MORPHOLOGY, SEISMIC STRATIGRAPHY, AND FLEXURE MODELING OF SELECTED GUYOTS IN THE MARSHALL ISLANDS

A DISSERTATION SUBMITTED TO THE GRADUATE DIVISION OF THE UNIVERSITY OF HAWAII IN PARTIAL FULFILLMENT OF THE REQUIREMENTS FOR THE DEGREE OF

DOCTOR OF PHILOSOPHY

IN

GEOLOGY AND GEOPHYSICS

DECEMBER 1993

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ACKNOWLEDGEMENTS

Funding for the research presented in this dissertation came from a number of sources, including the National Science Foundation, the Joint Oceanographic Institutes - U. S. Science Advisory Committee, the Ocean Drilling Program, and Mokes on Spokes. Many of the lithologic and age relationships discussed in Chapter 3 result directly from discussions with participants on ODP Leg 144, and the data collected during this cruise will continue to shape our understanding of the Marshall Islands in the years to come. I look forward to collaborating with this group in the future.

On a more personal note, I am indebted to the late William Coulbourn who first brought me out to Hawaii to study submarine canyons off the coast of Peru and Chile. Having been introduced to the complexities associated with sediment transport along active margins, I decided to broaden my experience by dabbling in the realm of mid-plate volcanism and shallow-water carbonate accumulation. Guidance for this undertaking was provided by Fred Duennebier, who proved to be as excellent a dissertation advisor as he is a poker player. The whole Duennebier clan has made my stay in Hawaii all that more enjoyable. I also extend a heartfelt "mahalo" to my committee members, whose insightful reviews and timely comments kept my neophyte wanderings in check most of the time: John Sinton, Roy Wilkens, Sandy Shor, Paul Wessel, and Rick Grigg. Finally, to the many colleagues, office-mates, and Ultimate players I've met over the past few years, and especially to Barbara, "thank you" for your friendship and companionship. I hope our paths continue to cross in the years to come.
ABSTRACT

The Marshall Islands, located in the western Pacific, record the history of Cretaceous volcanism and shallow-water carbonate platform development. Common guyot features include flank ridges, terraces, flank channels, and perimeter ridges and cones across flat summit plateaus. Analogous features observed on modern-day volcanic islands and atolls include volcanic rift zones and remnants of landslides (flank ridges), fault-blocks (terraces), and reef tracts (perimeter ridges). On those guyots with drowned carbonate platforms (e.g., Wodejebato and Limalok), sediment units typically onlap a central basement high. On those guyots without carbonate platforms (e.g., Lo-En) basement topography disrupts the flat summit plateau.

Edifice size and plate flexure from nearby volcanic eruptions influence the morphologic and lithologic differences observed on Limalok, Lo-En, and Wodejebato. Drowned edifices are smaller than their attached atoll pairs, suggesting that truncation depth during sea level low-stands affects carbonate platform survival. Simple flexure models suggest Late Cretaceous volcanism at Anewetak caused net subsidence (~100 m) across the mid-Cretaceous platform of Lo-En, thereby inhibiting coral colonization. Flexure models for Pikinni and Mili suggest that the atoll-guyot volcanic platforms were constructed at approximately the same time because the guyots do not tilt towards that atolls.

Plate rotations show that the Marshall Islands were located near French Polynesia during their construction. Misfits between Cretaceous hot spot tracks and the observed Marshall Islands topography and ages result from: 1) more hot spots existing in French Polynesia than currently observed, or 2) wander of French Polynesia hot spots relative to other Pacific hot spots. Model observations include:
1. The Ralik chain formed in the Late Cretaceous, but not over any of the existing French Polynesian hot spots.

2. Macdonald hot spot must wander northeast ~4 mm/yr to align with the observed mid-Cretaceous edifices.

3. The Ratak chain formed either from the Rurutu or Mehetia hot spots.

4. The Anewetak-Ujlan volcanic cluster correlates with the Rarotonga hot spot and a change in plate motion.

From the modeling it may be possible to define a pre-100 Ma stage pole for Pacific plate motion.
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CHAPTER 1
INTRODUCTION

One of vexing problems still facing earth scientists today is the interpretation of features and the comprehension of events recorded across older portions of the Pacific basin. Since the inception of this research project back in 1988, one of its primary goals has been to improve understanding of the sequences of volcanism, uplift, and carbonate platform growth occurring across the western Pacific during the Cretaceous. As the seamounts, guyots, and atolls composing the Marshall Islands record episodes of volcanism and tectonism during this time period, expanding the data base over this island group was a logical step towards addressing this objective. Such was the justification for R/V *Moana Wave* cruise MW8805.

In a temporal sense, the Marshall Islands fall within a period of increased ocean crust production and extensive volcanism occurring across a very large portion of the Pacific plate [e.g., Menard, 1964; Schlanger et al., 1987; Larson; 1991]. Menard [1964] named the area of shallower than expected seafloor in the Pacific basin the Darwin Rise, and postulated that the volcanism recorded in the form of seamounts, islands, atolls, and mid-ocean plateaus resulted from fracturing of the plate over a large mantle plume (Figure 1.1.). Schlanger [1981] shared a similar view of long-term (40 m.y.) sea floor uplift in his interpretation of sediments recovered from drilling in the Nauru Basin, although other schools of thought resulting from deep-sea drilling across the Pacific included regional plate uplift but with two sequences of volcanism separated by a period of relative quiescence [e.g., Rea and Vallier, 1983], and local plate uplift associated with a more conventional mechanism of volcanic activity (e.g., hot spots). The latter hypothesis formed the core of the MW8805 surveying and sampling prospectus, and influential in shaping the studies presented in this dissertation.
Prior to MW8805, data across the Marshall Islands were limited to a couple of old U.S.G.S. drill holes on Anewetak and Pikinni atolls, a couple of Deep Sea Drilling Program holes in basins to the west and east of the Marshall Islands (Sites 462 and 169, respectively), and isolated dredge samples across a few edifices separated by considerable distances (e.g., von Valtier, Wodejebato, Lokkworkwor, and Erikub). Since MW8805, this region has received considerable more attention in the form of various surveying and sampling cruises (e.g., the U.S.G.S. F10-89-CP cruise, Hein et al., 1990; the University of Hawaii MW9009 cruise, Bergersen and Smith, 1990; the Scripps Institute of Oceanography TUNE06 cruise; Staudigel et al., 1992), including a pair of Ocean Drilling Program legs (Legs 143 and 144; Winterer, Sager et al., 1993; Haggerty, Premoli-Silva, 1993). The end result of these cruises has been a veritable explosion of information concerning the gross morphology, internal structure, and age of edifices in this region, and a vastly enlarged set of data upon which new interpretations for the Cretaceous history of volcanism, tectonism, and sedimentation are emerging. In this dissertation, a considerable amount of space is devoted to consolidating relatively recent observations into a reference frame ammenable to other events occuring across the Pacific.

The diversity of morphologic, stratigraphic, and petrologic relationships observed across edifices in the Marshall Islands during MW8805 foreshadowed the complications associated with studying this area of the Pacific. Within a relatively confined area one might find a living, thriving atoll attached to a 1400 m deep drowned carbonate platform, which in turn might lie adjacent to shallower edifice whose summit plateau appears barren of shallow-water carbonate sediments. Identifying simple relationships between island groups, much less features common to all islands, appeared to be an imposing task. Towards this end though, the analysis of Wodejebato Guyot presented in Chapter 2 was undertaken to help understand how a typical volcanic and carbonate platform within the Marshall Islands evolved through time.
The insights gathered from the analysis on the gross morphology and internal structure of this edifice proved useful both for comparative purposes with other guyots in the Marshall Islands and for identifying drill sites on Wodejebato during Leg 144. In Chapter 3, the three edifices drilled during Leg 144 in the Marshall Islands (Limalok, Lo-En, and Wodejebato guyots) are compared to one another and to possible modern-day counterparts in French Polynesia. In addition to these morphologic and lithologic comparisons, an analysis of how plate subsidence, edifice size, and plate flexure may have influenced summit plateau morphology and ultimately platform drowning is presented in this chapter.

Throughout the dissertation, similarities between edifices in the Marshall Islands and those in French Polynesia become more obvious. Indeed, plate rotations show that the region of the Pacific plate upon which the Marshall Islands are built was in the vicinity of French Polynesia during the Cretaceous. Previous attempts at identifying hot spot trends through the Marshall Islands have focused on tracking individual edifices back to their presumed hot spot source, with limited success. In Chapter 4 Cretaceous hot spot trends through the Marshall Islands are analyzed by plotting the topography and available ages for this region in oblique mercator projections oriented about stage poles defining Pacific plate motion. While the majority of edifices in the Marshall Islands can be explained by hot spots in French Polynesia, the projected topography and edifice age relationships relative to modeled hot spot tracks raise some important questions concerning the duration, number, and fixity of hot spots in the southeast Pacific. Finally, the implications of the models presented in this dissertation for future work in the Marshall Islands are discussed within the Conclusions chapter.
CHAPTER 2
GEOLOGY AND GEOMORPHOLOGY OF WODEJEBATO (SYLVANIA) GUYOT

2.1. Abstract

For much of his life, Sy Schlanger's work centered around understanding the affects of Cretaceous volcanism on such global phenomena as eustatic changes in sea-level and ocean-wide anoxia events. The Marshall Islands, which typify the complex history of volcanism in the western Pacific Ocean, pose a number of vexing tectonic and sedimentologic questions essential to our understanding of Cretaceous events in the Pacific. Within this island group, clusters of atolls and guyots lie in roughly linear chains, but the transposition of edifice ages in the direction of plate motion argues against a simple hot spot model of formation. An important step towards deciphering the geologic history of these island chains, and this region as a whole, is to more fully understand the history of their individual components. In this paper, the morphology and internal structure of Wodejebato Guyot (formerly Sylvania) is examined and the constraints these observations place on the history of this drowned atoll are discussed.

Wodejebato Guyot lies ~74 km northwest of Pikinni (formerly Bikini) Atoll. A volcanic ridge 8.7 km wide and 1600 m deep attaches the two edifices. Shallow-water, Cretaceous carbonate sediments overlie volcanic basement, and in turn are overlain by 100 m of pelagic sediment. Along the edge of the summit plateau, the carbonate platform crops out from the pelagic sediments. Acoustically massive perimeter ridges, interpreted as reefs, bound the presumed lagoon sediments along the east, north and west flanks. Along the south flank, a distinct 100 m to 1000 m wide terrace lies outside of the carbonate platform. The absence of a perimeter ridge on this flank and the strong correlation between the stratigraphy of the platform sediments and that of the terrace sediments in seismic profiles suggests large-scale block-faulting affects this area on Wodejebato. High backscatter
channels, presumably of debris-flow origin, mar the flanks of the guyot outside the carbonate platform.

Given the morphologic and stratigraphic observations presented in this paper, models examining the evolution of Wodejebato must take into account not only the existing radiometric and fossil evidence for two episodes of uplift, but also such information as the depth of volcanic basement along the guyot's flanks and beneath the carbonate cap, the presence of perimeter ridges along the west and north flanks, and the presumed block-faulting along the south flank. Volcanic basement lies at a depth of ~1600 m along most of the guyot flanks and an isopach map of carbonate platform thickness shows that basement shallows to the northeast across the summit plateau. Paired perimeter ridges along the north and northeast flank ridges suggest more than one episode of platform growth. Modification of the guyot flanks through block-faulting is apparently a process which continues to occur well after the edifice has moved away from hot spot swell.

2.2. Introduction

For much of his life, Sy Schlanger's work centered around trying to understand the world-wide distribution of Cretaceous volcanic features and events, including episodes of volcanism and carbonate platform growth in the Marshall Islands. Beginning early in his career with the U.S.G.S. sponsored Operation Crossroads, through a number of ship-based sampling and surveying programs, and culminating with his leadership in securing drilling time with the Ocean Drilling Program, Sy attempted to explain how Marshall Islands volcanic events (and those of a more regional scale) related to such global phenomena as eustatic sea level rises and ocean-wide anoxia events [e.g. Schlanger and Premoli Silva, 1981; Schlanger et al., 1981; Schlanger et al., 1984; Schlanger and Moberly, 1985]. As part of his ever-diligent efforts to expand the data base upon which the regional history of the western Pacific is built, Sy obtained funding to conduct a
detailed surveying and sampling program across selected guyots in the Marshall Islands. This program, conducted during a R/V Moana Wave cruise (MW8805), consisted of side-scan sonar and swath bathymetry mapping, 3.5 kHz and seismic profiling, gravity and magnetic measurements, and dredging. A subsequent cruise aboard the R/V Moana Wave (MW9009) collected additional 6-channel seismic data over those guyots selected as potential drill sites. In anticipation of drilling on ODP Leg 144, this paper presents a portion of the data collected over Wodejebotso Guyot (formerly Sylvania), and from these data proposes a model outlining the geologic history of this drowned atoll.

2.3. Background

The Marshall Islands, located between 5° and 18° N latitude and 160° and 170° E longitude, represent a complex aggregate of mid-plate volcanoes comprising at least 3 separate island chains generally trending in a northwest direction (Figure 2.1). The multiple episodes of mid-plate volcanism in and around the Marshall Islands indicated by drilling results [e.g. Emery et al., 1954; Winterer, Ewing et al., 1973; Larson and Schlanger, 1981; Moberly, Schlanger et al., 1986] are difficult to delineate by such simple physical traces as hot spot chains. Radiometric ages of basalt samples from various guyots in the Marshall Islands emphasize the complications associated with identifying hot spot chains (Figure 2.1). Some of this scatter may result from sampling biases or analytical errors, but a simple age-progression of volcanoes in the direction of plate motion is not readily apparent. For example, even within the geographically well-defined Ratak Chain, seamount ages fail to exhibit a linear correlation to seamount position [Davis et al., 1989]. In this chain, Woden-Kopakut Guyot (formerly Ratak), the younger edifice, lies "down-drift" from Lokkworkwor Guyot (formerly Erikub).

Radiometric ages of basalt samples collected from the various guyots provide, at present, evidence for a single episode of constructional volcanism for each edifice.
Figure 2.1. Regional bathymetry of the Marshall Islands. Guyots are labeled by their Marshallese name, and only those guyots with radiometric ages are listed. Radiometric ages come from Pringle [1992], Davis et al. [1989], and Moberly et al. [1986]. Figure revised from Hein et al. [1990].
Additional evidence concerning the history of volcanism and carbonate platform growth on these islands comes in the form of dredged limestone. Early dredges on Wodejebato Guyot (the "MP" labeled dredges shown in Figure 2.2) recovered only basaltic rocks and tuff breccia [Hamilton and Rex, 1959; Duennebier and Peterson, 1982]. Eocene foraminifers fill the cracks of these rocks. This information, along with the Early Eocene limestone recovered from Limalok Guyot (formerly Harrie) to the southeast and the results from drilling on Anewetak Atoll to the west, led Schlanger et al. [1987] to propose that the latest stage of volcanism in the Marshall Islands was Eocene in age.

During MW8805, four more dredges attempted across Wodejebato recovered basalt, basalt breccia, basalt conglomerate, and limestone (dredges labeled "RD" in Figure 2.2). The fossil content of the limestone indicates two distinct periods of carbonate platform growth across many of the summits in the Marshall Islands [Lincoln et al., 1989]. On Wodejebato Guyot, RD50 recovered rudist fragments within a Late Cretaceous foraminifer matrix. The rudists, Albian in age, and the foraminifer matrix, Campanian to Maastrichtian in age, suggest two sequences of shallow-water carbonate accumulation separated by approximately 30 m.y. [Lincoln, 1990]. Radiometric dating of a basalt from dredge MP43-D along the north flank of Wodejebato places the latest stage of volcanism at ~86 Ma [Pringle, 1992]. Thus, while the fossil assemblage in the limestone suggests that this edifice existed during the mid-Cretaceous, the basalt date reveals an episode of volcanism at least 20 m.y. later. Evidence for similar episodes of volcanism and carbonate platform growth exists on Lo-En Guyot (formerly Hess) and Lobbadege Guyot west and northwest of Wodejebato, respectively [Lincoln, 1990]. According to Lincoln [1990] the history of volcanism in this area of the Pacific extends back to the late Jurassic/early Cretaceous with secondary phases during Albian/Cenomanian time (105 Ma to 95 Ma), Santonian/Campanian time (85 Ma to 75 Ma), and possibly into the late Paleocene/early Eocene (60 Ma to 50 Ma).
Figure 2.2. Contoured SeaMARC II bathymetry over Wodejebato Guyot. Dredge locations are shown by triangles. The dredges represent work done during two cruises: the RD labeled dredges were collected during MW8805 [Schlanger and Duennebier, 1988] and the MP labeled dredges were collected during the Mid-Pacific Expedition [Hamilton and Rex, 1959].
Contour Interval = 100 m

Lithified pelagic sediments
Limestone
Breccia (basalt)
Conglomerate (basalt)
Basalt
Given that volcanic overprinting of the initial sea floor spreading event occurred in this area, it's likely that the volcano summits were subject to a complex history of subaerial exposure in response to local plate flexure from volcanic loading, regional plate flexure from mantle upwelling (the hot spot swell), and eustatic rises and falls in sea level. Two models can explain the age data and inferred tectonic and carbonate platform accumulation history for these guyots. The models differ by the degree to which a second episode of uplift affected the existing carbonate cap.

1. In the first model, uplift during the Late Cretaceous resulted in erosion of the mid-Cretaceous carbonate platform and transportation of debris into a younger platform growing primarily outside of the existing complex [Lincoln, 1990]. In this model, the older carbonate platform, lying topographically higher than the younger platform, should presumably show signs of subaerial exposure (e.g., meteoric oxygen isotope values or pronounced karst topography).

2. In the second model, substantial erosion of the older carbonate platform during the Late Cretaceous uplift left a thin veneer of carbonate sediments across the summit. The younger carbonate platform recolonized the existing topographic highs, burying the remnants of the mid-Cretaceous complex.

In light of these models, it is worthwhile to examine how the surface morphology and subsurface structure of a typical drowned atoll in this area constrain the proposed episodes of volcanism, uplift, and carbonate platform growth and erosion. The original subsurface mapping of Wodejebato Guyot was based on a limited number of digital single-channel seismic lines crossing the summit [Figures 8 and 9 of Schlanger et al., 1987]. From Operation Crossroads, estimates of Pikinni's basement depth range between 1300 m and 1600 m [Raitt, 1954; Alldredge et al., 1954]. The single-channel profiles across
Wodejebato show several acoustic units typical of a completely developed atoll, specifically "a high massive rim enclosing bedded lagoonal deposits" [Schlanger et al., 1987], but the paucity of both dredge data and geophysical data across this guyot precluded any detailed geologic interpretations. With the wealth of new information collected during MW8805 and MW9009, this paper provides detailed geomorphic descriptions of Wodejebato Guyot using side-scan images, swath bathymetry, 3.5 kHz profiles, selected seismic lines, and dredge data. These new data apply important constraints to the geologic history of this drowned atoll by showing how events after the Late Cretaceous uplift are responsible for most of the existing morphology.

2.4. Methods

Data used in this paper include side-scan images, swath bathymetry, 3.5 kHz profiles, single- and 6-channel digital seismic data, dredge descriptions, and radiometric and fossil ages. Swath mapping of the sea floor was accomplished with SeaMARC II, a shallow-towed (50 to 150 m) side-scan sonar mapping system which collected acoustic backscatter data in swaths up to 10 km wide and bathymetric data with a swath width equal to 3.4 times the water depth. The system operated at frequencies of 11 kHz on the port side and 12 kHz on the starboard side. A detailed description of the system can be found in Blackinton [1986]. At typical survey speeds of about 8 kts, over 3000 km² of sea floor could be surveyed per day.

Side-scan resolution is a function of many variables and is both range- and orientation-dependent. The resolution of the SeaMARC II system is on the order of several tens of meters [Johnson and Helferty, 1990]. Corrections applied to the backscatter data compensate for errors in slant range (based on water depth below the tow vehicle), beam pattern, bottom tracking, and gain variations. The data are then displayed as gray-scale images. Individual image swaths for the various tracks are assembled into a mosaic of the
survey area. Dark gray to black tones represent areas of high backscatter and shades of light gray to white represent areas of low backscatter and acoustic shadows. During MW8805 the starboard transducer arrays malfunctioned, causing clipping of the data at both the high and low ends of the recording spectrum and resulting in images with few data values represented by mid-range gray tones. Further distortion of the images occurs by the application of a "flat-bottom" assumption (i.e. the sea floor is flat beneath the tow-vehicle) prior to data recording. The result is an incorrect geographical positioning of features where the bottom topography slopes in relation to the tow-vehicle track (e.g., when surveying parallel to the slope of the guyot, features on the up-dip side of the tow-vehicle are positioned anomalously far from the ship track, and vice-versa).

The measurement of phase differences between incoming acoustic signals at a pair of transducer arrays mounted on either side of the SeaMARC II tow vehicle helps determine bathymetry. Comparison of the resulting acoustic angle to a "lookup" table (generated from data collected over flat sea floor) results in an estimate of depth. Bathymetric resolution is about 2% of the water depth [Blackinton, 1986].

A single-channel streamer using a Masscomp-HIGHRES digitizing system collected digital seismic data during MW8805. When the SeaMARC II tow-vehicle was in the water, the source was a 120 cubic inch air gun; at other times, an 80 cubic inch water gun was used. Processing of the data using the SIOSEIS software package included bandpass filtering from 10 to 100 Hz, automatic gain control, deconvolution, and muting of water noise. During MW9009, a 6-channel streamer was used with the Masscomp system. In addition to the processing routines applied to the MW8805 data, the 6-channel data were gathered, stacked, and migrated.

The preliminary dredge descriptions made during MW8805 were augmented by more detailed fossil descriptions and thin-section analyses [Lincoln, 1990]. Radiometric dating was attempted on basalts recovered during the Mid-Pacific Expedition [Hamilton and
Rex, 1959] and MW8805. Fossil ages presented in this paper are from Lincoln [1990], while radiometric ages are from Pringle [1992].

2.5. Geomorphic Description

Early measurements on Pikinni Atoll and Wodejebato Guyot come from the late 1940's Operation Crossroads [Emery et al., 1954]. Pikinni Atoll, 42 km long from east to west and 24 km wide from north to south, lies around 12° N and 165° E. Its 26 islands, elongated in a northwest direction, encompass approximately 440 km². Subsurface information on Pikinni comes primarily from the four holes bored on the northeast side of the atoll. The deepest of these holes penetrated 780 m of shallow-water carbonate sediments of Early Miocene to Oligocene age [Cole, 1954; Emery et al., 1954]; none of the holes encountered basalt. As the subsidence histories of Pikinni and Anewetak Atoll, lying over 300 km to the west, were considered coeval based on the correlation of their stratigraphy [Schlanger, 1963], the basement depth of Pikinni was inferred to lie between 1300 m and 1600 m.

The center of Wodejebato Guyot lies approximately 74 km to the northwest of Pikinni Atoll, a distance slightly less than that separating Hawaii (Kohala volcano) from East Maui Volcano in the Hawaiian Islands. A volcanic ridge 20 km long, 8.7 km wide and ~1600 m deep connects the two edifices. The summit of Wodejebato is about 43 km long (three-fourths the size of East Maui Volcano) and increases in width from less than 12 km in the southeast to greater than 25 km in the northwest (Figures 2.2, 2.3, and 2.4). Four volcanic ridges project from the northwest half of the edifice, and along with the volcanic spur attaching Wodejebato to Pikinni give the guyot a distinct "starfish" appearance. A drowned carbonate platform, slightly inset from the steep flanks of the guyot, covers the summit plateau (Figures 2.3 and 2.4). The regional gradient divides the flanks of the guyot into two provinces: a steep upper slope (~20°-24°) gives way to a more
Figure 2.3. Side-scan mosaic of Wodejebato Guyot showing the gross morphology of the guyot and the drowned Cretaceous carbonate platform (innermost dark band) capping the volcanic edifice. The boxes with figure labels indicate areas chosen for detailed discussion. Thick black lines show the location of seismic profiles across the south flank ridge (A-A'), the north flank ridge (B-B'), the south flank (C-C'), and two profiles used to interpret the drowned carbonate platform (C-C'' and D-D').
Figure 6

Wodejebato Guyot

Figure 7

Figure 8

Figure 5

Figure 6
Figure 2.4. Geologic interpretation of the side-scan mosaic based on rocks recovered during dredging (compare to Figures 2 and 3) and selected seismic profiles. Although pelagic sediments in this figure are shown only across the summit plateau, they also cover the shelves formed by the flank ridges.
Wodejebato Guyot
Geologic interpretation of side-scan images

Pelagic Sediments
Exposed Carbonate Platform
Slumped Platform Sediments?
Volcanic Features?
Channels
gently inclined lower slope (≈7°). In general, the transition depth between upper and lower slopes is around 2500 m. The following subsections describe and discuss the importance of various areas and features across Wodejebato for constraining the evolution of this drowned atoll (see Figure 2.3 for area locations).

