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Climate of the Past

Tightened constraints on the time-lag between Antarctic temperature and CO_2 during the last deglaciation

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Abstract. Antarctic ice cores provide clear evidence of a close coupling between variations in Antarctic temperature and the atmospheric concentration of CO_2 during the glacial/interglacial cycles of at least the past 800-thousand years. Precise information on the relative timing of the temperature and CO_2 changes can assist in refining our understanding of the physical processes involved in this coupling. Here, we focus on the last deglaciation, 19 000 to 11 000 yr before present, during which CO_2 concentrations increased by ~80 parts per million by volume and Antarctic temperature increased by ~10 °C. Utilising a recently developed proxy for regional Antarctic temperature, derived from five near-coastal ice cores and two ice core CO_2 records with high dating precision, we show that the increase in CO_2 likely lagged the increase in regional Antarctic temperature by less than 400 yr and that even a short lead of CO_2 over temperature cannot be excluded. This result, consistent for both CO_2 records, implies a faster coupling between temperature and CO_2 than previous estimates, which had permitted up to millennial-scale lags.

1 Introduction

The last deglaciation is the largest naturally-forced global climate change in Earth's recent climate history. Its initial impetus, as with the sequence of glacial to interglacial transitions that came before it, is most commonly attributed to orbitally induced variations in insolation at high northern latitudes (Hays et al., 1976). Amplification through climate feedback processes of the relatively weak orbital signal is then required to explain the full magnitude of the glacial to interglacial climate change (Lorius et al., 1990). Changes in the carbon cycle play a central role among these feedbacks (Lorius et al., 1990; Shackleton, 2000; Shakun et al., 2012). Mounting evidence attributes a large component of the deglacial CO_2 increase to release of old CO_2 from the deep Southern Ocean through changes in its biogeochemistry and physical circulation (Anderson et al., 2009; Sigman et al., 2010; Skinner et al., 2010; Schmitt et al., 2012). However, there are open questions about the exact mechanisms involved and their relative contributions (Fischer et al., 2010; Sigman et al., 2010; Toggweiler and Lea, 2010). Determining the timing of the increase in atmospheric CO_2 with respect to the increase in temperature (hereafter "the lag"), is crucial for the models seeking to discriminate between mechanisms (Schmittner and Galbraith, 2008; Ganopolski and Roche, 2009; Lee et al., 2011).

Antarctic ice cores are unique in preserving a record of both temperature variation and atmospheric CO_2 concentration. Water stable isotope ratios ($\delta^{18}O_{ice}$ and δD_{ice}) from the ice are proxies for temperature above the inversion layer at the time of snow formation (Jouzel et al., 1997), while CO_2 is preserved in air bubbles in the ice. The transformation of snow to glacial ice isolates these bubbles at a depth of 50–100 m, leaving them younger than the surrounding ice by an amount Δage. The Δage must therefore be known for the lag between temperature and CO_2 to be accurately determined.

The most commonly cited studies constraining the CO_2 lag during deglaciation have used ice cores from the East

