

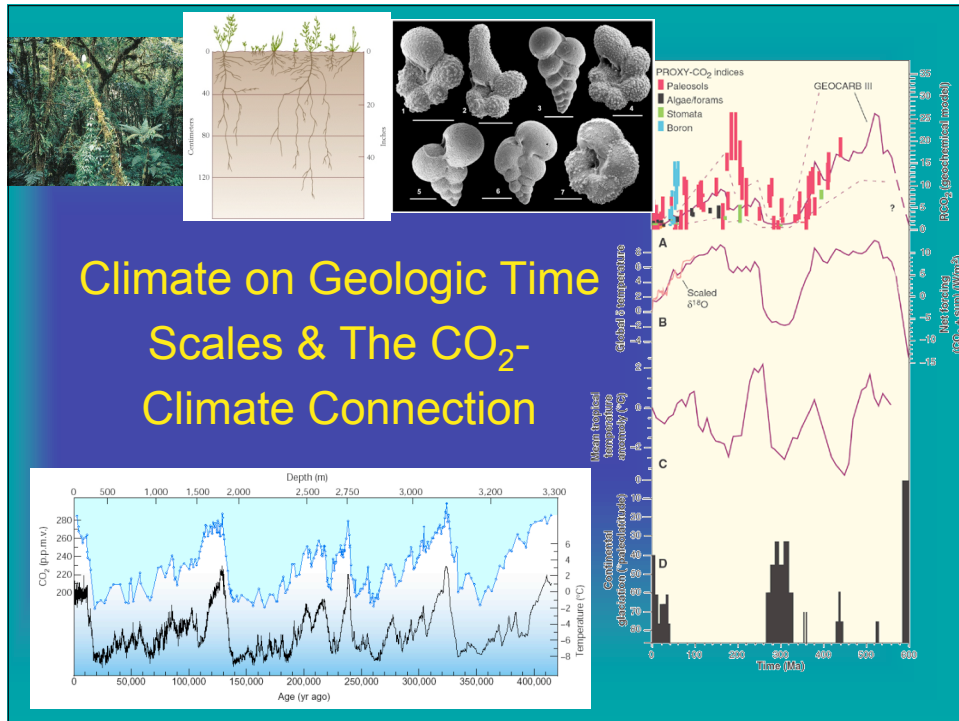
# Earth's Climate: the Last 550 Myr

OCEAN 355  
Prof. Julian Sachs  
Lecture Notes #7  
Autumn 2008

## Where We've Been & Where We Will Go

- Reviewed what processes control CO<sub>2</sub> greenhouse effect over geologic time (i.e., geochem C cycle)
- And what negative feedbacks (e.g., T-weathering, CO<sub>2</sub>-weathering) might keep climate system from reaching &/or remaining in extreme states
- Applied these concepts to understand the Neoproterozoic glaciations (SBEs)
- But the inferences we've made for a strong control of climate by CO<sub>2</sub> (e.g., during Faint Young Sun, SBE) have not been backed by actual data on CO<sub>2</sub> levels \*
- Now turn to geologic evidence for CO<sub>2</sub>-climate link during last 500 Myr

\* Prior to ~550 Ma the lack of animals with hard skeletons & vascular plants results in little or no fossil evidence of atmospheric CO<sub>2</sub> levels.

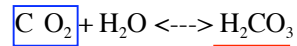


## Climate Controls - Long & Short Timescales

- Solar output (luminosity):  $10^9$  yr
- Continental drift (tectonics):  $10^8$  yr
- Orogeny (tectonics):  $10^7$  yr
- Orbital geometry (Earth -Sun distance):  $10^4$ - $10^5$  yr
- Ocean circulation (geography, climate):  $10^1$ - $10^3$  yr
- Composition of the atmosphere (biology, tectonics, volcanoes):  $10^0$ - $10^5$  yr



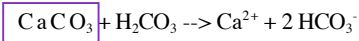
**Chemical Weathering** = chemical attack of rocks by dilute acid



The **Geochemical**  
(or non-biological  
part of the)  
**Carbon Cycle**

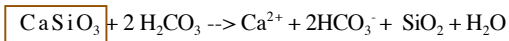
**1. Carbonate Weathering:**

Carbonate Rocks (e.g., limestone)



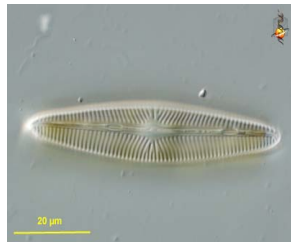
**2. Silicate Weathering:**

Silicate Rocks (most of the  
mantle & crust. E.g., granite)

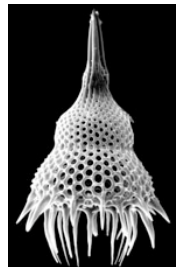


- 2x CO<sub>2</sub> consumption for silicates
- Carbonates weather faster than silicates

[http://en.wikipedia.org/wiki/Image:Yosemite\\_20\\_bg\\_090404.jpg](http://en.wikipedia.org/wiki/Image:Yosemite_20_bg_090404.jpg) [http://en.wikipedia.org/wiki/Image:Burren\\_karst.jpg](http://en.wikipedia.org/wiki/Image:Burren_karst.jpg)



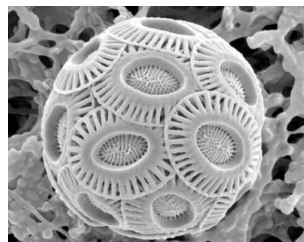
Diatom  
(SiO<sub>2</sub>)



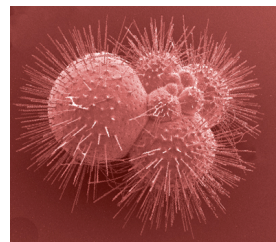
Radiolarian  
(SiO<sub>2</sub>)

Products of  
weathering  
precipitated as  
CaCO<sub>3</sub> & SiO<sub>2</sub> in  
ocean

R, Protozoans  
L, Eukaryotic Phytoplankton

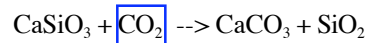


Coccolithophorid  
(CaCO<sub>3</sub>)



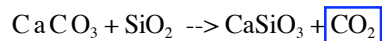
Foraminifer  
(CaCO<sub>3</sub>)

Net Reaction of Rock Weathering  
+  
Carbonate and Silica Precipitation in Ocean



- CO<sub>2</sub> consumed (~ 0.03 Gt C/yr)
- Would deplete atmospheric CO<sub>2</sub> in 20 kyr
- Plate tectonics returns CO<sub>2</sub> via Volcanism and Metamorphism

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Carbonate Metamorphism

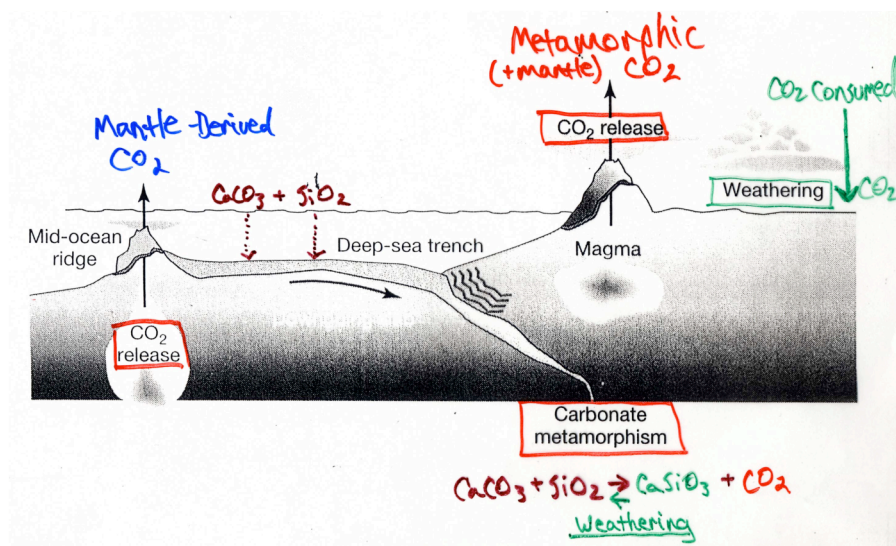


- CO<sub>2</sub> produced from subducted marine sediments

**Net reaction of  
geochemical  
carbon cycle  
(Urey Reaction)**

- On geologic time scales, rock weathering balanced by carbonate metamorphism
- Any *imbalance* can cause changes in atmospheric CO<sub>2</sub>

**Carbonate-Silicate Geochemical Cycle**



Kump et al. (1999)

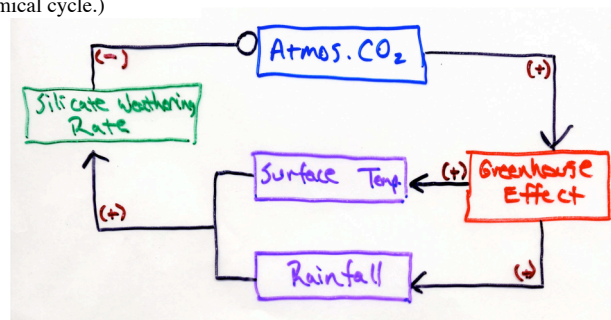
- Geologic record indicates climate has rarely reached or maintained extreme Greenhouse or Icehouse conditions....
- Negative feedbacks between climate and Geochemical Carbon Cycle must exist
- Thus far, only identified for Carbonate-Silicate Geochemical Cycle:

Temp., rainfall enhance weathering rates  
(Walker et al, 1981)

(I.e., no obvious climate dependence of tectonics or organic carbon geochemical cycle.)

How are CO<sub>2</sub> levels kept in balance?

Feedbacks



Adapted from Kump et al. (1999)

#### →Facts:

- Trace atmospheric gas that efficiently traps outgoing IR

#### →Hypotheses and theories:

- Solution to FYSP
- Through influence on CO<sub>2</sub>: weathering, tectonics and organic carbon burial/oxidation control climate on geologic timescales
- Negative feedbacks:
  1. Temp. – Weathering
  2. CO<sub>2</sub> – Weathering

#### →Tests:

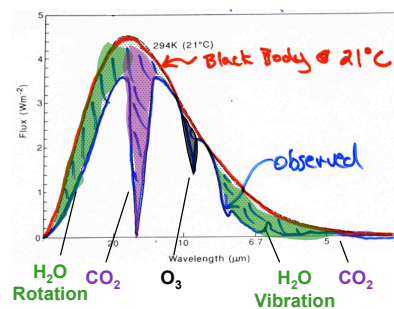
- Comparisons between "proxies" for CO<sub>2</sub> and T

#### →State of the science:

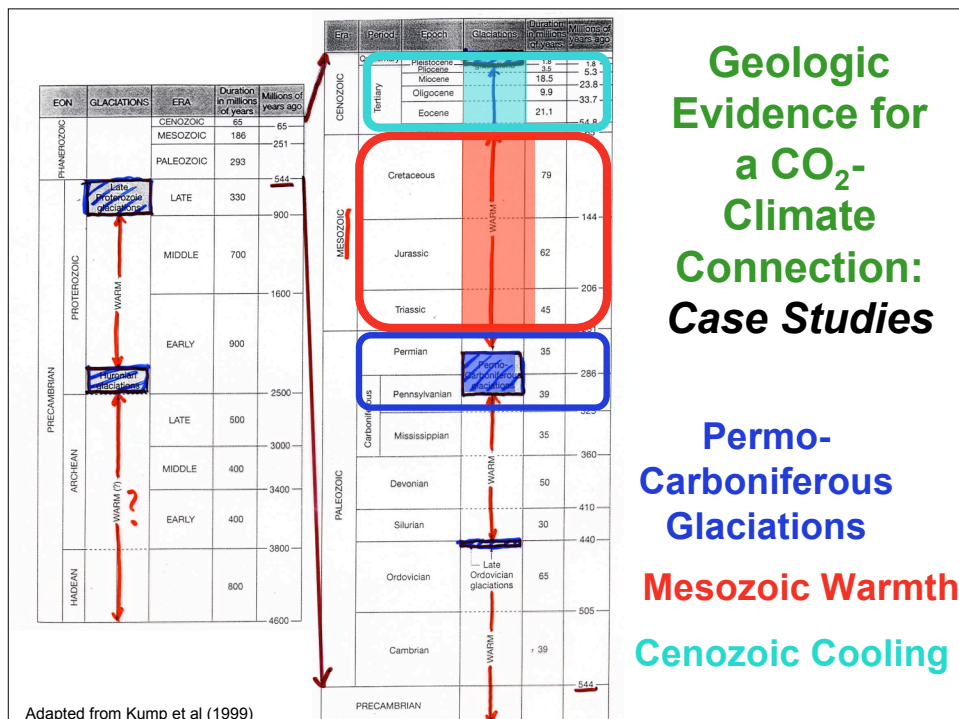
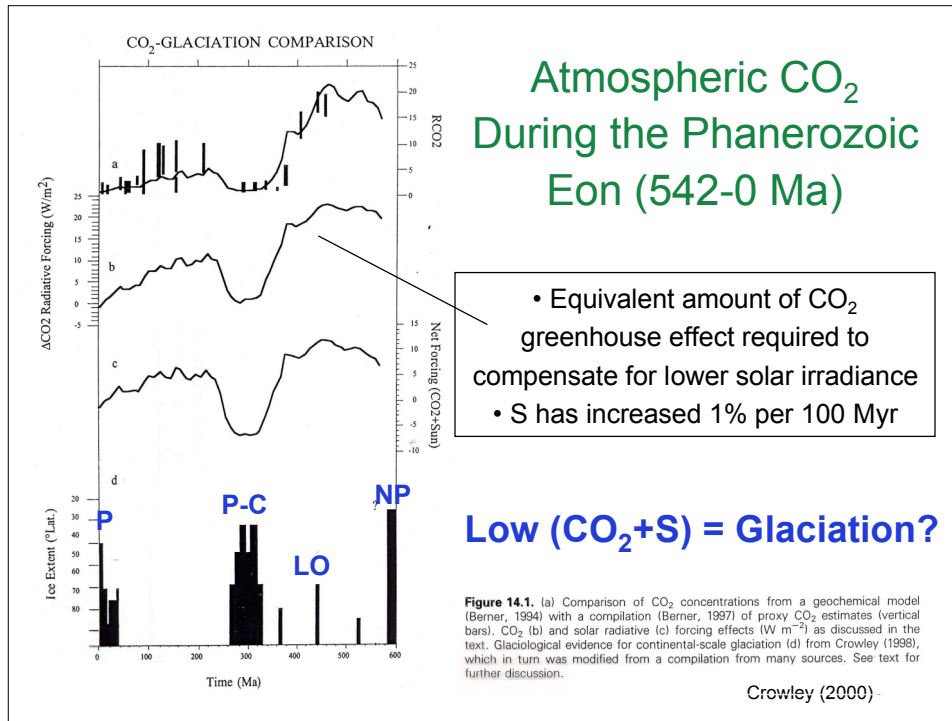
- Substantial support for close link... with notable exceptions....

## The CO<sub>2</sub>-Climate Connection

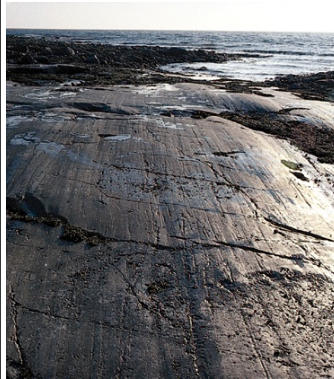
Earth's Surface Radiation Spectrum @ Top of Troposphere



Adapted from Bigg (1996)



# Permo-Carboniferous Glaciations (~300-275 Ma)



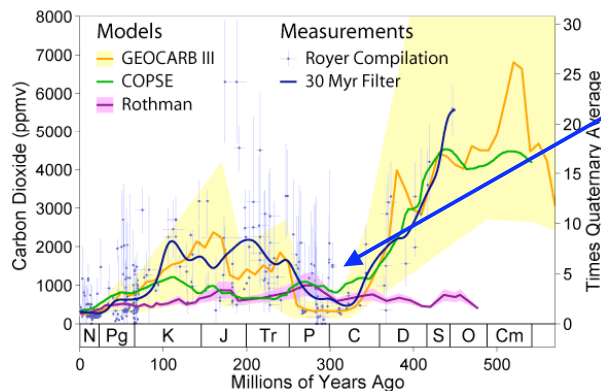
Stanley (2000)



Glacial Striations  
→ Origin of Permo-Carboniferous Glaciation  
COLD

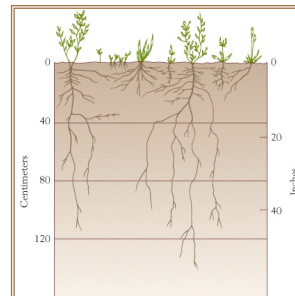
Fig. 11.11. Generalized diagram illustrating evidence for origination of Permo-Carboniferous glaciation on one large landmass. Arrows indicate direction of glacial flow. X = Permian pole position. [After Sullivan, 1974] Reproduced by permission from W. Sullivan, "Continents in Motion: The New Earth Debate," copyright 1974, McGraw-Hill Publishing Co.

## Phanerozoic CO<sub>2</sub> Evolution



• Permo-Carboniferous Glaciations Followed a period of marked CO<sub>2</sub> decline

- CO<sub>2</sub> decline likely resulted from spread of rooted vascular plants in the Devonian, 400-360 Ma.
- Dissolution of bedrock (weathering) from: secreted acids, metabolic CO<sub>2</sub> from C<sub>org</sub> decomposition, & anchoring of clay-rich soil to rock (which retains water).

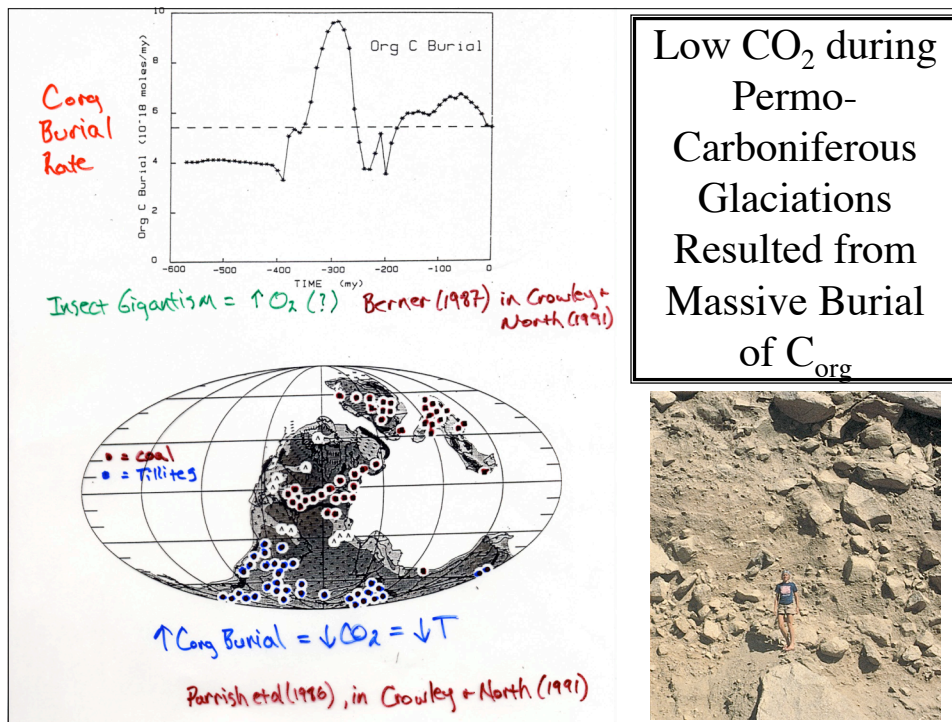


Stanley (2000)

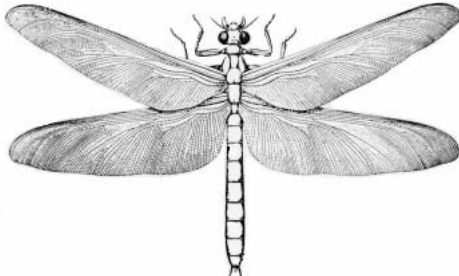

Theory: Berner (1997) *Science*, Vol. 276: 544-547  
[http://www.globalwarmingart.com/wiki/Image:Phanerozoic\\_Carbon\\_Dioxide\\_png](http://www.globalwarmingart.com/wiki/Image:Phanerozoic_Carbon_Dioxide_png)


\*\*\* Ended Here 11/19/08 \*\*\*





## Insect Gigantism: *Meganeura monyi*

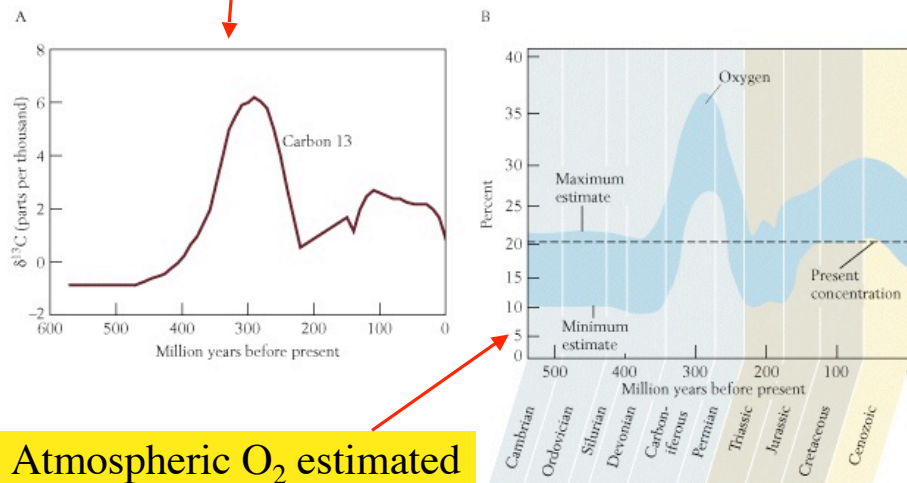

  
 > 0.75 m

*Meganeura* in BBC's Walking With Monsters

- *Meganeura monyi* was a prehistoric insect of the Carboniferous period (300 million years ago), resembling and related to the present-day dragonfly. With a wingspan of more than 75 cm (2.5 ft) wide, it was the largest known flying insect species to ever appear on Earth. (The Permian *Meganeuropsis permiana* is another contender). It was predatory, feeding on other insects and even small amphibians.

