Oxygen isotopes in corals and their use as proxies for El Niño

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English abstract

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Abstract: El Niño and the Southern Oscillation (ENSO) have a great impact on global climate, often with disastrous effects in many parts of the world. The carbonate skeleton of corals record El Niño activity by tracing sea surface temperature and salinity in the oceans where they grow. As El Niño's core region is the equatorial Pacific Ocean, this is where the most direct consequences of ENSO are seen, commonly associated with temperature anomalies and changes in rain patterns. These changes affect the isotopic and elemental composition of surface ocean waters, which are in turn observed in corals in the affected regions. This review presents how oxygen isotope ratios ($\delta^{18}O$) in coral skeleton can be used as a proxy to determine climatic trends of El Niño in the past, mainly with a focus on the Holocene. I explain the main features of ENSO in the present, including its different phases and their effects. The function of corals and their oxygen isotopes is described, providing detailed information on what determines the $\delta^{18}O$ values in seawater and carbonates, as well as the practical uses of coral $\delta^{18}O$ and what it can tell us about paleoclimate. Further, I present and analyse the information obtained from coral $\delta^{18}O$ on ENSO patterns in the past. Lastly, I suggest some improvements for future use of paleoclimatic methods for studying El Niño Southern Oscillation in order to give a more holistic understanding of past climate.

Keywords: El Niño, ENSO, oxygen isotopes, coral, sea surface temperature, paleoclimate, Pacific Ocean, $\delta^{18}O$

Supervisors: Helena L. Filipsson

Subject: Quaternary Geology

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Svenskt abstrakt

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Sammanfattning: El Niño Southern Oscillation (ENSO) har stor inverkan på det globala klimatet, ofta med katastrofala följder i många delar av världen. Korallers karbonatskelett kan ge information om El Niño-aktivitet genom att spåra havsytans temperatur och salthalt där de växer. Eftersom El Niños kärnområde är ekvatoriala Stilla Havet är det där de mest direkta konsekvenserna av ENSO kan ses, mestadels är de förknippade med temperaturavvikelser och förändringar i nederbörd. Dessa förändringar påverkar isotop- och grundämnessammansättningen av havets ytvatten, som i sin tur observeras i koraller i de drabbade regionerna. Denna litteraturstudie visar hur relationen mellan tunga och lätta syreisotoper ($\delta^{18}O$) i korallskelett kan användas för att avgöra El Niños klimattrender i det förflutna, framförallt med fokus på Holocen. Jag förklarar ENSO i nuet, inklusive dess olika faser och deras effekter. Funktionen av koraller och deras syreisotoper beskrivs, samt vad som avgör $\delta^{18}O$-värden i havsvatten och karbonater. Jag beskriver också hur korallers $\delta^{18}O$ kan användas praktiskt och vad det kan berätta om paleoklimat. Vidare presenterar jag och analyserar information från korall-$\delta^{18}O$ vad gäller ENSOs trender i det förflutna. Slutligen föreslår jag vissa förbättringar för framtida användning av paleoklimatiska metoder för att studera El Niño Southern Oscillation i syfte att ge en bättre förståelse av paleoklimat.

Nyckelord: El Niño, ENSO, syreisotoper, korall, havsytetemperatur, paleoklimat, Stilla Havet, $\delta^{18}O$

Handledare: Helena L. Filipsson

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1 Introduction

1.1 Importance of the study

El Niño has gained considerable attention in the mainstream media during the last few decades following a couple of intense episodes with major climatic consequences, for instance the events of 1982 and 1998. Numerous natural disasters have been attributed to El Niño as it has often led to floods, droughts and other extreme weather occurrences (Corrège 2006). El Niño is essentially a warming of surface waters in the eastern equatorial Pacific Ocean that normally lasts for a year. Its occurrence is irregular but typically stays within the range of 2-7 years and is strongly associated with fluctuations in atmospheric pressure called the Southern Oscillation. The system as a whole is thus known as El Niño Southern Oscillation (ENSO) and comprises both the atmospheric and oceanic phases (Carriquiry et al. 1994). Since ENSO has its origin in the equatorial Pacific Ocean, it is important to understand how tropical oceans vary in thermal and hydrological states in order to comprehend the fundamentals of El Niño and how it impacts the global climate system (Gagan et al. 2000).

The oceans play an immense role in regulating temperature, pressure, wind and precipitation. Waters around the Equator are particularly substantial to climate since they on average receive 2.5 times more solar radiation than the poles. A large part of the insolation that the equatorial zones receive is transported to higher latitudes (Corrège 2006). It is estimated that without this heat transport, the equatorial regions would be 14°C warmer while the polar regions would be 25°C colder than presently (Barry & Chorley 1998). Not only does El Niño have a large impact on Pacific equatorial waters, but its influence expands over vast parts of the globe outside of the tropics (Philander 1990).

The possibility that El Niño has held a major role in shaping past climatic patterns is probable and makes it an interesting topic to study. Unfortunately, instrumental measurements in tropical oceans older than 50 years are scarce (Liu et al. 2013) and virtually non-existent before the late 19th century (Brohan et al. 2006). Therefore, to determine the effects of El Niño in earlier times, it is necessary to resort to geologic proxy records instead. A proxy method often used for marine environments is the oxygen isotope ratio in carbonate marine organisms such as foraminifera, molluscs and corals. The deviation of the given sample's oxygen isotope ratio from a standardised ratio is referred to as the $\delta^{18}O$. Corals have been proven to provide reliable and high-resolution records of sea surface temperature (SST) and sea surface salinity (SSS) variability (Wu & Grottoli 2009). Their skeletal $\delta^{18}O$ signature relates to seawater temperature and seawater $\delta^{18}O$, which in turn is associated with the hydrologic cycle, temperature, salinity and precipitation (Cole & Fairbanks 1990). In many cases it is possible to achieve results as accurate as monthly to weekly temperature variability from coral $\delta^{18}O$, making it a unique and valuable proxy for El Niño (Giry et al. 2012).

1.2 Aim of the study

The aim of this study is to review how coral $\delta^{18}O$ can be used to reconstruct and retrieve information about El Niño that would not be possible to attain with instrumental measurements. This will include a description of causes and effects of El Niño, how corals grow and live, and their relationship with $\delta^{18}O$ in the surrounding waters, as well as some ways of measuring and analysing results. Furthermore, the way El Niño has affected the global climate in the past will be examined with a particular focus on the Holocene and predictability for the future will be briefly discussed. Lastly, some uncertainties will be analysed, and possible improvements will be proposed.

2 El Niño in the present

2.1 Causes for El Niño

In order to grasp the importance that El Niño has on global climate, it is first essential to understand what normal conditions are like in the Pacific Ocean, as well as the mechanisms that drive atmospheric and oceanic currents. Solar heating is the main driving force of all atmospheric circulation. Since insolation is on average greatest at the equator and smallest at the poles, warm air rises near the equator and is transported polewards near the tropopause. At approximately the latitudes 30°N and 30°S the air cools, sinks and is transported near the Earth's surface back toward the low-pressure equatorial regions, forming an atmospheric circulation called a Hadley cell (Figure 1).

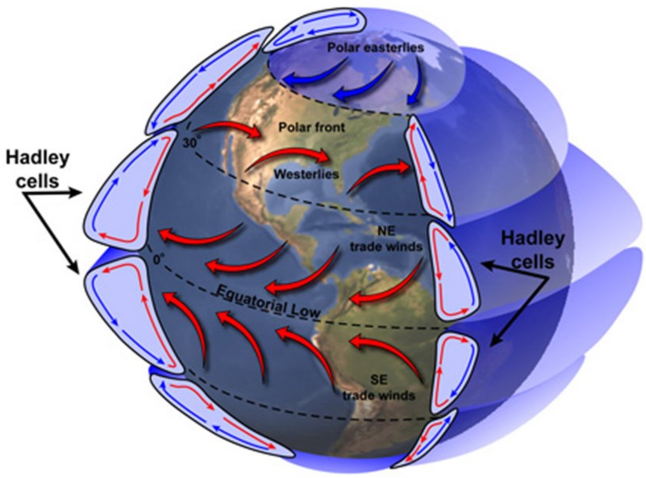

Fig. 1. Warm equatorial air rises and is transported towards the poles. At approximately 30°N and 30°S the air sinks and is transported back near the Earth's surface toward the equator. This circulation forms what is known as Hadley cells. *Source: NASA Earth Observatory.*

As the winds blow towards the equator at low altitudes, the Earth's eastward rotation causes the winds to deflect to the west. This is known as the Coriolis effect. These winds that blow from the northeast in the Northern Hemisphere and from the southeast in the Southern Hemisphere are called trade winds or easterlies and are the basis for a vast array of tropical climatic features (Persson 2006).

During non-El Niño conditions, the strong south-eastern trade winds over the Pacific Ocean bring warm surface water from the western coast of South America westwards. As this warm water pool of ~28° C moves towards the Australian east coast, cold bottom waters upwell in the east Pacific, near the Peruvian coast, where the thermocline is shallow, (~50 m; compare to the 200 m deep thermocline in the west) (Martinez 2009). The upwelling generates the Humboldt Current, a northward flow of cold, low salinity ocean water from Chile's southern tip to northern Peru (Thatje et al. 2008). Consequently, surface waters are 4-10°C colder in the east Pacific than in the west. The sea surface temperature difference also means a higher atmospheric pressure in the east. Since air moves down the pressure gradient, this results in stronger trade winds that keep pushing surface waters westwards, and thus a mechanism of positive feedback occurs (Bjerknes 1969).

