Palaeoclimatology Research Letters, Vol. 12, No. 2, 2019
DOI:10.1029/2019PA004404

Ocean modelling improves the use of coral-derived seawater δ^{18}O as a hydrological proxy in the tropical Pacific

Isma Abdelkader Di Carlo[1,2]*

Abstract
Oxygen isotopic ratios can be measured in corals in order to reconstruct passed climates. The measured coral δ^{18}O depends not only on the temperature in which the coral grew but also the δ^{18}O value of the surrounding seawater. The latter is not well constrained and has been used widely as a salinity proxy. Its relationship with salinity and its contribution to coral δ^{18}O variance has not been properly assessed. Here I investigate how seawater δ^{18}O variability can be explored with the isotope-enabled Regional Ocean Modeling System (isoROMS). Spatiotemporal variations of the contribution of seawater δ^{18}O to coral δ^{18}O variance and the salinity-$\delta^{18}O_{sw}$ relationship play a huge role and need to be taken into account when reconstructing the $\delta^{18}O_{sw}$ signal using coral records.

Keywords
Seawater δ^{18}O — Ocean model — Tropical Pacific

[1] *European Institute for Marine Studies, Brest, France*
[2] *Department of Geosciences, Ecole Normale Supérieure, Paris, France*

Introduction

The tropical Pacific Ocean is a highly dynamic region : it is the center of action of El Niño Southern Oscillation (ENSO), the dominant mode of tropical climate variability [Vecchi and Wittenberg, 2010]. This phenomenon changes temperature and weather patterns via atmospheric teleconnections, thus affecting ecosystems and human societies around the world [Collins et al., 2010]. Studies have used atmosphere-ocean coupled General Circulation Model (GCM) scenarios to assess ENSO's behavior under greenhouse warming through radiative forcing but as of yet, there is little consensus. Consequently, to better grasp the response of the tropical Pacific Ocean to anthropogenic climate change, long observational time periods are needed. Because of sparse historical data before 1980, long-term paleoclimatic records must be used.

The stable oxygen isotopic composition of tropical fast-growing shallow coral aragonite ($\delta^{18}O_{coral}$) has been thoroughly used to infer past sea surface temperature (SST) variability associated with ENSO, e.g. over the Holocene [Cobb et al., 2003] and Last Glacial Maximum [Tudhope et al., 2001]. Although coral $\delta^{18}O_{coral}$ reflects changes in SST, it also depends on the stable oxygen isotopic composition of the surrounding water ($\delta^{18}O_{sw}$). Indeed, $\delta^{18}O_{sw}$ is fractionated during calcification via a temperature-dependent process [Epstein et al., 1953] and the resulting $\delta^{18}O_{coral}$ is recorded in the coral aragonitic skeleton [Weber and Woodhead, 1970], albeit with some species-dependent vital effects [McConnaughey, 1989a, McConnaughey, 1989b]. This dual control on $\delta^{18}O_{coral}$ varies across locations and makes it difficult to interpret coral records. To better constrain the relative contributions from SST and $\delta^{18}O_{sw}$, previous studies have explored forward modelling [Evans et al., 2013], which converts SST and sea surface salinity (SSS) into synthetic corals, i.e. pseudo-corals ($\delta^{18}O_{pseudo-coral}$) [Thompson et al., 2011]. The temperature component is straightforward and uses the known temperature-dependent fractionation during coral growth whilst the salinity component assumes that $\delta^{18}O_{sw}$ correlates linearly with SSS.

Although SSS and $\delta^{18}O_{sw}$ are similarly influenced by freshwater fluxes, $\delta^{18}O_{sw}$ also depends on the δ^{18}O value of moisture sources in the atmosphere. Diffusion, horizontal advection and vertical entrainment of water masses may play a big role too [Stevenson et al., 2015]. Due to sparse continuous paired in situ measurements of SSS and $\delta^{18}O_{sw}$ [LeGrande and Schmidt, 2006, Conroy et al., 2014, Conroy et al., 2017], the relationship between the two variables has been assumed constant across space and time. Consequently, empirically derived region-wide slopes have been used ~~in order~~ to forward model $\delta^{18}O_{sw}$ with salinity. The $\delta^{18}O_{sw}$ contribution can be extracted by removing the influence of temperature from $\delta^{18}O_{coral}$, i.e. by measuring temperature-dependent Sr/Ca trace metal in the corals [Ren et al.,

