Milankovitch Forcing of the Last Interglacial Sea Level

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During the last interglacial, sea level was as high as present, 4000 to 6000 years before peak Northern Hemisphere insolation receipt 126,000 years ago. The sea-level results are shown to be consistent with climate models, which simulate a 3° to 4°C July temperature increase from 140,000 to 130,000 years ago in high latitudes, with all Northern Hemisphere land areas being warmer than present by 130,000 years ago. The early warming occurs because obliquity peaked earlier than precession and because precession values were greater than present before peak precessional forcing occurred. These results indicate that a fuller understanding of the Milankovitch-climate connection requires consideration of fields other than just insolation forcing at 65°N.

The theory of orbital forcing of Pleistocene climates (*1*) has received a substantial level of empirical support in the last two decades (*2*). However, two recent pieces of evidence indicate complications in understanding the orbital-climate linkage. Coral records indicate that sea level was as high as present by 130,000 years ago (*3*) and perhaps as early as 132,000 years ago (*4*). These values precede by 4000 to 6000 years the peak Northern Hemisphere insolation receipt at 65°N. This latter index is the standard measure for evaluating the orbital-climate connection. Additionally, a record from Devils Hole, Nevada, suggests significant deglaciation as early as 140,000 years ago (*5*). Although the very early warming at Devils Hole may reflect some regional changes in climate (*6*), the global sea-level record from coral reefs indicates that important elements are missing in our understanding of the orbital-climate link.

In this report, we use a climate model to examine the system response to Milankovitch forcing at 2000-year intervals covering the critical time period of the last interglacial. Our objective is to determine whether the geographic response to total insolation changes provides the necessary insight to explain differences between the insolation record at 65°N and the sea-level record. The model (*7*) is a two-dimensional energy balance model (EBM) that resolves the temperature response to seasonal insolation forcing as it is modified by geography. Although it is a simple model in the hierarchy of climate models, repeated comparisons of different versions of the EBM with atmospheric and oceanic general circulation models indicate a temperature sensitivity comparable to that of the larger models (*8*). To consider a minimum of adjustable parameters, we use the linear version of the EBM so that the only response we obtain is due to orbital forcing and the modifying effects of geography. For the same reason, we do not consider additional changes that can result from ice-age CO_2 changes. Thus, the responses we illustrate are minimum responses due to Milankovitch forcing. Orbital insolation values determined by Berger and Loutre (*9*) were used for our calculations.

We examined the July temperature fields over the interval bracketing the penultimate deglaciation (Fig. 1) because Northern Hemisphere summer insolation is considered critical for glacial inception and disintegration (*1*). Although peak July warming in Eurasia of 5° to 6°C occurs at the standard time of 126,000 years ago, warming in highest latitudes exceeds the present by 1.0° to 1.5°C as early as 134,000 years ago. Mid-latitude landmasses were cooler than present at this time in the centers of North America and Eurasia. The different responses by region and latitude occurred because obliquity peaked significantly earlier than precession at the time of the last interglacial (Fig. 2) and because the obliquity effect is largest in high latitudes (*10*). The low- and mid-latitude response is dominated by precession, which lags obliquity over this interval and is manifested most strongly on large landmasses (*10*).

Despite the lag between precession and obliquity, precession values were greater than at present as early as 134,000 years ago and equal to the early Holocene peak by 131,000 years ago (Fig. 2). The joint effect of precession and obliquity forcing on temperature is that by 130,000 years ago, simulated warming was more uniformly spread over Northern Hemisphere landmasses and high latitudes. Zonally averaged (Fig. 3) temperatures over critical Northern Hemisphere latitudes with ice cover (50° to 70°N) were about 2.5°C warmer than present by 130,000 years ago.

After peak warming 126,000 years ago, simulated temperatures started to wane in high latitudes. By 120,000 years ago, July temperatures were cooler than present in the highest latitudes. For the patterns at 122,000 and 120,000 years ago, the highest latitudes were cooler than they are now and the mid-latitude land areas were warmer than at present. Again, this pattern reflects the different contributions of obliquity and precession, as modified by the land-sea distribution. The cooling in high latitudes 120,000 years ago is consistent with the coral record of lower sea level after this time (*3*). The results also support evidence that

