Causes of ice age intensification across the Mid-Pleistocene Transition


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During the Mid-Pleistocene Transition (MPT; 1,200–800 kya), Earth’s orbitally paced ice ages cycles intensified, lengthened from ~40,000 (~40 ky) to ~100 ky, and became distinctly asymmetrical. Testing hypotheses that implicate changing atmospheric CO2 levels as a driver of the MPT has proven difficult with available observations. Here, we use orbitally resolved, boron isotope CO2 data to show that the glacial to interglacial CO2 difference increased from ~43 to ~75 μatm across the MPT, mainly because of lower glacial CO2 levels. Through carbon cycle modeling, we attribute this decline primarily to the initiation of substantive dust-borne iron fertilization of the Southern Ocean during peak glacial stages. We also observe a twofold steepening of the relationship between sea level and CO2-related climate forcing that is suggestive of a change in the dynamics that govern ice sheet stability, such as that expected from the removal of subglacial regolith or interhemispheric ice sheet phase-locking. We argue that neither ice sheet dynamics nor CO2 change in isolation can explain the MPT. Instead, we infer that the MPT was initiated by a change in ice sheet dynamics and that longer and deeper post-MPT ice ages were sustained by carbon cycle feedbacks related to dust fertilization of the Southern Ocean as a consequence of larger ice sheets.

Significance

Conflicting sets of hypotheses highlight either the role of ice sheets or atmospheric carbon dioxide (CO2) in causing the increase in duration and severity of ice age cycles ~1 Mya during the Mid-Pleistocene Transition (MPT). We document early MPT CO2 cycles that were smaller than during recent ice age cycles. Using model simulations, we attribute this to post-MPT increase in glacial-stage dustiness and its effect on Southern Ocean productivity. Detailed analysis reveals the importance of CO2 climate forcing as a powerful positive feedback that magnified MPT climate change originally triggered by a change in ice sheet dynamics. These findings offer insights into the close coupling of climate, oceans, and ice sheets within the Earth System.


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evidence that CO₂ decline was most pronounced during glacial stages. Here, we build on that work with the aim to resolve the coupling of CO₂ and climate on orbital timescales to address major unanswered questions regarding the role of CO₂ change in the MPT.

To better quantify the role of CO₂ during the MPT, we present two orbitally resolved, boron isotope-based CO₂ records generated using the calcite tests of surface-dwelling planktonic foraminifera from Ocean Drilling Program (ODP) Site 999 in the Caribbean (Fig. 3 and Figs. S1 and S2). Boron isotopes (δ¹¹B) in foraminifera have proven to be a reliable indicator of past ocean pH (18, 19) and with appropriate assumptions regarding a second carbonate system parameter (Materials and Methods and Fig. S3), allow reconstruction of atmospheric CO₂ levels. Site 999 likely remained near air–sea CO₂ equilibrium through time (20), and this is further supported by agreement of our data (blue and red in Figs. 1A and 3) with published low-resolution δ¹¹B-derived CO₂ data from ODP Site 668 in the equatorial Atlantic (11) (purple squares in Figs. 1A and 3B) and with the ice core CO₂ compilation (14).

![Fig. 1. Climate records across the MPT.](https://example.com/fig1)

Fig. 1. Climate records across the MPT. (A) CO₂ records are shown as follows: black line, ice core compilation (14); blue, our δ¹¹B-based LP260 data; red, our δ¹¹B-based eMPT data; and purple squares, low-resolution MPT δ¹¹B record of ref. 11 (all with 2σ error bars/envelopes). The range of ice core CO₂ measurements (17) from stratigraphically disturbed blue ice and their approximate ages are indicated. (B) SL records, where orange indicates the Red Sea record (21), dark blue represents Mg/Ca-based deconvolution of deep sea benthic foraminiferal oxygen isotope data (3), and pink shows a record from the Mediterranean Sea (4). (C) Dust mass accumulation rate (MAR) in a sub-Antarctic site ODP 1090 on the southern flank of the Agulhas Ridge (24). (D) LR04 benthic foraminiferal oxygen isotope stack (28). Warm intervals are highlighted by gray bars.

