
Preface

An Introduction to the KIHZ Project

The description of the climate system and the quantification of its natural variability and dynamics is essential to assess an ongoing anthropogenic climate change and to validate climate and biogeochemical models to allow for reliable projections into the future. Because the spatio-temporal coverage of direct meteorological observations is rather limited, high-resolution and absolutely dated climate archives represent the only key to a quantification of seasonal to millennial climate variations in the past. Furthermore, climate models provide insights into the major processes and causes relevant for climate variability on these time scales.

Both approaches represent one side of the same medal, however melting both sides down to one combined effort is often hampered by obstacles defined by the different nature of the approaches. For instance, General Circulation Models (GCMs) per se deal with spatially resolved data representing real climate variables in the model world (such as temperature or precipitation) with each model run reflecting one possible realization of climate history under given boundary conditions. In contrast, the records of natural climate archives are influenced by climate variations as they took place in reality, however, are often representative of local climate conditions only. Moreover, the climate information deduced from natural archives is in nearly all cases based on climate proxies, whose relationship to real climate variables, the so called transfer function, has to be established beforehand.

In order to bridge the gap between climate models and archives and to join these different aspects within the German climate research community, the project "Climate in historical times" (KIHZ) has been initiated in 1998 by geoscientists and climate modelers from five German National Research Centers, members of the Helmholtz association (HGF), as depicted in the inner ring in Fig. 1. Those starting members contributed lake sediments, tree rings, polar ice cores and climate information from permafrost regions as well

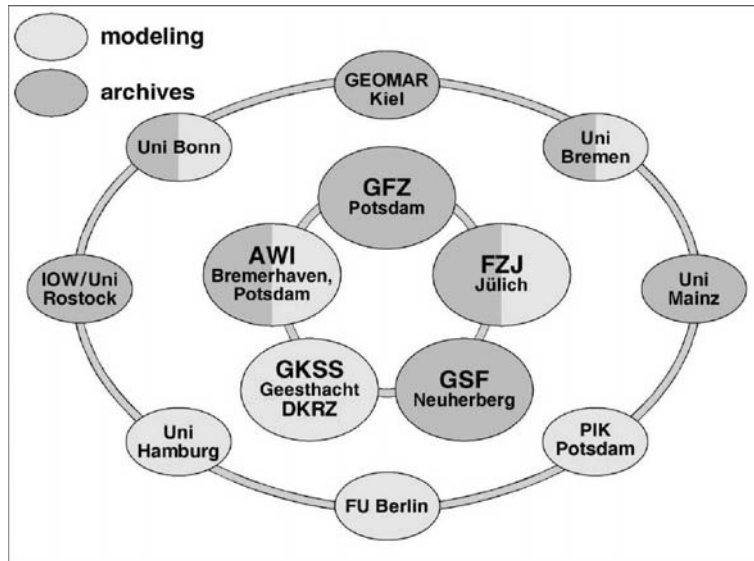


Fig. 1. Structure of the KIHZ project listing all participants. The inner ring summarizes the founding HGF institutes while the outer ring shows the complementary university and Leibniz Association institutes. Dark grey circles indicate institutes contributing natural climate archives, light grey circles indicate climate modeling groups.

as GCMs, high-resolution regional atmospheric models and models of intermediate complexity to KIHZ. In the year 2000, nine additional groups from German universities and institutes from the Leibniz association joined KIHZ (outer ring in Fig. 1) adding further competence e.g. on marine sediments, corals, lake sediments as well as on solar variations, coupled circulation models and earth climate system models of intermediate complexity.

Starting from joined meetings presenting the disciplinary work done prior to and within KIHZ, a very fruitful discussion took place between the diverse groups which did not only result in a common language and a better understanding of the different approaches but also in cross-disciplinary working groups attempting to fill the gaps between climate archives and models. Among others by upscaling studies, relating local climate proxy data to large scale teleconnection patterns using meteorological data and multivariate time series analysis, by improving transfer functions based on empirical and experimental evidence, by aiding the interpretation of climate records using climate models to identify processes relevant for the observed climate variations or by splicing climate reconstructions from climate archives into circulation models using nudging techniques.

The results of these projects are summarized in this volume which is a spin off of the final KIHZ meeting held in Wilhelmshaven in September 2001.

