
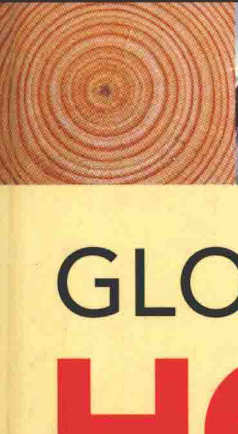



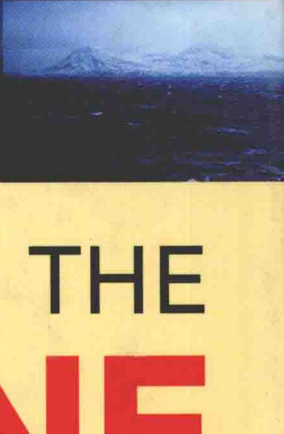

GLOBAL CHANGE IN THE HOLOCENE

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CHAPTER

1

INTRODUCTION: THE HOLOCENE, A SPECIAL TIME

Frank Oldfield

1.1 The Holocene in Temporal Perspective

For anyone raised in the tradition of field-based Quaternary studies in northwestern Europe, the transition from the end of glacial times to the beginning of the Holocene is one of the most notable and readily detectable of stratigraphic boundaries. Almost everywhere, it is marked by evidence for a dramatic shift in surface processes, denoting a major climate change that in turn triggered a whole sequence of responses in both abiotic and biotic ecosystem components. Evidence for the precise age, suddenness and synchroneity of the transition has gradually accumulated over the last 70 years until now, we have remarkably precise chronological control on its timing, the pace of change and the extraordinary spatial and temporal coherence of response over large areas of the globe. Indeed, were it not that colleagues dealing with **contemporary transformations** of the Earth System had placed their concerns so firmly under the heading of 'Global Change', that term could serve perfectly for the opening of the Holocene.

The isotopically-inferred temperature record in the GRISP/GISP ice cores from Central Greenland points up the sharp contrast between late Pleistocene and Holocene in terms both of mean values and of the amplitude of variability (Dansgaard *et al.*, 1993). To a large extent, the shift to the Holocene appears to be a rapid switch in mode from low mean temperatures and extreme variability on all time-scales from decadal to millennial, to one of much higher mean values and lower variability. Thus, if the record from Central Greenland were the only template for our interpretation of Holocene environmental change, we would be considering a period of rather remarkable invariance relative to that which preceded it. But the empirical evidence from other parts of the world, as well as our knowledge of the changing patterns of, and interactions between, external forcings and feedbacks reveal this as a serious oversimplification.

The orbitally driven changes that appear to have triggered the Pleistocene–Holocene transition were relatively gradual. Moreover, orbitally driven changes in solar irradiance have continued throughout the Holocene and they have had different expressions at different latitudes. Only in the second half of the Holocene, roughly the last 6000 years, have they been broadly comparable to those prevailing today. Thus, at the opening of the Holocene, the effects of smoothly changing external forcing were mediated by internal system dynamics to generate a range of abrupt and synchronous changes in many parts of the world, but not all the responses were immediate and speedily accomplished. Ice takes time to melt and the great northern hemisphere polar ice did not disappear overnight. Nor did it simply wane smoothly and continuously everywhere. In consequence, eustatic sea-level too took several millennia to reach its mid-Holocene levels. Not only did many physical responses to Holocene warming and deglaciation take place over several thousands of years, biotic responses too were not completed instantaneously. Migration, soil development, competition and succession all played a part in modulating ecological responses to the major changes in the Earth System that marked the opening of the Holocene. We may therefore think of this latter shift as the beginning of a longer, complex period of transition as well as a sharp boundary between Earth System regimes. Depending on our research focus and on where and how we look, it was both.

Did these transitional changes during the first half of the Holocene play out against the backdrop of a global climate as relatively invariant as the Central Greenland temperature record suggests? Undoubtedly not. High latitude temperature variability may have been reduced, but there were still major changes, especially during the early Holocene. Elsewhere, at lower latitudes and notably in tropical regions, hydrological variability was extreme over the same period, with dramatic changes in lake level well documented in Africa and South America. To some extent, the climatic variability that is recorded during the first half of the Holocene may be attributed to the sequence of changes taking place in the wake of deglaciation and to the way in which the changes interacted with the prevalent patterns of orbital forcing, but these factors alone fail to account for all the changes observed. Changes in ocean currents and land biota also appear to have influenced climate, at least on a continental scale.

Even during the second half of the Holocene, when orbitally driven external forcing was broadly similar to today, ice had melted to a minimum and eustatic sea-level had recovered, there is strong evidence for significant climate variability in all areas and on all time-scales. Such variability, as well as having had important effects on past hydrological regimes and ecosystems, is of outstanding interest at the present day. It is against this background variability that we must seek to detect and characterize the imprint of human-induced climate change resulting from ever increasing atmospheric greenhouse gas concentrations. Moreover, future climate change will be, in part, an expression of similar variability as it plays out in the future and interacts with the effects of any human-induced climate change.

