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Recent changes in climate in the far western equatorial Pacific and their relationship to the Southern Oscillation; oxygen isotope records from massive corals, Papua New Guinea

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Abstract

The western equatorial Pacific warm pool is believed to be the source region of a substantial proportion of the world's inter-annual climate variability, including the globally significant El Niño-Southern Oscillation. Here new data are presented on changes in the climate of this region over the past 70 yr, based on analysis of the stable oxygen isotopic composition of annually banded, massive corals living on the north coast of Papua New Guinea. In this area, the coincidence of abundant and isotopically light rainfall with small seasonal and inter-annual changes in temperature means that the coral skeletal $\delta^{18}O$ record may be reliably interpreted in terms of rainfall variation. Since the abundance of rainfall is governed by the intensity of deep atmospheric convection, the coral records provide regionally significant, and regionally reproducible, climatic indices. In particular, the coral records indicate larger changes in the degree of coupling of the climate of the region with the Southern Oscillation than have hitherto been described. These changes in coupling, which accompany shifts in the dominant periodicities of inter-annual climate variation, are indicative of rapid and large-scale reorganisation of ocean–atmosphere interactions. The data indicate that from the 1920s to 1950s the western equatorial Pacific was less important in modulating Pacific and global inter-annual climatic variability than it has appeared to be subsequently. The reproducibility of the coral climate records, combined with the presence on the north coast of Papua New Guinea of extensive and well preserved raised reef terraces, leads to the potential for extending understanding of climatic variability in this globally important region back into the late Quaternary.

1. Introduction

Regions of particularly high sea surface temperature (SST) in the tropical oceans play an important role in modulating global climate. The influence of these so-called 'warm pools' is largely attributable to the substantial release of energy to the atmosphere from latent, sensible and radiative heating, and to the sensitivity of the energy transfer process to small changes in SST. Over the warm pools, energy transfer is accompanied by deep tropospheric convection, any variations in which can substantially perturb the

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general circulation of the atmosphere. The sensitivity of the system arises from the non-linear relationship between SST and latent heat flux, such that changes in warm-pool SST are likely to induce much larger global climatic responses than equivalent SST variations in middle or high latitudes [1]. In this context, the western equatorial Pacific warm pool is particularly significant, having the warmest sea surface temperatures, highest precipitation and the greatest latent heat transfer to the atmosphere of any open ocean area on Earth. The region is now generally regarded as the source area for a substantial proportion of the Earth's inter-annual climate variability, including the globally significant El Niño-Southern Oscillation (ENSO) phenomenon [1–3].

The term Southern Oscillation (SO) describes an oscillation in sea level atmospheric pressure between the central Pacific and the north Australian/Indonesian region which has a period of between 2 and 7 yr. Relatively high atmospheric pressure in the central Pacific and low pressure over Indonesia is indicative of the La Niña phase of the Southern Oscillation. The opposite pressure regime describes the El Niño phase. The SO is one manifestation of a periodic, large-scale reorganisation of atmospheric and ocean surface circulation. Although centred in the Pacific region, the vigorous atmospheric teleconnections result in a near-global influence on inter-annual climate variability [4–7]. For example, such diverse climatic 'events' as drought in parts of Africa, the Indian subcontinent and Australia, cyclones in the central Pacific, exceptionally heavy rainfall in north-western South America, and severe winter storms in California and the USA Gulf States have been attributed to the El Niño phase of the SO.

Over the past decade, there has been concerted interdisciplinary and international research effort to investigate the origins of the SO; for example, through the Tropical Ocean–Global Atmosphere (TOGA) Program [1,8]. One of the prime motivations behind this work has been the desire to develop a predictive capability for modelling the likely course of future climate on time scales ranging from seasons to centuries. This research has made major advances in identification of the critical components of the equatorial Pacific climate system. However, the mechanisms responsible for observed variability in intensity, spatial manifestation and periodicity of SO events remain poorly understood. One significant obstacle to progress in this field has been the scarcity of long (more than a few decades) instrumental records of climate in the Pacific region on which to base or test conceptual and numerical models of ocean–atmosphere interactions. This has lead to the search for alternative, natural, archives of climatic information from the region. In this respect, the aragonitic skeletons of massive reef-building corals have been shown to hold considerable potential.

As they grow, corals lock into the structure and chemical composition of their skeleton information relating to the temperature and composition of the surrounding sea water. This attribute, combined with rapid growth rates, the presence of annual bands and the longevity of individual colonies, make corals excellent sources of palaeoenvironmental information, capable of yielding records with a temporal resolution of a few months over several centuries. The stable oxygen isotopic composition of the aragonitic skeletons has proved to be one of the most useful of the chemical tracers; with the caveat that, as with other geochemical climate proxies, care must be taken to consider the possible effects of large changes in coral growth rate [9–11]. In regions where water oxygen isotopic composition remains reasonably constant, the skeletal $\delta^{18}O$ may be converted into an estimate of past SST, with a resolution down to about 0.5°C [9]. Alternatively, in regions where there is little variation in SST, the skeletal $\delta^{18}O$ may be converted into a record of changes in seawater oxygen isotope composition, usually related to variations in the input of isotopically light rainfall and/or run-off [12,13]. Recently, these two palaeoenvironmental facets of coral stable oxygen isotope composition have been exploited within the Pacific region to produce long (a few decades–a few centuries) proxy records of inter-annual climate variability from the Galápagos Islands in the far eastern equatorial Pacific [14,15], and from Tarawa Atoll in western-central equatorial Pacific [12,13]. However, despite its apparently central role as the source region for much of the observed climatic variability, the far western equatorial Pacific remains poorly served by instrumental or proxy climate records extending beyond the past few decades. Here, we present evidence for hitherto undescribed climatic variations of the far western equatorial Pacific region over the past 70 yr based on analysis of massive corals from the north coast of Papua New Guinea.

2. Field area

2.1. Regional attributes

Samples for this study were collected from near Madang (5° 13' S; 145° 49'E) and at Laing Island (4° 9' S; 144° 53' E) on the Bismark Sea coast of Papua New Guinea (Fig. 1). Before describing the sampling locations in more detail, we review the factors which combine to make this region a suitable area for coral palaeoclimatic study.

2.1.1. Rainfall

The climate of the Bismark Sea coast of Papua New Guinea is warm and humid [16]. Coastal rainfall averages about 3,000 mm/yr, with a relatively dry season (about 100–200 mm/month) from June to September [17]. There is an annual excess of rainfall over evaporation of between about 1,000 mm and 1,500 mm [18–20]. The rainfall is isotopically light with respect to oxygen, averaging ca. $-8‰$ $\delta^{18}O_{SMOW}$ at Madang [21]. Inter-annual variations in rainfall for the whole Indonesia/Papua New Guinea/northern Australia area are correlated with the SO, with relative drought prevailing during the El Niño phase [5]. This relationship also holds true for individual coastal stations in Papua New Guinea, including Madang [17].

The combination of abundant isotopically light rainfall and the relationship between rainfall and large-scale reorganisation of atmosphere–ocean circulation, leads to the possibility of using the stable oxygen isotopic composition of corals in the region as a proxy for rainfall and, hence, as a regionally significant palaeoclimatic indicator.