The volume is divided into two parts. Part I provides an overview of the basics of the climate system and its dynamics and of the tools used for climate reconstruction, i.e. natural climate archives and climate models. In Part II individual results achieved during the KIHZ project in the last 5 years are presented which are aimed "Towards a synthesis of Holocene proxy data and climate models".

The Editors. Bremerhaven, November 2003

The Holocene: Considerations with Regard to its Climate and Climate Archives

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Summary. The Holocene as the youngest time of Earth's history is a warm period belonging to the last cold phase (ice-age) since 55 million years. The climate evolution since 11,600 yr B.P. is basically described by using temperature estimations from Greenland ice cores and vegetation changes (pollen assemblage zones) oscillating around an average temperature with an interval of around $\pm 1-3^{\circ}\text{C}$. These main climatic oscillations can be refined by using multi-proxy data on an annual-decadal resolution revealing trends, fluctuations and short events, giving hints to possible pace makers. However, we have to keep in mind that these climatic descriptions are generalizations due to the sensitivities of the used proxy and may reflect only a local instead of a regional or global signal. Therefore, an integration of many proxy-records is necessary to reconstruct the dominating climate regime.

1.1 Introduction

The Holocene (Holozän)¹ is the youngest part of the Earth's history and its definition is given geologically. It comprehends the postglacial warm period, commencing with the glacier retreat from the moraines of central Sweden and starts according to varve- and ice-chronology [631] between 11,590 and 11,550 yr B.P. It is characterized by the development of the vegetation and in Europe additionally by transgressions into the North and Baltic Seas. It is the time of evolution of mankind since the young Palaeolithic. The subdivision of the

¹ According to Klebelsberg, 1949 "Handbuch der Gletscherkunde und Glaziologie", Bd. II, S. 407/ 408, Wien, 'The Recent' was introduced by Ch. Lyell (1833) for youngest deposits, which was renamed by P. Gervais as Holocene ('completely new') following the philosophy of subdividing the Tertiary due to the new nomenclature at this time (e.g., Oligocene etc.).

Holocene² is based on pollen assemblages. The following stratigraphic units were defined in central Europe [306], [785] (Table 1.1):

Table 1.1. Pollen-assemblage zones of Meerfelder Maar (MFM 6) according to Kubitz [559] with varve-ages after Endres [277] in comparison to pollen zones of Firbas [306], Overbeck [785] and the prehistoric evolution.

PAZ MFM	Pollen distribution	Relative Varve age	Duration in years	Historical phases	Pollenzones OVERBECK (1975)	Pollenzones FIRBAS (1949)
9	ash, alder			Modern Age	Subatlantic XII	Subatlantic X
8c	beech, hornbeam - Gramineae	- 1540 B.P.		Middle Age 1500 A.D. - 475 A.D.	Subatlantic XI	Subatlantic IX
8b				VWZ 475 A.D. - 400 A.D.		
8a				Roman Age 400 A.D. - 25 B.C.		
7	Iron Age 25 B.C. - 700 B.C.					
6	alder, birch, oak	2000 - 2750 B.P.	750	Bronze Age 700 B.C. - 1800 B.C.	Subboreal X	Subboreal VIII
5	beech, birch, alder	2750 - 3800 B.P.	1050	Neolithic 1800 B.C. - 5500 B.C.	Subboreal IX	Atlantic VI / VII
4	hazel, alder, oak	3800 - 6200 B.P.	2400		Atlantic VIII	
3	oak, elm, lime, ash [mixed oak forest]	6200 - 8550 B.P.	2350		Mesolithic - 5500 B.C.	Boreal VI / VII
2	hazel, oak, elm	8550 - 10.200 B.P.	1650	Preboreal V		Preboreal IV
1	birch, pine, hazel	10.200 - 10.800 B.P.	600			
	birch, pine	10.800 - 11.000 B.P.	200			

To understand the climate of today and the climate of the Holocene warm period we have to recognize that our 'epoch' has its anchor in the climate history of the Earth over the last 600 million years. Only by understanding the climate variability of the past we are able to discriminate the driving mechanisms for global climate change. Since this time, Earth has experienced 4 cold phases with and 4 warm phases without polar icecaps and high sea level [325]. Obviously, these phases are basically a result of the coupling of several geological processes with orbital and solar irradiance changes. It can be demonstrated during Earth's history that continental drift (plate tectonics and sea floor spreading), distribution of continents and oceanic gateways, mountain building, formation of polar ice caps and mountain glaciers, falling sea levels and following extensions of flat continental areas are coupled. Referring to the last cold phase, Zachos et al. [1178] have made the comment:

