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# Past climates inform our future: Review Summary

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**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 analogue 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 novel 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<sub>2</sub>, 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<sub>2</sub> 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<sub>2</sub> concentrations. Simulations of past warm climates such as the Eocene further highlight the role that cloud feedbacks play in contributing to high ECS under elevated CO<sub>2</sub> 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, analysis 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 sub-seasonal 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 timescale 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. Finally, 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.

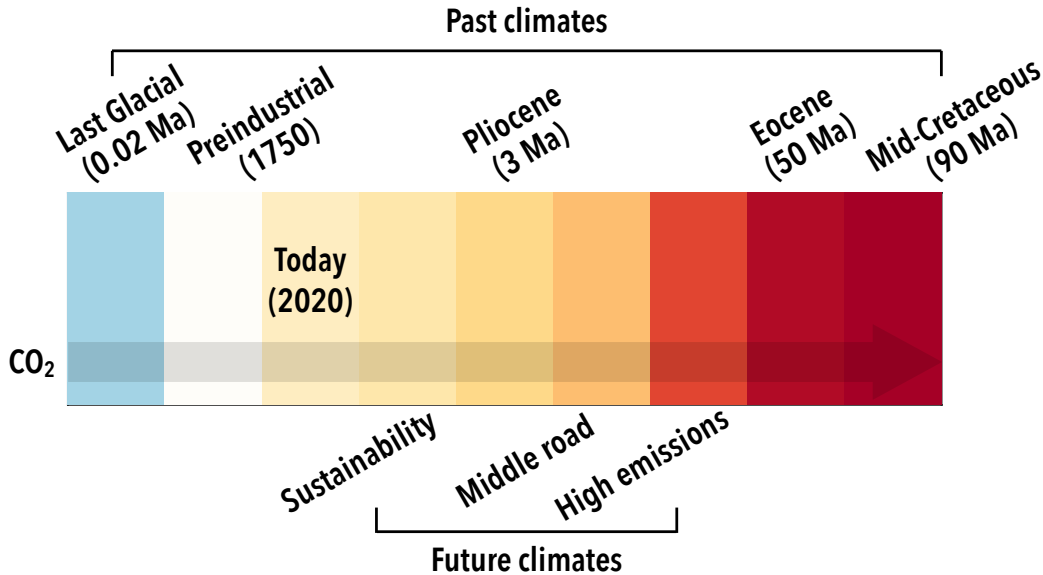


Figure 1: Past climates (denoted on top) provide context for future climate scenarios (at bottom). Ma = millions of years ago. Both past and future climates are colored by their estimated change in global mean annual surface temperature relative to preindustrial conditions. “Sustainability”, “Middle road”, and “High emissions” represent the estimated global temperature anomalies at 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<sub>2</sub>.

**Outlook:** A common concern with using paleoclimate information as model targets is that non-CO<sub>2</sub> forcings, such as aerosols and trace greenhouse gases, are not well known, especially in the distant past. While evidence thus far suggests that such forcings are secondary to CO<sub>2</sub>, future improvements in both geochemical proxies and modeling are on track to tackle this issue. New and rapidly evolving geochemical techniques have 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 inform our future

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1 **As the world warms, there is a profound need to improve projections**  
2 **of climate change. While the latest Earth system models offer an un-**  
3 **precedented number of features, fundamental uncertainties continue to**  
4 **cloud our view of the future. Past climates provide the only opportu-**  
5 **nity to observe how the Earth system responds to high CO<sub>2</sub>, underlining**  
6 **a fundamental role for paleoclimatology in constraining future climate**  
7 **change. Here, we review the relevancy of paleoclimate information for**  
8 **climate prediction and discuss the prospects for emerging methodologies**  
9 **to further insights gained from past climates. Advances in proxy methods**

10 **and interpretations pave the way for the use of past climates for model**  
11 **evaluation – a practice we argue should be widely adopted.**

## 12 **1 Introduction**

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

20 While paleoclimatology continues to narrate the history of Earth’s climate, it also plays an  
21 increasingly central role in understanding future climate change. The study of past climate has  
22 never been more relevant than now, as anthropogenic activities increase atmospheric greenhouse  
23 gas concentrations and modify the land surface and ocean chemistry at a rate and scale that  
24 exceed natural geologic processes. CO<sub>2</sub> levels are higher now than at any point in at least the  
25 last three million years and, at the current rate of emissions, will exceed concentrations typical  
26 of the last 30 million years by 2300 (Fig. 1). In this context, past climates are windows into  
27 our future (3) – the geological record is the only observational source of information for how the  
28 climate system operates in a state much warmer than the present.

29 The challenge for paleoclimatology is that there are few direct quantitative records of past  
30 climate (e.g. temperature, precipitation). Instead, we make use of “proxies,” surrogates for  
31 climate variables that cannot be measured directly. In some cases, the physical occurrence  
32 (or absence) of a proxy (like glacial deposits) reveals information about past environmental  
33 conditions. More often, geochemical data (such as elemental and stable isotope ratios) stored in  
34 fossils, minerals, or organic compounds, are used to infer past conditions. The discovery of new  
35 proxies, improvements in modeling and analytical techniques, and the increasing number of proxy  
36 records are actively expanding the utility of paleoclimate information. These innovations are  
37 refining our understanding of how the climate system responds to large changes in atmospheric  
38 CO<sub>2</sub>, and provide insights into aspects of past climates (such as seasonality and interannual  
39 variability) that were heretofore unknowable.

