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Sea surface temperature and salinity reconstruction from coral geochemical tracers

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Abstract

Massive scleractinian corals secrete an aragonitic skeleton which incorporates a large array of chemical tracers. Corals present several advantages for palaeoclimate research: they grow continuously, and can live up to 1000 years; they are easy to date; and they can be sampled at high resolution (weekly to monthly resolution). Both live and fossil corals can be collected in the field. In the past two decades, significant efforts have been made to identify robust tracers of sea surface temperature (SST) and salinity (SSS) in corals. To date, Sr/Ca and δ^{18}O are considered to be the most reliable SST tracers, although changes in seawater δ^{18}O can significantly alter SST reconstructed from coralline δ^{18}O. Because these variations in seawater δ^{18}O can be linked to SSS changes, this initial problem can in fact be turned into an advantage and provide us with an SSS tracer. The SST component in the coral δ^{18}O signal can either be evaluated through Sr/Ca measurements, or in some case simply filtered out. However, there is still much uncertainty concerning the exact mode of incorporation of trace elements and stable isotopes into the coral skeleton. The effects of growth rate, light intensity, feeding habits, pH and water chemistry are still poorly documented. A review of the strength and weaknesses of Sr/Ca and δ^{18}O is presented, together with some examples of SST and SSS reconstructions. Other potential SST tracers are also reviewed. It is expected that the ability to grow corals in aquarium under controlled conditions, and that the development of sophisticated analytical techniques at the micrometric level should help us understand better the robustness of each tracers and the factors controlling their incorporation in coral aragonite.

Keywords: Corals; Geochemical tracers; Temperature; Salinity

1. Introduction

Ever since life started to flourish on Earth, coral reefs have been an important part of the biosphere. Reef communities have evolved through geological time, and their study has always been a source of invaluable information for palaeoenvironmental scientists. The present-day reef building corals, of the order Scleractinia, have been present since the Triassic, and their modern distribution reflects the combined influence of climate change and plate tectonics through time (Veron, 1995). Since these environmental factors play such an important role in the development of corals, the idea emerged early to use them to gain information on past climate changes. From the pioneering work of the early 1970s to the CLIVAR/PAGES program ARTS (Annual Records of Tropical Systems; see the web site http://pangea.stanford.edu/Oceans/ARTS/), in which coral studies have a prominent role, many geochemical tracers have been developed in corals. As research pro-

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gressed, earlier reviews gave an overview of the state of research at particular times (Shen and Sanford, 1990; Dunbar and Cole, 1993; Druffel, 1997; Fairbanks et al., 1997; Gagan et al., 2000; Grottoli, 2001; Felis and Pätzold, 2003). The purpose of the present review is to give an up-to-date status on the various tracers used to gain information on the two main oceanic parameters: sea surface temperature (SST) and sea surface salinity (SSS).

2. Interest of Scleractinia for palaeoclimatologists

2.1. Geographic location and habitat

Although some corals can withstand SST as low as 11 °C (Veron, 2000), most coral reefs are found within the tropical zone, where SST rarely drops below 18 °C. This restriction of coral reefs to the tropics makes them ideal tools to study this zone which is of paramount importance to the climate of our planet. On average, the Equator receives 2.5 times more solar energy than the poles. A large part of this excess heat is transported from the tropics to the mid and high latitudes, as latent and sensible heat. Without this meridional heat transport, the equatorial region would be 14 °C warmer, and the poles 25 °C cooler (Barry and Chorley, 1998). There is also a zonal transport of heat in the tropics (the Walker circulation), which can vary on an interannual timescale. The most prominent phenomenon affecting zonal heat transport is the El Niño Southern Oscillation (ENSO). There is now ample evidence to suggest that ENSO alters the climate outside the tropics through teleconnections, over large parts of the globe (Philander, 1990).

In this context, and given the shortness of most instrumental records in the tropics, Scleractinia corals have a key role to play (e.g. Carriquiry et al., 1994; Cole and Fairbanks, 1990). They are present in the Pacific, Indian and Atlantic oceans, although their distribution is far from uniform (Fig. 1). The Indo-Pacific warm pool, the Red Sea and the Caribbean are the zones where most reefs are found. To date, most studies have concentrated on the largest and most diverse coral province in the world, the Indo-Pacific Warm Pool (IPWP). The IPWP is the area where ENSO events originate (Philander, 1990), and is therefore a key location to study in order to improve our understanding of interannual climate variability.

In terms of habitat, most scleractinian corals live in shallow waters (typically <50 m). They will therefore yield information on surface waters and on the upper thermocline. Although they prefer clear waters with little salinity variations, they can be found near river mouths, where they will potentially record run off and drought episodes. Note that some azooxanthellate scleractinians can live in very deep (i.e. more than 1000 m) and cold water. Such corals can form large deep-sea mounds (Mortensen et al., 1995) and their potential as tracers of past deep ocean circulation is increasingly gaining attention (Adkins et al., 1998; Smith et al., 2000; Adkins et al., 2003).

2.2. Mode of growth

Corals are formed of polyps, which secrete an aragonitic skeleton. Scleractinia have various growth forms: branching, tabulate, massive, columnar, foliaceous, encrusting or free-living. All Scleractinia can in theory be used for geochemical studies, however palaeoclimatologists have concentrated on massive corals for several reasons. First, massive corals are very resistant to mechanical breakage and erosion (as opposed to branching and foliaceous corals for example), and second, can form colonies 5 to 10 m high (Fig. 2 (2)). Growth rates are variable, and range from a few millimeters per year to up to 3 cm per year for some

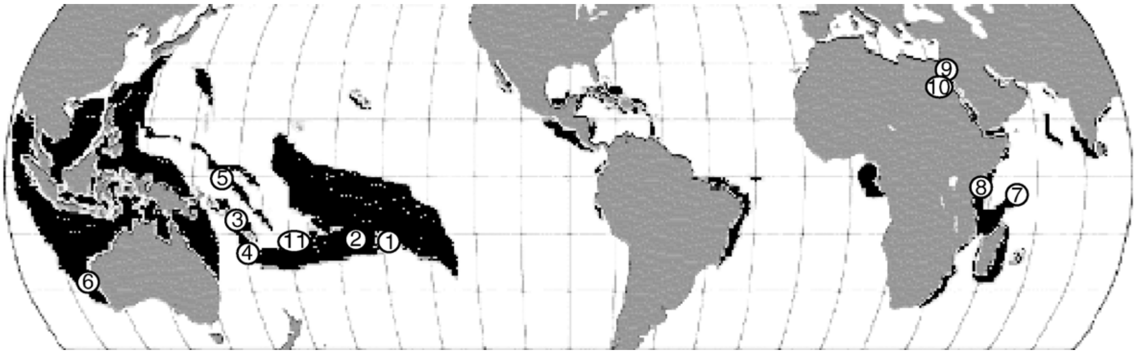

Fig. 1. Distribution of coral reefs in the world. Numbers refer to sites discussed in the text and on the other figures. (1) Mooréa; (2) Rarotonga; (3) Santo; (4) New Caledonia; (5) Maiana; (6) Abrohlos; (7) Seychelles; (8) Malindi; (9) Aqaba; (10) Ras Umm Sidd; (11) Fiji.

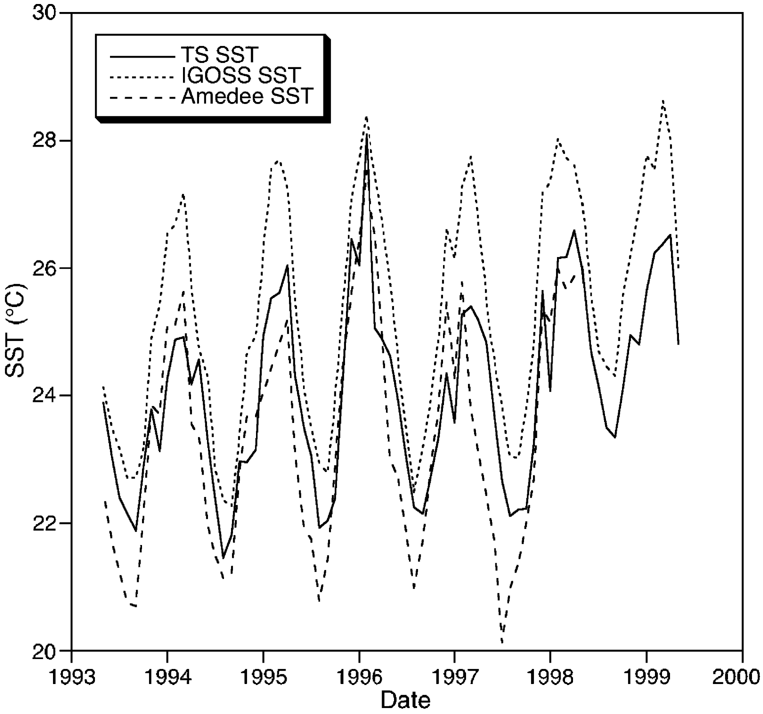

Fig. 2. Plot of the three SST data sets from the south west of New Caledonia.

