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REPORT

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The Holocene sea-level highstand in the equatorial Pacific: analysis of the insular paleosea-level database

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Abstract A review of the literature provides 92 estimates of the middle to late Holocene sea-level highstand on Pacific Islands. These data generally support geophysical model calculations that predict a +1 to 3 m relative sea-level highstand on oceanic islands due to the Earth's rheological response to the melting of the last continental ice sheets and subsequent redistribution of meltwater. Both predictions and observations indicate sea level was higher than present in the equatorial Pacific between 5000 and 1500 y B.P. A non-linear relationship exists between the age and elevation of the highstand peak, suggesting that different rates of isostatic adjustment may occur in the Pacific, with the highest rates of sea-level fall following the highstand near the equator. It is important to resolve detailed sea-level histories from insular sites to test and refine models of climatic, oceanographic, and geophysical processes including hydroisostasy, equatorial ocean siphoning, and lithospheric flexure that are invoked as mechanisms affecting relative sea-level position. We use a select subset of the available database meeting specific criteria to examine model relationships of paleosea-surface topography. This new evaluated database of paleosea-level positions is also validated for testing and constraining geophysical model predictions of past and present sea-level variations.

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Introduction

More than 65% of Earth's human population inhabit low coastal regions and islands. It is important, for the sake of these societies, that research continue to improve understanding of the rate and magnitude of sea-level movements past, present, and future (Houghton et al. 1996). The geologic record of relative sea level as recorded in fossil shorelines provides data for improving our understanding of the natural variability of the sea surface. Important periods of time that serve as models of higher sea level include the last interglacial and the middle Holocene. There is general agreement that global sea level was above present during the last interglacial. However, because of regional variation in the geoid and crustal motion, it is not widely understood that the equatorial Pacific also experienced a relative sea-level highstand above present in the middle to late Holocene. This history has important implications for interpreting the evolution of reefs (Davies and Montaggioni 1985), atolls (McLean and Woodroffe 1994), and atoll islands (Richmond 1992), as well as prehistorical settlement patterns within many low-lying coastal plains of the Pacific (Dickinson et al. 1994). Here we review the literature of Holocene sea level in the Pacific Ocean and use specific criteria to compile a select set of those data that are validated as paleosea-level proxies for better understanding the character of the postglacial sea-level maxima.

Background

Postglacial and Holocene sea level

Detailed reviews of postglacial sea-level movements and their relationship to climate and solid Earth processes are available in Devoy (1987), Kidson (1986), and van de Plassche (1986). The concept of a postglacial

(Holocene) sea level higher than present was introduced by Daly (1920, 1934) in his study of emerged coral platforms in the Indian and Pacific Oceans and is supported by observations of raised shorelines worldwide (e.g., Wentworth and Palmer 1925; Stearns 1935).

Recognizing that regional differences in sea levels are manifested by changes in the geoid (Mörner 1976; Bloom 1977) and in response to climatic, dynamic isostatic, and tectonic processes, sea-level researchers today focus on reconstructing relative sea-level histories with regional validity. Fundamental to this is the recognition that no part of Earth's crust can be assumed to be entirely stable and that regional differences in Earth's surface response to geoid variations preclude the use of sea-level curves outside their region of validity. The oceanic geoid ('equipotential sea surface') is a highly uneven surface that varies by as much as 180 m with respect to Earth's center (Mörner 1976). Because of these gravitational variations, mean sea level (msl) in the world's oceans more closely resembles the pitted surface of a golf ball rather than a smooth sphere.

Climatic and geophysical processes

The dominant forces that influence Holocene sea-level movements are: (1) climatic and oceanographic variation; (2) glacio- and hydroiostatic redistribution of Earth's mass in response to ice-sheet advance and retreat; (3) paleogravitational variation in the geoid; and (4) changes in the morphology of ocean basins and margins due to tectonism.

Climatic and oceanographic variation

Climate change and its impact on terrestrial ice reserves has played the greatest role in affecting early Holocene and postglacial sea levels, bringing about the deterioration of the last great ice sheets and supplying melt water to the ocean basins. Reef-accretion records from Barbados (Fairbanks 1989) and New Guinea (Shackleton 1987) indicate that as global mean temperature warmed ~ 4–$5°C$ and climate shifted toward the present interglacial condition, global sea level rose ~ 110–120 m (at times episodically) toward its present position. Periods of reduced rates of rise resulting from short-term shifts to cooler climatic conditions (Younger Dryas) and reduced glacial wastage are separated by episodes of rapid rise associated with accelerated glacial melting or ice sheet instability (MacAyeal 1993; Blanchon and Shaw 1995).

Meltwater pulses may drive episodic sea-level events, but the exact timing, magnitude, and number are still debated (Bard et al. 1996; Clark 1995; Montaggioni et al. 1997). Nevertheless, submerged paleoshorelines (Anderson and Thomas 1991), marine planation terraces, and submerged intertidal notches (Blanchon and Jones 1994; Blanchon and Shaw 1995; Fletcher and Sherman 1995; Locker et al. 1996) presumably the result of sudden drowning, are reported by workers. Additionally, seasonal sea-level variation associated with historic El Niño events (Jacobs et al. 1994) certainly occurred throughout the middle and late Holocene; in some low-lying regions possibly influencing the fossil record of past sea-level movements.

Variation in Earth's mean global temperature of the order of ~ 1–$2°C$ during the middle and late Holocene (Folland et al. 1990), may have influenced sea-level position. Thermal expansion of the upper water column associated with warming of 0.6–1.0°C could raise sea level 4–8 cm according to Wigley and Rapier (1987). Increased rates of ice accumulation during the late Holocene leading to Antarctic ice-volume increases may account for $\sim 1 \pm 0.2$ m sea-level lowering (Goodwin 1998). This would require a revision of the glacio-eustatic histories prescribed by current geophysical models.

Crustal response to deglaciation and meltwater load

It has been shown by Chappell (1974) that islands far from glaciated regions experience a hydroisostatic adjustment to the meltwater mass added to the ocean basins following deglaciation. The increased load on ocean floors induces mantle flow below the sea floor towards an island that will lift it relative to sea level, leaving a shoreline stranded above present day msl.

Combining a model of Earth's rheology with a deglaciation history, Nakada (1988) and Nakada and Lambeck (1989) predicted emerged Holocene shorelines on islands larger than 10–50 km in radius far from glaciated regions. Their ARC3 + ANT3a ice history assumed that the main deglaciation of the Northern Hemisphere and Antarctic ice sheets ended about 6000 y ago and that a portion of Antarctic ice continued melting thereafter.

Bumps and depressions in the sea surface modeled by early workers (Walcott 1972; Chappell 1974; Farrell and Clark 1976) have been surveyed and are believed to migrate over space (10^3 km) and time (10^{3-4} y), affecting the relative spatial and temporal position of sea level. The present geodetic sea level or 'equipotential surface of the geoid' varies with respect to Earth's center by as much as 180 m and is not stable, but must have changed with variation in gravity and other factors (Mörner 1976; Kidson 1986). Although also related to crustal movements of tectonic or isostatic origin, the oceanic geoid coincides with the sea surface and is hence distinct from instability of the ocean floor. Mörner (1976) argued that recognition of geoidal variation meant that sea-level changes could no longer be taken exclusively as evidence of glacial ice-volume changes.

Early model predictions for a higher than present middle Holocene shoreline 5000 y ago were made along longitude 165°W between 70°S and 45°N (Clark et al. 1978) and were generally supported by geologic observations (Nunn 1994). The "equatorial ocean siphoning" models ICE-3G (Mitrovica and Peltier 1991) and ICE-4G (Peltier 1994, 1996) proposed that middle to late Holocene shoreline emergence was due to the migration of the mantle away from the central equational ocean basin interiors toward regions that experienced maximum glaciation (Fig. 1). During a glacial period, the weight of continental ice sheets causes the downward deformation of the crust, forcing sublithospheric flow toward the equator, and creating lithospheric forebulges adjacent to the ice sheet margins. A gravitational anomaly develops in the low latitudes and creates an associated high in the oceanic geoid. With deglaciation, continents rebound viscoelastically, causing the gravity anomaly to decay and the swollen equatorial ocean geoid to migrate toward regions of glacioisostatic rebound and collapsing forebulges at higher latitudes. Using a deglaciation history that assumes Northern Hemisphere ice ceased melting by 8000 y ago while Antarctic ice-melt continued until 4000 y ago, ICE-3G predicted relative sea-level highstands that approximately matched observed histories of 15 small islands in the Pacific Ocean (Mitrovica and Peltier 1991).

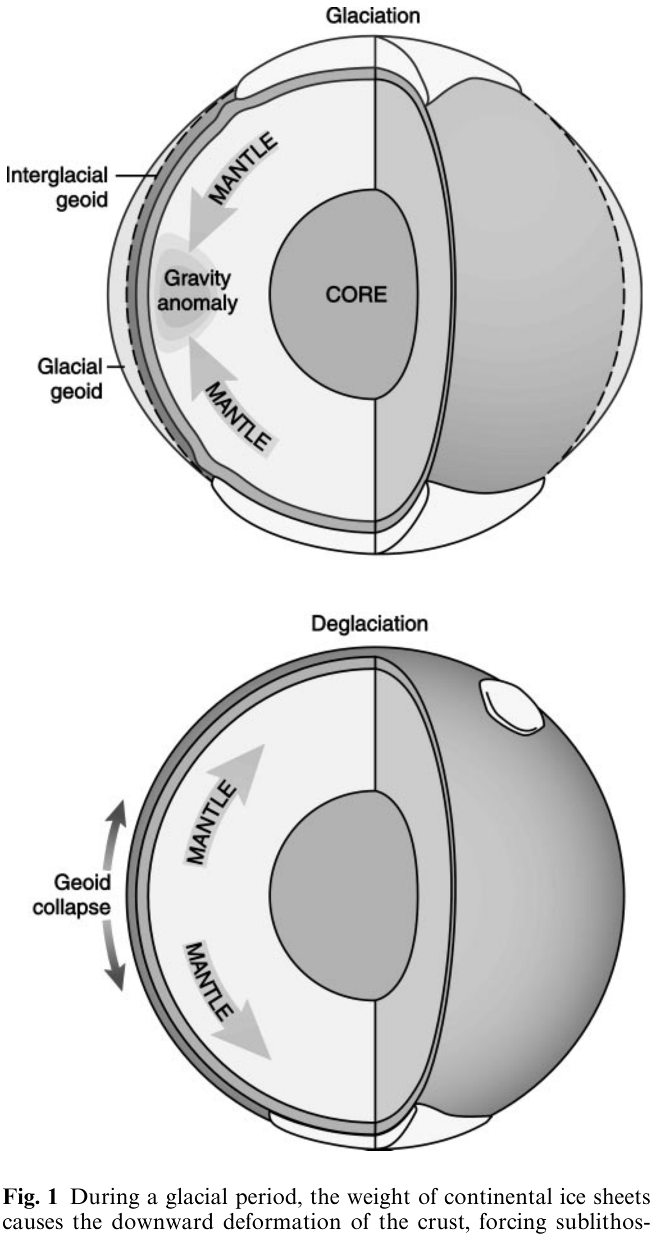

Fig. 1 During a glacial period, the weight of continental ice sheets causes the downward deformation of the crust, forcing sublithospheric flow toward the central equatorial ocean basins. A gravitational anomaly develops in the low latitudes creating an associated high in the oceanic geoid. At the end of an ice age, continents rebound viscoelastically, causing the gravity anomaly to decay and the oceanic geoid to migrate from lower to higher latitudes and toward the continents. This process, called equatorial ocean siphoning, may have caused a fall in late Holocene relative sea level among low-latitude oceanic islands

Tectonism and lithospheric flexure

On long time scales and at slow rates, sea floor spreading driven by plate tectonics alters the shape and storage volume of the world's oceans. The contribution of these ocean basin volume changes to Holocene sea-level movements is assumed to be small. Local-scale tectonic events, however, can be significant to relative sea level determinations (Chappell and Pollach 1991). In some instances, coseismic events have been disaggregated to resolve a detailed sea-level history (Kayanne et al. 1993).

