A Gcm Comparison of Plio-Pleistocene Interglacial-Glacial Periods in Relation to Lake El'gygytgyn, Ne Arctic Russia

Anthony J. Coletti
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A GCM Comparison of Plio-Pleistocene Interglacial-Glacial Periods in Relation to Lake El’gygytgyn, NE Arctic Russia

A Thesis presented

By

ANTHONY J. COLETTI

Submitted to the Graduate School of the University of Massachusetts Amherst in partial fulfillment of the requirements for the degree of

MASTERS OF SCIENCE

SEPTEMBER 2013

Program in Geosciences
A GCM Comparison of Plio-Pleistocene Interglacial-Glacial Periods in Relation to Lake El’gygytgyn, NE Arctic Russia

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ACKNOWLEDGMENTS

I would like to sincerely thank Robert M. DeConto for guiding me through my research project and having the patience while I learn and enhance my knowledge of MATLAB and using climate models. I would also like to thank Julie Brigham-Grette for having the patience as I learned the specific tools to carry out my research. She has instilled an immense amount of interest in the Arctic for me and I hope to continue research within that region. Finally, I’d like to thank Alan Condron for taking a chance by joining my committee about a research project that he did not know a lot about and giving me MATLAB advice during my first semester in graduate school. I strongly thank all my committee members for the invaluable help they have given me throughout my years doing my Master’s Degree.

I thank all my family and friends especially my mother, Eileen. I thank her for raising two boys on her own for 26 years, and giving me the drive and strength to continue my education by getting my Master’s Degree. All her hard work and dedication to my brother and I has taught me an invaluable lesson from day one and I hope to be as strong as she with my family and career.

Thank you, Mom.
ABSTRACT

A GCM COMPARISON OF PLIO-PLEISTOCENE INTERGLACIAL-GLACIAL PERIODS IN RELATION TO LAKE EL’GYGYTGYN, NE ARCTIC RUSSIA

SEPTEMBER 2013

ANTHONY J. COLETTI, B.S. RUTGERS UNIVERSITY
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Directed by: Professor Robert M. DeConto

Until now, the lack of time-continuous, terrestrial paleoenvironmental data from the Pleistocene Arctic has made model simulations of past interglacials difficult to assess. Here, we compare climate simulations of four warm interglacials at Marine Isotope Stage (MIS) 1 (9ka), 5e (127 ka), 11c (409 ka), and 31 (1072 ka) with new proxy climate data recovered from Lake El’gygytgyn, NE Russia. Climate reconstructions of the Mean Temperature of the Warmest Month (MTWM) indicate conditions 2.1, 0.5 and 3.1 °C warmer than today during MIS 5e, 11c, and 31 respectively. While the climate model captures much of the observed warming during each interglacial, largely in response to boreal summer orbital forcing, the extraordinary warmth of MIS 11c relative to the other interglacials in the proxy records remain difficult to explain. To deconvolve the contribution of multiple influences on interglacial warming at Lake El’gygytgyn, we isolated the influence of vegetation, sea ice, and circum-Arctic land ice feedbacks on the climate of the Beringian interior. Vegetation-land surface feedback simulations during all four interglacials show expanding boreal forest cover with increasing summer insolation intensity. A deglaciated Greenland is shown to have a minimal effect on Northeast Asian temperature during the warmth of stage 11c and 31 (Melles et al., 2012). A prescribed
enhancement of oceanic heat transport into the Arctic ocean has some effect on Beringian climate, suggesting intrahemispheric coupling seen in comparisons between Lake El’gygytgyn and Antarctic sediment records might be related to linkages between Antarctic ice volume and ocean circulation. The exceptional warmth of MIS 11c remains enigmatic however, relative to the modest orbital and greenhouse gas forcing during that interglacial. Large Northern Hemisphere ice sheets during Plio-Pleistocene glaciation causes a substantial decrease in Mean Temperature of the Coldest Month (MTCM) and Mean Annual Precipitation (PANN) causing significant Arctic aridification. Aridification and frigid conditions can be linked to a combination of mechanical forcing from the Laurentide and Fennoscandian ice sheets on mid-tropospheric westerly flow and expanded sea-ice cover causing albedo-enhanced feedback.
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CHAPTER 1

INTRODUCTION

Knowledge of our planet’s climate history has increased dramatically over the past three decades but there is much that remains poorly understood. It is important to understand the effects of rapid warming in the Arctic and systemic teleconnections to other latitudes. Within the last 100 years, global temperature has risen approximately 0.47 °C (1.33 °F) with eleven out of the last twelve years (1995-2006) being the warmest years in the instrumental record of global surface temperature (since 1850) (AASA, 2012). Increased seal level rise at rates of 3.1 mm year\(^{-1}\) with contributions from thermal expansion, melting glaciers and ice caps and polar ice sheets demonstrate further evidence of extreme global warming (AASA, 2012). Past warm periods known as Interglacials over the past 2.8 million years provide a means of studying global changes during a climate warmer than today giving us possible outcomes of trends seen in the modern world.

In 2009, a team of scientists from the United States, Germany, Russia and Austria drilled a 355 meter (1,165 ft.) sediment core from an 11 mile wide impact crater lake named “Lake El’gygytgyn” (alternatively, Lake “E”), in northeast Siberia. The recovered core contained the longest Arctic terrestrial record, extending back ~ 3.5 million years. The sediment core revealed evidence for exceptionally warm periods in the Arctic; each lasting tens of thousands of years. These warm periods are marked by relatively large negative excursions of δ\(^{18}\)O and are previously seen in ocean sediment cores. It has been shown that Marine Isotope Stage(s) 5e, 11c and 31 were some of the warmest interglacials in the Pleistocene Arctic (Melles et al., 2012) and are of important interest as
they can be considered an analogue for a warmer, future Arctic that our climate may be heading toward.

An interglacial period (alternatively called an interstadiul) is a period in geological time that is marked by warmer than average global temperatures that last thousands of years (fig 1.1). Evidence of such periods lie within many geological records such as deep-sea sediment cores, ice cores and speleothem analysis and are marked by large negative oxygen isotope ($\delta^{18}$O) excursions in the oxygen isotope record obtained from the composition of foraminiferal carbonate (Shackleton, 1967).

![Ice Age Temperature Changes](image_url)

**Figure 1.1:** Ice Age temperature differences and global ice volume for the first 450 kyr. The blue and green curves show temperature from two sites in Anatarctica derived from Deuterium measurements ($\delta D$) on ice cores (Augustin et al., 2004; Petit et al., 1999) and the red curve is global ice volume derived from $\delta^{18}$O measurements on benthic foraminifera from globally distributed ocean sediment cores (Lisiecki & Raymo, 2005). Notice that low and high global ice volume correlates with peaks and troughs in global average temperature showing interglacial and glacial periods, respectively. *(This image is under the GNU Free Documentation license)*
1.1 Marine Isotope Stage 1

Marine Isotope Stage (MIS) 1 represents the last 11,000 years and is marked by its onset near the end of the Younger-Dryas. Peak insolation anomalies occur ~ 9 kyr, when summer insolation was ~510 Wm$^{-2}$ at 65 $^\circ$N. Conditions were warmer than present (+1.6 $^\circ$C over western Arctic and +2 to 4$^\circ$C in circum-Arctic) with lush birch and alder shrubs (Melles et al., 2012) dominating vegetation in the lake region. This period, known as the Holocene Climate Optimum (HCO), was mostly spatially and temporally variable affecting the high latitudes while there was minimal warming in the mid-latitudes and tropics (Kitoh & Murakami, 2002).

1.2 Marine Isotope Stage 5e

Interglacial 5e is one of the warmest interglacials of the Pleistocene and lasted roughly ~12-10 kyr (130 to 116 kyr). High obliquity, eccentricity and precession allowed for very warm summer orbit and insolation intensity maximum at around 127 kyr. During the Last Interglacial (LIG), a warmer climate throughout the Arctic possibly caused a size reduction of the Greenland Ice Sheet (GIS). Studies involving Sr – Nd – Pb isotope ratios of silt-sized sediment discharged from southern Greenland suggest that no single southern Greenland geologic terrain was completely deglaciated during the LIG, however, greater southern GIS retreat was evident (Colville et al., 2011). Additional analysis of MIS 5e by Yin & Berger (2011) involved running a model of intermediate complexity to test contributions of Greenhouse Gas (GHG) and insolation forcing. It was found that GHGs play a dominant role on the variations of annual mean temperature of both the globe and the southern high latitudes, whereas insolation plays a dominant role
on the variations of tree fraction, precipitation of the northern high latitudes, temperature and sea ice (Yin & Berger, 2011). Focusing on the combined effect of GHG and insolation forcing, MIS 5e is one of the warmest interglacials of the Pleistocene.

