Paleoclimatology

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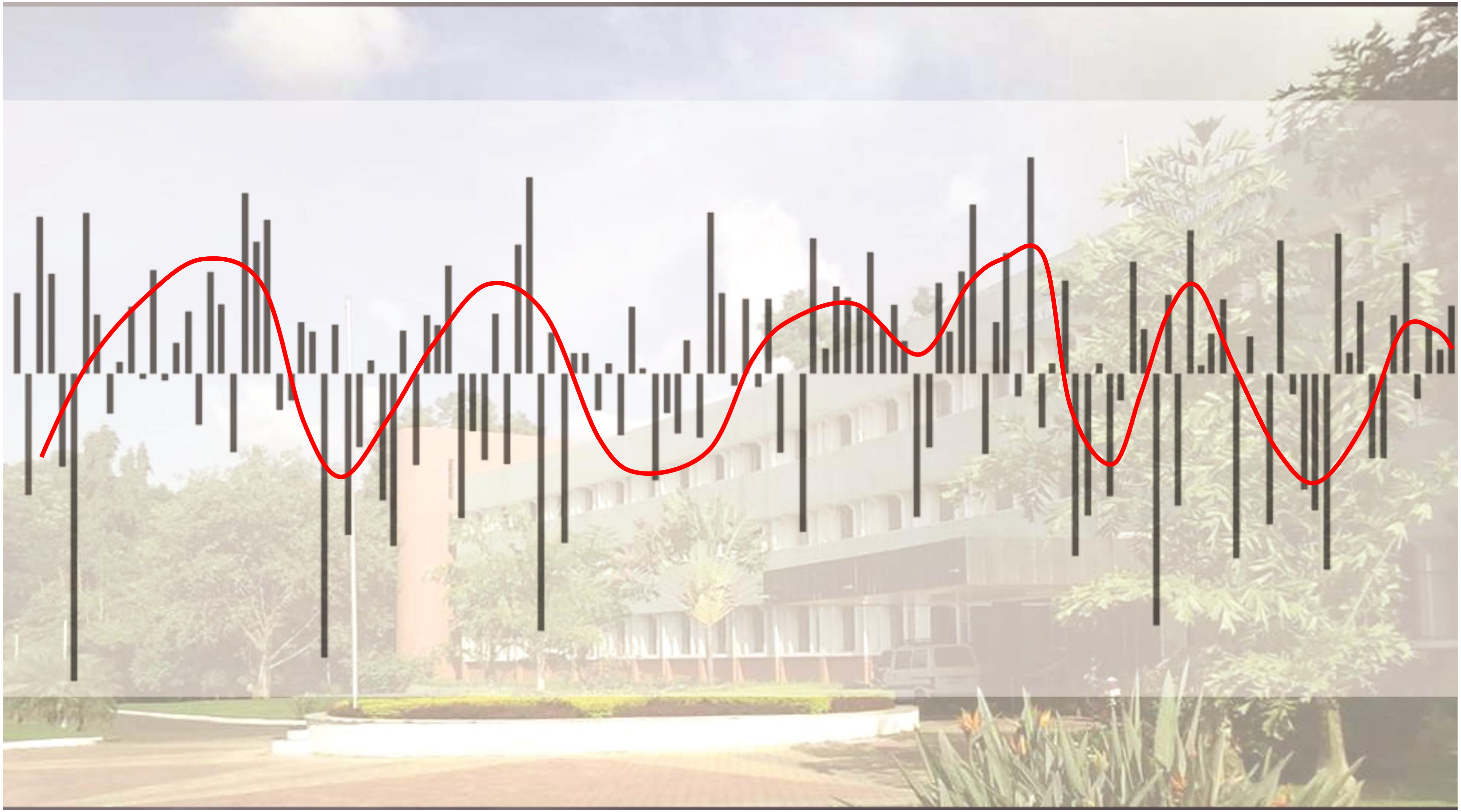
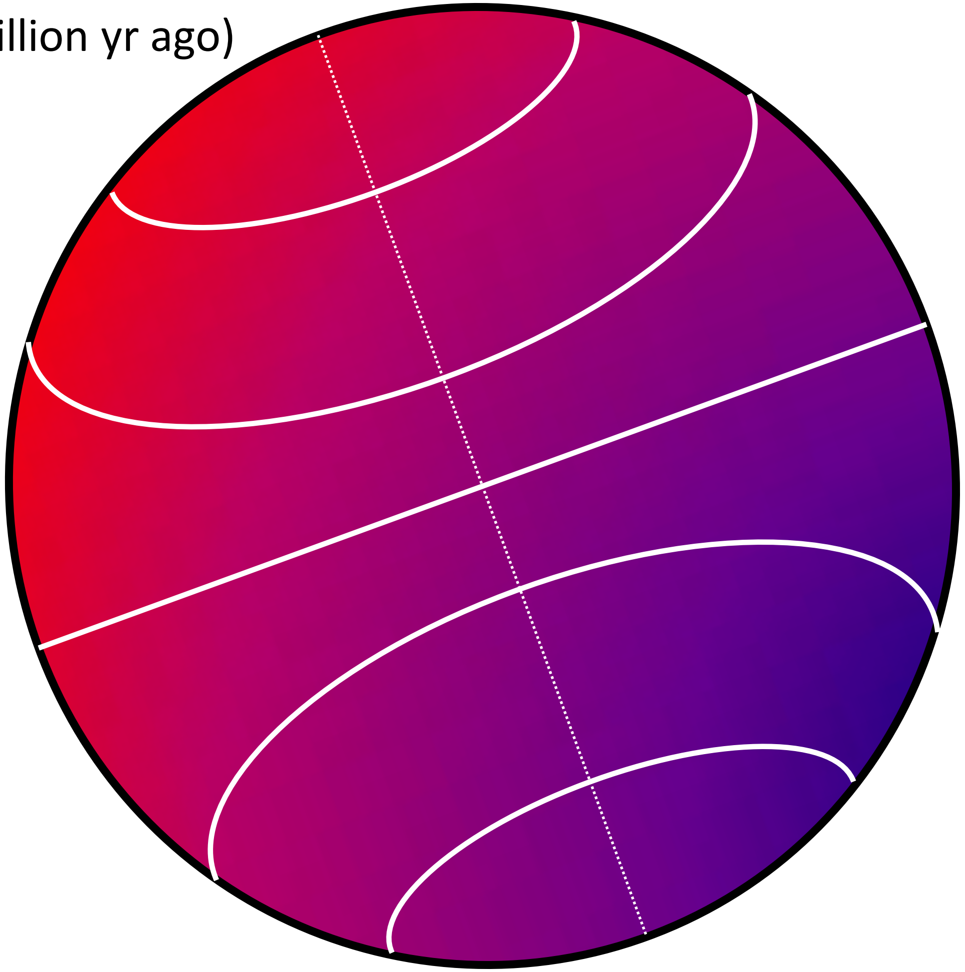
A schematic representation of Indian monsoon rainfall variability based on instrumental data



Instrumental records available from 1841

Climate Change Geologic time

Early Pliocene (3-5 million yr ago)



~3-4 ^{o}C warmer

~10 ^{o}C warmer

Under commonly assumed greenhouse forcing scenarios, atmospheric carbon dioxide concentrations of 500–600 ppmv — roughly twice the preindustrial level — would be required to produce the climate of the Pliocene.

ERA	PERIOD	AGE (Myr)	ICE AGES	EPOCH
PALEOZOIC	Quaternary			Holocene
		0.01		Pleistocene
	Tertiary	2		Pliocene
		5		Miocene
		26		Oligocene
		37		Eocene
		53		Paleocene
MESOZOIC	Cretaceous	65		
	Jurassic	136		
	Triassic	190		
CENOZOIC	Permian	225		
	Carboniferous	280		
	Devonian	345		
	Silurian	395		
	Ordovician	430		
	Cambrian	500		
		570		

How humans are reversing climate clock by 50m years

Washington: Humans are reversing a long-term cooling trend tracing back at least 50 million years, and it has taken us just two centuries to do so, according to a study.

By 2030, Earth's climate is expected to resemble that of the mid-Pliocene, going back more than three million years in geologic time, according to the study published in the journal 'PNAS'.

Without reductions in greenhouse gas emissions, our climates by 2150 could compare to the warm and mostly ice-free Eocene, an epoch that characterised the globe 50 million years ago, said researchers from the University of

AP

HEAT IS ON

More glaciers in Antarctica losing ice in last 10 years

Times of India : 12-12-18

Climate proxies

Climate archives contain many indicators of past climate

Proxy analysis involves: understanding the mechanism by which climate signals are recorded by proxy indicators in order to decipher climate changes.

Two kind of proxies used are:

1. Biotic proxies

(changes in the composition of plant and animal group)

2. Geological-geochemical proxies

(quantifies mass movement of Earth's materials through the climate system, either as discrete (physical) particles or in dissolved (chemical) form.

Climate proxies (cont.)

Biotic proxy

(plant fossil, spores/pollen, plankton, cocoliths, diatom)

Geological/Geochemical proxy

(major/minor/complex ions/magnetic property /isotopic ratio/elemental ratio/

- Morphology
- Pollen count
- Stomatal index/density
- **Ring width**
- Physical property
- Elemental ratio
- **Isotopic ratio** (Stable and Radio isotope)

^{16}O, ^{18}O, ^{13}C, ^{12}C, ^{2}H, H

TREE-RING BASED CLIMATE RECONSTRUCTION: INDIAN PERSPECTIVE

Dendrochronology: dendron (= “tree”)+chronos (= “time”)+- logy (= the study of) Dendrochronology:

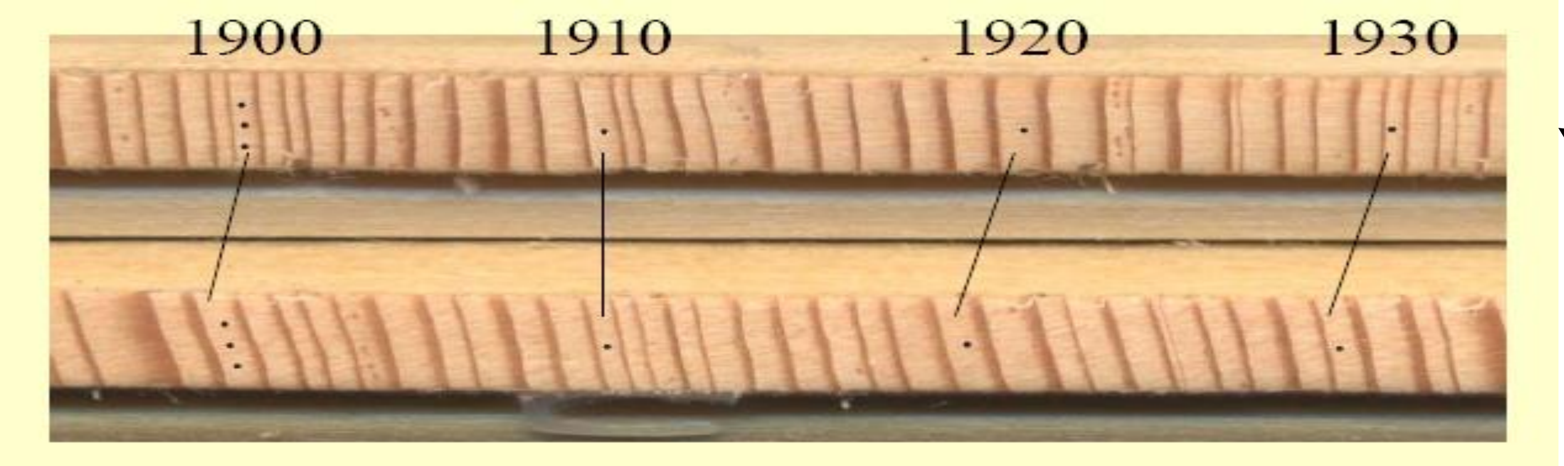
Dendroclimatology: The science of reconstructing past climate by using tree-rings is known as Dendroclimatology,

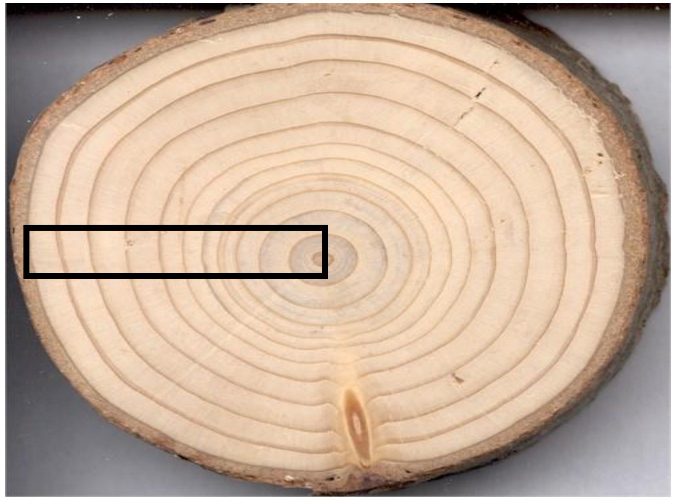
Tree rings are annually resolved

A calendar year can be assigned to each ring

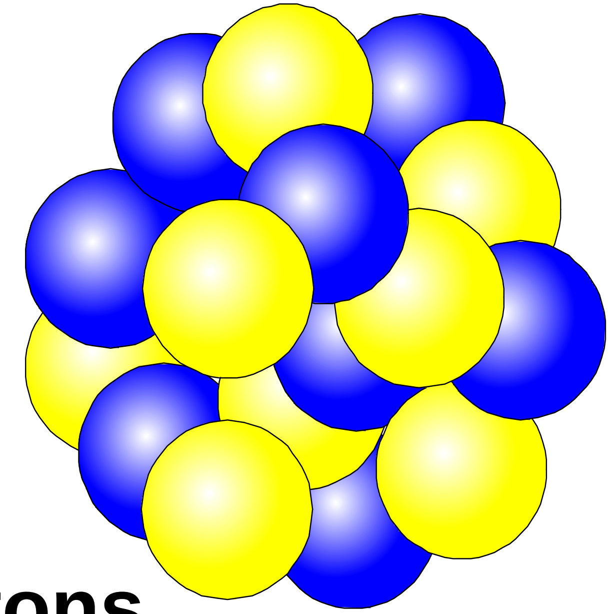
Show a continuous record; have widespread distribution

Credit: Dr. H.P. Borgaonkar





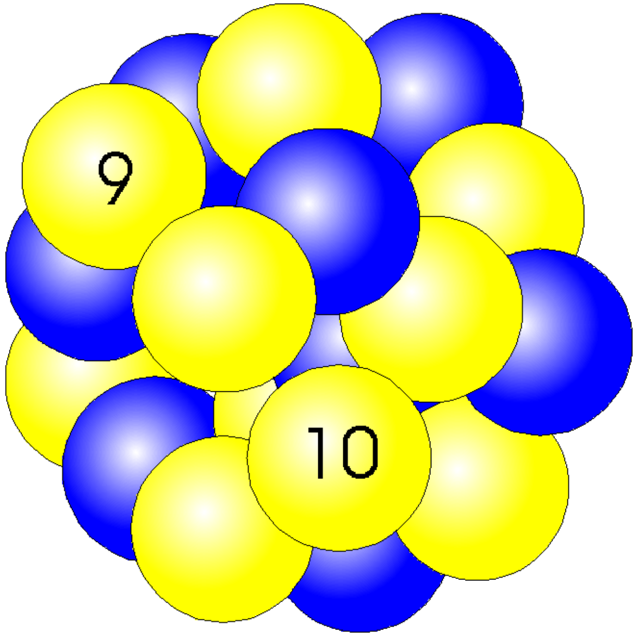
Climatic information recorded in atomic domain



8 protons

8 neutrons

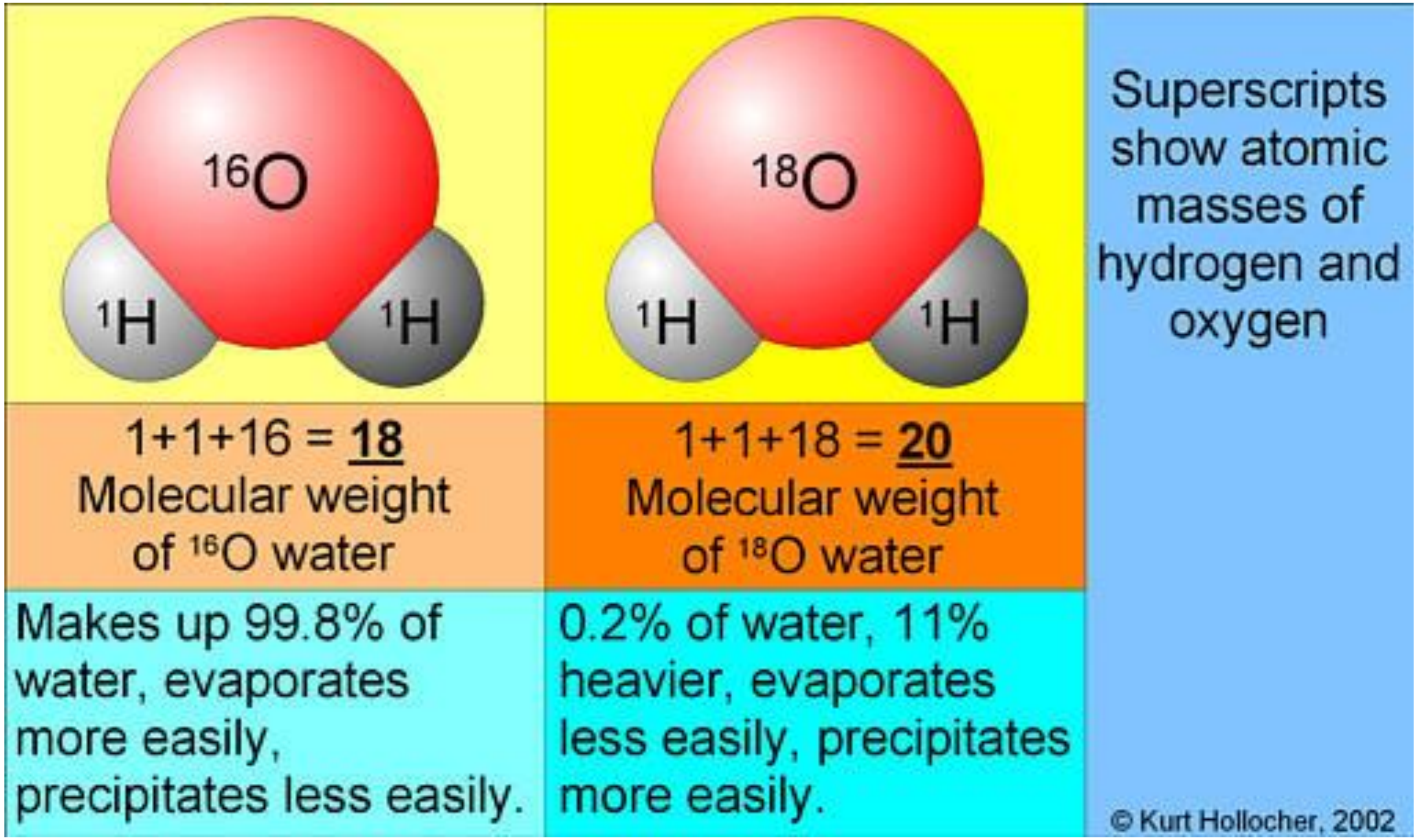
^{16}O



8 protons

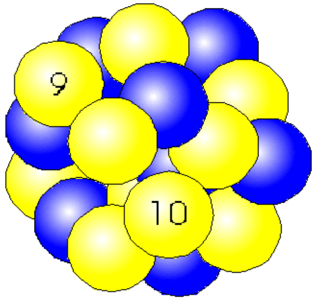
+ 10 neutrons

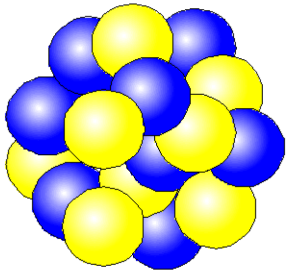
^{18}O



$^{1}H^{1}H^{16}O$, $^{1}H^{1}H^{18}O$; $^{1}H^{2}H^{16}O$, $^{1}H^{2}H^{18}O$, $^{1}H^{3}H^{16}O$

Element	Isotope	Atomic Weight (Amu)	Abundance (atom %)
Hydrogen (Z=1)	^{1}H (Protium) ^{2}H (Deuterium)	1.0078 2.0141	99.985 0.015
Carbon (Z=6)	^{12}C ^{13}C	12.000 13.033	98.90 1.10
Nitrogen (Z=7)	^{14}N ^{15}N	14.0030 15.0001	99.63 0.37
Oxygen (Z=8)	^{16}O ^{17}O ^{18}O	15.9949 16.9991 17.9991	99.76 0.04 0.20





