12

Solving the Ice Age Mystery: The Deep-Ocean Solution

12.1 Astronomical Drivers

By the time of the Pleistocene, the Earth had cooled to its lowest point since the Carboniferous glaciation 300 Ma ago. During the Pleistocene, much of the variation in the Earth's climate was due not to the plate tectonic processes that were important in older times, but to celestial mechanics. This led to cycles of 100 Ka in the eccentricity of the Earth's orbit, of 41 Ka in the Earth's axial tilt (which dominates radiation at high latitudes) and of 22 Ka in the precession of the equinoxes (which dominates radiation at low latitudes)[1]. As we saw in Chapter 6, Milutin Milankovitch set out the basis for our understanding of this process between 1920 and 1941.

In 1945, Frederick Zeuner (1905–1963) of London's Institute of Archaeology tested Milankovitch's theory by examining how it applied to what was known of the Pleistocene period on land[2]. Finding a close match between the sequence of Ice Age strata and the variations in insolation, he concluded, '*no objection can be raised against the astronomical theory of the glacial and interglacial phases of the Pleistocene*'[2]. Among those intrigued by what controlled the changes between glacial and interglacial periods was Richard Foster Flint (1901–1976) of Yale University, who, along with other honours, would be awarded the Prestwich Medal by the Geological Society of London in 1972 for his contributions to our understanding of the Ice Age. Flint was one of the most influential figures in Quaternary science in the 20th century, much admired for his seminal 1957 text *Glacial and Pleistocene Geology*[3]. This book concluded, '*the geometric scheme of distribution of insolation heating must be considered inadequate in itself to explain the Pleistocene climatic changes*'[3].

Things have changed since then. As Mike Walker of the University of Wales and John Lowe of Royal Holloway College London pointed out in 2007, Flint's approach was rooted in glacial geology. Since his day, those investigating Quaternary science have moved '*away from Flint's somewhat narrow glacial-geological paradigm towards the multi- and inter-disciplinary approach to the study of recent Earth history that is practiced today*'[3]. With this new approach, we can now analyse the '*rich and often readily accessible Quaternary record … at a level of detail not normally possible for older geological periods*'[3]. Milankovitch's theory has come to stay[4].

Milankovitch lacked computers. André Léon Georges Chevalier Berger (1942–) (Box 12.1) used them to refine his calculations[5–8]. Figure 12.1 provides an introduction to Berger's findings, which we explore in more detail in Chapter 13.

Full comprehension of Ice Age climate change hinges on novel studies of deep-ocean sediments collected by piston cores and deep-ocean drilling, as we see in this chapter, and of ice cores, which we examine in Chapter 13. These studies fall into the science of the Quaternary, which comprises the Pleistocene, starting at 2.6 Ma ago, and the Holocene – the last 11 700 years[3]. To help move the field forward, scientists formed the International Quaternary Union (INQUA) in 1928[3]. Several

Earth's Climate Evolution, First Edition. Colin P. Summerhayes.

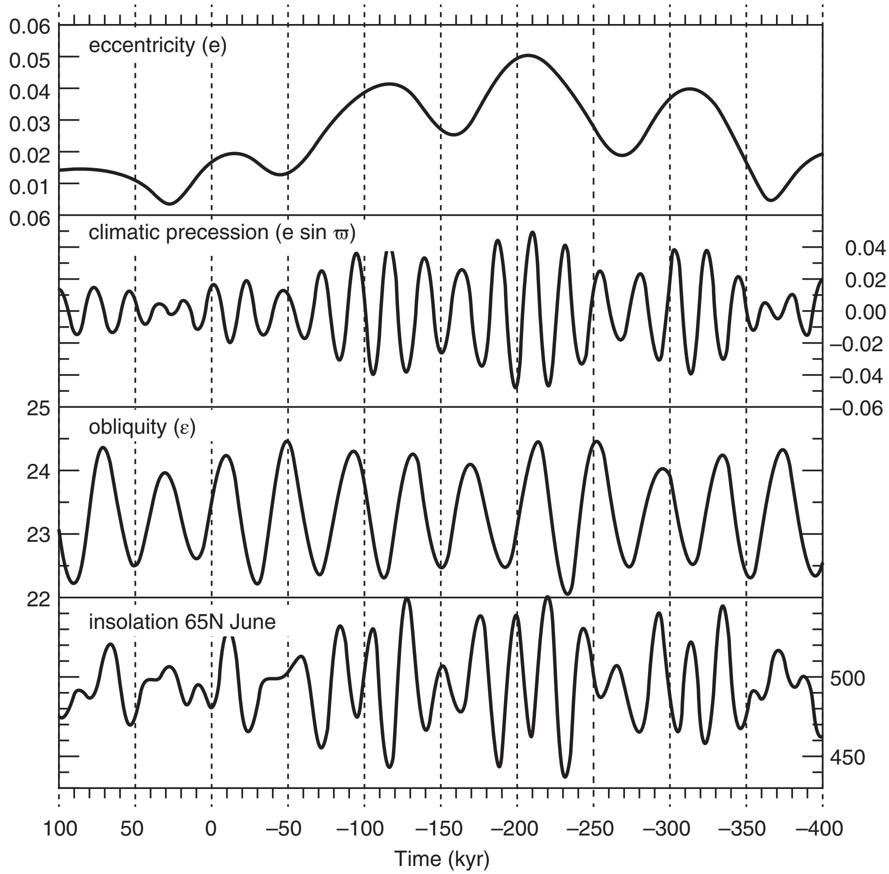

Figure 12.1 *The Berger Astronomical Model of Orbital Variability Present and Future. These curves have been produced in numerous formats in several publications by André Berger and Marie-France Loutre. Obliquity is expressed in degrees of tilt of the Earth's axis. Insolation at the summer solstice at 65° N is expressed in W/m^2.*

Box 12.1 André Léon Georges Chevalier Berger.

André Berger has a master's degree in meteorology from MIT (1971) and a doctorate from the Catholic University of Louvain, Belgium (1973). He is renowned for contributing to the renaissance of Milankovitch's theory of climate change, for making major contributions to simulating future climate change and for working on the first Earth model of intermediate complexity. He was professor of meteorology and climatology at Louvain, and then director of the Institute of Astronomy and Geophysics Georges Lemaître from 1978 to 2001, where he now has emeritus status. He has served as president or chairman of several national and international scientific organisations and committees and was honorary president of the European Geosciences Union. He was on the steering committee for the International Geosphere-Biosphere Programme (IGBP) and initiated the Palaeoclimate Modeling Intercomparison Project (PMIP). He has received many honours for his discoveries, including the Milutin Milankovitch Medal of the European Geophysical Society, and in 1996 he was made a knight of the realm by King Albert II.

scientific journals emerged to meet their needs: *Quaternary Research* (1971), *Boreas* (1972), *Quaternary Science Reviews* (1982), *Journal of Quaternary Science* (1985), *Quaternary International* (1989) and *The Holocene* (1991). Researchers also make good use of journals like *Paleoceanography* and *Palaeogeography, Palaeoecology, Palaeoclimatology*, as well as the four-volume *Encyclopaedia of Quaternary Science* (2006)[9].

12.2 An Ice Age Climate Signal Emerges from the Deep Ocean

As we saw in Chapter 7, the first to exploit the new technology of piston coring in order to examine the history of climate recorded in deep-sea sediments was Gustaf Arrhenius. In 1952, Arrhenius attributed alternations between carbonate-rich and carbonate-poor sediments in east Pacific cores to changes in the 'aggressiveness' of polar bottom waters. During the Ice Age, he thought, large volumes of bottom water were derived from the polar regions. Being cold, they carried large amounts of dissolved CO_2, which enabled them to dissolve, or to prevent the deposition of, deep-sea carbonates. Bottom waters of intervening warm periods carried less dissolved CO_2, so were less 'aggressive'. At the time, Arrhenius, like Flint, dismissed Milankovitch's ideas[10], although he later adopted them.

At about the same time, in the early 1950s, Lamont began its routine collection of long piston cores from the world's oceans. David Ericson, who was in charge of the new Lamont core store, found that although the distribution of the planktonic foraminiferan *Globorotalia menardii* indicated warm conditions, it was also influenced by ocean currents. Another species, *Globigerina pachyderma*, was a cold-water indicator, as were *Globigerina inflata* and *Globigerina bulloides*. Changes from warm to cold were also indicated by changes in the coiling direction of *Globorotalia truncatulinoides*. From the distribution of these species down-core, Ericson built a Quaternary stratigraphy incorporating the Holocene, the last glaciation and the previous interglacial[11].

Ericson and Arrhenius's interests were shared by Cesar Emiliani, whom we met in Chapters 6 and 7. In 1955, Emiliani analysed the oxygen isotopes in planktonic foraminifera collected from eight piston cores from the Swedish Deep-Sea Expedition and four from the Lamont core store. Finding fluctuations in the $\partial^{18}O$ ratio with time, he interpreted them as representing the variations in climate between glacial and interglacial periods[12]. By analysing the same species of surface-dwelling foraminiferan, he eliminated the effect of metabolic differences between different species. The resulting variations were due to differences in either the temperature or the isotopic composition of seawater, the latter representing the amount of water tied up as ice on land. Emiliani '*guessed that 60% of the signal was due to the temperature effect, 40% to the ice effect*'[13]. Following Zeuner's reasoning[2], he thought that the variations in $\partial^{18}O$ with time represented changing insolation[12]. He invented the nomenclature that is still in use today, in which each warm or cold period is identified as a marine isotope stage (MIS). MISs with even numbers are cold stages, those with odd numbers are warm stages. Some can be subdivided into substages (e.g. 5a, 5b, 5c).

In March 1961, Emiliani teamed up with Flint to link the Pleistocene record in continental and deep-sea sediments[14]. They considered that MIS 1 was the Holocene, MIS 2 was the main Würm glacial stage on land, MIS 3 was the early to main Würm glacial interval on land, MIS 4 was the early Würm on land and MIS 5 was the last interglacial. Thinking about the possible drivers for the Ice Age, they favoured a model of glaciation based on Milankovitch's concept that ice accumulated during cool summers, driven by insolation in the Northern Hemisphere. Evidently, Flint had changed his mind about Milankovitch since 1957. They discounted Plass's idea that ice ages were in some way controlled by atmospheric CO_2, because they thought, wrongly, that Revelle and Suess[15] had concluded, based on studies of ^{14}C, that atmospheric CO_2 would be rapidly taken up by the ocean. It would be a while before the role of CO_2 in the Ice Age climate would be fully understood.