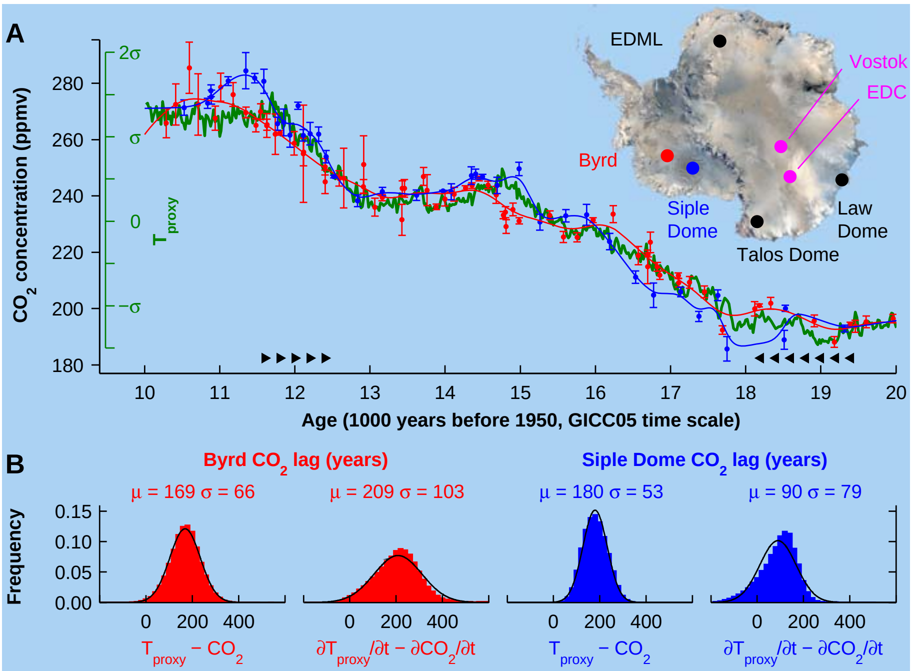

Fig. 1. The phase relationship between regional Antarctic temperature and atmospheric CO_2. **(A)** Antarctic temperature proxy T_{proxy} (green) and CO_2 data (1σ error bars) from Byrd (red) and Siple Dome (blue) on the GICC05 timescale. In the example shown, the CO_2 curves have been smoothed with a Gaussian filter of width 105 yr. **(B)** Lag histograms for the two lag determination techniques (direct correlation and correlation of derivatives) using each of the two CO_2 data sets (red, blue) and the best fit of Gaussian distributions (black curves and μ and σ values). The histogram widths reflect each lag determination's sensitivity to the degree of data smoothing, CO_2 measurement uncertainties, and different choices of lag calculation data interval start and end points (black triangles in part **(A)**). The map shows the location of the Antarctic ice core sites mentioned in the text (source: NASA).

Antarctic Plateau, including Vostok and EPICA Dome C (EDC); locations are marked on the map in Fig. 1. The low accumulation rates at these sites (2–3 $\mathrm{g\,cm^{-2}\,yr^{-1}}$) lead to lengthy intervals between snow deposition and bubble close off, and consequent high values for Δage: 2400–3300 yr for recent times and 4500–5200 yr at the last glacial maximum. A lag (CO_2 behind temperature) of 400–1000 yr is reported for Vostok and Taylor Dome (Fischer et al., 1999) and 800 ± 600 yr for EDC (Monnin et al., 2001). In interpreting these values it is important to be aware of the different methods and uncertainty terms that were applied. Fischer et al. (1999) used spline approximations to obtain the timing of the long-term minima and maxima in δD_{ice} and CO_2 before and after the past three deglacial transitions. Their 400–1000 yr range includes uncertainty in picking the timing of these features but excludes Δage uncertainty (estimated at between 100 and 1000 yr for modern and glacial conditions at that site). The authors state that the Δage uncertainties prevent any firm conclusion about the lag at the onset of deglaciation, but that a real lag is supported at the end of deglaciation/start of interglacials. Monnin et al. (2001) focused specifically on the onset of the last deglaciation and selected the points at which CO_2 and δD_{ice} began to rise by means of the crossing points of linear fits to the respective records. The 800 ± 600 yr range includes uncertainty in picking the timing of features and also nominally in Δage. However, the age model used by Monnin et al. (2001) has since been called into question. With the aid of new dating constraints in the ice and gas phase, Loulergue et al. (2007) argued that the model substantially overestimates Δage for the last glacial period and deglaciation, implying that the lag given by Monnin et al. (2001) is also too large.

A different approach to determining temperature and CO_2 phasing involves shifting the CO_2 and stable isotope records relative to one another in time until the optimum correlation between the series is reached. This method was applied to the 390–650 ka interval of the EDC core (Siegenthaler et al., 2005), and a similar method was applied to the full 420 ka of the Vostok core (Mudelsee, 2001), yielding optimum correlations at lags of 1.9 ka and 1.3 ± 1.0 ka, respectively. Both studies list Δage uncertainty as the major source of error in their estimates; a specific error term is not provided by Siegenthaler et al. (2005).

The contribution to the lag uncertainty from Δage can be minimised by considering ice core records from higher accumulation sites. Of the currently available CO_2 records for the last deglaciation, those with the lowest Δages are the Siple Dome (Ahn et al., 2004) and Byrd (Neftel et al., 1988; Staffelbach et al., 1991) ice cores. The Δages at these sites are 200–300 yr for recent times and 500–800 yr at the last glacial maximum, an order of magnitude smaller than records from the East Antarctic Plateau. The lag between temperature and CO_2 was previously investigated in the Siple Dome core by Ahn et al. (2004). The method used in this case was to determine the lag which gave the optimum correlation between the derivatives of the CO_2 and δD_{ice} records over the entire deglaciation (the potential advantages and disadvantages of using derivatives for such analyses are considered further below). Optimum correlation was observed at a CO_2 lag of ∼ 150–400 yr. After accounting for additional uncertainties, including in Δage, the authors main conclusion was that a lag of CO_2 versus Siple Dome temperature was likely and a lead of CO_2 versus Siple Dome temperature was unlikely.