Porites specimens. This means that pluri-metric colonies are several centuries old. A 2-m-high fossil colony of the slow-growing *Diploastrea heliopora* recovered from Vanuatu (southwest Pacific) spanned more than 700 years (Burr et al., 1998; Corrège et al., 2004), and a 3-m-high live colony of the same species from New Caledonia is believed to be more than 1000 years old.

In these colonies, annual growth bands consisting of a couplet of high and low density bands are often easy to identify through X-radiographs of the aragonitic skeleton (Knutson et al., 1972). Extensive discussions on the origin and significance of these density bands can be found in Barnes and Lough (1993), and although we still do not fully understand how density bands are formed, they are a useful tool for sclerochronology (the coral-based equivalent of dendrochronology; see for example Lough and Barnes, 1997), and for choosing sampling resolution and optimal sampling tracks on X-radiographs.

Compared to sediment cores, which are often affected by bioturbation, corals offer a continuous and usually undisturbed record. Potential disturbances to coral records include the physical destruction of the skeleton by grazing fish or boring organisms and bleaching events during which the coral stops growing. Grazer marks or bore holes are usually visible on X-radiographs and can be avoided when sampling. In area where grazers and borers proliferate, one should look for corals not affected by these problems, such as *D. heliopora* (Veron, 2000). Bleaching events are sometimes more difficult to identify, and it is not possible to know precisely how many months it took the coral to recover and to start growing again.

2.3. Precipitation and chemical composition of the coral skeleton

The aragonite secreted by coral polyps is an orthorhombic form of calcium carbonate ($CaCO_3$) which is less stable than calcite. It can contain various minor and trace elements, mainly divalent cations such as Sr, Ba, Mg, Zn, Pb and Mn.

Despite more than 120 years of research in this field, the exact process by which corals form their skeleton is still not fully understood (see for example Gattuso et al., 1999; Cohen and McConnaughey, 2003; Allemand et al., 2004). The classical view is that skeletogenesis is initiated in “centers of calcification” by precipitation of granular crystals near the calicoblastic ectoderm of the polyps. From these centers, crystalline fibers called sclerodermites radiate and fill the gaps. For a long time, the latter process was seen as an inorganic precipitation (Bryan and Hill, 1941; Barnes, 1970), although organic compounds were known to occur in coral skeletons (Wainwright, 1963; Young, 1971; Johnston, 1980; Constantz and Weiner, 1988). Recent investigations on the fine-scale structure of the supposedly "pure" aragonitic sclerodermites have proved the presence of various organic compounds as fine micrometric layers separating mineral phases (Cuif et al., 1999, 2003). In total, the non-carbonate phase of coral skeletons (i.e. organic material and water) accounts for 2.5% to 3% of its total weight (Cuif et al., 2004). The contribution of organic compounds represents about 1% in weight (Cohen and McConnaughey, 2003; Cuif et al., 2004). These findings are essential as they might have profound implications for the processes involved in the fractionation of trace and minor elements into the aragonitic skeleton. However, much more work is needed to fully understand the process of calcification. For example, it is still not clear whether the initial calcification takes place intracellularly, or just outside the calicoblastic ectoderm.

The pathways of the various chemical elements involved in calcification is also an area full of controversy (Gattuso et al., 1999; Allemand et al., 2004). To form its skeleton, the polyp, which has a low intracellular Ca^{2+} concentration, needs to transport a huge amount of Ca^{2+} to the calcification sites (up to 1700 nmol cm^2 h^{-1}; Wilbur and Simkiss, 1979). This transport is believed to involve several steps, and to be carrier mediated (Chalker, 1981; Wright and Marshall, 1991; Marshall, 1996; Tambutté et al., 1996; Marshall and Wright, 1998; Ferrier-Pagès et al., 2002; Clode and Marshall, 2002; Al-Horani et al., 2003; Zoccola et al., 2004). The incorporation of Ca^{2+} in corals also displays saturation kinetics and interaction with Sr^{2+} in yet another controversial fashion. Some authors postulate that Sr^{2+} and Ca^{2+} are transported by different mechanisms (Ip and Krishnaveni, 1991; Ip and Lim, 1991), while others invoke a common pathway for both ions (Goreau, 1977; Chalker, 1981; Wright and Marshall, 1991; Ferrier-Pagès et al., 2002).

Another complication in skeletogenesis is the fact that hermatypic corals possess endosymbiotic, unicellular algae (zooxanthellae) inside their endodermal tissues (e.g. Johnston, 1980). The early realization that calcification varies with the amount of light (Kawaguti and Sakumoto, 1948; Goreau, 1959) lead to the concept that zooxanthellae play an active role in calcification. However, it has been demonstrated that the sites of photosynthesis and calcification are different (Vandermeulen and Muscatine, 1974). The role of zooxanthellae in calcification is therefore not a simple one, and probably involves complex mechanisms (see Gattuso et al., 1999; Allemand et al., 2004 for a review). The ratio of light to dark calcification (i.e. calcification occurring during daytime vs. nighttime) is highly variable, ranging from slightly negative values to 127, but a compilation of 26 publications indicates that the median value is 3 (Gattuso et al., 1999). Using isotopic tracers, Marshall and Wright (1998) showed that the sites of light and dark calcification are distinct, a result of importance for the use of palaeoenvironmental tracers in coral skeletons. Although most authors consider that calcification is light enhanced due to the presence of zooxanthellae, Marshall (1996) put forward the controversial hypothesis that calcification might be in fact dark-repressed.

This short overview of skeletogenesis in hermatypic corals highlights the fact that despite notable progress in our understanding of the process, we still lack a complete theory to explain how major and trace elements are incorporated in coralline aragonite. When available, more specific information will be given for each tracer in the following chapters.

2.4. Age determination

One major interest of corals is the possibility to date them rather precisely for the period spanning the last 400 000 years. Corals contain trace elements such as uranium and lead, which make them datable with techniques such as ^{210}Pb, ^{14}C and U/Th. The ^{210}Pb technique is applied to live corals, as it can give ages ranging from 0 to 100 years. On the other hand, ^{14}C and U/Th can be applied to both modern and fossil material, roughly up to 40 000 years old and 400 000 years old, respectively. The combination of U/Th and ^{14}C dating on corals has been successfully applied to decipher ^{14}C production variability through time in the upper atmosphere and has improved and extended our calibration of the ^{14}C timescale (Bard et al., 1990).

Corals can also be dated relatively by density band counting, although the annual character of density bands is sometimes questionable. In corals growing near a river mouth, fluorescent bands can be found and used to date some flood events, helping in the establishment of an overall chronology (Isdale et al., 1998; Barnes et al., 2003).

Conducting geochemical work on corals is time and money consuming, and very often, only one core from a specific location is used. However, a recent study (Hendy et al., 2002), building on the work on growth bands developed by Lough and Barnes (1997), has successfully used up to eight cores from the Great Barrier Reef spanning the past 500 years to reconstruct a composite record of SST and SSS. This work uses techniques of cross dating between cores which are similar to those employed in dendrochronology, and paves the way for future studies with a more regional character.

An exciting new development of dating techniques has recently been published by Cobb et al. (2003a,b). Inspired once again by cross-dating work on trees, they have shown that fossil corals from the past 1000 years collected on Palmyra Island (central equatorial Pacific) can be dated by the U/Th technique with enough precision to overlap several records and create a splice. This important piece of work demonstrates the potential for reconstructing continuous, multi-millennial time series from different corals.

3. Oceanographic physical parameters and their tracers in corals

3.1. Sea surface temperature

Sea surface temperature (SST) is one of the most important parameters in oceanographic and climate studies and is certainly the one that has received the most attention. The behaviour of many chemical elements, from isotopes to trace metals, is believed to be more or less linked to change in temperature, thus providing us with a large array of potential SST tracers. The difficulty however, is to understand the exact nature of the SST influence, and to separate it from other potential sources of variation affecting a tracer. In the following sections, the present state of knowledge on the two main SST tracers: Sr/Ca and $\delta^{18}O$, will be presented, and other potential tracers such as Mg/Ca, U/Ca and some less familiar ones will be briefly discussed.

