## REVIEW SUMMARY

### CLIMATE CHANGE

# Past climates inform our future

Jessica E. Tierney*, Christopher J. Poulsen, Isabel P. Montañez, Tripti Bhattacharya, Ran Feng, Heather L. Ford, Bärbel Hönisch, Gordon N. Inglis, Sierra V. Petersen, Navjit Sagoo, Clay R. Tabor, Kaustubh Thirumalai, Jiang Zhu, Natalie J. Burls, Gavin L. Foster, Yves Goddéris, Brian T. Huber, Linda C. Ivany, Sandra Kirtland Turner, Daniel J. Lunt, Jennifer C. McElwain, Benjamin J. W. Mills, Bette L. Otto-Bliesner, Andy Ridgwell, Yi Ge Zhang

**BACKGROUND:** Anthropogenic emissions are rapidly altering Earth's climate, pushing it toward a warmer state for which there is no historical precedent. Although no perfect analog exists for such a disruption, Earth's history includes past climate states—"paleoclimates"—that hold lessons for the future of our warming world. These periods in Earth's past span a tremendous range of temperatures, precipitation patterns, cryospheric extent, and biospheric adaptations and are increasingly relevant for improving our understanding of how key elements of the climate system are affected by greenhouse gas levels. The rise of new geochemical and statistical methods, as well as improvements in paleoclimate modeling, allow for formal evaluation of climate models based on paleoclimate data. In particular, given that some of the newest generation of climate models have a high sensitivity to a doubling of atmospheric $CO_2$, there is a renewed role for paleoclimates in constraining equilibrium climate sensitivity (ECS) and its dependence on climate background state.

**ADVANCES:** In the past decade, an increasing number of studies have used paleoclimate temperature and $CO_2$ estimates to infer ECS in the deep past, in both warm and cold climate states. Recent studies support the paradigm that ECS is strongly state-dependent, rising with increased $CO_2$ concentrations. Simulations of past warm climates such as the Eocene further highlight the role that cloud feedbacks play in contributing to high ECS under increased $CO_2$ levels. Paleoclimates have provided critical constraints on the assessment of future ice sheet stability and concomitant sea level rise, including the viability of threshold processes like marine ice cliff instability. Beyond global-scale changes, analyses of past changes in the water cycle have advanced our understanding of dynamical drivers of hydroclimate, which is highly relevant for regional climate projections and societal impacts. New and expanding techniques, such as analyses of single shells of foraminifera, are yielding subseasonal climate information that can be used to study how intra- and interannual modes of variability are affected by external

climate forcing. Studies of extraordinary, transient departures in paleoclimate from the background state such as the Paleocene-Eocene Thermal Maximum provide critical context for the current anthropogenic aberration, its impact on the Earth system, and the time scale of recovery.

A number of advances have eroded the "language barrier" between climate model and proxy data, facilitating more direct use of paleoclimate information to constrain model performance. It is increasingly common to incorporate geochemical tracers, such as water isotopes, directly into model simulations, and this practice has vastly improved model-proxy comparisons. The development of new statistical approaches rooted in Bayesian inference has led to a more thorough quantification of paleoclimate data uncertainties. In addition, techniques like data assimilation allow for a formal combination of proxy and model data into hybrid products. Such syntheses provide a full-field view of past climates and can put constraints on climate variables that we have no direct proxies for, such as cloud cover or wind speed.

**OUTLOOK:** A common concern with using paleoclimate information as model targets is that non-$CO_2$ forcings, such as aerosols and trace greenhouse gases, are not well known, especially in the distant past. Although evidence thus far suggests that such forcings are secondary to $CO_2$, future improvements in both geochemical proxies and modeling are on track to tackle this issue. New and rapidly evolving geochemical techniques have the potential to provide improved constraints on the terrestrial biosphere, aerosols, and trace gases; likewise, biogeochemical cycles can now be incorporated into paleoclimate model simulations. Beyond constraining forcings, it is critical that proxy information is transformed into quantitative estimates that account for uncertainties in the proxy system. Statistical tools have already been developed to achieve this, which should make it easier to create robust targets for model evaluation. With this increase in quantification of paleoclimate information, we suggest that modeling centers include simulation of past climates in their evaluation and statement of their model performance. This practice is likely to narrow uncertainties surrounding climate sensitivity, ice sheets, and the water cycle and thus improve future climate projections. ∎
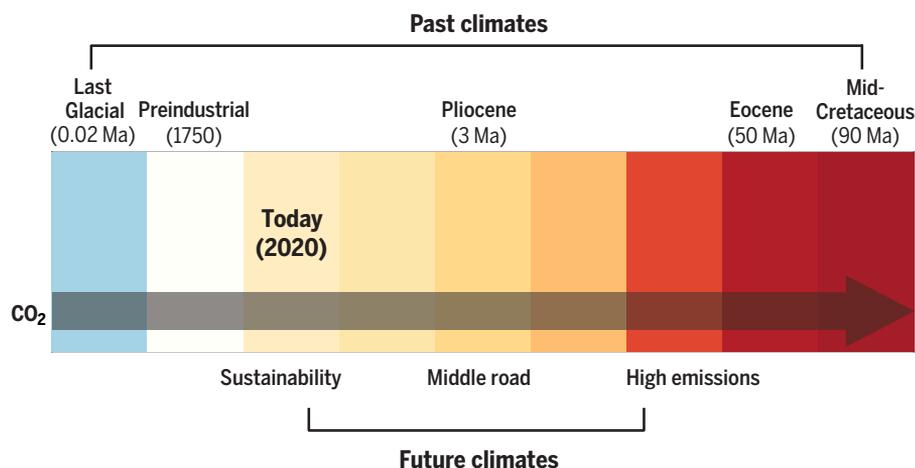


**Past climates provide context for future climate scenarios.** Both past (top) and future (bottom) climates are colored by their estimated change in global mean annual surface temperature relative to preindustrial conditions, ranging from blue (colder) to red (warmer). "Sustainability," "Middle road," and "High emissions" represent the estimated global temperature anomalies at year 2300 from the Shared Socioeconomic Pathways (SSPs) SSP1-2.6, SSP2-4.5, and SSP5-8.5, respectively. In both the past and future cases, warmer climates are associated with increases in $CO_2$ (indicated by the arrow). Ma, millions of years ago.

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As the world warms, there is a profound need to improve projections of climate change. Although the latest Earth system models offer an unprecedented number of features, fundamental uncertainties continue to cloud our view of the future. Past climates provide the only opportunity to observe how the Earth system responds to high carbon dioxide, underlining a fundamental role for paleoclimatology in constraining future climate change. Here, we review the relevancy of paleoclimate information for climate prediction and discuss the prospects for emerging methodologies to further insights gained from past climates. Advances in proxy methods and interpretations pave the way for the use of past climates for model evaluation—a practice that we argue should be widely adopted.

The discipline of paleoclimatology is rooted in the peculiarities of the geological record, which has long hinted that Earth's climate can change in profound ways. In possibly the first paleoclimate study, the 17th-century English physicist Robert Hooke concluded, based on observations of large turtles and ammonites in Jurassic rocks, that conditions in England had once been much warmer than now (1). Since then, paleoclimate studies have revolutionized our view of the climate system (2), documenting both warm and cold worlds much different from the one we inhabit and establishing the link between atmospheric $CO_2$ and global temperature (Fig. 1).

Although paleoclimatology continues to narrate the history of Earth's climate, it also plays an increasingly central role in understanding future climate change. The study of past climate has never been more relevant than it is now, as anthropogenic activities increase atmospheric greenhouse gas concentrations and modify the land surface and ocean chemistry at a rate and scale that exceed natural geologic processes. $CO_2$ levels are higher now than at any point in at least the past 3 million years and, at the current rate of emissions, will exceed concentrations typical of the past 30 million years by 2300 (Fig. 1). In this context, past climates are windows into our future (3)—the geological record is the only observational source of information for how the climate system operates in a state much warmer than the present.

The challenge for paleoclimatology is that there are few direct quantitative records of past climate (e.g., temperature, precipitation). Instead, we make use of "proxies," surrogates for climate variables that cannot be measured directly. In some cases, the physical occurrence (or absence) of a proxy (like glacial deposits) reveals information about past environmental conditions. More often, geochemical data (such as elemental and stable isotope ratios) stored in fossils, minerals, or organic compounds are used to infer past conditions. The discovery of new proxies, improvements in modeling and analytical techniques, and the increasing number of proxy records are actively expanding the utility of paleoclimate information. These innovations are refining our understanding of how the climate system responds to large changes in atmospheric $CO_2$ and provide insights into aspects of past climates (such as seasonality and interannual variability) that were heretofore unknowable.

Among the most important contributions that paleoclimatology can make is the evaluation of Earth system models that we rely on for projecting future climate change. The phys-

ical parameterizations in these sophisticated models are often tuned to fit the preindustrial or historical record (4). However, the latter is short in duration and samples a single climate state with a narrow $CO_2$ range. The performance of climate models under extreme forcing very different from the present is not commonly assessed, even though the models are used to project changes under high-emissions scenarios. When these models are used to simulate past warm climates, they often predict surface temperatures that are too cold and pole-to-equator temperature gradients that are too large (5). However, a new generation of models, alongside developments in proxy techniques and analysis, now provides opportunities to more fully exploit past climates for model evaluation and assessment of key metrics of the climate system.

## Past climates inform key processes

Earth's paleoclimate record contains tremendous variability. Over the past 100 million years, the climate gradually transitioned from an ice-free world of exceptional warmth [the mid-Cretaceous, 92 million years ago (Ma); Fig. 1] to the cold ice ages of the past few million years, glacial worlds with kilometers-thick ice caps covering one-fourth of the land surface [such as the Last Glacial Maximum (LGM), 21,000 years ago; Fig. 1]. Between Cretaceous and LGM extremes lie intermediate warm climates such as the early Eocene (53 to 49 Ma) and Pliocene (5.3 to 2.6 Ma) (Fig. 1). This long-term climate transition was far from steady—short-lived hyperthermal events (6) and cold stadials (7) punctuated the slower trends.

