On the beginnings of palaeoceanography: Foraminifera, pioneers and the *Albatross* Expedition

WOLFGANG H. BERGER

Geosciences Research Division, Scripps Institution of Oceanography, University of California, San Diego, California 92093-0244, USA (e-mail: wberger@ucsd.edu)

Abstract: The field of studies labelled palaeoceanography integrates research on foraminifers and other microfossils within their environment on deep-sea sediments raised by cores taken from ships. The first large-scale project deserving the label was the Swedish Deep-Sea Expedition, which used the sailing vessel *Albatross* and the then newly developed Kullenberg coring machine to raise hundreds of long cores from around the world during 1947–1948. Thus, just as the *Challenger* Expedition (1872–1876) serves as a marker point for the beginnings of oceanography, the *Albatross* represents the beginnings of palaeoceanography. The unique defining element – the link of micropalaeontology to ocean circulation and climate – came to dominate discussions in the 1950s with the reports by Gustaf Arrhenius and Cesare Emiliani. At the same time, important steps in the evolution of the new field depended on the expansion of the inventory of cores by major oceanographic institutions (especially Lamont Geological Observatory in Palisades, New York), and on the introduction of new dating methods. Subsequently, the field benefited from the increased use of various statistical techniques, as well as of numerical modelling, all of which profited greatly from the rapid growth of computing power.

As a field of study, palaeoceanography (ocean history) is older than its label, which gained general acceptance in the 1970s (e.g. Hay 1974). Much of the earlier work might be categorized as 'marine micropalaeontology'. Whatever the name, the history of palaeoceanography is not a subject that attracts much attention. In the recent *Encyclopedia of Paleoclimatology and Ancient Environments* (Gornitz 2009), for example, the entry for 'paleoceanography' (p. 690) has two citations to pre-1970 articles (out of 53): Budyko (1969) and Sellers (1969). Both are important papers on the radiation balance of the planet; neither of the authors is an ocean historian. The entry for 'marine biogenic sediments', in compensation, has eleven citations to pre-1970 papers (out of 47). The pre-1970 authors cited are G. O. S. Arrhenius, W. A. Berggren, M. N. Bramlette, D. Bukry, R. A. Daly, C. Emiliani, M. P. Luterbacher, I. Premoli-Silva, M. Milankovitch, E. Olausson and F. L. Parker. All except Milankovitch qualify as pioneers of palaeoceanography. Significantly, six of the ten palaeoceanographers are pioneers of biostratigraphy, reflecting the strong link to that field. Four of the pioneer palaeoceanographers cited worked on cores taken by the *Albatross* Expedition (1947–1948) as did W. Schott, the geologist who might be considered the founding figure of the field. Six of them (and also Schott) worked on foraminifers.

Within the sciences of the ocean, marine micropalaeontology and palaeoceanography together constitute perhaps 2–3% of the body of knowledge judging by the entries in the comprehensive *Encyclopedia of Ocean Sciences* (Steele *et al.* 2001). These fields of study have clearly been of central importance in the great geological revolutions experienced in the last 50 years, however, revolutions associated with the names of Wegener, Dietz, Hess, Alvarez and Milankovitch. Biostratigraphy based on planktonic remains in deep-sea sediments allowed the testing of seafloor spreading (during Deep Sea Drilling Project or DSDP Leg 3) and the timing and effects of the Cretaceous–Tertiary (K/T) impact (in the Gubbio section and in deep-sea cores). Quantitative micropalaeontology permitted the elucidation of the orbitally driven climate variations over the last several million years.

On the whole, the vital contributions of micropalaeontologists in providing the basic tools for the reconstruction of ocean history have been less visible than the contributions from geophysics, isotope chemistry and mathematical modelling. The reason may be the exploratory nature of much of micropalaeontogy where, throughout the twentieth century, data tended to be gathered in the manner of mapping unknown continents rather than for testing specific hypotheses.

Invariably, beginnings are the stuff of myth. Where to put the beginnings of a given field is a matter of definition; that is, of personal preference. For some, the speculative paper by Chamberlin (1906) on possible reversals in the ocean's deep

From: Bowden, A. J., Gregory, F. J. & Henderson, A. S. (eds) 2013. *Landmarks in Foraminiferal Micropalaeontology: History and Development*. The Micropalaeontological Society, Special Publications, Geological Society, London, 159–179.

circulation might signal (correctly) the beginnings of palaeoceanographic thinking. If we allow thought rather than data to lead the way, we might even end up with James Croll (1875) as the original pioneer. However, this approach would have the embarrassing consequence that the beginnings of palaeoceanography are earlier than those of the *Challenger* Expedition, commonly taken as the beginning of oceanography.

One strikingly bald possibility entertained in the *Encyclopedia of Ocean Sciences* (Steele *et al.* 2001, p. 2078) is that palaeoceanography is defined by whatever was done on Lamont cores in the 1950s (a viable point of view, perhaps, but somewhat narrow). In the list of pioneers of the related field of palaeoclimatology in the *Encyclopedia* edited by Gornitz (2009, p. 428 ff), we therefore find entries for two palaeoceanographers – Cesare Emiliani (1922–1995) and Nicholas Shackleton (1937–2006) – who both used cores from Lamont in their work. However, Emiliani worked in Chicago and Miami, and Shackleton worked at Cambridge University, UK. In the same volume, we find an article by James D. Hays of Lamont listing the fundamental contributions of the Climate: Long range Investigation, Mapping and Prediction (CLIMAP) project to palaeoceanography, a project that relied on the core collections at Lamont. In many ways, that project and its successor (SPECtral MAPping, SPECMAP) indeed mark the beginnings of modern palaeoceanography.

Another way to look at the history of palaeoceanography is to see the type of large-scale collaborative project described by Hays as the end of the time of the pioneers; this sets the beginnings to the time before 1970, with fully-fledged palaeoceanography operating afterwards. This view (the one I prefer) seems more in line with that of W. W. Hay (Miami, Colorado and Kiel) who wrote a thorough review of palaeoceanography for the Centennial of the Geological Society of America. Hay (1988) emphasized the pioneer role of Cesare Emiliani. In fact, since the history of the ocean resides in deep-sea cores, we would have to set the start of the story somewhat earlier with the recovery of cores from the deep-sea floor, that is, with the German *Meteor* Expedition (1927–1929) and with the long cores studied by Bradley *et al.* (1938, 1942) at the US Geological Survey. The Swedish *Albatross* Expedition (1947–1948) had cores from all ocean basins and a large number of scientists (including Emiliani) working up the materials. It can therefore serve as a convenient tie-point for the 'beginnings' of palaeoceanography, much as the British *Challlenger* Expedition (1872–1875) does for the field of oceanography, even though there were numerous earlier expeditions carrying out oceanographic research.

On the crucial role of Foraminifera

Foraminifera (now usually spelled 'foraminifers') were the main tool in the reconstruction of ocean history from the beginning. Other marine organisms also provide important information, of course, including nannofossils, radiolarians, diatoms, silicoflagellates, dinoflagellates and ostracods (Funnell & Riedel 1971). The study of foraminifers entirely dominated the field in the beginning, however.

This intimate connection of palaeoceanography to fossils links the true beginnings of the field with the beginnings of palaeontology: the discovery of the meaning of fossils as remains of once-living organisms, as markers of geological time and as clues to past environmental conditions. The book entitled *Paleoceanography* by Schopf (1980) illustrates this link to palaeontology, as do the palaeoecological portions of the *Principles of Paleontology* by Raup & Stanley (1978) and Valentine's *Evolutionary Paleoecology of the Marine Biosphere* (1973), for example. However, the modern usage is reflected in the treatment by Kennett (1982: 'Marine Geology'), who puts palaeoceanography in the context of the history of the deep ocean. In some ways this link goes back to the *Challenger* Expedition. It was an interest in evolution, stimulated by the work of Charles Darwin, that helped motivate the expensive, three-year *Challenger* Expedition across the world's seas. The search for 'living fossils' on the deep-sea floor proved futile. The biosphere of the deep sea is thoroughly modern, with benthic foraminifers holding the clues to the main reason for this fact: an overall cooling of deep waters since the end of the Eocene and associated environmental changes at the surface (e.g. eastern boundary upwelling).

Such insights came much later, thanks to a thorough probing of the memories stored in deep-sea sediments by drilling and the concomitant rise of the palaeoceanography of the Cenozoic (Savin *et al.* 1975; Shackleton & Kennett 1975; Kennett 1977; Douglas & Woodruff 1981; Vincent & Berger 1981; Miller *et al.* 1987).

For obvious reasons, the work of the pioneers was mostly restricted to speculating about the most accessible parts of ocean history, that is, the Late Quaternary. Already in the course of the *Challenger* Expedition, the Scotsman John Murray made the discovery that at least one-half of the deep-sea floor is covered with the remains of shell-bearing plankton, mainly microscopic unicellular algae and protozoans, with foraminifers as the most conspicuous component in the sand fraction of widespread calcareous ooze (Fig. 1). By sampling the water, he established that it is planktonic foraminifers rather than benthic species that deliver the bulk of shells on the sea floor (Murray 1897). The proportion

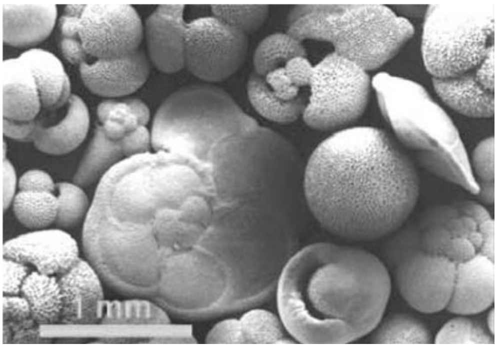

Fig. 1. Planktonic foraminifer fossils in the sand fraction of deep sea sediments. SEM M. Yasuda.

delivered by benthic foraminifers is comparably minor. However, the benthic forms are vastly more diverse than the planktonic forms (Brady 1884), an observation that is still unexplained (although not for a shortage of suggestions concerning controls on diversity; e.g. see Hessler & Sanders 1967; Dayton & Hessler 1972).

Murray realized that the sea floor collects a memory of the conditions of growth in surface waters (temperature and productivity) in the shape of shells of the organisms that once lived and reproduced there. However, sampling techniques were inadequate for providing sufficient material for the reconstruction of ancient conditions.

It took another half-century before a systematic recovery of sediment sequences from the deep-sea was attempted. It was accomplished during the *Meteor* Expedition (1925–1927) in the central and southern Atlantic. The planktonic foraminifers were studied by Wolfgang Schott, a sedimentologist. He was able to recognize postglacial sediments from the presence of *Globorotalia menardii* (a widely used synonym for *G. cultrata* (d'Orbigny)). Schott discovered that *G. menardii* is absent in Atlantic sediments of glacial age. The species invades the tropical Atlantic during glacial–postglacial transitions (presumably in warm-water eddies coming around the Cape of Good Hope from the Indian Ocean).

Schott was therefore able to map the thickness of postglacial sediments in the region (Fig. 2). With the benefit of hindsight informed by radiocarbon measurements (e.g. Rubin & Suess 1955), we can estimate the corresponding sedimentation rates as between 2 and 2.5 cm ka^{-1}.

Schott's timescale was too long by a factor of 2 (being based on information then available), but his rate determinations did correctly reflect the extremely slow accumulation of foraminifer tests and of deep-sea sediments in general. His discovery of the significance of *G. menardii* delivered an important tool for the correlation of cores in the central Atlantic and for an attempt, in the 1950s, to link deep-sea Quaternary sequences into the ice-age schemes derived from studies on land (Ericson *et al.* 1956; Ericson & Wollin 1964).

Schott introduced the concept of quantitative palaeontology by counting the relative abundance of each foraminifer species present in a standardized set of several hundred specimens. A change in the relative abundances of the species reflects with some precision corresponding changes in temperature and productivity of overlying waters. When fed to simple computer algorithms 40 years after the *Meteor* Expedition, Schott's data yielded the insight that glacial ocean temperatures had been much the same as today in the great desert region in the North Atlantic known as the 'central gyre', but that surface waters had been much colder than today close to Africa and along the Equator (Berger 1968). Indications of increased glacial productivity along NW Africa could be detected in Schott's data, as well as a hint that the deep circulation during glacial time was quite different from the patterns we see today. (In Hannover in 1970, Schott was happy but not surprised to learn that his data were still useful decades after his pioneering work.)

