Sensitivity of the Palaeocene–Eocene Thermal Maximum climate to cloud properties

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The Palaeocene–Eocene Thermal Maximum (PETM) was a significant global warming event in the Earth’s history (approx. 55 Ma). The cause for this warming event has been linked to increases in greenhouse gases, specifically carbon dioxide and methane. This rapid warming took place in the presence of the existing Early Eocene warm climate. Given that projected business-as-usual levels of atmospheric carbon dioxide reach concentrations of 800–1100 ppmv by 2100, it is of interest to study past climates where atmospheric carbon dioxide was higher than present. This is especially the case given the difficulty of climate models in simulating past warm climates. This study explores the sensitivity of the simulated pre-PETM and PETM periods to change in cloud condensation nuclei (CCN) and microphysical properties of liquid water clouds. Assuming lower levels of CCN for both of these periods leads to significant warming, especially at high latitudes. The study indicates that past differences in cloud properties may be an important factor in accurately simulating past warm climates. Importantly, additional shortwave warming from such a mechanism would imply lower required atmospheric CO₂ concentrations for simulated surface temperatures to be in reasonable agreement with proxy data for the Eocene.

1. Introduction

Simulating warm climates of the Earth’s past has been a challenge for the climate modelling community for decades [1,2]. Using proxy estimates of atmospheric greenhouse gas concentrations in fully coupled climate models yields simulated polar temperatures that are
often too cold, see, for example, Lunt et al. [3] for a multi-model comparison with proxy data. An additional challenge is to warm the polar regions without excessively warming the equatorial region [2], a problem that exists in spite of polar feedback processes operating in the climate system. Simulating past warm climates with a fully coupled general circulation model (GCM) is of great importance given the current and projected rise in atmospheric carbon dioxide (CO2). Projections indicate that if humans continue to burn fossil fuels at the current rate, then atmospheric CO2 levels will reach 800–1100 ppmv by the year 2100 [4,5]. It has been tens of millions of years since these concentrations of CO2 have existed in the Earth’s atmosphere. There are certainly clear differences between the Palaeocene–Eocene Thermal Maximum (PETM) climate state and the present and projected near-future climate warming. Despite these significant differences, there is still a need to better understand how the Earth’s climate processes function in differing climate regimes. Thus, studying the Earth’s warm past climates such as the PETM provides rich observational and modelling opportunities to better understand how the Earth operates in a warm climate regime [6,7].

Over the years, many physical mechanisms have been proposed to solve the low Equator-to-Pole thermal gradient problem in climate models. The basic challenge has been to find means of warming the polar regions more than warming the tropics under Eocene conditions (i.e. in the absence of strong snow and sea-ice and terrestrial ice feedbacks), since solely increasing greenhouse gases significantly warms both equatorial and the polar regions (e.g. [1]). Past proposed climate mechanisms include: increased ocean heat transport [8], polar stratospheric clouds related to enhanced atmospheric methane (CH4) [9], increased deep cloud convection at high latitudes with associated longwave cloud radiative forcing [10] and opening passageways in the Arctic [11], to name a few. It has also been argued that tropical temperatures may have been higher than previously considered, which would allow for a purely enhanced greenhouse gas explanation for warmer climates [2,12,13].

This study explores the role of another mechanism that may have operated in the deep past. It has been pointed out that aerosol properties were no doubt significantly different in deep time [14]. Specifically, Kump & Pollard [14] considered the role of reduced cloud condensation nuclei (CCN) for the warm equable climate of the Cretaceous. They found that lower levels of CCN led to a considerable warming of polar regions relative to warming in the tropics owing to associated changes in cloud properties. This study presents results from a coupled climate model that explores the possible roles of enhanced greenhouse gas concentrations and sensitivity to reduced CCN for pre-PETM and PETM climates. The exact properties of aerosols in the Eocene relative to the present are unknown. However, given that the climate of the Eocene was very different from that of today, it is probable that aerosol properties were different, since aerosol properties are tied to phenomena such as: vegetation type and distribution (organic aerosols, biomass burning), desert regions (dust aerosols), surface wind patterns (sea-salt aerosols) and the ocean productivity (emissions of dimethylsulfide). Thus, the motivation here is to perform a sensitivity study to see whether aerosol–cloud effects could play an important role for the climate of the Eocene.

Here, present-day observations from pristine regions (i.e. regions far from human pollution, but still affected by natural sources) are used to constrain cloud properties, given that currently there are no observations to accurately constrain Eocene aerosol properties. The study also explores the sensitivity of the results to the assumed drop number concentration over continents, since, even in present-day conditions, there is an observed difference in natural aerosols from marine to continental regions. Note that we are not suggesting that the Eocene had aerosol properties identical to the modern pristine values. We are using the modern pristine aerosol properties as a sensible starting point for this sensitivity study.

Additionally, reduction in low-level cloud cover because of lower CCN leads to increased shortwave heating of the Earth’s surface, which has important implications for estimates of global carbon cycle budgets for past warm climates. Presently, models assume high CO2 concentrations that enhance the Earth’s greenhouse effect and warm the climate system. If aerosol–cloud effects
were significantly different in the past, then less CO2 would be required to create a similar warm climate state. This is significant, since assumed CO2 levels for past warm climates are often higher than those estimated by carbon cycle modelling [15].