Atmospheric $CO_2$ concentrations generally follow these swings in global temperature (Fig. 1). Geochemical modeling demonstrates that the balance of geological sources (degassing through volcanism) and sinks (weathering and sedimentation) explains the general features of $CO_2$'s trajectory (8) and establishes causality—high $CO_2$ leads to high temperatures. The apparent exceptions to this rule, including the end-Cretaceous and early Paleocene (70 to 60 Ma) and the Miocene (23 to 5.3 Ma), are areas of active research. One explanation for the decoupling of $CO_2$ and temperature is that uncertainties associated with the proxies blur the relationship. Estimation of past $CO_2$ is challenging. Beyond the ice core record (9), $CO_2$ information comes from geochemical data, such as isotope ratios of boron and carbon, or paleobotanical indicators such as the density of leaf stomata. All of these proxies require assumptions about the physical, chemical, and biological state of the past that are not completely understood, sometimes leading to misinterpretations of the signal (10). Proxy methodologies and assumptions continue to be refined, and there is some indication that $CO_2$ at the end of the Cretaceous may have been higher than that shown in Fig. 1 (11). It

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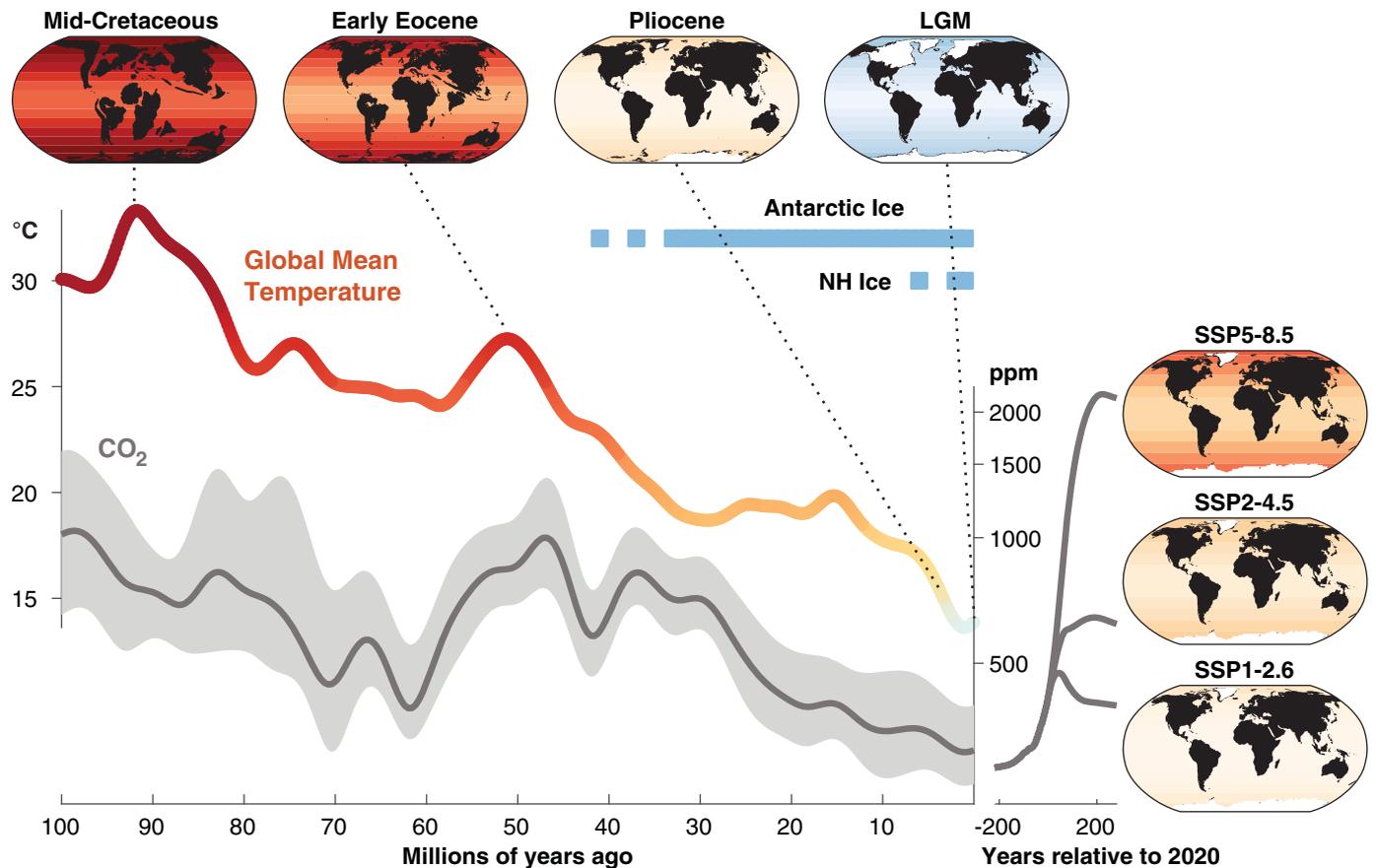
**Fig. 1. Paleoclimate context for future climate scenarios.** Global mean surface temperature for the past 100 million years is estimated from benthic $\delta^{18}O$ (*2*, *106*) using the method of (*104*). $CO_2$ is estimated from the multiproxy dataset compiled by (*105*) with additional phytane data from (*107*) and boron data from (*108*) and (*11*). Data with unrealistic values (<150 ppm) are excluded. The $CO_2$ error envelopes represent 1σ uncertainties. Note the logarithmic scale for $CO_2$. Gaussian smoothing was applied to both the temperature and $CO_2$ curves to emphasize long-term trends. Temperature colors are scaled relative to preindustrial conditions. The maps show simplified representations of surface temperature. Projected $CO_2$ concentrations are from the extended SSP scenarios (*109*). Blue bars indicate when there are well-developed ice sheets (solid lines) and intermittent ice sheets (dashed lines), according to previous syntheses (*2*). NH, Northern Hemisphere.

is also possible that these discrepancies have another explanation, such as a greater-than-expected role for non-$CO_2$ forcings and feedbacks. If the paleoclimate record has taught us anything, it is that the more we probe, the more we learn.

Past climate states were profoundly different from today. Their global mean temperatures, latitudinal temperature gradients, polar ice extents, regions of deep-water formation, vegetation types, patterns of precipitation and evaporation, and variability were all different. These differences are invaluable because they provide rich evidence of how climate processes operated across states that span the range of $CO_2$ concentrations [400 to 2000 parts per million (ppm)] associated with future emissions scenarios [the Shared Socioeconomic Pathways (SSPs); Fig. 1]. Under the sustainable SSP1-2.6 scenario, in which emissions are curtailed and become net negative by the end of the 21st century, $CO_2$ concentrations

would be stabilized near Pliocene levels (Fig. 1). By contrast, under the fossil-fuel intensive SSP5-8.5 scenario, $CO_2$ concentrations would approach or even exceed Eocene or mid-Cretaceous levels (Fig. 1). These past warm climates therefore serve as targets against which to measure the increasingly complex generation of climate models that are used for future climate prediction.

Past climates are not perfect analogs for future states—continental configurations are increasingly different with age, and they often represent equilibrium climates as opposed to transient changes associated with rapid greenhouse gas emissions (*12*). But as benchmarks for climate models, ancient climates need not be perfect analogs. Indeed, the differences are advantageous; they provide true out-of-sample validation for the strength and stability of key feedbacks, large-scale responses of the hydrological cycle, and the most ubiquitous metric of all, climate sensitivity.

**Paleoclimate constraints on climate sensitivity**

Equilibrium climate sensitivity (ECS) has been widely adopted as a simple metric of how responsive the Earth's climate system is to radiative forcing. It is defined as the change in global near-surface air temperature that results from a sustained doubling in atmospheric $CO_2$ after the fast-acting (time scales of years to decades) feedback processes (water vapor, clouds, snow) in the Earth system reach equilibrium. The fifth assessment report of the Intergovernmental Panel on Climate Change (IPCC) determined that ECS was likely between 1.5° and 4.5°C, a large range that has remained essentially unchanged for 40 years (*13*). Because the environmental impacts, socioeconomic implications, and mitigation time scales are very different for a low versus a high ECS (*14*), narrowing its range has always been a high priority.

That models with either a low or high present-day ECS can match historical observations (*15*)
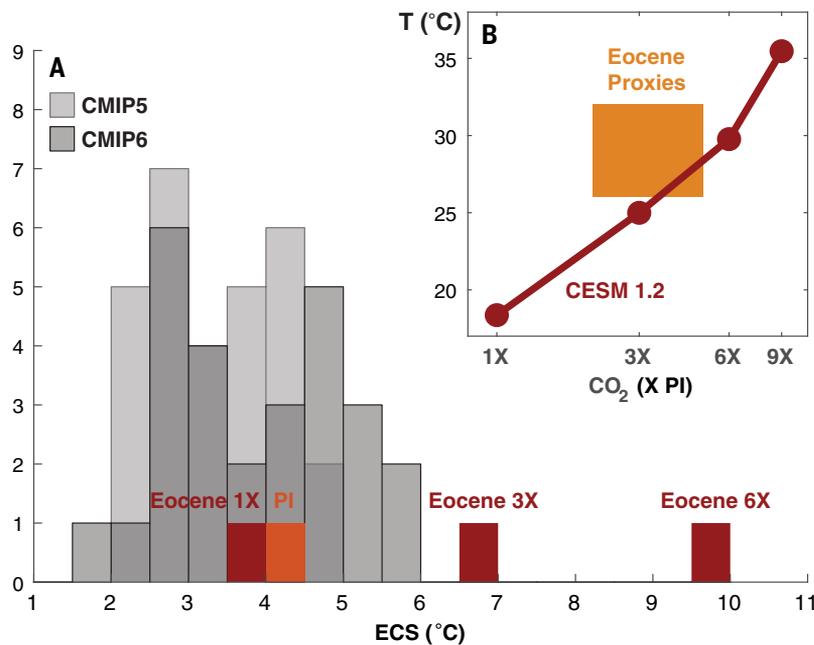
**Fig. 2. Constraining ECS through simulation of the early Eocene.** (**A**) ECS in CMIP5 and CMIP6 models [gray bars; (*15*)] compared with ECS in the CESM1.2 preindustrial (PI, orange bar) and Eocene simulations with 1X, 3X, and 6X preindustrial $CO_2$ levels (red bars). (**B**) $CO_2$ concentrations [times preindustrial level (X PI)] versus global mean temperature according to early Eocene proxies (yellow region) compared with the results from the CESM1.2 Eocene simulations. Proxy $CO_2$ estimates are a $2\sigma$ range derived from the collection plotted in Fig. 1. See (*19*) for details of the Eocene climate simulations and proxy global mean temperature estimation.