A caveat associated with the Siple Dome result and other lag assessments based on individual ice cores relates to the effects of local and/or non-climatic influences on stable isotope records (e.g. very complex ice flow, changes in surface elevation, and changes in the seasonal distribution of snow fall, Jones et al., 2009). Signals caused by such variability would not be expected to correlate with CO_2 evolution and therefore may have a confounding influence on lag assessments.

Our approach is to compare the Siple Dome and Byrd CO_2 records to an index of regional Antarctic temperature, T_{proxy}, derived from a composite of high-resolution water stable isotope records from the Law Dome, Siple Dome, Byrd, EPICA Dronning Maud Land (EDML) and Talos Dome ice cores; core locations are marked on the map in Fig. 1 (Pedro et al., 2011). Previous studies have demonstrated that local and/or non-climatic signals in individual ice cores are reduced in multi-core composites, leading to more robust representations of regional climate trends (Fisher et al., 1996; White et al., 1997). Since T_{proxy} contains ice cores representing the Indian, Atlantic, and Pacific coastal Antarctic regions, it is also expected to provide a better representation of high-latitude Southern Ocean processes than records from single sites. This approach requires that T_{proxy} and the two CO_2 records share a common timescale. This is achieved by synchronising all records to the Greenland Ice Core Chronology 2005 (GICC05) using the rapid and effectively globally synchronous variations in CH_4 concentrations found in both Antarctic and Greenland ice cores.

We do not use the EDC CO_2 record in this analysis since there are unresolved uncertainties in the currently available gas age timescales for the core (see, e.g. Loulergue et al., 2007; Lemieux-Dudon et al., 2010; Parrenin et al., 2012).

2 Materials and methods

2.1 Record synchronisation

The T_{proxy} record is already available on GICC05, standardised to unit variance and zero mean (Pedro et al., 2011). We synchronise the timescales of the Siple Dome and Byrd CO_2 records to GICC05 using previously published gas–age depth ties from CH_4-based synchronisations of the Byrd (Blunier and Brook, 2001) and Siple Dome (Brook et al., 2005) records with Greenland records. The Byrd CO_2 data on the GRIP ss09 timescale (Johnsen et al., 1997) are transferred via GRIP depth (Blunier and Brook, 2001) to GICC05 ages by linear interpolation using stratigraphical markers (following Rasmussen et al., 2008, the markers are listed in their Table 2 and at http://www.iceandclimate.nbi.ku.dk/data/). Similarly, the Siple Dome CO_2 data on a GISP2 timescale (Ahn et al., 2004) are transferred via GISP2 depth (Meese et al., 1997) to GICC05 ages, again, by linear interpolation using stratigraphical markers (Rasmussen et al., 2008). The dating error introduced in the transfer from the GRIP ss09 and GISP2 time scales to the GICC05 timescale is negligible compared to Δage uncertainties and the uncertainties in the original methane synchronisations. All GICC05 dates mentioned in the text hereafter use the convention "ka b1950" referring to thousands of years before 1950 AD.

2.2 Lag calculation

Visual inspection of the relative timing of the deglacial CO_2 and T_{proxy} increase in Fig. 1a suggests little or no lag of CO_2 after temperature. We determine the lag quantitatively by maximising the time-lagged correlation between the deglacial temperature and CO_2 curves throughout the entire deglaciation. We consider this approach to be more robust than comparing dates of events (e.g. exclusively the start of the deglaciation) because its estimation is based on a much greater number of data points (see, e.g. Mudelsee, 2001). As a first approximation, setting aside Δage and CO_2 measurement uncertainties and prior to any smoothing/filtering of the data, the maximum of the time-lagged correlation suggests that CO_2 lagged temperature by one to two centuries for both CO_2 records. To refine this approximation and explore its sensitivity to various parameters, we apply a Monte Carlo-style sensitivity analysis.