Similarly, simulations of Arctic warmth during the LIG using the National Center for Atmospheric Research’s (NCAR) Community Climate System Model (CCSM) ver. 2, without vegetation feedbacks yielded solar anomalies that lead to significant summer (June, July and August) Siberian warming (Otto-Bliesner, 2006). Additional warming during this period supports the notion of a significantly reduced Greenland Ice Sheet.

1.3 Marine Isotope Stage 11c

Interglacial 11c is another exceptionally warm interglacial that lasted from 428 to 383 kyr (~45 kyr). Sediment records containing information on MIS 11 are very uncommon in the Arctic and temperature data are inadequate (Miller et al., 2010). Unlike the other interglacials, MIS 11c was remarkably long with two insolation maxima anomalies at ~ 409 kyr and 423 kyr, creating extensive warmth throughout the Arctic. Similar to MIS 5e, there is evidence that the GIS may have been reduced (Raymo & Mitrovica, 2012; Willerslev et al., 2007) with lush boreal forest covering most of southern Greenland (de Vernal & Hillaire-Marcel, 2008). Particularly warm conditions are also suggested by pollen records analyzed from Lake Biwa (Tarasov et al., 2011). Likewise, a similar study done with a record from Lake Baikal also shows warmer than modern temperatures with a “conifer optimum” suggesting not only warmer conditions but also less continentality and higher sea levels than present (Prokopenko et al., 2010).
1.4 Marine Isotope Stage 31

MIS 31 lasted from 422 to 395 kyrs. Unfortunately, data has not been identified in most Arctic records, but what little evidence there is suggests a warmer climate relative to today at the poles. Interglacial 31 is best known for changes within Antarctica such as collapse of the marine based West Antarctic Ice Sheet (WAIS) (Pollard & DeConto, 2009) and a poleward shift, or deterioration, of the Polar Front (Scherer et al., 2008) which allowed the intrusion of warm surface waters onto Antarctic continental shelves reducing Antarctic sea ice. On Ellesmere Island, Fosheim Dome includes terrestrial deposits that date to ~1.1 Ma, which contains fossil beetle assemblages for MIS 31 suggesting temperatures of 8 to 14 °C above modern values (Elias & Matthews Jr., 2002). It is speculated, like that of MIS 11c and 5e, the Arctic may have been too warm to support a Greenland Ice Sheet therefore, the Greenland Ice Sheet may have been severely reduced in size, or possibly nonexistent (Raymo & Mitrovica, 2012).

1.5 Background

1.5.1 Setting and Today’s Climate

Lake El’gygytgyn is an impact crater located in northeast Siberia, 150 km southeast of Chaunskaya Bay at 67.30° N and 172° E and ~ 100 km to the north of the Arctic Circle (fig 1.2). Lake El’gygytgyn resides in an area of the western arctic known as Beringia, which is bounded by the 140° E meridian to the west and the Alaska/Canada border to the East, the 76° N parallel to the North, and the 50° N parallel to the South (Mock et al., 1998). Beringia is separated into two distinct geographical regions. Western Beringia, where the lake is located, is west of the Bering/Chukchi Sea and
Eastern Beringia is east of the Bering Sea and incorporates all of Alaska and the Yukon. January monthly mean temperatures range from -47 to -5 °C in Western Beringia to -30 to 0 °C in Eastern Beringia (Mock et al., 1998). Averaged July monthly temperatures tend to increase with decreasing latitude, with values ranging from 0 to 16 °C (Mock et al., 1998). For most of Beringia, precipitation is at maximum during mid- to late-summer with averaged July precipitation amounts varying from 50-100 mm (NCDC, 1999). Combined, both Western and Eastern Beringia have July averaged monthly precipitation in the range of 25 to 225 mm including both Western and Eastern Beringia.

![Figure 1.2: Location of Lake El’gygytgyn within the Beringian region. Exact location is within the orange box. (This image is under public domain from NASA).](image)

Today, mean annual air temperature at the lake is -10.4 ± 1.1 °C with daily average temperatures during the summer (JJA) ranging from 0 to 12 °C (Nolan & Brigham-Grette, 2006). Extremes in 2002 ranged from -40 °C in winter to as high as
+26 °C in the summer with occasional mid-winter warming approaching 0 °C (Nolan & Brigham-Grette, 2006). Precipitation amounts at the lake are rather small indicating a dry environment typical of the Arctic (<300 mm year^{-1}). Weather stations implemented around the lake in 2002 recorded 70 mm of rainfall all between mid-May and late-September with transient snowfall greater than 5 cm beginning in mid-July and lasting the rest of the summer and 178 mm of precipitation from the end of summer 2001 to end of summer 2002 (Nolan & Brigham-Grette, 2006).

1.6 Research Statement

The primary goal of this investigation is to study Arctic climate variability and sensitivity to the exceptionally warm interglacials during the past 2 million years (Pleistocene) and correlate the modeling data with the Lake El’gygytgyn multiproxy analysis. By studying the interglacial and glacial climates, Arctic variability can be assessed and quantified with an aim toward studying the teleconnections associated with them. This is especially important as the Earth continues to warm due to anthropogenic emission of Greenhouse Gas (GHG). The work discussed in this thesis will advance on the work already done on MIS 5e, 11c and 31. Such advances on the original work include 3 simulations with high-resolution interactive vegetation to show 1) biome regimes and vegetation feedbacks, 2) the sensitivity of the circum-Arctic to the Greenland Ice Sheet, 3) the sensitivity of the lake region to the reduced Arctic sea ice owing to the intrusion of warm north Pacific waters into the Arctic Ocean and 4) circum-Arctic and Lake El’gygytgyn sensitivity to major northern hemisphere ice sheets.
The document presented here is a progression from the description of the various tools and methods used, while also describing the model boundary conditions and details of the model simulations (Chapter 2). This is followed by a presentation of the results from each model simulation discussed in detail (Chapter 3). Discussion of simulations and overall climatic patterns are explained in (Chapter 4). Chapter 5 discusses the glacial paleo-arctic environment before the interglacial transition. Conclusion of analysis and research statement summary will be displayed in (Chapter 6). All figures and tables embedded throughout the text will be labeled and noted in the table of contents.
CHAPTER 2

METHODS

2.1 Mathworks MATLAB R2011b

Mathworks MATLAB is a high-level language and interactive environment for matrix-based numerical computation. All post-processing and analysis was conducted using computer-based programs (alternatively called scripts) executed in MATLAB. MATLAB provides both high and low-end access to NetCDF (Network Common Data Form) files, which are produced by the model, allowing read and writes capabilities from the MATLAB interface to the NetCDF format. MATLAB also contains low-level access to common NetCDF functions in order to access NetCDF libraries. The ability for MATLAB to access such libraries with large amounts of data makes it an ideal program to visualize and compute model output data.

2.2 Global Climate Model: GENESIS Version 3.0

All global climate simulations discussed in this thesis were performed using the Global ENvironmental and Ecological Simulation of Interactive Systems (GENESIS) Global Climate Model (GCM) version 3.0 (Thompson & Pollard, 1997). The GCM is written in Fortran and ran in parallel on a Silicon Graphics (SGI) computer running 10 years per day. GENESIS is an atmosphere, land-surface, ocean, snow, sea-ice, ice sheet and vegetation coupled model. Spectral resolution of the 3-D atmosphere GCM (AGCM) within GENESIS is T31 resolution (3.75° lat. x 3.75° long.) with 18 vertical levels (Thompson & Pollard, 1997). The AGCM is coupled to the land-surface model by a Land-Surface-Transfer scheme (LSX), which computes fluxes through the vegetation
model (Pollard & Thompson, 1995). In addition to a coupled AGCM and land-surface scheme, GENESIS allows for a 50-meter, mixed-layer, non-dynamical, slab ocean model that incorporates heat transfer, calculations of sea-surface temperatures (SST) and feedbacks operating between ocean surface and sea-ice. This version of the GCM has sensitivity to 2xCO₂ of 2.9 °C, without GHGs, vegetation or ice sheet feedbacks.

