Late Miocene climate cooling and intensification of southeast Asian winter monsoon

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The late Miocene offers the opportunity to assess the sensitivity of the Earth's climate to orbital forcing and to changing boundary conditions, such as ice volume and greenhouse gas concentrations, on a warmer-than-modern Earth. Here we investigate the relationships between low- and high-latitude climate variability in an extended succession from the subtropical northwestern Pacific Ocean. Our high-resolution benthic isotope record in combination with paired mixed layer isotope and Mg/Ca-derived temperature data reveal that a long-term cooling trend was synchronous with intensification of the Asian winter monsoon and strengthening of the biological pump from ~7 Ma until ~5.5 Ma. The climate shift occurred at the end of a global δ13C decrease, suggesting that changes in the carbon cycle involving the terrestrial and deep ocean carbon reservoirs were instrumental in driving late Miocene climate cooling. The inception of cooler climate conditions culminated with ephemeral Northern Hemisphere glaciations between 6.0 and 5.5 Ma.
The late Miocene (~11.6 to 5.3 Ma) stands out as a period of exceptional interest within the long-term Cenozoic cooling trend toward icehouse conditions, as it represents a geologically recent interval of relative global warmth that was marked by profound environmental change in both terrestrial and marine ecosystems (e.g., refs. 1,2). This interval provides a unique opportunity to document climate-carbon cycle dynamics on a warmer-than-modern Earth and, thus, to help guide models and constrain predictions of climate change and sensitivity. The detailed sequence of climate events and the range of natural climate variability through the late Miocene remain, however, poorly understood, mainly due to the scarcity of continuous, high-resolution climate archives. Most available records are adequate for characterizing long-term trends or mean states, but do not capture short-term climate events and orbital-scale phase relationships required to assess, for example, changes in ice volume, monsoon intensity, and carbon fluxes.

Multiproxy temperature reconstructions indicated that a reduced sea surface temperature (SST) zonal gradient generally prevailed in the tropical Pacific Ocean during the late Miocene, in contrast to the sharper gradient that developed during the late Pliocene to Pleistocene3,4. This mean state, typically referred to as “permanent El Niño-like conditions” or “El Padre”5,6, exerts a fundamental impact on regional and global climate because it is dynamically linked to the weakening of the Hadley and Walker circulation and the state of upper-ocean stratification4,7,8. However, it is difficult to reconcile the late Miocene warmth with inferred low atmospheric pCO2 levels, close to preindustrial values (e.g., ref. 9). This apparent decoupling between climate warmth and atmospheric pCO2 variations has prompted intense debate about the dynamics of warm climates and the role of pCO2 as driver of climate variations under different background states (e.g., refs. 10,11). The widely held view of sustained equable warmth through the late Miocene was recently challenged by SST reconstructions, which revealed that a prolonged global cooling spell occurred between ~7 and ~5.5 Ma2,12. During this period, SST dipped below early Pliocene values in the Mediterranean, the gradient intensifying at higher latitudes. Site 1146 is located at the northwestern edge of the western Pacific warm pool (WPWP) and is also ideal to constrain meridional variations in the extent of the WPWP and to monitor changes in southeast Asian monsoon climate. The extended, carbonate and clay-rich succession recovered at this site20, thus, provides an outstanding archive of subtropical climate variations, allowing new insights into the dynamics and forcing processes of late Miocene climate evolution.

Results
Late Miocene astronomically tuned chronology. The 1146 benthic foraminiferal stable isotope records based on Cibicidoides wuellerstorfi and/or Cibicidoides munda (5 cm sample spacing along a composite sequence or splice from Holes 1146A and 1146C) were tuned to an eccentricity-tilt (ET) composite target generated from the La2004 orbital solution21 (Supplementary Note 1; Supplementary Figs. 1–4). The tuned series exhibits a

