Climate, vegetation, and CO_2 dynamics during the Eemian interglacial (MIS 5e) in Europe

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1. Introduction

We are currently in an interglacial period, the Holocene, and it is predicted that this current interglacial will be exceptionally long (Berger and Loutre, 2002). In order to get a better understanding of our warm climate and its future, past interglacials are particularly relevant. Climate variations of the last 3 million years are characterized by glacial-interglacial cycles, which are generally believed to be driven by astronomically induced insolation changes. When a change in orbital parameters takes place this affects summer insolation at high northern latitudes and thus influences temperature conditions at Earth's surface. These changed temperature conditions lead to a reorganization of the global carbon cycle and on its turn amplifies the temperature response.

During the current Holocene interglacial, the cultural evolution of humans has accelerated considerably. This increased civilization has probably only been possible under the mild and relatively stable climatic conditions, with little land ice, and largely elevated temperatures at mid-high latitudes.

The palaeoclimatic record teaches us that these relatively mild and stable conditions cannot be taken for granted; all past interglacials terminate after a few thousand to a few tens of thousand years (Tzedakis et al., 2012). Moreover, the relatively mild and stable climate that we experienced the past 11,000 years has shown to be only prevailing of no more than 15 % of the last half million years (Kiefer & Kull, 2007). In fact, on a 10,000-year timescale interglacials appear to be rather unstable. For instance, late Holocene climate fluctuations on a shorter timescale (millennial to decadal) such as the Little Ice Age (Mörner, 2012) associated with the Maunder Minimum (Yamaguchi et al., 2010), and the abrupt 8.2 kyr cooling event (Alley et al., 2005), prove that the current interglacial climate has not been entirely stable and responds to even subtle changes in radiative forcing.

Given the fact that interglacial climates are not really stable, and it is likely for the global population to experience climatic fluctuations in the near future, with either minor or major impact on our socioeconomic system, it is of major importance to learn more about the sensitivity, thresholds and feedbacks of the climate-environment system, in order to develop possible adaptation and mitigation strategies. One of the possibilities to study these mechanisms is by studying the climate and environment of past periods where boundary conditions were highly similar to today. The quasi-cyclic reoccurrence of interglacials during the late Pleistocene (Petit et al., 1999) gives the opportunity to get more insight in the climate dynamics during different interglacials, and to study similarities and dissimilarities.

During previous years and decades scientific research focused mainly on the large-amplitude climate variations during glacials (Sejrup et al., 2009; Marks, 2012), and between glacials and interglacials (Sigman & Boyle, 2000; Fujita et al., 2010), thereby disregarding interglacial global change issues. One of the reasons is that the density and temporal resolution of records we had access to are insufficient to deduce reliable conclusions on high frequency or regional-scale climate variability during past interglacials. An important target for the present palaeoclimate community is thus to reconstruct and simulate centennial to decadal scale variability of interglacial climate.

The Eemian interglacial is the penultimate interglacial and approximately corresponds to marine isotope stage (MIS) 5e, which was dated to c. 130,000 - 116,000 years before present (BP) (Kukla et al., 2002). There is a variety of evidence that indicates conditions during the Eemian interglacial were as warm as during the current interglacial (GRIP

Members, 1993). At the end of the Eemian, the climate changed to cold glacial conditions. Similar transitions have been linked to changes in the orbit of the Earth around the Sun.

Since the Eemian interglacial is the most recent example in the geological record of a termination of a major warm episode, temperatures during the Eemian were comparable to the present interglacial, and the present orbital changes are similar to that of the end of the last interglacial (Kukla & Gavin, 1992), the Eemian interglacial is a prime target for palaeoclimate reconstruction. Moreover, the last interglacial-glacial transition provides the nearest natural analogue for the end of the present interglacial, and hence understanding the nature, timing and environmental consequences of climatic change at the end of the previous warm stage is an essential prerequisite for the development of models of future climatic scenarios in which human influence is likely to be an additional and possibly crucial factor.

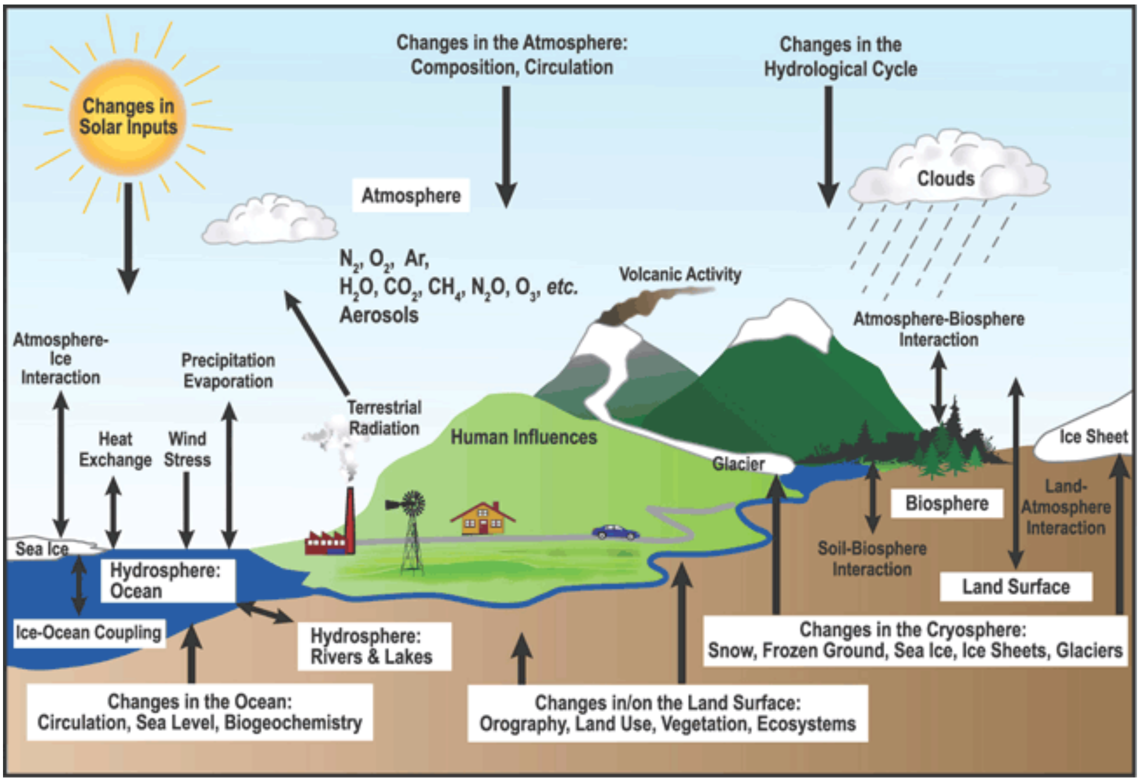

Figure 1. Schematic representation of the components of the climate system and their processes and interactions. Figure taken from IPCC Working Group I (2007).

2. The climate system

2.1 Components of the climate system

When interpreting palaeoclimate records often the questions arises what drives or forces the climate, which factors are external and which are internal. The classic definition of climate stated by Hann (1883) is the sum of all meteorological phenomena that characterize the mean state of the atmosphere. However, for understanding climate dynamics this definition is too restrictive, since the mean state of the atmosphere is affected by more than just meteorological phenomena. The atmosphere's mean state and its variations depend on dynamics of the climate and on the interaction between the various components of this climate system (Fig. 1).

