Lecture 26

The Carbon Cycle and Earth's Climate

Reading: (a) Ch12 (sec 12.9 or digital 607- end) (Mon)
(b) Ch 18 Langmuir/Broecker, "How to Build a Habitable Planet" (Wed)
(c) Ch 20 Langmuir/Broecker, "How to Build a Habitable Planet" (Fri)

Turn in Journals on Friday Nov. 1

Last time
1. Stable Isotopes

Today
2. a. Paleoclimate records and global climate controls
   b. The Carbon Cycle

Next time
3. Future Climate in the near term

Paleo-climate records:

As we saw last time, δD and δO^{18} of meteoric waters depend on temperature:

More fractionation at lower temperature, less at higher temperature

Relatively cold eras in the past of Earth history would have thus experience enhanced global O^{18}-O^{16} (and H-D) fractionation.

Relatively warm eras would have seen subdued fractionation.

In other words. the fresh surface hydrosphere is:

more fractionated (isotopically lighter)	during cold epochs
less fractionated (isotopically heavier)	during warm epochs (such as we are presently enjoying)

Paleo-climate records:

In essence, the O and H isotopic composition of the hydrosphere is a thermometer.

Two big questions are:

a. do we "know the code" for deciphering the thermometry information from materials in the geologic record?

b. how certain are we that other non-temperature related affects haven't also played a role, and if they have, can we correct for them?

Paleo-climate records:

The combination of oxygen isotope fractionation that occurs during evaporation of water from the oceans and during rain-out over land produces:

- ❖ atmospheric water with low δO^{18} and therefore relatively high δO^{18} surface oceans at relatively cold times.
- ❖ atmospheric water with high δO^{18} and relatively low δO^{18} surface oceans at relatively warm times.

Figure 9.8. V ariation in $\delta^{18}O$ in precipitation as a function of mean annual temperature

from White, "Geochemistry"

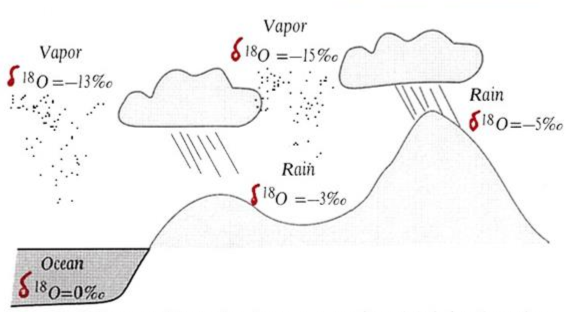


Figure 9.9. Cartoon illustrating the process of Rayleigh fractionation and the decreasing $\delta^{18}O$ in rain as it moves inland.

Paleo-climate records:

A second important consideration is relative sea water and ice cap volume fluctuations over Earth history.

During cold times, more H_2O is stored in ice caps and less in the oceans.

During warm times, less H_2O is stored in ice caps and less in the oceans.

Evaporation of seawater feeds the ice caps.

The resulting warm-cold era fluctuation in surface seawater isotopic composition enhances the temperature dependent O^{18}-O^{16} fractionation signature when water evaporates from it.

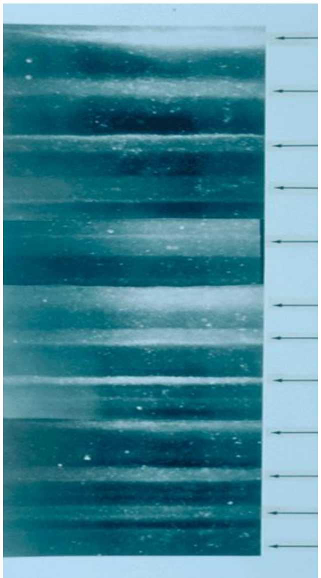
> i.e., during cold times seawater itself is isotopically heavier than it is in warm times due to the ice volume and the T-dependent effects.

Ice cores and marine sediments

The ***simplest*** O and H isotopic record to interpret is in the composition of ***ice layers in the polar ice caps***, which are the second largest reservoir of the hydrosphere (2.05% today).

The growth of the ice caps occurs primarily at the expense of the oceans, and vice versa.

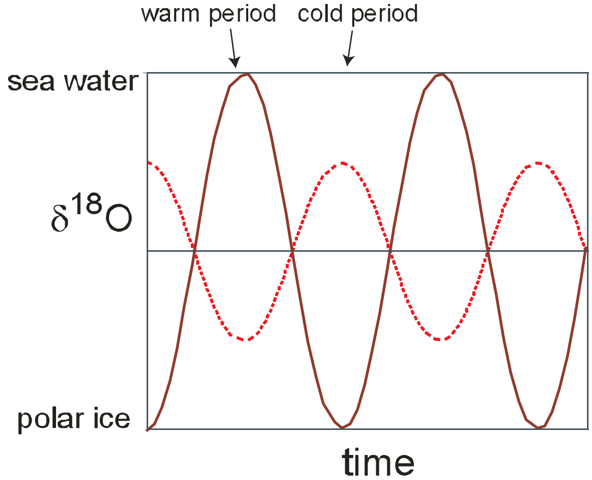
As more H_2O is stored in the ice caps and less is left in the oceans, the surface oceans show a complementary isotopic shift.



19 cm long section of Greenland Ice Sheet Project 2 ice core from 1855 meters showing annual layer structure illuminated from below by a fiber optic source. Section contains 11 annual layers with summer layers (arrowed) sandwiched between darker winter layers. *Image source: Wikimedia Commons*

Ice cores and marine sediments

The ice core record and records of surface sea water δO^{18} are negatively correlated (mirror images of each other)



Marine organisms take on isotopic compositions that reflect changes in the water with time, so that the fossil record represents a second, less straight-forward, proxy record.

Paleoclimate records – Ice Cores

Both the ice core and the sediment records demonstrate that global mean temperatures have varied greatly over the past few Ma.

This ice core record is from Vostok, Antarctica.

The longest ice cores generally only go back to about 200 ka.

