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Timing and warmth of the Last Interglacial period: new U-series evidence from Hawaii and Bermuda and a new fossil compilation for North America

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Abstract

The timing and duration of the Last Interglacial period have been controversial, with some studies suggesting a relatively short duration that is orbitally forced and others suggesting a long duration that is at most only partly related to orbital forcing. New, high-precision thermal ionization mass spectrometric (TIMS) U-series ages of Last Interglacial corals from Hawaii and Bermuda test these competing hypotheses. Waimanalo Formation corals from slowly uplifting Oahu, Hawaii range in age from ~ 134 to ~ 113 ka, with most ages between ~ 125 and ~ 115 ka. Combined with published U-series ages from nearby Lanai, the data suggest a long Last Interglacial period that may have occurred from ~ 136 to at least 115 ka. The results indicate that orbital forcing may not have been the only control on ice sheet growth and decay, because sea level would have been high at times of relatively low Northern Hemisphere summer insolation. On tectonically stable Bermuda, deposits from the ~ 200 ka (penultimate interglacial period), ~ 120 ka (peak Last Interglacial period) and ~ 80 ka (late Last Interglacial period) high sea stands have been newly dated. Fossil corals on Bermuda are derived from patch reefs that likely were “catch-up” responses to sea level rise. It is expected that U-series ages of Last-Interglacial corals on Bermuda should overlap with, but not be as old as the range of corals on Oahu. Last-Interglacial corals on Bermuda give a range of ~ 125 –113 ka, which supports this hypothesis.

A large number of emergent marine deposits on Hawaii, Bermuda and along coastal North America have now been dated to the Last Interglacial period. Both Oahu and Bermuda have marine invertebrate faunas with a number of extralimital southern species of mollusks, suggesting warmer-than-present waters during the Last Interglacial period. Warmer waters are also suggested for Last-Interglacial localities around most of North America, from Florida to Canada and Greenland and Baja California to Alaska. These observations are consistent with similar warm-water faunas of Last-Interglacial age reported from Japan, the Mediterranean basin and Western Australia. It is likely that significant changes in ocean currents took place during the Last Interglacial period, with a movement of relatively warm waters to higher latitudes than is the case today. © 2002 Elsevier Science Ltd. All rights reserved.

1. Introduction

The precise timing, duration and paleoclimatic conditions of the peak of the Last Interglacial period have been actively debated. It is an important time period to study because understanding its causes, duration and paleoclimatic conditions may provide clues to climates of the future (Broecker, 1998). Interpretation of the oxygen isotope record of foraminifera in deep-sea cores

suggests that the Last Interglacial period (known as substage 5e) began around 127 ka, was relatively short (on the order of a few thousand years) and was orbitally forced (CLIMAP Project Members, 1984; Martinson et al., 1987). However, records from deep-sea cores (McManus et al., 1994), pollen from Grand Pile, France (Kukla et al., 1997), vein calcite from Devils Hole, Nevada (Winograd et al., 1992, 1997), Antarctic ice cores (Lorius et al., 1985; Jouzel et al., 1993) and emergent reefs on Hawaii (Szabo et al., 1994) suggest that the Last Interglacial period could have been much longer, on the order of 20 ka. In addition, Henderson and Slowey (2000) directly dated marine sediments that have an interglacial oxygen isotope signature off the

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Bahamas using a U–Th isochron method. They proposed that the Last Interglacial could have begun as much as 8 ka earlier than the ~127 ka starting date previously suggested (CLIMAP Project Members, 1984 and Martinson et al., 1987). If this earlier age estimate is correct, it challenges the hypothesis that the Last Interglacial period was orbitally forced.

Despite the challenges to the widely accepted deep-sea record of the Last Interglacial period, there are problems with many of the studies cited above. The Grand Pile pollen record is not dated directly and the chronology proposed by Kukla et al. (1997) is based on a correlation to the deep-sea record. Direct U–Th dating of sediments off the Bahamas by Henderson and Slowey (2000) is complicated by relatively high amounts of inherited ^{230}Th and therefore requires a model for correction of apparent ages. As pointed out by McCulloch and Esat (2000), many of the U-series ages of corals from Hawaii reported by Szabo et al. (1994) have relatively large analytical uncertainties. Antarctic ice cores contain neither direct records of global ice volume, nor are they dated directly. The Devils Hole, Nevada vein calcite record has a robust suite of U-series ages for chronological control, but the oxygen isotope record in this calcite does not measure global ice volume even indirectly. Indeed, Winograd et al. (1997) have pointed out that the oxygen isotope record at Devils Hole is a temperature record that may well precede that of the deep-sea oxygen isotope record.

Sea level records on different coastlines could very well differ in their estimates of the duration of the Last Interglacial, due both to tectonics and the rate of reef growth. Edwards et al. (1987) pointed out that the timing of reef growth during an interglacial period is a function not only of the rate of sea level rise, but also of the rate of uplift (Fig. 1). On a tectonically active, rapidly rising coastline, vertical reef accretion can take place when the rate of sea level rise slightly exceeds the rate of uplift. To this can be added the complicating effects of reef type, which consist of “keep-up,” “catch-up” or “give-up” types (Neumann and MacIntyre, 1985). This concept was applied by Neumann and MacIntyre (1985) to Atlantic and Caribbean reefs, but has also been useful in understanding reefs in the Indo-Pacific province (Cabioch et al., 1999). Sea level rise in a deglacial period can be rapid. Rates of 2–7 mm/yr occurred during most of the last deglacial period on Barbados and meltwater “pulses” of up to 24 m occurred in less than 1000 radiocarbon years (Fairbanks, 1989). Certain types of corals may have matching growth rates and others may not. Fringing or barrier reefs dominated by *Acropora palmata* communities (such as those on uplifting Barbados or Haiti) can “keep up” with rising sea level and reef growth may continue after sea level has stabilized. Whereas most corals grow at rates of 1–10 mm/yr (Grigg, 1997),

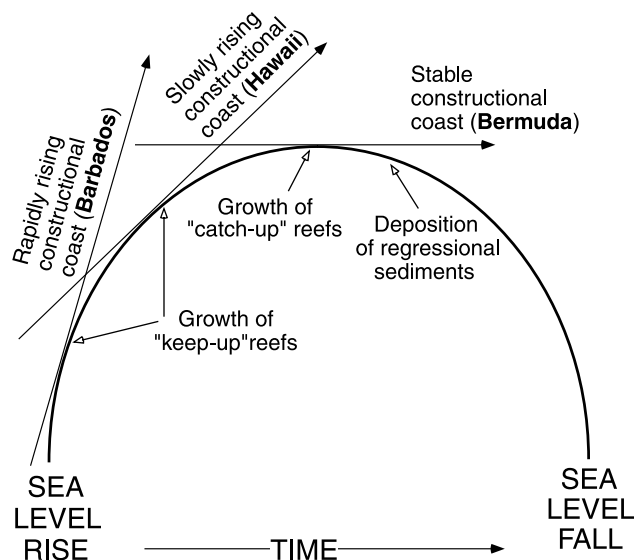


Fig. 1. Model of interglacial sea level rise and fall with hypothesized timing of different reef records. Modified from Bradley and Griggs (1976) and Edwards et al. (1987) using reef classification nomenclature of Neumann and MacIntyre (1985).

Acropora palmata has measured rates of 13–15 mm/yr (Buddemeier and Smith, 1988). Other coral communities, such as patch reefs, are referred to as “catch-up” types, whose development may be delayed until sea level has stabilized. Patch reefs are the predominant types on Bermuda (Garrett et al., 1971). On Oahu, Hawaii, fringing reefs are present. During the mid-to-late Holocene, vertical accretion rates of Hawaiian reefs were about 3 mm/yr (Easton and Olson, 1976). Given the model of Neumann and MacIntyre (1985), fossil reefs on Oahu may provide a fairly complete record of the Last Interglacial high sea stand. On Oahu, it can be hypothesized that reef development may keep pace with rising sea level during at least part of a deglacial period (thus potentially allowing a record of the early part of a sea-level rise). In addition, relatively slow uplift (Muhs and Szabo, 1994) may allow for maintenance of reef structures after sea level has stabilized.

It is important to note, however, that in both fringing reef and barrier reef environments, there is likely a lag between the time of initial sea level rise and vigorous, upward reef accretion. Lack of a suitable substrate, absence of larval replenishment centers, restricted circulation, poor water quality (highly turbid conditions), proximity to terrestrial influences and other factors can all result in a delay between the time of sea level rise and vertical reef accretion (Hopley, 1982, 1994). For example, in the Great Barrier Reef, Davies and Hopley (1983) report a hiatus of up to 2000 yr between the time of Holocene sea level rise and the time of reef colonization. Many Holocene reefs thus date

their initial growth to 7000–8000 radiocarbon yr BP. In the Caribbean Sea and western Atlantic Ocean, growth of corals in the early Holocene was terminated by stress conditions induced by flooding of shelves as sea level rose; coral growth resumed when conditions improved around 7000 radiocarbon yr BP (MacIntyre, 1988). Thus, the oldest Holocene *Acropora palmata* (a “keep-up” species) on fringing reefs off Panama and St. Croix date to around 5000–7000 radiocarbon yr BP. Studies of a Holocene reef on Oahu show that reef growth also began about 7000 radiocarbon yr ago, and was characterized by its most rapid vertical accretion from about 5800–3500 radiocarbon yr ago (Easton and Olson, 1976). The period of rapid vertical accretion was followed by a period of seaward accretion (since 3000 radiocarbon yr ago).

In this paper, we present new thermal ionization mass spectrometric (TIMS) U-series ages of pristine corals from marine deposits on Oahu, Hawaii and the tectonically stable island of Bermuda. We hypothesize that on a slowly rising, fringing-reef island setting such as Oahu, there should be a fairly complete record of a high sea stand such as the Last Interglacial period. We further hypothesize that on a tectonically stable island such as Bermuda, characterized mostly by patch reefs, past sea level records should overlap with, but not be as old as those of tectonically rising islands with fringing reef communities. Although there have already been extensive alpha-spectrometric U-series dating efforts on Oahu (Ku et al., 1974; Muhs and Szabo, 1994) and Bermuda (Harmon et al., 1978, 1983), testing of these hypotheses requires high-precision TIMS U-series dating. Szabo et al. (1994) reported a large suite of TIMS U-series ages of corals from Oahu, but two-sigma analytical uncertainties of these ages range from 1 to 6 ka. A rigorous testing of the hypotheses regarding the duration of the Last Interglacial period requires U-series ages with smaller analytical uncertainties. The new TIMS U-series ages reported here are from corals either not previously analyzed by Szabo et al. (1994) or analyzed only by alpha spectrometry by Muhs and Szabo (1994); all have improved analytical uncertainties.

We also compiled fossil faunal data (mostly mollusks) from marine deposits of Oahu and Bermuda, and compared these with their modern geographic ranges to assess the degree of zoogeographic shift during the Last Interglacial period. The paleoclimatic interpretations made from the Last-Interglacial faunal data of Oahu and Bermuda are also compared to those from marine deposits around North America. Examination of this compilation allows a test of the hypothesis that the Last Interglacial period was only about as warm as the present interglacial (CLIMAP Project Members, 1984; McCulloch and Esat, 2000) or at most slightly warmer (Montoya et al., 1998).

2. Study areas

2.1. Oahu, Hawaii

Oahu is part of the Hawaii-Emperor volcanic chain that extends across 6000 km of the Pacific Ocean. The best estimates of the age of the main volcanic rocks on Oahu are 2.6 and 3.7 Ma for the Koolau and Waianae Ranges, respectively (Clague and Dalrymple, 1987). Although the surficial geology of Oahu is dominated by Tertiary basalt, Pleistocene reef limestones fringe much of the island (Fig. 2). Some, although not all, of this limestone belongs to the Waimanalo Formation of Lum and Stearns (1970), which dates to the Last Interglacial period (Ku et al., 1974; Muhs and Szabo, 1994; Szabo et al., 1994). The Waimanalo Formation consists of corals in growth position, usually overlain by poorly sorted marine conglomerate with coral and basalt clasts. In some places, the reef facies is not present and limestones consist only of coral-basalt conglomerate resting on a marine bench or shore platform cut into Tertiary basalt.

The “Big Island” of Hawaii is situated near the still-active volcanic hot spot that created the Hawaii-Emperor island chain. Crustal subsidence of this island due to volcanic loading has been well documented (Szabo and Moore, 1986; Moore, 1987; Watts and ten Brink, 1989; Ludwig et al., 1991). This subsidence is hypothesized to have created a compensatory, upward lithospheric flexure on Oahu (Muhs and Szabo, 1994). Thus, Waimanalo Formation reefs, conglomerates and beach deposits at some localities on Oahu are somewhat higher (up to 13 m above sea level) than Last-Interglacial localities on tectonically quiescent islands and coastlines, such as Bermuda or Australia (Muhs and Szabo, 1994).

2.2. Bermuda

Bermuda is situated in the Atlantic Ocean, ~1000 km to the east of North America (Fig. 3). The island has a core of pre-Quaternary volcanic rocks derived from the Mid-Atlantic Ridge, but its surficial deposits are Quaternary carbonates, mostly eolianite (Fig. 4). Isolated sublittoral marine and beach deposits and some emergent, fossil patch reefs are present also as marine facies, along with the volumetrically larger eolian facies (Vacher et al., 1989). Hermatypic corals are present around Bermuda because the island is bathed by the warm Gulf Stream current.

Detailed mapping and stratigraphic studies in the past 15 years have clarified the Quaternary stratigraphy of Bermuda (Vacher and Hearty, 1989; Vacher et al., 1989, 1995; Hearty et al., 1992; Hearty and Kindler, 1995; Vacher and Rowe, 1997). Most of the marine deposits are assigned to the Belmont Formation (~200 ka) or the

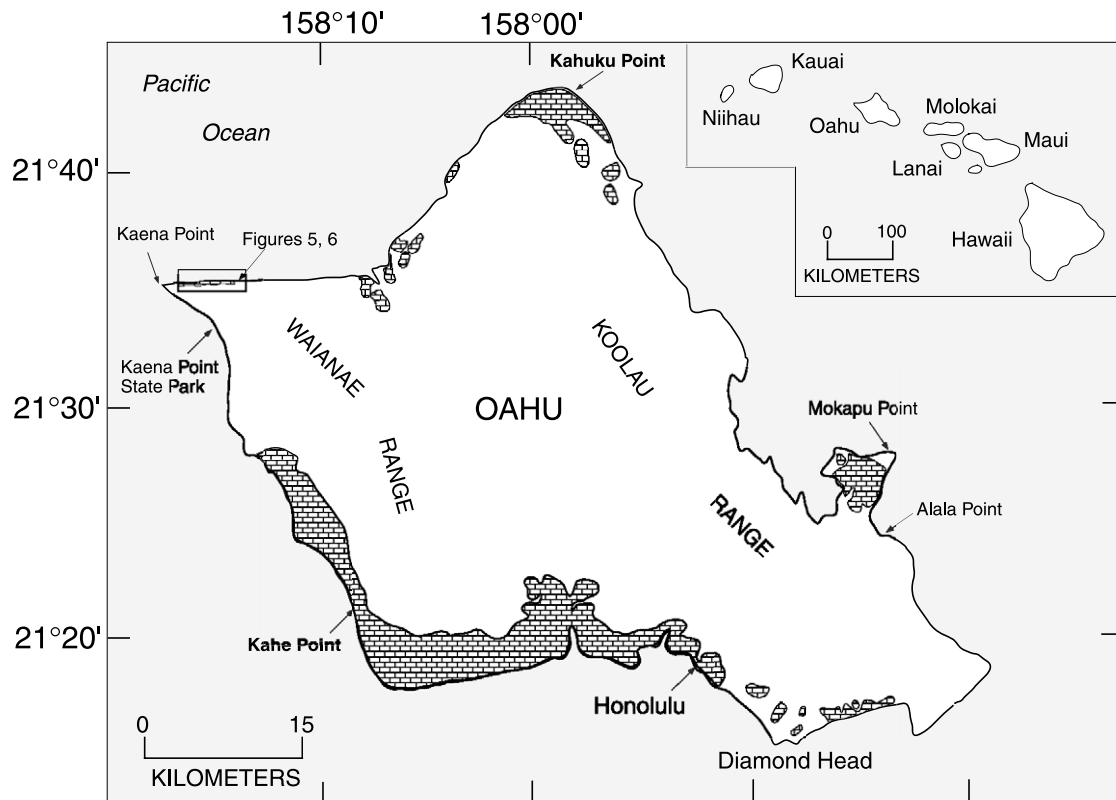


Fig. 2. Map of the island of Oahu, distribution of reef limestone (from Stearns, 1974, 1978, 1985) and fossil coral localities of the Waimanalo Formation referred to in the text. Place names in bold are confirmed ~ 120 ka, mollusk-bearing localities of the Waimanalo Formation (from this study and Szabo et al., 1994) studied by Ostergaard (1928) and Kosuge (1969).

