Amplified Late Pliocene terrestrial warmth in northern high latitudes from greater radiative forcing and closed Arctic Ocean gateways

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Key words

- Late Pliocene climate
- Glacial-interglacial variation in insolation
- Cluster analysis
- Terrestrial temperature reconstructions
- Polar amplification

Abstract

Proxy reconstructions of the mid-Piacenzian warm period (mPWP, between 3.264 and 3.025 Ma) suggest terrestrial temperatures were much warmer in the northern high latitudes (55° – 90 °N, referred to as NHL) than present-day. Climate models participating in the Pliocene Model Intercomparison Project Phase 1 (PlioMIP1) tend to underestimate this warmth. For instance, the underestimate is ~10 °C on average across NHL and up to 17 °C in the Canadian Arctic region in the Community Climate System Model version 4 (CCSM4). Here, we explore potential mPWP climate forcings that might contribute to this mPWP mismatch. We carry out seven experiments to assess terrestrial temperature responses to Pliocene Arctic gateway closure, variations in CO₂ level, and orbital forcing at millennial time scale.

To better compare the full range of simulated terrestrial temperatures with sparse proxy data, we introduce a pattern recognition technique that simplifies the model surface temperatures to a few representative patterns that can be validate with the limited terrestrial proxy data. The pattern recognition technique reveals two prominent features of simulated Pliocene surface temperature responses. First, distinctive patterns of amplified warming occur in the NHL, which can be explained by lowered surface elevation of Greenland, pattern and amount of Arctic sea ice loss, and changing strength of Atlantic meridional overturning circulation. Second, patterns of surface temperature response are similar among experiments with different forcing mechanisms. This similarity is due to strong feedbacks from changes in surface albedo and troposphere water vapor content as a result of sea ice responses, which overwhelm distinctions in contributions from changes in insolation, CO₂ forcing, and Arctic gateway closure.

By comparing CCSM4 simulations with proxy records, we demonstrate that both model and proxy records show similar patterns of mPWP NHL terrestrial warmth, but the model

underestimates the magnitude. High insolation, greater CO₂ forcing, and Arctic gateways closure each contributes to reduce the underestimate by enhancing the Arctic warmth of 1 - 2 °C. These results highlight the importance of considering proxy NHL warmth in the context of Pliocene Arctic gateway changes, and variations in insolation and CO₂ forcing.

Introduction

The mid-Piacenzian warm period (mPWP, between 3.264 and 3.025 Ma, Dowsett et al., 2010) features a distinctive climate state. Global geography and surface elevations are similar to present-day, yet climate as a whole is warmer by 1.9 - 3.6 °C based on combined proxy and model estimates (e.g. Haywood and Valdes, 2004; Masson-Delmotte et al., 2013). One remarkable feature of the mPWP is the strongly amplified arctic warming despite moderate CO₂ levels. The reconstructed CO₂ levels for mPWP range from pre-industrial to 500 ppm, with a best estimate of 350 to 450 ppm (e.g. Masson-Delmotte et al., 2013; Haywood et al., 2016a). Fossil pollen, plant records, and insect assemblages suggest boreal forests and warm conditions across the Canadian Arctic and Eastern Russia during the Pliocene (5.33 – 2.58 Ma) (Ballantyne et al., 2010; 2013; Elias and Matthews Jr., 2002; Salzmann et al., 2008; 2013), which are currently occupied by tundra and permafrost. Quantitative reconstructions further suggest mean annual temperatures of northern high latitudes are above freezing (Salzmann et al., 2013; Ballantyne et al., 2010; 2013). While independent proxy estimates converge on the strong Arctic warming, much greater than the global mean warming during the mPWP, coupled atmosphere-ocean climate simulations reproduce only a small amount of this amplified warming (Haywood et al., 2013). Temperature underestimates range from a few to up to 18 °C across northern high latitudes in model simulations with PlioMIP1 boundary conditions (Dowsett et al., 2010; Haywood et al., 2010), including HadCM3, MIROC, and CCSM4 (Haywood et al., 2013;

Salzmann et al., 2013). This large discrepancy between model results and proxy data may limit confidence in model predictions of future warming across northern high latitudes.

Uncertainties in geological reconstructions of boundary and forcing conditions used by climate models have been proposed to resolve the underestimates of Pliocene circum-Arctic warming in the coupled models. Reported terrestrial biomes for the mPWP represent a time slab that overlaps, but is much longer than, the targeted mPWP. As a result, these records could correspond to a variety of insolation and CO₂ forcings that are not represented by the prescribed orbital parameters (modern) and CO₂ level (405 ppm) in PlioMIP1 (Haywood et al, 2010). Within the mPWP time slab, the summer insolation varies by up to 89 W/m² and the annual insolation varies by 1.6 W/m² at 65°N (Laskar et al., 2005). Variation of summer and annual insolation at millennial time scale is known to be important for high latitude climates (e.g. Tabor et al., 2015; Prescott et al., 2014). This insolation variation may account for some of the proxy warming in the northern high latitudes, especially if terrestrial proxies are sampling towards high productivity periods (Prescott et al., 2014; Salzmann et al., 2013). The CO₂ level declines from over 500 ppm to pre-industrial levels through the Pliocene and varies by reconstruction methods (Martínez-Botí et al., 2015). The uncertainty range is over 100 ppm at almost any given Pliocene time period (See discussion of records by Haywood et al., 2016a).

