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The challenge of simulating warmth of the mid-Miocene Climate Optimum in CESM1

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Abstract

The mid-Miocene Climatic Optimum (MMCO) is an intriguing climatic period due to its above-modern temperatures in mid-to-high latitudes in the presence of close-tomodern CO_2 concentrations. We use the recently released Community Earth System

- ⁵ Model (CESM1.0) with a slab ocean to simulate this warm period, incorporating recent Miocene CO₂ reconstructions of 400 ppm. We simulate a global mean annual temperature (MAT) of 18°C, ~4°C above the pre-industrial value, but 4°C colder than the global Miocene MAT we calculate from climate proxies. Sensitivity tests reveal that the inclusion of a reduced Antarctic ice sheet, eastern equatorial Pacific Ocean tempera-
- ¹⁰ ture anomalies, increased CO₂ to 560 ppm, and variations in obliquity only marginally improve model-data agreement. All MMCO simulations have an equator to pole temperature gradient which is at least ~ 10 °C larger than the reconstruction from proxies. The MMCO simulation most comparable to the proxy records requires a CO₂ concentration of 800 ppm. Our results illustrate that MMCO warmth is not reproducible using
- the CESM1.0 forced with CO₂ concentrations reconstructed for the Miocene or including various proposed Earth system feedbacks; the remaining discrepancy in the MAT is comparable to that introduced by a CO₂ doubling. The models tendency to underestimate proxy derived global MAT and overestimate the equator to pole temperature gradient suggests a major climate problem in the MMCO akin to those in the Eocene.
- ²⁰ Our results imply that this latest model, as with previous generations of climate models, is either not sensitive enough or additional forcings remain missing that explain half of the anomalous warmth and pronounced polar amplification of the MMCO.

1 Introduction

The Middle Miocene Climatic Optimum (MMCO 17–14.50 Ma) (Zachos et al., 2008) is a period in Earth's history in which temperatures were significantly warmer in the deep ocean and in mid-to-high-latitudes (Böhme et al., 2007; Pound et al., 2012; Zachos et



al., 2008; Shevenell et al., 2008). These warm extra-tropical temperatures have been hard to reconcile with reconstructed below-modern tropical sea surface temperature (SST) records and boron and alkenone CO_2 reconstructions of 200–280 ppm levels (Pagani et al., 2005; Pearson and Palmer, 2000).

- Recent re-evaluation of the proxy records has led to advancement in our understanding of MMCO warmth. First, the MMCO tropical SST records showing below-modern levels (Savin, 1977; Nikolaev et al., 1998; Bojar et al., 2005) are now understood to have a cool diagenetic bias (Stewart et al., 2004). Excluding these records indicates that tropical SSTs in the Miocene were above modern (Shevenell et al., 2004; You et al., 2004; You et al., 2004).
- al., 2009; LaRiviere et al., 2012). Second, recent leaf stomatal studies reconstruct CO₂ concentrations at the MMCO to be 400–500 ppm (Kürschner et al., 2008) and these results have been confirmed in boron isotope-based reconstructions (Foster et al., 2012) and updated alkenone reconstructions (Zhang et al., 2013).

Nevertheless, even with higher CO₂ concentrations MMCO warming has been difficult to reproduce in an intermediate complexity Earth system model (Henrot et al., 2010), atmosphere and slab ocean models (Tong et al., 2009; You et al., 2009), and fully coupled atmosphere ocean models (Herold et al., 2011; Krapp and Jungclaus, 2011). For example, Herold et al. (2011) found that the Community Climate System Model (CCSM3.0) was ~ 10 °C too cold compared to proxy records in high latitude

²⁰ regions like Alaska and Antarctica. In this study, we implement boundary conditions from Herold et al. (2011) within the National Center for Atmospheric Research (NCAR) Community Earth System Model (CESM1.0) using the Community Atmosphere Model (CAM4) framework to simulate the MMCO. This allows for a clean comparison with previous simulations done with CCSM3.0, using a latest generation model included in the Coupled Model Intercomparison Project (CMIP5).