Vertical displacement of shorelines due to lithospheric flexure in the Pacific has been observed in Hawaii (Muhs and Szabo 1994) and Tahiti (Pirazzoli and Montaggioni 1988), where the horizontal movement of the Pacific Plate carries these islands across active volcanic hotspots. Loading of lithosphere at the Hawaiian hotspot (Fig. 2) creates an inner zone of subsidence proximal to the loading point, and a distal arch that circumscribes the region of greatest subsidence (Watts and ten Brink 1989). Uplift, followed by subsidence, is associated with the movement of an island across the arch. On Oahu, Hawaii, it has been estimated that uplift due to lithospheric flexure during the past 125 000 y ranged between 0.02–0.06 mm/y Jones 1993; Muhs and Szabo 1994). Similar rates have been found associated with hotspot loading near Mehetia in the Society Islands (Pirazzoli and Montaggioni 1988).

Materials and methods

The Holocene sea-level database

We review 92 published estimates of Pacific Holocene sea-level movements and use specific criteria to cull a validated set of the

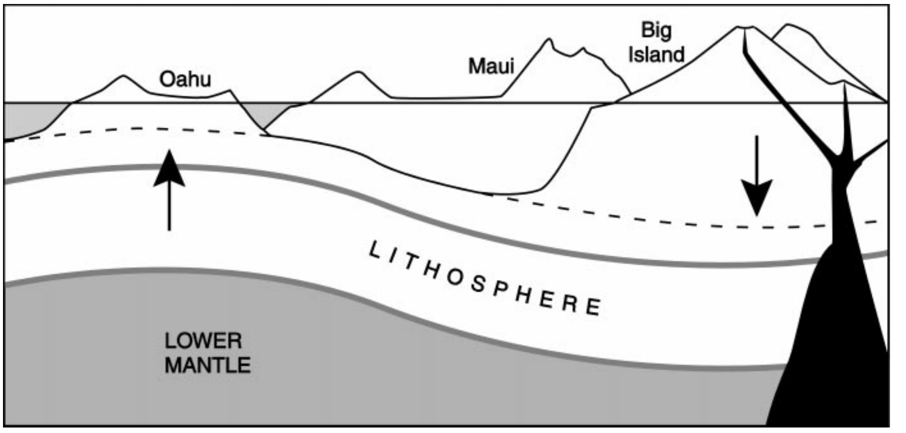

Fig. 2 In Hawaii, lithospheric flexure under the massive volcanic accumulations at the hotspot causes compensatory arching at a radius of approximately 400 km (dependent upon lithospheric rheology), resulting in uplift of Oahu

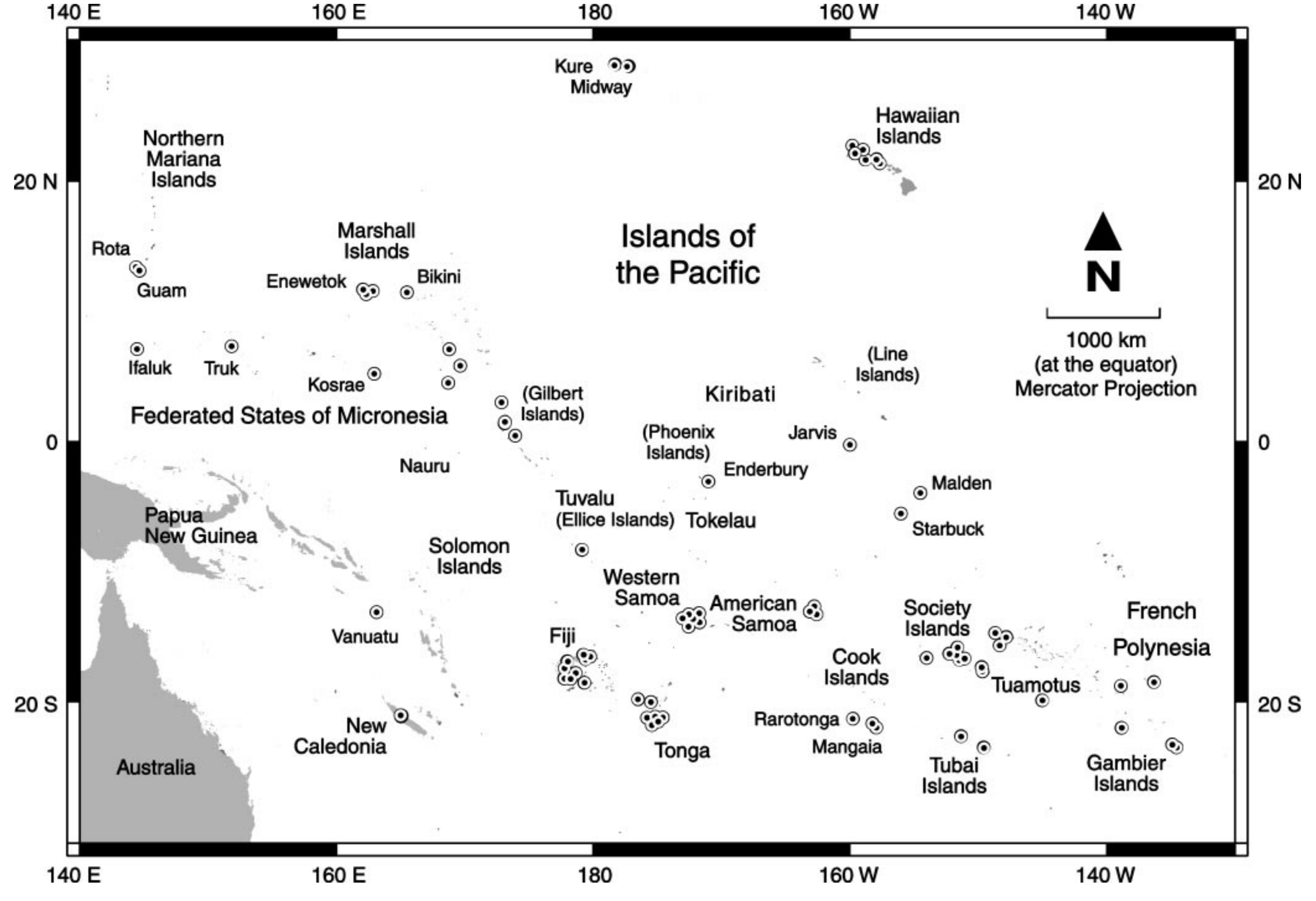

Fig. 3 The Pacific Ocean and its islands. Locations with published evidence of a higher than present middle to late Holocene relative sea-level highstand

paleosea-level proxy data for testing models of the postglacial sea-level maxima. The data are derived from numerous islands (Fig. 3) and are compiled in Table 1 by geographical region, counterclockwise from the Marianas to the Hawaiian Islands. This review intends to regionally update the *Atlas of Holocene Sea-Level Changes* (Pirazzoli 1991) and provide researchers with a comprehensive database with which to explore middle to late Holocene sea level in the Pacific. In addition, Table 1 reports site locations (latitude and longitude), island size (radius in km), estimates of the maximum paleosea-level elevation or maximum shoreline feature height (m above present msl), the corresponding radiometric (^{14}C) age of paleosea level or shoreline feature sampled (^{14}C y BP 1950), the earliest age of reconstructed or observed emergence, and its proxy.

Table 1 Elevations and ages of middle to late Holocene sea-level estimates and shoreline features on Pacific Islands

Location	Latitude	Longitude	Island size (km)	Maximum elevation (m)	(1σ)	Apex age (1000 ^{14}C y BP)	(1σ)	Age range (1000 ^{14}C y BP)	Limit	PSP	Source	Reference
Mariana Islands												
Rota	14 10 N	145 10 E	6	0.00	NA	4.20	0.10	6.0–4.2	min/msl	1	N, Reef	Kayanne et al. (1993)
Guam	13 27 N	144 40 E	15	0.00	NA	4.20	0.10	6.0–4.2	min/msl	1	N, Reef	Kayanne et al. (1993)
Guam	13 17 N	144 41 E	15	0.60	NA	2.88	0.11	NA	min	2	C	Curray et al. (1970)
Guam	13 20 N	144 38 E	15	1.60	0.20	3.40	0.25	NA	min	2	M	Tracey et al. (1964)
Guam	13 20 N	144 38 E	15	1.80	NA	3.60	0.25	5.12–2.88	min	3	Reef, C, M	Easton et al. (1978)
Caroline Islands												
Ifaluk[b]	7 15 N	144 27 E	3	1.00	NA	4.00	0.25	NA	min	2	MA, Reef	Tracey (1968)
Truk	7 27 N	151 51 E	25	0.00	NA	0.00	0.27	6.5–1	min	3	Peat	Bloom (1970)
Truk	7 27 N	151 51 E	25	0.00	NA	0.00	NA	6.5–0	min	3	C, Peat	Matsumoto et al. (1986)
Truk	7 27 N	151 51 E	25	0.30	NA	1.27	0.10	NA	min	3	C	Curray et al. (1970)
Ponape	65 1.6 N	158 18.2 E	8	0.00	NA	0.00	0.27	6.5–1	min	3	Peat	Bloom (1970)
Ponape	65 1.6 N	158 18.2 E	8	0.00	NA	0.00	NA	6.5–0	min	3	C, Peat	Matsumoto et al. (1986)
Kosrae	5 17 N	162 58.7 E	7	0.00	NA	0.00	0.27	6.5–1	min	3	Peat	Bloom (1970)
Kosrae[b]	5 17 N	162 58 E	7	1.10	NA	3.28	0.07	3.28–1.29	min	2	C, Reef	Athens (1995)
Marshall Islands												
Enewetok[a]	11 30 N	162 20 E	22	1.50	NA	2.75	0.15	4–1	min	1	Reef	Buddemeier et al. (1975)
Enewetok	11 30 N	162 20 E	22	1.50	NA	2.75	0.18	4.0–0.75	min	1	Reef	Tracey and Ladd (1974)
Bikini[a]	11 36 N	165 30 E	12	1.50	NA	2.80	0.30	4.0–0.75	min	1	Reef	Tracey and Ladd (1974)
Jaluit	5 54.5 N	169 39 E	22	0.50	NA	2.29	0.10	NA	min/msl	3	C	Curray et al. (1970)
Ailinglaplap	7 16.6 N	168 48 E	20	0.70	NA	2.79	0.10	NA	min/msl	3	C,M	Curray et al. (1970)
Ebon	4 35 N	168 43 E	8	0.75	NA	2.78	0.10	2.92–2.58	min/msl	3	C,M	Curray et al. (1970)
Kiribati (Gilbert Islands)												
Butaritari[a]	3 05 N	172 50 E	20	2.40	NA	2.68	0.09	2.87–2.11	min	1	C,M	Schofield (1977)
Tarawa[a]	1 21 N	173 07 E	15	2.40	NA	2.68	0.07	3.87–2.11	min	1	C, M	Schofield (1977)
Tarawa	1 30 N	173 00 E	15	0.00	NA	NA	NA	8.0–5.6	min	1	Reef growth	Marshall and Jacobson (1985)
Abemama[a]	0 25 N	173 55 E	14	2.40	NA	2.68	0.06	3.87–2.11	min	1	C, M	Schofield (1977)
Tabiteuea[a]	1.25 S	173 07 E	25	2.40	NA	2.68	0.07	3.87–2.11	min	1	C, M	Schofield (1977)
Kiribati (Phoenix Islands)												
Enderbury[b]	3 08 S	171 00 W	8	2.00	NA	4.89	0.10	4.89–2.17	min	2	Reef	Tracey (1972)
Tuvalu (Ellice Islands)												
Funafuti[a]	8 31 S	179 12 E	11	2.40	NA	2.68	0.07	3.87–2.11	min	1	M, Reef	Schofield (1977)
New Caledonia												
New Caledonia	21 00 S	165 00 E	60	1.00	NA	5.50	NA	7.3–0	min	1	Reef growth	Cabioch et al. (1995)
New Caledonia	21 00 S	165 00 E	60	>1	NA	3.70	0.10	4.4–3	min	1	Reef growth	Coudray and Delibrias (1972)
Fiji												
Vanua Levu	16 33 S	179 20 E	37	2.00	0.50	2.65	0.20	5–1	msl	1	FB	Nunn (1995)
Vanua Levu	16 45 S	179 20 E	37	1.50	0.50	N/A	NA	NA	min/msl	4	N, T	Berryman (1979)
Vanua Levu	16 45 S	179 20 E	37	1.50	0.50	N/A	NA	NA	min	4	Reef	Rodda (1986)
Vanua Levu[a]	16 45 S	179 25 E	37	2.00	0.50	4.00	0.20	6–3.4	min/msl	1	MA, N	Miyata et al. (1990)
Viti Levu	18 10 S	178 15 E	50	1.30	0.30	3.50	NA	NA	msl	2	Review	Nunn (1990b)
Viti Levu	18 10 S	178 15 E	50	0.60	0.10	4.00	0.07	NA	msl/min	3	R, P	Shepard (1998)
Viti Levu	17 25 S	177 45 E	50	0.45	0.35	5.30	0.08	5.8–1.4	min	2	C	Ash (1987)
Viti Levu	18 10 S	178 15 E	50	1.00	NA	4.00	0.36	7.6–1.6	min	3	ROB	Sugimura et al. (1988b)
Wailevu	16 40 S	179 48 E	37	1.00	NA	4.00	0.08	NA	min	2	C	Roy (1988)
Naroi	18 30 S	179 00 E	10	2.00	NA	4.00	NA	NA	min	3	M	Nunn (1988)
Yanuca	18 20 S	178 00 E	2	2.00	NA	4.00	NA	NA	msl	3	Ch	Green (1979)