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**Fig. 1** Location of ODP Site 1146 within a slope basin at the northern margin of the South China Sea. Satellite image and bathymetry from ref. 75. Paleogeographic reconstruction at 10 Ma (simplified from ref. 19).
mean sedimentation rate of ~3 cm kyr$^{-1}$ with a maximum of 5 cm kyr$^{-1}$ and a minimum of 1 cm kyr$^{-1}$ and a mean temporal resolution of ~2 kyr over the interval 9–5 Ma (Supplementary Fig. 4B). We note that the short eccentricity (100 kyr) period is prominent in the untuned and tuned benthic $\delta^{18}$O records from 9.0 to 7.9 Ma and that the low amplitude of short eccentricity and high amplitude of obliquity between ~7.7 and 7.2 Ma are clearly reflected in the benthic $\delta^{18}$O series, which exhibits pronounced 41 kyr variability over this interval (Fig. 2c; Supplementary Figs. 1–4). The interval 6–5 Ma includes prominent benthic $\delta^{18}$O maxima, identified as T8, TG4, TG12, TG14, TG20, and TG22. These globally traceable $\delta^{18}$O enrichments provide additional stratigraphic control. Superimposed on higher frequency variations (mainly 41 kyr), the untuned benthic and planktic $\delta^{13}$C series display low-frequency oscillations that broadly relate to the ~400 kyr long eccentricity cycle (Fig. 2a; Supplementary Fig. 1). Comparison of benthic and planktic $\delta^{18}$O and $\delta^{13}$C records plotted in the depth and time domains shows that the original spectral characteristics are preserved following the tuning procedure.

**Temporal trends in benthic and planktic stable isotopes.** Benthic and planktic $\delta^{18}$O exhibit different long-term trends and short-term variability from ~9 to ~5 Ma (Fig. 2b, c; Supplementary Fig. 2A, B). Between 9.0 and 7.3 Ma, mean benthic $\delta^{18}$O varies between 2.5 and 2.3‰ and displays an overall decreasing trend of ~0.2‰ (Fig. 2c, Supplementary Fig. 4B) with standard deviations (SDs) ranging between 0.20 and 0.08‰ (Supplementary Fig. 2E). Lowest mean benthic $\delta^{18}$O values of 2.3‰ are reached between 7.7 and 7.3 Ma during an interval of pronounced 41 kyr variability (SD mainly between 0.16 and 0.12‰) (Supplementary Fig. 2B, E). In contrast, planktic $\delta^{18}$O oscillates around a mean of ~2.0‰ from 9.0 to 7.7 Ma and exhibits a slightly increasing trend to ~1.9‰ from 7.7 to 7.3 Ma (Fig. 2b; Supplementary Fig. 2A). From 9.0 to 7.3 Ma, SDs in planktic $\delta^{18}$O fluctuate between 0.09 and 0.19‰ around a mean of 0.13‰ (Supplementary Fig. 2C).

Between 7.3 and 6.9 Ma, the Site 1146 high-resolution benthic and planktic $\delta^{18}$O records reveal a series of previously unrecognized short-term climate events (Fig. 2b, c). The benthic $\delta^{18}$O curve resolves a ~80 kyr long positive excursion (~0.3‰ amplitude) centered at 7.2 Ma followed by a rebound before a stepwise increase at 7.1–7.0 Ma (0.2‰ mean increase) (Fig. 2c; Supplementary Fig. 2B). The benthic $\delta^{18}$O shift was coupled to a stepwise increase in planktic $\delta^{18}$O between 7.2 and 7.1 Ma (Fig. 2b, c; Supplementary Fig. 2A, B), which marked the onset of a long-term trend of substantially heavier values (0.3‰ mean
increase) and overall higher amplitude variability (mean SD 0.18‰ after 7.0 Ma vs. 0.13‰ prior to 7.2 Ma) (Supplementary Fig. 2A, C).

From ~7 to 5.2 Ma, mean benthic δ¹⁸O oscillates around 2.5‰ (Supplementary Fig. 2B). Amplitude variability is relatively low until 6.1 Ma except for a few transient minima (mean SD 0.11‰ fluctuating between 0.07 and 0.16‰), but it increases markedly from 6.1 to 5.5 Ma (mean SD 0.14‰), culminating in the high-amplitude maxima TG12, TG14, TG20, and TG22, when peak benthic δ¹⁸O values reach ~3‰ (Fig. 2c; Supplementary Fig. 2E, F). Mean planktic δ¹⁸O shows a slight increasing trend (~0.2‰ mean increase to a maximum of ~1.5‰) after ~7 Ma, followed by a decrease to mean values between −1.8 and −1.6‰ after 5.5 Ma (Supplementary Fig. 2A). In contrast to benthic δ¹⁸O, the amplitude variability of planktic δ¹⁸O (Supplementary Fig. 2C, D) increases markedly after ~7 Ma and generally remains high until 5.2 Ma (mean SD 0.18‰, fluctuating between 0.10 and 0.24‰).