Variations in both CO₂ and orbital forcing have been shown to be important for explaining the PlioMIP1 model-proxy data mismatches. Salzmann et al. (2013) and Howell et al. (2016b) investigated terrestrial surface temperature sensitivity to extremes of mPWP insolation and CO₂ forcing of northern high latitudes; their results highlight the importance of these forcings in improving the model-proxy data comparison. Further, Prescott et al. (2014) performed a series of simulations using HadCM3 to simulate interglacial periods of mPWP

climate. The results show high sensitivity of Arctic surface temperatures to orbitally-driven insolation variations, especially at seasonal time scale (up to 20°C). Finally, Willeit et al. (2013) investigated variability of Pliocene climate to orbital forcing in a transient simulation with dynamical ice sheet and vegetation using an intermediate complexity model. In their results, simulated summer temperatures in the high latitudes approach proxy temperatures only during the interglacial periods of Pliocene.

Uncertainties in mPWP geography, in particular whether Arctic Ocean gateways are open or closed, can have strong effects on mPWP climate by changing ocean circulation and freshwater transport (Otto-Bliesner et al., 2016). Published simulations from PlioMIP1 use boundary conditions that were derived from Pliocene Research, Interpretation and Synoptic Mapping paleoenvironmental reconstructions (PRISM3D, Dowsett et al., 2010). Many modeling groups adopted modern geography and bathymetry since it is difficult for some complex climate models to alter the land/sea mask. During the second phase of PlioMIP (PlioMIP2), two ocean gateway changes around the Arctic are highlighted (i.e. closure of both the Bering Strait and Straits in the Canadian Arctic Archipelago) (Dowsett et al., 2016). Additional, but less constrained changes to Pliocene paleogeography may come from replacing the Barents Sea with land (Butt et al., 2002; Knies et al., 2009), a deeper undersea Greenland-Scotland-Iceland ridge (Robinson et al., 2011), and an extended drainage basin of the Hudson Bay (Duk-Rodkin and Hughes, 1994). Individual or a combination of these paleogeographic changes are shown to strengthen the Atlantic Meridional Overturning Circulation (AMOC) and/or redirect the north Atlantic currents, which subsequently warm the northern North Atlantic in excess of 5 °C (Robinson et al., 2011; Hill et al., 2015).

In this study, we provide a systematic evaluation of northern high latitude (NHL, stands for 55 ° - 90°N region) terrestrial climate responses to mPWP variations in orbital and CO₂ forcing, and Pliocene Arctic gateway changes using the CCSM4 model. A separate study by Otto-Bliesner et al. (2016) evaluates oceanic responses to Pliocene Arctic gateway closure. Although climate responses to some of these forcings have been reported for the HadCM3 model, a systematic synthesis and comparison is lacking, and model-proxy data comparison is mainly based on individual sites (Salzmann et al., 2012; Hill et al., 2015; Howell et al., 2016b). The use of CCSM4 and our new approach of model-proxy data comparison utilizing pattern recognition techniques provide independent and comprehensive evaluations of the gap between model and proxy data for the NHL terrestrial temperatures. A series of model experiments are conducted to quantify terrestrial temperature and sea ice responses to individual forcing mechanisms. We have also compiled NHL terrestrial proxy temperatures to provide a benchmark for our CCSM4 experiments by integrating previous compilations of Salzmann et al. (2013), Ballantyne et al. (2010, 2013), and more recent records (Pound et al., 2015; Panitz et al., 2016; Fletcher et al., 2017, personal communication). We further analyze and discuss large-scale results, patterns of simulated changes in surface temperature responses, comparison with terrestrial proxy data, and responses of NHL energy budget to these imposed Pliocene climate forcings.

Methods

1. Experiments

Seven experiments are conducted with the coupled CCSM4 model, including the upper end of Pliocene variations in CO₂ level (450 ppm: PlioHCO₂) (Seki et al., 2010; Haywood, et al.,

2016a), extremes in orbital forcing (65 °N July maximum: PlioMaxInso_{July}; 65 °N July minimum: PlioMinInso_{July}; annual maximum: PlioMaxInso_{ann}) (Laskar et al., 2005), and closure of the Bering Strait (PlioBSC), the Canadian Arctic Archipelago (PlioCAA), and combined closure of both gateways (PlioCAA+BSC) (Otto-Bliesner et al., 2016). Changes in boundary conditions and abbreviations for individual experiments are listed in Table 1. CCSM4 (Gent et al., 2011) is comprised of four dynamical model components: Community Atmospheric Model version 4 (CAM4), the Community Land Model version 4 (CLM4), Community Ice Code version 4 (CICE4), and Parallel Ocean Program version 2 (POP2). As in Rosenbloom et al. (2013), all model components are coupled. Pliocene vegetation types are prescribed based on the Salzmann et al. (2008) compilation. The carbon-nitrogen biogeochemistry component of CLM4 allows the vegetation phenology to be impacted by the modeled climate conditions.