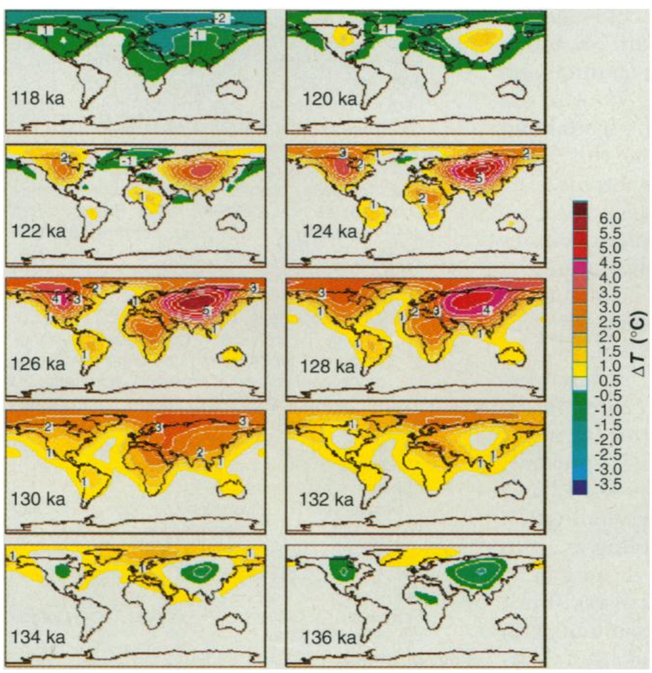

Fig. 1. Calculated temperature differences from present for July for 136,000 to 118,000 years ago (Ka). Note that color coding is for 0.5°C increments except between −0.5° and 0.5°C (white coding).

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the first phase of ice growth occurred in the high Arctic (*11*).

To obtain further insight into factors responsible for the sea-level rise, we examined the simulated temperature fields over the interval of 110,000 to 140,000 years ago at three key locations. Two of the sites are from high latitudes, with one from Arctic Canada and the other from the Barents Sea. For the last glacial maximum, ice sheets may have been grounded on the extensive continental shelves of the Barents and Kara seas (*12*). The third location is from midlatitude North America, near the southern boundary of the Laurentide Ice Sheet. Results (Fig. 4) indicate a total trough-to-peak warming of 3° to 4°C, with warming for the Barents Sea site peaking 130,000 years ago and leading the other two sites by 4000 years. Although it is at the same latitude as the Barents Sea site, peak warming at the Canadian Arctic site occurred 4000 years later. This response reflects the greater importance of the precession-controlled continentality effect on the Canadian site. However, all sites were warmer than they are now by 130,000 years ago. Although we have no information on the regional patterns of deglaciation for the penultimate interglacial, several lines of evidence (*13*) for the most recent deglaciation indicate initial melting in the highest latitudes. This evidence is in accord with some of our results.

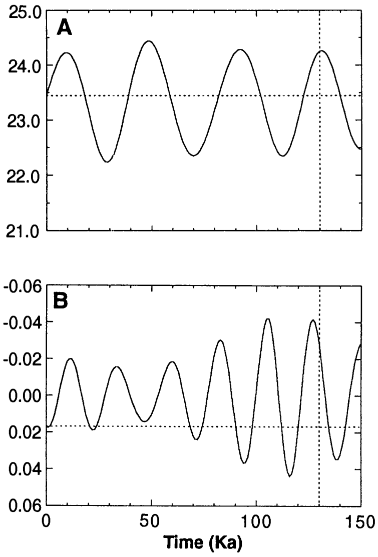

Fig. 2. (**A**) Obliquity and (**B**) precession indices for the last 150,000 years, as computed by Berger and Loutre (*9*). Vertical dashed lines are coincident with 130,000 years ago; horizontal dashed lines are present orbital values.

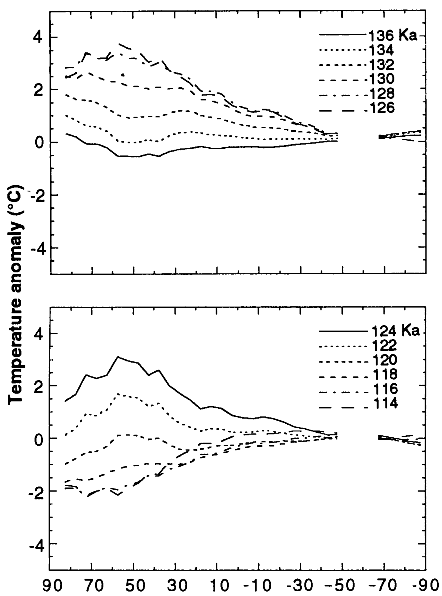

Fig. 3. Simulated zonal average temperature differences (land) from present for 12 intervals bracketing the last interglacial.