Every 2-8 years, however, the easterly trade winds weaken (McGregor et al. 2013). The warm water pool is no longer pushed westward, but approaches the South American west coast. Upwelling diminishes, surface water temperatures increase and rain patterns change (Figure 2). These conditions normally prevail for at least nine months to up to two years and are known as El Niño. Peruvian fishermen have long used the term "El Niño" (Spanish for "The Boy", in reference to baby Jesus) to refer to the annual warming of coastal waters that happens around Christmas (Cane 2005). Nowadays however, scientists limit this term to the warming of waters combined with high sea level, deep thermocline and increased rainfall in the eastern equatorial Pacific (Carriquiry et al. 1994).

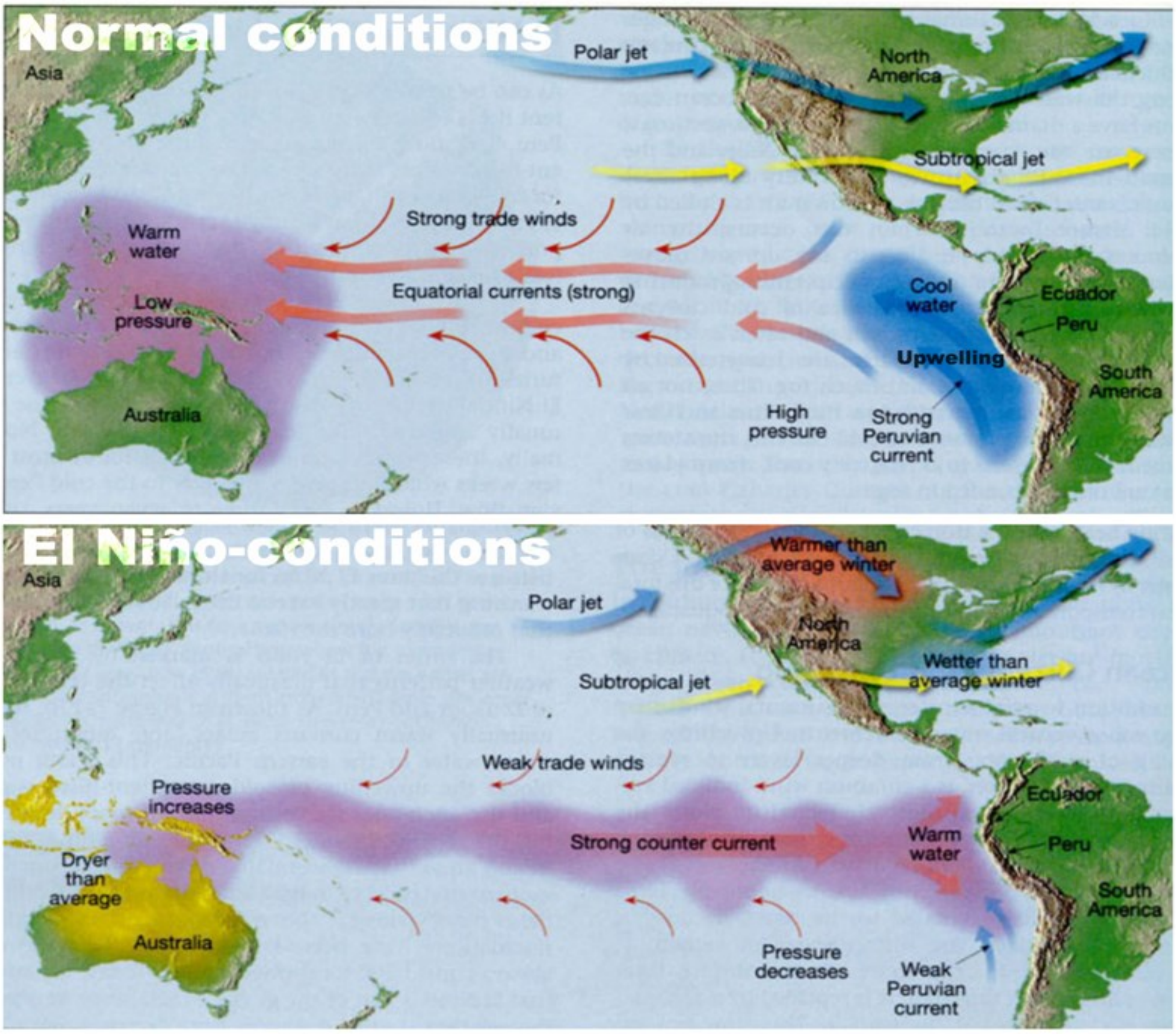

Fig. 2. Normal conditions are characterized by strong easterly trade wind, cold water upwelling in the east and a warm water pool in the west pacific. During El Niño the winds weaken and the warm water flows eastward, resulting in cooler surface waters in the west and warmer surface waters in the east. *Modified after Madl 2000.*

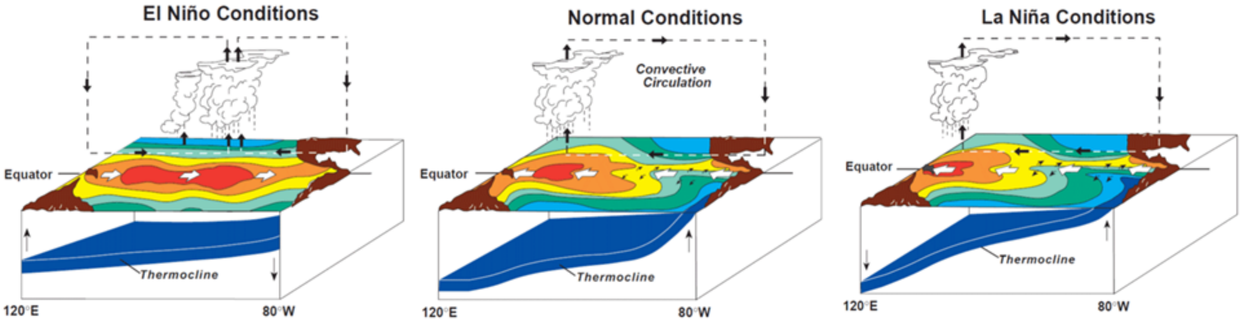

Fig. 3. The three stages of ENSO. During El Niño the thermocline in the east is deep and sea surface temperatures are warm. Normal conditions occur when trade winds are strong and warm water is pushed westward. During La Niña the trade winds are further strengthened and cold anomalies occur in the eastern Pacific, shallow thermocline in the east and deep thermocline in the west. *Source: NOAA/PMEL/TAO Project Office, Dr. Michael J. McPhaden.*

2.2 The Southern Oscillation

The changes are brought about by the Southern Oscillation (SO), a large-scale fluctuation in the surface air pressure between the tropical eastern and western Pacific. When the sea level pressure is below normal in the southeast tropical Pacific region, it follows that the pressure is above normal values in the Australian-Indonesian region, and vice versa. These pressure differences alternate on a timescale of monthly to yearly changes. The Southern Oscillation Index defines its strength by measuring the sea level pressure difference between Papeete, Tahiti and Darwin, Australia (Enfield 1992). The Southern Oscillation was first documented by Sir Gilbert Walker in 1923 (Walker 1923), but its association with SSTs in the Pacific remained undiscovered until the late 1950s (Berlage 1957)..

During a SO high phase, the high pressure of the southeast Pacific is higher than normal while the low pressure in the Indonesian region is lower than normal. This generates an increased pressure gradient and strong easterly trade wind. Conversely, during SO low phases the mechanism is reversed resulting in weakened trade winds, which corresponds with the SO warm phase El Niño (Enfield 1992).

2.3 Seasonal evolution of ENSO

ENSO comprises three phases as seen in Figure 3:

1. normal conditions during which there is upwelling of cold waters in the eastern Pacific and the easterly trade winds push the Pacific warm water pool westward. SSTs are colder in the eastern Pacific and warmer in the western Pacific Ocean;
2. ENSO warm phase or El Niño conditions, during which the SST in the eastern Pacific increase due to the warm water pool flowing eastward and upwelling lessening; and
3. a third, atypically cold phase that many times follows El Niño, when eastern Pacific SSTs drop below normal values. This ENSO cold phase is commonly called La Niña ("The Girl" in Spanish) because it is the antithesis of El Niño (Martinez 2009).

While ENSO does not have a set series of events that are consistent in time and space, there are some general characteristics that exemplify a typical ENSO cycle. El Niño anomalies typically start to emerge in June-August, increase in September-November and peak in December-February, after which it starts to fade. In total it lasts approximately a year. Although ENSO does not take place annually, it is interesting to note that when it does occur, the events are phase-locked according to the climatological year (McGregor et al. 2013). Another intriguing trait of the ENSO cycle is that a cold phase often is observed to follow a warm phase in the subsequent year. Likewise, a warm phase frequently follows a cold phase. The warm El Niño and cold La Niña events seemingly represent the two extremes of the ENSO cycle, as the system oscillates from one end to the other (Diaz & Markgraf 1992).

