

Temperatures at the last interglacial simulated by a coupled ocean-atmosphere climate model

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Abstract. The last interglacial (Eemian, 125,000 years ago) has generally been considered the warmest time period in the last 200,000 years and thus sometimes been used as a reference for greenhouse projections. Herein we report results from a coupled ocean-atmosphere climate model of the surface temperature response to changes in the radiative forcing at the last interglacial. Although the model generates the expected summer warming in the northern hemisphere, winter cooling of a comparable magnitude occurs over North Africa and tropical Asia. The global annual mean temperature for the Eemian run is 0.3°C cooler than the control run. Validation of simulated sea surface temperatures (SSTs) against reconstructed SSTs supports this conclusion and also the assumption that the flux correction, fitted for the present state, operates satisfactorily for modest perturbations. Our results imply that contrary to conventional expectations, Eemian global temperatures may already have been reached by the mid 20th century.

1. Introduction

The last interglacial (Eemian; 125 ka) has generally been considered the warmest time period of the last 200,000 years [Kellog, 1977; Hansen *et al.*, 1988; Budyko and Izrael, 1990; Zubakov and Borzenkova, 1990]. In addition to widespread evidence for warmth, global sea level may have been 6 m higher than at the present [Mesolella *et al.*, 1969; Bloom *et al.*, 1974; Ku *et al.*, 1974; Dodge *et al.*, 1983]. Estimates of global temperature increase for the Eemian range from 0.5° to 2.0°C greater than the present level. The high end of these estimates is comparable to the most recent best guess estimate [Houghton *et al.*, 1996] of the global temperature increase near the end of the next century by the Intergovernmental Panel on Climate Change (IPCC).

However, assessments of global sea surface temperature (SST) estimates for the last interglacial [Climate: Long-Range Investigation, Mapping, and Prediction (CLIMAP) Project Members, 1984] suggest that values were not significantly different from present ones. This is consistent with energy balance model calculations [Crowley, 1990] and observations [Barnola *et*

al., 1987], indicating that CO₂ levels were not significantly higher than in the Holocene and that on average, orbital insolation forcing causes only small changes in net global insolation receipt. The reason why the widespread evidence for warming did not translate into higher global average temperatures is because of seasonal and geographic biases in proxy climate indices (e.g., vegetation may be more sensitive to summer warming due to Milankovitch forcing, whereas in the winter, plants are almost dormant) and because the timing of peak warmth was not synchronous in all regions; for example, peak warming in southern and northern hemisphere records is almost 180° out of phase in some sectors [Crowley, 1990].

Despite prior work, conclusions about last interglacial warmth remain open to challenge. To address this problem further, we have used a coupled ocean-atmosphere general circulation model (GCM) to simulate climatic conditions at the Eemian. Although there have been a few attempts to model the climate of this epoch [Prell and Kutzbach, 1987; Crowley and Kim, 1994; de Noblet *et al.*, 1996], to our knowledge this has not yet been done with a coupled GCM. In the subsequent sections we describe the climate model and boundary conditions used in this experiment, study the change in surface temperature, and compare the simulated and CLIMAP reconstructed change in SST. We focus ex-

clusively on changes of surface temperature; the examination of other fields is beyond the scope of this paper and is part of a more complete description of the experiment which is in preparation.

2. Model Description

The climate model used in this experiment (hereafter, the Eemian experiment) is the ECHAM-1 T21/LSG coupled GCM, which has been described by Cubasch *et al.* [1992] and used in several climate change experiments [Bakan *et al.*, 1991; Cubasch *et al.*, 1992, 1994]. The atmospheric component (ECHAM-1) is a low resolution version of the numerical weather forecasting model of the European Centre for Medium Range Weather Forecasts which has been modified in Hamburg for climate simulation purposes [Roeckner *et al.*, 1992]. It is a spectral model with a horizontal resolution given by a triangular cutoff at zonal wave number 21 (T21) which is transformed into a Gaussian grid of about 5.625° and a vertical hybrid σ -p coordinate system with 19 levels. The oceanic component is the large scale geostrophic (LSG) model [Maier-Reimer *et al.*, 1993]. It has 11 variably spaced levels in the vertical and two overlapping $5.6^\circ \times 5.6^\circ$ horizontal grids corresponding to an effective grid size of 4° which are interpolated onto the Gaussian grid used in ECHAM-1 T21. The atmospheric and oceanic components are coupled synchronously with a flux correction in order to minimize a climate drift of the coupled system away from the climatologies simulated by the uncoupled models [Sausen *et al.*, 1988]. The validation of model results against an independent data set from a different climate state can be used as one test of the validity of the flux correction procedure, which has been criticized when it is employed in greenhouse runs.

