A MARINE RESERVOIR CORRECTION FOR THE HOUTMAN-ABROLHOS ARCHIPELAGO, EAST INDIAN OCEAN, WESTERN AUSTRALIA

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ABSTRACT. High-precision analysis using accelerator mass spectrometry (AMS) was performed upon known-age Holocene and modern, pre-bomb coral samples to generate a marine reservoir age correction value (ΔR) for the Houtman-Abrolhos Archipelago (28.7°S, 113.8°E) off the Western Australian coast. The mean ΔR value calculated for the Abrolhos Islands, 54 ± 30 yr (1σ) agrees well with regional ΔR values for Leeuwin Current source waters (N-NW Australia-Java) of 60 ± 38 yr. The Abrolhos Islands show little variation with ΔR values of the northwestern and north Australian coast, underlining the dominance of the more equilibrated western Pacific-derived waters of the Leeuwin Current over local upwelling. The Abrolhos Islands ΔR values have remained stable over the last 2896 cal yr BP, being also attributed to the Leeuwin Current and the El Niño Southern Oscillation (ENSO) signal during this period. Expected future trends will be a strengthening of the teleconnection of the Abrolhos Islands to the climatic patterns of the equatorial Pacific via enhanced ENSO and global warming activity strengthening the Leeuwin Current. The possible effect upon the trend of future ΔR values may be to maintain similar values and an increase in stability. However, warming trends of global climate change may cause increasing dissimilarity of ΔR values due to the effects of increasing heat stress upon lower-latitude coral communities.

INTRODUCTION

Surface ocean radiocarbon levels are largely representative of the proportionate contributions of atmospheric CO_2 and ^{14}C-depleted subsurface water DIC, thus making them lower than atmospheric and terrestrial [14]C. The influence of the "older" ^{14}C-depleted carbon source incorporated by marine organisms creates a differential between the marine organism's ^{14}C age and the ^{14}C age of its terrestrial contemporary. The average offset between the marine and terrestrial ^{14}C age (~400 yr) is referred to as the marine reservoir age (*R*) and is described and accounted for by the application of atmospheric modeling. The marine reservoir age also has regional variation referred to as the marine reservoir correction or ΔR value and represents the local influence of ocean and atmospheric mixing due to currents, gyres, upwelling and climatic variations, which vary on seasonal to multidecadal scales (Stuiver and Reimer 1986; Stuiver et al. 1986; Stuiver and Braziunas 1993; Reimer and Reimer 2001).

Around the Australian coastline, especially towards the remote north and northwestern regions and along the East Indian Ocean, there are many gaps in the record of ΔR values with low spatial resolution (Reimer and Reimer 2001; Southon et al. 2002; Ulm 2006). In addition, there are several factors that limit the range of existing ΔR values available for any chosen analysis. To avoid potential errors when selecting ΔR values (or material to sample) for calculating site or regional averages, one must consider beforehand criteria such as: the trophic level and particular feeding habits of sampled organisms; the physicochemical nature of the aquatic environment and habitat of collection; the possibility of diagenetic alteration; and the dating/calibration methods, calculations, and corrections

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used (Reimer and Reimer 2001; Scholz et al. 2004; Ulm 2006; Nothdurft et al. 2007; Nothdurft and Webb 2009).

Corals represent ideal candidates for calibrating ΔR values due to their ability to fix dissolved inorganic carbon (DIC) from the surrounding water column (Grottoli and Eakin 2007). The ^{14}C signatures of their carbonate skeletons reflect the oceanographic conditions integrated over time and are a good proxy for oceanic circulation and climatic processes at subannual to millennial scales (Grottoli and Eakin 2007; McGregor et al. 2011). This paper presents a local marine reservoir correction for the southernmost coral reef complex in the Indian Ocean and evaluates this in context with ΔR values calculated for the source waters of influential currents including local values in the broader region.