Before Emiliani's isotopic analyses of deep-sea cores, it was thought that there were only four major glacial periods during the Ice Age[2]. Emiliani found more than twice that! While this shocked classical Quaternary geologists, physicist Nicholas J. Shackleton (Box 7.3) agreed with Emiliani[16]. The fly in the ointment was the inadequacy of methods for accurately dating Emiliani's cores. Emiliani guessed that his main $\partial^{18}O$ cycles were 41 Ka long, but they were later found to last about 100 Ka.

Subtle differences in the ways in which Ericson and Emiliani interpreted their data meant that, for a while, they disagreed about the precise sequence of cold and warm events. When they did agree, we had two independent means of mapping changes between glacial and interglacial periods in deep-sea cores – one palaeontological,

Figure 12.2 *John Imbrie.*

from microscopic fossils of marine plankton (microfossils), and the other geochemical, from oxygen isotopes in those same fossils.

By the mid 1960s, two new figures occupied centre stage: Nick Shackleton and the micropalaeontologist and stratigrapher John Imbrie (1925–), of Brown University (Figure 12.2, Box 12.2). Our understanding of Ice Age climate owes a great deal to these two men.

Box 12.2 John Imbrie.

Imbrie obtained a BA from Princeton in 1948, after serving with the US 10th Mountain Division in Italy during the Second World War. Having obtained a PhD from Yale in 1951, he taught at Columbia University until 1967. Joining Brown University, he held the Henry L. Doherty Chair in Oceanography, and he now holds emeritus status there. He pioneered the use of computers to demonstrate the relation of assemblages of plankton to the temperature of surface waters, thus providing palaeoceanography with one of its key tools. His book *Ice Ages: Solving the Mystery*[18], written with his daughter Katherine, won the 1976 Phi Beta Kappa Prize. Imbrie was co-author with Hays and Shackleton of the 1976 paper in *Science* that linked Milankovitch variations to the sediment record. He was elected to the US National Academy of Sciences in 1978 and received the Maurice Ewing Medal of the American Geophysical Union in 1986, the Twenhofel Medal of the Society of Sedimentary Geology, the Lyell Medal of the Geological Society of London in 1991 and the Vetlesen Prize in 1996.

Using his refined mass-spectrometric technique (see Chapter 7), Shackleton showed by 1967 that a distinctive ∂^{18}O pattern is evident in the benthic foraminifera that live on the seabed, which are bathed in cold oxygen-rich bottom water sinking from the surface in the polar regions[17]. Most of this water is colder than 4 °C, so short-term variations in the ∂^{18}O ratio in these organisms tell us more about changing seawater temperature than changing ice volume. Shackleton found that planktonic foraminifera growing in surface waters displayed much the same signal. He calculated that about 66% of the ∂^{18}O shift between glacial and interglacial periods was due to changes in ice volume, not to the influence of temperature, which was the opposite of what Emiliani had concluded. This revelation caused a paradigm shift in our understanding of ∂^{18}O ratios in the service of palaeothermometry. Urey had assumed that the oxygen isotopic composition of seawater would be invariant[19]. Clearly it was not. Today, the ∂^{18}O ratios of benthic foraminifera are taken as representing the temperatures of polar surface waters.

By 1969, Imbrie realised that because the total assemblage of planktonic foraminifera in surface waters should reflect the environment in which they lived, and multivariate statistical analyses could quantify that relationship, statistical analyses of faunal assemblages down-core could be used to ascertain past climate change[20]. Reanalysing the Caribbean cores analysed by Ericson and Emiliani, he and Nilva Kipp showed that the temperatures of surface waters of glacial periods there fell by just 2 °C, not 6 °C. Because the fluctuations in their data agreed with those determined by Emiliani from isotopes, it was clear that *Globorotalia menardii*, which Ericson used to identify warm periods, fluctuated in ways unrelated to temperature, explaining the discrepancy between Ericson and Emiliani's data[21,22]. This was a breakthrough.

By the late 1960s, as the Deep Sea Drilling Project (DSDP) got underway, analyses of ∂^{18}O changes down piston cores from different parts of the ocean showed that sediments could be routinely subdivided into Emiliani's MISs representing glacial and interglacial periods. These stages coincided with intervals defined by Imbrie's assemblages of microfossils and could be correlated from one core to another over vast oceanic distances, suggesting planetary control.

Was that planetary control the same as Milankovitch's astronomically controlled insolation? To find out, geologists needed closely spaced dates down-core. In the late 1960s to early 1970s, microfossils could tell us about environmental change from cold to warm and back, but not about the ages of cold and warm stages. As we saw in Chapter 6, radiocarbon dating was useful, but only in sediments less than 50 Ka old. Layers of volcanic ash older than 100 Ka could be dated by the potassium–argon (K-Ar) method. Changes in the Earth's magnetic field could be used to date specific sedimentary horizons, but only in sediments older than 780 Ka. Together, these independent techniques provided a crude means of dating sediment layers in cores. Over time, more techniques would become available. A major breakthrough in radiocarbon (^{14}C) dating came about in 1977, when a new technique – accelerator mass spectrometry (AMS) – enabled us to count ^{14}C atoms, as opposed to measuring ^{14}C decay. AMS can date samples as small as a pinhead-sized microfossil. The technique is fast and cheap, but it is still limited to sediments less than about 50 Ka old.

Determining the ages of the glacial and interglacial sedimentary stages identified by ∂^{18}O and microfossil analyses became a major objective of the international Climate Long Range Investigation, Mapping and Prediction (CLIMAP) project. Founded by Imbrie, Shackleton and others, CLIMAP began in spring 1971 as part of the International Decade of Ocean Exploration[23]. The project aimed to establish average boundary conditions for the Last Glacial Maximum at 18 Ka ago. Those conditions included the geography of the continents, the albedo of land and ice surfaces, the extent and elevation of permanent ice and the sea surface temperature. Modellers would use those conditions in atmospheric General Circulation Models (GCMs) to map the climate of the Last Glacial Maximum. CLIMAP would then test model outputs against palaeoclimate data. The first simulation was for August 18 Ka ago[23]. Sea surface temperature values were derived from ∂^{18}O data and from Imbrie's statistical analyses of planktonic faunal assemblages. The extent of sea ice in the polar regions was estimated from the presence or absence of diatomaceous sediments, with absence indicating ice.

The CLIMAP data showed that extensive cooling at the poles and an expanded area of land and sea ice steepened the thermal gradient between the Equator and the poles, strengthening the winds. In the Southern Ocean, the Antarctic polar front moved north, along with Antarctic sea ice. The Subtropical Front moved far enough north to limit the passage of warm Indian Ocean water around South Africa and into the South Atlantic. This cooled the South Atlantic and created a closed anticlockwise gyre in the Indian Ocean. Upwelling increased where it is found today, along continental margins and along the Equator, as the winds that drove it increased in strength. On land, grasslands, steppes, deserts and ice spread at the expense of forests, increasing the Earth's albedo.

Palaeoclimatologists noticed that the wiggles in the ∂^{18}O curves down sediment cores seemed to mimic the wiggles in the patterns of Earth's insolation through time. If the match was real then it offered an opportunity to date the age of the sediments from the pattern of wiggles in the oxygen isotope data. Starting with just a few radiometric dates as tie points, the CLIMAP scientists assumed that the rates of sedimentation in different MISs were constant down-core. This enabled them to estimate the age of each wiggle on the curve of variation in ∂^{18}O back through time. Applying spectral analysis to the ∂^{18}O curve dated in this way, James Hays (1938–) of Lamont, together with Imbrie and Shackleton, demonstrated in a landmark paper in 1976 that in cores where sedimentation was undisturbed, the variations in ∂^{18}O over the past 450 Ka accurately mimicked the orbital signals calculated by André Berger[24]. Furthermore, climate changes in the Northern Hemisphere were essentially synchronous with those observed in the Southern Hemisphere. The correlation between the astronomical variables and ∂^{18}O told them that '*changes in the earth's orbital geometry are the fundamental cause of the succession of Quaternary Ice Ages*'[24]. The three Milankovitch mechanisms – eccentricity, tilt and precession – worked in unison to provide the 'pacemaker of the ice ages'.

Not only that, but the fact that the same pattern of wiggles occurred everywhere meant that even cores without precise radiometric dates could be dated by reference to a standard ∂^{18}O curve derived by merging data from several well-dated cores. Wiggle matching offered an incredible opportunity to date intervals of time as small as about 1000 years long. Patterns of tree rings back through time offer much the same possibility, provided they come from much the same area and so experienced more or less the same climate changes through time. While this is very helpful for the past 11 Ka, tree rings do not provide us with a lengthy and globally distributed data base of the kind provided by deep-sea cores. Wiggle matching of oxygen isotope curves from core to core does contain the assumption that the section has not been disturbed by either erosion or the lateral introduction of material by turbidity currents, but these possibilities are identified or eliminated by comparing individual cores with the global standard. Indeed, wiggle matching can identify how much section has been removed! Nowadays, wiggle matching has enabled palaeoclimatologists to push the ∂^{18}O-calibrated time scale back into the Oligocene, about 30 Ma ago[13]. Amazing!

This incredible breakthrough refined the resolution of the geological time scale beyond anything previously imaginable, except where annual layering was preserved in tree rings, corals, stalagmites or lake sediments, most of which did not allow dating back beyond about 2000 years ago. Palaeoclimatologists who started their careers in the 1980s and later take these advances for granted, but they were astonishing developments to the geologists of my age group. As Mike Leeder pointed out in 2011, this new understanding was '*arguably as big an earth sciences discovery as that of plate tectonics*'[25]. Mike Walker and John Lowe agreed, citing the 1976 Hays paper as '*perhaps the most important Quaternary paper of the past 50 years*'[3]. It definitively overturned the long-held perception that there had been four major Quaternary glacial periods[26], by showing that there had been many more cycles of climate, ice volume and sea level in the late Cenozoic, and that these cycles formed in response to variations in Earth's orbital parameters. As Nick McCave and Harry Elderfield point out in Shackleton's obituary, '*This clear recognition of orbital control is also now revolutionizing the whole of stratigraphy (the study of geological strata) because it provides in principle a means of correlating beds at separated parts of the Earth to a precision of 20 000 years at a time of hundreds of millions of years ago, and of determining precise "orbitally tuned" age-calibrated stratigraphies back to about 250 Ma ago*'[27].