The main components of the climate system are the atmosphere, hydrosphere, cryosphere, marine and terrestrial biosphere, pedosphere, and the lithosphere. The atmosphere component is the envelope of gases that surrounds the Earth. The hydrosphere consists off all water bodies, mainly the ocean, but also rivers, lakes and rain- and groundwater are part of this climate component. The cryosphere is the portion of the Earth's surface where water is in solid form, and thus encompasses all ice sheets, glaciers, sea ice, snow, and permafrost. All living organisms that inhabit either water or on land define the marine and terrestrial biospheres. The pedosphere consists of the Earth's soils and rocks, and the lithosphere is the upper mantle of the Earth's interior. The climate system in its total is powered by the solar radiation. The climate system also involves processes at work within it, such as precipitation, evaporation and winds.

As shown in figure 1 these components and processes interact with each other, hence the climate system evolves in time under influence of its own internal dynamics but also by changes in external forcing factors.

2.2 External climate forcing

Four fundamental kinds of external climate forcing exist in the natural world; changes in the strength of the sun, Earth-orbital changes, tectonic processes, and the effect of humans on climate.

As can be seen in figure 1 the climate system in its total is powered by solar radiation, and therefore changes in the strength of the Sun affect Earth's climate system, by changes in the amount of solar radiation that reaches the Earth. The Sun becomes steadily hotter, and therefore the strength of the Sun has also steadily increased which leads to a slow increase in solar radiation, about 10 % per billion years. In addition, shorter term variations that occur over decades and or longer have been found (Lean et al., 1995; Bard et al., 2000).

Earth-orbital changes result from variations in Earth's orbit around the Sun. Changes in obliquity, eccentricity, and precession alter the amount of solar radiation received on Earth by season and by latitude. Orbital changes occur tens to hundred of thousands years.

Tectonic forcing on the Earth's climate is driven by tectonic processes generated by Earth's internal heat, and affect Earth's surface by means of processes that alter its basic geography. Among tectonic forcing are the spreading of oceanic crust and subduction of continental plates, which are processes that operate very slowly over millions of years. Another important example of tectonic forcing is volcanic eruption. Volcanic explosions affect the chemistry of the atmosphere by expelling sulphate aerosols in the atmosphere. If this happens in a substantial amount, this has a negative influence on the radiation balance of the Earth, a few years after the explosion a cooling on global scale can be detected (Jones & Mann, 2004). The frequency of outbursts determines the long-term climate variation (Briffa et al., 1998).

Humans inhabit the terrestrial realm, and it could be said that they are part of the terrestrial biosphere component of the climate system, however, they are usually regarded as an external forcing, because human activity, economics, culture, and values cannot be accessed by tools to simulate natural components of the climate system (Claussen, 2007). Humans affect the climate system by changing Earth's land-surface structure, and by altering the chemical composition of the atmosphere. For instance, by burning fossil fuels, humans have emitted a large amount of CO_2 into the atmosphere. The amount of other greenhouse gases also have been altered by humans, and furthermore, humans have created new chemical substances that even have a stronger potential as greenhouse gas compared to carbon dioxide and methane. Another anthropogenic activity that leads to a change in the Earth's climate system is land use. For instance by deforestation, the Earth's surface reflects more solar radiation (i.e. albedo).

2.3 Climate system feedbacks

An important kind of interaction in the climate system is the operation of feedbacks, processes that alter climate changes that are already underway, either by amplifying them (i.e. positive feedbacks) or by suppressing them (negative feedbacks). A change in the climate by external forcing will consist of many different responses among the various internal components of the climate system. The changes in some of these components will then further alter climate through the action of feedbacks.

For example, an increase in greenhouse gases in the atmosphere caused by anthropogenic forcing leads to increasing annual mean temperature, which on its turn leads to more evaporation. This increased evaporation is associated with an increase in water vapour, one of the most important greenhouse gases, and leads to even a further increase of annual mean temperatures (IPCC, 2007). Therefore, the initial warming is amplified by a positive feedback mechanism.

Another important player in the climate system and in feedback mechanisms is the ocean. For instance, the ocean circulation (thermohaline circulation) is mainly driven by salinity and sea temperature, since these factors determine the density of seawater. Both salinity and sea temperatures change when global temperatures change, which leads to a change in the ocean circulation. The thermohaline circulation on its turn is responsible for bringing heat from the tropics to the high latitudes, and thus has an impact on the global climate. This is one of the most important feedback mechanisms in climate.

3. Introduction to the Eemian interglacial

3.1 History of Eemian research

The last interglacial is among the best-documented warm periods that are similar to the current warm Holocene. This warm period is often referred to as marine oxygen-isotope stage 5 (MIS 5), or as the Eemian. However, the exact timing of the Eemian interglacial in Europe is still under debate, as the following short history of Eemian research reveals.

The first evidence for this warm period was found c. 160 years ago in European sediments. Harting (1852) was the first to apply the term Eemian as a stratigraphic unit, when he was examining the subsoil of the Amsterdam and Amersfoort areas, and noticed the consistent occurrence of clays and sands that were rich in Mediterranean and Lusitanean (Portugese/Spanish) mollusc and diatom fossils (Bosch et al., 2000). Following this, pollen analysis from Danish sediments, revealed that during the Eemian there was a widespread *Quercus* (oak), *Corylus* (hazel), and *Carpinus* (hornbeam) woodland (Jessen and Milthers, 1928). In 1961, Zagwijn showed a similar expansion of deciduous forest in Germany.

Figure 2. The stacked marine isotope record for the past 130,000 years (after Martinson et al., 1987).

Emiliani (1955, 1966) analysed oxygen isotope data and made a temperature reconstruction for the central Caribbean surface waters around the same time, and this reconstruction formed the basis of the marine oxygen isotope stratigraphy. The temperatures were deduced from oxygen isotope data. Seventeen isotopic stages altering between warm and cool periods were identified, corresponding to glacials and interglacials. Marine isotope stage 5 (MIS 5) was found to be corresponding to the last interglacial (Fig. 2). The stratigraphy was later refined and based on the correlation between marine and terrestrial records, it was found that only the first $\delta^{18}O$ minimum values had to be considered as the equivalent of the Eemian as identified on land (Shackleton, 1969), which was named MIS 5e. Subsequent Turon (1984) established the first direct though low-resolution correlation between pollen and benthic isotopic data in a core from the northwestern Iberian margin, and this research confirmed that the Eemian corresponded only with the first minimum values in the oxygen isotope record. The Eemian was found to last for 11,000 years between 126 and 115 ky BP.

It is known that MIS 5 was a period of minimum ice volume that expanded from c. 130 to 75 ky BP (Imbrie et al., 1989). MIS 5e was the first and warmest interval, characterized by sea levels 0 to 6 meter higher than those of the present day (Stirling et al., 1998; Schellmann and Radtke, 2004). This was followed by two cold periods in the ocean (MIS 5d and MIS 5b) alternating with two warm ones (MIS 5c and MIS 5a).

Direct correlation between high resolution isotopic and pollen data shows that in western Europe, MIS 5 encompasses the Zeifen interstadial, a short stadial event before the Eemian, followed by four cold/warm cycles; Mélisey I/Saint Germain I1, Montaigu/Saint Germain Ic, Mélisey II/Saint Germain II, and Stadial I/Ognon I (Sánchez Goñi et al., 1999, 2005).

