Supporting Online Material for

2.8 Million Years of Arctic Climate Change from Lake El’gygytgyn, NE Russia


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Materials and Methods

Drilling operation

Scientific deep drilling in El’gygytgyn crater, in remote northeastern Russia (cf. Fig. 1), was performed in winter 2008/09 (9), following a comprehensive geophysical site survey and shallow coring (11, 57-59). One borehole was drilled into permafrost-rich alluvial fan deposits in the western lake catchment (ICDP Site 5011-3) and three holes were drilled in the center of the lake (ICDP Site 5011-1, A, B, and C) (Table S1). Because casing was set into the upper portion of the lake sediment sequence, a stratigraphic gap exists between the top of the lake drill cores and the sediment-water interface, which is successfully filled by a percussion piston core (Lz1024) taken during site survey work in 2003.

Table S1. Penetration, drilling and core recovery at ICDP Sites 5011-1 and 5011-3 in the El’gygytgyn Crater (all data given in field depth; mbfl = meters below lake floor).

<table>
<thead>
<tr>
<th>Site</th>
<th>Hole</th>
<th>Type of Material</th>
<th>Penetrated (mbfl)</th>
<th>Drilled (m)</th>
<th>Recovered (m)</th>
<th>Recovery (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>5011-1</td>
<td>1A</td>
<td>lake sediment</td>
<td>146.6</td>
<td>143.7</td>
<td>132.0</td>
<td>92</td>
</tr>
<tr>
<td></td>
<td>1B</td>
<td>lake sediment</td>
<td>111.9</td>
<td>108.4</td>
<td>106.6</td>
<td>98</td>
</tr>
<tr>
<td></td>
<td>1C</td>
<td>total</td>
<td>517.3</td>
<td>431.5</td>
<td>273.8</td>
<td>63</td>
</tr>
<tr>
<td></td>
<td></td>
<td>lake sediment</td>
<td></td>
<td></td>
<td>225.3</td>
<td>52</td>
</tr>
<tr>
<td></td>
<td></td>
<td>impact rocks</td>
<td></td>
<td></td>
<td>207.5</td>
<td>76</td>
</tr>
<tr>
<td>5011-3</td>
<td></td>
<td>permafrost deposits</td>
<td>141.5</td>
<td>141.5</td>
<td>129.9</td>
<td>91</td>
</tr>
</tbody>
</table>

Creation of the composite profile

The composite profile from ICDP Site 5011-1 (given in corrected depths below lake floor) was constructed from the best-preserved sediment intervals in overlapping core sequences based upon visual core inspection supported by variations in the proxy data. Layers originating from volcanic ashes or mass movement events in Lake El’gygytgyn were generally omitted if their thicknesses exceeded 5 cm, producing small gaps within the composite proxy data sets. These layers usually have sharp lower and upper boundaries and can unequivocally be distinguished from the pelagic sediments; moreover, results from earlier pilot cores suggest that most mass movement events in the central part of the lake are not associated with erosion of previously deposited sediments (59). The uppermost 5.66 m of the composite depth were taken from pilot core Lz1024. Below 5.66 m, to a depth of 104.8 m, alternating sediment intervals from parallel cores 5011-1A and -1B were spliced together. Between 104.8 m and 145.7 m, cores 1A, 1B and 1C contributed to the composite. Below 145.7 m down to the sediment impact breccia interface at 318 m composite depth, the record relies solely on core 5011-1C.

Laboratory analyses

Magnetic (volume) susceptibility (MS) (10^-6 SI) was measured in 1-mm steps on core halves using an automated split-core logger. Data from core Lz1024 were acquired with a Bartington MS2E spot-reading sensor in combination with a MS2 control unit. MS from ICDP
5011-1 cores were obtained using a Bartington MS2E sensor first attached to a MS2 control unit, which was later replaced by a technically improved MS3 control unit. The response function of the MS2E sensor with respect to a thin magnetic layer is equivalent to a Gaussian curve with a half-width of slightly less than 4 mm. The amplitude resolution of the sensor is $10 \times 10^6$ in combination with the MS2 unit and $2 \times 10^6$ with the MS3 unit. During data acquisition, blank readings against air were obtained after every 10 measurements in order to correct the data for subtle shifts in the sensor's background due to temperature drift.

Acquisition of magnetostratigraphic data (inclination and declination) was mainly performed on sediments sampled in U-channels (2×2 cm in cross-section) at 2 cm intervals. Below 145.7 m bfl, the sediments were too stiff for extracting U-channels so sampling was accomplished with discrete samples every 10 cm, using 2×2×1.5 cm plastic boxes. From the lowermost core sections, irregular but stiff pieces of core were directly placed into the magnetometer's sample holder. In all cases, magnetic polarity was determined from the results of stepwise alternating field demagnetization of the natural remanent magnetization up to 100 mT and subsequently applied principle component analysis (60).

