Dating of Sediment Cores:

If we have drilled a deep core of sediment from the ocean, how do we get the timing? How does one relate depth in core to time in the past? What do you think?

1) One way is to assume a constant sedimentation rate tagged down periodically by magnetic reversals whose dates are known, or that at least give a way of comparing cores with different sedimentation rates(Berggren, et al. 1985). How deeply into past time we can probe is governed by how deep a core we can dig out and how fast the sediment accumulated. To get high time resolution one must drill where sedimentation rate is high, but to get deep history with poorer temporal resolution one must drill where the sedimentation rate is slower.

2) Another way is to use the radioisotope 14C. But this only takes you back 40,000 years or so and is inexact because the source of 14C varies with solar activity and Earth's magnetic field.

3) Other isotopic decay methods like U/Th, or K/Ar.

4) Tune the time scales constrained by infrequent magnetic reversals with orbital parameter periodicities. This has the disadvantage of presupposing the orbital parameter theory for climate is essentially correct.

The Cretaceous Period (144-65Mya):

The Cretaceous Period is very interesting climatologically because the climate was warm everywhere from the equator to the poles. Dinosaurs ranged above the Arctic Circle, for example, and large trees and ferns were there too. Dinosaurs ruled the earth and reptiles were the largest marine predators. The sea level was higher, and the continents had shallow seas on them. The interior of the USA was flooded, for example. During this period the continents were beginning to move toward their present positions, but not exactly. The Atlantic ocean was widening as Gondwana broke up. There were no maritime glaciers, so the deep ocean was probably stagnant. This period was marked by black shale deposits which are an accumulation of organic-rich black muds in deep water, where the ocean was probably anoxic. Since there were no maritime glaciers to provide a source of cold water at the surface, the thermohaline circulation was probably not operating.

During this period plankton in the ocean diversified and diatoms and coccolithophores became abundant. Coccoliths (shells) accumulated in large numbers on the ocean floor forming chalks, such as the white cliffs of Dover, England. The word Cretaceous means 'chalk bearing'. Most of the top predators at this time have since become extinct. During this period flowering plants first appeared, and conifers became the dominant gymnosperm. There was a rapid rate of insect diversification during the Cretaceous. The mass extinction at the end of the Cretaceous wiped out the dinosaurs, ammonoids (having a coiled shell like a nautilus), large marine reptiles (plesiosaurs and monosaurs) and rudists(a kind of asymmetrical reef-building bivalve) and many other invertebrate taxa. Drastic reductions in the number and diversity of small marine organisms such as Coccolithophores, planktonic foramanifera, radiolarians, and belemnoids (dart-shaped, like a squid). The extinction affected both sea and land species. On land, only small animals weighing less than about 50 pounds survived. There are many theories for these extinctions, including; meteorite impact, climatic change, interspecies competition, disease, etc.

Cenozoic Decline:

So lets' take a look at a really slow sedimentation rate, where we can go back about 70 million years and look at $\delta^{18}O=[^{18}O/^{16}O\text{sample}/^{18}O/^{16}O\text{standard} -1]\times1000$. What do we see? $\delta^{18}O$ is increasing with time over this period, especially during certain intervals. This is from a paper by Raymo and Ruddiman(1992). Shown is a record of $\delta^{18}O$ from ocean sediment cores reconstructing the era from about 70 million years ago to present.

Early in the record during the Paleocene about 60Mya, we know it was warmer because of evidence of subtropical flora in Alaska and such things.

About 55Mya during the Eocene, we see evidence in the $\delta^{18}O$ and other records of a marked period of cooling.

Further cooling followed until about 36Mya, during the Oligocene, when the first major ice event occurred in Antarctica. Still a temperate climate in Alaska.

Then during the last several million years, there was a decline to the glacial conditions of the Plio-Pleistocene, which we are still experiencing

Theories for this cooling period.