²⁵ the Coupled Model Intercomparison Project (CMIP5).

To explore if the modelling framework is able to match MMCO warmth we conduct a pointwise model data comparison using proxy records compiled for the MMCO (Tables S1 and S2). The MMCO is a good choice for climate model validation because the continental configuration is relatively close to modern (Herold et al., 2008) although





differences exist (Potter and Szatmari, 2009). Additionally, the CO₂ levels during the MMCO are in the range of values for the next century, and paleoclimate records are better constrained compared to earlier warm periods such as the Eocene (\sim 56–33.9 Ma) where there is large uncertainty in the CO₂ (Pagani, 2002; Pearson and Palmer, 2000; Royer et al., 2012) and temperature records.

2 Methods

2.1 Modelling framework

A series of MMCO global climate simulations are conducted using components of the NCAR CESM1.0 (Gent et al., 2011). The Community Atmospheric Model (CAM4) is run at 1.9° × 2.5° horizontal resolution with 26 vertical levels and coupled to the Community Land Model (CLM4) (Lawrence et al., 2012), the Community Sea-Ice Model (CICE4) (Hunke and Lipscomb, 2008) and the slab ocean model, described below (Bitz et al., 2012). This model simulates modern surface temperature distributions and equator to pole temperature gradients well (Gent et al., 2011), although biases exist (Kay et al., 2012; Neale et al., 2013).

2.2 Experimental design

The control Pre-industrial (PI) simulation employs the modelling components described above in standard configuration and with CO₂ concentrations set at 287 ppm. The slab ocean forcing file for the PI case has heat fluxes, salinity, and density inputs from a ²⁰ fully coupled atmosphere, ocean, ice, and land simulation (Bitz et al., 2012). Additionally we run a PI simulation at 400 ppm CO₂ (PI400) to compare with our MMCO simulation (also at 400 ppm CO₂). This high CO₂ PI configuration allows us to isolate the temperature effect of including MMCO boundary conditions at constant CO₂.

The MMCO simulation has vegetation cover and topography described in Herold et al. (2011). Previous slab ocean and atmosphere MMCO simulations have been con-





ducted within the CCSM3 framework (Tong et al., 2009; You et al., 2009), but here we improve upon their methodology by using ocean heat fluxes derived from a coupled ocean-atmosphere simulation of the Miocene. To create the Miocene slab ocean forcing file we use a previous CCSM3.0 Miocene simulation (Herold et al., 2012). This

- includes mix-layer depth and ocean heat transport from a fully coupled model. Using slab fluxes from CCSM3.0 is not an issue because we find no substantial differences in SST (Fig. S1) or climate between CCSM3.0 and CESM1.0 for deep paleoclimate simulations such as the Eocene as the ocean component biases are very similar between the two modeling frameworks (Danabasoglu et al., 2012).
- Our use of the slab ocean model in this study, as opposed to a fully dynamic ocean 10 model, is justified given that (1) we are interested in simulating a large number of sensitivity experiments which demand already intensive computational resources. (2) Experience from modern and Eocene studies show that this slab ocean approach produces very similar answers to those from coupled models (Gettelman et al., 2012; Bitz et
- al., 2012). (3) We can run the slab ocean simulations with higher resolution in the at-15 mosphere $(1.9^{\circ} \times 2.5^{\circ})$ than is standard for most paleoclimate studies because of the reduced computational requirement.

The simulations conducted are run for over 60 yr with the last 20 used for analysis. The simulations here are well equilibrated as evidenced by the radiative balance statistics found in Table S3. Additional MMCO CO₂ sensitivity experiments were run at 20 560 ppm CO₂ (MMCO560) to account for the uncertainty in Miocene CO₂ reconstructions and the model data comparisons for this experiment is described in Fig. 3b. We also run a simulation at 800 ppm (MMCO800) CO₂ (Fig. 3c) to explore a wide range of CO₂ values although we note that this is well outside the range of the reconstructed CO₂ levels for the MMCO.

