27 May 1983, Volume 220, Number 4600 SCIENCE

Insular Erosion, Isostasy, and Subsidence

H. W. Menard

Normal oceanic crust is created at the shallow crest of a midocean ridge and then, as it ages and cools, it subsides at a rate that is proportional to the square root of time. The subsidence can be quantified because both the depth and magnetic age of the sea floor have been determined in millions of places, producing a standard relation between depth and age. Anomalously shallow regions 1 to 2 kilometers high and 500 to 1000 km in diameter have been identified, and the origin and history of these vast midplate swells are controversial. Competing hypotheses at present include the following: (i) dynamic elevation over rising mantle convection (*1*), (ii) compositional changes that decrease the density of the lithosphere (*2*), and (iii) heating that decreases the thickness and density of the lithosphere (*3, 4*).

Summary. Organic reefs and shore erosion record the intersection of sea level with islands. From this record it is possible to reconstruct the history of vertical movement of the islands and the adjacent deep sea floor, including midplate swells. As judged by coral thickness, islands with barrier reefs sink as though they were on thermally youthful crust regardless of the actual age. Reefless islands do not sink until truncated by erosion. Apparently, thermal subsidence is balanced by isostatic uplift in response to erosion. Barrier reefs prevent wave erosion of encircled volcanoes and capture products of stream erosion so that isostatic uplift is eliminated. Insular shelves widen initially at rates of 0.6 to 1.7 kilometers per million years; the rates decrease with time. Thus the subsidence of islands depends on the size of the island and the presence of reefs, and it may not always be the same as that of the surrounding oceanic crust.

All the factors may contribute to the origin of midplate swells, but the importance of the last hypothesis, thermal rejuvenation, is supported by a few dozen observations of the rate of subsidence of the swells. The deep ocean basins contain roughly a thousand drowned volcanic islands including guyots, which are merely submerged, and atolls, which are submerged and capped with coral reefs. Almost all active volcanoes in ocean basins are either on ridge crests or midplate swells, and the low relief of the drowned ancient volcanoes shows that they too were active in relatively shallow water and thus in the same tectonic environments (*5*). From the present depth of drowned islands it is possible to determine the rudiments of the subsidence history of midplate swells. Detrich and Crough (*3*) and Crough (*3*) observed that some atolls and guyots on Mesozoic crust have subsided as though on normal oceanic crust 15 to 25 million years old. Such crust is often described as having an equivalent thermal age of 15 to 25 million years. Guyots on younger crust subside as though their equivalent thermal ages were zero (*4*).

Information on atoll and guyot subsidence, however, is sparse, particularly data on the early stages of subsidence, which are critical to establishing the equivalent thermal age. The morphology of volcanic islands and the stratigraphy of fringing and barrier reefs are too complex to be reconstructed from a few dozen drill holes and a few hundred dredge hauls. However, the study of the hundreds of islands that are now in the initial stages of subsidence may be a useful guide in interpreting the ancient geologic history.

The principal factors that influence the relative vertical movement of a volcanic island are thermal subsidence of the underlying crust, eustatic changes in sea level, and isostasy. Crustal subsidence may be as much as 2000 meters and occurs at a rate that varies inversely with the square root of the equivalent thermal age. Eustatic changes in sea level may cause either relative emergence or submergence in a range of a few hundred meters. These changes, although clearly important, are considered as of second order here because they are relatively small and of varying sign. Isostatic effects result from compensation for thermal subsidence, eustatic fluctuations, loading by growth of coral, and unloading by erosion. The importance of these effects depends on the density contrasts between air, water, volcanic rock (2.4 grams per cubic centimeter) and coral reef (1.8 to 2.2 g/cm^3). The effect on subsidence of loading by coral growth in a barrier reef, for example, is relatively small because the density contrast between coral and the water it displaces is only 0.8 to 1.2 g/cm^3. Moreover, subsidence is opposed by isostatic compensation for the submergence of the volcanic island encircled by the reef.

By far the most important effect may be expected from erosion. The importance of erosion-isostasy in the history of epeirogeny (broad, gentle warping) of islands is the focus of this article. Because of a bountiful sample of islands in different circumstances, it is possible to isolate the effects of thermal subsidence and erosion-isostatic uplift to a degree that is unusual in geology by (i) studying islands encircled by barrier reefs to measure thermal subsidence when the erosion-isostasy effect is largely eliminated; (ii) analyzing the subsi-

The author is professor of geology in the Institute of Marine Resources and Scripps Institution of Oceanography, University of California, La Jolla 92093.

dence of large reefless islands to determine whether erosion-isostasy can balance thermal subsidence under the most favorable circumstances; (iii) studying the transitional effects when a balance does not occur; and (iv) measuring rates of erosion to evaluate the persistence of reefless islands before submergence begins.