A commonly used index to determine what qualifies as El Niño conditions, is measuring the sea surface temperature and sea level pressure in a region defined as NINO3 (5S-5N, 90W-150W). Temperature anomalies have been between 0.5°C and 1.5°C warmer during weak El Niños, and around 2.5°C during strong El Niño conditions for the past 45 years (Kerr 1983).

Conversely, La Niña is characterised by unusually cold surface waters in the same region. This happens when the cold-water pool (normally ~18°C) becomes cooler and the trade winds strengthen. La Niña conditions are defined as surface waters in the NINO3 region reaching temperatures of at least 0.5°C colder than normal. Since 1970 there have been several La Niñas reaching temperatures colder than 1.5°C and one passing 2°C colder than normal ("El Niño y la Niña", Comunidad Andina). It is largely the opposite of El Niño both in causes and effects.

2.4 Effects of El Niño

The changes that the equatorial Pacific

Ocean sees during the different phases of ENSO have profound ramifications both for the core region and for more distant areas. Remote atmospheric and oceanic anomalies are linked to ENSO through teleconnections, i.e. the mechanism of responses outside of the equatorial Pacific Ocean to the fluctuation of sea level pressure and sea surface temperature (Diaz & Kiladis 1992). Oceanic changes include a rise in sea level near the shore in the eastern Pacific, a deepening of the thermocline, changes in currents and advection and an increase in dissolved oxygen concentration where there is normally hypoxia (i.e. a concentration of <2 mg O_2/l) (Tarazona & Arntz 2001).

The warming of surface waters off the coast of South America has a direct impact on fishing in the area. The cold waters that usually flow upward bring nutrients to the surface, making it an ideal environment for marine fauna. As upwelling decreases the surface waters become nutrient poor. This causes problems for the coastal populations that rely on local fishing industry, as there is a substantial reduction in fish (Thatje et al. 2008). In the lower elevations of Ecuador, Peru and Bolivia, El Niño causes increased rainfall, floods and mass wasting (such as landslides and mudslides), while Colombia and the Altiplano of Peru and Bolivia see increased droughts and forest fires. These effects translate into important losses in fishing industry, agriculture, infrastructure and homes. A particularly strong El Niño in 1997-1998 led to losses in Colombia and Bolivia worth more than 500 million USD ("El Niño y la Niña", Comunidad Andina).

Change in precipitation is a key feature of El Niño. Evaporation of warm surface water leads to cloud formation, so when the Pacific warm water pool moves eastward, precipitation patterns change. Instead of abundant rain over the west Pacific Ocean, South East Asia and eastern Australia, these regions often experience periods of dry conditions during El Niño (Chen & Cane 2008). For instance, in 1982-1983, Australia and Africa experienced the most severe droughts ever recorded (Arntz & Tarazona 1990). In fact, it has been observed since the early 20th century that droughts in Australia often coincide with similar conditions in India, northern China, and parts of Africa and America, and occur during El Niño episodes (Nicholls 1992).

Over the Atlantic Ocean, Caribbean Sea and Gulf of Mexico there is usually a decrease in hurricanes during El Niño, but an increase over the United States and west coast of Mexico as well as more typhoons in the central Pacific Ocean ("Effects of El Niño on World Weather", RNMI). Various biological and physical markers such as corals, marine molluscs, glacial ice, tree rings and sedimentary layers record many of the effects that have their origin in El Niño.

3 Corals

3.1 Location and habitat

Corals are animals of the phylum *Cnidaria* that typically live in colonies in marine environments. Many corals form hard exoskeletons that support the soft tissues (polyps) inside and build reefs that can last for thousands of years if undisturbed. These large structures are mostly found in shallow tropical waters (Spalding 2001) where sea surface temperatures stay above 18°C, although some corals can survive temperatures as low as 11°C (Veron 2000). Stony corals (formally called *Scleractinia*) are used in paleoclimatology due to their many advantageous characteristics:

1. they grow continuously and can reach an age of 1000 years,
2. they are easy to date and can be sampled with a weekly to monthly resolution,
3. they incorporate elemental and isotopic tracers such as Sr/Ca (strontium-calcium) and $\delta^{18}O$ that can be used to reconstruct paleoceanography and climate,
4. their wide distribution throughout the tropics makes them easily obtainable,
5. it is fairly simple to interpret their proxy records (Grottoli & Eakin 2007).

Corals that are used for paleoclimatic studies are usually hermatypic, which means that the coral polyps live symbiotically with zooxanthellae (unicellular algae). These algae are photosynthetic and have specific demands on their environment since they depend on sunlight for their processes. As the coral polyps in turn are dependent upon the zooxanthellae for their food (organic carbon diffused by the algal cells), they live in environments suitable for photosynthesis. Corals thus tend to grow in depths between 0-20 m in warm waters, generally between the latitudes 30°N and 30°S (Bradley 1999).

Coral reefs are abundant in the Indo-Pacific warm pool, the Red Sea and the Caribbean, but can be found in other areas of the Pacific, Indian and Atlantic oceans (Corrège 2006). Corals in diverse areas of the world record various effects of El Niño and serve in the reconstruction of its different aspects. As El Niño originates in the equatorial Pacific, however, this is one of the more important areas to study for a good understanding of ENSO (Philander 1990).

3.2 Composition and growth

Polyps secrete an aragonite skeleton, which is what makes up the hard parts of the coral. The aragonite, a carbonate mineral consisting of calcium carbonate ($CaCO_3$), can also contain trace elements such as strontium (Sr), barium (Ba), magnesium (Mg), zinc (Zn), lead (Pb), uranium (U) and manganese (Mn). The Pb and U and make them dateable as far back as 400 000 years for fossil corals, and 40 000 years for modern ones (Bard et al. 1990). The exact mode in which corals incorporate trace elements into their skeleton is riddled with uncertainty and a subject for controversy. It is clear, however, that the development of corals is strongly connected to environmental factors, which is why they prove so valuable for climate change information (Corrège 2006).

Scleractinia corals can grow in different forms: branching, tabulate, columnar, foliaceous, encrusting, free-living or massive (Figure 4). Although all aforementioned types can provide information on geochemistry, massive corals are the most useful. Their colonies can grow as high as 10 m and they do not break or erode easily (Corrège 2006). They grow between a few millimetres to up to 3 cm per year, typically forming one high density and one low density growth band annually (Grottoli 2001), which can be used for schlerochronology, a dating technique comparable to dendrochronology in trees (Lough & Barnes 1997). Corals of the genera *Porites*, *Pavona* and *Montastraea* are the most commonly used for climate reconstruction (Grottoli & Eakin 2007).

Corals can potentially be disturbed by animals in their environment that destroy their skeleton by grazing or boring, but this can often be identified when studying them with instruments such as X-radiographs. Bleaching events, on the other hand, can be more problematic as they halt the growth of corals. This can happen when coral reefs are subjected to stress, such as significant changes in water temperature, salinity, water chemistry and oxygen starvation.

Fig. 4. Different forms in which corals can grow (from top left to bottom right): columnar, branching, encrusting, massive, free-living, tabulate and foliaceous. *Modified after Laboratoire Informatique & Systématique; Australian Institute of Marine Science; Kadalaut; University of Leeds; Hindustani Times.*

The zooxanthellae, which are what give corals their colour, are then either expelled or lose their pigmentation, leaving the coral reef white (Hoegh-Guldberg 1999). It can be difficult to recognize when these events have taken place and determine how long it took for the corals to regain their normal growth rate (Corrège 2006). Some El Niños have been known to augment coral growth and reef development, whereas others have seriously disturbed reefs in the eastern Pacific (Carriquiry et al. 1994).

3.3 Collection and methods

Corals are gathered from shallow waters, usually by a team of scuba divers. To obtain a continuous record of past climate from a coral, a core should be extracted with a drill along its major axis of growth, which means from the head or tip of the coral to its initial point of growth. The result will be a vertical core with horizontal growth bands (Figure 5). The core is then cut longitudinally into slices of 0.5-1 cm in thickness, which are then washed and dried (Grottoli & Eakin 2007). To reveal the skeletal density bands and determine growth rates, the dry coral is X-radiographed. Parts with unclear banding patterns that cannot be used to obtain growth rate information are pulverised to aragonite powder, which is then analysed for $\delta^{18}O$ with a mass spectrometer (Carriquiry et al. 1994).

4 $\delta^{18}O$ as proxy

4.1 Oxygen isotope ratios

Oxygen occurs as three stable isotopes, ^{16}O, ^{17}O and ^{18}O. The most common type is ^{16}O, which accounts for the vast majority of oxygen in the atmosphere and oceans. The heavy isotope ^{18}O only comprises a small part but is crucial to understanding paleoclimate. It is the ratio between ^{16}O and ^{18}O that is commonly measured and used in climate reconstruction (Brenninkmeijer et al. 2003). Different environmental factors determine the isotopic ratio of oxygen, making it far from constant and sometimes cumbersome to analyse. It has thus been necessary to establish a standard ratio in order to facilitate comparison between samples. The sample's deviation from the established standard $^{18}O/^{16}O$ is termed $\delta^{18}O$ (delta-O-18) and counted per mil (‰). If the $\delta^{18}O$ value is positive, the sample contains more heavy ^{18}O isotopes than the standard, and vice versa. To calculate the $\delta^{18}O$ of a sample, the following equation is used (Ravelo & Hillaire-Marcel 2007):

$$\delta^{18}O = [(^{18}O/^{16}O_{sample} - {}^{18}O/^{16}O_{standard}) / {}^{18}O/^{16}O_{standard}] \times 1000 \text{‰}$$

4.2 $\delta^{18}O$ in seawater

The standard $^{18}O/^{16}O$ used for water is defined as the Vienna Standard Mean Ocean Water (VSMOW). All $\delta^{18}O$ values for H_2O are thus reported relative to VSMOW, including seawater, groundwater, rain, ice and snow (Ravelo & Hillaire-Marcel 2007). Henceforth, the $\delta^{18}O$ in seawater will be referred to as $\delta^{18}O_{sw}$ in order to distinguish it from $\delta^{18}O_c$ in the coral carbonate skeleton.

