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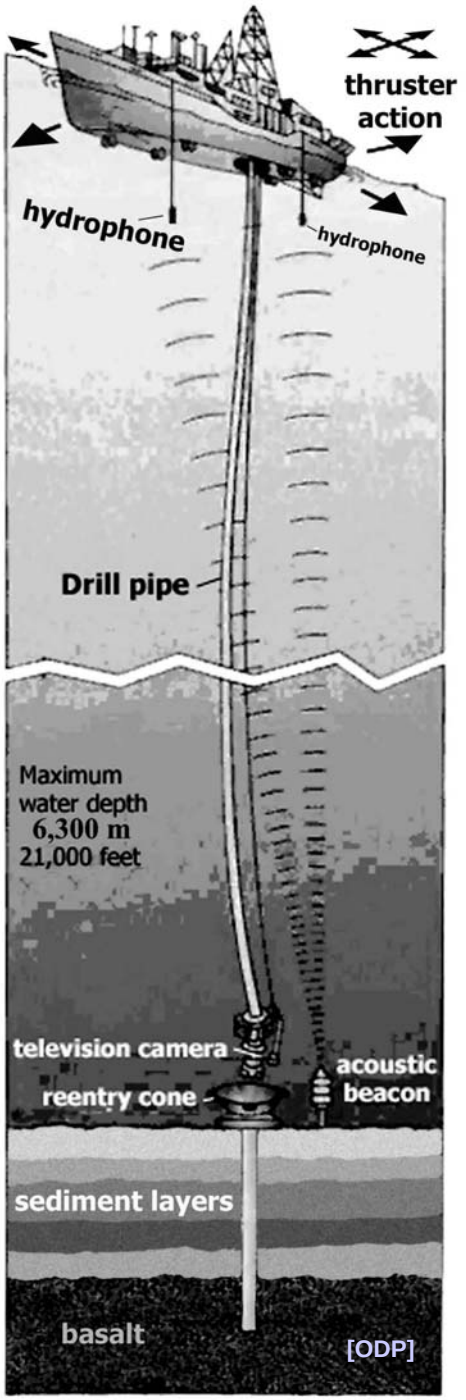
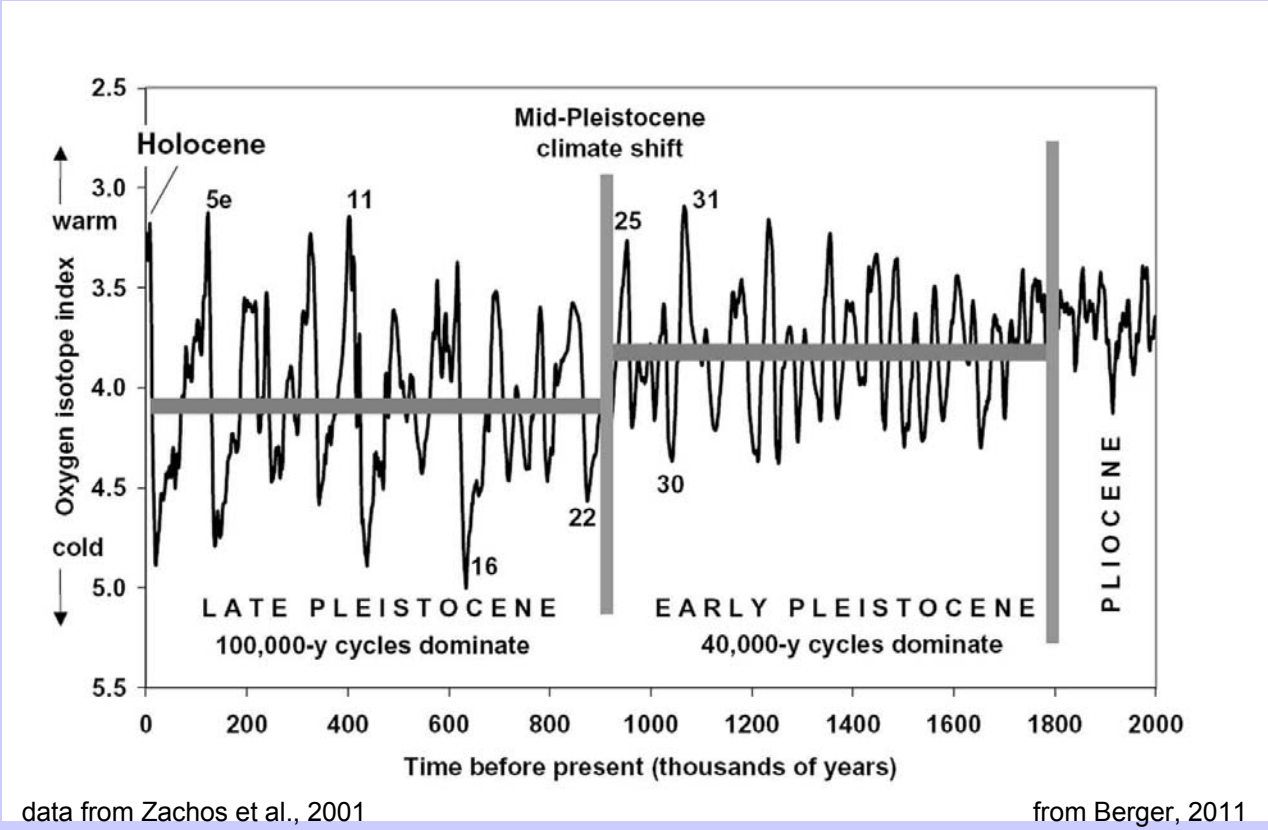


CLIMATE CHANGE IN THE CENOZOIC AS SEEN IN DEEP SEA SEDIMENTS

AMSTERDAM LECTURES MARCH 2011

EXCERPTS



data from Zachos et al., 2001

from Berger, 2011

Wolf Berger, Ph.D.,
Scripps Institution of Oceanography
University of California, San Diego

OUTLINE

1 QUATERNARY

2 LESSONS FOR GLOBAL WARMING

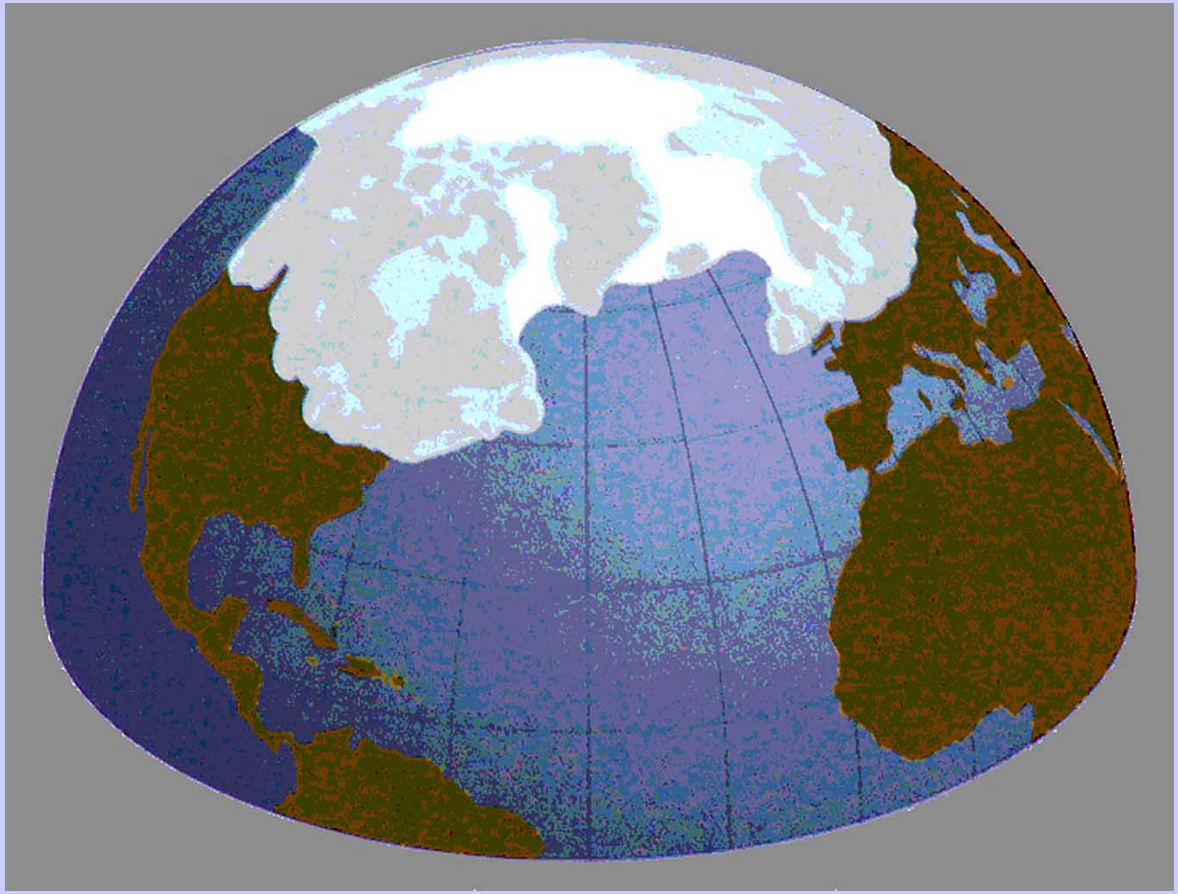
3 NEOGENE

4 PALEOGENE

5 REFERENCES

1 QUATERNARY

It is the waxing and waning of northern ice sheets that are the chief manifestation of climate change in the Quaternary.

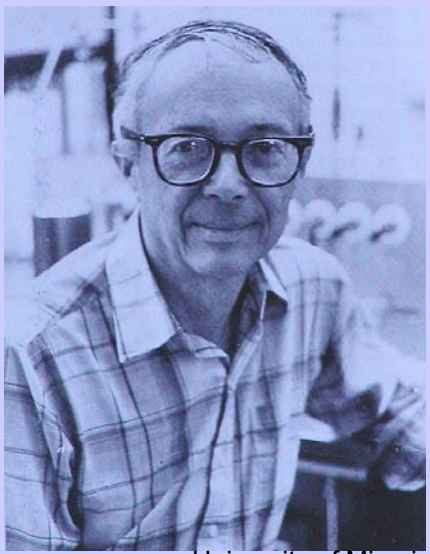


After Imbrie and Imbrie, 1979.

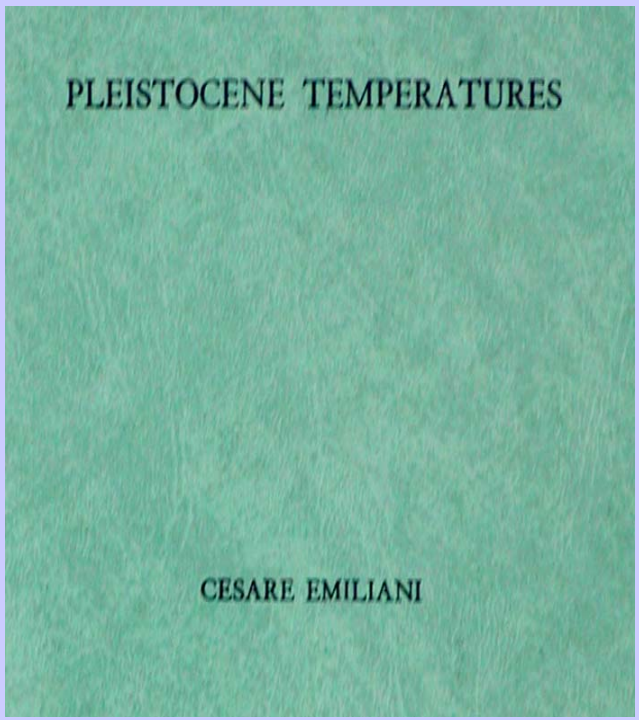
The buildup and decay of large ice masses takes time. Thus, there are certain limits on the rate at which the overall climate can respond to forcing when ice masses are at stake (rather than winds or currents).

The first person to systematically investigate the ocean's memory system on the sea floor for its record on ice-age fluctuations was Cesare Emiliani (1922-1995), an Italian-American geologist, with training in both paleontology (Italy) and isotopic chemistry (Chicago).

Emiliani obtained samples from Hans Pettersson (1888-1966), leader of the Swedish Albatross Expedition (1947-1948). He analyzed foraminifers for oxygen isotopes and wrote the classic paper "Pleistocene temperatures" (1955), one of the first important articles (and a founding feat) in the branch of science called "paleoceanography." (Other important pioneers were Wolfgang Schott, Gustaf Arrhenius, Fred Phleger, Frances Parker, and David Ericson.)

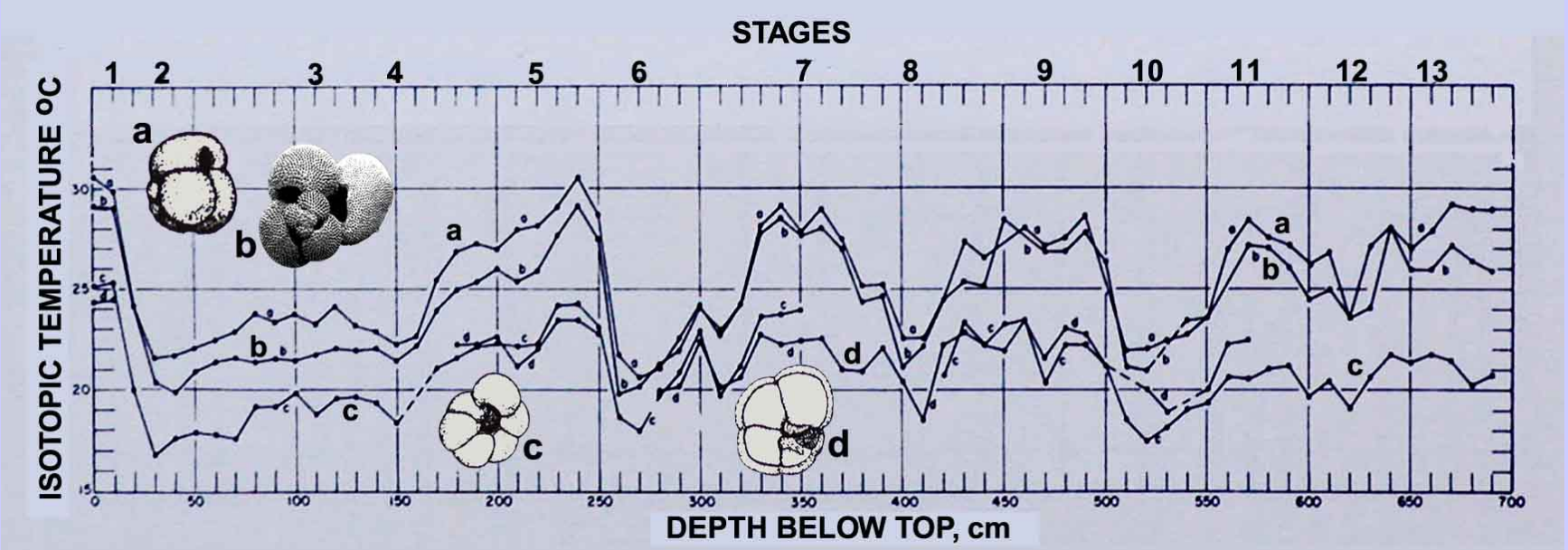


University of Miami



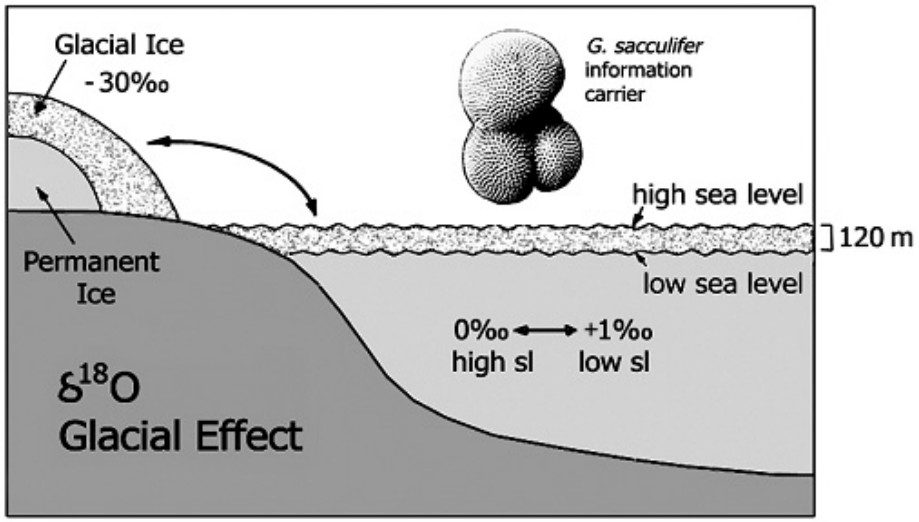
In the paper, he showed that the cycles found in deep-sea sediments are of the correct sign and length to qualify as records of the ice ages.

Emiliani also used cores from Lamont, obtaining samples from David Ericson. Emiliani analyzed a number of species of foraminifers, establishing that they give different results, presumably owing to depth stratification. In the process he showed that the Penck-Brueckner scheme of multiple ice ages (adopted by Ericson) was probably incorrect.

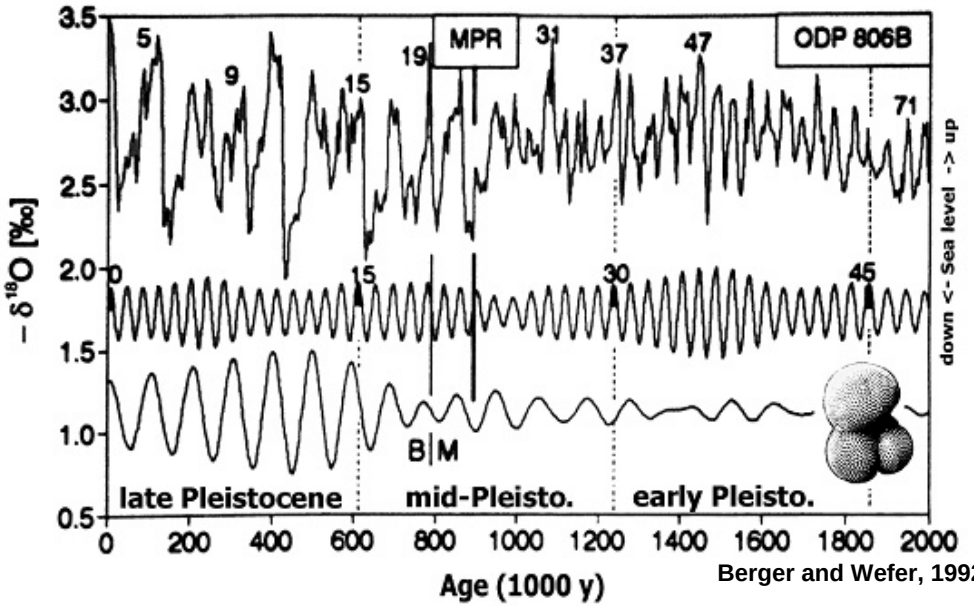


after C. Emiliani, 1955

Analysis of the record to get variations in sealevel rise begins with the isotopic analysis of foraminifers. It is used to determine the mass of polar ice extant.



Seibold and Berger, 1996

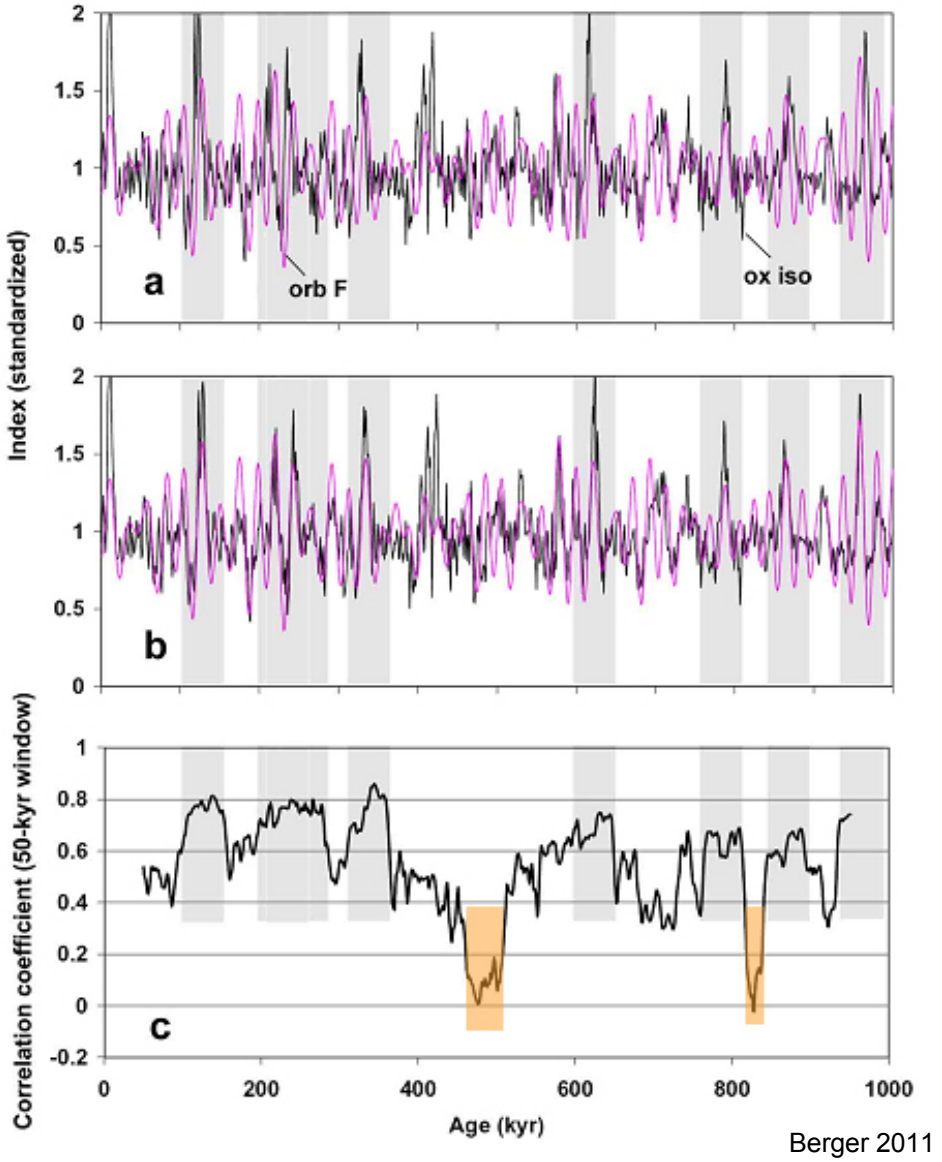


Berger and Wefer, 1992

The principle is shown at left (upper graph): water evaporates from the sea and ends up in glacial ice. It is enriched in the isotope O-16. Correspondingly, the ocean is depleted in that isotope. The change is reflected in the isotope index values (-30 and +1 permil, respectively).

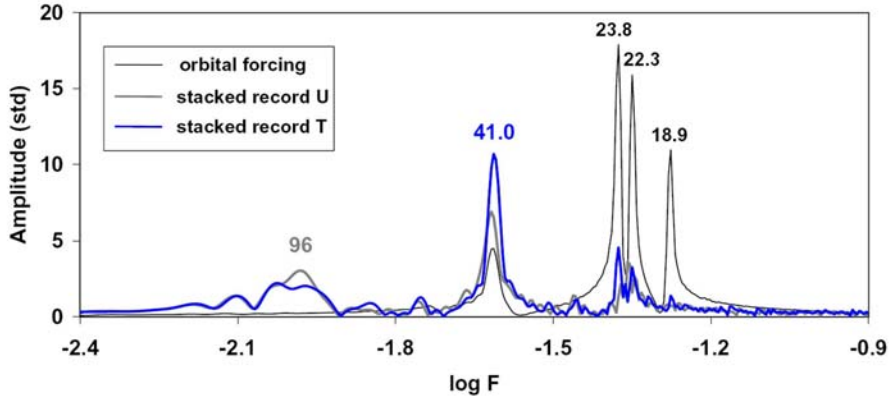
The record in deep-sea sediments (here: ODP 806, western equatorial Pacific) yields changes in isotope values (y-axis) dominated by ice-mass changes. These show cycles at 41,000 y throughout the Quaternary, and at 100,000 y for the last 650,000 years, in the manner indicated.

MPR, mid-Pleistocene Revolution; B/M, Brunhes- Matuyama boundary (790 ky).



Berger 2011

Data Zachos et al. 2001, Lisiecki & Raymo, 2005



forcing: A,Berger; U, untuned; T, tuned record

Precise age assignment relies on the reliability of Milankovitch forcing, and a match of the forcing to the first derivative of the oxygen isotope record.