Subsidence of Islands with Barrier Reefs

The fact that islands with barrier reefs subside can hardly be doubted. Darwin's elegant hypothesis of the origin of atolls (*6*) by subsidence of volcanic islands even accounts for many of their morphologic characterisics that were unknown to him (*7*), as well as their age sequence (*8*). Some of the pertinent characteristics of dated islands with barrier reefs in the main ocean basins are summarized in Table 1. Note especially the characteristics of sea cliffs. Volcanic island cliffs form penecontemporaneously as the islands grow; immediately after the volcano becomes inactive, the cliffs commonly are 300 to 400 m high. Thus the 300-m cliffs of southeastern Tahiti (Tahiti-Iti, 0.4 million years) are normal. Northwestern Tahiti (Tahiti-Nui, 0.5 to 1.2 million years) is largely ringed by cliffs, although the vicinity of the harbor of Papeete is the exceptional area seen by Darwin. The sea cliffs are 60 to 150 m high and generally removed from wave erosion by a narrow coastal plain, a narrow lagoon, and a barrier reef. Moorea (1.5 to 1.6 million years) still has cliffs but only on one side. Except for 15-m cliffs on adjacent sides of the paired islands of Raiatea (2.4 to 2.6 million years) and Tahaa (2.6 to 2.9 million years), sea cliffs are not found on older islands with barrier reefs. Darwin's hypothesis offers a simple explanation of why the cliffs become lower and disappear with age.

The existence and temporal distribution of other morphological features are also revealing. Embayments are a consequence of submergence of valleys, and they are rare or minor in reefless islands but conspicuous in those with barrier reefs. Deltas partially fill some bays in younger islands that are subject to vigorous stream erosion, but older islands cannot supply adequate sediment and the bays are open. The relief of volcanic islands also indicates subsidence. The maximum heights of the peaks tend to decrease with age, but the relation may be influenced by initial size and by continuing erosion as well as by subsidence. The distribution of tiny, off-lying volcanic islands is more significant. Large volcanoes in the tropics tend to be carved by rivers into fantastic ridges and spires. Ridges on Moorea and in the Marquesas Islands are so thin and steep-walled that great windows are eroded through. The narrow saddles between spires are lower on older islands, and after about 2.5 million years some saddles have vanished leaving small outlying islands (Fig. 1). This disappearance continues until at Mangareva and Truk the once connected peaks have become widely separated, small, steep-sided, islands. Darwin's hypothesis explains this sequence.

The rate of subsidence is conjectural since drilling data and even geophysical profiles are lacking, but morphology may lend support to the conjecture. The original shape of most oceanic shield volcanoes is easily reconstructed for 2.5 million years after subsidence begins, because the style of cliff erosion by streams preserves interfluves. From topographic and bathymetric maps, the shape of a young volcanic platform can usually be distinguished from the shape of a reef built on it (Fig. 2), and the thickness of the reef can then be estimated. Davis (*7*) used this technique for subaerial slopes, but it is questionable for older islands where the original interfluves have been eroded into ridges. The submarine slopes are preserved, however, and thus the thickness of coral can be estimated in many places (Fig. 1 and Table 1).

Examination of nine islands with barrier reefs protecting them from erosion shows that they have subsided rapidly just as the subaerial morphology suggests. The subsidence of six of these islands with ages between 0.5 and 4.0 million years ranges randomly from about 250 to 350 m, but the older islands of Mangareva, Ponape, and Truk have subsided 600 to 1100 m. Standard subsidence curves can be fitted to the observations on the assumption that subsidence is caused by cooling. The curves indicate an equivalent thermal age of 0 to 3 million years (Fig. 3) despite the fact that the actual age of the underlying oceanic crust is early Tertiary or Mesozoic.

This morphologic analysis of islands with barrier reefs shows that some oceanic volcanoes tend to sink as though they were on thermally rejuvenated crust. But because barrier reefs protect islands from waves and capture most sand and mud from stream erosion, isostatic compensation for erosional losses is largely eliminated. A question thus arises about how this compensation would compare with thermal subsidence when both processes are active.