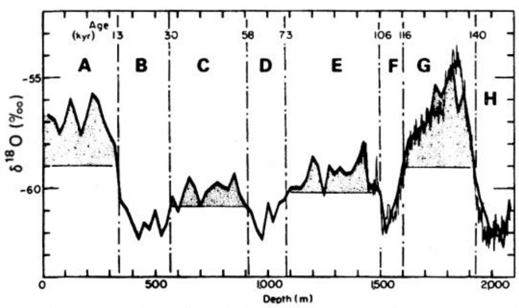
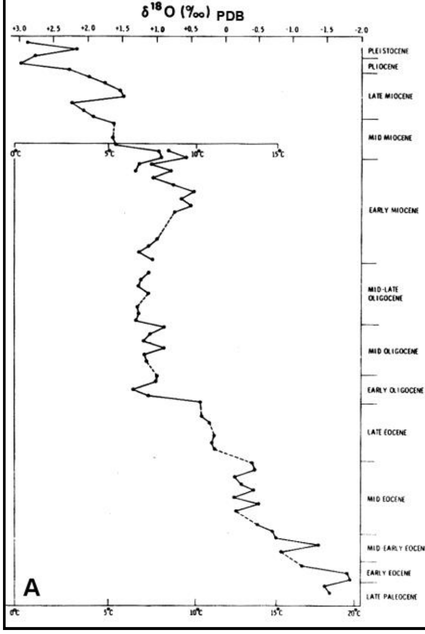
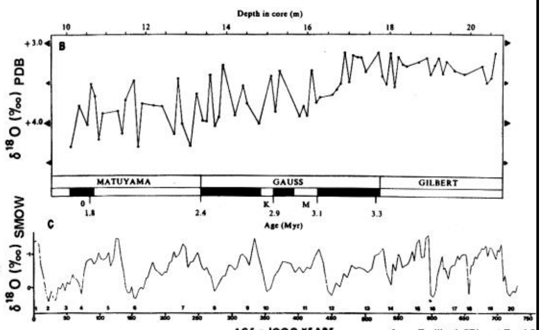


Figure 24.10. Oxygen-isotopic record from the Vostok ice core in Antarctica (78°28'S, 106°48'E, 2,083 m long), subdivided into warmer (A, C, E, G) and colder (B, D, F, H) stages. The record is continuous from the penultimate glaciation (H) to the present. The bottom scale gives the depth in meters below the core top. The top scale gives the age in thousands of years. Because the ice below compresses as more ice accumulates on top, stage G has a thickness similar to that of stage A, but it represents almost twice the time. The isotope peak of the last interglacial age (G) is higher than that of the present interglacial (A), reflecting the additional amount of ice that melted at that time and the concomitant high sea level of +6 m (see Fig. 16.10). The isotopic trend of the present interglacial (A) indicates that we have passed the peak. (Lorius et al., 1985, p 592, Fig. 1.) *from Emiliani, "Planet Earth"*

Paleoclimate records - The sediment record

Biogenic $CaCO_3$ in marine sediments are a good proxy record of sea water δO^{18} that go much further into the past.





from Emiliani, "Planet Earth"

Figure 24.7. The oxygen-isotopic record of planktic foraminifera. All values refer to the Chicago standard PDB-1. (A) Isotopic record, 60·10⁶ y long, showing the increasing amount of ice on Earth and the accompanying temperature decrease. Approximate temperatures are shown only for Cenozoic time before the Middle Miocene, because after that time the Antarctic ice significantly expanded, and northern glaciations began. The isotopic composition of seawater was drastically altered, making temperature estimates uncertain. (Shackleton and Kennett, 1975.) (B) Isotopic record from 4·10⁶ to 1.6·10⁶ y B.P. from the equatorial Atlantic. The isotopic minimum at 2.36·10⁶ y represents the first major northern glaciation. (Shackleton and Opdyke, 1977.) (C) Isotopic record of the past 750,000 y from the tropical Atlantic and the Caribbean, showing the alternation of glacial and interglacial ages. (Emiliani, 1978, p. 353, Fig. 2.)

Paleoclimate records - The sediment record

δO^{18} in $CaCO_3$ will reflect δO^{18} of the H_2O it was formed in. There is also a temperature dependence to the fractionation.

$T = 16.9 - 4.2(\delta_c - \delta_w) + 0.13(\delta_c - \delta_w)^2$ for δ relative to smow

Fortunately, surface-temperature variations are minimal near the Equator between warm and cold times.

Thus, equatorial planktonic tests and shells yield a record that's relatively independent of water temperature.

$\delta^{18}O$ variations in equatorial planktonic tests and shells mainly reflect fractionation during *evaporation/precipitation of H_2O and formation/melting of ice caps*.

There are also biological (species-specific) effects on fractionation.

In addition, post-depositional alteration and the reworking of sediments by benthic organisms can change $\delta^{18}O_{CaCO3}$.

Paleoclimate records - The sediment record

The Quaternary planktonic foraminifera record shows significant cyclicity

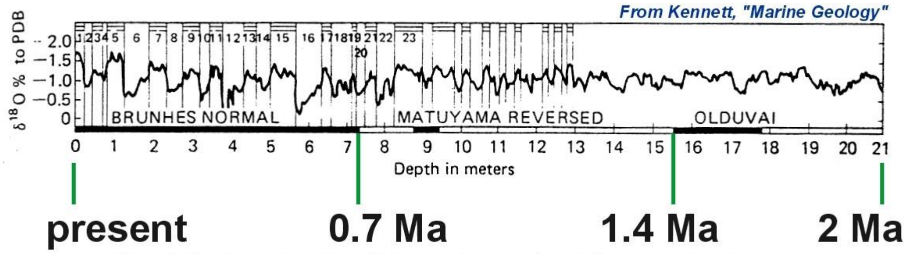


Figure 3–12 Oxygen isotopic oscillations in planktonic foraminifera expressed as deviation from Emiliani B1 standard for last 2 m.y. scaled against paleomagnetic record in piston core V28–239. Stages in oxygen isotopic record (1 through 23) are numbered after Emiliani [1955; 1966] and Shackleton and Opdyke [1973]. The last 0.7 Ma (Brunhes) contains glacial stages represented by amplitudes in excess of 1 per mil and with periodicities of about 100,000 years. Isotopic minima are approximately the same in the different "core stages." The mid-Matuyama (~1.4–0.7 Ma) interval contains isotopic fluctuations with approximately 40,000-year periodicities. Amplitudes are lower (0.7 per mil) than during the Brunhes. (After N. J. Shackleton and N. D. Opdyke, 1976, courtesy The Geological Society of America)

Paleoclimate records - The sediment record

Biogenic $CaCO_3$ from planktonic organisms in sediments records surface sea water δO^{18}. Benthic organisms ones tell about deeper waters.

