Chapter 6

A Geological History of Climate Change

Jan Zalasiewicz and Mark Williams

Department of Geology, University of Leicester, University Road, Leicester LE1 7RH, UK

1. INTRODUCTION

Earth's climate is now changing in response to an array of anthropogenic perturbations, notably the release of greenhouse gases; an understanding of the rate, mode and scale of this change is now of literally vital importance to society. There is presently intense study of current and historical (i.e. measured) changes in both perceived climate drivers and the Earth system response. Such studies typically lead to climate models that, in linking proposed causes and effects, are aimed at allowing prediction of climate evolution over an annual to centennial scale.

However, the Earth system is complex and imperfectly understood, not least as regards resolving the effect of multiple feedbacks in the system, and of assessing the scale and importance of leads, lags and thresholds ('tipping points') in climate change. There is thus a need to set modern climate studies within a realistic context. This is most effectively done by examining the preserved history of the Earth's climate. Such study cannot provide precise replicas of the unplanned global experiment that is now underway (for the sum of human actions represents a geological novelty). However, it is providing an increasingly detailed picture of the nature, scale, rate and causes of past climate change and of its wider effects, as regards for instance sea level and biota. Imperfect as it is, it provides an indispensable context for modern climate studies, not least as a provision of ground truth for computer models (see below) of former and present climate.

Aspects of climate that are recorded in strata include temperature and seasonality [1,2], humidity/aridity [3], and wind direction and intensity [4]. Classical palaeoenvironmental indicators such as glacial tills, reef limestones and desert dune sandstones have in recent years been joined by a plethora of other proxy indicators. These include many biological (fossilised pollen, insects, marine algae) and chemical proxies (e.g. Mg/Ca ratio in biogenic carbonates). Others are isotopic: oxygen isotopes provide information on temperature and ice volume; carbon isotopes reflect global biomass and inputs (of methane or carbon dioxide) into the ocean/atmosphere system; strontium and osmium are proxies for weathering, and the latter, with molybdenum also, for oceanic oxygenation levels. Other proxies include recalcitrant organic molecules: long-chain algal-derived alkenones as sea temperature indicators [5] and isorenieratane as a specific indicator of photic zone anoxia [6]. These and many other proxies are listed in Ref. [7]. Levels of greenhouse gases such as carbon dioxide and methane going back to 800 ka can be measured in ice cores [8]. For older times, (somewhat imprecise) proxies have been used, such as leaf stomata densities [9,10], palaeosol chemistry [11] or boron isotopes [12]; estimates of greenhouse gas concentrations in the atmosphere have also been arrived at by modelling [13,14].

2. CLIMATE MODELS

Since the 1960s, computer models of climate have been developed that provide detailed global and regional projections of future climate and reconstructions of deep time climate. Some of these models are used to simulate conditions during icehouse climates, for example, of the Late Proterozoic [15], whilst others simulate warm intervals of global climate, such as during the Mesozoic greenhouse [16]. The most widely applied computer simulations of palaeoclimate are general circulation models (GCMs). The increasing complexity of these models has followed the exponential growth in computer power.

GCMs divide the Earth into a series of grid boxes. Within each of these grid boxes, variables important for the prediction of climate are calculated, based upon the laws of thermodynamics and Newton's laws of motion. At progressive time steps of the model the reaction between the individual grid boxes is calculated. GCM simulations rely on establishing key boundary conditions. These conditions include solar intensity, atmospheric composition (e.g. level of greenhouse gases), surface albedo, ocean heat transport, geography, orography, vegetation cover and orbital parameters. Whilst solar luminosity can be estimated with a high degree of confidence for different time periods, some of the other boundary conditions are much more difficult to establish, and the magnitude of the problem increases with greater age. Thus, models of Late Proterozoic climate can establish solar luminosity as 93% of present, but the geography of Proterozoic palaeocontinents is much more controversial [15]. It depends on geological data, in this example from remnant magnetism, preserved within rocks and placing the continents in their ancient position according to the Earth's magnetic field.

Geological data (e.g. sedimentology, palaeontology) are essential to 'ground truth' climate models, to establish whether they are providing a realistic reconstruction of the ancient world, and also to provide data for calibrating boundary conditions for the models. Of major importance for GCM palaeoclimate reconstructions is accurate information about sea surface temperatures (SSTs), as this provides a strong indication of how ocean circulation was working. The most extensive deep time reconstruction of SSTs is that of the United States Geological Survey PRISM Group, based on a global dataset of planktonic foraminifera [17]. This dataset has been used for calibrating a range of climate model scenarios for the 'mid Pliocene warm period' and also includes an extensive catalogue of terrestrial data [18]. This time interval is used for potential comparison with the path of future global warming [19].

3. LONG-TERM CLIMATE TRENDS

Earth's known climate history, as decipherable through forensic examination of sedimentary strata, spans some 3.8 Ga (billion years), to the beginning of the Archaean (Fig. 1). The previous history, now generally assigned to the Hadean Eon, is only fragmentarily recorded as occasional ancient mineral fragments contained within younger rocks – particularly of highly resistant zircon dated to nearly 4.4 Ga ago [20] and thus stretching back to very nearly the beginning of the Earth at 4.56 Ga ago [21]. The chemistry of these very ancient fragments hints at the presence of a hydrosphere even at that early date, though one almost certainly disrupted by massive meteorite impacts [22]. Certainly, by the beginning of the Archaean, oceans had developed, and an atmosphere sufficiently reducing to allow the preservation of detrital minerals such as pyrite and uraninite that would not survive in the presence of free oxygen [23].