² The Holocene time scale is based on a combination of historical and geological calendars both referring to solar planetary motion (year: orbital motion of the Earth around the sun; lunar month: orbital motion of the moon around the Earth; day: spin of the Earth). The used pollen zones have to be dated by, e.g., counting annual layers (tree-rings, varves, speleothemes, ice-accumulation layers) or by radiocarbon (¹⁴C) dating calibrated by wood of known dendrochronological age [1023], [1025].

Since 65 million years ago (Ma), Earth's climate has undergone a significant and complex evolution, [...] This evolution includes gradual trends of warming and cooling driven by tectonic processes on time scales of 10^5 to 10^7 years, rhythmic or periodic cycles driven by orbital processes with 10^4 to 10^6 years cyclicity, and rare rapid aberrant shifts and extreme climate transients with durations of 10^3 to 10^5 years.

Since the last $\sim 800,000$ years, the Milankowitch cycles (Fig. 1.1) appeared to be the dominant pace makers or the main causes for the climate variability producing the 'interglacials' having a duration of normally 15,000–20,000 years within the $\sim 110,000$ years eccentricity cycles. However, according to the longer eccentricity cycle of 400 kyr having its lowest value³ since the beginning of the Holocene (Fig. 1.1) – the shape of Earth's orbit around the sun is near circular – this condition with similar energy (temperature) input will prevail for a longer period than usual, possibly causing a duration of the Holocene of up to 50,000 years, due to models of Loutre and Berger [653] and Paillard [799], and additionally triggered by anthropogenic greenhouse gases. The last interglacial, the Eemian, therefore would be no analogue for the Holocene.

Besides these orbital parameters, variabilities of the shorter sub-Milankowitch types are dominant during the Holocene and are only recorded in annual sequences like tree rings, ice-cores, sediment varves, and speleothemes.

1.2 The Reconstruction of Holocene Climate

There are few direct records of palaeoclimate prior to the last ~ 300 years such as meteorological measurements and historical documents about remarkable weather or climate extremes (floods, droughts etc.). Thus we have to rely on proxies – measurable parameters that provide quantitative and qualitative information about variables like temperature, precipitation and humidity in order to reconstruct the palaeoclimate. For the purpose of synchronization of these different multi-proxy data networks these reconstructions of global, regional or local variability patterns have to be based on independent and reliable time scales of annual resolution. This is mostly in contradiction to geological subdivisions based on principles of stratigraphy – bio- or lithostratigraphy – lacking a true calendar age model, hence being often diachronous.

The annual resolution is necessary to calibrate these multi-proxy data networks with meteorological data sets of the last 100 years with the prevailing climate regime. A climate regime like the North Atlantic Oscillation is a pattern of coherent climate fluctuations caused by the dynamics of the climate system and the complex interplay of its subsystems (ocean, atmosphere, cryosphere, lithosphere, biosphere) with important contributions by

³ Today's value is 0,0167 (ϵ) showing that eccentricity drops to zero because the orbit is circular ($a^2 - b^2 = 0$) ($\epsilon = \sqrt{a^2 - b^2}/a$)

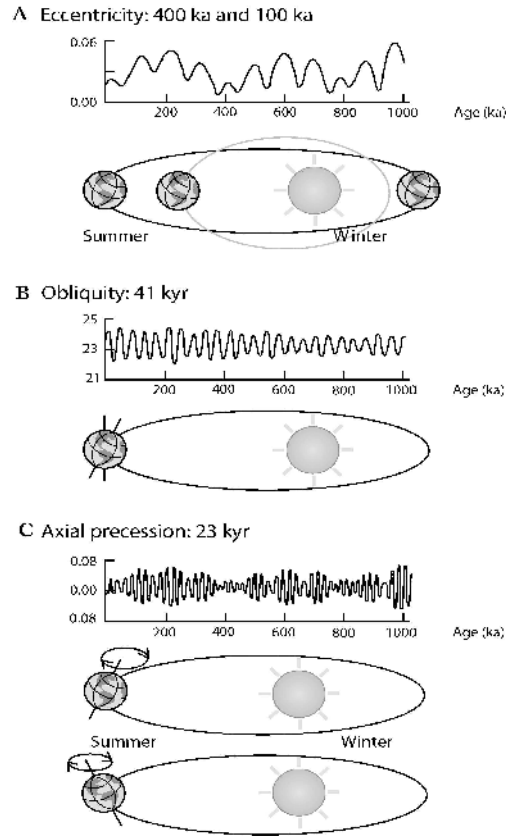


Fig. 1.1. Astronomical control of solar radiation. Orbital cycles that overlap and cause the Milankowitch-cycles (according to Zachos et al. [1178]).