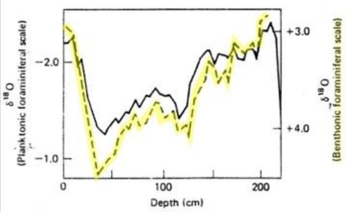


Figure 3–11 Oxygen isotopic records of planktonic versus benthonic foraminifera in piston core V28-238. Note scale differences at left and right figure margins. (After N. J. Shackleton and N. D. Opdyke, 1973)

From Kennett, "Marine Geology"

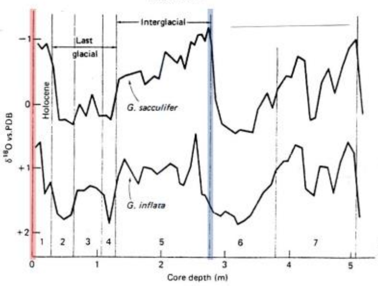
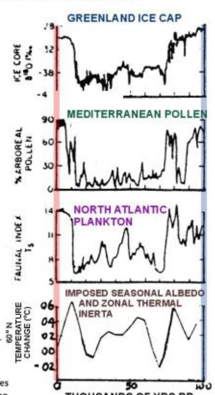


Figure 17–13 Oxygen isotopic record of two planktonic foraminiferal species through the late Quaternary (approximately 260,000 years) in the southern Indian Ocean. Pleistocene stage numbers are shown at bottom. (After J. C. Duplessy, 1978)

From Kennett, "Marine Geology"



From Schneider and Thompson, Quat. Res., 12, 188-203 (1979)

Notice the benthic-planktonic offset in Fig 3-11

Even when foraminifera grow in the same H_2O, not all species fractionate O^{16}-O^{18} the same (Fig 17-13).

Other climate proxies vary in sync with these records over the Quaternary

Other Paleoclimate records

Temperature swings during glacial-interglacial cycles are also recorded in species assemblages in deep sea cores. Depth variations within a core at a given latitude indicate conditions in the overlying water column were changing .

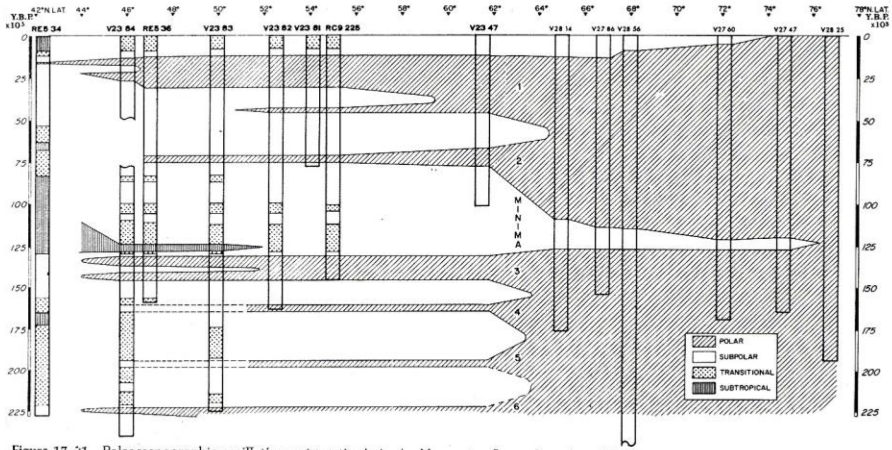


Figure 17-21 Paleoceanographic oscillations with latitude in the Norwegian Sea and northern North Atlantic during the late Quaternary. Plotted are planktonic foraminiferal-coccolith assemblages specific to particular water masses. Note that subpolar faunas have penetrated into Norwegian Sea only twice during the last 150,000 years: at present and about 120,000 years ago. (From T. B. Kellogg, in *Investigation of Late Quaternary Paleoceanography and Paleoclimatology*, ed. R. M. Cline and J. D. Hays, GSA Memoir 145, p. 105, 1976, courtesy The Geological Society of America)

From Kennett, "Marine Geology"

13

Paleoclimate records combined

All this information can be compiled to produce maps of estimated surface temperature variations on Earth between for instance the last glacial maximum and today, as in this example for the north Atlantic.

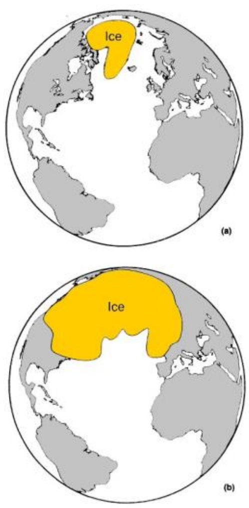


FIGURE 10.4
(a) Earth and its coverage of ice in the northern hemisphere today, compared with (b) its coverage 11,000 years ago. White areas are the coverage of ice.
(After Imbrie and Imbrie, 1986.)

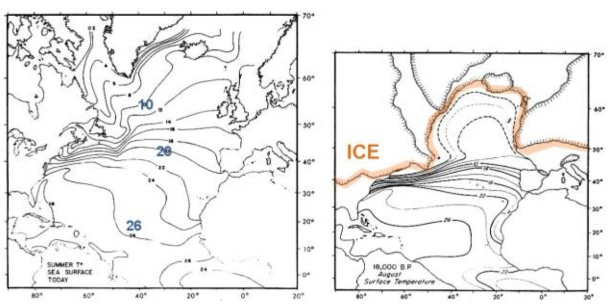


Figure 17-1 Surface-water isotherm maps for August today and August 18,000 years ago in the North Atlantic. (After A. McIntyre and others, 1976, courtesy The Geological Society of America)

From Kennett, "Marine Geology"

Notice that the temperature gradient from the poles to the equator was steeper in glacial times.

