Role of deep sea temperature in the carbon cycle during the last glacial

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To characterize the response of Earth’s climate system to increases in atmospheric CO$_2$, climate modelers define climate sensitivity as the change in global mean temperature in response to prescribed forcing. Here we turn this approach around and use estimates of ocean temperature change to investigate the mechanisms driving CO$_2$ variations over the last glacial. New records provide evidence of a link between deep ocean temperature and atmospheric CO$_2$ over the last glacial cycle. Two mechanisms simultaneously couple pCO$_2$ and deep ocean temperature: the temperature-dependent solubility of CO$_2$ in seawater and the atmospheric CO$_2$-dependent radiative forcing of temperature. Each of these forcing mechanisms leaves a unique slope of covariation between CO$_2$ and deep ocean temperature, which we estimate using numerical models of climate and the carbon cycle. The pCO$_2$/T slopes derived from paleoclimate data differ between the deglaciation and shorter 5-kyr duration events in marine isotope stage 3 (MIS 3), revealing different mechanisms driving atmospheric CO$_2$ variability.

The amplitude of changes over the deglaciation coincides with estimates for CO$_2$ forcing of temperature; however, CO$_2$ changes during MIS 3 can be explained solely by temperature-dependent solubility driving variations in atmospheric pCO$_2$. The deep water temperature changes during MIS 3 may reflect changes in the temperature or relative contribution of Antarctic Bottom Water and play a role in the “bipolar seesaw.”

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1. Introduction

Two factors relate atmospheric CO$_2$ and the temperature of the ocean. The first is the solubility of CO$_2$ gas in sea water, which increases with decreasing temperature, and the second is the radiative, or greenhouse, forcing of the temperature of Earth as a function of the concentration of CO$_2$ in the atmosphere. These two constraints in the climate system, CO$_2$ solubility and radiative equilibrium, are not independent and must be satisfied simultaneously. Their interaction can be visualized as lines on a plot of CO$_2$ and ocean T (Figure 1a). Following an arbitrary perturbation to either component of the climate system, feedbacks governed by solubility and radiative equilibrium will cause a series of changes to both CO$_2$ and mean ocean temperature. The CO$_2$ concentration will relax toward solubility equilibrium (vertically on the plot) and ocean temperature will be forced toward radiative equilibrium (horizontally). For example, a slight increase in atmospheric CO$_2$ will warm the ocean, which will lead to a small increase in CO$_2$ due to outgassing, which in turn leads to additional warming. If the slope of the radiative equilibrium relation is steeper than that of the solubility relation, as model results described below predict, then the resulting trajectory will take atmospheric CO$_2$ and deep sea temperature to the intersection of the two lines (Figure 1a). This will occur on the $10^3$-year timescale for heat and CO$_2$ to invade the deep sea.

Much of the speculation about glacial/interglacial pCO$_2$ cycles has centered on ways of changing the distribution of CO$_2$ between the atmosphere and the ocean, independent of mean ocean temperature changes. Mechanisms include the biological pump, which exports carbon from surface waters to depth in the form of sinking particles [Volk and Hoffert, 1985], and the CaCO$_3$ cycle, which determines ocean pH [Archer and Maier-Reimer, 1994; Berger, 1978]. By forcing CO$_2$ independently of T, these scenarios displace the solubility relation vertically (Figure 1b). Vertical displacement of the solubility line moves the stable intersection between solubility and radiative equilibrium (the point to which the system relaxes), amplifying the primary CO$_2$ forcing. The signature of CO$_2$ forcing in climate records would be the temporal covariation of atmospheric CO$_2$ and deep water T quantitatively following the radiative relation.