We run CCSM4 with a finite volume dynamical core at $0.9 \times 1.25^\circ$ horizontal resolution of the atmosphere and the displaced Greenland pole grid with an average $\sim 1.11 \times 0.54^\circ$ horizontal resolution of the ocean (Gent et al., 2011). The model features 26 (surface to 3.5 hPa) and 60 vertical levels of the atmosphere and ocean, respectively. The model resolution is the highest among all the participating models in PlioMIP1 (Haywood et al., 2013). The experiments are branched from CCSM4 PlioMIP1 simulations (Rosenbloom et al., 2013) and run for at least 200 model years each, until the experiments reach near equilibrium with top of the atmosphere energy imbalance below 0.1 W/m^2 and trends of global surface temperatures below $0.1^\circ\text{C/century}$. Simulations are averaged over the final 50 model years to produce climatologies.

For present-day simulations, CCSM4's pattern and magnitude of northern high latitude surface temperatures match well with observations (Gent et al., 2011). In contrast to previous versions of the atmosphere model, CAM4 includes a freeze-dry modification in the low cloud

parameterization, which leads to reduced wintertime Arctic low clouds and improved match with observation data (Gent et al., 2011). Major improvements have also been made to the shortwave physics in the sea ice component, which is able to predict albedo of snow, ice, and melt ponds based on simulated environmental conditions and optical properties of snow and ice (Holland et al., 2011). This prognostic albedo scheme differs from many other models for which snow and ice albedo are prescribed. The improvements made in both cloud and sea ice albedo schemes lead to realistic simulation of present-day Arctic sea ice extent, thickness, and distribution (Gent et al., 2011; Howell et al., 2016a). The model, however, still has a few known biases, including an overestimate of Arctic sea ice volume (Shu et al., 2015), and an underestimate of SSTs (up to ~ -2 °C) and terrestrial temperatures at the periphery of the Arctic Ocean and northern Greenland (up to -4 °C) (Gent et al., 2011). Yet, among atmosphere-ocean coupled models, CCSM4 shows superior skill in simulating Arctic sea ice extent, volume, and seasonal cycle (Table 2 in Shu et al., 2015).

2. Updated proxy terrestrial temperature compilation and model-proxy data comparison using cluster analysis

For model comparison, we create a new compilation of NHL terrestrial proxy temperatures records for the mPWP using proxy records with published ages that overlap the mPWP. Our compilation combines Salzmann et al. (2013) and Ballantyne et al. (2010) compilations with additional records from White Channel Gravels, Yukon, Canada (Pound et al., 2015), ODP642 in the Norwegian Sea (Panitz et al., 2016), and recent estimates from multiple sites of Canadian Arctic Archipelago (Fletcher et al., 2017, personal communication). Additionally, Tjörnes bed record of the northern Iceland (66°N , 17°W) was originally dated to

be middle to late Pliocene (Akhmetiev, 1991) and was included in the previous compilations, but has since been revised to be late Miocene (6 Ma) (Mattingsdal et al., 2014) to early Pliocene (4.5 Ma) (Verhoeven et al, 2011). This record is removed from our revised compilation. Finally, we evaluate confidence levels of newly added sites based on age control and resolution, completeness of fossil information, and uncertainties in temperature estimates consistent with Salzmänn et al., (2013). Details of the data compilation and our evaluation of confidence levels are provided in the Supplemental Materials.

We conduct model-proxy data comparison based on surface temperature responses (ΔT -model) in Pliocene simulations by comparing against the preindustrial (PI) control simulation. Values of ΔT -model are quantified as annual mean 2-meter air temperatures above the surface in Pliocene simulations minus those in PI. The PI control simulation uses the same CCSM4 model configuration and resolution. This differencing approach removes systematic model biases that occur in both PI and Pliocene simulations. Accordingly, we estimate ΔT -proxy by differencing proxy estimates of Pliocene terrestrial temperatures and gridded (~50 km horizontal resolution) observations of mean surface air temperature from 1901-1930 (Willmott and Matsuura, 1995), the earliest period for which high-resolution gridded observations (not the data assimilation product) are available. We group ΔT -proxy by reconstruction method and confidence level (Table S1).

Using cluster analysis, we extract representative patterns of NHL ΔT -model and utilize these patterns to conduct our model-proxy data comparison. We use cluster analysis, instead of site-by-site comparison, because there are only 31 published Pliocene terrestrial temperatures from the NHL (Table S1). Therefore, a site-by-site comparison only evaluates a tiny fraction of model results. Furthermore, among our simulations, cluster separations reflect regional

heterogeneity of climate responses to different forcings (described in the Results), serving a physical basis for model-proxy data comparison.

Under present-day climate conditions, the broad-scale nature of surface energy balance and mean circulation state provides physical basis for organized patterns of surface temperatures. For example, temperature contrasts between the East Coast of the U.S. and Europe and between Europe and East Eurasia result from cold and warm air intrusion along the planetary wave path (e.g., Seager et al., 2002). Moreover, temperature contrasts between the Canadian Arctic and Siberia are due to the existence of thick sea and land ice adjacent to northern Canada and over Greenland. Surface temperatures are also dependent on topography. Topographic changes lead to organized regional responses of atmospheric circulation and surface energy budget, and hence, surface temperatures (e.g., Feng et al., 2013; Feng and Poulsen, 2014; 2016). Similar physical processes are expected to affect ΔT -model and may lead to organized patterns in our simulations.