This was not all. Given that the climate was governed by celestial mechanics, and that Berger's data projected Earth's orbital properties and insolation far into the future, Hays, Imbrie and Shackleton deduced that '*the long term trend over the next 20,000 years is towards extensive Northern Hemisphere glaciation and a cooler climate*'[24]. That's not quite the picture we have today, but it's close. Berger's data show that because the Earth's orbit is at present close to circular, the present warm Holocene interglacial should last some 30–50 Ka, of which we have already experienced 10 Ka[28,29]. So, we should have ~20 Ka more relatively warm climate before the next glacial period. For today, Berger's data show that Earth should be experiencing a slight cooling trend, which started around 10 Ka ago (Figure 12.1). We look at that more closely in later chapters. These various dramatic developments were ably summarised by John and Katherine Imbrie in their 1979 book *Ice Ages: Solving the Mystery*[18]. It remains a classic.

In the 1980s, the CLIMAP project was succeeded by the SPECMAP (Spectral Mapping) project, designed by Imbrie, Shackleton and others to produce continuous time series of Ice Age climate change from deep-sea sediments and to facilitate studying their spectral properties. A key SPECMAP achievement was publication of a time scale for the last 780 Ka, based on a ∂^{18}O reference curve compiled by stacking together planktonic foraminiferal ∂^{18}O records from five low- and middle-latitude sites. Stacking avoids local 'noise' interfering with the underlying signals[30]. Known as the SPECMAP stack, this curve, which was tuned (phase-locked) to the oscillations of precession and obliquity, provided a continuous geological time scale for the late Pleistocene, the divisions of which were accurate to within ±5000 years, an astoundingly high resolution for geological records. The SPECMAP ∂^{18}O stack was improved and extended over time[31,32]. In 2005, Lorraine Lisiecki, then at Brown University, and Maureen Raymo, then at Boston University, replaced it with a stack made from combinations of benthic foraminiferal data (Figure 12.3)[33,34], which show less variability than planktonic data. Over 100 MISs have been identified, going back some 6.6 Ma. The stack is a 'type section' against which new core measurements are compared.

The Last Glacial Maximum, or MIS 2, extended from 30 to 15 Ka ago. On average, compared with today, it was thought to be 5–6 °C colder globally. It was much colder in the Arctic, with temperatures in central Greenland depressed by as much as 20 °C[35]. At that time, much Arctic land lay beneath continental ice sheets, and the Arctic Ocean was mantled by continuous sea ice and entrapped icebergs. The lack of northward transport of warm, salty water during the winters made them exceptionally cold. Polar desert replaced tundra. Ice volume peaked about 21 Ka ago, after which rising insolation caused ice sheets and glaciers to melt back, with most coastlines becoming ice-free before 13 Ka ago[35]. We will discuss deglaciation later.

The climate signal in cores was not smoothly varying, unlike the variation in orbital and axial properties (Figure 12.1). As we can see from Figure 12.3, it was saw-toothed – something first pointed out by Broecker

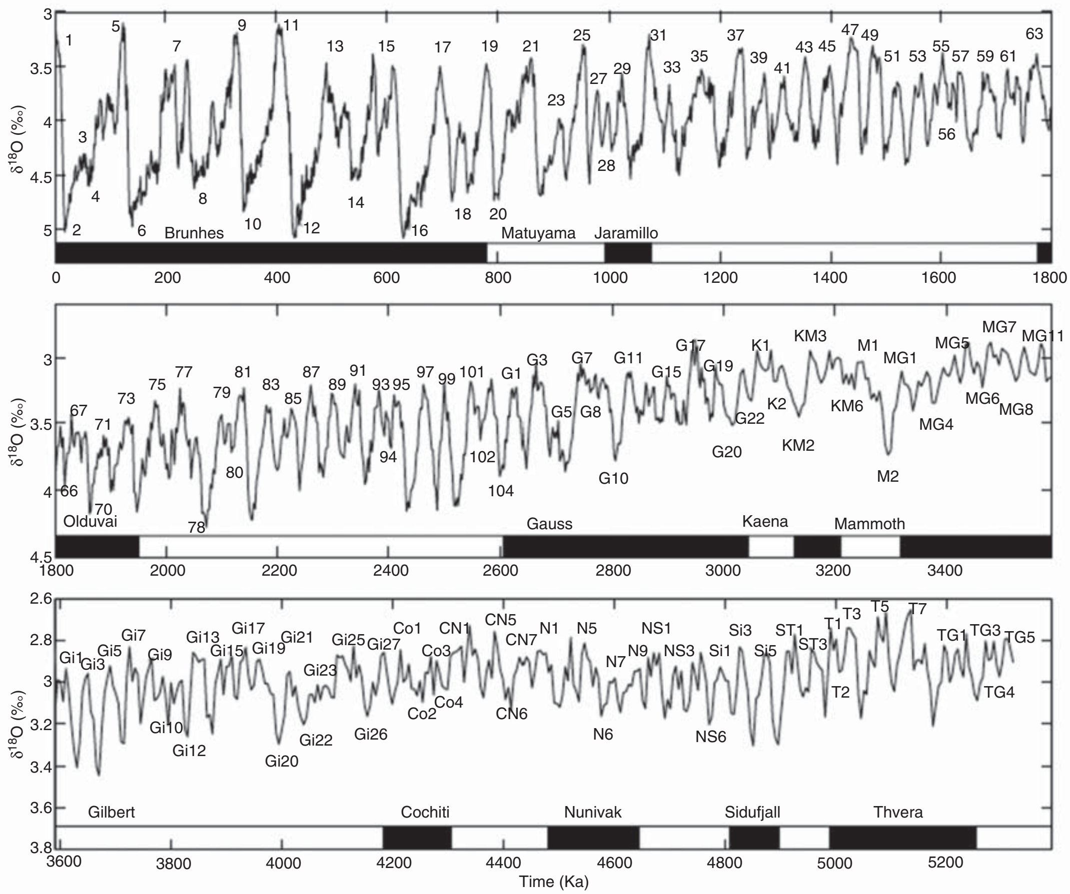

Figure 12.3 *Benthic oxygen isotope stack, constructed by the graphic correlation of 57 globally distributed benthic ∂^{18}O records covering 5.3 Ma. Note that the scale of the vertical axis changes from panel to panel. From this stack, a number of new MISs were identified in the early Pliocene. MISs are identified by number back to 2.6 Ma ago; before that, the lettering refers to the name of the magnetic chron in which the isotope peaks appear (e.g. Si = Sidufjall, Co = Cochiti etc).*

and Van Donk[36]. The saw-tooth shape represents the slow growth of ice followed by rapid deglaciation, calling for strong positive-feedback mechanisms to accelerate melting. We examine this in more detail in Chapter 13.

CLIMAP's maps of sea surface temperature for the winter and summer of the Last Glacial Maximum (21.5–18.0 Ka ago) were later updated by the Glacial Atlantic Mapping Project (GLAMAP)[37], which reconstructed the glacial Atlantic Ocean from 275 deep-sea sediment cores[38]. During the northern winter, sea ice extended south as far as about 50° N, close to the latitude of Cork, on the south coast of Ireland. During the northern summer, warm surface waters moved northwards into the Norwegian-Greenland Sea much as they do today, but they were between ice sheets on both sides, meeting sea ice at about latitude 70° N. As today, there was a return flow from the Arctic down the east side of Greenland. A proto-Gulf Stream marched eastward from Labrador to Lisbon, where the sea surface temperature averaged about 16 °C.

In parallel with GLAMAP, a comparable project began to examine Environmental Processes of the Ice Age:

Land, Oceans, Glaciers (EPILOG). It focused on the 21 Ka interval, this being the age of minimal summer insolation at 65 °N, and the time when ice sheets reached their maximal volume, as represented by sea level fall[39]. GLAMAP and EPILOG were followed by the Multiproxy Approach for the Reconstruction of the Glacial Ocean project (MARGO)[40]. These various studies and others concluded that the last deglaciation began between 22 and 18 Ka ago. Two schools of thought emerged, one suggesting that deglaciation began in the Southern Hemisphere, with surface and deep-ocean warming followed by tropical sea surface temperatures and by atmospheric CO_2, and the other suggesting that it began in the Northern Hemisphere, where summer insolation at high northern latitudes was the trigger for ice-sheet decay and sea level rise[41]. We explore these issues in Chapter 13.

The MARGO project team constructed maps of sea surface temperature to provide constraints on ocean cooling at the Last Glacial Maximum[42]. Like CLIMAP, they found that the strongest mean annual cooling (−10 °C) occurred in the mid-latitude North Atlantic, extending into the western Mediterranean, but unlike CLIMAP they found that this cooling was most pronounced in the east. Indeed, most ocean basins had cooler eastern than western sides. This eastern cooling was probably due to increased upwelling of cold water forced by stronger coastal winds. It was not replicated by existing GCMs for the Last Glacial Maximum, for reasons not then fully understood (in 2009). In contrast with CLIMAP, the MARGO team found that conditions were ice-free in the summer in the Nordic seas, and that the tropics were on average 1.7 °C cooler than CLIMAP had thought. In the Southern Ocean, the polar front shifted north from near 60 to 45° S, associated with a cooling of 2–6 °C in the austral winter.

One of the key features evident from the stack of $\partial^{18}O$ data (Figure 12.3), noted in 1976 by Shackleton and Opdyke[43], is that climate variability has grown with time[19]. In the earlier part of the Pleistocene, the signal comprised relatively small glacial–interglacial changes, in which signals of precession (22 Ka cycles) and obliquity (41 Ka cycles) predominated. The signals became much larger at about 900 Ka ago, after which a signal with a spacing of 100 Ka intervals predominated. The 100 Ka signal itself became larger with time, especially from around 430 Ka ago (MIS 11) onwards. The change at 900 Ka ago formed the Mid Pleistocene Transition (MPT). What did it represent?

Harry Elderfield and colleagues used Mg/Ca ratios to establish what part of the $\partial^{18}O$ signal at the transition was due to ice volume rather than water temperature[44]. Changes in ice volume from glacial to interglacial were much smaller before the transition than since, presumably because the older ice sheets were smaller in area and/or thickness. The transition was a sudden jump, not the result of a long-term trend towards increased ice volume and colder temperatures. Elderfield's team concluded that it represented '*an abrupt reorganization of the climate system*'[44]. The trigger seemed to be a brief period of anomalously low summer insolation in the Southern Hemisphere during the warm MIS 23. This suppressed the melting of ice formed previously in cold MIS 24, allowing unusually extended ice growth in the following cold MIS 22, at 900 Ka ago, to yield a very large ice sheet, associated with a lowering of sea level of about 120 m[44].