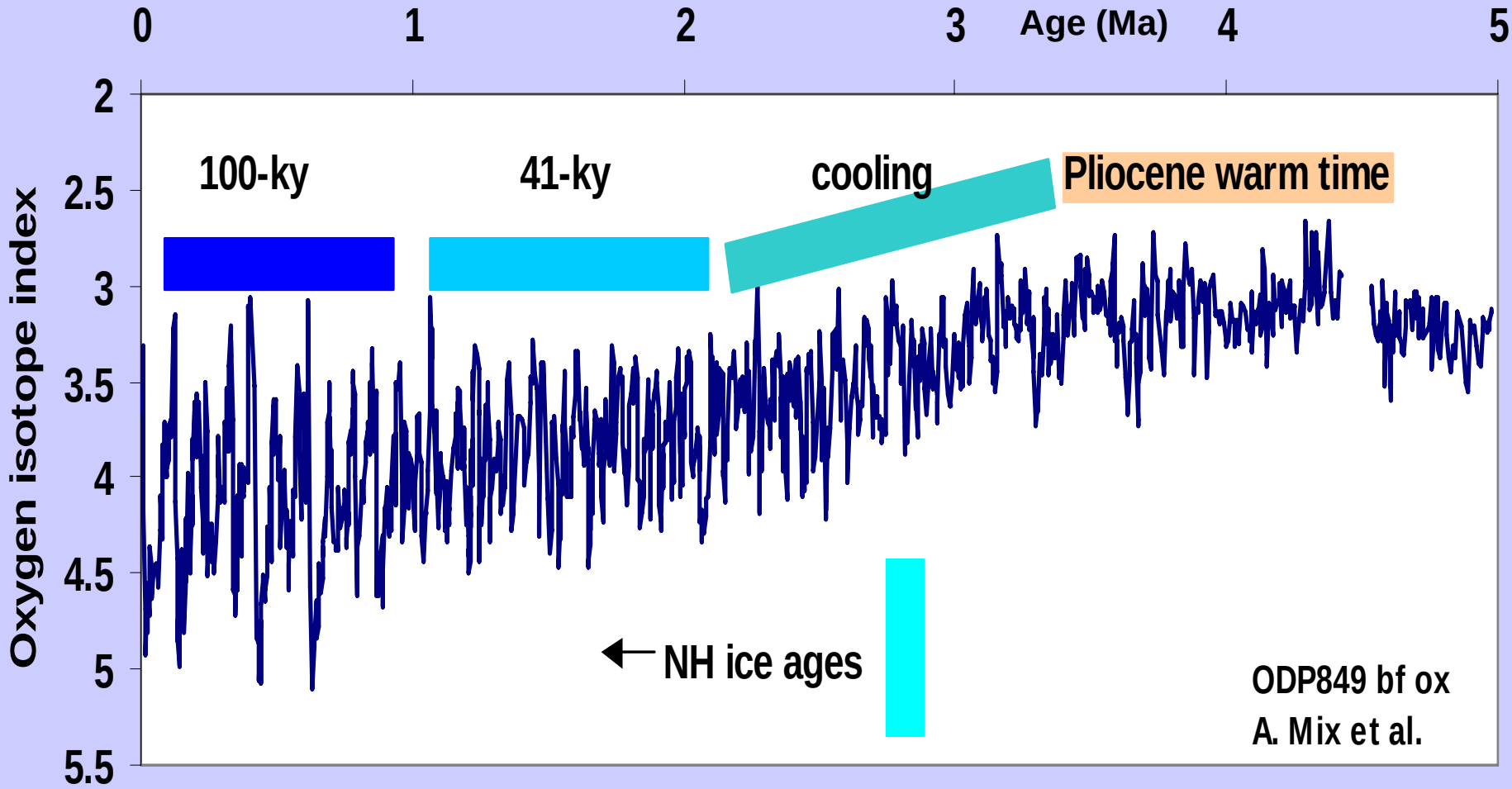
(a) before tuning

(b) after tuning (spectra to right)

(c) corr. coeff. forcing-response, 50-kyr window

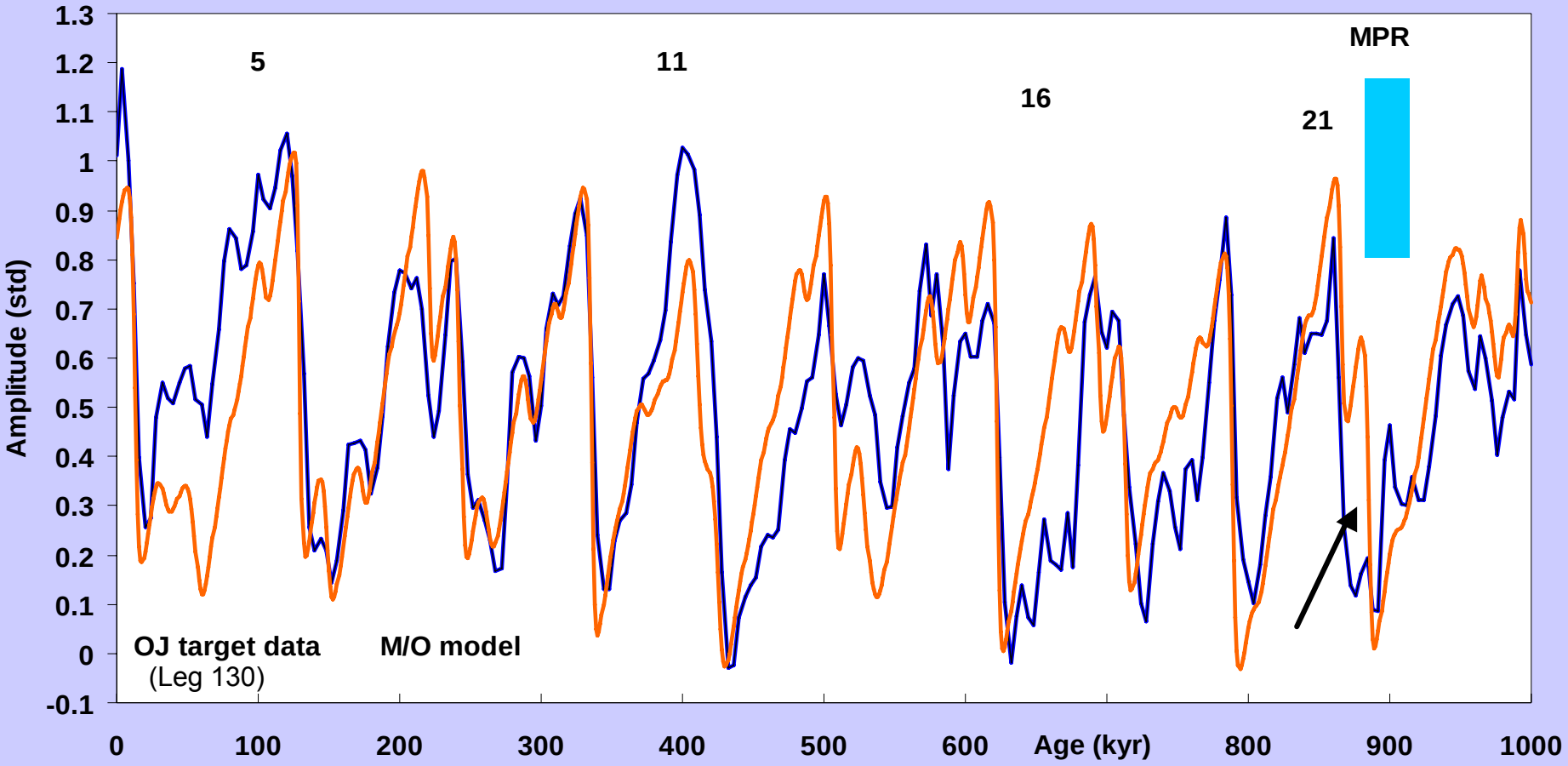
note lack of response near 480 and near 840 kyr: deaf zones (orange)

The reason for the change from a dominant obliquity-driven cycle to a dominant cycle near 100 kyr at 0.9 Ma is not clear. That the transition has multiples of 41 kyr (rather than 100 kyr cycles) is probably significant.



The Mid-Pleistocene Revolution is a natural consequence of an overall cooling trend, with larger ice masses generating larger climatic variation in BOTH directions.

Strictly speaking, 100-kyr cycles start at Stage 16 (650 kyr ago). Does this mean the system changed its mode of operation at the time? Apparently not: the same "rules" of modeling apply right back to the Mid-Pleistocene Revolution (MPR).



Rules have not changed since the MPR: The onset of 100-kyr cycles (at 650 ka) does NOT indicate a change in rules at that time.

The model uses Milankovitch forcing (A. Berger and Loutre, 1991) for input and has an inbuilt oscillation slightly larger than 100 kyr: Berger et al. 1996. A notable discrepancy in the placement of a termation appears near 0.9 Ma (arrow in graph).

2 LESSONS

ARE THERE LESSONS ABOUT CLIMATE CHANGE THAT EMERGE FROM THE STUDY OF THE ICE AGES AND THE CENOZOIC IN GENERAL, LESSONS THAT ARE APPLICABLE TO THE PREDICAMENT OF ONGOING AND FUTURE GLOBAL WARMING?

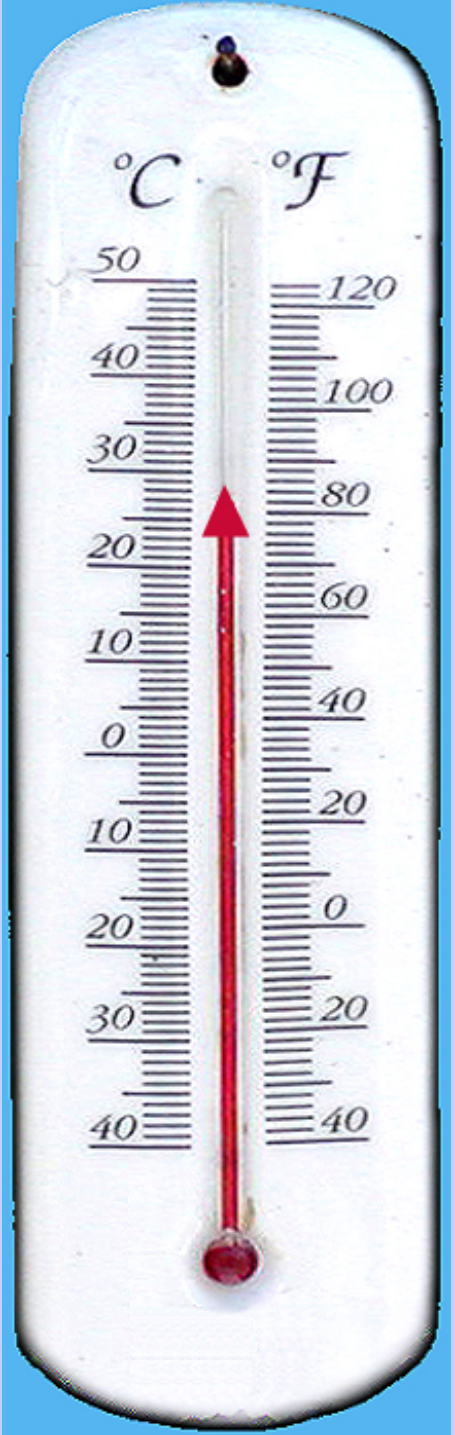
YES, WITHIN LIMITS.

THE LIMITS HAVE TO DO WITH THE FACT THAT CENOZOIC CLIMATE CHANGE IS STUDIED ON TIME SCALES OF THOUSANDS OF YEARS, WHEREAS TIME SCALES OF DECADES ARE RELEVANT TO THE IMPACT OF CLIMATE CHANGE ON HUMAN WELL-BEING.

General rules that transcend time scales about the ocean's heat budget and about changes in sea level are of special interest.

Other items are the uptake of carbon dioxide by the ocean, changes in the productivity of upwelling regions, and predictability of the climate.

Great potential for damage is associated with vast deposits of methane ice, but the likely response of methane clathrates to global warming is entirely obscure, and hardly elucidated by the ocean's record.



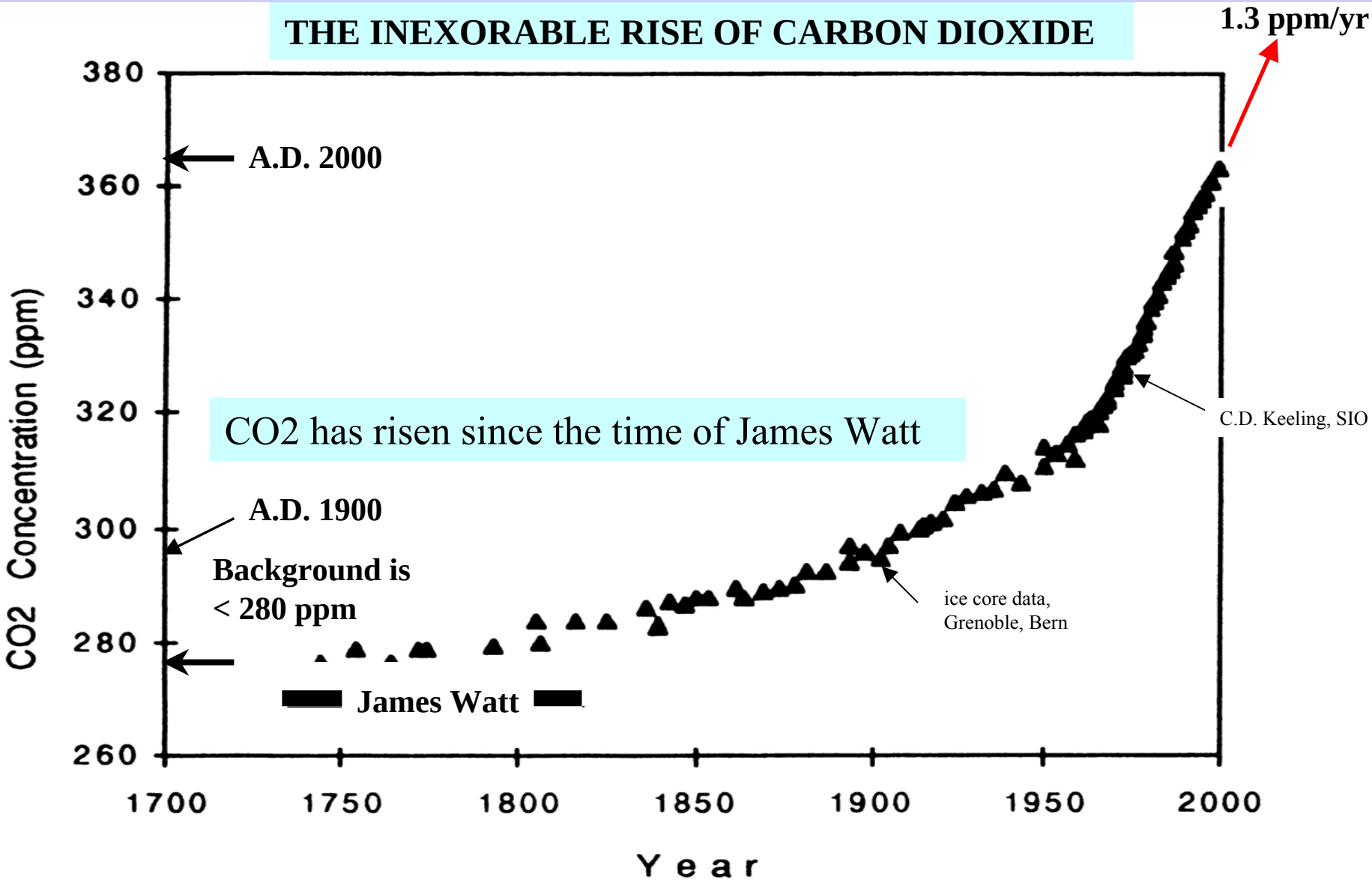
First of all: Global Warming is for real. It is getting warmer.

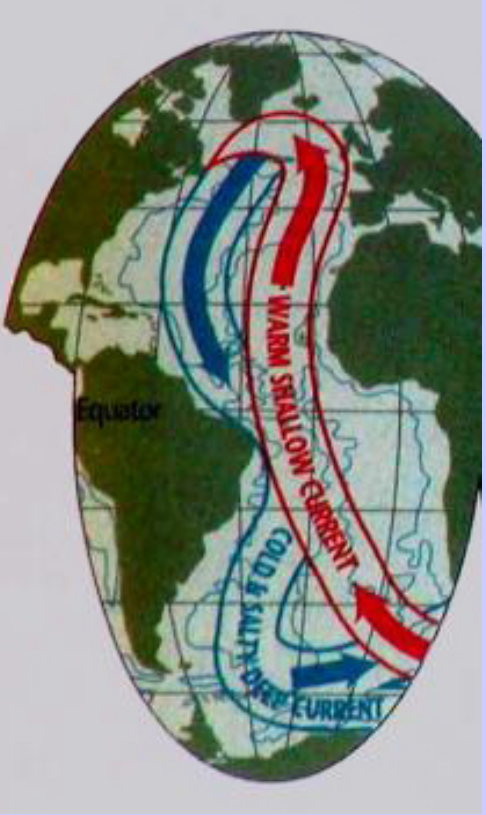
Meteorologists tell us that 2010 was one of the warmest years on record (perhaps the warmest), when considering average temperatures for the entire planet.

In fact, record-setting years occurred all through the past decade, during a time of modest to low solar activity.

Practically all working climate scientists agree that measurable planetary warming has resulted from the release of greenhouse gases to the atmosphere, notably carbon dioxide.

- **We know the cause of much of the warming:**





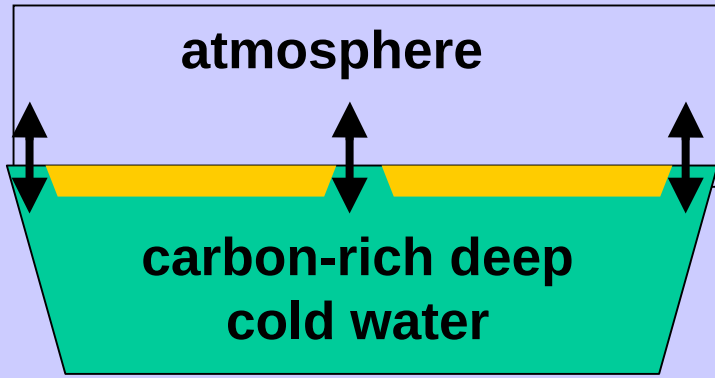
HOW WILL THE OCEAN RESPOND? WE DO NOT KNOW. BUT THESE ARE CRUCIAL TOPICS IN RELEVANT RESEARCH:

•PHYSICS

The ocean redistributes heat. How will this change?

•CHEMISTRY

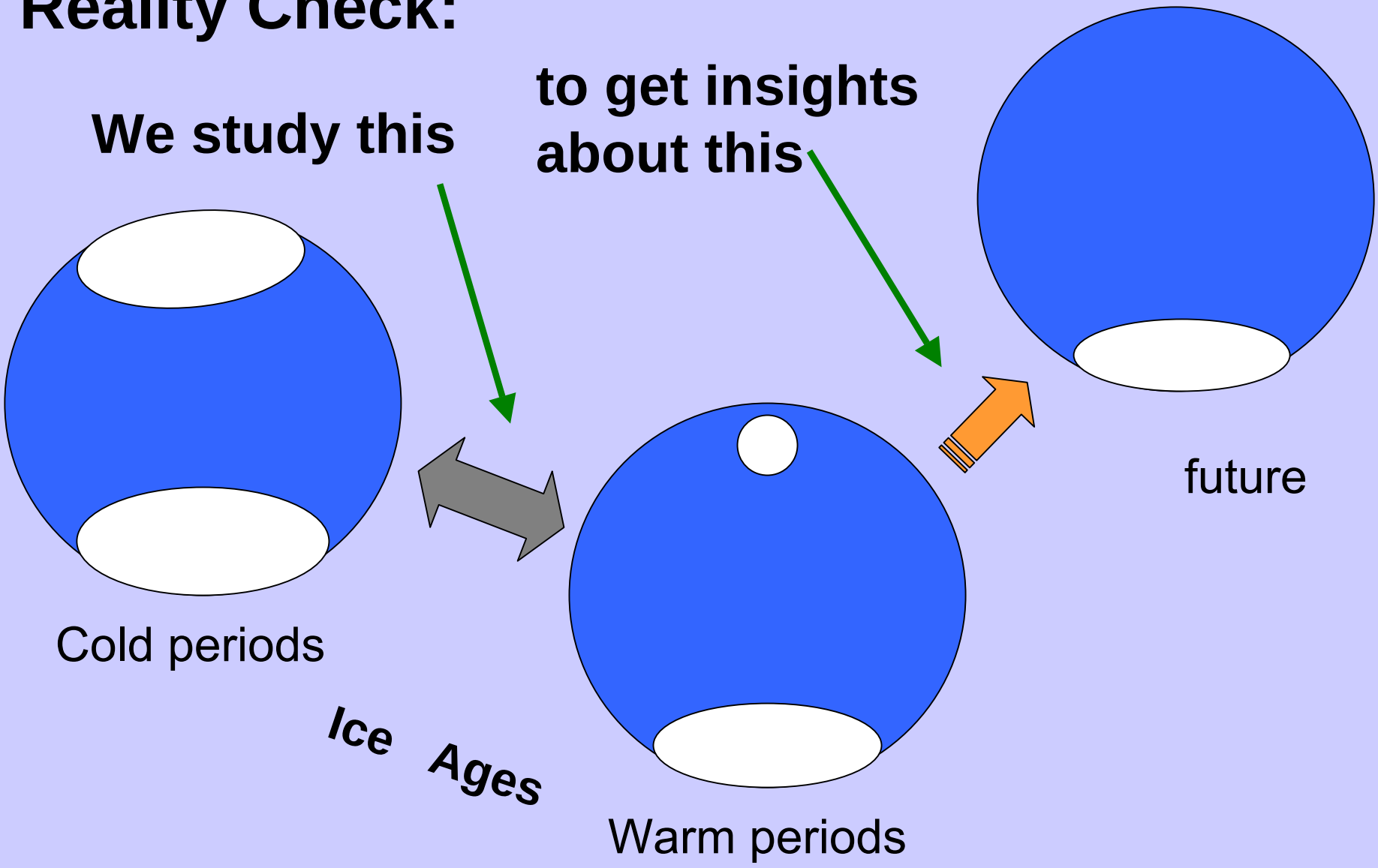
The ocean takes up excess carbon from the air. How will this change?



•BIOLOGY

The ocean produces food for people. How will this change?

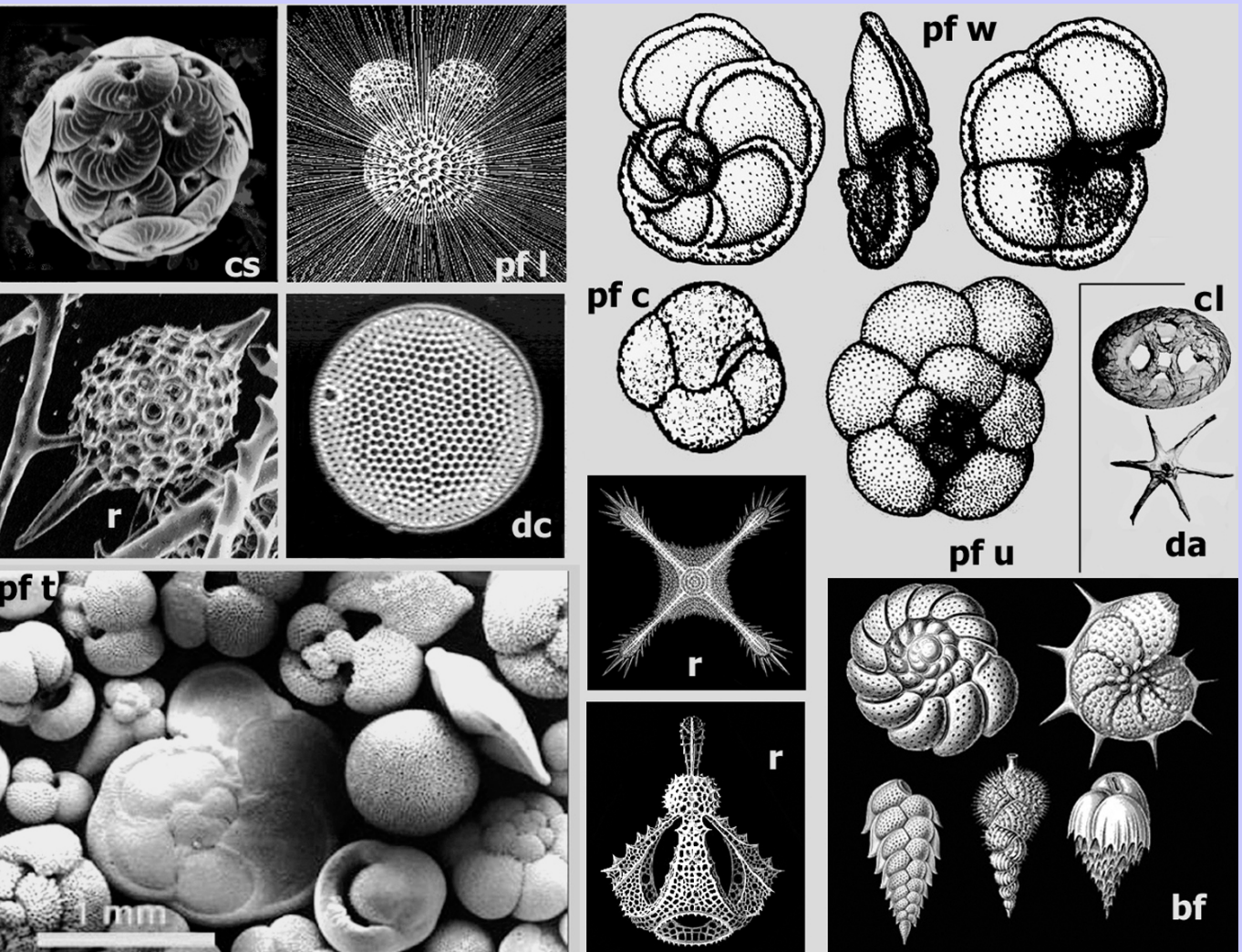
Reality Check:



The future has no analogs in the past.

In addition to the oxygen isotope record, there is ocean history information in the abundance distributions of microfossils and nannofossils and also (as discovered more recently) in that of various organic compounds.

Reality check: fossils are not ideal recording devices, but are biased reporters telling us only what they care about.



cs, coccosphere; pf l, planktonic foraminifer live; r, radiolarian; dc, diatom centric; pf t, planktonic foraminifers thanatocoenosis (shell assemblage); w, warm; c, cold; u, upwelling; cl, coccolith; da, discoaster; bf, benthic foraminifer.

Microfossils: coccoliths, foraminifers, radiolarians, diatoms.

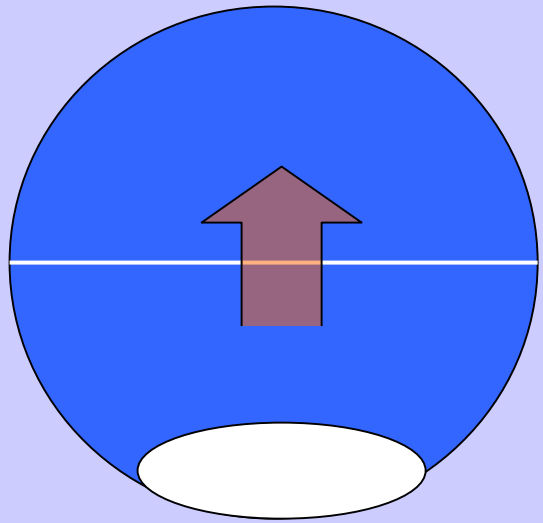
Sources: S.I.O., M. Yasuda, Challenger Reports, E. Haeckel, F.L. Parker, M.N. Bramlette, D. Bukry.

How will the ocean respond regarding heat distribution?

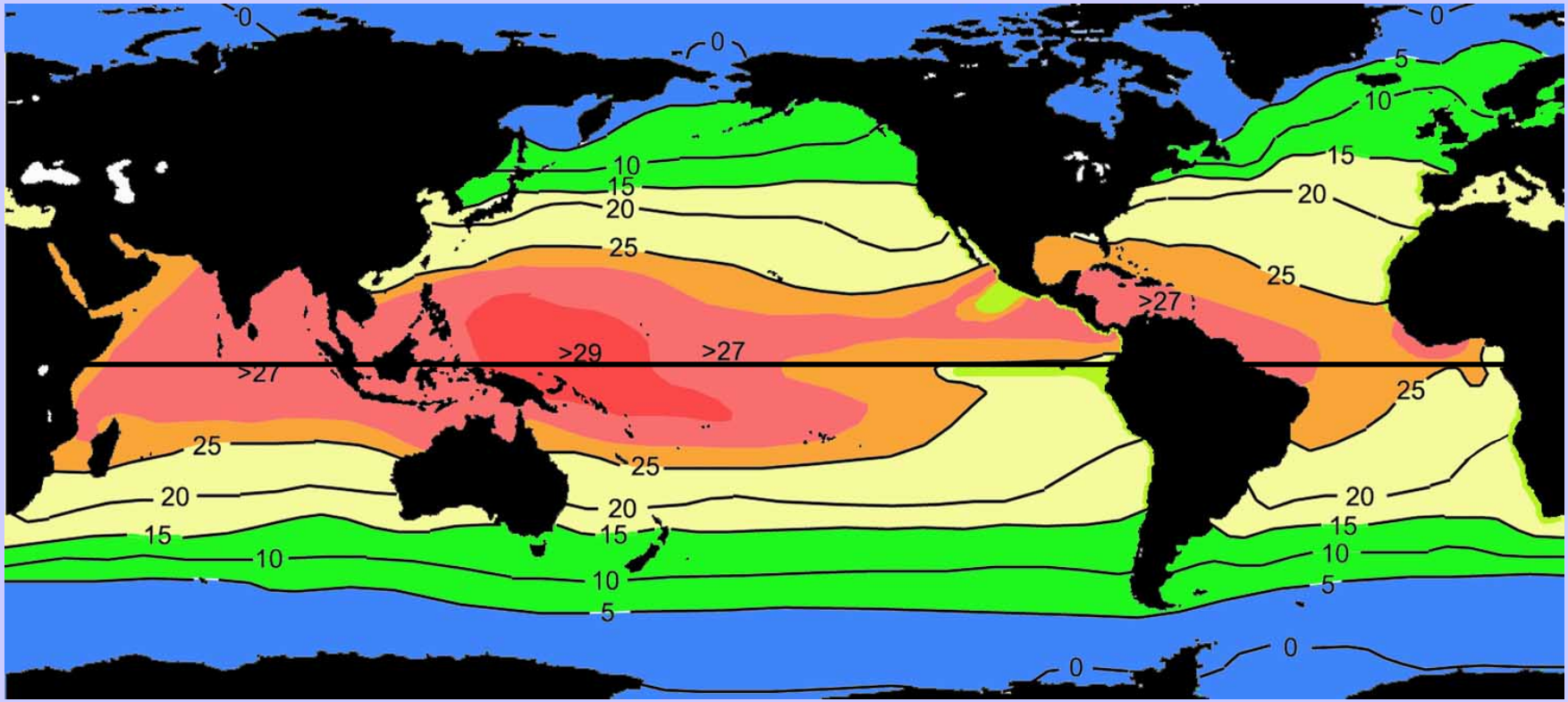
- The North Atlantic will gain *more* heat at the expense of the South Atlantic.
- This is the reverse of the usually seen break-down-of-the-NADW-production scenario.