2.5.1 Southern Flank

The carbonate platform cropping out in this area appears in the side-scan images as a fairly continuous band of high-backscatter material disrupted along-strike by thin acoustic shadows (Figure 2.5a). Small-scale changes in topography related to normal faulting or slumping of the carbonate sediments cause these acoustic shadows. The scarp truncating the carbonate platform is about 90 m high. In the side-scan images, the band of exposed carbonate sediments is between 500 m to 2000 m wide along-strike. Seismic data across this flank (Figure 2.5b) suggest the exposed sediments are lagoonal in origin; horizontal reflectors typical of lagoon sediments across the summit plateau do not appear to truncate against an acoustically-massive perimeter ridge.

Seaward of the carbonate platform lies a terrace, shown in the side-scan images as a region of low-backscatter between the high-backscatter carbonate sediments and the channels marking the upper slope. The terrace lies at a depth of around 1600 m in this area, but shallows to the northwest. It widens from less than 100 m in the southeast to over 1 km in the northwest, where it disappears along the southern flank ridge. In the seismic data, the stratigraphy of the presumed lagoon sediments matches that of the terrace (Figure 2.5b). Consequently, the terrace represents a down-dropped block of lagoon sediments.

Dredge hauls along the slopes below and above the terrace recovered basalt and limestone, respectively (Figure 2.2). RD49, sampling the upper slope of the edifice (~1800 m water depth), collected a mixture of basalt conglomerates containing rounded
Figure 2.5. Side-scan image (a) and seismic profile (b) across the south flank of Wodejebato. Along this flank, the exposed carbonate sediments are interpreted as lagoon or back reef deposits. In the 6-channel seismic profile, these presumed lagoon sediments (represented by the horizontal reflectors) do not truncate against an acoustically-massive perimeter ridge. Dredge RD50 sampled the scarp above the terrace and recovered only carbonate sediment. Dredge RD49 recovered only basalt from the scarp below the terrace in ~1800 m water depth. The upper 0.1 sec of terrace reflectors correlate with those of the topographically higher carbonate platform, suggesting that the terrace is a down-dropped block of carbonate sediments.
basalt pebbles, hyaloclastites, and pieces of previously formed conglomerates, but no reef debris. Lincoln [1990] noted that Mn coatings 2 to 3 cm thick on earlier-generation breccias and basalt pebbles suggest considerable hiatuses between erosional events. RD50, sampling the carbonate scarp at ~1500 m water depth, recovered about 300 lbs of Mn-encrusted limestone. The limestone contained Albian-age whole and fragmented rudists (along with other shell fragments) within a Campanian- to Maastrichtian-age foraminifer matrix [Lincoln, 1990].

2.5.2 Flank Ridges

As mentioned previously, four volcanic ridges extend from the northwest half of Wodejebato. These volcanic ridges are probably equivalent to the flank rift zones noted by Vogt and Smoot [1977] on seamounts and guyots in the north Pacific. Their relationship to the zones of active volcanism observed on the islands of Hawaii (i.e., loci of vents fed by dikes) is unclear. In this paper I refer to the volcanic ridges extending radial to the edifice as flank ridges; an examination of their origin follows their description.

The southern and northern flank ridges are the most prominent, appearing as broad (~4.5 km wide), relatively flat shelves extending ~15 km outward from the summit plateau (Figure 2.3). Although less broad than their northern and southern counterparts, the western and northeastern flank ridges extend outward about the same distance.

The southern flank ridge deepens away from the summit by 500 m, at a gradient of about 2°. In the side-scan images it is difficult to discern where the terrace, so prominent along the south flank, merges with the flank ridge (Figure 2.6a). The high backscatter band marking the areas where the carbonate platform crops out from the overlying pelagic sediments coalesces from two narrow bands in the east (lower half of the swath) to a single band in the west (upper half of the swath). No apparent perimeter ridge bounds the presumed lagoon sediments (Figure 2.6b). The first break in slope down from the summit
Figure 2.6. Side-scan image (a) and seismic profile (b) across the south flank ridge.

Profile A-A' shows that the block-faulting noted along the south flank continues to erode the carbonate platform. The reflectors labeled in the seismic profile represent the base of pelagic sediments (PR), the base of Limestone Unit 1 (LR1), and the base of Limestone Unit 2 or the top of volcanic basement (VB). Figures 9 and 10 show how these units vary across the summit of Wodejebato. The vertical exaggeration of profile A-A' = 6.5X.
High-backscatter cone
Flank ridge
Summit plateau

High-backscatter cones
Edge of carbonate platform
Terrace

Scale (km)

0 5

0200 Z 0230 Z 0300 Z 0330 Z

A'

Data gap

Slumped lagoonal sediments
Faults
LR1
PR

PR PR PR

0200 Z 0230 Z 0300 Z 0330 Z

2.7
is around 1500 m, suggesting that either the fault-block terrace shallows towards the northwest or portions of the carbonate platform slump on to the true terrace. Given the faulting to the east of this flank ridge, it is likely that the two narrow bands seen in the side-scan images are fault-related and that the area to the west of profile A-A’ is relatively unaffected by the erosion occurring along the southern flank. Along the edge of the flank ridge, away from the summit plateau, high-backscatter features crop out from the overlying pelagic sediment (Figure 2.6a). A few of these features resemble cones with diameters around 2 km. The height of the cones is unknown because the swath width of the SeaMARC II bathymetry was too narrow to provide coverage at this depth. A fathogram collected during the Mid-Pacific expedition [Emery et al., 1954, Figure 57, profile 9] possibly crosses one of the high backscatter cones. If so, then the cone in the area of profile 9 is less than 60 m high. Dredge MP43-DD, from about 1600 m water depth in this area, recovered 200 lbs of coarse-grained olivine basalt tuff. In contrast, RD47 sampled the lower slope of this flank ridge and recovered basalt, hyaloclastic conglomerates, and lithified pelagic sediments from a water depth of around 3250 m. Seismic data across this flank ridge show an undulating layer of pelagic sediment covering what may be a thin layer of slumped sediment or volcaniclastics (Figure 2.6b). As alluded to previously, horizontally-bedded, presumably lagoonal sediments crop out along the edge of the carbonate platform and an acoustically-massive perimeter ridge is absent. As neither dredge recovered rocks directly related to the drowned carbonate platform, the shallow-water carbonate sediments do not appear to extend on to the flank ridge shelf. If indeed the carbonate platform is eroding through block-faulting, some of the high backscatter features seen across the flank ridge near the summit edge may simply be slumped carbonate sediment. On the basis of the basalt tuff recovered in dredge MP43-DD, at least some of the cones visible across the flank ridge shelf must be volcanic in origin.
Moving clockwise around the guyot, side-scan data over the western flank ridge only partially map this feature, limiting the available geomorphic information (Figures 2.3 and 2.4). As stated previously, its length is comparable to the other flank ridges but it does not appear to be as planar as its northern and southern counterparts. North of the flank ridge, well-developed channels along the western flank of Wodejebato are noticeably absent in the side-scan images (Figure 2.3). This absence of channels could result either from a low-gradient upper slope hindering channel formation or from "recent" large-scale erosion along this flank erasing the acoustic record of such channels. Regional bathymetry in this area reveals the gradient to be a fairly uniform -6°, much less than the 20° to 24° upper slope along the southern and northern flanks. The cause for this lower gradient is unclear at this time.

The northern flank ridge is similar in length and width to its southern counterpart, but it does not appear to dip away from the summit. The 1600 m contour outlines the flank ridge shelf in the bathymetry (Figure 2.2). A dredge haul (MP43-A) across the upper slope (1900 m to 1600 m) of this ridge recovered tuff breccia [Emery et al., 1954], again suggesting that the carbonate platform does not extend on to the shelves. Unlike the southern flank ridge, the side-scan data do not show high-backscatter cones across the shelf (Figure 2.3). Instead, two perimeter ridges appear as high-backscatter bands across the area where the flank ridge intersects the volcanic edifice (Figure 2.7a). In contrast to the flank ridges which extend radial to the edifice, the perimeter ridges parallel the summit plateau edge. In the seismic data, the perimeter ridges in this area appear as low-relief (~75 m) mounds covered by a thin layer of pelagic sediment (Figure 2.7b). The inner ridge is slightly shallower (~45 m) and less broad than the seaward ridge. The two ridges extend around the summit along the northern flank, becoming more broad across areas where the shelf margin widens (i.e., adjacent to flank ridges) and finally disappearing on the volcanic ridge attaching Wodejebato to Pikinni. Given their location along the edge of the summit
Figure 2.7. Side-scan image (a) and seismic profile (b) across the north flank ridge. The side-scan image shows two perimeter ridges (represented by the high backscatter bands) cropping out from the pelagic sediments. These ridges are interpreted to be carbonate buildups. The inner perimeter ridge is relatively continuous around most of the guyot, while the outer ridge only appears on the north and northeast flank ridges, and the volcanic ridge attaching Wodejebato Guyot to Pikinni Atoll. The reflectors labeled in B-B' are the same as those listed in Figure 6. Vertical exaggeration of B-B' = 4.3X.
plateau, their relatively continuous distribution around the summit, and their mound-like
morphology in seismic profiles, these features probably represent reefs.

The northeast flank ridge exhibits a unique morphology in comparison to the other
flank ridges. As opposed to extending from the summit as a broad shelf, a narrow (~1.7
km) volcanic spur connects a 300 m high hill to the main edifice (Figure 2.2). The hill,
elongate in the direction of the summit, is about 2.2 km in diameter. As in the case of the
north flank ridge, the perimeter ridges broaden across the region where the flank ridge
intersects the edifice (Figure 2.8).

Vogt and Smoot [1984] discuss factors controlling the number and length of flank
rift zones on edifices in the Emperor, Geisha, Michelson, Dutton, and Mid-Pacific chains.
They conclude that the height of the edifice correlates with the length of the rift zones and is
independent from the number of rift zones. The flank ridges on Wodejebato Guyot are
comparable in length and number to those observed in the Geisha guyots, although no
direct evidence exists in either the seismic data or the dredge data confirming a rift zone
origin for these features. In the Hawaiian chain, volcanoes generally exhibit only two rift
zones, although sometimes a third smaller rift exists [J. Sinton, personal communication].
Consequently, the four flank ridges on Wodejebato Guyot could be attributed to: 1) some
feature unrelated to Hawaiian-type volcanic rift zones; 2) a magma plumbing system which
behaved differently from that of Hawaiian volcanoes, or 3) a multiple-volcano platform.
From the side-scan and bathymetry data it appears all the flank ridges are attached to the
main volcanic platform of Wodejebato and hence it is unlikely that any of the ridges are
remnants of large-scale slumps or debris flows. It is possible that the northeast flank ridge
represents a small seamount attached to the main edifice by a narrow volcanic ridge but the
absence of age information or geophysical data directly over this feature reduces such a
hypothesis to mere speculation. In a similar vein, there is not enough information at
present to discern how the magma plumbing systems of Cretaceous volcanoes may have
Figure 2.8. Side-scan image showing a portion of the northeast flank ridge and northern flank. The two perimeter ridges visible in the side-scan images adjacent to the north and northeast flank ridges are indistinguishable from one another along the north flank. Seismic data across the north flank (not presented) do not show an outer perimeter ridge, suggesting either that this feature is unique to areas where a broad shelf extends away from the guyot flank (i.e., the flank ridges) or that it has eroded.
behaved differently from those associated with the two- to three-rift zone volcanoes in the Hawaiian chain. Perhaps the simplest explanation for the four flank ridges on Wodejebato is that they represent the rift zones from a multiple-volcano platform.

2.5.3 Northern Flank

Along the northern edge of the summit plateau, the two distinct perimeter ridges observed across the north and northeast flank ridges merge to become essentially indistinguishable from one another in the side-scan images (Figure 2.8), and barely identifiable in the seismic data. Near the northeast flank ridge the inner perimeter ridge is ~75 m high and extends ~35 m shallower than the outer perimeter ridge. The outer ridge is ~115 m high. Although not visible in the SeaMARC II data, a narrow basement shelf (<100 m wide) lies outside of the carbonate platform at ~1600 m depth [H. Staudigel, unpublished data]. Channels, presumably from debris flows, incise the flanks outside the reefs. Upper and lower slope gradients are comparable to those measured along the south flank (>20° and around 7°, respectively).

2.5.4 Drowned Carbonate Platform

A layer of pelagic sediment covers the drowned carbonate platform and hence the summit plateau appears fairly featureless in the side-scan images (Figure 2.3). Seismic profiles provide information on the internal structure of the carbonate platform and the depth to volcanic basement. Figures 2.9 and 2.10 show two representative profiles across the summit. The migrated 6-channel line (Figure 2.9) was used as a tie-line for the single-channel profiles.

This paper identifies two major lagoon sediment units (Figure 2.9). The lowermost unit (Limestone 2) exhibits a maximum thickness of 0.11 sec (two-way travel time, TWT) or 165 m assuming a velocity of 3.0 km/sec, and constitutes the majority of
Figure 2.9. 6-channel seismic profile extending from the south flank to the northeast flank ridge (profile location shown in Figure 3). The reflectors labeled in C-C" are the same as those listed in Figure 6. The reflector interpreted as volcanic basement shallows towards the center of the guyot, almost extending into the pelagic sediments. Subunits within Limestone Unit 2 onlap the basement high. The packet of strong reflectors forming Limestone Unit 1 lies conformably over the lower carbonate unit and the inner perimeter ridge. The basement reflector beneath the perimeter ridge is located around 2.0 sec. The nature of the deep reflector shown in this figure is unclear; it may represent a flow unit overlain by volcaniclastic sediments. Vertical exaggeration of C-C" = 7.5X.
Wodejebato Guyot
Migrated 6-channel seismic profile

Interpretation

1.7
Limestone Unit 2 Limestone Unit 1 Pelagic cap
Basement

Pelagic Sediments Limestone Unit 1 Inner Ridge
Limestone Unit 2

TWT (secs)
2.7

Deep Reflector
Basement

TWT (secs)
Figure 2.10. Seismic profile extending from west of the south flank ridge to east of the northeast flank ridge (profile location shown in Figure 3). The peak in volcanic basement is less pronounced and displaced to the northeast relative to profile C-C". A perimeter ridge appears to exist along the southwest flank of Wodejebato, and it may be a continuation of the inner ridge noted along the north and northeast flank ridges. It is difficult to determine in the single-channel profiles the depth to which the perimeter ridge extends. Limestone Unit 2 appears to partially onlap the basement high. Vertical exaggeration of D-D' = 11X.
carbonate sediment deposited on this guyot. Within this unit, horizontal reflectors define subunits which onlap the basement high. Age information on these subunits awaits drilling during ODP Leg 144. Overlying the lower limestone unit are a series of closely spaced reflectors (Limestone 1). The reflector defining the bottom of this unit could be a diagenetic horizon or a sequence of overbank sediments deposited shortly before the carbonate platform drowned. Both limestone units show relatively little relief (essentially horizontal reflectors). As noted previously, faulting exposes these units along the southern flank of Wodejebato. The overlying pelagic sediments attain a maximum thickness of ~100 m near the center of the summit plateau and thin towards its edges.

The basement reflector identified in this paper may represent a layer of weathered volcanioclastics, as opposed to "true" volcanic flows, but it still marks the lowermost extent of the carbonate sediments. In figures 2.9 and 2.10, the basement reflector shallows towards the center of the summit plateau. Profile C-C" (Figures 2.3 and 2.9) shows that the basement high extends almost into the pelagic sediments. The two perimeter ridges along the northeast flank ridge are prominent features in this profile. Referred to earlier in the paper as being "acoustically massive", it is apparent from Figure 2.9 that the ridges exhibit some internal structure. A 0.03 sec (TWT) packet of strong reflectors overlies 0.12 sec of less-defined reflectors. The relatively thin packet of strong reflectors appears to extend ~2 km behind the inner ridge, and may be related to the Limestone 1 unit identified across the interior portions of the platform. The seaward ridge is 125 m high whereas the inner ridge, extending about 50 m shallower than the seaward ridge, is around 90 m high. In profile B-B' (Figures 2.3 and 2.10), the basement peak is offset to the northeast. The single, 150 m to 200 m high perimeter ridge shown in this profile suggests that this flank is unaffected by the faulting occurring along the south flank farther to the southeast.

The isopach map of carbonate thickness shown in Figure 2.11 maps basement topography. Areas where the carbonate platform sediments thin represent basement highs.
Figure 2.11. Isopach map of total carbonate thickness (Limestone Units 1 and 2) over the summit plateau of Wodejebato. The thin black contours represent carbonate thickness in millisec, while the thicker gray contours represent bathymetry (showing the geographic extent of the summit plateau). As the interface between the carbonate and pelagic sediments is a relatively flat surface, the carbonate thickness contours actually map the basement topography. Volcanic basement shallows to the northeast, and apparently has two peaks displaced from one another by ~5 km.
Volcanic basement shallows to the northeast and appears to have two peaks. Although the apparent peaks in basement provide additional support for the hypothesis that two volcanoes compose the main edifice of Wodejebato, the size of these volcanoes must have been small in comparison to Hawaiian volcanoes as the maximum summit diameter of Wodejebato is only about three-fourths the size of East Maui Volcano. Alternatively, the trough separating the apparent peaks in basement could be an artifact of the wide ship-track spacing.

2.6. Discussion

An important step towards deciphering the volcanic and tectonic history of the Marshall Islands as a whole is to more fully understand the history of its individual components. With the wealth of information now available through dredging and geophysical surveying (and soon drilling), it's inevitable that the existing interpretations for this area of the Pacific plate will undergo revisions. An example of such revisions is the reclassification of the carbonate platform capping Wodejebato Guyot. The Eocene age proposed by Schlanger et al. [1987] has been replaced by a Cretaceous age derived from the MW8805 dredge data [Schlanger and Duennebier, 1988]. This preliminary observation has in turn been revised to include two stages of carbonate platform growth over the course of two episodes of tectonic uplift [Lincoln et al., 1989; Lincoln, 1990]. The data presented in this paper provide a more detailed picture of Wodejebato's morphology, and furnish some additional insights to its geologic history. Any model explaining the evolution of Wodejebato must take into account not only the radiometric and fossil evidence for two episodes of uplift and volcanism, but also such geomorphic information as the depth of volcanic basement along the guyot's flanks and beneath the carbonate cap, the presence of the perimeter ridges along the eastern, northern, and western flanks, and the existence of the terrace along the southern flank.
2.6.1 Implications of Guyot Morphology

The depth of volcanic basement along the flanks is about 1600 m. This contour outlines most of the flank ridges on this guyot, and marks the first break in slope down from the drowned carbonate platform along margins where an acoustically massive perimeter ridge exists. Along the southern flank of Wodejebato, volcanic basement can not be deeper than 1800 m as shown by RD49. From the isopach map of carbonate thickness (Figure 2.11), it's apparent that volcanic basement shallows towards the northeast, although the implications of this shallowing are unclear.

The perimeter ridges extending around the northern flank of Wodejebato may represent the two episodes of carbonate platform accumulation suggested by fossil and radiometric data [Lincoln, 1990], but they could just as well have formed during a single episode of uplift. From the seismic data alone it is unclear to what depth the reefs extend, or upon what type of substrate they grew (i.e., volcanic basement or existing carbonate sediments). The strong reflector at ~2.01 sec (TWT) in Figure 2.9 may be volcanic basement, but testing of this hypothesis awaits drilling during Leg 144. Without drilling information, it's also impossible to determine the age of lagoon sediments identified in the seismic profiles. In lieu of this additional information, three different growth models may apply to Wodejebato's drowned carbonate platform:

1. The perimeter ridges mark two distinct periods of platform growth (i.e. mid-Cretaceous and Late Cretaceous) with the younger ridge lying outside of and deeper than an older carbonate platform [Lincoln 1990].

2. The perimeter ridges again mark 2 distinct periods of platform growth, but in this case the younger ridge (and carbonate platform) lies inside of and shallower than the older carbonate platform.
3. Both ridges (and the carbonate platform as a whole) grew during the last stage of uplift but are offset from one another because of some mechanism other than regional uplift (e.g., a eustatic sea level change or plate flexure from loading on nearby volcanoes).

If the carbonate platform grew during separate episodes of hot spot induced uplift, then the first or second growth models apply. The absence of a pronounced karstic surface in the seismic profiles weakens support for the first model. It also seems unlikely that subsidence rates after the Late Cretaceous uplift event were so great that a younger carbonate platform was unable to grow on top of the existing platform. The second growth model conforms better to observed reef growth patterns along modern shelves, where younger reefs tend to colonize preexisting highs formed by older reefs [e.g. Purdy, 1974; Hopley, 1982], but the third growth model is equally applicable given the current data. The ultimate test for these models will come with the drilling. If the first model applies, then the lagoon sediments should be primarily mid-Cretaceous in age and should contain evidence for extensive subaerial exposure and diagenesis. If the second or third models apply, Late Cretaceous lagoon sediments should unconformably overlie a thin veneer of mid-Cretaceous sediments. In this case, drilling directly behind the perimeter ridges may determine whether the ridges formed contemporaneously or during two separate uplift events.

For this study, it's important not to focus too much on when the perimeter ridges grew or what mechanism was responsible for their growth (i.e., tectonic or eustatic). The important point here is that a complete atoll existed across the summit at the time of platform drowning. Alterations to the expected atoll morphology (i.e., a massive reef rim enclosing layered lagoon deposits) place constraints on the timing and the extent of selected tectonic events. For example, the 100 to 1000 m wide terrace seaward of the carbonate platform on the southern flank and its proposed fault-block origin suggest that relatively
large-scale submarine landslides continue to modify the flanks of volcanic islands well after they have moved away from the hot spot. In the case of Wodejebato, the faulting had to occur after the atoll drowned otherwise a new perimeter ridge would presumably have grown on top of the lagoon sediments. As shown in Figure 2.5b, smaller faults continue to disrupt the lagoon sediments along this flank.

2.6.2 Model for the Geologic Evolution of Wodejebato Guyot

Although the sequence of events responsible for the formation of Wodejebato Guyot was undoubtedly complex, Figure 2.12 shows a simple model of guyot evolution favored in this paper. The model combines some of the hypotheses of Lincoln [1990] with the geomorphic observations given in this paper. Information on the early history of Wodejebato comes exclusively from fossil ages; it is impossible to deduce anything except the most recent episode of uplift and island formation from the geomorphic information provided in this paper. As suggested by Lincoln [1990], the edifice first formed in the mid-Cretaceous, possibly in the vicinity of the Macdonald hot spot. Cooling and subsidence of the plate away from the hot spot, along with wave erosion, eventually formed the summit plateau of the volcano. A carbonate platform established itself on top of the wave-planated summit. At some point during its history, the rates of subsidence and sea level rise surpassed the rate of carbonate platform growth, and the developing platform drowned.

Uplift of the plate (and edifice) as it passed over the Rurutu hot spot during the Late Cretaceous [Lincoln, 1990] resulted in restricted volcanism across the edifice and extensive erosion of the mid-Cretaceous carbonate platform. As the plate cooled and subsided again (along with the edifice) a Late Cretaceous carbonate platform recolonized the existing veneer of carbonate sediments. The carbonate platform morphology was similar to many modern atolls: raised perimeter ridges bounded relatively flat-lying lagoon deposits. The
Figure 2.12. A simple model depicting the evolution of Wodejebato Guyot. The cross-section shown in this cartoon is comparable to the profile shown in C-C" (excluding the flank ridge). Details of the model are provided in the text.
Evolution of Wodejebato Guyot

**t1**

Area viewed in subsequent time slices

Volcanic edifice

Sea Level

Sea Floor

**t2**

Establishment and drowning of mid-Cretaceous carbonate platform

Drowned Albian carbonate platform

Sea Level

**t3**

Late Cretaceous uplift and erosion of mid-Cretaceous carbonate platform

Erosion surface

Remnants of mid-Cretaceous carbonate platform

Sea Level

**t4**

Subsidence of edifice and establishment of Late Cretaceous carbonate platform

Late Cretaceous carbonate platform

Remnant of mid-Cretaceous carbonate platform

Sea Level

**t5**

Drowning of Late Cretaceous carbonate platform

Sea Level

**t6**

Normal faulting of edifice (south flank)

Fault block

Pelagic sediments

Erosion
two perimeter ridges along the northern and eastern flanks of Wodejebato suggest that more than one episode of carbonate platform growth occurred during this last stage of uplift. After full atoll development a large-scale submarine landslide, possibly comparable to those noted along the Hawaiian Islands [Moore et al., 1989], removed a portion of the southern flank including the perimeter ridge. Smaller-scale faults continue to erode blocks of basement and lagoon sediments, one of which forms the terrace along the southern flank.

2.7. Conclusions

By examining the morphology and subsurface structure of Wodejebato guyot in detail, and through the analysis of basalt and limestone samples collected from its slopes, a relatively complete history for this volcano is emerging. Initial edifice construction occurred during the mid-Cretaceous, with a later stage of volcanic activity around 86 Ma. Carbonate platform accumulation across the summit apparently accompanied each period of plate uplift. In areas where shelves exist along the flanks of the edifice (e.g., next to flank ridges), the perimeter ridges broaden. The southern flank of this guyot exhibits extensive erosion. Lagoon sediments crop out where normal faulting has removed the reef rim.

