PERGAMON

Quaternary Science Reviews 19 (2000) 45–64

QSR

New views of tropical paleoclimates from corals

M.K. Gagan[a,*], L.K. Ayliffe[a,1], J.W. Beck[b], J.E. Cole[c], E.R.M. Druffel[d], R.B. Dunbar[e], D.P. Schrag[f]

[a]*Research School of Earth Sciences, Australian National University, Canberra, ACT 0200, Australia*
[b]*NSF AMS Facility, Department of Physics, University of Arizona, Tucson, AZ 85721, USA*
[c]*Department of Geological Sciences/INSTAAR, Box 450, University of Colorado, Boulder, CO 80309, USA*
[d]*Department of Earth System Science, University of California, Irvine, CA 92697-3100, USA*
[e]*Department of Geological and Environmental Sciences, Stanford University, Stanford, CA 94305, USA*
[f]*Department of Earth and Planetary Sciences, Harvard University, Cambridge, MA 02138, USA*

Abstract

Corals offer a rich archive of past climate variability in tropical ocean regions where instrumental data are limited and where our knowledge of multi-decadal climate sensitivity is incomplete. In the eastern equatorial Pacific, coral isotopic records track variations in ENSO-related changes in sea-surface temperature; further west, corals record variability in sea-surface temperature and rainfall that accompanies zonal displacement of the Indonesian Low during ENSO events. These multi-century records reveal previously unrecognised ENSO variability on time scales of decades to centuries. Outside the ENSO-sensitive equatorial Pacific, long-term trends towards recent warmer/wetter conditions suggest the tropics respond to global forcings. New coral paleothermometers indicate that surface-ocean temperatures in the tropical southwestern Pacific were depressed by 4–6°C during the Younger Dryas climatic event and rose episodically during the next 4000 yr. High temporal-resolution measurements of Sr/Ca and $\delta^{18}O$ in corals provide information about the surface-ocean hydrologic balance and can resolve the seasonal balance between precipitation and evaporation. Radiocarbon measurements in corals, coupled with ocean circulation models, may be used to reconstruct near-surface ocean circulation, past mixing rates, and the distribution of fossil fuel CO_2 in the upper ocean. Most recently, seasonal to interannual variations in the radiocarbon of corals from the equatorial Pacific have been linked to the redistribution of surface waters associated with the ENSO.

1. Introduction

Massive corals growing in the reef ecosystems of the tropics provide some of the richest paleoclimate archives in the world. Corals are particularly useful paleoclimate recorders because they are widely distributed, can be accurately dated, and contain a broad array of geochemical tracers within their skeletons. During the last decade, there has been a concerted effort to identify new climatic tracers in corals and develop more sophisticated techniques for data extraction and measurement. As a result, a multi-proxy approach to coral-based paleoclimatology is emerging that is yielding new insights into tropical paleoclimates.

Coral records with seasonal to annual resolution can contribute to resolving key uncertainties in our knowledge of tropical climate on two main fronts. First, continuous coral records spanning several centuries are revealing the natural limits and behaviour of the tropical ocean-atmosphere, at time scales that are relevant to society. The aragonite skeletons of reef-building corals carry isotopic and chemical indicators that track water temperature, salinity, and isotopic composition as well as site-specific features including turbidity, runoff, and upwelling intensity. By documenting the natural behaviour of these systems, we can assess their sensitivity to various natural phenomena, such as solar and volcanic changes, and anthropogenic inputs such as increasing greenhouse gas concentrations and land-use changes. A high priority in coral research is to produce quantitative indicators of specific aspects of climate that can be integrated with other high-resolution paleoclimate data derived from tree rings, ice cores, and varved sediments. The climate

* Corresponding author.
[1] Present address: Laboratoire des Sciences du Climat et de l'Environnement (LSCE), CNRS-CEA, F-91198 Gif-sur-Yvette, Cedex, France.

0277-3791/99/$ - see front matter
PII: S0277-3791(99)00054-2

indices generated via this network of proxy climate records can then be compared with climate model output to help assess model performance.

The second area in which corals can contribute important climate understanding is through the reconstruction of tropical climate variability during times of altered background climate, such as the Last Glacial Maximum (LGM) and the mid-Holocene. At least two attributes of fossil corals make them particularly well suited for defining the natural bounds and sensitivity of tropical climate. First, corals have the ability to track climate changes within the annual cycle, allowing sea-surface conditions to be reconstructed with reference to specific seasons. Seasonally resolved paleoclimate data are essential for reconstructing dynamic climate systems such as the El Niño-Southern Oscillation (ENSO), the monsoon, and wind-driven oceanic upwelling. Secondly, very precise dates for fossil corals may be obtained by measuring $^{230}Th/^{234}U$ ratios with thermal ionisation mass spectrometry (Edwards et al., 1987). Errors associated with ^{230}Th age determinations are generally less than 1% for samples that are 100 to 200,000 yr old. Density band couplets, or annually varying geochemical tracers, can then be used to achieve seasonal time control within individual "floating" chronologies. These data sets will be particularly useful for understanding the sensitivity of climatic processes to global climate change.

The purpose of this paper is to summarise recent advances in three promising coral research streams including: (1) the development of pan-tropical reconstructions of multi-decadal climate variability and the ENSO; (2) progress in high-resolution coral paleothermometry; and (3) the use of radiocarbon in corals as a tracer of ocean circulation. A synthesis is presented of the available multi-century coral oxygen isotope and radiocarbon records as they relate to changes in climate and ocean circulation in the tropical Pacific and Indian Oceans. The second section of the paper summarises high-resolution coral paleothermometry and oxygen isotope data compiled for the southwestern Pacific region, which reveal new information about sea-surface temperature (SST) and the surface-ocean hydrologic balance since the LGM. We then synthesise new accelerator mass spectrometric (AMS) measurements of radiocarbon in corals from the equatorial Pacific, at sub-seasonal resolution, which resolve interannual and decadal variations in upper ocean circulation that are related to changes in ENSO.

We deliberately focus on the three main streams of coral research that are concentrating primarily on the tropical Pacific region to demonstrate the importance of developing a multi-proxy approach to coral paleoclimatology. Indeed, the suite of geochemical tracers in corals presently exceeds that developed in any other biogenic mineral phase, including marine organisms such as foraminifera and radiolaria, which have contributed so much to the field of paleoceanography. Therefore we refer interested readers to thorough reviews of the status of geochemical tracers in corals (Shen, 1993; Dunbar and Cole, 1993; Druffel, 1997a). These publications also summarise aspects of coral biology, density band formation, and factors affecting the uptake of isotopic and chemical signals in coral skeletal aragonite, all of which are important in understanding corals as environmental proxies.

2. Tropical climate over the past several centuries

Worldwide, 15 to 20 coral-based climate records extending back to at least the mid 1800s are currently available or nearing completion (Fig. 1). The published records have provided new information on environmental changes in the tropical surface-ocean over the past several centuries (Pätzold, 1986; Druffel and Griffin, 1993; Quinn et al., 1993; Heiss, 1994; Linsley et al., 1994; Dunbar et al., 1994, 1996; Charles et al., 1997; Crowley et al., 1997; Lough and Barnes, 1997; Boiseau et al., 1998; Isdale et al., 1998; Quinn et al., 1998). An additional 20 to 30 records extend back 50 to 100 yr from the present. Since reliable instrumental records of past SST variability in the tropics rarely extend prior to about 1950, these coral records are important for studies of decadal climate variability and linkages among climate systems at interannual time scales.

Most massive reef corals live at water depths of < 40 m and grow continuously at rates of 6–20 mm yr^{-1}, producing annual density band couplets that provide time markers for the development of long chronologies (Knutson et al., 1972). For many coral records, density bands provide an inexpensive, fast, and precise chronology of skeletal growth. However, some corals produce complicated or ambiguous density patterns resulting in the assignment of errors of about 1% to age estimates. Where banding is absent or poorly defined, the seasonal cycling of detailed oxygen or carbon isotope records can be used to fill gaps, or even to establish relatively long chronologies (Fairbanks and Dodge, 1979; Cole et al., 1993; Gagan et al., 1996; Evans et al., 1998). Application of cross-dating procedures should allow most coral records to achieve true annual chronologic precision.

Many coral studies have utilized measurements of oxygen isotope ratios ($\delta^{18}O$) in coral aragonite because they are readily available and relatively straightforward to interpret. In oceanic settings where the oxygen isotopic composition of seawater is constant, coral skeletal $\delta^{18}O$ records SST variability, usually according to the standard paleotemperature relationship for carbonates (Epstein et al., 1953). The isotopic composition is offset by a biological non-equilibrium component that appears to be stable through time, as long as a consistent, maximum growth axis is sampled within a coral colony (Weber and Woodhead, 1972; Dunbar and Wellington, 1981;

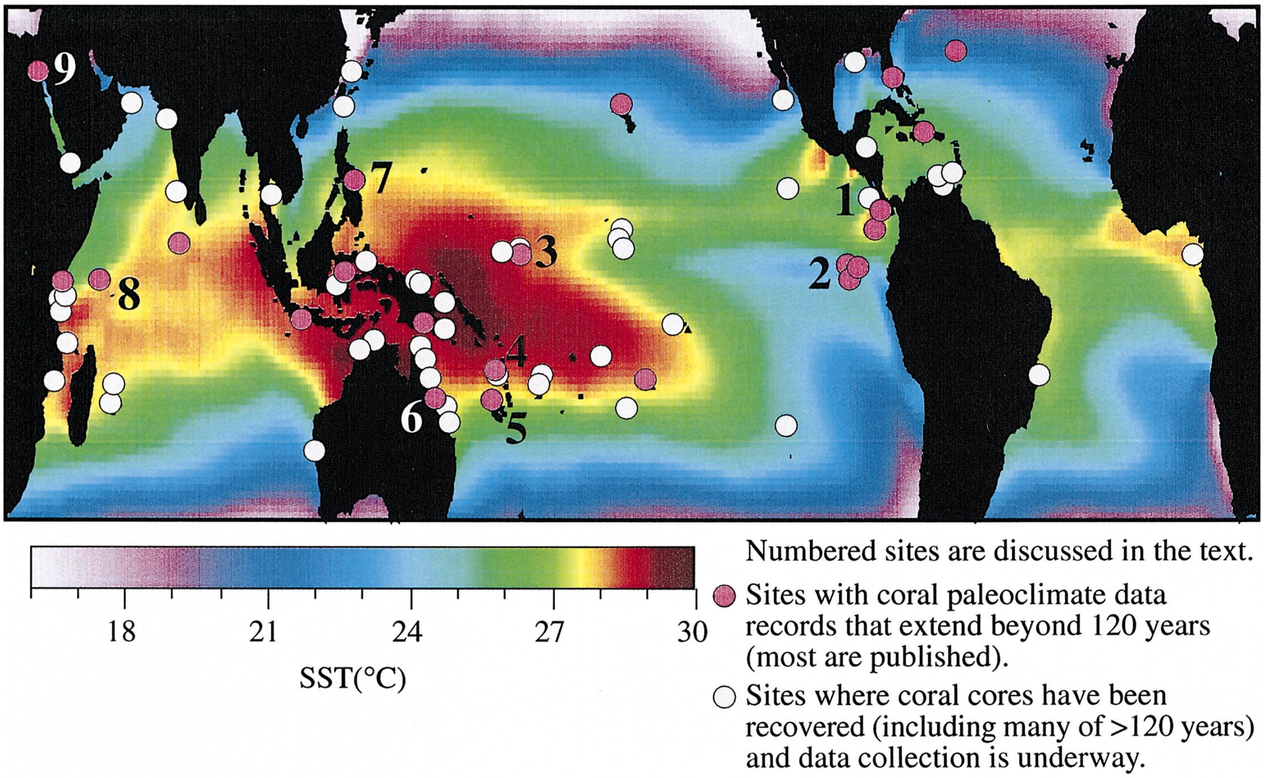

Fig. 1. Distribution of annual mean tropical SSTs and approximate locations of current coral paleoclimate research. SST data are from the National Meteorological Center and are available at: http://www.ingrid.ldgo.columbia.edu. The coral sites represent the work of many investigators and may be incomplete. Locations of coral $\delta^{18}O$ records described in Fig. 2 are numbered (1–9).

