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Delta18O signatures from modern and fossil equatorial Pacific microatolls: indicators of mid-late Holocene ENSO variability

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δ¹⁸O signatures from modern and fossil equatorial Pacific microatolls:
Indications of mid-late Holocene ENSO variability.

A thesis submitted in fulfillment of the requirements for the award of the degree

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Matthew Robert Beech

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2002
Abstract

This thesis examines the skeletal oxygen isotope ratios of modern and fossil massive Porites microatolls from the reef flat and interior of Christmas Island, central equatorial Pacific Ocean. It has been known since the 1970s that the calcium carbonate skeletons of corals retain an oxygen isotopic ratio that is a function of the temperature and the isotopic signature of the sea water in which they live. Sea Surface Temperature (SST) and rainfall variation in the equatorial Pacific are major components of the El Niño Southern Oscillation (ENSO) phenomenon that dominates the ocean-atmosphere interactions of the equatorial Pacific. However, a limiting factor in the understanding of ocean-atmosphere variability is the short length of instrumental records of oceanic and atmospheric processes. To unravel the complexities of ENSO variability in the pre-instrumental past, it is necessary to find high-resolution proxy records whose reliability can be tested using modern proxy and instrumental data. Palaeoclimatic proxies provide the opportunity to test the results of sophisticated modelling techniques and together a history of variability can be constructed.

Christmas Island is located near the equator in the central Pacific in a narrow, dry zone where climate is dominated by the alternation of the warm and cold phases of ENSO. It captures a large proportion of the El Niño-La Niña SST variation as recorded by oceanic observing systems and experiences extremes of rainfall and drought associated with the respective warm and cold phases of ENSO. The island is surrounded by reef flats that are inhabited by massive and branching corals and the interior is dominated by a network of hypersaline lakes that were previously interconnected, around which are numerous fossil microatolls from the mid and late Holocene. Comparison of the skeletal oxygen isotope ratios of two modern microatolls and instrumental SST and rainfall data, demonstrates the reliability of these coral proxies to reflect a pattern of ENSO variation that is equivalent to that of instrumental records. The analysis of fossil microatoll density structures provided a method from which to base internal skeletal chronologies of four fossil microatolls. These chronologies were applied to the results of oxygen isotope analysis and subsequently provided a rare insight into ENSO variability from the core region of the central equatorial Pacific.
The comparative histories indicate that ENSO SST and precipitation variation was less intense between 3.8 and 2.9 ka and that La Niña events were more pronounced in the ENSO signature in the mid and late Holocene, than they are at present. Intensity had increased to an extent that exceeds modern ENSO by the late Holocene and is consistent with a peak in frequency and intensity predicted by modelled experiments based on precessional changes in insolation seasonality, at ~2 ka. However, the magnitude of the increase in variation is somewhat greater than the models predict and reflects stronger rainfall teleconnections through enhanced interaction between the Southern Oscillation and the Intertropical Convergence Zone.
Acknowledgements.

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Chapter 1

Introduction.

1.1. Introduction.

The El Niño-Southern Oscillation (ENSO) apparent in the surface wind strength, Sea Surface Temperature (SST) and convective system of the central equatorial Pacific has far reaching effects on tropical and extra tropical climate. Current understanding of ENSO events is based on instrumental records since the 1970s (Trenberth and Hoar, 1996, McPhaden, 2001). Typical events last from twelve to eighteen months and are generally detected in the Northern Hemisphere autumn and peak in winter, although each event is unique in its behaviour. Events can have a fast development and can peak quickly (El Niño event of 1982-83) while others develop slowly over longer periods. The El Niño event of 1997-98 took twelve months to reach its peak, starting in December 1996 and reaching a maximum by the same time the following year. What actually causes the ENSO cycle cannot be isolated to one single process because how this coupled atmosphere-ocean system evolves is not well understood (Cane and Zebiak, 1985). Modern ENSO oscillates between two phases of warm and cold events that have an average frequency of 4.5 years, with a range of 2-10 years (Hall et al., 2001).

The importance of ENSO as a major climate driver has been established in the last thirty years of research. Measurements have shown that it effects the entire tropical Pacific and beyond (Rasmusson and Carpenter, 1982). Teleconnections are connections between spatially separated atmospheric systems and were first discussed in terms of ENSO by Bjerknes (1969). These ideas initiated the development of atmosphere-ocean models whose main objective is to simulate and predict ENSO events and their impacts. These prediction schemes are based on statistical techniques or coupled ocean-atmosphere models of varying degrees of complexity, all of which use parameters based on the instrumental recording of ENSO processes (SST, winds, rainfall, thermocline etc). In a review of ENSO prediction Latif et al (1994) conclude that ENSO is predictable at least one year in advance, although their conclusion applies to the gross indices of ENSO such as the Southern Oscillation Index (SOI) and not more specific features such as Indian Monsoon rainfall and South African drought (Latif et al, 1994).
Recent ENSO behaviour is considered to be becoming more intense and frequent and is possibly responding to global warming (Timmermann et al., 1999). However, instrumental records are short (approximately 150 years) and conclusions should be made with caution. It is vital to find proxies, that can be easily retrieved, reliable and reproducible for the understanding of ENSO variation over longer periods than decades. Corals from oceanic areas that are otherwise devoid of suitable proxy material are potentially the most useful proxies (Dunbar et al., 1994, Urban et al., 2000). They can extend our knowledge of ENSO back in time so that the understanding and prediction of ENSO variation for the future can be improved.

This study analyses isotopic ratios from two modern coral microatolls from Christmas Island and establishes their ability to reproduce the modern ENSO variation recorded in instrumental data. The relationship subsequently forms the basis for the interpretation of ENSO variability from isotopic ratios of fossil microatolls from the mid and late Holocene.

1.2. Christmas Island.

Christmas Island is situated in the Line Islands of Kiribati, at 1°52’N, 157°20’W. The most southern of the Line Islands, north of the equator, it lies in the narrow, dry zone of the eastern and central Pacific. Shaped like a lobster claw, it is the world’s largest atoll with a surface area of 360 km² (Trichet et al., 2001). The interior is dominated by a network of more than 500 interconnected or isolated lakes with salinities varying from brackish (17 %o; Trichet et al., 2001) to hypersaline (≥ 3000 %o; Valencia, 1977). Connected to some of the more westerly lakes there is a large, shallow lagoon that covers an area of 190 km² (Woodroffe and McLean 1998). Scrub and ground cover of various species (see, Chock and Hamilton, 1982, Wester, 1985) dominate the majority of the exposed interior and the exterior is surrounded by narrow reef flats covered with branching and massive corals. The reef flats that are generally shallowly submerged at the macrotidal low tide are subject to a tidal range of <1 m and interannual sea level variation associated with ENSO (Woodroffe and McLean, 1998).

The islands climate is also subject to ENSO variation due to its close proximity to the equator and its location in the central Pacific (Figure 1.). The mean annual
rainfall for the island is 936.3 mm for the period 1951-1999, with a range of 177 mm (1951) – 3,686 mm (1997). Isolated from other landmasses the island lies in a position that experiences a see-saw like oscillation of SST driven by the prevailing trade winds. It is therefore, well placed to capture a large proportion of the El Niño-La Niña SST variation as described by the TOGA-TAO Array (Figure 1), recording a decrease in SST of ~5°C (c) from extreme El Niño conditions of December 1997 (a) to extreme La Niña conditions of December 1998 (b).

![Figure 1](image)

*Figure 1.* The equatorial Pacific and location of Christmas Island (triangle) during (a) El Niño conditions of December 1997, (b) La Niña conditions of December 1998 and (c) the difference between a and b.

The δ¹⁸O signature of Christmas Island *Porites* has been shown to correlate well with the NINO3 SST anomaly estimates and used to reconstruct recent SST. Evans *et al.*, (1999) made the comparison of δ¹⁸O from a *Porites* coral from South West Point on Christmas Island and the NINO3 area index of the eastern equatorial Pacific SST anomaly. They termed the δ¹⁸O signature, the NINO-C index and showed that it shared over 60% variance with the instrumental NINO3. Furthermore, a study to determine optimal sites for coral based reconstructions of SST variability determines that Christmas Island, in the central and Galapagos in the eastern Pacific, minimise the error in the reconstruction of the global SST field (Evans *et al.*, 1998 b). These correlations are significant as numerous fossil microatolls that are relatively easy to collect occur at the edges of the saline lakes of the interior of Christmas Island. They range from 0.5 to 1.0 m above those found on the modern reef flat, and although there is no geographical pattern or elevation trend in the distribution of fossil microatolls, their existence implies
that past sea levels were higher (Woodroffe and McLean 1998). Therefore, the lakes were probably more extensive and incorporated into one reticulated lagoon that was less isolated from open ocean conditions than they are at present (Figure 2). The correlation of a *Porites* microatoll (CW3) $\delta^{18}O$ with instrumental SST and rainfall data, and analysis of the $\delta^{18}O$ signature of a fossil microatoll implies that these fossil microatolls are the possible containers of historical records of ENSO variability (Woodroffe and Gagan, 2000). They also originate from a site in the central equatorial Pacific that has been described as 'an optimum location for the monitoring of the thermal oceanographic signal of the full ENSO cycle' (Evans, 1999).

![Figure 2. Christmas Island and the locations of modern and fossil microatolls. Based on locations of fossils and current lagoons, the dashed line represents a hypothetically more extensive lagoon that was less isolated from the open ocean conditions and less saline than the present day lagoons.](image)

This project will endeavor to substantiate the correlation between CW3 $\delta^{18}O$ and instrumental SST and rainfall data from Christmas Island in the analysis of a second modern microatoll. In doing so, the ability of individual microatolls from the same fields to reproduce variation of the same oceanographic processes will be established. It will then be possible to interpret the $\delta^{18}O$ signatures from fossil microatolls in terms of pre-instrumental ENSO variability.
Chapter 1

1.3. Thesis outline.

The thesis begins with a literature review that introduces the El Niño-Southern Oscillation and the current consensus on its variation and existence through time. Corals as environmental proxies and *Porites* microatolls are introduced and examples of *Porites* as proxies are given.

Methods of collection and sampling are explained (conducted before the project began) before the methods of analysis. These methods comprise of the X-radiography, Computerised Tomography (CT), image and stable isotope analysis that has been conducted on modern and fossil microatoll samples. Oceanic data comes from a number of sources and an explanation of what these sources are and how they are measured precedes the explanation of statistical methods.

The results (Chapters 4 and 5) explain the δ¹⁸O records of two modern microatolls and how they are used to reconstruct the pattern of ENSO variation given by instrumental SST and rainfall data. The density structures of the modern microatolls are then examined, in search of a method to determine an internal skeletal chronology for four slices of fossil microatolls, so that their δ¹⁸O signatures can be explained. The results of radiocarbon dating of fossil microatoll samples are given and the implications of the temporal distribution of the fossils are considered. The internal skeletal chronologies for the fossil samples are then determined using their density structures, and the δ¹⁸O signatures are examined with respect to ENSO. The annual variation of ENSO using running averages of the δ¹⁸O data sets are examined in all the fossil records and compared with that of the modern. This variation is then used in describing typical ENSO event characteristics for each of the time periods involved.

A discussion of the results and their implications is subsequently given, along with how they fit in to the current understanding of ENSO variability over recent millennia. Conclusions of the project and suggestions for further research are presented at the end.
El Niño Southern Oscillation, corals as proxies and *Porites* microatolls.

2.1. Southern Oscillation and the tropical circulations.

The failure of the monsoon in 1899 and the subsequent famine prompted Sir Gilbert Walker to embark on predicting interannual variations in the monsoon. He described that when pressure is high in the Indian Ocean, it is low in the Pacific and termed the oscillation in pressure difference, the Southern Oscillation (Philander, 1990). In an effort to document the full scope of the Southern Oscillation and subsequently find the key to monsoonal prediction, Walker published papers in the period 1923-1937 that synthesised his results and established that the Southern Oscillation is correlated with major changes in rainfall patterns over the Indian and tropical Pacific Oceans. He used an average pressure difference and the standard deviations away from this average, from records of Darwin, Australia, Canton Island, central Kiribati and Santiago, Chile, and called it the Southern Oscillation Index. However, due to anomalous conditions associated with the Southern Oscillation and his failure to identify them, the scientific world was sceptical and interest in the Southern Oscillation became dormant.

Sir Gilbert Walker would not be credited for his work until 1969 when Jacob Bjerknes described a physical relationship between interannual variations in the atmosphere and ocean and called it the Walker Circulation (Bjerknes, 1969). He explained how the cold, dry air from the upper atmosphere sinks over the eastern equatorial Pacific and flows west with the trade winds when it reaches the ocean surface. The cold air heats up as it blows over an ever-increasing sea surface temperature gradient, giving rise to towering rain clouds in the west. This cycle is thought to be completed in the upper atmosphere as it decreases in temperature in a west-east direction, although this process has not been confirmed (Figure 3).

![The Walker Cell driven by SST gradient of the equator](image)

**Figure 3.** Intensity of the zonal Walker Circulation depends on the strength of equatorial easterly winds that are strong in La Niña events and weak in El Niño events.
The Hadley Circulation is the meridional counterpart of the Walker Circulation, where warm air rises at the equator and flows North or South in the upper atmosphere, sinking again in the belt of subtropical highs in each respective hemisphere (Bjerknes, 1966). On reaching the cooler ocean surface the air flows back, over an increasing sea surface temperature gradient, towards the equator in the northeasterly trade winds in the Northern Hemisphere and the southeasterly trade winds in the Southern Hemisphere (Figure 4). The Hadley Circulation tends to be strongest in the Northern Hemisphere during December, January and February and strongest in the Southern Hemisphere in June, July and August, the respective winters of each hemisphere (Trenberth, 1991).

Figure 4. Intensity of the Hadley Circulation also depends on the strength of the equatorial easterlies but in the opposite direction to the Walker Circulation. The Hadley Circulation is strongest when easterlies are weak and convection along the width of the equatorial Pacific is high, such as during an El Niño event.

2.2. El Niño Southern Oscillation.

The term ‘El Niño’ is Spanish for Christ (The Boy) and derives from a description used by the Peruvian fishermen that experience periodic warm water seasons off their coast at Christmas time. During these seasons the warm water replaces the cold, nutrient rich upwelling water, causing the decline of the Peruvian fishery as the fish stocks move to richer hunting grounds or die. Not only does El Niño cause a decline in fisheries, but it also brings with it greater amounts of rainfall causing floods and together, El Niño seasons are potentially disastrous for the Peruvian economy.

In 1969, Bjerknes coupled the zonal and meridional atmospheric cycles of tropical circulation and the temporal and seasonal variations in SST of the equatorial
Pacific and termed it El Niño-Southern Oscillation (ENSO), referring to the Peruvian fishermen and to Sir Gilbert Walker’s Southern-Oscillation.

The El Niño Southern Oscillation has a warm and a cool phase, El Niño and La Niña respectively, and is measured using the Southern Oscillation Index. The rising air in the west and the sinking in the east creates the respective low and high-pressure systems apparent in the equatorial Pacific, the former being termed The Indonesian Low. It is the magnitude of this pressure difference, measured between Darwin and Tahiti (not Darwin, Canton and Santiago, as Walker had suggested) that determines the value of the SOI and therefore the state of ENSO.

The Southern Oscillation Index is the standard deviation about the long term mean sea-level pressure difference between Darwin and Tahiti. Negative values signify the warm phase (El Niño) of ENSO and positive values signify the cool phase (La Niña). La Nina is apparent when the sea level pressure at Darwin is low and sea level pressure at Tahiti is high. This pressure difference allows a strengthening of the trade winds in the east, which drives the westward migration of the warm pool. In the eastern Pacific, upwelling increases as surface water moves to the west and the thermocline rises. The westward migration of the warm water causes the sea surface temperature gradient to increase in an easterly direction, thus allowing the pressure difference between Tahiti and Darwin to exist (Figure 5).