Investigating the behaviours of ice volume and temperature, Elderfield's team confirmed that ice volume followed a saw-toothed pattern, growing steadily from low amounts during interglacials to high amounts during glacials, then suddenly retreating. In contrast, bottom water temperatures followed a square wave pattern, falling to a certain level as ice volume grew, then staying more or less constant, before rising again as ice volume decreased. The temperatures of bottom waters during glacial periods remained constant at −1.5–2.0 °C, because, once the temperature of surface water in the source region fell to about the freezing point of salt water, it would fall no further. Bottom water temperatures warmed to 3 °C during interglacial periods.

Prior to the transition, sea level fell to 70 m below present levels. The drop by a further 50 m at the transition exposed continental shelf sediments to erosion, transferring marine organic matter rich in ^{12}C to the deep sea and lowering the $\partial^{13}C$ ratio of bottom water and benthic organisms. Elderfield's team calculated that about half of the fall in $\partial^{13}C$ at 900 Ka was due to this change in carbon reservoirs, with the other half coming from a reduction in the influence of North Atlantic Deep Water[44].

As the volume of ice increased across the transition, the supply of aeolian dust, represented by the rates of accumulation of sedimentary iron and terrestrial leaf waxes, doubled in the Southern Ocean[45]. The increase in the dust supply tells us that the surrounding lands dried out as the globe cooled. The increase in iron helped to cool the globe further, via positive feedback, because an increase in iron as a key nutrient stimulates productivity, drawing CO_2 from the atmosphere, as we see in more detail later[45]. This interpretation is supported by an increase in the sedimentation of opal, representing diatom productivity, at the same time. The rise in productivity drove a 30 ppm reduction in CO_2 across the transition. By driving a descent into deep, cold, glacial periods, the insolation/dust/CO_2 feedback may have initiated the strong 100 Ka periodicity that characterised subsequent climate change. On the global scale, an increase in the supply of dust also coincided with the start of the major Northern Hemisphere glaciation at about 2.6 Ma ago, which drew down CO_2 in much the same way.

The notion of using a single stack of $\partial^{18}O$ values to represent Earth's recent glacial history would have seemed odd to Joseph Adhémar and James Croll, whom we met in Chapter 3, because they thought that cooling related to precession would alternate between the two hemispheres. As we now know, glaciers and ice sheets in Patagonia and Antarctica actually advance and retreat at more or less the same times as those in the Northern Hemisphere. Why? The answer lies in those feedbacks to which Croll first introduced us. Antarctica is an ice-covered continent surrounded by ocean – there is nowhere for its land ice to expand into. When Antarctic ice is at a maximum, global ice can only increase by growing on the Northern Hemisphere continents. This growth lowers sea level, exposing Antarctica's continental shelf and so providing space for yet more ice growth. Sea level links ice growth on Antarctica to that on the northern continents. Thus, glaciation tends to become more or less synchronous in both hemispheres, even while the insolation is opposite[13]. Besides that, insolation has certain seasonal characteristics that help to align glaciation in the two hemispheres even though their annual insolation signal is opposed, as we see in Chapter 13.

The latest view of the temperature of the Last Glacial Maximum is that it was probably 4.0 ± 0.8 °C cooler than the modern preindustrial climate[46].

The growing literature on orbital variations and their record in cores from ice, sediments, corals and stalactites fuelled intensive discussion about the precise mechanisms underlying the climate changes of the Ice Age, which we review in Chapter 13.

12.3 The Ice Age CO_2 Signal Hidden on the Deep-Sea Floor

Could carbon isotopes from marine sediments tell us about the abundance of CO_2 in the Ice Age atmosphere? Wally Broecker suggested that the atmospheric CO_2 signal could be represented by the difference in $\partial^{13}C$ ratios between surface planktonic foraminifera and bottom-dwelling benthic foraminifera, an idea followed up by Nick Shackleton and colleagues in 1983[47]. The variations they detected in CO_2 by this means matched those found in ice cores by Oeschger, as mentioned in Chapter 9. Clearly, CO_2 rose and fell with rises and falls in temperature during the Ice Age. How reliable was the association between estimated CO_2 and the $\partial^{18}O$ values used to estimate temperature? In 1985, working with Nick Pisias of Oregon State University to analyse the $\partial^{13}C$ and $\partial^{18}O$ profiles through the past 340 Ka in a core from 3091 m depth in the Pacific, Shackleton found that CO_2 closely followed temperature[48]. It slightly lagged orbital insolation in June at 65° N, led the response in ice volume (documented by variations in $\partial^{18}O$) by about 2500 years and was closely linked to variations in axial tilt, which dominates the insolation signal at middle to high latitudes. Shackleton and Pisias concluded that the CO_2 signal was forced by high-latitude orbital insolation through '*a mechanism at present not fully understood*'[48] – probably the effect of that insolation on ocean circulation. As changes in CO_2 led changes in ice volume in the North Atlantic, the CO_2 must have contributed to the forcing of changes in ice volume there. It was a forcing factor. Insolation warmed the ocean, which released CO_2, which enhanced temperature, stimulating an eventual decrease in ice volume.

These conclusions were consistent with the proposal by Wally Broecker and Tsung-Hung Peng in their 1982 classic *Tracers in the Sea* that, since the ocean contains about 60 times more carbon than the atmosphere, the glacial–interglacial change in atmospheric CO_2 content must have been driven by changes in ocean chemistry[49]. Broecker seems to have been the first to suggest that a glacial increase in the strength of the biological pump drove down CO_2 levels. His initial ideas about the biological pump revolutionised the field of chemical oceanography[50].

Broecker's ideas got John Martin (1935–1993) thinking. Director of the Moss Landing Marine Laboratories in California, and crippled by polio when he was 19, Martin set himself the task of figuring out the role of phytoplankton – the grass of the sea – in the global climate system[51]. Phytoplankton use CO_2 for photosynthesis. When their remains sink to the seafloor and decompose, the CO_2 is returned to deep-ocean waters or trapped in sediment and so can no longer contribute to warming the planet. In order to determine how much plankton sank to the seafloor in a given time, Martin organised the Vertical Transport and Exchange of Oceanic Particulate Program (VERTEX) in 1981, placing sediment traps across the North Pacific to sample the flux and composition of settling particulates. Among other things, he discovered that the parts of the ocean that are high in nutrients but low in chlorophyll were depleted in iron (Fe)[52]. Joseph Hart, an English scientist, had speculated in the 1930s that this might be the case, but was unable to prove it. Martin proposed in 1990 that Fe was a limiting nutrient and that production of phytoplankton could be negatively affected by its supply, for example in airborne dust[53].

During the Last Glacial Maximum, the supply of dust was 50 times greater than today, enhancing productivity enough to draw CO_2 out of the air. Lack of Fe-rich dust during interglacials slowed productivity, leaving CO_2 in the air. Martin suggested testing his hypothesis through Fe-enrichment experiments even at the scale of the whole Southern Ocean. Several such experiments were carried out, the first in 1993, although none at the scale of an entire ocean. The early experiments found that the excess organic matter created by the addition of Fe was recycled in the water column; it did not settle to the seabed as Martin had imagined[54]. That changed in 2012, when Victor Smetacek of Germany's Alfred Wegener Institute for Polar and Marine Research performed a 5-week-long Fe-fertilisation experiment in the Antarctic Circumpolar Current with colleagues and discovered that at least half of the diatom bloom caused by the fertilisation sank far below 1000 m, with much being likely to have reached the deep-sea floor[55]. This confirmed the view that '*iron-fertilized diatom blooms may sequester carbon for timescales of centuries in ocean bottom water and for longer in the sediments*'[55]. Here was support for the geoengineering notion that iron fertilisation of the ocean could transport CO_2 out of the surface waters and into the deeps, thus drawing down atmospheric CO_2.

The abundance of CO_2 in the air between glacial and interglacial times was also governed by the presence or absence of sea ice[56]. Growing sea ice placed a lid on polar surface waters, preventing them from absorbing CO_2 from the air. As a result, CO_2 would gradually accumulate in the air, causing it to warm. Melting of sea ice exposed the cold ocean, enabling it to absorb CO_2 from the air, contributing to eventual cooling.

12.4 Flip-Flops in the Conveyor

Palaeoceanographic studies radically changed our understanding of the variability of Ice Age climate and the role of the ocean in climate change. One key result was the realisation that the circulation of the ocean had different stable states for glacials and for interglacials. During interglacials like the Holocene, in which we live now, ocean circulation was much as it is today (Figure 12.4).

Warm, salty surface water is drawn towards the Arctic through the northern branch of the Gulf Stream, losing heat to the atmosphere en route. This heat warms northwest Europe. By the time the salty water reaches the Norwegian-Greenland Sea, it has cooled to the point of becoming dense enough to sink and form North Atlantic Deep Water, which moves south towards Antarctica and fills the mid-water depths of the Atlantic, Indian and Pacific Oceans. The strong westerly winds blowing around Antarctica towards the east force these northern-sourced deep waters to the surface through the process of upwelling. The newly upwelled surface waters then return to the North Atlantic to close the cycle through two pathways. First, under the influence of Antarctica's coastal easterly winds, some water moves south on to the Antarctic continental shelf, where the excretion of salt from sea ice forming at the ocean's surface makes it dense enough to sink to the deep-ocean floor. This deep, cold Antarctic Bottom Water moves back to the north through the Atlantic, Indian and Pacific Oceans. Because cold water dissolves larger amounts of oxygen from the atmosphere than does warm water, these deep waters rich in oxygen aerate the bottom of the world's oceans. Second, much of the rest of the Circumpolar Deep Water eventually wells up to the surface in the Pacific. There, it becomes entrained in the major surface currents that move west from the North Pacific through the Indonesian archipelago, across the Indian Ocean, down the East African coast in the Agulhas Current and across the South Atlantic to the Equator in the Benguela Current, gaining salt and heat along the way. This water ends up feeding in to the southern end of the Gulf Stream, to repeat the cycle. This global pattern of southward-moving North Atlantic Deep Water and northward-moving warm salty surface water forms the so-called Thermohaline Conveyor Belt (from '*thermo*', meaning heat, and '*haline*', meaning salt), which moves heat and salt around the globe. Wally Broecker is credited with devising this cartoon of ocean circulation[57]. As we'll see later, there is a third pathway not shown in Figure 12.4. Northward-moving surface water in the Southern Ocean eventually sinks at the polar front near 60° S to form Antarctic Intermediate Water, which circulates through the world ocean at depths of 600–1000 m.