Cryosphere today, for comparison

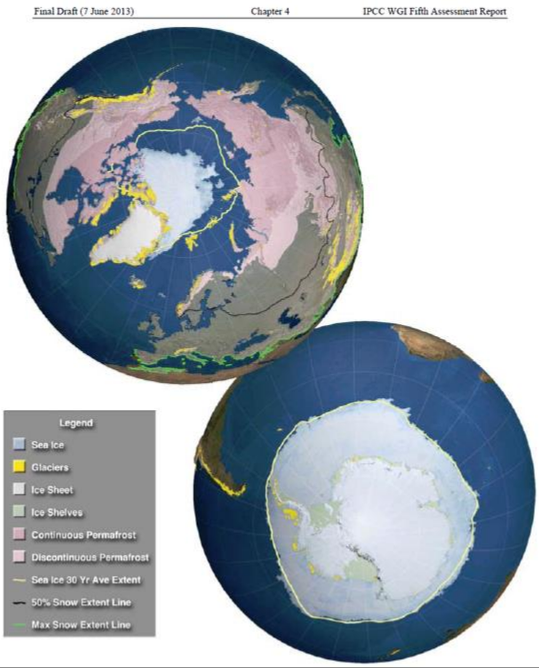


Figure 4.1: The cryosphere in the Northern and Southern Hemispheres in polar projection. The map of the Northern Hemisphere shows the sea ice cover during minimum summer extent (13th September 2012). The yellow line is the average location of the ice edge (15% ice concentration) for the yearly minima from 1979 to 2012. Areas of continuous permafrost (see Glossary) are shown in dark pink, discontinuous permafrost in light pink. The green line along the southern border of the map shows the maximum snow extent while the black line across North America, Europe and Asia shows the contour for the 50% snow extent. The Greenland ice sheet (blue/grey) and locations of glaciers (small golden dots) are also shown. The map of the Southern Hemisphere shows approximately the maximum sea ice cover during an austral winter (13th September 2012). The yellow line shows the average ice edge (15% ice concentration) during maximum extent of the sea ice cover from 1979 to 2012. Some of the elements (e.g., some glaciers and snow) located at low latitudes are not visible in this projection (see Figure 4.8). The source of the data for sea ice, permafrost, snow and ice sheet are datasets held at the National Snow and Ice Data Center (NSIDC), University of Colorado, on behalf of the North American Atlas, Instituto Nacional de Estadística, Geografía e Informática (Mexico), Natural Resources Canada, U.S. Geological Survey, Government of Canada, Canada Centre for Remote Sensing and The Atlas of Canada. Glacier locations were derived from the multiple datasets compiled in the Randolph Glacier Inventory (Arendt et al., 2012).

Past Climates on Earth

We have a reasonably detailed picture of past climates on Earth from:

Stable isotopic O and H records for the hydrosphere (marine and freshwater $CaCO_3$ fossils, ice caps only back to 200 ka), which provide high fidelity records of past temperature fluctuations and hydrosphere volume changes because isotopic fractionation is sensitive to such changes

Plus

a. the distribution and abundance in marine and terrestrial sedimentary records of:
- ✓ aquatic fossils
- ✓ pollens
- ✓ sediment types

b. Ice core records

c. the distribution of glacial deposits and landforms

These types of measures indicate paleo temperatures swings of 5-7° C or so between recent glacial and interglacial times on Earth.

Past Climates on Earth

Mean global temperature has varied remarkably little (~25°C or so) over most of Earth history but it has been enough to allow periods of ice accumulation at the poles and times where there was little or no ice there.

extensive glaciation

- The Pleistocene (10ka - 2 Ma)
- The Permian through Devonian periods (240 x 10^6 to 400 x 10^6 yrs ago)
- Parts of the early Proterozoic through mid Archaen (2100 x 10^6 to 2500x 10^6)

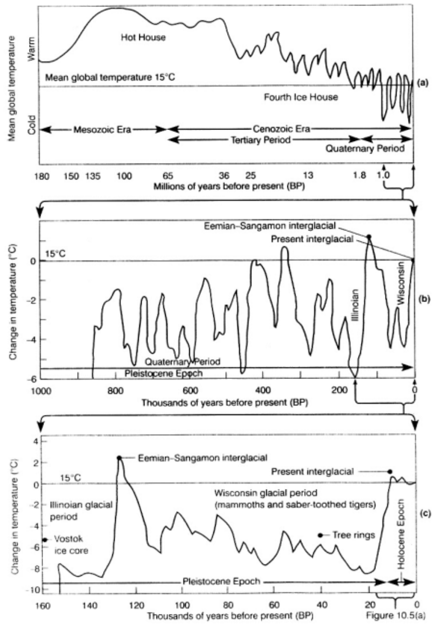
No Glaciation

- The Cretaceous (65 x 10^6 to 136 x 10^6).

The Earth has glaciers today so we are in a glacial epoch. However, within the Quaternary periods of more or less glaciation have occurred frequently.

Relatively warm periods like today are known as *interglacials*.

True *glacial* periods are much colder.



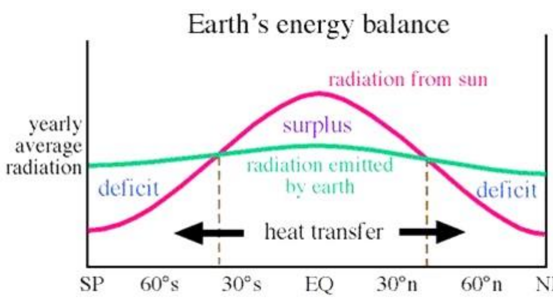
(a) is the temperature record of Earth during the past 180 million years; (b) is an expanded representation of the last one million years; and (c) is an expanded view of the last 160,000 years. (After UCAR/OIES, 1991a.)

What regulates Earth's climate system?

Before we ask why climate has changed, let's consider the many variables that affect it.