3.2 Timing and duration of the last interglacial

It is well accepted that climate variations during MIS5 were largely determined by insolation, however other factors and feedback mechanisms played an important role in determining the timing and duration of the last interglacial. A continuous debate concerns the timing and duration of the full last interglacial.

The oxygen isotope signal from the deep-ocean floors is the standard for global correlation. The large number of isotopic records that are now available for the last 130,000 years have been normalized to produce a generalized or standard curve with a timescale derived by orbital tuning (Fig. 2). The isotopic record is essentially a reflection of changes in global ice volumes, and thus indicative for past climate conditions. From this curve it becomes clear that there is a marked shift from isotopically heavy to lighter values around 130 ky BP, that characterizes the transition from glacial to interglacial conditions (Termination II). MIS 5e is marked by the peak of isotopically light values. After this peak isotopic values gradually shift to heavier values throughout MIS 5d to MIS 3. MIS 5c and MIS 5a are episodes with isotopically lighter values, indicative for warmer conditions. From this record MIS 5 in its total has a duration of c. 55,000 years between c. 130 to 75 ky BP. MIS 5e started around 130 ky BP and ended around 115 ky BP, which reveals a total duration of c. 15,000 years.

Dates that are in broad agreement with this chronology have been obtained from the Vostok ice-core (Petit et al., 1990), from coral reef sequences and marine speleotherm records from both the Caribbean and the Pacific (Harmon et al., 1983; Bard et al., 1990, Lundberg & Ford, 1994), and from the last interglacial soil in the Baoji loess-soil sequence on the Loess Plateau of China (Ding et al., 1994).

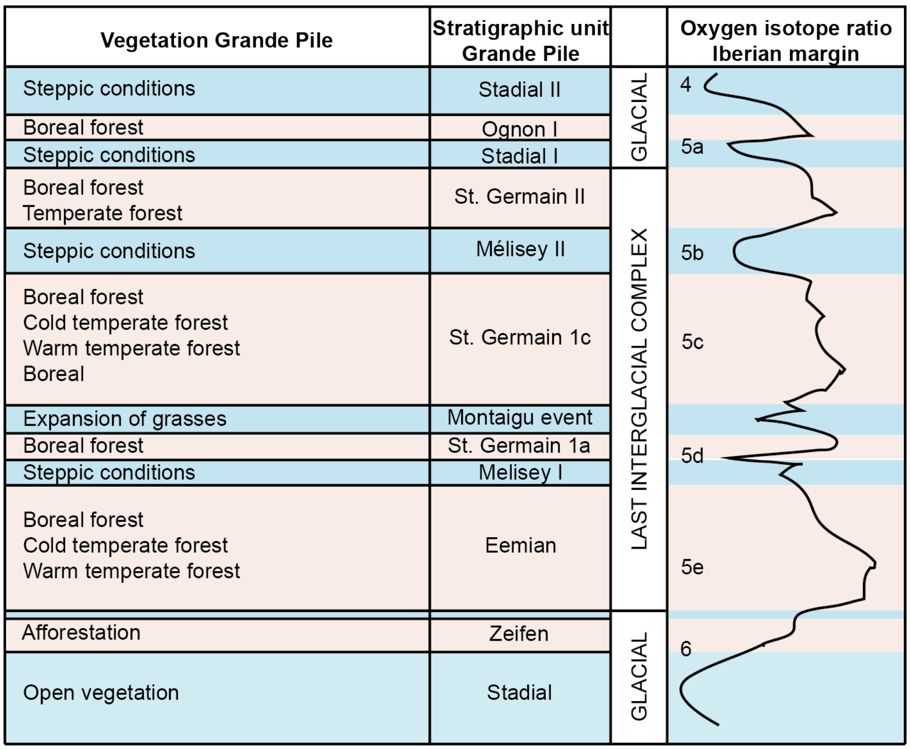

Figure 3. Vegetational succession and recognized units of pollen diagram x at Grande Pile (Woillard, 1978) and the planktonic isotopic curve from the northwestern Iberian margin (Sánchez Goñi et al., 1999).

3.3 *Eemian climatic variability*

The Eemian interglacial has been traditionally seen as a period characterized by a rather uniform warm climate, as observed for the Holocene and predicted by astronomical parameters. This idea of climatic stability was contradicted when Dansgaard et al. (1993) identified high amplitude changes in the atmosphere of Greenland as detected by the GRIP isotopic record (Fig. 4). An even further subdivision of MIS 5 was suggested by dividing MIS 5e into three principal warm substages (MIS 5e5, MIS 5e3, and MIS 5e1), separated by two sustained cool periods (MIS 5e4 and MIS 5e2) with duration up to 5000 years (GRIP Members, 1993).

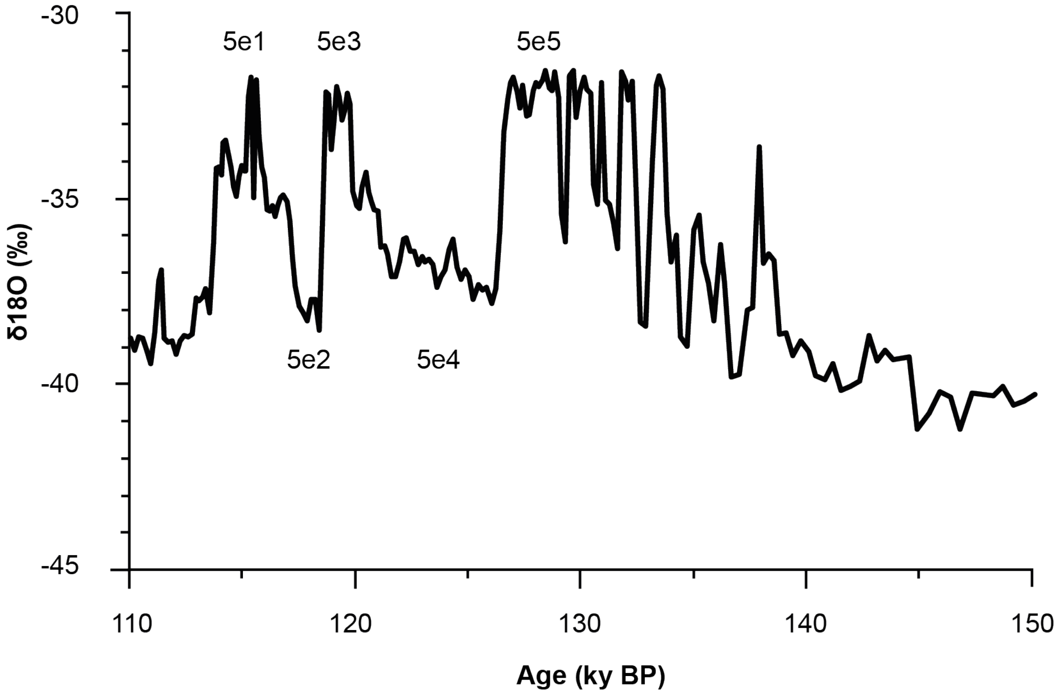

Figure 4. Oxygen isotope record of the Eemian (MIS 5e) section of the GRIP Summit core (after GRIP Members, 2003).