McConnaughey, 1989; Winter et al., 1991; Shen et al., 1992; Gagan et al., 1994; Leder et al., 1996; Swart et al., 1996; Wellington et al., 1996; Cohen and Hart, 1997). When seawater $\delta^{18}O$ varies in response to changes in the balance between precipitation, evaporation, and water advection, the coral $\delta^{18}O$ changes accordingly (Swart and Coleman, 1980; Dunbar and Wellington, 1981; Cole et al., 1993; Gagan et al., 1994; Linsley et al., 1994). In oceanic settings where seawater $\delta^{18}O$ correlates with rainfall, long records of coral $\delta^{18}O$ have been used to reconstruct precipitation (Cole et al., 1993; Linsley et al., 1994).

2.1. Punctuated long-term trends

Nine of the longest published coral $\delta^{18}O$ records from pan-tropical sites in the Pacific and Indian Oceans are shown in Fig. 2. Several of these long coral time-series have only recently become available and the following short discussion is not meant to forecast the results of a comprehensive synthesis. However, the results thus far are intriguing enough to warrant further data acquisition and initial attempts at quantitative analysis. For example, seven of the records show a warming/freshening trend toward the 20th century. With the exception of the Galapagos record, those spanning more than 200 yr indicate that the warming/freshening trend began during, or before, the 19th century (Cole, 1996). The authenticity of the relatively stable Galapagos $\delta^{18}O$ record is indicated by recent analysis of eastern Pacific SSTs, which shows a slight cooling trend during the 20th century (Cane et al., 1997). If the shift in $\delta^{18}O$ is due entirely to warming of the surface ocean, it is equivalent to 0.3–2°C since the early 1800s. However, the magnitude of surface-ocean warming indicated by the coral $\delta^{18}O$ is large compared with the ~ 0.5°C warming indicated for the tropical Pacific since the 1850s, based on analysis of the available instrumental data (Cane et al., 1997; Kaplan et al., 1998). Therefore, it is important to establish the temperature component of the $\delta^{18}O$ shift using other coral paleothermometers, to reveal the magnitude of any 20th century freshening of the ocean surface.

Superimposed on this long-term warming/freshening trend are abrupt, spatially coherent shifts in the coral $\delta^{18}O$ values. The six records spanning more than 200 yr show a distinctive cool period ($\sim$ 1800–1840 AD) that may be related to enhanced volcanism that occurred during this time (Crowley et al., 1997; Quinn et al., 1998). Following this cool interval, four of the six records (Panama, Vanuatu, Great Barrier Reef, Red Sea) display an

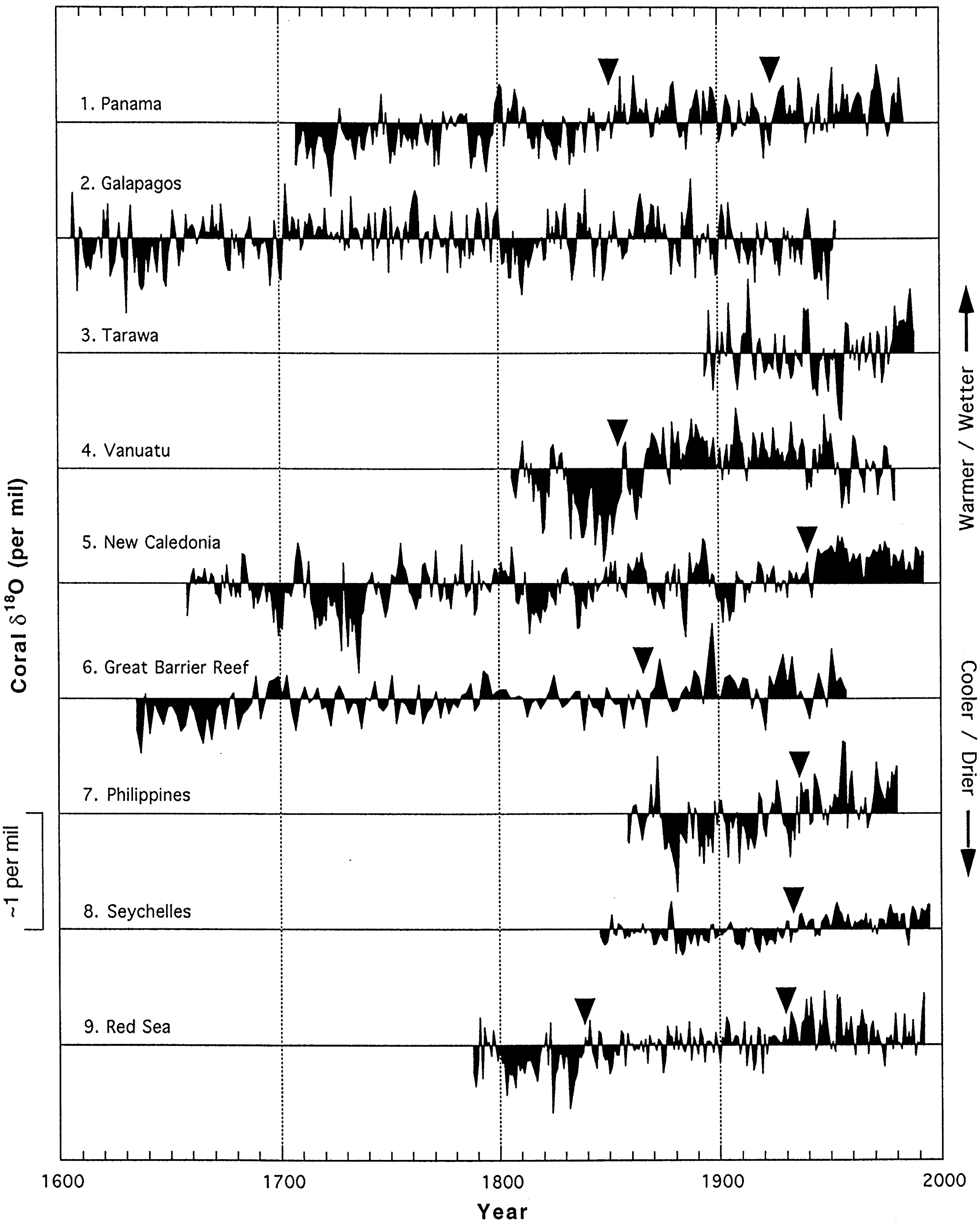

Fig. 2. Comparison of annual mean coral $\delta^{18}O$ records in the Pacific and Indian Ocean region extending back more than 100 years (locations of coral cores are shown in Fig. 1). Horizontal lines show the mean $\delta^{18}O$ value for each site. Black triangles mark approximate times of abrupt shifts in $\delta^{18}O$ heralding warmer/wetter conditions. Data are from the World Data Center-A for Paleoclimatology, NOAA/NGDC Paleoclimatology Program, Boulder, Colorado, USA (http://www.ngdc.noaa.gov/paleo/coral) and the original references. Details of cores (locality, species name, time-span of record, original reference) are as follows: 1. Gulf of Chiriqui, Panama (8°S, 82°W; *Porites lobata*; 1708–1984; Linsley et al., 1994); 2. Urvina Bay, Galapagos, Ecuador (0°S, 91°W; *Pavona clavus* and *P. gigantea*; 1607–1981; Dunbar et al., 1994); 3. Tarawa Atoll, Republic of Kiribati (1°N, 172°E; *Porites spp.*; 1893–1989; Cole et al., 1993); 4. Espiritu Santo, Vanuatu (15°S, 167°E; *Platygyra lamellina*; 1806–1979; Quinn et al., 1993); 5. Amedee Lighthouse, New Caledonia (22°S, 166°E; *Porites lutea*; 1657–1992; Quinn et al., 1998); 6. Abraham Reef, Great Barrier Reef, Australia (22°S, 153°E; *Porites australiensis*; 1635–1957; Druffel and Griffin, 1993); 7. Cebu, Philippines (10°N, 124°E; *Porites lobata*; 1859–1980; Pätzold, 1986); 8. Mahe Island, Seychelles (5°S, 55°E; *Porites lutea*; 1846–1995; Charles et al., 1997); 9. Aqaba, Red Sea (29°N, 35°E; *Porites sp.*; 1788–1992; Heiss, 1994; data for vertical core).

abrupt shift to warmer/wetter conditions at around 1840–1860 AD (Fig. 2). Another significant shift to warmer/wetter conditions occurs between 1925 and 1940 in five records from the Pacific and Indian Ocean regions (Panama, New Caledonia, Philippines, Seychelles, Red Sea). The timing of the early 20th century warming matches a similar trend in instrumental SST records for the tropical Pacific (Kaplan et al., 1998).

Intriguing as the long-term and abrupt shifts in coral $\delta^{18}O$ may be, they must be interpreted with caution. Potential biological causes of long-term variability in coral $\delta^{18}O$ have yet to be determined via observation or experiment. However, there are no a priori reasons to expect a strictly biological aging effect that would produce spurious results at multi-decadal to century time scales. A non-biological cause of long-term warming in coral $\delta^{18}O$ records may result from the coral surface growing at shallower water depths (by several meters) as the coral grows upwards. This may expose the coral growth surface to waters of slightly different temperature, salinity, and light intensity. However, the seven records showing similar long-term trends were produced from coral colonies with upper growth surfaces terminated in a range of water depths (1–7 m). Moreover, the abrupt shifts in coral $\delta^{18}O$ (ca 1840–1860 and 1925–1940) are observed at widely separated sites (Fig. 2), suggesting they may be responding to global climate forcing.

The pattern of warming/freshening in the tropical surface ocean indicated by the coral $\delta^{18}O$ records is consistent with recent mid-latitude and high-latitude temperature reconstructions suggesting that the 20th century could be the warmest of the millenium (Overpeck et al., 1997; Mann et al., 1998, 1999). The growing number of coral records showing a significant warming/freshening trend points to the involvement of the tropical oceans in enhancing the hydrological cycle and the warming of higher latitudes, as suggested by Graham (1995) from theoretical constraints.

