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Climate modelling and deep-time climate change

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Abstract

Detailed and reliable understanding of past climate change is a key ingredient in unravelling how climate has influenced life on earth and will continue to do so in the future. Palaeoclimatology and climate modelling have both made rapid strides over the past decades, and there has been fruitful two-way interaction between the two fields. The application of climate models to palaeoclimates has proved useful both in interpreting palaeoclimate proxy data and in testing the robustness and generality of climate models. Here, we give an overview of the current state of climate modelling and review recent progress in understanding deep-time climate change, with emphasis on problems where climate models have played a salient role. By suitably adjusting the concentration of atmospheric greenhouse gases, climate models can be made to replicate many key climatic transitions in the earth's history. However, important discrepancies remain between modelled climates and proxy reconstructions, particularly on the warm end of the spectrum.

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2.1 Introduction

Climate science deals with reconstructing and explaining the long-term mean and variability of physical conditions in the earth's envelope. A striking feature emerging from such analysis is the vast range of timescales on which there is significant variability. Part of this variability, including the diurnal and annual cycles, is periodic and predictable, but mostly it is random and unpredictable. We know from direct experience that the weather changes from hour to hour, from day to day, and from year to year. Analysis of long-term instrumental records and of palaeoclimatological proxy data shows that the same is true on longer timescales; conditions change randomly from century to century, millennium to millennium, and from aeon to aeon. Figure 2.1 shows a composite spectrum of surface temperature and temperature proxies spanning timescales from a few days up to about 1 million years (Huybers and Curry, 2006). There are sharp peaks at the annual frequency and subannual harmonics, as expected. There are also some more poorly defined peaks at the 41 000-year and 100 000-year orbital variability frequencies (see Wunsch, 2003, 2004, for a discussion of the statistical significance of these peaks). However, most of the spectrum shows very little structure, and is best approximated by straight lines of constant slope. Thus, at least on timescales up to around 1 million years, climate variability is mostly red noise: fully stochastic, unpredictable variability skewed towards low frequencies.

Time series whose spectra are red up to the lowest frequencies, such as those shown in Fig 2.1, are said to possess 'long memory' or 'long-range correlation' (Beran, 1994). A feature of long-memory processes is that any time-series segment of finite length will typically show a trend, because the segment can be thought of as being 'embedded' in a longer-term fluctuation. Figures 2.2 and 2.3 show temperature reconstructions over the Cenozoic (up to 65 million years ago - mya - Zachos et al., 2001) and the Phanerozoic (up to 545 mya - Royer et al., 2004). Interestingly, both these time series show clear trends (for the Phanerozoic time series in Fig 2.3, this is only the case if the oxygen isotope data are appropriately corrected for pH effects - Royer et al., 2004). Although this trend by itself is not enough to infer long memory, the important implication is that the climate system appears never to have been in a steady state, even over time spans comparable to the age of the earth. Life has coevolved with climate over this time, and much remains to be understood about the way in which life and climate have mutually conditioned each other's evolution. This chapter will focus on climatic aspects, providing a backdrop for the more biologically oriented chapters in the rest of the book.

Mitchell (1976) and Kutzbach (1976) provide classic accounts of the mechanisms underlying long-term climate variability. The climate system contains two fluid components, the atmosphere and ocean, which are strongly heated in the tropics and more weakly heated in the high latitudes. This heating gradient

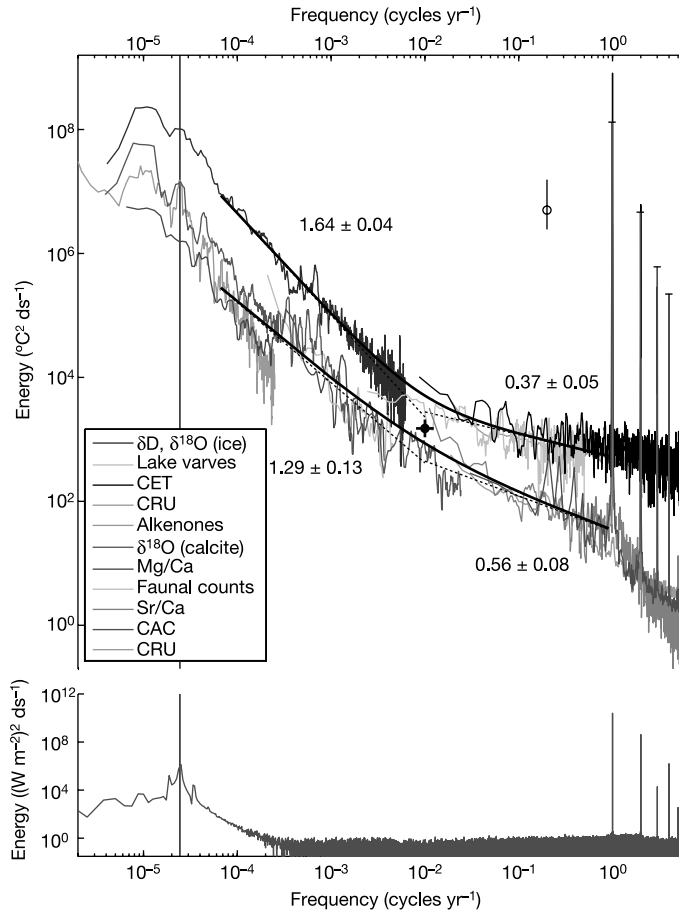


Figure 2.1 Patchwork spectral estimate using instrumental and proxy records of surface temperature variability. The more energetic spectral estimate is from high-latitude continental records and the less energetic estimate from tropical sea-surface temperatures. Power-law estimates for 1.1–100-year and 100–15 000-year periods are listed along with standard errors, and are indicated by the dashed lines. The sum of the power laws fitted to the long and short period continuum is indicated by the black curve. The vertical line segment indicates the approximate 95% confidence interval, where the circle indicates the background level. The mark at 1/100 yr indicates the region midway between the annual and Milankovitch periods. At the bottom is the spectrum of insolation at 65° N sampled monthly over the past million years plus a small amount of white noise. The vertical black line indicates the 41-kyr obliquity period. Reproduced with permission from Huybers and Curry (2006). *See colour plate section.*

triggers internal instabilities leading to ceaseless, chaotic motion in the ocean-atmosphere system with timescales from days to a few thousand years and spatial scales from hundreds to many thousands of kilometres (the fact that heating is