Chief argument:

Increased N-S asymmetry favors northern heat piracy.



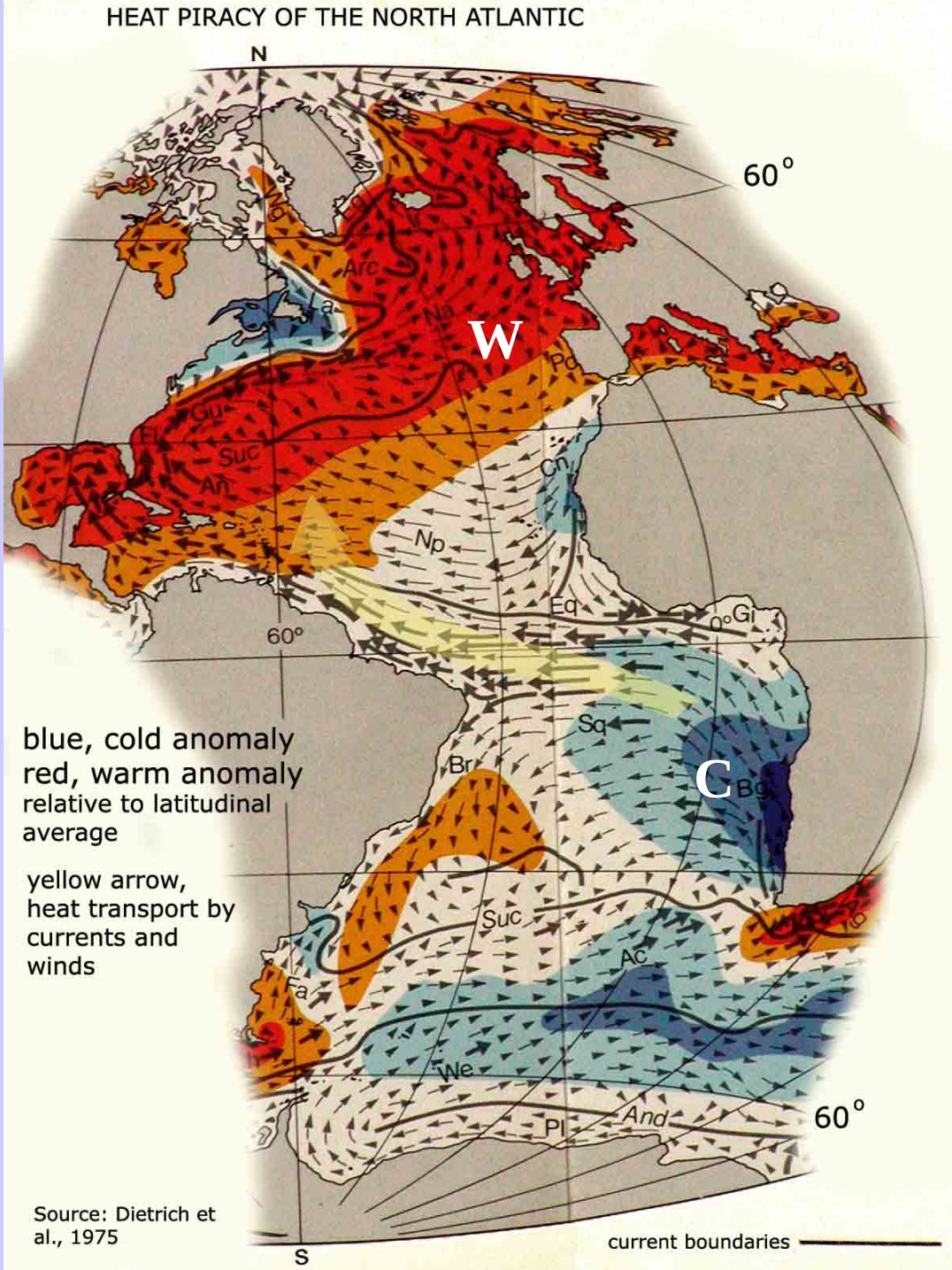
Heat distribution in the present ocean is asymmetric: Note that well over one half of the area of warmest surface waters is north of the equator.



Base map from F. Bryan

- **THE BEST EXAMPLE FOR NORTHERN HEAT PIRACY IS WITHIN THE ATLANTIC OCEAN.**

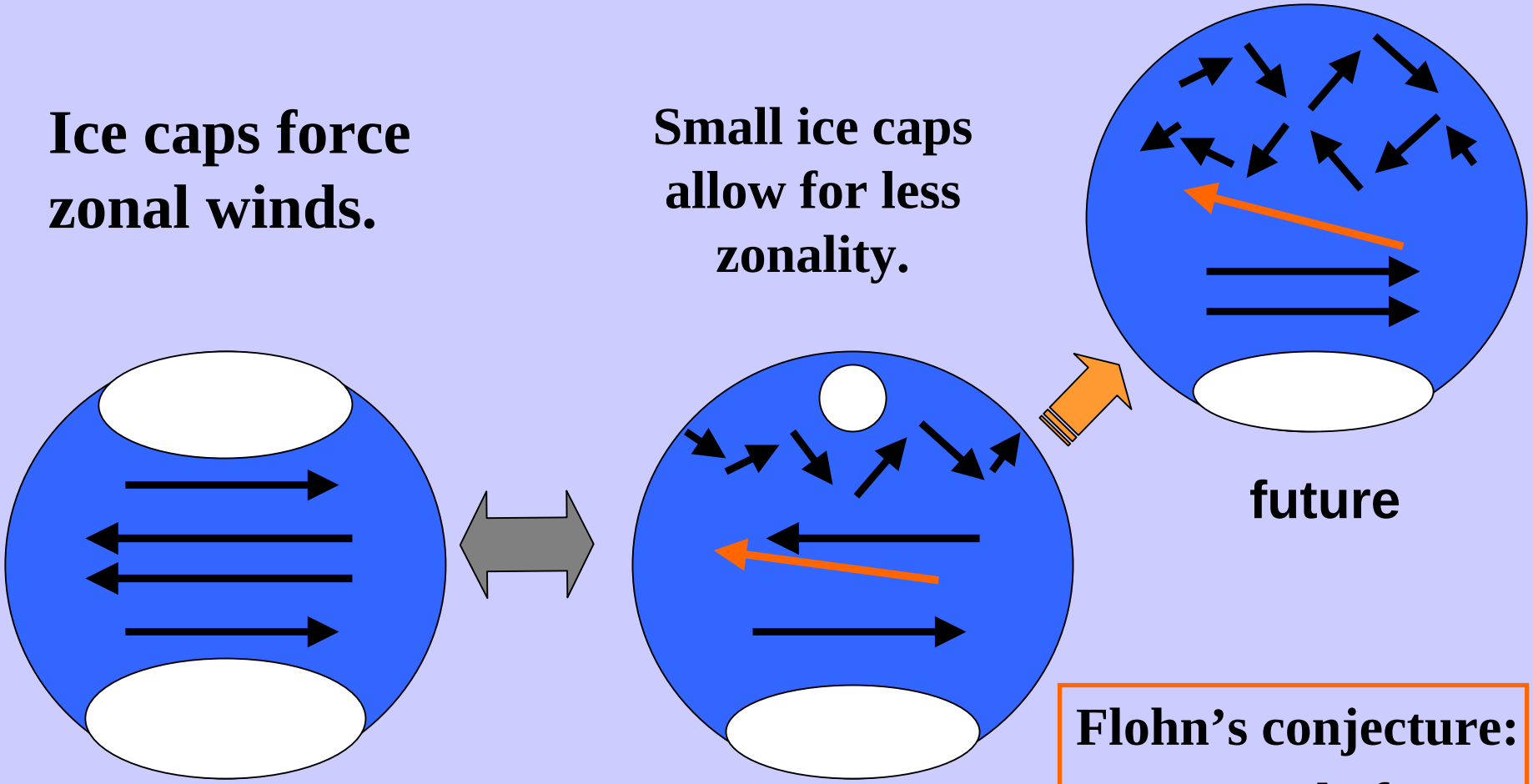
We have known, for more than half a century (when a figure similar to that on the right was first drawn) that heat moves across the equator from south to north, both with surface currents (as shown) and with warm moist winds (into the Intertropical Convergence Zone). The heat then largely moves westward, carried along with trade winds and currents.



Winds feed on temperature gradients.

Ice caps force zonal winds.

Small ice caps allow for less zonality.



future

Cold periods

Warm periods

present

Flohn's conjecture:

Removal of northern ice favors N-S heat transfer

How will the wind patterns change? Expect: changes toward less predictable patterns.

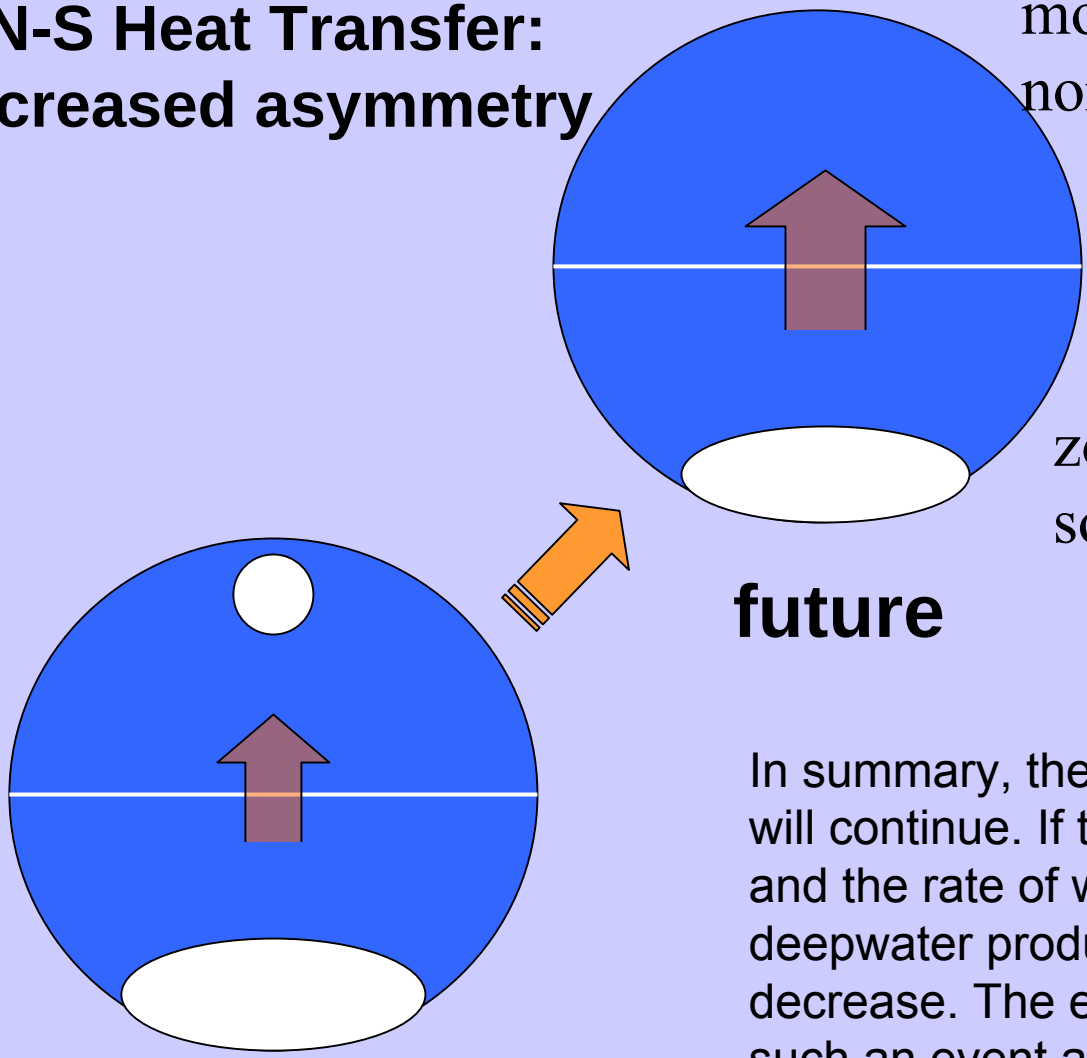
The nature of the wind patterns in the northern hemisphere will change, from strongly zonal to strongly monsoonal.

Monsoonal winds attract heat from south of the equator (most of the planet's land is in the northern hemisphere).

Nature of the evidence: the early Pliocene was warm in northern latitudes (no large ice masses). NADW production was strong.

N-S Heat Transfer: increased asymmetry

monsoonal winds gain north of the equator



zonal winds remain strong south of the equator

future

present

zonal winds strong in both hemispheres

In summary, the south-to-north heat transfer will continue. If the production of meltwater and the rate of warming overwhelm the deepwater production, it may greatly decrease. The effects on heat transport of such an event are not clear, from the record; presumably transport is diminished. While the effect of deepwater shutdown may be transient geologically, it may dominate on a scale of centuries.

The ocean's productivity response to warming depends on what happens to upwelling, which controls maximum (and overall) production.

Will upwelling increase or decrease?

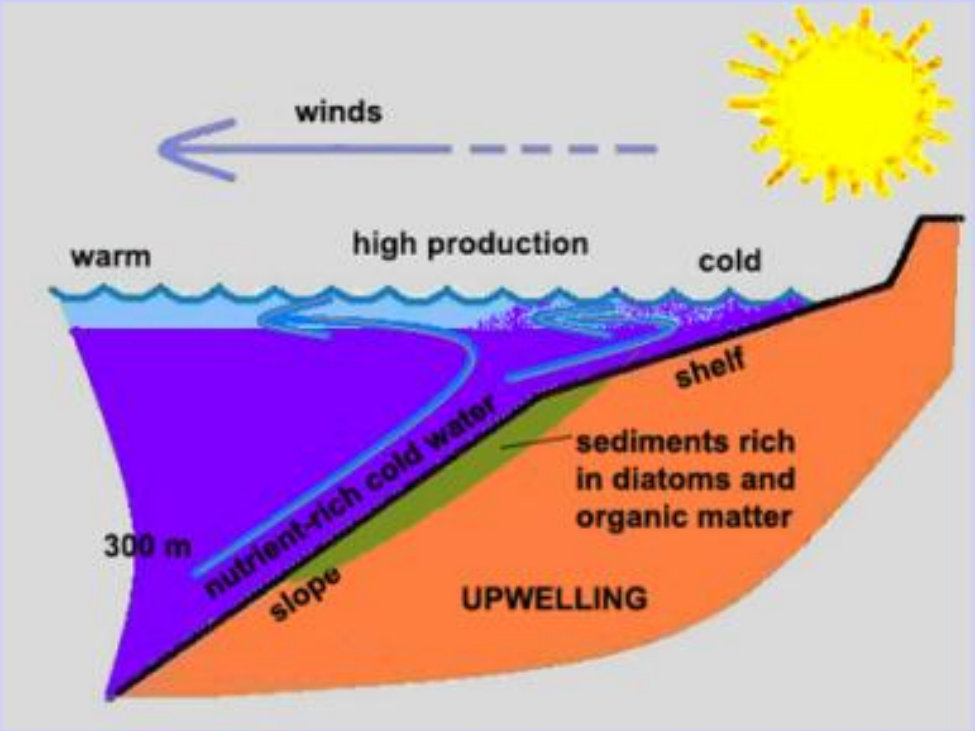
The answer depends largely on the future strength of zonal winds. North of the equator, one would expect a decrease of the planetary temperature gradient and a corresponding decrease in the strength of zonal winds (trades) that drive much of the upwelling. South of the equator, because of the relative stability of the polar front around Antarctica, zonal winds might not decrease at all, for centuries . Upwelling (and high production) will then persist.

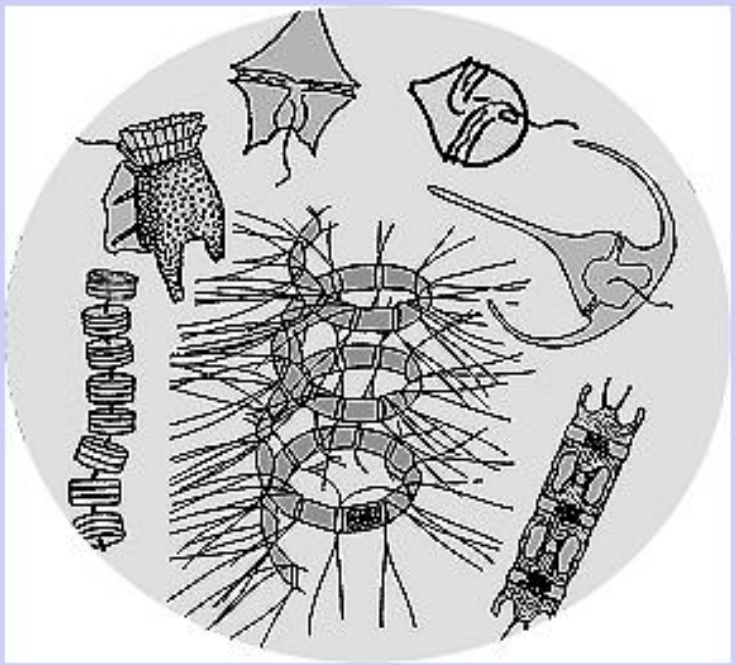
Gannett on the prowl for fish, Helgoland Photo W.H.B.

Sea birds and sea mammals, with their high energy requirements, depend on upwelling and the high productivity it provides.

Conclusion: the North is more vulnerable than the South.

Why is the wind field of crucial importance?





Sources: Ocean (left) and Sverdrup et al. 1942 (right).

Importantly, a weakening of zonal winds (from a decrease in meridional temperature gradients) is likely to result in a weakening of coastal and equatorial upwellling. Such upwelling provides nutrients to phytoplankton growing in sunlit surface waters (right graph). The phytoplankton is at the base of the food chain to fish and all large organisms. A weakening of upwelling, therefore, would affect the food supply of all creatures in the sea.

Quaternary: measuring changes in paleoproductivity poses some challenges. Clearly, in the western Pacific, PP changed by more than a factor of 1.5, based on bf flux studies on box cores (graph).

Paleoproductivity from benthic foraminifera abundance: Glacial to postglacial change in the west-equatorial Pacific

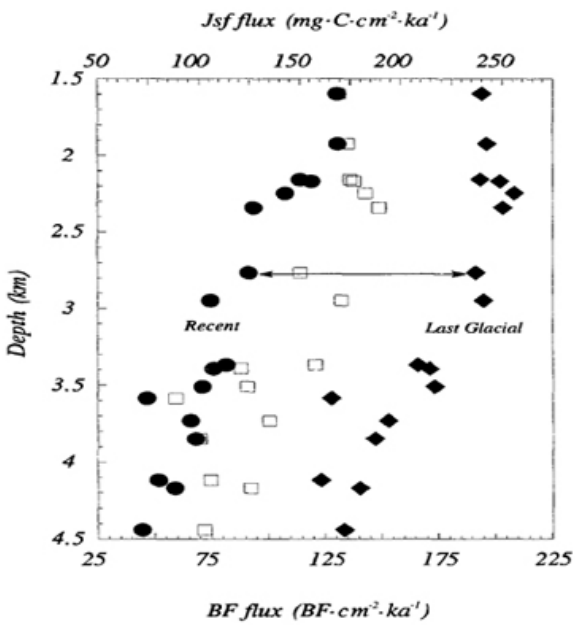
Geology 1991;19;1173-1176

Juan C. Herguera, W. H. Berger
S.I.O./UCSD
La Jolla, California 92093

ABSTRACT

Surface productivity is correlated with the rate of accumulation of benthic foraminifera on the deep-sea floor. As a rule of thumb, for each 1 mg of organic carbon arriving at the sea floor, one benthic foram shell >150 μm is deposited. The correlation can be used to reconstruct organic flux to the sea floor in the past, and hence the productivity of past oceans. Applying the appropriate equations to box core data from the Ontong Java Plateau in the western equatorial Pacific, we found that productivity during the last glacial maximum exceeded present productivity by a factor of between 1.5 and 2.0, with intermediate values for the mid-transition period. Accumulation of benthic foraminifera was depressed on top of the plateau during the glacial and transitional period, presumably because increased winnowing removed part of the food supply.

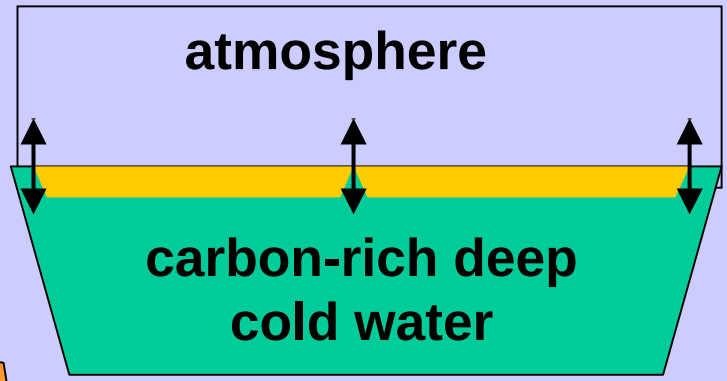
Figure 1. Accumulation rate of benthic foraminifera (BF) as function of depth, Ontong Java Plateau. Circles = core-top samples; squares = mid-transition; diamonds = late glacial maximum. Bottom scale: estimated *Jz* flux (in mg C · cm^{-2} · ka^{-1}).



Accumulation rates are not necessarily readily available, so ratios of species have to be used in cases to obtain at least qualitative information (example: "BoBu index" [Bo+Bu]/total bf; Bo, *Bolivina*; Bu, *Bulimina*).

How will the uptake of carbon dioxide change? Expect: less uptake in the future.

weakened zonal winds, rain in high latitudes, warming of surface waters

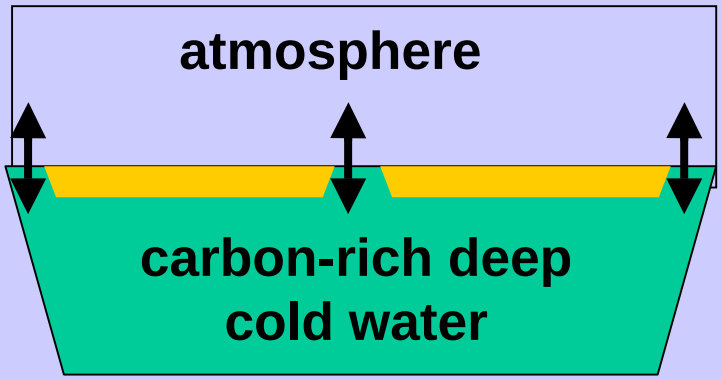


exchange through mixing and convection

future

access to deep water reduced

reduced exchange

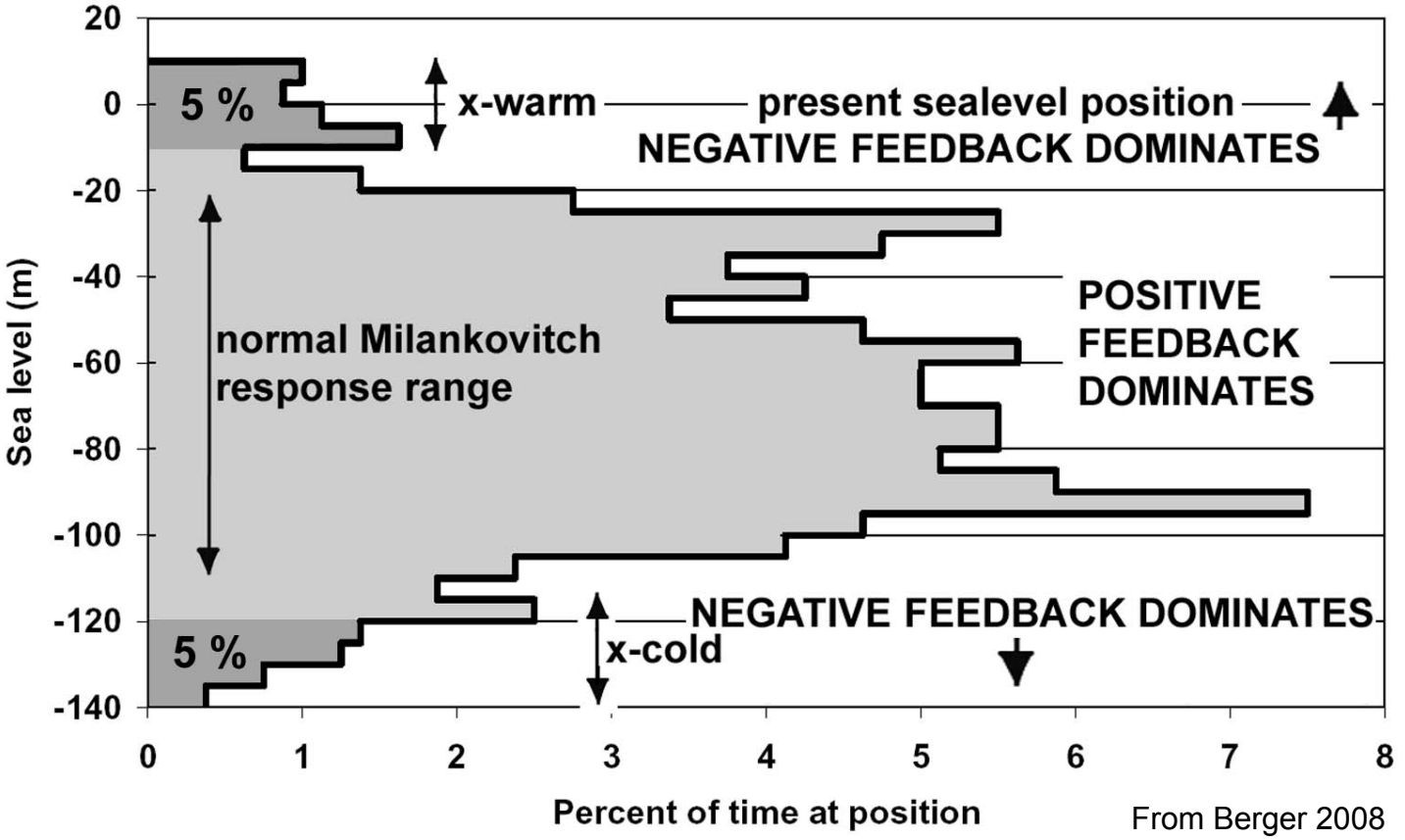


present

deep water accessible

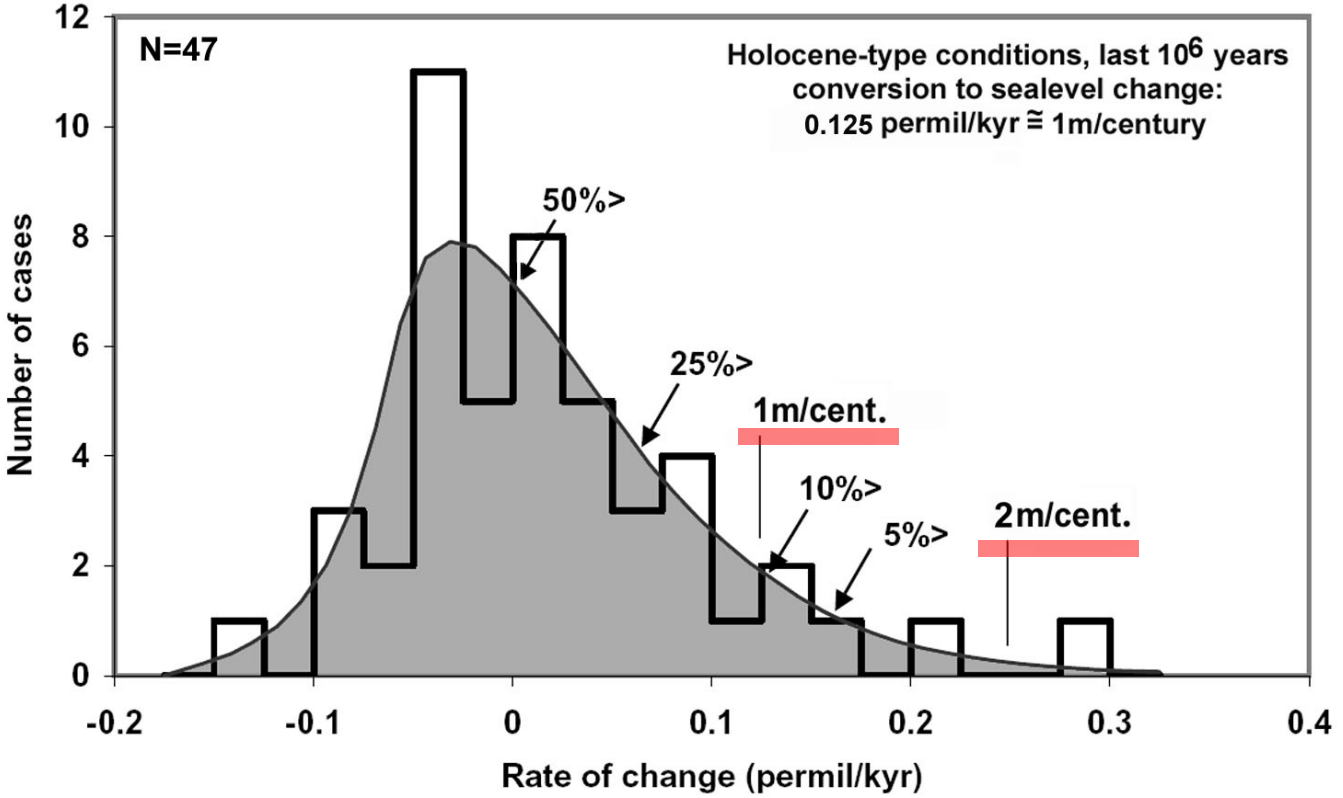
The ocean will become less helpful in sequestering industrial carbon dioxide, and not just from acidification.