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2.3 MMCO terrestrial and sea surface temperature compilation

For the model data comparison we update the compilation of terrestrial and SST proxy records described in Pound et al. (2012), Herold et al. (2011), and others (Tables S1





and S2). We present the longitudinal and spatial distribution of the proxy records in Fig. 1. The proxy reconstruction spans over the MMCO (17–14.50 Ma), however, because of the sparseness of data over this period we include records that have an average age between 20 and 13.65 Ma, where they fill spatial gaps (i.e. Southern Hemisphere). This data compilation can be used as a reference data set for future MMCO model data comparisons.

We update the minimum error in our compiled terrestrial proxy records for a number of reasons. Firstly, recent work suggests that for physiognomic leaf-climate methods there should be a minimum error of ±5 °C (Royer, 2012). Secondly, studies have suggested that there is large uncertainty in estimating MAT (Grimm and Denk, 2012) using the coexistence approach (Mosbrugger and Utescher, 2007). For our intended purposes increasing the minimum proxy record uncertainty should make matching the simulations more obtainable. If our model still fails to match proxy data even with generous error bars this merely proves our main results further.

The SST records are compiled from available published data in the literature and we describe these records in detail in Table S2. We leave out some tropical SST records which may have a diagenetic bias as described in (Sexton et al., 2006; Huber, 2008). Tropical SSTs are few and far between for the MMCO, but more common in the mid-to-late Miocene, thus we may omit proxy records from over almost half the surface area of the planet (30° N and 30° S) or utilize data from intervals slightly outside the

- MMCO. Because there is a lack of tropical SST data points for the MMCO we compile SSTs from the late Miocene and justify this based off the minimal change between middle and late Miocene SSTs at other locations (LaRiviere et al., 2012). Given that the Pliocene tropical SSTs were ~ 4-6 °C (Brierley et al., 2009; Dekens et al., 2007;
- Ravelo et al., 2006; Fedorov et al., 2013) above modern and the late Miocene were ~7–9°C above modern (LaRiviere et al., 2012) it is reasonable to conjecture MMCO tropical SSTs were this warm or warmer. Either approach introduces potential errors in interpretation and here we choose to utilize SST estimates in data sparse regions that lie generally within the early to middle Miocene, but may be outside the MMCO. Our





updated minimum error bars are large enough to encompass the temporal variation in these records.

Previous work has discussed the importance of including orbital variations when quantifying uncertainty in model data comparisons (Haywood et al., 2013). To quan-

- tify the possible error introduced by aliasing of orbital variability in our interpretation of model data mismatch, we conduct two sensitivity experiments varying obliquity to minimum and maximum Miocene values (22° and 25° respectively). We then calculate the maximum and minimum model-derived temperatures at each proxy location from both extreme orbit simulations and use this absolute anomaly as an estimate of orbitally
 induced variance. These maximum and minimum values are plotted as vertical error bars on the modelled MAT in our pointwise model data comparisons (Figs. 2, 3, 4, 5)
- bars on the modelled MAT in our pointwise model data comparisons (Figs. 2, 3, 4, 5, 6).

3 Results

3.1 Proxy derived MAT value

- ¹⁵ To determine the difference in global MAT between Miocene and pre-industrial climate we take the proxy records and perform a pointwise anomaly of proxy-derived MAT compared to modern observed MAT at paleo-latitudes and paleo-longitudes. We split the resulting anomalies into tropical (30° N to 30° S) mid-latitude (30° N/S to 60° N/S) and polar (60° N/S to 90° N/S) regions and conduct a weighted average anomaly over each
- ²⁰ latitudinal region. This latitudinal binning and area weighting addresses issues of having more proxy records in certain regions (i.e. the mid-latitudes). Using proxy records for the MMCO (Table S1 and S2) we calculate a global MAT change of ~ 7.6 °C ± 2.30 (We report two standard errors from the mean) compared to PI. The proxy-derived temperatures compared against modern observations (ECMWF 40 Year Reanalysis).
- ²⁵ Project) is $6.8^{\circ}C \pm 2.20$ as there is ~ $1.0^{\circ}C$ of warming between modern observations and PI climate.