STUDY SITE

The Abrolhos Archipelago (between 28°04′S and 29°27′S) is a series of coral reef islands along the passive shelf margin off the Western Australian coast in the Indian Ocean approximately 70 km from Geraldton (Figure 1). It is the southernmost coral reef complex in the Indian Ocean, comprising 122 islands in 3 distinct groups: the Pelsaert Group in the south, the Easter Group in the center, and the Wallabi Group in the north. The age structure of the emergent coral reef shingle pavements, underlying reefs, and coral-rubble storm ridges have been reported previously but not extensively or in great detail until sea-level studies performed by Collins et al. (2006) and paleo-tempestological work by Scheffers et al. (2012). The region is considered to be "far field" in terms of tectonism and isostatic adjustment (Collins et al. 2006).

The long-term dominance of the warm, tropically derived waters of the Leeuwin Current is the reason permanent coral communities are established at the Abrolhos Islands in such high latitudes. Compared to other boundary currents around the world, little detailed or extensive study of its hydrodynamics, physicochemical, or ecological aspects occurred in the last 2 decades until the research voyage of the *Southern Surveyor* in 2003, the findings of which are published in the *Deep Sea Research II* Special Issue 2007. The Leeuwin Current (LC) is a pole-ward flowing eastern boundary surface current transporting warm, oligotrophic, low-salinity water along the Western Australian seaboard, creating a predominantly downwelling coastline along its route (Pearce and Phillips 1988; Pearce and Pattiaratchi 1999; Hanson et al. 2005; Waite et al. 2007). Source waters for the LC are derived from the Western Pacific Warm Pool (WPWM) via the Indonesian throughflow, the South Equatorial Current, and the Java Current including contributions from the Indian Ocean (Pearce and Phillips 1988; Pearce and Pattiaratchi 1999; Domingues et al. 2007). These source waters are considered to be well mixed and equilibrated with the atmosphere having typically low ΔR values around or below 60 yr (Reimer and Reimer 2001; Southon et al. 2002; Hua et al. 2004; Ulm 2006). The strength of the LC varies with seasonal shifts in prevailing wind fields, the effect of which is magnified by ENSO oscillation. During the austral winter, the LC experiences less resistance from opposing south to southwesterly winds arising from the Southern Ocean and is at its peak velocity. During summer, the opposing southerly wind gradient is at its strongest, slowing the LC significantly and occasionally reversing flow direction in some instances. Similarly, when La Niña periods develop the LC flow is enhanced and during El Niño events, subdued (Pearce and Phillips 1988; Pearce and Pattiaratchi 1999; Li and Clarke 2004; Woo et al. 2006; Domingues et al. 2006, 2007; Feng et al. 2007; Waite et al. 2007). During the early to mid-Holocene up to 5000 yr BP, there was a rising ENSO signal with a strengthening of the LC due to a more effective summer monsoon and higher frequency of La Niña events. Since then, there has been a rising El Niño frequency and resulting decline in precipitation and increase in arid conditions. With lower precipitation and surface flows, the vegetation down the NW-W coast of Australia, where the LC operates, has gradually
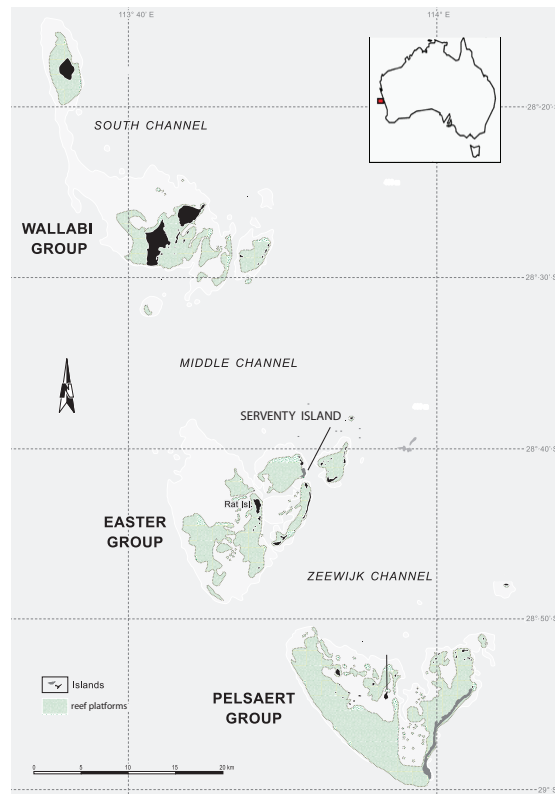
Figure 1 Map of the Abrolhos Islands. Coral samples were taken from Serventy Island. For further details, see Collins et al. (2006).

tended towards semi-desert, giving rise to the modern conditions observed today (van der Kaars and De Deckker 2002; McGregor et al. 2008; McCulloch and Mortimer 2008; Wyrwoll et al. 2009).