Evidence on deep circulation came from the study of preservation patterns. Schott had pointed out that the boundary between the two main, deep-water masses in the central Atlantic – Antarctic Bottom Water and the overlying North Atlantic Deep Water – marks a preservation boundary; samples from above the level of separation of the water masses show good preservation while those from below that level do not. In times past, the boundary between well-preserved and poorly preserved fossils moved up and down along the sloping sea floor, reflecting changes in the depth level of the boundary between the two main, deep-water masses (Ruddiman & Heezen 1967). It was subsequently revealed that there is a counterpoint pattern of preservation on the deep-sea floor in the Pacific basin, so that the deep preservation boundaries in the Atlantic and Pacific move in see-saw fashion (Berger 1973). It is now thought that the preservation see-saw mainly reflects the changing intensity of the flushing of the Atlantic basin by North Atlantic Deep Water.

On the arrival of abundant long cores and subsequent developments

Comparisons between different basins first became possible through the great number of long cores retrieved by the Swedish Deep Sea Expedition (1947–1948) using the research vessel *Albatross*

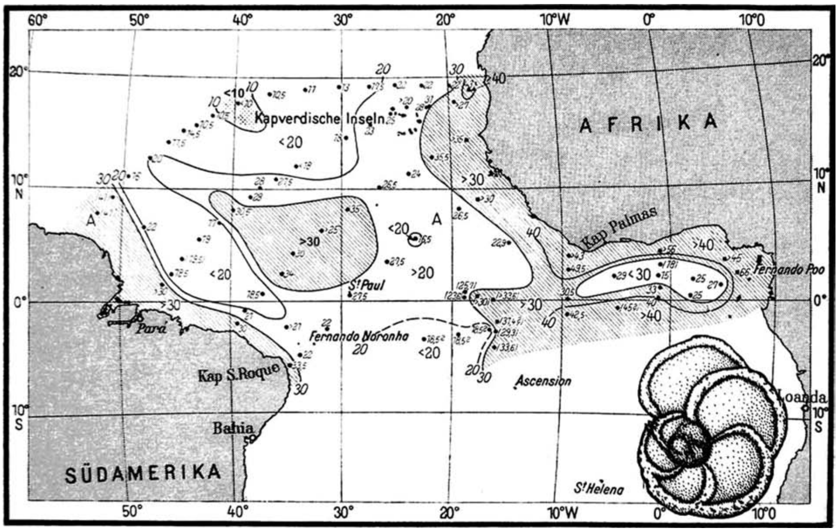

Fig. 2. Thickness of the postglacial sediment in the central Atlantic, identified by noting the presence of *Globorotalia cultrata* (inset, drawing by F. L. Parker). Map from Schott (1935). An expanded version is in Schott (1966: fig. 54), with additional data from *Albatross* samples.

(Fig. 3). The total number of cores was 299 with a combined length of 1.6 km (Kullenberg 1955; Olausson 1996). The most important technical innovation that provided for the success of the expedition was the Kullenberg piston corer (Kullenberg 1955). This is a clever modification of the steel tube principle used for decades earlier, a modification that prevents water pressure from hindering the entry of sediment into the tube. Kullenberg's device allowed the taking of long cores, routinely 10 m long or more. The 'memories' locked in the sediments therefore reached back 0.5–1 Ma. The piston core technology was soon widely adopted by the major oceanographic institutions collecting cores. Modified versions also eventually entered the realm of drilling, with pressure-assisted coring through the hole at the drill bit. A major concern in post-*Albatross* developments was to correct the tendency for piston cores to bring up sequences with significant distortions caused by pressurized filling of the steel tubes.

The marine geologists working on the cores recovered by the *Albatross* established the new field of 'ocean history' as read from deep-sea sediments, the field now known by the term 'palaeoceanography'. Concepts studied include: (1) the response of the ocean to glacial conditions; (2) the nature of the transition from glacial to postglacial conditions; and (3) the nature of sediment cycles and their connection to Milankovitch forcing. As is true for the *Challenger* Expedition some 80 years earlier, the *Albatross* Expedition established a plethora of questions and not necessarily answers.

The project was international in scope. Based on the study of the *Albatross* cores, the Swedish geochemist Gustaf Arrhenius considered the evidence for great changes in productivity in the eastern tropical Pacific and argued for corresponding changes in upwelling from a change in the strength of trade winds (Arrhenius 1952; see Fig. 4). The Italian-American palaeontologist and isotope chemist Cesare Emiliani adduced evidence for the cyclicity of the ice-age oxygen isotope record (Emiliani 1955), and argued for the importance of orbital forcing in climatic change (in the sense proposed by the Serbian astronomer Milutin Milankovitch 1930). The Swedish geochemist Eric Olausson proposed that carbonate preservation patterns contain clues to changes in deep-sea circulation (Olausson 1965, 1967). Finally, the American palaeontologists Frances L. Parker (1906–2002) and Fred B. Phleger (1909–1993) produced evidence for large-scale shifts in climatic zones in the North Atlantic within the last several hundred thousand years (Phleger *et al.* 1953; portraits in Fig. 5).

A number of developments greatly furthered the rapid expansion of palaeoceanographic studies following that expedition: (1) increased awareness of oceanographic concepts among geologists, thanks to the milestone text by Sverdrup *et al.* (1942); (2) a substantial increase in the number of long deep-sea cores raised by various oceanographic institutions (especially at Lamont of Columbia University, thanks to rules laid down by its director Maurice Ewing); (3) a surge of interest in the interpretation of deep-sea sediments raised by

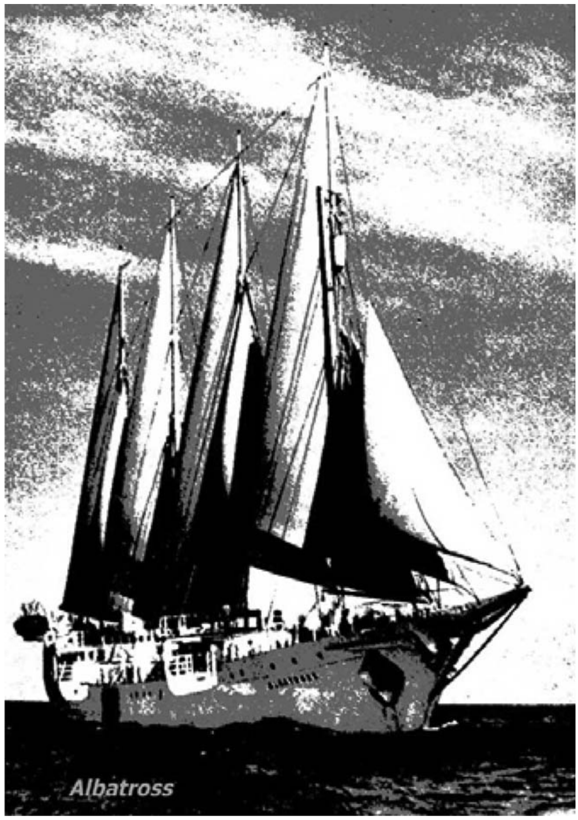

Fig. 3. The *Albatross* under sail. From Hans Petterson's book on the expedition (1953), with modifications.

drilling, which started in 1968; and (4) a boost to precise age assignments from the application of palaeomagnetic dating to deep-sea sediments.

Other important developments in the early 1970s concerned the improvement of analytical equipment, including mass spectrometry (an endeavour led by Nicholas Shackleton at Cambridge; Shackleton 1965) and the introduction of sophisticated statistical methods to the interpretation of fossil assemblages using ever more powerful computing devices (a task initiated by John Imbrie and Nilva Kipp at Brown University in Rhode Island; Imbrie & Kipp 1971).

A new emphasis also arose on climatic change from concerns about the possible impact of human activities on climate (i.e. the threat that the increase in carbon dioxide in the atmosphere will produce profound and unwelcome changes). It soon became clear to a number of Earth scientists that the predictive powers of climate models emerge when testing the models in the reconstruction of past conditions. In addition, the study of the history of climatic change recorded in deep-sea sediments could considerably broaden the horizons of the researchers involved. These aims gave great impetus to the field.

Palaeoceanography is an observational science based on the study of fossils and sediments. The clues to history therefore reside within the properties of these objects. To use fossils, we need to know how organisms lived before they became fossils, and what happened to the shells on the sea floor. The task requires observing the relevant processes in the present ocean, making such studies part of palaeoceanography even though they do not address history.

Emphasis has traditionally been on distribution patterns, along with environmental corollaries (Cushman & Parker 1931; Natland 1933; Parker 1954; Bandy 1956; Bé 1959; Funnell 1967). Palaeoceanography became the focus with invoking oceanographic processes and climate cycles (for Quaternary sediments) in explaining the patterns found (Schott 1935; Cushman & Henbest 1940; Arrhenius 1952; Phleger *et al.* 1953; Emiliani 1955, 1966; Ericson & Wollin 1956, 1964; Ericson *et al.* 1956; Boltovskoy 1959; Bradshaw 1959; Phleger 1960; Olausson 1961; Lidz 1966; Berger

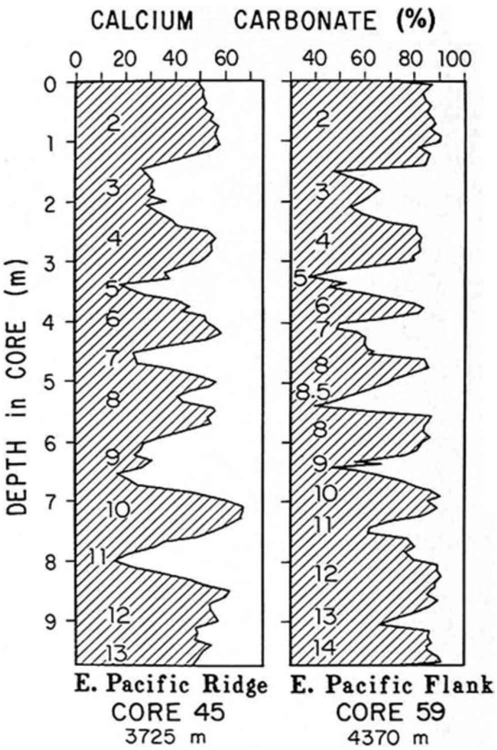

Fig. 4. Carbonate cycles in deep-sea sediments of the eastern equatorial Pacific. From the Reports of the Swedish Deep-Sea Expedition (Arrhenius 1952).

1968; Seibold 1970; Funnell & Riedel 1971). The transition was in essence completed by 1965, judging from the papers presented at the International Union for Quaternary Science (INQUA) in 1965 (Sears 1967). There is one paper in the Sears volume that attempts to 'explain' changes in deep-sea fauna in the North Pacific by proposing vertical movements of the sea floor of more than one kilometre over large regions (rather than by calling on changing bottom water properties and productivity in surface waters). Incidentally, the same approach using phantom tectonics is last seen in 1970, in an attempt to explain changes in Cenozoic patterns of carbonate sedimentation in the South Atlantic.

One major discovery of the *Albatross* Expedition was the presence of carbonate cycles that could be correlated over vast distances (Arrhenius 1952). The nature of these cycles has been the subject of much discussion. All sedimentation processes involving fossils have at least two aspects: shell production and shell preservation. A third one, mixing on the sea floor, is usually also present. A fourth aspect, diagenesis, must be considered in the older sediments where time has been sufficient to change the morphology and chemical properties of fossils. The latter process is commonly ignored for Quaternary sediments.

Production (i.e. supply of shells) is not a number but a complex process involving sunlight, nutrient supply, stratification of the upper ocean and the nature of the food web, in addition to an array of life-styles exhibited by the various species destined to leave a record. Preservation, in turn, is impacted by the nature of the shells, the relevant chemical state of the ocean, depth of deposition, bottom currents and local rates of deposition of the various sediment components (especially of organic carbon but also of inert materials). With so many unknowns, it is advisable to control reconstruction by several different proxies; this principle is reminiscent of elementary algebra, which requires several equations to solve for several unknowns.

Differences of opinion readily arise whenever there is insufficient proxy information available for reliable reconstruction, which is commonly the situation. The most familiar proxies are carbonate and opal percentages, species content, fragmentation of shells and the isotopes of oxygen and carbon within the calcareous fossils (see Fischer & Wefer 1999). Other proxies include a number of trace metals and an array of organic compounds said to reflect the temperature and chemistry of the seawater they originated in. Unfortunately, neither average nor seasonal temperature is usually sufficiently well constrained in palaeoceanographic reconstruction to make the reconstruction of (carbon-) chemistry highly reliable.