This study uses the fully coupled atmosphere–ocean–land–sea ice Community Climate System Model (CCSM) configured for the Early Eocene palaeoclimatic conditions. Simulations are presented for climate conditions representative of both pre-PETM and PETM time periods. The PETM simulation represents climatic conditions for the peak of the warming event, while we use the term pre-PETM for the simulation representing conditions prior to the warming event. The present study does not attempt to simulate the temporal transition across the event owing to computational limitations in carrying out a simulation extending for tens to hundreds of thousands of simulated years.

The benefit of considering the Early Eocene compared with periods in the deeper past is that palaeoproxy data for temperatures for this time period cover a significant latitudinal range. These data provide extensive evidence for extreme warmth at high latitudes [16], with a very active hydrological cycle [15]. All of these reconstructions are signatures of a very warm climate regime due to elevated greenhouse gases. The data are also suggestive of mechanisms—positive feedbacks—that amplify the initial greenhouse radiative forcing [17,18].

The study is organized as follows: §2 describes the model configuration and experimental design of the PETM and pre-PETM simulations, §3 presents the results from these simulations and compares simulated surface temperatures with various palaeo proxy data, §4 explores the implications of the present work for understanding the Earth’s sensitivity to increased levels of CO2 and §5 summarizes the findings of the present study.

2. Model configurations and experimental design

This work employs CCSM v. 3 (CCSM3) [19], a fully coupled climate model with active atmosphere, ocean, sea-ice and land components. Lengthy near-steady-state simulations require considerable computational resources so the low-resolution version of CCSM3 [20] is used for all simulations. The atmospheric and land components of the CCSM3 employ an Eulerian spectral dynamical core of T31 (implying an equivalent horizontal resolution of $3.75^\circ \times 3.75^\circ$) with 26 vertical levels in the atmosphere. The ocean and ice components use a nominal $3^\circ$ horizontal resolution with 25 oceanic depth levels. Further modifications include a marginal sea parametrization over the Arctic Ocean basin to ensure reasonable salinity values over long equilibrium runs (see [21] for further details). The equilibrium climate sensitivity of this version of the model is 2.5°C warming for a doubling of CO2 from present-day CO2 concentrations. This sensitivity is at the lower end of the canonical range of 2.1–4.5°C [4] in climate sensitivity. Using a model with higher climate sensitivity would require lower greenhouse concentrations to arrive at a similar climate simulation.

The model configuration employs recent reconstructions [21,22] of the Middle Eocene palaeogeography, palaeotopography and ocean bathymetry. Specification of the spatial distribution of vegetation follows [22]. The vegetation specification is the same for both PRE-PETM and PETM simulations, i.e. there is no vegetation feedback across the PETM event. Recent studies suggest CO2 levels for the PETM that may lie in a range from approximately 1700 to 2250 ppmv [23,24]. But it is fair to state that wide uncertainty exists in the actual CO2 concentrations during both pre-PETM and PETM climate states. Increases in atmospheric CH4 concentrations have also been proposed for the PETM because of the observed negative carbon isotope excursion in $\delta^{13}C$ (see fig. 3 in [25]). The assumed PETM atmospheric concentrations of CO2, CH4 and N2O are 2250 ppmv, 16 ppmv and 275 ppbv, respectively. The level of atmospheric CO2 is based on the work by Panchuk et al. [24] (L. Kump 2009, personal communication), which used a geochemical model of intermediate complexity to infer atmospheric CO2 levels consistent with geochemical markers. The level of CH4 employed for the present work comes from a modelling study on the effects of CH4 release during the PETM [26]. The pre-PETM atmospheric concentrations of CO2, CH4 and N2O are
1375 ppmv, 760 ppbv and 275 ppbv, respectively. The pre-PETM level of CO₂ was obtained by taking the PETM simulation and reducing CO₂ levels until the global annual mean temperature was reduced by approximately 5°C in order to agree with the global estimate of observed temperature change. Here the use of a pre-industrial level of atmospheric CH₄ is no doubt low for the warm Early Eocene, given the moist environment that would have allowed for more wetland regions. Given that there are no observational data to constrain CH₄ concentrations for this time period, a conservative assumption is made concerning the pre-PETM CH₄ levels. Note that sustained levels of CH₄ require a continual release source of CH₄ into the atmosphere given the relatively short CH₄ lifetime of approximately 12 years. This issue is addressed in §4.

