Climate forced atmospheric CO₂ variability in the early Holocene: A stomatal frequency reconstruction

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Abstract

The dynamic climate in the Northern Hemisphere during the early Holocene could be expected to have impacted on the global carbon cycle. Ice core studies however, show little variability in atmospheric CO₂. Resolving any possible centennial to decadal CO₂ changes is limited by gas diffusion through the firm layer during bubble enclosure. Here we apply the inverse relationship between stomatal index (measured on sub-fossil leaves) and atmospheric CO₂ to complement ice core records between 11,230 and 10,330 cal. yr BP. High-resolution sampling and radiocarbon dating of lake sediments from the Faroe Islands reconstruct a distinct CO₂ decrease centred on ca. 11,050 cal. yr BP, a consistent and steady decline between ca. 10,900 and 10,600 cal. yr BP and an increased instability after ca. 10,550 cal. yr BP. The earliest decline lasting ca. 150 yr is probably associated with the Preboreal Oscillation, an abrupt climatic cooling affecting much of the Northern Hemisphere a few hundred years after the end of the Younger Dryas. In the absence of known global climatic instability, the decline to ca. 10,600 cal. yr BP is possibly due to expanding vegetation in the Northern Hemisphere. The increasing instability in CO₂ after 10,600 cal. yr BP occurs during a period of increasing cooling of surface waters in the North Atlantic and some increased variability in proxy climate indicators in the region.

The reconstructed CO₂ changes also show a distinct similarity to indicators of changing solar activity. This may suggest that at least the Northern Hemisphere was particularly sensitive to changes in solar activity during this time and that atmospheric CO₂ concentrations fluctuated via rapid responses in climate.

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1. Introduction

On sub-millennial timescales, changes in global biomass and ocean temperatures/salinities control exchanges between the main global carbon reservoirs (oceans, terrestrial biosphere and atmosphere). In the Northern Hemisphere, the early Holocene (ca. 11,600–8000 cal. yr BP) was a climatically dynamic period with, for example, meltwater pulses, vegetational colonisation, soil formation and oceanic circulation reorganisations on decadal to centennial timescales. It is unlikely therefore, that atmospheric CO₂ concentrations were stable during this period. However, ice core measurements of CO₂ during the early Holocene show little variability (Monnin

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et al., 2001). Atmospheric CO₂ is a well-mixed atmospheric trace gas responding to carbon flux variations in both the Southern and Northern hemispheres. Carbon exchanges are both directly and indirectly related to climate and many proxy records register abrupt climatic events during the early Holocene (Behre, 1978; Björck et al., 1997; Hald and Hagen, 1998; Björck et al., 2001; Sejrup et al., 2001). One of the most widespread cooling events occurred a few hundred years after the end of the Younger Dryas (YD) cold event and is generally known as the Preboreal Oscillation (PBO) (Björck et al., 1996; Björck et al., 1997). The PBO is difficult to ¹⁴C date reliably but most probably lies ca. 11,300–11,150 cal. yr BP (Björck et al., 1997) and has been recognized for a number of years. Other events, however, of varying magnitude, spatial scale and timing are now emerging from ice cores (Johnsen et al., 1992; O’Brien et al., 1995), terrestrial (Björck et al., 2001) and marine (Hald and Hagen, 1998) sediment proxies.

Measurements of past atmospheric CO₂, and therefore indications of changing fluxes, are obtained directly from air bubbles trapped in Antarctic ice cores and provide excellent records over long time periods (Fischer et al., 1999; Indermühle et al., 1999; Petit et al., 1999; Monnin et al., 2001). The recognition of higher frequency changes in atmospheric CO₂ is limited by the inherent smoothing of ice core gas records by diffusion (Trudinger et al., 2003). To obtain a better time resolution of CO₂ changes it would be preferable to use the higher accumulation rates of Greenland sites but this is generally avoided due to the dust related in situ production of CO₂ (Anklin et al., 1995; Tschumi and Stauffer, 2000). To investigate how strongly the records are smoothed, Spahni et al. (2003) applied a diffusion and enclosure model to a decrease in CH₄ through the cooling event dated to ca. 8200 yr BP in both GRIP (Greenland) and Dome C (Antarctica) ice cores. The model estimated that only 34–59% of the original atmospheric amplitude shows up in the Dome C record for an event of ca. 100–200 yr and that only events of more than 500- (Dome C) and 62- (GRIP) year duration will register 95% of the actual atmospheric gas variation. Diffusion smoothing mechanisms affect CH₄ and CO₂ in the same way and it is therefore possible that any early Holocene short-term CO₂ variations lasting less than 1–2 centuries are strongly damped in the Antarctic ice core records.