Solar radiation is one key parameter:

The angle of the suns rays on Earth and Earth's dynamic atmosphere-hydrosphere system play critical roles in determining the retention and geographical distribution of solar heat fluxes to Earth



(Recall the importance of oceanic and atmospheric circulation to flow within the hydrologic cycle)

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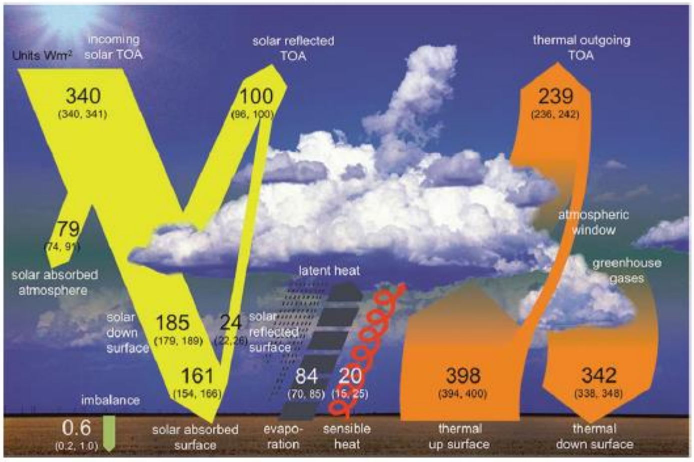


Figure 2.11: Global mean energy budget under present day climate conditions. Numbers state magnitudes of the individual energy fluxes in W/m², adjusted within their uncertainty ranges to close the energy budgets. Numbers in parentheses attached to the energy fluxes cover the range of values in line with observational constraints. Figure adapted from Wild et al. (2013).

Atmospheric gasses modulate the planet's temperature through the absorption of IR radiation that would otherwise be reflected to space from Earth's surface.

H_2O absorption bands dominate atmospheric IR transparency.

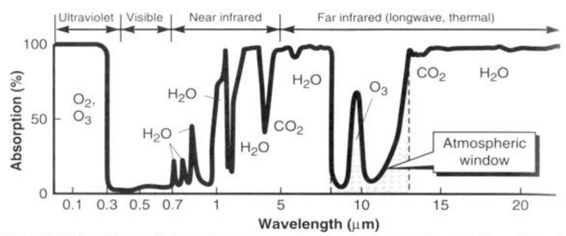


Figure 11.10 The efficiency of radiation absorption by the atmosphere as a function of wavelength through the solar and terrestrial regions of the spectrum (Appendix C). The absorption spectrum is expressed as the percentage of the sunlight incident at the top of the atmosphere that is absorbed by air before reaching the ground. An absorption of 100 percent means that no radiation at that wavelength can penetrate the atmosphere, because of absorption by molecules in the atmosphere. Regions of high absorption associated with specific molecules are identified on the plot. The major spectral ranges are shown: The region from about 8 to 13 micrometers is referred to as the longwave *atmospheric window*. (Data from R.M. Goody, *Atmospheric Radiation: I. Theoretical Basis*, London: Oxford University Press [1964], adapted from data originally presented in Figure 1.1, p. 4)

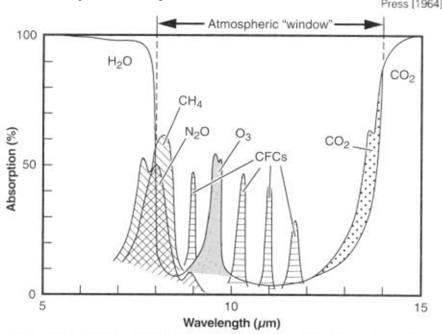


Figure 11.12 The absorption spectrum of the atmosphere in the "atmospheric window" region. The principal absorbing molecules and the relative sizes of their absorption bands are illustrated.

Other gasses play a secondary role:

- ➢CO_2
- ➢followed by CH_4, O_3, N_2O and CFC's.

abundances of all but CFCs have varied for natural reasons over Earth history (CFCs are a human invention)

Climate Variations on Earth

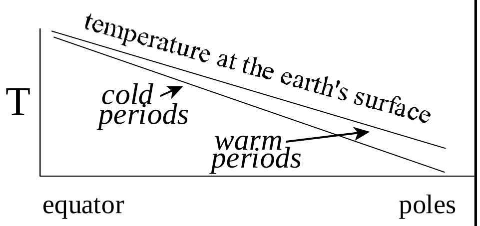
Global temperature swings also reflect a change in Earthly heat budget from:
☼ the amount of incident sunlight to the Earth

☼ the amounts of light reflected to space (both albedo and IR via heat trapping of heat by the hydrosphere/atmosphere system, which is related to atmospheric greenhouse gas composition).

external controls on climate:
☼ solar intensity
☼ Earth's orbital fluctuations
☼ meteor/comet impacts

internal controls on climate:
☼ global tectonism
☼ volcanism
☼ the biosphere

Regardless of the specific controls, we can think of past global climate variations to a large extent as variations in the temperature *gradient* between the poles and Equator; cold and warm periods have their greatest effects upon polar climate and little effect on equatorial climate.



Climate variations and controls

What are the chief controls on climate? As we've seen, the Quaternary record shows a considerable amount of ***cyclicity*** in the 10^3-10^5 year range.

None of the internal controls are cyclic on time scales that can affect climate over periods of 10^3-10^6 years.
- The *Wilson Cycle* of plate tectonics is a few hundred million years in length
- other tectonic cycles are in the >10 Myr range.
- *Volcanism* normally affects climate for just a few years at a time.
  - *Flood basalt events* can have dramatic climatic effects for much longer periods, but are rare
  - none have occurred in the Quaternary.

Among external factors, large meteorite and comet impacts are also rare, and not cyclic. Fluctuations in solar intensity no doubt occur on long time scales, but there's little evidence for significant variations in the Quaternary.

The amount of sunlight the earth receives as a function of changes in its orbit around the sun varies cyclically on the right time scale.

The primary cause of geologically recent (Pleistocene to present) global T fluctuations

Earth's orbital parameter's about the sun vary regularly and on the right time scale to explain most observed phenomena.