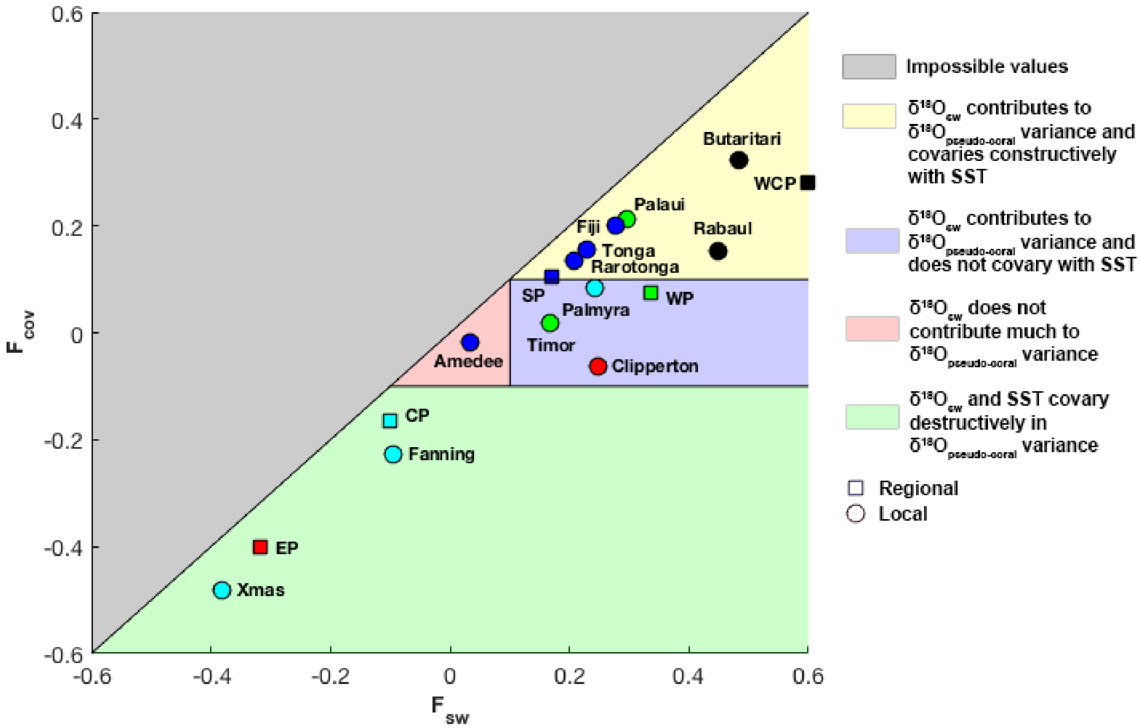

Figure 1. Local and regional interpretation of F_{sw} and F_{cov} metrics. Squares indicate regional averages explained in *Methods* on page 3 and circles are local estimates for the most well-known coral sites as seen in fig 2. The shapes are color-coded by region with red = Eastern Pacific (EP) , cyan = Central Pacific (CP), green = Western Pacific (WP), blue = South Pacific (SP) and black = Western-Central Pacific (WCP). Areas in pastel colors indicate the 'interpretation' associated with a combined set of F_{sw} and F_{cov} values.

2003, Cahyarini et al., 2008]. Results showed that the processes affecting the $\delta^{18}O_{sw}$-SSS relationship may be distinct, leading to uncertainties of more than 50% when using forward modeling [Stevenson et al., 2013].

In this study, I use an isotope-enabled Regional Ocean Modelling System (isoROMS) ~~in order~~ to assess the variability of $\delta^{18}O_{sw}$ as a hydrological proxy in the tropical Pacific Ocean. Its configuration and validation are explained in *Methods* ~~on page 5;~~ the equations used for $\delta^{18}O_{sw}$ and $\delta^{18}O_{pseudo-coral}$ can also be found there.

I first explore how $\delta^{18}O_{sw}$ contributes to $\delta^{18}O_{pseudo-coral}$ variance spatially and temporally. Then, I investigate how spatiotemporal variability affect the $\delta^{18}O_{sw}$-SSS relationship. The goal here is to show how reconstructing $\delta^{18}O_{sw}$ from coral records can be troublesome and may lead to high uncertainties if not properly assessed beforehand. Hereafter, all the results come from isoROMS.

1. Variability of seawater δ^{18}O contribution to pseudo-coral δ^{18}O variance

I define two metrics ~~for this part~~ taken from ~~[~~Russon et al.~~,~~ 2013], whose equations can be found in *Methods* ~~on page 6~~ :

- F_{sw} represents the fraction of $\delta^{18}O_{pseudo-coral}$ variance, i.e. var($\delta^{18}O_{pseudo-coral}$), explained by $\delta^{18}O_{sw}$.
- F_{cov} is the fraction of var($\delta^{18}O_{pseudo-coral}$) explained by both $\delta^{18}O_{sw}$ and SST, i.e. cov(SST,$\delta^{18}O_{sw}$).

1.1 Spatial patterns

The overall spatial distributions of F_{sw} and F_{cov} in figures 1 and 2 agree with the ones in ~~[~~Russon et al.~~,~~ 2013], but my values are higher in magnitude. In the tropical Pacific, the portion of var($\delta^{18}O_{pseudo-coral}$) explained by $\delta^{18}O_{sw}$ is substantial, i.e. $|F_{sw}| > 0.1$. Most of the coral records cannot therefore be interpreted ~~only~~ as palaeo-thermometers. The highest values of $|F_{sw}|$ do not exceed 0.6. For that reason, coral records cannot be used ~~only~~ as hydrological proxies too. ~~Looking at~~ F_{cov} spatial distribution in figure 2 shows a similar pattern to F_{sw}, albeit with a lower magnitude.