From then until the present, Earth's climate has remained within narrow temperature limits that have allowed the presence of abundant liquid water, water vapour and variable amounts of water ice, the last of these (when present) generally accumulating at high latitudes and/or high altitudes. This is despite widely accepted astrophysical models suggesting that the sun has

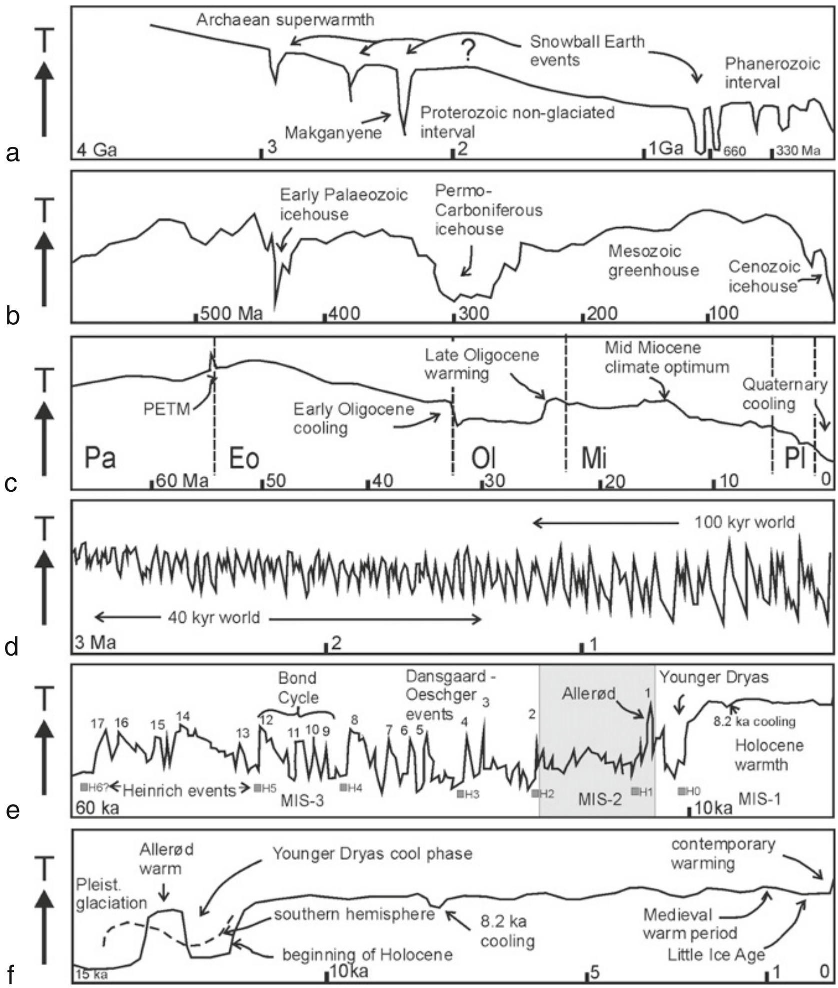

FIGURE 1 Global climate variation at six different timescales. Data adapted from sources including [7,27,48,71,95,100]. On the left side of the figure, the figure 'T' denotes relative temperature. Note that the line denoting 'T' is derived from $\delta^{18}O$ from benthic foraminifera for the Cenozoic time slices (c–e), but for the intervals with polar ice this line will record a combination of ice volume and temperature change.

increased its luminosity by some 20% since the early Archaean, and contrasts sharply with the history of our planetary neighbours: Venus now having a surface temperature of ca. 400 °C with a dense anhydrous atmosphere dominated by carbon dioxide (representing approximately the amount of carbon that on Earth is bound up in rock form as carbonates and hydrocarbons); and Mars with an early history of running surface water (roughly during the Earth's Hadean Eon) and subsequently being essentially freeze-dried.

Hypotheses to explain the Earth's climate stability (that has allowed *inter alia* a continuous lineage of living organisms) have included such as the Gaia hypothesis [24], in which the totality of the Earth's biota operate to maintain optimum conditions for their existence (via feedback mechanisms that involve such factors as albedo and atmospheric composition). Currently, it is thought that terrestrial silicate weathering (a largely abiotic mechanism) is an important factor in Earthly homeostasis [25]. Thus, as temperatures rise through an increase in greenhouse gases, increased reaction rates of rainwater (i.e. dilute carbonic acid) with rock – allied to increased humidity from enhanced evaporation rates – will cause drawdown of carbon dioxide, thus lowering temperatures [26]. Similarly, as greenhouse gas levels and temperatures fall, diminished rates of weathering will allow carbon dioxide levels to rise, and so warm the Earth's climate. The silicate weathering mechanism operates on timescales of hundreds of thousands to millions of years, with greenhouse gas levels having fallen throughout Earth history as the sun's luminosity has increased. At shorter timescales, this mechanism may be over-ridden by other factors, to allow the production of climate states that are hotter or colder than the long-term average.

4. EARLY CLIMATE HISTORY

At long time scales, Earth's (post-Hadean) climate history can be broadly divided into: *greenhouse* (or *hothouse*) states, when the Earth's climate was generally warm, with little or no polar ice; and *icehouse* states with substantial high/mid (and sometimes low latitude) ice masses over land and ocean. Ability to resolve the duration and timing of these states becomes increasingly better as the geological record becomes younger, with a gulf, in particular, between a Phanerozoic record (from 0.542 Ga) that is highly resolved because of an abundant fossil content and a Precambrian record in which dating and correlation are based upon sporadic radiometric dates and, increasingly, chemical and event stratigraphy. Similarly, the Quaternary glaciation is much better resolved than previous Phanerozoic glaciations.

The earliest reasonable indications of climate in the Archaean hint at a very warm world (Fig. 1a): silicon and oxygen isotopes in Archaean and early Proterozoic rocks suggest temperatures of some 50–80 °C, before temperatures declined to 20–30 °C by 1.5 Ga [27,28]. Most of the post-Hadean Precambrian seems to have roughly equated to a greenhouse world in general, and high-carbon dioxide and high-methane atmospheres have both been suggested as a means of maintaining high temperatures in the face of a faint early sun [29,30].

