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Glacial–interglacial contrast in climate variability at centennial-to-millennial timescales: observations and conceptual model

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Abstract

A set of published paleoclimate proxy records from the northern hemisphere, capturing different climate processes, is used to study glacial–interglacial differences in climate variability at centennial-to-millennial timescales during the past fifty thousand years. These proxy records reveal the existence of distinct oscillatory modes of the climate system. Glacial climate variability is dominated by a single mode, the Dansgaard–Oeschger cycles, composed of stadial and interstadial states. This glacial mode results in well-expressed covariations of the proxies, which are paced by a *fundamental* 1470-year signal. In contrast, there is no compelling evidence for a dominant and persistent centennial-to-millennial climate cycle during the Holocene. Interglacial climate variations seem to covary less pronounced than those of the last glacial period, suggesting the simultaneous activity of independent climate modes, each characterized by its own natural periods, between approximately 400–3000 years. A conceptual model is introduced to interpret this contrast in covariation at glacial–interglacial timescales. It is assumed that different climate modes can be represented by relaxation oscillators with different natural periods in the centennial-to-millennial band. Interactions among such oscillators may lead to a phase-synchronization and the development of a new climate mode with a joint frequency. We suggest that the coupled state with its synchronized dynamics resembles a glacial whereas the decoupled state represents an interglacial with its reduced covariations of climate fluctuations. The synchronization greatly enhances the frequency stability of the coupled system, and has the potential to reconcile the stability of the glacial 1470-year pacing cycle with an origin within the Earth's climate system.

1. Introduction

Late Pleistocene climate variability at centennial-to-millennial timescales differs significantly between glacials and interglacials. Whereas the former are dominated by rapid transitions between cold stadials and warm Dansgaard–Oeschger (DO) interstadials, the latter show only little variability around the mean state (McManus et al., 1999; Helmke et al., 2002; Pahnke et al., 2003). Despite intensive research efforts, the origin of the glacial 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; Schulz et al., 2002; Timmermann et al., 2003) over periodic calving of the Greenland ice sheet (van Kreveld et al., 2000), to external forcing mechanisms (Ganopolski and Rahmstorf, 2001). In addition, it remains also unresolved whether Holocene climate variability merely corresponds to extremely damped DO-type variations or represents a truly different state of the climate system.

Another complication in understanding late Pleistocene climate variations arises from the fact that the onset of DO-interstadials appears to be paced by a fundamental period of 1470 years, that is, they are separated by multiples of 1470 years (Schulz, 2002; Rahmstorf, 2003). Neither the origin of the pacing nor

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its apparent frequency stability, which has been estimated to range from ±12% to ±20% (Schulz, 2002; Rahmstorf, 2003), have been determined as yet. It has been suggested that this inferred stability of the pacing period makes an origin within the climate system rather unlikely (Wunsch, 2000) and that the regularity of the DO cycles would be easier to reconcile with an extraterrestrial cause (Berger and von Rad, 2002; Rahmstorf, 2003).

Using existing paleoclimate records we show that Holocene paleoclimate proxy records are incompatible with the derived glacial 1470-year pacing cycle. We introduce a conceptual model in which different interactions of oscillatory modes of the climate system are utilized to explain the apparent contrast between covarying glacial and largely independent interglacial climate fluctuations. Moreover, our conceptual model offers a means to reconcile the high frequency stability of the glacial 1470-year pacing with an Earth-bound origin.

2. Glacial–interglacial contrast in climate variability

It is our main interest to understand climate variability at centennial-to-millennial timescales at large spatial scales (i.e., regional to global) and across different components of the climate system. To investigate differences in climate variability between glacials and interglacials we choose two paleoclimate proxy records from the same archive covering a significant portion of a full glacial–interglacial cycle at multidecadal resolution. This approach minimizes any bias due to stratigraphic uncertainties as well as to possible offsets between different archives across glacial-to-interglacial transitions.

Our analysis is based on the $\delta^{18}O$ and potassium records from the Greenland GISP2 ice core (Mayewski et al., 1997; Stuiver and Grootes, 2000). To a first order $\delta^{18}O$ composition of the ice scales linearly with the air temperature above Greenland (Stuiver and Grootes, 2000), whereas the potassium concentration (Mayewski et al., 1997) is a proxy for the strength of the Siberian High (Meeker and Mayewski, 2002). (Since dry deposition at the GISP2 ice-core location appears to be negligible (Meeker et al., 1997), it follows that variations in ice-core ion concentration are proportional to changes in atmospheric ion concentration (cf. Alley et al., 1995). Accordingly, we follow Mayewski et al. (1997) and use potassium concentration instead of the accumulation rate to infer changes in atmospheric potassium loading.)

During the last glacial period both proxies recorded the sequence of DO events in a very similar manner (Fig. 1a), resulting in a high correlation between the two time series ($R^2 = 0.73$). Specifically, the proxy data suggest a strengthening of the Siberian High during stadials and a weakening during interstadials. In contrast, during the Holocene the records show no discernable similarity (Fig. 1b) and their correlation is greatly reduced ($R^2 = 0.08$) compared to the glacial interval. Even the largest event recorded in the $\delta^{18}O$ series during the Holocene, at ~8.2 ky BP (thousand years before present), has no corresponding counterpart in the potassium record. Nevertheless, both proxy records show clearly discernable variability at centennial-to-millennial timescales during the entire Holocene, which, however, seems to be unrelated between the records. Taken together, these observations suggest that the strength of the Siberian High and temperature above Greenland covary during the course of DO events, but fluctuate independent of each other in the absence of DO events.

The pronounced covariation of glacial DO-type climate fluctuations at a global scale is further supported by a recent compilation of glacial paleoclimate records (Voelker et al., 2002). According to this study, DO-type climate-change signals can be found in proxies characterizing the state of the atmosphere (e.g., large-scale circulation patterns; Mayewski et al. (1997)), the ocean (e.g., thermohaline circulation; Sarnthein et al. (2001)), and the carbon cycle (e.g., atmospheric methane concentration; Brook et al. (1999)). The notion of an ubiquitous DO-signal in the climate system does neither imply that all proxies record identical "patterns" of climate histories nor that all time series vary in phase which each other. As an example, consider the modified "see-saw" concept (Stocker and Johnsen, 2003), in which temperature in the North and South Atlantic Ocean *change* simultaneously, but in opposite direction. This leads to the generation of two signals with identical "patterns" which are out of phase by 180°. The delayed response of temperature in Antarctica results in a modified pattern of the temperature evolution in this region compared to the South Atlantic (Stocker and Johnsen, 2003). Despite these regional differences in the actual course of climate change, the entire system considered in this conceptual model is pervaded by the DO events.