The Holocene period thus emerges not as a bland, pastoral coda to the contrasted movements of a stirring Pleistocene symphony; rather we now see it as a period of continuous change, the documenting and understanding of which becomes increasingly urgent as our concerns for future climate change grow. All the foregoing serves to reinforce the importance of the Holocene as a major research challenge; but there is an additional element that may be of even greater importance, for it is during the Holocene, and especially the later part, that human activities have begun to reshape the nature of the Earth System not only through systemic impacts on the composition and concentrations of atmospheric trace gases, but through the cumulative effects of land clearance, deforestation, soil erosion, salinization, urbanization, loss of biodiversity and a myriad other impacts that have transformed our environment at an ever accelerating rate. These processes began many thousands of years ago at local and regional levels in long settled areas of the globe, but over the last two centuries and at an accelerating rate in the last few decades, the impacts have become global and the implications for rapidly increasing human populations a cause for growing anxiety. It follows from all the above that the themes of this book are of major relevance to our present-day environmental concerns (cf. Oldfield and Alverson, 2003).

1.2 THE DEMISE OF THE 35-YEAR MEAN

One of the cornerstones of climatology 50 years ago was the notion of the 35-year mean. This purported to encapsulate an adequate first approximation to the climate of a station or region. At the same time, it was recognized that climate had changed in the past, as witness the sequence of glaciations and the climate oscillations they implied. It is doubtful whether any conflict was perceived between these two perspectives as they were the concerns of quite different scholarly communities. Reconciling the notion of the 35-year mean with the realization that climate had changed was, in any case, quite easy if one took the view that past change entailed a switch between distinctive episodes, each of relative constancy. The 'post-glacial' period in northwest Europe for example, was one that could be divided into a suite of phases – Pre-Boreal, Boreal, Atlantic, Sub-Boreal and Sub-Atlantic. From around 500 BC, we had been in the cool, wet Sub-Atlantic phase and, by implication, the 35-year mean could serve to describe the climate regime typical of that period for any given location. It took the work of scholars like Gordon Manley (1974) and Hubert Lamb (1963) to bridge the gap between instrumental records and the longer time-scales of climate change. In so doing, they helped to show that climate variability was continuous on all time-scales, that short-term changes were nested within longer-term trends and that there was no such thing as a mean value that could serve for any time interval other than that for which it was calculated.

Put another way, change is the norm. This has important implications for almost every aspect of environmental science, for it shifts our perspective away from any static descriptor of a relatively constant state to an acknowledgement that for any place and over any time-scale there has been an envelope of variability which changes with the time-span which it represents. Characterizing and understanding the processes contributing to and modulating past variability on a wide range of time-scales constitutes a major scientific challenge, but one that is of vital interest at the present day and for the future.

1.3 LESSONS FROM THE PAST

When the threat of future greenhouse warming was first identified and clearly stated (see summary in Oeschger, 2000), it was tempting to turn to the past for analogues. There had been warmer worlds in the past; what were they like? Could they provide a partial template for a possibly warmer world of the future? Quite quickly, this rather simple way of using hindsight was seen to be seriously flawed. We cannot hope to find analogues with any useful degree of realism by turning to periods when the external boundary conditions and the very configuration of the planet were different. Instead, palaeo-scientists began to interrogate the past record of environmental change with questions about processes, rates of change, long-term Earth System dynamics, non-linear responses to external forcing, feedback mechanisms involving the hydrosphere and biosphere and a myriad other similar issues (see e.g. Alverson *et al.*, 2000, 2003).

In adopting this much more realistic research agenda, the main focus in palaeo-science has been on the late Quaternary period. Indeed, the record of the last four glacial cycles spanning the last 430,000 years from Vostok in Antarctica (Petit *et al.*, 1999) has come to serve as an almost universal template for this type of research. The Holocene represents no more than the last 2.7 per cent of this time interval. What are the special qualities of the period that make it of compelling interest? What are the key questions we can address by using the record from the Holocene and what are the key issues that improved knowledge of the Holocene may help us to resolve?

1.4 THE SPECIAL INTEREST OF THE HOLOCENE

The realization that the isotopically inferred temperature record from Central Greenland was not a template for all aspects of Holocene climate everywhere has been noted above. Nevertheless, the contrast between late Pleistocene and Holocene variability in the ice core record has strongly influenced thinking in the research community. It has, for example, added special point to questions about climate variability in warm, interglacial intervals. Evidence for climate variability in the Eemian interglacial (Marine Isotope Stage 5e) has evoked a good deal of interest, but continuous, well dated, fine resolution records from the Eemian are rare. It is to the Holocene itself that we must turn for the bulk of our evidence for 'warm climate' variability. The paragraphs that follow seek to highlight some of those qualities of the Holocene that make the record of environmental change during the period of such special interest and value.

1.4.1 Boundary Conditions, External Forcing and Internal Feedbacks

As already hinted at above, the Holocene as a whole is the period for which we have the most information about climate variability during warm, interglacial times. Significant changes in temperature that were certainly synchronous between Greenland and Europe have been well documented for the early Holocene (Alley *et al.*, 1997; von Grafenstein *et al.*, 1998). Even more dramatic in human terms were the widescale changes in lake levels, plant cover and soil moisture that took place at lower latitudes and continued at least until around 4000 years ago (Gasse and Van Campo, 1994). Less dramatic, but nevertheless highly significant, changes in hydrology have also been recorded throughout the second half of the Holocene (see e.g. Verschuren *et al.*, 2000).

