Hydrological impact of Middle Miocene Antarctic ice-free areas coupled to deep ocean temperatures

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Oxygen isotopes from ocean sediments (δ18O) used to reconstruct past continental ice volumes additionally record deep water temperatures (DWTs). Traditionally, these are assumed to be coupled (ice-volume changes cause DWT changes). However, δ18O records during peak Middle Miocene warmth (~16–15 million years ago) document large rapid fluctuations (~1–1.5‰) difficult to explain as huge Antarctic ice sheet (AIS) volume changes. Here, using climate modelling and data comparisons, we show DWTs are coupled to AIS spatial extent, not volume, because Antarctic albedo changes modify the hydrological cycle, affecting Antarctic deep water production regions. We suggest the Middle Miocene AIS had retreated substantially from previous Oligocene maxima. The residual ice sheet varied spatially more rapidly on orbital timescales than previously thought, enabling large DWT swings (up to 4 °C). When Middle Miocene warmth terminated (~13 million years ago) and a continent-scale AIS had stabilized, further ice-volume changes were predominantly in height rather than extent, with little impact on DWT. Our findings imply a shift in ocean sensitivity to ice-sheet changes occurs when AIS retreat exposes previously ice-covered land; associated feedbacks could reduce the Earth system’s ability to maintain a large AIS. This demonstrates ice-sheet changes should be characterized not only by ice volume but also by spatial extent.

Knowledge of Earth’s glacial history and evolution through past warm periods is crucial for understanding cryosphere dynamics and future ice-sheet stability. However, the magnitude and timing of ice-sheet variations remains uncertain, even for the largest Cenozoic shifts1. Glacial history is commonly reconstructed from the oxygen isotope composition of fossil calcareous benthic foraminifera shells (δ18O), a proxy for seawater temperature and ice volume. A rapid coeval increase in global δ18O records is indicative of major ice-growth events. Over the past 40 million years, rapid expansion of the Antarctic ice sheet (AIS) during the Middle Miocene climatic transition (MMCT; ~14–13.8 million years ago (Ma)) stands out as one of the three periods of major ice growth in the δ18O record1.

The MMCT is particularly fascinating because of the hypothesized transition from a less-stable small wet-based AIS1 where meltwater encourages basal sliding and fast-moving ice to a more-stable large dry-based AIS where the base is frozen to the bedrock2. The major ice-growth event is well marked in palaeorecords by an ~1‰ increase in δ18O (Fig. 1), and thereafter, δ18O values have remained at, or above, these levels to the present day1 because a climatic threshold was crossed3. From δ18O, inferred MMCT ice growth is equivalent to the size of the entire present-day AIS, or larger1–4, but ice-sheet isotopic composition changes accounts for some of the amplitude1,13. Sequence stratigraphic estimates of sea level (independent from δ18O) indicate ~20–60 m changes4–17. Previous studies concluded this magnitude of ice growth implies the pre-MMCT AIS volumes must have been very small1,4,5. There is little evidence for notable contemporary Northern Hemisphere glaciation8, and although Antarctic topography has changed with time because of tectonics, isostatic adjustments and glacial erosion, topographic changes probably account for ~8 m sea-level equivalent (SLE) greater magnitude of ice growth for the same forcing1, leaving an additional 12–52 m necessary to explain observations.

In stark contrast to the MMCT glaciation, the preceding Miocene Climatic Optimum (MCO; ~16.8–14.8 Ma) contains the lowest δ18O values of the past 25 million years and fossil evidence for notable tundra and woody Antarctic vegetation11–12 and thereby the globally warmest period/least amount of continental ice. Evidence points to a much reduced size of the dynamic wet-based AIS in the MCO compared with its early Oligocene counterpart14–17, and large-amplitude δ18O fluctuations combined with sea-level estimates imply a highly dynamic cryosphere (Fig. 1 and Extended Data Fig. 1).

Differing ice growth–DWT relationships

Some key observations from the Middle Miocene Antarctic cryosphere still require explanation. There is a long-held assumption that continental ice volume is inherently coupled to deep water temperatures (DWTs) because expanding ice sheets are assumed to cool high-latitude regions of deep convection14,15,23,44. We would therefore expect both the MCO and MMCT to be associated with DWT changes, yet the δ18O, record, combined with independent temperature reconstructions (Fig. 1), reveals some challenging observations. During the MCO, both δ18O and DWT were highly variable (70% of δ18O variability is attributed to changes in DWT18). During the MMCT, δ18O was highly variable but DWT variations reduced in amplitude. During the MMCT glaciation, δ18O was highly variable but DWT variations were small (70% of the δ18O variability is attributed to changes in ice volume18,19). After the MMCT, δ18O and DWT were variable but less variable than during the MCO.