40 Among the most important contributions that paleoclimatology can make is the evaluation  
41 of Earth system models that we rely on for projecting future climate change. The physical  
42 parameterizations in these sophisticated models are often tuned to fit the preindustrial or his-  
43 torical record (4). However, the latter is short in duration and samples a single climate state  
44 with a narrow CO<sub>2</sub> range. The performance of climate models under extreme forcing very  
45 different than present is not commonly assessed, despite the fact that the models are used to  
46 project changes under high-emissions scenarios. When these models are used to simulate past  
47 warm climates, they often predict surface temperatures that are too cold and pole-to-equator  
48 temperature gradients that are too large (5). However, a new generation of models, alongside  
49 developments in proxy techniques and analysis, now provide opportunities to more fully exploit  
50 past climates for model evaluation and assessment of key metrics of the climate system.

## 51 2 Past climates inform key processes

52 Earth’s paleoclimate record contains tremendous variability. Over the last 100 million years, the  
53 climate gradually transitioned from an ice-free world of exceptional warmth (the mid-Cretaceous,  
54 92 Ma, Fig. 1) to the cold ice ages of the past few million years, glacial worlds with kilometers-  
55 thick ice caps covering one-fourth of the land surface (such as the Last Glacial Maximum (LGM),  
56 21 ka, Fig. 1). Between Cretaceous and LGM extremes lie intermediate warm climates such as  
57 the early Eocene (53–49 Ma) and Pliocene (5.3–2.6 Ma) (Fig. 1). This long-term climate transi-  
58 tion was far from steady – short-lived hyperthermal events (6) and cold stadials (7) punctuated  
59 the slower trends.

60 Atmospheric CO<sub>2</sub> concentrations generally follow these swings in global temperature (Fig. 1).  
61 Geochemical modeling demonstrates that the balance of geological sources (degassing through  
62 volcanism) and sinks (weathering and sedimentation) explains the general features of CO<sub>2</sub>’s  
63 trajectory (8) and establishes causality – high CO<sub>2</sub> leads to high temperatures. The apparent  
64 exceptions to this rule, including the end-Cretaceous and early Paleocene (70–60 Ma) and the  
65 Miocene (23–5.3 Ma), are areas of active research. One explanation for the decoupling of CO<sub>2</sub>  
66 and temperature is that uncertainties associated with the proxies blur the relationship. Esti-  
67 mation of past CO<sub>2</sub> is challenging. Beyond the ice core record (9), CO<sub>2</sub> information comes  
68 from geochemical data, such as isotope ratios of boron and carbon, or paleobotanical indicators  
69 such as the density of leaf stomata. All of these proxies require assumptions about the phys-  
70 ical, chemical, and biological state of the past that are not completely understood, sometimes  
71 leading to misinterpretations of the signal (10). Proxy methodologies and assumptions continue  
72 to be refined, and there is some indication that CO<sub>2</sub> at the end of the Cretaceous may have  
73 been higher than shown in Fig. 1 (11). It is also possible that these discrepancies have another  
74 explanation, such as a greater-than-expected role for non-CO<sub>2</sub> forcings and feedbacks. If the  
75 paleoclimate record has taught us anything, it is that the more we probe, the more we learn.

76 Past climate states were profoundly different from today. Their global mean temperatures,  
77 latitudinal temperature gradients, polar ice extents, regions of deep-water formation, vegetation  
78 types, patterns of precipitation and evaporation, and variability were all different. These dif-  
79 ferences are invaluable as they provide rich evidence of how climate processes operated across  
80 states that span the range of CO<sub>2</sub> concentrations (400–2000 ppm) associated with future emis-  
81 sions scenarios (the Shared Socioeconomic Pathways (SSPs), Fig. 1). Under the sustainable  
82 SSP1-2.6 scenario, in which emissions are curtailed and become net-negative by the end of the  
83 21st century, CO<sub>2</sub> concentrations would be stabilized near Pliocene levels (Fig. 1). In contrast,  
84 under the fossil-fuel intensive SSP5-8.5 scenario, CO<sub>2</sub> concentrations would approach or even  
85 exceed Eocene or mid-Cretaceous levels (Fig. 1). These past warm climates therefore serve as  
86 targets against which to measure the increasingly complex generation of climate models that  
87 are used for future climate prediction.

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

94 climate sensitivity.

### 95 **3 Paleoclimate constraints on climate sensitivity**

96 Equilibrium climate sensitivity (ECS) has been widely adopted as a simple metric of how re-  
97 sponsive the Earth’s climate system is to radiative forcing. It is defined as the change in global  
98 near-surface air temperature resulting from a sustained doubling in atmospheric CO<sub>2</sub> after the  
99 fast-acting (timescales of years to decades) feedback processes (water vapor, clouds, snow) in  
100 the Earth system reach equilibrium. The 5th assessment report of the IPCC determined that  
101 ECS was likely between 1.5 and 4.5°C, a large range that has remained essentially unchanged for  
102 40 years (13). Because the environmental impacts, socio-economic implications, and mitigation  
103 timescales are very different for a low versus a high ECS (14), narrowing its range has always  
104 been a high priority.