1. Continental drift – Piling ice on the pole: Continents drifting toward poles may foster the development of an ice age. Polar continents provide a platform for continental ice in refrigerator of polar regions. Problem: Continents didn't drift much in last 70M years.

2. Continental drift – Oceanic Pathways: Critical ocean pathways or shallow seas can be opened or closed by continental drift. For example, when S. America separated from Antarctica and made possible the circumpolar current, this greatly reduced the ocean heat transport to high latitudes. This could have led to cooling of Antarctica and buildup of the Antarctic ice sheet. A counter argument is that Antarctica was probably already cold enough for ice to form. Further cooling just reduces the supply of snowfall. GCM experiment suggests that more heat transport leads to more snowfall (Oglesby and Park 1989). The Isthmus of Panama is also considered to be a critical geographical feature for climate. It formed at some point and cut off the flow of water to the west. This strengthened flow of Gulf Stream. Its formation seems to preceed glaciation by 0.5 to 2 MYr. (Raymo and Ruddiman 1992)

3. Uplift and Erosion – The rise of the Himalaya: Pretty good evidence suggests that the Tibetan Plateau rose up during the Cenozoic. Most of Tibet was below sea level before about 70My. Big changes hypothesized for around 8±3Mya.(Molnar and England 1990, Molnar, et al. 1993). A change in the Indian Monsoon circulation is also evident at that time. What effects could this have on global climate.

1) Effects on global circulation and climate. GCM experiments. The rise of the Himalaya affects not only Asia, but also Mediterranean dryness fed by subsidence caused by Tibetan Plateau (Rodwell and Hoskins 1996). Adding Tibet to a GCM produces some changes that are consistent with observed paleoclimatic changes, e.g. more regional variation, more dry areas, some high latitude cooling, but not quite enough (Kutzbach, et al. 1997, Rind, et al. 1997, Ruddiman, et al. 1997a, Ruddiman and Prell 1997, Ruddiman, et al. 1997b, Ruddiman 1998).

2) Effects the Indian Summer Monsoon - puts more rain on land: A lot of warm rain in uplifted crustal rock affects weathering which can influence atmospheric CO2. On time scales of millions of years, atmospheric CO2 levels are controlled by the balance between the input of CO2 from volcanoes and the rate of chemical weathering of rocks at the earth's surface. Weathering reactions can be simplified to:

$$CaSiO_3 + CO_2 \rightarrow CaCO_3 + SiO_2$$

to the right is weathering, to the left is metamorphosis

Weathering: Rock plus carbon dioxide goes to calcium carbonate plus silica. Calcium carbonate goes into the ocean and is fixed in shells of small animals.

The reverse process is *metamorphosis*. Silica and calcium carbonate under pressure and heat are metamorphosed back into rock in the mantle.

So you could get variations in atmospheric CO2 either through variations in CO2 input, or through variations in weathering. So if volcanoes declined over this period, we could explain the temp decline. Or if weathering increased. In the first case, weathering would be high when CO2 was high, in the latter case weathering would be high when CO2 was low.

4. Sea floor spreading and CO2: Increased sea-floor spreading gives more volcanism/CO2 release, greater sea area (rising of ridges), goes to less weathering and more CO2 to explain Cretaceous climate (Berner, et al. 1983). Problem: decline is too soon 100-50 Mybp rather than 60-0Mybp. An alternative is to say that weathering is really proportional to relief, which exposes new rock at a greater rate, than to area (Raymo and Ruddiman, 1992). Then mountain uplift is an important factor in global CO2 budget. More silicates rather than carbonates, more rain, more rapid exposure.

Himalaya are the big player in weathering. Data show that 25% of the dissolved sediment load reaching the oceans comes from 5% of Earth's land surface. Tibet gives a large mountain near a warm ocean, and the heavy rainfall over the mountains during the summer monsoon gives weathering maximum.(see topography of Tibet).