<http://en.wikipedia.org/wiki/Meganeura>

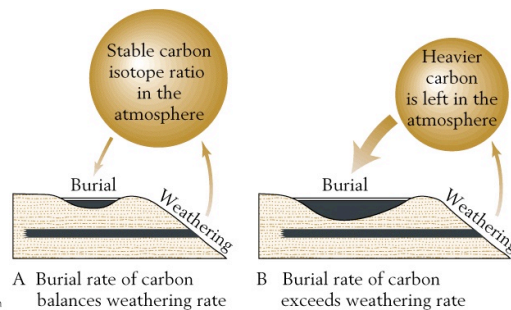
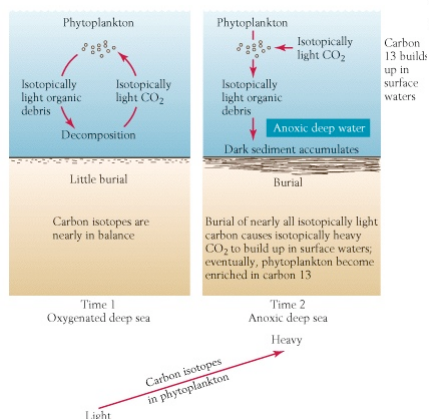
## $C_{org}$ burial rate estimated from $\delta^{13}C$ in $CaCO_3$



Atmospheric  $O_2$  estimated from  $C_{org}$  burial rate

Stanley (2000)

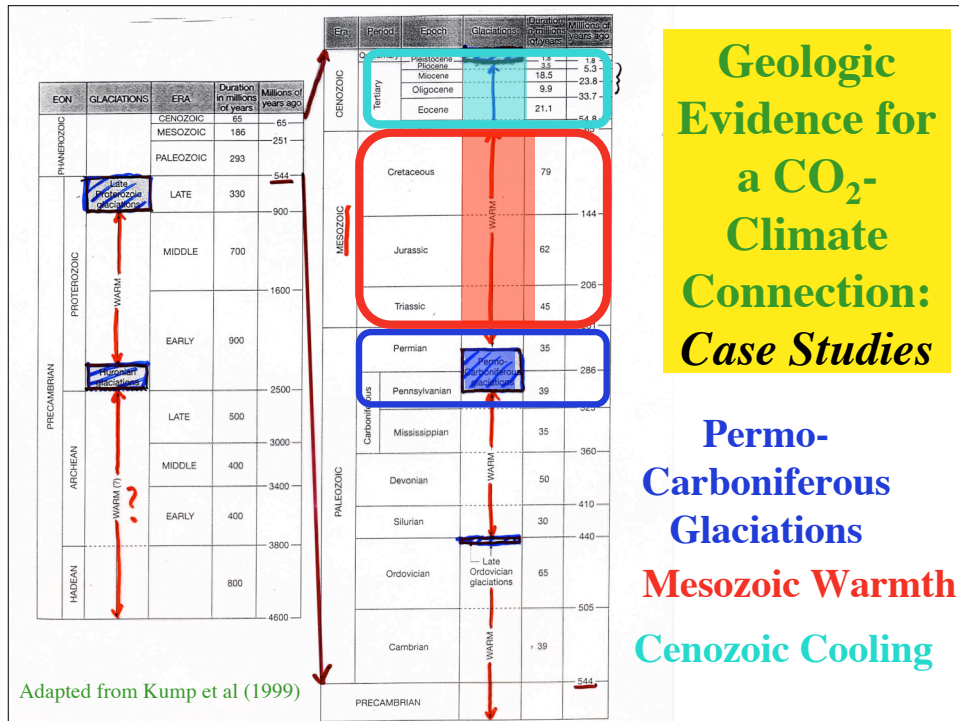
High  $C_{org}$  Burial Results in High  $^{13}C/^{12}C$  in Seawater &  $CaCO_3$



On longer time scales, higher burial rates of C, relative to weathering rates, results in elevated  $^{13}C/^{12}C$  in atmos/ocean

Stanley (2000)





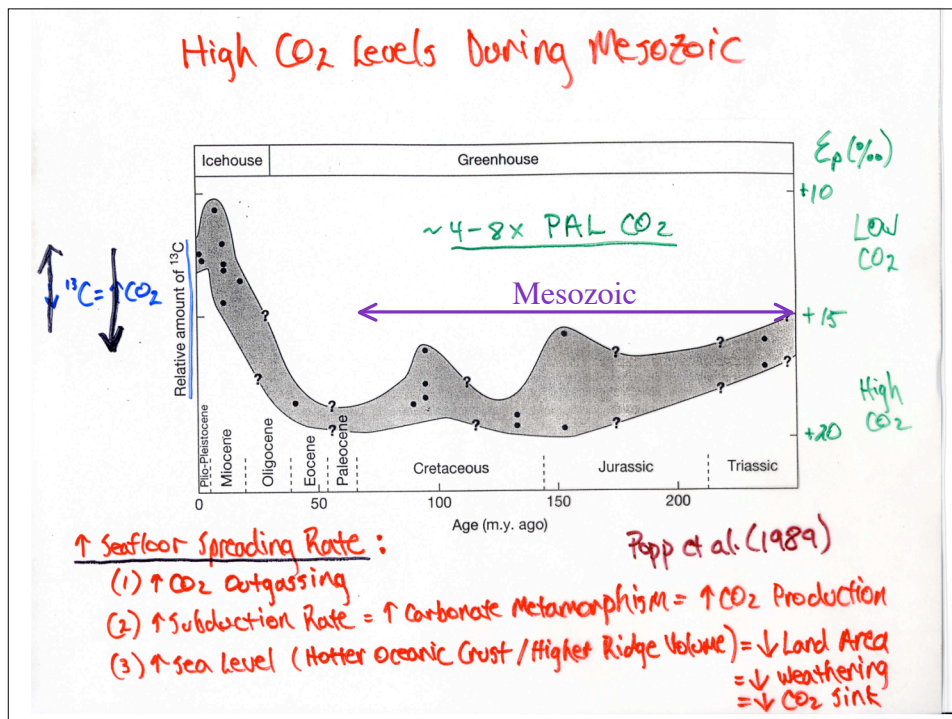
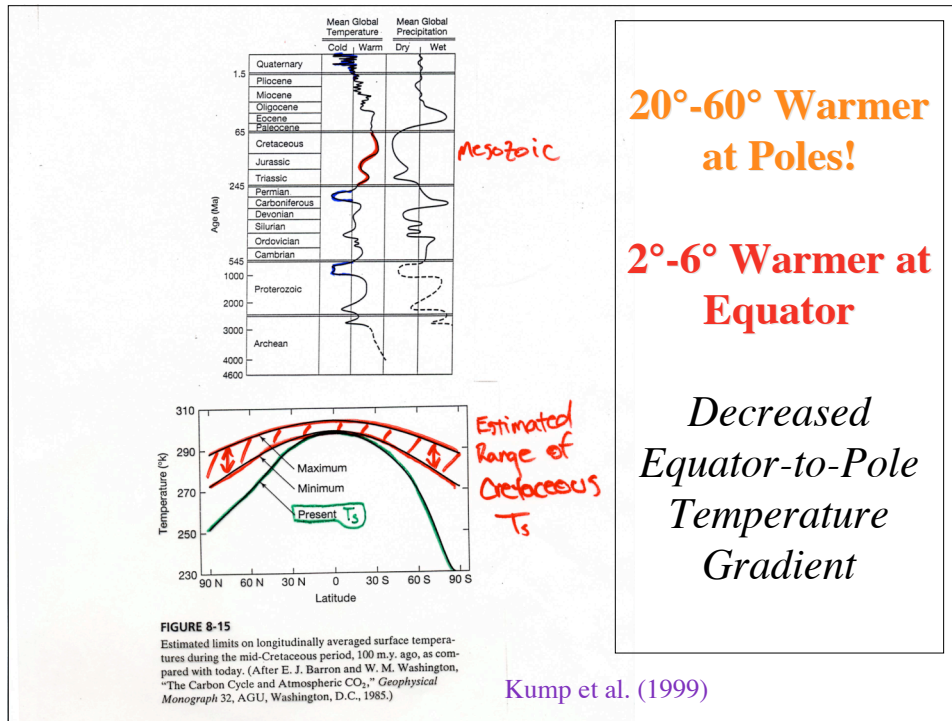
**Mesozoic Warmth**

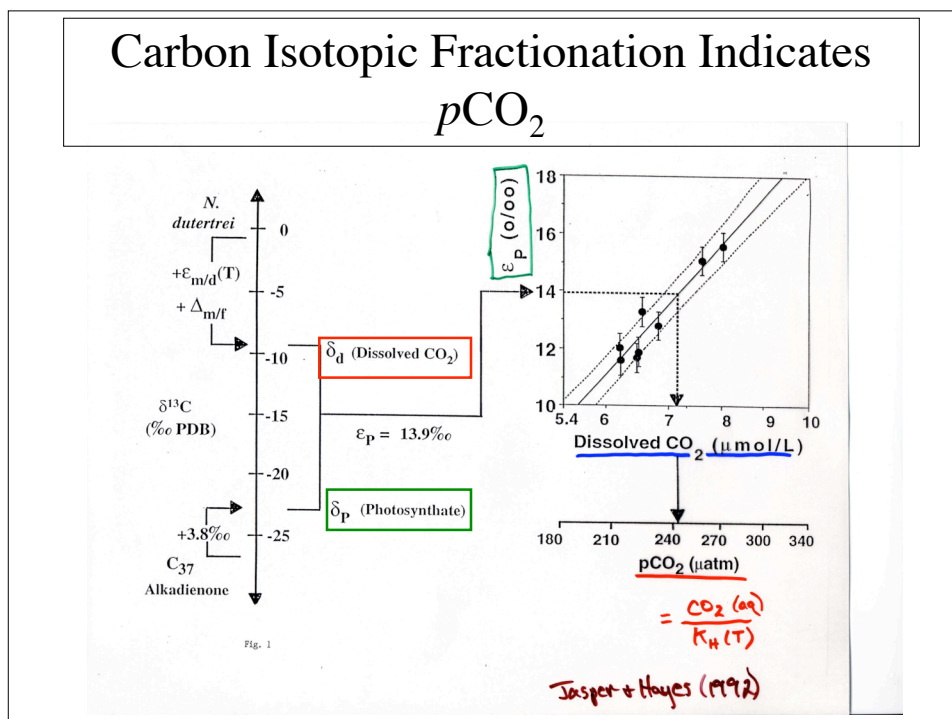
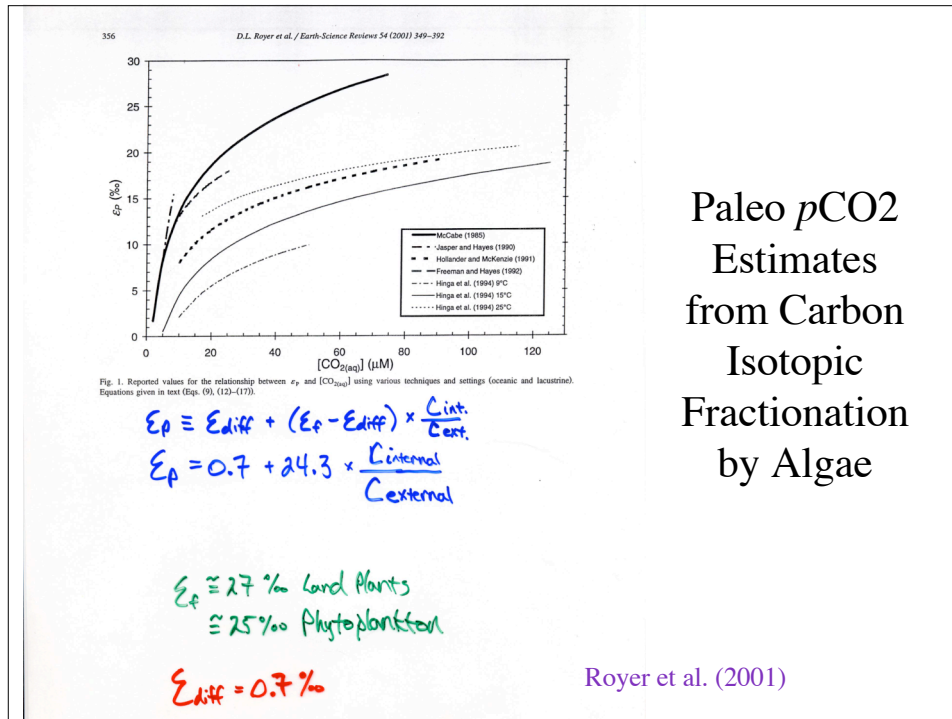
**Jurassic 220-140 Ma**

**Cretaceous 140-65 Ma**

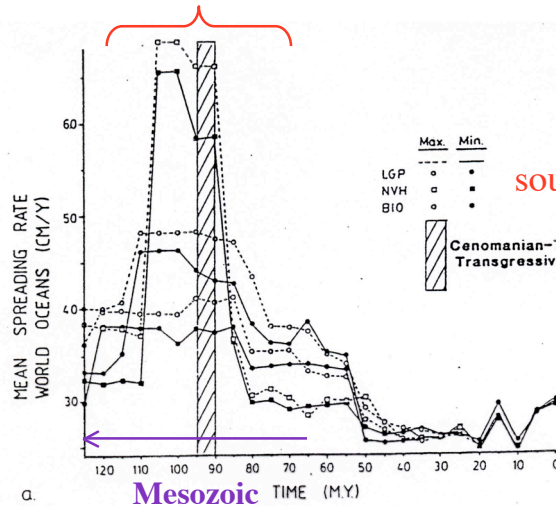
Stanley (2000)

- Ferns & alligators in Siberia
- Dinosaur bones in AK (N of Arctic Circle)





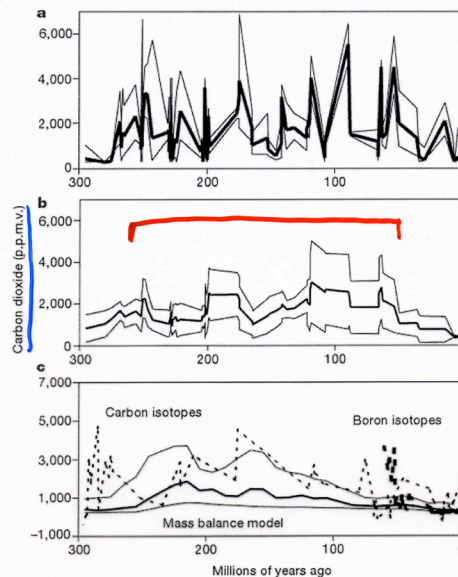
## Seafloor Spreading Rates Appear to have Been High in Latter Half of Mesozoic



- High seafloor spreading rates\* provide potential source of elevated CO<sub>2</sub>

\* Determined by thickness of seafloor magnetic stripes between dated magnetic reversals

Fossil Leaf Cuticles



Fossil leaf cuticles provide evidence for elevated CO<sub>2</sub> during Mesozoic

3-6 x PAL during Mesozoic

$$SI(\%) = \frac{SD}{(SD+ED)} \times 100\%$$

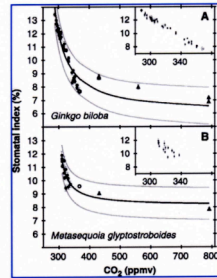
SD= stomatal density  
ED=epidermal cell density

(I.e., the proportion of epidermal cells that are stomata)

Retallack (2001), *Nature*, Vol. 411, pp. 287-290

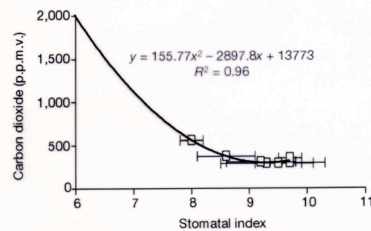


## CO<sub>2</sub> from Leaf Stomatal Indices



Royer et al (2001), *Science*, Vol. 292, pp. 2310-2313

## Calibrating the Leaf Stomatal “Paleo-barometer”



Retallack (2001), *Nature*, Vol. 411, pp. 287-290

*Extrapolation to high pCO<sub>2</sub> not (yet) supported by data...*

## Response of stomata to [CO<sub>2</sub>] is species-dependent

*Limiting SI-derived paleo-CO<sub>2</sub> estimates to times and places when fossilized leaves from extant species exist...*

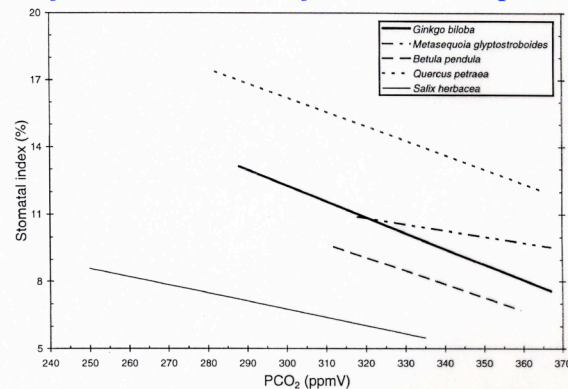


Fig. 9. Stomatal index response of five species to PCO<sub>2</sub>. Data compiled from herbarium sheets and altitudinal transects. Sources are as follows: *Ginkgo biloba* and *Metasequoia glyptostroboides* (Royer, unpublished data); *Betula pendula* (Wagner et al., 1996); *Quercus petraea* (Kürschner et al., 1996); *Salix herbacea* (Rundgren and Beerling, 1999).

Royer et al. (2001)

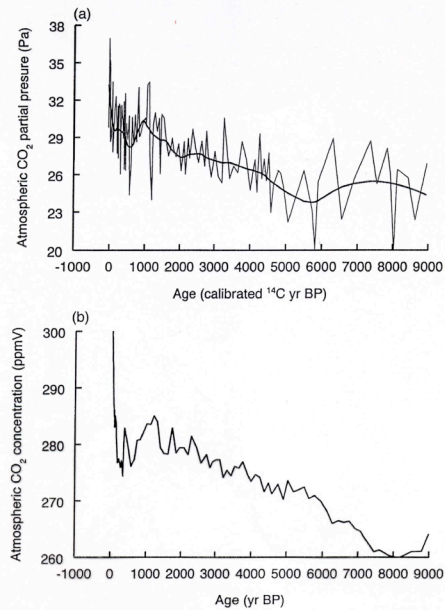
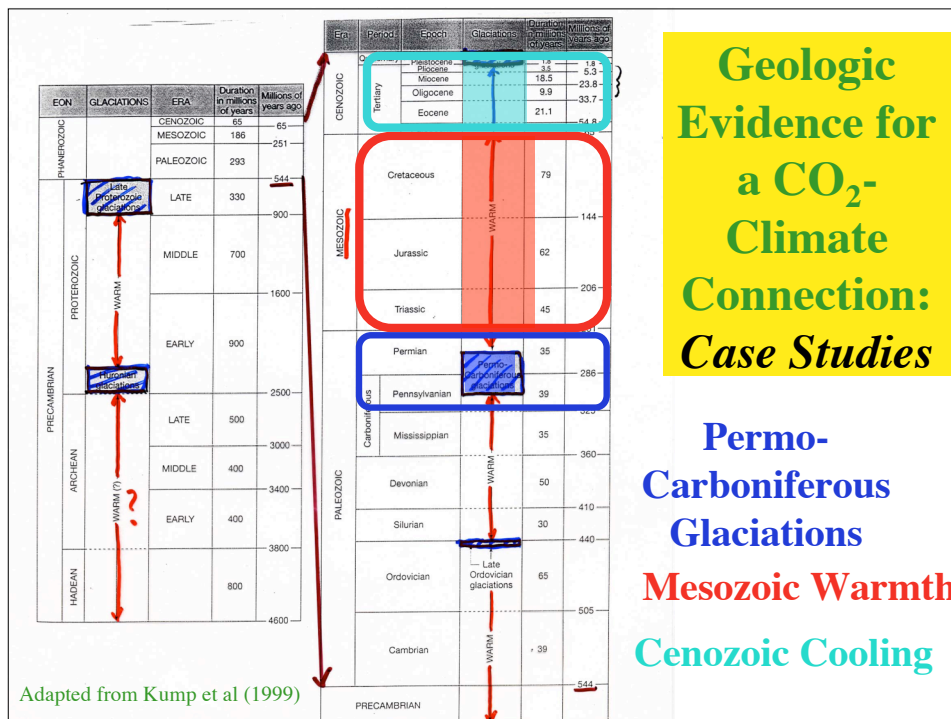


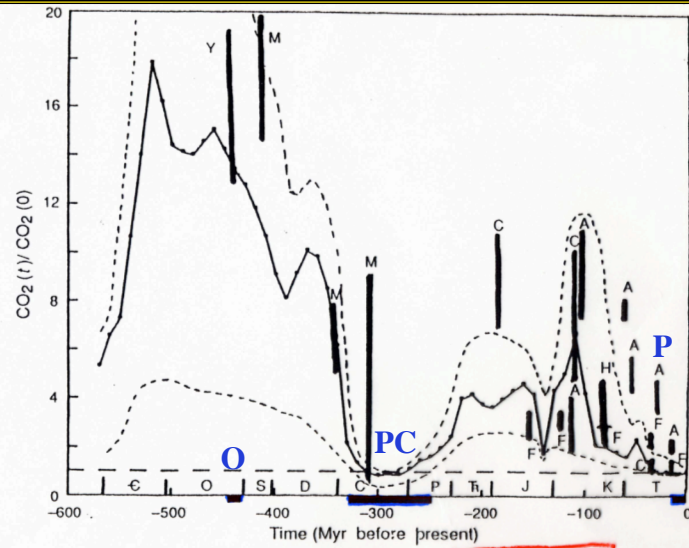
Fig. 12. Holocene reconstructed variations in (a) the partial pressure of atmospheric  $\text{CO}_2$  using fossil *Salix herbacea* leaves (after Rundgren and Beerling, 1999) and (b) measurements of atmospheric  $\text{CO}_2$  from ice cores (after Indermühle et al., 1999b).

