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Analysis of Partially Emerged Corals and Reef Terraces in the Central Vanuatu Arc: Comparison of Contemporary Coseismic and Nonseismic With Quaternary Vertical Movements

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In the central Vanuatu arc, living and recently deceased reef corals act as natural tide gauges which have allowed us to map vertical tectonic deformation patterns. As corals grow, the density of the aragonite coral skeletons varies on an annual cycle, producing annual growth bands similar to tree rings. Using coral growth bands, we can determine the year coral surfaces died due to emergence. We interpret four major coral emergence events as coseismic uplifts that occurred near the epicenters and times of large shallow earthquakes on January 5, 1946 (M_S = 7.3), August 11, 1965 (M_S = 7.5), October 27, 1971 (M_S = 7.1) and December 29, 1973 (M_S = 7.5). The 1965 and 1973 events caused maximum uplifts of 120 and 60 cm, respectively, in the frontal arc. Also related to these events are uplifts of 10 cm and 6 cm in the back arc on Pentecost and Maewo islands, which lie east of the volcanic chain and the primary forearc zones of uplift and subsidence. Similar secondary zones of uplift occurred with the great 1960 Chile and 1964 Alaska earthquakes. The amplitude of these secondary uplifts is significantly larger than that predicted by models having a single fault in an elastic half-space. However, the amount of secondary uplift is comparable to that predicted if the fault occurs in a plate of constant thickness overlying a viscoelastic half-space. At various places in 1957, 1969-1970, 1977, and 1978-1981 there was about 5-10 cm of emergence not associated with major earthquakes, which may indicate nonseismic tectonic uplift. However, oceanographically lowered sea levels, as in El Niños, may have determined the times when corals died and recorded these events. Nevertheless, the accumulation of emergence, its persistence, the limited geographic extent of each event, and occurrence in areas of rapid Holocene uplift suggest that the causes of the uplifts are tectonic. These events suggest that in some areas a third or more of the total accumulated uplift in central Vanuatu takes place as aseismic motion. However, in some areas we find only coseismic emergence. In central Vanuatu, contemporary coseismic vertical deformation, Holocene uplift, and topography have remarkably similar patterns. This suggests that the mechanisms and processes causing vertical deformation have varied little over the last 10^6 years. Apparently, the topography, structure, and seismotectonics are controlled by the subduction of the d'Entrecasteaux ridge, a major bathymetric feature underthrusting this part of the arc. The influence of this ridge may have been especially extensive because it migrates very slowly along the arc trend, and thus it interacts for a long time with a single portion of the arc system. Our previous studies of reef terraces indicated the existence of at least four seismotectonic arc segments or blocks along the Santo-Malekula interval of the arc, and our present results further support this conclusion. Each block has uplifted at different times, by different amounts, at different rates, and tilted in a different direction. Boundaries between the north Santo and the south Santo segment and between the north Malekula and the south Malekula segment correlate with the north and south flanks of the d'Entrecasteaux ridge, as does the absence of a physiographic trench west of Santo.

INTRODUCTION

Vertical deformation is the most visible expression of the tectonic processes acting at convergent plate boundaries such as the Vanuatu (formerly New Hebrides) arc (Figure 1). Over long periods these motions provide the topographic and bathymetric relief characteristic of convergent margins, and over short periods these motions are dramatically evident as uplift or subsidence associated with individual large earthquakes. For example, careful measurements of the pattern of vertical deformation for the two largest earthquakes in this century, the great 1960 Chilean and the 1964 Alaskan earthquake, found vertical motions of up to 10 m (Figure 2) extending over thousands of square kilometers [*Plafker and Savage*, 1970; *Plafker*, 1972]. Similar studies in Japan [e.g, *Fitch and Scholz*, 1971; *Yonekura*, 1975; *Matsuda et al.*, 1978] also have important implications for earthquake recurrence intervals and prediction.

Subduction, seismicity, seismic and nonseismic vertical movements, and tectonic evolution of arc systems are all closely related. However, a detailed history of vertical deformation on a variety of time scales is difficult to acquire. Generally, even for relatively large earthquakes, seismologists can only make rather crude estimates of the spatial extent of the earthquake rupture

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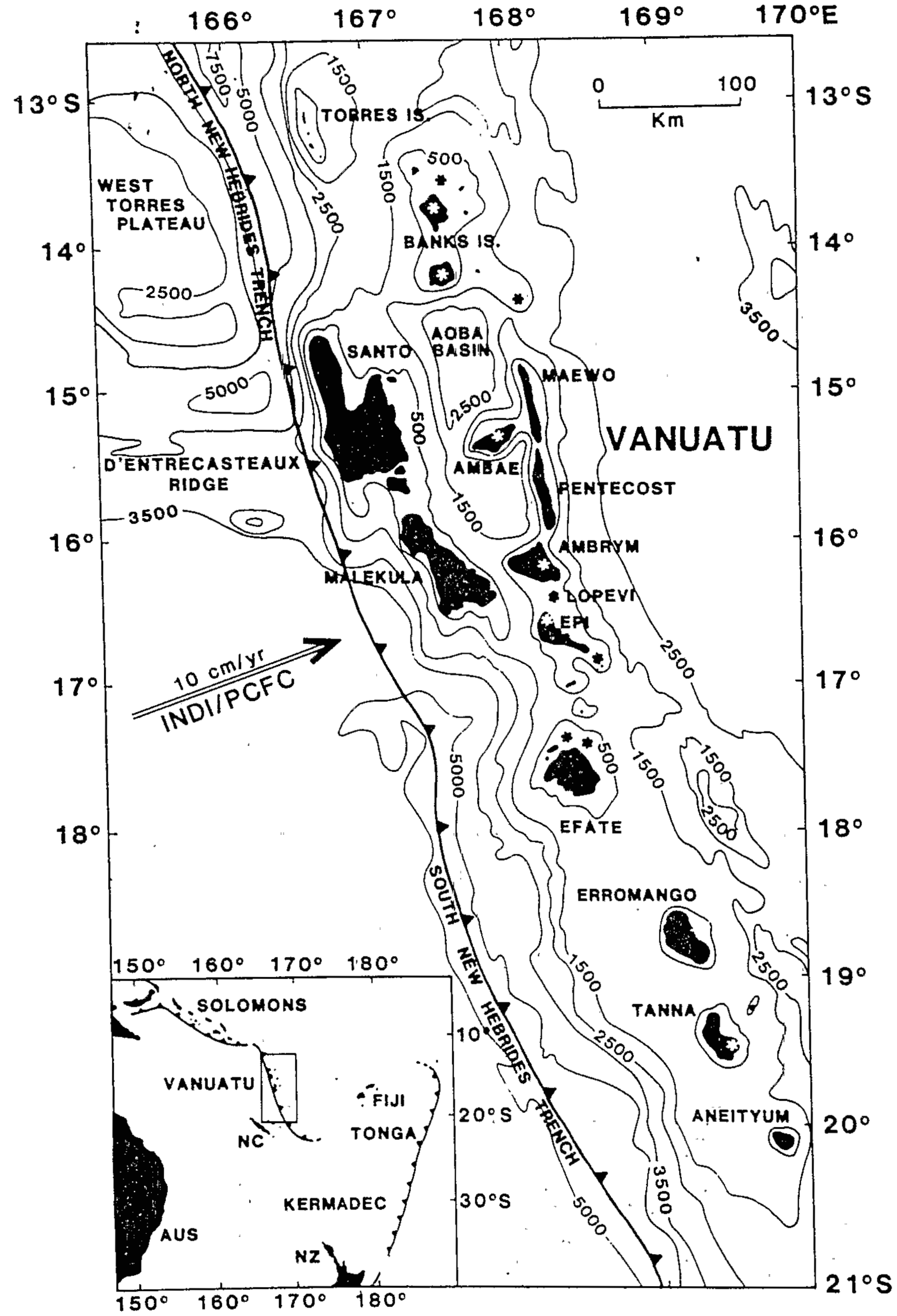


Fig. 1. Bathymetry of the Vanuatu (New Hebrides) region. Volcanoes are indicated by asterisks. Note that where the d'Entrecasteaux ridge (DR) intersects the arc, the physiographic trench is absent west of the nonvolcanic islands of Santo and Malekula. The nonvolcanic back arc islands of Maewo and Pentecost lie east of the volcanic chain. The contours in this and all subsequent figures are in meters and are taken from *Kroenke et al.* [1983].

process based on the directivity of seismic waves and the earthquake moment [e.g., *Beck and Ruff*, 1985]. Furthermore, it is possible that a substantial fraction of the fault motion occurs as interseismic creep, as has been observed along strike-slip margins [e.g., *Schulz et al.*, 1982]. Leveling data from Japan, and Alaska to a lesser extent, are of sufficient quality to compare coseismic, interseismic, and longer term net vertical deformation patterns [e.g., *Fitch and Scholz*, 1971; *Plafker*, 1972; *Thatcher*, 1984*a*, *b*]. However, comparable data for other arc systems are of interest because the mechanics of the subduction process may depend on numerous factors such as convergence rate, the topography and age of the subducted lithosphere, the degree of coupling between plates [e.g., *Uyeda and Kanamori*, 1979; *Uyeda*, 1982], and the amount of sediment in the trench [*Cloos*, 1982; *Shreve and Cloos*, 1986].

For several reasons, the Vanuatu arc is ideal for study of the relationships between vertical tectonism, plate convergence, and earthquake activity. Emerged coral reef terraces on many islands provide a record of late Quaternary uplift, tilting, and faulting [*Neef and Veeh*, 1977; *Taylor et al.*, 1980, 1981, 1985; *Jouannic et al.*, 1980, 1982; *Gilpin*, 1982]. Local seismograph and tiltmeter networks operated by Cornell University and the Office de la Recherche Scientifique et Technique Outre-Mer (ORSTOM) [*Marthelot et al.*, 1980, 1985] offer a detailed view of ongoing plate interaction to compare with the coral record of vertical tectonism.

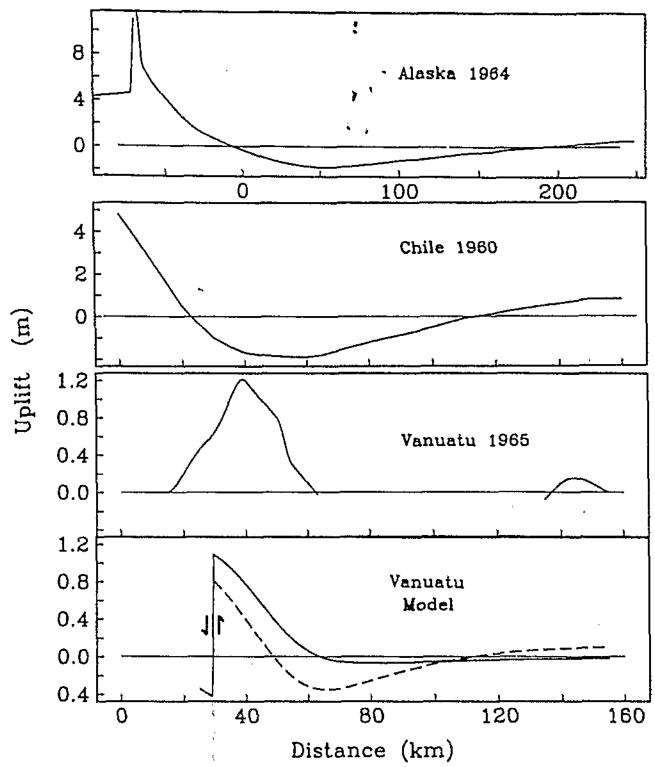


Fig. 2. Uplift profiles observed for the Vanuatu earthquake of August 11, 1965 (M_S = 7.5), the Chile earthquake of May 22, 1960 (M_W = 9.5), and the Alaska earthquake of March 28, 1964 (M_W = 9.2). The data for the Alaska and Chile events are from *Plafker* [1972] and *Plafker and Savage* [1970]. All profiles trend approximately perpendicular to the arc. Note that all three earthquakes have substantial uplift in the back arc, away from the main uplift zone. For comparison, at the bottom we show the uplift predicted with a two-dimensional model having slip on a fault in an elastic half-space (solid line), and for slip in an elastic plate overlying a viscoelastic half-space (dashed line). For the modeling, we use the expressions of *Savage and Gu* [1985]. Following *Ebel* [1980], we assume a fault with dip 50° rupturing to a depth of 25 km. Vertical exaggeration makes the 50° fault appear vertical. The slip is 2 m, the plate thickness is 30 km, the Young's modulus is 70 GPa, and the density is 3.8 g cm^3.

Quaternary uplift has been particularly rapid in central Vanuatu where the d'Entrecasteaux ridge (DR) underthrusts the arc (Figures 1 and 3). Coral limestone covers parts of most islands to elevations of tens or hundreds of meters and shows that both fore arc and back arc islands have uplifted (Figure 3). Living and recently emerged corals demonstrate that vertical deformation has continued up to the present. Indeed, we shall show that vertical deformation took place near the times of four large shallow earthquakes occurring in January 1946, August 1965, October 1971, and December 1973 (Figure 4). The events of 1965, 1971, and 1973 have the highest seismic moments of all earthquakes occurring in central Vanuatu since 1964 [*Ebel*, 1980; *Chinn and Isacks*, 1983].

One of the primary objectives of the present paper is to present all the available information concerning the amount and extent of the vertical motions associated with these events, as determined from corals in the epicentral regions. Very few of these data have been published previously. Uplift of the frontal arc had been reported in association with the August 11, 1965, earthquake, but no precise measurements had been made [*Mitchell*, 1968; *Benoit and Dubois*, 1971]. *Taylor et al.* [1980] used emerged corals to precisely measure and map the 1965 uplift on north Malekula. However, additional field observations now demonstrate that the 1965 uplift area is larger than previously reported. As with the great 1960 Chile and 1964 Alaska earthquakes [*Plafker and Savage*, 1970; *Plafker*, 1972], the 1965 and December 1973 events have a secondary zone of uplift occurring approximately 100 km east of the primary uplift zone, and somewhat east of the zone of coseismic subsidence (Figure 2). In addition, this paper reports several additional coral emergence events which may be caused by nonseismic tectonic processes or by sea level changes associated with oceanographic processes, such as the El Niño/Southern Oscillation (henceforth, El Niño). In this paper we shall use the adjective contemporary to describe any of these events which have occurred in the twentieth century in contrast with similar events which may have occurred throughout the Holocene.

A second important objective of this paper is to describe in detail how we use corals to measure the extent and amount of vertical motion. A comprehensive presentation of these methods, developed primarily by the first author while studying the effects of the 1965 earthquake [*Taylor et al.* , 1979; 1980], has not appeared previously [e.g., *Taylor et al.*, 1980, 1981; 1985; *Buskirk et al.*, 1981]. Because a substantial fraction of all subduction zones occur in tropical regions, our approach will be useful for measuring vertical motions in regions other than Vanuatu. These relatively inexpensive methods can indicate the best places to measure future vertical deformation using strainmeters, tiltmeters, the Global Positioning System, or other geodetic methods and provide a long-term vertical deformation history for comparison with patterns and rates of short-term movements.

Discovery, mapping, and measurement of previously unrecognized contemporary deformation patterns is possible because living corals act as tide gauges that record the amount and time of occurrence of vertical tectonism. Where coseismic uplift is small, only the upper part of a coral head is killed by emergence. Because some coral heads have annual growth bands more-or-less analogous to tree rings, they contain an internal record of when the emerged part of a coral head died.

There are several advantages of using corals rather than mechanical tide gauges to measure and date relative sea level changes. Because corals exist along nearly every interval of tropical coast, one can reconstruct the local pattern of relative sea level change. This helps discriminate between tectonic motions and sea level changes caused by oceanographic effects, which usually occur over much larger geographic areas than tectonic processes. Tide gauges are immovable, expensive, and difficult to operate in large numbers. When there are data from only a single location, it is often impossible to deduce the cause for an observed sea level change, whether it is recorded by a tide gauge or a coral head. For example, in Japan a tectonic uplift inferred from tide gauge data has been shown to be caused by steric sea level changes [e.g., *White et al.*, 1979]. A major advantage of the coral method is that a record of past events can be recovered, whereas man-made instruments can only begin to record after they are emplaced.

The final objective of this paper is to compare the coseismic vertical deformation with the total accumulated vertical deformation in Vanuatu recorded by isotopically dated Holocene and Pleistocene coral reefs, which have average uplift rates as great as 0.6-0.7 cm yr^{-1} on southern Santo [*Jouannic et al.*, 1980, 1982; *Gilpin*, 1982; *Urmos*, 1985; *Taylor et al.*, 1985]. From this comparison it is obvious that the contemporary events that we document need to continue for only a few hundred thousand years to influence profoundly the structure and morphology of the arc. Indeed, we suggest that these processes have already operated for several hundred thousand years and that they are responsible for many of the anomalous topographic features of the central Vanuatu

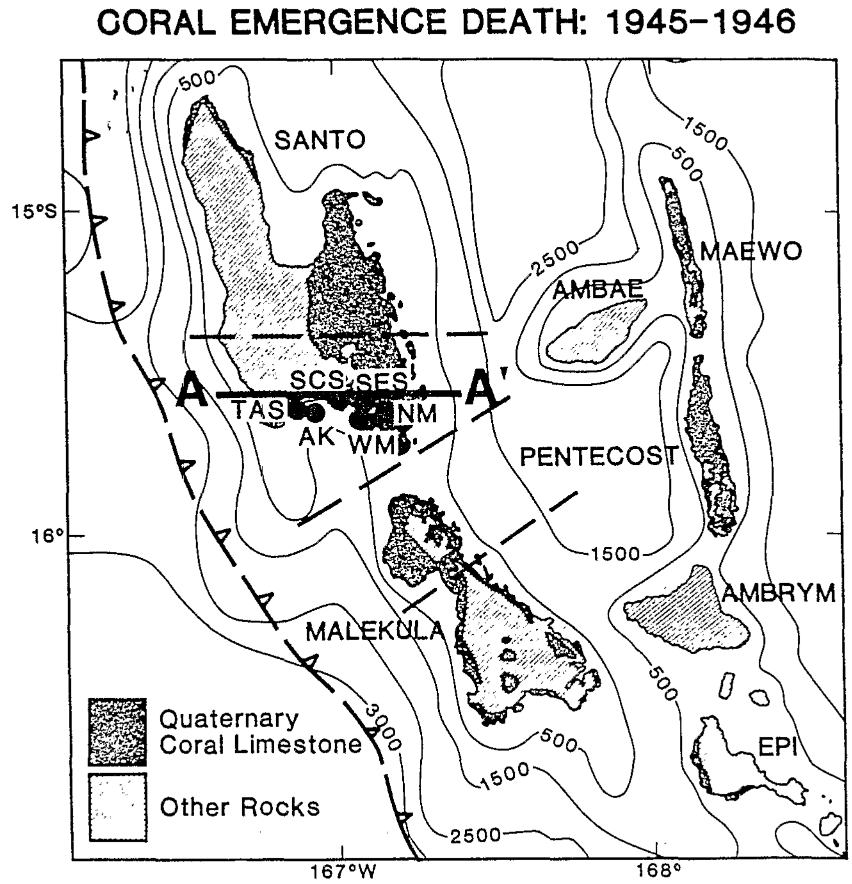


Fig. 3. Distribution of Quaternary coral limestone and central Vanuatu bathymetry. Other rocks include all ages and lithologies that are not coral limestone. Dashed lines indicate previously identified seismotectonic block boundaries, identified using reef terraces, seismology, and other geological indicators [*Taylor et al.*, 1980]. The presence of emerged coral reefs demonstrates Quaternary emergence of most islands. However, even though it has emerged tectonically, western Santo generally lacks coral limestone because large amounts of sediment shed from the high mountains inhibit coral reef growth. Where reefs may have grown on western Santo, they would have uplifted above sea level too rapidly for a significant thickness of coral limestone to accumulate. The solid circles are localities where corals show evidence of emergence death in 1945-1946. The event is dated by coral growth bands at localities AK and SES. Elsewhere, the time of this event was determined from coral head morphology by assuming that corals grow about 1 cm yr^{-1} (Table 2). A-A' is the location for a profile of the 1945-1946 uplift (Figure 19).

arc in the vicinity of Santo and Malekula islands (Figure 3). The individual coseismic movements that we document from corals are the small incremental vertical changes that sum to create the present morphologic and structural character of the Vanuatu arc. These movements are not superficial oscillations superimposed on preexisting arc structure. Instead, they represent the long-term process of tectonic evolution of the arc.