HOW TYPICAL IS THE PRESENT POSITION OF SEA LEVEL, SEEN AGAINST THE TIME SINCE THE MPR? NOT TYPICAL AT ALL.



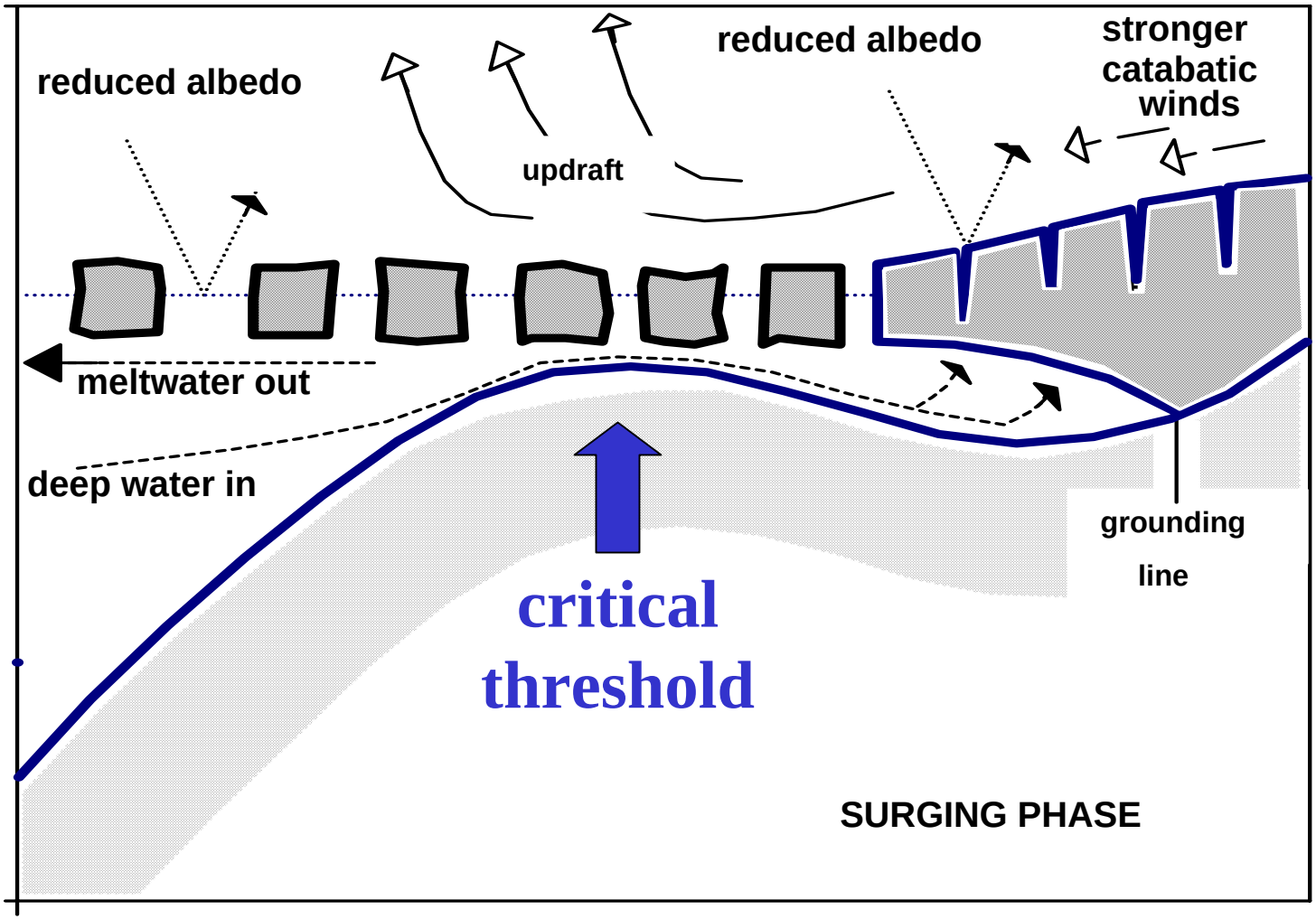
Distribution of sealevel position in the late Quat.

Obviously, the most interesting point is how fast sea level rose during times that were much like the present. For this to emerge, we need to restrict analysis to the five percent of the record when sea level stood highest ("x-warm"). We chose the conditions when sea level was higher than -10 m, which is clearly offset from the "regular" pattern, as shown in the histogram of sealevel abundances. Note that there is evidence that sea level did not rise above +10 m during the ice ages, which suggests strong negative feedback on further rise at this level (since there is still plenty of ice around in AA).



Abundance distribution of rates of change of oxygen isotopes in the late Quaternary, and implications for rates of rise of sea level.

The number of points in the uppermost 5% of the range, for the last million years, is 47; roughly 5% of the total. Fifty percent of these points are end points (change = 0) of a 1 kyr rise (as shown). Fast rise (>1m/century) is restricted to the top 10 percent of instances (5 cases). What this suggests is that, in the past, sea level rose by > 1 m/century without much provocation about 10 percent of the time under conditions similar to the Holocene ones. With ongoing warming, the chance for such rise therefore should be substantially higher than ten percent. In fact, it may be possible to reach the 2 m/century level of action with sufficient warming. Note that on the whole the melting of ice is faster than buildup.



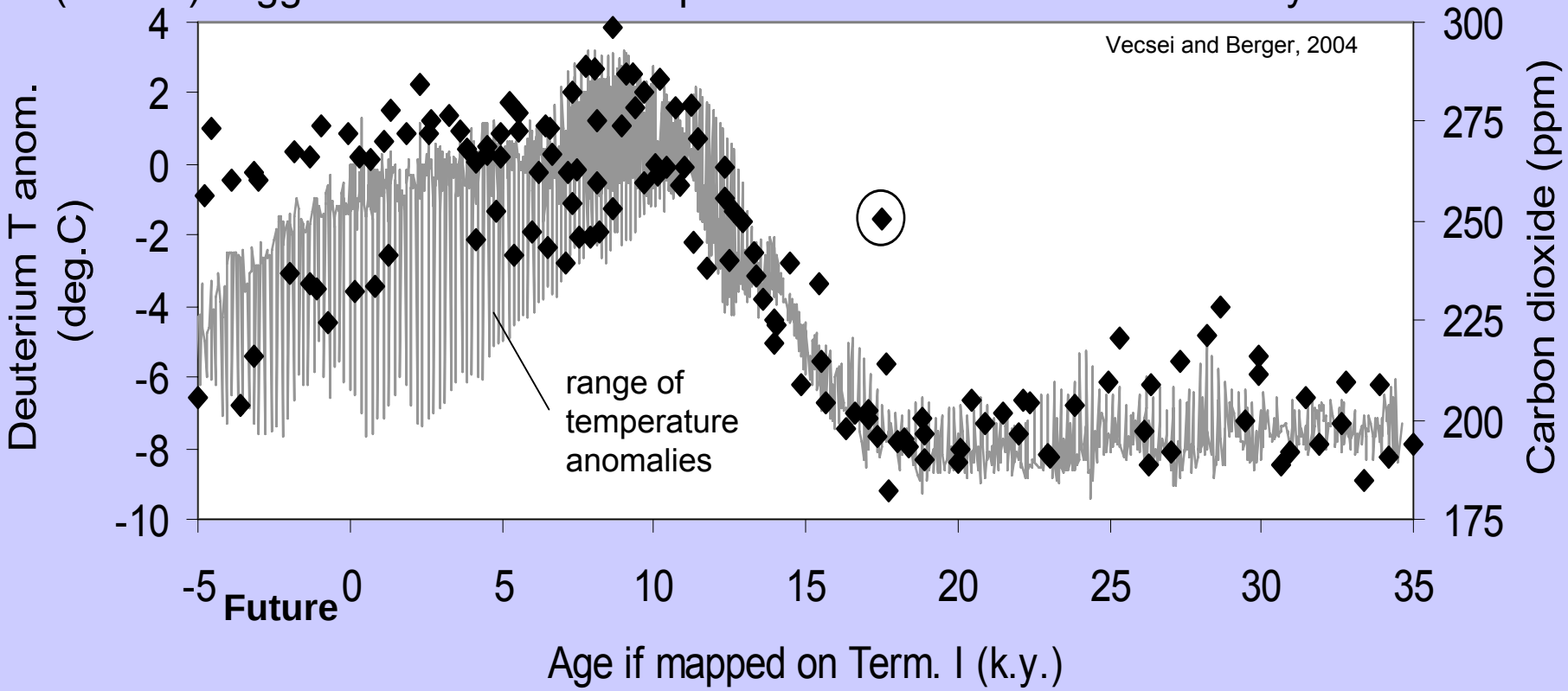
Berger and Jansen, 1995.

Why should glacial ice masses melt so fast within the transition to postglacial condition? The answer is unstable ice.

Collapse (for Step 1 of deglaciation) is triggered when seawater is able to invade below marine-based ice and melt it from the bottom.

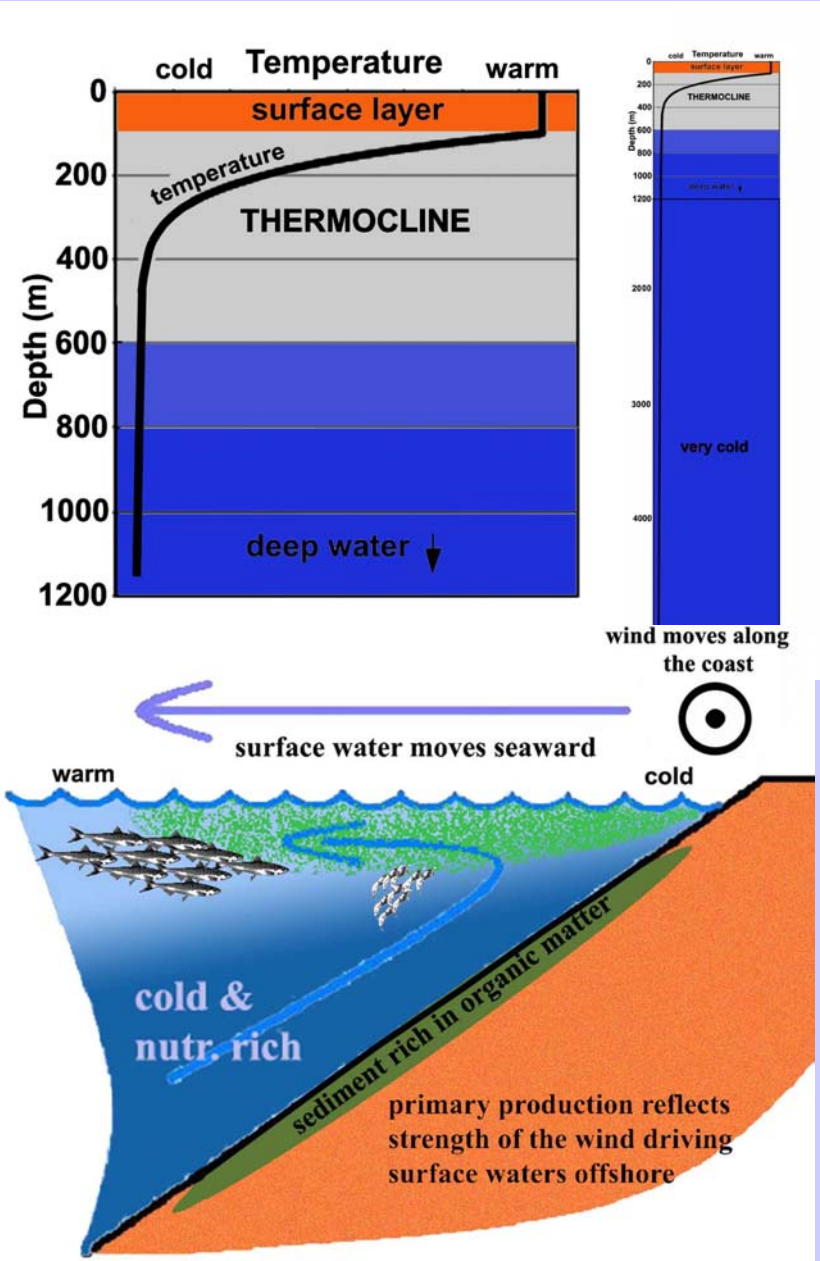
What about feedback from a change in carbon dioxide in the air?

The ice record suggests that a rapid rise in sea level is invariably associated with a rapid rise in carbon dioxide, such that positive feedback prevails. The terminations are stacked in this graph, to show their agreement in terms of deuterium content, and in terms of the associated carbon dioxide content in the air samples extracted. (Data from Petit et al., 1999.) The anomalous point (circled) suggests caution with respect to confidence in the accuracy of the data.



Whether temperature change leads change in carbon dioxide is not clear and is largely irrelevant to the problem. The two factors change together, and small discrepancies in phasing are unlikely to affect the nature of the feedback, which is not dependent on the mode of initiation.

CENOZOIC OCEAN HISTORY: THE NEOGENE



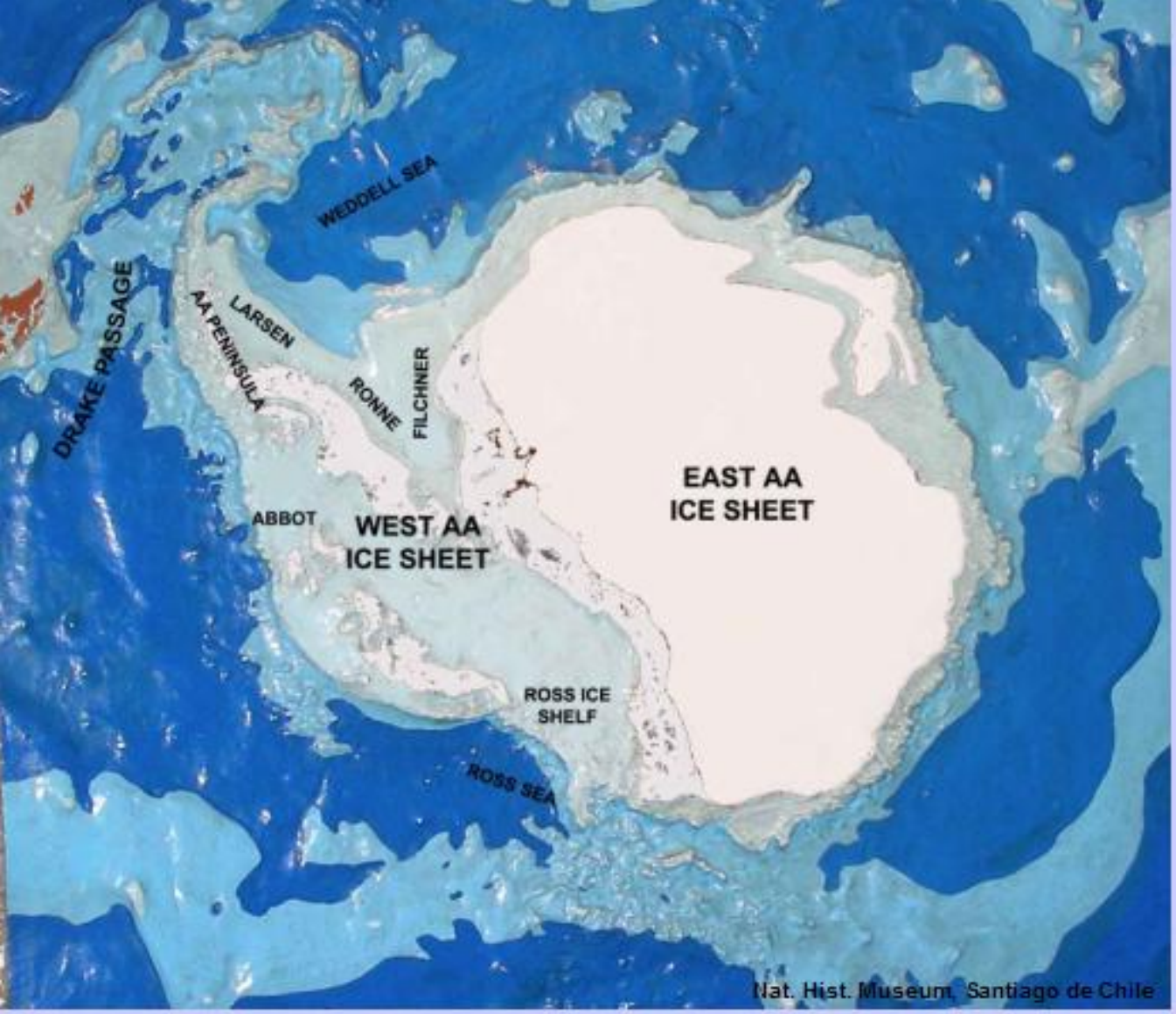
The Neogene (~24-0), just over one third of the Cenozoic, comprises the Miocene, Pliocene, and Quaternary. About 80 percent of the time is taken up by the Miocene, leaving a fifth for the other two.

The most remarkable historical events in ocean history have to do with the origin of the *THERMOCLINE* and the rise of both vast deserts and of *PRIVILEGED PRODUCTION* regions in the sea, which profoundly affected evolution. A *MAJOR BUILDUP OF ICE* occurred in the *MIDDLE MIOCENE*, presumably *MAINLY IN ANTARCTICA*. With a buildup of ice in the southern hemisphere, we have *MAJOR ASYMMETRIES* sustained by unequal ice distribution. A striking enhancement of asymmetry occurred toward the end of the Middle Miocene: *THE GREAT SILICA SWITCH.* It signaled the beginnings of the *modern ocean* (ca. 12 million years ago).

The central theme of the Neogene is a general cooling and the associated buildup of ice.

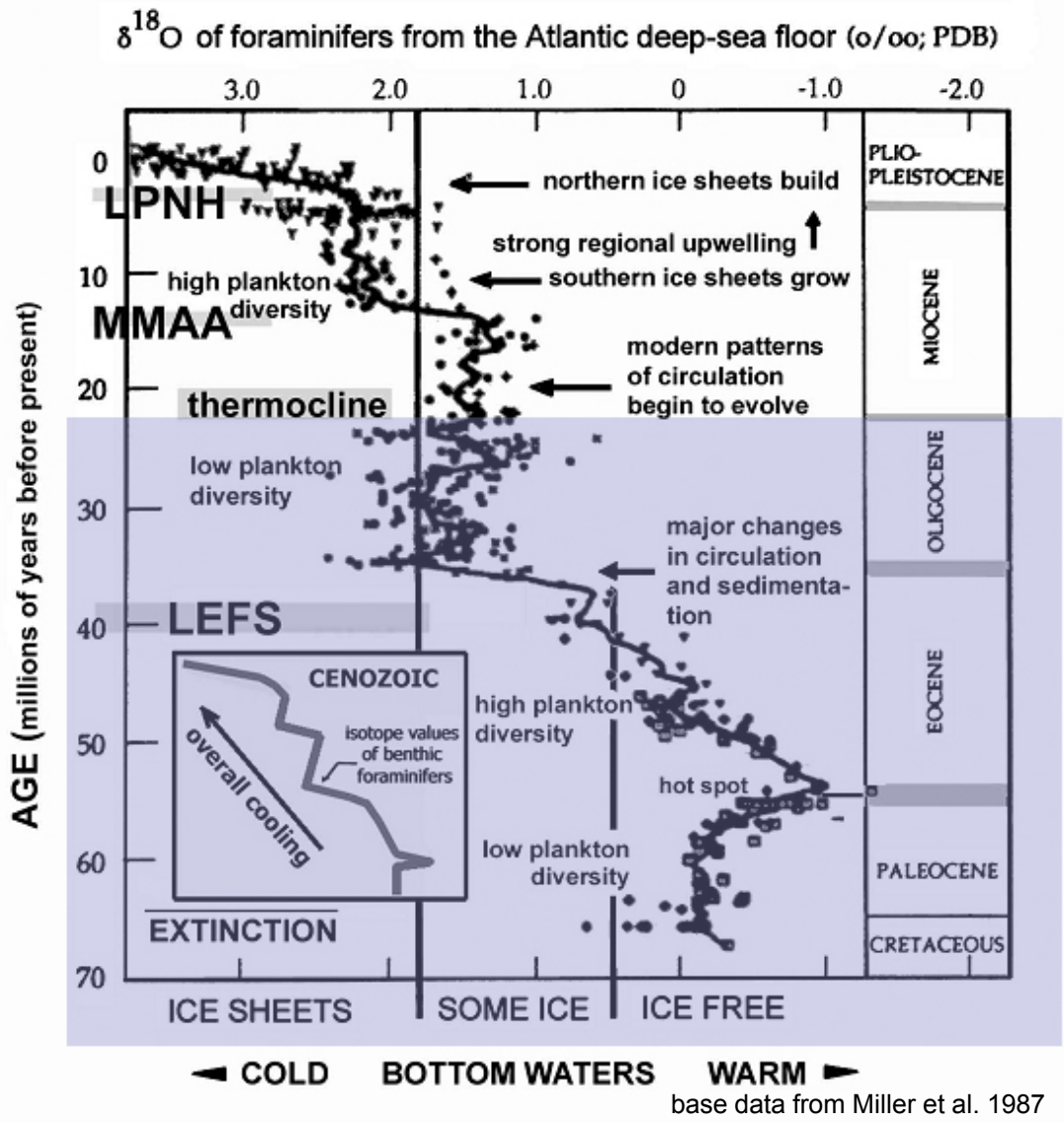
The ice sheets in Greeland (upper photo) and in Antarctica (lower photo), are self-stabilizing. They build the land to high elevation (cold!) and they reflect sunlight (staying cold).

However, once melting starts for any reason, this stability is decreased. (At lower elevations the air is warmer. Rocks and open water are dark and absorb sunlight. Meltwater within and below the ice facilitates downhill motion.)



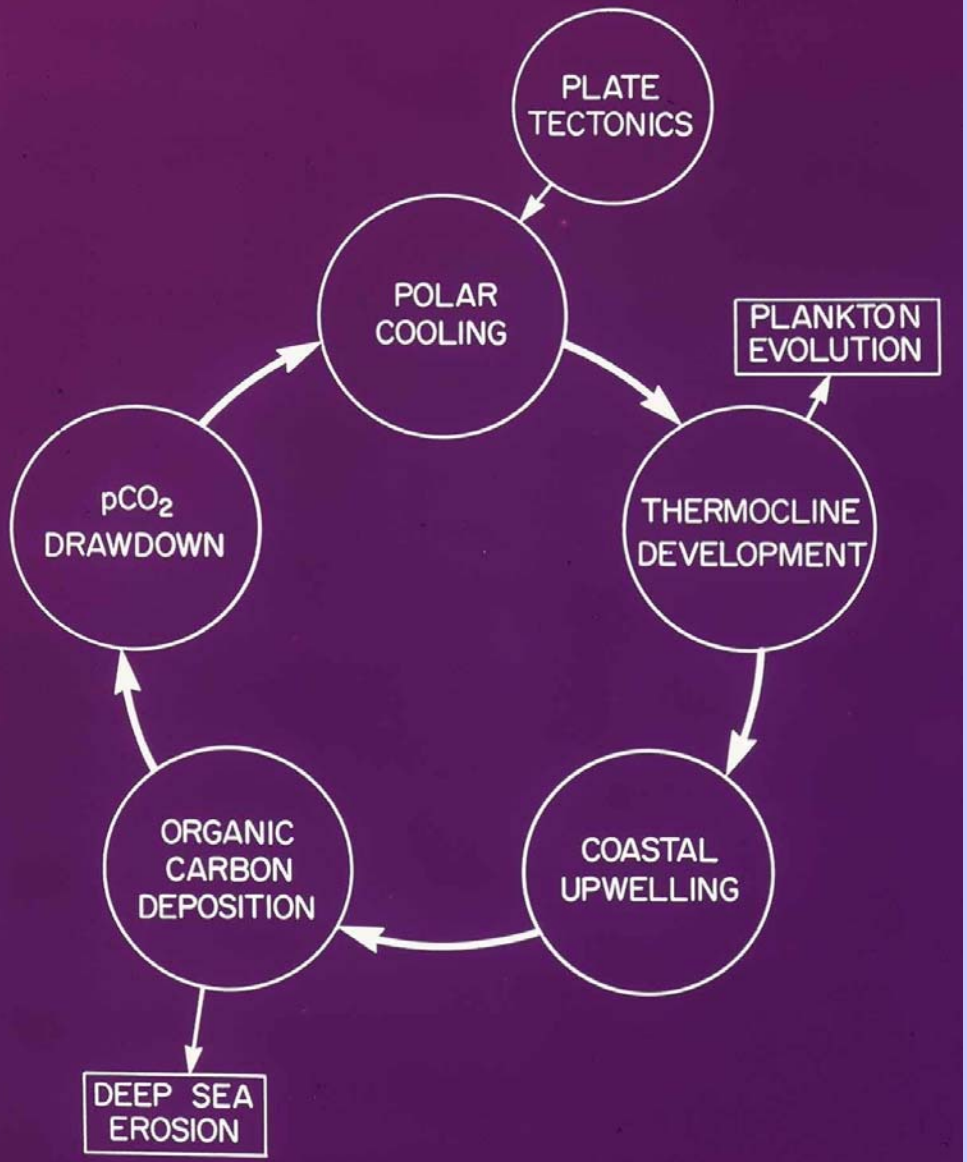
In the Neogene, substantial amounts of water were locked up in ice sheets in Antarctica. This, and the opening and deepening of Drake Passage just before the Neogene, had important consequences for the power of the great Ring Current traveling around Antarctica.

Our planet has an ice-covered continent centered on the South Pole. The ice, if melted, would raise sea level by about 60 m (200 feet). In the "West Antarctic," much of the ice is marine-grounded (light blue).



base data from Miller et al. 1987

The theme of transition from a cool-planet state (Oligocene and early Miocene) to a cold-planet state (Pleistocene) is clearly seen in the oxygen isotopes of benthic foraminifers from the Atlantic, in the compilation by Miller et al. (1987). Cooling in the Neogene occurs in two major steps -- in the middle Miocene (MMAA) and in the late Pliocene (LPNH), with ice buildup on Antarctica and in the northern hemisphere, respectively.



