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ABSTRACT

Considerable uncertainty exists as to whether the last interglacial was relatively "short" (~10 ka) or "long" (~20–60 ka), although most investigators generally agree that the last interglacial correlates with all or part of deep-sea oxygen-isotope stage 5. A compilation of reliable U-series ages of marine terrace corals from deposits that have been correlated with isotope stage 5 indicates that there were three relatively high sea-level stands at ca 125–120 ka, ca. 105 ka, and ca. 85–80 ka, and these ages agree with the times of high sea level predicted by the Milankovitch orbital-forcing theory. At a number of localities, however, there are apparently reliable coral ages of ca. 145–135 ka and ca. 70 ka, and the Milankovitch theory would not predict high sea levels at these times. These ages are at present unexplained and require further study.

The issue of whether the last interglacial was "short" or "long"can be addressed by examining the evidence for how high sea level was during the stands at ca. 125 ka, ca. 105 ka, and ca. 80 ka, because sea level is inversely proportional to global ice volume. In tectonically stable areas such as Bermuda, the Bahamas, the Yucatan peninsula, and Florida, there is clear evidence that sea level at ca. 125 ka was +3 to +10 m higher than present. During the ca. 105 ka and ca. 80 ka high sea-level stands, there is conflicting evidence for how high sea levels were. Studies of uplifted terraces on Barbados and Haiti and most studies of terraces on New Guinea indicate sea levels considerably lower than present. Studies of the terraces and deposits on the east and west coasts of North America, Bermuda, and the Bahamas, however, indicate sea levels close to, or only slightly below, the present at these times. Thus, data from Barbados, Haiti, and New Guinea indicate a "short" last interglacial centering ca. 125 ka, but data from the other localities indicate that sea level was high during much of the period from 125 to 80 ka, and that there were two minor ice advances in that period.

If it is accepted that the last interglacial period was relatively "long" and ended sometime after ca. 80 ka, then coastal deposits on the California Channel Islands record a shift in the nature of sedimentation at the interglacial/glacial transition. Marine terraces that are ca. 80 ka are overlain by two eolianite units separated by paleosols. U-series ages of the terrace corals and carbonate rhizoliths indicate that eolian sedimentation occurred between ca. 80 and 49 ka, and again between ca. 27 and 14 ka. Eolian sands were apparently derived from carbonate-rich shelf sediments during glacially-lowered sea levels, because there are not sufficient beach sources for calcareous sediment at present. The times of eolian sedimentation agree well with times of glaciation predicted by the Milankovitch model of climatic change.

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INTRODUCTION

Accurate information about major climatic shifts of the past is critical to the prediction of future climates because these can be used to test alternative models of climate change. The last interglacial/glacial transition is an important interval in the late Quaternary history of North America. Less is known about the last interglacial/glacial transition than the last glacial/Holocene transition, because the former occurred at least 75 ka ago and perhaps as much as 115 ka. The record of the last interglacial/glacial transition is not as well preserved as the last glacial/Holocene transition, there are fewer dating methods available for developing an accurate chronology, and fewer investigators have studied this time interval.

There seems to be agreement among a great many investigators that the last interglacial is found in various geologic records, but there is considerable disagreement as to the chronology of this interval of time. In the midcontinent, the last interglacial is represented stratigraphically by the Sangamon soil (Follmer, 1978). Many investigators in the midcontinent have suggested that the period during which this well-developed soil formed was warmer than the present, but perhaps no longer than the Holocene, or about 10 ka (see review in Boardman [1985]). Fulton and Prest (1987) and St.-Onge (1987) suggest that the "Sangamonian Stage" in Canada lasted from 130 to 80 ka, and it is considered to be correlative with all of deep-sea oxygen-isotope stage 5. However, they regard the "Sangamon interglaciation" to be a time-transgressive term for geologic-climate or event units. Richmond and Fullerton (1986) also correlated the glacial record with the deep-sea oxygen-isotope record, but restricted the term "Sangamon" to substage 5e of the deep-sea record (ca. 132–122 ka) and referred to substages 5a–5d as the "Eowisconsin." In describing the Clear Lake, California, pollen record, Adam (1988) correlated a general period of high oak pollen with all of deep-sea isotope stage 5, but pointed out that the early part of this period, which he correlated with substage 5e, is the only time period that is truly "interglacial" in temperature terms. In their major study of the marine record from deep-sea cores, the CLIMAP Project Members (1984) restricted their definition of the "last interglacial" to substage 5e on the grounds that this was the last time that glacial ice volume was as low as or lower than present.

There are several major challenges to the concept that the last interglacial had a duration of only about 10 ka and is correlative only with isotope substage 5e, as suggested by the studies cited above. From the pedologic record, Ruhe (1974) and Ruhe and others (1974) proposed that the well-developed (compared to modern soils) Sangamon soils in the midwestern United States are the result of a warmer, wetter climate in that region during the period of soil formation. In contrast, Follmer (1982), Muhs (1983a), and Boardman (1985) argued that characteristics of the Sangamon soil could also be explained by a period of pedogenesis significantly in excess of 10 ka. Speleothem growth periods in Canadian mountains and Iowa, dated by U-series methods, suggest that the last interglacial period may have had a duration of 60 ka or more (Harmon and others, 1977; Lively, 1983). Oxygen-isotope data from an ice core taken from Devon Island off northern Canada also suggest a last interglacial much longer than 10 ka (Koerner and Fisher, 1985). Thus, two schools of thought have emerged from studies of proxy climate data: those paleoenvironmental records suggesting a "short" (ca 10 ka) last interglacial and those suggesting a "long" (ca 20–60 ka) last interglacial.

In this chapter I review age estimates and paleo-sea levels of the last interglacial derived from marine terraces found on the east and west coasts of North America and on islands close to North America such as Bermuda, the Bahamas, Barbados, and Haiti, particularly with regard to the record of the timing of the last interglacial. Terrace ages on the Huon peninsula of New Guinea are also examined, because this area has been studied carefully and has given one of the most detailed sea-level records of the late Quaternary. Finally, some preliminary results of studies of marine and eolian sediments on the Channel Islands of California and their implications for the record of the last interglacial/glacial transition are presented.

AGE ESTIMATES OF DEEP-SEA OXYGEN-ISOTOPE STAGE 5

A common ground in almost all of the studies cited above is correlation to the oxygen-isotope record in deep-sea cores, which is considered to be a much more complete record than those of glacial deposits, buried soils, pollen, speleothems, or ice. It follows, therefore, that one of the most important questions concerning the record of the last interglacial in deep-sea cores is the method(s) by which it has been dated. Most investigators agree that all or part of isotope stage 5 represents the last interglacial, so it is pertinent to examine the chronological data for this interval of time.

A compilation of data from some of the better-studied deep-sea cores reveals a range of age estimates for both the beginning and end of oxygen-isotope stage 5 (Table 1). The age estimates, derived by a variety of methods, are in general agreement, but there are significant differences when they are examined in detail. Both the ^{230}Th excess and Al accumulation methods used for core V28-238 indicate that stage 5 could have begun significantly earlier (by 10–15 ka) than the "orbital tuning" and ^{231}Pa/^{230}Th/sedimentation-rate age estimates suggest for core RC11-120. It is also possible to compare two different cores (V28-238 and CH72-02) dated by the same method (^{230}Th excess). These two cores differ in their starting dates for stage 5 by 20 ka and in their ending dates for stage 5 by almost 13 ka. Given all of the uncertainties in the dating methods that have been used on deep-sea cores, it is not possible, at the present time, to favor one chronology over another. Collectively, these data suggest that further refinements in independent dating of deep-sea cores is desirable.

Because of the problems in dating cores directly, a number of investigators have attempted to develop time scales for the deep-sea record by "tuning" the oxygen-isotope record to the

TABLE 1. ESTIMATES OF THE DEEP-SEA OXYGEN ISOTOPE STAGE 4/5 AND STAGE 5/6 BOUNDARIES

Core	4/5 Boundary (ka)	5/6 Boundary (ka)	Method(s) of Age Determination	Reference*
V28-238	77.9†	133.9†	Paleomagnetism/ sedimentation rate	1
V28-238	81.6	144.7	^{230}Th excess	2
V28-238	80.4	138	Aluminum accumulation	2
CH72-02	68.7	124	^{230}Th excess	3
RC11-120	73	127	^{231}Pa/^{230}Th/sedimentation rate	4
RC11-120	73.9	129.8	Orbital tuning	5
RC14-37	73.5	132	K/Ar/sedimentation rate	6

*1 = Shackleton and Opdyke (1973); 2 = Kominz and others (1979); 3 = Southon and others (1987); 4 = Hays and others (1976); 5 = Martinson and others (1987); 6 = Ninkovich and others (1978).
†Calculated by Muhs using core data from Shackleton and Opdyke (1973) and 730 ka as the time of the Brunhes-Matuyama boundary (Mankinen and Dalrymple, 1979).

time scale of changes in the distribution of solar radiation over the surface of the Earth (Morley and Hays, 1981; Johnson, 1982; Martinson and others, 1987). This approach assumes that the major factors driving climate changes are changes in the Earth's orbital parameters, as outlined by Milankovitch (1941). Unfortunately, this method is somewhat circular in its reasoning: it was the coincidence of independent U-series age estimates of deep-sea cores (using the ^{230}Th excess and ^{231}Pa excess methods) and uplifted coral reefs (using the ^{230}Th/^{234}U method) that was the original basis of support for the Milankovitch orbital-forcing theory (Broecker and others, 1968). This concern is magnified in light of recent climatic modeling evidence of an uncertain role for orbital factors in the growth and maintenance of continental ice sheets (Rind and others, 1989, this volume).

Until new methods for dating cores become available, some insight can be gained by examining the sea-level highstands recorded by marine terraces and dated by U-series analysis of fossil corals associated with these terraces. Marine terraces can yield data about both the timing of interglacial periods and the amount of global glacial ice, as recorded by the magnitude of sea-level rise.

MARINE TERRACES AS INDICATORS OF THE TIMING OF INTERGLACIAL SEA-LEVEL HIGHSTANDS

The importance of marine terraces as indicators for the timing of interglacial periods stems from the fact that heights of sea level have an inverse relation to amounts of global ice. Investigators generally agree that emergent marine terraces (whether coral reefs or cliff-platform types) represent highstands of sea level. On coasts undergoing tectonic uplift, terraces may also form during lowstands of sea level, but such terraces are not now exposed and many that formed were probably largely eroded away by higher stands of sea level. However, terraces that formed during highstands on tectonically active coasts can be uplifted sufficiently during the succeeding intervals of lowstands that they are elevated and hence preserved above sea level during subsequent highstands. For both coral-reef growth and cliff-platform cutting on tectonically emergent coasts, the optimum time of formation will be when the rate of sea-level rise matches the rate of uplift. This will occur during stage 1, 2, or 3 (as shown in Fig. 1), depending on whether the coast has a high, moderate, or low rate of uplift, respectively. On stable coasts, terrace formation will occur when sea level is not changing, shown as stage 4 in Figure 1. On both uplifting and stable coasts, terraces will be abandoned when sea level falls (stage 5 in Fig. 1). Thus, on coasts where tectonic uplift has been occurring throughout the Quaternary, a series of emergent terraces may be present which record the major highstands of sea level that correspond to times of minimal global ice, or interglacial and interstadial intervals (Fig. 2).

Under the assumption that, on uplifting coasts, marine terraces correspond to some part of a sea-level highstand, it follows that the intervals between times of successive terrace formation must represent lowstands that occur during buildup and maintenance of glacial ice. Thus, flights of marine terraces have a record of ice-volume fluctuations. The best method for determining the

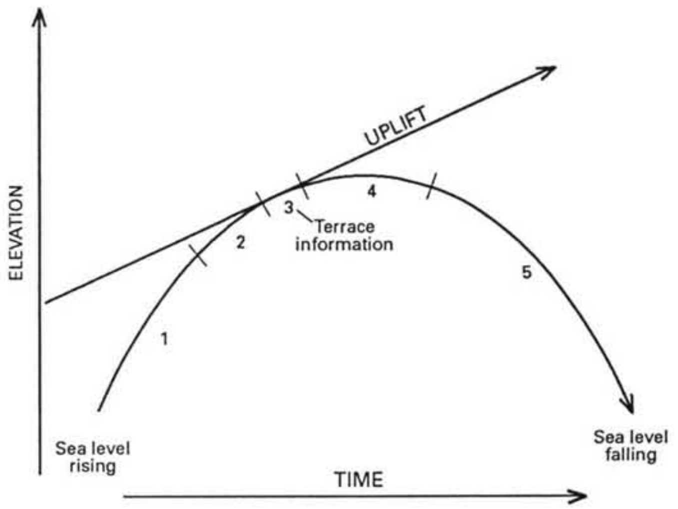

Figure 1. Model of marine terrace formation as a function of sea-level change and tectonic uplift. The diagonal line marked "uplift" represents a constant rate of uplift through time on a tectonically active coastline. The curved line represents the rise and fall of sea level during an interglacial sea-level highstand. The numbered segments of the sea-level curve represent different time periods during the history of sea-level rise and fall. On a tectonically active coast, terrace formation begins in stage 2 (rising sea) and is completed in stage 3, when the rate of uplift matches the rate of sea-level rise (shown here by the constant uplift rate line as a tangent to the sea-level curve in stage 3). The terrace is abandoned during stages 4 and 5. On a tectonically stable coast, terrace formation would take place during stage 4, when sea level is high, but not changing. Modified from Bradley and Griggs (1976).