Based on the chronology of the GISP2 $\delta^{18}O$ record (Stuiver and Grootes, 2000), the recurrence times of the onset of DO-interstadials occur as multiples of 1470 years (Schulz, 2002; Rahmstorf, 2003). Taken together with the well-expressed covariations of glacial climate fluctuations this finding implies commensurate recurrence times of climate events, which are linked to DO-type climate variability, irrespective of the paleoclimate proxy being considered. The estimated pacing period of the DO interstadials and its stability depend critically on the chronology of the GISP2 ice-core (Meese et al., 1997). Any inaccuracy of the ice-core chronology will likely affect both estimates. Unfortunately, available

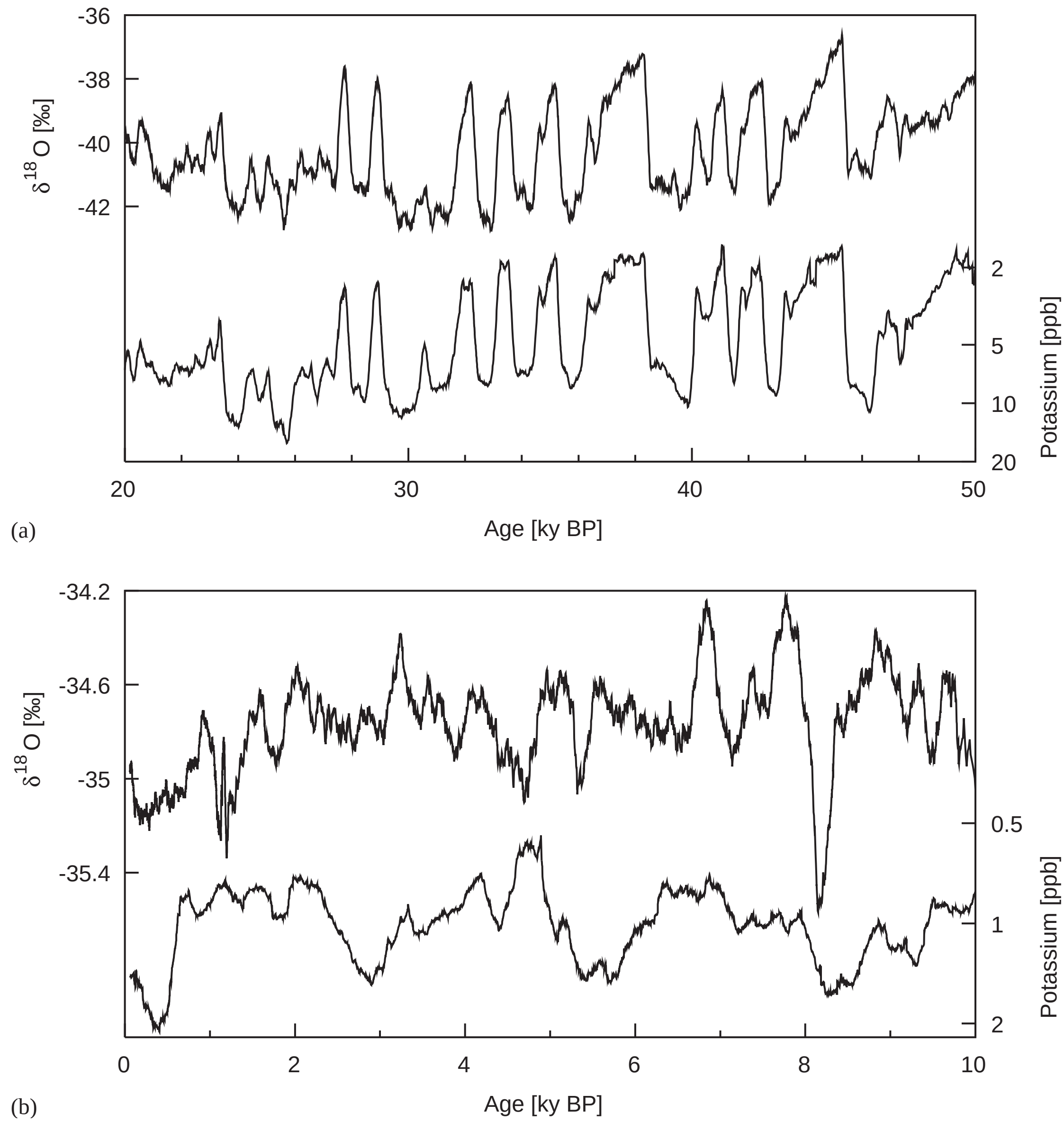

Fig. 1. Glacial–interglacial contrast in covariations of climate fluctuations. (a) GISP2 $\delta^{18}O$ (Grootes and Stuiver, 1997) and potassium concentration (Mayewski et al., 1997) during the last glacial period. Since our focus is on variations at centennial-to-millennial timescales, both series were smoothed with a 200-year running mean filter. Both records depict the same course of events. (b) as (a) but for the Holocene. In contrast to the glacial interval (a) the time series indicate no obvious covariations. Note that the length of the ages axes differs between (a) and (b).

radiometric datings (Wang et al., 2001; Spötl and Mangini, 2002; Burns et al., 2003; Genty et al., 2003) are inconclusive in this respect due to their uncertainties. With respect to possible modifications of the GISP2 chronology, we feel confident that the inferred high degree of covariations of glacial DO-type climate fluctuations is a robust feature.