The sea-ice model is based upon the sea ice component used in Washington & Meehl (1996) in a fully coupled atmosphere, ocean and sea ice model. Sea ice dynamics are based upon the cavitating fluid method derived by (Flato & Hibler, 1992). The sea ice component is driven by momentum; heat and freshwater fluxes provided at the upper and lower ice boundaries from the atmospheric and oceanic model components. A Flux Coupler (Bryan et al., 1996) facilitates and manages the exchange of fluxes between the model components and conserves equations of heat, momentum and freshwater within the model climate system (Weatherly et al., 1996). Sea ice is able to drift in the model using the shear stress of the wind across the upper boundary of the ice.

2.2.1 AGCM Overview

The 3-D atmosphere model (AGCM) is a modified version of the NCAR Community Climate Model Version 1 (CCM1). Solar radiation calculations are performed every 1.5 hours, which includes a diurnal cycle, and solar infrared radiation calculations are executed every 0.5 hours. Calculations such as absorptivities and emissivities of H₂O, CO₂ and O₃ gasses are done every 24 hours due to computationally intense calculations. GENESIS uses an adapted version of the NCAR CCM3 solar and thermal infrared radiation code. The solar radiation scheme of (Thompson et al., 1987) is
included, which combines all clouds into a single effective layer for solar computations and allows for effects of atmospheric aerosols.

Clouds in the GCM are parameterized similar to Slingo & Slingo (1991) for three different clouds types: stratus, anvil cirrus, and convective (Thompson & Pollard, 1995). A constraint on stratus clouds is used when specific humidity is very low permitting sensible amounts of clouds to form over Polar Regions during winter (Thompson & Pollard, 1995). Clouds in the GCM are formed using a plume model similar to (Kreitzberg & Perkey, 1976) but does not include cloud microphysics. At each horizontal gridpoint and at each time step, a column model of subgrid rising thermals is solved including saturation and precipitation (Thompson & Pollard, 1995). From the previously solved equations, large-scale vertical fluxes, latent heat and precipitation can be inferred. Similarly, the planetary boundary layer (PBL) is calculated using the same model by initiating dry PBL thermals at the center of the lowest model layer (Morton, 1968; Telford, 1966). Surface (2° lat. x 2° long.) and AGCM (3.75° lat. x 3.75° long.) fields are coupled to one another by two different model schemes: bilinear interpolation (AGCM to surface) and forward area averaging (surface to AGCM) at each time step.

2.2.2 BIOME4 Vegetation model Overview

GENESIS GCM is fully coupled to the BIOME4 (Kaplan, 2003) interactive vegetation model that was developed from the BIOME3 model of (Haxeltine & Prentice, 1996). BIOME4 is a coupled carbon and water flux model that predicts equilibrium vegetation distribution, structure and biogeochemistry. Vegetation distributions take the form of 27 plant biomes including 12 plant functional types (PFTs) that represent broad,
physiologically distinct classes ranging from cushion-forbs to tropical rain forest trees (Kaplan, 2003). Each PFT is assigned limits in relation to climate, which would define whether or not the functional type exists within that grid cell. Identification of the biome in each grid cell is determined by the ranking of the PFTs, given by the model. The ranking is based on biogeochemical variables, such as leaf area index (LAI), the monthly mean climatology and mean annual soil moisture, which determine the appropriate biome.

Simulated BIOME4 vegetation distributions in mid to high latitude compare favorably with standard potential natural vegetation maps (fig 2.1), pollen surface samples, field-based maps of vegetation for modern day (Bigelow, 2003; Kaplan, 2003; Wohlfahrt et al., 2004; Koenig et al., 2012) and for the past with available pollen data for the mid-Holocene (Koenig et al., 2012; Wohlfahrt et al., 2008, 2004), Last Glacial Maximum (Kaplan, 2003), the Pliocene (Salzmann et al., 2009; Salzmann et al., 2008) and the late Paleozoic (Horton et al., 2010). Uncorrected distributions over boreal high latitudes are close to observed allowing us to run paleoclimatic simulations without bias corrections (Koenig et al., 2012).
Figure 2.1: Comparison of vegetation simulations to present-day potential natural vegetation. a) Present-day natural potential vegetation, b) vegetation simulated by BIOME4 north of 55° N, with legend (c). Map sectors are labeled in (d). (Figure and caption from (Kaplan, 2003)).
<table>
<thead>
<tr>
<th>Biome #</th>
<th>Biome Type</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Tropical evergreen forest</td>
</tr>
<tr>
<td>2</td>
<td>Tropical semi-deciduous forest</td>
</tr>
<tr>
<td>3</td>
<td>Tropical deciduous forest/woodland</td>
</tr>
<tr>
<td>4</td>
<td>Tropical xerophytic shrubland</td>
</tr>
<tr>
<td>5</td>
<td>Temperate xerophytic shrubland</td>
</tr>
<tr>
<td>6</td>
<td>Tropical grassland</td>
</tr>
<tr>
<td>7</td>
<td>Temperate grassland</td>
</tr>
<tr>
<td>8</td>
<td>Temperate conifer forest</td>
</tr>
<tr>
<td>9</td>
<td>Warm mixed forest</td>
</tr>
<tr>
<td>10</td>
<td>Cool mixed forest</td>
</tr>
<tr>
<td>11</td>
<td>Cool conifer forest</td>
</tr>
<tr>
<td>12</td>
<td>Cold mixed forest</td>
</tr>
<tr>
<td>13</td>
<td>Temperate deciduous forest</td>
</tr>
<tr>
<td>14</td>
<td>Evergreen taiga/montane forest</td>
</tr>
<tr>
<td>15</td>
<td>Deciduous taiga/montane forest</td>
</tr>
<tr>
<td>16</td>
<td>Tropical Savannah</td>
</tr>
<tr>
<td>17</td>
<td>Temperate broadleaved savanna</td>
</tr>
<tr>
<td>18</td>
<td>Open conifer woodland</td>
</tr>
<tr>
<td>19</td>
<td>Temperate sclerophyll woodland</td>
</tr>
<tr>
<td>20</td>
<td>Boreal Parkland</td>
</tr>
<tr>
<td>21</td>
<td>Steppe Tundra</td>
</tr>
<tr>
<td>22</td>
<td>Shrub Tundra</td>
</tr>
<tr>
<td>23</td>
<td>Dwarf shrub tundra</td>
</tr>
<tr>
<td>24</td>
<td>Prostrate shrub tundra</td>
</tr>
<tr>
<td>25</td>
<td>Cushion-forbs lichen and moss</td>
</tr>
<tr>
<td>26</td>
<td>Desert</td>
</tr>
<tr>
<td>27</td>
<td>Barren</td>
</tr>
<tr>
<td>28</td>
<td>Land ice</td>
</tr>
</tbody>
</table>

Table 2.1: The 28 biome types calculated by BIOME4. Each biome is defined by PFT and leaf area index (LAI) based on environmental factors and mean climatology.
2.3 GENESIS Experimental Setup

Milankovitch orbital parameters, such as Eccentricity, Obliquity and Precession, for each model simulation can be prescribed through the model namelist parameters allowing full control of Earth’s orbit. The orbital values, affect insolation at the top of the atmosphere (TOA) at each model timestep. All orbital parameters used here are based upon the astronomical solutions of Berger (1978). Precessional values need to be converted from longitude of precession defined as $\Omega (°)$, to the definition of precession used by the GCM using the equation below:

\[
360 - \Omega - 180 \quad if \quad \Omega < 180 \\
360 - \Omega + 180 \quad if \quad \Omega > 180
\]

**Equation 2.1**: Model equation to convert omega, to PRECC value.

In addition to control over Milankovitch parameters, GENESIS also allows Greenhouse gasses to be prescribed uniformly. GENESIS’s namelist parameters allow changes in Carbon Dioxide ($pCO_2$), Methane (CH$_4$), Nitrous Oxide (N$_2$O) and Chlorofluorocarbons (CFCs). Simulations of each interglacial were run with the proper GHG concentrations from the literature and orbital parameters from Berger’s algorithm.

2.4 GENESIS Boundary Conditions

Boundary conditions in GENESIS were initiated on a 2° lat. x 2° long. surface grid (90 rows x 180 columns). Conditions upon startup are default values within each topography and surface input file. The input parameters read by the model include surface type, gravity-wave roughness, topography, vegetation, ocean-lake fraction,
atmospheric ozone distribution, soil texture and depth (Thompson and Pollard, 1995[guide]).