$$\frac{^{18}O}{^{16}O} = \frac{1}{500}$$

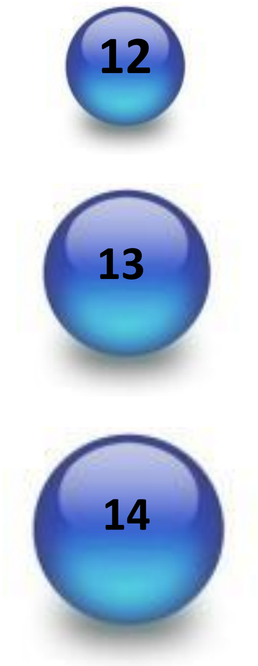
RATIO (R) : Heavier to light isotope

$$\delta (‰) = (R_{sample}/R_{standard} - 1) \cdot 1000$$

R_x = $^{2}H/^{1}H$ or $^{18}O/^{16}O$ (for Hydrogen or Oxygen)

Isotope fractionation

Carbon atom



Because of difference in size and atomic weight, different isotopes can react at slightly different rates.

Isotope Fractionation

Isotopic fractionation involves the partial separation of isotopes during physical, chemical or biological processes.

Isotope fractionation (mechanism)

No. of electrons control the chemical reaction of the atom undergoes

Atomic mass determines the vibrational energy of the nucleus

So differences in mass $\longrightarrow$ Reaction Rate and Bond Strength

Why mass difference leads to difference in physical behavior?

K.E. is constant for a given element in a given environmental condition (const. T)

$$KE = \frac{1}{2} mv^2$$

So higher mass (isotope) possesses lower velocity

Isotope fractionation (mechanism – contd.)

If L stands for $H_2{}^{16}O$ and H for $H_2{}^{18}O$

Then $v_L/v_H = (m_H/m_L)^{1/2}$

So $v_{^{16}O} / v_{^{18}O} = (20/18)^{1/2}$

So at any temp the velocity of $H_2{}^{16}O$ is 1.05 times faster than $H_2{}^{18}O$

Characteristic physical properties of $H_2{}^{16}O$, $D_2{}^{16}O$ and $H_2{}^{18}O$

Property	$H_2{}^{16}O$	$D_2{}^{16}O$	$H_2{}^{18}O$
Density (20°C, in g cm^{-3})	0.997	1.1051	1.1106
Temperature of greatest density (°C)	3.98	11.24	4.30
Melting point (760 Torr, in °C)	0.00	3.81	0.28
Boiling point (760 Torr, in °C)	100.00	101.42	100.14
Vapour pressure (at 100°C, in Torr)	760.00	721.60	
Viscosity (at 20°C, in centipoise)	1.002	1.247	1.056

Isotope fractionation processes (contd.)

Types of fractionation

1. Isotope exchange reaction (equilibrium isotope distribution)
2. Kinetic processes (depends primarily on differences in reaction rates of isotopic molecules)

Isotope exchange is used for all situations in which there is no net reaction, but in which the isotope distribution changes between different chemical phases

$$H_2{}^{18}O + C^{16}O_2 \longleftrightarrow H_2{}^{16}O + C^{18}O^{16}O$$

^{18}O forms a stronger covalent bond with carbon than does ^{16}O

Isotope fractionation processes

Kinetic isotopic fractionation results when **rates** of reactions or physical processes **differ**.

It also results from irreversible i.e., one way physical or chemical processes

Example:

- ❑ Evaporation of water with immediate withdrawal of the vapor
- ❑ Absorption and diffusion of gases
- ❑ Bacterial decay of plants
- ❑ Rapid calcite precipitation

Isotope fractionation processes

Kinetic isotope effects: Kinetic isotope effects generally relate to difference in the dissociation energies of molecules composed of different isotopes.

For example, the rate determining step in a set of chemical reactions might involve the breakage of a bond. It is substantially easier to break the bonds of molecules that contain the lightest isotopes, which is plausible because the vibrational frequency of such bonds will tend to be higher, then the lighter isotopes will be preferentially incorporated in the products of incomplete reactions, while the heavy isotopes will become enriched in the unreacted residue.

Evaporation: It is an unidirectional, non-equilibrium processes that can cause isotope fractionation. In this case higher translational velocities of molecules containing the *lightest* isotopes may allow them to preferentially break through the liquid surface and escape into the atmosphere.

Isotope fractionation factor

$$\alpha_{A\text{-}B} = \frac{R_A}{R_B}$$

R_A = ratio of the heavy isotope to light isotope in phase A

R_B = the same in phase B

Calcium carbonate precipitation and paleo-thermometer

$$\alpha_{calcite\text{-}water} = (^{18}O/^{16}O)_{CaCO3} / (^{18}O/^{16}O)_{H2O} = 1.0286 \text{ at } 25^oC$$

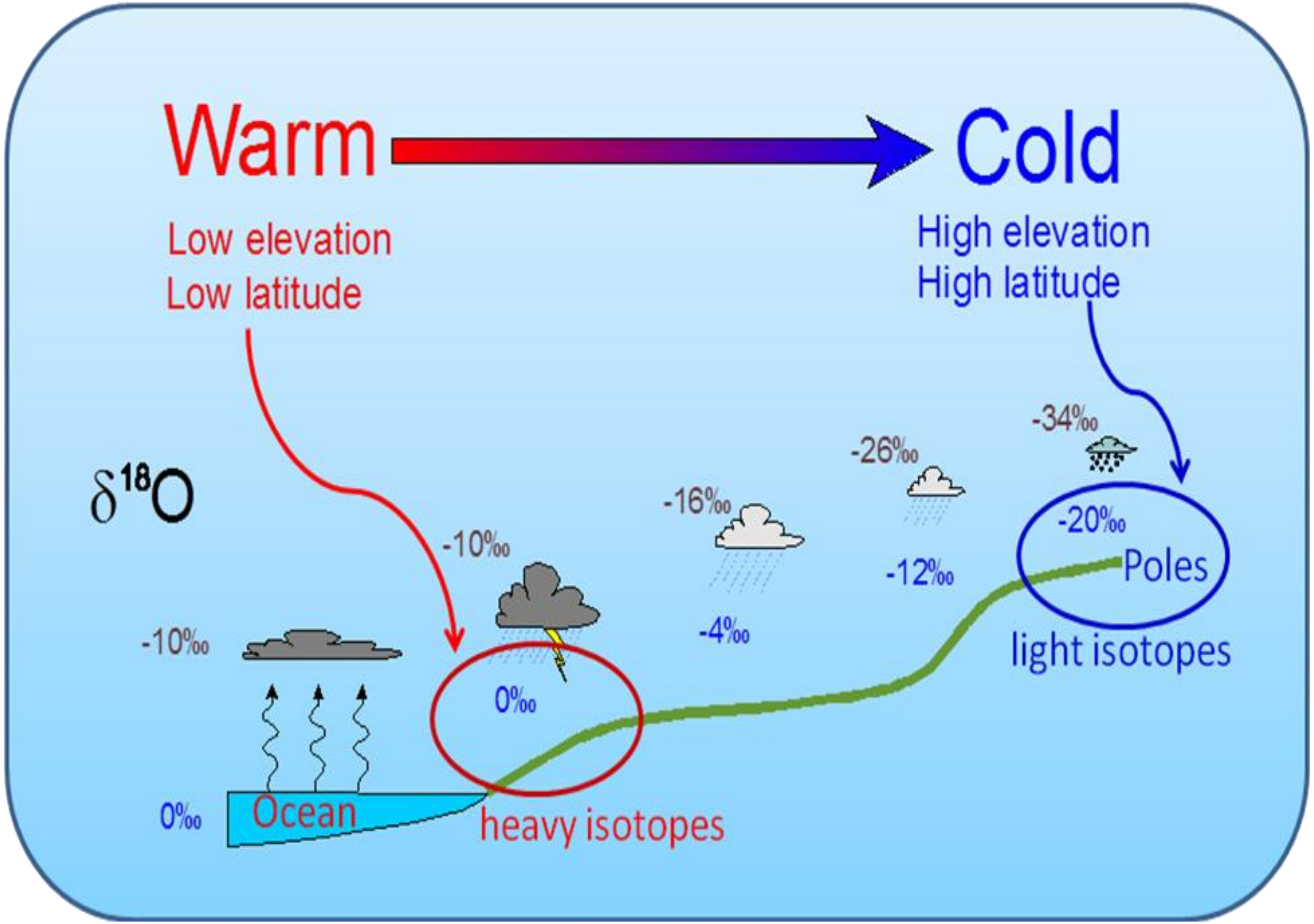
$$H_2{}^{18}O + \tfrac{1}{3}CaC^{16}O_3 \leftrightarrow H_2{}^{16}O + \tfrac{1}{3}\,CaC^{18}O_3$$

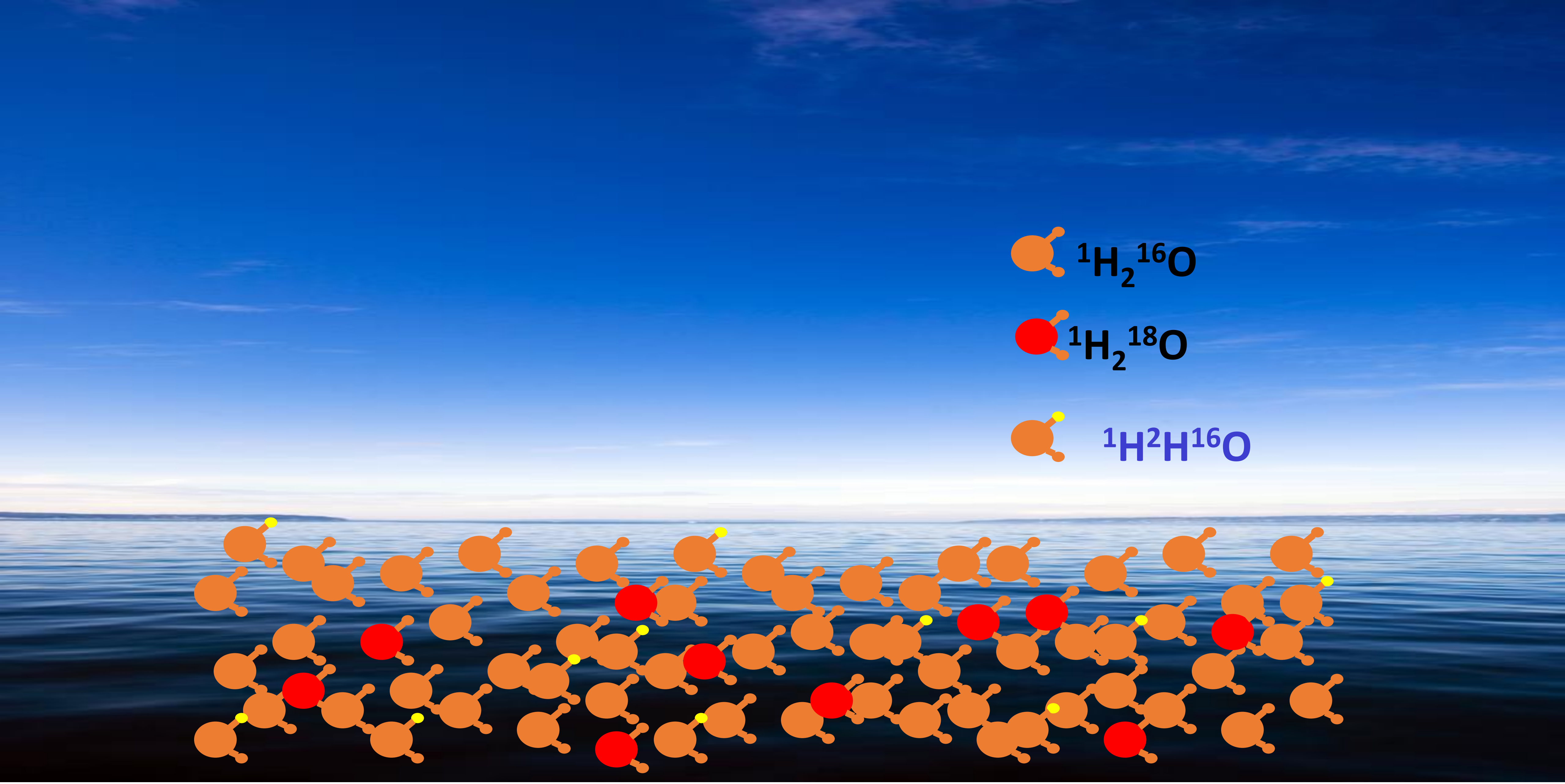
$$t^oC = 16.9 - 4.2\,(\delta_c\text{-}\delta_w) + 0.13\,(\delta_c\text{-}\delta_w)^2$$

$$\delta_{A\text{-}B} = [\,(R_A/R_B)_{sample} - (R_A/R_B)_{reference}\,]\,/\,(R_A/R_B)_{reference} * 1000$$

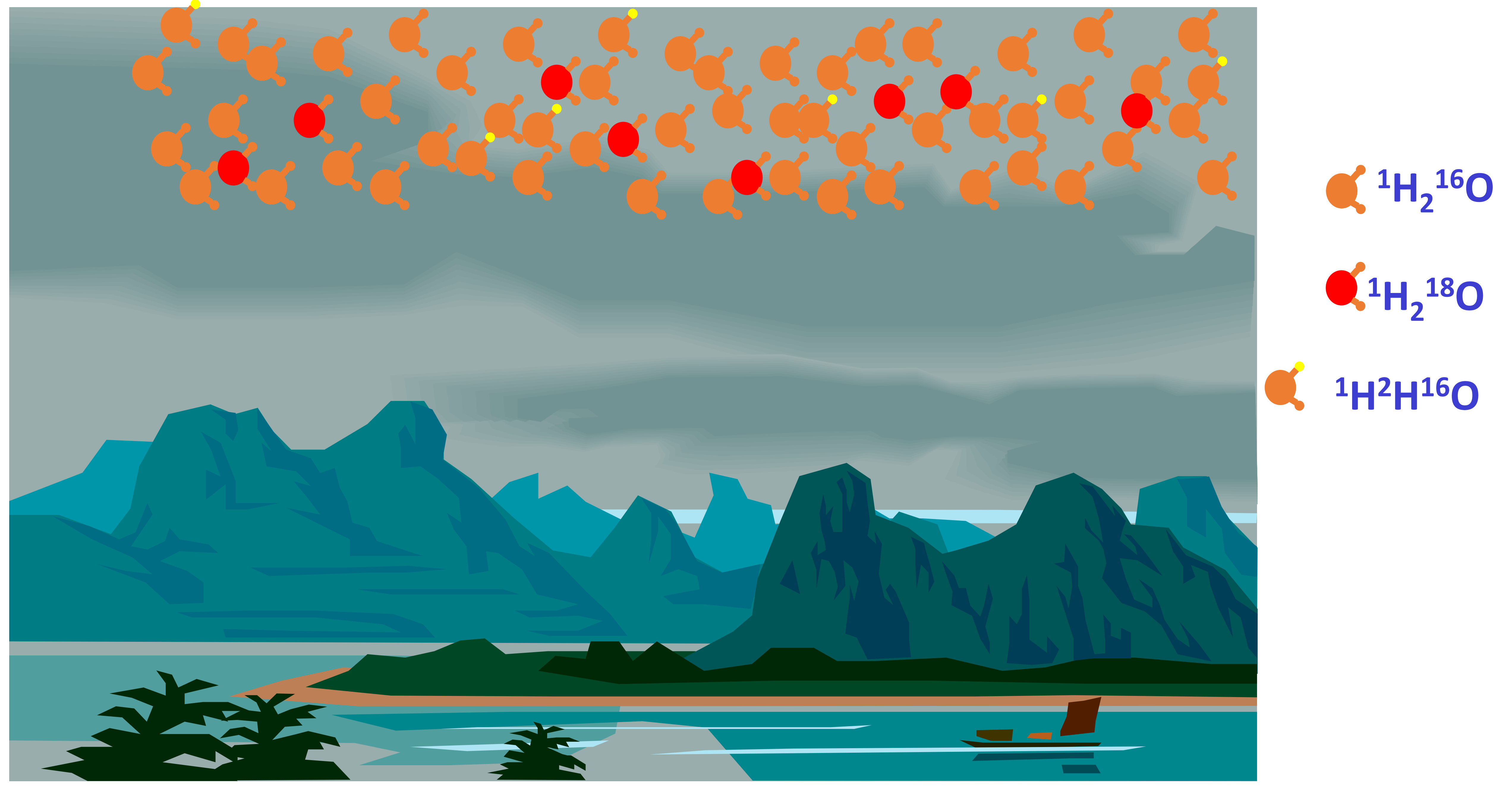
$$\alpha_{A\text{-}B} = \frac{1000+\delta_A}{1000+\delta_B}$$

How do the isotopes "carry" the signature of rainfall?