Table 1 (continued)

Location	Latitude	Longitude	Island size (km)	Maximum elevation (m)	(1σ)	Apex age (1000 ^{14}C y BP)	(1σ)	Age range (1000 ^{14}C y BP)	Limit	PSP	Source	Reference
Tonga												
Tongatapu	21 10 S	175 12 W	13	1.60	NA	3.50	NA	4.5–1.5	msl	1	Review	Nunn (1995)
Ha'apai Group	19 47 S	173 30 W	2	1.50	NA	3.25	0.07	NA	msl/mhw	2	Ch, Pro	Dickinson et al. (1994)
Eua	21 22 S	174 58 W	10	1.65	NA	5.90	0.20	NA	min	3	C, Reef, T	Taylor and Bloom (1977)
Tongatapu	21 10 S	175 12 W	13	2.20	NA	5.90	0.20	NA	min	3	C, Reef, T	Taylor (1978)
Tongatapu	21 10 S	175 12 W	13	1.00	NA	3.09	NA	NA	msl	3	A	Davidson (1979)
Tongatapu	21 10 S	175 12 W	13	1.00	NA	3.13	NA	NA	msl	3	A	Poulsen (1967)
Western Samoa												
Savai'i	13 28 S	172 30 W	30	1.50	NA	1.30	NA	4–0	msl	1	Review	Nunn (1991)
Savai'i	13 28 S	172 45 W	30	2.30	0.50	NA	N/A	N/A	msl	4	BR	Sugimura et al. (1988a)
Savai'i	13 47 S	172 23 W	30	0.92	N/A	1.85	0.33	N/A	msl	3	C, S	Grant-Taylor and Rafter (1962)
Upolu	13 55 S	171 58.5 W	25	0.95	N/A	N/A	N/A	N/A	msl	4	BR	Sugimura et al. (1988a)
Upolu	13 56 S	171 58 W	25	1.50	N/A	1.52	0.33	N/A	msl	3	C, S	Grant-Taylor and Rafter (1962)
Upolu	13 56 S	171 58 W	25	> 1	N/A	2.30	NA	NA	msl	2	Notch seds	Rodda (1988)
America Samoa												
Manu'a	14 13 S	169 30 W	4	1.50	N/A	N/A	N/A	N/A	min	4	B, BR	Stice and McCoy (1968)
Niue												
Niue	19 02 S	169 56 W	12	1.50	N/A	N/A	N/A	N/A	min	4	Reef	Schofield (1959)
Cook Islands												
Mangaia	21 54 S	157 58 W	4	1.10	NA	5.50	NA	6.5–4.5	min	2	Cp, P, Po	Ellison (1994)
Mangaia[a]	21 54 S	157 58 W	4	1.70	NA	3.40	0.26	5–3.15	min	1	Reef growth	Yonekura et al. (1988)
Rarotonga[b]	21 14 S	159 47 W	5	1.00	NA	2.03	0.06	NA	min	2	Reef	Stoddart (1972)
Suwarrow	13 14 S	163 05 W	10	0.50	NA	3.00	NA	NA	min	3	Reef	Scoffin et al. (1985)
Manu	13 15 S	163 09 W	10	0.50	NA	4.65	NA	NA	min	3	Reef	Scoffin et al. (1985)
New	13 20 S	163 05 W	10	0.50	NA	2.40	NA	NA	min	3	C	Scoffin et al. (1995)
One Tree	13 13 S	163 07 W	10	0.50	NA	1.70	NA	NA	min	3	MA	Scoffin et al. (1985)
Kiribati (Line Islands)												
Jarvis[b]	0 15 S	160 00 W	3	1.50	NA	3.98	0.25	3.98–1.8	min	2	Reef	Tracey (1972)
Starbuck[b]	5 37 S	156 00 W	4	1.50	NA	2.40	NA	2.94–1.23	min	2	Reef	Tracey (1972)
Malden[b]	4 03 S	154 30 W	6	1.00	NA	3.55	NA	NA	min	2	Reef	Tracey (1972)
French Polynesia												
Society Islands												
Mauipiti	16 25 S	152 15.5 W	4	0.73	0.23	3.25	0.06	5–1.25	min	1	BR, C	Pirazzoli and Montaggioni (1988)
Bora Bora	16 26.5 S	151 44.5 W	5	0.60	0.13	3.39	0.07	5–1.25	min	1	BR, C	Pirazzoli and Montaggioni (1988)
Bora Bora	16 26.5 S	151 44.5 W	5	1.00	0.10	2.25	0.07	5–1.25	min	1	Reef	Guilcher et al. (1969)
Mopelia	16 45 S	154 00 W	4	0.80	0.10	3.45	0.13	5–1.25	min	1	Reef	Guilcher et al. (1969)
Raiatea	16 44.5 S	151 25.5 W	25	0.63	0.28	3.65	0.08	5–1.25	min	1	BR, C	Pirazzoli and Montaggioni (1988)
Huahine	16 43 S	151 00 W	6	0.60	0.25	3.75	0.08	5–1.25	min	1	BR, C	Pirazzoli and Montaggioni (1988)
Moorea	17 30 S	149 46 W	6	0.55	0.35	4.10	0.13	5–1.25	min	1	BR	Pirazzoli and Montaggioni (1998)
Tahiti	17 40 S	149 37 W	16	0.35	0.10	3.10	0.06	5–1.25	min	1	BR, Reef	Pirazzoli and Montaggioni (1988)
Tuamotu Islands												
Mataiva	14 53 S	148 39 W	6	0.48	0.28	3.80	0.13	5–1.25	min	1	BR, C	Pirazzoli et al. (1985)
Rangiroa	15 13 S	147 48 W	22	0.83	0.18	2.63	0.10	5–1.25	min	1	C, M	Pirazzoli et al. (1985)
Makatea[a]	15 49 S	148 16 W	5	0.76	0.15	4.44	0.13	5–1.25	min	1	C, N	Pirazzoli et al. (1985)

Table 1 (continued)

Location	Latitude	Longitude	Island size (km)	Maximum elevation (m)	(1σ)	Apex age (1000 ^{14}C y BP)	(1σ)	Age range (1000 ^{14}C y BP)	Limit	PSP	Source	Reference
Vahitahi	18 46 S	138 50 W	7	0.68	0.18	2.75	0.11	5–1.25	min	1	BR, C, Reef	Pirazzoli et al. (1988)
Reao[a]	18 30 S	136 20 W	10	0.85	0.23	4.25	0.10	5–1.25	min	1	C, Al	Pirazzoli et al. (1988)
Hereheretue	19 51.5 S	145 00.5 W	8	1.00	0.10	3.20	0.70	5–1.25	min	1	BR, C, Reef	Pirazzoli et al. (1988)
Mururoa[a]	21 49 S	138 47 W	10	0.94	0.00	3.32	0.20	5–1.25	min	1	C	Pirazzoli et al. (1988)
Gambier Islands												
Gambier	23 06 S	134 52 W	13	0.90	0.30	1.95	0.07	5–1.25	min	1	BR, C	Pirazzoli (1987)
Temoe	23 20 S	134 29 W	8	0.55	0.15	2.60	0.27	5–1.25	min	1	BR, C	Pirazzoli (1987)
Tubai (Austral) Islands												
Rurutu	22 30 S	151 20 W	8	1.15	0.55	1.80	0.10	5–1.25	min	1	C, N	Pirazzoli (1987)
Tubuai	23 21.5 S	149 32 W	9	0.38	0.28	1.90	0.06	5–1.25	min	1	BR, C	Pirazzoli (1987)
Hawaiian Islands												
Kauai[b]	22 13 N	159 30 W	20	1.80	0.30	3.70	0.08	4.2–3.2	min	2	C	Jones (1992)
Kauai	22 13 N	159 30 W	20	1.75	0.25	3.50	0.10	3.1–2.4	max	2	S	Calhoun and Fletcher (1996)
Kauai	22 06 N	159 18 W	20	1.30	NA	3.90	0.22	4–3.8	min	2	M, ROB	Matsumoto et al. (1988)
Oahu	21 16.5 N	157 42 W	25	0.00	NA	3.50	0.22	7–0.5	min	1	Reef growth	Easton and Olson (1976)
Oahu	21 16.5 N	157 42 W	25	1.80	0.30	3.49	0.16	NA	msl/hw	3	Ca	Stearns (1974)
Oahu	21 25 N	157 45 W	25	NA	NA	3.10	0.10	4.0–2.2	NA	4	S	Athens and Ward (1991)
Oahu	21 28 N	158 48 W	25	1.50	0.45	3.80	0.12	5–2	msl	2	C, M, S	Fletcher and Jones (1996)
Oahu[a]	21 28 N	158 48 W	25	2.00	0.35	3.70	0.11	5–2	msl	1	FB	Grossman and Fletcher (1998)
Midway[b]	28 13 N	177 22 W	15	1.50	NA	2.42	0.30	2.4–1.28	min	2	Reef	Ladd et al. (1970)
Kure[b]	28 25 N	178 20 W	12	1.50	0.50	1.48	0.25	NA	min	2	Reef	Gross et al. (1969)

[a] Primary dataset
[b] Secondary dataset
Island size is island/atoll radius. Maximum elevation is in m above present msl. Apex age is the age contemporaneous with the maximum elevation of paleosea-level estimate or shoreline feature. Age range represents ages of samples studied. Limit signifies sea-level proxy function as an upper, mid, or lower limit to sea level. PSP is ranking scheme for proxies of paleosea-level position (see text for details). Sources include: (A) archaeological site; (Al) algal crust; (B) bench; (BR) beachrock; (C) coral in situ; (Ca) carbonate, (CC) coral clast; (Ch) charcoal; (FB) fossil beach; (M) mollusc; (MA) microatoll; (N) notch; (P) peat; (Po) pollen; (Pro) prograded beach; (R) beach ridges; (Reef) reef, reef flat, reef platform; (Reef growth) Reef growth or accretion history; (Review) review of other work; (ROB) regressive overlap boundary; (S) marine sand; (T) coastal terrace, coastal plain.