James Croll first suggested that solar irradiance variations affected global climate in 1870. It was not until 1930 however that the variability in Earth's orbital parameters were quantified by Milankovich. For this reason, the theory that orbital cyclicity dominates global climate is known as Milankovich theory. The basics are given in the figure at right.

Combining all of the predicted solar insolation effects arising from the various orbital fluctuations gives a curve of relative insolation into the past that matches numerous proxy records.

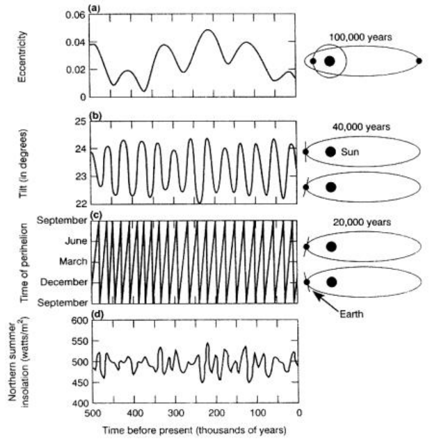


FIGURE 10.6
The Milankovitch theory of climatic change during the Pleistocene. The onset of ice ages is due to variations in three orbital parameters of Earth. (a) The eccentricity is the degree to which Earth's orbit departs from a circle. Times of maximum eccentricity are separated by roughly 100,000 years. (b) The tilt angle is the angle between Earth's axis and a line perpendicular to the plane of the orbit of the planet. (c) The time of perihelion involves the tilt of Earth's axis at its closest approach to the sun. The cycles of tilt and time of perihelion are roughly 40,000 and 20,000 years, respectively. (d) The calculated amount of sunlight received at 60° to 70° north latitude during the summer (summer insolation, July), based on the cycles of variation of Earth's orbital parameters. One watt = 0.0569 British thermal units (Btu) per minute = 14.28 calories per minute.
(After Covey, 1984.)

This periodicity shows up in various climate proxy records

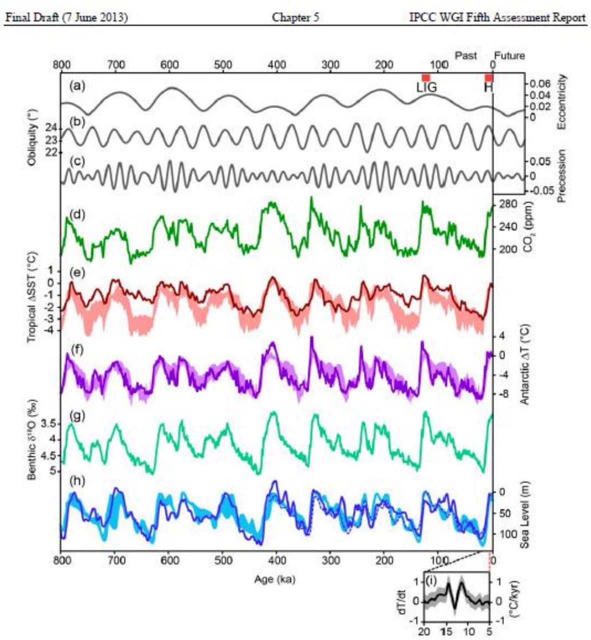


Figure 5.3: Orbital parameters and proxy records over the past 800 kyr. (a) Eccentricity, (b) obliquity, (c) precessional parameter (Berger and Loutre, 1991), (d) atmospheric concentration of CO2 from Antarctic ice cores (Petit et al., 1999; Siegenthaler et al., 2005; Ahn and Brook, 2008; Lüthi et al., 2008), (e) tropical SST stack (Herbert et al., 2010), (f) Antarctic temperature stack based on up to seven different ice cores (Petit et al., 1999; Blunier and Brook, 2001; Watanabe et al., 2003; EPICA Community Members, 2006; Jouzel et al., 2007; Stenni et al., 2011), (g) stack of benthic $\delta^{18}O$, a proxy for global ice volume and deep-ocean temperature (Lisiecki and Raymo, 2005), (h) reconstructed sea level (dashed line: Rohling et al., 2010; solid line: Elderfield et al., 2012). Lines represent orbital forcing and proxy records, shaded areas represent the range of simulations with climate (GENIE-1, Holden et al., 2010a; Bern3D, Ritz et al., 2011), climate-ice sheet models of intermediate complexity (CLIMBER-2, Ganopolski and Calov, 2011) and an icesheet model (IcIES, Abe-Ouchi et al., 2007) forced by variations of the orbital parameters and the atmospheric concentrations of the major greenhouse gases. (i) Rate of changes of global mean temperature during Termination I based on Shakun et al. (2012).

Contents of greenhouse gasses CO_2 and CH_4 "follow" the pattern of orbital forcing variations so gasses probably rise and fall primarily because of externally forced climate fluctuations, rather than vice-versa.

As gas concentrations rise, they enhance the magnitude of temperature increase due to insolation alone and as they fall, they enhance the magnitude of temperature decrease.

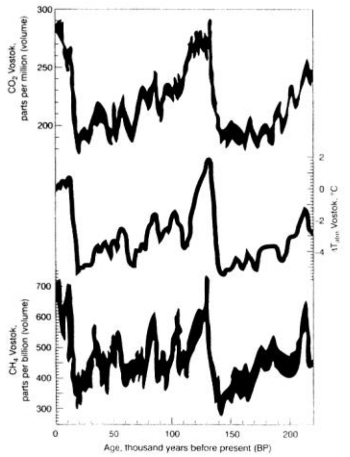


FIGURE 10.7
The trend of atmospheric CO_2, CH_4, and temperature as recorded in the Vostok, Antarctica, ice core. The atmospheric temperature is plotted as a deviation from present-day mean temperature (ΔT_{atm})
(After Jouzel et al., 1993)

These records "wiggle" in the same places but magnitudes of temperature changes are not always the same.

This is because many forces work together to "set" Earth's thermostat and the feedbacks between systems are affected in ways that alter the magnitudes of their effects.

Other internal factors such as sudden changes in volcanic or biogenic input of S to the atmosphere can also alter the system.