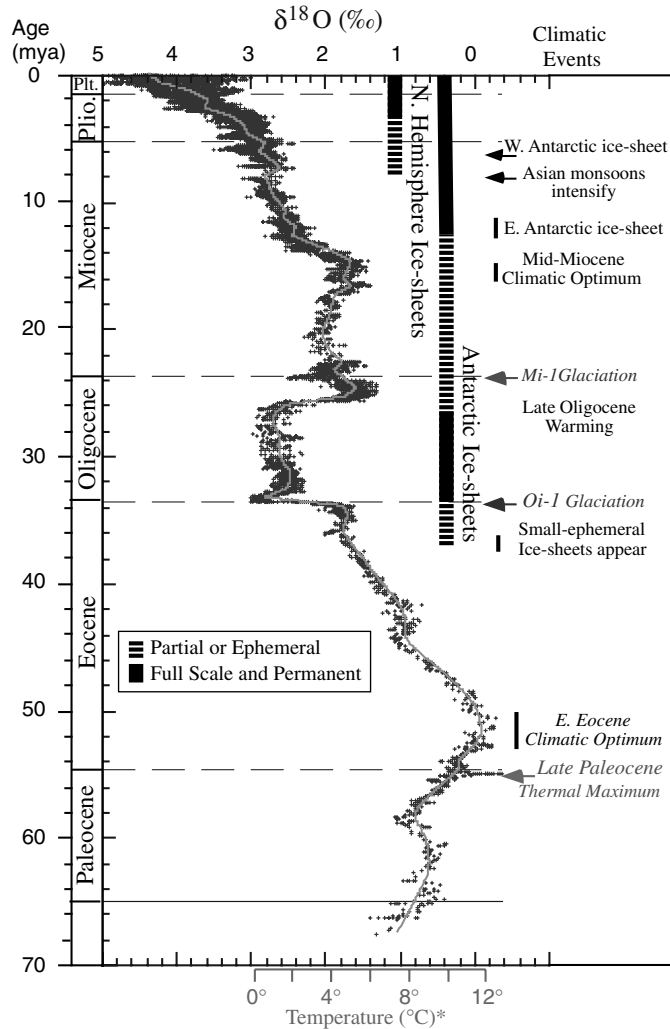


Figure 2.2 Global deep-sea oxygen isotope records for the Cenozoic era, based on data compiled from more than 40 ocean drilling sites. The raw data were smoothed using a five-point running mean and curve fitted with a locally weighted mean. The oxygen isotope temperature scale was computed for an ice-free ocean, and thus only applies to the time preceding the onset of large-scale glaciation on Antarctica (35 mya). From the early Oligocene to the present, much of the variability in the oxygen isotope record reflects changes in Antarctica and northern hemisphere ice volume. The vertical bars provide a rough qualitative representation of ice volume in each hemisphere relative to the Last Glacial Maximum (21 000 years ago), with the dashed bar representing periods of minimal ice coverage (less than 50%), and the full bar representing close to maximum ice coverage (more than 50% of present). Some key tectonic and biotic events are listed as well. Adapted from Zachos et al. (2001). *See colour plate section.*

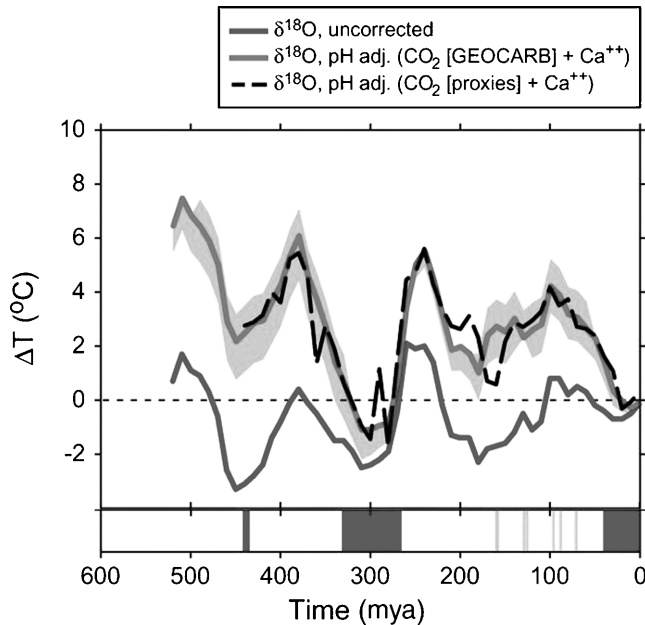


Figure 2.3 Shallow marine carbonate oxygen isotope record over the Phanerozoic eon. The blue curve corresponds to temperature deviations relative to today calculated by Shaviv and Veizer (2003). In the two remaining curves, data from the blue curve have been adjusted for pH effects due to changes in seawater calcium ion concentration and CO_2 based either on model (GEOCARB) or proxy reconstructions. Blue bands in the strip along the bottom indicate icehouse intervals, with extensive, permanent continental ice sheets (dark blue) or cool climates, with modest, ephemeral ice sheets (light blue). Adapted from Royer et al. (2004). See colour plate section.

stronger near the surface than aloft also triggers convective motions in the atmosphere with much shorter space and time scales). The atmosphere and ocean also interact with non-fluid components having much longer timescales. The build-up of large continental ice sheets requires in the order of 10^5 years. The long-term carbon cycle, which controls atmospheric carbon dioxide (CO_2) concentrations and thus the greenhouse effect and surface temperature, involves the slow weathering of silicate rocks (Walker et al., 1981), which sets a timescale in the order of 10^5 years for the drawdown of atmospheric CO_2 . Tectonic processes, including the creation of large mountain ranges, the opening/closing of ocean seaways and the establishment of major igneous provinces, have timescales of 10^6 – 10^7 years. The formation and breakup of supercontinents, loosely speaking the Wilson cycle (Wilson, 1966), have timescales of 500 million years. At the very longest timescales, the sun's luminosity has been steadily increasing, by perhaps 40% over geologic time. The 'faint young sun' paradox – the problem of reconciling evidence for

liquid water early in the earth's history with the low insolation then prevalent, first pointed out by Sagan and Mullen (1972) – has yet to be fully resolved. Finally, the appearance and evolution of life on earth has also had a major impact on climate, the details of which have been only partially unravelled.

By comparison with the earth's previous history, the past 10 000 years (the Holocene) have been relatively quiescent, with no major swings in ice volume or sea level and relatively mild changes in global and regional temperatures. This is also the period in which agriculture and settled human society first arose. Human population and its economic activities have now reached a scale such that they may strongly impact climate at the regional and global scales. Since human society has evolved and is adapted to a particular set of climatic conditions, any change is likely to result in widespread disruption and hardship. Perhaps the most alarming possibility is of a major melting of the Greenland and/or Antarctic ice sheets. Total collapse of the Greenland ice sheet alone would cause sea-level rise of around 7 m (IPCC, 2007), potentially submerging many of the world's most populous cities and driving hundreds of millions of people into destitution.

The problem of understanding the climate system and predicting its evolution in the near future has gained much prominence in scientific and public debate. The study of palaeoclimates helps, amongst other things, to answer questions about the extent and speed of climate change. It also serves as a test bed for theories of climate. Quantitative assessment of future climate change relies very heavily on global climate models, which contain a detailed description of the physics governing climate. Application of climate models to palaeoclimates helps solve some of the puzzles posed by these climates, and provides a particularly stringent test of model robustness and reliability. In the following sections, we provide a brief review of climate modelling and of long-term climate change. We emphasise time periods where the interplay of models and palaeoclimate proxy data has been particularly illuminating; these fall mostly (but not entirely) in the Cenozoic, for which proxy data are most abundant.

2.2 Climate modelling

2.2.1 The physical basis of climate modelling

The earth's climate is governed by fundamental and well-understood principles of physics. In weather prediction and climate models, these principles are expressed as mathematical equations which are then solved by numerical means. Washington and Parkinson (2005) provide an excellent introduction to climate modelling. For an account of the evolution of numerical weather prediction models, see Lynch (2006).

The motions of the atmosphere and oceans evolve according to the laws of dynamics first established by Newton in the seventeenth century. These laws relate

changes in the motion to the forces applied to the system: if we can identify and specify the forces, we can use the laws of motion to deduce how the system will evolve. Thus, the dynamics can be described in precise terms. However, dynamics alone are insufficient. We also need to consider the effects of heating and cooling. These effects are described by the laws of thermodynamics and electromagnetic radiation, which were brought to light in the nineteenth century. Climate is driven by solar energy. Because of the spherical shape of the earth and the configuration of its orbit around the sun, the amount of incoming solar energy is highly variable, depending strongly on the time of day, the season and the latitude. The climate system can be viewed as an enormous thermodynamic engine, with the atmosphere and oceans playing the role of the working fluid, transporting heat from source to sink regions and performing work in the process.