2.2. Decadal variability and ENSO

The other interesting aspect of these records is the large magnitude of variability they reveal at decade to century time scales, both in terms of SST and rainfall. In the eastern Pacific, a comprehensive network of coral climate reconstructions is developing (Druffel, 1981; Druffel et al., 1990; Linn et al., 1990; Shen et al., 1991, 1992; Dunbar et al., 1994, 1996; Carriquiry et al., 1994, 1998; Linsley et al., 1994, 1999; Wellington and Dunbar, 1995; Shen, 1996; Wellington et al., 1996). The 280-yr $\delta^{18}O$ record from a rainfall-sensitive site near Panama indicates decadal periods in the strength/position of Intertropical Convergence Zone (ITCZ) precipitation (Fig. 2). Decade-century variance in the 370-yr Galapagos $\delta^{18}O$ record appears to correlate inversely with the Panama record, consistent with the southward displacement of the ITCZ during warmer periods in the Galapagos (Dunbar et al., 1996). In the western Pacific, decadal variability of SST and/or rainfall is implicated at Vanuatu by a persistent 14-yr period in the $\delta^{18}O$ record (Quinn et al., 1993). The 3-century record of $\delta^{18}O$ (and radiocarbon) from Abraham Reef, Great Barrier Reef, has been interpreted as reflecting changes in circulation off northeast Australia (Druffel and Griffin, 1993). SST reconstructions from Indian Ocean corals reveal substantial variability on decadal time scales, which is poorly documented by the very limited instrumental SST record. In a coral from the Seychelles, decadal patterns of variation in $\delta^{18}O$ correspond with Indian monsoon rainfall indices, suggesting that long-term regional rainfall variability may originate at least in part from the ocean (Charles et al., 1997).

Some regional isotopic events and trends can be explained in terms of variability in key climatic systems of the Pacific. For example, the driest/coolest interval at Tarawa (early 1950s) corresponds with a very dry/cool period at Vanuatu (southwest Pacific) and a warm/wet period in the Philippines (Fig. 2). These observations are consistent with a westward contraction of the warm pool and consequent local anchoring of the Indonesian Low in the far western Pacific. Cook (1995) discusses another multi-decadal inverse relationship between SSTs derived from the Galapagos coral record (Dunbar et al., 1994), a Great Barrier Reef SST index derived from coral growth band thickness (Lough and Barnes, 1997), and the Abraham Reef $\delta^{18}O$ record (Druffel and Griffin, 1993), all of which span the past four centuries. These results, and others, suggest that ENSO-type variability may be modulated over long time scales.

The tropical Pacific has been a common focus of many coral paleoclimate studies, due to the importance of ENSO in global climate variability. Corals from sites in the equatorial Pacific record past ENSO variability through its impact on SST, surface water salinity, or a combination of the two. Work on corals in the ENSO-sensitive central Pacific has focused on the islands of Kiribati (near 1°N, 172°E; Cole and Fairbanks, 1990; Cole et al., 1993, Shen et al., 1992) and the Line Islands ($\sim$160°W; Druffel, 1987; Evans et al., 1998). In this region, warm and wet anomalies during ENSO warm phases work together to generate lower coral skeletal $\delta^{18}O$ values. An oxygen isotope record from Tarawa Atoll correlates highly with standard indices of ENSO, as locally intense rainfall and warmer SSTs decrease the coral $\delta^{18}O$ values during ENSO warm phases. Evolutionary spectral analysis of this record shows that seasonal and interannual variability have varied together over the past century (Fig. 3). For example, in the mid-20th century, stronger seasonal cycles accompany a

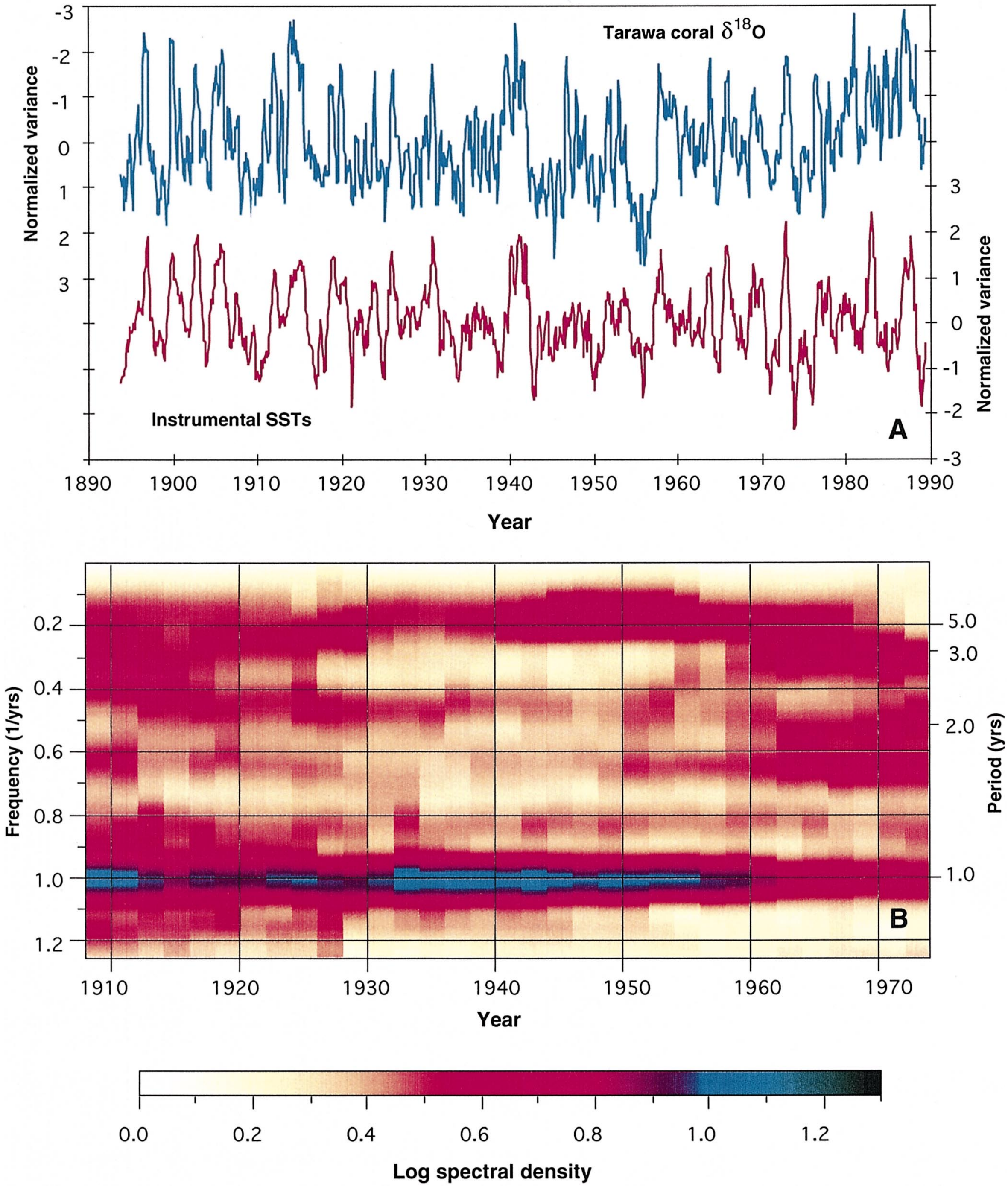

Fig. 3. Panel A: Monthly record of normalised coral $\delta^{18}O$ (Cole et al., 1993) compared with an index of SST in the central equatorial Pacific (Nino 3.4; Kaplan et al., 1998). ENSO warm phases are revealed by upward excursions, and match well between records, although the signal at Tarawa is primarily due to salinity changes associated with rainfall. Panel B: Results of evolutionary spectral analysis using Blackman–Tukey methods and the ARAND software package (courtesy P. Howell, Brown University). Concentrations of variance at ENSO periods change over the course of this century; for example, between 1925 and 1950, the three-year period of ENSO is weaker and the seasonal cycle has intensified. This period corresponds with a time of weaker ENSO variability within the Pacific and weaker teleconnections outside this region.

period of reduced ENSO amplitude seen both here and in other records within and beyond the tropical Pacific. In addition, corals from Papua New Guinea and Indonesia reveal interannual variability in $\delta^{18}O$ that is strongly correlated with the Southern Oscillation (Tudhope et al., 1995; Fairbanks et al., 1997).

3. Exploring The Natural Bounds Of Tropical Paleoclimate

3.1. Paleothermometry

The reconstruction of past SSTs is a classical problem in paleoceanography and is one which is fundamental to understanding the mechanisms of climate change. Towards this end, recent research on several coral paleothermometers has yielded some promising results. Most of these are based on the temperature-dependent incorporation of divalent cations into coral skeletal aragonite. The use of coralline Sr/Ca ratios as a temperature proxy has received the most attention (Smith et al., 1979; Beck et al., 1992) though several other coral temperature proxies also have been proposed, most notably Mg/Ca ratios (Oomori et al., 1983; Mitsuguchi et al., 1996) and U/Ca ratios (Min et al., 1995; Shen and Dunbar, 1995). Other potential coral paleothermometers include B and F (Hart and Cohen, 1996) which occur as trace elements in seawater and coral (< 1 ppm).

One of the most promising of these ocean temperature proxies involves the precise measurement of Sr/Ca ratios in corals using thermal ionisation mass spectrometry (TIMS). The groundwork for Sr/Ca thermometry was laid by Smith et al. (1979) who showed that, within measurement resolution ($\pm 2°C$), the Sr/Ca ratio in coralline aragonite correlated with temperature. An important improvement was made to this technique by Beck et al. (1992), who showed that by using the TIMS technique the signal-to-noise ratio for this proxy could be dramatically increased. Using TIMS, the current measurement precision of 0.03% means that, theoretically, temperature changes of 0.05°C can be distinguished. TIMS measurements are relatively time consuming, however, which limits the applicability of the technique, and makes the U/Ca proxy very attractive because it can be measured by the more rapid, but less precise, inductively coupled plasma mass spectrophotometry (ICP-MS) analytical technique. A new high precision method of Sr/Ca and Mg/Ca determination via inductively coupled plasma atomic emission spectroscopy (ICP-AES) involving frequent comparison of samples to reference solutions (Schrag, 1999), greatly increases sample throughput and may allow these paleothermometers to be used routinely.

There is now abundant evidence that Sr/Ca in coral can be a high-fidelity temperature proxy (deVilliers et al., 1994; McCulloch et al., 1994; Min et al., 1995; Shen et al., 1996; Alibert and McCulloch, 1997; Gagan et al., 1998; Ayliffe et al., submitted). Nevertheless, several important issues remain to be resolved which could impact the implementation of this or the other potential paleothermometers. For example, very high spatial resolution studies of the distribution of Sr/Ca in corals using ion microprobe (Allison, 1996; Hart and Cohen, 1996) and laser ablation ICP-MS techniques (Sinclair et al., 1998) reveal micron-scale variability apparently unrelated to temperature. Measurement reproducibility on coral samples suggests that these inhomogeneities in the distribution of Sr/Ca can produce uncertainties of around $\pm 0.3°C$ in fine-scale measurements ("weekly" increments). In addition, there is some question that the incorporation of Sr (deVilliers et al., 1994) or Mg (Oomori et al., 1983) may respond to growth rate variations, as well as temperature.

