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Cross-references

Carbon Isotopes, Stable
Heat Transport, Oceanic and Atmospheric
Ocean Paleocirculation
Ocean Paleotemperatures
Paleoclimate Modeling, Pre-Quaternary
Paleoclimate Modeling, Quaternary
Paleoceanography
Plate Tectonics and Climate Change
Thermohaline Circulation

PALEOCEANOGRAPHY

Introduction

Paleoceanography is the study of the history of the oceans. It encompasses aspects of oceanography, climatology, biology, chemistry and geology. The main sources of information are biogenic and inorganic marine sediments, as well as corals. Biogenic sediment includes planktonic and benthic fossils whereas inorganic sediment includes ice-rafted debris and dust. On land, paleo-shorelines and erosional features as well as outcroppings of paleomarine sediments are the principal sources of proxy data. Glaciological records can also give indirect information about paleoceans. The ocean's high heat capacity and its ability to transport energy and to sequester and release greenhouse gases give it an important role in helping to determine the state of the planet's climate. Thus, paleoceanographic research is also intimately linked to paleoclimatology.

Methods

The reconstruction of paleocean characteristics and dynamics requires climatic detective work. It involves the dating and interpretation of paleoclimatic records as well as the definition of physical and dynamical constraints which specify possible circulation patterns and characteristics.

Reconstructions (Proxy Data)

Direct measurements of the quantities of interest to oceanographers extend only into the relatively recent past and in most cases do not go further back than the mid-nineteenth century. To study the ocean during periods for which there are no direct measurements one must rely on indirect evidence. Historical documents can be used as sources of data. Ship logbooks and sailing times across frequently traveled routes have provided estimates of the directions and strengths of past prevailing winds (Brazdil et al., 2005). The frequency and intensity of El Niño events since the 1500s have been reconstructed based partially on historical accounts of large floods and crop losses (Quinn and Neal, 1987). This type of analysis furnishes qualitative descriptions of the past.

Quantitative reconstructions are possible by proxy, where a quantity which is preserved in a natural archive and can be measured, stands as a surrogate of the parameter of interest. A basic requirement is that the relationship between the proxy parameter and the quantity of interest has to be known. When this is the case, the history of the proxy variable can be converted into the history of the variable of interest by the use of mathematical expressions of the type:

$$Int_t = f(Prx_t) \tag{1}$$

which state that the parameter of interest, *Int*, is a function of the proxy quantity, *Prx*. The *t* index refers to time. Equations of this kind are commonly called *transfer functions*. The confidence with which *Int* can be estimated will depend on a series of factors, starting with the quality of the proxy measurements. Also relevant, and a common source of uncertainty, is how well *f* represents the relationship between *Prx* and *Int*.

In most cases, transfer functions are obtained empirically by comparing directly measured values of the quantity of interest to a pertinent set of proxy data. A potential source of error is that the function obtained by this procedure might not be general, but in fact could represent a relationship between *Prx* and *Int* that is peculiar to the data sets used to generate *f*. This problem can be minimized by expanding the spatial and temporal coverage of the data used to establish the transfer function. Still, even assuming that *f* is a perfect representation of how the proxy and the quantity of interest are connected to each other in the present, there is no guarantee that the relationship between them was the same in the past.

Another source of error can be easily understood by rewriting Equation (1) so that it expresses the proxy quantity as a function of the variable of interest. It is reasonable to assume (and in many cases it has been demonstrated) that *Int* is not the only factor controlling *Prx*, so that in fact we end up with:

$$Prx_t = f^{-1}(Int_t, E_{1t}, E_{2t}, \ldots, E_{nt}) \tag{2}$$

where $E_1, E_2, \ldots, E_n$ represent environmental parameters that also influence the proxy variable but are independent of the quantity of interest. An immediate conclusion is that reconstructions of *Int* based on *Prx* will be "contaminated" by other factors so that part of the variability observed in the proxy quantity is not related to changes in the parameter of interest. Comprehensive analysis of the relationships between proxies and a series of observed parameters can offer some insight into how to remove part of the undesired influence of other factors from the reconstruction.