In order to identify patterns of simulated Pliocene surface warmth, we perform cluster analysis on gridded data of ΔT -model using the Gaussian Mixture Model (GMM) (R package ‘Mclust’, Fraley et al., 2012). The GMM assumes that sample data can be described with a selection of Gaussian distributions, which is appropriate for surface temperatures (e.g., Hansen et al., 2012). We determine the number of clusters by maximizing Bayesian Information Criterion (BIC: a combined measure of likelihood and model complexity) and minimizing the number of clusters needed (see Supplement Material for details).

We assume that ΔT -model and ΔT -proxy share similar geographic boundaries of cluster patterns, and conduct the model-proxy data comparison based on cluster patterns derived from ΔT -model. This assumption can be evaluated qualitatively. If the cluster partitioning of ΔT -model is similar to ΔT -proxy, increasing (decreasing) medians of ΔT -proxy from one cluster to

another will follow increasing (decreasing) medians of ΔT -model. However, if the cluster partitioning of ΔT -model differs from ΔT -proxy, changing medians of ΔT -model from one cluster to another may correspond to little or opposite changes in the medians of ΔT -proxy. This approach is further illustrated in Fig. S4 of the Supplemental Materials.

In order to quantify the model-proxy data difference, the medians of ΔT -proxy are calculated for sites located within the spatial range of each cluster, and grouped by reconstruction method, proxy source, and confidence level. For collocating records, if all records have the same reconstruction method, proxy source, and confidence level, we only use one median value. If not, different medians are calculated for records with different reconstruction methods, proxy sources, or confidence levels. Based on these medians, confidence-level weighted (weights are: 4-very high confidence, 3-high confidence, 2-medium confidence, 1-low confidence) mean differences between ΔT -model and ΔT -proxy are first calculated for each cluster, then averaged over all clusters. We choose the median over the mean due to its robustness to outliers.

3. Energy balance calculation

We adopt the method outlined by Lunt et al. (2012a, b) and Hill et al. (2014) to calculate contributions to Pliocene NHL surface warming from different radiative components. In this method, zonal mean surface temperature (T) at each model latitude is approximated using $S_0 \times (1 - \alpha) + H = \varepsilon \sigma T^4$, where S_0 is the incoming solar radiation at the top of the atmosphere, H is zonal mean heat convergence, α is planetary albedo, ε is top of atmosphere emissivity, and σ is Planck's constant. ε is calculated using the model surface and top of the atmosphere outgoing longwave radiation fluxes and separated between clear- and cloudy-sky conditions.

We have made two improvements upon previous implementations of this method. First, in the original calculation, cloudy-sky albedo is determined by differencing radiation fluxes between all-sky and clear-sky. The accuracy of this calculation is compromised in polar regions, where changes in surface reflection between experiments can be mistaken for changes in cloud reflection. When comparing PI and Pliocene cases, this ambiguity may lead to incorrect results, because the cloudy-sky albedo will seem to increase with a decrease of surface snow/ice area even without any changes in cloud field. In order to avoid this error, we apply the Approximate Partial Radiative Perturbation (APRP) method to separate planetary albedo changes due to changes in surface, clear atmosphere, and cloud reflection using a simplified explicit radiative transfer model for shortwave radiation (Taylor et al., 2007). Specifically, we use APRP to calculate contributions to α from changes in surface and clear-sky reflection using the model outputs. Contribution from cloudy-sky reflection is treated as a residual. This modification to the original APRP method was implemented because the estimates of cloudy-sky radiative fluxes in the original APRP model require known in-situ cloud fraction, which cannot be accurately represented by the model outputs of averaged cloud fractions.

Second, we separate contributions to the latitudinal temperature changes from zonal mean heat convergence (H) into those from the atmosphere (H-atm) and ocean (H-ocn). H is calculated from the zonal mean energy imbalance at the top of the atmosphere between net incoming solar radiation and net outgoing longwave radiation: $H = S_{\text{LW-TOA}} - S_{\text{SW-TOA}}$. H-ocn is then determined using the top of the ocean (surface) energy imbalance ($S_{\text{LW-SRF}} - S_{\text{SW-SRF}}$), and latent (LHF) and sensible (SHF) heat fluxes: $H\text{-ocn} = -(S_{\text{SW-SRF}} - S_{\text{LW-SRF}} - \text{LHF} - \text{SHF})$. The difference between H and H-ocn gives the H-atm.

Table 1. Statistics of annual mean 2-meter air temperatures (T_{2m}), sea ice, and strength of Atlantic overturning circulation (AMOC) in preindustrial simulation (PI) and Pliocene simulations minus PI. Ecce: eccentricity angle; obli: obliquity angle; lonP: the angle between the earth at autumnal equinox and the earth at perihelion. Sv: Sverdrup ($1 \text{ Sv} = 10^6 \text{ m}^3/\text{s}$). Changes of sea ice and AMOC in Pliocene experiments are reported as percentage changes relative to the PI.