Devonshire marine member of the Rocky Bay Formation (~ 125 ka) (Vacher et al., 1989). Age estimates of these two formations are based on alpha-spectrometric U-series dating of corals (Harmon et al., 1978, 1983). Both formations have eolianite facies that overlie the marine facies. A younger unit, the Southampton Formation, can be differentiated from the Rocky Bay Formation on the basis of stratigraphic relations (Vacher et al., 1989, 1995) and amino acid racemization data from mollusks and whole-rock bioclastic sediment (Hearty et al., 1992). In addition, Harmon et al. (1983) reported a single alpha-spectrometric U-series age of 85 ± 12 ka (2 sigma) for a coral (*Oculina* sp.) collected from the Southampton marine facies near Fort St. Catherine. Ludwig et al. (1996) also reported four TIMS U-series ages of ~ 80 ka from *Oculina* corals from Fort St. Catherine.

3. Materials and methods

On Oahu, we visited most known localities of the Waimanalo Formation, described and measured coastal sections using hand level and tape, and collected corals for U-series dating. Some of the stratigraphic data are reported in Muhs and Szabo (1994) and other data are

new. We also studied the stratigraphic relations of marine and eolian deposits at a number of localities on Bermuda and collected corals for U-series dating from the Devonshire marine member at Grape Bay and from the Southampton Formation at Fort St. Catherine. Additional corals from both the Devonshire marine member and the Belmont Formation at Grape Bay, collected by H. Lowenstam in the 1950s, were also analyzed.

Analyzed corals were cleaned mechanically, washed in distilled water and X-rayed for aragonite purity. Oahu samples HA-13C, HA-19A, HA-19B, and HA-20B were 96–97% aragonite; all others were 99–100% aragonite. For Bermuda, sample GB-3 was 95% aragonite and GB-10 was 97% aragonite; all other samples were 99–100% aragonite. Sample preparation followed methods outlined by Ludwig et al. (1992) and is summarized briefly here. Cleaned corals were dissolved in HNO_3 , spiked with ^{229}Th , ^{233}U , and ^{236}U and purified with ion exchange methods. Purified U and Th were loaded with colloidal graphite on separate Re filaments; isotopic abundances were determined by thermal ionization mass spectrometry. Ages were calculated using a half-life of 75,381 yr for ^{230}Th and 244,600 yr for ^{234}U (Tables 1 and 2). Cheng et al. (2000) re-determined the half-life of ^{230}Th to be 75,690 yr and the half-life of ^{234}U to be

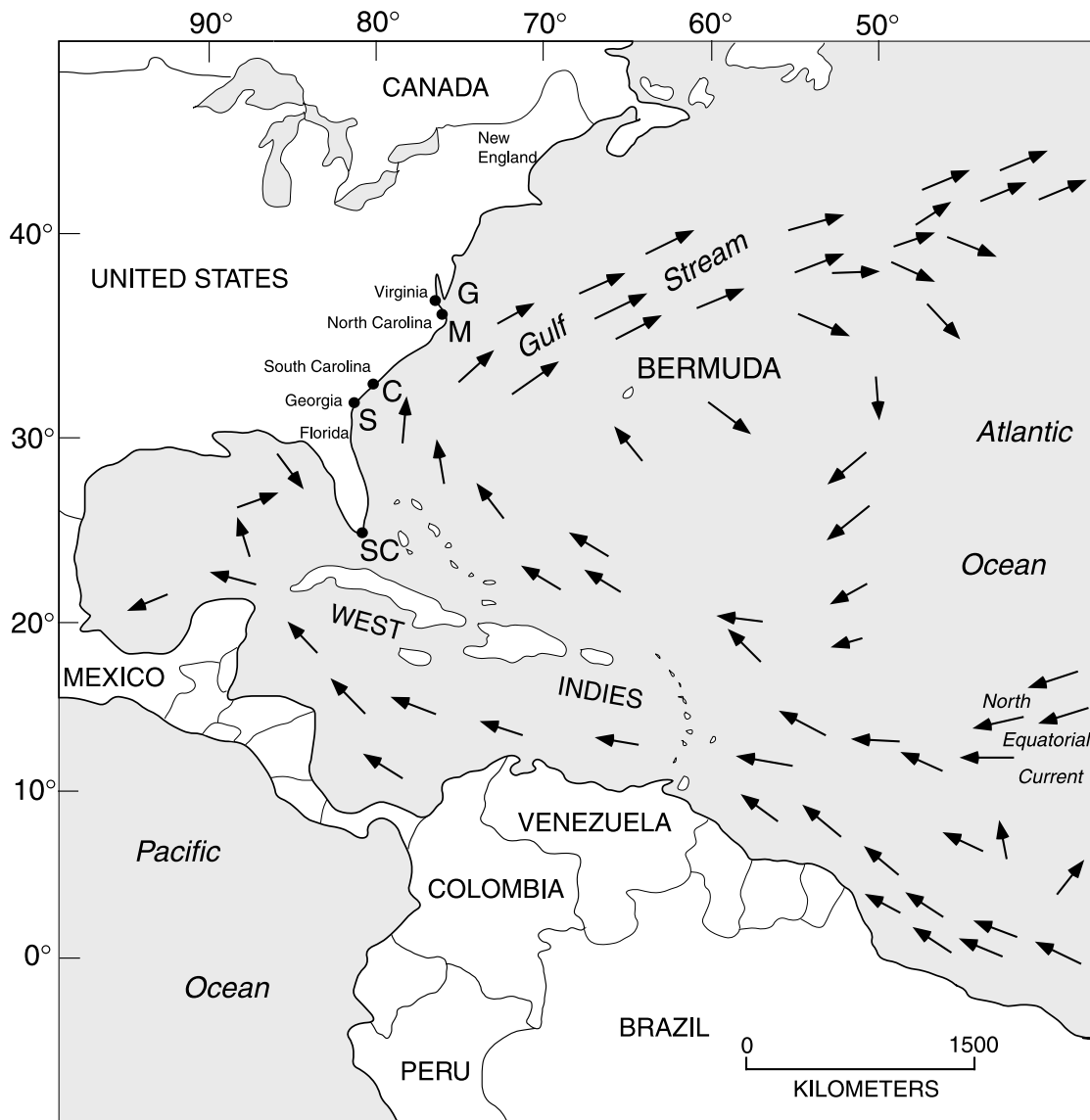


Fig. 3. Map of the western Atlantic Ocean adjacent to North and South America, location of Bermuda, major ocean currents, and ~80 ka localities referred to in the text. SC, Sand Key, Florida; S, Savannah (Skidaway Island), Georgia; C, Charleston (Mt. Pleasant), South Carolina; M, Moyock, North Carolina; G, Gomez, Virginia.

245,250 yr. Recalculation of our U-series data using these new half-lives results in ages that are in agreement, within analytical uncertainty, with ages calculated using the older half-lives (Table 3). Duplicate analyses of individual coral samples generally show excellent agreement, within limits of analytical uncertainty (Tables 1 and 2).

A consensus has developed among U–Th geochronologists that several conditions must be met in order for U-series ages of corals to be considered reliable. These include: (1) little or no evidence of recrystallization to calcite, (2) U concentrations similar to those present in living specimens of the same genus, (3) presence of little or no “inherited” ^{230}Th , as indicated by low concentrations of ^{232}Th and high $^{230}\text{Th}/^{232}\text{Th}$ values, and (4) a

calculated initial $^{234}\text{U}/^{238}\text{U}$ value in agreement, within analytical uncertainty, with that measured in modern sea water (activity ratio of about 1.148; see Chen et al., 1986). In most U-series studies of corals the usual problem encountered is a calculated initial $^{234}\text{U}/^{238}\text{U}$ value that is higher than that measured in modern sea water. Empirical observations from the Bahamas, New Guinea, Australia, and Barbados (Chen et al., 1991; Stein et al., 1993; Stirling et al., 1995, 1998; Edwards et al., 1997; Esat et al., 1999) and modeling studies (Gallup et al., 1994) have shown that higher-than-modern calculated initial $^{234}\text{U}/^{238}\text{U}$ values bias corals to older apparent ages. Esat et al. (1999) used an upper limit of initial $^{234}\text{U}/^{238}\text{U}$ activity of 1.157 and a lower limit of 1.149 as a range for corals that were considered

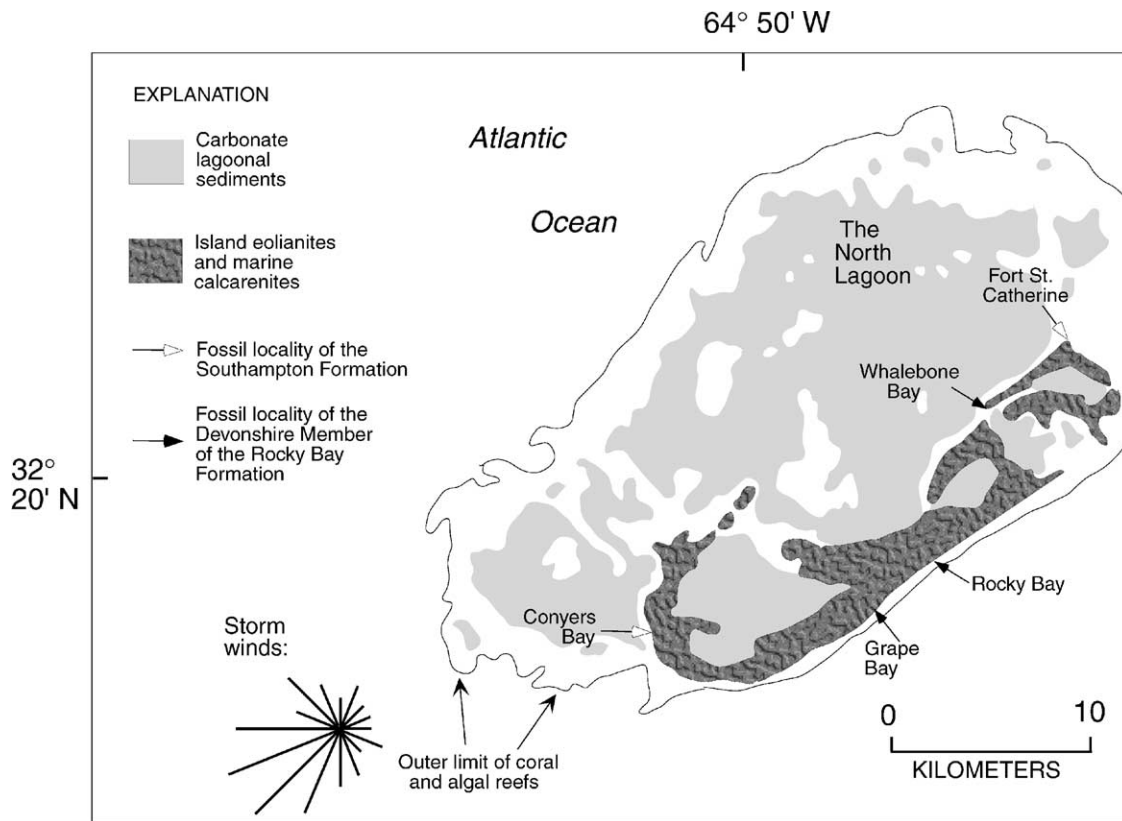


Fig. 4. Map of Bermuda showing islands, lagoonal sediments, outer limits of coral reefs, localities studied and those referred to in the text. Base map from Vacher et al. (1989).

to have experienced closed-system conditions. On the other hand, Edwards et al. (1997) showed that corals with initial $^{234}\text{U}/^{238}\text{U}$ activity values as high as 1.166 showed concordance between ^{230}Th and ^{231}Pa ages. In this study, we consider as “reliable” any corals with initial $^{234}\text{U}/^{238}\text{U}$ activity values from 1.145 to 1.159, similar to the range used by Stirling et al. (1998).

4. Uranium-series ages

4.1. Oahu

The Waimanalo Formation is well exposed at a number of localities around Oahu and corals were collected from many exposures. On the north shore of Oahu, east of Kaena Point, extensive outcrops of the Waimanalo Formation are easily mapped over more than 3 km of coastline (Figs. 5 and 6). Corals in the Waimanalo Formation are generally well preserved, show little recrystallization to calcite, and their potential for yielding reliable ages is high. On Oahu, approximately half of the analyzed corals met all closed system requirements (described above) and therefore can be considered to have reliable ages. These corals have a range of ages that spans ~133–113 ka (Table 1), which

is very similar to the range reported by Szabo et al. (1994).

Corals in growth position in the lower, reef facies of the Waimanalo Formation are found at Mokapu Point, Kahuku Point, and east of Kaena Point (Fig. 7). Two samples of growth-position *Porites* from Mokapu Point yielded relatively old apparent ages of ~143 and ~130 ka, but both had higher-than-acceptable initial $^{234}\text{U}/^{238}\text{U}$ activity values. At Kahuku Point, a previous study showed that growth-position *Porites* samples yielded acceptable initial $^{234}\text{U}/^{238}\text{U}$ activity values (Szabo et al., 1994), and ages that ranged from 114.1 ± 1.8 to 121.7 ± 1.5 ka. New analyses of Kahuku Point corals agree with the earlier data: two samples of growth position *Porites* yielded ages of 113 ± 0.7 and 121 ± 1.3 ka and acceptable initial $^{234}\text{U}/^{238}\text{U}$ activity values.

Some of the best exposures of the Waimanalo Formation are present along the north shore of Oahu, east of Kaena Point (Fig. 5). This area was mapped in detail by Muhs and Szabo (1994) and is shown in Fig. 6. Three new analyses of growth-position *Porites* from these exposures gave ages ranging from ~126 to ~118 ka, all with acceptable initial $^{234}\text{U}/^{238}\text{U}$ activity values. This range of values is somewhat older than the ~110–121 ka range reported by Szabo et al. (1994), but

Table 1

U and Th concentrations, isotopic activity ratios, and ages of corals from Oahu, Hawaii. “Reliable” ages and their initial U isotopic ratios are given in bold type