With the planned ODP drilling during Legs 143 and 144, answers to many of the vexing tectonic, sea level, and carbonate platform-growth questions still posed by this complex chain of atolls and guyots should be forthcoming. Future generations of scientists involved in these areas of research will undoubtedly benefit from this new information, thanks in part to Sy Schlanger's perseverance in furthering our understanding of Cretaceous events.
CHAPTER 3
MORPHOLOGY OF MARSHALL ISLANDS GUYOTS DRILLED DURING ODP LEG 144: GEOPHYSICAL CONSTRAINTS ON PLATFORM DROWNING

3.1. Abstract

Recent drilling during Ocean Drilling Program Leg 144 has provided much needed stratigraphic control on selected guyots in the Marshall Islands. The combined information from geophysical surveys conducted prior to drilling and the preliminary ages and lithologies assigned to the carbonate and volcanic sediments sampled during the drilling makes it possible to compare the drowned platforms of Limalok, Lo-En, and Wodejebato with various islands and atolls in the Hawaiian and French Polynesian chains. As a whole, the gross morphology and stratigraphy of the platforms drilled during Leg 144 appear quite similar to modern islands and atolls. The number of flank ridges extending from each edifice generally conforms to the number of rift zones observed for Hawaiian volcanoes. Fault-blocks perched along the flanks of the guyots suggest that the edifices remain unstable even after moving away from the hot spot swell. Perimeter ridges bounding lagoon sediments, transition zones of clay and altered volcaniclastic sediments separating the shallow-water platform carbonates from the underlying volcanic flows, and basement highs towards the center of the platform all appear similar to atolls and uplifted atolls studied in French Polynesia. The platforms drilled in the Marshall Islands are also similar to French Polynesian islands in the wide geomorphic and stratigraphic variations they exhibit between hot spot chains: 230 m of Paleocene to middle Eocene platform carbonates top Limalok, a 100 m-thick Late Cretaceous carbonate platform caps Wodejebato, and essentially no shallow-water carbonates cover Lo-En. Given the complex history of volcanism in the Marshall Islands, it is not surprising that some observations across the summit plateaus do not adhere to traditional models of slow plate subsidence and edifice
truncation. Cone- and lobe-shaped volcanic features cropping out across the relatively flat summit plateau of Lo-En and the absence of shallow-water carbonates covering this volcanic platform are difficult to explain without invoking some mechanism to raise sea level at a relatively fast rate. One mechanism capable of causing relatively rapid subsidence or uplift on an edifice is plate flexure from the construction of nearby volcanoes. The preliminary ages and basement depths supplied by Leg 144 make it possible to examine the role of plate flexure in creating the morphologic and stratigraphic differences observed across the summit plateaus of Lo-En, Wodejebato, and Limalok.

The best age control for the atoll and guyot pairs discussed in this paper exists on Anewetak and Lo-En. Flexure modeling on these two edifices suggests Lo-En experienced little if any uplift or subsidence in response to volcanism on Anewetak. These results may explain the negligible difference between the basement depths of Anewetak and Lo-En, and the high-backscatter cone- and lobe-shaped features cropping out along the flat summit plateau of Lo-En. Age control on the remaining two atoll and guyot pairs, Pikinni-Wodejebato and Mili-Limalok, is less constrained. Flexure modeling on these two pairs of edifices suggests Pikinni and Wodejebato were probably constructed at approximately the same time, and that construction of a volcano to the south of Limalok may explain the sequence of prograding basement reflectors. The models fail to establish the cause of carbonate platform drowning (or survival), but illustrate the importance of better understanding the sequence and timing of volcanism in the Marshall Islands. Attaining such an understanding is critical towards using the drowned carbonate platforms drilled during Leg 144 as "dipsticks" for recording the history of Cretaceous sea level fluctuations.
3.2. Introduction

In recent years the western portion of the Pacific plate has become the focus of a number of ship-based surveying and sampling programs designed to gather information on Cretaceous volcanism and sea level changes. This region is notable for an anomalously shallow sea floor depth in comparison to other ocean basins of the world, and for a relatively high concentration of seamounts, atolls, guyots, and mid-ocean plateaus. The late H. W. Menard, an early pioneer of Pacific marine geology surveys who coined the term "Darwin Rise" for the area extending from the Mid-Pacific Mountains to the Ontong-Java Plateau, postulated that the excessive volcanism resulted from fracturing of the plate over a mantle bulge [Menard, 1964]. With the advent of plate tectonics and subsequent age estimates for volcanic features within the Darwin Rise, the portion of the Pacific plate upon which Marshall Islands lie was shown to be near the present-day location of French Polynesia at the time of the volcanism [e.g., Smith et al., 1989; Lincoln, 1990; Pringle, 1992]. Accordingly, the islands composing French Polynesia supply important clues for deciphering the tectonic history of the Marshall Islands (Figure 3.1).

French Polynesia, or today the Polynesian Plume Province, consists of five island chains trending in a northwest direction. From northeast to southwest these chains are the Marquesas Islands, the Gambier-Pitcairn Islands, the Tuamotu Islands, the Society Islands, and the Austral-Cook Islands. Four of these five chains can be directly linked to mid-plate volcanism associated with hot spots; the Tuamotu Islands are the exception, and appear to be a series of uplifted atolls superimposed on an aseismic ridge [e.g., Duncan and McDougall, 1974, 1976; Turner and Jarrard, 1982]. Geophysical and geochemical studies in and around the French Polynesian islands show that this region is similar to the Marshall Islands in such respects as anomalously shallow sea floor and multiple, linear island chains [McNutt and Fischer, 1987]. Other characteristics of French Polynesia may also apply to the Marshall Islands, including an apparent overprinting of volcanic events on selected
Figure 3.1. Location map for the Marshall Islands and French Polynesia. Boxes mark the guyots discussed in this paper. The dates shown in parantheses represent radiometric ages of basalts recovered during dredging or drilling; these dates come from Davis et al. [1987] and Pringle [1992, unpublished data]. Bathymetry map revised from Hein et al. [1990]
Republic of the Marshall Islands

North Wod-En (86 Ma)

Lobbedede (82 Ma)

Woden-Kopakut (82 Ma)

Look (138 Ma)

Anewetak (76 Ma)

Pikinni

Wodejebato (86 Ma)

Lo-Én

Lallihjet (80 Ma)

Mlj-Leep (110 Ma)

Hole 462A (110 Ma)

Lokkworkwor (87 Ma)

Limalok

160°E 164° 168° 172°
islands and uplift of neighboring islands through volcanic loading and plate flexure [Duncan and McDougall, 1974, 1976; Turner and Jarrard, 1982; McNutt and Menard, 1978]. These observations combined with the distinct geochemical signature of basalt samples [Hart, 1984], the shallow low-velocity zone for Love waves [Nishimura and Forsyth, 1985], and the small estimates for the effective elastic plate thickness [Calmant and Cazenave, 1987] led to the hypothesis that a large mantle plume, or "superswell", perturbs normal plate subsidence in this region [McNutt and Fischer, 1987; McNutt and Judge, 1990]. Such a plume may also have been active during the Cretaceous, and thus responsible for the apparent similarities in volcanism and subsidence across French Polynesia and the Darwin Rise [e.g., McNutt, 1990; Larson, 1992].

The results from recent Ocean Drilling Program (ODP) cruises Leg 130 (Ontong-Java Plateau), Leg 143 (Mid-Pacific Mountains), and Leg 144 (Marshall Islands, Wake Islands, and Japanese Seamounts) continue to refine the genetic links established between the Darwin Rise and French Polynesia. The primary focus of the atoll and guyot drilling legs (143 and 144) was to better understand the Cretaceous history of sea level changes, coral growth patterns, and episodes of volcanism across the Pacific plate as recorded by the shallow-water carbonate and volcanogenic sediments deposited on top of and adjacent to selected edifices. Inherent in this drilling strategy was the assumption that a simple subsidence history for these guyots could be modeled and removed from the sediment record to give an indication of eustatic changes in sea level during the Cretaceous. Such a simple subsidence pattern may not exist in the Marshall Islands, as suggested by the uplifted atolls and islands in French Polynesia. The first requirement for an accurate interpretation of the drilling data collected during Leg 144 is a good understanding of the factors controlling uplift and subsidence in this region. In light of the preliminary results from Leg 144, the objectives of this paper are: 1) to review the existing models for the evolution of the Marshall Islands in relation to the drilling data, 2) to examine physical
differences between the platforms drilled during Leg 144 and to discuss the significance of
these differences on the proposed evolutionary models, and 3) to model the effects of plate
flexure from volcanic loading on adjacent atoll and guyot pairs. Descriptions of the
geological and geophysical data serve the simple purpose of putting the drilling results in
perspective with the overall structure of the guyot. Ideally, features noted on the Marshalls
guyots should be similar to features observed on subaerial volcanic islands which have
been studied in more detail. Obvious candidates for such a comparative study include the
islands composing Hawaii and French Polynesia (e.g., the Society and Austral chains).

When comparing the Marshall Islands to the Hawaiian Islands, one of the problems
encountered is a difference in scale; the islands in Hawaii are typically at least twice as large
as the edifices drilled in the Marshall Islands. On the other hand, while islands within the
Society and Austral chains are more comparable in size to guyots in the Marshall Islands,
geological information from French Polynesia is less available. Accordingly, this paper
uses examples from both regions to better understand similar features noted in the Marshall
Islands.

3.3. Background

The Marshall Islands can be subdivided into 3 geographic provinces: to the east lies
the Ratak chain, to the west lies a cluster of edifices grouped around Anewetak and Ujlan
atolls, and between these two groups lies the Ralik chain. With each successive surveying
and sampling program across this region, the history of plate motion, plate uplift and
subsidence, and mid-plate volcanism has become more clear. Invariably, the information
derived from each program has resulted in a new set of questions concerning the magnitude
and timing of Cretaceous volcanic events in the western Pacific. This cycle of information
 gained versus questions raised does not exclude the data collected during Leg 144. As a
prelude to insights resulting from the drilling, this section reviews the models developed from the earlier sampling programs.

3.3.1. Results from Atoll Drilling

Early models explaining the evolution of volcanic chains within the Marshall Islands were derived from drilling on Anewetak (formerly Enewetak) and Pikinni (formerly Bikini) atolls during Operation Crossroads [Emery et al., 1954; Schlanger, 1963]. The atoll drilling provided the first estimates of volcanic platform age and depth in the Marshall Islands, and allowed comparisons between two shallow-water carbonate platforms (e.g., stratigraphy, thickness, and age of carbonate sediments; depth to various solution horizons). Two deep holes drilled on Anewetak (Figure 3.2) reached volcanic basement between 1405 m (Elugelab Island in the northwest) and 1282 m (Parry Island in the southeast; Emery et al., 1954; Schlanger, 1963). Drilling on Pikinni Atoll consisted of four holes (Figure 3.2), the deepest of which penetrated 779 m of shallow-water reef sediments but did not encounter basalt basement [Cole, 1954; Emery et al., 1954]. The average depth to basement on Pikinni was estimated by Raitt [1954] from seismic refraction data to lie around 1300 m, although he noted the data support a depth range between 600 m and 2100 m. On Anewetak Atoll, shallow-water carbonate sediments of Eocene age lay directly over the basalt, originally dated by K/Ar methods as being in the range of 51.4 to 61.4 Ma. No estimate for the age of the volcanic platform beneath Pikinni was proffered at this time but it seems reasonable, given the depth estimate for basement on this edifice, that it could be similar to the one beneath Anewetak.

The model developed for the Marshall Islands based on the results of the atoll drilling was that shallow-water carbonate platforms of Eocene age grew on top of Cenozoic-age volcanic platforms. These platforms subsided at anomalously fast rates given the age of the plate upon which they were built [Detrick and Crough, 1978]. The
Figure 3.2. Stratigraphy of cores drilled on Anewetak and Pikinni atolls during Operation Crossroads and on Limalok, Lo-En, and Wodejebato guyots during Leg 144. Site locations shown in subsequent figures.
shallow-water carbonate sediments recorded a series of relative sea level falls in the form of solution unconformities, and the good correlation between the depths of these unconformities across the two carbonate platforms suggested these sea level changes were eustatic in nature. Accordingly, atolls and guyots became viewed as mid-ocean "dipsticks" upon which eustatic changes in sea level were recorded.

3.3.2. Results from Ship-based Sampling Programs

Questions concerning the history of volcanism in the Marshall Islands, and the Cretaceous in general, began to arise with the drilling results from Deep Sea Drilling Program (DSDP) Legs 61 and 89. Both legs spent time drilling or deepening a hole west of the Marshall Islands in the Nauru Basin. Hole 462 penetrated a total of 599 m of sediment and 18 m of basalt; Hole 462A extended "basement" penetration an additional 451 m into the basalt complex [Larson, Schlanger et al., 1981; Moberly, Schlanger et al., 1986]. The lowermost unit at this site consists of basalt sheet flows overlain by single and multiple basalt sills, all of which are normally magnetized (presumably erupted during the Cretaceous normal superchron). Microfossils in the sediment layer separating the flow complex from the overlying sill units suggest the youngest flows are ~112 m.y. old [Moberly, Schlanger et al., 1986], while radiometric dating of the oldest flows shows an age range between 127 m.y. and 131 m.y. [Takigami et al., 1986]. Farther up in the hole, redeposited shallow-water sands, reef limestones, phosphorite, and subaerial basalt indicate nearby edifice-building during the Campanian and Maestrichtian [e.g., Premoli-Silva et al., 1981]. The results from Site 462 provided a better perspective, in comparison to that afforded by the Operation Crossroads drilling, on the extent and types of volcanic events affecting this portion of the Pacific plate during the Cretaceous. From the drilling data, Schlanger and Premoli-Silva [1981] postulated that plate uplift and major mid-plate volcanism began in Barremian to early Aptian time (~115 Ma) and continued intermittently.
through the Maestrichtian. Alternatively, Moberly and Jenkyns [1981] proposed two distinct episodes of mid-plate volcanism, one at 118 +/- 6 Ma and another at 74 +/- 2 Ma. Both models extended the beginning of edifice building in the Marshall Islands back to the mid-Cretaceous.

Dredging across the guyots composing the Marshall Islands during the 1988 R/V _Moana Wave_ cruise MW8805 provided the next influx of information concerning their evolution. Samples collected during this cruise varied widely in composition (e.g., basalt, basalt conglomerates, shallow-water carbonates, etc.) and age (mid-Cretaceous to middle Eocene), and were the basis for a more-refined model detailing the Cretaceous history of volcanism and carbonate platform growth in this region. Lincoln [1990] used the age information from the fossil constituents of shallow-water carbonates and carbonate cements and the radiometric dating of basalts to develop a multiple hot spot model for the evolution of the Marshall Islands. He proposed that as the portion of the Pacific plate containing the Marshall Islands moved away from the East Pacific Rise during the Cretaceous, it was influenced by the swells from at least two aligned hot spots, similar to the present situation in French Polynesia (e.g., the Society and Rurutu hot spots; Turner and Jarrard, 1982). Uplift and volcanism associated with these hot spot swells resulted in two sequences of reef growth across many of the edifices in the Marshall Islands.

The fossil content of limestones recovered in rock dredge (RD) 50 from Wodejebato Guyot during MW8805 was central to the development of this model. These limestones contain an assemblage of rudists (identified as Albian in age) in a matrix of shallow-water foraminifera (Campanian to Maestrichtian in age). Radiometric dating of a basalt sample collected along the north flank of Wodejebato during the Mid-Pacific Expedition (MP43-D; Hamilton and Rex, 1959) places the latest stage of volcanism at ~86 Ma [M. Pringle, unpublished data]. Thus, while the fossil assemblage in the limestone suggests that the summit plateau of this edifice was at sea level during the Albian and late
Cretaceous, the radiometric age provides support for only a late Cretaceous episode of volcanism (and presumably edifice uplift). This episode of volcanism is at least 20 m.y. later than the main edifice-building stage (presumably > 105 Ma) as proposed by Lincoln [1990]. Other edifices from which mid-Cretaceous carbonate sediments were dredged during MW8805 include Lo-En, Lewa, Lobbadede, and Ruwituntun guyots, with ages ranging between latest Albian to earliest Cenomanian (Figure 3.1; Lincoln, 1990). Lo-En Guyot is of interest here because of its proximity to Anewetak Atoll. Recent $^{40}$Ar/ $^{39}$Ar dating of the basalts recovered during drilling on Anewetak produces a total fusion age of 75.9 +/- 0.6 Ma for the volcanic platform [Pringle, 1992]. The contents of RD33 along the southwest flank of Lo-En (~2400 m depth) consists of breccia in the form of angular fragments of vesicular basalt floating in a calcareous matrix of shallow-water debris and planktonic foraminifers. The planktonic foraminifers within the matrix suggested a middle to late Albian age for the rocks. Two other dredges on this guyot (RD34 and RD35) sampled shallower portions of the slope (~1900 m and ~1775 m, respectively) and recovered rounded basalt cobbles and basalt pebbles in matrices containing planktonic foraminifers of Paleogene age. Although no direct evidence in the MW8805 data supported an episode of Late Cretaceous volcanism or reef growth on Lo-En, Lincoln [1990] postulated that uplift and subaerial exposure of this edifice was likely around the time of volcanic activity on Anewetak Atoll (~76 Ma). Radiometric ages for basalts dredged from guyots throughout the Marshall Islands appear to contradict this "mid-Cretaceous/Late Cretaceous" model for plate uplift and volcanism. Pringle and Staudigel [1992] show that the majority of basalts dated from this region fall between 90 and 75 m.y. old. Only two guyots, Look and Mij-Lep, exhibit ages older than 100 Ma (Figure 3.1), and neither of these edifices are capped by a carbonate platform. Consequently, very few radiometric ages support extensive edifice-building, or even the existence of edifices, in the Marshall Islands during the mid-Cretaceous.
The Cenozoic development of carbonate platforms in the Marshall Islands is best represented by Limalok Guyot. Dredges across Limalok during a 1981 R/V Kana Keoki cruise (KK810626, leg 2) returned a mixture of shallow-water carbonates within a planktonic foraminifer matrix, indicating a complex history of platform growth and subsequent redeposition in a deeper-water environment [Schlanger et al, 1987]. Large, benthic foraminifers within the limestones suggest the carbonate platform was established by early Eocene time. Schlanger et al. [1987] postulated that Eocene sea level changes, apparently derived from the eustacy curves of Vail [1977], resulted in emergence and eventually drowning of the carbonate platform. Lincoln [1990] attempted to constrain the history of emergence and subsidence across this carbonate platform by examining the carbon and oxygen isotope compositions of the limestones recovered in dredges KK81-4 and KK81-6. Analyses on both the primary fossil components (foraminifers and red algae) and the cements failed to support the existence of a subaerial exposure event affecting the limestones capping Limalok, nor did they provide any additional information on when or why this carbonate platform drowned.

3.3.3. Preliminary Results from ODP Leg 144

As illustrated above, the Cretaceous history of the Marshall Islands prior to drilling during Leg 144 had been deduced mainly from a relatively small number of dredge samples collected from a few scattered guyots. Leg 144 attempted to collect a relatively complete stratigraphic section extending from the early Cenozoic to presumably the mid-Cretaceous by drilling the platforms of Limalok (Site 871), Lo-En (Site 872), and Wodejebato (Sites 873 through 877) guyots, or one edifice from each of the three island chains forming the Marshall Islands. Although the observations and interpretations made during the cruise continue to be refined, the preliminary ages assigned to the basement basalts and the shallow-water carbonates provided immediate constraints on the duration of carbonate
platform growth across these edifices [Haggerty, Premoli-Silva, et al., in press] and a reevaluation of the existing models explaining the evolution of the Marshall Islands.

The deepest hole on Limalok Guyot at Site 871 penetrated a total of 500 m of sediments and altered basalt (Figure 3.2). The four lithologic units identified at Site 871 [Haggerty, Premoli-Silva et al., in press] consist of pelagic sediments (133.7 m), middle Eocene to early late Paleocene shallow-water platform carbonates (~290 m), variegated clays (~30 m), breccias, and basalts marking volcanic basement (penetration of 50 m). An iron-manganese oxide and phosphatic hardground marks the contact between the pelagic sediments and the underlying platform carbonates. Pelagic sediments of late early Oligocene age fill borings in the hardground and mark the approximate time the atoll drowned. The carbonate platform is almost entirely Eocene in age, although some Paleocene sediments lie directly over the 30 m thick clay layer separating the platform carbonates from the volcanic edifice. This clay layer weathering horizon changes color from dark-gray at the bottom to light-gray at the top, showing a slow progression towards a higher energy (more open marine) depositional environment [Haggerty, Premoli-Silva et al., in press]. Volcanic basement on this edifice consists of ~6 m of volcanogenic breccias overlying a sequence of massive flows ranging in thickness from 1 to 7 m. Magnetic inclination data suggest the edifice was at a paleolatitude of ~10° S during the eruption of these flows [Nakanishi, Gee et al., 1992].

Although the primary objective at Site 872 on Lo-En Guyot was to sample the pelagic sediments capping the summit plateau, drilling proceeded until penetration of volcanic basement (Figure 3.2). Of the four lithologic units recognized at this site, pelagic sediment and basalt form the bulk of the samples recovered [Haggerty, Premoli-Silva et al., in press]. The ~142 m of pelagic sediment sampled at this site (Unit I) is lithologically similar to the pelagic unit sampled at Site 871 on Limalok Guyot. These sediments range in age from Pleistocene to late Oligocene. Units II and III form a thin transition layer (~10 cm
each) between the pelagic sediments and the underlying volcanic platform (Unit IV) and consist of subangular basalt fragments, phosphatized lithoclasts, and volcaniclastic sands within a pelagic limestone matrix ranging in age from middle-Eocene to late early Santonian. Drilling at Hole 872B penetrated 57 m into volcanic basement, sampling a series of differentiated alkali olivine basalts as massive flows and flow-top breccias. Preliminary analysis of the foraminifer filling the cracks in these flows places the age of volcanism as pre-late Turonian (~90 Ma) [Haggerty, Premoli-Silva et al., in press]. Magnetic inclination estimates suggest that the edifice was at a paleolatitude of 30° S during these eruptions, a latitude consistent with plate-rotation estimates based on the fossil-age for the volcanics [Nakanishi, Gee et al., 1992]. The absence of shallow-water carbonates, along with the composition and age of the transitional units II and III, suggest that a depositional hiatus of > 50 m.y. (pre-Turonian to late Oligocene) exists across this summit plateau [Haggerty and Premoli-Silva, in press]. Surprisingly, the only sample recovered on this guyot which appears related to the ~76 Ma volcanic event recorded at Anewetak Atoll was a small pebble containing glass shards in a Campanian-age pelagic matrix.

Drilling on the summit plateau of Wodejebato Guyot, consisting of 5 sites and nine holes, sampled a variety of depositional environments. Holes at Site 873 cored lagoon sediments from the south-central portion of the plateau, while holes at Sites 874 through 877 sampled a pair of perimeter ridges along the north edge of the plateau (Figure 3.2). The six lithologic units recognized at Site 873 [Haggerty, Premoli-Silva et al., in press] consist of pelagic sediments (54 m thick), a thin veneer of manganese-oxide encrusted, phosphatized limestone conglomerate (0.6 m), shallow-water platform carbonates (~100 m thick), ferrigenous clays and olivine claystone (~20 m), altered basalt (~30 m), and poorly sorted, angular to subangular volcanic breccia (penetration of ~30 m). Preliminary analysis of the fossils composing the shallow-water carbonate sediments suggests the platform was established by the late Campanian and survived until late Maestrichtian time. Changes in
faunal diversity mark alternating phases of open marine and more restricted conditions within the shallow lagoon and represent a number of flooding events across the platform [Haggerty, Premoli-Silva, et al., in press]. These flooding events appear in the seismic data as reflector-bounded subunits onlapping a central basement high [Bergersen, 1993]. Drilling at Sites 874 through 877 sampled a pair of ridges along the perimeter of the summit plateau. The inner ridge appears to be a fairly continuous feature along the edge of the plateau, and its sediments record two episodes of bioherm growth separated by a major flooding event. The outer ridge, on the other hand, only appears along shelves formed by the flank ridges extending away from the edifice, and consists primarily of skeletal grainstones and packstones. A clay weathering horizon lies beneath both inner ridge sites. Preliminary interpretations of the sediments recovered at these sites suggest the inner ridge may represent a true reef tract, whereas the outer ridge appears to be an accumulation of shell debris redeposited from another location, perhaps akin to a fore-reef deposit [Haggerty, Premoli-Silva, et al., in press]. Reworked Cenomanian nannofossils found in the overlying Campanian clay layer provide a minimum biostratigraphic age for the basement unit on this guyot. The magnetic inclination data are consistent between all the sites and suggest that the basalts were erupted at a paleolatitude of ~10° S [Nakanishi, Gee et al., 1992]. It is also apparent at all the sites that the carbonate platform subsided into the pelagic realm before the late Paleocene.