In addition to dynamics and thermodynamics, there is the principle of conservation of mass, which provides a quantitative relationship called the continuity equation. The physical variables are linked through a relationship known as the equation of state. Finally, the physical principles of how matter and radiation interact, and how water changes between the solid, liquid and gaseous phases, complete the picture. We can specify the state of the ocean-atmosphere system at a particular time by providing the values of the physical variables everywhere throughout the system. Thus, if we give the pressure, temperature, density and (for the ocean) salinity, and the velocity in each direction and at every point, the configuration of the system is determined. Such data are called the initial conditions. Given these conditions, the physical principles may now be used to deduce the future evolution of the system.

Each physical law has a precise mathematical expression, and the assemblage of laws provides us with a closed set of equations. Mathematicians describe these as non-linear partial differential equations. They cannot be solved by analytical means; the solution in each case must be calculated by numerical approximations. The complete system of equations was first assembled around 1900 by scientists interested in forecasting weather by rational means. However, 50 years had to pass before the goal of numerical weather prediction began to become a reality.

In 1950 a drastically simplified mathematical model of the atmosphere was used to simulate the evolution of the weather over a one-day period (Charney et al., 1950). A few years later, Phillips carried out the first long-range simulation of the general circulation of the atmosphere. He used a two-level model with simplified equations and geometry, and with rudimentary physics. The computation used a spatial grid of 16×17 points, and the simulation was for a period of about one month. Idealised initial conditions were used and, during the simulated month, a state developed having some of the key features of the observed atmosphere. Starting from a zonal (westerly) flow with small random perturbations, a wave disturbance with wavelength of 6000 km developed. It had the

characteristic westward tilt with height found in baroclinic waves, the large-scale unstable waves found in the mid-latitude atmosphere, and the disturbance moved eastward at about 20 m s^{-1} , which is comparable to observed atmospheric waves. Phillips examined the energy exchanges of the developing wave and found good qualitative agreement with observations of wave systems in the atmosphere. He also examined the flow averaged around parallels of latitude, the mean meridional flow, and found circulations corresponding to the three atmospheric cells.

Phillips completed his experiment in 1955 and communicated the results to John von Neumann, who immediately recognised their significance and arranged a conference in Princeton in October 1955: Application of Numerical Integration Techniques to the Problem of the General Circulation (see Lewis, 1998). Following this conference, Phillips entered the research for the first Napier Shaw Memorial Prize. On 20 June 1956, the adjudicators recommended that the prize be given to Phillips for his essay ‘The general circulation of the atmosphere: a numerical experiment’, which had been published in the *Quarterly Journal of the Royal Meteorological Society* (Phillips, 1956).

Phillips’s work had a galvanising effect on the meteorological community. Within 10 years, there were several major research groups modelling the general circulation of the atmosphere, some of the leading ones being at the Geophysical Fluid Dynamics Laboratory in Princeton, New Jersey, USA, the National Center for Atmospheric Research in Boulder, Colorado, USA, the Meteorological Office’s Hadley Centre in Exeter, UK, and the Max Planck Institute for Climate Research in Hamburg, Germany.

2.2.2 Current global models

Following the seminal work of Phillips, the development of general circulation models (GCMs) advanced rapidly, hand in hand with increasing computer power. Standard current-generation GCMs include a detailed description of the dynamics and physics of both atmosphere and ocean (these are referred to as ‘coupled’ GCMs), and also incorporate sophisticated treatment of the hydrological cycle over land, and of the formation and dynamics of sea ice. For palaeoclimate applications, a standard GCM can be coupled to a dynamic ice-sheet model. Ice sheets evolve over timescales of 10^4 – 10^5 years, far in excess of what is feasible for a continuous GCM simulation given current computer capabilities, so in this case the coupling is performed asynchronously, i.e. in a sequence of quasi-equilibrium steps. Much current effort is being devoted to incorporating aspects of the carbon cycle, including interaction with land and ocean biota, to more refined atmospheric chemistry and aerosol schemes, and to incorporation of oxygen isotope fractionation.

As an example of a current climate model, we consider HadCM3, developed at the Hadley Centre. Many earlier coupled models needed a flux adjustment (additional artificial heat and moisture fluxes at the ocean surface) to produce good

simulations. The higher ocean resolution of HadCM3 was a major factor in removing this requirement. To test its stability, HadCM3 has been run for over 1000 years' simulated time and shows minimal drift in its surface climate. The atmospheric component of HadCM3 has 19 levels and a latitude/longitude resolution of 2.5×3.75 degrees, with grid of 96×73 points covering the globe. The resolution is about 417×278 km at the equator. The physical parameterisation package of the model is very sophisticated. The radiative effects of minor greenhouse gases as well as CO_2 , water vapour and ozone are explicitly represented. A parameterisation of background aerosol is included. The land surface scheme includes freezing and melting of soil moisture, surface runoff and soil drainage. The convective scheme includes explicit down draughts. Orographic and gravity wave drag are modelled. Cloud water is an explicit variable in the large-scale precipitation and cloud scheme. The atmospheric component of the model allows the emission, transport, oxidation and deposition of sulphur compounds to be simulated interactively. The oceanic component of HadCM3 has 20 levels with a horizontal resolution of 1.25×1.25 degrees, permitting important details in the oceanic current structure to be represented. The model is initialised directly from the observed ocean state at rest, with a suitable atmospheric and sea-ice state. HadCM3 is being used for a wide range of climate studies, which will form crucial inputs to the forthcoming Fifth Assessment Report (AR5) of the Intergovernmental Panel on Climate Change (IPCC), which will be finalised in 2014.

2.2.3 Downscaling and regional climate models

Global GCMs are run at coarse spatial resolution, typically using a 100 km grid. They are unable to resolve many important subgrid-scale features such as clouds and topography. As a result, GCMs are not suitable for regional impact studies. To address this problem, various downscaling methods are employed to study local-scale climate impacts. The outputs of the global models are used as inputs to the regional models. The values generated during the execution of the global model are saved, and are used to specify the values around the boundaries of the limited domain, that is, the lateral boundary conditions. Wilby and Wigley (1997) identified four categories of downscaling: regression methods, weather pattern-based approaches, stochastic weather generators, and limited-area modelling.

In the regression method, the fine details of the past climate are related to the coarse representation provided by the global model. Assuming the relationship holds under a different climate regime, the regional details of the future climate can be deduced from coarse-grain predictions generated by the global model. This statistical downscaling has been used with some success. However, the hypothesis that the relationship between coarse- and fine-grain structures remains unchanged under a changing climate is open to serious question. We therefore seek alternative approaches.

A regional climate model (RCM) uses the same physical principles and mathematical equations as global models, but covers only a limited geographical domain. This allows the use of a much finer computational grid. Whereas a global model might have points separated by 100 km, a regional model would typically have a 10 km grid. Thus, it is capable of resolving a large range of atmospheric phenomena that ‘fall between the gaps’ of a GCM. In particular, the representation of mountains is much better in the RCM. Since temperature and precipitation are so strongly coupled with elevation and slope, this is a major advantage.

