The tempo of climate change during Dansgaard-Oeschger interstadials and its potential to affect the manifestation of the 1470-year climate cycle

Michael Schulz

Institut für Geowissenschaften, Universität Kiel, Germany

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[1] The oxygen-isotope record from the GISP2 ice core serves a air-temperature proxy to estimate cooling rates during Dansgaard-Oeschger interstadials between 15–53 thousand years before present. The duration of individual Dansgaard-Oeschger interstadials scales inversely with the cooling rate and high cooling rates are associated with large continental ice volume and vice versa. Cooling rates determine the extent to which the postulated 1470-year pacing cycle of the Dansgaard-Oeschger interstadials is visible in paleoclimatological records: Only cooling rates associated with intermediate ice volume (approx. -84 ± 20 m relative sea-level equivalent) allow the climate system to respond to succeeding trigger events of the 1470-year cycle and result in stadials/interstadials of nearly the same duration. *INDEX TERMS:* 3344 Meteorology and Atmospheric Dynamics: Paleoclimatology, 4267 Oceanography: General: Paleoceanography; *KEYWORDS:* climate change, cooling rates, ice core

1. Introduction

[2] The climate of the last glacial period (~10–80 ky BP; thousand years before present) is characterized by variability on millennial timescales with rapid transitions between cold stadials and warm Dansgaard-Oeschger interstadials (DOI; Figure 1a). These climate fluctuations were first recognized in the ice-core records from Greenland [*Dansgaard et al.*, 1993], but since then their global significance has been revealed [e.g. *Clark et al.*, 1999].

[3] No consensus exists with respect to the nature of the timing of the DOI onsets. Spectral analysis of the oxygen-isotope ($\delta^{18}O$) record from the GISP2 ice core in Greenland shows a prominent spectral peak at a periodicity of 1470 years (y), which is thought to be an expression of the recurrent switching between Dansgaard-Oeschger stadials and interstadials [*Grootes and Stuiver*, 1997]. In contrast, *Wunsch* [2000] reasoned that the sharp 1470-y spectral peak is caused by aliasing spectral variance from "annual" cycles to frequencies in the millennial band, implying that no periodic pacing of the DOI onsets exists. Considering the sampling properties of the GISP2 ice core, this contention was convincingly rejected by *Meeker et al.* [2001]. Further re-analysis of the GISP2 $\delta^{18}O$ series revealed that the statistical significance of the 1470-y spectral peak depends critically on the presence of DOI 5, 6, and 7 between 31–36 ky BP in the record [*Schulz*, accepted]. Yet, it was also demonstrated in this study that the timing of DOI onsets during the last 50 ky is consistent with a 1470-y pacing cycle. Another unresolved aspect of the DOI concerns the duration of individual DOI. *Bond et al.* [1993] noted that long DOI tend to occur after Heinrich events, i.e., massive ice-discharge events from the Laurentide ice sheet, while shorter DOI are more likely to occur prior to Heinrich events.

[4] The origin of the Dansgaard-Oeschger-type climate variability remains controversial, ranging from internal oscillations of the ocean-atmosphere system [e.g. *Sakai and Peltier*, 1997] to periodic calving of the Greenland ice sheet [*van Kreveld et al.*, 2000], and to external forcing mechanisms [*van Geel et al.*, 1999]. A recent modeling study [*Ganopolski and Rahmstorf*, 2001] corroborated the earlier suggestion [*Berger and Jansen*, 1995] according to which DOI are perturbations of a relatively stable stadial climate mode — a view adopted in this study. In these model experiments DOI are triggered by prescribed low-amplitude freshwater-forcing cycles during periods of negative anomalies in the northern North Atlantic. Since the DOI mode is unstable, it gradually decays until a threshold is reached and the climate system jumps back into the stable stadial mode. Accordingly, the tempo at which climate cools during DOI seems to control the duration of the DOI. Here it is shown that this rate is proportional to continental ice volume. This implies that the expression of the 1470-y pacing cycle of the DOI in paleoclimatic proxy records parallels changes in ice volume.

2. Estimating the tempo of interstadial climate change

[5] The $\delta^{18}O$ record from the GISP2 ice core (Figure 1a) reflects atmospheric temperature variations of global significance [*Grootes and Stuiver*, 1997; *Stuiver and Grootes*, 2000]. In this record, DOI are characterized by a rapid $\delta^{18}O$ increase at their onset, a gradual $\delta^{18}O$ decrease during the interstadial, and finally a fast return to low stadial $\delta^{18}O$ values. To a first order the $\delta^{18}O$ values scale linearly with air temperature above Greenland [*Grootes and Stuiver*, 1997]. Accordingly, the average cooling rate during each DOI was estimated from the slope of a straight line, obtained by a least-squares fit to the $\delta^{18}O$ data (Figure 1a). Onset and termination of each DOI was determined based on the occurrence of rapid transitions in the $\delta^{18}O$ record. The analysis is restricted to the interval 14–53 ky BP, where the timescale uncertainty is 2% for ages up to ~40 ky BP and 5–10% for older ages [*Meese et al.*, 1997]. This makes the time series well suited for estimating rates of change.

