Amplitude variations of 1470-year climate oscillations during the last 100,000 years linked to fluctuations of continental ice mass

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Abstract. We describe amplitude variations of a 1470-year (y) signal in the oxygen-isotope record from the GISP2 ice-core as a function of continental ice mass as recorded in sea-level variations. The 1470-y signal is closely tied to Dansgaard-Oeschger interstadial events, which perturbed the climate system on a global scale. Large amplitude 1470 y oscillations occur only if continental ice exceeds a threshold level, equivalent to sea level at approximately ~45 m and are tied to times of change in ice mass. Minima of the 1470-y signal in the Greenland ice are associated with short time intervals of relatively stable phases of sea level and with reduced climate variability in the entire periodicity range between approximately 550–5000 y.

Introduction

Oxygen isotope (δ¹⁸O) data from Greenland ice cores reveal that large and rapid temperature fluctuations (warming by up to 7 °C within a few decades) dominated climate in Greenland between ~11–74 thousand years before present (ky BP) [Johnsen et al., 1992; Dansgaard et al., 1993; Grootes et al., 1993]. Grootes and Stuiver [1997] demonstrated that warm peaks (Dansgaard-Oeschger interstadials) occurred periodically every 1470 years (y). The significance of this periodicity is corroborated by similar climate variations on a global scale which can be correlated to the Dansgaard-Oeschger events [e.g. Stuiver et al., 1993; Behl and Kinnen, 1996; Mayewski et al., 1997; Schulz et al., 1998; Sarnthein et al., in press]). We investigate a possible link between the amplitude of the 1470-y signal, associated with Dansgaard-Oeschger events, and variations in the amount of continental ice, for the last 100 ky. Our analysis is intended to provide semi-quantitative insights into the dynamics which control the evolution of the 1470-y signal.

Temporal Evolution of the 1470-y Signal

Based on time-frequency analysis (Fig. 1a, c) the temporal evolution of the 1470-y signal component in the GISP2 δ¹⁸O record [Grootes et al., 1993, Grootes and Stuiver, 1997] shows tight links to the Dansgaard Oeschger interstadials: In the interval 15–78 ky BP each δ¹⁸O and temperature maximum coincides with an amplitude maximum of the 1470-y signal. Moreover, the long-lasting Dansgaard-Oeschger events 8, 12, 14, 19 and 20 comprise two 1470-y cycles. The amplitude of the 1470-y signal varied notably, with a maximum around 31 ky BP, a distinct minimum during the last glacial maximum (18–21.5 ky BP) and small amplitudes after 11 and before 78 ky BP (Fig. 2a).

Amplitude maxima of the 1470-y signal cluster around the last glacial stages. It is plausible, therefore, to relate their occurrence to the amount of continental ice that existed at any given time. Using marine δ¹⁸O as proxy for sea level and global ice volume [Berger et al., 1996], it is obvious that maxima of the 1470 y signal amplitude are confined to times when sea level was more than ~45 m below its present level (Fig. 2a), suggesting threshold behavior of a system responsive to ice mass. Minima occur at 21, 50, 63 and 79 ky BP and are coeval to local minima or maxima of sea level (Fig. 2a), that is, times of little change in ice mass.

The temporal pattern of the 1470-y signal amplitude is not necessarily restricted to this very frequency but maybe found over a wider range. Time-frequency analysis of the GISP2 δ¹⁸O data in the periodicity range between ~550–10000 y (Fig. 3) reveals distinct time intervals (~12, 16–24, 48–51, 75–82, 86–100 and 103–108 ky BP) during which amplitudes drop in the entire periodicity range between ~550–5000 y. Since the GISP2 δ¹⁸O series at least records climate in the North Atlantic region [Grootes and Stuiver, 1997] and to some extent global climate variability [Blunier et al., 1998], these time intervals mark periods when the climate system showed subdued centennial-to-millennial-scale variability. Thus low-agility intervals (“agility gaps”) in the temporal behavior of the climate system are not restricted to the 1470-y signal, but appear to be a pervasive characteristic of climate variability during the last 110 ky BP. An exception is the minimum of the 1470-y signal around 63 ky BP, which does not coincide with a well defined “agility gap”.