In the Western, Western-Central and South Pacific, regional averages of F_{sw} and F_{cov} match rather well with their local counterparts in figure 1. The portion of var($\delta^{18}O_{pseudo-coral}$) explained by $\delta^{18}O_{sw}$ in these regions is significant (not counting Amédé). Coral records there can be used as hydrological proxies, i.e. $|F_{sw}| > 0.1$ and $F_{cov} > -0.1$. Reconstructing the $\delta^{18}O_{sw}$ signal from $\delta^{18}O_{coral}$ is thus possible.

In the Central and Eastern Pacific, the covariation between $\delta^{18}O_{sw}$

and SST is destructive, i.e. $F_{cov} < -0.1$; this renders the extracting of $\delta^{18}O_{sw}$ from $\delta^{18}O_{coral}$ difficult. Moreover, the regional averages and local sites have large offsets, e.g. Palmyra and Clipperton show promise to be good hydrological proxies compared with the regions they are located in.

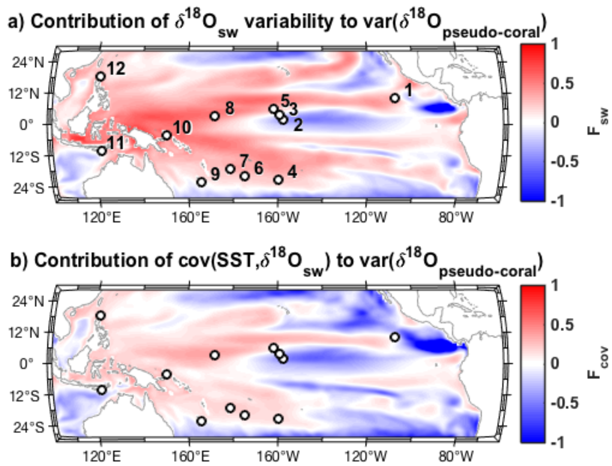

Figure 2. Spatial patterns across the tropical Pacific Ocean of (a) the F_{sw} metric, i.e. the contribution of total $\delta^{18}O_{sw}$ variability to var($\delta^{18}O_{pseudo-coral}$) and (b) the F_{cov} metric, i.e. the contribution of the $\delta^{18}O_{sw}$ variability arising from cov(SST,$\delta^{18}O_{sw}$) to var($\delta^{18}O_{pseudo-coral}$). The black-contoured and white-filled circles represent the coral locations with the corresponding names : (1) Clipperton, (2) Christmas, (3) Fanning, (4) Tonga, (5) Palmyra, (6) Rarotonga, (7) Fiji, (8) Butaritari, (9) Amédée, (10) Rabaul, (11) Timor and (12) Palaui.

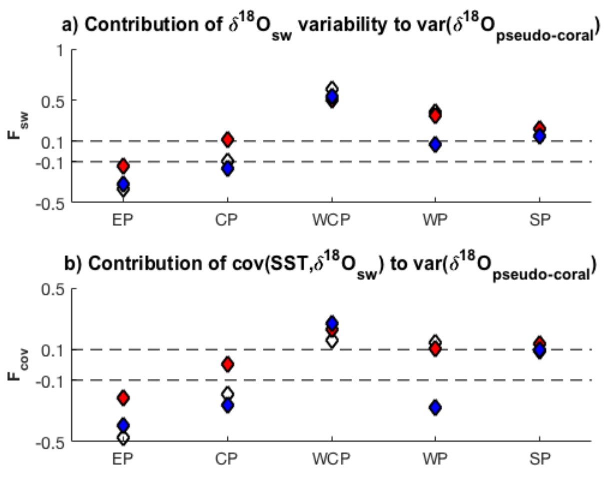

Figure 3. Interannual variability of a) F_{sw} and b) F_{cov} metrics for the different regions. Red = El Niño, blue = La Niña and white = neutral periods. The cold, warm and neutral phases of ENSO are plotted following the definition of the Oceanic Niño Index (ONI) by NOAA for the period 1965-2009.

1.2 Interannual variability

In the Western-Central and South Pacific, F_{sw} and F_{cov} remain the same during neutral or ENSO periods in figure 3. In those regions, the interpretation of F_{sw} and F_{cov} metrics are temporally stable.

The Western Pacific is unaffected by El Niño phases whilst La Niña events highly affect F_{sw} and F_{cov}. La Niña leads to a lower contribution of $\delta^{18}O_{sw}$ in cov($\delta^{18}O_{pseudo-coral}$) because ~~of~~ destructively covarying with SST (green area in figure 1).

The Eastern and Central Pacific are more sensible to El Niño events. For both regions, El Niño diminishes the contribution of $\delta^{18}O_{sw}$, rendering cov(SST,$\delta^{18}O_{sw}$) ~~more~~ negligible (red area in figure 1). In those regions, $\delta^{18}O_{coral}$ can potentially be interpreted ~~only~~ as a palaeothermometer.