METHODS

Living Corals as Recorders of Relative Sea level Change

Scoffin and Stoddart [1978] and *Stoddart and Scoffin* [1979] realized that coral colonies living in very shallow water adjust their level of growth to changes in relative sea level. They also noted that terraced coral colonies indicate emergence and that upward growth of a colony where upward growth had formerly been limited must indicate submergence. Based on these principles, it is possible to predict the morphologies of emerged coral heads such as we have found in Vanuatu. However, no previous studies other than our own preliminary results [*Taylor et al.*, 1980, 1981; *Buskirk et al.*, 1981] have used corals to document the timing and amount of vertical tectonic movements.

Shallow-living coral heads behave as natural tide gauges whose surface morphology, and annual density bands will record the timing of even a few centimeters of persistent relative sea level change (Figure 5). At successively higher levels in the intertidal zone, the duration and intensity of exposure of corals to sunlight and air during low tides increase. While some coral species thrive in an intertidal environment, there is always a maximum level above which they cannot survive and grow. We call this the highest level of survival (HLS). Corals living beneath the HLS grow both horizontally and upward until the tops of the coral heads intersect the HLS. Subsequently, their tops die, and they are limited to horizontal growth.

Emergence of a coral above the HLS will kill only the emerged portion, while the coral surface below the HLS continues to live and grow (Figures 5*a* and 6). If partial emergence persists for several years or more, continued growth below the HLS causes miniature terraces to form on the coral surface. Successive emergence events may be recorded as a series of small terraces on the surface of a coral head.

Corals can record a rise of the HLS by growing upward toward the risen HLS (Figures 5*b* and 6*c*). However, because coral heads typically grow about 1 cm yr^{-1}, a coral may require many years to reveal the total amount of submergence, whereas it may die down to a lower HLS and record emergence after a few

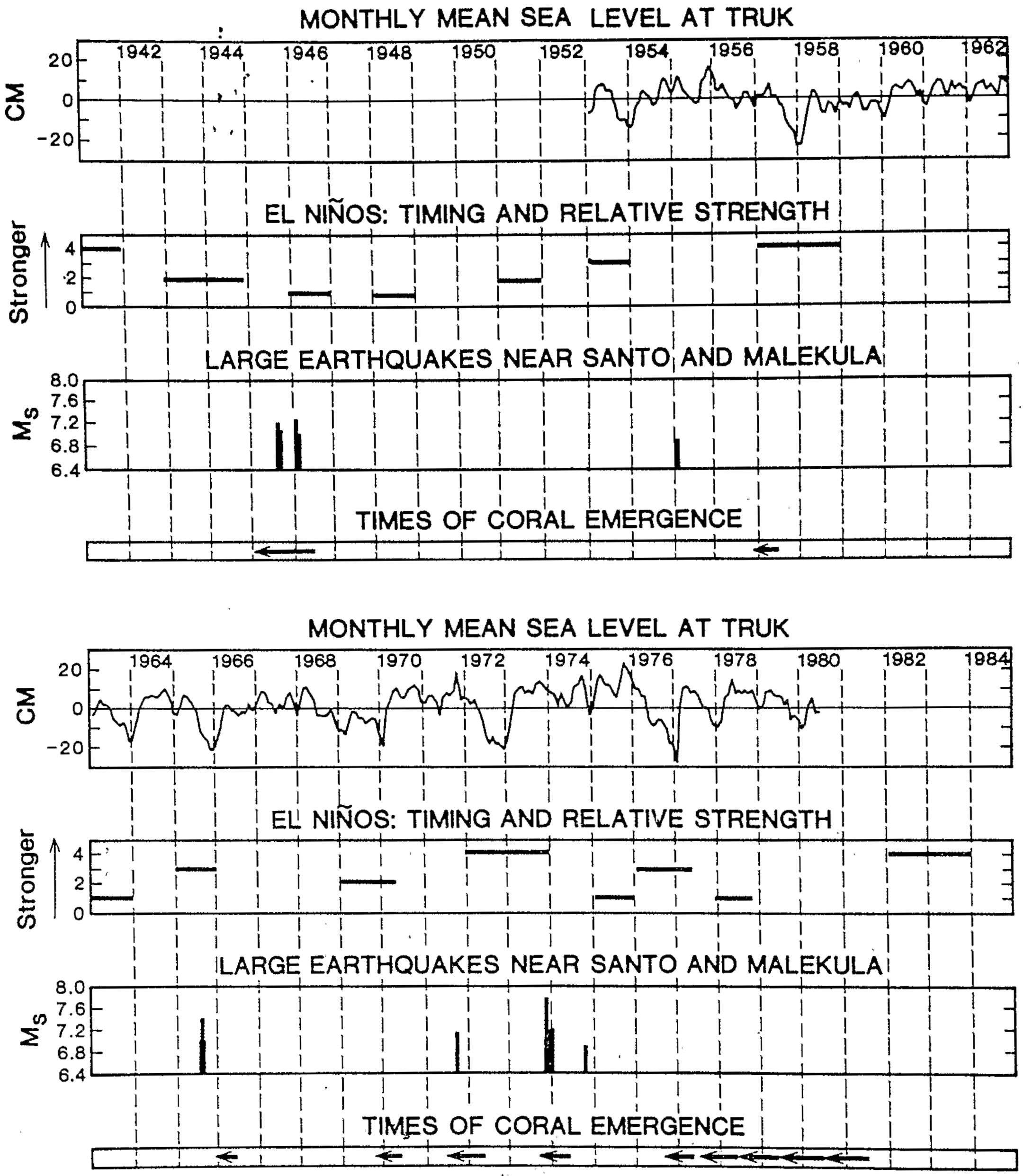


Fig. 4. Comparison of oceanographic events in the western Pacific, large earthquakes in central Vanuatu, and times of coral emergence death. The earthquakes are from *Gutenberg and Richter* [1954] and *Habermann* [1984], and the times of coral emergence are from this study. The times and relative strengths of El Niños are from *Quinn et al.* [1978]. The monthly mean sea level record at Truk is after *Meyers* [1982] and was chosen because of its length and the apparent close relationship between the sea level lows and El Niño events. Similar sea level changes may accompany El Niño events in Vanuatu, but there is no long tide record to show if the relationship is clear there as well. A similar relationship between El Niños and sea level may have occurred near Vanuatu, although the timing and amplitude may be slightly different than at Truk. The arrows showing times of coral emergence death point left to indicate that emergence causing coral death may have occurred a year or more before exposure finally killed the living coral surface. Note that coral emergence death occurred soon after the earthquakes of January 1946, August 1965, October 1971, and December 1973. The 1955 earthquakes were of relatively smaller magnitude, and no associated emergence is known. Emergence events not clearly associated with large earthquakes are usually 10 cm or less and are often associated with the times of El Niños.

days. A few years time is sufficient to indicate that the HLS has risen because the living coral surface will grow upward and often will overgrow the dead coral surfaces that are higher than the former HLS.

We determine the timing of vertical changes of HLS in coral heads by counting density growth bands that form the coral skeleton (Figure 7). Analogous to rings in trees, the density of the aragonite added to a living coral surface varies seasonally [*Knutson et al.*, 1972]. In Vanuatu, *Goniastrea retiformis* adds a high-density (HD) aragonite layer during summer and a lower-density (LD) layer in winter [*Buskirk et al.*, 1981; *Taylor et al.*, 1981]. These density variations are observed by sawing a slice of coral about 0.5 cm thick parallel to the growth direction. X radiography of the slab (Figure 7) reveals the density variations described in more detail elsewhere [e.g., *Buddemeier* , 1978; *Dodge*, 1980; *Taylor et al.*, 1981]. The year in which each dead surface died is determined by counting how many annual HD-LD band pairs lie between the living and dead coral surfaces.

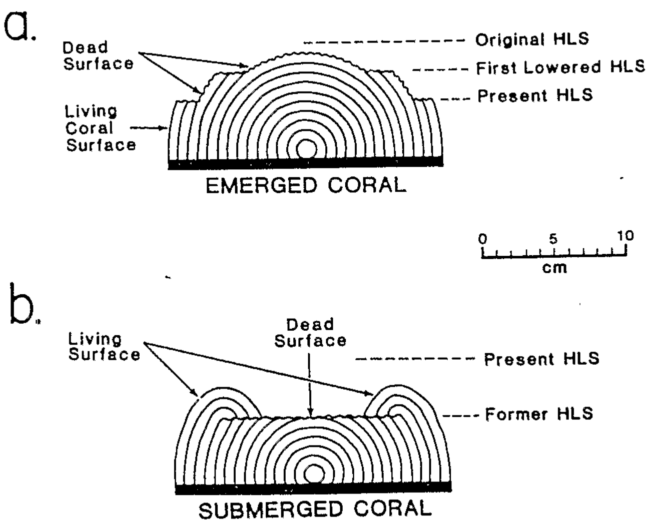


Fig. 5. Schematic cross sections of coral heads showing their growth in response to (*a*) falling or (*b*) rising sea levels. The roughly parallel lines represent annual skeletal growth increments. The highest level of survival (HLS) is the maximum height at which the coral can live and grow. Horizontal growth at the HLS results in a dead upper coral surface marking the path of the HLS through time. (*a*) This coral formerly lived deeper than the HLS. Subsequently, two emergence events lowered the HLS. Continued coral growth at about 1 cm yr^{-1} below the HLS followed each emergence event and results in a terraced morphology on the coral head. (*b*) This coral originally colonized a shallow substrate and grew upward until its upper surface reached the (former) HLS. Upward growth ceased at the HLS and the top surface died, but the sides continued growing outward. When the HLS rose due to a relative sea level rise, the living coral on the sides began to grow upward at about 1 cm yr^{-1} toward the present, higher HLS. Upward growth will continue until living coral reaches the present HLS. *Scoffin and Stoddart* [1978] present additional scenarios for water level control of coral growth.

How Reliable Are Corals as Tide Gauges?

Before one can use corals for tectonic studies there are several obvious questions to consider:

1. How accurately do corals record the amount of relative sea level change?
2. How small a relative sea level change can corals record?
3. How accurately do corals record the time of a relative sea level change?
4. How long must a relative sea level change persist before corals will record it?

One problem in assessing the accuracy of the coral method is that in Vanuatu there are no independent measurements of vertical deformation. Tide gauge, leveling, and tiltmeter measurements [*Marthelot et al.*, 1980; *Bevis and Isacks*, 1981; *Isacks et al.* , 1981] began after most of the events recorded by the corals. Moreover, the instrumental records represent relatively small parts of coastal areas where block faulting is common.

However, the corals themselves offer some insights into their precision as recorders of reative sea level changes. The upper surfaces of corals in Vanuatu and elsewhere [e.g., *Scoffin and Stoddard*, 1978] may form flat surfaces after several years of growth, indicating that the HLS has remained the same year after year. This observation suggests that the HLS is well defined and persistent under conditions of stable relative sea level. We also can estimate the precision of corals as recorders of relative sea level changes by comparing them with each other. For branching *Acropora* species, *Taylor et al.* [1980] found an average standard deviation of ±10 cm for suites of corals emerged 50 cm or more. For 12 coral heads at a locality on south Santo where coral growth bands show the mean 1965-1966 emergence to be 23 cm, the standard deviation was ±6 cm. Where emergence is less than 10 cm one can probably only conclude that there has been roughly a few centimeters of emergence.

At least two types of nontectonic phenomena cause changes of the HLS of corals. The first of these is annual cyclic tidal variation. The second is change in sea level caused by oceanographic/climatic phenomena (Figure 4) such as El Niño, which occurs at intervals of 2-10 years with an average recurrence interval of about 4 years [*Cane*, 1983].

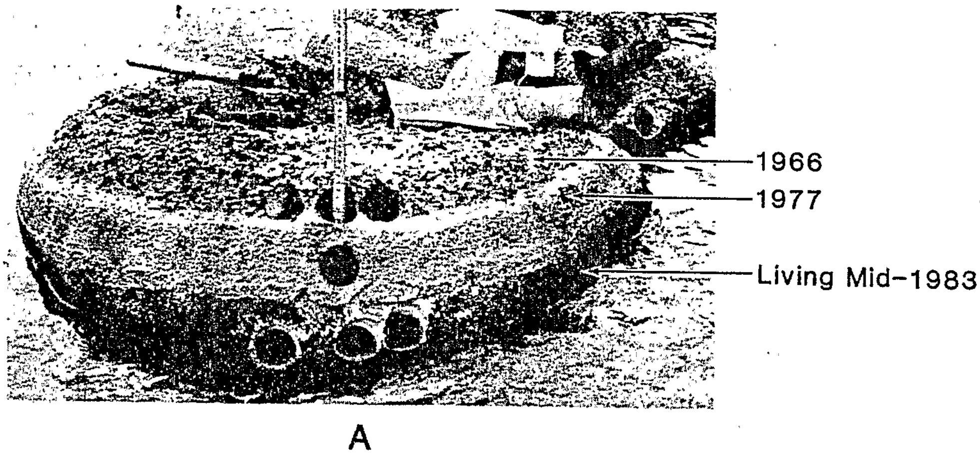
Vanuatu has semidiurnal mixed tidal cycles, each day having a period of high-high water (HHW), low-high water (LHW), high-low water (HLW), and low-low water (LLW). In the summer, LLW occurs at night and HLW during daytime, and then after the autumnal equinox the timing reverses [*Hydrographer of the Navy*, 1977]. The average difference between LLW and HLW is about 30 cm, with a maximum of nearly 40 cm when spring tides occur near solstices. In *Goniastrea retiformis*, we discovered heliotropism [*Buskirk et al.*, 1981], or orientation with respect to the sun angle, and showed that the HLS is probably controlled by exposure to direct sunlight. Therefore *G. retiformis* colonies are exposed to harmful amounts of direct sunlight only in midwinter when the exceptionally low LLW spring tides occur near noontime. If corals were at higher levels, they could live during most of the year. However, exposure of a colony is greatest during a few winter days of very low tides, and this controls the HLS. Because corals grow very slowly, the lowest HLSs of each year or every few years control the highest levels of living coral.

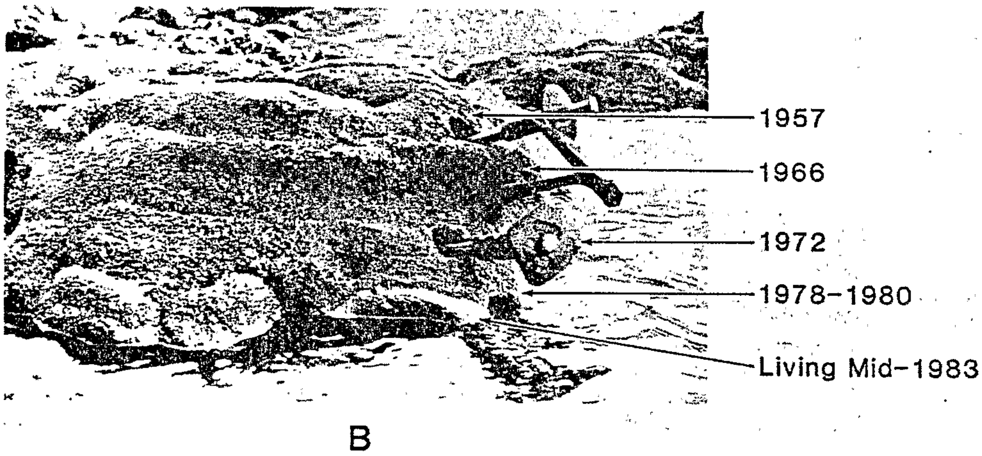
Relative sea level depressions smaller than about 30 cm must persist through the midwinter period of lowest tides to lower the HLS and affect coral growth. A temporary 30 cm fall of sea level probably would not affect coral growth if it occurred during summertime because the coral would be excessively exposed above water level only at night. Also, if a coral emerges 30 cm in the spring season, the coral surface will not begin to die until the following winter when LLW begins to occur near midday. If the sky happens to be cloudy during times of the LLW spring tides, then the damage to exposed coral surfaces may be minimal and coral death may be delayed about a year until the next period of vulnerability. Nearly all emerged corals that we sampled died after the deposition of the summer HD band and before the deposition of the winter LD band was complete.

How Can We Distinguish Tectonic From Other Causes for Relative Sea level Change?

As with tectonic emergence, to lower the HLS, nontectonic emergence must persist through the winter months, causing exposure of living coral to sunlight during the lowest low tides of the year. Regional tidegauge networks demonstrate that nontectonic sea level changes are common. Some of the largest sea level changes are associated with times of El Niño [e.g., *Wyrtki*, 1977, 1979, 1985; *Meyers*, 1982]. Sea level anomalies of tens of centimeters, persisting for months have been recorded (Figure 4). In locations where the uplift rate is low, these oceanographically related sea level variations cause fluctuations in the HLS which strongly influence the time when emerged coral surfaces die.

When either tectonic or oceanographic changes in HLS cause emergence of living corals, the occurrence and timing of coral death usually depends markedly on both the uplift rate and the rate





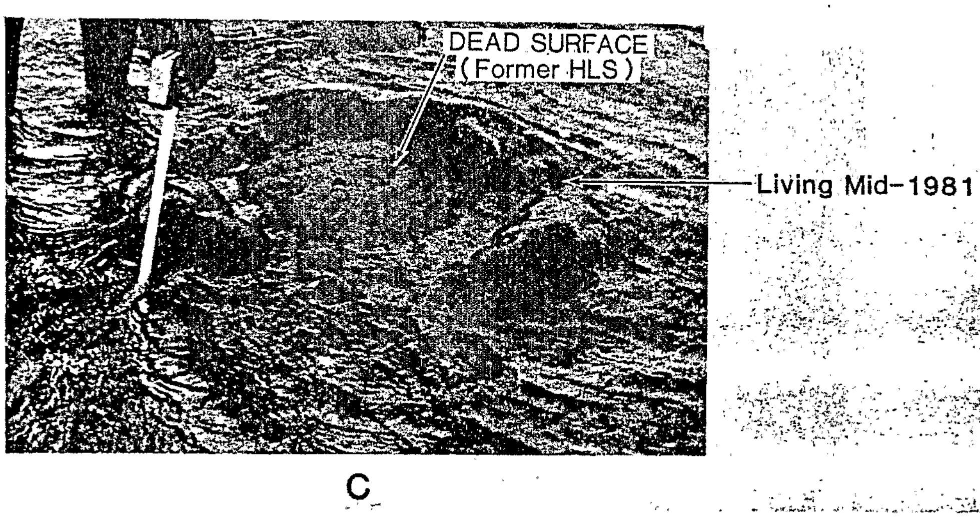


Fig. 6. (*a*) *Goniastrea retiformis* coral head from the west coast of Pentecost records two emergence events of about 10 cm each. The first event caused the upper horizontal band of coral to die in mid-1966, and the second caused the intermediate band of coral to die in mid-1977. The darker lower band of coral was living in this August 1983 photograph. The holes are the locations of the cores shown in Figure 7. (*b*) *G. retiformis* coral head from southwest Malekula records four coral emergence events: (1) 1957, (2) 1966, (3) 1972-1973, and (4) 1978-1980. The lowermost fringe of coral was still living when this colony was sampled and photographed in September 1983. (*c*) *G. retiformis* coral head at the western end of Ambae in the middle of the Ambae basin. This colony indicates recent submergence as illustrated in Figure 5 *b* in contrast to corals from islands on the margins of the basin that usually are like those above and in Figure 5 *a*. We found no emerged reefs or partially emerged coral colonies on Ambae.