The climate scientists at University College Dublin and Met Éireann have been collaborating for several years on a project called the Community Climate Change Consortium for Ireland (C₄I). The objective is to develop regional climate models and apply them to the question of Ireland’s future climate. The main results to date are summarised in the report *Ireland in a Warmer World* (McGrath and Lynch, 2008). The work has employed a range of regional models.

2.2.4 Uncertainty of the predictions

Climate models are the best means we have for predicting future changes in our climate. They have a solid scientific basis in the principles of physics and are increasing rapidly in sophistication and in the accuracy with which they can simulate the climate. However, the atmosphere is chaotic, that is, highly sensitive to very small changes, so that if the starting conditions are changed in a small way, the subsequent evolution will be completely different. Thus, no matter how good the models are, there will always be a degree of uncertainty in the predictions. There are many sources of uncertainty. Firstly, the initial state of the system is not known precisely; this is particularly the case for the deep ocean. Secondly, the external forcings cannot be known exactly; for example, the level of CO₂ 20 or 40 years hence depends on population growth and on energy use, which may be strongly influenced by technological developments. Thirdly, the models themselves have imperfections, and many approximations must be made in constructing practical models. Fourthly, the physical processes in the atmosphere and ocean are complex and not completely understood. Finally, there are many feedbacks in the climate system that make it very sensitive to small changes.

Among the processes that can act as positive feedbacks, we may consider the level of water vapour, the ice albedo effect, cloud amount, carbon uptake by forest, and emission of methane from melting tundra. An increase of cloudiness may have either a cooling or a warming effect on climate, depending on the cloud height and structure. In general, low clouds act to cool the climate and high clouds have a warming effect. But in reality things are much more complicated. For climate predictions over the next century, cloud feedbacks remain the greatest source of uncertainty (e.g. Soden and Held, 2006).

2.2.5 Probabilistic prediction: the ensemble approach

In view of the many uncertainties in predicting future climate, scientists now approach the problem from a stochastic point of view. For example, instead of trying to predict the increase of temperature by 2100 in a deterministic manner, we aim to predict both the most probable change and also the probable range of changes that may be observed. In essence, we try to forecast the probability density function rather than just a single deterministic value. To do this, we need to assess the sensitivity of the predictions to various errors in the initial state, in the model itself and in the specified external forcings. Considering a range of values of initial conditions and other parameters, we perform a collection of numerical predictions, called an ensemble. Depending on the available computer power, the ensemble may contain anything from a few predictions to as many as a hundred.

Using the ensemble outputs, we can examine how the predictions differ, and the extent to which they spread out from each other. If they form a compact set of closely related values, we can be confident that the average of these values is a reliable guide to what may happen. If, however, they are widely dispersed, we may have relatively little confidence in the prediction. The IPCC predictions are generally based on ensembles of model runs, and are given in a probabilistic form. For example, the mean temperature increase under the assumptions of a particular emission scenario (A1b) is given as a 'best estimate' of 2.8 K and a 'likely range' of between 1.7 K and 4.4 K.

2.3 Long-term climate change

2.3.1 Greenhouse and icehouse climates

A useful classification of the earth's climate types is into icehouse climates, which feature large-scale continental ice sheets and widespread perennial sea ice at high latitudes, and greenhouse climates, which do not. We are currently in an icehouse world, but the earth has switched between the two states several times during its history. Based mostly on the presence or absence of geological evidence for large-scale glaciation, Frakes et al. (1992) identify five icehouse periods separated by four greenhouse periods during the past 600 million years, each interval lasting 50–100 million years. However, not all the cool periods identified were of the same intensity, as indicated in Fig 2.3. There is abundant evidence for two full-blown glaciations, during the early Carboniferous to mid-Permian (340–250 mya) and the late Cenozoic (34 mya to present), while the late Jurassic to early Cretaceous (183–105 mya) and late Ordovician to early Silurian (458–421 mya) cool periods may have seen only ephemeral ice sheets and seasonal sea ice.

Understanding how earth shifts between greenhouse and icehouse is a central question in palaeoclimatology. Changing levels of atmospheric greenhouse gas concentrations, especially CO₂, appear to play an important role in these transitions. CO₂ reconstructions over the Phanerozoic are subject to much uncertainty, but they generally show reduced CO₂ levels during the Carboniferous–Permian and the late Cenozoic, qualitatively matching the glaciation record (Royer et al., 2004; Royer, 2006). There is also considerable modelling evidence that continental glaciation is controlled by CO₂ levels. The association between climate conditions and CO₂ is somewhat problematic for the end-Ordovician glaciation (440 mya, see Fig 2.3), which occurred during a period of apparently high CO₂. The discrepancy may be due to the short duration of this glacial interval, which is possibly not well resolved by the CO₂ data. It should also be recalled that insolation at that time was around 3–4% lower than today.

The question of what ultimately caused the CO₂ fluctuations over geological time is less settled. Long-term atmospheric CO₂ levels are controlled by the balance between the rate of volcanic outgassing and the rate of drawdown due to chemical weathering of silicate rocks (Walker et al., 1981; Berner et al., 1983). Plate tectonics can affect both source and sink: increased seafloor spreading rates or the subduction of carbonate-rich sediments will lead to increased outgassing, while the breakup of a supercontinent, continental drift from a temperate to a tropical zone and mountain uplift will all increase weathering rates and lead to reduced CO₂ levels. As a specific example, Kent and Muttoni (2008) suggest that high CO₂ during the early Cenozoic was due to the closing of Tethys (see Rodríguez-Sánchez and Arroyo, Chapter 13), a relatively shallow tropical ocean with a deep carpet of carbonate sediments, whose subduction led to a period of increased outgassing. This period ended with the collision of India and Asia, which also brought the highly weatherable Deccan Traps into the equatorial humid belt, explaining the subsequent decline in CO₂ through the Eocene and Oligocene. The evolution of land plants over geological time also had a major impact on the long-term carbon cycle, notably through the effect of root systems on weathering rates (Berner, 2004) and potentially through a host of other feedback loops (Beerling and Berner, 2005).

An important related question concerns the role of climate change in driving mass extinction events, particularly the ‘big five’ extinction events at the end-Ordovician (446 mya), late Devonian (371 mya), Permian–Triassic (PT) boundary (251 mya), Triassic–Jurassic (TJ) boundary (200 mya) and Cretaceous–Tertiary (KT) boundary (65 mya) (Sepkoski, 1982). The PT, TJ and KT extinctions coincided with intense volcanism events, forming the Siberian Traps, Central Atlantic Magmatic Province and the Deccan Traps respectively, and the PT and TJ were also periods of rapidly rising CO₂ and temperature (McElwain et al., 1999; Benton and Twitchett, 2003). It is not clear, however, whether global warming by itself is sufficient to explain the mass extinctions (Mayhew et al., 2008), or whether other

'kill mechanisms' need to be invoked. It is widely accepted that the KT extinction was associated with a large bolide impact (Alvarez et al., 1980), but there is scant evidence for such impacts during the other extinctions. Global warming can lead to sluggish oceanic circulation and widespread anoxia, which could explain extinction in the oceanic realm (Kiehl and Shields, 2005). It has also been hypothesised that the emission of sulphur dioxide (SO₂) and other pollutants from flood basalts may have had a directly noxious effect on biota, especially plants (van de Schootbrugge et al., 2009).