Two methods of correlation are used for the sensitivity analysis: first, as above, by direct correlation between T_{proxy} and CO_2 ("direct method", similar to Siegenthaler et al., 2005); and second, by correlation between the corresponding derivative curves, $\partial T_{proxy}/\partial t$ and $\partial CO_2/\partial t$ ("derivative method", similar to Ahn et al., 2004). The derivative method has smaller sensitivity to misidentification of the pre- and post-transition levels at the expense of increased sensitivity to measurement noise, especially in the sections of sparse data. Since it is not clear a priori which approach is superior, we apply both methods in parallel. Prior to the analyses, the CO_2 data are first linearly interpolated to 20-yr resolution over the 9–21 ka b1950 interval to match the 20-yr resolution of T_{proxy}. Derivatives, $\partial T_{proxy}/\partial t$ and $\partial CO_2/\partial t$, are calculated on the same grid. For both methods the data are then smoothed and the time-lagged correlation determined for lags in the −400 to +1000 yr range. The robustness of the lag calculation to CO_2 data uncertainty is evaluated by repeating the lag determination 100 times using random realisations of the CO_2 measurements, each sampled using the published uncertainties (represented by the error bars in Fig. 1a). As the lag shows some sensitivity to the degree of smoothing applied and the overall time interval chosen for assessment, we repeated the analysis using 19 different degrees of smoothing and 35 different choices of time interval. The different degrees of smoothing were applied by convolution with a Gaussian filter with a standard deviation ranging from 105–375 yr in steps of 15 yr; the minimum width of the smoothing corresponds to the approximate sampling rate of CO_2 data in the sparsely sampled parts of the data set, and the maximum was chosen to preserve a clear Antarctic Cold Reversal signature. Similar results were obtained by replacing the Gaussian filter with a simple running mean filter of equivalent half-width. The choices of the time interval start and end points for the lag calculation were varied within reasonable limits that encompass the overall deglaciation; specifically, the start and end ages were varied in 200 yr steps with the start spanning 19.4–18.2 ka b1950 (7 options) and end spanning 12.4–11.6 ka b1950 (5 options), as illustrated by the black triangles of Fig. 1a. This covers the start and end of the deglaciation as defined from the perspective of the onset and end of the warming trend in T_{proxy} (Pedro et al., 2011), but excludes the early Holocene during which the correlation between Antarctic temperature and CO_2 appears to deteriorate (both CO_2 curves have high early Holocene values that have no counterpart in T_{proxy}). Given the close agreement between CO_2 and temperature throughout the deglacial warming and Antarctic Cold Reversal, the weaker relationship in the early Holocene may imply that different processes (e.g. processes operating outside the Southern Ocean) became more important in controlling atmospheric CO_2. A recently produced carbon-stable isotope record from the EDC core depicts a positive excursion in $\delta^{13}C_{atm}$ beginning around 12.2 ka, which the authors attribute to regrowth of the terrestrial biosphere (Schmitt et al., 2012).

A caveat associated with our lag determination method is that it implicitly assumes that the maximum of the lagged correlation does in reality provide a valid estimate of the lag. Simple linear models of this form are widely used in previous studies of temperature and CO_2 phasing (e.g. Ahn et al., 2004; Siegenthaler et al., 2005; Shakun et al., 2012); however, the result may somewhat change systematically if other models are used.

3 Results and discussion

Close correspondence between T_{proxy} and both CO_2 records is observed (Fig. 1a), supporting the hypothesis that marine processes at high southern latitudes are linked to the deglacial CO_2 increase.

Quantitative results on the relative timing of the deglacial CO_2 and T_{proxy} increase are summarised by the four histograms in Fig. 1b. Each distribution comprises 66 500 optimal lag values generated from the sensitivity analysis. The widths of the distributions thus reflect the sensitivity of the lag values to the lag determination method (direct or derivative), the CO_2 measurement uncertainties, differences between the two CO_2 data sets, smoothing filter widths and the start and end points used in the analysis. The distributions show larger spread for the derivative method, but overall consistency between the two CO_2 data sets and the two methods. Since the individual distributions are not completely independent, we take a conservative approach and pool all of the results. Applying a Gaussian best-fit to the pooled results suggests a mean ($\pm 1\sigma$) for the lag of 162 ± 89 yr.

