Are there pre-Quaternary geological analogues for a future greenhouse warming?

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Given the inherent uncertainties in predicting how climate and environments will respond to anthropogenic emissions of greenhouse gases, it would be beneficial to society if science could identify geological analogues to the human race’s current grand climate experiment. This has been a focus of the geological and palaeoclimate communities over the last 30 years, with many scientific papers claiming that intervals in Earth history can be used as an analogue for future climate change. Using a coupled ocean–atmosphere modelling approach, we test this assertion for the most probable pre-Quaternary candidates of the last 100 million years: the Mid- and Late Cretaceous, the Palaeocene–Eocene Thermal Maximum (PETM), the Early Eocene, as well as warm intervals within the Miocene and Pliocene epochs. These intervals fail as true direct analogues since they either represent equilibrium climate states to a long-term CO2 forcing—whereas anthropogenic emissions of greenhouse gases provide a progressive (transient) forcing on climate—or the sensitivity of the climate system itself to CO2 was different. While no close geological analogue exists, past warm intervals in Earth history provide a unique opportunity to investigate processes that operated during warm (high CO2) climate states. Palaeoclimate and environmental reconstruction/modelling are facilitating the assessment and calculation of the response of global temperatures to increasing CO2 concentrations in the longer term (multiple centuries); this is now referred to as the Earth System Sensitivity, which is critical in identifying CO2 thresholds in the atmosphere that must not be crossed to avoid dangerous levels of climate change in the long term. Palaeoclimatology also provides a unique and independent way to evaluate the qualities of climate and Earth system models used to predict future climate.

Keywords: palaeoclimate; proxies; climate models; climate sensitivity; Earth System Sensitivity; analogue

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1. Introduction

(a) Extending uniformitarianism

Uniformitarianism, the observation that, fundamentally, the same geological processes that operate today also operated in the distant past is a key principle of modern geology. The methodological significance of the principle is summarized in the statement the present is the key to the past. Uniformitarianism originated with the work of the geologist James Hutton, but was refined by John Playfair and popularized by Charles Lyell’s Principles of geology [1]. Can, though, the principle be reversed and extended so that the past becomes the key to the future?

The search for geological analogues for future climate change is not new, but what is new is a realization that, given current and projected levels of greenhouse gases in the atmosphere, we must travel far into the geological past to find intervals of time in which greenhouse gases, as well as global temperatures, were comparable to what climate models predict will occur by the end of the twenty-first century [2]. Trace gas records from ice cores indicate that atmospheric concentrations of CO₂ are already higher than at any time during the last 800,000 years [3–5]. Evidence from new alkenone-based, boron isotope-based and stomatal density-based CO₂ proxy data indicate that the current (2010) concentration of CO₂ (390 ppmv recorded in 2009; data from the National Oceanic and Atmospheric Administration’s Earth System Research Laboratory, Mauna Loa Observatory) in the atmosphere may not have been reached in the last 3 million years. A doubling of pre-industrial concentrations of CO₂ to 560 ppmv, as expected later this century, has not been reconstructed for the last 20 million years [6–11].

In 1991, Thomas J. Crowley published what the authors consider to be a landmark paper in the Journal of Climate [12]: he examined, for the first time, suggestions made by Kellogg [13], Hansen & Lebedeff [14], Budyko & Sedunov [15] and Zubakov & Borzenkova [16] that warm intervals in Earth history could be used as a frame of reference or even possible analogues for future atmospheric CO₂-induced warming. Existing climate-model predictions of the climatological pattern and response times predicted a doubling of atmospheric CO₂ were compared with early, and in many ways limited, palaeoclimate modelling case studies available at the time from the Pleistocene (Early Holocene and Last Interglacial) and pre-Pleistocene (Early Pliocene, Early Eocene and Mid-Cretaceous). For various reasons, many of which we will directly or indirectly re-examine in this paper, it was concluded that there may be no satisfactory geological analogue and therefore that future discussions on geological analogues should be ‘restricted to the study of processes operating in the climate system, and that continued use of the term for past warm periods be abandoned’. Yet, there are many examples of scientific papers published since 1991 that continue to refer to time periods in Earth history as being geological analogues, or at least partial analogues for the current anthropogenic climate perturbation (e.g. [17,18]). It is interesting to consider why this is the case. Is it because the science has progressed since Crowley [12], and that clear examples of past warm intervals as analogues for future greenhouse-gas-induced warming are now available? Or is it because Crowley’s [12] analysis was limited by the modelling tools and case studies available at the time?

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In this paper, we revisit the pre-Quaternary case studies examined by Crowley to explore the possibility of geological analogues for future greenhouse warming (from the oldest to the youngest). We do this by using outputs from fully coupled ocean–atmosphere climate models that are able to overcome some of the limitations of the early climate models used by Crowley [12], and through reference to the published literature. We consider a new potential analogue that has arisen since 1991, the Palaeocene–Eocene Thermal Maximum (PETM), and conclude with a discussion on the relevance and usefulness of an examination of Earth history for future climate change. We answer the question are there any satisfactory pre-Quaternary geological analogues for a future greenhouse warming?