The $\delta^{18}O_{sw}$ in the oceans varies both on a global and local scale. The chief factor that influences $\delta^{18}O_{sw}$ globally is the amount of water locked in ice caps. Precipitation and condensation play vital roles in regulating the oxygen isotope ratios and it is important to understand this mechanism. Since ^{16}O is lighter than ^{18}O, water molecules that contain light oxygen isotopes ($H_2{}^{16}O$) evaporate more easily than heavy molecules ($H_2{}^{18}O$) and as a result, water vapour and cloud droplets become ^{18}O-depleted (low $\delta^{18}O$ value) compared with seawater. Accordingly, $H_2{}^{18}O$ condenses more easily than $H_2{}^{16}O$, making the former more disposed to rain and to be removed from the clouds (Ravelo & Hillaire-Marcel 2007). As the remaining clouds travel from lower latitudes towards the poles they become increasingly depleted in ^{18}O, causing polar snowfall to have very low $\delta^{18}O$ values, usually ranging between -30 to -50‰. In fact, the lowest value recorded in our present climate was snow from Antarctica, reaching $\delta^{18}O$ values as low as -58‰ (Bradley 1999). In this fashion, during colder periods when more ^{18}O-depleted water is stored in ice sheets, the ^{18}O concentration in the world's oceans is higher than during warmer periods. For example, during the Last Glacial Maximum (LGM) the average $\delta^{18}O_{sw}$ in the global oceans was ~1.1‰ higher than today (Ravelo & Hillaire-Marcel 2007).

Additionally, local processes contribute to $\delta^{18}O_{sw}$ in oceans, and these values change on a shorter timescale than the aforementioned global aspects. Some of the factors that determine smaller-scale $\delta^{18}O_{sw}$ are temperature, salinity, local evaporation and freshwater input. As discussed, water that is evaporated has a relatively low $\delta^{18}O$ value. The remaining surface waters in high evaporation areas therefore have relatively high $\delta^{18}O$ values as well as increased salinity, since the latter also relates to evaporation and precipitation. This is usually true for tropical, low-latitude surface waters that receive abundant solar radiation and have a high evaporation rate. Areas with heavy rainfall generally become depleted in ^{18}O and salinity, which tend to be mid and high-latitude regions with less evaporation (Grottoli & Eakin 2007). Besides the evaporation/precipitation balance, other processes that affect local $\delta^{18}O_{sw}$ are sea-ice growth and melting, advection, upwelling and freshwater input from rivers, since the influence of other bodies of water with different $\delta^{18}O_{sw}$ values can alter the local oceanic conditions and the sea surface $\delta^{18}O_{sw}$ (Ravelo & Hillaire-Marcel 2007).

4.3 $\delta^{18}O$ in carbonates

When using corals to infer the oceans' $\delta^{18}O_{sw}$ in the past, it is necessary to use a different standard

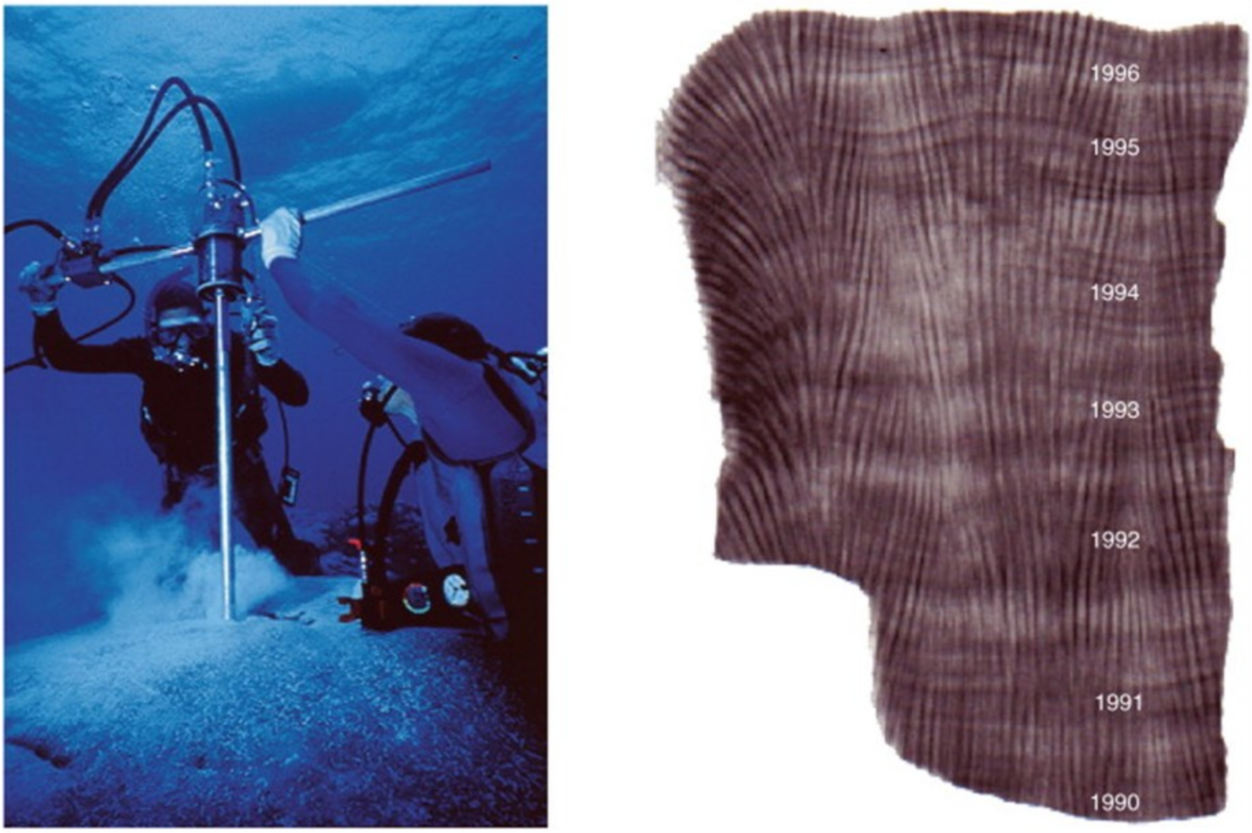

Fig. 5. **Left**: Scuba divers collecting coral cores with a drill. **Right**: A *Porites spp*. coral showing growth bands that have been dated from 1990 to 1996. *Modified after Grottoli 2001*.

for $\delta^{18}O_c$ than for water. The $\delta^{18}O_c$ of all carbonates is defined as the deviation of a sample's $^{18}O/^{16}O$ relative to the Vienna Pee Dee Belemnite (VPDB), a *Belemnitella americana* from the Peedee formation in South Carolina, USA. Oxygen isotope ratios in marine carbonates depend on two main factors: the $\delta^{18}O_{sw}$ in the surrounding waters and the thermodynamic isotopic fractionation between the water and carbonate (Bowen 1991).

The process of enrichment or depletion of ^{18}O (or any other isotope) is called isotope fractionation. When studying processes that involve changes of state of the same substance, such as water evaporation and condensation in the hydrological cycle, it is kinetic fractionation that is most relevant. By this process, as previously mentioned, heavier isotopes tend to be more concentrated in liquids whereas lighter isotopes are more abundant in gas and vapour because the latter move and react at a faster rate than the former (Zhou & Zheng 2003).

Oxygen isotope fractionation between aragonite (a type of carbonate) and seawater can be a complex matter since there are several factors that have to be taken into consideration, but the most important one is the temperature in which calcification occurs in water, for instance by aragonite secretion in coral skeleton (Epstein et al. 1953). If the two substances are in chemical or isotopic equilibrium ("a state in which a process and its reverse are occurring at equal rates so that no overall change is taking place"; Oxford English Dictionary) the process is called equilibrium fractionation. Assuming this is the case, fractionation between aragonite and seawater ($\delta^{18}O_c - \delta^{18}O_{sw}$), is inversely related to temperature (Ravelo & Hillaire-Marcel 2007). In summary, coral carbonate $\delta^{18}O_c$ depends on water $\delta^{18}O_{sw}$ combined with water temperature, and carbonates tend to become more ^{18}O enriched in colder water (Epstein et al. 1953).

Additionally, seawater salinity and pH as well as light intensity and coral growth rate can also influence coral $\delta^{18}O_c$ (Wang et al. 2013). While these have a lesser effect than temperature and can sometimes be considered next to negligible (with the exception of salinity which can in fact be a dominating factor in some cases due to its relation to $\delta^{18}O_{sw}$), they might play a pertinent role in very slow-growing corals and those that live in low light conditions (Grottoli & Eakin 2007). Calculations containing several different factors tend to become rather complex, which is inevitable since the aim is a holistic understanding of processes in nature. Nonetheless, if assuming that the seawater – aragonite fractionation only depends on the surrounding seawater temperature, the following estimates can be used: for every 1°C of seawater cooling, the $\delta^{18}O_c$ increases by 0.22‰ (Epstein et al. 1953).