Magnetic anomaly estimates suggest that the Indian subcontinent struck Eurasia about 10Mya. This is about the right time to explain the final stage of the Cenozoic decline (Molnar and England, 1993). So it seems like the raising of Tibet could be a cause of the late Cenozoic decline, but something else must have contributed to the early decline out of the Cretaceous.

Evidence of Monsoon switch-on:

G. Bolloides from Ocean Cores

Currently there is a very strong Asian monsoon. One of the places that this is reflected is off the Somali coast. During NH summer strong winds blow along the Somali Coast (~20N, 60E) from the SW at low levels (Hartmann(1994) fig. 6.19) . These surface winds drive the surface drift to the SE, and cause strong upwelling along the coast of Somalia. This upwelling brings cold nutrient-rich water to the surface. During the winter, the winds are reversed and the upwelling stops. Certain small organisms with shells are well adapted to this upwelling environment and their population swells when upwelling occurs. One example is Globigerina Bolloides. Current data show large increases in G. Bulloides in the Arabian Sea during summer when upwelling occurs compared to winter. So current data that indicate G. Bulloides is a good indicator of upwelling and this in turn is a good indicator of the winds associated with the summer monsoon caused by the heating of Tibet.

One can then consider a long record of G. Bulloides fractional abundance from ocean cores. Sure enough, their fractional abundance increases about 8.5 Mya at about the time we think that that Tibet shot up a lot. Thus monsoon onset was rather abrupt.

Some Background Material on Photosynthesis

Photosynthesis can be represented by the chemical reaction:

$$6CO_2 + 6H_2O + PAR \dashrightarrow C_6H_{12}O_6 + 6O_2$$

So plants take in CO_2 and water and make glucose plus molecular oxygen as a byproduct. The water normally comes up from the roots, but the CO_2 must be gathered from the air through pores in the leaves called stomata. The H_2O molecule is lighter than the CO_2 molecule, so when the stomata are open, more water diffuses out than CO_2 diffuses in. This is particularly true when the sun is shining, since then the leaves are warmer than the air and the vapor pressure inside the leave is greater than that of the air. Lots of water is given lost in the process of taking in the CO_2. This is called evapotranspiration. In conditions where water is scarce, this can be a problem.

Plants have developed several adaptations to overcome the problem of water loss. First, leaves absorb photosynthetically active radiation (PAR) efficiently, but reflect much of the near IR radiation from the sun that cannot be used for photosynthesis (see chapter 3 of Hartmann, 1994). Plants can close their stomata when conditions are excessively dry, such as during a drought, or in the heat of the day, and then open them when conditions are more favorable. They have developed numerous leaf adaptations for conserving water such as hairy, textured or waxy surfaces. In CAM (Crassulacean Acid Metabolism) photosynthesis, some plants take in CO_2 during the night, store it, close their stomata, and still do photosynthesis during the day. This type of photosynthesis strategy is mostly found in desert succulents. The two most common forms of photosynthesis are C3 and C4 photosynthesis. C4 is a more recent adaptation and is more efficient in its use of water.

C4 plants fix CO_2 into 4-carbon carboxylic acid, which reduces the effective concentration of CO_2 within the leaf, creating a steeper gradient for CO_2 into the leaf compared to the gradient for H_2O out from the leaf. The result of this is higher water use efficiency. C4 plants can do as much as twice the fixation of carbon per unit of water used than C3 plants. The downside is that C4 photosynthesis becomes inefficient in shady sites or at low temperatures. So factors that favor C4 plants are: high PAR, high temperature, low water availability(at least at some time of the life cycle), soil salinity.

C3 plants: Wheat, rye, field beans, French bean, white clover, alfalfa, oaks, beech, birch, pines.