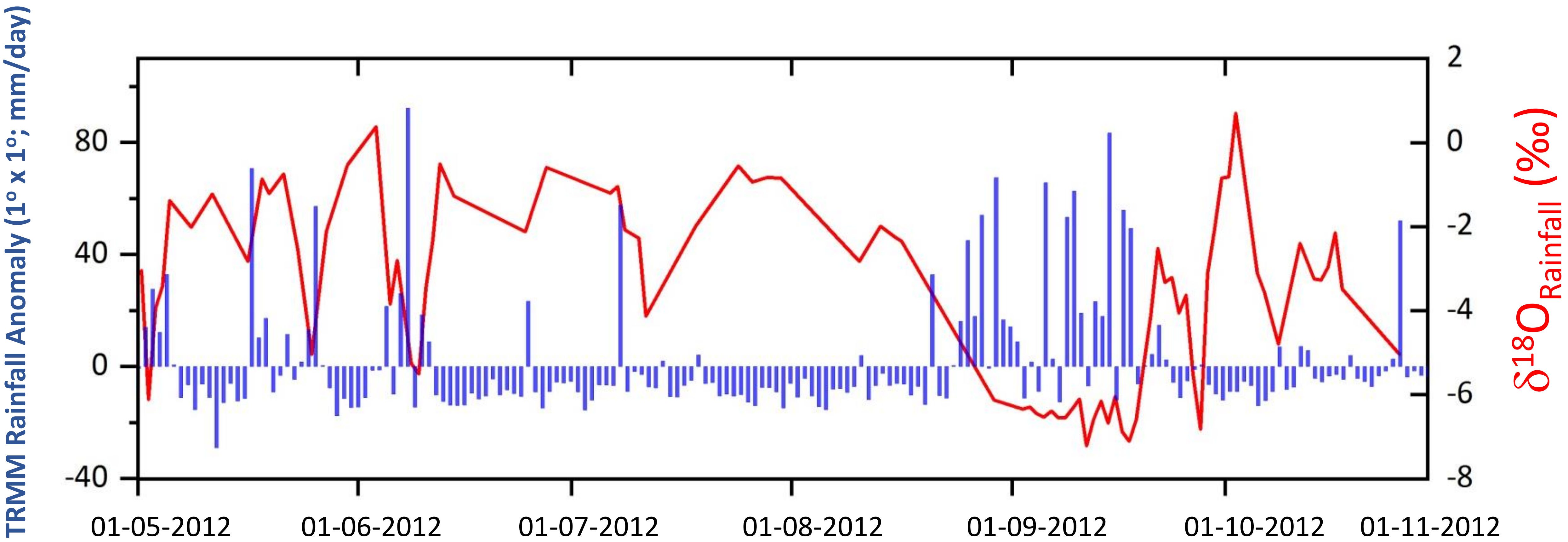




Graphics credit: Trina Bose



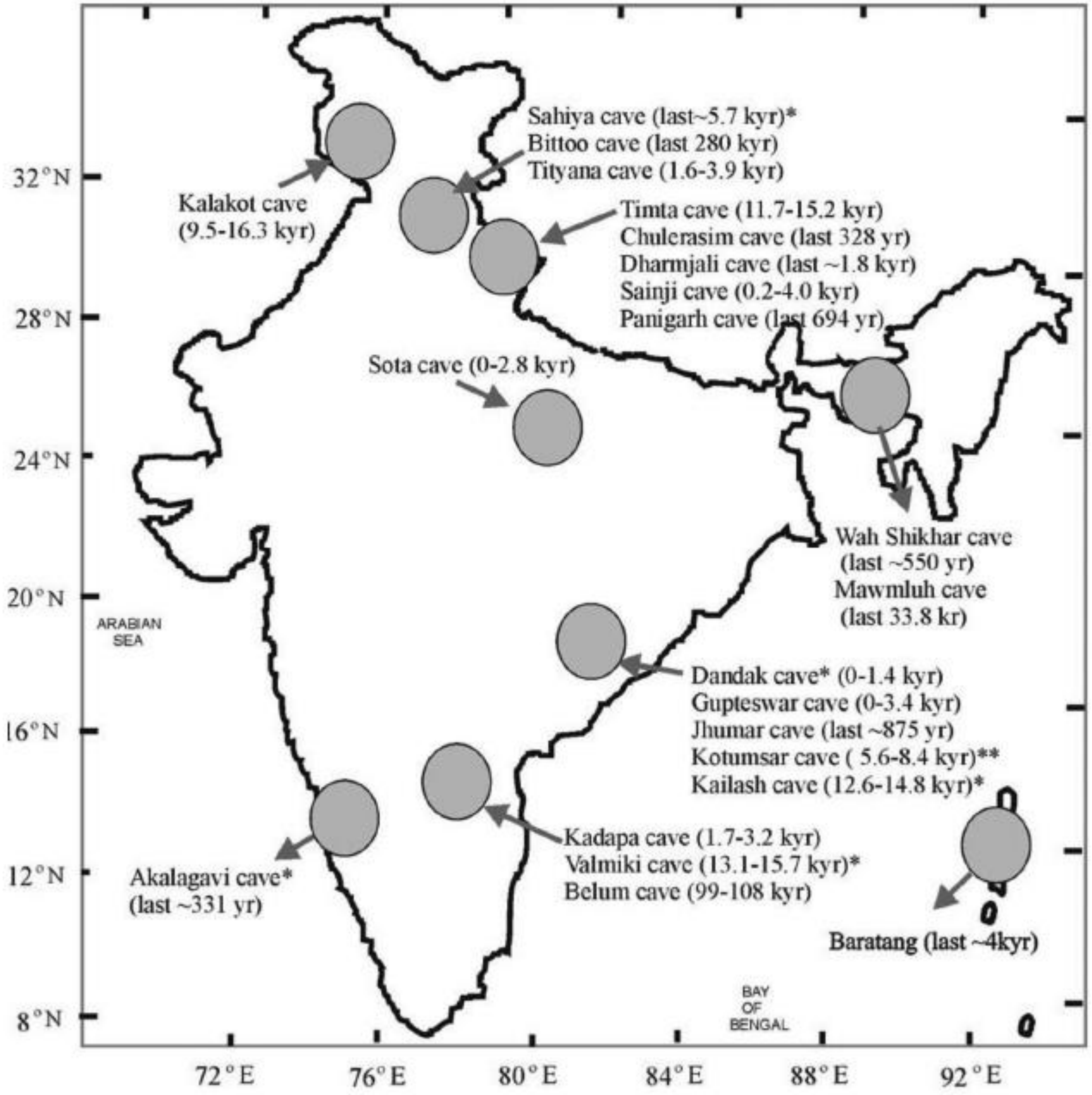
$^1H_2{}^{16}O$
$^1H_2{}^{18}O$
$^1H^2H^{16}O$



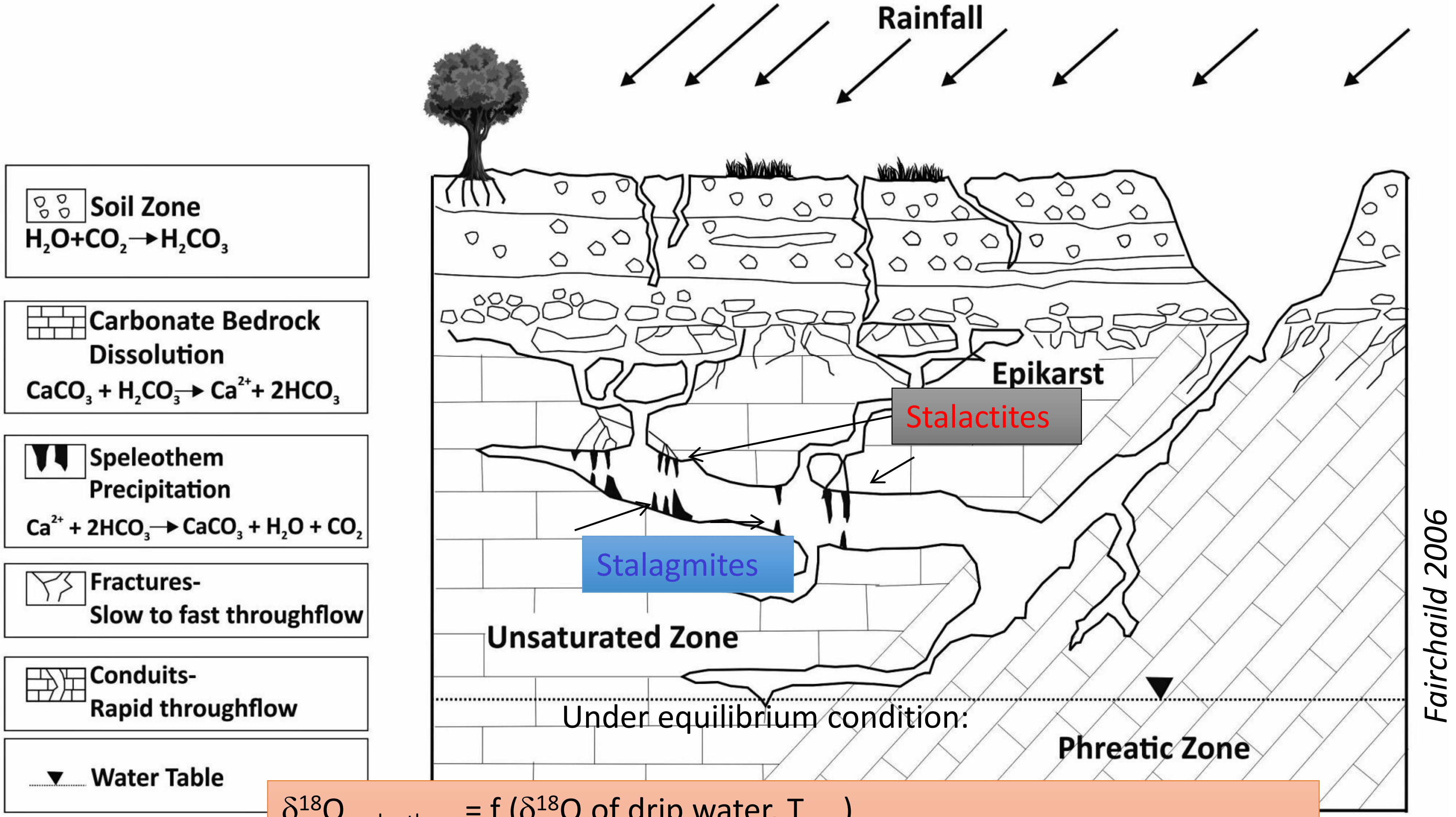
Intra-seasonal variation of rainfall and its oxygen isotopic composition at Port Blair – Andaman Islands

Chakraborty et al. (2016)