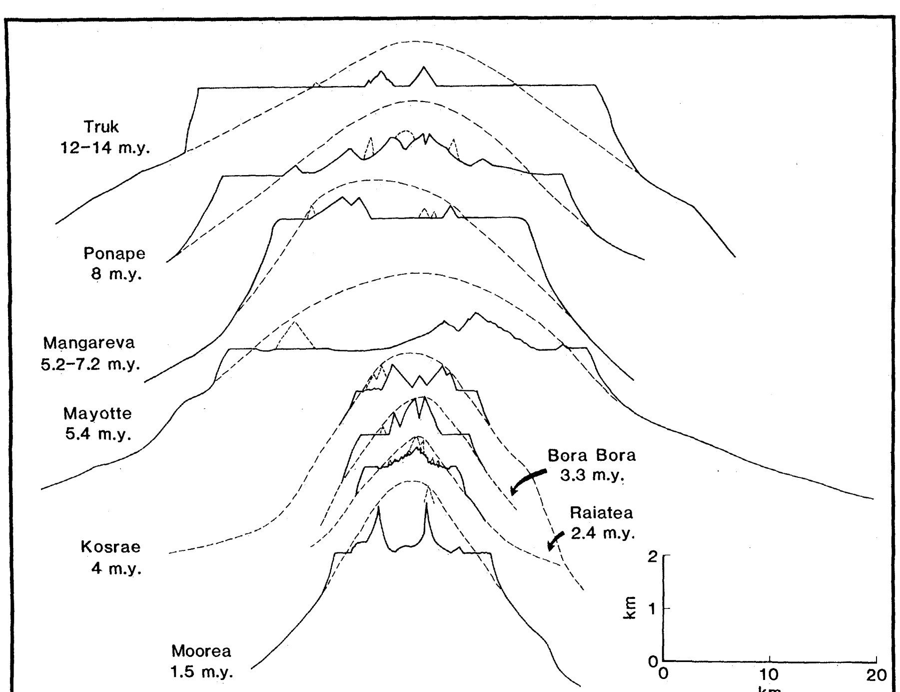


Fig. 1. Morphology of dated oceanic volcanoes with barrier reefs. Short dashed lines indicate relief interpolated into profile, and long dashed lines represent assumed subaerial and probable relief buried under reefs. Ages are from (*8, 32–34*). Vertical exaggeration, ×5; *m.y.*, million years.

Subsidence of Reefless Islands

The net subsidence of a reefless island relative to sea level can be determined by measuring the depth of the break between the insular shelf and slope—the shelf break. The fact that shelves sur-

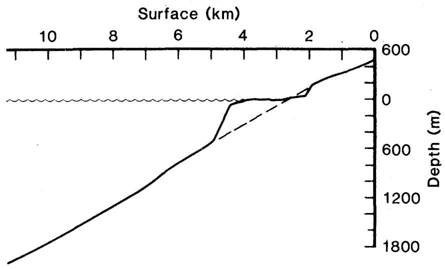


Fig. 2. Profile on the west side of Tahiti showing initial surface constructed by the volcano and later modified by a sea cliff and barrier reef. Dashed line is assumed initial surface.

rounding islands are dominantly erosional is shown by the towering cliffs of adjacent shorelines, the prevalence of marine stacks near shore, and abundant notations of rocky bottom on nautical charts. The shelf break, at the outer edge of the erosional terrain, is the product of vigorous wave erosion, which does not occur in water deeper than about 10 m (*9*). Thus the break is a relatively sensitive indicator of sea level.

If the epeirogeny of reefless islands is, like that of barrier reefs, largely influenced by thermal subsidence, the depth of the shelf break should be directly related to the age of the volcano and inversely related to the age of the crust. Thus the depth of the shelf break of an old volcano on young crust should be greater than that of a young volcano on old crust. No such relation is apparent. Shelf-break depth shows no significant variation with age on 23 islands on crust 3.5 to 125 million years old. This might be the result of thermal rejuvenation, which could produce a relatively uniform rate of subsidence. If so, the increase in the depth of the shelf break with volcano age should be more obvious: it is not (Fig. 4).

The depth of the shelf break in clustered islands may be affected by crustal warping related to sequential loading (*10*); organic reefs may also complicate the picture. Consequently it is desirable to consider only relatively isolated and reefless islands. The shelf break off seven active or very young islands where coastal rocks have been dated ranges from about 80 to 180 m deep. Five islands, 1 to 3 million years old, have shelf breaks from 80 to 130 m; for three islands, 4 to 12 million years old, the breaks are between 70 and 120 m. Isolated volcanic islands as old as 12 million years apparently have not subsided relative to active volcanoes.

Although the significance of the relation is less clear because of clustering, the depth of the shelf break in 25 places in the Hawaiian and Canary islands also shows little correlation with age. A possible exception is the break at 220 m off one sector of Gran Canaria Island, which is between 10.3 and 13.6 million years old (*11*), but off two other sectors of similar age the depth is only 120 m, which is close to the 90 and 110 m depths of breaks off two sectors 2 to 4 million years old.

Reefless islands of all ages appear to remain at sea level until they are almost completely truncated. Likewise, guyot platforms are generally smooth, indicating that they were stable long enough for such truncation. The contrast with the rapid subsidence of barrier reef islands of comparable ages is striking (Fig. 5). The abyssal sea floor sinks as it cools, barrier reefs sink, atolls sink, guyots (once truncated) sink, but reefless islands do not sink until truncated. Why not? Some force is required that acts on volcanic islands but not other parts of ocean basins and that is adequate to balance thermal subsidence. If isostasy applies throughout, the only significantly different influence must be that of erosion, and, in order to preserve isostatic balance, the crustal column under an island must rise as the island is eroded away.