There were, though, the striking exceptions of the 'Snowball Earth' glaciations (Fig. 1a). There are possible representatives (certainly glacial if not 'Snowball') in the early Proterozoic at ca. 2.5–2.0 Ga [31]. But, they are most typical of the late Proterozoic within the 'Cryogenian Period' (now widely used as a geological time period, but not yet properly defined and ratified). Stratigraphic and palaeomagnetic evidence suggests widespread icesheets in at least two pulses (Sturtian 740–660 Ma ago and Marinoan 660–635 Ma ago [32] that reached into tropical latitudes, with ice present on all main continents. Budyko [33] suggested a theoretical basis for a snowball glaciation, showing that if ice extended to within 30° latitude of the equator, the ice albedo effect would produce a positive feedback mechanism allowing ice sheets to grow to the equator. It has been proposed, controversially, that ice encased the entire globe (the 'hard snowball' variant: [34]), preventing exchange between land/oceans and atmosphere. This has been disputed, with opponents preferring 'slush ball', 'zipper-rift' or 'high tilt' Earth models [32] leaving significant areas of ocean ice-free.

Whichever version is nearer the truth, these appear to have been extreme excursions of the Earth system, with deglaciation being rapid, perhaps 'catastrophic', and marked by the deposition of unique 'cap carbonate' deposits – dolomites and limestones that, worldwide, immediately overlie the glacial deposits. Deglaciation mechanisms commonly involve crossing thresholds in greenhouse gas concentrations. In the 'hard snowball' model this takes the form of volcanic carbon dioxide being prevented from dissolving in the ocean or reacting with rock (because of their carapace of ice), and hence building up to levels high enough to cause rapid ice melt, with acid rain then reacting rapidly with newly exposed bedrock to generate alkalinity that precipitated as carbonates. In the 'slush ball' model, deglaciation hypotheses include massive release of methane, with at least local isotopic evidence of methane release accompanied by ice-melt [35]. Perhaps in support of a 'slush ball' or alternative glacial hypothesis, some GCMs do not replicate the conditions in which a 'hard' Snowball Earth could develop even with very low levels of atmospheric carbon dioxide prescribed [15].

5. PHANEROZOIC GLACIATIONS

Phanerozoic time has also been dominated overall by 'greenhouse' conditions [7]. Glaciations during the Phanerozoic were less extreme, neither reaching the equator nor being associated with post-glacial cap carbonates. Three main glaciations took place (Fig. 1b): a late Ordovician/early Silurian 'Early Palaeozoic Icehouse' (ca. 455–425 Ma) [36], with an end-Ordovician glacial maximum [37] that collapsed in a rapid deglaciation; a long-lived Permo-Carboniferous glaciation (ca. 325–270 Ma) [7] with ice covering much of the palaeocontinent Gondwana (leaving widespread traces in South America and Africa, then over the South Pole); and the current glaciation, that began in the southern hemisphere through the Eocene–Oligocene Epoch boundary interval (ca. 35 Ma) with ice growing on Antarctica [38], and developed into a full-scale bipolar glaciation around the beginning of the Quaternary Period, at ca. 2.6 Ma, with the significant expansion of northern hemisphere ice.

Each of these glaciations took place in different contexts, particularly as regards the carbon cycle. The Early Palaeozoic Icehouse took place in the effective absence of either a terrestrial flora or of widespread well-developed (and hence carbon-rich) soils. Hence, the oceans and marine sediments were of prime importance in carbon storage, with the intermittent anoxia of those oceans perhaps playing a key role as thermostat, episodically enhancing carbon sequestration that led to cooling [36]. In the Carboniferous, the explosive growth and widespread burial of plants on deltaic/coastal plain sediments (subsequently becoming coal) has long been considered key in driving down atmospheric carbon dioxide and leading to glaciation [39]. Other mechanisms have been invoked, such as continental rearrangement to alter patterns of ocean currents and hence global heat transport [40].

6. THE MESOZOIC–EARLY CENOZOIC GREENHOUSE

These early Phanerozoic switches between greenhouse and icehouse give invaluable (and increasingly well resolved) information on the mode and rate of climate change. However, it is the temporal background to, and the development of, the current glaciation that offers the most resolved history and the best clue to causal and controlling mechanisms. This is partly because of a biota that is closer to the present one and hence more interpretable, but crucially because there is a widespread oceanic record (buried under the present ocean floors) to accompany that from land and continental seas; Palaeozoic ocean deposits, by contrast, have almost all been obliterated through subduction, with only rare fragments being preserved by obduction on to destructive continental margins.

Mesozoic and early Tertiary climate was generally in 'greenhouse' mode with little (but generally some) polar ice, widespread epicontinental seas and ocean circulation driven by salinity rather than temperature differences (and hence more sluggish than today's, with a tendency to anoxia). Within this broad pattern, there were warmer and colder intervals [7]. Fossil evidence shows that high latitudes, in particular were considerably warmer during this interval, with extensive near-polar forests [16].

This interval includes brief (0.1–0.2 Ma) climate 'spikes' in which sudden temperature rises were accompanied by biotic changes and marked changes in carbon isotopes. These changes suggest massive (thousands of gigatonnes) transfer of carbon from rock reservoirs to the atmosphere/ocean system with the consequence of ocean acidification as well as warming [41]. The best-known of these [42,43] were in the Toarcian Age of the Jurassic Period (ca. 183 Ma) and at the boundary of the Paleocene and Eocene epochs (ca. 55 Ma). The most likely mechanism seems to be some initial warming (perhaps from volcanic carbon dioxide) that triggered large-scale dissociation of methane hydrates from the sea floor [44], although the baking of coal basins by igneous intrusions [45] may also be implicated. By whichever mechanism, the relevance for contemporary global warming is clear as, while humankind has not yet released as much carbon (ca. 600 Gt (gigatonnes)), it has done it much more quickly [46]. Re-equilibration of climate following the spikes was likely achieved via silicate weathering [26,47].