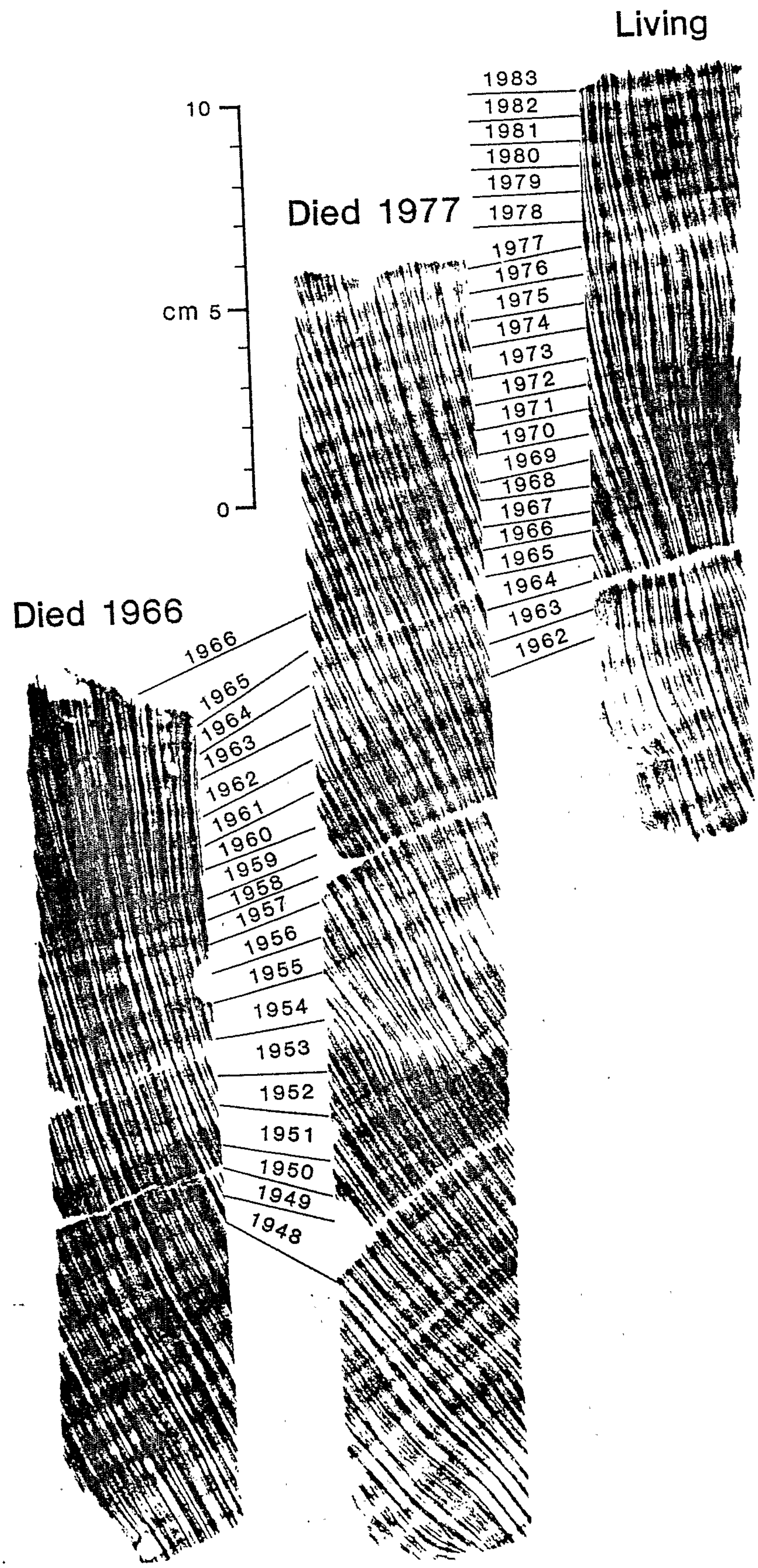


Fig. 7. X radiograph positive prints of samples drilled from the living surface, and two dead surfaces of the coral in Figure 6 *a* on Pentecost. The dark bands represent denser parts of the coral skeleton which form during summers [*Buskirk et al.*, 1981]. Lines between cores mark the approximate beginning of growth for each year. By matching the three cores we can count annual growth increments and specify the time of emergence death for each of the two dead emerged coral surfaces. The cores are easy to match because we know their spatial relationships in the coral and because the band sequences are similar from core to core.

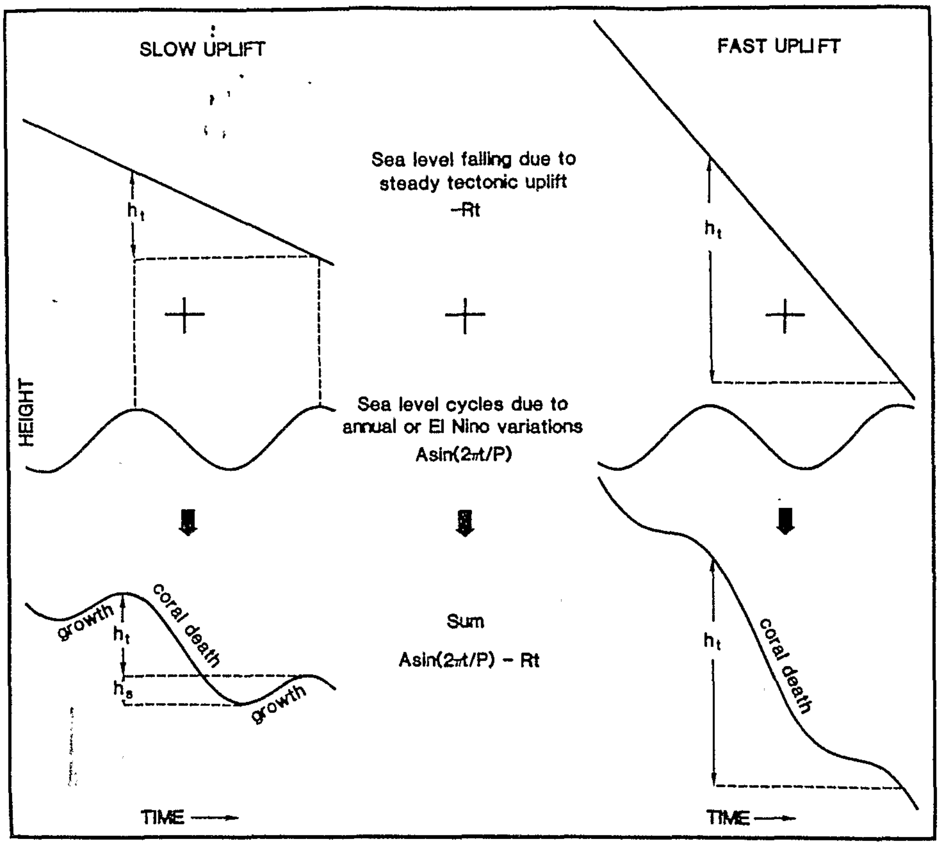


Fig. 8. Modeling of the effects on coral growth and death of (*a* slow and (*b*) fast uplift superimposed on fluctuating sea level, which is represented as a sine wave. (a) If the uplift rate is similar to or less than the amplitude and rate of cylic variation in sea level, coral surfaces will die down during cyclic falls of sea level, even though uplift is continuous. Thus coincidence of coral death and El Niño sea level lows does not preclude tectonic uplift as the underlying cause of coral emergence death. Even while uplift occurs, corals may grow upward (<1 cm yr^{-1}) during portions of the cycle where the cyclic sea level rise rate equals the uplift rate. Thus, the maximum possible vertical extent of coral emergence death ($h_s + H_t$) may exceed the tectonic uplift h_t. (*b*) If the uplift rate is greater than the amplitude and rate of cyclic sea level change the surface of the coral will die downward continuously, even when sea level is rising.

and amplitude of cyclic sea level variations. In particular, if uplift occurs with rate R cm yr^{-1}, and if there are periodic cyclic sea level variations with a peak to peak amplitude of $2A$ cm and period P yr, then the relative sea level (RSL) at time t is

$$\mathrm{RSL} = A \sin(2\pi t/P) - Rt$$

The slope of this curve is

$$\frac{d\mathrm{RSL}}{dt} = \frac{2\pi A}{P} \cos(2\pi t/P) - Rt$$

We can distinguish two important cases:

Rapid uplift. If $R > 2\pi A/P$ (alternatively, $(RP/2\pi A) > 1$), the slope is always negative, because the cosine term can never exceed 1.0. This means that the corals are continually emerging, and thus the death is not strongly emergent on the cyclic variations. For example, during the 1965 earthquake, uplifts of up to 1.0 m occurred over a day or less (0.003 year). The coral death was not strongly influenced by the annual cycles with amplitude of about 50 cm, since $RP/2\pi A = 10.0$.

Slow uplift. If $RP/2\pi A < 1$, corals will actually be submerging and growing during a portion of the cycle (Figure 8) when the slope is positive. Indeed, the maximum amount E_{max} of coral that emerges each cycle can be as large as

$$E_{max} = RP + 2\left[A \sqrt{1 - (RP/2\pi A)^2} - \frac{RP}{2\pi} \sin^{-1}(RP/2\pi A) \right]$$

where the first term is the tectonic emergence and the second term is the maximum possible amount of upward growth during the submerging part of the cycle. Note that if RP is very small, the maximum possible emergence E_{max} is just $2A$, as expected. As an example, suppose a steady uplift of 5 cm yr^{-1} is superimposed on an El Niño with a period of 4 years and an A of 10 cm. Then $RP/2\pi A = 0.32$, and $E_{max} = 35.9$ cm. Of course, in practice the emergence would be somewhat less because the coral cannot grow the required 15.9 cm during the submergence period of less than 2 years.

The above results show why slowly uplifting corals die during years of oceanographically lowered HLS and only during the winter seasons of those years. However, large uplifts impose a short-term rapid average uplift rate which may cause immediate death for the part of the uplift exceeding the short-term HLS oscillations. Presumably, nontectonic sea level variations have affected coral growth in Vanuatu. Unfortunately, the Port Vila tide gauge has operated only since 1977, so we have no long-term record of the amplitude and timing of local sea level variations. Since 1957, sea level may have been anomalously low in Vanuatu in association with El Niño-type events of 1957-1958,

TABLE 1. Contemporary Coral Emergence Inferred From Growth Bands

Location	Year Measured	Total Corals Analyzed	Date of Coral Death	Number of Coral Recording Date	Inferred Date of Emergence	Mean Emergence, cm
South Santo						
SES	1981	4	1945 1/2-1946 1/2	1	1944 1/2-1946 1/2	9
			1966 1/2	4	1965 1/2-1966 1/2	11
			1972 1/2	3	1971 1/2-1972 1/2	8
			1979 1/2	3	1978 1/2-1979 1/2	5
SCS	1981	3	1966 1/2	3	1965 1/2-1966 1/2	17
			1972 1/2	3	1971 1/2-1972 1/2	12
			1978 1/2	2	1977 1/2-1978 1/2	4
TAN	1979	5	1966 1/2	2	1965 1/2-1966 1/2	≥ 6
			1972 1/2	5	1971 1/2-1972 1/2	6
			1979 1/2	1	1978 1/2-1979 1/2	4
AK	1981	2	1945 1/2-1946 1/2	1	1944 1/2-1946 1/2	≥ 18
			1966 1/2	2	1965 1/2-1966 1/2	18
TAS	1981	1	1965 1/2-1966 1/2	1	1965 1/2-1966 1/2	25
WM	1979	10	1966 1/2	6	1965 1/2-1966 1/2	13 ± 2
			1972 1/2	10	1971 1/2-1972 1/2	9 ± 3
NM	1981	1	1966 1/2	1	1965 1/2-1966 1/2	≥ 10
			1972 1/2	1	1971 1/2-1972 1/2	15
SWS	1981	2	1979 1/2	2	1978 1/2-1979 1/2	5
North Santo						
OLP	1981	3	1974 1/2	3	1973 1/2-1974 1/2	see Table 4
NWP	1979	4	1974 1/2	4	1973 1/2-1974 1/2	see Table 4
NES	1981	1	1974 1/2	1	1973 1/2-1974 1/2	≥ 5
			1978 1/2	1	1977 1/2-1978 1/2	5
Malekula						
M-AB	1977	1	1966	1	1965 1/2-1966 1/2	≥ 15
SWM	1981/1983	7	1957 ± 1	2	1956 1/2-1958 1/2	10
			1966 1/2	6	1965 1/2-1966 1/2	9
			1970 1/2-1971 1/2	5	1969 1/2-1971 1/2	7
			1972 1/2	3	1971 1/2-1972 1/2	5
			1978 1/2-1980 1/2	1	1971 1/2-1980 1/2	4
AKAM	1983	2	1979 1/2-1980 1/2	2	1977 1/2-1980 1/2	4
Epi						
EPI	1983	7	1966 1/2	2	1965 1/2-1966 1/2	7
			1977 1/2	3	1976 1/2-1977 1/2	5
			1978 1/2	2	1976 1/2-1978 1/2	6
			1980 1/2	2	1978 1/2-1980 1/2	4
			1977 1/2-1983 1/2	1	1976 1/2-1983 1/2	7
			1977 1/2-1980 1/2	1	1976 1/2-1980 1/2	≥ 5
Pentecost						
MLS	1983	6	1966 1/2	4	1965 1/2-1966 1/2	
			1967 ± 1	1	1965 1/2-1968 1/2	
			1968 1/2	1	1966 1/2-1968 1/2	8 ± 4
			1977 1/2	4	1976 1/2-1977 1/2	7 ± 2
			1978 1/2	2	1977 1/2-1978 1/2	
			1980 1/2	1	1979 1/2-1980 1/2	3
MEL	1983	4	1966 1/2	2	1965 1/2-1966 1/2	8
			1967 1/2	1	1965 1/2-1967 1/2	
			1977 1/2	4	1976 1/2-1977 1/2	6 ± 1

TABLE 1. (continued)

Location	Year Measured	Total Corals Analyzed	Date of Coral Death	Number of Coral Recording Date	Inferred Date of Emergence	Mean Emergence, cm
Maewo						
NAS	1983	7	1965 1/2	1	1965 1/2-1966 1/2	
			1964 1/2-1967 1/2	1	1963 1/2-1967 1/2	6
			1966 ± 1	1	1964 1/2-1967 1/2	
			1969 1/2-1970 1/2	2	1968 1/2-1969 1/2	4
			1973 1/2	2	1972 1/2-1973 1/2	4
			1974 1/2-1975 1/2	6	1973 1/2-1974 1/2	6 ± 3
			1977 1/2-1979 1/2	7	1976 1/2-1978 1/2	5 ± 2
			1981 1/2	1	1980 1/2-1981 1/2	3

1963-1964, 1965-1966, 1969-1970, 1972-1973, 1976-1977, and 1982-1983 [*Quinn et al.*, 1978, Figure 4].

However, for several reasons it is often possible to infer tectonic causes for relative sea level changes:

1. Changes of HLS near the epicenter and the time of major earthquakes are likely to be tectonic.

2. Changes of HLS which are large (10 cm) or which vary rapidly over small distances are more likely to be tectonic. Oceanographic changes in HLS should occur throughout Vanuatu at about the same time.

3. When changes are found in areas of rapid Quaternary tectonism and are not found in known stable areas, the event is likely to be tectonic.

4. When a change of HLS persists for years with no tendency for coral colonies to grow upward to their former levels the event is more likely to be tectonic. For example, when each emergence event adds to the total emerged and killed coral surface, tectonic uplift is likely even if each level of coral surface died during a time of oceanographically lowered sea level. Oceanographic changes of HLS are usually transient events with durations of a few months to a few years.

Living Coral Record of Emergence and Submergence

Data from living and recently dead emerged corals provide a detailed account of emergence and submergence during recent decades in central Vanuatu (Tables 1-3). In this section we present detailed descriptions of coral emergence events in order of their estimated time of emergence.

1945 or 1946 Event

The oldest emergence event whose age we could determine (1945 or 1946) was found only on south Santo (Figure 3 and Tables 1 and 2). It was not found on Malekula or north Santo, although well-preserved coral heads exist that surely would have recorded the event. We found this event in six corals at four different locations (Tables 1 and 2). Growth bands in *Goniastrea retiformis* colonies at AK and SES show a minimum of 18 cm of uplift in 1945 or 1946 (Table 1), causing death between about mid-1945 and late-1946. The growth bands in the four undated *Porites lobata* corals were difficult to interpret because as is common in this species, the growth bands are vague (Table 2). However, there was approximately 20 cm of coral growth between an emergence event in 1965 (see below) and the previous emergence event. As corals typically grow about 1 cm yr^{-1}, we propose that the previous event was the 1945-1946 event. The emergence has persisted, and there is no evidence of upward growth of coral polyps higher than the coral surface that died in 1945 or 1946. Evidence of the 1945-1946 event probably is scarce because (1) emergence was less than 20 cm so that the dead corals remained in the intertidal zone where bioerosion is intense and (2) two subsequent emergence events in 1965 and 1971 probably killed other corals that recorded the 1945-1946 event without raising them above the zone of intense bioerosion.

1957-1958 Event

Only two coral heads on southwest Malekula record some emergence at this time (Figures 6 *b* and 9 and Table 1), but nearly all corals from that area display especially dense growth bands from 1957 until 1959-1960 (Table 1 and Figure 10). However, the corals indicate only a few centimeters of temporary emergence after which most grew back upward to their pre-1957 levels. No other areas of 1957-1958 emergence were discovered in central Vanuatu, but 1957-1959 growth bands in corals from other areas are often denser and thinner than usual, suggesting that some environmental change slowed growth.

1965-1968 Event

This was the largest and most widespread emergence event of recent decades in Vanuatu. After *Mitchell* [1971] and *Benoit and Dubois* [1971] noted this event, *Taylor et al.* [1980] mapped its extent. More recent results show that the emerged area also included south Santo, an area of southwest Malekula, and the back arc island of Pentecost (Figure 11 and Tables 1 and 2).

Where the 1965-1968 coral emergence is greater than about 30 cm, corals died in late 1965. However, where emergence was less than 30 cm, most corals died in 1966, and a few in 1967 and even 1968. It is possible that all died due to a single event, but where emergence was slight some corals survived for 1-3 years. Growth bands in those that died in 1967 or 1968 often show increased density and slower growth after the mid-1966 HD band.

Maximum emergence was 120 cm along a SW-NE axis across north Malekula (Figure 11) [*Taylor et al.*, 1980]. Emergence decreases to nil at the east coast and to about 60 cm across a zone of steeply dipping faults on the west coast of north Malekula. Emergence decreases gradually to zero southward along the west coast of central and south Malekula. Most of the south and southeastern coast did not emerge, and even a slight submergence may have occurred.

TABLE 2. Contemporary Coral Emergence Based on Growth Bands and Growth Rates

Location	Year Measured	Dates of Emergence	Total Corals Recording Date	Range of Emergence, cm	Mean Emergence, cm	Total Contemporary Emergence, cm
South Santo						
	1979/1981	1946	3	20-48	29	
		1965	5	13-25	20	55
		1971	4	5-8	6	
AK	1981	1946	1		≥ 18	
		1965	2		18	41
		1971	1		5	
TAN	1979	1965	2		≥ 6	
		1971	5		6	≥ 16
		1978-1979	1		4	
SCS	1981	1946	2		14	
		1965	6		12	
		1971	6		11	42
		1978-1979	2		5	
DEL	1981	1965	2		10	
		1971	2		11	≥ 27 +
		1978-1979	1		6	
RAT	1981	1946	2	9-10	10	
		1965	7	6-18	11	
		1971	6	5-12	7	33
		1978-1979	4	4-5	5	
WM	1979	1946	1		10	
		1965	6	9-15	13	
		1971	10	5-18	9	38
		1978-1979	1	–	6	
NM	1981	1946	1	–	≥ 5	
		1965	2	4-14	9	
		1971	2	4-10	7	24
		1978-1979	1		3	
CSW	1979	1965	1		10	≥ 10
SWS	1981	1978-1979	2		10	5
Pentecost						
BAT	1983	1965-1968	2	5-8	7	
		1976-1977	7	6-11	8	15
LON	1983	1965-1968	2		5	
		1976-1977	3		5-6	11
MAR	1983	1965-1968	observed		10	10
Malekula						
SBN	1981	1965	observed		10	10

Along part of the west coast of Malekula, submergence appears to have preceded the 1965 emergence. Just north of Dixon Reef are many *Goniastrea retiformis* colonies that had emerged about 35 cm in 1965-1968. However, these colonies have a morphology as in Figure 5*a* that records a period of submergence that began about 20 years before the 1965-1968 emergence event. Unfortunately, we were unable to sample these corals to obtain a more precise relative sea level history. Near Dixon Reef, emerged *G. retiformis* colonies had colonized subtidal beachrock. Normally, beachrock forms intertidally. Perhaps subsidence lowered the beachrock to subtidal levels where corals could colonize it, unless the beachrock formed prior to 6000 years B.P. when "eustatic" sea level was lower.