Figure 5. A La Niña event where equatorial easterly winds, the intensity of the Walker Circulation and the sea surface gradient are high. A low pressure system resides over the warmest water which is ‘piled up’ in the west, causing enhanced rainfall in the western Pacific region. A high pressure system over the coolest water in the east causes drought in the eastern Pacific region as the ‘cold tongue’ advances towards the west.
In a La Niña event, the strength of the Walker Circulation increases driven by the strengthening of the equatorial easterlies whereas the Hadley Circulation decreases as convection throughout the central and eastern Pacific decreases.

El Niño is the warm phase of ENSO and has the opposite effects to La Nina. The sea level pressure difference between Darwin and Tahiti is reversed, causing a weakening and sometimes a reversal of the trade winds and a decline of the sea surface temperature gradient. This allows the eastward migration of the warm pool and the subsequent depression of the ‘cold tongue’. In the eastern Pacific the thermocline falls to deeper depths and upwelling decreases, reducing surface mixing and promoting nutrient depleted water at the coasts of Peru (Figure 6).

**Figure 6.** An El Niño event where equatorial easterly winds and the sea surface gradient decrease and the intensity of the Hadley Circulation increases. The low pressure system follows the warmest water into the central and eastern Pacific and is replaced by high pressure. Rainfall is enhanced through the central and eastern Pacific as the thermocline retreats to deeper depths.

### 2.3. Tropical convergence zones.

There has been a collaborative scientific project to examine seasonal and interannual variability and the regional and global climate variability associated with the phenomenon since Bjerknes (1966, 1969, 1972) described and named ENSO. The regional variability of ENSO is briefly described above and is a function of the strength of the Walker Circulation in the zonal field and the Hadley Circulation in the meridional field, depending on the state of ENSO. In an El Niño event the warm pool expands eastwards across the Pacific, bringing with it the warm, moist air and associated rain clouds. The Indonesian Low migrates southeast, causing drought in the western Pacific.
and over the northern Australian and eastern Indonesian landmasses (Philander, 1990). The central Pacific through to equatorial South America experiences greater amounts of rainfall and flooding; annual rainfall can be four to ten times higher than average during El Niño events (Holmgren et al., 2001). In a La Nina event the reverse is observed. The strengthening of the equatorial trade winds causes an increase in the sea surface temperature gradient and the Indonesian Low and the warm pool migrate further west than is usual, causing heavy floods in northern Australia and eastern Indonesia and droughts in equatorial South America (Glantz et al. 1991).

Rainfall variation in the equatorial Pacific is a response to convection caused by the position of the warmest water which also controls the position of the Intertropical and South Pacific Convergence Zones (ITCZ and SPCZ respectively). Convergence is intensified over warm water as more air rises than it does over cold water. At the equator the north and southeasterly trade winds converge over the warm central Pacific region resulting in the position of the ITCZ between 0° and 10°N. South of the Equator between 150° and 160°E the south easterlies converge with the north easterlies associated with the Indonesian Low pressure system, thus resulting in the position of the SPCZ.

These convergence zones prevail over water that is warmer than 27°C (Trenberth, 1991) thus, moving accordingly with the warm SST field that varies with the state of ENSO. During an El Niño event the warm pool extends across the Pacific, which results in the convergence of the ITCZ and SPCZ. The former moving further south, towards the Equator and the latter moving north, towards the Equator (Figure 7). In an extreme El Niño event such as that of 1982-83, the convection that is formed over the warm pool might be interpreted as one massive convergence zone, incorporating both the ITCZ and the SPCZ (Trenberth, 1991).

![Figure 7. Positions of the tropical convergence zones during El Niño.](image-url)
As the two zones get closer the amount of warm, rising air increases, subsequently strengthening the magnitude of the Hadley Circulation, described above. The Walker Circulation is weakened due to the decrease of rising warm air in the west, sinking cool air in the east and a decline of the sea surface temperature gradient.

In a La Niña event, the reverse happens. The cold tongue extends westward along the equator, piling up the warm water in the western Pacific, which decreases the amount of rising air in the central Pacific, resulting in the separation of the convergence zones. The ITCZ moves north and the SPCZ moves south, leaving room for the strengthening of the described Walker Circulation and the westward propagation of the cold tongue (Figure 8).

![Figure 8. The position of the tropical convergence zones during La Niña.](image)

Since the 1970s, ENSO has been at the forefront of ocean-atmosphere research. Variability has increased in intensity over the last few decades and modelling results suggest that it will further do so in a greenhouse warmed future (Timmerman et al., 2001). Key to the prediction of future ENSO variability is the understanding of its history and corals are consistently providing proxies for this. Emanating from research into coral proxies and comparison with proxies of terrestrial origin, is an understanding of the millennial scale ENSO variability that is described below.

### 2.4. ENSO in the pre-instrumental past.

The pre-instrumental history of the Pacific is documented in proxies distributed throughout the Pacific rim. A simple conceptual model of climate change based on changes in the overall circulation intensity and in the Pacific Walker Circulation describes an enhancement of the Walker Circulation around 5 ka (Shulmeister, 1999).
An increase in the Walker Circulation would be driven by cooler SSTs throughout the tropical Pacific, consistent with current understanding of deceased SSTs throughout the Pacific since the last glacial maximum (LGM). However, there is considerable debate in the magnitude of this cooling as mountain areas in New Guinea (Western Pacific) have shown a mean annual temperature of at least 6°C cooler at the LGM (Hope, 1989) supported by δ¹⁸O values from corals from the Huon Peninsula showing a similar decrease (Aharon, 1983). In contrast to these estimates, McCulloch et al (1996) recorded SSTs at ~8.9-7.4 ka of 2-3°C cooler than present in *Porites* corals from the Huon Peninsula and are more consistent with the conclusions of the CLIMAP project (1981) that estimated cooler than ~1°C at the LGM. More recent research has indicated mean SSTs varied between 1.5 and 3.6°C cooler than present between 4.2 and 1.8 ka (Abram et al., 2001). However, other authors have described a warmer LGM mean SST (Anderson, 1989) and a 1°C warmer mean SST in the western Pacific region at ~5.4 ka (Gagan et al., 1994). Although the debate as to how much cooler the surface ocean has been since the LGM continues, most studies are in agreement that it was cooler and has increased in temperature since the LGM.

Cooler conditions in the tropical Pacific would increase the Walker Circulation as suggested by Shulmeister (1999), subsequently enhancing the cool water input through enhanced Ekman transport off the coast of South America. A persistent cool water input to the tropical Pacific would decrease average SST throughout and the climate of the equatorial Pacific would be more similar to the modern La Niña than El Niño as some early-mid-Holocene model results suggest (Bush, 1999). It would also cause a westward migration of the western Pacific warm pool. A mid-Holocene precipitation maximum (5-4 ka) at Groote Eylandt, Gulf of Carpenteria supports this, as convection maxima would also be displaced to the west with the warm pool (Shulmeister and Lees, 1995).

Inorganic laminae younger than 200 years, deposited by storm induced debris flows in Ecuadorian Alpine lakes match historic records of El Niño events. Older records from the lakes dating back as far as 15 ka show periodicities of greater than 15 years between 15 ka and 7 ka. At approximately 5 ka this periodicity decreased to between 2 and 8.5 years, indicating the on-set of modern El Niño (Rodbell et al, 1999).
These authors inferred that the modern El Niño periodicities reflect the onset of a steeper zonal SST gradient caused by an enhanced Walker Circulation. Sandweiss et al. (1996) present geoarchaeological data from Peru that also indicates the onset of El Niño which is supported by Tudhope et al. (2001) who present coral data from New Guinea indicating considerably weaker or non-existent ENSO activity between 8.8 and 5.8 ka.

Collectively, these proxies give an idea of equatorial Pacific conditions throughout the Holocene but they all originate from the continental shores of the Pacific or from terrestrial sites of the Pacific coasts and evidence from the central Pacific is limited in the literature.

2.5. Corals as climate proxies.

Instrumental records of tropical oceanic processes rarely exist prior to 1950 (Fairbanks et al., 1997). The use of tree rings in the science of dendrochronology has provided windows into past climatic conditions but is limited to terrestrial and atmospheric observations. In a similar way, corals have been shown to successfully record recent oceanic variability, based on the cross dating with instrumental records (Cole et al., 1993, Gagan et al., 1994, McCulloch et al., 1996, Evans, 1999, Linsley et al., 1999). Considerable research has assessed the extent and accuracy of corals as climate recorders since Knutson et al., (1972) described density bands in corals and their annual and interannual variability. It is vital to discover a method of successfully correlating variations in coral make up with present climatic conditions, in order to determine climatic conditions of the past. Modern corals are scattered over an ocean area of around 190 x 10^6 km^2 within the tropical belt (Aharon, 1991) and are the most promising containers of historical oceanic data available. By determining the relationships between modern corals and ocean-atmosphere conditions it will be possible to use the same relationships to determine past ocean-atmosphere conditions from information found in fossil corals.

**Density Bands.**

Ma (1933, 1937) described a cyclical pattern of coral growth related to environmental fluctuations, but it was not until 1972 that density bands in massive coral species were analysed in detail. Knutson et al., (1972) described how the growth rates
in massive corals result in bands of varying density perpendicular to the axis of growth, which can be revealed by X-radiography. They determined that the bands were annual by comparing the density bands with autoradiographs of the skeleton that revealed the radionuclide signature produced by atomic weapons testing (Knutson, et al. 1972). Since 1972 many methods of demonstrating the annual nature of density banding have been described (Dodge and Thompson, 1974, Dodge and Vaisnys, 1975, Hudson et al., 1976) using various species of coral. The annual density pattern is almost universally described as comprising one dense and one less dense band (Barnes et al., 1989). However, 12 alternating density bands have been identified to represent one year, or one month for each respective band (Barnes and Lough., 1989). Although some corals show complicated and sometimes unidentifiable density banding, others show very clear bands, which can be examined by different methods of densitometry (optical, X-ray or gamma), and chronologies can be produced. There are however, limitations to these methods which can produce errors in the chronologies determined. The calcification during growth of corals is poorly understood, making it difficult to associate specific bands with real-time and instrumental data sets, collected to ‘ground truth’ the results. However, Lough and Barnes (2000) have shown that calcification and extension rates in massive coral *Porites* from the Great Barrier Reef are significantly and directly related to SST. More often than not, a linear growth rate (the length of the growth axis / the amount of annual bands observed) is used, but this is not always the case and they may vary between coral species. Massive corals usually grow relatively uniformly making them the optimal species for this type of study (Land et al., 1975).

A further problem with the corals is the three-dimensional manner in which they grow and the two-dimensional picture produced by X-radiography. Neighbouring coral polyps secrete and live in small calices of calcium carbonate, together forming colonial aggregations which make up the coral skeleton. They have an average diameter of 1mm for massive corals (Veron, 1986) and a slice of massive coral for x-radiography is likely to be no less than 5 mm thick and no more than 10 mm thick, due to the difficulties in cutting the delicate skeleton. Although this is optimum thickness, a slice this thick is likely to include several, overlapping calices. X-radiography is the measure of density through the medium being observed, in this case the coral slice. The X-radiograph will average the density through the thickness of the slice, subsequently distorting the picture of the true density banding due to overlapping calices, this distortion was termed
the ‘Venetian blind effect’ by Barns and Devereux, (1988). Although the density bands can be used to approximate climate variability, they cannot be treated as simple black box recorders of environmental conditions and must be verified with other tracers (Barnes and Lough, 1996).

**Stable Isotopes.**

Weber and Woodhead (1970) began looking at the ratios of oxygen and carbon isotopes in corals and found that the most important factors causing isotope ratio variations in the skeletal carbonate of corals are zooxanthellar activity and temperature. Since then, considerable research has been done to determine the exact cause of these variations and what can be learnt of pre-instrumental variability (Weber and Woodhead, 1972, Erez, 1978, Swart and Coleman, 1980, Druffel, 1985, Cole and Fairbanks, 1990, Gagen et al., 1994). Such studies have resulted in the conclusions that $\delta^{18}O$ ratios reliably track variations in sea surface temperature (Druffel, 1885, Winter et al., 1991, Woodroffe and Gagan 2000), seawater compositions (Dunbar and Wellington, 1981), light intensity (Swart and Coleman, 1980) and salinity and precipitation (Swart and Coleman, 1980, Cole and Fairbanks, 1990).

Oxygen isotopes are a commonly used tool for coral paleoceanography. Calcium carbonate precipitated from water by marine organisms has an oxygen isotope composition, which is temperature dependent (Epstein et al., 1951 and Epstein et al., 1953). Variations of $\delta^{18}O$ in coral skeletons are a combined signal of the temperature and isotopic signature of the water in which they live (Druffel, 1985). $\delta^{18}O$ decreases with a rise in SST due to removal of $^{18}O$ during kinetic effects, decreasing by 0.18-0.22 %o for a rise of 1°C in SST (Gagan et al., 1994) depending on species. It increases with a rise in salinity due to the removal of the lighter isotope $^{16}O$, increasing by 0.06 %o with a rise of 0.1 %o in salinity (Druffel, 1985). As the isotopes are stable and remain in the calcium carbonate skeleton in the same concentrations in which they were deposited, they are therefore historical records of environmental information. The climatic variations associated with variations in SST and salinity are extremes of rainfall and positions of low and high pressure systems (Cole and Fairbanks, 1990), flooding and land runoff (Isdale, 1984), positions of warm pools and areas of upwelling (Lea et al., 1989). $\delta^{18}O$ is a record of SST in the open ocean as there is often little influence on
salinity variability (Druffel, 1985). However, in ENSO dominated areas of the Pacific, such as Tarawa, $\delta^{18}O$ gives a salinity signature due to greater amounts of rainfall experienced during El Nino (Cole et al., 1993) and in areas such as Canton, both the SST and precipitation effects of ENSO are experienced (Fairbanks et al., 1997). Cole et al (1992) demonstrate how a collection of corals from sites that span the equatorial Pacific can be used to monitor ENSO dynamics from different localities. Christmas Island, located on the eastern edge of the warm pool is close to the maximum thermal signal associated with ENSO and ENSO induced warm and cool phase SST anomalies are prominent (Bjerknes, 1969), whereas precipitation anomalies are moderate (Ropelewski and Halpert, 1987). Paired strontium/calcium and $\delta^{18}O$ ratios from Christmas Island corals have been shown to be driven by SST, with rainfall related salinity variation a secondary influence (Evans 1999, see Trace Elements).

$\delta^{13}C$ isotopes can also be used as climate recorders although they are harder to decipher (Dunbar and Cole, 1993). Interpretations of $\delta^{13}C$ variability have thus far, been concentrated on specific processes or conditions, rather than by a thorough understanding of all relevant processes (Dunbar and Cole 1993). $\delta^{13}C$ has been shown to provide reconstructions of historical variations in photosynthesis and the associated light intensity (Emiliani et al., 1978), insolation (Fairbanks and Dodge, 1979) and coral growth rates (Goreau, 1977, Dunbar and Wellington, 1981). $\delta^{13}C$ variations are caused by the coral itself and derived form the water in which it lives. Photosynthesis and respiration are important factors as they modify the isotopic composition of the seawater in the region of the coral polyps (Aharon, 1991). Photosynthetic rates cause depletion in $\delta^{13}C$ when they are high, thus allowing growth rates of corals to be determined by $\delta^{13}C$. Accelerated growth rates in Porites from the Great Barrier Reef have been determined by depletion in $\delta^{13}C$, it was also observed that depletion was greatest in the fastest growing parts of a single coral colony (Dunbar and Wellington, 1981 Aharon, 1991,). The relationship between coral growth vigour and $\delta^{13}C$ has been recognised as a useful indicator of coral growth hiatuses (Cole and Fairbanks, 1990, Cole et al., 1993). However, Gagan et al (1994), suggest that $\delta^{13}C$ is not governed by growth rate but by changes in coral metabolism associated with gamete production, based on the results from Porites from the Great Barrier Reef. It was also recognised that these corals spawn at predictable times associated with lunar and tidal cycles and
that the timing of $\delta^{13}$C peaks coincided with these spawning events. It was then possible to accurately establish a time scale on the basis of spawning events (Gagan et al., 1994).