The change in CCN was incorporated into the atmospheric model by altering both the liquid cloud drop number and the effective cloud drop radii. As noted, there are no observations of CCN or cloud microphysical properties for deep time periods. Thus, the present study should be viewed as a sensitivity study with regards to the effects of cloud microphysical properties on past climates. Framed as a sensitivity study, this work will make simple assumptions about aerosol and cloud properties for the Eocene. Given this assumption, an observational composite (see fig. 5 in [27]) of cloud drop number for present-day remote pristine regions is used to set the cloud drop number in the simulations. For present-day pristine regions, the observed cloud drop number concentration is around 50 drops per cm³ for liquid water clouds. Lower CCN leads to fewer cloud drops that grow to larger sizes. Observations indicate that effective cloud drop radii for pristine clouds are approximately 17 μm. For the sensitivity studies, the model configurations assume that all liquid clouds have present-day pristine cloud drop properties. Once these properties are prescribed the cloud microphysical and radiative parametrizations in the atmospheric model respond to these cloud drop properties, i.e. cloud rainout processes and shortwave absorption change owing to the change in cloud drop properties. Decreasing cloud drop number leads to increased precipitation rate and shorter cloud lifetime, which in the time mean implies a reduction in cloud cover. Increased cloud drop size leads to more shortwave absorption in clouds, which dissipates clouds. Fewer low-level clouds results in more shortwave radiation reaching the surface. Since low-level liquid water stratiform clouds predominate at high latitudes, the reduction in these types of clouds leads to a preferential warming of polar regions.

To summarize, for pre-PETM and PETM simulations, liquid water cloud properties are changed as follows: the cloud drop density is set to 50 cm⁻³ everywhere, as compared with the present-day prescription of 400 cm⁻³ for continental regions, 150 cm⁻³ for ocean regions and 75 cm⁻³ over sea-ice and snow-covered regions. The effective liquid cloud drop radius is set to 17 μm everywhere, as compared with the present-day assumed values of 8 μm over land and 14 μm over ocean, sea-ice and snow-covered regions. The role of continental versus marine cloud drop differences is explored in a companion PETM simulation in which the cloud drop density is set to 400 cm⁻³ and the cloud drop effective radius to 10 μm over continental regions.

All simulations employ a constant uniform pre-industrial aerosol optical depth representing a general background aerosol concentration. The simulations also assume a fixed geothermal heat flux into the oceans of 0.088 W m⁻². All simulations assume a 0.487% reduction in solar luminosity for the Early Eocene time period and the orbital parameters are those used in [21].

(a) Slab ocean model sensitivities

Based on these modelling assumptions, four factors account for the warm simulated climate of the PETM: enhanced CO₂ concentrations, enhanced CH₄ concentrations, the absence of ice sheets and a reduction in CCN, i.e. lower liquid cloud drop number and larger cloud drop effective radius. In order to assess the relative warming contribution from three of these factors—CO₂, CH₄ and CCN effects—sensitivity climate simulations were carried out with a version of the CCSM3 that uses a slab mixed layer ocean component in place of the full dynamical ocean. Note that all of these simulations assume the absence of terrestrial ice sheets and use the palaeogeography of the Middle Eocene. In this version of the model the ocean heat transport for the mixed layer model is based on the fully coupled PETM simulation. The advantage of the slab ocean model
Figure 1. Zonal mean change in: (a) PETM annual surface air temperature (°C) owing to changes in CO₂ from 280 ppmv to 2250 ppmv (dashed line), changes in CH₄ from 760 ppbv to 16 ppbv (dotted line) and changes in cloud properties (solid line); (b) annual surface air temperature normalized by respective global mean change in surface air temperature due to changes in CO₂ from 280 ppmv to 2250 ppmv (dashed line), changes in CH₄ from 760 ppbv to 16 ppbv (dotted line) and changes in cloud properties (solid line). Normalized value = [(Tₛ)exp - (Tₛ)control] / ⟨Tₛ⟩control, where [] indicates zonal mean and ⟨⟩ indicates global mean.

for sensitivity studies it can be run to a steady state within only 40 simulated years. In the first simulation, only CO₂ levels were decreased to a pre-industrial value of 280 ppmv. In the second simulation, only CH₄ levels were decreased to a pre-industrial level of 760 ppbv. In the third calculation, cloud drop number and effective drop radii were changed to present-day (i.e. polluted) values. The change in annual zonal mean surface air temperature from these three simulations (figure 1a) indicates that the largest warming effect is because of CO₂ with polar warming of 12°C. The second largest warming is because of changes in CCN-induced liquid water cloud properties and yields a 7–9°C warming at the Poles, while the third largest contributor to warming arises from increased CH₄ with a modest 4°C warming at the Poles. In order to eliminate the effects of a different base state for these results, the normalized change in zonal mean surface air temperature is shown (figure 1b), in which the zonal mean changes are normalized by their respective global mean changes. If similar amplification processes are present in all three simulations, then the three curves should cluster together. However, in the Northern Hemisphere polar regions, there are still significant differences between the CCN simulation and CO₂ and CH₄ simulations. These differences are due to the inherent differences in radiative forcing for the CCN simulation compared with the greenhouse gases simulations. The radiative forcing from the CCN sensitivity simulation arises from two factors: (i) changing the cloud drop number to 50 cm⁻³ lowers the liquid water path in the clouds and decreases cloud area (this effect arises from an increase in precipitation efficiency and hence a decrease in cloud lifetime) and (ii) increasing the cloud drop size to 17 μm decreases the single scattering albedo of the clouds, which leads to more shortwave absorption in the clouds. This, in turn, leads to a burn-off of low cloud cover. Both of these effects result in more shortwave radiation reaching the surface. Additional simulations have been performed to isolate these two cloud effects to see which dominates high-latitude warming. These simulations show that the change in cloud drop size—enhanced shortwave cloud absorption—dominates the CCN forcing simulations. At high latitudes, the CCN effect will play a major warming role during late spring to early autumn, i.e. when shortwave radiative forcing is high. However, at high latitudes during local winter conditions, the shortwave CCN effect will not be active. Analysis of the simulations indicates that high-latitude local winter warming is due to three factors: (i) an overall increase in tropospheric water vapour leading to an enhanced
greenhouse warming, (ii) an increase in upper tropospheric cloud cover leading to an increase in longwave cloud forcing, and (iii) an open Arctic ocean basin that stores more energy through the winter than an ice-covered Arctic, which does not store energy through local winter. These results indicate that there are differences in response from the CCN effect compared with the standard greenhouse effect. This is essentially because the CCN effect is affecting low-level stratiform cloud cover, which predominates the high latitudes coupled with the seasonal asymmetry in solar radiation reaching high latitudes. These two factors lead to higher forcing at high latitudes than what is obtained from greenhouse forcing.