The manganese/iron (Mn/Fe) and silicon/titanium (Si/Ti) ratios were determined on core halves and U-channels using an X-ray fluorescence (XRF) core scanner (ITRAX, Cox Ltd., Sweden), equipped with a Mo-tube, which was set to 30 kV and 30 mA. XRF scanning was performed at 2-mm resolution using an integration time of 10 s per measurement. Since the intensities derived from wet sediments, especially those of light elements such as Si, might be influenced by effects of the sediment matrix (61), matrix-corrected Si counts ($Si_{mc}$) were calculated from the raw Si integrals ($Si_{raw}$) using the empirically determined formula

$$Si_{mc} = \frac{Si_{raw}}{3.2994 \times e^{-0.505^{inc/coh}}}.$$  

The formula is based on an exponential attenuation function between the ratio of wet and dry element intensities and the ITRAX-derived ratio of Compton and Rayleigh scattering (inc/coh ratio) of 340 samples from the Lake El’gygytgyn record. The inc/coh ratio in general is dependent on the average atomic number of the sample and thus has been used as an indicator for organic matter content and/or matrix-induced density variations (62).

For biogeochemical analyses, one of the core halves was continuously sub-sampled at 2-cm spacing. The samples were freeze-dried and ground to < 63 µm, then biogenic silica (BSi) was quantified every 2 cm using Fourier Transform Infrared Spectroscopy (FTIRS) by Bruker (ltd.) IFS 66/v and Vertex 70 spectrometers. Analyses, sample preparation, and calibration of the resulting infrared spectral data followed the methods described in (63), with the exception that only samples from Lake El’gygytgyn were included in the FTIRS-BSi calibration model. The content of total organic carbon (TOC) in the 5011-1 core composite was determined in steps of 2 cm based upon the difference between total carbon and total inorganic carbon measured with a DimaTOC carbon analyzer (Dimatec Corp.) in aqueous suspension. TOC of percussion piston core Lz1024 was measured with a METALYT CS 1000S analyzer (ELTRA Corp.), following sediment pretreatment with 10% HCl in order to remove calcium carbonate.

Palynological investigations of the ICDP 5011-1 core composite are currently restricted to selected intervals. These intervals include core Lz1024 down to 16.6 m (corresponding to the past 340 ka), composite intervals 16.91-19.33 m and 43.55-44.40 m (encompassing MIS 11c and
MIS 31), and the sediments below 101.90 m composite depth (deposited between 2.15 and 3.6 Ma ago). Average sample resolutions throughout these intervals to date are at 6, 5, and 60 cm, respectively. Pollen sample processing followed standard techniques used for organic-poor sediments (64). Water-free glycerol was used in sample storage and preparation of the microscopic slides. Pollen and spores as well as a number of non-pollen palynomorphs were identified and counted at magnifications of 400x and 1000x, with the aid of published pollen keys and atlases. Tablets containing Lycopodium marker spores were added to the samples to allow for calculation of pollen concentrations (65). If available, at least 250 pollen grains were counted in each sample. The relative frequencies of pollen taxa were calculated from the sum of the pollen taxa. Spore percentages are based on the sum of pollen and spores. The percentages of non-pollen palynomorphs are based on the sum of the pollen and non-pollen palynomorphs, and the percentages of algae are based on the sum of pollen and algae. The Tilia/TiliaGraph/TGView software (66) was used for the calculation of percentages and for drawing initial diagrams.

Development of the age model

The development of the age model for the composite core ICDP 5011-1 followed a 3-step approach (see Fig. S1 for example; note that for reasons of clarity not all resulting correlation tie points of that interval are displayed).

First order tie points are provided by the magnetostratigraphic results, which show 16 polarity reversals for chron, subchrons, and cryptochrons (e.g., red dashed line in Fig. S1). Fourteen of these reversals are well defined in the El’gygytgyn lake sediment record (Table S2). Only the top of the Kaena and the base of the Mammoth, both within the Gauss Chron, are somewhat ambiguous, when only paleomagnetic information is considered. The ages of nine reversals were assigned according to (12), who provide ages of the major reversals of the Earth's magnetic dipole field based on their time frame for the LR04 stack. The ages of the remaining seven reversals were derived from the 2nd and 3rd order tie points (see below). This includes the boundaries of the short-term Cobb Mountain event (67), which is clearly linked to MIS 35 (68). Our age range for the Réunion subchron is in good agreement with the placing of this event into MIS 80 and 81 (69) and radiometric dating by (70) and (71). We also give an age range for a precursor prior to the Olduvai subchron, which previously was detected in sediments in the North Atlantic (69). According to our study, the Matuyama/Gauss reversal occurred at 2.588 Ma, in the middle of MIS 103. The younger age of 2.581 Ma, given by (72), places the Matuyama/Gauss reversal in the late MIS 103, whereas the older age of 2.608 Ma, given by (12), places this major reversal into MIS 104. Our position within MIS 103 is supported by respective findings in the sediment record of Lake Baikal (73). For the base of the lacustrine sediment section the age of the impact at 3.58±0.04 Ma was adopted from (7) as another 1st order tie point, and we assume that limnic sedimentation started sometime shortly after the impact event.