105 The fact that models with either a low or high present-day ECS can match historical ob-  
106 servations (15) suggests that preindustrial and industrial climatic changes provide insufficient  
107 constraints on this important metric. Furthermore, the emerging view is that ECS is depen-  
108 dent on, and changes with, the background climate state – specifically, it increases in warmer  
109 climates (16–19). Past warm climates therefore provide key constraints on the range of plausi-  
110 ble ECS values as well as the strength of feedbacks involved. Simulations of the early Eocene  
111 provide a salient example. Figure 2 shows a comparison between the ECS of Coupled Model In-  
112 tercomparison Project (CMIP) Phase 5 and Phase 6 models (used in the last and the upcoming  
113 IPCC assessments) and the ECS of preindustrial and Eocene simulations conducted with the  
114 National Center for Atmospheric Research (NCAR) Community Earth System Model (CESM)  
115 version 1.2 (19). Doubling CO<sub>2</sub> in an Eocene experiment with preindustrial CO<sub>2</sub> (285 ppm;  
116 1X) yields an ECS similar to the preindustrial experiment and overlaps with the CMIP range  
117 (Fig. 2). This indicates that non-CO<sub>2</sub> Eocene boundary conditions, including the position of  
118 the continents and the absence of continental ice sheets, do not have a large effect on ECS in  
119 CESM1.2. In contrast, raising CO<sub>2</sub> levels elevates ECS in the Eocene simulations to values above  
120 6°C (Fig. 2). This increase in ECS with increasing temperature results in accurate simulation  
121 of Eocene temperatures at CO<sub>2</sub> concentrations that agree with proxy estimates (Fig. 2, inset).  
122 The elevated ECS under high CO<sub>2</sub> in CESM1.2 is due to improved representation of clouds in  
123 the CAM5 atmospheric model, which drives a strong low-cloud positive feedback under high  
124 CO<sub>2</sub> (19) – a finding in agreement with the emerging recognition that cloud feedbacks are a  
125 key component of warm climates (20, 21). The fact that CESM1.2 simulates Eocene proxy tem-  
126 peratures within the bounds of proxy CO<sub>2</sub> estimates provides support for its cloud physics and  
127 increases our confidence in the model’s state-dependent ECS. The Geophysical Fluid Dynamics  
128 Laboratory (GFDL) CM2.1 model can also simulate the large-scale features of Eocene proxy  
129 temperatures (22), likewise suggesting that its ECS is reasonable. On the other hand, CESM2  
130 (the newest version of the NCAR model) estimates Eocene temperatures that exceed the upper  
131 bound of proxy constraints at low CO<sub>2</sub> levels (23), suggesting that its modern ECS of 5.3°C  
132 is too high. A little more than a third of the newest-generation CMIP6 models have an ECS  
133 higher than 4.5°C (15). From historical observations alone, it is very challenging to assess the  
134 plausibility of these higher ECS values. As the Eocene example highlights, warm paleoclimates  
135 are key in this respect.

136 The early Eocene provides an important constraint on model ECS but samples a single  
137 high-CO<sub>2</sub> climate state. Given the dependence of ECS on the background climate state, other  
138 past climates are critical to constraining ECS and relevant physics under both lower (e.g. LGM,  
139 Pliocene) and higher (e.g. Eocene, Cretaceous) background CO<sub>2</sub> levels. One concern about using  
140 past climates as model targets is that the forcings, especially aerosol and non-CO<sub>2</sub> greenhouse  
141 gas concentrations, are uncertain and increasingly so in the distant past. While important, it is  
142 worth noting that these forcings are secondary to CO<sub>2</sub> (e.g. (24)) and, for extreme climates like  
143 the Eocene and Cretaceous, may largely fall within the climate proxy uncertainties. Moreover,  
144 this concern can be mitigated by examining model responses to the potential range of under-  
145 constrained forcings and, as is increasingly done, by incorporating biogeochemical cycles and  
146 the simulation of aerosol production and transport into the models.

#### 147 4 Paleoclimate perspectives on the stability of the cryosphere

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

158 Paleoclimates also provide critical insights into processes that drive destabilization of ice  
159 sheets. Of particular relevance for future projections is assessing the likelihood of marine ice-  
160 cliff instability (MICI), a rapid collapse of coastal ice cliffs following the disintegration of an  
161 ice shelf, which has the potential to contribute to substantial sea level rise by the end of the  
162 21st century (32, 33). The record of sea level change from past warm climates offers a way to  
163 test this hypothesis. Recent work has focused on the Pliocene, given that CO<sub>2</sub> concentrations  
164 during this time were similar to current anthropogenic levels (Fig. 1). A new reconstruction of  
165 global mean sea-level during the mid-Pliocene warm period indicates a rise of ~ 17 m, implying  
166 near-to-complete loss of Greenland and the West Antarctic Ice Sheet with some additional  
167 contribution from East Antarctica (34). While this represents an outstanding loss of ice, MICI  
168 is not necessarily needed to explain it (33, 34). However, simulated changes in sea level are highly  
169 dependent on each model’s treatment of ice sheet stability (35), and paleoclimate investigations  
170 of warmer climates, such as the early Pliocene and the Miocene, indicate larger magnitudes of  
171 ice loss, thermal expansion, and consequent sea level rise (34, 36). Moving forward, refining our  
172 understanding of threshold behavior in ice sheets, and thus improving projections of future sea  
173 level rise, will require a synergistic approach that leverages paleoclimate estimates from multiple  
174 warm climates alongside solid Earth, ice sheet, and climate modeling (31).



## 5 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, as 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,  $\delta^{18}\text{O}$  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). Since individual planktic foraminifera live for about a month, analyses of single shells yields 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, as proxy signals tend to be more complex; however, even basic directional information (wetter vs. drier) can be used to test 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 analogue for future warming.

Studies of past warm climates have the potential to provide even more insights into the

218 behavior of regional climate in a warming world. Future model projections broadly simulate a  
219 pattern of subtropical drying, while the deep tropics and high latitudes get wetter (40). Recently,  
220 however, researchers have argued that subtropical drying is transient and might not persist in  
221 equilibrium with higher radiative forcing (59, 60). Indeed, several paleoclimatic intervals (61, 62)  
222 suggest that a warmer world could feature a different pattern, with wetter conditions in both the  
223 subtropics and high latitudes (50). This pattern is especially evident in western North America,  
224 where widespread Pliocene lake deposits suggest much wetter conditions (63). This evidence  
225 stands in stark contrast to future projections for this region, which overwhelmingly predict drier  
226 conditions and more intense droughts (64), and suggests that paleoclimates can provide vital  
227 constraints on the response of arid lands to higher CO<sub>2</sub> concentrations.