3.2. Sr/Ca thermometer calibration

Coral Sr/Ca–SST relations, based on TIMS measurements of Sr/Ca in the coral genus *Porites*, have now been published for nine sites in the Pacific and Indian Oceans and applied in paleoclimatic reconstructions (Beck et al., 1992; deVilliers et al., 1994; Shen et al., 1996; Alibert and McCulloch, 1997; Gagan et al., 1998; Ayliffe et al., submitted). Despite these studies, there is still no single accepted coral Sr/Ca-SST relation. Examination of Fig. 4 shows that there are offsets among the regression lines amounting to 3.5°C. However, with the exception of the relation produced by deVilliers et al. (1994), the slopes of the calibration equations are remarkably similar (mean $= 0.062 \pm 0.014$, 2σ).

The observed dispersion in the Sr/Ca-SST relations may be an experimental artifact related to differences in coral collection and sampling protocols or standardisation of instrumental SSTs among calibration sites, or it may reflect some real variability. Inter-laboratory differences in Sr/Ca spike calibrations also may produce errors. Two key assumptions in applying the Sr/Ca thermometer are that biological controls on skeletal Sr/Ca up-take are not important, or are controlled by temperature alone, and secondly that changes in the Sr/Ca content of seawater are negligible. As pointed out by Smith et al. (1979), there is an offset between coralline aragonite Sr/Ca ratios at a given temperature and aragonite precipitated inorganically (see Kinsman and Holland, 1969, Fig. 4). Based on this and other experimental observations, deVilliers et al. (1994, 1995) suggested temperature may not be the only important control on coralline Sr/Ca ratios. In particular, they suggested that precipitation rate or other biological processes also may be important. Other studies have, however, demonstrated that changes in mean annual calcification rate spanning the spectrum of growth expected for the coral genus *Porites* produce offsets of less than 1°C among the calculated Sr/Ca-SST relations (Shen et al., 1996; Alibert and McCulloch, 1997; Gagan et al., 1998).

Equally important, all of the temperature proxies require the oceanic concentration ratio of the appropriate metals to remain invariant. Measurements of Sr/Ca in seawater for modern oligotrophic reef settings in the Pacific and Atlantic Oceans indicate ocean water Sr/Ca

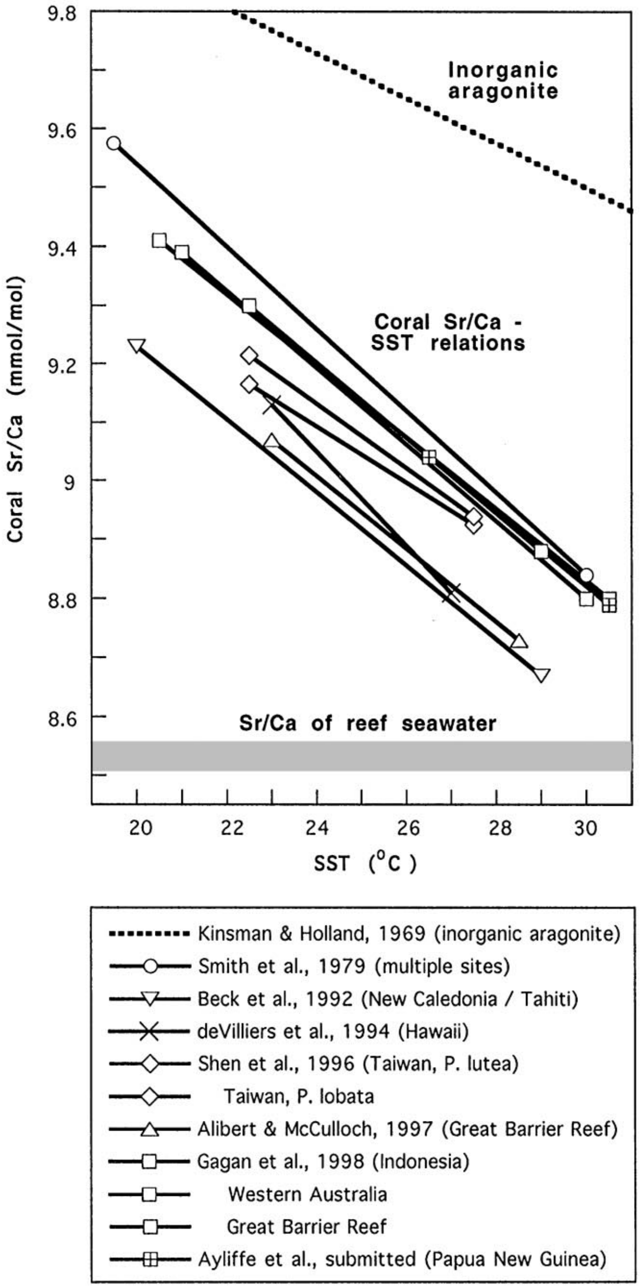

Fig. 4. Summary of coral Sr/Ca-temperature relations derived from seven different studies (1979–1999). The study of Smith et al. (1979) used the ICP-AES technique to measure coral Sr/Ca ratios, whereas the more recent studies (1992–1999) used TIMS. The length of the regression lines indicates the temperature range over which the Sr/Ca–SST relation was tested. The known range of Sr/Ca in oligotrophic reef seawater (deVilliers et al., 1994; Shen et al., 1996) and the inorganic aragonite precipitation line derived by Kinsman and Holland (1969) are shown for comparison.

ratios are not invariant as once supposed, but exhibit a range of 8.51 to 8.55 mmol/mol (deVilliers et al., 1994; Shen et al., 1996). Such differences could account for about 0.6°C of the variation among calibration sites, based on an average sensitivity in the Sr/Ca–SST relations of 0.062 mmol/mol per°C. Furthermore, recent modelling studies by Stoll and Schrag (1998) indicate that dissolution of Sr-enriched aragonite exposed on continental shelves during sea-level lowstands (i.e. the last glacial maximum) will increase the Sr/Ca ratio of glacial seawater. Such variations are sufficient to produce "cool" artifacts of up to 1.5°C in paleotemperatures reconstructed from corals living during glacial maxima. This finding is yet to be reproduced in other studies but, if correct, it may be possible to reconstruct past ocean Sr/Ca variations by measuring the Sr/Ca ratio of porewater trapped in the sediment of deep-sea cores, and correct for any drift in the Sr/Ca thermometer.

3.3. Reconstructing SST and seawater $\delta^{18}O$

In principle, coupled measurements of coral skeletal Sr/Ca and $\delta^{18}O$ should make it possible to reconstruct the $\delta^{18}O$ of seawater, as well as SSTs, by removal of the temperature component of the coral $\delta^{18}O$ signal (McCulloch et al., 1994; Gagan et al., 1998; Ayliffe et al., submitted). The changes in seawater $\delta^{18}O$ defined by the differences between coral Sr/Ca and $\delta^{18}O$ curves (residual $\delta^{18}O$) can be used to make inferences about the tropical hydrological cycle because ^{18}O is fractionally distilled from ^{16}O during the evaporation and precipitation of water vapour between the ocean and atmosphere. Ideally, the technique should be capable of yielding information about SST, precipitation, evaporation, and water mass transport, even in oceanic settings where both temperature and seawater $\delta^{18}O$ change in unexpected ways.

The utility of the coupled Sr/Ca–$\delta^{18}O$ technique is illustrated for four oceanic settings encompassing seasonal extremes in SST, salinity, and turbidity spanning much of the survival range for *Porites* (Figs. 5 and 6). Recent calibration experiments demonstate that the same Sr/Ca–SST relation can be used for all four sites with no significant difference between instrumental and coral-derived SSTs (Gagan et al., 1998; Ayliffe et al., submitted). Regional differences in seawater $\delta^{18}O$ should reflect regional changes in sea-surface salinity (SSS) because tropical rainfall is depleted in ^{18}O (IAEA, 1981), relative to seawater, while evaporation tends to enrich the surface ocean in ^{18}O.

Regional differences in the $\delta^{18}O$ of seawater among sites can be determined using the $\delta^{18}O$ residuals ($\Delta^{18}O$ in Fig. 6) for the austral winter dry season when changes in SSS, and presumably seawater $\delta^{18}O$, are negligible. As a starting point, the $\delta^{18}O$-SST calibration for the Great Barrier Reef site is centred on a dry-season salinity of 35.2 p.s.u. ($\delta^{18}O$ defined 0). By comparison, the dry-season coral $\delta^{18}O$ residuals for Java and Papua New Guinea are $-0.40‰$ and $-0.35‰$, respectively. The negative $\delta^{18}O$ residuals reflect the relatively low salinity water mass encompassing the Warm Pool region, where

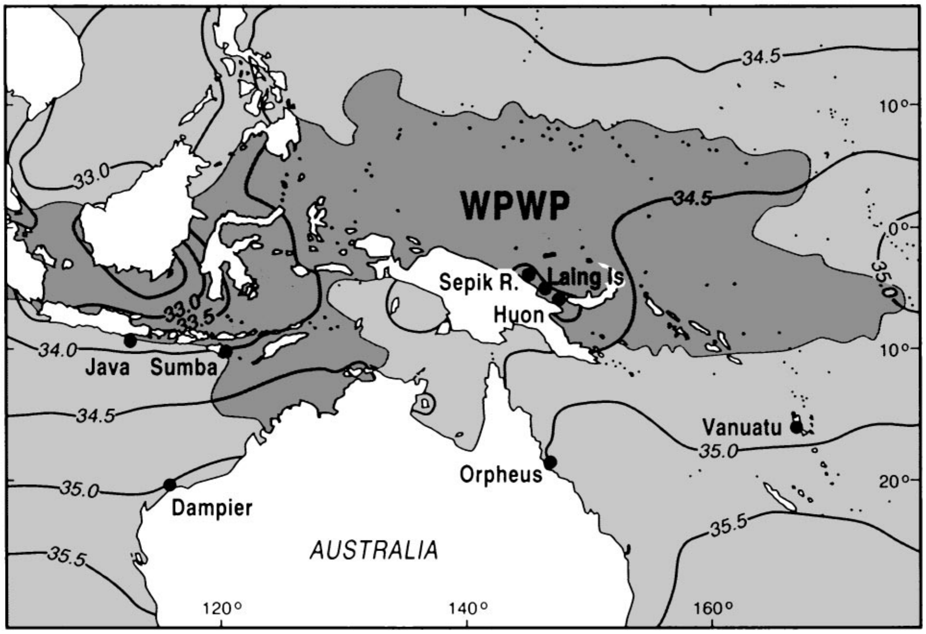

Fig. 5. Locations of coral Sr/Ca and $\delta^{18}O$ records described in Figs. 6–8 superimposed on the distribution of sea-surface salinity (compiled by T. Barrows, ANU, using data of Levitus et al., 1994). Dark stippling marks the average extent of the Western Pacific Warm Pool (*WPWP*, mean annual SST $> 28°C$), as defined by Yan et al., 1992.

the input of ^{18}O depleted precipitation greatly exceeds evaporation. Given that the average difference in salinity between the two Warm Pool sites and Orpheus Island in the Great Barrier Reef is ~ 1.4 p.s.u. (Fig. 5), the slope of the $\delta^{18}O$/salinity relationship is ~ 0.25–0.29‰ per p.s.u. This $\delta^{18}O$/salinity slope is essentially the same as that derived recently for a network of water sampling sites near reefs of the tropical Pacific (0.27‰ per p.s.u.; Fairbanks et al., 1997). The result suggests that the coral residual $\delta^{18}O$ technique is capable of yielding accurate reconstructions of relative differences in seawater $\delta^{18}O$ which, at least for Indo-Pacific reef settings, can be used to make inferences about salinity. The lowest $\delta^{18}O$ residuals for the Dampier coral (-0.15‰; ~ 34.6 p.s.u.) reflect the advection of low salinity Warm Pool water into the eastern Indian Ocean, via the Indonesian Throughflow, where strong seasonal evaporation produces high annual mean salinities near the coast of northwestern Australia (Fig. 6).