3. Boundary Condition

In the Eemian experiment the ECHAM-1 T21/LSG climate model was integrated for 510 years starting at year 600 of a control experiment, which is our definition of the present. Boundary conditions of the Eemian experiment differ from those of the control run in two ways (Table 1) (the proposed 6 m sea level rise for the Eemian results in negligible changes in the model land-

sea grid). First, the Earth's orbital parameters have been set to their values at 125 ka [Berger, 1978]. The greater obliquity and eccentricity and the fact that perihelion occurred in northern hemisphere summer (rather than winter, as today) caused an amplification (reduction) of the seasonal cycle of insolation in the northern (southern) hemisphere. Second, the CO_2 concentration in the Eemian experiment has been set to 267 ppmv. This value is slightly lower than the mean CO_2 levels for the Eemian as estimated from ice core measurements (~ 270 -275 ppmv) [Barnola *et al.*, 1987], but the difference in terms of radiative forcing is small enough to consider the effect negligible. Our run should be considered as indicative of mid-Eemian conditions: although the standard oxygen isotope chronology [Imbrie *et al.*, 1894] for the last interglacial places 125 ka at the very end of the major deglaciation preceding the interglacial, high-precision uranium series dates from coral reef terraces [Edwards *et al.*, 1987; Chen *et al.*, 1991; Gallup *et al.*, 1994] and deep-sea sediments [Slowey *et al.*, 1996] suggest that the age of the last interglacial is $\sim 120 - 130$ ka.

The precession of the equinoxes and the change in eccentricity lead to the dilemma that equal dates at 125 ka and present do not correspond to the same value of the celestial longitude Φ (angle measured from the vernal equinox). Therefore it may make no sense to compare monthly or seasonal mean fields of both experiments [Kutzbach and Gallimore, 1988; Jossaume and Braconnot, 1997]. Since our interests focus on summer and winter mean fields, we have considered the orbit of the Earth [Monin, 1986] to redefine northern hemisphere summer and winter at 125 ka as those periods which correspond to the same interval in Φ as the present June-July-August and December-January-February, namely, as periods between days 150 and 231 and days 320 and 59, respectively, in our 360 day model. The shorter summer and longer winter at 125 ka are due to the fact that at this epoch, perihelion occurred in northern summer rather than in northern winter as today. Our results show that in the case of surface or near-surface temperatures the difference between strict seasonal means and our redefined summer and winter means is small compared to the actual difference fields between the Eemian and the present.

Table 1. Boundary Conditions

	CO_2 , ppmv	Eccentricity	Obliquity, deg.	Angle of Perihelion, deg.
Control run	330	0.017	23.45	282.16
125 ka	267	0.040	23.79	127.27

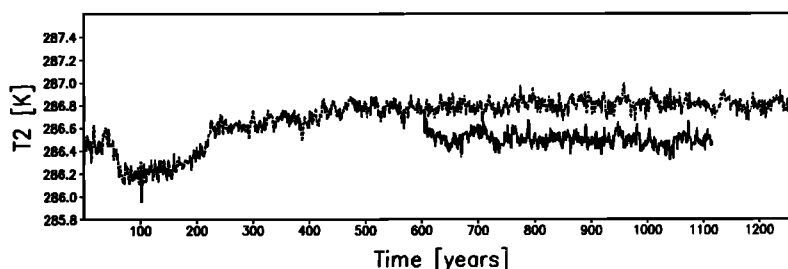


Figure 1. Time series of the annual mean globally averaged near-surface (2 m) temperature for the control (dashed line) and Eemian run (solid line), in degrees Celsius.

4. Results

4.1. Description of Model Run

The temporal evolution, the time-mean states, and the variations of the quasi-stationary atmosphere-ocean system in the control run have been analyzed in previous works [Cubasch *et al.*, 1992; von Storch, 1994; von Storch *et al.*, 1997]. Both the control and Eemian run are quasi-stationary with respect to surface and near-surface temperature approximately from years 450 and 750, respectively (Figure 1). The strong drift shown by the near-surface temperature in the control run in the first years subsequent to the coupling is caused by the drift of sea ice, which, in turn, is related to the simplified sea ice model and the applied flux correction [Cubasch *et al.*, 1994; von Storch *et al.*, 1997] and is also manifested in a drift of the globally averaged temperature of the upper layers of the ocean (not shown). This drift ends after ~ 450 years, when the atmosphere, sea ice, and upper ocean reached quasi-stationary equilibrium. In the Eemian run the globally averaged temperature of the upper layers of the ocean has not completely reached equilibrium after 510 years, but the trend is below $-0.1^\circ\text{C}/400$ years at all levels down to 1000 m depth. The situation in the deep ocean is different, where a nearly constant trend in globally averaged temperature is found in both the Eemian and the control run. A possible reason for this drift would be that the deep ocean had not reached equilibrium before its coupling to the atmosphere model [von Storch *et al.*, 1997].

We consider differences in simulated mean seasonal and annual surface temperatures (Eemian minus control run) averaged over 300 years at the end of the Eemian run (Plates 1a-1c). These differences reflect, in large part, the expected changes due to the amplification (reduction) of the seasonal cycle of insolation in the northern (southern) hemisphere. In northern summer (Plate 1a), there is an overall increase of the surface temperature in the northern hemisphere over land except for cooling of as much as $2^\circ\text{--}3^\circ\text{C}$ in some regions in central Africa and south Asia that are