To contrast the pronounced covariations of glacial climate fluctuations with the more independent interglacial temporal patterns further, we turn to a selection of Holocene paleoclimate records (Fig. 2). These records are thought to reflect various climate processes in the Atlantic region and encompass atmospheric and marine proxies. It is obvious from the inspection of Fig. 2 that most time series exhibit no obvious common denominator in terms of variability at centennial-to-millennial timescales. At first glance this lack of low-frequency climate variability is surprising, since some of the proxies are considered to be closely coupled due to their link to changes in North Atlantic deep-water production: drift-ice record (Bond et al., 2001); sea-surface temperature off the western Barents Sea (Sarnthein et al., 2003); $\delta^{13}C$ of overflow water (Oppo et al., 2003). The overall lack of a common temporal pattern is especially obvious, when focusing on different climate-proxy records from the same archive

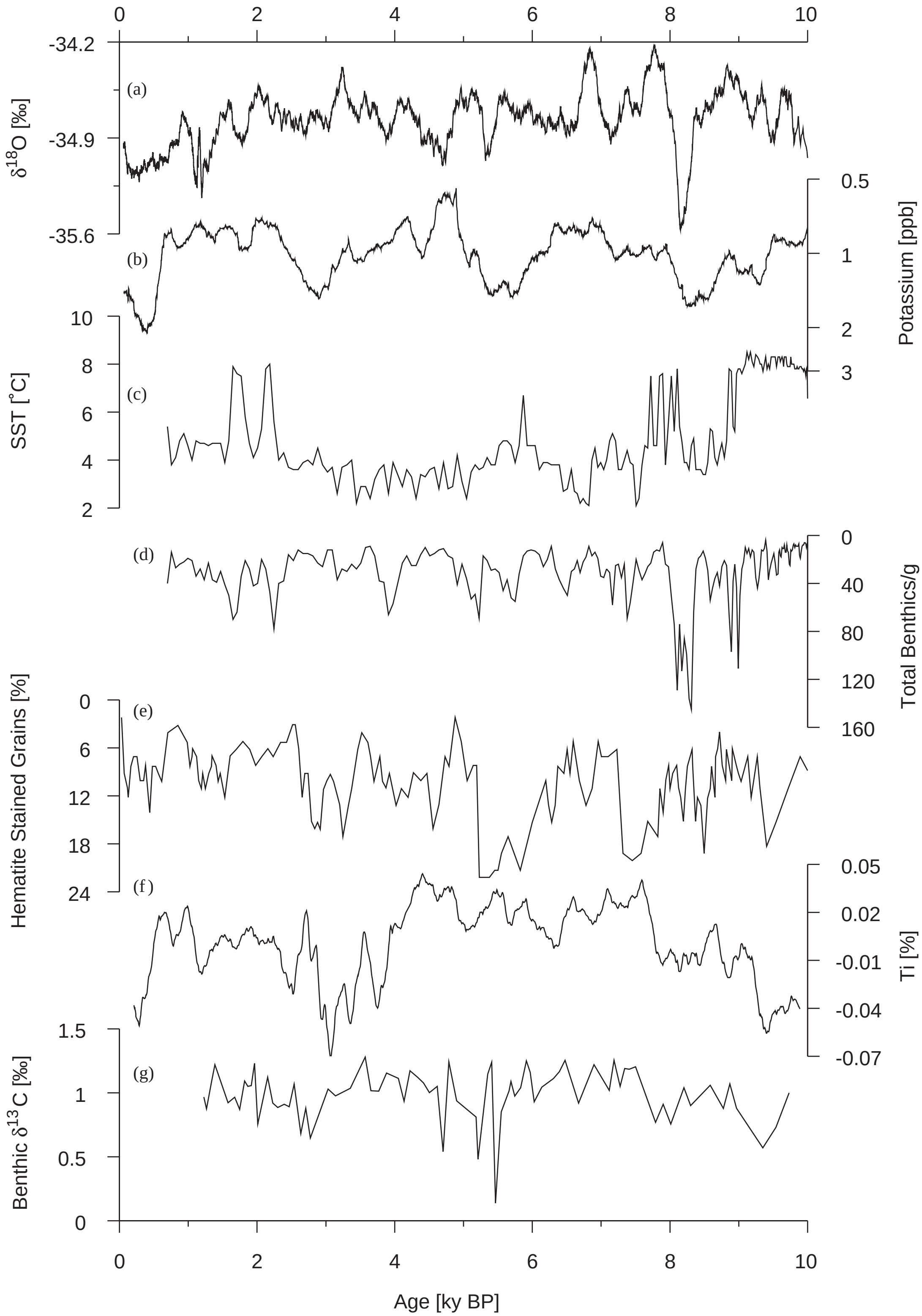

Fig. 2. Selection of Holocene paleoclimate records from the Atlantic region, indicating largely independent interglacial climate variations. (a) GISP2 δ^{18}O (Grootes and Stuiver, 1997), a proxy for air temperature above Greenland. (b) GISP2 potassium concentration (Mayewski et al., 1997), indicating the strength of the Siberian High (Meeker and Mayewski, 2002). (c) Reconstructed sea-surface temperature (SST) off the western Barents shelf (Sarnthein et al., 2003). (d) Total number of benthic foraminifera in a core off the western Barents shelf (Sarnthein et al., 2003), documenting winter storm activity. (e) Hematite stained grains in a sediment core off Ireland (Bond et al., 2001), measuring the amount of drift ice. (f) Titanium concentration (linearly detrended) in a Cariaco–Basin record, indicating changes in the hydrological cycle and river runoff (Haug et al., 2001). (g) Benthic δ^{13}C from south of Iceland (Oppo et al., 2003), interpreted to reflect changes in North-Atlantic deep-water formation. Time series in (a), (b) and (f) were smoothed with a 200-year running mean filter. Orientation of ordinate scales is such that inferred cold intervals point downward.

(Fig. 2a/b and c/d), that is, for situations in which any stratigraphic bias is rather small. Low signal-to-noise ratios and stratigraphic uncertainties could potentially hamper the comparison of low-frequency climate variability between the Holocene records. Whenever these obstacles are minimal, individual records documenting coupled climate processes may indeed reveal similar variations (e.g., the log-transformed potassium concentration (Mayewski et al., 1997; Fig. 2b) and titanium-concentration record (Haug et al., 2001; Fig. 2f) exhibit significant [$\alpha = 0.01$] cross-spectral coherency at a period of approximately 2500 years; not shown).