Royer et al. (2001)

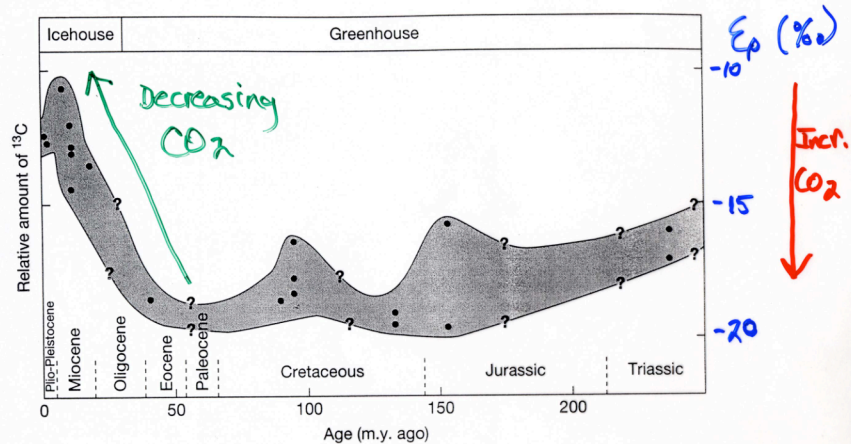
**Nevertheless,  
calibrations of  
the SI appear  
accurate for at  
least the last 9  
kyr**



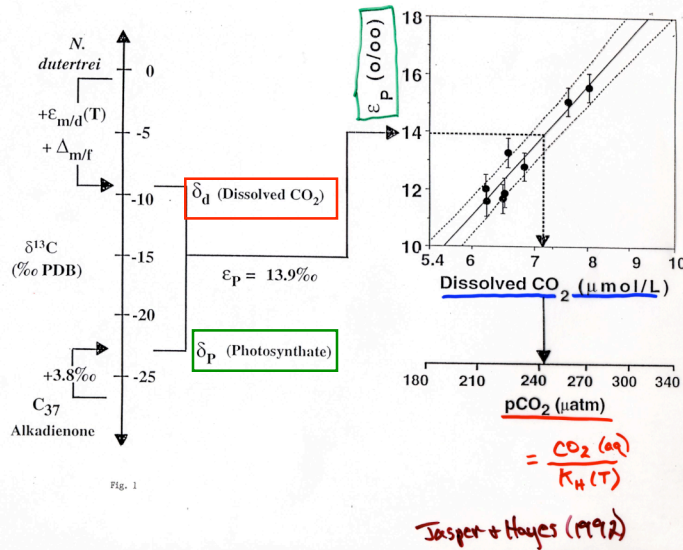
## Phanerozoic CO<sub>2</sub> and Climate



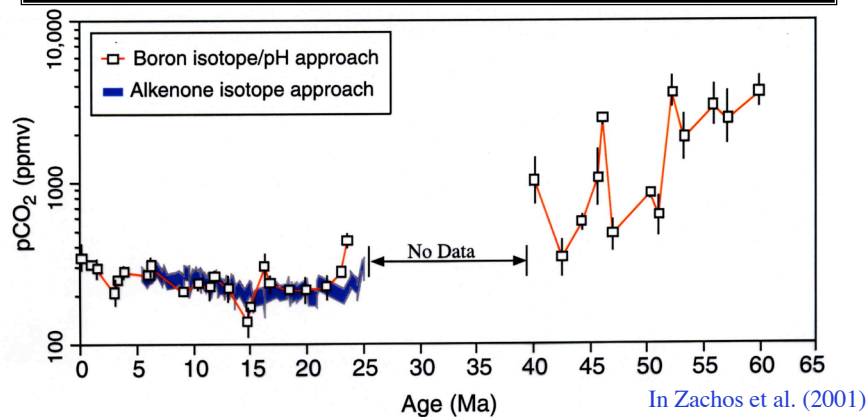
## Cenozoic CO<sub>2</sub> Decrease



## Carbon Isotopic Fractionation Indicates $p\text{CO}_2$

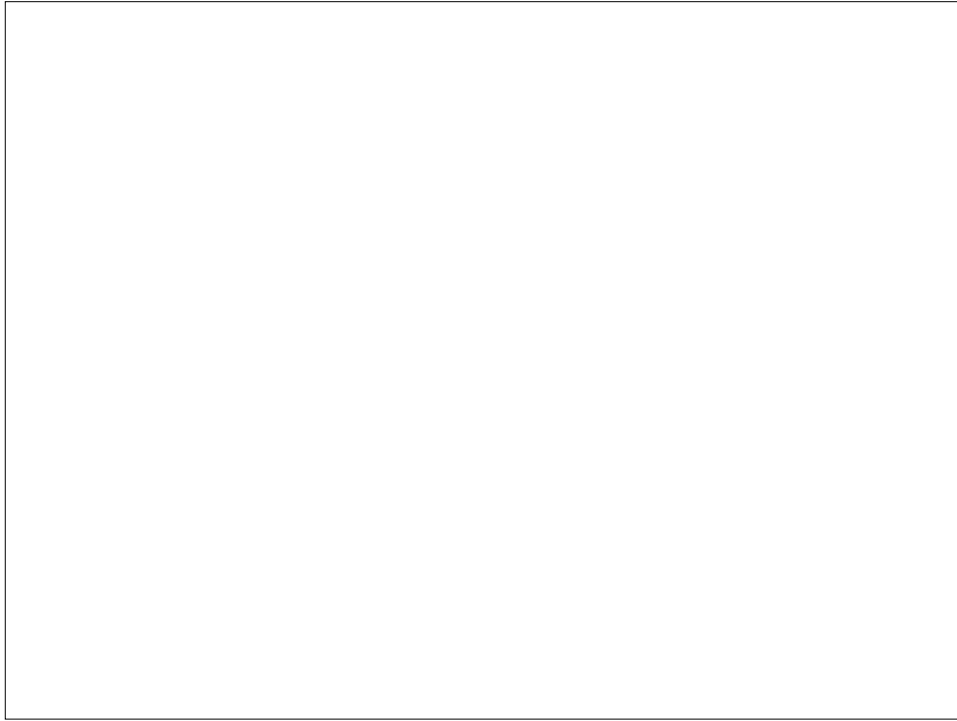


## Boron Isotopes in Seawater Also Indicate Large Cenozoic $\text{CO}_2$ Decline



- B in seawater:  $\text{B}(\text{OH})_3$ ,  $\text{B}(\text{OH})_4^-$
  - Relative abundance controlled by pH
  - B incorporated into calcite:  $\text{B}(\text{OH})_4^-$
  - Strong isotopic fractionation between  $^{10}\text{B}$  &  $^{11}\text{B}$ :  
 $^{10}\text{B}$  = tetrahedral coordination, -19.8‰ relative to  $^{11}\text{B}$
- $$\delta^{11}\text{B} = \left[ \frac{(^{11}\text{B}/^{10}\text{B})_{\text{sample}}}{(^{11}\text{B}/^{10}\text{B})_{\text{std}}} - 1 \right] \times 1000\text{‰}$$

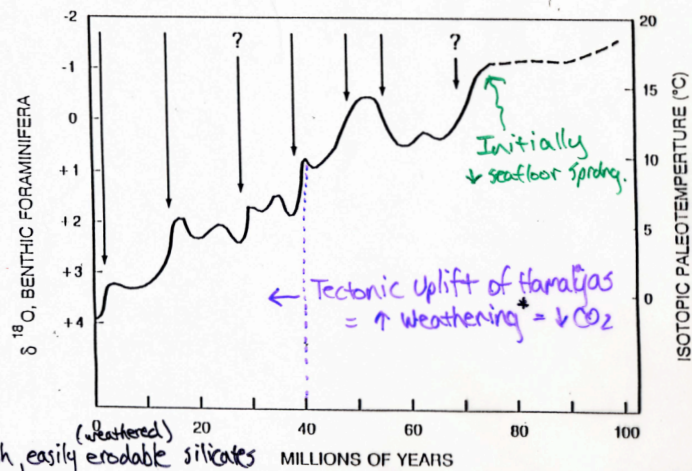




\*\*\* Ended Here 11-21-08; Quiz  
#3 covers through here \*\*\*

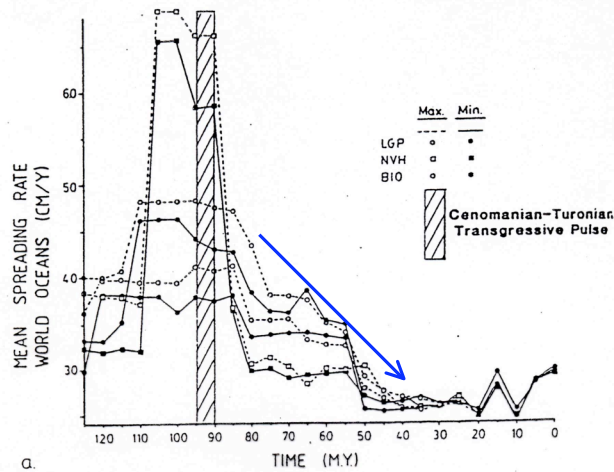


## Cenozoic Cooling 80-0 Ma

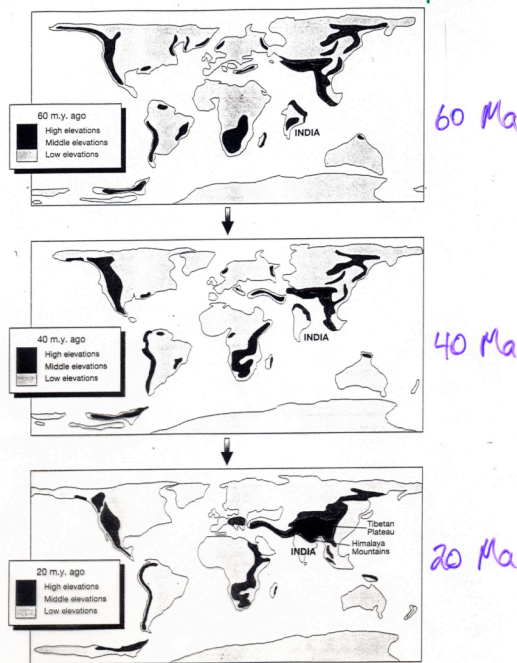


- \* Fresh, easily erodable silicates
- \* Orographic concentration of rain on mts. After Roymo et al. (1988, 1991, 1992)
- \* Steep slopes to rapidly flush weathering products

## Declining Seafloor Spreading Rates 80-40 Ma

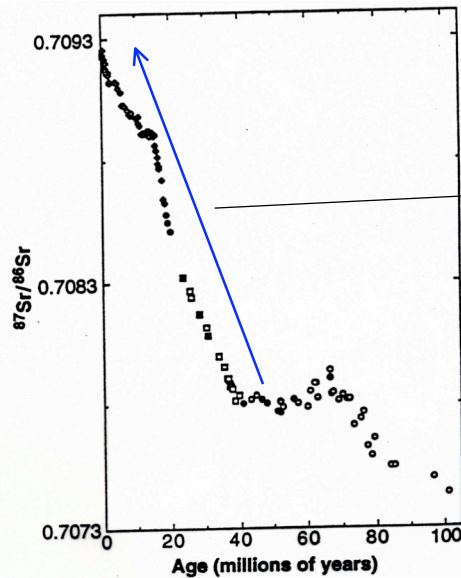


•Declining s.f. spreading rates are consistent with decreasing CO<sub>2</sub> in early Cenozoic



Himalayan Orogeny & Uplift of Tibetan Plateau

## Increasing Strontium Isotopic Composition of Seawater During Cenozoic Implies Increasing Weathering Rates



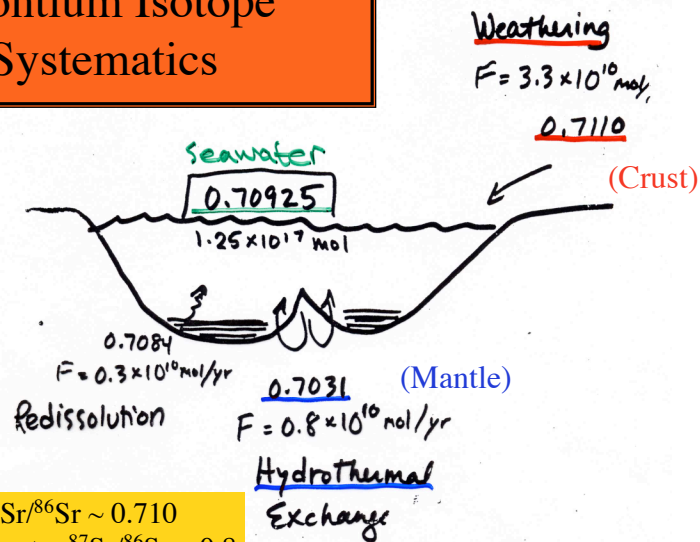
SW  $^{87}\text{Sr}/^{86}\text{Sr}$  is balance between:

1. Deep-sea hydrothermal input of non-radiogenic Sr (0.7035)
2. More radiogenic input riverine flux from continental weathering (0.712)

Abyssal carbonate  $^{87}\text{Sr}/^{86}\text{Sr}$   
 $^{87}\text{Rb} \rightarrow ^{87}\text{Sr}$ ,  $t_{1/2} \sim 48$  Gyr

DePaolo & Ingram (1985)  
 in Edmond (1992)

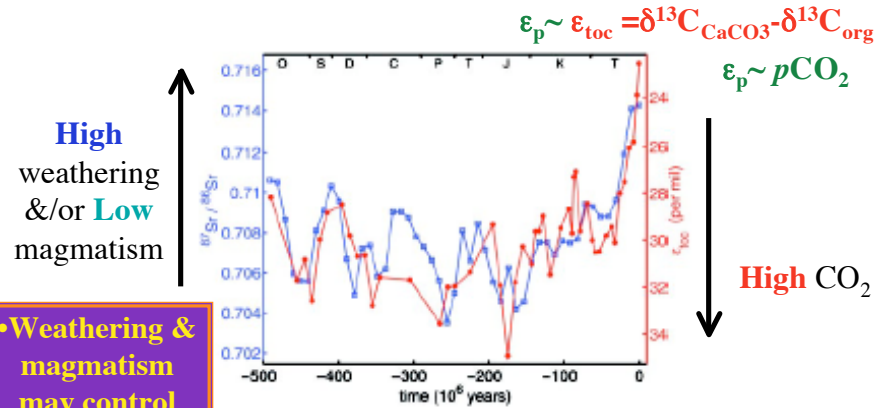
## Strontium Isotope Systematics



Normal River  $^{87}\text{Sr}/^{86}\text{Sr} \sim 0.710$   
 Ganges-Brahmaputra  $^{87}\text{Sr}/^{86}\text{Sr} \sim 0.8$

Raymo (1998)

## Co-Variation of $^{87}\text{Sr}/^{86}\text{Sr}$ & $\text{CO}_2$ through the Phanerozoic

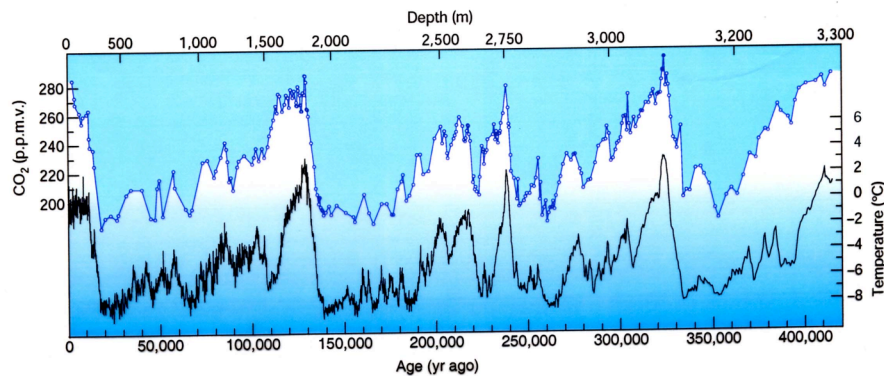


•Weathering & magmatism may control  $\text{CO}_2$ , but does  $\text{CO}_2$  control climate?

Fig. 3. The function  $g$  (blue squares) obtained by removing the memory flux from the  $^{87}\text{Sr}/^{86}\text{Sr}$  data of Fig. 1, along with  $\epsilon_{\text{toc}}$  (red circles). Note that the range of the  $^{87}\text{Sr}/^{86}\text{Sr}$  curve is approximately five times greater than in Fig. 1. The data are plotted such that the mean of both time series lies on the same horizontal line and their rms fluctuations have the same vertical extent.

Rothman (2002) *PNAS*, Vol 99(7):4167

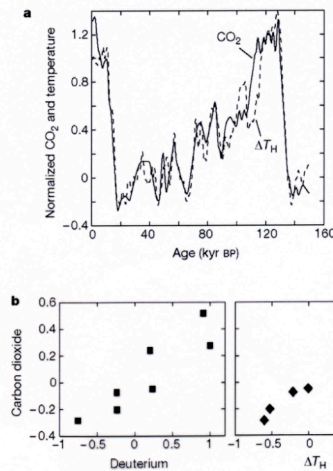
## $\text{CO}_2$ During the last 450 kyr from the Vostok, Antarctica Ice Core



Petit et al (1999) in Kump (2002), *Nature*, vol. 419:188-190.

## Covariation Between CO<sub>2</sub> and T in Pleistocene

Vostok Ice Core, Central East Antarctica



Cuffey and Vimeux (2001), *Nature*, Vol. 412, pp. 523-527

**What caused glacial-interglacial  
CO<sub>2</sub> variations?  
(a still-unanswered question!)**

### Possible Scenario for lower glacial CO<sub>2</sub>

- Increased:  
Equator-Pole T gradient, Wind strength, Dust flux to ocean, Iron flux to ocean
- 50% of global 1° production occurs in ocean
- Ocean 1° production is limited by iron
- Higher 1° production draws CO<sub>2</sub> out of atmosphere & sequesters it in the deep ocean & sediments
- Colder seawater dissolves more CO<sub>2</sub>

## letters to nature

1. J. A. C. Cook, A. J. C. Cook, & J. A. C. Cook. The significance of the relationship between atmospheric CO<sub>2</sub> and global temperature. *J. Chem. Phys.* 46, 194–196 (1967).
2. M. A. S. Coelho, L. J. Lima, & A. S. Coelho. A. S. Coelho. The relationship between atmospheric CO<sub>2</sub> and global temperature. *J. Chem. Phys.* 46, 194–196 (1967).
3. M. A. S. Coelho, L. J. Lima, & A. S. Coelho. A. S. Coelho. The relationship between atmospheric CO<sub>2</sub> and global temperature. *J. Chem. Phys.* 46, 194–196 (1967).
4. M. A. S. Coelho, L. J. Lima, & A. S. Coelho. A. S. Coelho. The relationship between atmospheric CO<sub>2</sub> and global temperature. *J. Chem. Phys.* 46, 194–196 (1967).
5. M. A. S. Coelho, L. J. Lima, & A. S. Coelho. A. S. Coelho. The relationship between atmospheric CO<sub>2</sub> and global temperature. *J. Chem. Phys.* 46, 194–196 (1967).
6. M. A. S. Coelho, L. J. Lima, & A. S. Coelho. A. S. Coelho. The relationship between atmospheric CO<sub>2</sub> and global temperature. *J. Chem. Phys.* 46, 194–196 (1967).
7. M. A. S. Coelho, L. J. Lima, & A. S. Coelho. A. S. Coelho. The relationship between atmospheric CO<sub>2</sub> and global temperature. *J. Chem. Phys.* 46, 194–196 (1967).
8. M. A. S. Coelho, L. J. Lima, & A. S. Coelho. A. S. Coelho. The relationship between atmospheric CO<sub>2</sub> and global temperature. *J. Chem. Phys.* 46, 194–196 (1967).
9. M. A. S. Coelho, L. J. Lima, & A. S. Coelho. A. S. Coelho. The relationship between atmospheric CO<sub>2</sub> and global temperature. *J. Chem. Phys.* 46, 194–196 (1967).
10. M. A. S. Coelho, L. J. Lima, & A. S. Coelho. A. S. Coelho. The relationship between atmospheric CO<sub>2</sub> and global temperature. *J. Chem. Phys.* 46, 194–196 (1967).
11. M. A. S. Coelho, L. J. Lima, & A. S. Coelho. A. S. Coelho. The relationship between atmospheric CO<sub>2</sub> and global temperature. *J. Chem. Phys.* 46, 194–196 (1967).
12. M. A. S. Coelho, L. J. Lima, & A. S. Coelho. A. S. Coelho. The relationship between atmospheric CO<sub>2</sub> and global temperature. *J. Chem. Phys.* 46, 194–196 (1967).
13. M. A. S. Coelho, L. J. Lima, & A. S. Coelho. A. S. Coelho. The relationship between atmospheric CO<sub>2</sub> and global temperature. *J. Chem. Phys.* 46, 194–196 (1967).
14. M. A. S. Coelho, L. J. Lima, & A. S. Coelho. A. S. Coelho. The relationship between atmospheric CO<sub>2</sub> and global temperature. *J. Chem. Phys.* 46, 194–196 (1967).
15. M. A. S. Coelho, L. J. Lima, & A. S. Coelho. A. S. Coelho. The relationship between atmospheric CO<sub>2</sub> and global temperature. *J. Chem. Phys.* 46, 194–196 (1967).
16. M. A. S. Coelho, L. J. Lima, & A. S. Coelho. A. S. Coelho. The relationship between atmospheric CO<sub>2</sub> and global temperature. *J. Chem. Phys.* 46, 194–196 (1967).
17. M. A. S. Coelho, L. J. Lima, & A. S. Coelho. A. S. Coelho. The relationship between atmospheric CO<sub>2</sub> and global temperature. *J. Chem. Phys.* 46, 194–196 (1967).
18. M. A. S. Coelho, L. J. Lima, & A. S. Coelho. A. S. Coelho. The relationship between atmospheric CO<sub>2</sub> and global temperature. *J. Chem. Phys.* 46, 194–196 (1967).
19. M. A. S. Coelho, L. J. Lima, & A. S. Coelho. A. S. Coelho. The relationship between atmospheric CO<sub>2</sub> and global temperature. *J. Chem. Phys.* 46, 194–196 (1967).
20. M. A. S. Coelho, L. J. Lima, & A. S. Coelho. A. S. Coelho. The relationship between atmospheric CO<sub>2</sub> and global temperature. *J. Chem. Phys.* 46, 194–196 (1967).
21. M. A. S. Coelho, L. J. Lima, & A. S. Coelho. A. S. Coelho. The relationship between atmospheric CO<sub>2</sub> and global temperature. *J. Chem. Phys.* 46, 194–196 (1967).
22. M. A. S. Coelho, L. J. Lima, & A. S. Coelho. A. S. Coelho. The relationship between atmospheric CO<sub>2</sub> and global temperature. *J. Chem. Phys.* 46, 194–196 (1967).
23. M. A. S. Coelho, L. J. Lima, & A. S. Coelho. A. S. Coelho. The relationship between atmospheric CO<sub>2</sub> and global temperature. *J. Chem. Phys.* 46, 194–196 (1967).