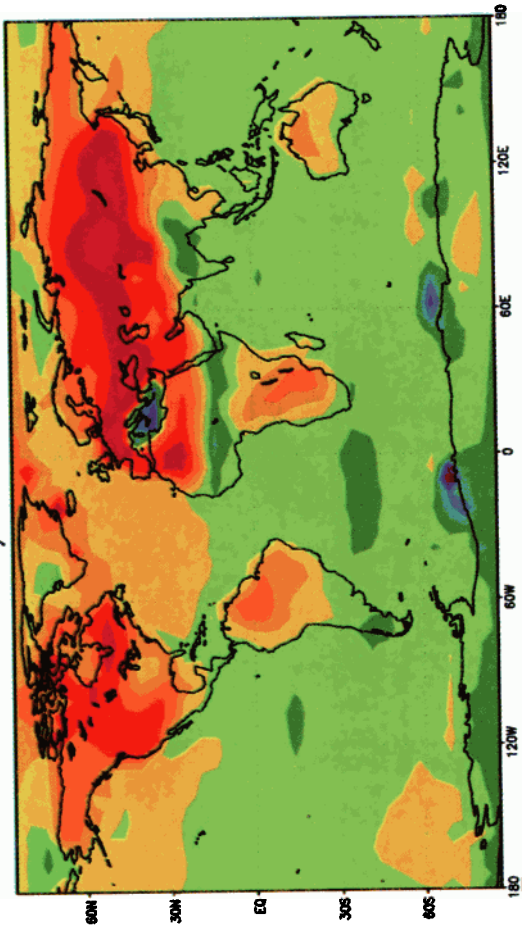
related to enhanced precipitation in these areas (not shown). As Kutzbach *et al.* [1996] have demonstrated that vegetation/soil feedbacks can modify the precipitation response in such regions, the potential for additional cloudiness/temperature changes in the Sahara may further modify the response we obtain.

Maximum surface temperature increases above $4^\circ\text{--}5^\circ\text{C}$ occur in the centers of the continents (North America, North Africa, central Europe, the Middle East, and Siberia). The warming is approximately the same as that found in an atmospheric general circulation model (AGCM) experiment with prescribed SSTs [Prell and Kutzbach, 1987] and an energy balance model calculation [Crowley and Kim, 1994] but slightly less than that obtained by Harrison *et al.* [1995]. By contrast, modeled northern hemisphere summer SSTs show enhancements below 1°C . These different results reflect the significant increase in northern hemisphere summer insolation and the different heat capacities of land and ocean. For the same period in the southern hemisphere, surface temperature increases in the centers of the subtropical continents (South America, South Africa, and Australia), while SSTs show a slight cooling. This response reflects the fact that even at 30°S , northern hemisphere summer insolation during the Eemian exceeds the present by about 21 W/m^2 (Figure 2a), and this difference is again manifested on land because of the different heat capacities of land and ocean.

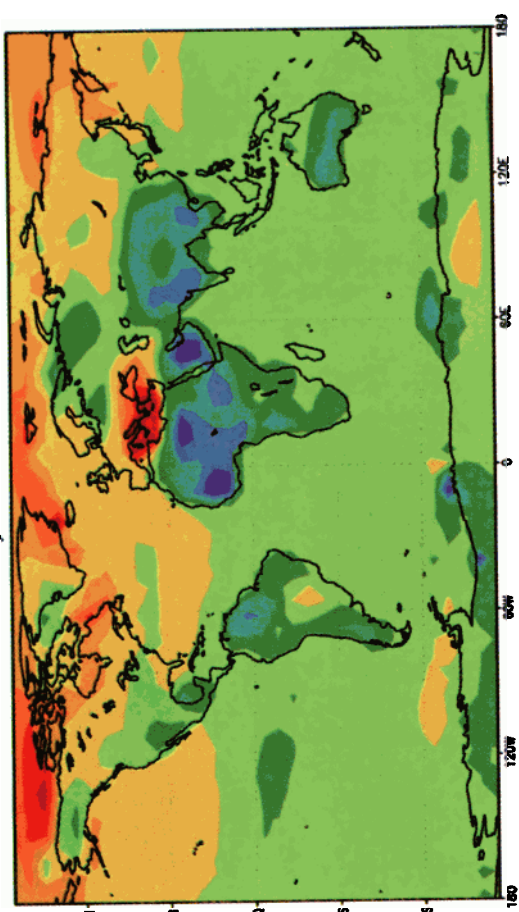
In the northern winter (Plate 1b), warming up to 4°C takes place in high northern latitudes (North America and the Arctic Ocean). This warming reflects a decrease in sea ice thickness and extent due to enhanced summer insolation and is also smaller than that modeled by Harrison *et al.* [1995] (warming up to 8°C in the Arctic). However, their response could be unrealistically large because the lack of a dynamic ocean model leads to too large a sea ice cover in the Norwegian Sea in their control run.

Surface temperatures show a slight cooling in northern winter, which is enhanced south of 30°N . Extreme cooling with differences $> 3^\circ\text{C}$ occurs mainly in south