2.1.2. SST and surface mixed layer

Mean annual SST in the Bismark Sea is generally close to 29°C, with an annual range of 0.5–1.5°C in monthly means (COADS SST and station data [Climatic Research Unit, University of East Anglia, pers. commun., 1992]). The period of warmest SST broadly coincides with the southern hemisphere summer, which is also the rainy season. Inter-annual variations in SST are generally less than 0.5°C but show a weak correlation with the SO, the cooler temperatures prevailing during the El Niño phase [2]. These small variations in SST suggest that temperature-related inter-annual variations in coral $\delta^{18}O$ will be small ($< 0.1‰$ $\delta^{18}O$).

The surface salinity of the Bismark Sea region is relatively low, averaging about 34.25–34.5‰ [18]. Recent oceanographic studies have revealed that the surface mixed layer throughout the western equato-

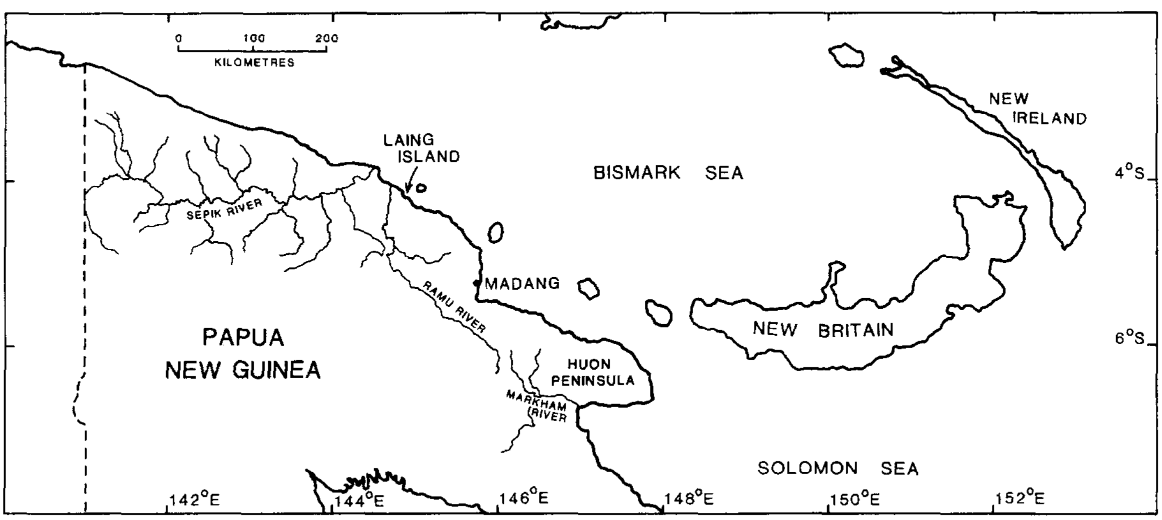


Fig. 1. Location of coral sampling sites at Madang and Laing Island.

rial Pacific (including the Bismark Sea) is much shallower than previously thought, possibly averaging about 30 m [22,23]. It is believed that this relatively low salinity and buoyant mixed layer owes its origins to the abundant rainfall and generally low wind speeds. These observations suggest that the rainfall is initially mixed within only a relatively small volume of sea water. This, in turn, suggests that the rainfall, with its light isotopic composition, is likely to have a large impact on the $\delta^{18}O$ composition of the sea water in which reef-building corals live.

2.1.3. Corals

Coral reefs fringe much of the north coast of the Papua New Guinea mainland and offshore islands [24]. Massive *Porites* corals are a major component of these reefs, occurring in colonies up to about 2 m in diameter, and growing at rates of between 8 and 16 mm/yr. *Porites* is one of the most commonly used genera of coral for palaeoenvironmental reconstruction in the Indo-Pacific region, due to its ubiquity, longevity, rapid growth rate and apparent reliability of isotopic records.

2.2. Sampling sites

2.2.1. Madang

The main area sampled for corals in the present study was the Madang Lagoon, immediately north of the town of Madang (Fig. 1). The lagoon is 15 km long, up to 4 km wide, has an average mid-lagoon depth of about 30 m and is bounded on its seaward side by a living barrier reef, whose top lies about 1 m below mean low water, spring tide level. Although the term lagoon often carries a connotation of an enclosed body of water, this is not the case at Madang. The lagoon is flushed by strong tidal and wind-driven currents, which flow through 3 deep passes as well as over the reef top. The passes are as deep as the lagoon itself ensuring no stagnation of water. In addition, a temperature monitoring programme (3 underwater temperature loggers recording temperature every 2 h since 1991; precision better than 0.2°C; accuracy better than 0.5°C) has revealed no significant differences in temperature (absolute or variability) either across the lagoon, or between the lagoon and the top of the barrier reef. In addition, we can be certain from documentary records (annual company reports from periods of colonisation) that there have been no major tectonic movements or changes in land use this century which could have affected lagoon circulation. Within the lagoon, living reefs occur as an almost continuous fringing reef along the mainland shore (except in the immediate vicinity of stream mouths), and as patch reefs reaching up to within a few metres of sea level. The lagoon has an area of 46 km^2, and an additional terrestrial catchment area of 240 km^2. Surface freshwater run-off into the lagoon is restricted to a few small streams (< 5 m wide).

2.2.2. Laing Island

One of the coral cores from this study was collected from a reef (Duranguit Reef) situated 1 km north of Laing Island, within the 7 km diameter embayment of Hansa Bay (Fig. 1). This reef is totally submerged, reaching to within 3 m of sea level. It is situated 2.5 km from the mainland shore and lies southeast of the mouths of two large rivers, the Sepik River (50 km distant) and the Ramu River (25 km distant).

3. Materials and methods

3.1. Sample collection

Cores were collected from 4 corals for this study. All of these were massive *Porites* corals, with brown/yellow pigmentation, between 1.5 and 2.0 m diameter, and with similar massive growth form. Corals Madang-1 and Madang-2 were situated at the break in slope on the seaward edge of a 50 m wide reef flat. These colonies were about 100 m apart. The tops of both of these colonies reached to about 0.2 m below low spring tide level. Coral Madang-3 was on the top of a mid-lagoon patch reef (Deplik Tabat) where the water was 5 m deep. Coral Laing-1 was in 3 m water depth.

Five coral cores are discussed in this study, two from coral Madang-1 (labelled 1a and 1b) and one each from the other 3 colonies. These cores were retrieved in 1991 and 1992. Core Madang-1a has been analyzed to produce a 69 yr long stable oxygen isotope record. Cores Madang-1b, Madang-2, Madang-3 and Laing-1 were collected in order to assess within-colony, between-colony, between-site (local) and between-site (regional) reproducibility of stable oxygen isotope records over a 10 yr growth period (1981–1991).