Surface topography (2° long. x 2° lat.) editing is done through data input files that are interconnected to the AGCM (T31 resolution). Default values in the topography files are derived from the U.S. Navy FNOC global elevation dataset at 10 min. resolution (Cuming & Hawkins, 1981; Kineman & Hastings, 1992) in Fortran I5 and are measured in mean sea level in meters. Surface files are coded with 1=land, 2=ice sheet and 3=ocean in Fortran A1 format. Ice sheet areas were superimposed using Cogley’s 1° x 1° Global Hydrographic Dataset (Cogley, 1991; Pollard and Thompson, 1995[guide]). Greenhouse Gasses are prescribed during initial startup using the model namelist parameters in the configuration file.

2.4.1 Paleoclimate Boundary Conditions

Topography during this study was changed remained largely unchanged, except for simulations of an ice free Greenland where exceptionally warm conditions in the Arctic during interglacials 31 and 11c (Elias & Matthews Jr., 2002; Melles et al., 2012; Raymo & Mitrovica, 2012) prevailed, and a change in Greenland topography and surface type is required if simulations are to be accurate. Removing Greenland’s ice sheet requires changing the surface type and topography input files. Such an edit necessitates a change in elevation of Greenland’s topography by +6 meters to simulate glacial isostatic adjustment (GIA) in each grid cell and a surface type change from ice to land. Similarly, edits of topography and sea level were also needed in paleoclimate simulations with large
Northern Hemisphere ice sheets. Such ice sheet data was extracted from ICE4G (Peltier, 1994) dataset of ice and water cover since the Last Glacial Maximum (LGM).

In simulations when vegetation is not interactive, and is prescribed rather than simulated, vegetation and biome distribution input files, similar to those of topography and surface, must be edited to the correct biome. Biome designations are labeled 1-12 based on (Dorman & Sellers, 1989) vegetation type and designate a single globally uniform vegetation type for all land points (Thompson and Pollard, 1995 [guide]).

2.4.2 Greenhouse Gas concentrations

Greenhouse Gas concentrations were prescribed uniformly. Since MIS 31 lies beyond the age of the oldest ice core record, atmospheric Carbon Dioxide ($pCO_2$) concentrations were prescribed from boron isotopic compositions of foraminifera shells (Honisch et al., 2009) (Table 2.3). For interglacials 1, 5e and 11c, $pCO_2$ was prescribed from high resolution measurements from EPICA Dome C ice core (Lüthi et al., 2008; Yin & Berger, 2011) (Table 2.3). Methane (CH$_4$) and Nitrous Oxide (N$_2$O) were also prescribed from EPICA Dome C ice cores for all marine isotope stages (Loulergue et al., 2008; Schilt et al., 2010) (Table 2.3). Chlorofluorocarbons were not present during the time of this paleoclimate study; therefore it was disregarded. Simulations of glacial inception at 2.7 Ma were run using 300 ppm $pCO_2$.

Greenhouse Gas forcing contributions (Table 1) were calculated using the IPCC simplified calculations for radiative forcing due to CO$_2$, CH$_4$, N$_2$O, and halocarbons, the latter omitted from our experiments (Smithson, 2002).
# Table 2.2: Expressions for calculations of radiative forcing (DF) due to CO$_2$, CH$_4$, and N$_2$O.

The equation for CO$_2$ is from WMO (1999), based on Hansen et al. (1998). *(Table adapted from Smithson, 2002).*

<table>
<thead>
<tr>
<th>Trace Gas</th>
<th>Simplified expression Radiative forcing, $\Delta F$ (Wm$^{-2}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>CO$_2$</td>
<td>$\alpha = 3.35$</td>
</tr>
<tr>
<td></td>
<td>$\Delta F = \alpha (g(C) - g(C_0))$</td>
</tr>
<tr>
<td></td>
<td>$g(C) = \ln (1 + 1.2C + 0.005C^2$</td>
</tr>
<tr>
<td></td>
<td>$\quad + 1.4 \times 10^{-6}C^3)$</td>
</tr>
<tr>
<td>CH$_4$</td>
<td>$\alpha = 0.036$</td>
</tr>
<tr>
<td></td>
<td>$\Delta F = \alpha (\sqrt{M} - \sqrt{M_0}) - (f(M,N_0))$</td>
</tr>
<tr>
<td></td>
<td>$\quad - f(M_0,N_0))$</td>
</tr>
<tr>
<td>N$_2$O</td>
<td>$\alpha = 0.12$</td>
</tr>
<tr>
<td></td>
<td>$\Delta F = \alpha (\sqrt{N} - \sqrt{N_0}) - (f(M,N))$</td>
</tr>
<tr>
<td></td>
<td>$\quad - f(M_0,N_0))$</td>
</tr>
</tbody>
</table>

The equation for CH$_4$ is

$$f(M,N) = 0.47 \ln[1+2.01 \times 10^{-5} (MN)^{0.75}+5.31 \times 10^{-15} M(MN)^{1.52}]$$

C is CO$_2$ in ppm

M is CH$_4$ in ppb

N is N$_2$O in ppb

## 2.5 General Experimental Setup

All simulations of paleoenvironments were run using the GENESIS GCM coupled to BIOME4 vegetation model. Target intervals include Marine Isotope Stage(s)
MIS 1 (11 kyr), 5e (127 kyr), 11c (409 kyr) and 31 (1072 kyr), all corresponding to the timing of peak summer warmth and identified as “super-interglacials” by Melles et al. (2012). It was noted during analysis that Lake El’gygytgyn was recording peak summer temperatures allowing our focus to be on the summer months, especially July, for surface temperature. Simulations of present day and pre-industrial climate were run as the control experiments for future comparisons. The control run simulates atmospheric conditions of the present day (1950 AD concentrations) while the Pre-Industrial run (1750 AD concentrations) simulates conditions at the onset of anthropogenic emissions.

<table>
<thead>
<tr>
<th>Run Name</th>
<th>CO₂ (ppm)</th>
<th>CH₄ (ppbv)</th>
<th>N₂O (ppbv)</th>
<th>Eccentricity</th>
<th>Obliquity</th>
<th>Precession</th>
<th>Temp. (°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</td>
<td>~611</td>
<td>~263</td>
<td>0.01928</td>
<td>24.29</td>
<td>311.26</td>
<td>12.4</td>
<td>438</td>
</tr>
<tr>
<td>MIS 5e-with GIS</td>
<td>287</td>
<td>724</td>
<td>262</td>
<td>0.039378</td>
<td>24.04</td>
<td>275.42</td>
<td>14.5</td>
<td>401</td>
</tr>
<tr>
<td>MIS 11c-with GIS</td>
<td>285</td>
<td>713</td>
<td>285</td>
<td>0.019322</td>
<td>23.781</td>
<td>276.67</td>
<td>12.2</td>
<td>475</td>
</tr>
<tr>
<td>MIS 31-with GIS</td>
<td>325</td>
<td>800</td>
<td>288</td>
<td>0.05597</td>
<td>23.898</td>
<td>289.79</td>
<td>13.8</td>
<td>438</td>
</tr>
<tr>
<td>MIS11c-no GIS</td>
<td>285</td>
<td>713</td>
<td>284</td>
<td>0.019322</td>
<td>23.781</td>
<td>276.67</td>
<td>12.5</td>
<td>438</td>
</tr>
<tr>
<td>MIS11c-no GIS-10Wm⁻²</td>
<td>285</td>
<td>713</td>
<td>284</td>
<td>0.019322</td>
<td>23.781</td>
<td>276.67</td>
<td>13.2</td>
<td>475</td>
</tr>
</tbody>
</table>

Table 2.3: Overview of interglacial simulations performed during this study. Orbital configurations (Berger, 1978) and greenhouse gas (GHG) concentrations (Honisch et al., 2009; Loulergue et al., 2008; Lüthi et al., 2008; Schilt et al., 2010). Modern GHG concentrations are taken from 1950 AD; obliquity is given in degrees and precession is Ω. Temperatures are mean July temperatures. (Table from (Melles et al., 2012) supplemental).

## 2.5.1 Experimental Run – Marine Isotope Stage (MIS) 5e, 127 kyr

Marine Isotope stage 5e was run with uniformly prescribed greenhouse gas concentrations of: 287 ppmv, pCO₂, 724 ppbv, CH₄ and 260 ppbv N₂O. Obliquity and Eccentricity will be set to a value of 24.04°, 0.039378 (Berger, 1978) respectively, and precession (omega, Ω), 274.41 (Berger, 1978) converted to 264.59 using equation 1 (see table 2.3). The purpose of this simulation will be to simply simulate the paleoenvironment in the Arctic during this period and investigate temperature, vegetation
and precipitation and correlate the data to pollen proxy analysis. Orbital and GHG values are estimated for 127 kyr; peak warmth during MIS 5e.