Superimposed on the regional differences in the $\delta^{18}O$ of seawater are short-term, seasonal changes reflecting the local balance between precipitation and evaporation. The distinct lack of runoff from 1992-1994 in the Great Barrier Reef testifies to the severity of the ENSO-induced drought that gripped northeast Australia (Fig. 6). In Papua New Guinea, a shorter period of reduced runoff from the Sepik River in 1992-1993 marks the warm phase of the ENSO. The 1992-1994 ENSO had little effect on monsoonal rainfall in eastern Java, as reflected by the persistance of the monsoon in the coral record. On the hot, dry coast of northwestern Australia, the timing and magnitude of seasonal evaporation are accurately recorded by the Dampier coral. The $\delta^{18}O$ residuals become progressively more positive in the austral spring and reach maxima in November–December. This pattern coincides with increasing air temperature and solar radiation, both of which would serve to increase ocean surface evaporation and the concentration of ^{18}O in seawater.

3.4. Post-glacial tropical SSTs

Geochemical tracers also can be applied to fossil corals to gain new perspectives of tropical climate variability throughout the late Quaternary. Coral paleothermometers, combined with oxygen and carbon isotope ratios, are allowing us to explore the natural bounds in tropical SSTs, the hydrological cycle, and ocean circulation during the last full glacial cycle. So far, several studies have yielded information about past changes in tropical SST (Beck et al., 1992, 1997; Guilderson et al., 1994; McCulloch et al., 1996), global ice volume (Guilderson et al., 1994), surface-ocean hydrologic balance (Klein et al., 1990; Gagan et al., 1998), and ocean mixing (Edwards et al., 1993).

New paleotemperature estimates from the Sr/Ca ratio of corals from the Caribbean and western Pacific indicate that tropical SSTs were 5–6°C cooler than today during the mid-stages of the last deglaciation (10–14 ka; Guilderson et al., 1994; Beck et al., 1997). These results

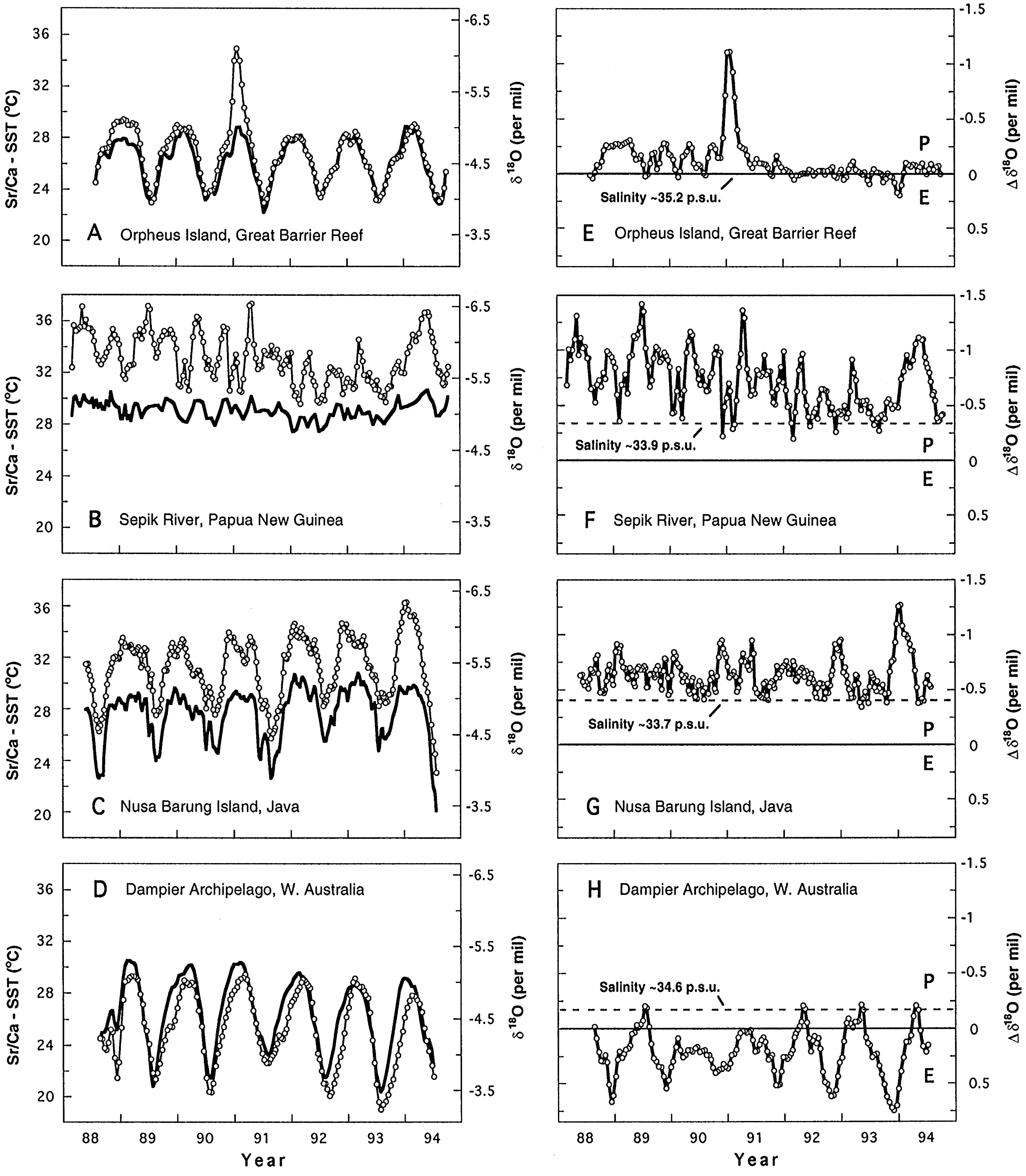

Fig. 6. Panels A–D: Comparison of Sr/Ca–SSTs (solid curves) and $\delta^{18}O$ values (curves with circles) for *P.lutea* from Orpheus Island, central Great Barrier Reef (18°45′S, 146°29′E); Sepik River, Papua New Guinea (03°31′S, 144°35′E); Nusa Barung Island, Java (8°31′S, 113°22′E); and the Dampier Archipelago, Western Australia (20°36′S, 116°45′E). The raw data have been filtered by applying a two-point running mean. The Sr/Ca–SST relation is: $T_{Sr/Ca} = 168.2 - 15674(Sr/Ca)$ of Gagan et al. (1998). For comparison "apparent" SSTs derived from the coral $\delta^{18}O$ values, which include offsets due to changes in seawater $\delta^{18}O$, are given by the relation: $T_{\delta^{18}O} = 1.02 - 5.56\,\delta^{18}O$ (Gagan et al., 1994; Gagan and Chivas, 1995). Panels E–H: Comparison of differences in seawater $\delta^{18}O$ ($\Delta\delta^{18}O$) after subtracting the temperature component of the $\delta^{18}O$ signal. $\Delta\delta^{18}O$ values are obtained using: $\Delta\delta^{18}O = \partial\delta^{18}O/\partial T[T_{\delta^{18}O} - T_{Sr/Ca}]$, where $\partial\delta^{18}O/\partial T$ is the empirically derived temperature-dependent oxygen isotope fractionation ($-0.18‰$ per degree Celsius) for the *Porites* investigated in this study. The solid horizontal lines define the mean, dry-season $\Delta\delta^{18}O$ value (defined zero) for the Orpheus coral, centred on a salinity of 35.2 p.s.u. Dashed horizontal lines estimate regional differences in salinity and the balance between precipitation (P; negative $\Delta\delta^{18}O$) and evaporation (E; positive $\Delta\delta^{18}O$), relative to Great Barrier Reef waters (solid horizontal lines). See text for discussion.

suggest that the envelope of potential SST change in the tropics may be large, in contrast to the CLIMAP SST reconstruction for the tropics during the last glacial maximum, which indicates little or no change (CLIMAP, 1981). For several regions of the globe, including the tropics, it is now clear that the transition to modern climate following the last glacial maximum was punctuated by a number of rapid and substantial climatic oscillations (Alley et al., 1993; Gasse and Vancampo, 1994; Thompson et al., 1998). In contrast, relatively little is known about how the tropical ocean responded to the deglaciation because few high resolution records are available from these regions (Hughen et al., 1996).

It is now possible to investigate the potential for abrupt climate change in the tropical surface-ocean via several deglacial to late Holocene coral paleotemperature records available from the southwestern Pacific region. Fig. 7 shows a compilation of paleo-SST estimates for corals collected from Papua New Guinea, the Australian Great Barrier Reef, Vanuatu, and Tahiti. The corals were dated using TIMS ^{230}Th, AMS ^{14}C, or U/Th measured by alpha spectrometry. Differences in paleo-SSTs were estimated, relative to modern values, using Sr/Ca, U/Ca, and δ^{18}O paleothermometers (see equations in Fig. 7). The Sr/Ca and U/Ca paleo-SSTs are estimated assuming the ocean water values have remained constant through time although there may be a slight bias toward cooler SSTs for those based on Sr/Ca ratios (Stoll and Schrag, 1998). Paleo-SST estimates based on δ^{18}O have been corrected for the ice volume effect, where necessary (Fairbanks, 1989). However, the paleo-SSTs based on δ^{18}O could be minimum estimates if the ice volume effect is smaller than the conventional figure (Schrag et al., 1996) or the carbonate ion concentration in the late-glacial ocean was different from today (Spero et al., 1997).

The U/Ca temperatures from Vanuatu corals show SSTs oscillating $\approx$1–3°C below modern values just prior to 12 ka BP. Immediately afterward, during the Younger Dryas climatic event, SST decreased dramatically to about 4–6°C below present and apparently remained there until about 10.3 ka BP. The remaining data are based on Sr/Ca or δ^{18}O proxy-temperature measurements which show that SST rose episodically during the next 4000 yr. This rise was interrupted by a moderate temperature decrease around 8 ka BP and culminated with an optimum temperature 1.2°C above present values at about 6 ka BP. By 4 ka BP SSTs had cooled to near modern temperatures.