![Fig. 2. Changing relationship between CO₂ climate forcing and ice sheet size.](https://example.com/fig2)

Fig. 2. Changing relationship between CO₂ climate forcing and ice sheet size. Three scenarios (A–C) for the MPT intensification of glacial cycles compared with observations (D). Reconstructed SL is taken here to reflect continental ice sheet size in relationship to CO₂ climate forcing (ΔRCO₂) calculated (33) from our orbitally resolved CO₂ data. In all panels, red and blue represent conditions during our two sampling intervals before and after the MPT (i.e., eMPT and LP260), respectively. The end member scenarios posit (A) a change in ice sheet dynamics, causing ice volume to become more sensitive to unchanged G−IG climate forcing, and (B) an unchanged sensitivity of ice sheet size to forcing, with glacial intensification driven by additional CO₂ drawdown. Neither one of these two scenarios adequately describes both observed changes of increased ice sheet sensitivity (greater slope) and additional glacial CO₂ drawdown (more negative climate forcing). Here, we argue for a hybrid scenario with a change in ice sheet dynamics (possibly caused by regolith removal of ref. 8 or ice sheet phase-locking of ref. 10), allowing ice sheets to grow larger and to trigger a positive ice–dust–CO₂ feedback that promotes additional glacial intensification. In D, the regression confidence intervals account for uncertainty in both SL and ΔRCO₂ (SI Forcing to SL Relationship), but to avoid clutter, we only display the regression based on the Mediterranean SL reconstruction (4) and the uncertainty on the slope rather than the individual data points. We refer the reader to SI Forcing to SL Relationship and Fig. S7 for other SL records and full treatment of data uncertainties.
Results

Our two datasets span an early portion of the Mid-Pleistocene Transition (eMPT) from 1,080 to 1,250 kya \((n = 51)\) and for validation against the ice core CO2 record, the Pleistocene interval from 0 to 260 kya \((LP260; n = 59, including 32 recalculated data points from ref. 18)\), yielding a similar median sampling interval of \(3.5-4.5\) ky for both records. Our LP260 CO2 dataset has a confidence interval of \(\pm 20\ \mu\text{atm}\) \((2\sigma)\) and is offset by a mean of \(+7\ \mu\text{atm}\) from the ice core CO2 data when accounting for both CO2 and age uncertainties \((21)\) (Fig. 3B and SI Methodology). Comparison between our two CO2 records reveals that eMPT glacial on average experienced higher CO2 levels than LP260 (CO2: \(241 \pm 21\ \mu\text{atm}\) vs. LP260: \(203 \pm 14\ \mu\text{atm}\); \(2\sigma\)), whereas interglacial levels were indistinguishable between the two time slices (eMPT: \(284 \pm 17\ \mu\text{atm}\) vs. LP260: \(277 \pm 18\ \mu\text{atm}\); \(2\sigma\)).

Ocean and Atlantic mechanisms thought to have contributed to the most recent Late Pleistocene G-IG CO2 cycles \((22)\). For this, we force the CYCLOPS carbon cycle model \((23)\) with ODP 1090 sedimentary iron mass accumulation rates \((24)\), ODP 1094 Ba/Fe ratios \((25)\), and ODP 982/U1313 \((Fig. S1)\) benthic \(\Delta^{13}C\) variations \((26, 27)\) to represent, respectively, \(i\) sub-Antarctic dust-borne iron fertilization; \(ii\) combined changes in polar Antarctic stratification, nutrient drawdown, and export production; and \(iii\) transitions in the geometry and depth structure of the Atlantic Meridional Overturning Circulation (AMOC) \((Fig. S5)\). These mechanisms and their model sensitivities have been documented elsewhere \((23)\). Here, we invert the model and the forcing to minimize the mismatch between simulated atmospheric CO2 levels and the ice core CO2 record of the last 800 ky \((residual rms error of 12.3 \mu\text{atm})\) \((SI Carbon Cycle Modeling)\) and then, to predict atmospheric CO2 levels back to 1,500 ky \((Fig. S5)\) for comparison with our data.