The pattern of orbitally driven solar forcing changes relatively slowly and continuously, but over the last 6000 years, which is to say during the second half of the Holocene, it has not differed greatly from the pattern that prevailed during the centuries immediately before the human-induced increase in atmospheric greenhouse gas concentrations began. By the middle of the Holocene, other aspects of the Earth System that influence climate significantly – polar ice cap and sea-ice extent, sea-level and major terrestrial biomes, for example – had all achieved states within an envelope of variability not significantly different from that typical of the last millennium. Thus the main patterns of forcing and feedbacks that characterized the period immediately before human activities began to modify the atmosphere significantly were, broadly speaking, in place by the middle of the Holocene. Anything that we can learn about variability and environmental change since then thus has special relevance for understanding the processes operating now and in the most recent past.

As Bradley (pp. 10–19 in this volume) points out, solar irradiance reaching the outer edge of the earth's atmosphere varies on many time-scales and is modulated by processes some of which are quite independent of orbital changes. The role of these shorter-term variations in solar activity as drivers of global climate has recently received increasing attention. In part, this is due to the fact that for the Holocene period it appears possible to reconstruct a detailed and well dated proxy record of variations in received solar irradiance by measuring deviations in the relationship between the decline in radiocarbon concentrations with age in tree-rings and true dendrochronological age (Stuiver *et al.*, 1991). Where records of past climate variability have been sufficiently well and independently dated, this opens up the possibility of exploring the extent to which the climate changes recorded are coherent with inferred variations in solar activity. Our growing knowledge of the Holocene thus provides key information for testing hypotheses about climate forcing.

1.4.2 Modes of Variability

One of the ways in which climatologists make sense of climate variability on a global scale is by identifying and characterizing relatively distinct modes of variability. The El Niño Southern Oscillation (ENSO) and the North Atlantic Oscillation (NAO) are well-known examples. Other modes currently recognized include a decadal oscillation in the North Pacific and an Arctic Oscillation that interacts with the NAO. One of the key findings of recent research on late Holocene records is that these modes of variability are remarkably protean. Over a period of decades and centuries, their amplitudes, frequencies and spatial domains change (see e.g. Cole and Cook, 1998; Markgraf and Diaz, 2001). This knowledge presents both a contemporary caution and a future challenge. In our present state of knowledge it reduces the confidence with which predictions of the long-term incidence and effects of these modes in the future can be made. At the same time, it challenges us to discover the factors responsible for the decadal- and century-scale variability. Only by understanding these and incorporating them in model simulations will there be any realistic chance of improving future predictability. Once more, the Holocene period is the crucial time interval for exploring these issues, though longer-term insights into the nature of ENSO variability, for example, also shed important light on the range of possible behaviours ENSO may assume (Tudhope *et al.*, 2001).

1.4.3 Continuity and Overlap with the Present Day

Many of the archives and proxies that form the toolkit of the palaeo-scientists can bring the Holocene record of variability right up to the present day. Tree-rings are still being formed, lake sediments continue to accumulate, corals and speleothems still grow. This allows Holocene research to reap multiple benefits. The insights gained contribute to our understanding of present-day ecosystems and environmental processes that have been in part conditioned by their antecedents. By bringing records of climate change from the centuries well before significant human impact right through to the short period of instrumental records (Jones and Thompson, pp. 140–158 in this volume), palaeoclimatology makes a crucial contribution to resolving the questions of detection and attribution raised by global warming in recent decades. The same kind of temporal overlap allows direct comparison between recent instrumental records of the amplitude, duration and recurrence intervals of extreme events and their longer-term history (Page *et al.*, 1994; Knox, 2000).

The above examples stress only one facet of the importance of continuity and overlap, for the points made would count for little were it not possible to translate proxy records of environmental change into inferences sufficiently quantitative to permit comparison with direct measurements. This requires calibration (Birks, pp. 107–123 in this volume) and, in this regard, the period of overlap between past proxy records and present-day measurements is crucial. Calibration, whether achieved by comparing directly measured sequences with proxy records covering the same time interval, or by linking proxies to a spatial array of contemporary measurements spanning a range of variability, is at its most robust for situations where the processes, biological communities or geochemical signatures in which the proxy signals reside lie within or as close as possible to the variability encompassed by the calibration process. For time intervals in which past biological communities lack present-day analogues, or abiotic proxies have values that can only be matched by significant extrapolation of a calibration function, the inferences become less secure and the statistical uncertainties much greater. Once more, the Holocene, and especially the late Holocene, have important advantages. As we move further back in time, confidence in quantitative reconstructions of climate often decline quite steeply (see e.g. Bigler *et al.*, 2002).