These short cold periods, however, are neither recorded in the central North Atlantic (McManus et al., 1994), nor by high-resolution terrestrial and marine records from western Iberian margin deep sea cores (Sánchez Goñi et al., 1999), nor in European pollen and diatom archives (Cheddadi et al., 1998; Frogley et al., 1999; Rioual et al., 2001). The high amplitude changes in the GRIP isotopic signal during MIS 5e are now believed to be an artefact of iceflow (Chappellaz et al., 1997; Landais et al., 2003). The first major cold episode in the GRIP record associated with substantial iceberg discharges was around 117 – 110 ky BP, and coincides with the Mélisey I steppic period on land (Shackleton et al., 2002, 2003; Sánchez Goñi et al., 2005).
However, Eemian cooling was also detected in the North Atlantic (Cortijo et al., 1994), and this was associated with southward displacement of the Polar Front, caused by a reduction of the upper North Atlantic deep-water ventilation (Maslin et al., 1998). Furthermore, several weak cooling episodes during the Eemian were also detected in a high Alpine stalagmite, and it is thought that these episodes were caused by changes in the solar activity (Holzkämper et al., 2004). The NGRIP record indicates warmer temperatures from 123 to 120 ky BP, followed by a cooling from 120 to 115 ky BP

(NGRIP Members, 2004). This cooling episode coincides with the first southward displacement of vegetation belts in western Europe and of the North Atlantic current in response to decreasing summer insolation at higher latitudes (Müller and Kukla, 2004).
In conclusion, it seems that the Eemian climate may have been relatively stable, but there are some short cooling periods visible in several palaeorecords, indicating periods of climate instability. The next three chapters will take a closer look on vegetation, temperature, and CO_2 dynamics during the Eemian in Europe.

4. Vegetation and temperature dynamics during the Eemian

4.1 Introduction

Correlations in terms of vegetational facies can easily be made because of a marked uniformity of vegetational development across much of northern Europe during the Eemian. These pollen-zonation schemes, whether local or regional, are essentially based on successive acme zones for different temperate tree taxa. However, because a precise dating for the Eemian is absent, the exact temporal relationships between parallel pollen sequences along a transect across Europe can not be determined with certainty, because of the leads and lags of rates of plant migration and establishment.
Several studies concerning vegetational succession during the Eemian have received much attention, because they demonstrate a more or less complete vegetational sequence through the interglacial. One of these studies concerns the Grande Pile site in the Vosges mountains in France (Woillard, 1975; De Beaulieu & Reille, 1992), which was not only revealing a complete vegetational succession throughout the Eemian, but also, for the first time, provided a convincing correlation with the deep-ocean oxygen isotope record and clearly provided the equivalence of the Eemian with MIS 5e of the marine stratigraphy. Another important research concerned sequences from Amersfoort and Amsterdam in the Netherlands (Zagwijn, 1961; Cleveringa et al., 2000; Van Leeuwen et al., 2000). Furthermore, a number of sites in northern Germany have yielded pollen diagrams covering the complete last interglacial cycle (Hahne et al., 1994; Litt, 1990). Also in Denmark and Poland pollen diagrams covering the full Eemian interglacial have been analysed (Andersen, 1965, 1966; Mamakowa, 1989). Table 1 gives an overview of the pollen zonation from a few of these studies, zone names were based either on specific taxa, or to most abundant characteristic taxa.

4.2 Vegetation succession during the Eemian interglacial

The transition from glacial to interglacial conditions is marked in the pollen records by the expansion of woodland and forest, caused by the warming at the beginning of the Eemian. The increase of *Betula* (tree birches) and *Pinus* (pine) pollen has a dominant role in this afforestation.
In almost all records *Ulmus* (elm) was the first temperate deciduous taxon to appear and expand, followed by *Quercus* (oak). *Quercus* then became dominant and after a while *Corylus* (hazel) took over. During this early-temperate phase *Alnus* (alder) was often abundant as well. The major regional differences during this early-temperate phase were caused by the presence of *Taxus* (yew) and *Tilia* (lime). *Taxus* was present in sites in France, Denmark and Poland, but sparser or even absent in sites in other sites. This difference in presence of yew is thought to be caused by local geological conditions, and also by failure to determine this taxon adequately, especially in older palynological

Table 1. Eemian pollen-zonation schemes for several areas of western and north-central Europe, showing formal zone names (in Bold), and characteristic pollen taxa (after Turner, 2000**).**

Turner & West (1968)	**Massif Central** De Beaulieu & Reille (1992)	**The Netherlands** Zagwijn (1961)	**Northern Germany** Menke & Tynni (1984)	**Poland** Mamakowa (1989)
Early glacial	**22 NAP** *Artemisia*, Chenopodiaceae	**EW1 NAP** Ericales	**WF1 NAP** Ericales	**EV1** Graminae-*Artemisia*-*Betula nana*
Post-temperate	**21 *Pinus*** *Pinus-Picea-Betula*	**E6b *Pinus*** *Pinus-Betula*	**VII *Pinus***	**E7 *Pinus***
	20 *Picea* *Picea-Carpinus-Abies*	**E6a *Picea*** *Pinus-Picea-Abies-Alnus*	**VI *Pinus-Picea-Abies***	**E6 *Picea-Abies-Alnus***
Late-temperate	**19 *Abies*** *Carpinus-Corylus-Abies-Picea*	**E5. *Carpinus*** *Pinus-Picea-Abies-Alnus*	**V *Carpinus-Picea***	**E5 *Carpinus-Corylus-Alnus***
	18 *Carpinus* optimum *Carpinus-Corylus-Quercus*			
	17 *Carpinus* *Corylus-Quercus-Carpinus-Taxus*			
Early-temperate	**16 *Taxus*** *Quercus-Corylus-Taxus-Fraxinus*	**E4b *Taxus*** *Quercus-Corylus-Ulmus-Fraxinus-Tilia*	**IVb *Corylus-Taxus-Tilia***	**E4 *Corylus-Quercus-Tilia***
	15 *Corylus* *Corylus-Quercus-Fraxinus*	**E4a *Corylus*** *Corylus-Quercus-Alnus*		
		E3b *Quercus-Corylus* *Quercus-Corylus*	**Iva *Quercetum* mixtum-*Corylus***	
	14 *Quercus* *Quercus-Ulmus-Fraxinus-Corylus*	**E3a *Quercus*** *Quercus-Ulmus-Fraxinus*		**E3 *Quercus-Fraxinus-Ulmus***
		E2b *Pinus-Quercus* *Pinus-Quercus-Alnus*	**III *Pinus-Quercetum* mixtum**	
		E2a *Pinus-Ulmus* *Pinus-Ulmus*		**E2 *Pinus-Betula-Ulmus***
Pre-temperate	**13 *Betula*** *Betula-Pinus-Ulmus*-NAP	**E1 *Betula-Pinus*** *Pinus-Betula*	**II *Pinus-Betula***	**E1 *Pinus-Betula***
			I *Betula*	
Late-glacial	**12** *Artemisia*-Poaceae-*Juniperus*	**LS**	**SSC** *Hippophae-Juniperus*	**LG MPG** NAP-*Hippophae-Juniperus*

investigations (Turner, 2000). The distribution of *Tilia* during the early-temperate phase of the Eemian does, in contrast to *Taxus*, show a geographical pattern. Lime was rare or even absent in western European areas like France and Britain, and only present in low percentages in the Netherlands, but much higher values at some sites in northern Germany, Denmark, and Poland (Andersen, 1975; Menke & Tynni, 1984; Mamakowa, 1989), where lime must have dominated the forest for a period.