Sample No.	Genus	Locality	U (ppm)	²³² Th (ppm)	²³⁴ U/ ²³⁸ U	Error	²³⁰ Th/ ²³⁸ U	Error	²³⁰ Th/ ²³² Th	²³⁰ Th/ ²³⁸ U Age (ka)	Error	²³⁴ U/ ²³⁸ U init	Error
HA-7-X	<i>Pocillopora</i>	Kahe Point	2.41	0.0101	1.1154	0.0015	0.7508	0.0022	542	118.2	0.7	1.1613	0.0019
HA-7-X dup	<i>Pocillopora</i>	Kahe Point	2.40	0.0093	1.1163	0.0025	0.7474	0.0025	584	117.0	0.9	1.1620	0.0032
HA-7-Y	<i>Porites</i>	Kahe Point	2.71	0.0038	1.1052	0.0019	0.7478	0.0046	1621	119.5	1.4	1.1477	0.0025
HA-7-Y dup	<i>Porites</i>	Kahe Point	2.68	0.0038	1.1063	0.0032	0.7489	0.0028	1595	119.6	1.1	1.1492	0.0042
HA-7-Y dup #2	<i>Porites</i>	Kahe Point	2.69	0.0034	1.1083	0.0018	0.7497	0.0015	1797	119.4	0.6	1.1519	0.0023
HA-3-A	<i>Porites</i>	Alala Point	2.69	0.0002	1.1077	0.0013	0.7491	0.0106	33430	119.4	3.0	1.1511	0.0022
HA-3-A dup	<i>Porites</i>	Alala Point	2.73	0.0001	1.1092	0.0018	0.7578	0.0015	83510	121.5	0.6	1.1541	0.0023
HA-3-B	<i>Porites</i>	Alala Point	2.76	0.0006	1.1085	0.0016	0.7952	0.0021	11224	133.0	0.8	1.1582	0.0021
HA-3-B dup	<i>Porites</i>	Alala Point	2.79	0.0003	1.1084	0.0043	0.7997	0.0026	20369	134.5	1.4	1.1587	0.0058
HA-3-C	<i>Porites</i>	Alala Point	2.79	0.0006	1.1104	0.0014	0.7925	0.0020	11606	131.7	0.7	1.1603	0.0019
HA-3-D	<i>Porites</i>	Alala Point	2.64	0.0002	1.1109	0.0020	0.7871	0.0023	37321	129.9	0.9	1.1603	0.0027
HA-9-A	<i>Pavona</i>	Kaena Pt. State Park	2.86	0.0006	1.1063	0.0021	0.7757	0.0018	11897	127.5	0.8	1.1525	0.0028
HA-9-B	<i>Pocillopora</i>	Kaena Pt. State Park	2.71	0.0001	1.1089	0.0018	0.7372	0.0019	48562	115.8	0.6	1.1512	0.0023
HA-9-C	<i>Pocillopora</i>	Kaena Pt. State Park	2.49	0.0002	1.1114	0.0013	0.7468	0.0029	33663	117.9	0.8	1.1556	0.0018
HA-9-E	<i>Pocillopora</i>	Kaena Pt. State Park	2.47	0.0002	1.1178	0.0060	0.7276	0.0017	27087	111.5	1.2	1.1615	0.0078
HA-9-E dup	<i>Pocillopora</i>	Kaena Pt. State Park	2.47	0.0001	1.1191	0.0017	0.7439	0.0034	41277	115.5	1.0	1.1652	0.0022
HA-9-F	<i>Pocillopora</i>	Kaena Pt. State Park	2.36	0.0002	1.1138	0.0026	0.7517	0.0018	24238	118.8	0.7	1.1593	0.0033
HA-9-F dup	<i>Pocillopora</i>	Kaena Pt. State Park	2.34	0.0003	1.1126	0.0019	0.7601	0.0026	19466	121.4	0.9	1.1588	0.0025
HA-9-G	<i>Porites</i>	Kaena Pt. State Park	2.58	0.0002	1.1063	0.0022	0.7457	0.0027	29897	118.7	0.9	1.1487	0.0029
HA-9-H	<i>Porites</i>	Kaena Pt. State Park	2.86	0.0004	1.1037	0.0020	0.7461	0.0031	16043	119.4	1.0	1.1455	0.0026
HA-9-H dup	<i>Porites</i>	Kaena Pt. State Park	2.69	0.0001	1.1084	0.0014	0.7466	0.0018	49215	118.5	0.6	1.1517	0.0019
HA-13-B	<i>Porites</i> , in situ	Kahuku Point	2.71	0.0001	1.1121	0.0021	0.7292	0.0020	40543	113.0	0.7	1.1544	0.0027
HA-13-C	<i>Porites</i> , in situ	Kahuku Point	2.72	0.0020	1.1105	0.0041	0.7571	0.0033	3048	121.0	1.3	1.1557	0.0054
HA-15-A	<i>Pocillopora</i>	N. shore, E. of Kaena Pt.	2.38	0.0002	1.1143	0.0028	0.7539	0.0018	31459	119.3	0.8	1.1603	0.0036
HA-15-B	<i>Pocillopora</i>	N. shore, E. of Kaena Pt.	2.22	0.0001	1.1348	0.0025	0.7853	0.0023	59947	123.6	0.9	1.1914	0.0033
HA-15-C	<i>Pocillopora</i>	N. shore, E. of Kaena Pt.	2.38	0.0001	1.1608	0.0017	0.8992	0.0026	47952	152.8	1.1	1.2480	0.0024
HA-15-D	<i>Pocillopora</i>	N. shore, E. of Kaena Pt.	2.18	0.0002	1.1218	0.0018	0.7549	0.0022	28759	118.0	0.7	1.1702	0.0023
HA-15-(1)	<i>Porites</i>	N. shore, E. of Kaena Pt.	2.94	0.0002	1.1055	0.0014	0.7707	0.0021	42531	126.2	0.7	1.1508	0.0019
HA-15-(2)	<i>Pocillopora</i>	N. shore, E. of Kaena Pt.	2.41	0.0001	1.1200	0.0020	0.7538	0.0019	41802	118.0	0.7	1.1677	0.0026
HA-15-3	<i>Pocillopora</i>	N. shore, E. of Kaena Pt.	2.29	0.0002	1.1143	0.0020	0.7392	0.0032	30636	115.2	0.9	1.1585	0.0026
HA-15-(4)	<i>Porites</i>	N. shore, E. of Kaena Pt.	3.01	0.0002	1.1032	0.0019	0.7626	0.0039	40686	124.3	1.2	1.1468	0.0025
HA-15-(4) dup	<i>Porites</i>	N. shore, E. of Kaena Pt.	2.75	0.0004	1.1030	0.0030	0.7550	0.0023	16812	122.1	0.9	1.1455	0.0039
HA-20-B	<i>Porites</i> , in situ	N. shore, E. of Kaena Pt.	2.58	0.0002	1.1088	0.0017	0.7448	0.0018	32624	117.9	0.6	1.1520	0.0022
HA-20-(2)	<i>Porites</i> , in situ	N. shore, E. of Kaena Pt.	2.72	0.0004	1.1067	0.0019	0.7483	0.0024	15907	119.3	0.8	1.1496	0.0025
HA-21-(1)	<i>Porites</i> , in situ	N. shore, E. of Kaena Pt.	2.67	0.0010	1.1066	0.0015	0.7717	0.0021	6451	126.2	0.7	1.1524	0.0021
HA-19-A	<i>Porites</i>	Mokapu Point	2.68	0.0004	1.1119	0.0012	0.7886	0.0019	15320	130.1	0.7	1.1618	0.0016
HA-19B-(1/2)	<i>Porites</i>	Mokapu Point	2.48	0.0034	1.1104	0.0018	0.8273	0.0028	1809	143.2	1.1	1.1656	0.0025

Table 2

U and Th concentrations, isotopic activity ratios, and ages of corals from Bermuda. “Reliable” ages and their initial U isotopic ratios are given in bold type

Sample No.	Locality	Species	U (ppm)	²³² Th (ppm)	²³⁴ U/ ²³⁸ U	Error	²³⁰ Th/ ²³⁸ U	Error	²³⁰ Th/ ²³² Th	²³⁰ Th/ ²³⁸ U age (ka)	Error	²³⁴ U/ ²³⁸ U init	Error
<i>Southampton formation</i>													
FSC-OC-1-A	Fort St. Catherine	<i>Oculina</i> sp.	2.44	0.0061	1.1137	0.0017	0.5939	0.0023	719	81.5	0.5	1.1432	0.0020
FSC-OC-1-B	Fort St. Catherine	<i>Oculina</i> sp.	2.44	0.0088	1.1149	0.0018	0.5865	0.0020	493	79.9	0.4	1.1441	0.0022
FSC-OC-1-C	Fort St. Catherine	<i>Oculina</i> sp.	2.71	0.0065	1.1200	0.0024	0.5911	0.0019	751	80.2	0.5	1.1506	0.0028
FSC-OC-1-D	Fort St. Catherine	<i>Oculina</i> sp.	2.27	0.0061	1.1153	0.0029	0.6034	0.0031	676	83.3	0.7	1.1459	0.0035
FSC-OC-1-E	Fort St. Catherine	<i>Oculina</i> sp.	2.33	0.0035	1.1226	0.0020	0.5927	0.0024	1,205	80.2	0.5	1.1539	0.0024
FSC-OC-1-F	Fort St. Catherine	<i>Oculina</i> sp.	3.67	0.0055	1.1144	0.0017	0.5880	0.0018	1,184	80.3	0.4	1.1436	0.0020
FSC-OC-1-G	Fort St. Catherine	<i>Oculina</i> sp.	2.93	0.0050	1.1171	0.0018	0.5834	0.0020	1,034	79.0	0.4	1.1465	0.0022
FSC-OC-1G dup	Fort St. Catherine	<i>Oculina</i> sp.	2.92	0.0059	1.1154	0.0017	0.5841	0.0020	884	79.4	0.4	1.1445	0.0020
FSC-OC-1-H	Fort St. Catherine	<i>Oculina</i> sp.	2.31	0.0040	1.1139	0.0017	0.6065	0.0033	1,050	84.0	0.7	1.1445	0.0021
FSC-OC-1-I	Fort St. Catherine	<i>Oculina</i> sp.	3.06	0.0020	1.1170	0.0017	0.5851	0.0014	2,779	79.4	0.3	1.1466	0.0020
FSC-OC-1-J	Fort St. Catherine	<i>Oculina</i> sp.	2.23	0.0013	1.1164	0.0016	0.5827	0.0019	2,974	79.0	0.4	1.1456	0.0019
FSC-OC-1-K	Fort St. Catherine	<i>Oculina</i> sp.	2.52	0.0023	1.1165	0.0017	0.5801	0.0018	1,949	78.4	0.4	1.1456	0.0020
FSC-OC-1-L	Fort St. Catherine	<i>Oculina</i> sp.	2.37	0.0021	1.1130	0.0014	0.5821	0.0019	2,000	79.2	0.4	1.1414	0.0018
FSC-OC-1-M	Fort St. Catherine	<i>Oculina</i> sp.	2.58	0.0017	1.1161	0.0017	0.5789	0.0017	2,629	78.3	0.4	1.1450	0.0020
FSC-OC-1-N	Fort St. Catherine	<i>Oculina</i> sp.	2.54	0.0019	1.1177	0.0021	0.5925	0.0040	2,431	80.8	0.8	1.1480	0.0026
FSC-OC-1-O	Fort St. Catherine	<i>Oculina</i> sp.	2.76	0.0023	1.1078	0.0017	0.5791	0.0014	2,108	79.2	0.3	1.1350	0.0020
FSC-OC-1-P	Fort St. Catherine	<i>Oculina</i> sp.	2.77	0.0029	1.1139	0.0017	0.5929	0.0012	1,721	81.3	0.3	1.1434	0.0020
FSC-OC-1-Q	Fort St. Catherine	<i>Oculina</i> sp.	2.95	0.0032	1.1146	0.0018	0.5874	0.0016	1,639	80.1	0.4	1.1438	0.0022
<i>Rocky bay formation</i>													
GB-2A	Grape Bay	<i>Siderastrea</i> sp.	3.08	0.0172	1.1101	0.0018	0.7572	0.0022	412	121.1	0.7	1.1553	0.0024
GB-3	Grape Bay	<i>Oculina</i> sp.	1.44	0.0015	1.1248	0.0021	0.7683	0.0063	2,300	121.0	1.8	1.1758	0.0029
GB-4	Grape Bay	<i>Oculina</i> sp.	1.54	0.0029	1.1743	0.0022	0.7103	0.0058	1,143	97.9	1.3	1.2300	0.0029
GB-6	Grape Bay	<i>Colpophyllia</i> sp.	2.82	0.0012	1.1073	0.0017	0.7389	0.0030	5,067	116.6	0.9	1.1493	0.0022
GB-7	Grape Bay	<i>Montastrea cavernosa</i>	2.75	0.0009	1.1077	0.0021	0.7274	0.0030	6,834	113.4	0.9	1.1485	0.0027
GB-9	Grape Bay	<i>Siderastrea siderea</i>	3.02	0.0009	1.1103	0.0017	0.7477	0.0029	7,982	118.4	0.9	1.1543	0.0022
GB-10	Grape Bay	<i>Colpophyllia natans</i>	2.81	0.0020	1.1073	0.0017	0.7471	0.0019	3,152	118.9	0.6	1.1503	0.0022
GB-11	Grape Bay	<i>Montastrea cavernosa</i>	2.75	0.0013	1.1093	0.0013	0.7715	0.0026	4,851	125.5	0.8	1.1560	0.0018
<i>Belmont formation</i>													
GB-12	Grape Bay	<i>Oculina</i> sp.	2.44	0.0003	1.0889	0.0020	0.9314	0.0027	26,892	198.8	2.0	1.1562	0.0030
GB-13	Grape Bay	<i>Oculina</i> sp.	2.54	0.0002	1.0842	0.0020	0.9299	0.0027	30,458	200.8	2.0	1.1487	0.0030

Table 3

Comparison of U-series ages and U activity ratios using old and new half-lives. “Reliable” ages and their initial U isotopic ratios are given in bold type

Sample#	$^{230}\text{Th}/^{238}\text{U}$ age (ka) (calculated using old half-lives)	Error	$^{234}\text{U}/^{238}\text{U}$ init	Error	$^{230}\text{Th}/^{238}\text{U}$ age (ka) (calculated using new half-lives)	Error	$^{234}\text{U}/^{238}\text{U}$ init	Error
<i>Waimanalo formation, Oahu, Hawaii</i>								
HA-7-X	118.2	0.7	1.1613	0.0019	117.8	0.7	1.1610	0.0019
HA-7-X dup	117.0	0.9	1.1620	0.0032	116.7	0.9	1.1617	0.0032
HA-7-Y	119.5	1.4	1.1477	0.0025	119.1	1.4	1.1474	0.0025
HA-7-Y dup	119.6	1.1	1.1492	0.0042	119.2	1.1	1.1489	0.0042
HA-7-Y dup #2	119.4	0.6	1.1519	0.0023	119.0	0.6	1.1516	0.0023
HA-3-A	119.4	3.0	1.1511	0.0022	119.0	3.0	1.1508	0.0022
HA-3-A dup	121.5	0.6	1.1541	0.0023	121.1	0.6	1.1538	0.0023
HA-3-B	133.0	0.8	1.1582	0.0021	132.5	0.8	1.1579	0.0021
HA-3-B dup	134.5	1.4	1.1587	0.0058	134.0	1.4	1.1583	0.0058
HA-3-C	131.7	0.7	1.1603	0.0019	131.2	0.7	1.1599	0.0019
HA-3-D	129.9	0.9	1.1603	0.0027	129.4	0.9	1.1599	0.0027
HA-9-A	127.5	0.8	1.1525	0.0028	127.1	0.7	1.1522	0.0028
HA-9-B	115.8	0.6	1.1512	0.0023	115.4	0.6	1.1509	0.0023
HA-9-C	117.9	0.8	1.1556	0.0018	117.5	0.8	1.1553	0.0018
HA-9-E	111.5	1.2	1.1615	0.0078	111.1	1.2	1.1612	0.0078
HA-9-E dup	115.5	1.0	1.1652	0.0022	115.2	1.0	1.1649	0.0022
HA-9-F	118.8	0.7	1.1593	0.0033	118.4	0.7	1.1590	0.0033
HA-9-F dup	121.4	0.9	1.1588	0.0025	121.0	0.9	1.1585	0.0025
HA-9-G	118.7	0.9	1.1487	0.0029	118.3	0.9	1.1484	0.0029
HA-9-H	119.4	1.0	1.1455	0.0026	119.0	1.0	1.1452	0.0026
HA-9-H dup	118.5	0.6	1.1517	0.0019	118.1	0.6	1.1514	0.0019
HA-13-B	113.0	0.7	1.1544	0.0027	112.7	0.7	1.1541	0.0027
HA-13-C	121.0	1.3	1.1557	0.0054	120.6	1.3	1.1554	0.0054
HA-15-A	119.3	0.8	1.1603	0.0036	118.9	0.8	1.1600	0.0036
HA-15-B	123.6	0.9	1.1914	0.0033	123.2	0.9	1.1910	0.0033
HA-15-C	152.8	1.1	1.248	0.0024	152.1	1.1	1.2472	0.0024
HA-15-D	118.0	0.7	1.1702	0.0023	117.6	0.7	1.1699	0.0023
HA-15-(1)	126.2	0.7	1.1508	0.0019	125.7	0.7	1.1505	0.0019
HA-15-(2)	118.0	0.7	1.1677	0.0026	117.7	0.7	1.1673	0.0026
HA-15-3	115.2	0.9	1.1585	0.0026	114.9	0.9	1.1582	0.0026
HA-15-(4)	124.3	1.2	1.1468	0.0025	123.9	1.2	1.1465	0.0025
HA-15-(4) dup	122.1	0.9	1.1455	0.0039	121.7	0.9	1.1452	0.0039
HA-20-B	117.9	0.6	1.152	0.0022	117.5	0.6	1.1517	0.0022
HA-20-(2)	119.3	0.8	1.1496	0.0025	119.0	0.8	1.1493	0.0025
HA-21-(1)	126.2	0.7	1.1524	0.0021	125.8	0.7	1.1521	0.0021
HA-19-A	130.1	0.7	1.1618	0.0016	129.6	0.7	1.1614	0.0016
HA-19B-(1/2)	143.2	1.1	1.1656	0.0025	142.6	1.1	1.1652	0.0025
<i>Southampton formation, Bermuda</i>								
FSC-OC-1-A	81.5	0.5	1.1432	0.0020	81.4	0.5	1.1430	0.0020
FSC-OC-1-B	79.9	0.4	1.1441	0.0022	79.7	0.4	1.1440	0.0022
FSC-OC-1-C	80.2	0.5	1.1506	0.0028	80.1	0.5	1.1505	0.0028
FSC-OC-1-D	83.3	0.7	1.1459	0.0035	83.1	0.7	1.1458	0.0035
FSC-OC-1-E	80.2	0.5	1.1539	0.0024	80.1	0.5	1.1538	0.0024
FSC-OC-1-F	80.3	0.4	1.1436	0.0020	80.1	0.4	1.1435	0.0020
FSC-OC-1-G	79.0	0.4	1.1465	0.0022	78.9	0.4	1.1463	0.0022
FSC-OC-1G dup	79.4	0.4	1.1445	0.0020	79.2	0.4	1.1443	0.0020
FSC-OC-1-H	84.0	0.7	1.1445	0.0021	83.9	0.7	1.1444	0.0021
FSC-OC-1-I	79.4	0.3	1.1466	0.0020	79.2	0.3	1.1464	0.0020
FSC-OC-1-J	79.0	0.4	1.1456	0.0019	78.8	0.4	1.1454	0.0019
FSC-OC-1-K	78.4	0.4	1.1456	0.0020	78.3	0.4	1.1454	0.0020
FSC-OC-1-L	79.2	0.4	1.1414	0.0018	79.1	0.4	1.1413	0.0018
FSC-OC-1-M	78.3	0.4	1.1450	0.0020	78.1	0.4	1.1448	0.0020
FSC-OC-1-N	80.8	0.8	1.1480	0.0026	80.6	0.8	1.1478	0.0026
FSC-OC-1-O	79.2	0.3	1.1350	0.0020	79.1	0.3	1.1348	0.0020
FSC-OC-1-P	81.3	0.3	1.1434	0.0020	81.1	0.3	1.1433	0.0020
FSC-OC-1-Q	80.1	0.4	1.1438	0.0022	80.0	0.4	1.1436	0.0022
<i>Rocky bay formation, Bermuda</i>								
GB-2A	121.1	0.7	1.1553	0.0024	—	—	—	—
GB-3	121.0	1.8	1.1758	0.0029	120.7	1.8	1.1755	0.0029

