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Climate bifurcation during the last deglaciation

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Abstract

The last deglaciation was characterised by two abrupt warming events, at the start of the Bølling-Allerød and at the end of the Younger Dryas, but their underlying causes are unclear. Some abrupt climate changes may involve gradual forcing past a bifurcation point, in which a prevailing climate state loses its stability and the climate tips into an alternative state, providing an early warning signal in the form of slowing responses to perturbations. However, the abrupt Dansgaard-Oeschger (DO) events during the last ice age were probably triggered by stochastic fluctuations without bifurcation or early warning, and whether the onset of the Bølling-Allerød (DO event 1) was preceded by slowing down or not is debated. Here we show that the interval from the Last Glacial Maximum to the end of the Younger Dryas, as recorded in three Greenland ice cores with two different climate proxies, was accompanied by a robust slowing down in climate dynamics and an increase in climate variability, consistent with approaching bifurcation. Prior to the Bølling warming there was a robust increase in climate variability but no consistent slowing down signal, suggesting this abrupt change was probably triggered by a stochastic fluctuation. The Bølling warming marked a distinct destabilisation of the climate system, which excited an internal mode of variability in Atlantic meridional overturning circulation strength, causing multi-centennial climate fluctuations. There is some evidence for slowing down in the transition to and during the Younger Dryas. We infer that a bifurcation point was finally approached at the end of the Younger Dryas, in which the cold climate state, with weak Atlantic overturning circulation, lost its stability, and the climate tipped irreversibly into a warm interglacial state. The lack of a large triggering perturbation at the end of the Younger Dryas, and the fact that subsequent meltwater perturbations did not cause sustained cooling, support the bifurcation hypothesis.

CPD

8, 321–348, 2012

1 Introduction

Human civilisations have developed in the stable climate of the Holocene interglacial epoch, but now there is concern that human activities could tip the climate out of its current regime (Alley et al., 2003; Lenton et al., 2008; Lenton, 2011). An important question is whether any early warning can be provided of such an approaching threshold change or "tipping point" in the climate system (Lenton, 2011). In general, for a system slowly approaching a threshold where its current state becomes unstable, and it transitions to some other state, one can expect to see it become more sluggish in its response to small perturbations (Scheffer et al., 2009). Mathematically speaking, for systems gradually approaching a co-dimension-1 bifurcation point in their equilibrium solutions, the real part of their leading eigenvalue tends toward zero, indicating a tendency toward infinitely slow recovery from perturbations. This phenomenon, known as "critical slowing down" in dynamical systems theory, has long been appreciated in physics (Wiesenfeld and McNamara, 1986) and ecology (Wissel, 1984), but has only recently been applied to climate dynamics, where it can be detected as increasing auto-correlations in time-series data approaching abrupt transitions (Held and Kleinen, 2004; Livina and Lenton, 2007; Dakos et al., 2008; Lenton, 2011).

Paleo-data approaching past abrupt climate changes provide a testing ground for this proposed early warning indicator (Livina and Lenton, 2007; Dakos et al., 2008), in particular, the last deglaciation was characterised by several abrupt climate changes, notably warming at the start of the Bølling-Allerød period, cooling into the Younger Dryas, and warming at the end of the Younger Dryas (into the Preboreal). The first identification of early warning signals in paleo-climate data was found across the entire deglacial sequence in Greenland ice core (GISP2) data, and tentatively associated with the Preboreal onset at the end of the Younger Dryas (Livina and Lenton, 2007) (although it could equally have been linked to the Bølling warming). Subsequent work found an early warning signal prior to the Bølling warming in the same GISP2 data, and also discovered early warning prior to the end of the Younger Dryas in marine sediment data from the tropical Atlantic (Cariaco basin) (Dakos et al., 2008). However, a recent analysis of higher-resolution North Greenland (NGRIP) ice core data, failed to find early warning signals prior to any of the abrupt Dansgaard-Oeschger (DO) events during the last ice age, including the Bølling warming (Ditlevsen and Johnsen, 2010). These events have been characterised instead as noise-induced transitions between pre-existing states (or attractors) in the climate system (Ditlevsen and Johnsen, 2010; Livina et al., 2010) (see Fig. 1 of Lenton, 2011). Whereas slow forcing past a bifurcation point should show the early warning signal of critical slowing down, rapid forcing (stochastic or otherwise) between attractors is not expected to (Ditlevsen and Johnsen, 2010; Lenton, 2011).

Rising variance has also been suggested as an early warning signal prior to abrupt transitions (Carpenter and Brock, 2006), but there is an ongoing debate as to how robust an indicator it is (Dakos et al., 2012; Ditlevsen and Johnsen, 2010). Following the fluctuation-dissipation theorem, it has been argued that both rising variance and rising autocorrelation must be detected together to have a robust early warning signal of approaching bifurcation, and that the ratio of variance to correlation time should remain constant, set by the noise intensity (assuming it is constant) (Ditlevsen and Johnsen, 2010). However, several counter-examples have recently been provided, where systems show rising autocorrelation prior to bifurcations, but variance does not rise, and indeed may decrease (Dakos et al., 2012).

Here we attempt to resolve existing ambiguity regarding the nature of abrupt climate changes during the last deglaciation by applying these early warning methods to two different proxies from three Greenland ice-cores: GRIP (Greenland ice core project) (Dansgaard et al., 1993), NGRIP (North Greenland ice core project) (NGRIP, 2004), and GISP2 (Greenland ice sheet project 2) (Alley, 2004). In particular, we examine whether slowing down detected across the whole deglaciation (Livina and Lenton, 2007) is robust and present in multiple Greenland ice-core records. We also try to resolve whether slowing down precedes the Bølling warming (Dakos et al., 2008), or not (Ditlevsen and Johnsen, 2010), and whether slowing down detected during the Younger

CPD
8, 321–348, 2012

Dryas in the tropical Atlantic (Dakos et al., 2008) is also present in Greenland. We use two different measures of increasing correlation (across different timescales) to look for critical slowing down, and also monitor variance, conducting a sensitivity analysis for the statistical parameters in our methods, to assess the robustness of our early warning signals. In the Discussion we relate our findings to other paleo-data and to dynamical models, which also disagree over the mechanisms underlying abrupt deglacial climate changes (Weaver et al., 2000; Liu et al., 2009; Ganopolski and Rahmstorf, 2001).