7. DEVELOPMENT OF THE QUATERNARY ICEHOUSE

The development of the Tertiary/Quaternary icehouse took place as a series of steps (Fig. 1c), with relatively rapid transitions between one climate state and the next, strongly suggesting the common operation of thresholds or ‘tipping points’ [48]. The early Oligocene inception is clearly seen as an isotopic and Mg/Ca signal, in benthic foraminifera [49], of ocean cooling and de-acidification [50] linked to the growth of substantial ice on Antarctica. Two mechanisms have been invoked, that in reality were likely inter-related: the separation of South America from Antarctica to open the Drake Passage and hence to allow a continuous circum-Antarctic cold current [51]; and a steep drop in carbon dioxide levels from about $\times 4$ to $\times 2$ present-day levels [38].

Subsequent Tertiary history includes Mid Miocene warming, possibly associated with release of carbon dioxide to the atmosphere via volcanism or meteorite impact (see Ref. [52] for an overview of possible causes) during which tundra conditions were developed at high southern latitudes within 1500 km of the South Pole [53], and late mid-Miocene cooling (often termed the ‘Monterey event’ [54]), which may have been influenced by drawdown of carbon dioxide from the atmosphere or by changes to ocean heat transport that triggered ice sheet growth and cooling [55].

The subsequent Pliocene Epoch marks the final phase of ‘late Tertiary’ climate. The Early and Mid Pliocene represent conditions that overall were somewhat warmer than present, with global ice volumes smaller, and global surface temperatures perhaps 2–3 °C warmer [56]. The last phase of this warmer world was the ‘mid Pliocene warm period’ some three million years ago [17]. Following this interval, global temperatures decreased, ice volumes increased, and the amplitude of glacial-interglacial oscillation also increased [57] heralding the intensification of Northern Hemisphere Glaciation (NHG). As the last interval of warmth, the ‘mid Pliocene warm period’ has received growing attention as a possible comparison for the path of future global warming [58].

The intensification of NHG that is characteristic of the Quaternary Period (sensu [59]) was marked by the growth of substantial ice in the northern polar region [60]. It is associated with ice-rafted debris appearing in North Atlantic Ocean floor deposits, together with the beginning of substantial loess accumulation in central Europe and China, the drying of Africa to create extensive savannah areas, and other global phenomena. This event may partly reflect a further carbon dioxide threshold [61], with strontium isotope evidence of increased rock weathering, not least from uplift phases of the Himalayas [62]. However, there is strong evidence to suggest the importance of enhanced ice growth rather than simply temperature, with the development of the 'snow gun' hypothesis [63] in which the bringing of a warm moisture-laden ocean current against a cold north American continent led to increased snow precipitation and ice formation on that continent, and hence (via increased albedo and other feedbacks) to further cooling.

8. ASTRONOMICAL MODULATION OF CLIMATE

Over the last 40 years, an astronomical pacemaker for the Quaternary 'Ice Age' has been established beyond doubt, comprising variations in orbital eccentricity ('stretch'), axial tilt and precession ('wobble') with dominant periodicities of roughly 100, 40 and 20 ka, respectively [48]. These produced small variations in the amount and seasonal distribution of sunlight reaching the Earth that, when amplified by various feedback mechanisms – notably via variations in atmospheric greenhouse gas concentrations – led to the well-established pattern of Quaternary glacial/interglacial and stadial/interstadial changes. This mechanism was famously championed in the early twentieth century by Milutin Milankovitch [64], fell out of favour because the timing of individual glaciations as deduced from the fragmentary terrestrial record did not seem to fit, and then was triumphantly vindicated by analysis using oxygen isotopes from fossil foraminifera, that reflected temporal variations in ambient temperature and ice volume of the more complete ocean record [65,66].

The exploitation of Milankovitch cycles has subsequently developed in various directions. It has become a stratigraphic tool for dating and correlation, not only in the Quaternary, but in Tertiary and yet older strata [67], where a longer, 400 ka, orbital 'stretch' cycle is used as a more or less invariant 'pulse' that can be exploited stratigraphically and even quasi-formalised [68]. This in turn has led to the realisation that climate in greenhouse as well as icehouse times was modulated by astronomical forcing, with variations in humidity/aridity and biological productivity producing patterns that, although more subtle than those produced by large ice volume changes, are nonetheless recognisable.

Also, the detailed expression of Milankovitch cycles has come under scrutiny. Astronomical calculation can precisely reveal insolation variations and hence predict the climate patterns that should result. The observed patterns from the stratal record depart from this in several ways. Firstly, they typically show a 'sawtooth' pattern rather than the predicted temporally symmetrical one: thus, individual glacial phases tend to develop slowly but finish abruptly. Secondly, the periodicity that is expected to be dominant is not always so, as will be seen below. Thirdly, and particularly in cold phases, there are marked, higher-frequency 'sub-Milankovitch' climate cycles that have been well-described (also see below) but have not yet had adequate explanation.

9. MILANKOVITCH CYCLICITY IN QUATERNARY (PLEISTOCENE) CLIMATE HISTORY

The Quaternary displays a marked progression of overall climate state that may be regarded as an intensification of the glacial signature through time. The early Quaternary is dominated by the 40 ka axial tilt signal. About a million years ago, this gave way to dominance by the 100 ka orbital eccentricity cycle that has persisted to the present (Fig. 1d). This dominance has yet to be explained satisfactorily, for it would not be predicted from consideration of calculated insolation patterns over this interval, in which the eccentricity signal should be small. Suggested explanations have included the evolution of the ice-sheet/substrate system to resonate (i.e. most easily grow and decay) to a 100 ka periodicity [69,70]; these explanations are tentative, for detailed models linking ice volume to insolation remain elusive [71]. The dominance by eccentricity has been accompanied by colder glacial maxima and warmer interglacial peaks, and it is this interval that has seen the greatest advances of ice, and in general represents the 'ice ages' of vernacular usage.

The past million years includes a detailed record of atmospheric composition as well as temperature, in the form of the ice core data extracted from Greenland and Antarctica (with some ice core data of shorter duration from mountain glacier ice elsewhere) [72]. The longest current record is from Antarctica, extending to ca. 800 ka [73,74] and planned drilling is aimed at extending the record to beyond a million years ago, and so into the 'forty kiloyear world'. The Greenland record goes back to little more than 130 ka, and so just into the last interglacial phase; but it is of high-resolution, because of a greater rate of snowfall, and is of great value in also allowing detailed comparison with the southern hemisphere, given the different climate behaviour of the hemispheres at short time scales (discussed below).