We suggest that the magnitude of simulated temperature changes may have been sufficient to trigger deglaciation by about 134,000 years ago. Increasing CO_2 levels (*14*) would augment such melting trends. We further suggest that, because simulated temperatures for all Northern Hemisphere land areas were warmer than at present by 130,000 years ago, the results are consistent with near-total melting and high sea level by that time. It might be argued that the simulated temperature changes would be muted by a higher albedo of snow and ice in high latitudes. However, ice core records (*15*) indicate that atmospheric dust levels were substantial during the penultimate and last glacial maxima, and dust significantly reduces the albedo of snow (*16*). Eventually, a climate model must be coupled with an ice sheet model to determine whether the responses we simulate are sufficient to trigger deglaciation of a dust-covered ice sheet.

The coral reef sea-level records (*3*, *4*) are at variance with the record of Imbrie and colleagues (*2*) of the $\delta^{18}O$ of global ice volume, which does not indicate sea levels as high as those at present until 124,000 years ago. Some of the differences may reflect stipulation of the midpoint of the penultimate deglaciation 128,000 years ago (*2*). If this number were now shifted to about 134,000 years ago, much of the discrepancy would disappear. But the duration of the event would still be shorter in the deep-sea record than in the coral reef record. It is possible that bioturbation and compositing of records from different deep-sea cores may have resulted in an apparent shortening of the length of the last interglacial with respect to the coral reef record. Finally, the time scale of Imbrie and co-workers (*2*) is based on specifying a constant phase offset between orbital forcing and $\delta^{18}O$. Because the amplitude of precession varies significantly over time (Fig. 2), it may be necessary to retune the $\delta^{18}O$ record as a function of both amplitude and phase of orbital forcing. Although these ideas require testing, they indicate that no fundamental problems necessarily exist in the interpretation of the deep-sea $\delta^{18}O$ record.

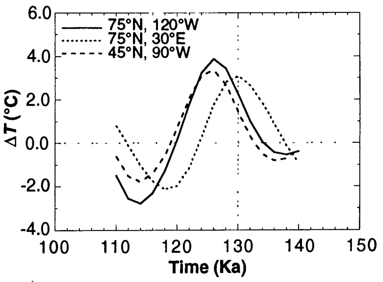

Fig. 4. Time series of model-simulated July temperature for three sites representative of high-latitude and low-latitude ice sheets. The vertical dashed line is coincident with 130,000 years ago; the horizontal dashed line is the present simulated temperature. *T*, temperature.

Despite remaining questions, our results indicate that early sea-level peaks 130,000 to 132,000 years ago may be entirely consistent with Milankovitch forcing. The results also indicate that the warming signatures from obliquity and precession can sometimes be out of phase even in the same hemisphere. These results indicate that, although Milankovitch forcing may still be the dominant factor in driving Pleistocene climate, a fuller understanding of the orbital-climate connection requires examination of fields other than just insolation values at 65°N.

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Carbon Dioxide Supersaturation in the Surface Waters of Lakes

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Data on the partial pressure of carbon dioxide (CO_2) in the surface waters from a large number of lakes (1835) with a worldwide distribution show that only a small proportion of the 4665 samples analyzed (less than 10 percent) were within ±20 percent of equilibrium with the atmosphere and that most samples (87 percent) were supersaturated. The mean partial pressure of CO_2 averaged 1036 microatmospheres, about three times the value in the overlying atmosphere, indicating that lakes are sources rather than sinks of atmospheric CO_2. On a global scale, the potential efflux of CO_2 from lakes (about 0.14 × 10^{15} grams of carbon per year) is about half as large as riverine transport of organic plus inorganic carbon to the ocean. Lakes are a small but potentially important conduit for carbon from terrestrial sources to the atmospheric sink.

Processes that add and remove CO_2 occur simultaneously in the surface waters of lakes. Lakes can thus act either as sources or as sinks for CO_2. Earlier studies have shown that Arctic lakes are strongly supersaturated in CO_2 and therefore are sources to the overlying atmosphere (*1*, *2*). In the Arctic the transport of tundra organic matter to surface waters leads to CO_2 supersaturation (*1*, *2*). Other regions that lack the vast soil carbon storage of the Arctic may behave differently. In fact, detailed studies on a limited number of temperate and boreal lakes have suggested that these lakes are net sinks for atmospheric CO_2 (*3*), but no comprehensive analysis exists. We report data from lakes with a worldwide distribution that show that boreal, temperate, and tropical lakes are typically supersaturated with CO_2 and thus are net sources to the atmosphere.