This picture of largely independent climate variability at centennial-to-millennial timescales during the Holocene is further supported by a large range of recurrence times between climate events that have been reported for different proxy records from the North-Atlantic region (e.g., O′Brien et al., 1995; Bond et al., 1997; Bianchi and McCave, 1999; Chapman and Shackleton, 2000; Schulz and Paul, 2002; Risebrobakken et al., 2003; Sarnthein et al., 2003; Hall et al., 2004). While the recurrence times range from approximately 400–3000 years, they appear to be clustered around 400–500 years and 900–1100 years, but not at approximately 1500 years or multiples thereof (Table 1). It should be kept in mind that the notion of recurrence time only reflects the fact that a record contains a distinct temporal pattern which is repeated after some time. Neither the exact repetition of such a pattern nor an exact timing of its recurrence is implied. Furthermore, reported recurrence times during the Holocene vary by as much as approximately $\pm 50\%$ within the same record (Bond et al., 1997; Sarnthein et al., 2003). In summary, analyses of many Holocene paleoclimate proxy records failed to confirm an ubiquitous glacial-type "1500-year cycle" (or multiples thereof) and display variations at centennial-to-millennial timescale that seem to be largely unrelated to each other.

Table 1
Selected Holocene (quasi)-periods from paleoclimate proxy records from the North Atlantic region (NATL = North Atlantic Ocean)

Proxy record	Period(s) (years)	Reference
NATL, drift ice	400–500 and 900–1100	Bond et al. (2001)
NATL, benthic foraminifera abundance	~400–1300	Sarnthein et al. (2003)
NATL, sediment color	550, 1000, ~1600 (?)	Chapman and Shackleton (2000)
NATL, planktonic foraminifera $\delta^{18}O$	550 and 1150	Risebrobakken et al. (2003)
NATL, mean size of sortable silt	400 and ~1000	Hall et al. (2004)
GISP2, $\delta^{18}O$	900	Schulz and Paul (2002)
GISP2, potassium	2600	O′Brien et al. (1995)

3. Glacial-type pacing of Holocene climate variations?

A record of hematite-stained grains from the northeastern Atlantic Ocean (Bond et al., 1997), which has been interpreted to reflect the amount of drift ice reaching this area, is of great importance for understanding climate variability at centennial-to-millennial timescales. The glacial variability in this record is tightly linked to the occurrence of DO events and, therefore, also reflects the fundamental 1470-year pacing associated with these events. In contrast to other climate proxy data sets, the millennial-scale climate variability documented by this record is unaffected by the last glacial–interglacial transition giving rise to a Holocene recurrence time of inferred cold events of 1470 ± 500 years (Bond et al., 1997). Moreover, it has been postulated that the Holocene variations seen in this data set are controlled by variations in solar output (Bond et al., 2001).

If the Holocene drift ice-events are indeed manifestations of muted glacial-type DO stadials (Bond et al., 1997), one would expect the Holocene recurrence time of drift-ice anomalies to be consistent with the fundamental 1470 year pacing of the glacial DO events. To conclude for such a consistency two aspects have to be fulfilled: (i) a match of the period of the fundamental pacing cycle across the glacial termination and (ii) a consistent phase of a pacing cycle between glacial and interglacial.

To test if the drift-ice anomalies are consistent with the estimated glacial 1470 year pacing cycle, we extrapolate the 1470 year template for the onset of glacial DO interstadials (Schulz, 2002) into the Holocene (Fig. 3). If the timing of events is unaffected by the last deglaciation, the predicted times from the template should coincide with transitions from drift-ice maxima to minima, i.e., inferred warmings in the North Atlantic. For this comparison we take into account the $\pm 12\%$ uncertainty estimated for the glacial pacing period (Rahmstorf, 2003). During the last ~10 ky, the drift-ice record shows nine clearly discernable "warmings" (Fig. 3). However, a match between the actual timing of the "warmings" and those predicted by the template occurs only in three cases. Based on a binomial distribution (cf. Schulz, 2002) the probability for three matches out of nine "trials" by chance is 22%. Hence, the occurrence of Holocene climate events, as recorded by the drift-ice proxy, is consistent with a random origin and does not require pacing by the glacial 1470 year cycle. This suggests that the Holocene climate events recorded in this proxy record may not

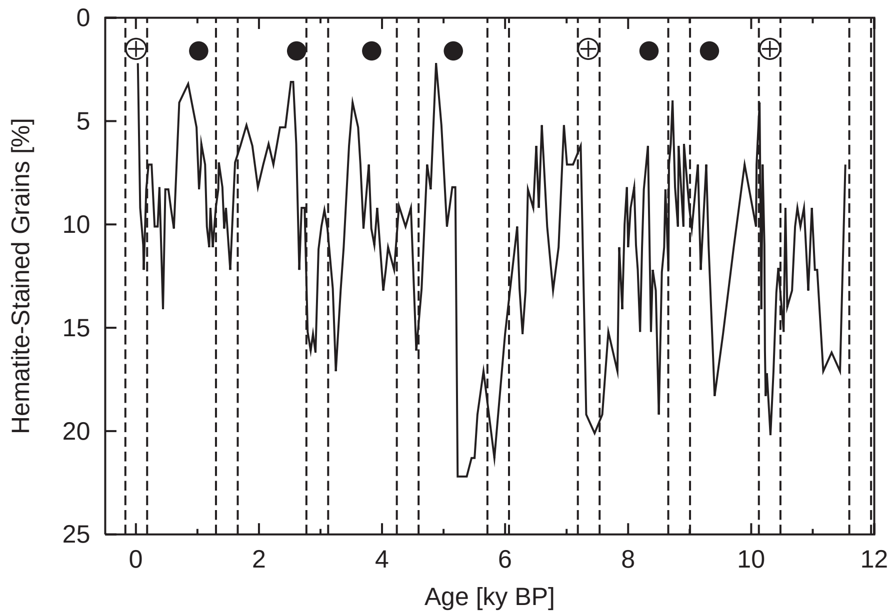

Fig. 3. Drift-ice record from site VM 29-191 from the northeastern North Atlantic (Bond et al., 2001, solid line) and 1470-year pacing template (dashed lines) extrapolated from the last glacial period. Adjacent dashed lines indicate the $\pm 12\%$ uncertainty associated with the pacing (Rahmstorf, 2003). Orientation of the drift ice-curve and the template are such that warm events appear at the top of the graph. Encircled plus-signs indicate where the onset of warming events in the drift-ice record coincides with the onset predicted by the template, whereas filled circles indicate mismatches. The number of matches is too small to exclude a random origin. The pacing template was constructed by fitting a trapezoidal time-series model to Dansgaard–Oeschger events 5–7 in the GISP2 $\delta^{18}O$ series (see Schulz, 2002 for details).

be a simple (muted) continuation of glacial Dansgaard–Oeschger events. (This finding is robust with respect to the technique used to estimate the glacial pacing of the DO events (Schulz, 2002; Rahmstorf, 2003). While both studies suggested identical periods of the pacing cycle, the actual templates are offset by 166 years.)