1.4.4 Chronology

Not only are Holocene palaeoarchives much more abundant than those for earlier time intervals, especially in continental situations, they can generally be dated much more accurately and precisely. Radiocarbon dating, whether used conventionally or by 'wiggle-matching' (van Geel and Mook, 1989), tephra analysis, varve counting and use of annual speleothem growth increments are all most effective for the Holocene and immediately pre-Holocene period. Tree-rings, with relatively few exceptions, are pretty well limited in their use both for direct chronologies and as climate proxies, to the late Holocene (Baillie and Brown, pp. 75–91 in this volume). Refinements in chronology are vital for addressing many of the most urgent questions in palaeoenvironmental research and recent advances should not blind us to the need for ever better chronologies.

1.4.5 Testing Models

All the above advantages ascribed to the Holocene reinforce its value for testing climate models. Rigorous tests require that the empirical basis for the 'ground truth' against which model simulations are compared be as secure and well constrained as possible. Because models provide the only way of developing scenarios of future conditions other than expert opinion or simple extrapolation, testing their output against known conditions in the past is of prime importance. If models are unable to replicate a known set of conditions or sequence of events in the hind-cast mode, they can have little credibility as predictors of future changes. This realization has led to a whole range of model–data interactions using palaeo-research both to improve parameterization and to provide the basis for testing. The synergy between data acquisition and model refinement is well illustrated in Claussen (pp. 422–434 in this volume), mainly using output from EMIC (Earth Models of Intermediate Complexity). Model–data comparison also plays an important role in ascribing recent climate variability to different forcing mechanisms (e.g. Crowley and Kim, 1999), in testing time-slice simulations performed by more complex global circulation and coupled ocean–atmosphere models (Kohfeld and Harrison, 2000; Valdes, pp. 20–35 in this volume), in exploring the implications for past climates of proxies such as stable isotope signatures (Hoffmann, 2002) and reconstructing biomes representing both present and past climate conditions (Prentice *et al.*, 1992, 1993). In all these roles, well-dated and calibrated proxy records from the Holocene are of major importance.

1.5 PREREQUISITES FOR RESEARCH INTO HOLOCENE VARIABILITY

In the above section we seek to identify some of the key reasons why research on Holocene environmental history is of such outstanding importance. Below, we consider briefly what is required for carrying out effective research on the history of environmental change during the Holocene. These are the 'tools' that allow us to extend the record of past change beyond the instrumental period. Clearly, in many parts of the world, documentary records span a much longer period than do instrumental records and these have been used with increasing skill for reconstructing past climate variability and the impacts on human populations and ecosystems of extreme events such as major floods and persistent droughts (Brimblecome, pp. 159–167 in this volume). Below, we concentrate on the evidence available from environmental archives rather than documentary sources.

1.5.1 Environmental Archives

Many of these have already been referred to in the foregoing paragraphs. In essence they are environmental contexts that preserve one or more types of decipherable record of past environmental conditions. They are as diverse as trees, both living and dead, lake and marine sediments, peat bogs, corals, glaciers and ice fields, and speleothems. In many cases a single archive will contain several possible proxies that can be translated, through calibration, into well validated information on the nature of past environmental conditions. Over the last decade, the main thrust of this type of research has been in reconstructing some aspect of past climate.

1.5.2 Proxies

Proxies are components within an archive that can be extracted, identified and quantified in such a way that their implications for past environmental conditions can be reliably and consistently inferred. The current emphasis on climate reconstruction has led to an incredible diversity of proxy climate signatures. In most cases, these refer to temperature, either annual or, more often, seasonal. Overall, there is a tendency for the majority of proxies to reflect spring and/or summer temperatures: for example, biological proxies are usually calibrated to growing-season conditions and glacier melt layers reflect summer warmth. Reconstructions of palaeo-precipitation are less common, but they can be found in sources as contrasted as the stratigraphic record from ombrotrophic peat bogs (Barber and Charman, pp. 210–226 in this volume) and both the geomorphological and stratigraphic evidence for lake level variations. Biological remains often record a range of influences. Aquatic organisms such as diatoms or chironomids, respond to lake chemistry as well as water temperature. Pollen-producing plants that provide the source of the palynological record in peats and sediments reflect human activities such as deforestation and agriculture as well as changes in climate. In these cases, calibration to some aspect of climate is, in effect, imposing a filter on the full range of information intrinsic to the biological record.

Just as many sub-fossil remains contain a range of potentially calibratable signatures, some types of proxy can be identified within a wide and diverse range of contexts. The demonstration that the ratio between the stable isotopes of oxygen ($\delta^{18}O$) and hydrogen (δD) in rain water reflects air temperature has led to the use of stable isotope records as (often rather indirect) palaeoclimateic proxies in a wide range of archives – ice cores, tree-rings, carbonates, both marine and fresh water, speleothems and, more recently, sedimentary plant cellulose and diatom silica. The use of stable isotopes thus constitutes a versatile methodology applicable to a wide range of archives and components, both biotic and abiotic, within them (Leng, pp. 124–139 in this volume).

All the proxies currently available pose challenges in interpretation and no single one alone can be relied on universally to provide a complete and secure record of past climate change. This has led to the frequent use of what is often termed a multi-proxy approach to climate reconstruction (e.g. Lotter, pp. 373–383 in this volume). Where used, this allows the mutually independent records to act as constraints on each other, either reinforcing inferences where they are in agreement, or posing new questions where they differ significantly.