The combination of atmospheric composition records with climate proxy records (through hydrogen and oxygen isotope data, dust concentrations and so on) is extremely powerful (indeed, unique in the geological record); but, it is not precisely calibrated because ice data directly relates to deposition, while the gas data relates to the time of final closure of air bubbles in the ice, some distance down in the snow pack. The uncertainty that stems from this is small but important, because the correlation of carbon dioxide and methane levels with temperature is so close that questions of cause-and-effect have arisen. The consensus now is that astronomically driven insolation thresholds lead to small temperature rises, leading to carbon dioxide/methane increases that then strongly amplify the temperature rises [72].

The glacial-to-interglacial difference seen in the ice core records is about 100 ppm (from ca. 180 to ca. 280 ppm $p\mathrm{CO_2}$, respectively), representing several hundred gigatonnes of carbon that must be stored somewhere during glacial phases. Terrestrial storage via increased plant growth is unlikely, given the diminution of vegetated land during glacials, though storage in carbon-rich permafrost soils (‘yedoma’) has been mooted [75]. Ocean storage is generally considered more likely, and it is tempting to link this with the enhanced dust supply noted in the ice core records, that would fertilise open ocean waters and enhance carbon drawdown via increased plankton growth. However, ocean sediment records of barium (a proxy for plankton productivity) do not generally show increases during glacial phases. One means of combining low plankton productivity and increased trapping of carbon dioxide is to have a more stratified glacial ocean, limiting nutrient supply from below because of a stronger surface water ‘lid’ and also storing more dissolved carbon dioxide at depth [76]. There is evidence for such a model in the form of glacial-phase benthic foraminifer tests containing excess ‘old’ (i.e. radiocarbon-poor) carbon [77].

10. QUATERNARY SUB-MILANKOVITCH CYCLICITY

Examination of high-resolution Quaternary records suggests significant climate variability that takes place on a sub-Milankovitch scale, a variability that is particularly marked in the cold phases that make up the bulk of the record (Fig. 1e). Thus, the cold phase that separates the present interglacial and the preceding (Eemian) one comprises not only five precession cycles, but also 26 well-marked temperature oscillations, termed Dansgaard–Oeschger (D–O) cycles. These are most clearly expressed in the northern hemisphere, where they comprise rapid warming (of 8–16 °C over Greenland) followed by slower cooling [78], to produce what are essentially a succession of interstadial and stadial units that average some 1470 a in duration [79]. The D–O cycles may be grouped into larger Bond cycles, terminated by intermittent (every several 1000 years) Heinrich events [80]: iceberg ‘armadas’ released from the Laurentide and Scandinavian ice sheets marking episodes of partial collapse (Fig. 1e). The Heinrich events led to distinctive gravel-rich layers within sea floor sediments (brought in from melting icebergs), metre-scale rises in global sea level and rapid northern hemisphere cooling. The D–O cycles have one-to-one counterparts in the southern hemisphere, but more muted ones (about 1–3 °C in Antarctica) that are in partial antiphase (Fig. 1f), being offset from the northern D–O events by about 90° (northern cold coinciding with southern warming) rather than in ‘see-saw’ fashion [81,82]. The causal mechanisms of the D–O cycles and related phenomena remain unclear, having been ascribed to changes in solar luminosity [83] and also to ‘binge-purge’ cycles of the great ice sheets [84].

The transition into the current Holocene interglacial was complex: thus, glacial conditions in the northern hemisphere were terminated at ca. 14.5 ka, with rapid deglaciation ushering in the millennial-scale Allerød warm phase, itself terminated by rapid cooling into the Younger Dryas cold interval, also lasting about a thousand years. This finished abruptly at 11.8 ka, when temperatures in the northern hemisphere rose by ca. 5 °C in about a decade, ushering in the warm and relatively stable conditions of the Holocene.

The reversal into the Younger Dryas has been ascribed to a major meltwater flood from the Laurentide ice-sheet into the north Atlantic, putting a low-salinity 'lid' on the north Atlantic, hence stopping the formation of the cold dense (high-salinity) North Atlantic Deep Water current and its ultimate return flow, the north Atlantic Drift ('Gulf Stream'); eventual re-start of this oceanic circulation pattern brought warmth once more back to the region. As with the D–O cycles, correlation with the southern hemisphere was complex, partly out-of-phase, and it is debated whether the climate changes were driven from the north or the south [85,86].

11. THE HOLOCENE

The Holocene is simply the latest of the many interglacial phases of the Quaternary; it is now longer than the preceding three interglacials by some 2000–3000 a [73], but only one-third of the length of the preceding one, OIS 11 [87], that lasted one-and-a-half, rather than half a precession cycle; it is still unclear to which style of interglacial the Holocene 'naturally' belongs to on astronomical grounds (and thus what its 'natural' duration might be). Its duration to date has also been linked with the slow rise in atmospheric carbon dioxide levels from 260 to 280 ppm, ascribed (controversially) to pre-industrial forest clearance by humans [88].

To date, though, other than a brief northern-hemisphere cooling event at 8.2 ka (also ascribed to a meltwater pulse from the decaying ice-sheets: [89], the Holocene has seen remarkable stability of temperature and sea level, even when compared with other interglacials. Climate variation within it includes subdued, millennial-scale temperature oscillations of 1 °C or so, examples being the 'Medieval warm period' and succeeding 'Little Ice Age', with sea level variations of perhaps 1 or 1–2 m [90]. As with the D–O cycles, their cause is obscure. Other shorter-period variations include the ENSO/El Nino events and the North Atlantic Oscillation (NAO) (of a few years periodicity each); as with the millennial scale variations, these have far-reaching global impacts on such factors as regional rainfall patterns via a series of global teleconnections [91].