2.5.2 Experimental Run – Marine Isotope (MIS) 11c, 409 kyr

In this section, there will be three different simulations to test the sensitivity of the lake region during MIS 11c. The first simulation will be done with default boundary conditions, including a Greenland Ice Sheet and will be referred to as MIS11GIS. The second simulation will test the sensitivity of the Arctic to an ice-free Greenland, hereafter known as MIS11NG. The scientific literature shows that during interglacial 11c, the Greenland Ice Sheet was significantly reduced and warm boreal forests (spruce, alder, etc.) covered parts of the island (Raymo & Mitrovica, 2012; Willerslev et al., 2007). Consequently, the GIS was removed and topography of Greenland was corrected for glacial isostatic adjustment (GIA) within the appropriate model topography files. The final sensitivity experiment involved an increase in sub-oceanic surface heat flux from 2 Wm$^2$ in our modern control, to 10 Wm$^2$ (additional +8 Wm$^2$) to test the Beringian sensitivity to an ice-free Arctic Ocean. Today, the Bering Strait is limited to ~ 50 m in depth with a northward transport of ~ 0.8 Sv (Woodgate, et al., 2010). The increase heat flux assumes an extreme 3 Sverdrup (Sv) increase in Bering Strait throughflow and a 4 °C temperature contrast between North Pacific and North Polar surface water (Melles et al., 2012, supplemental). The additional heat flux convergence was used to simulate increases in energy flux through a wider Bering Strait during times of higher sea level. Using BIOME4, comparison of Arctic vegetation within the Beringian region can be analyzed in order to compare model and pollen proxy data that were collected from Lake
El’gygytgyn. Furthermore, fixed vegetation studies using BIOME4 will isolate and quantify the forcing effect of vegetation on surface temperatures around the lake region. Concentrations of GHGs will be prescribed as: 285 ppm \( pCO_2 \), 713 ppb, \( CH_4 \), and 285 ppb, \( N_2O \). In terms of orbital parameters, Obliquity and Eccentricity will be set to a value of 23.78°, 0.019322 (Berger, 1978) respectively and omega as 276.67 converted to 263.33 (eq. 1)(see table 2.3).

2.5.3 Experimental Run – Marine Isotope Stage (MIS) 31, 1072 kyr

This period, in addition to MIS 11c, was also speculated to be too warm for a Greenland Ice sheet to exist (Melles et al., 2012). Therefore, model runs with and without a Greenland ice sheet (including glacial isostatic adjustment) were executed to show sensitivity and forcing feedback for these scenarios.

Concentrations of GHGs will be prescribed as: 325 ppm \( pCO_2 \), 800 ppb, \( CH_4 \), and 288 ppb, \( N_2O \). Orbits from Yin and Berger (2011) show a very warm orbit with Obliquity and Eccentricity of 23.89° and 0.05597, respectively. Precessional value is rather large, with a value of 289.79 after conversion to model specific value (eq. 1).

2.6 Experimental Analysis

In order to test the sensitivity of the Arctic, especially the Western Arctic to changing boundary conditions, data, such as temperature, precipitation and vegetation, were plotted and compared to control runs (Pre-Industrial and Modern control). The goal of this study is to observe the Arctic’s climatic and terrestrial response to high levels of greenhouse gasses and warm orbits that coincide with the interglacial periods. Such
responses being studied are the effects on temperature, precipitation and vegetation in and around the lake region. Moreover, analyses of atmospheric properties such as sea level pressure and geopotential heights were analyzed to show pressure anomalies that may be linked to changes in topography and ice sheets in the circum-arctic. Using these data, comparisons of model output temperatures and precipitation relative to Pre-Industrial and Modern control runs can be studied. More importantly, pollen analysis done on the lake core (Melles, Brigham-Grette et al., 2012) can be validated by analyzing surface temperatures and precipitation whereas also validating plant assemblages by using vegetation output from the BIOME4 interactive vegetation component of GENESIS, or vice-versa. Possible changes in atmospheric circulation, temperature and precipitation due to regional changes in topography and ice sheets will also be considered and associated to control scenarios.

2.7 Model Output Post-Processing

GENESIS surface-model history files (LSX) contain 54 variables in monthly mean data sets. Analysis using these history files will focus mainly on 2-meter surface temperature and precipitation. Likewise, AGCM history files contain 38 variables in monthly data sets. The variables used here during the study were 500 hPa geopotential heights, surface and sea-level pressure and insolation at top of the atmosphere (TOA). Simulations of the specific time periods were ran for 30 to 40 years to ensure model climate equilibration with initial conditions and a 50-meter slab ocean. For analysis, the last 10 years of data (20-30; 30-40) was extracted and averaged over a 180 x 90 x 12 grid. This grid represents 180 degrees of longitude, 90 degrees of latitude containing 12
months (one year) of data. On this grid, averaged monthly or yearly data can be plotted on a map projection allowing it to be visually attractive for publishing and easily examined for data analysis.

2.7.1 BIOME4 Output Processing

Biome vegetation output was analyzed by accessing the last year of vegetation in the equilibrated run. Additionally, averaging biome data becomes a programming challenge. This is due to the arrangement of the 28 biomes and the fact that, for example, the 11th biome type may not all be related to the 12th biome type (Tracey, 2012, Master’s Thesis). Hence, vegetation is represented by the last year and month of simulation.
<table>
<thead>
<tr>
<th>Biome #</th>
<th>Biome Type</th>
<th>Color</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Tropical evergreen forest</td>
<td></td>
</tr>
<tr>
<td>2</td>
<td>Tropical semi-deciduous forest</td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>Tropical deciduous forest/woodland</td>
<td></td>
</tr>
<tr>
<td>4</td>
<td>Tropical xerophytic shrubland</td>
<td></td>
</tr>
<tr>
<td>5</td>
<td>Temperate xerophytic shrubland</td>
<td></td>
</tr>
<tr>
<td>6</td>
<td>Tropical grassland</td>
<td></td>
</tr>
<tr>
<td>7</td>
<td>Temperate grassland</td>
<td></td>
</tr>
<tr>
<td>8</td>
<td>Temperate conifer forest</td>
<td></td>
</tr>
<tr>
<td>9</td>
<td>Warm mixed forest</td>
<td></td>
</tr>
<tr>
<td>10</td>
<td>Cool mixed forest</td>
<td></td>
</tr>
<tr>
<td>11</td>
<td>Cool conifer forest</td>
<td></td>
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<tr>
<td>12</td>
<td>Cold mixed forest</td>
<td></td>
</tr>
<tr>
<td>13</td>
<td>Temperate deciduous forest</td>
<td></td>
</tr>
<tr>
<td>14</td>
<td>Evergreen taiga/montane forest</td>
<td></td>
</tr>
<tr>
<td>15</td>
<td>Deciduous taiga/montane forest</td>
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<tr>
<td>26</td>
<td>Desert</td>
<td></td>
</tr>
<tr>
<td>27</td>
<td>Barren</td>
<td></td>
</tr>
<tr>
<td>28</td>
<td>Land ice</td>
<td></td>
</tr>
</tbody>
</table>

**Table 2.4: BIOME4 biome/vegetation color key.** Each biome is plotted in a different color in order to make biomes discernible from one another.
CHAPTER 3

RESULTS

3.1 Control Simulations

3.1.1 Modern Simulation

3.1.1.1 Temperature

In order to test the model’s ability to accurately simulate 2-meter (2-m) surface temperature at the lake, the data were compared to modern day temperature observations. This test allows confidence in surface temperature simulations and calculations that do not involve actual observations. The control simulation yielded mean annual 2-m air temperature (MAAT) of -9 °C, which is within error-range of MAAT of -10.3 ± 1.1 °C, recorded by Nolan and Brigham-Grette (2006) in 2002, using weather station measurements from around the lake. Mean summer (JJA) and Mean Temperature of the Warmest Month (MTWM; July) surface temperatures were simulated to be 10.2 and 12 °C, respectively (fig 3.1 A) which is on par with the current climatology of the region based on reanalysis. Warm summer temperatures are owed to a modern orbit consisting of large obliquity (~24.438°) creating warm summers and allowing an annual insolation of 212 Wm⁻². Modern summer insolation anomalies were calculated to be 421 Wm⁻² and July insolation alone was 445 Wm⁻² at the top of the atmosphere indicating peak insolation anomalies during this month (fig 3.2 D).