3.2. Subsampling and date allocation

There are a variety of different types of annual banding in coral skeletons which may be used to provide a chronology; for example: density banding, fluorescent banding and seasonality in $\delta^{13}C$ and $\delta^{18}O$ records. The best choice of parameter(s) depends upon the specific reef site and corals concerned. In our corals we used a combination of distinctive skeletal fluorescent banding (visible under illumination with longwave UV light; Fig. 2) to provide an initial chronology. The accuracy of this chronology was subsequently verified on a year-by-year basis by consideration of the annual cycle in skeletal $\delta^{18}O$ (Fig. 3). The full procedure was as follows: We used vital staining with Alizarin Red S to ascertain the periodicity and timing of fluorescent band deposition in our Papua New Guinea field area. Cores collected from previously stained coral heads confirmed that major bright–dull fluorescent band couplets represent approximately 1 year's growth, and that bright band deposition occurs during the wet season, and dull band deposition during the dry season. This pattern conforms with the situation on Australia's Great Barrier Reef, where similar banding has been shown to result from the incorporation into the skeleton of terrestrially derived humic acids during wet season rainfall and run-off [25,26]. We opted for a sampling frequency of 4 samples/annual growth increment to facilitate high-resolution analysis of inter-annual variations (our principal interest), whilst maintaining some ability to investigate seasonality. This sampling was achieved by cutting each annual growth increment (as indicated by fluorescent banding) into 4 equally sized samples using a hand-held miniature saw (blade thickness of 0.2 mm). Following isotopic analysis of these samples (see below), we constructed an independent chronology on the basis of the isotopic and growth rate data. In our corals, the annual cycle in skeletal $\delta^{18}O$ appears to be the most reliable of the chemical chronometers, with the isotopically heavy skeleton being deposited during the dry, cool season (June–September) and isotopically light skeleton during the warm and wet season (Fig. 3A). A double isotopic minima in some years reflects the usual bimodal distribution of wet season rainfall in this region [16]. Skeletal linear extension rate (measured parallel to the local growth axis) was used to constrain identification of annual growth increments on the basis that this rate is extremely unlikely to change by more than 50% in succeeding years. Skeletal $\delta^{13}C$ has a strong annual cycle in most years but in other years this cycle is much subdued and so can only be used as a chronometer with caution. The independent chronology based on skeletal $\delta^{18}O$ agreed ($\pm$ a few months) with the initial time-scale based on skeletal fluorescent banding.

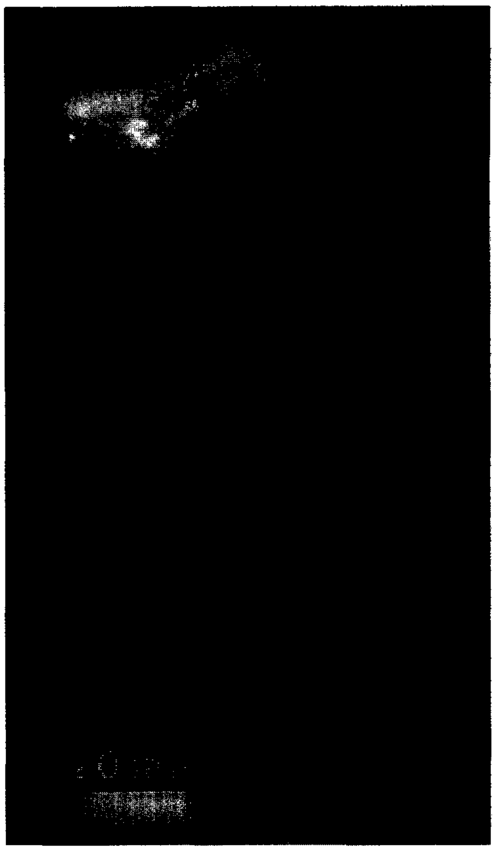


Fig. 2. Photograph of a slab of core from coral Madang-3 taken under illumination with longwave ultraviolet light to reveal fluorescent banding. Each major bright + dull band couplet represents approximately 1 year's growth (a single year's growth is represented by the increment between the tips of the two black arrows).