The Thermohaline Conveyor, these days referred to by physical oceanographers as the Meridional Overturning Circulation (MOC), is geologically young. It did not exist before the opening of the Drake Passage. Robbie Toggweiler from Princeton and H. Bjornsson from Iceland used experiments with an ocean model in 2001 to show that, prior to the opening of the passage, ocean temperature should have been symmetric about the Equator, with meridional overturning being driven by deep-water formation at the poles in both hemispheres[58]. With the passage open, the overturning took the form of an interhemispheric conveyor, with deep-water formation primarily in the Northern Hemisphere. The conveyor made temperatures rise in the Northern Hemisphere and fall in the Southern Hemisphere, as the ocean transported heat north across the Equator, especially in the Atlantic. The high salt content of the warm surface water allowed northern waters to become dense when cooled, thus driving the return flow at depth as North Atlantic Deep Water. While salinity differences are obviously important in driving the conveyor, Toggweiler's model showed that the conveyor could not be entirely driven by buoyancy. The westerly winds funnelled through Drake Passage do more than 'set the stage' for the work of the buoyancy forces in the North Atlantic: they are an indispensible part of the conveyor circulation, because they drive the upwelling of deep water around Antarctica to bring North Atlantic Deep Water to the surface[58].

In glacial times, the power of the Thermohaline Conveyor was much reduced, as we can see from the geochemistry of microfossils from cores collected at different water depths. In 1982, Ed Boyle from MIT and Lloyd Keigwin from Woods Hole Oceanographic Institution (WHOI) found that the shells of benthic foraminifera contained variations down-core in the ratio of cadmium (Cd) to calcium (Ca)[59]. The ocean distribution of Cd follows that of nutrients, so the Cd/Ca ratio in these shells records variations in the nutrient content of bottom waters. North Atlantic Deep Water turned out to be relatively poor in Cd compared with deep water of southern origin. The Cd/Ca evidence told Boyle and Keigwin that the intensity of the northern source relative to the southern one diminished by a factor of two during severe glaciations. $\partial^{13}C$ values also document the distribution of nutrients, being low in old waters containing abundant dissolved organic carbon rich in ^{12}C. Both Cd/Ca and $\partial^{13}C$ distributions suggest a strong stratification in the North Atlantic at the Last Glacial Maximum, with a low-nutrient, high-$\partial^{13}C$ water

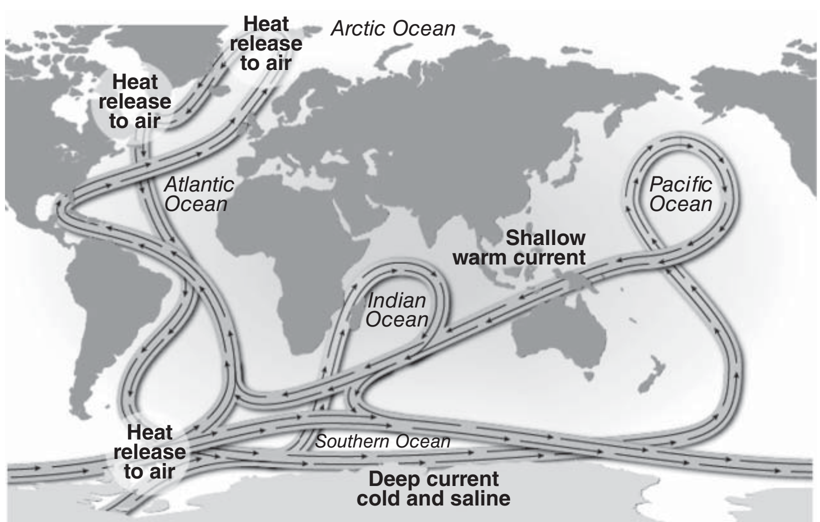

Figure 12.4 *Ocean thermohaline conveyor belt, showing the directions and depths of cold, salty, oxygen-rich deep currents, warm surface currents and vertical connections from deep to shallow and vice versa.*

mass (Glacial North Atlantic Intermediate Water) occupying depths down to around 2000 m, and a high-nutrient, low-$\partial^{13}C$ water mass of southern origin below that[60]. Laurent Labeyrie of the Centre National de la Recherche Scientifique (CNRS) laboratory in Gif-sur-Yvette, southwest of Paris, a winner of the European Geophysical Union's Hans Oeschger Medal in 2005, confirmed that the overturning circulation of the glacial Atlantic was shallower and weaker than today's[61]. Production of North Atlantic Deep Water was much reduced during the Last Glacial Maximum.

With the development of vast ice sheets on western Europe and North America, the sea's surface froze during the winter as far south, at times, as 40° N – the latitude of Boston in the west and Lisbon in the east. With the freezing over of the Norwegian-Greenland Sea, there was no longer a significant source for North Atlantic Deep Water[62,63]. Boyle realised that this icing up prevented deep-water formation[64]. The Cd/Ca signal showed strong periodicity at 41 Ka, synchronous with changes in the tilt of the Earth's axis, confirming that Northern Hemisphere ice cover (at least in the Norwegian-Greenland Sea) was controlled by insolation at high latitudes. Increasing tilt produced higher summer insolation and less ice cover.

The icing up of the North Atlantic north of 40° N at the Last Glacial Maximum switched off the branch of the Gulf Stream that extended into the Nordic Seas and deflected the Gulf Stream east between New York and Lisbon. Sea ice also extended 10° closer to the Equator in the Southern Hemisphere, putting a lid on Southern Ocean processes. With the growth of ice on land, sea level fell 120–130 m. The growth of land ice and sea ice increased Earth's albedo significantly, helping to cool the planet.

12.5 A Surprise Millennial Signal Emerges

In the 1970s, marine geologists dredged large angular boulders of continental rocks such as granite from the Mid-Atlantic Ridge in the North Atlantic, and some people unwisely took this to mean that the ridge was made of continental rock[65]. By 1977, Bill Ruddiman of Lamont-Geological Observatory had mapped wide swathes of ice-rafted glacial debris over much of the North Atlantic seafloor north of a line connecting Boston with Lisbon, and it was realised that the angular boulders were also ice-rafted[66–68]. More detailed studies of piston cores from the North Atlantic by the German geologist Hartmut Heinrich (1952–), of the Deutsches Hydrographisches Institut in Hamburg, found ice-rafted debris concentrated in six layers deposited half a precession unit (11 Ka) apart[69]. Heinrich's discovery, in 1988, aroused intense interest, and an international team formed under the leadership of Gerard Clark Bond (1940–2005) of Lamont in order to investigate these sediments, which Bond's group named 'Heinrich layers'[70].

The layers formed between 10 and 60 Ka ago, apparently in response to surging within the Canadian ice sheet, which led to it breaking up into myriads of icebergs – an armada carrying rocks across the ocean. Charles Lyell would have been pleased, since this was the mechanism he had proposed back in the 1830s to explain the widespread distribution of glacial debris across western Europe. However, as we saw in Chapter 2, it is not reasonable to call on drifting icebergs to cover Europe – there, we must call on a grounded ice sheet to explain the origin of the boulders and associated boulder clay. Along with the melting icebergs came vast volumes of cold, fresh water, cooling the North Atlantic. Each event lasted around 750 years and began suddenly – within about a decade. These outbreaks reached the continental margin off Portugal, as I discovered in 1995, when I went to sea with Nick Shackleton on the maiden voyage of the new French research vessel *Marion Dufresne*, with Yves Lancelot as chief scientist. My goal was to collect samples for a project on 'Northeast Atlantic Palaeoceanography and Climate Change' that I had cooked up with geochemist John Thomson, from my institute at Wormley in Surrey. We cored the Portuguese continental margin en route from the Azores to Marseilles. I can heartily recommend French research cruises to those with a taste for cordon bleu cuisine and fine wines.

Much to our delight, Thomson and I found in the *Marion Dufresne*'s 40 m-long giant piston cores certain layers rich in magnesium derived from dolomite rock deposited as ice-rafted debris by the armadas of melting icebergs from Canada[71,72]. Later, I spent a short sabbatical at Lancelot's laboratory at Aix-en-Provence, working up some of the data from our long cores. I remember his kindness in loaning me a car to help me get around during my month there. It was also a pleasure to meet other prominent French palaeoclimatologists during my stay, among them Edith Vincent and Edouard Bard.

The periodic irruption of ice-rafting in Heinrich events showed that the climate of the Ice Age was variable at the millennial scale, as well as at Milankovitch's orbital frequencies. Bond was keen to find out more about these millennial-scale processes, and in 1995, with Rusty Lotti, carried out close-spaced analyses down two cores collected west of Ireland and spanning the past 9–38 Ka[73]. Between each of the Heinrich layers they found yet more layers of ice-rafted debris, but of lesser magnitude. They deduced that iceberg calving had recurred at intervals of 2000–3000 years. Examination of the rock grains in these layers showed that, while the carbonate-rich ones came from Canada and were concentrated in Heinrich layers, red (hematite)-stained rock grains from multiple sources were common, along with grains of volcanic glass from Iceland or Jan Mayen Island, in the other millennial-scale layers. By 1997, yet more detailed studies by Bond and his team suggested that the red (hematite)-stained rock grains in these layers came from East Greenland or Svalbard[74]. The ice-rafting took place on average every 1536 ± 563 years. This cyclicity persisted into the Holocene, where the frequency was 1374 ± 502 years – statistically indistinguishable from the cyclicity of the glacial period. Averaging the Holocene and glacial signals gave a frequency of 1470 ± 532 years, which, they thought, '*reflects the presence of a pervasive, at least quasi-periodic, climate cycle occurring independently of the glacial-interglacial climate state*'[74]. Furthermore, '*ocean circulation … [is implicated] as a major factor in forcing the climate signal and in amplifying it during the last glaciation … with a single mechanism – an oscillating ocean surface circulation, we can explain at once the synchronous ocean surface coolings, changes in IRD [ice-rafted debris] … and foraminiferal concentrations, and changes in petrologic tracers*'[74]. Laurent Labeyrie's French team distilled this jargon-rich waffle down to the conclusion that a 'climate oscillator' caused the millennial oscillations[75].

