Relaxation oscillators in concert: A framework for climate change at millennial timescales during the late Pleistocene

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[1] Using a box model of the North Atlantic Ocean and a parameterization of Heinrich Events, we suggest that self-sustained oscillations of the large-scale oceanic circulation provide a framework to accommodate crucial elements of late Pleistocene climate variability: (*i*) Dansgaard-Oeschger-style oscillations with varying interstadial length, (*ii*) synchronization between Dansgaard-Oeschger stadials and Heinrich Events, and (*iii*) Younger Dryas-type events. The latter result from the restart of the oscillations after a glacial maximum and can be regarded as Dansgaard-Oeschger stadials, overprinted by rapidly changing boundary conditions. *INDEX TERMS:* 3344 Meteorology and Atmospheric Dynamics: Paleoclimatology. **Citation:** Schulz, M., A. Paul, and A. Timmermann, Relaxation oscillators in concert: A framework for climate change at millennial timescales during the late Pleistocene, *Geophys. Res. Lett.*, *29*(24), 2193, doi:10.1029/2002GL016144, 2002.

1. Introduction

[2] The climate of the last glacial period (~10–80 kyr BP; thousand years before present) was characterized by variability at millennial timescales with rapid transitions between cold stadials and warm Dansgaard-Oeschger (DO) interstadials. The origin of the DO-type climate variability remains controversial. Suggested hypotheses range from internal oscillations of the ocean-atmosphere system [*Broecker et al.*, 1990; *Winton*, 1993; *Sakai and Peltier*, 1997] over periodic calving of the Greenland ice sheet [*van Kreveld et al.*, 2000], to external forcing mechanisms [*Ganopolski and Rahmstrof*, 2001]. DO-type climate variations appear to be linked to continental ice volume, occurring only if ice-volume equivalent sea level is in an intermediate range [*McManus et al.*, 1999; *Schulz*, 2002a]. Within this range the length of individual DO interstadials varies inversely with ice volume [*Schulz*, 2002a].

[3] A second key feature of the last glacial period were massive ice-discharge events from the Laurentide ice sheet. These Heinrich Events (HE) coincide with DO stadials and are followed by relatively long DO interstadials [*Bond et al.*, 1993]. Internal oscillations of the Laurentide ice cap are a potential mechanism for generating HEs [*MacAyeal*, 1993]. However, this concept requires an additional process to synchronize HEs and DO stadials. To this end, contrasting conjectures exist about the role of atmospheric cooling during DO stadials for triggering HEs [*Oerlemans*, 1993; *Clarke et al.*, 1999]. Moreover, since potentially destabilizing sea-level rise seems to occur during but not before HEs [*Chappell*, 2002], identification of the synchronization mechanism remains elusive.

[4] A third aspect of millennial-scale climate variability involves the intermittent return to glacial conditions during glacial terminations, such as the ~1300-yr (year) long Younger Dryas (YD) during the last deglaciation. Similar cold events have also been documented for older terminations [*Sarnthein and Tiedemann*, 1990] and thus appear to be a general feature of late Pleistocene climate. The physical mechanisms responsible for these YD-type events are also not yet uncovered [e.g. *Sarnthein et al.*, 2001].

[5] A common aspect of these three types of climate variations is their close link to changes in deep-water formation in the North Atlantic Ocean and the associated poleward heat transport. Compared to present-day, deep-water formation was most likely reduced during DO stadials and the YD and ceased during HEs [*Sarnthein et al.*, 2001].

2. Conceptual Ocean and Ice-Sheet Oscillators

[6] *Winton* [1993] used an ocean general circulation model to study internal oscillations of the thermohaline circulation with centennial-to-millennial periods. In this model, continued surface freshening causes a polar halocline to form, which eventually suppresses deep-water formation. The Atlantic Ocean enters a "deep-decoupled phase" with reduced meridional circulation and heat transport. Subsequent import of heat and salt by advection and diffusion weakens the polar halocline, which finally breaks down. The concomitant onset of convection in high latitudes triggers a "deep-coupled phase" with active deep-water formation and strong meridional circulation and heat transport. The reestablishment of the halocline completes a "deep-decoupling oscillation" [e.g. *Paul and Schulz*, 2002 for details]. The essence of these results can be distilled into a box model (Figure 1), which represents the competing effects of surface freshening in polar regions and subsurface meridional heat transfer [*Winton*, 1993]. Similar relaxation oscillations were also found in other models [e.g. *Weaver et al.*, 1993; *Paul and Schulz*, 2002; *Timmermann et al.*, submitted].