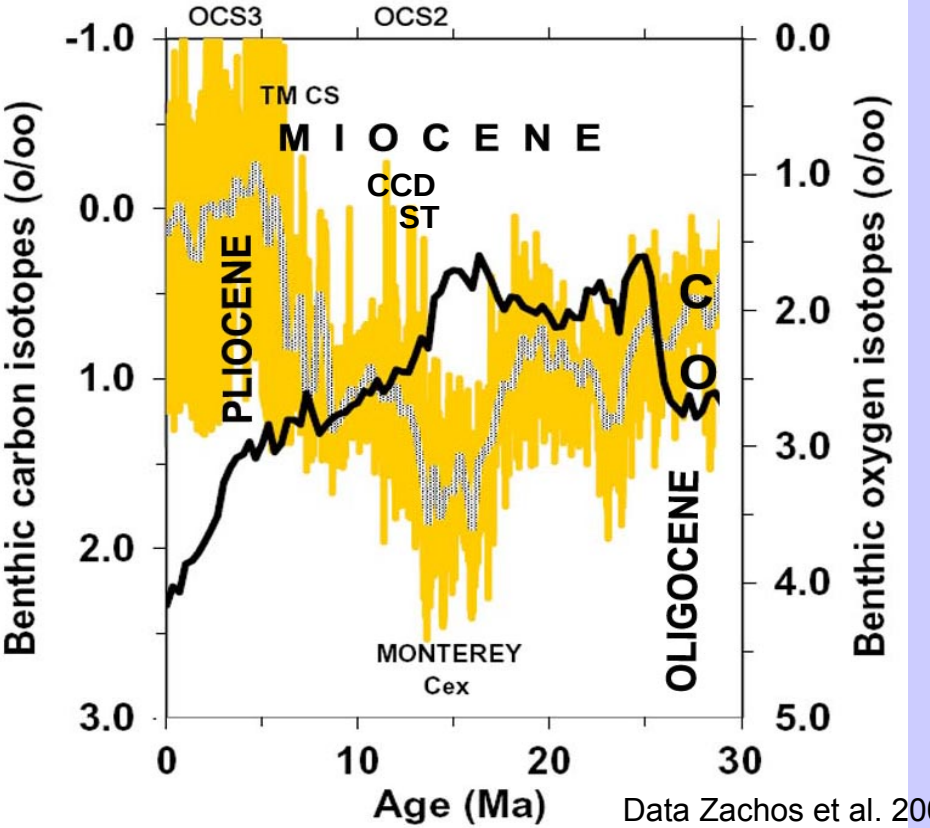
Vincent and Berger (1985)

The Mid-Miocene cooling step is preceded by a global carbon isotope excursion suggesting the large-scale burial of organic carbon ("Monterey Event" of Vincent and Berger, 1985).

A typical feedback loop for the Neogene (graph) might start with plate tectonics stimulating polar cooling, which changes thermocline development and coastal upwelling, resulting in the deposition of organic carbon and drawdown of carbon dioxide, favoring further cooling.

Mountain building, as well, would result in burial of organic carbon, through increased sedimentation rates in the continental margins. In this case, much of the organic matter could be of terrestrial origin.

Let's look at the Monterey Event, the excursion of C isotopes toward heavy values. The most parsimonious explanation is buildup of C12-rich stores of biogenic C (resulting in C12-poor waters, as seen in the foraminifers, "Monterey Cex") preceding a Seravallian-Tortonian series of excursions to negative values ("ST"). Presumably, after the buildup of negative C stores (including methane ice?) there is a major release of C, presumably as the result of a sharp drop in sea level.

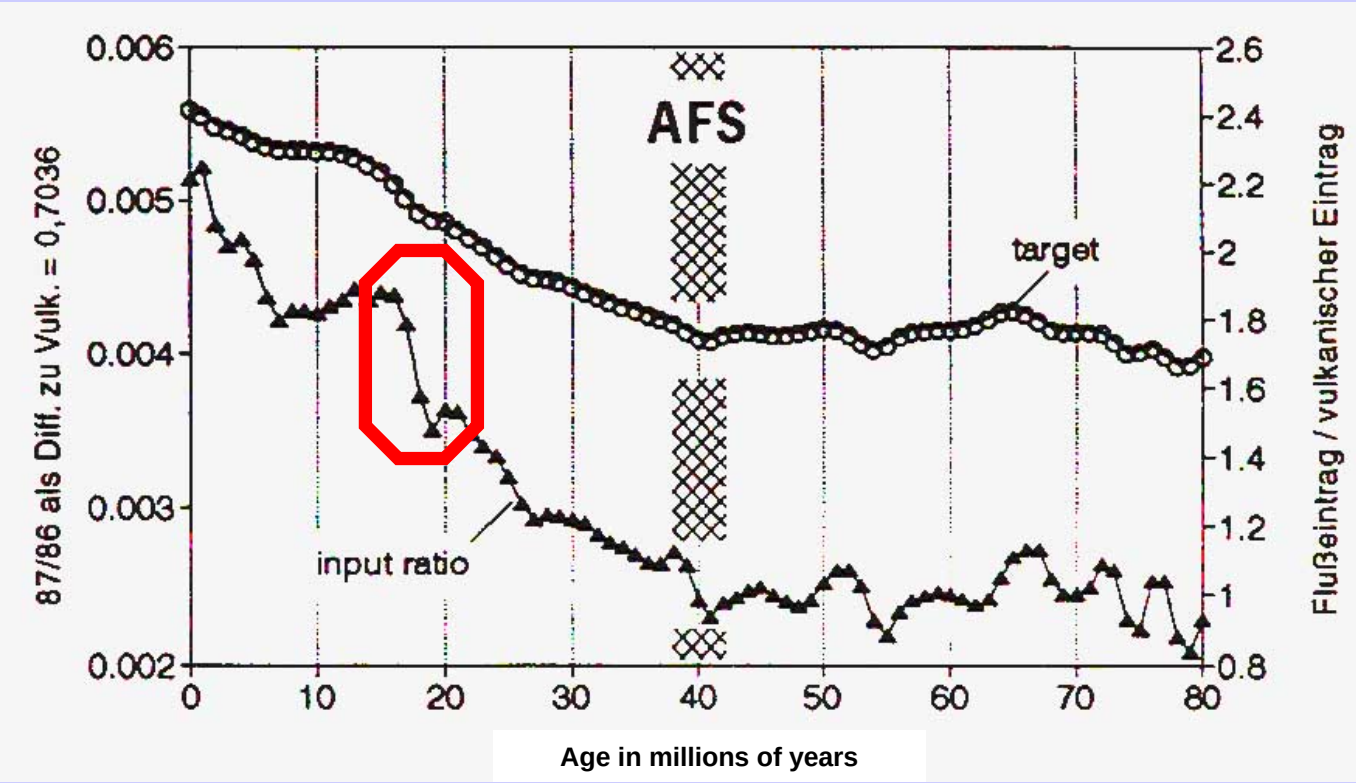


Data Zachos et al. 2001

Associated with the Sera-Torto negative excursion (ST in graph) there is an event marked by a rise of the CCD (also known as "carbonate crash") suggesting that the postulated major release of C resulted in dissolution of deep-sea carbonate.

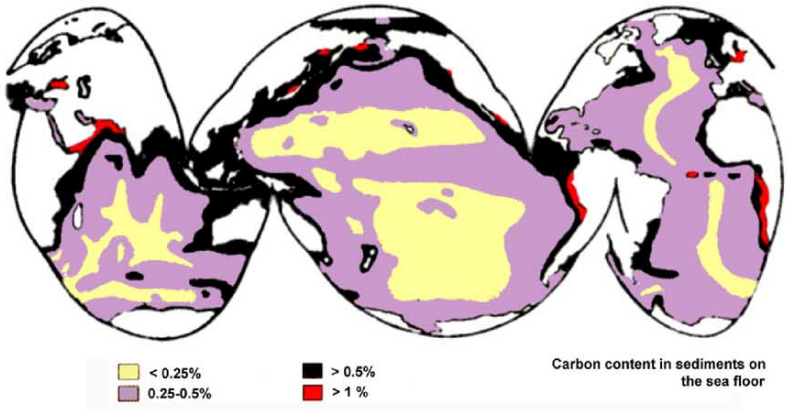
There is no cooling seen in deep-sea forams at the time of the onset of the Monterey C excursion. To invoke the growth of methane ice, we may need to call on an increase in upwelling in production-privileged regions.
TM CS, Tortonian-Messinian Carbon (isotope) step.

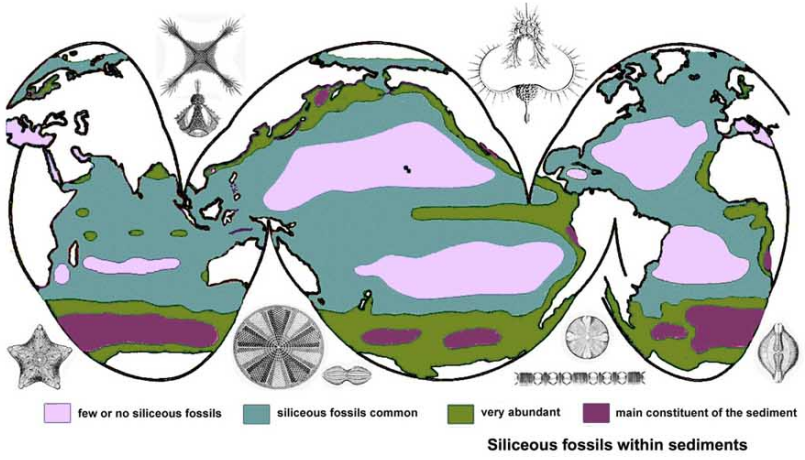
Effects from mountain building presumably are reflected in the record of strontium isotopes in marine fossils. The expectation is that the erosion of granitic continental crust will increase the ratio of isotopes 87 to 86 in the sea, since Strontium-87 is a product of decay of Rubidium-87, an element enriched in continental crustal rocks.

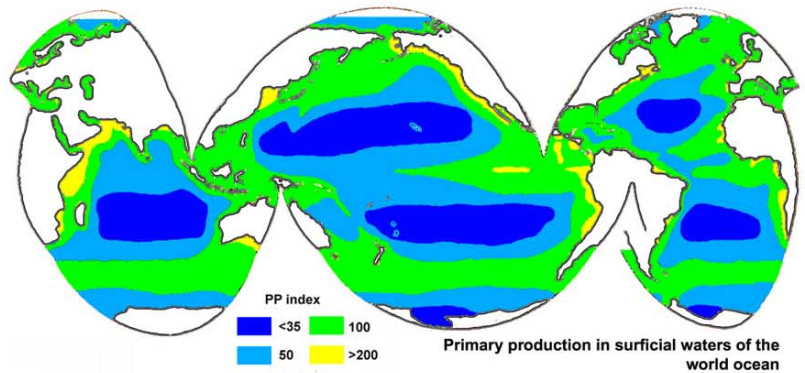


Available data suggest that a drastic step in the input of radiogenic strontium occurred in the middle Miocene, just before the buildup of large ice masses in Antarctica.

Strontium isotope stratigraphy of the Cenozoic. Left y-axis: 87/86 ratio as difference to volcanic standard (0.7036). Right y-axis: ratio of suppy from rivers vs. supply from volcanism. Note the great increase in that ratio in the middle Miocene (suggesting the rise of mountains exposing granitic crust). From Berger and Wefer, 1996. Data from Elderfield, 1986. AFS, Auversian Facies Shift in the late Eocene (large drop of the CCD).







The quality of upwelling waters (and the reliability of the upwelling itself) varies through time. This is very important when looking at biodiversity in the sea.

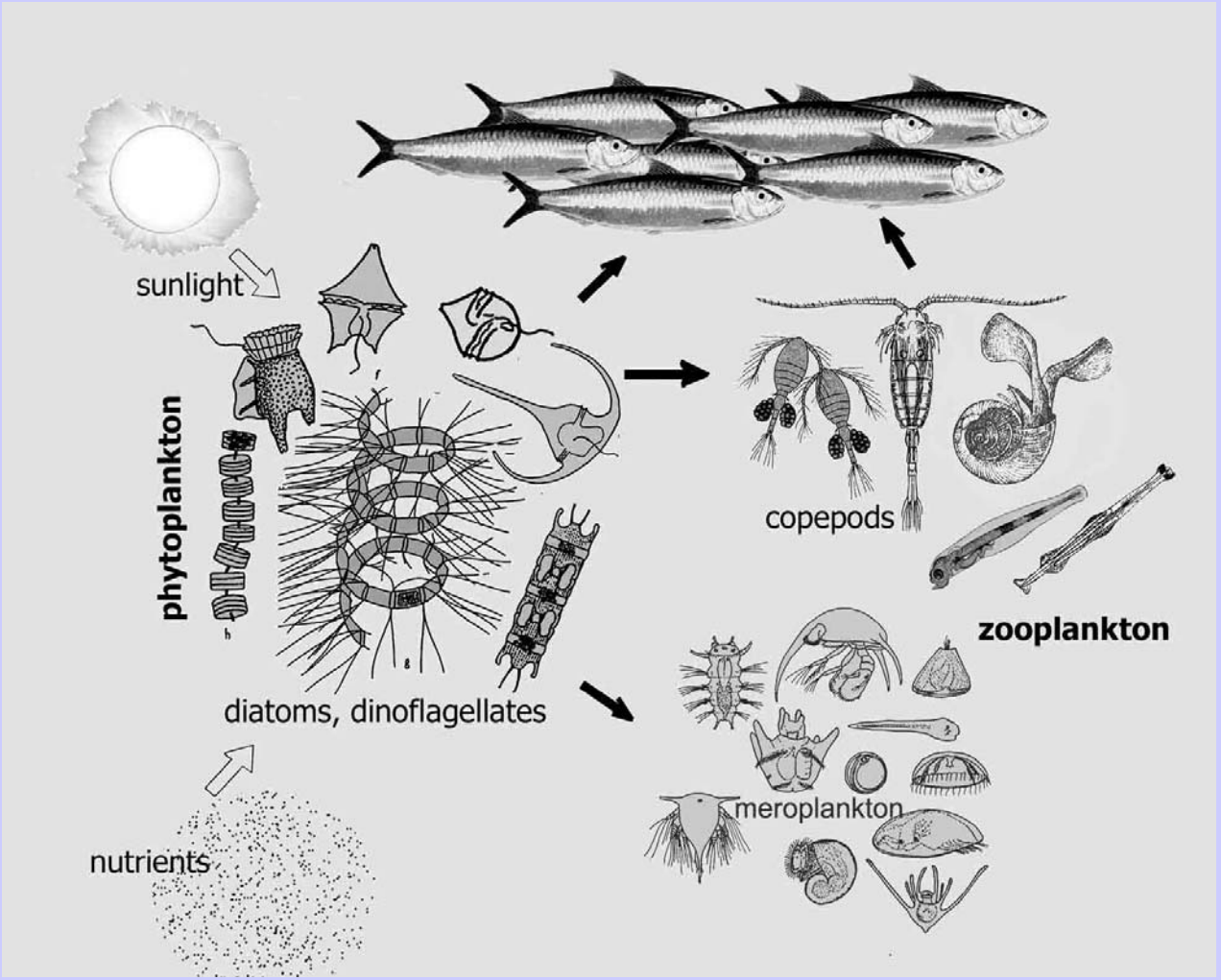
What controls the nutrient quality factor?

Probably largely the AA Ring system, which collects, sequesters, and redistributes silicate into thermocline waters.

Note the difference in deposition patterns of organic C (top panel) and silica (middle panel).

Traditional productivity maps (bottom panel) ignore differences in quality. For this reason, and others (residence times of fixed C), such maps offer no more than an impressionistic view of production.

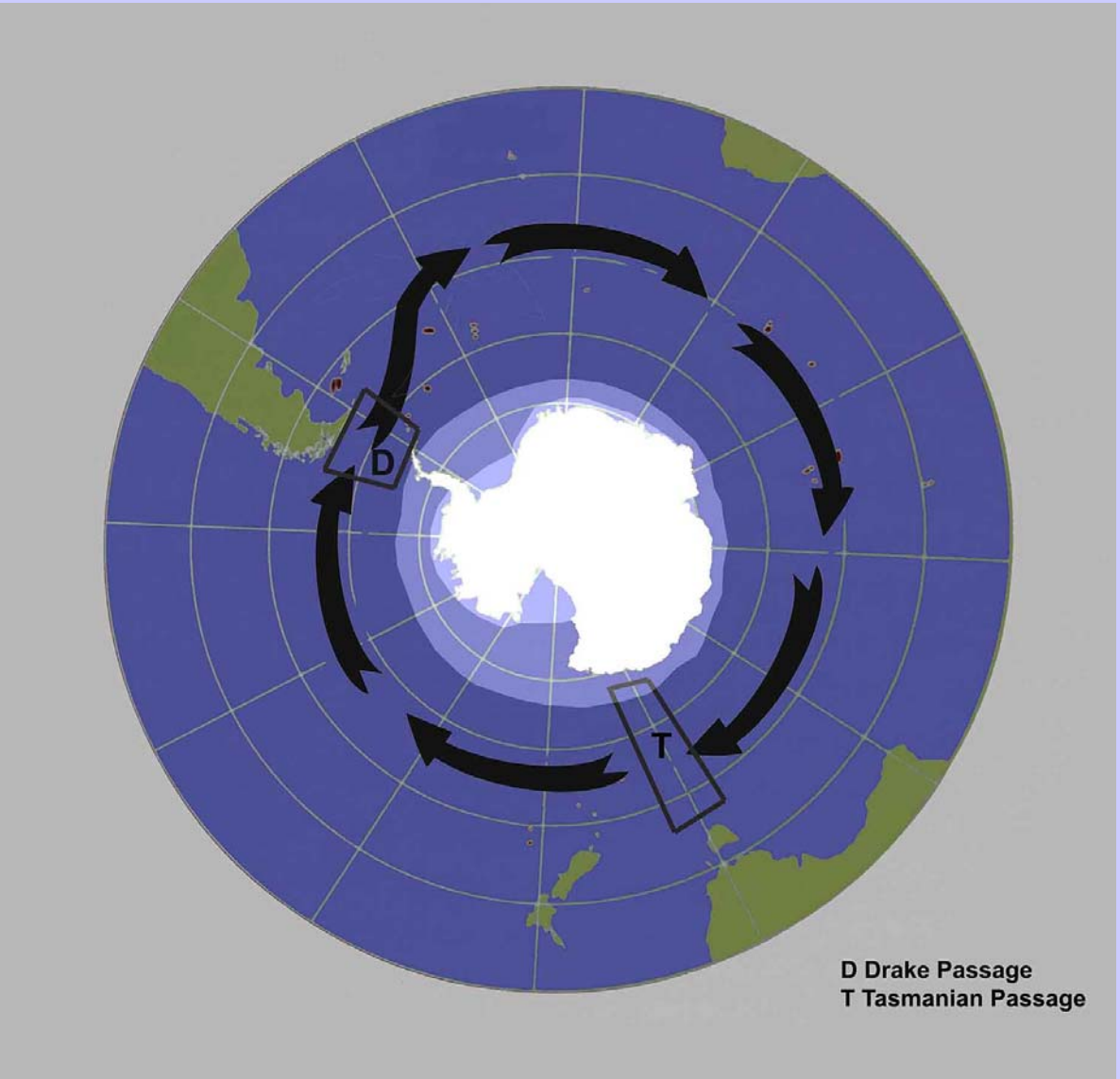
A CRUCIAL CRITERION OF PRODUCTIVITY IS THE LENGTH OF THE FOOD CHAIN (as emphasized by the oceanographer J.H. Ryther, in 1969).



WHEN AND HOW DOES THE SHORT FOOD CHAIN ARISE?

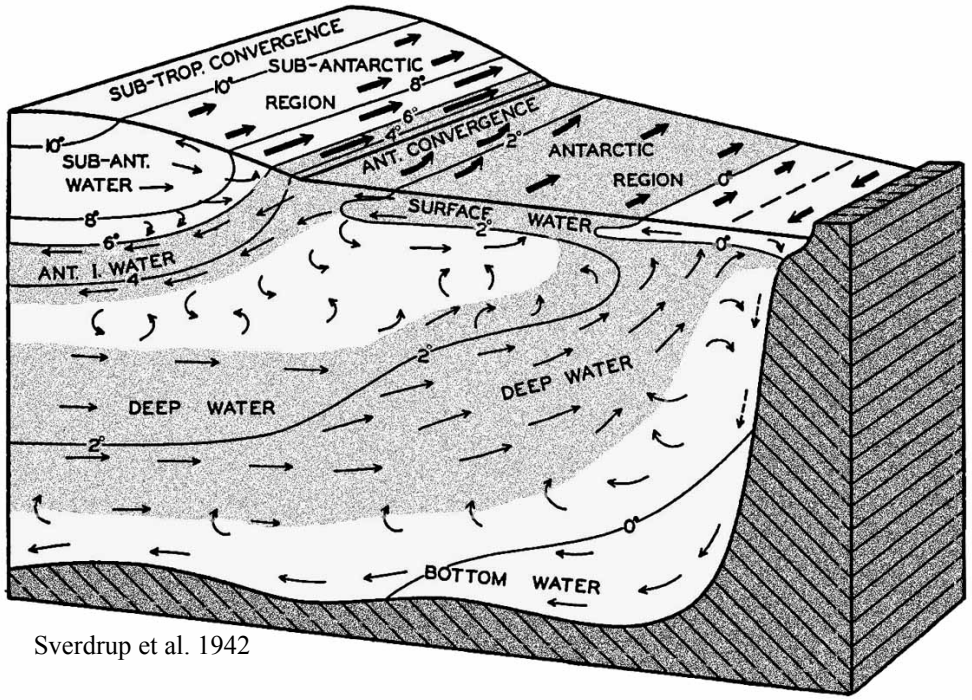
The **Short Food Chain** in high production areas is the source of whale food. (The long food chain starts with bacteria and nannoplankton and leads nowhere in terms of apex consumers such as fish, mammals and birds.)

The Great Ring Current, the circumpolar current around a white Antarctic continent surrounded by sea ice is a major feature in present-day planetary conditions. It was crucially important from the middle Miocene on, after the buildup of substantial AA ice (with Tasman and Drake Passages open for some time).

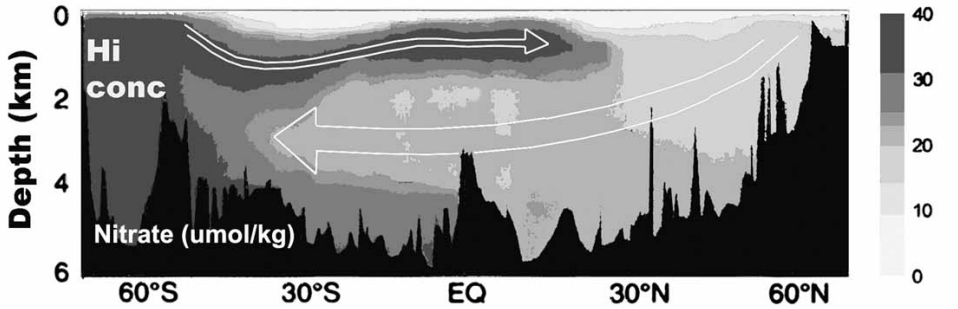


Two items are of crucial importance:
(1) AA as a source of deep cold water (in part from sea ice formation), and
(2) the trapping of silica in, and the redistribution from, the Great Southern Ring Current (arrows in graph).

The presence of silicate in upwelling is of special significance in the quality of production: silicate permits the growth of diatoms, thus favoring a short food chain. A short food chain supports high-energy animals (mammals, birds).



Sverdrup et al. 1942



Schlitzer, Wittheit Bremen

profile through the Atlantic Ocean

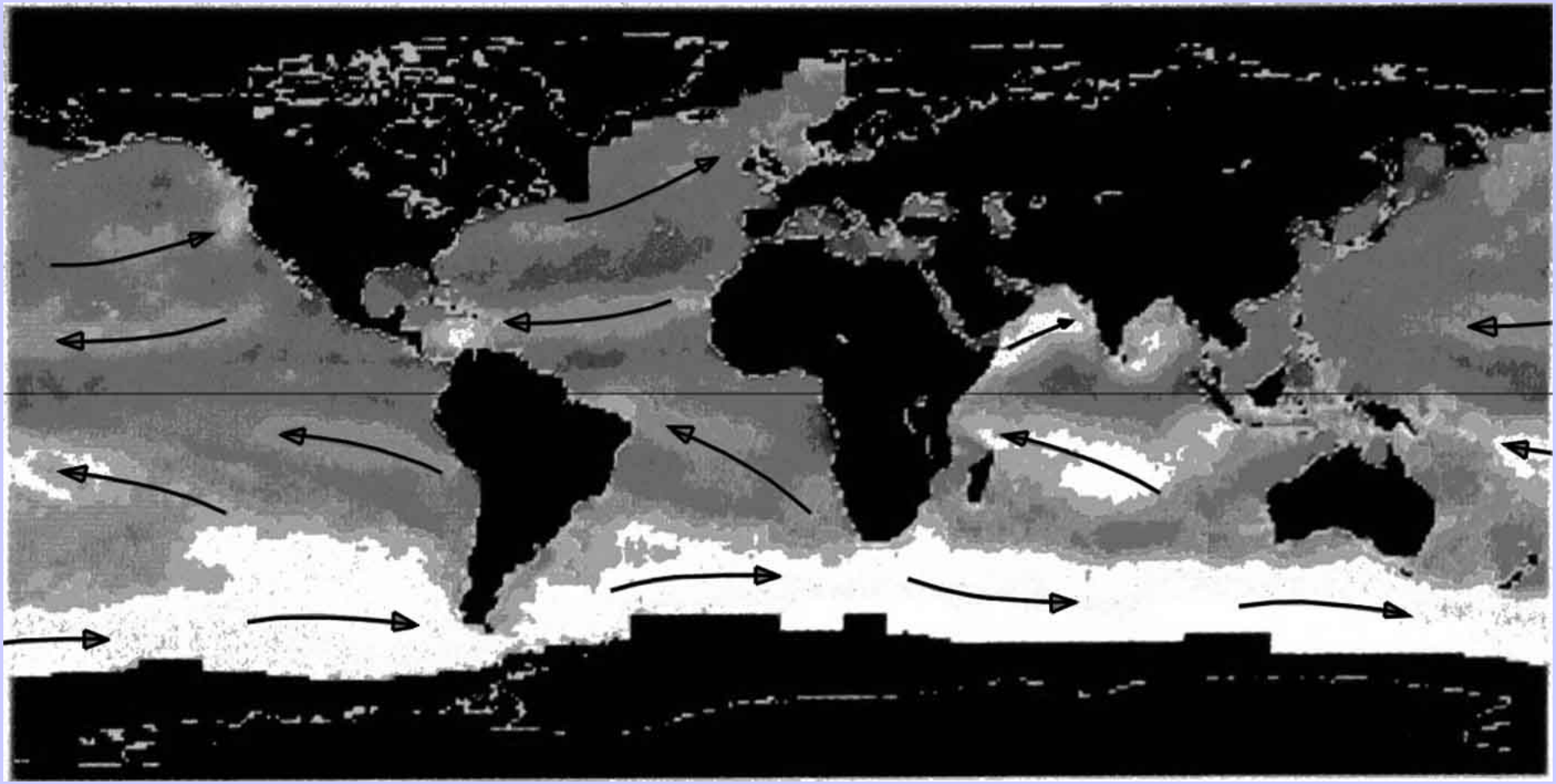
Without deep mixing, silica and other nutrients would collect in deep waters, brought there by falling diatom frustules and organic matter.

Vertical density stratification poses a major obstacle to the re-cycling of the nutrients.