Age ranges representing source material are given for studies based on multiple sample dates. Measurement errors are given when provided by the original author(s).

Mariana Islands

Tracey et al. (1964) dated a *Tridacna* shell (3400 ^{14}C y BP) at an elevation of 1.6 $\pm$ 0.2 m in growth position imbedded in a coral head whose top was truncated. Because truncation occurs today at mean lower low water (mllw) they proposed that sea level at this time was 1.4–1.8 m above present. Curray et al. (1970) concluded that a coral head "apparently in situ" at an elevation of 0.6 m with a radiocarbon age of 2880 ^{14}C y BP indicated uplift "because of numerous other reef and terrace levels."

Kayanne et al. (1993) provided the only sea-level curve for the Mariana Islands, which shows a 1.8 m rise of the sea between 6000 and 4200 ^{14}C y BP. Using 54 radiocarbon dates of surface and cored samples from emergent and modern reefs on Rota and Guam, they described changes in vertical and lateral reef accretion and a maximum height of the transgression apparently assuming "keep-up" reef conditions. Their sea-level maximum during this period does not exceed present msl because they attributed the observed emergence entirely to tectonic uplift to account for elevation differences between their sea-level reconstructions on Rota and Guam and to changes they see in reef accretion and progradation. Subtracting the influence of uplift events, Kayanne et al. (1993) concluded that the hydroisostatic models of Nakada (1988) and Nakada and Lambeck (1989) best match their reconstructed sea-level history of the Marianas. They also acknowledged that a portion of the observed shoreline emergence may correspond to the highstand predicted by the 'equatorial ocean siphoning' model of Mitrovica and Peltier (1991), because their calibration depends on how much tectonic uplift they subtract from their observations.

Caroline Islands

From cores in tidal swamps on Truk, Ponape, and Kosrae (eastern Caroline Islands), Bloom (1970) found a history of shoreline progradation associated with decelerating submergence. Curray et al. (1970) and Matsumoto et al. (1986) reported inconclusive evidence of a higher than present Holocene sea level in the Caroline Islands from studies of reefs and swampy coastal plains, respectively. Curray et al. (1970) attributed the platforms observed on Truk, Ponape and Kosrae to lithified storm deposits formed after 3000 ^{14}C y ago and not indicative of paleosea-level position. This view was shared by Shepard et al. (1967) and Newell and Bloom (1970). Curray et al. (1970) also reported one coral head sample on Truk Island with a radiocarbon age of 1270 $\pm$ 95 ^{14}C y B.P. standing at 0.3 m amidst the rubble platform, which "could be storm rubble or could indicate higher sea level". In a recent study, Athens (1995) reported an *in situ* homogeneous reef platform on Kosrae in addition to storm rubble deposits. He suggested that the storm rubble platforms deposited after 3000 ^{14}C y BP "preserved the highstand reef formation at this location". Athens (1995) used archaeological evidence and two radiocarbon dates from the emerged reef flat to conclude that sea level stood 0.5–1.1 m above present between 2800–800 ^{14}C y BP on Kosrae.

Marshall Islands

In the Marshall Islands, Buddemeier et al. (1975) found evidence of a sea-level highstand on Enewetok Atoll from analysis of reef flat pavements, reflecting rapid carbonate deposition up until 2000 ^{14}C y BP, followed by erosion. Assuming a continuous subsidence rate of 0.02–0.04 mm/y (Menard 1964) for Enewetok during the late Holocene, Buddemeir et al. (1975) reconstructed sea level "significantly more than 1 m above present" between 3500 and 2000 ^{14}C y BP. Their results are supported by Tracey and Ladd (1974) who found in situ corals near Aranit and Biijiri Islands (Enewetok Atoll) at 1 m dating 3300–1900 ^{14}C y BP and truncated microatolls between 0.5–1.0 m above mllw on Bikini Island.

Curray et al. (1970) also found emerged corals and mollusc shells ranging from 0–0.75 m and dating 2780–2660 ^{14}C y BP on Ailinglaplap and 2920–2580 ^{14}C y BP on Ebon, but concluded that these samples were from storm rubble platforms that do not indicate evidence for a highstand. Their eustatic sea-level curve proposed sea level rose at a rate in excess of 10 mm/y between 10 000 and 6000 y BP., and then decreased to an average of $\pm$ 1 mm/y after 4000 y BP. Buddemeier et al. (1975) argued against this, stating that under a stabilization of sea level following a rapid rise, a progression of communities, environments, and net depositional rates would be recorded in the reef sediments. The environmental and biological discontinuities actually observed instead indicate a shift from a net depositional environment to a net erosional environment sometime after about 3000 ^{14}C y BP, which Buddemeier et al. (1975) attributed to either a fall in sea level, a change in wave regime, or both.

Kiribati (Gilbert and Phoenix Islands)

Schofield (1977) used radiocarbon dates of corals and *Tridacna* shells sampled from emerged and often eroded reefs on Abemama, Onotoa, and Tabiteuea atolls to suggest that sea level stood between 1.6–2.4 m above present levels in the Gilbert and Phoenix Islands between 3000 and 1000 ^{14}C y BP. Schofield's (1977) dates are based on the actual ^{14}C half-life (5730 y) instead of the Libby half-life (5568 y). His chronology has been corrected here for comparison by dividing the age by 1.029 (5730 y/5568 y) as suggested by Stuiver and Reimer (1993).

On Tarawa Atoll in the Gilbert Islands, Marshall and Jacobson (1985) used cores drilled into the reef on Tarawa to depict a minimum sea-level history significantly above the history inferred by Bloom (1970). Their curve reached its present height by 5500 ^{14}C y ago, much earlier than previously suggested. Only one core showed evidence of reef rock above present sea level, and it could not be shown to be in situ. On Enderbury Atoll in the Phoenix Islands, Tracey (1972) found evidence of a sea-level highstand that peaked at + 2 m from dates of emerged reef rock ranging 4890–2170 ^{14}C y BP.

Tuvalu (Ellice Islands)

Schofield (1977) reported emerged corals in growth position and emerged and eroded reef platforms on Funafuti Atoll in Tuvalu. While some of the exposed reef platforms are storm deposited boulders, others were believed to be *in situ*. In addition, in situ reef rock was found stranded inside a dammed intertidal mangrove swamp. It was thought that the reef grew in what was a back lagoon that received free exchange of water and nutrients by tidal changes under a 1.65 m sea-level highstand. At ca. 1570 ^{14}C y BP as sea level fell, the lagoon was cut off from the sea by the formation of a storm-built rampart. He reconstructed sea level about + 2.4 m in Tuvalu about 2680 ^{14}C y BP.

New Caledonia

Baltzer (1970) identified a middle to late Holocene sea-level highstand from radiocarbon-dated samples of *Rizophora* peat, an indicator of the higher part of the intertidal zone. Cabioch et al. (1989) obtained elevations and dates of corals from 39 cores drilled into the fringing reef which also support a higher sea-level in New Caledonia about 6000 ^{14}C y BP and lasting until about 3000 ^{14}C y BP. Coudray and Delibrias (1972) originally proposed that sea level in New Caledonia stood 1 m higher than present between 4400 and 3000 ^{14}C y BP, and lasted above 0.7 m until 770 ^{14}C y BP, based on studies of emerged reef rock. Cabioch et al. (1995) summarized the preferred sea-level curve for New Caledonia which reached a peak of $\sim$ 1 m above present about 5500 ^{14}C y BP.

Fiji

Radiocarbon ages of emerged beach sediments at 0.75–0.95 m on Vanua Levu (Nunn 1990a) support numerous studies, reviewed in Nunn (1990b, 1991, 1994, 1995), that suggest Fiji experienced a sea-level highstand 1.5 m higher than present between 3000 and 1000 ^{14}C y BP. However, dates and elevations obtained by the HIPAC Team from emerged reefs, notches, and fossil corals in growth position (Miyata et al. 1988, 1990) on Vanua Levu, push the peak elevation up to 2 m and the age of the highstand peak back to 4000 ^{14}C y BP. This history is supported by elevations and ages obtained from emerged coastal terraces and notches (Berryman 1979), corals sampled from the islands of Nacilau and Rabulu (Ash 1987) and Wailevu (Roy 1988), shells from Naroi (Nunn 1988), and charcoal on Yanuca Island (Green 1979). Sugimura et al. (1988b) provided a time-transgressive reconstruction of the boundary between marine and terrigenous deposition and infer a 1 m highstand before 3000 y BP. However, they had no sample elevations or ages within the peak to control the maximum elevation or timing of the highstand apex.

Tonga

In central Tonga, Dickinson et al. (1994) obtained dates ranging from 6000–3000 ^{14}C y BP from charcoals excavated from abandoned and buried Holocene coastal archaeological sites on various islands within the Ha'apai Group. Post-highstand accretion of the coastal flat seaward of several of these sites indicates that progradation of the beach as a result of a relative sea-level fall has left a late Holocene shoreline emerged 1–2 m above msl.

Nunn (1991, 1994) reviewed several studies that point to late Holocene shoreline emergence in Tonga. Taylor and Bloom (1977) and Taylor (1978) found radiometric dates of 5900 ^{14}C y BP from emerged corals on 'Eua which they believed has been stable throughout the last 125 000 y. Similar findings by Davidson (1979) and Poulsen (1967) on Tongatapu and 'Eua suggest these islands have experienced late Holocene emergence. In the Nomuka Group and on Niuatoputapu, Cunningham et al. (1985) speculated that an emerged reef, believed to have doubled Niuatoputapo's land area since initial human occupation about 2750 ^{14}C y BP (Kirch 1978), is also of late Holocene age. Ellison (1989) documented sudden changes in mangrove zonation that suggest episodic uplift events have influenced late Holocene emergence on Tongatapu. Nunn (1995) provided a sea-level envelope for Tongatapu, with a "likely course of sea level" that rose to 1.6 m about 3500 ^{14}C y BP.

Western Samoa

Several researchers have found evidence of a highstand in Western Samoa. Emerged beachrock between 0.8–0.95 m (msl) on the islands of Savai'i and Upolu (Sugimura et al. 1988a), and a conglomerate-filled wave-cut notch in basalt at 2.3 m on Upolu (Rodda 1988) point to a recent highstand. Coral sands and fragments sampled from $\sim$ 1–2 m on both Sava'i and Upolu furnished radicarbon ages of 1850–1200 $\pm$ 70 ^{14}C y BP (Grant-Taylor and Rafter 1962), pointing to a late Holocene highstand of regional extent. Nunn (1991) stated that despite only few reliable dates, a 2.1 m sea-level highstand about 1300 ^{14}C y BP in Western Samoa is relatively certain, but the nature of sea-level rise to it is highly dependent on assumed subsidence rates affecting peat accumulation (Bloom 1980).

American Samoa

In the Manu'a Islands of American Samoa, Stice and McCoy (1968) found numerous features reminiscent of the $\sim$ 2 m shorelines reported throughout the Pacific Ocean, but did not obtain any dates for their formation. On Tau and the seaward side of Nu'utele Islet, they found benches carved into weak tuff, which they suggest may represent a short still stand at $\sim$ 1–2 m. They also found beachrock exposed at similar heights, which was reported as out of equilibrium with present sea level and like the benches, formed under a higher than present sea level.

Niue

Schofield (1959) found numerous emerged reefs on Niue (Savage) Island coincident with the $\sim$ 1–2 m shoreline noted throughout the Pacific. He speculated that a 5-foot notch is not explained by modern sea level. While he obtained no dates, Nunn (1994) showed that it plots close to the predicted 5000 y BP shoreline (Clark et al. 1978) in that region of the Pacific.