During the late-temperate phase of the Eemian interglacial *Carpinus betulus* (hornbeam) had its entrance into the forest and rapidly expanded, hence though other temperate trees like *Quercus, Fraxinus* (ash), and *Taxus*, were still present, hornbeam became dominant in the forest.

The post-temperate phase of the Eemian is characterized by the development of boreal forest, dominated by *Pinus, Betula,* and *Picea.* Furthermore, the presence of *Calluna* pollen and *Sphagnum* spores indicate the presence of heath and bog, probably caused by high rainfall and leaching and acidification of soils.

4.3 Temperature development during the Eemian as indicated by vegetation

Aalbersberg & Litt (1998) did an extensive study to temperature changes during the Eemian as indicated by vegetational succession based on 106 of the many vegetation reconstructions that were carried out in Europe. By using the modern analogue method, plant species present during the Eemian provide minimum mean monthly temperatures and daily maximum or minimum temperatures for the last interglacial.

The early-temperate phase of the Eemian, marked by Aalbersberg & Litt (1998) as the Eemian *Pinus-Quercetum*-mixtum-*Corylus* phase, is considered to be the period in which the climate had reached its full temperate character. During this phase most of northwestern Europe had a subcontinental climate regime, and it is commonly regarded as the time when summer temperatures reached their highest values. During this phase a northwest to southeast temperature gradient was inferred (Fig. 5), with values up to 18 °C for the minimum mean July temperature in northwestern Europe, and high values up to 20 °C in the southeast part of Europe. Almost the entire transect through Europe reveals minimum mean January temperatures of -2 °C. Only the western part of the Netherlands, northwestern Germany, and England reveal values around 0 °C. Hence, if a gradient existed it was directed southeast to northwest, opposite to that of the minimum mean July temperatures.

The late-temperate phase of the Eemian, or *Carpinus-Picea* phase, is suggested to have a more oceanic climate, based on the temperature amplitude between summer and winter inferred temperatures. Although summer temperatures seem to be lower compared to the early-temperate phase, the climate in general was milder, especially during winter. During this phase minimum mean July temperatures of at least 18 °C are reached everywhere throughout Europe. Minimum mean January temperatures also show a uniform pattern with values of 0 °C and 1 °C throughout Europe.

The post-temperate phase of the Eemian, or *Pinus-Picea-Abies* phase, marks the deteriorating climate towards the Weichselian glacial, the climate is changing to more boreal conditions. During this phase, mean minimum July temperatures have dropped to values of 15 °C and 16 °C. A gradient existed from the south-southeast to north-northwest. The mean minimum January temperature does not show a real pattern, since several warm species still occur, while in others those species are absent, pointing to colder conditions. It is thought that overall the temperature during the coldest month must have been around 0 °C for the entire transect throughout Europe.

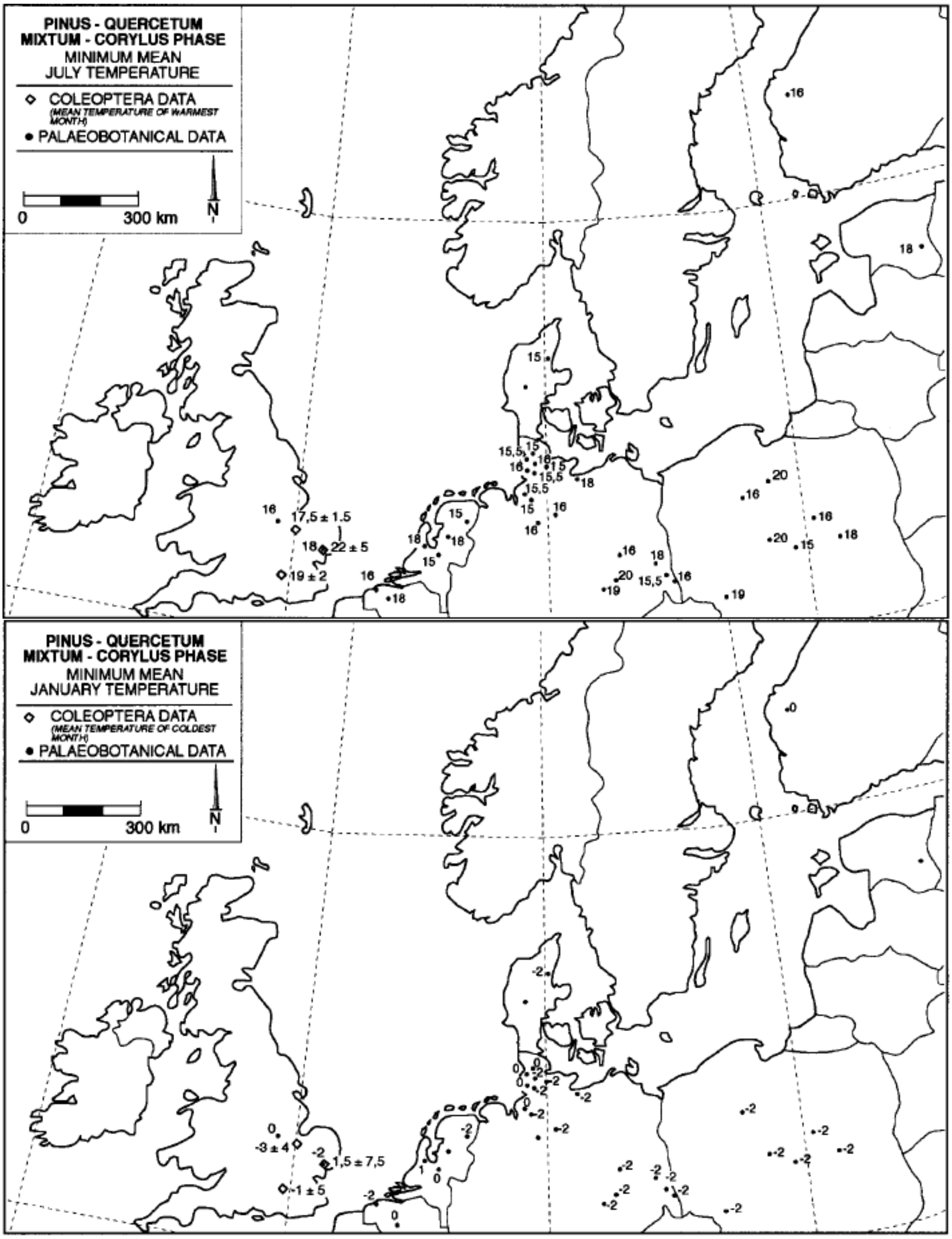

Figure 5. Reconstructed minimum mean July temperatures (top) and minimum mean January temperatures (bottom) for the early-temperate phase of the Eemian. Figure taken from Aalbersberg & Litt (1998).

In conclusion, these results suggest a rather gradual evolution of vegetation and temperatures during the Eemian, hence based on the vegetation development, a relatively stable climate evolution is expected to have existed during the Eemian. The beginning of the Eemian is characterized by a subcontinental climate where *Pinus*, *Quercus* and *Corylus* florish. The second half of the Eemian is considered as an oceanic climate, where *Carpinus* and *Picea* became dominant. At the end of the Eemian the climate became more boreal, and *Pinus, Abies* and *Picea* became dominant.