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Fluxes in the arctic, which is a particularly sensitive part of the system

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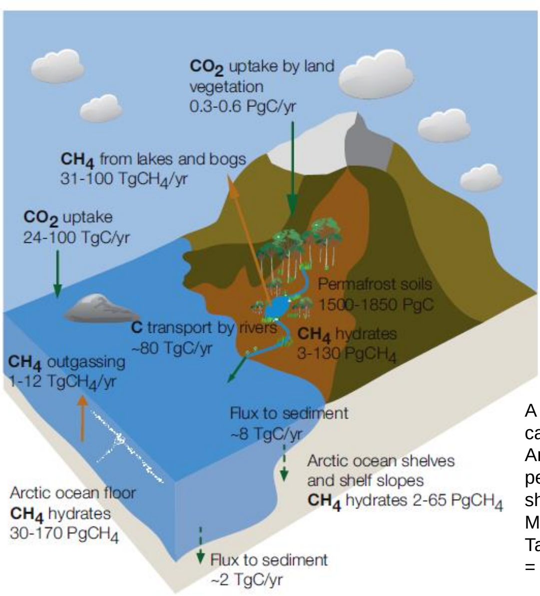

Can you imagine how these would be different during glacial times?

A simplified graph of current major carbon pools and flows in the Arctic domain, including permafrost on land, continental shelves and ocean (adapted from McGuire et al. (2009) and Tarnocai et al. (2009)). TgC $= 10^{12}$ gC, and PgC $= 10^{15}$ gC.

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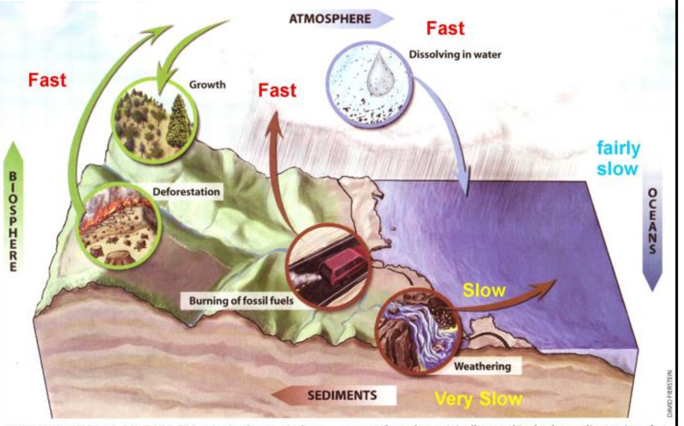
Temporal aspects of these phenomena (some of which are shown in the figure):

Long term controls:

☼ CO_2 consumption and release via silicate rock weathering: Variable proportions of CO_2 in the atmosphere and in the hydrosphere + carbonate rocks over geological time has resulted during Earth's history from changes in plate tectonic parameters.

Shorter term controls:

☼ amount of photosynthesis and respiration
☼ oceanic circulation.
☼ Variations in global volcanism (for both green house gasses and in albedo from stratospheric particulates).

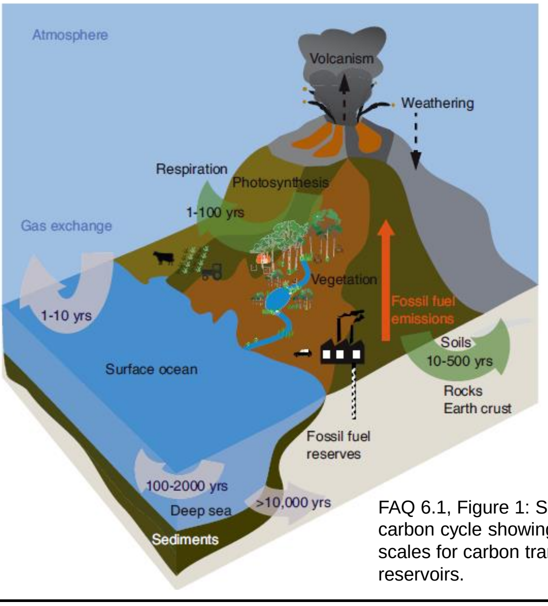


NATURAL STORES OF CARBON exist in the atmosphere, oceans, sediments and biosphere; exchange between these reservoirs occurs in a variety of ways. When humans burn fossil fuels, we transfer carbon originally stored in the deep sediments into the atmosphere. The goal of carbon sequestration is to redirect carbon from the atmosphere into one of the other three reservoirs.

GG325 L26, F2013

Another look at Carbon Turnover Timescales

Final Draft (7 June 2013) Chapter 6 IPCC WGI Fifth Assessment Report



FAQ 6.1, Figure 1: Simplified schematic of the global carbon cycle showing the typical turnover time scales for carbon transfers through the major reservoirs.

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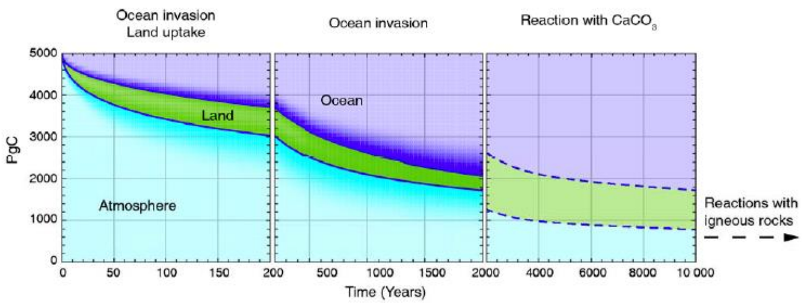
Final Draft (7 June 2013) Chapter 6 IPCC WGI Fifth Assessment Report



FAQ 6.1, Figure 2: Decay of a CO_2 excess amount of 5000 PgC emitted at time zero into the atmosphere, and its subsequent redistribution into land and ocean as a function of time, computed by coupled carbon–cycle climate models. The size of the colour bands indicate the carbon uptake by the respective reservoir. The first two panels show the multi-model mean from a model intercomparison project (Joos et al., 2013). The last panel shows the longer term redistribution including ocean dissolution of carbonaceous sediments as computed with an Earth System Model of Intermediate Complexity (after Archer et al., 2009b).