2 Methods

2.1 Data

The GRIP, NGRIP and GISP2 ice cores have recently been synchronised on the Greenland Ice Core Chronology 2005 (GICC05) time scale through the last deglaciation (Rasmussen et al., 2008) and these data form the core of our analysis. Critical slowing down, if it occurs, is a property of the slowest decay mode of a system, so we concentrate primarily on 20-yr resolution ice core records, which aggregate over shorter timescale variability and fast decay modes. We also reanalysed an earlier version of the GISP2 δ^{18}O record on a different timescale (Alley, 2004), as used in previous work (Livina and Lenton, 2007; Dakos et al., 2008), and this gives similar results (not reported here).

For each ice core we examined two proxy records. We focus first on the δ^{18}O water isotope record, which is a proxy for past air-temperature, but can also be influenced by changing water source temperatures and snowfall seasonality. Secondly, we examined the [Ca^{++}] record, which represents dust from soil-derived carbonates, and thus can capture changes in climate aridity, winds or dust source regions. [Ca^{++}] is greatest in cold, dry intervals and fluctuates over orders of magnitude, falling to very low levels in warm, wet intervals such as the Bølling-Allerød. Hence it is common to consider $\log_e([Ca^{++}])$, which shows a good anti-correlation with δ^{18}O and exhibits

CPD
8, 321–348, 2012

comparable fluctuations. We follow the convention of showing $-\log_e[Ca^{++}]$ for ease of visual comparison with $\delta^{18}O$.

The interval comprising the Bølling-Allerød and Younger Dryas periods, is too short ($n \sim 150$ points) to get reliable results with 20-yr resolution data. Hence we also examined annual resolution $\log_e(Ca)$ data from GRIP, in order to get a denser time-series. This dataset is on a different timescale and uses different units (but these changes are not critical to the results).

For each dataset, we extracted two different indicators of slowing down; the AR1 coefficient (ACF-indicator), and a rescaled DFA scaling exponent (DFA-indicator), and also monitored changes in variance.

2.2 Autocorrelation function (ACF-indicator)

Slowing down is measured by an increase in lag-1 autocorrelation, estimated by fitting an autoregressive model of order 1 (linear AR(1)-process) of the form; $y_{t+\Delta t} = c \cdot y_t + \sigma\eta_t$, using an ordinary least-squares fitting method, where η_t is a Gaussian white noise process of variance σ^2, and c is the autoregressive coefficient; $c = \exp(-\kappa\Delta t)$, where κ is the decay rate of perturbations. The decay rate of the major mode, κ, tends to zero (i.e. $c \to 1$) as bifurcation is approached (Held and Kleinen, 2004). The changing estimated value of the AR1 coefficient, c, as one moves through a time-series is referred to here as the "ACF indicator" – previously termed the "propagator" (Held and Kleinen, 2004).

2.3 Detrended fluctuation analysis (DFA-indicator)

Slowing down causes an increase in short-term memory, which is measured using detrended fluctuation analysis. DFA extracts the fluctuation function of window size s, which increases as a power law if the data series is long-term power-law correlated; $F(s) \propto s^\alpha$, where α is the DFA scaling exponent. We consider only the short-term regime, in which as $c \to 1$ and the data approach critical behaviour, the slowing

CPD
8, 321–348, 2012

exponential decay is well approximated by a power law in which $\alpha \rightarrow 1.5$ (corresponding to a random walk). The DFA exponent is rescaled to give a "DFA indicator" that has been calibrated against the ACF indicator for direct comparison, and reaches value 1 (rescaled from 1.5) at critical behaviour (Livina and Lenton, 2007).

2.4 Variance

We monitor variance, calculated as standard deviation, for comparison with previous work (Ditlevsen and Johnsen, 2010). If the fluctuation-dissipation theorem is applicable, then as a bifurcation is approached (and $\kappa \rightarrow 0$), variance of the system is expected to increase according to $\mathrm{Var}(y) = \sigma^2/2\kappa$, where σ^2 is the variance of the noise (Ditlevsen and Johnsen, 2010). However, variance may not increase if, for example, critical slowing down reduces the capacity of a system to follow high frequency fluctuations, or if a system becomes less sensitive to stochastic fluctuations as it approaches a threshold (Dakos et al., 2012).

2.5 Detrending

Due to non-stationarities in the paleo-climate records, it is necessary to remove trends before estimating the slowing down indicators or variance. DFA includes an inherent, internal detrending routine (Livina and Lenton, 2007), which is of low-order here and equivalent to simple linear detrending. Before calculating the ACF-indicator or variance we examined several de-trending approaches (Lenton et al., 2012) and chose Gaussian filtering, which fits a Gaussian kernel smoothing function across the whole record prior to transition (Dakos et al., 2008). Results are similar when Gaussian filtering is applied only within the sliding window (Lenton et al., 2012). The fit is subtracted from the record to obtain the residual data series. Bandwidth for the kernel determines the degree of smoothing, and should be chosen such that it neither over-fits the data nor filters out low frequencies in the record. Here a default bandwidth of 25 is typically used, but as part of our sensitivity analysis the bandwidth size of the Gaussian filtering is varied over 5–250 (Dakos et al., 2008).

CPD
8, 321–348, 2012

2.6 Sliding window length

All our indicators are estimated within a sliding window over a time series preceding the onset of a transition. The choice of the sliding window length is a trade-off between time-resolution (data availability), and reliability of the estimate for the indicators. Here, a default value of half the record length is used (Dakos et al., 2008), but sensitivity analyses were performed where the length of the sliding window was varied from 25 % of the record length up to 75 % using increments of 20 points.