To validate our approach for estimating proxy derived MAT we calculate a resampled MAT using our methodology and compare against a globally weighted MAT (we will call this true MAT) from both model runs and modern observational datasets. The globally weighted true MAT value of the MMCO simulation is 18.00 °C (Table 1) whereas our calculation for MAT resampled over the proxy record regions using the methodology from above is 17.12 °C. The calculated standard error from the mean including proxy record uncertainty is 1.33 °C, which illustrates that our resampled MAT value is well within the calculated standard error. We also calculate the resampled MAT using modern observations and with other Miocene simulations and find that all the resampled MAT estimates fall within two standard errors of the true MAT. For all intended purposes we are confident that our approach for reconstructing global MAT from our proxy record compilation is a valid estimate.

3.2 MMCO simulation compared against the proxy records

The MMCO simulation is 4.04 °C warmer than the control PI simulation, but the simulation is about 4 °C cooler than globally averaged MMCO proxy temperature reconstructions (Table 1). The MMCO simulation generally captures the tropical and mid-latitude temperature distribution of the proxy records, but fails to achieve above-freezing temperatures in the high latitudes (Fig. 2b, Table 2). The nature of this discrepancy can be clarified by examining the equator to pole surface temperature gradient. It is 17 °C

- ²⁰ larger in the MMCO simulation than in the proxy records (Table 1). Using the methods described in Lunt et al. (2012), the equator to pole temperature gradient is calculated by averaging the mean annual temperatures over the absolute latitudes of (60–80°) minus (0–30°); except here we use 80° because this the maximum latitudinal extent of proxy records. Additionally, an error weighted best fit line for the pointwise comparison
- ²⁵ reveals a root mean square (RMS) error of ~6°C and y-intercept of -6°C, although the slope of the regression line is close to 1 (Table 1). In summary, the MMCO simulation (at 400 ppm CO₂) is unable to produce high latitude warmth or a sufficiently warm global mean temperature compared to the paleo temperature records.



3.3 Effect of MMCO boundary conditions and CO₂ sensitivity experiments

We find that our MMCO simulation is 2.43 °C warmer compared to the PI simulation run at 400 ppm CO₂ (PI400). Thus 2.43 °C of the temperature difference between our MMCO and PI simulations are a result of changes in continental positions, topography, and vegetation. This change is consistent with late Miocene modelling which finds 3.0 °C of warming due to changes in vegetation and topography (Knorr et al., 2011).

A CO₂ sensitivity experiment run at 560 ppm CO₂ (above most reconstructed CO₂ records) is also too cold at high latitudes compared to proxy records (Fig. 3b) and the equator to pole temperature difference is still too large by ~ 13 °C (Table 1). This simula-

- tion has a global MAT 5.89 °C higher than the control PI simulation, and is ~ 2 °C colder than the proxy-derived global MAT. The error weighted best fit line for the MMCO560 pointwise comparison gives a y-intercept of ~ −2.5 °C, but the calculated RMS error is still 5.7 °C (Table 1). The MMCO800 simulation has a MAT 7.26 °C above PI (Table 1), which is our best comparison with the proxy derived MAT value. The error weighted
- ¹⁵ best fit line is also very close to the one to one line and has a y-intercept close to zero (Fig. 3d). Overall MMCO800 matches the proxy compilation the best and we use this comparison to prove that matching global MMCO warmth can be accomplished, but at CO₂ concentrations approximately twice that reconstructed from proxies. These results are very similar to those found in the Eocene (Huber and Caballero, 2011; Lunt et al., 2012)

Below, we test hypotheses that have been proposed to explain Miocene warmth, with the goal of improving the model data comparison without having to increase CO_2 above reconstructed levels.