(b) Fully coupled simulations

The fully coupled PETM simulations involved a multi-stage spin-up process. First, an initial PETM simulation was carried out by running the fully coupled CCSM3 for 600 years from an ocean-only simulation that had been run in an accelerated mode for 8300 years. This was done in order to obtain a more representative deep ocean state. The pre-PETM fully coupled model was run from this simulation with the pre-PETM CO$_2$ and CH$_4$ concentrations and pristine CCN conditions. The pre-PETM state was run for another 1400 years. The ocean state for this simulation has vigorous ventilation, which means the deep ocean comes into an approximate steady state. CO$_2$ and CH$_4$ concentrations were then set to their PETM levels and the PETM simulation was initialized from the end of pre-PETM simulation and run for another 1660 years, after which the net energy imbalance of the coupled PETM climate system was less than 0.4 W m$^{-2}$. Thus, the deepest layers of the ocean in this simulation are still not in a steady state. However, the surface temperatures at the end of this simulation have reached near-steady-state conditions with trends in zonal mean surface temperature less than $10^{-4}$ °C per year. All results are based on 50 year averages from the end of the pre-PETM and PETM CCSM3 simulations.

3. Model results

(a) Simulated global mean state

The global annual mean sea surface temperature (SST) for the fully coupled PETM simulation is 32.3°C. To obtain an estimate of the observed PETM global mean SST, it is assumed that the zonal surface temperature can be represented by the function $A + B \cos(\varphi_{\text{lat}})$, where $\varphi_{\text{lat}}$ is palaeolatitude [5]. Two observational points [28,29] at 36° N and 75° N palaeolatitude with SSTs of 33°C and 25°C, respectively, determine the coefficients A and B. Analytically integrating the expression yields an observed estimated global annual mean SST of 33°C for the PETM. Thus, the simulated global annual mean SST is in good agreement with the first-order observational estimate. Note that the simulated global mean surface temperature (land plus ocean) is 31.9°C, so the SST value provides a very good estimate for the global mean. This agreement between ocean-only and global mean values is applicable in a warm world with little snow or ice cover.

(b) Simulated ocean state

Zonal annual mean SSTs from the PETM and pre-PETM model simulations exhibit a reduced Equator-to-Pole temperature gradient compared with the present-day simulation (figure 2). Note that the present-day CCSM3 simulation is in good agreement with observed SSTs with a slight cold bias in the polar regions. As noted from the slab ocean sensitivity simulations, the dominant factors contributing to the amplified polar warmth are increased atmospheric CO$_2$ concentration and the change in cloud properties associated with reduced CCN (see figure 1). Increased levels of atmospheric CH$_4$ contribute one-third of the warming relative to warming from increased CO$_2$. Zonal mean PETM SSTs at approximately 70° north and south are approximately 20°C, in good
Figure 2. Zonal annual mean SSTs (°C) from the modern (black line), PETM (red line) and pre-PETM (blue line) CCSM3 simulations.

Table 1. Comparison of modelled PETM surface temperatures (°C) for various locations including marine and terrestrial sites. Data are from table 1 in [28–31], and [32]. Numbers in parentheses () are mean annual model values and values in square brackets [ ] are summer model values.

<table>
<thead>
<tr>
<th>Palaeo latitude</th>
<th>Surface temperature (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>75° N (Arctic)</td>
<td>25 (17) [28]</td>
</tr>
<tr>
<td>47° N (Big Horn, WY)</td>
<td>20–26 (25) [40]</td>
</tr>
<tr>
<td>~ 36° N (NJ coast)</td>
<td>33 (32) [37]</td>
</tr>
<tr>
<td>~ 6° N (Columbia)</td>
<td>38–40 (38) [37]</td>
</tr>
<tr>
<td>55° S (New Zealand)</td>
<td>33 (23) [28]</td>
</tr>
<tr>
<td>65° S</td>
<td>25 (23) [28]</td>
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*Terrestrial reconstructions.*

agreement with the proxy estimates at these latitudes (see table 1 for point-wise comparisons with proxy data). Tropical temperatures for the PETM are approximately 40°C. Currently, there are no proxy estimates of SSTs for the deep equatorial marine environment, but tropical and subtropical temperatures approaching the PETM indicate a warm climate with surface temperatures in the range 35–41°C (see figure S1 in [2]).