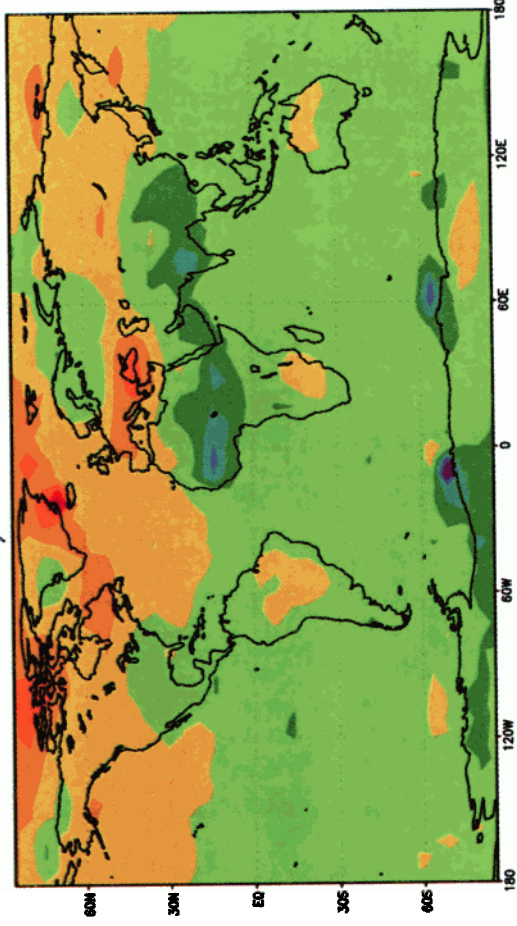
a) NH Summer



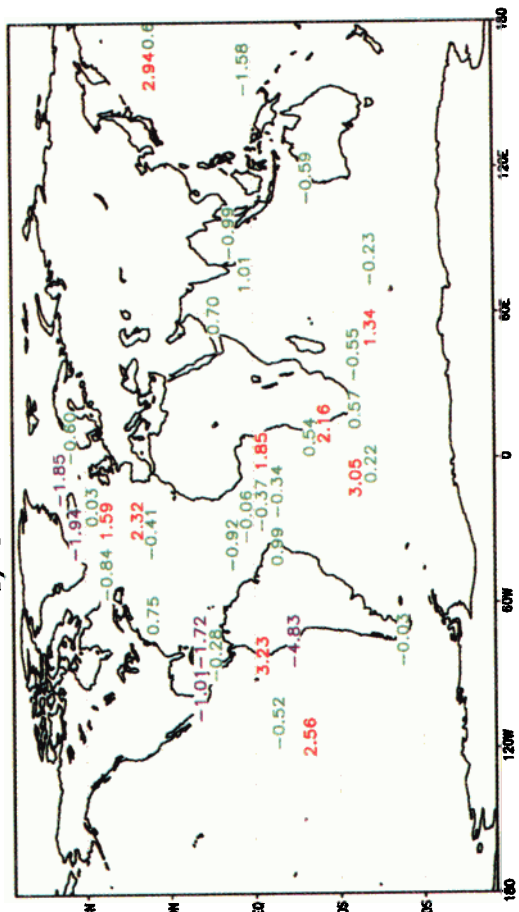
b) NH Winter



c) Annual Mean



d) CLIMAP - Model



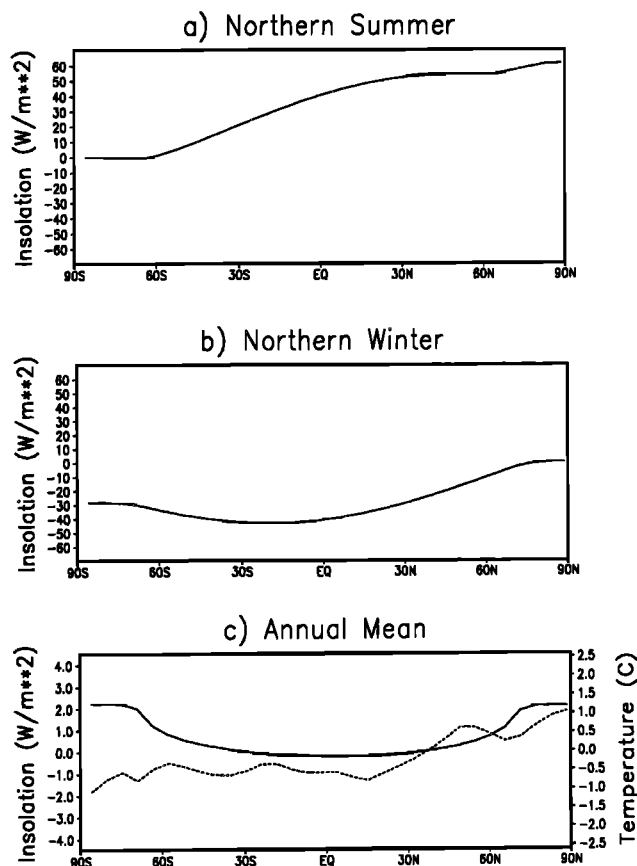


Figure 2. Zonally averaged incoming insolation departures at the Eemian with respect to the present for (a) northern summer, (b) northern winter, and (c) annual mean (in W/m^2). In figure 2c the dashed line shows the zonally averaged mean annual surface temperature differences (Eemian minus control run) in degrees Celsius.

Asia and central and North Africa. Again, this very large cooling reflects the facts that the northern hemisphere summer insolation increases in the Eemian are more or less offset by significant decreases in winter insolation and that the difference in heat capacity between land and sea amplifies the response between land and sea. The effect is most pronounced in low latitudes because of the larger amount of radiation in the tropics. The large-scale pattern on land again reflects the linear energy balance modeling results for the last interglacial (S.-J. Kim and T. J. Crowley, unpublished results, 1997). The Mediterranean Sea shows strong cool-

ing in summer and heating by almost the same amount in winter. This may be due to the fact that it is closed according to the topography of the climate model; for this reason we do not further discuss the results from this region.