MATERIAL AND METHODS

The sample material used in this study was obtained from the collection of the Collins et al. (2006) sea-level study. The material comprised of coral framework and individual coral samples (mostly *Acropora*) from storm rubble ridges along the HST2 Sample Traverse on Serventy Island (28.682°S, 113.835°E). The representative age span of the sample material was about 0.116–7 U-Th ka cal BP. In order to reduce noise or error from interannual variability, sample material for AMS was obtained from coral sections long enough to cover several years of growth along identifiable growth axes (>7 cm). Suitable candidates were then selected across the available 7000-yr age span.

Prior to sampling for AMS, the selected coral pieces were screened for diagenetic alteration at the Earth, Environmental and Biological Sciences, Queensland University of Technology. The coral sections were cut in longitudinal and transverse sections, polished, etched, and carbon coated or left uncoated for observation using an FEI Quanta™ 200 environmental scanning electron microscope (SEM) with energy dispersive spectroscopy (EDS). SEM observations were made in several operation modes, including operation at high vacuum (15 and 20 kV) on samples coated with carbon and at 2–10 kV for uncoated samples using both secondary and backscattered electron imaging. Secondary electron images were made from freshly broken and external surfaces, with some specimens

etched with dilute formic acid (2%) for 20 s prior to being carbon coated. Carbonate polymorphs and other minerals were identified *in situ* on broken, sawn, and polished sections (Nothdurft and Webb 2009). The sample material deemed suitable for AMS analysis after screening had a resultant age span of 0.116–4.334 U-Th ka cal BP (Table 1).

Sample material for AMS analysis was carefully removed from the larger central coral mass along the growth axes with a dental drill. The samples were cleaned with deionized water in an ultrasonic bath twice for 40 min each to remove any surface contamination. They were then dried in an oven at 60 °C for 2 days before hydrolysis. The cleaned samples were hydrolyzed to CO_2 using 85% phosphoric acid. The CO_2 samples were then converted to graphite using the H_2/Fe method (Hua et al. 2001). A small portion of graphite from each sample was employed in the determination of $\delta^{13}C$ using the Micromass IsoPrime elemental analyzer/isotope ratio mass spectrometer (EA/IRMS) at ANSTO. AMS ^{14}C measurements were performed using the STAR facility at ANSTO (Lawson et al. 2000; Fink et al. 2004) with a precision of 0.3–0.4%.

RESULTS

The U-series isotopic data from Collins et al. (2006) correspond with our AMS samples and are listed in Table 1. Results of the AMS ^{14}C analysis of our samples are shown in Table 2. The ΔR values for the Abrolhos Islands were calculated as the difference between the conventional ^{14}C age after correction for $\delta^{13}C$ (Stuiver and Polach 1977) and the model age (Marine09; Reimer et al. 2009) of U/Th age (cal yr BP). When calculating a mean ΔR correction, in order to be compatible with the online marine reservoir correction database (Reimer and Reimer 2001), the uncertainty associated with an error-weighted mean ΔR was chosen as the larger of the error of the mean and the standard deviation.

Coral samples 5000 U-Th cal BP and older did not pass the initial diagenetic screening and were rejected for AMS analysis. There were 3 anomalous results for samples OZO421 (HST1-PAVE), OZO422 (HST2-R3), and OZO426 (HST2-R9) with large, negative ΔR values (Table 2). Low $\delta^{13}C$ values present in these results may indicate exposure of sample material to freshwater (Southon et al. 2002) and subsequent diagenetic effects and/or geological alteration—a common occurrence in emergent pavements and coral rubble ridges exposed above sea level (Nothdurft et al. 2007; Nothdurft and Webb 2009). Nevertheless, low $\delta^{13}C$ values seem not to consistently affect the AMS results, suggesting that poor sample quality is related to other forms of degradation or a combination of both. The 3 anomalous samples showed advanced signs of microboring, the presence of gypsum, and some sponge chips at the surface margins. To avoid any diagenetically altered material, samples were carefully extracted from the central mass of the coral. Unfortunately, as large (>7 cm) sections of material were sampled, the possibility of contamination was greatly enhanced. These data were subsequently treated as outliers and were excluded from the mean ΔR calculation.