Monsoon reconstruction using speleothem

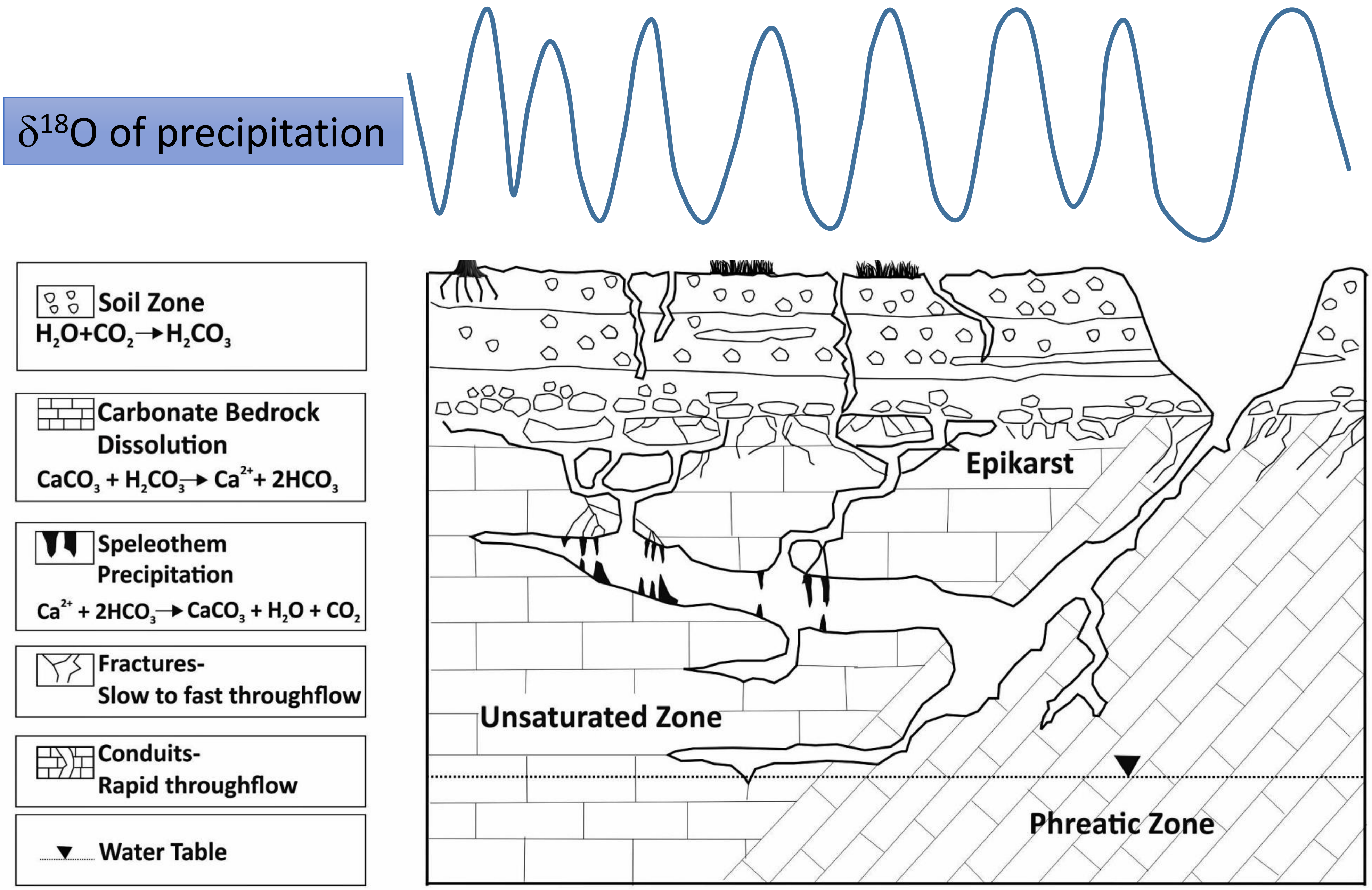


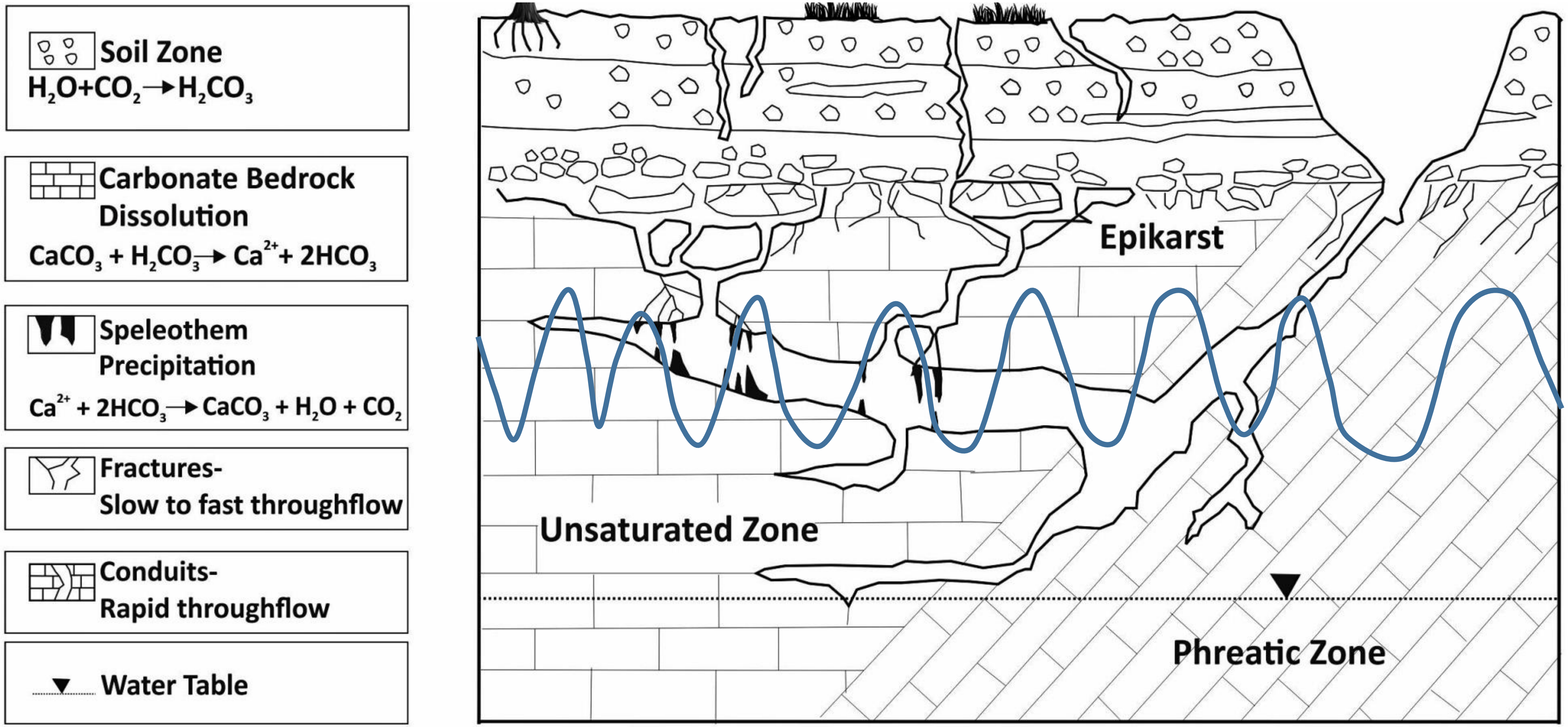
Band and Yadava 2020

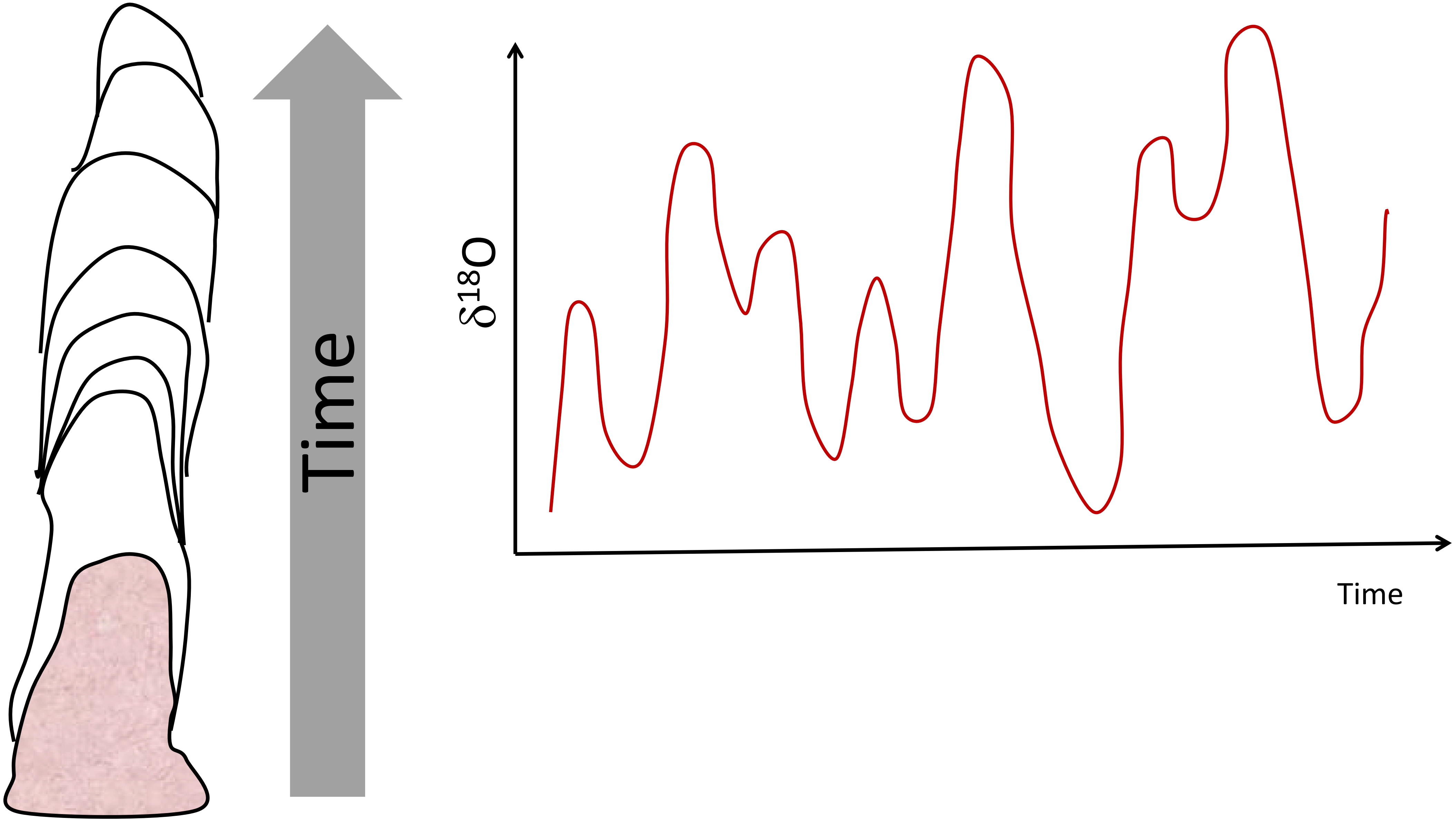


Under equilibrium condition:

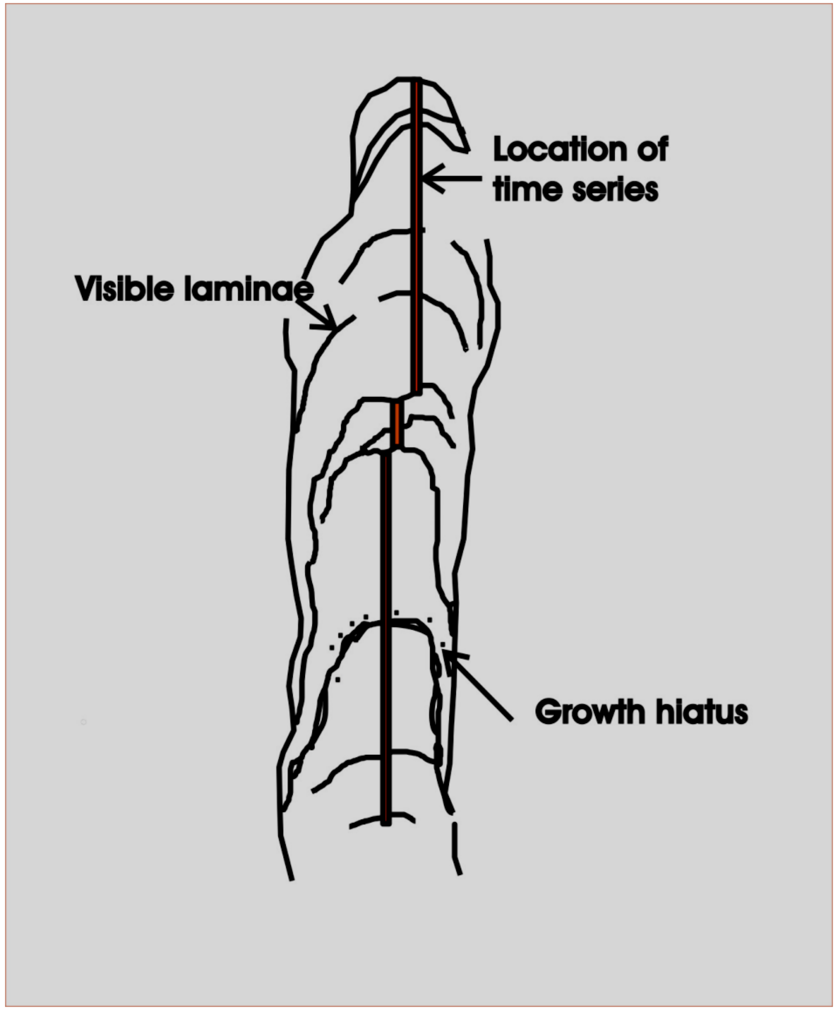
$\delta^{18}O_{speleothem} = f\ (\delta^{18}O \text{ of drip water}, T_{cave})$


δ18O of precipitation
Soil Zone
H2O+CO2→H2CO3
Carbonate Bedrock Dissolution
CaCO3 + H2CO3→ Ca2+ + 2HCO3
Speleothem Precipitation
Ca2+ + 2HCO3→CaCO3 + H2O + CO2
Fractures-
Slow to fast throughflow
Conduits-
Rapid throughflow
Water Table
Epikarst
Unsaturated Zone
Phreatic Zone


Soil Zone
$H_2O + CO_2 \rightarrow H_2CO_3$
Carbonate Bedrock Dissolution
$CaCO_3 + H_2CO_3 \rightarrow Ca^{2+} + 2HCO_3$
Speleothem Precipitation
$Ca^{2+} + 2HCO_3 \rightarrow CaCO_3 + H_2O + CO_2$
Fractures-
Slow to fast throughflow
Conduits-
Rapid throughflow
Water Table
Epikarst
Unsaturated Zone
Phreatic Zone


Time
δ18O
Time

Dating of speleothems



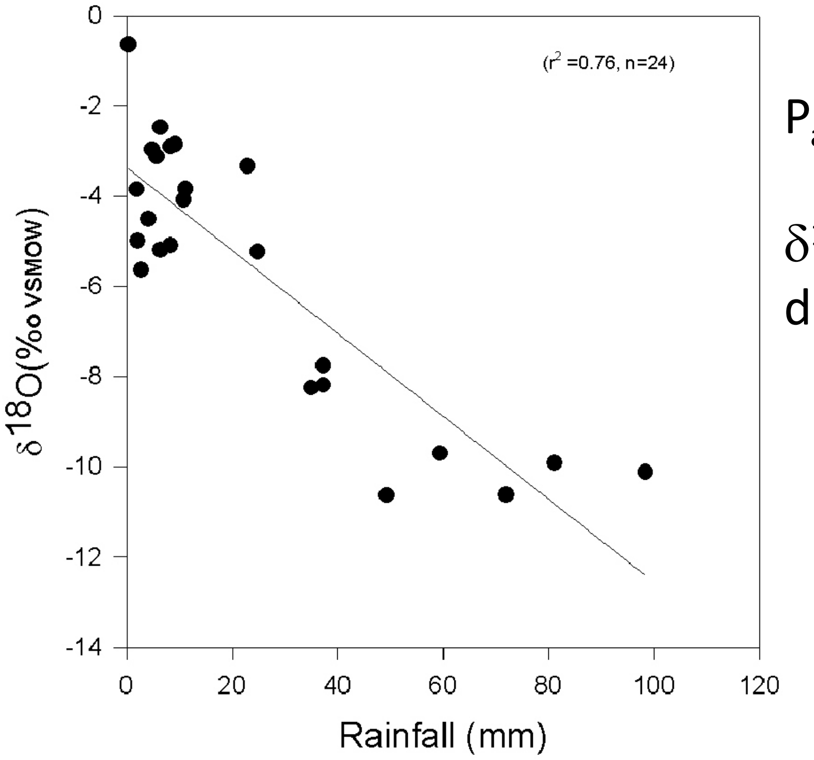
- **C**ounting of laminations
- **^{14}C** method

Uncertainty in the proportion of "dead" carbon limits utility of the method

- **U-Th** method

Decay of ^{234}U to ^{230}Th is the principal method

Amount effect in the S-W monsoon, Akalgavi cave, Karnataka



$P_a = (100/1.5) * 5.3 * (\delta^{18}O_{tip} - \delta^{18}O_i) + 3257.2$

$\delta^{18}O_i$, $\delta^{18}O_{tip}$ -> Oxygen isotopic values of speleothem at any depth and at the tip resp.

In the equatorial belt

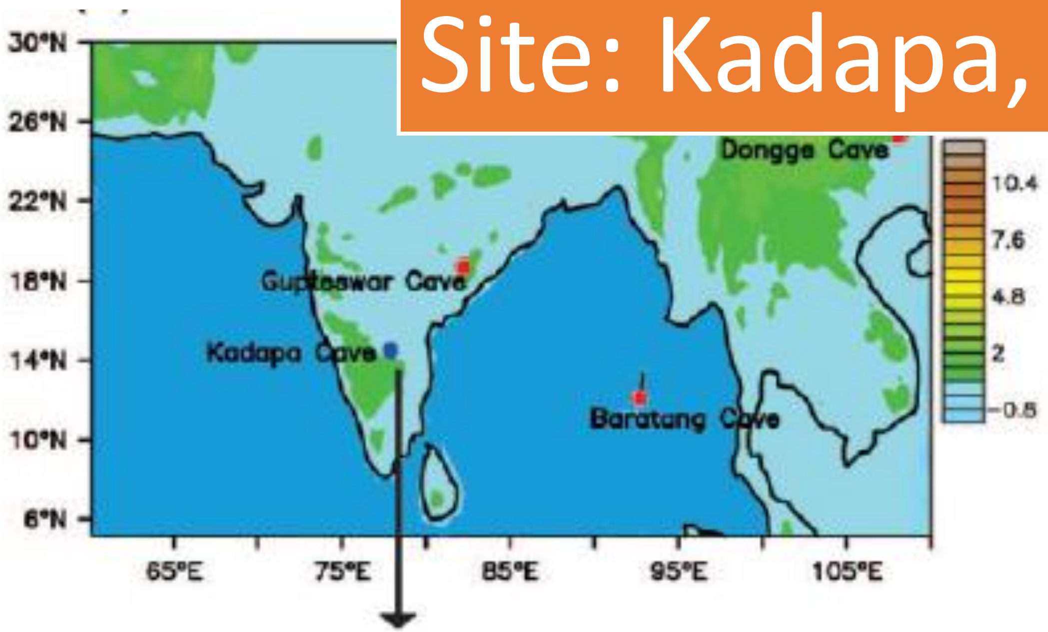
$\delta^{18}O_m = (-0.015 \pm 0.002)P_m - (0.47 \pm 0.42)$

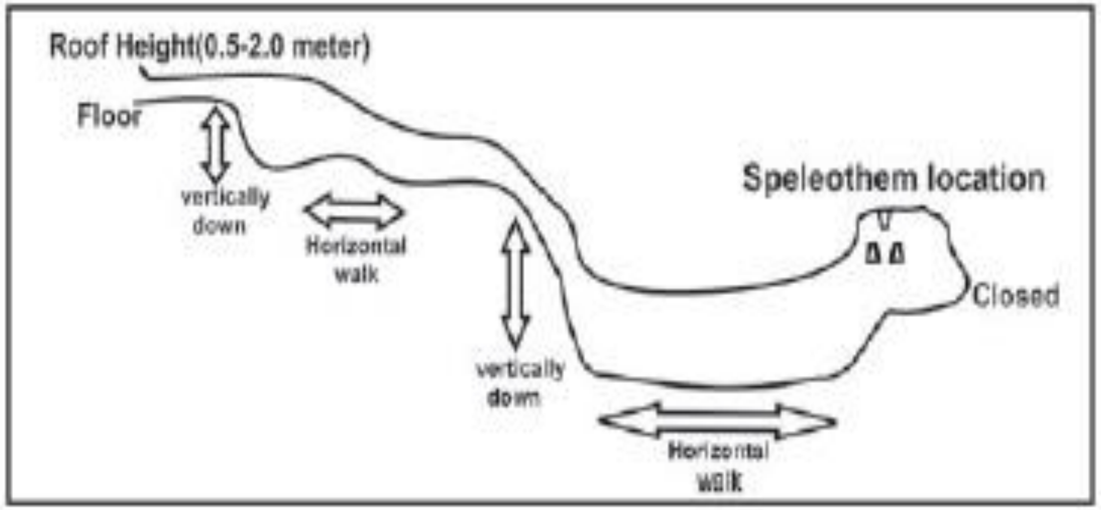
$\delta^{18}O_m$ = isotopic value- monthly mean of precipitation