## acknowledgments

This research was supported by the NSF Division of Earth and Planetary Sciences, and the Petroleum Research Fund administered by the American Chemical Society. We thank M. D. Lipin for helpful discussions and E. S. Sarnoff for assistance.

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## Evidence for decoupling of atmospheric CO<sub>2</sub> and global climate during the Phanerozoic eon

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Atmospheric carbon dioxide concentrations are believed to drive climate changes from glacial to interglacial modes, although geological and astronomical oscillations have been invoked as ultimate causes. Additionally, it is unclear whether the changes between cold and warm modes should be regarded as a global phenomenon, affecting tropical and high-latitude temperatures alike<sup>1–3</sup>, or if they are better described as an expansion and

contraction of the latitudinal climate zones, keeping equatorial temperatures approximately constant<sup>4–6</sup>. Here we present a reconstruction of tropical sea surface temperatures throughout the Phanerozoic eon (the past ~550 Myr) from our database<sup>7</sup> of oxygen isotopes in calcite and argonite shells. The data indicate large oscillations of tropical sea surface temperatures in phase with the cold-warm cycles, thus favouring the idea of climate variability as a global phenomenon. But our data conflict with a temperature reconstruction using an energy balance model that is forced by reconstructed atmospheric carbon dioxide concentrations<sup>8</sup>. The results can be reconciled if atmospheric carbon dioxide concentrations were not the principal driver of climate variability on geological timescales for at least one-third of the Phanerozoic eon, or if the reconstructed carbon dioxide concentrations are not reliable.

Calculated partial pressures of CO<sub>2</sub> in the atmosphere<sup>9,10</sup> (pCO<sub>2</sub>) are indeed relatively low for the Permian/Carboniferous and Cretaceous icehouse episodes, times of predominantly cold climates lasting tens of millions of years<sup>11</sup>, but for the other two Phanerozoic icehouses—the late Ordovician/Carboniferous and the late Jurassic/Cretaceous—all global biogeochemical models<sup>12–15</sup> predict high pCO<sub>2</sub> levels. In an attempt to resolve this discrepancy, particularly for the late Ordovician glaciation(s), theoretical models advocating the development of permanent high-latitude ice sheets<sup>16</sup> at more than 10 times present-day CO<sub>2</sub> levels have been proposed. As an alternative, we present here experimental evidence that suggests large variations in tropical sea surface temperatures (SSTs) up to 9°C between the peaks of greenhouse/icehouse modes. The testing of model predictions on Phanerozoic timescales has been hampered by the lack of high-resolution isotope databases for ancient sea water, as reflected in (bio)chemical sediments. This has been particularly the case for oxygen isotopes, where the depletion in <sup>18</sup>O for progressively older sediments was mostly considered to be a product of post-depositional diagenetic overprint. However, new databases have been published<sup>17</sup> recently—based on ~500 low-Mg calcite and 125 argonite shells—for δ<sup>18</sup>O, δ<sup>13</sup>C and <sup>87</sup>Sr/<sup>86</sup>Sr, establishing the baseline for Phanerozoic δ<sup>18</sup>O (Fig. 1) in ref. 17. Most studied samples originate from the photic zone of palaeotropical seas (30°S–30°N). The entire datasets are available at [http://www.sediment.com/geology/geology\\_data\\_reference\\_17](http://www.sediment.com/geology/geology_data_reference_17), because of the large number of measurements, dealt only with the documentation of data and their quality, and with the arguments for the primary nature of the δ<sup>18</sup>O record.

The secular trend for the δ<sup>18</sup>O values of Phanerozoic carbonates, Fig. 1 in ref. 17, consists of a long-term rising linear trend that exceeds the duration of planetary greenhouse/icehouse modes. This trend is tectonically controlled, a proposition supported by the strong correlation of δ<sup>18</sup>O with the <sup>87</sup>Sr/<sup>86</sup>Sr signal for the Phanerozoic sea water (Figs 16 and 17 in ref. 17). δ<sup>18</sup>O isotope is a proxy for terrestrial tectonic processes, because seawater <sup>87</sup>Sr/<sup>86</sup>Sr is controlled principally by inputs of it from rivers (resulting of <sup>87</sup>Se-enriched continental crust) and from hydrothermal circulation cells at mid-ocean ridges (interaction with <sup>87</sup>Se-depleted basalt). Such long-term tectonic signals have to be subtracted from the superimposed higher-order oscillations before any discussion of climatic consequences, and we therefore derived the data by removing the least-squares linear fit.

The first-order δ<sup>18</sup>O oscillations around the least-squares fit correlate well with the palaeoclimate as established from sedimentological criteria<sup>18</sup>. The icehouses usually coincide with <sup>18</sup>O-enriched values and greenhouses with <sup>18</sup>O-depleted values (Fig. 1). We propose therefore that the Phanerozoic δ<sup>18</sup>O oscillations reflect variations in SSTs. Future improvements on the Phanerozoic database may result in amelioration of the amplitudes, or in partial temporal shifts of some of the peaks. This is particularly the case for the Neoproterozoic interval, where the rising mean is forced to higher values by the inclusion of some isotopically heavy planktonic

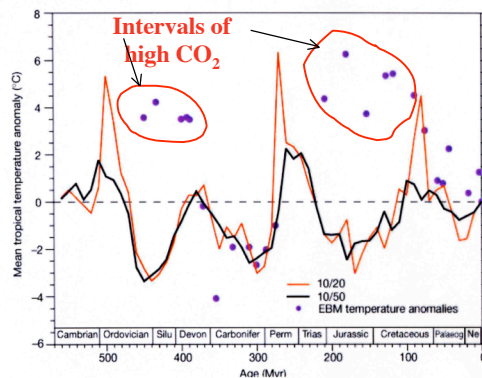
While a large and growing body of evidence indicates that CO<sub>2</sub> and climate co-vary, there is some indication that the two may not be closely linked at all times....

(& it is always important to remember that correlation does not always mean causation)

Tuesday, September 4, 2001

Veizer\_Nature\_13

## Cool Tropics - High CO<sub>2</sub> (?)



Veizer et al. (2000), *Nature*, Vol. 408, pp. 698-701

(or diagenetic alteration of CaCO<sub>3</sub>?)

## Model-Data SST Comparison

### Tropical SST anomaly (Data)

- Assumes 2‰ of 3-5‰ δ<sup>18</sup>O range due to ice volume (2x present ice volume in “icehouse”; No ice in “greenhouse”).
- Leaves ~2‰, or ~9°C of SST change

### Simple E Balance Model

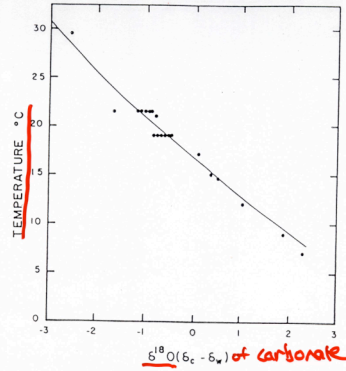
- CO<sub>2</sub> (Berner, 1992)
- Solar constant increasing by 5% over Phanerozoic

$$T_s - \Delta T_g = T_{eff}$$

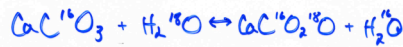
$$\sigma T_{eff}^4 = S/4 * (1 - A)$$



# T-dependent Isotopic Fractionation



Natural Abundances  
 $^{16}\text{O} = 99.756\%$   
 $^{17}\text{O} = 0.039$   
 $^{18}\text{O} = 0.205$



$$T = 16.9 - 4.2 (\delta^{18}\text{O}_c - \delta^{18}\text{O}_w)$$

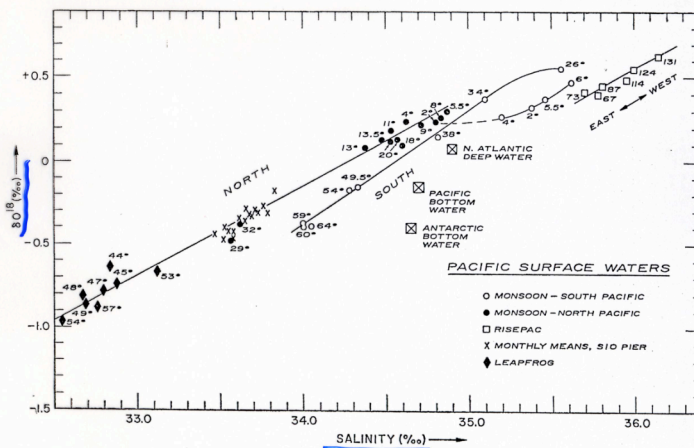
$$\delta^{18}\text{O}_w = f(\text{Salinity, ice volume})$$

$$\delta^{18}\text{O} = \left[ \frac{(^{18}\text{O}/^{16}\text{O})_{\text{sample}}}{(^{18}\text{O}/^{16}\text{O})_{\text{standard}}} - 1 \right] \times 1000\%$$

Epstein et al. (1953)

A  
 Temperature-  
 Dependent  
 Isotopic  
 Fractionation  
 of  $^{18}\text{O}$  from  $^{16}\text{O}$   
 Occurs During  
 Calcification

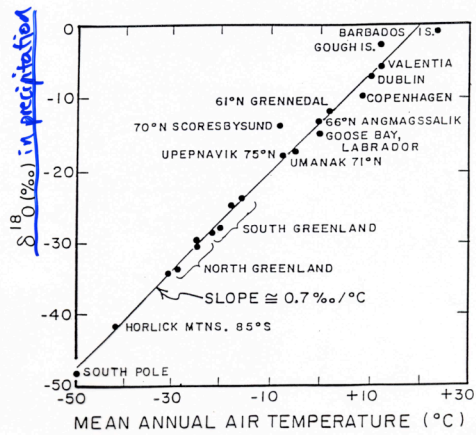
## $\delta^{18}\text{O}$ of Seawater varies w/ Salinity



H. Craig and L. I. Gordon (1965)

FIGURE 10. — Oxygen 18-salinity relationships in surface and deep Pacific Ocean samples. Latitudes are shown for Monsoon and Leapfrog samples; the figures on the RISEPAC expedition points are station numbers shown in figure 6.





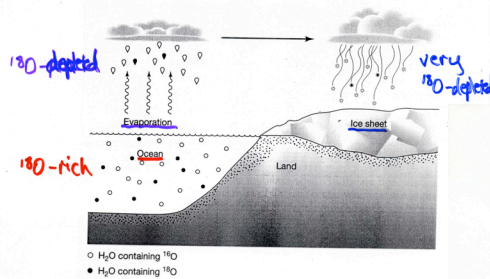
Dansgaard (1964)

•  $\delta^{18}\text{O}$  of precipitation varies with latitude & altitude.

• Water derived from melting snow & ice is highly depleted in  $^{18}\text{O}$ .

• Airmass trajectory influences  $\delta^{18}\text{O}$  of precipitation

$\text{H}_2^{18}\text{O}$  has 1% lower vapor P  
 $\therefore$  Cold Air (High latitude +/or Altitude)  
 has low  $\delta^{18}\text{O}$



Influence of Ice Volume on Ocean  $\delta^{18}\text{O}$

1.3 ‰ Decrease of Ocean  $\delta^{18}\text{O}$  from Melting Glaciers

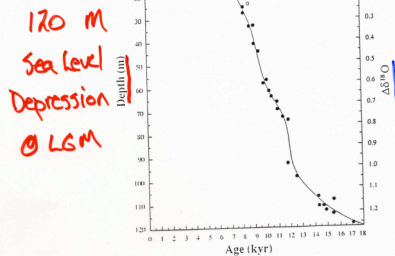
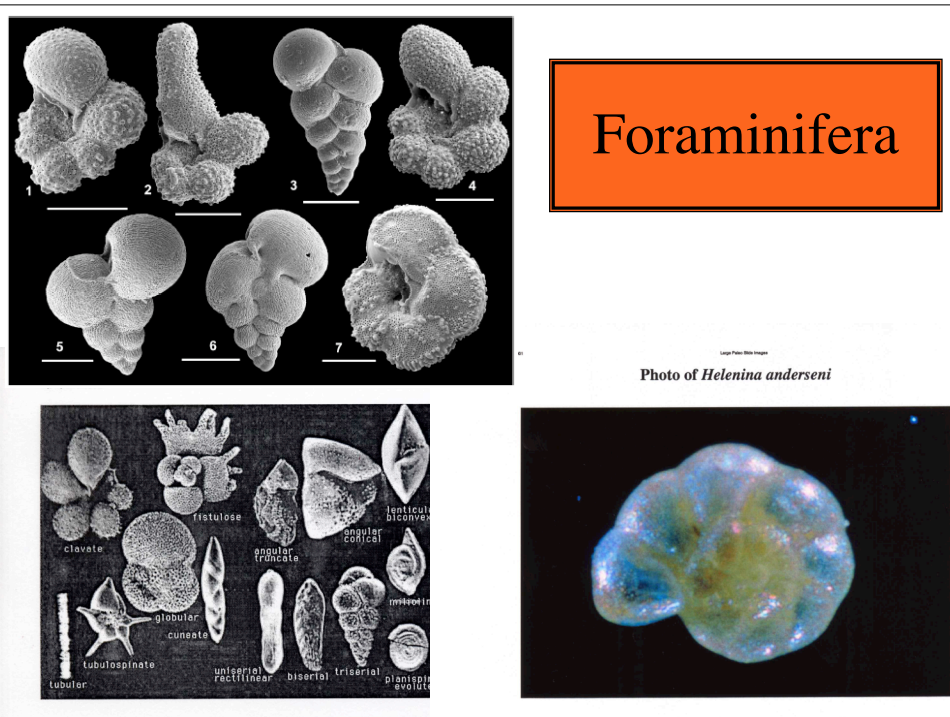


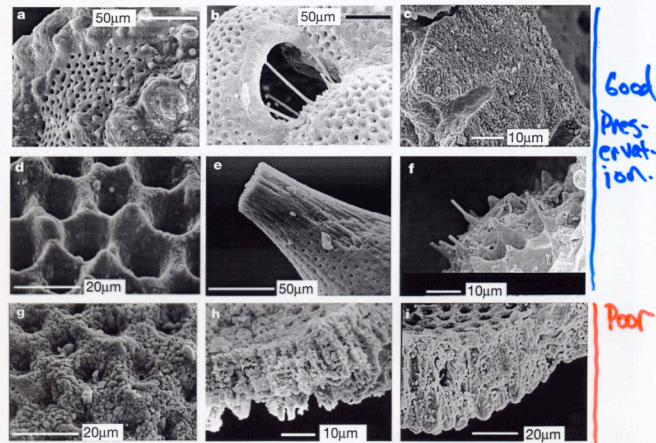
FIG. 2 Barbados sea level curve based on radiocarbon-dated *A. palmata* (filled circles) compared with *A. palmata* age-depth data (open circles) for four other Caribbean island locations. All radiocarbon ages in this figure are corrected for local seawater  $\Delta^{14}\text{C}$  by subtracting 400 yr from the measured radiocarbon ages<sup>11</sup> but they are not corrected for secular changes in atmospheric  $^{14}\text{C}$  levels<sup>12</sup>. The Barbados data are corrected for the estimated mean uplift of 34 cm kyr<sup>-1</sup>. The right-hand axis of the Barbados sea level curve (solid line) is scaled to the estimated  $\delta^{18}\text{O}$  change of mean ocean water using the calibration in ref. 5.

Fairbanks (1989), Nature, 342: 637-642

# Diagenetic Alteration of $^{18}\text{O}/^{16}\text{O}$ in $\text{CaCO}_3$



## Recrystallization of Foramin Calcite

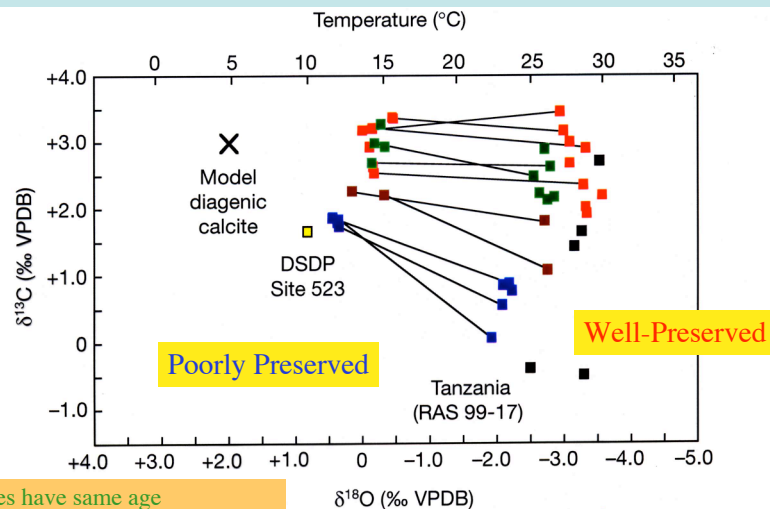


Good  
Pres-  
ervation.

Poor

Pearson et al. (2001),  
Nature, 413: 481-487

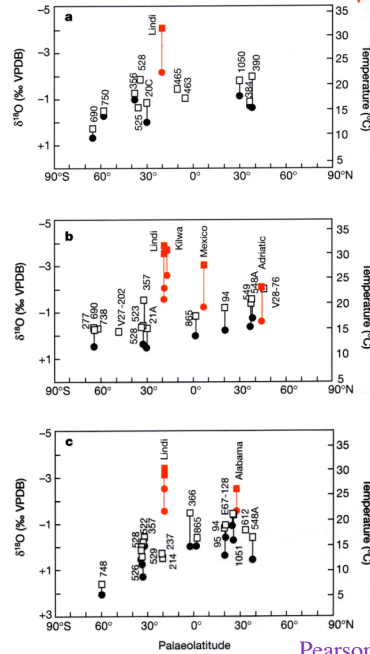
## Large Effect of Diagenesis Makes T Estimates Biased to Cold



- Samples have same age
- Lines connect same species of foraminifera

Pearson et al. (2001) *Nature* Vol. 413: 481-487.

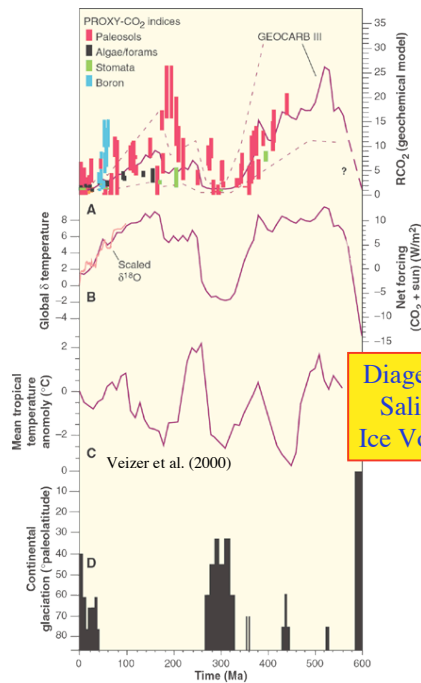
Well-Preserved Forams Indicate ↑ Tropical SST



Well-Preserved Fossils are Essential!

Pearson et al. (2001) *Nature* Vol. 413: 481-487.

## CO<sub>2</sub> & Climate



Diagenesis?  
Salinity?  
Ice Volume?

**Records of change.** (A) Comparison of CO<sub>2</sub> concentrations from the GEOCARB III model (6) with a compilation (9) of proxy-CO<sub>2</sub> evidence (vertical bars). Dashed lines: estimates of uncertainty in the geochemical model values (6). Solid line: conjectured extension to the late Neoproterozoic (about 590 to 600 Ma). RCO<sub>2</sub>, ratio of CO<sub>2</sub> levels with respect to the present (300 parts per million). Other carbon cycle models (21, 22) for the past 150 million years are in general agreement with the results from this model. (B) Radiative forcing for CO<sub>2</sub> calculated from (23) and corrected for changing luminosity (24) after adjusting for an assumed 30% planetary albedo. Deep-sea oxygen isotope data over the past 100 Ma (13, 14) have been scaled to global temperature variations according to (7). (C) Oxygen isotope-based low-latitude paleotemperatures from (5). (D) Glaciological data for continental-scale ice sheets modified from (7, 8) and based on many sources. The duration of the late Neoproterozoic glaciation is a subject of considerable debate.

Crowley & Berner (2002) *Science*, Vol. 292:870.

However, even with well-preserved foraminifera, Cretaceous warmth may have commenced *after* large (inferred) CO<sub>2</sub> emission by seafloor spreading...