3.1.1. Sr/Ca

Although historically $\delta^{18}O$ has been first used as a SST tracer, Sr/Ca is in theory a much "cleaner" tracer, simply because variability in seawater Sr/Ca is far less than that of $\delta^{18}O$. Indeed, one of the prerequisites for a temperature tracer is that its concentration in seawater is stable through time. $\delta^{18}O$ is strongly influenced by the precipitation/evaporation ratio, whereas Sr/Ca is believed to be more stable in seawater. Subtle changes in the seawater Sr/Ca ratio through space (de Villiers et al., 1994; de Villiers, 1999) and time (Stoll and Schrag, 1998) can however affect the accuracy of reconstructed SST.

Early work on Sr/Ca in coral skeletons failed to recognize any temperature effect (Thompson and Chow, 1955; Bowen, 1956; Thompson and Livingston, 1970), mainly because of the sampling procedures used and the poor analytical precision available in those days. Weber (1973) was the first one to acknowledge a possible temperature component in Sr/Ca ratios, but he concluded that the effect of growth rate was almost an order of magnitude larger than that of temperature. Controlled laboratory growth experiments and field observations led Houck et al. (1977) and Smith et al. (1979) to reject the influence of growth rate and to propose the first equations linking Sr/Ca ratios to SST in three coral genera, including the now widely used *Porites* genus. It is interesting to note that more than 30 years after Weber's (1973) publication, the debate on the importance of growth rate is still very active. After Smith et al. (1979), different authors investigated the use of Sr as a palaeothermometer (e.g. Swart, 1981; Schneider and Smith, 1982; Oomori et al., 1982; Lea et al., 1989), but it is the article by Beck et al. (1992) that really started the widespread interest in Sr/Ca as a reliable tracer of SST. In this article, Beck et al. demonstrated that the use of high precision Thermal Ionisation Mass Spectrometry (TIMS) to measure Sr/Ca in coral skeleton at near monthly resolution permitted to reconstruct SST with an apparent accuracy of around 0.5 °C, sufficient to document tropical seawater annual SST variations. With this precision, and the realization that Sr/Ca could be coupled to $\delta^{18}O$ to gain information on sea surface salinity (SSS) variability (see Section 3.2), the way was paved to use corals to study the most important interannual source of climate variability on the planet: the El Niño Southern Oscillation (ENSO).

As always when a promising new tracer is discovered, the generation of long series of Sr/Ca (e.g. Linsley et al., 2000; Corrège et al., 2000, 2001; Hendy et al., 2002) was accompanied by extensive work aimed at testing the robustness of the tracer and at better understanding the mechanisms of Sr incorporation in coralline aragonite (e.g. Alibert and McCulloch, 1997; de Villiers et al., 1994, 1995; Hart and Cohen, 1996; Allison et al., 2001; Cohen et al., 2001, 2002; Ferrier-Pagès et al., 2002, Finch et al., 2003; Finch and Allison, 2003). It is now well established that Sr^{2+} is strongly bound in coralline aragonite, and probably substitutes for Ca^{2+} (Amiel et al., 1973; Mitsuguchi et al., 2001; Watanabe et al., 2001; Allison et al., 2001; Finch et al., 2003). A compilation of all the published Sr/Ca vs. SST calibration for the genus *Porites* (Table 1), however, clearly indicates that there are some discrepancies, and that no "universal" calibration exists. To illustrate this fact, we calculated the Sr/Ca corresponding to a SST of 25 °C (which is the

Table 1
Summary of all the Sr/Ca vs. SST calibrations published to date using the genus *Porites*

Reference	*Porites* species	Location	*a*	*b*	*n*	r^2	SST range (°C)	Source of SST	SST$_{9.035}$
Smith et al., 1979	*Porites* sp.	Various locations	10.94	−0.07	22	0.5	20–30	Local SST	27.22
Beck et al., 1992	*Porites lobata*	Amédée, NewCal	10.479	−0.06245			20–27	Oxygen 18	23.13
de Villiers et al., 1994	*Porites lobata*	Hawaii	10.956	−0.07952	26	0.95	23–27	Local SST	24.16
Shen et al., 1996	*Porites lutea* et *lobata*	Taiwan	10.286	−0.0514		>0.9	22–28	Local SST	24.34
Mitsuguchi et al., 1996	*Porites lutea*	Ryukyu, Japan	10.5	−0.0608	112	0.73	20–30	SST 30 km away	24.10
Alibert and McCulloch, 1997	*Porites mayeri*, *Porites lobata* et *lutea*	GBR, Australia	10.48	−0.0615	157	0.96	23–29	Local SST	23.50
Bessat, 1997	*Porites* sp.	Mururoa	11.302	−0.0815	116	0.66	23–28.5	Local SST	27.82
Heiss et al., 1997	*Porites lutea*	La Reunion	10.577	−0.061	73	0.84	23–29	COADS SST	25.28
Gagan et al., 1998	*Porites lutea*	Java and Australia	10.73	−0.0638			20–30	IGOSS	26.57
Sinclair et al., 1998	*Porites mayeri*	GBR, Australia	10.8	−0.07		0.68	23–29	Local SST	25.22
Evans et al., 1999	*Porites* sp.	Kiritimati	0.01[a]	−0.0614		0.59		IGOSS anomaly	
Fallon et al., 1999	*Porites lobata*	Shikoku, Japan	10.76	−0.063		0.59	18–28	Local SST	27.39
McCulloch et al., 1999	*Porites* sp.	Huon peninsula	10.7	−0.062			28–30	Local SST	26.86
Corrège et al., 2000	*Porites lutea*	Amédée, NewCal	10.73	−0.0657	85	0.64	20–26	Local SST	25.80
Linsley et al., 2000[b]	*Porites lutea*	Rarotonga	11.12	−0.065		0.75	22.5–28.5	IGOSS	32.08
Wei et al., 2000	*Porites lutea*	South China Sea	10.6	−0.0504	147	0.56	20–30	SST 90 km away	31.06
Marshall and McCulloch, 2001	*Porites* sp.	Xmas Isld (Indian)	10.375	−0.0593	350	0.69	24–30.5	IGOSS	22.60
Quinn and Sampson, 2002	*Porites lutea*	Amédée, NewCal	10.073	−0.052	212	0.84	20–27	Local SST (OLS)	19.97
Quinn and Sampson, 2002	*Porites lutea*	Amédée, NewCal	10.12	−0.057	212	0.84	20–27	Local SST (RMA)	19.04
Quinn and Sampson, 2002	*Porites lutea*	Amédée, NewCal	10.383	−0.061	293	0.84	20–27	GISST (OLS)	22.10
Quinn and Sampson, 2002	*Porites lutea*	Amédée, NewCal	10.524	−0.067	293	0.84	20–27	GISST (RMA)	22.23
Marshall and McCulloch, 2002	*Porites lutea*	GBR, Australia	10.4	−0.0575	112		23–29	Local SST	23.74
Marshall and McCulloch, 2002	*Porites* sp.	GBR, Australia	10.4	−0.0587	66		22.6–29.6	SST 70 km away	23.26
Fallon et al., 2003	*Porites* sp.	GBR, Australia	10.11	−0.0412		0.25	21–30.5	Local/IGOSS	26.10
Fallon et al., 2003	*Porites* sp.	GBR, Australia	10.62	−0.0652		0.69	21–30.5	Local/IGOSS	24.31
Fallon et al., 2003	*Porites* sp.	GBR, Australia	10.23	−0.052		0.62	21–30.5	Local/IGOSS	22.99
Fallon et al., 2003	*Porites* sp.	GBR, Australia	10.39	−0.0602		0.71	22.5–29.5	Local/IGOSS	22.51
Fallon et al., 2003	*Porites* sp.	GBR, Australia	10.73	−0.0707		0.83	22.5–29.5	Local/IGOSS	23.98
Fallon et al., 2003	*Porites* sp.	GBR, Australia	10.42	−0.0565		0.62	22.5–29.5	Local/IGOSS	24.52
Fallon et al., 2003	*Porites* sp.	GBR, Australia	10.35	−0.0595		0.67	25–29.5	Local/IGOSS	22.11
Allison and Finch, 2004	*Porites lobata* (COC)	Hawaii	10.965	−0.067	8	0.87	23–28	SST10 km away	28.81
Allison and Finch, 2004	*Porites lobata* (fasciculi)	Hawaii	10.861	−0.08	8	0.92	23–28	SST10 km away	22.83
Linsley et al., 2004	*Porites lutea*	Fiji	10.65	−0.053		0.77	24.5–29.5	IGOSS	30.48
Felis et al., 2004	*Porites* sp.	Eilat, Red Sea	10.781	−0.0597		0.78	21–26.5	Local SST	29.25
Zinke et al., 2004	*Porites lobata*	Madagascar	10.348	−0.05		0.89	22–28.4	GISST	26.27
Le Bec et al., submitted for publication	*Porites australiensis*	Fiji	10.18	−0.047		0.42	25–29	IGOSS	24.37
This paper	*Porites* sp.	Uitoe, NewCal	10.407	−0.0576	95	0.91	22–28	TS SST	23.82
This paper	*Porites* sp.	Uitoe, NewCal	10.19	−0.0465	117	0.85	22.5–28	IGOSS	24.85