Given the complexity involved in developing skillful transfer functions as well as in identifying and correcting for potential sources of error, a common strategy is to reconstruct the same parameter of interest using different proxies. Such analyses are known as multi-proxy reconstructions (Fischer and Wefer, 1999).

Types of proxies:
The systematic use of proxies in quantitative reconstructions of past oceanic environments originated in the second half of the twentieth century. Since then, a large number of proxy techniques have been established and more are constantly being developed. Proxies can be grouped in six broad categories, based on the type of direct measurement (Fischer and Wefer, 1999). These are listed below, together with brief descriptions of the main variables of interest associated with each proxy. The following chapters on paleotemperature, paleoproductivity and paleocirculation present in more detail the proxies relevant to each of these fields. Comprehensive discussions of oceanographic proxies can be found in the literature (Bradley, 1999; Fischer and Wefer, 1999; Henderson, 2002).

- *Microfossil assemblages*. The relative abundance of planktonic and benthic species of foraminifera, coccoliths, radiolaria, diatoms and other organisms can be used to estimate past ocean temperature, productivity and sea ice distribution. This proxy type was used for the CLIMAP project, which produced the first global distribution of sea surface temperature for the Last Glacial Maximum (CLIMAP – Project Members, 1976).
- *Stable isotopes* are based on the ratio between different isotopes of an element. The ratios are usually standardized by a reference value and named after the heavier isotope. The ratio between ^{18}O and ^{16}O, for example, is represented by $\delta^{18}O$. Isotope readings are retrieved mainly from foraminifera skeletons (tests), organic matter or other sources (e.g., water molecules in continental ice sheets). The amount of ^{18}O incorporated by organisms like foraminifera and corals increases as temperature decreases. Continental ice is relatively depleted in ^{18}O compared to sea water. This makes $\delta^{18}O$ a proxy for both temperature and the extent of continental ice sheets. δ^{11} B is used as a proxy for pH. Productivity, nutrient concentration and past circulation can be reconstructed from δ^{15} N and δ^{13} C (^{12}C is taken up with slight preference to ^{13}C during photosynthesis). Together with microfossil assemblages, $\delta^{18}O$ and δ^{13} C are the paleoceanographic proxies with the most widespread use.
- *Radiogenic isotopes*. The different solubilities of uranium and two products of its naturally occurring decay, thorium (Th) and protactinium (Pa) can be used to estimate the rate of deep water flow and the flux of particles from the water column to the sediments. This flux can also be used as a productivity estimate. ^{14}C preserved in organic matter is used to estimate the age difference between near surface and deep waters and hence, ventilation rates.
- *Biogenic compounds*. The concentrations of some compounds, mainly organic carbon, calcium carbonate and opal, are used as estimates of past productivity. Calcium carbonate is also an indicator of the calcite compensation depth. Alkenones are long chained organic molecules resistant to degradation. The alkenones produced by some coccolithophors can have two or three double bonds in their structure. The ratio between molecules with two and three double bonds reflect the temperature at the time of synthesis.
- *Elements*. The concentrations and ratios of certain elements in the sediment, organic remains, tests and corals are also used as proxies. The ratios of strontium to calcium (Sr/Ca) and magnesium to calcium (Mg/Ca) in biologically precipitated marine carbonates are temperature dependent. The cadmium to calcium (Cd/Ca) ratio is used for nutrient reconstructions. Barium concentration and the Ba/Ca ratio are proxies for productivity and alkalinity, respectively.
- *Sedimentology*. Grain size distribution can provide qualitative information about bottom current speeds and act as an indicator of ice rafted debris. Information about past tides can be inferred from layered sediments called rhythmites. The mineralogy of the sediments can be used to establish source areas and direction of transport for both water and wind borne sediments.