The simulated spatial distribution of PETM tropical and sub-tropical SSTs (figure 3a) is similar to the modern-day pattern with a warm pool of water in the Indian and western Pacific Oceans and a cold tongue of water in the eastern Pacific. Warm pool temperatures in excess of 40°C exist in the palaeo Tethys region on either side of India, indicating the lack of an ocean thermostat to keep these waters close to present-day values. Warm waters extend far into the extra-tropics with 32°C water off the coast of present-day New Jersey, in good agreement with the proxy data for this region [28] (table 1). At higher latitudes, the PETM Arctic is slightly cold by approximately 8°C compared with the observational value for this region [29]. Note that simulated PETM summer temperatures are in better agreement with the observational estimate (table 1). The largest discrepancy between simulated SSTs and reconstructions occurs at 55° S latitude [30]. Here the model is colder than the proxy data, but again the simulated summer temperature is closer to the observations. Further south at 65° latitude the model agrees with the proxies to within 2°C [31].

The change in SST (figure 3b) (PETM minus pre-PETM) is in good agreement with observed changes at all latitudes (see table 1 in [33] for a compilation of proxy changes in SSTs). Given that the pre-PETM model atmospheric CO₂ level of 1375 ppmv was chosen to ensure an
Figure 3. Geographical distribution simulated: (a) PETM SSTs (°C) and (b) change in SST (°C) from pre-PETM to PETM climate. Numbers in boxes are observed range in temperature changes [33].

approximately 5°C global annual mean change in surface temperature (see §2), this agreement may seem unsurprising. However, the choice of this CO2 level does not guarantee that the spatial distribution of change in temperature will agree with reconstructions at specific geographical locations.

The sea surface salinity distribution for the PETM simulation (figure 4a) shows extremely fresh waters in the Arctic basin, which is in agreement with recently published proxies [15]. One important regional feature of the PETM simulation is high salinity located in the Turgay strait between present-day Europe and Asia. As will be shown below, this feature increases surface water density relative to the pre-PETM simulation, causing sinking in this region. The
Figure 4. PETM simulated (a) sea surface salinity (practical salinity units), (b) evaporation minus precipitation (mm day$^{-1}$) and (c) surface run-off (mm day$^{-1}$).
overall salinity distribution—determined by local water balance between evaporation minus precipitation and run-off (figure 4b,c)—exhibits enhanced net fresh water input at high latitudes compared with the present-day climate and enhanced tropical precipitation. Continental run-off plays an important role in determining salinity levels in coastal regions (compare figure 4a,c), where the largest run-off occurs into the Arctic basin and in the tropical regions.

The location and strength of deep water formation is important for understanding the deep ocean circulation. In particular, changes in deep water formation due to increasing levels of greenhouse gases may have important implications for carbon cycle processes [34,35]. The seasonal cycle of maximum mixed layer depth (figure 5) is an informative measure of the location of deep water formation. Other measures were also used to identify regions of deep water formation occurring in the pre-PETM and PETM simulations, including the seasonal cycle of surface potential density and isopleths of zonal potential temperature and salinity. The conclusions concerning deep water formation using maximum mixed layer depth are supported

Figure 5. Simulated maximum mixed layer depths (m) for (a) January pre-PETM, (b) February pre-PETM, (c) July pre-PETM, (d) August pre-PETM, (e) January PETM, (f) February PETM, (g) July PETM and (h) August PETM.
by all of the above metrics. Two winter months of maximum mixed layer depth for each hemisphere are shown (figure 5), since these seasons are when the densest waters sink to maximum depth. Northern Hemisphere winter pre-PETM maximum mixing (figure 5a, b) occurs in the north Pacific, where water formed in this region penetrates to 4000 m depth. Mixing is very vigorous, as indicated by ideal water ages of less than 50 years at these depths, where ideal age measures the time in years that water at a specific depth was last exposed to the ocean surface. In the Southern Hemisphere winter pre-PETM (figure 5c,d), maximum mixing occurs off the coasts of Australia and Antarctica. Again, ocean ventilation is very efficient for this region. This efficient ventilation for the pre-PETM climate means that the deep ocean is strongly coupled to the ocean surface in high-latitude regions. It also means that the equilibrium time scale of ocean circulation for the pre-PETM ocean state is much shorter than that of the PETM world. The maximum mixed layer depths for the PETM simulation (figure 5e,h) exhibit a very different configuration for deep water formation. The high-latitude Pacific formation sites for both hemispheres no longer exist. Surface warming and fresh water input at high latitudes have significantly reduced the specific density gradients and stratification has essentially shut off high-latitude deep water pathways. However, one region of maximum mixed layer depth remains in the Turgay straight. Here, high surface salinities (figure 4a) cause surface waters to sink to approximately 1000 m depth. This water then spreads out from the Tethys region, forming intermediate waters into the wider Pacific region. Thus, the source location of deep water formation switches in going from the pre-PETM into the PETM climate state, as first suggested by the modelling results of Bice & Marotzke [36] and proxy studies [37–39].