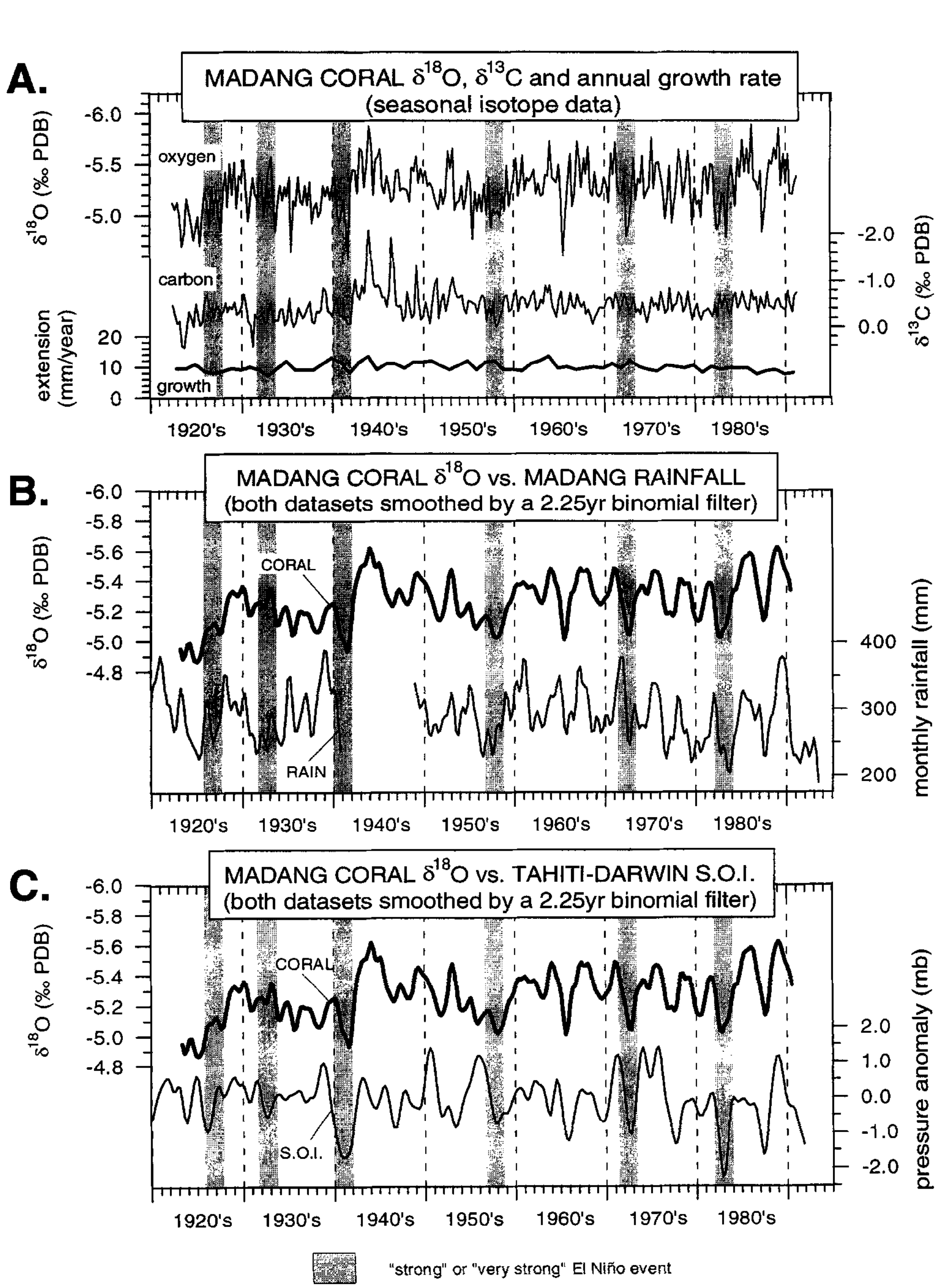


Fig. 3. The stable oxygen isotope record from coral core Madang-1a compared with some climatic indices. Years when there were 'strong' or 'very strong' El Niño events off the Peruvian coast [28] are indicated by stipple. (A) Unsmoothed Madang-1a $\delta^{18}O$ and $\delta^{13}C$ records (seasonal resolution) and Madang-1a annual linear extension rate. Although skeletal $\delta^{13}C$ displays a strong annual cycle in most years, records of inter-annual variability are often not reproducible between corals. Therefore, we do not consider $\delta^{13}C$ further in this paper. (B) Madang-1a $\delta^{18}O$ record and Madang rainfall. Both records have been smoothed by a 9-point, 2.25 yr, binomial filter. The gap in the rainfall curve in the 1940s reflects missing data. (C) Madang-1a $\delta^{18}O$ record and the Tahiti–Darwin Southern Oscillation Index. Both records have been smoothed by a 9-point, 2.25 yr binomial filter.

Date allocation for each sample was made by assuming that the bright-to-dull fluorescent band transition (which coincides with the isotopically light-to-isotopically heavy $\delta^{18}O$ transition to within one sample in all years) represented end May–start June (supported by the results of the Alizarin stain experiments and by consideration of local rainfall records) and that each increment represented 3 months' growth. Since the precise timing of fluorescent band deposition may vary from year to year, we must expect our date allocations to contain a small degree of error, possibly up to ± 2 months.

3.3. Isotopic analysis

Samples from the top of the coral cores were treated with commercial bleach to remove coral tissue prior to isotopic analysis. All other samples were untreated. The cut coral pieces were crushed in an agate mortar and pestle immediately prior to analysis of 0.5–1.0 mg subsamples. Subsamples were reacted with 100% orthophosphoric acid at 90°C in an automatic carbonate preparation system. The resulting CO_2 was analyzed on a VG Isogas Prism mass spectrometer. The standard deviation for 50 analyses of a coral powder (COR1B) run as a sample over a 6 month period was ± 0.076‰ for $\delta^{18}O$. All carbonate isotopic values are quoted relative to PDB.

4. Results

The results of our study are split into several subsections. We start by making a visual inspection of the 69 yr long record from core Madang-1a, making comparisons between it and local and regional climatic indices. Before analysing these coral and climate records in more detail, we discuss evidence for the reproducibility of the coral record on a variety of spatial scales. Statistical treatment of the data begins with linear regression analysis. Spectral analysis is then used to investigate variations and relationships in the frequency domain. Finally, aspects of seasonality are explored further using the time series technique of complex demodulation. Isotope and chronology data are available, on request, from the senior author.

4.1. General character of the 69 year coral record

The $\delta^{18}O$ record from core Madang-1a displays marked seasonality, with an annual range averaging 0.3–0.4‰ and reaching up to 0.8‰ (Fig. 3A). Light isotopic values correspond with the rainy (and slightly warmer) season. Once the effects of seasonality are removed from the coral $\delta^{18}O$ record by smoothing with a 2.25 yr (9-point) binomial filter (Fig. 3B), the nature and extent of inter-annual variability become more apparent. From about 1960 onwards, these variations appear to be fairly regularly spaced in time, with a periodicity of about 4 yr and an amplitude of 0.3–0.5‰. Prior to 1960, the variability is less regular, and the skeleton precipitated in the 1920s and 1930s is slightly isotopically heavier (ca. 0.1–0.2‰ $\delta^{18}O$) than that deposited subsequently. The coral had a mean growth rate (linear extension) of 9.9 mm/yr, with no significant trends through time (Fig. 3A). In corals growing faster than about 4–5 mm/yr, variations in coral growth rate are believed to have only a small effect on the kinetic fractionation of oxygen isotopes into the coral skeleton [9,11]. This, combined with the lack of any significant correlation between coral growth rate and skeletal $\delta^{18}O$ in our records (Fig. 3A), leads us to conclude that the relatively small inter-annual variations in growth of our coral had little influence on the observed variations in skeletal $\delta^{18}O$.

There is a good visual match between the smoothed coral record and the similarly smoothed Madang rainfall record for the period from the late 1950s onwards (Fig. 3B). Periods of low rainfall coincide with deposition of isotopically heavy skeleton. This is the relationship that was anticipated from consideration of the regional climate and oceanography. The match between rainfall and coral $\delta^{18}O$ in the 1920s and 1930s is not so good, although there are still similarities between the two parameters.

In Fig. 3C, the smoothed coral isotopic record is compared with the C.A.C. version of the Tahiti–Darwin Southern Oscillation Index (T–D SOI) [27]. This Tahiti–Darwin atmospheric pressure index is one of the most commonly used, and easy to update, instrumental measures of SO activity. Negative values of the index indicate relatively low pressure in Tahiti and high pressure in Darwin, and this state is diagnostic of the El Niño phase of the SO. The coral

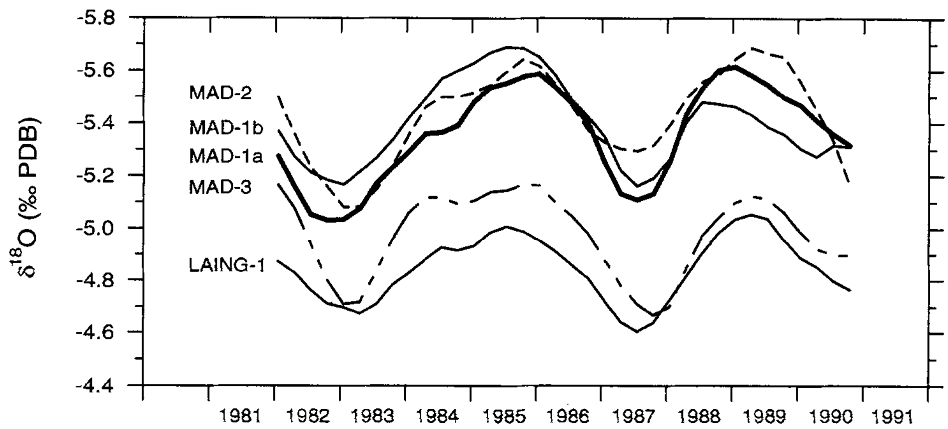


Fig. 4. A comparison of stable oxygen isotope records from 5 *Porites* coral cores over the period 1981–1991. Seasonal data from each core has been smoothed by taking an annual running mean to remove the effects of seasonality. Cores Madang-1a and Madang-1b (MAD-1a and MAD-1b) are from the same coral colony; core Madang-2 (MAD-2) comes from a colony 100 m distant from Madang-1; core Madang-3 (MAD-3) comes from a coral situated 3.5 km distant from corals Madang-1 and Madang-2; and core Laing-1 (LAING-1) come from a coral 150 km northwest of Madang.