solar irradiance changes, explosive volcanic eruptions and possible cosmic dust input.

Reconstructed palaeoclimate variabilities based on proxy-data, model generated variabilities as well as variabilities generated by proxy-data assimilated models are now available (e.g., [18], [692], [1143]). In Europe, instrumental data and historical climate observations [381], [1129] are available besides the proxy-data obtained from ice-cores, tree rings, lake sediments, marine sediments, corals, glacial deposits and marine terraces (e.g., this volume, [536]).

A short general description of archives (ice-cores, lake sediments, tree rings, corals, marine sediments) that are used for the reconstruction of the Holocene climate within the framework of the KIHZ-project follows [1178]:

Ice-cores:

Polar ice sheets and mountain glaciers are structured by snow accumulation

rates. They produce annual layers with subseasonal to decadal resolution. Dating is done by layer counting and ice flow models which are accurate back to 40,000 yr B.P. (Greenland). In general the record can be extended back to $\sim 440,000$ yr B.P. Ice-cores provide direct archives for the investigation of former atmospheres, containing CO_2 and methane (CH_4) in gas bubbles, Ca^{2+} from dust (wind strength), Cl^- , Na^+ from sea salt, pollen, $\delta^{18}\text{O}$ isotope ratios from precipitation related to surface temperatures, ^{10}Be and other cosmogenic isotopes tracing solar activity. Additionally, SO_4^{2-} originating from volcanic emissions permits the detection of teleconnections with distant archives.

Lake sediments:

Pure or mixed clastic, organic (diatomaceous gyttja) or evaporitic (marl, limestone, gypsum, salt) records in relation to the climatic regime are subdivided by annual layers (varves) often with seasonal resolution. Dating is done by varve counting of several parallel cores controlled by AMS ^{14}C -analysis, tephralayers and geomagnetic secular variation curves. Pollen distribution as well as diatom analysis indicate palaeoenvironmental and palaeoclimatic shifts and reveal information about temperature and moisture. $\delta^{13}\text{C}$ is temperature-sensitive. Geochemical parameters like TOC (total organic carbon), TIC (total inorganic carbon), biogenic silica, several oxides and biomarkers are indicators for changes in the local or regional environment (precipitation).

Varve thickness variations of e.g., organic varves reveal sunspot cyclicities whereas tephra layers are ideal markers for correlation of continental records with marine sequences. The time range is from present to 150,000 yr B.P. comprehending the last glacial cycle including the last interglacial. Varved records are also known from Permian or, for example, from young Precambrian times.

Tree rings:

Sequences of tree rings provide terrestrial records with annual resolution permitting the establishment of an 'absolute' chronology and the calibration of radiocarbon measurements for dating purposes back to $\sim 14,000$ yr B.P. These are most informative records for the last millennium if extreme climatic sites are chosen. $\delta^{13}\text{C}$, $\delta^{18}\text{O}$, $\delta^2\text{H}$ as well as ring width indices are indicative for rainfall and temperature. Moreover ^{14}C values reveal sunspot periodicities.

Corals:

Currently corals are the best tropical records showing seasonal changes in ocean systems with annual banding and monthly to weekly resolution. Dating is achieved by layer counting and ^{14}C and U/Th dating. The records cover presently the last 800 years. In several cases fossil corals have been used for climate reconstruction going back as far as 130,000 yr B.P., but they only cover short time intervals. Sea surface temperature and salinity can be derived from

oxygen isotopes and/or element ratios like Sr/Ca. Corals can also be used for the detection of sea level changes.