On the nature of the *Albatross* carbonate cycles

The *Albatross* carbonate cycles (Fig. 4) perhaps represent the simplest case history for the use of clues in ocean reconstruction. In his original write-up, Arrhenius related the cycles to changing productivity in the eastern equatorial Pacific, invoking strengthened trade winds during glacial times which would have increased the rate of upwelling along the Equator. Arrhenius cited a number of clues suggesting the contrast in productivity of overlying waters, including evidence from diatom deposition (obtained from R. W. Kolbe). Two decades later I suggested that the carbonate cycles mainly represent dissolution cycles (Berger 1973). Fifteen years later, Arrhenius (1988) defended his original position. The discussion continues (e.g. see Mekik & Raterink 2008). In recent decades, Arrhenius' preference for large changes in the intensity of production between glacial and interglacial times has been greatly strengthened by the analysis of ODP Site 805 in terms of foraminifer content (Yasuda *et al.* 1993) and by the chemical analysis

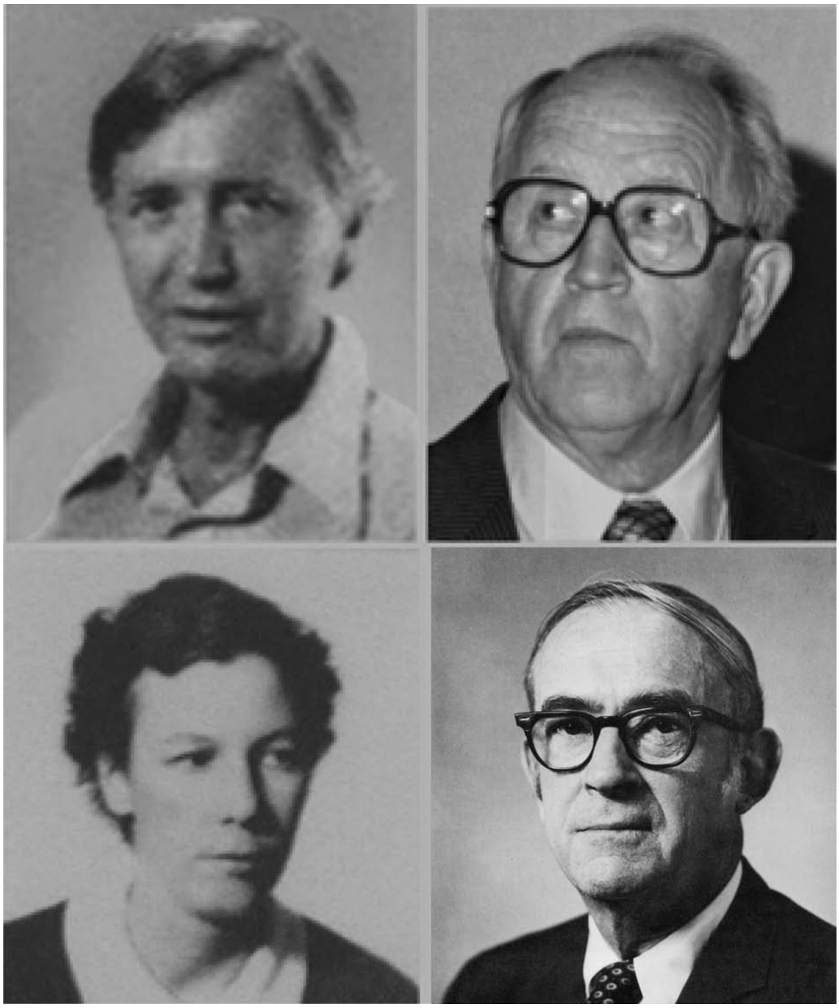

Fig. 5. Pioneers who worked on *Albatross* cores: upper: G. O. S. Arrhenius (*c.* 1970) and E. Olausson (*c.* 1990), lower: F. L. Parker (*c.* 1950) and F. B. Phleger (*c.* 1970). Sources: S. I. O. archives and Jan Backman (E. O.).

of cores (for combustible component, Perks & Keeling 1998; for barite, Paytan *et al.* 1996). In any case, there can be no question whatever that the supply of biogenous sediments to the sea floor at the Pacific Equator is greater than in regions to the north and south (as originally pointed out by Arrhenius), supporting a strong link of sedimentation to upwelling. It now seems well established that dissolution varies together with export production in response to the ice-age cycles (e.g. Murray *et al.* 2000).

The question is therefore not whether the *Albatross* carbonate cycles are production cycles or dissolution cycles. Rather, the question is to what degree they are one and the other. The problem can be constrained by thought experiment. First, based on abundant observations going back to John Murray of the *Challenger* Expedition (summarized in Parker & Berger 1971; Berger 1973), we can erect a schematic diagram of present-day carbonate preservation and associated states of the foraminifer shells (Fig. 6). The relevant portion of this graph is the gradient below 3.5 km, which depicts an additional loss of 6% of original carbonate supply for each 100 m depth increment. This is in good agreement with various other determinations of carbonate loss with depth (e.g. van Andel *et al.* 1975).

The low values of carbonate percent in Core 45 (Fig. 4) make it likely that the sediments are affected by dilution from opal supply at this rather eastern location, a factor that would complicate matters. Furthermore, entire tests per gram of

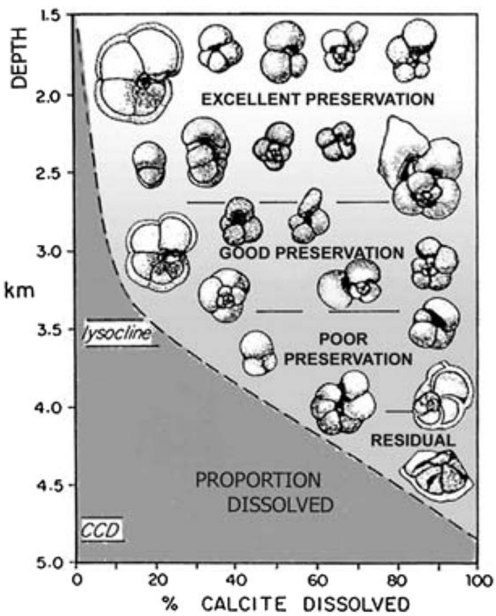

Fig. 6. Schematic of carbonate dissolution with depth and associated preservation of foraminifer tests. From Berger & Wefer (2009). Drawings by F. L. Parker, modified.

sediment seem low (Arrhenius 1952, fig. 2.45.1) and more than half of the samples have a benthic content of >10% of the total fauna (Arrhenius 1952, fig. 2.45.7.2), commonly a sign of considerable carbonate dissolution (Parker & Berger 1971). Let us therefore focus on Core 59. This core, taken at a depth of 4370 m, has variations of carbonate content in the range *c.* 50–80%. What is the vertical displacement of the profile in Figure 6 that would produce such a range?

The answer depends somewhat on assumptions about initial conditions. In the high-carbonate samples there are 80 parts of carbonate for 20 parts of non-carbonate. Getting to 50% implies reducing the 80/20 ratio to 20/20; that is, a removal of 60 parts of carbonate. If the initial condition was 100 parts of carbonate, this requires a vertical depth change of the dissolution profile of 1 km. In other words, the calcite compensation depth (CCD) would have to have moved by that much between cold and warm stages of the ice ages. If the initial condition was 80 parts of carbonate, removing 60 parts implies dissolving 75% for a corresponding depth difference of 1250 m. The data suggest that in fact the 80% carbonate state represents partial dissolution already (based on the spread in foraminifer content per gram and maximum carbonate values; Arrhenius 1952, fig. 2.59.7.1). The 1 km vertical motion of dissolution levels therefore has the greater merit as an explanation for the carbonate variations in Core 59. In addition, a vertical displacement of the profile by *c.* 1 km also appears reasonable in light of other information in equatorial Pacific sediments (vertical range of lower boundary of *G. rubescens* in box cores: Berger *et al.* 1977, fig. 18; vertical range of carbonate percentages in Pacific cores: Farrell & Prell 1989).

It appears from this thought experiment that the dissolution hypothesis is viable and in agreement with observations.

Now let us turn to the production hypothesis. The fundamental observation is that the CCD is lowered at the equator and that this lowering must reflect the high equatorial production, which is also seen in increased rates of accumulation of both carbonate and silica (Arrhenius 1952). It is entirely reasonable to expect that wherever production is anomalously high, it should also be subject to substantial changes through time. If we assume that the diluting material is abiogenic, a variation between 80% and 50% carbonate implies a factor of change in the supply of carbonate of 4 (80/20 v. 20/20). Part of the diluting material is opal however, and its supply grows along with the supply of carbonate. Thus, the factor-of-4 estimate is a lower boundary. If only one-half of the diluting material is independent of production (i.e. we go with 20/(10 + 10) for low production), the required change in export is a factor of 10 (i.e. $200/(10 + 40)$, with the 40 parts of opal assumed conservatively, to get 80/20).

While the factor of change of production between glacial and postglacial is not known with great confidence, it is a safe assumption that it is lower than the range 4–10 (Herguera & Berger 1991; Herguera 2000). If we admit a more conservative factor of 2 or 3 (still in strong support of the basic Arrhenius hypothesis), we then need to make up the missing difference by dissolution effects. The action necessary to explain the residual change of carbonate percentages turns out to imply a vertical CCD range of *c.* 400–700m, depending on assumptions. No problem.

It was Sir John Murray (1841–1914) of the *Challenger* Expedition who first realized that carbonate dissolution is pervasive. Schott (1935), Revelle (1944) and Phleger *et al.* (1953) (among others) pointed out its various effects on foraminifer assemblages on the sea floor. That the intensity of dissolution varies locally through time is no longer in doubt. Through effects from differential dissolution, it puts severe constraints on the interpretation of foraminifer assemblages in terms of changes in surface water conditions (Fig. 7). For example, the warm-water species *Globigerinoides ruber* is quite susceptible to dissolution, compared with the more robust species *Globorotalia menardii*, *Globorotalia*

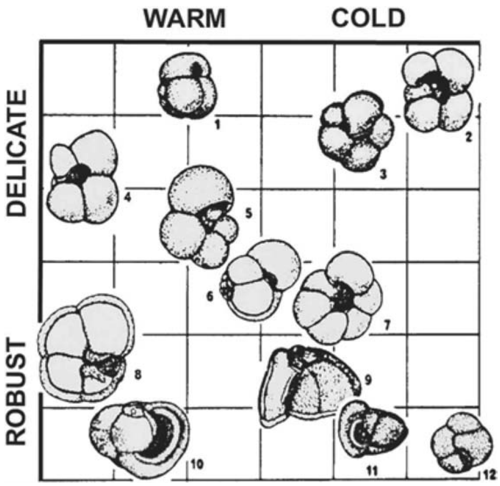

Fig. 7. Placement of common planktonic foraminifers in a two-dimensional framework of temperature preference and dissolution susceptibility. 1, *G. ruber*; 2, *G. bulloides*; 3, *G. quniqueloba*; 4, G. *sacculifer*; 5, *G. aequilateralis*; 6, *G. glutinata*; 7, *N. dutertrei*; 8, *G. menardii*; 9, *G. truncatulinoides*; 10, *P. obliquiloculata*; 11, *G. inflata*; 12, *N. pachyderma*. Drawings modified from F. L. Parker (1962).

tumida and *Pulleniatina obliquiloculata.* Low abundance of *G. ruber* within an assemblage does not necessarily signify cool waters, but may simply indicate preferential post-depositional removal. These principles were well recognized by the late 1960s (e.g. Olausson 1967; Ruddiman & Heezen 1967; Berger 1968).

On the importance of oxygen isotopes

Oxygen isotopes, determined for the calcareous shells of planktonic and benthic foraminifers, have become the master signal of ocean history, comparable in importance to the inventory of the ocean's temperature distributions in oceanographic studies. The fundamental pioneer work was completed by Cesare Emiliani (1922–1995), who began his studies by analysing the composition of foraminifers from the *Albatross* Expedition.

Emiliani started his remarkable career as a micropalaeontologist, earning a doctorate in geology from the University of Bologna in 1945. He obtained his PhD (in 1950) with isotopic studies in Chicago, working in the laboratory of Harold Urey. Regarding his main line of research, Emiliani wrote to me as follows (pers. comm. 1994; see Berger 2002*a*): 'I was drafted into Urey's paleotemperature project by Heinz Lowenstam who wanted to know the isotopic composition of planktic forams (I was the only one in Chicago at that time who knew anything about forams)'.

Emiliani happened to be in the right place at the right time, with the right kind of background. The *Albatross* Expedition had just completed its tour around the world and had come back with a rich treasure of long cores. As Emiliani put it (pers. comm. 1994; see Berger 2002*a*): 'When, in 1951, Hans Pettersson [the leader of the *Albatross* Expedition] came to Chicago and gave a talk about his famous expedition, I saw that deep-sea cores was the way to go. I visited him in Stockholm the following summer and he was most generous with samples from his cores'.