We must also consider the relative dating uncertainty between T_{proxy} and the two CO_2 records. This uncertainty comprises a component associated with synchronising CH_4 records between multiple cores (synchronisation error) and a component associated with the Δage uncertainties of the individual cores (Δage error). The dating uncertainty in T_{proxy} (relative to GICC05) ranges from 199–384 yr over the time interval considered, with a mean of 269 yr (Pedro et al., 2011). The relative dating uncertainty between T_{proxy} and the two CO_2 records is smaller than the uncertainty in T_{proxy}, because the synchronisation error applies to both T_{proxy} and the CO_2 data and thus partially cancels (as seen in the case where data from only one core is used and the synchronisation error cancels completely). Furthermore, the similarity of the results from the lag calculation using Siple Dome and Byrd CO_2 data supports that the records are well-synchronised on centennial scales or better. On this basis, we estimate an overall relative dating uncertainty (nominal 1σ) of 200 yr. This term is independent of the 89-yr uncertainty captured by the sensitivity analysis. Taking the central estimate of the lag from the pooled sensitivity analysis and combining the two uncertainty terms in quadrature leads to a likely range for the CO_2 lag of −56 to 381 yr. From this we arrive at our main conclusion that the deglacial CO_2 increase likely lagged regional Antarctic temperature by less than 400 yr and that even a short lead of CO_2 over temperature cannot be excluded.

We conducted a series of jack-knife tests to assess the sensitivity of our lag estimate to the inclusion/exclusion of any of the individual stable isotope records used in T_{proxy}. By excluding in turn each one of the 5 records and using each of the two CO_2 data sets and correlation methods, the 20 histograms shown in Fig. 2 are produced. The mean lag value of these 1.33 million realisations is just one year larger than

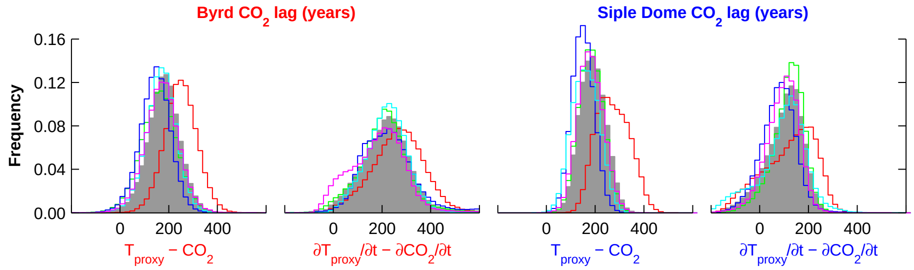

Fig. 2. Lag histograms for the two methods of determining the lag of atmospheric CO_2 after regional Antarctic temperature changes (direct correlation and correlation of derivatives), using each of the two CO_2 data sets (Byrd and Siple Dome). The gray background histograms are based on the complete T_{proxy} composite, the same as in Fig. 1b. The superimposed curves show the corresponding lag histograms when excluding in turn each of the 5 records from the T_{proxy} composite (jack-knifing): excluding Siple (red), excluding Law Dome (green), excluding Byrd (blue), excluding EDML (cyan), and excluding Talos Dome (magenta).

the value obtained using all the records, but comprises some systematic differences. Excluding Law Dome, Byrd, Talos Dome or EDML records reduces the mean lag values by up to a couple of decades with a mean value of 12 yr, while there is somewhat larger sensitivity to exclusion of the Siple Dome record, which shifts the lag higher by on average 55 yr. A possible explanation for the somewhat larger sensitivity for Siple Dome may be an underestimate of Δage at that site, or the influence of local and/or non-climate signals. The abrupt increase in δD_{ice} at around 15 ka in the Siple Dome record may be an example of such a local signal (as discussed previously, Severinghaus et al., 2003; Taylor et al., 2004). Note that an error in Δage would not affect the dating of the Siple Dome CO_2 record since it is directly synchronised using methane records to Greenland.

Direct intercomparison of our result with prior estimates is complicated by the different time intervals considered and different ways to estimate uncertainties applied in the respective studies. Our result is in broad agreement with Ahn et al. (2004), which suggested short lag at Siple Dome, also determined over the entire deglaciation. As mentioned earlier, Fischer et al. (1999) were confident in a real lag between temperature and CO_2 for Vostok only at the start of the interglacials. There is a suggestion in both the Byrd and Siple Dome records of a larger lag at the onset of the Holocene (Fig. 1a), but this period is not included in the time window of our lag assessment as we assume it is likely influenced by processes acting on the carbon cycle outside the Southern Ocean. Monnin et al. (2001) defined the 800 ± 600 yr lag at EDC at the onset of deglaciation. The Byrd and Siple Dome CO_2 data is not sufficiently dense to constrain the lag at the same point, but our overall result is not inconsistent with the Monnin et al. (2001) estimate within its uncertainty range. Confidence in our result is provided by the sensitivity analysis and the consistent results for two different CO_2 records and two different correlation methods. The conclusion of Loulergue et al. (2007) that Δage (and thus the CO_2 temperature lag) was significantly overestimated during the deglacial onset by Monnin et al. (2001) also lends support to our result.