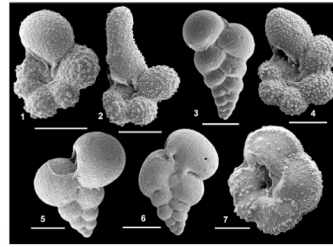
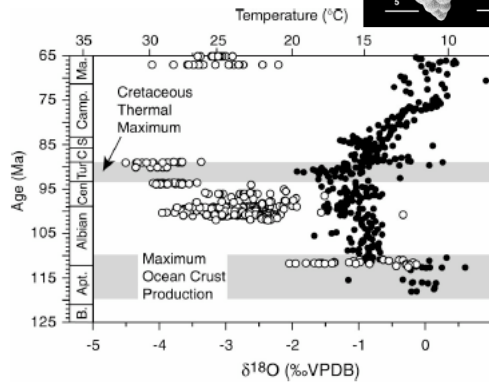


Figure 3. Comparison of Cretaceous  $\delta^{18}\text{O}$ -temperature records indicates 20–40 m.y. mismatch between peak Cretaceous-Cenozoic global warmth (Cretaceous thermal maximum) and peak Cretaceous-Cenozoic tectonic CO<sub>2</sub> production inferred from ocean-crust cycling (Larson, 1991). Glassy planktic foraminifera (open symbols) from low-latitude western tropical Atlantic (plus Maastrichtian data from Tanzania and Gulf Coast) vs. bulk carbonate (solid symbols) from high-latitude Southern Indian Ocean (Clarke and Jenkyns, 1999; Erbacher et al., 2001; Wilson and Norris, 2001; Pearson et al., 2001; Norris et al., 2002). All temperatures calculated in same way as conservative temperatures in Figure 2 and would be 3–6 °C higher if modern latitudinal trends in  $\delta w$  (global mean Cretaceous seawater) were applied (see text).



Wilson et al. *Geology* (2002)

## Other Evidence for Weak CO<sub>2</sub> - Climate Connection during Phanerozoic

$$\epsilon_p \sim p\text{CO}_2$$

$$\epsilon_p \sim \epsilon_{\text{toc}} = \delta^{13}\text{C}_{\text{CaCO}_3} - \delta^{13}\text{C}_{\text{org}}$$

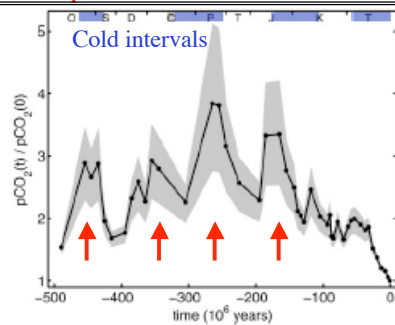
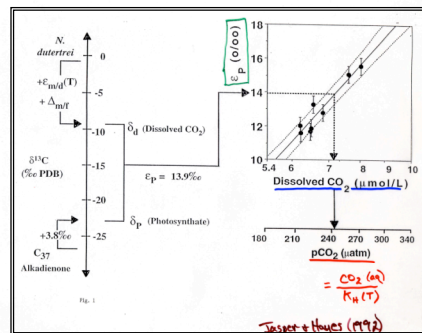
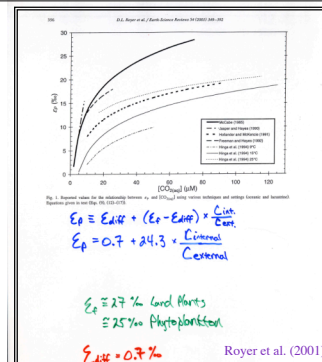


Fig. 4. Fluctuations of  $p\text{CO}_2$  for the last 500 My, normalized by the estimate of  $p\text{CO}_2$  obtained from the most recent value of  $\epsilon$ . The solid line is obtained from Eq. 12 by using  $\epsilon_0 = 36\%$ . The lower and upper limits of the gray area surrounding the  $p\text{CO}_2$  curve result from  $\epsilon_0 = 38$  and  $35\%$ , respectively. The gray bars at the top correspond to periods when Earth's climate was relatively cool; the white spaces between them correspond to warm modes (18).

Rothman (2002) *PNAS*





## But different CO<sub>2</sub> proxies lead to different results....

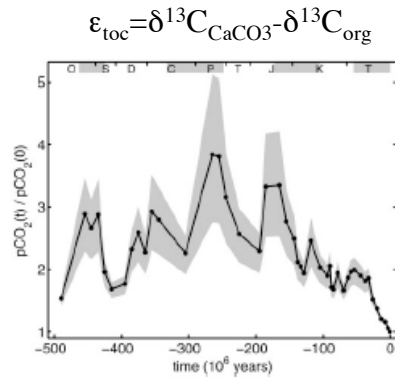


Fig. 4. Fluctuations of  $p\text{CO}_2$  for the last 500 My, normalized by the estimate of  $p\text{CO}_2$  obtained from the most recent value of  $\epsilon$ . The solid line is obtained from Eq. 12 by using  $\epsilon_0 = 36\text{‰}$ . The lower and upper limits of the gray area surrounding the  $p\text{CO}_2$  curve result from  $\epsilon_0 = 38$  and  $35\text{‰}$ , respectively. The gray bars at the top correspond to periods when Earth's climate was relatively cool; the white spaces between them correspond to warm modes (18).

Rothman (2002)

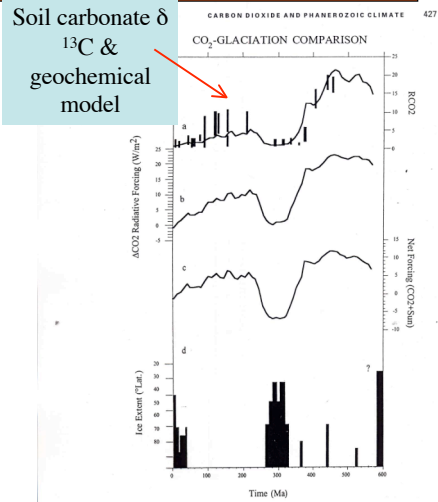
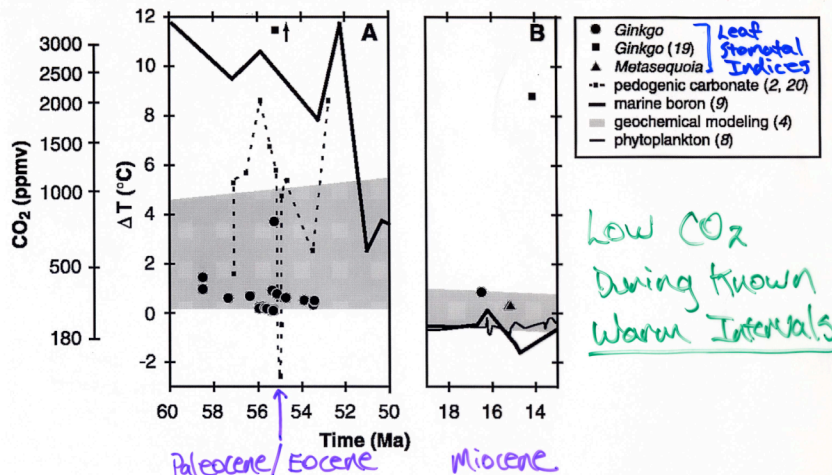


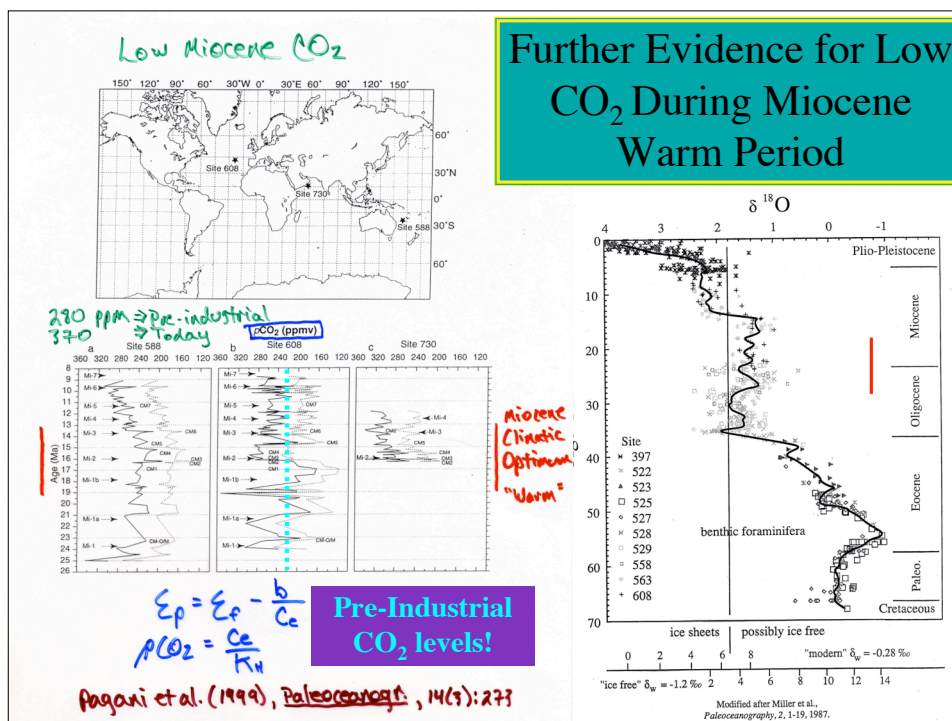
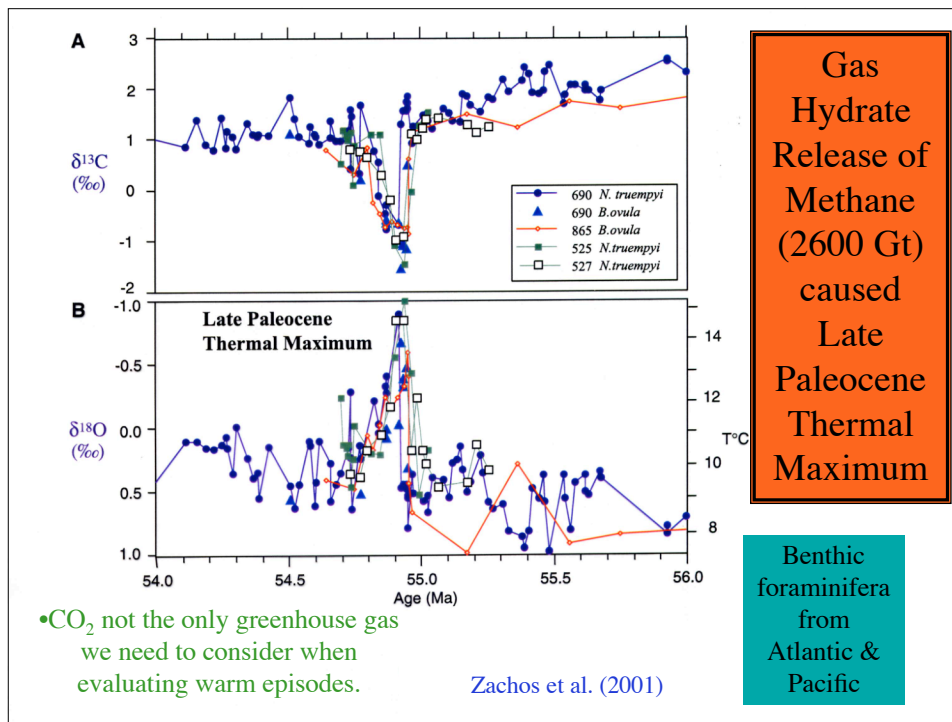
Figure 14.1. (a) Comparison of  $\text{CO}_2$  concentrations from a geochemical model (Berner, 1994) with a compilation (Berner, 1997) of proxy  $\text{CO}_2$  estimates (vertical bars).  $\text{CO}_2$  (b) and solar radiative (c) forcing effects ( $\text{W m}^{-2}$ ) as discussed in the text. Glaciological evidence for continental-scale glaciation (d) from Crowley (1998), which in turn was modified from a compilation from many sources. See text for further discussion.

Crowley (2000)

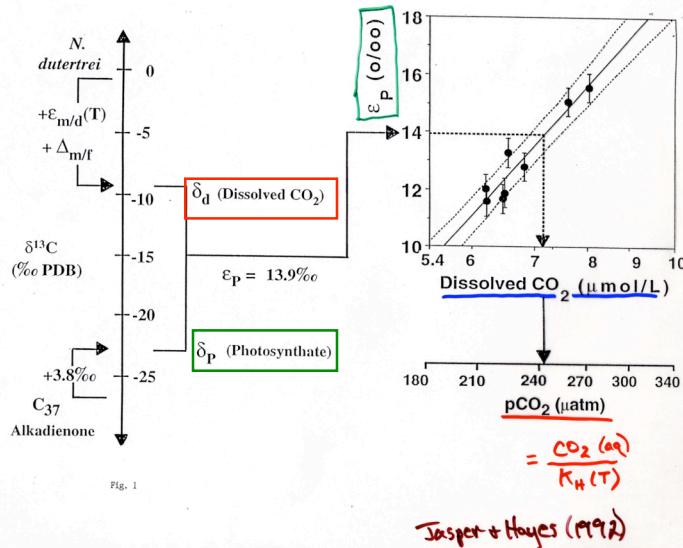
## Other evidence for Low CO<sub>2</sub> During Known Warm Periods



Royer, et al. (2001)  
Science, Vol. 292, p. 2330



## Carbon Isotopic Fractionation Indicates $p\text{CO}_2$

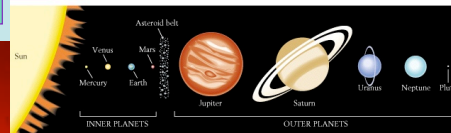


**Substantial evidence exists for a strong link between  $\text{CO}_2$  & climate on a variety of timescales.... With some notable exceptions!**

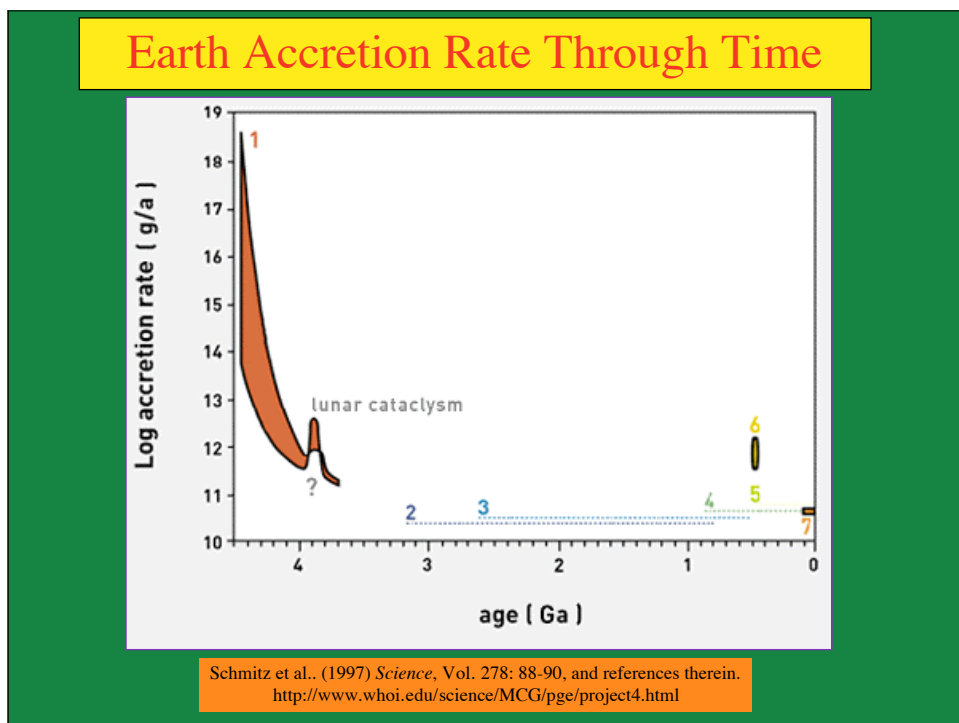
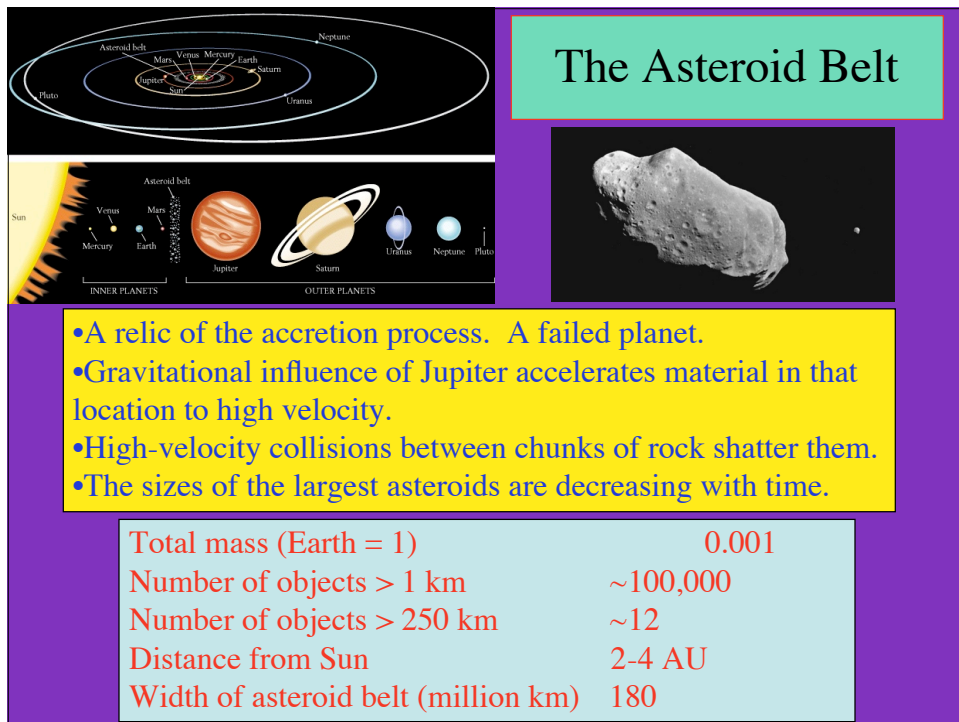
Additional paleoclimate reconstructions & numerical model simulations are necessary. But the biggest (non-controlled) experiment ever attempted is now underway...



## A dramatic illustration of a celestial collision. On the left, a large, dark, cratered sphere, resembling a protoplanet or the Moon, is shown in profile. To its right, a smaller, intensely bright orange and yellow sphere, likely a protostar or young star, is engulfed or about to collide with it. A massive, glowing orange and yellow cloud of dust and gas surrounds the point of impact, radiating outwards. The background is a deep black space filled with numerous small, distant stars.



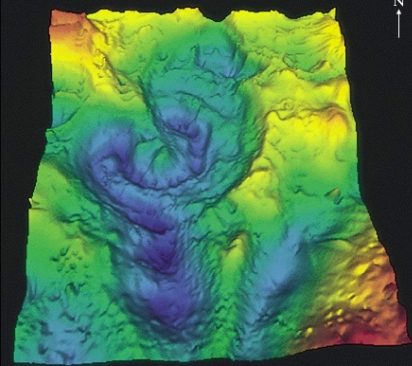
35



## Accretion continues...

### Chicxulub Crater, Gulf of Mexico

- 200 km crater
- 10-km impactor
- 65 Myr BP
- Extinction of 75% of all species!

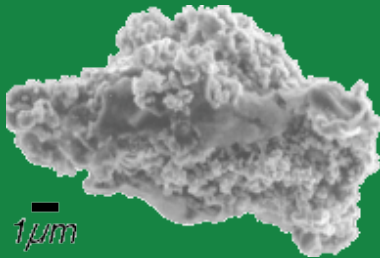


### Meteor (Barringer) Crater, Arizona

- 1 km diam. Crater
- 40-m diam Fe-meteorite
- 50 kyr BP
- 300,000 Mton
- 15 km/s

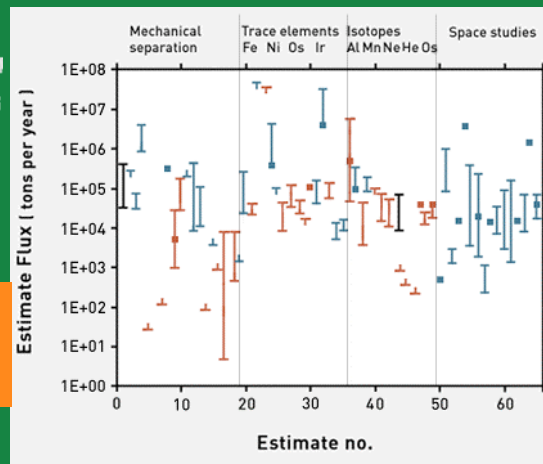
<http://www.gi.alaska.edu/rem sensing/features/impactcrater/imagexplain.htm>

## Interplanetary Dust Accumulation

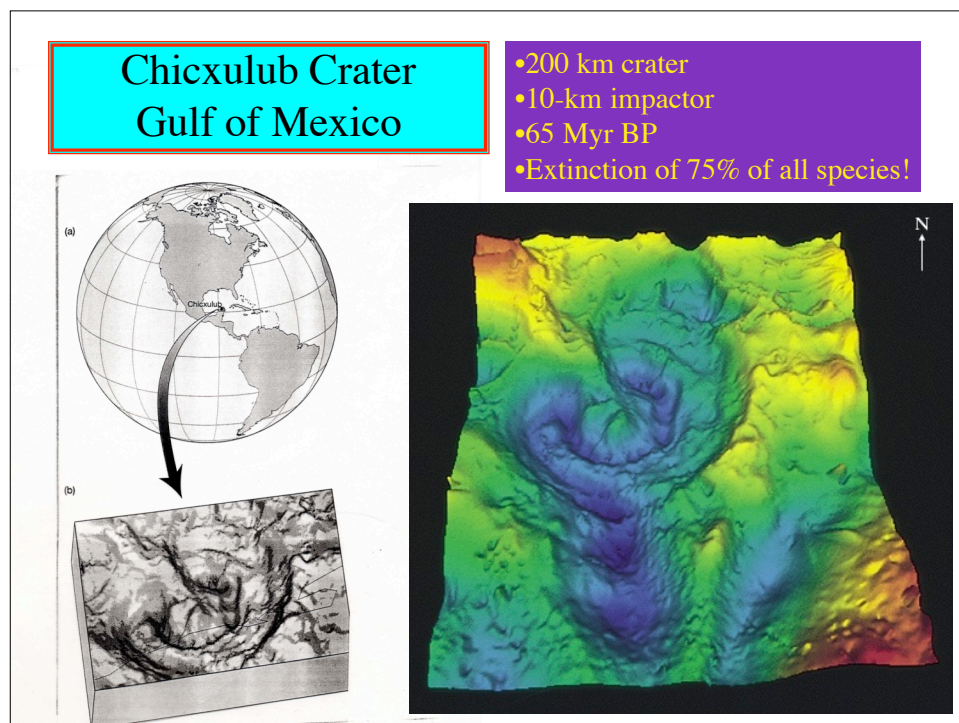
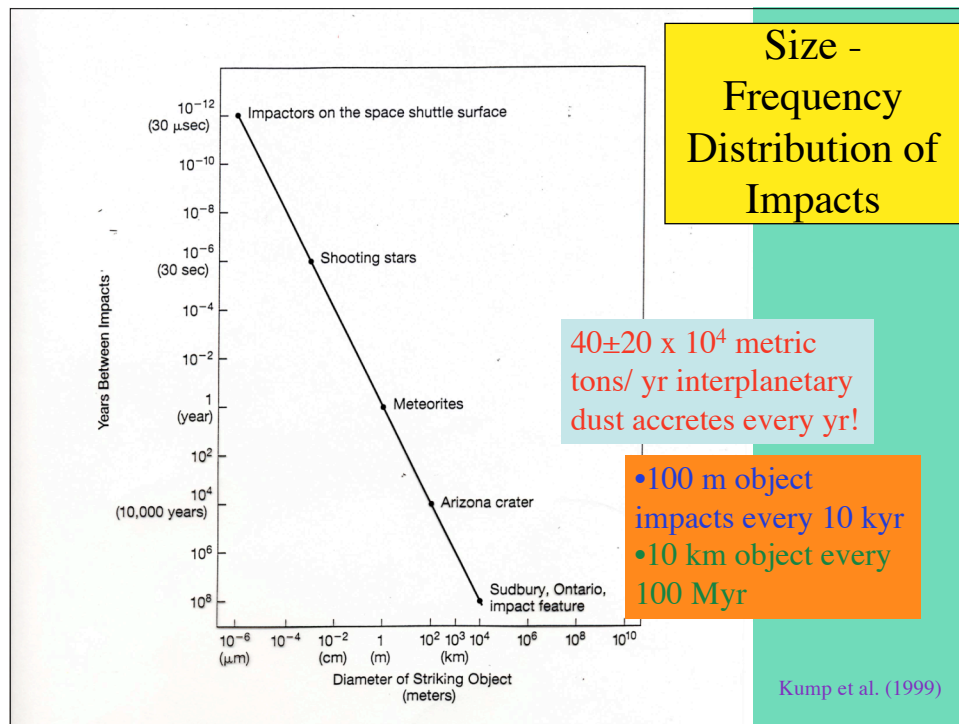


<http://presolar.wustl.edu/work/idp.html>

$40 \pm 20 \times 10^4$  metric tons/ yr  
( $40 \times 10^{10}$  g) interplanetary  
dust accretes every yr!