Benthic and planktic δ¹³C exhibit consistent long-term (400 kyr) and short-term (41 kyr) variability from ~9 to ~5 Ma (Fig. 2a; Supplementary Figs. 5, 6A). Between 9 and 7.7 Ma, benthic and planktic δ¹³C oscillate between 1.1 and 0.1‰ (mean 0.73‰, SD 0.19‰) and between 2.8 and 1.6‰ (mean 2.21‰, SD 0.24‰), respectively. Between ~7.7 and ~7.0 Ma, a characteristic feature of the benthic and planktic δ¹³C records is the massive, long-term decrease of ~1‰, from 1.0 to ~0.3‰ (benthic) and 2.7 to 1.3‰ (planktic), which corresponds to the global decline in δ¹³C known as the late Miocene carbon isotope shift (LMCIS, Fig. 2a). The final phase of the LMCIS at 7.2–7.0 Ma coincides with a distinct sharpening of the gradient between planktic and benthic δ¹³C (Δδ¹³C) between 7.1 and 7.0 Ma, which lasts until ~5.5 Ma (Figs. 3a and 4b), and with a stepwise increases in benthic and planktic δ¹⁸O (Figs. 2b, c, and 4e, f). Between 7 and 5 Ma, following the end of the LMCIS, benthic and planktic δ¹³C fluctuate between 0.47 and −0.74‰ (mean −0.05‰, SD 0.20‰) and between 2.16 and 0.88‰ (mean 1.59‰, SD 0.21‰), respectively (Fig. 2a).

Coherence (k) between benthic and planktic δ¹⁸O remains overall lower than between benthic and planktic δ¹³C over the interval 9–5 Ma, with a maximum of 0.90 on the obliquity band at a frequency of 0.02455 cycles kyr⁻¹ and a second maximum on the precession band of 0.83 at a frequency of 0.0425, whereas coherence on the 400 and 100 kyr eccentricity bands remains insignificant at the 80% level (k below 0.4) (Supplementary Note 2; Supplementary Figs. 5, 6A, B). In contrast, benthic and planktic δ¹³C exhibit high coherence both in their long-term (400 kyr) and short-term (predominantly 41 kyr) variability throughout the interval 9–5 Ma (Supplementary Note 2; Supplementary Fig. 6A, B), implying that both are influenced by changes in the global carbon cycle. Coherence on the long eccentricity band (frequencies of 0.0023–0.0029 cycles kyr⁻¹) fluctuates around 0.9, while coherence on the short eccentricity (0.0097 cycles kyr⁻¹) and obliquity (0.247 cycles kyr⁻¹) bands reaches maximum of 0.98 and 0.96, respectively.

**Evolution of mixed layer temperatures.** Reconstructed mixed layer temperatures based on Mg/Ca ratios in the planktic
foraminifer *Globigerinoides sacculifer* (without correction for secular changes in seawater Mg/Ca concentrations) vary between 22 and 28 °C from 8.2 to 5 Ma (Fig. 3b). These relatively low values are most likely due to long-term secular changes in seawater Mg/Ca concentrations (e.g., ref. 29). A correction for seawater Mg/Ca concentration, using a latest Miocene seawater Mg/Ca ratio of ~4.5 mol mol$^{-1}$ and a modern day seawater Mg/Ca of ~5.1 mol mol$^{-1}$ following the calculation outlined in ref. 31 results in a temperature increase of ~1.5 °C. This is consistent with estimates of a 4.3 mol mol$^{-1}$ Pliocene Mg/Ca ratio and 0.9–1.9 °C Pliocene Mg/Ca ocean temperatures, deducted from coupled seawater-test Mg/Ca temperature laboratory calibrations of *Globigerinoides ruber* (ref. 32). Corrected temperatures (Figs. 3b, 4a, and 5a) are close to the range of modern seasonal variability in the area of Site 1146, between 24.7 and 27.8 °C at 25 m water depth (Supplementary Note 3; Supplementary Fig. 7) with the warmest reconstructed temperatures exceeding modern mixed layer temperatures by almost ~2 °C, whereas the lowest temperatures during transient cold events remain ~1 °C below modern winter temperatures.