A brief comparison with the recent work by Shakun et al. (2012) is also warranted. Their study evaluates the phasing between the EDC CO_2 record and multi-proxy hemispheric and global (rather than exclusively Antarctic) temperature reconstructions. They report a CO_2 lag behind their Southern Hemisphere temperature reconstruction (620 ± 660 yr), a *lead* of CO_2 over their Northern Hemisphere reconstruction (720 ± 330 yr), and a short *lead* of CO_2 over their full global reconstruction (460 ± 340 yr). The southern lag and northern lead is attributed to an anti-phased hemispheric temperature response to ocean circulation changes (as also discussed further below) superimposed on globally in-phase warming driven by the CO_2 increase. This emphasises the role of CO_2 as both feedback and forcing in the deglacial warming. Within the quoted uncertainty bounds, the 620 ± 660 yr lag for the Southern Hemisphere is not inconsistent with our Antarctica-based result; also, considering the aforementioned Δage issues for EDC, their Southern Hemisphere lag is likely somewhat overestimated (and the northern and global lead are likely underestimated). The larger uncertainty range around the Shakun et al. (2012) result must be expected given the challenges of synchronising records from multiple proxy types. In our view, the remarkable similarity of the Antarctic temperature and CO_2 curves and the independent evidence that the high latitude Southern Ocean was a centre of action in the deglacial CO_2 release make the lag determination from an Antarctic perspective critical for constraining the mechanisms involved in the CO_2 increase.

Additional observational constraints on the processes involved in the CO_2 increase are obtained by considering in more detail the millennial and sub-millennial trends in CO_2 throughout the deglaciation. The CO_2 increase, as noted previously (Fischer et al., 1999; Monnin et al., 2001), occurs in two main steps. These steps coincide with the two periods of significant warming in T_{proxy} (pink bands, Fig. 3) and are separated by a step down in CO_2 concentration as T_{proxy}