To prevent our investigation of analogues becoming unmanageable, we have established criteria that restrict the number of case studies we must discuss. For the pre-Quaternary, we have only selected intervals when atmospheric CO₂ concentrations and global mean temperatures were higher than pre-industrial values, as we consider this a pre-requisite for a time interval considered as a potential analogue. We have also restricted our discussions to intervals of time when unaltered marine sediments are present. This ensures that we have primary information available from both the marine and terrestrial realms to support the published palaeoclimate reconstructions and to constrain our climate simulations (i.e. back to 100 million years ago (Ma)). We have also attempted to examine palaeoclimates that may have been in equilibrium with ambient CO₂ concentrations as well as aberrant events when greenhouse-gas forcing and climate change was transient. Combined, these criteria dictate a focus on greenhouse states of the Cretaceous and Eocene, the PETM as well as warm intervals within the Miocene and Pliocene epochs.

2. Predictions for future greenhouse-gas-induced warming

(a) Predictions of policy-relevant climate change

Before we begin our exploration of geological case studies described above, it is first necessary to briefly review the latest model predictions for ‘climate policy relevant’ future climate change (i.e. end of this century) so that reconstructions and model predictions of past climates can be placed in the appropriate context. The Intergovernmental Panel on Climate Change (IPCC) summary [19] of post twenty-first century climate change will also be considered. Climate-model predictions forced with different CO₂ emissions scenarios (Special Report on Emissions Scenarios (SRES) B1, A1T, B2, A1B, A2 and A1FI), presented in Working Group I of the IPCC [19], project global average surface warming and sea-level rise at the end of the twenty-first century (temperature change in °C at 2090–2099 relative to 1980–1999) to range from 1.1–6.4°C and 0.18–0.59 m, respectively [20]. The precise nature of regional climate responses to CO₂ emissions depends on the selection of a CO₂ emission scenario, but certain climate trends are seen to be independent of the choice of scenario. Warming is expected to be greatest over land and at most high northern latitudes, and least over the Southern Ocean (near Antarctica) and the northern North Atlantic. Sea ice is projected to shrink in both the Arctic and Antarctic under all SRES scenarios. In some projections, Arctic late-summer sea ice disappears almost entirely by the
latter part of the twenty-first century. Based on a range of models, it is probable that future tropical cyclones (typhoons and hurricanes) will become more intense, with larger peak wind speeds and more heavy precipitation associated with ongoing increases of tropical sea-surface temperatures (SSTs). Extra-tropical storm tracks are projected to move poleward, with consequent changes in wind, precipitation and temperature patterns. Increases in precipitation are very likely at high latitudes, while decreases are likely over most subtropical land regions [20].

(b) Climate predictions beyond the twenty-first century

Climate change may continue to be induced by emissions of CO$_2$ long after these emissions have been stabilized, owing to the inertia of the climate system. This makes examination of climate trends after 2100, along with any similarities of consistencies between such trends during warm intervals in Earth history, necessary and advantageous [19–21]. If radiative forcing were to be stabilized, keeping all the radiative forcing agents constant at B1 or A1B levels in 2100, model experiments show that a further increase in global average temperature of about 0.5$^\circ$C would still be expected by 2200. In addition, thermal expansion alone would lead to 0.3–0.8 m of sea-level rise by 2300 (relative to 1980–1999). Thermal expansion would continue for many centuries, owing to the time required to transport heat into the deep ocean. Both past and future anthropogenic CO$_2$ emissions will continue to contribute to warming and sea-level rise for more than a millennium, owing to the time scales required for the removal of this gas from the atmosphere [20].

(c) HadCM3 and predictions of equilibrium global climate with instantaneous CO$_2$ forcing

Model predictions of the climate response to a doubling of atmospheric CO$_2$ have been performed using instantaneous forcing and transient forcing-type experiments. In the former, atmospheric CO$_2$ partial pressure is instantaneously increased from a pre-industrial value, with the simulation then run to equilibrium. In the latter, forcing is gradually introduced by increasing CO$_2$ concentrations in each year of the simulation, which is a more realistic approach. However, palaeoclimate experiments are normally conducted using the instantaneous forcing approach. Therefore, to maintain consistency in our experimental design, we present results of a doubled CO$_2$ experiment using instantaneous forcing.

The future and palaeoclimate simulations presented in this paper were all performed using the Hadley Centre for Climate Prediction and Research fully coupled ocean–atmosphere climate model v. 3 (HadCM3 and HadCM3L). The particulars of this model are well documented (e.g. [22,23]). HadCM3 and HadCM3L were one of the first coupled atmosphere–ocean climate models that required no flux corrections to be made, even for simulations of 1000 years or more [24]. The general circulation model consists of a linked atmospheric model, ocean model and sea-ice model. In HadCM3 and HadCM3L, the horizontal resolution of the atmosphere model is 2.5$^\circ$ in latitude by 3.75$^\circ$ in longitude. This gives a grid spacing at the equator of 278 km in the north–south direction and 417 km east–west, and is approximately comparable to a T42 spectral model resolution. The atmospheric model consists of 19 layers. The spatial resolution over the ocean in HadCM3 is 1.25$^\circ$ × 1.25$^\circ$ and the model has 20
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layers, while in HadCM3L, the horizontal resolution of the ocean is the same as the atmosphere. The atmospheric model has a time step of 30 min and includes a radiation scheme that can represent the effects of minor trace gases [25]. A parametrization of simple background aerosol climatology is also included [26]. The convection scheme is that of Gregory et al. [27]. A land-surface scheme includes the representation of the freezing and melting of soil moisture. The representation of evaporation includes the dependence of stomatal resistance on temperature, vapour pressure and CO₂ concentration [28].