Mixed layer temperature estimates at Site 1146 exhibit a slight warming trend between 7.3 and 7.1 Ma, increasing from (uncorrected) mean values of 25.0 °C ($n$ = 122, SD 0.7 °C) between 8.1 and 7.3 Ma to 25.8 °C ($n$ = 36; SD 0.9 °C) between 7.3 and 7.1 Ma (Fig. 3b). This transient warming is interrupted by a pronounced cooling step of ~2 °C between 7.1 and 6.9 Ma, previously documented at this site by a low-resolution study (34). This cooling step marks the onset of a long-term trend of cooler temperatures, lasting until ~5.5 Ma, which coincides with a long-term increase in the mean and amplitude variability of seawater δ$^{18}$O (Supplementary Figs. 8, 9A, C) and with an intensification of the gradient between planktic and benthic δ$^{13}$C (Figs. 3a and 4b). Between 6.9 and 5.7 Ma, uncorrected temperatures fluctuate between 21.9 and 26.8 °C with a mean of 24.2 °C ($n$ = 199, 1.0 °C SD). After ~5.7 Ma, uncorrected temperatures generally oscillate between 22.1 and 27.0 °C and exhibit a sustained increase of 0.8 °C in their mean value ($n$ = 146, 0.9 °C SD). Between 6.0 and 5.5 Ma, transient episodes of intense mixed layer cooling by ~2 to ~3 °C coincide with sharp increases in planktic δ$^{18}$O and the prominent benthic δ$^{18}$O maxima TG12, TG14, TG20, and TG22 (Fig. 5a, e, f).
In this study, we discuss relative changes in mixed layer temperatures rather than absolute values, since the history of Miocene seawater Mg/Ca composition is still poorly constrained (Supplementary Note 3). Furthermore, our interpretations are based on relatively short-term changes in mixed layer temperatures, which are not affected by the long-term variability of seawater Mg/Ca concentration with changes in the order of millions of years due to the long residence time of these elements in the ocean.

Discussion
The planktic and benthic δ\(^{18}\)O signals at Site 1146 differ markedly in their long- and short-term trends between 9 and 5 Ma, pointing to a decoupling of regional hydrology and the evolution of the Antarctic ice sheet, which formed the main component of the cryosphere during the middle and late Miocene (e.g., refs. 35,36). Mixed layer temperature and seawater δ\(^{18}\)O reconstructions at Site 1146 additionally support that substantial changes in southeast Asian hydroclimate occurred after ~8 Ma, which accelerated at ~7 Ma, but do not appear closely connected with Southern Hemisphere high-latitude climate (benthic δ\(^{18}\)O) trends.

Between 7.1 and 6.9 Ma, upper-ocean temperatures at Site 1146 document a sustained cooling (~2 °C mean cooling), which persisted until ~5.7 Ma (Figs. 3b and 4a). This cooling was associated with a long-term increase in the mean and amplitude variability of seawater δ\(^{18}\)O (Supplementary Note 3; Supplementary Fig. 9A, C), as indicated by a previous low-resolution study. These trends signal a change in the amount and/or δ\(^{18}\)O composition of precipitation and runoff, likely associated with changes in the provenance and/or seasonality of precipitation toward a more pronounced monsoonal seasonality and a more temperature-controlled seasonality of rainwater (i.e., δ\(^{18}\)O depleted winter precipitation). We attribute these hydrological changes in the northern South China Sea after ~7 Ma to cooling and drying of the Asian landmass and a related southward shift of the average summer position of the Intertropical Convergence Zone (ITCZ), resulting in decreased influence of tropical convection and intensified dry winter monsoon over southeast Asia. Drying and cooling on the Asian continent at ~7 Ma are supported by independent lines of evidence including enhanced dust accumulation rates in northern China, vegetation change in central China and an increase in the mean grain size of the terrigenous...
Fig. 6 Comparison of late Miocene inter-basinal benthic δ¹⁸O and δ¹³C gradients. **a** Vertical distribution of δ¹³C in world’s oceans following late Miocene δ¹³C shift. Averaged values over interval 7–5 Ma for key sites in the Pacific, Atlantic, Indian, and Southern Oceans compiled from refs. 22,54,55,76. NADW: North Atlantic Deep Water, PCW: Pacific Central Water. **b** Late Miocene evolution of inter-basinal benthic δ¹⁸O and δ¹³C gradients: comparison of Pacific ODP Site 1146 and Atlantic ODP Sites 926 and 99922,23 and equatorial Pacific IODP Site U133855 over interval 9–5 Ma. Stable isotope data from Sites 926 and 999 are plotted on originally published age models. Over the interval 8.2–7.5 Ma, the Site U1338 age model was adjusted to that of Site 1146 by tuning the δ¹³C records. Lilac shading marks global δ¹³C decline coincident with planktic δ¹⁸O increase and high-amplitude obliquity modulation of benthic δ¹⁸O. Blue shading marks final stage of global δ¹³C decline. Light orange shading marks climate warming after 5.5 Ma. Smooth curve in **b** fitted using locally weighted least squared error (Lowess) method.
sediment fraction at Site 1146.  In addition, the predominance of a mollusc group preferring cold-arid conditions in the loess and paleosol layers of central China between 7.1 and 5.5 Ma is indicative of a dominant winter monsoon regime over this period.