[7] A deep-decoupling oscillation is reminiscent of a DO-cycle, with the deep-coupled phase representing the DO interstadial and the decoupled phase characterizing the DO-stadial mode with reduced deep-water formation [*Winton*, 1993; *Timmermann et al.*, submitted]. In the box model the period of the deep-decoupling oscillation depends on the freshwater forcing (F_w) and varies between approximately 1.3 and 4.1 kyr (Figure 2). For F_w < 0.704 m/yr

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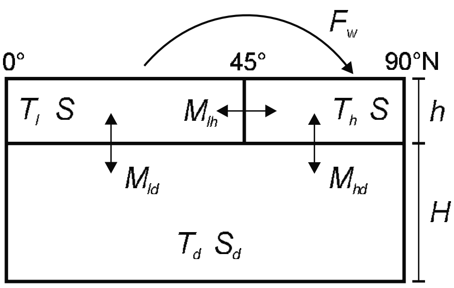

Figure 1. Three-box model of the North Atlantic Ocean (after [*Winton*, 1993]) consisting of low (subscript l) and high (subscript h) latitude surface boxes (height h = 100 m) and deep (subscript d) box (height H = 1000 m). Longitudinal extent of all boxes is 60°. Temperatures in surface boxes are constant (T_l = 15°C; T_h = 0°C) whereas temperature in deep box (T_d) and salinities (S) in all boxes are computed. Mixing M_{ij} between boxes i and j occurs at timescale τ_{ij} such that $M_{ij} = d_{ij}/\tau_{ij}$, where d_{ij} is the exchange-area-weighted distance between box centers ($\tau_{ld} = \tau_{hd}$ = 400 yr and τ_{lh} = 5 yr). Using a quadratic equation of state, M_{hd} is increased by a factor C_h = 10 if the high latitude water column is gravitationally unstable; else C_h = 3 is adopted to account for partial shutdown of multiple convection sites. F_w is low-to-high latitude freshwater flux.

the model stays permanently in the "interstadial" deep-coupled phase. For F_w > 0.875 m/yr a permanent halocline develops in the high-latitude surface box, leading to a continuous "stadial" deep-decoupled phase. The implied freshening of the high-latitude surface box along with increasingly colder climate conditions is consistent with enhanced runoff from growing ice-sheets [*Marshall and Clarke*, 1991] and is also supported by results from coupled ocean-atmosphere models [*Ganopolski and Rahmstorf*, 2001; *Khodri et al.*, 2001]. For intermediate F_w values, the length of the interstadials decreases with increasing F_w while that of the stadials increases (Figure 2). A similar relationship between climatic background state (represented by the magnitude of F_w) and the "interstadial-stadial ratio" was derived from palaeoclimatic proxy data [*Schulz*, 2002a].

[8] To represent HEs in the model, only the unstable portion of the continental ice involved in HEs is considered. HEs are parameterized as slow build-up of this ice (time-scale 10 kyr) and, after reaching a threshold, a rapid removal of the unstable ice within 500 yr. A single link exists between the ensuing ice-volume oscillation and the deep-decoupling oscillations: if a HE coincides with a stadial phase in the ocean model, high-latitude convection is suppressed (C_h = 0) until all unstable ice is lost. Conversely, HEs occurring during an interstadial have no effect on high-latitude convection, because the stability of the thermohaline circulation increases with its strength [*Lohmann and Schulz*, 2000; *Tziperman*, 2000]. With this model configuration, predicted HEs (Figures 3a and 3b) do neither occur near the end of a stadial, nor do they last for a full stadial period. Indeed, the varying phase relationship between deep-decoupling oscillations and HEs depends entirely on the periods of deep-decoupling and ice-volume oscillations and the initial conditions. Hence, this model

