A method for using a fully coupled climate system model to generate detailed surface boundary conditions for paleoclimate modeling investigations: an early Paleogene example

Jacob O. Sewall\textsuperscript{a,}\textsuperscript{*}, Matthew Huber\textsuperscript{b}, Lisa Cirbus Sloan\textsuperscript{a}

\textsuperscript{a}Department of Earth Sciences, University of California, Santa Cruz, CA 95064, United States
\textsuperscript{b}Department of Earth and Atmospheric Sciences, Purdue University, West Lafayette, IN, United States

Accepted 8 March 2004

Abstract

Although fully coupled models of the earth system are now common, simpler model architectures maintain significant utility, and scientific investigations aimed at understanding paleoclimates are frequently conducted with fixed sea surface temperature (SST) or slab ocean modeling experiments. One of the challenges facing the paleoclimate community is that the proxy data used to generate SST boundary conditions exist at a finer resolution and with very limited spatial coverage when compared to a climate model. In addition, SST proxy estimates often represent a single season or annual average conditions. This mismatch in coverage and resolution frequently results in paleoclimate modelers using SST distributions that have very limited spatial and temporal variability. In many regions, a spatially and temporally detailed SST distribution may be necessary for the accurate reproduction of paleoclimatic conditions. Here we borrow from the concept of flux correction and, using available proxy estimates of SST as our guide, force a fully coupled earth system model to produce a spatially and temporally detailed SST distribution for the paleoclimate of the early Paleogene (~45–65 Ma). The SST values we produce represent a conservative estimate of early Paleogene high latitude SSTs and match tropical temperatures for this time period well. In addition to matching proxy estimates, our model-derived SST distribution has spatial and temporal variability that meshes well with global climate model resolution. This detailed SST distribution is now available to us as we investigate the causes and sensitivities of early Paleogene climate in fixed SST and slab ocean modeling experiments. The method we used to generate this spatially and temporally detailed SST distribution may prove useful for those investigating other time periods in the past, or the future, for which detailed model boundary conditions are unavailable.

\textsuperscript{*} Corresponding author. Tel.: +1 831 459 3504; fax: +1 831 459 3074.
\textit{E-mail address:} jsewall@es.ucsc.edu (J.O. Sewall).

© 2004 Elsevier B.V. All rights reserved.

Keywords: Cenozoic; Paleoclimatology; Ocean temperature; Modeling
1. Introduction

The investigators of deep paleoclimates (those more than 10 Ma) frequently encounter challenges not seen by those who investigate more recent climates and climate change. One of the more persistent of these challenges is the limited knowledge of boundary and initial conditions (e.g. sea surface temperatures (SST)) at sufficient spatial and temporal resolutions to allow reproduction of climate characteristics and variability as indicated by data. One time period where this challenge presents itself is the early Paleogene (~45–65 Ma). We know from proxy climate indicators that the early Paleogene was a time period of global warmth with global mean annual temperatures of 17–21 °C (Basinger, 1991; McIntyre, 1991; Wing and Greenwood, 1993; Greenwood and Wing, 1995; Wilt, 2000; Johnson and Ellis, 2002) and exceptional warmth from 50 to 53 Ma (Zachos et al., 2001). To date, paleoclimate scientists have not completely deciphered how early Paleogene climate achieved such warmth; modeled early Paleogene climate and proxy data climatic estimates frequently disagree (e.g. Sloan and Barron, 1992; Sloan, 1994; Sewall et al., 2000; Huber and Sloan, 2001; Shellito et al., 2003). While this outstanding question remains the focus of a significant body of research (Sloan and Pollard, 1998; Sloan et al., 2001; Shellito et al., 2003), the early Paleogene, as a greenhouse period, has the potential to provide a wealth of information as to how the earth system is affected by and functions under such climatic conditions.