There is evidence that the south tropical Atlantic Ocean also experienced anomalously cold SST at $\sim$9 ka BP, with mean February and August SST 2.25°C and 1.6°C below present temperatures (Ruddiman and Mix, 1993). Likewise, oxygen isotope temperature estimates from ice cores in the central Andes also appear to show unusually cold and rapidly rising atmospheric temperatures between 8 and 10 ka BP (Thompson et al., 1995). This view is reinforced by pollen records from tropical lake sediments in Madagascar, which show a substantial temperature depression occurring midway during the last deglaciation (Gasse and Vancampo, 1998). Such results suggest that the southwestern Pacific corals may be truly representative of the tropics in recording a widespread temperature depression and rapid temperature rise during the late deglaciation.

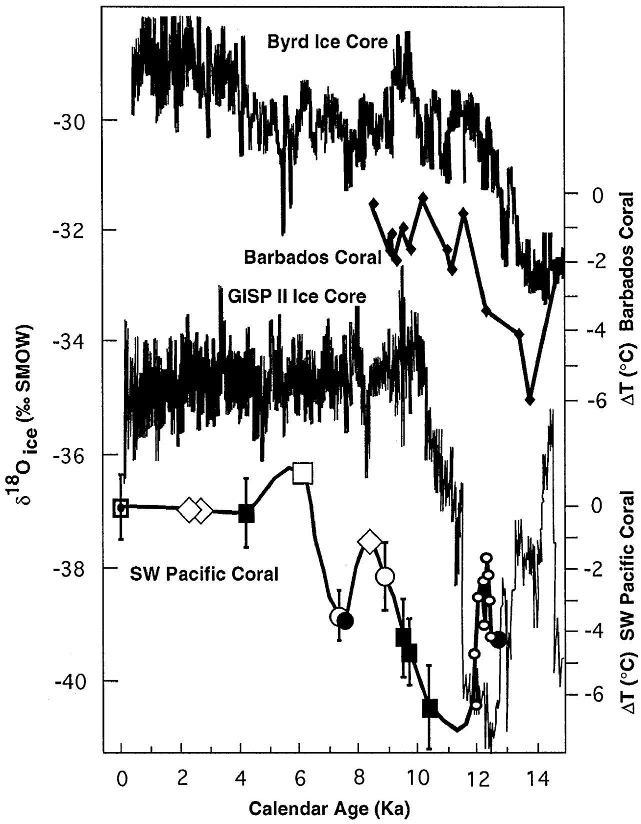

Fig. 7. Comparison of reconstructed SSTs for the southwestern Pacific and the tropical Atlantic (after Guilderson et al., 1994). Also shown for comparision are the δ^{18}O variations observed in the GISP II and Byrd polar ice cores (after Sowers and Bender, 1995). The Sr/Ca–SST relations used include: T(°C) $= 137.8 - 12\,575$(Sr/Ca) for Papua New Guinea; $167.8 - 16{,}013$(Sr/Ca) for Vanuatu, Seychelles, and Tahiti; and $168.2 - 15\,674$(Sr/Ca) for the Australian Great Barrier Reef. The U/Ca–SST relation is $T(°C) = 48.0 - [21.5 \times 10^6 \times (^{238}U/^{40}Ca)]$, derived by Min et al. (1995). The slope of the δ^{18}O–SST relation is $-0.23‰$ per degree celsius, as defined by Epstein et al., (1953). ○, Sr/Ca Papua New Guinea (McCulloch et al., 1996); ■, Sr/Ca Vanuatu (Beck et al., 1997), ●, Sr/Ca Seychelles, Tahiti (Castellaro et al., 1997); □ Sr/Ca Great Barrier, Australia (Gagan et al., 1998); ◇, δ^{18}O Papua New Guinea (Tudhop, unpublished data), ○, U/Ca, this study.

Not all records indicate that the tropics behaved uniformly during the deglaciation, however. In particular, although the Barbados coral SST reconstruction (Guilderson et al., 1994) indicates a rapid 5–6°C SST rise in the western tropical Atlantic during the late deglaciation, the timing appears to be shifted by nearly 3 ka relative to that seen in the southwest Pacific, with most of the tropical Atlantic warming occurring between 14 and 11.5 ka BP (Fig. 7). It is interesting that a phase shift in high latitude deglacial warming has also been determined (Sowers and Bender, 1995) based on a comparison of the GISP II and Byrd polar ice cores. In that case, however, the inter-hemispheric phasing is reversed, with warming in the Antarctic beginning approximately 3 ka before the warm Bølling period in Greenland. As seen in Fig. 7, the large rise in SST observed in the SW Pacific much more closely matches the timing of the post-Younger Dryas warming seen in the GISP II record than it does the Antarctic Byrd ice core record.

3.5. *Mid-Holocene SSTs and hydrologic balance*

The global climate change debate has led to renewed interest in analysing corals that grew during times when the earth was warmer than today, or warming rapidly. Although these climates of the past are not analogues for a CO_2-warmed Earth (Crowley, 1990), such records will certainly yield perspectives on processes driving the climate system (Rind, 1993). Well dated, seasonally resolved paleoclimate records extracted from corals are capable of revealing subtle changes in SSTs, rainfall variability, and evaporation within the annual cycle, all of which could provide clues to large-scale changes in the tropical ocean and atmosphere.

To illustrate the potential of fossil corals to yield detailed paleoclimate information, the paleo-SSTs and oxygen isotopic composition of seawater have been estimated for a period near the end of the deglaciation using tandem Sr/Ca and $\delta^{18}O$ measurements for *Porites* corals growing in the Australian Great Barrier Reef and at Sumba, Indonesia. The fossil corals have ocean reservoir-corrected radiocarbon ages of 5350 ± 60 yr BP (Great Barrier Reef) and 4330 ± 80 yr BP (Sumba). In both cases, we compare the fossil coral data directly with calibrated data for live *Porites* colonies growing adjacent to the fossil corals (Fig. 8). Details of the methodologies are described by Gagan et al. (1998).

The Sr/Ca paleotemperature estimate for the Great Barrier Reef coral indicates that the mean SST at about 5350 yr BP was 27.0°C, which is 1.2°C warmer than the mean SST for the early 1990s (Fig. 8). The difference between the coral Sr/Ca and $\delta^{18}O$ curves represents the residual $\delta^{18}O$ signal ($\Delta^{18}O$) that can be used to define the oxygen isotopic composition of seawater. The mean oxygen isotopic composition of ambient seawater can be determined using the $\delta^{18}O$ residuals for the austral winters, when transient changes in seawater $\delta^{18}O$ are negligible. The lines of best fit through the winter $\delta^{18}O$ residuals define an enrichment in ^{18}O of 0.47‰, relative to modern values. A similar picture is emerging from our work in Sumba. The lines of best fit through the winter $\delta^{18}O$ values, when transient changes in seawater $\delta^{18}O$ are negligible, define an enrichment in seawater ^{18}O of 0.54‰, relative to modern values. Preliminary Sr/Ca measurements indicate that SSTs in Sumba at about 4330 yr BP were 0.5°C warmer than today. If this is the case, the ^{18}O enrichment could be as large as 0.63%.

The ^{18}O enrichment in the mid-Holocene tropical western Pacific may be driven by enhanced surface-ocean evaporation in response to the higher SSTs at the end of the deglaciation (Gagan et al., 1998). Today, about 11 sverdrup (1 sverdrup = $10^6\ m^3\ s^{-1}$) of water evaporates from the warm surface of the tropics and subtropics, of which about 1.5 sverdrup is transported by the atmosphere and precipitated poleward of $\pm 40°$ (Broecker and Denton, 1989; Norton et al., 1997). Recent work has shown that even a small increase in tropical SST, such as the 0.5–1°C rise in SST reported here, leads to a marked increase in surface-ocean evaporation and water vapour in the atmosphere (Flohn et al., 1990; Graham, 1995). Enhanced evaporation is evident in the Orpheus Island fossil coral record where the $\delta^{18}O$ residuals become progressively more positive in spring and reach maxima between November to December, immediately before the onset of monsoonal rainfall (Fig. 8). Transport of some of this extra water vapour to higher latitudes, or elsewhere in the tropics, could tip the hydrologic balance toward evaporation and concentrate ^{18}O in the surface waters of the southwestern Pacific.

Theoretical and general circulation model studies also indicate that even subtle changes in the distribution of tropical heating, particularly warming off the equator, can profoundly alter the intensity of the Hadley circulation and the transport of water vapour and heat into the extratropics (Hou, 1993; Lindzen and Pan, 1994; Rind, 1998). Although the spatial distribution of the warmer mid-Holocene SSTs needs to be established, it is possible that the apparent drying of the tropical southwestern Pacific, and equable climates in the extratropical latitudes at this time (COHMAP, 1988; McGlone et al., 1994) are linked, in part, to an increased latent heat flux from the tropics. Such a redistribution in atmospheric moisture could produce large and abrupt changes in global climate (Broecker, 1996) by altering the high-latitude surface ocean salinity field and structure of the oceanic thermohaline circulation (Kerr, 1998).

4. Using Radiocarbon in Corals to Constrain Ocean Circulation

The circulation of the world oceans plays a fundamental role in human activity through modulating climate, distributing nutrients essential for biological productivity, and controlling cycling of carbon including excess CO_2 from fossil fuel combustion. To study how the ocean circulates, oceanographers have used chemical tracers to complement observations of fundamental