Exp ID	PI	PlioMIP1	PlioH-CO2	PlioMin-Inso _{July} (3.049 Ma)	PlioMax-Inso _{July} (3.037 Ma)	PlioMax-Inso _{ann} (3.159 Ma)	PlioBS-C	PlioCAA	PlioCAA+BSC
ΔBoundary conditions	1850 reconstructions	PlioMIP I boundary conditions	450 ppm CO2	ecce = 0.0545	ecce = 0.0511	ecce = 0.0450	Closed Bering strait	Closed Canadian Arctic Archipelago Straits	Closed Bering strait and Canadian Arctic Archipelago Straits
				obli = 23.1432	obli = 23.6424	obli = 23.8868			
				lonP = 107.3686	lonP = 303.929	lonP = 186.3326			
	Climatology	Pliocene minus PI simulation results							
T _{2m} -global (°C)	13.26	1.83	2.21	2.12	1.91	1.88	1.85	1.5	1.88
T _{2m} -(70 ° - 90 °N)	-16.58	4.63	5.76	4.15	6.06	6.12	5.30	1.16	4.67
NHL sea ice Vol (10 ⁴ km3)	3.75	-53%	-63%	-39%	-71%	-70%	-58%	+20%	-54%
AMOC strength (Sv)	25.7	-4%	-5%	-4%	-9%	-9%	+6%	-16%	+14%

Results

1. Mean climate responses

Global mean temperature responses to Pliocene variations in orbital, CO₂ forcing, or Arctic gateway changes are small (≤ 0.5 °C). Temperature responses are amplified in the Arctic (70 – 90 °N) with 1-2 °C of warming in response to a higher CO₂, insolation forcing, or a closed Bering Strait comparing to PlioMIP 1 (Table 1). Across Pliocene simulations, mean ΔT -model of the circum-Arctic region (55 ° - 70 °N) shows a good correlation of -0.91 with mean Arctic sea ice volume. This result corroborates the findings by Ballantyne et al. (2013) and Howell et al. (2016a), confirming the strong connection between simulated Pliocene Arctic sea ice state and

surface temperatures of the circum-Arctic region, where most of the NHL terrestrial proxy records are located.

2. Patterns of Northern high latitude (NHL) surface temperature responses to Pliocene climate forcings

Our calculations of BIC as a function of the number of clusters show that beyond five, increasing the number of clusters does not increase BIC of ΔT -model in any simulation (Fig. S1). Therefore, five clusters are sufficient for the GMM model to describe the annual mean ΔT -model of NHL (see methods and Fig. S1). Spatial distributions of these clusters reveal zonally heterogeneous, yet self-coherent, structures (Fig. 1 and 2) that are similar among experiments, but cannot be captured using zonal averages. A chi-square test is performed to test the independence of cluster patterns between experiments; the calculated p-values are below 0.01, suggesting strong connection between the patterns. Excluding PlioCAA, most experiments feature similar cluster partitioning across the Arctic and adjacent continental region. We further explore the physical basis behind the partitioning of ΔT -model of NHL into different clusters.

Fig. 1. Spatial distribution of ΔT -model across the northern high latitudes (north of 55 °N). ΔT -model: Surface temperature differences of Pliocene minus Pre-industrial experiments. Insignificant values (below 98% confidence level) are hatched.

Fig. 2. Patterns of simulated northern high latitude surface temperature responses (north of 55 °N) to Pliocene climate forcings. Also shown on PlioMIP1 map are terrestrial proxy sites. Each proxy record is mapped according to the site location, confidence level (confidence increases with bubble size increase), reconstruction method and proxy source (indicated by colors). Records from the same sites are displaced ($\leq 2^\circ$ longitudes/latitudes) for display purpose.

Fig. 3. Median of ΔT -model (simulated Pliocene surface temperatures minus pre-industrial) (bars) and ΔT -proxy (Pliocene terrestrial proxy temperatures minus collocated 1901-1930 surface temperature observations) (filled squares) located geographically within each cluster mapped in Fig. 2. A vertical line on top of each bar spans the 5th and 95th percentile of ΔT -model of each cluster. Values of ΔT -proxy are grouped by cluster separation in space, confidence level (marker size increases with confidence), reconstruction method and proxy source (indicated by marker colors as shown in Fig. 2). If multiple ΔT -proxy values locate within the same cluster, having the same reconstruction method, proxy source and confidence level, a vertical line is drawn through the ΔT -proxy median to represent the maximum and minimum. The confidence level weighted mean difference (Diff) between ΔT -model and ΔT -proxy is shown in the center of each panel.

Cluster 5 represents the greatest warming of ΔT -model over Greenland for all experiments except PlioCAA (Fig. 3). This warming corresponds to reduction in thickness and extent of the Greenland ice sheet. In the PlioCAA, advection of cooler air from the North Atlantic and Labrador Sea suppresses the warming in the Greenland interior by 3 – 5 °C. This connection between the Greenland interior temperatures and temperatures of the North Atlantic and Labrador Sea is revealed by cluster analysis. The GMM model finds the highest possibility for ΔT -model of Greenland interior coming from the same Gaussian distributions as ΔT -model of the North Atlantic and Labrador Sea (cluster 1) instead of forming a separate distribution.