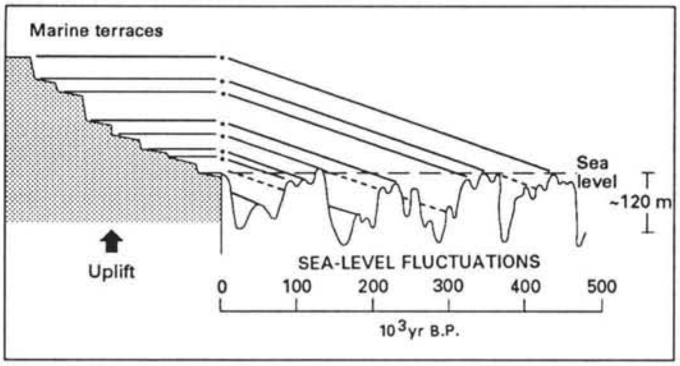

Figure 2. Model of successive marine-terrace formation on a tectonically active coast. Terraces form during highstands of sea level and are uplifted and preserved during succeeding lowstands of sea level. Slopes of diagonal lines represent uplift rates. Modified from Lajoie (1986).

time of terrace formation comes from uranium-series ($^{230}Th/^{234}U$, $^{234}U/^{238}U$, and $^{231}Pa/^{235}U$) dating of unaltered, aragonitic corals, either colonial or solitary forms.

URANIUM-SERIES AGES OF MARINE TERRACES

Marine terrace record

Detailed studies on the tectonically emergent islands of Barbados, New Guinea, and Haiti have identified three uplifted coral-reef terraces that record three highstands of sea level correlated, by many investigators, with some or all of deep-sea oxygen-isotope stage 5. The terraces are listed in order of ascending elevation and increasing age: on Barbados, the Worthing, Ventnor, and Rendezvous Hill terraces (Bender and others, 1979; also called Barbados I, II, and III, respectively, by Mesolella and others, 1969); on New Guinea, reef complexes V, VI, and VII (Chappell, 1974; Bloom and others, 1974); and on Haiti, the Mole, Saint, and Nicolas terraces (Dodge and others, 1983). These terraces have been dated by U-series methods as ca. 80 ka, 105 ka, and 125 ka, and have been correlated with, from youngest to oldest, isotope substages 5a, 5c, and 5e, respectively, by Shackleton and Opdyke (1973), Shackleton and Matthews (1977), Fairbanks and Matthews (1978), Bender and others (1979), Aharon (1983), Dodge and others (1983), Matthews (1985), and Chappell and Shackleton (1986). Even with the uncertainties in the age assignments for isotope stage 5 (Table 1). these are probably reasonable correlations. On the tectonically active west coast of North America, two and sometimes three terraces have been correlated with the Barbados, New Guinea, and Haiti terraces described above by Ku and Kern (1974), Kern (1977), Kennedy and others (1982), Muhs and Szabo (1982), Muhs and others (1987b, 1988, 1990), and Rockwell and others (1989). Terraces on the west coast of North America are not emergent coral reefs, but rather are uplifted shore platforms that are overlain by marine sands and gravels. These marine-terrace deposits frequently contain solitary corals or colonial hydrocorals that allow U-series dating.

On coastlines thought to be tectonically stable (i.e., not undergoing significant uplift or subsidence), such as the Bahamas, Bermuda, the Yucatan peninsula, and the east coast of the United States, there is a far less-clear picture of the sea-level record. Marine deposits on these coasts generally do not form geomorphically distinct terraces as they do on uplifting coasts, and there is considerably more controversy concerning their interpretation. However, a number of these emergent marine deposits contain corals, and U-series ages have been reported by Broecker and Thurber (1965), Osmond and others (1965), Neumann and Moore (1975), Szabo and others (1978), Harmon and others (1983), Szabo (1985), and Szabo and Halley (1988).

Uranium-series systematics

The reliability of a U-series age of a fossil coral is contingent on that organism having incorporated a small amount of U (but no Th) from sea water during growth, and having maintained a closed system with respect to U and its daughter products after death. The most rigorous test for closed-system conditions is agreement between $^{230}Th/^{234}U$ and $^{231}Pa/^{235}U$ ages, but few laboratories analyze samples for ^{231}Pa abundances. However, a $^{230}Th/^{234}U$ age can be interpreted to be reliable if the fossil coral passes the following tests: (1) the sample is 95%–100% aragonite; (2) the $^{230}Th/^{232}Th$ value is high (>20), indicating no incorporation of Th-bearing minerals; (3) the U concentration is 2–3 ppm (typical for colonial corals; solitary corals appear to be higher), indicating no post-mortem gains or losses of U; and (4) the $^{234}U/^{238}U$ value is consistent with the $^{230}Th/^{234}U$ age, assuming that the initial $^{234}U/^{238}U$ value in sea water at the time of coral growth was 1.14–1.15, and that this ratio has decreased with ^{234}U decay toward an equilibrium value of 1.00. Measurements of modern sea water have given $^{234}U/^{238}U$ values of 1.14–1.15 (Ku and others, 1977; Chen and others, 1986).

In Table 2, I have compiled all published U-series ages of corals of which I am aware from the study areas described above that are from marine deposits or emergent reefs that have either been correlated by various investigators with oxygen-isotope stage 5 or have ages that are in the estimated age ranges for isotope stage 5 given in Table 1. In making this compilation I have included only those samples that meet the closed-system criteria described above. Because of methodological differences between laboratories (Harmon and others, 1979; Ivanovich and others, 1984), I have recalculated all of the ages for the samples in Table 2 using half-lives of 75,200 yr and 244,000 yr for ^{230}Th and ^{234}U, respectively, as well as giving the original ages reported by the authors. All uncertainties given in Table 2 and in the text are 1 sigma.

Uranium-series ages of terrace corals: A general picture

When all closed-system U-series ages of terrace corals that have been correlated with isotope stage 5 are plotted as a frequency histogram, there is evidence of two distinct highstands of sea level ca. 125–120 ka and 85–80 ka and a less-distinct highstand ca. 105 ka (Fig. 3). The three age modes agree well with the peaks of summer insolation at high latitudes as calculated by Berger (1978) and therefore it can be said that in general the sea-level record supports the Milankovitch orbital-forcing theory of climatic change (Fig. 3). However, when this plot is examined in some detail, there are a number of problematic ages. There are several coral ages between 120 and 105 ka, although most of these permit agreement with the 125–120 ka or 105 ka ages when their analytical errors are considered. There are also several ages in the range 145–135 ka that are analytically distinguishable from the ca. 125–120 ka ages. Some workers have argued that there may be two distinct highstands of sea level in the time interval from 120 to 140 ka (Chappell and Veeh, 1978; Moore, 1982). Only a single ice-volume minimum in this time period is recorded in deep-sea cores, but it is interesting that ages of ca. 145–135 ka for terrace corals are in agreement with some of the age estimates for the start of substage 5e based on direct analyses of the cores (Table 1). Winograd and others (1988) cited the ca. 140 ka ages as supportive evidence for a start of substage 5e that is considerably earlier than that used by most investigators. The 145–135 ka ages for corals and for the start of substage 5e in cores present a challenge to the Milankovitch orbital-forcing theory, because insolation during summer at high latitudes was relatively low around this time (Fig. 3). Further discussion of this problem can be found in Chappell and Veeh (1978), Moore (1982), Kaufman (1986), and Pillans (1987).

In addition to uncertainties about when isotope stage 5 began, there are problems in interpreting U-series data from terraces to determine when isotope stage 5 ended. As discussed earlier, some workers regard the whole of deep-sea stage 5 as the last interglacial period; others would argue that only substage 5e can be considered "last interglacial." If the latter interpretation is used, then the data plotted in Figure 3 imply that the last interglacial ended ca. 120–115 ka. If the former interpretation is used, then the close of the last interglacial period, on the basis of what is portrayed in Figure 3, is much harder to define. There are a number of terrace corals in Table 2 and Figure 3 with ages ranging from 75 to 60 ka from deposits that have been correlated with isotope substage 5a. As with the 145–135 ka corals, these ages represent a challenge to the Milankovitch theory, because summer insolation at high latitudes was quite low at this time. Thus, there are a number of problems and inconsistencies in the data given in Table 2 and plotted in Figure 3, even though all these samples passed the geochemical and isotopic screening described above. It is useful to discuss these problems on a study area by study area basis.

Barbados

The coral terraces on Barbados have probably received more attention for U-series dating than any other terrace sequence in the world, and as a result, the data tabulated in Table 2 allow for detailed comparisons. Most of the ages for the Worthing terrace cluster around 80 ka. This age was used by Broecker and others (1968) and Mesolella and others (1969) to provide support for the orbital-forcing theory of climatic change, because this postdates a time (ca. 85 ka) of high summer insolation at high latitudes (Fig. 3). However, two recent suites of analyses of corals from this terrace disagree with the ca. 80 ka age estimates. Radtke and others (1988) analyzed three corals from the Worthing terrace and obtained ages of 66 ± 2 ka, 70 ± 2 ka, and 72 ± 2 ka (Table 2). These dates are problematic because they indicate a time of formation of the Worthing terrace when insolation was very low at high latitudes (Fig. 3). A second problematic set of ages was reported by Edwards and others (1987). These workers obtained mass-spectrometric ages of 87 ka (with an uncertainty of only 200–300 yr) for the Worthing terrace, which would indicate that sea level was high at a time just *before* the high-latitude insolation maximum (Fig. 3). This would require that melting of most glacial ice occurred well before summer radiation was at a maximum at high latitudes. It is perhaps more reasonable to think that sea level would be at a maximum sometime after the solar-radiation maximum.

A similar problem can be identified with the mass spectrometric ages of 111–112 ka for the Ventnor terrace reported by Edwards and others (1987). These ages precede, by a significant amount, an insolation maximum that occurred at about 103 ka. Other workers who have analyzed corals from this terrace have obtained ages that average about 105 ka, but which range from 100 to 111 ka, with an uncertainty of 2–6 ka (Table 2). Edwards and others (1987) concluded that factors other than orbital forcing could have caused a high sea level at 111–112 ka, but the other ages of the Ventnor terrace would still permit an orbitally-induced sea-level highstand.

There are also problems associated with the highstand of about 120–129 ka recorded by the Rendezvous Hill terrace on Barbados. Recent U-series ages of coral from this terrace reported by Ku and others (1990) are mostly 120–125 ka, in agreement with previous analyses, but three of their samples gave ages of 99 ± 3 ka, 103 ± 2 ka, and 106 ± 2 ka. These ages are in the general range of ages for the younger Ventnor terrace (Table 2). Whereas it is possible for older terrace corals to be reworked onto a younger terrace surface, it is difficult to envision how younger terrace corals could be reworked onto an older, topographically higher terrace. In addition, these corals do not show any evidence of significant recrystallization, secondary U additions, or other open-system effects. Therefore, it is difficult to explain the younger-than-expected ages in these samples. All of the age discrepancies discussed for the Worthing, Ventnor, and Rendezvous Hill terrace corals could be the result of subtle diagenetic processes, as suggested by Ku and others (1990), but there are at present no clues as to what these processes might be.