(continued on next page)

Table 1 (*continued*)

Reference	*Porites* species	Location	*a*	*b*	*n*	r^2	SST range (°C)	Source of SST	$SST_{9.035}$
Mean			**10.553**	**−0.0607**					**25.00**
Kinsman and Holland, 1969	Inorganic aragonite		10.99	−0.045			16–96		
Dietzel et al., 2004	Inorganic aragonite		11.27	−0.043			10–50		

Calibrations are of the type: Sr/Ca (mmol/mol) = $a + b * \text{SST}$ (°C). r^2 is the correlation coefficient; *n* is the number of points used in the regression. $SST_{9.035}$ is the calculated SST for a Sr/Ca of 9.035 mmol/mol, which is the value corresponding to a SST of 25 °C in the mean *Porites* equation. Inorganic aragonite values are calculated for a seawater Sr/Ca value of 8.54 mmol/mol (de Villiers, 1999).

[a] Intercept (*a*) calculated from SST anomalies.

[b] Corrected in Linsley et al. (2004).

middle of the coral living temperature range) from the mean equation (i.e. Sr/Ca = 10.553 − 0.0607 * SST). This Sr/Ca value equal to 9.035 mmol/mol was then used to calculate SST from all the equations given in Table 1. Although the majority of reconstructed $SST_{9.035}$ fall within 2 °C of the 25 °C value, extreme values of 19 °C or 32 °C are also given by some equations. Apart from *Porites*, it should be noted that other corals have been investigated for Sr palaeothermometry. Table 2 gives an overview of the different genera used and of the published equations linking SST to Sr/Ca. Quite amazingly given the rather large variation found in the slopes and intercepts of these equations, the mean values are very similar to those found for *Porites*. However, it is important to note that the average slope for all coral species, which is close to −0.06 mmol/mol $°C^{-1}$, is different from the slopes calculated for the temperature dependence of Sr into inorganic aragonite (Kinsman and Holland, 1969; Dietzel et al., 2004; see Table 1). To fully assess the robustness of the Sr/Ca palaeothermometer, we must examine three major questions:

1) What are the pathways for the incorporation of Sr^{2+} in coralline aragonite
2) What is the relative (if any) importance of factors other than temperature
3) How are calibrations constructed.

We already saw that the question of the mode of Sr incorporation in coralline aragonite is a controversial one. Some authors suggest that Sr^{2+} follows a passive pathway (Kinsman, 1969; Ip and Krishnaveni, 1991; Ip and Lim, 1991), whereas others propose that Sr^{2+} and Ca^{2+} follow a similar, active transport pathway (Goreau, 1977; Chalker, 1981; Wright and Marshall, 1991; Ferrier-Pagès et al., 2002; Reynaud et al., 2004). Building on Sr/Ca analyses obtained at the micrometric level in various parts of the coral skeleton (Cohen et al., 2001, 2002), Cohen and McConnaughey (2003) proposed a theory in which passive transport of Sr^{2+} dominates at nighttime, whereas active transport takes place during daytime. Thus, during nighttime, it is hypothesised that centers of calcification (COC) are accreted from a fluid with a Sr/Ca similar to seawater, and that the chemistry of the nighttime skeleton is almost the same as inorganic aragonite. In contrast, sclerodermites deposited during the day will be strongly affected by kinetic processes linked to the carrier-mediated transport of Sr^{2+} (Ferrier-Pagès et al., 2002; Al-

Table 2
Sr/Ca vs. SST calibrations published for corals other than *Porites*

Reference	Coral species	Location	*a*	*b*	*n*	r^2	SST range (°C)	Source of SST
Smith et al., 1979	*Pocillopora damicornis*	Hawaii	11.01	−0.071	32	0.59	18.5–30	Local SST
Smith et al., 1979	*Montipora verrucosa*	Hawaii	11.64	−0.089	18	0.4	18.5–30	Local SST
de Villiers et al., 1994	*Pocillopora eydouxi*	Hawaii	11.004	−0.0763	14	0.93	24–27	Local SST
de Villiers et al., 1994	*Pavona clavus*	Galapagos	10.646	−0.0675	29	0.68	20–25	Local SST
Cardinal et al., 2001	*Diploria labyrinthiformis*	Bermuda	10.03	−0.045	50	0.86	19–28.5	COADS
Swart et al., 2002	*Montastraea annularis*	Florida	9.994	−0.0377	98	0.88	22–31	Local SST
Corrège et al., 2004	*Diploastrea heliopora*	Indonesia/New Cal	10.57	−0.06	25	0.75	22–30	Local and GISST
Yu et al., 2004	*Goniopora* sp.	South China Sea	9.6	−0.0305	48	0.99	18–30	??
Mean for all corals			**10.562**	**−0.0596**				

See Table 1 for explanation.

Horani et al., 2003). Although appealing, this theory still needs to be refined in view of recent work (Allison and Finch, 2004) that shows the Sr/Ca ratio in COC are much more variable than previously measured by Cohen et al. (2001). Recent results also show that at constant temperature, nighttime uptake of Sr^{2+} varies as a function of the amount of light received during the day (Reynaud et al., 2004), possibly in association with nutritional status (Houlbrèque et al., 2003, 2004).

More work with micro-probes is also needed to document further the large pseudo-monthly oscillations in Sr/Ca seen by Meibom et al. (2003) in a *Porites* specimen from Kenya, but which appear minimal (Cohen and Sohn, 2004) or absent (Allison and Finch, 2004) in Pacific corals. The work of Allison and Finch (2004), however, is promising in that it shows that the large small-scale heterogeneity of Sr/Ca, whatever its cause is, can be potentially smoothed out in favour of the SST factor when the sampling resolution is decreased, providing an explanation for the good calibration curves of Sr/Ca vs. SST produced over the years (Allison and Finch, 2004).

The question remains, however, of what are the other important parameters potentially influencing the Sr/Ca ratio of aragonitic coral skeletons. Growth rate (or calcification rate, or extension rate) has been the preferred candidate, stirring a lot of controversy in the scientific community. First, it should be pointed out that there is often a confusion between extension rate (i.e. the linear growth rate of a coral colony in one direction) and calcification rate (i.e. the total amount of skeleton deposited per unit area per unit time). Although these two parameters are often correlated in *Porites* corals (Lough and Barnes, 2000), calcification rate is generally considered a more relevant parameter in the study of geochemical proxies (Mitsuguchi et al., 2003). Looking either at extension or growth rates, a number of studies have concluded that these parameters do not affect the Sr/Ca ratio of coral skeleton, providing sampling is carried out along the main growth axis of corallite fans (Shen et al., 1996; Alibert and McCulloch, 1997; Gagan et al., 1998; Wei et al., 2000; Swart et al., 2002, Mitsuguchi et al., 2003; Corrège et al., 2004; Allison and Finch, 2004). In contrast, a few studies (de Villiers et al., 1994, 1995; Cohen and Hart, 2004) have found a growth rate effect on Sr/Ca ratios. A recent study on the effect of extension rate on $\delta^{18}O$ showed that this was strong for *Porites* with extension rates lower than 0.6 cm $year^{-1}$, but negligible above that value (Felis et al., 2003). From a theoretical point of view, growth rate is most probably affected by physiological processes (Weiner and Dove, 2003) involving algal symbionts (Weber, 1973; McConnaughey, 1989; Cohen et al., 2002; Cohen and McConnaughey, 2003; Rosenfeld et al., 2003) or feeding regimes (Reynaud et al., 2002; Houlbrèque et al., 2003, 2004). It would appear therefore that stable isotopes, which follow a more complex pathway than Sr^{2+} during the calcification process (Cohen and McConnaughey, 2003; Allemand et al., 2004), should be in theory more affected by variation in growth rate than Sr/Ca, as reflected in the existing literature. A recent study (Reynaud et al., 2004) demonstrates that Sr^{2+} uptake is strongly influenced by light level, temperature and growth rates. However, more work is needed is this area before the role of calcification rate on Sr/Ca ratios can be fully assessed (Ferrier-Pagès et al., 2002; Reynaud et al., 2004).