Building a 1470-y Amplitude Template

Sea level h is converted into a dimensionless proxy for continental ice mass M = (h – h_min) / (h_max – h_min), where subscripts indicate minimum and maximum values. We use the sea-level curve of Berger et al. [1996] which is based on a stacked planktonic δ¹⁸O-record from the western Pacific and modeling using summer insolation at 65°N as input and agrees with measurements of Fairbanks [1989] and Chappell et al. [1996]. We denote absolute rate of ice-mass change by r = dM/dt, where t is time, and normalize to R = (r – r_min) / (r_max – r_min). The amplitude of the 1470-y signal A is then expressed as function of ice mass M and absolute rate of change R

A(t) = F (M, R, ⟨R⟩),

(1)

where ⟨R⟩, is the average of R in the interval [t, t–τ]. Eq. (1) is based on the inferred relation between 1470-y signal amplitude and sea level (Fig. 2a): (1) Beyond some threshold value for M, the amplitude becomes rather insensitive to the actual ice mass and (2) amplitude minima are more likely to occur if ice mass is relatively stable. Accordingly, there is no obvious way to describe the amplitude maximum around 31 ky BP, which coincides with little ice mass variations. Since this peak occurs at the end of a prolonged period with relatively stable ice conditions (Fig. 2a) and is restricted to the 1470-y signal (Fig. 3), we assume that it is caused by an additional mechanism, which is linked to the long-term stability of ice mass. To account for this mechanism, we include “system memory” into eq. (1) by the

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Figure 1. (a) Bottom: Oxygen Isotope (δ¹⁸O) record from Greenland (GISP2 ice core [Grootes et al., 1993; Grootes and Stuiver, 1997]). Numbers above δ¹⁸O maxima denote Dansgaard-Oeschger interstadial events. Top: Temporal changes of 1470-y signal component in the δ¹⁸O record estimated by a harmonic-filtering algorithm [modified after Ferraz-Mello, 1981] using a sliding rectangular window of width 4 × 1470 y (half-amplitude bandwidth = 0.2 ky). (b) Power spectrum of the 15–80 ky BP section of the δ¹⁸O time series (bar indicates resolution bandwidth, BW). Integration of the autospectrum reveals that the 1470-y signal accounts for ~17% of the variance of the δ¹⁸O data. (c) Detailed view of the time series in (a), showing the tight phase coupling between the Dansgaard-Oeschger events and the 1470-y signal. Signal maximum near 31 ky BP results from the averaging effect of the sliding window (a matching small δ¹⁸O maximum can be recognized in both the GISP2 and GRIP record).

variable \( \langle R \rangle_c \). Numerical experiments suggest the following form of eq. (1):

\[
\Lambda(t) \text{[\%]} = F_1(M) \left( a F_2(R) + b \langle R \rangle_c^{-1} \right),
\]

with \( F_1(M) = 0.5 \tanh(5(M - c)) + 0.5, F_2(R) = 0.5 \tanh(15(R - d)) + 0.5 \) and parameters \( \tau = 7 \) ky, \( a = 0.38, b = 0.06, c = 0.41 \) and \( d = 0.17 \), estimated by a least-squares procedure (Fig. 2b). Functions \( F_1 \) and \( F_2 \) act as switches and introduce threshold behavior into eq. (2). Whereas \( F_1 \) modulates signal amplitude \( A \) by ice mass \( M \), function \( F_2 \) provides for “agility gaps” by reducing \( A \) for small values of \( R \) near local minima and maxima of \( M \). The term involving \( \langle R \rangle \) increases \( A \) whenever \( R \) remains close to zero for \( \tau \) years. We refer to eq. (2) as template in order to distinguish it from models, which treat the involved processes explicitly.