The ocean model includes the use of the Gent–McWilliams mixing scheme [29]. There is no explicit horizontal tracer diffusion in the model. The horizontal resolution allows the use of a smaller coefficient of horizontal momentum viscosity leading to an improved simulation of ocean velocities. The sea-ice model is a simple thermodynamic scheme and contains parametrizations of ice drift and leads (Polynyas; [30]).

3. Geological case studies

(a) The Cretaceous and Early Eocene

We begin our analysis of potential pre-Quaternary analogues by examining some of the most extreme and long-term warm events in the last 100 million years, specifically the so-called ‘greenhouse intervals’ of the Cretaceous and Early Eocene (100–50 Ma). Reconstructing atmospheric CO₂ this far back in time is extremely challenging and fraught with uncertainties. However, an overview of all available Cretaceous and Early Eocene CO₂ proxy data, along with mass-balance calculations, indicates a picture of higher ambient CO₂ concentrations, albeit with significant variability during this interval. This variability is probably present within a single geological stage, such as the Maastrichtian. For example, available estimates for palaeo CO₂ during the Maastrichtian from the δ¹³C fractionation of pedogenic carbonates [31–33] indicate that CO₂ may have ranged from pre-industrial concentrations (or even lower) to approximately 1500 ppmv. This dynamic range encompasses any SRES emissions scenario for CO₂ and therefore the Cretaceous is often referred to as a potential analogue for future climate change (e.g. [17]).

We compare the response of precipitation and evaporation (and precipitation minus evaporation) predicted by HadCM3L for the Late Cretaceous (Maastrichtian stage) and present-day, both with higher than pre-industrial atmospheric CO₂ concentrations (figure 1a–d). In a control Maastrichtian, simulation with a CO₂ concentration twice that of the pre-industrial level (2 × PAL), the average global annual mean precipitation rate equals 3.32 mm d⁻¹. In our future climate-change experiment, with 2 × PAL CO₂, the average global mean precipitation rate equals 2.95 mm d⁻¹. Initially, this appears to support the utility of the Late Cretaceous as an analogue for the future, since in both cases, the model predicts an enhanced global precipitation rate compared with the pre-industrial simulation (2.83 mm d⁻¹ global mean average total precipitation rate). However, total precipitation rates in the Late Cretaceous are greater than the future climate-change experiment. A Late Cretaceous simulation using 1 × PAL CO₂ reveals that even in the absence of higher than pre-industrial levels of CO₂, the global annual mean precipitation rate is still greater than the pre-industrial
or even future climate-change experiment by approximately 10 per cent and 5 per cent, respectively. More significantly, the magnitude of the response in total precipitation rate to a doubling of CO$_2$ in the future climate-change and Cretaceous scenarios also differs (4.2% and 7.1%, respectively).

This altered intensity of precipitation-rate response between the future and Cretaceous experiments, with or without an increase in CO$_2$ to 2 × PAL, can be ascribed to a number of factors. Firstly, in the Cretaceous scenario, there is no ice at either pole, hence the high-latitude, subpolar and temperate regions are warmer than the present-day and thus evaporation and precipitation are enhanced. Continental position and the area of epeiric seas, which are currently

Figure 1. HadCM3 predictions for a future climate-change experiment and the Maastrichtian experiment: (a) annual mean evaporation rate (mm d$^{-1}$) for 2 × PAL CO$_2$ experiment; (b) annual mean evaporation rate (mm d$^{-1}$) for 2 × PAL CO$_2$ Maastrichtian experiment; (c) difference in annual mean precipitation minus evaporation rate (P – E; mm d$^{-1}$) between a 2 × PAL CO$_2$ experiment and a pre-industrial experiment; (d) difference in annual mean precipitation minus evaporation rate (P – E; mm d$^{-1}$) between a 2 × PAL CO$_2$ Maastrichtian experiment and a pre-industrial CO$_2$ Maastrichtian experiment; (e,f) prescribed bathymetry (m) in HadCM3 for the modern and Maastrichtian simulations. Note the increased area of shallow marine seas (epeiric seas) in the Maastrichtian case. (Online version in colour.)
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...at an all-time Phanerozoic low [34], also appear to play a role. Owing to the continental configuration during the Maastrichtian stage, the area of epeiric seas is greatly expanded compared to today (figure 1e,f). Given that tropical temperatures are high, this has the effect of changing the amount of water evaporated from the oceans in the tropics, introducing a large additional flux of water vapour in the Late Cretaceous scenario that cannot be matched in the pre-industrial or future climate-change experiments (figure 1a–d). Equivalent simulations of the Early Eocene using HadCM3L demonstrate a similar pattern of precipitation-rate enhancement, even in the absence of higher CO2 [35]. Greenhouse worlds of the Cretaceous and Eocene represent equilibrium climate states to long-term CO2 forcing (rather than a transient CO2 forcing), and the sensitivity of the climate system to CO2 during these intervals was apparently different, which is very significant for their utility as analogues.