To further test the validity of the GCM temperatures, a comparison was made with National Center for Environmental Prediction (NCEP) Reanalysis data. The difference indicates that GENESIS is only +0.5 °C warmer than the modern reanalysis data in the lake region signifying relatively reliable results when doing calculations with
July surface temperatures (fig 3.1). Yet, GENESIS presents a warm bias over Greenland and parts of Northeastern Canada, and a cold bias in central, interior Russia compared to NCEP data.

![GENESIS minus NCEP Reanalysis 2-m Modern July surface Temperature](image)

**Figure 3.1: GENESIS 2-m Modern average July surface temperature difference from NCEP Reanalysis data.** The green star marks Lake El’gygytgyn location. The lake region is within a zone of white indicating very little difference between our GENESIS temperatures and NCEP’s Reanalysis. Areas of no shading (white) roughly correspond to statistically significant anomalies at the 95% confidence interval.

### 3.1.1.2 Precipitation

Control simulation of Mean Annual Precipitation (PANN) (fig 3.1 B) was rather high, indicating ~ 475 mm year\(^{-1}\) of liquid precipitation. This is exceptionally greater than Nolan & Brigham-Grette (2006) analysis of 178 mm year\(^{-1}\) from a summer-to-
summer yearly measurement and may be associated with interannual variability within the model simulations. The lowest amounts of precipitation are seen directly at the poles (~ 200 mm year\(^{-1}\)) with higher amounts of precipitation seen over the Bering Sea, Northwestern Yukon and the North Pacific, in some cases exceeding 1100 mm year\(^{-1}\). Mean summer precipitation in the lake region is ~ 63 mm month\(^{-1}\) indicating similar summer precipitation conditions around the lake (Melles, Brigham-Grette et al., 2012; Nolan & Brigham-Grette, 2006). Moreover, winter (December, January, and February) was rather dry, showing precipitation amounts of ~ 26 mm month\(^{-1}\). It is important to reiterate that Lake El’gygytgyn climatology is fairly dry, on the order of 178 mm year\(^{-1}\) of liquid precipitation. Our simulations suggest + 297 mm year\(^{-1}\) relative to annual precipitation amounts from observations and + 225 mm year\(^{-1}\) relative to annual precipitation amounts derived from analysis of pollen proxy. In other words, GENESIS exhibits somewhat of an overall wet bias in regards to annual precipitation in our study region.

### 3.1.1.3 Vegetation Distribution

Modern model simulations of biome distribution (fig 3.2 C) show the lake region and most of the Beringian interior is covered by Evergreen taiga/montane forest, with some exception along the coasts (East Siberian Sea, Chukchi Sea, Beaufort Sea), where Dwarf and Shrub tundra are dominant. Additionally, deciduous taiga/montane forests heavily dominate interior Siberia and northern coast with a few areas of shrub tundra and grassland mixed in. Warm and cool mixed forests seem to dominate further South, on Kamchatka Peninsula. However, vegetation in this control scenario does not match up
with modern observations and field data of modern, Arctic vegetation in the region (Kolosova, 1980; Viereck & Little Jr, 1975). With this said, I hypothesize that the vegetation is not in full balance with the environment suggesting it is still transitioning into equilibrium.
Figure 3.2: Simulated modern (control) mean summer 2-m surface temperature and PANN. A) Modern mean summer (JJA) 2-m surface temperature (°C) and B) Modern mean annual precipitation (mm year$^{-1}$). Black and Red stars indicate location of Lake El’gygytgyn. Mean summer air temperature in our Modern control run is \(~12\) °C.
Figure 3.3: Simulated modern vegetation in the Arctic and monthly insolation averaged over latitude. C) BIOME4 Modern vegetation in Beringia and D) Monthly modern insolation anomalies at top of the atmosphere (Wm$^{-2}$). Y-axis is latitude and X-axis is months (1 – 12; Jan. – Dec.). Red star denotes location of Lake El’gygytgyn. Please note modern simulated vegetation around the lake is not in equilibrium and suggest conifer forests instead of shrub tundra, biome #22.
3.1.2 Pre-Industrial

3.1.2.1 Temperature

Simulations of pre-industrial 2-m MAAT and MTWM at Lake El’gygytgyn are -12 and 10.3 °C, respectively. This is to be expected, as pre-industrial GHG levels are lower than those of present day. Thus, lake regional annual air and July temperatures are -3°C and -1.7°C cooler than those of the modern simulations, respectively (fig 3.4 D, C). Similarly, summer temperatures are cooler as well, on the order of -2.2 °C cooler (8°C) (fig 3.3 A) than modern temperatures. Although Earth’s orbit, specifically obliquity, has not changed in 120 years, temperatures are still cooler than modern temperatures. This is largely due to the fact that low CO₂ emissions during this period attenuate the effect of yearlong radiation being transferred from the atmosphere back to the surface. GHG radiative forcing from a combination of CO₂, CH₄, and N₂O atmospheric mixing ratios determined from the literature indicates a -1.8 Wm⁻² change relative to modern GHG radiative forcing. CO₂ radiative forcing contribution alone is the largest contributor to the decrease in forcing feedback (-1.3 Wm⁻²), all contributing to the cooler surface temperatures.

3.1.2.2 Precipitation

Generally, PANN values in the pre-industrial simulation showed slightly lower values than that of our modern precipitation values. Annual precipitation was around 438 mm year⁻¹ (+122 mm year⁻¹ relative to obs.) (fig 3.3 B) indicating slightly drier conditions in the lake region coinciding with a cooler, pre-anthropogenic warming environment. The same can be said for precipitation amounts in Northwest Yukon and
North Pacific where amounts prior to this run were +200 mm year\(^{-1}\) higher. Mean winter precipitation range was about 25 mm month\(^{-1}\), while mean summer precipitation was 43 mm month\(^{-1}\), indicating -1 and -20 mm month\(^{-1}\) less precipitation relative to modern control, respectively. Most of the circum-Arctic experiences drier conditions during the seasons, with wetter conditions prevailing in the modern runs.

### 3.1.2.3 Vegetation Distribution

Though modern vegetation distributions are not in equilibrium with the environment, pre-industrial vegetation distributions are in equilibrium. Shrub Tundra dominates most of Beringia and the lake region with lingering evergreen taiga and deciduous forests in interior Siberia and Yukon (fig 3.5). Likewise, evergreen taiga dominates southern Alaska and most of the southern half of the Yukon. Dwarf shrub tundra dominates along the coasts bordering northern Alaska and eastern portions of the Beringian coast bordering the Chukchi Sea. Biome distributions are similar to modern day vegetation described by Kolosova (1980) and Viereck & Little Jr (1975) indicating accurate near-modern biome distributions. The switch from evergreen taiga dominating most of interior Beringia to dominate shrub tundra can be attributed to a decrease in CO\(_2\) coinciding with drier, Arctic conditions.
Figure 3.4: Simulated Pre-Industrial (control) mean summer 2-m temperature and PANN. A) Pre-Industrial mean summer (JJA) 2-m surface temperature and B) Pre-Industrial annual precipitation (mm year⁻¹). Red and Black stars denote the location of Lake El’gygytgyn.
Figure 3.5: Simulated temperature difference between Pre-Industrial and Modern temperatures. C) 2-m annual MTWM surface temperature difference (Pre-Ind. minus Modern) and D) 2-m mean annual summer surface temperature difference (Pre-Ind. minus Modern).
Figure 3.6: Simulated Pre-Industrial BIOME4 vegetation around the circum-Arctic with emphasis on Beringia. Vegetation is very similar to modern vegetation in the Arctic. Red star designates Lake El’gygytgyn location.
3.2 Paleoclimate simulations

3.2.1 Marine Isotope Stage 1 (9 kyr); Holocene Thermal Maximum

3.2.1.1 Temperature

Marine Isotope Stage 1 simulation yielded similar mean July temperatures (12.4 °C), only 0.4 °C warmer than modern July temperatures (12.0 °C) in the lake region (fig 3.7 C). Average summer temperatures around the lake are about +1.6 °C (fig 3.6 A) warmer than pre-industrial temperatures, with an overall warming of interior Siberia of >5 °C, but only -0.3 °C (fig 3.7 D) cooler than modern summer temperatures. July temperatures relative to pre-industrial (fig 3.6 B) exceed >+2 °C around most of the lake and Beringia.