1.5.3 Chronology

The need for chronological control on all records of past environmental change has already been stressed. Ideally, chronological control in Holocene records should achieve decadal or sub-decadal resolution, with annually or seasonally resolved records highly desirable at least for the last millennium. This degree of control permits close comparison between sites and archives, greatly reduces the scope for miscorrelation, allows characterization of short-term, transient responses to events such as volcanic eruptions and provides evidence that can be smoothly linked to instrumental time series. Even where absolute dates are not known, a finely resolved 'floating' chronology makes possible calculations of rates of change as well as the precise sequence of ecosystem responses to perturbations. Tree-rings (Baillie and Brown, pp. 75–91 in this volume), varved lake (Zolitschka, pp. 92–106 in this volume), or marine sediments and speleothems (Lauritzen, pp. 242–263 in this volume) all provide the opportunity for annual dating, but in the many studies, this level of accuracy and precision is unattainable. Nevertheless, increasingly effective use of AMS radiocarbon dating, often coupled with tephra recognition, is providing an ever firmer chronological framework for Holocene research (Pilcher, pp. 63–74 in this volume).

1.6 CONCLUDING OBSERVATIONS

Some of the special qualities as well as the crucial significance of research into Holocene environmental variability have been identified above. This volume seeks to provide an authoritative guide to these, through chapters on methodologies, selected archives and proxies, as well as illustrative examples of applications and case studies. The special attention devoted to the Holocene record from lower latitudes reflects the relative paucity of evidence from these parts of the world (Bush, pp. 384–395 in this volume; Scott, pp. 396–405 in this volume), in comparison with the wealth of information available from North America and Europe.

The interaction between Holocene and immediately pre-Holocene environmental change and a crucial step in the development of human societies is the subject of the chapter by Wright and Thorpe (pp. 49–62 in this volume) and broader issues surrounding the role of climate change in societal development are addressed by Shennan (pp. 36–48 in this volume). These chapters give some sense of the interactions between past climate variability and social organization. Human societies have never been so insulated from environmental processes as to escape vulnerability to major shifts in climatic regime, extreme events or persistent droughts. Recognizing the role of environmental change in human affairs does not imply a return to simple-minded environmental determinism, for the impacts of environmental change on human societies are mediated through all manner of cultural and socio-economic processes. Nevertheless, they are far from insignificant.

This still applies today, especially for the poorest societies, which, in many cases, are the ones deemed most likely to be negatively impacted by future climate change (Houghton *et al.*, 2001). Even for the most technologically advanced societies, evidence from the past can highlight future threats of extreme gravity, as for example, in the case of rapidly dwindling groundwater resources in parts of the western United States. There, as in many other parts of the world, these are largely fossil waters formed during moister periods either in the first half of the Holocene or earlier. They are now being mined at a pace greatly in excess of any foreseeable rate of recharge. The palaeo-record thus provides both a glimpse into history and a dire warning for the future. Possible future impacts are also implicit in the last two chapters. As Claussen (pp. 422–434 in this volume) shows, where models and data are used together to shed light on climate system dynamics, feedbacks between the different components of the Earth System often give rise to non-linear responses disproportionate to the original forcing – an important point to bear in mind when we consider the future implications of the huge global experiment initiated by greenhouse gas enrichment of our atmosphere. Goodwin (pp. 406–421 in this volume) focuses on sea-level variability during the late Holocene and cites empirical evidence pointing up the links between climate variability and sea-level over the last two millennia, a period when the amplitude of global climate variability was much less than even the most modest projections for the next century. Much more work is needed to quantify possible links between recent secular variations in climate and sea-level, but the evidence so far gives no grounds for complacency in a world of densely settled shorelines and numerous coastal conurbations.

Almost all the major changes in human societies that have formed mileposts on the way from hunter-gatherer to the complex civilizations existing today have taken place during the Holocene period, the time since the end of the last Ice Age. Throughout the Holocene, there have been major environmental changes in every part of the world, but the rate of change has accelerated during the last 50 to 100 years, largely as a result of human activities. It is essential that we place our concerns for the future of the environment in the context of the changes that have occurred during the Holocene as a whole. We need to know how quickly ecosystems have responded to changes in the past and how resilient they may be to changes forced by current and future threats, for, if our Holocene past is anything to go by, we may expect 'surprises' – environmental responses well beyond any change in the external processes that provoked them and extremes well outside the range documented during the short period of instrumental records. The history of environmental change during the Holocene provides us with essential knowledge about the way the climate system and the biosphere work, rapidly growing insight into processes and rates of environmental change, a temporal context within which to place many of our present-day observations and a test bed for models designed to look into the future – for if the models cannot create scenarios to match the reality of the past, we must remain sceptical of their power to shed realistic light on the future.

CHAPTER

2

CLIMATE FORCING DURING THE HOLOCENE

Raymond S. Bradley

Abstract: The role of several important factors that have played a role in Holocene climate change is examined. These forcing factors operate on different time-scales: lower frequency (millennial-scale) climate changes associated with orbital forcing, century-scale variability associated with solar forcing, and annual- to decadal-scale variability associated with volcanic forcing. Feedbacks within the climate system may involve non-linear responses to forcing, especially if critical thresholds are exceeded. In addition, there may be distinct regional climate anomaly patterns that result from certain types of forcing. Other anomalies that appear in Holocene paleoclimatic records may be unrelated to external forcing factors, but reflect conditions entirely within the climate system. General circulation model simulations play an important role in helping to understand how these various factors interact to produce the observed changes in Holocene climate.