The paleomagnetic age/depth record from Lake El’gygytgyn then allowed for 2nd and 3rd order comparisons of climate events using additional proxies. The Si/Ti ratio most resembles the LR04 isotope stack (12), while both MS and TOC rather resemble the Northern Hemisphere (67.5°N, El’gygytgyn latitude) cumulative spring/summer insolation (May to August) according to (13). Fluctuations in the Si/Ti ratio, therefore, were correlated with LR04, providing 2nd order
Fig. S1. Chronostratigraphic plot for the time window from 650 ka to 900 ka, illustrating the definition of 1st, 2nd, and 3rd order tie points used for creating the composite age model for ICDP Site 5011-1. (A) Lithological facies occurring in the time interval, (B) Si/Ti ratio derived from X-ray fluorescence (XRF) scanning, (C) cumulative May, June, July, August insolation for 67.5°N (13), (D) LR04 oxygen isotope stack ($\delta^{18}$O) (12), (E) magnetic susceptibility (MS), (F) total organic carbon (TOC), and (G) bars indicating normal (black) and reversed (white) polarity. Geomagnetic field reversals are defined as 1st order tie points, correlation of the Si/Ti ratio to the LR04 stack as 2nd order tie points, and correlation of MS and TOC to insolation patterns as 3rd order tie points.

tie points (dark green dashed lines in Fig. S1). These correlations are inherently associated with at least the uncertainties in the LR04 age model, which are estimated to be 6 ka for the time interval 3 to 1 Ma and 4 ka since 1 Ma (12). Fluctuations of MS and TOC were correlated with the regional insolation, providing 3rd order tie points (blue dashed lines in Fig. S1). The 2nd and 3rd order tie points take advantage of the observed environmental changes and coincident Northern Hemisphere summer insolation, which shows a stronger variability than the LR04 stack. For instance, the onsets of interglacials (e.g., MIS 17, 19, and 21 in Fig. S1) are linked with pronounced insolation maxima. During these interglacials one (e.g., MIS 17 and 19) or even two (e.g., MIS 21) insolation minima occur. These minima and all other minima in insolation are obviously linked to the onsets of lows in MS and highs in TOC when applying 3rd order tie points.
for correlation. The onset of maxima in MS and minima in TOC, on the other hand, coincide well with the steepest gradients of increasing insolation after the preceding insolation minima. Exceptions to this common correlation scheme only occur for TOC in the most pronounced interglacials (Facies C), when particularly high primary production, reflected in distinct peaks in Si/Ti, leads to distinctly enhanced TOC content despite oxic bottom waters and decomposition throughout most of the year.

**Table S2.** Comparison of magnetostratigraphic boundaries for geomagnetic chron, subchrons, and cryptochrons according to (12, 70, 105, 106) with those taken in this study. We adopted all ages from (12) except the ones highlighted by a star, which were retrieved from the age model of ICDP 5011-1 (o = onset, t = termination, the Gauss/Gilbert boundary is not documented in the lake sediment record).

<table>
<thead>
<tr>
<th>Boundary</th>
<th>Ma (105)</th>
<th>Ma (106)</th>
<th>Ma (70)</th>
<th>Ma (12)</th>
<th>Ma (this study)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Brunhes/Matuyama</td>
<td>0.780</td>
<td>0.781</td>
<td>0.780</td>
<td>0.780</td>
<td></td>
</tr>
<tr>
<td>Jaramillo (t)</td>
<td>0.990</td>
<td>0.988</td>
<td>0.991</td>
<td>0.991</td>
<td></td>
</tr>
<tr>
<td>Jaramillo (o)</td>
<td>1.070</td>
<td>1.072</td>
<td>1.075</td>
<td>1.075</td>
<td></td>
</tr>
<tr>
<td>Cobb-Mountain (t)</td>
<td>1.201</td>
<td>1.173</td>
<td></td>
<td></td>
<td>1.1858*</td>
</tr>
<tr>
<td>Cobb-Mountain (o)</td>
<td>1.211</td>
<td>1.185</td>
<td></td>
<td></td>
<td>1.1938*</td>
</tr>
<tr>
<td>Olduvai (t)</td>
<td>1.770</td>
<td>1.785</td>
<td>1.778</td>
<td>1.781</td>
<td>1.781</td>
</tr>
<tr>
<td>Olduvai (o)</td>
<td>1.950</td>
<td>1.942</td>
<td>1.945</td>
<td>1.968</td>
<td>1.968</td>
</tr>
<tr>
<td>Olduvai precursor (t)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>1.9782*</td>
</tr>
<tr>
<td>Olduvai precursor (o)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>1.9815*</td>
</tr>
<tr>
<td>Réunion (t)</td>
<td>2.140</td>
<td>2.129</td>
<td>2.128</td>
<td>2.126</td>
<td>2.126*</td>
</tr>
<tr>
<td>Réunion (o)</td>
<td>2.150</td>
<td>2.149</td>
<td>2.15</td>
<td>2.1384*</td>
<td></td>
</tr>
<tr>
<td>Matuyama/Gauss</td>
<td>2.581</td>
<td>2.582</td>
<td>2.581</td>
<td>2.608</td>
<td>2.588*</td>
</tr>
<tr>
<td>Kaena (t)</td>
<td>3.040</td>
<td>3.032</td>
<td>3.032</td>
<td>3.045</td>
<td>3.045</td>
</tr>
<tr>
<td>Kaena (o)</td>
<td>3.110</td>
<td>3.116</td>
<td>3.116</td>
<td>3.127</td>
<td>3.127</td>
</tr>
<tr>
<td>Mammoth (t)</td>
<td>3.220</td>
<td>3.207</td>
<td>3.207</td>
<td>3.210</td>
<td>3.210</td>
</tr>
<tr>
<td>Gauss/Gilbert</td>
<td>3.580</td>
<td>3.596</td>
<td>3.596</td>
<td>3.588</td>
<td>3.588</td>
</tr>
</tbody>
</table>