Conspicuously absent from the quantities of interest listed above is salinity, a fundamental parameter which influences many aspects of the ocean environment. There is, at the moment, no independent proxy for this quantity. As salinity also influences $\delta^{18}O$, an indirect measurement can be obtained by using independent estimates of temperature (alkenones, for example) to remove the temperature signal from existing $\delta^{18}O$ series. Attempts have also been made to infer salinity from microfossil assemblages. Unfortunately, both approaches generate errors of $\sim$1 psu, very large compared to the range of salinity variability in the oceans (Fischer and Wefer, 1999).

Reconstructions (Models)

The theoretical approach to paleoceanography uses quantitative ocean and climate models to reconstruct paleoconditions of the ocean and interpret observations. A number of different ocean and coupled climate models have been used in the past and we will give a very short overview of the hierarchy and use of these models in the broad research area of paleoceanography.

It is, of course, impossible to build an exact model of the climate system; for the ocean alone, the position and momentum of approximately 5×10^{46} molecules would have to be calculated at each instant of time. As an alternative, the ocean, atmosphere, sea ice, or land surface is split into discrete macroscopic elements with measurable characteristics such as temperature, density, velocity, etc. The state of and exchange between these discrete elements follow physical laws and can therefore be determined with numerical models. Small scale processes within each element can also influence the large scale pattern and therefore have to be parametrized. Existing ocean (and climate) models differ in regards to:

- their temporal and spatial resolution (resolution is defined as the spatial scale which defines the boundary between processes that are resolved by the model and those which are parametrized)
- the nature of processes which are resolved (for example, some models include biogeochemical cycles whereas other models resolve physical processes only)

- the number of subsystems taken into account (for example, an ocean model needs boundary conditions at the surface which can either be provided by data, or by an atmosphere model which is physically interacting with the ocean model)

As the different subsystems (ocean, atmosphere, sea ice, continental ice sheets, vegetation, etc.) of the climate system interact with each other in a complex manner and on a very broad range of timescales, climate modelling reduces to the process of identifying isolable subsystems and processes that are relevant to the problem at hand. While identifying these subsystems and processes, the researcher has also to keep in mind that these processes have to be suitable to be simulated by limited mathematical models and will provide results in a reasonable computational time (Crowley and North, 1991).

The simplest class of climate models includes one-dimensional Energy Balance Models which were first developed by Budyko (1969) and Sellers (1969). It is interesting to note that these simple models yield two stable solutions under present day boundary conditions: the present day climate and a completely frozen Earth (also called "Snowball Earth").

Today's physical ocean models can be classified based on the following categories:

- geography (regional models, global models)
- physics (hydrodynamic, thermodynamic or hydrothermodynamic models)
- surface approximation (free surface, rigid lid)
- vertical discretization (fixed level, isopycnal, sigma-coordinate, semi-spectral)
- density variation (barotropic, baroclinic)

Because the boundary data for paleoclimatic simulations tend to contain large uncertainties, global models are better suited for this research area than regional models. The most comprehensive results are given by global ocean general circulation models (OGCMs). These models can be driven with reconstructed data specifying the boundary conditions. For example, numerous modelling studies restored the ocean surface characteristics to the CLIMAP data set (CLIMAP – Project Members, 1976) for simulations of the Last Glacial Maximum.

A better approach than using ocean-only models is to use coupled ocean-atmosphere models; by computing the surface boundary conditions, one can bypass the data problem. However, coupled ocean-atmosphere GCMs often need flux adjustments. Flux adjustments balance surface fluxes at the ocean-atmosphere interface to avoid a numerical drift of the coupled system. As flux adjustments have been "tuned" to the present day climate, the use of these adjustments to simulate past climates is not very reliable. However, some recent studies use coupled atmosphere-ocean GCMs which do not need artificial flux adjustments (e.g., LeGrande et al., 2006).

Some studies use simple atmosphere models which still provide reasonable boundary conditions for the ocean model (Weaver et al., 2001). Other climate subsystems may also have an important influence on the state of the ocean (continental ice sheets, sea ice and land surface processes for example). The class of Earth System Models has been developed recently and comprise all models which take into account more than two subsystems of the climate system. Today, Earth System Models are widely used for paleoceanographic and paleoclimate simulations. There is also growing evidence that geochemical interactions between the subsystems are more important than initially thought (carbon cycle, nitrogen cycle, methane, etc.) and some initial attemps in integrating these cycles in Earth System Climate models for paleoclimatic simulations have been made (Crucifix, 2005). The modelling community is continuously integrating new processes and subsystems in their models to obtain a better representation of the climate system dynamics.