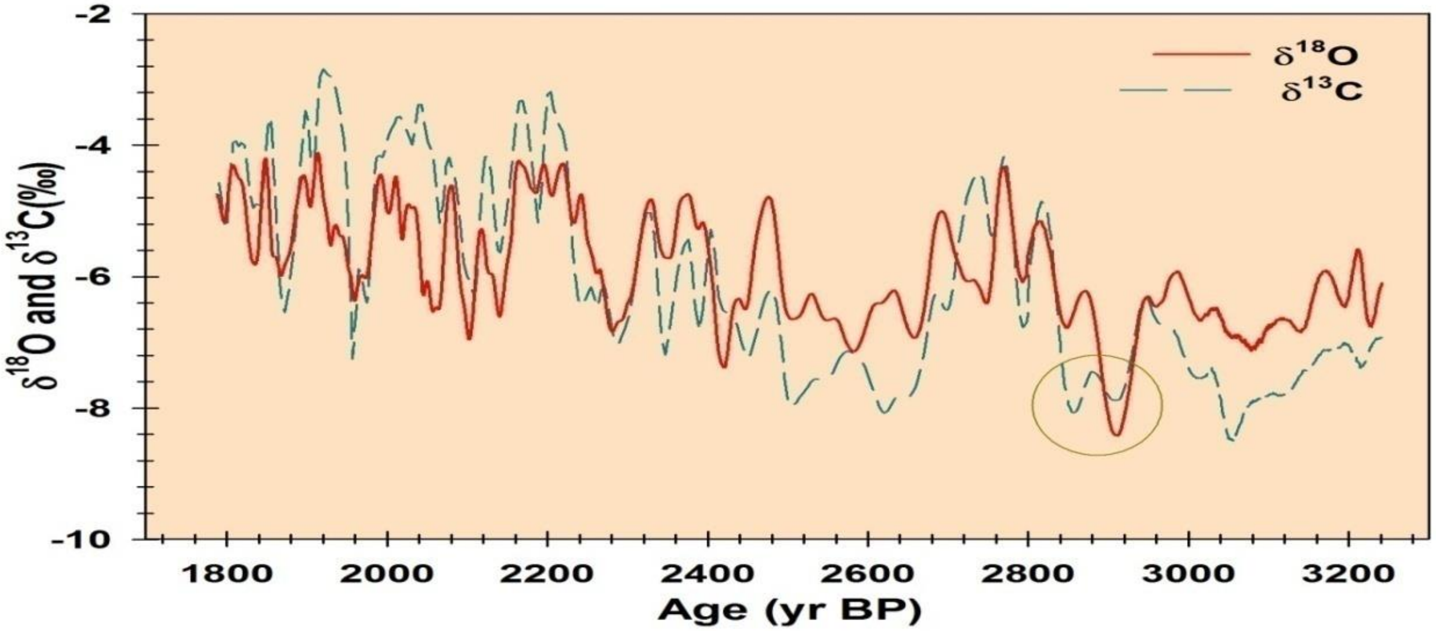
P_m = monthly mean rainfall amount

Yadava et al. 2004

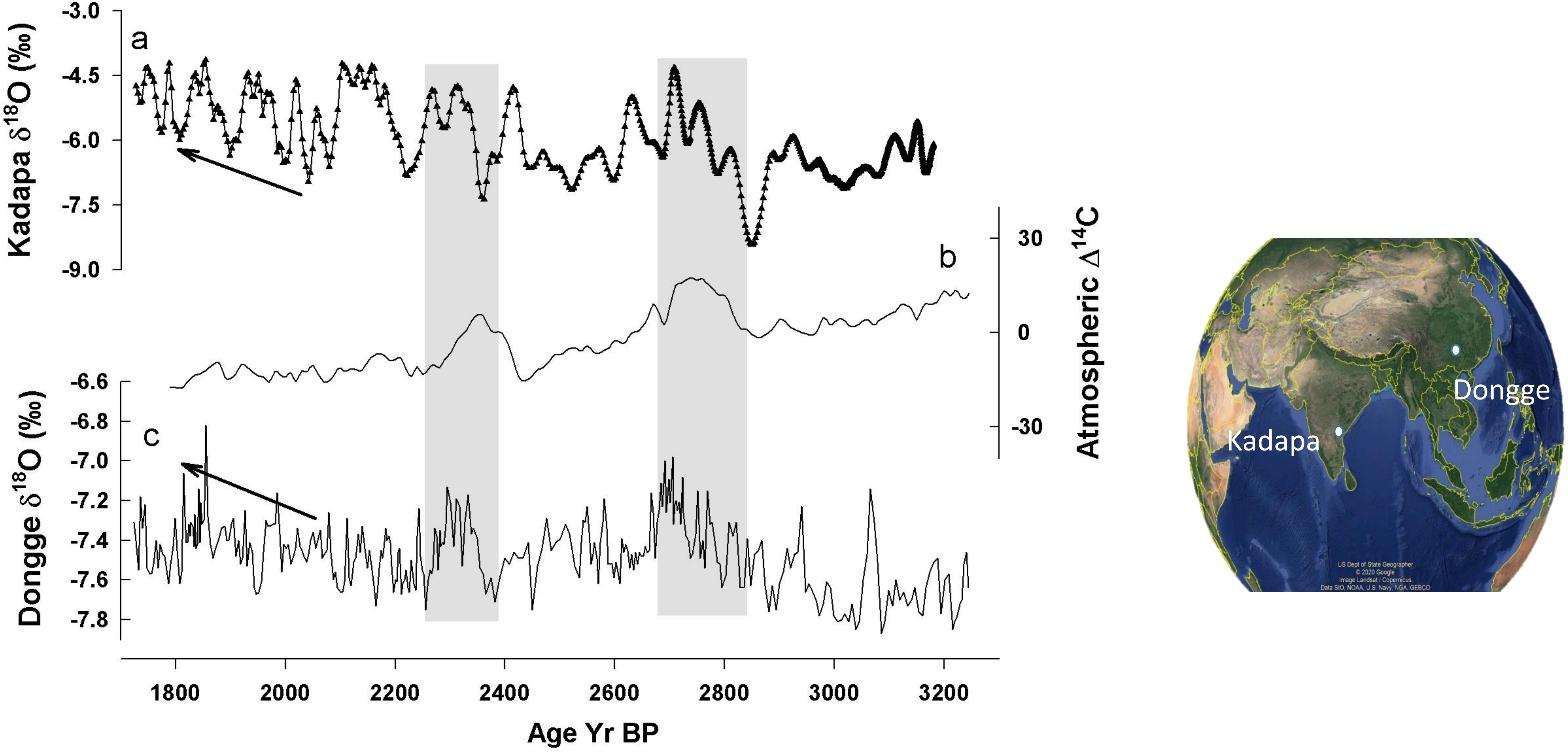
Site: Kadapa, Andhra Pradesh



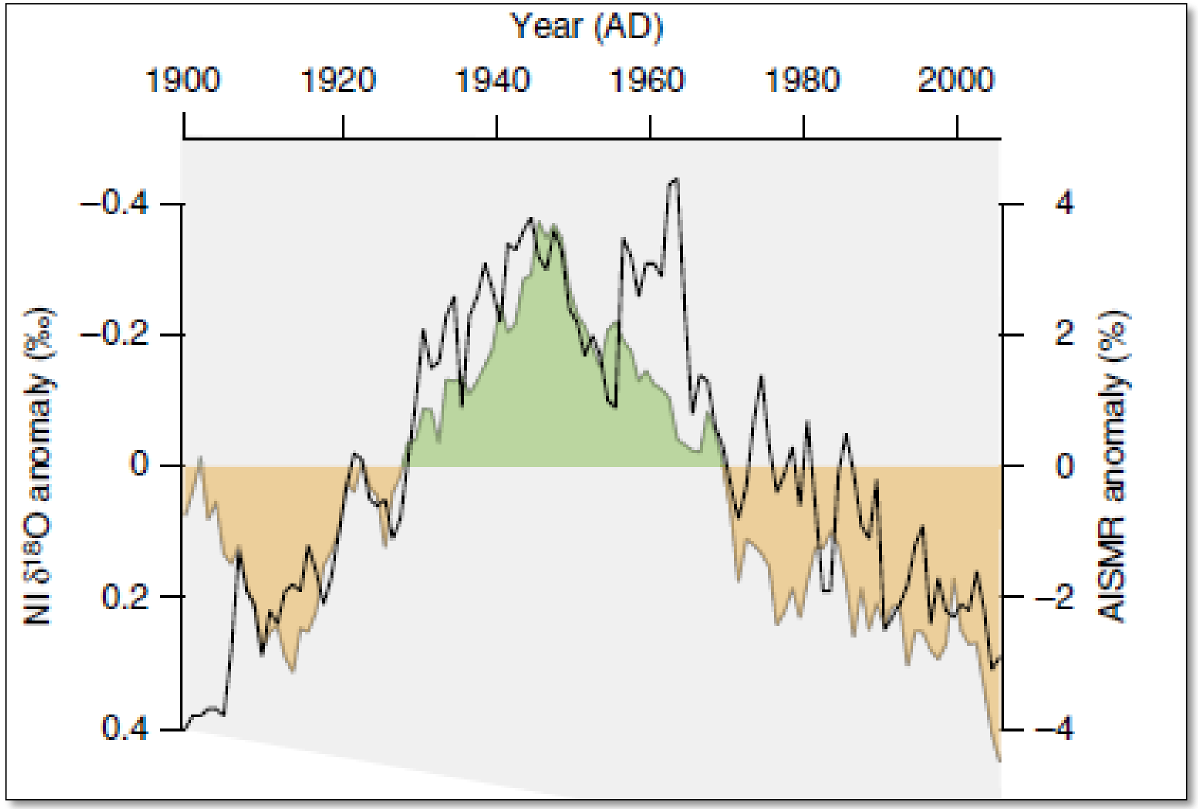


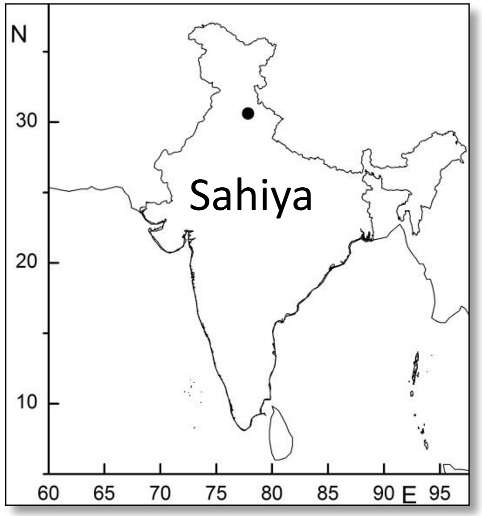


Sinha et al. 2018



Kadapa δ18O (‰)
-3.0
-4.5
-6.0
-7.5
-9.0
a
b
Atmospheric Δ14C
30
0
-30
c
Dongge δ18O (‰)
-6.6
-6.8
-7.0
-7.2
-7.4
-7.6
-7.8
1800
2000
2200
2400
2600
2800
3000
3200
Age Yr BP
Dongge
Kadapa
US Dept of State Geographer
© 2020 Google
Image Landsat / Copernicus
Data SIO, NOAA, U.S. Navy, NGA, GEBCO

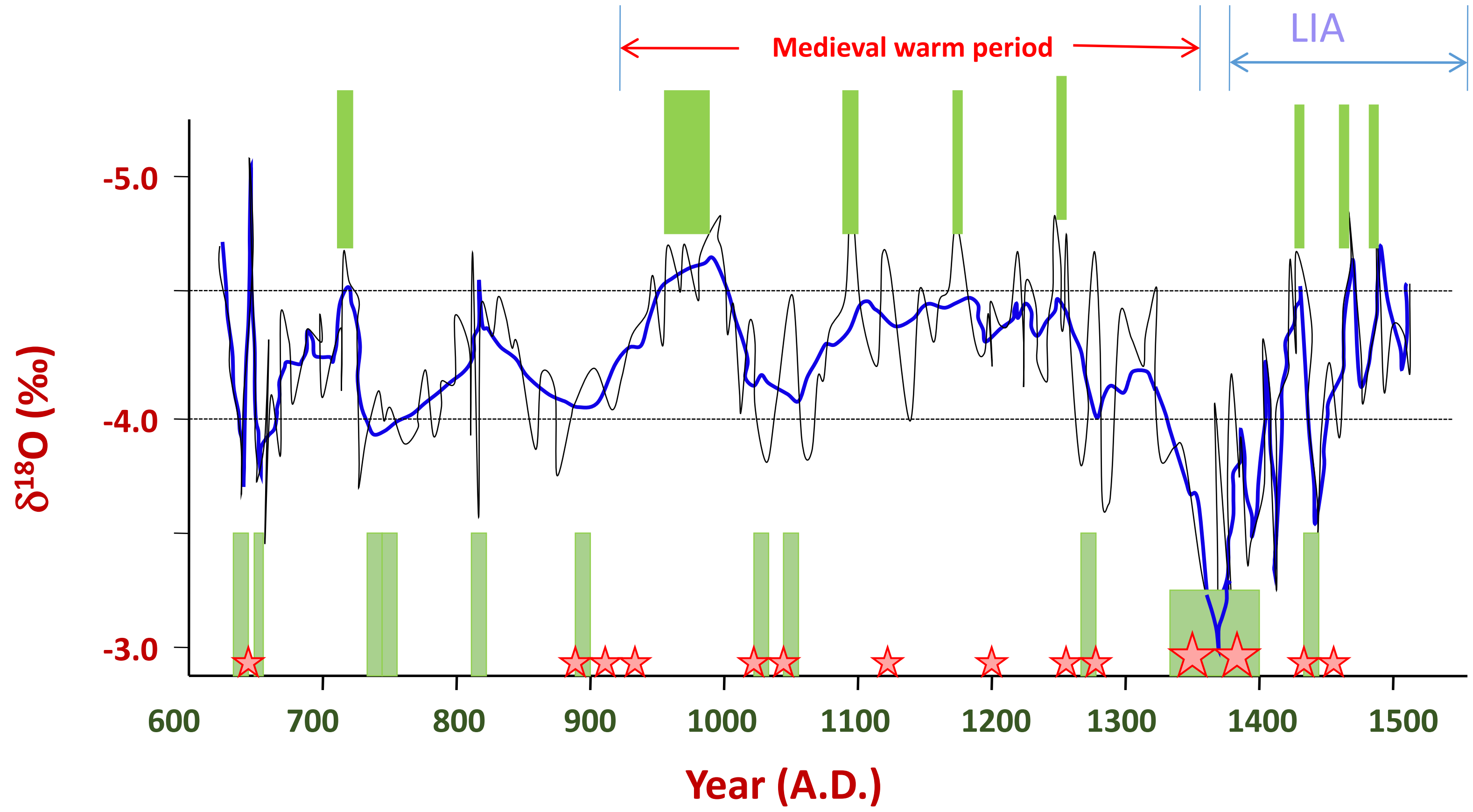




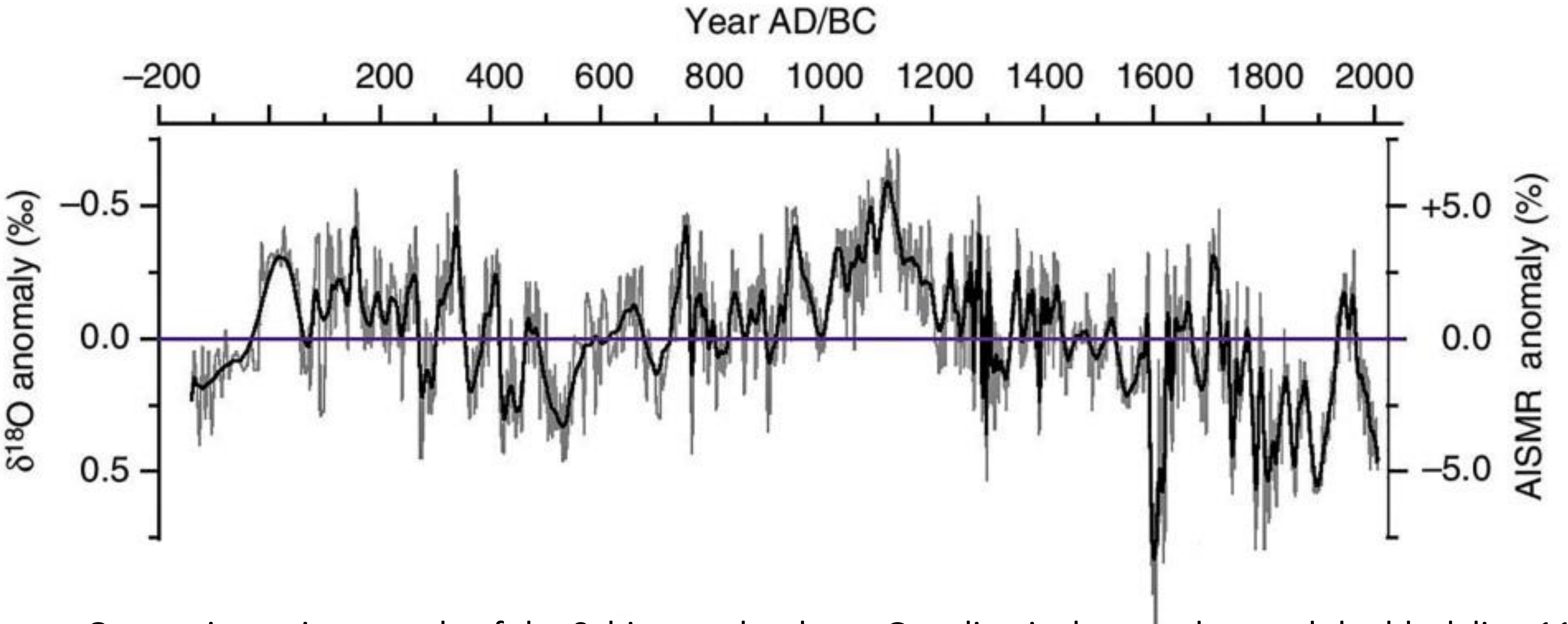
Sinha et al. 2015

Speleothem time series δ^{18}O record from north India (shading).
Black line: All India Summer Monsoon Rainfall anomaly.

Sinha et al. 2007

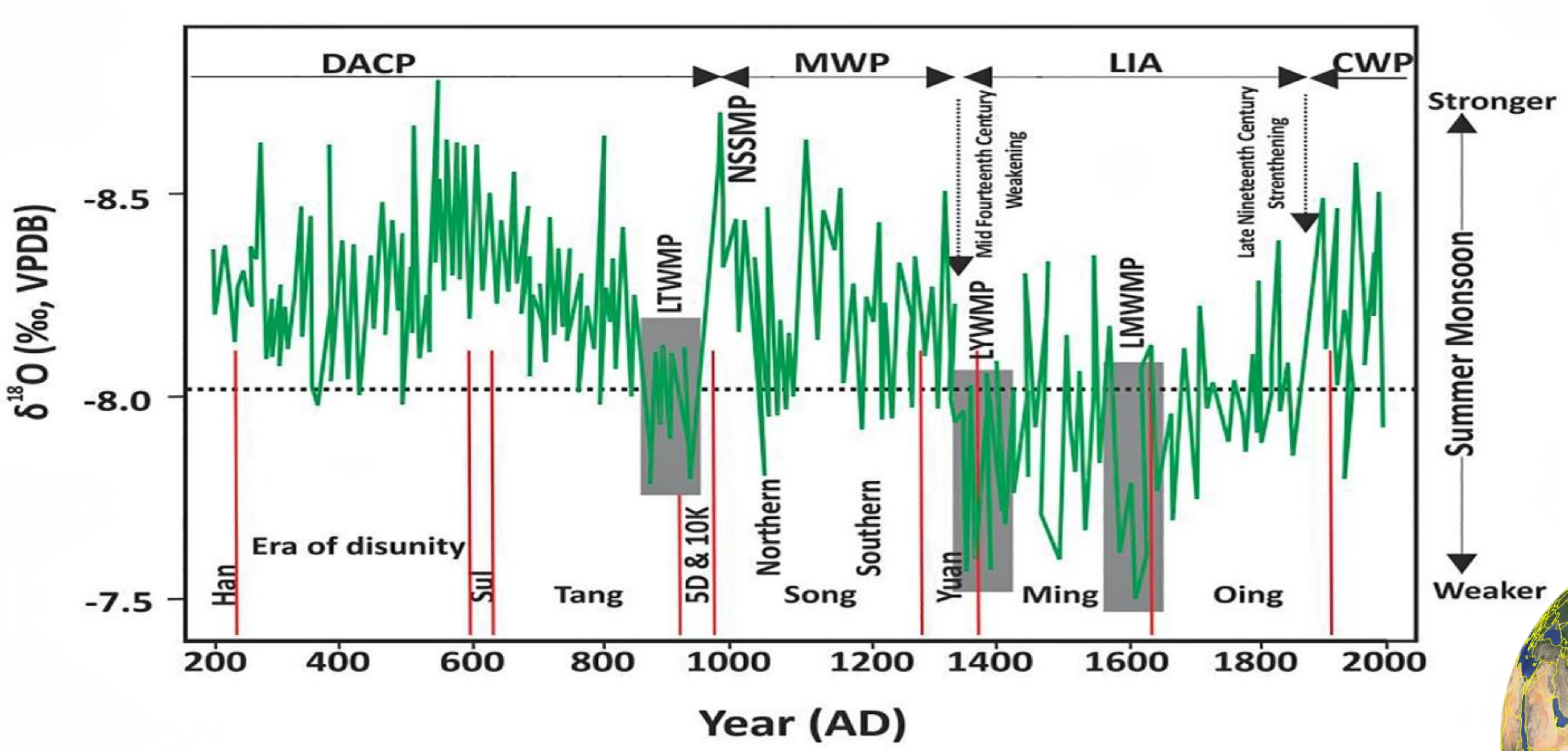


$\delta^{18}O$ record of cave stalagmite from Dandak (19ºN, 82ºE). Dashed lines indicate range of $\delta^{18}O$ during the modern instrumental period. Green and brown shades represent stronger and weaker monsoon respectively. Stars are historical records of famine in India. A 1.5‰ change in $\delta^{18}O$ during the fourteenth century is interpreted as 30% reduction in monsoon rainfall.

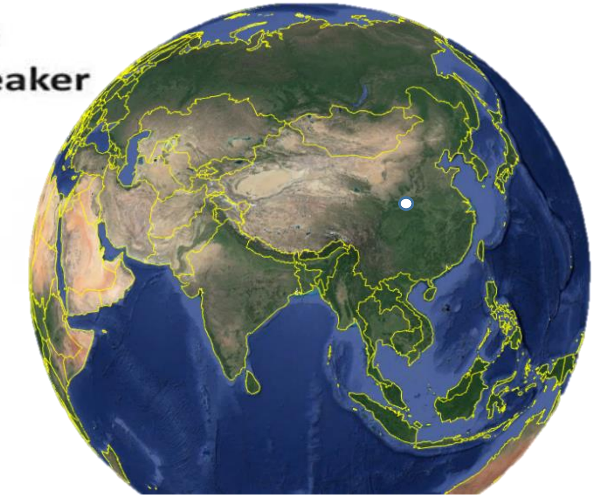


Oxygen isotopic anomaly of the Sahiya speleothem. Grey line is the raw data and the black line 11 yr running mean.