Table 3 (continued)

Sample#	$^{230}\text{Th}/^{238}\text{U}$ age (ka) (calculated using old half-lives)	Error	$^{234}\text{U}/^{238}\text{U}$ init	Error	$^{230}\text{Th}/^{238}\text{U}$ age (ka) (calculated using new half-lives)	Error	$^{234}\text{U}/^{238}\text{U}$ init	Error
GB-4	97.9	1.3	1.2300	0.0029	97.7	1.3	1.2297	0.0029
GB-6	116.6	0.9	1.1493	0.0022	116.2	0.9	1.1490	0.0022
GB-7	113.4	0.9	1.1485	0.0027	113.0	0.9	1.1482	0.0027
GB-9	118.4	0.9	1.1543	0.0022	118.0	0.9	1.1540	0.0022
GB-10	118.9	0.6	1.1503	0.0022	118.5	0.6	1.1500	0.0022
GB-11	125.5	0.8	1.1560	0.0018	125.1	0.8	1.1556	0.0018
<i>Belmont formation, Bermuda</i>								
GB-12	198.8	2.0	1.1562	0.0030	197.3	2.0	1.1553	0.0030
GB-13	200.8	2.0	1.1487	0.0030	199.3	2.0	1.1478	0.0030

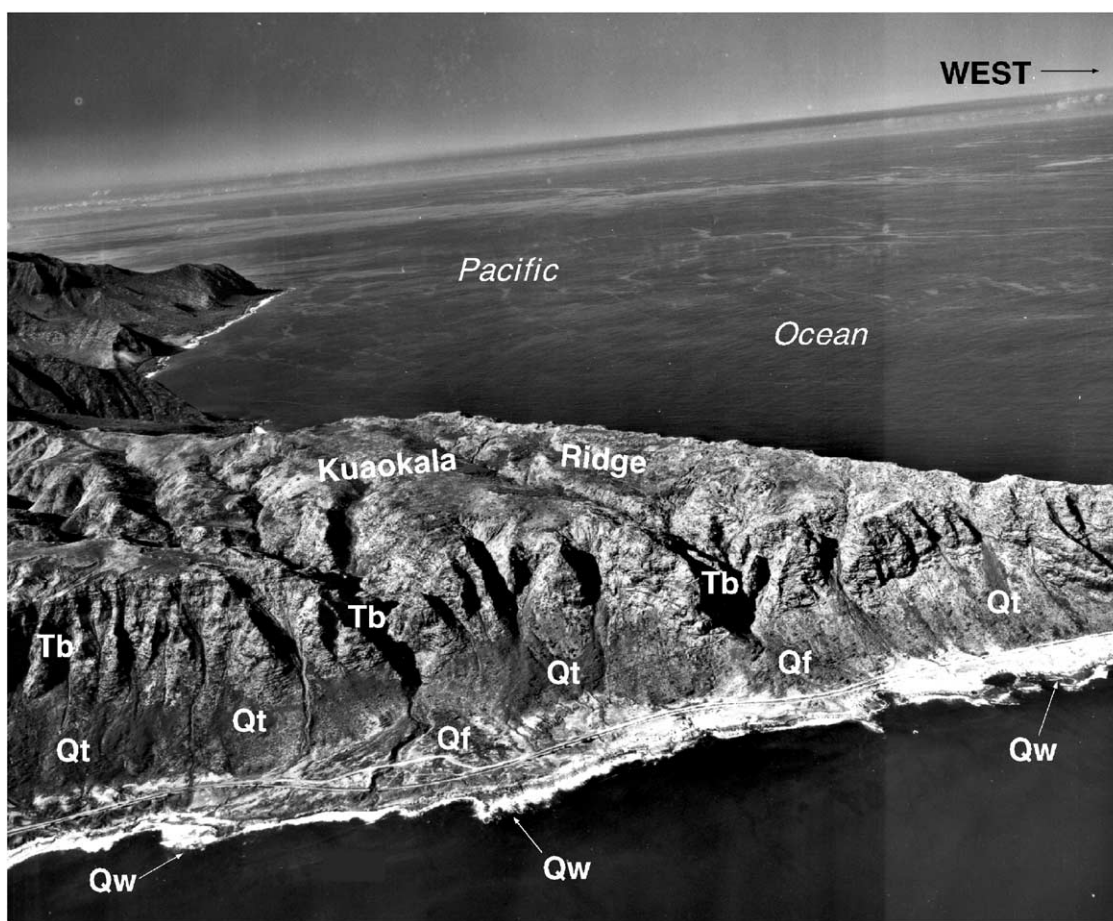


Fig. 5. Oblique aerial photograph of the Kuaokala Ridge-Kaena Point area, looking south, showing the Waimanalo Formation (Qw), overlying talus deposits (Qt) and alluvial fan deposits (Qf) and Tertiary basalt (Tb).

their analyses had larger (up to 3.8 ka) analytical uncertainties. The latter workers also did not analyze corals from the conglomerate facies of the Waimanalo Formation east of Kaena Point. Although we analyzed numerous specimens from the conglomerate facies, only one *Pocillopora* specimen and two *Porites* yielded acceptable initial $^{234}\text{U}/^{238}\text{U}$ activity values. The *Pocillopora* clast gave an age of ~ 115 ka, and the *Porites* clasts

gave ages of ~ 122 and ~ 126 ka. Corals are better preserved in the conglomerate facies of the Waimanalo Formation southeast of Kaena Point, near Kaena Point State Park. Six specimens from the latter locality gave ages with acceptable initial $^{234}\text{U}/^{238}\text{U}$ activity values. Two *Porites* clasts both gave ages of ~ 118 – 119 ka and four *Pocillopora* clasts gave ages ranging from ~ 116 to ~ 127 ka.

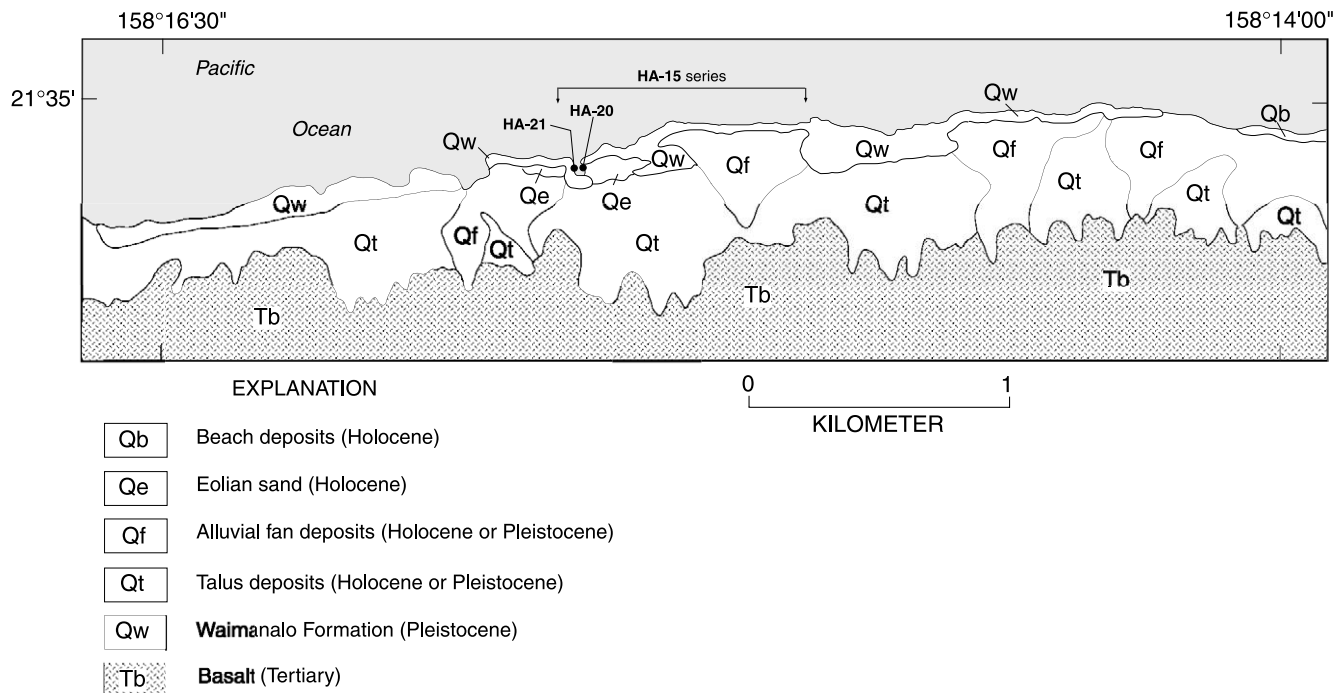


Fig. 6. Surficial geologic map of a part of the north shore of Oahu illustrating the distribution of the Waimanalo Formation shown in Fig. 5 and adjacent areas and fossil coral localities. Redrawn from Muhs and Szabo (1994).

Szabo et al. (1994) reported some of the oldest Waimanalo Formation corals at a locality near Alala Point. New analyses agree with the earlier results. Two *Porites* from the reef limestone at Alala Point yielded ages of ~121 and ~133 ka. A third sample gave an age of ~132 ka, with a slightly higher-than-acceptable initial $^{234}\text{U}/^{238}\text{U}$ activity value of 1.160. This age may be ~2 ka too old if the model of Gallup et al. (1994) is correct.

U-series ages of corals from Kahe Point were reported by Easton and Ku (1981), Muhs and Szabo (1994) and Szabo et al. (1994). Kahe Point is important because it has one of the richest Last-Interglacial molluscan faunas on Oahu (Kosuge, 1969). The locality is also of interest because older ages (> 130 ka), done by alpha spectrometry, were reported by Easton and Ku (1981) and Muhs and Szabo (1994). A *Pocillopora* from this locality (also analyzed by Muhs and Szabo, 1994) gave an unacceptably high initial $^{234}\text{U}/^{238}\text{U}$ activity value, but duplicate analyses of a *Porites* sample analyzed by Muhs and Szabo (1994) gave an age of ~119 ka with an acceptable initial $^{234}\text{U}/^{238}\text{U}$ activity value. This latter age obviously fails to confirm the possible older ages suggested by the lower-precision alpha-spectrometric ages.

4.2. Bermuda

Corals from Bermuda generally also show excellent preservation, and the majority of studied specimens,

even those from the penultimate interglacial Belmont Formation, met all requirements for closed-system conditions (Table 2). Two *Oculina* specimens from the Belmont Formation at Grape Bay (Fig. 8) yielded ages of ~199 and ~201 ka, both with acceptable initial $^{234}\text{U}/^{238}\text{U}$ activity values. These ages agree well with alpha-spectrometric U-series ages of ~200 ka (± 40 ka at 2-sigma) for the Belmont Formation at other localities (Harmon et al., 1983). The new ages for the Belmont Formation also agree well with ages of *Acropora palmata* cobbles from a 91-m-high beach deposit in the Clermont Nose area of Barbados (Gallup et al., 1994; Edwards et al., 1997). Their reported ages of ~194, 199, and 201 ka are not significantly different from the new ages (Table 2) for the Belmont Formation.

At Grape Bay, the Devonshire marine member overlies the Belmont Formation and marks the Last Interglacial high stand of sea. The contact between the two units is well illustrated in Figs. 5–12 of Meischner et al. (1995). Soil pipes (solution channels filled with red clay) are found in the Belmont Formation at this locality (Herwitz and Muhs, 1995). The stratigraphic significance of the soil pipes is that, although truncated, they mark a period of subaerial exposure that occurred after regression of the high stand of sea represented by the Belmont Formation and before deposition of the Devonshire marine member.

Most previously published U-series analyses of Bermuda corals are from either the Devonshire marine member of the Rocky Bay Formation or what

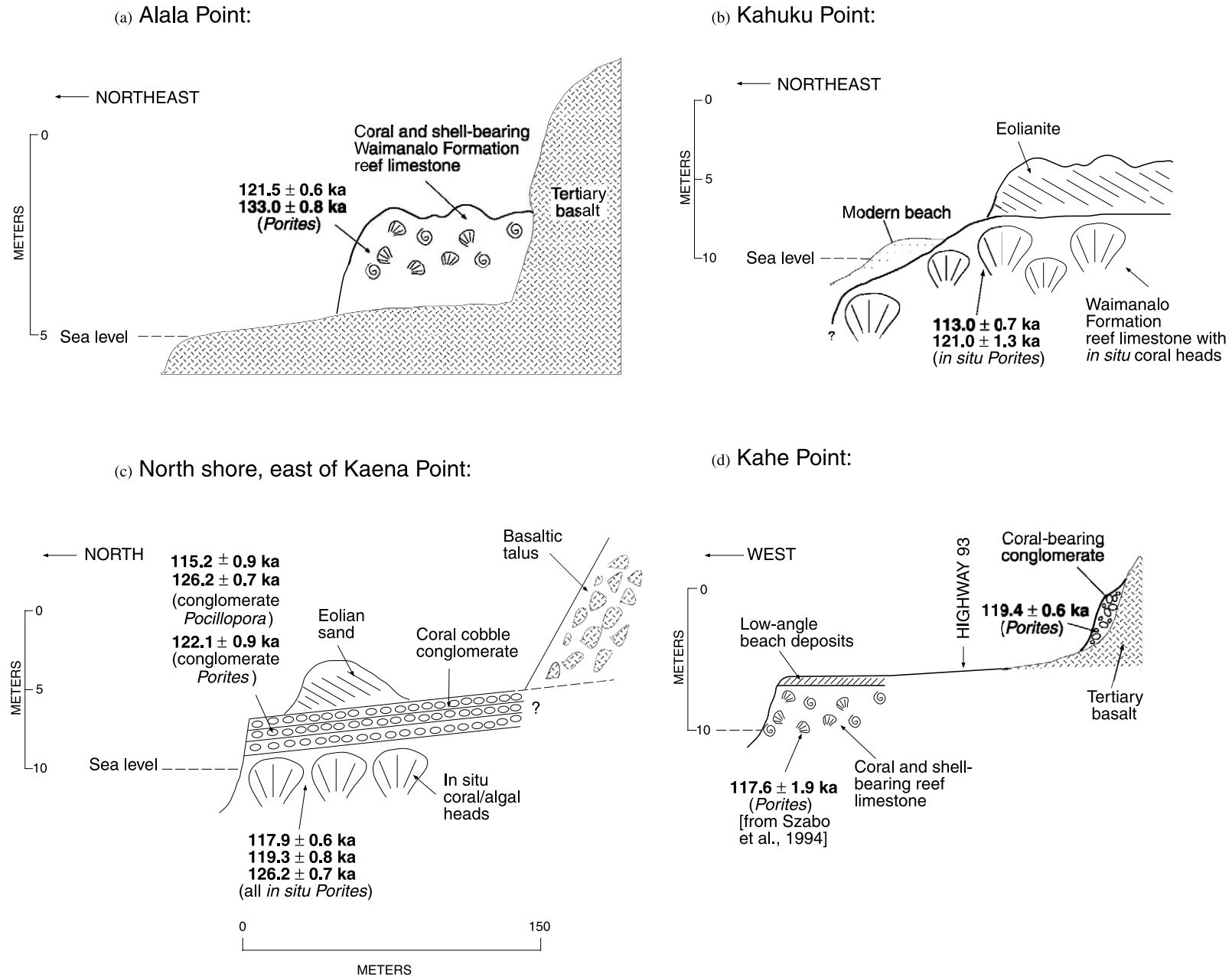


Fig. 7. Cross-sections of outcrops of the Waimanalo Formation exposed on (a) Alala Point, (b) Kahuku Point, (c) the north shore of Oahu east of Kaena Point and (d) Kahe Point (see Figs. 2 and 6 for locations), along with reliable U-series ages of corals (Table 1). Stratigraphy at Kaena Point and Kahe Point redrawn from Muhs and Szabo (1994).