4 Further sensitivity studies

4.1 Reducing Antarctic ice-sheet volume

Recent work estimates the volume of the middle Miocene Antarctic Ice Sheet (AIS) to be $\sim 30-50~\%$ less than modern (Shevenell et al., 2008). Consequently the Herold et

- al. (2008) reconstruction for AIS elevation and extent is likely too large (Fig. 2a). To correct this, we utilize a new AIS reconstruction derived from a fully interactive terrestrial ice and atmosphere model (Pollard personal communication) (Fig. 4a). We introduce an AIS that is half the volume of that used in Herold et al. (2011) (Fig. 2a) from the offline interactive ice sheet simulation. This new AIS volume is within the range of esti-
- mates from proxy records (Pekar and DeConto, 2006; Billups and Schrag, 2003). We also reduce the area of glacier albedo over Antarctica by half and replace it with a combination of unvegetated and tundra-like land cover. We introduce this new AIS topography and vegetation cover (Fig. 4a) into the MMCO boundary conditions described in Herold et al. (2008) and denote this simulation LOW AIS. The difference in surface
- ¹⁵ albedo over the AIS between these two simulations ends up being similar as snow (also with a high albedo) ends up covering the areas that were once glacier because Antarctica stays below freezing year round.

The LOW AIS simulation is 4.15 °C warmer than PI and 0.10 °C warmer than the previously described MMCO simulation with a high AIS (Fig. 4c). Thus, there is no sig-

- nificant global mean temperature impact from decreasing the size of the AIS, consistent with previous work (Goldner et al., 2013). Although recent coupled MMCO simulations have found warmer and wetter conditions regionally over Europe due to reducing ice extent in Antarctica highlighting the importance of including ocean feedbacks for resolving regional temperature distributions (Hamon et al., 2012). The temperature difference
- ²⁵ between LOW AIS and the MMCO simulation is largest over Antarctica (Fig. 4b) because of the imposed elevation and surface albedo changes. Although lowering the AIS warms the Antarctic continent, the Miocene LOW AIS simulation results in negligible improvement in matching proxy records elsewhere in the high latitudes (Table 2).





A slight warming occurs in the Ross Sea between the LOW AIS simulation and the MMCO simulation, but overall there is minor improvement in the model data comparison (Fig. 4c) by lowering the height and reducing glacier extent of the AIS (Fig. 4a)

4.2 El Padre

- It has been hypothesized that pre-Quaternary climates were characterized by a reorganization of tropical ocean-atmosphere circulation inducing a permanent El Niño-like SST distribution (Philander and Fedorov, 2003; Lyle et al., 2008; Ravelo et al., 2006) which has been called El Padre. A reduced temperature gradient in the eastern equatorial Pacific (EEP) should induce high latitude warming in Alaska and other high latitude
- regions, because this is a standard teleconnected response during modern El Niño's (Molnar and Cane, 2007). Prior modelling studies have demonstrated the effectiveness of this mechanism (Barreiro et al., 2006; Vizcaíno et al., 2010; Bonham et al., 2009; Haywood et al., 2007; Goldner et al., 2011), although no modeling study has explicitly studied its impacts with realistic MMCO boundary conditions.
- ¹⁵ To explore the impacts of an El Padre SST anomaly in our simulations, we take the heat convergence and mixed layer depths derived from a fully coupled Miocene simulation (Herold et al., 2012) and zonally average these quantities across the Equatorial Pacific (10° N and 10° S of the equator). We introduce the zonally averaged ocean heat convergence and mixed layer depths into a new slab ocean forcing file and simulate
- the MMCO with a low AIS at 400 ppm CO₂. The resulting surface temperature anomaly is El Padre like (Fig. 5a) and the simulation is called EP. We are confident the CAM4 CESM1.0 framework reproduces modern day observational teleconnections patterns induced by El Niño forcing as described in detail in other studies (Wang et al., 2013; Shields et al., 2012). Although an interesting question for past warm periods like the MMCO is how these rights and regional reserves to ENCO have varied through but through a strength regional reserves to ENCO.
- MMCO is how these global and regional responses to ENSO have varied throughout geologic time, as modelling of the late Miocene has shown that ENSO teleconnections can be modified from modern (Galeotti et al., 2010; Tang et al., 2013).