suggests that preindustrial and industrial climatic changes provide insufficient constraints on this important metric. Furthermore, the emerging view is that ECS is dependent on, and changes with, the background climate state—specifically, it increases in warmer climates (*16–19*). Past warm climates therefore provide key constraints on the range of plausible ECS values as well as the strength of feedbacks involved. Simulations of the early Eocene provide a salient example. Figure 2 shows a comparison between the ECS of Coupled Model Intercomparison Project (CMIP) phase 5 and phase 6 models (used in the last and the upcoming IPCC assessments) and the ECS of preindustrial and Eocene simulations conducted with the National Center for Atmospheric Research (NCAR) Community Earth System Model (CESM) version 1.2 (*19*). Doubling $CO_2$ in an Eocene experiment with preindustrial $CO_2$ (285 ppm; labeled 1X) yields an ECS similar to the preindustrial experiment and overlaps with the CMIP range (Fig. 2). This indicates that non-$CO_2$ Eocene boundary conditions, including the position of the continents and the absence of continental ice sheets, do not have a large effect on ECS in CESM1.2. By contrast, raising $CO_2$ levels increases ECS in the Eocene simulations to values above 6°C (Fig. 2). This increase in ECS with increasing temperature results in accurate simulation of Eocene temperatures at $CO_2$ concen-

trations that agree with proxy estimates (Fig. 2B). The increased ECS under high $CO_2$ in CESM1.2 is due to the improved representation of clouds in the CAM5 atmospheric model, which drives a strong low-cloud positive feedback under high $CO_2$ (*19*)—a finding in agreement with the emerging recognition that cloud feedbacks are a key component of warm climates (*20*, *21*). That CESM1.2 simulates Eocene proxy temperatures within the bounds of proxy $CO_2$ estimates provides support for its cloud physics and increases our confidence in the model's state-dependent ECS. The Geophysical Fluid Dynamics Laboratory (GFDL) CM2.1 model can also simulate the large-scale features of Eocene proxy temperatures (*22*), likewise suggesting that its ECS is reasonable. Conversely, CESM2 (the newest version of the NCAR model) estimates Eocene temperatures that exceed the upper bound of proxy constraints at low $CO_2$ levels (*23*), suggesting that its modern ECS of 5.3°C is too high. A little more than a third of the newest-generation CMIP6 models have an ECS higher than 4.5°C (*15*). From historical observations alone, it is very challenging to assess the plausibility of these higher ECS values. As the Eocene example highlights, warm paleoclimates are key in this respect.

The early Eocene provides an important constraint on model ECS but samples a single

high-$CO_2$ climate state. Given the dependence of ECS on the background climate state, other past climates are critical to constraining ECS and relevant physics under both lower (e.g., LGM, Pliocene) and higher (e.g., Eocene, Cretaceous) background $CO_2$ levels. One concern about using past climates as model targets is that the forcings, especially aerosol and non-$CO_2$ greenhouse gas concentrations, are uncertain and increasingly so in the distant past. Although important, it is worth noting that these forcings are secondary to $CO_2$ [e.g., (*24*)] and, for extreme climates like the Eocene and Cretaceous, may largely fall within the climate proxy uncertainties. Moreover, this concern can be mitigated by examining model responses to the potential range of underconstrained forcings and, as is increasingly done, by incorporating biogeochemical cycles and the simulation of aerosol production and transport into the models.

## Paleoclimate perspectives on the stability of the cryosphere

Future projections of sea level rise have large uncertainties, mainly owing to unknowns surrounding the stability and threshold behavior of ice sheets (*25*). The paleoclimate record furnishes true "out-of-sample" tests for understanding the sensitivity of the cryosphere to warming that can lower these uncertainties. The past few years have seen a number of advances on both data and climate modeling fronts to understand past changes in ice sheets and connect these to the future. Advances in the generation and interpretation of proxy indicators of ice sheet size, shape, and extent (*26–28*) are helping to refine our understanding of cryosphere dynamics in warmer climates. Improvements in modeling the effects of dynamic topography and glacial isostatic adjustment are continually reducing uncertainties associated with estimates of past global sea level (*29*, *30*), providing more accurate benchmarks for model simulations (*31*).

Paleoclimates also provide critical insights into processes that drive destabilization of ice sheets. Of particular relevance for future projections is assessing the likelihood of marine ice-cliff instability (MICI), a rapid collapse of coastal ice cliffs that follows the disintegration of an ice shelf, which has the potential to contribute to substantial sea level rise by the end of the 21st century (*32*, *33*). The record of sea level change from past warm climates offers a way to test this hypothesis. Recent work has focused on the Pliocene, given that $CO_2$ concentrations during this time were similar to current anthropogenic levels (Fig. 1). A new reconstruction of global mean sea level during the mid-Pliocene warm period indicates a rise of ~17 m, implying near-to-complete loss of Greenland and the West Antarctic Ice Sheet with some additional contribution from East

Antarctica (*34*). Although this represents an outstanding loss of ice, MICI is not necessarily needed to explain it (*33, 34*). However, simulated changes in sea level are highly dependent on each model's treatment of ice sheet stability (*35*), and paleoclimate investigations of warmer climates, such as the early Pliocene and the Miocene, indicate larger magnitudes of ice loss, thermal expansion, and consequent sea level rise (*34, 36*). Moving forward, refining our understanding of threshold behavior in ice sheets, and thus improving projections of future sea level rise, will require a synergistic approach that leverages paleoclimate estimates from multiple warm climates alongside solid Earth, ice sheet, and climate modeling (*31*).

### Regional and seasonal information from past climates

Future warming will shift regional and seasonal patterns of rainfall and temperature, with dramatic consequences for human society (*37, 38*). Regional changes in the land surface (reduced snow cover, melting permafrost, greening, desertification) can further trigger biogeochemical feedbacks that could dampen or amplify initial radiative forcing, with implications for climate sensitivity (*39*). Unfortunately, climate models disagree about the direction and magnitude of future regional rainfall change (*40*). Improving future predictions of regional climate requires separating internal variability in the climate system (i.e., interannual-centennial oscillations) from externally forced changes (i.e., from greenhouse gases or aerosols). Regional and seasonal paleoclimate data are critical in this respect, because they provide long, continuous estimates of the natural range of variation, augmenting the relatively short observational record (*41, 42*).

Subannually resolved paleobiological and sedimentary archives, made more accessible by recent advances in geochemical techniques, allow for the study of seasonal-scale variations in both temperature and hydroclimate. For example, oxygen isotope ($\delta^{18}$O) measurements of fossil bivalves can be used to gain insights into the drivers of seasonal variability during the Eocene greenhouse climate (*43, 44*) (Fig. 3A). Because individual planktic foraminifera live for about a month, analyses of single shells yield subannual sea-surface temperature (SST) data from ancient climates (*45*). This can be leveraged to reveal past changes in key seasonal phenomena such as the El Niño–Southern Oscillation (ENSO) (*46*) (Fig. 3). Proxy data can even provide records of changes in the frequency or intensity of extreme events like hurricanes (*47*).

Reconstructions of hydroclimate are considerably more challenging than temperature, because proxy signals tend to be more complex; however, even basic directional information (wetter versus drier) can be used to test
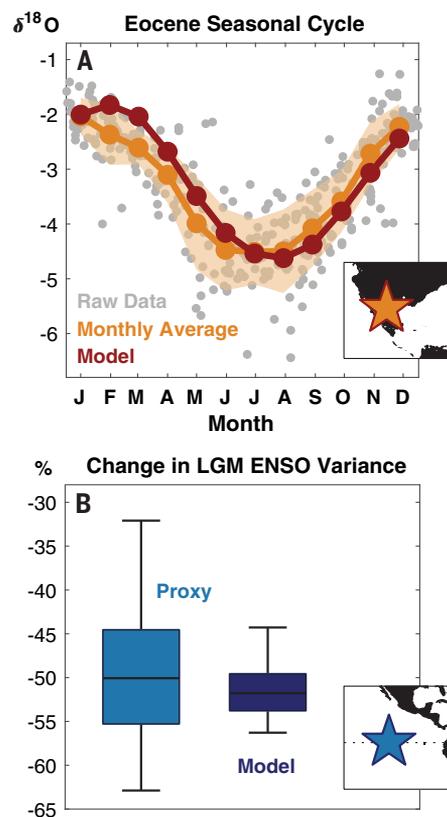


**Fig. 3. Examples of seasonal and interannual paleoclimate data and comparison to models.**
(**A**) Seasonally resolved $\delta^{18}$O carbonate from the shells of a fossil bivalve, *Venericardia hatcheplata*, from the early Eocene Hatchetigbee Formation (orange star in inset) (*43, 44*). Monthly averaged data (orange, with 1$\sigma$ uncertainty bounds) are compared with predicted $\delta^{18}$O-carbonate seasonality at the same grid point from an isotope-enabled Eocene model simulation (*19*) (red) [using modeled $\delta^{18}$O of seawater and SST, and the calibration of (*110*)]. (**B**) Mg/Ca measurements of individual planktic foraminifera *Trilobatus sacculifer* from the eastern equatorial Pacific (blue star in inset) provide proxy evidence of a reduction in ENSO variability during the LGM relative to preindustrial conditions (*46*) (light blue). The magnitude of reduction agrees with simulations using CESM1.2 (dark blue) (*111*). Error bars represent the 95% confidence interval.

spatial patterns in models [e.g., (*48*)]. Past warm climates allow us to test the extent to which the thermodynamic "wet-gets-wetter, dry-gets-drier" response broadly holds with warming (*49*) or if dynamical changes, such as shifts in the Hadley or Walker cells, play more of a key role in the regional water cycle response to changes in surface temperature gradients (*48, 50*).