Marine sediments:

Marine sediments are the most prominent sediments covering much of the Earth's surface providing the basic records for the stratigraphy of the last 600 million years. Continuous records exist in the marine realm since Cretaceous times (~ 180 million years). However, the corresponding time resolution is mostly poor, i.e. 1000–100 years, in rare cases annual to seasonal resolution can be gained. Isotopes and element ratios ($\delta^{18}\text{O}$, $\delta^{13}\text{C}$, $\delta^{11}\text{B}$, $\delta^{13}\text{C}_{\text{org}}$, Cd/Ca, Mg/Ca, Sr/Ca) from planktonic and benthic microfossils (foraminifera, diatoms, radiolarians and ostracodes) reveal sea surface temperatures (SST), sea surface salinity (SSS), ice volume, atmospheric CO_2 and ocean circulation. Alkenone thermometry is used for the determination of the sea surface temperature. Radiocarbon dating for fossils is done up to $\sim 50,000$ yr B.P. Further dating is accomplished using correlations of the oxygen isotope curve to regular changes in Earth's orbit going back millions of years by orbital tuning (Marine Oxygen Isotope Stages [OIS]).

1.3 The Holocene Climate in Northwest Europe and Adjacent Areas

The evolution of the Holocene climate, as elaborated with records in Europe, depends on the climate regime of the North Atlantic Oscillation. Climate variability of the Holocene is, therefore, excellently and reliably documented in this region and provides a basic tool for a description worldwide. The compilation in this article is therefore restricted to Europe, but further information is available for the African–Arabian desert in [840], for the Americas by Grimm et al. [398], Fritz et al. [347] and for other areas in the world by Alverson et al. [18], Bradley [104].

Considering the variability pattern of multi-proxy data networks in space and time originating from various archives, it can be observed that the sensitivities of proxies are different. However, in general it is possible to describe trends, oscillations and aberrations, keeping in mind that the resolution of proxies varies from years, e.g. for trees, to ± 500 years, e.g. for lake level changes.

To gain a first overview about the climate evolution during the Holocene, relative temperature reconstructions of Greenland ice-cores are the most instructive. In Fig. 1.2 the climate variability is demonstrated by indicating relatively warm (O) and relatively cold (P) periods. Generally, the evolution can be described as follows:

Preboreal:

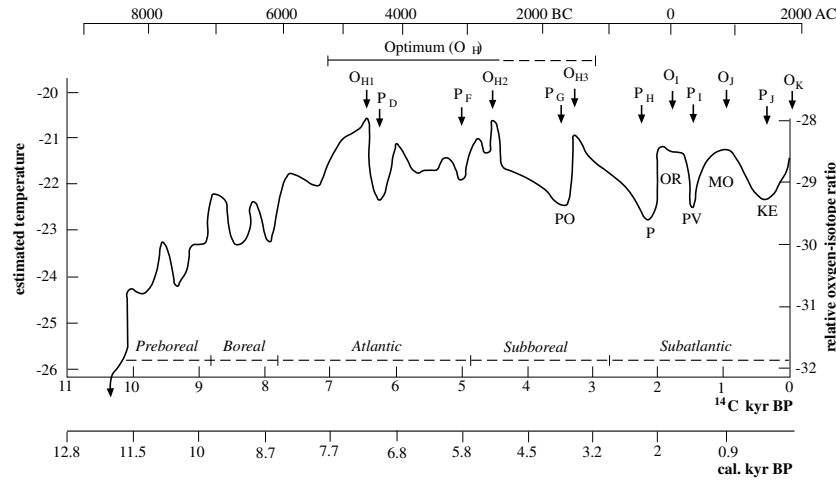


Fig. 1.2. Estimated temperature changes during the Holocene according to Greenland ice-cores after Schönwiese [944] and Dansgaard et al [217].

O: temperature optima, P: temperature pessima, PO: Piora-oscillation, P: main pessimum, OR: optimum of the Roman time; PV: pessimum migration time, MO: Medieval optimum, KE: Little Ice Age, OH₁, OH₂: maxima of the mid-Holocene 'climate optimum'.

The onset of this warm period following the cold Younger Dryas, lasted around 60 years. It is anticipated that the summers have been similarly warm like today, however with very cold winters due to the existence of some of the great ice shields.

Boreal:

In general, summers have been warmer than today with mild winters but single cold winters. Relatively low precipitation is dominant.

Atlantic:

This period is the warmest of the whole last glacial cycle, and therefore named 'optimum' with very mild winters. The last great ice shields disappeared and precipitation increased in the final stage (wet period).

