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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$^\circ$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$_{\text{ice}}$ and $\delta$D$_{\text{ice}}$) 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...
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 g cm\(^{-2}\) yr\(^{-1}\)) lead to lengthy intervals between snow deposition and bubble close off, and consequent high values for \(\Delta\)age: 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 \(\Delta_{\text{D}_{\text{ice}}}\) 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 \(\Delta\)age uncertainty (estimated at between 100 and 1000 yr for modern and glacial conditions at that site). The authors state that the \(\Delta\)age 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 \(\Delta_{\text{D}_{\text{ice}}}\) 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 \(\Delta\)age. 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 \(\Delta\)age 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 (Mudselee, 2001), yielding optimum correlations at lags of 1.9 ka and 1.3 ± 1.0 ka, respectively. Both studies list \(\Delta\)age 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 \(\Delta\)age 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 \(\Delta\)ages are the Siple Dome (Ahn et al., 2004) and Byrd (Neftel et al.,

![Fig. 1. The phase relationship between regional Antarctic temperature and atmospheric CO\(_2\). (A) Antarctic temperature proxy \(T_{\text{proxy}}\) (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 \(\mu\) and \(\sigma\) 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).](https://www.clim-past.net/8/1213/2012/)

1998; 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 CO₂ 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 CO₂ 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 CO₂ lag of ~150–400 yr. After accounting for additional uncertainties, including in Δage, the authors main conclusion was that a lag of CO₂ versus Siple Dome temperature was likely and a lead of CO₂ 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 CO₂ evolution and therefore may have a confounding influence on lag assessments.

Our approach is to compare the Siple Dome and Byrd CO₂ 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 both CO₂ and δD values are strongly controlled by a variety of factors including precipitation, temperature, snow fall, sea level change, atmospheric circulation, and possibly volcanic events (see e.g. Jones et al., 2009). 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 CO₂ 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 isotope variations in CH₄ concentrations found in both Antarctic and Greenland ice cores.

We do not use the EDC CO₂ 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 CO₂ records to GICC05 using previously published gas–age depth ties from CH₄-based synchronisations of the Byrd (Blunier and Brook, 2001) and Siple Dome (Brook et al., 2005) records with Greenland records. The Byrd CO₂ 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 CO₂ 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 deglaciation CO₂ and T_proxy increase in Fig. 1a suggests little or no lag of CO₂ after temperature. We determine the lag quantitatively by maximising the time-lagged correlation between the deglacial temperature and CO₂ 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 CO₂ measurement uncertainties and prior to any smoothing/filtering of the data, the maximum of the time-lagged correlation suggests that CO₂ lagged temperature by one to two centuries for both CO₂ 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 CO₂ (“direct method”, similar to Siegenthaler et al., 2005); and second, by correlation between the corresponding derivative curves, ∂T_proxy/∂t and ∂CO₂/∂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 
CO2 data are first linearly interpolated to 20-yr resolution 
over the 9–21 ka b1950 interval to match the 20-yr resolution of 
Tproxy. Derivatives, ∂Tproxy/∂t and ∂CO2/∂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 CO2 data uncertainty is evaluated by repeating 
the lag determination 100 times using random realisations of 
the CO2 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 ap-
plied 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 CO2 data in 
the sparsely sampled parts of the data set, and the maximum 
was chosen to preserve a clear Antarctic Cold Reversal signa-
ture. 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 span-
ing 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 Tproxy (Pedro et al., 2011), but excludes the 
early Holocene during which the correlation between Antarct-
ic temperature and CO2 appears to deteriorate (both CO2 
curves have high early Holocene values that have no counter-
part in Tproxy). Given the close agreement between CO2 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 con-
trolling atmospheric CO2. A recently produced carbon-stable 
isotope record from the EDC core depicts a positive excursion in δ13C 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 previ-
ous studies of temperature and CO2 phasing (e.g. Ahn et al., 
2004; Siegenthaler et al., 2005; Shakun et al., 2012); how-
ever, the result may somewhat change systematically if other 
models are used.

3 Results and discussion

Close correspondence between Tproxy and both CO2 records is 
observed (Fig. 1a), supporting the hypothesis that marine 
processes at high southern latitudes are linked to the deglacial 
CO2 increase.

Quantitative results on the relative timing of the deglacial 
CO2 and Tproxy increase are summarised by the four his-
tograms in Fig. 1b. Each distribution comprises 66 500 op-
timal 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 deriva-
tive), the CO2 measurement uncertainties, differences be-
tween the two CO2 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 con-
sistency between the two CO2 data sets and the two meth-
ods. Since the individual distributions are not completely in-
dependent, we take a conservative approach and pool all of 
the results. Applying a Gaussian best-fit to the pooled results 
suggests a mean (± 1σ) for the lag of 162 ± 89 yr.

We must also consider the relative dating uncertainty be-
tween Tproxy and the two CO2 records. This uncertainty 
comprises a component associated with synchronising CH4 
records between multiple cores (synchronisation error) and a 
component associated with the Δage uncertainties of the in-
dividual cores (Δage error). The dating uncertainty in Tproxy 
(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 Tproxy and the 
two CO2 records is smaller than the uncertainty in Tproxy, be-
cause the synchronisation error applies to both Tproxy and the 
CO2 data and thus partially cancels (as seen in the case where 
data from only one core is used and the synchronisation er-
or cancels completely). Furthermore, the similarity of the 
results from the lag calculation using Siple Dome and Byrd 
CO2 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 CO2 lag of −56 to 381 yr. From this we arrive at 
our main conclusion that the deglacial CO2 increase likely 
lagged regional Antarctic temperature by less than 400 yr and 
that even a short lead of CO2 over temperature cannot be 
excluded.

We conducted a series of jack-knife tests to assess the sen-
sitivity of our lag estimate to the inclusion/exclusion of any 
of the individual stable isotope records used in Tproxy. By ex-
cluding in turn each one of the 5 records and using each of 
the two CO2 data sets and correlation methods, the 20 his-
tograms shown in Fig. 2 are produced. The mean lag value 
of these 1.33 million realisations is just one year larger than
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 $\Delta$age at that site, or the influence of local and/or non-climate signals. The abrupt increase in $\delta$D$_{\text{ice}}$ 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 $\Delta$age 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 $\Delta$age (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 $\Delta$age 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 $T_{\text{proxy}}$ (pink bands, Fig. 3) and are separated by a step down in CO$_2$ concentration as $T_{\text{proxy}}$.
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\textsubscript{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 (\(1^{4}\)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\textsubscript{2} response identified here proposes that the increases in upwelling were responsible for the simultaneous delivery of both sequestered heat and CO\textsubscript{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\textsubscript{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 \(T_{\text{proxy}}\) and the major millennial and sub-millennial climate events recorded by the North GRIP \(\delta^{18}\)O\textsubscript{ice} record (Fig. 3); i.e. the timing of the well-known bipolar seesaw (Broecker, 1998; Stocker and Johnsen, 2003). Since \(T_{\text{proxy}}\) and the CO\textsubscript{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\textsubscript{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 CO2. 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 CO2 release from the high-latitude Southern Ocean, corresponding to an atmospheric CO2 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 CO2 and exposing the Southern Ocean surface waters to the action of the westerlies. As above, the strengthened westerlies promote upwelling and ventilation of CO2 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 CO2 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 CO2 in Schmittner and Galbraith (2008) is attributed to CO2 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 CO2. 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 CO2. 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 CO2 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 CO2 measurements from high-accumulation ice core sites may assist with constraining the time evolution of the Antarctic temperature and CO2 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 CO2 ventilation are needed.

The Tproxy series and CO2 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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