Comparisons of proxies and models can also be used to identify the processes that are critical for accurate simulation of regional shifts in the water cycle, where local moisture

and energy budgets exert an important control (*51*). The processes that drive these budgets—i.e., land surface properties and clouds—must be parameterized in global climate models and are often poorly understood yet have huge consequences for predicted patterns in humidity and rainfall (*52–55*). Past changes in Earth's boundary conditions offer a much broader set of scenarios where observations can be used to evaluate the performance of parameterization schemes. In particular, paleoclimates spanning the last glacial cycle have helped us better understand the role of land-atmosphere feedbacks in determining hydroclimatic response. Analyses of LGM proxies for SST and water balance in Southeast Asia suggest a direct relationship between convective parameterization and model skill at capturing regional hydroclimate (*48, 56*). Studies of the mid-Holocene "Green Sahara" highlight the importance of vegetation and dust feedbacks in accurately simulating the response of the west African monsoon to radiative forcing (*57, 58*). These examples demonstrate the value of hydroclimate proxy-model comparison even if the past climate state is not a direct analog for future warming.

Studies of past warm climates have the potential to provide even more insights into the behavior of regional climate in a warming world. Future model projections broadly simulate a pattern of subtropical drying, whereas the deep tropics and high latitudes get wetter (*40*). Recently, however, researchers have argued that subtropical drying is transient and might not persist in equilibrium with higher radiative forcing (*59, 60*). Indeed, several paleoclimatic intervals (*61, 62*) suggest that a warmer world could feature a different pattern, with wetter conditions in both the subtropics and high latitudes (*50*). This pattern is especially evident in western North America, where widespread Pliocene lake deposits suggest much wetter conditions (*63*). This evidence stands in stark contrast to future projections for this region, which overwhelmingly predict drier conditions and more intense droughts (*64*), and suggests that paleoclimates can provide vital constraints on the response of arid lands to higher $CO_2$ concentrations.

### Climatic aberrations

Among the most important discoveries in paleoclimatology is the occurrence of climatic "aberrations"—extraordinary transient departures from a background climate state. Such events are distinguished by radical changes in temperature, precipitation patterns, and ocean circulation that often leave distinctive marks in the geological record, like the pervasive black shales of the mid-Cretaceous Ocean Anoxic Events (*65*). An aberration typically occurs in response to a short-lived perturbation to the climate system, such as a sudden release of
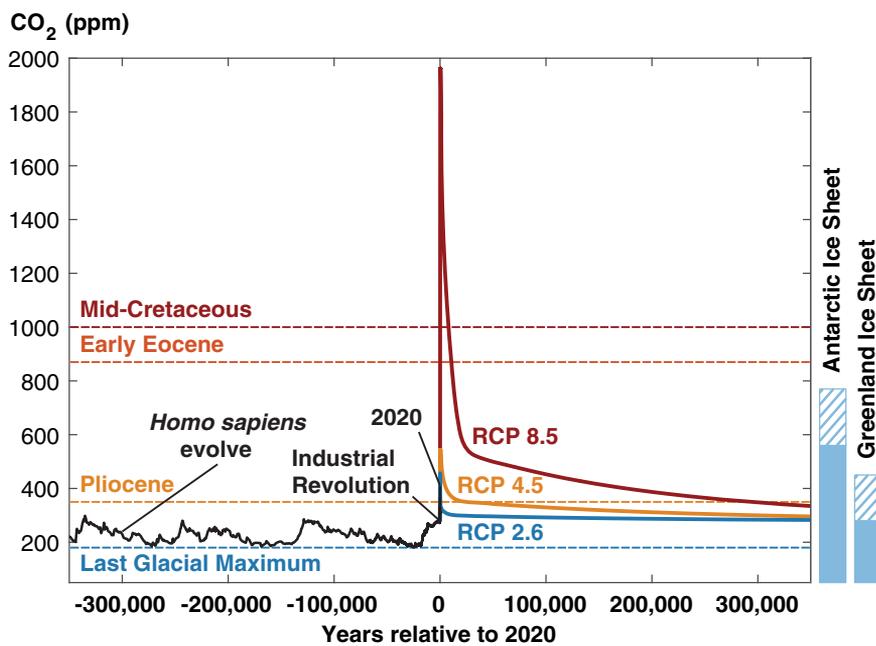
**Fig. 4. The anthropogenic climate aberration.** The black line shows $CO_2$ measured in ice cores for the past 350,000 years (*9*). Solid colored lines show future $CO_2$ concentrations for the IPCC Fifth Assessment Report RCPs, run out to 350,000 years in the future with the cGENIE model. Dashed lines indicate average $CO_2$ for key time periods in the geologic past. Bars at the right indicate $CO_2$ concentrations under which there are well-developed ice sheets (solid areas) and intermittent ice sheets (hatched areas), based on geologic evidence and ice sheet modeling (*112*).

greenhouse gases (e.g., from volcanoes, methane clathrates, or terrestrial organic deposits). Aberrations need not be "abrupt" in the sense that the rate of climate change must exceed the rate of forcing, and they can potentially last for a long time [for example, the Sturtian Snowball Earth lasted 55 million years (*66*)]. They are instructive because they provide information on extreme climate states and the ability of the Earth system to recover from such states.

One of the most notable aberrations in the paleoclimate record, the Paleocene-Eocene Thermal Maximum (PETM), may foreshadow future changes that Earth will experience owing to anthropogenic emissions. The PETM, which occurred 56 million years ago, was triggered by rapid emission of greenhouse gases; proxy and model estimates suggest that $CO_2$ doubled or even tripled from a background state of ~900 ppm (*67–69*) in less than 5000 years (*70, 71*). In response, global temperatures spiked by 4° to 6°C (*72*). The surface ocean rapidly acidified (*68, 73*) and seafloor carbonates dissolved (*74*), resulting in dramatic biogeographic range shifts in plankton and the largest extinction in deep-sea calcifying benthic foraminifera ever observed (*75*). Precipitation patterns changed dramatically, with much more rain falling at the high latitudes (*76*). It took Earth ~100,000 years to recover from this perturbation (*68, 77*).

Although the PETM stands out starkly in the geologic record, the rate of $CO_2$ release was still 4 to 10 times slower than current anthropogenic emissions (*71, 78*). Indeed, the geological record leaves no doubt that our current rate of global warming, driven by anomalous (anthropogenic) forcing, is an exceptional aberration—the rate and magnitude of change far exceed the typical multi–thousand year variability that preceded it (Fig. 4). In the past 100 million years, $CO_2$ has ranged from maximum values in the mid-Cretaceous to minimum levels at the LGM (Fig. 1). Going forward, we are on pace to experience an equivalent magnitude of change in atmospheric $CO_2$ concentrations, in reverse, over a period of time that is more than 10,000 times shorter (Fig. 4). In just over 150 years, we have already raised $CO_2$ concentrations (now at 410 ppm) to Pliocene levels (Fig. 4). Under a middle-of-the-road emissions scenario such as SSP2-4.5 (or the CMIP5 equivalent, RCP4.5), $CO_2$ will approach 600 ppm by year 2100, and if we follow the high-emissions SSP5-8.5 (or RCP8.5), $CO_2$ will rise beyond mid-Cretaceous concentrations (around 1000 ppm) by year 2100 (Fig. 4). In comparison, over the past 800,000 years of geologic history, $CO_2$ only varied between 180 and 280 ppm (*9*) (see also Fig. 4).

How long will it take for Earth to neutralize anthropogenic $CO_2$ and return to preindustrial levels? Earth has the ability to recover from a rapid increase in atmospheric $CO_2$ concentration—the PETM is a textbook example of this process. Indeed, in every case of past $CO_2$ perturbation, the Earth system has compensated to avoid a runaway greenhouse or a permanent icehouse. Yet the natural recovery from aberrations takes place on geological, not anthropogenic, time scales (Fig. 4). Some of the processes that remove $CO_2$ from the atmosphere occur on relatively short (100 to 1000 year) time scales (e.g., ocean uptake), but others take tens to hundreds of thousands of years (e.g., weathering of silicate rocks) (*79*). Using the intermediate-complexity Earth system model cGENIE, we can estimate how long the recovery process takes under different future forcing scenarios. Under an aggressive mitigation scenario [Representative Concentration Pathway (RCP) 2.6], $CO_2$ concentrations remain at Pliocene-like concentrations (>350 ppm) through the year 2350, but it still takes hundreds of thousands of years for concentrations to return to preindustrial levels (Fig. 4). Under a middle-of-the-road scenario (RCP 4.5), $CO_2$ peaks around 550 ppm and remains above Pliocene levels for 30,000 years. Under a worst-case scenario (RCP 8.5), atmospheric $CO_2$ will remain at mid-Cretaceous (>1000 ppm) concentrations for 5000 years, at Eocene concentrations (~850 ppm) for 10,000 years, and at Pliocene concentrations (>350 ppm) for 300,000 years (Fig. 4). It will be at least 500,000 years, a duration equivalent to 20,000 human generations, before atmospheric $CO_2$ fully returns to preindustrial levels. Our planet will recover, but for humans, and the organisms with which we share this planet, the changes in climate will appear to be a permanent state shift.