The disparate ages, depths, and thicknesses of carbonate and volcanic platforms in the Marshall Islands illustrate the complexity of volcanism and tectonism which affected this region of the Pacific plate. Preliminary results from Leg 144 fail to clearly establish the degree to which hot spot swells in the mid- and Late Cretaceous influenced the history of volcanism, uplift, reef growth, and subsidence across these guyots. For example, the Campanian-age, ferrigenous clay and claystone layer beneath the shallow-water carbonate sediments at Site 873 on Wodejejebato Guyot contains reworked Cenomanian nannofossils,
placing a minimum age on the volcanism of between 91 and 97.5 Ma according to the time scale of Kent and Gradstein, 1985. All the basalt recovered on this guyot is reversely magnetized with a paleolatitude of 10° S [Nakanishi, Gee et al., 1992], suggesting that the volcanism occurred either before or after the Cretaceous Long Normal. The reversed polarity of the basalt places volcanic activity either during Anomaly 34 (< 84 Ma) or during M0 (> 118.7 Ma). While the reworking of Cenomanian nannofossils into Campanian clays argues for the older episode of volcanism, the radiometric age obtained by Pringle [1992] and the paleolatitude estimates of Nakanishi, Gee et al. [1992] conform better to the younger episode of volcanism. Radiometric dating of the basalt obtained during Leg 144 will undoubtedly help clarify the history of volcanism for Wodejebato (and the other guyots sampled during this leg), but a more complete understanding of how this island group formed requires additional analyses of hot spot trends. As a working model for this paper, the history of volcanism and uplift proposed by Lincoln [1990] appears consistent with the multiple episodes of volcanism and plate uplift observed in French Polynesia [e.g., Duncan and McDougall, 1974; Turner and Jarrard, 1982]. Comparisons between the morphology and seismic stratigraphy of guyots drilled during Leg 144 should provide information on the nature and timing of volcanism on these three guyots and their atoll pairs. Indeed, if volcanic overprinting is prevalent throughout the Marshall Islands then plate flexure produced from such loading should be evident in the seismic profiles across the summit plateaus and may play an important role in the subsidence history of these guyots.

3.4. Methodology

The new data presented in this paper consists primarily of SeaMARC II side-scan sonar images and swath bathymetry, SeaBeam multibeam bathymetry, analog and digital single-channel seismic profiles, 6-channel seismic profiles, and 3.5 kHz echosounder profiles. Observations from these data, along with various observations from French
Polynesia, are used to model the flexure caused by volcanic loading around the edifices of Limalok, Lo-En, and Wodejebato guyots. A later section in this paper examines the effects of plate flexure on these three guyots.

3.4.1. Side-scan Sonar Images

Detailed surface mapping of Limalok, Lo-En, and Wodejebato guyots was accomplished with the shallow-towed side-scan sonar system SeaMARC II. This system, prior to its loss at sea, was capable of collecting acoustic backscatter data in swaths up to 10 km wide and bathymetric data with a swath width equal to 3.4 times the water depth. The system operated at frequencies of 11 kHz on the port side and 12 kHz on the starboard side. This paper presents the side-scan images and bathymetry collected by SeaMARC II, and hence a detailed discussion of the system is not given here. Blackinton [1986] provides a thorough description of the system and its capabilities.

Side-scan resolution is a function of many variables and is both range- and orientation-dependent. The resolution of the SeaMARC II system is on the order of several tens of meters [Johnson and Helferty, 1990]. Corrections applied to the backscatter data compensate for errors in slant range (based on water depth below the tow vehicle), beam pattern, bottom tracking, and gain variations. The data are then displayed as gray-scale images. Individual image swaths for the various tracks are assembled into a mosaic of the survey area. In this paper, dark gray to black tones represent areas of high backscatter and shades of light gray to white represent areas of low backscatter and acoustic shadows.

KK820402 was one of the first test cruises for the SeaMARC II system, and during this particular leg only the starboard-side transducer arrays were connected. During MW8805 the starboard arrays malfunctioned, and although it was still possible to collect backscatter data from both sides of the tow-vehicle, the starboard-side data are clipped at both the high and low ends of the recording spectrum. This clipping resulted in images with few data.
values in the mid-range gray tones. Further distortion of the images produced from both cruises occurs by the application of a "flat-bottom" assumption (i.e. the sea floor is flat beneath the tow-vehicle) prior to data recording. The result is an incorrect geographical positioning of features where the bottom topography slopes in relation to the tow-vehicle track (e.g., when surveying parallel to the slope of the guyot, features on the up-dip side of the tow-vehicle are positioned anomalously far from the ship track, and vice-versa).

The measurement of phase differences between incoming acoustic signals at a pair of transducer arrays mounted on either side of the SeaMARC II tow vehicle determine bathymetry. Comparison of the resulting acoustic angle to a "lookup" table (generated from data collected over flat sea floor) results in an estimate of depth at various distances away from the tow-vehicle. These depth estimates, when co-registered with the depths recorded beneath the ship, result in a swath of bathymetry whose resolution is about 2% of the water depth [Blackinton, 1986].

3.4.2. Digital Seismic Data

A single-channel streamer was used to collect digital seismic data during MW8805. When the SeaMARC II tow-vehicle was in the water, the source was a 120 cubic inch air gun; at other times, an 80 cubic inch water gun was used. Processing of the data included band-pass filtering from 10 to 100 Hz, automatic gain control, deconvolution, and muting of water noise. During MW9009, a 6-channel streamer was used with a pair of synchronized 80 cu. in. water guns. In addition to the processing routines applied to the MW8805 data, the 6-channel data were gathered, stacked, and migrated.
3.4.3. Flexure Model and Model Parameters

A number of geophysics and geodynamics texts review in detail the concepts important to evaluating plate elasticity and flexure [e.g., Menke and Abbot, 1990]. The following discussion is but a brief overview of this information.

The most critical and least known piece of information necessary for building a well-constrained flexure model in the Marshall Islands is the time of seamount construction (or plate loading) in relation to existing features on the plate. The scattered and seemingly random distribution of radiometric and fossil ages illustrates that such information on a regional scale is sorely lacking in this area (Figure 3.1). Fortunately, the preliminary age information derived from the Leg 144 drilling provides good age control on the volcanic platforms beneath Lo-En, Wodejebato, and Limalok, especially in relation to previous drilling results on Anewetak and Pikinni atolls. The Leg 144 drilling data show that the volcanic platform of Lo-En is much older than that of Anewetak, and that the carbonate platform on Wodejebato drowned before the end of the Late Cretaceous. By making some assumptions on the age of the edifice beneath Pikinni, the implications of flexure on these two atoll and guyot pairs can be examined. Further extrapolation of the results on Pikinni and Wodejebato allow flexure modeling for the edifices around Limalok.

In addition to the timing of loading, the following plate and load parameters must be specified in order to build a flexure model:

1. Type of plate behavior in response to a surface load (i.e., elastic or viscoelastic).
2. Effective elastic thickness of the plate and the viscous relaxation time.
3. Dimensions of the load (e.g., height, basal radius, slope).
4. Mantle, load, infill sediment, and water densities.
Calment and Cazenave [1986] show that little evidence exists in the geoid data over the Cook-Austral and Society chains to suggest that a viscoelastic model is appropriate for the plate beneath these islands (i.e., plate thickness apparently decreasing with age) and hence an elastic plate model will be used in all flexure calculations. Assuming that 1) the earth's lithosphere acts as an elastic shell overlying a fluid medium, 2) the flexural rigidity of the plate is constant, and 3) there are no horizontal or tensile stresses within the plate prior to the flexural deformation, then the differential equation for three-dimensional plate flexure simplifies to:

$$D \nabla^4 w + (\rho_m - \rho_i) gw = q \quad (1)$$

where $D$ is the flexural rigidity of the plate ($D = E h^3 /[12(1-\nu^2)]$), $w$ is the vertical displacement in response to a load, $\rho$ is the density of the mantle and infill (respectively), $g$ is the gravitational constant (9.8 m/s$^2$), and $q$ is the load ($[(\rho_l - \rho_w)gh]$. The term $(\rho_m - \rho_i) gw$ accounts for the buoyancy force acting beneath the load as material filling the moat displaces mantle material within the region of deflection. The "h" used in the definition for the flexural rigidity of the plate ($D$) represents the elastic thickness of the plate ($T_e$), whereas the "h" in the definition for the load ($q$) represents the height of the load.

Often, it is faster to estimate three-dimensional plate flexure by finding the response function of a constant rigidity plate to a point load (the Green's function of the system). Since a unit point load on a homogeneous plate produces an axisymmetrical deflection, the response (Green's) function can be written in polar coordinates which allows the derivation of the Green's function in spatial terms:

$$w(r) = \frac{q_0 \gamma^2}{2\pi} \frac{K_{ei}(\gamma r)}{(\rho_m - \rho_i) g} \quad (2)$$
where $\gamma = 3$-dimensional flexure wave number = $\left[ (\rho_m - \rho_i) g / D \right]^{1/4}$

$\text{Kei}(\gamma r)$ = the zero-order Kelvin-Bessel function describing axisymmetric deflections about a point load.

Numerical evaluation of the deflection for any load is possible by approximating the load as a number of point loads and by summing the deflections from all these points loads as observed at various points across an area (i.e., convolving the load with the Green's function).

The remaining plate and load parameters used in this paper are approximated from flexure modeling in French Polynesia. Models of SEASAT satellite-derived geoid height data over the southeast Pacific provide estimates of the effective elastic plate thickness, which appears to vary between ~8 km beneath the Society, Marquesas, and Pitcairn seamounts and ~13 km beneath the Cook-Austral chain [Calmant and Cazenave, 1986; 1987]. More recent studies using ship-based gravity measurements suggest elastic plate thickness may be on the order of 23 +/- 2 km around the Society Islands and 18 +/- 2 km around the Marquesas Islands [Filmer et al., 1993]. The only estimate of plate thickness in the Marshall Islands comes from gravity measurements over Woden-Kopakut Guyot (formerly Ratak) which suggest the volcano was constructed on a plate < 15 km thick [Smith, 1990]. Lambeck [1981] modeled flexure in the Society Islands using a number of different flexure indicators (e.g., arch distance and amplitude, geoid measurements, regional bathymetry) and different load and fill geometries. He arrived at preferred estimates for the density of the load around 2500 kg/m$^3$ and the flexural rigidity of the plate $\sim 3 \times 10^{29}$ N m. These values are very similar to those used by McNutt and Menard [1978]: 2800 kg/m$^3$ and $2 \times 10^{29}$ N m. Given that this paper does not establish independent estimates of elastic plate thickness for the Marshall Islands, the models presented here vary
the plate thickness between 10 km and 20 km. The models also use a density of 3300 kg/m³ for the mantle, 2700 kg/m³ for the load, 2500 kg/m³ for the infill, and 1000 kg/m³ for water. More extreme values for the load and infill densities (maximum values of 3000 kg/m³ and 2500 kg/m³, and minimum values of 2500 kg/m³ and 2300 kg/m³, respectively) cause a change flexure wavelength of <10 %.

Modeling the original size and shape of ancient volcanic loads requires a number of assumptions. The atolls of Anewetak and Pikinni presumably grew on top of the eroded summits of shield volcanoes. Consequently, simple axisymmetrical cones of a given height and basal diameter can be used to represent the volcanic platforms (Figure 3.3). The height of each load can be estimated by summing the depth of the undeformed plate at the time of loading with the height of the volcano above sea level. Seismic data between Anewetak and Pikinni, well away from the flexural deformation caused by these loads, show three distinct reflectors beneath the 5120 m deep sea floor (Figure 3.3). The lowermost reflector is assumed to define the undeformed plate depth, which appears around 7.6 sec (TWT). Drilling at Site 462 in the Nauru Basin sampled a similar sequence of reflectors and presumably sediments [Larson, Schlanger et al., 1981], and hence velocities of 1700 m/sec and 2300 m/sec were assumed for the two major units overlying the deep reflector. The total sediment thickness assuming these velocities is ~775 m, which produces an undeformed plate depth of 5895 m. This depth estimate does not account for the isostatic effects of sediment loading. Using a 600 m/sec correction value to compensate for the sediment loading [Crough, 1983] and subtracting the amount of subsidence brought about by thermal cooling of the plate (~1300 m for both Anewetak and Pikinni as shown by the depth to top of the volcanic platforms beneath these two atolls), the corrected plate depth appears to be ~4125 m at the time of loading. This depth is consistent with the 4000 m to 4300 m sea floor depths around present-day French Polynesia [e.g., Cheminee et al., 1991; Stoffers et al., 1991]. As the volcanic platforms were presumably shield volcanoes
Figure 3.3. Pictorial representation of load parameters used for flexure modeling in the Marshall Islands. Attempts to model plate flexure from volcanic loading require a good understanding of the load dimensions. In the case of the Marshall Islands, number of assumptions must be made about the load because of erosion and sediment deposition subsequent to the time of load emplacement. Estimated parameters include the original height of the volcano (the observed height plus the amount buried by sediment and the amount removed during subaerial exposure), the density of the edifice, the density of the sediment filling the moat, the density of the plate, and the density of the mantle.
\[ D \nabla^4 w + (\rho_m - \rho_i) g w = q \]

- Physical strength of the plate
- Bouyancy force
- Load
- Surface load \( \rho_1 \)
- Moat infill \( \rho_i \)
- Load infill \( \rho_{li} \)
- Level of undeformed plate

**Notations:**
- \( D \): Differential operator
- \( \nabla^4 \): Laplacian operator
- \( w \): Displacement
- \( \rho_m \): Mantle density
- \( \rho_i \): Infill density
- \( \rho_{li} \): Load infill density
- \( q \): Load

**Layering:**
- Volcanic edifice
- Marine sediments
- Sediments filling moat
- Buried or eroded portions of edifice

**Diagram Elements:**
- Estimated erosion
- Observed load height
- Plate
- Mantle
and not flat-topped edifices at the time of loading, some mass must be added to the tops of the observed platforms to account for subaerial erosion. The island in French Polynesia which appears most comparable in size to both Anewetak and Pikinni is Tahiti-Nui (Table 1), the larger of the two cones forming the island of Tahiti in the Society chain [Duncan and McDougall, 1974]. Tahiti-Nui rises 2241 m above sea level, and hence an additional 2300 m is added to the height of the model loads resulting in a grand total of 6425 m.

The basal diameter of the volcanoes, buried beneath ~775 m of sediment, can not be measured directly from bathymetry maps. For the modeling presented in this paper the basal diameter was estimated by measuring the slope of the flank between two widely-spaced contours encircling the edifice (e.g. 1000 m and 2000 m contours) and then applying this slope to the portion of the edifice buried beneath the sediments. This method resulted in basal diameters of 100 km for Anewetak Atoll and 90 km for Pikinni Atoll. The shape and size of the loads representing Lo-En and Wodejebato guyots are not important as the modeling conducted in this paper only examines the affects of loading by Anewetak and Pikinni.

Prior to presenting the results from the flexure modeling, it is critical to describe the morphology and seismic stratigraphy of the drowned platforms drilled during Leg 144 as these observations will be used in part to test the model results. As discussed later in the text, the flexure models also help explain some of the anomalous features observed on Limalok, Lo-En, and Wodejebato.

3.5. Guyot Descriptions

Limalok Guyot, located between 5° 30' and 5° 45' N, and 172° 10' and 172° 30' E, lies in the southernmost portion of the Ratak chain (Figure 3.1). Nakanishi et al. [1992] identified Anomaly M22 near this portion of the Marshalls, making the crust beneath Limalok ~155 m.y. old. The summit plateau narrows slightly from 27 km in the north to
Table 3.1: Comparisons between selected Pacific island chains.

<table>
<thead>
<tr>
<th></th>
<th>Summit Plateau</th>
<th>Volcanic Episodes</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Area (km²)</td>
<td>Length (km)</td>
<td>Ages (m.y.)</td>
</tr>
<tr>
<td><strong>Marshall Islands</strong></td>
<td></td>
<td>Width (km)</td>
<td></td>
</tr>
<tr>
<td>Limalok</td>
<td>636</td>
<td>36</td>
<td>&gt; Paleocene (f)</td>
</tr>
<tr>
<td>Lo-En</td>
<td>823</td>
<td>42</td>
<td>pre-Turonian (f)</td>
</tr>
<tr>
<td>Wodejebato</td>
<td>505</td>
<td>38</td>
<td>Cenomanian (f) ~86 (r)</td>
</tr>
</tbody>
</table>

|                          |                |                  |                           |                                                                 |
| **French Polynesia**     |                |                  |                           |                                                                 |
|                          |                |                  |                           |                                                                 |
| **Cook Islands**         |                |                  |                           |                                                                 |
| Raratonga                | 65             | 11.0             | 1.6-2.3 c                 | Geomorphically young island with very little reef development along its shores. Located SE of numerous makatea islands. |
| Aitutaki                 | 14             | 7.8              | 7.39 to > 8.73 c          | Very nearly an atoll, with carbonate platform area approaching 95 km². Appears to have two episodes of volcanism separated by ~6 m.y. |

|                          |                |                  |                           |                                                                 |
| **Austral Islands**      |                |                  |                           |                                                                 |
| Rimataro                 | 35             | 11.0             | > 4.78 +/- 0.42           | Uplifted island with 83 m high volcanic summit and 11 m high makatea. |
| Rurutu                   | 35             | 11.0             | 1.06-1.12                 | Uplifted remnant of single shield volcano with limestone cliffs up to 100 m high surrounding much of the island. Clay transition zone up to 3 m thick. |

|                          |                |                  |                           |                                                                 |
| **Society Islands**      |                |                  |                           |                                                                 |
| Tahiti-Nui               | 1014           | 37.9             | 36.2                      | Larger of the two cones forming the island of Tahiti. Reaches a maximum elevation of 2241 m. Resembles the size of the Marshall Islands. |
| Tahiti-Iti               | 309            | 26.6             | 17.5                      | Smaller of the two cones forming Tahiti. Also referred to as Taiarapu. |
| Moorea                   | 128            | 14.6             | 14.2                      | Lies ~16 km NW of Tahiti. Approximately 1/3 of the cone appears to have eroded away. |
| Huahine                  | 50             | 10.7             | 5.7                       | Consists of twin islets encircled by a common barrier reef. The islets represent the remnants of a single cone split by faulting. |

|                          |                |                  |                           |                                                                 |
| **Gambier Islands**      |                |                  |                           |                                                                 |
| Mururoa                  | 154            | 27.5             | 11.1                      | Atoll on which volcanic basement shallows by over 200 m towards the center of cone, and a clay transition zone reaches thicknesses of up to 90 m. |

|                          |                |                  |                           |                                                                 |
| **Hawaiian Islands**     |                |                  |                           |                                                                 |
| Maui                     | 1887           | 77               | 67                        | Perhaps the Hawaiian Island most similar to the Marshall Islands atoll and guyot pairs. |
| Mauna Kea                | 2382           | 82               | 40                        | One of the shield volcanoes composing the island of Hawaii. Note the size difference between this one shield and the above listed Marshall Islands. |
15 km in the south while measuring ~36 km in length (Figure 3.4, Table 2). A volcanic ridge extends ~67 km to the northwest, attaching Limalok to Mili and Knox atolls. In the side-scan images (Figures 3.5 and 3.6), a high-backscatter (dark) band along the perimeter of the summit plateau marks areas where the carbonate complex crops out from the overlying pelagic sediment cap. Broad terraces, 1 to 4 km wide and up to 12 km long, lie outside the scalloped portions of the band (Table 2). These terraces are most prominent along the south and west sides of the plateau.

Lo-En Guyot, centered about 10° 10' N and 163° 52' E, lies ~165 km southwest-southeast of Anewetak Atoll in the cluster of volcanoes marking the western boundary of the Marshall Islands (Figure 3.1). Extrapolation of magnetic lineations identified to the south of these two edifices [Nakanishi et al., 1992] produces an estimated plate age of >160 m.y.. Prior to the site survey work conducted during MW8805 little was known about the detailed morphology of Lo-En. Bathymetry maps constructed from echo-sounder profiles showed a ridge apparently connecting Anewetak and Lo-En but it was not possible to resolve distinct features until the collection of side-scan sonar images and swath bathymetry during MW8805 (Figure 3.7). The summit plateau of Lo-En varies between 30 km and 40 km in diameter, with the 1400 m contour generally defining the first major slope break (Table 1). In the side-scan images the summit plateau appears as a region of low-backscatter disrupted occasionally by high-backscatter cone- and lobe-shaped features (Figures 3.8 and 3.9). The low-backscatter values across the plateau result from a thick layer of pelagic sediments. One of the cone-shaped features crossed on the plateau has a summit depression filled with pelagic sediment (Figure 3.10) and a relatively large-amplitude magnetic signature. Dredging on this cone during MW8805 recovered manganese-encrusted basalt breccia containing altered clasts of basalt with oxidized phenocrysts of olivine. Additional dredging during MW9009 recovered only manganese crusts.
Figure 3.4. Bathymetry map of Limalok Guyot constructed from echosounder depths. Shallow-water carbonates recovered from this guyot are primarily Eocene in age. The small triangles and accompanying circles show the location and lithology of these dredge samples. Grey lines mark the location of seismic profiles collected during KK810626 (leg 2) and Leg 144, with the thicker, annotated lines showing profiles presented in this paper.
Figure 3.5. Side-scan sonar images over Limalok Guyot, collected during KK820402.

The images show a scalloped high-backscatter (dark tones) band inset from the edge of the summit plateau. The band marks the edge of the carbonate platform capping this edifice, and the scalloped configuration of the band results from block-faulting of the platform (see Figure 3.6 for geologic interpretation).
Figure 3.6. Geologic interpretation of Limalok side-scan images showing the major morphologic features as related to their backscatter characteristics.
Limalok Guyot
Geologic interpretation of side-scan images

- Pelagic sediment
- Exposed carbonate platform
- Channels

 Fault block (carbonates)
Exposed edge of carbonate platform
Fault blocks (carbonates)
Fault scarp (??)
Table 3.2. Flank Ridge and Fault Block Dimensions

<table>
<thead>
<tr>
<th>Flank Ridge Dimensions</th>
<th>Fault Block Dimensions</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Area (km²)</strong></td>
<td><strong>Length (km)</strong></td>
</tr>
<tr>
<td>Limalok</td>
<td>-------</td>
</tr>
<tr>
<td>Lo-En</td>
<td></td>
</tr>
<tr>
<td><strong>South FR</strong></td>
<td>181</td>
</tr>
<tr>
<td>Wodejebato</td>
<td></td>
</tr>
<tr>
<td><strong>South FR</strong></td>
<td>73</td>
</tr>
<tr>
<td><strong>West FR</strong></td>
<td>~23</td>
</tr>
<tr>
<td><strong>North FR</strong></td>
<td>54</td>
</tr>
<tr>
<td><strong>Northeast FR</strong></td>
<td>12</td>
</tr>
<tr>
<td>~29</td>
<td>~6.6</td>
</tr>
</tbody>
</table>
Figure 3.7. SeaMARC II bathymetry over Lo-En Guyot. The side-scan sonar images and swath bathymetry collected during MW8805 provided the first detailed coverage over this edifice. Superimposed on the SeaMARC II bathymetry are the location and lithology of dredge samples recovered during MW8805. The area shown in subsequent side-scan images (Figure 3.8) does not include the small unnamed guyot to the west-southwest of Lo-En.
Foraminifer limestone

Conglomerate (basalt)

Breccia (basalt)

Basalt

MW8805

MW9009

Lo-En Guyot

Contour interval = 100 m

Ridge extending towards Anewetak Atoll

Small seamount straddling ridge

Trough separating two edifices
Figure 3.8. Side-scan images over Lo-En Guyot. The images show a number of high-backscatter cone- and lobe-shaped features cropping out from the pelagic sediments covering the summit plateau (see Figure 3.9 for the location of these features). Relatively small-scale terraces lie perched along the southwest portion of the edifice, and a flank ridge extends to the south.
Figure 3.9. Geologic interpretation of the side-scan images over Lo-En showing the major morphologic features as related to their backscatter characteristics.
Lo-En Guyot
Geologic interpretation of side-scan images

- Pelagic sediments
- Channels
- Lobate features

Seamount straddling ridge extending towards Anewetak Atoll
Low-backscatter areas
Exposed portions of volcanic ridge
Volcanic cones (??)
Volcanic flows or sediment debris flows (??)
Volcanic ridge (??)
Fault blocks
Acoustic shadow
Flank ridge

10° 00'
10° 30' N
162° 30 E'
163° 00'
Figure 3.10. Sidescan image (a) and echosounder profile (b) across the summit plateau of Lo-En Guyot. MW8805 crossed directly over one of the cone-shaped, high-backscatter features shown in the side-scan images. In the 3.5 kHz echosounder profile, the cone shows a summit depression filled with pelagic sediments. Basalt breccia recovered in RD36 and the magnetic high over this cone suggests it is volcanic in origin.
Wodejebato Guyot, located at 12° 00' N and 164° 50' E, lies in the northernmost portion of the central Ralik Chain, ~60 km northwest of Pikinni Atoll (Figure 3.1). As in the case of Lo-En, the plate age beneath Wodejebato must be derived through the extrapolation of magnetic anomalies identified in the southern Marshall Islands [Nakanishi et al., 1992], resulting in an estimated plate age of > 160 m.y.. The summit plateau of Wodejebato is about 43 km long and increases in width from less than 12 km in the southeast to greater than 25 km in the northwest (Figure 3.11, Table 1). A change in the slope gradient divides the flanks of the guyot into a steep upper slope (~20° -24°) which gives way to a more gently inclined lower slope (~7°). In general, the transition depth between upper and lower slopes is around 2500 m. Four flank ridges extend from the main body of the edifice, and along with the volcanic spur attaching Wodejebato to Pikinni give the guyot a distinct "starfish" shape. Side-scan images over the summit plateau (Figures 3.12 and 3.13) show a high-backscatter band inset slightly from the upper slope of the guyot. Similar to Limalok Guyot, this high-backscatter band represents areas where the carbonate platform crops out from the overlying pelagic sediments. Bergersen [1993] provides a detailed discussion of the morphology, internal structure, and geologic history of Wodejebato.