Subboreal:

It is mainly a warm period similar to the beginning of the Atlantic, however, with a high variability but less precipitation than in the Subatlantic. At the beginning of the Subboreal, a very cold period, named Piora-oscillation, with glacier advances is documented.

Subatlantic:

The beginning is characterized by a pronounced cold period, the Main Pessimum. Mean annual temperatures in Europe were 1–2°C lower than today caused by cool summers with high precipitation and well known glacier advances.

- ‘Optimum’ of the Roman time: It was a period warmer than the ‘Medieval Optimum’ and several Alpine ‘passes’ were free even in winter. High precipitation is documented like in North Africa.
- ‘Pessimum’ of the Migration time: It was a cool and wet period with glacier advances in various areas.
- Medieval Optimum: This medieval warm period faced mean annual temperatures of 1–1.5°C higher than today, allowing wine-growing in NW-Europe. At the beginning it was drier, later on wetter.
- Transition period: It is a period of change from a warm period to the so-called ‘Little Ice Age’ with a pronounced cooling, high precipitation and increase of storm-tracks.
- Little Ice-Age: This cold period is characterized by a mean annual temperature of 1°C less than today with severe cold winters and extreme variations.
- Modern Optimum: It is a warm period that commenced around 1940 and that increases since.

Prentice et al. [843] have used pollen assemblages to reconstruct biomes and Huntley [483] has explained the Holocene climatic history recorded by changes in the overall vegetation patterns on the continent:

- At 10,000 ¹⁴C yr BP (ca. 9400 BC), forest cover was extensive only in mainland Europe, whereas most of the British Isles and Fennoscandia was occupied by treeless tundra-like vegetation.
- Much of the area occupied by forest biomes at 10,000 ¹⁴C yr BP (ca. 9400 BC) had vegetation that is either absent or uncommon in Europe at the present day, notably ‘cold mixed forest’ and ‘cold deciduous forest’ (for descriptions of the biomes see [842], [843]).
- Although ‘temperate deciduous forest’ was extensive by 8000 ¹⁴C yr BP (ca. 6950 BC), the forest patterns were quite different from those of the present day; ‘taiga’ was principally found only around the Urals, whereas Fennoscandia was occupied mainly by ‘cold mixed’, ‘cool conifer’ and ‘cold deciduous’ forests.
- The extent of ‘temperate deciduous forest’ was greatest in the mid-Holocene, ca. 6000 ¹⁴C yr BP (ca. 4850 BC); since that time it has been reduced in extent along its northern, eastern and southern limits. ‘Taiga’ has become extensive in Fennoscandia since the mid-Holocene, its expansion westwards being paralleled by the expansion of ‘cool mixed forest’ in eastern Europe and in the Alpine region.

- Although ‘broad-leaved evergreen/warm mixed forest’ has been present locally in southern Europe since the early Holocene, it has expanded considerably in the late Holocene, replacing ‘temperate deciduous forest’ in many parts of southern Europe.
- The expansion of treeless grassland and heathland vegetation in western Europe since the mid-Holocene is reflected by an increasing incidence of misclassification of pollen spectra from these regions as representing the ‘tundra’ biome.

These vegetation changes reflect independent changes in three principal aspects of the climate, the annual temperature sum, the minimum temperature and the moisture availability. Additionally, they document that pollen sequences cannot be used simply for climate reconstructions since the human influence increased.

Huntley et al. [485] have documented this human influence in Europe and the shifts of the cultural periods from South to North. Further on Wefer et al. [1143] present several papers that demonstrate the Holocene climate evolution using evidence from different archives, ice shields [423], marine sediments [545], lake sediments [754], lake hydrology [431], Baltic sea cores [274], vegetation changes [483], glacier fluctuations in Norway [761], palaeoenvironmental changes [485], sea level change [61], tree ring narrowest ring events [39].

Huntley et al. [485] have published a compilation of several of these information as “principal Holocene climatic changes and events, and major changes in human settlement patterns and activities”.

1.4 Holocene Climatic Trends, Fluctuations and Events

Huntley et al. [485] have described the Holocene palaeoenvironmental history of NW-Europe by distinguishing between long-term trends, fluctuations of 100–1000 years and multi-annual to multi-decadal events.