The obstacle is overcome by injecting nutrients into the thermocline, laterally, in high latitudes.

The result of the high-latitude thermocline injection is seen in high nutrient content of AA Intermediate Waters.

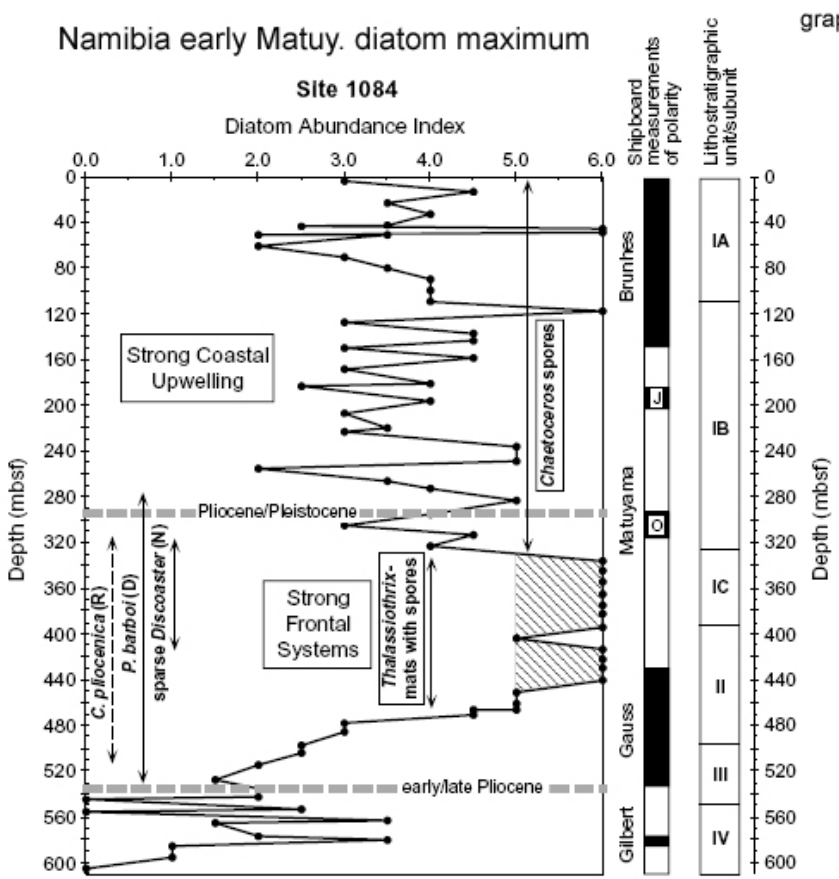
What sets the AA Circumpolar Current apart: ***the depth of mixing***, from immensely strong zonal winds (white in graph)



base map: K.Kuenzi 2002

To find clues to possible causes of varying silicate supply to the thermocline, we check how the systems reacted to the buildup of northern ice masses, in the beginning of the northern ice ages.

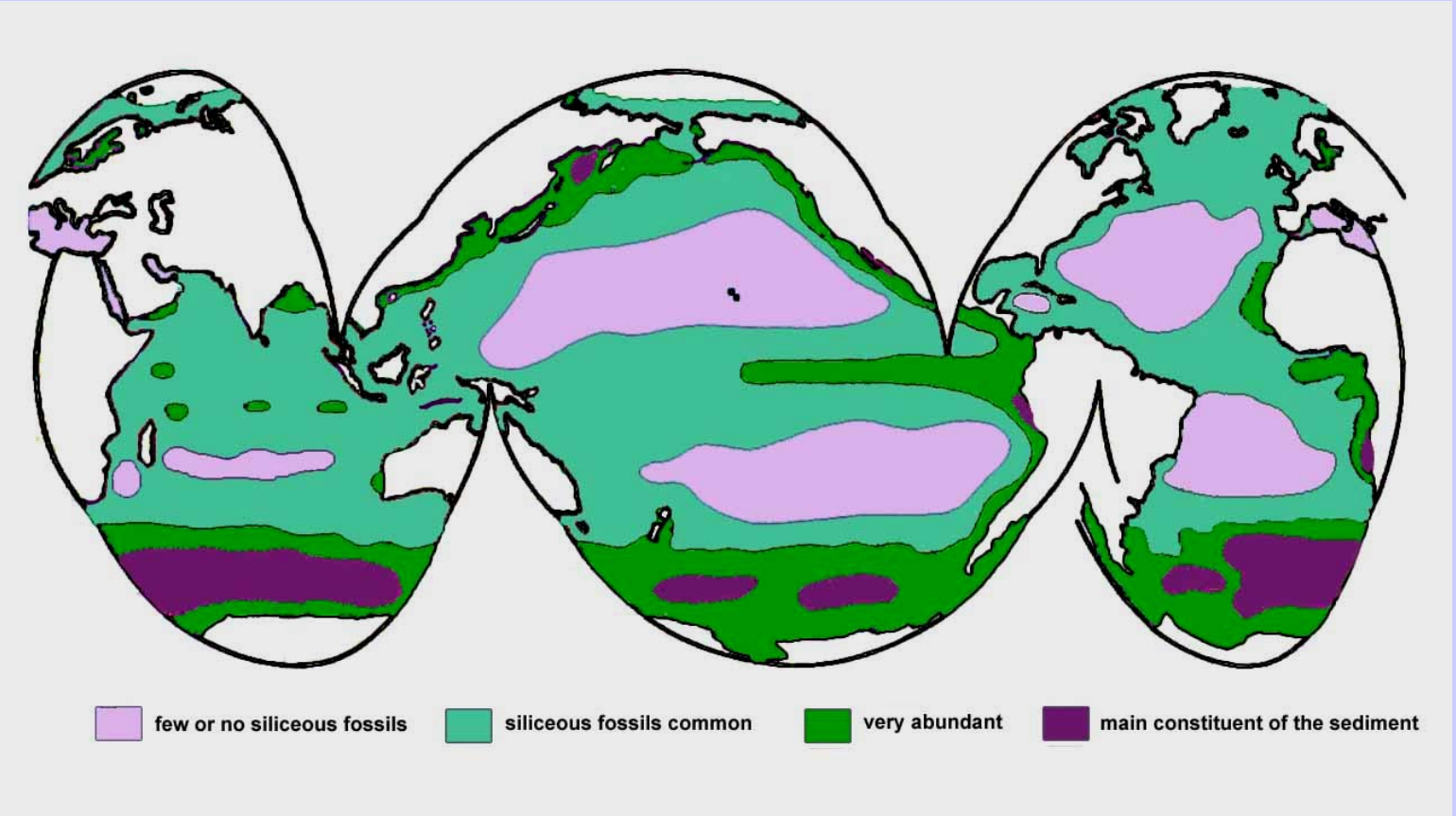
Figure F25 (continued). Site 1084. The graphs are from Shipboard Scientific Party, 1998h, 1998j). J = Jamarillo, O = Olduvai. R = radiolarian, D = diatom, N = nannofossil.



ODP Leg 175

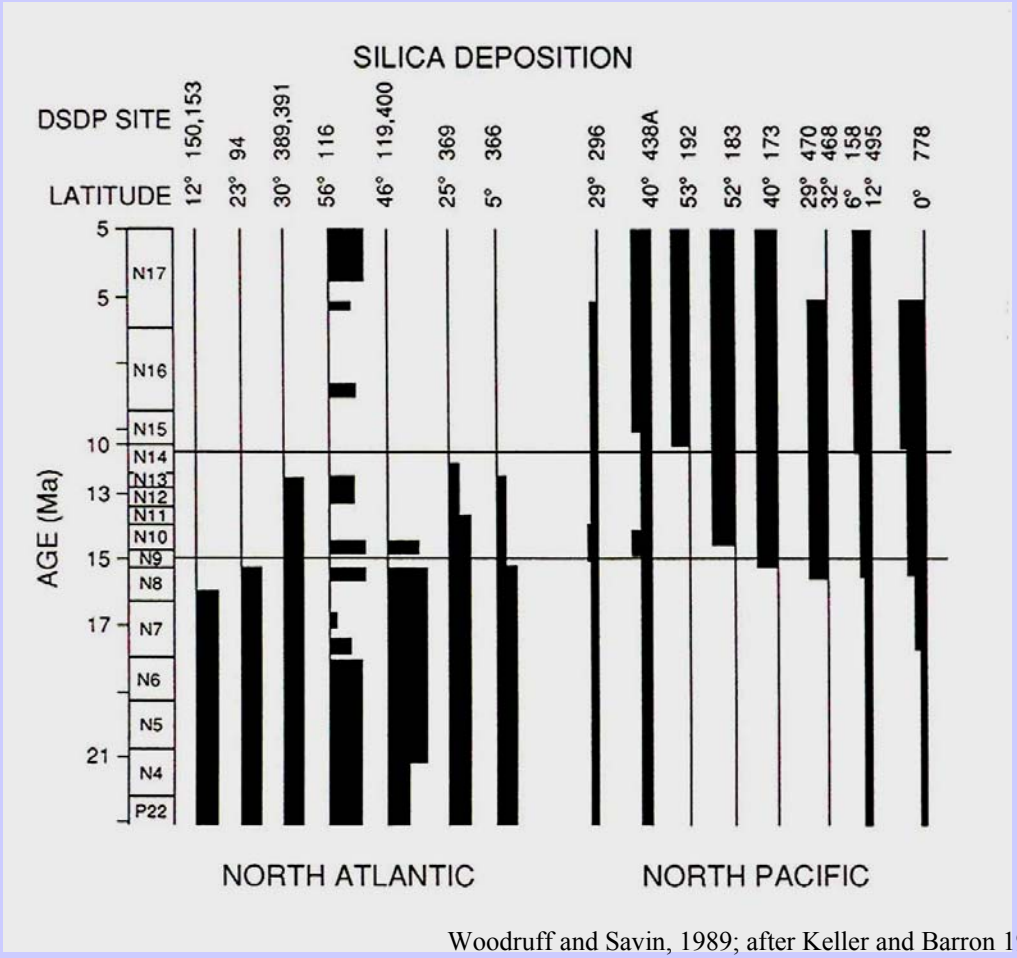
Off Namibia, the maximum supply of diatoms occurred at the end of the Pliocene, in the transition to the northern ice ages (the “Early Matuyama Diatom Maximum”). Subsequently, upwelling became stronger, but the supply of diatoms diminished (see graph). The cause, presumably, is sequestration of sililcate below regions of upwelling.

With respect to the evolution of high-energy organisms (whales, seals, seabirds) the development of the silica cycle is of utmost importance. At present, the ocean is starved of silicate except where silicate-rich waters from Antarctica enter the system. Thus, the most production-privileged systems are in the southern hemisphere.



Perhaps it is no coincidence that we here find a great diversity of marine mammals and seabirds, including penguins: Upwelling presumably has the longest history in the southern hemisphere.

With the mid-Miocene cooling step, Si-rich upwelling became important in the North Pacific.



Woodruff and Savin, 1989; after Keller and Barron 1983

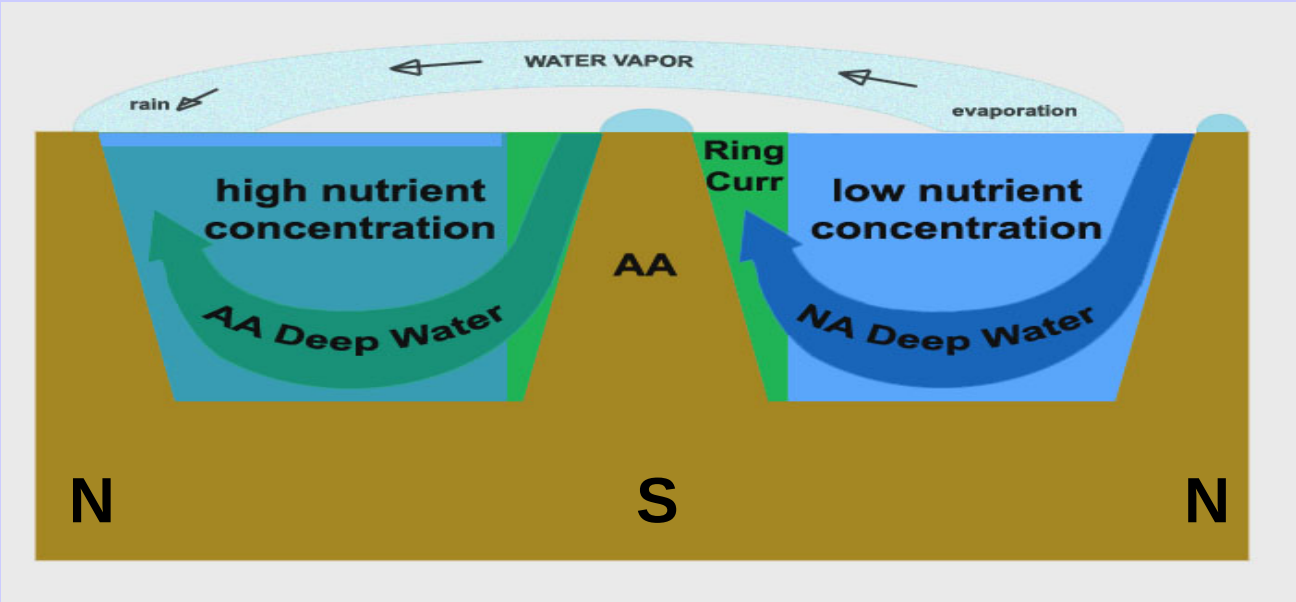
The Atlantic-Pacific "silica switch" seen in the distribution of siliceous fossils on the deep-sea floor, was completed about 11 million years ago (graph).

The switch marks the time when production of North Atlantic Deep Water became continuous and effective in re-arranging the ocean's silica cycle.

From there on, the AA ring of cold deeply mixed water masses was reliably high in silicate content (as it received the deep waters of the Atlantic).

Thus, the AA became a distributor of silicate to upwelling regions after 11 million years ago. Diatom-based food chains became a dominant part of the ocean's productivity pattern. Food chains shortened accordingly and food supply to apex consumers doubled, tripled or quadrupled, with interesting consequences for the evolution of whales and seals and seabirds.

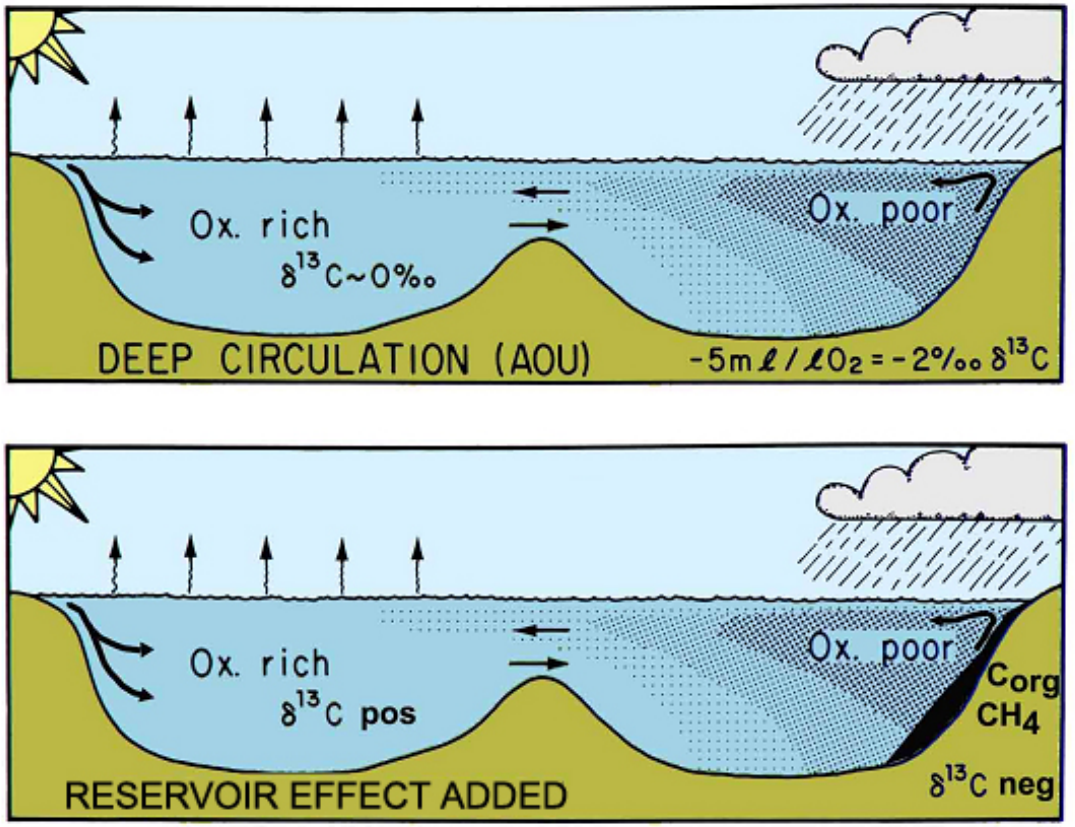
The great Atlantic-Pacific asymmetry that characterizes nutrient distributions (and biogenic sedimentation) of the present ocean was initiated at the time of the silica switch. It signals the beginning of the modern mode of basin-basin fractionation and thus documents the time of first NADW production.



The deep Atlantic feeds nutrients into the AA Ring Current; the deep Pacific is recipient of nutrients from the Ring.

Basin-basin fractionation. The nutrient content in deep waters of the Pacific is markedly higher than that of the Atlantic.

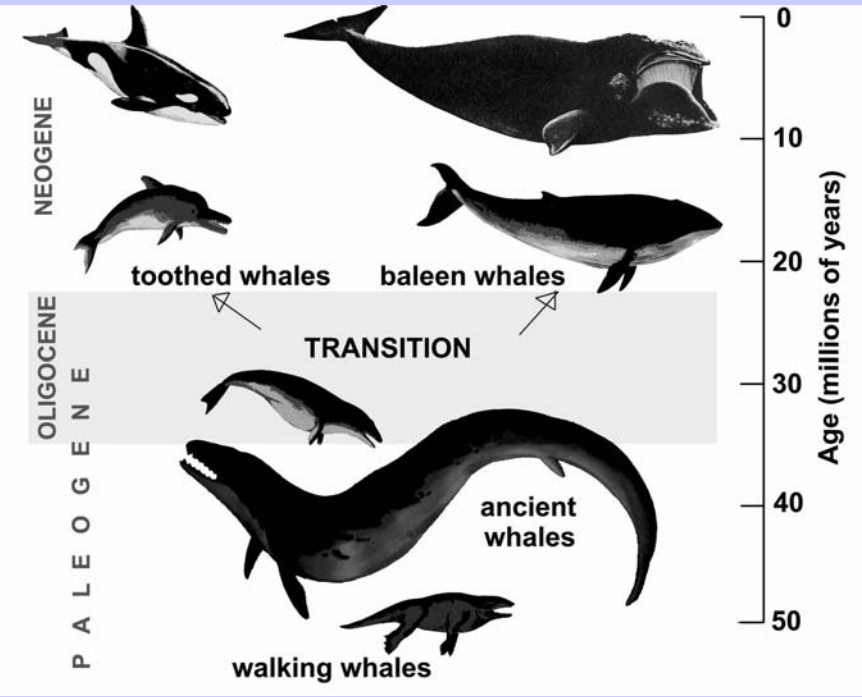
Basin-basin fractionation is reflected in in the carbon isotope patterns. When interpreting carbon isotope records, it is useful to keep in mind several important ways in which C isotopes change in the sea: (1) the surface layer is depleted in C-12, from biological extraction; (2) C-12 is enriched where oxygen content is low, from the oxidation of C-12 rich organic matter; (3) there is a global effect, from building up sediments rich in C-12 (especially when involving buildup of methane ice).



There is an asymmetry effect from (1) and (2) between communicating basins (top panel, graph).

There is a global effect from (3), which is superposed on the asymmetry effect (bottom panel, graph).

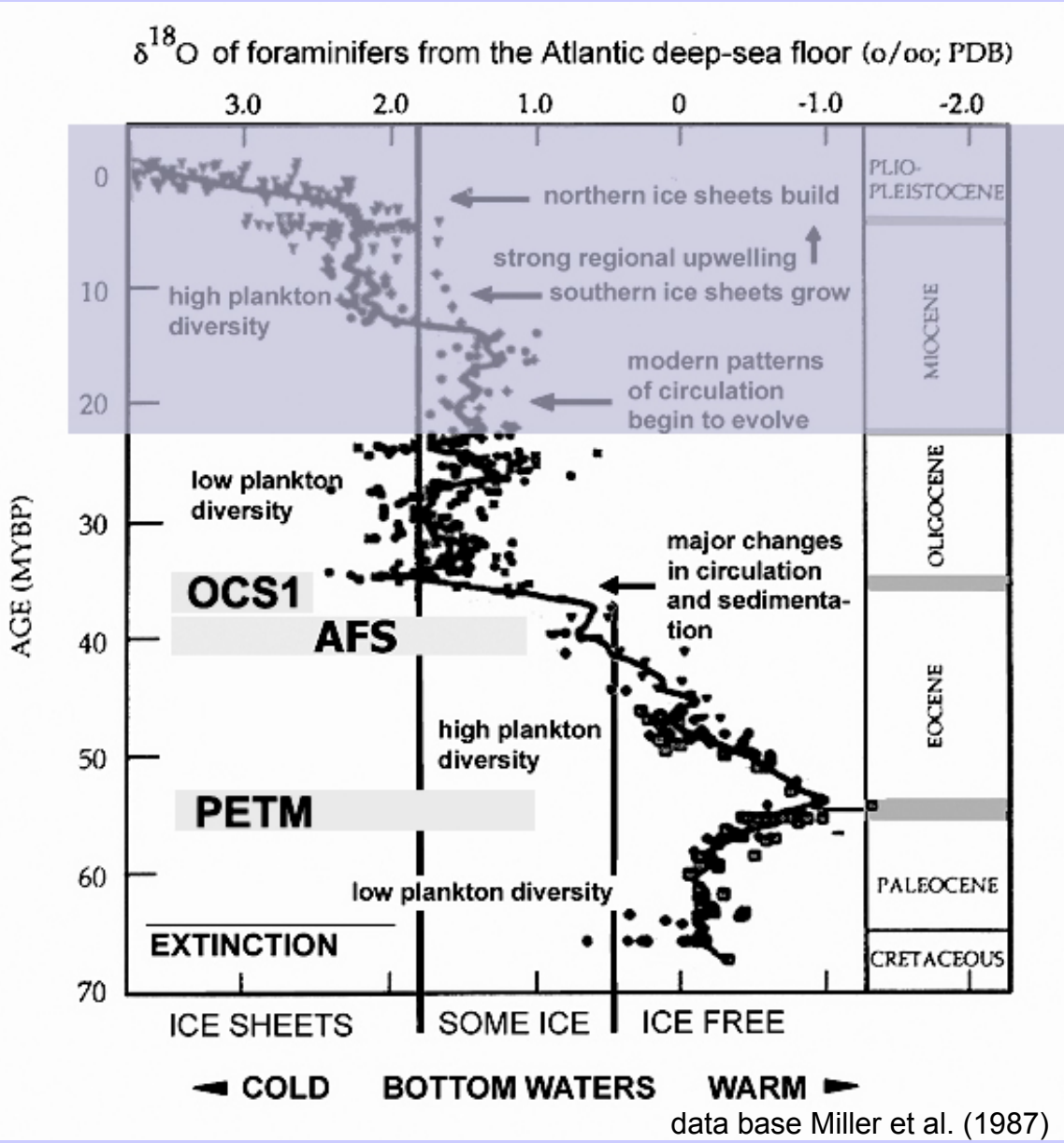
CENOZOIC OCEAN HISTORY: THE PALEOGENE



The Paleogene sets the stage for the transition into the modern ocean with deep cold waters and wind-driven upwelling. It is a period of ancestral evolution of whales, within a warm setting, but with cooling since the end of the Paleocene (55 Ma).

The Paleogene (~65-24), almost two thirds of the Cenozoic, comprises the Paleocene, Eocene, and Oligocene. About one half of the time is taken up by the Eocene, leaving the other half for the other two.

The most remarkable historical events in the Paleogene have to do with the carbon isotope spike at the end of the Paleocene (the *PETM;* that is, the Paleocene-Eocene temperature maximum), the great change in global sediment distribution in the late Eocene (the *AFS;* that is, the Auversian facies shift), and the drastic drop of oxygen isotopes at the end of the Eocene (the *OCS1*; that is, the Ocean Cooling Step 1). The OCS1, initially interpreted largely in terms of production of cold deep water, signals the transition of the planet's climate into the present cold phase, starting with ice on Antarctica.

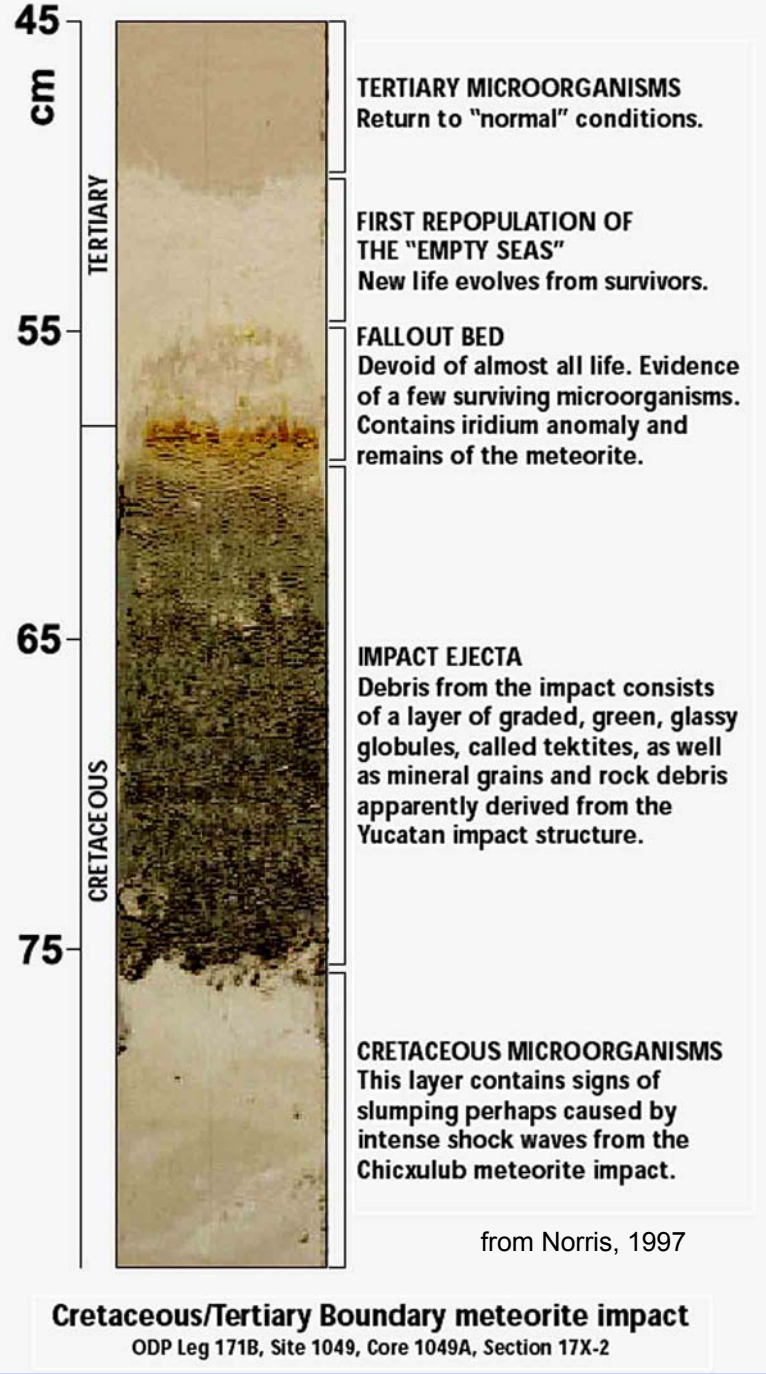


data base Miller et al. (1987)

Three major events in the Paleogene

The theme of a general cooling from a temperature maximum at the beginning of the Eocene, with a dramatic step at the end of that period, is clearly seen in the oxygen isotopes of benthic foraminifers from the Atlantic, in the compilation by Miller et al. (1987).