12. CLIMATE OF THE ANTHROPOCENE

About two centuries ago, human population rose above a billion (it is now over six billion). Widespread industrialisation, powered by fossil fuels, also started then and continues to this day – indeed, is currently accelerating [92,93]. The sum total of physical, chemical and biological changes associated with this has led to the concept of the Anthropocene, a geological interval dominated by human activity [94]; if considered as a formal stratigraphic unit at an Epoch level [95], it follows that the Holocene has terminated.

Climate drivers of the Anthropocene are already well outside Holocene norms, for instance in: the marked increase in greenhouse gas levels (now higher than in pre-industrial times by the amount separating glacial and interglacial phases of the past, the change being considerably more rapid than either glacial-to-interglacial changes [96] or those associated with, say, the Toarcian event [46]; the changing nature of carbon sinks associated with land-use changes; and, as we write, the diminishing albedo associated with rapidly waning Arctic sea ice. The current greenhouse warming is acting on an already warm phase, and hence is bringing in a novel environmental state. The long-term effect, if median predictions of the Inter-Governmental Panel on Climate Change [97] come to pass, may resemble the brief 'super-interglacial' suggested by Broecker [98], the normal Quaternary Milankovitch cyclical climate changes subsequently resuming. Alternatively, modelled changes to the long-term carbon balance, together with threshold effects, suggest perturbation to at least several glacial cycles [99]. In whichever scenario (but particularly in the latter), the effects of the current warming will have geologically long-lasting effects.

13. CONCLUSIONS

The history of Earth's climate system, as deduced from forensic examination of strata, has shown a general very long-term stability, which has probably been maintained by a complex interaction between the biosphere, atmosphere, hydrosphere, cryosphere and lithosphere. Superimposed on this overall stability has been a variety of climate perturbations on timescales ranging from multi-million year to sub-decadal, inferred to have been driven, amongst others, by variations in palaeogeography, greenhouse gas concentrations, astronomically forced insolation and inter-regional heat transport. Current anthropogenic changes to the Earth system, particularly as regards changes to the carbon cycle, are geologically significant. Their effects may likely include the onset of climate conditions of broadly pre-Quaternary style such as those of the 'mid-Pliocene warm period', with higher temperatures (particularly at high latitudes), substantially reduced polar ice cover, and modified precipitation and biotic patterns.

ACKNOWLEDGEMENT

We thank Philip Gibbard, Alan Haywood and Euan Nisbet for their comments on earlier drafts of this paper.

REFERENCES

1. A.L.A. Johnson, J.A. Hickson, J. Swan, M.R. Brown, T.H.E. Heaton, S. Chenery, P.S. Balson, in: E.M. Harper, J.D. Taylor, J.A. Crame (Eds.), The Evolutionary Biology of the Bivalvia, vol.177, Geological Society, London, 2000, pp. 425–439, Special Publications.

2. D.S. Jones, W.D. Allmon, Lethaia 28 (1995) 61–74.
3. B. Jenny, B.L. Valero-Garcés, R. Villa-Martínez, R. Urrutia, M. Geyh, H. Veit, Quat. Res. 58 (2002) 160–170.
4. C.M. Rowe, D.B. Loope, R.J. Oglesby, R. Van der Voo, C.E. Broadwater, Science 318 (2007) 1284.
5. K.T. Lawrence, T.D. Herbert, P.S. Dekens, A.C. Ravelo, in: M. Williams, A. Haywood, F.J. Gregory, D.N. Schmidt (Eds.), The Micropalaeontological Society, The Geological Society, London, 2007, pp. 539–562, Special Publication.
6. F. Kenig, J.D. Hudson, J.S. Sinninghe Damsté, B.N. Popp, Geology 32 (2004) 421–424.
7. A.P.M. Vaughan, in: M. Williams, A. Haywood, F.J. Gregory, D.N. Schmidt (Eds.), The Micropalaeontological Society, The Geological Society, London, 2007, pp. 5–59, Special Publication.
8. E. Brook, Nature 453 (2008) 291–2.
9. J.C. McElwain, W.G. Chaloner, Ann.Bot. 76 (1995) 389–395.
10. W.M. Kürschner, J. van der Burgh, H. Visscher, D.L. Dilcher, Mar. Micropaleontol. 27(1996) 299–331.
11. G.J. Retallack, Science 276 (1997) 583–585.
12. P. Pearson, M. Palmer, Nature 406 (2000) 695–699.
13. R.A. Berner, Geochimica et Cosmochimica Acta 70 (2006) 5653–5664.
14. D.L. Royer, R.A. Berner, J. Park, Nature 446 (2007) 530–32.
15. L.E. Sohl, M.A. Chandler, in: M. Williams, A. Haywood, F.J. Gregory, D.N. Schmidt (Eds.), The Micropalaeontological Society, The Geological Society, London, 2007, pp. 61–80, Special Publication.
16. B. Sellwood, P. Valdes, in: M. Williams, A. Haywood, F.J. Gregory, D.N. Schmidt (Eds.), The Micropalaeontological Society, London, 2007, pp. 201–224, Special Publication.
17. H.J. Dowsett, in: M. Williams, A. Haywood, F.J. Gregory, D.N. Schmidt (Eds.), The Micropalaeontological Society, The Geological Society, London, 2007, pp. 459–480, Special Publication.
18. U. Salzmann, A. Haywood, D. Lunt, P. Valdes, D. Hill, Global Ecol. Biogeogr. 17 (2008) 432–447.
19. E.E. Jansen, J. Overpeck, K.R. Briffa, J.-C. Duplessy, F. Joos, V. Masson-Delmotte, D. Olago, B. Otto-Bliesner, W.R. Peltier, S. Rahmstorf, R. Ramesh, D. Raynaud, D. Rind, O. Solomina, R. Villalba, D. Zhang, in: S. Solomon, D. Qin, M. Manning, Z. Chen, M. Marquis, K.B. Averyt, M. Tignor, H.L. Miller (Eds.), Cambridge University Press, Cambridge, UK and New York, NY, USA, 2007.
20. S.E. Wilde, J.W. Valley, W.H. Peck, C.M. Graham, Nature 409 (2001) 175–178.
21. J. Baker, M. Bizzarro, N. Wittig, J. Connelly, H. Haack, Nature 436 (2005) 1127–1131.
22. E.G. Nisbet, N.H. Sleep, Nature 409 (2001) 1083–1091.
23. B. Rasmussen, R. Buick, Geology 27 (1999) 115–118.
24. J.E. Lovelock, Nature 344 (1990) 100–102.
25. M.J. Bickle, Terra Nova 8 (1996) 270–276.
26. A.S. Cohen, A.L. Coe, S.M. Harding, L. Schwark, Geology 32 (2004) 157–160.
27. F. Robert, M. Chaussidon, Nature 443 (2006) 969–972.
28. L.P. Knauth, Palaeogeogr. Palaeoclimatol. Palaeoceanogr. 219 (2005) 53–69.
29. H. Ohmototo, Y. Watanabe, K. Kumazawa, Nature 429 (2004) 395–399.
30. T.W. Lyons, Nature 429 (2004) 359–360.
31. I.A. Hilburn, J.L. Kirschvink, E. Tajika, R. Tada, Y. Hamano, S. Yamamoto, Earth Planet. Sci. Lett. 409 (2005) 175–178.
32. I.J. Fairchild, M.J. Kennedy, J. Geol. Soc. Lond. 164 (2007) 895–921.
33. M.I. Budyko, Tellus 21 (1969) 611–619.
34. P.F. Hoffman, A.J. Kaufman, G.P. Halverson, D.P. Schrag, Science 281 (1998) 1342.
35. M. Kennedy, D. Mrofka, C. von der Borch, Nature 453 (2008) 642–645.