5. Eemian CO_2 dynamics

5.1 Introduction

The current interglacial is irreversible different from the past interglacials because humans have their influence on atmospheric CO_2 concentrations. By burning fossil fuels, CO_2 levels have been rising rapidly since the industrial revolution, therefore CO_2 is nowadays an important forcing factor that is not (only) influenced by climate. Hence, a direct comparison between the last interglacial and the current interglacial can not be made, however getting more information on the last transition from interglacial to glacial is important to get a better understanding on how the climate system operates without human interference, and may also help in a better understanding of the impact of the current anthropogenic influence.

In this study we focus on the role of atmospheric CO_2 during the last interglacial. The role of atmospheric CO_2 during climate changes is not yet fully understood and under constant debate. The number of CO_2 reconstructions from different proxies is very limited, most reconstructions are based on atmospheric gas trapped in ice cores from Antarctica (Barnola et al., 1987, Petit et al., 1999, Indermühle et al., 1999), and therefore many models are made in order to understand and predict changes in atmospheric carbon dioxide and their influence on other climate parameters. In order to construct a reliable model to simulate climate changes it is important to get a full understanding of the mechanisms involving climate change.

Furthermore, it is already known that the current interglacial differs from previous interglacials (Ruddiman, 2003) when looking at atmospheric CO_2 evolution. Previous interglacials show a pronounced peak at the beginning (Petit et al., 1999). However, Holocene carbon dioxide levels show an increasing trend up to the industrialization (Indermühle et al., 1999), this increasing trend is not seen during previous interglacials, where there was no reverse trend towards higher CO_2 concentrations.

5.2 The global carbon cycle

The inventory of global carbon is constant. Yet carbon is exchanged between the different reservoirs (atmosphere, biosphere, ocean and lithosphere) over different processes. This is referred to as the global carbon cycle.

For the global climate, the atmosphere is the most important reservoir where carbon is occurring mainly (99%) in the compound CO_2, which is a powerful greenhouse gas (IPCC, 2007). Its concentration varied over time, especially within glacial-interglacial cycles. The residence time of carbon in the atmosphere is relatively short.

Carbon in the terrestrial biosphere on land is another important component of the global carbon cycle. In general, the metabolism in plants, algae and many bacteria uses solar energy to transform atmospheric CO_2 into organic matter (photosynthesis). The residence time of carbon in the terrestrial biosphere is relatively short, which makes the terrestrial biosphere a relatively fast reacting system to climate change. The global vegetation distribution varied considerably from glacial to interglacial conditions as

shown by pollen and plant macrofossil analysis (Novenko et al., 1993; Velichko et al., 2005).
The ocean contains about 60 times the amount of carbon in the atmosphere (IPCC, 2007). In contrast to the terrestrial carbon cycle, which is mainly influenced by the formation, decomposition and recycling of organic carbon, the oceanic cycle is dominated by inorganic carbon chemistry. Several different physical and biological processes are acting together to reduce the surface concentration of inorganic carbon relative to the deep ocean.

5.3 Atmospheric CO_2 concentrations in the Vostok ice core record

When snow accumulates at melt-free zones of ice sheets, air becomes isolated from the surrounding atmosphere and is trapped in the pores of newly formed ice. After closure of the pores the gas remains stored within the ice sheet. In this way continuous samples of the atmosphere at the surface of ice sheets are taken by nature throughout the ages. Therefore, past CO_2 changes in the atmosphere can be determined by analyzing the enclosed air in the pores of ice.
Barnola et al. (1987) analyzed an ice-core taken at Vostok (East Antarctica) and reconstructed CO_2 concentrations during the last 160,000 years (fig. 6). The record includes the Holocene, the last glaciation, the Eemian interglacial, and the end of the penultimate glaciation.
The two major changes in this record correspond to the transitions from glacial conditions to interglacial conditions. In both cases, CO_2 levels change from low concentrations (190 - 200 ppmv) to highest values (260 - 300 ppmv). The high atmospheric CO_2 concentrations observed during the last interglacials are comparable with pre-industrial CO_2 concentrations that prevailed c. 200 years ago, before the anthropogenic forcing of atmospheric CO_2 concentrations started. The low levels of atmospheric CO_2 observed during glacial conditions are among the lowest values of the known geological history of atmospheric CO_2 over the last 10^8 year.
A spectral analysis was done, to investigate the potential link between astronomical cyclic forcing of climate and CO_2. The spectra obtained were dominated by 100-kyr components, suggesting a 100,000 year cyclicity caused by the Earth's eccentricity. Furthermore peaks were found at 21 and 40 kyr, corresponding to the precession and obliquity cycles.
The Vostok CO_2 record was compared to a climatic record based on the same core (Jouzel et al., 1987). A continuous deuterium (δD) profile was constructed and interpreted in terms of atmospheric temperature changes. In general it can be said that high δD values are associated with warm temperatures, and low δD values indicate colder temperatures. In general both records show the same trend throughout the last 160,000 years, high temperatures accompany high CO_2 concentrations. During the last interglacial both CO_2 and δD peak simultaneously at c. 135 kyr. However, where the high CO_2 concentrations remain high for at least 10 kyr, δD values decreases over the same time interval. It seems that the transition from interglacial to glacial conditions is recorded simultaneously in both temperature and CO_2 records. However, the trend within the last interglacial is different between temperature and CO_2, changes in atmospheric carbon dioxide lag changes in temperature. Barnola et al. (1987), suggest also a lag of atmospheric carbon dioxide changes to temperature changes during the transition from interglacial to glacial conditions, whereas the transition from interglacial to glacial conditions in both proxies are recorded simultaneously. It is suggested that

during the transition from interglacial to glacial, orbitally derived forcing could have influenced the marked cooling trend in the Antarctic with a dominant contribution to temperature decrease, whereas CO_2 concentrations are not affected yet.