2.7 Indicator trends

Here we consider upward trends in both ACF and DFA indicators as sufficient to indicate critical slowing down, regardless of the trend in variance. We use our sensitivity analyses to assess the robustness and strength of the trends in all three indicators. Indicator trends are quantified using the nonparametric Kendall τ rank correlation coefficient (Kendall, 1948). This measures, in the range -1 to $+1$, the strength of the tendency of an indicator to increase (positive values) or decrease (negative values) with time, against the null hypothesis of randomness for a sequence of measurements against time (value approximately zero). We focus on trends in the indicators, because their absolute values are affected by the ratio between the timescale of the dynamics and the frequency of measurements. We plot each indicator value at the end of the sliding window from which it was calculated, to mimic the situation of not being able to access data from the future. However, if one were trying to estimate when critical behaviour will be reached in the future (indicators tending to 1), results should instead be plotted in the middle of the sliding window, and any trend extrapolated forward (with an appropriate error range).

CPD
8, 321–348, 2012

3 Results

We focus here on the interval from the end of DO event 2 ~ 22.9 ka (22 880 yr b2k on the GICC05 timescale), approaching the Last Glacial Maximum, to the abrupt warming at the end of the Younger Dryas ~11.7 ka (11 740 yr b2k). We have also examined an earlier interval of DO events (47–27 ka), but typically obtain ambiguous results, with no consistent trends in the indicators (results not shown). This is not surprising as the sliding window typically spans both cold stadial and warm interstadial states, and previous work suggests that these two states had conflicting stability trends through the interval 47–27 ka, the cold state becoming increasingly stable, whilst the warm interstadial state was losing stability (Livina et al., 2010). The warm climate state then regained stability during, or prior to, the deglaciation (Livina et al., 2010). With the exception of the Bølling-Allerød, our analysis here, over 22.9–11.7 ka, is dominated by the cold state and its changing stability properties.

3.1 The deglaciation as a whole

Simply detrending the δ^{18}O data from any of the three ice cores over 22.9–11.7 ka and examining the residuals, one can see by eye that over time, Greenland climate becomes prone to larger and longer fluctuations, particularly after the Bølling warming (Fig. 1). In other words the climate system becomes more sluggish in response to perturbations as it proceeds through the deglaciation. This signal of critical slowing down is examined in detail in Figs. 2 and 3.

The AR1 coefficient (ACF indicator) and standard deviation rise together and roughly in proportion in the GRIP δ^{18}O data (Fig. 2a) and in the GISP2 δ^{18}O data (Fig. 2b), as expected from the fluctuation-dissipation theorem (Ditlevsen and Johnsen, 2010). There is some decoupling between the rises in the AR1 coefficient and standard deviation in the NGRIP data (Fig. 2c), in particular, a downward trend in the variance after the Bølling warming, whilst the AR1 coefficient continues to rise. In GRIP, NGRIP and GISP2 δ^{18}O data, for both the ACF and DFA indicators of slowing down, and the variance, positive trends are robust across a wide range of window lengths and filtering bandwidths used in the analysis (Fig. 2, colour contour plot insets).

Analysis of the $\log_e[Ca^{++}]$ data from the three ice cores over 22.9–11.7 ka shows even more robust positive trends in the ACF and DFA indicators of slowing down, and the variance (Fig. 3). Even if one considers raw $[Ca^{++}]$ data there remain robust positive trends in ACF and DFA indicators of slowing down, but there are ambiguous trends in variance because $[Ca^{++}]$ is much less variable in the warm Bølling-Allerød interval which is toward the end of the time series (results not shown).

Looking across all the results for both proxies (Figs. 2 and 3), increases in the indicators are typically concentrated toward the end of the timeseries (associated with Bølling-Allerød and Younger Dryas), with upward jumps typically associated with the Bølling warming.

3.2 The run up to the Bølling warming

As existing studies disagree over whether slowing down precedes the Bølling warming (Dakos et al., 2008), or not (Ditlevsen and Johnsen, 2010), we analysed just the interval from the end of DO event 2 ~ 22.9 ka up to the abrupt warming marking the start of the Bølling period (DO event 1) ~14.7 ka (14 740 yr b2k).

In the run up to the Bølling warming in GRIP $\delta^{18}O$ data, the ACF indicator shows consistent speeding up whilst the DFA indicator shows ambiguous trends (Fig. 4a). In GISP2 $\delta^{18}O$ data (Fig. 4b), the ACF indicator generally shows slowing down, consistent with previously reported results (Dakos et al., 2008) (which are from a shorter interval of GISP2 $\delta^{18}O$ data on a different timescale), but the DFA indicator shows consistent speeding up. In NGRIP $\delta^{18}O$ data (Fig. 4c), the DFA indicator shows consistent speeding up whilst the ACF indicator gives ambiguous trends, consistent with previous results (Ditlevsen and Johnsen, 2010). All three $\delta^{18}O$ records show robustly rising variance (Fig. 4a–c), whereas previous analyses of NGRIP $\delta^{18}O$ suggested no trend in variance (Ditlevsen and Johnsen, 2010).

Analysis of the $\log_e[Ca^{++}]$ data from the three ice cores over 22.9–14.7 ka also shows mixed results for the ACF and DFA indicators (Fig. 5), with the example indicators (righthand panels) showing a general increase up to ∼17 ka but often a decline after that. Variance is also generally increasing when considering the data up to ∼17 ka but declines after that, leading to some ambiguity in the indicators.

Looking across all the results for both proxies (Figs. 4 and 5), there is no robust, widespread slowing down prior to the Bølling warming, but there is a robust increase in $\delta^{18}O$ variability (Fig. 4), and $\log_e[Ca^{++}]$ variability also increases up to ∼17 ka (Fig. 5).

3.3 The Bølling-Allerød and Younger Dryas in high resolution

If the slowing down detected across the deglaciation (Figs. 2 and 3) is not present in the run up to the Bølling warming (Figs. 4 and 5) then it must be associated with the interval containing the Bølling-Allerød and the Younger Dryas. However, the results (Figs. 2 and 3) may be influenced by inadequate detrending of the abrupt transition at the onset of the Bølling-Allerød. To address this, we initially analysed just the interval after it, from the start of the Bølling-Allerød to the end of the Younger Dryas ∼14.6–11.7 ka (14 600–11 740 yr b2k). However, this interval is too short ($n = 144$ points) to get reliable results with 20-yr resolution data (results not shown). Furthermore, although the onset of the Younger Dryas is not viewed as an abrupt transition in the NGRIP $\delta^{18}O$ data (Steffensen et al., 2008), the Bølling-Allerød and Younger Dryas may represent two different climate states, in which case including the transition between them will involve a shift in dynamics and influence the results.