(b) The Palaeocene–Eocene Thermal Maximum and carbon cycling

A substantial transient warming of the Earth’s surface occurred approximately 55.5 Ma (the PETM); a range of temperature (and hydrological cycling) proxies were recorded [36] and synchronous with a pronounced negative carbon isotopic (\(\delta^{13}C\)) excursion [37]. This isotopic excursion has been observed globally: in marine cores from the Arctic through Equatorial Pacific to the Southern Ocean [38,39], in shelf sediments [40], as well as in those from terrestrial deposits [39,41], and occurs in both biogenic carbonates and organic matter [37]. To explain the event’s global nature, as well as large (ca 4‰) magnitude compared with general Cenozoic variability [42], requires a release of isotopically light carbon to the ocean and/or atmosphere. This, in turn, strongly supports a primary driving role for increased atmospheric greenhouse-gas concentrations in the observed warming and hence highlights the PETM as a potential analogue for future climate change. Of particular future relevance is whether methane (CH4) hydrate destabilization was involved, particularly, in terms of what triggered it and how strong the feedbacks between global warming and CH4 release are. The PETM also represents a past case study into the impacts of acidification of the surface ocean on calcifying organisms such as foraminifera and coccolithophores [43]. However, to draw out any analogue with the future, we need to constrain the source, amount and fate of the carbon release, as these bear on the nature of carbon-cycle feedbacks operating, the degree of ocean acidification and climate sensitivity.

The recorded magnitude of the excursion alone can, in theory, be used to infer the carbon release. The relationship between carbon release (\(\Delta M_{\text{new}}\)) and observed excursion (\(\Delta\delta^{13}C\)) can be approximated by

\[
\Delta M_{\text{new}} = \frac{M_{\text{initial}}}{(\delta^{13}C_{\text{new}}/\Delta\delta^{13}C) - 1},
\]

where \(M_{\text{initial}}\) is the initial inventory of carbon in the readily exchangeable surface reservoirs (atmosphere, ocean and terrestrial biosphere) and \(\delta^{13}C_{\text{new}}\) is the isotopic composition of the carbon source. However, two main problems are encountered in this. Firstly, the full magnitude of the carbon isotopic excursion is rarely (if at all) recorded in marine carbonates owing to concurrent dissolution (a consequence of ocean acidification and reduced carbonate preservation [44]).
while records contained within terrestrial plants may be affected by changes in atmospheric CO$_2$ concentrations, local temperature and precipitation (e.g. [39]). The second problem is the multiplicity of potential sources of carbon for the PETM event, as different carbon sources have different isotopic signatures. Sources hypothesized to date include: comet impact [45], volcanic carbon release associated with the North Atlantic Igneous Province [46,47], large-scale peat fires [48], weathering of organic-rich desiccated epeiric sea sediments [49] and CH$_4$ release, either thermogenic CH$_4$ associated with North Atlantic volcanism [50], or biogenic CH$_4$ from destabilized methane hydrate within ocean sediments [51]. The sensitivity of the total carbon release to the source (equation (3.1)) is illustrated in figure 2.

A second independent constraint can be placed on the total carbon release by interpreting the variation across the event in the relative abundance of calcium carbonate (CaCO$_3$) mineral particles in deep-sea sediments. The basis for this is the tendency for CaCO$_3$ to dissolve and not be preserved in accumulating
seds as CO₂ is added to the ocean and sea water becomes increasingly acidified [53]. Estimates for the total carbon release that best explains observed changes in the CaCO₃ content of marine sediments are somewhat model dependent (depending on e.g. how much CaCO₃ is assumed to be present in sediments globally immediately prior to the event) and range from 3000 [54] to 6800 Pg of carbon (PgC) [55]. This would imply either a dominant biogenic CH₄ source (3000 PgC at −50‰) or a dominant organic matter (or thermogenic CH₄) contribution (6800 PgC at −22‰). This calculation is also influenced by the assumed time scale of carbon release [55]. As the duration of carbon release approaches and exceeds the carbonate buffering time scale of the ocean (ca 2–8 kyr [53]), more total carbon is required to create the same degree of CaCO₃ dissolution in deep-sea sediments. The much longer (ca 100 kyr) carbon isotopic response time of the system dictates that slower, larger releases must have a less isotopically depleted source. Other effects such as changes in ocean circulation [56] and mixing of the sediments by bioturbation [44] can also affect the calculation.

The time scale of the carbon release at the PETM is particularly important in the degree to which the PETM can provide analogue insights into anthropogenic change. For instance, it is notable that marine organisms have experienced substantial secular variability in their environment over geological time scales, with surface ocean, pH conditions likely to have been some approximately 0.6–0.7 pH units lower during the Cretaceous and Jurassic compared with modern [43,57], a pH change comparable to or exceeding projected future ocean acidification [58]. Yet, calcareous plankton originated, diversified and proliferated during this time. However, this is a false comparison as not only can individuals and ecosystems adapt and evolve to changing climate and ocean geochemistry on ‘long’ (million year) time scales, but also the carbon saturation state of the ocean, which determines the thermodynamic ‘ease’ by which marine organisms can produce carbonate shells and skeletons, is generally well regulated on geological time scales [57]. In contrast, only events involving geologically abrupt (less than 10 kyr) releases of CO₂ can (temporarily) overwhelm the buffering afforded by marine sediments to produce a coupled decline in both pH and saturation state (as is occurring now).