In contrast to these major hydrological changes in the Northern Hemisphere, mean benthic δ^{18}O suggests only a relatively modest, stepwise glacial expansion of the Antarctic ice sheet and/or deep water cooling at ~7 Ma (Figs. 2c and 4f). However, the intensification of the southeast Asian winter monsoon after ~7 Ma was associated with a long-term trend toward heavier benthic δ^{18}O maxima, which culminated in the most intense maxima (TG22, 20, 14, and 12 between 5.8 and 5.5 Ma) within the entire late Miocene, before reversing in the
early Pliocene (Figs. 2c and 5f). During these extreme events, benthic δ¹⁸O hovered close to 3‰ (~0.4–0.6‰ increase), which is in the range of late Pliocene values and of intermediate values between peak Holocene and glacial levels at the same location 18. A previous study 42 related these intense δ¹⁸O maxima to episodes of Antarctic ice volume increase. However, the Site 1146 records show that benthic δ¹⁸O maxima (TG22, 20, 14, 12, 4, and T8) coincide with planktic δ¹⁸O maxima between 6.0 and 5.0 Ma, indicating concomitant variations in deep water δ¹⁸O and regional hydrology, which is closely linked to extra-tropical Northern Hemisphere climate variations (Fig. 5e, f). Mixed layer temperatures additionally show concurrent sharp decreases of 2–3 °C during these events (Fig. 5a), implying massive Northern Hemisphere cooling down to subtropical latitudes. The occurrence of ice rafted debris in North Pacific 43 and North Atlantic 44 sediment cores further indicates Northern Hemisphere ice builds between 6 and 5 Ma. Expansion of Arctic sea ice during these intense cold spells would have increased the positive albedo feedback, amplifying cooling and favoring ice growth. Together these lines of evidence support the development of ephemeral Northern Hemisphere ice sheets (e.g., Greenland, Alaska, Labrador) between 6.0 and 5.5 Ma that were highly susceptible to insolation forcing.

The Site 1146 records additionally reveal that climate cooling and intensification of the winter monsoon at ~7 Ma coincided with the final stage of a long-term, global benthic, and planktic δ¹³C decline 27,28 (LMCIS, Figs. 2a and 4c, d). This major shift of ~1‰, which started close to 7.8 Ma, has been interpreted as a global decrease in the δ¹³C of the dissolved inorganic carbon pool, although its causes remain debated (e.g., refs. 22,45,46). A long-held view among contending hypotheses is that this global δ¹³C decrease was linked to the late Miocene spreading of C4 grasslands, which are better adapted to low pCO₂ and to reduced seasonal precipitation. This large-scale expansion is thought to have resulted in a transfer of ¹³C from the marine to the terrestrial carbon pool 17–49. A decrease in atmospheric pCO₂, linked, for instance, to long-term changes in the oceanic and/or terrestrial carbon inventories, could explain climate cooling after ~7 Ma associated with equatorward migration of the ITCZ and contraction of the WPWP.

The gradient between benthic and planktic δ¹³C additionally provides insights into changes in atmospheric pCO₂, since it is influenced by two main factors: the sequestration efficiency of the biological pump and equilibration processes between the upper ocean and the atmosphere (Supplementary Note 4; Supplementary Fig. 10). The equilibration time for δ¹³C in the mixed surface layer of the ocean exhibits a linear correlation to the ratio of dissolved inorganic carbon to pCO₂, which leads to slow equilibration and elevated δ¹³C in the mixed layer of the ocean with respect to the atmosphere under low pCO₂. Recent model simulations showed that accelerated equilibration under elevated atmospheric pCO₂ decreases the isotopic disequilibrium, leads to lower upper ocean δ¹³C and, thus, decreases the gradient between the δ¹³C of surface and deep water masses 50. Consequently, the vertical δ¹³C gradient in the ocean exhibits a gentler slope under high atmospheric pCO₂ and steepens during intervals of declining pCO₂.

Steepening of the gradient between planktic and benthic δ¹³C after ~7 Ma at Site 1146, when mixed layer temperatures also declined (Fig. 3a, b), suggests that pCO₂ levels decreased, eventually reaching levels that enabled the formation of transient Northern Hemisphere ice sheets between 6.0 and 5.5 Ma. This steeper gradient also denotes a prolonged interval of substantially enhanced marine productivity and accumulation rates of biogenic components (“biogenic bloom” originally described in ref. 51) at numerous locations in the Pacific, Indian, and Atlantic Oceans (e.g., ref. 52 and references therein). In the eastern equatorial Pacific Ocean, opal, and carbonate deposition reached a maximum between 7.0 and 6.4 Ma during the peak of the biogenic bloom in the region 46. Thus, a plausible scenario is that changes in nutrient supply and/or pathways stimulated marine productivity after ~7 Ma. Steepening of the equator to pole temperature gradient associated with global cooling after ~7 Ma (Fig. 3a, b; ref. 2) promoted intensified of the Hadley and Walker circulation with repercussions on the wind-driven circulation and precipitation patterns (e.g., ref. 53). The strengthening of winds may have in turn fostered upwelling and ocean fertilization, helping to drive intense biogenic blooms through the Pacific Ocean, which enhanced carbon storage and decreased pCO₂ in the ocean in a positive feedback loop.