Although a complete assessment of the causes of earlyPaleogene climate will doubtless require investigations with fully coupled earth system models, many of the consequences and dynamics of early Paleogene climate can be investigated with simpler model architectures. Those architectures most commonly used are experiments with prescribed sea surface temperatures (SSTs) (e.g. Sloan and Barron, 1992; Sewall et al., 2000; Sloan et al., 2001) and those incorporating a slab ocean wherein SSTs are not specified but the rate of heat transfer in the surface ocean is (Sloan, 1994; Sloan and Rea, 1995; Shellito et al., 2003). Both of these are frequently used, powerful methods for investigating paleoclimates (Vavrus, 1999; Gibbs et al., 2002; Haywood et al., 2002a; Haywood et al., 2002b); however, both also require some foreknowledge of global sea surface conditions in the past. As noted above, detailed global datasets for deep paleoclimate are nonexistent and early Paleogene SSTs are no exception. The same is true for the oceanic heat flux values needed to run a slab ocean model; in fact, those fluxes are traditionally derived from detailed SST datasets (e.g. Haywood et al., 2002a; Maloney and Kiehl, 2002; Vettoretti and Peltier, 2003).

What we do know about early Paleogene SSTs is very general. From terrestrial and marine proxy climate indicators, we may deduce that Arctic coastal high latitude temperatures were seasonally warm; they hovered near the freezing point of sea water (~1.8 °C) during winter months (Greenwood and Wing, 1995; LePage, 2003) and reached 10–15 °C during the spring, summer, and fall (Tripati et al., 2001). North Atlantic temperatures were significantly warmer than present (winter SST up to 18 °C; Andreasson and Schmitz, 2000) and southern high-latitude SSTs ranged from 10 to 15 °C (Zachos et al., 1994; Dutton et al., 2002). Early Paleogene tropical and subtropical SSTs were similar to or warmer than those of today by 4–12 °C (Thomas et al., 1999; Andreasson and Schmitz, 2000; Kobashi et al., 2001; Pearson et al., 2001; Tripati et al., 2003; Zachos et al., 2003).

As we noted previously, modeled early Paleogene climate does not precisely match proxy estimates. One of the areas of mismatch is Arctic winter temperatures which modeling experiments with a variety of climate forcings consistently underpredict even while matching SSTs at other latitudes reasonably well (Sloan et al., 1999; Huber and Sloan, 2001; Shellito et al., 2003). We are, however, certain that warm Arctic temperatures (0 to −2 °C; (Greenwood and Wing, 1995; Tripati et al., 2001; LePage, 2003) did exist and that relatively warm Arctic Ocean SSTs probably played a critical role in maintaining mild circum-Arctic temperatures and may have had significant influence on extrapolar climate as well. While most previous early Paleogene studies have utilized zonally constant SST distributions (Sloan and Barron, 1992; Sewall et al., 2000; Sloan et al., 2001), if we wish to either investigate the influence of such warm SSTs or explore the response of a climate with such SSTs to other climatic forcings, we must have a detailed SST distribution for the early Paleogene. Given the limited number of data tie points for such a SST distribution, there are many methods by which we could generate
our early Paleogene SSTs. In this paper, we present one such method where we used a fully coupled earth system model to produce one possible early Paleogene SST distribution that agrees with the broad constraints provided by proxy climate data.

2. Methods

While proxy data give us a general idea of what early Paleogene SSTs should be, we cannot achieve those values in a fully coupled climate model experiment (Sloan et al., 1999; Huber and Sloan, 2001; Shellito et al., 2003). Similarly, in many coupled climate system models, modern SST distributions cannot be achieved purely via integrating model physics; in those models, flux corrections are applied to return the modeled state to the known state. In these modern cases, the known state is frequently better defined than that of the early Paleogene, however, we can employ a similar technique. The main differences between our technique and traditional flux corrections are that (1) we do not know precisely what flux is necessary to achieve early Paleogene conditions so we must determine possible values empirically; thus, our method is more accurately “flux addition,” and (2) we strive not to perturb the overall energy balance of the model as we do not know beforehand the exact magnitude or origin of the missing flux. As a result of knowing only the result and not the origin of the missing flux, we choose to simplify the correction process by applying additions to fluxes from only one component model, the atmosphere. We apply additions to the atmospheric sensible heat flux to both the ice and ocean component models.

Sensible heat flux between the component models can be generally formulated as:

\[ Q_{\text{sen}} = \rho_a c_p^a C_{H} W_{10}(T_a - \theta_1), \]

where \( Q_{\text{sen}} \) is the sensible heat flux, \( \rho_a \) is the air density, \( c_p^a \) is the specific heat of air, \( C_{H} \) is the transfer coefficient for heat, \( W_{10} \) is the wind speed at 10 m and equals \((U_{10} + V_{10})^{1/2}\) where \( U_{10} \) and \( V_{10} \) are the zonal and meridional winds at 10 m, \( T_a \) is the 2-m air temperature, and \( \theta_1 \) is the sea surface temperature.