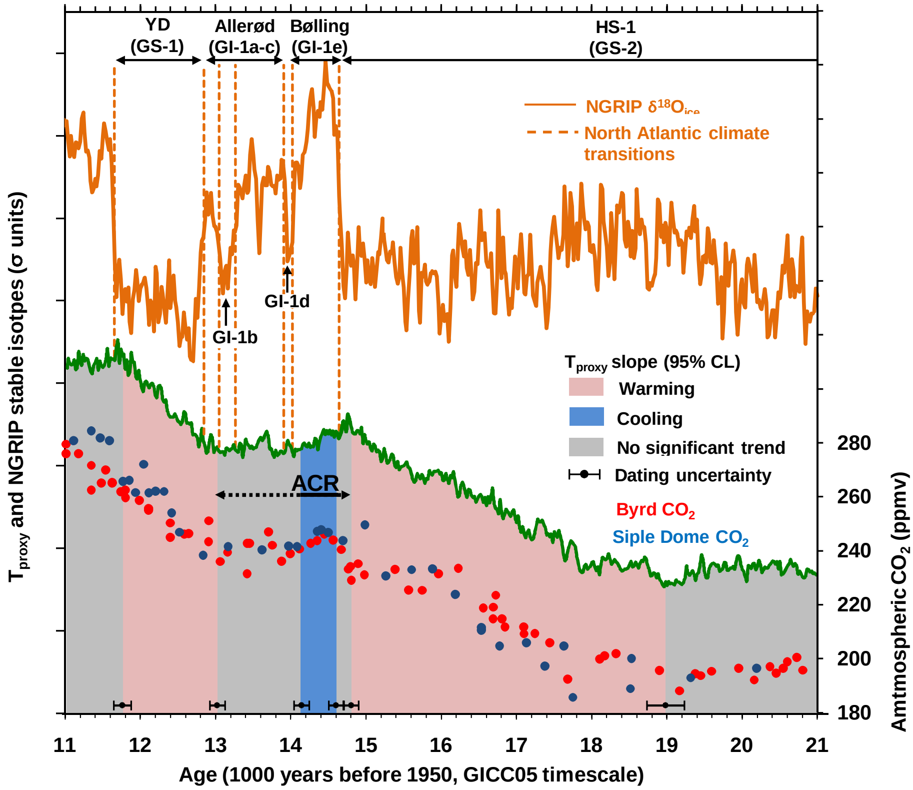

Fig. 3. Atmospheric CO_2 and the bipolar seesaw on the GICC05 timescale. Significant warming and cooling trends in T_{proxy} are represented by coloured vertical bands, adopted from a previous study (Pedro et al., 2011). Climate in the North Atlantic region is represented by the NorthGRIP ice core $\delta^{18}O$ record (Lowe et al., 2008). Changes in the slope of T_{proxy} are synchronous with climate transitions in the North Atlantic (vertical orange lines) within relative dating uncertainties (horizontal error bars). The deglacial increase in CO_2 occurs in two steps, corresponding to the significant warming trends in T_{proxy} (pink bands). A pause in the CO_2 rise is aligned with a break in the warming trend (central grey band) during the Antarctic Cold Reversal (ACR). Within the core of the Antarctic Cold Reversal, significant cooling in T_{proxy} (dark blue band) coincides with an apparent decrease in CO_2. Fast-acting inter-hemispheric coupling mechanisms linking Antarctica, Greenland and the Southern Ocean are required to satisfy these timing constraints.

exhibits significant cooling within the core of the Antarctic Cold Reversal (dark blue band, Fig. 3).

Evidence from Southern Ocean marine sediment cores directly links each of the two warming steps with release of CO_2 accumulated in the deep Southern Ocean during the last glacial period: cores from south of the Antarctic Polar Front show pulses in upwelling (represented by opal fluxes) synchronous with each warming step (Anderson et al., 2009) and coincident with negative excursions in atmospheric $\delta^{13}C$ (Schmitt et al., 2012), while cores from the Atlantic sector of the Southern Ocean identify a source of old (^{14}C-depleted) carbon-rich deep water that dissipated over two corresponding intervals (Skinner et al., 2010). An explanation consistent with the near-synchronous temperature and CO_2 response identified here proposes that the increases in upwelling were responsible for the simultaneous delivery of both sequestered heat and CO_2 to the atmosphere around Antarctica (Anderson et al., 2009; Skinner et al., 2010). This does not imply that the Southern Ocean is the only important CO_2 source during deglaciation; in a coupled system it is plausible that other processes operating outside the region may also correlate with Antarctic temperature.

We now consider the time relationship between T_{proxy} and the major millennial and sub-millennial climate events recorded by the North GRIP $\delta^{18}O_{ice}$ record (Fig. 3); i.e. the timing of the well-known bipolar seesaw (Broecker, 1998; Stocker and Johnsen, 2003). Since T_{proxy} and the CO_2 records are synchronised to GICC05, the records from both hemispheres can be directly compared. Fig. 3 reveals that there is also little or no time lag separating the major millennial and sub millennial climate transitions in Greenland (Lowe et al., 2008) from the onsets and ends of the warming and cooling trends in Antarctica (Pedro et al., 2011). A picture thus emerges from the ice cores of rapid (centennial-scale or faster) coupling between Antarctic temperatures, Greenland temperatures and atmospheric CO_2.

We give attention here to two potentially complimentary mechanisms which appear consistent with these observations. One mechanism supports a greater role for atmospheric pathways (Anderson et al., 2009; Toggweiler and Lea, 2010;