Dome C is generally considered the best resolved record of atmospheric CO₂ over the early Holocene period (Monnin et al., 2004). However, it indicates maximum amplitude of variation of only 5–6 ppmv over this time period (i.e. similar to the modern annual variation). This limited variability is likely to partly reflect the relatively large age distribution per air sample (ca. 200 yr) for this ice core, and the difficulties in resolving changes on centennial or sub-centennial timescales create a need for other complementary methods. Less direct, proxy reconstructions may therefore be the more sensitive archives when short-term, rapid changes are of interest. The relatively new method of stomatal frequency analysis applies the physiological response of certain C₃ plants to changing concentrations of CO₂ in the atmosphere (Woodward, 1987). Rapidly accumulating lake sediments, if supplied with sufficient numbers of leaves from the surrounding vegetation, allow the reconstruction of atmospheric CO₂ changes on centennial to decadal timescales by applying its inverse relationship to stomatal frequency. This paper presents a stomatal frequency reconstruction of atmospheric CO₂ at decadal resolution over the time period 11,230 to 10,330 cal. yr BP, supported by a highly resolved AMS radiocarbon chronology from lake sediments of the Faroe Islands (situatated in the North Atlantic between south west Norway and Iceland). As our atmospheric CO₂ estimates should reflect the global net effect of carbon transfers between reservoirs responsive on sub-millennial to decadal timescales, we also discuss possible causes for the changes in reconstructed CO₂ in relation to evidence of climate variation and solar activity during the early Holocene.

2. Site and stratigraphy

Lykkjøvøtn is a small (100×25 m), shallow lake situated ca. 1.5 km west of Skopun on Sandoy, Faroe Islands (61°54’37”N, 6°54’30”W) at 52 m.a.s.l. Nine parallel sediment cores were extracted from the deepest part of the lake (50 cm water depth) using a Russian corer. The deepest 100 cm of sediment extracted consisted of silt gyttja with gyttja clays and silty clays at the base (Fig. 1a). The gyttja sediments contained abundant Salix herbacea leaf macrofossils. The Saksunarvatn Ash, a common isochron in early Holocene lake sediments from the Faroe Islands (Wastegård et al., 2001), is identified as a dark ca. 1 cm layer at 373 cm depth and is dated to 10,180 yr BP (1950 AD) in the GRIP ice core (Grönvold et al., 1995) and later revised to 10,240 yr BP (Björck et al., 2001) (Fig. 1). Magnetic susceptibility measurements from each of the nine sediment cores provided distinctive profiles enabling a good correlation and indicate a change in sedimentation rate at 414 cm.

3. Chronology

Twelve AMS radiocarbon dates were obtained from S. herbacea leaves from the 68 cm of gyttja below the
Saksunarvatn Ash (Table 1, Fig. 1). In an attempt to date the base of the sequence, 3 additional bulk samples from the silt gyttja and a single bulk sample from the clays were AMS radiocarbon dated (Table 1). Calibration was performed using the INTCAL 04 data set (Reimer et al., 2004) and the OxCal v.3.10 calibration program (Bronk Ramsey, 1995; Bron Ramsey, 2001). Three of the bulk radiocarbon dates resulted in slightly younger than expected age ranges and lie on the 9600 14Cy rB P plateau (Fig. 1b). These dates were excluded from the age–depth model (Fig. 1a). The application of the calibrated age–depth model with linear segments uses minimum 90% probability age ranges (Fig. 1a) and takes into account the sedimentation rate change at 414 cm as indicated by magnetic susceptibility and other proxy measurements (Jessen et al., in preparation). The resulting accumulation rate of sediments is 0.47 mm per yr from the basal clays to 414 cm and 1.18 mm per yr from 414 cm to the Saksunarvatn Ash.