In the EP simulation, high latitude regions warm, especially Alaska and Antarctica (Fig. 5a). The pointwise model data comparison for the EP simulation is plotted in Fig. 5b. This simulation is ~4.6 °C warmer in global mean than the PI simulation and ~0.5 °C warmer than the MMCO and LOW AIS simulations. Warming due to adding EI

- ⁵ Padre is largest in regions where the model previously performed the worst (Fig. 5a). Roughly 2°C of warming occurs in Alaska, but the simulation is still ~ 8.5°C too cold in this region (Table 2) and still has a ~ 13°C larger equator to pole surface temperature gradient compared to the proxy records (Table 1). Imposing an El Padre illustrates a mechanism capable of warming the high latitudes without elevating CO₂ consistent with
- the results of (LaRiviere et al., 2012; Sriver and Huber, 2010; Brierley et al., 2009). Nevertheless this change does not reconcile the warmth of the MMCO, as temperatures are still ~ 2 °C too cool globally and ~ 8.5 °C too cool in the high latitudes.

Adding EP and increasing obliquity to 25° results in a simulation that is 5.64°C warmer than PI (Fig. 6). This MAT anomaly compared to PI is similar to the warm-

ing found in the MMCO560 simulation. The MMCO560 simulation does not include any of the boundary condition changes aimed at increasing high latitude warmth. Interestingly the EP, AIS, and obliquity forcing results in a 4°C improvement in simulating the equator to pole temperature gradient compared to MMCO560 (Table 1). Both comparisons are too cold compared to the proxy derived global MAT value as matching the proxy records in high latitudes requires a CO₂ concentration double what is predicted in the reconstructions.

5 Discussion

5.1 Comparison with previous MMCO CCSM3.0 simulations

The most comparative study to the experiments presented here are the CCSM3.0 ²⁵ MMCO simulations described in Herold et al. (2011) (Table 1). The CESM1.0 Miocene simulations are ~2.0 °C warmer than the Miocene CCSM3.0 simulations (Herold et



al., 2011) at the same CO₂ levels. CAM4 is warmer than CAM3 at the same CO₂ concentrations, in large part because it is a more sensitive model to background CO₂ concentrations. CCSM3.0 had a 2.5 °C change in global mean surface temperature to a doubling of CO₂ (Kiehl et al., 2006), while CSEM1.0 has a 3.5 °C temperature change

⁵ to a doubling of CO₂ (Gettelman et al., 2012), roughly a 1 °C higher climate sensitivity. Imposing large forcings such as an El Padre anomaly, reducing AIS topography and albedo, or including uncertainty of orbital forcing failed to produce suitably warm temperatures in the global mean (Table 1) and at high latitudes (Table 2).

Additional simulations in past warm climates exploring precession (Sloan and Huber, 2001; Lawrence et al., 2003), eccentricity (Westerhold et al., 2005), vegetation (Knorr et al., 2011), and using a dynamic ocean could be important in our understanding of MMCO warmth. We point out that previous fully coupled ocean atmosphere simulations at 560 ppm CO₂ were unable to reproduce MMCO warmth (Herold et al., 2011). In fact, the CCSM3.0 MMCO simulation at 560 ppm CO₂ performs worse against the proxy records than our MMCO CAM4 simulation at 400 ppm CO₂ (Table 1). We also reiterate that the temperature effect of including MMCO boundary conditions induces 2.43 °C of warming compared to the PI400 simulation. This is roughly a third of the warming needed to explain the MMCO warmth of ~ 7.6 °C ± 2.30.

5.2 Comparison with other fully coupled MMCO simulations

- Krapp and Jungclaus (2011) simulated the MMCO and found a MAT of 17.1 °C at 480 ppm CO₂ and 19.2 °C at 720 ppm CO₂. These simulations are roughly 4 °C and 2 °C colder compared to the MAT calculated from the proxy records presented here. This study also comes to similar conclusions about their model's inability to reproduce reconstructed warmth in the high latitude regions especially in the Southern Hemisphere. Hamon et al. (2012) also conducted fully coupled MMCO simulations under a
- variety of different changes in boundary conditions. Comparison to this study is difficult because results focused on regional temperature changes to AIS forcing and they did not report global MAT values. Henrot et al. (2010), using an intermediate complex-





ity Planet Simulator explored changes in topography, sea-ways, CO_2 , and vegetation across the MMCO and these simulations are too cold in the mid-latitudes compared to the records. They simulate warming of 2.9 °C and 3.4 °C above PI when CO_2 concentrations are increased to 500 ppm and vegetation is altered, which is half of the temperature change needed to explain the proxy derived MAT.