Keywords: Orbital forcing, Solar forcing, Volcanic forcing

Why did climate change during the Holocene? The paleoclimatic records that are discussed at length in other chapters reflect, to a large extent, the composite effects of external factors operating on the climate system, plus feedbacks within the climate system that were triggered by these factors. Climate forcing (external factors that may cause climate to change) can be considered on several time-scales, ranging from very long-term (multi-millennial) to interannual. The resulting climate in any one region is the consequence of variability across all time-scales, but breaking the spectrum of climate variability down, from lower to higher frequencies, provides a useful way of assessing different forcing mechanisms. Here some of the main forcing factors during the Holocene are considered, beginning with factors operating at the lower frequency end of the spectrum.

2.1 Orbital Forcing

On the very longest, multi-millennial time-scales, the main factors affecting Holocene climate change are related to orbital forcing (changes in obliquity, precession and eccentricity). These changes involved virtually no change in overall global insolation receipts (over the course of each year) but significant re-distribution of energy, both seasonally and latitudinally. Representing the time- and space-varying nature of orbitally driven insolation anomalies is difficult in a single diagram, but Plate 1 shows these changes schematically for each month, with each panel representing the time-varying anomaly pattern over the last 10,000 years, with respect to latitude. In the Early Holocene, precessional changes led to perihelion at the time of the northern hemisphere summer solstice (today it is closer to the winter solstice). This resulted in higher summer insolation in the Early Holocene at all latitudes of the northern hemisphere (ranging from ~40°W/m^2 higher than today at 60°N to 25°W/m^2 higher at the Equator). Thus, July insolation (radiation at the top, or outside, the atmosphere) has slowly decreased over the last 12,000 years (See Plate 1). Anomalies during southern hemisphere summers were smaller, centred at lower latitudes, and they were opposite in sign (that is, insolation increased over the course of the Holocene). For example, January insolation anomalies were ~30°W/m^2 below current values at 20°S in the Early Holocene.

What impact did such changes have on climate in different regions? Unfortunately, it is not a simple matter to translate insolation anomalies of solar radiation entering the atmosphere into radiation receipts at the surface, and it is even more difficult to then infer the effect of such changes on climate. Radiation passing through the atmosphere is reflected and absorbed differently from one region to another (depending to a large extent on the type and amount of cloud cover). Furthermore, surface albedo conditions also determine how much of the radiation reaching the surface will be absorbed. There may also be complexities induced in the local radiation balance. For example, Kutzbach and Guetter (1986) found that a 7 per cent increase in solar radiation at low latitudes, outside the atmosphere, at 9 ka BP was associated with 11 per cent higher net radiation at the surface due to a decrease in outgoing long-wave radiation (because of increased evaporation and higher water vapor levels in the atmosphere, which absorb long-wave radiation). However, this amplification of solar radiation effects was not as important at higher latitudes (where precipitation amounts are much less a function of solar radiation anomalies). Differences in surface properties may lead to changes in regional-scale circulation; for example, differential heating of land versus ocean (with the same insolation anomaly) could lead to land–sea circulation changes. Indeed, this effect served to drive an enhanced monsoon circulation over large parts of the northern continents in the Early Holocene, leading to wetter conditions and consequent changes in vegetation (see below). Finally, on an even larger scale, differential radiation anomalies from the Poles to the Equator may have led to changes in temperature gradients and consequent changes in the overall strength of atmospheric circulation, with associated shifts in the Hadley and extra-tropical circulation (Rind, 1998, 2000).

To assess such complexities under a constantly varying insolation regime requires general circulation model simulations. Numerous studies have examined the effects of orbital forcing for selected time intervals during the Holocene, initially using atmosphere global circulation models (GCMs; with fixed sea-surface temperatures: e.g. Kutzbach and Guetter, 1986; Hall and Valdes, 1997), then models with interactive ocean–atmosphere systems (e.g. Kutzbach and Liu, 1997; Hewitt and Mitchell, 1998) and, more recently, fully coupled ocean–atmosphere–biosphere models, where the interactions between the atmosphere and land surface hydrology and vegetation is treated explicitly (Brovkin *et al.*, 1998; Kutzbach, 1996; Coe and Bonin, 1997; Broström *et al.*, 1998). Most of these studies focus in particular on northern Africa where Early Holocene conditions were much wetter than the Late Holocene, and the transition between these states was quite abrupt, at around 5500 calendar years BP (deMenocal *et al.*, 2000a). All model simulations demonstrate that increased summer insolation in northern hemisphere summers, in the Early Holocene, caused a stronger monsoon circulation and increased precipitation in sub-Saharan Africa. However, unless vegetation and hydrological feedbacks are incorporated into the models, the precipitation amounts simulated are well below those that would have been necessary to support the lakes and vegetation changes that are known (from the paleoclimatic record) to have occurred (Foley, 1994).