C4 plants: Corn, sugar cane, millet, sorghum, amaranths

Cerling et al.(1998) have proposed that the decline of C3 plants and the ascent of C4 plants during the last 65 million years was related to CO_2 starvation of C3 plants. They argue that this has had an effect on the evolution of mammals during this period. The relative abundance of C3 and C4 plants was inferred for the ^{13}C in mammoth teeth. C3 plants incorporate more CO_2 per photon than C4 plants, but C4 plants avoid a photorespiration step. At low CO_2 concentrations, the quantum yield of C3 plants is reduced by photorespiration, in which O_2 is absorbed and CO_2 is lost. High temperatures and low CO_2 lead to increased photorespiration in C3 plants. About 8Mya, CO_2 decreased, C4 plants became more numerous, significant changes in fauna occurred, and the climate got colder and drier. How are these changes related? Cerling, et al(1998) say that CO_2 starvation of C3 plants was important. They argue that the rise of C4 plants between 8 and 6 Mya could have been caused by CO_2 starvation as much as by climate changes. The change in plant abundance changed the fauna, because herbivores tend to specialize in certain plants. C4 plants do better at warmer temperatures because of their more efficient use of water.

Because of their different photosynthesis pathways, C3 and C4 plants have different selectivity for isotopes of carbon. Because they lower their internal CO_2 vapor pressure, C4 plants are more likely to take in the heavier carbon isotope than C3 plants.

Sediments and Paleosols and Asian Monsoon onset

More evidence of monsoon onset comes from 1) increased silaceous sedimentation rates in the Indian Ocean, about 8-10Mya. Also paleosols (ancient soils) from Pakistan show an increase in $\delta^{13}C$ in pedogenic(generated in soils) carbonates. There are two main types of plants C3 plants and C4 plants.

C4 plants use C4 photosynthesis and absorb more ^{13}C than C3 plants. These types of plants are grasses and herbs that favor warm growing seasons.

C3 plants use C3 photosynthesis and fix less 13C. C3 plants include most trees and shrubs and grasses and herbs that favor colder growing seasons.

Paleosol data show an increase in $\delta^{13}C$ around about 5-10 Mya, which suggests a change to C4 plants at this time. This could be explained by the monsoon rains, which shifted the growing season from cool temperature of winter to warm, wet summers that occurred after uplift. Without the topography of the Himalaya, India/Tibet would be a subtropical semiarid lowland, with maximum precipitation in winter, rather than summer.

The last 3 million years of ice volume record

OK, next let's go to finer resolution of the last 3My (Raymo 1992) Let's take a look at a 3Myr record of $\delta^{18}O$ from site 607 at 41N, 33W, in 3424 meters of water just west of the mid-Atlantic Ridge. The samples were derived from benthic forams, Cibicidoides and Uvigerina. This record is primarily a record of global ice volume, although ocean temperature also enters. What do we see:

1) Global ice volume has almost always been greater than at present during the last 2.5 Myr. Sea level was about 120 meters lower 20,000 years ago. We are poised at the extreme warm end of a climate system that has persisted for 2.5 Myr.

2) Between 3.1 and 2.6 Mya there was a gradual cooling, with embedded oscillations, that is perhaps the end of the Cenozoic Decline. Around 3Mya the cold extremes of the oscillations were warmer than the warm extremes of more recent times. This means that Antarctica had to be largely unglaciated. Greenland is too small to do it, unless sea temps were also much higher, 3.5C. Deep SST can't go up until marine ice disappears, however, as we shall explain later.

3) about 2.7Mya we crossed a threshold from a largely ice free world into the glacial ages of the last 2.5 My. We are at about the same point in global ice volume now. Can we make a transition in the opposite direction?

Let's next divide the sample up into 4 characteristic periods.

a) 0-0.4 Mya Late Pleistocene
b) 0.8-1.2 Mya
c) 1.8.2.2 Mya Mid Pleistocene
d) 2.3-2.7 Mya Late Pliocene

In the late Pleistocene variance is dominated by 100Kyr variations. In the late Pliocene variance is dominated by 40kyr periods. The presence of 100kyr variations in the mid-Pleistocene when maximum ice values were smaller, suggests that a critical threshold for total ice volume is not necessary to produce the 100kyr oscillation.

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