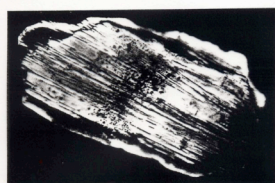
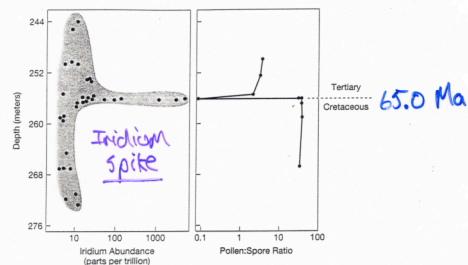


<http://www.whoi.edu/science/MCG/pge/project4.html>

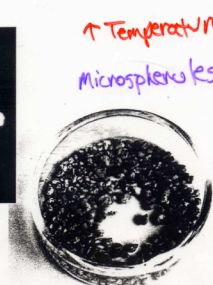




## Evidence for Meteorite Impact @ K-T Boundary



↑ Pressure  
Shocked Quartz



↑ Temperature  
Microspores

(b)

## Phanerozoic History of Extinctions

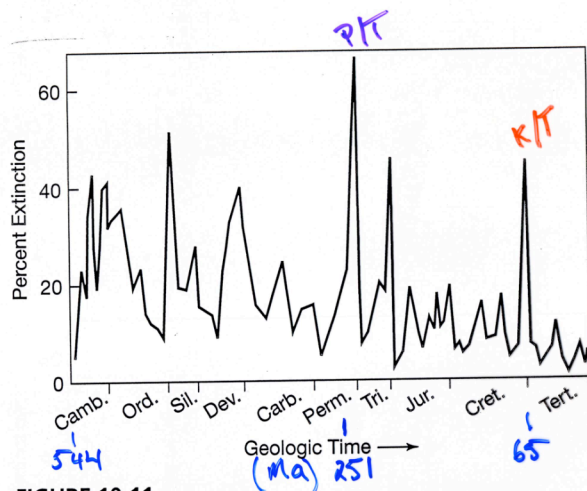
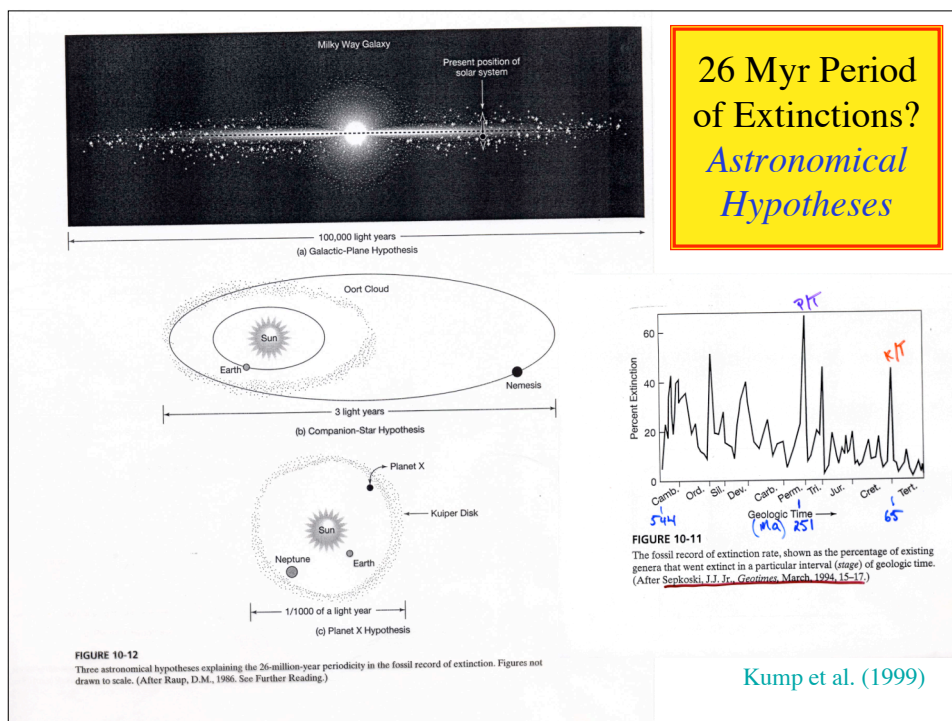
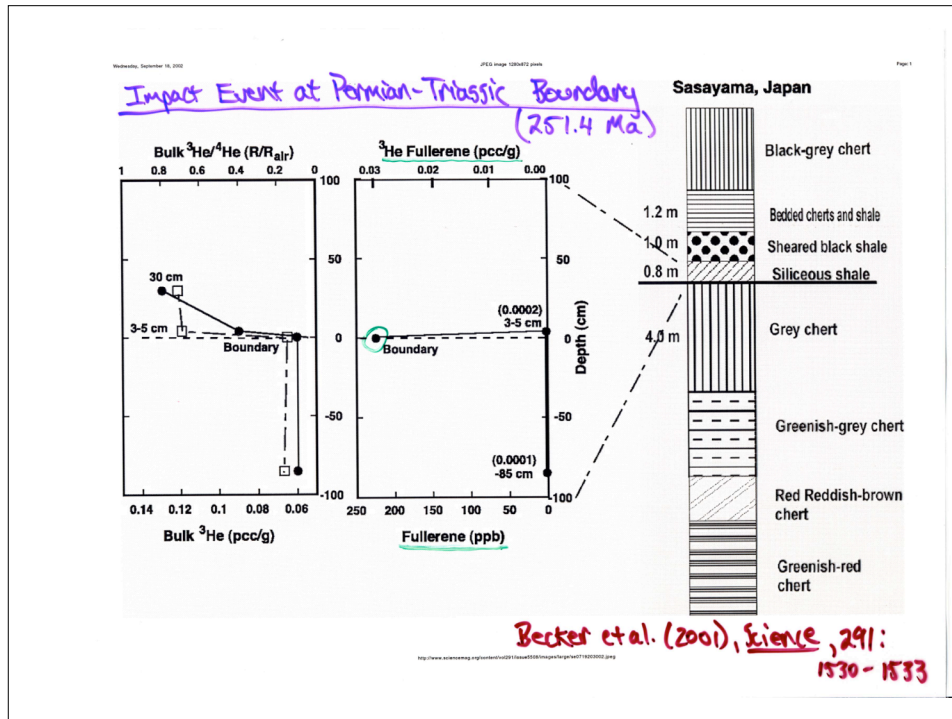
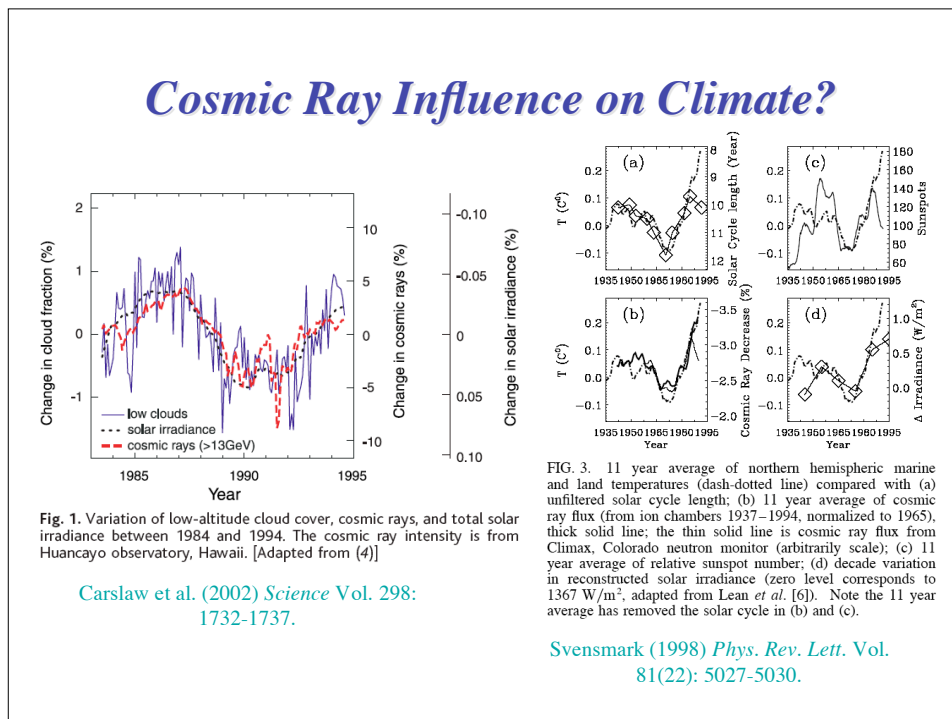
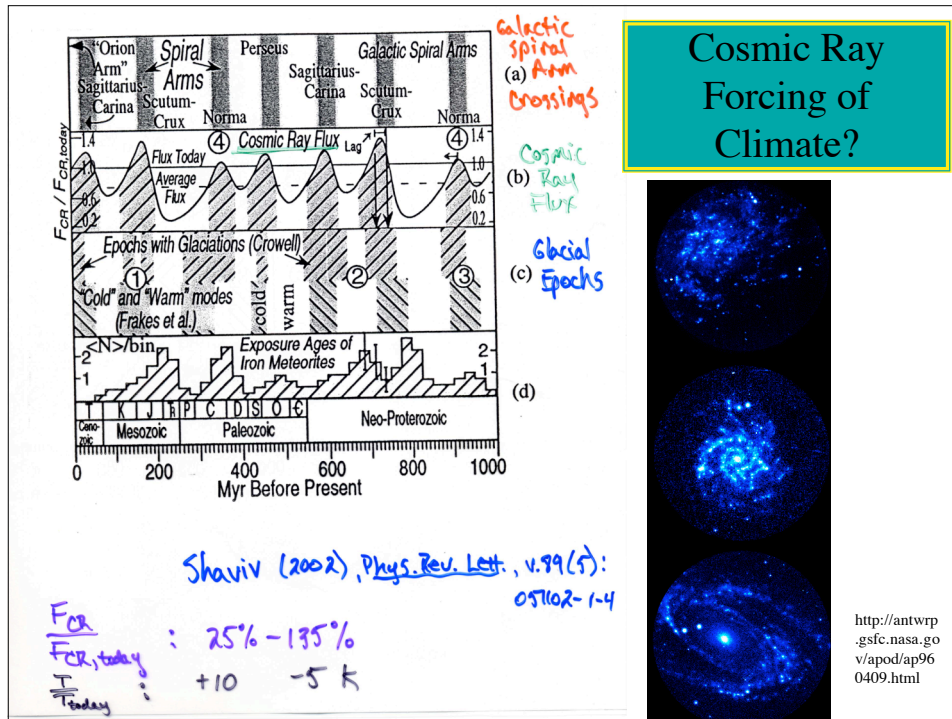


FIGURE 10-11

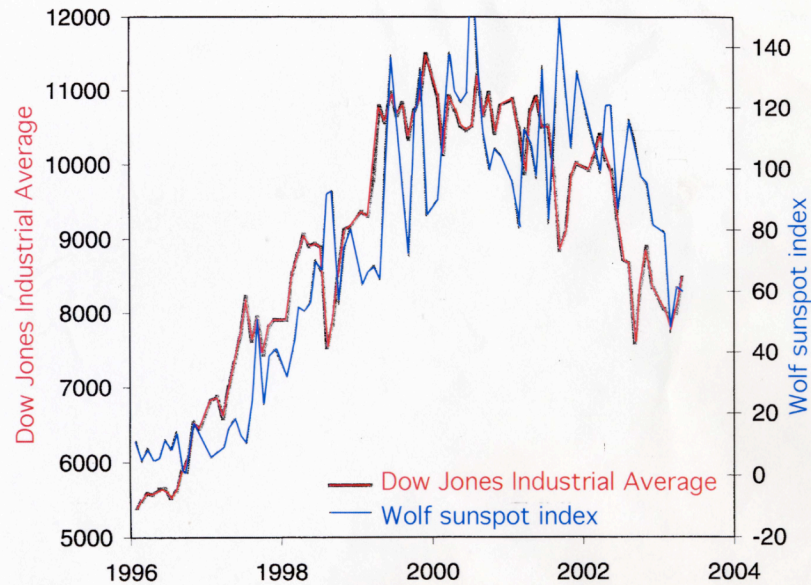
The fossil record of extinction rate, shown as the percentage of existing genera that went extinct in a particular interval (*stage*) of geologic time. (After Sepkoski, J.J. Jr., *Geotimes*, March, 1994, 15-17.)





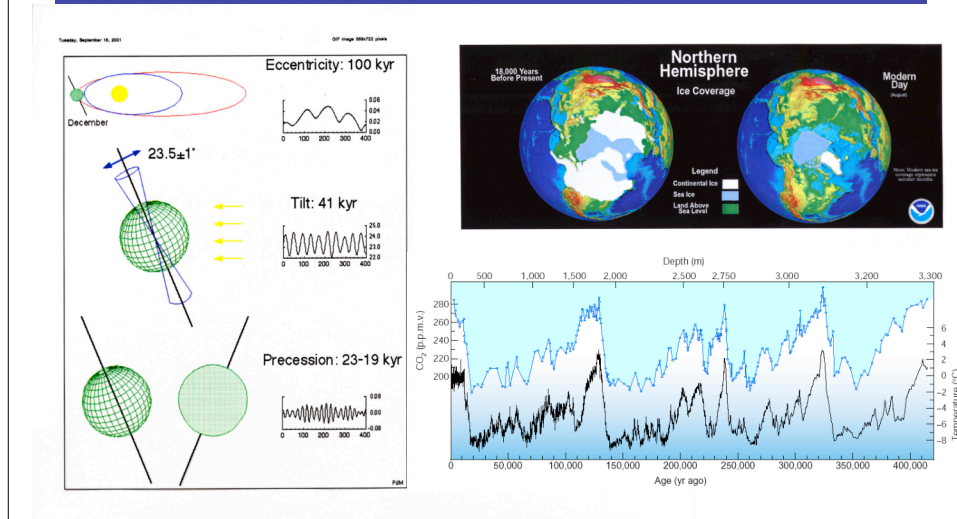


## *Correlation Is Not Causation*



## *Earth's Climate Part 4*

# Pleistocene Glaciations



## Earth's Orbital Geometry:

### The Milankovitch Hypothesis & the Pacing of Pleistocene Ice Ages



## Milankovitch Hypothesis: Historical Perspective

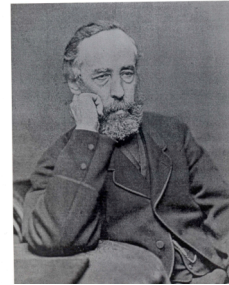
**What:** Astronomical theory of Pleistocene ice ages.

**How:** Varying orbital geometry influences climate by changing seasonal & latitudinal distribution of solar radiation incident at top of atmosphere (insolation).

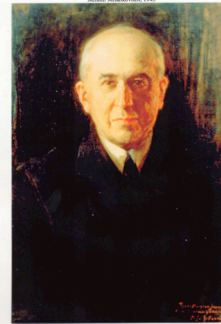
### Milestones: Hypothesis

- **Croll (1864, 1875):** Proposed that variations in seasonal influx of energy--the cumulative affect of eccentricity, obliquity & precession--could trigger large climate response.
- **Milankovitch (1920, 1941):** Combined laws of radiation with planetary mechanics to derive insolation curves as function of time (600 kyr) and latitude. Concluded summer insolation at high N. lat. ( $65^{\circ}\text{N}$ ) critical to growth/decay of ice sheets. "The Milankovitch Hypothesis".

James Croll, 1896



Milutin Milankovitch, 1943



## Eccentricity, Obliquity (tilt), Precession

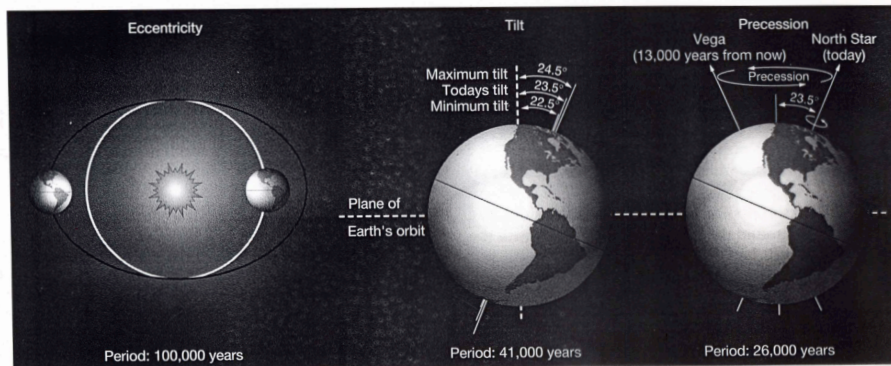


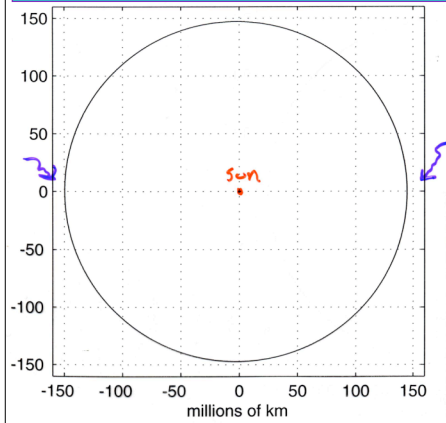
FIGURE 11-5

Aspects of Earth's orbit around the Sun that have implications for climate change. (a) The elliptical nature of the orbit (eccentricity) changes on 100,000-year time scales. (b) The tilt of the spin axis with respect to the plane of Earth's orbit around the Sun (obliquity) changes on a 41,000-year time scale. (c) The orientation of the spin axis in space wobbles (precesses) with periodicities of 19,000 and 23,000 years. (From J.P. Davidson, W.E. Reed, and P.M. Davis: *Exploring Earth: An Introduction to Physical Geology*, 1997. Reprinted by permission of Prentice Hall, Upper Saddle River, N.J.)

Kump et al. (1999)



## Eccentricity of Present Earth Orbit Around Sun (to Scale)



Present eccentricity = 0.017  
Range: 0 - 0.06  
100 & 400 kyr periods

Muller & MacDonald (2000)

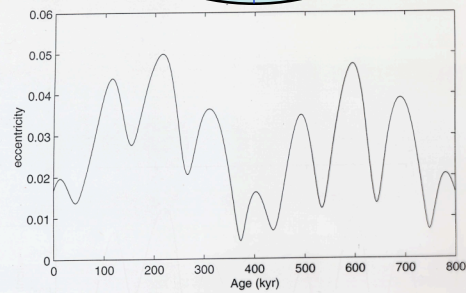
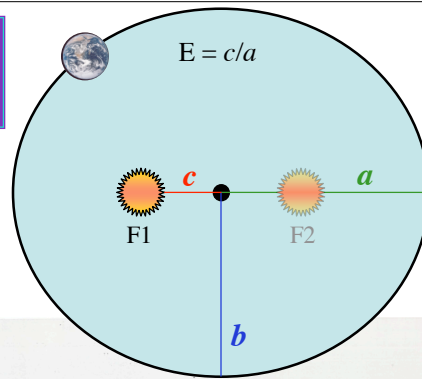


Fig. 2.6. The eccentricity of the Earth's orbit.

## Obliquity

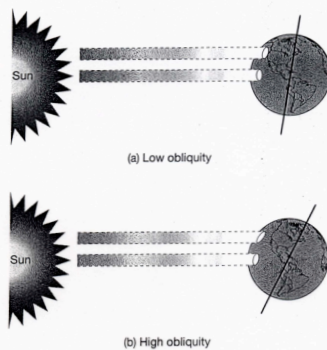


FIGURE 11-7  
(a) At low obliquity, Earth has less contrast in insolation between the seasons. (b) At high obliquity, the seasonal contrast is greater.