In real world conditions the slope can vary depending on location, depth and what species of coral is used (Weber & Woodhead 1972). For example, in some eastern Pacific corals, $\delta^{18}O_c$ can decrease between 0.16‰ and 0.53‰ per 1°C increase, even within the same species (Wellington & Dunbar 1995). Moreover, measurements on the species *Porites lobata* show that a 1°C temperature increase, decreases the coral $\delta^{18}O_c$ by 0.19‰ in the Gulf of Panama (Wellington & Dunbar 1995), 0.22‰ in the Galápagos (Wellington et al. 1996) and 0.23‰ at Clipperton Atoll (Linsley et al. 1999). It is therefore essential to be aware of the uncertainties that may arise if one fails to consider the contributing factors when attempting to accurately reconstruct paleoclimatic conditions.

4.4 SST and SSS from $\delta^{18}O$

Since corals normally dwell in the upper layer of the tropical oceans, they are used for deducing sea surface conditions. The most important information for reconstructing ENSO is that of sea surface temperature (SST) and sea surface salinity (SSS). Pacific Ocean temperature anomalies and changes in rain patterns are the most direct consequences of El Niño episodes, both of which affect the $\delta^{18}O$ of surface water and hence of corals inhabiting the affected regions (Cole et al. 1992).

As rain is ^{18}O-depleted compared to the ocean, periods of heavy precipitation lead to a decrease of $\delta^{18}O_{sw}$ in the sea surface, as well as a reduced SSS due to the input of fresh water. According to Fairbanks et al. (1997), a salinity decrease of 1.0 is equivalent to a $\delta^{18}O_{sw}$ decrease of approximately 0.27‰, depending on latitude, depth and ocean basin. Corals that live in locations where SSS is relatively constant throughout the year are used to infer SST but if, on the other hand, SSS varies significantly it can come to dominate the $\delta^{18}O_{sw}$ signal. In cases where both SST and SSS are highly variable, the coral $\delta^{18}O_c$ interpretation is often complicated, especially if their relationship in unclear (Grottoli & Eakin 2007).

Awareness about changes in SST due to seasonal and interannual variability is vital when identifying El Niño or La Niña events in the coral record. If knowledge about currents and seasonal temperature changes is limited, a coral $\delta^{18}O$ signal that represents a temperature fluctuation unrelated to ENSO, can easily be mistaken for an anomaly. One must therefore be able to distinguish between normal sea surface conditions and those connected to El Niño, keeping in mind that its effects differ depending on geographic location, season and water depth among other parameters (Hillaire-Marcel & Vernal 2007).

As the case is that there are coral reefs in distinct locations equally spread across the Pacific in the ENSO core region, we have a source of concurrent data by which we can verify its related effects in order to obtain a holistic view of ENSO behaviour. Cole et al. (1992) studied three key coral sites that are sensitive to upwelling, SST, rainfall and winds across the tropical Pacific Ocean (Figure 6): Galápagos Islands (east Pacific, 1°S, 91°W), Tarawa Atoll (central Pacific, 1°N, 173°E) and Bali (west Pacific, 8°S, 115°E). During El Niño phases, SST and rainfall in the Galápagos Islands region increase while upwelling is su-

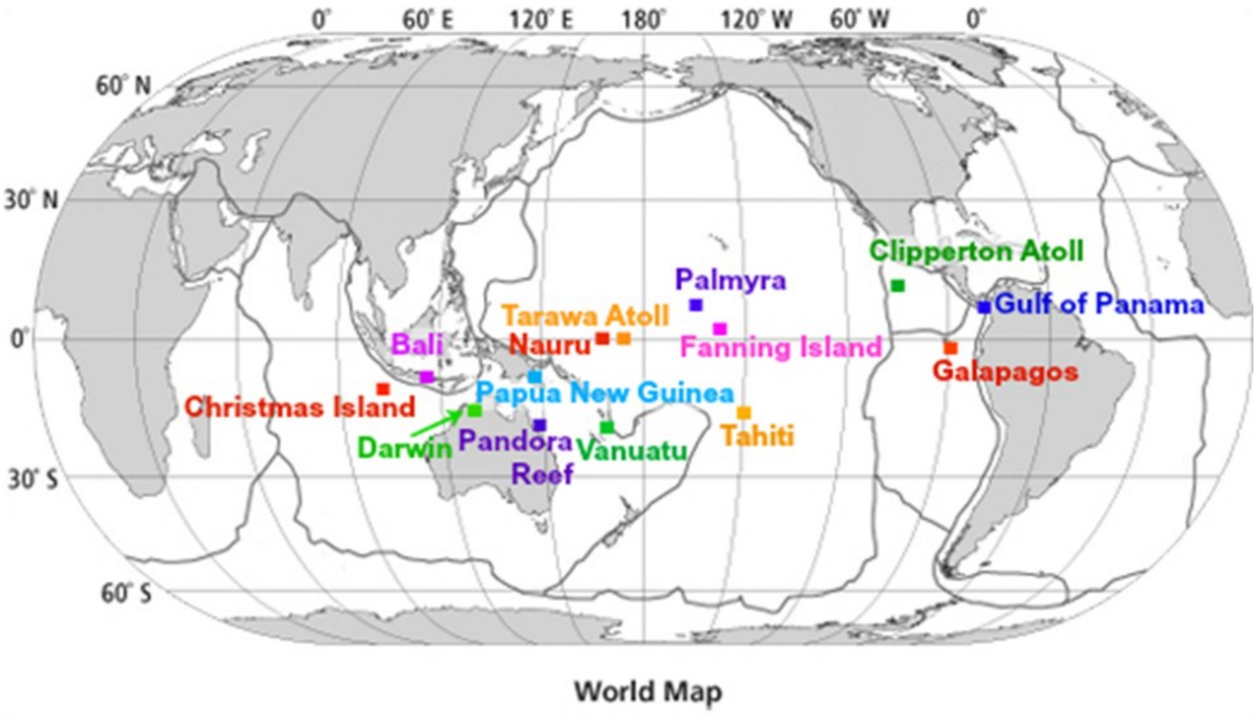

Fig. 6. A world map showing the locations mentioned. Note that Palmyra and Fanning Island are parts of the Line Islands, which also include Kiribati.

pressed. Corals here record an average SST rise of 1-2°C, which is reflected in its $\delta^{18}O$. Similarly, the Tarawa region sees intense rainfall, resulting in ^{18}O-depleted surface waters and corals, but the thermal response is slight. Bali, on the other hand, experiences drier conditions during El Niño, as the rain system migrates eastward. Moreover, cool SST anomalies occur together with more saline waters, and produce more positive $\delta^{18}O$ values in the corals of that area. Conversely, during La Niña conditions, temperatures around Bali rise as they decrease in Galápagos Islands (Cole et al. 1992). By understanding the climatic responses to the different phases of present-day ENSO in different geographic areas, we can verify and contrast past ENSO events as well as eliminate unrelated effects in the coral record.

Additionally, some factors that have not yet been discussed are also highly relevant to obtain the information necessary for paleo-ENSO reconstructions. Oxygen isotope data should be complemented with other proxies in order to achieve a well-rounded set of data and more accurate results. Coral skeletal strontium-calcium (Sr/Ca) ratios, for instance, are commonly used in conjunction with oxygen isotopes to help accurately decipher the code of past surface temperatures in the Pacific Ocean. Along with coral records, there are also numerous other paleoclimatic tools to study ENSO activity. Various marine carbonates such as foraminifera, ostracodes and molluscs can be used in a similar fashion as corals (Rosenthal 2007). Varves in sedimentary layers (Patterson et al. 2013), tree rings (Xu et al. 2013), ice-cores (Bowen 1991), fossil pollen and historical records can further be merged to give a multi-faceted history of El Niño (Mock 2007). The advantages of a multi-proxy approach will be further discussed in a later chapter.

5 El Niño in the paleoclimate record

5.1 Origin of El Niño

In 1986 it was proposed that El Niño started around 5 000 years BP (before present) (Rollins et al. 1986). Several later studies however, have shown that it originated much earlier. Fossil coral $\delta^{18}O$ and Sr/Ca records from the Huan Peninsula (Papua New Guinea; Figure 6) reveal that El Niño has been active for at least the last 130,000 years (Tudhope et al. 2001; Martínez 2009; Vásquez et al. 2010). The oxygen isotope records show variations in rainfall, which in this area occurs together with warmer temperatures, happening every 2-7 years as is typical for ENSO (Cane 2005).

Clement et al. (2001) argue that it is probable that ENSO has existed since 500 ka BP (ka = thousand years) and has had two interruptions, one around 412 ka BP and one during the Younger Dryas (ca. 12.8 ka BP – 11.5 ka BP). There are further indications from $\delta^{18}O$ in bivalves and foraminifera indicating that there might have been an even earlier origin, or perhaps permanent El Niño-like conditions (higher SST and decreased upwelling in the eastern Pacific) during the Pliocene ~3 Ma BP (Ma = million years). The lack of high-resolution marine proxies, however, makes it difficult to establish reliable evidence of seasonal changes during this time (Vásquez et al. 2010).