Simulated seasonal changes almost cancel out in the annual mean (Plate 1c). Surface temperatures show little change, especially SSTs. Over land, surface temperatures increase significantly for some points at high northern latitudes but decrease in central Africa and south Asia. These results are consistent with land records from Europe which suggest that slightly higher mean temperatures existed in the Eemian [van der Hammen *et al.*, 1971; Woillard, 1978] and also with oceanic evidence suggesting that the ocean at the last interglacial was generally similar to the present, although conditions might have been slightly warmer in parts of the North Atlantic and North Pacific [Ruddiman and McIntyre, 1976; Crowley, 1981; CLIMAP Project Members, 1984].

The zonal mean annual temperature response (Figure 2c) indicates warming north of $\sim 30^\circ\text{N}$ and slight cooling south of that point. The zonal mean response in highest latitudes may reflect the $2 \text{ W}/\text{m}^2$ difference in forcing due to obliquity changes, which are symmetric in both hemispheres. Because sensitivity to mean annual changes is about an order of magnitude greater than sensitivity to seasonal changes [North *et al.*, 1984] and $2 \text{ W}/\text{m}^2$ is comparable to the ice age CO_2 radiative perturbation, the perturbation in highest latitudes could have a significant climate effect. However, the SST response between 70°N and 70°S departs from the near linearity of the response of the seasonal temperature field on land and indicates changes in either ocean heat transport or clouds as likely explanations. In-depth analyses of these fields will be presented in the paper now under preparation that describes the responses of these and other model fields.

Despite the large seasonal changes the simulated mean global average temperature change is very small. The change in mean globally averaged near-surface (2 m) temperature (Eemian run minus control run) is $\Delta T_0 = -0.32^\circ\text{C}$; that is, we obtain a decrease of 0.32°C at the Eemian with respect to the control run. This result is to be expected because of the small difference in mean annual insolation ($0.23 \text{ W}/\text{m}^2$) and the fact that the CO_2 concentration at the Eemian was lower

Plate 1. Mean surface temperature changes (Eemian minus control run) for (a) northern hemisphere summer, (b) northern hemisphere winter, (c) annual mean (300-year averages; from 801 to 1100), and (d) Climate: Long-Range Investigation, Mapping, and Prediction (CLIMAP) Project Members minus model-simulated sea-surface temperature (SST) anomalies (difference between Eemian and present summer and winter average SSTs) at CLIMAP core locations (blue indicates that the CLIMAP anomaly is colder than predicted by the model; red indicates that it is warmer; green indicates that there is no significant difference at $1-\sigma$ level (in degrees Celsius)).

than for the control run. A local t test was applied to individual points and showed that differences in mean surface temperatures (seasonal and annual mean) are statistically significant at the 5% significance level at almost all points (not shown).

4.2. Comparison With Observations

We have tested the climate model's response against the CLIMAP Project Members [1984] SST field. Although the CLIMAP SST fields for the last glacial maximum have been challenged on a number of grounds [e.g., Guilderson *et al.*, 1994], the general similarity of the last interglacial and the present biota suggests that the transfer function technique can be applied with more confidence to the last interglacial. We compared simulated and CLIMAP SST anomalies (Eemian minus present mean SSTs, which are estimated by averaging winter and summer SSTs at each epoch) only at those sites, discretized as in the climate model, where CLIMAP cores are located. Since the informational value of a grid box value is somewhat questionable [von Storch, 1995], making use of simulated SSTs at single grid points is of limited value. Moreover, the size of the area represented by a single grid point is larger than that of a core. These problems are usually dealt with by the use of dynamic and empirical downscaling [Zorita and von Storch, 1997]. However, in our case this is precluded by the lack of observational data and regional climate model capabilities.

Core top estimates (instead of present SST data) were chosen as modern CLIMAP SST estimates. In this way systematic errors in SST estimates cancel out when computing the anomalies. This forced us to select those cores for which such estimates exist (a total of 41). At those cores where a biotic census for different faunal groups exists we took the average of the SST estimates for all of them. Errors for the CLIMAP reconstructed SST anomalies were calculated from the standard errors of seasonal estimates. These represent a lower estimate of the total uncertainty in the SST estimates since they only include transfer function errors which are between 1°–2°C with an average value of 1.5°C.

While the control run is representative of the mid to late 20th century (CO₂ levels of 330 ppmv were reached about 1960), the core tops represent average climatic conditions of the last circa 1500 years: the core tops have been regressed against SST data sets approximately representative of 1960 [CLIMAP Project Members, 1976], that is, a time interval that was probably warmer than the mean value for the last 1500 years. Even though there is a slight bias in this regression, utilization of the same regression for the Eemian maintains the same bias for that time interval, so the CLIMAP reconstructed SST anomalies represent differences with respect to the average conditions of the last 1500 years.

In order to define the same reference interval for the SST simulated anomalies the latter must be adjusted by the calculated CO₂-induced temperature warming between preindustrial and mid 20th century levels ($\sim 0.3^\circ\text{C}$) [Houghton *et al.*, 1996].