Farther south on Malekula we discovered an isolated zone of 1965-1968 coral emergence (Figure 11). Of seven samples collected, six died down an average of 10 cm in 1966-1968 (Table 1). This zone of emergence is separated from the north Malekula 1965-1968 emergence by an area near Southwest Bay where there was no Holocene or contemporary coral emergence.

Emerged corals across the south coast of Santo died in mid-1966 from emergence that occurred between mid-1965 and mid-1966. Maximum 1966 emergence of about 25 cm occurred at

TABLE 3. Contemporary Emergence Based on *Acropora* sp. Corals

Location	Year Measured	Total Number of Corals Measured	Emergence Range, cm	Mean Emergence, cm
North Santo				
NWP	1981	7	50-65	58±6
SWP	1979	12	46-69	59±7
OLP	1981	10	40-70	58±9
TMT	1979	observed		0
WAS	1979	observed		0
MAT	1981	observed		0
NES	1981	observed		~10
South Santo				
PAL	1979	observed	–	5
RAT	1981	17	10-25	16±5
DEL	1981	14	22-50	34±7
RPT	1981	33	22-42	34±5
TAN	1979	7	23-38	29±6
AK	1981	9	20-32	26±5
CSE	1979	3	24-30	27
CSW	1979	observed		~10
CLS	1979	observed		0
TSK	1981	observed		~10 (post-1965)
Malekula				
DRS	1981	19	14-37	26±9

Tasmaloum with a rapid decrease to nil emergence near the west coast. Emergence decreases very gradually toward the east (Figure 11). Emergence of this age does not extend up the east or west coasts of Santo.

Most intertidal corals on Pentecost died from emergence of about 10 cm in 1966 with some surviving until 1967 and 1968. We documented emergence in 1966-1968 for a total of eight out of 10 living corals at two localities on Pentecost (Table 1). Other localities were not sampled, but we infer emergence at about the same time based on the amount of growth since emergence killed part of the colonies.

We sampled only one locality on Maewo, but of seven corals, four died down slightly between 1964 and 1966. On Maewo, emergence and death of some corals preceded that of 1966 on Pentecost, Santo, and Malekula. Moreover, three of the four Maewo corals grew upward following emergence in mid-1964 to mid-1967. This suggests a different cause for emergence or a different sequence of vertical movements for Maewo. A few coral heads from the west coast of Epi also record slight emergence and death in 1966-1968.

Other areas clearly did not emerge near 1965-1968. These include the northern three fourths of Santo and southeast Malekula. Ambae and Ambrym may have submerged slightly, but these areas also did not emerge at any time in recent decades.

Although the times of coral death ranged from 1965 until 1968, most coral growth bands exhibit increased density after the early 1966 HD growth band. The increase in density may be related to emergence. We infer a time of emergence between mid-1965 and mid-1966. Only on Maewo is there any evidence of upward coral growth following emergence in 1965-1968, and here the emergence event seems to have occurred prior to 1965.

1969-1970 Event

A few corals died down slightly in 1969-1970 on southwest Malekula (four of seven samples), Tangoa off the south coast of Santo (only one of many), on Maewo (two of seven), and on Epi (one coral; see Table 1 and Figure 12). Only a few corals record this event in areas where it is found except on southwest Malekula where most sampled corals record the event. Whatever the cause of this event, it occurred between mid-1969 and mid-1970. Only the 1969-1970 emergence of southwest Malekula has persisted without evidence of the living coral surface regaining the former higher HLS.

1971-1972 Event

Nearly every coral that we sampled along the south coast of Santo died down in mid-1972 from partial emergence (Figure 13 and Tables 1 and 2). Maximum emergence of about 15 cm occurred near the center of the south Santo coast line (Tables 1, 2, and 3). Toward southwest Santo and eastern Malo, emergence decreases to nil. This event is not found on the west or east coasts of Santo, on north Santo, or on north Malekula. However, three partially emerged coral heads on southwest Malekula also died down in 1972 (e.g., Figure 7). This may have been delayed coral death related to the 1969-1970 event on southwest Malekula. One sample from Epi records an event in 1972, but the emergence was temporary and coral grew back up to its pre-1972 level by about 1976.

The main 1971-1972 event is limited to south Santo and neighboring small islands. Emergence could have occurred any time from the last two thirds of 1971 until mid-1972. On south Santo the emergence has persisted with no tendency for corals to grow up to the levels they reached before the emergence event.

1973-1975 Event

The maximum uplift measured for this event was 59 ±7 cm, determined from *Acropora sp.* colonies at locality SWP (Table 3 and Figure 14). This event may not extend very far south because there is no apparent emergence at small patch reefs farther south

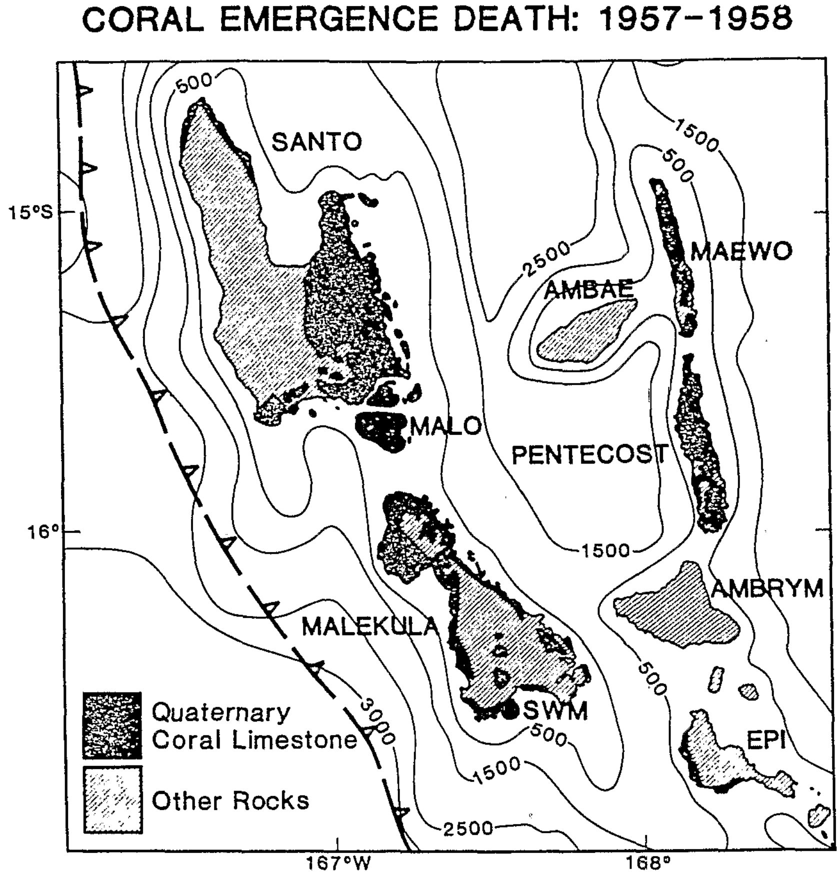


Fig. 9. Location of emerged coral death in 1957-1958. We found exposed coral surfaces that died in 1957 only in one coral on southwest Malekula (Figures 6 *b* and 10). However, at other locations on Maewo and Pentecost there is a clear increase in skeletal density and slower growth for the 1957-1958 growth bands. A temporary density increase also occurred in 1957-1958 in corals to the south on Efate Island (Figure 1).

on the west coast of Santo. However, these localities are on little peninsulas that may be separated from the mainland by active normal faults. All of the emerged corals sampled on northwest Santo died in early 1974 (Table 1), usually after the HD band was deposited. Unfortunately, we could not visit sites on the east side of the northwest peninsula, but we expect that coral reefs in that area recorded the 1973-1974 event. On the southwest side of the northeast peninsula there was no 1974 emergence at MAT. However, a few kilometers to the north slightly emerged corals begin to appear (NES; Table 3). At locality NES we collected an emerged coral sample which clearly died down after the 1974 HD band (Table 3). Other coral heads emerged at about the same time, but we did not acquire samples.

On Maewo Island there is strong evidence of about 6 cm of emergence coral death in 1973-1975 (locality NAS; Figure 4 and Table 1). Five samples died after the 1975 (locality NAS; Figure 4 and Table 1) HD band, one died after the 1973 HD band, and one after the 1974 HD band. Bioerosion may have removed the 1974 growth band from the one coral whose last band grew in 1973. No coral sample from Pentecost, Malekula, or islands other than Maewo, records emergence death between 1973 and 1975. Emergence occurred sometime between the last two thirds of 1973 and mid-1974.

1977 Event

Coral death in mid-1977 is particularly prominent on the west coast of Pentecost (Figures 6*a*, 7, and 15 and Table 1). Of the 10 corals sampled, eight died in mid-1977 and two in mid-1978. Average emergence was 6 cm. The 1977 and 1978 events probably are related since no coral records both events. Three of eight coral samples from Epi also record a 1977 emergence event. However, emergence that began to kill coral surfaces in 1977 on Epi continued to lower their HLS a little each year until about 1980 or 1981.

1978-1981 Event

On Maewo, coral surfaces died from mid-1978 through mid-1979. Several died down a few centimeters each year, whereas others died down all in 1 year. Death of these coral surfaces may have a single cause, whose effect was delayed in some corals. Average total emergence for 1978-1979 is only 5 cm.

A few corals from other islands died during this period (Table 1 and Figures 6*b*, 10, and 16). Of 11 corals sampled at WM off south Santo, one coral died in 1978 and one in 1979 (Table 1). Nearby SCS records one event occurring in 1978 or 1979. TAN has only one 1979 emergence among six coral heads, SWS has two 1979 emergence events from three samples, and the coral at NES died down a few centimeters in 1978 (Table 1 and Figure 16). However, besides Pentecost and Maewo, this event is clearly represented only on southwest Malekula, where most corals have denser bands in 1979-1981 and seven of nine died down about 4 cm (e.g., Figure 11). As mentioned previously, during this period the gradual emergence on Epi that started in 1977 continued, causing several corals to die down in 1978, 1980, and 1981 (Table 1).

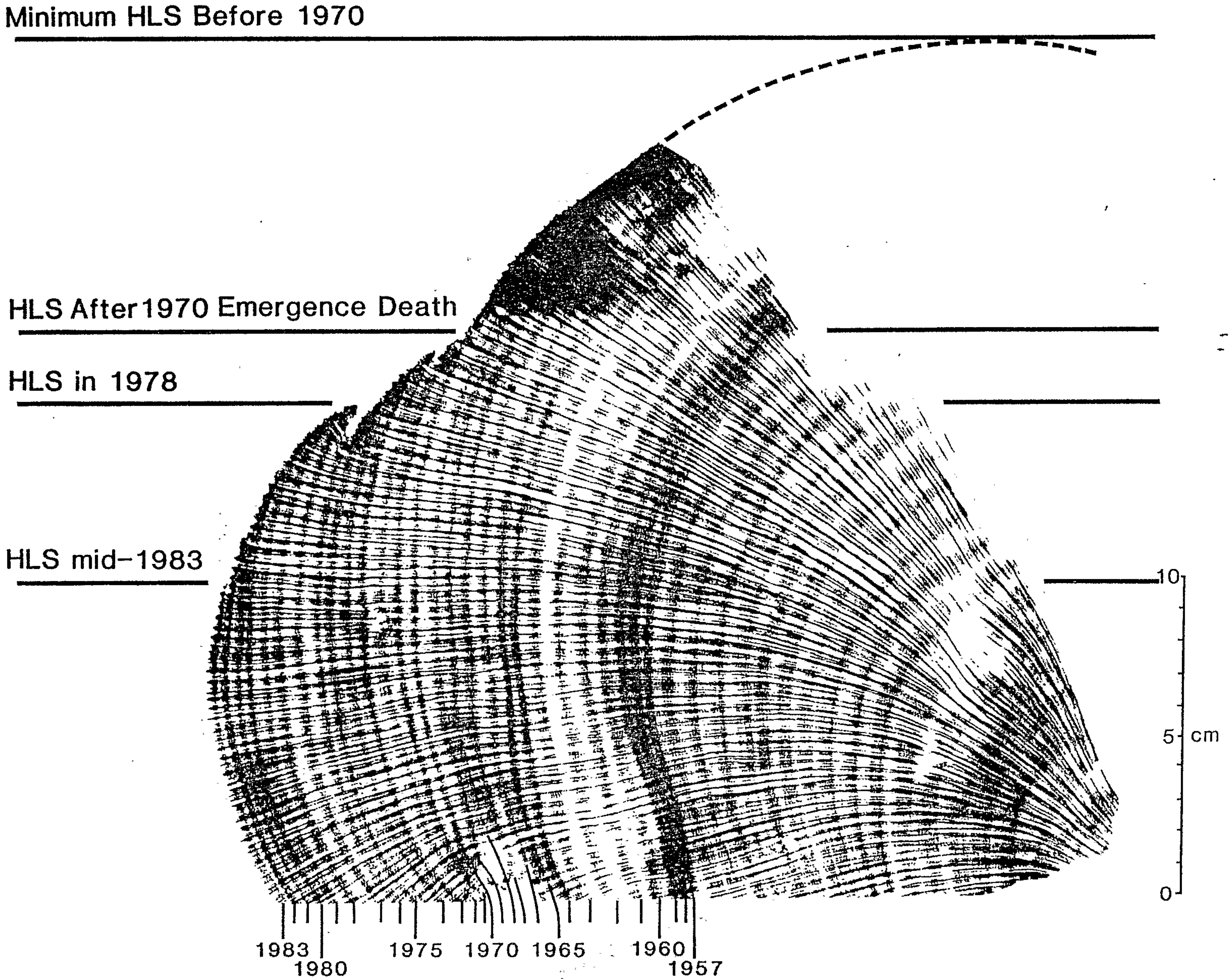


Fig. 10. X radiograph positive of part of a coral head from southwest Malekula (Table 1 and Figure 9) that died downward during several emergence events beginning in 1970. Note years indicated at the bottom. Slight additional death occurred between 1972 and 1975 with about one centimeter of upward growth in 1976 and 1977. The level of living coral fell several centimeters from 1978 through 1981. The 1957-1959 growth bands are particularly narrow and dense and may be related to a major El Niño event at that time (Figure 4 *a*).

CORAL EMERGENCE DEATH: 1965-1968

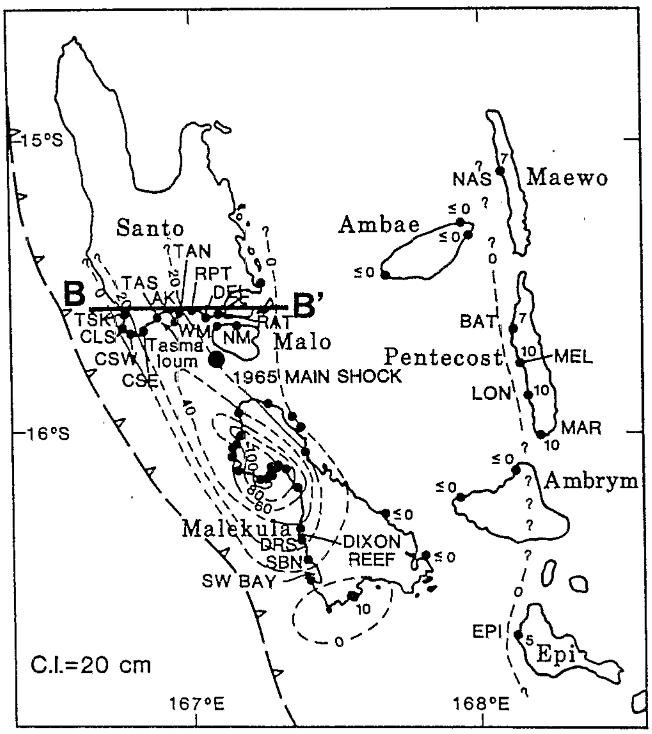


Fig. 11. Locations of coral emergence death in 1965-1968. Unlabeled localities are from Table 2 and Figure 8 of *Taylor et al.* [1980] or are for localities where no evidence of emergence exists. New data are shown in Tables 1, 2, and 3. Most emerged coral surfaces died in 1966, but a few colonies that were only slightly emerged died down in 1967 or 1968. There is a clear absence of emergence at this time on north Santo, Ambae, and Ambrym, although these and other areas were examined. On north Malekula and south Santo, this emergence was clearly coseismic. The epicenter of the main 1965 shock is shown as a solid circle south of Malo Island. See Figure 18 for the aftershock zone. B-B' is the location for a profile of these data (Figure 19). We have no proof that emergence on southernmost Malekula, Pentecost, and Epi was coseismic, but it is spatially and temporally related to that on north Malekula and south Santo.

Between the 1978-1981 emergence event or events until mid-1983, no significant events causing additional coral death are presently known. Of course, we have not visited some areas since 1979. A few the samples show continued lowering of their HLS into 1980 or 1981, but the process began in 1977 or 1978. Surprisingly, there was no evidence of coral death associated with the strong 1982-1983 El Niño [*Wyrtki*, 1985] when we left Vanuatu in September 1983, nor when we visited Efate in 1986.

Evidence of Recent Submergence on Ambae

No emerged coral limestone occurs on either Ambae or Ambrym (Figure 3). We examined living corals on both the eastern and western ends of Ambae and found numerous corals bearing evidence of having been at their HLS, then submerging at least a few centimeters (Figure 6*c*). A stand of dead coconut palms near the shore and severe coastal erosion also indicates submergence. The dead trees stand in a pond of water on the low coastal plain because the water table has risen. The height of the trees indicates they had grown there at least 30 years before the area became so swampy that they died. We did not attempt to determine the time of submergence, but preservation of the trees indicates that they died within the past 10 years.

INTERPRETATION OF CORAL EMERGENCE EVENTS

There are eight time intervals from 1945-1980 when the HLS of coral heads lowered in the central Vanuatu islands. Only four of these events are definitely related to large shallow earthquakes in 1946, 1965, 1971, and 1973 (Figures 17 and 18). If the remaining events are tectonic, then they imply that nonseismic uplifts contribute significantly to the tectonics in Vanuatu. However, during recent decades there have been several temporary oceanographic lowerings of sea level. In this section we shall evaluate the reliability of each coral emergence event and consider its cause.

1945-1946 Event

This event, with a maximum emergence of 20 cm, was probably tectonic and very likely coseismic. Only the limited area of south Santo was affected, and the emergence has persisted. We are certain that there was little if any uplift of north Malekula or of northwest Santo. Numerous colonies in these areas would have recorded emergence, but none do except in 1965 and 1973. We do not know whether the back arc islands uplifted at this time.

There were two major earthquakes in 1946 (Figure 17). *McCann* [1980] relocated the first (January 5, 1946; $M_s = 7.3$) under south Santo and the other (January 20, 1946; $M_s = 7.0$) beneath north Malekula. At least one of these earthquakes must have ruptured the south Santo arc segment. This indicates that McCann's location beneath south Santo is correct and suggests that the second 1946 event (15 days later) may have also been in the south Santo segment, rather than under north Malekula, possibly as a downdip aftershock.

1957-1958 Event

There was a major El Niño event in 1957-1958 which may have lowered sea level in the Vanuatu region. However, this emergence event caused persistent emergence of some corals only in a zone of documented rapid Holocene uplift on southwest Malekula. Perhaps corals in this area were vulnerable for a nontectonic reason, or perhaps the sea level lowering was superimposed on corals that had already accumulated a few centimeters of uplift. We favor the second hypothesis which explains why corals living on southwest Malekula died without later growing upward to their former levels.

1965-1968 Event

We know that Santo and Malekula uplifted coseismically during the August 11, 1965, earthquakes (Figure 18) because of eyewitness accounts and a visit to Malekula by A. H. G. Mitchell about 2 weeks after the event [*Mitchell*, 1968, 1971, personal communication, 1978]. Numerous additional witnesses from Santo and Malekula described events on the morning of the main shock. The details differ, but some common observations are probably accurate. Combined with the accounts from *Benoit and Dubois* [1971] and *Taylor et al.* [1980], these previously unpublished comments provide a more complete review of events accompanying the August 1965 earthquakes.