[6] The results indicate an inverse relationship between cooling rate and duration of the DOI (Figure 2). For example, the slow tempo of the coolings around 37, 44, and 50 ky BP resulted in relatively long time spans until a threshold for the interstadial-to-stadial transition were crossed, ensuing the long-lasting DOI 8, 12, and 14 (Figure 1a, b). In contrast, high cooling rates between 22–29 and 40–42 ky BP led to the short DOI 2, 3, 4, 9, and 10.

3. Linking the tempo of climate change to ice volume

[7] Changes in DOI cooling rates are paralleled by variations in sea level (Figure 1c), which is a proxy for global ice volume: Larger ice volume seems to favor more rapid cooling

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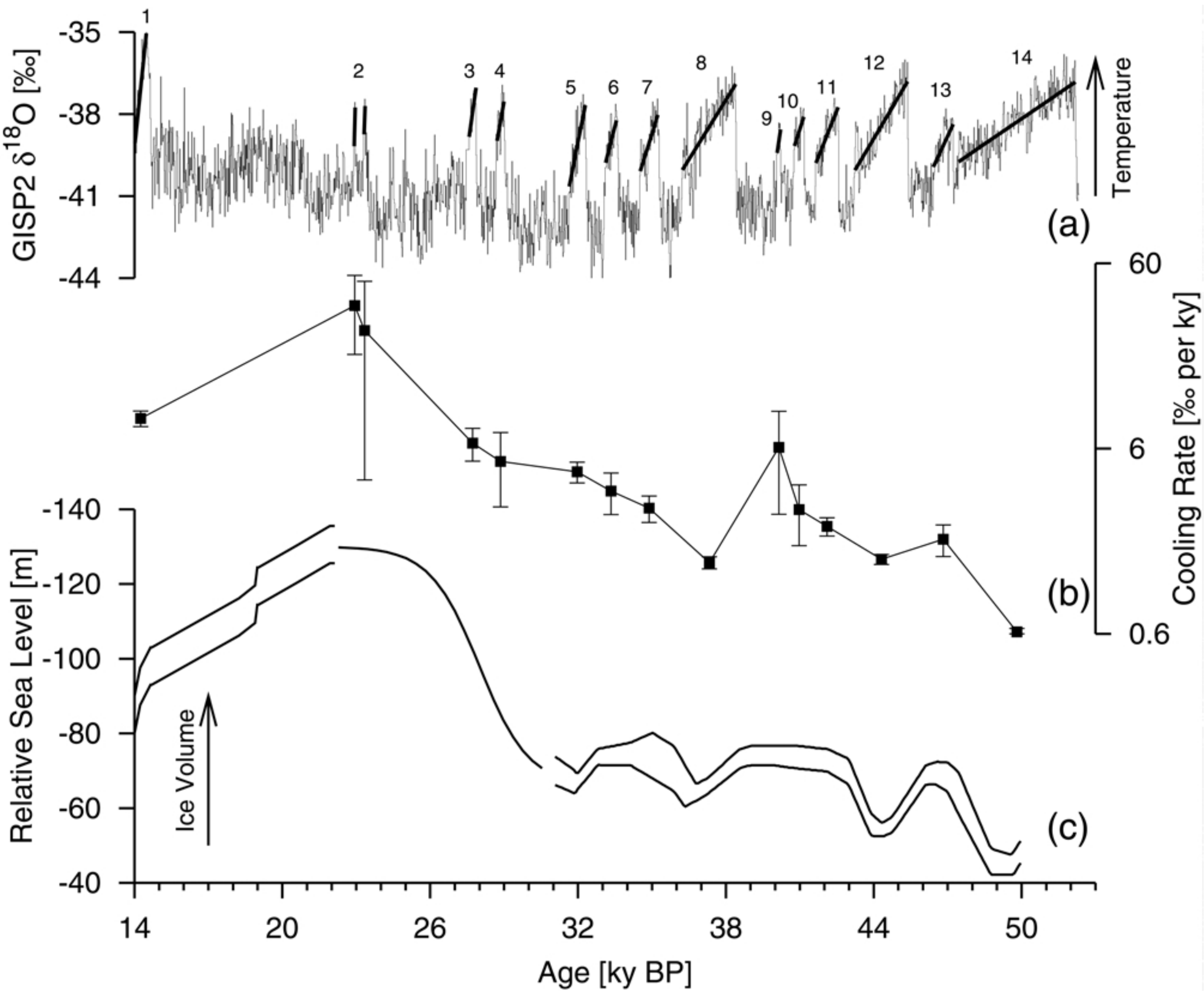