Loss or gain of initial uranium or daughter isotopes from the sample material, due to means other than radioactive decay, may lead to under- or overestimation (bias) of age determinations. Uranium levels (U ppm, Table 1) were elevated as may be expected for scleractinian corals. However, bias is still possible via U^{234} and ^{234}Th-Th^{230} mobility during late diagenesis of rubble ridges and porous coral framestone pavement exposed to cementation processes in semi-arid conditions (Zhu et al. 1993; Stirling et al. 1995; Hillaire-Marcel 2009). Though the XRD calcite screening threshold (with associated sensitivity-precision issues) was at 2% (Collins et al. 2006) instead of the recommended $\leq 1\%$ (Hughen et al. 2004; Robinson et al. 2006; Hillaire-Marcel 2009), the well-constrained errors observed in the U-Th dates suggest authigenic U-buffering against open system behavior and provide a good minimum age estimate.

Table 1 U-Th age and isotopic data for the corals used in AMS analysis. Errors are at 2 σ.

Lab code	Sample name[a]	Material	Genus	U (ppm)	^{232}Th (ppb)	$(^{230}$Th/^{232}Th)	$(^{230}$Th/^{238}U)	$(^{234}$U/^{238}U)	Uncorrected ^{230}Th age (yr)	Corrected ^{230}Th age (yr)	yr BP	Corrected initial $(^{234}$U/^{238}U)
OZO422	HST2-R3	Coral - storm ridge	Acropora	4.0358 ± 0.0019	1.114	502.52	0.04570 ± 0.0001	1.1528 ± 0.0014	4395 ± 12	4388 ± 13	4334 ± 13	1.1547 ± 0.0014
OZO421	HST1-PAVE	Coral - framestone	Porites	4.1419 ± 0.0035	0.99	505	0.03997 ± 0.0003	1.1519 ± 0.0025	3838 ± 33	3831 ± 33	3778 ± 33	1.1536 ± 0.0026
OZO423	HST2-R4	Coral - storm ridge	Acropora	3.6256 ± 0.0036	1.224	276.28	0.03074 ± 0.0001	1.1449 ± 0.0019	2959 ± 13	2950 ± 14	2896 ± 14	1.1461 ± 0.0019
OZO424	HST2-R7	Coral - storm ridge	Acropora	3.8162 ± 0.0037	3.04	95.44	0.02506 ± 0.0002	1.1511 ± 0.0028	2393 ± 17	2372 ± 20	2318 ± 20	1.1521 ± 0.0028
OZO425	HST2-R8	Coral - storm ridge	Acropora	3.8622 ± 0.0049	0.813	304.18	0.02109 ± 0.0002	1.1456 ± 0.0044	2020 ± 23	2015 ± 24	1961 ± 24	1.1464 ± 0.0044
OZO426	HST2-R9	Coral - storm ridge	Acropora	3.8194 ± 0.0023	1.681	102.87	0.01492 ± 0.0001	1.1506 ± 0.0014	1419 ± 14	1408 ± 15	1354 ± 15	1.1512 ± 0.0015
OZO427	HST2-R10	Coral - storm ridge	Acropora	3.5301 ± 0.0023	0.644	88.62	0.00533 ± 0.0001	1.1485 ± 0.0025	506 ± 7	501 ± 8	447 ± 8	1.1487 ± 0.0025
OZO428	HST2-R11	Coral - storm ridge	Acropora	3.9751 ± 0.0039	1.075	53.23	0.00474 ± 0.0001	1.1509 ± 0.002	449 ± 6	442 ± 7	388 ± 7	1.1511 ± 0.002
OZO429	HST2-R12	Coral - storm ridge	Acropora	3.8107 ± 0.0037	2.609	8.74	0.00197 ± 0?	1.1462 ± 0.0022	187 ± 3	170 ± 9	116 ± 9	1.1463 ± 0.0022

[a]Collins et al. 2006.