Confirmation that the uplift associated

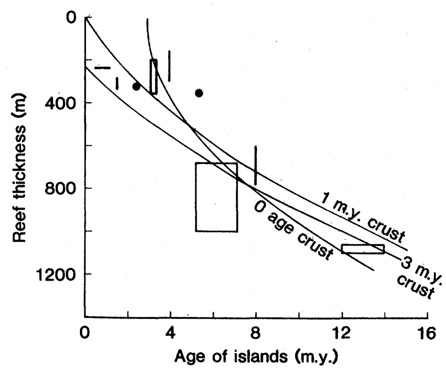


Fig. 3. Estimated thickness of barrier reefs around dated islands on old crust compared with thickness that would result from subsidence of crust with young equivalent thermal ages. Crust with an equivalent thermal age of as little as 3 million years (*m.y.*) would subside too slowly to fit the data (*35*). Size and shape of symbols indicate ranges of age or thickness.

with erosion-isostasy can balance thermal subsidence in favorable circumstances can be derived from the relation between island age and the width of insular shelves. If an island tends to remain at sea level, the width of the shelf should increase with age. However, if vulcanism occurs late in the shield-building stage or in the posterosional stage, lava commonly flows across the shelf and resets the relation between width and age. Thus it is more accurate to compare the width of a shelf to the age of the adjacent coast than to the age of the whole island, if data are available.

The ages and shelf widths of eight isolated Atlantic islands and eleven major sectors of the western Canary Islands (Fig. 6) have been compiled. The simplest interpretation of the data is a linear rate of shelf widening of 0.6 to 0.7 km every million years for 16 million years. A similar analysis for the relatively reefless Hawaiian and Marquesas islands indicates a faster rate of widening, 1.1 to 1.7 km every million years for 5 to 7 million years. The width of the shelf for

Table 1. Morphological evidence for rates of subsidence of islands with barrier reefs.

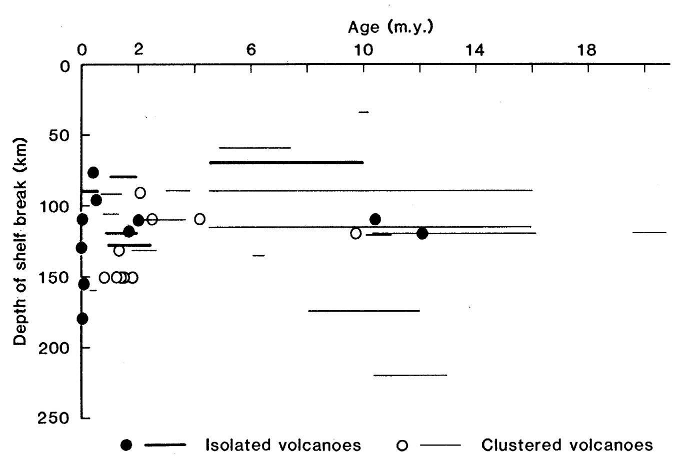
Island	Age (million years)	Cliff height (m)	Mountain height (m)	Coral thickness (m)	Embayed	Outlying volcanic islands	Cliffs
Tahiti-Iti	0.4 (*8*)	300 (*7*)	1315		Filled (*2*)	No	Widespread
Tahiti-Nui	0.5 to 1.2	60 to 150	2224		Filled	No	Widespread
Moorea	1.5 to 1.6	60 to 150	1207	280 to 340	Yes	No	North side
Huahine	2.0 to 2.6		669		Yes	Yes	
Raiatea	2.4 to 2.6	15	1035	320 to 340	Yes	Small	
Tahaa	2.6 to 2.9	15	590		Yes	Small	
Bora Bora	3.1 to 3.4		723	200 to 360	Yes	Yes	
Kosrae	4.0 (*32*)		625	160 to 300	Filled	One	
Mayotte	5.4 (*33*)		660	360	Yes	Yes	
Mangareva	5.2 to 7.2 (*34*)		425	680 to 1000	Yes	Many	
Ponape	8.0		786	600 to 800	Yes	Yes	
Truk	12.0 to 14.0		440	1060 to 1100		Many	

older islands in the Hawaiian chain was excluded from consideration because of upward growth of coral.

More complex relations are suggested if only dated coasts of all islands, including active volcanoes, are examined. The average rate of shelf widening appears to be relatively fast at first and then declines after 5 to 6 million years. This seems reasonable in that, other things being equal, the later stages of marine peneplanation presumably are relatively slow. However, the data are much more scattered for older coasts than for younger ones, suggesting that variable forces are at work. In part this reflects variations in the intensity of waves and swells not unlike the variations in stream erosion on the weather and lee sides of high islands. For example, Lord Howe Island is not in the center of its shelf but is 14 km from the shelf break on the weather side and 5 km away on the lee side. The great variation in shelf width of older islands may also be related to the intensity of subaerial erosion, which produces deep valleys on the weather side. Waves can erode headlands between valleys to cut a broad shelf much faster than they can the continuous high cliffs of the lee side.