Kump et al. (1999)

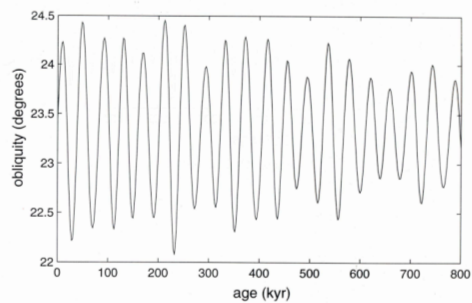
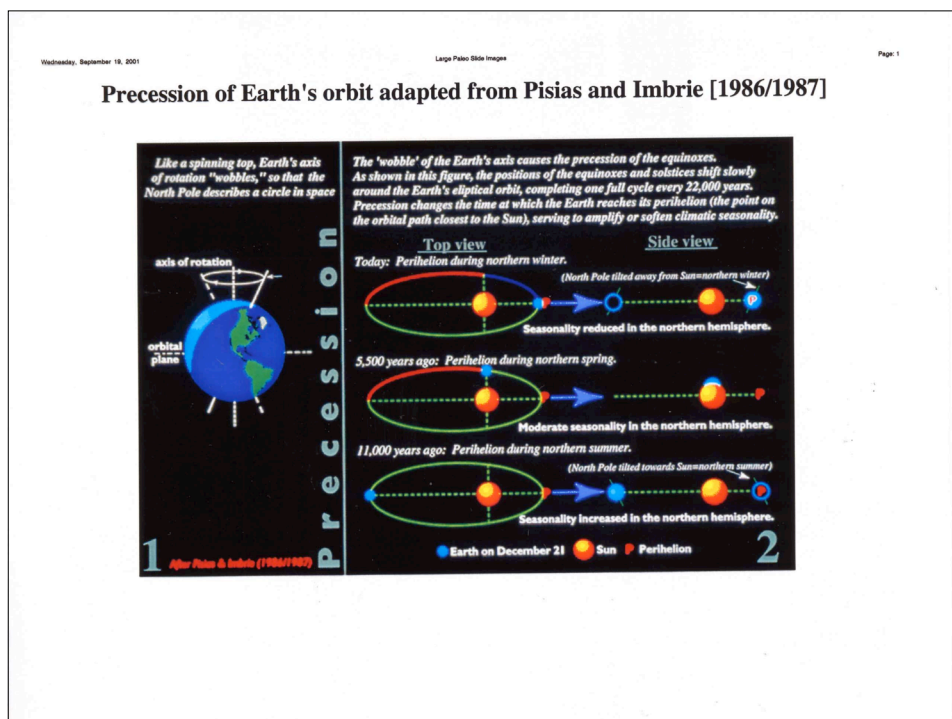
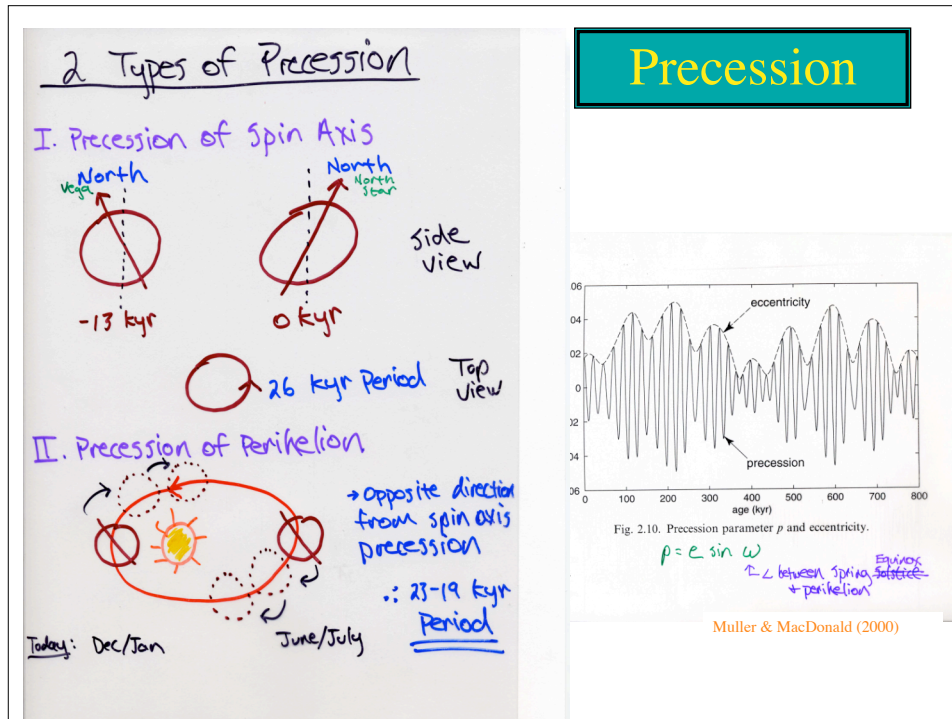
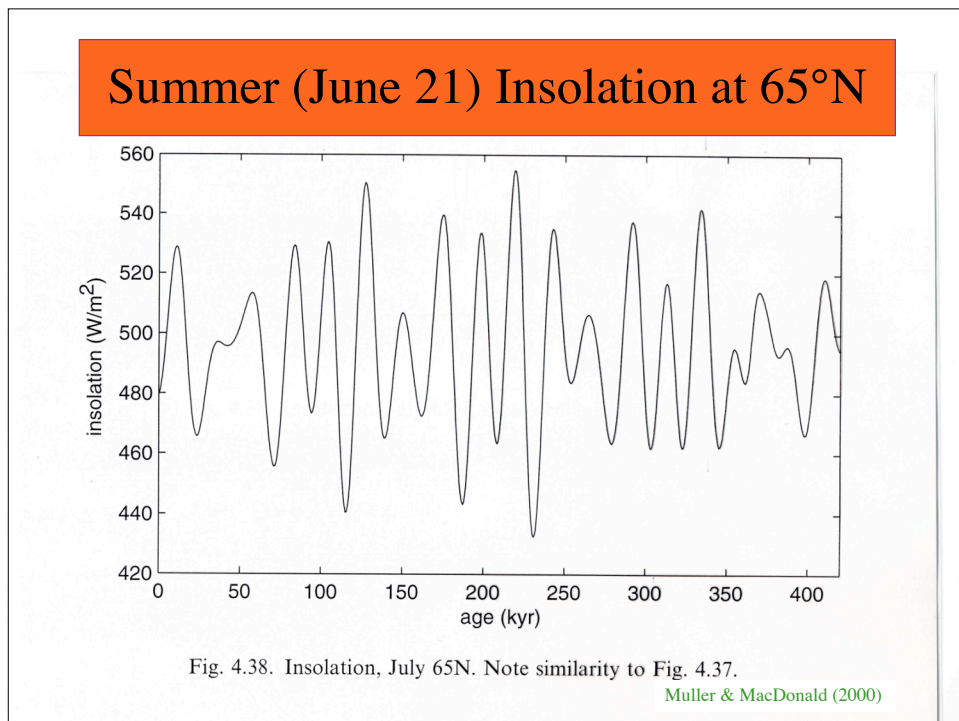
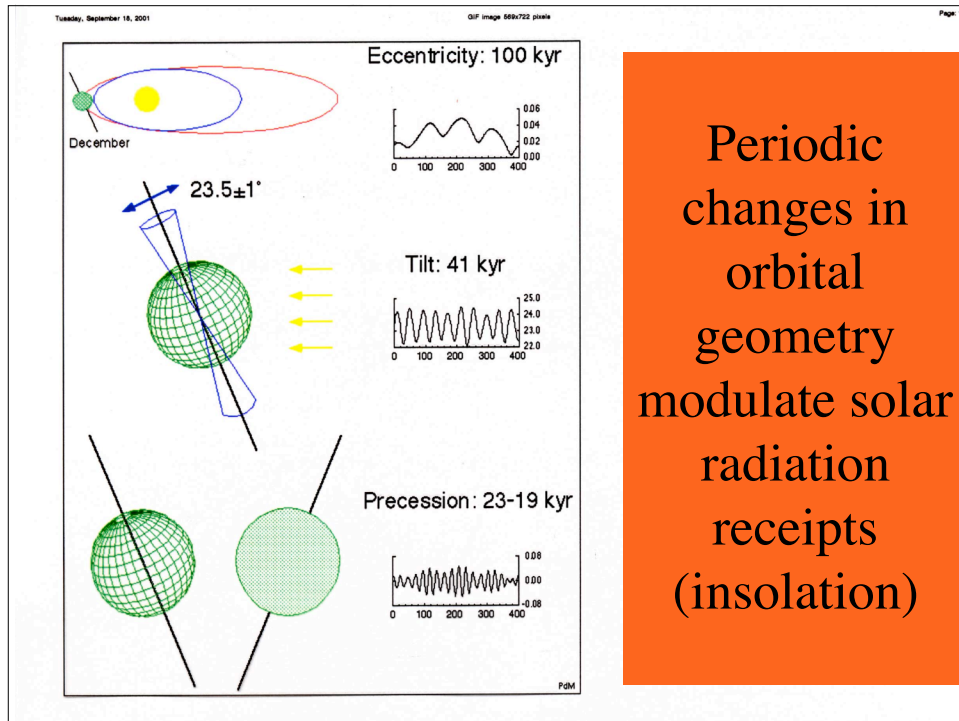
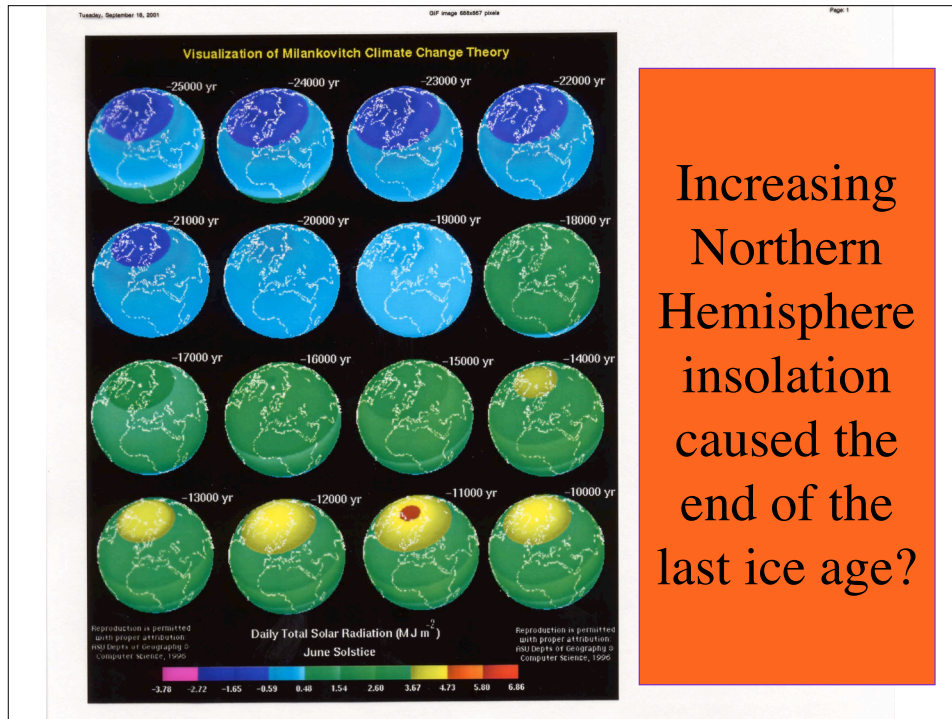


Fig. 2.12. Obliquity.

Muller & MacDonald (2000)



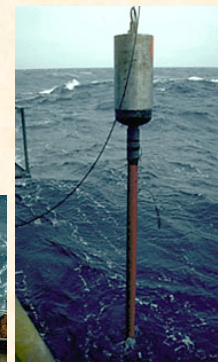




## Milankovitch Hypothesis: Milestones & Support

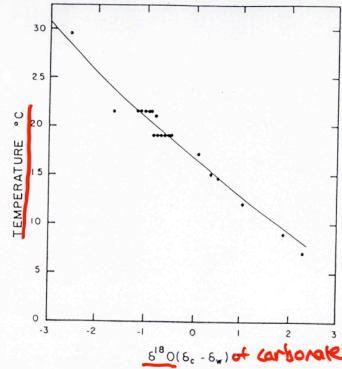
- Kullenberg (1947): Invented deep-sea piston coring.  
→ Recovery of long, continuous climate records possible.
- Emiliani (1955): Pioneered use  $^{18}\text{O}/^{16}\text{O}$  ratio of fossil foraminifera in sediment cores as climate (temp.) proxy.
- Olausson (1965); Shackleton (1967): Interpret foram  $\delta^{18}\text{O}$  changes as whole-ocean isotopic shifts caused by ice sheet growth/decay.
- 1960's: Recognition of magnetic stripes on ocean floor (geomagnetic field reversals) as global stratigraphic markers.
- Johnson (1982); Shackleton et al (1990): Use astronomically-driven insolation variations (E, T, P) to derive timescales for deep-sea cores.  
→ Accurately predict age of Brunhes-Matuyama (B/M) magnetic reversal, 780-790 kyr BP. (K/Ar dates for B/M incorrectly placed it at 730 kyr BP.)
- Baksi et al (1992):  $^{40}\text{Ar}/^{39}\text{Ar}$  date for B/M = 783 kyr BP.
- Raymo (1997): Multiple  $\delta^{18}\text{O}$  records on 'simple' timescale supports link between N. Hemisphere summer insolation and glacial terminations.

→ Strong support for astronomical influence on climate.  
(The magnitude of which remains debated...)





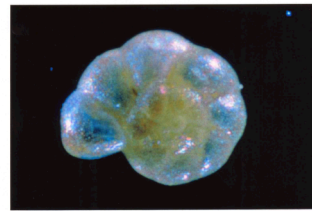
# T-dependent Isotopic Fractionation



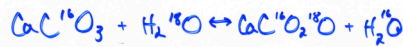
Natural Abundances  
 $^{16}\text{O} = 99.756\%$   
 $^{17}\text{O} = 0.039$   
 $^{18}\text{O} = 0.205$

A Temperature-Dependent Isotopic Fractionation of  $^{18}\text{O}$  from  $^{16}\text{O}$  Occurs During Calcification

Photo of Helenina anderseni



Foraminifera



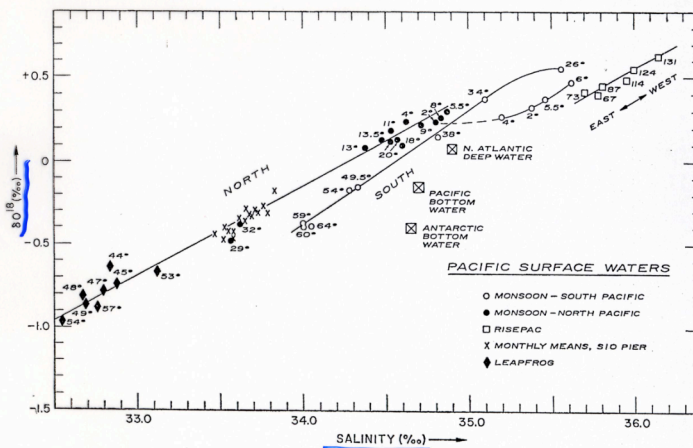
$$T = 16.9 - 4.2 (\delta^{18}\text{O}_c - \delta^{18}\text{O}_w)$$

$$\delta^{18}\text{O}_w = f(\text{Salinity, ice volume})$$

$$\delta^{18}\text{O} = \left[ \frac{(^{18}\text{O}/^{16}\text{O})_{\text{sample}}}{(^{18}\text{O}/^{16}\text{O})_{\text{standard}}} - 1 \right] \times 1000\text{‰}$$

Epstein et al. (1953)

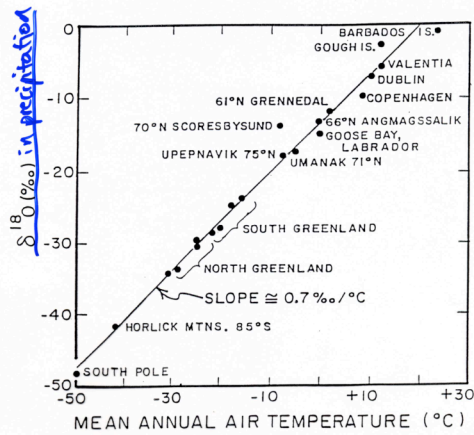
# $\delta^{18}\text{O}$ of Seawater varies w/ Salinity



H. Craig and L. I. Gordon (1965)

FIGURE 10. — Oxygen 18-salinity relationships in surface and deep Pacific Ocean samples. Latitudes are shown for Monsoon and Leapfrog samples; the figures on the RISEPAC expedition points are station numbers shown in figure 6.





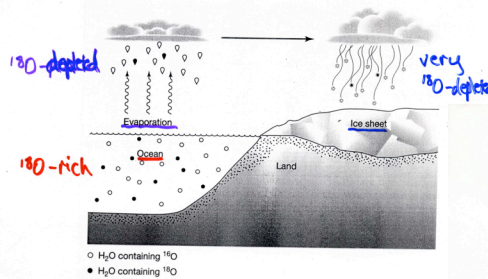
Dansgaard (1964)

•  $\delta^{18}\text{O}$  of precipitation varies with latitude & altitude.

• Water derived from melting snow & ice is highly depleted in  $^{18}\text{O}$ .

• Airmass trajectory influences  $\delta^{18}\text{O}$  of precipitation

$\text{H}_2^{18}\text{O}$  has 1% lower vapor P  
 $\therefore$  Cold Air (High latitude +/or Altitude)  
 has low  $\delta^{18}\text{O}$



Influence of Ice Volume on Ocean  $\delta^{18}\text{O}$

1.3‰ Decrease of Ocean  $\delta^{18}\text{O}$  from Melting Glaciers

120 m Sea Level Depression @ LGM

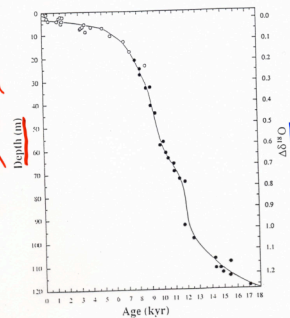
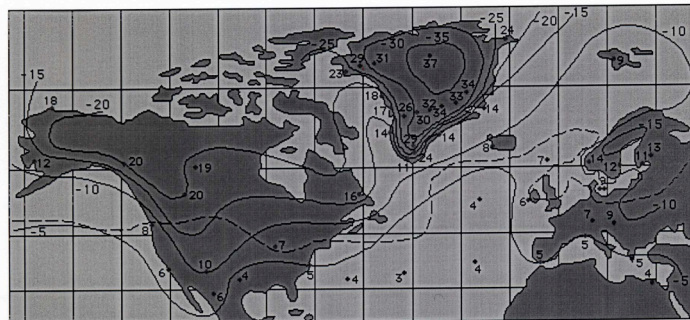


FIG. 2 Barbados sea level curve based on radiocarbon-dated *A. palmata* (filled circles) compared with *A. palmata* age-depth data<sup>3</sup> (open circles) for four other Caribbean island locations. All radiocarbon ages in this figure are corrected for local seawater  $\Delta^{14}\text{C}$  by subtracting 400 yr from the measured radiocarbon ages<sup>3</sup> but they are not corrected for secular changes in atmospheric  $^{14}\text{C}$  levels<sup>11</sup>. The Barbados data are corrected for the estimated mean uplift of 34 cm kyr<sup>-1</sup>. The right-hand axis of the Barbados sea level curve (solid line) is scaled to the estimated  $\delta^{18}\text{O}$  change of mean ocean water using the calibration in ref. 5.

Fairbanks (1989), Nature, 342: 637-642

## Global Variation of $\delta^{18}\text{O}$ in Precipitation



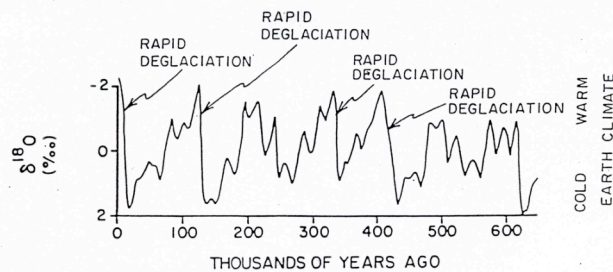
Numerical mean  $\delta$  values of annual precipitation, covering periods of 2 to 10 years, and approximate iso- $\delta$ -lines;  $\delta$  values are relative per mille deviations of  $\text{O}^{18}/\text{O}^{16}$  ratios from Standard Mean Ocean Water <sup>1</sup>. The influence of the present Gulf Stream is seen very clearly in the course of the isoline for  $\delta = -10$  per mille. The dashed line indicates the maximum extension of continental ice and permanent pack ice during the last glacial maximum according to Flint <sup>2</sup>.

<sup>1</sup>Craig, H. (1961) *Science* **133**, 1833.

<sup>2</sup>Flint, R.F. (1957) *Glacial and Pleistocene Geology*. New York: Wiley, plate 3.

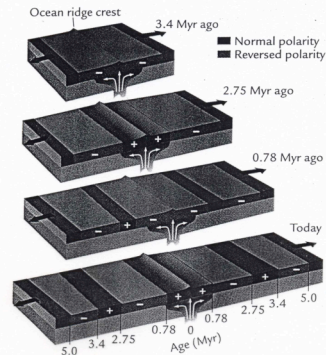
Adapted from: Danagard, W. (1969) Glacier Oxygen-18 Content and Pleistocene Ocean Temperatures, *Science* **166**, 3904, pp. 499-502.

Note: All values should be *negative*.