Previous work showed that the amplitude of the LMCIS differs in ocean basins (e.g., ref. 54). In particular, a comparison of benthic δ¹³C profiles indicates that the gradient between the Pacific and Atlantic Oceans intensified during the final stage of the LMCIS (Fig. 6b; ref. 34). The steeper inter-basin gradient after ~7 Ma cannot be explained by increased production and southward advection of North Atlantic Deep Water, as this relatively warm and/or fresh (lighter δ¹⁸O) and δ¹³C-enriched water mass appears not to have spread into the South Atlantic and Southern Ocean, which remained influenced by colder, denser (heavier δ¹⁸O) and δ¹³C-depleted water masses through the late Miocene (Fig. 6a, b; ref. 34). Alternatively, the steeper inter-basin δ¹³C gradient after ~7 Ma may be driven by increased export of nutrient enriched waters with a lower preformed δ¹³C from the Southern Ocean into the Pacific Ocean (e.g., ref. 54) and/or to enhanced primary productivity and nutrient regeneration in the low-latitude Pacific Ocean.

Comparison of benthic δ¹³C profiles from Site U1338 in the abyssal equatorial Pacific Ocean 52 and the shallower Site 1146 in the northwestern subtropical Pacific Ocean (Fig. 6b) shows that the composition of Pacific water masses changed after 7.2 Ma. The divergence of the δ¹³C records after 7.2 Ma indicates expansion of a δ¹³C-depleted central Pacific deep water mass into shallower depths during the peak of the biogenic bloom. If primarily driven by increased productivity and nutrient regeneration in the Pacific and Indian Oceans, the expansion of a δ¹²C-enriched deep water mass after 7.2 Ma also implies increased carbon storage in the deep Pacific Ocean. The global efficiency of the biological pump reflects a balance between high- and low-latitude regions with different

![Fig. 7](image-url) Middle to late Miocene climate cooling steps coincident with unusual congruence of the Earth’s orbit. a Miocene to Pleistocene (16–0 Ma) benthic δ¹⁸O and δ¹³C records from ODP Site 1146, compiled from refs. 15–18 and this work. Blue arrows mark main phases of glacial expansion/deep water cooling; 3 pt smooth: 3 pt moving mean. Light orange shading marks global δ¹³C decline coincident with planktic δ¹⁸O increase and high-amplitude obliquity modulation of benthic δ¹⁸O. Blue shading marks final stage of global δ¹³C decline. Light orange shading marks climate warming after 5.5 Ma. b Comparison of benthic (C. wuellerstorfi and/or C. mundulus) δ¹³C from ODP Site 1146 (15–17 and this work) with orbital parameters (eccentricity and obliquity from ref. 21) reveals similar sequence of climatic events during three Miocene cooling episodes with strikingly similar background orbital configuration. Blue shading marks cooling episodes following an extended period of high-amplitude variability in obliquity congruent with low variability in short eccentricity (gray shading). Light orange shading marks transient warming episodes coincident with high-amplitude variability in short eccentricity.
sequestration efficiency. Thus, enhanced productivity and organic matter export in the tropical and subtropical ocean may increase global sequestration efficiency and lower atmospheric pCO2, even when deep water formation occurs in high-latitude areas with an inefficient biological pump.

The integrated benthic isotope data in Site 1146 provide the first continuous, highly resolved time series from a single site that span the last 16.4 Myr (Fig. 7a). These extended records track the transition from a warmer middle Miocene climate phase with a reduced and highly dynamic Antarctic ice cover (until ~14 Ma) to an increasingly colder mode with more permanent and stable ice sheets in the late Miocene. These records additionally allow us to evaluate the long-term relationship between radiative forcing and the response of the ocean/climate that is imprinted on the benthic δ18O signal. For instance, the 41 kyr obliquity cycle is especially prominent in the benthic δ18O series between 7.7 and 7.2 Ma (Fig. 2c), during a configuration of the Earth’s orbit, when high-amplitude variability in obliquity is congruent with extremely low-amplitude variability in short eccentricity (Supplementary Figs. 3, 4A, E). The onset of the ~80 kyr long positive excursion in benthic δ18O centered at 7.2 Ma notably coincides with minima in obliquity (41 kyr) and in eccentricity (100 kyr, 400 kyr, and 2.4 Myr amplitude modulation) (Fig. 7b; Supplementary Fig. 3).