Though ENSO was active during the Last Glacial Maximum (LGM, ~26 – 12 ka BP), as is evident from coral and foraminifera $\delta^{18}O$ as well as other proxies, it was weakened during the glacial ages and probably displayed La Niña-like conditions (colder SST in the eastern Pacific) (Martínez 2009). It is in the Holocene (11.7 ka BP – present) that patterns in ENSO become easier to study using coral records, and resemble modern climatic trends (Carré et al. 2012). It should be noted that when studying climatic patterns far back in time the only available coral records are fossilized, and the investigations rely on the assumption that the geographic patterns of El Niño have remained similar over the last few thousand years (Cobb et al. 2003).

5.2 Early and mid-Holocene

Gagan et al. (2000) review tropical SST variations from fossil coral $\delta^{18}O$ and Sr/Ca measurements throughout the Holocene and conclude that temperatures in the tropical western Pacific Ocean surface were 5-6°C colder during the mid-stages of the last deglaciation (14 – 10 ka BP) compared to today. During the Younger Dryas (12.8 – 11.5 ka BP), normal SSTs in Vanuatu (western Pacific) were 4-6°C lower than at present until 10.3 ka BP. $\delta^{18}O$ measurements show that temperatures then rose for the next 4000 years, reaching 1.2°C warmer than present at 6 ka BP and cooling to modern temperatures by 4 ka BP. Although these values in themselves do not give direct information about ENSO since they do not show annual or seasonal variations, they do demonstrate that temperatures in Pacific surface waters have fluctuated during the Holocene, which has possibly had an effect on or been a result of the fluctuations in strength and amplitude of El Niño (Gagan et al. 2000). As the early Holocene marks the end of the last glacial period, the shift in global climate is likely to have been a trigger for changes in ENSO. Ice reduction, sea level rise and temperature increase had a considerable impact on ocean circulation (Markgraf & Diaz 2000).

McGregor & Gagan (2004) use fossil *Porites* coral $\delta^{18}O$ from Papua New Guinea in the Western Pacific Warm Pool to investigate climate during the mid- to late Holocene. Since this area under normal conditions has a warm surface ocean and high rainfall it is closely linked to ENSO variability, as dry conditions coupled with reduced SSTs indicate El Niño events (Tudhope et al. 2001). The eight corals studied have annual resolutions and make it possible to reconstruct ENSO quite precisely. In order to define moderate and strong El Niño events in the mid-Holocene, the authors compared the results from the fossil corals to the average modern $\delta^{18}O$ anomaly for El Niños 1950-1997, which is 0.18‰ ± 0.06‰. The $\delta^{18}O$ threshold for moderate El Niño events in the fossil record was thus defined as 0.12‰ (0.18‰ minus 0.06‰). Values higher than 0.24‰ (0.18‰ plus 0.06‰) were defined as strong to very strong. The coral records were classified into three periods: 7.6 – 7.1 ka, 6.1 – 5.4 ka and 2 ka BP (McGregor & Gagan 2004).

During the early periods El Niño events seem to have been less frequent than at present. From 7.6 – 7.1 ka BP, there was an average of 12 events per century and 6.1 – 5.4 ka BP shows an average of 8 events per century (Table 1). These are significantly low numbers compared to the average of 23 events in the 20th century (as calculated from modern coral $\delta^{18}O$ in the same region). Furthermore, the amplitude or strength of El Niño was reduced to approximately 85% of modern values in 7.6 – 7.1 ka and 6.1 – 5.4 ka BP (McGregor & Gagan 2004). According to another study of Papua New Guinean fossil coral $\delta^{18}O$, the least tendency for extreme or high amplitude El Niño events is at ~6.5 ka BP (Tudhope et al. 2001). A 4300-year-old fossil *Porites* coral from Christmas Island continuously records a 175-year long period with monthly resolution by $\delta^{18}O$. It indicates that El Niño was less frequent in 4.3 ka BP than presently, occurring every 6-8 years (~14 events/century), had weaker anomalies and also had a delayed seasonal characteristic, starting in September-November and peaking in February.

The weakness and low frequency of El Niño in the mid-Holocene has also been confirmed by other proxies. For example, sedimentary deposits in Peru record no severe flood or debris flow events from 8.4 – 5.3 ka BP (Keefer et al. 2003). The overall reduction in ENSO seems to derive from changes in insolation due to the orientation of the Earth. An altered tilt of its axis affected ocean heating in the Pacific, and additionally increased warming during the northern summer months in higher latitudes, which is thought to have strengthened trade wind circulations, thus supressing El Niño (McGregor et al. 2013).

Table. 1. A summary of the frequency of El Niño events during the early to mid-Holocene and during the last millennium, as described in chapters 5.2 and 5.3. A trend towards closer (more frequent) events can be seen from the 16th century to the present. The 20th century experienced 23 El Niños, an average of ~4.3 years between events.

Period	**Average time between events (years)**
7600 – 7100 BP	8.3
6500 BP	16.7
6100 – 5400 BP	12.5
4300 BP	6 – 8
1100 – 1200 AD	16.7
1300 – 1500 AD	14.3
1500 – 1600 AD	5.9
1600 – 1700 AD	4.6 – 7
1700 – 1750 AD	3 – 4.6
1800 – 1850 AD	3.5
1900 – 2000 AD	4.3

5.3 Late Holocene

After the mid-Holocene the frequency of El Niño increased steadily. The amplitude however, has been consistently reduced during the mid to late Holocene. In 3.7 – 2.7 ka BP the anomalies of El Niño events indicate 70-80% of modern values. Nevertheless, longer and more severe El Niños were common during 2.5 – 1.7 ka BP reaching twice the amplitude of the strong 1997-1998 event in 1.7 ka BP and lasting as long as 7 years in 2 ka BP. This is longer than any other El Niño event recorded in the Holocene or modern time (McGregor & Gagan 2004).

Cobb et al. (2003) present ENSO reconstructions of the last millennium using fossil *Porites* coral δ^{18}O from Palmyra Island (6°N, 162°W, part of the Line Islands). In this region conditions are warmer and wetter during El Niño, which results in more negative coral δ^{18}O. During La Niña colder and drier conditions persists which results in a more positive coral δ^{18}O. Several of the coral records overlap and they were dated with U/Th to five intervals: AD 928 – 961, 1149 – 1220, 1317 – 1464, 1635 – 1703 and 1886 – 1998. The δ^{18}O values show that ENSO has had an ample variation in strength and frequency over the last millennium (Figure 7). In the 17th century ENSO was stronger and more frequent than in the late 20th century while both the 12th and 14th centuries saw a greatly reduced activity (Cobb et al. 2003). Furthermore, in a later study Cobb et al. (2013) analysed fossil corals from Fanning Island (4°N, 160°W) in the northern Line Islands spanning the last 7000 years and conclude that ENSO strength has varied significantly throughout the Holocene, being strongest in the 17th and 20th centuries (Cobb et al. 2013).

Fossil corals provide good records of Pacific climate before 1600 AD but when reconstructing climate variations of the last few centuries modern corals are more often used than fossil ones due to their higher resolution and more accurate results (Cobb et al. 2003). Grottoli & Eakin (2007) present coral δ^{18}O from the Pacific, Atlantic and Indian oceans in order to reconstruct ENSO from the mid 1800s until present. From 1860 to 1990, Pacific coral δ^{18}O decreased by 0.17‰, corresponding to a ~0.79°C warming in SST. Between the same years corals from the Red Sea and Indian Ocean had a δ^{18}O decrease of 0.15‰, which corresponds to ~0.68°C warming, while further south in the Indian Ocean the decrease in δ^{18}O is 0.82‰, equivalent to an SST increase of ~3.5°C. This warming trend is also seen in other areas, most dramatically in Nauru (0.5°S, 167°E) which displays a δ^{18}O decrease of 1.1‰, a warming of up to ~5°C. Grottoli & Eakin argue that ENSO has been a dominating pattern in the Pacific for hundreds of years and that it is possible that the frequency of El Niño has increased during the 20th century. This research further shows that ENSO has a great effect on areas far from the tropical Pacific, for example corals in the Red Sea, which have an important coherence with patterns of El Niño (Grottoli & Eakin 2007).

In the central Pacific region warm and wet conditions, in the form of heavy rainfall and higher SSTs, during El Niño cause low skeletal δ^{18}O. Coral records from the island of Kiribati (1°N, 172°E) and other Line Islands (~160°W) show that ENSO had reduced amplitude in the mid 1900s but stronger seasonal cycles (Gagan et al. 2000), while Clipperton Atoll corals reveal a period of lower frequency prior to that (1925 – 1940) (Grottoli 2001). These significant shifts in ENSO frequency dominate the last 300 years, as measured from a Galápagos coral. During the 1600s the time between cycles was 4.6 – 7 years. 1700 – 1750 saw an increased frequency, when El Niño events occurred as close as 3 – 4.6 years from each other, continuing at 3.5 years from 1800 – 1850. These findings indicate important changes in Pacific climate over the last three centuries (Grottoli 2001).