The concentration of both Sr^{2+} and Ca^{2+} in seawater also affects the Sr/Ca measured in corals. It is in fact a prerequisite when using Sr/Ca for palaeothermometry to assume that the Sr/Ca ratio of seawater is constant over space and time. It has been shown that this is probably not the case, and that seawater Sr/Ca ratio can vary slightly in the modern ocean (de Villiers et al., 1994; Shen et al., 1996). Over glacial–interglacial cycles, the erosion of exposed carbonate shelves can also potentially alter the seawater Sr/Ca (Stoll and Schrag, 1998). The effect of the variability of seawater Sr/Ca in space and time on reconstructed SST could be 0.2 °C to 2 °C. However, these estimates should be re-evaluated because they do not take into account the fact that Sr^{2+} incorporation in corals depends on the absolute Ca^{2+} concentration in seawater (Ferrier-Pagès et al., 2002).

The final question regarding the robustness of the Sr/Ca palaeothermometer is the fact that, as shown in Table 1, there is a large array of published calibration equations linking Sr/Ca to SST. It can be argued that part of this discrepancy is due to the lack of a common standard for the precise measure of Sr/Ca between different laboratories (equivalent to NBS-19 for stable isotopes standardization), or to the use of different analytical methods (TIMS, ICP-MS, ICP-AES, etc.). However, as pointed out by others (e.g. Marshall and McCulloch, 2002; Quinn and Sampson, 2002; Solow and Huppert, 2004a), the way calibrations are constructed is highly variable and could account for most of the discrepancy. One crucial point is the choice of the SST database. Table 1 indicates that some calibrations are based on SST measured in situ, while others rely on SST measured in locations sometimes distant by 90 km. Finally, some calibrations are based on larger-scale SST data sets, the most commonly used being the mostly

Table 3
Correlation (R^2) amongst the various SST data sets from New Caledonia, and between Sr/Ca and each SST data set

	TS SST	IGOSS SST	Amédée SST	Sr/Ca
TS SST		0.88	0.79	0.92
IGOSS SST	0.88		0.88	0.88
Amédée SST	0.79	0.88		0.74

See Figs. 2 and 4 and text for details.

satellite-based 1°×1° IGOSS data set (Reynolds and Smith, 1994). To assess the potential difference introduced in using various SST data sets, a *Porites* coral growing near a thermosalinograph (TS) was sampled, offshore Noumea, New Caledonia. The coral grows on the ocean side of the barrier reef. It was collected in February 2000, at 10 m water depth. It was sampled and analyzed following standard procedures described elsewhere (Le Cornec and Corrège, 1997; Corrège et al., 2000), over a 6-year period running from May 1993 to May 1999. Three SST data sets were used in the comparison: the in situ TS data set, a data set from Amédée Island (located 45 km south of the coral, but inside the lagoon), and the IGOSS data set (Reynolds and Smith, 1994). Notable differences are present amongst the various SST data sets, the IGOSS SST being the warmest and the Amédée SST the coolest (Fig. 2). The correlation between the various data sets is however acceptable (Table 3). The Sr/Ca data were fitted to the in situ TS SST using the Analyseries program (Paillard et al., 1996), and resampled at monthly intervals. Fig. 3 shows the excellent fit obtained between the Sr/Ca and TS SST data. Fig. 4 displays the three regressions calculated between Sr/Ca and the SST data sets. The slope of the regression varies from 0.0391 to 0.0549 mmol/mol °C^{-1}, clearly indicating that the range of slopes found in the literature and displayed in Table 1 is strongly influenced by the choice of the SST data set. This comparative exercise highlights several key points to consider when attempting coral-based SST reconstruction. First, it is always advisable to have local SST, because as shown in Fig. 2, close stations, especially when they are located in different settings (inside vs. outside the lagoon in our case) yield significantly different calibrations that are not as reliable as those based on in situ or satellite data sets. Second, it is important to check that the local SST is well correlated with the larger-scale satellite-based SST, because the usefulness of a coral record lies in its ability to represent regional climate variability. Thus, when local SST are available and meet this criteria, it is advisable to always provide the calibration based on the satellite-based SST, to enable a more thorough assessment of the natural variability of the Sr/Ca vs. SST regression. When possible, a normalization to a fixed

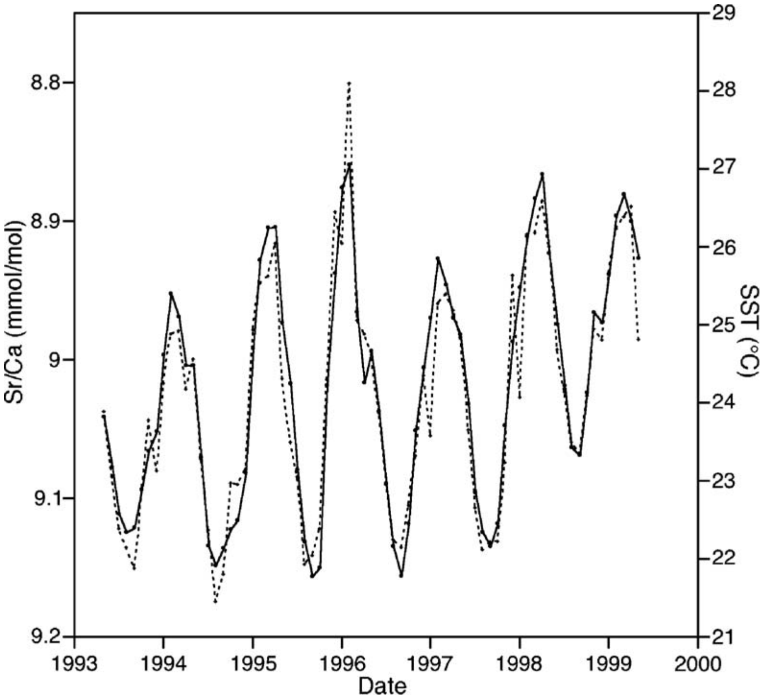

Fig. 3. Plot of Sr/Ca (plain line) from a *Porites* coral growing near the thermo-salinograph (TS) located at Uitoe, New Caledonia, compared to the TS SST record (dotted line).

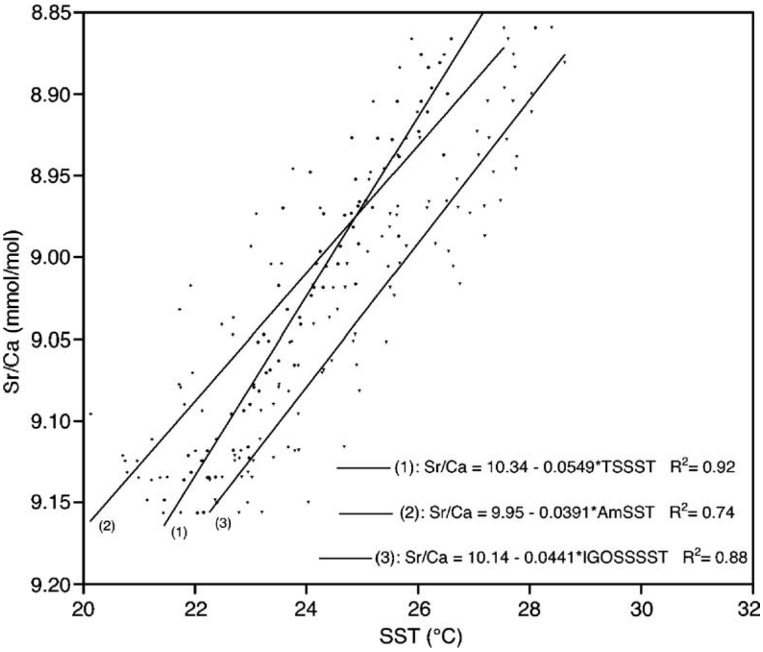

Fig. 4. Regression between Sr/Ca and the three SST data sets from New Caledonia. TSSST: Thermo-salinograph; AmSST: Amédée; IGOSSSST: Blended satellite. See text for details.

value of seawater Sr/Ca (e.g. 8.54 mmol/mol; de Villiers, 1999) is also needed.