A final problem with the terrace sequence on Barbados concerns the newly-named Maxwell terrace reported by Ku and others (1990). This terrace actually appeared on the reef trend map of Mesolella and others (1969, Fig. 6), but was not mapped as a separate reef crest by Bender and others (1979, Fig. 2). Ku

TABLE 2. ISOTOPIC ACTIVITY RATIOS AND URANIUM-SERIES AGES OF CORALS FROM MARINE DEPOSITS THAT ARE THOUGHT TO BE CORRELATE WITH ISOTOPE STAGE 5

Sample Number	$^{234}U/^{238}U$ Activity Ratios	$^{230}Th/^{234}U$ Activity Ratios	Age* (ka)	Age† (ka)	Ref.§
NORTH CAROLINA AND VIRGINIA (NORFOLK AND KEMPSVILLE FORMATIONS)					
C-2	1.10 ± 0.02	0.50 ± 0.02	74 ± 4	74 ± 4	1
C-18	1.11 ± 0.02	0.52 ± 0.02	79 ± 5	78 ± 5	1
C-36A	1.10 ± 0.02	0.48 ± 0.02	69 ± 4	70 ± 4	1
C-36B	1.09 ± 0.02	0.46 ± 0.02	67 ± 4	66 ± 4	1
C-15	1.10 ± 0.02	0.49 ± 0.02	72 ± 4	73 ± 4	1
C-34	1.12 ± 0.02	0.49 ± 0.02	72 ± 4	72 ± 4	1
SOUTH CAROLINA ("LATE WANDO" FORMATION)					
	1.09 ± 0.02	0.56 ± 0.01	87 ± 4	88 ± 3	1
FLORIDA (KEY LARGO LIMESTONE)					
KL1	1.13 ± 0.01	0.75 ± 0.02	145 ± 14	144 ± 8	2
765A	1.12 ± 0.01	0.61 ± 0.03	95 ± 9	100 ± 8	3
801D	1.09 ± 0.02	0.71 ± 0.04	130 ± 20	131 ± 15	3
801F	1.09 ± 0.01	0.71 ± 0.03	130 ± 15	131 ± 11	3
801C	1.10 ± 0.02	0.75 ± 0.03	140 ± 15	145 ± 13	4
KL-5	1.14 ± 0.03	0.69 ± 0.04	125 ± 15	122 ± 14	4
KL-8	1.09 ± 0.02	0.68 ± 0.03	125 ± 11	121 ± 10	4
KL-10	1.08 ± 0.03	0.70 ± 0.05	133 ± 19	128 ± 18	4
KL-10	1.15 ± 0.08	0.74 ± 0.05	145 ± 14	139 ± 21	4
BERMUDA (SOUTHAMPTON FORMATION)					
790515-6	1.08 ± 0.02	0.54 ± 0.02	85 ± 6	83 ± 5	5
PG-218	1.08 ± 0.01	0.53 ± 0.02	83 ± 5	81 ± 5	5
BERMUDA (SPENCER'S POINT FORMATION)					
75012	1.11 ± 0.02	0.60 ± 0.03	97 ± 6	97 ± 8	5
780524-14	1.11 ± 0.03	0.62 ± 0.02	106 ± 6	103 ± 5	5
780522-5	1.05 ± 0.01	0.63 ± 0.02	108 ± 6	107 ± 6	5
BERMUDA (DEVONSHIRE FORMATION)					
75014	1.10 ± 0.02	0.67 ± 0.02	118 ± 6	117 ± 7	5
PG-321	1.06 ± 0.02	0.67 ± 0.04	118 ± 11	118 ± 14	5
PG-4	1.09 ± 0.01	0.67 ± 0.02	118 ± 6	118 ± 6	5
780522-6	1.10 ± 0.02	0.68 ± 0.02	121 ± 6	120 ± 7	5
75013	1.11 ± 0.02	0.69 ± 0.02	124 ± 6	123 ± 7	5
016	1.10 ± 0.03	0.70 ± 0.03	124 ± 8	127 ± 11	5
75020	1.09 ± 0.02	0.69 ± 0.02	124 ± 6	124 ± 7	5
PG-324	1.11 ± 0.03	0.69 ± 0.03	124 ± 8	123 ± 11	5
PG-322	1.10 ± 0.02	0.69 ± 0.04	124 ± 12	124 ± 14	5
75015	1.09 ± 0.01	0.68 ± 0.01	125 ± 6	121 ± 3	5
790513-2	1.06 ± 0.02	0.70 ± 0.03	127 ± 9	128 ± 11	5
PG-304	1.07 ± 0.02	0.70 ± 0.03	127 ± 9	128 ± 11	5
780521	1.11 ± 0.01	0.70 ± 0.02	127 ± 6	127 ± 7	5
PG-220	1.07 ± 0.01	0.71 ± 0.02	131 ± 7	131 ± 8	5
PG-210	1.12 ± 0.02	0.72 ± 0.02	134 ± 8	133 ± 7	5
NORTHERN BAHAMAS					
717F	1.14 ± 0.02	0.56 ± 0.04	80 ± 3	87 ± 10	3
20	1.08 ± 0.02	0.68 ± 0.02	120 ± 6	121 ± 7	6
21	1.10 ± 0.02	0.70 ± 0.02	128 ± 6	126 ± 7	6
5-7	1.13 ± 0.03	0.67 ± 0.02	125 ± 6	117 ± 6	6
5-8	1.13 ± 0.03	0.67 ± 0.02	125 ± 6	117 ± 6	6
6-3	1.13 ± 0.03	0.71 ± 0.03	140 ± 9	129 ± 11	6
36-C	1.11 ± 0.03	0.74 ± 0.03	146 ± 9	141 ± 12	6
43-1	1.09 ± 0.03	0.68 ± 0.02	121 ± 6	121 ± 7	6
43-3	1.06 ± 0.03	0.70 ± 0.02	121 ± 6	128 ± 7	6
46-2B	1.13 ± 0.03	0.69 ± 0.02	132 ± 7	123 ± 6	6
47-D	1.07 ± 0.03	0.66 ± 0.02	112 ± 6	115 ± 6	6
SAN SALVADOR, BAHAMAS					
GB-3	1.10 ± 0.02	0.69 ± 0.02	123 ± 9	124 ± 7	7
HAITI (MOLE TERRACE)					
E12	1.10 ± 0.02	0.53 ± 0.01	80 ± 3	81 ± 3	8
02	1.12 ± 0.02	0.54 ± 0.01	82 ± 2	82 ± 3	8
K1	1.10 ± 0.02	0.51 ± 0.01	77 ± 3	76 ± 2	8
P3	1.10 ± 0.02	0.54 ± 0.01	83 ± 3	83 ± 2	8
G1	1.09 ± 0.02	0.70 ± 0.02	128 ± 6	127 ± 7	8
HAITI (SAINT TERRACE)					
Q5	1.10 ± 0.01	0.64 ± 0.02	108 ± 5	108 ± 6	8
HAITI (NICOLAS TERRACE)					
B2	1.10 ± 0.02	0.71 ± 0.02	130 ± 6	130 ± 8	8
B11	1.12 ± 0.02	0.71 ± 0.01	132 ± 5	130 ± 4	8
C1 (α-spec)	1.12 ± 0.02	0.71 ± 0.02	130 ± 6	130 ± 7	8
C4 (α-spec)	1.12 ± 0.02	0.70 ± 0.02	126 ± 6	126 ± 7	8
C1 (mass-spec)	1.107± 0.002	0.684 ± 0.002	121.9 ± 0.5	121.5 ±0.7	9
C4#1 (mass-spec)	1.109± 0.004	0.690 ± 0.002	123.6 ± 0.8	123.4 ±0.7	9
BARBADOS (WORTHING TERRACE)					
AGA-1	1.11 ± 0.02	0.52 ± 0.02	79 ± 4	78 ± 5	10
OC-26	1.13 ± 0.01	0.53 ± 0.01	82 ± 2	80 ± 2	10
AEH-1	1.12 ± 0.01	0.53 ± 0.02	82 ± 4	80 ± 5	10
FS-3	1.13 ± 0.02	0.54 ± 0.02	84 ± 4	83 ± 5	10
1152C	1.11 ± 0.01	0.52 ± 0.01	79 ± 2	78 ± 3	11
B-4	1.11 ± 0.01	0.46 ± 0.01	67 ± 2	66 ± 2	12
B-5	1.13 ± 0.02	0.48 ± 0.01	70 ± 3	70 ± 2	12
B-7	1.12 ± 0.02	0.49 ± 0.03	73 ± 6	72 ± 6	12
OC-51A	1.126± 0.003	0.560 ± 0.001	87.5 ± 0.3	87.3 ±0.3	13
OC-51B	1.126± 0.003	0.562 ± 0.001	87.9 ± 0.4	87.8 ±0.2	13
FS-50A	1.127± 0.008	0.519± 0.006	78 ± 1	78 ± 1	14
FS-51	1.134± 0.008	0.508± 0.006	76 ± 1	76 ± 2	14
OC-50	1.126± 0.007	0.551± 0.006	85 ± 1	85 ± 2	14
OC-51	1.131± 0.007	0.542± 0.006	83 ± 1	83 ± 1	14
OC-53	1.13 ± 0.01	0.543± 0.007	83 ± 2	83 ± 2	14
BARBADOS (VENTNOR TERRACE)					
1144C	1.13 ± 0.01	0.63 ± 0.01	105 ± 3	105 ± 3	11
FT-1	1.11 ± 0.02	0.62 ± 0.02	104 ± 4	103 ± 6	10
FT-1	1.12 ± 0.02	0.61 ± 0.02	100 ± 4	100 ± 5	10
FT-1	1.11 ± 0.02	0.61 ± 0.02	100 ± 4	100 ± 5	10
AFZ-1	1.10 ± 0.01	0.62 ± 0.02	104 ± 4	103 ± 5	10
AFZ-1	1.11 ± 0.01	0.62 ± 0.02	104 ± 4	103 ± 6	10
AFK-1	1.10 ± 0.01	0.62 ± 0.02	104 ± 6	103 ± 5	10
AEG-2	1.10 ± 0.02	0.65 ± 0.02	111 ± 6	111 ± 6	10
X-5	1.13 ± 0.02	0.65 ± 0.02	111 ± 6	111 ± 6	10
B59a	1.13 ± 0.02	0.63 ± 0.01	100+6/-3	105 ± 3	12
FT-50A	1.126± 0.003	0.653 ± 0.001	112.0 ± 0.5	111.6 ±0.3	13
FT-50B	1.127± 0.002	0.653 ± 0.002	111.8 ± 0.6	111.6 ±0.6	13
FT-50B	1.124± 0.003	0.654 ± 0.001	112.3 ± 0.5	112.0 ±0.3	13
ANM-21	1.116± 0.007	0.620± 0.007	102 ± 2	102 ± 2	14

TABLE 2. ISOTOPIC ACTIVITY RATIOS AND URANIUM-SERIES AGES OF CORALS FROM MARINE DEPOSITS THAT ARE THOUGHT TO BE CORRELATE WITH ISOTOPE STAGE 5 (continued)