4. Stomatal frequency method and calculations

All the 1 cm sub-samples from the silt gyttja contained well-preserved whole or partial S. herbacea leaves and care was taken to ensure that fragments were not parts of the same leaf (Table 1). Counting of stomata and epidermal cells was conducted according to standard procedures (Poole and Kürschner, 1999) using an epifluorescence microscope and digital imaging software. In most levels, the optimal 5 individual leaves were counted per sub-sample with 5 fields from each of the abaxial and adaxial surfaces (maximum of 50 count fields per sub-sample level). Fields were limited to interveinal areas. A few small leaf fragments were observed in the basal clays but these were not sufficiently preserved to allow stomatal index (SI) determination. A total of 42 levels were counted (Table 1).

The influence of environmental variables other than atmospheric CO2 can be a consideration when quantifying stomatal frequency by stomatal density (number of stomata per unit area). However, many studies testify to the fact that the calculation of stomatal index (the number of stomata proportional to the sum of stomata and epidermal cells) minimises effects other than CO2 (Beerling, 1999; Royer, 2001). In this study stomatal index was calculated and calibrated using inverse regression with a previously produced modern training set for S. herbacea (Rundgren and Beerling, 1999). Errors produced therefore reflect those of the calibration and the number of leaves counted.

5. CO2(SI) reconstruction and centred comparison

The reconstructed stomatal-based CO2 (CO2(SI)) concentrations are shown in Fig. 2a and b with detailed results listed in Table 1. It is expected that stomatal-based data include a certain amount of biological scatter and smoothing with a 5 pt running mean is commonly used. This method does not however, account for any changes in sedimentation rate and it is therefore inconsistent over time. Additionally, it gives no
| AMS radiocarbon dates for Lykkjuvøtn calibrated using the IntCal04 data set (Reimer et al., 2004) and OxCal v.3.10 (Bronk Ramsey, 1995; Bronk Ramsey, 2001) |
|---|---|---|---|---|---|---|---|---|---|
| Mid-sample depth (cm) | Lab no. | Material analyzed | Weight (mg) | Reported ¹⁴C age (±1σ, 1σ) | Calibrated age with 95% probability (cal. BP) | SI | Calibrated CO₂ (ppmv) | Upper 95% confidence interval (ppmv) | Lower 95% confidence interval (ppmv) |
| 375 | LuA5760 | S. herbacea leaves | 3 | 9110±60 | 10,490–10,180 | 2 | 5.26 | 343.61 | 372.29 | 309.89 |
| 376 | LuA5761 | S. herbacea leaves | 4 | 9060±60 | 10,410–9940 | 5 | 5.10 | 348.29 | 368.02 | 322.98 |
| 377 | LuA5762 | S. herbacea leaves | 6 | 9090±110 | 10,600–9900 | 5 | 5.54 | 335.33 | 354.38 | 312.21 |
| 378 | LuA5763 | S. herbacea leaves | 4 | 9310±60 | 10,680–10,290 | 5 | 5.60 | 326.01 | 344.80 | 304.22 |
| 379 | LuA5764 | S. herbacea leaves | 10 | 9320±50 | 10,690–10,300 | 5 | 5.53 | 351.50 | 373.83 | 309.84 |
| 380 | LuA5765 | S. herbacea leaves | 6 | 9360±50 | 10,720–10,420 | 5 | 6.07 | 319.89 | 338.63 | 298.86 |
| 381 | LuA5766 | S. herbacea leaves | 4 | 9300±60 | 10,660–10,280 | 5 | 5.54 | 335.33 | 354.38 | 312.21 |
| 382 | LuA5767 | S. herbacea leaves | 4 | 9400±60 | 10,800–10,400 | 5 | 5.68 | 320.51 | 341.26 | 301.17 |
| 383 | LuA5772 | Silt gyttja | ca.4 | 9460±60 | 11,100–10,550 | 5 | 6.04 | 320.63 | 339.38 | 299.51 |
| 384 | LuA5768 | S. herbacea leaves | 8 | 9500±50 | 11,080–10,580 | 5 | 5.66 | 331.22 | 351.06 | 309.48 |
| 385 | LuA5769 | S. herbacea leaves | 10 | 9610±60 | 11,180–10,750 | 5 | 5.54 | 343.49 | 362.93 | 319.03 |
| 386 | LuA5770 | S. herbacea leaves | 4 | 9580±50 | 11,140–10,730 | 5 | 6.35 | 311.70 | 330.54 | 291.52 |
| 387 | LuA5771 | S. herbacea leaves | 6 | 9830±60 | 11,400–11,150 | 5 | 5.57 | 305.74 | 324.77 | 286.06 |
| 388 | LuA6008 | Silty clay | ca.5 | 9490±70 | 11,100–10,550 | 5 | 5.65 | 330.96 | 349.86 | 308.49 |

indication of errors. Low-pass filtering of the data (80-year cut-off) with errors calculated using Monte Carlo simulation, takes account of changing sedimentation rates and produces errors on the smoothed curve but does not assume any biological scatter. We acknowledge however that the higher data resolution in the upper part of the core may influence variability in both smoothing methods.