The erosion and subsidence of Iceland are particularly difficult to interpret because the island has been sequentially produced by ridge-crest volcanism during late Cenozoic time. However, it is also particularly interesting with regard to the balance between thermal subsidence and the erosion-isostasy uplift. At a ridge crest, the sea floor should sink about 350 m in 1 million years and more than 1100 m in 10 million years, but the voluminous volcanism has built an enormous island that is being eroded rapidly by glaciers, great rivers, and the heavy swell of the North Atlantic. Despite the complexities of Icelandic geology, the depth of the shelf break seems to indicate what has occurred in these extreme conditions. According to the most detailed published charts, the shelf break does not deepen with age but the shelf grows wider. However, the Icelandic shelf has been prograded by voluminous sediment (*12*), and the depth of the break probably reflects the last lowering of sea level. Even so, the mere existence of older subaerial outcrops shows that even on a ridge crest a large volcanic island persists while the deep sea floor around it subsides rapidly.

This is also true for the early history of many pairs of aseismic ridges which, like Iceland, are the products of profuse volcanism at spreading centers (*13*). Most of these active volcanoes were large islands that ultimately were truncated and then subsided. The shape of the subsidence curve is comparable to normal oceanic crust, but the depth is always less than that expected for the age (*14*). This has been explained by observations that the volcanic peaks were far above sea level, and it was proposed that because they maintain that relief they are always above normal depth. The relation discussed above provides another explanation; indeed it predicts a depth above normal. A large island stays at sea level until it is truncated by erosion although the surrounding sea floor sinks as it cools. Then they subside together at rates determined by the equivalent thermal age. This explanation makes the subsidence of the Iceland–Faroe Ridge predictable instead of anomalous. The aseismic ridge was derived from the voluminous Iceland hot spot. Drilling has shown that near Deep-Sea Drilling Project site 336 the ridge was an island (*15*) which is 1100 m shallower than expected for its age (*14*). The sedimentary record suggests that the site was above sea level for about 15 million years. The depth, as opposed to the shape, of a subsidence curve is a function of the size of an island. Thus if the island at site 336 approached the size of Iceland it would be expected to persist for a long period,



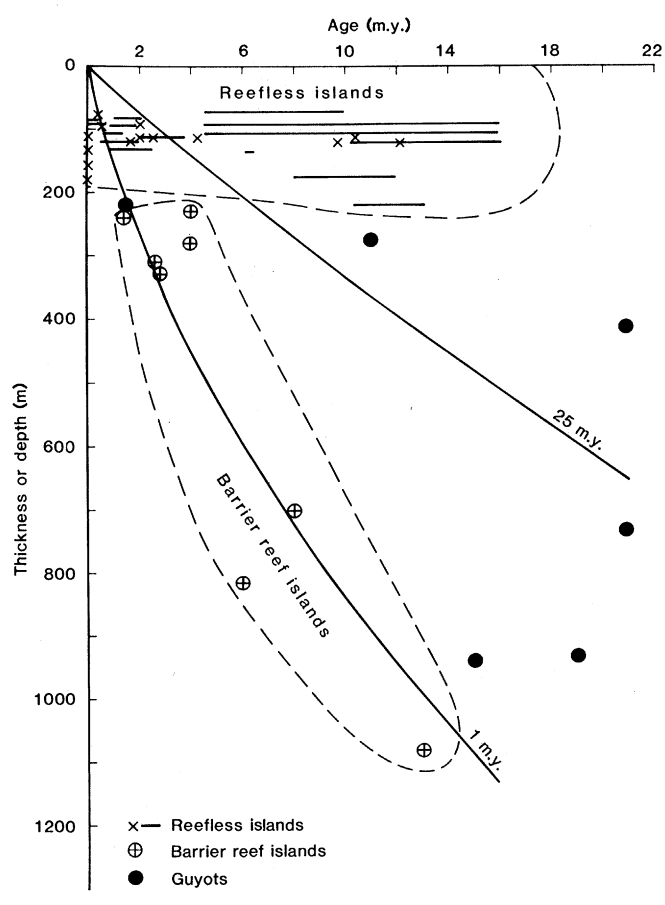


Fig. 4 (left). Lack of correlation between island age and the depth of its shelf break. Isostatic response to erosion seems to balance thermal subsidence until islands are truncated. Because sea level was 150 m lower only 12,000 years ago, only depths below that indicate subsidence; *m.y.*, million years. Fig. 5 (right). Comparison of subsidence of barrier reef islands, reefless islands, and reefless guyots. Presumably the guyots remained at sea level, like reefless islands, until truncated and then sank like barrier reef islands of the same age. The 1- and 25-million-year (*m.y.*) lines indicate the subsidence expected for oceanic crust with these equivalent thermal ages.

as observed. The larger the island the longer it should stay at sea level.