Finally, it should be noted that there is an ongoing debate as to which method should be used to regress Sr/Ca against SST. Most studies used Ordinary least Square regression (OLS), but some authors recommend using Reduced Major Axis regression (Shen and Dunbar, 1995; Quinn and Sampson, 2002), although this approach has been criticized lately (Solow and Huppert, 2004a). Whatever regression method is used, it should always be quoted to avoid confusion.

3.1.2. $\delta^{18}O$

δ^{18}O has been widely used as a tracers of SST, despite the fact that variability in seawater δ^{18}O is much larger than that of Sr/Ca, and can affect SST reconstructions. Pioneering work on the use of δ^{18}O as a palaeothermometer was published by Epstein et al. (1951, 1953), and its application to corals was the seminal work by Weber and Woodhead (1972). These authors analyzed hundreds of corals belonging to 44 genera from 29 locations worldwide. They confirmed that corals form their skeleton out of isotopic equilibrium with ambient seawater, and provided equations linking δ^{18}O and SST for the genera studied. Following pioneering sub-annual work by Goreau (1977) and Emiliani et al. (1978), Fairbanks and Dodge were the first in 1979 to show that coral δ^{18}O could be used to resolve monthly SST variability. A decade later, McConnaughey (1989) proposed a kinetic model to account for the isotopic disequilibrium in biogenic carbonates. In short, the kinetic model states that isotopic fractionation occurs during the hydration and hydroxylation of CO_2. Consequently, CO_3^{2-} will more or less equilibrate isotopically with the water present in the calcifying fluid (which has the same isotopic composition as seawater) before it is incorporated in the newly formed skeleton. The quicker the calcification rate, the larger the isotopic disequilibrium with seawater will remain. Details on this model, and on alternative views (e.g. Spero et al., 1997; Adkins et al., 2003; Rollion-Bard et al., 2003a,b) can be found in McConnaughey (2003), Swart and Grottoli (2003), Cohen and McConnaughey (2003) and Allemand et al. (2004).

Essentially, the discussion presented in the previous chapter on the reliability of Sr/Ca as a palaeothermometer could be repeated for δ^{18}O. The major difference between the two tracers, however, is the way they are incorporated in the coral skeleton. Although both mechanisms are still not fully understood (Allemand et al., 2004), it is clear that transport of oxygen is far more complex and involves different chemical forms from CO_2 to CO_3^{2-}. These forms are present in seawater, but their isotopic composition can be modified by respiration from the zooxanthellae and during their transport to calcification sites (McConnaughey, 1989; Cohen and

McConnaughey, 2003; Allemand et al., 2004). However, most of the work published so far on fractionation of stable isotopes during skeletogenesis deals mainly with δ^{13}C rather than δ^{18}O, despite the fact that δ^{18}O is a more commonly used tracer. This is unfortunate because micrometric scale analyses of both isotopes indicate that δ^{18}O variability is much higher than that of δ^{13}C (Rollion-Bard et al., 2003a,b), possibly suggesting different fractionation mechanisms.

Apart from temperature (and assuming a constant seawater δ^{18}O value), the other factors potentially affecting δ^{18}O in coral skeleton that have been studied are growth rates, variation in pH near the calcification sites, light levels, productivity and feeding habits, with the former generating the most literature.

The effect of growth rate, or of the more easily measured extension rate on δ^{18}O has been assessed to be either significant (e.g. Land et al., 1975; McConnaughey, 1989; de Villiers et al., 1995; Allison et al., 1996; Rosenfeld et al., 2003; Felis et al., 2003; Maier et al., 2004) or inconclusive (Leder et al., 1996; Cohen and Hart, 1997; Gagan et al., 1998; Guilderson and Schrag, 1999; Watanabe et al., 2003). Some authors found that this effect only becomes significant when corals grow under a threshold extension rate (McConnaughey, 1989; Felis et al., 2003; Rosenfeld et al., 2003). Resolving these discrepancies will require a better understanding of calcification processes, and of the exact relationship between calcification rate and extension rate. It seems however, that correcting δ^{18}O records from the variability in extension rate can already produce time series that reflect more faithfully environmental parameters (Maier et al., 2004).

Recent advances in analytical techniques (i.e. Secondary Ion Mass Spectrometry) now allow precise measurements of stable isotopes at micrometric scales (Rollion-Bard et al., 2003a,b). Analyses on *Porites* samples indicate that, similar to Sr/Ca ratios, δ^{18}O variability is high at micrometric levels and cannot be explained by temperature fractionation. Coupled boron analyses, which are believed to be a proxy for pH, indicate that rapid variations in pH could control the kinetics of the isotopic equilibration of the various carbonate species (Rollion-Bard et al., 2003a,b). This controversial work opens a new sector of investigations which should undoubtedly be developed in the future.

Finally, there is also controversy in the possible effect of light level, feeding habits and metabolic activity on δ^{18}O. Fairbanks and Dodge (1979) found no change in δ^{18}O with depth in a thermally homogeneous environment, and concluded that δ^{18}O was not affected by light level. Similar conclusions were reached by Weil et al. (1981) from an aquarium study. On the contrary, another aquarium study (Reynaud-Vaganay et al., 2001) found that the average δ^{18}O of two coral genera was more depleted under low light than under high light conditions, consistent with other studies (Weber and Woodhead, 1970; Erez, 1978; Grottoli and Wellington, 1999). Because light induces photosynthesis, and photosynthesis is believed to stimulate calcification, the findings of Reynaud-Vaganay et al. (2001) are at odds with McConnaughey's (1989) kinetic model which predicts that δ^{18}O depleted coral skeleton should be associated with high calcification rates (and therefore high light levels). In another aquarium experiment, the Monaco team (Reynaud et al., 2002) also showed that fed coral colonies had lower δ^{18}O than starved colonies, potentially affecting the δ^{18}O thermometer (i.e. fed corals would give higher SST than starved corals). On the other hand, Grottoli and Wellington (1999) found no difference in δ^{18}O between fed and starved colonies, a difference that Reynaud et al. (2002) attribute to a possible indirect effect of calcification rate. Finally, Juillet-Leclerc et al. (1997) found no relationship between δ^{18}O and productivity levels, concluding that δ^{18}O is not controlled by metabolic activity.

The complexity of the various factors discussed above, and the fact that the δ^{18}O value of seawater is not always properly known may explain the variation in calibration of coral δ^{18}O vs. SST published to date. Usually, the value of the slope varies between 0.18‰ $^{\circ}\mathrm{C}^{-1}$ and 0.22‰ $^{\circ}\mathrm{C}^{-1}$. The general conclusion of the review presented above is that, like Sr/Ca, changes in δ^{18}O are most probably largely driven by SST variations. However, many other factors can influence δ^{18}O in coral skeleton, and before we understand their relative importance in the final measured δ^{18}O, we should bear in mind the potential error they can introduce in reconstructed SST.

3.1.3. Other SST tracers

Apart from Sr/Ca and δ^{18}O, a few other SST tracers have been investigated in corals. The most popular tracers in that sense are uranium (U/Ca) and magnesium (Mg/Ca), and to a lesser extent boron (B/Ca).

Livingston and Thompson (1971) stated that SST had no effect on the incorporation of uranium in coral skeleton. It took almost 25 years, and some analytical improvements to realize that U/Ca ratios could indeed potentially give information on past SST (Min et al., 1995; Shen and Dunbar, 1995). However, despite a sensitivity to SST about 5-fold that of Sr/Ca, it was suggested that other environmental parameters could influence the incorporation of uranium in coral aragonite (Min et al., 1995; Shen and Dunbar, 1995). This is mainly due to the complex chemical behaviour of uranium in seawater and to its relatively unknown mode of incorporation in aragonite (Min et al., 1995; Pingitore et al., 2002, Lazar et al., 2004). Despite some successes in correlating SST derived from U/Ca and Sr/Ca (e.g. Corrège et al., 2000), there is still a large spread in the published U/Ca vs. SST regressions (Min et al., 1995; Sinclair et al., 1998; Fallon et al., 1999, 2003; Corrège et al., 2000; Wei et al., 2000; Quinn and Sampson, 2002).