**BMA climate reconstruction**

The best modern analogue (BMA), also known as modern analogue approach (28, 74-76), assumes that pollen assemblages with a similar composition of taxa are produced by compositionally and structurally similar vegetation communities, which grow under similar climatic conditions. This assumption makes it possible to identify the closest modern analogues for each analyzed fossil sample by comparison with modern pollen samples included in the reference dataset. The modern climate parameters affiliated with the sites of modern pollen samples serve as the closest analogues and these values are then assigned to the analyzed fossil samples and considered as reconstructed values of the past climate. To identify modern analogues we used the squared-chord metric of dissimilarity (SCD) following the methods of (74).

In order to calculate the modern climate at the reference pollen sampling sites, we used the free-access software packages Polation and Polygon (77, 78). The inverse distance weighting interpolation within 1° interpolation radius was applied to a global high-resolution dataset of surface climate averaged over the 1961-1990 interval (56), in order to calculate modern climate at the surface pollen sampling sites included in the reference dataset (76).
The BMA approach was tested comparing reconstructed values for the surface-pollen dataset to the interpolated climate measurements. In this testing mode a modern sample is not allowed to serve as the best analogue (the 'leave-one-out' method); therefore the reconstructions are distinct from and can be assessed against the observed values. Two spectra were judged analogous if their SCDs were less than a threshold value, “T”. Experimenting with the threshold T and the number of closest modern analogues, (77) found that these parameters only slightly influenced the generally good correlation between estimated and reconstructed climatic variables. After rechecking this conclusion with the extended dataset, we came to the same result keeping T=0.4 maximum and N=8 minimum. Analogue selection was not geographically restricted.

Over the last two decades, various quantitative approaches have been developed and used to evaluate climate reconstruction results derived from pollen records (28). Table S3 shows performance statistics for the BMA approach used in our study, calculated using C2 version 1.7.2 software package (79).

**Table S3.** Performance statistics of the BMA climate reconstruction method applied in this study, giving the coefficient of determination (R²), the average and maximum bias of the reconstruction, and the root mean square error of prediction (RMSEP) for annual precipitation (PANN), mean temperature of the warmest month (MTWM), mean temperature of the coldest month (MTCO), and mean annual temperature (TANN).

<table>
<thead>
<tr>
<th>Variable</th>
<th>R²</th>
<th>Average bias</th>
<th>Maximum bias</th>
<th>RMSEP</th>
</tr>
</thead>
<tbody>
<tr>
<td>PANN</td>
<td>0.72</td>
<td>-5.89</td>
<td>106.44</td>
<td>92.54</td>
</tr>
<tr>
<td>MTWM</td>
<td>0.75</td>
<td>-0.008</td>
<td>10.02</td>
<td>1.63</td>
</tr>
<tr>
<td>MTCO</td>
<td>0.77</td>
<td>-0.45</td>
<td>7.56</td>
<td>5.02</td>
</tr>
<tr>
<td>TANN</td>
<td>0.81</td>
<td>-0.23</td>
<td>5.59</td>
<td>2.84</td>
</tr>
</tbody>
</table>

Another way to evaluate possible reconstruction errors for the reconstructed climate variables is by calculating the uncertainty range (75, 77). When applied to the pollen-based reconstruction results, this method usually produces much larger errors. The error ranges defined by the climatic variability in the set of analogues are frequently asymmetric, because they take into account the selected analogue values not normally distributed around the most probable value (76). The uncertainty ranges representing minimum and maximum values in the range of best modern analogues are provided for modern and fossil climate reconstructions. Table S4 shows results of the pollen-based BMA climate reconstruction for the modern pollen spectra from Lake El’gygytgyn. The pollen-derived climatic ranges for the four climatic variables are reasonably close to the estimated variables derived from the modern climate averages around Lake El’gygytgyn (56).

**Table S4.** Comparison between the pollen-based BMA climate reconstruction (expressed as the range of the closest analogues) for the modern pollen spectra from Lake El’gygytgyn and those estimated from modern climate data (56). For acronyms see Table S3.

<table>
<thead>
<tr>
<th></th>
<th>PANN (mm/yr)</th>
<th>MTWM (°C)</th>
<th>MTCO (°C)</th>
<th>TANN (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Reconstructed max</td>
<td>280.6</td>
<td>9.9</td>
<td>-31.0</td>
<td>-13.4</td>
</tr>
<tr>
<td>Reconstructed min</td>
<td>253.3</td>
<td>8.8</td>
<td>-35.5</td>
<td>-14.4</td>
</tr>
<tr>
<td>Estimated from climate data</td>
<td>235</td>
<td>8.8 to 9</td>
<td>-31 to -32</td>
<td>-13.5 to -14</td>
</tr>
</tbody>
</table>
Climate-vegetation modeling