The effects of this shift in water mass formation is reflected in the ocean circulation at 1000 m depth between the PETM state (figure 6a) and the pre-PETM (figure 6b). These figures show a shift from a circulation with stronger and larger gyres to a weaker circulation for the warmer climate. Note that the gyre circulation is also due to a shift in the atmospheric circulation in moving into a warmer climatic state. Figure 6a also exhibits evidence for the flow of warmer water from the Turgay straight out to the Pacific in the PETM state compared with the pre-PETM. The near-surface PETM circulation (figure 7) exhibits surface flow from the Atlantic into the Pacific through the open Panama strait. The simulation also shows signatures of both Kurishio- and Gulf-like currents along the eastern boundaries of present-day Asian and North American continents, respectively. Coastal upwelling driven by along-shore atmospheric circulations is also apparent.

One proposed mechanism for maintaining a low Equator-to-Pole thermal gradient was related to an increase in ocean heat transport [8]. The present simulations find no evidence for this hypothesis (figure 8), in agreement with other modelling studies that have used fully coupled GCMs to study the warm Eocene [40]. Indeed, the warmer climate state of the PETM exhibits less Northern Hemisphere ocean heat transport than either the pre-PETM climate or the present-day simulated climate. Peak PETM ocean heat transport at 20–30° N is approximately 30% less than pre-PETM transport and approximately 45% less than present-day peak transport. Note, however, that ocean heat transport in the Southern Hemisphere is greater than the present-day simulated transport and may contribute to warmer high-latitude temperatures in this region.

Finally, the zonal annual mean vertical thermal structure of oceans for both PETM and pre-PETM simulations (figure 9a,b) is significantly different from the present-day thermal structure (figure 9c). In general, the warm Eocene simulations are more stratified than the present-day ocean structure, in agreement with previous studies (e.g. [34,40]). This stratification is more evident when considering vertical profiles of potential temperature (figure 10a–d) for tropical and northern and southern high-latitude regions. For Northern Hemisphere winter conditions (figure 10a,b), the pre-PETM simulation shows a region in the north Pacific of near-constant temperatures through the depth of the ocean column, indicative of vigorous mixing in this region. In the PETM state, this region of mixing is suppressed. Similarly, for the Southern Hemisphere winter (figure 10c,d), a region of well-mixed water exists in the southern Pacific, which has been suppressed in the PETM
climate state. The PETM ocean temperatures at depths ranging from approximately 1300 to 3400 m for specific locations are generally in good agreement with proxy data (table 2). However, it must be remembered that at these depths the ocean state is still warming in the PETM simulation.

(c) Simulated terrestrial state

The simulated high-latitude terrestrial annual mean surface air temperatures for the PETM (figure 11a) are 15–20°C, in reasonable agreement with the limited proxy data. For example, temperatures for the Wyoming Bighorn Basin are in excellent agreement with reconstructions (table 1) [42]. Simulated surface temperatures in the tropical regions of South America and Africa are excessively high, in excess of 48°C at some points. Such high temperatures would imply severe conditions for the existence of life in these regions. The change in surface air
temperature from pre-PETM to PETM conditions (figure 11b) shows warming of 5–10°C for most regions. This 5–10°C warming in middle North America is in very good agreement with data [6]. For the northernmost region of North America, minimum January temperatures are 8°C, which is supported by the limited palaeobotanical evidence of palm trees at these high latitudes. In general, for the PETM simulation cold month mean surface air temperatures are above freezing at all locations (figure 12a), while for the pre-PETM intercontinental region cold month mean temperatures are below freezing (figure 12b). A comparison of simulated terrestrial surface temperatures with the proxy compilation of Huber & Caballero [12] (figure 13) indicates
general agreement across a range of locations. Given that the reconstructions span a wide range of Early Eocene time periods, with few data representative of the PETM event, agreement is expected to be better for the pre-PETM simulation than for the PETM and indeed this is the case (compare figure 13a, b, respectively).

As noted in the Introduction, specification of a continental cloud drop size density of 50 cm$^{-3}$ may be too low, given the diversity of aerosol sources from vegetation (secondary organics, biomass burning) and changes in surface conditions, e.g. dust loading. An additional sensitivity study was carried out to test how the PETM results depend on this assumption. Using a cloud drop size of 400 cm$^{-3}$ and drop effective radius of 10 μm over continental regions (figure 14) leads to a 2–3°C reduction in continental surface temperature. Thus, inclusion of a difference between continental and marine CCN conditions slightly cools the continents relative to the simpler approach of uniform CCN or cloud drop properties.