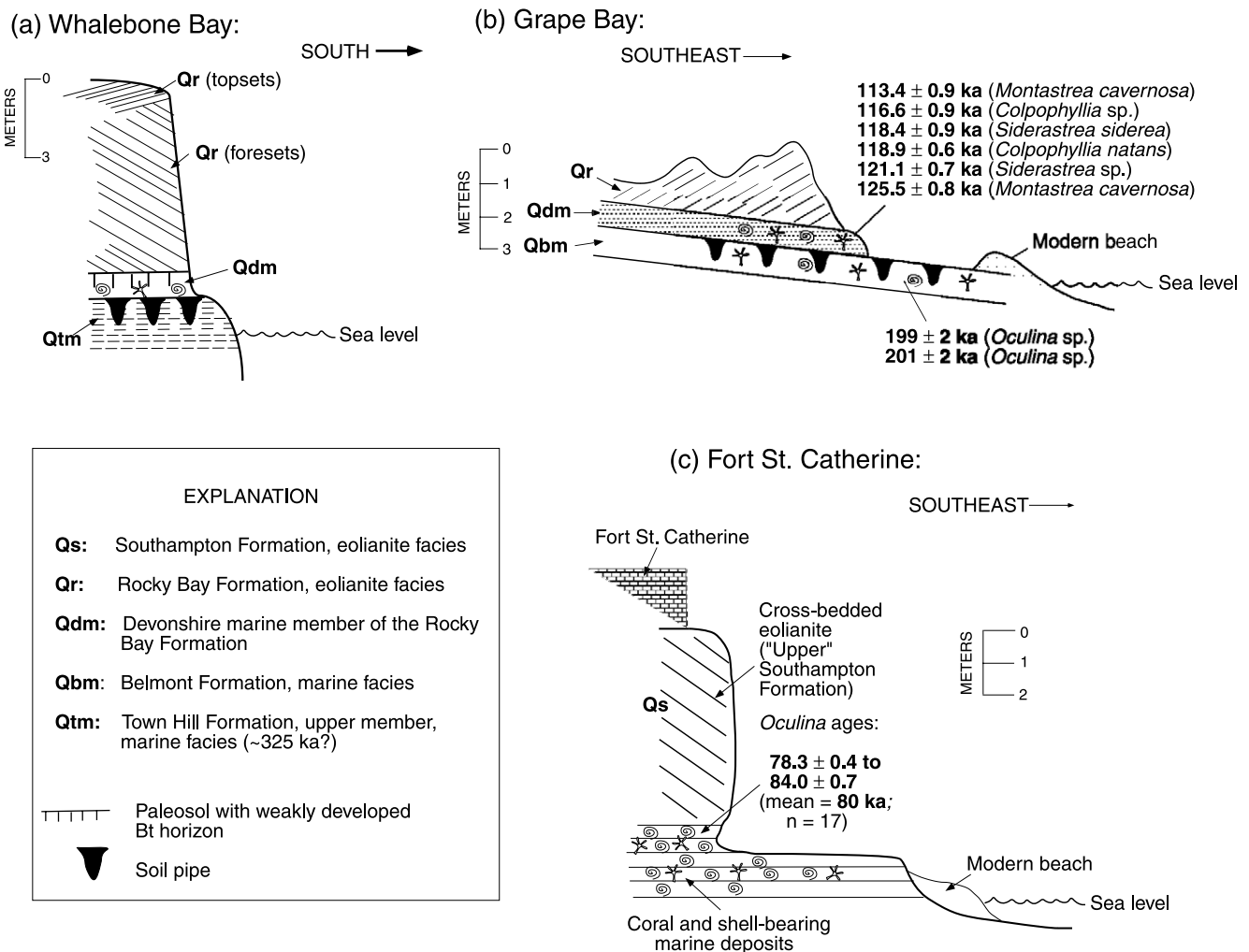


Fig. 8. Stratigraphy of sections at (a) Whalebone Bay, (b) Grape Bay, and (c) Fort St. Catherine, Bermuda and U-series ages (Table 2). See Fig. 4 for localities. Stratigraphy at Whalebone Bay and Grape Bay from Herwitz and Muhs (1995); stratigraphy at Fort St. Catherine from this study.

was previously referred to as the “Spencer’s Point Formation” of Harmon et al. (1978, 1983). Previously reported ages of corals from the Devonshire member, all done by alpha spectrometry, range from 118 ± 12 to 134 ± 8 ka (2 sigma errors). Corals from the “Spencer’s Point Formation” range from 97 ± 12 to 108 ± 16 ka. Subsequent detailed mapping and stratigraphic studies suggest that the “Spencer’s Point Formation” is equivalent to the Devonshire member of the Rocky Bay Formation (Vacher and Hearty, 1987; Vacher et al., 1989, 1995; Hearty et al., 1992; Hearty and Kindler, 1995; Vacher and Rowe, 1997), and the term “Spencer’s Point Formation” is no longer used in Bermudan stratigraphy. Unfortunately, we did not find any corals at localities previously assigned to the Spencer’s Point Formation. However, our collections at Grape Bay, as well as the earlier collections of H. Lowenstam, yielded several corals that allow new age determinations for the Devonshire member.

New U-series analyses of corals from the Devonshire member at Grape Bay agree reasonably well with those of Harmon et al. (1978, 1983). Six specimens, two each of *Montastrea*, *Siderastrea*, and *Colpophyllia*, meet all our criteria for reliable ages. With one exception, these corals yielded U-series ages ranging from 116.6 ± 0.9 to 125.5 ± 0.8 ka, which overlap with previous alpha-spectrometric ages reported by Harmon et al. (1978, 1983). One sample of *Montastrea cavernosa* yielded a younger age of 113.4 ± 0.9 ka. Two *Oculina* samples had U contents of less than 2 ppm, despite having little or no evidence of recrystallization to calcite. Most colonial corals have U contents of 2–4 ppm, and analyses of two modern *Oculina* specimens from Bermuda showed U contents of 3 ppm (Ludwig et al., 1996). In addition, the two *Oculina* specimens from Grape Bay yielded higher-than-acceptable initial $^{234}\text{U}/^{238}\text{U}$ values and we consider them to have experienced open system conditions.

The youngest Pleistocene deposit on Bermuda with both an eolian and marine facies is the Southampton

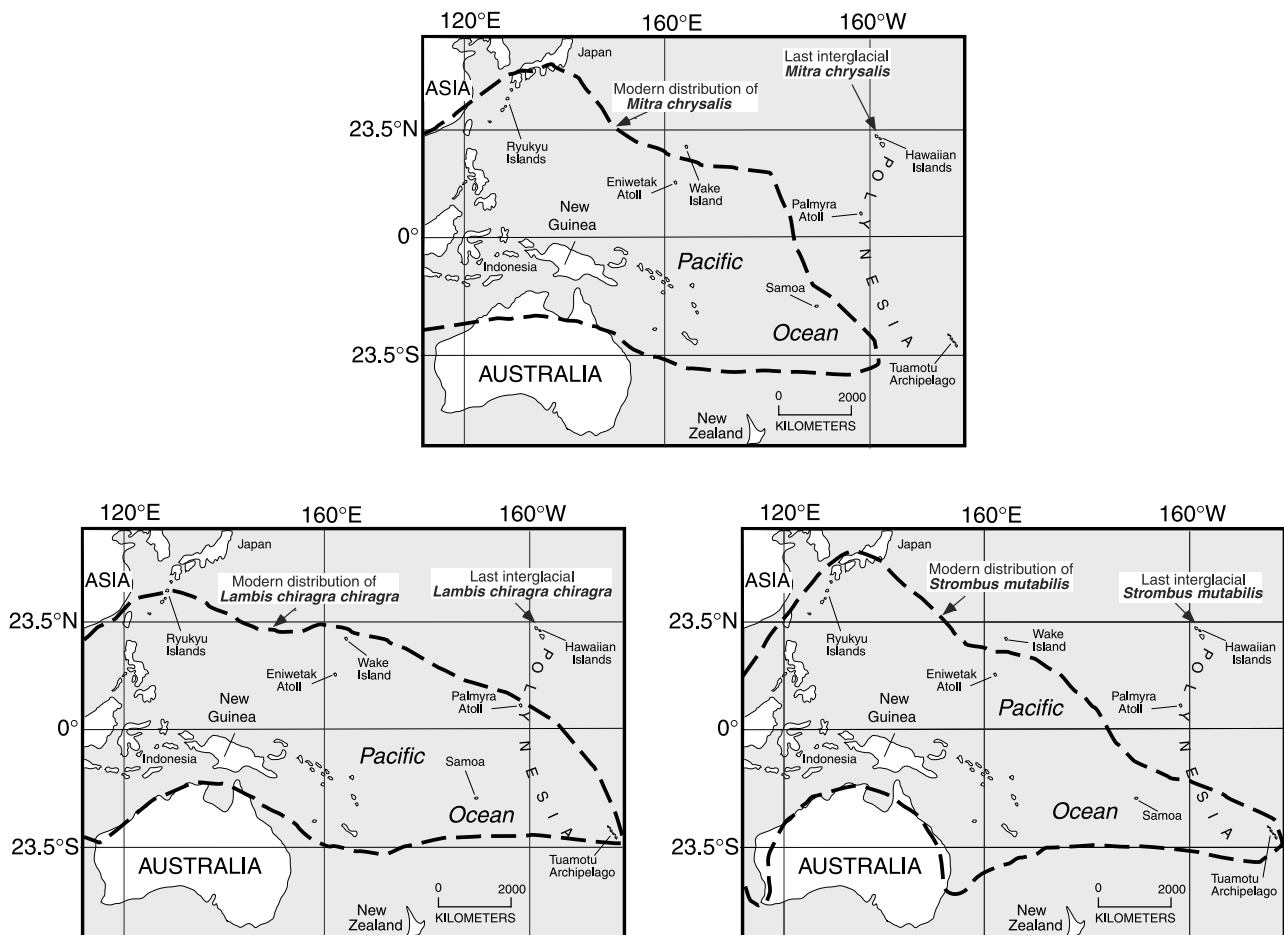


Fig. 9. Modern geographic ranges of *Lambis chiragra chiragra* (redrawn from Abbot, 1961), *Strombus mutabilis* (redrawn from Abbot, 1960), and *Mitra* (*Nebularia*) *chrysalis* (redrawn from Cernohorsky, 1976) in the Indo-Pacific region.

Formation (Vacher et al., 1989). A single alpha-spectrometric U-series analysis of an *Oculina* from the marine facies at Fort St. Catherine gave an age of 85 ± 12 ka (Harmon et al., 1983). They considered the deposit to have been storm-reworked and did not regard the age as an indication of a higher-than-present sea level at ~ 85 ka. Vacher and Hearty (1989), however, challenged the storm-emplacement hypothesis of Harmon et al. (1983) with new stratigraphic and amino acid data and suggested that Fort St. Catherine was an unlikely locality for deposition by storm waves. They proposed that the deposit could indeed represent a higher-than-present sea level at ~ 80 ka. Ludwig et al. (1996) presented new TIMS U-series ages of 77–82 ka for four *Oculina* specimens from Fort St. Catherine and reiterated the arguments of Vacher and Hearty (1989), agreeing that the deposit could represent a high sea stand at ~ 80 ka. Toscano and Lundberg (1999), however, rejected the arguments of Vacher and Hearty (1989) and Ludwig et al. (1996) and returned to the storm emplacement hypothesis of Harmon et al. (1983). The possibility that

the storm emplacement event occurred during the Holocene was also proposed by Toscano and Lundberg (1999).

Because of the controversy about this deposit, and because of the significant implications it has for the magnitude of past sea level, we analyzed a large suite of corals from Fort St. Catherine. Our goal was to provide a rigorous test of the storm emplacement hypothesis. If, as Toscano and Lundberg (1999) suggested, the storm emplacement event occurred during the Holocene, then the deposit at Fort St. Catherine should contain corals of any age that are found on the Bermuda Platform. Obviously, this could include corals of Holocene age, but also older corals, such as those dated at ~ 87 , 94 and 103 ka from the upper parts of cores taken on the platform to the northwest of Fort St. Catherine (Vollbrecht, 1990).

Results of analyses of 17 *Oculina* fragments from Fort St. Catherine show a narrow range of ages. All but one sample (FSC-OC-1-O, which shows a lower-than-acceptable initial $^{234}\text{U}/^{238}\text{U}$ value) show excellent preservation and near-ideal criteria for closed-system

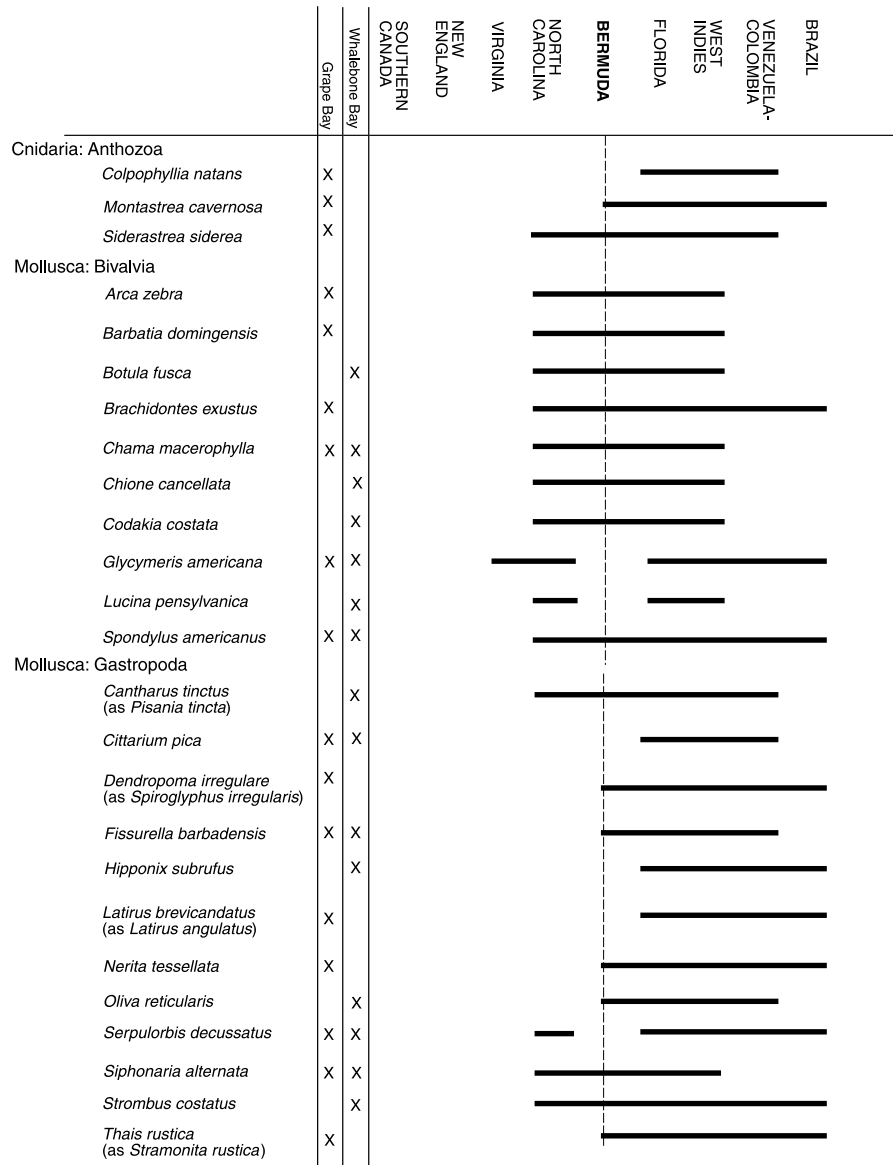


Fig. 10. Modern geographic ranges of fossils found in the Devonshire marine member of the Rocky Bay Formation of Bermuda at localities with either U-series ages of ~ 120 ka (from Harmon et al., 1983 and this study) or amino acid data which indicate an age of ~ 120 ka (Vacher and Hearty, 1989; Hearty et al., 1992). Fossil data from Epstein and Lowenstam (1953), Richards et al. (1969), and this study; see text for sources of modern geographic ranges.

conditions. All had U contents of 2–3 ppm, no evidence of inherited Th, and initial $^{234}\text{U}/^{238}\text{U}$ values that were indistinguishable from modern sea water values. Thus, we regard the ages from this suite of corals as very reliable. Ages range from 78.3 ± 0.4 to 84.0 ± 0.7 ka and have a mean value of ~ 80 ka. There is no evidence of reworking of either older (~ 100 ka) or younger (Holocene) corals that are known to be present on the Bermuda Platform to the northwest of Fort St. Catherine. Thus, we reject the Holocene storm emplacement hypothesis. Because the corals we observed at Fort St. Catherine were not in growth position and there was not evidence of in situ patch reefs at that locality, it is

possible that the deposit originated by a storm emplacement event or events that occurred *during* the ~ 80 ka high sea stand. However, the lack of older (~ 100 or ~ 125 ka) corals would tend to mitigate against this. Finally, it is possible that the deposit does not have a storm-emplacement origin but also does not represent a sea stand near present at ~ 80 ka. Such an origin is possible if islands and coastlines at mid-latitudes, closer to the positions of full-glacial ice sheets than those in tropical latitudes, experience isostatic rebound that elevates marine deposits to positions higher than where they formed initially (Stirling et al., 1998).