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Fig. 1 | Middle Miocene benthic (Cibicidoides spp.) oxygen isotope, DWT and sea-level changes. a, $\delta^{18}O_c$ splice48. b, Site 1171 DWT25, Southern Ocean (Antarctica-proximal). c, Site 761 DWT8, Indian Ocean (Antarctica-distal but in AABW path). Data locations shown on right of panels. DWTs are Mg/Ca reconstructions (uncertainty ±4 °C; relative values are considered more robust than absolutes). Other available DWT records are too low resolution/short. Data are plotted on their respective age models (full details in Supplementary Table 1). d, Sea level13,14 (eustatic estimates ×1.48 cf. ref. 49). Shading: MCO, yellow; MMCT, blue and grey; major ice-growth event (MMCT, blue). Vertical lines are indicative of the typical maximum $\delta^{18}O_c$/DWT amplitudes during the MCO (4), MMCT before the major ice-growth event (3), MMCT major ice-growth event (2) and post-MMCT (1). Since the data cover different times and resolutions, these lines are not coincident in time for panels a–c.
Interpreting $\delta^{18}$O is complicated because both temperature and the ambient seawater isotopic composition ($\delta^{18}$O$_{sw}$) are recorded. $\delta^{18}$O$_{sw}$ itself depends on global continental ice volume, the isotopic composition of this ice and localized salinity effects. Paired independent reconstructions can isolate the temperature signal, and analysis of spatially distributed $\delta^{18}$O records can reduce the salinity component. However, for the MMCT glaciation there remains the observation of a large ice increase but little DWT cooling, raising the question of why, if there is a strong coupling between ice volume and DWT as assumed, did DWT vary so much less during the MMCT when ice-sheet growth was most rapid? Here we present new climate model results assessing the impact of ice-sheet size on DWT across the MCO and MMCT. Our results confirm the findings of a previous modelling study that DWT is insensitive to ice-sheet growth at the MMCT. While the previous study explains the MMCT ice volume–DWT decoupling in terms of strong feedbacks in the coupled atmosphere–ocean–sea ice system, our study provides further mechanistic understanding of the differing degrees of Middle Miocene ice volume–DWT coupling by proposing a key role for the hydrological cycle. We here advance our understanding of the paradigm by highlighting the important role not only of ice-sheet volume but also of spatial ice-sheet coverage in determining the DWT response to glaciation during the MMCT and the MCO.

We use a fully coupled atmosphere–ocean–vegetation general circulation model, HadCM3LB-M2.1aE, configured with Middle Miocene palaeogeography (see Methods and Fig. 2). For our initial assessment using preindustrial CO$_2$ concentrations, we find that AIS expansion from ice-free (ICEFREE) to the 22 m SLE regional-scale ice-sheet configuration (ICE$_{PART22m}$) and from ICE$_{PART22m}$ to the 55 m SLE continental ice-sheet configuration (ICE$_{FULL55m}$) reduces DWT by ~0.5 °C for each step; thus, here, ice growth and DWT are coupled (Fig. 3a). However, AIS expansion from ICE$_{FULL55m}$ to the 90 m SLE continental-scale ice-sheet configuration (ICE$_{FULL90m}$) does not cause further deep ocean cooling (by contrast, a slight temperature increase is seen); thus, here, ice growth and DWT are decoupled (ice-volume changes do not affect DWT).

We propose that coupling between ice-sheet volume and DWT occurs only until the ice sheet reaches the coast because the ice-albedo feedback mechanism and vegetation–climate interactions invoke additional feedback processes identified here. To demonstrate it is ice-sheet spatial extent (rather than height/volume) that is
coupled to DWT, we carry out a non-realistic sensitivity study imposing AIS configurations spanning extreme endmembers from ice-free to ice covered but keep ice volume constant. We assume the ice sheet is of ‘skin thickness’ (no effective change in elevation as compared with the ice-free state, nominally 1 m SLE when fully ice covered; ‘ICE FULL 1m’) and vary the ice extent longitudinally, latitudinally and topographically. We use preindustrial CO2 concentrations throughout and conduct an additional high-CO2 sensitivity test (~850 ppm; Fig. 4). Combining our results, we find a strong relationship between ice-free extent and DWT, with no evidence of nonlinearity (Fig. 3b).