In general, the hydrological cycle over land is enhanced in the pre-PETM and PETM simulations compared with the present-day simulation as reconstructions suggest [43]. Run-off in the Arctic also increases from the pre-PETM to the PETM, resulting in very fresh water for the entire basin—a feature of the Arctic supported by proxy data [15]. Overall the hydrological cycle is more vigorous for the warm Eocene simulations than for the present-day simulation. The change in precipitation over land (figure 15) from the pre-PETM to the PETM climate state indicates that most regions experience an increase in rainfall. In the northern part of North America, precipitation increases by 20–50%. However, there are locations that experience a slight reduction in precipitation, e.g. the North American southwest. Reconstructions of changes in precipitation

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**Figure 9.** Zonal annual mean ocean potential temperature (°C) for simulated: (a) PETM, (b) pre-PETM and (c) present climate simulations.
Figure 10. Vertical profiles of potential temperature (°C) for December–January–February average: (a) tropical, (b) northern high latitudes, and for June–July–August average (c) tropical and (d) southern high latitudes. PETM case is solid line; pre-PETM case is dashed line.

Table 2. Comparison of modelled PETM ocean temperatures (°C) with reconstructions for three locations and ocean depths (m). Model results are given in parentheses. Data are from [41].

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<thead>
<tr>
<th>Palaeo latitude</th>
<th>depth (m)</th>
<th>temperature (°C)</th>
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<tbody>
<tr>
<td>∼ 30° S (S. Atlantic)</td>
<td>∼3400</td>
<td>14–15 (15)</td>
</tr>
<tr>
<td>∼ 10 – 15° N (Pacific)</td>
<td>∼2400</td>
<td>13–21 (15)</td>
</tr>
<tr>
<td>∼ 2° N (Eq. Pacific)</td>
<td>∼1300</td>
<td>13–17 (17)</td>
</tr>
</tbody>
</table>

[42] in North America offer a complex picture of regions of decreased precipitation early after the event, but a general increase in precipitation at the peak of the warm PETM. The seemingly large percentage change in precipitation in the central African region is somewhat misleading given that both the pre-PETM and PETM simulated precipitation in this region is very low (less than 1 mm d<sup>−1</sup>). The percentage change in this general region of low precipitation appears to be related to the poleward shift in the zonal mean circulation, which affects moisture transport into this region.
4. Implications for climate sensitivity

Considerable attention has recently been given to what deep time climates tell us about the Earth’s climate sensitivity [5,44–46]. The traditional or Charney climate sensitivity is defined as the equilibrium warming due to a doubling of CO$_2$ about the modern climate state and assumes that the factors amplifying the initial greenhouse radiative forcing take place on time scales of days to decades. Thus, this sensitivity includes feedback mechanisms involving changes in processes such as water vapour, lapse rate, clouds and sea ice. One measure of the strength of these feedback processes is the so-called climate feedback factor defined as the ratio of the doubled CO$_2$ equilibrium warming (approx. 2.5°C) to the radiative forcing due to a doubling of CO$_2$ (approx. 3.7 W m$^{-2}$), i.e. approximately 0.68°C (W m$^{-2}$)$^{-1}$.
Figure 12. Simulated cold month mean (°C): (a) PETM and (b) pre-PETM.

Figure 13. Comparison of simulated terrestrial surface air temperatures (°C) with the Huber & Caballero [12] proxy database for the: (a) PETM and (b) pre-PETM climates.
Figure 14. Difference in annual mean PETM surface temperature (°C) between a simulation assuming a continental cloud drop density of 400 cm\(^{-3}\) and drop effective radius of 10 μm and a simulation assuming a continental cloud drop density of 50 cm\(^{-3}\) and effective radius of 17 μm.

Figure 15. Per cent change in terrestrial precipitation between the PETM and pre-PETM climate simulations.

As discussed in §3a, the present study arrives at a proxy estimate for the global mean PETM surface temperature of approximately 33°C. Given that the Earth’s pre-industrial temperature was approximately 15°C, the PETM was warmer by 18°C compared with the pre-industrial time period. Assuming that the PETM CO\(_2\) concentration (2250 ppmv) was eight times larger than pre-industrial levels implies a CO\(_2\) radiative forcing of 14.5 W m\(^{-2}\), where the CO\(_2\) forcing is obtained by the method described in [47] employing the atmospheric version of CCSM3 in a fixed SST
configuration. The shortwave forcing from pre-industrial to the PETM time period (55 Ma), owing to the change in solar luminosity, is a forcing of −1.2 W m$^{-2}$, assuming a planetary albedo of 0.27 derived from the PETM simulation. Thus, the net forcing (CO$_2$ greenhouse + solar luminosity) from pre-industrial to the PETM is approximately 13 W m$^{-2}$. Using these estimates for the PETM implies a feedback factor of approximately 18°C/(13 W m$^{-2}$) or approximately 1.4°C (W m$^{-2}$)$^{-1}$.

Thus, the climate sensitivity considering a change in state from the modern to the PETM is two times larger than the modern Charney sensitivity. This larger climate sensitivity has been called the Earth system sensitivity (ESS) [45] and includes feedback processes operating on time scales of decades to many millennia, e.g. ice sheet destabilization, possible methane hydrate release, changes to vegetation and alterations to the global carbon cycle. Note that if a lower PETM CO$_2$ concentration were assumed, then the deduced ESS would be even higher.