Sample Number	$^{234}U/^{238}U$ Activity Ratios	$^{230}Th/^{234}U$ Activity Ratios	Age* (ka)	Age† (ka)	Ref.§
BARBADOS (VENTNOR TERRACE) continued					
AM-22	1.10 ± 0.01	0.615± 0.008	102 ± 2	102 ± 3	14
FT-51(1)	1.117± 0.006	0.65 ± 0.01	110 ± 2	111 ± 3	14
BAB-10	1.125± 0.008	0.623± 0.007	103 ± 2	103 ± 2	14
BARBADOS (MAXWELL TERRACE)					
AEJ-5	1.12 ± 0.01	0.69 ± 0.02	124 ± 6	123 ± 7	10
AEJ-21	1.105± 0.007	0.702± 0.007	128 ± 2	128 ± 3	14
AEJ-22(2)	1.12 ± 0.01	0.683± 0.008	121 ± 3	121 ± 3	14
AGP-10	1.111± 0.007	0.661± 0.007	114 ± 2	114 ± 2	14
AGP-11	1.099± 0.008	0.662± 0.008	118 ± 2	118 ± 2	14
AGP-12(1)	1.10 ± 0.01	0.662± 0.008	115 ± 2	115 ± 3	14
BARBADOS (RENDEZVOUS HILL TERRACE)					
1152E	1.10 ± 0.01	0.69 ± 0.01	125 ± 5	125 ± 4	11
AFS-1	1.12 ± 0.03	0.69 ± 0.02	124 ± 6	123 ± 7	10
AFM-T-2	1.11 ± 0.01	0.70 ± 0.02	127 ± 6	127 ± 7	10
AFM-B-1	1.09 ± 0.02	0.70 ± 0.02	127 ± 6	127 ± 7	10
ADR-1	1.11 ± 0.02	0.70 ± 0.02	127 ± 6	127 ± 7	10
B66	1.13 ± 0.02	0.70 ± 0.02	128 ± 7	125 ± 7	12
AFM-20	1.111± 0.002	0.707 ± 0.002	129.2 ± 0.7	129.0 ±0.7	13
R-52	1.115± 0.003	0.704 ± 0.002	128.1 ± 0.8	127.9 ±0.7	13
AFS-10	1.110± 0.002	0.696 ± 0.001	125.7 ± 0.6	125.4 ±0.4	13
AFS-11	1.114± 0.003	0.687 ± 0.001	122.6 ± 0.7	122.3 ±0.7	13
AFS-12A	1.109± 0.002	0.685 ± 0.001	122.1 ± 0.5	121.8 ±0.3	13
AFS-12B	1.111± 0.003	0.687 ± 0.001	122.7 ± 0.6	122.4 ±0.3	13
AFS-12C	1.109± 0.003	0.692 ± 0.001	124.5 ± 0.6	124.1 ±0.4	14
AFM-20A(1)	1.099± 0.009	0.673± 0.009	118 ± 3	118 ± 3	14
AFM-20A(2)	1.114± 0.009	0.680± 0.008	120 ± 3	120 ± 3	14
AFM-22A(1)	1.103± 0.008	0.619± 0.007	102 ± 2	102 ± 2	14
AFM-22A(2)	1.111± 0.009	0.608± 0.008	99 ± 2	99 ± 3	14
AFM-23(1)	1.116± 0.009	0.659± 0.008	114 ± 2	114 ± 3	14
AFM-23(2)	1.11 ± 0.01	0.631± 0.008	106 ± 2	106 ± 2	14
R-51(2)	1.111± 0.008	0.688± 0.007	123 ± 2	123 ± 3	14
AFS-10	1.097± 0.009	0.685± 0.008	122 ± 3	122 ± 3	14
AFS-11	1.096± 0.009	0.690± 0.008	124 ± 3	124 ± 3	14
AFS-12	1.12 ± 0.01	0.682± 0.009	121 ± 3	120 ± 3	14
AFM3#1	1.108± 0.004	0.694 ± 0.002	125.1 ± 0.8	124.8 ±0.7	9
YUCATAN PENINSULA					
355-11	1.10 ± 0.01	0.68 ± 0.02	121 ± 6	120 ± 7	15
355-12	1.10 ± 0.01	0.69 ± 0.02	123 ± 6	124 ± 7	15
355-14	1.10 ± 0.01	0.68 ± 0.02	120 ± 6	120 ± 7	15
293-9	1.11 ± 0.01	0.69 ± 0.02	120 ± 6	123 ± 7	15
C-22	1.10 ± 0.01	0.68 ± 0.02	121 ± 6	120 ± 7	15
OREGON (WHISKY RUN TERRACE)					
M2798	1.10 ± 0.01	0.54 ± 0.02	83 ± 5	83 ± 5	16
CALIFORNIA (POINT ARENA)					
M7824	1.12 ± 0.02	0.51 ± 0.02	76 ± 4	76 ± 4	16
CALIFORNIA (POINT LOMA, NESTOR TERRACE)					
2577	1.11 ± 0.01	0.64 ± 0.02	109 ± 6	108 ± 6	17
2577	1.12 ± 0.02	0.71 ± 0.02	131 ± 8	130 ± 7	17
2577	1.12 ± 0.01	0.69 ± 0.02	124 ± 7	123 ± 7	17
BAJA CALIFORNIA (PUNTA BANDA, LIGHTHOUSE TERRACE)					
PB-8	1.11 ± 0.01	0.52 ± 0.01	78 ± 3	78 ± 3	18
BAJA CALIFORNIA (PUNTA BANDA, SEA CAVE TERRACE)					
PB-3	1.11 ± 0.01	0.68 ± 0.01	120 ± 3	120 ± 3	18
PB-10	1.10 ± 0.01	0.69 ± 0.01	124 ± 4	124 ± 4	18
BAJA CALIFORNIA (MAGDALENA TERRACE)					
AO-3	1.08 ± 0.02	0.66 ± 0.02	116 ± 8	115 ± 6	19
AO-6	1.14 ± 0.03	0.66 ± 0.04	118 ± 12	113 ± 13	19
BAJA CALIFORNIA (MULEGE TERRACE)					
3258	1.10 ± 0.03	0.75 ± 0.02	144 ± 7	145 ± 8	20
3261	1.11 ± 0.03	0.69 ± 0.02	124 ± 5	123 ± 7	20
BAJA CALIFORNIA (LA PAZ AREA)					
C-A	1.12 ± 0.02	0.75 ± 0.02	146 ± 9	144 ± 7	21
C-B	1.12 ± 0.02	0.72 ± 0.02	135 ± 6	133 ± 7	21
NEW GUINEA (REEF COMPLEX V)					
8	1.13 ± 0.02	0.43 ± 0.02	61 ± 4	60 ± 4	22
12	1.10 ± 0.02	0.54 ± 0.02	84 ± 4	83 ± 5	22
38	1.11 ± 0.02	0.55 ± 0.02	86 ± 4	85 ± 5	22
NEW GUINEA (REEF COMPLEX VI)					
14b	1.12 ± 0.02	0.63 ± 0.03	107 ± 9	105 ± 9	22
20	1.09 ± 0.02	0.63 ± 0.02	107 ± 6	106 ± 6	22
NEW GUINEA (REEF COMPLEX VIIB)					
NG618	1.11 ± 0.01	0.67 ± 0.02	116 ± 7	117 ± 6	23
NG618	1.11 ± 0.01	0.68 ± 0.02	119 ± 7	120 ± 7	23
NEW GUINEA (REEF COMPLEX VIIA)					
NG616	1.10 ± 0.01	0.74 ± 0.03	140 ± 10	141 ± 12	23
NG616	1.12 ± 0.02	0.72 ± 0.03	133 ± 10	133 ± 11	23
15	1.08 ± 0.02	0.74 ± 0.02	142 ± 8	142 ± 8	22

*Age reported by authors in reference cited.
†Age calculated by Muhs using half-lives of ^{230}Th and ^{234}U of 75,200 and 244,000 years, respectively.
§1 = Szabo (1985); 2 = Szabo and Halley (1988); 3 = Broecker and Thurber (1965); 4 = Osmond and others (1965); 5 = Harmon and others (1983); 6 = Neumann and Moore (1975); 7 = Szabo and others (1988); 8 = Dodge and others (1983); 9 = Bard and others (1990); 10 = Mesolella and others (1969); 11 = Ku (1968); 12 = Radtke and others (1988); 13 = Edwards and others (1987); 14 = Ku and others (1990); 15 = Szabo and others (1978); 16 = Muhs and others (1990); 17 = Ku and Kern (1974); 18 = Rockwell and others (1989); 19 = Omura and others (1979); 20 = Ashby and others (1987); 21 = Szabo and others (1990); 22 = Bloom and others (1974); 23 = Veeh and Chappell (1970).

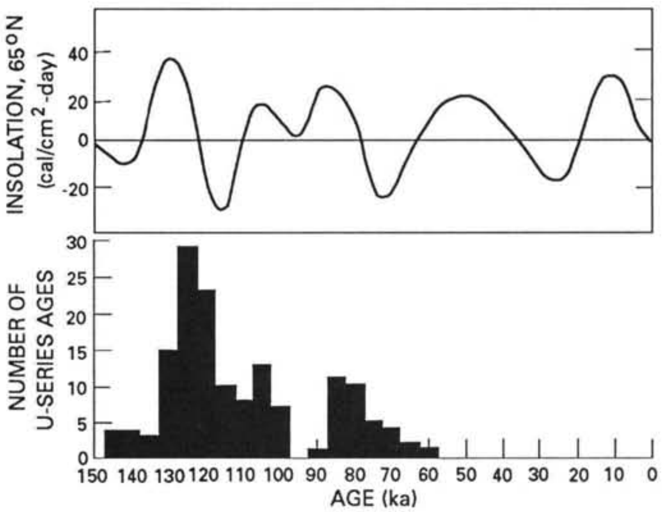

Figure 3. Histogram of U-series ages of corals from terraces that have been correlated with deep-sea oxygen-isotope stage 5. Ages plotted are those calculated by Muhs from data in Table 2, rounded off to the closest 5,000 years. Also shown for comparison is average insolation received at the top of the atmosphere at lat 65°N for the summer half year, expressed as deviations from the present (A.D. 1950) value. Insolation data are from Berger (1978).

and others (1990) tested the hypothesis that these two terraces could represent the "dual peak" proposed by Chappell and Veeh (1978), Aharon and others (1980), and Moore (1982) and alluded to earlier in this chapter. They confirmed the earlier results of Mesolella and others (1969) that there is no significant age difference between these two terraces and thus they do not support the hypothesis of bipartite sea levels in the time period 120–140 ka. However, what is not explained by these data is why there are *two* geomorphically distinct terraces, both with *Acropora palmata* reef-crest corals, that are the same age.

East Coast of the United States

At localities in Virginia and North Carolina, several corals from the Norfolk and Kempsville formations (or their equivalents), meeting all of the closed-system requirements, give ages ranging from 66 to 78 ka and averaging about 72 ka (Szabo, 1985; Table 2). On the basis of their fossil assemblages (equivalent to modern ones), and their estimated paleo-sea level (+4 to +10 m), Mixon and others (1982) thought that these deposits should be about 125 ka, and speculated that the corals giving ages of 66–78 ka had undergone postdepositional loss of ^{230}Th. Given the relative immobility of Th in most near-surface environments, ^{230}Th loss does not seem very likely, and Szabo (1985) later showed that the ^{230}Th/^{234}U ages were reliable, on the basis of concordant ^{231}Pa/^{235}U ages. As for the ca. 70 ka corals from the Worthing terrace on Barbados, these ages suggest a very high sea level at a time when summer insolation at high latitudes was quite low (Fig. 3), and thus they are in conflict with the orbital-forcing theory.

Localities with 145–135 ka terrace corals

Certain terrace corals from New Guinea, Baja California, Florida, and the Bahamas have given ages in the range of 145–135 ka (Table 2). As discussed earlier, these ages are problematic in that they are in a time period of relatively low summer insolation at high latitudes (Fig. 3), yet there is no apparent evidence for open-system conditions. In Baja California, Florida, and the Bahamas, at least some of these corals are apparently found in the same deposits that yielded corals with 120–125 ka ages (Table 2). These data suggest that subtle diagenetic processes, for which we have little or no geochemical or isotopic evidence, may be responsible for bringing about open-system conditions. Geomorphic, stratigraphic, U-series, and stable-isotope data from the rapidly uplifting Huon Peninsula of New Guinea, however, indicate that reef complex VIIa is distinct from reef complex VIIb; the former has been dated at 142–133 ka and the latter has been dated as 120–117 ka (Veeh and Chappell, 1970; Chappell, 1974; Bloom and others, 1974; Aharon, 1983). It is possible that on tectonically stable coasts or coasts with very low uplift rates such as Baja California Sur, Florida, and the Bahamas), one would expect a mixture of corals of two different ages if there were two highstands of sea level of about the same magnitude, not greatly separated in time. It is possible that future high-precision mass-spectrometric U-series analyses of terrace corals will shed further light on this issue.

MAGNITUDE OF SEA-LEVEL RISE DURING ISOTOPE STAGE 5

Sea level at ca. 125 ka

In tectonically stable areas distant from plate boundaries, U-series dating of corals indicates that sea level ca. 125 ka was about 5–6 m higher than present. Patch reefs and marine calcarenites on Bermuda dated ca. 125 ka indicate a former sea level 5 m higher than present (Harmon and others, 1983). In the northern Bahamas, coral-bearing marine conglomerates and in situ reefs that have been dated ca. 125 ka occur at 0.2 to 2.3 m above current sea level; bioerosional notches that are thought to have formed during the same high sea-level stand occur at 5 to 6 m above sea level (Neumann and Moore, 1975). Elsewhere in the Bahamas, ooids in beach deposits dated ca. 125 ka indicate a possible paleo-sea level of up to 10 m higher than present sea level (Garrett and Gould, 1984; Muhs and others, 1987a). In Florida, the Key Largo limestone has a maximum elevation of 6 m above sea level on Windley Key (Szabo and Halley, 1988), and ages of ca. 125 and 145 ka have been reported for corals from this locality (Broecker and Thurber, 1965; Osmond and others, 1965; Szabo and Halley, 1988; Table 2). Corals dated ca. 125 ka on the Yucatan Peninsula are associated with beach deposits that indicate a paleo-sea level of 3 to 6 m higher than present (Szabo and others, 1978). Thus, there is convincing evidence that sea level ca. 125 ka stood at 6 ± 4 m above the present, and there is general agreement among most investigators that this marks a time of significantly lower global ice volume compared to the present.