Cook Islands

In the southern Cook Islands, Ellison (1994) found evidence of a sea-level highstand in samples of pollen and charcoal from five coastal clay-filled swamps on the Island of Mangaia. Dates of samples of Cyperaceae and *Rhizophora*, indicate that between 6500 and 4500 ^{14}C y BP these swamps (presently 1.1 m above msl) were coastal embayments inundated by the sea. This is in agreement with Yonekura et al. (1988) who obtained ages ranging from 5000–3400 ^{14}C y BP from raised microatolls at a maximum elevation of 1.7 m, and reconstructed the age of the highstand apex at 3400 ^{14}C y BP. Stoddart (1972) reported raised coral reef rock at 1 m with an age of 2030 $\pm$ 60 ^{14}C y BP. In the northern Cook Islands, Scoffin et al. (1985) found evidence of a short-lived sea-level peak between 4000 and 2000 ^{14}C y BP among emerged microatolls and reefs on Manu, New Island and One Tree Island on Suwarrow Atoll. Variation in their elevations and large cracks in the exposed reef have been attributed to differential vertical tectonic motions.

Kiribati (Line Islands)

Tracey (1972) found evidence of a 1–2 m highstand from studies of emerged coral reef platforms and lagoonal environments on Jarvis, Starbuck, and Malden Islands. Maximum ages range between 3980 and 2940 ^{14}C y BP. The ages for Starbuck and Jarvis have been misreported in the literature (Stoddart 1972; Nunn 1994), thus the original elevations and ages as reported in (Tracey, 1972) and the *US Geological Survey*, *Virginia*, *Radiocarbon Dates XIV* report (Kelley et al. 1978) are documented here.

French Polynesia

Throughout French Polynesia, Pirazzoli and Montaggioni (1988) found that sea level stood 0.8 to 1.0 m above present between 5000 and 1250 ^{14}C y BP with a maximum about 2000–1500 ^{14}C y BP. Evidence from the Society Islands, Tuamotus, Gambier and Tubai (Austral) Islands include exposed corals, abandoned algal ridges and reef frameworks in growth position, emerged intertidal notches, and skeletal reef conglomerates in which the position of the former low water level at the time of cementation was determined by petrological analysis. Furthermore, they have shown with numerous dated samples ranging between 4500 and 1250 ^{14}C y BP, that sea level could not have dropped below $+$ 0.7 m for any appreciable time. The gradual drop in sea level since 1500 ^{14}C y BP however, is not explained by global isostatic model predictions which predict a highstand in this region during the middle Holocene. Pirazzoli and Montaggioni (1988) proposed that a gradual cooling of deep water in the region may explain the observed sea-level fall better than geoidal migration because the sea-level curve of the neighboring Tuamotus show a similar Holocene history and present-day configuration of contours in the geoid.

Hawaiian Islands

In the Hawaiian Islands, Stearns (1935) originally proposed that wave abrasion under a higher sea level about 5000 y BP, the Kapapa Stand of the Sea, was responsible for the formation of an emerged intertidal platform $\sim$ 2 m above msl on Kapapa Island, Oahu, and numerous other islands throughout the Pacific. He later based this age on a radiocarbon date of fossil coral (GX 2673, 3485 $\pm$ 160 ^{14}C y BP) found at 1.95 $\pm$ 0.45 m in Hanauma Bay, Oahu (Stearns 1974). Ku et al. (1974) observed modern wave overwash of the Hanauma bench and concluded that there is no evidence that sea level has been higher than present on Oahu since the last interglacial period. Bryan and Stevens (1993) proposed that salt weathering of the Hanauma Bay coastal cliffs and periodic wave overwash of the bench surface itself (restricting salt crystallization), are responsible for the Hanauma Bay bench. They did not, however, rule out the occurrence of a middle to late Holocene sea level above present.

Easton and Olson (1976) used 63 ^{14}C dates from cores in the Hanauma Bay reef to show that Holocene reef growth commenced about 7000 ^{14}C y BP, vertical accretion ensued at 3.3 mm/y between 5800 and 3500 ^{14}C y BP, and lateral accretion at 22.2 mm/y has dominated since 3000 ^{14}C y BP. Assuming reef growth kept pace with sea-level rise, they concluded that Holocene sea level was never higher than present despite evidence of reef truncation (reeftop age

~ 2500 ^{14}C y BP), fringing reef growth 2.4–3.0 m below msl (i.e., growth post-dated sea-level rise), and that reef accretion records the lower portion of the tidal range (Hopley 1986). In addition, the Hanauma reef-accretion curve contains a depth bias of 0.5 to 1 m too low due to the placement of "recovered" samples lacking depth control (cored open cavities) at the base of cored intervals. According to Montaggioni (1988), sea level would have been 2 to 3 m higher at the reef crest and up to 15 m higher at the reef front. In a comment on the work of Easton and Olson (1976), Stearns (1977) criticized their conclusion that there has not been a middle Holocene highstand and argued that their curve ignores an important outcrop of beachrock coincident with the elevation of the prominent 2 m bench found in Hanauma Bay and elsewhere in Hawaii and the Pacific.

Recently, we have shown that the stratigraphy of a fossil beach on Kapapa Island records a detailed depositional history between 5000 and 2000 y ago under a sea-level highstand that reached a maximum of 2.0 ± 0.35 m above present about 3500 y ago (Grossman and Fletcher 1998). Our reconstructed sea-level highstand is supported by an emerged fossil breccia cemented in a fossil intertidal notch on S. Mokulua Island, Oahu (Fletcher and Jones 1996), a fossil marine embayment buried under the Kailua and Hamakua, Oahu, coastal plains (Athens and Ward 1991, 1993), a regressive marine contact within the coastal plains in Kealia, Kauai (Matsumoto et al. 1988) and Hanalei, Kauai (Calhoun and Fletcher 1996), and a stranded coral reef in the present day Hanalei River (Jones 1992). Calibrating the Hanauma Bay reef-accretion chronology of Easton and Olson (1976) to calendar years, correcting sample depths for Holocene uplift and incorporating Montaggioni's (1988) 2.5 m reef-crest habitat correction, provides a new reconstruction of Holocene sea-level movements on Oahu (Grossman and Fletcher 1998).

To the north of the main islands in the Hawaiian chain, Ladd et al. (1970) described an emergent reef on Midway with dates ranging from 2400–1280 ^{14}C y BP as evidence of a 1–2 m higher sea level. It appeared to have been significantly eroded, so its maximum extent may have been even higher. Gross et al. (1969) found similar emerged reefs on Kure Atoll and an age of 1480 ^{14}C y BP, indicating that a relative sea-level highstand of 1–2 m had a regional extent throughout the northern Hawaiian Islands.

Criteria for data selection

The source materials used in these paleosea-level studies (Table 1) include emerged and submerged archeological sites, algal crusts, intertidal benches, in situ corals, carbonate conglomerates, marine-cemented conglomerates, coral reef flats, in situ molluscs, fossil beaches, microatolls, notches, peat, pollen, beach ridges and progradation complexes, regressive overlap boundaries, marine sand lithosomes, and coastal terraces or plains. Their utility as sea-level indicators and their relation to sea-level datums have been proven in various studies (van de Plassche 1986; Pirazzoli 1991) with a wide range of confidence levels. Depending on local tectonic and depositional environments, these source materials may better represent limits to paleosea level rather than estimates of it. Inter-sample and inter-reconstruction variations can be subtle such that criteria for classifying and assessing their utility as sea-level indicators are necessary before they can be used to analyze processes acting within spatial scales of meters. Some of these data are simply surveyed heights and ages of shoreline features, while others have been carefully studied and proven as vital components in deriving a sea-level curve. We consider the suitability of data as a sea-level proxy, the extent of analysis (e.g., sea-level reconstruction versus field sampling of an elevation/age data pair), and island stability as important criteria for selecting the most adequate sea-level proxy data from Table 1 for model comparisons.

Of the 92 highstand estimates in Table 1, 91 have elevation assessments, 85 have radiometric ages, and 84 have a combined elevation-age pair determination representing the highstand apex. We give proxies of paleosea-level position (PSP) that include an assessment for environmental factors (e.g. tide range, habitat range, relation to modern analogs, local tectonics) and that define published sea-level reconstructions highest priority with a ranking (1) in the "PSP" column ($n = 38$, Table 1). Proxies or shoreline features that are not proven components of sea-level curves but that are corrected for environmental uncertainties and serve as supporting evidence of paleosea-level position ($n = 22$) are ranked (2). Sea-level proxies or shoreline features that include uncertainties in their relation to a sea-level datum or require corrections for (un)known tectonics, compaction, and/or possible reworking ($n = 25$) are ranked (3), while features that lack elevation or age control ($n = 7$) are ranked (4).

To address island stability during the Holocene, we apply the criteria of "effectively stable" islands in Nunn (1994, Table 8.5, p. 285), excluding the data from New Caledonia and Western Samoa for which subsidence rates remain uncertain, and include a wide range of estimates. We have included Bikini Atoll (Tracey and Ladd 1974), Butaritari, Tarawa, Abemama, Tabiteuea, and Funafuti (Schofield 1977), Ifaluk (Tracey 1968), and Kosrae (Athens 1995) where sea-level estimates have accounted for steady subsidence or where aseismic and coseismic movements have not been identified (Nunn 1994). In addition, we include data from Oahu, Hawaii and Makatea, Tuamotu Islands, which have undergone slow, steady uplift during the late Quaternary and Holocene, probably due to lithospheric flexure, based on the work in Hawaii by Muhs and Szabo (1994) and in French Polynesia by Pirazzoli and Montaggioni (1988).

Our final step in culling proxies of paleosea level from Table 1 consists of eliminating replicate sea-level estimates for each island meeting the effectively stable criteria and ranked with a PSP of 1 or 2. This includes Enewetok, Vanua Levu, Oahu, and Kauai. The duplicate estimates for Enewetok are nearly identical, and we have selected the more recent of the two studies that elaborated on the earlier work. We have selected the sea-level reconstruction of Miyata et al. (1990) for Vanua Levu, for its detailed treatment of emerged notch and microatoll elevations and tectonic reconstruction. On Oahu, we have selected the reconstruction of Grossman and Fletcher (1998) over the minimum curve based on reef accretion (Easton and Olson, 1976). For Kauai we use the estimate of Jones (1992) even though the fossil reef could not be confirmed (Calhoun and Fletcher, 1996), because the conclusion based on emerged coral contains fewer corrections than the other studies and is supported by Matsumoto et al. (1988) and Calhoun and Fletcher (1996). The testable data set culled from Table 1 that meet the criteria set forth here are considered best proxies for paleosea-level analysis and are divided into a primary set ($n = 13$, Table 1 locations distinguished by superscript (a)) and a secondary set of supporting data ($n = 10$, Table 1 locations distinguished by superscript (b)).

Results and discussion

The criteria used, which give precedence to sea-level proxy data derived from detailed sea-level reconstructions and effectively stable islands, enable a refinement of the estimated position of relative sea level during the middle to late Holocene. Through least squares curve fitting techniques, we show (Figs. 4–9) that a steady improvement is made in fitting the criteria-based "primary" data set to non-linear relationships as proposed for late-Holocene sea level by geophysical models (Nakada 1986; Nakada and Lambeck 1989; Mitrovica and Peltier 1991; Peltier 1994, 1996). Correlation coefficients are provided to show the improvement in goodness of fit, not to statistically validate these models.