## 228 6 Climatic aberrations

229 Among the most important discoveries in paleoclimatology is the occurrence of climatic “aber-  
230 rations” – extraordinary transient departures from a background climate state. Such events are  
231 distinguished by radical changes in temperature, precipitation patterns, and ocean circulation  
232 that often leave distinctive marks in the geological record, like the pervasive black shales of  
233 the mid-Cretaceous Ocean Anoxic Events (65). An aberration typically occurs in response to  
234 a short-lived perturbation to the climate system, such as a sudden release of greenhouse gases  
235 (e.g., from volcanoes, methane clathrates, or terrestrial organic deposits). Aberrations need not  
236 be “abrupt” in the sense that the rate of climate change must exceed the rate of forcing, and  
237 they can potentially last for a long time (for example, the Sturtian Snowball Earth lasted 55  
238 million years (66)). They are instructive because they provide information on extreme climate  
239 states, and the ability of the Earth system to recover from such states.

240 One of the most striking aberrations in the paleoclimate record, the Paleocene-Eocene Ther-  
241 mal Maximum (PETM), may foreshadow future changes that Earth will experience due to  
242 anthropogenic emissions. The PETM, which occurred 56 million years ago, was triggered by  
243 rapid emission of greenhouse gases; proxy and model estimates suggest that CO<sub>2</sub> doubled or  
244 even tripled from a background state of ~900 ppm (67–69) in less than 5,000 years (70, 71). In  
245 response, global temperatures spiked by 4–6°C (72). The surface ocean rapidly acidified (68, 73),  
246 and seafloor carbonates dissolved (74), resulting in dramatic biogeographic range shifts in plank-  
247 ton and the largest extinction in deep-sea calcifying benthic foraminifera ever observed (75). Pre-  
248 cipitation patterns changed dramatically, with much more rain falling at the high latitudes (76).  
249 It took the Earth ~100,000 years to recover from this perturbation (68, 77).

250 Although the PETM stands out starkly in the geologic record, the rate of CO<sub>2</sub> release was still  
251 4–10 times slower than current anthropogenic emissions (71, 78). Indeed, the geological record  
252 leaves no doubt that our current rate of global warming, driven by anomalous (anthropogenic)  
253 forcing, is an exceptional aberration – the rate and magnitude of change far exceeds the typical  
254 multi-thousand year variability that preceded it (Fig. 4). In the last 100 million years, CO<sub>2</sub>  
255 has ranged from maximum values in the mid-Cretaceous to minimum levels at the Last Glacial  
256 Maximum (Fig. 1). Going forward, we are on pace to experience an equivalent magnitude  
257 of change in atmospheric CO<sub>2</sub> concentrations, in reverse, over a period of time that is over  
258 10,000 times shorter (Fig. 4). In just over 150 years, we have already raised CO<sub>2</sub> concentrations  
259 (currently at 410 ppm) to Pliocene levels (Fig. 4). Under a middle-of-the-road emissions scenario

260 such as SSP2-4.5 (or the CMIP5 equivalent, RCP4.5), CO<sub>2</sub> will approach 600 ppm by Year 2100,  
261 and if we follow the high-emissions SSP5-8.5 (or RCP8.5), CO<sub>2</sub> will rise beyond mid-Cretaceous  
262 concentrations (ca. 1000 ppm) by Year 2100 (Fig. 4). In comparison, over the past 800,000  
263 years of geologic history CO<sub>2</sub> only varied between 180 and 280 ppm (9) (see also Fig. 4).

264 How long will it take for the Earth to neutralize anthropogenic CO<sub>2</sub> and return to pre-  
265 industrial levels? The Earth has the ability to recover from a rapid increase in atmospheric CO<sub>2</sub>  
266 concentration – the PETM is a textbook example of this process. In fact, in every case of past  
267 CO<sub>2</sub> perturbations, the Earth system has compensated in order to avoid a runaway greenhouse  
268 or a permanent icehouse. Yet the natural recovery from aberrations takes place on geological, not  
269 anthropogenic, timescales (Fig. 4). Some of the processes that remove CO<sub>2</sub> from the atmosphere  
270 occur on relatively short (100–1000 yr) timescales (e.g. ocean uptake), but others take tens to  
271 hundreds of thousands of years (e.g. weathering of silicate rocks) (79). Using the intermediate  
272 complexity Earth system model cGENIE, we can estimate how long the recovery process takes  
273 under different future forcing scenarios. Under an aggressive mitigation scenario (RCP 2.6), CO<sub>2</sub>  
274 concentrations remain at Pliocene-like concentrations (>350 ppm) through Year 2350, but it still  
275 takes hundreds of thousands of years for concentrations to return to preindustrial levels (Fig. 4).  
276 Under a middle-of-the-road scenario (RCP 4.5), CO<sub>2</sub> peaks around 550 ppm and remains above  
277 Pliocene levels for 30,000 years. Under a worst-case scenario (RCP 8.5) atmospheric CO<sub>2</sub> will  
278 remain at mid-Cretaceous (>1000 ppm) concentrations for 5,000 years, at Eocene concentrations  
279 (~850 ppm) for 10,000 years, and at Pliocene concentrations (>350 ppm) for 300,000 years (Fig.  
280 4). It will be at least 500,000 years, a duration equivalent to 20,000 human generations, before  
281 atmospheric CO<sub>2</sub> fully returns to preindustrial levels. Our planet will recover, but for humans,  
282 and the organisms with which we share this planet, the changes in climate will appear to be a  
283 permanent state shift.