The two modes of smoothing are illustrated in Fig. 2a and b. There are clear similarities between the two smoothed curves and, with the exception of the extreme earliest point, the running mean values fall within the Monte Carlo simulated errors. The running mean smoothing is considered the most appropriate when comparing to other stomatal-based CO2 records and assessing the rates of change over short periods of time, because it accounts for biological scatter. When considering the whole time period and longer term trends however, concentrations inferred from low-pass filtering may be the most reliable representation of the raw data. Detailed discussion of the stomatal index based results uses the different filters accordingly.

Variations in atmospheric CO2 concentration are reconstructed (in both the smoothed and unsmoothed records) with distinct low values centred at ca. 11,050 cal. yr BP, 10,600 and 10,400 cal. yr BP and high values at ca. 10,900 and 10,550 cal. yr BP. Rates of change (based on running mean smoothing) are variable; the decrease in values to the oldest minimum at ca. 11,050 cal. yr BP indicates ca. 15 ppmv in 100 yr (ca. 1.5 ppmv per decade). This is slightly slower than the decrease to the youngest minimum at ca. 10,450 cal. yr BP, ca. 20 ppmv in ca. 100 yr (ca. 2.0 ppmv per decade) but both are much faster than the decrease of ca. 20 ppmv over ca. 250–300 yr (ca. 0.8–0.7 ppmv per decade) between ca. 10,900 and 10,600 cal. yr BP. This compares to ca. 6 ppmv in ca. 40 yr (1.5 ppmv per decade) measured in the Law Dome ice core at the end of the sixteenth century AD (Etheridge et al., 1996). Because atmospheric gas records from ice...
cores are inherently smoothed by diffusion during bubble closure this is a minimum estimate of the actual atmospheric change (Spahni et al., 2003; Trudinger et al., 2003). The contribution of differential smoothing to the difference in amplitude between ice core (CO2_{(ice)}) and CO2(SI) records was investigated by van Hoof et al. (2005). Application of differential smoothing, as constrained by the site-specific parameters for the D47 ice core, to a stomatal reconstruction between 1000 and 1500 AD, reduced the CO2(SI) amplitude by 25%, resembling the 12 ppmv range recorded in D47 (van Hoof et al., 2005). Clearly, the amplitudes of change throughout the Lykkjuvøtn record are realistic however it is still apparent, that although stomatal change throughout the Lykkjuvøtn record are realistic 1000 and 1500 AD, reduced the CO2(SI) amplitude by 25%, resembling the 12 ppmv range recorded in D47 (van Hoof et al., 2005). The centred values of all the stomatal index records show a reduction in CO2(SI) beginning a few decades before 11,100 cal. yr BP equivalent to 20–30 ppmv (Fig. 2c). Concentrations remain low for ca. 150 yr followed by a rise reaching pre-event values at ca. 10,900 cal. yr BP. Thereafter the Lykkjuvøtn and Borchert records both indicate stable or decreasing values whereas the Madtjärn record suggests rising values. Because this represents the very end of the smoothed Madtjärn record, it must be interpreted with caution and may not be a persistent trend.

6. Discussion

6.1. Potential early Holocene climate–carbon cycle connections

A difference in the CO2 partial pressure (pCO2) of air and water allows the exchange of carbon. In simple terms, surface ocean temperatures (SST’s) and salinities (SSS’s) control pCO2 via their effect on dissolved inorganic carbon (DIC) concentrations and therefore the efficiency in sequestering carbon from the atmosphere (the solubility pump). A 1 °C lowering of SST’s reduces pCO2 by ca. 4.2% (Takahashi et al., 1993), sequestering more carbon and reducing atmospheric concentrations. Although less efficient, decreasing salinities also reduce concentrations in the atmosphere (Broecker, 2002). Sea ice obstructing transfer (Stephens and Keeling, 2000) is another factor and the loosely constrained, poorly understood effects of wind speeds must also be considered in relation to climate change (Ito and Follows, 2003). Oceanic biomass uptake of carbon via fluctuations in the supply of limiting nutrients (Fe) is especially important in the main HNLC (High Nutrient Low Chlorophyll) regions where nitrate and phosphate are in excess due to upwelling or vertical mixing (Pedersen et al., 2002). The effects of this ‘biological pump’, especially in the Southern Oceans (due to its proximity to carbon in the deep ocean), have been mainly discussed relative to the 80–100 ppmv reduction of CO2_{(ice)} concentrations during glacial periods. It is
not considered to be a major influence on Holocene CO₂ concentrations (Pedersen et al., 2002).