$\delta^{18}O$ and the T–D SOI records show strong similarities, with large negative values of the index coinciding with isotopically heavy coral skeleton. Once again, the correlation between the two records appears to be best from the late 1950s onwards. Years of 'strong' or 'very strong' occurrence of the warm El Niño current off the coast of Peru [28] tend to coincide with large negative values of the T–D SOI, low rainfall at Madang, and the deposition of isotopically heavy coral skeleton.

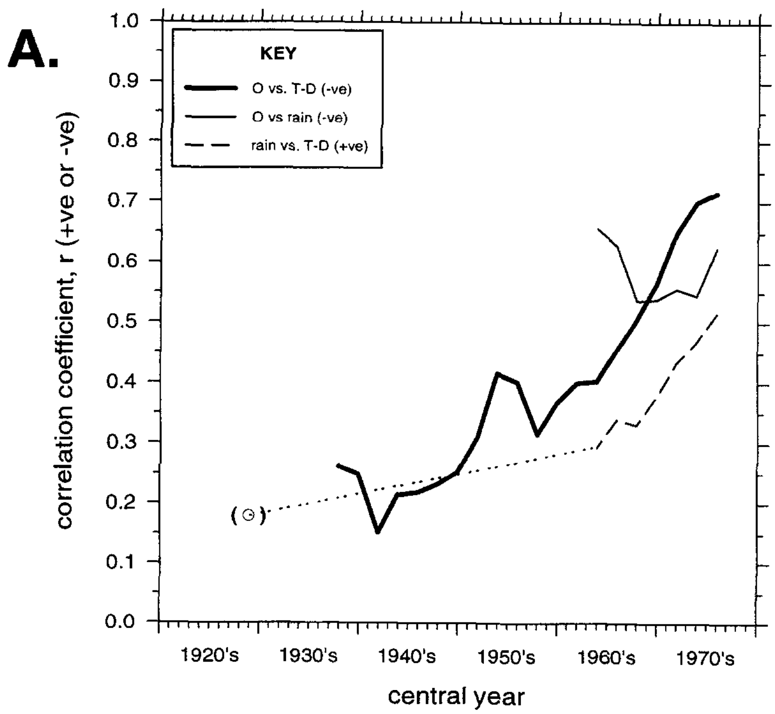
In summary, visual inspection of the coral record supports the hypothesis that the isotopic composition of corals from the region can reveal information on past rainfall and SO activity, but there is some indication of a change in the coral–rainfall–SO relationships around 1960.

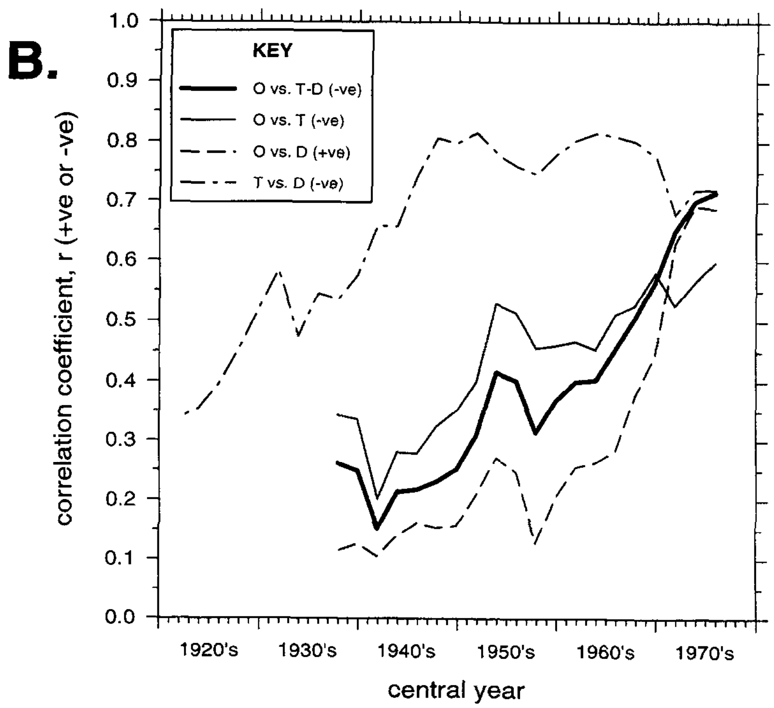
4.2. Reproducibility of coral records

Before exploring these coral–climate relationships in more quantitative terms, it is appropriate to consider the degree of reproducibility of the coral records, both within and between corals and within and between sites. Fig. 4 presents the results of analysis of 5 cores over the same 10 yr growth increment, June 1981–May 1991. In the figure, the data have been smoothed by taking an annual (4 sample) running mean to remove the effects of seasonality. The similarity of the inter-annual signals in corals separated by up to 150 km (e.g., a correlation coefficient, r value, of 0.919 for linear regression of annual mean $\delta^{18}O$ between corals Laing-1 and Madang-3) indicates the dominant regional climatic control on skeletal $\delta^{18}O$. In addition, since any slight inaccuracies in subsampling would translate to decreased correlation between the records, the true degree of correlation between the corals will almost certainly be higher than indicated by Fig. 4.

Similarity in the timing and magnitude of inter-annual variations between all the records, including that of Laing-1, which is relatively close to the mouths of the Sepik and Ramu rivers, argues for a dominant direct rainfall control on inter-annual $\delta^{18}O$ changes, with the effects of variations in the intensity

Fig. 5. Sliding correlation plots generated by performing linear regression analysis on overlapping 30 yr time slices, with a 2 yr step between adjacent slices. The correlation value has been plotted opposite the mid-point of each time slice. All analyses were performed on annual mean data, with the year running from start December to end November. O = $\delta^{18}O$ record from coral core Madang-1a; T–D = Tahiti–Darwin Southern Oscillation Index; rain = Madang rainfall; T = Tahiti sea-level atmospheric pressure; D = Darwin sea level atmospheric pressure. The point in parenthesis in (A) is for a 24 yr long time slice.


A.
KEY
O vs. T-D (-ve)
O vs rain (-ve)
rain vs. T-D (+ve)
(○)
correlation coefficient, r (+ve or -ve)
1.0
0.9
0.8
0.7
0.6
0.5
0.4
0.3
0.2
0.1
0.0
1920's
1930's
1940's
1950's
1960's
1970's
central year


B.
KEY
O vs. T-D (-ve)
O vs. T (-ve)
O vs. D (+ve)
T vs. D (-ve)
correlation coefficient, r (+ve or -ve)
1.0
0.9
0.8
0.7
0.6
0.5
0.4
0.3
0.2
0.1
0.0
1920's
1930's
1940's
1950's
1960's
1970's
central year

and distribution of river run-off playing a subsidiary role. However, differences in absolute $\delta^{18}O$ composition of corals at Madang most probably reflect the different locations with respect to freshwater run-off (coral Madang-3 was the furthest from the mainland and was isotopically the heaviest).