The potential sensitivity of Mg/Ca to SST was first described by Chave (1954). Subsequent work by Amiel et al. (1973), Weber (1974), Goreau (1977), Swart (1981) and Oomori et al. (1982) reached diverse conclusions concerning the parameters ruling the incorporation of Mg in coral aragonite. Mitsuguchi et al. (1996) published a study that showed an extremely good correlation between Mg/Ca and SST. However, subsequent work and calibrations from other locations supported the contention made by Amiel et al. (1973) that 20% to 30% (up to 40% according to Watanabe et al., 2001) of the magnesium is located in exchangeable sites, or associated with the organic component of the skeleton. The reliability of the Mg/Ca thermometer is therefore greatly questionable (Sinclair et al., 1998; Fallon et al., 1999, 2003; Wei et al., 1999, 2000; Mitsuguchi et al., 2001; Watanabe et al., 2001; Quinn and Sampson, 2002; Marshall, 2002; Mitsuguchi et al., 2003).

Boron as a palaeothermometer was first investigated by Hart and Cohen (1996). Then, the ANU group (Sinclair et al., 1998; Fallon et al., 1999, 2003) investigated further this tracer by laser ablation ICP-MS. The current consensus is that there are variable calibrations depending on the sites studied, and that more work is required to test the full potential of B/Ca to reconstruct past SST.

3.1.4. Decadal variability of tropical SST in the past 300 years

Both δ^{18}O and, to a lesser extent, Sr/Ca have been analyzed in massive corals to produce long series of past SST. Most work has concentrated on the past three centuries, but significant results have also been published from fossil corals ranging in age from the mid-Holocene (e.g. Gagan et al., 1998; Corrège et al., 2000; McGregor and Gagan, 2004) to the last interglacial (Hughen et al., 1999) or even to 350 ka (Kilbourne et al., 2004a). In those studies, emphasis is generally put on the presence or absence of ENSO, and when present, on its strength and frequency. Coral studies are in particular important to test hypothesis linking ENSO strength to orbitally driven insolation changes (Clement et al., 1999). Compared to proxies such as pollens or planktonic organisms retrieved in continental or marine sediments, corals, with their higher time resolution (i.e. typically monthly), can provide more reliable information on El Niño events that usually last for a few months only. For example, earlier claims that ENSO started to operate in the mid-Holocene (Shulmeister and Lees, 1995; Sandweiss et al., 1996) were later proved wrong, in particular by coral studies that showed that ENSO was weaker but still active in the early Holocene (Tudhope et al., 2001; Gagan et al., 2004; McGregor and Gagan, 2004).

One of the great successes of corals has also been to provide physical oceanographers and modellers with a longer-term perspective on ENSO. Instrumental records are too short to thoroughly address the problem of the low frequency (i.e. decadal to multidecadal) modulation of ENSO. The generation of multi-century coral records is a first step forward in that direction. Fig. 5 shows the longest, best dated and continuous δ^{18}O and Sr/Ca coral records taken as SST proxies, together with a tree-ring based reconstruction of the Pacific Decadal Oscillation (PDO; D'Arrigo et al., 2001). The chosen records are from Mooréa (Boiseau et al., 1998), Rarotonga (Linsley et al., 2000), Santo (Quinn et al., 1993), New Caledonia (Quinn et al., 1998), Maiana (Urban et al., 2000), Abrohlos (Kuhnert et al., 1999), Seychelles (Charles et al., 1997), Malindi (Cole et al., 2000), Aqaba (Heiss, 1994), and Ras Umm Sidd (Felis et al., 2000). To highlight the decadal variability, each record was detrended, normalized to unit variance and filtered with a 7-year Hanning filter (to eliminate ENSO and higher frequencies). Except for Rarotonga which is a Sr/Ca record (and therefore considered to be a better proxy for SST), all the other coral series are δ^{18}O records. They could therefore be also affected by changes in seawater δ^{18}O. Nevertheless, in most of the cases presented here, SST is the prime factor ruling the changes in coral δ^{18}O. The 10 coral records presented in Fig. 5 are plotted from east (top) to west (bottom) starting in the central Pacific. The background colours emphasize the southern hemisphere (where 8 of the 10 records are from) warming trend which started circa 1910 (Jones et al., 2001). Correlation coefficients between the Rarotonga record (taken as a reference) and each series was calculated prior to and after 1910.

From Fig. 5, it is obvious that all the records contain a significant decadal to multi-decadal component. In the Pacific basin, this low frequency variability seems complex in its spatial pattern, as shown by the various

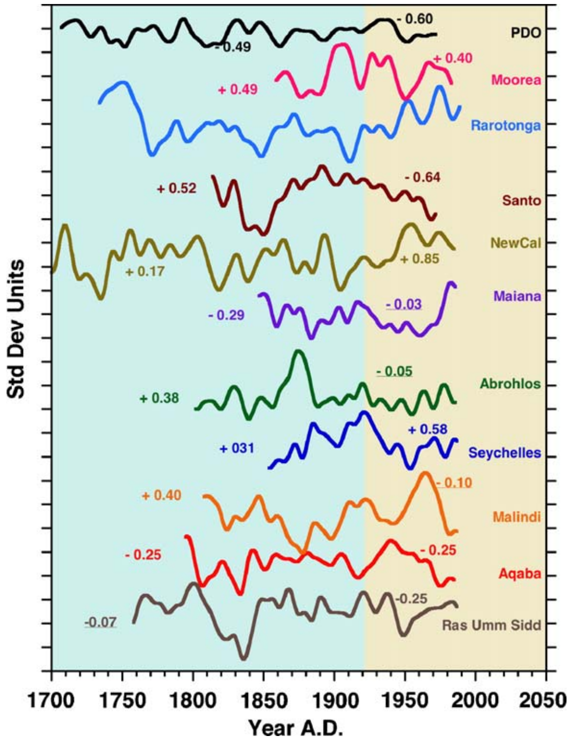

Fig. 5. Normalized and filtered proxy-SST time series extracted from corals compared with a proxy index of the Pacific Decadal Oscillation. The background colours indicate the southern hemisphere warming trend which started circa 1910. For each series, the correlation coefficient with the Rarotonga SST record is shown before and after 1910. All coefficients are significant at the 99% level except those underlined. Site location can be found in Fig. 1. See text for further details. Most series used here are available on the web from: http://www.ngdc.noaa.gov/paleo/.

correlations. Two sites relatively close by like Santo (Vanuatu) and New Caledonia behave differently, especially since 1910. It is also interesting to note that there is an overall good correlation between the central Pacific and the Indian Ocean and Red Sea. Before 1910, the correlation between Rarotonga and the Indian Ocean seemed to be consistent, whereas after 1910, only the Seychelles keep a good correlation with Rarotonga. This could suggest that global warming is affecting the teleconnections between the oceans.

One of the key questions being debated today concerns the tropical or extra-tropical origin of the low frequency modulation of ENSO. The correlation between central Pacific SST and the PDO (which is defined in the north Pacific) is good (Fig. 5), and suggests a strong link between the low and mid to high latitudes. Work has started in France and the USA to assimilate coral data into models and to test the various scenarios put forward to explain decadal climate variability. Parallel to that effort, it seems necessary to produce additional long coral series, particularly in the tropical Atlantic Ocean.

3.2. Sea surface salinity

Like SST, sea surface salinity (SSS) is an extremely important physical parameter in climate studies. Together with temperature, salinity is the driving force of the thermohaline circulation. Salinity also plays a crucial role in the generation of ENSO events (Delcroix et al., 1998; Maes et al., 2002). And because changes in SSS can often be related to changes in precipitation/evaporation regimes, a good tracer of salinity has the potential to give information on atmospheric components of climate.

Despite some claims that U/Ca ratios could be influenced by SSS variations (Shen and Dunbar, 1995) and therefore provide a potential SSS tracer, most of the work to date has focused on δ^{18}O, either as a sole tracer, or coupled with Sr/Ca.

3.2.1. The ^{18}O/^{16}O cycle and its relationship with salinity

Within the hydrological cycle, the ^{18}O/^{16}O ratio is affected by fractionation processes. Evaporation tends to discriminate against the heavier ^{18}O isotope, whereas the reverse happens during precipitation. Salinity changes in surface waters are also largely related to the balance of precipitation and evaporation, thus providing the basis for a strong relationship between SSS and seawater δ^{18}O (hereafter $\delta^{18}O_{sw}$). This relationship has been studied for 40 years (Craig and Gordon, 1965; Ostlund et al., 1987; Bigg and Rohling, 2000), but there are still some significant gaps in the spatial coverage, as shown in the NASA/GISS Global Seawater Oxygen Isotope Database at: http://www.giss.nasa.gov/data/o18data (Schmidt et al., 1999). It has been shown that there is not one $\delta^{18}O_{sw}$–SSS relationship in the oceans, but rather multiple relationships depending on the latitude and the ocean basin considered (e.g. Bigg and Rohling, 2000). Even in the tropics, the $\delta^{18}O_{sw}$–SSS relationship can vary because the isotopic composition of rainfall and river water is not constant (Benway and Mix, 2004). The published slope of the regression between $\delta^{18}O_{sw}$ and SSS for the tropical Pacific Ocean varies between 0.27‰ psu^{-1} and 0.42‰ psu^{-1} (Fairbanks et al., 1997; Morimoto et al., 2002). Work in progress shows that there is a possible seasonal variability in this slope as well (Corrège and Stievenard, unpublished data). All in all, the variability in the $\delta^{18}O_{sw}$–SSS relationship will introduce an error when inferring past salinity from $\delta^{18}O_{sw}$. This error has been estimated to lie between 0.1 and 2 psu (Rohling and Bigg, 1998; Benway and Mix, 2004).