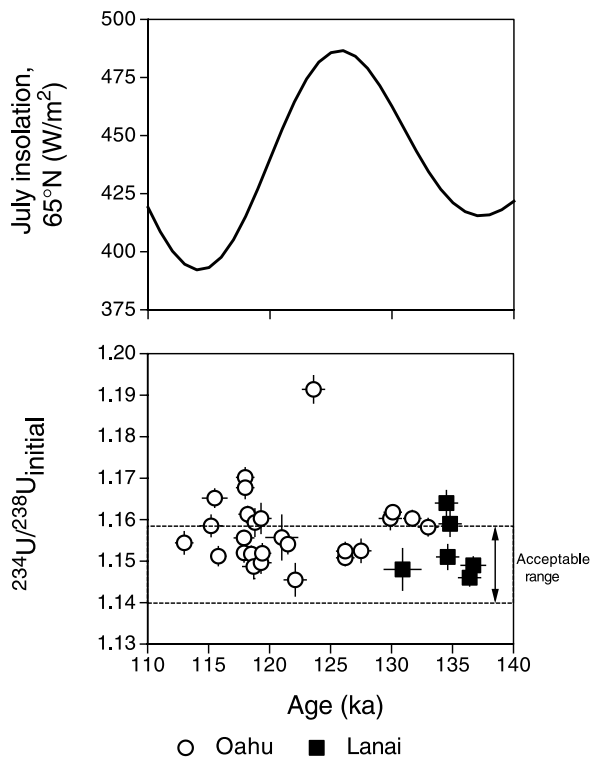


Fig. 11. Calculated initial $^{234}\text{U}/^{238}\text{U}$ activity ratios for Hawaiian corals from the Waimanalo Formation on Oahu (open circles) and from marine terrace deposits on Lanai (solid squares) plotted against their $^{230}\text{Th}/\text{U}$ ages with 2-sigma error bars; where no bars appear, errors are smaller than the plot symbol. Also shown are July insolation values at the top of the atmosphere at 65°N for the period from 140 to 110 ka (insolation data from Berger and Loutre, 1991). "Acceptable range" of initial $^{234}\text{U}/^{238}\text{U}$ activity ratios is 1.140 to 1.159 (see Discussion in text). Lanai U-series ages from Rubin et al. (2001); Oahu U-series ages from this study (Table 1). Where duplicate analyses were conducted, the sample with the smallest error is plotted.

5. Paleozoogeography and ocean temperatures

5.1. Approaches

Temperatures of the Last Interglacial period have been simulated using atmospheric general circulation models. Montoya et al. (1998) used such a model to infer significantly higher (up to 4°C) summer temperatures and similar or slightly warmer winter temperatures over continental North America during the Last Interglacial period. In contrast, Last Interglacial sea surface temperatures are inferred to have increased only slightly ($\sim 1^\circ\text{C}$) relative to present around much of North America, with little or no difference at low latitudes and slightly greater differences at high latitudes (Montoya et al., 1998).

Paleotemperatures of the Last-Interglacial ocean also have been inferred from analysis of foraminifera in deep-sea cores. Using a global suite of cores, CLIMAP Project Members (1984) suggested that the Last-Inter-

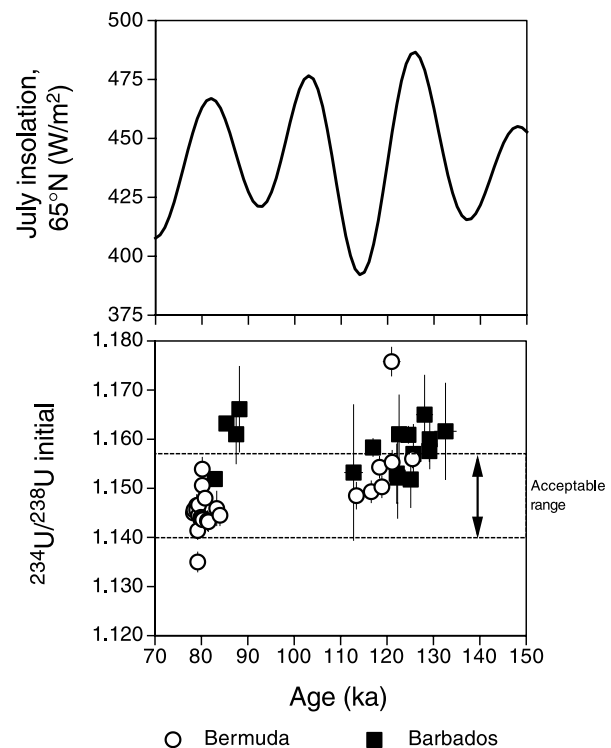


Fig. 12. Calculated initial $^{234}\text{U}/^{238}\text{U}$ activity ratios for Bermuda corals from the Devonshire marine member of the Rocky Bay Formation and the Southampton Formation plotted against their $^{230}\text{Th}/\text{U}$ ages with 2-sigma error bars; where no bars appear, errors are smaller than the plot symbol. Also shown are similar data from the Rendezvous Hill ("Barbados III") and Worthing ("Barbados I") terraces from Barbados and July insolation values at the top of the atmosphere at 65°N for the period from 150 to 70 ka. Barbados data from Edwards et al. (1987, 1997), Bard et al. (1990), Hamelin et al. (1991), and Gallup et al. (1994); insolation data from Berger and Loutre (1991). "Acceptable range" of initial $^{234}\text{U}/^{238}\text{U}$ activity ratios as in Fig. 13.

glacial ocean may have been, overall, very similar to that of the present, but they also reported numerous anomalies, some due to sampling or data biases and others unexplained. In the CLIMAP study, however, few cores were studied from the western Atlantic, eastern Pacific, and Arctic Oceans; thus, little could be inferred about sea surface temperatures around North America. Cortijo et al. (1999) estimated sea surface temperatures from planktonic foraminifera in cores from the western Atlantic Ocean, including two from around the latitude of Bermuda. However, because few Holocene calibration cores exist in this region, uncertainties in sea surface temperature estimates are as much as 3.5°C . In the tropical region of northwestern Australia, McCulloch and Esat (2000) used Sr/Ca values in corals to infer Last-Interglacial sea surface temperatures little different from today's.

We use a zoogeographic approach (Kennedy, 1978, 2000; Kennedy et al., 1982, 1992) to estimating relative sea surface temperatures around Hawaii, Bermuda and North America during the Last Interglacial period. This

method utilizes the fact that modern geographic ranges of nearshore marine invertebrates, such as mollusks, ostracodes and corals, are mainly a function of the latitudinal temperature gradient between equatorial and polar regions. Species that presently live only to the north of any fossil locality are referred to as extralimital northern species and would be indicative of cooler-than-present marine conditions. Conversely, any deposit with extralimital southern species can be inferred to have experienced warmer-than-present ocean temperatures. Kennedy (2000) points out that inferences of relatively warmer or cooler waters can also be made from a dominance of species that are not strictly extralimital, but whose modern ranges have endpoints at or near the fossil locality of concern.

5.2. Oahu, Hawaii

Fossil mollusks are abundant in Pleistocene limestones of Oahu (Ostergaard, 1928; Kosuge, 1969; Kohn, 1980). The U-series ages of corals reported by Ku et al. (1974), Szabo et al. (1994), Muhs and Szabo (1994), and the present study allow identification of those fossil mollusk localities on Oahu that date to the Last Interglacial period. Of the numerous localities reported by Ostergaard (1928) and Kosuge (1969), only Kahuku Point and Kahe Point (Fig. 2) are confidently dated to the Last Interglacial period. Nevertheless, the faunas at Kahe Point and Kahuku Point are diverse enough that some inferences can be made about oceanic paleotemperatures.

Kosuge (1969) reported 69 species of mollusks, mostly gastropods, from the Kahe Point and Kahuku Point localities. Although some of these species are extinct and others are endemic to the Hawaiian Islands, the majority of species live today around Hawaii (Kay, 1979), as well as throughout much of the Indo-Pacific region. Kosuge (1969) interpreted the faunas from Oahu as indicative of warmer-than-present waters in the past, but some species distributions were not well known at the time of his study. Based on a re-examination of his species lists and newer studies of Indo-Pacific molluscan zoogeography (Cernohorsky, 1976, 1978; Kay, 1979; Warén, 1980; Kilburn, 1983; Bernard et al., 1993; Kreipl and Poppe, 1999; Abbott and Dance, 2000), we have determined that eight species do not live in Hawaiian waters today, but range elsewhere in warmer-water regions of the Indo-Pacific (Table 4). *Isognomon marsupialis* is known to occur today in the South China Sea and as far north as Okinawa (Bernard et al., 1993), but does not occur in Hawaii at present (Kay, 1979). Its occurrence at both Kahe Point and Kahuku Point on Oahu indicates a Last-Interglacial range extension of ~7500 km. *Lambis chiragra* occurs in the northern and eastern Indian Ocean, the Ryukyu Islands, northern Australia and east to the Tuamotu Archipelago (Fig. 9).

The closest documented modern record for *L. chiragra* is Palmyra Atoll, more than 1500 km to the south of Hawaii. The same is true for *L. truncata*, which is reported for an unspecified locality in the Waimanalo Formation by Kay (1979). Kosuge (1969) also reported the species as a fossil on Oahu, but did not specify a locality. *Vexillum semifasciatum* (as *Pusia semifasciata*) ranges only from the Red Sea to Samoa. *Scalenostoma sublata* (as *Stilifer speciosa*) has a circumtropical distribution, but is presently unknown from Hawaii. *Strombus mutabilis* ranges throughout the Indo-Pacific, but not in Hawaii nor any part of eastern Polynesia (Fig. 9). The closest modern record for this species is Samoa, more than 4000 km to the south of Hawaii. *Mitra chrysalis* ranges from the Red Sea and Madagascar, through the Indian Ocean to the Ryukyu Islands and Cook Islands (Fig. 9). The closest modern record of *M. chrysalis* is Eniwetok Atoll, more than 4000 km to the southwest of Hawaii. Finally, *Acar plicatum* ranges from South Africa, through the Indian Ocean to the Philippines, Ryukyu Islands, and northwestern Australia, but not to Hawaii. Thus, we interpret all species cited above as warm-water indicators. Two additional species reported by Kosuge (1969) from Kahe Point are *Cypraea erosa* Linné and *Conus tulipa* Linné. Although both species are found in Hawaii today (Kay, 1979), they are extremely rare. *Cypraea erosa* is widely distributed in the Indo-West Pacific (Kay, 1979), and although uncommon, *C. tulipa* is also found in the broader Indo-Pacific (Cernohorsky, 1978). Thus, both species can also be regarded as warm-water indicators, as they are apparently at or near their northern distributional limits in Hawaii. We conclude from this compilation that water temperatures around Oahu during the Last Interglacial period were at least as warm as present and probably, based on species that range only to the south and southwest of Oahu today, warmer than present, as concluded by Kosuge (1969).

5.3. Bermuda

The molluscan faunas from Last-Interglacial deposits on Bermuda have been well documented by Epstein and Lowenstam (1953), Richards et al. (1969) and Hearty et al. (1992). We compiled the species identified by these workers from two localities, Grape Bay and Whalebone Bay, where there are either well-documented U-series ages on corals (from Harmon et al., 1978, 1983 and this study) or amino acid ratios in mollusks similar to those in mollusks from localities with dated, co-existing corals (Vacher and Hearty, 1989 and Hearty et al., 1992). We also present the modern geographic ranges of these taxa, taken from the most recent data available in Sterrer (1986), Abbott (1974), Weisbord (1974), Abbott and Morris (1995) and web data bases from the Academy of

Table 4
Extralimital southern species from the Waimanalo Formation, Oahu, Hawaii

Species	Other species name	Localities of Kosuge (1969)	Modern geographic range	Source
<i>Isognomon marsupialis</i> Röding	<i>Isognomon marsupiale</i> Röding	Kahe Point, Kahuku Point	South China Sea, Okinawa	Bernard et al. (1993)
<i>Lambis chiragra</i> Linné	n/a	Kahe Point	Northern and eastern Indian Ocean (Sri Lanka to Indonesia), Ryukyu Islands to northern Australia and Tuamotu Archipelago	Abbot (1961), Kreipl and Poppe (1999)
<i>Lambis truncata sebae</i> (Kiener)	n/a	Waimanalo high sea stand from unspecified locality Kosuge, 1969; Kay, 1979)	Red Sea and Persian Gulf; Borneo, Philippine Islands, Ryukyu Islands to northern Australia and Tuamotu Archipelago	Abbot (1961)
<i>Mitra chrysalis</i> Reeve	n/a	Kahuku Point	Red Sea and Madagascar, Indian Ocean to Ryukyu Islands and Cook Islands	Cernohorsky (1976)
<i>Pusia semifasciata</i> Lamarck	<i>Vexillum semifasciatum</i>	Kahe Point	Red Sea to Samoa	Abbott and Dance (2000)
<i>Stilifer speciosa</i> (A. Adams)	<i>Scalenostoma sublata</i> (Broderip)	Kahe Point	Circumtropical, but not the eastern Pacific	Warén (1980)
<i>Strombus mutabilis</i> Swainson	n/a	Kahe Point, Kahuku Point	Entire Indo-Pacific except for Hawaii and eastern Polynesia	Abbot (1960); Kreipl and Poppe (1999)
<i>Acar plicatum</i> (Dillwyn)	n/a	Kahe Point	South Africa, throughout Indian Ocean, Philippines, Ryukyu Islands to north-western Australia	Kilburn (1983)
<i>Conus tulipa</i> Linné	n/a	Kahe Point	Uncommon in Indo-Pacific; very rare in Hawaii	Cernohorsky (1978); Kay (1979)
<i>Cypraea erosa</i> Linné	n/a	Kahe Point	Common in Indo-West Pacific; very rare in Hawaii	Kay (1979)

Natural Sciences (Philadelphia) and the Florida Museum of Natural History. Geographic ranges of mollusks in these various sources sometimes differ; where this is the case, we use the geographic range given in the more recent reference. All these sources update the modern geographic ranges of Bermudan fossils given in Richards et al. (1969).

A warm-water aspect is present in the faunal assemblage from the Last-Interglacial Devonshire marine member of Bermuda (Fig. 10). The Devonshire contains three extralimital southern gastropods (*Cittarium pica*, *Hipponix subrufus*, and *Latirus brevicandatus* [as *Latirus angulatus*]) that do not live around Bermuda today. Modern northern range endpoints of these three species are in the Florida Keys, although there is a single, poorly documented (1938) report of *C. pica* near St. Petersburg, Florida in the records of the Florida Museum of Natural History. The southern range endpoints of these species are either Venezuela or Brazil. Sterrer (1986) points out that in unpublished data of A.E. Verrill, *C. pica* has twice been introduced to

Bermuda and failed to survive. If the Florida Keys are the northernmost limit for these species, their presence in Last-Interglacial deposits of Bermuda indicates northward range extensions of ~800 km. In addition to extralimital southern mollusks, the coral *Colpophyllia natans*, found in the Devonshire marine deposits at Grape Bay, presently lives only as far north as the southern Florida Keys (Weisbord, 1974; Shinn et al., 1989), but as far south as Venezuela (Weisbord, 1974; Chiappone et al., 1996). Although Weisbord (1974) listed *C. natans* from Bermuda, recent study shows that it does not live around Bermuda at present (Sterrer, 1986). Thus, its presence in the Devonshire marine member indicates a Last-Interglacial northward range extension of ~800 km. Six other species, five gastropods and a coral, live no farther north than Bermuda today and have southern range endpoints off either Venezuela or Brazil. All the coral and mollusk species (with the exception of *Glycymeris americana*, which ranges as far north as Virginia), range no farther north than North Carolina and have southern range endpoints in the West

Indies or off Venezuela or Brazil. We therefore interpret the Devonshire fauna to have a warmer-than-modern aspect.

6. Discussion

6.1. Timing of the peak of the Last Interglacial period

Our new U-series data for Oahu and Bermuda help evaluate hypotheses about the duration of the peak of the Last Interglacial period. The deep-sea oxygen isotope record, as discussed earlier, would suggest a *peak* Last Interglacial period, when global ice volume was similar to or less than present, of less than 3 ka (based on the estimated time period from events 5.53 to 5.51 of Martinson et al., 1987). On the other hand, other records, such as Antarctic ice cores, Devils Hole calcite, Grand Pile pollen, and certain emergent reefs would indicate a much longer peak Last Interglacial duration (e.g., ~10 ka estimated from Devils Hole; Winograd et al., 1997). The Oahu U-series data presented here have an age range of ~134–113 ka, which is not significantly different from that reported by Szabo et al. (1994). McCulloch and Esat (2000) suggested that the Oahu data set of Szabo et al. (1994) was of too-low a precision to allow an estimate of the duration of the Last Interglacial period. However, our data set (different samples but from the same localities on Oahu) is of very high precision. We conclude that the concept of a long Last Interglacial period on Oahu, as proposed by Szabo et al. (1994) is still valid. Our new estimates indicate that *peak* Last-Interglacial conditions, when global ice volume was less than present, could have been many times longer than that estimated from the deep-sea oxygen isotope record.