At obliquity and eccentricity minima, lower summer insolation at high latitudes inhibits melting of snow and ice. This conjunction of climatic forcing factors likely fostered a sustained cold phase in the high latitudes that lasted through two consecutive obliquity cycles, resulting in an extended benthic δ18O positive excursion. Renewed high-amplitude variations in eccentricity and precession together with maximum amplitude variability in obliquity probably drove the successive rebounds between 7.2 and 7.0 Ma.

This sequence of climatic events, as well as their background orbital configuration were strikingly similar during two previous Miocene cooling episodes: the middle Miocene climate transition at ~13.9 Ma, which resulted in major expansion of the East Antarctic ice sheet and the less pronounced late Miocene cooling step at ~9.0 Ma (Fig. 7b). In all three instances, the 41 kyr cycle initially stands out in the benthic δ18O signal during a protracted period of high-amplitude variability in obliquity, congruent with low variability in short eccentricity. A marked enrichment in benthic δ18O (0.2–0.3‰), indicative of ice growth and/or deep water cooling toward the end of this interval, coincides with prolonged minima in eccentricity, lasting ~100–200 kyr. Subsequent rebounds at peak insolation, linked to changes in eccentricity cadence (from 400 to 100 kyr variability), indicate episodes of transient ice sheet disintegration and deep water warming. This unusual orbital congruence appears propitious to high-latitude cooling in the Northern and Southern Hemispheres, although boundary conditions differed markedly during these three intervals of climate change. The middle Miocene cooling step occurred in a much warmer climate phase, characterized by substantially lighter mean benthic δ18O (Fig. 7a, b). At this time, the less extensive ice cover over Antarctica was likely more dynamic and highly responsive to Southern Hemisphere summer insolation, in contrast to the more expanded Antarctic ice sheet during the late Miocene. This long-term perspective illustrates the nonlinear response of the ocean/climate system to orbital forcing and the role of internal feedback processes including ice sheet hysteresis, latitudinal temperature gradients, ocean circulation and CO2 exchange between terrestrial, atmospheric and oceanic reservoirs.

Arguably, the uncertainty of the CO2 forcing during the Miocene remains a major challenge for defining the characteristics and dynamics of warmer climate states. Although, current pCO2 reconstructions show no significant change through the late Miocene, with levels staying close to or slightly above preindustrial levels, uncertainties in excess of 200 p.p.m. (see compilations in refs. 59,60) preclude assessment of variability and sensitivity to CO2 forcing within the critical preindustrial to modern range. To test the sensitivity of outputs to pCO2 uncertainties, some simulations of late Miocene climate using coupled atmosphere-ocean circulation models have applied atmospheric pCO2 concentrations in the preindustrial range (~280 p.p.m.), as well as more elevated levels of 400–450 p.p.m. (e.g., refs. 60,61). These studies indicated major changes in vegetation distribution, and sea ice cover61 in the Northern Hemisphere under these different pCO2 states. In particular, forest areas decreased and the albedo of the Eurasian and North American landmasses increased under lower pCO2, due to the markedly lower (by 4–10 °C) mean air temperatures and reduced precipitation during boreal winter. These findings are in agreement with a previous modeling study62, which found that vegetation changes were more important than palaeogeography in determining late Miocene climate. Simulated mean summer SST and sea ice concentrations in the Arctic Ocean61 also showed that the region is highly sensitive to relatively small changes in pCO2, as a year-round sea ice cover prevails in the central Arctic Ocean at preindustrial levels, whereas summer conditions are ice-free at concentrations of 450 p.p.m. This difference in seasonal ice cover is critical because it implies very different feedbacks in terms of albedo and heat exchange with far-reaching repercussions for global climate61.