About 3 days before the sequence began, A. Stevens at Tasmaloum on south Santo (Figure 11) reported that a fish about 3 m long swam into Tasmaloum Bay and beached itself. This unfamiliar type of fish had a long muscular snout or head and no discernible eyes. The next day two large manta rays beached

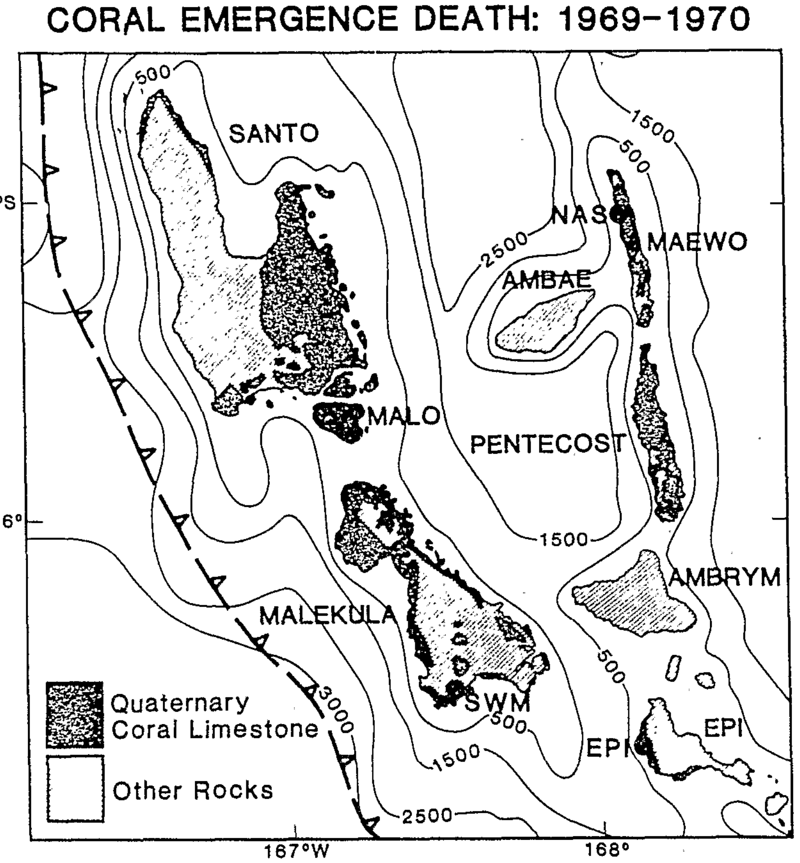


Fig. 12. Locations of coral emergence death in 1969-1970. The event was small and temporarily affected only a few corals on Maewo. However, on southwest Malekula several centimeters of 1970 emergence have persisted. On Erromango at 18.5 °S (Figure 1), nearly every colony records a temporary emergence event at this time. The time of coral death is probably controlled by the 1969-1970 El Niño, but the persistence of emergence on south Malekula most strongly suggests tectonic uplift there.

themselves. There were also large fish kills around Santo and Malekula reported near the time of the main shock [*Taylor et al.*, 1980; A. H. G. Mitchell, personal communication, 1978]. During late morning on August 11, roughly at the time of the main shock, sea level on south Santo slowly fell at least 2.5 m. Here and on north Malekula, people ran out and collected fish that were either stranded or stunned. Stevens estimated that the water stayed out no more than 5-8 min. The sea then surged to approximately 1 m above normal high-tide level. The sea made some noise going out and returning. Other accounts from north Malekula differ in that the sea stayed out for times estimated at 15 min to 1 hour. One witness stated that sea level rose 2 m for 30 min before going out for 30 min after which it rose again to well above high-tide level for about 30 min. One witness stated that the water simply returned to its normal level.

From these stories it is clear that (1) sea level fell as much as several meters, (2) the sea stayed low long enough for people to walk out and collect fish, and (3) fish behaved abnormally even before the foreshocks and main earthquakes began. The retreat of the sea suggests either a long-period tsunami or that south Santo and north Malekula temporarily uplifted several meters and then rapidly subsided. The precise relationship of the main shock and the 1965 uplift pattern to these events is not clear. However, we assume that the 1965 coseismic emergence portrayed in Figure 11 occurred during these events.

Following the preliminary map of *Benoit and Dubois* [1971], we mapped this deformation pattern on north Malekula [*Taylor et al.*, 1980]. However, we have modified our previous map based on new data (Tables 1, 2, and 3 and Figure 11). Emerged coral heads at many locations on south Santo show partial death in mid-1966 (Tables 2 and 3). As described above, the main contiguous area of emergence of Malekula and south Santo was certainly uplifted coseismically in August 1965. A witness claimed that north Malekula uplifted 2-3 m during the 1965 earthquake but that most of this uplift was recovered within 1 year. If this is so, the recovery must have taken place within days or weeks of the initial uplift. Otherwise the uplift would have killed all corals down to at least 2 m deeper than the highest corals living before the 1965 earthquake, whereas in 1976 we observed only a maximum of 1.2 m of coral death on Malekula.

Except on Malekula and south Santo we can not assume that the emergence was coseismic. Emergence death of most coral samples from Pentecost in 1966-1968 may also be tectonic because it occurred near the time of the large 1965 uplift in the west and has persisted. A few corals on Epi and Maewo also record slight emergence near this time. El Niño events in 1963-1964 and 1965-1966 [*Quinn et al.*, 1978] may have dictated when corals died but are not the cause of persistent emergence. The absence of this event elsewhere suggests that slight emergence of corals on Epi and Maewo may be related to that on Pentecost.

The locations of the five main earthquakes of the August 1965 sequence (Figure 18) indicate that both the south Santo and north Malekula segments ruptured [*Pascal et al.*, 1978; *Taylor et al.*, 1980]. Our results show that both north Malekula and south

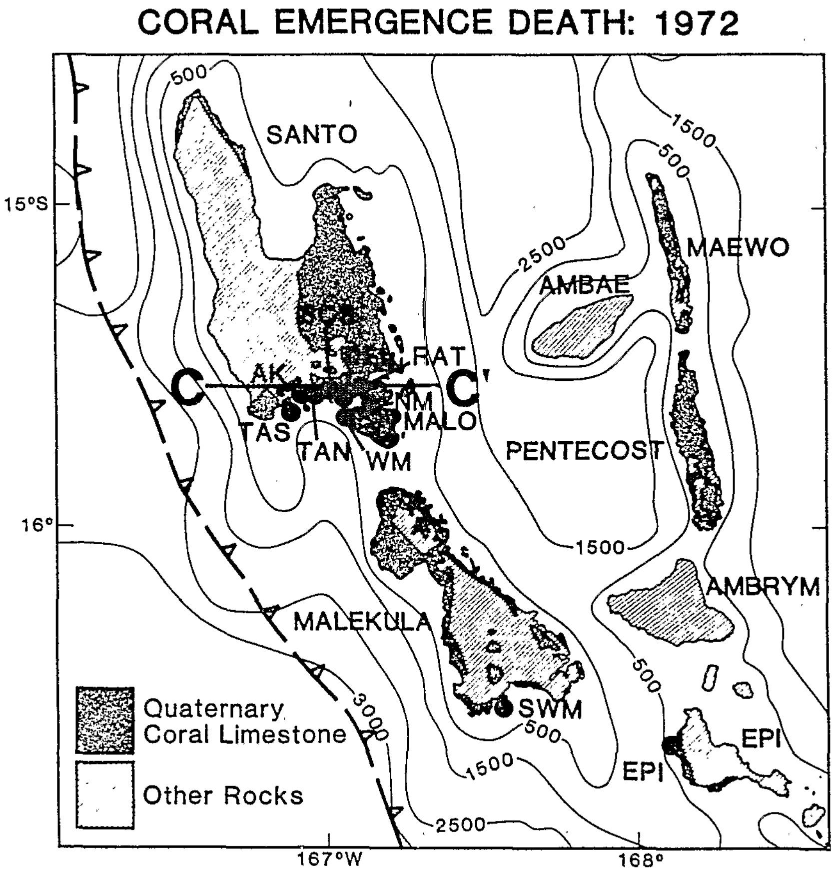


Fig. 13. Locations of emerged coral death in 1972. This event is recorded by virtually every coral head on south Santo and is probably related to the October 27, 1971, earthquake (Figure 8). It affected about half the colonies sampled on southwest Malekula (*e.g.*, Figure 6*b*) and only one sample from Epi. The persistence of emergence of these corals suggests a tectonic cause, but not necessarily related to the distant 1971 earthquake. C-C' is the location for a profile of these data (Figure 19).

Santo uplifted. However, south Santo responded to seismic rupture differently than north Malekula. Although maximum Holocene emergence for south Santo is about twice that of north Malekula, maximum coseismic uplift on south Santo was only about 25 cm whereas north Malekula uplifted as much as 120 cm.

South Malekula also responded to the 1965 event differently than north Malekula. Southward from central Malekula along the west coast 1965 uplift decreases to zero at Southwest Bay. However, net Holocene uplift along this coast decreases even more quickly to zero, while 1965 uplift is still about 30 cm. There is no evidence of preseismic subsidence on north Malekula, but we did find such evidence along this coast. Uplift does not accumulate in this area If the 1965 uplift is typical, then the area must oscillate with preseismic subsidence and coseismic uplift.

The zone of 1966-1967 emergence on southwest Malekula is separated from the main uplift zone to the north by a graben at Southwest Bay. The significance of this isolated uplift zone is not clear. Here Holocene uplift has averaged 0.25 cm yr^{-1} in the most emerged area, and there is no evidence of preseismic emergence. The area uplifted much less in 1965 than did parts of north Malekula that have the same average Holocene uplift rate. The presence of other emergence events which may be tectonic in origin suggest that this area may uplift more frequently than north Malekula.

Corals from south Malekula suggest that the 1965 rupture extended beneath Malekular either coseismically or postseismically. A few aftershocks south of Malekula led *McCann* [1980] and *Habermann* [1984] to extend the rupture zone south of Malekula. However, given the location and magnitude of the main shocks, others consider it unlikely that the rupture zone extended beyond central Malekula [e.g., *Ebel*, 1980; *Isacks et al.*, 1981]. The presence of a few 1965 aftershocks under southwesternmost Malekula does not prove that the 1965 rupture and coseismic uplift extended this far south. The 1965 earthquake may be atypical because it ruptured at least the south Santo and north Malekula segments, unlike the 1946 and 1971 events confined to south Santo.

1969-1970 Event

We hesitate to assign a tectonic origin to this event, as few corals record it and it is not a major event. Moreover, this was a time of El Niño when sea level was lower elsewhere in the western Pacific [*Meyers*, 1982] and possibly in Vanuatu as well. However, on southwest Malekula it is possible that it was caused by the superposition of lowered regional sea level on gradually uplifting corals.

1972 Event

Coral death in mid-1972 on south Santo is surely related to the shallow earthquake ($M_s = 7.1$; Figure 18) which occurred beneath southeastern Santo on October 27, 1971. This earthquake had the third largest moment of all earthquakes occurring in central Vanuatu since 1964 [*Chinn and Isacks*, 1983]. Unlike the 1965 event, uplift associated with this earthquake occurred almost

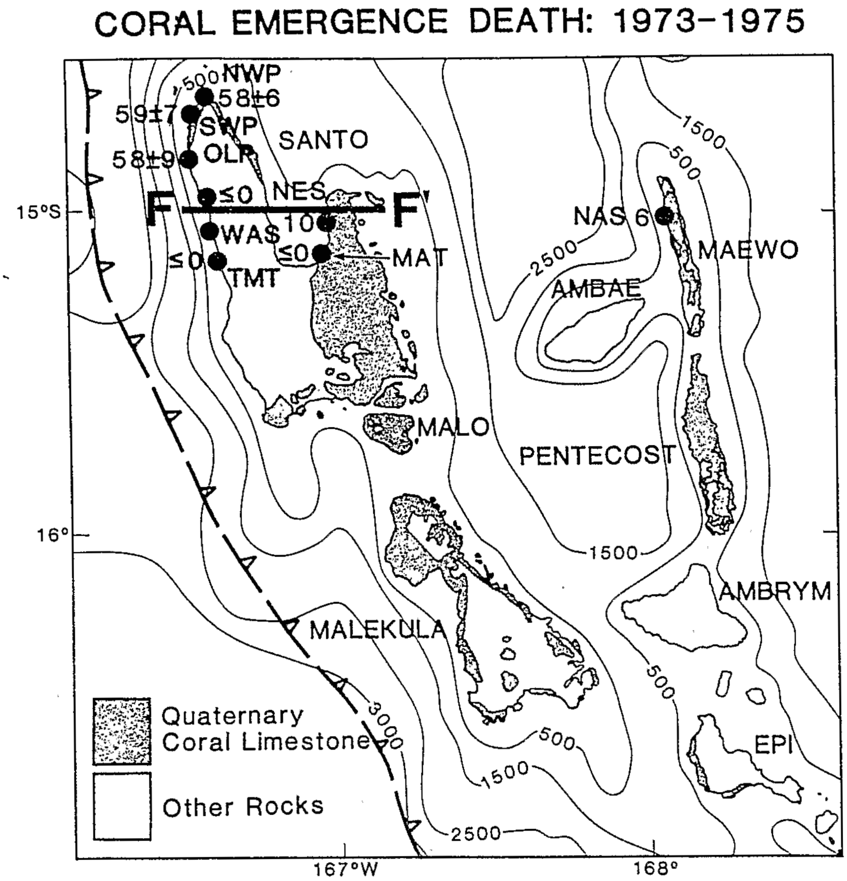


Fig. 14. Locations of coral emergence death death in 1973-1975. Nearly every coral sampled at these localities recorded this event (Table 1). Maximum emergence amounts of nearly 60 cm are shown on northwest Santo. F-F' locates topographic, and uplift profiles shown in Figure 21. The 1973-1975 emergence is probably due to the large earthquakes of December 1973 (Figure 18). The numbers to the left of the stations are the centimeters of emergence measured. The zero values on the west coast of Santo are for corals on small peninsulas that appear to be separated from the island by normal faults. However, it is possible that uplift decreases rapidly southward from the areas of emerged coral on the northern end of northwest Santo.

exclusively on south Santo and nearby small islands (Figures 13 and 19). The timing of coral death requires that emergence occurred anytime from late mid-1971 until mid-1972 and persisted until we sampled coral heads in 1979 and 1981. It is unlikely that this event affected the emerged corals found on southwest Malekula and on Epi (Figure 16). An El Niño regional sea level lowering was reported for 1972 [*Wyrtki*, 1977] and a few corals that were not uplifted at all could have been affected. Nevertheless, the 1972 emergence of southwest Malekula has persisted and a component of tectonic uplift there is certainly possible.

For the 1971 earthquake, most of the uplift is toward the eastern side of the 1965 uplift zone on south Santo (Figure 19), and the October 1971 epicenter lies slightly east of the August 1965 epicenters [*Pascal et al.*, 1978]. The 1971 event has a thrusting focal mechanism and probably released residual stress on the interplate thrust downdip of the 1965 rupture zone. Other, smaller magnitude events in this area in 1966, 1972, and 1973 may also have released downdip stress, although they did not cause vertical deformation.

1973-1975 Event

The 60 cm of emergence concentrated on the western part of north Santo (Table 1 and Figure 14) is the second largest emergence event observed in central Vanuatu, after the 1965 event. This event is certainly related to the December 1973 to January 1974 earthquakes (Figure 18) and was probably coseismic. In accord with the large observed uplift, the December 29, 1973, earthquake had a higher seismic moment than any other central Vanuatu earthquake since the 1965 event [*Chinn and Isacks*, 1983]. In the area of the greatest 1973 uplift, large corals live immediately below those killed by emergence and preclude greater uplift than we measured unless rapid subsidence followed and was complete by mid-1974. Many of the corals were only partially emerged and partially killed and record no post-1973 upgrowth. Coral emergence decreases rapidly toward the east. No 1973 uplift was found at locality MAT, although it is near the epicenter of the largest aftershock located by the International Seismological Centre (M_S = 7.2; depth = 30 km). At NES there is slight 1973 uplift, but the maximum 1973 uplift is skewed toward the west. Because northeast Santo has more than 15 m of Holocene emergence, one wonders how it comes to be uplifted if the 1973 event is typical. An earthquake occurred on northeast Santo on November 20, 1974 (M_S = 6.9; depth = 66km), but we find no associated uplift.

In the back arc, 1973-1975 emergence death is limited to Maewo. The emergence is less than 10 cm, which may account for some coral surfaces dying in 1975 instead of 1974. Two corals that partially died in mid-1973 must have emerged before the December 1973 earthquake and suggest preseismic uplift. Neither of these coral heads record El Niño events nor did sea level anomalies coincide with this emergence event.

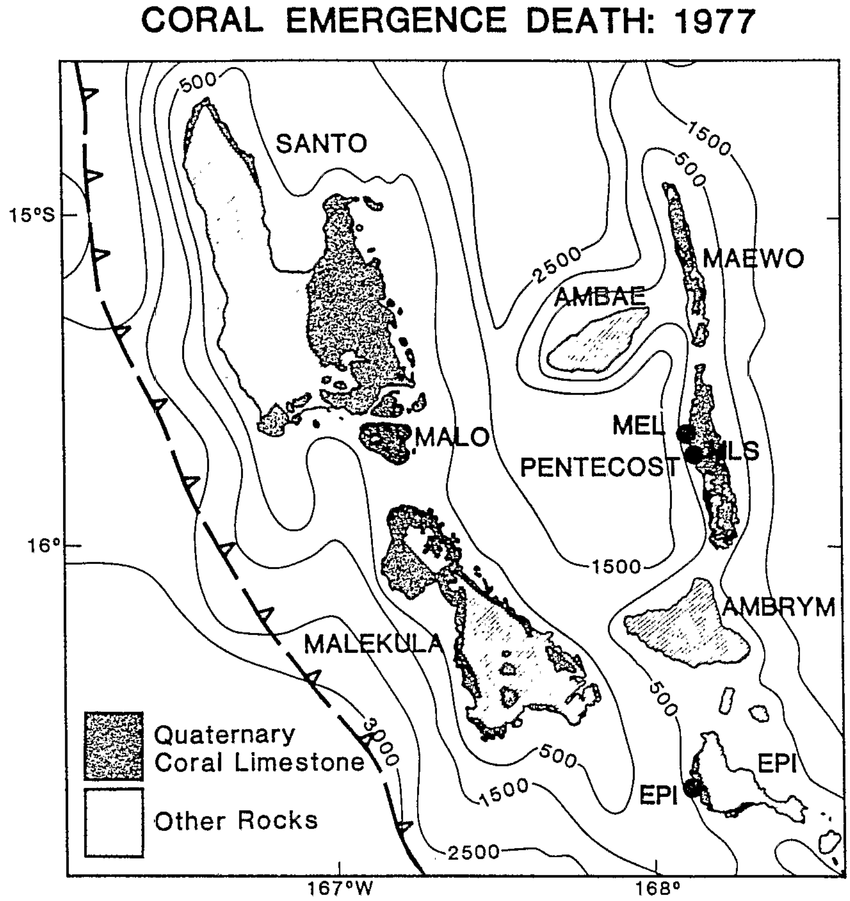


Fig. 15. Locations of coral emergence death in 1977. Nearly every sample from Pentecost recorded this event, which otherwise appeared in only a single sample from Epi. This event is apparently limited to Pentecost and is thought to be tectonic, but not coseismic, because of its persistence and geographic limits.

1977 Event

The mid-1977 coral death on Pentecost where nearly every sample records 2-10 cm of emergence is probably related to tectonic uplift. There was a regional oceanographic sea level depression in 1976-1977 [*Wyrtki*, 1979]. If this depression included Vanuatu, it affected coral growth in 1977 only on Pentecost and in two samples from Epi. This seems unlikely, so we suggest that a sea level low in 1976-1977 may have been superimposed on nonseismic tectonic uplift that occurred before mid-1977.

1978-1981 Event

Perhaps several discrete emergence events occurred during the 1978-1981 period. Emergence is very common among corals on Maewo and south Malekula but is rare on Santo (Figure 16). It occurs on Epi but then continues gradually at about 1 cm yr^{-1} until 1981-1983. Some samples from southwest Malekula also show a continuation of coral death to progressively lower levels for several years after 1978 (e.g., Figure 10). The 1978-1979 emergence death occasionally found on Santo is smaller than the 1977 event on Pentecost or the 1978 event on Maewo, where nearly every sample records 2-10 cm of emergence. Although 1977 coral death on Pentecost may be related to the mid-1978 or 1979 death on Maewo, the consistent difference in timing suggests otherwise. On Epi emergence was common in 1977 or 1978 but seems to continue gradually until 1981-1983.