Current estimates of the time scale of carbon release at the PETM are of a carbon input that was essentially complete within 10 kyr [36,59], with much of this potentially released within just a few kiloyears [54]. Placing further constraints on the rate of release and warming at the onset of the PETM is particularly difficult because changes in climate (and weathering) and carbonate dissolution associated with the onset of the PETM strongly affect sedimentation rates and hence available age models. However, a main release taking place as rapid as over 1 kyr (or less) cannot be ruled out geologically, although most of the hypothesized carbon sources would be unable to deliver sufficient carbon to the ocean or atmosphere this rapidly.

In conclusion, while the PETM represents a greenhouse-gas-driven global warming and ocean acidification transient event, with a total carbon release comparable to available fossil-fuel reserves, the rate of climatic and ocean geochemical changes is likely to have been an order of magnitude slower. Hence, it would be prudent to take the observed biotic impacts and ecosystem disruption at the PETM as a lower estimate of potential future impacts rather than a direct analogue.

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(c) Miocene warm intervals

There are a number of intervals within the Miocene epoch that are generally considered to have been warmer than the present-day (i.e. the Mid-Miocene Climatic Optimum and the Late Miocene; [60–63]). Continental position and topography were changing during the epoch, developing towards a modern configuration/condition by the Late Miocene. These changes/developments included the presence of a Central American Seaway (CAS), a marine encroachment from the south into Argentina, a large extension of Eurasia into the Arctic Sea to approximately \(80^\circ\) N, the large Pannonian Lake in central Europe and a wider Indonesian Seaway [64].

The Miocene represented a crucial epoch of uplift and the generation of arid regions [65,66]. The uplift of the Himalayas from a relatively low Tibetan Plateau (1–3 km) in the Late Oligocene to an average height of 4–5 km in the Late Miocene (approx. 9 Ma) had effects on global atmospheric circulation, weathering rates and the Asian Monsoon [65,67–70]. The Andes may have been at half their modern height at 10.7 Ma (approx. 1800 m) and have since been uplifting at 0.2–0.3 mm yr\(^{-1}\) [71]. The Rocky Mountains of western North America are a product of several orogenic events, the most recent of which was the Laramide Orogeny that is dated to the Late Cretaceous to Palaeocene [72]. Subsequent to this major event, the Colorado Plateau has been uplifted by nearly 2 km since the Cretaceous [73]. Estimates of the exact timing of the uplift and the rate are still unresolved, but recent work focusing on the Colorado Plateau suggests that a change in the dynamic topography of 400–1100 m has occurred in the last 30 million years [74,75]. The Alps in the Early to Mid-Miocene were merely islands between the Paratethys and western Tethys Seas, being at an estimated height of less than 1800 m, then major uplift occurred after 14 Ma until present [76].

CO\(_2\) levels for the Miocene have been estimated using boron isotopes [77], alkenones [8], stomatal indices [6,78], pedogenic carbonate [33] and GEOCARB mass-balance modelling [79]. These techniques estimate CO\(_2\) to range between Last Glacial Maximum values and approximately \(2 \times \) PAL [6,8,77–79], although pedogenic carbonates used to estimate CO\(_2\) go as high as 1170 ppmv at 10 Ma [33]. It is important to note that this dynamic range is not supported by all CO\(_2\) proxies. Most notably, alkenone-based CO\(_2\) reconstructions for the Miocene reach approximately pre-industrial concentrations by the beginning of the epoch and maintain this concentration throughout, even through notable warm intervals such as the Mid-Miocene Climatic Optimum, which was arguably the warmest phase in Earth history over the last 15 million years. This has led some to suggest that Miocene climate change was not related to atmospheric CO\(_2\) variations [8,80,81], which, if true, would automatically disqualify the epoch for consideration as an analogue. However, this is far from clear, since newer CO\(_2\) estimates, such as those derived from stomatal indices, are consistent with the coevolution of Miocene climate and CO\(_2\) [6]. The apparent discrepancy/inconsistencies between CO\(_2\) proxies for the Miocene require resolution before warm intervals in the Miocene can be fully explored in terms of analogue issues (also see caveats associated with Pliocene interglacials in §3d that are relevant to the Miocene as well).
(d) Pliocene warm intervals

Most recent climate-model predictions [21,82] suggest that during warm ‘interglacials’ of the Pliocene epoch, global annual mean temperatures were 2–3°C higher than during the pre-industrial interval. During these warm interglacials, sea levels were higher than today (estimated to be 10–30+ m), meaning that global ice volume was reduced [83]. There were large fluctuations in ice cover on Greenland and West Antarctica, and during the interglacials, they may have been largely free of ice [84–86]. Some ice may also have been lost from around the margins of East Antarctica, especially in the Aurora and Wilkes sub-glacial basins [87]. Coniferous forests replaced tundra in the high latitudes of the Northern Hemisphere [88], and the Arctic Ocean may have been seasonally free of sea ice [89–92]. The most recent and detailed estimates of the CO₂ concentrations in the atmosphere range between 280 and 450 ppmv [7,9]. The Mid-Pliocene warm period (mPWP; 3.26–3.025 Ma BP; time scale of Lisiecki & Raymo [93]) is a particularly well-documented interval of warmth during the Pliocene, with global datasets of multi-proxy SSTs, bottom water temperatures, vegetation cover, topography and ice volume readily available as boundary conditions for global climate models [94,95].