3.2.1.2 Precipitation

PANN values are very similar to pre-industrial values and only – 37 mm year\(^{-1}\) drier compared to the modern control simulation (fig 3.8 E). As expected, the Arctic Ocean basin is very dry, averaging about 200 mm year\(^{-1}\) of liquid precipitation. Wetter conditions prevail over high topography and southern latitudes below the Arctic Circle. Somewhat drier conditions prevail in interior Siberia and may be linked to lack of moisture and enhanced continentality.

3.2.1.3 Vegetation Distribution

Most of Alaska is covered with evergreen taiga forest with deciduous toward the north coast. Lake El’gygytgyn is in a transition zone with dominant shrub tundra to the east and deciduous forest to the west (fig 3.9). Most of interior Siberia is deciduous forest
with some desert in the central part of Siberia. Edges of the GIS invoke areas of dwarf and shrub tundra that extend all around the ice sheet in areas that are not glaciated. Most of interior Yukon during this period is evergreen forest with only small amounts of temperate grassland and small deserts. Evergreen and deciduous forests remain the dominant interior biome and shrub tundra remains the dominant coastal biome.
Figure 3.7: MIS 1 2-m summer and July temperature differences from Pre-Industrial (control). A) Mean summer temperature anomalies (MIS 1 – Pre-Ind.) and B) Mean July temperature anomalies (MIS 1 – Pre-Ind.). Anomalies are within the 95% confidence interval.
Figure 3.8: MIS 1 2-m summer and July temperature differences from Modern (control). C) Mean summer temperature anomalies (MIS 1 – Mod.) and D) Mean July temperature anomalies (MIS 1– Mod.)
Figure 3.9: Simulated (PANN) and insolation forcing (relative to Modern) averaged over latitude for MIS 1.
E) Mean annual precipitation and F) Difference of insolation anomalies at the top of the atmosphere (MIS 1 – Mod.). Y-axis is latitude and X-axis is month (1-12; Jan-Dec.).
Figure 3.10: BIOME4 simulated vegetation for MIS 1. Red star denotes location of Lake El’gygytgyn. Vegetation correlates well with increase of trees and shrubs in the Lake El’gygytgyn multiproxy record during peak insolation anomalies.
3.2.2 Marine Isotope Stage 5e (127 kyr)

3.2.2.1 Temperature

Overall warming of the Beringian interior was +5 (±1) °C relative to modern temperatures. Mean annual summer and July temperatures during interglacial 5e show 11 and 14.5 °C, respectively (fig 3.10 A, B). The net effect of this orbital configuration produces high intensity insolation anomalies of >50 Wm⁻² (roughly 60 - 75 Wm⁻²) (fig 3.13 F) at the top of the atmosphere, relative to a modern orbit. Carbon Dioxide (CO₂) concentrations during this period were about 287 ppmv, contributing -1.17 Wm⁻² of surface radiative forcing with total GHG (CO₂, CH₄, and N₂O) contributions of -1.79 Wm⁻² forcing relative to modern GHG ratios.

Average summer warmth and MTWM maximum temperatures around the lake were simulated to be +0.6 and +2.1°C warmer than modern, respectively (fig 3.12 G, H). Comparisons with pre-industrial control simulations show differences of summer and MTWM maxima temperatures (+2.5, +4.2 °C) are similar to comparisons of the modern control simulation, with the exception of July being warmer (fig 3.11 C, D). Mean summer warming over the GIS reflects +5 °C warmer than pre-industrial and only roughly +1 °C warmer than modern simulations.

3.2.2.2 Precipitation

Mean annual precipitation during MIS 5e is about 401 mm year⁻¹ (fig 3.13 E), which is -74 and -37 mm year⁻¹ less than modern and pre-industrial levels, respectively. Overall, similar precipitation patterns are seen over the Arctic between MIS 5e and the pre-industrial control scenario.
3.2.2.3 Vegetation Distribution

Most of Beringia and the lake region is covered by deciduous taiga (fig 3.14) and evergreen taiga biome distributions with evergreen taiga being the most dominant in Alaska and Yukon while deciduous taiga being more dominant around the lake region and the northern coast of interior Siberia. The southern coast near Kamchatka Peninsula contains mostly evergreen taiga biome with some shrub tundra overlapping each other. Most of the Beringian coasts bordering the Bering Strait and Arctic Ocean are dominated by scattered patches of dwarf and prostrate shrub tundra biomes. Both dwarf and shrub tundra biomes mainly dominate Greenland’s coast, with the center being an exception as a near-modern ice sheet covers it.
Figure 3.11: Simulated mean summer and MTWM (July) 2-m temperatures for MIS 5e. A) Mean summer annual temperature and B) Mean July temperatures. GCM temperatures are warmer than pollen proxy assemblages from Lake El’gygytgyn analysis.
Figure 3.12: Simulated 2-m summer and July MIS 5e interglacial warming comparisons with Pre-Industrial. C) Mean summer temperature anomalies (MIS 5e – Pre-Ind.) and D) Mean July temperature anomalies (MIS 5e – Pre-Ind.)
Figure 3.13: Simulated 2-m summer and July MIS 5e interglacial warming comparisons with Modern (control).  
G) Mean summer temperature differences (MIS 5e – Modern) and 
H) Mean July temperature differences (MIS 5e – Modern).
Figure 3.14: Simulated PANN and insolation forcing (relative to Modern) averaged over latitude for MIS 5e. E) MIS annual precipitation and F) MIS 5e insolation anomalies (MIS 5e – Modern orbit config.)
**Figure 3.15: BIOME4 simulated vegetation for MIS 5e.** Red star denotes the location of Lake El’gygytgyn. Like the other interglacials, simulated vegetation correlates well with Lake El’gygytgyn multiproxy analysis.
3.2.3 **Marine Isotope Stage 11c (409 kyr)**

Marine Isotope Stage 11c is a long interglacial compared to the other interglacials in this study. We assume an ice-free Greenland in our MIS 11c simulations, with the ice sheet removed and replace with isostatically equilibrated (ice-free) land elevations. Additional experiments involving sea-ice extent will also be mentioned with the results outlined.

3.2.3.1 **Temperature**

Contribution of summer insolation forcing during this period ranges from +45 – 55 Wm$^{-2}$ (fig 3.17 F) allowing temperatures over the lake region during July (month of maximum insolation) to increase about +0.5 °C (fig 3.16 D) relative to modern and +2.2 °C (fig 3.15 B) relative to pre-industrial. In general, mean annual summer temperatures over the circum-Arctic are not much different than modern summer temperatures, with the exception of interior Siberia where warming relative to modern is about +3 – 4 °C.

In similar simulations performed with a modern Greenland Ice Sheet (GIS), temperature difference with and without a modern GIS was negligible, as the loss of the ice sheet only created July warming of ~0.3 °C around the lake. Warming, albeit slight, was present when comparing geopotential height anomalies around the lake. Anomalies of +4 – 10 meters indicate warming of the column of air above the lake, with negative height anomalies to the west of the lake.

The warmer climate across the Arctic and reduced GIS was thought to have increased sea levels by as much as >11 meters (Raymo & Mitrovica, 2012) with little sea
ice extent. In order to test high sea levels and an ice-free Arctic Ocean around Lake El’gygytgyn, increased subsurface heat flux convergence from 2 Wm⁻² to 10 Wm⁻² was initiated. The resulting reductions in sea ice and warmer Arctic SST’s produced negligible warming in the Beringian interior around the lake (< 0.7 °C). Interestingly, boreal forest biome forcing on surface temperatures was quantified around the lake region presenting a net cooling of -2 °C rather than warming.

3.2.3.2 Precipitation

Precipitation amounts at the lake during MIS11GIS are very similar to modern precipitation amounts of 475 mm year⁻¹ (fig 3.17 E). Also, MIS11NG exhibits exact precipitation amounts as our pre-industrial control run (~438 mm year⁻¹). Rainfall conditions directly in the Arctic Ocean basin are very dry, ~200 mm year⁻¹, which is expected based on Arctic climatology for the region. On the contrary, simulations of MIS11NG show lessened precipitation amounts of ~37 mm year⁻¹ relative to MIS11GIS however, runs with increased heat flux balanced out the loss of precipitation and demonstrated values exactly matching rainfall rates of modern control values (~478 mm year⁻¹).

3.2.3.3 Vegetation Distribution

The Lake El’gygytgyn region during MIS 11c is on the border of evergreen taiga and shrub tundra biomes (fig 3.18 G). Most of interior Siberia is deciduous forest and temperate grassland, similar to MIS 5e and 1. Interior Alaska and Yukon are mostly evergreen taiga and some deciduous forest toward the northern shore of Alaska, with
incorporated sporadic shrub tundra mixed in. With the loss of the GIS, Greenland is now predominantly shrub tundra with dwarf shrub tundra along the northern shore.