Bond et al. (2001) reconstructed variations in solar output by means of ^{10}Be and ^{14}C measurements and argued that the Holocene climate variations are controlled by fluctuations in solar forcing. We use a combined ^{10}Be record (Finkel and Nishiizumi, 1997; Yiou et al., 1997; converted to fluxes as in Bond et al. (2001)) to test if the reconstructed Holocene variations in solar forcing are compatible with the glacial-type 1470 year pacing, following the same approach as before. Between 3 and 11 ky BP the ^{10}Be accumulation rate shows four transitions between high and low values, which are thought to correspond to climate "warmings" (Bond et al., 2001). Only one out of four of these transitions is consistent which the glacial 1470 year pacing cycle (Fig. 4). This finding strongly suggests that the glacial 1470 year pacing cycle is not controlled by variations in solar forcing, unless one accepts the unlikely scenario that variations in solar-controlled climate pacing change across glacial–interglacial transitions. Hence, we have to seek for an alternative origin of the glacial 1470 year pacing cycle and its perplexing frequency stability.

4. Modeling the synchronization of interacting oscillators

The question arises whether the difference between covarying glacial climate variations and more independently fluctuating interglacial climate variations at centennial-to-millennial timescales holds the key for explaining the regularity of the glacial pacing cycle. To address this question, we consider a collection of oscillators within the climate system, which are thought to represent various oscillatory modes at centennial-to-millennial timescales. Examples for such oscillators include ice caps (e.g., MacAyeal, 1993) or the ocean–atmosphere system (e.g., Winton, 1993). Our working hypothesis is that during glacials a physical process is present which couples these oscillators, whereas the coupling is absent or insignificant during interglacials. Such a mechanism may then explain the lack of covariance in low-frequency climate fluctuations among climate proxies during the Holocene, which is indicative of the simultaneous presence of independent climate modes (see Sections 2 and 3).

To elucidate this contrast in synchronization, we explore the potential of the so-called mean-field coupling of oscillators (e.g., Pikovsky et al., 2001). We illustrate this concept by considering three different oscillatory modes that are represented by relaxations oscillators with different periods of approximately 400, 800 and 2400 years (Fig. 5; see Appendix A for details). These periods were chosen to represent the periods of

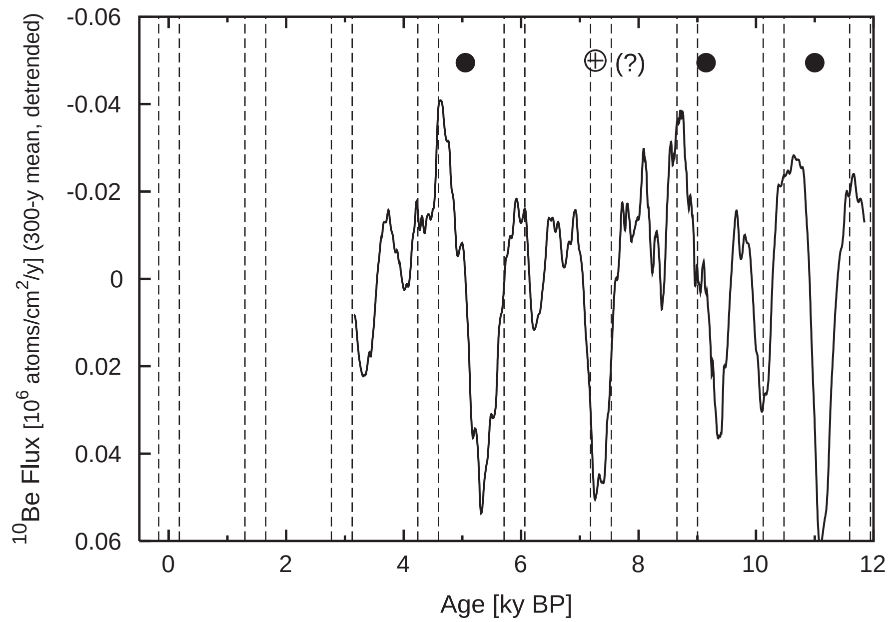

Fig. 4. Reconstructed solar forcing (Bond et al., 2001, solid line) versus 1470-year pacing template (as in Fig. 3). Changes in solar forcing are inferred from the beryllium-10 flux derived from Greenland ice-core data (Bond et al., 2001 and refs. therein). Encircled plus-sign indicates a match between the onset of a "warming event" in the ^{10}Be record with the onset predicted by the template, whereas filled circles indicate mismatches. The overall mismatch makes a solar origin of the 1470-year pacing cycle unlikely. The ^{10}Be data were smoothed with a 300-year running-mean filter and subsequently detrended. Orientation of ^{10}Be series and template is such that warm events appear at the top of the graph.