Sea levels at ca. 105 and 80 ka

Whether the relatively high sea-level stands ca. 80 and 105 ka can be regarded as "interglacial" depends in part on how high the sea stood at these times, which in turn has an inverse relation to the volume of global ice. Studies of terraces on the tectonically emergent islands of Barbados, New Guinea, and Haiti all suggest that sea levels were significantly below the present at these times, based on elevations of the ca. 80, 105, and 125 ka terraces, an assumption of a +6 m highstand of sea level ca. 125 ka, and an assumption of a constant rate of uplift in the past 125 ka. The rate of uplift is calculated from the present elevation of the 125 ka terrace (–6 m) divided by its age. Using the ages and present elevations of the 80 and 105 ka terraces, it is then possible to calculate the paleo-sea levels at the time of formation of these terraces. Using the latest measurements of the elevations of reefs reported by Chappell and Shackleton (1986) for New Guinea, Bender and others (1979) for Barbados, and Dodge and others (1983) for Haiti, sea level is estimated to have been at –16 to –18 m at 105 ka and –15 to –16 m at 80 ka, relative to present. By the use of an alternative method, Bloom and Yonekura (1985) recalculated paleo-sea levels on New Guinea with the surprising conclusion that sea level at 80 ka is estimated to have been ~7 m lower relative to present and sea level at 105 ka is estimated to have been about the same as the present (paleo-sea levels estimated for terraces younger than ca. 80 ka gave about the same results by both methods). Unfortunately, there are too few elevation data for terraces on Barbados and Haiti to make comparative paleo-sea-level estimates using the Bloom and Yonekura (1985) method. It is not known why the two methods should yield such significantly different results, but it is clear that the elevations of the 80 and 105 ka sea-level highstands are not yet firm.

Data from the tectonically stable coasts of Bermuda, the Bahamas, and the eastern United States and from the tectonically uplifting coast of western North America, indicate sea levels near to, or even higher than, present at 80 and 105 ka. On Bermuda, there are several patchy, coral-bearing marine conglomerates that Harmon and others (1983) have dated ca. 80 and 105 ka. These investigators interpreted these marine conglomerates as storm deposits and said that currently submerged speleothems (also dated by U series) on Bermuda indicate that sea level was significantly lower than present at 80 and 105 ka. Vacher and Hearty (1989) pointed out that the conglomerates with the 80 ka corals were from localities on the protected side of Bermuda. They questioned the storm origin of both deposits, and suggested instead that the 105 ka ages represented scatter about true ages of ca. 125 ka, but the 80 ka corals represented a different highstand, at about 1 m higher than present, at that time.

Neumann and Moore (1975) reported U-series ages of corals from low-elevation marine deposits in the tectonically stable northern Bahamas. Among these were two samples of *Diploria* from the Berry Islands, collected from marine conglomerates found at elevations of 1.0–1.5 m above sea level. The ages of these samples are reported to be 103 ± 5 ka and 105 ± 5 ka. Unfortunately, isotopic data for these samples were not given, so it is not possible to evaluate whether these ages are reasonable. If they have experienced closed-system conditions, the data indicate that sea level ca. 105 ka was as high or higher than present. Supporting data for this idea are found on New Providence Island in the northern Bahamas, where there is a complex sequence of oolitic eolianites described by Garrett and Gould (1984). Muhs and others (1987a) analyzed some of the six eolianite units described by Garrett and Gould (1984) and reported preliminary U-series ages of ca. 105 ka for one of the units. Note that the U-series ages of ooids date the time of ooid formation, not the time of eolian deposition, although it is possible that these two events may not be greatly separated in time. In any case, the times of ooid formation on the Bahamas platform are significant for sea-level studies. Ooids form only in very shallow waters on the Bahamas platform today; most of this platform is less than about 6 m deep at present. If sea level dropped more than about 6 m below the present, most of the platform would be subaerially exposed, and ooid formation would not take place. Hence, U-series ages of ooids of ca. 105 ka imply a sea level at that time within about 6 m of the present. A sea-level stand higher than about –10 to –15 m, relative to present, ca. 100 ka is also implied by high-precision U-series ages of a submerged speleothem on Grand Bahama Island (Li and others, 1989). Broecker and Thurber (1965) reported a U-series age of 80 ± 8 ka (recalculated to be 87 ± 10 ka; see Table 2) for a coral found on a marine terrace cut into oolitic eolianite from the Berry Islands of the northern Bahamas. The elevation of this terrace is about 1 to 2 m above present sea level (N. Newell, 1990, personal commun.) and it clearly implies a sea level higher than present at the time of its formation. Thus, in the northern Bahamas, a tectonically stable setting, there is evidence for sea levels close to the present ca. 105 ka and possibly also ca. 80 ka.

The east coast of the United States is apparently undergoing a rather low long-term rate of uplift, ~0.02 m/1000 yr (Cronin, 1981), so that for the late Pleistocene, it can be considered to be relatively stable. The ages of certain marine deposits on the east coast of the United States remain controversial (Cronin and others, 1981; McCarten and others, 1982; Szabo, 1985; Corrado and others, 1986; Wehmiller and others, 1988; Hollin and Hearty, 1990), but ages of other marine deposits are agreed upon by most workers who have studied them. In Virginia and North Carolina, the Norfolk and Kempsville formations have corals in them that have been analyzed by Szabo (1985). He reported U-series ages of corals from these deposits that ranged from 66 ± 4 ka to 78 ± 5 ka (see Table 2), and correlated these units with oxygen-isotope substage 5a. Amino-acid data reported by Wehmiller and others (1988) support this correlation. These formations are currently 4–10 m above sea level (Cronin and others, 1981). Thus, if the correlation of these deposits with substage 5a is accepted, then the data imply a sea level higher than present from 66 ± 4 ka to 78 ±

5 ka. The major uncertainty here is the correlation of the deposits with substage 5a when the ages of the corals could be as young as 66 ka (see earlier discussion in this paper on the timing of sea-level highstands). In the Charleston, South Carolina, area, Szabo (1985) analyzed several corals from what he described as "late Wando" deposits. The top of this formation is currently 3–10 m above sea level (McCarten and others, 1982). Corals from this unit have relatively low $^{230}Th/^{232}Th$ values, but Szabo (1985) used several corals to derive an isochron plot that showed excellent linearity and gave an age of 88 ± 3 ka (Table 2). Szabo (1985) correlated this unit with isotope substage 5c, but a correlation with substage 5a may be more appropriate. Wehmiller and others (1988) used amino-acid ratios in shells from the same deposits to estimate an age of ca. 200 ka; they thought that the disagreement with Szabo's (1985) U-series ages was due to problems in the U-series systematics of the corals. A simpler explanation is that the shells with apparent amino-acid age estimates of ca. 200 ka are reworked from an older deposit, such as those for which Szabo (1985) reported U-series ages of coral of around 200 ka. If the ca. 80 ka age and correlation to isotope substage 5a are accepted, the elevation of the late Wando deposits implies a higher-than-present sea level at 88 ka.

The west coast of North America is tectonically active, and paleo-sea-level estimates must be made in a manner similar to that used for Barbados, New Guinea, and Haiti. Marine terraces in this area contain solitary corals or hydrocorals that have been dated by the U-series method by Ku and Kern (1974), Muhs and Szabo (1982), Muhs and others (1987b, 1988, 1990), and Rockwell and others (1989). Three localities (Point Loma and San Nicolas Island, California, and Punta Banda, Baja California, Mexico) have confidently dated 80 ka and 125 ka terraces. In addition, Punta Banda has a terrace of intermediate elevation (between the 80 ka and 125 ka terraces) that Rockwell and others (1989) interpreted to be equivalent to the 105 ka terrace found on Barbados, New Guinea, and Haiti. Assuming a sea level 6 m higher than present at 125 ka and a constant uplift rate, the elevations of the shoreline angles of these terraces indicate sea levels ~1 m lower than present at 105 ka and ~4 m lower than present at 80 ka.

Thus, data from two tectonically stable areas off North America (Bermuda and the Bahamas) and the east and west coasts of North America indicate that sea levels may have stood relatively high at both 80 ka and 105 ka, when compared to the New Guinea, Barbados, and Haiti terrace data. If sea levels at these times were in fact close to the present, the implication is that isotope substages 5a and 5c were really periods of an "interglacial" character, at least in terms of global ice volume. With relatively high sea levels (and thus low ice volumes) during isotope substages 5a, 5c, and 5e, isotope stage 5 may be viewed as a rather long interglacial period that was interrupted by two short-lived and spatially limited glaciations during substages 5d and 5b. These two short periods of ice growth are recorded as the sea-level lowstands that produced initial emergence of the 125 ka and 105 ka terraces. What is not clear, however, is why paleo-sea levels calculated from Barbados, New Guinea, and Haiti differ so significantly from those calculated in the same manner from terrace data on the west coast of North America. Resolution of this issue will require more studies on both stable and tectonically active coasts.

IS THE LAST INTERGLACIAL-GLACIAL TRANSITION FOUND IN THE COASTAL SEDIMENTARY RECORD?

Coastal stratigraphy on the California Channel Islands

Recent studies conducted on the California Channel Islands (Fig. 4) indicate that the last interglacial/glacial transition may be recorded by a shift from marine to eolian sedimentation in some coastal areas. Marine terraces on the California Channel Islands are locally overlain by calcareous eolian sand, some of which has been weakly lithified into eolianite (Fig. 5) (Orr, 1960; Vedder and Norris, 1963; Johnson, 1977; Muhs, 1983b). This cemented, carbonate-rich sand commonly occurs in areas where there is no present sand source. Even where potential beach sources exist, the eolianite has a much higher carbonate content than the adjacent beaches (Vedder and Norris, 1963; Johnson, 1972). Instead, carbonate content of the eolian sand and eolianite is closer to that of modern, sandy, insular shelf sediments of the Channel Islands. On San Nicolas Island, for example, modern shelf sediments average 49% $CaCO_3$, whereas modern beach sediments on the island average only 13% $CaCO_3$ (Vedder and Norris, 1963). In addition, the eolian deposits do not grade laterally into marine-terrace deposits, but are found on top of them, separated by a paleosol. This stratigraphic relation is significant, because it indicates that eolian sedimentation occurred after sea level was lowered. Radiocarbon ages of charcoal found in paleosols (locations 178 and 180 of Johnson [1977]) that underlie and overlie eolianite on San Miguel Island (Fig. 4) indicate that eolian sedimentation took place close to the last glacial maximum, between about 18 and 20 ka (Johnson, 1977). Collectively, these observations indicate that eolian sands were deposited during glacial periods, when sea level was significantly lower and calcareous shelf sediments were exposed to the wind (Fig. 6). The best estimate of the magnitude of sea-level lowering during the last glacial maximum comes from a recent study of submerged corals found off the island of Barbados reported by Fairbanks (1989). His data indicate that the last glacial maximum ca. 21–22 ka (Bard and others, 1990) had a sea level about 121 m lower than present. I mapped the extent of San Nicolas and San Clemente islands at the time of the last glacial maximum using the present 120 m isobath. The results indicate that large insular shelf areas would have been exposed 21–22 ka (Figs. 7 and 8).