Physical and compositional differences between these three guyots relate directly to differing episodes of volcanism, uplift, subsidence, and erosion. For example, the shallow-water carbonates across the summit plateaus of Limalok and Wodejebato suggest a much different history than the cone- and lobe-shaped features cropping out from the pelagic sediments on Lo-En. The degree to which the history of volcanism, uplift, and subsidence coincides across these three edifices can be examined by comparing the following features found in common: ridges extending from the flanks of the edifice, terraces or benches located down-slope from the first slope break, and relatively flat
Figure 3.11. SeaMARC II bathymetry over Wodejebato Guyot. The location and lithology of dredge samples collected during MW8805 and the Mid-Pac Expedition are shown by the accompanying triangles and circles. Lincoln [1990] used the fossil components of dredge RD50 to construct his model for a mid-Cretaceous/Late Cretaceous evolution of the Marshall Islands (see text). The radiometric age for volcanism on this guyot (~86 Ma) comes from the basalt recovered in dredge MP43.D [Pringle, 1992]. The bathymetry clearly shows the five flank ridges extending from the central edifice. Note that the northeast ridge consists of a topographic high detached from the main edifice.
Figure 3.12. Side-scan images over Wodejebato Guyot. The images show a high-backscatter band slightly inset from the edge of the summit plateau. As in the case of Limalok, this band marks areas where the carbonate platform crops out from the overlying pelagic sediments. A second high-backscatter band appears along the shelves formed by the north and northeast flank ridges and the ridge extending towards Pikinni Atoll. In multibeam bathymetry maps and seismic profiles, these bands appear as topographic highs.
Wodejebato Guyot

12° 15'

164° 30' E

165° 00'
Figure 3.13. Geologic interpretation of the side-scan images over Wodejebato showing the major morphologic features as related to their backscatter characteristics.
Geologic interpretation of side-scan images

Wodejebato Guyot

North Flank Ridge

West Flank Ridge

Outer Perimeter Ridge

Northeast Flank Ridge

Inner Perimeter Ridge

South Flank Ridge

Legend:
- Polagic Sediments
- Exposed Carbonate Platform
- Slumped Platform Sediments?
- Volcanic Features?
- Channels

Coordinates:
- 12° 15'
- 11° 45'
- 164° 30'
- 165° 00'

98
3.5.1. Flank ridges

Rift zones as observed in the Hawaiian Islands are zones of volcanic features associated with underlying dike complexes [Bates and Jackson, 1980]. Vogt and Smoot [1984] introduced the concept that ridges extending from the flanks of seamounts and guyots (their "flank rift zones") are analogous to rift zones observed on volcanic islands. The only evidence they cite supporting this relationship is the morphology of the ridges as observed in multibeam bathymetric maps. Seamounts located off the flanks of a guyot or the headwall scarps formed by large-scale landslides may also appear in bathymetry maps as relatively continuous ridges, thus complicating the apparently simple correlation between volcanic rift zones (or zones of active volcanism) and morphologic rift zones (or ridges extending away from a central edifice). Consequently, this paper uses the term "flank ridge" to avoid any confusion between the assumed and known origin for these features.

Well-defined ridges extend between all three of the atoll and guyot pairs discussed in this paper. On Limalok and Lo-en guyots, these ridges extend from the north-northwest flanks towards Mili and Anewetak atolls, respectively (Figures 3.4 and 3.7). The distance between Limalok and Mili (67 km) is comparable to the total size of the two cones forming the island of Tahiti in French Polynesia (65 km, Table 1). Lo-En and Anewetak, on the other hand, are separated by a distance of approximately 167 km. High-backscatter features crop out from the pelagic sediments along the top of this ridge, and a small seamount straddles the ridge at about its mid-point (Figures 3.8 and 3.9). A second, smaller ridge extends ~14.5 km from the south flank of Lo-En, dipping away from the edifice at an average of 6° (Table 2). High-backscatter features also crop out from the summit plateaus. The information derived from these features sheds additional light on the evolution of each edifice and on the Marshall Islands as a whole.
pelagic sediments along the top of this ridge, forming what appears to be a north-south line of volcanism across the summit of Lo-En (Figures 3.8 and 3.9).

The number and morphology of flank ridges on Wodejebato Guyot are distinctly different from those observed on Limalok and Lo-En. Five ridges extend from the main edifice of Wodejebato, with the southeast ridge attaching to Pikinni Atoll over a distance of 64 km (Figure 3.11). All the other flank ridges are 11 km to 13 km long, but the width and the area of each varies considerably (Table 2). The southern and northern flank ridges form broad shelves which deepen slightly away from the summit plateau (the inclination of the southern ridge is < 2°, while that of the northern ridge is even less). The shelf formed by the northern ridge is relatively featureless in the side-scan images, unlike the southern ridge where a number of high-backscatter cone- and lobe-shaped features crop out from the thin covering of pelagic sediments (Figures 3.12 and 3.13). Bergersen [1993] interpreted these features as a combination of volcanic cones and redistributed summit plateau sediments as dredges along the distal edges of the ridges recovered only basalt and basalt breccia. The absence of shallow-water carbonates in these dredges suggests that the carbonate platform does not extend out on to the shelves. The western flank ridge was only partially mapped in the side-scan images collected during MW8805, but it does not appear as wide or as planar as its southern and northern counterparts. Along the northeast flank, a narrow volcanic spur (~1.7 km wide) attaches a 300 m high "hill" to the main edifice (Figure 3.11). This particular flank ridge may represent a small seamount connected to the main edifice by volcanic flows and volcaniclastic sediments deposited during the constructional phase of Wodejebato [Bergersen, 1993].

Vogt and Smoot [1984] discuss factors controlling the number and length of flank rifts zones on edifices in the Emperor, Geisha, Michelson, Dutton, and Mid-Pacific chains, placing major emphasis on the pressure differential between the magma source and the conduits through which the magma flows. They conclude that the height of the edifice
correlates with the length of the rift zones and is independent of the number of rift zones. In Hawaii, volcanoes usually possess only 2 or 3 rift zones (J. Sinton, personal communication). Recent SeaBeam surveying over Society and Austral hot spot volcanoes [Cheminee et al., 1989; Stoffers et al., 1989; Binard et al., 1991] shows 5 to 8 flank ridges extending away from these edifices over distances of up to 20 km, although it is debatable whether all these ridges are true rift zones without additional evidence to corroborate the bathymetry observations (e.g., samples from dredging, modeling of gravity over the ridges). Until such data show that the plumbing systems of volcanoes formed in French Polynesia behave differently than those in Hawaii, it seems reasonable to assume a 2 or 3 rift zone system applies to the shield volcanoes forming the Marshall Islands. Accordingly, the number of flank ridges extending away from an edifice may provide information on the number of ancient shield volcanoes composing that edifice, assuming these flank ridges are identified correctly as rift zones. In the cases of Limalok and Lo-En, only 1 or 2 ridges are visible. The atoll-linked ridges extending from the north flanks of Limalok and Lo-En and the southeast flank of Wodejebato probably do not represent true rift zones. They more likely result from an accumulation of volcanic flows and volcaniclastic sediment shed from the two volcanoes during their shield-building stages (similar in a sense to the twin volcanoes forming such islands as Maui in the Hawaiian chain and Tahiti in the Society chain). Wodejebato Guyot, with four flank ridges (excluding the atoll-attached ridge), holds the most promise for being the product of more than one volcano, although if the northeast flank ridge is a separate seamount then the number of "rift zones" observed on this edifice conforms to the Hawaii model. An alternative "origin" for these ridges is that they are not constructional features at all, but rather simply erosional remnants.
3.5.2. Terraces

Terraces or benches observed in seismic profiles along the flanks of guyots have traditionally been thought to be wave-cut features related to relative highs or lows in sea level [e.g., Pratt, 1963; Budinger, 1967]. Side-scan sonar images and swath bathymetry maps often show that the lateral extent of these benches is discontinuous around the edge of an edifice [e.g., Lonsdale, 1972; Bergersen, 1993], an observation inconsistent with a formative mechanism presumably affecting all sides of the island. On the south flank of Wodejebato Guyot a large block of platform sediments lies down-slope from the edge of the summit plateau. In the side-scan images over this area, a low-backscatter terrace separates the edge of the carbonate platform (high-backscatter band) from the guyot flank (Figure 3.14). Seismic profiles show that the flat-lying lagoonal sediments crop out along this scarp, as opposed to truncating against a perimeter ridge present along all the other flanks. The correlation between the seismic stratigraphy of the plateau sediments and that of the terrace sediments suggests the low-backscatter terrace represents a down-dropped block of lagoonal sediments. Apparently faulting and subsequent erosion have removed any perimeter ridge that once existed in this area.

Observations over Limalok and Lo-En guyots provide additional support for the assertion that faulting continues to shape the morphology of volcanoes even after they have moved away from the hot spot swell. For example, on Limalok Guyot arc-shaped "terraces" (visible in side-scan images) lie outside of the high-backscatter band marking the exposed edge of the carbonate platform (Figures 3.5 and 3.6). Seismic profiles across these "terraces" show large fault-blocks down-dropped relative to the summit plateau (Figure 3.15 and 3.16). In some cases (Figure 3.16) the ship crossed a fault-block at an oblique angle, resulting in the appearance of a "ridge" outside of the down-dropped block. Although the resolution of the single-channel data is too poor to make a direct correlation between the seismic stratigraphy of the fault-blocks and that of the platform carbonates (as
Figure 3.14. Side-scan image (a) and 6-channel seismic profile (b) over the south flank of Wodejebato Guyot. A low-backscatter terrace visible in the side-scan images along this flank appears as a fault-block in the seismic data. The seismic stratigraphy of the fault-block matches that of the lagoon sediments. Faulting also appears to disrupt the carbonate platform farther back from the edge of the summit plateau. Side-scan image location shown in Figure 3.11.
Figure 3.15. Single-channel seismic profile (a) and interpretation (b) over the west and south flanks of Limalok Guyot. The faulting noted along the south flank of Wodejebato also occurs on the south and west flanks of Limalok Guyot. This particular profile only shows a small portion of the faulting along the west flank (left side of figure). Smaller faults offset platform reflectors along the south flank. Some of the platform reflectors appear to onlap the reflector interpreted as volcanic basement. Profile location shown in Figure 3.4.
Limalok Guyot
(Single-channel seismic profile)

Interpretation

Pelagic sediments
Carbonate platform
Onlapping units
Fault block (?)
Volcanic basement

Faults
Carbonate platform
Volcanic basement
Figure 3.16. Single-channel seismic profile (a) and interpretation (b) across the summit plateau of Limalok Guyot. This particular seismic profile crossed one of the terraces seen in the side-scan images along the west flank of the guyot, showing it to be a large fault-block lying down-slope from the edge of the carbonate platform. The topographic high outside of the downdropped fault-block results from an oblique ship crossing across a portion of the plateau unaffected by the faulting. Note what appears to be a perimeter ridge along the east side of the plateau. As in the case of profile LimA-LimA', some of the carbonate platform reflectors onlap the basement reflector. Profile location shown in Figure 3.4.
Limalok Guyot
(Single-channel seismic profile)

Interpretation

Pelagic sediments
Perimeter ridge
Carbonate platform
Onlapping carbonate unit or prograding volcanicslastic unit
Fault block
Volcanic basement
Volcanic basement

0900 Z 0930 Z 1000 Z 1030 Z
in the case of Wodejebato), the side-scan images show that the blocks are discrete, discontinuous features along the edge of the guyot and hence not wave-cut features. The scarps adjacent to the fault-blocks range between 75 m and 150 m high. Similar to Wodejebato, smaller-scale faulting disrupts the platform carbonates landward of the down-dropped blocks. On Lo-En Guyot, the low-backscatter terraces visible in side-scan images along the west flank (Figures 3.8 and 3.9) are presumably fault-related. Faulting also disrupts the more central portions of the summit plateau across this guyot (Figure 3.17). The absence of platform carbonates on Lo-En suggests that such block-faulting is not unique to edifices capped by a drowned carbonate platform, although the size of the blocks appears smaller (Table 2).

The paucity of side-scan sonar data and detailed bathymetric maps over sunken islands in French Polynesia prohibits a direct comparison with the fault-blocks observed in the Marshall Islands. Given that well-documented large-scale landslides occur off the islands of Hawaii [Moore et al, 1989] and that similar faulting (and presumably landslide generation) appears common throughout the subaerial islands of French Polynesia (e.g., Moorea and Huahine islands; Table 1) a continuation of this mass-wasting process seems the most likely mechanism responsible for the fault-blocks perched along the flanks of Marshall Islands guyots. The timing and triggering mechanism of the faulting remains less clear. It evidently happened after the growth of the carbonate platform as these sediments are part of the fault-blocks. In the case of Wodejebato, the absence of a perimeter ridge along the south flank suggests that the faulting occurred after the platform was drowned, otherwise a new perimeter ridge would presumably have formed on the older carbonate sediments. One possible triggering mechanism is sediment overburden as supplied by the carbonate platform, although this mechanism apparently does not apply to Lo-En. Another possibility is that the volcanic flows and volcaniclastic sediments forming the original shield volcano are inherently unstable and therefore susceptible to slope oversteepening.
Figure 3.17. Single-channel seismic profiles across the summit plateau of Lo-En Guyot.

Profile LoenA-LoenA' shows the volcanic cone which crops out from the pelagic sediments, and profile LoenB-LoenB' is a portion of the seismic line from which Site 872 was originally selected. Faulting clearly offsets reflectors across the plateau. The deep reflector annotated in B-B' was first identified in a migrated 6-channel profile (see Figure 3.18) and suggests a change in the character of volcanic basement. Profile locations shown in Figure 3.7.
Obviously fault-blocks of this sort supply no information on the history of volcanism experienced by a particular edifice, but they give some indication of flank stability away from the hot spot swell. Unfortunately, the paucity of high-resolution bathymetry and side-scan sonar images over old seamounts makes it difficult to determine the prevalence of similar features on other edifices.

3.5.3. Summit Plateau

As mentioned previously, one of the keys to reconstructing a Cretaceous sea level curve based on the sediments drilled during Leg 144 is the separation of sea level fluctuations of a eustatic origin from those brought about by such tectonic forces as plate uplift and plate flexure. The sequences of sediment sampled across the summit plateaus of Limalok, Lo-En, and Wodejebato represent our best record of the sum effects of volcanism, uplift, and subsidence. Consequently, the first step towards removing, or at least better understanding, the tectonic sea level signal associated with these guyots is to examine the topography and the internal structure of the summit plateaus. Comparisons between the drowned platforms in the Marshall Islands and subaerial islands in the Hawaiian and French Polynesian chains should help identify subsidence and erosion patterns typical for volcanic islands, atolls, and uplifted atolls.

Lo-En Guyot appears to be the most unique platform drilled during Leg 144 in that the summit plateau is apparently devoid of shallow-water carbonate sediments. Preliminary interpretations of the basalt units sampled at Hole 872B suggest that the flows forming the volcanic platform represent the very latest shield stage or alkalic-cap stage of hot spot volcanism, and probably were erupted in a subaerial environment [Haggerty, Premoli-Silva et al., in press]. Core samples at Site 872 also suggest that a long hiatus in sediment deposition (> 50 m.y.) exists between the late Oligocene to Pleistocene pelagic sediments and the pre-Turonian volcanic platform [Haggerty, Premoli-Silva et al., in press]. Single-
and multi-channel seismic data across Lo-En support the drilling results with regards to the apparent absence of a carbonate platform beneath the pelagic sediments (Figures 3.17 and 3.18). The high-backscatter cone- and lobe-shaped features appearing in the side-scan images (Figure 3.8) are most likely volcanic in origin as suggested by the high magnetic signature and the basalt breccia recovered from the cone surveyed during MW8805 (RD36). This particular cone is roughly the size of Diamond Head, a well-known post-erosional cone in the Hawaiian Islands. Unfortunately, the high-degree of alteration of the breccia recovered in RD36 precludes any detailed interpretation of the cone's origin (e.g., a tuff cone formed during secondary eruptions as in the case of Diamond Head). Both the absence of a carbonate platform and the presence of cones across the relatively flat summit plateau contradict traditional views of island formation followed by slow plate subsidence, edifice truncation, and shallow-water carbonate accumulation. Lo-En must have subsided at such a rate, or been constructed at such a latitude, that extensive erosion of the cone, weathering of the volcanic platform, and reef growth across the summit plateau were prohibited. Alternatively, the volcanism responsible for the cones could have occurred in a submarine environment, thus circumventing subaerial erosion.

The stratigraphic units encountered on the drowned carbonate platforms of Wodejebato and Limalok are very similar to those observed on uplifted or drilled atolls in French Polynesia. For example, Rurutu Island in the Austral chain is an uplifted, deeply dissected shield volcano with younger flows (1 to 2 m.y. old) superimposed on an older edifice (between 8.6 and 12.5 m.y. old; Turner and Jarrard, 1982). Shallow-water carbonate sediments form 90 to 100 m high cliffs around the entire island, and a 3 m thick stratified layer of clay marks a transition zone between the volcanic and limestone units [Duncan and McDougall, 1976; Bardintzeff et al., 1985]. This stratigraphic sequence resembles the sequence of sediments drilled at Site 873 on Wodejebato Guyot, although the transitional clay layer is much thicker (20 m) than on Rurutu. The clay transition zone is
Figure 3.18. 6-channel seismic profile over northern half of Lo-En Guyot. This profile shows the "sub-basement" structure of this guyot, although the significance of the "deep reflector" is still poorly understood. It may represent some type of flow unit overlain by volcaniclastic sediment. A similar reflector also appears on Wodejebato Guyot. Profile location shown in Figure 3.7.
Lo-En Guyot
6-channel seismic profile

Interpretation

Pelagic sediments
Volcaniclastic layer (?)
Deep reflector
also much thicker on Limalok Guyot (30 m). Drilling on Muruoa Atoll in the Gambier-Pitcairn chain provides another example of how French Polynesian islands are similar to the volcanic platforms beneath Wodejebato and Limalok (Table 1). Muruoa Atoll was extensively sampled through drilling along an ENE transect [Buigues, 1985]. The depth at which these drill holes encountered basalt shows that volcanic basement on this edifice shallows towards the center of the summit plateau. In a similar fashion, the thickness of the carbonate platforms on Limalok and Wodejebato increase towards the edges of their summit plateaus (Figure 3.19). On Limalok, the basement high traverses the central portion of the summit plateau, possibly merging with the ridge extending towards Mili Atoll (Figure 3.19a). On Wodejebato, the basement high appears to consist of two separate peaks offset to the northeast, although these peaks may be an artifact of widely-spaced seismic profiles (Figure 3.19b). Seismic profiles over both Limalok and Wodejebato show subunits within the carbonate platform onlapping these central basement highs (e.g., Figures 3.15 and 3.20).

The similarities noted between the volcanic platforms of modern atolls and those beneath Wodejebato and Limalok also extend to the morphology and seismic stratigraphy of the carbonate platforms, at least at the resolution offered by the existing data. In modern atoll environments, a raised reef rim encircles a central lagoon, commonly dotted with patch reefs and small islands. Presumably the lagoon sediments slowly covered basement highs as the edifice subsided with the plate, and the coral organisms kept pace with this relative rise in sea level. Relief across the top of the drowned platforms of Wodejebato and Limalok (5 m and 10 m) is typical of depth ranges for the lagoons in most atoll settings [e.g., Weins 1962]. Seismic profiles across the carbonate platform of Wodejebato (Figure 3.20) show two primary sediment packets: a lower unit consisting of subunits onlapping a central basement high, and an upper unit consisting of sediments deposited after a major flooding event [Bergersen, 1993; Haggerty, Premoli-Silva et al., in press]. The quality
Figure 3.19. Isopach maps of carbonate platform thickness over Limalok (a) and Wodejebato (b) guyots. The maps were generated by subtracting a grid of pelagic-platform reflector depths from a grid of platform-basement reflector depths. Profile locations used to construct these maps are shown by thin dashed and solid ship tracks. Contour interval is 10 milliseconds. In this figure the contours truncate against the bathymetric contour marking the approximate edge of the summit plateau. In reality the isopach contours should converge to 0 at the edge of the platform. On both guyots, the shallow-water carbonate sediments thicken towards the edges of the summit plateau. As the top of the platforms show no preferential tilt, changes in platform thickness represent changes in basement topography. Consequently, these isopach maps show the location basement highs. The basement high on Limalok merges with the ridge extending towards Mili Atoll (to the north), whereas the basement high on Wodejebato appears slightly offset to the northeast across the summit plateau. The two peaks in basement on Wodejebato probably result from the wide-spacing of ship tracks.
Isopach Maps
Carbonate Platform Thickness

(a) Limalok Guyot

(b) Wodejebato Guyot
Figure 3.20. 6-channel seismic profile (a), interpretation (b), and drilling results (c) across the summit plateau of Wodejebato Guyot. Two primary sediment units compose the shallow-water platform carbonates (denoted by the reflectors LR1 and LR2). The upper unit appears relatively consistent in thickness across the entire plateau (possibly extending over the inner perimeter ridge) whereas reflectors within the lower unit truncate against the central basement high. A sub-basement "deep reflector" (similar to the one noted on Lo-En Guyot) shallows towards the center of the edifice. Site 873 sampled the lagoon sediment above the shallowest portion of the "deep reflector", and Sites 874 and 875 sampled the two perimeter ridges along the northeast edge of the summit plateau. Profile location shown in Figure 3.11.
Wodejebato Guyot

6-channel seismic profile

Sites drilled during Leg 144
of single-channel records over the carbonate platform of Limalok prohibit a detailed interpretation of its seismic stratigraphy. The Leg 144 profiles show a series of basement reflectors deepening and prograding to the south (Figure 3.21). Additional analyses of the logging data and the core samples from Site 871 will undoubtedly clarify the subsidence history of this platform and may allow mapping of the Eocene erosional surface proposed by Schlanger et al [1987].

The perimeter ridge encircling the summit plateau of Wodejebato Guyot appears to be equivalent to the raised reef rim observed on modern atolls. In the case of Wodejebato though, two ridges appear in the side-scan images along the north and northeast flank ridges, and along the ridge extending towards Pikinni Atoll. Drilling at Sites 875 and 876 showed that the outer ridge consists primarily of redeposited shell-debris. From the side-scan images, it is apparent that the outer ridge broadens and becomes more distinct across the shelves formed by the flank ridges (Figures 3.22 and 3.23). SeaBeam bathymetry collected over the northeast flank ridge [H. Staudigel, unpublished data] shows that the outer perimeter ridge loses relief and narrows towards the edge of the shelf (Figure 3.23). A small trough lying between the two ridges also becomes more prominent towards the edge of the shelf. The outer ridge reaches a maximum height of ~135 m while the inner ridge, extending ~45 m shallower than the outer ridge and 40 m shallower than the lagoon sediments, remains fairly consistent in width and height over the area surveyed during Leg 144. In the seismic profiles, the ridges consist of a relatively thin (0.03 sec) packet of closely-spaced reflectors overlying a thicker, more chaotic unit (Figure 3.23). The upper sediment packet may be related to the major flooding event noted in the lagoon sediments. On Limalok, a perimeter ridge may exist on the east and north sides of the summit plateau (Figure 3.16; north half of Leg 144 profile). Given the amount of block-faulting affecting this platform, it would not be surprising if such features have been eroded away from the south and west margins.
Figure 3.21. Single-channel seismic data collected across the south flank of Limalok during Leg 144. Basement reflectors on this guyot appear to prograde and deepen to the south, and some carbonate platform reflectors onlap the central basement high. Small-scale faulting appears to offset reflectors across the summit plateau. Profile location shown in Figure 3.4.
Interpretation

Site 871

Pelagic sediments

Carbonate platform

Clay layer (?)