These conclusions have been obtained from model simulations, generally centred at specific time intervals (snapshots at 3000-year intervals) through the Holocene (e.g. Kutzbach, 1996; Joussaume *et al.*, 1999). With complex GCMs, given computational constraints, it is not feasible to run long, multi-millennia simulations to examine transient changes. However, the paleoclimatic record suggests that the gradual changes in insolation were not matched by equally gradual changes in surface climate over North Africa. Rather, the transition from arid to humid conditions was abrupt, both at the onset of wetter conditions (~14,800 calendar years BP and at its termination ~5500 calendar years BP). Both transitions correspond to summer (June–July–August (JJA)) insolation levels of ~4 per cent greater than today (outside the atmosphere) at 20°N. To investigate this, transient model simulations have been made for the last 9000 years, using a much lower resolution zonally-averaged model, (but with a coupled ocean–atmosphere system and vegetation feedbacks) (Claussen *et al.*, 1998, 1999; Ganopolski *et al.*, 1998a). These point to the importance of vegetation feedbacks as critically important; vegetation changes abruptly amplified the linear orbital influence on precipitation over North Africa, to produce an abrupt, non-linear change at ~5440 BP, corresponding to the paleoclimatic field evidence (Fig. 2.1). Although the North African case may be an extreme example of the role of vegetation feedbacks, other model simulations also suggest that vegetation changes at pronounced ecotones (such as the tundra–boreal forest interface) may also play a strong role in modifying initial forcing factors (Foley *et al.*, 1994; TEMPO, 1996; Texier *et al.*, 1997).

2.2 SOLAR FORCING

Orbital forcing involves the redistribution of incoming solar energy, both latitudinally and seasonally. Thus there are differential effects on the climate system that can lead to circulation changes, and there may be different responses to the forcing in the northern and southern hemispheres. Changes in solar irradiance (energy emitted by the sun) might be expected to affect all parts of the earth equally. However, this is not so because the response to solar irradiance forcing is amplified regionally, as a result of feedbacks and interactions within the atmosphere (Rind, 2002).

Until quite recently it was assumed (based on measurements in dry, high-altitude locations) that total irradiance did not vary, at least not on inter-annual to decadal scales – hence the term 'solar constant' was coined to describe the energy that is intercepted by the atmosphere when the sun is overhead (1368°W/m^2) (National Research Council, 1994; Hoyt and Schatten, 1997). Satellite measurements over the last ~25 years tell a different story – total solar irradiance (TSI; that is, integrated over all wavelengths) varies by ~0.08 per cent over a Schwabe solar cycle (average length of ~11 years), with maximum values at times of maximum solar activity (when there are many sunspots and bright solar faculae; Lean, 1996; Fröhlich and Lean, 1998) (Fig. 2.2). Furthermore, irradiance changes at very short (ultraviolet) wavelengths vary even more over a solar cycle (Lean, 2000). Such changes have significance because an increase in UV radiation causes more ozone (O_3) to be produced in the upper stratosphere; ozone absorbs radiation (at UV wavelengths of 200–340 nm) so heating rates in the upper atmosphere are increased during times of enhanced solar activity. This then affects stratospheric winds (strengthening

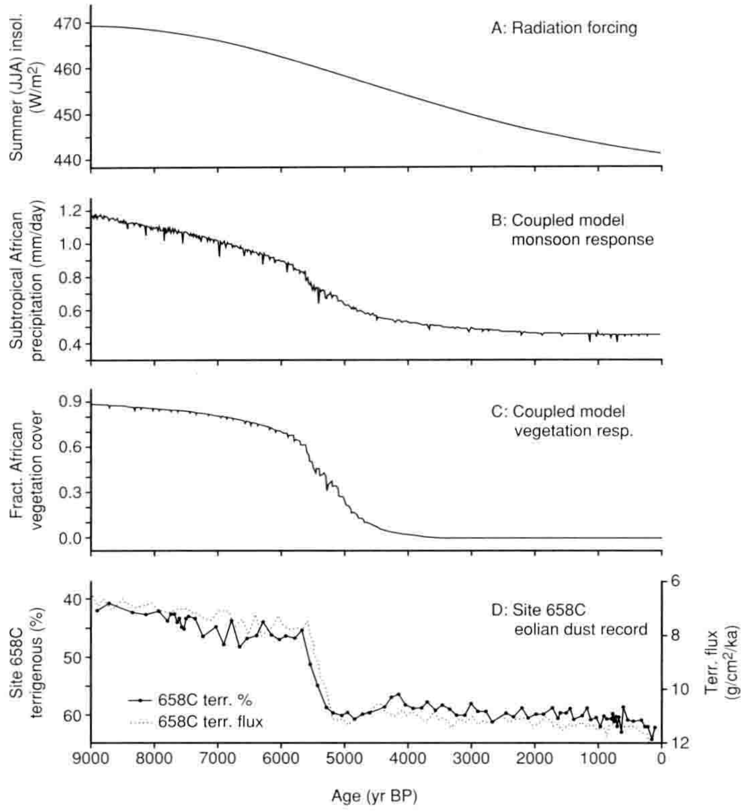

Figure 2.1 Model simulations (Claussen *et al.*, 1999) of the response of North African precipitation (b) and fractional vegetation cover (c) compared to radiation forcing – summer (JJA) radiation at 20°N (a) and the record of eolian (Saharan) dust in a sediment core from off the west coast of North Africa (deMenocal *et al.*, 2000a). The model incorporates vegetation feedbacks that seem to be important in generating a non-linear response to orbital forcing, at ~5500 BP. Reprinted from deMenocal *et al.* (2000a), with kind permission from Elsevier Science.