As ever, the science marched on. By 1999, Bond's team had confirmed the persistence of the 1470-year cycle back to 80 Ka ago[76]. Why weren't each of these ice-rafting events associated with massive iceberg outbreaks of the kind forming Heinrich events? Bond thought that after a massive purge of ice during a Heinrich event, it took a few thousand years for the ice at source to grow back and reach the unstable conditions needed for another massive discharge[76]. Noting that Heinrich events got closer together between MIS 4 and the Last Glacial Maximum, he thought that this might reflect deterioration of global climate after the last interglacial, with recovery of the ice in Hudson Strait taking progressively less time as the climate cooled. Heinrich events were not strictly periodic, and the oscillations were internal to the system.

Bond's team found signs of oceanic cooling leading up to Heinrich events[76], including a southward extension of polar surface waters and disruption of the North Atlantic Current, the branch of the Gulf Stream that transports warm waters north[75]. Mark Maslin of the Environment Change Research Centre at University College London suggested that some of Bond's 1500-year cooling events may have produced ice surges from Iceland and East Greenland that were large enough to raise sea level to the extent that it undercut and destabilised the edges of the Laurentide Ice Sheet, precipitating a full-scale Heinrich event[77].

Following up Bond's work, Bill Curry and his team from the WHOI, on Cape Cod, used $\partial^{18}O$, $\partial^{13}C$, and Cd/Ca ratios from benthic foraminifera from a core near Iceland to show that these cooling events occurred at more or less the same time as decreases in the production of North Atlantic Deep Water and cooling of surface waters in the western equatorial Atlantic[78]. After each Heinrich event, warm, saline surface water re-entered the area, and thermohaline circulation resumed[75,78]. Labeyrie and his French team agreed that millennial ice-rafting events were associated with cooling of the surface waters and southward extension of cold and low-salinity Arctic waters, but found that the widespread low-salinity meltwater accompanying these events delayed resumption of the production of deep water by several hundred years by forming a lid on the ocean[75].

Changes at the millennial scale, such as Heinrich events, Bond cycles and the Younger Dryas cold event at the end of the last glaciation (discussed later), are visible in the reconstructions of past sea surface temperatures made for the Mediterranean and North Atlantic from alkenone palaeothermometry[79]. The Canary Current transports these millennial signals south into the tropics along the coast of northwest Africa, where, as a result, the coldest time at the sea surface in the past 80 Ka (−12 °C) was not at the Last Glacial Maximum, at around 20 Ka ago, but during Heinrich event 2, just before the Last Glacial Maximum, and Heinrich event 1, just after the Last Glacial Maximum[80]. Surface currents also transported these signals, including all those seen in Greenland ice cores for the past 50 Ka, into the Mediterranean, affecting the climate there, too[81]. As off northwest Africa, the temperatures of these events were colder in the Mediterranean than were those of the Last Glacial Maximum.

Some of the millennial changes were quite fast, and coincided with large, rapid climate changes recorded in ice cores from Greenland. The discovery of these large and sudden changes in piston cores and ice cores was a revelation, proving that the glacial period was far from being as stable as was once supposed. It now appeared that slow, steady changes in the climate of that time led eventually to 'tipping points' at which the climate changed to a different state, before eventually tipping back to its previous one[77]. Some of the most pronounced changes, especially around the northern North Atlantic, occurred within a few decades, or even just a few years. These sudden step-like transitions would have had significant effects on human life at the time, making it prudent for us to reflect on what caused them in the past and what might do so in the future as global warming continues. Was this 'flickering' between one state and another typical just of glacial periods, or might it occur also in interglacials like the one we are now living through? We revisit this question in Chapters 13 and 14, when we explore these rapid changes in some detail.

Before concluding this section, we should note the conclusion of Julian Dowdeswell, of Scott Polar Research Institute, concerning the behaviour of ice sheet margins. The response of an ice sheet to a climatic event or a rise in sea level is not necessarily uniform from one ice margin to another[82]. Random processes could have led to surges in Northern Hemisphere ice sheets during the last glacial period, which means that even if some external forcing agent is involved, there may be a random component to Heinrich events. Readers who want more detail on the competing models explaining periodic surging by ice sheets may find it useful to consult Cronin's *Paleoclimates*[41].

12.6 Ice Age Productivity

Prior to my *Marion Dufresne* cruise, I had been trying to test the hypothesis that the increase in the steepness of the thermal gradient between the Equator and the poles during glacial times increased the strength of the Trade Winds and so enhanced upwelling on continental margins. I did so in 1995, using a piston core that I had collected back in 1973 from the continental slope off Namibia when I was chief scientist on one leg of RV *Chain* cruise 115 between Dakar and Cape Town[83]. I wanted to know whether the upwelling associated with the Benguela Current changed with time, and, if so, how and when. The answer required assembling a multidisciplinary team.

To assess the temperature history, I needed alkenone data. Fortunately, I knew Geoff Eglinton well, having first met him in the late 1970s, when we were both members of the Organic Geochemistry Panel advising the DSDP. Geoff agreed to provide the alkenone data we needed to determine sea surface temperatures over the past 70 Ka[83]. To our surprise, we found that they were coldest and most productive during MIS 3 (60–24 Ka ago), a warm interstadial during the last glacial period. We deduced that the alongshore Trade Winds had been strongest during stage 3, thus driving more upwelling of cold, nutrient-rich and highly productive water. In the colder isotopic stages above and below (MIS 2 and MIS 4), waters were slightly warmer and slightly less productive. While this could indicate reduced wind strength, the evidence suggested that wind directions might have changed, there being more winds rich in desert dust blowing directly offshore, and fewer of the alongshore Trades that drove upwelling currents. In contrast, today's sea surface along that margin is very much warmer and less productive than it was in glacial times, although upwelling still prevails there, and surface waters are still highly productive – this is one of the world's great fishing grounds. Our organic carbon signal fluctuated through time on a cycle of about 22 Ka, evidently driven by variations in the precession of the Earth's orbit.

Several other researchers were extracting climate signals from piston cores from close by in the southeastern Atlantic at the same time, and we pooled our resources to show that the Heinrich events, when icebergs were most abundant in the North Atlantic, were represented by warming oceanographic signals in the South Atlantic[84]. We explore the reasons for this unexpected hemispheric climatic connection in Chapter 13.

Looking at the Benguela Current system in rather more detail, a group of German researchers used the alkenone method to show that the warming characteristic of the Last Glacial Maximum began before it, probably in response to a change in the winds that allowed subtropical surface waters to move south down the coast from Angola[85]. This coincided with conditions less favourable for upwelling, which helps to explain the decrease in organic carbon accumulation we found on the continental slope off Walvis Bay. Timothy Herbert of Brown University found much the same thing off southern California – just a slight cooling at the Last Glacial Maximum, close to the coast. In both of these environments, the cores from the open ocean farther offshore contained temperature profiles typical of those seen in the global SPECMAP stack, with the coldest sea surface temperatures at the time of maximum ice volume – the Last Glacial Maximum[79]. Along the Benguela and California coast, then, upwelling was stronger and more productive than during the Holocene during glacials, including at the Last Glacial Maximum, but was not as strong or productive at the Last Glacial Maximum as it was in the interstadial period (MIS 3).

Was our finding that upwelling had decreased during peak glacial times (MIS 2) typical, I wondered? Yes. As Sigman and Haug explained, the coastal upwelling zones off California and Mexico in the north and off Peru in the south were less productive during the recent glacial period[50]. Sigman and Haug attribute this to the effect of continental cooling (and a large North American ice sheet, in the case of the California Current) on the winds that currently drive coastal upwelling. Upwelling associated with monsoonal circulation in the Somali Current of the western Indian Ocean also decreased, because the cooling of the Tibetan Plateau weakened the southwest monsoonal winds. In contrast, upwelling was strengthened in the equatorial Indian Ocean, where the northeast monsoonal winds remained strong; the same applied in the South China Sea. It was also increased in the eastern equatorial Pacific[86] and the equatorial Atlantic[87].

Yet another geochemical technique helped to ascertain the history of productivity in the Southern Ocean during the Ice Age. Because the element thorium (Th) is rapidly adsorbed from the ocean on to sinking particles, and there is little lateral transport of dissolved Th from its site of production to its site of deposition, an isotope of thorium (^{230}Th) can be used as a proxy for the vertical downward flux of sediment. Use of this technique helped to determine the vertical fluxes of opal, barium, organic carbon and other proxies for palaeoproductivity in the Southern Ocean[88]. Compared with the Holocene, productivity was lower south of the polar front and higher north of the polar front during the Last Glacial Maximum in the Atlantic and Indian Ocean sectors[51, 90]. The main planktonic organisms in these cold waters are siliceous diatoms. Diatom production shifted north as temperatures cooled. While this applied in the Atlantic and Indian Ocean sectors, it did not apply in the Pacific sector, where productivity was lower in the Last Glacial Maximum than in the Holocene.

The northward shift in productivity reflects northward migration of the oceanic fronts and their accompanying sea ice during glacial times. The absence of high productivity in the Pacific sector was probably due to its excessive distance from the westerly sources of dust that transported Fe to fertilise the ocean. The lower productivity of Antarctic waters during the recent glacial period was most likely due to decreasing supply of deep water to the surface, resulting from the diminished supply of North Atlantic Deep Water, which would have driven a relative fall in atmospheric CO_2. In addition, the prevailing westerly winds shifted northwards as the Hadley Cell shrank, reducing upwelling in the Antarctic coastal sector. Besides that, more extensive cover of sea ice in the Southern Ocean limited the exposure of the ocean to the air, contributing to a fall in atmospheric CO_2[89–91]. At the Last Glacial Maximum, sea ice was double its present extent, both in winter and in summer[91]. Sigman and Haug suggest that salinity stratification associated with sea ice was a major limiting factor on CO_2 exchange with the air during glacial times[51]. Whether there was a net change in total productivity of the Southern Ocean from the Last Glacial Maximum to the Holocene remains a topic for debate[88]. It seems more likely that marine productivity stayed the same but underwent a lateral shift from south to north in the Last Glacial Maximum. Regardless of what happened in the polar regions, studies of ^{230}Th in the equatorial Pacific show little change in productivity from glacial to interglacial[88].