36. A.A. Page, J.A. Zalasiewicz, M. Williams, L.E. Popov, in: M. Williams, A. Haywood, F.J. Gregory, D.N. Schmidt (Eds.), The Micropalaeontological Society, The Geological Society, London, 2007, pp. 123–156, Special Publication.
37. P.J. Brenchley, J.D. Marshall, G.A.F. Carden, D.B.R. Robertson, D.G.F. Long, T. Meidla, L. Hints, T.F. Anderson, Geology 22 (1994) 295–298.
38. R.M. DeConto, D. Pollard, 2003, Nature 421 (2007) 245–249.
39. T.D. Frank, L.P. Birgenheier, I.P. Montañez, C.R. Fielding, M.C. Rygel, in: C.R. Fielding, T.D. Frank, J.L. Isbell (Eds.), Geological Society of America Special, 2008, pp. 331–342, Paper 441.
40. A.G. Smith, K.T. Pickering, J. Geol. Soc., Lond. 160 (2003) 337–340.
41. J. Zachos, U. Röhl, S.A. Schellenberg, A. Sluijs, D.A. Hodell, D.C. Kelly, E. Thomas, M. Nicolo, I. Raffi, L.J. Lourens, H. McCarren, D. Kroon, Science 308 (2005) 1611–1615.
42. A.S. Cohen, A.L. Coe, D.B. Kemp, J. Geol. Soc., Lond. 164 (2007) 1093–1108.
43. J.C. Zachos, G.R. Dickens, R.E. Zeebe, Nature 451 (2008) 279–283.
44. D.J. Thomas, J.C. Zachos, T.J. Bralower, E. Thomas, S. Bohaty, Geology 30 (2002) 1067–1070.
45. J.C. McElwain, J. Wade-Murphy, S.P. Hesselbo, Nature 435 (2005) 479–482.
46. D.B. Kemp, A.L. Coe, A.S. Cohen, L. Schwark, Nature 437 (2005) 396–399.
47. M.E. Smith, A.R. Carroll, E.R. Mueller, Nat. Geosci. 1 (2008), 370–374.
48. J. Zachos, M. Pagani, L. Sloan, E. Thomas, K. Billups, Science 292 (2001) 686–693.
49. C. Lear, in: M. Williams, A. Haywood, F.J. Gregory, D.N. Schmidt (Eds.), The Micropalaeontological Society, The Geological Society, London, 2007, pp. 313–322, Special Publication.
50. A. Merico, T. Tyrrell, P.A. Wilson, Nature 452 (2008) 979–982.
51. J.P. Kennett, J. Geophys. Res. 82 (1977) 3843–3859.
52. G.J. Retallack, Palaeogeogr. Palaeoclimatol. Palaeoecol. 214 (2004) 97–123.
53. A.R. Lewis, D.R. Marchant, A.C. Ashworth, L. Hedenäs, S.R. Hemming, J.V. Johnson, M.J. Leng, M.L. Machlus, A.E. Newton, J.I. Raine, J.K. Willenbring, M. Williams, A.P. Wolfe, PNAS 105.
54. B.P. Flower, J.P. Kennett, Geology 21 (1993) 877–880.
55. A.E. Shevenell, J.P. Kennett, D.W. Lea, Science 305 (2004) 1766–1770.
56. A.M. Haywood, P. Dekens, A.C. Ravelo, M. Williams, Geochem. Geophys. Geosyst. 6 (2005) 1–20.
57. L. Liesecki, M.E. Raymo, Paleoceanography 20 (2005), 17, PA1003.
58. A.M. Haywood, H.J. Dowsett, P.J. Valdes (Eds.), Phil. Trans. R. Soc. A 367 (2009), 3–204.
59. P.L. Gibbard, A.G. Smith, J.A. Zalasiewicz, T.L. Barry, D. Cantrill, A.L. Coe, J.C.W. Cope, A.S. Galp, F.J. Gregory, J.H. Powell, P.F. Rawson, P. Stone, C.N. Waters, Boreas 34 (2005) 1–6.
60. M. Mudelsee, M.E. Raymo, Paleoceanography 20 (2005) PA-4022.
61. D.J. Lunt, G.L. Foster, A.M. Haywood, E.J. Stone, Nature 454 (2008) 1102–1105.
62. J.M. Edmond, Science 258 (1992) 1594–1597.
63. G.H. Haug, A. Ganopolski, D.M. Sigman, A. Rosell-Mele, G.E.A. Swann, R. Tiedemann, S.L. Jaccard, J. Bollman, M.A. Maslin, M.J. Leng, G. Eglinton, Nature 433 (2005) 821–825.
64. M. Milankovitch, Kanon der Erdbestrahlungen und seine Anwendung auf das Eiszeitenproblem, R. Serbian Academy Belgrade, 1941 Special Publication 132.
65. N. Shackleton, Nature 215 (1967) 15–17.
66. N.J. Shackleton, N.D. Opdyke, Quat. Res. 3 (1973) 39–55.
67. A. Gale, J. Hardenbol, B. Hathway, W.J. Kennedy, J.R. Young, V. Phansalkar, Geology 30 (2002) 291–294.
68. B.S. Wade, H. Pälike, Paleoceanography 19 (2004) PA001042.
69. P.U. Clark, R.B. Alley, D. Pollard, Science 286 (1999) 1104–1111.
70. R. Bintanja, R.S. van der Wal, Nature 454 (2008) 869–872.
71. M. Raymo, P. Huybers, Nature 451 (2008) 284–285.
72. E.W. Wolff, Episodes 31 (2008) 219–221.