Our additional heat flux is applied as a correction to \( T_a \) prior to the calculation of \( Q_{\text{sen}} \). This correction is applied only in Arctic regions (those poleward of 60°N) where values of \( T_a \) below freezing are increased to just above freezing.

We make slight adjustments to \( T_a \) until we achieve a SST distribution that is consistent with proxy constraints. Our main consistency check is the seasonality, thickness, and extent of sea ice cover. Atmospheric temperatures such as those indicated by proxy data would correspond to a highly seasonal sea ice regime with ice free conditions for much of the year and limited seasonal ice cover during peak ice months (February–May). Such sea ice would have reduced concentrations (ice fractions under 100%) and limited thickness due to the lack of multi-year pack ice. Once we have achieved a stable sea ice regime, we consider our SSTs to be equilibrated.

We conducted our research with the National Center for Atmospheric Research (NCAR) Community Climate System Model version 1 (CCSM1) (Journal of Climate, v.6 1998; (Boville et al., 2001) at a spectral resolution of T31 (~3.75°lat×3.75°lon). Due to the risk of perturbing the overall planetary energy budget and limited constraints on the early Paleogene energy budget, we determined the magnitude of our fluxes in a present-day ground truth experiment. We initialized our ground truth experiment from an equilibrated modern-day scenario (Otto-Bliesner et al., 2002) and, over the course of 150 years, determined an adequate \( T_a \) adjustment for achieving a nearly seasonally ice-free Arctic Ocean under present-day climate conditions. Our final \( T_a \) adjustment values are 1 °C for calculation of fluxes to the ice component model and 2 °C for calculation of fluxes to the ocean component model. Our \( T_a \) adjustment is equivalent to adding an additional annual averaged flux of 200–225 W/m² over most of the central Arctic. We applied this flux for an additional 54 years to ensure that the ice cover and SSTs were in equilibrium. Though the ice conditions in our ground truth experiment represent temperatures cooler than early Paleogene proxy data indicate, our modern experiment does not have the elevated greenhouse gas concentrations that many researchers believe were characteristic of the early Paleogene (Pearson and Palmer, 2000; Retallack, 2001; Shellito et al., 2003). Consequently, we deem this flux adequate to produce the desired Arctic ice cover and SSTs under early Paleogene conditions. Comparisons between residual energy at the top of the model in our ground truth experiment and the modern
energy balance, we applied this flux, once again in the form of adjusted values for $T_a$, to an early Paleogene experiment (henceforth known as WARC—Warm ARctic Case). We initiated WARC from the final year of an equilibrated (trends in deep ocean temperature are negligible), fully coupled early Paleogene model run (our Cold ARctic Control case, henceforth known as CARC) that incorporates realistic early Paleogene vegetative and topographic boundary conditions (Sewall et al., 2000) and elevated greenhouse gas concentrations (1120 ppm CO$_2$ and 700 ppb CH$_4$). WARC differs from CARC only in that our additional heat flux is applied to northern high latitudes in WARC. Given the increased greenhouse gas concentrations in WARC, our baseline values for $T_a$ are significantly higher than those in our ground-truth modern-day experiment. Consequently, our adjusted $T_a$ values are equivalent to an annual average additional heat flux of only up to 60 W/m$^2$ over Arctic regions. We integrated WARC for 200 years until there was no trend in sea ice concentration or thickness (Fig. 1) and averaged the final 50 years of WARC for comparison to 50-year average fields from CARC.
3. Results

3.1. Sea ice

Arctic sea ice shows a significant decrease between CARC and WARC. Winter ice concentration decreases by 25–100% (Fig. 2) and ice thicknesses decrease by 0.25–0.6 m. Absolute winter ice thicknesses in WARC are less than 0.5 m (Fig. 3). March, April and May (MAM), the peak ice season, shows a 25–100% decrease in ice concentration (Fig. 2). WARC has 0.5–0.8 m thinner MAM ice than CARC (Fig. 2) with actual ice thicknesses of 0.5 m or less (Fig. 3). Ice melt off in WARC is complete by June and the Arctic Ocean remains ice free through December (Fig. 3). This is in contrast to CARC where ice cover remains through July and reappears in November (Fig. 3). WARC has 5 more ice-free months than CARC.