Bevis and Isacks [1981] detected a tilt anomaly on SE Santo in 1979-1980 that they term "marginally significant" [*Isacks et al.*, 1981]. There was a relative sea level low in 1978-1979 at Efate (K. Wyrtki, personal communication, 1985), which may have had an oceanographic cause.

QUATERNARY VERTICAL DEFORMATION OF THE VANUATU ARC

A series of highly visible emerged coral reefs forms the coastline throughout much of central Vanuatu (Figure 3). Mapping and isotopic dating of the coral reef terraces provide an excellent record of the Quaternary vertical tectonic history of central Vanuatu [*Launay and Recy*, 1972; *Neef and Veeh*, 1977; *Taylor et al.*, 1980, 1981, 1985; *Jouannic et al.*, 1980, 1982; *Gilpin*, 1982; *Urmos*, 1985]. Santo and Malekula both uplifted while tilting down to the east with maximum average late Quaternary rates as high as 0.6-0.7 cm yr^{-1} along the topographic/uplift axis paralleling the west coast of Santo. West of this axis, uplift rates of north Malekula and south Santo derease rapidly across a zone of north-south trending, steeply dipping faults [*Mitchell*, 1968; *Robinson*, 1969; *Mallick and Greenbaum*, 1977]. Coral limestone to heights of 602 m on north Malekula and 783 m on south Santo suggest that the present episode of tectonic uplift has lasted for at least 300,000 years.

The ^{14}C and $^{230}Th/^{234}U$ ages from approximately 200 samples collected over the past 10 years document the past 200,000 years of uplift and tilting particularly well. Based on tilt directions

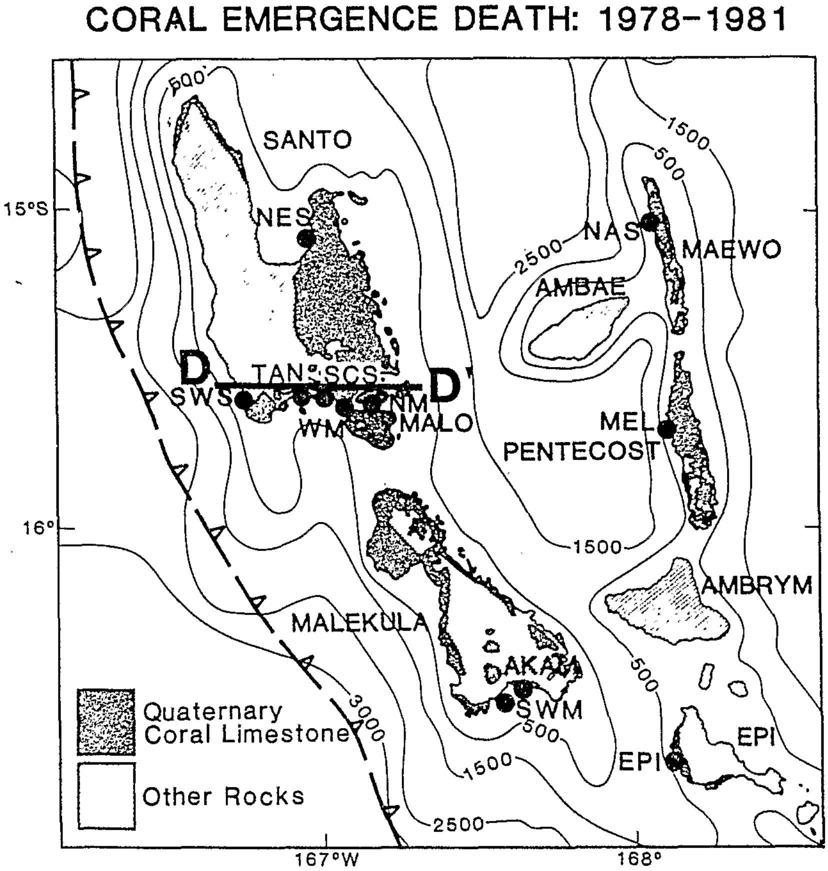


Fig. 16. Locations of coral emergence death from 1978 through 1981. A few corals at many locations record slight emergence during this interval, but in many areas, emergence of this age is absent. It is not clear why the timing of coral death is so variable, but the wide geographic distribution argues against a single tectonic cause. However, slight gradual uplift may contribute in some areas, especially Santo and southwest Malekula where Holocene uplift rates are rapid. D-D' is the location for a profile of these data (Figure 19).

from the Pleistocene reef terraces, *Taylor et al.* [1980] divided Santo and Malekula into at least four seismotectonic arc segments or blocks (Figure 3). Although there are faults and blocks within each segment, each is essentially coherent and separated from adjacent blocks by major faults or by significant changes in tilt directions [*Taylor et al.*, 1980]. Independent determinations of seismic rupture zone boundaries and barriers to rupture for the 1965 and 1973 earthquakes [*Pascal et al.*, 1978; *Ebel*, 1980; *Isacks et al.*, 1981; *Habermann*, 1984] coincide with the segment boundaries based on reef terraces.

The Holocene uplift history is more complete for central Vanuatu, and it is more relevant to contemporary tectonic movements. Very few data for this epoch have been presented previously, although we have numerous isotopic ages and have systematically mapped the emerged Holoene reefs (Figure 20).

In the Vanuatu area, sea level was at or near its present level by about 6000 years B.P. [e.g., *Baltzer*, 1970; *Thom and Chappell*, 1975], and it may even have reached a meter or two higher than its present level [*Clark et al.*, 1978]. Although the exact paleosea level for any particular time in the region is not well known or is controversial, emergence due to tectonic uplift is usually so great that 1 or 2 meters of uncertainty in Holocene sea level history is insignificant. Throughout this paper we have calculated uplift rates by the simple relationship:

$$\text{uplift rate} = (H - \text{PSL})/\text{age}$$

where H is the altitude of the coral sample above sea level, PSL is the paleosea level at the time when the coral lived relative to present sea level, and age is the isotopic age of the coral in years before present. For the period 0-6000 years B.P., we assume that PSL is the same as present sea level.

The anomalous topography of central Vanuatu appears to be due to inteaction of the DR and the arc. The Holocene uplift and tilt pattern is similar to that of the late Quaternary, but there are some important differences. *Jouannic et al.* [1980, 1982] and *Gilpin* [1982] observed that the Holocene uplift rates for most of Santo and north Malekula are twice the average rates for the past 100,000 years. We suspect that this increase is related to subduction of the DR because Holocene rates are not faster than Quaternary rates on the Torres Islands [*Taylor et al.*, 1985].

In late Quaternary time there was a change in the uplift pattern for south Malekula, as revealed by the isolated zone where there is at least 16 m of Holocene emergence (Figure 20). If the Holocene uplift rate here were extrapolated back in time, this large feature suggests that Pleistocene uplift should be represented here by a continuous series of reef terraces extending to high altitudes. However, while isolated patches of coral limestone at high levels indicate late Cenozoic uplift, there is no series of terraces indicating that this uplift was continuous until late Quaternary time (Figure 3). Emerged coral limestone occurs on the eastern coast of south Malekula, but it is quite weathered and does not indicate ongoing uplift. Indeed, the extremely wide modern reef flat and the fiordlike shape of Port Sandwich on the southeast coast (Figure 3) provide classic geomorphologic evidence of drowning due to tectonic subsidence. In summary, our observations suggest there was an early phase of uplift on south Malekula, followed by

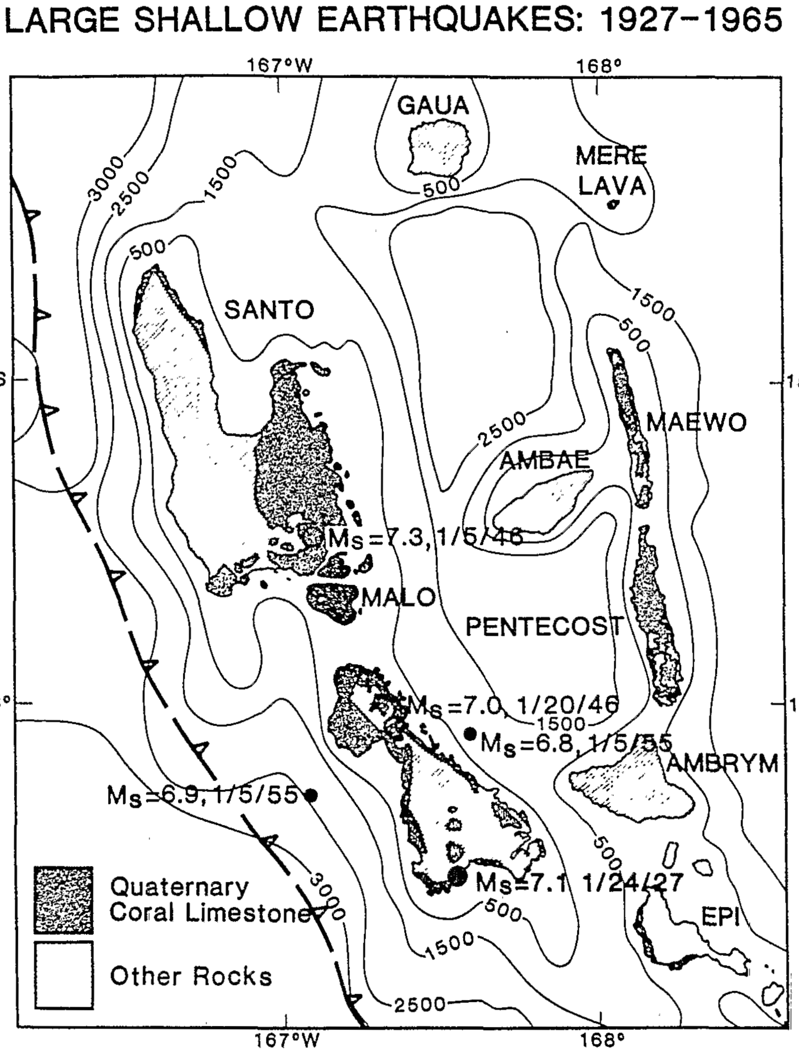


Fig. 17. Large shallow earthquakes from 1927 to 1965. Solid symbols for locations from *Gutenberg and Richter* [1954] and *Rothe* [1969]. These older events may have location errors of more than 100 km. Open circles indicate events relocated by *McCann* [1980]. Emergence of south Santo in 1945-1946 suggests coseismic rupture and uplift of the south Santo arc segment, probably during the January 5, 1946, earthquake. Corals show no indication of uplift associated with the two 1955 and the 1946 events near Malekula.

a period of relative stability and erosion. In late Quaternary time vertical deformation recommenced but with a quite different distribution of uplift and subsidence, which has only begun to alter the island morphology.

Discussion

Comparison of Contemporary Vertical Movements With Quaternary Deformation Patterns

The correlation among topography, Holocene reef emergence, and coral emergence of recent decades is remarkable (Figures 11, 19, 20, and 21). In general, coasts that have emerged Holocene reef terraces (Figure 20) also have recently emerged coral heads. The parts of south Santo and north Malekula having the most rapid Holocene uplift rates also uplifted most in 1965 (Figures 11 and 20). The tilt directions determined from Holocene reefs, and from reef terraces 200,000 years B.P. and younger, are similar to tilts imposed in 1965 [*Taylor et al.*, 1980]. It seems clear that the emerged Holocene reef represents an accumulation of discrete vertical movements such as we show to have occurred in recent decades. Indeed, the morphology and great topographic relief of Santo and north Malekula can be explained by vertical movements such as we document accumulated over roughly 500,000 years. Similarly, on Ambae and Ambrym, which do not have emerged reefs, we found no partially emerged corals and Ambae even has indications of contemporary submergence (Figure 6*c*).

These observations suggest that the pattern of vertical deformation has changed only slightly during late Quaternary time. *Taylor et al.* [1985] identified topographic features that suggest that the zones of most rapid uplift may have undergone a discrete westward shift on Santo and Malekula roughly 500,000 years ago. However, on north Santo the 1973 uplift pattern differs significantly from topography and the Holocene uplift pattern (Figure 21). Uplift occurred mainly along the northwest peninsula with hardly any uplift of the northeast peninsula even though it has emerged more than 15 m in Holocene time (Figures 14 and 20). To have the same proportion of 1973 to Holocene emergence as did northwest Santo, northeast Santo should have uplifted about 35 cm, instead of the observed 10 cm. This suggests that either the 1973 uplift pattern is not typical, or that additional slip will occur downdip from the 1973 rupture zone and produce more uplift of the northeast peninsula.

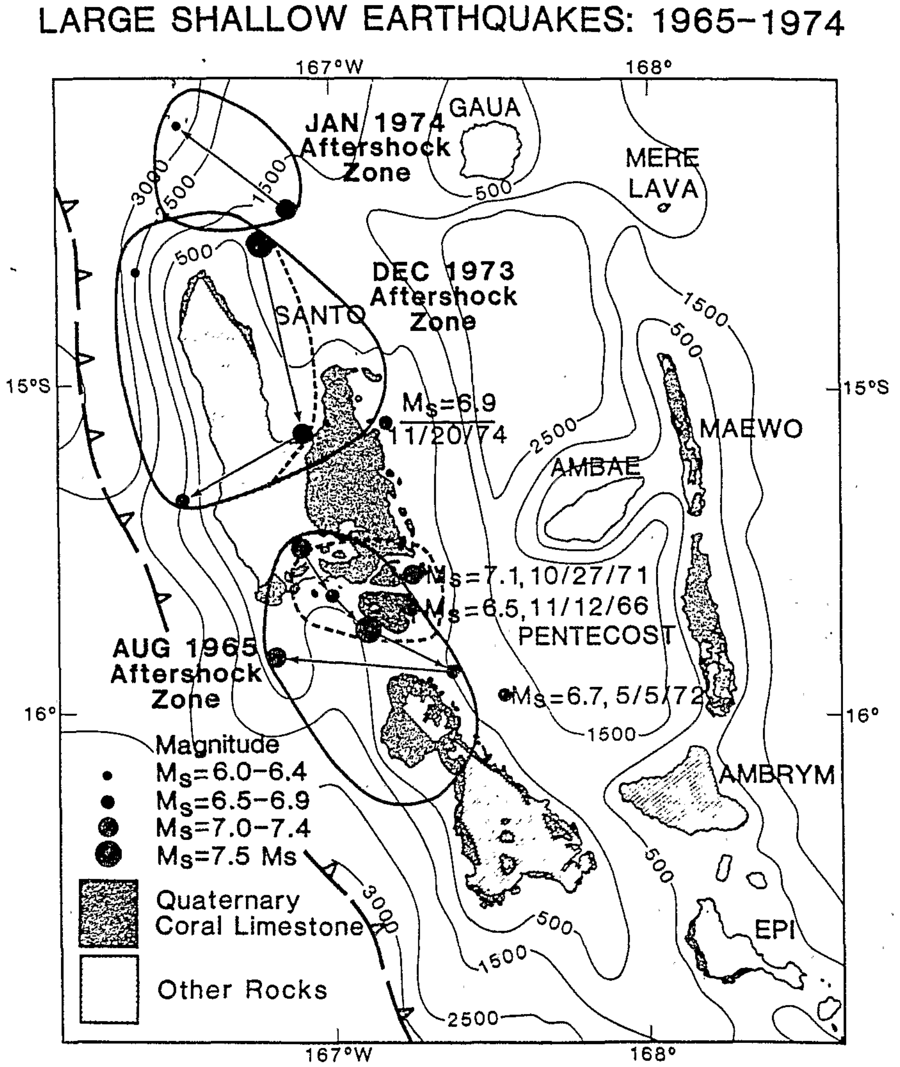


Fig. 18. Large shallow earthquakes since 1964. Solid lines show the rupture zones inferred from aftershocks [after *Isacks et al.*, 1981]. Other seismologists propose extending the aftershock zone to include southernmost Malekula [*McCann*, 1980; *Habermann*, 1984]. The October 27, 1971, event had few aftershocks to define a rupture surface, but we suggest the dashed rupture zone because north Malekula and central Santo have no uplifted corals of this age. We also suggest reducing the area of the 1973 rupture zone because of the distribution of emerged corals on the northeastern peninsula of Santo (Figure 14). Perhaps a future downdip earthquake will impose uplift mainly on the northeastern peninsula, just as the 1971 event uplifted the region to the east of the 1965 rupture.

South of the tilt discontinuity separating north and south Malekula the relationship between Holocene and recent emergence is more complex. The 35-45 cm of 1965 uplift along the central west coast (Figure 11) was very large in proportion to the 2-4 m of net Holocene emergence (Figure 20). For comparison, there was only about 1 m of 1965 uplift on north Malekula, where net Holocene emergence is about 20 m. However, on the central west coast of south Malekula, emerged 1965 coral heads show clear evidence of submergence in the years preceding 1965 emergence. If these movements are typical then this region (Figure 3) undergoes cycles of interseismic subsidence and coseismic uplift and accumulates negligible net uplift. In contrast, there is no evidence of preseismic submergence of north Malekula. At Southwest Bay, which is a narrow northwest trending graben at the southern end of the central west coast region, both Holocene and contemporary emergence decrease to zero.

South of Southwest Bay, recent coral emergence increases toward a small zone where there is up to 16 m of Holocene emergence (Figure 20). We do not know why this isolated area has uplifted while adjacent areas have not. Here, the 1965 emergence of only 11 cm is small compared with net Holocene emergence of at least 16 m. The 1965 emergence may not have been coseismic. In addition, nonseismic emergence occurring since 1965 has contributed an additional 11 cm per coral sampled. We suspect that this additional emergence was tectonic, even though, as explained in the methods section, its timing coincides with suspected regional sea level lows. To make the proportion of contemporary to Holocene uplift similar to that observed on north Malekula, the total nonseismic uplift on southwest Malekula will have to accumulate to 70 cm.

Along the west coasts of Epi, Maewo, and Pentecost, total Holocene emergence varies but averages about 5 m and suggests an average Holocene uplift rate of about 0.1 cm yr^{-1}. In comparison, emergence of living corals in recent years has been rather large (Table 1), particularly on Pentecost. Possibly, episodes of tectonic subsidence or quiescence occur between episodes of uplift. Otherwise, the total emergence rate of the past 20 years is about 10 times the average uplift rate.