## 284 7 Bridging the gap between paleoclimate data and models

285 Climate models provide direct estimates of quantities like temperatures, wind speed, and precip-  
286 itation. In contrast, paleoclimate information is indirect, filtered through a proxy – a physical,  
287 chemical, and/or biological entity that responds to climate – such as foraminifera, algae, or  
288 the chemical composition of sediments. Proxies are imperfect recorders of climate; they have  
289 inherent uncertainties associated with, for example, biological processes and preservation. Thus,  
290 while proxy data can be transformed into climate variables for direct comparison with models  
291 using regression, transfer functions, and assumptions, if these structural uncertainties are not  
292 accounted for they can lead to unclear or erroneous interpretations. This creates a “language  
293 barrier” between model output and proxy data that has limited the use of paleoclimate infor-  
294 mation to infer past climate states and evaluate climate models. Three key innovations are  
295 now breaking down this barrier, allowing paleoclimate information to directly constrain model  
296 performance: 1) the inclusion of chemical tracers relevant to proxies directly in Earth system  
297 models; 2) the creation of robust proxy system models that explicitly encode processes, uncer-  
298 tainties, and multivariate sensitivities; and 3) the development of statistical methods to formally  
299 combine proxy and model data.

300 As far as chemical tracers are concerned, the single most important advance has been the  
301 increasingly routine incorporation of water isotopes in model simulations. The stable isotopes

302 of water –  $\delta^{18}\text{O}$  and  $\delta\text{D}$  – and their incorporation into natural archives are the foundation of  
303 modern paleoclimatology (80). A large number of paleoclimate proxies record water isotopes,  
304 including foraminifera, stalagmites, leaf waxes, soil carbonates, and ice cores. Water isotope  
305 composition, however, reflects multiple processes including changes in temperature, moisture  
306 source, evaporation, precipitation, and convection. Including water isotopes in models gener-  
307 ates simulated isotope fields that are consistent with the model’s treatment of these processes,  
308 eliminating the need to independently conjecture how these various factors may have influenced  
309 the proxy data. This creates an “apples to apples” comparison between proxy information and  
310 model output that can be used to evaluate model performance and diagnose climatic processes  
311 (e.g. (81). For example, using the water-isotope-enabled CESM1.2 (iCESM) (82), it is possi-  
312 ble to directly compare carbonate  $\delta^{18}\text{O}$  data from Eocene fossil bivalves to model-simulated  
313  $\delta^{18}\text{O}$  (43, 44, 83) (Fig. 3a). The model predicts a roughly 3‰ annual range in carbonate  $\delta^{18}\text{O}$ ,  
314 in good agreement with observed proxy data (Fig. 3a). The match with the  $\delta^{18}\text{O}$  data builds  
315 confidence that the model can correctly simulate climatology in this location, and allows us to  
316 deconvolve the contribution of SST and  $\delta^{18}\text{O}$  of seawater. The site-specific seasonality in SST is  
317 8–10°C, while  $\delta^{18}\text{O}$ -seawater has a seasonal range of 0.6–0.8‰. This indicates that temperature  
318 is primarily responsible for the large seasonal range in carbonate  $\delta^{18}\text{O}$  during this greenhouse  
319 climate state.

320 One aspect of paleoclimate information that has traditionally limited its use in model eval-  
321 uation is an inability to precisely quantify uncertainties surrounding the proxies. However, in  
322 the last decade, increasingly detailed proxy system models (84) have been developed to address  
323 this issue (e.g.,) (85–87). Many of these use Bayesian inference to quantify uncertainties in the  
324 sensitivity of proxies to environmental parameters, which can then be used for probabilistic as-  
325 sessments of past climate states, model-proxy agreement, and model evaluation (88). These have  
326 helped to transform proxy-model comparisons from qualitative statements (“they look similar”)  
327 to quantitative statements (“there is a 90% probability that the data and the model agree”).

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

344 In sum, the disintegration of the model-proxy language barrier has narrowed uncertainties  
345 in proxy interpretation. Recent studies have been able to use proxy data to infer key cli-

346 matic processes and evaluate models across multiple time periods, including the LGM (91), the  
347 Pliocene (88), and the Eocene (19, 22). This opens the door for explicit use of paleoclimates to  
348 assess and improve model physics.

## 349 8 Moving Forward

350 Past climates will continue to provide insights into the range, rate, and dynamics of climate  
351 change. Over the last decade, we have witnessed breakthroughs in proxy development and  
352 refinement as well as the generation of many new high-resolution marine and terrestrial paleo-  
353 oclimate records. In addition to continued advances, the collection of additional temperature  
354 and CO<sub>2</sub> proxy records at higher resolution will be paramount for developing better estimates  
355 of climate sensitivity. Future proxy collection efforts should also focus on hydroclimate proxies,  
356 given the large spread in model projections (40). These reconstructions will help us refine our  
357 understanding of the response of atmospheric circulation and rainfall to climate change.

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

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

378 Looking ahead, there are many outstanding process-based uncertainties associated with fu-  
379 ture climate change that paleoclimatology can help constrain. For example, paleobotanical  
380 records can inform plant physiological responses to changes in CO<sub>2</sub> (95), which remain highly  
381 uncertain (96) but important for quantifying evapotranspirative and surface runoff fluxes. Sim-  
382 ilarly, past vegetation reconstructions can assess dynamic vegetation models and simulated  
383 changes in the hydrologic cycle through time (97). Moreover, additional quantitative reconstruc-  
384 tions of hydroclimate, in combination with better constraints on plant physiological functioning  
385 in the past, will help refine our understanding of the regional water cycle and its dependence on  
386 local energy fluxes and large-scale circulation.