3.2.2. Reconstruction of SSS from coral $\delta^{18}O$

As a first approximation, $\delta^{18}O$ in coral aragonite is dependent upon SST and $\delta^{18}O_{sw}$. Therefore, it is in theory possible to assess past changes in SSS from coral $\delta^{18}O$ providing the local $\delta^{18}O_{sw}$–SSS relationship is known, and the SST component of the $\delta^{18}O$ signal is either known or negligible.

When Beck et al. (1992) published their work on the Sr/Ca thermometer, they stressed the fact that the coupling of Sr/Ca and $\delta^{18}O$ measurements had the potential to yield a reliable palaeo-salinometer. This fact was successfully demonstrated in several papers by the ANU group (McCulloch et al., 1994; Gagan et al., 1998; Hendy et al., 2002; see also Gagan et al., 2000 for a review). By subtracting the SST component given by Sr/Ca from the $\delta^{18}O$ signal, these authors were able to document major changes in surface hydrology (and therefore salinity, although they did not give absolute SSS values). Ren et al. (2003) proposed a different approach, but upon close examination, it appears to be essentially the same as the ANU one (Huppert and Solow, 2004; Kilbourne et al., 2004b). These methods have now been used by different groups around the world to reconstruct past time series of $\delta^{18}O_{sw}$ (Corrège et al., 2004; Kilbourne et al., 2004b; Linsley et al., 2004).

In areas where SST variability is small the $\delta^{18}O$ signal has been used directly to infer past changes in $\delta^{18}O_{sw}$ (Cole and Fairbanks, 1990; Linsley et al., 1994; Tudhope et al., 1995). However, there is always a small seasonal SST component that distorts the reconstructed $\delta^{18}O_{sw}$. To overcome this problem, a different approach was pioneered by Le Bec et al. (2000). It relies on the fact that on some timescales, the SST component might become negligible. This situation occurs in Fiji (Le Bec et al., 2000), where SST variability is fairly constant from year to year whereas interannual variability of SSS is large. Therefore, by filtering the $\delta^{18}O$ signal with a 25-month window, the SST is virtually eliminated and the residual $\delta^{18}O$ is strongly correlated to filtered SSS variations ($r=0.71$; see Fig. 6). Although limited in its application, this method can give better results than the coupled Sr/Ca–$\delta^{18}O$ method in specific areas (Kilbourne et al., 2004b).

It is clear at this stage that a better understanding of the $\delta^{18}O_{sw}$–SSS relationship and of its potential variability through time is needed before precise palaeo-SSS estimates can be extracted from corals.

4. Future directions and conclusions

There is little doubt that corals are formidable archives of past oceanic conditions. Their aragonitic skeleton contains an array of tracers, and our task is to decipher them and to understand exactly how they are incorporated. In this review, information is pro-

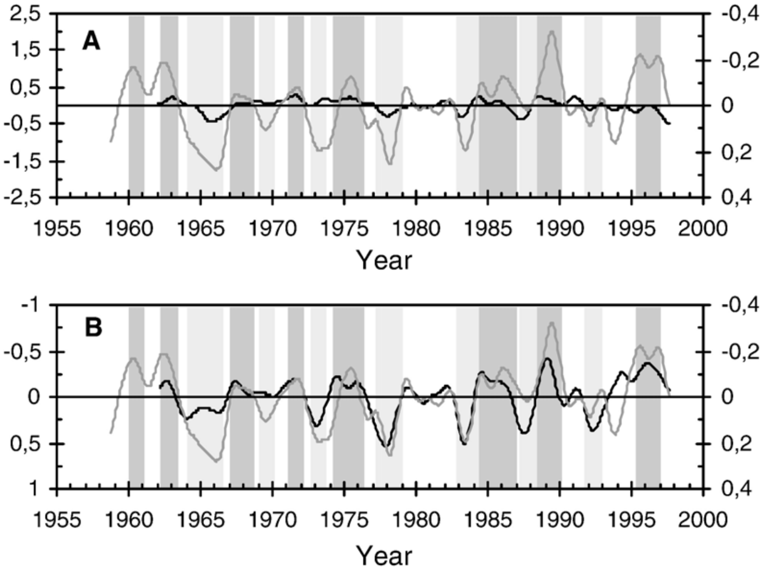

Fig. 6. 25-month Hanning filtered time series of coral $\delta^{18}O$ (light line), instrumental SST (A) and SSS (B) (dark lines) from Fiji. Data are normalized to the 1960–1997 period. The light shaded bars represent El Niño events and the dark shaded ones, La Niña events. SST and SSS are scaled to express their respective effects on $\delta^{18}O$ (modified from Le Bec et al., 2000).

vided pertaining to the reconstruction of the two main oceanic parameters: temperature and salinity. However, other tracers in corals such as barium, cadmium and ^{14}C have been used to infer past changes in upwelling activity and oceanic circulation (Shen et al., 1987; Lea et al., 1989; Guilderson and Schrag, 1998). Barium can also be a useful tracer of river discharge (McCulloch et al., 2003). Studies have also been published on manganese (Shen et al., 1991, 1992), on the isotopic composition of elements such as boron (Gaillardet and Allègre, 1995) and osmium (Levasseur et al., 1999), and on the most controversial tracer of all: $\delta^{13}C$ (see for example Swart et al., 1996; Heikoop et al., 2000). Similarly to SST and SSS proxies, little is known about the pathways of incorporation of these elements in the coral skeleton, and how strongly they are bound to the aragonitic crystalline structure. This is clearly an area of research where more effort should be expended, especially now that controlled aquarium growth of corals is done routinely (Reynaud-Vaganay et al., 1999), and that microprobes allow chemical investigations of trace metals and isotopes at micrometric levels (Hart and Cohen, 1996; Allison et al., 2001; Rollion-Bard et al., 2003a,b).

Investigations should also focus on biochemical tracers. Corals contain about 1% of organic matter in their skeleton (Cohen and McConnaughey, 2003; Cuif et al., 2004), and a better characterization of it would undoubtedly reveal potential new tracers of past environmental conditions. Likewise, a chemical footprint of fluorescent bands that are sometimes present in corals (Isdale et al., 1998; Barnes et al., 2003) could yield further insight into their origin.

Downstream of the work on the tracers themselves, there is a need to improve our use of the various tracers, bearing in mind their current imperfections. Some recommendations in that sense are given by Lough (2004) in a recent paper. Her main concern is the need for a level of replication of coral records, a feat difficult to achieve with fossil corals. In such corals, which are often unique (e.g. Corrège et al., 2004), it seems more realistic to use multiple tracers. Even though it has been proven that the use of multiple tracers does not necessarily improve the precision of reconstructed parameters (Solow and Huppert, 2004b), it undeniably improves the overall confidence in the reconstruction, because tracers usually have a different response to diagenesis (Müller et al., 2001; McGregor and Gagan, 2003). However, it must be pointed out that in some locations, like Palmyra island in the central tropical Pacific, it has been possible to recover, to precisely date and to overlap multiple fossil corals from the same time interval of the past millennium (Cobb et al., 2003a,b). It is therefore suggested that sites similar to Palmyra should be sought to collect multiple coral records that could be spliced to eventually provide reliable and continuous records for the past centuries.

Finally, there should be a concerted effort to link geochemists generating palaeorecords from corals with modellers. The full potential of coral records will only be fulfilled when they are assimilated in climate models. Corals have the ability to complement back in time the instrumental record, and to improve our understanding, and ultimately prediction of phenomena such as El Niño Southern Oscillation or the Pacific Decadal Oscillation. The assimilation of proxy data in models is by no mean a simple feat, but work is in progress in various groups and its outcome is eagerly awaited by the climate scientific community.

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