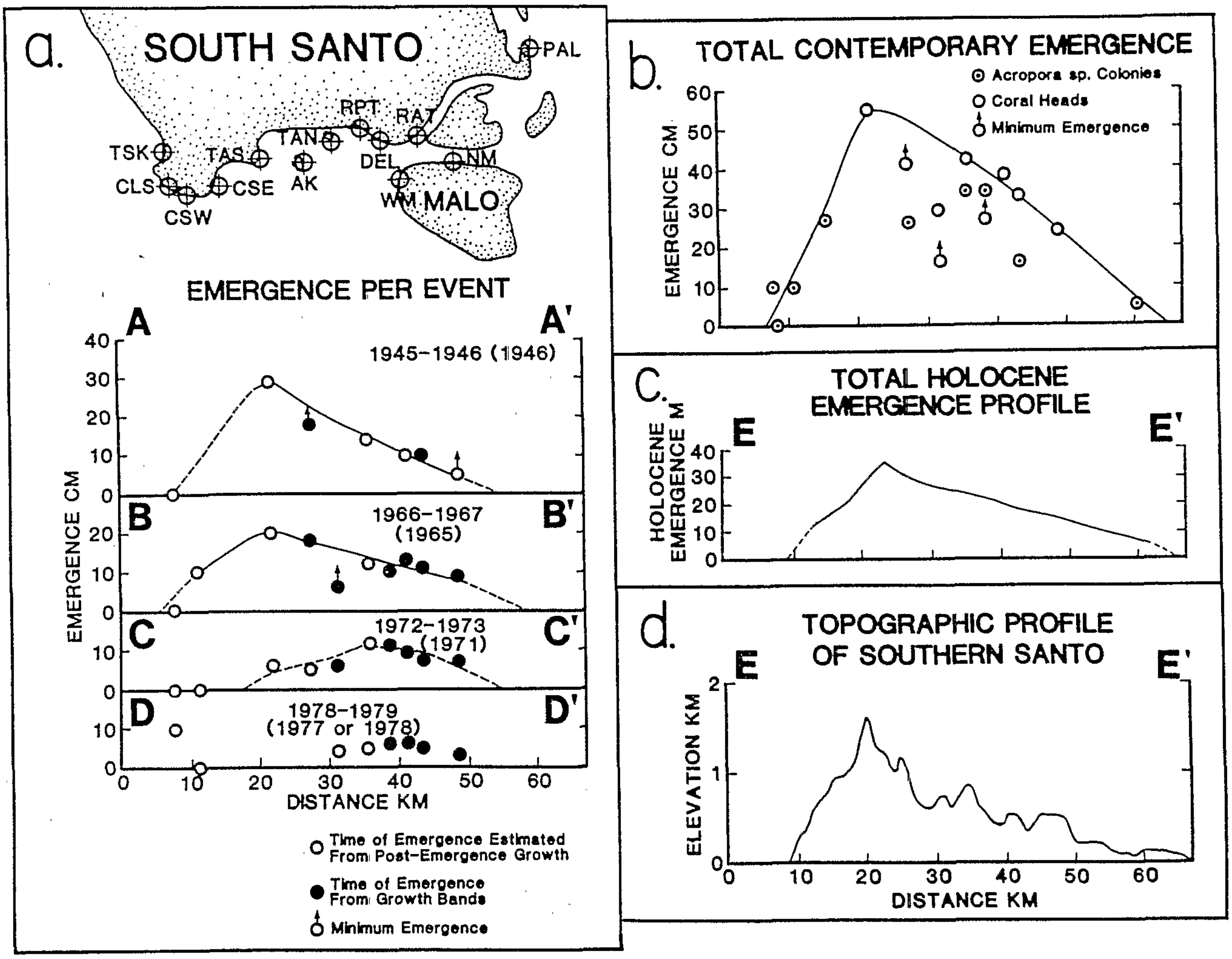


Fig. 19. Comparison of various east-west profiles across south Santo: (*a*) contemporary emergence for several periods projected along east-west lines shown in Figures 3, 11, 13, and 16; (b) sum of emergence for recent decades; (*c*) total Holocene emergence along line E-E' in Figure 20; (*d*) topography along line E-E'. The symbols show locations where measurements of emergence and, usually, coral growth band dates of emergence were determined (Tables 1, 2, and 3). Figure 19 *a* includes all emergence recorded by relatively well preserved corals. Total emergence recorded by *Acropora sp.* colonies (Table 3) is less because their fragile branching morphology is destroyed more quickly by bioerosion than that of massive coral heads. Thus *Acropora sp.* records only the 1965 and later events.

Subduction of the d'Entrecasteaux Ridge and Quaternary Vertical Deformation of the Central Vanuatu Arc

The Vanuatu arc has evolved in response to the subduction of the Indian plate beneath the Pacific plate, with a present relative velocity of about 10 cm yr^{-1} [*Dubois et al.*, 1977; *Minster and Jordan*, 1978] (Figure 1). However, the presence of sea floor spreading on the Fiji Plateau behind the arc [*Falvey*, 1975, 1978; *Malahoff et al.*, 1982] may make the total convergence rate several centimeters per year greater. The present arc-polarity geometry has prevailed since a post-middle Miocene arc-polarity reversal [*Mitchell*, 1971; *Chase*, 1971; *Karig and Mammerickx*, 1972; etc.].

Several peculiar features of the present-day Vanuatu arc may be caused by the subduction of the DR, a prominent bathymetric feature on the Indian plate on the west side of Santo and Malekula. There is no trench along the central part of the arc (Figure 1), and the western parts of Santo and Malekula seem to lie on the plate boundary, where the inner trench slope is normally located. Santo and Malekula may have migrated westward by opening of the Ambae basin [*Karig and Mammerickx* , 1972]. Alternatively, the extreme western position of emerged islands may reflect great uplift of a former inner trench slope in response to underthrusting of the DR [*Chung and Kanamori* , 1978*a*, *b*; *Taylor et al.*, 1985]. Two other unusual features of central Vanuatu include (1) the presence of the uplifted back arc islands of Pentecost and Maewo and (2) the Ambae (Aoba) basin centered on the volcanic chain (Figures 2 and 3). The absence of emerged coral limestone on the volcanic islands of Ambae and Ambrym indicates continuing subsidence in the Ambae basin. This contrasts with abundant emerged coral limestone recording uplift of the back arc and frontal arc islands. The Ambae basin, whose margins are defined by stepfaults and flexures, has formed since the post-middle Miocene arc-polarity reversal [*Ravenne et al.*, 1977; *Carney and Macfarlane*, 1980]. As the basin subsided, the islands on its eastern and western flanks uplifted. Several investigators have suggested that subsidence of the Ambae basin and uplift of its flanks are related to subduction of the DR [e.g, *Luyendyk et al.*, 1974; *Taylor et al.*, 1980; *Collot et al.*, 1985].

Contemporary and Holocene vertical movements documented here lend further support to the division of Santo and Malekula

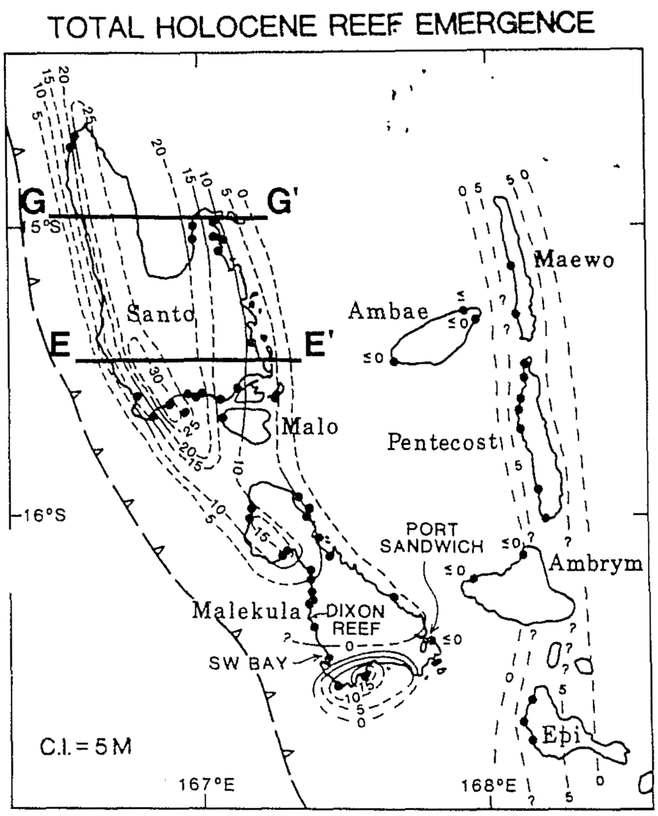


Fig. 20. Contour map of total Holocene reef emergence in meters. Solid circles indicate stations where the elevation of a Holocene reef surface was determined. Numerous ^{14}C and $^{230}Th/^{234}U$ ages document the Holocene age of the emerged reefs used for this figure [*Taylor et al.*, 1980; *Jouannic et al.*, 1980, 1982; *Gilpin*, 1982; *Urmos*, 1984; and unpublished data, 1987]. The absence of emerged coral reefs on Ambae and Ambrym is known from published reports as well as personal observations. The isolated zone of rapid Holocene uplift on southern Malekula apparently began in latest Quaternary time. Note the scarcity of coral limestone inland in this area on Figure 3. E-E' and G-G' are locations of topographic and total Holocene uplift profiles shown in Figures 19 and 21.

(Figure 3) into four seismotectonic blocks. As noted by *Taylor et al.* [1980], the segment boundaries between north and south Santo and north and south Malekul coincide roughly with the steep north and south flanks of the DR, respectively. In summary, the segment-by-segment characteristics in terms of contemporary and Holocene vertical movements are as follows:

1. In north Santo, the maximum total Holocene emergence of about 25 m (~0.4 cm yr^{-1}) compares with 60 cm of 1973 uplift (Figure 21). Coseismic uplift of northwest Santo between 1930 and 1973 is unlikely because of the absence of historical earthquakes in the area.

2. In south Santo, maximum Holocene uplift of about 35 m compares with about 20 cm of 1946 uplift, 25 cm of 1965 uplift, and perhaps a few centimeters of 1971 uplift (Figure 19).

3. In north Malekula, maximum Holocene emergence of about 20 m compares with 120 cm of 1965 coseismic uplift.

4. In south Malekula, maximum Holocene emergence of 2-4 m on the middle west coast compares with 35-45 cm of 1965 coseismic uplift, preceded by submergence. Farther south on southwest Malekula is an area with about 16 m of total Holocene emergence, and contemporary emergence of 10 cm or less in 1957-1958, 1966-1967, 1970, 1972, and 1978-1981, respectively.

From these observations we infer that each block has responded differently to underthrusting. North Santo and north Malekula appear to undergo infrequent coseismic uplifts which are large in proportion to total Holocene emergence. Uplift of north Malekula in 1965 mimicked the Holocene emergence pattern, but the 1973 uplift of north Santo was skewed to the west (Figures 19 and 21). South Santo appears to undergo more frequent seismic events, each having a maximum uplift of no more than 25 cm. Thus south Santo may achieve nearly twice the uplift rate of north Malekula with only 20% as much uplift per earthquake. Note that south Santo and north Malekula both

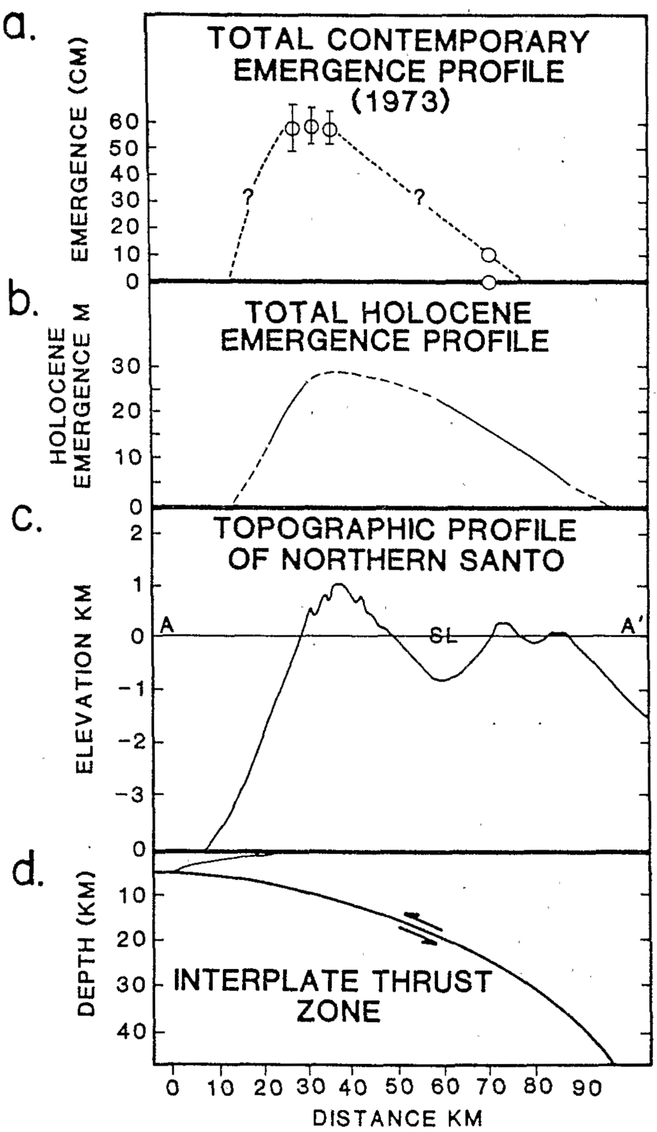


Fig. 21. Comparison of total contemporary (all December 1973) emergence in north Santo (a) along profile F-F' of Figure 14 with (b) total Holocene emergence along line G-G' of Figure 20; and (c) an east-west topographic profile across north Santo (also line F-F' of Figure 14). The interplate thrust zone (d) beneath Santo is defined by carefully relocated hypocenters [*Chinn and Isacks*, 1983]. Note that these profiles are disimilar, unlike the south Santo profiles. We assume a steady decrease in total Holocene emergence across the bay on Figure 20 (dashed center part of profile in Figure 21*b*), but the existence of the deep bay could indicate a zone of slower uplift. The data for the the contemporary emergence profile are relatively few, but they indicate that most of the December 1973 uplift is in the extreme west and decreases rapidly toward the northeast peninsula. Thus the northeastern peninsula would require about 30 cm of additional uplift to make the proportions of Holocene and recent emergence similar on both the northwest and northeast peninsulas.

uplifted during the 1965 earthquakes but by a quite different amount. On south Malekula, the western coast underwent pre-1965 nonseismic subsidence, then uplifted about 40 cm in areas where total Holocene emergence is almost nil. Yet in the isolated zone of uplift farther south on Malekula where there has been 16 m of Holocene emergence, the 1965 uplift is not more than 10 cm.

Whereas earthquakes released elastic strain for the north Santo block in 1973 and for the south Santo and north Malekula blocks in 1965, on south Malekula the paucity of 1965 aftershocks and the relationship between Holocene and recent uplifts suggests there is a different type of slip regime. Perhaps interplate coupling is relatively weak beneath south Malekula. When south Santo and north Malekula ruptured in 1965, interplate slip may have propagated southward beneath south Malekula in part because of weak coupling and lack of resistance to underthrusting. Continuing uplift of south Malekula may be due to continuing aseismic slip and adjustment to the 1965 coseismic slip of the adjacent arc segment. On the remaining three blocks, all documented vertical movements have occurred at or very near the times of large earthquakes. However, our data cannot completely disprove the existence of important nonseismic vertical movements there. If the 1973 and 1965 uplifts are typical, north Santo and north Malekula must either have long seismic recurrence intervals or undergo interseismic subsidence. However, subsidence would have to occur rapidly within a period of 1-3 years prior to coseismic uplift or the corals would record evidence of preseismic submergence.

Although investigators agree that convergence of the DR profoundly affects the vertical tectonics of Vanuatu, there is not total agreement about the mechanics of the interaction process. Factors affecting the influence of the DR include its topographic relief, the direction and rate of underthrusting, and the relatively low rate of migration of its intersection with the plate boundary, only a few centimeters per year. All models agree that the absence of a physiographic trench, the vertical deformation pattern, the arc segment boundaries, and the absence of intermediate depth seismicity beneath central Vanuatu [*Marthelot et al.*, 1985] can be attributed to the subduction of the DR. One class of proposed models explains these features in terms of the flexure of an elastic plate having a single fault along the western edge of the arc. For example, the model of *Chung and Kanamori* [1978*a*, *b*] assume that uplift of Santo and Malekula is due to upward loading of the leading edge of the upper plate by the DR, with the resulting plate flexure and isostatic effects causing the subsidence of the Ambae basin. *Chung and Kanamori* [1978*b*] do not mention that plate models will also predict a secondary zone of uplift farther east of the zone of subsidence (Figure 1). *Moretti and Ngokwey* [1985] also apply a plate model, although they allow for the possibility that horizontal forces prevent the plate from achieving isostatic equilibrium.

A significant uncertainty concerns the timing of the uplifts in the back arc relative to the time of occurrence of major earthquakes. Over hours or days, the mantle is elastic, whereas over long periods it is viscoelastic. Thus the initial response to motion on a lithospheric fault will be like that of an elastic halfspace, and then subsequently the mantle beneath the lithosphere will flow to adjust to the redistribution of mass occurring in the earthquake [*Savage and Gu*, 1985]. While *Taylor et al.* [1980] could explain the uplift pattern in the fore arc with a faulted elastic half-space model, such models seriously underestimate the amount of uplift in the back arc (Figure 1). However, while the elastic plate/viscoelastic mantle models can explain the deformation in the back arc, there is considerable uncertainty about the characteristic time for the viscoelastic deformation to take place. Estimates for this characteristic time range from 1 to 100 years, depending on the viscosity of the mantle. In Vanuatu, the fact that the uplift of Pentecost and Maewo apparently occurred near the time of the 1965 and 1973 frontal arc uplifts suggests a direct link between fore arc and back arc tectonism. This implies that the characteristic mantle relaxation time beneath Vanuatu is very short, perhaps a year or less.

A second class of models requires at least two faults to explain the observed deformation. *Isacks et al.* [1981] and *Collot et al.* [1985] propose that horizontal loading of the upper plate due to impingement of the DR against the arc has caused faulting and uplift in the back arc. This view is supported by focal mechanisms of four recent shallow earthquakes occurring along the Pentecost-Maewo trend, reported by *Marthelot et al.*, [1985]. These events indicate compressional deformation in the upper plate near the back arc uplift that is consistent with reverse faulting on west-northwest trending steeply dipping nodal planes. The absence of the Coriolis trough rift system east of Santo and Malekula [*Dubois et al.*, 1977] also suggests compression that contrasts with arc segments to the north and south where the extensional Coriolis rift system exists.

In some models the amount of vertical deformation of the upper plate is independent of whether interplate thrusting is dominantly seismic or aseismic [e.g., *Cloos*, 1982]. However, others require strong coupling, implying seismic slip for significant vertical deformation of the upper plate [e.g., *Dewey*, 1980]. The area above the DR has the largest seismic stress drops of the arc [*Wyss et al.*, 1983; *Habermann*, 1984], and we have concluded that most uplift of this area is coseismic. There may be ongoing aseismic uplift of south Malekula, but this is south of the intersection of the DR. In any case, the upward displacement occurs as the upper plate rides over buoyant underthrusting topography. This may include an "underplating" of material to the base of the upper plate, or some other process that alters the structure of the upper plate.

Clearly, seismic slip and high levels of shallow seismic activity accompany at least part of the underthrusting of the DR beneath Vanuatu. However, along the Tonga-Kermadec arc there is much less shallow seismic activity associated with underthrusting of the Louisville ridge and the rest of the Pacific plate. This indicates weaker interplate coupling than in Vanuatu and suggests that the Tonga-Kemadec upper plate may not respond to aseismic ridge subduction as strongly as in Vanuatu. Absence of late Quaternary uplift of the northern Tonga frontal arc [*Taylor and Bloom*, 1977; *Taylor*, 1978] is consistent with the low level of shallow seismicity. However, the Louisville ridge does appear to have modified the structure of the upper plate that it has underthrust [*Dupont and Herzer*, 1985].

Evidence for Aseismic Tectonic Uplift

At least some of the minor emergence events not associated with large earthquakes probably are a result of aseismic tectonic uplift. There is indisputable evidence from witnesses and corals that coseismic uplift occurred during the 1965 sequence of large shallow earthquakes and strong circumstantial evidence that uplift occurred near the times and epicenters of large earthquakes occurring in 1946, 1971, and 1973 (Figure 4). Although some of the remaining uplift events may be influenced by El Niños-type sea level variations, these variations are so common [*Quinn et al.*, 1978] that they are especially likely to influence the time of death of shallow-living corals undergoing small tectonic uplifts.

For several reasons, we suggest that most of the remaining events are of tectonic origin:

1. These events occur only in areas that have a history of rapid uplift.

2. Most of the emergence has persisted. For example, corals that partially died in 1965-1966 and 1969-1970 on southwest Malekula remained emerged. There was still further emergence in 1972, so that many corals in this zone of rapid Holocene uplift are mostly dead because of the series of cumulative emergence events (Figure 6*b*).

3. From one area to the next there is little synchrony among the times of emergence death of corals. An El Niños-type event will cause similar sea level variations over larger areas than we are considering [e.g., *Cane*, 1983; *Wyrtki*, 1985].

4. We observed no emergence death for any corals on the subsiding island of Ambae.

The observations of coral emergence in central Vanuatu suggest that for the area as a whole a third or more of all the interplate motion occurs as aseismic slip. We obtain this estimate by comparing the total mean emergence (353 cm), as determined by adding all the entries in the last column of Table 1, to the aseismic emergence (103 cm), as determined by adding those entries for events inferred for years other than 1946, 1965, 1971, or 1973. This estimate is rather crude and is probably somewhat too low, as it would not include aseismic uplift which occurred during the days or months following an earthquake.

There is evidence from other plate margins that some interplate motion occurs as aseismic slip. For example, for the Kuriles and northern Japan, *Kanamori* [1977] compared plate convergence rates to coseismic slips determined from the moment of great earthquakes and concluded that seismic slip made up only about one quarter of the total slip. Elsewhere, *Brown et al.* [1977] evaluated leveling surveys undertaken between 1966 and 1975 in the vicinity of the 1964 Alaska earthquake, and found indirect evidence suggesting that aseismic creep occurred on the downdip extension of the fault plane. In South America, near the border of Ecuador and Peru where no great earthquake has occurred for at least 400 years [*Kelleher*, 1972], there exist marine terraces up to 300 m above sea level, indicative of continuing Quaternary uplift. In North America there are direct measurements showing that aseismic creep occurs along the strike-slip margin in California [e.g., *Burford and Harsh*, 1980; *Schulz et al.*, 1982; *Louie et al.*, 1985].