Climate simulations of interglacials use the current (2011) version of the GENESIS v. 3.0 GCM (80) interactively coupled to the BIOME4 equilibrium vegetation model (81). The model has been tested extensively in present day and paleoclimate scenarios and simulates modern Arctic climate and distributions of potential vegetation close to observations (44). Simulated mean temperature of the warmest month (MTWM) at Lake El’gygytgyn for preindustrial and modern conditions (10.3 and 12°C, respectively) compare favorably with 20th century reanalysis products (14), but are slightly warmer than limited ground observations, which range from ~8.3-10°C (8, 56, 104). As used here, the atmospheric component has 18 vertical layers, a spectral resolution of T31 (~3.75°), and uses an adapted version of the NCAR CCM3 solar and thermal infrared radiation code (82). The model atmosphere is coupled to 2°×2° surface models including a 50-meter slab ocean model with prognostic sea surface temperatures, diffusive heat transport, and dynamic-thermodynamic sea ice. Terrestrial land surface components include multi-layer snow and soil models, and a land-surface-transfer scheme (LSX) that calculates momentum transfer and fluxes of energy and water between the atmosphere and ice, snow, soil surfaces, and upper and lower vegetation canopies. This version of the GCM has a sensitivity to 2×CO₂ of 2.9 °C, without vegetation, GHG, or ice sheet feedbacks.

Potential equilibrium vegetation distributions are predicted by BIOME4, which is interactively coupled to the GCM. The model predicts the distribution, community structure, and biogeochemistry of 27 biomes using monthly climatologies of temperature, precipitation, and clouds simulated by the GCM. In turn, the simulated vegetation provides the physical land-surface attributes in LSX. The vegetation model was run without bias corrections in both the modern control and paleoclimate simulations. Interglacial climate-biome simulations were run for 50 years to allow complete equilibration. Climatological means were calculated from the last 10 years of each simulation.

Model boundary conditions and interglacial forcing scenarios

Model boundary conditions and primary inputs are summarized in Table S5. Land surface elevations and ice sheets are the same as modern in MIS 1 and MIS 5e simulations. MIS 11c and MIS 31 simulations use both modern and modified land surface boundary conditions with the Greenland Ice Sheet removed. In those simulations, all Greenland ice is removed and the sub-glacial land surface is isostatically restored to its ice-free elevation. Prescribed CO₂, CH₄, and N₂O atmospheric mixing ratios in MIS 1, 5e, and 11c are based on ice core observations (40, 83-86). Atmospheric CO₂ in the MIS 31 simulation is based on a geochemical proxy reconstruction (87). The combined radiative forcing potential (in W m⁻²) of prescribed CO₂, CH₄, and N₂O values used in each simulation were calculated using standard expressions (88, 89).

Prescribed orbital parameters (Table S5) assume the same timing of Northern Hemispheric interglacial forcing used by (40). For MIS 1, 5e, and 11c, values of orbital eccentricity, obliquity, and the longitude of perihelion are from (86). For MIS 31, orbital parameters correspond to the timing of peak July insolation at the latitude of Lake El’gygytgyn (1072 ka) (90). Orbital and greenhouse gas values are fixed during each simulation, representing a quasi-equilibrated snap-
shot of each timeslice and the maximum potential response to the imposed forcing while ignoring the transient nature of forcing which is different for each interglacial.

### Table S5. Overview of interglacial simulations performed for this study, with orbital and GHG forcing according to (40, 83-87, 90). For modern CO$_2$ the 1950 AD concentration is taken; obliquity (tilt) is given in degrees and precession is $\Omega$.

<table>
<thead>
<tr>
<th>Run Name</th>
<th>CO$_2$ (ppm)</th>
<th>CH$_4$ (ppbv)</th>
<th>N$_2$O (ppbv)</th>
<th>Eccentricity</th>
<th>Obliquity</th>
<th>Precession</th>
<th>Temp. ($^\circ$C)</th>
<th>Prec. (mm/yr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>pre-industrial</td>
<td>280</td>
<td>801</td>
<td>289</td>
<td>0.016706</td>
<td>23.438</td>
<td>102.94</td>
<td>10.3</td>
<td>438</td>
</tr>
<tr>
<td>modern</td>
<td>355</td>
<td>1748</td>
<td>311</td>
<td>0.016706</td>
<td>23.438</td>
<td>102.94</td>
<td>12.0</td>
<td>475</td>
</tr>
<tr>
<td>MIS 1-with GIS</td>
<td>~260 ~611</td>
<td>~263</td>
<td>0.01928</td>
<td>24.29</td>
<td>311.26</td>
<td>12.4</td>
<td>438</td>
<td></td>
</tr>
<tr>
<td>MIS 5e-with GIS</td>
<td>287 724</td>
<td>262</td>
<td>0.039378</td>
<td>24.04</td>
<td>275.42</td>
<td>14.5</td>
<td>401</td>
<td></td>
</tr>
<tr>
<td>MIS 11c-with GIS</td>
<td>285 713</td>
<td>285</td>
<td>0.019322</td>
<td>23.781</td>
<td>276.67</td>
<td>12.2</td>
<td>475</td>
<td></td>
</tr>
<tr>
<td>MIS 31-with GIS</td>
<td>325 800</td>
<td>288</td>
<td>0.05597</td>
<td>23.898</td>
<td>289.79</td>
<td>13.8</td>
<td>438</td>
<td></td>
</tr>
<tr>
<td>MIS11c-no GIS</td>
<td>285 713</td>
<td>284</td>
<td>0.019322</td>
<td>23.781</td>
<td>276.67</td>
<td>12.5</td>
<td>438</td>
<td></td>
</tr>
<tr>
<td>MIS11c-no GIS-10Wm$^2$</td>
<td>285 713</td>
<td>284</td>
<td>0.019322</td>
<td>23.781</td>
<td>276.67</td>
<td>13.2</td>
<td>475</td>
<td></td>
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</tbody>
</table>