stratospheric easterlies), which can in turn influence surface climate via dynamical linkages between the stratosphere and the troposphere (Shindell *et al.*, 1999; Baldwin and Dunkerton, 2001; O'Hanlon, 2002). Model simulations of these effects indicate that there is a poleward shift of the tropospheric westerly jet and a poleward extension of the Hadley circulation, by ~70 km from solar minimum to solar maximum, in the summer hemisphere (Haigh, 1996; Larkin *et al.*, 2000). Although such changes are small, if irradiance changes in the past were larger and more persistent than solar cycle variability, the effects may have been quite significant.

How much has irradiance changed over longer time-scales? Satellite measurements are too short to shed light on longer-term irradiance changes, so these must be inferred from other lines of evidence. Lean *et al.* (1992) examined variations in brightness of stars similar to our own sun, concluding that present-day solar activity is at relatively high levels. By analogy with the range

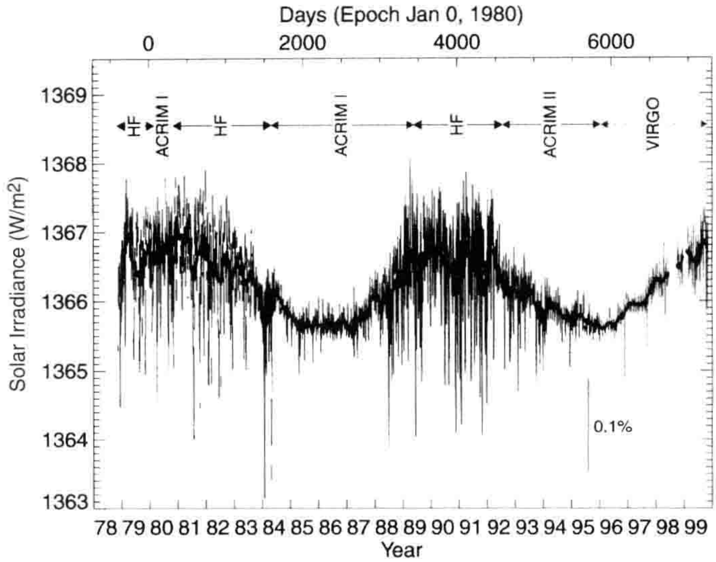

Figure 2.2 Total solar irradiance as recorded by satellites since 1979 (Fröhlich, 2000). This energy is distributed only over the illuminated half of the earth which intercepts it over a circular area ('circle of illumination'). Considering the area of a sphere (4 πr^2) versus that of a circle (πr^2), the average energy impinging at the top of the atmosphere is 1368/4, or ~342°W/m^2 (Hoyt and Schatten, 1997). Hence a variation of 0.08% over an ~11-year (Schwabe) solar cycle is equivalent to a mean forcing of 0.27°W/m^2; this is further reduced (by ~30%) due to planetary albedo effects (scattering, reflection) to ~0.2°W/m^2. It should be born in mind, however, that the earth can not reach radiative equilibrium in relation to forcing over an 11-year solar cycle, as the ocean has great thermal inertia, which smoothes out the effects of rapid changes in external forcing. Reprinted from Fröhlich (2000) with kind permission from Kluwer Academic Publishers.

of brightness in stars like the sun, with or without activity cycles, they inferred that the historical range of TSI varied by ~0.24 per cent, from the time of minimal solar variability at the end of the 17th century (the 'Maunder Minimum', ~AD 1645–1715) to the present (i.e. the mean of the most recent solar cycle). Shorter-term (~11 years) variability of ~0.08 per cent is superimposed on the lower frequency changes (Fig. 2.3). Simple comparisons with long-term temperature estimates suggest that changes in TSI of ~0.24 per cent were associated with mean annual surface temperature changes over the northern hemisphere of 0.2–0.4 °C (Lean *et al.*, 1995) and such changes have also been simulated in GCM and energy balance solar forcing experiments (Rind *et al.*, 1999; Crowley, 2000; Shindell *et al.*, 2001). Indeed, Crowley (2000) concluded that much of the low frequency variability in northern hemisphere temperatures over the last millennium (prior to the onset of global anthropogenic effects) could be explained in terms of solar and volcanic forcing. It is also interesting that distinct patterns of regional temperature change may be associated with solar forcing, as seen in both empirical and modelling studies, due to complex interactions between the circulation in the stratosphere and the troposphere (Shindell *et al.*, 2001; Waple *et al.*, 2001). Prolonged periods of reduced solar activity, like the Maunder Minimum, are associated with overall cooler conditions, but cooling is especially pronounced over mid- to high-latitude continental interiors, and warmer