The early Clark et al. (1978) model of the migrating geoid predicted that the 5000 y BP shoreline along the

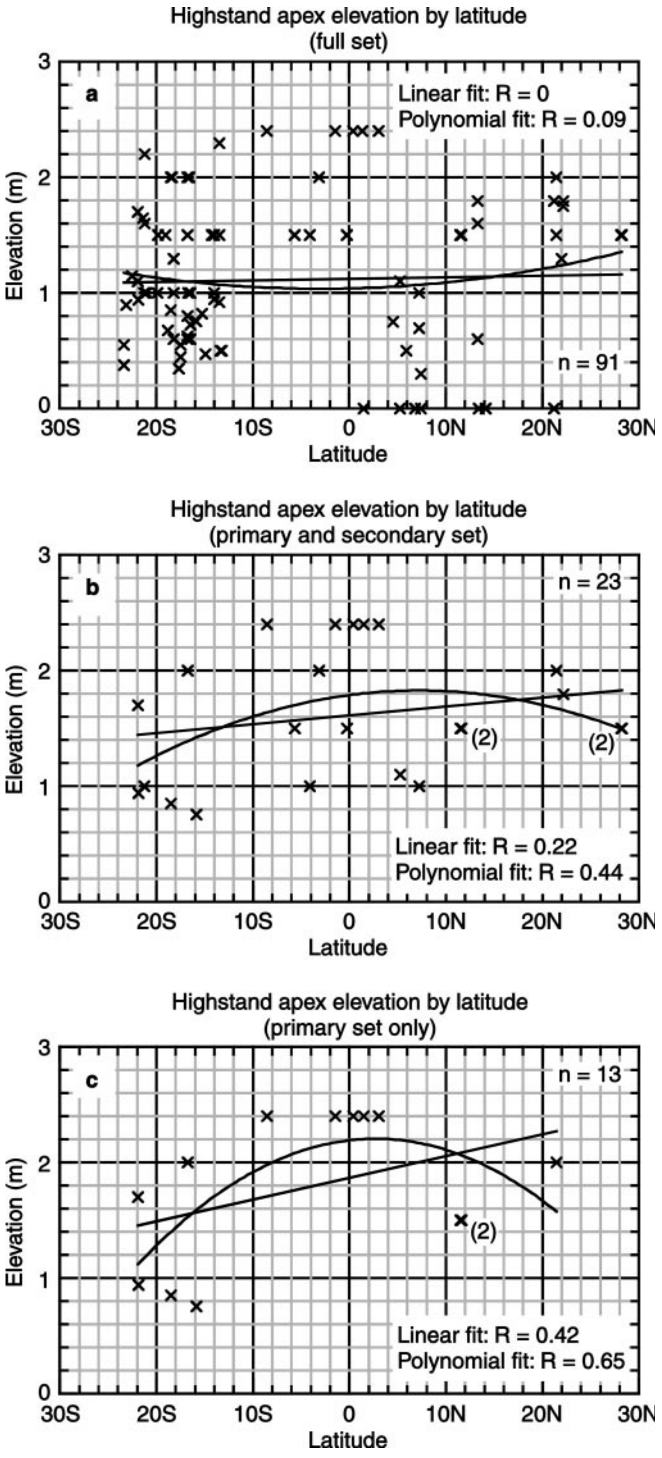

Fig. 4a–c Linear and 2nd-order polynomial best fits to the maximum highstand elevation plotted as a function of latitude for **a** the full data set of Table 1, **b** the primary and secondary set, and **c** primary data set alone satisfying our selection criteria

165° W meridian was highest near 50°S and decreased northward. This was only partly supported by 5 data points (Nunn 1994). A slight increase in paleoshoreline height to the south has also been found in Polynesia (Pirazzoli and Montaggioni 1988). Figure 4 suggests that the best linear fit to the highstand apex elevations plotted as a function of latitude is poor using our full data set (Fig 4a), but steadily improves with the application of our selection criteria (Fig. 4b, c). These suggest higher paleosea-level positions between 10°N and 10°S, unlike Clark et al. (1978). The mechanism of equatorial ocean siphoning (Mitrovica and Peltier 1991) requires that the maximum geoid anomaly centered in the equatorial basin decays toward the continents and especially regions previously ice-covered. As a result, the greatest emergence should occur in the central Pacific, roughly near the equator. Hence, the paleogeoid would be parabolic in shape rather than linear and maximum emergence would occur in the equatorial region, as shown in Fig. 4c.

An inverse relationship exists between highstand apex age and latitude (Fig. 5). Although the best linear fit of the age data to latitude improves with the application of our selection criteria and shows older highstand ages in the Southern Hemisphere, the better 2nd-order polynomial fit suggests that emergence near the equator is younger than emergence at higher latitudes. The parabolic relationship between apex age and latitude indicates that differential rheological adjustment occurs across latitudes with time. In light of higher paleosea level near the equator (Fig. 4), younger emergence there requires a higher rate of sea-level fall.

Fig. 6 shows an improvement in fitting linear and polynomial models to the highstand apex elevations relative to their ages as we apply our selection criteria. In each step the linear fit is similar to the polynomial relation. However, the best fits using our selection criteria (Fig. 6b, c) suggest that higher elevations are younger. This strong inverse relationship implies that different rates of isostatic adjustment have occurred over latitude and through time.

The inverse relationship between the highstand apex elevations and ages (Fig. 6c) supports a postglacial history in the Pacific characterized by a highstand above present. In addition, assuming the data are sufficiently robust, the elevation-age pairs represent minimum elevations and maximum ages of the highstand peak. In response to this transgressive-regressive history, three modes of deposition based on different relative rates of subsidence and comparable sediment supply are apparent. Each has a distinct impact on the development of islands, shorelines, and reefs (Richmond 1992; Woodroffe 1992). Deposition would ensue placing younger sediments above older sediments up to a discrete elevation (depth) primarily controlled by the maximum highstand position. Beyond the inflection point, however, different relative rates of subsidence and sea-level fall, will result in distinctly different deposits. Where subsidence is less than the rate of sea-level fall, a lack of deposition and/or erosion could leave apex-age deposits stranded highest above present sea level and create a void in the fossil record of emerged sediments younger than the apex age (Fig. 6c).

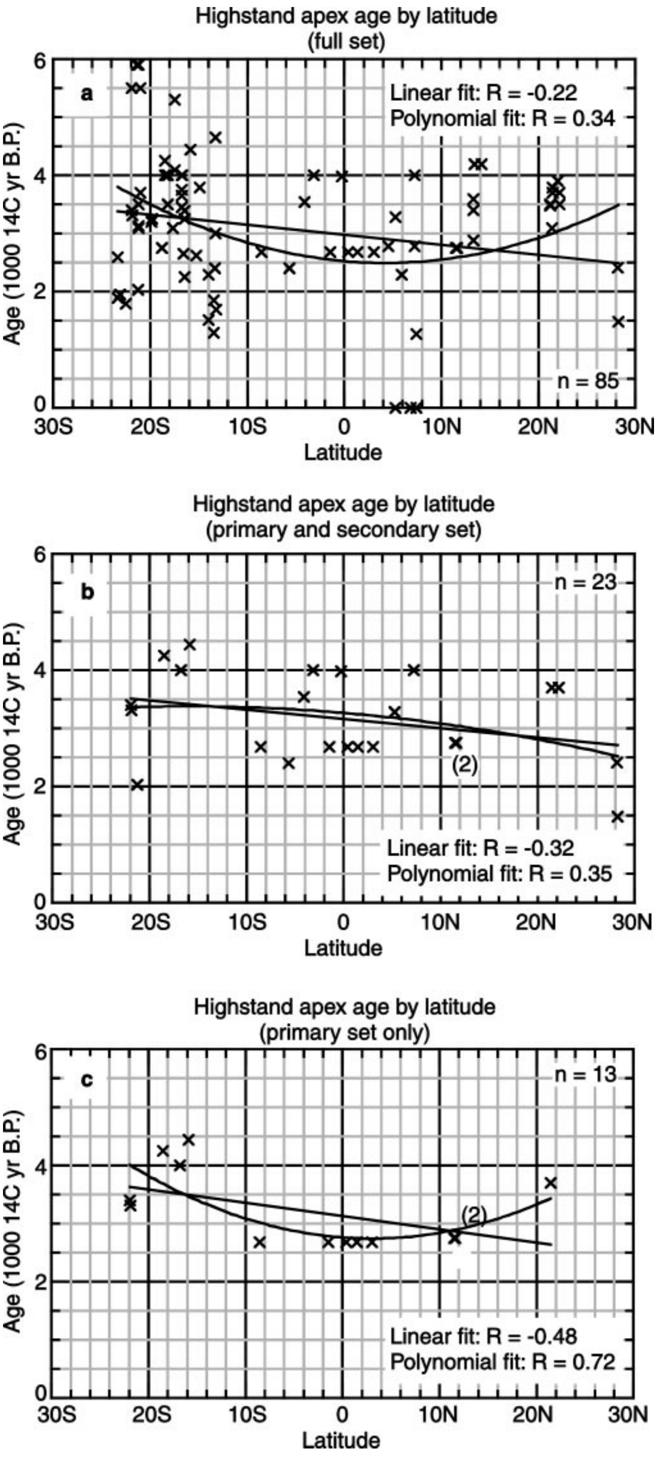

Fig. 5a–c Linear and 2nd-order polynomial best fits to the age of highstand apex plotted as a function of latitude for **a** the entire data set of Table 1, **b** the primary and secondary set, and **c** primary data set alone that satisfy our selection criteria

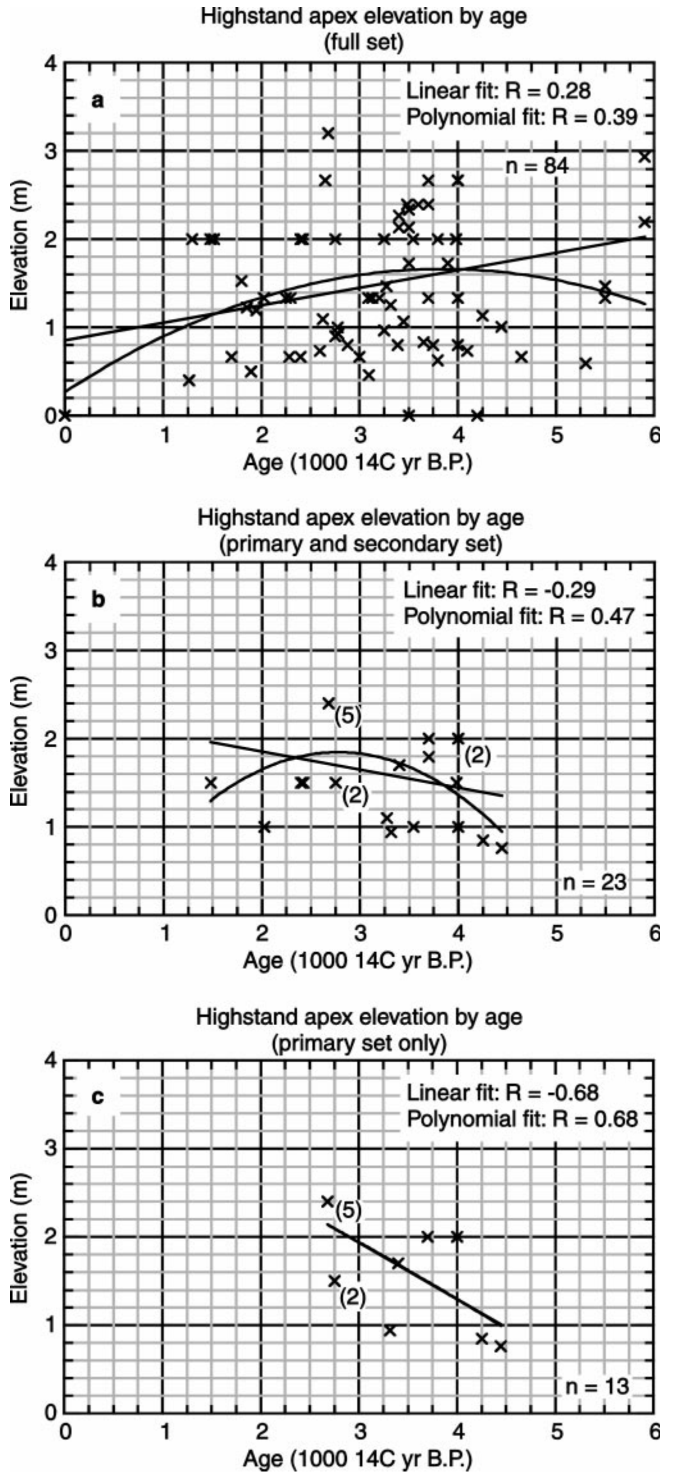

Fig. 6a–c Linear and 2nd-order polynomial best fits to the maximum highstand elevation plotted as a function of highstand apex age for **a** the entire data set of Table 1, **b** the primary and secondary set, and **c** primary data set alone that satisfy our selection criteria

Where subsidence is greater than the rate of sea-level fall, deposition could continue to the present position of sea level, possibly reaching it in the recent (common on subsiding coasts). Where subsidence equals rate of sea-level fall, deposition could reach present sea-level position prior to the highstand peak, but the subsequent sea-level fall would appear as an apparent stillstand; the top of the deposit would be truncated close to present msl with ages close to (or just older than) the highstand peak.