On a point-by-point basis the model agrees with the observations within 1 standard deviation for 63% of the cores (Plate 1d) and within 2 standard deviations for 92% of them. Large differences between the climate model and observations primarily occur in regions along eastern boundary currents and subtropical/subpolar frontal zones. These differences could reflect inadequacies in the model both in terms of physics and resolution (e.g., problems in the simulation of sea ice, strong gradients in these regions which are not resolved by the model, etc.), but they may also be an indication of problems with transfer functions in such regions [CLIMAP Project Members, 1984; Ravelo *et al.*, 1990], long timescale (> 2000 years) leads and lags in the climate system that are not captured by an equilibrium simulation [CLIMAP Project Members, 1984; Crowley, 1990], or nonsynchronous stratigraphic picks, bioturbation, dissolution, and a possible biasing of some samples by a brief 290 ppmv CO₂ excursion in the early Eemian [Barnola *et al.*, 1987].

The overall mean anomalies for the model (adjusted), calculated over the CLIMAP sample locations, and for the observations are 0.04°C and 0.19°C, respectively. These values agree within the pooled error of the SST CLIMAP anomaly (0.22°C). Therefore, in mean terms the results of the climate model are indistinguishable from the CLIMAP data, and both of them indicate that within the uncertainty of the sampling, the data, and the model, Eemian values had already been reached by the mid 20th century.

To address the potential bias in global temperatures due to the small size of the samples used in the comparison, we have compared the mean simulated anomaly over grid points containing CLIMAP cores with the globally averaged simulated anomaly. The unadjusted (adjusted) mean simulated temperature anomaly based on the 41 CLIMAP sample locations is -0.26°C (0.04°C), which is very close to the change in mean global SST of -0.30°C (0.00°C) and to the change in winter and summer global means of -0.14°C (0.16°C). This close agreement supports arguments [Shen *et al.*, 1994] that even relatively sparse data sets can sometimes yield accurate estimates of the global mean. In fact the good agreement with the sparse data set was obtained with a sampling density about two-thirds that estimated by Shen *et al.* [1994] and with a non-optimal data distribution. The good correlation occurs because there is a very high correlation (0.91) between the simulated zonally averaged SSTs in the Atlantic and the Pacific, the latter basin of which is significantly less

sampled than the Atlantic. This high correlation reflects the zonal nature of the insolation perturbation.

The above exercise therefore suggests that although some regions may have had warmer temperatures in the Eemian than the present, there are compensating changes elsewhere, with the net effect being only minor changes in global average temperature. The exercise also pinpoints regions that could be the subject of further modeling studies, for example, regions of the midlatitude northern hemisphere where paleo data are slightly warmer than the climate model [Terasmae, 1960; van der Hammen et al., 1971; Woillard, 1978; Miller et al., 1983; de Vernal et al., 1986; MacCracken et al., 1990; Mangerud and Svendsen, 1992; Pons et al., 1992; Guiot et al., 1993]. Note that the stratigraphic position of some of these sites are still open to question.

5. Discussion and Conclusions

To summarize, our results show significant changes in mean seasonal (winter and summer) simulated surface temperatures during the Eemian with respect to the present. Such changes are due to the different orbital parameters which translate into an amplification of the seasonal cycle during the Eemian, modulated in each hemisphere by the asymmetric landmass distribution and the date of perihelion. Seasonal changes are, in large part, compensated when we consider annual mean surface temperature changes, which also reflect the previous asymmetries.

Our results suggest that the control run, which is representative of the mid to late 20th century, is already warmer than the Eemian in terms of globally averaged temperature by 0.32°C. This result should not be surprising since the difference in terms of orbital forcing

is very small and the CO₂ level in the Eemian is lower than the control run and mid 20th century observations. However, fuller comparison of the Eemian with the present will eventually require inclusion of the effects of vegetation changes [Harrison et al., 1995] on global temperatures. Although the very large vegetation changes during the last glacial maximum sometimes resulted in regional temperature changes of several degrees, global temperatures varied < 0.2°C [Crowley and Baum, 1997]. We would not expect future assessments of Eemian vegetation on temperature to differ greatly from this number. We therefore anticipate that our conclusions about the stability of Eemian global temperatures (with respect to the present) will be robust.

Finally, surface temperatures might, in turn, be constrained by the flux correction procedure used. This is an open question which cannot be answered until atmosphere and ocean models are improved to the point that flux correction terms are not needed in their coupling. However, the very fact that the model calculations with flux correction agree with an independent data set is one line of evidence in support of the assumption that the use of the flux correction for modest perturbations from the basic state is valid.