**Bridging the gap between paleoclimate data and models**

Climate models provide direct estimates of quantities like temperatures, wind speed, and precipitation. By contrast, paleoclimate information is indirect, filtered through a proxy—a physical, chemical, and/or biological entity that responds to climate—such as foraminifera, algae, or the chemical composition of sediments. Proxies are imperfect recorders of climate; they have inherent uncertainties associated with, for example, biological processes and preservation. Thus, although proxy data can be transformed into climate variables for direct comparison with models using regression, transfer functions, and assumptions, if these structural uncertainties are not accounted for, they can lead to unclear or erroneous interpretations. This creates a "language barrier" between model output and proxy data that has limited the use of paleoclimate information to infer past climate states and evaluate climate models. Three key innovations are now breaking down this barrier, allowing paleoclimate
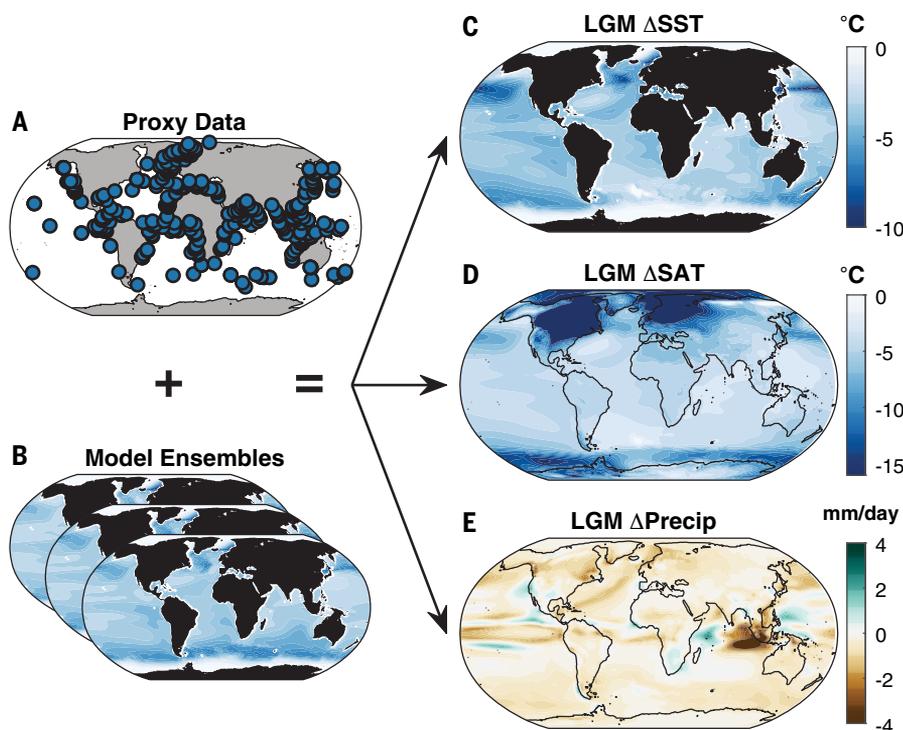
**Fig. 5. An example of paleoclimate data assimilation.** (**A** to **E**) Marine SST proxy data from the LGM and the Late Holocene (A) are combined with an ensemble of model simulations (B) that contain multiple climatic variables. The results [(C) to (E); LGM–Late Holocene differences for SST (ΔSST), surface air temperature (ΔSAT), and mean annual precipitation (ΔPrecip)] include all the variables in the model prior, which are influenced by the assimilated SST proxy data. Proxy data, model fields, and assimilated results are from (*91*).

climate states, model-proxy agreement, and model evaluation (*88*). These have helped to transform proxy-model comparisons from qualitative statements ("they look similar") to quantitative statements ("there is a 90% probability that the data and the model agree").

A final component of the "language barrier" is the fact that proxy data are sparse in both space and time, because they are fundamentally dependent on the presence and preservation of their archives. Yet proxy data are real-world estimates of the "true" climate state. By contrast, climate model information is spatially and temporally continuous and physically self-consistent—but is only a best "guess" at what did or what will happen. One solution to bridge these fundamentally different pieces of information is to formally combine them in a statistical framework and thus leverage their respective strengths. Reduced space methods, commonly used to produce historical reconstructions of climate, can be used to infill missing data and produce maps of paleoclimate states (*88*, *89*). Recently, weather-based data assimilation techniques have been adapted for paleoclimate applications (*90*). The resulting products are spatially complete reconstructions of multiple climate variables that represent a balance between the proxy information and the physics and covariance structure of the climate model. This allows local paleoclimate proxy information to be used to infer global metrics of climate, such as global mean air temperature (*91*). It also allows for the recovery of climatic variables that are consistent with the proxy information but for which we have no direct proxies, such as cloud cover, wind patterns, or precipitation (Fig. 5).

In sum, the disintegration of the model-proxy language barrier has narrowed uncertainties in proxy interpretation. Recent studies have been able to use proxy data to infer key climatic processes and evaluate models across multiple time periods, including the LGM (*91*), the Pliocene (*88*), and the Eocene (*19*, *22*). This opens the door for explicit use of paleoclimates to assess and improve model physics.

**Moving forward**

Past climates will continue to provide insights into the range, rate, and dynamics of climate change. Over the past decade, we have witnessed breakthroughs in proxy development and refinement as well as the generation of many new high-resolution marine and terrestrial paleoclimate records. In addition to continued advances, the collection of additional temperature and $CO_2$ proxy records at higher resolution will be paramount for developing better estimates of climate sensitivity. Future proxy collection efforts should also focus on hydroclimate proxies, given the large spread in model projections (*40*). These

information to directly constrain model performance: (i) the inclusion of chemical tracers relevant to proxies directly in Earth system models; (ii) the creation of robust proxy system models that explicitly encode processes, uncertainties, and multivariate sensitivities; and (iii) the development of statistical methods to formally combine proxy and model data.

As far as chemical tracers are concerned, the single most important advance has been the increasingly routine incorporation of water isotopes in model simulations. The stable isotopes of water—$\delta^{18}O$ and hydrogen isotopes ($\delta D$)—and their incorporation into natural archives are the foundation of modern paleoclimatology (*80*). A large number of paleoclimate proxies record water isotopes, including foraminifera, stalagmites, leaf waxes, soil carbonates, and ice cores. Water isotope composition, however, reflects multiple processes, including changes in temperature, moisture source, evaporation, precipitation, and convection. Including water isotopes in models generates simulated isotope fields that are consistent with the model's treatment of these processes, eliminating the need to independently conjecture how these various factors may have influenced the proxy data. This creates an "apples to apples" comparison between proxy information and model output that can be used to evaluate model

performance and diagnose climatic processes [e.g., (*81*)]. For example, using the water isotope–enabled CESM1.2 (iCESM) (*82*), it is possible to directly compare carbonate $\delta^{18}O$ data from Eocene fossil bivalves to model-simulated $\delta^{18}O$ (*43*, *44*, *83*) (Fig. 3A). The model predicts a roughly 3 per mil annual range in carbonate $\delta^{18}O$, in good agreement with observed proxy data (Fig. 3A). The match with the $\delta^{18}O$ data builds confidence that the model can correctly simulate climatology in this location and allows us to deconvolve the contribution of SST and $\delta^{18}O$ of seawater. The site-specific seasonality in SST is 8° to 10°C, whereas the $\delta^{18}O$ of seawater has a seasonal range of 0.6 to 0.8. This indicates that temperature is primarily responsible for the large seasonal range in carbonate $\delta^{18}O$ during this greenhouse climate state.

One aspect of paleoclimate information that has traditionally limited its use in model evaluation is an inability to precisely quantify uncertainties surrounding the proxies. However, in the past decade, increasingly detailed proxy system models (*84*) have been developed to address this issue [e.g., (*85*–*87*)]. Many of these use Bayesian inference to quantify uncertainties in the sensitivity of proxies to environmental parameters, which can then be used for probabilistic assessments of past

reconstructions will help us refine our understanding of the response of atmospheric circulation and rainfall to climate change.

On the modeling side, the inclusion of chemical tracers, such as water and carbon isotopes, within many of the newly developed CMIP6 (*92*) models offers more robust means of data-model comparison. With these new model tools, we anticipate the rapid development and improvement of data-model synthesis products (*90*, *91*) and more focused proxy collection efforts to help reduce model uncertainties. In addition, evaluating CMIP6 models using both the historical and paleoclimate record will result in a more comprehensive and robust approach to understanding the climate system (*93*). We recommend widespread adoption of this practice, so that model ECS and other emergent properties are constrained by paleoclimate data as well as observations. We suggest that weighting or ranking models that perform well over multiple past climate states is a crucial way to constrain the response of the model to changing background conditions and the validity of simulated climate changes under various emissions scenarios. In general, climate models should be able to accurately simulate multiple extreme paleoclimate states—warm and cold—before being trusted for future climate projection.

Despite promising CMIP6 model advances, maintaining a variety of models with different levels of complexity is important. Not all climate questions require high levels of model complexity, and sometimes complexity is so great that interpretation becomes limited (*94*). In paleoclimatology, complexity can also lead to prohibitive computational expense. Maintained support for lower-resolution and variable-resolution configurations is vital for better interpreting model results and performing long, transient simulations that can address fundamental aspects of paleoclimatology such as glacial cycles and carbon cycle changes.

Looking ahead, there are many outstanding process-based uncertainties associated with future climate change that paleoclimatology can help constrain. For example, paleobotanical records can inform plant physiological responses to changes in $CO_2$ (*95*), which remain highly uncertain (*96*) but important for quantifying evapotranspirative and surface runoff fluxes. Similarly, past vegetation reconstructions can assess dynamic vegetation models and simulated changes in the hydrologic cycle through time (*97*). Moreover, additional quantitative reconstructions of hydroclimate, in combination with better constraints on plant physiological functioning in the past, will help refine our understanding of the regional water cycle and its dependence on local energy fluxes and large-scale circulation.