In the course of his early studies, published in his famous paper on Pleistocene temperatures (1955), Emiliani found evidence that the ice ages of the last half million years were cyclic in nature, which gave strong support to Milankovitch's (1930) hypothesis that the ice ages reflect orbital forcing. In addition to samples from cores from the *Albatross* Expedition, he used samples provided by David Ericson at Lamont (who knew which cores correlated well with others and were therefore trustworthy). In subsequent decades he used cores raised by expeditions out of Miami. Emiliani's method, adapted from his mentors Harold Urey, Sam Epstein and Heinz Lowenstam and discussed with his colleagues at Chicago (Harmon Craig, Toshiko Mayeda), became the standard procedure for interpreting the deep-sea record in terms of ocean and climate history. Emiliani introduced a timescale suggesting that the cycles are typically 40 ka in duration, and he defended this scale for almost 20 years after first proposing it. He also thought that temperature was a more important influence on oxygen isotope variations of the ocean than the build-up and decay of Northern Hemisphere ice sheets. Both the notion about timescale and that about temperature proved incorrect. However, his insistence that the Milankovitch mechanism, in conjunction with ice dynamics and crustal response to loading, is the driving force behind the climate cycles of the Quaternary proved viable.

One immediate benefit of isotope stratigraphy, regardless of the precise meaning of the isotopes and regardless of the accuracy of the age scale applied, was the opportunity for detailed correlation between cores separated by long distances (Fig. 8). Emiliani (1958) pointed out an important corollary: while the sedimentation rates differed markedly from one place to the next, they seemed to stay roughly constant in any one location. (The stratigraphies shown are based on the study of *Albatross* cores from the North Atlantic, the Caribbean and the Mediterranean, and also include two Lamont cores, prefixed by the letter 'A'.) The

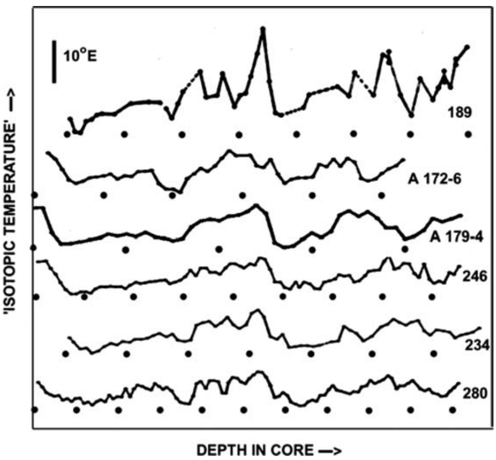

Fig. 8. Correlation of six cores from the North Atlantic, the Mediterranean and the Caribbean using isotope stratigraphy. Isotopic scale in terms of 'Emiliani temperature', with 10° corresponding to a change in $\delta^{18}O$ of *c.* 2.75 per mille. Depth in core shown as bullets 1 m apart, starting with zero to the left. Modified after Emiliani (1958).

sequences comprise the last two 100 ka cycles, back to glacial Stage 8.

Emiliani applied one sedimentation rate throughout for each core to produce the correlation seen. It is readily noted that the amplitudes within the Mediterranean (Core 189) are higher than elsewhere, reflecting the fact that the sea is surrounded by land and therefore has a high glacial–interglacial temperature range (Emiliani 1958).

Since the 1970s, the oxygen isotope composition of planktonic and benthic foraminifers provides the guiding stratigraphy for the study of Quaternary deep-sea sediments, as well as for older sequences. Isotope stratigraphy has revolutionized our understanding of climate change on long time scales. In particular, it is now firmly established that ice-age fluctuations are cyclic, and that they are driven by changes in seasonal insolation in high northern latitudes, as outlined by Milankovitch. It is also now generally accepted that the deep waters of the ocean cooled in response to planetary cooling in the Cenozoic, a cooling that eventually led into the northern ice ages about 3 Ma ago (Berggren 1972). A general Cenozoic cooling trend in deep waters had been proposed by Emiliani (1961) on the basis of very few samples. His guess proved astute.

Emiliani introduced a concept to deep-sea studies he labelled 'isotopic temperature'. It is based on the discovery by Urey that, when in thermodynamic equilibrium, calcite ($CaCO_3$) precipitating from seawater (H_2O) differs in its ratio of ^{18}O to ^{16}O from the ratio in the water, and this difference decreases with increasing temperature. Shells precipitated in equilibrium with seawater are enriched in ^{18}O relative to seawater, but less so at higher temperatures. The equation relating temperature of precipitation to the oxygen isotopic composition of mollusc shells was found to be:

$$T = 16.5 - 4.3(\delta - A) + 0.14(\delta - A)^2. \quad (1)$$

where δ is the difference in ‰ in the isotope ratios of sample and standard of the standard:

$$\delta = 1000 * (R \text{ of sample} - R \text{ of standard})/(R \text{ of standard})$$

where R is the ratio between the two isotopes $^{18}O/^{16}O$, and A is the δ of the seawater within which the sample (calcitic shell, foraminifer) was precipitated.

Equation (1), based on the work by Epstein *et al.* (1953), was used by Emiliani (1955). It is still in use today, with minor adjustments based on additional experiments using foraminifers. A is usually not known and must be estimated before the temperature T can be calculated. The standard is usually taken as 'PDB', for a belemnite from the Pee Dee formation in South Carolina that was originally analysed by Urey's group.

During the ice-age climate fluctuations, the term A varies because polar ice caps are impoverished in ^{18}O (or enriched in ^{16}O, which means the same thing). Piling up large ice masses has the effect of changing the isotopic composition of seawater, since the overall global average ratio between the two isotopes is preserved (isotopes being neither made nor destroyed). Emiliani was well aware of this effect, as well as of more local effects of evaporation and precipitation. He carefully listed the various problems that interfere with reading the oxygen isotope record in terms of palaeotemperature (the ice effect, geographical variation in isotopic composition, vital effects, seasonal growth of shells, growth of foraminifers at various depths in the water). But once he settled on the preferred species for analysis (*G. rubra*, *G. sacculifer*) to exclude the depth-of-growth effect, he presented his data as indices of 'isotopic temperature' in his graphs, thus implying that these various other interfering factors do not matter all that much and can be captured by a single number for maximum A. Although the size of the maximum A had to be adjusted in subsequent studies, his assessment that A may be taken as a constant proportion of the overall change in oxygen isotopes was largely adopted and is widely used.

It was precisely the term A (i.e. the amount of global change in δ-value of seawater as a consequence of the build-up and decay of polar ice) that proved to be the bugbear of Emiliani's interpretation. Emiliani estimated an overall increase of about 0.4‰ over present values for this effect. With an isotopic range near 2‰ for the contrast between warm and cold periods in the Caribbean, this left 1.6‰ to be explained by temperature (assuming all other factors did not change the range, but merely offset values to a different mean). From this emerges the 7° range in temperature seen in the surface-water species (*Globigerinoides rubra*, *Globigerinoides sacculifer*) in some of Emiliani's graphs. In the abstract of 'Pleistocene temperatures', Emiliani (1955, p. 538) wrote: 'Oxygen isotopic analyses of pelagic Foraminifera from Atlantic, Caribbean, and Pacific deep-sea cores indicate that the temperature of superficial waters in the equatorial Atlantic and Caribbean underwent periodic oscillations during the Pleistocene with an amplitude of about 6 °C'.

The error regarding *A* was eventually recognized and corrected, but not by Emiliani who was locked into a limited range by the results from his Pacific studies on unsuitable material. The correct value turns out to be near 1.1 or 1.2‰, as first suggested by Eric Olausson (1965) and N. J. Shackleton (1967), and as confirmed by the results from ice core studies by Dansgaard & Tauber (1969). Ironically, Emiliani's (1955) data for benthic foraminifers could have suggested the correct answer for maximum *A*, assuming that abyssal temperatures did not change much. There is, however, the significant problem that different benthic species yield different results (Duplessy *et al.* 1970; Woodruff *et al.* 1980; Vincent *et al.* 1981), which invalidates comparison of average isotope values between assemblages from different horizons, other than assemblages restricted to the same species.

It is perhaps unfortunate that Emiliani decided to plot his fascinating new data in terms of an idiosyncratic temperature scale (hence the term 'Emiliani temperature' in Fig. 8). However, many of his contributions are unaffected by this questionable decision. The stages he assigned to glacial periods and to interglacials, back to Stage 17, have become part of the master stratigraphy of Quaternary history, regardless of the size of the glacial effect on isotopes. Also, the fact that correlation between cores from different regions is excellent is not a function of assigning magnitude to this effect, although admittedly a large glacial effect on isotopic values makes the results more plausible. Such correlation has become indispensable in palaeoceanography, including the production of synoptic maps for 'time slices'.

Emiliani's other chief problem – the one regarding the timescale – stemmed largely from his conviction that Milankovitch theory would provide a template for the chronology (a concept now widely accepted and used). Thus, whenever someone questioned his timescale, he reacted as though they were attacking Milankovitch theory. Unfortunately, his original attempts to extrapolate sedimentation rates down-core from radiocarbon dates had a low chance of success because the uppermost portions of cores (where radiocarbon dates are made) tend to be disturbed in different ways during the coring process. In the end, the correct timescale was a matter of co-ordinating isotope stratigraphy with the results from palaeomagnetism, applying the date found in basalt layers for the Matuyama–Brunhes boundary to cores with known magnetic stratigraphy (as in Shackleton & Opdyke 1973). The agreement of dating by that method and by Milankovitch tuning (urged by Shackleton *et al.* 1990) is the strongest argument yet for the correctness of Milankovitch theory.

Emiliani realized that Milankovitch theory is not in fact capable of describing the observed cycles in simple fashion, being incomplete as a mechanism for explaining ice ages (Emiliani & Geiss 1959). The solar radiation schedule emphasized by Milankovitch represents driving forces, but the ice-age history represents integration of Earth's response to such forces over many millennia, a response that is modulated by crustal reaction to loading and by ice dynamics. It took some time for these ideas to enter into the appropriate models. We have to look to the late 1970s and mid-1980s for a competent treatment of the matter. One such is in the two-volume Milankovitch symposium edited by A. Berger *et al.* (1984): a model by Pollard (1984). Also in this symposium we find an isotopic stratigraphy for the Late Quaternary, known as the SPECMAP timescale, which is reliable back to *c.* 650 ka (Imbrie *et al.* 1984). Well-dated isotope stratigraphies are now available for the entire Quaternary and into the Late Tertiary (Zachos *et al.* 2001; Lisiecki & Raymo 2005). A table for the last million years based on these published data, re-tuned assuming Milankovitch forcing of the differentials of the isotope record, can be found in Berger (2011).

On the nature and timing of deglaciation

As a period of pronounced global warming, the time span of deglaciation (*c.* 20–7 ka) has taken centre stage in many discussions as a source of analogue situations for the type of response that may be expected of the Earth's climate system to human-forced warming from the rise of carbon dioxide and other greenhouse gases in the atmosphere. The overall difference in average global temperatures between glacial and postglacial conditions is somewhere near 5 °C (the exact value is not known). The last time the global temperature range changed over this interval, it did so over a time span of roughly 10 ka. The value of 5 °C is at the upper end of a range of estimates for future warming *by the end of the present century* (IPCC 2007). Clearly, even should the real value of warming by the end of the century turn out to be five times smaller (as urged by some), overall deglaciation cannot provide a valid analogue to such a fast rate of change. However, quite possibly there are times of extremely rapid change within spans of less than a century, embedded within the period of deglaciation (Dansgaard *et al.* 1993; Grootes *et al.* 1993; Severinghaus & Brook 1999). The evidence for such abrupt warming comes mainly from ice cores and is therefore largely restricted to high latitudes.

In any case, the recorded sequences describing deglaciation history are of great interest in the context of current human impacts on climate change. Such interest focuses, for example, on the potential rates of ice sheet disintegration (Mercer 1978; Hughes 1987; Fleming *et al.* 1998; Zwally *et al.* 2002; Berger 2008) and on the relationships between general and deep-sea warming and ice mass and the rise of carbon dioxide (Denton & Hughes 1983; Fastook 1984; Oeschger *et al.* 1984; Stott *et al.* 2007).

One major enigma is the problem of the Younger Dryas episode, a return to glacial conditions in the Northern Hemisphere in the middle of the deglaciation interval (Dansgaard *et al.* 1971; Duplessy *et al.* 1981; Ruddiman & McIntyre 1981; Lehman & Keigwin 1992; Broecker 1997; Bradley & England 2008). The reversal of the general postglacial warming that is represented by the Younger Dryas has led to much discussion, including speculations about the turning on and off of the production of North Atlantic Deep Water and changes in deep circulation in general (Jansen & Erlenkeuser 1985; Broecker & Denton 1989; Keigwin *et al.* 1991; Berger & Jansen 1995; McManus *et al.* 2004; Carlson *et al.* 2008). Many of the suggestions regarding the origin of the Younger Dryas are based on information from the isotopic stratigraphy of benthic foraminifers (the shells of which carry clues about the ventilation of deep waters in their chemistry; e.g. Jansen & Erlenkeuser 1985; Boyle & Keigwin 1987). Additional clues come from the study of sea-level changes based on coral (Fairbanks 1989; Bard *et al.* 1990). Other constraints are based on modelling of climate change and inferred response of the circulation.