This led us to focus on annual resolution GRIP $\log_e(Ca)$ data. However, we first need to examine whether it shows comparable behaviour to the 20-yr resolution data. Figure 6a shows analysis of the entire deglacial interval for comparison to Fig. 3a. The upward trend in variance is comparable and the DFA-indicator still shows robust slowing down (albeit with a weaker trend), however lag-1 autocorrelation (the ACF-indicator) shows different behaviour in annual data. Speeding up is generally detected, with only large filtering bandwidths revealing an overall slowing down signal. We infer that fast decay modes sped up whilst slower decay modes were slowing down, as seen in some model results (Lenton et al., 2012). Thus, lag-1 autocorrelation in annual data does not capture behaviour of the slowest decay mode that is pertinent to bifurcation detection.

With this caveat in mind, we looked at the Bølling-Allerød and Younger Dryas intervals (Fig. 6b–d). When we analyse the Bølling-Allerød and Younger Dryas together, including the transition between them, we find a robust upward trend in the DFA indicator (and also the ACF indicator) and a robust decline in variance (Fig. 6b). When we analyse the intervals separately, both are characterised by robustly declining variance (Fig. 6c, d). The DFA indicator shows no clear trend during the Bølling-Allerød suggesting unchanging stability properties (Fig. 6c). This is consistent with the onset of the Younger Dryas being caused by a perturbation. During the Younger Dryas there is some upward trend in the DFA indicator (albeit sensitive to the window length chosen) (Fig. 6d).

4 Discussion

Our core result is a robust overall slowing down in climate dynamics, accompanied by an increase in variability, through the last deglaciation. This strengthens an earlier result detecting critical behaviour in one Greenland ice core proxy record of the deglaciation (Livina and Lenton, 2007). The proportional rises in AR1 coefficient and variance, especially at the summit of Greenland (GRIP and GISP2 records), are consistent with the behaviour expected from the fluctuation-dissipation theorem on approaching a bifurcation (Ditlevsen and Johnsen, 2010). As the interval considered is dominated by a cold climate state, we infer that this is the state approaching a bifurcation during the deglaciation. However, locating the timing of bifurcation is more challenging and necessarily imprecise, because internal climate variability can trigger abrupt transitions before a bifurcation is reached. A simplified schematic model of our interpretation is given in Fig. 7.

Previously reported results showing critical slowing down prior to the Bølling warming in the GISP2 core δ^{18}O record (Dakos et al., 2008) are not supported by an alternative DFA indicator of critical slowing down in the record we analysed here, which is on a different timescale. Nor are they found in the GISP2 $\log_e$[Ca^{++}] record with either ACF or DFA indicators, or in the neighbouring GRIP core δ^{18}O record, or in the more distant NGRIP core δ^{18}O record. There is therefore no widespread, robust slowing down tendency prior to the Bølling warming, and no evidence that this abrupt climate change was caused by a bifurcation (in which the cold, dry climate state would have to have lost stability). This is consistent with the return to cold, dry conditions in the Younger Dryas, which suggests an alternative cold climate state remained stable during the early part of the deglaciation even though it was not being sampled during the Bølling-Allerød.

Instead we adopt the hypothesis that the abrupt Bølling warming event was caused by a stochastic fluctuation (Ditlevsen and Johnsen, 2010). This fluctuation could have taken the form of a rapid, large perturbation. The Atlantic meridional overturning circulation (AMOC) had been shut down (McManus et al., 2004) from ~17.5 ka, in response to catastrophic iceberg discharge into the North Atlantic (Heinrich event H1), and it resumed abruptly at the Bølling warming (McManus et al., 2004). The triggering fluctuation may have been a sudden cessation of meltwater discharge into the North Atlantic (Clark et al., 2001; Liu et al., 2009), or meltwater pulse 1A originating from Antarctica (Weaver et al., 2000). However, models require much larger freshwater perturbations to affect a transition than data constraints allow (Valdes, 2011), and they generally exhibit much lower internal variability than the real glacial climate. An alternative interpretation, following (Ditlevsen and Johnsen, 2010), is that the level of glacial climate variability was such that it could (very occasionally) tip the system into an alternative warm state. This interpretation is helped by the observation that the strongest signal prior to the Bølling warming is an increase in paleo-temperature variability as recorded by δ^{18}O (Fig. 4). The $\log_e$[Ca^{++}] records also show increasing variability up to the time of Heinrich event H1, but become more muted thereafter. We conclude that a switch

CPD
8, 321–348, 2012

between co-existing cold and warm climate states occurred at the Bølling warming, rather than a bifurcation involving the disappearance of the cold climate state.

The noise-induced switch at the Bølling warming may have been preceded by a bifurcation that re-created a stable, warm climate state, which had lost its stability at the last glacial maximum (Livina et al., 2010). However, the methods applied here would not be able to detect that, as the climate system was only sampling the cold climate state 22.9–14.7 ka, and they focus on deducing changes to its stability properties.

The Bølling warming marked a distinct destabilisation of the climate system. On going through the transition, autocorrelation and variance generally increase in the six ice core records (Figs. 2 and 3). This suggests the warm climate state that had been entered was less stable than the preceding cold state of the last glacial maximum. There is a clear shift to lower frequency fluctuations consistent with critical slowing down, with the Bølling-Allerød period being characterised by multi-centennial climate fluctuations, which can be seen in the $\delta^{18}O$ and $\log_e[Ca^{++}]$ data (Figs. 2 and 3). Their timescale is consistent with an internal low-frequency mode of variability of the AMOC found in models (Mikolajewicz and Maier-Reimer, 1990; Park and Latif, 2008). Furthermore, decreases in AMOC strength centred on ∼14.1 ka, ∼13.8 ka and ∼13.3 ka have been detected in proxy data (Hughen et al., 2000; Obbink et al., 2010), and linked to temperature minima in Greenland (Obbink et al., 2010), North America and Europe (Yu and Eicher, 2001) (known as the intra-Bølling cold period, Older Dryas, and intra-Allerød cold period, respectively). Fluctuations in the Eastward routing of freshwater from the Laurentide ice sheet occurred at these times (Clark et al., 2001; Obbink et al., 2010) and may have contributed to AMOC weakening (Obbink et al., 2010), but could equally be viewed as the result of fluctuations in AMOC strength affecting temperatures over the ice sheet (Clark et al., 2001). Thus, we hypothesise that the Bølling warming excited oscillations in the AMOC, coupled to the Laurentide ice sheet.