To test the potential effect of a warmer Arctic Ocean with reduced sea ice on the Beringian interior, the slab ocean model component was modified by adding an additional 8 W m$^{-2}$ of ocean heat convergence under sea ice, in addition to the 2 W m$^{-2}$ used in modern control simulations (10 W m$^{-2}$ total). This mimics the potential impact of a substantial enhancement of oceanic heat flux to the Arctic basin at times of high sea level (MIS 11c). The increased heat flux is based on a simple calculation assuming an extreme 3 Sv increase in Bering Strait throughflow and a 4 $^\circ$C temperature contrast between North Pacific and North Polar surface water. This should be considered a simplistic sensitivity test that should be constrained by future ocean modeling studies.

### SOM Text

**Additional information on facies interpretation**

Sedimentation in Lake El’gygytgyn is highly variable. Distinct lithofacies of the pelagic sediment record were defined based on the physical characteristics of the sediments, including color, particle size, and the presence or absence of various sedimentary structures as visually observed in the split core halves and high-resolution radiographs (100 $\mu$m resolution) obtained using an ITRAX core scanner (Fig. S2). Fine-scale details of characteristic type-sections were further investigated using thin-sections prepared from epoxy-impregnated sediment slabs. High-resolution digital images and backscattered Scanning Electron Microscope (SEM) images were used to evaluate the thin sections.

**Facies A** is defined by the presence of fine clastic laminations (<5mm in thickness; average is ~0.2 mm). Sediments of Facies A are predominantly dark gray to black in color. Laminations have distinct lower bounding surfaces and grade upwards from silt to clay before repeating. Median particle size ranges from approximately 4 to 6 $\mu$m with clay-sized particles comprising less than 30 % of the sediment. The alternating pattern of silt and clay-sized particles results in
clearly observed density variations in radiographs. A variation of Facies A consists of fine wavy laminations with an abundance of elongate sedimentary “pellets” composed of silt-sized and coarser particles. Individual pellets are typically several millimeters thick and have long axes up to ~1cm lying parallel to the bedding plane.

This facies is also present in a 13 m long pilot core (PG1351) that represents the last 250 kyrs of depositional history in Lake El’gygytgyn. Based on presence/absence of pellets, and additional geochemical data, two different subfacies distinguished by (11), both interpreted as reflecting pronounced glacial/stadials and perennial lake-ice cover. These settings explain most of the characteristics of the combined Facies A, as they exclude density-driven or wind-induced mixing of the water column. As a consequence, physical disturbance of bottom sediments is limited and oxygen depletion of the lower water column, which is indicated by the dark sediment color and lack of bioturbation, is enhanced. The laminae characteristics argue for fluvial sediment input, which is possible under perennial ice cover, as only small moats around the margins of the lake are needed to connect inflowing streams with the lake basin. Grading from silt to clay are likely pulses in meltwater production in the arctic watershed, with fluvial input leading to silt deposition, followed by clay when the inflow diminishes and the remaining suspension load settles (91). The pellets occurring in parts of Facies A were interpreted by (11) as cryoconites, which are formed by agglomeration of aeolian material that is transported through the ice along vertical conduits (92). Another possibility is formation by lake-ice rafting of shoreline material entrained in the base of lake ice (93).
Facies B is the most abundant pelagic facies found in the Quaternary sediments of Lake El’gygytgyn, including modern sedimentation (11). While Facies B is variable in nature and ranges in color from olive gray to brown, with frequent greenish bands usually 1-3 cm in thickness, its defining characteristic is a lack of distinct sedimentary structures. Facies B consists of massive to faintly banded silt with common presence of diagenetic vivianite crystals. Transitions between this unit and the other pelagic sediments are typically gradational in nature. Median particle size ranges from 3 to 7 µm with approximately 25 to 35 % of the sediment composed of clay-sized particles.

The variable color of Facies B suggests alternating sedimentary sources or depositional processes as well as variable post-depositional alteration of the sediments. The latter includes redox layers rich in iron and manganese oxides, which are formed today 20 to 30 cm below the sediment surface, and whose fossil remnants form the characteristic greenish bands (59). The overall massive nature of the sediments indicates mixing or minor disturbance of sediment through bioturbation (11). Along with the olive gray to brownish color this implies oxygenated bottom water, due to complete mixing of the water column during ice-free summers, when shelf waters may warm to 4°C, thus leading to wind and density-driven mixing (8). Oxygen in the bottom water is only partly consumed by organic matter decomposition. Being most abundant in the record, Facies B is interpreted as representing the predominant, albeit widely varying climate and environmental conditions at Lake El’gygytgyn.