In all these discussions, questions raised include the possible effects of high rates of meltwater input and warming of sea surface waters, and the possible effects of the ensuing increased stratification on deep circulation. The debate spans many decades (Olausson 1965; Berger *et al.* 1977; Broecker & Denton 1989). Of special interest is the possibility that meltwater was stored in periglacial lakes and released to the sea in pulses (Emiliani *et al.* 1975; Kennett & Shackleton 1975; Rooth 1982, 1990; Broecker *et al.* 1989; Broecker 2006). The discovery of the relatively young age of the last glacial maximum in the 1950s, implying a short time span for melting enormous masses of ice and for moving from glacial conditions into postglacial conditions, is an achievement that owes much to radiocarbon dating of deep-sea sediments during pioneer time (Rubin & Suess 1955; Suess 1956), with strong connections to the study of oxygen isotopes in foraminifers.

That the time of deglaciation was potentially a period of abrupt climate change was realized at the end of the 1950s (Broecker *et al.* 1960). Emiliani was reluctant to accept the concept, however. Apparently he thought that sediment disturbance might play an important role in at least some of the instances of sudden change reported by others. Thus, regarding the abrupt transition from glacial to postglacial conditions described by Broecker *et al.* (1960), he wrote as follows (Emiliani 1961, p. 531): ‘Broecker *et al.* (1958, 1960) have marshaled evidence suggesting that an abrupt temperature rise may have occurred within a few hundred years about 11 000 years ago. The deep-sea cores, generally, do not show such an abruptness, probably a result of sediment reworking by bottom animals’.

It is not clear to me whether he means to say that the abruptness might be real, but would be prevented from showing itself by reworking if the core were of acceptable quality, or whether he means to draw attention to the creation of apparent abruptness by burrowing. Both are possible (one from mixing, the other from juxtaposing sediments derived from different levels in the core). Regardless of this complication in interpreting Emiliani’s comments, good cores do not in fact show such abruptness as he is objecting to. A likely cause in instances of abrupt change in deep-sea cores is that a portion of the sediment is missing; confirmation from other sequences is therefore essential.

Even as Emiliani expressed his scepticism regarding evidence for abruptness, he also realized that rapid change clearly needs to be addressed, given the evidence from the oxygen isotope stratigraphies (Emiliani 1961): ‘Disappearance of sea ice in the northern North Atlantic and elsewhere, causing a decrease of albedo, may explain the rapidity of the temperature rise (Emiliani & Geiss 1959). Sea ice, in fact, is thin and can be melted away very rapidly’. Perhaps his familiarity with the problems arising when reconstructing the last deglaciation prevented him from pursuing this idea further. The last deglaciation is of course a poor example for overall sudden change, since it shows a complicated isotopic signal with two or three steps (two of which have acquired labels since). Concerning sudden change, Emiliani apparently erred on the side of caution. It was left to Broecker & van Donk (1970) to boldly draw the 'terminations' on top of Emiliani's stratigraphy, thus introducing this very fruitful concept into the thinking about the ice ages.

At the heart of the concern with the rates of meltwater release is the rise of sea-level itself, in addition to any effects on deep circulation. It has become reasonably clear that the behaviour of disintegrating ice includes elements of mass wasting (e.g. Weertman 1957; Hughes 1987; Alley *et al.* 2005) in addition to the well-prescribed plastic flow depicted in diagrams illustrating compression of the stratigraphy with depth. The question arising is therefore: just how fast can major ice sheets disintegrate? The answer from geology is fairly straightforward: at a rate corresponding to an average rise of sea-level of 1 m per century for thousands of years. This answer results directly from dividing the overall eustatic rise of 120 m by the time available for melting; that is, *c.* 120 centuries. Presumably, within some portions of a dozen consecutive millennia, rates of sea-level rise can be substantially higher, since there are other portions when sea-level remains more or less fixed (as in the Younger Dryas). In addition, we cannot assume that during any given millennium the rise was steady.

There are two ingredients to calculating the overall rate in question: the difference in sea-level between glacial maximum and postglacial time, and the time available. The first value can be taken as the average depth of the shelf edge (Shepard 1963), apparently a good guess in the light of later studies (e.g. Fairbanks 1989). The second, as mentioned, is obtained from radiocarbon dating, more recently calibrated to more reliable radioisotope methods (Bard *et al.* 1990).

Emiliani's pioneering studies offer an early glimpse of what the transition looked like, and his data readily yield the overall sea-level rise when applying Shepard's criterion for change of sea-level (Fig. 9). Clearly, the dating still depended on the direct application of radiochemistry at the time. However, once oxygen isotope profiles were dated by such constraints, they themselves were then useful in further dating, an opportunity that has been widely used and appreciated. In any case, the oxygen isotopes specified the location of the glacial maximum and the length of the transition to the postglacial, and made the information portable between cores.

Emiliani's work (1966) puts the central age of the last glacial near 25 ka before present (Fig. 9), and defines the time span of deglaciation as *c.* 10 ka.

It may be noted that the same result is obtained when plotting Emiliani temperature values. The exact ratio of the influence of ice mass and temperature on the isotope index does not matter for the message in the graph, as long as it is assumed that ice mass and temperature run parallel to each other. This is in fact the common assumption applied when interpreting isotope records.

The CLIMAP reconstruction of the Last Glacial Maximum (CLIMAP Project Members 1976)

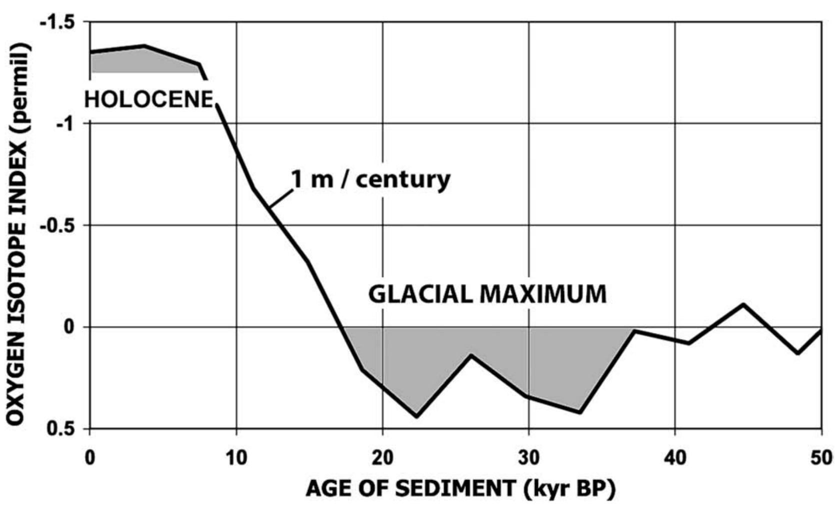

Fig. 9. Oxygen isotope stratigraphy of the uppermost part of Atlantic Core P6304–9, based on analysis of *G. sacculifer*, as published by Cesare Emiliani in 1966. Sedimentation rate is near 2.7 cm ka^{-1}, taken as invariable for this graph. Rate of sea-level rise (*c.* 1 m/century) based on taking the deglaciation period as 10 ka, and the total rise as *c.* 100 m.

is summarized in a map commonly referred to as the '18-k map,' which assigns an age of 18 ka to the Last Glacial Maximum.

Emiliani's data (Fig. 9) suggest that an age near 20 ka indeed seems appropriate for these sediments. The end of the last glacial may be marked by an unusual cooling, so that the coldest portion would be at the end of the interval marked 'Last Glacial Maximum' in the graph.

Later studies by a number of workers, but especially those of Fairbanks (1989) and Bard *et al.* (1990), have confirmed a strong two-step nature of the transition suspected earlier by researchers studying underwater terraces, physicists studying ice cores and palaeoceanographers studying foraminifer sequences in deep-sea cores from the North Atlantic. The reasons for the steepness of the two steps, the reasons why there are steps at all or why there are two steps rather than three or four, and the reasons for the duration of the cold spell separating the two steps of warming are as yet obscure. From this circumstance, it might be concluded that we do not understand the climate machine very well, especially where large ice masses are involved. This means, for the sake of clarity, that we do not have much of a grip on how the ice masses will react to future global warming.

While understanding is limited, it is possible to hold opinions as to what happened in a general way as outlined in the following (largely taken from Berger & Jansen 1995). The first of the two great meltwater pulses can be ascribed to Milankovitch forcing, which is centred near 14 ka, when July insolation in high northern latitudes was at its maximum. The implication is that ice sheets (or rather *some* ice sheets) were ready to go, and they did so when the summer sunlight was of sufficient intensity. The ice to go first, presumably, was that which was farthest from the poles and most vulnerable to being invaded by the sea, that is, the ice that was grounded below sea-level, for example in the North Sea. As sea-level rose, it lifted more of the ice off the ground, both in northern shelf seas (e.g. Barents Sea) and around Antarctica. A run-away situation ensued. When this rather vulnerable ice was used up, the process stopped. During a pause, while the remaining glacial ice warmed in its interior (heated by the refreezing of meltwater trickling down), the climate went back to the state it had left during the first melting step (i.e. it went into the condition known as the 'Younger Dryas'). There is nothing mysterious about the cold spell, in fact. After all, large ice sheets still dominated the scene with their whiteness and high elevation, so the arrival of a cold spell is not a surprise and no special explanations are needed.

The mystery, such as it is, is about the warming. What made the ice decay in such a short time? As suggested, a likely answer is that ice grounded below sea-level is especially vulnerable to collapse, as emphasized by the geologist T. J. Hughes (Hughes 1987; also see Alley *et al.* 2005). In addition, as noted by Olausson (1965) and many since, the addition of large amounts of meltwater to the northern North Atlantic interferes with the formation of deep water and thus has many ramifications for the global thermohaline circulation and the heat budget of the planet, as noted by Broecker & Denton (1989). Such interference with deep-water formation is difficult to demonstrate in any detail, because the foraminifer shell chemistry reflects *conditions* near the sea floor, not the rates of flow of deep water. Besides, a straight application of the principle of North Atlantic Deep Water shutdown from direct input of meltwater suffers from the fact that the initial warm spell (the Alleroed) lasted many centuries. Why did the system not shut down when meltwater production was at a maximum? Presumably, storage of meltwater in periglacial lakes and pulsed release (Rooth 1982, 1990) can deal with this problem by delaying the expected shutdown action and overwhelming the system upon release of the meltwater.

Mediterranean sapropels, Foraminifera and climate cycles

The *Albatross* Expedition retrieved a large number of cores in the Mediterranean Sea. This started the interesting process of making the basin the home of palaeoceanographic thought experiments, with considerable study and discussion. Of special interest are the black organic-rich layers called 'sapropels' (Kullenberg 1952; Olausson 1961) and the more general evidence on long-term warm–cold cycles seen in foraminifers (Parker 1958). The evidence for the uppermost sapropel horizon (tentatively dated by Olausson 1961, p. 353, as starting *c.* 11 ka ago and lasting 3–4 ka) suggested that deglaciation played a major role in changing the circulation in the Mediterranean. Thus it readily emerged as the problem that is first and foremost concerning palaeoceanography. Subsequent dating puts the event at the very end of deglaciation and the beginning of the Holocene (10–7 ka BP, according to Emeis 2009). This position favours a palaeoclimatic interpretation involving monsoon rains in Africa (as proposed by Rossignol-Strick 1983).