6 Conclusions

Paleoclimate modeling studies need to conduct a pointwise model data comparison to be confident that their modelling results match proxy records and consequently we will make the presented MMCO temperature data set available for these types of compar¹⁰ isons. Simulating the MMCO at 400 ppm CO₂ using the CAM4 CESM1.0 framework produces a significant model data mismatch in global MAT and in high latitudes. The discrepancy in the MAT comparison is equal to that introduced by a full doubling of CO₂, as the model matches the data best at 800 ppm CO₂. A similar conclusion about climate model sensitivity to background CO₂ forcing was reached based on fully coupled ocean
¹⁵ atmosphere Eocene simulations where a CO₂ level nearly double the reconstructions was required to match the proxy records (Huber and Caballero, 2011). It is interesting

- to note that the reconstructed CO_2 used in this study of 400 ppm is equivalent to the concentration used in simulations of the Pliocene, where global temperatures were not as warm as the Miocene.
- Including two of the most discussed Earth system feedbacks (El Padre and reduced ice volume) had small impacts on improving the model predictions even when we included uncertainty associated with time varying and possible aliasing of orbital forcing. Like previous fully coupled atmosphere ocean efforts (Herold et al., 2011; Krapp and Jungclaus, 2011) matching proxy records at the MMCO is challenging even in the latest approximate of model with a simple possible and with a simple possible and possible at the simple possible at the simple possible at the simple possible at the simple possible possible at the simple possible possible
- 25 generation of models and using a model with a climate sensitivity near the median of Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report (AR4) estimates. Given the variety of methods used for reconstructing Miocene climate (Ta-





bles S1 and S2) we are confident in the broad trends reflected in the proxy record. Thus, explaining the warming will require additional incremental changes in boundary conditions (such as an even higher CO_2), a more sensitive model to background CO_2 concentrations, and/or identification of some – as yet unknown – process or forcing that accounts for almost half of the difference in temperature between today and the MMCO.

Although some terrestrial CO_2 proxies suggest CO_2 was higher than 500 ppm, this would not solve the data model mismatch, as increasing CO_2 past 560 would likely make the tropics too warm (e.g. Fig. 3b, d). Ultimately, our inability either to identify a missing paleoclimate forcing or formulate models with sufficient positive feedbacks to

¹⁰ missing paleoclimate forcing or formulate models with sufficient positive feedbacks to recreate substantial increases in global mean temperature with strong polar amplification represents a persistent weakness of climate models.

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Table 1. Compilation of model and proxy MAT values	, equator to pole temperature gradient
values, and model data point wise comparison statistics	3.

Simulation Name and Records	MAT (°C)	Miocene minus PI (°C)	Equator to Pole Temperature Gradient (°C) ^a	Slope ^b	Y-intercept of best fit line ^b	RMS Error
MMCO Records	21.89 ± 2.2	_	24.50	-	-	-
	13.95	1 62	43.84 42.16	1.29	-13.73	10.12
MMCO	18.00	4.04	41.79	1.11	-5.91	6.05
MMCO560	19.84	5.89	37.50	1.09	-2.39	5.72
MMCO800	21.19	7.26	33.00	0.95	-0.21	4.81
LOW AIS	18.10	4.15	39.08	1.01	-5.54	5.23
EP	18.68	4.66	37.89	1.08	-5.32	5.99
EP+ORB CCSM3.0 T31	19.66	5.64	33.79	1.06	-3.86	5.19
355 ppm CO ₂ CCSM3.0 T31	15.38	1.43	37.00	_	_	-
560 ppm CO ₂	16.94	2.99	35.00	-	-	-

^a The equator to pole surface temperature gradient is calculated by averaging the mean annual temperatures over the absolute latitudes of (60-80°) minus (0-30°); 80° is the maximum latitudinal extent of proxy records. ^b The slope and y-intercept of the best fit line for the pointwise model and proxy comparisons in Figs. 2, 3, 4, 5, 6. The best fit line is weighted to include the error uncertainty found in the proxy records (Tables S1, S2).