Facies C does not occur in the pilot core investigated by (11). It is the most irregularly distributed and least common pelagic facies found in the core composite from ICDP Site 5011-1. Facies C is defined by its distinctly reddish-brown color and the presence of faint pale, millimeter to centimeter-scale laminations. Median particle size ranges from ~2 to 10 µm with a clay content between ~20 and 45 %. The banding in Facies C is highly distinct from the clastic laminations characteristic of Facies A and in contrast shows only subtle density variations in radiographs. SEM images indicate that the 1-2 mm thick pale bands delineating the laminations in Facies C reflect a transition from poorly sorted silty sediment to a nearly homogenous band of finer particles with only a few scattered silt-sized clasts. The basal contact of the pale bands is typically sharp, although not planar when viewed at the grain scale. The pale bands gradually transition into the background style of sediments, which persists until the next band occurs.

The reddish-brown color that is diagnostic of Facies C suggests oxidation of bottom sediments by a well-ventilated water column. As for Facies B, this can be traced back to high summer temperatures and a seasonal ice cover. The origin of the banding in Facies C, reflecting the lack of bioturbation, as yet is not fully understood. One possibility is variable fluvial sediment supply and the absence of endobenthic fauna due to seasonal oxygen depletion of the bottom water. This could take place in wintertime, when degradation of organic consumes oxygen in a stratified water column beneath the lake ice.

Additional information on proxy interpretation

MS values in lake sediments reflect the concentration of magnetic mineral phases. In sediments of Lake El’gygytgyn, MS fluctuations are mainly controlled by the preservation rather than by the input of magnetite (94). MS minima during glacial periods are primarily due to
magnetite dissolution in anoxic bottom and pore waters beneath perennial lake ice. Better preservation takes place during interglacial and interstadial periods, when lake-ice cover decreases and the lake becomes fully ventilated. Due to its obvious dependence on temperature and the duration of lake-ice cover, fluctuations in MS, like those of TOC (see below), are predominantly controlled by changes in regional insolation.

Fe and Mn are sensitive to redox changes in aquatic environments (95), with oxic forms of both elements being mostly soluble, whereas Mn and Fe oxyhydroxides are insoluble. Nevertheless, both elements have different $E_h$ stability fields, and Fe(II) is more rapidly oxidized (96). Thus, the Mn/Fe ratio can be used as a tracer for syn- and post-depositional redox conditions in lacustrine sediments (e.g., 97), with low Mn/Fe ratios implying more reduced condition.

Ti is a relatively immobile element that occurs in a variety of mineral phases, widely independent on grain size. Therefore, it is commonly used as a proxy for clastic sediment supply, e.g., of fluvial (98) or eolian (99) origin. The amount of Si is also related to clastic input, but to a large extent is derived from biogenic silica (BSi), which varies in amounts from less than 5% to up to 56.1% in the Quaternary sediment record of Lake El’gygytgyn (Fig. S3). BSi in lakes is formed by siliceous microfossils, generally diatom frustules, and hence is often used as a proxy for primary production (e.g., 100). In Lake El’gygytgyn, diatoms function as the main primary producers and are relatively well preserved in the sediment record (101, 102). As in other lakes (e.g., 103), the Si/Ti ratio in Lake El’gygytgyn strongly correlates with the BSi content ($r^2 = 0.88$; Fig. S3), suggesting the record of Si/Ti ratio is primarily modulated by varying BSi content reflecting aquatic primary production. The amount of aquatic primary production is not only controlled by the duration of ice cover, which is strongly dependent on insolation, but also controlled by weathering in the catchment and nutrient supply, both of which vary with long-term temperature changes. Therefore, fluctuations in the Si/Ti ratio correlate with the regional insolation and with the global MIS isotope record.

**Fig. S3.** Regression plots illustrating that the Si/Ti ratio in the core composite ICDP 5011-1 predominantly reflects biogenic silica (BSi). (A) Conventionally measured BSi content versus BSi determined by FTIRS on 470 validation samples, (B) BSi content from FTIRS versus Si/Ti ratio determined by XRF scanning of 635 validation samples. Dashed red lines indicate 2s (95%) confidence intervals.
The concentration of TOC in the sediments also depends on lacustrine primary production, but is additionally controlled by organic matter supply from the catchment and decomposition in the lake \((/1/\)). TOC maxima during glacial periods in El’gygytgyn sediments are primarily due to restricted decomposition in anoxic bottom waters. Conversely, enhanced decomposition takes place during interglacial periods with oxygenated bottom waters. Thus, high amounts of TOC in Facies C, especially during super interglacials, are caused by enhanced organic matter fluxes that distinctly exceed decomposition.

Palynological data from Lake El’gygytgyn sediments reflect the response of the vegetation on regional climate change, leading to significant differences between the interglacials MIS 1 and 5e and the super interglacials MIS 11c and 31 (Fig. S4) and a distinct change at the Pliocene/Pleistocene boundary.

MIS 1 and 5e evolved from typical open tundra-steppe vegetation during MIS 2 and 6 \((/104/\)) (Fig. S4A and B). Climate amelioration is reflected by gradual increases in shrub pollen, at MIS 1 accompanied by \textit{Sphagnum} spores, which are dominant from \(~13.5\) and \(~12.5\) ka BP, respectively. The increase in birch and especially alder pollen coincident with the decrease in
herb pollen ~12.5-13 ka BP correlates well with the Allerød warming. Low contents of shrub pollen in the sediments dated ~12.5-11 ka BP suggest that climatic conditions became colder and dryer during the Younger Dryas event. At the onset of the Holocene forest-tundra communities became common in the area.