In each of these depositional settings, a plot of maximum elevations as a function of age would show an inverse relationship as in Fig. 6c. However, only the first model could leave deposits stranded above present msl and only if there was sufficient deposition (assuming stable tectonics). In addition, because the probability is high that post-highstand erosion has affected the various emerged deposits remaining in the Pacific, the elevation-age pairs in Fig. 6c may underestimate the maximum position of the highstand and pre-date the timing of the apex.

We can use the primary set of sea-level reconstructions (Table 1) to test if published sea-level histories following the sea-level highstand apex support greater and more recent emergence near the equator. Table 2 lists the elevation, $Elev_{(max)}$, and associated age, $Age_{(max)}$, of the maximum position of the Holocene highstand, the age when the sea began to fall from the highstand peak, $Age_{(t_0)}$, as well as the youngest available age, $Age_{(t_1)}$, and associated position of sea-level, $Elev_{(t_1)}$, derived from the originally published sea-level curves. Fig. 7 shows that rates of sea-level fall from the highstand toward its present position are greater near the equator. These estimates are largely based on the variables $Age_{(t_0)}$, $Age_{(t_1)}$, and $Elev_{(t_1)}$ which are independent of $Age_{(max)}$ and $Elev_{(max)}$. As such, these results are consistent with the analysis of Fig. 4 and 5, and suggest that regional sea level fell most recently, and most rapidly, in the equatorial Pacific. This implies an asymmetric development and decay of the equatorial highstand with the younger history falling steeply towards its present msl position.

Hydroisostasy predicts shoreline emergence independent of latitude because on the time scale of Earth's viscoelastic adjustment (10^3–10^4 y), the addition of meltwater to the world's oceans occurred simultaneously (10^1–10^2 y) across all latitudes. According to hydroisostasy, islands with radius < 10 km submerge with the surrounding seafloor as it is depressed under the added load of ~ 120 m of postglacial meltwater. As the seafloor is depressed, it induces sublithospheric flow away from the load and toward the underside of islands, such that islands with radius > 10 km (more buoyant) will emerge. As stated by Nakada (1986), larger islands should undergo greater emergence. This is supported by our filtered data, which show a positive linear relationship between highstand apex elevation and island radius (Fig. 8b,c) that is an improvement

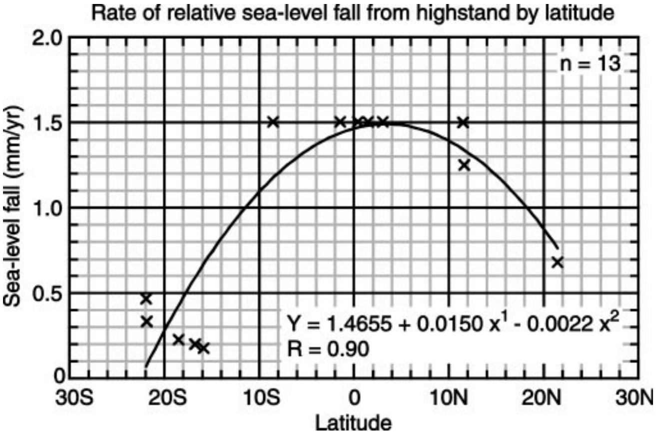

Fig. 7 Rates of relative sea-level fall following the highstand derived from the primary data set plotted as a function of latitude

Table 2 Maximum elevations of the Holocene sea-level highstand and rates of post-highstand sea-level fall derived from primary data set

Location	$Elev_{(max)}$ (m)	$Age_{(max)}$ (1000 ^{14}C y BP)	$Age_{(t_0)}$ (1000 ^{14} C y BP)	$Age_{(t_1)}$ (1000 ^{14}C y BP)	$Elev_{(t_1)}$ (m)	Rate sea-level fall (mm/y)	Reference
Enewetok	1.5	2.75	2.0	1.00	0	1.50	Buddemeier et al. (1975)
Bikini	1.5	2.75	2.2	1.00	0	1.25	Tracey and Ladd (1974)
Butaritari	2.4	2.68	2.68	1.75	1.0	1.51	Schofield (1977)
Tarawa	2.4	2.68	2.68	1.75	1.0	1.51	Schofield (1977)
Abemama	2.4	2.68	2.68	1.75	1.0	1.51	Schofield (1977)
Tabiteuea	2.4	2.68	2.68	1.75	1.0	1.51	Schofield (1977)
Funafuti	2.4	2.68	2.68	1.75	1.0	1.51	Schofield (1977)
Vanua Levu	2.0	4.0	4.0	3.00	1.8	0.20	Miyata et al. (1990)
Mangaia	1.7	3.4	3.4	2.33	1.2	0.47	Yonekura et al. (1988)
Makatea	0.76	4.444	1.25	0.50	0	0.18	Pirazzoli and Montaggioni (1988)
Reao	0.85	4.25	1.25	0.50	0	0.23	Pirazzoli and Montaggioni (1988)
Muruoa	0.94	3.315	1.25	0.50	0	0.33	Pirazzoli and Montaggioni (1988)
Oahu	2.02	3.70	3.70	2.00	1.0	0.68	Grossman and Fletcher (1998)

over our full data set (Fig. 8a). The 2nd-order polynomial fit to the highstand apex elevation and island radius data (Fig. 8b,c) is only a marginal improvement over the linear model.

Fig. 9 shows that there is no linear relationship between the highstand apex age and island radius. However, the subtantially improved 2nd-order polynomial fit and application of our criteria (Fig. 9c) suggests that a nonlinear relationship exists for these data. In light of emergence trends with latitude (Figs. 4, 5), these results indicate that variation across latitude is important in addition to island size for determining the timing of emergence.

Fig. 10 shows contour plots of the paleosea-surface topography at 4000 ^{14}C y BP (Fig. 10a) and 2500 ^{14}C y BP (Fig. 10b) based on elevation-age pairs obtained from the original published sea-level curves and provided in Table 2. Fig. 10c shows the change in sea level between 4000 and 2500 ^{14}C y BP with cool colors signifying a fall and warm colors representing a rise. At some island locations (e.g. Enewetok, Tarawa, Oahu) these snapshots may capture the timing of the highstand apex; however, at others (e.g. Midway and Kure), they may post-date it. The contours are only constrained between the sample sites (red dots), thus we do not speculate about the exterior portion of the map. We also focus our discussion on the general trend of the topography.

According to our primary data set, sea level about 4000 ^{14}C y BP (Fig. 10a) was 1 to 2 m higher throughout most of the western central Pacific except in the vicinity of French Polynesia where it ranged between 0.8 and 1.0 m. The highest topography about 4000 ^{14}C y BP was found near Hawaii north of the equator and near Fiji to the south. At 2500 ^{14}C y BP (Fig. 10b), sea level ranged between 1 and 2 m, but the region of highest sea level shifted toward the west. Sea level was generally lower in the eastern portion of the region and slightly higher in the west than at 4000 ^{14}C y BP. The greatest change in sea level between 4000 and 2500 ^{14}C y BP. (Fig. 10c) occurred in the central low latitudes near the Gilbert Islands where it rose ~ 0.7 m and in the eastern region near French Polynesia where it fell ~ 0.1 to 0.2 m. The general change between 4000 and 2500 ^{14}C y BP is a sea-level fall in the central Pacific (east of the dateline) and a relative rise in the western Pacific. This generally supports the equatorial ocean siphoning model which predicts greatest emergence in the central ocean basin interior (the eastern central Pacific).

Northern Hemisphere ice sheets are thought to have grown larger in area and mass than their southern counterparts, and to have produced larger troughs and forebulges around their boundaries. The northern ice sheets are also believed to have melted earlier and faster (ice-melt histories prescribe this), thus fostering a mechanism for more rapid migration of the geoid northward to fill the larger forebulge voids there. The question arises, what is the behavior of the migrating geoid? Does it simply relax or does it migrate either as a single peak or an oscillating wave series from its maximum position in the central ocean basin interior?

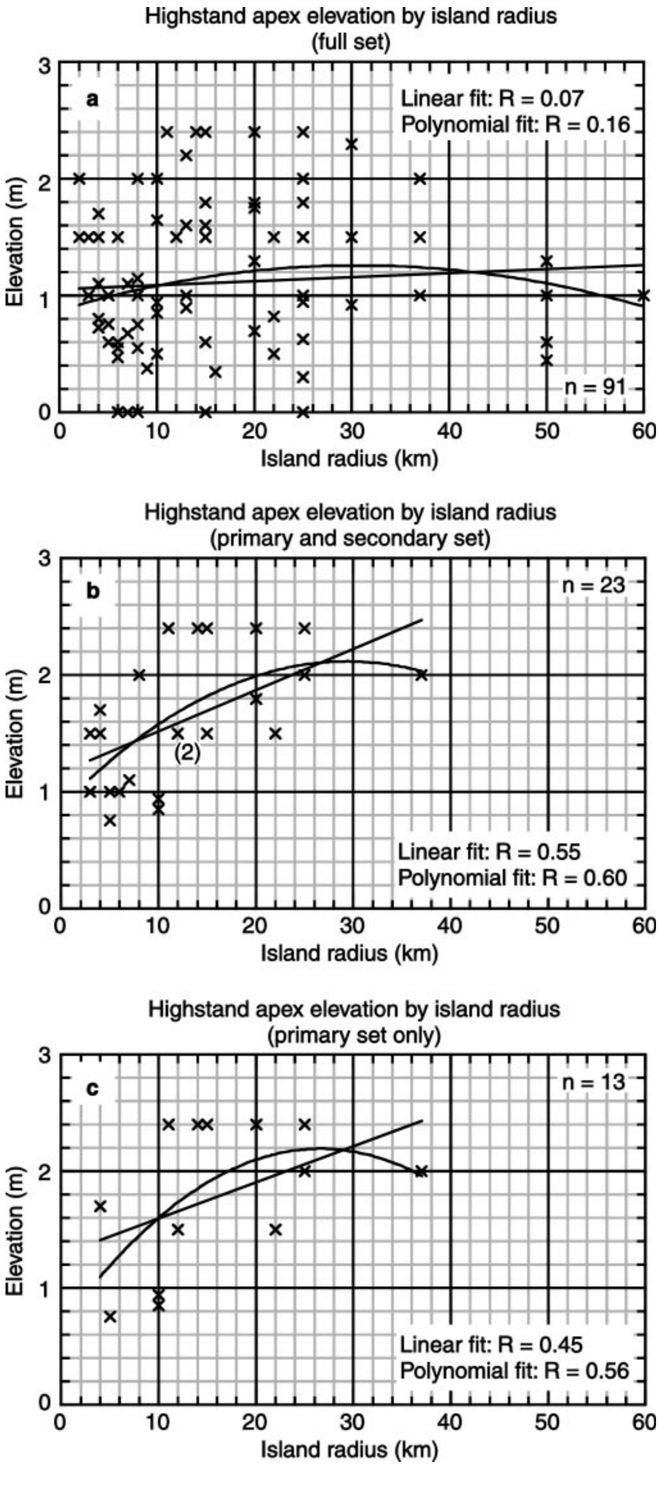

Fig. 8a–c Linear and 2nd-order polynomial best fits to the maximum highstand elevation plotted as a function of island radius for **a** the entire data set of Table 1, **b** the primary and secondary set, and **c** primary data set alone that satisfy our selection criteria