Recent work on Lanai (Fig. 2) by Rubin et al. (2001) has yielded additional high-precision TIMS U-series ages of marine terrace corals. These deposits were originally considered to have been deposited by a giant wave generated by a submarine landslide (Moore and Moore, 1984, 1988), but are now interpreted to be marine terrace deposits that reached their present elevations by uplift (Rubin et al., 2001). If this interpretation is correct, then these U-series ages mark the time of relatively high sea levels during past interglacial events, including the Last Interglacial. Five corals analyzed by Rubin et al. (2001) have ages of ~137–131 ka with acceptable initial $^{234}\text{U}/^{238}\text{U}$ values. The Lanai coral ages overlap our oldest Oahu coral ages slightly but most are older (Fig. 11). It is important to note that the Lanai marine terrace deposits mostly have elevations that are higher than those on Oahu; one locality is at 3.8 m, but the others range from 19 to 23 m above modern sea level (Rubin et al., 2001). Thus, although Oahu may be experiencing slow uplift (Muhs

and Szabo, 1994), Lanai may be experiencing uplift rates that are higher. If Lanai is indeed experiencing uplift at a higher rate than Oahu, then somewhat older coral ages might be expected there.

Bradley and Griggs (1976) and Edwards et al. (1987) presented models of reef or terrace formation on uplifting coasts during the high-sea-level phase of an interglacial–glacial cycle. We have modified these models to incorporate the concepts of “keep-up,” “catch-up” and “give-up” reef types, as outlined by Neumann and MacIntyre (1985). Lanai, with its higher uplift rate, should record an interglacial high sea stand sooner than Oahu as uplift keeps pace with rapidly rising sea level. The results from the present study seem to support this model. If so, then we might also expect to see similar differences on other coasts as well. To test this hypothesis, we compared the Bermuda U-series ages with those for Barbados. The difference in coral ages might be expected to be even greater than those for Oahu–Lanai, not only because Bermuda is tectonically stable compared to uplifting Barbados, but primarily because Bermuda patch reefs are “catch-up” types whereas *Acropora palmata*-dominated reefs on Barbados are “keep-up” types. A comparison with both the peak-Last Interglacial (~120 ka) corals (Devonshire member on Bermuda; Rendezvous Hill terrace on Barbados) and late-Last Interglacial (~80 ka) corals (Southampton Formation on Bermuda and Worthing terrace on Barbados) supports the hypothesized lag of Bermudan coral growth (Fig. 12). Some of the analytical uncertainties for the Barbados corals are larger than what would be desirable for a rigorous test of the hypothesis; nevertheless, mass-spectrometric $^{231}\text{Pa}/^{235}\text{U}$ dating of Barbados corals indicates that the $^{230}\text{Th}/^{234}\text{U}$ ages are probably valid (Edwards et al., 1997). Thus, we conclude that the sea-level model presented by Edwards et al. (1987) is supported by the comparison of U-series ages. If so, then the Last Interglacial high stand of sea recorded on the Hawaiian Islands may have begun around ~135 ka or even slightly before and remained high until at least 117 ka or even slightly later. Sea level may have been high before ~135 ka because of a lag between the time of sea level rise and reef growth as observed by Hopley (1982, 1994) and Davies and Hopley (1983). We do not see any evidence in the array of U-series ages to infer a precipitous drop in sea level of ~60–80 m at around 130 ka, as suggested in the studies of Esat et al. (1999) on New Guinea. We also do not see any persuasive evidence for two separate high stands of sea as interpreted by Sherman et al. (1993) from elsewhere on Oahu. However, the U-series data do support an interpretation that global warming with near-or-higher-than-present sea level could have occurred well before the northern high-latitude summer insolation peak of ~126 ka. Such a hypothesis is also suggested by the Devils Hole, Nevada calcite record

(Winograd et al., 1992, 1997) and the oxygen isotope record in Vostok, Antarctica ice cores (Lorius et al., 1985; Jouzel et al., 1993). The recent direct dating of marine sediments off the Bahamas islands by Henderson and Slowey (2000) also supports the concept of interglacial conditions occurring well before the northern high-latitude summer insolation peak at ~ 126 ka.

6.2. The dilemma of a high sea stand at ~ 80 ka

Although much more attention has been given to the ~ 125 ka high stand of sea, the magnitude of sea level rise at ~ 80 ka has also been actively debated, and careful studies show conflicting results. Indeed, on a global basis, there are about equal numbers of studies showing that this high stand was either relatively low or was relatively high. On Bermuda, as discussed earlier, a potential record of this high sea stand is the marine facies of the Southampton Formation found at Fort St. Catherine. Whereas Vacher and Hearty (1989) and Ludwig et al. (1996) interpreted this deposit to be indicative of an ~ 80 ka high sea stand close to the present, Harmon et al. (1983) and Toscano and Lundberg (1999) interpreted the marine sediments at Fort St. Catherine to be storm deposits. Because the deposits are not in situ patch reefs but bedded, poorly sorted, coral-bearing marine sands and gravels, a storm-emplacement origin cannot be excluded. Based on the large suite of ~ 80 ka coral ages (and *only* ~ 80 ka coral ages), our interpretation is that the marine sediments at Fort St. Catherine do not represent a Holocene storm emplacement event, as hypothesized by Toscano and Lundberg (1999). It is possible that the deposits represent one or more storm emplacement events that occurred during the ~ 80 ka high sea stand, but the lack of ~ 100 ka corals, documented elsewhere on the Bermuda Platform northwest of Fort St. Catherine, may not support this interpretation.

Tectonically stable (or only slowly uplifting) coastlines around the world show conflicting records of sea level at ~ 80 ka. On the tectonically stable east coast of the United States, a shallow submerged reef at Sand Cay off the Florida coast dated to ~ 80 ka has been interpreted both as indicating a sea level near present (Ludwig et al., 1996) or several meters below present (Toscano and Lundberg, 1999). On the tectonically stable Coastal Plain of the eastern United States, corals from emergent marine deposits a few meters above modern sea level (Savannah, Georgia; Charleston, South Carolina; Moyock, North Carolina; and Gomez, Virginia: see Fig. 3), have been dated at 75–80 ka using alpha-spectrometric $^{230}\text{Th}/^{234}\text{U}$ methods, supported by concordant $^{231}\text{Pa}/^{235}\text{U}$ ages (Szabo, 1985). Redating of these corals using TIMS U-series methods has confirmed the 75–80 ka alpha-spectrometric ages (Wehmiller et al., 1997). The tectonically stable Bahamas show

mixed evidence of sea level position at this time. U-series dating of submerged speleothems suggests that sea level was at least 10–15 m below present at ~ 80 ka (Lundberg and Ford, 1994; Richards et al., 1994). Conversely, ooids (which can form only when sea level is near present on the Bahamas platform) show evidence of a high sea stand younger than ~ 120 ka, but not as young as Holocene (Hearty and Kaufman, 2000). The tectonically stable coast of Western Australia does not show any evidence of an ~ 80 ka high sea stand near present sea level (Stirling et al., 1995, 1998), nor does Isla de Guadalupe off the coast of Baja California (Muhs et al., 1994). On Oahu, a single U-series age of ~ 80 ka from a growth-position *Porites lobata*, at a depth of ~ 30 m, has been reported (Sherman et al., 1999). However, *P. lobata* can live from the intertidal zone to depths of ~ 60 m below sea level (Maragos, 1977).

There are also conflicting estimates of paleo-sea level at ~ 80 ka from tectonically active coastlines. Terraces dated to ~ 80 ka on the tectonically rising islands of Barbados, Haiti and New Guinea (Mesolella et al., 1969; Bloom et al., 1974; Dodge et al., 1983) are estimated to have formed at a sea level 10–20 m below present, using the 120 ka terrace to calculate an uplift rate and assuming linear tectonics over time. However, other tectonically rising coasts (Japan and California) suggest that sea level was slightly below, although close to present level at ~ 80 ka, when calculated using the same methods (Ota and Omura, 1992; Muhs et al., 1994).

Despite the abundance of well-dated corals from both tectonically stable and rising coastlines and islands, we cannot come to any firm conclusions about paleo-sea level at ~ 80 ka. One observation that can be made, however, is that those localities (Bermuda, eastern US Coastal Plain, California, Japan and perhaps the Bahamas) showing evidence of a near-present sea level at ~ 80 ka belong to what has been termed “intermediate-field” localities with respect to northern high-latitude ice sheets (Stirling et al., 1998). These localities are closer to limits of the penultimate Laurentide ice sheet and are hypothesized to have experienced post-glacial isostatic rebound that elevated ~ 80 and ~ 105 ka marine deposits to positions above modern sea level. In contrast, “far-field” coastlines (Australia, New Guinea and perhaps Barbados and Oahu), distant from northern high-latitude ice sheets, show no evidence of ~ 80 and 105 ka marine deposits near present sea level. Nevertheless, an argument against the isostasy hypothesis is the observation that ~ 125 ka deposits on tectonically stable, intermediate-field localities (the Bahamas, Bermuda and localities in Baja California: see Harmon et al., 1983; Chen et al., 1991; Muhs et al., 1994) are not significantly higher than ~ 125 ka deposits on tectonically stable, far-field localities (Australia).

With regard to sea level at ~ 80 ka, it is perhaps worth noting that summer insolation at 60°N was actually

higher at ~80 ka than at 11 ka (Berger and Loutre, 1991). At 65°N, summer insolation at ~80 ka was only slightly below the 11 ka maximum (Fig. 12). One of the most important centers of growth for the Laurentide ice sheet at these latitudes was Baffin Island, Canada. Recent studies of pollen and diatoms from widely separated lakes on Baffin Island show that climate at ~80 ka (estimated from concordant TL and IRSL ages) was warmer than at any time in the Holocene (Miller et al., 1999; Wolfe et al., 2000). One of the inferences from the Baffin Island studies is that under such conditions, there must have been essentially complete deglaciation of the Laurentide ice sheet at ~80 ka (Miller et al., 1999). If this conclusion is correct, then evidence of a high sea stand near present at ~80 ka should be expected.

6.3. Paleozoogeography of North American coastal sites of Last-Interglacial age

We compiled published, Last-Interglacial marine faunal data from around North America that can be compared to the faunas for Hawaii and Bermuda. In making this compilation, we used mostly mollusks, but also corals and ostracodes where such data were available. We limited our selection, as much as possible, to those localities with reliable U-series ages on corals. For those deposits lacking corals (mostly Greenland, Canada and Alaska) we consider as “Last Interglacial” those deposits that meet the following four criteria: (1) nonfinite radiocarbon ages, (2) a stratigraphic position that allows an inference of Last-Interglacial age (e.g., sediments directly overlain by dated Wisconsin-age deposits), (3) deposits that lie at elevations slightly above modern sea level in tectonically stable settings, and (4) amino acid ratios on mollusks that permit an interpretation of Last-Interglacial age. Some localities, such as those on Greenland, also have luminescence ages that support a Last-Interglacial age. Geographic ranges were updated from sources such as Keen (1971), Abbott (1974), Weisbord (1974), Kay (1979), Sterrer (1986), Abbott and Morris (1995), Coan et al. (2000), and web sites available at the Academy of Natural Sciences (Philadelphia), the Bernice P. Bishop Museum (Hawaii), and the Florida Museum of Natural History.

Results of this compilation show that high-latitude waters around North America and nearby islands had Last-Interglacial oceanic temperatures warmer than present (Fig. 13). Coastal waters around Alaska, in both the Arctic Ocean and the Bering Sea, show warmer-than-modern temperatures from both mollusks in Last Interglacial (Pelukian) marine terrace deposits and marine ostracodes collected from cores taken in Prudhoe Bay (McDougall et al., 1986; Brigham-Grette and Hopkins, 1995). For example, the bivalve *Mya elegans* (as *Mya baxteri*) has a modern northern range

limit near Nome, Alaska (~64.5°N) on the Seward Peninsula (Coan et al., 2000), but is found in Last-Interglacial marine deposits on the Arctic Coastal Plain of Alaska (Brigham-Grette and Hopkins, 1995), a northward range extension of more than 600 km. Marine deposits from the Last Interglacial (Kogalu aminozone) of Baffin Island (Miller et al., 1992) contain subarctic faunas well north of their modern northern range endpoints (Miller et al., 1977; Nelson, 1981; Brigham, 1983). Two of the most common extralimital southern (subarctic) species of mollusks that have been reported in Last-Interglacial deposits are *Chlamys islandica* and *Colus spitzbergensis*, neither of which live in the colder, Arctic waters off eastern Baffin Island today. Mode (1985) also reported *Astarte subaequaliter*, which ranges today from Florida north only to Labrador (Abbott and Morris, 1995). Its presence in Last-Interglacial deposits of eastern Baffin Island indicates a northward range extension of as much as 1500 km. Subarctic waters reached farther north than they do at present along the coasts of both West and East Greenland. In East Greenland, extralimital southern mollusks, such as *C. islandica*, today found only in subarctic waters, are present in emergent marine deposits of the Langelandselv interglaciation in the Jameson Land area (Funder et al., 1991, 1998; Hansen et al., 1994; Landvik et al., 1994; Tveranger et al., 1994). Luminescence age estimates suggest that these deposits date to the Last Interglacial period (Landvik et al., 1994; Mejdahl and Funder, 1994; Funder et al., 1998). In West Greenland, near Thule, warmer-than-present waters are also inferred for Last-Interglacial deposits, based on the presence of subarctic mollusks and barnacles that are at or north of their modern limits (Funder et al., 1991; Kelly et al., 1999).

Farther south along the Atlantic coast of Canada and into New England, warmer-than present waters are also inferred for the Last Interglacial period. Deposits in Nova Scotia have mollusks (Clarke et al., 1972) with amino acid ratios similar to those from coral-bearing deposits with a Last-Interglacial age on Nantucket Island, Massachusetts, based on U-series dating (Oldale et al., 1982; Wehmiller et al., 1988). The Last-Interglacial deposits on Nova Scotia have a “neutral” fauna, without extralimital southern or northern species (Clarke et al., 1972). The same is true for the Last-Interglacial mollusks on Nantucket Island. However, the zoogeographic ranges of ostracodes in the marine deposits on Nantucket Island show a transition from warmer-than-modern conditions at the peak of the Last Interglacial period to cooler-than-modern conditions at the end of this period (Oldale et al., 1982).

Marine deposits from lower-latitude localities in Florida, the Bahamas, and the West Indies generally show “neutral” faunal thermal aspects (terminology of