2.3.2 Neoproterozoic snowball earth

In the most extreme case, the earth can exist in a stable equilibrium where the oceans are entirely covered in ice. Whether this snowball earth scenario has ever actually occurred remains contentious. A brief paper by Kirschvink (1992), noting evidence for low-latitude glaciation during the Neoproterozoic (*c.* 750 mya), suggested a snowball earth could have existed at that time. The idea was promoted powerfully by Hoffman et al. (1998), who provided more extensive evidence, as well as a coherent narrative, of how the earth could enter and exit such a state. The paper triggered a spurt of publications on the topic. While there is general consensus that the Neoproterozoic was an extreme icehouse period featuring low-latitude glaciation, the community is divided as to whether the planet was truly covered in ice, or whether the sea-ice line reached only to the subtropics, leaving a broad swath of open ocean in the tropics (so-called 'slushball earth'). The two hypotheses are qualitatively different as regards their impacts on life: a true snowball would have killed off most marine life, since no light would have reached the underlying ocean through the thick layer of ice, while in a slushball scenario life could have continued relatively unperturbed in the tropics.

Climate model simulations suggest that a transition to a snowball earth can indeed occur under Neoproterozoic conditions, though this is somewhat sensitive to the precise level of atmospheric CO₂ (e.g. Donnadieu et al., 2004). A greater difficulty with the snowball hypothesis concerns the exit mechanism. The basic idea is that, since the hydrological cycle comes to a halt during a snowball episode and the oceans are isolated from the atmosphere, there are no sinks for atmospheric CO₂, so given enough time subaerial volcanism will gradually build up large atmospheric CO₂ concentrations. Eventually, a threshold is crossed where tropical temperatures are high enough to melt the ice, at which point a catastrophic meltback ensues as the surface albedo feedback works in reverse (Caldeira and Kasting, 1992; Hoffman et al., 1998). However, attempts to reproduce this process using a detailed climate model (Pierrehumbert, 2004, 2005) show that the system remains far short of deglaciation even at extremely elevated CO₂, in the order of 20% of the total mass of the atmosphere. Thus, either the earth was never in a snowball state, or other as yet unknown mechanisms explain the transition out of that state.

2.3.3 The early Cenozoic greenhouse

The early Eocene climate optimum

The early Cenozoic was a time of extreme global warmth, culminating in the early Eocene climate optimum (EECO) around 50 mya. Oxygen isotope proxies indicate that benthic temperatures (which reflect high-latitude annual mean ocean surface conditions) could have been as high as 12 °C (Zachos et al., 2001). Direct reconstruction of Arctic ocean surface temperatures using the TEX86 proxy suggest values as high as 19 °C (Sluijs et al., 2006). High-latitude continental interiors were also much warmer than today (Greenwood and Wing, 1995). In the tropics, recent sea-surface temperature reconstructions using $\delta^{18}\text{O}$ in well-preserved foraminifera indicate temperatures in excess of 30 °C, independently confirmed by TEX86 estimates (Pearson et al., 2007). Overall, these reconstructions suggest a global mean temperature during the EECO of perhaps 25 °C, around 10 °C higher than today.

Simulating the EECO poses a major challenge to climate models. Atmospheric CO_2 concentration during the EECO is known to have been much higher than today, with estimates ranging from around 1000 to 4000 ppm (compared to 280 ppm in modern pre-industrial times). Setting model CO_2 to values within this range, it is possible to match the 10 °C increase in global mean temperature (Shellito et al., 2003), but it has proved impossible to capture the very weak equator-pole temperature gradient indicated by the proxy reconstructions. This 'low-gradient paradox' suggests that current climate models may be missing some key physical ingredient responsible for the reduction in meridional temperature gradient.

Two types of explanation are possible: either some unknown or poorly represented radiative feedback acts to limit temperatures in the tropics or boost them at the poles, or dynamical atmosphere/ocean poleward heat transport increases much more rapidly with temperature than predicted by the models. Clouds play a major role in most explanations of the first type, either cooling the tropics (Lindzen et al., 2001) or warming the high latitudes (Sloan and Pollard, 1998; Abbot and Tziperman, 2008). As for the second category, it appears that poleward atmospheric heat transport in models cannot increase above a certain limit, for dynamical reasons that remain unclear (Caballero and Langen, 2005; O'Gorman and Schneider, 2008). Model ocean transports also do not increase much in coupled Eocene simulations (Huber and Sloan, 2001). Recent work has focused on the role of tropical cyclones, which are unresolved in current climate models and are therefore an ideal candidate for a missing process. Tropical cyclone frequency and intensity are predicted to increase in a warmer world; by increasing vertical mixing in the ocean, this could lead to stronger poleward heat transport (Emanuel, 2001, 2002). However, direct implementation of a temperature-dependent ocean mixing rate in a climate model, designed to mimic

the effect of tropical cyclones, failed to show much increase in heat transport to the high latitudes (Korty et al., 2008).

More recently, it has become apparent that much of the model data discrepancy may actually be due to deficiencies in the data. While older temperature reconstructions from pelagic foraminifera (e.g. Zachos et al., 1994) showed Eocene tropical temperatures comparable to or even lower than today, more recent studies using exceptionally well-preserved forams indicate much higher temperatures (Pearson et al., 2007), suggesting that the colder estimates are biased low due to diagenesis. With these new tropical temperature estimates, the low-gradient paradox becomes much less severe.

Hyperthermals

The early Eocene was also exceptional in that it was punctuated by a series of hyperthermals, short-lived events of extreme global warmth. Several such events have been documented (Lourens et al., 2005). The best-known hyperthermal, the Palaeocene–Eocene thermal maximum (PETM), saw global temperature soar by over 5°C in less than 10 000 years (Zachos et al., 2001), and then revert to pre-event levels over a longer period of about 150 000 years. The warming was accompanied by a sharp negative carbon isotope excursion (a reduction of the $^{12}\text{C}/^{13}\text{C}$ ratio) of global extent, and rapid shoaling of the calcite compensation depth indicative of ocean acidification (Zachos et al., 2005). The generally accepted interpretation of these facts is that a rapid injection of isotopically light carbon into the atmosphere caused a large, transient increase in the greenhouse effect, producing the rapid warming. The carbon anomaly was subsequently reabsorbed on the typical timescale associated with silicate rock weathering and oceanic deposition.

The origin of the requisite large, rapid carbon injection remains enigmatic. Several hypotheses have been put forward. The carbon could have been ‘baked’ out of deep rocks by volcanic intrusions (Svensen et al., 2004), although this is hard to reconcile with the short-lived and recurrent nature of the events (Lourens et al., 2005). Catastrophic methane release from clathrate deposits has also been invoked (Dickens, 2003), though building large deposits under warm early Eocene conditions is difficult (Buffett and Archer, 2004). Finally, the isotopically light carbon could have come from the large-scale respiration of organic matter, possibly through the dessication of epicontinental seas (Higgins and Schrag, 2006).

Whatever its origin, the rapid carbon injection during hyperthermals is the closest known analogue to the ongoing anthropogenic carbon release. In this context, hyperthermals provide a perspective on the kind of climate changes we may expect in future, as well as a unique opportunity to test the climate models used to predict these changes. The most recent estimate puts the total carbon release during the PETM at no more than 3000 Gt (Zeebe et al., 2009), which would have resulted in somewhat less than a doubling of the pre-existing atmospheric CO_2

concentration. This is insufficient to explain the estimated global warming of 5°C during the PETM, given the typical sensitivity of current climate models of around 3°C per doubling of CO₂. Thus, if we believe these estimates to be correct, either some unknown positive feedbacks acted during the PETM to amplify the warming, or the sensitivity of current climate models is biased towards low values.