Before leaving this subject it may be useful to estimate how rapidly waves can erode an island. Prince Edward Island lies in the Indian Ocean at 46°37′S latitude, in the "roaring forties" of nautical lore, where strong westerly gales prevail and great swells roll eternally from the west and southwest. The island is one of two rising from a broad volcanic platform a few hundred meters deep. Brief volcanism ending 215,000 years ago (*16*) built a broad gently sloping shield 550 m above present sea level. A detailed topographic survey (*17*) indicates minimal stream erosion, but the great swells truncated the western half of the island and produced a cliff 500 m high. Roughly 15,000 years ago, small volcanoes rebuilt part of the shelf above sea level, but the eroded volume out to the present 70-m depth contour can be reconstructed with more than the usual confidence. The average thickness of rock removed was 172 m in 215,000 years, and a shelf 2 to 3 km wide was cut. Like the Hawaiian Islands, the lee (eastern) half of Prince Edward Island has been virtually undissected. Meanwhile, cliffs 150 to 200 m high have already been cut into the small 75,000-year-old cones on the windward side. Without the latest volcanism, presumably the whole island would have been truncated in roughly 0.5 million years, leaving a bank about 8 km by 12 km.

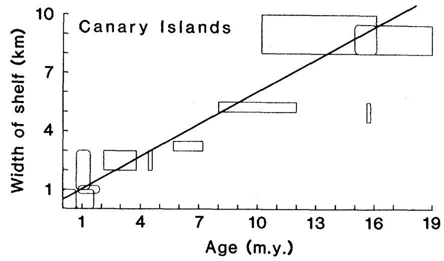


Fig. 6. Increase of shelf width with age (*11*) of coastal sectors of the Canary Islands. Size and shape of symbols indicate ranges of age or width; *m.y.*, million years.

Deep Terraces and Shape of Guyot Platforms

In some circumstances, thermal subsidence may be balanced by erosion-isostasy, but otherwise it may only be slowed. Thus terrains that differ from normal insular shelves may be eroded in the case of small islands on young crust or during the initial stage of rapid thermal subsidence of large islands before river erosion becomes widespread. A combination of subsidence and wave erosion may produce either a series of narrow terraces separated by cliffs or a relatively smooth shelf that is steeper near the shelf break than in the interior.

An ideal place to study the possibility of the predicted deep terraces would be Easter Island where rocks only 1 million years younger than the surrounding young crust are exposed (*18*). But it has not been dredged or surveyed appropriately to test for the presence of terraces. The only other small, isolated volcano known to have grown upward fast enough to become an island next to a ridge crest (*19*) is now the submerged Cobb bank on the west flank of the Juan de Fuca Ridge. It is 1.5 million years old (*20*), lies on 3-million-year-old crust, and thus was probably active on crust less than 1.5 million years old. It has submerged terraces at two general levels, 823 to 1189 m and 183 to 220 m, a broad summit platform at 82 m, and a central pinnacle rising to 34 m (*21*). From detailed soundings it appears that a simple volcanic cone has been notched and cliffed at the two deep levels and almost truncated on top. Rounded basalt cobbles, characteristic of beaches, support the interpretation that the terraces were formerly at sea level (*22*). Cobb bank thus has the morphology predicted for a very small island and a very young equivalent thermal age.

In order to predict morphology associated with midplate vulcanism it is necessary to have an independent measure of thermal rejuvenation. This can be obtained from a guyot by examining its relief, which indicates the water depth when it was truncated, and the subsidence rate. This information is available, for example, for Welker Guyot in the Gulf of Alaska (*23*). The age of the crust is 26 million years and of the volcano 15 million years (*24*). The guyot has a summit platform about 8 km in diameter—equivalent to an insular shelf 4 km wide. It was a small island and probably persisted, in these stormy seas, for less than 4 million years. The platform is at 940 m, and there is a narrow terrace at 1400 m from which the rounded cobbles of a beach deposit were dredged by H.W.M. The terrace is 2620 m above the volcanic basement, suggesting that the equivalent thermal age is close to zero, although the crests of ridges at this latitude are generally as shallow as 2200 m (*25*). The rate of subsidence is complicated by thick turbidite deposits, but like most small guyots on young crust, it apparently subsided as though it had an equivalent thermal age of zero (*4*).

The existence of the deep terrace would be expected under these circumstances as, indeed, would the observed morphology of the summit. Summit morphologies can be predicted for different equivalent thermal ages and erosion rates if only thermal subsidence and wave truncation are considered and isostasy is ignored. The summit surface that is cut is a mirror image of a subsidence curve. The summits of Welker, Miller, and Quinn guyots in the Gulf of Alaska are all known from detailed topographic surveys (*26*) and air-gun profiles. They are all small and have relief and subsidence records indicating very young equivalent thermal ages. Only Welker is known to have a deep terrace, but only a few deep dredge hauls have been made on other guyots. The model presented here, which includes isostasy, predicts that the slopes of the outer zones of the guyot summits should be controlled by rapid thermal subsidence. They, in fact, correspond to the predicted shape for an