3.2. Sea surface temperatures

WARC Arctic SSTs have annual average values of 0–5 °C, winter (December, January, February; DJF) values of −0.5 to 2.5 °C, and summer (June, July, August; JJA) values of up to 7.5 °C (Fig. 4). JJA values are 2–4 °C warmer than SSTs in CARC, DJF values are 1–2.5 °C warmer and the annual average difference between WARC and CARC Arctic SSTs is 1.5–2.5 °C (Fig. 5).

WARC tropical SSTs range from 25 to 35 °C in all seasons and show no difference from CARC tropical SSTs (Figs. 4 and 5). Mid-latitude SSTs also show no appreciable difference between WARC and CARC (Fig. 5). High latitude southern hemisphere SSTs are similar in WARC and CARC and, even in austral winter (JJA), never reach the freezing point of seawater.

3.3. Atmospheric temperatures

DJF surface air temperatures in WARC are 1–10 °C warmer than those in CARC over the Arctic Ocean.
and Arctic maritime locations (those within 1000 km of the coast) (Fig. 6b). Temperature differences over the rest of the planet are negligible (Fig. 6b). JJA surface temperatures show a similar pattern with circum-Arctic warming of 1–6 °C (Fig. 6c). Annual average temperature differences between WARC and CARC are 1–6 °C and exist only in Arctic regions (Fig. 6a).

3.4. Other climate variables

Other than sea ice, SSTs, and surface air temperatures, differences between WARC and CARC are negligible. Circum-Arctic 500 mb geopotential heights in WARC increase slightly over those in CARC with annual average values 0.1–0.4 km higher in WARC (not shown). Annual average precipitation (not shown) shows no difference between the two cases. Relative humidity (not shown) is slightly lower in WARC compared to CARC and low-level Arctic clouds in WARC decrease by up to 15% compared to CARC. Those differences are most pronounced in DJF but persist in all seasons. Incoming JJA short-wave radiation at the Arctic surface and at the top of the atmosphere is up to 20 W/m² greater in WARC than in CARC (not shown). Global, annual averaged
residual energy at the top of the model is 0.007 W/m² larger in WARC than it is in CARC.

4. Discussion

Overall, our imposed additional heat flux at high northern latitudes results in circum-Arctic warming and a reduction in sea ice coverage with no significant impact on the remainder of the global climate system. There is no extrapolar warming; warming, both due to our imposed heat flux and positive feedbacks, is limited to Arctic locations (Fig. 6). We note here that the absence of extrapolar warming, particularly in northern hemisphere continental interiors, is counter to expectations that researchers have previously presented (e.g. Sewall and Sloan, 2001; Jahren and Sternberg, 2002) and supports the conclusion drawn from other work that warm ocean temperatures have little influence on early Paleogene continental interior climates (e.g. Sloan, 1994; Huber et al., 2003). Unlike modern climate scenarios where spatial patterns of surface pressure changes can alter mid-latitude climates (Thompson and Wallace, 1998), there is no significant change in surface pressure between WARC and CARC and temperature and geopotential height anomalies are zonally symmetric. This lack of spatial character in temperature and geopotential height anomalies, as well as the lack of a surface pressure anomaly, may well explain the lack of extrapolar response to our imposed high latitude forcing. In addition, our imposed forcing is applied only in winter months over the Arctic and, as such, is a relatively small perturbation to the global energy budget, therefore, we might expect that extrapolar responses would be small.

With regards to sea ice, there is a dramatic increase in the seasonality of Arctic sea ice in WARC as compared to CARC (Fig. 3). The resulting sea ice distribution with thin, limited winter ice cover is in agreement with the most conservative estimates of early Paleogene DJF Arctic temperatures (−1 to −2 °C; Greenwood and Wing, 1995; LePage, 2003); such winter temperatures hover near the freezing point of sea water and would produce thin, discontinuous ice cover just as we see in our experiment (Fig. 3). The increase in the number of ice-free months reduces Arctic surface albedo and contributes to an increase in JJA and mean annual solar insolation absorbed at the Arctic surface (up to 20 W/m²). Increased solar insolation can also be partially attributed to a cloudiness feedback. Warmer Arctic temperatures (Fig. 6) drive a slight decrease in relative humidity (not shown) that in turn reduces low-level clouds (not shown) by up to 15%. While this decrease is most pronounced in DJF when temperature differentials between WARC and CARC are greatest (Fig. 6a), it persists in all seasons and can thus contribute to increased JJA solar insolation at the Arctic surface. Increased solar insolation can, in turn, contribute to surface warming and a further reduction in albedo and, thus, represents a positive feedback in the system as the Arctic warms. In spite of pronounced Arctic warming, the difference in annual averaged residual energy at the top of the model varies between WARC and CARC by 0.007 W/m², significantly less than the interannual variability within either WARC or CARC (~14.82 W/m²).