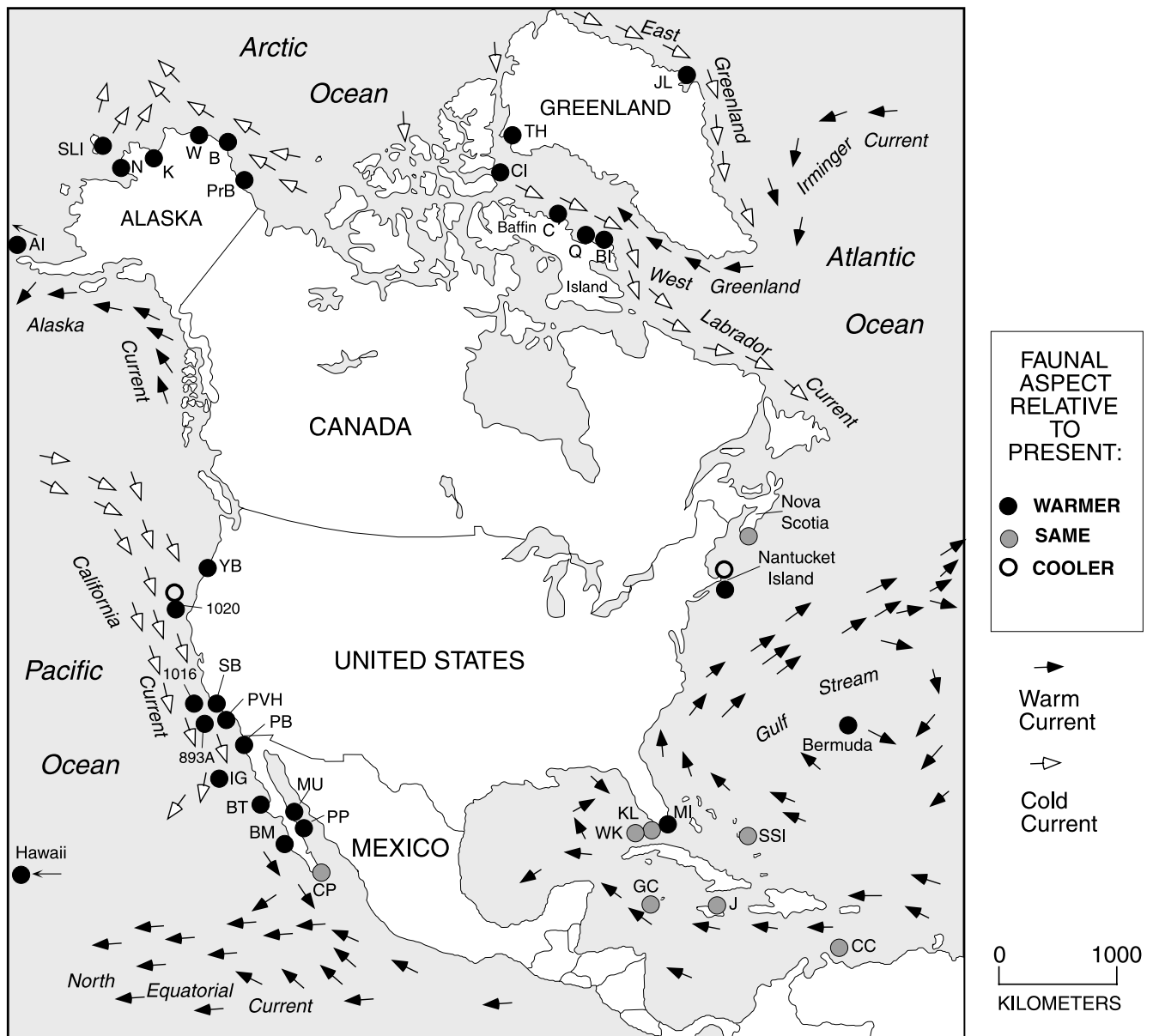


Fig. 13. Compilation of fossil localities that date (or *probably* date) to the peak of the last interglacial period around North America, their temperature aspects relative to modern waters based on presence or absence of extralimital southern or northern species in the fossil assemblages, and modern warm and cold ocean currents. Most localities shown are for mollusk data, but one has ostracodes (PrB), another has both mollusks and ostracodes (Nantucket Island), two others use mollusks and corals (Bermuda and Windley Key) and Ocean Drilling Program (ODP) sites off California use alkenone and foraminifera data. Correlation to the last interglacial period based on either U-series ages of corals (most localities in the continental United States and Mexico), amino acid ratios in mollusks (most localities in Canada and Alaska), amino acid ratios in mollusks and luminescence dating of sediments (localities in Greenland), or oxygen isotope stratigraphy (ODP sites off California). Abbreviations: PrB, Prudhoe Bay, Alaska; B, Barrow, Alaska; W, Wainwright, Alaska; K, Kotzebue, Alaska; SLI, St. Lawrence Island, Alaska; N, Nome, Alaska; AI, Amchitka Island, Alaska; YB, Yaquina Bay, Oregon; 1020, ODP Hole 1020, northern California; 1016, ODP Hole 1016, off Point Conception; SB, Santa Barbara, California; 893A, ODP Hole 893A, Santa Barbara Basin; PVH, Palos Verdes Hills, California; PB, Punta Banda, Baja California; IG, Isla de Guadalupe, Baja California; BT, Bahía Tortugas, Baja California Sur; BM, Bahía de Magdalena, Baja California Sur; CP, Cabo Pulmo, Baja California Sur; PP, Playa La Palmita, Baja California Sur; MU, Mulegé, Baja California Sur; CC, Curaçao; GC, Grand Cayman Island; J, Jamaica; WK, Windley Key, Florida; KL, Key Largo, Florida; MI, Miami, Florida; SSI, San Salvador Island, Bahamas; CI, Coburg Island, Canada; BI, Broughton Island, Canada; Q, Qivitu Peninsula, Canada; C, Clyde Foreland, Canada; TH, Thule area, Greenland; JL, Jameson Land, Greenland. Fossil and age data compiled by the authors from the following sources: *Alaska*: Allison (1973), Szabo and Gard (1975), McDougall et al. (1986), and Brigham-Grette and Hopkins (1995); *Oregon and California*: Kennedy (1978, 1992), Kennedy et al. (1982), Muhs et al. (1992, 1994), Kreitz et al. (2000); Yamamoto et al. (2000); Poore et al. (2000); Herbert et al. (1995); Woodring et al. (1946); *Baja California*: Rockwell et al. (1989), Omura et al. (1979), Libbey and Johnson (1997); Ortlieb (1984); Ashby et al. (1987); Emerson et al. (1981), and Squires (1959); *Curaçao*: Buissonjé (1974), Schubert and Szabo (1978); *Grand Cayman Island*: Cerridwen and Jones (1991), Vézina et al. (1999); *Jamaica*: Szabo (1979), Donovan and Littlewood (1993); *San Salvador Island*: Hagey and Mylroie (1995) and Chen et al. (1991); *Florida*: Stanley (1966), Hoffmeister et al. (1967), Hoffmeister and Multer (1968), Halley and Evans (1983), and Fruijtier et al. (2000); *Nantucket Island*: Oldale et al. (1982); *Nova Scotia*: Clarke et al. (1972) and Wehmiller et al. (1988); *Baffin Island and Coburg Island, Canada*: Miller et al. (1977, 1992), Miller (1985), Mode (1985), Nelson (1981), and Brigham (1983); *Greenland*: Funder et al. (1991, 1998), Tveranger et al. (1994), Hansen et al. (1994), Landvik et al. (1994), Mejdaal and Funder (1994) and Kelly et al. (1999).

Kennedy et al., 1982) in that they lack any extralimital species. On San Salvador Island in the Bahamas, the Cockburn Town reef has been well dated to the Last Interglacial period (Chen et al., 1991). The extensive molluscan fauna in this reef (Hagey and Mylroie, 1995) has species that all live in the region today. The same is true for molluscan faunas from low-elevation, emergent reefs on Curaçao, Grand Cayman Island, and Jamaica (Buissonjé, 1974; Cerridwen and Jones, 1991; Donovan and Littlewood, 1993), all of which have been dated to the Last Interglacial period (Schubert and Szabo, 1978; Szabo, 1979; Vézina et al., 1999). Thus, we interpret the faunas from the Caribbean to indicate water temperatures little different from those of today. In Florida, the Key Largo Limestone, dated to the Last Interglacial period (Fruijtier et al., 2000), also has a coral and mollusk fauna very similar to that found today in the Florida Keys, except for the absence of *Acropora palmata* (Stanley, 1966; Hoffmeister et al., 1967; Hoffmeister and Multer, 1968). However, in the Last-Interglacial Miami Limestone, an oolitic facies contemporary with the Key Largo Limestone, Halley and Evans (1983) report the gastropod *Cittarium pica* (also found in the Devonshire member of Bermuda) near Miami. Its presence in the Miami Limestone indicates that, during the Last Interglacial period, it was at least 150 km north of its northernmost, marginal modern range endpoint.

Oregon, California and Baja California have Last-Interglacial faunas that mostly show warmer than present conditions in eastern Pacific waters, a conclusion reached by Kennedy et al. (1982) based on an earlier compilation. There are many localities that have either reliable Last-Interglacial U-series ages on coral (Omura et al., 1979; Rockwell et al., 1989; Muhs et al., 1994) or amino acid ratios in mollusks similar to those in nearby U-series-dated deposits (Kennedy et al., 1982, 1992; Muhs et al., 1992). These confidently dated deposits show warm-water signatures (Woodring et al., 1946; Valentine, 1961; Kennedy, 1978, 1992; Lindberg et al., 1980; Emerson et al., 1981; Kennedy et al., 1982; Ashby et al., 1987; Rockwell et al., 1989; Libbey and Johnson, 1997). Peak Last-Interglacial localities from the Gulf of California and along the Pacific coast from southern Baja California Sur to central Oregon have faunas with a distinctly southern, warm-water aspect.

In Baja California Sur, hermatypic colonial corals such as *Porites* and *Pocillopora* today range only as far north as La Paz in the Gulf of California (Johnson and Libbey, 1997) and Bahía Magdalena on the Pacific Coast (Squires, 1959). However, corals that date to the Last Interglacial period are known as far north as Isla de Guadalupe off the Pacific Coast (Lindberg et al., 1980) and Mulegé and Punta Chivato (Ashby et al., 1987; Libbey and Johnson, 1997) in the Gulf of California (Fig. 13). This indicates a Last-Interglacial, northward

range extension of more than 300 km in the Gulf of California and more than 600 km on the Pacific Coast. Isla de Guadalupe's Last-Interglacial marine deposits also have at least ten extralimital southern species of mollusks (Lindberg et al., 1980). Six species from Last-Interglacial deposits on Isla de Guadalupe, *Purpura pausa*, *Thais planspira*, *Mitrella guttata*, *Mitra rupicola*, *Conus purpurascens*, and *Codakia distinguenda*, presently live no farther north than southern Baja California Sur and range as far south as Panama, Colombia, Ecuador or Peru (Keen, 1971). This indicates Last-Interglacial northward range extensions of 600–800 km. The Last Interglacial deposits on Isla de Guadalupe include at least one Indo-Pacific faunal element, the cowrie *Cypraea cernica* (Lindberg et al., 1980), which is known today at least as far east as Hawaii (Burgess, 1970; Kay, 1979), but is not otherwise known from the tropical eastern Pacific (Keen, 1971). If Hawaii is the closest modern record of *C. cernica*, its presence on Isla de Guadalupe represents a northeastern range extension of more than 4000 km. In northern Baja California near Ensenada, the third emergent marine terrace at Punta Banda has confidently dated solitary corals of Last-Interglacial age (Muhs et al., 1994). Extralimital southern mollusks such as the gastropods *Macron aethiops* and *Stramonita biserialis* in these deposits (Rockwell et al., 1989) do not range today any farther north than Isla Cedros in Baja California (Keen, 1971), indicating a northward range extension of more than 350 km.

Farther north along the Pacific Coast of North America, there are numerous localities with emergent marine deposits or deep-sea sediments from cores that show warm-water signatures during the Last Interglacial period. Off the coast of California, alkenone data from the portions of Santa Barbara Basin ODP core 893A and Point Conception ODP core 1016 that date to oxygen isotope substage 5e indicate surface water temperatures as much as 4°C warmer than present (Herbert et al., 1995; Yamamoto et al., 2000). This higher-than-modern paleotemperature estimate is consistent with a warm-water molluscan fauna from a Last-Interglacial marine terrace near Santa Barbara (Kennedy et al., 1992). Alkenone data from ODP core 1020 off northern California also indicate that sea surface temperatures were warmer than present during substage 5e (Yamamoto et al., 2000), which is consistent with a warm-water molluscan fauna from marine deposits on the coast of central Oregon (Kennedy, 1978; Kennedy et al., 1982). However, foraminifera data from ODP core 1020 indicate that offshore only the earliest part of substage 5e was warmer than present; cooler-than-present temperatures off northern California are inferred for the latter part of this substage (Poore et al., 2000).

The picture that emerges from Hawaii, Bermuda, and the compilation of published data for localities around

North America is that of a warmer-than-present Last Interglacial period. Northward movement of warmer waters likely played an important role in generating the faunas (Fig. 13). On Baffin Island, increased movement of warmer Atlantic water into the Labrador Sea and Baffin Bay might explain the warmer faunas there, as may also have been the case at ~80 ka (Miller et al., 1999). A stronger (warm) West Greenland Current could deliver relatively warmer Atlantic waters to both the northeastern coast of Baffin Island and the western coast of Greenland. Conversely, a weaker (cold) East Greenland Current could allow relatively warmer Atlantic waters to reach east Greenland. A stronger or warmer Gulf Stream may explain the warm faunas around Miami, on Bermuda, and perhaps Nantucket Island. On the west coast of North America, a weaker (cold) California Current with replacement by greater northward movement of warm, equatorial waters may explain the southern extralimital species in Baja California, California and Oregon. All modern corals in the Hawaiian Islands have their origins in the Indo-West Pacific region and the western boundary current system of the Pacific (Kuroshio and Subtropical Counter-Current) is the primary source for transport of coral larvae across the western part of the Pacific Ocean (Grigg, 1997). Thus, a stronger Kuroshio Current may also explain the presence of extralimital Indo-West Pacific mollusks around Oahu during the Last Interglacial period. Around Alaska, warmer waters from the North Pacific Ocean reached farther north and winter sea ice apparently did not extend south of Bering Strait during the Last Interglacial period (Brigham-Grette and Hopkins, 1995). In addition, the Last-Interglacial ostracode faunas from the Prudhoe Bay area indicate that more saline Atlantic water was present in that region and Arctic Ocean sea-ice cover must not have been perennial (Brigham-Grette and Hopkins, 1995).

The reconstruction of warmer-than-present Last-Interglacial oceans around North America agrees with observations of Last-Interglacial conditions in other regions, such as Japan, the Mediterranean basin and Western Australia. Marine terrace deposits dated to the Last Interglacial period fringe much of the coastline of the Japanese Islands (Ota and Omura, 1991). On the Noto Peninsula of central Honshu, marine terrace deposits contain solitary corals that have been dated by U-series methods to 115–125 ka, which corresponds to the peak of the Last Interglacial period (Omura, 1980). Last-Interglacial marine terrace deposits on the Noto Peninsula contain warm-water mollusks, including the subtropical species *Apollon natator*, *Trisidos kiyonoi*, and *Dendostrea paulucciae* (Matsuura, 1985). Matsuura (1985) interprets these to indicate that a warm current flowed into the Sea of Japan, including the embayment area of the Noto Peninsula, during the Last Interglacial period. Kitamura et al. (2000) have pointed out that a

major source of heat for the Sea of Japan is the warm Tsushima Current, a branch of the Kuroshio Current, and that warm-water faunas characterize interglacial periods when this current has a strong presence in the Sea of Japan.

It has long been recognized that low-elevation marine deposits around the Mediterranean Sea, assigned to the “Tyrrhenian II” sea-level high stand, contain a distinctive warm-water “Senegalese” fauna and could date to the Last Interglacial period (Butzer and Cuerda, 1962). Corals from localities in Mallorca, Sardinia, southern Italy and Tunisia have now been dated to the peak of the Last Interglacial period at around 120–130 ka by U-series methods (Hearty et al., 1986). Using amino acid ratios in mollusks, localities elsewhere in the region also have been correlated to this high stand of sea by the same workers. Many of the localities studied by Hearty et al. (1986) have the distinctive Senegalese molluscan fauna that contains the gastropods *Strombus bubonius* (as *S. latus*) and *Conus testudinarius*, and the bivalve *Cardita senegalensis*. These species presently live only off west Africa (Abbott and Dance, 2000). Thus, the presence of Senegalese mollusks in Mediterranean waters during the Last Interglacial period represents a northward range extension of over 2000 km.

In Western Australia, deposits of Last-Interglacial age have been studied along almost the entire coastline (Kendrick et al., 1991; Collins et al., 1993; Zhu et al., 1993; Johnson et al., 1995; Stirling et al., 1995, 1998; McCulloch and Esat, 2000). McCulloch and Esat (2000), using Sr/Ca values in the coral *Porites*, suggested that sea surface temperatures off Western Australia were not significantly different from modern values. Although reef-building corals presently live only as far south as ~24°S (Veron, 1995), Johnson et al. (1995) report an emergent, *Acropora*-dominated reef of Last-Interglacial age as far south as ~28.5°S, a southward range extension of more than 500 km (Johnson and Libbey, 1997). Faunas from Last-Interglacial deposits of Western Australia also contain extralimital northern (i.e., warmer) mollusks (Kendrick et al., 1991). For example, *Astrarium rhodostomum* is found in Last-Interglacial deposits that are ~400 km south of its present southern range limit. Thus, although the “core” tropical region of Western Australia may not have warmed appreciably during the Last Interglacial period (on the basis of the Sr/Ca values in corals), the subtropical region must have been considerably warmer than present.

7. Conclusions

New, higher-precision TIMS U-series ages of Wai-manalo Formation corals from slowly uplifting Oahu, Hawaii range from ~134 to ~113 ka, with most ages

between ~125 and ~115 ka. These ages are very similar to those reported earlier by Szabo et al. (1994). When combined with published U-series ages from nearby Lanai, which may be experiencing faster uplift, the data suggest a long Last Interglacial period that may have existed from ~136 to at least 115 ka. The results suggest that solar forcing may not have been the only control on ice sheet growth and decay, because summer insolation was at relatively low values at northern high latitudes at ~136 ka and at its lowest values of the past 200 ka at ~115 ka (Berger and Loutre, 1991). Thus, the Oahu data continue to challenge the Milankovitch theory of climate change. On Bermuda, corals that date to the Last Interglacial period are derived from patch reefs that likely were “catch-up” responses to sea level rise. As a consequence, coral colonization may have occurred later on Bermuda (~125–113 ka), but overlapped the time of coral growth on Oahu and Barbados. A logical next step in studies of the duration of the Last Interglacial period would be to date the Hawaiian and Bermuda corals using ^{231}Pa (Edwards et al., 1997), as a further check on the reliability of the ages presented here.

A large number of emergent marine localities with faunal data on Hawaii, Bermuda and around North America have now been dated to the Last Interglacial period. Thus, it is possible to generate a compilation of faunal records from which Last-Interglacial ocean temperatures can be inferred. Both Oahu and Bermuda have a number of extralimital southern species of mollusks, suggesting warmer-than-present waters during the Last Interglacial period. Warmer waters are also suggested for Last-Interglacial localities around North America, including Alaska, Greenland, Baffin Island, Nantucket Island, southern Florida, Baja California, California, and Oregon. Studies in Japan, the Mediterranean Basin and Western Australia also show evidence of warmer-than-modern Last-Interglacial waters. Thus, it is likely that during the Last Interglacial period, there was a greater movement of warm waters to higher latitudes than is the case today.

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