Combining models and proxy data

The interpretation of paleoproxy data is an ongoing challenge for paleoclimate scientists. As a striking example, one could compare the studies of Clark et al. (2002) and Bond et al. (2001). Both papers are highly regarded, yet draw opposite conclusions from the atmospheric $\Delta^{14}C$ record. Whereas Bond et al. (2001) relate the variability of atmospheric $\Delta^{14}C$ to changes in solar radiation, Clark et al. (2002) interpret the same type of record as a signature of variability in the thermohaline circulation and ocean heat transport. On the other hand, model simulations of past climates depend strongly on boundary conditions, assumptions, and the model used for the study. The simulated climate for a certain time span can be radically different depending on the model and boundary conditions used. Interpretation of measured paleoclimate data is thus urgently needed through collaboration between modellers and observers.

One possible approach to combine proxy data with climate models is that of "data assimilation." Data assimilation involves the construction of a field that accommodates best the information obtained from paleoproxies with the physical (and dynamical) constraints of the climate system using coupled climate models (e.g., Paul and Schäfer-Neth, 2005).

Another possible approach is to incorporate paleoproxy data (e.g., $\delta^{18}O$, deuterium excess, $\Delta^{14}C$, $\delta^{13}C$, $\delta^{10}Be$, etc.) as prognostic active tracers in climate models. Perturbations (such as meltwater events or changing solar activity) and other climate states (e.g., the Last Glacial Maximum) can then be simulated and the behaviour of these simulated proxies can be compared to observed proxy data obtained from ice cores, marine sediments and other records. This has been done with uncoupled ocean (or atmosphere) general circulation models (e.g., Schmidt, 1999; Werner et al., 2000; Butzin et al., 2005), vegetation models (Kaplan et al., 2002) and continental ice sheet models (e.g., Clarke et al., 2005). However, the importance of interactions between atmosphere, oceans and other systems such as the biosphere and the cryosphere point to the necessity of using coupled models. To date, there have been only a few studies simulating paleoproxy data with either coupled ocean-atmosphere GCMs or Earth System Models (e.g., Stocker and Wright, 1996; Meissner et al., 2003; Roche et al., 2004; Crucifix, 2005; LeGrande et al., 2006).

A third way to bridge the gap between the modelling and proxy data approach is to find locations of proxy data records of special interest with the use of climate models. A simulation including prognostic paleoproxy tracers can determine the geographical region of greatest impact on a given paleoproxy data during a given climate event.

In conclusion, large amounts of paleoproxy data have been retrieved from various types of archives, but attempts to use numerical models for verification and interpretation of this data are sparse. The science of using three-dimensional climate models to interpret paleo records is still in its infancy.

Processes

In this Section, we give a short overview of the processes and boundary conditions which might have influenced the oceans in the past.

Paleotides

The history of tides over geological time is associated with the evolution of the Earth-Moon system and the shape of ocean basins. Tidal currents generate friction at the bottom of the ocean resulting in the transfer of energy and angular momentum associated with the Earth's rotation to the Moon's orbital motion. This process has been gradually slowing the Earth's spin and increasing the radius of the Moon's orbit. According to some estimates, at ~620 million years ago (Ma), days were approximately 22 h long and the Earth-Moon distance was about 96% of its present day value (Williams, 2000). The main proxy for tidal (and Earth-Moon system) changes over periods of millions of years is based on the analysis of tidal rhythmites, laminated sediments whose deposition is associated with tidal currents (Williams, 2000).

Tidal dissipation depends strongly on the shape of ocean basins. As tectonic processes cause the basins to change, the effects of tidal friction should also change. There are many indications that this is the case. For example, the present rates of dissipation appear to be higher than the average rates over the planet's history (Kagan, 1997; Gills and Ray, 1999).