12.7 Observations on Deglaciation and Past Interglacials

The last deglaciation was the most massive change in Earth's climate in the past 25 Ka. The Northern Hemisphere ice sheets began to melt back around 21 Ka ago, as insolation and CO_2 began to rise. Rising seas contributed to the rapid decay of those ice sheets, encouraging an increase in the rate of flow of ice streams draining the interior, thus thinning the ice sheets and facilitating their collapse[92]. Increased melting formed large meltwater lakes on the southern fringes of the ice sheets, especially in North America, where Lake Agassiz covered an area about the size of the Black Sea. Its remnants today form Lakes Winnipeg and Manitoba. The sudden drainage of the lake put a freshwater cap on the North Atlantic, shutting down the northern arm of the Thermohaline Conveyor. This cap was probably responsible for the Younger Dryas cold period or stadial interrupting the deglaciation between 12 800 and 11 500 years ago, which caused temperatures to drop 5 °C in the United Kingdom, for instance. A further drainage from the lake gave rise to a brief cooling 8200 years ago, which we examine in Chapter 14. Other meltwater pulses occurred at around 14 200 and 11 000 years ago[93]. In many respects, the cold Younger Dryas period represented a temporary return to the glacial circulation pattern of reduced North Atlantic Deep Water. Surprising though it may seem, it was primarily a Northern Hemisphere phenomenon, although the alkenone data show that sea surface temperatures fell by some 12 °C off western North America, and demonstrate cooling of the same age in the South China Sea, the Indian Ocean and the South Atlantic[79]. The puzzle of how Lake Agassiz drained into the ocean was eventually solved. From gravels and a regional erosion plain in northern Canada, Julian Murton and colleagues at the University of Sussex showed in 2010 that it discharged along the path of the Mackenzie River[94].

Temperatures derived from alkenone data can also be used to check the temperatures obtained for the Last Glacial Maximum by CLIMAP researchers. The alkenone data show that the surface ocean was cooler then than CLIMAP researchers thought, but that the tropics cooled much less than the high latitudes, perhaps by only 1 °C[79].

Are past interglacials analogues for the Holocene – the interglacial we are now living in? The simple answer is: no. Interglacials are not all alike. The modulating effect of the roughly 400 Ka cycle of eccentricity means that the interglacial most similar to our own is that from roughly 400 Ka ago, during MIS 11[95]. Analyses of the ∂^{18}O ratios in samples of the right-coiling planktonic foraminifera *Neogloboquadrina pachyderma* from stage 11 in deep-sea drill cores in the northeast Atlantic show that sea surface temperatures varied by less than ±1 °C from the long-term mean for at least 30 Ka[96]. The near-circular orbit of the Earth at the time prevented the 20 Ka precession signal from having much effect within this isotope stage. In effect, the Milankovitch cycle 'missed a beat', prolonging the interglacial to close on 50 Ka. MIS 11 was in effect about two precession cycles long, instead of one.

As André Droxler of Rice University in Houston pointed out, MIS 11 and the present interglacial are similar because their orbital variables are almost identical. According to Droxler and colleagues, '*both interglacials correspond to times when the eccentricity of the Earth orbit was at its minimum, so that the amplitude of the precessional cycle was damped*'[97]. The strongest and longest Pleistocene interglacial, stage 11, had prolonged intense warmth, sea level stands up to perhaps 13–20 m above present levels[98] and significant poleward penetration of warm waters. It lasted twice as long as more recent interglacial stages. The Holocene is likely to be just as long. In Chapter 13 we will look at possible explanations for these patterns.

The warming in stage 11 was important for the establishment of coral reefs. Wolf Berger of Scripps and Gerold Wefer of the University of Bremen used deep-sea drill core data to show that the western Pacific warm pool of surface water expanded dramatically some 400 Ka ago, helping to explain the growth of Australia's Great Barrier Reef[99]. At that time, shallow carbonate platforms grew to the point where they clogged the flow of surface water through the Indonesian islands between the Pacific and Indian Oceans. The warming also triggered the establishment of other barrier reefs, like that off Belize[97]. These reefs grew when the large rise in sea level at the end of the previous glacial maximum extensively flooded fluvial plains, preventing the former supply of riverborne silt and sand from reaching offshore reef sites.

'*Will such warm conditions be replicated as the Holocene continues?*' asked Droxler and colleagues[97]. Yes, they concluded, '*we can expect another ~20,000 years of interglacial conditions, independent of any anthropogenic forcing*'[97]. Is MIS 11 an exact analogue for the Holocene? Not according to David Hodell of the University of Florida at Gainsville, who found maxima in the ∂^{13}C of the planktonic foraminifera *Globigerina bulloides* and in fragmented foraminiferal remains in stage 11 sediments from a deep-sea drill core from the Cape Basin off South Africa[100]. The same indicators in sediments from the last 100 Ka and the Holocene have much lower values, however, and the carbonate compensation depth was 600 m shallower in stage 11 than it is now. These patterns suggest a lowering in the concentration of carbonate ions in the ocean at that time, possibly related to the massive building of barrier reefs in shallow waters. Atmospheric and oceanic feedbacks are not operating in exactly the same way today as they were then. Even so, a comparison of sea surface temperatures and of ∂^{18}O ratios from benthic foraminifera in the southeast Atlantic showed that the 11.7 Ka of the Holocene are indeed comparable to the first 12 Ka of MIS 11[101].

What about the last interglacial, the Eemian, during MIS 5, which began at around 135 Ka ago and lasted until around 110 Ka ago? Alkenone and Mg/Ca data from marine sediments suggest that it was warmer than the late Holocene by up to 3 °C – consistent with stage 5 experiencing significantly higher orbital insolation[79]. These new data improve on the CLIMAP data, which suggested that there was little difference between stage 5 and today. It now seems likely that stage 5 was as warm as the early

Holocene climatic optimum, when insolation was much higher than it is today. We can extract more evidence of Eemian climate change from pollen and lake records in central Europe, loess sediments from central China and marine sediment cores from the eastern subtropical Atlantic. These data show evidence for a single, sudden cool event in the middle Eemian at about 122–120 Ka ago, showing that short, sharp cold periods can occur in interglacials.

Compared with today, global ice volumes were smaller and solar radiation was 13% stronger over the Arctic in summer during the Eemian interglacial. According to Gifford Miller and his team, sea ice and permafrost were vastly reduced, boreal forest expanded to the Arctic shore and most Northern Hemisphere glaciers melted[35]. Summer temperature anomalies over Arctic lands were 4–5 °C above present values, especially in the Atlantic sector. Northern Canada and parts of Greenland were 5 °C warmer than today in summer, but Alaska and Siberia were only about 2 °C warmer. Interpretation of marine data is complicated by the stratification of the Arctic Ocean, which commonly has a cool relatively fresh cap (<-1 °C) overlying warmer subsurface waters (>1 °C).

A final point to bear in mind is that interglacials as warm as the present one occurred for only about 10% of the time in the late Quaternary[102]. The climate of the past 800 Ka was predominantly cold.

12.8 Sea Level

Rising sea level is one of the most highly visible results of a warming world, driven by the melting of ice on land and the expansion of warming seawater. These changes are termed 'eustatic'[102]. Although changes in sea level provide us with yet another proxy for past climate change – especially for ice volume – the relation between climate and sea level is not simple, as we saw in Chapter 11. As R. Lawrence Edwards of the University of Minnesota and his colleagues remind us[103], sea level can also change as the result of tectonic uplift or sinking of the Earth's surface, or of isostatic changes, through which land sinks beneath ice sheets but rises around their periphery to form a 'fore-bulge' – a process that reverses when the ice sheets melt. Only eustatic change is truly global. Tectonic and isostatic adjustments cause local or regional changes that complicate the extraction of a global sea level signal. Lyell knew all about tectonic effects, having observed that the so-called Temple at Serapis in Italy had first been partly drowned and then uplifted. These competing signals must be unravelled to separate the local from the global signal, as we see in some detail in Chapter 14.

There is also the question of rates. For example, the Scandinavian ice sheet melted by about 6000 years ago, but Scandinavia is still slowly rising. So too is Scotland, which lost its ice long ago. In contrast, southern England, the southern edge of the Baltic and the west coasts of Germany and the Netherlands, which were on the fore-bulge around the European ice sheets, are still slowly subsiding. Similarly, the parts of the northernmost United States and Canada that lay beneath the Laurentide Ice Sheet are now rising, while the southern United States, which formed the fore-bulge area outside that ice sheet, is slowly sinking. Within the area of the former Laurentide Ice Sheet, its core region – Hudson's Bay – is still depressed below sea level, although it is rising slowly.

Past sea levels can be determined directly by using the ^{14}C or other radiometric techniques to date carbonates such as reefs and other features that formed at or very close to sea level[103]. These techniques include U/Th dating, which involves calculating ages from radioactive decay relationships between ^{238}U, ^{234}U and ^{230}Th; this is also known as ^{230}Th dating. A further check on accuracy can be obtained from U/Pa dating, in which ages are calculated from the relationship between Uranium-235 (^{235}U) and its daughter, Protoactinium-231 (^{231}Pa). U/Th and U/Pa dating extend the range of ^{14}C dating (maximum 50 Ka) to 250 Ka (^{230}Pa) and 600 Ka (^{230}Th). These techniques, like AMS ^{14}C dating, came into their own after the mid 1980s, with the development of mass spectrometric measurements that reduced sample size and increased the speed and precision of analysis. Even so, despite the accuracy of the dates, all estimates of past sea level come with some uncertainty.

In this section, we focus on how high sea level might have been during past warm interglacials. The data available to Edwards in 2003 suggested that sea levels were up to 20 m above today's level in MIS 11 (400 Ka ago); up to 29 m above in MIS 9 (330 Ka ago); up to 9 m above in MIS 7 (240 Ka ago); and around 5 ± 3 m above in MIS 5, the last interglacial (100 Ka ago)[103]. The rise in sea level from a low point of about -130 m following the Last Glacial Maximum is known in some detail, thanks to comparable data from the New Guinea's Huon Peninsula, Tahiti, South East Asia's Sunda Shelf and northwest Australia's Bonaparte Gulf. These are far-field sites remote from polar ice sheets. Isostatic adjustments are unimportant, and the data reflect a true global signal.

These estimates were refined for the last interglacial (MIS 5) by a team led by Robert Kopp of Princeton, who in 2009 compiled a large number of indicators of local sea level change and applied a statistical approach to estimating global sea level[104]. They found a 95% probability that global sea level peaked at least 6.6 m higher than today, and a 67% probability that it exceeded 8 m, but only a 33% likelihood that it exceeded 9.4 m. Rates of sea level rise could have varied between about 56 and 92 cm per century. For comparison, the present rate of sea level rise is around 33 cm per century. The last interglacial was only slightly warmer than the present – by about 2 °C. Achieving a sea level rise in excess of 6.6 m higher than present '*is likely to have required major melting of both the Greenland and West Antarctic ice sheets*', they concluded[104].