The debate about origins has focused largely on the question of whether the organic-rich layers reflect a lack of oxygen (at depth) or a marked spike in production (within surface layers). Given that, at present, anaerobic or dysoxic conditions tend to develop in oceanic regions of high

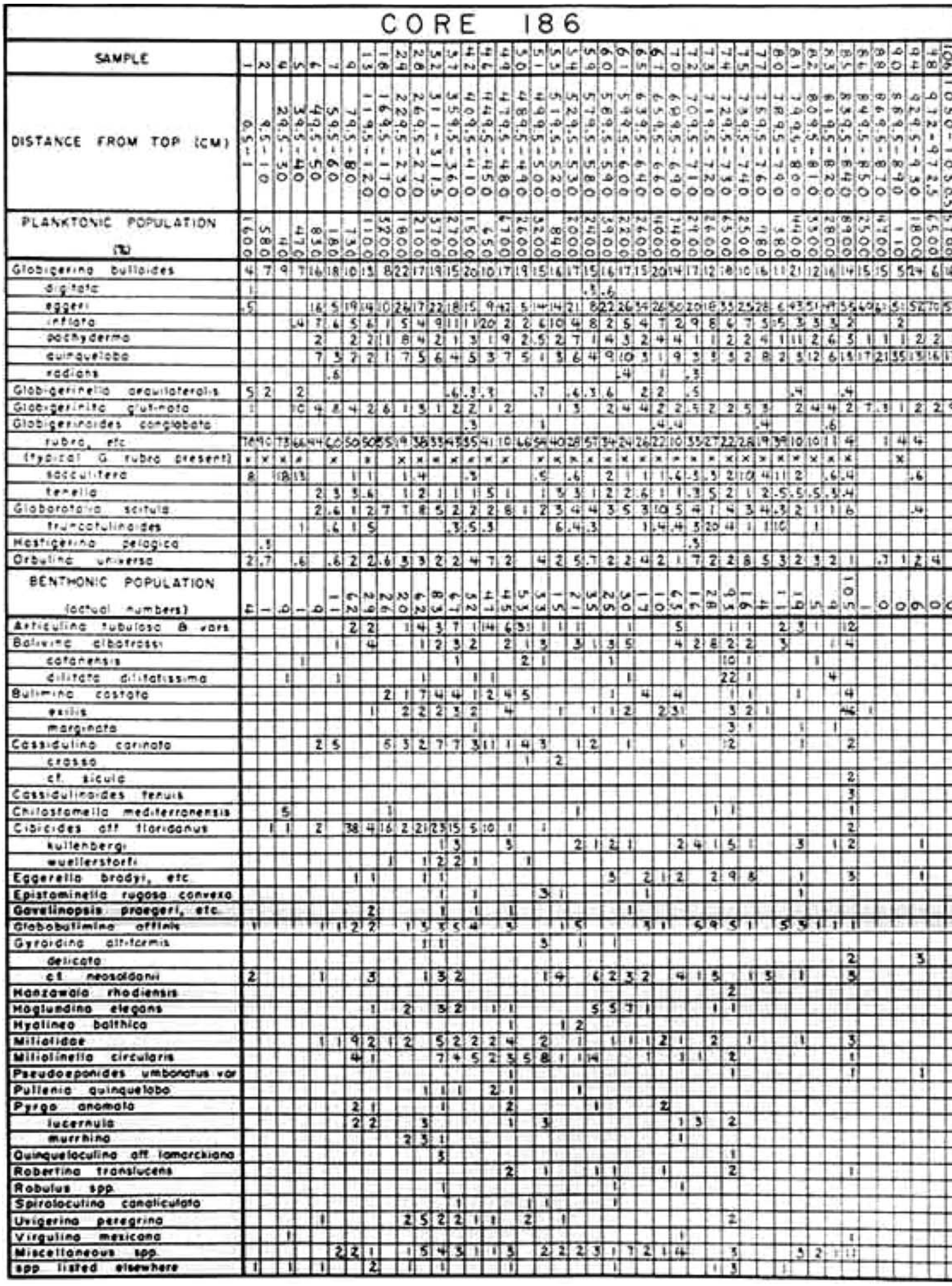

Fig. 10. Example of foraminifer inventory for an *Albatross* core taken in the eastern Mediterranean, counted and tabulated by Parker (1958). Such counts are the basis for reconstruction of climate cycles.

production, there might be some doubt that these are separable causes. More likely, they are aspects of the same change in overall circulation within the basin; that is, a shift from anti-estuarine to estuarine overall circulation. Such a change, in principle, increases production within the basin and decreases the oxygen level at depth. It is brought about by increasing the supply of fresh water to a given body of water in a basin (Seibold 1970), either by opening access to a new source in an adjacent basin (such as the Black Sea; Olausson 1961) or by changing the climate and input from rivers in the region, or both. The ensuing change in conditions is readily seen in the foraminifer assemblages (Muerdter *et al.* 1984), although such clues do not necessarily indicate causes. Identifying climate change (precipitation, river supply) as the main factor links the sapropel origin to orbitally driven climate variation (Cita *et al.* 1977; Rossignol-Strick 1983). Because of the link to orbital driving, the spacing of sapropel layers has provided a base for an astronomical timescale back into the Pliocene (Hilgen 1991).

An early attempt to document the back and forth between 'cold' and 'warm' times, based on the abundance of different types of foraminifers, is seen in graphs by Cushman & Henbest (1940, p. 36 ff.) but the assessments are not based on Schott-type counts (and are therefore qualitative). Also, somewhat surprisingly, Cushman and Henbest give separate determinations of 'colder' and 'warmer' conditions both for planktonic and for benthic foraminifers. Their choices are 'based on the temperature conditions exhibited by the Foraminifera in the top sample of each core and thus represent[ing] the present-day temperature of the water at that collecting station' (Cushman & Henbest 1940, p. 38). Calibration of planktonic distributions against present surface temperatures has, of course, been the norm for decades. However, one might question why benthic foraminifers living at great depth would care at all about the temperature of overlying waters, other than through a changing supply of organic matter.

A supremely pragmatic approach was taken by Frances Parker (see Berger 2002*b*). In her report on *Albatross* cores from the eastern Mediterranean, Frances L. Parker returned to the careful quantitative counting of foraminifers that she had introduced earlier in the 1930s roughly at the same time as Schott (Parker 1958; see Fig. 10). Such counts provide a quantitative basis for the reconstruction of climate cycles. From her discussions, her chief motivation in attempting such reconstruction appears to have been the desire for accurate correlation, both between the cores and into then prevalent schemes of ice-age sequences on land, as published by Penck & Brűckner (1909), Flint (1947) and Zeuner (1952). She considered that Emiliani's scale was quite incorrect, preferring the interpretations of Ericson and Wollin.

Unlike later practice she did not count samples taken at standardized intervals, but used subjective judgment on what constitutes significant change from one sample to the next, restricting counts to levels considered crucial: 'The foraminiferal assemblages of all the core samples were examined but population counts were made only at approximately 50 cm intervals, including top and bottom samples, unless intervening ones showed faunal change by gross examination' (1958, p. 223). She routinely counted both planktonic and benthic foraminifers, a feat requiring much experience.

Her treatment of benthic foraminifers was somewhat cursory, reflecting the general lack of understanding about the controls on distributions at the time (a lack that she mentioned). She emphasized a relationship to depth zones and associated water layers, but without further comment. Her data for depth distribution of benthic foraminifers (Parker 1958, table 3) show that the genera *Bolivina*, *Bulimina* and *Uvigerina* preferably occur in the uppermost thermocline zone (Fig. 11), presumably a reflection of their status as indicators of high food supply (not then an issue). Later work (e.g. Aksu *et al.* 1995) provided much evidence that foraminifers – both in the plankton and on the sea floor – are very useful indeed in providing clues to the environment of sapropel deposition. This type of integration was not yet common at the time of the pioneers.

Along with a lack of methods for integrating environmental information with the distribution of fossils, there was a lack of methods for the treatment of time series, such as represented by

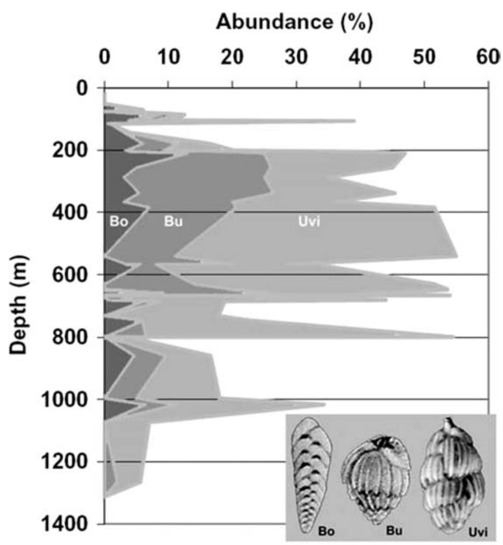

Fig. 11. Depth distribution of the genera *Bolivina*, *Bulimina* and *Uvigerina* in the eastern Mediterranean, based on surface samples. Data and inset illustrations from Parker (1958).

down-core abundance data. The methods were in existence, but they were only introduced to the field in the 1970s following the classic paper by Hays *et al*. (1976). Applying the correct mathematical treatment was later celebrated by Imbrie & Imbrie (1979) as the path to the mysteries of the ice ages.

Conclusions

Palaeoceanography – the way it is understood today – is linked to foraminifers, long cores and international collaboration. The first project bringing all these together was the Swedish *Albatross* Expedition which circled the world in the years 1947–1948, took 299 long cores in all major ocean basins in the tropics and subtropics and made them available to competent researchers in many countries. Just as the *Challenger* Expedition (1872–1876) serves as a marker point for the beginnings of oceanography, the *Albatross* readily serves the same purpose for palaeoceanography. This does not mean that there were no pioneer studies before the *Albatross* set out on her voyage. There were, and they include work on (short) cores from the central Atlantic (by W. Schott) taken during the *Meteor* Expedition (1927–1929) and the several long cores taken with the Piggott device in the North Atlantic by the *Lord Kelvin*, cable ship of the Western Union Telegraph Co. (1936) (Bradley *et al.* 1938, 1942; Cushman & Henbest 1940). Also, many of the pre-*Albatross* workers in biostratigraphy and micropalaeontology in general were certainly interested in problems we would now assign to palaeoceanography. The unique defining element – the link of micropalaeontology to ocean circulation and climate – only became prominent in the 1950s, however, with concepts discussed by Gustaf Arrhenius and Cesare Emiliani. These concepts guided much of the proliferation of studies in the 1960s, a time of transition to full development of the field, from about 1970. Important steps in the evolution of the new field were provided by the expansion of dating methods and of various statistical techniques, as well as modelling, all of which benefited greatly from the rapid growth of computing power.

References

Aksu, A. E., Yasar, D., Mudie, P. J. & Gillespie, H. 1995. Late glacial–Holocene paleoclimatic and paleoceanographic evolution of the Aegean Sea: micropalentological and stable isotope evidence. *Marine Micropaleontology*, **25**, 1–28.

Alley, R., Clark, P. U., Huybrechts, P. & Joughin, I. 2005. Ice sheets and sea-level changes. *Science*, **310**, 456–460.

Arrhenius, G. O. S. 1952. Sediment cores from the east Pacific. *Reports of the Swedish Deep-Sea Expedition 1947–1948*, **5**, 1–227.

Arrhenius, G. O. S. 1988. Rate of production, dissolution and accumulation of biogenic solids in the ocean. *Palaeogeography, Palaeoclimatology, Palaeoecology*, **67**, 119–146.

Bandy, O. L. 1956. Ecology of Foraminifera in northeastern Gulf of Mexico. *US Geological Survey, Professional Paper*, **274 G**, 179–204.

Bard, E., Hamelin, B., Fairbanks, R. G., Zindler, A., Arnold, M. & Mathieu, G. 1990. Calibration of the 14C timescale over the past 30 000 years using mass spectrometric U–Th ages from Barbados corals. *Nature*, **345**, 405–410.

Bé, A. W. H. 1959. Ecology of Recent planktonic Foraminifera. *Micropaleontology*, **5**, 77–100.

Berger, W. H. 1968. Planktonic foraminifera: selective solution and paleoclimatic interpretation. *Deep-Sea Research*, **15**, 31–43.

Berger, W. H. 1973. Deep sea carbonates: Pleistocene dissolution cycles. *Journal of Foraminiferal Research*, **3**, 187–195.

Berger, W. H. 2002*a*. Cesare Emiliani (1922–1995), pioneer of ice age studies and oxygen isotope stratigraphy. *Comptes Rendus Palevol*, **1**, 479–487.

Berger, W. H. 2002*b*. Frances Lawrence Parker (1906–2002), micropaleontologist and pioneer of paleoceanography. *Comptes Rendus Palevol*, **1**, 471–477.

Berger, W. H. 2008. Sea level in the Late Quaternary: patterns of variation and implications. *International Journal of Earth Sciences (Geologische Rundschau)*, **97**, 1143–1150.

Berger, W. H. 2011. Geologist at sea: aspects of ocean history. *Annual Review of Marine Science*, **3**, 1–34.

Berger, W. H. & Jansen, E. 1995. Younger Dryas episode: Ice collapse and super-fjord heat pump. *In*: Troelstra, S. R., van Hinte, J. E. & Ganssen, G. M. (eds) *The Younger Dryas*. North-Holland, Amsterdam, 61–105.

Berger, W. H. & Wefer, G. 2009. Marine Biogenic sediments. *In*: Gornitz, V. (ed.) *Encyclopedia of Paleoclimatology and Ancient Environments*. Springer, Dordrecht, 525–533.

Berger, W. H., Johnson, R. F. & Killingley, J. S. 1977. 'Unmixing' of the deep-sea record and the deglacial meltwater spike. *Nature*, **269**, 661–663.