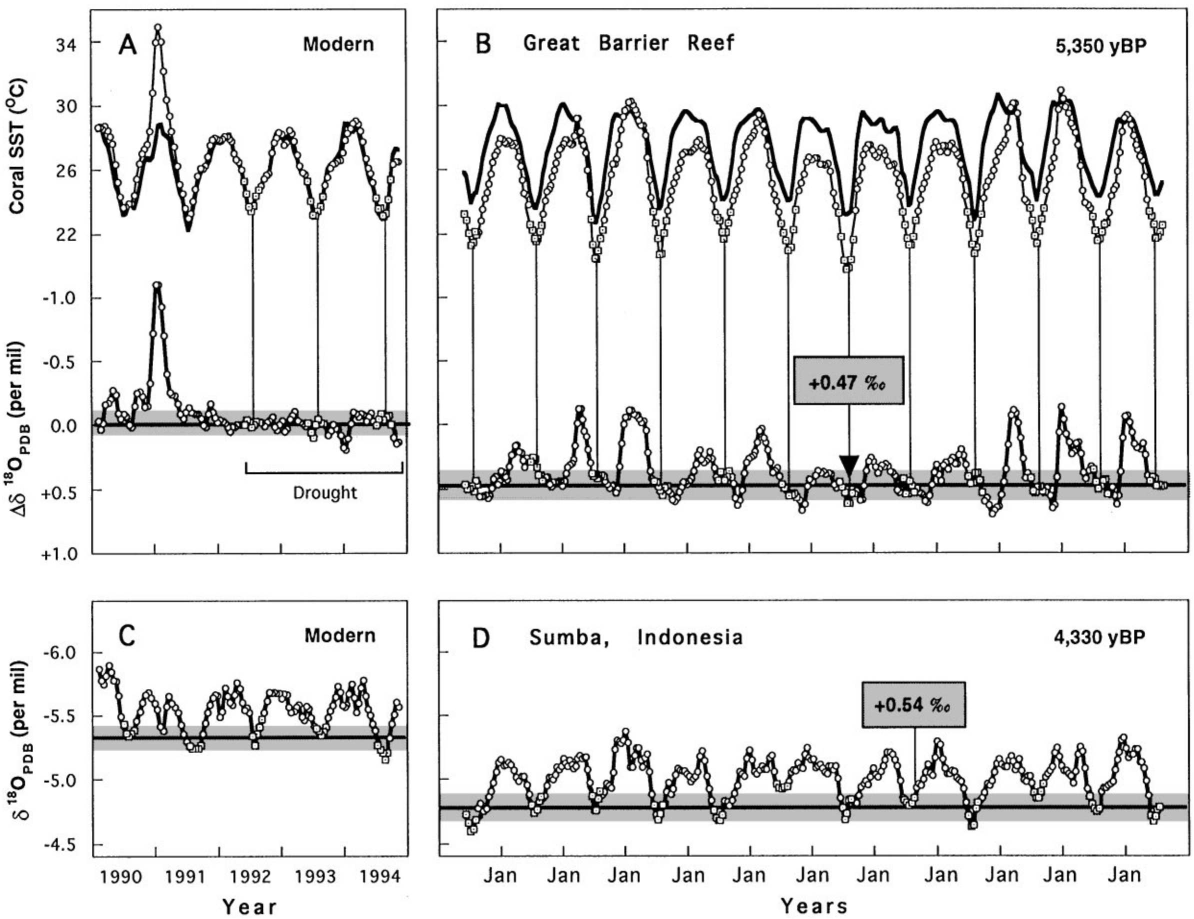

Fig. 8. Panels A–B: Comparison of calculated coral Sr/Ca (solid curves) and $\delta^{18}O$ temperatures (upper curves with circles) for the modern and 5350 yr BP Porites from Orpheus Island, central Great Barrier Reef. Differences in seawater $\delta^{18}O$ ($\Delta\delta^{18}O$, lower curves with circles), relative to the modern mean, are obtained by removal of the temperature component of the $\delta^{18}O$ signal using the equation defined in Fig. 6. The horizontal lines show the mean $\Delta\delta^{18}O$ of seawater, as defined by the seven $\Delta\delta^{18}O$ values (squares) falling in the austral winters (vertical lines). Panels C–D: Comparison of coral $\delta^{18}O$ values for the modern and 4330 yr BP Porites from Sumba, Indonesia. The horizontal lines estimate the mean $\Delta\delta^{18}O$ value of seawater as defined by the five $\delta^{18}O$ values (squares) falling in the winters, when transient changes in seawater $\delta^{18}O$ are negligible. The raw data have been filtered by applying a two-point running mean.

properties such as temperature and salinity. Among the transient tracers that are most useful for studying ocean circulation is radiocarbon measured as $\Delta^{14}C$ of the dissolved inorganic carbon (DIC) in seawater. Radiocarbon is produced both naturally in the upper atmosphere and as a result of thermonuclear testing in the 1950s and early 1960s. Post-bomb atmospheric ^{14}C levels increased by nearly 100% (1000‰) from 1950 to 1963. The spike of bomb $^{14}CO_2$ has decreased in the atmosphere during subsequent years, mostly as a result of isotopic exchange with $^{12}CO_2$ in the surface ocean. The penetration of bomb ^{14}C into the water column has been measured by several investigators over the years as a component of water-sampling programs such as the Geochemical Ocean Sections Program (GEOSECS) and the World Ocean Circulation Experiment (WOCE) (Linick, 1980; Östlund and Stuiver, 1980; Quay et al., 1983; Broecker et al., 1985). These ^{14}C data have been important for critical tests in ocean circulation models (Toggweiler et al., 1991).

Within the past two decades, several workers have reported that hermatypic coral skeletons growing in the upper 40 m of the temperate and tropical oceans (32°N to 32°S) record the input and spatial distribution of ^{14}C in the oceans. From these records, researchers have uncovered evidence of past changes in surface-subsurface mixing (Nozaki et al., 1978; Druffel, 1989; 1997b), major current shifts (Druffel and Griffin, 1993; Guilderson et al., 1998), and changes in thermocline depth (Guilderson and Schrag, 1998). These coral records supply information unavailable either from modern instrumental records or from tracer studies based on water-sampling efforts, which provide only a snapshot of ocean conditions. By incorporating the radiocarbon time series into ocean circulation models (e.g., Rodgers et al., 1997), coral records can be used to study variability in ocean circulation over a range of time scales.

4.1. Decadal to centennial variations in upper ocean radiocarbon

Over longer time scales of several centuries, records of $\Delta^{14}C$ in corals have been used to reconstruct past mixing rates between surface and subsurface waters in an ocean basin (Druffel, 1989; 1997b), detect the presence of fossil fuel-derived CO_2 in the upper ocean (Druffel and Linick, 1978; Nozaki et al., 1978), and help determine the decadal variability of ocean circulation (Druffel and Griffin, 1993). A biennial $\Delta^{14}C$ record for a coral from Abraham Reef in the southern Great Barrier Reef of Australia is shown in Fig. 9 (Druffel and Griffin, 1993). This high precision $\Delta^{14}C$ record demonstrated variations on an interannual time-scale that were particularly large between AD 1680 and 1730. By comparison with tree ring $\Delta^{14}C$ records (Stuiver and Quay, 1980), it was clear that these shifts were not caused by changes in the $\Delta^{14}C$ of atmospheric CO_2. Changes in vertical mixing and large-scale, advective changes involving source waters to the western Coral Sea region are likely processes that could have accounted for these large $\Delta^{14}C$ variations. It was also observed that most low $\Delta^{14}C$ values for the period AD 1635–1875 coincided with El Niño events (Fig. 9). However, during 1875–1920, $\Delta^{14}C$ values remained high despite several El Niño events. Cross-spectral analysis of the early half of the $\Delta^{14}C$ and $\delta^{18}O$ records (AD 1635–1795) revealed that the 6-yr period was coherent; however, the coherency was not present in the latter half (AD 1797–1957) of the isotope records. These data support the concept of century time-scale changes in the nature of El Niño, as it is manifest in the southwest Pacific. Interestingly, the Abraham Reef coral $\Delta^{14}C$ record did not show a Suess effect (the lowering of $\Delta^{14}C$ from late 1800s through 1955 due mainly to CO_2 input from fossil fuel burning). This is coincident with the change that was observed in the nature of El Niño and is further evidence that a long-term change in mixing of upper waters occurred in this region.

4.2. Bomb radiocarbon in the tropical thermocline

In the post-bomb era, the radiocarbon spike in the atmosphere amplified the contrast in radiocarbon content between the surface and deep ocean, rendering the patterns of radiocarbon variability in the surface ocean very sensitive to vertical mixing. The evolution of radiocarbon in the surface ocean in response to nuclear testing was first described by Nozaki et al. (1978) and Druffel and Linick (1978) using corals sampled at annual resolution. More recently, large seasonal and interannual variability (i.e., 30–100‰), of the same order as the spatial variability of radiocarbon across the surface ocean ($\sim$ 100‰), was identified in corals sampled at sub-annual resolution at a variety of locations (Druffel, 1987; Brown et al., 1993; Moore et al., 1997; Guilderson et al., 1998; Guilderson and Schrag, 1998). In addition, Takahashi et al. (1991, unpublished data) and Masiello et al. (1998) have observed seasonal variability in surface water radiocarbon of similar amplitude in water samples from the open Pacific. Because air–sea isotopic equilibration (gas exchange) occurs too slowly to account for large variability over time scales shorter than one year (it is roughly an order of magnitude slower than the seasonal cycle), a dynamical process internal to the ocean must be responsible for the large seasonal ^{14}C variability. Rodgers et al. (1997) used an ocean circulation model to show that most of the seasonal and interannual

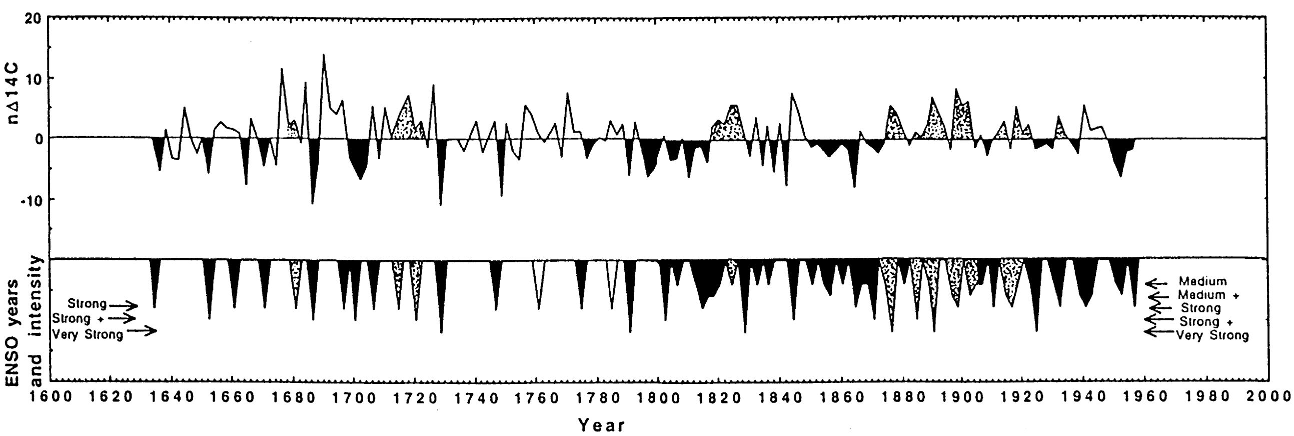

Fig. 9. Biennial Australian Great Barrier Reef coral $\Delta^{14}C$ values (detrended with respect to a third-order polynomial fit) plotted with incidences of strong and very strong ENSO events before AD 1800 and including medium ENSO events after AD 1800 (plotted biennially) as reported by Quinn et al. (1987). Solid peaks represent correlation between ENSO events and coincident ($\pm$ 2–4 yr) $\Delta^{14}C$ lows, stippled areas represent correlation between ENSO events and coincident $\Delta^{14}C$ highs, and open areas indicate no correlation (from Druffel and Griffin, 1993).

variability in the tropical Pacific reflected variability in equatorial upwelling in the eastern Pacific and lateral redistribution of radiocarbon by shifting surface currents. They suggested that comparison of coral radiocarbon records with ocean circulation models could be used to improve the skill of the models as well as to study patterns of ocean circulation including upwelling dynamics and movements of tropical currents, particularly at times when instrumental data are not abundant (i.e., prior to the deployment of stationary moorings in the 1980s).