Estimates of the development of global terrestrial biomass and therefore its carbon storage are problematic and involve large uncertainties. Globally, palaeodata are fragmentary in both space and time and inundation of coastal zones due to rising sea levels, spreading forests in recently deglaciated areas, soil development, permafrost fluctuations and especially the initiation of peatlands all potentially change the size of the terrestrial carbon reservoir. Biospheric carbon storage in vegetation and soils is dependant upon climatic conditions and thus fluctuations in total vegetation cover will affect atmospheric–biosphere carbon exchanges. Although some coupled atmosphere–ocean–vegetation models have suggested that the impact of changing carbon storage in the terrestrial biosphere on the global cycle is minimal (Brovkin et al., 2002), recently it has been proposed that during stable periods such as the Holocene there may have been a larger terrestrial impact than previously thought (Köhler and Fischer, 2004). Indermühle et al. (1999) performed inverse modelling on Holocene records of CO₂(ice) concentration and δ13C of CO₂ from the Taylor Dome ice core. They concluded that the terrestrial biosphere made a substantial contribution to global CO₂ variations during this period, supporting modelling experiments of its importance also on shorter timescales during the Holocene (Trudinger et al., 1999; Gerber et al., 2003). During unstable periods such as deglaciation, oceanic processes are most likely to exert the largest effect (Smith et al., 1999). Generally, simulations are severely hampered by the lack of high-resolution δ13C records indicating the major sources of carbon fluxes into the atmosphere on shorter timescales, but it is not difficult to envisage terrestrial biosphere changes contributing to fluctuations in atmospheric CO₂ during the climatically dynamic early Holocene in the Northern Hemisphere.

In this context, the disparity between the ‘ocean dominated’ Southern Hemisphere and the ‘land dominated’ Northern Hemisphere raises the question of their relative importance for changes in atmospheric CO₂ at different times. The correlation of polar ice cores, based on the matching of CH₄ records, suggests that the Northern and Southern hemispheres were climatically asynchronous through the last deglaciation (Blunier et al., 1997; Blunier et al., 1998). Although there has been some debate (Steig et al., 1998), the consensus of opinion and additional terrestrial evidence (Turney et al., 2003) support an asynchrony during deglaciation. During the Northern Hemisphere YD cold event, Antarctic temperatures were undergoing a rise out of the last glaciation at the same time as atmospheric CO₂ (ice) records show an increase in concentrations (Monnin et al., 2001). This may suggest that, during an asynchronous deglaciation, CO₂ concentrations can be most adequately explained by climatic changes in the Southern Hemisphere (Marchal et al., 1999). If southern ocean warming partially compensated for Northern cooling, essentially obscuring any Northern Hemispheric changes in carbon storage visible in CO₂, other proxies for carbon cycle variation would be expected to indicate a similar pattern. A clear YD signal is recorded in atmospheric CH₄ (mainly controlled by the northern hemisphere and tropical regions) (Blunier et al., 1997), and both CH₄ and CO₂(ice) register an accelerated rise at the end of the YD. During the early Holocene, a substantial contribution to CH₄ emissions by the Northern Hemisphere is suggested by the ice core derived Interpolar Methane Gradient (IPG) during the early Holocene (Chappellaz et al., 1997; Brook et al., 2000). The large scale initiation of peat formation in the Western Siberian Lowlands has been identified as a possible major CH₄ source at this time (Smith et al., 2004). Peat formation on this scale would constitute an important CO₂ sink in the Northern Hemisphere. This suggests therefore, that during the early Holocene the dynamic climate in the Northern Hemisphere was influential on global atmospheric CO₂. Higher resolution reconstructions of CO₂(SI) also suggest a Northern Hemisphere influence on the carbon cycle during deglaciation (Rundgren and Björck, 2003). In this study a rapid decrease of over 40 ppmv was reconstructed shortly after the onset of the YD, recovering to stable values over a further ca. 300 yr.