Over the Quaternary, the shape of ocean basins was altered due to changes in sea level and the presence or absence of ice shelves. Numerical models show that in the Labrador Sea, tidal amplitudes during the last glacial were about twice as large compared to present day conditions. This has led to the suggestion that these higher tides could have destabilized floating ice shelves and caused Heinrich events (Arbic et al., 2004). Another connection between tides and climate relates to a millennial tidal cycle. It has been proposed that very high tides, occurring every 1,800 years, can cause increased mixing and cooling of the sea surface. This cooling would be related to abrupt climate change observed with similar periodicity (Keeling and Whorf, 2000). Some researchers contest this hypothesis, arguing that tidal forcing at these frequencies is very weak (Munk et al., 2001).

Radiation

Incoming short wave radiation from the Sun is the ultimate source of energy for ocean dynamics, the hydrological cycle and life in the oceans. Geographic and seasonal variations in the intensity of insolation result in temperature and pressure gradients which have an important influence on ocean dynamics and the climate system. Although the ocean surface circulation is mostly wind driven, the winds themselves are the result of the uneven distribution of energy on Earth. Density driven currents on the other hand depend strongly on temperature (and thus indirectly on energy distribution) and salinity gradients (which result from precipitation/evaporation patterns and are thus closely linked to the hydrological cycle).

Incoming solar radiation changes over a range of very different timescales. Firstly, the solar luminosity has gradually increased throughout the Earth's lifetime. It is estimated that, during the early days of our planet, the solar luminosity was 25–30% weaker than today's value. According to model results, such a reduction in incoming shortwave radiation should result in a completely frozen planet ("Snowball Earth"). However, even if evidence exists for snowball Earth conditions in early Earth's climate history, there is also counter evidence that during long periods of time the paleoceans were ice-free. Thus, it is inferred that, at these times, greenhouse gas concentrations must have been higher than present day levels in order to prevent the system from slipping into an icehouse state (see *Faint young Sun paradox*, this volume).

On shorter timescales (order of tens to hundred thousands of years), the incoming solar radiation is modulated by changes in the Earth's orbital parameters which describe the character of the Earth's orbit around the Sun. These three parameters change continuously, causing a variation in the total amount of solar radiation received on Earth as well as the seasonal and latitudinal distribution of insolation. The first parameter, called eccentricity, describes the degree of ellipticity in the Earth's orbit around the Sun and hence the shape of its orbit. The characteristic periods of changes in eccentricity are 95,000, 131,000, 413,000, and 2,100,000 years. Eccentricity is the only parameter which modulates the total global amount of solar energy received at the top of the atmosphere. The change in Earth's axial tilt through time is described by obliquity and has a distinct period of 41,000 years. Finally, the precession of the equinoxes combines the axial precession (wobbling of the axis) and the precession of the ellipse (rotation of the elliptical shape of Earth's orbit) and consists of a strong cycle with a 23,000 year period and a weaker one with a 19,000 year period (Ruddiman, 2000). Precession and obliquity do not alter the total amount of solar radiation received, however, they change the distribution of incoming radiation by latitude and by season. All the frequencies of these parameters can be found in climatological records of the paleocean. Thus, orbital parameters play an important role in driving the climate system and ocean dynamics.

Finally, short-term variability in solar luminosity (which is correlated with changes in the number of sunspots visible on the Sun's surface) acts on timescales of decades to millenia. Over the last hundred years, the global mean temperature has followed a trend similar to the sunspot record. Some climate scientists have hypothesized that periods of global cooling (e.g., the Little Ice Age (~1560–1850)) have been partly caused by a minimum in sunspots (Sporer sunspot minimum (1460–1550) and Maunder sunspot minimum (1645–1715)).