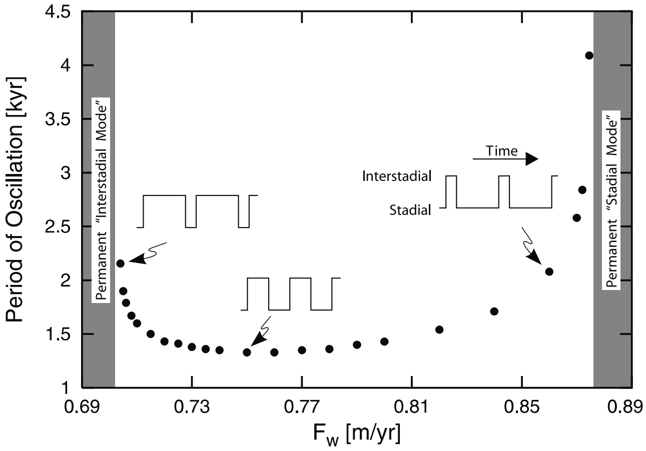

Figure 2. Period of the oscillation in the model as function of freshwater flux F_w (dots). Ratio between durations of coupled and decoupled phases decreases with increasing F_w, as indicated by square waves (not to scale).

setup fails to reproduce a crucial aspect of the palaeoclimatic evidence.

3. Relaxation Oscillators in Concert

[9] Synchronization of deep-decoupling oscillations and HEs can be achieved by adding an interaction mechanism between both relaxation oscillators [*Pikovsky et al.*, 2001]. Doubling the stadial ice-growth rate compared to the interstadial rate represents one possible feedback mechanism, and is in accordance with modeling results [*Marshall and Clarke*, 1999]. Adding this process to the model phase-locks the calving events to stadial intervals of the deep-decoupling oscillations (Figures 3c and 3d). Moreover, the duration of intervals without convection in high latitudes increases compared to the previous experiment and is as long as the

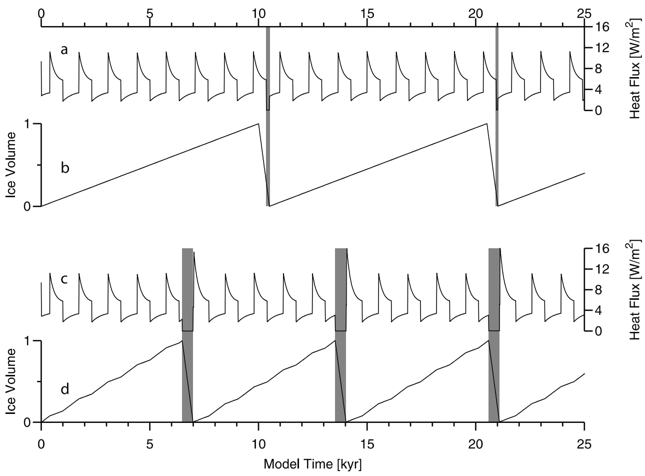

Figure 3. Synchronization of DO stadials and HEs. (a) Heat flux from deep to high-latitude surface box for F_w = 0.75 m/yr. (b) Free ice-volume oscillation. Concurrence of rapid ice volume decrease (HE) with phases of weak heat flux lead to shut down of convection in high latitudes (gray bars). (c, d) As (a, b) but with doubled ice-growth rate during stadials, leading to a synchronization between both free oscillations.