Causes of early Holocene abrupt climatic events are often thought to relate to meltwater outbursts from the remaining, rapidly disintegrating ice sheets in the Northern Hemisphere (Björck et al., 1996). Meltwater outbursts are difficult to date reliably but decreasing salinities and lower sea surface temperatures in the North Atlantic ocean may not only have resulted in lower pCO₂ but also a slowdown in THC and a climatic cooling and/or hydrological changes in neighbouring terrestrial regions, all of which could contribute to a lowering of global atmospheric CO₂.

The time lag between forcing and CO₂ response has been investigated in relation to meltwater outbursts. Experiments with models simulating the response of surface ocean pCO₂ (Marchal et al., 1999) and the terrestrial carbon reservoir (Scholze et al., 2003) to a freshwater pulse give an indication of the meltwater forcing-atmospheric CO₂ response time lag. In the oceanic model experiment (Marchal et al., 1999), THC
was switched off by a freshwater influx and a first response of atmospheric CO₂ began ca. 250 yr later, recovering after ca. 650 yr. A faster first response of ca. 100 yr was simulated in terrestrial carbon stocks with maximum response of vegetation and soils after ca. 250 yr and 400 yr respectively (Scholze et al., 2003). Both of these simulations positioned a freshwater influx into the northern North Atlantic, known to elicit a more widespread effect on climate (Manabe and Stouffer, 2000). However, between non-meltwater forced climate (temperature and precipitation) change and CO₂ response, no major time lag between climatic cooling and CO₂ decrease is found. During the so-called Little Ice Age (LIA), no significant time lag between climatic cooling and CO₂ can be observed in Antarctic ice cores (Etheridge et al., 1996).

6.2. The Lykkjuvøtn CO₂ record and early Holocene climate change

Ice core (Masson et al., 2000), marine (Bianchi and Gersonde, 2004) and terrestrial records (Heusser, 1998; Markgraf et al., 2003) indicate that the Southern Hemisphere was undergoing a relatively stable thermal optimum during the Early Holocene (ca. 12,000–9000 cal. yr BP). Regions around the North Atlantic were, in contrast, subject to rapid climate changes through continued ice sheet melting and pulses of freshwater with associated changes in thermohaline circulation (THC) (Björck et al., 1996). The variability in our CO₂ record suggests an influence of these dynamic changes in Northern Hemisphere climate on the carbon cycle (Fig. 3). The distinct decrease centred at ca. 11,050 cal. yr BP has been reconstructed previously (Fig. 2c) and suggested to be a carbon cycle response to the PBO (Wagner et al., 2004). This cooling event, a few centuries after the end of the YD and occurring at a period with significant Δ¹⁴C changes, is suggested to have been triggered by freshwater outbursts from Lake Agassiz and the Baltic Ice Lake into the North Atlantic (Björck et al., 1996; Hald and Hagen, 1998; Fisher et al., 2002; Teller et al., 2002). The Preboreal is also characterized by a rapid rise in global sea level (MWP 1B) (Bard et al., 1996). Signs of a Preboreal cooling are seen over a large geographic area and range from Greenland temperatures (Johnsen et al., 1992), Nordic Sea surface temperatures (Hald and Hagen, 1998), Icelandic vegetation changes and glacier advances (Björck et al., 1997), Norwegian glacier advance (Bakke et al., 2005), and Northern and Central European vegetation (Björck et al., 1996; van der Plicht et al., 2004) and hydrological changes (Bos, 2001;
Magny and Bégeot, 2004). Although radiocarbon dating of this period is problematic, terrestrial records generally suggest a humid, cooler climate in the North Atlantic region between ca. 11,300 and 11,150 cal. yr BP (Björck et al., 1997). GRIP $\delta^{18}O$ infers that Greenland temperatures decreased between ca. 11,400 and 11,200 yr BP (Johnsen et al., 1992) and recent evidence suggests a possible two phase (cool/dry followed by more humid) event beginning ca. 11,400 cal. yr BP (Sejrup et al., 2001; Björck et al., 2002; van der Plicht et al., 2004). The CO$_2$(SI) decrease begins shortly after the end of inferred climatic cooling at ca. 11,150 cal. yr BP (Fig. 3). If this is a meltwater forced climatic cooling, the CO$_2$ response could be due to a combination of a cooler/fresher surface ocean and a cooling forced increased carbon storage in soils and vegetation as is suggested for the LIA (Gerber et al., 2003). Additional storage in Western Siberian peats which were initiated around this time (Smith et al., 2004) could also be a contributing factor, but the slight reduction of CH$_4$ in the Dome C ice core (Flückiger et al., 2002) may indicate a short-term suppression of these new carbon sinks (Fig. 3). Taking into account dating uncertainties, it is possible that the reconstructed CO$_2$(SI) decrease partly overlaps with the PBO and therefore may have acted as a positive feedback on the meltwater induced climatic cooling.