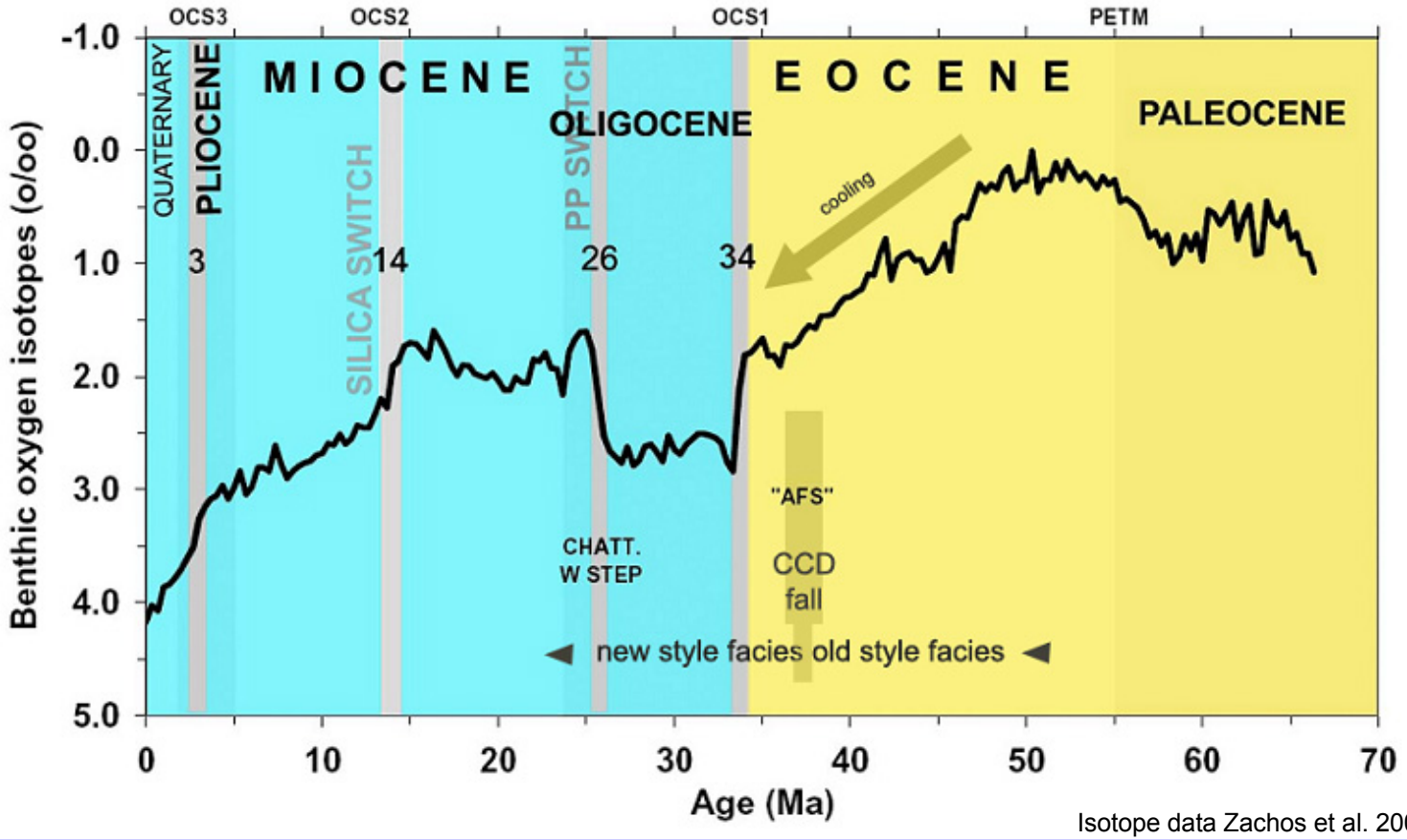
Cooling in the Paleogene thus comprises a general phase in the Eocene, following the PETM, and a major cooling step as a culmination (the OCS1). Just before that step, there is a major shift in sediment distribution (the AFS, Auversian Facies Shift).



from Norris, 1997

Cretaceous/Tertiary Boundary meteorite impact
ODP Leg 171B, Site 1049, Core 1049A, Section 17X-2

The Paleogene begins after the great extinctions at the end of the Cretaceous, an event that is generally attributed to the consequences of an impact by a large bolide coming into Earth from space. Evidence for such an impact has been gathered both in sections on land (e.g., Gubbio in Italy) and through drilling into the sea floor (image to the left).

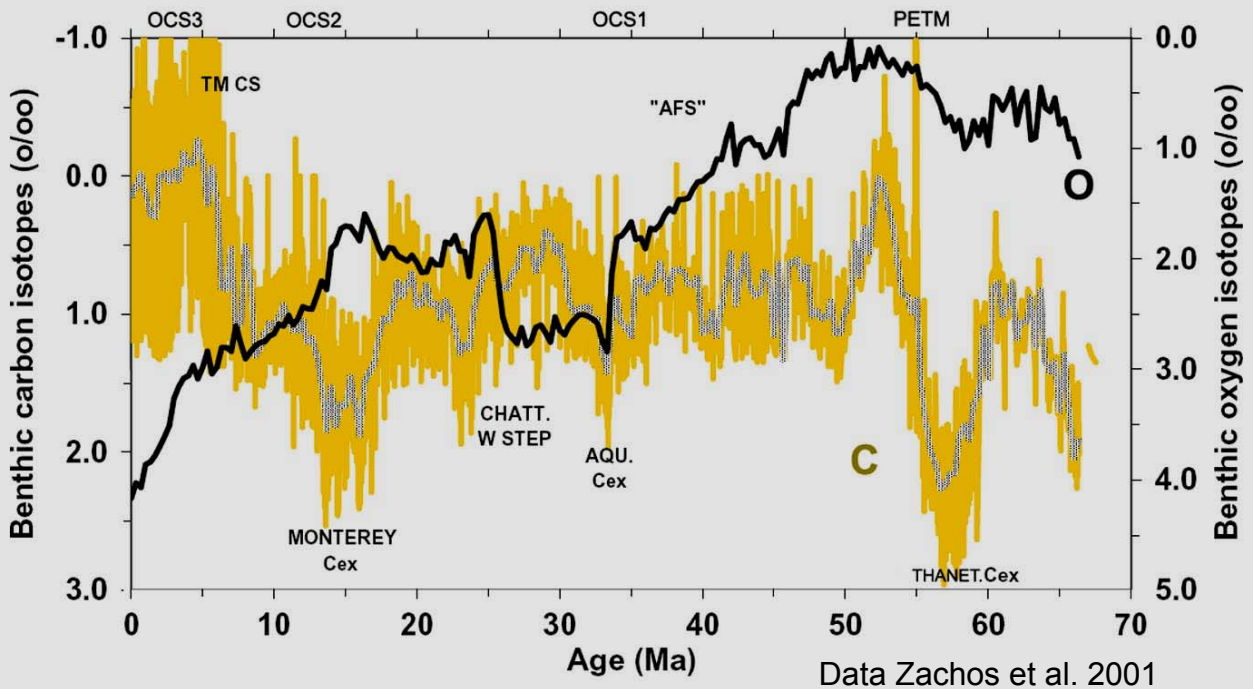


Isotope data Zachos et al. 2001

The Paleogene is a time of transition from a warm-planet state (Cretaceous & Paleocene) to a cold-planet state (Oligocene & Neogene). Cooling occurs all through the middle and late Eocene and ends with a major shift in deep-sea facies (AFS) followed by a sudden drop in temp. (Ox Iso Cooling Step 1) - a threshold event, with strong positive feedback from buildup of ice and snow in AA highlands. The last major event within the warm-planet state, and perhaps the first sign of the coming of a new regime is at the end of the Paleocene, the "PETM."

The PETM (near 55 Ma, Thanetian) is associated with a uniquely strong excursion in carbon isotope values toward low values -- in fact, that is how it was discovered, and that is why it is assumed that there was major release of organic carbon from a sediment reservoir to the ocean-atmosphere system. Dissociation of methane clathrates is commonly held responsible (Dickens et al., 1995; 1997; Dickens, 2000). The associated warming exceeds 5°C.

It is of great interest that the spike toward negative values is preceded by a long-term strong peak in positive values, presumably the inverse of methane release (that is, methane buildup in a reservoir of clathrates). It is not clear that the minor cooling accompanying the excursion of carbon isotopes toward positive values is strong enough to serve as a cause for the buildup.

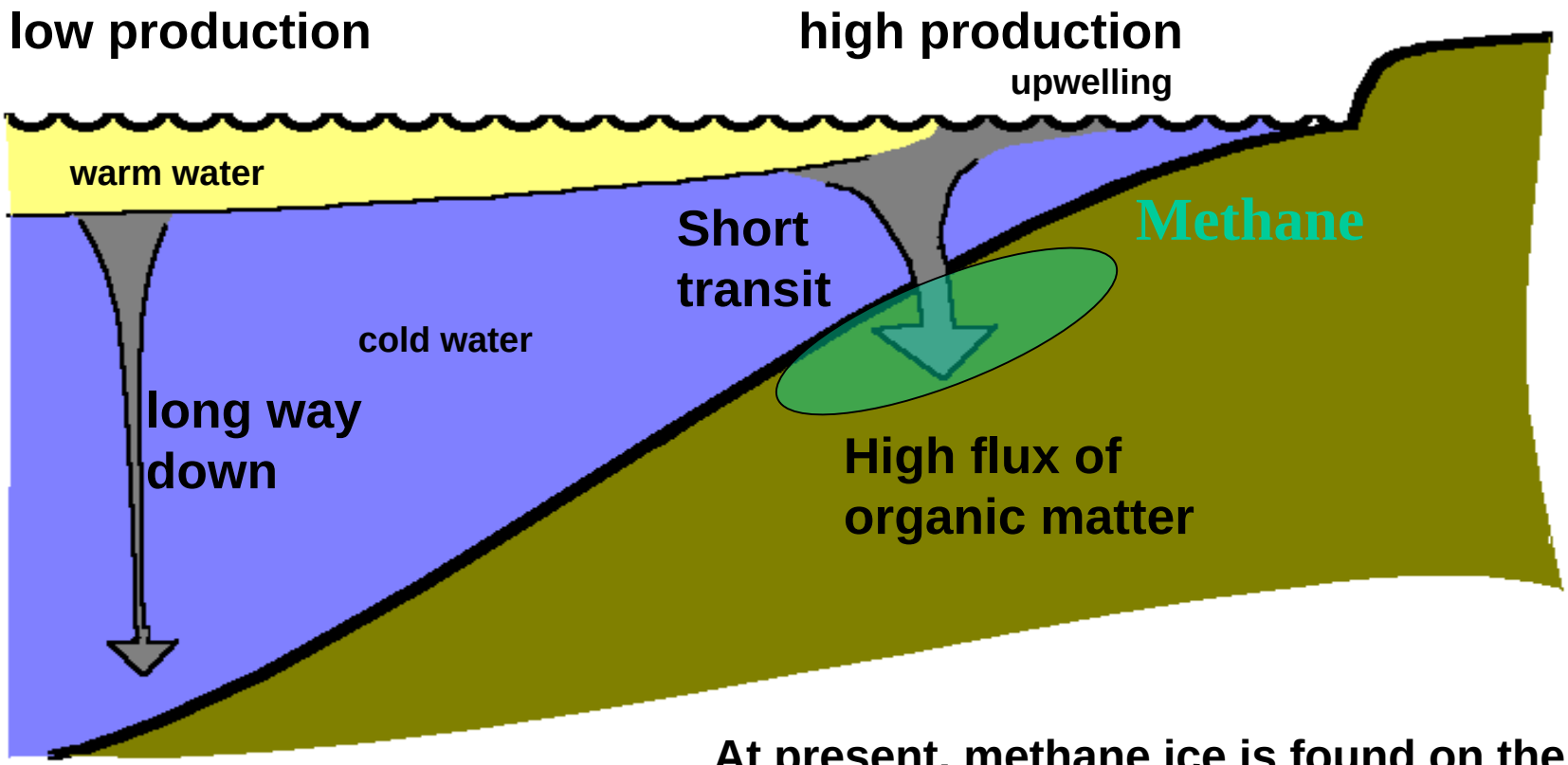


Data Zachos et al. 2001

Subsequent much stronger cooling (after the PETM) seems to have no notable effect in moving carbon indices toward positive values, until one gets to the very end of the Eocene.

Thus, reasons for any buildup and release of methane from clathrates are as yet obscure.

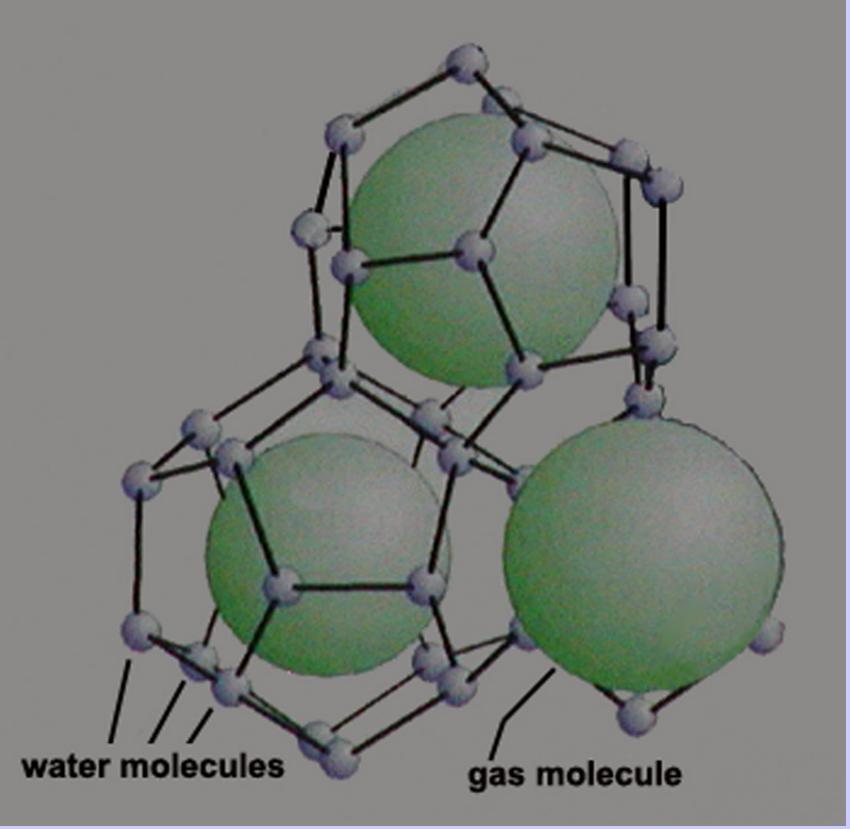
The methane story is relevant to ongoing global warming. Is there a danger of sudden and sustained release?



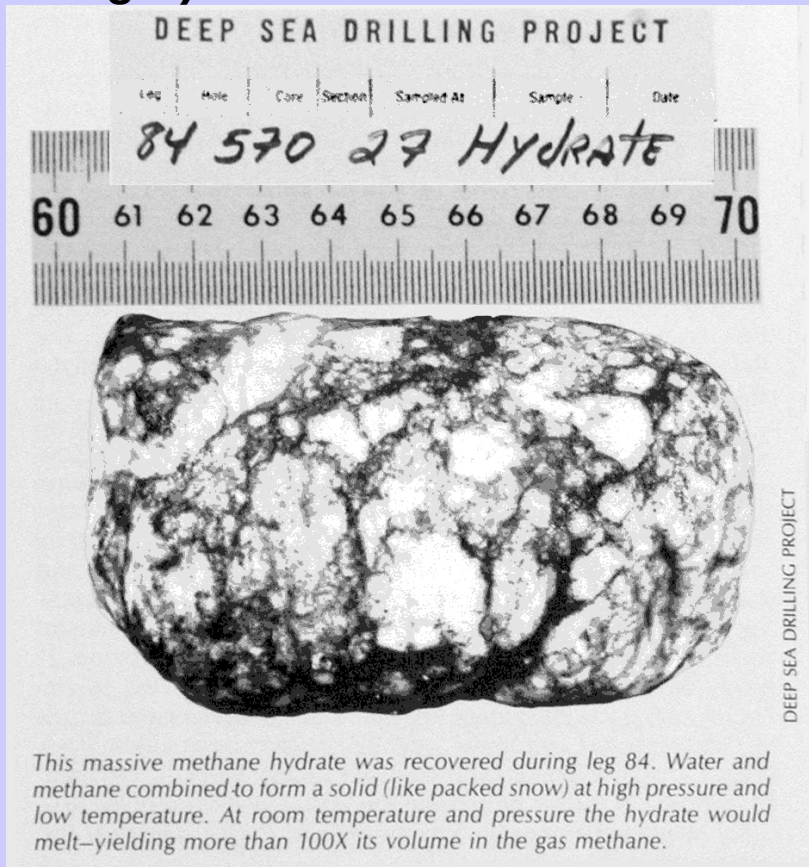
At present, methane ice is found on the continental margins of the world ocean, in areas of high productivity.

Methane, in the ice cores, tends to track carbon dioxide rather well during the ice ages.

Apparently, climate change produces a positive feedback from methane, a powerful greenhouse gas.



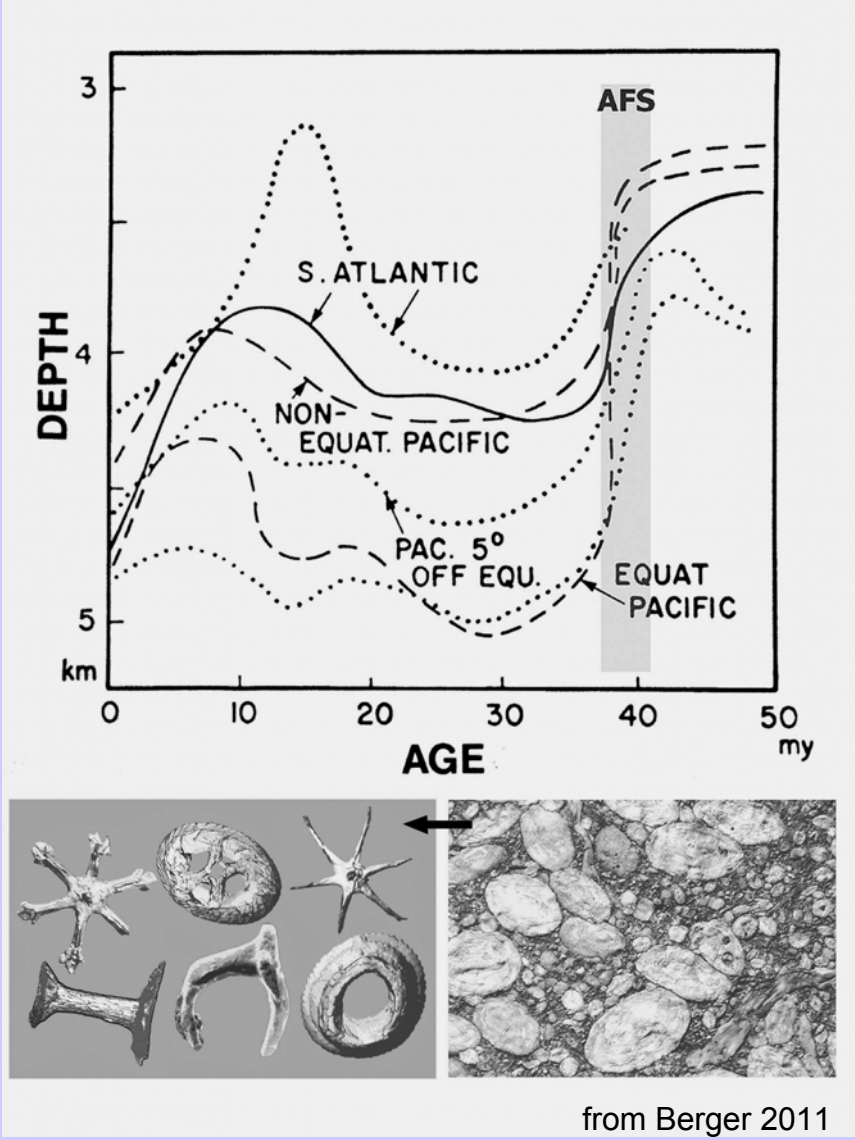
The typical clathrate structure consists of cages of water molecules, cages that contain gas molecules such as methane ("natural gas") (top graph, after E. Suess et al.). Similar structures accommodate carbon dioxide. In margin sediments of the sea floor, the ice looks white, and is commonly embedded in mud (DSDP photo, lower right).



This massive methane hydrate was recovered during leg 84. Water and methane combined to form a solid (like packed snow) at high pressure and low temperature. At room temperature and pressure the hydrate would melt–yielding more than 100X its volume in the gas methane.

When methane ice is lit, the ice melts and the escaping methane burns.

Photo Geomar; courtesy G. Bohrmann



from Berger 2011

A drastic drop in the CCD (the carbonate compensation depth) in the deep sea near the end of the Eocene (here in an older time scale) seems intimately related both to a general reorganization of deep sea sediment facies (AFS, Auversian Facies Shift), and to a shift from carbonate-rich shelf sediments (symbolized by Catalunyan nummulite limestone in the graph) to vast sequences of deep-sea carbonate (symbolized by nannofossils).

One way to reconstruct the timing of the collision between India and Eurasia, assuming it produced major uplift, is from the record of strontium isotopes. The expectation is that the erosion of granitic continental crust increases the ratio of isotopes 87 to 86 in the sea, since Strontium-87 is a product of decay of Rubidium-87, an element enriched in the crustal rocks of continents.

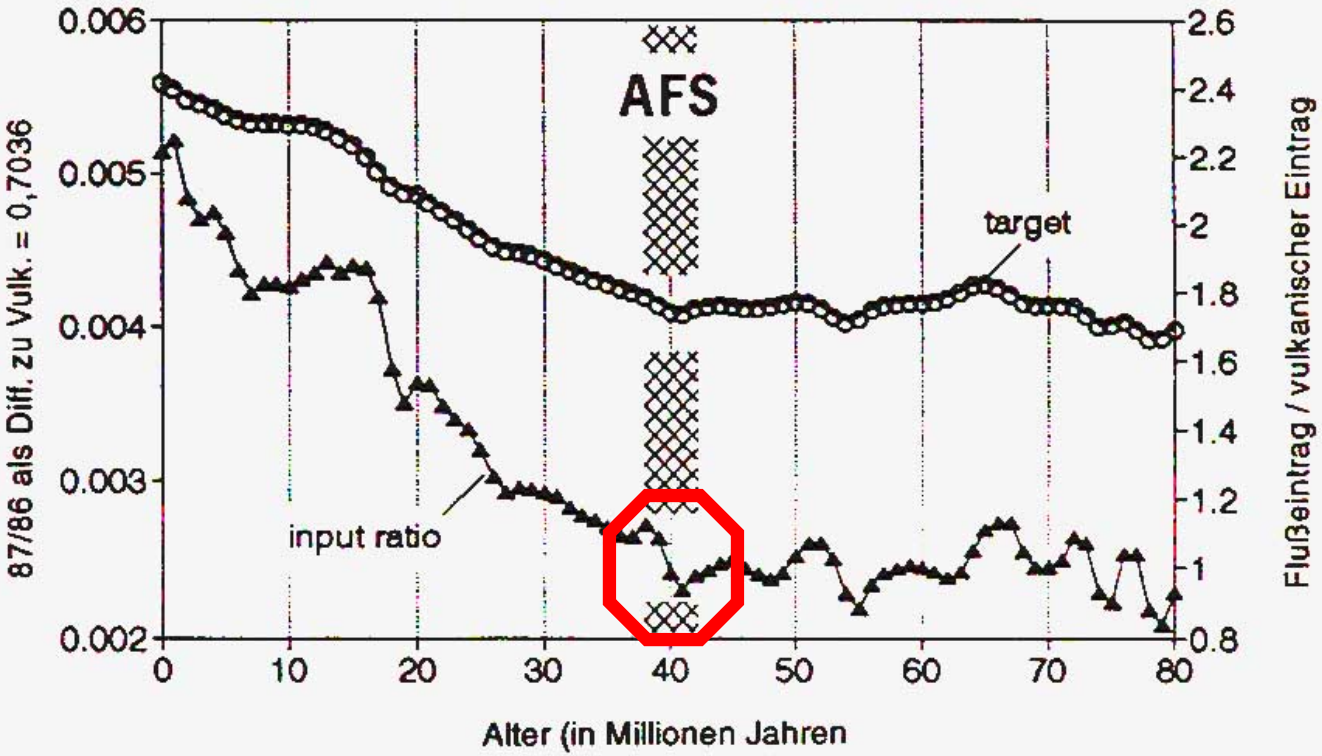
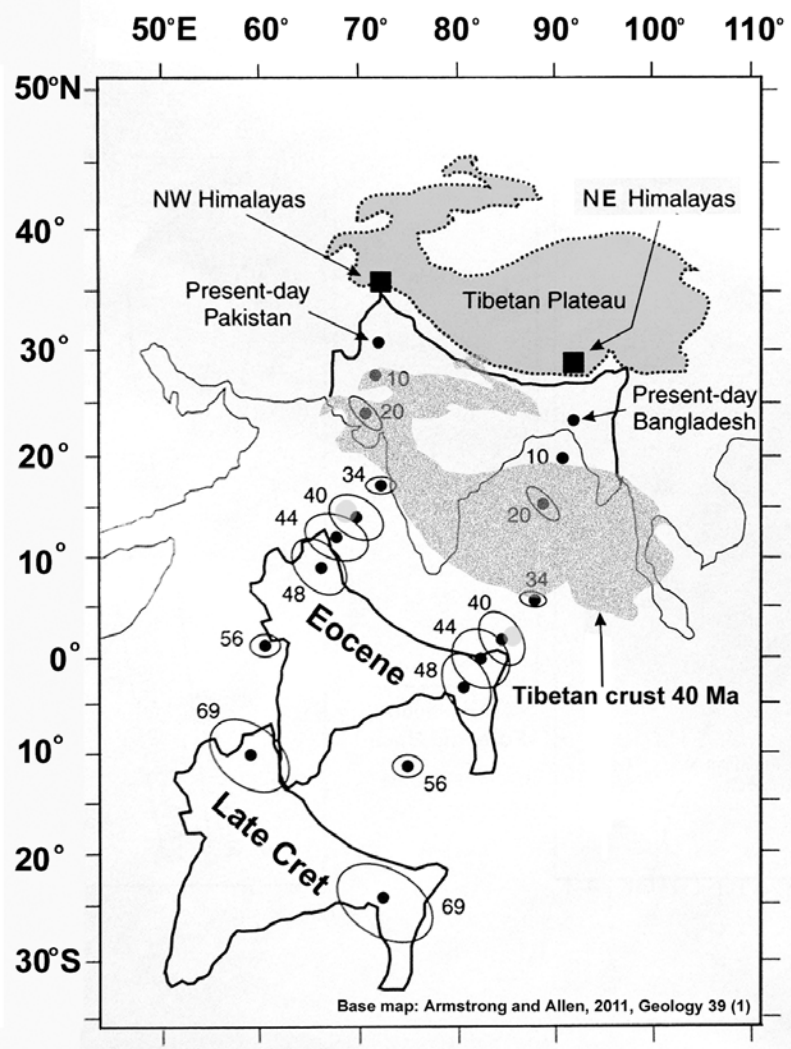


Abb. 3.25. Strontium-Isotopen-Kurve für das Känozoikum aufgrund der Analyse von marinen Karbonaten. *Target* = $^{87}Sr/^{86}Sr$-Verhältnis als Differenz zum vulkanischen (hydrothermalen) Wert von 0,7036. *Input ratio* = Verhältnis zwischen Flußeintrag und vulkanischem Eintrag. dabei wird ein Fluß-Wert von 0,7116 angenommen. *AFS* (Auversian Facies Shift): Beginn der stärkeren Faziestrennung der Tiefseesedimente (aus Berger und Wefer 1996, Daten aus Elderfield 1986)

Available data suggest that a long-term change in the ratio of Sr isotopes entering the sea started in the Late Eocene, roughly coincident with the great facies shift (AFS) seen on the deep ocean floor.