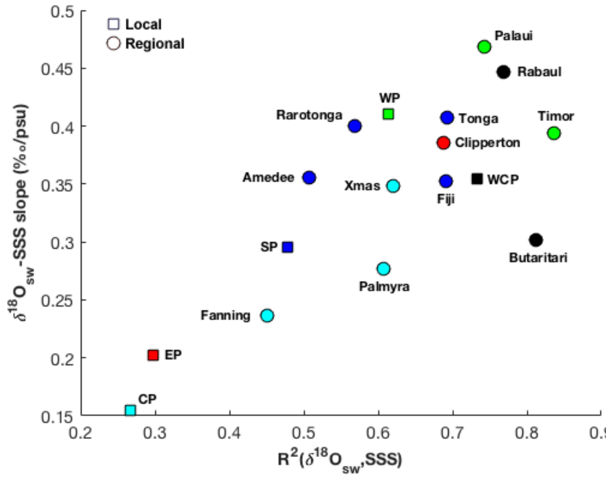

Figure 4. Local and regional $\delta^{18}O_{sw}$-SSS slopes as a function of their R-values for the period 1965-2009. The shapes are color-coded by region with red = Eastern Pacific (EP), cyan = Central Pacific (CP), green = Western Pacific (WP), blue = South Pacific (SP) and black = Western-Central Pacific (WCP).

2. Seawater δ^{18}O-SSS relationship spatiotemporal varibility

Coral-derived $\delta^{18}O_{sw}$ variations are interpreted as SSS variations in palaeoclimatology. Low $\delta^{18}O_{sw}$ values correspond to saline waters and higher values to fresh ones.

Forward modelling of $\delta^{18}O_{coral}$ uses SSS as an approximation for the $\delta^{18}O_{sw}$ contribution. This approximation is a regional $\delta^{18}O_{sw}$-SSS slope derived from the little paired in situ data there is. The slopes used in previous studies do not account for spatiotemporal variability.

The $\delta^{18}O_{sw}$-SSS slopes are steeper in the Eastern and Central Pacific in figure 4. The higher their R-values, the greater the value of the slopes are.

Local slopes do not reflect regional slopes. There is a considerable offset between them. However, the local slopes from the

Western-Central, Western and South Pacific remain somewhat coherent ~~to~~ their regional counterpart. This is not the case for the Eastern and Western Pacific.

[Conroy et al., 2017] measured paired in situ $\delta^{18}O_{sw}$-SSS data during short periods of time ~~as well as~~ computed the slopes and R-values. The sites of interest include Christmas and Manus (closest to Rabaul). $\delta^{18}O_{sw}$-SSS slope values were respectively 0.20 ± 0.05 and 0.17 ± 0.19 ‰/psu. I overestimated both slopes. Furthermore, isoROMS reproduces well observed R-values at Manus (~ 0.7) but not at Christmas (~ 0.14).

The spatial variability of the $\delta^{18}O_{sw}$-SSS slope is on par with the temporal variability, at least on interannual scale, in figure 5.

The slopes are more temporally stable in the Western and South Pacific. A regional slope approximation would lead to less uncertainties in these regions.

The Eastern, Central and Western-Central Pacific have more spread in their temporal slopes, seemingly due to the impact of ENSO.

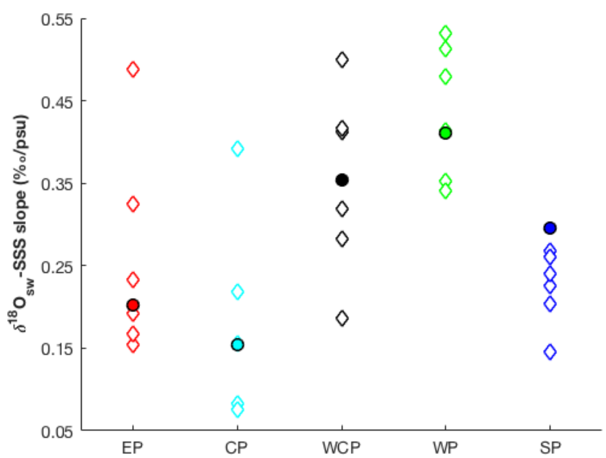

Figure 5. Spatiotemporal variability of the $\delta^{18}O_{sw}$-SSS slope for the different tropical Pacific regions. Red = Eastern Pacific (EP), cyan = Central Pacific (CP), black = Western-Central Pacific (WCP) and blue = South Pacific (SP). Filled circles are spatial slopes (for the period 1965-2009) and the diamonds are temporal slopes. The latter ones are computed with an algorithm that chooses a random 5-year window during the period 1965-2009 and calculates the slope. Six non-overlapping slopes are given for each region.

3. Discussion and conclusions

$\delta^{18}O_{coral}$ is both a function of SST and $\delta^{18}O_{sw}$. The latter is a key factor in better understanding $\delta^{18}O_{coral}$ records since its variability, primarily on interannual timescales, is not well constrained.

Previous studies have used forward modelling ~~in order~~ to understand the contribution of $\delta^{18}O_{sw}$ in the $\delta^{18}O_{coral}$ signal [Thompson et al., 2011, Evans et al., 2013]. In this study, I looked at the variability of $\delta^{18}O_{sw}$ contribution to var($\delta^{18}O_{coral}$). The variability of the $\delta^{18}O_{sw}$ contribution arises either from $\delta^{18}O_{sw}$ itself or from the covariation with SST.