Ocean basin changes

The shape of the ocean basins sets the boundary conditions for the ocean circulation. Different continent configurations result in different flow patterns. Understanding how the boundaries influenced past circulation patterns might offer insight into the dynamics and other important processes of the present day oceans. For example, the existence of an unobstructed low latitude passage in the Tethys Ocean (~160–14 Ma) has been associated with increased poleward heat transport (Hotinski and Toggweiler, 2003). Of course, both reconstructions and simulations of ocean currents this far into the past are subject to many uncertainties. For example, even the existence of a prominent, large scale feature such as the Tethys circumequatorial current is still not unequivocally accepted (Barron and Peterson, 1989; Bush, 1997; Poulsen et al., 1998; Hotinski and Toggweiler, 2003).

A number of proxy and modelling studies show that the opening and closing of passages between two basins can have a large impact on ocean circulation and climate. The closing of the isthmus of Panama which is the gap between the North and South American continents at ~5–3 Ma is thought to have intensified the Gulf Stream and the associated northward transport of heat and salt into the North Atlantic (Haug and Tiedemann, 1998). The closure of the isthmus of Panama has also been associated with the onset of Northern Hemisphere glaciation (Keigwin, 1982; Lear et al., 2003; Mudelsee and Raymo, 2005) as well as with changes in water mass properties and the overturning circulation of the North Pacific (Motoi et al., 2005). The opening of the Drake passage between Antarctica and South America (~28–33 Ma) and subsequent establishment of the Antarctic Circumpolar Current is associated with the glaciation of the Antarctic continent (Kennett, 1977; Toggweiler and Bjornsson, 2000). It has been proposed that the relative stability of the Atlantic Meridional Overturning Circulation (MOC) during the Holocene is related to the open connection between the Arctic and Pacific Oceans provided by the Bering Strait (De Boer and Nof, 2004). According to this scenario, low salinity anomalies in the North Atlantic make the MOC unstable when the strait is blocked (as during the last glacial). On the other hand, the flux of low salinity water from the Pacific into the Arctic through a wide open Bering Strait during conditions with higher sea levels (last interglacial, ~115–130 ka BP) has been linked to the more unstable MOC (Shaffer and Bendsten, 1994).

Oscillations in sea level result in changes of the surface's land to ocean ratio which in turn influences the planetary albedo. According to model results, alteration of the planet's albedo caused by a sea level drop of approximately 400 m during the late Ordovician (~455–445 Ma) was one of the factors that could have driven the climate into a cooler state (Herrmann et al., 2004). By submerging vegetated areas or making new land available to plants, changes in sea level can also impact carbon storage. The amount of carbon present on shelves inundated by the rise in sea level since the Last Glacial Maximum appears to be equivalent to the increase in the atmospheric stock of carbon during the same period (Montenegro et al., 2006).

Ice

A gradual cooling over the last 55 million years led to the presence of extensive continental ice sheets in both the Southern and Northern Hemisphere. Intensification of Northern Hemisphere glaciation in Eurasia and North America occurred between 2.7 and 2.5 Ma. With increasing terrestrial ice sheets, sediment records show an intensification of climate oscillations between extreme cold glacial maxima and interglacial warm periods on timescales related to the orbital parameters. The boundary conditions for paleoceans varied dramatically between glacial and interglacial periods. The formation and melting of extensive ice sheets in North America and Eurasia, important variations in atmospheric CO_2 concentrations as well as shifts in the extent of sea ice changed the radiation balance, salinity distribution, dynamical forcing and productivity of the oceans.

Continental ice sheets

Continental ice sheet growth and decay during glacial cycles may have influenced the ocean circulation in several ways:

- Sea level changes due to the storage of freshwater on continents and bedrock depression may have led to different circulation patterns in the ocean.
- Extraction of freshwater from the ocean to build up continental ice sheets led to changes in the global average and distribution of ocean salinity. These changes in salinity (and therefore density) may in turn have influenced the circulation and heat transport (Meissner and Gerdes, 2002).
- Changes in land surface albedo altered the local radiation balance over continental ice sheets, which in turn must have changed sea surface temperatures and circulation.
- Elevation changes in regions of continental ice sheets could also have caused changes in atmospheric circulation (Lehman, 1993) by affecting the atmosphere's stationary wave pattern (Jackson, 2000) and therefore the dynamic forcing of the ocean circulation.
- Changes in global atmospheric temperatures resulting from the presence or absence of ice sheets led to changes in ocean temperatures which may in turn have affected the circulation and heat transport.