The template correctly predicts the major features of the observed signal amplitude (Fig. 2b): (1) A rapid increase at ~78 ky BP when ice mass reached a critical value, corresponding to sea level approximately 45 m below its present value. (2) “Facilitating” behavior between ~80–40 ky BP driven largely by alternating intervals of unstable sea level and short periods with relatively stable ice mass, which result in minima of \( F_2(R) \) at 38, 50, 63 and 81 ky. (3) A distinct peak at ~31 ky BP, subsequent to a 7 ky long period with little fluctuations in ice mass resulting in a maximum of \( \langle R \rangle_c^{-1} \). (4) A minimum during the last glacial maximum, due to the increase in ice mass which results in a drop of \( F_2(R) \) and \( \langle R \rangle_c^{-1} \). During the last deglaciation, amplitude increases until continental ice mass reaches the critical level beyond which \( F_1(M) \) goes to zero, taking \( A \) with it.

Discussion and Conclusions

Possible sources of discrepancies between template and data-derived amplitude are: (1) It is unlikely that the response of the system, represented by the various parameters in eq. (2), has been constant throughout time. (2) Periodicities < 8 ky are suppressed in the employed sea-level record [Berger et al., 1996], making its resolution considerably lower than the decadal-to-centennial resolution of the GISP2 δ¹⁸O data [Grootes et al., 1993; Grootes and Stuiver, 1997]. Hence, any possible link between short-term sea-level fluctuations [Chappell et al., 1996] and 1470-y signal amplitude variations cannot be resolved. The amplitude drop predicted by the template during the last deglaciation occurs approximately 1300 y earlier than the decline estimated from the data. This offset in time is identical to the time shift between the sea level histories of Fairbanks [1989] and Berger et al. [1996] for deglacial sea level at ~45 m and, thus, likely to be a deficit of the sea-level history used.
favor the existence of a second threshold, which reduces millennial-scale climate variability in the presence of large continental ice mass. If the system were controlled by two thresholds only (at ~ 45 and ~95 m; the latter to account for amplitude minima around 21 and 63 ky BP; Fig. 2a) the minima of the 1470-y signal amplitude near 50 ky BP cannot be explained (Fig. 2a). Since the total time interval covered by the GISP2 δ18O record encompasses only one interval when sea level was below ~100 m, we cannot completely rule out the existence of a second threshold. Despite this limitation, we regard it as unlikely that the existence of two thresholds alone is sufficient to explain the amplitude variations of the 1470-y signal.

Our results contrast with the inference that amplitude maxima show a better correlation to decreasing northern hemisphere insolation than to ice volume changes [Mayewski et al., 1997]. The delayed response of ice volume to insolation [e.g., Berger et al., 1996] leads instead to a lagged correlation between insolation change and 1470-y amplitude maxima.

The apparent insensitivity of the 1470-y signal amplitude to the sign of the ice growth rate and its sensitivity to the absolute value of the ice growth rate calls for some rectifying mechanism in the underlying physical system. The saw-tooth shape of the δ18O record during the course of Dansgaard-Oeschger cycles (Fig. 1a) is reminiscent of the 100-ky ice-age cycles [e.g., Berger et al., 1996] with its slow build-up and sudden "release" of ice-sheet instability. Accordingly, we surmise that this basic mode of operation applies also to Dansgaard-Oeschger cycles. To account for the different time scales of variability (100 ky vs. ~1.5 ky), we assume that the fast response of ice-sheet margins and small ice caps to perturbations is important for creating instability and for enhancing Dansgaard-Oeschger events. During a phase of ice growth, both southward expansion of continental ice sheets into more temperate climate zones and the erosion of older periglacial sediments will tend to destabilize ice margins and thereby increase the sensitivity of the climate system to perturbations. Moreover, ice growth may disturb the thermal balance of an ice cap, resulting in a larger probability of basal sliding [MacAyeal, 1993]. During intervals of ice-sheet decay, the accompanying rise in sea level has probably the largest potential to add instability to the system by destabilizing marine based ice sheets.