Warm intervals of the Pliocene, and particularly the mPWP, are attractive targets to examine analogue issues, and many parallels have been drawn between the predicted Pliocene and twenty-first century climate anomalies from present-day, particularly in terms of (i) the change in annual mean global temperature [96,97], (ii) changing meridional surface-temperature profiles showing a strong polar amplification of the warming [88,90,98,99], (iii) changing precipitation patterns and storm tracks [100], and even (iv) hurricane intensity and El Niño Southern Oscillation (ENSO)-event frequency/extra-tropical teleconnections [18,101]. This attraction is made more intense by the fact that the continents had essentially reached their modern position, and the greater confidence in the modern analogue techniques used in palaeoclimatereconstructions, because the biotic assemblages and relationships are much more similar to modern than in previous warm intervals [88,102].

However, caveats remain in using warm periods of the Pliocene as a guide to future climate and environmental change. Although the continents had reached their modern position, there may still have been significant differences in topography that would have a bearing on the local and regional, perhaps even global, sensitivity of the climate system to higher than pre-industrial concentrations of CO₂ in the atmosphere. For example, there is uncertainty surrounding the precise nature and timing of the evolution of the western Cordillera of North and South America (e.g. [103,104]). The uplift of the Andes and Rockies through the Pliocene had significant effects on local and regional climate and environments in affecting atmospheric circulation (figure 3a).

A number of critical ocean gateways/sills and throughflows also evolved during the Pliocene (Bering Strait, the Greenland–Scotland Ridge/Indonesian through flow and the CAS). Arguably the most significant of these is the CAS, which shoaled and finally closed during the Pliocene, cutting off the tropical connection between the Atlantic/Caribbean Sea and the Eastern Equatorial Pacific. The precise timing of this closure, and whether or not it was a simple history of closure or more a complex history with numerous breaching events, remains an
area of ongoing research (e.g. [105]). What is more certain, from an ocean and coupled ocean–atmosphere modelling perspective, is the significant impact that an open CAS would have had on the meridional overturning circulation and global climate during the Pliocene. In HadCM3, the response of opening the CAS is bipolar with a strong cooling in the Northern Hemisphere matched by an equal warming of the Southern Hemisphere ([106]; figure 3b). This is linked to a dramatic change (weakening) in the meridional overturning circulation in the Atlantic with an open CAS. Such a response has been noted in a number of other models of various complexities (e.g. [107]). If the CAS was open during much of the Pliocene, the regional effects of higher than pre-industrial concentrations of CO₂ could then be expressed differently to the future climate scenario when the

Figure 3. HadCM3 sensitivity tests for the mPWP: (a) difference in annual mean surface air temperature (SAT; °C) between two experiments, one with the Western Cordillera of North and South America at modern height, the other reduced by 50%; (b) difference in annual mean surface air temperature between an open and closed Central American Seaway experiment; (c) difference in annual mean surface air temperature between a 560 and 280 ppmv CO₂ simulation; (d) difference in annual mean surface air temperature between a future climate-change experiment (560 ppmv CO₂) and an experiment for the pre-industrial era (280 ppmv CO₂); and (e) difference between (c) and (d). (Online version in colour.)
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meridional overturning circulation is more vigorous. This is a further example of how the sensitivity of the climate system to CO$_2$ forcing in the past may have changed.

An obvious solution would be to examine only the interglacials of the later Pliocene when we can be confident that the CAS had closed. Sensitivity experiments with HADCM3 indicate a very similar anomaly in mean annual temperature to a doubling of CO$_2$ in both Pliocene and pre-industrial scenarios (with both pre-industrial and Pliocene simulations initialized with pre-industrial and 2 × PAL CO$_2$; figure 3c,d). However, there are subtle regional variations in the pattern of mean annual warming (figure 3e). For example, the amount of SST warming in response to a doubling of CO$_2$ in the Nordic Seas is subdued in the Pliocene case. The reason is intuitive: even though both Pliocene and pre-industrial 560 ppmv simulations are compared with control simulations using pre-industrial levels of CO$_2$, the Pliocene 280 ppmv scenario is still warmer than the pre-industrial equivalent in this region because of the altered high-latitude vegetation and ice-sheet coverage. This reduces the amount of sea ice in the Pliocene 280 ppmv simulation (compared with the pre-industrial experiment), meaning that when the additional CO$_2$ forcing is applied, the response in the Arctic cannot be strong, since there is less sea ice in the Pliocene case compared with the pre-industrial scenario in the first place; the sea-ice albedo feedback mechanism cannot work as efficiently in the Pliocene scenario.

In the subpolar North Atlantic, the opposite is found, with the Pliocene showing a higher sensitivity to CO$_2$ increases (figure 3e). This results from the changes in the Rocky Mountains that alter atmospheric and ocean circulation patterns [97]. Increases in the subpolar zonal winds increase export of Labrador Sea surface waters and air masses into the North Atlantic. These seem to be particularly sensitive, as this one basin contains both the regional sea-ice winter and summer limits, which both retreat in response to CO$_2$ increases. This high sensitivity is largely confined to the immediate Labrador Sea area in the modern. However, with greater export in both the atmosphere and ocean, the Pliocene North Atlantic also exhibits this increased sensitivity.