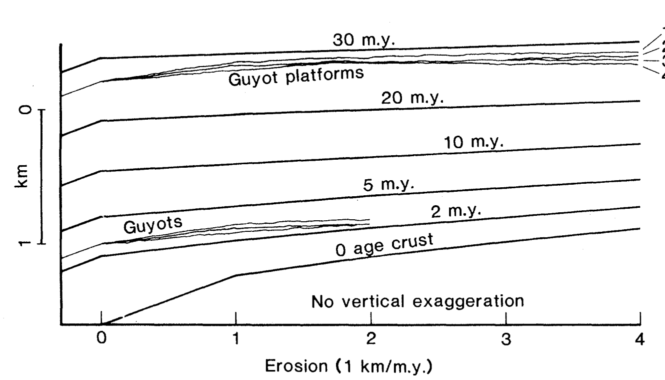


Fig. 7. Profiles of eroded basement summits of four small guyots in the Gulf of Alaska compared with theoretical shapes produced by shelf-widening at 1 km per million years (*m.y.*) and thermal subsidence at different equivalent thermal ages. The outer annulus of each summit can be modeled by subsidence of very young crust. The central region of each shelf is flatter, as expected if isostatic uplift was partially balancing thermal subsidence.

equivalent thermal age of 2 to 5 million years (Fig. 7) and wave erosion at 1 km per million years. The inner slopes however are much gentler (*27*) and indicate either that erosion was much faster or subsidence much slower.

The required change in rate of erosion has the opposite sign from that observed on modern islands and is thus improbable, as is slow, purely thermal subsidence, because it would be too slow for the actual age of the crust. Thus, no simple erosion-subsidence curve fits both inner and outer slopes. However the model predicts qualitatively that isostasy will increasingly balance subsidence as the island ages, and thus the inner truncated slopes should be gentle, as they are.

Insular Erosion by Rivers

The persistence of islands and particularly the duration of any transitional period when isostasy only partially balances thermal subsidence depends on the erosional history of islands—river erosion as well as wave erosion. The relative importance of these two erosion processes depends on the radius of the island because the rate of wave erosion varies with the perimeter and river erosion with the area (*7*).

Small islands tend to be eroded by waves, large ones by streams. This relation is reinforced by the influence of height on rainfall. High islands intersect the trade wind clouds in the tropics and cause orographic precipitation at any latitude. The consequences are obvious in the Hawaiian Islands. Consider, for example, Kohala volcano on the northern end of Hawaii. The 0.4-million-year-old volcano (*28*) is deeply dissected on the northeastern side where the trade winds deliver frequent rains, but it is hardly eroded on the lee side. A comparison of the present volume of the volcano with its reconstructed original form (*29*) indicates an average of 112 m of erosion to windward and only 2.4 m to leeward.

Although the results are approximate, similar studies indicate that small, low islands and the lee sides of large, young islands generally are eroded very slowly by streams. Meanwhile the windward sides of high islands are eroded relatively rapidly—perhaps 30 to 40 times as fast—for a few million years. After 3 million years, however, erosion of the lee side may begin to catch up, as is the case on Kauai. It appears that, once the windward side is eroded away, rainfall may be centered on the summit of the lee slopes, which are then cut by deep canyons. This is already occurring on Tahiti (*30*), which is only 1.2 million years old.

Origin and Subsidence of Midplate Swells

The study of existing islands indicates that the subsidence of ancient islands is more complex than has been surmised. Both the size of an island and the presence or absence of encircling reefs must be taken into account in calculating the history of subsidence. The best data for measuring subsidence rates during the critical initial stages appear to come from islands with barrier reefs and from small, young, reefless guyots such as those in the Gulf of Alaska. Both groups subside as though the crust had an equivalent thermal age of 0 to 3 million years, regardless of its actual age. Midplate swells on young crust have a depth similar to ridge crests, but those associated with isolated, older volcanoes have depths that appear to be proportional to the age of the crust (*4*). Large volcanoes and atolls on older crust appear to sink at rates expected of an equivalent thermal age of about 25 million years (*3*). Presumably the thermal subsidence rates for the larger guyots will be found to be faster after correcting for erosion and isostasy. The appropriate corrections, if any, for subsidence derived from atoll stratigraphy are more of an enigma. Unconformities (*31*) indicate intermittent uplift relative to sea level, which may reflect eustatic lowering, but gaps in the record do not affect the average rate of subsidence, merely the rate at shorter intervals.

Drill core and morphologic evidence need not be in conflict if the drill did not sample the barrier reef that existed when subsidence was rapid. The outer reefs of atolls are not vertical walls, and reefs do not appear to grow outward very much on the steep submarine slopes of volcanic islands. Thus the size of an atoll decreases as it sinks. Accordingly, it is possible that drilling on a modern island at the edge of an atoll would not penetrate the initial barrier reef without slant drilling. Indeed, the prevalence of lagoonal facies in atoll drilling (*32*) suggests that the oldest parts of the reef may not have been sampled.