The mechanism linking ice cover to DWTs

Our modelling results suggest summer ‘ice-free’ Antarctica (ICEFREE, Fig. 3, column 1) would be warm and wet because the land–sea thermal contrast drives monsoon winds, which transport moisture into the Antarctic continental interior from the Southern Ocean (Fig. 5a–c). This moisture falls over the relatively warm continent as rain, not snow, during the summer months (Fig. 5b) and over much of the continent during the winter months for the two highest CO2 scenarios. Summer Antarctic temperatures and precipitation are similar to proxy reconstructions for a vegetated Antarctica4,21,22,29–31. A comprehensive model–data comparison (Supplementary Note) indicates peak CO2 would need to be >850 ppm for a complete overlap with proxy reconstructions, in agreement with recent MCO reconstructions32. ICE FREE also results in the warmest, freshest deep ocean of all the simulations (Fig. 5d–e). Surface runoff from the active hydrologic cycle, being less saline and thus less dense than the seawater it drains into, forms a polar halocline at the surface. This halocline reduces ventilation of the deep ocean (Fig. 5d), weakening overturning. In our simulations, deep water is in all cases produced primarily in the Southern Ocean; thus, DWTs are determined by southern sinking regions. Antarctic Bottom Water (AABW) production never ceases completely in the model for any scenario (Supplementary Discussion B).

In ICEFULL1m, ICEFULL55m and ICEFULL90m (Fig. 5, columns 3–5), cold surface temperatures near the ice sheet and the large increase in albedo cause localized radiative cooling of the air column and a reduction in vapour holding capacity. The land–sea thermal contrast reduces (Fig. 5a), and the summer monsoon system ceases to operate. Katabatic winds form as the cold dense air flows away from the elevated areas towards the coast (Fig. 5c). The interaction between the winds and sea ice is complex and depends on background CO2 (Fig. 5b and Supplementary Discussion A). Reduced precipitation (Fig. 5b) and subsequent runoff reduce...
ocean stratification (Fig. 5d), permitting the cold surface waters to sink more freely from the continental shelf into the abyss (Fig. 5e). Increased AABW production invigorates ocean ventilation.

Empirical studies show a clear relationship between ice-sheet volume and spatial extent\(^\text{33}\), implying ice-sheet thickness is limited by spatial extent. Therefore, to grow vertically, an ice sheet must also...
We infer a consequent ~0.8 °C mean temperature change in the more details of the boundary conditions used. However, our results show that ice extent had a larger impact on DWT than did CO2. For the MMCT glaciation, the most-recent CO2 reconstructions show at most a 170 ppm reduction from ~570 to 400 ppm, for which we infer from our ice-covered model simulations a temperature drop of 0.5–0.8 °C at the two sites (Supplementary Discussion E). This is consistent with the reconstructions (Fig. 1) if we assume Antarctica was ice covered before the MMCT glaciation (that is, little DWT change occurred as a result of increasing ice-sheet extent). In the absence of ice-sheet changes, we find a minimal effect of orbital configuration on DWT (Supplementary Discussion F).

From thin and vulnerable to thick and established

We introduce the hydrological cycle as a crucial mechanism mediating the link between the DWT and the ice spatial extent (rather than absolute volume), thus explaining the different degrees of coupling between ice-sheet changes and DWT during the MMCT and MCO. Our new results lead us to propose that DWT varied by up to 4 °C during the MCO because the spatial extent of ice and vegetation rapidly altered. Taken together with existing δ18Oc, temperature, vegetation and CO2 reconstructions, this implies the AIS had retreated substantially during the MCO, when average CO2 concentrations were probably 470–630 ppm, reaching 780–1,100 ppm at times. Previous work clearly demonstrates the dynamic behaviour of a small AIS when driven by CO2 changes combined with orbital forcing. How far exactly the ice sheet retreated during these warmest intervals, however, is unknown. Ice-sheet modelling suggests a retreat exposing 60–70% of the Antarctic land surface is consistent with the palaeorecord. Other work concludes a retreat even greater than this, perhaps even ice-free. The evidence for vegetation, including trees, growing on the continent throughout the MCO implies both warm and wet conditions, and it is suggested the moisture supply derived from the Southern Ocean. To achieve this, our results indicate a greater reduction of ice is needed than the ICEFULL,22m scenario ice-sheet extent because the Wilkes Land winds are directed landward in ICEFREE but seaward for ICEFULL,22m. We suggest these monsoon moisture-carrying winds induced by spatial ice retreat could provide an explanation for major ice advance onto the continental shelf in the Ross Sea, during the MCO occurring at the same time as open water and woody vegetation in the Wilkes Land.