My results are similar to [Russon et al., 2013] with higher contributions of $\delta^{18}O_{sw}$ to var($\delta^{18}O_{pseudo-coral}$) in the Western, Western-Central and South Pacific. Site-specific contributions occur in the Eastern and Central Pacific because of more spatial variability in their contribution.

The contribution of $\delta^{18}O_{sw}$ changes substantially during El Niño events in the Eastern and Central Pacific whilst it does in the Western Pacific during La Niña events. The South and Western-Central Pacific show the most stable contributions. This means that generally those regions would make great hydrological proxies (over the time period concerned) as they are not affected as much by ENSO.

SSS has been used widely as a proxy for $\delta^{18}O_{sw}$, assuming a linear relationship between the two. Because of the lack of paired in situ measurements, their relationship cannot be constrained locally and as such, regional $\delta^{18}O_{sw}$-SSS slopes are used. In order to study $\delta^{18}O_{sw}$ and the SSS-$\delta^{18}O_{sw}$ relationship in a direct manner, isotope-enabled models have been specifically designed to incorporate seawater oxygen isotopes [LeGrande and Schmidt, 2011, Russon et al., 2013, Stevenson et al., 2015, Stevenson et al., 2018]. I showed with isoROMS that regional patterns are quite similar to observations [LeGrande and Schmidt, 2006, Conroy et al., 2014, Conroy et al., 2017] and model results [Russon et al., 2013, LeGrande and Schmidt, 2011] with higher slopes in the Western, Western-Central and South Pacific. However, the model overestimates the SSS-$\delta^{18}O_{sw}$ slopes locally, where data were available [Conroy et al., 2017].

Regional slopes do not make good averages due to site-specific relationships leading to great offsets. This leads to more errors when forward-modelling, on top of the fact that SSS and $\delta^{18}O_{sw}$ do not covary strongly at some sites [Stevenson et al., 2018].

SSS-$\delta^{18}O_{sw}$ temporal slope variability is on the same magnitude as spatial slope variability. The error of reconstructing phenomena like ENSO will then have a temporal offset as high as the spatial pattern observed.

In this study, I used modelled $\delta^{18}O_{sw}$ data, which does not equate to using coral-derived $\delta^{18}O_{sw}$. A further investigation on this subject is needed.

Palaeo-reconstruction of $\delta^{18}O_{coral}$ lead to great uncertainties if one does not properly assess variability of the $\delta^{18}O_{sw}$ contribution.

Salinity-based forward-modelling of $\delta^{18}O_{coral}$, a tool of utmost importance in palaeoclimatology, also lead to great uncertainties if the variability of the SSS-$\delta^{18}O_{sw}$ relationship is not properly assessed. The impacts of temporal and site-specific variations have indeed been shown to be important.

More model-based studies of local scale processes and continuous in situ measurements are needed in order to fully appreciate coral-derived $\delta^{18}O_{sw}$ as a hydrological proxy.

Acknowledgments

I would like to thank Dr. Kim Cobb for giving the opportunity to work in her wonderful lab and hone my skills as a paleoclimatologist. I would also like to thank Dr. Hussein Sayani for being the best person ever and helping with everything I needed. To Aaron Jones and Shellby Miller, you literally put a smile on my face every day that I came to work and made my life much better now that you are in it. Hudson Moss, I do not forget you, you're the best. To Samantha Stevenson, thank you for giving me access to isoROMS.

Methods

isoROMS configuration

IsoROMS is a modified version of ROMS [Shchepetkin and C. McWilliams, 2005], to which [Stevenson et al., 2015, Stevenson et al., 2018] and I added oxygen isotope tracers.

The model allows the simulation of $\delta^{18}O_{sw}$ by adding $H_2^{16}O$ and $H_2^{18}O$ as passive tracers in the water column. Initial and boundary conditions are taken from the German Contribution to the Estimating of the Circulation and Climate of the Ocean (GECCO2) [Armin, 2015] for oceanic conditions, the isotope-enabled Community Atmosphere Model v5 (iCAM5) [Nusbaumer et al., 2017] for precipitation δ^{18}O values, the global gridded dataset from [LeGrande and Schmidt, 2006] for seawater oxygen isotopes and the Common Ocean-Ice Reference experiment (CORE2) [Large and Yeager, 2009] for atmospheric forcings. The model encompasses a fine meridional and zonal resolution, especially near the equator which allows for a study of local processes affecting $\delta^{18}O_{sw}$ compared to prior GCM experiments that have been used for isotope simulations. isoROMS spans from 1965 to 2009.

IsoROMS only gives us water isotopes of the first 50 m in the ocean but not the direct $\delta^{18}O_{sw}$ values. This can be done using the formula of $\delta^{18}O_{sw}$:

$$\delta^{18}O_{sw} = \left(\frac{\left(\frac{^{18}O}{^{16}O}\right)_{water}}{\left(\frac{^{18}O}{^{16}O}\right)_{standard}} - 1 \right) * 10^3 \qquad (1)$$

with $\left(\frac{^{18}O}{^{16}O}\right)_{water}$ the oxygen isotopic ratio in the water and $\left(\frac{^{18}O}{^{16}O}\right)_{standard}$ the oxygen isotopic ratio in the standard, i.e. Vienna Standard Mean Ocean Water (V-SMOW).