New geochemical techniques will also refine our understanding of the Earth system. Devel-opment of radiation (*98*), biogenic aerosol (*99*), and dust (*100*) records has the potential to help constrain past aerosol and cloud radiative effects, which are arguably the most uncertain component of Earth system models (*101*). In addition, new geochemical tracers for methane cycling (*102*) and upwelling, which is important for $N_2O$ production (*103*), will provide insights into trace greenhouse gases during past climate states. The combination of these new techniques will allow the paleoclimate community to better quantify biogeochemical feedbacks and climate sensitivity to greenhouse gas forcings across a range of climate states—and ultimately improve climate forecasts for the coming decades to millennia.

The paleoclimate record is the basis for how we understand the potential range and rate of climate change. Past climates represent the only target for climate model predictions at $CO_2$ concentrations outside of the narrow historical range and, for this reason, are vital tools for evaluating the newest generation of Earth system models. The study of past climates continues to reveal key insights into Earth's response to increased concentrations of greenhouse gases. Innovations in Earth system models, geochemical techniques, and statistical methods further allow for a more direct connection from the past to the future—worlds for which the preindustrial and industrial climate states provide limited guidance. The future of paleoclimatology is to incorporate past climate information formally in model evaluation, so that we can better predict and plan for the impacts of anthropogenic climate change.

## REFERENCES AND NOTES

1. R. Hooke, *The Posthumous Works of Robert Hooke... Containing His Cutlerian Lectures, and Other Discourses, Read at the Meetings of the Illustrious Royal Society...Illustrated with Sculptures* (The Royal Society, London, 1705), pp. 342–343.
2. J. Zachos, M. Pagani, L. Sloan, E. Thomas, K. Billups, Trends, rhythms, and aberrations in global climate 65 Ma to present. *Science* 292, 686–693 (2001). doi: 10.1126/science.1059412; pmid: 11326091
3. K. D. Burke et al., Pliocene and Eocene provide best analogs for near-future climates. *Proc. Natl. Acad. Sci. U.S.A.* 115, 13288–13293 (2018). doi: 10.1073/pnas.1809600115; pmid: 30530685
4. G. A. Schmidt et al., Practice and philosophy of climate model tuning across six U.S. modeling centers. *Geosci. Model Dev.* 10, 3207–3223 (2017). doi: 10.5194/gmd-10-3207-2017; pmid: 30595813
5. M. Huber, R. Caballero, The early Eocene equable climate problem revisited. *Clim. Past* 7, 603–633 (2011). doi: 10.5194/cp-7-603-2011
6. V. Lauretano, J. C. Zachos, L. J. Lourens, Orbitally paced carbon and deep-sea temperature changes at the peak of the early Eocene climatic optimum. *Paleoceanogr. Paleoclimatol.* 33, 1050–1065 (2018). doi: 10.1029/2018PA003422
7. D. A. Hodell, J. E. T. Channell, J. H. Curtis, O. E. Romero, U. Röhl, Onset of "Hudson Strait" Heinrich events in the eastern North Atlantic at the end of the middle Pleistocene transition (∼640 ka)? *Paleoceanography* 23, PA4218 (2008). doi: 10.1029/2008PA001591
8. J. C. Walker, P. Hays, J. F. Kasting, A negative feedback mechanism for the long-term stabilization of Earth's surface temperature. *J. Geophys. Res. Oceans* 86, 9776–9782 (1981). doi: 10.1029/JC086iC10p09776

9. B. Bereiter et al., Revision of the EPICA Dome C $CO_2$ record from 800 to 600 kyr before present. *Geophys. Res. Lett.* 42, 542–549 (2015). doi: 10.1002/2014GL061957
10. W. Konrad, D. L. Royer, P. J. Franks, A. Roth-Nebelsick, Quantitative critique of leaf-based paleo-$CO_2$ proxies: Consequences for their reliability and applicability. *Geol. J.* 10.1002/gj.3807 (2020). doi: 10.1002/gj.3807
11. M. J. Henehan et al., Rapid ocean acidification and protracted Earth system recovery followed the end-Cretaceous Chicxulub impact. *Proc. Natl. Acad. Sci. U.S.A.* 116, 22500–22504 (2019). doi: 10.1073/pnas.1905989116; pmid: 31636204
12. Y. Goddéris, Y. Donnadieu, G. Le Hir, V. Lefebvre, E. Nardin, The role of palaeogeography in the Phanerozoic history of atmospheric $CO_2$ and climate. *Earth Sci. Rev.* 128, 122–138 (2014). doi: 10.1016/j.earscirev.2013.11.004
13. J. G. Charney et al., "Carbon dioxide and climate: A scientific assessment" (Consensus Study Report, National Academy of Sciences, The National Academies Press, Washington, DC, 1979).
14. C. Hope, The $10 trillion value of better information about the transient climate response. *Philos. Trans. A Math. Phys. Eng. Sci.* 373, 20140429 (2015). doi: 10.1098/rsta.2014.0429; pmid: 26438286
15. M. D. Zelinka et al., Causes of higher climate sensitivity in CMIP6 models. *Geophys. Res. Lett.* 47, e2019GL085782 (2020). doi: 10.1029/2019GL085782
16. R. Caballero, M. Huber, State-dependent climate sensitivity in past warm climates and its implications for future climate projections. *Proc. Natl. Acad. Sci. U.S.A.* 110, 14162–14167 (2013). doi: 10.1073/pnas.1303365110; pmid: 23918397
17. K. Meraner, T. Mauritsen, A. Voigt, Robust increase in equilibrium climate sensitivity under global warming. *Geophys. Res. Lett.* 40, 5944–5948 (2013). doi: 10.1002/2013GL058118
18. T. Friedrich, A. Timmermann, M. Tigchelaar, O. Elison Timm, A. Ganopolski, Nonlinear climate sensitivity and its implications for future greenhouse warming. *Sci. Adv.* 2, e1501923 (2016). doi: 10.1126/sciadv.1501923; pmid: 28861462
19. J. Zhu, C. J. Poulsen, J. E. Tierney, Simulation of Eocene extreme warmth and high climate sensitivity through cloud feedbacks. *Sci. Adv.* 5, eaax1874 (2019). doi: 10.1126/sciadv.aax1874; pmid: 31555736
20. C. S. Bretherton, Insights into low-latitude cloud feedbacks from high-resolution models. *Philos. Trans. A Math. Phys. Eng. Sci.* 373, 20140415 (2015). doi: 10.1098/rsta.2014.0415; pmid: 26438280
21. T. Schneider, C. M. Kaul, K. G. Pressel, Possible climate transitions from breakup of stratocumulus decks under greenhouse warming. *Nat. Geosci.* 12, 163–167 (2019). doi: 10.1038/s41561-019-0310-1
22. D. J. Lunt et al., DeepMIP: Model intercomparison of early Eocene climatic optimum (EECO) large-scale climate features and comparison with proxy data. *Clim. Past Discuss.* 10.5194/cp-2019-149 (2019). doi: 10.5194/cp-2019-149
23. J. Zhu, C. J. Poulsen, B. L. Otto-Bliesner, High climate sensitivity in CMIP6 model not supported by paleoclimate. *Nat. Clim. Chang.* 10, 378–379 (2020). doi: 10.1038/s41558-020-0764-6
24. D. J. Beerling, A. Fox, D. S. Stevenson, P. J. Valdes, Enhanced chemistry-climate feedbacks in past greenhouse worlds. *Proc. Natl. Acad. Sci. U.S.A.* 108, 9770–9775 (2011). doi: 10.1073/pnas.1102409108; pmid: 21628580
25. J. L. Bamber, M. Oppenheimer, R. E. Kopp, W. P. Aspinall, R. M. Cooke, Ice sheet contributions to future sea-level rise from structured expert judgment. *Proc. Natl. Acad. Sci. U.S.A.* 116, 11195–11200 (2019). doi: 10.1073/pnas.1817205116; pmid: 31110015
26. M. G. Wise, A. J. Dowdeswell, M. Jakobsson, R. D. Larter, Evidence of marine ice-cliff instability in Pine Island Bay from iceberg-keel plough marks. *Nature* 550, 506–510 (2017). doi: 10.1038/nature24458; pmid: 28972274
27. S. P. S. Gulick et al., Initiation and long-term instability of the East Antarctic Ice Sheet. *Nature* 552, 225–229 (2017). doi: 10.1038/nature25026; pmid: 29239353
28. A. Rovere et al., The analysis of Last Interglacial (MIS 5e) relative sea-level indicators: Reconstructing sea-level in a warmer world. *Earth Sci. Rev.* 159, 404–427 (2016). doi: 10.1016/j.earscirev.2016.06.006
29. A. Dutton et al., Sea-level rise due to polar ice-sheet mass loss during past warm periods. *Science* 349, aaa4019 (2015). doi: 10.1126/science.aaa4019; pmid: 26160951

30. J. Austermann et al., The impact of dynamic topography change on antarctic ice sheet stability during the mid-pliocene warm period. Geology 43, 927–930 (2015). doi: 10.1130/G36988.1

31. B. P. Horton, R. Kopp, A. Dutton, T. Shaw, Geological records of past sea-level changes as constraints for future projections. PAGES Magazine 27, 28–29 (2019).