387 New geochemical techniques will also refine our understanding of the Earth system. Devel-

388 opment of radiation (98), biogenic aerosol (99), and dust (100) records have the potential to  
389 help constrain past aerosol and cloud radiative effects, which are arguably the most uncertain  
390 component of Earth system models (101). In addition, new geochemical tracers for methane  
391 cycling (102) and upwelling, which is important for N<sub>2</sub>O production (103), will provide unique  
392 insights into trace greenhouse gases during past climate states. The combination of these new  
393 techniques will allow the paleoclimate community to better quantify biogeochemical feedbacks  
394 and climate sensitivity to greenhouse gas forcings across a range of climate states, and ultimately  
395 improve climate forecasts for the coming decades to millennia.

396 In summary, the paleoclimate record is the basis for how we understand the potential range  
397 and rate of climate change. Past climates represent the only target for climate model predic-  
398 tions at CO<sub>2</sub> concentrations outside of the narrow historical range and, for this reason, are vital  
399 tools for evaluating the newest generation of Earth system models. The study of past climates  
400 continues to reveal key insights to the Earth’s response to elevated concentrations of greenhouse  
401 gases. Innovations in Earth system models, geochemical techniques, and statistical methods  
402 further allow for a more direct connection from the past to the future – worlds for which the  
403 preindustrial and industrial climate states provide limited guidance. The future of paleoclima-  
404 tology is to incorporate past climate information formally in model evaluation, so that we can  
405 better predict and plan for the impacts of anthropogenic climate change.

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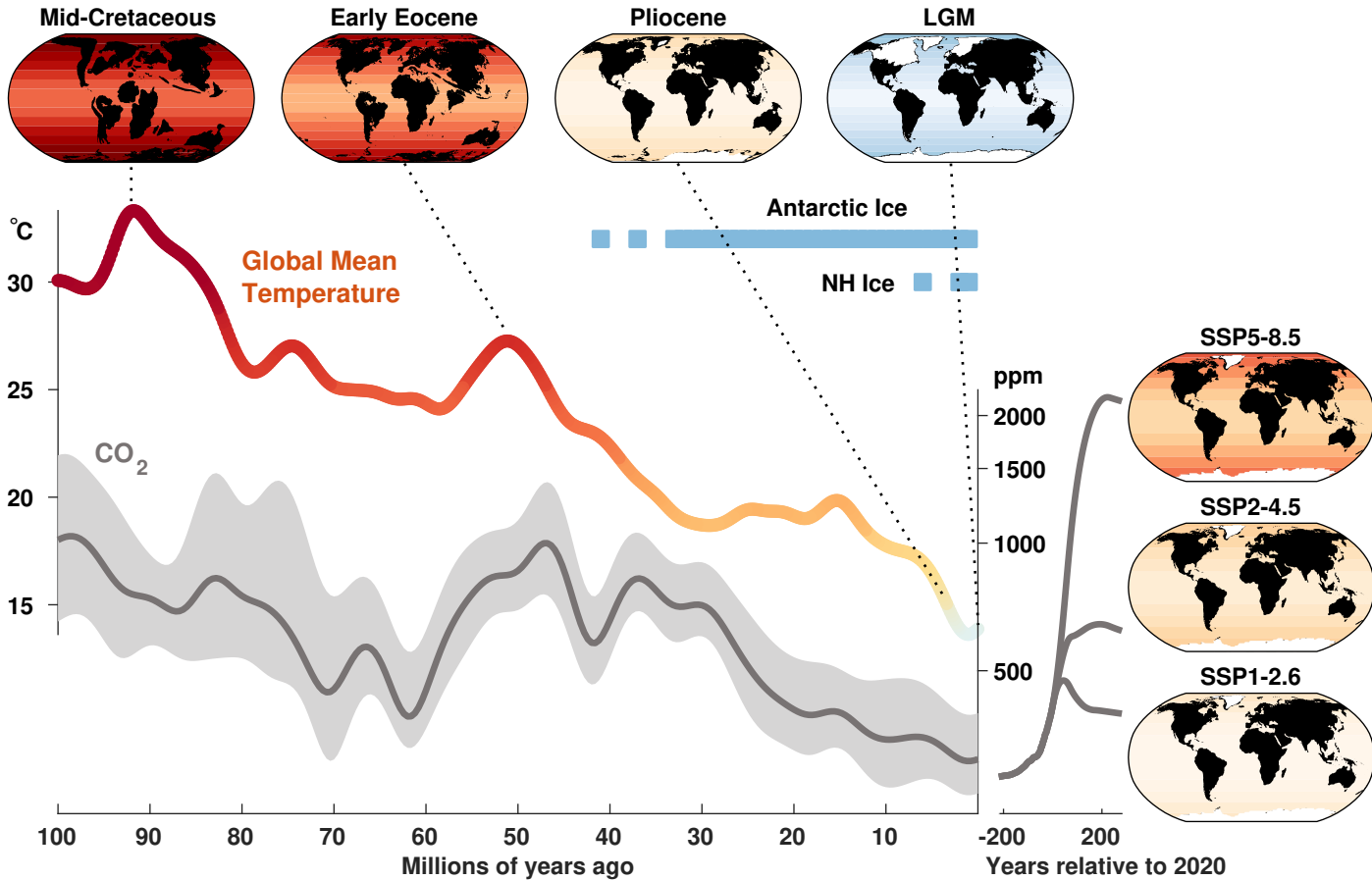


Figure 1: **Paleoclimate context for future climate scenarios.** Global mean surface temperature for the past 100 million years is estimated from benthic  $\delta^{18}\text{O}$  (2, 106) using the method of (104).  $\text{CO}_2$  is estimated from the multi-proxy data set compiled by (105) with additional phytane data from (107) and boron data from (108) and (11). Data with unrealistic values ( $<150$  ppm) are excluded. The  $\text{CO}_2$  error envelopes represent  $1\sigma$  uncertainties. Note logarithmic scale for  $\text{CO}_2$ . Gaussian smoothing was applied to both the temperature and  $\text{CO}_2$  curves in order to emphasize long-term trends. Temperature colors are scaled relative to preindustrial conditions. The maps show simplified representations of surface temperature. Projected  $\text{CO}_2$  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).