In order to compare the model to $\delta^{18}O_{coral}$ values, we need to compute pseudo-coral data using forward modeling on isoROMS outputs. The following equation, used in [Russon et al., 2013], takes into account any vital effects [McConnaughey, 1989a, McConnaughey, 1989b] :

$$\delta^{18}O_{pseudo-coral} = \delta^{18}O_{sw} + \gamma * (SST - C) \qquad (2)$$

with $\delta^{18}O_{pseudo-coral}$ (‰) the modeled coral δ^{18}O (‰) and C (°C) the variable taking into account 'vital-effects' whose value is estimated to be $C \sim 4$°C [Erez, 1978].

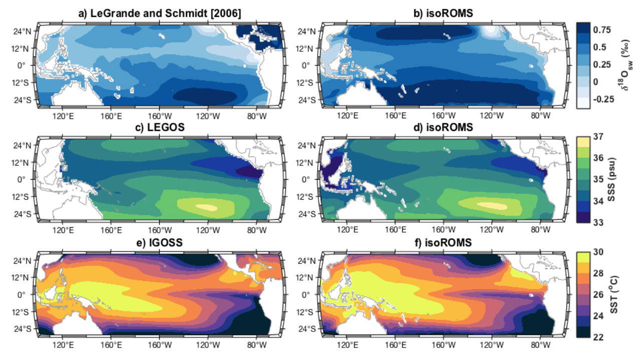

Figure 6. Climatological mean of $\delta^{18}O_{sw}$ in (a) [LeGrande and Schmidt, 2006] and (b) isoROMS, SSS in (c) LEGOS and (d) isoROMS, SST in (e) IGOSS and (f) isoROMS. The means are computed over the period 1981-2010 when satellite data started being available. The tool m_map for MatLab was used to create the maps.

isoROMS validation

The model validation is done with the following data :

- The gridded $\delta^{18}O_{sw}$ product from [LeGrande and Schmidt, 2006] (interpolated data from observations) for spatial structures, i.e. in figure 6.
- IGOSS SST from 1981 to 2009 for spatial structures.
- Monthly blended ship and satellite Laboratoire d'Études en Géophysique et Océanographie Spatiales (LEGOS) [Delcroix et al., 2011] SSS dataset from 1981 to 2009 for spatial structures and from 1965 to 2009 for local sites.
- Monthly blended ship and satellite Extended Reconstruction Sea Surface Temperature v5 (ERSSTv5) [Huang et al., 2017] SST product from 1965 to 2009 for local sites.

The model does very well at representing the main spatial structures of SSS and SST observational data, as seen in figure 6, such as the Western Pacific Warm Pool (WPWP) pattern showing for the SST maps and the less saline waters in the East due to the upwelling of deep waters for the SSS maps. However, for $\delta^{18}O_{sw}$, although isoROMS does share similarity with [LeGrande and Schmidt, 2006], the clear spatial pattern in the northern tropical Pacific appearing in isoROMS is not present in the observations. Furthermore, the model values are overall a bit higher.

Spatial averages

For regional patterns discussed in this study, I refer to spatial averages made using isoROMS, with these corresponding definitions :

- Eastern Pacific (EP) : 12°N−8°S, 120°W−80°W

- Central Pacific (CP) : 8°N−8°N, 170°W−120°W
- Western-Central Pacific (WCP) : 8°N−8°S, 140°E−180°E
- Western Pacific (WP) : 20°N−10°S, 120°E−140°E
- South Pacific (SP) : 10°S−22°S, 160°E−140°W

For local sites, I took the coordinates of the most well-known coral sites :

- Clipperton in EP
- Christmas, Fanning and Palmyra in CP
- Butaritari and Rabaul in WCP
- Timor and Palaui in WP
- Tonga, Rarotonga, Fiji and Amédée in SP

F_{sw} and F_{cov} metrics

Metrics were made by [Russon et al., 2013] in order to study the contribution of $\delta^{18}O_{sw}$ to $\delta^{18}O_{pseudo-coral}$ variance. From eq. 2, we can see that the contribution of $\delta^{18}O_{sw}$ to $\delta^{18}O_{pseudo-coral}$ not only depends on $\delta^{18}O_{sw}$ itself but also on the covariance of both $\delta^{18}O_{sw}$ and SST. The fraction of var($\delta^{18}O_{pseudo-coral}$) from $\delta^{18}O_{sw}$ variability is given by the F_{sw} metric defined as follows :

$$F_{sw} = \frac{var(\delta^{18}O_{sw}) + 2\gamma * cov(SST, \delta^{18}O_{sw})}{var(\delta^{18}O_{pseudo-coral})} \quad (3)$$