Cluster 3 and 4 represent warming of ΔT -model for the Arctic Ocean and the circum-Arctic regions including areas around the Bering Strait. The circum-Arctic warming, represented by Cluster 3, is up to 2 °C less than the Arctic warming represented by cluster 4. Examinations of changes in Arctic sea ice concentration (per grid cell) and absolute sea ice thickness suggest that both cluster 3 and 4 overlap with regions showing substantial reduction of sea ice concentration (over 20%) (Fig. 4a). Additionally, cluster 4 overlaps areas with extensive loss of ice thickness (Fig. 4b). Within cluster 4, thinner ice enhances the warming of surface air due to its warmer ice temperature and enhanced ablation of highly reflective snow (up to ~10% reduction of snow on thinned ice between Pliocene experiments and PI). In PlioCAA, the reduction of sea ice is restricted to the ice margin in the western Arctic, adjacent to the North American landmass. Consequently, the maximum warming of ΔT -model shifts towards the North America-western Arctic region (represented by cluster 5).

Clusters 1 and 2 (clusters 2 and 3 in PlioCAA) are composed of ΔT -model between 55 ° and ~65 °N (Fig. 2). ΔT -model represented by these clusters scales well with anomalies in sea ice concentration around the Arctic sea ice margins (Fig. 5a and b), highlighting the importance

of Arctic sea ice retreat in defining these cluster patterns. ΔT -model in cluster 1 (cluster 2 for PlioCAA) show less warming than ΔT -model in cluster 2 (cluster 3 for PlioCAA). This distinction reflects influences from high topography (Fig. S2) and Atlantic meridional overturning circulation (AMOC).

Weaker surface warming in cluster 1 (cluster 2 for PlioCAA) collocate with high topography of western North America and eastern Eurasia (Fig. S2). Due to a moister Pliocene atmosphere, annual mean snowfall and surface accumulation enhance around these high terrains especially during boreal spring and winter. The resulted greater surface snow-cover (Fig. S3) increases the heat capacity and albedo of the land surface, reducing warming of ΔT -model around these high terrain features. ΔT -model shows little correspondence to changes in geopotential height (both 600 hPa and 300 hPa) at either annual or seasonal time scales, indicating an unimportant role from atmospheric dynamical changes.

Fig. 4. Differences in Northern Hemisphere annual grid-mean sea ice concentration (a) and absolute sea ice thickness (absolute sea ice thickness = grid-mean sea ice thickness/grid-mean sea ice concentration) (b) between Pliocene and Pre-Industrial simulations.

A weaker surface warming represented by cluster 1 also occurs in the North Atlantic in experiments except PlioBSC, PlioCAA and PlioCAA+BSC (Fig. 1). This weaker warming results from a stable AMOC in these cases (Table 1). In contrast, a stronger AMOC in PlioBSC and PlioCAA+BSC (Table 1) leads to a greater warming in the North Atlantic, which is grouped into cluster 2 (greater warming); whereas, a noticeably weaker AMOC in PlioCAA leads to cooling of the North Atlantic grouped in cluster 1. AMOC responses to gateway closure are results of changing pathways of freshwater transport between the Arctic and the North Atlantic as reported in Otto-Bliesner et al., (2016).

Fig. 5. Regression of annual medians of ΔT -model located geographically within clusters 1 (a) and 2 (b) with annual differences in Arctic sea ice concentration (%) between Pliocene and PI simulations.

3. Comparison with terrestrial proxy data from NHL

We compare NHL ΔT -model with NHL ΔT -proxy using the cluster patterns of individual experiments shown in Figure 2. Medians of ΔT -model (bar chart) and ΔT -proxy of proxy sites (filled squares) geographically located within each cluster are shown in Figure 3. ΔT -proxy is first grouped based on reconstruction method and proxy source (different colors in Figs. 2 and 3) and confidence levels (different marker sizes in Figs. 2 and 3) before calculating the medians. Notice that although the general cluster pattern remains similar among experiments, the spatial

coverage by individual clusters shows subtle differences, which alters the partitioning of ΔT -proxy and hence the calculated ΔT -proxy medians. Changes in the partitioning of ΔT -proxy are the clearest in PlioMinInso_{July} and PlioCAA. In PlioMinInso_{July}, the warming represented by cluster 3 is limited to the periphery of Arctic Ocean and continental margins, whereas cluster 3 covers most of Siberia in other experiments. In PlioCAA, the overall cluster pattern rotates clockwise towards the western Arctic region while clusters center around the Greenland interior and Atlantic side of the Arctic in other experiments. The distinctive cluster partitioning in PlioMinInso_{July} and PlioCAA is unsupported by terrestrial proxy data, as shown by different shapes of changes of ΔT -proxy and ΔT -model medians following the same cluster partitioning from cluster 1 to 5 (Figure 3).

The cluster partitioning of ΔT -proxy is fairly similar among the remaining experiments. The resulting medians of ΔT -proxy share a similar first-order characteristic with the medians of ΔT -model, showing strong regional amplification of NHL surface warming. Based on cluster patterns, we estimate ΔT -model of PlioMIP1 in the NHL is on average $\sim 10^\circ\text{C}$ colder than ΔT -proxy (the calculation of mean difference between ΔT -proxy and ΔT -model is described in the Method). A higher CO_2 level (450 ppm), higher insolation forcing, or closed Arctic gateways reduces this cold bias by a 1.3°C , a 1.8°C , and a 1.3°C , respectively (Fig. 3).