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Table 2. High latitude model proxy data comparison for the Alaskan and Antarctic records. The simulations in the comparison include CESM1.0 and CCSM3.0 (Herold et al., 2011) model runs.

	Latitude (°)	Proxy (°C)	Error (±°C)	MMCO (°C)	MMCO 560	LOW AIS	EP	EP+ ORBI TAL25	CCSM3.0 T31 355 ppm CO ₂	PI
Porcupine River 90-1, Organic bed	68.19° N	8.00	8.00	-7.00	-3.7	-7.40	-5.20	-3.0	-6.80	-10.81
Nenana Coal Fm	65.11° N	7.50	8.00	0.00	2.9	-0.50	1.30	3.20	-5.59	-10.35
Coal Creek	64.99° N	8.00	8.00	0.00	2.9	-0.50	1.30	3.20	-5.59	-10.35
Cook Inlet	62.00° N	11.00	3.00	2.10	4.60	1.30	3.10	4.90	1.39	-9.95
AND-2A (Ross Sea)	–77.00° S	5.50	5.00	-1.50	0.00	-1.43	-1.40	-0.25	-1.72	-1.73







Fig. 1. (a) Longitudinal distribution of MMCO terrestrial temperatures (red diamonds) and SST (blue crosses) with proxy record error plotted as the vertical bars and described in Tables S1 and S2. (b) The spatial distribution of the terrestrial and SST proxy records used in the model data comparisons overlain onto the Miocene topography (Herold et al., 2008).







Fig. 2. (a) High AIS topography used in Herold et al. (2011), **(b)** Pointwise MAT comparison between the MMCO simulation and proxy records (Tables S1 and S2). Vertical error bars are the modelled pointwise maximum and minimum temperatures from the extreme obliquity simulations (see Methods Sect. 2.3) and methodological error is plotted as the horizontal error bars. The best fit line (black dashed) is weighted to include proxy uncertainty and is fitted across all points. The weighting for each proxy record is calculated by 1/(error²). The y-intercept and slope are reported in Table 1.







Fig. 3. (a) Modelled temperature anomaly for the MMCO560 (560 ppm CO_2) simulation minus the MMCO simulation (°C). **(b)** Pointwise MMCO560 simulated global MAT compared against the proxy record MAT (°C). **(c)** Modelled temperature anomaly for the MMCO800 (800 ppm CO_2) simulation minus the MMCO simulation (°C). **(d)** Pointwise MMCO800 simulated global MAT compared against the proxy record MAT (°C). These are the same terrestrial and SST records described in Fig. 1. Vertical error bars indicate the uncertainty recorded by maximum and minimum temperatures of extreme orbital obliquity parameters (see Methods Sect. 2.3). The best fit line (black dashed) is weighted to include error uncertainty is fitted across all points and the y-intercept and slope reported in Table 1







Fig. 4. (a) LOW AIS topography based on offline ice-sheet modeling (David Pollard, personal comms), **(b)** modelled temperature anomaly (°C) between the LOW AIS simulation and the MMCO simulation with the high AIS. **(c)** Pointwise MAT comparison between the LOW AIS simulation and proxy records (Tables S1 and S2). The best fit line (black dashed) is weighted to include error uncertainty and is fitted across all points and the y-intercept and slope are reported in Table 1.













Fig. 6. (a) Modelled temperature anomaly for the EP+ORBITAL25 simulation minus the LOW AIS simulation (°C), **(b)** Pointwise EP+ORBITAL25 case global mean MAT compared against the proxy record MAT (°C). These are the same terrestrial and SST records described in Fig. 1. Vertical error bars indicate the uncertainty recorded by maximum and minimum temperatures of extreme orbital obliquity same as Fig. 1. The best fit line (black dashed) is weighted to include error uncertainty is fitted across all points and the y-intercept and slope reported in Table 1.