Alternatively, variations in mean ocean temperature can drive atmospheric pCO$_2$ by the solubility mechanism. Examples of temperature-forcing mechanisms include heat transport in the North Atlantic [Stocker et al., 1992a], the albedo effect of continental ice sheets [Manabe and Broccoli, 1985], and direct orbital forcing [Milankovitch, 1930]. The effect would be to move the radiative relation horizontally (Figure 1c). As above, atmospheric pCO$_2$ and deep ocean T would relax to the new intersection of the two constraints,
We examined Mg/Ca variation in the benthic foraminifera *Uvigerina sp.* in two deep sea cores in the eastern tropical Pacific. Because the deep Pacific dominates the ocean by volume, changes in deep water temperature in this region should be reflective of mean ocean changes. Martin et al. presented a ~250-kyr Mg-derived temperature record from deep eastern tropical Pacific core TR163-31P (3°35'S 83°57' W; 3205 m) [Martin et al., 2002b]. We augment this record with a second Mg-derived temperature record for the deglaciation from nearby core TR163-20B (3°30'N 90°47' W; 3200 m). Depths in core TR163-31P were converted to age using a revised age-depth model constructed by comparison of the benthic oxygen isotope records from TR163-31P to the orbitally tuned oxygen isotope record from V19-30. The age model for TR163-20B was derived by comparison to TR163-31P (see supplementary data for both age models). Samples were prepared following the full trace metal cleaning protocol [Boyle and Keigwin, 1985] which has been tested for minor elements [Martin and Lea, 2002] and analyzed by ICP-MS for Mg/Ca following the procedure in the work of Martin et al. [2002b]. Precision is estimated to be better than 3%. Benthic δ^18O data for core TR163-20B were generated in J. Kennett’s laboratory at the University of California, Santa Barbara following standard protocol. New data are archived in Table 1.

2. Methods

We base our analysis on reconstructions of deep sea temperature using the Mg/Ca ratio of benthic foraminifera and various estimates of glacial-interglacial deep sea temperature derived from foraminiferal δ^18O [Adkins and Schrag, 2001; Chappell, 2002; Chappell and Shackleton, 1986; Labeyrie et al., 1987; Schrag et al., 1996; Shackleton, 2000]. Mg/Ca is a direct recorder of temperature based on an empirical relationship between water temperature and shell Mg/Ca, which increases exponentially with increasing temperature [Lear et al., 2000; Martin et al., 2002b; Rosenthal et al., 2000]. Foraminiferal δ^18O records changes in both temperature and δ^18O_H2O and the two signals must be separated to estimate temperature. The general approach to separating the signals has been to estimate glacial/interglacial changes in δ^18O_H2O (predominately an ice volume signal in deep waters) and assign the residual change in foraminiferal δ^18O to temperature. Methods used to characterize δ^18O_H2O include quantifying site to site differences in foraminiferal δ^18O, estimating changes in sea level from coral terraces, and measuring pore water δ^18O [Adkins and Schrag, 2001; Chappell, 2002; Chappell and Shackleton, 1986; Labeyrie et al., 1987; Schrag et al., 1996; Shackleton, 2000].
deep tropical Atlantic core implies a slightly larger change in shell Mg/Ca per °C for *Uvigerina sp.* \((\text{Mg/Ca}_{\text{Uvi}} = 0.87e0.15*T)\) which we apply directly in this study. Martin et al. [2002b] previously used the same equation adjusted to core top Mg/Ca values from core TR163-31P but our new data from core TR163-20B suggest that this probably led to underestimates of glacial cooling in the deep Pacific. Error in calculating absolute temperatures estimated from the global calibration exceeds 1.2 °C. Systematic downcore variations, however, indicate that Mg/Ca can resolve much smaller changes in temperature at a single site (14).

3. Results

3.1. Deep Sea Temperatures Over the Last 90 kyr

The warming accompanying deglaciation is recorded by both benthic foraminiferal Mg/Ca records presented here (Figure 2). Both records show similar trends of increasing Mg/Ca culminating in a maximum at the end of the deglaciation; however, higher Mg/Ca values in core TR163-20B over the latter half of the deglaciation imply a larger deep Pacific glacial-interglacial temperature change (Figure 2). Evidence in the core top calibration suggests that intense dissolution (or a fractionation effect during calcification related to saturation state) may lower the Mg/Ca of benthic foraminifera [Martin et al., 2002a]. We suspect the Mg/Ca values in core TR163-31P over the latter half of the deglaciation and early Holocene may have been lowered by dissolution. If dissolution has led to lower Mg/Ca temperatures (and thus lower reconstructed temperatures) in the youngest section of TR163-31P, we might also expect that the values for the oldest 5–10 kyr of the record here (corresponding to MIS 5) underestimate absolute temperatures. Although the two cores are relatively close in geographic setting and water depth, TR163-31P is located much closer to the continental margin while TR163-20B is on the Galapagos platform; differences in the productivity of the overlying waters and the flux of remineralizable carbon to both sites could likely account for differences in dissolution intensity. Our most compelling evidence that TR163-20B is a better recorder of Holocene bottom water temperature is that applying the *Uvigerina sp.* calibration derived directly from the Atlantic yields a temperature for the core top close to the in situ bottom water temperatures in that region and a temperature estimate for the LGM comparable to estimates derived from pore water data [Schrag et al., 1996]. In addition, the coldest temperatures derived using the new calibration yield temperatures only slightly lower than the freezing point of seawater and exceed freezing in only a few intervals. We also note the young age of core top planktonic foraminifera from TR163-20B relative to TR163-31B, the companion core to TR163-31P (1640 versus 5340 radiocarbon years) [Brown and Elderfield, 1996; Dekens et al., 2002; Le et al., 1995; Levitus and Boyer, 1994; Martin et al., 2002b; Rosenthal et al., 1997]. Despite the presence of this artifact, both records show a similar pattern of variance during the deglaciation. These new estimates of temperature change over the deglaciation are ~1 °C larger than temperatures estimated from regional comparison of benthic δ¹⁸O records [Labeyrie et al., 1987] but comparable to estimates of deep sea cooling inferred from pore water δ¹⁸O changes [Adkins and Schrag, 2001; Schrag et al., 1996].