Climate variability on millenial time scales has been more important during glacial periods than during interglacials. Dansgaard-Oeschger temperature oscillations, which seem to be part of slow-cooling cycles occurring every 10–15 kyr (Dansgaard et al., 1993), have been related in several studies to changes in the strength of the meridional overturning (MOC, see *Thermohaline Circulation*, this volume). Local sea surface salinity perturbations at high latitudes due to meltwater and/or iceberg discharges (Heinrich Events) may have weakened the MOC (e.g., Rahmstorf, 1995 and references therein) leading to abrupt climate change (Broecker, 1994; Manabe and Stouffer, 1995).

Sea ice

Sea ice regulates exchanges of heat and freshwater between ocean and atmosphere and can change sea surface salinity through melting or freezing (brine rejection). It insulates the relatively warm ocean water from the cold polar atmosphere, changes the surface albedo (and therefore the local radiation balance) dramatically and influences evaporation and therefore local cloud cover and precipitation. By changing the surface characteristics of the oceans, sea ice plays an important role in deep water formation and meridional heat transport in the ocean. The positive ice-albedo climate feedback might have led to a "run away" icehouse feedback in the Precambrian ("Snowball Earth," Hoffman et al., 1998).

Conclusion

Paleoceanography is an exciting research area that unifies specialists with a broad range of backgrounds. Scientists in each of the two classical schools of paleoceanography (reconstructions through proxy data and theoretical quantitative analysis using models) are more and more exchanging expertise and working together for a better understanding of the past oceanographic environment. However, inherent problems associated with the poor spatial and temporal resolution of proxy data as well as uncertainties related to the proxy data themselves make their interpretation difficult and sometimes impossible. At the same time, models are limited by resolved processes, resolution and the quality of boundary conditions. Overall, there is much progress to be made in both fields. In spite of these problems, the geological data as well as model simulations provide a substantial set of results which gives us some insight on how the ocean and the whole climate system functions. This knowledge is of ultimate importance to understand and predict future climate changes due to anthropogenic perturbations.

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Cross-references

Alkenones
Astronomical Theory of Climate Change
Atmospheric Circulation During the Last Glacial Maximum
CLIMAP
Dating, Radiometric Methods
Eolian Dust, Marine Sediments
Faint Young Sun Paradox
Geochemical Proxies (Non-Isotopic)
Ice-Rafted Debris (IRD)
Little Ice Age
Marine Biogenic Sediments
Maunder Minimum
Ocean Paleotemperatures
Organic Geochemical Proxies
Paleocean Modeling
Paleoclimate Proxies, an Introduction
Plate Tectonics and Climate Change
Quaternary Climate Transitions and Cycles
Snowball Earth Hypothesis
Thermohaline Circulation

PALEOCENE-EOCENE THERMAL MAXIMUM

Approximately 55.5 Ma ago in marine and terrestrial sections all around the globe, there was an exceptionally large carbon isotope (δ^{13}C) excursion (CIE) which is unique in the Tertiary for its size and rapidity of its onset. This event, which was first recognized in multiple cores by Kennett and Stott (1991), is contemporaneous with a significant warm anomaly (and oxygen isotope anomaly) and has thus been named the Paleocene-Eocene Thermal Maximum (PETM) (Figure P8). Due to the singular nature of this perturbation and the ecological, faunal and physical gradients across it, the base of this event is now used to define the boundary of the Paleocene-Eocene transition. The previous definition of the P/E boundary was significantly later (at the base of the Ypresian section in Europe), and thus placed the CIE in the late Paleocene. Previously authors had therefore referred to this event as the Late (or Latest) Paleocene Thermal Maximum (LPTM). Some authors have since referred to it as the Initial Eocene Thermal Maximum (IETM). The CIE is contemporaneous with the Clarkforkian/Wasatchian transition, long accepted as the Paleocene-Eocene boundary in the North American Land Mammal Series.