We further infer DWT varied so much less during the MMCT when the AIS volume was growing rapidly because it had already extended to cover most of the continent before the major ice-growth event, in agreement with previous findings. Thus, the ice sheet subsequently increased mainly in thickness, not area, and so DWTs were largely unaffected because, without the additional ice–albedo feedback, changes to the hydrological cycle were much smaller. Post-MMCT (label 1 in Fig. 1), both δ18Oc and DWT are variable, but less so than during the MCO. The exact degree of coupling, and its mechanism, needs to be explored in a model set-up that includes marine-based ice sheets and ice shelves, not included in this study. However, the physical limits on seawater temperatures (~1.8 °C) will set the lower boundary on possible temperature changes as climate cools.

Interpretation of our results leads us to support a highly dynamic MCO AIS, and state, alongside CO2, it was changes in ice-sheet area and proximity to the coast, not volume, that were of key importance for global DWTs. This fundamentally changes the way we should characterize ice-sheet changes and how we must view the long-term δ18Oc records spanning greenhouse–icehouse transitions.
In the absence of independent temperature proxies, it must not be assumed that DWTs scale with ice-volume changes. While we do not propose the MCO Antarctica was ever completely ice-free, our results demonstrate any spatial retreat of the AIS can increase precipitation, causing associated warming of the deep ocean—changes perhaps having the ability both to accelerate ice melt of ice shelves and glaciers through hydrodynamizing from increased precipitation falling into crevasses and to accelerate ice melt of marine-based subglacial basins. Although the temperature changes resulting from changing ice-sheet extent are similar to those resulting from CO2 changes, our study does not include feedbacks to the carbon cycle or to the ice sheet itself, and therefore the significance of our results could be greater than indicated here. Our non-realistic sensitivity studies using only a skin thickness of ice demonstrate the importance of both surface albedo and roughness for a hydrologic control on DWT evolution. It is therefore possible that our mechanism could operate in areas even without complete ice loss if these two factors change markedly, for example, in regions of debris-covered glaciers, rock glaciers, vegetation-covered rock glaciers and ‘glacier mice’, which all increase in the context of retreating ice glaciers, in regions of accumulating dark particles (dust and soot) and in regions of glacier algae, which bloom in supraglacial meltwater.

**Online content**

Any methods, additional references, Nature Research reporting summaries, source data, extended data, supplementary information, acknowledgements, peer review information; details of author contributions and competing interests; and statements of data and code availability are available at https://doi.org/10.1038/s41561-021-00745-w.

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Methods
The model used in these experiments is the fully coupled atmosphere–ocean GCM HadCM3LB-M2.1eA16 with the interactive vegetation model TRIFFID17. This is the low-resolution ocean version of HadCM318, and both the atmosphere and the ocean components have a resolution of 2.5° latitude by 3.75° longitude. The model is run without the need for flux adjustments in the modern configuration by the now-standard practice of removing Iceland from the land–sea mask53. Eddies in the model are parameterized using a spatially constant coefficient of 1,000 m s⁻¹ according to the Gent–Williams scheme54. Vertical diffusion is parameterized using the Richardson number-dependent formulation and a background diffusivity of 10⁻⁵ m² s⁻¹ at the surface, which increases linearly at a rate of 2.8 x 10⁻⁵ m² s⁻¹ with depth55. A linear mixing profile with depth has been shown to be able to capture that vertical mixing is strongest over topography56. Of all the Paleoclimate Modelling Intercomparison Project (PMIP) 1.5 and PMIP2 models, only the HadCM3-based models, which had such a linear scheme, managed to correctly simulate the Last Glacial Maximum inverse relationship between the volume and production rate of AABW56. As such, we have confidence in key processes determining the transfer of surface climate signals to depth are represented well in the ocean component of our model.

The background palaeogeography19 is representative of Middle Miocene conditions. Notable features as compared with modern are that the Panama Gateway and the Indonesian Seaway are open, Australia is located 4° farther south, and the Barents Sea and the Bering Strait are closed. Note that in all experiments, the Eastern Tethys Seaway is also closed. Land topography also differs from modern in that the Tibetan Plateau and the Andes are more than 2,000 m lower than at present, and the Rocky Mountains are about 1,000 m higher. Greenland is also more than 2,000 m lower than modern, and is ice-free.