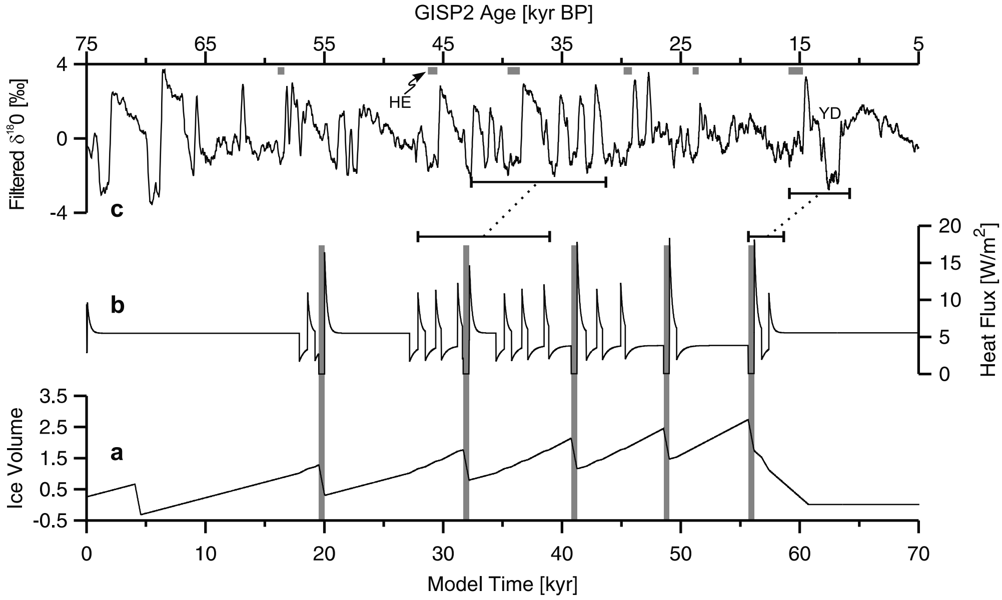

Figure 4. Modeled climate events and Greenland ice-core data. (a) Evolution of ice volume for an idealized glacial-interglacial cycle in the full model. (b) Heat flux from deep to high-latitude surface box. HEs associated with cessation of convection in high latitude are marked by gray vertical bars. (c) GISP2 oxygen-isotope record reflecting air-temperature above Greenland [*Stuiver and Grootes*, 2000]. For ease of comparison with time series in (b), data were band-pass filtered (0.2–12 kyr). Positions of YD and HEs (gray bars at top) are indicated. Horizontal lines denote sequences where modeled results closely resemble observed climate variability. Owing to the idealized model setup, no match is expected for the entire record.

calving events (500 yr). The higher stadial ice-growth rate leads to an earlier attainment of the threshold and, hence, a higher frequency of HEs, compared to the experiment with constant ice-growth rate (Figures 3b and 3d).

[10] Interruption of convection isolates the deep box from the high-latitude surface box, resulting in a warming of the deep box. In the synchronization experiment the 500-yr long interruption of convection leads to excess warming by about 2.5°C in the deep box (not shown). In contrast, the convection-free intervals in the experiment without synchronization are too short for a discernible excess heating of the deep box. Reestablishment of weak convection (C_h = 3) immediately after a HE results in a fast erosion of the halocline within ~20 yr. This is followed by the onset of strong convection (C_h = 10), which finally gives rise to a sudden release of the heat, accumulated in the deep box, to the high-latitude surface box. The heat flux during these flushing events exceeds the maximum values, which occur during the unperturbed deep-decoupling oscillations (Figure 3c). Thus, in this configuration, the model successfully simulates the occurrence of an anomalously large heat transfer at high latitudes following a HE.

4. Glacial Millennial-Scale Climate Oscillations

[11] To study the behavior of the full model during an idealized glacial cycle we assume an overall increase in ice volume over ~56 kyr, followed by a prescribed deglaciation during ~5 kyr. Ice-volume variations associated with HEs are superimposed on this long-term trend and include a doubled ice-growth rate during stadials. Freshwater forcing F_w depends on instantaneous ice volume as F_w [m/yr] = 0.5 + 0.2 × V_i, where V_i is dimensionless ice volume. Starting from V_i = 0.25, the initial value of F_w is too small to allow for deep-decoupling oscillations (cf. Figure 2) and the model remains in the interstadial mode (Figure 4). Around 18 kyr, F_w is sufficiently high to allow for the onset of deep-decoupling oscillations. Between ~20–27 kyr the reduction in F_w-values, associated with the loss of ice during the HE, leads to an intermittent cessation of the free deep-decoupling oscillation, giving rise to a long interstadial in this "early glacial" interval. From ~27 to 45 kyr, deep-decoupling oscillations with similar durations of stadials and interstadials appear.