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References

- Bakan, S., et al., Climate response to smoke from the burning oil wells in Kuwait, *Nature*, **351**, 367-371, 1991.
- Barnola, J. M., D. Raynaud, Y. S. Korotkevich, and C. Lorius, Vostok ice core provides 160,000-year record of atmospheric CO₂, *Nature*, **329**, 408-414, 1987.
- Berger, A. L., Long-term variations of daily insolation and Quaternary climatic changes, *J. Atmos. Sci.*, **35**, 2362-2367, 1978.
- Bloom, A. L., J. M. A. Chappel, W. S. Broecker, R. K. Matthews, and K. J. Mesolella, Quaternary sea-level fluctuations on a tectonic coast: New ²³⁰Th/²³⁴U dates from the Huon peninsula, *Quat. Res.*, **4**, 185-205, 1974.
- Budyko, M., and Y. A. Izrael (Eds.), *Anthropogenic Climate Changes*, Arizona Univ. Press, Tucson, 1990.
- Chen, J. H., H. A. Curran, B. White, and G. J. Wasserburg, Precise chronology of the last interglacial period: ²³⁴U - ²³⁰Th data from fossil coral reefs in the Bahamas, *Geol. Soc. Am. Bull.*, **103**, 92-97, 1991.
- Climate: Long-Range Investigation, Mapping, and Prediction (CLIMAP) Project Members, The surface of the ice-age Earth, *Science*, **191**, 1131-1137, 1976.
- CLIMAP Project Members, The last interglacial ocean, *Quat. Res.*, **21**, 123-224, 1984.
- Crowley, T. J., Temperature and circulation changes in the eastern North Atlantic during the last 150,000 years: Evidence from the planktonic foraminiferal record, *Mar. Micropaleontol.*, **6**, 97-129, 1981.
- Crowley, T. J., Are there any satisfactory geologic analogs for a future greenhouse warming? *J. Clim.*, **3**, 1282-1292, 1990.
- Crowley, T. J., and S. K. Baum, Effect of vegetation on an ice age climate model simulation, *J. Geophys. Res.*, **102**, 16,463-16,480, 1997.
- Crowley, T. J., and K.-Y. Kim, Milankovitch forcing of the last interglacial sea level, *Science*, **265**, 1566-1568, 1994.
- Cubasch, U., K. Hasselmann, H. Höck, E. Maier-Reimer, U. Mikolajewicz, B.D. Santer, and R. Sausen, Time-dependent greenhouse warming computations with a coupled ocean-atmosphere model, *Clim. Dyn.*, **8**, 55-69, 1992.
- Cubasch, U., B. D. Santer, A. Hellbach, G. C. Hegerl, H. Höck, E. Maier-Reimer, U. Mikolajewicz, A. Stössel, and R. Voss, Monte Carlo climate change forecasts with a global coupled ocean-atmosphere model, *Clim. Dyn.*, **10**, 1-19, 1994.
- de Noblet, N., P. Braconnot, S. Jossaume,