Values of F_{sw} vary between - 1 and 1. However, $|F|_{sw} < 0.1$ is considered to be insignificant, i.e. the fraction of var($\delta^{18}O_{pseudo-coral}$) from $\delta^{18}O_{sw}$ variability is insignificant. Another interpretation for the latter is that the contribution from $\delta^{18}O_{sw}$ variability is only 10% of the total var($\delta^{18}O_{pseudo-coral}$). This 10% is our baseline of insignificance for the rest of this study. The greater $|F_{sw}|$ is, the greater $\delta^{18}O_{sw}$ variability plays a role in var($\delta^{18}O_{pseudo-coral}$). Significant negative F_{sw} values, i.e. $F_{sw} < -0.1$, relate to the fact that $\delta^{18}O_{sw}$ variability has a positive influence on SST (since γ is negative in eq. 3, cov(SST,$\delta^{18}O_{sw}$) has to be positive for F_{sw} to be negative). The more positive the correlation between the two, the more var($\delta^{18}O_{sw}$) is lowered. The opposite reasoning is not true for significant positive F_{sw} values, i.e. $F_{sw} > 0.1$. The problem underlined by [Russon et al., 2013] is that significant positive values can result from constructive covariation (anti-correlation) between SST and $\delta^{18}O_{sw}$ but also if $\delta^{18}O_{sw}$ variability is very significant and substantially independent from SST variability. The latter means that $\delta^{18}O_{sw}$ could be considered an added 'white noise' contribution to the overall SST contribution. To make sure our interpretation is correct, we have to make sure that significant positive values are one or the other, thus we need to consider independently the fraction of var($\delta^{18}O_{coral}$) explained by cov(SST, $\delta^{18}O_{sw}$) alone. This metric is called F_{cov} and is given by :

$$F_{cov} = \frac{2\gamma * cov(SST, \delta^{18}O_{sw})}{var(\delta^{18}O_{pseudo-coral})} \quad (4)$$

Values of F_{cov} also vary between - 1 and 1. Furthermore, $|F_{cov}| < 0.1$ is considered to be insignificant as well, i.e. the fraction of var($\delta^{18}O_{pseudo-coral}$) explained by cov(SST,$\delta^{18}O_{sw}$) is insignificant, thus the contribution from $\delta^{18}O_{sw}$ variability arising from cov(SST,$\delta^{18}O_{sw}$) alone is only 10% of the total var($\delta^{18}O_{pseudo-coral}$). Significant positive F_{cov} values mean that $\delta^{18}O_{sw}$ and SST covary constructively (negative correlation between the two) whereas significant negative F_{cov} values signify a destructive covariation between SST and $\delta^{18}O_{sw}$ (positive correlation between the two).

References

[Armin, 2015] Armin, K. (2015). Evaluation of the GECCO2 ocean synthesis: transports of volume, heat and freshwater in the Atlantic. *Quarterly Journal of the Royal Meteorological Society*, 141(686):166–181.

[Cahyarini et al., 2008] Cahyarini, S. Y., Pfeiffer, M., Timm, O., Dullo, W.-C., and Schönberg, D. G. (2008). Reconstructing seawater δ^{18}O from paired coral δ^{18}O and Sr/Ca ratios: Methods, error analysis and problems, with examples from Tahiti (French Polynesia) and Timor (Indonesia). *Geochimica et Cosmochimica Acta*, 72(12):2841–2853.

[Cobb et al., 2003] Cobb, K. M., Charles, C. D., Cheng, H., and Edwards, R. L. (2003). El Niño/Southern Oscillation and tropical Pacific climate during the last millennium. *Nature*, 424:271–276.

[Collins et al., 2010] Collins, M., An, S.-I., Cai, W., Ganachaud, A., Guilyardi, E., Jin, F.-F., Jochum, M., Lengaigne, M., Power, S., Timmermann, A., Vecchi, G., and Wittenberg, A. (2010). The impact of global warming on the tropical Pacific Ocean and El Niño. *Nature Geoscience*, 3:391–397.

[Conroy et al., 2014] Conroy, J. L., Cobb, K. M., Lynch-Stieglitz, J., and Polissar, P. J. (2014). Constraints on the salinity-oxygen isotope relationship in the central tropical Pacific Ocean. *Marine Chemistry*, 161:26–33.

[Conroy et al., 2017] Conroy, J. L., Thompson, D. M., Cobb, K. M., Noone, D., Rea, S., and LeGrande, A. N. (2017). Spatiotemporal variability in the δ^{18}O-salinity relationship of seawater across the tropical Pacific Ocean. *Paleoceanography*, 32(5):484–497.

[Delcroix et al., 2011] Delcroix, T., Alory, G., Cravatte, S., Corrège, T., and McPhaden, M. (2011). A gridded sea surface salinity data set for the tropical Pacific with sample applications (1950-2008). *Deep Sea Research Part I: Oceanographic Research Papers*, 58(1):38–48.

[Epstein et al., 1953] Epstein, S., Buchsbaum, R., Lowenstam, H. A., and Urey, H. C. (1953). Revised carbonate-water isotopic temperature scale. *GSA Bulletin*, 64(11):1315.

[Erez, 1978] Erez, J. (1978). Vital effect on stable-isotope composition seen in foraminifera and coral skeletons. *Nature*, 273:199–202.