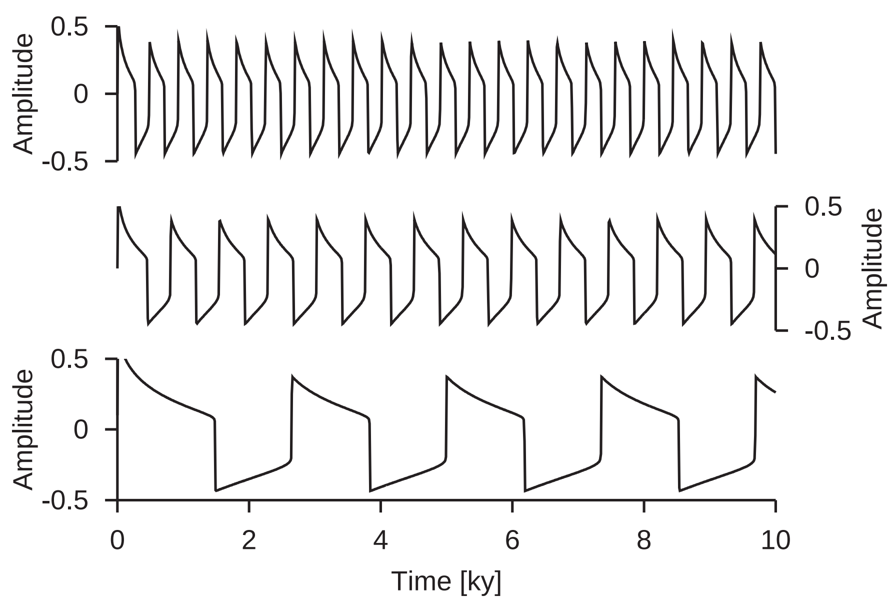

Fig. 5. Relaxation oscillations of three uncoupled oscillators with periods of approximately, 400, 800 and 2400 years (see appendix A for details).

the identified clusters of low-frequency climate variability during the Holocene (Table 1). It should also be noted that the actual number of oscillators is not critical for the following discussion, as long as at least two oscillators are considered. Initially, the oscillators are uncoupled, that is, they are not "aware" of each other and can thus oscillate at their "own" frequencies (Fig. 6). After 10 ky each oscillator is allowed to influence the remaining oscillators (see Appendix A for details). As a result, the three oscillations become synchronized within approximately 1000 years. Even more important is that the period of the synchronized oscillations of 1300 years differs from the periods of the uncoupled oscillators. In other words: a new timescale, corresponding to the natural period of a new oscillatory mode, arises from the synchronization of oscillators. (This behavior is not

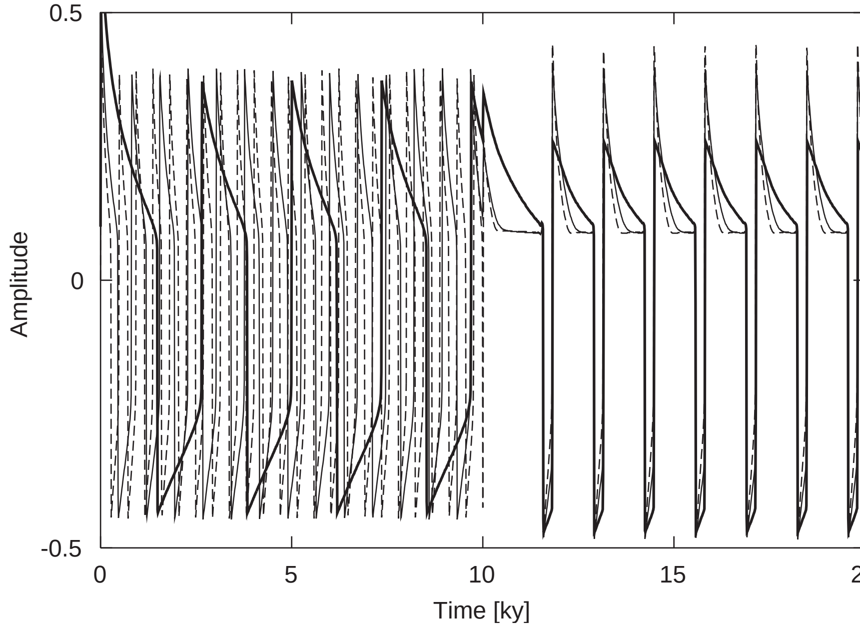

Fig. 6. Synchronization of relaxation oscillators. Between 0 and 10 ky the three oscillators are uncoupled ($\alpha = 0$), leading to independent oscillations (as in Fig. 5). At 10 ky the coupling strength is increased instantaneously to $\alpha = 0.3$, resulting in a synchronization of the oscillators within ~1 ky. The period of the synchronized oscillations is 1300 years.

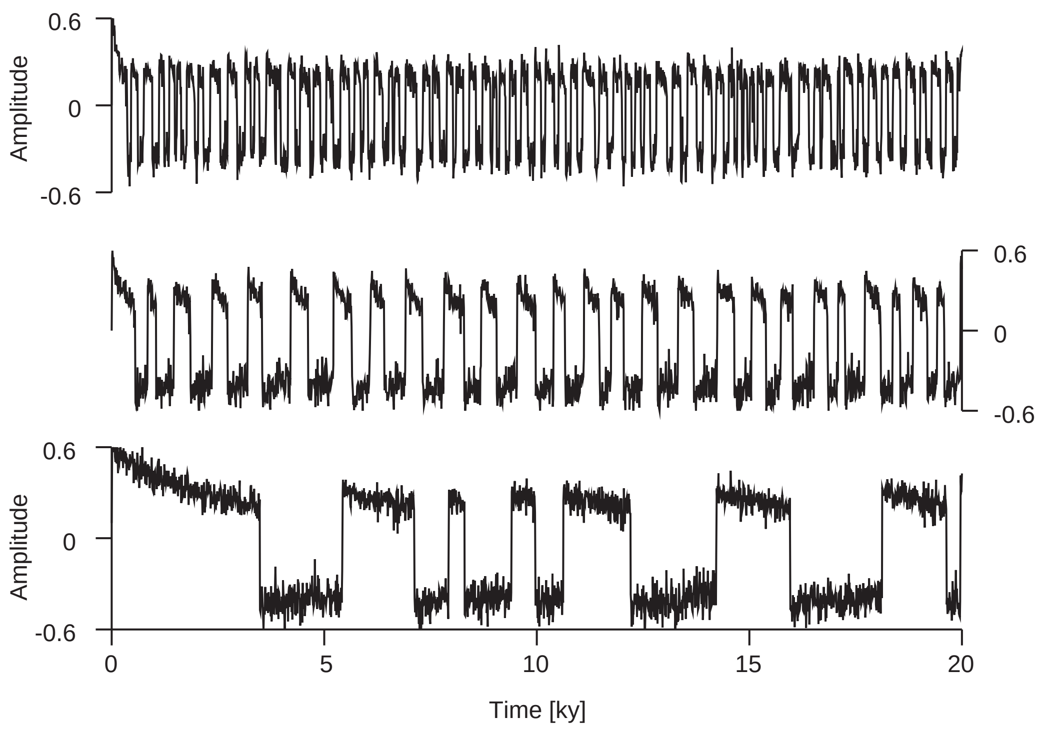

Fig. 7. Relaxation oscillations of three uncoupled oscillators with randomly perturbed periods of approximately 300 ($\pm$20%), 700 ($\pm$20%) and 2500 ($\pm$50%) years (see appendix A for details).

limited to three oscillators and can easily be achieved with a larger number of oscillators; not shown.)