The overall shift to a less stable climate state, may have facilitated further abrupt changes. We find no overall trend in climate stability during the Bølling-Allerød period itself, although variability in high resolution GRIP Ca data clearly declines (Fig. 6c).

This is consistent with the onset of the Younger Dryas being caused by a perturbation rather than a bifurcation. Indeed, the cooling that occurred at the onset of the Younger Dryas ~12.9 ka, has been linked to the Northward draining of Lake Agassiz into the Arctic Ocean (Murton et al., 2010). Ventilation of the deep Atlantic may have temporarily ceased (McManus et al., 2004; Hughen et al., 2000), but North Atlantic deep water formation soon resumed (Elmore and Wright, 2011). Overall, the Younger Dryas was characterised by a weak but not collapsed AMOC (McManus et al., 2004), somewhat analogous to its cold (stadial) mode of operation during the ice age. Thus, the transition into the Younger Dryas can be interpreted as a switch back into a co-existing cold climate state, triggered by a large perturbation. However, the onset of the Younger Dryas was not an abrupt transition in the same sense that the Bølling warming and the end of the Younger Dryas were (Steffensen et al., 2008).

The transition into the Younger Dryas appears to be associated with some slowing down in high-resolution GRIP Ca data (Fig. 6b). The Younger Dryas itself has previously been found to exhibit slowing down in a high-resolution productivity proxy from a sediment core in the tropical Atlantic (Dakos et al., 2008) (Cariaco basin). We also find some signs of slowing down in high-resolution GRIP Ca data during the Younger Dryas, although fast decay modes captured by lag-1 autocorrelation speed up (Fig. 6d). It is tempting to infer that the cold climate state of the Younger Dryas experienced some ongoing destabilisation that had begun much earlier. A much clearer signal is a decline in variability in GRIP Ca data during the Younger Dryas (Fig. 6d), which is also seen in the tropical Atlantic (Cariaco basin) (Lenton, 2011). The abrupt warming at the end of the Younger Dryas ~11.7 ka lasted ~60 yr, an order of magnitude slower than the Bølling warming (Steffensen et al., 2008), suggesting somewhat different underlying dynamics. But without an obvious proximate trigger perturbation, the cause of the end-Younger Dryas abrupt climate change has received less attention than the Bølling warming.

Based on our results here and elsewhere (Livina and Lenton, 2007; Dakos et al., 2008; Livina et al., 2010), we suggest that the trend of climate destabilisation during

CPD
8, 321–348, 2012

the whole deglaciation approached a bifurcation point at the end of the Younger Dryas, in which the cold (stadial) climate state lost its stability. If a bifurcation was being approached, it would have made it easier for the Younger Dryas to be ended by a stochastic fluctuation, so the observed abrupt transition likely preceded any bifurcation. A slowing of sea-level rise during the Younger Dryas (Bard et al., 2010) suggests freshwater input to the North Atlantic from ice sheet melt was declining. In models, such a reduction in freshwater input can cause the AMOC to pass a bifurcation point in which the weak (stadial) overturning state loses stability (Rahmstorf et al., 2005). In support of this mechanism, there is evidence for abrupt strengthening of the AMOC at the end of the Younger Dryas (McManus et al., 2004). Furthermore, bifurcations of the AMOC in models show the same generic early warning signal of slowing down (Lenton et al., 2012). Finally, the observation that the subsequent meltwater pulses 1B (Bard et al., 2010) and the 8.2 ka event caused only a transient weakening of the AMOC (McManus et al., 2004) and associated cooling, rather than a sustained collapse, is consistent with the cold (stadial) state having lost its stability at, or soon after, the end of the Younger Dryas.

5 Conclusions

We detect robust critical slowing down through the last deglaciation, recorded in two proxies from three different Greenland ice-cores. The Bølling warming event marked a distinct destabilisation of the climate system, followed by oscillatory dynamics. However, the climate system remained bi-stable (with warm and cold states co-existing under a range of boundary conditions) until at least the end of the Younger Dryas. We infer that a bifurcation point was finally approached at the end of the Younger Dryas, in which the prevailing cold climate state, with weak Atlantic overturning circulation, lost its stability, and the climate tipped irreversibly into a warm interglacial state. The lack of a large triggering perturbation at the end of the Younger Dryas, and the fact that subsequent meltwater perturbations did not cause sustained cooling, are consistent with the

CPD
8, 321–348, 2012

bifurcation hypothesis. This interpretation implies that the Holocene epoch, at least at its outset, was characterised by only one stable global climate state, which hence was quite resilient to perturbations, particularly in the cooling direction. Our analysis here however does not address how subsequent changes in natural climate forcing affected the stability properties of the climate as the Holocene proceeded. Now that human activities are forcing the climate system in the warming direction, its underlying stability properties are likely changing again. The outstanding question is whether we could tip a similarly abrupt and irreversible exit from the Holocene.