Data were obtained both from the literature and from our own direct measurements of the partial pressure of CO_2 (P_{CO_2}). The value of P_{CO_2} was calculated from pH and dissolved inorganic carbon (DIC) or acid-neutralizing capacity (ANC) with corrections for other physical and chemical variables (*4*). For the direct measurement of P_{CO_2}, we collected water from 0.1 to 0.25 m below the surface into a thermally insulated 2-liter glass bottle and equilibrated it with 50 ml of ambient air (*5*). Gas chromatography was used to measure CO_2 on the equilibrated head space. Simultaneously ambient air 1 m above the lake surface was collected for the measurement of atmospheric CO_2.

The lakes analyzed range in size from 8.2 × 10^6 ha (Lake Superior) to 0.4 ha, span latitudes from 60°S to 62°N, and include both hard and soft waters (Table 1). For each data set there are differences in the type of measurements made, the intensity of those measurements over time, and geographic location (Table 1). For these reasons we discuss the data in terms of a series of individual data sets.

Of the 37 lakes (390 samples) where direct measurements of P_{CO_2} were made, 16 lakes (43%) were supersaturated at all samplings. Data for a persistently supersaturated lake are shown in Fig. 1. The mean measured P_{CO_2} for the 37 lakes was 801 ± 67 μatm (mean ± standard error), about 2.2 times the measured atmospheric value (Fig. 2A). Only 7% of the samples were within ±20% of atmospheric equilibrium (Fig. 2A).

In the larger data sets for which P_{CO_2} was calculated, we also see a consistent tendency toward supersaturation. For the 1612 lakes sampled in autumn [eastern lakes survey (*6*) (Fig. 2B)], mean P_{CO_2} was 1031 ± 19.4 μatm or three times the atmospheric equilibrium value. Relatively few lakes (6.6%) were within ±20% of the atmospheric equilibrium value.

Because of thermal mixing, autumn may be a time of elevated P_{CO_2} in surface waters. Nevertheless, a similar pattern of supersaturation was seen for 69 lakes from all over the world with full seasonal data for the entire ice-free season (Fig. 2C). For lakes with ANC-based data, calculated mean P_{CO_2} averaged 1122 μatm; for lakes with DIC-based data, calculated mean P_{CO_2} was 1039 μatm. Both of these values are well above atmospheric equilibrium (*7*) and similar to that from the lakes sampled only in autumn (Fig. 2B). For 34 of these 69 lakes, 100% of the samples were supersaturated, and for every lake the time-weighted average was above the atmospheric value.

The 60 lakes (179 samples) sampled only during summer stratification, a time of lower-than-average P_{CO_2}, were also supersaturated. The distribution is broader and the mean lower (680 ± 65 μatm) than for the other data sets (Fig. 2D). For these samples also, only 12.8% were within ±20% of the atmospheric equilibrium value.

Tropical African lakes were strongly supersaturated with the mean P_{CO_2} being about six times (2296 ± 409 μatm) the atmospheric value; few samples (11%) were within ±20% of atmospheric equilibrium (Fig. 2E).

For the lakes for which we both directly measured and calculated P_{CO_2}, we found relatively good agreement between measured and calculated values (*8*), but such agreement

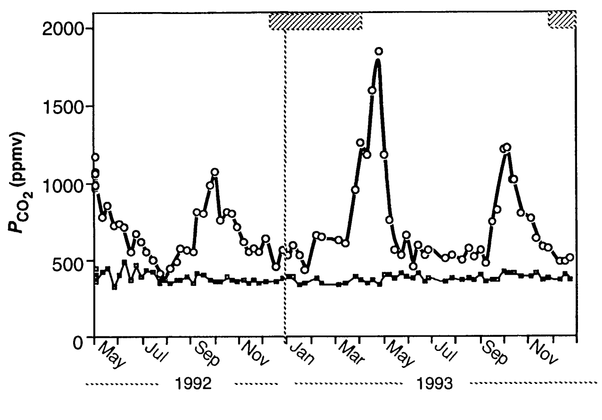

Fig. 1. Seasonal cycle of direct measurements of the P_{CO_2} in the surface water of Mirror Lake (circles) and in the overlying air (squares), showing persistent supersaturation. Mirror Lake is a soft water lake in New Hampshire (*15*); ppmv, parts per million by volume. The hatched areas represent ice cover.

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