Figure 1. Relation between temperature and global ice volume. (a) GISP2 oxygen-isotope (δ^{18}O) record [*Stuiver and Grootes*, 2000] (thin line) reflecting air-temperature and best-fit lines (bold line). Numbers above δ^{18}O maxima denote Dansgaard-Oeschger interstadials (DOI). (b) DOI cooling rates defined by slopes of best-fit lines to the δ^{18}O data in (a). Estimated uncertainties of the cooling rates measure the variability of the δ^{18}O data around the regression lines but do not take into account dating uncertainties (1σ error bars). Note logarithmic scale. (c) Most probable relative sea-level range based on Th/U-dated corals and reef-growth modeling (for ages >31 ky BP) [*Fairbanks*, 1990; *Lambeck and Chappell*, 2001]. Note inverted scale and interpolation in data gap between 22–31 ky BP.

and, hence, swifter return from the perturbed interstadial state to the stable stadial mode. Since the DOI cooling trends seen in the ice-core δ^{18}O record go along with decreasing deep-water export from the Norwegian-Greenland Seas [*Rasmussen et al.*, 1996], it is likely that these climate deteriorations are associated with a gradual reduction in convection in the Nordic Seas [*Ganopolski and Rahmstorf*, 2001]. A mechanism seems to operate, which accelerates this decline in deep-water formation and, therefore, the cooling rate in the presence of large ice volume.

[8] Possible mechanisms involve a change in mean sea-surface salinity in the North Atlantic, linked to variations in net precipitation in the catchment area of the North Atlantic and Arctic and/or changes in runoff from the surrounding ice sheets. Model simulations indicate a freshening of surface water in the North Atlantic for glacial boundary conditions [*Bush and Philander*, 1999]. This freshening is largely due to enhanced net precipitation, which is in turn sensitive to the height of the Laurentide ice sheet [*Kageyama and Valdes*, 2000]. Moreover, it is conceivable that an extensive Laurentide ice sheet results in large runoff to the North Atlantic [*Marshall and Clarke*, 1999] and therefore to reduced sea-surface salinity. In contrast, decreased runoff, associated with a small Laurentide ice sheet, would make the surface water of the Atlantic saltier. Paleoceanographic data [*Dokken and Jansen*, 1999; *Sarnthein et al.*, 2001] and modeling results [*Ganopolski and Rahmstorf*, 2001] indicate that during DOI near-surface waters from the North Atlantic penetrated into the Nordic Seas to feed deep-water formation. This circulation thus allows for the advection of salinity-anomalies from the North Atlantic to the convection sites within the Nordic Seas. The lower the salinity of the advected water, the faster the threshold will be reached upon which convection in the Nordic Sea ceases and the thermohaline circulation jumps back into its stadial mode with deep-water formation restricted to locations south of Iceland [*Ganopolski and Rahmstorf*, 2001]. Accordingly, the existence of a large ice volume seems to be consistent with the inferred fast cooling during DOI and can thus account for short DOI and vice versa. However, careful modeling

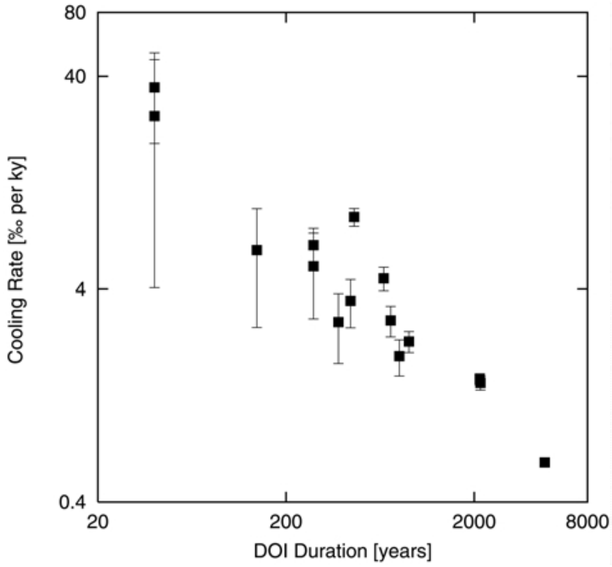

Figure 2. Relation between duration of Dansgaard-Oeschger interstadials and estimated cooling rate (as in Figure 1a). Note logarithmic scales.

studies are required to uncover the details of the link between ice volume and cooling rate.