With regard to the origin of midplate swells, the thermal rejuvenation hypothesis of Detrich and Crough (*3*) is again (*4*) supported. However, it is probable that the equivalent thermal age of rejuvenation in many places may be relatively young regardless of crustal age.

References and Notes

1. D. P. McKenzie, *J. Geophys. Res.* **72**, 6261 (1967).
2. T. H. Jordan, in *The Mantle Sample: Inclusions in Kimberlites and Other Volcanics*, F. R. Boyd and H. O. Meyer, Eds. (American Geophysical Union, Washington, D.C., 1979), pp. 1–14.
3. R. S. Detrick and S. T. Crough, *J. Geophys. Res.* **83**, 1236 (1978); S. T. Crough, *Geophys. J. R. Astron. Soc.* **55**, 451 (1978); *Tectonophysics* **61**, 321 (1979).
4. H. W. Menard and M. McNutt, *J. Geophys. Res.* **87**, 8570 (1982).
5. H. W. Menard, *Marine Geology of the Pacific* (McGraw-Hill, New York, 1964).
6. C. Darwin, *The Structure and Distribution of Coral Reefs* (John Murray, London, 1842).
7. W. M. Davis, *The Coral Reef Problem* (American Geographic Society, New York, 1928).
8. R. A. Duncan and I. McDougall, *J. Volcan. Geotherm. Res.* **1**, 197 (1976).
9. R. S. Dietz and H. W. Menard, *Bull. Am. Assoc. Petrol. Geol.* **35**, 1994 (1951).
10. M. McNutt and H. W. Menard, *J. Geophys. Res.* **83**, 1206 (1978); K. Lambeck, *Geophys. J. R. Astron. Soc.* **67**, 91 (1981).
11. A. Abdel-Monem, N. D. Watkins, P. W. Gast, *Am. J. Sci.* **271**, 490 (1971); *ibid.* **272**, 805 (1972).
12. P. Vogt, personal communication.
13. J. Tuzo Wilson, *Nature (London)* **207**, 907 (1965).
14. R. S. Detrick, J. G. Sclater, J. Thiede, *Earth Planet. Sci. Lett.* **34**, 185 (1977).
15. M. Talwani and G. Udintsev, *Init. Rept. Deep-Sea Drill. Proj.* **38**, 1213 (1976).
16. I. McDougall, in *Marion and Prince Edward Islands, Report on the South African Biological and Geological Expedition 1965–1966*, E. M. van Zinderen Bakker, J. M. Winterbottom, R. A. Dyer, Eds. (Balkema, Cape Town, 1971), chapter 5.
17. O. Langenegger and W. J. Verwoerd, in *ibid.*, chapter 2.
18. J. G. Clark and J. R. Dymond, *J. Volcan. Geotherm. Res.* **2**, 19 (1977).
19. H. W. Menard, *J. Geophys. Res.* **74**, 4827 (1969).
20. J. R. Dymond, N. D. Watkins, Y. R. Nayudu, *ibid.* **73**, 3977 (1968).
21. T. F. Budinger, *Deep-Sea Res.* **14**, 191 (1967).
22. Y. R. Nayudu [*Geophys. Monogr. Am. Geophys. Union* **6**, 171 (1962)] offers an alternative explanation.
23. H. W. Menard and R. S. Dietz, *Bull. Geol. Soc. Am.* **62**, 239 (1951).
24. B. Dalrymple, personal communication.
25. H. W. Menard and L. M. Dorman, *J. Geophys. Res.* **82**, 5329 (1977).
26. H. W. Menard, *Bull. Geol. Soc. Am.* **66**, 1149 (1955); N. C. Smoot, *Mar. Geol.* **43**, 1487 (1981).
27. H. H. Hess, *Am. J. Sci.* **244**, 772 (1946).
28. I. McDougall, *Earth Planet. Sci. Lett.* **46**, 31 (1979).
29. C. K. Wentworth, *J. Geol.* **35**, 117 (1927).
30. H. Williams, *Bernice P. Bishop Mus. Bull. 105* (1933).
31. H. S. Ladd and J. I. Tracey, Jr., *Deep-Sea Res.* **4**, 218 (1957).
32. H. S. Ladd and S. O. Schlanger, *U.S. Geol. Surv. Prof. Pap. 260-Y* (1960), p. 863.
33. B. Keating, D. Mattey, J. Naughton, D. Epp, C. E. Helsley, *Eos* **62**, 381 (1981).
34. C. M. Emerick and R. A. Duncan, *Earth Planet. Sci. Lett.*, in press.
35. R. Brousse, J.-C. Philippet, G. Guille, H. Bellon, *C. R. Acad. Sci.* **274**, 1995 (1972).
36. B. Parsons and J. G. Sclater, *Geophys. J. R. Astron. Soc.* **82**, 804 (1977).

37. I am indebted to L. M. Dorman, T. H. Jordan, M. K. McNutt, and R. P. Sharp for critical comments on the manuscript and to C. M. Emerick and R. A. Duncan for allowing me to read a paper in advance of publication.