2.3.4 Late Cenozoic glaciation

Glaciating Antarctica

Through the late Cretaceous and early Palaeogene, the earth was in a greenhouse state free of permanent continental ice sheets. Then, at the Eocene–Oligocene (EO) boundary (c. 34 mya), a large ice sheet abruptly formed covering Antarctica. The transition is marked by a step-like rise in the benthic δ¹⁸O record (Fig 2.2). Early work conjectured that the transition was tectonically driven, coinciding with the separation of Australia from Antarctica (Kennett, 1977). The opening of the Tasman Gateway created a continuous ocean belt around Antarctica, bringing the Antarctic Circumpolar Current into existence and supposedly interrupting the southward flow of warm water from the subtropics that existed previously. The resulting thermal isolation of Antarctica would then have led to its glaciation. A study using a simplified ocean–atmosphere model showed that opening a circumpolar seaway would indeed give a modest cooling over Antarctica (Sijp and England, 2004). On the other hand, simulations using a full physics coupled ocean–atmosphere climate model under late Eocene conditions showed that Antarctica was in fact thermally isolated even before the opening of the Tasman Gateway (Huber et al., 2004); no warm subtropical currents reached high southern latitudes even with Tasmania attached to Antarctica.

DeConto and Pollard (2003) proposed an entirely different scenario for the onset of Antarctic glaciation. They ran a global atmospheric model with late Eocene topography coupled to an ice-sheet model, and gradually reduced the level of atmospheric CO₂. They found that once the concentration dropped below a threshold level, an ice sheet abruptly and spontaneously appeared over Antarctica. Although the change in CO₂ was gradual, the build-up of the ice sheet was rapid due to the strong feedbacks at play. The most obvious feedback involves surface albedo: ice and snow are much more reflective than most other types of surface cover, so their presence reduces net absorbed insolation and leads to lower temperatures, promoting further ice formation. Additionally, increasing ice-sheet height enhances mass gain during winter and reduces mass loss during summer (because a high ice-sheet top is colder and therefore less subject to melting), so that the ice sheet's mass balance becomes increasingly positive the larger it grows. This positive feedback is finally halted by downhill ice flow once the ice sheet becomes sufficiently massive. Antarctica, having high orography near the

pole, is particularly sensitive to this mechanism, while the Arctic, with an ocean around the pole and no high orography on the surrounding continents, requires much lower levels of CO₂ to form an ice sheet.

Glaciating the Arctic

While the early Oligocene emplacement of a continental-scale Antarctic ice sheet is well established, the time of first appearance of ice sheets in the northern hemisphere remains controversial. Recent work shows evidence for ice-rafted debris in the northern hemisphere during the Eocene and early Oligocene (Eldrett et al., 2007; Tripathi et al., 2008). Moreover, palaeothermometry studies using magnesium/calcium isotopes suggest little benthic temperature change across the EO boundary, implying that the entire 1.5% shift in $\delta^{18}\text{O}$ is attributable to ice-sheet growth (Billups and Schrag, 2003; Coxall et al., 2005). A shift this large would require more ice growth than can be accommodated on Antarctica. Overall, this new evidence suggests that northern hemisphere ice sheets appeared in the early Oligocene and that the EO boundary glaciation was in fact bipolar. However, DeConto et al. (2008) found that Arctic glaciation requires CO₂ concentration to drop below 280 ppm in their atmosphere-ice-sheet coupled model, a level far below that which prevailed during the early Oligocene according to the palaeo-CO₂ reconstruction of Pagani et al. (2005). In addition, palaeotemperature reconstruction using the TEX86 proxy shows that high latitude sea-surface temperatures dropped by about 5°C across the EO boundary (Liu et al., 2009). This cooling can account for much of the observed oxygen isotope excursion, and the remaining fraction can be explained by an ice volume that can be comfortably accommodated on Antarctica alone. The evidence for ice-rafted debris can be explained by localised, ephemeral ice caps on the higher orography.

The final transition to the modern icehouse world, including orbitally paced waxing and waning of continental-scale northern hemisphere ice sheets, is thought to have occurred in the late Pliocene, *c.* 2.7 mya (Shackleton et al., 1984). Several mechanisms have been proposed for this transition, and there is as yet no clear consensus on which was the dominant one. The Panamanian hypothesis notes the rough coincidence between glacial onset and final closure of the Panamanian seaway (Keigwin, 1982). By preventing mixing of Atlantic and Pacific waters, this closure could have increased the strength of the Atlantic conveyor-belt circulation, increasing the supply of moisture to Greenland and ultimately feeding a larger ice sheet. Atlantic temperature proxy data provide some support for this scenario (Bartoli et al., 2005). The uplift hypothesis suggests that the rise of the Rockies and Himalayan Plateau during the Cenozoic would have altered atmospheric circulation patterns, making Greenland's climate cooler, moister and more conducive to ice-sheet formation (Ruddiman and Kutzbach, 1989). A more recent hypothesis points to evidence that during the late Miocene and early Pliocene

the tropical Pacific was in a permanent El Niño state, i.e. one with permanently warm conditions in the eastern basin and no east–west equatorial temperature gradient (Wara et al., 2005). Transition to the modern regime, with cooler average conditions in the east and interannual fluctuations between El Niño (warm) and La Niña (cool) events, roughly coincides with glacial onset. Noting the modern observed impacts of El Niño events on high-latitude climate, some authors have drawn a causal connection between the two transitions (Molnar and Cane, 2002; Philander and Fedorov, 2003). Finally, it is possible that CO₂ concentrations decreased through the Pliocene (Lunt et al., 2008), bringing the system across the threshold for Arctic glaciation (DeConto et al., 2008). A test of all these hypotheses using a comprehensive atmosphere–ocean ice-sheet model found that only a drop in CO₂ could fully account for glacial onset (Lunt et al., 2008). Furthermore, model date comparison by Huber and Caballero (2003) showed that an active El Niño–Southern Oscillation (ENSO) cycle probably existed during the Eocene, when all conditions for a permanent El Niño were satisfied, so it is difficult to see what could have induced such a state in the Miocene.

2.4 Conclusions

Earth's climate system fluctuates on all timescales, from hours to billions of years. Over its history, earth has gone through a succession of greenhouse and icehouse states, which correlate well with periods of high and low atmospheric CO₂ concentration. Climate modelling is playing an increasingly important role in unravelling the intricate details of these transitions. At the same time, palaeoclimate applications expose some of the greatest limitations of current models. In particular, palaeoclimate proxy data from the very warm climates of the early Eocene suggest that climate models may substantially underestimate the sensitivity of surface temperature to CO₂, particularly at high latitudes. While the long-standing low-gradient paradox may be nearing resolution due to improved palaeotemperature estimates, the emerging challenge is to explain how palaeoclimates may have become so warm with relatively modest levels of CO₂. Changes in clouds, perhaps linked to changes in ocean biota (Kump and Pollard, 2008), may provide part of the answer. This provides an exciting avenue for future research in bioclimate interactions.

References

- Abbot, D. S. and Tziperman, E. (2008). Sea ice, high-latitude convection, and equable climates. *Geophysical Research Letters*, **35**, L03702.
- Alvarez, L. W., Alvarez, W., Asaro, F. and Michel, H. V. (1980). Extraterrestrial cause for the Cretaceous–Tertiary extinction. *Science*, **208**, 1095–1108.