Long-term trends

Bore hole temperature logs from Greenland ice cores [485] revealed mean annual temperatures integrated over years to decades peaking around 5000 yr B.P. at a value of ca. 5°C higher than the present day value. A decrease started ca. 4100 yr B.P. and reached its minimum around 2200 yr B.P. Afterwards, smaller fluctuations are indicated with the Little Ice Age being some 1–1.5°C cooler than today whereas the MWP (*Medieval Warm Period*) was as warm as the present day.

This is not reflected in some of the new $\delta^{18}\text{O}$ Greenland ice core records and shows no early or mid-Holocene ‘thermal maximum’ which is documented in lake sediments [754] and other archives. The early to mid-Holocene ‘thermal maximum’ is, however, even documented in the altitudinal tree limit in Scandinavian mountains being 300m above the present tree limit [72]. Summer

temperatures have been higher than at present, an optimum occurred around 6000 to 5000 yr B.P. with values 1.5°C (± 0.5) higher than today.

In contrast to the tree limit maxima, glacier moraines (advances) mark the cool events (temperature minima), for example, in Norway [761], reflecting even the Bond cycles [90] and the Little Ice Age (Fig. 1.3).

Speleothemes from northern Norway reveal a Holocene temperature maximum (2.5°C warmer) around 9500 to 7500 yr B.P. [485]. This seems to support the evidence that the snow line in the Swiss Alps is higher at present than it was between 8000 and 6000 yr B.P.

The early to mid-Holocene ‘thermal maximum’ is a general feature in Europe and documented in Greenland, Scandinavia, Central Europe and the Atlantic summer SST (sea surface temperature) [103] and extends to the Cariaco-Basin (Fig. 1.3).

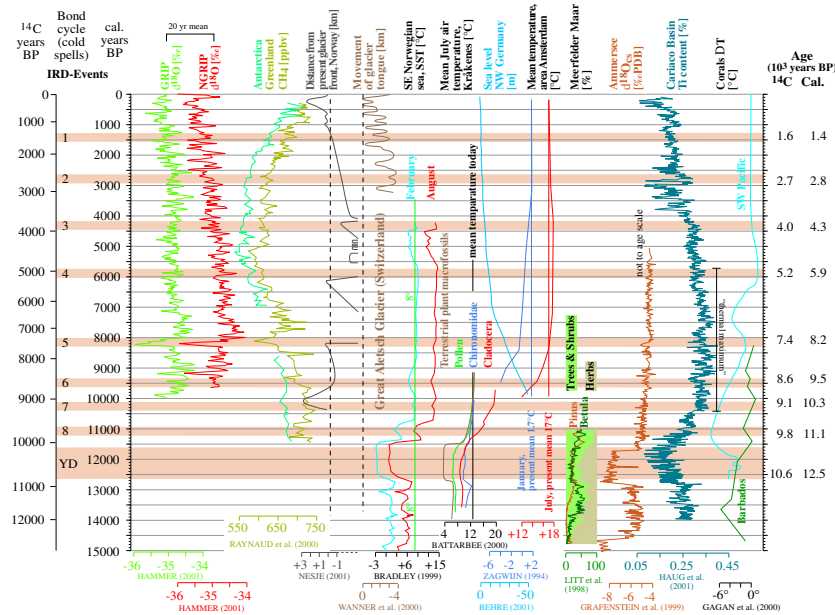


Fig. 1.3. Cold events marked by the Bond cycles 1–8 [90] due to ice rafted debris in the Atlantic and the Younger Dryas to identify possible cold spells in proxy records of different archives and climatic regions.

To explore the spatial pattern in the Holocene palaeoclimate of NW-Europe, only pollen data, lake level data and sea surface conditions are available to a sufficient degree. Huntley and Birks [486] and Huntley and Prentice [487] compiled pollen analytical data which revealed summer temperatures higher than present day temperatures north of 50°N with its maximum in the maritime area of NW-Europe for the time slice around 6000 yr B.P. In

winter, the temperatures in this region were lower than in the Mediterranean area today but warmer than presently north of the Alps. Moisture was higher in the southern North Sea area and in southern Europe but lower than today in other areas. A similar reconstruction for NW-Europe based on pollen assemblages was developed by Zagwijn [1179] (Fig. 1.3).