On San Clemente Island, Muhs (1983b) mapped two eolianite units (in addition to modern, active dune sand) that can be distinguished on the basis of stratigraphic relations and degree of soil development (Fig. 9). The older unit, mapped as "Qeo" in Figure 9, is not found on the first terrace (ca. 80 ka) or the second

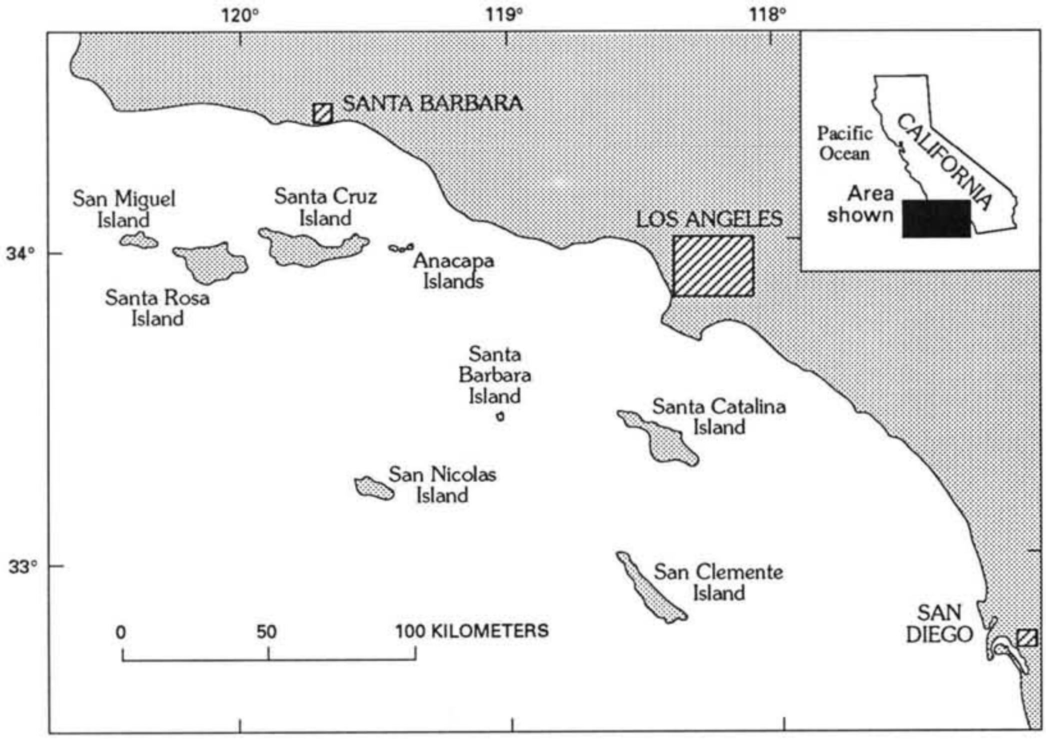

Figure 4. Map of part of the continental borderland and coastal area of southern California showing localities referred to in the text.

Figure 5. Cross-bedded eolianite exposed on Vizcaino Point, San Nicolas Island, California.

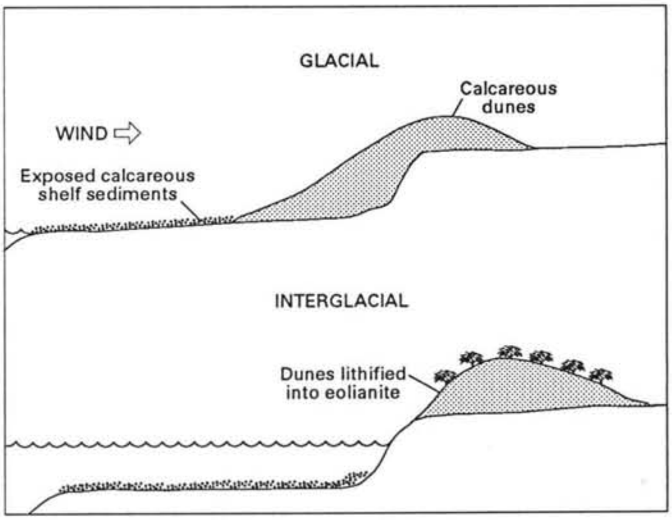

Figure 6. Model of eolianite formation during glacial periods on coasts lacking sandy beaches but which have abundant calcareous, sandy shelf sediments.

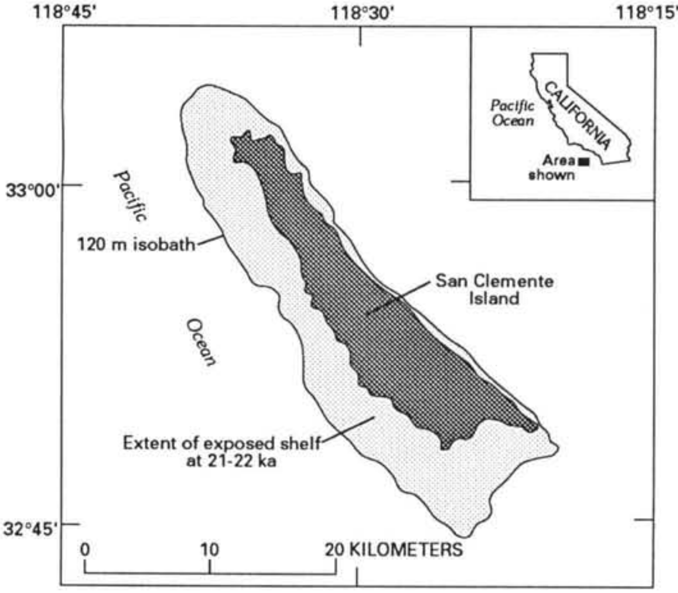

Figure 7. Shelf area on San Clemente Island, California, that would have been exposed during the last glacial maximum at 21–22 ka.

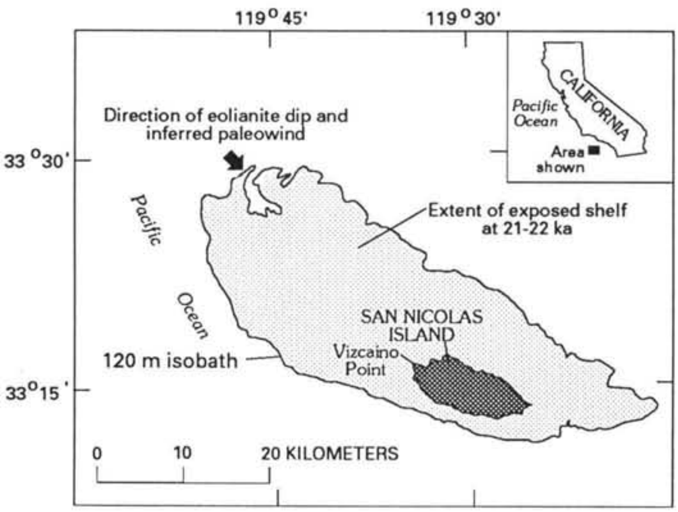

Figure 8. Shelf area on San Nicolas Island, California, that would have been exposed during the last glacial maximum at 21–22 ka.

terrace (ca. 127 ka) (Muhs, 1983b; Muhs and Szabo, 1982), but is found on many of the higher marine terraces, and is characterized by strongly developed soils with red, argillic B horizons. Thus, the stratigraphy and soil data suggest that this eolianite is older than 127 ka, perhaps significantly older. The younger eolianite (mapped as "Qey" in Fig. 9) is best exposed in modern sea cliffs on the west coast of San Clemente Island, where it is underlain by deposits of the first terrace, estimated by amino-acid ratios on shells to be about 80 ka (Muhs, 1983b). The Qey unit is subdivided into a number of members separated by paleosols (Fig. 10). The major members are shown as E2 and E3 (hereafter referred to as "lower" and "upper" members, respectively) in Figure 10. Where surface soils on the Qey unit are found, they are characterized by minimally developed B horizons that lack red colors and have significantly lower clay contents than the soils of the Qeo unit. Muhs (1983b) reported a ^{14}C age of ca. 22,000 yr B.P. for land snails found in the P3 paleosol, but this is probably a minimum age. Sands from all these eolian units were collected for thermoluminescence (TL) dating, but only the sediments from unit E3 had a sufficient amount of 4–11 μm material for analysis. This sample (Alpha #2611) yielded the following values: U, 1.3 ppm; Th, 1.0 ppm, K_2O, 0.37%; "a" value, 0.117; total dose rate, 0.95 rad/yr; equivalent dose, 1.168 krad; age estimate (using 75% of the laboratory measured saturation water content), 12,200 ± 1800 yr B.P. It is not known whether this silt-sized material was deposited at the same time as the more abundant sand-sized sediment, or if it is a secondary eolian silt blown in from a more distant source, such as the California mainland (Muhs, 1983c).

Calcareous eolian sands, locally lithified into eolianite, are also exposed in cliff sections around Vizcaino Point on San Nicolas Island (Fig. 8). Two eolian units overlie a paleosol developed on the first marine terrace (Fig. 11). On the basis of the similar stratigraphy, I correlate the two eolian units on San Nicolas Island with the two major eolianite units overlying the first marine terrace on San Clemente Island (units E2 and E3 in Fig. 10).

Uranium-series ages of rhizoliths found in California eolianites

On both San Nicolas Island and San Clemente Island, secondary carbonates are found that have a similar relation to the eolian sands. Near the upper part of each eolianite or eolian sand unit, rhizoliths (carbonate root casts) and horizontal pedogenic calcretes mark intervals of eolian sand stabilization and pedogenesis (Fig. 12). Eolian sedimentation should have ceased when sea-level rise progressed far enough to flood the shelf source areas. Because rhizoliths record times of vegetation stabilization when source areas had been cut off, they may record minimum ages for the underlying eolian sands.

Because rhizoliths on the California Channel Islands are relatively pure (up to 99% $CaCO_3$) secondary carbonates and are very dense, they are potentially suitable for U-series dating. X-ray diffraction analyses showed that the interiors of all but one of the rhizoliths studied are composed of calcite that lack primary (shell-derived) aragonite. This is significant because it indicates that, in all but one case, the rhizoliths are completely secondary and U-series ages should reflect the time of calcite precipitation. Only data for those rhizoliths that completely lack aragonite are reported here. Rhizoliths were collected for U-series dating from the three eolianite units on San Clemente Island (Qeo, and Qey, lower and upper) and the two eolian units that overlie the first marine terrace on San Nicolas Island. In addition, solitary corals (*Balanophyllia elegans*) were found in the first terrace deposits at

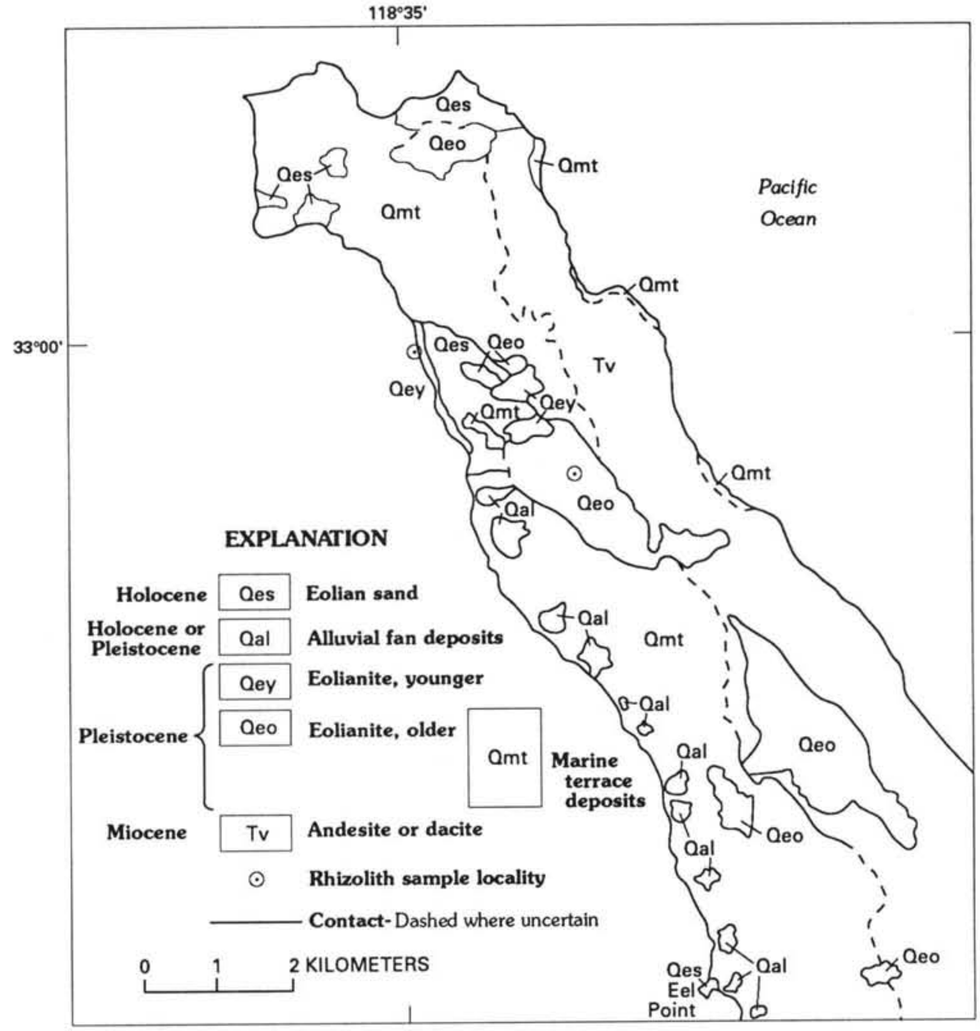

Figure 9. Geologic map of Quaternary deposits on northern San Clemente Island, California, showing rhizolith sample localities. Geologic map from Muhs (1983b).