- Bartoli, G., Sarnthein, M., Weinelt, M. et al. (2005). Final closure of Panama and the onset of northern hemisphere glaciation. *Earth and Planetary Science Letters*, **237**, 33–44.
- Berling, D. J. and Berner, R. A. (2005). Feedbacks and the coevolution of plants and atmospheric CO₂. *Proceedings of the National Academy of Sciences of the USA*, **102**, 1302–1305.
- Benton, M. J. and Twitchett, R. J. (2003). How to kill (almost) all life: the end-Permian extinction event. *Trends in Ecology and Evolution*, **18**, 358–365.
- Beran, J. (1994). *Statistics for Long-Memory Processes*. Monographs on Statistics and Applied Probability, Boca Raton, FL: Chapman and Hall/CRC.
- Berner, R. A. (2004). *The Phanerozoic Carbon Cycle: CO₂ and O₂*. Oxford: Oxford University Press.
- Berner, R. A., Lasaga, A. and Garrels, R. (1983). The carbonate-silicate geochemical cycle and its effect on atmospheric carbon dioxide over the past 100 million years. *American Journal of Science*, **283**, 641–683.
- Billups, K. and Schrag, D. P. (2003). Application of benthic foraminiferal Mg/Ca ratios to questions of Cenozoic climate change. *Earth and Planetary Science Letters*, **209**, 181–195.
- Buffett, B. and Archer, D. (2004). Global inventory of methane clathrate: sensitivity to changes in the deep ocean. *Earth and Planetary Science Letters*, **227**, 185–199.
- Caballero, R. and Langen, P. L. (2005). The dynamic range of poleward energy transport in an atmospheric general circulation model. *Geophysical Research Letters*, **32**, L02705.
- Caldeira, K. and Kasting, J. F. (1992). Susceptibility of the early Earth to irreversible glaciation caused by carbon dioxide clouds. *Nature*, **359**, 226–228.
- Charney, J. G., Fjørtoft, R. and von Neumann, J. (1950). Numerical integration of the barotropic vorticity equation. *Tellus*, **2**, 237–254.
- Coxall, H. K., Wilson, P. A., Pälike, H., Lear, C. H. and Backman, J. (2005). Rapid stepwise onset of Antarctic glaciation and deeper calcite compensation in the Pacific Ocean. *Nature*, **433**, 53–57.
- DeConto, R. M. and Pollard, D. (2003). Rapid Cenozoic glaciation of Antarctica induced by declining atmospheric CO₂. *Nature*, **421**, 245–249.
- DeConto, R. M., Pollard, D., Wilson, P. et al. (2008). Thresholds for Cenozoic bipolar glaciation. *Nature*, **455**, 652–656.
- Dickens, G. R. (2003). Rethinking the global carbon cycle with a large, dynamic and microbially mediated gas hydrate capacitor. *Earth and Planetary Science Letters*, **213**, 169–183.
- Donnadieu, Y., Godd eris, Y., Ramstein, G., N ed elec, A. and Meert, J. (2004). A snowball Earth climate triggered by continental break-up through changes in runoff. *Nature*, **428**, 303–306.
- Eldrett, J. S., Harding, I. C., Wilson, P. A., Butler, E. and Roberts, A. P. (2007). Continental ice in Greenland during the Eocene and Oligocene. *Nature*, **446**, 176–179.
- Emanuel, K. (2001). Contribution of tropical cyclones to meridional heat transport by the oceans. *Journal of Geophysical Research*, **106**, 14771–14781.
- Emanuel, K. (2002). A simple model for multiple climate regimes. *Journal of Geophysical Research*, **107** (D9), 4077.
- Frakes, L. A., Francis, J. E. and Syktus, J. I. (1992). *Climate Modes of the Phanerozoic*. Cambridge: Cambridge University Press.

- Greenwood, D. R. and Wing, S. L. (1995). Eocene continental climates and latitudinal temperature gradients. *Geology*, **23**, 1044–1048.
- Higgins, J. A. and Schrag, D. P. (2006). Beyond methane: towards a theory for the Paleocene-Eocene thermal maximum. *Earth and Planetary Science Letters*, **245**, 523–537.
- Hoffman, P. F., Kaufman, A. J., Halverson, G. P. and Schrag, D. P. (1998). A Neoproterozoic snowball Earth. *Science*, **281**, 1342–1346.
- Huber, M. and Caballero, R. (2003). Eocene El Niño: evidence for robust tropical dynamics in the ‘hothouse’. *Science*, **299**, 877–881.
- Huber, M. and Sloan, L. C. (2001). Heat transport, deep waters and thermal gradients: coupled climate simulation of an Eocene greenhouse climate. *Geophysical Research Letters*, **28**, 3841–3884.
- Huber, M., Brinkhuis, H., Stickley, C. E. et al. (2004). Eocene circulation of the Southern Ocean: was Antarctica kept warm by subtropical waters? *Paleoceanography*, **19**, PA4026.
- Huybers, P. and Curry, W. (2006). Links between annual, Milankovitch and continuum temperature variability. *Nature*, **441**, 329–332.
- Intergovernmental Panel on Climate Change (IPCC) (2007). *Climate Change 2007: the Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change*, ed. S. Solomon, D. Qin, M. Manning et al. Cambridge: Cambridge University Press.
- Keigwin, L. (1982). Isotopic paleoceanography of the Caribbean and East Pacific: role of Panama uplift in late Neogene time. *Science*, **217**, 350–353.
- Kennett, J. P. (1977). Cenozoic evolution of Antarctic glaciation, the circum-Antarctic Ocean, and their impact on global paleoceanography. *Journal of Geophysical Research*, **82**, 3843–3860.
- Kent, D. V. and Muttoni, G. (2008). Equatorial convergence of India and early Cenozoic climate trends. *Proceedings of the National Academy of Sciences of the USA*, **105**, 16065–16070.
- Kiehl, J. T. and Shields, C. A. (2005). Climate simulation of the latest Permian: implications for mass extinction. *Geology*, **33**, 757–760.
- Kirschvink, J. L. (1992). Late Proterozoic low-latitude global glaciation: the snowball earth. In *The Proterozoic Biosphere: a Multidisciplinary Study*, ed. J. W. Schopf and C. Klein. Cambridge: Cambridge University Press, pp. 51–52.
- Korty, R. L., Emanuel, K. A. and Scott, J. R. (2008). Tropical cyclone-induced upper-ocean mixing and climate: application to equable climates. *Journal of Climate*, **21**, 638–654.
- Kump, L. R. and Pollard, D. (2008). Amplification of Cretaceous warmth by biological cloud feedbacks. *Science*, **320**, 195.
- Kutzbach, J. E. (1976). The nature of climate and climatic variations. *Quaternary Research*, **6**, 471–480.
- Lewis, J. M. (1998). Clarifying the dynamics of the general circulation: Phillips’s 1956 experiment. *Bulletin of the American Meteorological Society*, **79**, 39–60.
- Lindzen, R. S., Chou, M. D. and Hou, A. Y. (2001). Does the earth have an adaptive infrared iris? *Bulletin of the American Meteorological Society*, **82**, 417–432.
- Liu, Z., Pagani, M., Zinniker, D. et al. (2009). Global cooling during the Eocene-Oligocene climate transition. *Science*, **323**, 1187–1190.