Volcanic basement

Prograding basement unit
Figure 3.22. Side-scan image (a), and single-channel seismic profile across the north flank ridge of Wodejebato Guyot. The two perimeter ridges along the northern edge of the summit plateau appear as high-backscatter (dark) bands in the side-scan image. Note how the outer perimeter ridge widens across the shelf formed by the flank ridge. Topographic relief on these two ridges in this location is small. For the reflectors annotated in the seismic profile, PR marks the bottom of the pelagic unit, LR1 marks the bottom of the upper limestone unit, and VB marks the top of volcanic basement. Side-scan image location shown in Figure 3.11.
Figure 3.23. Side-scan image (a), 6-channel seismic profile (b), and SeaBeam bathymetry (c) across the shelf formed by the northeast flank ridge. The perimeter ridges in this area are very distinct in both the multibeam bathymetry and the seismic profiles. Similar to the case on the north flank ridge, the outer perimeter ridge widens across the shelf formed by the northeast flank ridge. Note the channel separating the inner and outer perimeter ridges, and the loss of relief across the outer ridge towards the edge of the shelf. Side-scan image location shown in Figure 3.11. SeaBeam data used with permission from H. Staudigel [unpublished data].
3.5.4. Summary

From a morphologic and stratigraphic perspective, the volcanic edifices and carbonate platforms of the guyots drilled during Leg 144 in the Marshall Islands appear quite similar to modern islands and atolls in the Hawaiian and French Polynesian chains. The number of flank ridges extending from each edifice generally conforms to the number of rift zones observed for Hawaiian volcanoes, especially if the northeast flank ridge on Wodejebato is actually a small seamount. Fault-blocks perched along the flanks of the guyots suggest that the edifices remain unstable even after moving away from the hot spot swell, although the triggering mechanism for this faulting remains unclear (sediment overburden and slope oversteepening are two possibilities). The gross morphology and stratigraphy of the drowned carbonate platforms (e.g., perimeter ridges bounding lagoon sediments, a transition zone of clay and altered volcaniclastic sediments separating the shallow-water platform carbonates from the underlying volcanic flows) appear similar to atolls in French Polynesia. In the same respect, some features observed across the summit plateaus do not adhere to traditional models of slow plate subsidence and edifice truncation (e.g., the cones across the summit plateau of Lo-En).

Given the complicated history of volcanism and uplift displayed by many French Polynesian edifices (e.g., Rurutu, Makatea, Mauke, Mitiaro) and assuming a similar tectonic environment existed in the Cretaceous, it is no wonder the guyots in the Marshall Islands display such a wide variation in morphology and stratigraphy. Differences in carbonate platform thickness and age across the guyots drilled during Leg 144 clearly demonstrate differing subsidence histories: a 230 m thick sequence of Paleocene to middle Eocene platform carbonates on Limalok, a 100 m thick sequence of Late Cretaceous carbonates on Wodejebato, and essentially no shallow-water carbonates on Lo-En. Further analyses on the core samples collected during Leg 144 may produce a definitive answer to why some carbonate platforms drowned and others survived, although applying a single
drowning mechanism to the Marshall Islands on a regional scale is tenuous because so many factors influencing platform drowning remain poorly constrained. Among these poorly constrained factors are the Cretaceous environmental conditions and the timing and sequence of hot spot activity in this area. A discussion of the environmental conditions affecting rates of subaerial erosion and net carbonate accumulation is beyond the scope of this paper, but the preliminary platform age and depth estimates supplied by Leg 144 make it possible to examine the effects of plate flexure on atoll and guyot pairs.

3.6. Constraints on Platform Morphology and Platform Drowning

The existence of drowned carbonate platforms connected to living atolls has long been considered a paradox in geology as the conditions inhibiting coral growth on one platform should have inhibited growth on the other as well [e.g., Schlager, 1981]. Carbonate platform survival depends primarily on the balance established between the net vertical accretion rate of the sediments or coral reefs and the relative rate of sea level rise. This balance determines the phase of growth exhibited by the reef system ("start-up", "catch-up", "keep-up", or "give-up") and the resulting sedimentary facies deposited through time [Neumann and Macintyre, 1985; Davies et al, 1985].

Maximum net accumulation rates for modern reef systems range from approximately 1 mm/yr to 10 mm/yr, with maximum productivity occurring within the upper 5 m to 15 m of water [e.g., Buddemeir and Smith, 1988]. Below 30 m to 40 m water depth, referred to as the "critical depth", a reduction of up to 40 % in carbonate accumulation rates can occur [Baker and Weber, 1975; Grigg and Epp, 1989]. Temperature extremes, nutrient excesses, sedimentation, inimical bank water exposures, and storm events can also affect coral growth rates [e.g., Schlager, 1981; Grigg and Epp, 1989].
Rates of relative sea level rise, on the other hand, include the effects of plate subsidence away from a spreading ridge or hot spot, plate flexure caused by volcanic or sediment loading, and eustatic changes in sea level. As a plate moves away from a ridge or hot spot, thermal cooling of the plate results in subsidence which adheres to a (time)\(^{1/2}\) relationship for plates less than ~70 m.y. old [Parsons and Sclater, 1977]. Plate flexure occurs during construction of a volcanic edifice. Both the moat and arch formed by a new load influence the history of uplift or subsidence on adjacent, existing edifices [e.g., McNutt and Menard, 1978; Watts and ten Brink, 1989; Jones, 1993]. The effects from these tectonic factors must be removed before attempting to establish any sort of eustatic sea level curve. Platform drowning occurs when the rate of relative sea level rise exceeds the rate of net carbonate accumulation and subsequent falls in sea level fail to bring the carbonate bank up to a depth where the accumulation rate can once again keep pace with changes in sea level. Grigg and Epp [1989], noting the vast difference between rates of net carbonate accumulation and the tectonic factors controlling relative rates of sea level rise, postulated that the size of the summit plateau plays an important role in determining platform erosion and subsequently the survival or demise of coral islands. Edifice size is inversely related to the truncation depth of a platform during low sea level stands because wave erosion acts more quickly on reducing smaller edifices to sea level than on larger ones. During subsequent rises in sea level, the truncated platforms may be submerged to such a depth that coral growth rates are insufficient to allow a successful recolonization. The following sections examine how plate subsidence in response to thermal cooling, plate flexure brought about by volcanic loading, and differences in summit area and volcanic platform depth apply to the atoll and guyot pairs studied during Leg 144.
3.6.1. Plate Subsidence

Observations of subsidence rates on portions of plates passing over hot spots in comparison to portions of equal age unaffected by hot spots suggest that the former subside at anomalously fast rates [Detrick and Crough, 1978]. Comparisons of these rates to standard plate subsidence curves led to the hypothesis that hot spot interaction with a plate thins the lithosphere and resets the thermal age of the plate to some value less than its absolute age, thereby allowing it to subside at a faster rate [Detrick and Crough, 1978; Crough, 1983].

A number of authors have shown that plate subsidence in response to thermal cooling is an unsatisfactory explanation for platform drowning because the rates are generally too slow [e.g., Wilson, 1975; Schlager, 1981; Grigg and Epp, 1989]. For example, even if the thermal age of a plate was reset to zero (i.e. equivalent to a ridge crest) the subsidence rate (using the plate subsidence formula of Parsons and Sclater, 1977) would not exceed 0.35 mm/yr, or at least an order of magnitude less than present-day coral growth estimates. Assuming a more realistic thermal reset age of 25 m.y. for a plate passing over a hot spot [Detrick and Crough, 1978], the initial subsidence of the plate (and newly formed volcano) approaches 0.035 mm/yr, an amount clearly insufficient to surpass present-day net carbonate accumulation rates (assuming these rates apply to the Cretaceous platforms in the Marshall Islands).

Plate subsidence also fails to explain the topography observed across the summit plateau of Lo-En Guyot. Presumably subaerial erosion associated with such subsidence accounts for the overall flatness of Lo-En's summit, but not for the survival of the cone- and lobe-shaped features cropping out from the pelagic sediments. In fact, the apparent absence of a carbonate buildup or reef-like structure along the perimeter of the summit plateau precludes any barrier preventing wave-erosion across the interior regions of the summit. Equally puzzling is the absence of a thick clay weathering horizon beneath the
pelagic sediments drilled at Site 872. Any mechanism invoked to explain the subsidence history for this guyot must account for these two observations.

3.6.2. Summit Area and Volcanic Platform Depth

In testing their hypothesis that edifice size played a critical role in the survival or demise of coral islands in the Hawaiian chain, Grigg and Epp [1989] showed that in most cases islands at or near sea level possessed larger summit areas than submerged platforms. They surmised that during the last major sea level fall smaller edifices were truncated more quickly and more completely than larger platforms. After the Holocene sea level rise the smaller platforms were too deep for coral growth to keep pace with tectonic subsidence, whereas incompletely truncated edifices provided a portion of the platform shallow enough for coral organisms to colonize and survive. This model is consistent with the observed carbonate bank development on the Great Barrier Reef in Australia where Holocene reefs grow primarily on top of antecedent Pleistocene topography [e.g., Hopley, 1984].

The volcanic platforms associated with the atoll and guyot pairs discussed in this paper generally adhere to the Grigg and Epp model: platforms beneath living atolls are larger than drowned platforms. However, the degree to which differences in summit area explain all drowned carbonate platforms in the Marshall Islands is debatable for a number of reasons:

1. As mentioned previously, linear trends in radiometric ages for the islands, atolls, and guyots composing the Marshall Islands are noticeably absent even within fairly well-defined groups of islands [e.g., Davis et al., 1989]. This scattered distribution of ages is quite unlike the relatively simple Hawaiian chain and strongly suggests that multiple episodes of volcanism occurred in the Marshall Islands. Such events are relatively common throughout the island chains in French Polynesia [e.g., Turner and Jarrard, 1982]. Although a detailed study of Cretaceous hot spot tracks through
the Marshall Islands is currently lacking, recurrent volcanism would result in a complicated history of uplift and subsidence through the effects of plate flexure from volcanic loading.

2. Very few atoll and guyot pairs within the Marshall Islands have been surveyed or sampled to such a degree that accurate measurements of summit plateau size (based on a given truncation horizon) or volcanic platform depth are possible. Even on the well-studied atolls discussed in this paper the depth to basement across the entire summit plateau is not well known. As shown by the isopach maps for Limalok and Wodejebato guyots in the Marshalls and the drilling on Mururo Atoll in French Polynesia, basement commonly shallows towards the center of the summit plateau. Consequently, determining the depth to basement for a particular platform or the difference in basement depth between two adjacent platforms (e.g., an atoll and guyot pair) from drill data along the perimeter of a plateau may not be appropriate. Such information appears critical for conducting a rigorous test of the Grigg and Epp model.

3. It remains unclear at what point a difference in summit area (and presumably edifice size) becomes a significant factor in the survival or demise of a coral island, especially in relation to paired atolls and guyots. In the case of Wodejebato and Pikinni, the difference in summit plateau size is only ~100 sq. km. as measured from the 1600 m contour. This difference does not take into account the amount of material removed by block-faulting along the flanks of Wodejebato. Similar to determining differences in basement depth from drill holes along the perimeter of a summit plateau, it may be equally inappropriate to attribute platform drowning to some arbitrary difference in platform size.
In Hawaii, where Grigg and Epp developed their model, eustatic changes in sea level were apparently responsible for the demise of coral islands in this chain. In such areas as French Polynesia and presumably the Marshall Islands, plate flexure from volcanic loading is another mechanism which could affect the history of uplift and subsidence on individual edifices. While differences in the size of edifices is certainly a viable candidate for causing platform drowning in the Marshall Islands (along with changes in environmental conditions), this region requires a closer analysis of plate flexure before drawing any final conclusions.

3.6.3. Plate Flexure

Plate flexure from volcanic loading is an appealing mechanism for explaining platform drowning because flexure can cause relatively rapid changes in sea level on existing sea floor features as the moat and arch form around a load. The rates of subsidence within the moat region can surpass maximum coral growth rates and prevent extensive subaerial erosion, whereas tensional stresses and uplift in the arch region may be responsible for rejuvenescent volcanic events on existing islands. Flexure modeling for the island of Hawaii suggests that the moat formed by this load depressed the island of Maui between 1000 m to 2000 m [Watts and Ten Brink, 1989]. As the average construction time of a Hawaiian-type volcano is ~ 0.5 m.y. [Moore and Clague, 1992] the average rate of subsidence on Maui was at least 1 mm/yr to 2 mm/yr during the time of the loading. The closer the existing feature lies to the load, the greater the deflection associated with the moat and the closer the subsidence rate approaches the reef system's maximum rate of net accumulation. Hence, plate flexure is a viable mechanism for causing platform drowning on some islands.

Observations from French Polynesia are once again a key to understanding and modeling plate flexure in the Marshall Islands. McNutt and Menard [1978] present a
flexure model which explains not only the observed uplift on ancient volcanic islands and atolls in the Cook, Tuamotu, Marquesas, and Pitcairn groups but also provides a good estimate of flexural rigidity for the plate in this region (1.7 to 2.5 x 10^{22} \text{ N m}). As a whole, their model fits the observed data (primarily uplifted islands) quite well, although some of the model parameters may require adjustment. For example, while Jarrard and Turner [1979] agree that plate flexure plays an important role in the uplift of French Polynesian islands, they argue that McNutt and Menard [1978] underestimate the amount of uplift on various atolls and oversimplify the episodes of volcanism affecting these island chains. Such observations serve warning towards attempting flexure modeling in the Marshall Islands where data are more scarce and the episodes of volcanism are even less well-understood. The objective of this section is therefore not in establishing independent estimates of such plate parameters as the elastic thickness or flexural rigidity, nor even such load parameters as the density of the material filling the moat or of the load itself. Many of the parameter values used for modeling flexure in French Polynesia will be assumed to hold true for modeling attempts in the Marshall Islands. The primary focus of this section is to examine the wavelength and, to a lesser degree, the amplitude of flexure caused by volcanic loading on adjacent edifices. It is important to emphasize that great care must be taken when interpreting and presenting the results of flexure modeling in an area where so many of the flexure parameters are not well-constrained. For example, estimates of the original size of the load and when the load was added often require a number of assumptions. Given the current data base and the scant information on edifice ages within this region, the most that can be accomplished with the flexure modeling presented in this paper is to show the areas over which relative deformation occurred. Consequently, the models shown in Figure 3.24 use bands of relative deformation rather than individual contours. The innermost gray-bands in this figure represent the 300 m deflection contour for a plate varying in thickness from 10 km (inner edge) to 20 km (outer edge). In a similar
Figure 3.24. Results of flexure modeling for the atoll and guyot pairs of Anewetak and Lo-En (a), Pikinni and Wodejebato (b), and Mili and Limalok (c). Each map of the model results shows two gray-bands. The inner band marks the 300 m deflection contour for a plate varying thickness from 10 km (inner edge) to 20 km (outer edge). The outer band marks the 0 m deflection contour for a similarly varying plate. For Anewetak and Lo-En (a), the load added to the plate was the volcanic edifice of Anewetak. Lo-En lies squarely in the region of 0 deflection from this load. For Pikinni and Wodejebato (b), the load added to the plate was the volcanic edifice beneath Pikinni Atoll. Wodejebato lies slightly within the region of 300 m deflection, yet neither the volcanic platform nor the carbonate platform appear tilted towards Wodejebato. The lack of tilting suggests both edifices were erupted at approximately the same time. For Mili and Limalok (c), the load added to the plate was the volcanic edifice south of Limalok. Limalok lies slightly inside the region of 0 deflection, which may explain the pattern of prograding basement reflectors along the south flank of this guyot.
fashion, the outermost bands represent the 0 m deflection contour over the same range of plate thicknesses.

Anewetak and Lo-En exhibit the best age-control of the three atoll and guyot pairs discussed in this paper. Drilling results from these edifices suggest that construction of Lo-En occurred around 90 Ma while construction of Anewetak was around 77 Ma [Haggerty, Premoli-Silva et al., in press; Pringle, 1992]. As shown by Figure 3.24a, Lo-En lies squarely in the region of 0 deflection produced from loading on Anewetak. If the 10 km plate thickness is more accurate, then Lo-En should have been sitting on the flexural arch of Anewetak. Conversely, if the 20 km plate thickness is more correct, then Lo-En should have subsided up to 100 m and the summit plateau should be tilted towards Anewetak. The basement depths on these two edifices are approximately equal; drilling along the perimeter of Anewetak reached basalt between 1283 m and 1410 m, and drilling at Site 872 on Lo-En encountered basement rocks at ~1225 m. Figure 3-10 shows that the platform beneath the pelagic sediments tilts slightly in the direction of Anewetak, increasing in depth from 1267 m on the south side to ~1305 m on the north side. If Menard's [1983] estimate for the rate of shelf widening in the Marquesas and Hawaiian Islands (1.1 to 1.7 km/m.y.) accurately measures the rate at which volcanic islands were truncated during the Cretaceous in the Marshall Islands, then Lo-En should have been eroded to sea level 6 m.y. to 10 m.y. after its formation, well before eruptions at Anewetak took place. As noted previously, such erosion would explain the overall flatness of Lo-En's summit plateau but not the topography cropping out from the pelagic sediments. Presumably uplift occurred across Lo-En in the Late Cretaceous as it passed over the hot spot swell associated with the volcanism on Anewetak. Given the correspondence between basement depths on these two edifices and the slight tilt of Lo-En towards Anewetak (suggesting the edifice was on the inside of the flexural arch), it is possible the summit plateau of Lo-En remained submerged during this Late Cretaceous episode of volcanism. No age data exist for the cones cones
cropping out across Lo-En's summit, but one possibility is that tensional stresses
associated with the plate flexure resulted in rejuvenescent volcanism on the truncated,
possibly submerged plateau.

Age-control on Wodejebato is quite good, but unfortunately the same can not be
said for Pikinni. Drilling during Leg 144 on Wodejebato produced a preliminary basement
age of Cenomanian and a drowning age for the carbonate platform of Late Cretaceous
[Haggerty, Premoli-Silva et al., in press]. Dredging along the flanks of this guyot
recovered basalt which gave a radiometric age of 86 m.y. [M. Pringle, unpublished data].
Drilling during Operation Crossroads on Pikinni sampled Eocene platform carbonates but
failed to reach volcanic basement. The model shown in Figure 3.24b assumes Pikinni was
constructed at about the same time as Anewetak, and hence flexure from this load affected
the existing edifice of Wodejebato. This model essentially tests whether plate flexure was
responsible for the drowning of the carbonate platform on top of Wodejebato. Figure
3.24b shows that Wodejebato lies well inside the region of 0 deflection, and slightly inside
the 300 m region of deflection. Seismic profiles across the summit plateau of Wodejebato
show no pronounced tilting to the southeast of either the basement reflector or the top of the
carbonate platform. In fact, the opposite appears true: Figure 3.20 shows basement
shallowing to the northeast. Flexure modeling also suggests that the basement depth on
Wodejebato should lie at least 300 m deeper than that of Pikinni (Figure 3.24b). As
mentioned previously, Raitt [1954] estimated that the average depth to basement on Pikinni
is ~1300 m even though the data support a depth range between 600 m and 2100 m.
Basement depth at Site 873 is 1509 m but this site lies southwest of the basement high.

Hence there does not appear to be an appreciable difference between the basement depth on
these two edifices. Given the poor age- and depth-control on the volcanic platform of
Pikinni, another possibility is that Wodejebato could be younger than Pikinni. Such a
scenario does not appear likely given the apparently small difference between basement
depths and the fact that the carbonate platform on Pikinni survived whatever event caused the drowning of Wodejebato. One important point to note is that reducing the lag time between two presumed episodes of edifice construction also minimizes the differences between platform stratigraphy (at least relating to plate flexure). That is in fact what the observed data suggest: both edifices were probably built at approximately the same time as volcanic basement on Wodejebato does not appear to be tilted towards Pikinni. The basement depth at Site 873 (1490 m) is shallower than the best estimate of basement depth on Pikinni (~1300 m; Raitt, 1954), but this depth discrepancy may not be real because Site 873 is not located on the shallowest portion of Wodejebato's summit plateau and the refraction data upon which the depth of basement on Pikinni is estimated varies by up to 1500 m.

Given the results of the flexure modeling on Wodejebato, it is unlikely construction of the edifice beneath Mili Atoll occurred much later than the edifice beneath Limalok Guyot as neither the top of the volcanic platform nor the carbonate platform of Limalok tilt towards the north-northwest. If anything, basement reflectors on Limalok appear to prograde to the south (Figure 3.21). One way to cause such a pattern of progradation is to load the plate in this direction, and the third model constructed in this paper examines such a possibility with the guyot located south-southeast of Limalok. The model load for this "Southern Guyot" assumes a regional plate depth and an original volcano height (~2300 m) similar to the model loads used for Anewetak and Pikinni. The total height of the load (6125 m) differs slightly from that of Anewetak and Pikinni because the assumed platform depth corresponds to that beneath Limaok (Figure 3.24c). Flexure calculations from this load show Limalok straddling the 0 deflection contour for the 10 km plate, suggesting that it may have been on the inner side of the flexural bulge during construction of the edifice to the south. The modeled load lies ~130 km south-southeast of Limalok, and if we assume the plate was moving to the north-northwest at a rate of 10 cm/yr then this load was
constructed ~1.3 m.y. after Limalok. This time difference is too small to account for the drowning of the carbonate platform on top of Limalok but it may explain the prograding sequence of basement reflectors on Limalok. Unfortunately, the validity of this model is highly debatable without additional age control on Mili Atoll and the guyot to the south of Limalok.

3.6.4. Summary

Flexure modeling in the Marshall Islands is subject to a number of assumptions concerning plate and load parameters, the most important being the time of plate loading. The models presented in this paper, while not answering all the questions concerning the evolution of Limalok, Lo-En, and Wodejebato guyots, provide additional insights on the history of volcanism, uplift, and subsidence for this region.

The apparent time lag between volcanism on Lo-En Guyot and Anewetak Atoll is sufficiently long to have allowed complete truncation of the original shield volcano forming Lo-En. Magnetic inclination data suggest this edifice formed around 30° S, a latitude near the present-day limit for active coral growth [Grigg, 1982]. Consequently, even though the plate motion was carrying the volcano into more favorable latitudes for reef growth, unfavorable water temperatures and subsidence of the plate, along with subaerial erosion, probably submerged the truncated summit prior to the establishment of a carbonate platform. Without a protective fringing reef along the edges of the summit plateau it seems unlikely that any sort of erosional remnant would survive this period of subaerial erosion, unless of course sea level rose rapidly near the time of complete truncation. A second episode of uplift across Lo-En should have occurred during the Late Cretaceous as the plate passed over the hot spot swell presumably associated with the volcanism on Anewetak. Whether this uplift was sufficient to cause subaerial exposure across Lo-En’s summit is another matter. The flexure modeling presented in this paper suggests that the load of
Anewetak resulted in little if any uplift on Lo-En, the most likely case being subsidence of up to 100 m. Such subsidence may have been sufficient to prevent the truncated summit of Lo-En from reaching a depth where coral organisms could successfully colonize the volcanic platform, thus explaining the absence of shallow-water carbonates across the plateau. In addition, stresses associated with the plate flexure may have created an easier path for the penetration of magmas on Lo-En, thus producing a second phase of volcanism and possibly explaining the cones cropping out across the summit plateau. Currents associated with a moderately shallow-water environment could presumably erode the clays and other sediments marking the original truncation horizon without affecting the volcanic cones.

Flexure modeling around Wodejebato and Limalok is less-constrained than around Anewetak but still provides some useful insights to the evolution of these two guyots. The absence of tilted volcanic or carbonate platforms suggests that the original shield volcanoes beneath living atolls and the drowned carbonate platforms were constructed at approximately the same time. Flexure from loading to the south of Limalok may be responsible for the pattern of prograding basement reflectors seen in the seismic profiles across this guyot.

The simple models presented in this paper only describe the flexure associated with the construction of a single volcano. A more accurate model would incorporate the flexure associated with other nearby volcanoes, but the lack of good age-control on these edifices prevents such an analysis. Without additional age-control on surrounding edifices plate flexure can not be completely ruled out as a mechanism responsible for platform drowning. For example, although it appears Pikinni and Wodejebato were constructed at approximately the same time, loading on adjacent edifices (e.g. the Ronlap atoll complex) may have resulted in an additional episode of uplift or subsidence across these two edifices.
3.7. Conclusions

Limalok, Lo-En, and Wodejebato guyots are all comparable in some respects to modern islands and atolls in French Polynesia, and this area of Pacific holds perhaps the most promise for understanding the tectonic environment responsible for the formation of edifices within the Marshall Islands. A good understanding of the volcanism and tectonism affecting the Marshall Islands is crucial for any attempt to establish a eustatic sea level curve from the sediments cored during Leg 144. It is clear from the data presented in this paper that plate subsidence and edifice depth fail to adequately explain all the features noted across the summit plateaus and along the flanks of Limalok, Lo-En, and Wodejebato. Plate flexure associated with volcano construction has almost certainly influenced the history of uplift and subsidence experienced by edifices within the Marshall Islands, but with the paucity of age data from the various island groups only general models can be constructed showing relative regions of arching or down-warping. The flexure models presented here possibly explain the negligible difference between the basement depths of Anewetak and Lo-En, the high-backscatter cone- and lobe-shaped features cropping out along the flat summit plateau of Lo-En, and the sequence of prograding basement reflectors on Limalok. The models also suggest that construction of the volcanic platforms beneath Wodejebato and Pikinni, and possibly Limalok and Mili, occurred without a substantial time gap. What the models fail to establish is the cause of carbonate platform drowning (or survival).