4.3. Regression analysis of coral and climate data

Inspection of Fig. 3 suggested a correlation between the coral record and both local (Madang) rainfall and the Tahiti–Darwin pressure index. However, it was also apparent that the degree of correlation was not constant through the time interval from 1922 until 1992. Therefore, to investigate these relationships in more detail, linear regression analyses were performed between coral $\delta^{18}O$, Madang rainfall, the T–D SOI, Tahiti sea level pressure and Darwin sea level pressure in a series of overlapping 30 yr time slices, with a 2 yr step between slices. The results are illustrated in the form of sliding correlation plots in Fig. 5. The main points to emerge are as follows:

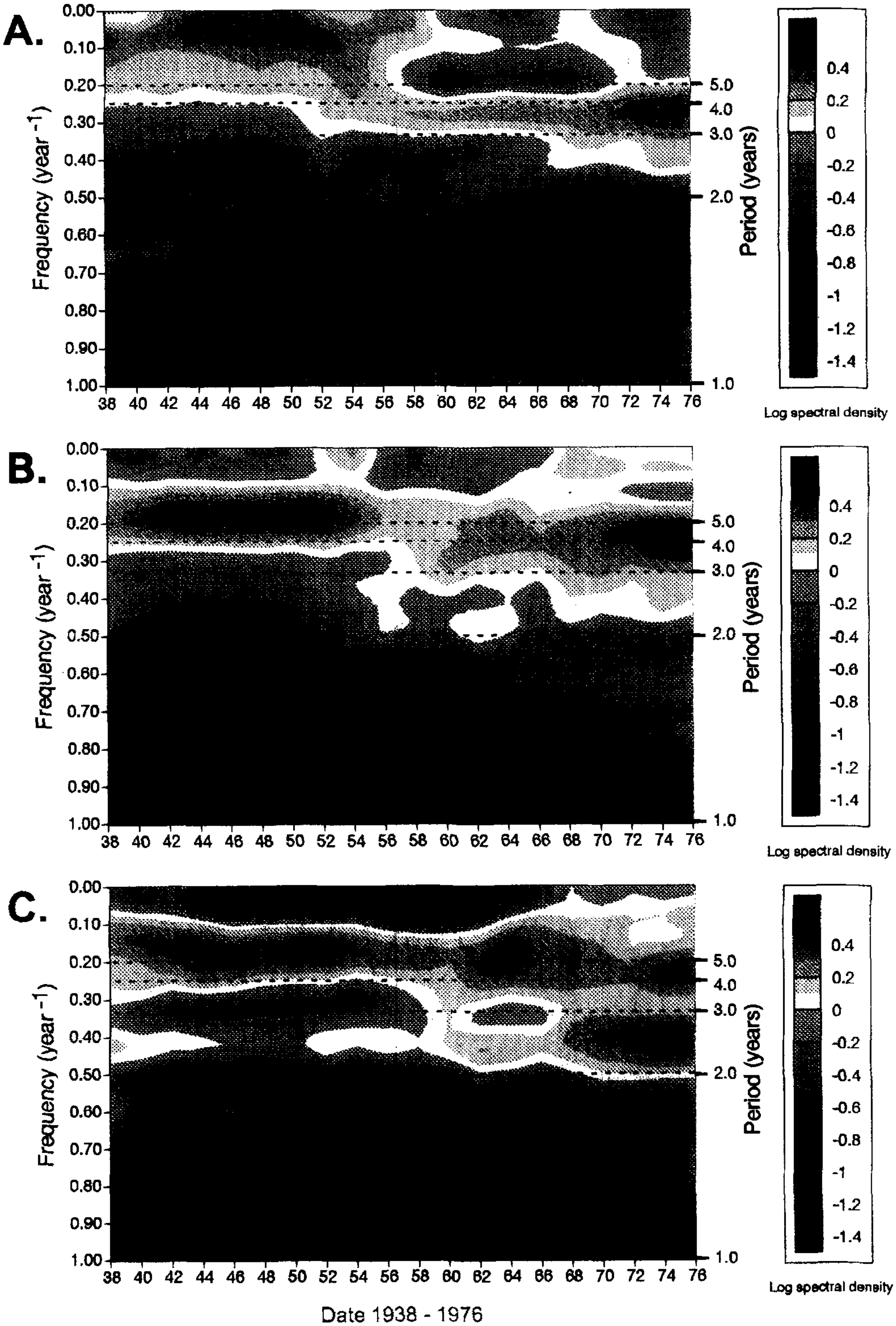
(a) The coral $\delta^{18}O$ record has a higher degree of correlation with the T–D SOI than does the Madang rainfall for all time slices. Although this may be, in part, an artefact caused by slight differences in the phase relationships between the records (see results of spectral analysis discussed below), there are two environmental factors which could contribute to the coral record being a better index of SO activity than the rainfall record. Firstly, the coral record may integrate rainfall over a larger, and therefore more representative, area than does a rain gauge. Secondly, inter-annual temperature variations are related to the SO activity, and, although relatively small, these will tend to magnify the SO signal in the coral $\delta^{18}O$. During the El Niño phase of the oscillation, SST tends to be cooler and rainfall less than on average, both of which will push the coral isotopic composition towards heavier values.

(b) Going back in time from the most recent 30 yr time slice (1961–1990), there is a substantial decrease in correlation between the coral $\delta^{18}O$ record and the T–D SOI. A similar decrease in correlation back through time is observed between the Madang rainfall record and the T–D SOI, although the data gap in rainfall records in the 1940s means that the continuance of this trend cannot be confirmed for the earlier part of the time period. This suggests a substantial change in the degree of coupling of the climate of the region with SO activity prior to about 1960, thus supporting the initial visual assessment of the records. The relatively constant degree of correlation between the coral $\delta^{18}O$ and Madang rainfall records over the period for which the regression analyses could be performed supports the idea that the principle change is in the correlation between the regional climate and the T–D SOI, rather than being a change in the relationship between the coral record and the climate.

(c) When the coral $\delta^{18}O$ record is compared separately with Tahiti pressure and Darwin pressure, both show decreasing correlation back through time, the decrease in correlation being particularly large and rapid between the coral $\delta^{18}O$ and Darwin pressure. This suggests a decoupling of the regional climate of Madang from that of both central equatorial Pacific and north Australian regions in the 1920s–1950s.

(d) The degree of correlation between Tahiti pressure and Darwin pressure is reasonably constant for the 30 yr time slices centred on the mid-1940s to mid-1970s but decreases prior to that time. The lack of shifts in correlation over the later part of the record suggests that the changing relationship be-

Fig. 6. Evolutionary spectra for: (A) Madang coral $\delta^{18}O$; (B) Darwin sea level atmospheric pressure; and (C) Tahiti sea-level atmospheric pressure. Each figure has been generated by performing spectral analysis (autocovariance) on overlapping 30 yr time slices, with a step length of 2 yr. Prior to analysis, each data set was standardised to remove seasonality and to generate a variance close to 1. Results of each analysis were plotted opposite the mid-point of the time slice, then the whole contoured. The figures illustrate concentrations of variance (in units of log spectral density; high positive values indicating strong concentration) as a function of frequency (vertical profiles across the diagrams), and changes in these relationships through time (moving horizontally across the diagrams). (The 80% confidence limits of log spectral density are 0.35 units above and 0.22 units below the indicated values; 40 lags were used for all analyses.)


