The phenomenon of El Niño
Consequences of El Niño
Climate records through isotope proxies
El Niño Southern Oscillation ENSO

ENSO is known to be a natural oscillation of the ocean-atmosphere system with El Niño as the ocean component of the interaction between the upper layers of the tropical Southern Pacific and the atmosphere. ENSO appears as enhanced sea surface temperature in the Easter Pacific near the Peruvian coast. The water temperature is usually dominated by the cold water flow of the Humboldt Current. The ENSO effect enhances temperatures and reduces biological productivity. The effect is observed around Christmas about every seven years and was therefore dubbed El Niño (the son) by the local fishermen.

The Southwestern Pacific hot ocean water region is typically contained by trade winds blowing from the east. These winds change directions every 7 years expanding the warm water zone eastwards.

L corresponds to low atmospheric pressure, H to high atmospheric pressure.
Normally low pressure region over Australia and Indonesia delivering rainfall, in El Nino years, there is high pressure over these regions and the low pressure area moves east and rainfall occurs over the Pacific Ocean.

The Southern Oscillation Index SOI is defined in terms of the Mean Sea Level Pressure $P$ between Tahiti and Darwin, Australia. SOI indicates El Nino episodes:

$$SOI = 10 \cdot \frac{(P_{Tahiti} - P_{Darwin})_{\text{month}} - (P_{Tahiti} - P_{Darwin})_{\text{mean}}}{\delta(P_{Tahiti} - P_{Darwin})_{\text{month}}}$$

**ENSO:** $P_{Darwin} > P_{Tahiti}$ SOI negative

**Non ENSO:** $P_{Darwin} < P_{Tahiti}$ SOI positive

Computed at a monthly basis SOI ranging from -3 to +3

http://www.ncdc.noaa.gov/teleconnections/enso/indicators/soi.php
There is a directly observable correlation between the SOI and the water surface temperature SST of the coast of South America!

La Niña is the name chosen for a situation where strong trade winds confine the hot water condition to the southwestern region of the Pacific.
1. The trigger of the El Niño event are recurrent spontaneous statistical fluctuations sea surface temperature reflecting different oscillation modes in the coupled ocean SST system.

2. This SST fluctuation causes a spontaneous reduction of the the east west pressure gradient in the atmosphere.

3. Changed atmospheric pressure gradient reduces wind (collapse of trade winds as already observed by Darwin on his journey on the Beagle) and wind driven drag of water westwards.

4. Reduced drag allows warm water to expand eastwards preventing cold water up-welling from Humboldt current.
The ENSO occurrence has global consequences, it changes rainfall patterns in Africa and North America and causes droughts in Australia. The ENSO event is recorded by tree ring, ice core, and coral reef analysis for millennia.

ENSO driven draught in Australia

ENSO driven rainstorm in Southern California
El Niño
There are statistics based arguments that the number of Tornados in the US tornado alley is correlated with El Niño and La Niña conditions in the Southern Pacific Ocean.
Climate Records from isotope and/or molecule distributions
The isotope chart

Each element with atomic number $Z$ (number of protons) has a number of stable and beta-unstable isotopes which are distinguished by a different number of neutrons $N$

- **Isotope**: $Z=\text{constant}$
- **Isobar**: $A=\text{constant}$
- **Isotone**: $N=\text{constant}$
Mass differences

\[ M^{(18}O) = 17.99916 \text{amu} \cdot 1.66 \cdot 10^{-24} \text{g} = 2.987860 \cdot 10^{-23} \text{g} \]

\[ M^{(16}O) = 15.99493 \text{amu} \cdot 1.66 \cdot 10^{-24} \text{g} = 2.655158 \cdot 10^{-23} \text{g} \]

\[ \Delta M = M^{(18}O) - M^{(16}O) = 3.327 \cdot 10^{-24} \text{g} \]

\[
F_{\text{Lorenz}} = m \cdot v \times B = \frac{m \cdot v^2}{r} = F_{\text{centrifugd}} \quad E_{\text{kin}} = \frac{1}{2} m \cdot v^2 = q \cdot U
\]

\[ r = \frac{m \cdot v}{q \cdot B} \quad v = \sqrt{\frac{2 \cdot q \cdot U}{m}} \quad \Rightarrow \quad r = \frac{m \cdot v}{q \cdot B} \sqrt{\frac{2 \cdot q \cdot U}{m \cdot B^2}} = C_{\text{separator}} \cdot \sqrt{\frac{m}{q}} \quad C_{\text{separator}} = \frac{\sqrt{2 \cdot U}}{B}
\]

\[
\frac{r^{18}O}{r^{16}O} = \sqrt{\frac{m^{18}O}{m^{16}O}} = 1.0608
\]
Variation and fractionation of isotopic abundances

The chemical behavior of the isotopes of a certain element are identical because the chemical interactions are controlled by \( Z \), the number and configurations of the electron in the atomic shell.