Finally, for theoretical reasons it is plausible that asthenospheric flow would contribute a long-term response to seismic motion on fault planes, even if all the motion on the fault plane itself were of seismic origin. As discussed previously, the mantle is an elastic solid over time periods of seconds appropriate to seismic slip and a viscoelastic material over time periods of years [*Savage and Gu*, 1985]. Thus if the time between events is much longer than the characteristic relaxation time of the mantle, we would expect to see aseismic deformation associated with particular seismic events. Otherwise, we would see slow and steady adjustment not attributable to particular earthquakes. The fact that we observe distinct regional episodes of uplift, sometimes in the same year as major earthquakes, suggests that the mantle relaxation time is relatively short.

CONCLUSIONS

1. In tropical regions, analysis of emerged coral limestone and the growth patterns of living corals provide valuable information about tectonic motions over periods from 1 year to many thousands of years. One can accurately determine the zone of deformation for major earthquakes years after they occur without deploying or maintaining instruments and without ever visiting the region prior to the earthquakes. Mapping and selectively dating older exposed corals allows one to compare contemporary vertical movements with those occurring throughout the Quaternary.

2. The coral method confirms that significant uplift occurred in association with earthquakes occurring in central Vanuatu in 1946, 1965, 1971, and 1973. For the 1965 earthquake the vertical deformation is remarkably similar to that observed for the great 1960 Chile and 1964 Alaska earthquakes. A peculiar feature of the Chile, Alaska, and 1965 Vanuatu events is a zone of secondary uplift occurring approximately 100-200 km arcward from the primary uplift zone. For the 1965 Vanuatu event the approximately 10 cm of secondary uplift exceeds the uplift predicted by models having a single fault in an elastic half-space. The secondary uplift can be explained by a model with faulting in an elastic plate overlying a viscoelastic half-space. However, in this case the data suggest that the characteristic relaxation time of the mantle is of the order of a year or less.

3. The coral method strongly suggests that aseismic vertical motions occur regularly in parts of the central Vanuatu arc and constitute a significant proportion of the total accumulated uplift. Oceanographic effects such as El Niños affect when the corals die and record these motions. However, because the coral death is persistent and does not occur simultaneously throughout the entire central Vanuatu arc, it cannot be solely of oceanographic origin.

4. There is generally remarkable agreement between the gross topography of the entire central Vanuatu region, the pattern of uplift that has taken place since 1946, the pattern accumulated over the past 6000 years of Holocene time, and late Pleistocene uplift and tilt patterns. Thus we infer that the same tectonic processes have been shaping the arc throughout most of the latter half of the Quaternary. In Vanuatu and elsewhere, geomorphological analysis of the topography may provide important information about the location and geographic extent of future earthquakes.

5. In central Vanuatu there is some evidence that during late Quaternary time the d'Entrecasteaux ridge began to control the topography and deformation. Where the ridge meets the arc, the uplift rate is higher, and there is little or no evidence of any subsidence following the uplifts associated with major earthquakes.

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REFERENCES

Baltzer, M. F., Datation absolue de la transgression holocene sur la cote ouest de Nouvelle-Caledonie sur des echantillons de tourbes a paletuviers: Interpretation tectonique, *C. R. Hebd. Seances Acad. Sci. Paris, 271*, 2251-2254, 1970.

Beck, S. L., and L. J. Ruff, The rupture process of the 1976 Mindanao earthquake, *J. Geophys. Res., 90*, 6673-6782, 1985.

Benoit, M., and J. Dubois, The earthquake swarm in the New Hebrides archipelago, August, 1965, Recent Crustal Movements, edited by B. W. Collins and R. Fraser, *Bull. 9, R. Soc. N. Z.*, pp. 141-148, 1971.

Bevis, M., and B. L. Isacks, Leveling arrays as multicomponent tiltmeters: Slow deformation in the New Hebrides Island arc, *J. Geophys. Res., 86*, 7808-7824, 1981.

Brown, L. D., R. E. Reilinger, S. R. Holdahl, and E. I. Balazs, Postseismic crustal uplift near Anchorage, Alaska, *J. Geophys. Res., 82*, 3369-3378, 1977.

Buddemeir, R. W., Coral growth: Retrospective analysis, in *Coral Reefs: Research Methods, UNESCO Monographs on Oceanographic Methodology*, edited by D. R. Stoddart and R. E. Johannes, pp. 551-571, Page Brothers, Norwich, England, 1978.

Burford, R. O. and P. W. Harsh, Slip on the San Andreas fault in central California from alinement array surveys, *Bull. Seismol. Soc. Am., 70*, 1233-1261, 1980.

Buskirk, R., F. W. Taylor, W. P. O'Brien, P. Maillet, and L. Gilpin, Seasonal growth patterns and mortality of corals in the New Hebrides (Vanuatu), in *Proceedings of the Fourth International Coral Reef Symposium, Manila*, vol. 2, pp. 197-200, Marine Sciences Center, Univ. Philippines, Quezon City, Philippines, 1981.

Cane, M. A., Oceanographic events during El Niño: *Science, 222*, 1189-1195, 1983.

Carney, J., and A. Macfarlane, A sedimentary basin in the central New Hebrides, *Tech. Bull. U. N. Econ. Soc. Comm. Asia Pac. Comm. Co-ord. Jt. Prospect. Miner. Resour. South Pac. Offshore Area, 3*, 109-120, 1980.

Chase, C. G., Tectonic histoy of the Fiji Plateau, *Geol. Soc. Am. Bull., 82*, 3087-3110, 1971.

Chinn, D., and B. L. Isacks, Accurate source depths and focal mechanisms of shallow earthquakes in western South America and in the New Hebrides island arc, *Tectonics, 2*, 529-563, 1983.

Chung, W. Y., and H. Kanamori, Subduction process of a fracture zone and aseismic ridges-The focal mechanism and source characteristics of the New Hebrides earthquake of January 19, 1969, and some related events, *Geophys. J. R. Astron. Soc., 54*, 221-240, 1978*a*.

Chung, W. Y., and H. Kanamori, A mechanical model for plate deformation associated with aseismic ridge subduction in the New Hebrides arc, *Tectonophysics, 50*, 29-40, 1978*b*.

Clark, J. A., W. E. Farrell, and W. R. Peltier, Global changes in postglacial sea level: A numerical calculation, *Quat. Res., 9*, 275-278, 1978.

Cloos, M., Flow melanges: Numerical modeling and geologic constraints on their origin in the Franciscan subduction complex, *Geol. Soc. Am. Bull., 93*, 330-345, 1982.

Collot, J. Y., J. Daniel, and R. V. Burne, Recent tectonics associated with the subduction/collision of the d'Entrecasteaux zone in the central New Hebrides, *Tectonophysics, 112*, 325-356, 1985.

Dewey, J. F., Episodicity, sequence, and style at convergent plate boundaries, The Continental Crust and Its Mineral Deposits, edited by D. W. Strangway, *Spec. Pap. Geol. Assoc. Can. 20*, 553-573, 1980.

Dodge, R. E., Preparation and examination of Skeletal material for growth studies, part B, Corals, Appendix, in *Skeletal Growth: Biological Records of Environment Change*, edited by D. C. Rhoads and R. A. Lutz, pp. 615-618, Plenum, New York, 1980.

Dubois, J., J. Launay, J. Recy and J. Marshall, New Hebrides trench: Subduction rate from associated lithospheric bulge, *Can. J. Earth Sci., 14*, 250-255, 1977.

Dupont, J. R., and R. H. Herzer, Effect of subduction of the Louisville ridge on the structure and morphology of the Tonga arc, in *Geology and Offshore Resources of the Pacific Island Arcs-Tonga Region, Earth Sci. Ser.*, vol. 2, edited by D. W. Scholl and T. L. Vallier, pp. 323-332, Circum-Pacific Council for Energy and Mineral Resources, Los Angeles, Calif., 1985.

Ebel, J. E., Source processes of the 1965 New Hebrides Islands earthquakes inferred from teleseismic wave forms, *Geophys. J. R. Astron. Soc., 63*, 381-403, 1980.

Falvey, D. A., Arc reversals and a tectonic model for the North Fiji Basin, *Bull. Aust. Soc. Explor. Geophys., 6*, 47-49, 1975.

Falvey, D. A., Analysis of paleomagnetic data from the New Hebrides, *Bull. Aust. Soc. Explor. Geophys.*, 9, 117-123, 1978.

Fitch, T. J., and C. H. Scholz, Mechanisms of underthrusting in southwest Japan: A model of convergent plate interactions, *J. Geophys. Res., 76*, 7260-7292, 1971.

Gilpin, L., Tectonic geomorphology of Santo Island, Vanuatu, M.S. thesis, 147 pp., Cornell Univ., Ithaca, N. Y., 1982.

Gutenberg, B., and C. F. Richter, *Seismicity of the Earth*, 2nd ed., pp. 161-234, Princeton University Press, Princeton, N. J., 1954.

Habermann, R. E., Spatial seismicity variations and asperities in the New Hebrides seismic zone, *J. Geophys. Res., 89*, 5891-5903, 1984.

Hydrographer of the Navy, Admiralty tide tables 1977, Pacific Ocean and adjacent seas, vol. 3, 484 pp., location, 1976.

Isacks, B. L., R. K. Cardwell, J. L. Chatelain, M. Barazangi, J. M. Marthelot, D. Chinn, and R. Louat, Seismicity and tectonics of the central New Hebrides Island Arc, in *Earthquake Prediction: An International Review, Maurice Ewing Ser., vol. 4*, edited by D. W. Simpson and P. G. Richards, pp. 93-116, AGU, Washington, D. C., 1981.

Jouannic, C., F. W. Taylor, A. L. Bloom, and M. Bernat, Late Quaternary uplift history from emerged reef terraces on Santo and Malekula Islands, central New Hebrides island arc, *Tech. Bull. U. N. Econ. Soc. Comm. Asia Pac. Comm. Co-ord. Jt. Prospect. Miner. Resour. South Pac. Offshore Areas, 3*, 91-108, 1980.

Jouannic, C., F. W. Taylor, and A. L. Bloom, Sur la surrection et la deformation d'un arc jeune: L'arc des Nouvelles Hebrides, *Trav. Doc. ORSTOM, 147*, 223-246, 1982.

Kanamori, H., Seismic and aseismic slip along subduction zones and their tectonic implications, in *Island Arcs, Deep Sea Trenchs, and Back-Arc Basins, Maurice Ewing Ser.*, vol. 1, edited by M. Talwani, C. G. Harrison, and D. E. Hayes, pp. 163-174, AGU, Washington, D. C., 1977.

Karig, D. E., and J. Mammerickx, Tectonic framework of the New Hebrides island arc, *Mar. Geol., 12*, 187-205, 1972.

Kelleher, J. A., Rupture zones of large South American earthquakes and some predictions, *J. Geophys. Res., 77*, 2087-2103, 1972.

Knutson, D. W., R. W. Buddemeir, and S. V. Smith, Coral chronometers: Seasonal growth bands in reef corals, *Science, 177*, 270-272, 1972.

Kroenke, L. W., C. Jouannic, and P. Woodward, Bathymetry of the southwest Pacific, *Geophysical Atlas of the Southwest Pacific*, chart 1, *UN IGCP 110*, United Nations ESCAP, New York, 1983.

Launay, J., and J. Recy, Variations relatives du niveau de la mer et neotectonique en Nouvelle Caledonie au Pleistocene superior et a l'Holocene, *Rev. Geogr. Phys. Geol. Dyn., 14*, 47-66, 1972.

Louie, J. N., C. R. Allen, D. C. Johnson, P. C. Haase, and S. N. Cohn, Fault slip in southern California, *Bull. Seismol. Soc. Am., 75*, 811-833, 1985.

Luyendyk, B., W. B. Bryan, and P. A. Jezek, Shallow structure of the New Hebrides island arc, *Geol. Soc. Am. Bull., 85*, 1287-1300, 1974.

Malahoff, A., R. H. Feden, and H. S. Fleming, Magnetic anomaly and tectonic fabric of marginal basins north of New Zealand, *J. Geophys. Res., 87*, 4109-4125, 1982.

Mallick, D. I. J., and D. Greenbaum, Geology of southern Santo, regional report, 67 pp., Geol. Surv., Vila, New Hebrides, 1977.

Marthelot, J. M., E. Coudert, and B. L. Isacks, Tidal tilt from localized ocean loading in the New Hebrides arc, *Bull. Seismol. Soc. Am., 70*, 283-292, 1980.

Marthelot, J. M., J. L. Chatelain, B. L. Isacks, R. K. Cardwell, and E. Coudert, Seismicity and attenuation in the central Vanuatu (New Hebrides) Islands: A new interpretation of the effect of subduction of the d'Entrecasteaux fracture zone, *J. Geophys. Res., 90*, 8641-8650, 1985.

Matsuda, T., Y. Ota, M. Ando, and N. Yonekura, Fault mechanism and recurrence time of major earthquakes in southern Kanto district, Japan, as deduced from coastal terrace data, *Geol. Soc. Am. Bull., 89*, 1610-1618, 1978.

McCann, W. R., Large and moderate size earthquakes: Their relationship to the tectonics of subduction, Ph.D. dissertation, Columbia Univ., New York, 1980.

Meyers, G., Interannual variation in sea level near Truk Island--A bimodal seasonal cycle, *J. Phys. Oceanogr.*, *12*, 161-168, 1982.

Minster, J. B., and T. H. Jordan, Present-day plate motions, *J. Geophys. Res.*, *83*, 5331-5354, 1978.

Mitchell, A. H. G., Raised reef-capped terraces and Plio-Pleistocene sea level changes, north Malekula, New Hebrides, *J. Geol.*, *77*, 56-67, 1968.

Mitchell, A. H. G., Geology of north Malekula, regional report, 56 pp., Geol. Surv., Vila, New Hebrides, 1971.

Moretti, I., and K. Ngokwey, Aseismic ridge subduction and vertical motion of overriding plate, *Symposium Geodynamique des Caraibes, Paris, 5-8 Fevrier 1985*, pp. 245-252, Editions Technip, Paris, 1985.

Neef, G., and H. H. Vech, Uranium series ages and late Quaternary uplift in the New Hebrides, *Nature*, *269*, 682-683, 1977.

Pascal, G., B. L. Isacks, M. Barazangi, and J. Dubois, Precise relocations of earthquakes and seismotectonics of the New Hebrides island arc, *J. Geophys. Res.*, *83*, 4957-4973, 1978.

Plafker, G., Alaskan earthquake of 1964, and Chilean earthquake of 1960: Implications for arc tectonics, *J. Geophys. Res.*, *77*, 901-925, 1972.

Plafker, G., and J. C. Savage, Mechanism of the Chilean earthquakes of May 21 and 22, 1960, *Geol. Soc. Am. Bull.*, *81*, 1001-1030, 1970.

Quinn, W. H., D. O. Zopf, E. S. Short, and R. T. W. Kuo Yang, Historical trends and statistics of the southern oscillation, El Niño, and Indonesian droughts, *Fish. Bull.*, *76*, 663-678, 1978.

Ravenne, C., G. Pascal, J. Dubois, F. Dugas, and L. Montadert, Model of a young intra-oceanic arc: The New Hebrides island arc, in *Geodynamics in the South-West Pacific*, pp. 63-78, Editions Technip, Paris, 1977.

Robinson, G. P., The geology of North Santo, regional report, 77 pp., Geol. Surv., Vila, New Hebrides, 1969.

Rothe, J. P., *The Seismicity of the Earth 1953-1965, Earth Sci.*, vol. 1, 336 pp., UNESCO, New York, 1969.

Savage, J. C., and G. Gu, A plate flexure approximation to postseismic and interseismic deformation, *J. Geophys. Res.*, *90*, 8570-8580, 1985.

Schulz, S. S., G. M. Mavko, R. O. Burford, and W. D. Stuart, Long-term fault creep observations in central California, *J. Geophys. Res.*, *87*, 6977-6982, 1982.

Scoffin, T. P., and D. R. Stoddart, The nature and significance of microatolls, *Philos. Trans. R. Soc. London, Ser. B*, *284*, 99-122, 1978.

Shreve, R. L., and M. Cloos, Dynamics of sediment subduction, melange formation, and prism accretion, *J. Geophys. Res.* , *91*, 10,229-10,245, 1986.

Stoddart, D. R., and T. P. Scoffin, Microatolls: Review of form, origin, and terminology, *Atoll Res. Bull.*, *228*, 17 pp., 1979.

Taylor, F. W., Quaternary tectonic and sea level history, Tonga and Fiji, Southwest Pacific, Ph. D. dissertation, Cornell Univ., Ithaca, N. Y., 1978.

Taylor, F. W. and A. L. Bloom, Coral reefs on tectonic blocks, Tonga island arc, in *Proceedings of the Third International Coral Reef Symposium, vol. 2*, pp. 275- 281, publisher, location, 1977.

Taylor, F. W., R. G. Fairbanks, C. Jouannic, and A. L. Bloom, Sclerochronolgy: A new method for absolute dating of paleoseismic uplift (abstract), *Eos Trans. AGU*, *60*, 384, 1979.

Taylor, F. W., B. L. Isacks, C. Jouannic, A. L. Bloom, and J. Dubois, Coseismic and Quaternary vertical tectonic movements, Santo and Malekula Islands, New Hebrides island arc, *J. Geophys. Res.*, *85*, 5367-5381, 1980.

Taylor, F. W., C. Jouannic, L. Gilpin, and A. L. Bloom, Coral colonies as monitors of change in relative level of the land and sea: applications to vertical tectonism, in *Proceedings of the Fourth International Coral Reef Symposium, Manila, vol. 2*, pp. 485-492, Marine Science Center, Univ. Philippines, Quezon City, Philippines, 1981.

Taylor, F. W., C. Jouannic, and A. L. Bloom, Quaternary uplift history of the Torres Islands, northern New Hebrides frontal arc: Comparison with Santo and Malekula Islands, Central New Hebrides frontal arc, *J. Geol.*, *93*, 419-438, 1985.

Thatcher, W., The earthquake deformation cycle at the Nankai trough, southwest Japan, *J. Geophys. Res.*, *89*, 3087-3101, 1984*a*.

Thatcher, W., The earthquake deformation cycle, recurrence and the time-predictable model, *J. Geophys. Res.*, *89*, 5674-5680, 1984*b*.

Thom, B. G., and J. Chappell, Holocene sea levels relative to Australia, *Search*, *6*, 90-93, 1975.

Urmos, J., Oxygen isotopes, sea levels, and uplift of reef terraces, Araki Island, Vanuatu, M. S. thesis, 123 pp., Cornell Univ., Ithaca, N. Y., 1985.

Uyeda, S., Subduction zones: An introduction to comparative subductology, *Tectonophysics*, *81*, 133-159, 1982.

Uyeda, S., and H. Kanamori, Back-arc opening and mode of subduction, *J. Geophys. Res.*, *84*, 1049-1061, 1979.

White, W., K. Hasunuma, and G. Meyers, Large-scale secular trend in steric sea level over the western North Pacific from 1954-1974, *J. Geod. Soc. Jpn.*, *25*, 49-55, 1979.

Wyrtki, K., Sea level during the 1972 El Niño, *J. Phys. Oceanogr.*, *8* , 779-787, 1977.

Wyrtki, K., The response of sea surface topography to the 1976 El Niño, *J. Phys. Oceanogr.*, *9*, 1224-1231, 1979.

Wyrtki, K., Sea level fluctuations in the Pacific during the 1982-83 El Niño, *Geophys. Res. Lett.*, *12*, 125-128, 1985.

Wyss, M., R. E. Habermann, and C. Heiniger, Seismic quiescence, stress drops, and asperities in the New Hebrides arc, *Bull. Seismol. Soc. Am.*, *73*, 219-236, 1983.

Yonekura, N., Quaternary tectonic monuments in the outer arc of southwest Japan with special reference to seismic crustal deformations, *Bull. Dep. Geogr. Univ. Tokyo*, *7*, 19-71, 1975.

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