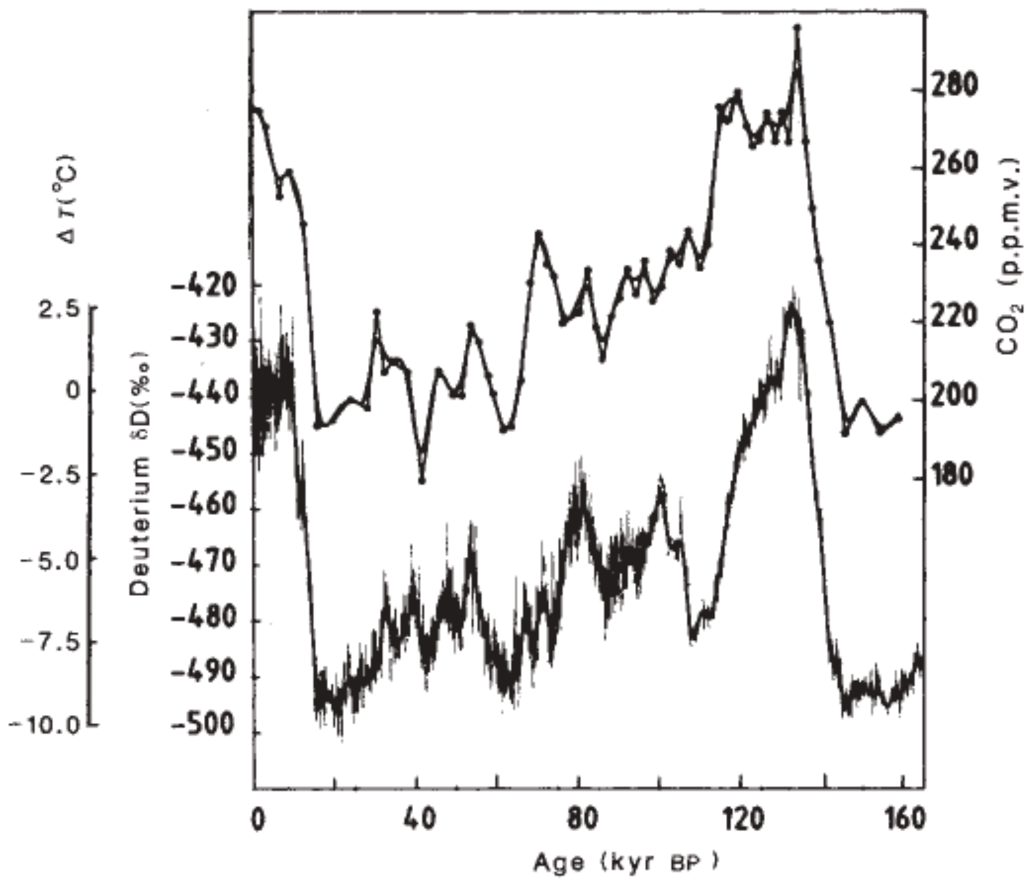

Figure 6. CO_2 concentrations (best estimates) and smoothed values (spline function) in ppmv plotted against age in the Vostok record (upper curves) and atmospheric temperature change derived from the isotopic profile (lower curve). The deuterium scale corresponds to values after correction for deuterium changes of oceanic water. Figure taken from Barnola et al. (1987).

Comparison with oxygen isotope data from the same core (Petit et al., 1999; Sowers et al., 1999) showed an even larger lag between CO_2 and global ice volume, indicating that the melting of icesheets was significantly delayed relative to increasing CO_2 and temperatures.

From this Vostok record it can thus be concluded that orbital forcing dominates the transitions between glacial and interglacial conditions. Due to these orbital changes, insolation changes and affects temperature. This on its turn lead to a reorganization in the global carbon cycle. Melting or growing of icesheets follows changing temperatures and carbon dioxide concentrations.

5.4 Atmospheric CO_2 concentrations from stomatal index data

It has already been shown that plants change their cuticle morphology in response to changing atmospheric carbon dioxide levels (Woodward, 1987). The level of CO_2 determines the number of stomata, which is set during the early stages of leaf

development. Under high carbon dioxide levels plants reduce their number of stomata, since CO_2 is highly available and water loss by transpirational cooling is in this way reduced. The stomatal index (SI) is the number of stomata relative to the sum of stomata and epidermal cells, and in this way carbon dioxide concentrations under which they grew can be reconstructed from SI. Fossil leaf cuticles can be found in lakes sediments and peat soils, and therefore it is a good proxy for past CO_2 conditions.
Rundgren et al. (2005) reconstructed atmospheric carbon dioxide concentration on a centennial resolution of the first 7400 years of the last interglacial in Hollerup (Danmark). Figure 7 shows the CO_2 reconstruction based on *Betula* and *Quercus* leaves. CO_2 levels range between 240 - 330 ppmv, which is comparable to the range of the Vostok record (Fig. 6). In general it can be seen that the first 3000 years of the Eemian were unstable, and were followed by more stable conditions over the next 3000 years. Carbon dioxide increases the first 3000 years after the Eemian onset, however a decrease is recorded between c. 2000 and 2700 years after the Eemian onset, and *Quercus* leaves indicate large-scale CO_2 shifts at 2700 - 3000 years after the onset of the interglacial.
A multi-proxy analysis was done on the same core by Björck et al. (2000), and therefore the Eemian CO_2 dynamics in the Hollerup record could be related to other climatic parameters. The transition from the relatively unstable start of the Eemian to the stable second 3000 years was accompanied by a change in the pollen stratigraphy from a *Corylus/Taxus* phase to a *Carpinus/Picea* dominated phase. Furthermore this transition coincides with the end of a climatic optimum defined by Eemian summer temperatures. These changes in vegetation also lead to a reorganization of the carbon cycle, therefore the variability during the Eemian may be linked to vegetation dynamics in Europe following deglaciation.

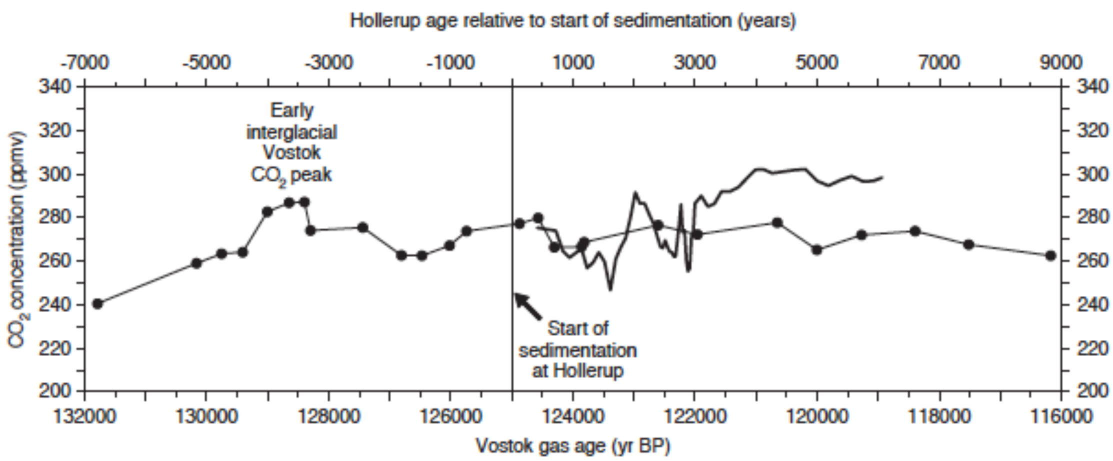

Figure 7. Tentative matching of the Vostok CO_2 record (Petit et al., 1999) and Hollerup CO_2 reconstruction (five-point running mean values). The proposed matching takes into account the 6-7 kyr lag of peak interglacial sea levels relative to the early interglacial CO_2 peak in Vostok indicated by $\delta^{18}O_{atm}$ measurements on atmospheric O_2 trapped in the Vostok core (Sowers et al., 1991) and the 3-4 kyr delayed attainment of maximum interglacial sea levels in the North Sea area relative to the onset of the Eemian in northwestern Europe (Zagwijn, 1996; Kristensen et al., 2000). Figure taken from Rundgren et al. (2005).

A comparison was between the Vostok record and the Hollerup record. Since reduces in global ice volume lagged increases in CO_2 and temperature in the Antarctic (Petit et al., 1999), it is likely that increases in CO_2 and temperature during the last interglacial lead promoted melting of the northern ice sheets. Hence, deglaciation in the Northern hemisphere lagged the peak in CO_2 and temperatures recorded during the early interglacial in the Antarctic region. Therefore, the Eemian interglacial in northern Europe is supposed to start 6-7 ky later compared to the southern hemisphere, around 126 ky BP. The Hollerup record shows a larger variability compared to the Vostok record, however it should be taken into account that the resolution of the Vostok record is coarser compared to that of the Hollerup record. The millennial scale trends in both records are similar.