**Figure 5.** The  $^{18}\text{O}$ - $^{16}\text{O}$  record for benthic foraminifera from deep sea sediments. The oxygen isotope results are given as percent difference in the  $^{18}\text{O}$  to  $^{16}\text{O}$  ratio from that in an international reference standard. The more positive the  $\delta^{18}\text{O}$  value the larger the  $^{18}\text{O}$  content of sea water. As the growth of ice caps enriches sea water in  $^{18}\text{O}$ , these high values correspond to times of large continental ice cover. The time scale was obtained from radioisotope measurements on the deep sea sediments. Although not a regular progression, the times of large ice cover follow one another at roughly 100,000 year intervals. Note also the rapidity with which the largest of these ice masses disappeared!

## Magnetic Stripes



Magnetic Stratigraphy

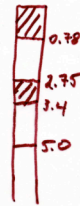


FIGURE 5-8 Magnetization of ocean crust Successive bands of ocean crust form as molten lava erupts at the seafloor, cools, and solidifies. The new crust is magnetized in the normal or reversed polarity prevailing at the time. As the plates move apart, equal amounts of magnetized crust are carried away from the ridge axis in both directions and can be used to date the seafloor. (Modified from F. Press and R. Siever, *Understanding Earth*, 2nd ed., © 1998 by W. H. Freeman and Company.)

- ①  $^{40}\text{Ar}/^{39}\text{Ar}$  - date lava flows
- ② Apply dates to downcore reversals

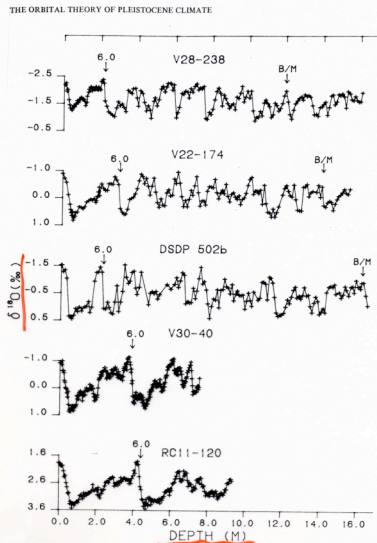


Figure 4 Variations in  $\delta^{18}\text{O}$  as a function of depth in five deep-sea cores. Two important stratigraphic levels are labeled as follows: 6.0 for the boundary between isotope stages 5 and 6; and B/M for the magnetic reversal at the Brunhes-Matuyama boundary. See Table 1.

Imbrie et al. (1984)

## Simple Timescale

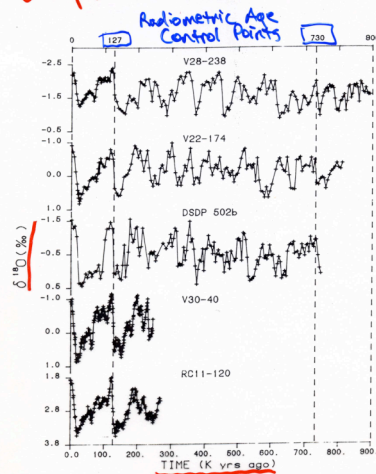


Figure 5 Variations in  $\delta^{18}\text{O}$  as a function of estimated time in five deep-sea cores. The time scale is derived by linear interpolation between (and extrapolation beyond) control points at 127 KY, 730 KY BP. For details, see text and Table 5.

Imbrie et al. (1984)

## Filter $\delta^{18}\text{O}$ Record at Obliquity + Precession Frequencies

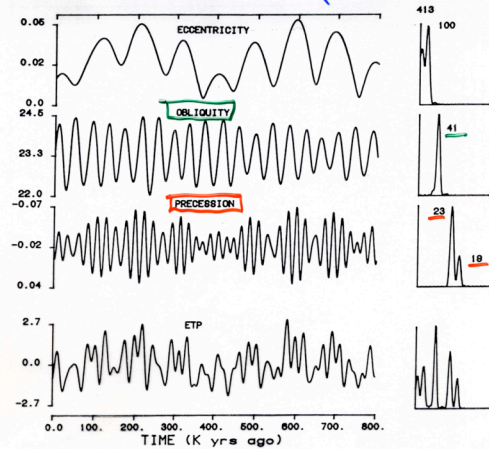


Figure 2 Variations in eccentricity, obliquity, and the precession index ( $\Delta\epsilon \sin \omega$ ) over the past 800 000 years. Left: The three upper time series are from the work of Berger (1). These have been normalized and added to form the curve labeled ETP. The scale for obliquity is in degrees; for ETP, in standard deviation units. Right: Variance spectra calculated from these time series, with the dominant periods (KY) of conspicuous peaks indicated.

Imbrie et al. (1984)



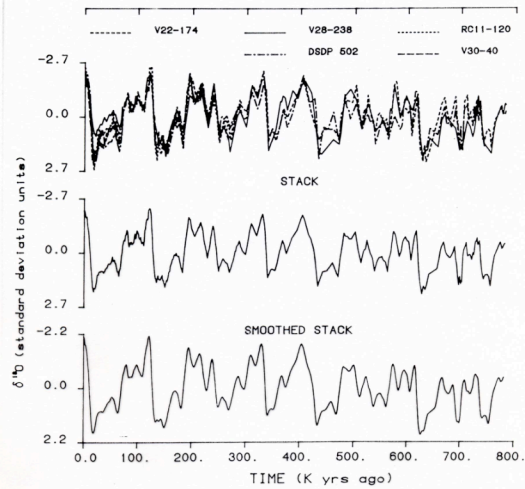


Figure 8  $\delta^{18}\text{O}$  variations in five deep-sea cores normalized and plotted on the SPECMAP time scale. In the top panel, data from each core has been normalized to zero mean and unit standard deviation. After interpolation at intervals of 1 KY, these curves have been averaged (middle panel), and smoothed with a 9-point Gaussian filter (bottom panel).

Imbrie et al. (1984)

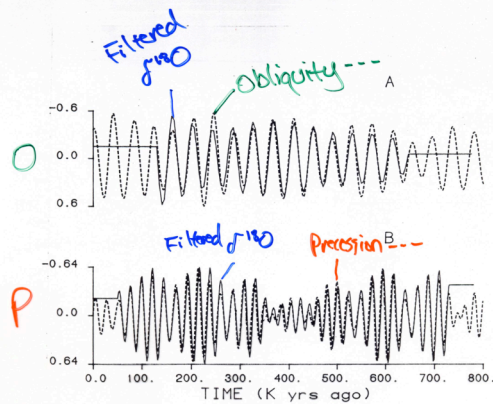


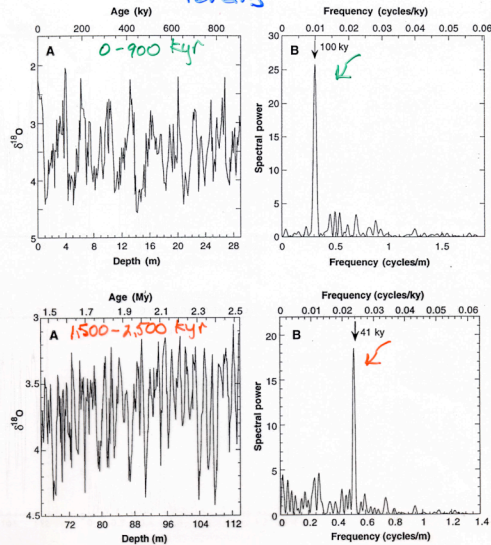
Figure 9 Variations in obliquity, precession, and the corresponding frequency components of  $\delta^{18}\text{O}$  over the past 800 KY. Dashed lines are phase-shifted versions of obliquity (A) and precession (B) curves. Solid lines are filtered versions of the stacked  $\delta^{18}\text{O}$  record plotted on the SPECMAP time scale. The filters used were centered on periods of 41 KY (A) and 22 KY (B). All curves have been transformed to have zero means with arbitrary scales.

High coherence (i.e., amplitude correlation) lends support to astronomical forcing of climate.

Imbrie et al. (1984)



## Narrow Spectral Peaks Support Astronomical Forcing



⇒ Linear sedimentation Rates Assumed!

Muller + MacDonald (1997)  
SCIENCE, 277: 215-218

### 1. Why is climate response in 100-kyr band so strong?

**Observation:** High correlation of  $\delta^{18}\text{O}$  cycles with astronomically-driven radiation cycles at E, T & P frequencies suggests causal link in all 3 bands.

**Problem:** Amplitude of insolation change ( $\sim 0.2\%$ ) is  $\sim 10\times$  smaller than in T,P bands.

**Possible Solution:** E modulates climatic effect of P. High E favors NH glaciation when P causes NH summer to occur at maximum Earth-Sun distance (i.e., Imbrie et al, 1993).

### 2. Why do glacial cycles switch from 41-kyr to 100-kyr period $\sim 700$ kyr BP?

**Possible solution:** L/T cooling trend, perhaps from tectonically-driven decrease in atmospheric  $\text{CO}_2$ , facilitates NH ice sheet growth beyond a critical threshold during insolation minima. These large ice sheets drive climate through feedbacks internal to the climate system (geo-, cryo-, atmo-, hydro-sphere).

### 3. Why do full glacial Terminations, and ensuing interglacial periods, occur $\sim 430$ and $\sim 15$ kyr BP when E is very low?

**Possible solution:** 100-kyr cycle of orbital inclination (Muller and MacDonald, 1995).

**Caveat:** no obvious mechanism linking climate to inclination.

## Milankovitch Hypothesis Challenges:

### 100-kyr Cycle Problems

## Eccentricity Modulates Precession

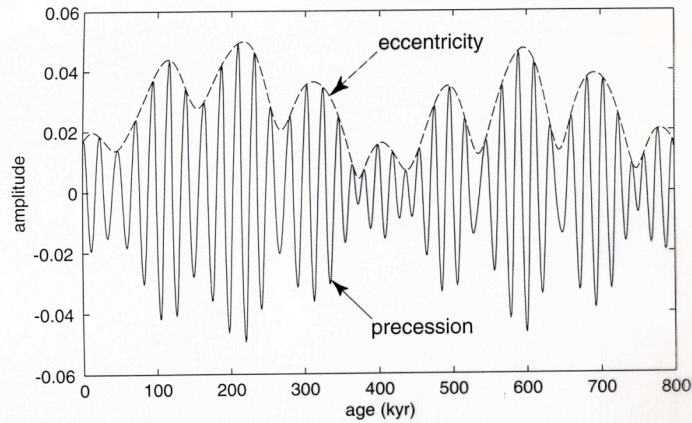


Fig. 2.10. Precession parameter  $p$  and eccentricity.

$$p = e \sin \omega$$

Muller & MacDonald (2000)

$\omega$  =  $\angle$  between Spring Equinox & perihelion

## 100-kyr Cycle Problem-1

Glacial Climate Feedbacks 219

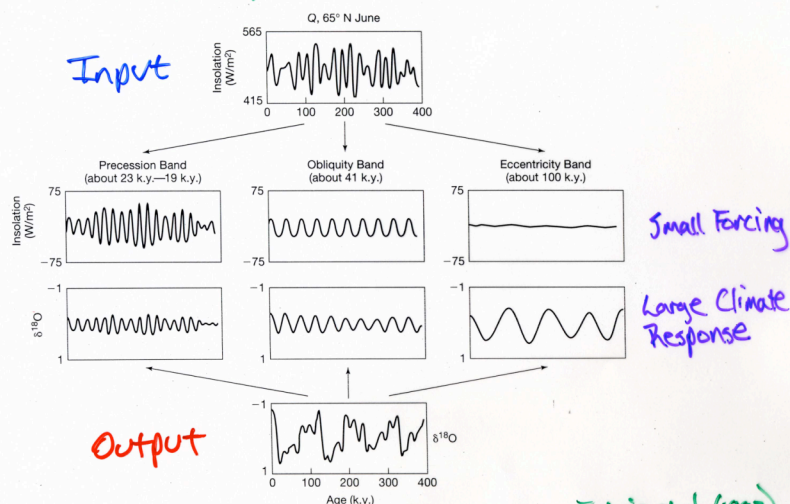
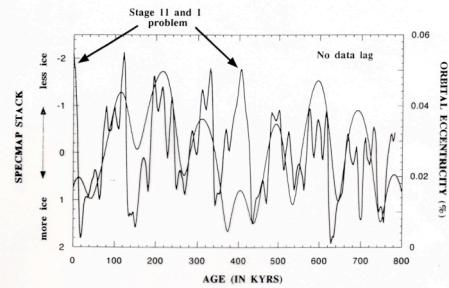
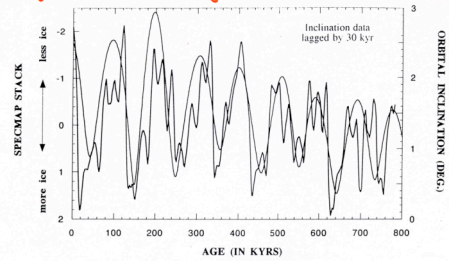


FIGURE 11-9

Northern Hemisphere June insolation ( $Q$ , the climate forcing) and marine oxygen isotopic composition ( $\delta^{18}O$ , the climate response) and their dominant periodic components. (After Imbrie et al., *Paleoceanography* 7:701–738, 1992.)

Imbrie et al. (1993)

# 100 kyr Climate Forcing from Orbital Inclination?



Muller + MacDonald (1995), *Nature*, v.377:108-108

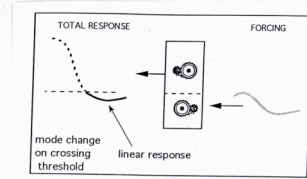
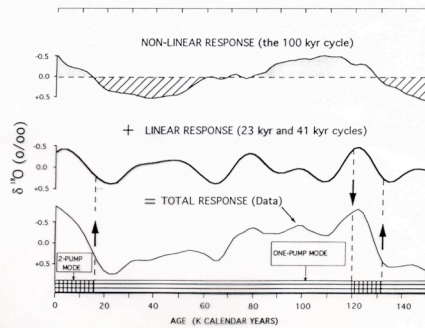
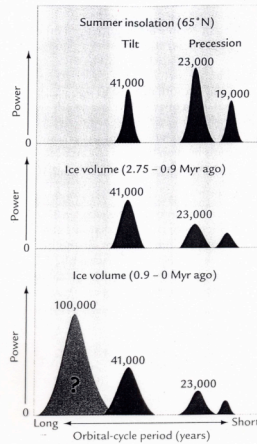


Fig. 17. Model of a nonlinear amplification mechanism. The system's gain is sharply increased when the linear response to an external forcing crosses a threshold.



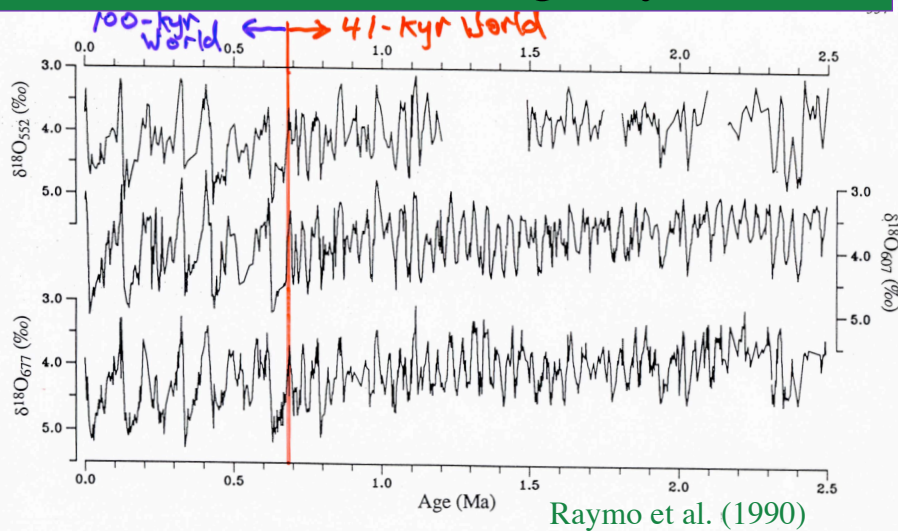
Imbrie et al. (1993)

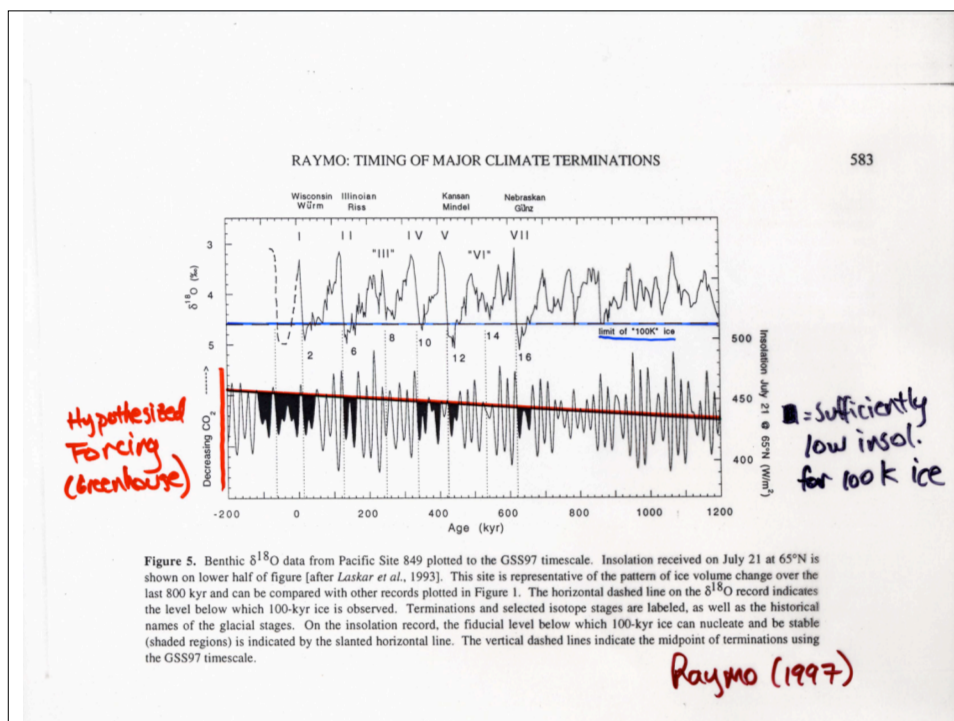
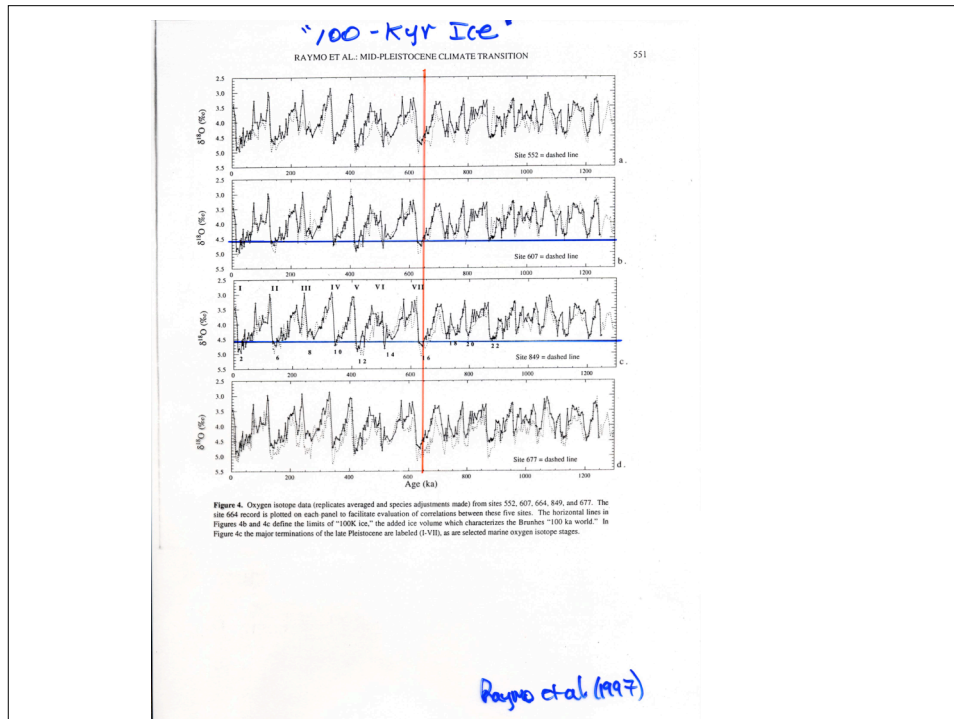
## The 100 kyr Cycle Problem-2



**FIGURE 10-17 Spectral analysis: insolation and ice volume**  
Summer insolation changes at 65°N occur only at the orbital rhythms of tilt (41,000 years) and precession (23,000 and 19,000 years) (top). Between 2.75 and 0.9 Myr ago, the  $\delta^{18}\text{O}$  signal (northern hemisphere ice volume and deep-ocean temperature) contains the same rhythms as orbital insolation (middle). The large 100,000-year cycle evident in the  $\delta^{18}\text{O}$  (ice volume) signal since 0.9 Myr ago is not present in the insolation signal (bottom).

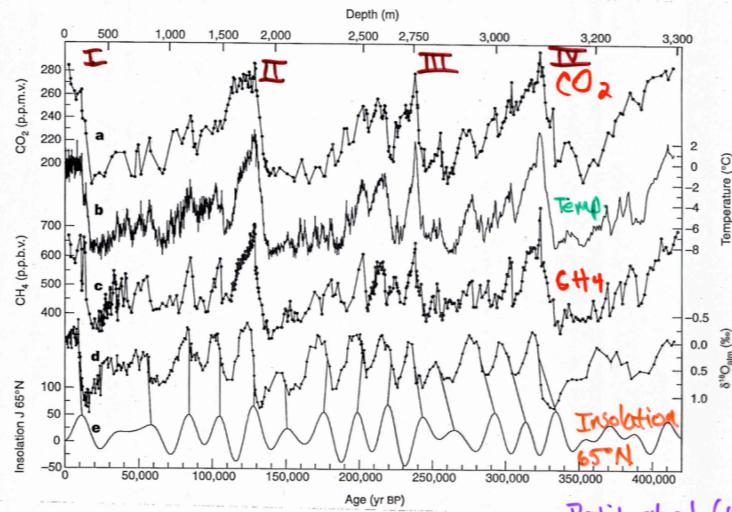
## Pleistocene Ice Age Cycles







## 420,000 Years of Climate at Vostok, Antarctica



Petit et al. (1999)