**Trace Elements.**

The first trace elements used to show oceanic variability were cadmium (Cd) and barium (Ba) (Shen et al., 1987 and Lea, et al., 1989). A cadmium/calcium (Cd/Ca) ratio was recorded from the reef-building coral *Pavona clavas* collected at Punta Pitt, San Cristobel Island, Galapagos Islands. Cadmium is found in phosphates and nitrates that are highly concentrated in upwelling, nutrient-rich water, thus signaling the cold phase of ENSO. Also variations in Cd have been correlated with aeolian fluxes of twentieth century, North American industrial cadmium to the western North Atlantic (Shen et al., 1987). Barium is also found in large concentrations in nutrient-rich-water, ie. upwelling or river runoff (Shen and Sanford, 1990). The Ba record from the Galapagos identified upwelling in the eastern equatorial Pacific. The Ba/Ca ratio can be seen to be at a maximum when SSTs are low and at a minimum when SSTs are high which implies that Ba concentrations increase as cold, nutrient rich water from depth, upwells to the surface. In turn, indicating periods of high and low SSTs and, therefore the state of ENSO (Lea et al., 1989).

Scleractinian corals secrete skeletons of aragonite (CaCO$_3$), which incorporates both strontium (Sr) and calcium (Ca) into its structure. The ratio of incorporation of Sr to Ca is controlled by two factors, the activity ratio of the ocean water and the Sr/Ca distribution coefficient between aragonite and seawater. The Sr/Ca distribution coefficient depends, mainly on the temperature of the seawater that the coral grows in, rather than the chemical composition of the seawater. As the residence times of Sr and Ca in seawater have probably remained constant over time scales of $10^5$ years, the Sr/Ca in coral structures can potentially be used as a monitor of ocean temperature on these time scales. Beck et al. (1992) showed that high precision mass spectrometric analysis of Sr/Ca ratios can be used to determine past SST values within a few tenths of a degree celsius. The analysis of both Sr/Ca and $\delta^{18}$O therefore enables the differentiation of the temperature and $\delta^{18}$O composition of sea water signals apparent in coral $\delta^{18}$O records (McCulloch et al., 1994). McCulloch et al. used this method to separate the combined
effects of temperature and seawater $\delta^{18}$O apparent in a coral $\delta^{18}$O record from the Great Barrier Reef, Australia and to estimate the flux of $^{18}$O depleted river runoff from the nearby Burdekin River. Evans et al. (1999) also adopted this method in a study of $\delta^{18}$O signature from Christmas Island corals. Their results suggest that interannual $\delta^{18}$O variability is primarily a function of SST variation, and sea water $\delta^{18}$O influences are greatest during large El Niño events when local SSTs exceed the convective threshold and extreme rainfall maxima are experienced. However, they also imply that the combined $\delta^{18}$O signal may make a better climate monitor than SST or rainfall alone, as the co-variance of SST and enhanced local rainfall at Christmas Island synthesises both the convective atmospheric and thermal oceanographic signal of ENSO (Evans et al., 1999).

Other trace elements recorded in aragonite composition can also be used for recording past changes in reef environments e.g., manganese reversals in the flow of the trade wind (Shen et al., 1992b), global augmentation of environmental lead by industrialization (Shen and Boyle, 1987) and fluvial discharge fluctuations of barium (Shen and Sanford, 1990).

**Fluorescence.**

In 1984, Peter Isdale first described the presence of fluorescent bands in massive coral slices from the Great Barrier Reef when put under long wave ultra violet light. The timing, width and intensity of these bands correlated strongly with summer, monsoonal rainfall and coastal runoff (Isdale, 1984). The cause of this intense fluorescence was the fluvial runoff of the Burdekin River, associated with extreme monsoonal rains over Northern Queensland (Isdale’s study area). He determined this by sampling corals restricted to within 20 km off the coast and mid- and outer shelf corals, the latter, being influenced by the rainfall but not by fluvial runoff. Later work inferred that the cause of the fluorescence to be the fulvic and humic acids found in fluvial runoff (Boto and Isdale, 1985, Coble et al, 1993, Matthews et al., 1996). However, other authors have reported fluorescence in corals that originate from regions that are far removed from land or fresh water sources that would give rise to fulvic and humic acids (Susic et al., 1991, Smithers 1997). Further more, Barnes and Taylor (2001) conducted an experiment to assess the nature and cause of luminescence in corals and
concluded that it was predominantly caused by skeletal architecture and not by terrestrial fulvic and humic acids (Barnes and Taylor, 2001).

2.6. Microatolls.

Corals of the genus *Porites* are widespread throughout the Indo-Pacific, forming massive domed colonies up to 3 m in diameter in deep water and microatolls in shallow-water reef environments such as occur on reef flats and in lagoons (Smithers, 1997). Microatolls are discoid coral colonies with flat tops, devoid of living tissue with rounded living edges that are formed by predominantly massive corals that live in intertidal reef flat and lagoonal environments (Stoddart and Scoffin, 1979).

Microatolls develop when the upward growth of the colony is limited and the polyps on the sides of the coral grow outwards. Excessive sedimentation has been scrutinised as the limiting factor (Dodge and Veisnys, 1977, Brown *et al.*, 1986), although the current consensus is that exposure at the low tide air-water interface is a more likely control (Woodroffe and McLean, 1990, Scoffin *et al.*, 1997, Smithers, 1997). As the coral polyps reach the air-water interface they cease to survive and retreat to lower levels on the side of the skeleton where they can survive. As long as the mean sea level remains constant, they will continue to grow outwards adopting a lateral growth form as seen in the ‘classic’ microatoll shape. Environments such as lagoons or moated regions where the tide falls to similar levels on each ebb provide ideal locations for the formation of microatolls. However, it has been observed that microatolls respond to sea level in locations where oceanic processes such as ENSO, dominate mean sea level variation (Woodroffe and McLean, 1990).

A series of possible sea level variations, and microatoll responses are shown in Figure 9. A massive coral will grow up from the sea bed until it reaches a limiting level (a), after this it will adopt a lateral growth form where the top will be near horizontal and devoid of living tissue and the edges will be rounded and colonised by the polyps (b). If the sea level then rises it will resume its vertical growth, in doing so constructing ridged features on its outer most rim (c). Alternatively, the sea level may gradually fall
causing the polyps to retreat to lower levels while retaining lateral growth and constructing an elongated dome shaped structure (d). Similarly, the sea level may undergo intermittent decreases causing the formation of terraces on the top of the microatoll (e). In a scenario where sea level rises and falls through time the microatoll will form a series of ridges along its top (f).

<table>
<thead>
<tr>
<th>a. Microatoll grows upwards to a limit to coral growth.</th>
<th>b. On reaching the surface lateral growth prevails.</th>
</tr>
</thead>
<tbody>
<tr>
<td><img src="image1" alt="Diagram" /></td>
<td><img src="image2" alt="Diagram" /></td>
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<tr>
<th>c. Vertical growth resumes as sea level rises.</th>
<th>d. Microatoll grows laterally but to lower elevation with a gradual sea level fall</th>
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<tr>
<td><img src="image3" alt="Diagram" /></td>
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<tr>
<th>e. Intermittent fall of sea level causes terracing</th>
<th>f. Peaks and troughs are formed as sea level rises and falls.</th>
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<tr>
<td><img src="image5" alt="Diagram" /></td>
<td><img src="image6" alt="Diagram" /></td>
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**Figure 9.** Growth characteristics of microatolls under different conditions of sea level variation (after Woodroffe and McLean, 1990).

Research into microatolls has shown that they are potential recorders of past sea levels as their upper surfaces vary in elevation with local sea level (Smithers and Woodroffe, 2000). Microatolls from the Maldives and Cocos Islands (Indian Ocean) and from Kiribati (Pacific Ocean) have been shown to record a slight fall in sea level over the 1980s (Woodroffe and McLean, 1990). Kiribati microatolls have also been observed to respond to the lowering of sea level at the end of the 1982-83, 1986-87 El Niño events (Woodriffe and McLean 1990). Due to the temporary rise and the subsequent fall of mean sea level during the passing of El Niño, ridges are apparent in
the microatopographies of the corals. These ridges were used to construct a growth rate related to the occurrence of El Niño events for a modern microatoll from the reef crest at Christmas Island (Woodroffe and Gagan, 2000).

Recent variability in the Southern Oscillation has also been described in the $\delta^{18}$O in coral samples from Tarawa Atoll, in the western central Pacific (Cole et al., 1993). Their study relates positive isotopic shifts in $\delta^{18}$O to intense rainfall that accompanies enhanced convection during El Niño extremes. Elsewhere in the Pacific, massive Porites from the Galapagos, eastern Pacific, appear to record El Niño events in their skeletal $\delta^{18}$O signatures (Linsley et al., 1999). Coupled St/Ca and $\delta^{18}$O analysis of Porites from Christmas Island has shown that the $\delta^{18}$O signature is primarily driven by SST rather than rainfall induced salinity variation, which is secondary (Evans, 1999).

There is an absence of data from the central equatorial Pacific region that can be used to extend current understanding of ENSO variation to pre-instrumental history. A modern $\delta^{18}$O record from a Christmas Island microatoll has been shown to correlate well with instrumental SST and rainfall data and is used as a basis for the interpretation of the $\delta^{18}$O signature from a late Holocene microatoll (Woodroffe and Gagan, 2000).
Methods

3.1. Collection and sampling.

Microatoll slices were collected by Colin Woodroffe and Roger McLean in 1991 and by Colin Woodroffe, John Morrison and David Kennedy in 1999 from a series of sites within the interior of Christmas Island. All fossil sites occurred at the edges of hypersaline lakes that were previously part of a more extensive lagoon system, and modern sites were on the reef flat surrounding most of the atoll (Figure 10). Vertical slabs of fossil and modern microatolls of up to 5 cm in width were cut in the field parallel to the growth axis. Two modern microatolls (CW3 and XM0) from just behind the reef crest at Northeast Point and four fossil microatolls (CW2, XM1, XM6, XM9) from the interior have been sliced and subsampled for several analyses (Figure 11).

Figure 10. Christmas Island and locations of modern and fossil microatolls.

The delicate nature of microatolls resulted in secondary fracturing of many slices, fossils CW2, XM6 and XM9 were broken. This has necessitated in the ‘stitching’ together of the X-ray images using the image analysis package, Adobe Photoshop.
Each slab of fossil coral was sliced to a thickness of 6 mm, the field sample of the modern coral, XM0, has a maximum and minimum thickness of 15 mm and 9 mm respectively and could not be feasibly cut further. Two subsampling strategies were employed. Initial subsampling followed the detailed milling of corals recommended by Gagan et al. (1994). In this procedure a ledge of 2 mm by 2 mm in cross-sectional area was milled from the lower edge of the slice. This microsampling procedure (applied to CW3 and CW2, at least 80 mm below the upper surface of the microatoll) was designed to provide a minimum of 50 samples per annual growth increment, despite large differences in the annual extension rate among coral colonies, to ensure that the full range of annual SST cycle was well defined (Gagan et al. 1998). Despite the microsampling, only every 4th or 8th sample was analysed, achieving approximately monthly sampling. Subsequently, samples were drilled directly from the lower edges of slices XM0, XM1, XM6 and XM9 at 1.5 mm and 3 mm intervals using an engraver.

3.2. Radiocarbon dating.

Radiocarbon ages for material from fossil microatolls were determined by liquid scintillation at the Radiocarbon Dating Laboratory at the Australian National University (ANU). Conventional dates are reported in $^{14}$C Years BP (Stuiver and Polach, 1977) and have been calibrated using the Calib program of the University of Washington at calib.org.com adopting a marine reservoir correction of $-450 \pm 35$ years (Gillespie and Polach, 1979, Woodroffe and Gagan, 2000). The large carbon reservoir of the oceans causes the conventional radiocarbon ages of oceanic samples, such as corals to be several hundred years older than their true age. It is therefore necessary to use a
correction to determine a more realistic age for marine organisms, the average global ocean reservoir correction is approximately 400 years and the Calib program uses this default value. However, due to complexities in ocean circulation the actual correction varies with location and is known as AR (Stuiver and Braziunas, 1993). The AR value used in this study was -50 ± 35 (Woodroffe and Gagan, 2000). The results of this calibration give 2 sigma maximum and minimum age ranges and a calibrated age between the two range values.

3.3. Isotope analyses.

Isotope measurements for all samples apart from XM6 were carried out at the ANU using an automated individual-carbonate reaction (Kiel) device coupled with a Finnigan MAT-251 mass spectrometer. Those of XM6 were carried out at the University of Wollongong (UoW) using a Multiprep coupled with a Fisions Prism III mass spectrometer. Isotope ratios are reported as per mil (‰) deviations relative to Vienna – Peedee belemnite (V-PDB) and calibrated using the National Bureau of Standards NBS19, with internal precision of 0.06‰ (ANU) and 0.05‰ (UoW) for a typical 150µg sample.

In the reaction device (Kiel or Multiprep) phosphoric acid reacts with the calcium carbonate sample producing CO₂, which contains the isotopes ¹⁶O, ¹⁸O, ¹³C and ¹²C. The CO₂ gas is sent to an ionisation chamber where it is bombarded by electrons before the ion CO₂⁺ is accelerated through an electromagnet and collected in collectors where the isotope ratios are recorded. Standardised with V-PDB and calibrated with NBS19, the subsequent δ¹⁸O and δ¹³C ratios are obtained.

3.4. X-radiography, computerised tomography and imaging.

X-radiography.

All samples have been X-rayed at the South Coast X-ray Unit in Wollongong using a Kodak T-mat G film in Kodak x-omatic cassettes with LANGX regular screens. Each sample needed a different setting due to the different cross-sectional thicknesses. These settings varied between 50-40 kv/100-200 mA for 0.03-0.01 s all at a focal
distance of 1 m. The X-rays were scanned using an AGFA Arcus II flat bed scanner with the lid of the scanner removed so that light can enter from above, a method used to scan transparencies. The images were then saved as TIFFs on to Compact Disc for image analysis. TIFFs are 8-bit images whose pixels are represented by unsigned integers ranging in value from 0-255. It is PC convention to display pixels with a value of 0 as white and pixels with a value of 255 as black. Pixel values between 0 and 255 are variations of grey. The pixel range 0-255 is known as the grey scale, a term used here after. X-ray measures the density through a medium and displays the variation as a black and white image where white areas are high density and dark areas are low density. The computerised negative of an image will be an image of density variation defined by the pixel values where lowest densities are white or 0 on the grey scale and highest densities are black or 255 on the grey scale. Thus, density variation is represented by the variation in the grey scale values of the image pixels.

**Computerised Tomography.**

The X-ray negatives were difficult to analyse due to variation in thickness in some of the samples (XM0 and XM1), thicker areas appeared too dark if the lighter areas were made visible. If the settings on the X-ray machine were changed, then the opposite was apparent where thinner areas appeared too light if the thicker areas were made visible. Clearer resolution was obtained for selected slices using a Computerised Tomography (CT) scanner, operated by the Department of Physics, University of Wollongong.