MIS 11c and MIS 31 (Fig. S4C and D) at Lake El’gygytgyn evolved from milder climates than MIS 1 and 5e. This is reflected in the pollen assemblages that accumulated ~428-425.5 ka BP and ~1088-1080 ka BP, which suggest forest-tundra and northern larch taiga environments with alder and birch shrubs. Increases in spruce pollen contents around ~424 and ~1078 ka BP point to significant climate amelioration, respectively. Both super interglacials are characterized by dark coniferous forest with a dominance of spruce, pine (probably originating from tree pine species), birch, alder, and larch. The onset of MIS 31 and the termination of both interglacials are associated with high amounts of *Sphagnum*, suggesting that boggy environments were common during these transitions. Gradual decreases in tree and shrub pollen contents between ~400 and ~390 ka BP (MIS 11c) and since ~1062.5 (MIS 31) reflect that forest-tundra and northern taiga environments again became common in the study area.

The Pliocene/Pleistocene boundary at 2.588 Ma (Table S2) coincides well with the first occurrence of Facies A (2.602 - 2.598 Ma). In addition, a decrease in pine and the disappearance of spruce pollen coincident with an increase in the pollen of shrubs and herbs reflects gradual climate deterioration after ~2.63. At ~2.598 Ma, this shift sharply culminates in environmental conditions unprecedented in the Pliocene. The conditions are similar to those of MIS 2, 4, 6, and 8 (Fig. S4) (104), marked by an absence of pine and spruce in the regional vegetation, but differ because of the dominance of open larch forests with shrubby alder and birch.

Additional information on modeling results

Distributions of modern terrestrial ecosystems simulated by the climate-biome model compare favorably with observations of modern vegetation, natural potential vegetation maps, and pollen data. According to kappa statistics, this is particularly true in the mid-high latitudes relevant to Lake El’gygytgyn (44). In the model, the boundary between boreal forest and tundra is a function of calculated net primary productivity, which is strongly correlated with growing degree-days above 0ºC. In our preindustrial control simulation (Fig. S5A), the region surrounding the lake is dominated by shrub tundra as it is today. The effects of interglacial forcing on distributions of terrestrial ecosystems (corresponding to the July temperature anomalies in Fig. 3I) can be clearly seen by the poleward spread of deciduous and evergreen taiga (Fig. S5B to E). Evergreen forest provides a strong amplifying warming effect on spring and early summer temperatures via reductions in albedo during snow-covered months with substantial insolation (44). The simulated advance of forest into western Beringia during super interglacials, as reflected in percentages of tree pollen (Fig. 3L), is captured by the model (Fig. S5D and E). The resulting positive vegetation-warming feedback accounts for some of the simulated surface warming at the location of the lake in the warmest simulations.
Fig. S5. Distributions of interglacial vegetation simulated by BIOME4 interactively coupled to the GCM. (A) Preindustrial control, (B) MIS 1 (9 ka) corresponding to summer temperature anomalies in Fig. 4A, (C) MIS 5e corresponding to summer temperature anomalies in Fig. 4B, (D) MIS 11c corresponding to summer temperature anomalies in Fig. 4C, and (E) MIS 31 corresponding to summer temperature anomalies in Fig. 4D. The location of Lake El’gygytgyn is shown with the red star. Note the differential poleward advance of taiga biomes into the lake region during each interglacial and near complete replacement of tundra with taiga over most of Beringia in MIS 11c and 31 simulations.

Comparison to the ANDRILL 1B record

Figure S6 illustrates the remarkable coincidence of the super interglacials at Lake El’gygytgyn with the diatomites in the Antarctic ANDRILL 1B record of the Ross Sea, despite some hiatuses by glacial erosion in the latter. These diatomites each represent a major retreat of the West Antarctic Ice Sheet (WAIS) (35). MIS 31 for example, is remarkably well documented in both records as a super interglacial at Lake El’gygytgyn and diatomite in the ANDRILL core. The super interglacial MIS 55 may be associated with a diatomite in the Ross Sea dated to the period MIS 54 to 58. Similarly, super interglacials MIS 77, 87, 91, and 92 may coincide with at least three diatomites described between MIS 74 and MIS 104. Super interglacials MIS 11c and 49 appear not to be correlative with diatomites in the Ross Sea, however, this could be due to the discontinuous record of ANDRILL 1B.
Fig. S6. Stratigraphic correlation of the super interglacials at Lake El’gygytgyn with the diatomites in the ANDRILL 1B record, the latter interpreted to mark major retreats in the West Antarctic Ice Sheet (WAIS) (35). (A) Facies succession in the El’gygytgyn record throughout the past 2.8 Ma. Peak glacial conditions are indicated by Facies A, intermediate glacial to interglacial conditions by Facies B, and super interglacials by parts of Facies C (see red bars). (B) LR04 global marine isotope stack (12) with selected numbers of isotope stages. (C) Lithology of the uppermost 215 m of the ANDRILL 1B record modified from (46) and (107) and correlated to the LR04 stack according to (35). Quaternary diatomites according to (35) are highlighted with orange arrows. Saw-toothed lines indicate larger (red) and smaller (black) unconformities.

References and Notes:


10. Materials and methods are available as supplementary material on *Science* Online.


SOM References:


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