- and V. Masson, Sensitivity of simulated Asian and African summer monsoons to orbitally induced variations in insolation 126, 115 and 6 kBP, *Clim. Dyn.*, **12**, 589-603, 1996.
- deVernal, A., C. Causse, C. Hillaire-Marcel, R. J. Mott, and S. Occhietti, Palynostratigraphy and Th/U ages of upper Pleistocene interglacial and interstadial deposits on Cape Breton Island, eastern Canada, *Geology*, **14**, 554-557, 1986.
- Dodge, R. E., R.G. Fairbanks, L.K. Benninger, and F. Maurrasse, Pleistocene sea levels from raised coral reefs of Haiti, *Science*, **219**, 1423-1425, 1983.
- Edwards, R. L., J. H. Chen, T.-L. Ku, and G. J. Wasserburg, Precise timing of the last interglacial period from mass spectrometric determination of Thorium-230 in corals, *Science*, **236**, 1547-1553, 1987.
- Gallup, C. D., R. L. Edwards, and R. G. Johnson, The timing of high sea levels over the past 200,000 years, *Science*, **263**, 796-800, 1994.
- Guilderson, T. P., R. G. Fairbanks, and J. L. Rubenstone, Tropical temperature variations since 20,000 years ago: Modulating interhemispheric climate change, *Science*, **263**, 663-665, 1994.
- Guiot, J., J. L. de Beaulieu, R. Cheddadi, F. David, P. Ponel, and M. Reille, The climate in western Europe during the last glacial/interglacial cycle derived from pollen and insect remains, *Paleogeogr., Palaeoclimatol., Palaeoecol.*, **103**, 73, 1993.
- Hansen, J. E., I. Fung, A. Lacis, D. Rind, S. Lebedeff, R. Ruedy, and G. Russell, Global climate changes as forecast by Goddard Institute for Space Studies three-dimensional model, *J. Geophys. Res.*, **93**, 9341-9364, 1988.
- Harrison, S. P., J. E., Kutzbach, C. E. Prentice, P. J. Behling, and M. T. Sykes, The response of northern hemisphere extratropical climate and vegetation to orbitally induced changes in insolation during the last interglaciation, *Quat. Res.*, **43**, 174-184, 1995.
- Houghton, J. T., L. G. Meira Filho, B. A. Callander, N. Harris, A. Katterberg, and K. Maskell (Eds.), *Climate Change 1995: The Science of Climate Change*, pp. 9-50, Cambridge Univ. Press, New York, 1996.
- Imbrie, J., J. D. Hays, D. G. Martinson, A. McIntyre, A. C. Mix, J. J. Morley, N. G. Pisias, and N. J. Shackleton, The orbital theory of Pleistocene climate: Support from a revised chronology of the marine $\delta^{18}\text{O}$ record, in *Milankovitch and Climate*, edited by A. Berger, et al., pp. 269-305, D. Reidel, Norwell, Mass., 1984.
- Jossaume, S., and P. Braconnot, Sensitivity of paleoclimate simulations results to season definitions, *J. Geophys. Res.*, **102**, 1943-1956, 1997.
- Kellogg, W., Effects of human activities on global climate, *Rep. 486*, World Meteorol. Org., Geneva, Switzerland, 1977.
- Ku, T.-L., M. A. Kimmel, W. H. Easton, and T. J. O'Neil, Eustatic sea level 120,000 years ago on Oahu, Hawaii, *Science*, **183**, 959-962, 1974.
- Kutzbach, J., and R. Gallimore, Sensitivity of a coupled atmosphere/mixed layer ocean model to changes in orbital forcing at 9000 BP, *J. Geophys. Res.*, **93**, 801-821, 1988.
- Kutzbach, J., G. Bonan, J. Foley, and S. P. Harrison, Vegetation and soil feedbacks on the response of the African monsoon to orbital forcing in the early to middle Holocene, *Nature*, **384**, 623-626, 1996.
- MacCracken, M. C., M. I. Budyko, A. D. Hecht, and Y. A. Izrael, *Prospects for Future Climate*, Lewis, Chelsea, Michigan, 1990.
- Maier-Reimer, E., U. Mikolajewicz, and K. Hasselmann, Mean circulation of the Hamburg LSG OGCM and its sensitivity to the thermohaline surface forcing, *J. Phys. Oceanogr.*, **23**, 731-757, 1993.
- Mangerud, J., and J. I. Svendsen, The last interglacial-glacial period on Spitsbergen, Svalbard, *Quat. Sci. Rev.*, **11**, 633-664, 1992.
- Mesolella, K. J., R. K. Matthews, W. S. Broecker, and D. L. Thurber, The astronomical theory of climatic change: Barbados data, *J. Geol.*, **77**, 250-274, 1969.
- Miller, G. H., H. P. Sejrup, J. Mangerud, and B. G. Andersen, Amino acid ratios in Quaternary molluscs and foraminifera from western Norway: Correlation, geochronology and paleotemperature estimates, *Boreas*, **12**, 107-124, 1983.
- Monin, A. S., *Introduction to the Theory of Climate*, pp. 10-29, D. Reidel, Mass., 1986.
- North, G. R., J. G. Mengel, and D. A. Short, On the transient response patterns of climate to time dependent concentrations of atmospheric CO_2 , in *Climate Processes and Climate Sensitivity*, *Geophys. Monogr. Ser.*, vol. **29**, edited by J. E. Hansen and T. Takahashi, pp. 164-170, AGU, Washington, D. C., 1984.
- Pons, A., J. Guiot, J.L. de Beaulieu, and M. Reille, Recent contributions to the climatology of the last glacial-interglacial cycle based on French pollen sequences, *Quat. Sci. Rev.*, **11**, 439-448, 1992.
- Prell, W. L., and J. E. Kutzbach, Monsoon variability over the last 150,000 years, *J. Geophys. Res.*, **92**, 8411-8425, 1987.
- Ravelo, A. C., R. G. Fairbanks, and S. G. H. Philander, Reconstructing tropical Atlantic hydrography using planktonic foraminifera and an ocean model, *Paleoceanography*, **5**, 409-431, 1990.
- Roeckner, E., et al., Simulation of the present day climate with the ECHAM model: Impact of model physics and resolution, *Rep. 93*, Max-Planck-Inst. für Meteorol., Hamburg, Germany, 1992.
- Ruddiman, W. F., and A. McIntyre, Northeast Atlantic paleoclimatic changes over the past 600,000 years, in *Investigation of Late Quaternary Paleoclimatology and Paleoclimatology*, edited by R. M. Cline et al., Mem. Geol. Soc. Am., **145**, 111-146, 1976.
- Sausen, R., K. Barthel, and K. Hasselmann, Coupled ocean-atmosphere models with flux corrections, *Clim. Dyn.*, **2**, 154-163, 1988.
- Shen, S. S., G. R. North, and K.-Y. Kim, Spectral approach to optimal estimation of the global average temperature, *J. Clim.*, **7**, 1999-2007, 1994.
- Slowey, N. C., G. M. Henderson, and W. B. Curry, Direct U-Th dating of marine sediments from the two most recent interglacial periods, *Nature*, **383**, 242-244, 1996.
- Terasmae, J., A palynological study of Pleistocene interglacial beds at Toronto, Ontario, *Geol. Surv. Can. Bull.*, **56**, 23-41, 1960.
- van der Hammen, T., T. A. Wijmstra, and W. H. Zagwijn, The floral record of the Late Cenozoic of Europe, in *The Late Cenozoic Glacial Ages*, edited by K. K. Turekian, pp. 391-424, Yale Univ. Press, New Haven, Conn., 1971.
- von Storch, H., Inconsistencies at the interface of climate impact studies and global climate research, *Meteorol. Z.*, **4 NF**, 72-80, 1995.
- von Storch, J.-S., Interdecadal variability in a coupled model, *Tellus, Ser. A*, **46**, 419-432, 1994.
- von Storch, J.-S., V. Kharin, U. Cubasch, G. C. Hegerl, D. Schriever, H. von Storch, and E. Zorita, A 1260-year control integration with the coupled ECHAM1/LSG general circulation model, *J. Clim.*, in press, 1997.
- Woillard, G. M., Grande Pile peat bog: A continuous pollen record for the last 140,000 years, *Quat. Res.*, **9**, 1-21, 1978.
- Zorita, E., and H. von Storch, A survey of statistical downscaling techniques, *Rep. 97/E/20*, GKSS-Forschungszentrum, Geesthacht, Germany, 1997.
- Zubakov, V. A., and I. I. Borzenkova, *Global Palaeoclimate of the Late Cenozoic*, 453 pp., Elsevier, New York, 1990.

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