After ca. 140 yr concentrations increase to pre-event values but following a period of high values, the data reflect a ca. 300 yr (ca. 10,900 to 10,600 cal. yr BP) consistent and steady decline in CO$_2$ concentrations (Fig. 3). Available palaeoclimatic records suggest that this time period was generally stable and warm in both the Northern and Southern Hemispheres (Johnsen et al., 1992; Steig et al., 2000). Minimum winter sea ice cover in the Atlantic sector of the Southern Ocean (Bianchi and Gersonde, 2004) and minimum percentage IRD in the North Atlantic (Bond et al., 2001) during this time both support this suggestion. There is little evidence suggesting large scale regional changes at this time and the decrease in CO$_2$(SI) concentrations are therefore most likely due to the general Northern Hemisphere expansion of forests and peatlands with an associated increase in terrestrial carbon storage.

Atmospheric CO$_2$(SI) demonstrates greater variability in the time period after ca. 10,600 cal. yr BP than in the previous ca. 250 yr (Fig. 3). The higher data resolution in this part of the reconstruction may have influenced variability however, it coincides with an increase in North Atlantic iceberg occurrence (Bond et al., 2001) (Fig. 3), a greyscale inferred decrease in North Atlantic SST's (Hughen et al., 1996) and an increase in meltwater outburst frequency (Björck et al., 1996; Teller et al., 2002). In addition, some increase in the amplitude of variability is seen in Greenland temperatures (NGRIP, 2004) and chironomid inferred evidence of lower summer temperatures in the Swiss Alps during ca. 10,700 to 10,500 cal. yr BP (Heiri et al., 2004). Bond et al. (1997) showed that in the North Atlantic, cooler waters with sea ice were periodically advected further.
south, demonstrating a cyclical pattern. The increase in CO$_2$(SI) variation begins during the cooling of the cycle culminating at ca. 10,300 cal. yr BP and may be connected to any associated instability in climate (Fig. 3).

7. Solar activity and atmospheric CO$_2$ — a striking similarity

Muscheler et al. (2000) reconstructed past production rates of radionuclides through the period 15,000 to 9300 cal. yr BP based on $^{10}$Be measurements in the GISP2 ice core. A comparison of modelled $^{10}$Be and $\Delta^{14}$C measured in tree rings with $^{10}$Be flux was used to assess the relative impact of changing THC on the carbon cycle. They concluded that during the early Holocene, $^{10}$Be and $\Delta^{14}$C variations on time scales shorter than 500 yr can be solely explained by changes in the radionuclide production rates most probably caused by changes in solar activity. Two of the largest and longest Holocene production rate maxima (i.e. solar activity minima) occur at ca. 11,250 and 10,300 yr BP. The former has been discussed as a trigger for the PBO (Bond et al., 2001; Björck et al., 2001) and the latter as a possible cause of a short cooling event dated to ca. 10,300 cal. yr BP in the Northern Hemisphere (Björck et al., 2001). Fig. 4 shows a comparison of our atmospheric CO$_2$(SI) with $\Delta^{14}$C and $^{10}$Be. The striking similarity between these solar activity proxies and changes in CO$_2$(SI) is clear during the CO$_2$(SI) minima centred at ca. 11,050 cal. yr BP, the slow decrease between ca. 10,900 and 10,600 cal. yr BP and the subsequent variability. Three possibilities are apparent: (1) coincidence; (2) a link between solar activity and SI; and (3) a link between solar activity and atmospheric CO$_2$ via climate. The similarities (even when chronological problems are considered) appear too obvious to allow for coincidence. Regarding the second option, any possible response in SI would most likely be via climate change and SI has been shown to eliminate or minimise climatic effects (water stress, light intensity, temperature, humidity etc.) (Royer, 2001). Although it could be argued that the decrease in stomatal frequency observed in herbarium specimens from the past ca. 150 yr (Woodward, 1987) coincides with a period of general solar activity increase (Hoyt and Schatten, 1998), this tentative link does not hold for the period after ca. 1950 AD when solar activity was relatively stable. In addition, modern experiments in controlled environments and CO$_2$ concentrations show a clear inverse response of SI to CO$_2$ and herbarium material demonstrates a decrease in SI at the same time as the post-industrial CO$_2$ increase (Woodward, 1987; Woodward and Bazzaz, 1988; Royer, 2001). This suggests that it is not likely that SI is responding directly or indirectly (via climate) to changes in solar activity.