Table 2 ΔR values and AMS results for the Abrolhos Archipelago (Serventy Island). Year of growth or collection is 2002 for all samples.

Lab code	Sample name[a]	U/Th age[b] before 1950 (cal yr BP) 2σ	Species	δ^{13}C (‰)	Conventional ^{14}C age (yr BP) 1σ	Model age (yr BP) 1σ	ΔR (^{14}C yr) 1σ
OZO422	HST2-R3	4334 ± 13	*Acropora*	−1	4085 ± 35	4215 ± 5	−130 ± 35[c]
OZO421	HST1-PAVE	3778 ± 33	*Porites*	−0.2	3695 ± 35	3820 ± 5	−125 ± 35[c]
OZO423	HST2-R4	2896 ± 14	*Acropora*	−1.9	3155 ± 35	3125 ± 5	30 ± 35
OZO424	HST2-R7	2318 ± 20	*Acropora*	−1.9	2690 ± 35	2635 ± 25	55 ± 43
OZO425	HST2-R8	1961 ± 24	*Acropora*	−1.3	2410 ± 35	2345 ± 5	65 ± 35
OZO426	HST2-R9	1354 ± 15	*Acropora*	−1.6	1740 ± 30	1825 ± 5	−85 ± 30[c]
OZO427	HST2-R10	447 ± 8	*Acropora*	−1.5	800 ± 30	790 ± 5	10 ± 30
OZO428	HST2-R11	388 ± 7	*Acropora*	0.3	805 ± 30	730 ± 5	75 ± 30
OZO429	HST2-R12	116 ± 9	*Acropora*	−1	575 ± 30	490 ± 5	85 ± 30
						Mean	54 ± 30

[a]Collins et al. 2006.

[b]Calculated using U-Th ages for the Abrolhos Islands (see Table 1).

[c]These values were treated as outliers and removed from the calculation of the mean ΔR.

Table 3 Pre-bomb regional and local ΔR values of previous studies. NB: Values in bold type denote regional and subregional averages.

Region or location	Lat (°S)	Long (°E)	References	ΔR (yr)
Pelabuhanratu, Java (n = 1)	7	106.5	Southon et al. 2002	42 ± 70
Torres Strait, Australia (n = 3)	10	143	Gillespie 1977; Gillespie and Polach 1979	**50 ± 45**
Northern Territory, Australia (n = 1)	11.3	132.4	Southon et al. 2002	59 ± 40
Cocos (Keeling Islands), Australia (n = 12)	12	97	Hua et al. 2004 and references therein	**64 ± 15**
Kimberley region, Australia (n = 9)	16.4–18.1	122.2–123	Bowman 1985; O'Connor et al. 2010	**58 ± 19**
Abrolhos Islands (Serventy Island), Australia (n = 6)	28	114	This study	**54 ± 30**
Rowley Shoals, Kimberley Region, Australia (n = 13)	17.3	119.15	Unpublished data - AINSE grant 11/085	**60 ± 44**
Southwestern Australia	32.3–35	115.7–116.5	Gillespie 1977; Gillespie and Polach 1979; Bowman and Harvey 1983; Bowman 1985	**71 ± 47**
Java - Cocos Islands - Torres Strait - Kimberley (n = 26)	7–18.1	97–123	References within this table	**60 ± 38**

Summarized in Table 3 and displayed graphically in Figures 2 and 3 are individual location and subregional mean ΔR values for the contributing source waters of the Leeuwin Current (LC). The areal coverage extends from Indonesia and the Cocos (Keeling) Islands to Torres Strait and across the Northern Territory to the Kimberley region. These end-members represent waters from the Western Pacific Warm Pool, i.e. the Indonesian throughflow and South Equatorial Current including also the West Indian Ocean and Java Current. Included for comparison are some of the southernmost regions influenced by the LC. Data were selected from the Reimer and Reimer (2001) database best reflecting the ambient conditions of oceanic/coastal waters. On this basis, corals (where possible) were prioritized and only bivalve shells of suspension-feeding species were selected. To avoid the problem of "older" depleted carbon signals (i.e. from contamination by terrestrial/local sediments), shell specimens were excluded from selection if they belonged to a deposit-feeding genus. Similarly, any material was rejected if originating from estuarine/enclosed waters, were of unknown feeding habit or from uncertain location. The only exception to this rule was a specimen of the *Thais* genus from