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CPD
8, 321–348, 2012

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CPD
8, 321–348, 2012

Climate bifurcation during the last deglaciation

T. M. Lenton et al.

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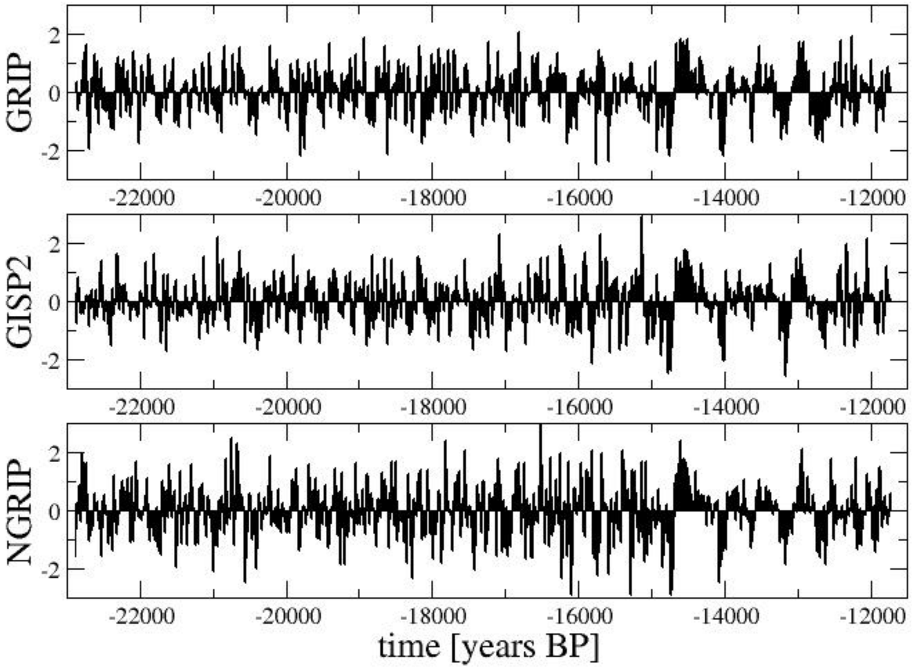

Fig. 1. Slowing down through the last deglaciation is visible by eye in detrended Greenland ice core paleo-temperature proxy data: **(a)** GRIP, **(b)** GISP2, **(c)** NGRIP. Example residuals after detrending δ^{18}O records 22.88–11.74 ka ($n = 558$), through the Last Glacial Maximum to the end of the Younger Dryas, using a filtering bandwidth of 25, showing pronouced slowing down, especially after the Bølling warming (~14.7 ka).

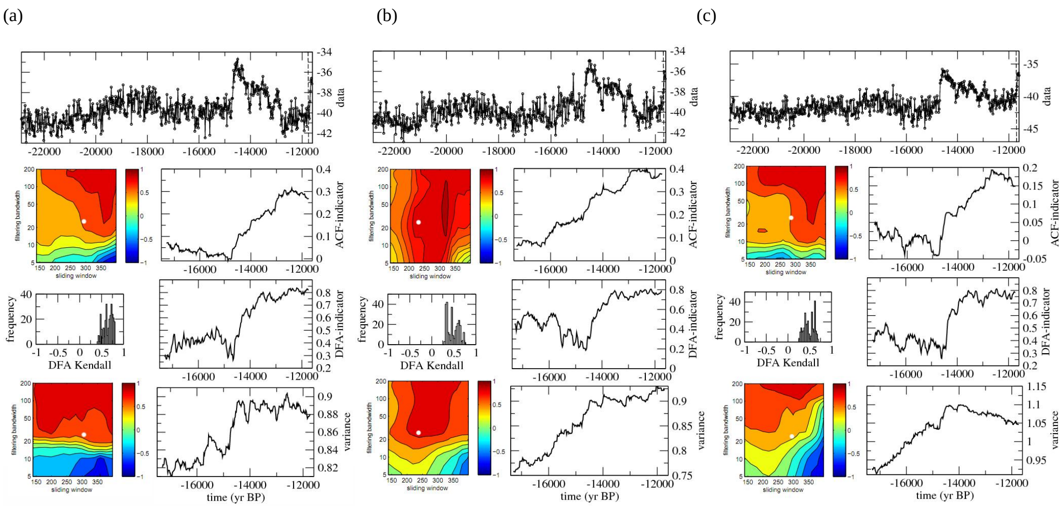

Fig. 2. Indicators of slowing down and changing variance in Greenland ice-core δ^{18}O paleo-temperature proxy records through the last deglciation: **(a)** GRIP **(b)** GISP2 **(c)** NGRIP. In each case: (Top panel) Timeseries of data, where analysis spans 22.88–11.74 ka ($n = 558$), through the Last Glacial Maximum, the Bølling-Allerød, and stopping at the vertical dashed line before the exit from the Younger Dryas. (Left panels) Sensitivity analysis showing values of the Kendall trend statistic for the indicators; (top) ACF, (middle) DFA, (bottom) variance, when varying the sliding window length and (for ACF and variance) the filtering bandwidth used in de-trending – white dots in the contour plots indicate the values of window length and filtering bandwidth used for the example indicators. (Right panels) Example indicators calculated after de-trending, using sliding window length of half the series and (for ACF and variance) a filtering bandwidth of 25; (top) ACF, (middle) DFA, (bottom) variance.

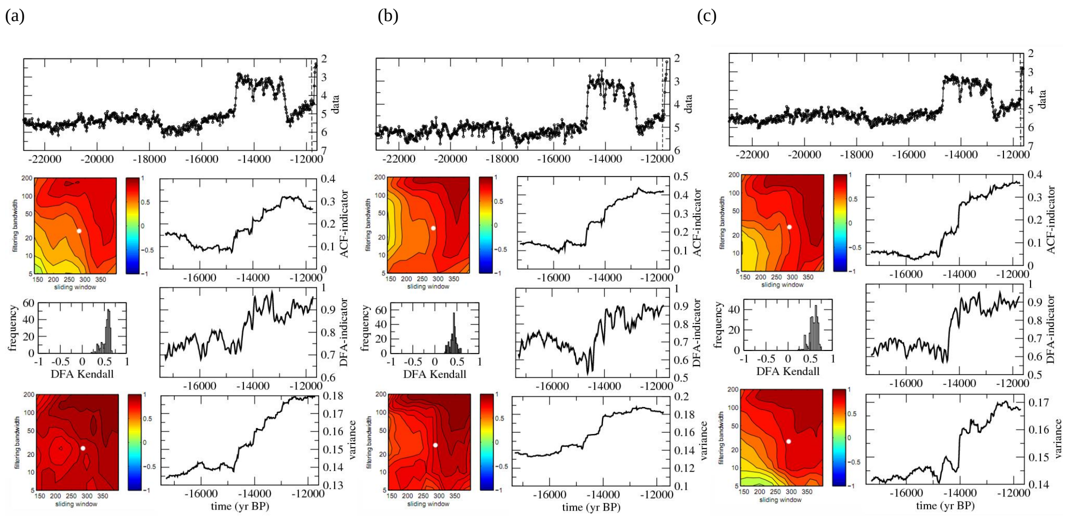

Fig. 3. Indicators of slowing down and changing variance in Greenland ice-core $\log_e[Ca^{++}]$ records through the last deglciation: **(a)** GRIP **(b)** GISP2 **(c)** NGRIP. Panel descriptions are as in Fig. 2.