A more subtle example is available from North America. Here, the difference in CO$_2$ sensitivity between the Pliocene and pre-industrial adopts a dipole pattern, with the Pliocene showing greater warming in the north and less warming in the south of the continent (figure 3e). This type of dipole pattern is classically observed over North America during an El Niño event. While this is unlikely to be the sole cause of the dipole, regression analyses confirm that teleconnections with ENSO over North America are expressed more clearly in a Pliocene world than a pre-industrial world owing to boundary condition changes in the Pliocene, specifically the lower Western Cordillera of North America (101; S. G. Bonham 2010, private communication). Therefore, it is logical to expect this trend to continue and perhaps to intensify with a doubling of CO$_2$.

The differences in climate sensitivity (Charney Sensitivity) described above are all based on analyses of differences in the global mean response. For warming events during a pre-Quaternary interval to be considered a good analogue for the future, climate sensitivity through the seasons must also be the same. Analysis of HADCM3-predicted seasonal differences between the Pliocene and modern sensitivity to 2 × PAL of CO$_2$ indicates regional differences of more than ± 10°C (figure 4).
Warm interglacials of the Pliocene have many advantages over other time intervals discussed as an analogue for future climate change. Even so, Pliocene interglacials presumably reflected long-term equilibrium to a given ambient CO₂ level, whereas current greenhouse-gas emissions provide a rapid transient forcing on climate. Warm events during the Pliocene are a more robust choice than earlier warm intervals of the pre-Quaternary, yet important unresolved questions over the nature of topography and critical ocean gateways must be resolved through further research. HADCM3 experiments indicate that while the global climate sensitivity of the Pliocene and modern worlds is almost identical, the regional and seasonal expression of the climate sensitivity may have been different.

4. A focus on understanding Earth System Sensitivity and on climate-model evaluation

(a) Understanding and calculating Earth System Sensitivity

The concept of climate sensitivity has been discussed extensively over the last 30 years, in particular, since the 1979 National Research Council report [108]. It is defined as the increase in global temperature owing to a doubling of CO₂ after ‘fast
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feedbacks’, typically acting on time scales of years to decades, in the atmosphere and upper ocean have had time to equilibrate [109]. These fast feedbacks correspond to the physics available in climate models ca 1980 (e.g. [110]), and include, for example, water vapour, snow albedo, sea-ice albedo and clouds. This sensitivity (Charney Sensitivity) remains a very useful benchmark for comparing different climate models in idealized circumstances, and has been one of the central concepts used by the IPCC in their assessments and scenarios for future climate change. The IPCC summarizes studies that have aimed to characterize the uncertainty in Charney Sensitivity (e.g. [111–113]) by stating that ‘[Charney Sensitivity] is likely to lie in the range 2°C to 4.5°C, with a most likely value of about 3°C’. More recent studies have agreed with this assessment (e.g. [114]). These estimates of climate sensitivity have also been used to determine the probable impacts, both environmental and social/economic, of various CO₂ stabilization scenarios (e.g. [115]), or the level of greenhouse-gas emissions consistent with stabilization of the global mean temperature below a certain value (e.g. [116]).

However, owing to insufficient understanding of key mechanisms, a political focus on the next few decades and a lack of the necessary computational resources, these estimates of climate sensitivity have all ignored important feedbacks in the Earth system, which act on longer time scales when the palaeoclimate case studies discussed here become more relevant (typically multiple centuries). For example, it is well known that vegetation changes can provide an important positive feedback on climate change, both in future projections (e.g. [117]) and in past climates (e.g. [118]). Similarly, ice sheets are a key player in long-term climate change, as in the continental-scale ice coverage of North America and northern Eurasia during the Last Glacial Maximum, 21,000 years ago [119]. Other processes such as dust–climate interactions, atmospheric chemistry and non-CO₂ greenhouse gases are also potentially important, but neglected in most current estimates of climate sensitivity owing to a lack of understanding of key mechanisms.

The fact that these additional feedbacks will modify the long-term climate sensitivity is now beginning to be appreciated in the climate-science community (e.g. [120]), and Lunt et al. [21] proposed a new term—that of ‘Earth System Sensitivity’ (ESS) as a long-term and more complete version of Charney Sensitivity. Lunt et al. [21] demonstrated that a combined palaeoclimate modelling and data approach can be used to investigate the concept and significance of ESS. Provided that the CO₂ forcing and palaeoenvironmental boundary conditions (e.g. vegetation and ice sheets) are sufficiently known, they can be prescribed in a climate model, allowing relatively short integration times to reach equilibrium. The model can therefore be used to provide a global estimate of long-term equilibrium surface temperature for the given CO₂ forcing. This temperature estimate can be assessed by additional proxy data (e.g. SST) for those locations where it is available. The initial study of Lunt et al. [21], focusing on the mPWP of the Pliocene epoch, showed that the approach above gave an estimate of ESS 30–50% greater than the traditional Charney Sensitivity.