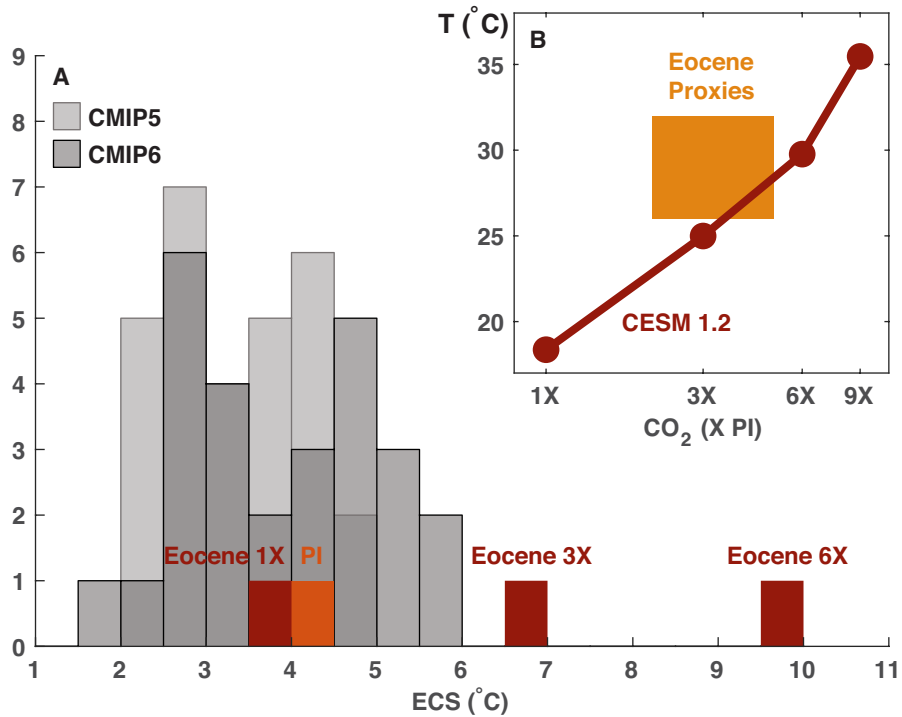


Figure 2: **Constraining equilibrium climate sensitivity (ECS) through simulation of the early Eocene.** a. ECS in CMIP5 and CMIP6 models (grey bars; (15)) compared to ECS in the CESM1.2 preindustrial (PI, orange bar) and Eocene simulations with 1X, 3X and 6X preindustrial CO<sub>2</sub> levels (red bars). b. CO<sub>2</sub> concentrations (times preindustrial level) vs. global mean temperature according to early Eocene proxies (yellow patch) compared to the results from the CESM1.2 Eocene simulations. Proxy CO<sub>2</sub> estimates are a 2 $\sigma$  range derived from the collection plotted in Figure 1. Readers are referred to (19) for details of the Eocene climate simulations and proxy global mean temperature estimation.



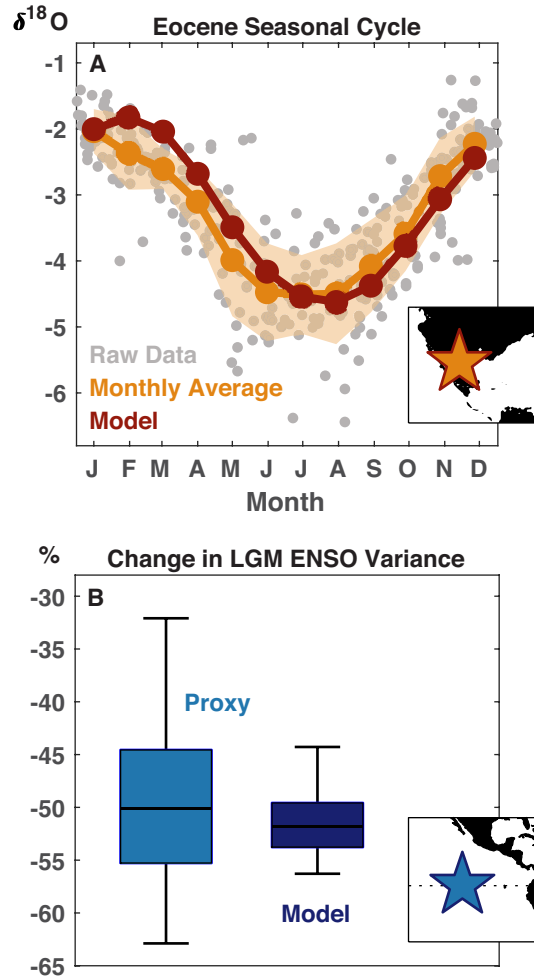


Figure 3: **Examples of seasonal and interannual paleoclimate data and comparison to models.** (a) Seasonally-resolved  $\delta^{18}\text{O}$  carbonate from the shells of a fossil bivalve, *Venericardia hatcheplata*, from the early Eocene Hatchetigbee Formation (orange star in inset) (43, 44). Monthly averaged data (orange, with  $1\sigma$  uncertainty bounds) are compared with predicted  $\delta^{18}\text{O}$ -carbonate seasonality at the same grid-point from an isotope-enabled Eocene model simulation (19) (red) (using modeled  $\delta^{18}\text{O}$  of seawater and SST, and the calibration of ref. (110)). (b) Mg/Ca measurements of individual planktic foraminifera *Trilobatus sacculifer* from an eastern equatorial site (blue star in inset) provide proxy evidence of a reduction in ENSO variability during the Last Glacial Maximum (LGM) relative to pre-industrial conditions (46) (lighter blue). The magnitude of reduction agrees with simulations using CESM1.2 (darker blue) (111).