CPD
8, 321–348, 2012

Title Page
Abstract
Introduction
Conclusions
References
Tables
Figures
◀
▶
Back
Close
Full Screen / Esc
Printer-friendly Version
Interactive Discussion

CPD
8, 321–348, 2012

Climate bifurcation during the last deglaciation

T. M. Lenton et al.

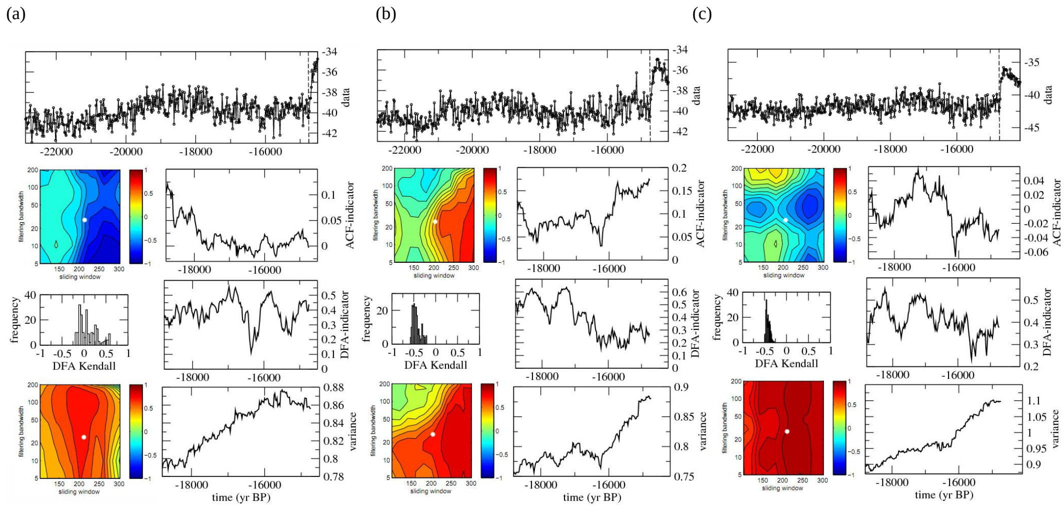

Fig. 4. Indicators of slowing down and changing variance in Greenland ice-core δ^{18}O paleo-temperature proxy records during the interval after DO event 2 to the Bølling warming (DO event 1): **(a)** GRIP **(b)** GISP2 **(c)** NGRIP. Here the analysis spans 22.88–14.74 ka ($n = 408$), stopping at the vertical dashed line before the Bølling warming. Otherwise panel descriptions are as in Fig. 2.

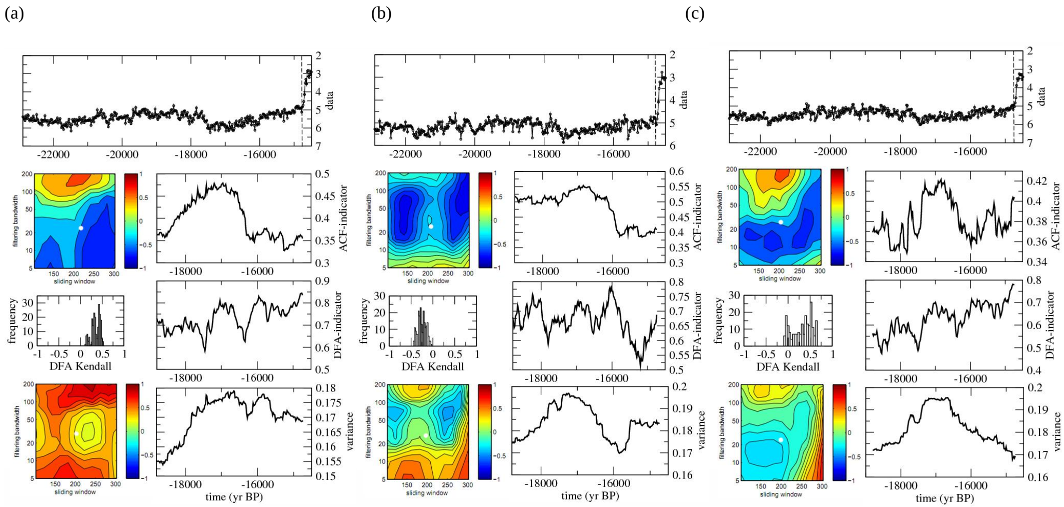

Fig. 5. Indicators of slowing down and changing variance in Greenland ice-core $\log_e[Ca^{++}]$ records during the interval after DO event 2 to the Bølling warming (DO event 1): **(a)** GRIP **(b)** GISP2 **(c)** NGRIP. Here the analysis spans 22.88–14.74 ka ($n = 408$), stopping at the vertical dashed line before the Bølling warming. Otherwise panel descriptions are as in Fig. 2.