The CT scanning technique measures the density across the width of the sample at pre-defined intervals throughout the depth of the medium being observed. The idea that CT might be useful for the study of coral densities was first introduced by Dodge (1980). Logan and Anderson (1991) used a CT scanner to establish whether or not banding was sufficiently identifiable to merit sectioning a brain coral sample from Bermuda for X-radiography. They scanned the coral at 10 mm intervals and concluded that their sample contained annual density bands and that slicing for X-radiography would be worthwhile. A substantial review of CT scanning corals is presented by Bosher (1992) who reduced the sampling interval of 10 mm to 2 mm. Bosher also concluded that CT is a valuable application for the preliminary analysis of coral
densities. Although X-radiography is more precise, CT scanning is not destructive and can be used to ascertain whether or not corals contain banding, before slices are made.

CT scanning has been used in this study in order to determine the accuracy of the X-radiograph of XM0. The density bands in the X-radiograph of XM0 appeared unexpectedly ambiguous at first sight. This was attributed to the uneven thickness of the slice that was possibly distorting the image. However, the CT image of uniform thickness confirmed the ambiguous density structure of the modern sample, XM0.

An interval of 1 mm was chosen along the length of the sample with a file generated for each cross section. These files are then compiled into a DICOM file from which the CT computer can create a three dimensional density structure displayed as an 8-bit image. The density of the sample can then be analysed along any plane using the CT computer. Incompatibility between CT software and other imaging programs limited the utility of CT images. However, despite complications in retrieving the data within the DICOM files, a method was eventually found that enabled the transformation of the DICOM file into a JPEG image. This was completed using the MIRcro program which converts the DICOM data files and displays the three dimensional image.

Image analysis was undertaken on selected slices using Abode Photoshop and Scion Image. Although the CT technique promises to be a useful application in determining density structure, the knowledge of how to use the machine and accessibility to it has been a defiant factor in this investigation. X-radiography has proved more user friendly after a means of digitally representing the image had been discovered.

Imaging.
Adobe Photoshop was used to manipulate images from the X-ray and CT scans. It can darken or brighten an image or alter the contrast between colours (black and white in this case); it can cut, crop and stitch together separate images to create one master image. Cropping and stitching has been done where some of the samples were broken (CW2, XM6 and XM9) and contrast alteration has been used where the contrast in the X-ray or CT images were too large for visual analysis. The relative density features of the image are not altered by using this method as long as the ratio between
high and low grey scale values remain the same. A line profile tool has been used to assess the variation in grey scale over a selected line in the image. As density variation is depicted in X-rays and CT scans and transferred in to grey scale images, the application of the line profile is used to describe relative density by way of the grey scale.

A simpler method of displaying grey scale across the entire image, rather than just one line was found in a program called Scion Image (http://www.scioncorp.com). Scion Image is an image processing and analysis program for the IBM PC. It has a number of applications used for microscopy but is used for its ‘surface plot’ in this investigation. The surface plot command creates a three dimensional representation of density over a defined rectangular area. The three dimensions are length (x) and width (y) of the rectangular area and grey scale (z) of the pixels within the selected area.

3.5. Instrumental oceanographic data.

There are many methods used in the collection and distribution of ocean data in the equatorial Pacific. Projects such as the World Ocean Circulation Experiment (WOCE) (1990 - 2002) and the Tropical Ocean/Global Atmosphere (TOGA) (1985-1995) program were the fore-runners in a global effort to understand and predict ocean processes and mechanisms. Emanating from these projects is a global data base on SST values and other oceanic parameters used for the understanding and prediction of ocean-atmosphere processes and interactions.

Southern Oscillation Index

The Southern Oscillation Index (SOI) is calculated from the monthly or seasonal fluctuations in the air pressure difference between Tahiti and Darwin. The pressure difference is standardised in various ways by different authors (Write, 1989, Troup 1965, Trenberth, 1976 and Ropelewski and Jones, 1987) Methods of SOI computation vary but after comparison most give similar results see Elliot and Angell (1988) for an examination of the relationship between a number of different SO indices. In this project the data set from the Climate Research Unit (CRU) at East Anglia University is used.
To form this SOI, the annual cycle of pressure at each station was removed by forming anomalies from the long-term monthly averages. These averages were then normalised by the appropriate monthly standard deviations and then the difference, Tahiti minus Darwin, was taken. The 1951-1980 period was used as the base period for computations of the means and standard deviations. The SOI is used in a number of scientific studies such as studies of Australian rainfall and weather patterns (McBride and Nicholls, 1983), the relationship between sea surface warmings off the coast of Peru (El Niños) and the Southern Oscillation (Deser and Wallace, 1987) and the examination and identification of El Niño-Southern Oscillation events (Rasmusson, 1985). In this study the SOI for the time period 1975-1999 is used as the approximate time period represented by the modern microatolls.

**Sea Surface Temperature**

Two SST data sets are used in this project, IGOSS data for the correlation of isotope data and TOGA-TAO Array data for the reconstruction of SST fields during extreme ENSO events. IGOSS gives data on a specified location and is derived from various different sources. The TOGA-TAO Array buoys are an array of moored buoys, spanning much of the width of the equatorial Pacific and can be used to create maps of SST or other physical parameters of the Pacific.

**IGOSS**

The Integrated Global Ocean Services System (IGOOS) consists of three components

1. IGOSS Observing System (IOS)
2. IGOSS Data processing and Services System (IDPSS)
3. IGOSS Telecommunications Arrangements (ITA)

The IOS consists of naval ships, research vessels and merchant ships that are members of the Ships-Of-Opportunity Program (SOOP) and fixed and floating buoys which all transmit oceanographic data such as surface temperature and salinity in near real time to the IDPSS. The IDPSS is a network of several types of national, specialised and world data centres for processing and dissemination of data and data products. Through the Global Telecommunication System (GTS) of the WMO the ITA transmits data reports with headers that correspond to the nation which made the observation and placed data on the GTS. The principle of free and open data exchange for all member
states is accomplished through GTS. IGOSS data is quality controlled at sea by member states who acquired the data and data centres where they are achieved. Monitoring is performed by the IGOSS-Operations Co-ordinator on a routine basis.

The purpose of IGOSS is to provide the necessary means for the observation, analysis and prediction of important ocean features, the results of which are made available to commercial fisheries, hydrology, marine exploration, disaster prevention, marine pollution and ocean modelling, on an operational basis normally with in 30 days or less. These results are used for both operational and research applications.

IGOSS data is available at Ingrid.ldeo.Columbia.edu/SOURCES/.IGOSS/ and consist of data sets dating back to November 1981. IGOSS SST data is used in this project to assess the $\delta^{18}\text{O}$ records from modern microatolls as oceanic proxies of ENSO.

**TOGA-TAO Array**

One of the main objectives of the ten year (1985-1995) Tropical Ocean Global Atmosphere (TOGA) project was to develop an ocean observing system to support studies of large scale ocean-atmosphere interactions on seasonal to interannual time scales. After the large El Niño event of 1982-83, which was neither predicted nor detected until nearly at its peak (McPhaden, 1995), a need for an observing system on an oceanic scale was acknowledged. The Tropical Atmosphere Ocean (TAO) Array is this system. Completed in the final year of TOGA, the TAO Array is an array of moored buoys that spans the equatorial Pacific between 8°N and 8°S. The array is made of three types of instruments. The Autonomous Temperature Line Acquisition System (ATLAS) moorings were the initial moorings used in the TAO project, developed by the Pacific Marine Laboratory (PMEL). The ATLAS moorings measure surface winds, air temperature, relative humidity, sea surface temperature and ten subsurface temperatures from a 500 m long thermistor cable. Surface Acoustic Doppler Profiler (ADCP) moorings are subsurface moorings that measure subsurface currents, temperature and pressure. The most recent addition to the array are the Triangular Trans-Ocean Buoy Network (TRITON) moorings maintained by the Japan Marine Science and Technology Centre (JAMSTEC). These buoys are deployed every 2-3° of latitude at intervals of 10-15° of longitude and measure the oceanographic and
meteorological variables of the Equatorial Pacific and telemeter real time data to shore via satellite relay. This data is made available for the scientific community at www.pmel.noaa.gov/tao and is used in this project to construct SST maps of large scale ENSO events and the difference between them. Data can be retrieved by entering the dates, variable and particular buoys of interest. These maps show why Christmas Island is an ‘optimum site for the monitoring of the thermal oceanographic signal of the full ENSO cycle’ (Evans et al. 1999).

Rainfall

Rainfall records for various islands in the Pacific have been compiled by Tony Falkland, of ACTEW Corporation (tony@ecowise.com.au). Monthly rainfall totals for Christmas Island are used for the period of interest and have been made available by a groundwater project on the atoll (Falkland and Woodroffe, 1997).

Sea level

Sea level data is recorded by the University of Hawaii Sea Level Centre/National Oceanographic Data Centre Joint Archive for Sea Level. The station, Christmas B is located at 157°28.3’W, 01°58.9’N and uses 3 instruments.

2. (11/1982 - present) Handar Encoder 436A (ENC)
3. (12/1988 - present) Handar Encoder 436B (ENB)

Monthly data is used and computed by a simple average of all daily values, calculated if seven of fewer days are missing. All heights have been referred to the station tide staff zero, which is linked to fixed bench marks. The daily and monthly means reveal no apparent changes in the reference level and agree with Canton.


A number of statistical methods have been explored during the completion of this project. Three programs are used (Excel, SPSS and SAS). Excel is used for simple descriptive statistics and graphing results of the other statistical packages. Descriptive or summary statistics such as the range, minimum, maximum, averages (mean) and standard deviations where calculated in Excel. As the project deals with time series the mean average is the state to which data would be expected to reach based on the history
of the data. These descriptive statistics are used for comparison of ENSO intensity (ranges), together with how far data ranges from its statistical norm (mean) and the variation (standard deviation) of ENSO through the time periods of interest.

Running averages can be computed in Excel and SPSS and are used to smooth data curves so that the underlying variation in the data can be observed without small, short term oscillation. If a data set has monthly values, a twelve point running average would represent the average annual variation. This can easily be carried out with the SST, rainfall and SOI data as intervals are monthly. To do the same for the isotope data, a chronology is required. After an internal skeletal chronology for the modern δ¹⁸O is defined, linear interpolation between δ¹⁸O values extends the data set to monthly intervals. A twelve point running average is then used to define annual variation. The fossil data sets were not extended using linear interpolation and the running averages are based on the growth rates defined by annual density banding. For example, if a growth rate of 24 mm yr⁻¹ is established for a sample and the sampling interval was 3 mm, then an eight point running average was used to define the annual variation. In this manner annual variation in the modern δ¹⁸O can be compared to that of the SOI and simulated in the fossil δ¹⁸O.
4.1. Introduction.

This chapter begins with a reconsideration of the $\delta^{18}O$ signature from a modern microatoll slice collected from Christmas Island in 1992 (CW3). Preliminary comparison of the $\delta^{18}O$ signature with ENSO variation, as seen in instrumental SST and rainfall data, indicates a strong correlation (Woodroffe and Gagan, 2000). A second modern microatoll, XM0 sampled in 1999 extends the record of CW3 and together, their $\delta^{18}O$ signatures show a pattern that reflects variation in instrumental SST and rainfall data associated with ENSO.

Annual signatures of $\delta^{18}O$ are constructed using linearly interpolated monthly values and 12 point running averages. They are subsequently compared to the annual signature (12 point running average) of the SOI. The comparison of the modern annual $\delta^{18}O$ and SOI signatures introduces the possibility of fossil $\delta^{18}O$ signatures being used as a proxy for pre-instrumental SOI.

The modern $\delta^{18}O$ record relies on a time scale that is developed using ENSO induced growth peaks in the microtopographies of the corals and instrumental data of SST and rainfall, both of which are non-existent in the fossil records. While observing the modern density structures it was noted that in some parts high density bands coincided with the ENSO induced elevations of surface morphology and that this relationship would be worth further investigation. Ridges in the surface morphology are related to local sea level rise caused by the decline of the equatorial sea level gradient during the warm El Niño phase of ENSO. Therefore, consideration was given to the relationship between high density and El Niño events. Using images provided by Computerised Tomography scanning and image analysis, a relationship between high density and El Niño events was observed and a method of describing annual density banding in the fossils was obtained.
4.2. **ENSO variability defined in $\delta^{18}O$ from modern microatoll, CW3.**

Preliminary analysis of a modern microatoll from Christmas Island indicated that a record of substantial sea surface temperature variations are preserved in the oxygen isotope geochemistry of the coral skeleton, and that these are related to ENSO (Woodroffe and Gagan, 2000). A modern microatoll sample CW3, from within a field of microatolls at Northeast Point, contained a $\delta^{18}O$ record for the period 1978-June 1991. Figure 12 demonstrates a correlation with the $\delta^{18}O$ and instrumental SST and rainfall data, establishing that the $\delta^{18}O$ variation from within these microatolls is one that captures the ENSO signal. The chronology for $\delta^{18}O$ in Figure 12 is based on an average extension rate of 18.5 mm yr$^{-1}$, tied to the surface morphology of the sample (sea-level depressions in 1983 and 1988).

![Figure 12. Comparison between the CW3 $\delta^{18}O$ record (a), IGOSS SST for the period November 1981-June 1991 (b) and monthly rainfall for the period January 1978-June 1991 (c) (after Woodroffe and Gagan, 2000).](image)

The rainfall record from January 1978-June 1991 shows a variation that is a convective response to the SST, recorded for the period November 1981-June 1991. The $\delta^{18}O$ signature from CW3 records the SST and rainfall variation remarkably well and both the large El Niños of the 1980s, 1982-83 and 1987-88 can be observed.
4.3. **ENSO variability defined in δ\(^{18}\)O from modern microatoll, XM0.**

Another microatoll was sampled and collected from the same location in 1999. This has been analysed as part of this study in order to extend the record preserved in CW3, and to give further support to the correlation between ENSO variation and the δ\(^{18}\)O signature apparent in Figure 12. The overlap between the two time periods represented by the corals provides the opportunity to examine the intercomparability of adjacent corals from within the same field.

Gagan *et al.* (1994) used δ\(^{13}\)C as an annual recorder linked to mass spawning events in corals from the western equatorial Pacific, a method adopted by Evans *et al.* (1998) for Christmas Island corals. However, the timing of mass spawning events in the central equatorial Pacific has not yet been established as they have for the western Pacific and the δ\(^{13}\)C and δ\(^{18}\)O records from these microatolls do not show an annual pattern (Figure 13 a and b), therefore this method has not been adopted.

![Figure 13](image)

**Figure 13.** Comparison of δ\(^{18}\)O (a), δ\(^{13}\)C (b) and density structure (c) of XM0. The sample points (upper line) can be seen in the X-ray, lower points represent intermediate samples not assessed in this study. Annual variation does not appear in any of the representations.

Annual density bands have been used to construct time scales for coral proxies since Knutson *et al.*, (1972). The density structure was therefore examined in XM0.
using X-radiography and CT scanning, in search of a pattern that could define a chronology for the δ¹⁸O record. Although present, density bands did not appear to be annual in their nature (Figure 13 c) and this method was subsequently rejected. However, some high density bands were noted to coincide with raised features in the microtopography of the sample (Figure 14).

Microtopographies have been observed to respond to ENSO induced sea level oscillation at this location (Woodroffe and Gagan, 2000), which provides one method to define the required chronology. A method of peak matching low δ¹⁸O with high SST and rainfall values was adopted and with the features from the microtopography an internal skeletal chronology for XM0 was constructed. Although density and isotope measurements do not allow reconstruction of years on their own, the time scale for this sample has been reconstructed using two methods.
First, raised features in the microtopography that correspond to ENSO induced sea level rise are used as indicators of timings for $\delta^{18}O$ sample points (small white dots) found on the X-radiograph negative of the slice. Following the path of a density band from the top of the slice to the sampling line, groups of points on the sampling line are assigned time boundaries related to three periods of sea level rise that are represented in the microtopography (Figure 14 a-c). The time boundaries are then traced to the sampling line of the XM0 slice where sample ID points are assigned these boundaries (Figure 14).