However, small differences due to physical conditions can lead to a deviation from the characteristic isotopic abundance ratio by fractionation processes. These variations are extremely slight, typically in the range of 0.01‰ to 0.1‰ and require sensitive equipment to be detected. The fractionation is expressed in terms of \( \delta \).

\[
\delta = 1000 \cdot \left( \frac{\text{sample isotope ratio} - \text{standard isotope ratio}}{\text{standard isotope ratio}} \right)
\]

\[
\delta^{13}C = 1000 \cdot \left[ \frac{^{13}C}{^{12}C}_{\text{sm}} - \frac{^{13}C}{^{12}C}_{\text{st}} \right]
\]

The isotope ratios are expressed with the heavier component in the numerator; \( \delta \) is a relative deviation from the standard ratio in units parts/mil or ‰!
Oxygen has three stable isotopes, $^{16}\text{O}$, $^{17}\text{O}$, $^{18}\text{O}$ with the average abundances of 99.756%, 0.039%, and 0.205%, respectively. The $^{16}\text{O}/^{18}\text{O}$ abundance ratio in a Jurassic limestone is 472.4335 the ratio in ocean water is 486.594. The fractionation can be determined with respect to the average abundance ratio (486.6146) or to a standard. If taking the ocean water as standard (standard mean ocean water, SMOW), the fractionation $\delta^{18}\text{O}$ in Jurassic lime stone is:

$$\delta^{18}\text{O}_{\text{SMOW}} = 1000 \cdot \left( \frac{\frac{^{18}\text{O}}{^{16}\text{O}}}{\text{sm}} - \frac{\frac{^{18}\text{O}}{^{16}\text{O}}}{\text{st}} \right) = 1000 \cdot \frac{0.002117 - 0.002055}{0.002055} = 29.97\%$$

The positive number indicates that the sample has a higher heavy isotope abundance as the standard. If the sample has a lower heavy isotope abundance the fractionation value $\delta^{18}\text{O}$ will be negative. Ocean water in Maxwell Bay, Antarctica has:  $\delta^{18}\text{O}_{\text{SMOW}} = -3.1$

$$\frac{^{18}\text{O}}{^{16}\text{O}}_{\text{MB}} = \delta^{18}\text{O}_{\text{SMOW}} \cdot \frac{^{18}\text{O}}{^{16}\text{O}}_{\text{SMOW}} + \frac{^{18}\text{O}}{^{16}\text{O}}_{\text{SMOW}} = -3.1 \cdot 0.002055 + 0.002055 = 0.002049$$
An extensive data bank on the $^{18}$O isotope distribution has been developed worldwide. It is useful for a wide range of applications, ranging from hydrology to forensics, from plant ecology to meteorite analysis, and from geology to climate studies.
Fractionation processes

There are a number of chemistry specific fractionation processes for the various elements. For oxygen, the chemical processes are mostly associated with geological fractionation observed for rocks, lime stones and carbonates. For plant, water, and ice related oxygen isotope analysis, the dominant fractionation processes are mostly associated with evaporation and precipitation processes. Therefore the oxygen fractionation is frequently correlated with hydrogen fractionation of the two table hydrogen isotopes, hydrogen $^1$H and deuterium $^2$H.
Chemical or physics processes for two different isotopes show subtle differences in the specific reaction rate $k$ for the lighter isotopes $k_L$ and the heavier isotopes $k_H$. The ratio of these rates defines the fractionation factor $\alpha$.

$$\alpha = \frac{k_H}{k_L}$$

Chemical or physical processes with light isotopes are typically faster than the ones involving heavier isotopes: $k_H < k_L$, $\alpha < 1$. This results in a fractionation of isotopes in the chemical process with an enrichment factor $\varepsilon = \alpha - 1$.