Full characterization of the event has been handicapped by the number of incomplete sections and frequent hiatuses found near the Paleocene-Eocene transition. The Global Stratigraphic Section and Point (GSSP) for this transition has been defined to be the Dababiya Section, 35 km south of Luxor, on the right bank of the Nile Valley. The event lies within Nano-Plankton zone 9 (NP-9) and reverse-polarity magnetic period Chron 24r. The dating of the PETM is estimated to be around 55.5 Ma, based on the chronology of (1996) using interpolation between well-dated ash layers.

Background

The Paleocene was a time of large shifts in the global environment following the disruptions that occurred at the Cretaceous-Tertiary boundary (65 Ma). Evolutionary increases in diversity, and increasing ocean productivity marked this epoch, leading to the most positive carbon isotope values in the Tertiary. Temperatures were generally warm, and the polar regions (such as Ellesmere Island) were replete with sub-tropical fauna and flora (crocodiles, ferns etc.). There is no evidence of extensive glacial ice, though evidence for absence of sea ice in high northern latitudes is not as convincing. Tropical temperature estimates do not show as much (if any) change compared to present, but tropical proxies are known to suffer more from diagenetic alteration and the possibly complicating factors relating to the tropical hydrological cycle.

Focusing more specifically on the ≈2 Ma late Paleocene to early Eocene transitional period, global temperatures increased (to an estimated 2–4 °C warmer than present day), deep ocean temperatures increased by approximately 3 °C and δ^{13}C values

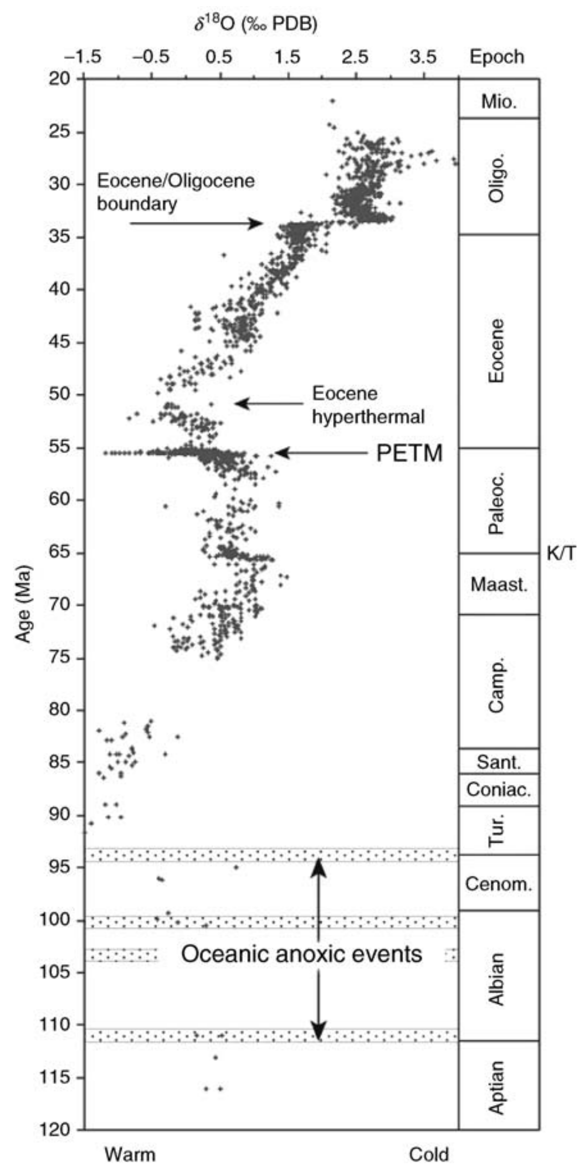

Figure P8 A compilation of δ^{18}O values in marine carbonate sediments since the Cretaceous showing the PETM anomaly in the context of Cenozoic variability (after Zachos et al., 2001).