The next step towards understanding the evolution of the Marshall Islands as a whole is the identification of hot spot trends within the various island chains. Additional analyses of the sediments recovered during Leg 144 will undoubtedly shed light on the drowning mechanisms for carbonate platforms in this region, be they environmental (cooling of water temperature, over-nutritification of surface waters, storm disturbances), biological (changes in faunal growth rates), or geological (uplift or subsidence brought about by plate flexure, secondary episodes of volcanism, erosional differences) in nature.
Until the sequence and timing of volcanism in the Marshall Islands is better understood, and hence the tectonic signal of sea level changes can be removed, using drowned carbonate platforms in the Marshall Islands as "dipsticks" for recording the history of Cretaceous sea level fluctuations is not possible.
CHAPTER 4
CRETACEOUS HOT SPOT TRACKS THROUGH THE MARSHALL ISLANDS

4.1. Introduction

The complicated history of Cretaceous volcanism across the Pacific plate is well-recorded by the numerous islands, atolls, guyots, seamounts, and mid-ocean plateaus occupying the western Pacific, including those composing the Marshall Islands. Early ideas explaining the preponderance of these volcanic features centered around such tectonic events as regional uplift and large-scale fracturing of the plate [Menard, 1964]. Age and depth information from several legs of the Deep Sea Drilling Project (DSDP) and from the dredging of seamounts throughout the Pacific basin confirmed that regional uplift and extensive volcanism occurred during the Cretaceous across an area extending from the Mid-Pacific Mountains and the Line Islands westward to the Marshall Islands and Nauru Basin [e.g., Winterer, Ewing et al., 1973; Larson, Moberly et al., 1974; Schlanger, Jackson et al., 1974; Larson, Schlanger et al., 1981; Moberly, Schlanger et al., 1986]. However, the duration of the volcanism as shown in the DSDP cores was still somewhat debatable. For example, after drilling at Site 462 Schlanger and Premoli Silva [1981] postulated that uplift and mid-plate volcanism began in Barremian to early Aptian time and continued through the Maestrichtian, whereas Moberly and Jenkyns [1981] argued for two distinct episodes of volcanism separated by approximately 40 m.y.

Additional dredging and drilling efforts over the past five years have resulted in a number of radiometric and fossil ages for the edifices composing the Marshall Islands, and suggest a complicated history of volcanism and uplift across this area (Figure 4.1, Table 4.1). Two island chains have been formally recognized within the Marshall Islands. The easternmost line of volcanoes, the Ratak chain, extends from around Limalok Guyot in the south to Woden-Kopakut Guyot in the north. The more centrally-located Ralik chain
Figure 4.1. Base map of the Marshall Islands showing selected edifice names and ages, and the two formally recognized volcanic chains in this area. Table 4.6 contains a complete list of edifice names and locations. Radiometric ages shown in this figure come from Davis et al. [1987] and Pringle [1992]. Bathymetry map revised from Hein et al. [1990].
Table 4.1. Radiometric ages for edifices in the Marshall Islands

<table>
<thead>
<tr>
<th>Edifice</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Age (m.y.)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lokkworkwor Guyot</td>
<td>8.85</td>
<td>169.75</td>
<td>86.7 - 87.9*</td>
</tr>
<tr>
<td>Woden-Kopakut Guyot</td>
<td>14.00</td>
<td>167.45</td>
<td>80.6 - 83.8*</td>
</tr>
<tr>
<td>Look Guyot</td>
<td>12.20</td>
<td>166.15</td>
<td>137.7 - 138.7 b</td>
</tr>
<tr>
<td>Wodejebato Guyot</td>
<td>11.92</td>
<td>164.85</td>
<td>85.0 - 87.0 c</td>
</tr>
<tr>
<td>Lobbadede Guyot</td>
<td>13.85</td>
<td>163.90</td>
<td>80.0 - 84.8 b</td>
</tr>
<tr>
<td>North Wod-En Guyot</td>
<td>16.10</td>
<td>163.05</td>
<td>84.9 - 85.7 b</td>
</tr>
<tr>
<td>Mij-Lep Guyot</td>
<td>8.78</td>
<td>163.15</td>
<td>109.0 - 111.0 c</td>
</tr>
<tr>
<td>Lalibjet Guyot</td>
<td>10.10</td>
<td>160.43</td>
<td>81.4 - 82.8 b</td>
</tr>
<tr>
<td>Anewetak Atoll</td>
<td>11.55</td>
<td>162.16</td>
<td>75.3 - 76.5 b</td>
</tr>
</tbody>
</table>
extends from Jalwoj Atoll in the south to around Lewa Guyot in the north. These two chains appear to intersect around 14° N, making the delineation between edifice groups virtually impossible. Farther to the west, a cluster of volcanoes centered about Ujlan and Anewetak atolls forms a third geographic province.

Edifices within these three provinces fail to show a simple, linear age progression in the direction of plate motion, unlike many other hot spot chains in the Pacific (e.g., the Hawaiian Islands, the Emperor Seamounts, the Society Islands). Davis et al. [1989] and Pringle [1992] have shown that younger edifices lie "downdrift" from older edifices in both the Ratak and Ralik chains (Figure 4.1, Table 4.1). "Downdrift" in this case means to lie farther away from the presumed hot spot source in the direction of plate motion. For example, Woden-Kopakut and Lobbadede guyots lie 5.5° and 2° northwest of Lokkworkwor and Wodejebato guyots, respectively. Yet both of these edifices are younger than their more southerly counterparts. The Ujlan-Anewetak cluster of volcanoes also exhibits a transposition of edifice ages where the preliminary results from drilling during Ocean Drilling Program (ODP) Leg 144 suggest that Lo-En Guyot may be over 14 m.y. older than the edifice beneath the more northwesterly Anewetak Atoll [Haggerty and Premoli Silva, in press]. Shallow-water limestones recovered from various summit plateaus in the Marshall Islands also suggest a complicated history of plate uplift and subsidence. Lincoln [1990] postulated that two distinct episodes of uplift and shallow-water exposure occurred across many of the platforms, resulting in two distinct episodes of coral growth: the first during the mid-Cretaceous and another during the Late Cretaceous (Table 4.2). Nearly all the volcanic rocks dated by radiometric means in the Marshall Islands are of Late Cretaceous age, including those samples from guyots identified by Lincoln [1990] as having mid-Cretaceous shallow-water carbonate sediments [Pringle and Staudigel, 1992]. Drilling on Limalok, Lo-En, and Wodejebato guyots during Leg 144 also failed to recover any mid-Cretaceous shallow-water carbonate sediments, although Lo-
<table>
<thead>
<tr>
<th>Edifice</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Fossil Evidence</th>
</tr>
</thead>
<tbody>
<tr>
<td>Wodejebato Guyot</td>
<td>11.92</td>
<td>164.85</td>
<td>Albian reef associated with Campanian- to Maastrichtian-age benthic foraminifer matrix</td>
</tr>
<tr>
<td>Ruwituntun Guyot</td>
<td>11.90</td>
<td>166.95</td>
<td>Albian reef fragments in Late Cretaceous foraminifer matrix</td>
</tr>
<tr>
<td>Lo-En Guyot</td>
<td>10.15</td>
<td>162.80</td>
<td>Albian- to Cenomanian-age planktonic foraminifer of near-reef affinity</td>
</tr>
<tr>
<td>Lalibjet Guyot</td>
<td>10.10</td>
<td>160.43</td>
<td>Basalt breccia within a Late Cretaceous foraminifer matrix</td>
</tr>
<tr>
<td>Lobbadede Guyot</td>
<td>13.85</td>
<td>163.90</td>
<td>Phosphatized limestones containing Cenomanian- to Campanian-age planktonic foraminifers</td>
</tr>
<tr>
<td>Lewa Guyot</td>
<td>14.00</td>
<td>163.20</td>
<td>Ooidic limestone conglomerate in a matrix of Albian- to Cenomanian-age foraminifers</td>
</tr>
<tr>
<td>Lomtal Guyot</td>
<td>14.78</td>
<td>163.50</td>
<td>Basalt conglomerate in a phosphatized Santonian-age (?) foraminifer (poorly preserved)</td>
</tr>
</tbody>
</table>
En Guyot appears to be >90 m.y. old [Haggerty and Premoli-Silva, in press]. Accordingly, a number of questions regarding the age, sequence, and duration of volcanism in the Marshall Islands still exist.

In an attempt to identify the hot spots responsible for the two apparent episodes of uplift in the Marshall Islands, Lincoln [1990] used a method similar to Duncan and Clague [1985] for rotating dated edifices back through time to their region of construction. In the case of the Marshall Islands, this region is French Polynesia. Smith et al. [1989] also utilized this technique to "backtrack" Himu and Hemler guyots in the Magellan Seamounts and Woden-Kopakut and Lokkworkwor in the Marshall Islands. Both Lincoln and Smith et al. agreed that the proposed hot spots beneath Rurutu and Rarotonga influenced the development of the Marshall Islands, although neither group went so far as to identify specific hot spot trends through either the Ratak or Ralik chains. For example, even though Smith et al. [1989] successfully tracked Woden-Kopakut Guyot back to an area near present-day Rurutu hot spot they were unable to identify a hot spot directly responsible for Lokkworkwor. In a similar manner, Lincoln [1990] suggested that the fossil-dated mid-Cretaceous edifices formed near the Macdonald hot spot, and that the Late Cretaceous volcanism and uplift was probably attributable to a combination of the Rurutu, Rarotonga, and Tahiti hot spots; he did not extend his analysis to show linear trends within the Marshall Islands.

All of the studies to date have focused on tracking individual edifices back to presently active hot spots in French Polynesia. Given the complicated history of volcanism in French Polynesia, especially within the Cook-Austral chain [McNutt and Menard, 1978; Turner and Jarrard, 1982] one can see how the transposition of edifice ages may occur and how the record of an older carbonate platform may be partially or totally erased through uplift, subaerial exposure, and erosion. Further complications to understanding the history of volcanism across ancient edifices arise from:
1. the uncertainties associated with such dating techniques as fossils vs. radioisotopes, and K/Ar dating vs. \(^{40}\text{Ar}/^{39}\text{Ar}\) dating [among others, Pringle, 1992].

2. the duration of volcanism beyond the shield-building stage [e.g., Clague and Dalrymple, 1988],

3. the length of time separating edifice construction, summit truncation, and coral colonization [e.g., Bardintzeff et al., 1985], and

4. the calculation of stage poles and rotation angles back through the Cretaceous [e.g., Morgan, 1972; Epp, 1978; Duncan and Clague, 1985; Engebretson et al., 1985; Yan and Kroenke, 1993].

This section attempts to identify Cretaceous hot spot tracks through the Marshall Islands by treating the assorted guyots, atolls, and islands as members of hot spot chains rather than individual edifices. After the hypothetical trends have been tested against the available age information, the focus of the paper shifts to understanding the implications of misfits between the modeled tracks and the observed data, in particular the longevity and the fixity of hot spots during the construction of the Marshall Islands.

4.2. Methodology

A number of assumptions are inherent in any model attempting to identify hot spot trends through the Marshall Islands. These assumptions include:

1. The edifices composing the Marshall Islands result from mid-plate volcanism over hot spots. This assumption excludes such catastrophic tectonic events as large-scale fracturing of the plate [Menard, 1964], or the simultaneous construction of volcanoes along "hot lines" [Bonatti and Harrison, 1976];

2. Hot spots in general are relatively long-lived, deep-seated mantle plumes capable of creating discrete chains of volcanoes. The geometry of these volcanic island chains records changes in plate motions through time, and the age progression along a
chain indicates the rate of plate motion relative to the fixed hot spot source. As such, hot spot chains continue to be a powerful tool in plate tectonic theory.

3. The four hot spots postulated to be influencing the Cook-Austral and Society chains also were active in the Cretaceous, and their geographic location has remained essentially the same (i.e., a fixed hot spot source).

The third assumption requires some additional discussion as the data supporting the existence of each French Polynesian hot spot varies substantially.

Two well-defined, northwest-trending chains of volcanoes, the Cook-Austral Islands and the Society Islands, mark the western extent of French Polynesia (Figure 4.2). Macdonald Seamount lies at the southeastern end of the Austral chain and marks the site of active volcanism in this island group (Table 4.3). Edifices within the Austral chain linearly increase in age up to Rurutu Island, where young volcanic rocks are superimposed on an older shield volcano [Turner and Jarrard, 1982]. The volcanic rocks of this island record two episodes of volcanism separated by ~10 m.y., whereas the island itself consists of a single shield volcano surrounded by 100 m high cliffs of shallow-water carbonates. A number of uplifted atolls lie to the north of Rurutu, forming for the most part the southern Cook Islands (Figure 4.2). Volcanic rocks dated from these uplifted edifices show a linear age progression related to the young volcanism recorded at Rurutu (Table 4.3). The uplifted atolls on the other hand suggest that the underlying volcanic platform may be more closely related to volcanism at Macdonald as sufficient time must have passed since the time of edifice construction for the summit plateau to subside and be colonized by shallow-water reefal organisms. Rarotonga Island lies to the southwest of the Cook Islands, and is yet another site of recent (1.1 m.y. to 1.7 m.y.; Table 4.3) volcanism [Jarrard and Turner, 1979; Matsuda et al., 1984]. The island appears to be geomorphically young with very little reef along its shores, but there does not appear to be any evidence in the sea floor topography for the passage of a hot spot through this area (neither to the northwest nor the
Figure 4.2. Bathymetry of the Cook-Austral Islands and the Society Islands. This map shows the location of hot spots discussed in this paper and the alignment of edifices associated with each hot spot.
### Table 4.3. Descriptions of Selected French Polynesia Hot Spot Chains

<table>
<thead>
<tr>
<th>Hot Spot Chain</th>
<th>Radiometric Ages (m.y.)</th>
<th>Distance (km)</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>Austral Islands</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mcdonald</td>
<td>0</td>
<td>0</td>
<td></td>
</tr>
<tr>
<td>Rapa</td>
<td>5.1 +/- 0.3 e</td>
<td>405.9</td>
<td>Horseshoe-shaped eroded caldera surrounded by very little reef.</td>
</tr>
<tr>
<td>Raivavae</td>
<td>5.5-7.3 d</td>
<td>888.0</td>
<td>Elongate island 460 m high surrounded by barrier reef.</td>
</tr>
<tr>
<td>Tubuai</td>
<td>8.6-10.4 d</td>
<td>1062.9</td>
<td>Elliptical island surrounded by barrier reef.</td>
</tr>
<tr>
<td>South Cook Islands</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Rurutu</td>
<td>1.06-1.12 d, 12.08-12.56 e</td>
<td>0</td>
<td>Uplifted remnant of shield volcano surrounded by 100 m high coral reef.</td>
</tr>
<tr>
<td>Rimatara</td>
<td>&gt; 4.78 +/- 0.42 c, 20.6-29.9 c</td>
<td>180.3</td>
<td>Uplifted atoll with 83 m high volcanic summit and 11 m high makatea.</td>
</tr>
<tr>
<td>Maria</td>
<td></td>
<td>374.7</td>
<td></td>
</tr>
<tr>
<td>Mauke</td>
<td>4.64 +/- 0.14 c, 6.3 +/- 0.2 c</td>
<td>671.7</td>
<td>Uplifted atoll with flat summit (29m) surrounded by 24 m high makatea.</td>
</tr>
<tr>
<td>Mitiaro</td>
<td>&gt; 12.3 +/- 0.42 c</td>
<td>716.1</td>
<td>Uplifted atoll with 12 m high volcanic hills and a 15 m high makatea rim.</td>
</tr>
<tr>
<td>Atiu</td>
<td>8.0-8.5 c, -10 c</td>
<td>743.7</td>
<td>Uplifted atoll showing 2 episodes of volcanism.</td>
</tr>
<tr>
<td>Aitutaki</td>
<td>7.39 to &gt; 8.73 c, 0.7-1.9 c</td>
<td>951.0</td>
<td>Very nearly an atoll with relatively large carbonate platform.</td>
</tr>
<tr>
<td>Mangaia</td>
<td>16.2-19.8 b,c</td>
<td>-191.4</td>
<td>Uplifted atoll with 169 m high volcanic summit and 90 m high limestone rim.</td>
</tr>
<tr>
<td>Raratonga</td>
<td>1.6-2.3 c, 1.4 +/- 0.3 a</td>
<td>0</td>
<td>Geomorphically young island showing very little reef development.</td>
</tr>
<tr>
<td>Society Isls.</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mehetia</td>
<td>0</td>
<td>0</td>
<td></td>
</tr>
<tr>
<td>Tahiti-Nui</td>
<td>0.4-1.2 d, f</td>
<td>39.0</td>
<td>Maximum elevation of 2241 m.</td>
</tr>
<tr>
<td>Tahiti-Iti</td>
<td>&lt; 0.5 d</td>
<td>69.3</td>
<td>Total island length for Tahiti is around 65 km.</td>
</tr>
<tr>
<td>Moorea</td>
<td>1.49-1.64 d</td>
<td>108.3</td>
<td>Approximately 1/3 of the cone appears to have eroded away.</td>
</tr>
<tr>
<td>Huahine</td>
<td>2.01-2.58 d</td>
<td>244.2</td>
<td>Consists of twin islets formed by the faulting of a single cone.</td>
</tr>
<tr>
<td>Raiatea</td>
<td>&gt; 2.45 d</td>
<td>283.2</td>
<td>Triangular in shape, rising to a maximum elevation of 1033 m.</td>
</tr>
<tr>
<td>Tahaa</td>
<td>2.56-3.16 d</td>
<td>294.3</td>
<td>Initiation of volcanism coincides with volcanism on Bora Bora.</td>
</tr>
<tr>
<td>Bora Bora</td>
<td>3.12-3.51 d</td>
<td>321.9</td>
<td>Type example of a barrier reef as described by Darwin.</td>
</tr>
<tr>
<td>Maupiti</td>
<td>3.94-4.34 d</td>
<td>391.2</td>
<td>Caldera may have lain to the south of the island.</td>
</tr>
</tbody>
</table>

References: a) Matsuda et al., 1984; b) Dalrymple et al., 1975; c) Turner and Jarrard, 1982; d) Duncan and Mcdougall, 1976; e) Krummenacher and Noetzlin, 1966. All ages based on K/Ar dating techniques with the exception of the one Ar$^{40}$/Ar$^{39}$ Matsuda et al. [1984] age.
southeast of this edifice). The fourth hot spot proposed for this region is located southeast of Tahiti in the Society chain (Figure 4.2). Mehetia hot spot consists of at least five centers of volcanic and hydrothermal activity encompassing an area of approximately 8000 km² [Stoffers et al., 1989; Cheminee et al., 1989]. The Society chain exhibits a simple linear age progression in the direction of plate motion (Table 4.3), extending from Mehetia Seamount in the southeast to at least Motu One Island in the northwest [Duncan and McDougall, 1976]. The Macdonald and Mehetia hot spots appear to have been active for at least the past 20 m.y. and probably longer [Duncan and McDougall, 1976; Turner and Jarrard, 1982]. Accordingly, they are given the highest priority when attempting to match hot spot tracks in the Marshall Islands to the observed topography and age data. The proposed Rurutu hot spot also seems to have played a relatively important role in the development of the Cook Islands, but the scarcity of data on and around this edifice make it less reliable than Macdonald and Mehetia. Rarotonga Island, albeit young, shows very little evidence for being directly related to hot spot activity, and hence receives the least consideration throughout the modeling. Other hot spots have been proposed for this area (e.g., Raevavae and Aitutaki; Fleitout and Moriceau, 1991) but these four appear to have the best support with respect to the recency of their volcanism.

If one accepts the assumptions as listed above, then a simple model for identifying hot spot tracks through the Marshall Islands can be constructed. The first step towards constructing this model is to plot the topography of the Marshall Islands in oblique mercator projections. In these projections, the poles of observation are the stage poles defining Pacific plate motion relative to hot spots. Such projections force features created during the time interval of the stage pole to lie on small circles about the pole. All volcanoes constructed over a stationary hot spot source during a particular stage pole interval will form horizontal lines in the projection. A number of stage poles describing Pacific plate motion relative to hot spots exist in the literature. This paper initially examines
three groups (Table 4.4). The poles listed in the unpublished manuscript of R. Gordon and L. Henderson were used by Lincoln [1990] to track guyots in the Marshall Islands, and are used in this paper to remain consistent with Lincoln's models. The Engebretson et al. [1985] poles were derived from global plate circuits as part of a project to describe the displacement history between western North America, eastern Eurasia, and adjacent oceanic plates (Pacific, Farallon, Izanagi, Kula, and Phoenix). A third group of stage poles results from an analysis of linear volcanic chains on the Pacific plate [Duncan and Clague, 1985].

All three groups of poles are in relative agreement back to about 78 Ma (Figure 4.3). The Hawaiian chain provides good control on Pacific plate motion from 0 Ma to 43 Ma [e.g., Morgan, 1972; Jarrard and Clague, 1977; Minster and Jordan, 1978; Turner et al., 1980]. In a similar fashion, the Emperor chain provides good contraints on Pacific plate motion relative to hot spots for the time period extending from 43 Ma to ~78 Ma [Clague and Jarrard, 1973; Epp, 1978; Henderson and Gordon, 1981]. Differences in the stage duration and the angle of rotation for this time interval (Table 4.4) result from choosing different edifices marking the change of plate motion in the northern Emperor chain. Gordon and Henderson [unpublished data] used Detroit Tablemount, whereas the other two groups used Tenji Seamount. The pre-78 Ma stage pole and angle of rotation used by Engebretson et al. [1985] differs substantially from those used by the other two groups (Figure 4.3, Table 4.4). Duncan and Clague [1985] and Gordon and Henderson [unpublished manuscript] used the stage pole of Epp [1978], found by fitting small circles to three northwest-striking seamount chains north of the Hawaiian chain. The stage pole listed by Engebretson et al. [1985, p. 7] is quite different from the Epp pole, as is the angle of rotation (almost 9° more than that prescribed by Epp [1978]). Consequently, the present analysis ignores this set of stage poles. The other two groups have only slight differences in the angle of rotation and duration of the stage period. Gordon and Henderson's
Figure 4.3. Map showing the variations in stage poles and angles of rotation used by Engebretson et al. [1985], Duncan and Clague [1985], and Gordon and Henderson [unpublished manuscript]. The yellow track represents the Engebretson et al. stage poles, the blue track represents the Duncan and Clague stage poles, and the green track corresponds to the Gordon and Henderson poles. All the poles are in agreement to approximately 78 Ma, after which the Engebretson et al. plate rotations diverge from other two sets of poles.
Variations in Stage Pole Tracks
<table>
<thead>
<tr>
<th>Stage Poles for Pacific Plate Motion Relative to Hot Spots</th>
<th>Gordon and Henderson Model</th>
<th>Engebretson et al. Model</th>
<th>Duncan and Clague Model</th>
</tr>
</thead>
<tbody>
<tr>
<td>Latitude</td>
<td>Longitude</td>
<td>Angle</td>
<td>Reference</td>
</tr>
<tr>
<td>0 - 5 Ma</td>
<td>61.7</td>
<td>277.5</td>
<td>4.8</td>
</tr>
<tr>
<td>0 - 43 Ma</td>
<td>67.0</td>
<td>287.0</td>
<td>34.0</td>
</tr>
<tr>
<td>43 - 78 Ma</td>
<td>17.0</td>
<td>253.0</td>
<td>19.0</td>
</tr>
<tr>
<td>78 - 97 Ma</td>
<td>36.0</td>
<td>284.0</td>
<td>15.5</td>
</tr>
<tr>
<td>97 - 130 Ma</td>
<td>70.0</td>
<td>289.0</td>
<td>13.0</td>
</tr>
<tr>
<td>0 - 5 Ma</td>
<td>57.0</td>
<td>285.0</td>
<td>4.7</td>
</tr>
<tr>
<td>5 - 43 Ma</td>
<td>69.0</td>
<td>289.0</td>
<td>18.2</td>
</tr>
<tr>
<td>43 - 74 Ma</td>
<td>22.0</td>
<td>269.0</td>
<td>20.2</td>
</tr>
<tr>
<td>74 - 100 Ma</td>
<td>48.0</td>
<td>299.0</td>
<td>24.0</td>
</tr>
<tr>
<td>100 - 115 Ma</td>
<td>68.0</td>
<td>162.0</td>
<td>8.0</td>
</tr>
<tr>
<td>115 - 135 Ma</td>
<td>75.0</td>
<td>273.0</td>
<td>10.0</td>
</tr>
<tr>
<td>0 - 42 Ma</td>
<td>68.0</td>
<td>285.0</td>
<td>34.0</td>
</tr>
<tr>
<td>42 - 65 Ma</td>
<td>17.0</td>
<td>253.0</td>
<td>14.0</td>
</tr>
<tr>
<td>65 - 74 Ma</td>
<td>22.0</td>
<td>265.0</td>
<td>7.5</td>
</tr>
<tr>
<td>74 - 100 Ma</td>
<td>36.0</td>
<td>284.0</td>
<td>15.0</td>
</tr>
<tr>
<td>100 - 150 Ma</td>
<td>85.0</td>
<td>165.0</td>
<td>24.0</td>
</tr>
</tbody>
</table>
unpublished rotation angle compromises between placing the Line Islands-Line Cross trend intersection over the Easter hot spot and the Hess Rise over the Marquesas hot spot, whereas Duncan and Clague [1985] propose slightly less rotation than Epp [1978]. The constraints on stage pole estimates for pre-100 Ma plate motions are very poor. The Mid-Pacific Mountains and Hess Rise appear to be the most reliable recorders of plate motion during this time interval, but the paucity of ages for edifices in these two volcanic provinces makes estimated stage poles and rotation angles dubious. Duncan and Clague [1985] used a single stage pole extending from 100 Ma to 150 Ma, whereas Gordon and Henderson defined two sub-stage poles at 97 Ma to 130 Ma and 130 Ma to 150 Ma (hence dividing the Mid-Pacific Mountains and Hess Rise into more realistic two trend volcanic chains). With the two sets of stage pole estimates of Duncan and Clague [1985] and Gordon and Henderson [unpublished manuscript], hypothetical hot spot tracks for Macdonald, Mehetia, Rurutu, and Rarotonga can be traced across the Pacific plate for the period extending from 42 Ma to > 100 Ma.