4. Implications for the manifestation of the 1470-y cycle

[9] As noted above, the occurrence of DOI is consistent with a pacing by a 1470-year cycle [*Schulz*, accepted] of hitherto unknown origin. Assuming that DOI are indeed linked to a dynamically unstable state of the climate system [*Ganopolski and Rahmstorf*, 2001], changing cooling rates during DOI have an important implication for the expression of this pacing cycle in the glacial geological record (Figure 3). If the cooling during a DOI proceeds too slow, the climate system has not yet returned to the stadial state when the pacing cycle tries to trigger the next DOI. Hence the duration of a DOI may be longer than 1470 y. Cooling rates associated with intermediate ice volume (approximately −64 to −104 m relative sea-level equivalent; cf. Figure 1) allow the climate system to react to succeeding trigger events of the 1470-year pacing cycle and result in stadials and interstadials of similar duration. Higher cooling rates, associated with relative sea-level below approximately −110 m, lead to pulse-like shaped DOI. At very large ice volume it may become impossible to trigger a DOI, leading to extended stadial conditions. This was probably the case during the last glacial maximum (18–21.5 ky BP), which is characterized by a lack of DOI. Accordingly, the seeming contradiction between the existence of a 1470-y pacing cycle and the duration of individual DOI, which may be longer or shorter than expected from the pacing period (cf. Figure 3), can be resolved. Moreover, the distinct pattern of the cooling rates during the DOI, together with the evidence for a pacing of the onsets of the Dansgaard-Oeschger interstadials by a 1470-year cycle [*Schulz*, accepted], suggests that Dansgaard-Oeschger interstadials do not represent random "flickers" of the climate system.

[10] For the last 100 ky *Schulz et al.* [1999] noted that the amplitude of the 1470-y signal component in the GISP2 $\delta^{18}O$ time series reached maximum values during times of ice buildup and ice decay as well as during time intervals with relatively stable ice volume at intermediate level (31–36 ky BP; Figure 1c). In view of the results presented here, these findings can be easily unified into a common principle: During the times of ice-volume change, the amount of ice passes through the discriminating range in which the associated DOI cooling rates allow for an optimal tracing of the 1470-y pacing cycle by paleoclimatic proxies (approx. -84 ± 20 m relative sea-level equivalent). The relatively stable sea-level between 31–36 ky BP falls into the same critical range (Figure 1c), leading to 1470-y climate cycles, characterized by stadials/interstadials of nearly equal length.

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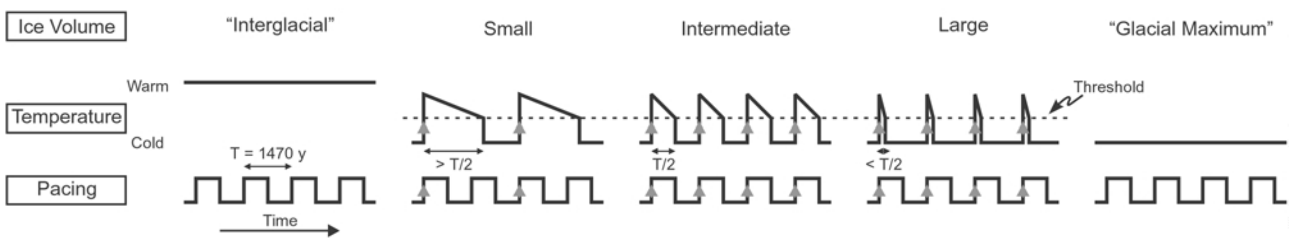

Figure 3. Manifestation of the 1470-y pacing cycle as function of continental ice volume. Actual shape of the pacing cycle (bottom row) is unimportant because this signal acts only as trigger (vertical arrowheads) for Dansgaard-Oeschger type temperature fluctuations (center row). Cooling rate during unstable DOI depends on ice volume (top row). After reaching a threshold the system jumps back into the cold stadial state. At "interglacial" ice volume climate remains in the stable warm mode. An ice volume corresponding to a relative sea-level equivalent of approximately −45 m [*Schulz et al.*, 1999] is required to make the climate systems susceptible for the 1470-y pacing. At such "small" ice volume, the cooling rate is too low to reach the threshold before the next pacing cycle starts, resulting in DOI durations which exceed 1470 y. At "intermediate" ice volume ($\simeq 84 \pm 20$ m relative sea-level equivalent; cf. Figure 1) the cooling rate is sufficiently high to produce interstadials and stadials of approximately the same duration. At even "larger" ice volume the DOI become pulse-shaped, while at the ice volume of the last glacial maximum the system looses its capability to undergo Dansgaard-Oeschger fluctuations.

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M. Schulz, Institut für Geowissenschaften, Universität Kiel, D-24098 Kiel, Germany. (mschulz@email.uni-kiel.de)