[Evans et al., 2013] Evans, M., Tolwinski-Ward, S., Thompson, D., and Anchukaitis, K. (2013). Applications of proxy system modeling in high resolution paleoclimatology. *Quaternary Science Reviews*, 76:16–28.

[Huang et al., 2017] Huang, B., Thorne, P. W., Banzon, V. F., Boyer, T., Chepurin, G., Lawrimore, J. H., Menne, M. J., Smith, T. M., Vose, R. S., and Zhang, H.-M. (2017). Extended Reconstructed Sea Surface Temperature, Version 5 (ERSSTv5): upgrades, validations, and intercomparisons. *Journal of Climate*, 30(20):8179–8205.

[Large and Yeager, 2009] Large, W. G. and Yeager, S. G. (2009). The global climatology of an interannually varying air-sea flux data set. *Climate Dynamics*, 33(2):341–364.

[LeGrande and Schmidt, 2011] LeGrande, A. and Schmidt, G. (2011). Water isotopologues as a quantitative paleosalinity proxy. *Paleoceanography*, 26:PA3225.

[LeGrande and Schmidt, 2006] LeGrande, A. N. and Schmidt, G. A. (2006). Global gridded data set of the oxygen isotopic composition in seawater. *Geophysical Research Letters*, 33(12).

[McConnaughey, 1989a] McConnaughey, T. (1989a). ^{13}C and ^{18}O isotopic disequilibrium in biological carbonates: I. Patterns. *Geochimica et Cosmochimica Acta*, 53(1):151–162.

[McConnaughey, 1989b] McConnaughey, T. (1989b). ^{13}C and ^{18}O isotopic disequilibrium in biological carbonates: Ii. In vitro simulation of kinetic isotope effects. *Geochimica et Cosmochimica Acta*, 53(1):163–171.

[Nusbaumer et al., 2017] Nusbaumer, J., Wong, T. E., Bardeen, C., and Noone, D. (2017). Evaluating hydrological processes in the Community Atmosphere Model Version 5 (CAM5) using stable isotope ratios of water. *Journal of Advances in Modeling Earth Systems*, 9(2):949–977.

[Ren et al., 2003] Ren, L., Linsley, B. K., Wellington, G. M., Schrag, D. P., and Hoegh-guldberg, O. (2003). Deconvolving the δ^{18}O seawater component from subseasonal coral δ^{18}O and Sr/Ca at Rarotonga in the southwestern subtropical Pacific for the period 1726 to 1997. *Geochimica et Cosmochimica Acta*, 67(9):1609–1621.

[Russon et al., 2013] Russon, T., Tudhope, A. W., Hegerl, G. C., Collins, M., and Tindall, J. (2013). Inter-annual tropical Pacific climate variability in an isotope-enabled CGCM: implications for interpreting coral stable oxygen isotope records of ENSO. *Climate of the Past*, 9(4):1543–1557.

[Shchepetkin and C. McWilliams, 2005] Shchepetkin, A. and C. McWilliams, J. (2005). The Regional Oceanic Modeling System (ROMS): a split-explicit, free-surface, topography-following-coordinate ocean model. 9:347–404.

[Stevenson et al., 2013] Stevenson, S., McGregor, H. V., Phipps, S. J., and Fox-Kemper, B. (2013). Quantifying errors in coral-based ENSO estimates: Toward improved forward modeling of δ^{18}O. *Paleoceanography*, 28(4):633–649.

[Stevenson et al., 2018] Stevenson, S., Powell, B. S., Cobb, K. M., Nusbaumer, J., Merrifield, M. A., and Noone, D. (2018). 20th century seawater δ^{18}O dynamics and implications for coral-based climate reconstruction. *Paleoceanogr. Paleoclimatol.*, 33(6):606–625.

[Stevenson et al., 2015] Stevenson, S., Powell, B. S., Merrifield, M. A., Cobb, K. M., Nusbaumer, J., and Noone, D. (2015). Characterizing seawater oxygen isotopic variability in a regional ocean modeling framework: Implications for coral proxy records. *Paleoceanography*, 30(11):1573–1593.

[Thompson et al., 2011] Thompson, D. M., Ault, T. R., Evans, M. N., Cole, J. E., and Emile-Geay, J. (2011). Comparison of observed and simulated tropical climate trends using a forward model of coral δ^{18}O. *Geophysical Research Letters*, 38(14).

[Tudhope et al., 2001] Tudhope, A. W., Chilcott, C. P., McCulloch, M. T., Cook, E. R., Chappell, J., Ellam, R. M., Lea, D. W., Lough, J. M., and Shimmield, G. B. (2001). Variability in the El Niño-Southern Oscillation through a glacial-interglacial cycle. *Science*, 291(5508):1511–1517.

[Vecchi and Wittenberg, 2010] Vecchi, G. A. and Wittenberg, A. T. (2010). El niño and our future climate: where do we stand? *Wiley Interdisciplinary Reviews: Climate Change*, 1(2):260–270.

[Weber and Woodhead, 1970] Weber, J. N. and Woodhead, P. M. (1970). Carbon and oxygen isotope fractionation in the skeletal carbonate of reef-building corals. *Chemical Geology*, 6:93–117.