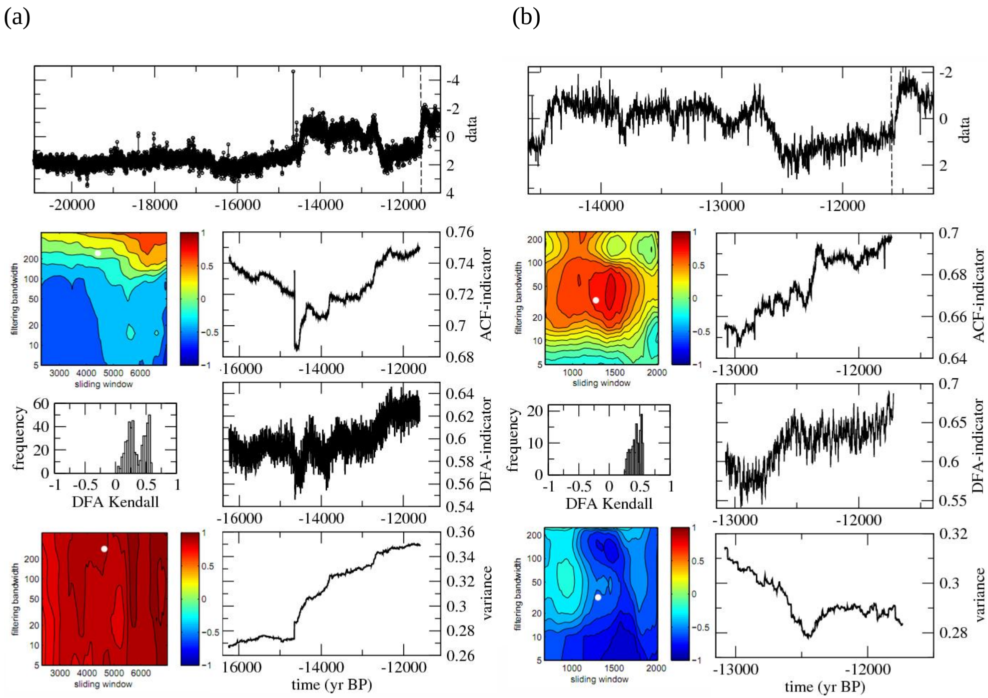

Fig. 6. Caption on next page.

CPD
8, 321–348, 2012

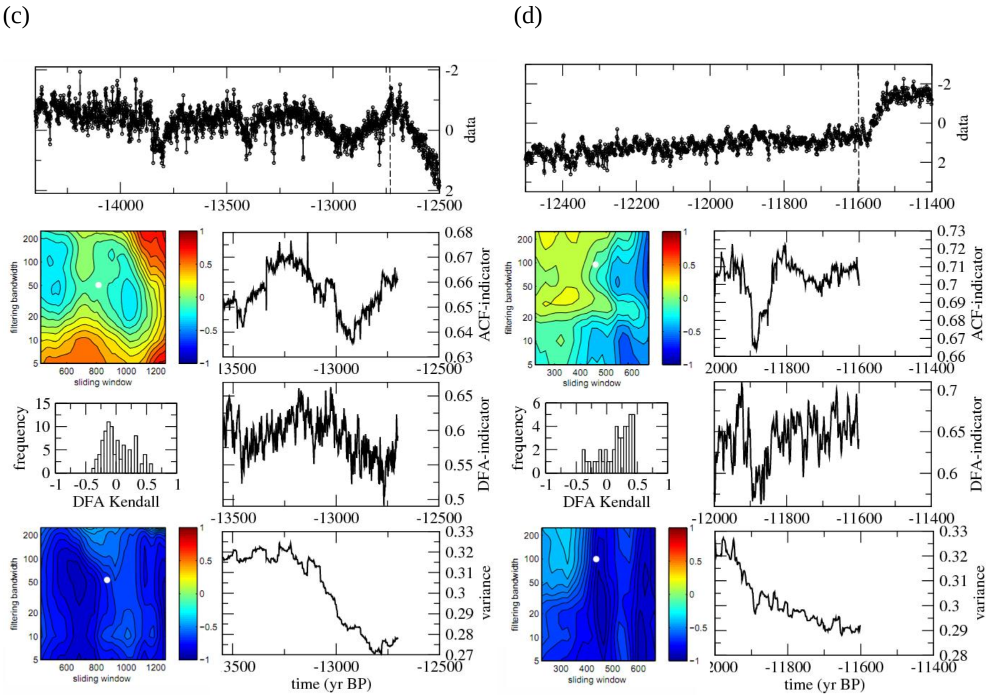

Fig. 6. Indicators of slowing down and changing variance in annual resolution GRIP ice core $\log_e$Ca records over four different time intervals: **(a)** From after DO event 2 to the end of the Younger Dryas (as in Figs. 2 and 3). **(b)** The Bølling-Allerød and Younger Dryas combined. **(c)** Bølling-Allerød. **(d)** Younger Dryas.

CPD
8, 321–348, 2012

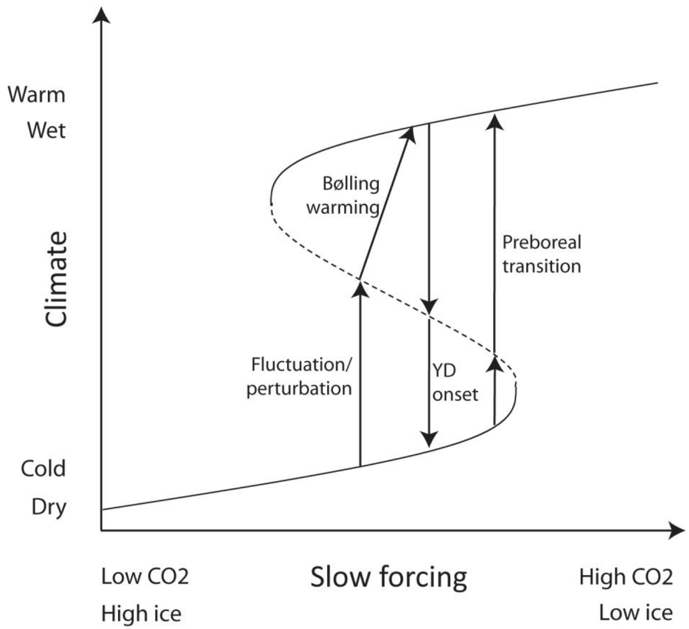

Fig. 7. Simple schematic diagram of our scenario for abrupt climate changes during the deglaciation, described in the Discussion.

CPD
8, 321–348, 2012

Climate bifurcation during the last deglaciation

T. M. Lenton et al.