A.
B.
C.
Frequency (year^{-1})
0.00
0.10
0.20
0.30
0.40
0.50
0.60
0.70
0.80
0.90
1.00
Period (years)
5.0
4.0
3.0
2.0
1.0
38 40 42 44 46 48 50 52 54 56 58 60 62 64 66 68 70 72 74 76
0.4
0.2
0
-0.2
-0.4
-0.6
-0.8
-1
-1.2
-1.4
Log spectral density
Date 1938 - 1976

tween the coral record and the T–D SOI cannot be simply attributed to a general decoupling of climate between the central equatorial Pacific and the north Australian regions.

4.4. Spectral analysis

Spectral analysis and cross-spectral analysis were used to explore relationships and changes through time in the frequency domain. We used the ARAND program (courtesy of Dr. Philip Howell, Brown University), which is based on the Blackman–Tukey FFT method. Due to the changing character of the coral isotope and climate records through time, spectral analysis of the whole 69 yr time interval proved to be unsatisfactory; changes in dominant periodicity were lost in a general broadening of variance peaks. Therefore, here we present the results of spectral analysis performed on overlapping 30 yr time slices, with a 2 yr step between each slice. Since we are primarily interested in inter-annual variations, prior to analysis we standardised the coral $\delta^{18}O$ and Tahiti and Darwin pressure records to a 1951–1980 reference period to remove the effects of seasonality (a similar method to that used in calculating the T–D SOI [27]).

Evolutionary spectra for coral $\delta^{18}O$, Tahiti pressure and Darwin pressure are presented in Fig. 6. The most striking feature of the coral $\delta^{18}O$ spectra is the change in dominant periodicity in the mid–late

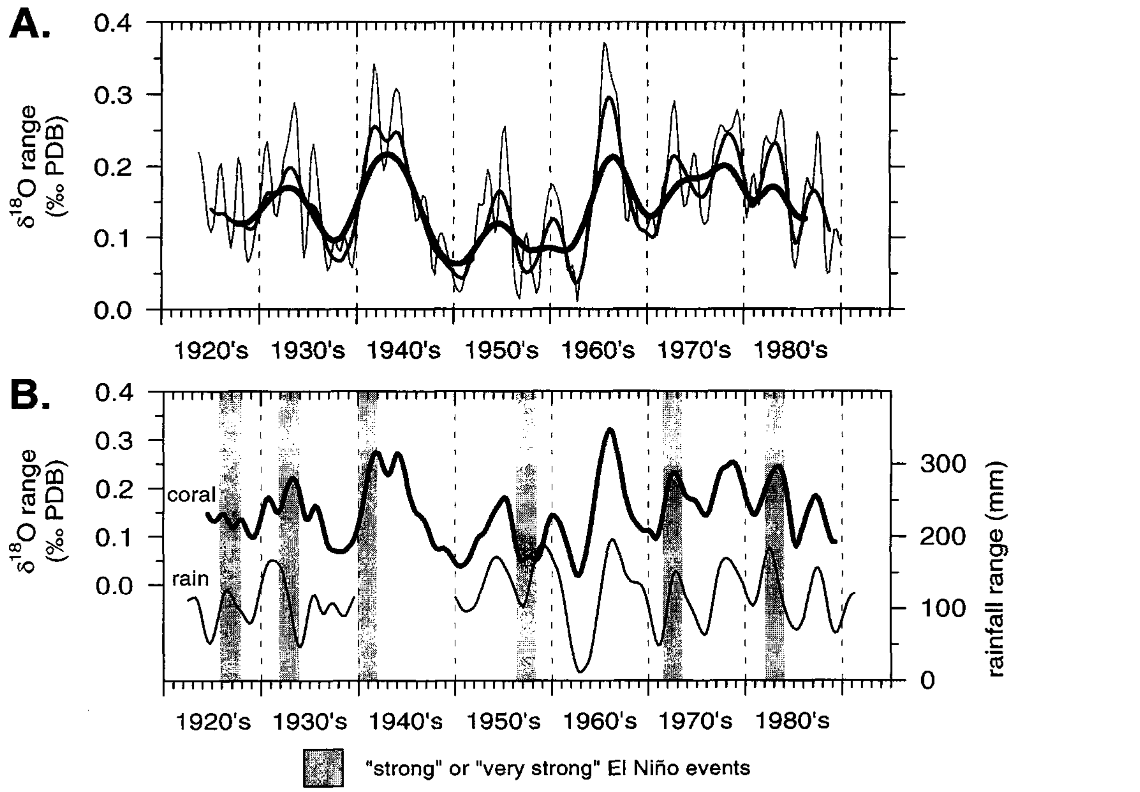


Fig. 7. Variations in the amplitude of the annual cycle measured by complex demodulation.
(A) Amplitude of the annual cycle in the $\delta^{18}O$ record from coral core Madang-1a calculated using 3 different levels of filtering.
(B) Amplitude of the annual cycle in coral $\delta^{18}O$ compared with the amplitude of the annual cycle in Madang monthly rainfall, both calculated by complex demodulation analysis using an intermediate degree of filtering (ca. 5 yr window length). The gap in the rainfall curve in the 1940s reflects missing data. Years when there were 'strong' or 'very strong' El Niño events off the Peruvian coast [28] are indicated by stipple.

1950s. Prior to that time, variance is long-period and diffuse; after that time variance becomes increasingly concentrated on a 3–5 yr periodicity, with a peak at about 3.5 yr. There is little evidence of a biennial periodicity, with the possible exception of a shoulder on the 3–5 yr peak in the later part of the record. Darwin pressure shows a strong concentration of variance at a 5–6 yr period for time slices centred in the 1940s to mid-1950s. After that time, variance becomes increasing concentrated at a 3–5 yr period. The evolutionary spectrum for Tahiti pressure also shows a change in character centred in the mid-late 1950s. For 30 yr time slices centred on about 1960 or later, there is concentration of variance at 2.5 yr and at 4–5 yr. The 2.5 yr periodicity appears to become stronger through time. For time slices centred before the late 1950s, the 2.5 yr periodicity is substantially weaker, and the 4–5 yr periodicity has been replaced by a concentration of variance at a 5–6 yr period.

All of the evolutionary spectra indicate a major change in the character of inter-annual variation centred on the mid–late 1950s. The main differences and similarities are as follows:

1. In the period prior to the late 1950s, the coral record displays a very diffuse concentration of variance at long periods, whereas both the Tahiti and Darwin pressure records show a focusing of variance at a period of about 5–6 yr.
2. All of the records show a concentration of variance at 3–5 yr period in the post-1960 time frame. The slightly shorter period displayed in the coral record compared to the pressure records may, or may not, be significant.
3. The change from > 5 yr period to 3–4 yr period is more rapid in the coral and Darwin records than in the Tahiti record.
4. Post-1960 there is a strong periodicity centred on 2.5 yr in the Tahiti pressure record, which is absent, or very much reduced, in the Darwin pressure and coral $\delta^{18}O$ records.