Pelebuhanratu (Java), Indonesia. Though an intertidal carnivore, this genus feeds mostly on herbivorous filter feeders and is also capable of directly sourcing food in suspension (Lau and Leung 2004). As this species is used as a bioindicator for heavy metals and various other marine contaminants accumulated via their prey, it must be selected with caution. However, because this specimen was collected in 1910 (Southon et al. 2002) and from a location near a historically remote fishing village known for its strong surf conditions, it may be assumed that the local well-mixed surface marine conditions are reflected, as its low ΔR value suggests. All calculated regional-subregional mean ΔR values are also error-weighted according to Reimer and Reimer (2001), where the uncertainty associated with an error-weighted mean ΔR was chosen as the larger of the error of the mean and the standard deviation.

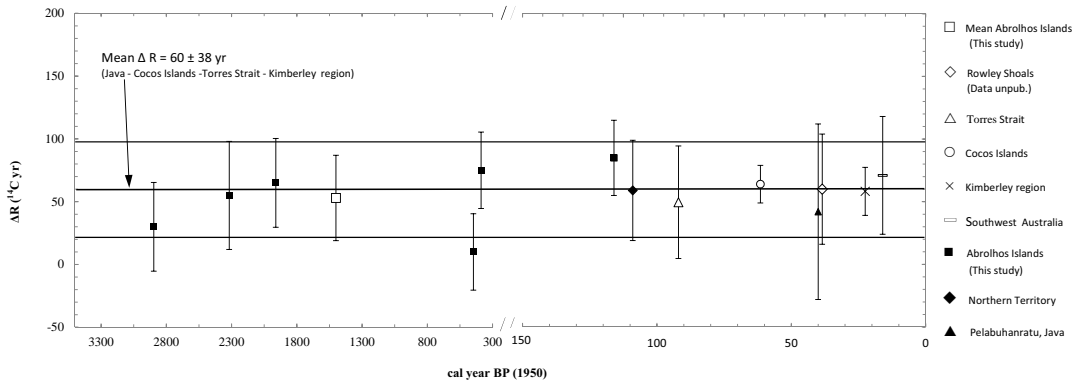Figure 2 Abrolhos ΔR values regional/subregional and individual ΔR values from Table 3. NB: For convenience of plotting, for data from studies using different dating methods, the dates of all ΔR values were converted to cal yr BP. The Abrolhos data were plotted against their U-Th dates (see Table 2).
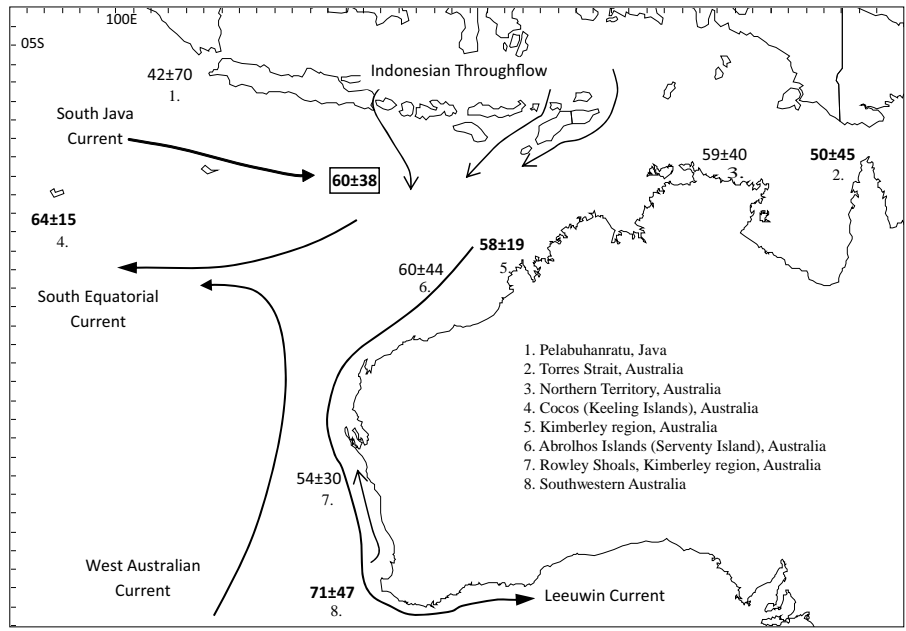

Figure 3 Location of ΔR values from Table 3 in relation to surface currents. Regional/subregional values are in bold type and the Leeuwin Current source-water mean is boxed.