Second, peak matching of negative $\delta^{18}O$ values to high IGOSS SSTs and high monthly rainfall amounts. Using the $\delta^{18}O$ values that are assigned time boundaries, the most negative values are matched with the largest monthly SST and rainfall values and linear interpolation between the remaining $\delta^{18}O$ values extends this chronology (Figure 15).

Figure 15 clearly demonstrates that the $\delta^{18}O$ signature from XM0 reflects the ENSO variation found in the instrumental SST and rainfall records for the period 1989-August 1999. The $\delta^{18}O$ record captures both El Niño (warm) and La Niña (cool) events that occur during this period. Towards the end of the period, $\delta^{18}O$ is falling from high...
values recorded in late 1988-early 1999 to lower values, as SST warms and rainfall increases after the large La Niña of 1998-99. The large El Niño event of 1997-98 where warm SSTs and high rainfall were apparent is recorded in low δ¹⁸O values for the same period as is the prolonged, low amplitude warm event of 1990-95. At the beginning of the period (January 1989), δ¹⁸O values are higher than at any other point indicating the La Niña event of 1988-89 that is recorded in significantly cooler SST and low rainfall at Christmas Island. The Pearson Correlation coefficient for the entire period is 0.83 (where p=0.01) supporting the evidence found in the graph. Using this chronology, a growth rate can be calculated using the temporal length of the record (10 years, 8 months) and the measured length of the sample (249 mm), the result of which is 23.3 mm yr⁻¹.


The time period for the CW3 record is January 1978-June 1991 (Woodroffe and Gagan, 2000) and that of XM0 is January 1989-August 1999, thus it is possible to extend the former record by 8 years, 2 months (Figure 16).

![Figure 16. Comparison between the CW3 and XM0 δ¹⁸O records (a), monthly IGOSS SST for the period November 1981-August 1999 (b) and monthly rainfall for the period January 1978-August 1999 (c).]
There appears to be a 2.5 year overlap between the two $\delta^{18}$O records where the Pearson correlation coefficient for the period is 0.69 (where $P=0.01$) showing a relatively good correlation between microatolls from the same field. The pattern of change is similar although XM0 reaches greater extremes of high and low. The XM0 data records both the dry, cool period of the La Niña event of 1988-89 and the warmer, wetter period between May 1990 and February 1991 better than the CW3 data. The CW3 data does not extend to such a positive $\delta^{18}$O extreme nor is the positive excursion in August 1990 recorded in the SST or rainfall values.

The mean $\delta^{18}$O value for the XM0 record (-5.17‰) is lower than that of CW3 (-5.02‰) indicating warmer SST and higher rainfall throughout the time period. However, if the SST and rainfall means for the respective time periods are examined the difference in the $\delta^{18}$O mean can be explained. The mean SST for the XM0 period is 27.2°C whereas that of CW3 is 27.1°C and the rainfall mean for XM0 is 95.1mm whereas that of CW3 is 87.1mm. These values support the lower $\delta^{18}$O mean for XM0 and are most likely the result of the prolonged low, amplitude El Niño of 1990-95.

The magnitude of ENSO variation is indicated by the range in $\delta^{18}$O from each sample. The range in XM0 (1.44‰) is larger than that of CW3 (1.03‰). Providing temperature is the controlling factor, this would represent a difference of about 2°C if a relationship between SST and $\delta^{18}$O where a fall of 1°C would cause a rise of 0.18-0.22 ‰ (Gagan et al., 1994) were to be adopted. This magnitude is reflected in the higher SST and rainfall values for the XM0 time period but is not reflected in the SST data (0.1°C for respective time periods) alone. Lending support to $\delta^{18}$O being a combined signal of SST and rainfall and that $\delta^{18}$O is substantially reduced during periods of rainfall extremes.

In both of these modern microatoll records there is a strong correlation between $\delta^{18}$O and SST. The Pearson correlation coefficient using linearly interpolated monthly $\delta^{18}$O is 0.80 for 1991-1983, decreasing to 0.66 for 1991-1981 for CW3 and 0.83 for the time period represented by XM0. These values (where $P=0.01$) support the evidence found in Figure 16 where correlation is strong in both records but better in XM0. The most recent $\delta^{18}$O record from XM0 also supports the correlation between the $\delta^{18}$O
record from CW3 and ENSO variation. Indicating that one, CW3 was not a coincidental recording and two, that microatolls from the same field consistently record a variation of ENSO that reflects that of SST and rainfall.

4.5. **Annual signatures of δ¹⁸O and the SOI.**

The Southern Oscillation drives the strength of the easterly trade winds and therefore the sea surface gradient and positions of the warm and cool water between the western and eastern Pacific. ENSO events are classified by their SOI values (negative for El Niños and positive for La Niñas). Comparison of the annual SOI and δ¹⁸O cycles will establish whether or not the annual δ¹⁸O captures the intensity of ENSO events as described by the SOI. Annual signatures are derived by calculating the twelve point running average of the monthly data sets, smoothing out monthly variation and describing the annual variation of ENSO since January 1978 (Figure 17).

![Figure 17](image)

*Figure 17.* The annual SOI for the period 1978-1999 (the y-axis is reversed for comparison with δ¹⁸O) and the annual δ¹⁸O signature for the same period given by microatolls CW3 (red) and XM0 (blue).

All major ENSO events according to the SOI are apparent in δ¹⁸O record. El Niño events of 1979-80, 1982-83, 1986-87, 1990-95 and 1996-97 and La Niña events of 1988-89 and 1998-99 are recorded as substantial excursions away from their means (horizontal lines) in all three records. By obtaining internal skeletal chronologies for
the fossil microatolls and calculating their annual signatures, a proxy for the SOI can subsequently be obtained using this method.

4.6. Density structure of modern samples.

X-radiography of corals provides an idea of density variation within the coral skeleton which reflects calcification rates. X-rays provide an insight into the growth pattern of the coral. The density structure in XM0 (Figure 13) is not banded in a manner that would represent annual intervals as defined by the chronology in Figure 15. Density variation does not appear to represent any seasonal pattern. This is also the case for SST and rainfall at this location (Figure 15), which indicates that density and SST and rainfall may be related. Two major features in the density structure of this sample are, one, the pattern of banding is not at all regular and two, darker (higher density) regions coincide with elevations of the microtopography implying that density is greatest during periods of high sea level or warmer surface temperature. Using the δ¹⁸O records from both XM0 and CW3 and the IGOSS SST curve for the period (November 1981-Aug 1999) a comparison of ENSO and density variation is made (Figures 18 and 19).

Figure 18 shows the CT scan and the grey scale surface plot of XM0 (a) the IGOSS SST curve (b) and the average density profile from the surface plot (c). The surface plot is a combination of line profiles across the width of the CT image measured at 1.5 mm intervals. Peaks can be seen to resemble each other through the width of the image representing high density banding although they are not aligned perpendicular to each other (a). The lack of alignment is due to the variation in density that can be observed in the CT image. El Niño and La Niña events can be observed in the surface plots and the density profile although they do not rise and fall in a manner that resembles the SST curve. The 1998-99 La Niña event is apparent at approximately 1-2 mm from the right hand side of the surface plot (first red line) and is followed by the El Niño of 1997-98 that is represented between 2-5 mm (second red line). This event is more obvious in the surface plot than it is in the average density profile because the peaks are not perpendicular to one another and are lost in averaging. The most obvious feature of the density structure is the period between 5 and 20 mm (third and fourth red lines) that is attributed to the prolonged El Niño of 1990-95.
Figure 18. CT images and Scion Image surface plots of grey scale for XM0 (a). The IGOSS SST curve (b) and the average of the density profiles from the surface plot (c).
Figure 19. CT images and Scion Image surface plots of grey scale for XM0 (a). The IGOSS SST curve (b) and the average of the density profiles from the surface plot (c).
Approximately 10 major peaks are apparent in the density structure through this distance, more obvious in the density profile than the surface plot. This implies two peaks per year in this high density region caused by the anomalous warming of SST through out the five year period. After this there is a fall in density (fifth red line) attributed to the decrease in SST of the La Niña event of 1988-89.

Figure 19 shows the same as Figure 18 for the microatoll CW3. The two El Niño events (1982-83 and 1986-86) for this record are recorded in the density structure (red lines). The surface plot and the density profile record both the El Niño events, although the recording of the 1982-83 event is better than that of 1986-87. Between the two events density decreases, which is represented in the surface plot better than the density profile. SST also decreases during this period and rises to two small consecutive peaks which are apparent in the surface plot (dashed lines), although not as obvious as other peaks. After approximately 16 mm there are four major peaks in the density structure that are attributed to the years 1978-81 where there is no SST record. This was a warm period of ENSO with an El Niño event in 1979-80. The four high density peaks are also observed in the CT image where high density bands are apparent.

Although the SST curve is not well represented in the density structure, prolonged ENSO events appear to be recorded by higher densities. Density appears to respond to periods of prolonged warming rather than short term, intense warming that is characteristic of severe El Niño events. An annual banding pattern is not apparent in these microatoll samples, the best example of which is the first four years of growth found in CW3 where annual density banding does appear to be represented. The analysis of the density structures in this manner has defined a method that enables a density pattern, that is not well represented in an X-ray or CT image, to be defined.
Analysis of fossil microatolls.

5.1. Introduction.

Analysis of the fossil density structures has been conducted in order to define a pattern that can be used to construct internal skeletal chronologies. The methods of image analysis used in Chapter 4 were applied to X-ray images of the fossil microatoll slices, CW2, XM1, XM6 and XM9 and chronologies for the δ¹⁸O data were subsequently defined.

Using these chronologies, analysis of the δ¹⁸O records in both intensity and temporal terms was possible. Annual patterns of variation in the δ¹⁸O signatures were calculated using the defined growth rates that allowed the periodicity of ENSO to be compared throughout the time periods involved.

Typical ENSO event characteristics were subsequently determined from each of the δ¹⁸O records, in terms of mean amplitude and duration. These models were used to describe what typical ENSO events may have been like in terms of mean amplitude and duration in the Holocene and how they compared to typical modern ENSO events.
5.2. Radiocarbon dating and ages of fossil microatolls.

Radiocarbon ages have been determined for 27 coral samples from the interior of Christmas Island, 21 of which are fossil microatolls (see Chapter 3). The conventional ages are given in $^{14}$C years before present (BP) and are converted into calibrated years BP (Table 1). The ages of these samples provides an interesting insight to past climatology of the island (Figure 20). By simply looking at this coral chronology some speculation can be made on the environmental conditions of this location by the life and death of these randomly sampled specimens. In the late mid-Holocene (~ 4-3 ka) the samples are relatively evenly spaced throughout the time period implying continuously suitable conditions for the growth of microatolls. At approximately 3 ka change is implied, resulting in the apparent death of microatolls in the lagoon on Christmas Island when conditions became unsuitable for microatoll growth. After this 1000 year absence of microatolls, a return to favourable conditions is indicated at around 2 ka, as six of the samples have been dated in this period. By analysing samples from these two time periods (before 2 ka and between 3 and 4 ka), an insight into the environmental conditions of the central Pacific may be inferred and some indication of ENSO variation might be sought from the $\delta^{18}$O. CW2 (1740 years BP) has been analysed for the late Holocene example and three samples (XM1, 2924 years BP, XM6 3258 years BP, and XM9 3571 years BP) are analysed for the mid-Holocene time period.

![Figure 20](attachment:image.png)

**Figure 20.** The temporal distribution of microatoll samples collected at Christmas Island that date through the mid-late Holocene.

5.3. Defining internal skeletal chronologies for the fossil samples.

The two methods used in defining an internal skeletal chronology for the modern samples are of no use in defining those of the fossil samples. The surface topography of the microatolls cannot be used as they have undergone between 1700 and 4000 years of weathering, rendering the interpretation of their surface microtopographies impossible.
Table 1. Radiocarbon dates, marine reservoir corrected age and calibrated age ranges of corals from Christmas Island (after Woodroffe and McLean, 1998).

<table>
<thead>
<tr>
<th>Lab Coad</th>
<th>Field I.D.</th>
<th>Material</th>
<th>Conventional Radiocarbon age (years BP)</th>
<th>$2\sigma$ maximum of calibrated age range (cal. age) and minimum cal. age (years BP)</th>
</tr>
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<tbody>
<tr>
<td>11211</td>
<td>XM2</td>
<td>Porites Microatoll</td>
<td>1970 ± 60</td>
<td>1620 (1477) 1307</td>
</tr>
<tr>
<td>7962</td>
<td>XI12</td>
<td>Porites Microatoll</td>
<td>1980 ± 60</td>
<td>1629 (1489) 1316</td>
</tr>
<tr>
<td>11212</td>
<td>XM3</td>
<td>Porites Microatoll</td>
<td>2130 ± 70</td>
<td>1839 (1664) 1479</td>
</tr>
<tr>
<td>7958</td>
<td>XI15</td>
<td>Porites Microatoll</td>
<td>2130 ± 90</td>
<td>1875 (1664) 1414</td>
</tr>
<tr>
<td>Beta12532</td>
<td>CW2</td>
<td>Porites Microatoll</td>
<td>2210 ± 80</td>
<td>1943 (1740) 1530</td>
</tr>
<tr>
<td>7956</td>
<td>XI18</td>
<td>Porites Microatoll</td>
<td>2230 ± 100</td>
<td>2005 (1780) 1519</td>
</tr>
<tr>
<td>11218</td>
<td>XM10</td>
<td>Porites Microatoll</td>
<td>2290 ± 50</td>
<td>1979 (1842) 1697</td>
</tr>
<tr>
<td>11210</td>
<td>XM1</td>
<td>Porites Microatoll</td>
<td>3200 ± 50</td>
<td>3105 (2924) 2771</td>
</tr>
<tr>
<td>11216</td>
<td>XM8</td>
<td>Porites Microatoll</td>
<td>3300 ± 60</td>
<td>3261 (3067) 2865</td>
</tr>
<tr>
<td>7954</td>
<td>XI20</td>
<td>Porites Microatoll</td>
<td>3360 ± 110</td>
<td>3422 (3158) 2839</td>
</tr>
<tr>
<td>11214</td>
<td>XM6</td>
<td>Porites Microatoll</td>
<td>3440 ± 60</td>
<td>3400 (3258) 3068</td>
</tr>
<tr>
<td>11219</td>
<td>XI36</td>
<td>Porites Microatoll</td>
<td>3460 ± 50</td>
<td>3404 (3298) 3131</td>
</tr>
<tr>
<td>7966</td>
<td>XI4b</td>
<td>Porites Microatoll</td>
<td>3600 ± 80</td>
<td>3636 (3434) 3242</td>
</tr>
<tr>
<td>7960</td>
<td>XI14a</td>
<td>Porites Microatoll</td>
<td>3670 ± 70</td>
<td>3694 (3505) 3342</td>
</tr>
<tr>
<td>7963</td>
<td>XI11</td>
<td>Tridacna</td>
<td>3700 ± 90</td>
<td>3812 (3552) 3370</td>
</tr>
<tr>
<td>11217</td>
<td>XM9</td>
<td>Porites Microatoll</td>
<td>3720 ± 70</td>
<td>3797 (3571) 3382</td>
</tr>
<tr>
<td>7960</td>
<td>XI4a</td>
<td>Porites Microatoll</td>
<td>3730 ± 70</td>
<td>3694 (3505) 3342</td>
</tr>
<tr>
<td>11213</td>
<td>XM4</td>
<td>Porites Microatoll</td>
<td>3750 ± 80</td>
<td>3832 (3618) 3394</td>
</tr>
<tr>
<td>7955</td>
<td>XI19a</td>
<td>in situ Millepora</td>
<td>3750 ± 120</td>
<td>3925 (3618) 3337</td>
</tr>
<tr>
<td>11215</td>
<td>XM7</td>
<td>Porites Microatoll</td>
<td>3780 ± 50</td>
<td>3818 (3637) 3474</td>
</tr>
<tr>
<td>7966</td>
<td>XI14b</td>
<td>Porites Microatoll</td>
<td>3810 ± 80</td>
<td>3907 (3681) 3462</td>
</tr>
<tr>
<td>7964</td>
<td>XI10</td>
<td>Porites Microatoll</td>
<td>4040 ± 70</td>
<td>4220 (3976) 3801</td>
</tr>
<tr>
<td>7965</td>
<td>XI5</td>
<td>Porites Microatoll</td>
<td>4150 ± 80</td>
<td>4398 (4139) 3891</td>
</tr>
<tr>
<td>7961</td>
<td>XI13a</td>
<td>Massive coral</td>
<td>5060 ± 100</td>
<td>5585 (5319) 5040</td>
</tr>
<tr>
<td>7957</td>
<td>16a</td>
<td>Massive coral</td>
<td>26100 ± 180</td>
<td>-</td>
</tr>
<tr>
<td>8202</td>
<td>BA1x</td>
<td>Tridacna</td>
<td>35810 ± 410</td>
<td>-</td>
</tr>
<tr>
<td>8203</td>
<td>BA1y</td>
<td>Coral</td>
<td>4082 ± 690</td>
<td>-</td>
</tr>
</tbody>
</table>