32. R. M. DeConto, D. Pollard, Contribution of Antarctica to past and future sea-level rise. Nature 531, 591–597 (2016). doi: 10.1038/nature17145; pmid: 27029274

33. T. L. Edwards et al., Revisiting Antarctic ice loss due to marine ice-cliff instability. Nature 566, 58–64 (2019). doi: 10.1038/s41586-019-0901-4; pmid: 30728522

34. O. A. Dumitru et al., Constraints on global mean sea level during Pliocene warmth. Nature 574, 233–236 (2019). doi: 10.1038/s41586-019-1543-2; pmid: 31471591

35. A. M. Dolan, B. de Boer, J. Bernales, D. J. Hill, A. M. Haywood, High climate model dependency of Pliocene Antarctic ice-sheet predictions. Nat. Commun. 9, 2799 (2018). doi: 10.1038/s41467-018-05179-4; pmid: 30022077

36. E. Gasson, R. M. DeConto, D. Pollard, R. H. Levy, Dynamic Antarctic ice sheet during the early to mid-Miocene. Proc. Natl. Acad. Sci. U.S.A. 113, 3459–3464 (2016). doi: 10.1073/pnas.1516130113; pmid: 26903645

37. A. J. McMichael, R. E. Woodruff, S. Hales, Climate change and human health: Present and future risks. Lancet 367, 859–869 (2006). doi: 10.1016/S0140-6736(06)68079-3; pmid: 16530580

38. R. L. Wilby, A review of climate change impacts on the built environment. Built Environ. 33, 31–45 (2007). doi: 10.2148/benv.33.1.31

39. A. Arneth et al., Terrestrial biogeochemical feedbacks in the climate system. Nat. Geosci. 3, 525–532 (2010). doi: 10.1038/ngeo905

40. R. Knutti, J. Sedláček, Robustness and uncertainties in the new CMIP5 climate model projections. Nat. Clim. Chang. 3, 369–373 (2013). doi: 10.1038/nclimate1716

41. C. Deser, A. Phillips, V. Bourdette, H. Teng, Uncertainty in climate change projections: The role of internal variability. Clim. Dyn. 38, 527–546 (2012). doi: 10.1007/s00382-010-0977-x

42. PAGES Hydro2k Consortium, Comparing proxy and model estimates of hydroclimate variability and change over the Common Era. Clim. Past 13, 1851–1900 (2017). doi: 10.5194/cp-13-1851-2017

43. L. C. Ivany et al., Intra-annual isotopic variation in Venericardia bivalves: Implications for early Eocene temperature, seasonality, and salinity on the US Gulf Coast. J. Sediment. Res. 74, 7–19 (2004). doi: 10.1306/052803740007

44. C. R. Keating-Bitonti, L. C. Ivany, H. P. Affek, P. Douglas, S. D. Samson, Warm, not super-hot, temperatures in the early Eocene subtropics. Geology 39, 771–774 (2011). doi: 10.1130/G32054.1

45. K. Thirumalai, P. N. DiNezio, J. E. Tierney, M. Puy, M. Mohtadi, An El Niño mode in the glacial Indian Ocean? Paleoceanogr. Paleoclimatol. 34, 1316–1327 (2019). doi: 10.1029/2019PA003669

46. H. L. Ford, A. C. Ravelo, P. J. Polissar, Reduced El Niño-Southern Oscillation during the Last Glacial Maximum. Science 347, 255–258 (2015). doi: 10.1126/science.1258437; pmid: 25593181

47. A. Frappier, T. Knutson, K.-B. Liu, K. Emanuel, Perspective: Coordinating paleoclimate research on tropical cyclones with hurricane-climate theory and modelling. Tellus A Dyn. Meteorol. Oceanogr. 59, 529–537 (2007). doi: 10.1111/j.1600-0870.2007.00250.x

48. P. N. DiNezio, J. E. Tierney, The effect of sea level on glacial Indo-Pacific climate. Nat. Geosci. 6, 485–491 (2013). doi: 10.1038/ngeo1823

49. I. M. Held, B. J. Soden, Robust responses of the hydrological cycle to global warming. J. Clim. 19, 5686–5699 (2006). doi: 10.1175/JCLI3990.1

50. N. J. Burls, A. V. Fedorov, Wetter subtropics in a warmer world: Contrasting past and future hydrological cycles. Proc. Natl. Acad. Sci. U.S.A. 114, 12888–12893 (2017). doi: 10.1073/pnas.1703421114; pmid: 29158397

51. C. Muller, P. O'Gorman, An energetic perspective on the regional response of precipitation to climate change. Nat. Clim. Chang. 1, 266–271 (2011). doi: 10.1038/nclimate1169

52. M. P. Byrne, P. A. O'Gorman, The response of precipitation minus evapotranspiration to climate warming: Why the "wet-get-wetter, dry-get-drier" scaling does not hold over land. J. Clim. 28, 8078–8092 (2015). doi: 10.1175/JCLI-D-15-0369.1

53. M. P. Byrne, P. A. O'Gorman, Trends in continental temperature and humidity directly linked to ocean warming. Proc. Natl. Acad. Sci. U.S.A. 115, 4863–4868 (2018). doi: 10.1073/pnas.1722312115; pmid: 29686095

54. E. Erfani, N. J. Burls, The strength of low-cloud feedbacks and tropical climate: A CESM sensitivity study. J. Clim. 32, 2497–2516 (2019). doi: 10.1175/JCLI-D-18-0551.1

55. S. Bony et al., Clouds, circulation and climate sensitivity. Nat. Geosci. 8, 261–268 (2015). doi: 10.1038/ngeo2398

56. P. N. DiNezio et al., Glacial changes in tropical climate amplified by the Indian Ocean. Sci. Adv. 4, eaat9658 (2018). doi: 10.1126/sciadv.aat9658; pmid: 30547084

57. J. E. Tierney, F. S. Pausata, P. B. deMenocal, Rainfall regimes of the Green Sahara. Sci. Adv. 3, e1601503 (2017). doi: 10.1126/sciadv.1601503; pmid: 28116352

58. W. R. Boos, R. L. Korty, Regional energy budget control of the intertropical convergence zone and application to mid-holocene rainfall. Nat. Geosci. 9, 892–897 (2016). doi: 10.1038/ngeo2833

59. P. Ceppi, G. Zappa, T. G. Shepherd, J. M. Gregory, Fast and slow components of the extratropical atmospheric circulation response to $CO_2$ forcing. J. Clim. 31, 1091–1105 (2018). doi: 10.1175/JCLI-D-17-0323.1

60. J. K. Sniderman et al., Southern Hemisphere subtropical drying as a transient response to warming. Nat. Clim. Chang. 9, 232–236 (2019). doi: 10.1038/s41558-019-0397-9

61. R. Feng, C. J. Poulsen, M. Werner, Tropical circulation intensification and tectonic extension recorded by Neogene terrestrial $\delta^{18}O$ records of the western United States. Geology 44, 971–974 (2016). doi: 10.1130/G38212.1

62. B. Carrapa, M. Clementz, R. Feng, Ecological and hydroclimate responses to strengthening of the Hadley circulation in South America during the Late Miocene cooling. Proc. Natl. Acad. Sci. U.S.A. 116, 9747–9752 (2019). doi: 10.1073/pnas.1810721116; pmid: 31036635

63. D. E. Ibarra et al., Warm and cold wet states in the western United States during the Pliocene–Pleistocene. Geology 46, 355–358 (2018). doi: 10.1130/G39962.1

64. B. I. Cook, T. R. Ault, J. E. Smerdon, Unprecedented 21st century drought risk in the American Southwest and Central Plains. Sci. Adv. 1, e1400082 (2015). doi: 10.1126/sciadv.1400082; pmid: 26601131

65. H. C. Jenkyns, Geochemistry of oceanic anoxic events. Geochem. Geophys. Geosyst. 11, Q03004 (2010). doi: 10.1029/2009GC002788

66. A. D. Rooney et al., Re-Os geochronology and coupled Os-Sr isotope constraints on the Sturtian snowball Earth. Proc. Natl. Acad. Sci. U.S.A. 111, 51–56 (2014). doi: 10.1073/pnas.1317266110; pmid: 24344274

67. R. E. Zeebe, J. C. Zachos, G. R. Dickens, Carbon dioxide forcing alone insufficient to explain Palaeocene–Eocene Thermal Maximum warming. Nat. Geosci. 2, 576–580 (2009). doi: 10.1038/ngeo578

68. M. Gutjahr et al., Very large release of mostly volcanic carbon during the Palaeocene-Eocene Thermal Maximum. Nature 548, 573–577 (2017). doi: 10.1038/nature23646; pmid: 28858305

69. Y. Cui, B. A. Schubert, Atmospheric $pCO_2$ reconstructed across five early Eocene global warming events. Earth Planet. Sci. Lett. 478, 225–233 (2017). doi: 10.1016/j.epsl.2017.08.038

70. S. Kirtland Turner, P. M. Hull, L. R. Kump, A. Ridgwell, A probabilistic assessment of the rapidity of PETM onset. Nat. Commun. 8, 353 (2017). doi: 10.1038/s41467-017-00292-2; pmid: 28842564

71. S. Kirtland Turner, Constraints on the onset duration of the Paleocene-Eocene Thermal Maximum. Philos. Trans. A Math. Phys. Eng. Sci. 376, 20170082 (2018). doi: 10.1098/rsta.2017.0082; pmid: 30177565

72. G. N. Inglis et al., Global mean surface temperature and climate sensitivity of the EECO, PETM and latest Paleocene. Clim. Past Discuss. 10.5194/cp-2019-167 (2020). doi: 10.5194/cp-2019-167

73. D. E. Penman, B. Hönisch, R. E. Zeebe, E. Thomas, J. C. Zachos, Rapid and sustained surface ocean acidification during the Paleocene-Eocene Thermal Maximum. Paleoceanography 29, 357–369 (2014). doi: 10.1002/2014PA002621

74. J. C. Zachos et al., Rapid acidification of the ocean during the Paleocene-Eocene thermal maximum. Science 308, 1611–1615 (2005). doi: 10.1126/science.1109004; pmid: 15947184

75. E. Thomas, N. J. Shackleton, The Paleocene-Eocene benthic foraminiferal extinction and stable isotope anomalies. Geol. Soc. London Spec. Publ. 101, 401–441 (1996). doi: 10.1144/GSL.SP.1996.101.01.20

76. M. J. Carmichael et al., Hydrological and associated biogeochemical consequences of rapid global warming during the Paleocene-Eocene Thermal Maximum. Global Planet. Change 157, 114–138 (2017). doi: 10.1016/j.gloplacha.2017.07.014