Pacific deep water temperature over the last glacial (10–90 ka BP) correlates with Antarctic air temperature and atmospheric pCO₂ (Figure 3). Variability in benthic Mg/Ca

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Figure 2. Mg/Ca, Mg-derived temperatures, and δ¹⁸O in cores TR163-31P and TR163-20B. The black solid lines and squares represent data from core TR163-31P. The gray dashed lines and triangles represent data from core TR163-20B. Temperature estimates are derived using the empirically derived equation for *Uvigerina sp.*: Mg/Ca = 0.87e0.15*T. Data from TR163-31P were previously published in the work of Martin et al. [2002b].
between 40 and 60 ka BP is consistent with instability identified in other components of the climate system during this period [Curry and Oppo, 1997; Dansgaard et al., 1993; Shackleton et al., 2000]. The broad pattern of deep water temperature variations (Mg/Ca) over the last 90,000 years recorded in TR163-31P (solid line) and TR163-20B (dashed line) resembles changes in Antarctic air temperature recorded in the Byrd ice core (\delta^{18}O) and atmospheric CO\textsubscript{2} concentration recorded in Taylor Dome. There is clear millennial-scale variability in deep water temperature between 40 and 70 ka BP (corresponding to MIS 3) but the moderate accumulation rate (~8 cm/kyr) and minimal age-depth tie points for the deep sea sediment record prevent us from distinguishing whether the events are synchronous with events in Greenland or Antarctica. Figure modeled after Blunier and Brook [2001]. The ice core records from Byrd and GISP are on the published GISP2 timescale, correlated using the methane variations. The Taylor Dome CO\textsubscript{2} record has also been converted to the GISP2 timescale by correlating the published methane record for the core with the GISP methane record (see supplementary data) [Indermuhle et al., 2000; Brook, 2000; Brook et al., 1999; Stuiver et al., 1988].

Together these records establish a temporal correlation between atmospheric CO\textsubscript{2} and the temperature of the deep sea over the last glacial cycle (Figures 3 and 4) and also provide independent estimates of temperature changes during MIS 3.

Assuming the four small, millennial-scale oscillations in bottom water temperature and CO\textsubscript{2} between 40 and 65 ka (MIS 3) are correlative events yields a CO\textsubscript{2}/T slope of approximately 12 \mu m\textsuperscript{2}/C (Figure 4). The amplitude of millennial-scale temperature events is more likely underestimated than overestimated. With a sedimentation rate of ~8 cm/kyr at the site of TR163-31P, the millennial-scale temperature changes may have been attenuated by ~40% or more which would imply larger temperature changes (and smaller CO\textsubscript{2}/T sensitivity) [Anderson, 2001]. For the deglaciation, the slope depends on the estimate of temperature change: a 3–4°C warming from the LGM to early Holocene (Figures 2 and 3), consistent
with new Mg/Ca data and pore water estimates, yields a slope of 17 or 26 μatm/°C. Using the lower estimates of Holocene temperature from core TR163-31P yields a slope of 26–35 μatm/°C. Therefore, in either case, the paleo-data indicate that the relationship between bottom water temperature changes and CO2 for the deglaciation (26 ± 9 μatm/°C) is steeper than the value defined by the events in MIS 3 (~12 μatm/°C).