Berger, A. L., Imbrie, J., Hays, J., Kukla, G. & Saltzman, B (eds) 1984. *Milankovitch and Climate: Understanding the Response to Astronomical Forcing*. Reidel, Dordrecht, **2**.

Berggren, W. A. 1972. Late Pliocene–Pleistocene glaciation. *Reports of the Deep-Sea Drilling Project*, **12**, 953–963.

Boltovskoy, E. 1959. Foraminifera as biological indicators in the study of ocean currents. *Micropaleontology*, **5**, 473–481.

Boyle, E. A. & Keigwin, L. D. 1987. North Atlantic thermohaline circulation during the last 20,000 years: link to high latitude surface temperature. *Nature*, **330**, 35–40.

Bradley, R. S. & England, J. H. 2008. The Younger Dryas and the sea of ancient ice. *Quaternary Research*, **70**, 1–10.

Bradley, W. H., Bramlette, M. N., Cushman, J. A., Henbest, L. G., Lohman, K. E. & Trask, P. D. 1938. The Geological Survey's work on the Piggot North Atlantic deep sea cores. *Proceedings of the American Philosophical Society*, **79**, 41–46.

Bradley, W. H., Piggot, C. S. *et al.* 1942. Geology and biology of North Atlantic deep-sea cores between Newfoundland and Ireland. *US Geological Survey, Professional Paper*, **196**, 1–163.

Bradshaw, J. S. 1959. Ecology of living planktonic Foraminifera in the North and Equatorial Pacific Ocean. *Contributions from the Cushman Laboratory for Foraminiferal Research*, **10**, 25–64.

Brady, H. B. 1884. Report on the Foraminifera dredged by the H.M.S. Challenger, during the years 1873–1876. *Reports of the Scientific Results of the Voyage of H.M.S. Challenger*, **9** (Zoology).

Broecker, W. S. 1997. Thermohaline circulation, the Achilles heel of our climate system: will manmade CO2 upset the current balance? *Science*, **278**, 1592–1588.

Broecker, W. S. 2006. Was the Younger Dryas triggered by a flood? *Science*, **312**, 1146–1148.

Broecker, W. S. & van Donk, J. 1970. Insolation changes, ice volumes, and the O-18 record in deep-sea cores. *Reviews of Geophysics and Space Physics*, **8**, 169–198.

Broecker, W. S. & Denton, G. H. 1989. The role of ocean-atmosphere reorganizations in glacial cycles. *Geochimica et Cosmochimica Acta*, **53**, 2465–2501.

Broecker, W. S., Turekian, K. K. & Heezen, B. C. 1958. The relation of deep sea sedimentation rates to variations in climate. *American Journal of Science*, **256**, 503–517.

Broecker, W. S., Ewing, M. & Heezen, B. C. 1960. Evidence of an abrupt change in climate close to 11 000 years ago. *American Journal of Science*, **258**, 429–448.

Broecker, W. S., Kennett, J. P., Flower, B. P., Teller, J. T., Trumbore, S., Bonani, G. & Wolfli, W. 1989. Routing of meltwater from Laurentide ice sheet during the Younger Dryas cold episode. *Nature*, **341**, 318–321.

Budyko, M. I. 1969. The effect of solar radiation variations on the climate of earth. *Tellus*, **21**, 611–619.

Carlson, A. E., Oppo, D. W., Came, R. E., LeGrande, A. N., Keigwin, L. D. & Curry, W. B. 2008. Subtropical Atlantic salinity variability and Atlantic meridional circulation during the last deglaciation. *Geology*, **36**, 991–994.

Chamberlin, T. C. 1906. On a possible reversal of deep-sea circulation and its influence on geologic climates. *Journal of Geology*, **14**, 363–373.

Cita, M. B., Vergnaud-Grazzini, C., Robert, C., Chamley, H., Diaranfi, N. & d'Onofrio, S. 1977. Paleoclimatic record of a long deep sea core from the eastern Mediterranean. *Quaternary Research*, **8**, 205–235.

CLIMAP PROJECT MEMBERS 1976. The surface of the ice-age earth. *Science*, **191**, 1131–1137.

Croll, J. 1875. *Climate and Time*. Daldy, Isbister & Co., London.

Cushman, J. A. & Parker, F. L. 1931. Recent Foraminifera from the Atlantic Coast of South America. *Proceedings of US Natural Museum*, **80**, 1–24, pls. 1–4. Smithsonian Institution, Washington D.C.

Cushman, J. A. & Henbest, L. G. 1940. Geology and biology of North Atlantic deep-sea cores between Newfoundland and Ireland. Pt. 2, Foraminifera. *US Geological Survey, Professional Paper*, **196-A**, 35–50.

Dansgaard, W. & Tauber, H. 1969. Glacier oxygen-18 content and Pleistocene ocean temperatures. *Science*, **166**, 499–502.

Dansgaard, W., Johnsen, S. J., Clausen, H. B. & Langway, C. C. 1971. Climatic record revealed by the camp century ice core. *In*: Turekian, K. K. (ed.) *The Late Cenozoic Glacial Ages*. Yale University Press, New Haven, 37–56.

Dansgaard, W., Johnsen, S. J. *et al.* 1993. Evidence for general instability of past climate from a 250-kyr ice-core record. *Nature*, **364**, 218–220.

Dayton, P. K. & Hessler, R. R. 1972. Role of biological disturbance in maintaining diversity in the deep sea. *Deep-Sea Research*, **19**, 199–208.

Denton, G. H. & Hughes, T. J. 1983. Milankovitch theory of ice ages: hypothesis of ice-sheet linkage between regional insolation and global climate. *Quaternary Research*, **20**, 125–144.

Douglas, R. & Woodruff, F. 1981. Deep sea benthic Foraminifera. *In*: Emiliani, C. (ed.) *The Oceanic Lithosphere*. John Wiley, New York, 1233–1237.

Duplessy, J. C., Lalou, C. & Vinot, A. C. 1970. Differential isotopic fractionation in benthic foraminifera and paleotemperatures reassessed. *Science*, **168**, 25–251.

Duplessy, J. C., Delibrias, G., Turon, J. L., Pujol, C. & Duprat, J. 1981. Deglacial warming of the northeastern North Atlantic Ocean: correlation with the paleo-climatic evolution of the European continent. *Palaeogeography, Palaeoclimatology, Palaeoecology*, **35**, 121–144.

Emeis, K.-C. 2009. Sapropels. *In*: Gornitz, V. (ed.) *Encyclopedia of Paleoclimatology and Ancient Environments*. Springer, Dordrecht, 875–876.

Emiliani, C. 1955. Pleistocene temperatures. *Journal of Geology*, **63**, 538–578.

Emiliani, C. 1958. Paleotemperature analysis of Core 280 and Pleistocene correlations. *Journal of Geology*, **66**, 264–275.

Emiliani, C. 1961. Cenozoic climatic changes as indicated by the stratigraphy and chronology of deep-sea cores of *Globigerina*-ooze facies. *Annals of the New York Academy of Sciences*, **95**, 521–536.

Emiliani, C. 1966. Paleotemperature analysis of Caribbean cores P6304–8 and P6304–9 and a generalized temperature curve for the past 425 000 years. *Journal of Geology*, **74**, 109–126.

Emiliani, C. & Geiss, J. 1959. On glaciations and their causes. *Geologische Rundschau*, **46**, 576–601.

Emiliani, C., Gartner, S. *et al.* 1975. Paleoclimatological analysis of Late Quaternary cores from the northeastern Gulf of Mexico. *Science*, **189**, 1083–1088.

Epstein, S., Buchsbaum, R., Lowenstam, H. A. & Urey, H. C. 1953. Revised carbonate-water isotopic temperature scale. *Geological Society of America Bulletin*, **64**, 1315–1325.

Ericson, D. B. & Wollin, G. 1956. Correlation of six cores from the equatorial Atlantic and the Caribbean. *Deep-Sea Research*, **3**, 104–125.

Ericson, D. B., Broecker, W. S., Kulp, J. L. & Wollin, G. 1956. Late-Pleistocene climates and deep-sea sediments. *Science*, **124**, 385–389.

Ericson, D. B. & Wollin, G. 1964. *The Deep and the Past*. Knopf, New York.

Fairbanks, R. G. 1989. A 17 000-year glacio-eustatic sea level record: influence of glacial melting rates on the Younger Dryas event and deep-ocean circulation. *Nature*, **342**, 637–642.

Farrell, J. W. & Prell, W. L. 1989. Climatic change and $CaCO_3$ preservation: an 800 000 year bathymetric reconstruction from the central equatorial Pacific Ocean. *Paleoceanography*, **4**, 447–466.

Fastook, J. L. 1984. West Antarctica, the sea-level controlled marine instability: past and future. *AGU Geophysical Monograph*, **29**, 275–287.

Fischer, G. & Wefer, G. (eds) 1999. *Use of Proxies in Paleoceanography: Examples from the South Atlantic*. Springer, Berlin Heidelberg.

Fleming, K., Johnston, P., Zwartz, D., Yokoyama, Y., Lambeck, K. & Chappell, J. 1998. Refining the eustatic sea-level curve since the Last Glacial Maximum. Using far-and intermediate-field sites. *Earth and Planetary Science Letters*, **163**, 327–342.

Flint, R. F. 1947. *Glacial Geology and the Pleistocene Epoch*. John Wiley & Sons, New York.

Funnell, B. M. 1967. Foraminifera and Radiolaria as depth indicators in the marine environment. *Marine Geology*, **5**, 333–347.

Funnell, B. M. & Riedel, W. R. (eds) 1971. *The Micropalaeontology of Oceans*. Cambridge University Press, Cambridge.

Gornitz, V. (ed.) 2009. *Encyclopedia of Paleoclimatology and Ancient Environments*. Springer, Dordrecht.

Grootes, P. M., Stuiver, M., White, J. W. C., Johnsen, S. & Jouzel, J. 1993. Comparison of oxygen isotope records from the GISP2 and GRIP Greenland ice cores. *Nature*, **366**, 552–554.

Hay, W. W. (ed.) 1974. *Studies in Paleo-oceanography*. Society of Economic Paleontologists and Mineralogists (SEPM) Special Publications, Tulsa, **20**.

Hay, W. W. 1988. Paleoceanography: a review for the GSA Centennial. *Geological Society of America Bulletin*, **100**, 1934–1956.

Hays, J. D., Imbrie, J. & Shackleton, N. J. 1976. Variations in the Earth's orbit: pacemaker of the Ice Ages. *Science*, **194**, 1121–1132.

Herguera, J. C. 2000. Last glacial paleoproductivity patterns in the eastern equatorial Pacific: Benthic foraminifera records. *Marine Micropaleontology*, **40**, 259–275.

Herguera, J. C. & Berger, W. H. 1991. Paleoproductivity from benthic foraminifera abundance: Glacial to postglacial change in the west-equatorial Pacific. *Geology*, **19**, 1173–1176.

Hessler, R. R. & Sanders, H. L. 1967. Faunal diversity in the deep sea. *Deep-Sea Research*, **14**, 65–78.

Hilgen, F. J. 1991. Astronomical calibration of Gauss and Matuyama sapropels in the Mediterranean and implication for the geomagnetic polarity time scale. *Earth and Planetary Science Letters*, **104**, 226–244.

Hughes, T. 1987. Ice dynamics and deglaciation models when ice sheets collapsed. *In*: Ruddiman, W. F. & Wright, H. E. (eds) *North America and Adjacent Oceans during the Last Deglaciation*. The Geology of North America, v. K-3. Geological Society of America, Boulder, Colorado, 183–220.

Imbrie, J. & Kipp, N. G. 1971. A new micropaleontological method for quantitative paleoclimatology: application to a Late Pleistocene Caribbean core. *In*: Turekian, K. K. (ed.) *The Late Cenozoic Glacial Ages*. Yale University Press, New Haven, 71–181.

Imbrie, J. & Imbrie, K. P. 1979. *Ice Ages: Solving the Mystery*. Enslow, Short Hills, New Jersey.

Imbrie, J., Hays, J. D. *et al.* 1984. The orbital theory of Pleistocene climate: support from a revised chronology of the marine δ^{18}O record. *In*: Berger, A. L., Imbrie, J., Hays, J. D., Kukla, J. & Saltzman, J. (eds) *Milankovitch and Climate (Part 1)*. Reidel, Dordrecht, 269–305.

IPCC 2007. *Climate Change 2007: Synthesis Report. Contributions of Working Groups I, II and III to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change*. IPCC, Geneva.

Jansen, D. & Erlenkeuser, H. 1985. Ocean circulation in the Norwegian Sea during the last deglaciation: isotopic evidence. *Palaeogeography, Palaeoclimatology, Palaeoecology*, **49**, 189–206.

Keigwin, L. D., Jones, G. A., Lehman, S. J. & Boyle, E. A. 1991. Deglacial meltwater discharge, North Atlantic deep circulation, and abrupt climate change. *Journal of Geophysical Research*, **96**, 16 811–16 826.