The peak matching technique is also impossible as the $\delta^{18}O$ records represent the non-existent instrumental data. Although annual banding was ambiguous in the modern density structures, first from X-ray and then CT scanning, this was not the case for the X-rays of the fossil samples. In contrast to the modern samples, annual banding patterns were more pronounced through several sections of the fossil corals and have been used to show that the $\delta^{18}O$ signatures have an annual nature in their oscillation. Where samples have been drilled from the fossils XM1, XM6 and XM9, holes have been left which can be seen in the X-ray negatives of the images. These sample points can be used to identify exactly where each $\delta^{18}O$ value is situated in the density structure. The annual pattern is still further shown by the surface plots from Scion Image. The surface plot is a compilation of grey scale profiles along lengths of the X-ray image at intervals of 1.5 mm across the width.
These plots show how relative density varies through the widths and the lengths of the selected rectangular areas of the microatoll slices. The selected area was chosen from just above the sampling line to the top of slice (blue box on X-ray image) so that the foremost grey scale profile represents density as close to the sampling line and therefore the $\delta^{18}$O signature as possible. The surface plots are compared to the $\delta^{18}$O signatures of each sample (Figures 21-22).

Annual banding patterns are used to determine growth rates for the samples and to determine years in the $\delta^{18}$O oscillation so that ENSO like variation can be scrutinised. Figure 21 shows the surface plots, X-ray images and $\delta^{18}$O records for CW2 (a) and XM1 (b). In CW2, the last (right hand side is the oldest point) eight years are quite clear in the surface plot and the X-ray image. After this however, the image becomes lighter (less dense) as height of the peaks in the surface plot fall. The surface plot is extremely useful as peaks and troughs are more easily recognised than difference in the grey scale through the X-ray image. The twelfth year is quite prominent followed by a trough, indicating a decrease in density, which may be associated with cool conditions represented by a positive excursion of $\delta^{18}$O at the same time. Darkness in the image and peaks in the grey scale increase after this point and five, high density bands can clearly be seen toward the beginning of the record (left hand side). Five years of annual oscillation in the $\delta^{18}$O signature also appear at the beginning of the record, reproducing the density structure. This 394 mm slice appears to represent nineteen years of growth, indicating a growth rate of 20.7 mm yr$^{-1}$ (Table 2). The record from XM1 is the shortest record of all the samples but shows a surface plot and a $\delta^{18}$O signature where nineteen years are defined, the same amount as that of CW2. The last five years are clear in both the X-ray image and the surface plot, before which contrast in the image diminishes. This is also represented in a smaller variation in the surface plot, although not to an extent where banding becomes ambiguous. Again, the surface plot is useful as banding is hard to distinguish in the X-ray image through this region of low contrast. A growth rate of 11.8 mm yr$^{-1}$ (Table 2) is calculated for this sample which is smaller than the growth rates found in both the modern and the late Holocene samples.
Figure 21. Surface plots, X-ray negatives and δ¹⁸O of CW2 (a) and XM1 (b). Annual density bands are indicated by red lines where peaks in the grey scale (high density) are apparent. Years are defined by the grey scale peaks (numbered) which are related to the oscillation in the δ¹⁸O records.
Figure 22 shows the surface plots, X-ray images and $\delta^{18}$O records for XM6 (a) and XM9 (b). XM6 represents a period of 26 years and annual banding is clear at the beginning (years 26–19) and end (years 4–1) of the record. Contrast between black and white in this image is large which can be seen in years 1, 14, 15 and 16 where the peaks are flat. In this 26 year record a growth rate of 14.3 mm yr\(^{-1}\) is implied (Table 2). The sample from XM9 is made of three slices, the first of which shows the best annual density banding in all of the samples. Twelve bands are clearly visible in this slice, as they are in the surface plot. The middle slice ends in a lighter region where two bands are visible in the X-ray image and less visible in the surface plot. Tracing this to the $\delta^{18}$O signature, years 13 and 14 are years of small variation, which might explain the decrease in density. After this, banding is better represented in the surface plot for the majority of the record. In this 33 year record a growth rate of 15.2 mm yr\(^{-1}\) is implied (Table 2).

Table 2 is a summary of the description above. The most obvious feature in this data is that growth rate appears to have increased from the mid to the late Holocene. After 2.9 ka (XM1) there is an apparent demise of microatolls in the lagoon of Christmas Island, lasting for approximately 1100 years (Figure 20), although this could be a result of random sampling. At around 1.8 ka (XM10, Table 1) there is a re-emergence of coral growth and it appears that they grew faster than they did before.

<table>
<thead>
<tr>
<th>Microatoll ID</th>
<th>Length of record (mm)</th>
<th>Number of years in record (years)</th>
<th>Growth rate (mm yr(^{-1}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>CW2</td>
<td>394</td>
<td>19</td>
<td>20.7</td>
</tr>
<tr>
<td>XM1</td>
<td>225</td>
<td>19</td>
<td>11.8</td>
</tr>
<tr>
<td>XM6</td>
<td>372</td>
<td>26</td>
<td>14.3</td>
</tr>
<tr>
<td>XM9</td>
<td>501</td>
<td>33</td>
<td>15.2</td>
</tr>
</tbody>
</table>
Figure 22. Surface plots, X-ray negatives and $\delta^{18}O$ of XM6 (a) and XM9 (b). Annual density bands are indicated by red lines where peaks in the grey scale (high density) are apparent. Years are defined by the grey scale peaks (numbered) which are related to the oscillation in the $\delta^{18}O$ records.
5.4. Mid–late Holocene ENSO events defined by fossil $\delta^{18}O$ signatures.

In the central equatorial Pacific the thermal (SST) and convective (rainfall) processes co-vary (see Chapter 4). Therefore, the variability in $\delta^{18}O$ can not be attributed to one process alone but is instead representative of the variability of ENSO which oscillates between El Niño and La Niña events (termed warm and cool events hereafter).

The fossil microatolls have similar mean $\delta^{18}O$ values ($CW2 = -4.54\%$, XM1 and XM6 = $-4.57\%$, XM9 = $-4.53\%$) that are significantly greater than that of the modern (modern mean $\delta^{18}O = -5.11\%$), indicating the possibility that Holocene mean SSTs may have been cooler. This is dealt with in greater detail in Chapter 6.4.

The annual oscillation is used to interpret the lengths of warm and cool (negative and positive excursions from the $\delta^{18}O$ mean) ENSO like events apparent in the $\delta^{18}O$ signatures. ENSO events are defined by excursions that exceed the lower (warm events) and upper (cool events) mean averages. These averages are calculated by finding the mean value for all the points that fall below (lower mean) or above (higher mean) the mean value for the entire data set.

In the late Holocene (Figure 23) the magnitude of $\delta^{18}O$ variation was similar as it is in modern times, varying by approximately 1.5%, although the pattern of variation can be seen to be much more regular as annual variation is well defined. During this 19 year period, 3 warm and 4 cool events are apparent. The first cool event (304-289 mm) happens after 5 years of low amplitude variation, rising to $-4.05\%$. This cooling sets off a 6 year period of intense ENSO activity (289–127 mm) where the magnitude of variation increases between large, positive ($-3.85\%$) and large, negative ($-6.03\%$) $\delta^{18}O$ excursions. 2 years of warm conditions (289–231 mm), comparable in magnitude ($-0.58$ and $-0.48\%$ away from their respective means) with the 1986-87 El Niño ($-0.55\%$ away from its mean) recorded in the modern record, precede the largest cool event in this record (232–218 mm rising to $-3.85\%$). After one year of insignificant oscillation the largest warm event in any of the records is encountered (198–152 mm). This is a 2 year period of 2 combined El Niño events, the first ($-5.15\%$) being smaller
than the second (-6.03%) and comparable with the previous two warm events (-0.61% away from its mean). The second year of this event reaches a negative $\delta^{18}O$ value that is unsurpassed in any of the records and indicative of a substantial rainfall peak. This may be caused by increased convection due to continued surface water heating that warmed SST to an extent that breached the thermal-convective threshold.

**Figure 23.** $\delta^{18}O$ record for the late Holocene microatoll, CW2. Annual variation is used to assess the duration of warm and cold events, indicated by excursion above or below the upper and higher mean averages (horizontal lines).

This substantial negative peak increases the range of the data by 0.8% and is considered when comparing variation magnitude with the modern signature. If ignored, variation magnitude would appear to be twice that of the present in the late Holocene, which is not the case for the remainder of the data, and this warm event is just one of three in this 19 year period that surpass modern $\delta^{18}O$ minima. After this large warm event, variation magnitude decreases in a four year period (139–79 mm) dominated by La Niña type events before returning to warmer conditions. The warm conditions last for a period of three years (78–40 mm), when a one year warm excursion of -0.26% comparable to the magnitude of the low amplitude event of 1990-95 (-0.29%) in the modern record is followed by a further two years of warming where $\delta^{18}O$ falls by -0.51 and -0.38% away from the mean. A one year cool event that rises to -3.98% is recorded before two years of insignificant ENSO oscillation complete the record.
Chapter 5

The mid-Holocene signatures all show variation at similar magnitudes, which are approximately half the magnitude found in the late Holocene (Table 3). In the youngest mid-Holocene record, XM1 (Figure 24) there appears to be two warm and two cool events of any significance. At the beginning of the record cool (216–207 mm) and warm (194–185 mm) events are apparently separated by one year of insignificant variation. There is then a two year period (180–161 mm) before $\delta^{18}O$ falls again in a warm period lasting for three years (157–127 mm). After a five year period of oscillation where no significant warming or cooling events are observed, a two year cool event of three years can be seen (64–43 mm).

![Figure 24. $\delta^{18}O$ record for the youngest mid-Holocene microatoll, XM1. Annual variation is used to assess the duration of warm and cold events, indicated by excursion above or below the upper and higher mean averages (horizontal lines).](image)

The $\delta^{18}O$ signature from XM6 (Figure 25) begins with two warm periods of one year in length each (368–326 mm). After a period of five years (322–270 mm) there is a significant cooling event that lasts for four years with two large cool events at either end of the period (270–217 mm). Immediately following this long cool period a long warm period begins, lasting for six years (217–159 mm) where $\delta^{18}O$ remains below the mean value for four years during this time (201–165 mm). The record then oscillates between warm and cool phases, increasing in cool phase amplitude over six years (154–80 mm). Warm conditions return for a period of three years (71–18 mm) before the end of the record.

![Figure 25. $\delta^{18}O$ record for the mid-Holocene microatoll, XM6. Annual variation is used to assess the duration of warm and cool events, indicated by excursion above or below the upper and higher mean averages (horizontal lines).](image)
The longest record from all of the microatolls comes from XM9 (Figure 26), which is also the oldest (3571 years BP). The period begins with a small warm event, falling by -0.14% away from the mean and lasting for one year (501-492 mm) before increasing to 0.24% above the mean for a cool period of one year (492-480 mm). A warm period which lasts for two years (474-453 mm) and exceeds the previous warm event, reaching -0.21% below the mean, follows. After this a warm event lasting for four years (437-387 mm) and falling to the lowest value of the record (-4.86%) is recorded. Over the next three years (384-345 mm) cool conditions prevail before a two year (345-312) warm period, similar in shape but at a reduced magnitude of the large warming in the late Holocene record (-0.25% as apposed to -1.49%). After an increase in δ¹⁸O (0.38% above the mean) lasting for one year (303-288 mm), variation becomes largely insignificant over the next nine years (288-159 mm). During this nine year period there is only one significant fall below the lower mean value, which is a sharp deviation due to an anomalous δ¹⁸O value of -4.81%.

![Figure 26](image)

**Figure 26.** δ¹⁸O record for the oldest mid-Holocene microatoll, XM9. Annual variation is used to assess the duration of warm and cool events, indicated by excursion above or below the upper and higher mean averages (horizontal lines).

After the long period of reduced ENSO activity the variation magnitude increases over a four year period (159-101 mm) where oscillation varies between cold (-4.17%) and warm (-4.72%) events in a see-saw like manner. This see-saw like oscillation leads to a two year cooling (96-81 mm) after which conditions develops into a large cool event, rising to -0.47% above the mean and lasting for two years (78-54 mm). In the final three years of the record (51-2 mm) conditions warm as δ¹⁸O falls in a period of oscillation below the mean value.

The variation in these records, measured by standard deviation (SD) of the data sets (Table 3), show a pattern that increases from the mid to late Holocene (mid-Holocene average SD = 0.15, late Holocene SD = 0.31) and decreases by present times.
(modern average SD = 0.22). However, the late Holocene SD is not representative of the whole data set as the large El Niño like event that extends to -6.03‰ substantially increases the measure of variability. If the δ¹⁸O values that cause this anomaly are removed, a late Holocene SD of 0.28 is calculated, which is closer but still larger than modern variation (0.22) recorded in the SD (Table 3).

Table 3. Summary of growth rate, mean, range and standard deviation of each sample.

<table>
<thead>
<tr>
<th>Microatoll I.D.</th>
<th>Growth rate (mm yr⁻¹)</th>
<th>δ¹⁸O mean (‰)</th>
<th>δ¹⁸O range (‰)</th>
<th>δ¹⁸O SD</th>
</tr>
</thead>
<tbody>
<tr>
<td>XM0</td>
<td>23.3</td>
<td>-5.20</td>
<td>1.41</td>
<td>0.21</td>
</tr>
<tr>
<td>CW3</td>
<td>28.5</td>
<td>-5.02</td>
<td>1.03</td>
<td>0.23</td>
</tr>
<tr>
<td>CW2</td>
<td>20.7</td>
<td>-4.54</td>
<td>2.18</td>
<td>0.31</td>
</tr>
<tr>
<td>CW2, large El Niño removed</td>
<td>20.7</td>
<td>-4.52</td>
<td>1.31</td>
<td>0.28</td>
</tr>
<tr>
<td>XM1</td>
<td>12.5</td>
<td>-4.57</td>
<td>1.02</td>
<td>0.17</td>
</tr>
<tr>
<td>XM6</td>
<td>14.3</td>
<td>-4.57</td>
<td>0.72</td>
<td>0.14</td>
</tr>
<tr>
<td>XM9</td>
<td>14.7</td>
<td>-4.53</td>
<td>0.80</td>
<td>0.14</td>
</tr>
</tbody>
</table>

5.5. Annual variation in ENSO, a proxy for the SOI.

The method used for annual variation of δ¹⁸O and the SOI in the modern time period is repeated using the growth rates defined for the fossil δ¹⁸O records so that an insight to past annual ENSO variation can be sought. The growth rates of the fossil microatolls are used to determine annual intervals in the δ¹⁸O data sets and running averages are subsequently computed to compare the annual nature of ENSO variation recorded in the δ¹⁸O signatures.