3.2. The pCO2/T Relation Derived From Numerical Experiments

For comparison with the paleo-data, we evaluated the pCO2 sensitivity to the mean temperature of the ocean through a series of ocean box and GCM carbon cycle model experiments (Figure 5) [Shin et al., 2003; Winguth et al., 2000]. The solubility of CO2 in a homogeneous seawater solution increases by about 4%/°C decreasing temperature due to changes in the equilibrium constants for CO2 solubility and pH reactions [Chipman et al., 1992]. Linearized about a midpoint of glacial-interglacial pCO2 change (~240 μatm/°C), this translates to a CO2/T relation of ~10 μatm/°C. It is tempting to use changes in sea surface temperatures to estimate the glacial/interglacial solubility effect; however, previous modeling experiments have shown that the solubility cannot be estimated from a simple weighted average of surface waters. Instead, the steady state CO2 concentration of the atmosphere is biased toward forcing in the high latitudes where deep waters form, an effect which is more pronounced in box models but still significant in general circulation models of the ocean carbon cycle [Archer et al., 2000a; Broecker et al., 1999].

To assess the effect of only changing the solubility of CO2 gas, we conducted two sets of numerical experiments. In the first set of experiments (labeled “Thermo” in Figure 5), we incrementally cooled mean ocean temperatures using a modern ocean circulation field. For simplicity, mean ocean temperature was changed in the “Thermo” GCM runs by uniformly changing surface water temperatures which led to a mean ocean temperature change. We experimented with incrementally lowering mean ocean temperature by independently lowering high-latitude and tropical sea surface temperatures using the Pandora 12-box model; we found a comparable pCO2 change per °C mean ocean temperature change for each scenario. Although records of tropical sea surface temperature change appear to be consistent with the 1–2°C oscillations during MIS 3.
that we observe in the deep sea record [Lea et al., 2001], this need not be the case to change pCO$_2$ through the solubility effect. Archer et al. previously demonstrated that the tropical sea surface temperatures are not an effective barrier to deep sea temperatures and have little direct effect on pCO$_2$ [Archer et al., 2000b].

[14] In a second set of experiments, (labeled “LGM Flow”) we allowed glacial temperature and climate forcing to drive the ocean circulation. Forcing for the large-scale geostrophic (LSG) general circulation model (GCM) was derived from CLIMAP sea surface temperatures and sea ice distributions and by an adjoint fit of sea surface salinity distribution to reproduce biochemical tracers $^{81}$C and Cd in deep sea CaCO$_3$. The Parallel Ocean Program (POP) primitive equation GCM was forced by sea surface temperature and salinities from a coupled atmosphere/ocean climate model forced by LGM pCO$_2$, orbital forcing, and land ice conditions. In spite of the complexities, all of the models, box models and GCMs alike, show a slighter lower sensitivity (smaller CO$_2$ change per °C) than the strict thermodynamic relation, in the range of 6–10 μatm/°C.

[15] Quantifying the change in deep ocean temperature resulting from radiative (greenhouse) forcing requires long integrations of coupled models of the atmosphere and ocean. A ~4000 year run of the GFDL coupled climate model predicts a mean ocean warming of 3.5°C in response to a doubling of CO$_2$, translating to ~75 μatm/°C (heavy dashed line, Figure 5) [Stouffer and Manabe, 1999]. This result may not apply to the deglaciation, however, because LGM cooling was greater than would be expected from CO$_2$ forcing alone, presumably because of the albedo of the continental ice sheets. Combined ice sheet, orbital, and CO$_2$ forcing in a coupled atmosphere/ocean model [Shin et al., 2003] resulted in a decrease in ocean temperature of 2.8°C (for LGM conditions), to give ~28 μatm/°C [Liu et al., 2002]. We also drove the POP ocean model (used in the experiments above) using sea surface forcing from the coupled simulation [Shin et al., 2003], and found a decrease in mean temperature by 3.2°C, to give 25 μatm/°C. The salient conclusions from these modeling exercises are (1) that the CO$_2$/T relation from radiative equilibrium is higher than that for solubility, and (2) its magnitude depends on whether the effects of ice sheet albedo are included in the definition of CO$_2$ climate forcing.