Fluctuations (100–1000 years)

The overall long-term trends of the Holocene are often superimposed by fluctuations in the order of 100 to 1000 years. One pronounced cooling event is found around 8200 yr B.P. [40] indicated in many proxies in Europe (e.g., [840]) and in the Cariaco-Basin [439] (Fig. 1.3). Additionally, 250–300 years lasting cool oscillations within the Preboreal and Boreal of the varve sequence of Holzmaar (Eifel) are equivalent to the Bond cycles 8 and 7 interpreted as cooler and wetter climate conditions [107]. Many glacier advances show fluctuations like the Little Ice Age during the Holocene.

The history of the Great Aletsch Glacier, for instance, appears to be representative of the history of the Alpine glaciers generally documenting the interruption of the Medieval Warm Period (MWP) by small short-lined advances [1129] (Figs. 1.3, 1.4).

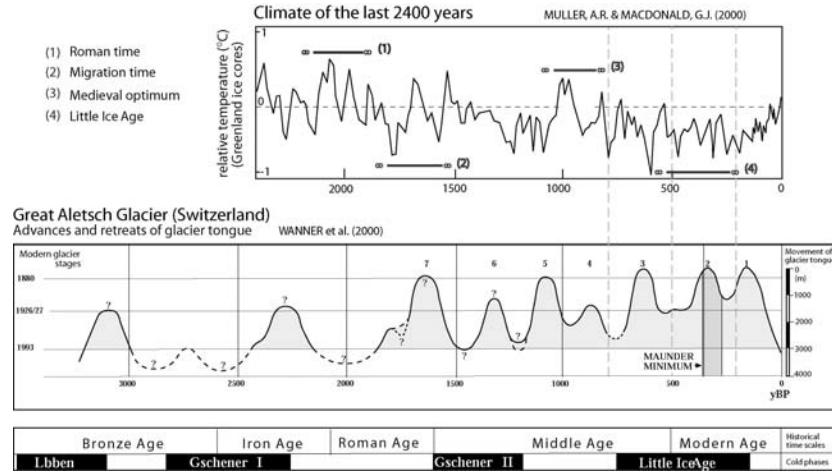


Fig. 1.4. Advances and retreats of the Great Aletsch Glacier tongue [1129] in comparison to relative temperatures of Greenland ice cores after [743]

Multi-annual to multi-decadal events

Events with a duration of less than a century are documented in records with annual resolution like ice cores, varved lake sediments and tree rings. These are mostly extreme cold events, some of them are apparently global in extent. Huntley et al. [485] state: These events (a decade in duration) are

marked as ‘narrowest ring events’ and are seen in tree ring sequences from many different parts of the world (e.g., Ireland, Germany, Scandinavia, North America, South America). A special case is the event assigned to the year 1410 B.P. which is possibly caused by volcanic eruptions⁴ and/or cosmic dust input. According to Baillie [39] the 1410 yr B.P. narrowest ring event corresponds to a period where Earth was at increased risk of meteoritic bombardment. A similar abrupt climate event was described by van Geel et al. [1086], [1087] in several archives in the Netherlands and documented in a sharp rise of ^{14}C around 800 cal. BC.

Time sequence analyzes of ice core data, ^{14}C date of tree rings as well as varve thickness variations reveal major periodicities of 11, 22, 88, 210, ~ 550 , ~ 900 , 1050 years [1118] documenting a variability that is not pronounced in other proxies, for example, within the pollen spectra. The cycles with 8 and more decades in duration are possibly responsible for the cooling events like the Spörer, Wolf and Maunder Minimum exhibiting the response to solar irradiance changes⁵.

Acknowledgements

I would like to thank two anonymous reviewers for their comments essentially improving the article by helpful corrections.

⁴ Since 818 B.C. massive volcanic eruptions have depressed the temperature in Greenland around 2°C . According to Briffa et al. [121] the northern hemisphere experienced temperature decreases around $0.2\text{--}0.4^\circ\text{C}$ for each volcanic eruption with the extreme at 1601 (0.8°C). In mid-latitudes it causes a warming of the winter and a cooling of the summer temperatures [39].

⁵ The response to solar forcing is calculated since 1600 up to $0.2\text{--}0.5\%$ which corresponds to a temperature less than 0.6°C . In gross, $20\text{--}35\%$ of the reconstructed climate variability for this time window is caused by solar forcing [39].