This leaves the third option, but mechanisms linking synchronous changes in global atmospheric CO$_2$ concentrations/climate and solar activity are still poorly
understood. Solar UV output shows a greater relative variation than total solar irradiance during the 11-year sunspot cycle and for this reason changes in UV radiation have been discussed as an important link between solar activity and climate (Haigh, 1996). The forcings are assumed to be relatively small however, and amplification would probably involve routes through the stratospheric heat budget to atmospheric circulation and in turn to the hydrologic cycle, (and/or the cryosphere) and thus also the carbon cycle.

If the time scales of Fig. 4 are correct, low solar activity results in decreased atmospheric CO2 probably via a rapid response in climate. This would suggest that there might have been a direct climatic response to changes in solar activity, to which the recently deglaciated Northern Hemisphere was particularly sensitive and which in turn has an influence on the global carbon cycle.

Climatic proxy responses to solar forcing have been suggested previously in both marine and terrestrial archives Björck et al., 2001; Hu et al., 2003; van der Plicht et al., 2004; Wang et al., 2005. Such a relationship can also partly be seen in records from the LIA (Fig. 5), a period of known climatic change. Shindell et al. (2001) modelled the global temperature response to the solar irradiance changes known as the Maunder Minimum (a similar forcing to the anomalies of the early Holocene) during the mid-17th to 18th centuries AD. Although global average temperature changes were minimal, the resulting 1–2 °C decrease in winter temperatures in North Atlantic regions is similar to known changes during the LIA. The Law Dome ice core records an atmospheric CO2 decrease of ca. 6 ppmv during the climatic cooling associated with the Maunder Minimum (Etheridge et al., 1996) but no clear decrease is registered during the Spörer Minimum a few centuries earlier (Fig. 5). Stomatal index based CO2 reconstructions covering this period presently either do not cover the whole time period (van Hoof, 2004), have poor chronological control (Rundgren and Beerling, 1999) or are based on stomatal density (Kouwenberg et al., 2005). To further investigate any possible correlation between changes in atmospheric CO2 and solar activity, reconstructions covering the last 1000 yr could prove useful and give a better insight into carbon cycle dynamics on these timescales.

8. Conclusions

By using a stomatal index based proxy for atmospheric CO2 concentration we have reconstructed changes during the early Holocene period (11,230–10,330). We have shown that to resolve changes in atmospheric CO2 concentration on timescales representative of its response rate, a proxy complementing ice core measurements is necessary. In contrast to ice core measurements, which cannot resolve CO2 variability on these timescales, our results indicate that the global carbon cycle was not in steady state during this period of the early Holocene. The main conclusions can be summarized thus:

- Stomatal index reconstructions of atmospheric CO2 concentrations can record variability and trends on decadal timescales, complementing the inherently smoothed ice core records. The trends of changing atmospheric CO2 during the early Holocene can be accounted for by the dynamic climatic changes in the Northern Hemisphere at this time.
- During the period ca. 11,230–10,330 cal. yr BP, stomatal frequency-inferred atmospheric CO2 concentrations demonstrate variability on centennial timescales. This could indicate that the global carbon cycle was not in steady state as is suggested by ice core CO2 measurements, which cannot resolve changes on these scales.
- During a climatically stable Southern Hemisphere in the early Holocene, the regional dynamic climate of the Northern Hemisphere is reflected in global CO2 as reconstructed from stomatal frequency and is therefore a major contributor to changes in the global carbon cycle.
- The CO2(SI) reconstruction through the early Holocene bears a striking similarity to reconstructed solar activity changes. This may suggest a rapid response of climate to minor changes in solar activity during this dynamic period, which in turn impacted the global carbon cycle. This can, to some extent, also be seen in the climatic responses associated with the Maunder Minimum in the mid-17th to early 18th centuries.

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