What do the palaeoclimate simulations of the pre-PETM and PETM in this study have to say concerning the issue of an enhanced ESS over the Charney sensitivity? In addition to CO$_2$ forcing, the absence of ice sheets and palaeogeography, this study considers two other factors to explain the warm simulated climates of the pre-PETM and the PETM relative to the modern climate state: (i) increased CH$_4$ concentrations and (ii) lower CCN. This sensitivity study suggests that these two additional processes may have played a critical role in enhancing climate sensitivity on long time scales relative to the present-day climate. Note that this argument assumes that CH$_4$ release and changes in aerosol–cloud interactions operate as feedback mechanisms and not forcing factors on these long time scales.

With regards to CH$_4$, Dickens [18] has argued that release of large reserves of CH$_4$ is quite feasible for the warm Eocene. Continued release of CH$_4$ from expanded wetland regions containing bacteria would also have led to sustained levels of atmospheric CH$_4$ [48]. Sustained emissions of CH$_4$ would be required to maintain elevated CH$_4$ levels in the presence of an otherwise short lifetime.

With regards to changes in CCN, this study uses present-day observations for very pristine regions to constrain the model cloud microphysical properties. This assumption is clearly simplistic given the temporal, spatial and chemical variability that exists in real aerosols. Thus, what is presented here is a sensitivity study, which uses present-day knowledge to link CCN and cloud drop number density. Kump & Pollard [14] provided arguments for why CCN would have been lower during the warm Cretaceous. There may be other reasons for lower CCN during past warm climate states, some of which may be linked to atmospheric chemistry. A recent study [49] found that the production of aerosols actually decreases in the presence of certain types of vegetation. This effect is actually opposite to what occurs where the production of secondary organic aerosols increases as a result of the emission of certain biogenic precursors from vegetation. The explanation of these new findings involves the effects of isoprene emissions from vegetation on the concentration of the atmospheric hydroxyl radical, which plays an important role in aerosol formation. In this case, increased warming initiated by increased greenhouse gases would lead to the migration of forests to higher latitudes [50] accompanied by a reduction in CCN with additional warming. This proposal is hypothetical, as such it would be of value to look for particular proxies that could either validate or invalidate this hypothetical biophysical feedback. Note that the conclusions presented here do not depend on this particular proposed mechanism, since this study considers the general sensitivity of the warm climate state to changes in cloud microphysical properties derived from present-day pristine conditions.

To date, modelling past warm climates, such as the Eocene, has focused mainly on two climate factors: greenhouse forcing and enhanced climate sensitivity. For sufficiently high CO$_2$ concentrations, enhanced greenhouse forcing yields a warm climate approximating reconstructions. However, the assumed level of CO$_2$ may be too high compared with palaeo pCO$_2$ proxies. Another solution to this dilemma is to use a model with a higher climate sensitivity, which means that a lower CO$_2$ concentration yields similar agreement with the reconstructions. This work explores another factor that is important to the climate system, i.e. shortwave feedback. This sensitivity study shows that including changes in shortwave forcing—via CCN–cloud interactions—results in additional heating of the climate system, which, in turn,
implies that less CO₂ is required to produce a similar warm climate state. In light of these results, a CCN–cloud mechanism could help alleviate current disparities between assumed CO₂ levels in climate models and those estimated from carbon cycle budget models for past warm climates.

5. Conclusions

This study finds that CCSM3 simulations of the PETM climate that includes enhanced levels of CO₂, CH₄ and lower cloud drop numbers are in good agreement with a wide range of palaeo temperature records. Along with the studies in [12,13], this is one of the few coupled climate model simulations that agrees with much of the proxy data, including polar regions. The simulations show that the climate system is sensitive to the specification of reduced levels of CCN and associated changes in cloud properties. Hence, this may be an important climate factor that needs to be accounted for in simulating past climates. Given these findings, it would be of great value to find means to quantify the aerosol properties that may have existed in past climate states.

In support of previous studies and reconstructions, the study finds that the ocean general circulation shifted between the pre-PETM climate to that of the PETM state. In particular, the sites of deep water formation shifted from the high polar regions in both hemispheres to the mid-latitudes upon entering the warmer PETM climate. Simulated deep ocean temperatures agree well with the limited available data for the PETM. In terms of the terrestrial sites, the PETM simulation is in good agreement with much of the proxy data. Surface temperatures are not excessively cold, with cold month mean temperatures staying above freezing. The simulations also indicate that the hydrological cycles of the PETM and pre-PETM were far more active than present day. In addition, the more active high-latitude hydrological cycle led to an increase in fresh water input into the Arctic basin. Remaining questions are related to the magnitude of surface warming for marine and terrestrial regions in the tropics between 30° S and 30° N. Few data exist to constrain surface temperatures in this region. The model simulations suggest surface temperatures in excess of 45°C for certain regions. It is important to find stricter proxy constraints for these tropical regions.

Finally, this study supports the finding of others that the Earth’s climate system is more sensitive to greenhouse forcing on longer time scales (e.g. [44–46]). In particular, this work finds that the ESS for the PETM or pre-PETM relative to the modern climate is twice as large as the traditional Charney climate sensitivity.

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References