4.3. Model Results

The Duncan and Clague poles are used to show general trends within the Marshall Islands. A number of observations are possible from these projections assuming completely stationary hot spots (Figure 4.4):

1. Rurutu hot spot appears to be responsible for the construction of much of the Ratak chain back through the 74 Ma to 100 Ma stage pole period;
2. The 100 Ma to 150 Ma portion of the Macdonald hot spot track crosses the northern Marshall Islands;
3. The Ujlan-Anewetak group of edifices may be linked by the Rarotonga hot spot; and
Figure 4.4. Oblique mercator projections using the stage poles of Duncan and Clague [1985]. The projections show the general correspondence between stationary hot spots in French Polynesia and the edifices composing the Marshall Islands.
4. Mehetia hot spot was apparently inactive or failed to penetrate the lithosphere at this time as no topography marks its passage.

5. A considerable number of edifices lie off the tracks from these four hot spots. The oblique projections shown in Figures 4.4a and 4.4b suggest the southern half of the Ratak chain, up to and including Maleolap Atoll, was constructed during the latest Cretaceous (pre-74 Ma stage poles). The preliminary fossil age assigned to the base of the carbonate platform capping Limalok Guyot (Paleocene; Haggerty and Premoli-Silva, in press) supports this observation. In the northern half of the Ratak chain though, Davis et al. [1989] note that the transposition of edifice ages between Lokokworkwor and Woden-Kopakut suggests a more complicated history of volcanism than usually associated with a single hot spot. While the 82 m.y. age appears appropriate for the hot spot track from Rurutu, the 87 m.y. age at Lokokworkwor is inconsistent with this track. Another inconsistency associated with the modeled tracks exists farther to the west where none of the tracks overlie the Ralik chain. The general parallelism of the chain of volcanoes from Jalwoj Atoll to North Wod-En with the 74 Ma to 100 Ma stage pole projection (Figure 4.5) suggests they formed during this time, as do the radiometric ages. As mentioned previously, fossils in the limestones recovered during dredging suggest that a number of the platforms in the northern Ralik chain also existed during the mid-Cretaceous [Lincoln, 1990]. Preliminary inversions of seamount magnetic data collected in this area also suggest that at least some of the edifices are mid-Cretaceous in age [P. Bryan and T. Shoberg, unpublished data], although the paleopole estimates (and therefore magnetic ages) are still questionable at this time because of the modeling program used (a least squares technique as opposed to the more generally accepted seminorm minimization technique; Parker, 1988, 1991). The mid-Cretaceous fossil and magnetic inversion ages for Lewa, Lobbade, and Aean-Kan appear consistent with the present position of the Macdonald hot spot track in the Duncan and Clague projections but the fit worsens with the presumably more realistic
Figure 4.5. Oblique mercator projections for the Late Cretaceous using the 78 Ma to 97 Ma stage pole of Gordon and Henderson [unpublished data]. (a). Hot spot location has remained stationary since the Cretaceous. (b). Hot spots have wandered relative to the Hawaiian hot spot at consistent rates to the northeast since the Cretaceous.
(a) Stationary Hot Spots

(b) Hot Spots Wandering at Variable Rates
Gordon and Henderson poles. Neither group of stage poles produces a track explaining the mid-Cretaceous fossil ages on Wodejebato and Ruwituntun guyots, nor do they account for the Late Cretaceous volcanic ages on Lobbadede or Wodejebato.

Inconsistencies between the model hot spot tracks and the observed topography and age data suggest that some of the initial assumptions require revision if we continue to assume that the Marshall Islands formed from French Polynesian hot spots. The two assumptions most amendable to revision are the duration of hot spot activity (or perhaps the number of hot spots) and the stationarity of hot spots relative to one another. Fluctuations in the strength of the mantle plume and changes in the thickness of the lithosphere influence the duration over which a hot spot leaves a topographic expression on the sea floor, and provides perhaps the easiest means of explaining anomalous linear trends of sea floor topography (or lack thereof). Under such conditions, every chain of volcanoes not accounted for by a presently active hot spot (e.g., the Ralik chain) could result from the penetration of a "new" hot spot through the lithosphere. In the same respect, every modeled hot spot track not showing a physical trace on the sea floor could simply result from the absence of a plume during that particular time or an impenetrable portion of the lithosphere. Under such conditions, the number of hot spots active beneath French Polynesia may be greater than currently accepted [Fleitout and Moriceau, 1991]. Needless to say though, arbitrarily varying the strength of mantle plumes or the lithosphere to match sea floor topography is a less than desirable mechanism for explaining misfits between the modeled tracks and the observed data because it diminishes their role as long-term recorders of plate motion history, and it appears to contradict observations from such well-established chains as the Hawaiian Islands and the Emperor Seamounts. Hot spots do tend to wax and wane through time [e.g., Detrick et al., 1986; Lonsdale, 1988; Sleep, 1990] so we can not exclude this possibility entirely. However, it is worthwhile to examine an
alternative means for matching poorly-fit linear trends of volcanic islands to existing hot spots: by allowing hot spots to wander relative to one another through time.

The wander of Atlantic Ocean hot spots relative to Indian Ocean hot spots has been examined in a number of papers [e.g., Morgan, 1981, 1983; Duncan, 1981; Duncan and Richards; 1991]. The general consensus from these studies is that the maximum rate of hot spot wander approaches 5 mm/yr over periods of 120 m.y., although constraints on the motion between Pacific hot spots are fewer because subduction zones isolate the plates in this basin from the global plate motion circuit except through Antarctica. Molnar and Stock [1987] postulated that the Hawaiian hot spot has moved southward 10-20 mm/yr with respect to hot spot traces in the Atlantic and Indian Oceans, but they did not account for deformation between East and West Antarctica. Including deformation in Antarctica reduces the amount of motion between hot spots to less than 10 mm/yr (J. Stock, as quoted by Duncan and Richards, 1991). For the initial models presented in this paper, the 5 mm/yr rate is viewed as an upper limit to inter-hot spot motion, and all motion is in reference to the hot spots defining the various poles of rotation for the Pacific plate (e.g., Hawaii). Wander rates above 5 mm/yr are considered suspiciously high.

Estimating rates and direction of wander requires that a well-constrained reference point be available for matching the modeled tracks with the available age data. As demonstrated previously the Hawaiian and Emperor chains provide good constraints on the stage poles and angles of rotation up to about 78 Ma. Consequently, the bend in the tracks corresponding to the change in plate motion at 78 Ma can be used as a reference point for where a particular track should lie with respect to a dated edifice in the Marshall Islands, assuming it is known which hot spots were responsible for which volcanic chains. The stage pole for the 78 Ma to 97 Ma time period is also probably valid, although the angle of rotation may be somewhat large given the additional age data obtained from the Musician
Seamounts [Pringle, 1992]. To simplify the hot spot wander models as much as possible, this paper applies two additional assumptions:

1. No "new" or unidentified hot spots existed during the construction of the Marshall Islands (i.e., the presently-active French Polynesian hot spots are responsible for the linear chains of volcanoes in the Marshall Islands), and

2. The reference frame defined by the four hot spots in French Polynesia remains essentially consistent through time. In other words, all the hot spots move in the same direction albeit at different rates with respect to the reference hot spot.

Allowing the French Polynesia hot spot reference frame to wander northeast through time (azimuth of 49°) such that the 82 Ma radiometric age on Woden-Kopakut in the Ratak chain aligns with the 86 m.y. age of the Mehetia track results in a similar alignment of the 78 Ma to 97 Ma portion of the Rurutu track with the Ralik chain, and the Macdonald track with the kink in the Ratak chain around Lobbadeede Guyot (Figure 4.5). The matching of the Woden-Kopakut age with the 86 m.y. age on the Mehetia track assumes: 1) that this hot spot constructed this edifice, and 2) that the main shield-building stage of Woden-Kopakut occurred ~4 m.y. before the dated volcanics. Even though the wander rates listed in Table 4.5 explain many of the observed features without having to invoke new hot spots, inconsistencies between the modeled tracks and the observed data still exist. For example, in order for the Mehetia track to align with the southern Ratak chain, it would have to wander at a rate of ~8 mm/yr. By doing so it would move well away from the dated edifice of Woden-Kopakut. Another example is the divergence of the Rurutu track from the northern portions of the Ralik chain (Figure 4.5). Finally, wander of the Macdonald hot spot fails to explain the mid-Cretaceous ages noted on Lewa and Lobbadeede guyots.

Perhaps these misfits are trivial considering the current data base, but one additional possibility which should be examined is a change in the direction of hot spot wander through time. Morgan [1981] notes that while the maximum rate of motion between
Table 4.5. Hot spot wander rates and directions.

<table>
<thead>
<tr>
<th>Location</th>
<th>Rate</th>
<th>Direction</th>
<th>Rate</th>
<th>Direction</th>
<th>Rate</th>
<th>Direction</th>
<th>Rate</th>
<th>Direction</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mehetia</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Model 1</td>
<td>Stationary</td>
<td></td>
<td>Stationary</td>
<td></td>
<td>Stationary</td>
<td></td>
<td>Stationary</td>
<td></td>
</tr>
<tr>
<td>Model 2</td>
<td>8.0</td>
<td>NE</td>
<td>6.0</td>
<td>NE</td>
<td>4.3</td>
<td>NE</td>
<td>1.5</td>
<td>NE</td>
</tr>
<tr>
<td>Model 3</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(0 - 78 Ma)</td>
<td>8.0</td>
<td>NE</td>
<td>6.0</td>
<td>NE</td>
<td>4.3</td>
<td>NE</td>
<td>1.5</td>
<td>NE</td>
</tr>
<tr>
<td>(78 - 130 Ma)</td>
<td>8.0</td>
<td>SW</td>
<td>6.0</td>
<td>SW</td>
<td>4.3</td>
<td>SW</td>
<td>1.5</td>
<td>SW</td>
</tr>
<tr>
<td>Macdonald</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Rurutu</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Rarotonga</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Atlantic and Indian ocean hot spots is around 5 mm/yr, this wander does not appear to be in a preferential direction. The modeling presented so far in this paper attempts to apply a consistent direction of wander among hot spots, primarily to bring the 78 m.y. track positions in line with the presumably appropriate Marshall Islands chains. One way to compensate for the divergence between the model track of Rurutu hot spot and the dated edifices in the northern Ralik chain is to reverse the wander direction at the 78 m.y. change in plate motion (Figure 4.6). Reversing the direction of wander is somewhat arbitrary, but its effect is to simplify the model. Such a change produces a better fit to the observed radiometric ages in the Ralik chain (Figure 4.6a). Applying a similar change in wander direction to the entire suite of French Polynesian hot spots again produces a mixed result. The Ratak chain and the Mehetia track fit better but this requires a high rate of hot spot wander (~8 mm/yr). A surprisingly good fit also occurs between the mid-Cretaceous edifices in the northern Ralik chain and the 97 Ma to 130 Ma portion of the Macdonald track. Unfortunately, such wander moves the Macdonald track away from the dated edifice of Lokworkwor. A higher rate of wander, comparable to the 8 mm/yr assigned to Mehetia, is necessary to make the track fit this edifice.

4.4. Discussion

At present, no single model adequately explains the observed distribution of edifice ages in the Marshall Islands without applying some tenuous assumptions regarding the longevity or fixity of hot spots. If we are willing to accept that different hot spots existed in French Polynesia during the Cretaceous, then matching ages to linear trends in the edifices is a fairly straightforward process: Rurutu hot spot formed the Ratak chain, Macdonald hot spot was responsible for mid-Cretaceous edifices in the northern Ralik chain (although even in this case some hot spot wander is necessary for the track of Macdonald to correspond with the dated edifices), the Rarotonga hot spot is somehow
Figure 4.6. Oblique mercator projections for the Late Cretaceous showing the affects of changing the direction of hot spot wander through time. (a) Reversing the direction that Rurutu wanders from southeast in the pre-78 Ma time frame to northeast during the post-78 Ma time frame appears to produce a better fit to the age data available for the Ralik chain. (b) Applying a similar change in the direction of hot spot wander for all the French Polynesian hot spots (using the rates of wander shown in Table 4.5 and Figure 4.5b) produces perhaps a better fit to all the tracks.
(a) Variable Wander Direction for Rurutu Hot Spot

(b) Variable Wander for all Hot Spots
linked to the volcanism in the Ujlan-Anewetak group, and an unidentified hot spot was responsible for the construction of the Ralik chain in the 78 Ma to 97 Ma stage period. If on the other hand the presently existing hot spot reference frame in French Polynesia applies to the Cretaceous then not only do anomalously fast rates of wander possibly apply to the Mehetia and Macdonald hot spots, but the direction of hot spot wander may also have changed through time.

With regards to the best-fitting model (i.e., one invoking high rates of wander for Mehetia and Macdonald hot spots and a reverse in wander direction at the 78 Ma change in plate motion) one must consider whether the assumptions applied to generate this model are reasonable. Certainly the direction of hot spot wander could have changed through time, and it seems reasonable that some link might exist between changes in plate motion and the direction a particular hot spot wanders. No preferential movement of hot spots has been noted in the literature, although the models presented here attempt to do so for the Marshall Islands. Changing the direction that a hot spot wanders by 180° seems suspicious but it simplifies the model considerably. Perhaps even more contentious than a complete reversal in wander direction is the requirement for anomalously fast rates of wander in the case of Mehetia and Macdonald. The rates used in the best-fitting model are approximately twice as fast as the rates assigned to Atlantic and Indian ocean hot spots, but well within the 10 mm maximum ascribed for the Hawaii hot spot by Stock and Molnar (as quoted in Duncan and Richardson, 1991). Our understanding of the South Pacific Superswell and its hot spots strongly influence many of the assumptions concerning hot spot longevity and stationarity in the Marshall Islands region. The lithosphere over the Superswell appears to be more susceptible to hot spot penetration [McNutt and Fischer, 1990]. Little information exists on the long-term stability of the French Polynesian hot spots as the oldest volcanoes dated in either the Cook-Austral or Society chains is < 30 m.y. old. Consequently, making direct links between volcanic chains in the Marshall Islands and hot spots in French
Polynesia requires some technique other than the physical traces recorded on the sea floor (perhaps the isotopic abundances or geochemistry of the volcanic rocks).

4.5. Conclusions

The results of modeling hot spot trends through the Marshall Islands are far from conclusive, but a tentative correspondence between volcanic chains in the Marshalls and hot spots in French Polynesia can be established subject to a number of assumptions. The modeling presented here illustrates that:

1. The Ralik chain formed during the 78 Ma to 97 Ma stage pole period, but does not correspond to any of the existing French Polynesian hot spots. Rurutu hot spot would have to wander in a northeast direction at a rate of ~4.3 mm/yr relative to the Hawaiian hot spot in order to account for this volcanic chain.

2. Macdonald hot spot requires a wander rate of ~4 mm/yr in a northeast direction to align its track with the fossil-dated mid-Cretaceous edifices in the northern part of the Ralik chain.

3. The Ratak chain could result from either the Rurutu or Mehetia hot spots, depending on whether these hot spots remained stationary or wandered through time. If the reference frame of French Polynesia maintained any sort of internal consistency since the Cretaceous, then Mehetia is probably responsible for the construction of most edifices in the Ratak chain.

4. The group of edifices clustered around Anewetak and Ujlan atolls appear to correlate with the Rarotonga hot spot and a change in plate motion around 78 Ma. This age appears to be consistent with the available radiometric ages from this group (76 Ma for Anewetak, 80 Ma for Lalibjet), although at least one older edifice resides in this area (Lo-En Guyot).
In general, the correlation between hot spots associated with the South Pacific Superswell and volcanic edifices in the Marshall Islands appears promising. In particular the mid-Cretaceous track of the Macdonald hot spot through this area may permit the calculation of a pre-100 Ma stage pole for Pacific plate motion relative to hot spots given that more age information can be obtained (perhaps through the inversion of magnetic data over various guyots). Allowing very simple hot spot wander improves the fit between modeled tracks and observed data while maintaining the relative hot spot reference frame in French Polynesia. This model seems more desirable than creating or removing unidentified hot spots. The anomalously high rates of wander required by the Mehetia hot spot to fit the Ratak chain suggest the model needs additional work before attaining a more complete understanding of hot spot trends through the Marshall Islands.
Table 4.6. Location of Marshall Island edifices.

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<tr>
<th>Edifice</th>
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Table 4.6. (Continued) Location of Marshall Islands edifices

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CHAPTER 5

CONCLUSIONS

The data collected during MW8805 serve as cornerstones for building a relatively comprehensive and realistic understanding of how and when the Marshall Islands formed, in addition to filling a substantial gap in our knowledge of Cretaceous volcanic events in the Pacific. The combined information from side-scan images, swath bathymetry, single- and 6-channel seismic profiles, echosounder profiles, gravity and magnetic measurements, and dredge samples permit accurate interpretations of the various morphologic and stratigraphic features observed on selected guyots. These data also illuminate similarities and differences between guyots within the Marshall Islands and various island and atoll counterparts scattered throughout the Pacific basin. For example, from the morphology and seismic stratigraphy of the sediments perched along the south flank of Wodejebato, especially in relation to the basalt and limestone samples recovered along this slope, it was apparent that portions of the south flank were eroding in the form of relatively large fault-blocks. Traditionally, terraces along the flanks of guyots were thought to be wave-cut features associated with low-stands of sea level. Similar fault-blocks are also observed along the flanks of Lo-En and Limalok guyots, the other two platforms drilled during Leg 144. Large-scale slope failure is common on volcanic islands in the Hawaiian and French Polynesian chains, but the mechanism responsible for continued flank erosion on ancient edifices remains unclear. Possibilities include sediment overburden associated with the shallow-water carbonate platform capping some of the truncated volcanic islands and inherently steep, unstable slopes associated with all volcanic islands. Another insight gained from the integration of the geophysical data with the rock samples was the nature of perimeter ridges observed along some summit plateaus. On Wodejebato, two perimeter ridges appear along the broad shelves formed by the north and northeast flank ridges. The inner perimeter remains as a relatively
continuous feature around the edge of the summit plateau, whereas the outer perimeter ridge loses relief away from the broad flank ridge shelves. Preliminary drilling results from Leg 144 (Sites 874 through 877) suggest that the inner ridge represents more of a true reef tract whereas the outer ridge may represent a fore-reef apron deposit.

Substantial differences exist in the morphology, lithology, and ages of guyots within the Marshall Islands. Perhaps the most notable difference between the three guyots drilled during Leg 144 is the absence of a shallow-water carbonate platform on top of Lo-En. The three guyots, Limalok, Lo-En, and Wodejebato, are all comparable in some respects to modern islands and atolls in French Polynesia, and this area of Pacific holds perhaps the most promise for understanding the early tectonic environment responsible for the formation of edifices within the Marshall Islands. Plate flexure associated with volcano construction has almost certainly influenced the history of uplift and subsidence experienced by edifices within the Marshall Islands, but with the paucity of age data from the various island groups only general models can be constructed showing relative regions of arching or down-warping. The flexure models presented in this dissertation are relatively simple and only examine the loading effects associated with the construction of a single nearby volcano. In the case of Lo-En and Wodejebato, the load added to the plate represents the volcanic edifices beneath Anewetak and Pikinni atolls, respectively. Flexure from Anewetak alone results in negligible uplift on Lo-En, and may in fact cause up to 100 m of subsidence. Such subsidence may have prevented the previously truncated summit of Lo-En from reaching a depth sufficiently shallow to allow the colonization of coral organisms. An echosounder profile across Lo-En shows that the summit plateau tilts gently towards Anewetak (deepening by ~35 m from south to north), thus supporting the model results. Plate flexure may also explain the high-backscatter cone- and lobe-shaped features cropping out along the flat summit plateau of Lo-En. In this case the Late Cretaceous volcanism occurring on Anewetak may have also
occurred on Lo-En as the result of tensional stresses producing "easier" paths for the magma to erupt. If Lo-En was still submerged at this time, then the volcanic cones would not be removed through subaerial erosion. The simple flexure models presented here also suggest that construction of the volcanic platforms beneath Wodejebato and Pikinni, and possibly Limalok and Mili, occurred without a substantial time gap as neither the volcanic nor the carbonate platforms tilt in the direction of the atolls (the assumption being that the volcanos beneath the atolls were constructed later than those forming the guyots). What the models fail to clearly establish is the cause of carbonate platform drowning (or survival). Edifice size is certainly a possibility (with smaller edifices truncating faster and therefore deeper during low-stands in sea level) but without better basement depth estimates for the atolls attached to the guyots drilled during Leg 144, unequivocal support for the Grigg and Epp [1987] model is lacking.

A good understanding of the volcanism and tectonism affecting the Marshall Islands is crucial for any attempt to establish a eustatic sea level curve from the sediments cored during Leg 144. Additional analyses of the sediments recovered during Leg 144 will undoubtedly shed light on the drowning mechanisms for carbonate platforms in this region, be they environmental (cooling of water temperature, over-nutriification of surface waters, storm disturbances), biological (changes in faunal growth rates), or geological (uplift or subsidence brought about by plate flexure, secondary episodes of volcanism, erosional differences) in nature. Until the sequence and timing of volcanism in the Marshall Islands is better understood, and hence the tectonic signal of sea level changes can be removed, using drowned carbonate platforms in the Marshall Islands as "dipsticks" for recording the history of Cretaceous sea level fluctuations is not possible. Consequently, the next step towards understanding the evolution of the Marshall Islands as a whole is the identification of hot spot trends within the various island chains.
The results of modeling hot spot trends through the Marshall Islands suggest that the majority of edifices within the various chains composing this island group can be explained by a few hot spots French Polynesia. The modeling presented in this dissertation illustrates that:

1. The Ralik chain formed during the 78 Ma to 97 Ma stage pole period, but does not correspond to any of the existing French Polynesian hot spots. Rurutu hot spot would have to wander in a northeast direction at a rate of ~4.3 mm/yr relative to the Hawaiian hot spot in order to account for this volcanic chain.

2. Macdonald hot spot requires a wander rate of ~4 mm/yr in a northeast direction to align its track with the fossil-dated mid-Cretaceous edifices in the northern part of the Ralik chain.

3. The Ratak chain could result from either the Rurutu or Mehetia hot spots, depending on whether these hot spots remained stationary or wandered through time. If the reference frame of French Polynesia maintained any sort of internal consistency since the Cretaceous, then Mehetia is probably responsible for the construction of most edifices in the Ratak chain.

4. The group of edifices clustered around Anewetak and Ujlan atolls appear to correlate with the Rarotonga hot spot and a change in plate motion around 78 Ma. This age appears to be consistent with the available radiometric ages from this group (76 Ma for Anewetak, 80 Ma for Lalibjet), although at least one older edifice resides in this area (Lo-En Guyot).

Given the misfits of the modeled tracks to the observed topography and available edifice ages, it is somewhat unclear the degree to which hot spots in French Polynesia were active in the Cretaceous. The simplest means of explaining the various model misfits is to either allow for more hot spots to have existed in French Polynesia than are now
presently active or to create or eliminate the physical trace of various hot spots during various periods of time. As shown in the discussion above though, it is possible to match tracks from presently active hot spots with the volcanic chains composing the Marshall Islands. Perhaps the most important implication of the modeling is the possibility that the mid-Cretaceous track of the Macdonald hot spot through this area may help define a pre-100 Ma stage pole for Pacific plate motion relative to hot spots.

Models depicting the evolution of the Marshall Islands and their relationship with the volcanic events occurring in French Polynesia will continue to undergo revisions as the drilling information acquired during Leg 144 is integrated with the existing geophysical data. Perhaps the final question which needs to be addressed is where should the emphasis of additional work in the Marshall Islands be placed? The sediment and rock samples recovered during Leg 144 will undoubtedly improve our understanding of Cretaceous sea level changes, but the drilling failed to clearly establish the presence of mid-Cretaceous volcanic platforms in the Marshall Islands. Given an improved understanding of features associated with guyots in conjunction with the model results presented in Chapter 4, I feel another surveying and sampling cruise should be in order within the next couple of years. Such a cruise should split time between collecting high-resolution geophysical data (including side-scan images, multibeam bathymetry, 6-channel seismic data, and selected long lines of gravity and magnetic measurements) and dredging basalts and limestones. Emphasis should be placed on identifying a mid-Cretaceous hot spot trend through the Marshall Islands as this will extend Pacific plate motion reconstructions beyond the pre-100 Ma period. If a more detailed analysis of the Cretaceous carbonate platforms appears favorable, I would suggest a series of submersible dives along the south flank of Wodejebato to sample the limestones cropping out along the fault scarp.
REFERENCES