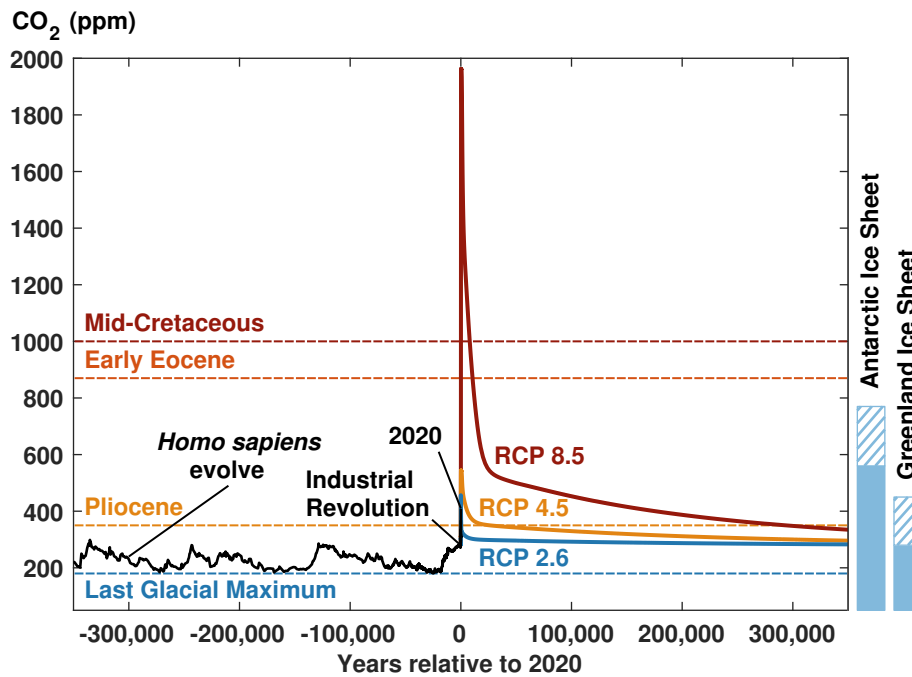


Figure 4: **The anthropogenic climate aberration.** Black line shows CO<sub>2</sub> measured in ice cores for the past 350,000 years (9). Solid colored lines show future CO<sub>2</sub> concentrations for the IPCC AR5 Representative Concentration Pathways, run out to 350,000 years in the future with the cGENIE model. Dotted lines indicate average CO<sub>2</sub> for key time periods in the geologic past. Bars at right indicate CO<sub>2</sub> 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).

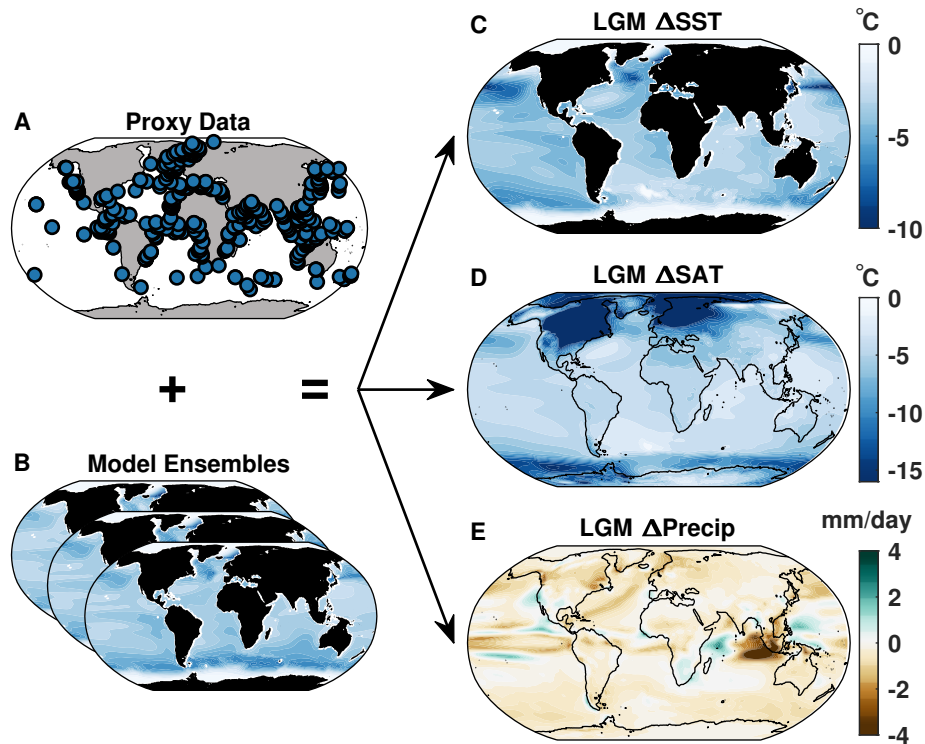


Figure 5: **An example of paleoclimate data assimilation.** Marine sea-surface temperature (SST) proxy data from the Last Glacial Maximum (LGM) and the Late Holocene (a) are combined with an ensemble of model simulations (b) which contain multiple climatic variables. The results (c-e; LGM - Late Holocene differences for sea-surface temperature ( $\Delta$ SST), surface air temperature ( $\Delta$ SAT), and mean annual precipitation ( $\Delta$ Precip)) 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 ref. (91).