Vizcaino Point on San Nicolas Island. No corals were found in the first terrace deposits on San Clemente Island.

Uranium-series ages were determined for the marine terrace corals and eolianite rhizoliths by isotope-dilution alpha spectrometry. Only the relatively pure interiors of rhizoliths were sampled for U-series analyses. All samples were cleaned mechanically prior to analysis to remove adhering detrital mineral grains. Samples were then powdered and heated to 900 °C for about 8 hrs to destroy organic matter and convert the $CaCO_3$ to CaO. The coral samples and some of the rhizoliths were then dissolved in 6*N* HCl and allowed to equilibrate with a combined ^{229}Th-^{236}U spike. After equilibration, a ferric nitrate carrier was added and the isotopes of U and Th were coprecipitated with NH_4OH. The precipitates were centrifuge washed, dissolved in 6*N* HCl, and loaded onto an ion-exchange column in chloride form to separate U from Th. The U and Th separates were taken to dryness and then dissolved in 8*N* HNO_3 and loaded onto ion-exchange columns in nitrate form for further purification. The purified U and Th separates were then taken to dryness. The thorium residue was then dissolved with about 0.2 ml of 0.1*N* HNO_3 and 0.1 ml of thenoyltrifluoroacetone (dissolved in benzene) was added to the solution. The solution was then agitated to transfer the Th into the organic phase; this was then transferred dropwise onto a heated stainless-steel disk. The U residue was dissolved in a drop of 6*N* HCl, mixed with about 1.5 ml of NH_4Cl buffer and adjusted to a pH of about 6 with NH_4OH. The solution containing the U was then added to a Teflon plating apparatus and electroplated onto a stainless-steel disk over a period of about 0.5 h at a current of 1.8 A. The disks containing the U and Th samples were then placed in an alpha spectrometer for counting. After some samples had been counted sufficiently, it was concluded that some detrital materials had remained in the samples after cleaning; this was apparent from $^{230}Th/^{232}Th$ activity ratios lower than about 20 (Table 3). In such cases, some of

Figure 10. Cliff exposure of eolianite at Vizcaino Point, San Nicolas Island, showing horizontal calcrete underlain by carbonate rhizoliths.

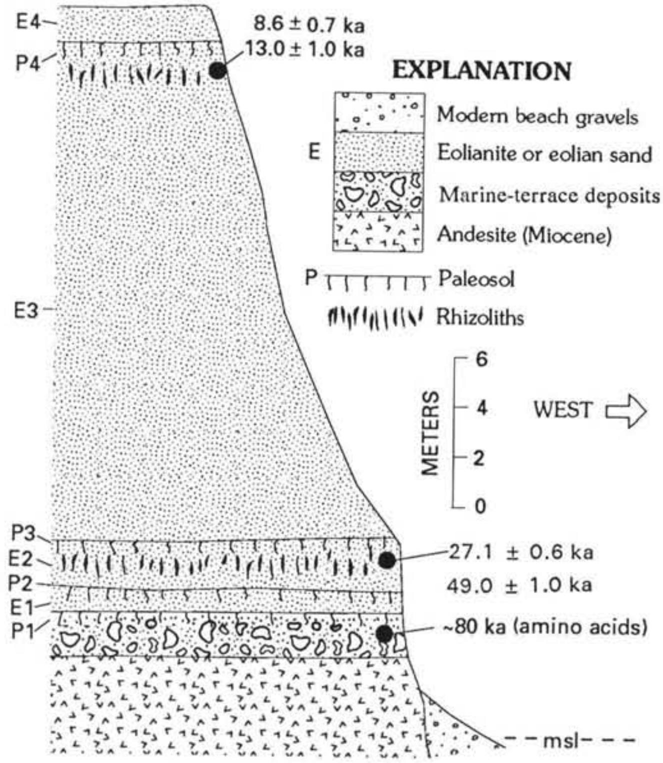

Figure 11. Cliff section exposed on the northwest coast of San Clemente Island, California, showing U-series ages of rhizoliths (see Table 3). Stratigraphy from Muhs (1983b).

the activity of ^{230}Th is contributed by the detrital minerals rather than all of the ^{230}Th activity being due to in situ radioactive decay of ^{234}U in the carbonate. This problem was minimized on all subsequent samples by gradual dissolution of the carbonate using dilute (0.25*N*) HNO_3. However, even with dilute HNO_3 leaching, some U and Th can still be leached from the detrital minerals. In the event that the ^{230}Th/^{232}Th ratios are still less than about 20 after dilute acid leaching, a correction can be made (using the method outlined by Ku and Liang, 1984) by a separate analysis of the residue (if a sufficient amount is present) or the host eolian sediment, which is the obvious source of the contaminating material. In the case of the San Clemente Island rhizoliths, only the samples from the youngest unit required corrections. Hence, samples of the appropriate host eolian sediment were analyzed, using a complete decomposition with an HNO_3-$HClO_4$-HF mixture; corrections were made using the mixing-line plot method of Ku and Liang (1984) with the isotopic composition of the host sediment as a proxy for the residue (Table 4). For all samples reported in Tables 3 and 4, errors for activity ratios and ages are based on counting statistics and are 1 sigma.

Uranium-series analysis of the fossil corals indicate that the first terrace on San Nicolas Island is 80 ± 2 ka and thus correlates with ca. 80 ka terraces reported from Barbados, New Guinea, Haiti, northern Baja California, and other areas discussed earlier. The 80 ka age is reasonable because these corals have (1) U concentrations that are typical for this species (Ku and Kern, 1974; Rockwell and others, 1989; Muhs and others, 1990); (2) ^{230}Th/^{232}Th ratios that are significantly greater than 20, thus requiring no corrections for detrital materials; and (3) ^{234}U/^{238}U ratios that are consistent with an 80 ka age and initial ^{234}U/^{238}U ratios in sea water of 1.14–1.15, within limits of analytical uncertainty.

Uranium-series ages of rhizoliths are in agreement with the stratigraphy and relative ages of the deposits and indicate that rhizoliths are suitable for U-series dating (Table 3 and Figs. 10 and 11). Within a given eolian unit, there is a significant range of rhizolith ages, but this is expected because a rhizolith can form during any time period that a surface is stabilized with vegetation. The oldest mapped eolianite on San Clemente Island is the Qeo unit, which is not found on the first (ca. 80 ka) or second (ca. 127 ka) marine terraces, but is found on most of the higher terraces. Rhizoliths B and D from this unit give ages of 152 ± 5 ka and 166

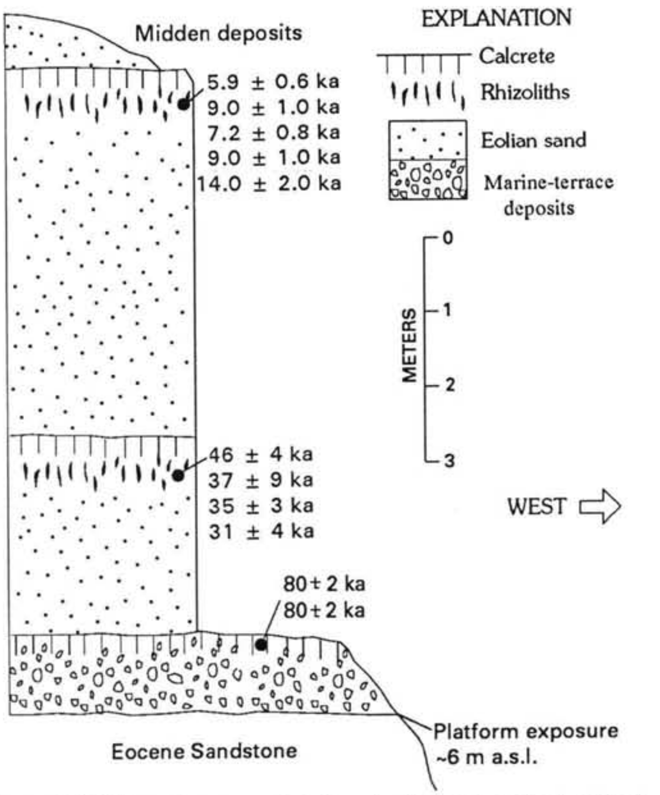

Figure 12. Cliff section exposed at Vizcaino Point, San Nicolas Island, California, showing U-series ages of rhizoliths and marine terrace corals (see Table 3).

± 10 ka (Table 3), in agreement with the stratigraphic relation to the marine terraces. Rhizolith A has an apparent age of 21 ka, but it also has a U concentration that is a factor of 4 to 5 greater than the other rhizoliths from the Qeo unit, and up to an order of magnitude higher than that found in the younger rhizoliths. Hence, it is likely that this rhizolith has undergone some recent, secondary U uptake. More U-series ages are needed from this unit, but these preliminary data indicate that eolian deposition could have taken place during the lowstands of sea level preceding the formation of the second terrace (ca. 127 ka) or the third terrace (ca. 200 ka) (Muhs and Szabo, 1982). The younger Qey eolianite on San Clemente Island is subdivided into "lower" and "upper" members on the basis of the stratigraphy exposed on the west coast (Fig. 10). U-series ages of rhizoliths support this relative age relation (Table 3 and Fig. 10). The lower member was apparently deposited between ca. 80 ka (the estimated age of the underlying marine terrace deposits) and 49 ± 1 ka, the age of the oldest rhizolith in this unit. The surface of this eolianite was apparently stabilized by vegetation at least until 27 ka, the age of the youngest rhizolith. Some time after 27 ka, the upper member of Qey (shown as E3 in Fig. 10) was deposited and stabilized by 13 ± 1 ka, the age of the oldest rhizolith from this unit, in agreement with the TL age estimate of 12 ± 2 ka of the 4–11 μm fraction of the deposit.

On San Nicolas Island, U-series ages of rhizoliths support the correlation of the two eolian sand units overlying the 80 ka marine terrace with the upper and lower members of the Qey eolianite on San Clemente Island (Table 3 and Fig. 11). The lower member on San Nicolas Island was deposited between 80 ± 2 ka and 46 ± 4 ka, or about the same time as the lower, E2 eolianite on San Clemente Island. The surface of this unit was stabilized by vegetation at least until about 31 ± 4 ka, the age of the youngest rhizolith. Eolian sedimentation began again sometime after ca. 31 ka and continued until 14 ± 2 ka, the age of the youngest rhizolith in the upper eolian unit.

Inferred late Quaternary sea-level history on the California Channel Islands

Uranium-series dating of these eolianites indicates a sequence of events that agrees with sea-level history recorded elsewhere. After the relatively high sea-level stand ca. 80 ka, sea level lowered and what is now the first marine terrace on both islands emerged. After initial emergence, and while sea level was still dropping, soils formed on the marine-terrace deposits (Figs. 10 and 11). When sea level had lowered sufficiently to expose calcareous shelf sands, eolian sedimentation began and resulted in the deposition of the lower eolian members. Sea level apparently rose sufficiently to cut off the source of the sands sometime around 46–49 ka, because this is when rhizolith formation, indicating vegetation stabilization, began on both islands. This period of stabilization continued until about 27–31 ka, the age of the youngest rhizoliths in the lower eolian members. On New Guinea, reef complexes IV (ca. 60 ka) IIIa (ca. 40–50 ka), IIIb (ca. 40 ka), and II (ca. 28 ka) record four relatively high sea-level stands during the interval from 60 to 28 ka. Calculations of the position of sea level at these times are in broad agreement by both the Bloom and Yonekura (1985) and Chappell and Shackleton (1986) methods, and indicate sea levels as high as −24 to −28 m, relative to present, ca. 60 ka, and as low as −35 to −44 m, relative to present, ca. 28 ka. If sea levels were this high, significant portions of the shelf areas of both islands would be submerged, and it is probable that most eolian sand sources would be cut off. Sometime after 27–31 ka, sea level dropped again, and eolian sedimentation was renewed and continued until 13–14 ka. These observations agree with those made by Johnson (1977) of an eolianite deposited between 18 and 20 ka on San Miguel Island, California. Recent studies by Fairbanks (1989) of submerged reefs off Barbados indicate that sea level was around 121 m lower than present at the last glacial maximum. The last glacial maximum is now estimated to be 21–22 ka by high-precision U-series dating of corals (Bard and others, 1990). The bracketing U-series ages of rhizoliths of 27–31 ka and 13–14 ka from San Nicolas and San Clemente islands are in broad agreement with these new age estimates of the last glacial maximum from Barbados, and indicate that major eolian sedimentation took place during the last glacial maximum, when sea level was lowered ~121 m and broad shelf areas were exposed on both islands (Figs. 7 and 8).