To investigate Middle Miocene climate, we used a suite of CO₂ and Antarctic ice-cover sensitivity studies. Four Antarctic ice-cover configurations were each simulated at five CO₂ concentrations: 180, 280, 400, 560 and ~850 ppm (in accordance with the uncertainty and temporal variability in CO₂ reconstructions for the Middle Miocene14). Each of these experiments was carried out in triplicate, giving a total of 20 simulations. The four ice-sheet configurations are defined as follows. First, an ice-free Antarctica using an Antarctic bedding configuration appropriate for the late Oligocene20 is referred to as ICE_FREE. Ice-free Antarctica is initialized in the Top-down Representation of Interactive Foliage and Flora Including Dynamics (TRIFFID) model as covered in the plant functional type ‘shrubs’. Although unrealistic for the Middle Miocene, as the extreme endmember scenario, this scenario helps to place our findings into context. Second, a skin-thickness ice-covered Antarctica (<1 m SLE), where the topography is kept the same as in the ice-free case in 1, is referred to as ICEFULL1m. Third, a modern-like ice-covered Antarctica (~55 m SLE), using the palaeogeographic configuration for the Middle Miocene21, is referred to as ICEFULL55m. Fourth, a larger-than-modern ice-covered Antarctica (~90 m SLE) is referred to as ICEFULL90m.

The second to fourth configurations have the same areal extent and differ in ice-sheet height only. In accordance with the proposed 90 m drop in sea level over the MMCT19, boundary conditions for an ice sheet of this proportion were developed by assuming one-third of the ice is associated with isostatic depression and then applying a uniform increase in elevation across the continent to obtain a 90 m SLE. Additionally simulated at 280 ppm (using the topography of the ICEFULL55m scenario) is the spatial extent of the regional-scale ice sheet simulated by a model that includes ice-shelf hydrofracture and ice-clip collapse of 22 m SLE, referred to as ICEPART22m. See Fig. 2 for geographic details of the Antarctic bedrock configuration appropriate for the late Oligocene21.

All of the Middle Miocene simulations continue from a 2,100-year integration under late Miocene boundary conditions22 and have been run for a further 2,000 years. Supplementary Figs. 6–9 show the evolution of DWTs throughout the simulations and give us confidence that they have stabilized sufficiently for us to draw conclusions.

Data availability
The climate model output data are available for analysis and download at https://www.paleo.bristol.ac.uk/umm/modelscripts/papers/Bradshaw_et_al_2021.html. It is possible to reproduce the information in Figs. 2, 3, 5 and 6 via this interface as well as download the data itself and the ancillary information (palaeogeography and ice-sheet configuration).

Code availability
The UK Met Office made available the source code of HadCM3 via the Ported Unified Model release (https://www.metoffice.gov.uk/research/approach/collaboration/unified-model/partnership). Enquiries regarding the use of HadCM3 should be directed in the first instance to the UM Partnership team, who can be contacted at um_collaboration@metoffice.gov.uk. The main repository for the Met Office Unified Model (UM) version corresponding to the model presented here can be viewed at http://cos.ncas.ac.uk/code_browsers/UM45/UMbrowser/index.html (registration required). The code detailing the changes required to update HadCM3 to HadCM3LB-M2.1 are available as a supplement to Valdes et al.19.

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Author contributions
C.D.B., C.H.L. and D.J.L. conceived the project and directed the research with the assistance of A.M.deB.; C.D.B. conducted and interpreted the modelling with the assistance of D.I.L., A.M.deB. and P.M.L.; C.D.B. compiled and interpreted the proxy records with the assistance of C.H.L., A.M.deB., H.K.C. and S.M.S.; C.D.B. led the writing of the paper. All authors contributed to writing the manuscript.

Competing interests
The authors declare no competing interests.

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Extended Data Fig. 1 | Benthic (Cibicidoides spp.) oxygen isotope δ¹⁸O, and deep water temperatures (DWT) changes from the Mg/Ca proxy through the Middle Miocene. a Site 761 in the Indian Ocean⁸, b Site 1171 in the Southern Ocean²⁵. Data are plotted on their respective age models. DWT uncertainty is ±4 °C; relative values are considered more robust than absolutes.