77. G. J. Bowen, J. C. Zachos, Rapid carbon sequestration at the termination of the Palaeocene–Eocene Thermal Maximum. Nat. Geosci. 3, 866–869 (2010). doi: 10.1038/ngeo1014

78. R. E. Zeebe, A. Ridgwell, J. C. Zachos, Anthropogenic carbon release rate unprecedented during the past 66 million years. Nat. Geosci. 9, 325–329 (2016). doi: 10.1038/ngeo2681

79. D. Archer, H. Kheshgi, E. Maier-Reimer, Multiple timescales for neutralization of fossil fuel $CO_2$. Geophys. Res. Lett. 24, 405–408 (1997). doi: 10.1029/97GL00168

80. C. Emiliani, Pleistocene temperatures. J. Geol. 63, 538–578 (1955). doi: 10.1086/626295

81. J. Zhou, C. Poulsen, D. Pollard, T. White, Simulation of modern and middle Cretaceous marine $\delta^{18}O$ with an ocean-atmosphere general circulation model. Paleoceanography 23, PA3223 (2008). doi: 10.1029/2008PA001596

82. E. Brady et al., The connected isotopic water cycle in the Community Earth System Model version 1. J. Adv. Model. Earth Syst. 11, 2547–2566 (2019). doi: 10.1029/2019MS001663

83. J. Zhu et al., Simulation of early Eocene water isotopes using an Earth system model and its implication for past climate reconstruction. Earth Planet. Sci. Lett. 537, 116164 (2020). doi: 10.1016/j.epsl.2020.116164

84. M. N. Evans, S. E. Tolwinski-Ward, D. M. Thompson, K. J. Anchukaitis, Applications of proxy system modeling in high resolution paleoclimatology. Quat. Sci. Rev. 76, 16–28 (2013). doi: 10.1016/j.quascirev.2013.05.024

85. S. Tolwinski-Ward, K. Anchukaitis, M. Evans, Bayesian parameter estimation and interpretation for an intermediate model of tree-ring width. Clim. Past 9, 1481–1493 (2013). doi: 10.5194/cp-9-1481-2013

86. C. N. Jex, S. Phipps, A. Baker, C. Bradley, Reducing uncertainty in the climatic interpretations of speleothem $\delta^{18}O$. Geophys. Res. Lett. 40, 2259–2264 (2013). doi: 10.1002/grl.50467

87. J. E. Tierney, M. P. Tingley, BAYSPLINE: A new calibration for the alkenone paleothermometer. Paleoceanogr. Paleoclimatol. 33, 281–301 (2018). doi: 10.1002/2017PA003201

88. J. E. Tierney, A. M. Haywood, R. Feng, T. Bhattacharya, B. L. Otto-Bliesner, Pliocene warmth consistent with greenhouse gas forcing. Geophys. Res. Lett. 46, 9136–9144 (2019). doi: 10.1029/2019GL083802

89. E. C. Gill, B. Rajagopalan, P. Molnar, T. M. Marchitto, Reduced-dimension reconstruction of the equatorial Pacific SST and zonal wind fields over the past 10,000 years using Mg/Ca and alkenone records. Paleoceanography 31, 928–952 (2016). doi: 10.1002/2016PA002948

90. G. J. Hakim et al., The last millennium climate reanalysis project: Framework and first results. J. Geophys. Res. D Atmospheres 121, 6745–6764 (2016). doi: 10.1002/2016JD024751

91. J. E. Tierney et al., Glacial cooling and climate sensitivity revisited. Nature 584, 569–573 (2020). doi: 10.1038/s41586-020-2617-x; pmid: 32848226

92. V. Eyring et al., Overview of the Coupled Model Intercomparison Project Phase 6 (CMIP6) experimental design and organization. Geosci. Model Dev. 9, 1937–1958 (2016). doi: 10.5194/gmd-9-1937-2016

93. G. Schmidt et al., Using palaeo-climate comparisons to constrain future projections in CMIP5. Clim. Past 10, 221–250 (2013). doi: 10.5194/cp-10-221-2014

94. I. M. Held, The gap between simulation and understanding in climate modeling. Bull. Am. Meteorol. Soc. 86, 1609–1614 (2005). doi: 10.1175/BAMS-86-11-1609

95. J. D. White et al., A process-based ecosystem model (Paleo-BGC) to simulate the dynamic response of Late Carboniferous plants to elevated $O_2$ and aridification. Am. J. Sci. 320, 547–598 (2020). doi: 10.2475/09.2020.01

96. G. Damour, T. Simonneau, H. Cochard, L. Urban, An overview of models of stomatal conductance at the leaf level. Plant Cell Environ. 33, 1419–1438 (2010). doi: 10.1111/j.1365-3040.2010.02181.x; pmid: 20545879

97. S. Hoetzel, L. Dupont, E. Schefuß, F. Rommerskirchen, G. Wefer, The role of fire in Miocene to Pliocene C$_4$ grassland and ecosystem evolution. *Nat. Geosci.* **6**, 1027–1030 (2013). doi: 10.1038/ngeo1984

98. B. H. Lomax *et al.*, Plant spore walls as a record of long-term changes in ultraviolet-B radiation. *Nat. Geosci.* **1**, 592–596 (2008). doi: 10.1038/ngeo278

99. S. Schüpbach *et al.*, Greenland records of aerosol source and atmospheric lifetime changes from the Eemian to the Holocene. *Nat. Commun.* **9**, 1476 (2018). doi: 10.1038/s41467-018-03924-3; pmid: 29662058

100. S. Kienast, G. Winckler, J. Lippold, S. Albani, N. Mahowald, Tracing dust input to the global ocean using thorium isotopes in marine sediments: Thoromap. *Global Biogeochem. Cycles* **30**, 1526–1541 (2016). doi: 10.1002/2016GB005408

101. J. H. Seinfeld *et al.*, Improving our fundamental understanding of the role of aerosol-cloud interactions in the climate system. *Proc. Natl. Acad. Sci. U.S.A.* **113**, 5781–5790 (2016). doi: 10.1073/pnas.1514043113; pmid: 27222566

102. G. N. Inglis, B. D. A. Naafs, Y. Zheng, J. Schellekens, R. D. Pancost, δ$^{13}$C values of bacterial hopanoids and leaf waxes as tracers for methanotrophy in peatlands. *Geochim. Cosmochim. Acta* **260**, 244–256 (2019). doi: 10.1016/j.gca.2019.06.030

103. Y. G. Zhang, M. Pagani, J. Henderiks, H. Ren, A long history of equatorial deep-water upwelling in the Pacific Ocean. *Earth Planet. Sci. Lett.* **467**, 1–9 (2017). doi: 10.1016/j.epsl.2017.03.016

104. J. Hansen, M. Sato, G. Russell, P. Kharecha, Climate sensitivity, sea level and atmospheric carbon dioxide. *Philos. Trans. A Math. Phys. Eng. Sci.* **371**,

20120294–20120294 (2013). doi: 10.1098/rsta.2012.0294; pmid: 24043864

105. G. L. Foster, D. L. Royer, D. J. Lunt, Future climate forcing potentially without precedent in the last 420 million years. *Nat. Commun.* **8**, 14845 (2017). doi: 10.1038/ncomms14845; pmid: 28375201

106. O. Friedrich, R. D. Norris, J. Erbacher, Evolution of middle to Late Cretaceous oceans—A 55 m.y. record of Earth's temperature and carbon cycle. *Geology* **40**, 107–110 (2012). doi: 10.1130/G32701.1

107. C. R. Witkowski, J. W. H. Weijers, B. Blais, S. Schouten, J. S. Sinninghe Damsté, Molecular fossils from phytoplankton reveal secular P$_{CO_2}$ trend over the Phanerozoic. *Sci. Adv.* **4**, eaat4556 (2018). doi: 10.1126/sciadv.aat4556; pmid: 30498776

108. S. M. Sosdian *et al.*, Constraining the evolution of Neogene ocean carbonate chemistry using the boron isotope pH proxy. *Earth Planet. Sci. Lett.* **498**, 362–376 (2018). doi: 10.1016/j.epsl.2018.06.017

109. M. Meinshausen *et al.*, The SSP greenhouse gas concentrations and their extensions to 2500. *Geosci. Model Dev. Discuss.* **2019**, 1–77 (2019).

110. E. L. Grossman, T.-L. Ku, Oxygen and carbon isotope fractionation in biogenic aragonite: Temperature effects. *Chem. Geol.* **59**, 59–74 (1986). doi: 10.1016/0168-9622(86)90057-6

111. J. Zhu *et al.*, Reduced ENSO variability at the LGM revealed by an isotope-enabled Earth system model. *Geophys. Res. Lett.* **44**, 6984–6992 (2017). doi: 10.1002/2017GL073406

112. R. M. Deconto *et al.*, Thresholds for Cenozoic bipolar glaciation. *Nature* **455**, 652–656 (2008). doi: 10.1038/nature07337; pmid: 18833277

# Science

## Past climates inform our future

Jessica E. Tierney, Christopher J. Poulsen, Isabel P. Montañez, Tripti Bhattacharya, Ran Feng, Heather L. Ford, Bärbel Hönisch, Gordon N. Inglis, Sierra V. Petersen, Navjit Sagoo, Clay R. Tabor, Kaustubh Thirumalai, Jiang Zhu, Natalie J. Burls, Gavin L. Foster, Yves Goddéris, Brian T. Huber, Linda C. Ivany, Sandra Kirtland Turner, Daniel J. Lunt, Jennifer C. McElwain, Benjamin J. W. Mills, Bette L. Otto-Bliesner, Andy Ridgwell and Yi Ge Zhang

### The future in the past

A major cause of uncertainties in climate projections is our imprecise knowledge of how much warming should occur as a result of a given increase in the amount of carbon dioxide in the atmosphere. Paleoclimate records have the potential to help us sharpen that understanding because they record such a wide variety of environmental conditions. Tierney *et al.* review the recent advances in data collection, statistics, and modeling that might help us better understand how rising levels of atmospheric carbon dioxide will affect future climate.
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