The occurrence of the amplitude maximum around 31 ky BP is seemingly at odds with the notion that ice-mass stability should result in an amplitude minimum. A possible explanation for this contradictory evidence would be amplification of the 1470-y signal by a particular configuration of continental ice sheets which requires about 10 years to establish, e.g., changes in the configuration of the Laurentide ice sheet. Reconstructions of this ice cap [Clark et al., 1993] include the possibility that Hudson Bay was not wholly ice-covered between 77–32 ky BP. A 2-dimensional model of the Laurentide ice sheet centered on Hudson Bay results in self-sustained 1500–5000 y oscillations [Greve and MacAyeal, 1996]. Accordingly, we can only surmise that build-up of an ice dome over Hudson Bay at ~32 ky BP provided the required amplification of the 1470-y signal. This interpretation disagrees with the occurrence of Heinrich events before 32 ky BP [Sarnthein et al., in press] and their assumed linkage to an ice dome over Hudson Bay [MacAyeal, 1993]. However, provenance analysis of ice-rafted debris deposited during Heinrich events only points to a source "surrounding Labrador Sea" and does "not require an ice stream surging through the Hudson Strait" [Hemming et al., 1998].

The small amplitude of the 1470-y signal during the Holocene (Fig. 1) supports the conclusion of Bond et al. [1997] that the smaller magnitude may be a consequence of the lack of large continental ice caps, which act as amplifiers of millennial-scale climate variability by adding instability to the climate system.

Figure 2. (a) Amplitude of the 1470-y signal in Fig. 1a; interpolated (cubic spline) to 0.1 ky intervals (thin line). Amplitude increases sharply as sea level [Berger et al., 1996] falls below ~45 m (dotted line). Pronounced amplitude minima between ~20–80 ky BP coincide with local minima or maxima of sea level (arrows). (b) Predicted temporal evolution of the 1470-y signal amplitude (solid line), using the template (eq. 2), versus data-derived amplitude (dashed line). For better clarity the estimated amplitude from (a) was smoothed using a 2000 y wide Hanning filter prior to fitting eq. (2) to the data. A fit to the unsmoothed data yields essentially the same results (not shown).

The template describes ~70% of the variance of the 1470-y amplitude, which makes us confident that we included into eq. (2) proxies for the most important processes, which modulated the 1470-y signal amplitude during the last 100 ky. The physics underlying these proxies, as well as the origin of the 1470-y cycle itself, are yet to be specified. Furthermore, proxies for local climate variability or certain physical aspects of the 1470-y oscillations (e.g. the trigger mechanism itself) may show a different temporal amplitude pattern than the GISP2 δ18O record.

Our results indicate that: (1) Significant 1470-y climate oscillations occur only if continental ice mass exceeds a threshold level, equivalent to sea level at ~45 m. For greater ice mass, values of amplitude maxima are largely insensitive to ice mass. (2) The 1470-y signal amplitude drops substantially whenever sea level stays close to a local minimum or maximum. (3) Time intervals with relatively stable sea level are accompanied by reduced climate variability in the entire periodicity range between ~550–5000 y ("agility gaps").

McManus et al. [1999] concluded that during the last 500 ky large-amplitude millennial-scale climate variability occurred only if sea level was below ~30 m. Our analysis confirms the existence of such a threshold, although at a value of ~45 m. This difference probably results from an offset in sea-level estimates for the time when the threshold is reached, but not from a difference in timing of the threshold crossing. McManus et al.
Figure 3. Time-frequency analysis of GISP2 818O data in Fig. 1a. Setting the width of the sliding window to 4 times the periodicity results in increasingly shorter filtered sequences towards lower frequencies. Half-amplitude bandwidth (BW) depends on periodicity (T) as BW = 1.2 / (4T). Horizontal line marks the time evolution of the 1470-y amplitude (cf. Fig. 2). Bars at top indicate “agility gaps”, characterized by amplitude drop in the entire periodicity range between ~550–5000 y.

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