The Marine Carbon Cycle

Remember, the marine biosphere and oceanic currents play a major role in the rates of carbon cycling. MOST inorganic and organic carbon is not associated with living organisms in the ocean (it is POC, PIC, DIC, & DOC)

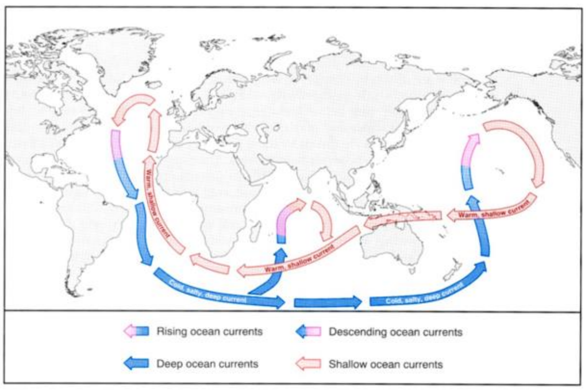


FIGURE 3.7
The conveyor belt circulation pattern of the world's oceans. Cold, salty water in the North Atlantic sinks to the deep ocean and moves southward to resurface and be warmed in the Indian and Pacific oceans. Surface currents then return the water to the Atlantic. A complete passage takes about one thousand years. Currently, this conveyor belt circulation pattern is driven to some extent by an imbalance between the loss of water from the Atlantic by evaporation and its gain by precipitation and continental runoff.
(After Dickinsen and Monastersky, 1991.)

Recall also from last week that.. Ocean basin differences in carbon utilization reflect the dominant current patterns and their effects on resident flora and fauna.

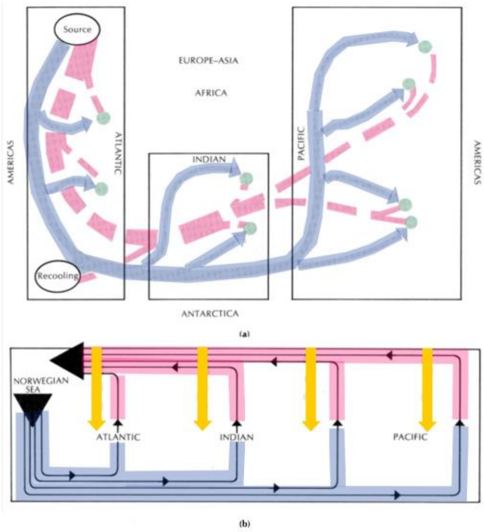


Figure 1-6 (a) is an idealized map of the patterns of deep water flow (solid lines) and surface water flow (dashed lines). The large circles designate the sinking of NADW in the Norwegian Sea and the recooling of water along the perimeter of the Antarctic Continent; the shaded circles indicate the distributed upwelling which balances this deep water generation. (b) is an idealized vertical section running from the North Atlantic to the North Pacific showing the major advective flow pattern (solid lines) and the rain of particles (wavy lines). The combination of these two cycles leads to the observed distribution of nutrients.

The oceans are an important “sink” for CO_2 because of the high solubility of CO_2 in seawater.

The oceans hold about **65x as much CO_2 as the atmosphere**.

The ocean’s role in the CH_4 budget and budgets of other biological gases *is more complex but still predictable*.

The slowness of the thermohaline circulation (in the present-day oceanic configuration one “cycle” takes about 1500-2000 years) means that the ***great bulk of ocean water is isolated from the atmosphere for centuries***, and thus cannot exchange gases with the atmosphere over shorter times.

One way to think of the forcing mechanisms in the recent geological past is that

the warm "interglacial" periods represent "normal" times and

their high CO_2 and CH_4 reflect "fully-functioning" biogeochemical exchanges.

When insolation goes down,
the planet cools,
starts to grow ice caps,
lowers sea level,
changes the carbon cycle,
and pulls CO_2 and CH_4 from the atmosphere.

The planet then grows ice caps.

Orbital fluctuations then progress and the system relaxes to an interglacial.

a detailed summary of the possible chain of events that occurs as Earth enters a glacial stage appears on the next slide. It includes many of the feedbacks we have discussed in class and a few we haven't.

TABLE 10.2
Possible series of events leading to development of a glacial stage in which changes in orbital parameters initiate the sequence of events.

(1. Start with the relatively low atmospheric carbon dioxide and temperature of this Ice House, which began about 35 million years ago.)

2. Milankovitch forcing leads to less insolation at northern hemisphere intermediate latitudes, giving rise to relatively warm winters and cool summers. Snow accumulates and is converted to ice in continental glaciers. There is a slight overall cooling of the Earth's surface due to a change in planetary albedo.

3. Sea level falls. There is more nutrient input to the ocean from erosion of fresh, organic-rich, continental margin sediments. Vertical mixing of the ocean is enhanced, leading to more rapid cycling of nutrients. Greater surface biological productivity draws down atmospheric carbon dioxide. Positive feedback leads to more cooling.

?

4. Oceanic conveyor belt slows and less heat is released by upwelling in the northern hemisphere. The northern hemisphere is further cooled. The conveyor belt circulation can turn on and off on short time scales, perhaps the decadel to century time scale.

5. Continental glaciers advance over North America and Europe, reaching thicknesses of 3 kilometers. Sea level falls about 120 meters.

6. Glaciers cover the tundra and higher latitude land ecosystems, and the flux of methane from boreal zones and tropical wetlands and that of nitrous oxide from terrestrial ecosystems decrease. This results in lower atmospheric methane and nitrous oxide concentrations; temperature falls—a positive feedback.

7. Stronger winds of the glacial stage promote more atmospheric dust transport and deposition and may be a source of iron for increasing glacial oceanic productivity.

8. The enhanced organic production of the glacial stage gives rise to higher dimethylsulfide emissions from the oceans and enhanced deposition of non-sea salt sulfate. This may be a positive feedback to cooling because of the link between emissions of dimethylsulfide, cloud condensation nuclei, sulfate aerosol, and cooling (pp. 286–287).

9. Orbital forcing relaxes, and the planet rapidly enters an interglacial stage.