m wide, 25 cm average depth) runs on the crest of the ridge line and appears as a rocky, washed-out depression with some boulders along the sides. A Cellana and a Cypraea shell are visible adjacent to the trail near the datum location.
This prehistoric trail route continues inland as a fairly definite, intermittent depression all the way to Site 1501, and presumably went much beyond.

1524-B This small, low and irregular, roughly oval shelter terrace (4.8 by 3.2 m, up to 45 cm high) has a collapsed C-shaped wall on its inland side, and is constructed of roughly stacked boulders and some cobbles. The sparse cultural remains visible on the surface include Conus, Cypraea, Thais, a water-worn basalt pebble, and flaked basalt. Some subsurface cultural deposit is likely within the interior soil floor area.
The feature is probably a temporary shelter structure associated with the ridge top trail route.

1524-C A rough and collapsed probable C-shaped shelter structure (2.6 by 2.5 m, up to 40 cm high) is just on the east side of the trail depression. It is composed of roughly stacked boulders and some cobbles and built against a bedrock boulder outcrop. The rather light surface cultural materials include Cypraea, Cellana, Nerita, Conus, coral fragments, a water-worn cobble and boulder, flaked basalt, and some natural volcanic glass nodules. Some cultural deposit is probable in the interior area under the wall collapse.
The structure is apparently a temporary shelter along the trail route.

1524-D This small, low, oval shelter terrace (2.5 by 2.1 m, up to 50 cm high) consists of roughly stacked boulders and a few cobbles. As it is located just 4 m to the northeast of Feature C, the surface cultural remains are the same as noted for that feature.
About 5 m upslope of this structure is a possible second small collapsed oval terrace with a water-worn boulder in the middle and a possible collapsed "cupboard" to its seaward side.

1524-E A small rockshelter with rough closing wall (sheltered area 1.8 by 1.2, 70 cm high; wall 2.1 m long, 40 cm high; natural terrace 5.2 by 4.0 m) is located on the steep rocky west slope of the large ridge line. A roughly stacked, low boulder wall is in front of and partly closes the small "cave" opening or cavity, which is in a fairly soft and clinkery bedrock ledge. Only a few water-worn basalt cobbles are visible on the surface. Some subsurface cultural deposit is quite likely in the soil floor interior of the cavity or the area in front of it.
This small cavity could have been cut into or enlarged in the rater soft cinder bedrock ledge, and may have been used as a temporary shelter or cache location, or possibly for burial purposes.

1524-F,G,H,I This group of four adjacent features forms a relatively discrete unit just to the west side of the trail depression running on the crest of the ridge. In general these features are constructed of roughly stacked boulders and cobbles with much use of exposed bedrock, and many portions appear quite collapsed. The moderate amount of surface cultural materials visible in the area includes _Conus_, _Cypraea_, _Drupa_, _Cellana_, _Nerita_, _Thais_, large _Triton_, coral fragments, water-worn basalt cobbles and pebbles, carbonate sandstone, a water-worn boulder, flaked basalt, a water-worn cobble hammerstone and some possible flaked volcanic glass with natural weathered glass nodules. These is a probablility for the presence of subsurface cultural deposits, especially in Feature I as well as G and H.

1524-F	This feature consists of three or four adjacent and connected small rough oval shelter terraces (10 by 5 m, up to 40 cm high).
1524-G	This feature is roughly rectangular, two level probable habitation terrace (9 by 6 m, up to 1.20 m high). About 8 m east of this structure is a section of roughly placed steppingstones in the trail.
1524-H	This feature is probably the collapsed remains of a low, rough, rectangular walled terrace with two floor units (7 by 3 m, up to 65 cm high).
1524-I	This feature is composed of a rectangular rock-filled terrace (6.0 by 2.6 m, up to 1.0 m high) with about five small rough oval shelter terraces adjacent just upslope (feature overall 13 by 12 m shelters up to 80 cm high). To the west side are two cairns constructed on top of bedrock boulders. The rectangular terrace has a facing of large upright boulders and is rock filled on the interior, and may possibly be a shrine or burial structure.
1524-J	A collapsed, roughly rectangular platform (6.2 by 5.1 m, up to 70 cm high) on the west side of the trail depression is composed on roughly stacked boulders and cobbles and includes some bedrock boulders. The sparse surface cultural remains consist of only a _Cypraea_ and a large _Triton_, one coral fragment, and a water-worn basalt cobble. The structure may be a collapsed burial platform or possible a trail-side shrine

1524-K A collapsed roughly rectangular terrace with two adjacent small, rough, oval shelter terraces (10 by 13 m; rectangular terrace 8.0 by 4.0 m, up to 1.15 m high) are built among and on the west side of outcropping bedrock boulders, and are constructed of roughly stacked boulders and cobbles. The moderate cultural materials visible on the surface include _Cypraea_, _Cellana_, _Nerita_, _Drupa_, _Conus_, _Thais_, fish bone, coral fragments, water-worn cobbles, flaked basalt, flaked volcanic glass, and a basalt core hammerstone. The presence of a subsurface cultural deposit is very probably within the rectangular terrace. The trail depression runs on the east side of the bedrock boulder outcrop, about 5 m from the feature. This feature probably consists of temporary use shelter structures associated with the ridge top trail route, although the larger rectangular terrace may be somewhat more substantial.

1524-L This small oval shelter terrace with C-shaped side wall (3.0 by 2.2 m, up to 80 cm high) is situated at the top of the west slope of the ridge line. It is composed of roughly stacked boulders and is built against a bedrock boulder. The sparse surface cultural remains include _Cellana_, _Drupa_, _Conus_, _Littorina_, _Nerita_, _Cypraea_, and flaked basalt. The feature is most likely a temporary use shelter structure. Another possible collapsed shelter terraces is located about 6 m to the northeast.

1524-M An irregular oval walled enclosure (21 by 13 m, up to 1.65 m high) includes two shelter compartments within its east side and a probable collapsed terrace in the northwest interior. It is constructed of rather roughly stacked boulders and some cobbles and is collapsed in parts. The trail depression appears to enter the enclosure through a break in the inland wall. The moderate to heavy cultural materials visible on the surface include _Cypraea_, _Drupa_, _Nerita_, _Tellina_, _Cellana_, _Thais_, _Conus_, _Nerita polita_, pearl shell, coral fragments, water-worn pebbles and cobbles and boulders, carbonate sandstone, flaked basalt, flaked volcanic glass, a coral abrader, and a fragment of a carbonate sandstone konane board. Subsurface cultural deposits are definite and potentially substantial in the area of this feature.
The enclosure wall may possibly be the remains of a historic animal pen, perhaps related to Feature D of Site 157 downslope, that was build on top of earlier prehistoric habitation structures.

1524-N The probable remnants of two or three rough and collapsed rectangular terraces (one is about 7 by 4 m, up to 60 cm high) are located downslope and outside of the Feature M enclosure wall. As the distribution of surface midden, etc. is continuous, the types of cultural materials present are largely the same as noted for Feature M. A subsurface cultural deposit is likely.
This feature consists of the collapsed remains of probable prehistoric habitation structures.

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