Why did sea level drop and expose shelf carbonates to erosion and redeposition at the end of the Eocene? There are two factors: (1) doubling the crust in the Tethys collision zone, (2) building up large ice masses on Antarctica.

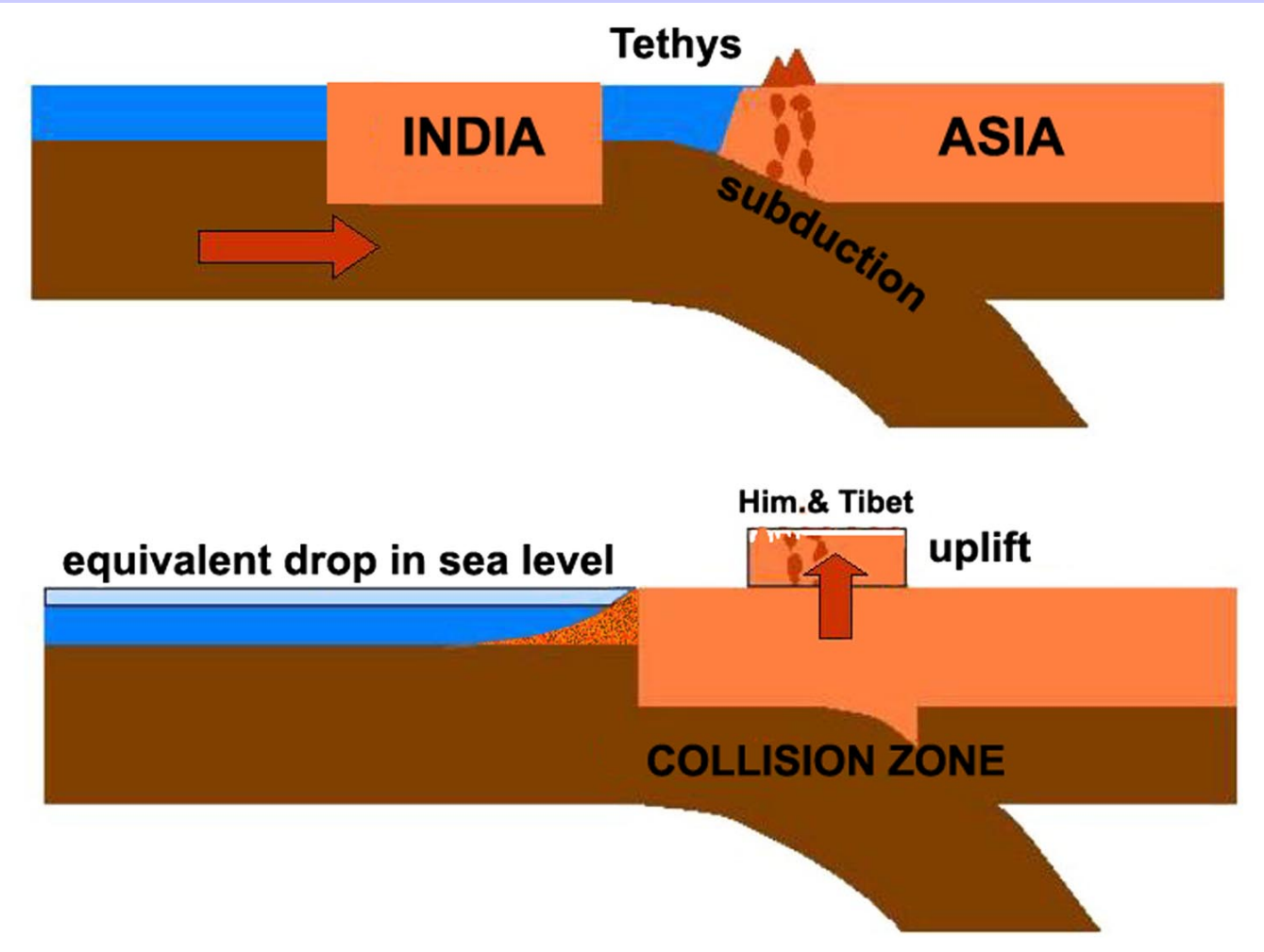


A recent reconstruction of the collision track India-into-Asia suggests that the end of the Eocene was indeed a good candidate for starting a doubling of the crust. At that time greater India first touched the rim of Asia, and its northernmost portion (the future deep Tibetan crust) started its disappearance act.

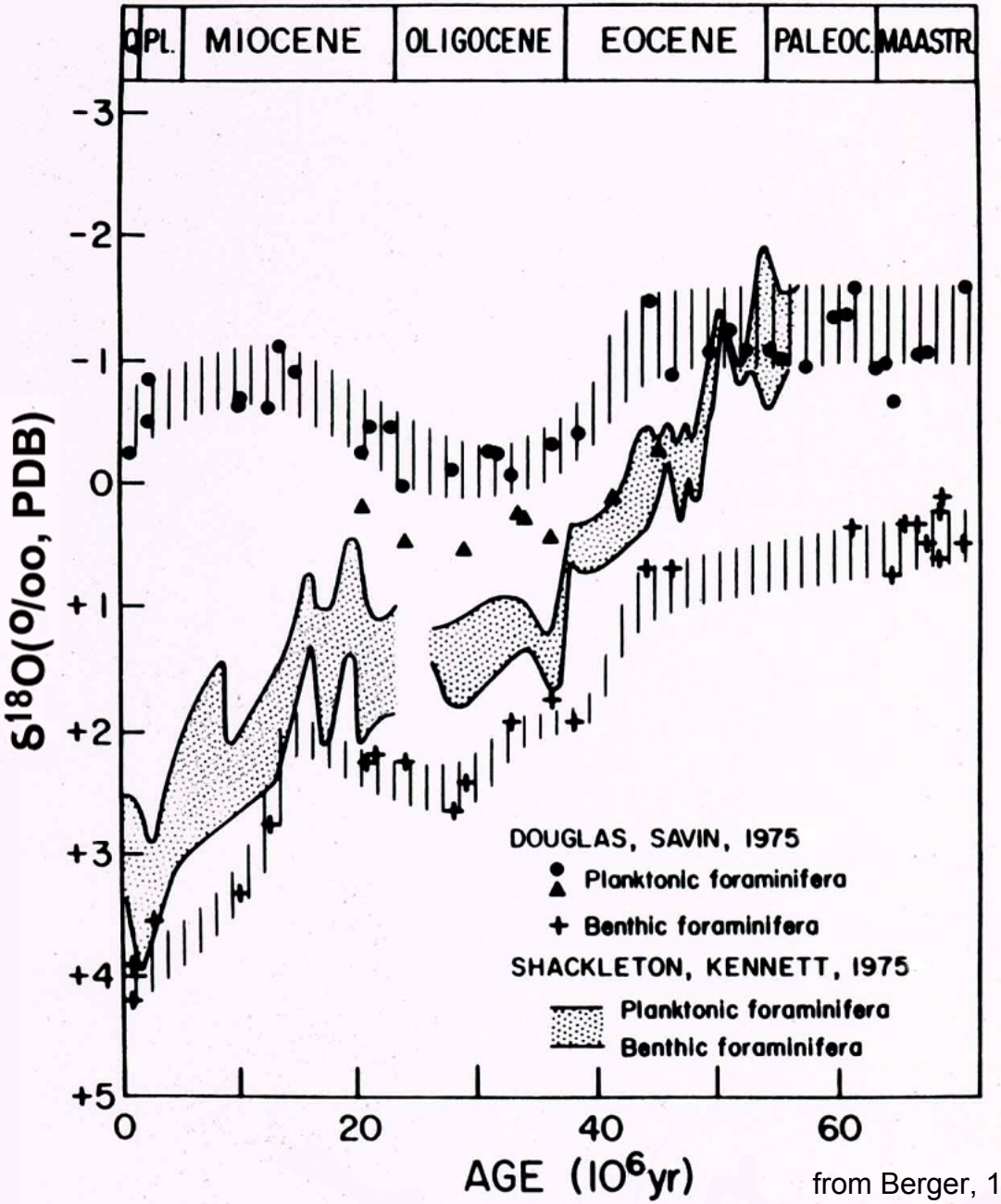
There is good evidence for ice buildup on Antarctica beginning in the late Eocene, from deep ocean drilling in Prydz Bay.

As a footnote to tectonic activity in the Paleogene, late Paleocene collision tectonics in the Tethyan realm might have caused exposure of methane clathrates that formed at great depth, e.g. through landslides.

The basic idea linking mountain building to a drop in sea level is simple: a doubling of the crust produces uplift, and the stacking of the crust results in a deficiency in the sea, which lowers sea level. The main signal in the record are products from erosion. A drop in sea level (roughly 50 m) changes the albedo of the planet: water is dark, shelf carbonates are bright.



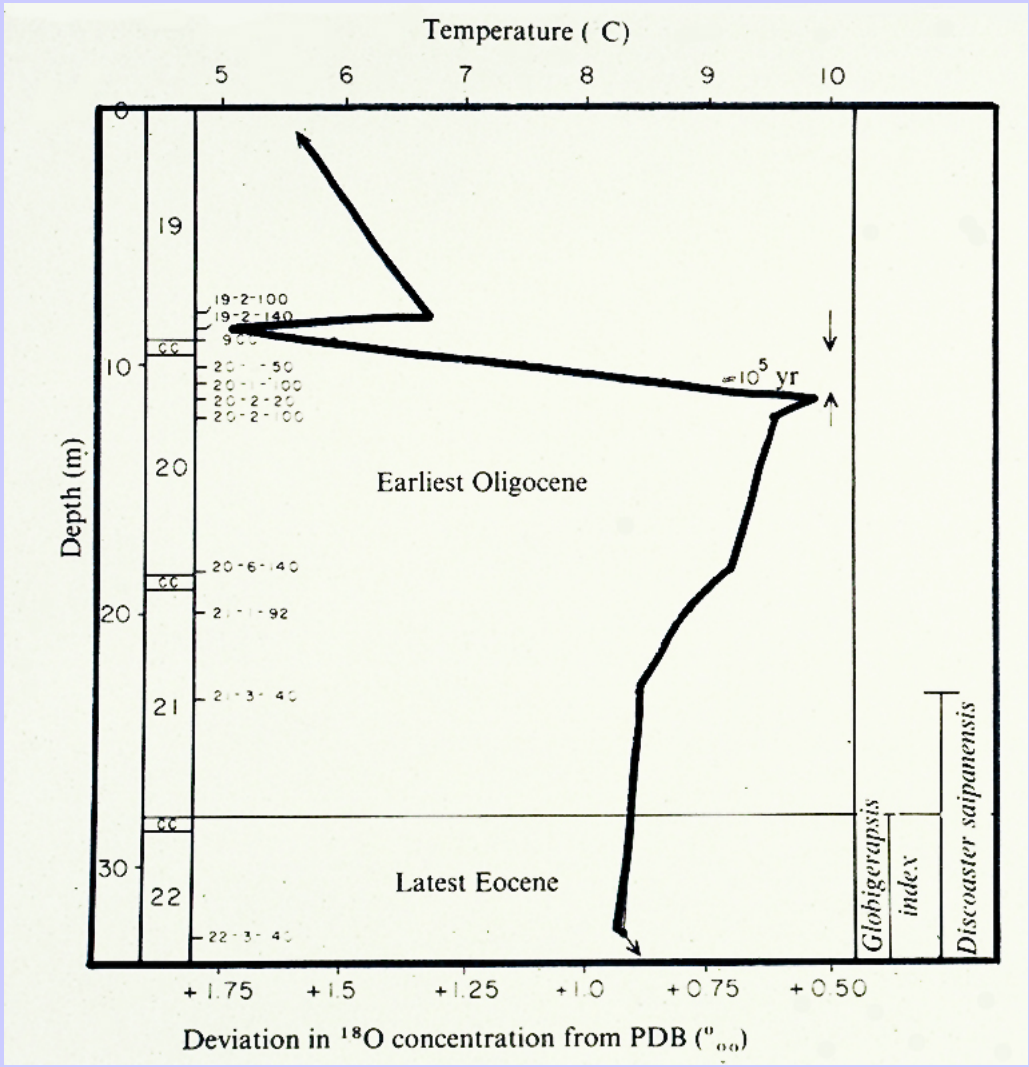
The change in albedo alters the radiation balance in favor of cooling.



from Berger, 1979

The End-of-Eocene cooling step was first seen in the high-latitude data published by Shackleton and Kennett (1975), a series that also established the general Cenozoic cooling trend, together with the information from Douglas and Savin (1975).

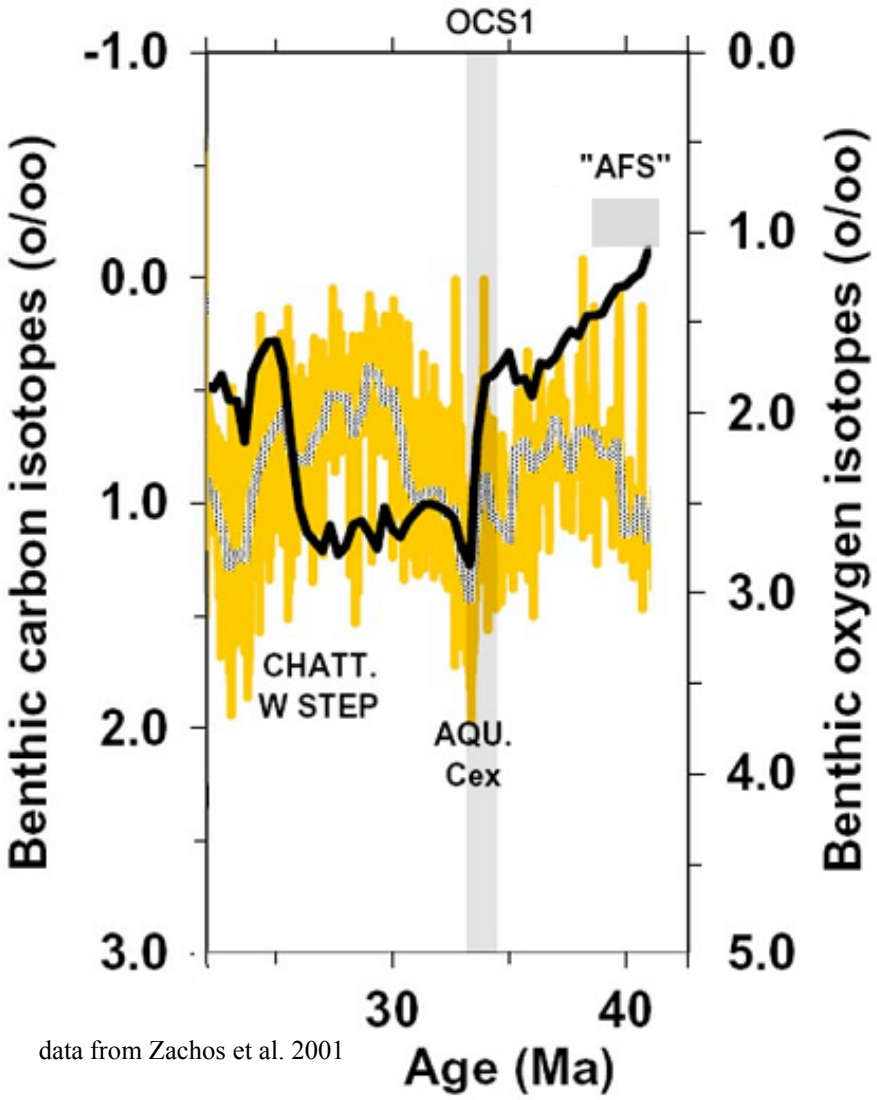
Comparison of the data sets showed that the planetary cooling was largely a high-latitude phenomenon, with low-latitude temperatures being much less affected.



from Kennett and Shackleton, 1976

Kennett and Shackleton (1976) put the End-of-Eocene T-shift (OCS1) into the earliest Oligocene, and interpreted it largely in terms of renewal of cold bottom waters from high-latitude sources. They proposed a cooling of around 5°C within the time span of ~ 100,000 years (graph).

Apparently, the cooling step at the beginning of the Oligocene was associated with a buildup of negative C storage (Aquitanian C excursion; methane ice?).

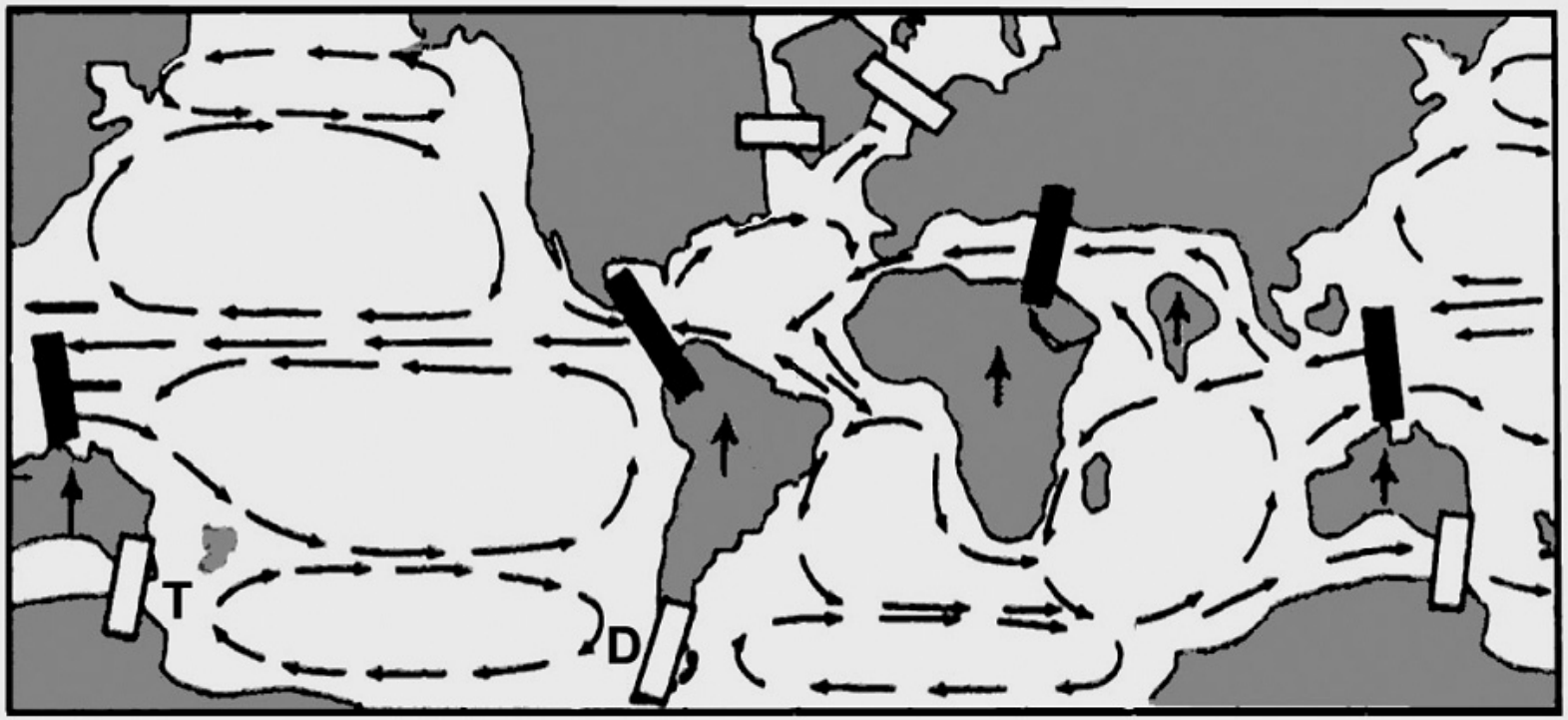


data from Zachos et al. 2001

When taking a closer look at the isotope data at the time of the major shift into the new “icehouse” regime (AFS and First Oxygen Cooling Step), we find that the entire Oligocene is unusually depressed, as far as deep ocean temperatures, and that there is a buildup of negative C stores precisely at the time of the major cooling step into the Oligocene.

The Oligocene ends with the Chattian warming step (unexplained). Interestingly, this warming is not associated with a negative carbon isotope spike (the absence being unexplained).

The closing of the Tethys seaway has many ramifications besides a drop in sea level and ice buildup on Antarctica, and the creation of a coldwater sphere in the deep ocean.

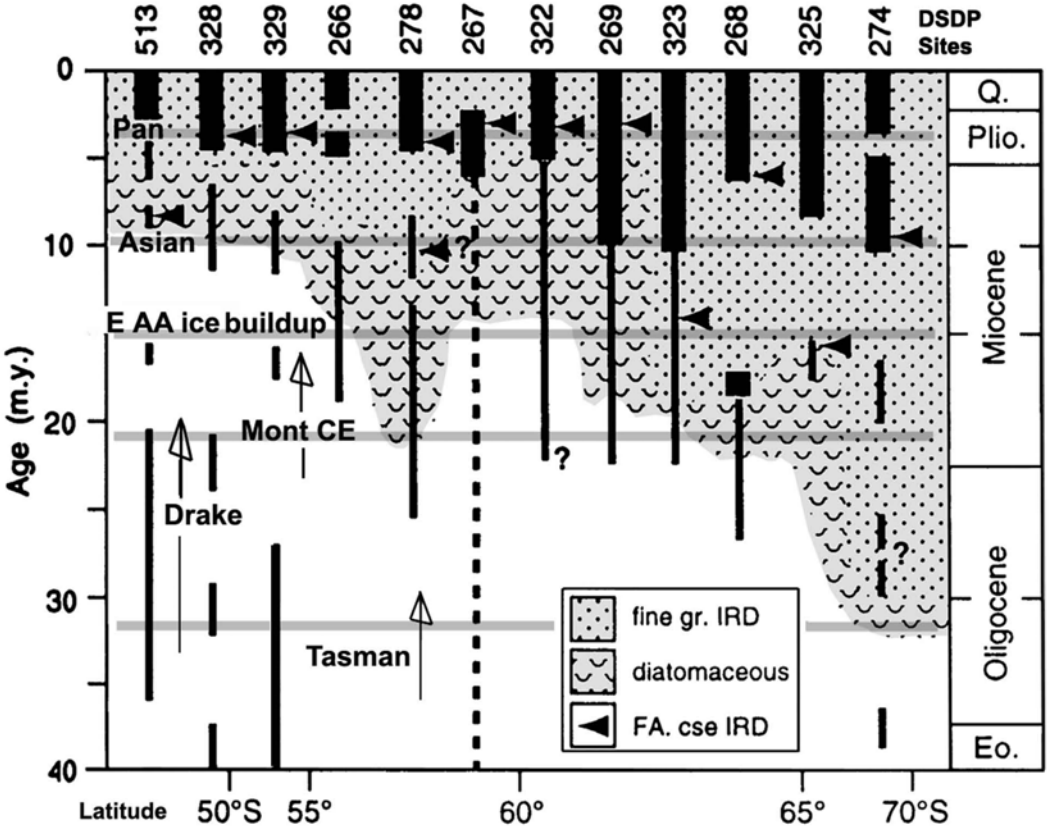


adapted from Haq (1981)

The rise of the circumpolar Ring Current is intimately linked to the replacement of tropical gateways by ocean gateways in the southern ocean (black bars, closing in the Cenozoic, white bars, opening).

In the Southern Ocean, the crucial choke points are the Tasmanian gateway (T, opening in the Eocene) and the Drake Passage (D, opening in the Oligocene).

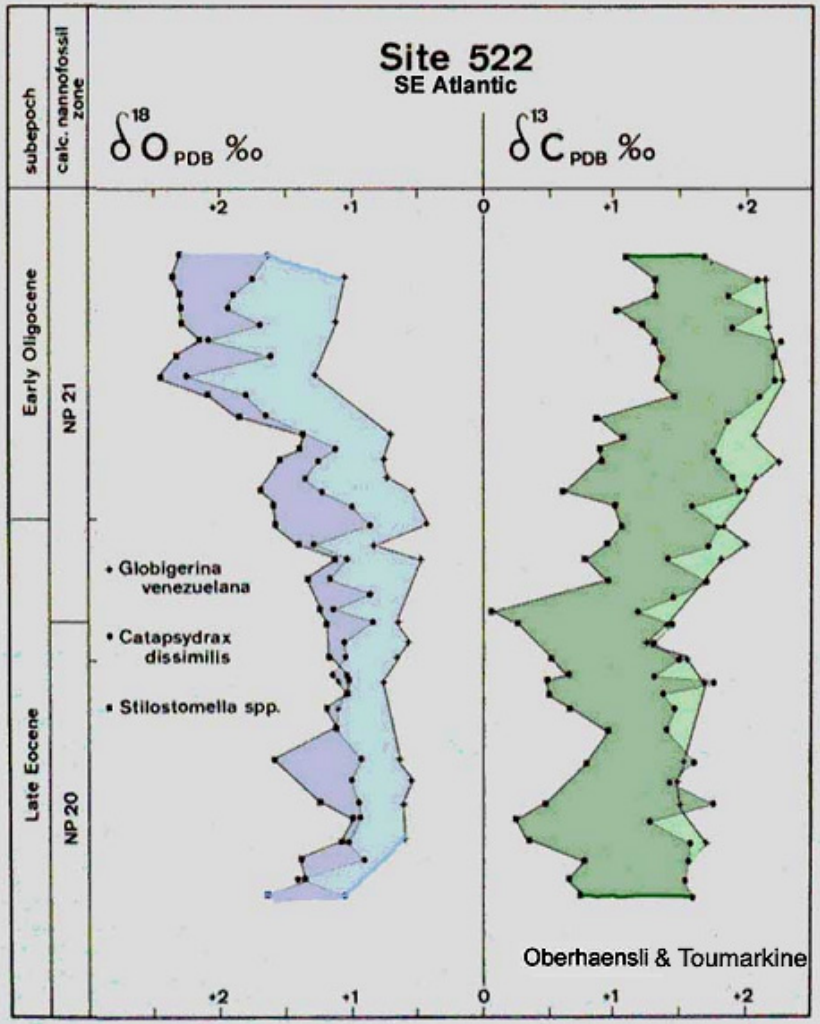
As far as the productivity of the world ocean, the fact that the Great Ring Current circling Antarctica became a major site of deep upwelling and a silica trap and silicate distribution center is the most important development in the Oligocene. It set the stage for major diversification of plankton and marine organisms in general, in the Neogene.



Ciesielski & Weaver 1983, modified

Facies developments in deep-sea sediments from around Antarctica support the conclusion that diatom deposits first arose in the Oligocene, suggesting the influence of upwelling in a coastal belt next to the Antarctic.

Among the various developments in physical oceanography resulting from the Oligocene revolution, none are more important than the evolution of the thermocline. Apparently the mixed layer greatly thickened in the early Oligocene, which made for poor conditions of production in the sea, owing to violation of the Sverdrup condition (which says that green organisms, to be productive, have to spend more time in the light than in the dark).



Stratification in upper waters are investigated by noting isotopes in various species of planktonic foraminifers (graph). However, the interpretation of results is quite difficult.

In summary, what is important about the history of the ocean (and highly relevant to future developments) is the evolution of the distribution of heat, and those elements of the circulation that greatly influence nutrient cycling and productivity.

The crucial items are the evolution of the great central gyres, of the replacement of deep waters and the vertical stratification in general, of upwelling and of the role of the Great Ring Current as a governing factor in nutrient recycling. In all of these, the wind field is of the essence.

All of the crucial items respond to tectonic forcing on long time scales, and to astronomic forcing on short ones. Evolution, in addition, has to cope with mass extinctions; that is, instantaneous deteriorations of life-sustaining conditions.

So far, Life has coped well.

The End

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