However, estimates of ESS are far from complete, and no study has attempted to explore the uncertainty or sensitivity of the ESS. Since stabilization targets are often defined in terms of absolute temperature thresholds, it is critical that the difference between ESS and Charney Sensitivity be properly quantified and
explored; this can only be achieved through reference to warm intervals in Earth history. Continental position and topography remain significant uncertainties in using the past to calculate ESS, since the effects of such changes must be quantified and removed from any calculation of ESS. Therefore, it is sensible to focus our efforts to calculate ESS on the warm intervals in Earth history when such changes were minimal, such as interglacial events of the latter part of the Pliocene epoch. Given the enormous amount of careful research and available proxy climate and environmental data, the mPWP is an ideal choice.

(b) Climate and Earth system model evaluation

Along with the assessment of ESS, the use of climate archives to confront climate and Earth system models highlights the value of palaeoclimatology. Climate-model predictions for the future cannot be confronted with observations. How well the models simulate modern climate is routinely evaluated (e.g. [121]), but it is possible to make a case that this does not provide a truly independent test of the models since observational datasets are used in the tuning of the model’s performance, and is implicit in the development of parametrization schemes included in all models. Therefore, an evaluation purely on the basis of observed climate could potentially yield an overly optimistic assessment of the predictive abilities of the models. Interestingly, there is frequently palaeoclimatic data in regions of the world where modern observations to test climate models are rare or absent (i.e. the Arctic and Antarctic), and this had led to some of the most surprising and persistent discrepancies between model predictions and geological reconstructions, for example, continental interior temperatures and high-latitude temperatures during the Cretaceous and Eocene greenhouses being too cold in the models (e.g. [122]). A pattern is emerging where the models generally appear to underestimate the magnitude of climate and environmental change compared with the data, which if true is cause for concern.

Model/data comparisons are not a panacea for pre-Quaternary palaeoclimatology; in fact, it is extremely challenging and technical work and demands the full quantification of uncertainty within the data as well as model estimates. In the pre-Quaternary, this is possible to achieve, but it has rarely been attempted. To be done well, such work requires a synergy, understanding and trust between modeller and data collector that remains rare in the community. To prove a model wrong, we must first prove that a model prediction lay clearly outside the envelope of data uncertainty, and then we must know that the error is due to the physics of the model itself, rather than a prescribed boundary condition. While the climate signal (anomaly to present) is large in pre-Quaternary climates, the further back in time, the larger the uncertainties become. Furthermore, the data community should appreciate that even if problems are identified in a single palaeoclimate simulation, this does not prove that the underlying physics of the models are flawed. A model can only be as good as the geological boundary conditions that drive it, and when data/model discrepancies occur, it is quite normal for the problem to be traced to an initial geological condition (e.g. CO₂, palaeogeography/topography/bathymetry). This is the real test and challenge to the usefulness of data/model comparisons in the pre-Quaternary; can the geological community provide sufficiently robust and
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constrained boundary conditions to the climate modellers so that the ‘blame’ for data/model mismatches can *unequivocally* be placed at the foot of the models and their physics?

Those who also model pre-Quaternary climates now recognize that more must be done to quantify uncertainty. The Quaternary modelling community has devoted time and effort to quantifying uncertainty in model predictions, but this is only in its infancy for the pre-Quaternary (e.g. [94,123]). Examples of multi-model ensembles (MMEs) and intercomparisons and perturbed physics ensembles (PPEs) are available in Quaternary Science (e.g. [124,125]), but with the exception of the Pliocene Model Intercomparison Project (PlioMIP; [94]) and Quantifying Uncertainty in Climate Model Predictions for the Pliocene (Plio-QUMP; [123]) are absent in the pre-Quaternary. Without MMEs and PPEs, combined with ensembles exploring the uncertainty generated by geological boundary conditions for the pre-Quaternary, it is not possible to assess the true predictive abilities of models.

5. Conclusions

Great societal benefit would come from the identification of a true pre-Quaternary geological analogue for future climate change. This would enable science to understand how regional climates and environments will respond to climate change during this century. It would provide a natural laboratory for the evaluation of numerical models of climate and the Earth system that we rely on to produce accurate predictions of future change. To be an analogue, a past warm interval in Earth history must be a result of increased concentrations of greenhouse gases in the atmosphere (compared with the pre-industrial era); the climate sensitivity of the Earth must also be demonstrably the same as predictions for the future, and the regional or global pattern of climate and environmental change in the past must not be modified (amplified/dampened) owing to changes in continental position or orography. The past warm interval must also be a response to a transient greenhouse-gas forcing rather than being a result of a long-term equilibrium condition to greenhouse-gas concentrations.

Through the application of coupled ocean–atmosphere models, the most probable candidates (warm intervals) with the pre-Quaternary record of the last 100 million years were tested, and we conclude that there is no satisfactory pre-Quaternary geological analogue to a future greenhouse warming. While Earth history fails to provide a true and direct analogue, it does provide many examples of how Earth responded to variations in atmospheric greenhouse-gas concentrations in the longer term (multiple centuries), and therefore illuminates the sensitivity of the Earth system, rather than simply the atmosphere and surface oceans in isolation, to greenhouse gases. Given the long residence time of CO$_2$ in the atmosphere, this is essential if we are to constrain reliable scenarios for emission reductions that enable us to avoid dangerous levels of climate change beyond 2100. Warm climates in Earth history also provide a tractable way of evaluating the predictive abilities of the current generation of climate and Earth system models. This is essential if we are to have confidence in their predictions of change in the future and for science to be able to place additional uncertainty limits on the predictions from such complex models.

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