4. Discussion

4.1. What Is Forcing the System?

[16] Several studies have noted a correlation between Antarctic air temperature and atmospheric CO$_2$, not only over the longer timescales of 100 kyr glacial cycles but also on the shorter millennial-scale characteristic of variability during the last glacial [Cuffey and Vimeux, 2001; Indermuhle et al., 2000; Petit et al., 1999; Stuuffer et al., 1998]. Our data extend this correlation to the temperature of the deep sea over the last glacial cycle (Figure 3). Comparison of the observed and modeled CO$_2$/deep ocean T relations, however, highlights a distinction between the deglaciation and stage 3 events. The data for the deglaciation lie close to the model-derived radiative equilibrium obtained by forcing CO$_2$ and allowing for changes in the extent of ice sheets (~30 μatm/°C; Figure 5). Some mechanism for changing the pCO$_2$ of the atmosphere over the deglaciation, such as a change in the biological pump or the pH of the ocean, is required. During stage 3, the ice sheet changes due to albedo forcing presumably remained nearly constant, and the radiative equilibrium CO$_2$/T relation should have been no less than the LGM climate sensitivity (~30 μatm/°C) and, more likely, closer to modern climate sensitivity (75 μatm/°C). The data are very close to the model-derived solubility relation of ~10 μatm/°C (Figure 4).
CO₂ correlates with ocean temperature, but changes in temperature exceed what would be expected from radiative forcing alone. Apparently, over these 5-kyr events, externally forced changes in temperature drive pCO₂ according to the solubility of CO₂ in seawater.

It is clear that the CO₂ changes from the effect of cooling on solubility itself cannot explain the amplitude of the LGM/Holocene CO₂ change. Given the remarkable linearity of CO₂ and T over the deglaciation [Cuffey and Vimeux, 2001; Monnin et al., 2001], however, it is tempting to invoke an “enhanced” CO₂ solubility effect. That is, we could infer that whatever mechanism pulls pCO₂ to low glacial levels responds causally to some aspect of glacial climate, generating a greater CO₂ sensitivity to temperature forcing than the solubility relation alone. Graphically, this could be envisioned as a change in the slope of the solubility response that yields a slope closer to the slope of the radiative response and amplifies the feedback (i.e., “enhanced solubility”; Figure 6a). It has been predicted, for example, that sea ice cover in the Southern Ocean might decrease atmospheric pCO₂ [Stephens and Keeling, 2000; Morales Maqueda and Rahmstorf, 2002; Archer et al., 2003]. If sea ice extent responds to deep sea temperature [Gildor and Tziperman, 2001; Keeling and Stephens, 2001], then this mechanism could enhance the solubility CO₂ response to changing ocean temperature. Or, it could turn out that subpolar surface waters, which may undergo a larger temperature change than the global mean, control CO₂. Any form of CO₂ forcing (ocean pH or biological pump changes, for example) could contribute to an “enhanced solubility relation,” as long as it responds causally to glacial climate. Known CO₂ forcings during the deglaciation include uptake by the terrestrial biosphere and increased solubility due to dilution of ocean salinity by fresh meltwater. Taking these factors into account, we would require a solubility relation of 25–36 m atm/K if we attribute the deglacial CO₂ rise to solubility alone. However, the solubility relation inferred from the stage 3 data is similar to our model derived value of 10 m atm/K. Therefore, whichever mechanism is invoked for enhancing the solubility effect, it must be active only over orbital timescales, not over the shorter timescales.

Similarly, considerations other than the direct effect of CO₂ changes the feedbacks between CO₂ and temperature. Ultimately, the effect of ice albedo responsible for much of the LGM cooling is not a forcing independent of CO₂, but must serve as an internal feedback, amplifying...
the radiative equilibrium (Figure 6b) over times of ice volume change and possibly leading to multiple equilibria in the climate system. The amplified radiative relation must be a nonlinear one. Graphically, the combined effect could be represented by a change in the slope of the simply modeled radiative relation and its representation as a straight line (Figures 1 and 6a) that we construct as follows. At pCO2 (and minimal ice volume) close to Holocene values the radiative relation is tangent to the climate model results of 60 μatm°C (dashed thin line in Figure 6b), smoothly connecting to the global warming forecast. At LGM pCO2, the temperature of the deep sea approaches freezing; constrained by freezing, the temperature becomes insensitive to further declines in pCO2. These two regions of low temperature sensitivity, constrained respectively by low ice volume and low temperature, must be connected by an interval of higher sensitivity, corresponding to the growth of ice sheets (and the ice albedo feedback). If we couple this relation with an enhanced solubility mechanism as described above (the thick dashed line in Figure 6b), then the system will contain two stable equilibria at the intersections, one at low pCO2 and one high. The only external forcing would be the thermal effect of orbital variations. Perhaps multiple equilibria of this type will ultimately explain the reproducibility of the glacial (200 μatm) and interglacial (280 μatm) pCO2 values recorded in ice cores.