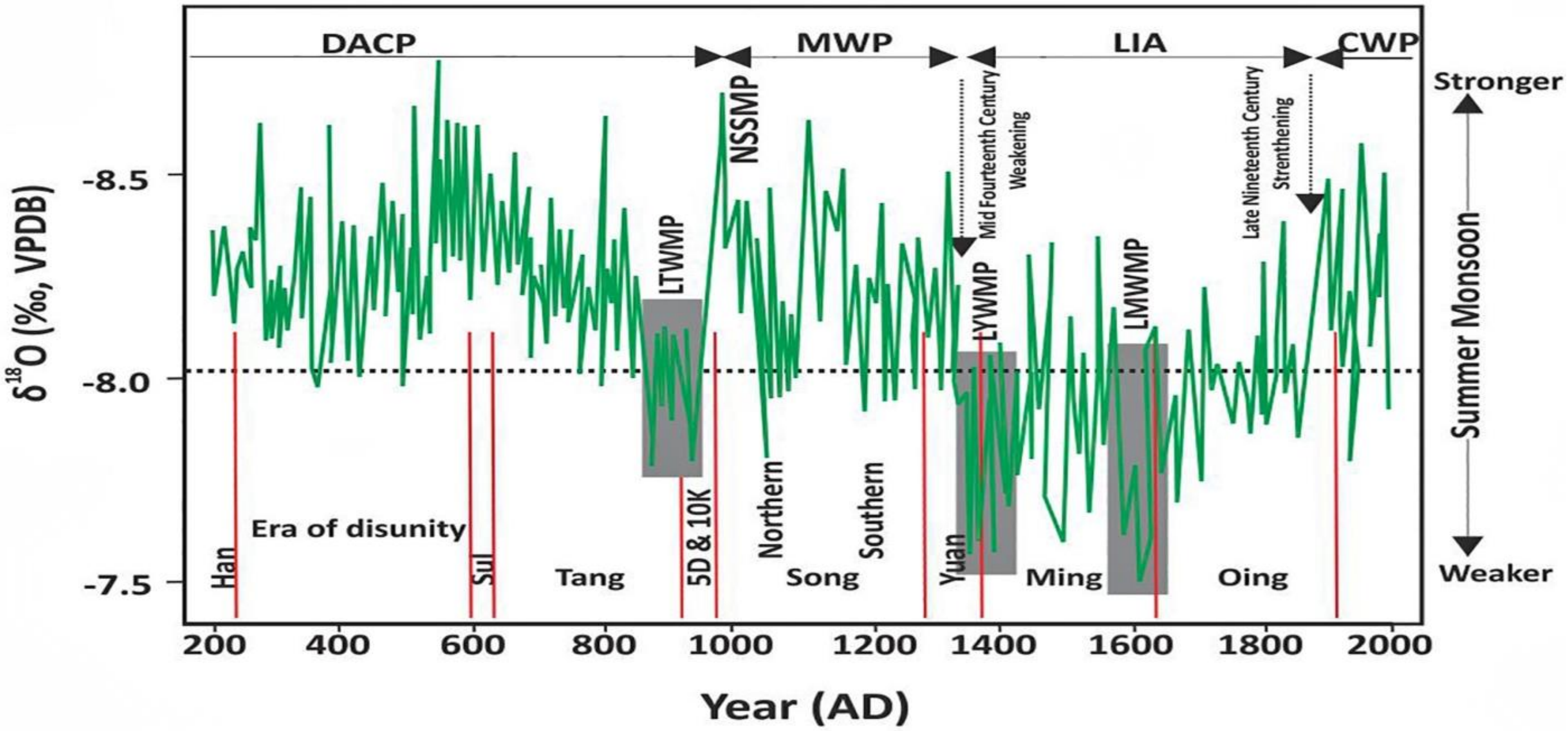
Sinha et al. 2015



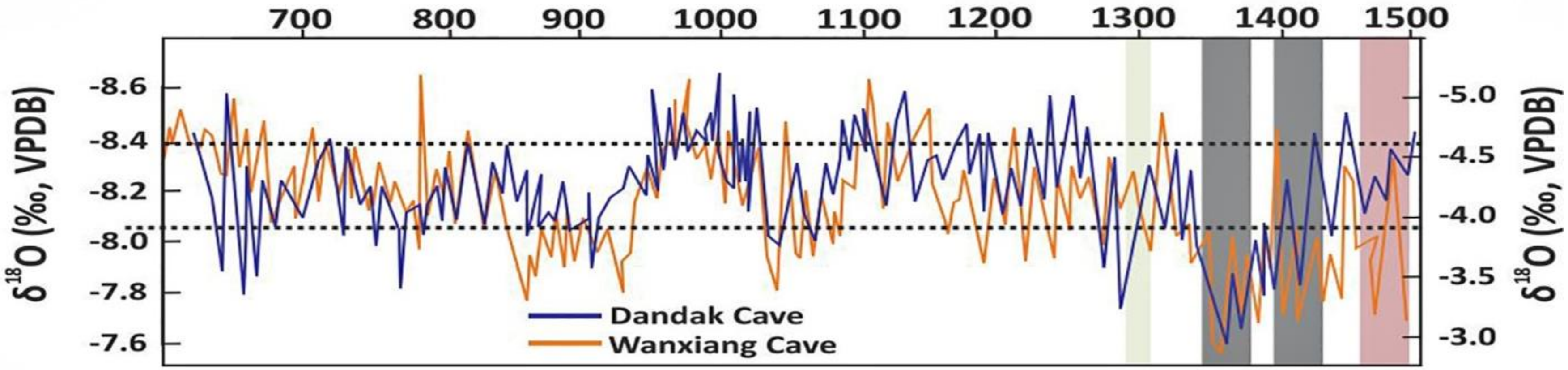
Zhang et al. 2008



Isotopic record from Wanxiang Cave, southern China (white dot in map)

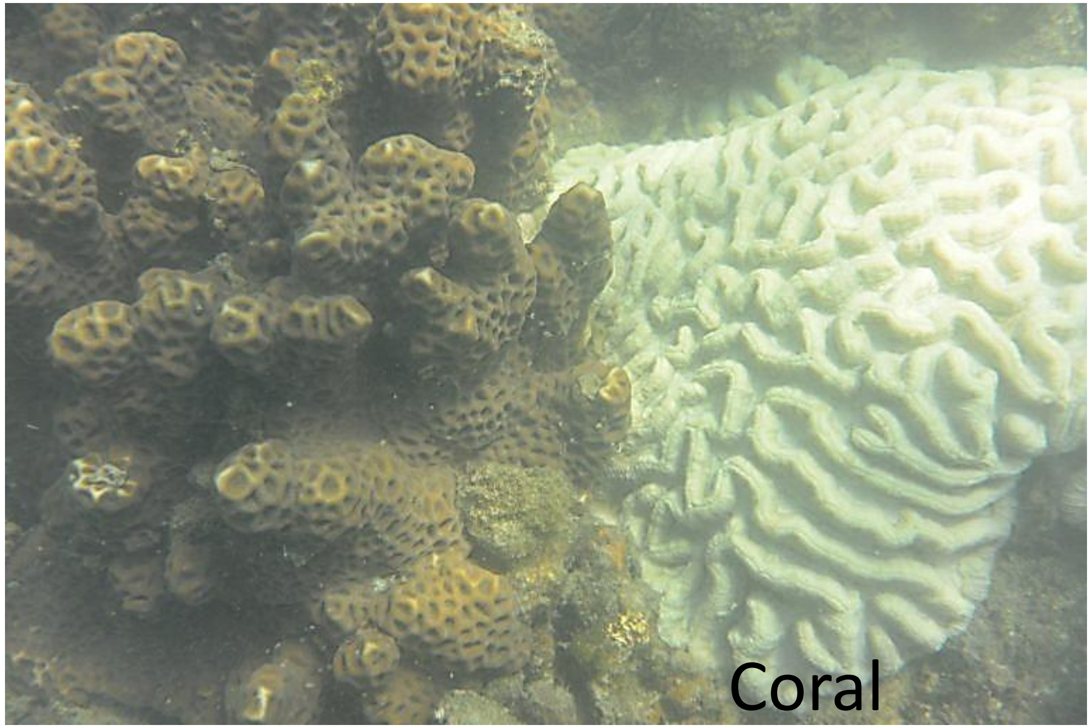


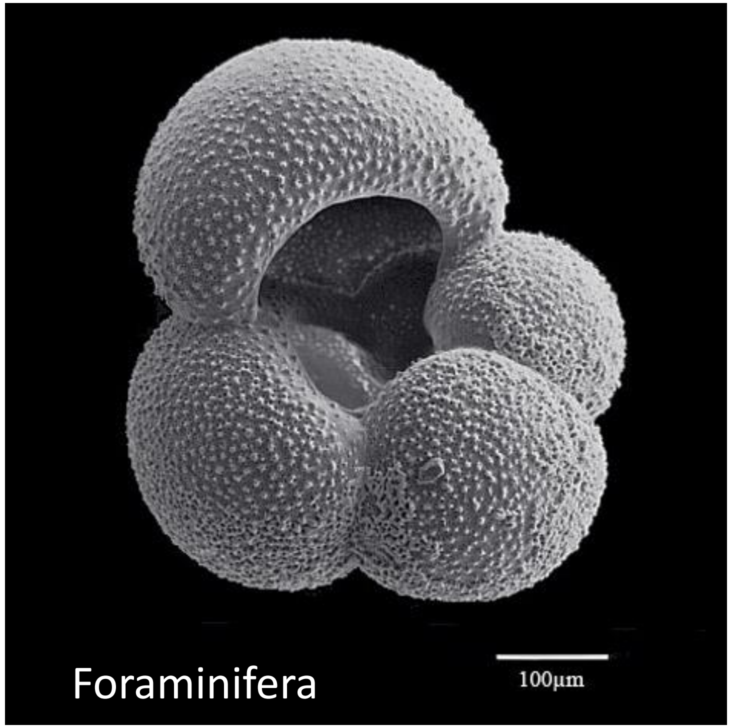
Zhang et al. 2008



Sinha et al. 2011

Paleoclimatic records from the marine environment



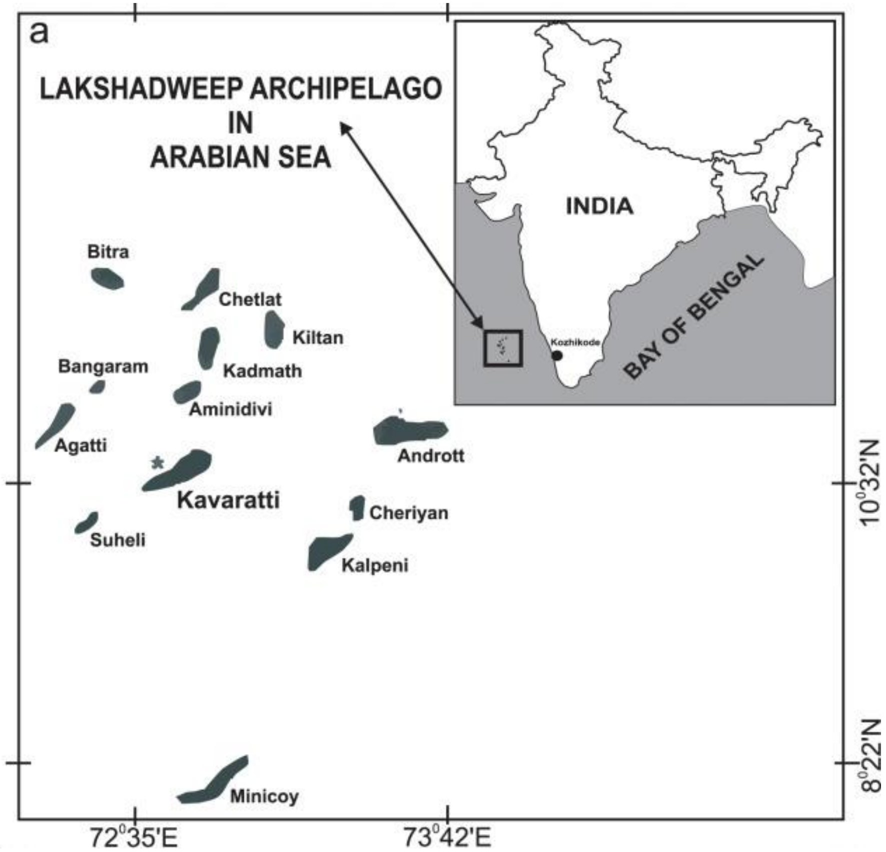


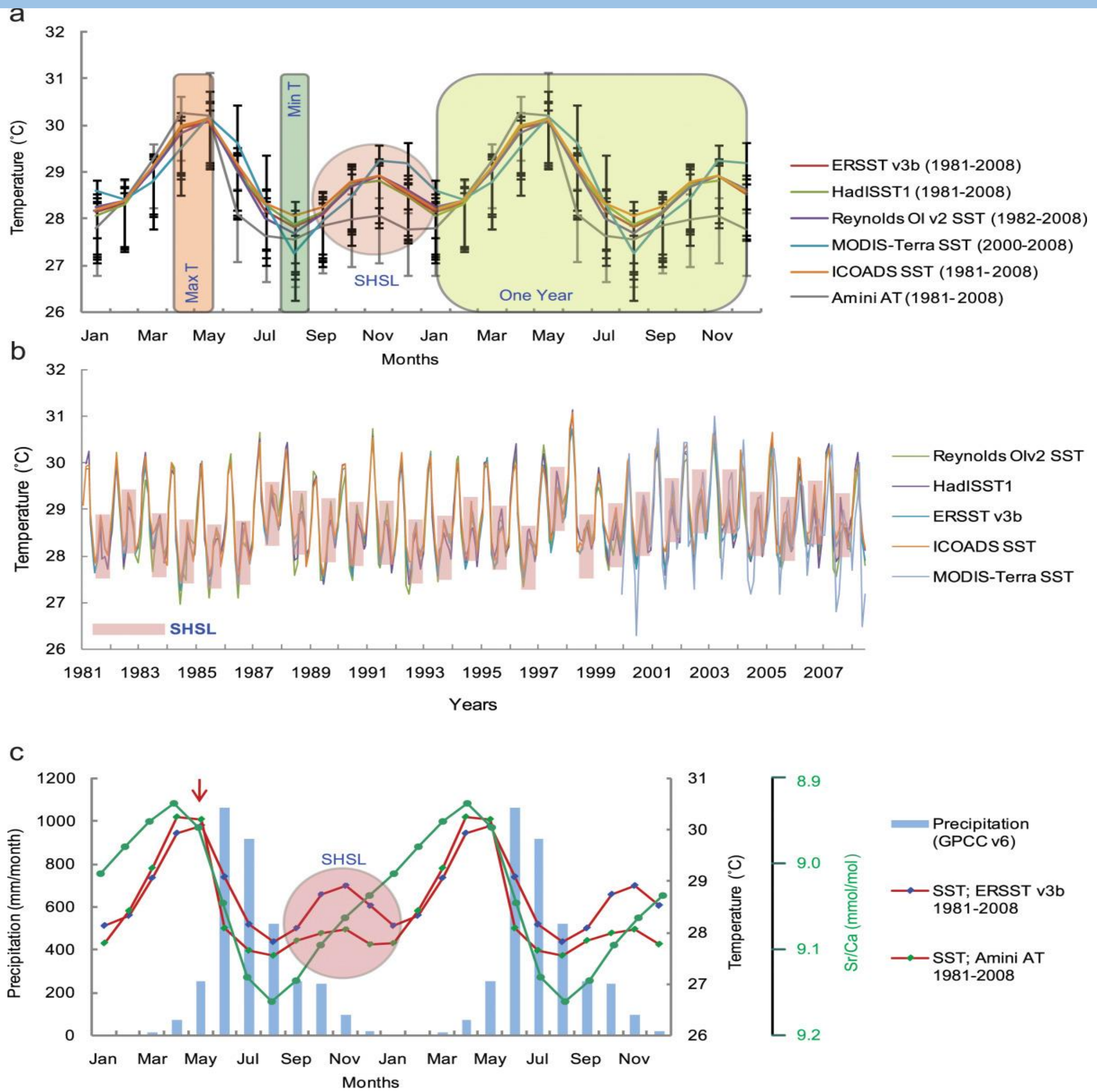
$\delta^{18}O$
$\delta^{13}C$
Mg/Ca
Sr/Ca
U/Ca

Marine carbonates

$$\delta^{18}O_{cal} = f\,(T, \delta^{18}O_{water})$$

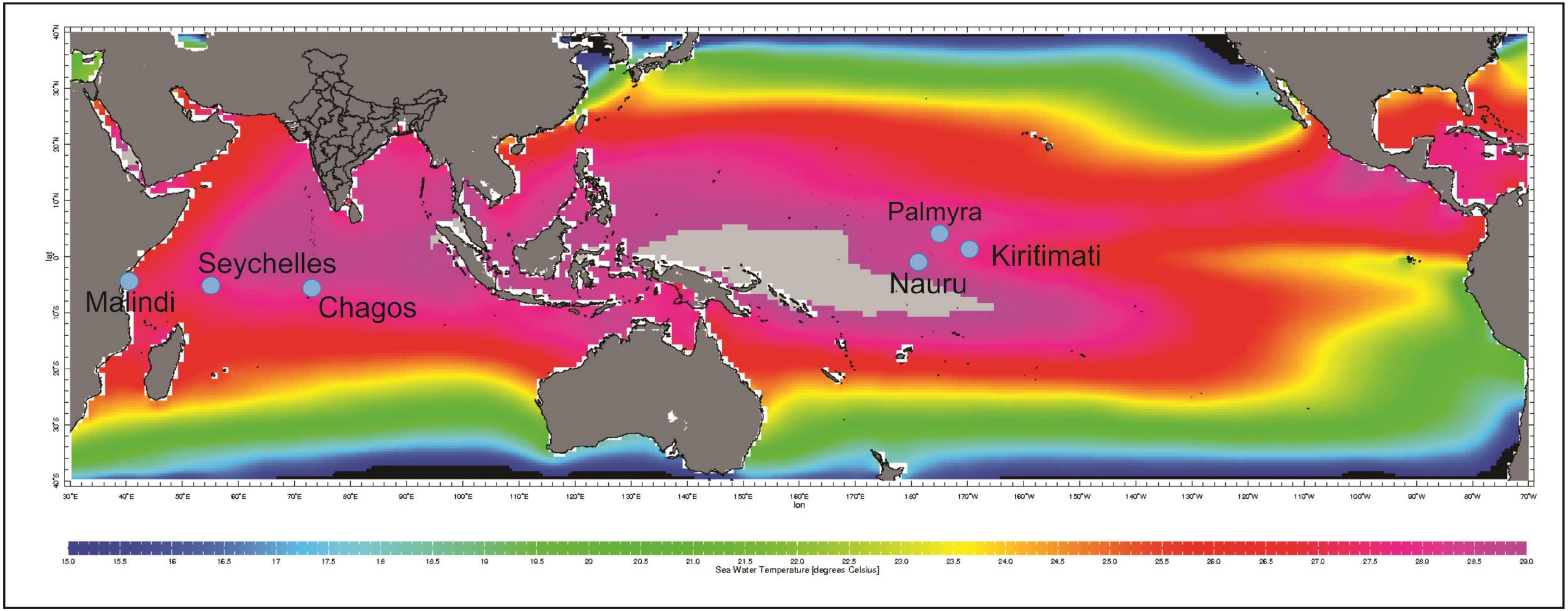
Trace element in corals: SST reconstruction