- Lourens, L. J., Sluijs, A., Kroon, D. et al. (2005). Astronomical pacing of late Palaeocene to early Eocene global warming events. *Nature*, **435**, 1083–1087.
- Lunt, D. J., Foster, G. L., Haywood, A. M. and Stone, E. J. (2008). Late Pliocene Greenland glaciation controlled by a decline in atmospheric CO₂ levels. *Nature*, **454**, 1102–1105.
- Lynch, P. (2006). *The Emergence of Numerical Weather Prediction: Richardson's Dream*. Cambridge: Cambridge University Press.
- Mayhew, P. J., Jenkins, G. B. and Benton, T. G. (2008). A long-term association between global temperature and biodiversity, origination and extinction in the fossil record. *Proceedings of the Royal Society of London B*, **275**, 47–53.
- McElwain, J. C., Beerling, D. J. and Woodward, F. I. (1999). Fossil plants and global warming at the Triassic–Jurassic boundary. *Science*, **285**, 1386–1390.
- McGrath, R. and Lynch, P., eds. (2008). *Ireland in a Warmer World: Scientific Predictions of the Irish Climate in the Twenty-First Century*. Dublin: Met Éireann.
- Mitchell, J. M. (1976). An overview of climatic variability and its causal mechanisms. *Quaternary Research*, **6**, 481–493.
- Molnar, P. and Cane, M. A. (2002). El Niño's tropical climate and teleconnections as a blueprint for pre-Ice Age climates. *Paleoceanography*, **17**, 1021.
- O'Gorman, P. A. and Schneider, T. (2008). The hydrological cycle over a wide range of climates simulated with an idealized GCM. *Journal of Climate*, **21**, 3815–3832.
- Pagani, M., Zachos, J. C., Freeman, K. H., Tipple, B. and Bohaty, S. (2005). Marked decline in atmospheric carbon dioxide concentrations during the Paleogene. *Science*, **309**, 600–603.
- Pearson, P. N., van Dongen, B. E., Nicholas, C. J. et al. (2007). Stable warm tropical climate through the Eocene Epoch. *Geology*, **35**, 211–214.
- Philander, S. G. and Fedorov, A. V. (2003). Role of tropics in changing the response to Milankovich forcing some three million years ago. *Paleoceanography*, **18**, 1045.
- Phillips, N. (1956). The general circulation of the atmosphere: a numerical experiment. *Quarterly Journal of the Royal Meteorological Society*, **82**, 123–164.
- Pierrehumbert, R. T. (2004). High levels of atmospheric carbon dioxide necessary for the termination of global glaciation. *Nature*, **429**, 646–649.
- Pierrehumbert, R. T. (2005). Climate dynamics of a hard snowball earth. *Journal of Geophysical Research*, **110**, D1111.
- Royer, D. L. (2006). CO₂-forced climate thresholds during the Phanerozoic. *Geochimica et Cosmochimica Acta*, **70**, 5665–5675.
- Royer, D. L., Berner, R. A., Montañez, I. P., Tabor, N. J. and Beerling, D. J. (2004). CO₂ as a primary driver of Phanerozoic climate. *Geological Society of America Today*, **14**, 5.
- Ruddiman, W. F. and Kutzbach, J. E. (1989). Forcing of late Cenozoic Northern Hemisphere climate by plateau uplift in southern Asia and the American West. *Journal of Geophysical Research*, **94**, 18, 409–418, 427.
- Sagan, C. and Mullen, G. (1972). Earth and Mars: evolution of atmospheres and

- surface temperatures. *Science*, **177**, 52–56.
- Sepkoski, J. J. (1982). Mass extinctions in the Phanerozoic oceans: a review. *Geological Society of America Special Papers*, **191**, 283–289.
- Shackleton, N., Backman, J., Zimmerman, H. et al. (1984). Oxygen isotope calibration of the onset of ice-rafting and history of glaciation in the North Atlantic region. *Nature*, **307**, 620–623.
- Shaviv, N. J. and Veizer, J. (2003). Celestial driver of Phanerozoic climate? *Geological Society of America Today*, **13**, 4–10.
- Shellito, C. J., Sloan, L. C. and Huber, M. (2003). Climate sensitivity to atmospheric CO₂ levels in the Early-Middle Paleogene. *Paleogeography, Paleoclimatology, Paleoecology*, **193**, 113–123.
- Sijp, W. P. and England, M. H. (2004). Effect of the Drake Passage throughflow on global climate. *Journal of Physical Oceanography*, **34**, 1254–1266.
- Sloan, L. C. and Pollard, D. (1998). Polar stratospheric clouds: a high latitude warming mechanism in an ancient greenhouse world. *Geophysical Research Letters*, **25**, 3517–3520.
- Sluijs, A., Schouten, S., Pagani, M. et al. (2006). Subtropical Arctic Ocean temperatures during the Palaeocene/Eocene thermal maximum. *Nature*, **441**, 610–613.
- Soden, B. J. and Held, I. M. (2006). An assessment of climate feedbacks in coupled ocean–atmosphere models. *Journal of Climate*, **19**, 3354–3360.
- Svensen, H., Planke, S., Malthe-Sørenssen, A. et al. (2004). Release of methane from a volcanic basin as a mechanism for initial Eocene global warming. *Nature*, **429**, 542–545.
- Tripathi, A. K., Eagle, R. A., Morton, A. et al. (2008). Evidence for glaciation in the Northern Hemisphere back to 44 Ma from ice-rafted debris in the Greenland Sea. *Earth and Planetary Science Letters*, **265**, 112–122.
- van de Schootbrugge, B., Quan, T. M., Lindström, S. et al. (2009). Floral changes across the Triassic/Jurassic boundary linked to flood basalt volcanism. *Nature Geoscience*, **2**, 589–594.
- Walker, J. C. G., Hays, P. B. and Kasting, J. F. (1981). A negative feedback mechanism for the long-term stabilization of the earth's surface temperature. *Journal of Geophysical Research*, **86**, 9776–9782.
- Wara, M. W., Ravelo, A. C. and Delaney, M. L. (2005). Permanent El Niño-like conditions during the Pliocene warm period. *Science*, **309**, 758–761.
- Washington, W. M. and Parkinson, C. L. (2005). *An Introduction to Three-Dimensional Climate Modeling*, 2nd edn. Sausalito, CA: University Science Books.
- Wilby, R. L. and Wigley, T. M. L. (1997). Downscaling general circulation model output: a review of methods and limitations. *Progress in Physical Geography*, **21**, 530–548.
- Wilson, J. T. (1966). Did the Atlantic close and then re-open? *Nature*, **211**, 676–681.
- Wunsch, C. (2003). The spectral description of climate change including the 100 ky energy. *Climate Dynamics*, **20**, 353–363.
- Wunsch, C. (2004). Quantitative estimate of the Milankovitch-forced contribution to observed Quaternary climate change. *Quarterly Science Reviews*, **23**, 1001–1012.

- Zachos, J. C., Stott, L. D. and Lohmann, K. C. (1994). Evolution of early Cenozoic marine temperatures. *Paleoceanography*, **9**, 353–387.
- Zachos, J. C., Pagani, M., Sloan, L., Thomas, E. and Billups, K. (2001). Trends, rhythms and aberrations in global climate 65 Ma to present. *Science*, **292**, 686–693.
- Zachos, J. C., Röhl, U., Schellenberg, S. A. et al. (2005). Rapid acidification of the ocean during the Paleocene–Eocene thermal maximum. *Science*, **308**, 1611–1615.
- Zeebe, R. E., Zachos, J. C. and Dickens, G. R. (2009). Carbon dioxide forcing alone insufficient to explain Palaeocene–Eocene Thermal Maximum warming. *Nature Geoscience*, **2**, 576–580.