From this study it can be concluded that the first 3000 years of the European Eemian interglacial were unstable, which may have to do with changes in vegetation following deglaciation leading to reorganization of the global carbon cycle. The second part of the European Eemian was relatively stable. However, the chronology of the Hollerup sequence based on pollen is a floating chronology, which raises a lot of uncertainty when fixing this record to the Vostok ice core record. The vegetation development at Hollerup showed the characteristic succession of trees (Turner, 2000), and includes all of the zones established for the Eemian in NW Europe, and therefore it is supposed that the Hollerup pollen record constitutes a complete Eemian sequence. Rundgren et al. (2005) fixed the floating Hollerup record to the Vostok ice core record, based on this characteristic vegetational development, and on lake level recontructions, which were synchronized with the known lag between CO_2 and global sea levels. This leads to a large uncertainty, since the sedimentation at Hollerup basin started later than the Eemian started, and the exact lag is not known. Furthermore, the lake level rise recorded in the Hollerup basin for the first 500 years of the record may be caused by other mechanisms than global sea level rise. Therefore, the fixation of this floating record to the Vostok ice core record has large uncertainties that have to be taken into account when discussing Eemian atmospheric carbon dioxide variability.

5.5 Holocene CO_2 dynamics compared to previous interglacial CO_2 dynamics

From the previous two carbon dioxide records it appears that the Eemian interglacial is characterized by high CO_2 levels, and is marked by an early peak in CO_2 concentrations and than in general shows a decreasing trend in carbon dioxide levels. Ruddiman (2008) compared this interglacial and 3 other interglacials (MIS 11, MIS 9, and MIS 7) to the current Holocene interglacial in terms of CO_2 and insolation trends (Fig. 8). The four past interglacials all have in common that they start with a pronounced peak in CO_2 and than gradually decrease over time, the current Holocene also shows a peak at the beginning, but shows an increasing trend over time, whereas in all cases the insolation decreases over time. A debate is going on whether the causes of the increasing CO_2 trend in the current interglacial are of natural or anthropogenic origin.

Several explanations have been proposed to explain this rise in CO_2 during the last 7000 years, of both natural and anthropogenic origin.

Among the natural CO_2 sources, an ocean carbonate compensation mechanism triggered by late-deglacial advance of forests into regions from which the melting ice sheets had retreated was proposed by Broecker et al. (1999). The theory is that as forests grew, they extracted a very large amount of carbon from the oceans. By this carbon removal from the ocean, the ocean chemistry was brought out of balance, leading to increased deposition of $CaCO_3$ on the seafloor. At the moment afforestation stopped c. 7000 years ago, the ocean gained acidity and dissolved the precipitated $CaCO_3$, which resulted in the re-entrance of the large amount of CO_2 in the atmosphere. However, if this hypothesis would be the case, than it is to be expected that the same would have happened during previous interglacials when icecaps melted, and afforestation took place. This is obviously not the case, since the other interglacial CO_2 concentrations all show a decreasing trend.

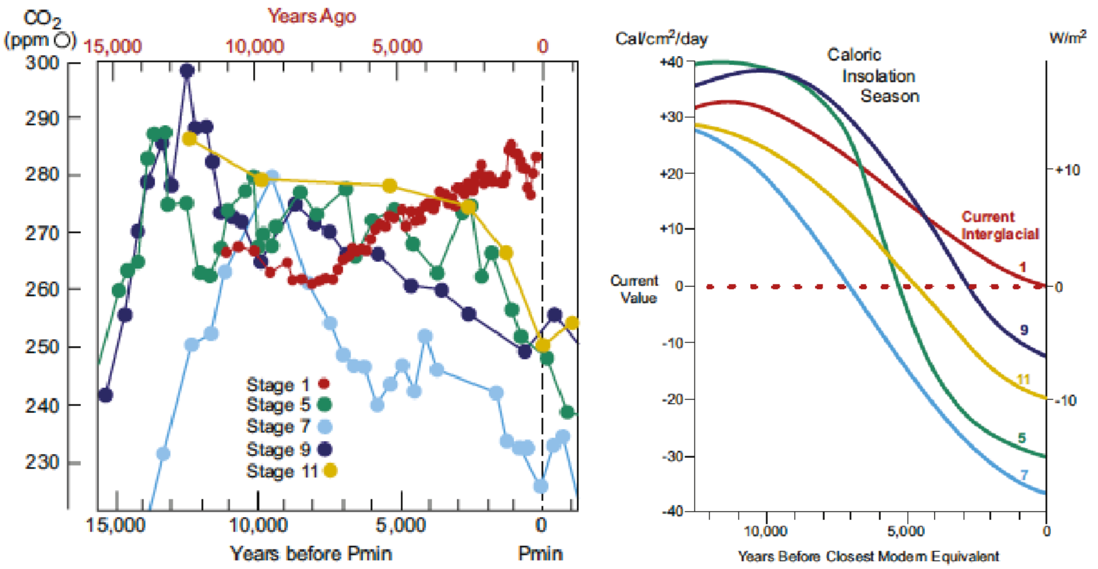

Figure 8. Comparison of atmospheric CO_2 and insolation trends for the last five interglaciations. Left: Early-interglacial CO_2 trends from Ruddiman (2007) plotted as years prior to the first precession insolation minimum in each interglaciation. Right: summer half-year insolation trends (based on Milankovich, 1941) for the northern hemisphere in cal/cm^2/day from Berger (1978), converted to W/m^2. Previous summer precession minima used as the closest equivalent of the modern precession minimum. Figure taken from Ruddiman, 2008.

Another natural CO_2 source was suggested by Ridgwell et al. (2003). It was proposed that reef construction was accelerated due to stabilization of the ocean around 7000 year ago, at peak-interglacial sea levels. Building reefs extract carbonate from the ocean, resulting in an enrichment of CO_2 in the ocean, and thus enrichment of CO_2 in the atmosphere. Again however, this should have also been the case during past interglacials.
From this we can conclude that thus far we have not been able to come up with a theory for a natural cause of the increasing carbon dioxide levels that would not have been the case during earlier interglacials. More research is needed to get a better understanding of the natural differences between the current and previous interglacials.

6. Conclusions

From the Vostok ice core record it appears that the Eemian interglacial is marked by a peak in carbon dioxide concentrations at the beginning of the interglacial, accompanied by increasing temperatures, whereas the mean global ice volume shows a later response to changes in the climate. The transition from glacial to interglacial conditions is very likely to be dominated by a change in the Earth's orbit, since a 100,000 year cyclicity is found in the CO_2 record which corresponds to the Earth's eccentricity.

The Hollerup record is to short to get more information on the cyclic behavior of glacial and interglacial carbon dioxide concentrations, nevertheless it shows a good correspondence with the Vostok ice core record in terms of atmospheric CO_2 concentrations. From this record it becomes clear that the first part of the Eemian in northern Europe was unstable. Changes in vegetation were thought to play a major role in the changing carbon dioxide levels.
Comparison with the Holocene record shows that the current interglacial is different from the previous ones, by showing a decreasing trend in carbon dioxide levels after reaching its peak early in the interglacial. Several studies were done to explain these differences in terms of changing natural sources of CO_2. However the increasing trend in carbon dioxide during the Holocene is not yet shown to be caused by natural changes in CO_2 sources.

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