[12] This sequence is interrupted by three HEs, which coincide with stadial phases and, therefore, lead to a shutdown of convection in high latitudes. Owing to the link between ice volume and F_w, the first interstadial after a HE lasts longer than the following warm phases, in full agreement with reconstructions (Figure 4c). Between ~45 and 56 kyr, high F_w values inhibit deep-decoupling oscillations and the model stays generally in the stadial mode, interrupted only by two HEs and associated interstadials. The deglaciation, which starts at 56 kyr, leads to a brief interval during which deep-decoupling oscillation become again possible. After ~61 kyr the model remains the interstadial mode.

[13] The appearance of deep-decoupling oscillations during the idealized glacial-interglacial cycle depends on the interaction between the timescale at which ice volume changes and the period of the deep-decoupling oscillations. The interaction works via the ice-volume dependency of F_w. During an interglacial-to-glacial transition, ice-volume changes at a timescale of $O(10^4$ yr) which is an order of magnitude larger than the period of the deep-decoupling oscillations (Figure 2). Hence, the system stays sufficiently long in a suitable F_w range to develop sustained deep-decoupling oscillations. A contrasting situation arises dur-

ing deglaciations when ice-volume varies at a timescale similar to the period of the deep-decoupling oscillations. Then, the system loses its capability to produce deep-decoupling oscillations shortly after the oscillations commence. Indeed, starting from a permanent stadial state, only a single oscillation may be generated, giving rise to the sequence interstadial – stadial, followed by a permanent interstadial. In the above experiment a HE is superimposed on this sequence (Figure 4b) and the resulting progression of events between $\sim$55 and 59 kyr is reminiscent of the sequence of events surrounding the YD (Figure 4c).

5. Implications

[14] If deep-decoupling oscillations are in fact representing DO-style oscillations, a variety of palaeoclimatic evidence can be accommodated within this framework: (*i*) The continuous shift from longer to shorter DO-interstadials during an interglacial-to-glacial transition [*Schulz*, 2002a]. (*ii*) Synchronization between DO-stadials and HEs by an ice-growth feedback and the occurrence of long DO-interstadials subsequent to HEs [*Bond et al.*, 1993]. (*iii*) YD-type events, resulting from the restart of the deep-decoupling oscillator after a glacial maximum. Accordingly, YD-type events are an expected response of the climate system during deglaciations and can be conceived as DO-stadials, overprinted by rapidly changing boundary conditions. Moreover, deep-decoupling oscillations can be paced by cyclic variations in high-latitude freshwater balance, which interrupt convection (not shown). This mechanism may account for the pacing of the Dansgaard-Oeschger interstadials by a basic period of 1470 yr [*Schulz*, 2002b].

[15] Although this conceptual framework captures important phenomena related to millennial-scale climate variability, it is not intended to disentangle the details of the underlying physical mechanisms. In particular, deep-decoupling oscillations require some substantial diapycnal mixing below the thermocline. Therefore, it is presently not known whether or not these oscillations have a counterpart in reality. Nevertheless, the concept of deep-decoupling oscillation leads to some testable predictions: Firstly, subsurface temperatures in the North Atlantic Ocean should increase by $\sim$2–3°C during DO-stadials, and should decrease during interstadials. Although no direct estimate of intermediate-water temperature variations exists as yet, benthic oxygen-isotope data [*Dokken and Jansen*, 1999; *van Kreveld et al.*, 2000] corroborate the inferred subsurface warming prior to or concurrent with stadial-to-interstadial transitions. Secondly, DO-style climate variations should be absent during interglacials and glacial maxima. This conjecture is supported by proxy records [*McManus et al.*, 1999; *Schulz*, 2002a].

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