In a second set of experiments, we explore whether the synchronization of the three oscillators works also in the presence of randomly driven frequency fluctuations. For this experiment the periods of the three oscillators are set to 300 ($\pm$20%), 700 ($\pm$20%) and 2500 ($\pm$50%) years (Fig. 7). Shortly after the oscillators have been coupled (at 20 ky), the three oscillations become again synchronized (Fig. 8). Thus, despite the presence of

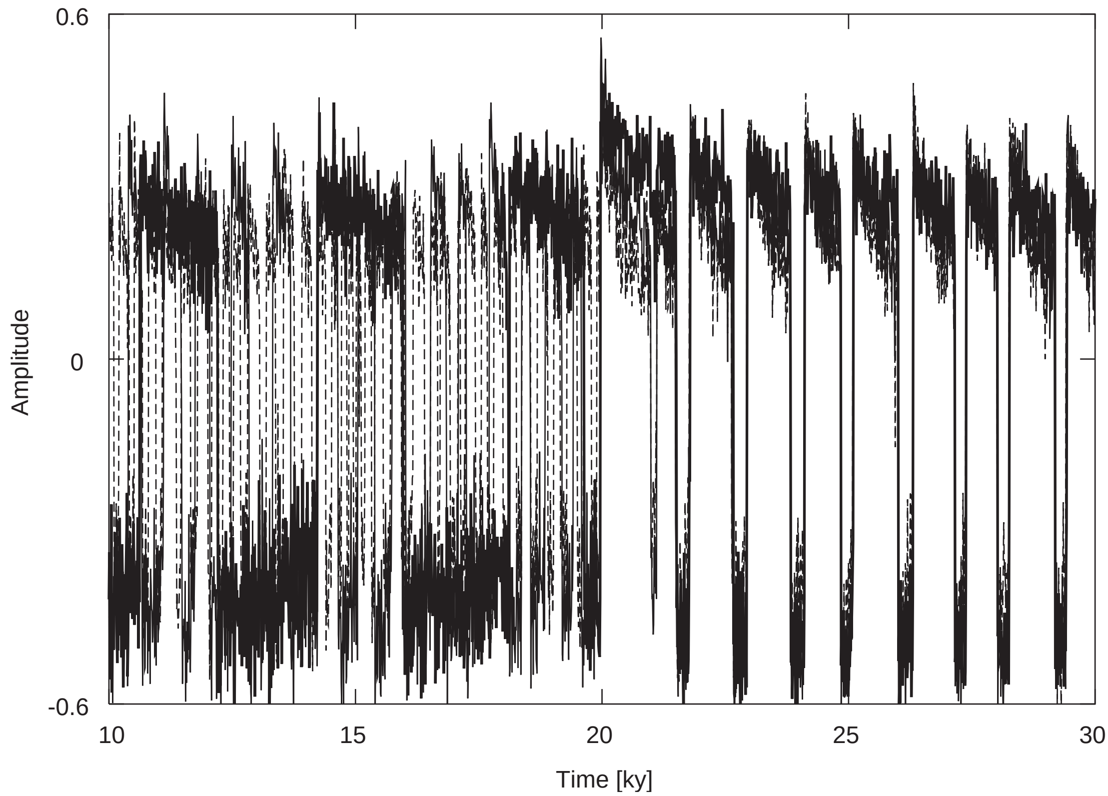

Fig. 8. Synchronization of relaxation oscillators in the presence of noise. Until 20 ky the three oscillators are uncoupled ($\alpha = 0$), leading to independent oscillations (as in Fig. 6; for clarity, only the portion after 10 ky is shown). At 20 ky the coupling strength is increased instantaneously to $\alpha = 0.6$, resulting in a synchronization of the oscillators within ~1 ky. The period of the synchronized oscillations is 1100 years with a variation of $\pm 10\%$. The frequency stability of the synchronized oscillations is thus higher than that of the uncoupled oscillators.

noise, the mean-field coupling mechanism is still capable of synchronizing the oscillators. As in the previous experiment, the mean period of the synchronized oscillations (1100 years; Fig. 8) has no direct counterpart in the periods of the non-synchronized oscillations. Even more striking is the fact that the period of the synchronized oscillation varies by only $\pm 10\%$, that is, the frequency stability of the synchronized oscillations is better than that of any of its constituents. Hence, this experiment strongly suggests that the synchronization of oscillators with time-varying periods can result in remarkable stable oscillations once the oscillators become synchronized.

5. Making sense of the glacial–interglacial contrast in climate variability—a concept

Based on the synchronization experiments we suggest that the inferred contrast in glacial–interglacial climate variability can be understood in terms of the presence or absence of interactions among different low-frequency climate modes. Accordingly, interglacials correspond to an uncoupled (or weakly coupled) state, whereas glacials provide a mechanism to couple low-frequency modes, thereby synchronizing their dynamics and giving rise to a new mode that dominates the entire climate system.

Our heuristic approach involving different interactions of oscillatory modes of the climate system between glacials and interglacials leads immediately to the following questions: (i) Which components of the climate system give rise to these modes and what are their "natural" periods? (ii) What controls the coupling strength between independent modes at glacial–interglacial timescales? (iii) What are the physical mechanisms that provide the actual coupling (communication) between the modes?