For evaporation:

$$\alpha = \frac{\delta_{product} + 1000}{\delta_{source} + 1000} \quad \delta_{product} = \alpha \cdot \delta_{source} - 1000 \cdot (1 - \alpha)$$

$k_{18O}$ is $\approx 1.2\%$ slower than $k_{16O}$

$$\alpha_{18O} = \frac{k_{18O}}{k_{16O}} = \frac{1 - 0.012}{1} = 0.988$$

$$\delta_{vapor} = \alpha \cdot \delta_{sea\ water} - 1000 \cdot (1 - \alpha) = -12\%$$
Fractionation in equilibrium

At isotopic equilibrium between two phases (e.g. liquid vapor), the forward and backward reaction rates of any particular isotope are equal. The isotopic compositions of two phases at equilibrium are not identical, but the ratios of the different isotopes in each phase are constant for a particular temperature. As water vapor condenses in rain clouds (equilibrium process), the heavier water isotopes ($^{18}$O and $^2$H) become enriched in the liquid phase, while the lighter isotopes ($^{16}$O and $^1$H) remain in the vapor phase.

\[
\alpha_{\text{Vapor} \leftrightarrow \text{Liquid}} = \frac{\frac{^{18}\text{O}}{^{16}\text{O}}_{\text{Vapor}}}{\frac{^{18}\text{O}}{^{16}\text{O}}_{\text{Liquid}}} \quad \alpha_{\text{Liquid} \leftrightarrow \text{Vapor}} = \frac{\frac{^{18}\text{O}}{^{16}\text{O}}_{\text{Liquid}}}{\frac{^{18}\text{O}}{^{16}\text{O}}_{\text{Vapor}}} \quad \text{and similarly} \quad \alpha_{\text{Vapor} \leftrightarrow \text{Liquid}} = \frac{\frac{^2\text{H}}{^1\text{H}}_{\text{Vapor}}}{\frac{^2\text{H}}{^1\text{H}}_{\text{Liquid}}}
\]

\[
\alpha_{\text{Liquid} \leftrightarrow \text{Vapor}} = \frac{1}{\alpha_{\text{Vapor} \leftrightarrow \text{Liquid}}}
\]

If the fractionation factor for evaporation is: \( \alpha_{^{18}\text{O}}^{\text{evaporation}} = 0.988 \)

The fractionation factor for condensation is: \( \alpha_{^{18}\text{O}}^{\text{condensation}} = 1.012 \)
All of the fractionation processes are temperature dependent! Evaporation driven fractionation is determined by the Clausius Clapeyron equation (lecture 7-2), stating that the saturated vapor pressure $SVP=P_{H_2O}$ increases exponentially with $T$.

$$k \equiv P$$

$$\frac{d \ln P}{dT} = \frac{dP}{P \cdot dT} = \frac{L}{R \cdot T^2}$$

$$\ln P = \frac{L}{RT} + C \quad \alpha = \frac{P_{H_2^{18}O}}{P_{H_2^{16}O}}$$

$$\ln \alpha = \ln P_{H_2^{18}O} - \ln P_{H_2^{16}O} \approx \frac{L_{H_2^{18}O} - L_{H_2^{16}O}}{RT}$$

$$\alpha = e^{\frac{L_{H_2^{18}O} - L_{H_2^{16}O}}{RT}}$$

The difference in evaporation process is due to differences in the latent heat (binding energy associated with the phase transition) for the water molecules with different oxygen (hydrogen) isotopes.

Typically modeled by functions like:

$$\ln \alpha = C_1 + \frac{C_2}{T} \left( + \frac{C_3}{T^2} \right) \quad \alpha = e^{\left( C_1 + \frac{C_2}{T} \right)}$$

for $T$ in Kelvin: $C_1 \approx 0.03 \ C_2 \approx 11.63$
Precipitation

Starting from the cloud fractionation value $\delta^{18}\text{O}=-12\%$ derived in previous example!

$$\delta_{\text{rain}} = \alpha \cdot (\delta_{\text{cloud}} + 1000) - 1000$$

for $\alpha = 1.00938$ at $T = 25^\circ C$

$$\delta_{\text{rain}} = 1.00938(-12 + 1000) - 1000 = -2.7\%$$

for $\alpha = 1.0117$ at $T = 0^\circ C$

$$\delta_{\text{rain}} = 1.0117(-12 + 1000) - 1000 = -0.44\%$$

Fractionation in condensation and precipitation processes are highly temperature dependent! This identifies these processes as possible tools for temperature measurements: $\delta = f(T)$!