The sampling interval for CW2 (2 mm) is smaller than those of XM1, XM6 and XM9 (3 mm) and growth rates for the fossils are different. Therefore, different running averages have to be used to define annual variation. CW2 has a growth rate of approximately 20 mm yr⁻¹ and a sampling interval of 2 mm, therefore a 10 point running average is used to describe annual variation. XM1 has a growth rate of approximately 12 mm yr⁻¹ and a sampling interval of 3 mm, therefore a 4 point running average is used. 5 point running averages are used for XM6 and XM9 as the growth rates are approximately 15 mm yr⁻¹ and the sampling intervals are 3 mm.
Figure 27. Annual variation in all records defined by the running averages and periods of El Niño (red boxes) and La Niña (blue boxes).
These running averages are compared to the modern record that has been shown to compare well with the SOI (Figure 27). The mid-Holocene records appear similar in their variation although the magnitude of cool events gradually decreases through the time period from 0.33‰ away from the mean (XM9) to 0.29‰ (XM6) to 0.23‰ (XM1).

The range of variation increases from the mid-Holocene (average = 0.86 ‰) to the late Holocene (2.18 ‰ or 1.31 ‰ if the extreme El Niño is neglected) about a mean δ¹⁸O value that remains essentially the same but average duration of events decrease from approximately two to one year. The late Holocene variation range (1.31 ‰) is similar to the modern (average = 1.22 ‰) with one extreme warm event, the largest in any of the records increasing the range to 2.18 ‰.

In the modern record there are seven recognised ENSO events in the 21.7 year period indicating the modern ENSO periodicity of 3.1 years. Periodicity decreased from the mid-Holocene (average = 4.2 years) to the late Holocene (3.2 years) implying an increase in ENSO frequency throughout the time period (Table 4).

The mean δ¹⁸O value decreases from the late Holocene to present indicating warmer and wetter conditions in modern times, however higher salinity in the more extensive lagoon would have caused a higher mean δ¹⁸O value that cannot be totally attributed to temperature.

<table>
<thead>
<tr>
<th>Microatoll ID</th>
<th>Number of years represented by the record</th>
<th>Number of ENSO events</th>
<th>Reoccurrence of ENSO</th>
</tr>
</thead>
<tbody>
<tr>
<td>Modern</td>
<td>21.7</td>
<td>7</td>
<td>3.1</td>
</tr>
<tr>
<td>CW2</td>
<td>19</td>
<td>5</td>
<td>3.8</td>
</tr>
<tr>
<td>XM1</td>
<td>19</td>
<td>6</td>
<td>3.2</td>
</tr>
<tr>
<td>XM6</td>
<td>26</td>
<td>5</td>
<td>4.3</td>
</tr>
<tr>
<td>XM9</td>
<td>33</td>
<td>8</td>
<td>5.2</td>
</tr>
</tbody>
</table>
5.6. Models of ENSO event amplitude and duration.

Averaging the described ENSO events within each of the three time periods (mid, late Holocene and modern) establishes typical event characteristics of the two ENSO phases. By averaging the maximum excursions (positive for La Niña and negative for El Niño) from the mean $\delta^{18}O$ value and the duration of each event, an El Niño and a La Niña that is representative of each record is described. This has been done for each of the six records and in the case of the modern and the mid-Holocene, these typical ENSO events have been averaged to produce events, representative of each time period (Figure 28).

![Figure 28](image)

**Figure 28.** ENSO events in the mid-Holocene (green line) were longer in duration but smaller in intensity than both the late Holocene (red) and modern (blue) times. Through the mid-late Holocene average warm and cold event magnitude approximately doubled (−0.26 to −0.53‰ in the warm and 0.36 to 0.60‰ in the cold events) and duration of events halved (~2.2 years to ~1.5 years).

In both of the fossil ENSO models, cool events have larger magnitudes than their warmer counterparts (0.36 and 0.60‰ for cold and 0.26 and 0.53‰ for warm in the mid and late Holocene respectively). In contrast, warm events are larger than the cool events in the modern model (0.51 and 0.43 respectively). This might be due to enhancement of the modern $\delta^{18}O$ signature due to a more convective period caused by warmer SST. In cooler and drier periods, salinity would vary less in a convective
system, giving rise to a stronger temperature signal in $\delta^{18}O$. Supporting this hypothesis is the large warm event found in the late Holocene, attributed to a substantial rainfall peak, which amplifies the average warm event to a greater magnitude. In cooler and less convective conditions, a large, prolonged SST increase would cause a sudden increase in convection and therefore rainfall and addition to the SST driven $\delta^{18}O$ signature.

These results demonstrate that the detailed analysis of fossil microatolls from Christmas Island provides a valuable insight to pre-instrumental ENSO variability. Four windows into the ENSO variability from the mid-Holocene to present have been demonstrated, and are discussed below.
Discussion, conclusions and suggestions for further research.

6.1. Modern $\delta^{18}O$ and ENSO Variability.

The modern microatoll, CW3 has been shown to contain a $\delta^{18}O$ signature that reflects the instrumental SST and rainfall records associated with the El Niño-Southern Oscillation (Woodroffe and Gagan, 2000). The consistency of this relationship has subsequently been tested by the stable isotope analysis of a further modern microatoll, XM0 sampled from within the same field eight years later. The comparison of the XM0 $\delta^{18}O$ signature with instrumental data showed an even higher correlation and substantiated the potential of modern Christmas Island microatolls to record a pattern of variation in their $\delta^{18}O$ composition that reflects the modern ENSO variability. Not only are the two records consistent, but they also share a similar pattern of variation with the Southern Oscillation Index and reinforce the potential that fossil microatolls from this location may possibly extend our understanding of ENSO variability into the pre-instrumental past.

Christmas Island is an extremely isolated landmass in a position that is crucial for the understanding of ocean-atmosphere interaction that is so far poorly understood. Its position is such that it experiences a SST and rainfall variation that co-varies and synthesises both the convective atmospheric and thermal oceanographic signal of ENSO. Other areas such as the western Pacific are of course equally important in unravelling the complexities of the ENSO phenomenon. However, they are often in regions where oceanographic and atmospheric processes are not so tightly phase locked. For example, Tarawa Atoll, approximately 3300 km to the west of Christmas Island, experiences high SSTs year round and relates an ENSO signal in the $\delta^{18}O$ signature to rainfall extremes. In a study of coral $\delta^{18}O$ from Tarawa, one of the largest El Niño events of the century (that of 1982-83) was not recorded (Cole et al., 1993) as the convective maximum bypassed Tarawa and moved out into the central and eastern Pacific (Rasmusson and Wallace, 1983).

The microatoll $\delta^{18}O$ signatures record all the recognised ENSO events, both cool and warm, that are apparent in the instrumental records. However, it appears that El
Niño events have a better representation than La Niña events which is attributed to the enhancement of the El Niño signal by extremes of rainfall. In a significantly dry period such as a La Niña event the δ18O signature is representative of SST alone whereas in an El Niño it will be a combined representation of SST and rainfall as the two processes co-vary.

6.2. Modern and fossil microatoll density structures.

The variation in δ18O from these modern Christmas Island microatolls has successfully reproduced a variation in ENSO that is evident in the instrumental data. However, the analysis of the density structures of these corals has not proved to be so successful. This analysis was undertaken in an effort to establish an internal skeletal chronology for the δ18O data as it was expected that an annual banding pattern would indicate the length of the record and annual intervals within it. It soon became clear that this assumption was premature because in fact, there were no annual indications and the banding that did occur was ambiguous with respect to an evenly spaced and distinct pattern. Fortunately, the surface morphology of these microatolls provided a basis on which to anchor specific points of the δ18O record and peak matching completed the required chronology. This approach was inappropriate for the fossil microatolls that had undergone several thousands of years of weathering. What did emanate from this procedure was a method that enables the visual interpretation of the density structure that is sometimes difficult to determine from X-radiographs or CT scans. This method of image analysis was subsequently used in the interpretation of the X-rays of the fossil samples that did in fact show a more definite pattern of annual density banding. The lengths of time represented by the fossil microatolls of between 19 and 33 years and their growth rates were subsequently defined. Together with the radiocarbon dates and the δ18O data, four windows into the pre-instrumental history of ENSO variability have been examined and are discussed below.

6.3. Mid and late Holocene δ18O signatures.

The calibrated radiocarbon ages of more than 20 fossil microatolls fall into two time periods, the mid-Holocene (4.1–3 ka) and the late Holocene (1.8-1.4 ka). Three
mid-Holocene samples and one late Holocene sample were analysed and have shown growth rates that differ between the two periods. In the mid-Holocene the average growth rate of the microatolls was 13.6 mm yr\(^{-1}\) but by the late Holocene growth rate had increased to 20.7 mm yr\(^{-1}\), a rate similar to the average of the modern samples (20.9 mm yr\(^{-1}\)). There is no clear evidence for the apparent demise of microatolls at \(~3\) ka and reappearance at around \(1.8\) ka, however the difference in growth rates may imply a change in conditions that govern microatoll growth. Alternatively, it may be coincidental that no microatolls of the age of \(~2\)\(-3000\) years were sampled. These two scenarios can only be tested by re-sampling at Christmas Island and obtaining further dates.

The magnitude of \(\delta^{18}O\) variation in the late Holocene is also different from that of the mid-Holocene although the pattern of this variation is similar and deserves some discussion. In the mid and late Holocene, \(\delta^{18}O\) sequentially rises and falls above and below the mean value but the mid-Holocene record has a smaller range (0.85% as apposed to 2.18 or 1.31% if the large El Niño event is neglected). The pattern of variation is more symmetrical in both time periods than it is in the modern and La Niña events are at least as large as El Niño events. This may imply that the enhancement due to rainfall extremes apparent in El Niño events in the modern record is not as prominent in the Holocene because conditions were cooler and drier. In the mid-Holocene the range of the variation was small compared to that of the late Holocene and the modern, implying the smallest range in temperature in a cooler/drier environment. By the late Holocene, temperature variation increased to an extent that exceeded the modern but retained its more symmetrical pattern, indicating an increase in the contrast between warm and cool conditions in a cooler and drier mid-late Holocene. Whether the transition between low and high temperature contrasting conditions between the mid and late Holocene was abrupt or gradual can not be estimated with certainty. However, it might be linked to the apparent absence of microatolls for the period and the increase in growth rate, apparent in the late Holocene.

The fossil microatolls originate from the lagoon of Christmas Island as apposed to the reef flat environment of the modern microatolls. Although the previously more extensive lagoon was likely to have been subject to conditions that were more representative of the open ocean than they are at present, conditions would have been
different to those of the modern reef flat. Salinity in the lagoon would almost certainly have been higher and circulation in the lagoon may have been dominated, to a greater degree, by tidal influences rather than open ocean circulations. These circumstances would give rise to the more regular oscillation of δ\(^{18}\)O about a higher mean value that is apparent in the fossil microatolls. However, there is considerable difference between the mid and the late Holocene examples, implying that the Holocene lagoon was subject to ENSO variability. Further more, the succession of precipitation of carbonate minerals due to heavy flooding during El Niño events and evaporitic minerals due to evaporation during La Niña is evidence that Christmas Island lagoons are indeed, modulated by ENSO (Trichet et al., 2001).

6.4. Mean SST variability through the Holocene.

Throughout the mid-late Holocene the mean δ\(^{18}\)O value (−4.5‰) remained essentially consistent, at a value of ~0.6‰ greater than the modern mean value (~5.1‰). This implies colder average SST and or higher salinities in the Holocene. Direct calculation of SST, using the fractionation of −0.18 to −0.22 ‰ per 1°C (Gagan et al., 1988), would imply that mean SST was ~2.5°C colder than present between 3.6 and 1.7 ka. It seems inappropriate that the difference in δ\(^{18}\)O mean values from the Christmas Island corals should be taken as a measure of an absolute temperature difference of ~2.5°C on two grounds. First, the fossil samples originate from a lagoon where salinity was almost certainly higher than that of the modern day reef flat with which comparison is made. Second, this value appears too large if compared to the findings of the CLIMAP project (1981) that suggest a North Australian tropical SST of ~1.5°C cooler than present at the Last Glacial maximum (LGM) of 18-20 ka. It is therefore inferred that the δ\(^{18}\)O difference between modern and mid-late Holocene is a combination of higher salinities in the lagoon and a colder average SST that was similar in magnitude to that of ~1.5°C at 1.8 ka described by Abram et al. (2001).

The results of CLIMAP are uncertain as δ\(^{18}\)O values from corals from the Huon Peninsula show a depression of up to 6°C in the last glacial (Aharon, 1983). Mountain areas in New Guinea (Western Pacific) have also shown mean annual temperature of at least 6°C cooler at the LGM (Hope, 1989). Coral Sr/Ca ratios from Vanuatu in the
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South western Pacific imply mean SSTs of ~6°C colder than present between 10.2 and 10.3 ka (Beck et al., 1992, Beck et al., 1997). Further Sr/Ca ratios from Papua New Guinea corals from between 8.9 and 7.4 ka indicate average SSTs of 2-3°C cooler than present (McCulloch et al., 1996). In contrast to colder SSTs, paired Sr/Ca and δ¹⁸O ratios from sites in the western Pacific region imply average SSTs of 1°C warmer than present approximately 3000 years later at 5.4 ka (Gagan et al., 1998). The difference between western and central Pacific mean SSTs is currently ~1°C warmer in the west and a stronger mid-Holocene Walker Circulation as hypothesised by Shulmester (1999) would serve to increase this difference and may be responsible for warmer western SSTs at 5.4 ka. Recent research describes an oscillation of mean SST from Porites corals from Japan that varies between 1.5, 3.6 and 1.5°C colder than present at 4.2, 3.4 and 1.8 ka respectively (Abram et al., 2001). It is clear that there is much debate on the mean temperature of the surface ocean since the LGM and it appears that there was a transition between warm and cold conditions some time between the early and mid-Holocene and since then mean SSTs have been colder but rising.

6.5. The onset of modern ENSO variability.

Evidence for the onset of modern El Niño characteristics comes from geoarchaeological evidence from Peru that is consistent with a major mid-Holocene climate change at ~5 ka (Sandweiss et al., 1996). Their study used faunal molluscan assemblages of tropical taxa to imply a warming of the eastern Pacific further north than 10°S only after 5 ka, which is inferred to indicate the timing for the onset of modern El Niño-like events. Consistent with this interpretation are the findings of Rodbell et al., (1999). They describe variations in inorganic laminae deposited by debris flows that correlate with El Niño-induced storm events over the last 200 years. A 15,000 year long record of the lake sediment was used to examine variation in ENSO over the Holocene. The results imply that prior to 7 ka, periodicity in ENSO reoccurrence was greater or equal to 15 years. Thereafter, periodicity progressively increased reaching 2-8.5 years (the modern day ENSO periodicity) by 5 ka (Rodbell et al., 1999). Periods calculated for the occurrence of ENSO from the Christmas Island δ¹⁸O records support this evidence, with periods ranging from an average of 4.2 years in the mid-Holocene, decreasing to 3.8 years in the late Holocene. The increase in ENSO frequency from the
mid-Holocene and the increase in intensity implied by the greater δ^{18}O range in the late Holocene may be attributed to the precession of the Earth's orbit and associated timing of solar radiative heat maxima that is currently the subject of experimental modelling predictions (Bush, 1999, Clement et al., 2001).