Coral δ^{18}O from the Pandora Reef (east coast of Australia) shows that SSTs in the area typically range between 21.5 and 28°C throughout the year during the period 1978 – 1984. An exception to this was in 1982 during El Niño when temperatures fell to 18.5°C, a significant anomaly that represents a very strong El Niño. In fact, the 1982 – 1983 event was a highly important climatic event of the 20th century (McCulloch et al. 1994), perhaps challenged only by the 1997-1998 El Niño, arguably the biggest event of the 20th century by several measures (Trenberth 2001).

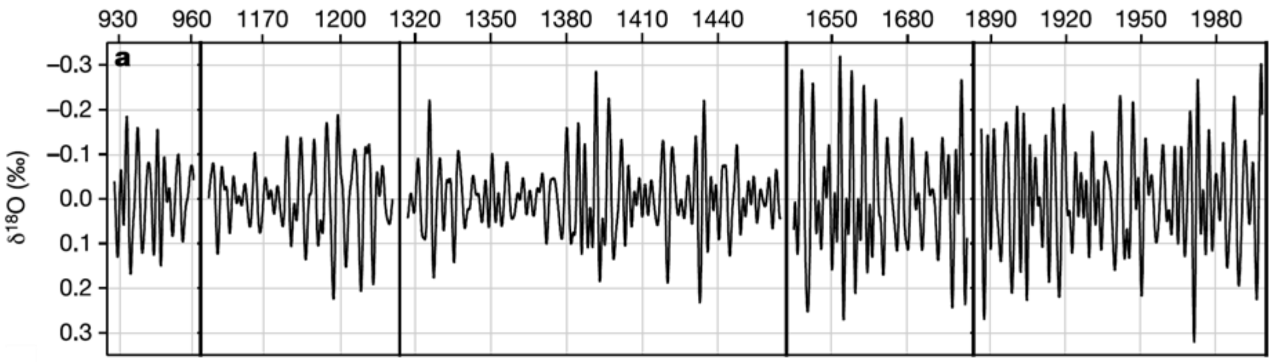

Fig. 7. Palmyra coral δ^{18}O anomaly values. Values less than –0.11‰ (i.e. above the –0.11‰ line) are defined as El Niño events and values more than 0.11‰ (i.e. below the 0.11‰ line) are defined as La Niña events. *Modified after Cob et al. 2003.*

If ENSO did indeed commence 130 ka BP, or at least take the form similar to how we now know it, its birth coincides with the start of the Eemian interglacial period. The fact that ENSO activity seems to pick up strength during interglacial and weaken during ice ages, suggests that warmer climate promotes stronger oscillations. It is thus plausible that the 21st century, having so far been warmer than any other recorded period ("GISS Surface Temperature Analysis (GISTEMP)", NASA) could see an increase in ENSO activity and perhaps stronger repercussions than in the past. The many factors that are involved in strengthening or suppressing El Niño activity, however, make it difficult to accurately analyse the past and certainly to predict its long-term future. Climate is undoubtedly a dynamic system with many parameters determining its development and it is therefore necessary to consult several sources before concluding how and why ENSO activity has varied.

5.4 ENSO predictability

Short-term prediction of El Niño is possible and often precise since ENSO dynamics are relatively well understood. Measurements of ocean and atmospheric pressure and temperature interactions facilitate the construction of forecast models (Chen & Cane 2008). The Pacific Ocean is regularly observed in order to detect trends toward El Niño-like conditions, mainly in the form of altered currents and rising SSTs in the eastern Pacific. As early as November 2013, the CPC (Climate Prediction Center) and IRI (International Research Institute) predicted that the possibility for El Niño to occur the following year was elevated compared to previous years (L'Heureux 2014). In March 2014 an "El Niño-watch" was announced after having observed an eastward migration of warm Pacific waters and increasing SSTs (Becker 2014).

By May 8th, the chance of El Niño developing in late autumn or early winter was as high as 80%. There are, however, still uncertainties due to the many factors that determine the development of El Niño (L'Heureux 2014), but the rise of a super-El Niño at the end of 2014 is nonetheless a real possibility as conditions similar to pre-1997 El Niño play out in the Pacific ("Is El Niño Developing?", NASA Earth Observatory). Although forecast models differ in their prediction of El Niño, they agree on that there is a trend towards warm anomalies that is most likely to continue for the rest of the year (Figure 8).

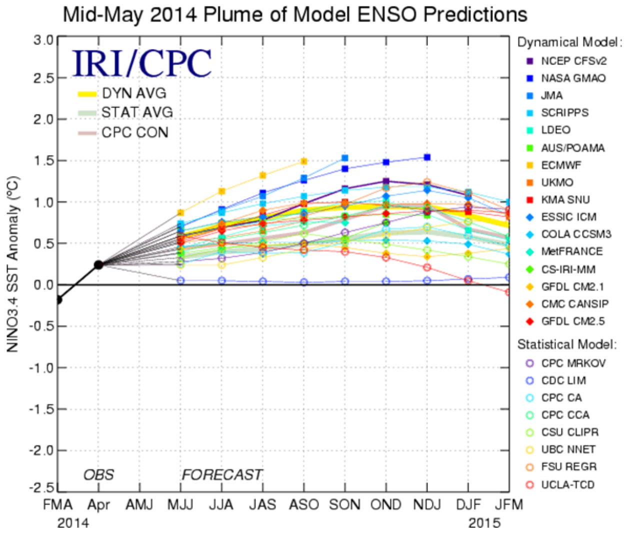

Fig. 8. 2014 prediction models of SST anomalies in the NINO3.4 region (5°S – 5°N, 120°W – 170°W), for nine overlapping 3-month periods (AMJ = April, May June; MJJ = May, June, July and so on) including both dynamical and statistical models. The yellow line represents the average SST anomalies of the dynamical models and the green line represents the average of the statistical models. The average of all models combined are as follows: MJJ: 0.5°, JJA: 0.6°, JAS: 0.7°, ASO: 0.8°, SON: 0.8°, OND: 0.8°, NDJ: 0.8°, DJF: 0.7°, JFM:0.6°. The differences between the models reflect both design dissimilarities and forecast uncertainties. *Source: IRI/CPC*

6. Limitations and possible improvements

6.1 Other paleoclimate tracers in corals

The information that coral $\delta^{18}O$ can provide about SST and SSS is certainly valuable for reconstructing ENSO, but the importance of a multi-proxy approach must be emphasized. In addition to oxygen isotopes, other tracers in corals such as trace elements and minor element compositions are also edifying. When combined they can yield a particularly comprehensive perspective on paleoceanography and paleoclimate.

Sr/Ca compositions in coral skeleton are reliable tracers of SST and are commonly combined with $\delta^{18}O$ to produce more accurate results (Grottoli 2001). Like $\delta^{18}O$ it correlates negatively with temperature (Rosenthal 2007) and by comparing the results from the two proxies, errors can be decreased as Sr/Ca helps solve problems that arise from uncertainties involving $\delta^{18}O$ in seawater and carbonate (McCulloch et al. 1994). Mg/Ca and U/Ca ratios serve the same purpose and can be used to verify temperature records in the case that Sr/Ca or $\delta^{18}O$ are not accurate enough. In a study of SSTs in the South China Sea from corals, Wei et al. (2000) conclude that the Sr/Ca and U/Ca ratios do not provide valid information of temperatures in that area, while Mg/Ca proves reliable for SST reconstruction due to its lack of sensitivity to environmental factors other than temperature.

Carbon isotope ratios are also sometimes used in combination with oxygen isotopes. $\delta^{13}C$ (a measure of $^{13}C/^{12}C$ ratios relative to the Vienna Pee Dee Belemnite; Carriquiry et al. 1994) is typically used to infer light intensity and nutrient availability as it is strongly related to the amount of zooxanthellae in symbiosis with the corals. Higher light levels lead to a higher rate of photosynthesis by zooxanthellae, which results in increased coral skeletal $\delta^{13}C$ values. ^{13}C-enriched corals thus suggest abundant sunlight. On the other hand, zooplankton tend to be ^{13}C-depleted, so when nutrient rich waters upwell, zooplankton levels increase and coral $\delta^{13}C$ values become lowered in the upwelling region (Grottoli 2001).

According Cole et al. (1992) the decrease in nutrients when upwelling is halted in the eastern Pacific during El Niño is also reflected in Galápagos corals' nutrient tracers cadmium (Cd) and barium (Ba), which increase in concentration with increased nutrient-rich upwelling water. Changes in manganese (Mn) concentrations in coral reefs north of Tarawa (Figure 6) also suggest diminished upwelling as well as trade wind reversal, a tell-tale sign of El Niño (Cole et al. 1992). Hence, since ENSO is strongly related to changes in upwelling in the western coast of South America, $\delta^{13}C$, Cd, Ba and Mn values of corals in that region can be used to help infer when events have taken place.

Low $\delta^{13}C$ values are moreover associated with stress in the coral environment, sometimes resulting in the expulsion of zooxanthellae, bleaching events and growth anomalies (Carriquiry et al. 1994). The knowledge of when these events have taken place can be greatly beneficial in order to avoid erroneous dating of corals and thus unreliable conclusions about the timing of ENSO occurrences. A list of proxies from corals and the environmental variable that they record can be seen in Table 2.

Table 2. Environmental variables that can be reconstructed from coral skeletal isotopes, minor elements, trace elements and growth record. Some variables, such as SST, SSS and upwelling associate more directly to ENSO. *Modified after Grottoli 2001*.