Results of cross-spectral analysis for the 1961–1990 time period indicated significant coherence between the coral $\delta^{18}O$, Madang rainfall and T–D SOI records for variance peaks with a 3–5 yr period. At this ENSO frequency, the T–D SOI is close to being in phase with, or slightly leads, the coral $\delta^{18}O$, and the coral $\delta^{18}O$ is close to being in phase with, or slightly leads, the rainfall. The rainfall appears to lag behind the T–D SOI by about 4 months. Weak spectral peaks in the biennial frequency band in the rainfall spectra and in the T–D SOI spectra are not coherent with one another.

4.5. Amplitude of annual cycle

Variations in the amplitude of the annual cycle in the coral $\delta^{18}O$ record through time were investigated using the complex demodulation technique. Complex demodulation is a form of localised harmonic analysis, particularly well suited to the examination of non-stationary time series [29]. The results of application of this technique to the coral record, with 3 different levels of pre-filtering of the data, are presented in Fig. 7A. This plot reveals apparent changes in the amplitude of the annual cycle on a variety of time scales. In Fig. 7B a comparison is made between the amplitude of the annual cycle in coral $\delta^{18}O$ with the amplitude of the annual cycle in Madang rainfall, both calculated using an intermediate degree of filtering. Periods of high amplitude of annual cycle tend to coincide in both records, and these periods tend to occur during the El Niño phase of the Southern Oscillation. These results suggest that, although relatively small in absolute terms, the observed variations in the amplitude of the coral $\delta^{18}O$ annual cycle (about 0.1–0.2‰) may be attributed mainly to variations in rainfall. In addition, they suggest that, in this region, the El Niño phase of the Southern Oscillation tends to manifest itself mainly as a reduction in dry season rainfall, rather than being a uniform decrease in rainfall for all seasons.

5. Discussion

A large degree of spatial and temporal variability in the ENSO phenomenon this century has been inferred by previous workers based on analysis of instrumental, documentary and proxy records from elsewhere in the Pacific region. Our results provide strong support for this general conclusion and, more importantly, provide unique detail on the nature of these changes in the far western equatorial region. The period from about 1920 to 1940 or 1950 has previously been described as one when there were relatively few 'moderate' or 'strong' El Niño events and when there were relatively weak correlations between various instrumental and proxy indices of the SO [30,31]. This interval broadly coincides with the period for which we infer that inter-annual climate variability in the Madang region was effectively decoupled from the climatic processes governing Tahiti and Darwin atmospheric pressure variation. In the post-1950 period, during which we infer a rapid coupling of the Madang climate to Tahiti and Darwin pressure oscillations, there were frequent occurrences of 'moderate' and 'strong' El Niño, generally strong correlations between various SO indices, and a strengthening of the trade winds in the eastern equatorial Pacific (up to 1980) [30–32]. However, these various changes do not appear to have occurred simultaneously around the Pacific [31]. The increase in degree of correlation between the coral $\delta^{18}O$ record and the T–D SOI (and Tahiti and Darwin pressures independently) occurred some 10–30 yr after the increase in correlation between Tahiti pressure and Darwin pressure (Fig. 5). This implies that the strengthening of the SO was not directly coupled to, or driven by, changes in the western equatorial Pacific. This, in turn, suggests that the far western equatorial Pacific was less important in modulating large-scale inter-annual climate variability in the 1920s–1950s than it appears to have been subsequently.

The changes in dominant periodicity of inter-annual variability centred on the late 1950s which we noted in the Madang coral and in the Tahiti and Darwin pressures, are broadly similar in timing and character to those described for rainfall in the western-central equatorial Pacific, inferred from isotopic analysis of massive corals at Tarawa Atoll [13]. Tarawa Atoll is 3,000 km ENE of Madang, effectively on the opposite side of the main ENSO rainfall anomaly region. That is, when the Indonesian low pressure migrates east during the El Niño phase of the Southern Oscillation, the north coast of Papua New Guinea experiences relative drought while the Tarawa region experiences particularly abundant rainfall. The coincidence in timing of changes in dominant periodicity with changes in Madang–SO climate correlation may indicate that these two features of the climate system are linked. That is, during the 5–6 yr period 'mode' of operation of the SO system, the far western equatorial Pacific may be less influential.

A significant feature of the $\delta^{18}O$ record from coral core Madang-1a is that the mean isotopic values are 0.1–0.2‰ $\delta^{18}O$ heavier in the 1920s and 1930s then in the rest of the record. Temporal variation in coral growth rate (Fig. 3A) is small and, therefore, unlikely to be a significant factor in these $\delta^{18}O$ changes [9,11]. Equally, there is no evidence of an increase in rainfall over this time frame (Fig. 3B). Variations in the isotopic composition of rain water are possible but, we believe, unlikely. Therefore, the most probable cause of the change in mean $\delta^{18}O$ is an increase in SST through time of about 0.5–1.0°C. Although the coral came from very shallow water where it could, conceivably, have experienced variations in diurnal heating related to changing cloud cover or wind speed, instrumental data provides some support for the idea of a more general rise in SST. COADS and station data for the 10° square box centred on 0°N, 150°E indicate a general temperature rise of about 0.4°C from the 1920s to the 1980s. Since small changes in SST in this region can translate to much larger and widespread changes in atmospheric circulation [1], even a temperature rise of 0.5°C could be related to the observed changes in climate variability which we have described.

6. Conclusions

From the above the following conclusions can be drawn:

1. Analysis of the $\delta^{18}O$ composition of cores from massive *Porites* corals on the north coast of Papua New Guinea can yield high-resolution and regionally reproducible records of climate. These records provide an insight into seasonal and inter-annual climate variability in a region acknowledged as critical for the understanding of the globally significant Southern Oscillation phenomenon.
2. Results from analysis of a 70 yr coral record have been used to extend the continuous climate record of the region by 26 yr, and have indicated significant changes in the mode of climate operation.
3. Changes in the magnitude of the annual cycle of

coral skeletal $\delta^{18}O$ and Madang rainfall are related to the Southern Oscillation, with strong seasonality generally occurring during the El Niño phase.

4. There is a marked change in dominant periodicity of the coral $\delta^{18}O$ record in the late 1950s. This change, from a rather diffuse concentration of variance at a >5 yr period, prior to the late 1950s, to an increasing concentration of variance within a 3–5 yr period after that time, is closely matched in timing, and to a large extent also in character, by changes in Tahiti pressure and Darwin pressure. Similar changes in rainfall at Tarawa Atoll, 3000 km to the ENE, have been documented [13].
5. At about the same time as the late 1950s change in periodicity, there are changes in the degree of coupling of the climate of the Madang region (as represented by local rainfall and coral skeletal $\delta^{18}O$ records) with the Southern Oscillation (represented by the T–D SOI and inter-annual atmospheric pressure fluctuations at Tahiti and Darwin). The post 1960 era is characterised by the highest degree of coupling of Madang climate with the Southern Oscillation. These data suggest that, during the 1920s–1950s, the far western equatorial Pacific region may have been less important in modulating Pacific and global inter-annual climate variability than it appears to have been subsequently.
6. The observed changes through time in the frequency domain and in the degree of coupling of regional climate to the SO appear to be part of a large-scale and complex evolution of the ENSO climate system. Our results emphasise the need to take account of the strongly non-stationary nature of the climate system when attempting to interpret instrumental or proxy records.

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