4.2. Mechanisms Driving Deep Sea Temperature Variations During MIS 3

[19] One potential mechanism to force the deep sea temperature is to alter the relative contributions of Antarctic Bottom Water (AABW) and North Atlantic Deep Water (NADW). The correlation between Antarctic air temperature [Indermühle et al., 2000] and deep water temperature suggests that the deep water warming events during MIS 3 may be linked to changes in deep water formation in the Southern Ocean. Modeling studies have demonstrated that overturning in the Southern Ocean is extremely sensitive to the density of surface waters, and that reduction in the export of a southern source bottom water leads to increased formation of North Atlantic Deep Water (NADW) [Seidov et al., 2001]. Reduced density of surface waters may have arisen from changes in fresh water forcing. Alternatively, it is also possible that AABW actually warmed during the 5-kyr events. Paired records of Mg/Ca and δ18O in planktonic foraminifera may help to distinguish between these hypotheses. However, in either case, an increase in the production of NADW relative to AABW would lead to warming of deep waters in the Atlantic and, presumably, an overall increase in mean ocean temperature.

[20] The fluctuations of deep ocean temperatures on short timescales during MIS 3 may play a role in the bipolar seesaw [Stocker et al., 1992b], the climate oscillations between North and South high latitudes. The Greenland and Antarctic ice core temperature records correlated to each other using bubble methane concentrations [Blunier and Brook, 2001] reveal a remarkably consistent series of temperature cycles. Each cycle begins with a slow warming in Antarctica while Greenland remains cold or undergoes a series of increasingly cool Dansgaard-Oeschger (D-O) temperature oscillations. Then, Greenland temperature rises abruptly, after which both Antarctica and Greenland begin a slow cooling. If the temperature of the deep sea is coupled to temperature in Antarctica, then warming in the deep sea may ultimately trigger convection in the North Atlantic resulting in sudden warming in Greenland. Thus the bipolar seesaw may be the surface manifestation of a surface/deep process.

5. Summary

[21] Measurements of benthic foraminiferal Mg/Ca from the deep eastern tropical Pacific provide estimates of the amplitude of changes in mean ocean temperature during the last glacial period. The Mg/Ca data presented here imply mean ocean temperature changes of ~3–4°C over the deglaciation, slightly larger than our previous estimates [Martin et al., 2002b] but consistent with recent estimates derived from pore water data [Adkins and Schrag, 2001; Schrag et al., 1996]. The Mg data also reveal several repeated excursions of 1 to 2°C during marine isotope stage (MIS) 3 on the timescale of thousands of years. Comparison of the Mg-derived millennial-scale temperature changes with (1) previously published high resolution benthic foraminiferal δ18O data, (2) estimates of sea-level changes from coral terraces (δ18Owater) and (3) Antarctic ice core records provide evidence of a temporal link between ocean temperature, Antarctic air temperature, and atmospheric CO2 during the last glacial cycle.

[22] Quantitative estimates of the CO2/T relation derived from the paleodata reveal two distinct slopes over the last 90,000 years, suggesting different mechanisms forced climate change on different timescales or during different climate states. Over the deglaciation, the paleodata show a large change in CO2 per degree change in mean ocean temperature consistent with previous model-derived estimates of CO2 forcing which account for the radiative effect of CO2 as well as changes in ice albedo. In this case, we still require external forcing (changes in ocean biology or pH, for example) to explain the full glacial-interglacial CO2 change. Over the ~5-kyr-long climate oscillations during the last glacial (MIS 3), however, the paleodata show a smaller change in CO2 and per °T (~12 μatm°C). This smaller change in sensitivity is consistent with new ocean GCM CO2 solubility results showing increases in atmospheric CO2 resulting from temperature-dependent solubility. The mean ocean temperature change forcing atmospheric CO2 during MIS 3 may have resulted from changes in the production of AABW, implying an active role of the Southern high latitudes in the climate oscillations and “bipolar seesaw” that operated during the last glacial.

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