6.6. The orbital precession and Holocene ENSO variability.

Orbitally forced modelling experiments suggest that the early to mid-Holocene Walker Circulation might have been stronger, driving an equatorial Pacific climate that was more similar to the La Niña phase of ENSO (Bush, 1999). Clement et al., (2000) also used an orbitally forced coupled ocean-atmosphere model to simulate ENSO variability over the Holocene and predicted that intensity and frequency of events was small in the mid-Holocene and increased, reaching a peak at ~3-1 ka. Supporting these modelling experiments is evidence from Papua New Guinea where coral δ^{18}O has been used to infer ENSO variability over the last 130,000 years (Tudhope et al., 2001). Their data imply that ENSO variability was weak at ~6.5 ka which agrees with evidence for an enhanced mid-Holocene Walker Circulation from North Australian pollen records (Shulmeister and Lees, 1995, Shulmeister, 1999). A stronger Walker Circulation in the mid-Holocene would give rise to colder/drier equatorial conditions, as convection would be inhibited by stronger easterly winds that drive a more prominent cold tongue. This is a possible cause for the more annually-driven ENSO signal apparent in the fossil δ^{18}O records from Christmas Island.

The orbital precession (earlier occurrence of the equinoxes each year) changes the timing of perihelion (the point at which the Earth is closest to the Sun in its solar orbit) and therefore solar radiative heat maxima received by the surface ocean. The heating of the surface ocean due to seasonal cycle of solar radiation is an important factor in the development of ENSO. The thermodynamic response to solar radiative heating of the ocean surface produces a uniform seasonal SST anomaly along the equator. As the SST gradient in the equatorial Pacific increases from east to west, the atmospheric heating response to the SST anomalies is greater in the west where the warm pool resides. Subsequently, the western atmospheric heating and the absence of it in the east drives an easterly wind anomaly due to the rising and sinking air over the
western and eastern Pacific respectively. Modern ENSO events tend to peak at the end of the year when SSTs in the eastern equatorial Pacific are seasonally warming and the convective system is enhanced by the intensification of the Hadley Circulation in the northern equatorial Pacific. Modern perihelion occurs at the beginning of the year, thus enhancing ocean surface heating at a time when El Niño events are at a peak. The timing of the wind anomaly subsequently assists the decline of the modern El Niño as the warm pool migrates back towards the west. In the mid-Holocene perihelion occurred in the middle of the year and the surface ocean would have been heated during the development of ENSO events rather than after their peaks.

A sequence of numerical simulations presented by Bush (1999) assessed the role of early to mid-Holocene orbital forcing on the ocean-atmosphere system. In two of these experiments the orbital parameters for 6 ka are applied to the Geophysical Fluid Dynamics Laboratory’s (GFDL) atmospheric General Circulation Model (GCM) and in the third, the atmospheric model is coupled dynamically and thermodynamically to the GFDL primitive equation global ocean model. The coupled ocean-atmosphere model allows the ocean to adjust to and possibly feed back upon changes in the atmospheric forcing induced by Holocene orbital parameters. This coupled model shows stronger monsoon winds than at present and increased upwelling along the equator, giving rise to stronger easterlies and the subsequent cooling by 0.5-1°C.

The coupled ocean-atmosphere model of Zebiak and Cane (1987) has also been used to simulate patterns of ENSO associated with the precession of the Earth’s orbit and the subsequent change in time of maximum solar forcing (Clement et al., 2000). In their mid-Holocene model when perihelion occurred in the middle of the year rather than at the beginning, the development of warm ENSO conditions was inhibited due to the timing of the easterly wind anomaly caused by the solar forcing maxima earlier in the year. In both the modern and the mid-Holocene models the development of the warm event began in the boreal spring. By summer the mid-Holocene warm event reached a greater degree of development than its modern counterpart due to additional warming from the solar forcing maximum. However, by autumn the solar forcing had also introduced the easterly wind anomaly that cools the eastern equatorial Pacific. At a time when El Niño develops quickly in the modern model the easterly wind anomaly impedes the development of the mid-Holocene El Niño resulting in the occurrence of
fewer large warm events at this time. By the late Holocene perihelion had moved into the autumn and the associated warming coincided with the time when modern El Niño events develop quickly. This causes a peak in the coupled model mean event amplitude at around 3-1 ka (Figure 29 c) when the largest El Niño event of the microatoll δ¹⁸O signatures occurs (Figure 29 b). However, the modelled experiment showed an increase in mean event amplitude of only 0.2°C from the mid-late Holocene whereas the difference in δ¹⁸O range (mid-Holocene average = 0.85 %‰, late Holocene = 2.18 %‰) indicates a much greater increase in event amplitude, implying that processes other than orbital precession may be involved.

Ocean-atmosphere models often remain constrained in regard to boundary conditions especially those that focus on the equatorial Pacific and interactions from processes that are not entirely associated with the region may be neglected. Poor resolution in oceanic data from regions such as southern South America and poor representation of geographical features such as the Andes may neglect the effects of the processes such as subtropical gyres and south easterly trade winds that play an important role in driving equatorial currents (Liu et al., 2000). Models may make unrealistic predictions of processes that subsequently effect other processes. For example, an overestimated thermocline gradient may increase the amount of upwelling in the eastern Pacific that will subsequently increase the strength and sensitivity of ENSO (Meehl et al., 2000) and coupled ocean-atmosphere models often produce a double ITCZ due to unrealistic symmetry about the equator (Mechoso et al., 1995). The ITCZ is an important process in modern ENSO variability as it plays a significant role in the magnitude of rainfall extremes during El Niño events. In such an event, the Walker Circulation decreases and the meridional Hadley Circulation increases, enhancing convection along the equator and inhibiting the easterly trade winds. The ITCZ subsequently migrates south, joining the SPCZ that migrates north, at the equator.

Evidence is emerging that there may have been major variation in the mean position of the ITCZ through the Holocene. Paleoclimatic data from South America provides evidence for an increase in the interaction between the Intertropical Convergence Zone (ITCZ) and the Southern Oscillation that drives the Walker Circulation (Haug et al., 2001). The Caraico Basin is located at ~10°N on the northern
shelf of Venezuela and is at the northern edge of the annual latitudinal range of the ITCZ. During the boreal summer season, rain falls in regions that drain directly into the basin and in winter when the ITCZ is further south local river runoff is diminished. Titanium concentrations in sediment from ODP site 1002 (10°42.73’N, 65°10.18’W) in the Cariaco Basin (Figure 29 a) indicate century-scale variations in river runoff during the period 3.8-2.6 ka and contrast with lower runoff and titanium input in the late Holocene as shown in Haug et al., (2001). These authors infer that the mid-Holocene variations are associated with a more northerly position of the ITCZ, which is consistent with a stronger equatorial Walker Circulation and a smaller degree of variation in ENSO as can be seen in the mid-Holocene δ18O. A decrease in river runoff in the late Holocene is inferred to imply a southward migration of the ITCZ during the mid-late Holocene period which would cause the apparent decrease in titanium variation.

Figure 29. Comparison of (a) titanium concentration in deep-sea sediment at Site 1002 as a proxy record of river runoff to the Cariaco Basin, South America, and (b) range (vertical bars) of δ18O for fossil and modern microatolls from Christmas Island. Horizontal line indicates 2 sigma range of calibrated radiocarbon ages. (c) Amplitude of modelled El Niño SST anomalies, based on precessional forcing (Clement et al., 2000), showing maximum at ~2 ka.
Figure 29 b shows the temporal positions of the microatoll records and the range in their $\delta^{18}$O that is equivalent to variation of ENSO extremes. Mid-Holocene variation is smaller than the late Holocene coinciding with a trend of increasing ENSO mean event amplitude (Figure 29 c) as perihelion moves into autumn (Clement et al. 2000). At this time Walker Circulation was strong in the equatorial Pacific, which might be responsible for a northerly ITCZ causing large variations in rainfall (Figure 34 a) in the Cariaco Basin (Haug et al., 2001). As the inferred ITCZ migrated south through the Holocene, its interaction with the Southern Oscillation increased at a time when perihelion was assisting the development of El Niño. This is consistent with the mean event amplitude in the coupled model that peaked at ~2 ka and the timing of the largest El Niño apparent in the $\delta^{18}$O signatures. By modern time, perihelion had moved into winter allowing El Niño events and the increased interaction between the convective zone and the thermal variability of the equatorial Pacific to develop.

These results demonstrate that variation in ENSO has increased from the mid-Holocene to present but the increase was not gradual with at least one large El Niño event in the late-Holocene that exceeds those recorded in the modern signatures. Conditions in the mid-Holocene were colder and driven by an ENSO that was more prone to La Niña than it is at present. A stronger Southern Oscillation, defined by a stronger Walker Circulation hindered the development of the convective system and the ITCZ prevailed further north. At around 2 ka there was a transition from reduced mid-Holocene El Niño amplitude to enhanced late Holocene amplitude, similar to that predicted in the coupled ocean-atmosphere model of Clement et al., (2000). This may have been caused by an increased interaction between the Southern Oscillation and the convective zone as the ITCZ moved further south due to increased solar radiative heating at a time when events were developing.

Timmermann et al., (1999) present results from a global climate model that predict more frequent El Niño and stronger La Niña in a greenhouse warmed future. Modern ENSO events have been shown to have increased in intensity since the mid-Holocene but were at a maximum in the late Holocene. This implies that even if the modern ENSO has been amplified by anthropogenic warming they have not been of an amplitude that exceed Holocene natural events. A variation between cold glacial and warm interglacial temperatures is also apparent in the literature that may have its role in determining patterns of ENSO variability. Christmas Island is located in an ideal position to study the ocean-atmosphere interaction of the central Pacific and provides a base from where to extend the understanding of ENSO variability into pre-instrumental history.
6.7. Conclusions.

Originating from the oceanic areas that cover the majority of the equatorial circumference of the globe, corals are extremely vulnerable to ocean-atmosphere interactions that govern global climate. They are observed to record the variation of processes that define the interaction between the ocean and atmosphere in locations that are often isolated from other suitable proxy materials. They can also be used as successful proxies for the study of pre-instrumental history of oceanic and atmospheric processes.

It has been demonstrated that modern Christmas Island microatolls contain in their skeletal $\delta^{18}O$ composition, a signature that reflects the modern variation of ENSO that is recorded in SST, rainfall data and the SOI, and are therefore suitable proxies for the reconstruction of central Pacific, pre-instrumental ENSO history. Until now, a major factor limiting our understanding of ENSO was an absence of proxy material from the core region of the central equatorial Pacific that enables the reliable reconstruction of pre-instrumental variability. Christmas Island provides an ideal location for the study of pre-instrumental ENSO history for several reasons. First, it is located in a region that experiences a large majority of the SST and rainfall variation associated with ENSO. Two, the presence of modern microatolls and indeed other species, allows the consistency of different corals to reflect instrumental recordings to be tested. Three, SST and rainfall co-vary in this location and their combined effect on $\delta^{18}O$ produces a signal of ENSO extremes. Four, its interior is scattered with an abundance of fossil microatolls that are generally more accessible than other fossil corals. Microatolls grow at or very close to sea level and are relatively easy to sample, as their lateral growth pattern does not necessitate coring.

Five fossil microatolls have been dated between ~5-3 ka and 2-1.5 ka and potentially contain $\delta^{18}O$ records that will enable the reconstruction of ENSO variability through much of the mid and the late Holocene.
The $\delta^{18}O$ proxy has been used to reconstruct four windows into ENSO variability in the mid and late Holocene and the conclusions are as follows:

ENSO variability in the central equatorial Pacific, defined by the $\delta^{18}O$ signatures from Christmas Island fossil and modern microatolls, was smaller in the mid-Holocene than the late Holocene and modern times. Due to a stronger Walker Circulation that drove ENSO processes at a time when perihelion occurred in the middle of the year, the development of El Niño would have been inhibited.

A transition, defined in the range of variability of fossil $\delta^{18}O$ signatures, between small and large ENSO variation occurred between the mid and late Holocene, which may be attributed to an increase in the interaction between the Southern Oscillation and the convective system of the ITCZ. As the strength of the Walker Circulation decreased when perihelion occurred in the autumn, the timing of the associated radiative solar heating maximum could possibly have assisted the development of El Niño to greater extremes.

Although anthropogenic influence may alter the natural cycle of ENSO, the $\delta^{18}O$ signature from the late Holocene implies that it has not yet altered it to an extent that exceeds Holocene natural variability.

ENSO variability appears to have increased through the Holocene and is likely to be further altered by global warming. However, longer term, millennial scale variability such as orbital precession should not be ignored when assessing the future impacts of ocean-atmosphere interactions on global climate.

It is vital in the understanding of current ENSO variability, to unravel the events of the past so that the knowledge gained can be applied to the future. Equatorial Pacific islands, such as Christmas Island, offer a rare insight into the pre-instrumental variability of ENSO and provide the locations from which to study the history of ocean atmosphere interactions that dominate the world’s climate. The corals that these islands sustain are potentially the most useful proxies available for the extension of the current monitoring systems of ENSO variability and provide the opportunity to test the
sophisticated modelling techniques used in the future prediction of global climate variability.

**6.8. Suggestions for further research.**

The difference between the mean $\delta^{18}$O values of the fossil and modern microatolls presents the question, how different were the conditions of the previously more extensive lagoon that give rise to such a difference in mean $\delta^{18}$O? The present day lagoon of Christmas Island is a small remnant of what previously existed and modern microatolls that live in this somewhat smaller lagoon may hold the answer to this question. The collection, analysis and subsequent comparison with the reef flat samples, of at least one of these microatolls will give an idea of how the $\delta^{18}$O signature of ENSO differs, if at all from that of the reef flat.

The $\delta^{18}$O precipitated in corals reflects both temperature and salinity and in a location such as a lagoon, that is subject to both processes, the interpretation of mean SST values becomes inaccurate. The Sr/Ca analysis of the modern microatolls would solve this problem, as this ratio is a reflection of temperature alone. The effect of rainfall on $\delta^{18}$O would then be possible to determine and a calibration of Sr/Ca to SST could be defined. Subsequently, the Sr/Ca analysis of the fossil microatolls would enable a mean SST for the central equatorial Pacific to be calculated.

A number of fossil microatolls have not yet been analysed that may further the reconstruction of ENSO variability. It would be sensible to suggest, first the analysis of more late Holocene microatolls, to test the interpretations made in this project and second, the analysis of the earlier mid-Holocene microatolls to extend the record of ENSO variability further. However, there remains an absence of fossil microatolls for the period of ~3-2 ka which would render the record incomplete. Resampling and dating may provide information for this period and help identify ENSO characteristics.

The incompleteness of the Christmas Island record leads to the final suggestion. Christmas Island is one of the Line Islands in the Republic of Kiribati, which consists of the Gilbert Islands (between 170°E and 180°), the Phoenix Islands (between 180° and
170°W) and the Line Islands (between 175°W and 150°W). Collection and analysis of corals from islands such as Tarawa and Tamana (northern and southern Gilbert Islands), Canton (Phoenix Islands) and Malden (southern Line Islands) or other such islands, could possibly cover an area of the western central Pacific between ~5°N and 5°S. Each site, would of course present a different history of variability depending on the processes that dominate each respective region. For example, Tarawa is located in the region of the western Pacific warm pool and would reflect extremes of rainfall rather than SST and rainfall, as Christmas Island does. However, comparison of the individual histories and synthesis of their results with respect to the processes that define ENSO, could potentially provide information on its variability. By widening the observational area, identifying large scale ocean-atmosphere interaction such as the southward migration of the ITCZ over the Holocene could possibly be achieved.
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