Current research efforts are underway to describe the spatial and temporal evolution of radiocarbon in the surface ocean using AMS measurements of $\Delta^{14}C$ in corals. As an illustration of the ongoing efforts, we consider records from the tropical and subtropical Pacific from Druffel (1987), Toggweiler et al. (1991), Guilderson et al. (1998), Guilderson and Schrag (1998), and Beck et al. (submitted) (Fig. 10). The patterns of variability from different records reflect the dynamic circulation of the tropical thermocline. In the subtropical coral records (Fiji and French Frigate Shoals), radiocarbon increases rapidly in the 1960s, and peaks in the early 1970s, reaching $\Delta^{14}C$ values in the North Pacific of 140–190‰ (Fig. 10). The relatively high peak reflects the stability of the subtropical gyres. In contrast, the coral record from the eastern tropics (Galapagos) reflects equatorial upwelling, which supplies water to the surface ocean from subduction of water in the subtropics and from deeper sources of water entrained in the tropical undercurrent. The transport time of water from the subtropics to the tropics (~ 10–20 yr), coupled with some amount of mixing, delays the peak in $\Delta^{14}C$ at Galapagos through the 1980s. Variability in the $\Delta^{14}C$ values in the eastern Pacific relate to seasonal and interannual variability in the equatorial upwelling. Coral radiocarbon records from the central and western tropical Pacific (Christmas and Nauru) reflect mixing between these two end-members, low-$\Delta^{14}C$ water from the eastern upwelling and higher $\Delta^{14}C$ water from the subtropics. Most of this mixing is driven by changes in wind-driven currents over interannual timescales (Rodgers et al., 1997). The peak in $\Delta^{14}C$ in the western Pacific occurs in the early 1980s, intermediate between the subtropics and the eastern tropics.

The general pattern of interannual radiocarbon variability along the equator is strongly associated with ENSO. Variability in thermocline depth in the eastern tropical Pacific causes large variability in radiocarbon at Galapagos, particularly during the upwelling season (Guilderson and Schrag, 1998). In the central and western tropical Pacific, ENSO warm events cause $\Delta^{14}C$ values at Nauru and Christmas to increase, often followed by

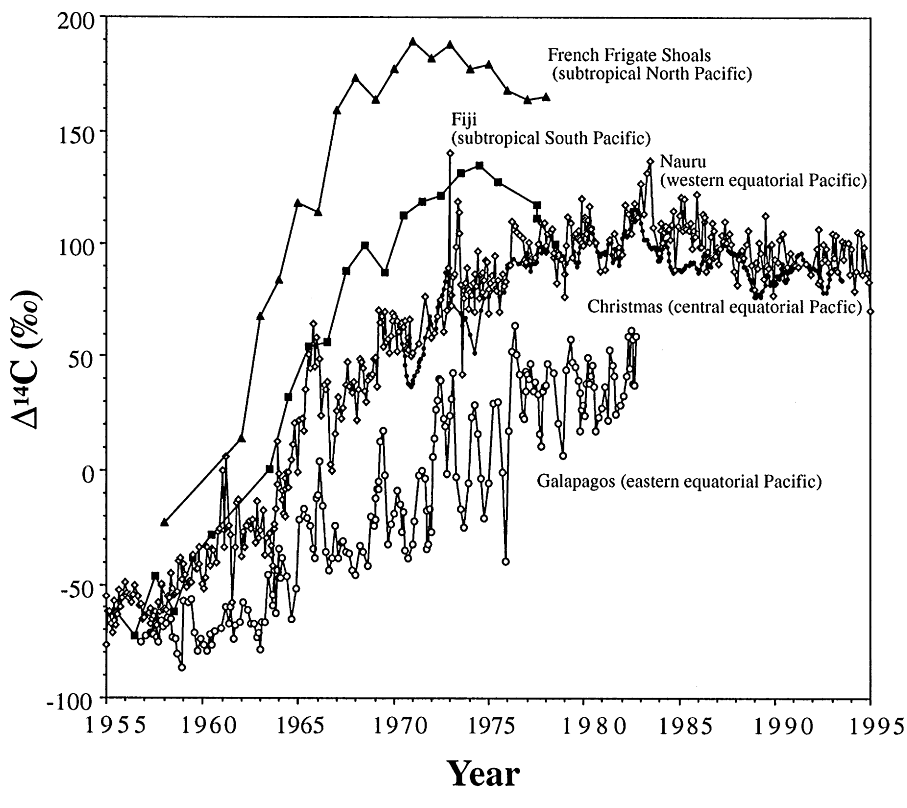

Fig. 10. Post-bomb radiocarbon records from five locations in the tropical and subtropical Pacific. Data are from French Frigate Shoals (Druffel, 1987), Fiji (Toggweiler et al., 1991), Nauru (Guilderson et al., 1998), Christmas Island (Beck et al., submitted), and Galapagos Islands (Guilderson and Schrag, 1998).

a large and rapid decrease, and then a return to mean values (Fig. 10). In a simple sense, this pattern can be understood from a conceptual model of ENSO. As a warm event begins, the upwelling of deep, radiocarbon-poor water from the upper thermocline is diminished, causing an increase in $\Delta^{14}C$ in the eastern Pacific. At the same time, the transport of low $\Delta^{14}C$ water westward slows down as tradewinds decrease in strength, allowing for more invasion of bomb-radiocarbon from the atmosphere, which also causes $\Delta^{14}C$ values to rise in the west. If westerlies develop along the equator at the western margin as occurs during some strong warm events, off-equatorial waters with much higher radiocarbon content are brought into the tropics, amplifying the increase in $\Delta^{14}C$ values. The large decreases in $\Delta^{14}C$ following ENSO warm events are consistent with the re-establishment of normal tradewinds as normal conditions return. In this situation, one expects an intensification of equatorial upwelling, and an increase in the strength of westward-flowing currents, bringing more water with lower radiocarbon into the eastern Pacific sea surface, and then transporting that water more rapidly to the west with less time to exchange with the atmosphere. In the subtropics, radiocarbon varies less with ENSO, and shows much stronger decadal variability (Druffel and Griffin, 1993). The processes that control this variability are not well understood.

An additional use of coral radiocarbon time series data is to identify unusual patterns that may reflect shifts in circulation patterns or upper ocean structure. For example, Guilderson and Schrag (1998), in a radiocarbon time series from a Galapagos coral, noted an abrupt increase in radiocarbon during upwelling seasons after 1976, a time when studies have noted that ENSO may have changed (Graham, 1994; Trenberth and Hurrell, 1994). This shift coincided with an abrupt increase in sea surface temperature during the upwelling season. The $\Delta^{14}C$ data require that the source of upwelling contained a greater proportion of water with bomb-radiocarbon after 1976. They suggested that the simplest way to satisfy the SST and $\Delta^{14}C$ constraints is if the average thermocline depth increased in 1976. This suggestion raises fundamental issues regarding decadal variability in tropical thermocline depth, many of which can be addressed with radiocarbon and other tracers in future studies.

5. Conclusions and future directions

The expanding network of coral oxygen isotope records is yielding new information on multi-decadal climate variability over the past several centuries. In the eastern equatorial Pacific, coral oxygen isotope records track ENSO-related changes in SST; further west, corals record variability in SSTs and rainfall that accompanies zonal displacement of the Indonesian Low during ENSO events. These multi-century records reveal previously unrecognized ENSO variability on time-scales of decades to centuries. Outside the ENSO-sensitive equatorial Pacific, a long-term trend towards recent warmer/wetter conditions is commonly punctuated by abrupt shifts in $\delta^{18}O$, which suggest that the tropics may be responding to global forcings. These changes may in turn contribute to long-term global climate variability via water vapour and dynamical feedbacks.

Sr/Ca and U/Ca ratios in well preserved fossil corals from the tropical western Pacific and Atlantic suggest temperatures were depressed by 4–6°C during the mid-stage of the last deglaciation (10–14 ka). Like high latitude ice core records, the coral proxies also show that several rapid climatic oscillations occurred during the transition to a Holocene temperature maximum $\sim$ 6000 yr ago. The advances in coral paleothermometry have opened the way to the use of coupled measurements of Sr/Ca and $\delta^{18}O$ in corals to uniquely reconstruct surface-ocean temperature and the $\delta^{18}O$ of seawater. The residual $\delta^{18}O$ signal provides information about the surface-ocean hydrologic balance and high temporal-resolution data sets can resolve the seasonal balance between precipitation and evaporation. With further refinement, these tools could be used to reconstruct important components of the global climate system such as the evolution of warm pools, monsoonal rainfall patterns, and the ENSO.

While shipboard water-sampling programs can provide a broad view of ocean ^{14}C concentrations in three dimensions, they cannot capture information on the temporal variability of this tracer. Radiocarbon measurements in corals provide continuous histories of ^{14}C variability extending back several centuries. The most recent studies have focused on AMS measurements of radiocarbon in corals at seasonal and interannual time-scales to constrain near-surface ocean circulation in the tropical Pacific. Some of these records reveal large interannual variability in mixed-layer $\Delta^{14}C$ which is strongly correlated with ENSO. These fluctuations are linked to the seasonal and interannual redistribution of surface waters in the tropical Pacific, as well as to ENSO-modulation of low-^{14}C water upwelling in the eastern and central equatorial Pacific. When coupled with ocean circulation models, coral radiocarbon records can be used to address key questions regarding interannual and decadal variations in ocean circulation, particularly with respect to the intensity and placement of zonal currents and their interaction with ENSO. The results of such studies will improve our understanding of circulation patterns in the tropical Pacific as well as improve the accuracy of ocean circulation models.

Despite past successes in the use of corals for paleoclimate reconstruction, several issues regarding their utility remain to be resolved. As new climatic tracers are identified there is an ongoing need to understand fully the underlying physiological processes controlling isotopic and elemental variations in corals. Uncertainties remain about the specific climate mechanisms that are responsible for producing the geochemical signals in corals. Replication of long isotopic and elemental records is also necessary to allow signal precision to be established based on differences between records. Toward this end, recent studies indicate that inductively coupled plasma mass spectrometry (ICP-MS; Le Cornec and Correge, 1997) and inductively coupled plasma atomic emission spectroscopy (ICP-AES; Schrag, 1999) can be used to greatly increase sample throughput for the analysis of elemental ratios, while maintaining appropriate data precision. The investigation of elemental ratios in corals that record purely temperature or salinity should be a high priority.

Over the longer term, it is important to expand the multi-proxy approach to explore other important paleoclimate archives and synthesise compatible data sets. Work is under way to establish three-dimensional reconstructions of oceanic processes by examining the distribution of light stable isotopes, elemental ratios, and radiocarbon in very long-lived, thermocline dwelling sclerosponges (Druffel and Benavides, 1986; Bohm et al., 1996) and deep-sea corals (Smith et al., 1997; Druffel, 1997b, Adkins et al., 1998). Terrestrial palaeoclimate reconstructions to be synthesised with coral data include annually resolved records from tropical ice cores, tree rings, varved sediments, speleothems, and lacustrine fish otoliths. Clearly we are still in the early and exciting stages of developing a comprehensive, multi-proxy approach to high-resolution tropical paleoclimatology.

Acknowledgements

The catalyst for this review paper came from the new PAGES/CLIVAR initiative: Annual Records in Tropical Systems (ARTS). We gratefully acknowledge financial support from the National Science Foundation (NSF), the National Oceanic and Atmospsheric Administration (NOAA), and the Australian National Greenhouse Advisory Committee. We thank the NOAA World Data Center-A for Paleoclimatology for archiving the coral oxygen isotope data presented in Fig. 2

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