4. CCSM4 sensitivity to prescribed Pliocene radiative forcing and Arctic gateway changes

Most of our mPWP experiments display similar patterns of terrestrial warming in the NHL. This observation holds true even for experiments with very different climate forcings. For example, between PlioMaxInso_{July} and PlioMIP1, 65°N summer insolation increases by 32 W/m^2 and is compensated by similar decreases in winter insolation; in PlioBSC and PlioCAA+BSC,

the main forcing is the change in freshwater transport into and out of the Arctic (Otto-Bliesner et al., 2016). In order to determine processes that are responsible for the similar patterns of temperature responses, we conduct energy balance calculations.

In all experiments, the energy balance analysis reveals a warming effect from reduction in top of the atmosphere clear-sky emissivity ($\varepsilon\text{-clr}$) (Fig. 6), which is further enhanced by reduction of cloudy-sky emissivity ($\varepsilon\text{-cld}$) and surface reflection ($\alpha\text{-srf}$) (except in the PlioCAA). Negative feedbacks to warming come from enhanced cloud reflection ($\alpha\text{-cld}$), notably in the maximum insolation simulations (PlioMaxInso_{July} and PlioMaxInso_{Ann}), and from a reduction in ocean heat transport (H-ocn), particularly in PlioHCO₂, PlioCAA, PlioMaxInso_{July}, and PlioMaxInso_{Ann}. The reduction of H-ocn can be further explained by the weaker AMOC in these experiments (Table 1). Changes in H-ocn, however, are mostly compensated by H-atm. The combined atmosphere and ocean heat transport into the NHL contributes little to the ΔT -model, even for experiments featuring major changes in ocean dynamics due to Arctic gateway closure (PlioCAA, PlioCAA+BSC, and PlioBSC). Closing the Arctic gateways does alter heat transport and surface temperatures (~ 3 to 4 °C warming of PlioBSC and PlioCAA+BSC, ~ -8 to -4 °C cooling in PlioCAA) in the mid-latitudes of North Atlantic (Otto-Bliesner et al., 2016). The effect of increased ocean heat transport in PlioBSC and PlioCAA+BSC, however, does not reach broad areas of NHL (does warm the North Atlantic in Fig. 1). And the effect of decreased ocean heat transport in PlioCAA is compensated by atmospheric heat transport (Fig. 6).

Reduction in surface reflection can largely be explained by sea ice melting in the Arctic Ocean, which is accompanied by the reduction of $\varepsilon\text{-clr}$ (Fig. 6). Changes in $\varepsilon\text{-clr}$ are determined by atmosphere CO₂ and moisture content. CO₂ concentration is the same for experiments except PlioHCO₂. The good correspondence between reduction in surface albedo and clear-sky

emissivity suggests a positive feedback from atmosphere moistening and enhanced downwelling longwave radiation in response to less sea ice, which is consistent with Ballantyne et al. (2013).

Fig. 6 Contributions to ΔT -model (raster color) from different radiative changes (x-axis) among experiments (y-axis) averaged between 55 ° to 90 °N of all grid points, land and ocean. ϵ : top of the atmosphere emissivity; α : planetary albedo; H-ocn and H-atm: ocean and atmosphere heat transport (integrated heat convergence of 55 ° - 90 °N); Inso: insolation; topo: topographic changes (reduced Greenland elevation due to removal of the icesheet according to PRISM3 reconstructions); srf: surface; clr: clear-sky; cld: cloudy-sky.

Given small contributions to NHL energy budget from direct changes in ocean dynamics and insolation forcing, it is unsurprising that even with quite different forcings, experiments show similar patterns of warming amplification (Fig. 2). This similarity highlights the strong feedbacks within climate system of NHL; energy perturbations, even at regional scale, can lead

to similar and broad-scale responses across the entire NHL region through positive albedo and water vapor feedbacks.

Discussion

1. Implications for proxy reconstructions

We demonstrate that model-proxy data mismatch in the default CCSM4 PlioMIP1 simulation ($\sim 10^\circ\text{C}$) is systematic rather than site-specific across the terrestrial environment of the NHL. Model-proxy data mismatch can be reduced by warming effects from closing Arctic gateways, higher CO_2 , and insolation forcings. Several studies have reviewed the evidence for closed Arctic gateways during the mPWP (Haywood et al. 2016b; Dowsett et al., 2016). However, the timing of gateway closure and opening (particularly, the Bering Strait) remain debated (Dowsett et al., 2016). For example, based on marine palynological record from the Iceland, cooling and development of a modern-like Nordic Sea ocean circulation starting at 4.5 Ma is linked to closure of the Central American Seaway and opening of the Bering Strait (De Schepper et al., 2015). If terrestrial proxies recorded climate conditions when the Bering Strait was open and Canadian Archipelago Straits was closed, a greater model-proxy data mismatch, similar to the case of PlioCAA, is expected.

Moreover, it is unclear whether or not NHL terrestrial proxies are biased towards periods of high CO_2 , high insolation, or even warm season signals. A data synthesis suggests that the mean CO_2 level is higher (up to 500 ppm) during the Early Pliocene (between $\sim 4.6 - 3.8$ Ma) but declines to between 280 to 400 ppm during the late Pliocene (since 4 Ma) (Haywood et al., 2016a). As a result, a 450 ppm $p\text{CO}_2$ may be too high for the mPWP. Nonetheless, due to uncertainties in CO_2 estimates for the mPWP (350 ppm to 450 ppm, Haywood et al., 2016a) and

poorly constrained ages of terrestrial temperatures (Supplementary Materials), it is difficult to rule out higher $p\text{CO}_2$ as a potential cause for some of the documented NHL warmth.