The range of ΔR values for the Abrolhos Islands (10–85 yr) shows little variation about a mean ΔR of 54 ± 30 over 2896 ± 14 cal yr BP to historical times (Table 2). The ΔR values overlap with each other at 1σ uncertainty except for the fourth oldest and lowest ΔR value (10 ± 30 yr) with the 2 youngest samples (Figure 2). All individual Abrolhos ΔR values overlap within 1σ uncertainty of the regional ΔR of 60 ± 38 yr calculated for LC source waters from NW Australia to Java (Figure 2). Apart from the minor deviation of this single value, the Abrolhos data is well constrained by this "regional" mean and its ΔR uncertainty.

DISCUSSION

The mean ΔR value for the Abrolhos Islands (54 ± 30) at 28.45°S is similar to those found in tropical latitudes to the north. For example, the Rowley Shoals, with its central position relative to the Leeuwin Current (LC) source water end-members, has a mean ΔR of 60 ± 44 yr, relatively equal to the calculated average regional ΔR of 60 ± 38 yr (Figure 2). The close agreement of mean ΔR values from the Abrolhos and tropical regions indicates the long-range influence of the well-equilibrated source waters of the LC (Figure 2, Table 3).

The ΔR values at the Abrolhos are relatively stable over the last ~3000 yr. This apparent stability occurs during the rise in frequency of El Niño versus La Niña patterns up to the present time in the trend towards modern conditions seen in the western Pacific with more protracted and higher amplitude El Niño events (e.g. McGregor et al. 2008; Yu et al. 2010). Factors responsible for the stability of the ΔR signal at the Abrolhos over the last 3000 yr also include minimal tectonic activity, oceanographic simplicity, and a smooth, mostly linear sea-level decline after the Holocene highstand (Collins et al. 2006; Ulm 2006; Burr et al. 2009; Petchey et al. 2009). The coastline of Western Australia has not been subject to the degree of local tectonism, hydro-isostatic variability (hence sea-level fluctuation), upwelling and mixing of waters of very different physicochemical nature as experienced by many islands and coastlines of the East Indian Ocean and eastern and central Pacific regions (Southon et al. 2002; Burr et al. 2009; Petchey et al. 2009).

The position of the Abrolhos Islands just before the continental shelf slope is another factor governing the long-term stability of ΔR values. The LC core is anchored to the shelf break and is centered on the 200-m isobath just seaward of the Abrolhos where there is an acceleration of current speed (Woo et al. 2006). This ensures that the Abrolhos Islands remain predominantly within the LC influence (and periodically engulfed by the LC) for much longer than experienced seasonally on the mainland coast and is more protected from upwelling events (Hanson et al. 2005). A strong meridional pressure gradient generates the LC southward flow against the southwesterly wind fields, suppressing Ekman transport and creating a predominantly downwelling coast from Cape Range south to Cape Leeuwin (Pearce and Pattiaratchi 1999; Waite et al. 2007). However, as the LC travels from the lower-latitude source waters, transporting heat and low-salinity water southwards, heat loss and salinity increases with the gradual geostrophic contribution of Indian Ocean surface waters. Potentially, during this process the heat loss may induce convective mixing of underlying waters below the LC (Feng et al. 2007; Waite et al. 2007).