Eelco Rohling of the University of Southampton used data from the Red Sea to suggest, like Kopp, that during the last interglacial, sea level reached a mean position of +6 m, with individual short-term peak positions up to about +9 m compared with today's level. The rate of rise of sea level, according to Rohling and colleagues, was about 1.6 m per century, which '*would correspond to disappearance of an ice sheet the size of Greenland in roughly four centuries*'[105]. This rate occurred when global mean temperature was 2 °C higher than today. Kurt Lambeck (1941–) (Box 12.3) of the Australian National University evaluated sea level for the last interglacial by using tectonically stable sites in the 'far field', estimating that it was 5.5–9.0 m above today's[106], which is consistent with the findings of both Kopp and Rohling.

Maureen Raymo of Boston University looked with a member of Kopp's team, Jerry Mitrovica of Harvard, into the contentious suggestion that Pleistocene shoreline features on the tectonically stable islands of Bermuda and the Bahamas were more than 20 m higher than today in MIS 11, some 400 Ka ago[107]. They found both sites to be located on the outer edge of the peripheral bulge of the Laurentide Ice Sheet. To account for postglacial crustal subsidence at these sites, their elevations were adjusted downwards by about 10 m. That reduced eustatic sea level rise to ~6–13 m above the today's level in the second half of MIS 11. This rise was caused by prolonged warmth leading to the collapse of both the Greenland and West Antarctic Ice Sheets. Given that the likely maximum rises in sea level for the melting of the Greenland and West Antarctic Ice Sheets are 7 and 5 m respectively, the change of 6–13 m suggests that changes in the volume of the East Antarctic Ice Sheet were minor.

Roland Gehrels of the University of Plymouth drew attention to other flaws in the analysis of sea level change, focusing on the rise of sea level since the Last Glacial Maximum[108]. Fairbanks' classic paper on sea level rise, published in 1989, and based largely on data from Barbados, suggested that sea level was 120 m below present at the Last Glacial Maximum[109]. Gehrels cited three possible sources of error in Fairbanks' data. First, Barbados lies in an active tectonic setting on the edge of the Caribbean Plate. Even slight tectonic changes could have affected the absolute amount of sea level change registered on the island. Second, Barbados lay on the trailing edge of the glacial fore-bulge pushed up around the margins of the Laurentide Ice Sheet, and the collapse of that feature would have created further vertical change. Third, Fairbanks' curve included data from other islands (Martinique, Bahamas, Puerto Rico and St Croix), '*thereby introducing errors resulting from differential isostatic movements and regional sea-level variations*'[108].

Focusing on far-field sites, Yokoyama and colleagues calculated in 2000 that global sea level was as low as 130–135 m below present levels at the Last Glacial Maximum[110]. Claire Waelbroeck and colleagues provided much the same picture, with a maximal lowering to −135 m at the Last Glacial Maximum[111]. They also calculated that sea level fell to about −125 m in MIS 6 (140 Ka ago) and MIS 10 (345 Ka ago), and to about −110 m during MIS 8 (250 Ka ago). Their values differ significantly from those derived by Nick Shackleton[112] and provide good reasons for discounting Shackleton's data. Lambeck recently estimated sea level lowering at the Last Glacial Maximum as −134 m[113], explaining that this was a measure of grounded ice volume, including ice grounded on shelves. Along far-field continental margins, the Last Glacial Maximum sea levels would generally be

Box 12.3 Kurt Lambeck.

Lambeck, professor of geophysics at the Australian National University in Canberra, was born in Utrecht in the Netherlands. From 2006 to 2010 he was president of the Australian Academy of Science. He has been honoured with several awards, among them fellowship in the French and US Academies of Science and the United Kingdom's Royal Society, the international Balzan Price (2012) and the Wollaston Medal of the Geological Society of London (2013).

less than this due to isostatic/gravitational effects, while in mid oceans they would exceed this (James Scourse, pers. comm.).

When it was published in 1989, the Barbados sea level curve[109] gained a great deal of attention because it showed evidence for episodes of very rapid sea level rise, most notably the event known as meltwater pulse 1a, dated to about 14 Ka ago, when sea levels rose by 15–25 m at rates of over 40 mm/yr[108]. The jury is still out with respect to the source of the meltwater pulse, which could have originated in surges of the Laurentide or the Antarctic Ice Sheets, or from the discharge of large glacial lakes.

Clearly, sea level has changed through time in response to the waxing and waning of ice sheets, and measurements of past sea level can be used as a proxy for ice volume change. In order to refine these calculations further, there is much still to learn about regional variations in sea level, which depend on local tectonics and glacial isostatic adjustments of the Earth's surface to the addition or removal of large masses of ice. Knowing that '*regional sea-level changes resulting from polar ice melt can depart by up to 30% from the global mean*', Gehrels concluded that '*regional sea-level variability precludes the use of the term "eustasy" in the traditional sense (i.e. global average sea-level change). The recognition that "eustasy" is only a concept should ... lead to [improved] regional sea-level predictions*'[108].

As we saw in Chapter 11, one way to avoid problems created by the use of past shorelines to establish past sea levels is to use the $\partial^{18}O$ composition of seawater, which is related to both ocean temperature and ice volume, both of which are, in effect, global. Subtracting the temperature signal enables us to determine ice volume and hence sea level[114, 115].

Using stable oxygen isotope analyses of planktonic foraminifera and bulk sediments from the Red Sea, Eelco Rohling and his team developed a relative sea level record for the past 520 Ka[98]. It shows a striking similarity to the record of Antarctic temperature, a relationship that remains the same regardless of whether the climate system is shifting towards glaciation or deglaciation, and which does not drift back through time. As this is a robust relationship within the climate system, it could be applied in estimating the effects of future climate change (see Chapter 16).

Jacqueline Austermann of Harvard agreed with the revisions to the Fairbanks model of sea level change[116]. Austermann and colleagues' model confirmed that at the Last Glacial Maximum, sea level should have been lowered to about −130 m, not −120 m as Fairbanks thought. That left a significant volume of ice in the Northern Hemisphere unaccounted for: it appeared from sea level data that more ice must have melted than had been available in the ice sheets of Laurentia (North America) and Fennoscandia. A joint team from Germany's Alfred Wegener Institute and the Korean Polar Research Institute discovered in 2013 that the furrows that arise when large ice sheets become grounded on the seabed are widespread on the seabed off the coast of northeast Siberia. The team estimated that the furrows represented the former existence of an Arctic ice sheet that covered an area at least as large as Scandinavia and was up to 1200 m thick. This previously missing ice may well explain the accounting discrepancy[117].

The study of sea level is worth an entire book. Readers wishing to probe further might like to start with a 2010 compilation of global data entitled *Understanding Sea-Level Rise and Variability*[118].

Returning to the astronomical calculations, it is now abundantly clear that regular changes in the Earth's orbit and axial tilt cause the amount of insolation we receive to vary within narrow limits, a discovery as influential in its own way as plate tectonic theory. The limits define a 'natural envelope' in which the maxima and minima are seldom if ever exceeded. These limits apply in turn to global temperature, which varied over the narrow global range of 4–5 °C between glacial and interglacial times. Back in 1982, Wolf Berger realised that this was '*a striking phenomenon, important especially for the survival of higher organisms*'[119]. This Ice Age natural envelope was superimposed on a background climate whose extremes varied within another natural envelope, in which, as we saw at the end of Chapter 9, the variation in CO_2 was driven by plate tectonic processes including the emission of CO_2 from volcanoes and its extraction by weathering, especially in mountainous areas, and by sedimentation in growing ocean basins. The natural envelope of CO_2 from 200 to 1000 ppm (Chapter 9) was only occasionally exceeded, as we saw in Chapter 10, when hothouse conditions prevailed.

To summarise, our view of Pleistocene climate changed dramatically from the mid 1960s onwards, when piston corers and deep-ocean drilling enabled us to study for the first time the climate history recorded over the 66% of the Earth's surface covered by water depths of more than 200 m. Application of novel palaeontological and geochemical techniques showed that Earth had experienced many more substantial variations in climate than was apparent from studies of glaciation on land, where the advances of later glaciers and ice sheets removed the records of earlier ones. The realisation that changes in insolation were intimately linked to changes in temperature and ice volume enabled palaeoclimatologists to tune their signals of climate change to orbital changes, thus deriving a novel method for dating core horizons to an unheard of accuracy of ±2000 years, over periods of more than 1 million years. Furthermore, as clockwork variations in insolation could be projected into the future, it became possible to estimate the extent, duration and timing of the next glaciation.

The growing global array of deep-ocean cores enabled comparisons to be made between glacial and interglacial conditions caused by changes in the extent of sea ice and in the Thermohaline Conveyor Belt. This confirmed the validity of orbital cycles and highlighted the saw-toothed pattern of actual climate change, reflecting the slow build-up of ice sheets and their rapid eventual demise, a pattern suggesting that once warming caused melting to reach some critical rate, land ice reservoirs collapsed to produce a glacial termination[119]. Carbon isotopes could be used to estimate the amount of CO_2 in the air, showing that CO_2 varied with temperature, presumably because carbon reservoirs slowly built up in peat beds, rain forest debris, fine-grained organic-rich sediments and deep-ocean waters as ice accumulated, before decaying rapidly and releasing CO_2 as ice melted and the climate warmed[119].

CO_2 provided one positive feedback, affecting temperature. Sea ice provided another, first through its affect on albedo, and second through governing the exchange of CO_2 between ocean and atmosphere[119]. Dust provided a third, increasing fertilisation of the ocean with iron in glacial times, thereby enhancing CO_2 draw-down; its absence in interglacials had the opposite effect. Water vapour provided a fourth, following CO_2, influencing temperature and governing change in the water cycle. Sea level provided a fifth, increasing or decreasing the area of ocean available for the exchange of CO_2 and water vapour between ocean and atmosphere. Sea levels in past interglacials may have been as high as 9 m above today's.

A millennial level of natural variability became apparent from concentrations of ice-rafted debris. Large glacial outbreaks formed Heinrich events; small ones formed 1500-year Bond cycles. They corresponded with cold periods in the North Atlantic and warm periods in the South Atlantic. Outbreaks from massive glacial lakes flooded the northern oceans from time to time, causing the northern freeze known as the Younger Dryas.

In the next chapter, we will look at the exciting discoveries that the ice core drillers were making on land, and compare them with those emerging from studies of the ocean floor. The history of fossil CO_2 in ice cores enables us to further explore the relationship between CO_2 and the curious 100 Ka climate cycle. We will also examine possible mechanisms for glacial–interglacial climate change.

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