The WARC SST distribution, the ultimate goal of this experiment, is in reasonable agreement with proxy climate indicators. Our tropical SSTs are ~3 °C warmer than modern values and differ not at all from our control case; they are also similar to SSTs predicted in previous modeling work (Huber and Sloan, 2001). These SST values, while warmer than present values, agree with new data that indicate warming of early Paleogene tropical SSTs (Pearson et al., 2001; Tripati et al., 2003; Zachos et al., 2003). Our southern high latitude temperatures also display no difference from CARC (Fig. 5), agree with both terrestrial and marine proxy data that indicate the absence of southern hemisphere sea ice (e.g. Ditchfield et al., 1994; Greenwood and Wing, 1995; Dingle et al., 1998; Francis and Poole, 2002) and match proxy derived estimates for the middle Eocene (~11 °C; Zachos et al., 1994) well although they are up to 5 °C cooler than estimates for the earliest Eocene (~14–15 °C; Zachos et al., 1994; Dutton et al., 2002). Our mid to high-latitude Atlantic SSTs also display no difference from CARC (Fig. 5) and match proxy estimates of winter (15–25 °C) summer (25–30 °C) and mean annual (26–27 °C) SSTs reasonably well (Andreaesson and Schmitz, 2000; Kobashi et al., 2001). Arctic SSTs do, however, display significant warming when WARC is compared to CARC (Fig. 5). These relatively warm Arctic SSTs are still 2.5–5 °C too cool in the summer when
compared with marine proxy data (Tripathi et al., 2001) but are a significant improvement over temperatures produced in other modeling work (Huber and Sloan, 2001) and agree well with the winter air temperatures predicted from some terrestrial proxies (Greenwood and Wing, 1995; LePage, 2003). In addition to the quantitative match with proxy data, our modeled global SST distribution displays significantly more spatial variability than the zonally averaged SST distributions that researchers have used in previous early Paleogene experiments (Sloan and Barron, 1992; Sewall et al., 2000; Sloan et al., 2001).

Although our experiment was not run for sufficient time to equilibrate the deep ocean, we are satisfied that the lack of a trend in either SSTs (not shown) or sea ice concentration and thickness (Fig. 1) indicates surface equilibrium. As the SST distribution is within the parameters suggested by proxy data in the tropics and, particularly, northern high latitudes, we believe that our modeled Arctic SSTs represent one possible early Paleogene distribution. In addition, given the insignificant perturbation to global climate, particularly the residual at the top of the model, that our additional flux engenders, we are confident that using this SST distribution to investigate the functioning, dynamics and consequences of early Paleogene climate is both appropriate and a significant step forward in understanding this fascinating time period.

5. Conclusions

By adding an additional high northern latitude heat flux to our simulation of early Paleogene climate, we generate early Paleogene SSTs, particularly in the Arctic Ocean, that agree well with proxy climate indicators yet retain the spatial and temporal variability characteristic of general circulation model results. These SSTs have substantially more detail in their character than the zonally averaged values traditionally used in investigations of early Paleogene climate (Sloan and Barron, 1992; Sewall et al., 2000; Sloan et al., 2001). In generating these SSTs, we do not significantly impact global climate outside of the Arctic nor do we perturb the energy balance of the coupled climate model. In addition, our experiment provides the first quantification of the flux necessary to achieve early Paleogene northern high latitude warmth (~60 W/m²). While the natural sources of this additional high northern latitude heat flux remain to be discovered, it is likely that the additional warmth is a result of trapping, rather than transporting, more heat in the region (e.g. Sloan et al., 1999). Although a fully coupled earth system model will be required to completely quantify how summer stored heat is maintained throughout the winter in northern high latitudes, our SST distribution will allow detailed investigations into the dynamics of early Paleogene climate with simpler, more efficient, model configurations.

Acknowledgements

The authors would like to thank NSF for funding support (NSF-OPP0116941) and NCAR for computing support. NCAR is funded by NSF.

References