Upwelling along the Western Australian coastline is mostly sporadic and located around topographical features of the shelf break occurring mainly south of 34° latitude. Studies have indicated that the Capes Current (CC) flowing northward from 34°S may be a source of upwelled water transported inside the LC over nearshore shelf waters to just north of the Abrolhos Islands chain near Shark Bay (Woo et al. 2006; Feng et al. 2007; Waite et al. 2007). Analysis of the composition of this water reveal that the CC is a relatively weak source compared to upwelling experienced in other eastern boundary currents (Woo et al. 2006). Often, by the time any upwelled water from the south arrives

near the Abrolhos it is well mixed with LC waters that have been subject to geostrophic contribution of surface central Indian Ocean water (Woo et al. 2006; Feng et al. 2007). Some upwelling has been reported off-shelf near the Abrolhos Islands and coastward shelf waters (Hatcher et al. 1987; Hatcher 1991) and possibly Ningaloo Reef to the far north but not directly observed or measured recently (Kuhnert et al. 1999, 2000; Woo et al. 2006; Feng et al. 2007; Waite et al. 2007). Two mesoscale counter-rotating cold- and warm-core eddies generated from detached LC meanders operate seasonally near the Abrolhos Islands, centered between 111°37.7′E, 31°19.20′S and 110°1.5′E, 30°56.7′S. These systems have the potential to advect deeper waters or cause localized upwelling, though detailed analysis of their structure revealed no upwelling (Feng et al. 2007; Waite et al. 2007).

Most of the $\delta^{18}O$ variability observed in coral cores taken from the Abrolhos Islands has been attributed to the localized influence upon sea surface temperatures by spatial LC mobility around the archipelago and the associated lateral advection of nearby off-shelf and shelf surface waters as a result of local weather and LC strength. The magnitude of variability in these factors is influenced predominantly by ENSO fluctuations (Kuhnert et al. 1999). If upwelling were significant for this region, then it would be expected that the depleted ^{14}C signal be apparent in the coral skeletons. During the austral summer, when corals at this high latitude are experiencing peak growth rates, the LC is usually at its least strength due to the opposing prevailing wind gradient. The potential for upwelling events is enhanced by these conditions and also the probability for the capture of these events within the coral skeleton. In tandem, the highest proportion of several years' worth of skeletal coral growth sampled is represented by summer growth, as the very low water temperatures of the Abrolhos in winter are responsible for little skeletal development (Kuhnert et al. 1999). If upwelling events were frequent, considering the above factors, it would be expected that the sample bias would be towards higher ΔR values over most of the last 3000 cal yr BP (especially as El Niño events became more frequent). Instead, the ΔR values observed are low and more similar to latitudes up to 20° further to the north. However, apart from the study site, there remains the possibility of spatial heterogeneity of ΔR values across the Abrolhos Islands due to localized influences of channel bathymetry and differences in shelter or exposure to offshore conditions including the seasonal and ENSO-related pentadal vagaries of the LC.

In the future, it is likely that the climatic teleconnection of the subtropical/warm temperate Abrolhos Islands with waters of lower latitudes around Rowley Shoals and further north to Indonesia will become stronger and reinforce the similarity of ΔR values. Review of $\delta^{18}O$ coral band records longer than 20 to 350 yr around the world (including the Abrolhos and Ningaloo reefs) suggest a warming and freshening trend of surface waters as ENSO and monsoonal patterns intensify, including also the impact of anthropogenic warming (Kuhnert et al. 1999, 2000; Grottoli and Eakin 2007; Wyrwoll et al. 2009). The effect of this would possibly be to strengthen the LC and increase its influence, enhancing the trend towards more Pleistocene-like conditions (Wyrwoll et al. 2009). It is possible that as more Equator-ward corals become stressed as they approach bleaching point that there may be a divergence of similarity of ΔR between the Abrolhos and northern waters even though the climatic connection becomes stronger.

CONCLUSIONS

A local average marine reservoir ΔR correction value was determined for the Abrolhos Archipelago (54 ± 30 yr). The ΔR value agrees well with the regional average for N-NW Australia and Java of 60 ± 38 yr. The Abrolhos Islands ΔR values have remained stable over the last 2896 cal yr BP due to the increasing strength of the LC and minimal upwelling. Potential future trends may be a

strengthening of the teleconnection of the Abrolhos Islands to the climatic patterns of the equatorial Pacific via enhanced ENSO and global warming activity strengthening the LC. The possible effect on future ΔR values at the Abrolhos would be to maintain similar low values and a likely increase in stability. However, warming trends of global climate change may cause increasing dissimilarity of ΔR values to regions of the north due to the effects of increasing heat stress on coral communities in lower latitudes.

ACKNOWLEDGMENTS

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