We find that changes in the periodicity of simulated CO2 levels closely match those in the ice core CO2 record, in the benthic foraminiferal oxygen isotope record, and in our \(\delta^{13}B\)-based CO2 reconstruction \((Fig. S6)\). Within the relative age uncertainty between the model forcing and our \(\delta^{13}B\) record, we find that the model explains more than 60% of the variance observed in our eMPT CO2 reconstruction, in line with model and reconstruction uncertainties. The model inversion does not include any secular change in the silicate weathering cycle \((11)\) \((SI Carbon Cycle Modeling)\), so that simulated CO2 change is exclusively related to carbon redistribution within the ocean-atmosphere system and associated CaCO3 compensation dynamics \((22, 23)\).
In good agreement with the δ^{11}B-based CO₂ reconstructions and the ice core CO₂ measurements, the model inversion yields (i) insignificant (−1 ± 3 μatm; 2σ) eMPT to LP260 interglacial CO₂ change and (ii) a −22 ± 5 μatm (2σ) eMPT to LP260 decline in glacial-stage CO₂ levels (Fig. 4 and Fig. S4). In the model, we can attribute most of the additional glacial CO₂ drawdown from 400 to 800 ky, because use of ODP 1094 Ba/Fe in the model inversion results in persistent polar Southern Ocean stratification as suggested previously (25). Through our eMPT sample interval, the model reproduces the −80-ky CO₂ periodicity that is evident in our eMPT δ^{11}B data (Fig. S6), mainly because of an ∼80-ky periodicity in eMPT polar Antarctic stratification and nutrient cycling recorded in ODP 1094 Ba/Fe (25). While all three forcings (iron fertilization, Atlantic circulation, coupled polar Antarctic changes) contribute to the simulated changes in CO₂ periodicitites that are highly coherent with the MPT change in rhythm of the climate system, the iron fertilization influence dominates the MPT intensification of ice age CO₂ drawdown (Fig. S5).

**Discussion**

MPT intensification of glacial-stage CO₂ drawdown is consistent with stabilization of continental ice sheets during increasing orbital obliquity by reduced greenhouse gas forcing, thereby helping ice sheets to grow larger and for periods longer than one obliquity cycle (scenario 2 in Fig. 2). However, when we directly compare changes in SL as a measure for ice volume against CO₂ climate forcing (∆RCO₂) from our records (Fig. 2D), we find that, between eMPT and LP260, ice sheet mass increased progressively more per CO₂ lowering, thereby increasing the SL–∆RCO₂ slope in Fig. 2. This suggests an increase in ice sheet sensitivity to CO₂ forcing across the MPT, with the caveat that eMPT may not fully capture pre-MPT conditions, although it agrees with the longer-term record of Hönisch et al. (11). This finding is robust, regardless of which SL reconstruction is used (Fig. 2D). In all cases, the SL to ∆RCO₂ relationships appear to be linear, with increasing slopes from eMPT to LP260. The steepening relationship is also evident when regressing δ^{11}B to δ^{30}P relationships, with both isotope ratios measured on the same sample material (Fig. S8). Using the SL record with the best coverage of both intervals, relative SL from the Mediterranean Sea (4), we estimate 25 ± 3 and 45 ± 5 m of SL lowering for each 1-Wm⁻² reduction in radiative forcing during eMPT and LP260, respectively. Such a pronounced increase in sensitivity implicates a change in ice sheet dynamics as predicted by the regolith hypothesis (8, 9) or the establishment of marine-based ice sheet margins in East Antarctica (10) (scenario 1 in Fig. 2).

The observed changes in the SL to ∆RCO₂ relationships contain elements of both end member scenarios shown in Fig. 2. A and B, in a greater slope is possibly related to changes internal to the ice sheets (scenario 1) and amplified glacial to interglacial CO₂ climate forcing is linked (this study) to increased glacial dustiness that causes enhanced Southern Ocean iron fertilization (scenario 2). Therefore, we propose a hybrid scenario (Fig. 2C) that incorporates both heightened ice sheet sensitivity to CO₂ forcing and dust-driven ocean sequestration of CO₂ to represent the observed system change across the MPT.

First, we propose that—indeed of orbital and CO₂ forcing—a process internal to the climate system yielded greater glacial buildup of ice sheets [e.g., regolith removal (8) or ice sheet phase-locking (10)]. Second, we infer that larger ice sheets led to increased glacial atmospheric dustiness (31, 32), either directly through SL lowering or indirectly because of atmospheric cooling, drying, and/or changes in surface winds. This, in turn, induced glacial iron fertilization of the Southern Ocean, thereby effecting the 20- to 40-μatm increase in the amplitude of the G-IG CO₂ cycles documented here (Fig. 4) (11). In our hybrid scenario, the positive climate–dust–CO₂ feedback is required to (i) drive additional ice sheet growth and (ii) stabilize those ice sheets during the critical orbital phase of rising obliquity, ensuring the survival of ice sheets beyond single obliquity cycles. Therefore, regardless of the mechanism that served as the initial MPT trigger, our findings further illustrate the exquisite coupling that exists in the Earth System between climate change, ice sheet mass, and the polar ocean mechanisms that regulate G-IG CO₂ change.
for analytical help; the Integrated Ocean Drilling Program (IODP) Gulf Atlantic Ocean drill site.


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