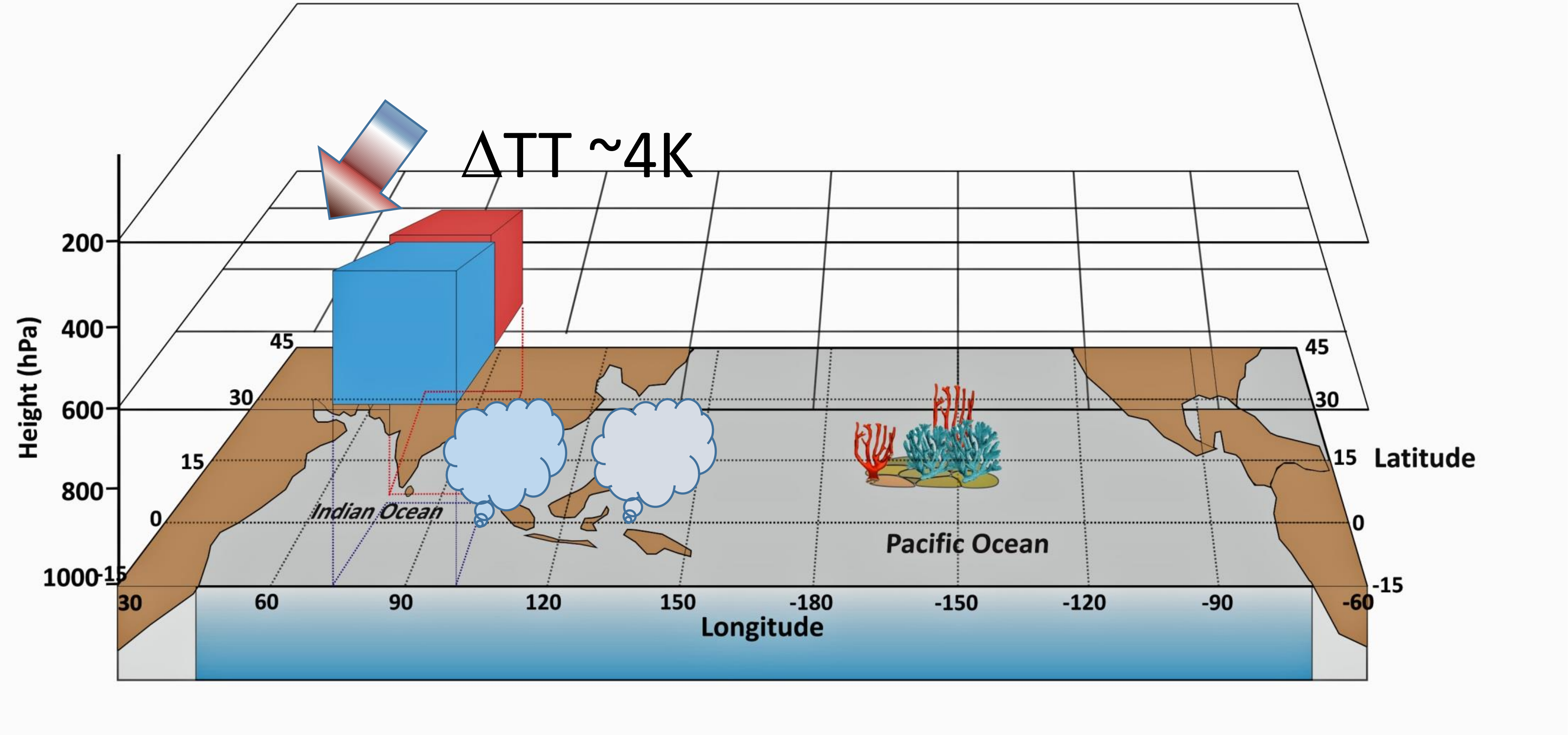


Sagar et al. 2016

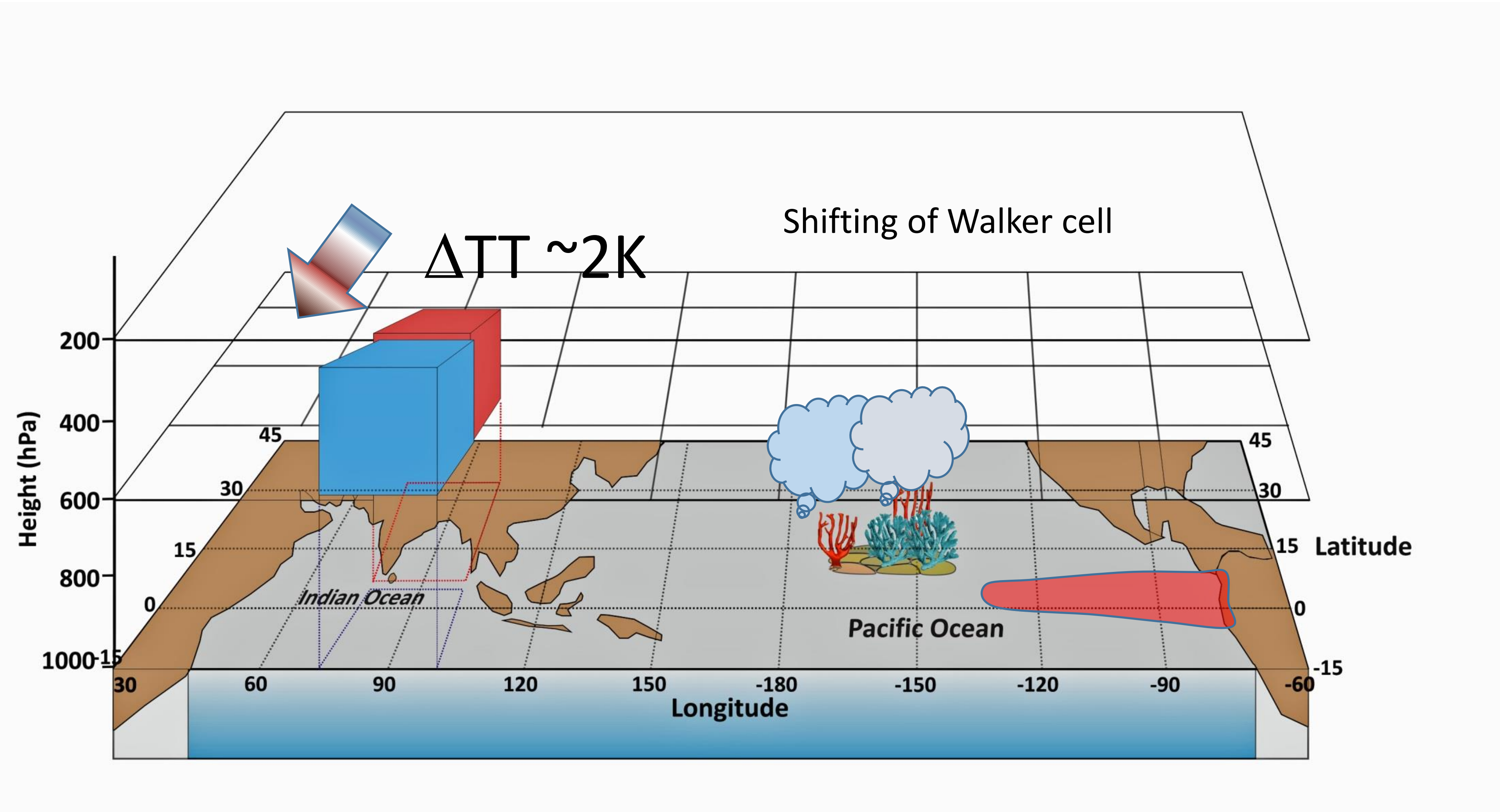
Rainfall reconstruction using coralline oxygen isotopes



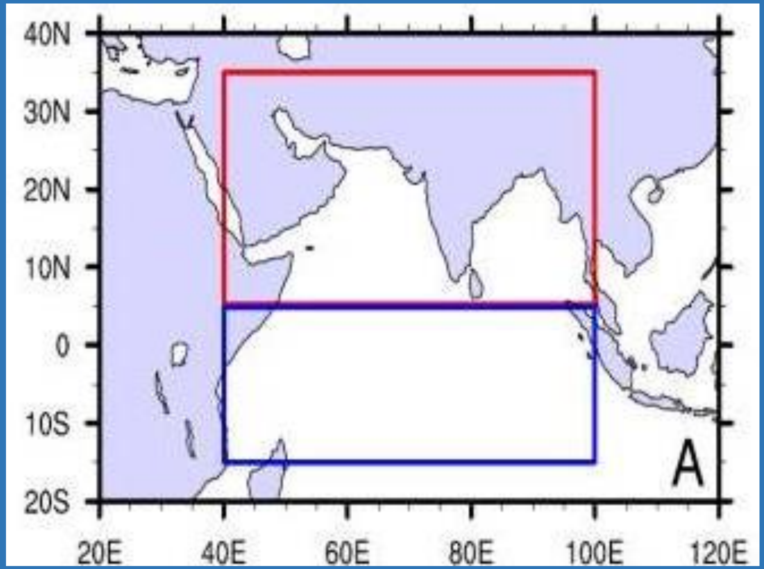
Tropical Pacific Climatology during a normal year

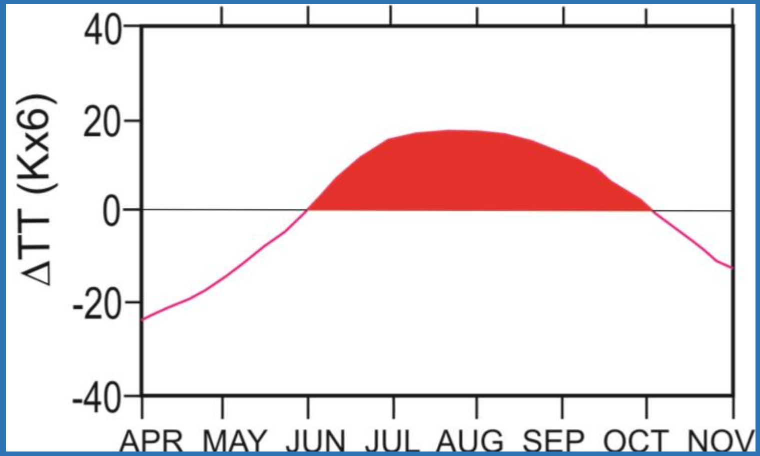


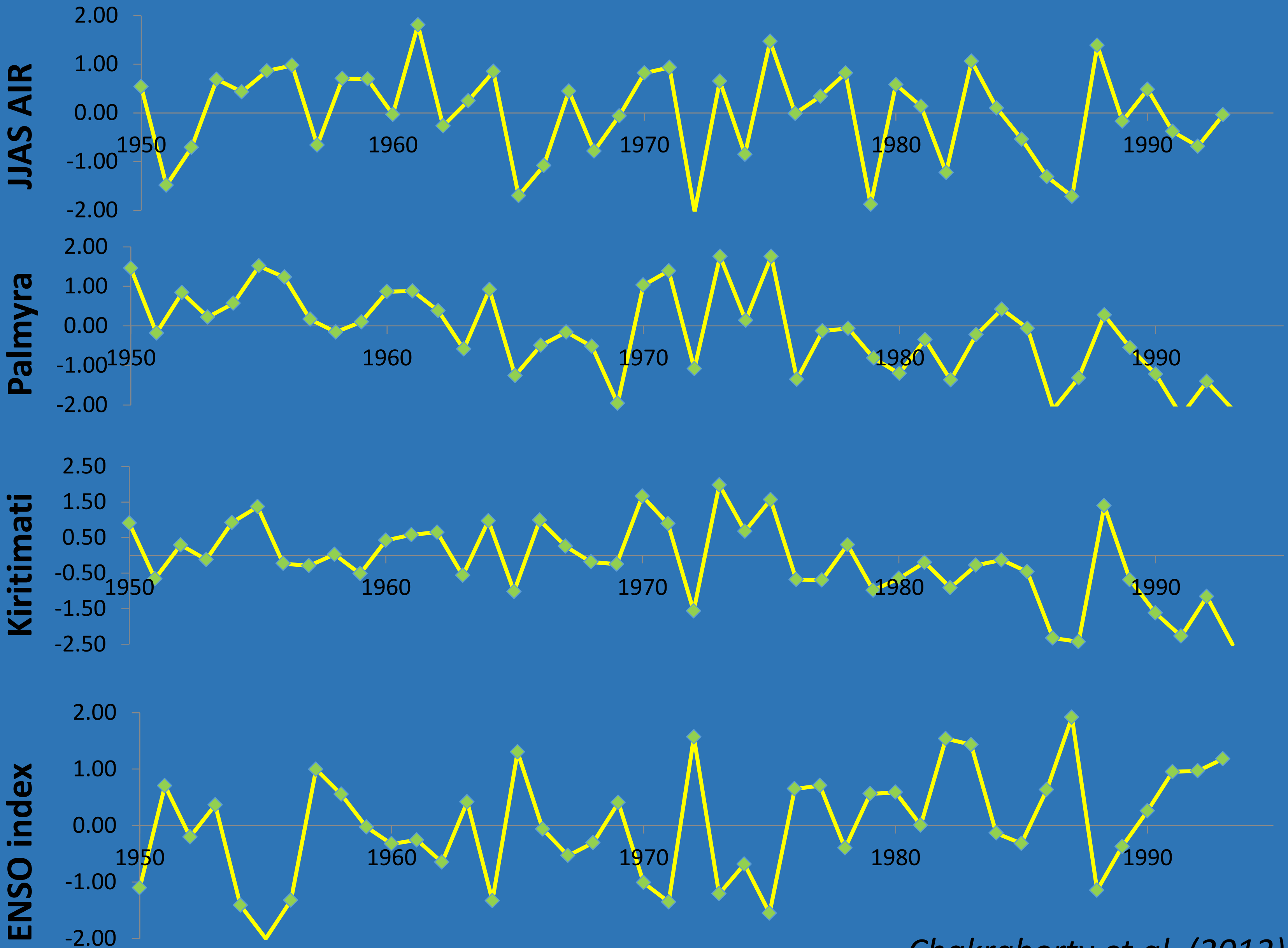
Tropical Pacific Climatology during an El nino year