Most high latitude proxy sites cannot resolve orbital time scale variability. One exception is the Lake El'gygytgyn core from northeastern Russia (55 °N), which contains high-resolution pollen records with clear 41-kyr cycles, indicative of changes in obliquity during the late Pliocene and early Pleistocene (Brigham-Grette et al., 2013). This record reveals ~2 °C millennial-scale variations in warm season temperatures that are strongly correlated with the benthic $\delta^{18}\text{O}$ stack (Lisiecki and Raymo, 2005) (Brigham-Grette et al., 2013). This magnitude of variation is comparable to our ~3 °C summer warming in the northeastern Russia between PlioMaxInso_{Ann} (mostly obliquity forced) and PlioMIP1. Orbital time scale variability in annual insolation is shown to be a strong driver of NHL bio-productivity (e.g. Tabor et al., 2014). Therefore, sampling bias towards periods of high annual insolation and high bio-productivity could be an important factor for interpreting the warm NHL environment documented by late Pliocene biomes.

Finally, new terrestrial fossil flora records from the Canadian Arctic suggest little difference in the surface temperatures of Meighen Island between the early and late Pliocene (Fletcher et al., 2017, personal communication). One explanation for this similarity is the taphonomic bias in proxy data that preferentially record warm and wet periods. Similarly, studies have pointed out possible warm season bias in productivity-based proxies such as tetraether composition, tree ring width, and isotopic compositions (e.g., Ballantyne et al., 2010). Additionally, warm- and cool-taxa may coexist in the Arctic (e.g. Ballantyne et al., 2010), leading to relatively large confidence intervals on temperature estimates derived from conventional coexistence approaches. As a result, quantification of true model-proxy data

mismatch in NHL requires better age constraints, high-resolution proxy data, and better understandings of fossil biome taphonomy in combination with advanced statistical techniques.

2 Implications for coupled Earth system model simulations

The model-proxy mismatch of NHL terrestrial temperatures may also come from model biases or climate forcings that are missing in our Pliocene simulations. Specifically, coupled models may produce overly resistant Arctic sea ice that fails to retreat with mPWP climate forcings. Quantitative reconstruction of mPWP sea ice distribution does not exist. However, terrestrial fossil assemblages and river deposits from the Canadian Arctic and marine fossil assemblages around Meighen Island and north of coastal Alaska all suggest a seasonally productive high Arctic environment that was likely seasonally sea ice free (Polyak et al., 2010). More recently, using the semi-quantitative biomarker IP₂₅ (Ice Proxy with 25 carbon atoms), Knies et al. (2014) reconstruct potential extents of annual Arctic sea ice during the middle/late Pliocene (~4 – 2.6 Ma). Their results suggest that Pliocene annual Arctic sea ice extent is close to present-day summer extent. Simulated Arctic sea ice is greatly reduced, and becomes seasonally absent, in 4 out of 8 participating models in PlioMIP1 (Howell et al., 2016a). However, among these 4 models, the simulated low Arctic sea ice state of the mPWP is more of a fortuitous fit given their substantial underestimation of present-day Arctic sea ice thickness (Shu et al., 2015; Howell et al., 2016a). Howell et al. (2016b) and Ballantyne et al. (2013) have explored whether the forced reduction of Arctic sea ice by lowering sea ice albedo or prescribing ice cover can help explain some of the terrestrial northern high latitude warmth of the mPWP. Their results are promising. However, the physical mechanisms and forcings responsible for a low sea ice state are still being investigated.

Conclusion

Motivated by the wide gap between proxy reconstructions and model simulations of NHL terrestrial temperatures in PlioMIP1 simulations, we conduct a series of sensitivity experiments using CCSM4 to explore potential contributors to NHL warmth during the mid-Piacenzian warm period (mPWP, 3.264 – 3.025 Ma) from variations in orbital forcing, CO₂ level, and closing of Arctic gateways.

Simulated surface temperature differences between the Pliocene and preindustrial can be explained by regional heterogeneity of surface topography, reduction in concentration and thickness of Arctic sea ice and the Greenland ice sheet, and changing strength of the North Atlantic meridional overturning circulation. The resulting patterns of surface temperature response reflect the amplification of NHL surface warming.

Radiative (changes in CO₂ level and orbital parameters) and ocean dynamical (gateway closure) forcings lead to similar patterns of terrestrial temperature responses in the NHL. This similarity can be explained by strong positive feedbacks from changes in emissivity and surface albedo, which distribute energy perturbations to the entire NHL climate system. Individual forcings of elevated CO₂ level (by 50 ppm), high summer/annual insolation of NHL, and closed Arctic gateways may explain 1 - 2 °C of the terrestrial model-proxy data mismatch in the NHL. These results highlight the importance of understanding/reconstructing Arctic warmth in the context of varying radiative forcings and changes in Arctic gateway configurations during the Pliocene.

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