Based on our evaluation of paleoclimate proxy data and our conceptual model we can give some tentative answers to these questions. We set out by assuming that the identified variance peaks at timescales of approximately, 400–500, 900–1100 and maybe 2600 years during the Holocene (Table 1), are the result of oscillatory climate modes. We note that neither the period of the oscillation is assumed to be stable during the Holocene nor that the oscillations are present throughout the entire Holocene. So far, our understanding of natural climate variability at centennial-to-millennial timescales seems to be too limited (cf. Folland et al., 2001) to pin down the actual sources of the variability. However, it is reasonable to assume that components of the climate system are involved that have a sufficiently large "inertia" for generating low-frequency climate oscillations. Such components include the deep ocean, ice caps and glaciers, groundwater storage, as well as vegetation (e.g., Saltzman, 2002). Using a continuation method, Weijer and Dijkstra (2003) demonstrated that the large-scale ocean circulation can exhibit low-frequency modes. Further modeling experiments are required to test the robustness of this result as well as to identify other potential mechanisms for low-frequency oscillations within the global climate system.

The most obvious factor controlling the coupling strength during a glacial–interglacial cycle are northern-hemisphere ice sheets. While this control of the coupling strength is a rather abstract concept, the physically more challenging questions surround the mechanisms which provide the potential coupling between the independent oscillatory modes of the climate system. Without detailed knowledge of how these oscillatory modes are generated we can offer only some general ideas along these lines. Atmospheric circulation and the associated heat and moisture transports offer an efficient and fast link to couple the climate modes mentioned above. Interhemispheric linkages could be accommodated by the oceanic thermohaline circulation, whereas an effective coupling between low and mid latitudes is provided by the ventilation of the thermocline. Finally, sea-level could communicate between climate modes at a global scale.

From the previous discussion it should be clear that oscillators and communicators may not always be clearly differentiated. For example, the oceanic thermohaline circulation may be an integral part of an oscillatory mode and at the same time provide the coupling between different modes. Moreover, one can anticipate that the existence of at least some oscillatory modes may depend on the general state of the climate system (e.g., DO-type events triggered by massive meltwater events in a cold climate (Paul and Schulz, 2002; Timmermann et al., 2003)).

In summary, the invoked mean-field coupling of interacting oscillators is a potential mechanism to explain the glacial–interglacial contrast in covariance of climate variations and to synchronize oscillatory climate modes over a wide range of natural frequencies. Moreover, this mechanism can account for a high degree of frequency stability in the coupled system and may hold the key to reconcile the stability of the glacial 1470 year pacing cycle with an origin within the Earth's climate system. It is tempting to speculate whether a combination of (i) the synchronization between Heinrich Events and DO-cycles (Schulz et al., 2002) and (ii) the triggering of a sequence DO-events by a Heinrich Event could lead to a self-synchronization, capable of generating the observed sequence of glacial climate events and a high degree of frequency stability of a fundamental pacing cycle of the DO events (this will be the subject of a forthcoming study).

6. Conclusions

Our analysis of climate variations at centennial-to-millennial timescales revealed that glacial climate changes covary to a high degree and are locked to a common pacing cycle. In contrast, we found no compelling evidence for a dominant interglacial climate cycle. Specifically, Holocene millennial-scale climate variability does not appear to be a (muted) continuation of glacial Dansgaard–Oeschger events. Moreover, a solar origin of the estimated glacial 1470 year pacing cycle is unlikely (keeping in mind that this inference is based on a rather limited set of observations).

The notion of oscillatory climate modes offers a framework to explain the glacial–interglacial contrast in climate covariations at multicentennial-to-millennial timescales by means of different coupling strength between these modes. While strong coupling during glacials leads to synchronized variations corresponding to a single dominating climate mode, weak or absent coupling during interglacials allows the climate modes to exist independently of each other and to result in rather independent climate variations.

Mean-field coupling is capable of synchronizing oscillators over a wide frequency range, even in the presence of noise. The synchronization greatly enhances the frequency stability of the coupled system, and has the potential to reconcile the stability of the glacial 1470 year pacing cycle with an origin within the Earth's climate system.

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Appendix A

The Morris–Lecar system (after Somers and Kopell, 1995) was used to generate relaxation oscillations. The equations for the ith, $i = 1, \ldots, N$ oscillator read

$$\begin{aligned} \mathrm{d}v_i/\mathrm{d}t = & -a_1 m_\infty(v_i)(v_i - b_1) - a_2 w_i (v_i - b_2) \\ & - a_3(v_i - b_3) + \Omega_i, \end{aligned} \tag{A.1}$$

$$\mathrm{d}w_i/\mathrm{d}t = \varepsilon_i[w_\infty(v_i) - w_i]/\tau_w(v_i), \qquad (A.2)$$

where

$$m_\infty(v_i) = 0.5[1 + \tanh\{(v_i - c_1)/c_2\}],$$

$$w_\infty(v_i) = 0.5[1 + \tanh\{(v_i - c_3)/c_4\}],$$

$$\tau_w(v_i) = 1/\cosh\{(v_i - c_3)/c_5\}$$

and t denotes time. The following parameter values were employed for all simulations: $a_1 = 1.0$, $a_2 = 2.0$, $a_3 = 0.5$, $b_1 = 1.0$, $b_2 = -0.7$, $b_3 = -0.4$, $c_1 = -0.01$, $c_2 = 0.15$, $c_3 = 0.1$, $c_4 = 0.145$, $c_5 = 0.29$. Furthermore $\Omega_i = 0.1\ \forall i$ was applied in the standard case, whereas random perturbations were achieved by setting $\Omega_i = \zeta_i(t)$ where ζ_i is an evenly distributed random number in [–1,1]. The value of ε_i in Eq. (A.2) controls the frequency of the oscillation. For the standard case we used $\varepsilon_i = \{0.0004, 0.0014, 0.0024\}$ for the $N = 3$ oscillators, whereas in case of random perturbations the following values were adopted: $\varepsilon_i = \{0.00008, 0.00058, 0.00108\}$. Global mean-field coupling of the oscillators was achieved by adding the term

$$-\alpha\kappa a_1(v_i - b_1)$$

to Eq. (A.1), where

$$\kappa = \frac{1}{N}\sum_{i=1}^{N} 0.5[1 + \tanh\{(v_i - c_6)/c_7\}].$$

Here α determines the coupling strength between the oscillators. Values of α ranged from 0.0 to 0.6 as stated in the figure captions. Additional parameter values are $c_6 = 0.05$ and $c_7 = 0.15$. A fourth-order Runge–Kutta scheme was used for integrating the system of equations.

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