Kennett, J. P. 1977. Cenozoic evolution of Antarctic glaciation, the Circum-Antarctic Ocean and their impact on global paleoceanography. *Journal of Geophysical Research*, **82**, 3843–3859.

Kennett, J. P. 1982. *Marine Geology*. Prentice-Hall, Englewood Cliffs, NJ.

Kennett, J. P. & Shackleton, N. J. 1975. Laurentide ice sheet meltwater recorded in Gulf of Mexico deep-sea cores. *Science*, **188**, 147–150.

Kullenberg, B. 1952. On the salinity of the water contained in marine sediments. *Meddelingen Oceanographic Institute Göteborg*, **21**, 1–38.

Kullenberg, B. 1955. Deep-sea coring. *Reports of the Swedish Deep-Sea Expedition, 1947–1948*, **IV**, 35–96.

Lehman, S. J. & Keigwin, L. D. 1992. Sudden changes in North Atlantic circulation during the last deglaciation. *Nature*, **356**, 757–762.

Lidz, L. 1966. Deep-sea Pleistocene stratigraphy. *Science*, **154**, 1448–1452.

Lisiecki, L. E. & Raymo, M. E. 2005. A Pliocene-Pleistocene stack of 57 globally distributed benthic δ^{18}O records. *Paleoceanography*, **20**, PA1003. http://dx.doi.org/10.1029/2004PA001071

McManus, J. F., Francois, R., Gherardi, J.-M., Keigwin, L. D. & Brown-Leger, S. 2004. Collapse and rapid resumption of Atlantic meridional circulation linked to deglacial climate changes. *Nature*, **428**, 834–837.

Mercer, J. H. 1978. West Antarctic ice sheet and CO_2 greenhouse effect: A threat of disaster. *Nature*, **271**, 321–325.

Mekik, A. F. & Raterink, L. 2008. Effects of surface ocean conditions on deep sea calcite dissolution proxies in the tropical Pacific. *Paleoceanography*, **23**, PA1216, http://dx.doi.org/10.1029/2007PA001433

Milankovitch, M. 1930. *Mathematische Klimalehre und astronomische Theorie der Klimaschwankungen. Handbuch der Klimatologie, Bd 1, Teil A*. Bornträger, Berlin.

Miller, K. G., Fairbanks, R. G. & Mountain, G. S. 1987. Tertiary oxygen isotope synthesis, sea-level history, and continental margin erosion. *Paleoceanography*, **2**, 1–19.

Muerdter, D. R., Kennett, J. P. & Thunell, R. C. 1984. Late Quaternary sapropel sediments in the eastern Mediterranean Sea: faunal variations and chronology. *Quaternary Research*, **21**, 385–403.

Murray, J. 1897. On the distribution of pelagic Foraminifera at the surface and on the floor of the ocean. *Nature Science*, **10**, 17–27.

Murray, R., Knowlton, C., Leinen, M., Mix, A. & Polsky, C. 2000. Export production and carbonate dissolution in the central equatorial Pacific Ocean. *Geochimica et Cosmochimica Acta*, **60**, 3869–3878.

Natland, M. L. 1933. The temperature and depth-distribution of some recent and fossil Foraminifera in the southern California region. *Bulletin of the Scripps Institution of Oceanography, Technical Series*, **3**, 225–230.

Oeschger, H., Beer, J., Siegenthaler, U., Stauffer, B., Dansgaard, W. & Langway, C. C. 1984. Late glacial climate history from ice cores. *AGU Geophysical Monograph*, **129**, 299–306.

Olausson, E. 1961. Studies of deep-sea cores. Sediment cores from the Mediterranean Sea and the Red Sea. *Reports of the Swedish Deep-Sea Expedition, 1947–1948*, **8**, 335–391.

Olausson, E. 1965. Evidence of climatic changes in North Atlantic deep-sea cores, with remarks on isotopic paleotemperature analysis. *Progress in Oceanography*, **3**, 221–252.

Olausson, E. 1967. Climatological, geoeconomical, and paleooceanographic aspects on carbonate deposition. *Progress in Oceanography*, **4**, 245–265.

Olausson, E. 1996. *The Swedish Deep-Sea Expedition with the Albatross 1947–1948: A Summary of the Sediment Core Studies*. Graabo, Sweden.

Parker, F. L. 1954. Distribution of the Foraminifera in the north-eastern Gulf of Mexico. *Bulletin of the Museum of Comparative Zoology*, **111**, 453–588.

Parker, F. L. 1958. Eastern Mediterranean Foraminifera. *Reports of the Swedish Deep-Sea Expedition, 1947–1948*, **8**, 219–283.

Parker, F. L. 1962. Planktonic foraminiferal species in Pacific sediments. *Micropaleontology*, **8**, 219–254.

Parker, F. L. & Berger, W. H. 1971. Faunal and solution patterns of planktonic foraminifera in surface sediments of the South Pacific. *Deep-Sea Research*, **18**, 73–107.

Paytan, A., Kastner, M. & Chavez, F. 1996. Glacial to interglacial fluctuations in productivity in the equatorial Pacific as indicated by marine barite. *Science*, **274**, 1355–1357.

Penck, A. & Brűckner, E. 1909. *Die Alpen im Eiszeitalter*. Tauchnitz, C.H., Leipzig.

Perks, H. M. & Keeling, R. F. 1998. A 400 kyr record of combustion oxygen demand in the western equatorial Pacific: evidence for a precessionally forced climate response. *Paleoceanography*, **13**, 63–69.

Pettersson, H. 1953. *Westward Ho with the Albatross*. Dutton & Co., New York.

Phleger, F. B. 1960. *Ecology and Distribution of Recent Foraminifera*. The Johns Hopkins Press, Baltimore, MD.

Phleger, F. B., Parker, F. L. & Peirson, J. F. 1953. North Atlantic Foraminifera. *Reports of the Swedish Deep-Sea Expedition, 1947–1948*, **7**, 3–122.

Pollard, D. 1984. Some ice-age aspects of calving ice-sheet model. *In*: Berger, A. L., Imbrie, J., Hays, J., Kukla, G. & Saltzman, B. (eds) *Milankovitch and Climate, Part 2, 541–564*. Reidel, Dordrecht.

Raup, D. M. & Stanley, S. M. 1978. *Principles of Paleontology*, 2nd edn. W. H. Freeman & Co., San Francisco.

Revelle, R. 1944. *Marine Bottom Samples Collected in the Pacific Ocean by the Carnegie on its Seventh Cruise*. Scientific Results of Cruise VII of the Carnegie, Carnegie Institute, Washington, Publication, **556**.

Rooth, C. G. H. 1982. Hydrology and ocean circulation. *Progress in Oceanography*, **11**, 131–149.

Rooth, C. G. H. 1990. Meltwater Younger Dryas upheld. *Nature*, **343**, 702.

Rossignol-Strick, M. 1983. African monsoon, an immediate climate response to orbital insolation. *Nature*, **304**, 46–49.

Rubin, M. & Suess, H. E. 1955. U.S. Geological Survey radiocarbon dates II. *Science*, **121**, 481–488.

Ruddiman, W. F. & Heezen, B. C. 1967. Differential solution of planktonic foraminifera. *Deep-Sea Research*, **14**, 801–808.

Ruddiman, W. F. & McIntyre, A. 1981. The North Atlantic during the last deglaciation. *Palaeogeography, Palaeoclimatology, Palaeoecology*, **35**, 145–214.

Savin, S. M., Douglas, R. G. & Stehli, F. G. 1975. Tertiary marine paleotemperatures. *Geological Society of America Bulletin*, **86**, 1499–1510.

Schopf, T. J. M. 1980. *Paleoceanography*. Harvard University Press, Cambridge, Mass.

Schott, W. 1935. Die Foraminiferen in dem äquatorialen Teil des Atlantischen Ozeans. *Wissenschaftliche Ergebnisse, Deutsche Atlantische Expedition Meteor, 1925–1927*, **3**, 43–134.

Schott, W. 1966. Foraminiferenfauna und Stratigraphie der Tiefsee-sedimente im Nordatlantischen Ozean. *Reports of the Swedish Deep-Sea Expedition 1947–1948*, **VII**, 5 tbls, 6 pls.

Sears, M. (ed.) 1967. *The Quaternary History of the Ocean Basins*. Progress in Oceanography. **4**, Pergamon Press, New York, 344 pp.

Seibold, E. 1970. Nebenmeere im humiden und ariden Klimabereich. *Geologische Rundschau*, **60**, 73–105.

Sellers, W. D. 1969. A global climate model based on the energy balance of the earth-atmosphere system. *Journal of Applied Meteorology*, **8**, 392–400.

Severinghaus, J. P. & Brook, E. J. 1999. Abrupt climate change at the end of the last glacial period inferred from trapped air in polar ice. *Science*, **286**, 930–934.

Shackleton, N. J. 1965. The high-precision isotopic analysis of oxygen and carbon in carbon dioxide. *Journal of Scientific Instruments*, **42**, 689–692.

Shackleton, N. J. 1967. Oxygen isotope analyses and Pleistocene temperatures reassessed. *Nature*, **215**, 15–17.

Shackleton, N. J. & Opdyke, N. D. 1973. Oxygen isotope and paleomagnetic stratigraphy of equatorial Pacific Core V28–238: oxygen isotope temperatures and ice volume on a 10^5-year and 10^6-year scale. *Quaternary Research*, **3**, 39–55.

Shackleton, N. J. & Kennett, J. P. 1975. Paleotemperature history of the Cenozoic and the initiation of Antarctic glaciation: oxygen and carbon isotope analyses in DSDP Sites 277, 279 and 281. *Deep Sea Drilling Project Initial Reports*, **29**, 743–755.

Shackleton, N. J., Berger, A. & Peltier, W. R. 1990. An alternative astronomical calibration of the lower Pleistocene timescale based on ODP Site 677. *Transactions of the Royal Society of Edinburgh-Earth Sciences*, **81**, 251–261.

Shepard, F. P. 1963. *Submarine Geology*, 2nd edn. Harper and Row, New York.

Steele, J. H., Thorpe, S. A. & Turekian, K. K. (eds.) 2001. *Encyclopedia of Ocean Sciences*. Academic Press, San Diego, **6**.

Stott, L., Timmermann, A. & Thunnell, R. 2007. Southern hemisphere and deep-sea warming led deglacial atmospheric CO_2 rise and tropical warming. *Science*, **318**, 435–438.

Suess, H. E. 1956. Absolute chronology of the last glaciation. *Science*, **123**, 355–357.

Sverdrup, H. U., Johnson, M. W. & Fleming, R. H. 1942. *The Oceans – Their Physics, Chemistry, and General Biology*. Prentice-Hall, Englewood Cliffs, NJ.

Valentine, J. W. 1973. *Evolutionary Paleoecology of the Marine Biosphere*. Prentice-Hall, Englewood Cliffs, NJ.

Van Andel, Tj. H., Heath, G. R. & Moore, T. C., Jr. 1975. Cenozoic history and paleoceanography of the central equatorial Pacific Ocean. *Geological Society of America, Memoirs*, **143**, 1–134.

Vincent, E. & Berger, W. H. 1981. Planktonic foraminifera and their use in paleoceanography. *In*: Emiliani, C. (ed.) *The Oceanic Lithosphere. The Sea, vol. 7*. Wiley, New York, 1025–1119.

Vincent, E., Killingley, J. S. & Berger, W. H. 1981. Stable isotope composition of benthic foraminifera from the equatorial Pacific. *Nature*, **289**, 639–643.

Weertman, J. 1957. On the sliding of glaciers. *Journal of Glaciology*, **3**, 205–214.

Woodruff, F., Savin, S. M. & Douglas, R. G. 1980. Biological fractionation of oxygen and carbon isotopes by recent benthic foraminifera. *Marine Micropaleontology*, **5**, 3–11.

Yasuda, M., Berger, W. H., Wu, G., Burke, S. & Schmidt, H. 1993. Foraminifer preservation record for the last million years: Site 805, Ontong Java Plateau. *Proceedings of the Ocean Drilling Program, Scientific Results*, **130**, 491–508.

Zachos, J., Pagani, M., Sloan, L., Thomas, E. & Billups, K. 2001. Trends, rhythms, and aberrations in global climate 65 Ma to present. *Science*, **292**, 686–693.

Zeuner, F. E. 1952. *Dating the Past – An Introduction to Geochronology*, 3rd edn. Methuen & Co., London.

Zwally, H. J., Abdalati, W., Herring, T., Larson, K., Saba, J. & Steffen, K. 2002. Surface melt induced acceleration of Greenland ice-sheet flow. *Science*, **297**, 218–220.