TABLE 3. URANIUM CONCENTRATIONS, ISOTOPIC ACTIVITY RATIOS, AND U-SERIES AGES OF CALIFORNIA CHANNEL ISLANDS RHIZOLITHS AND CORALS

Locality, Geologic Unit (Sample)*	U (ppm)	$^{234}U/^{238}U$	$^{230}Th/^{232}Th$	$^{230}Th/^{234}U$	Uncorrected Age† (ka)	Corrected Age§ (ka)
		Activity Ratios				
San Clemente Island, Qey, upper						
R, A (HNO_3 leach)	0.646 ± 0.008	1.07 ± 0.01	7.7 ± 0.4	0.110 ± 0.002	12.6 ± 0.3	8.6 ± 0.7
R, B (HNO_3 leach)	0.715 ± 0.009	1.08 ± 0.01	17 ± 1	0.127 ± 0.002	14.7 ± 0.3	13 ± 1
San Nicolas Island, upper eolian sand						
R, A (HCl leach)	0.86 ± 0.01	1.10 ± 0.01	2.2 ± 0.1	0.130 ± 0.004	15.1 ± 0.5	5.9 ± 0.6
R, A (HNO_3 leach)	0.82 ± 0.01	1.12 ± 0.01	3.4 ± 0.3	0.129 ± 0.007	15.0 ± 0.9	9 ± 1
R, B (HCl leach)	0.85 ± 0.01	1.08 ± 0.01	3.0 ± 0.2	0.114 ± 0.004	13.1 ± 0.5	7.2 ± 0.8
R, B (HNO_3 leach)	0.82 ± 0.01	1.10 ± 0.02	3.8 ± 0.2	0.121 ± 0.004	14.0 ± 0.5	9 ± 1
R, C (HNO_3 leach)	0.577 ± 0.008	1.09 ± 0.02	8 ± 1	0.142 ± 0.006	16.6 ± 0.8	14 ± 3
San Clemente Island, Qey, lower						
R, A (HNO_3 leach)	0.687 ± 0.008	1.128± 0.009	23 ± 2	0.222 ± 0.004	27.1 ± 0.6	27.1 ± 0.6
R, B (HCl leach)	0.726 ± 0.009	1.16 ± 0.01	21.6 ± 0.8	0.367 ± 0.005	49 ± 1	49 ± 1
San Nicolas Island, lower eolian sand						
R, A (HCl leach)	0.667 ± 0.008	1.11 ± 0.01	4.2 ± 0.1	0.417 ± 0.006	58 ± 1	46 ± 4
R, A (HNO_3 leach)	0.597 ± 0.008	1.13 ± 0.01	17 ± 2	0.304 ± 0.008	39 ± 1	37 ± 9
R, C (HCl leach)	0.86 ± 0.01	1.13 ± 0.01	7.1 ± 0.2	0.316 ± 0.005	41 ± 1	35 ± 3
R, C (HNO_3 leach)	0.77 ± 0.01	1.11 ± 0.01	17 ± 2	0.262 ± 0.006	33 ± 1	31 ± 4
San Nicolas Island, first marine terrace						
Coral, A (B. e.)	4.78 ± 0.07	1.14 ± 0.01	64 ± 7	0.530 ± 0.009	80 ± 2	80 ± 2
Coral, B (B.e.)	4.19 ± 0.06	1.12 ± 0.01	26 ± 1	0.529 ± 0.009	80 ± 2	80 ± 2
San Clemente Island, Qeo						
R, A (HNO_3 leach)	6.45 ± 0.08	1.046± 0.004	81 ± 5	0.178 ± 0.002	21.2 ± 0.3	21.2 ± 0.3
R, B (HNO_3 leach)	1.64 ± 0.02	1.042± 0.008	92 ± 6	0.76 ± 0.01	152 ± 5	152 ± 5
R, D (HNO_3 leach)	1.34 ± 0.02	1.12 ± 0.01	>800	0.80 ± 0.02	166 ±10	166 ±10

*R = Rhizolith. Corals are *Balanophyllia elegans* (B.e.); A was collected from Los Angeles County Museum of Natural History locality number 10621 and B was collected from locality number 11009.
†Calculated using half-lives of ^{230}Th and ^{234}U of 75,200 and 244,000 years, respectively.
§Corrected for inherited ^{230}Th using isotopic composition of the host sediment (Table 4) and two-point mixing-line plot (Ku and Liang, 1984) when ^{230}Th ratio is less than 20.

TABLE 4. URANIUM AND THORIUM ISOTOPIC COMPOSITION OF WHOLE-SEDIMENT EOLIAN SANDS USED FOR AGE CORRECTIONS

Locality, Geologic Unit U (ppm)	Th (ppm)	$^{234}U/^{238}U$	$^{230}Th/^{232}Th$ Activity Ratios	$^{230}Th/^{234}U$
San Clemente Island, Qey, upper eolianite				
0.82 ± 0.01	0.96 ± 0.03	1.03 ± 0.1	2.59 ± 0.09	0.95 ± 0.02
San Nicolas Island, upper eolian sands				
1.07 ± 0.01	1.72 ± 0.04	1.04 ± 0.01	1.45 ± 0.04	0.73 ± 0.02
San Nicolas Island, lower eolian sands				
0.90 ± 0.01	1.71 ± 0.03	1.07 ± 0.01	1.30 ± 0.02	0.74 ± 0.01

Thus, eolian sands on the southern Channel Islands record two periods of glaciation, an advance sometime between ca. 80 and 49 ka, and an advance sometime between 27–31 ka and 13–14 ka. Additional U-series dating of rhizoliths may narrow these age ranges significantly. Inferred times of low sea level (glacials) from the Channel Islands eolianite record agree well with times of lowered summer insolation in high latitudes, whereas times of marine-terrace formation (interglacials) agree well with periods of higher summer insolation in high latitudes (Fig. 13).

The marine-terrace record indicates that sea levels were low ca. 115 and 95 ka (Fig. 3). Were sea levels low enough at these times to expose the insular shelves of the Channel Islands and provide a source of sediment for eolianite formation? This question cannot be answered with the available data, but I have observed two eolianite units overlying the second, ca. 125 ka terrace at Vizcaino Point on San Nicolas Island. These two eolianites may correlate with the two eolian units overlying the ca. 80 ka terrace, or they may record earlier (ca. 115 and 95 ka) lowstands of sea level. Future U-series analyses should answer this question.

If the ca. 80 ka highstand of sea level is considered to be part of a "long interglacial" that includes all of deep-sea isotope stage 5, then the shift from marine to eolian sedimentation recorded by the coastal deposits on the California Channel Islands reflects the last interglacial/glacial transition. Such a shift in the type of sedimentation should be recorded on other coasts as well and invites comparison to the California Channel Islands. The model of eolianite formation during glacial periods is supported by observations in other localities, where eolianites overlie marine-terrace deposits and/or eolianites extend below sea level. These relations have been observed in South Africa (Marker, 1976), Australia (Fairbridge and Johnson, 1978; Sprigg, 1979), Hawaii (Stearns, 1978), Mallorca (Butzer and Cuerda, 1962; Butzer, 1975), Lebanon (Wright, 1962), and Puerto Rico (Kaye, 1959). There are, however, other localities where eolianites grade laterally into beach or marine deposits, such as in Bermuda and the Bahamas (Land and others, 1967; Garrett and Gould, 1984), indicating formation of eolianite during highstands of sea level. As Gardner (1983) pointed out, it is probably unwise to postulate a single relation of eolianite formation to sea level for all areas.

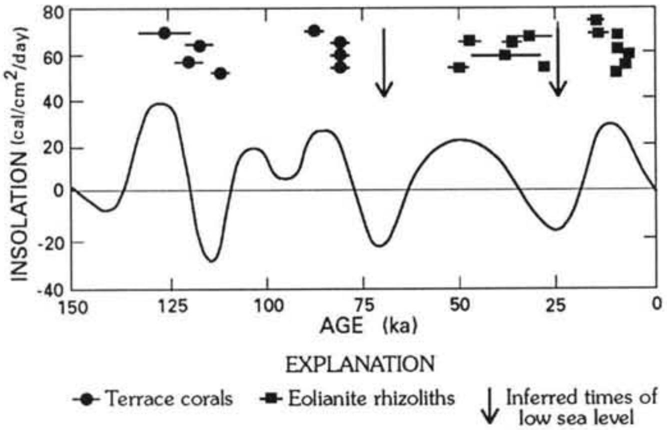

Figure 13. Plot of U-series ages of eolianite rhizoliths and marine-terrace corals from San Nicolas and San Clemente Islands, California (data from Table 3; Muhs and Szabo [1982]; and updated from Muhs and others [1987b]). Vertical arrows show times of inferred low sea level (and also continental glaciation) based on the distribution of U-series ages. Shown for comparison is a plot of average insolation received at the top of the atmosphere at lat 65°N for the summer half year, expressed as deviations from the present (A.D. 1950) value. Insolation data are from Berger (1978).

Nevertheless, the results presented here suggest that with careful field work and a rigorous dating program, a valuable record of the last interglacial/glacial transition could be found on many subtropical and tropical coasts.

CONCLUSIONS

Compilation of reliably dated corals from terraces that have been correlated with deep-sea oxygen-isotope stage 5 indicate, in general, three relatively high stands of sea level ca. 125–120 ka, 105 ka, and 85–80 ka. These times of high sea level are in agreement with calculated periods of high summer insolation at high latitudes, which would have favored global ice melting and resulted in relatively high sea levels. On both Barbados and the east coast of the United States, there are a number of corals from marine deposits that are apparently reliable and indicate a high sea level ca. 70 ka. Studies from both relatively stable coasts and tectonically emergent coasts indicate that there may have been a high sea level at 145–135 ka, but data are sparse and stratigraphic evidence is lacking, except on New Guinea. However, high sea levels at both 70 ka and 145–135 ka represent potential challenges to the Milankovitch orbital-forcing theory of climatic change, because summer insolation at high latitudes was low at these times, and would have favored glaciation and lowered sea levels. What is clearly needed to help resolve these issues are high-precision mass-spectrometric U-series analyses of corals for dating and careful studies of corals to investigate possible subtle diagenesis and U migration, using scanning-electron microscopy and fission-track mapping.

There is disagreement among investigators from different fields concerning the length of the last interglacial. Many glacial geologists and most marine scientists favor a "short" last interglacial (correlated to deep-sea substage 5e only), whereas some pedologists, speleothem investigators, and ice-core workers favor a "long" interglacial (most or all of deep-sea stage 5). The marine terrace record has added to the confusion, because whether the last interglacial was long or short depends on how high sea level was ca. 105 and 80 ka; records from different coasts disagree on the elevations reached by the sea at these times. Paleo-sea-level elevations generated by an assumption of constant uplift rate during the past 125 ka on Barbados, New Guinea, and Haiti all indicate sea levels significantly lower than present at 105 ka and 80 ka. New Guinea paleo-sea levels calculated by an alternative method indicate sea levels close to the present at these times. On the east and west coasts of North America and on Bermuda and the Bahamas, there is also evidence for sea levels close to the present or even higher ca. 105 and 80 ka. Taken as a whole, these results indicate that more studies are needed on other coasts before we can make confident statements about sea-level elevations during the last interglacial/glacial cycle.

Coastal deposits on the California Channel Islands record a shift from marine to eolian sedimentation at the last interglacial/glacial transition if it is assumed that the last interglacial was "long," and ended sometime after ca. 80 ka. Marine terrace deposits contain solitary corals that have been dated ca. 80 ka by U-series methods or have shells that give amino-acid ratios indicative of an age of ca. 80 ka. A paleosol has developed on these deposits, and this soil is in turn overlain by two eolian sand or eolianite units, separated by another paleosol. Carbonate rhizoliths in the eolian units give U-series ages that indicate two episodes of sedimentation, one between ca. 80 ka and 49–46 ka, and another between 31–27 ka and 14–13 ka. These eolian sands could only have been deposited when sea level was low and shelf sands were exposed, because there are not sufficient beach sediments at present to account for the volume of eolian sand. Greatly lowered sea levels imply glacial conditions, and the times of lowered sea level on the California Channel Islands, as indicated by the bracketing U-series ages, agree well with times of lower summer insolation at high latitudes. These observations indicate that marine terrace/eolianite sequences may provide a valuable record of the last interglacial/glacial transition, and studies should be initiated on other coasts where such deposits exist.

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