The physical basis

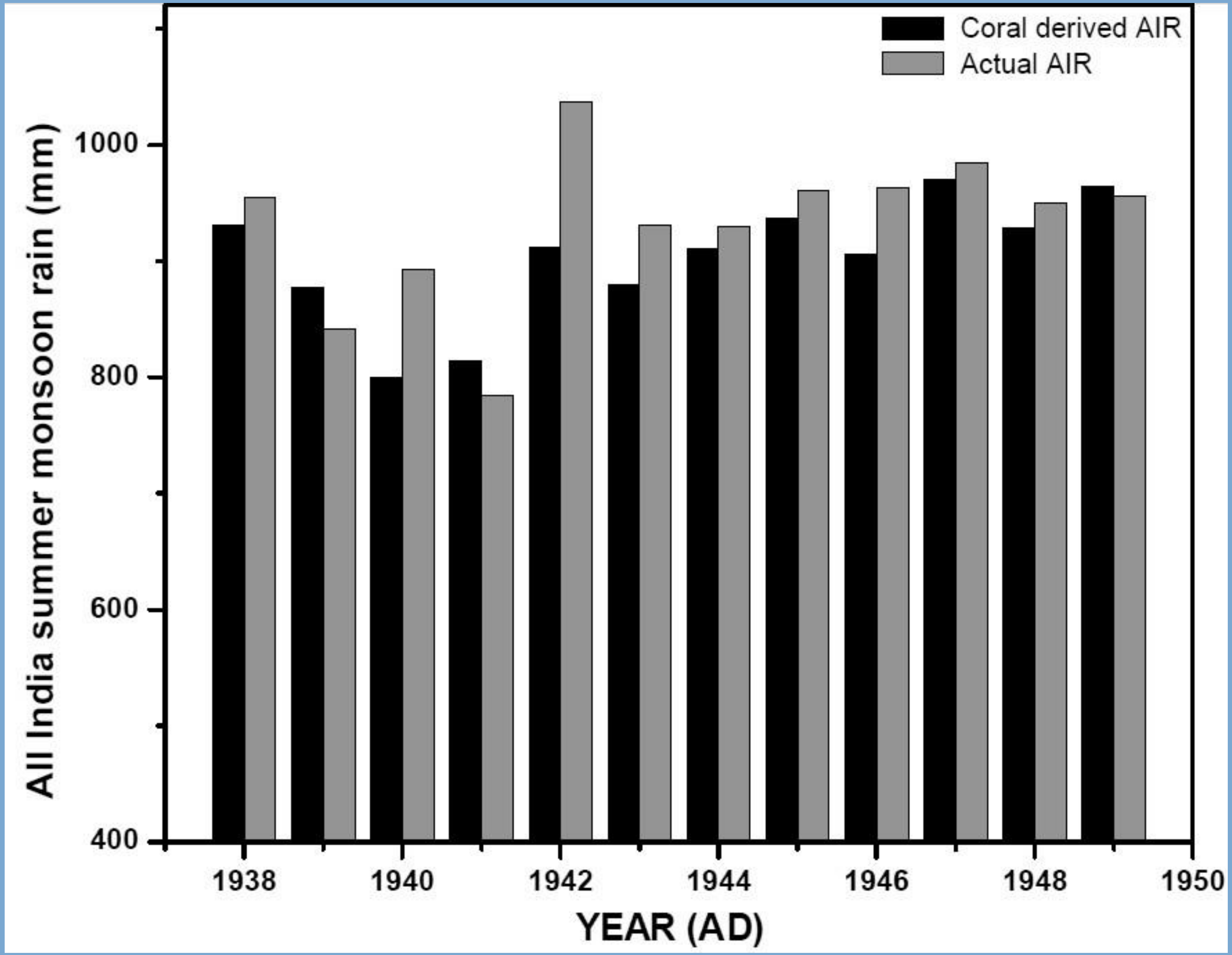






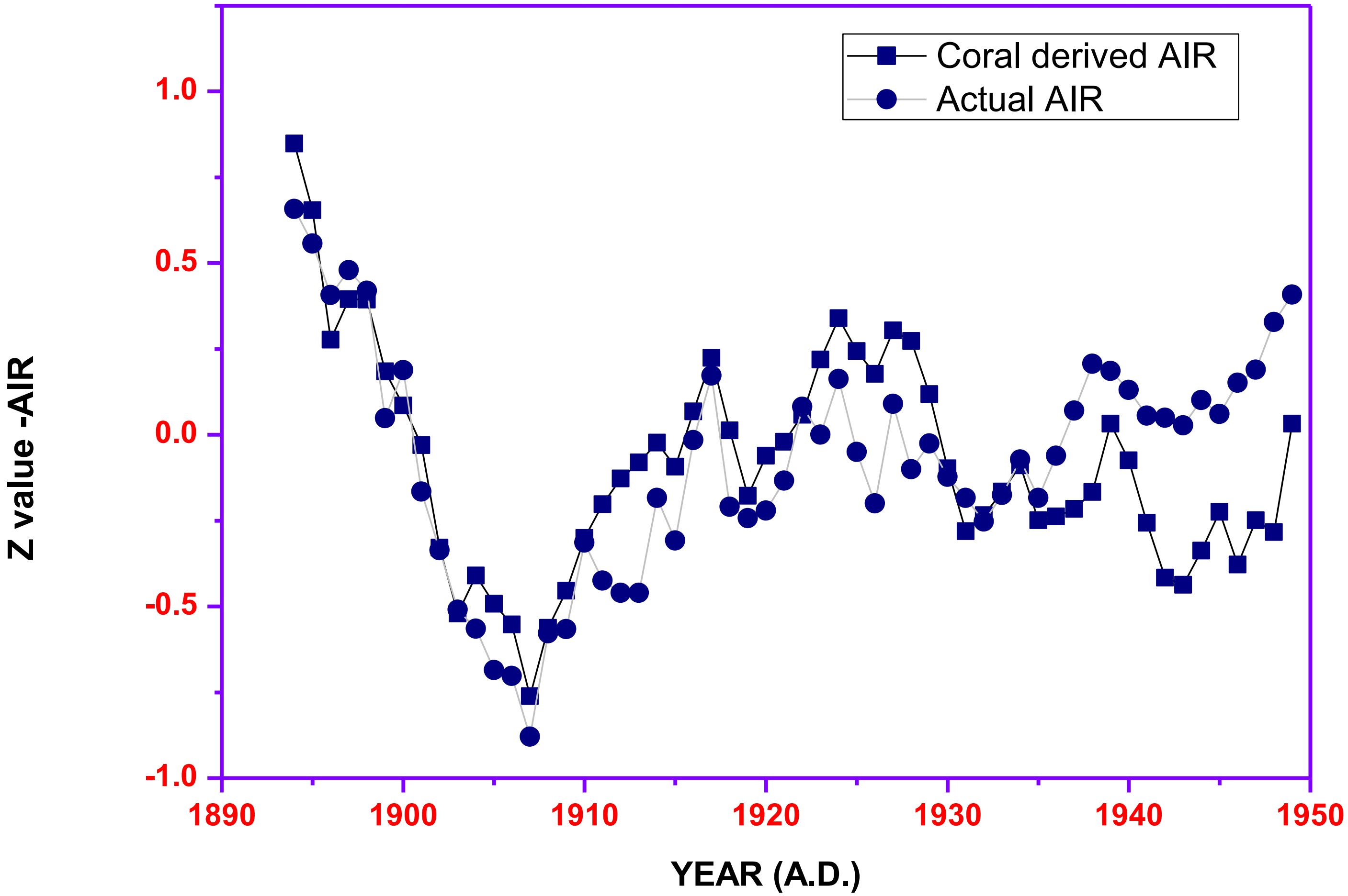
Chakraborty et al. (2012)

Rainfall reconstruction based on Kiritimati coral $\delta^{18}O$ between 1938-1949



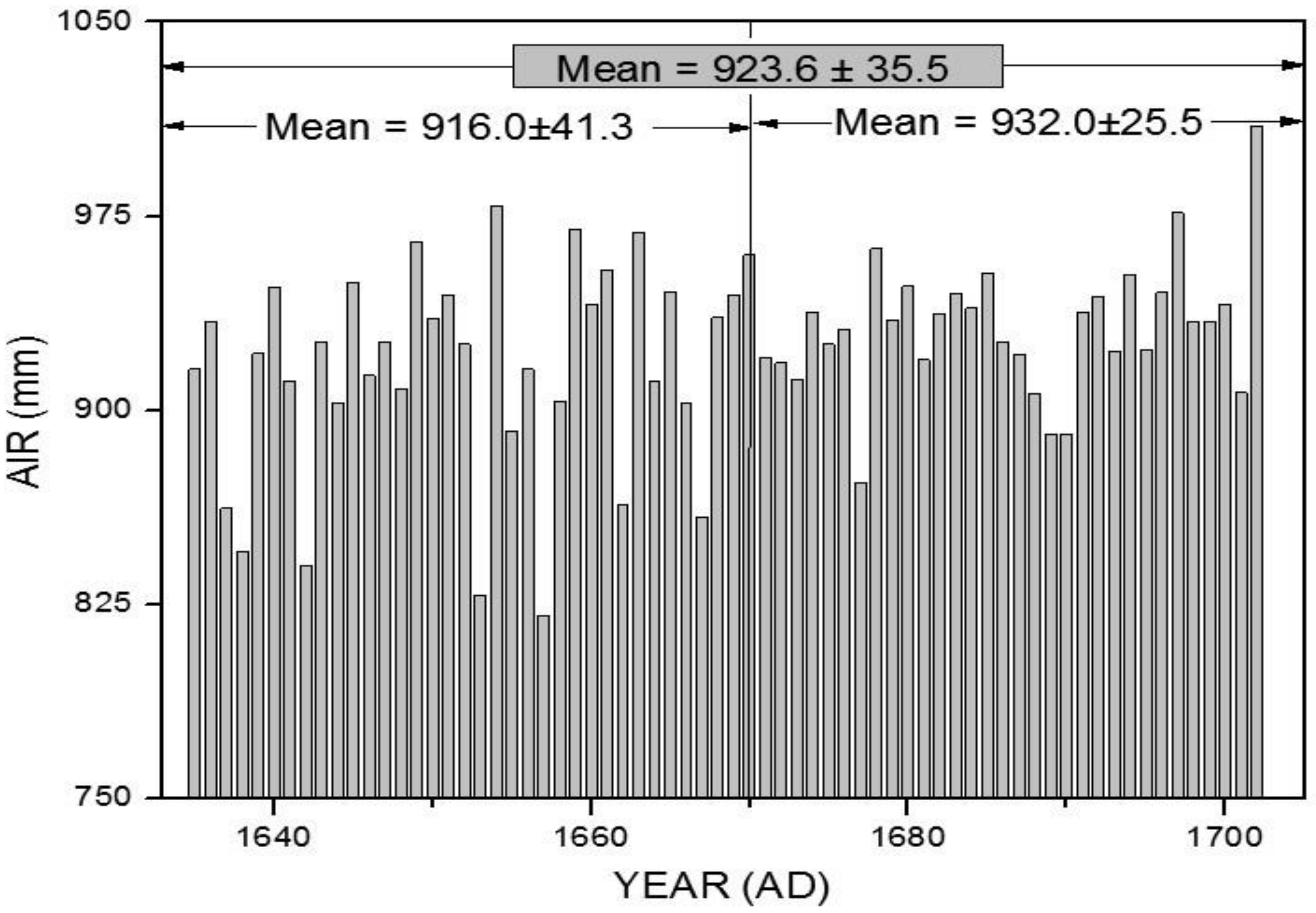
Chakraborty et al. (2012)

Comparison of the observed rain and coral derived rain in decadal time scale



The CC between reconstructed time series and observed rainfall is ($r=0.575$, $n=64$, $p<0.001$)

Chakraborty et al. (2012)

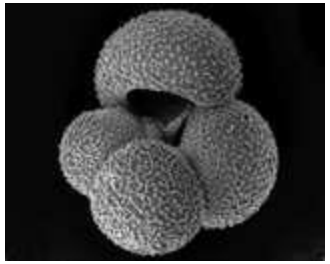


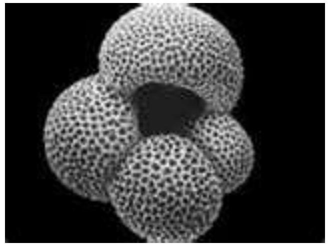
Coral based reconstruction of ISMR (1635-1670)

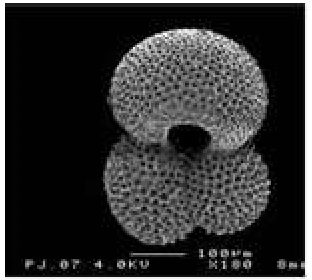
Reconstruction of the seventeenth century summer monsoon rainfall (mm)

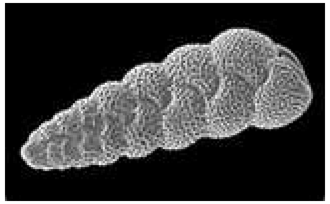
Chakraborty et al. (2012)

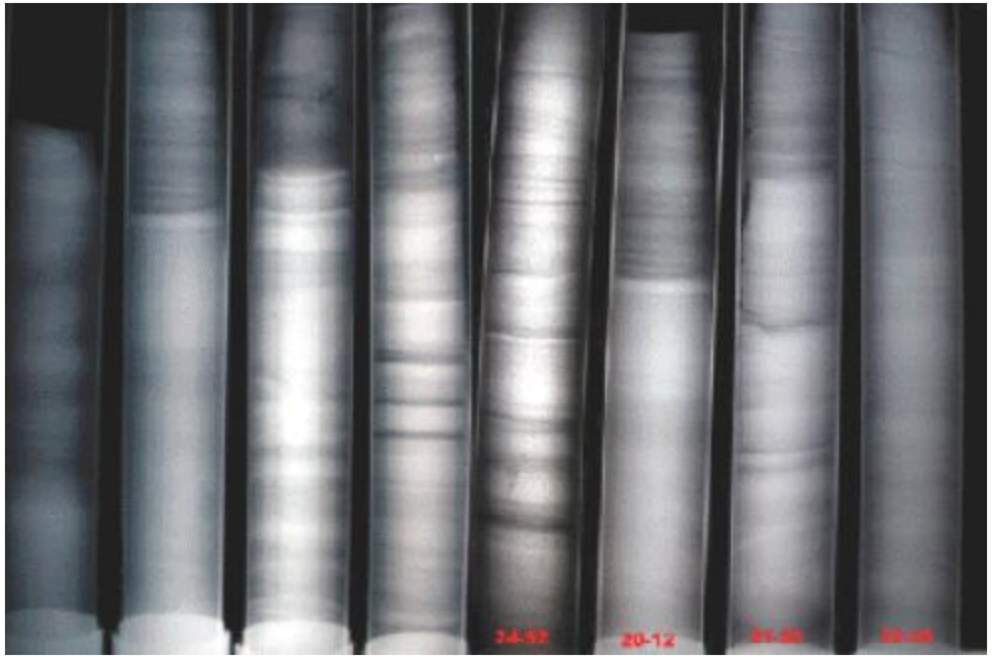
Foraminifera: a repository of past ocean variability





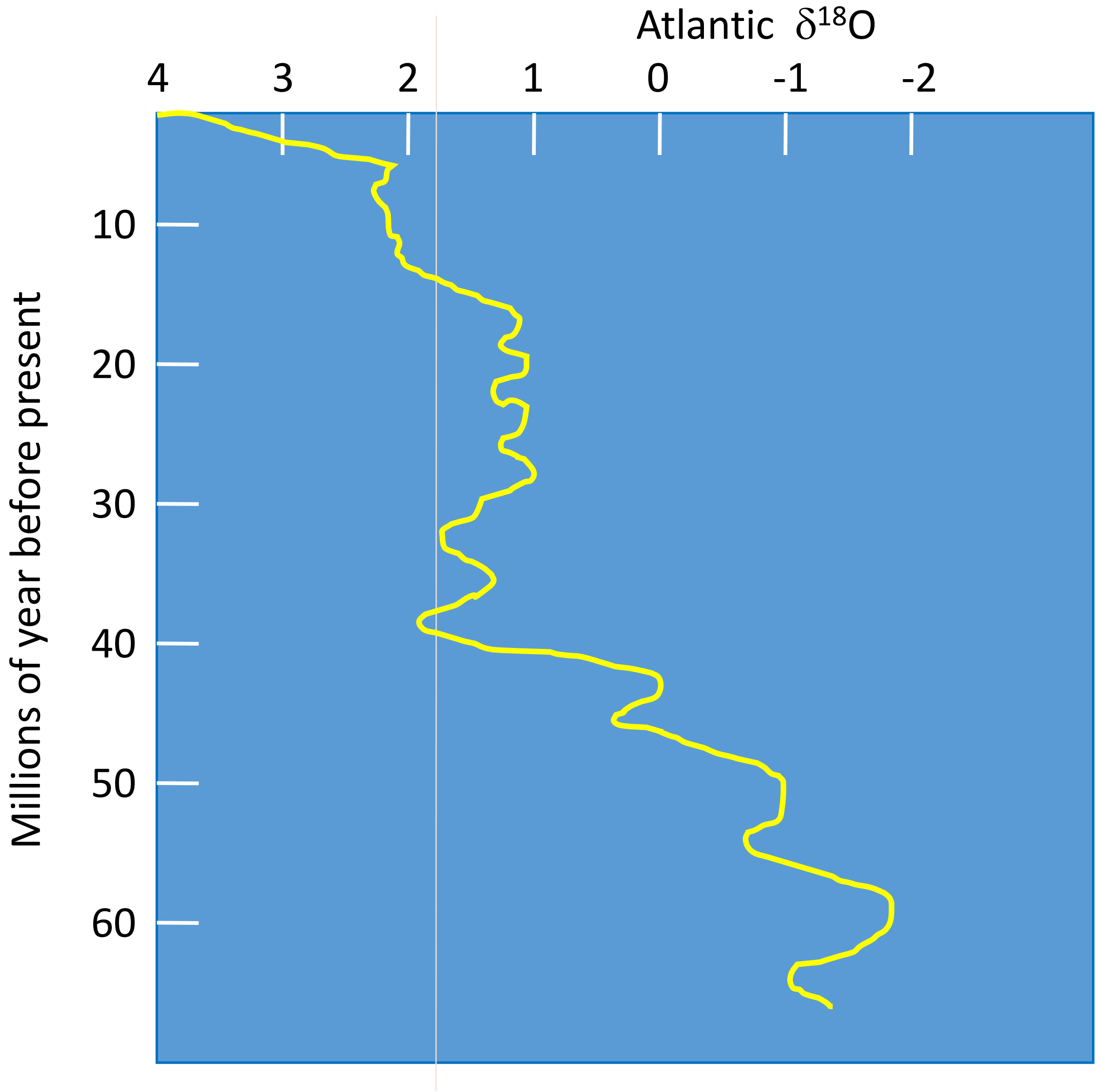






Deep sea sediment core

Paleotemperature over the past 70 million years: the $\delta^{18}O$ record of benthic foraminifera



14°C cooling of the ocean, or deep ocean temp. was about 16°C

Possible mechanisms

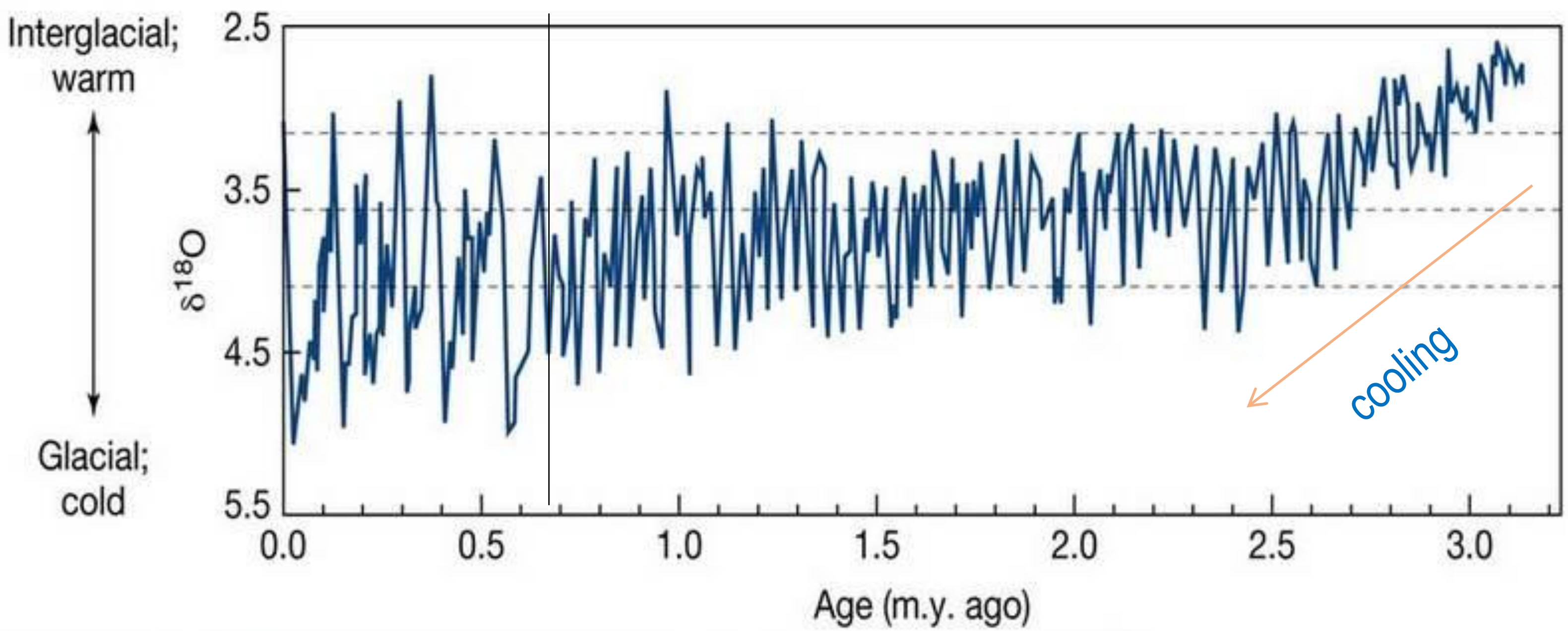
- Decreased input of CO_2 from the earth's interior
- Increased removal of CO_2 from the atmosphere due to enhanced weathering
- Progressive decrease in pole-ward heat transport arising due to the change in land-ocean distribution

Ice ages through geological time

First deep sea core recovered: 1950s

Continental records: showed four glaciations

Marine records → dozens of climate swings (over the Pleistocene)



Northern hemispheric glaciation – glacials – every 100 kyr in 700 kyr

Global avg T ~ 9-10 ^oC; pCO_2 ~200 ppm

Interglacials: Greenland, Antartica avg T ~ 15^oC, pCO_2 ~280 ppm


Thank You