Recent model simulations of Pliocene warmth additionally highlighted the importance of feedbacks associated with cloud albedo and ocean mixing in driving changes in meridional and zonal temperature gradients, despite relatively modest changes in pCO2.63–65. Data from this study support that subtropical climate cooling and intensification of the southeast Asian winter monsoon after ~7 Ma were synchronous with decreasing pCO2 (Figs. 3a and 4b) within a global context of steepening meridional thermal gradients.65 We speculate that this late Miocene climate shift was associated with a relatively small decline in pCO2, which was amplified by a conjunction of positive feedbacks. Variations in Northern Hemisphere sea ice cover and vegetation in concert with changes in ocean–atmosphere circulation were likely instrumental in driving late Miocene climate, as illustrated by recent modeling simulations of late Miocene climate60–62. The dynamic behavior of the ocean–climate system between 9 and 5 Ma suggests a tight coupling between carbon cycle variations and low-latitude climate evolution. In particular, our results show that changes in Antarctic ice volume were not the primary driver of late Miocene climate development and that low-latitude processes, including monsoonal wind forcing of upper-ocean circulation and productivity had a strong influence on climate-carbon cycle dynamics. Inception of colder climate conditions at ~7 Ma during the final stage of the LMCIS coincided with intensification of the Asian winter monsoon and strengthening of the Pacific Ocean’s biological pump, which persisted until ~5.5 Ma. This suggests that changes in the global carbon cycle involved transfer of terrestrial carbon in a cooling, drying climate, as well as fluctuations in the carbon storage capacity of the deep ocean and the sedimentary carbon sink. Ephemeral Northern Hemisphere glaciations between 6.0 and 5.5 Ma additionally indicate that atmospheric pCO2 levels hovered close to and occasionally reached the threshold necessary for Northern Hemisphere ice sheet growth during this period.
Methods

Revision of shipboard sediment splice. Cores were sampled in ~5 cm intervals (~2 ky time resolution) from a composite sequence (shipboard splice) of Holes 1146A and 1146C (Cores 1146C-30X to 1146C-39X). After comparison of the shipboard natural gamma ray, color reflectance, magnetic susceptibility data, and overlapping benthic isotope records over the splice tie points, we made the following modifications to the original shipboard splice: we defined a new tie point between Cores 1146C-38X and 1146C-39X (1146C-38X-4, 122 cm below sea floor (msbf) tie to 1146C-39X-2, 107 cm at 359.87 msbf), based on the match of isotopic data from Holes 1146A and C. This adjustment resulted in the addition of an 80 cm splice segment from Hole 1146A to the meter composite depth (mdc) scale.

Benthic and planktic foraminiferal stable isotopes. All samples were oven dried at 40 °C and weighed before washing over a 63 μm sieve. Residues were oven dried at 40 °C on a sheet of filter paper, then weighed and sieved into different size fractions. We measured δ18O and δ13C in the epifaunal benthic foraminifiers C. wuellerstorfi and/or C. mundulus and on the mixed layer foraminifer G. sacculifer. Well-preserved tests were broken into large fragments, cleaned in alcohol in ultrasonic bath, then dried at 100 °C. Stable isotopes were measured with a Finnigan MAT 253 mass spectrometer at the Leibniz Laboratory, University of Kiel. The instrument is coupled on-line to a Carbo-Kiel Device (Type IV) for automated CO₂ preparation from carbonate samples for isotopic analysis. Samples were reacted by individual acid addition (99% H₂PO₄ at 75 °C). On the basis of the performance of international and lab-internal standard, the precision is better than ±0.09‰. Paired measurements in middle Miocene samples from ODP Sites 1146 and 1237 previously indicated no significant offset in δ18O and δ13C between C. wuellerstorfi and C. mundulus [26]. Results were calibrated using the National Institute of Standard and Technology (Gaithersburg, MD) carbonate isotope standard and NBS (National Bureau of Standard) 19 and NBS 20, and are reported on the Vienna PeeDee Belemnite scale.

Astronomically tuned chronology. The chronology is based on minimal tuning [27] of the benthic oxygen isotope record to a computed orbital solution [21] (Supplementary Note 1). We used an ET composite with equal weight of eccentricity and obliquity as tuning target and tuned δ18Om into ET [8]. Fifty-four duplicate measurements show a precision of 0.0971. Prior to analysis, both time series were interpolated to 2 ky time steps and standardized (zero mean and standard deviation). Software is available at http://www.pol.ac.uk/home/research/waveletcoherence/

Data availability. Data are archived at the Data Publisher for Earth and Environmental Science (https://doi.pangaea.de/10.1594/PANGAEA.887393).

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**Author contributions**

A.E.H. and W.K. conceived the project and wrote the paper. A.E.H., S.C.C., and N.A. generated stable isotope measurements. K.G.D.K., J.J., and J.L. generated Mg/Ca measurements. A.E.H., W.K., and S.C.C. analyzed the data. All authors contributed ideas and discussed the paper.

**Additional information**

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