73. J.R. Petit, J. Jouzel, D. Raynaud, N.I. Barkov, J.M. Barnola, I. Basile, M. Bender, J. Chapellaz, M. Davis, G. Delaygue, M. Delmotte, V.M. Kotlyakov, M. Legrand, V.Y. Lipenkov, C. Lorius, L. Pépin, C. Ritz, E. Salzmann, M. Stievenard, Nature 399 (1999) 429–436.
74. E. Brook, Nature 453 (2008) 291–2.
75. S.A. Zimov, E.A.G. Schuur, F.C. Chapin III, Science 312 1612–1613.
76. S.L. Jaccard, G.H. Haug, D.M. Sigman, T.F. Pedersen, H.R. Thierstein, U. Röhl, Science 308 (2005) 1003–1006.
77. T.M. Marchitto, S.J. Lehman, J.D. Ortiz, J. Flüchtiger, A. van Gen, Science 316 (2007), 1456–1459.
78. C. Huber, M. Leuenberger, R. Spahni, J. Flückiger, J. Schwander, T.F. Stocker, S. Johnsen, A. Landais, J. Jouzel, Earth Planet. Sci. Lett. 245 (2006) 504–519.
79. G. Bond, W. Showers, M. Cheseby, R. Lotti, P. Almosi, P. deMenocal, P. Priore, H. Cullen, I. Hajdas, G. Bonani, Science 278 (1997) 1257–1266.
80. G. Bond, H. Heinrich, W. Broecker, L. Labeyrie, J. Mcmanus, J. Andrews, S. Huon, R. Jantschik, S. Clasen, C. Simet, K. Tedesco, M. Klas, G. Bonani, S. Ivy, Nature 360 (1992) 245–249.
81. EPICA community members, Nature 444 (2006) 195–198.
82. E.J. Steig, Nature 444 (2006) 152–153.
83. H. Braun, M. Christl, S. Rahmnstorf, A. Ganopolski, A. Mangini, C. Kubatzki, K. Roth, B. Kromer. Nature 438 (2005) 208–211.
84. D.R. MacAyeal, Paleoceanography 8 (1993) 775–784.
85. T.F. Stocker, Science 297 (2002) 1814–1815.
86. A.J. Weaver, O.A. Saenko, P.U. Clark, J.X. Mitrovica, Science 299 (2003) 1709–1713.
87. EPICA community members. Nature 429 (2004) 623–628.
88. W.F. Ruddiman, Clim. Change 61 (2003) 261–293.
89. J.D. Marshall, B. Lang, S.F. Crowley, G.P. Weedon, P. van Calsteren, E.H. Fisher, R. Holme, J.A. Holmes, R.T. Jones, A. Bedford, S.J. Brooks, J. Bloemendal, K. Kiriakovlakis, J.D. Ball. Geology 207 (2007) 639–642.
90. K-E. Behre, Boreas 36 (2007) 82–102.
91. T.M. Rittenour, J. Brigham-Grett, M.E. Mann, Science 288 (2000) 1039–1042.
92. W. Steffen, A. Sanderson, P.D. Tyson, J. Jäger, P.A. Matson, B. Moore III, F. Oldfield, K. Richardson, H.J. Schnellhuber, B.L. Turner II, R.J. Wasson, Global Change and the Earth System: A Planet under Pressure, Springer-Verlag, Berlin, 2004.
93. W. Steffen, P.J. Crutzen, J.R. McNeill, Ambio 36 (2008) 614–621.
94. P.J. Crutzen, Nature 415 (2002) 23.
95. J. Zalasiewicz, M. Williams, A. Smith, T.L. Barry, A.L. Coe, P.R. Brown, P. Brenchley, D. Cantrill, A. Gale, P. Gibbard, F.J. Gregory, M.W. Hounslow, A.C. Kerr, P. Pearson, R. Knox, J. Powell, C. Waters, J. Marshall, M. Oates, P. Rawson, P. Stone, GSA Today 18 (2) (2008) 4–8.
96. E. Mannin, A. Indermühle, A. Dällenbach, J. Flückiger, B. Stauffer, T.S. Stocker, D. Reynaud, J.M. Barnola, 2001, Science 291 (2001) 112–114.
97. IPCC (Intergovernmental Panel on Climate Change, Climate change 2007: synthesis report. (2007) Summary for policy makers. Available at: http://www.ipcc.ch/SPM2feb07.pdf.
98. W.S. Broecker, How to Build a Habitable Planet, Eldigio Press, NewYork, 1987, pp. 291.
99. T. Tyrrell, J.G. Shepherd, S. Castle, Tellus B 59 (2007) 664–672.
100. S. Rahmstorf, Nature 419 (2002) 207–214.