Proxy	Environmental variable
Corals:	
$\delta^{18}O$	Sea surface temperature, sea surface salinity
$\delta^{13}C$	Light intensity, nutrients/ zooplankton levels
$\Delta^{14}C$	Ocean ventilation, water mass circulation
Sr/Ca	Sea surface temperature
Mg/Ca	Sea surface temperature
U/Ca	Sea surface temperature
Mn/Ca	Wind anomalies, upwelling
Cd/Ca	Upwelling
$\delta^{11}B$	pH
F	Sea surface temperature
Ba/Ca	Upwelling, river outflow, sea surface temperatures
Skeletal growth bands	Light (seasonal changes), stress, water motion, sedimentation, sea surface temperature
Fluorescence	River outflow

6.2 Other paleoclimate proxies

If possible, corals isotope records should be combined with instrumental measurements. The case is, however, that the latter stretch back little more than a century. Nevertheless, in order to estimate the accuracy of coral $\delta^{18}O$, studies have been performed where modern corals from the 20th century are paired with instrumental measurements (Figure 9). The discrepancy between the two has then been established as the expected error from modern corals that have no instrumental counterparts (Cobb et al 2003). The fact remains, however, that $\delta^{18}O$ can differ by up to 0.3‰ (equivalent to a 1.5°C temperature uncertainty) between corals within the same area. Although the reasons for this are not clearly understood, a plausible explanation is that some of the sampled corals have grown in locations that do not represent regional climate, such as lagoons or sites with poor water mixing. Significant uncertainties can also arise if the corals are not properly sampled, or if the sampled coral skeleton has undergone post-depositional chemical changes (diagenesis). This is particularly common in fossil corals and can result in uncertainties equivalent to 1°C (Cobb et al. 2008). The aforementioned methods of using multiple proxies serve to reduce uncertainties such as these.

Further information about El Niño can be obtained by records other than corals, such as foraminifera, ostracodes and molluscs. These are used for paleothermometry in a similar fashion, as their shells have a composition comparable to that of coral skeleton (Rosenthal 2007). Patterson et al. (2013) found that ENSO teleconnections have a considerable effect in the northeast Pacific. By studying laminated sediments in Vancouver Island of British Columbia (Canada), they found that ENSO cyclicity can be observed by analyzing when a high number of a certain diatom species (unicellular algae) were present in the sediments, as these occurrences are attributed to warm El Niño events. Sediments can be used for a variety of reasons connected with the ENSO cycle, e.g. to identify when heavy rainfall or droughts have taken place in areas sensitive to precipitation changes, as Keefer et al. (2003) describe in their study of debris flow and flood deposits in Peruvian sediments.

Similarly, tree ring cellulose $\delta^{18}O$ yields evidence of abundant rainfall and dry seasons connected with El Niño, as wetter phases typically correspond with El Niño events if they occur in South America, and normal conditions or La Niña if they occur in Australia and Southeast Asia (Xu et al. 2013). Additionally, many other climate proxies can be used together with coral $\delta^{18}O$ to accurately reconstruct ENSO activity in the past.

7. Conclusions

Modern corals from the tropical Pacific ex-

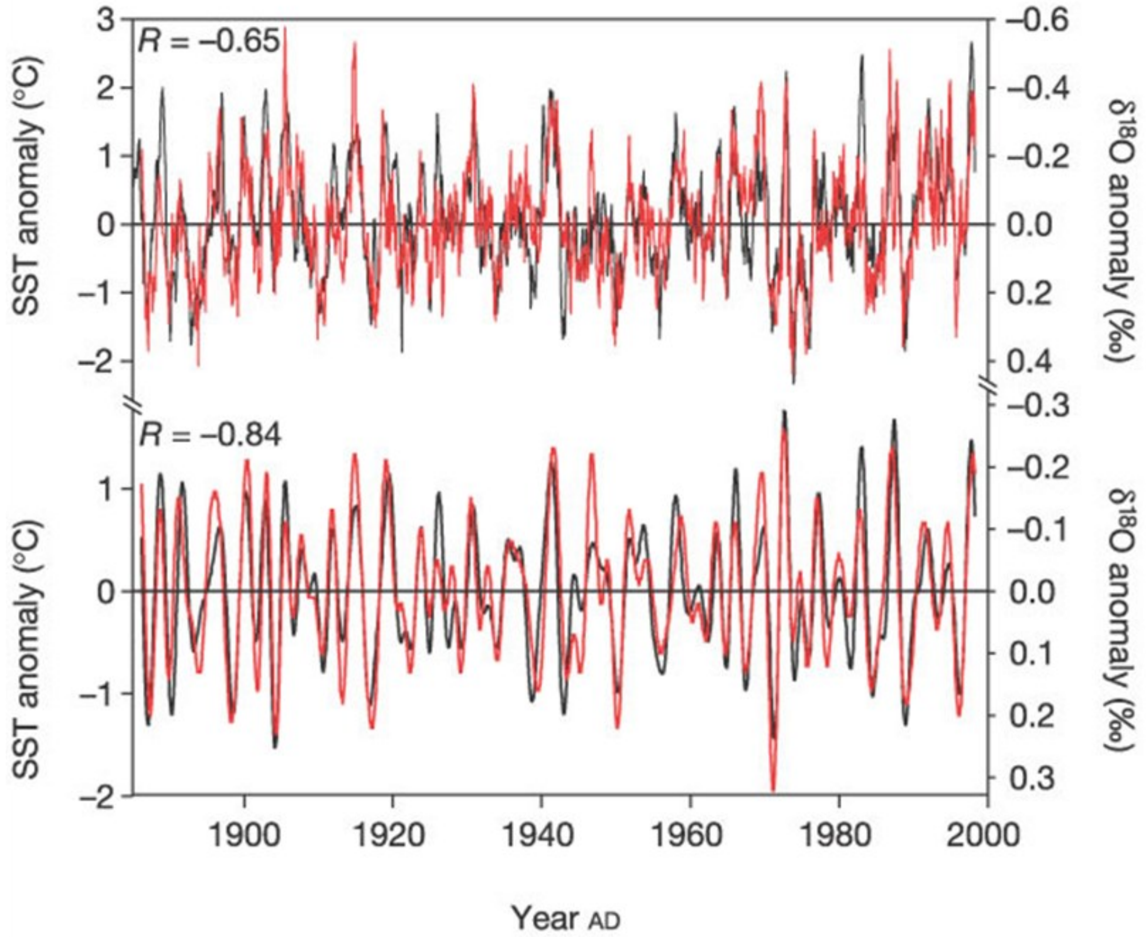

Fig. 9. Comparison of coral-based and instrumental records of central tropical Pacific climate during the 20th century. Top: Palmyra modern coral $\delta^{18}O$ anomalies (red) plotted with NIÑO3.4 SST anomalies (black; the average of SST anomalies from 5°N to 5°S, 120°W to 170°W). Bottom: same as top but a 2-7 yr bandpass filter has been applied to the record shown above. The Palmyra coral shown shares 72% of its interannual variance with the instrumental measurements, showing that corals from this site provide reliable proxies of ENSO activity. "R" refers to the relationship between the mean coral $\delta^{18}O$ and ENSO variance. *Source: Cobb et al. 2003*.

tend several centuries and can provide a high-resolution and continuous record of recent climate. In order to determine the climate in the distant past, fossil corals yield the necessary information in a similar way. Unfortunately, fossil corals of a significant length are not common and it can therefore be difficult to reproduce climate records. Moreover, their climate variability for a given period may not be as precise as for modern corals. Therefore, to reconstruct ENSO in the last few centuries, modern corals are preferred and give good results.

$\delta^{18}O$ values of fossil and modern corals serve as a reliable proxy for sea surface temperature, sea surface salinity and precipitation patterns, variables that are closely connected to the ENSO cycle. As corals can be accurately dated, the $\delta^{18}O$ variations in their skeleton reveal when El Niño and La Niña events have occurred as well as their amplitude. Low $\delta^{18}O$ values in eastern Pacific Ocean corals relate to increased SSTs, which is a sign of El Niño-like conditions. There are however, numerous variables that determine the coral $\delta^{18}O$, both on a global and regional scale, and care must therefore be taken to accurately evaluate the results.

In order to obtain a multi-faceted reconstruction of the past, it is preferable to use several different proxy methods. This decreases uncertainties and provides a better understanding of causes and effects both in the ENSO core region and its repercussions in other geographic areas through teleconnections. Though coral $\delta^{18}O$ is beneficial to deduce SST and SSS, other complementary tracers are commonly used, e.g. Sr/Ca. Furthermore, other proxies such as sediments, tree rings and marine organisms other than corals, also record events that can be related to ENSO.

Fossil coral records suggest that El Niño has been active for at least 130,000 years but could have an earlier origin. ENSO activity tends to be weaker during glacial eras and pick up strength when entering interglacials. In the Holocene El Niño has varied considerably in both frequency and amplitude, possibly as a result of earth axis and insolation changes leading to altered atmospheric and oceanic circulation patterns. Since the 16th century El Niño frequency has escalated, and the 20th and 21st centuries have seen an increased warming and seemingly stronger El Niño events than in most other periods on record. In fact, predictions have been made about an imminent El Niño at the end of 2014, which could reach strengths similar to the super-El Niño of 1997.

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