Lee et al., 2011) while the other supports a greater role for the ocean (Schmittner and Galbraith, 2008; Skinner et al., 2010). Neither mechanism directly specifies the proximal trigger of deglacial warming; indeed, this important point is not yet resolved. A possible forcing model suggests that orbitally-driven increases in boreal summer insolation during Greenland Stadial 2 initiate local warming and retreat of Northern Hemisphere ice sheets. Attendant freshwater release into the North Atlantic then weakens and displaces southward the Atlantic Meridional Overturning Circulation (AMOC) (Ganopolski and Rahmstorf, 2001; McManus et al., 2004). A feedback process has recently been proposed wherein the weakened overturning leads to a warming of North Atlantic subsurface waters (Marcott et al., 2011; Gutjahr and Lippold, 2011), destabilising ice shelves and in turn driving further ice and freshwater release (Álvarez-Solas et al., 2011). The atmospheric pathway proposes that surface cooling and sea ice expansion in the North Atlantic region initiates an atmospheric reorganisation in which both the Inter-Tropical Convergence Zone and the Southern Hemisphere mid-latitude westerlies are displaced south and/or intensified (Anderson et al., 2009; Toggweiler and Lea, 2010; Lee et al., 2011). The changed westerly regime over the Southern mid- and high-latitude oceans then displaces cold fresh surface waters, melts and dissipates sea ice, and draws up warmer deeper waters, which release heat and CO_2. A recent modeling study lends support to this hypothesis and also incorporates an important role for ocean biogeochemical feedbacks (Lee et al., 2011). In line with the constraints identified here, the model supports essentially simultaneous North Atlantic cooling and atmospheric CO_2 release from the high-latitude Southern Ocean, corresponding to an atmospheric CO_2 rise of 20–60 ppm where the range is dependent on the biological response. However, Lee et al. (2011) do not simulate the observed response of Antarctic temperatures, possibly due to fixed Southern Hemisphere sea ice.

The ocean pathway invokes the bipolar seesaw: weakening of the AMOC reduces northward ocean heat transport, thereby allowing heat to accumulate in the south (Broecker, 1998; Stocker and Johnsen, 2003). This melts back sea ice, removing a physical barrier to the release of CO_2 and exposing the Southern Ocean surface waters to the action of the westerlies. As above, the strengthened westerlies promote upwelling and ventilation of CO_2 from the deep waters (Skinner et al., 2010). The bipolar seesaw, as originally conceived (Broecker, 1998), operates on the millennial timescales of ocean heat transport and hence fails to meet the timing criteria identified here. However, an ocean wave-mediated seesaw (Stocker and Johnsen, 2003) could be capable of sufficiently rapid signal transmission. In one model study, the bipolar-seesaw changes in the circulation modes couple with changes in ocean biogeochemistry and carbon storage in the terrestrial biosphere to produce little or no lag between Greenland and Antarctic temperatures and CO_2 changes (Schmittner and Galbraith, 2008). Unlike Lee et al. (2011), this model does simulate the observed seesaw relationship between North Atlantic and Greenland temperatures. Notably, the initial (within the first 250 yr) rapid response in CO_2 in Schmittner and Galbraith (2008) is attributed to CO_2 release from the terrestrial biosphere (triggered by cooling associated with the AMOC reduction) and not to physical or biogeochemical components of the Southern Ocean carbon cycle. However, the response of terrestrial vegetation models and the interplay between ocean and vegetation responses is known to be highly model-dependent (Bouttes et al., 2012). Recent studies also emphasise the sensitivity of the timescales of ocean ventilation to model resolution; at least under modern boundary conditions, meso-scale eddy-resolving ocean models suggest much faster ventilation times than the coarser resolution models that are currently used for palaeo-simulations (Maltrud et al., 2010).

Current palaeo-modeling efforts appear to be accurately simulating key components of the deglacial sequence of events. However, there is not yet a simulation which fully captures all of the observed trends and phase relationships between Antarctic and North Atlantic temperatures and atmospheric CO_2. The results presented here, with their improved timing constraints over previous studies, provide a more stringent target for future modeling efforts.

4 Conclusions

The ice core observations point to a tightly-coupled system operating with little or no time delay between the onsets/terminations of North-Atlantic climate stages and near-simultaneous trend changes in both Antarctic temperature and atmospheric CO_2. As it stands, the observed timing of events lends support to the current concept of an atmospheric teleconnection between the northern and southern high-latitudes which forces wind-driven CO_2 release from the Southern Ocean. However, sorting out the relative roles of atmospheric and oceanic coupling mechanisms and the interactions between them remains a major challenge. More densely sampled and higher precision CO_2 measurements from high-accumulation ice core sites may assist with constraining the time evolution of the Antarctic temperature and CO_2 phasing during different stages of the deglaciation. Further progress in understanding these mechanisms requires more work at the interface between palaeoclimate observation and Earth system modeling; in particular, modeling efforts to more tightly constrain the potential timescales of bipolar-seesaw induced changes in CO_2 ventilation are needed.

The T_{proxy} series and CO_2 records on the GICC05 timescale are archived at the Australian Antarctic Data Center (http://data.aad.gov.au/) and at the World Data Center for Paleoclimatology (http://www.ncdc.noaa.gov/paleo/).

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