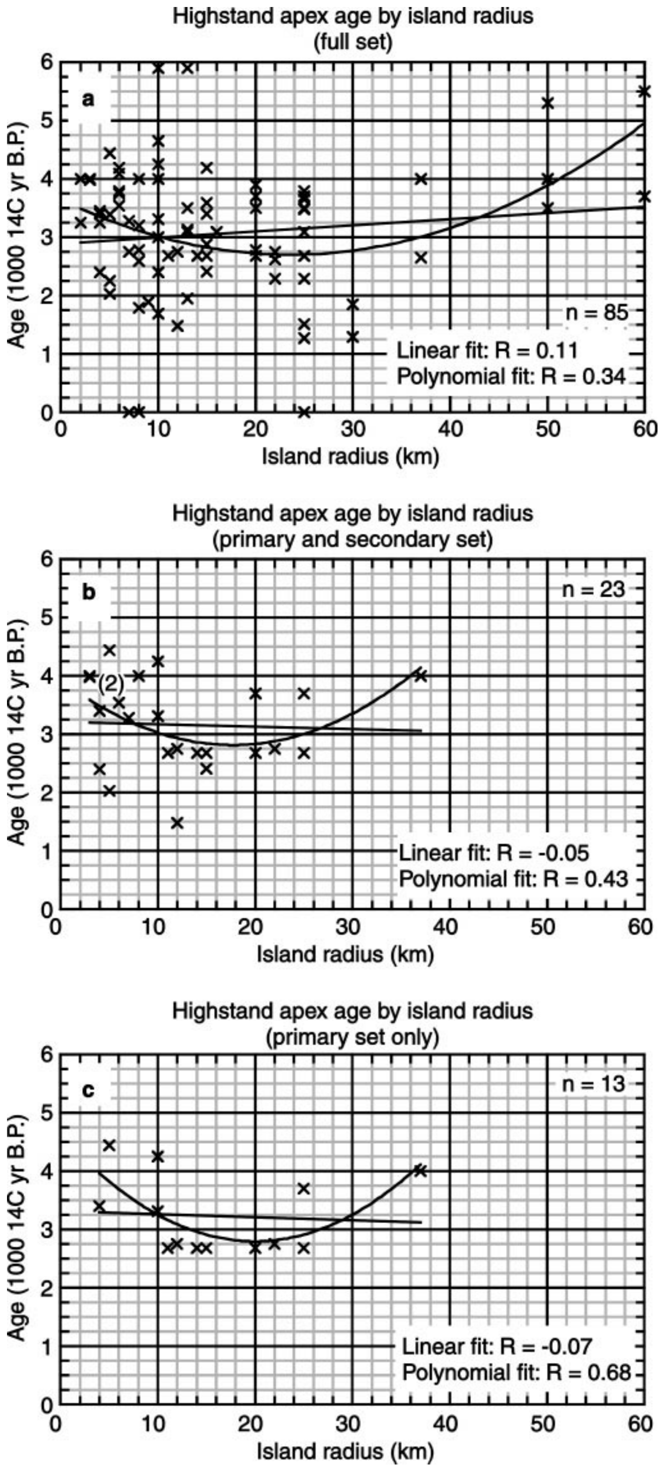

Fig. 9a–c Linear and 2nd-order polynomial best fits to the age of the highstand apex plotted as a function of island radius for **a** the entire data set of Table 1, **b** the primary and secondary set, **c** and primary data set alone that satisfy our selection criteria

As the geoid migrates to higher latitudes, do low- and mid-latitudes (5°–30°N/S) experience a slight, perhaps short-lived transgression and regression? Could hydroisostasy in concert with emergence due to the migrating geoid create a minor oscillation of sea level? These processes may disguise the timing and elevation of the eustatic peak of postglacial sea level or be hidden within a broad peak such as French Polynesia (Pirazzoli and Montaggioni 1988). How can we explain slightly higher (Fig. 4) and younger (Fig. 5) maximum paleosea-level positions in the Northern Hemisphere? Does equatorial ocean siphoning predict the geoid anomaly to decay in the western central Pacific toward the Asian continent (Fig. 10)?

The elevation and age data in Table 1 comprise a wide range of paleosea-level estimates from numerous island settings in the Pacific. Errors including those from measurements of elevation, age, tides, tectonics, wind and wave set-up, storm deposition, and environmental interpretation, make the uncertainty of the paleosea-level estimates comparable in magnitude to the marginal changes we want to understand. Many paleosea-level estimates are superior to others for testing and constraining model relationships and depicting the paleosea-surface topography. A significant barrier to making observational and model comparisons is the lack of standardization in treating and reporting paleoshoreline elevations and ages. Sea-level researchers are referred to van de Plassche (1986) and Pirazzoli (1991) for interpreting sea-level indicators, and Stuiver and Pollach (1977), Bard et al. (1990), Stuiver and Braziunas (1993), and Stuiver and Reimer (1993) for correcting and reporting ^{14}C ages.

In conclusion, using criteria that gives precedence to sea-level proxies derived from sea-level reconstructions, environmental considerations, and island stability, we have sorted a data set for improving our understanding of middle to late Holocene sea level on Pacific Islands. Our analyses of these data generally support geophysical model predictions of higher middle to late Holocene sea level in the central Pacific Ocean basin interior (near the equator) and generally higher sea level on islands of larger size. We find a non-linear relationship between the age and maximum paleosea-level elevation, which suggests that different rates of isostatic (vertical) adjustment may occur throughout the Pacific and during the Holocene, with highest rates of sea-level fall following the highstand near the equator. One mechanism invoked for different rates of paleosea-level movements include lateral variations in Earth's rheology which would influence the response to and adjustment of the load on the sea floor resulting from ~ 110 to 120 m of added meltwater during deglaciation.

Variation in the elevations and ages of the maximum position of Holocene sea level imply that the paleosea-surface, like the present configuration, is better represented by an irregular topography of regional-scale high points and depressions rather than a smooth spherical surface equal in height in all locations. This is due to regional-scale variation in the gravitational potential of the lithosphere that influence fluid behavior on the time scale of the late Holocene. As a result,

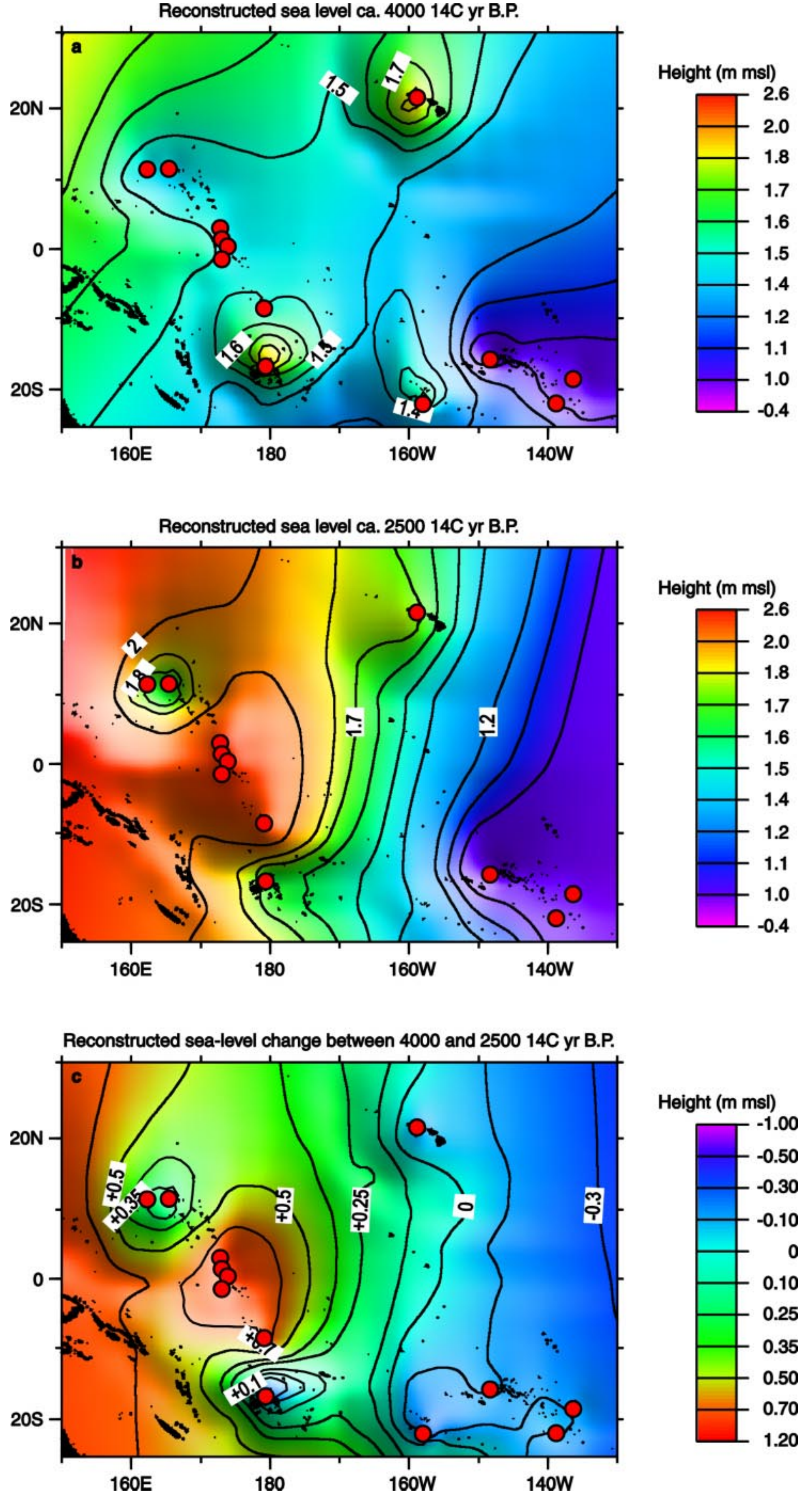

Fig. 10a–c Contour plot of the paleosea-surface topography at **a** 4000 ^{14}C y BP and **b** 2500 ^{14}C y BP and **c** the change in sea level between 4000 and 2500 ^{14}C y BP

detailed sea-level reconstructions must be treated as histories of only limited regional validity. Local sea-level histories, however, are vital for understanding natural rates of relative sea-level change that shaped coastal environments in the recent past.

The middle to late Holocene sea-level highstand peaked between 1 and 2 m above present between 5000 and 1500 y BP in the central equatorial Pacific. Knowledge of the behavior of this higher regional sea surface, especially with respect to local-scale tectonics, is important to better understand variation we see in coastal evolution, coral reef distribution, reef stratigraphy and geochronology, island formation, and human settlement patterns. Of particular importance to many Pacific Island societies are low-lying coastal plains that are underlain by unconsolidated marine sediments of middle to late Holocene age and owe their existence to shoreline progradation as sea level fell following the highstand. Often it is these same low-lying coastal environments which have experienced the greatest population increase and resource use during the last 100 years, that are now at greatest risk of inundation and modification by rising sea levels expected in the next 100 years. Knowing the natural rates of sea-level movement and shoreline adjustment in the past may enable us to better predict and respond to future changes.

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