Vegetation limits, such as tree lines, are slightly changed during our simulations with increased heat flux and a warmer, open Arctic Ocean. Evergreen forests around the lake region and in Alaska extend poleward toward the coast, and deciduous forest is replaced by shrub tundra on the northern coast of Alaska (fig 3.18 H). Evergreen forest in the Yukon continues to be dominant with an eastward migration of the tree line taking over some grassland, as Greenland remains unchanged.
Figure 3.16: Simulations of MIS11c 2-m summer and July temperature comparisons to Pre-Industrial (control). A) Mean summer temperature anomalies (MIS11 – Pre-Ind.) and B) Mean July temperature anomalies (MIS11 – Pre-Ind.). The model still fails to explain all of the warmth during this interglacial.
Figure 3.17: Simulations of MIS11c 2-m summer and July temperature comparisons to Modern (control). A) Mean summer temperature anomalies (MIS11 – Mod.) and B) Mean July temperature anomalies (MIS11 – Mod.)
Figure 3.18: MIS11c mean annual precipitation (PANN) and solar insolation anomalies. E) Mean annual precipitation in the Arctic and F) Latitudinal mean annual solar insolation anomalies at top of the atmosphere relative to Modern orbit (MIS11 – Mod.)
Figure 3.19: BIOME4 simulated MIS 11 vegetation. G) Regular (no heat flux increase) vegetation around the lake and H) Vegetation (with heat flux increase) around the lake. Similarly, vegetation correlates well with Lake El’gygytgyn pollen proxy assemblages.
3.2.4 Marine Isotope Stage 31 (1072 kyr)

3.2.4.1 Temperature

A very warm orbit with high obliquity, eccentricity and precession aligning perihelion with boreal summer allows insolation anomalies to be $> 50 \text{ Wm}^{-2}$ at the surface and $+60 - 80 \text{ Wm}^{-2}$ at the top of the atmosphere (fig 3.21 F). Average summer temperatures around the lake are about $+1.6 \degree\text{C}$ (fig 3.20 C) warmer than modern and $+3.6 \degree\text{C}$ (fig 3.19 A) warmer than pre-industrial. CO₂ forcing contributions of $+0.80 \text{ Wm}^{-2}$ relative to pre-industrial values, permit July temperatures to exceed $+5$ and $+3.5 \degree\text{C}$ warmer than pre-industrial and modern temperatures, respectively (fig 3.19 B; 3.20 D). Most summer warming is seen over Greenland and interior Siberia with temperatures over an ice-free Greenland of $+15 - 17 \degree\text{C}$ and interior Siberia, with temperatures $+6 - 8 \degree\text{C}$ warmer relative to pre-industrial and modern temperatures. July average temperatures are in similar agreement relative to modern and pre-industrial control mean summer temperatures with overall $>15 \degree\text{C}$ warming over Greenland during this period.

3.2.4.2 Precipitation

Overall precipitation in the Arctic during interglacial 31 is $\sim 438 \text{ mm year}^{-1}$, similar to that of interglacial 11c (fig 3.21 E). However, summer precipitation is similar to modern values; about $65 \text{ mm month}^{-1}$ indicating more water vapor in the air possibly correlated with increased temperatures.
3.2.4.3 Vegetation Distribution

Vegetation distribution is similar to most of the interglacials described here. Most of the Alaskan interior is dominated by evergreen taiga forest with only a few areas of shrub tundra on the coasts. Lake El’gygytgyn is dominated by deciduous taiga with evergreen dominating toward the eastern coast (fig 3.21). Most of interior Siberia shifted from once being predominantly deciduous forest to now being only half deciduous forest and an expanding area of temperate grasslands. Without a GIS dominating interior Greenland, the landscape has shifted from tundra in MIS 11c to mostly evergreen forest. Interior Yukon remains the similar to other interglacials with a mix of temperate grassland and evergreen forests.
Figure 3.20: Simulations of MIS31 2-m summer and July temperature comparisons to Pre-Industrial (control). A) Mean summer temperature anomalies (MIS31 – Pre-Ind.) and B) Mean July temperature anomalies (MIS31 – Pre-Ind.)
Figure 3.21: Simulations of MIS31 2-m summer and July temperature comparisons to Modern (control). C) Mean summer temperature anomalies (MIS31 – Mod.) and D) Mean July temperature anomalies (MIS31 – Mod.)
Figure 3.22: Simulated MIS31 mean annual precipitation (PANN) and solar insolation anomalies from Modern (control). E) Mean annual precipitation in the Arctic and F) Latitudinal mean annual solar insolation anomalies at top of the atmosphere relative to Modern orbit (MIS31 – Mod.)
Figure 3.23: BIOME4 simulated vegetation for MIS31. Red star denotes location of Lake El’gygytgyn.
CHAPTER 4
DISCUSSION

The exceptionally warm periods of Marine Isotope Stage(s) 1, 5e, 11c and 31 show significant, but similar changes in the Arctic, especially around Lake El’gygytgyn. Temperature reconstructions during the Holocene Thermal Maximum (9 kyr) indicate +1.6 (±0.8) °C warming in the western Arctic (Brigham-Grette et al., 2003) with an overall warming of 1.7 (±0.8) °C in the circum-Arctic (Miller et al., 2010), relative to modern temperatures. Though our model does not fully account for all the warming relative to modern temperatures during this period, it does reflect the important warming in the western Arctic that is similar to the aforementioned study. With the decrease in moisture in the Arctic and low CO₂, deciduous and evergreen forests dominate the Arctic landscape with tree species such as Alnus, Betula (nut bearing trees and fruits), Poaceae (grasses) and some birch and alder. MIS 1 demonstrates significant changes in Arctic climate and vegetation showing how sensitive the Arctic is to Milankovitch forcing.

Marine Isotope Stage 5e produced the greatest summer warming amongst all four interglacials studied here. Comparisons with pre-industrial control runs show that differences in MTWM maxima at Lake El’gygytgyn during MIS 1 and 5e (+2.1 and +4.2 °C) are similar range of MIS 11c and 31 (+2.2 and +3.5 °C). Furthermore, similar simulated temperature differences have been seen in studies using a model of intermediate complexity and had shown that a very warm orbit of high obliquity, eccentricity and precession aligning perihelion with boreal summer will give way to maximum insolation forcing playing a dominant role on tree fraction, precipitation,
temperature and sea-ice (Yin & Berger, 2011). Temperature reconstructions for MIS 5e thermal maximum show variable temperature reconstructions indicating +5 (±1) °C across the entire arctic, with smaller anomalies reconstructed for the Pacific sector (Miller et al., 2010). Powerful insolation forcing at these latitudes permits July maximum temperatures to exceed both pre-industrial and modern temperatures by at least >3 °C which is in agreement with the previous study. The 2 – 4 °C warming in Siberia and western Beringia in our results has been shown by simulations with a model without vegetation feedbacks and has been linked to strong summer insolation anomalies (Otto-Bliesner, 2006). Anomalous insolation forcing was shown between 130 and 127 kyr during the summer season, the maxima at which our GCM was simulated. Moreover, the exceptional summer warming compared to other interglacials was thought to have caused a reduction in the Greenland Ice Sheet adding 1.6 to 2.2 m of equivalent sea level rise (Colville et al., 2011). A more recent study conducted by the North Greenland Eemian Ice Drilling Project (NEEM) confirmed that the thickness of the Northwest sector of the GIS decreased by 400 ± 250 meters reaching surface elevations of 130 ± 300 meters lower than present (Dahl-Jensen et al., 2013). This indicates that our simulations of MIS 5e with a near-modern GIS are a good approximation for this period. Increased warmth allows almost a full replacement of shrub tundra with deciduous forest in and around the lake region. Pollen analysis during this period show tree species of birch, alder, pine and spruce (Melles et al., 2012). However, multiproxy studies of MIS 5e show a change in MTWM of only +2 °C warming at the lake compared to modern temperatures (Melles et al., 2012). I conclude that a warm summer orbit with only moderate GHG concentrations does account for exceptionally warm temperatures in Beringia however, the particularly
muted response in the Lake El’gytgyn proxy record to summer insolation forcing cannot be fully explained (fig 4.1 I).

Simulations of 11c exhibit another very warm interglacial in the Arctic around the lake with MTWM maxima approaching +2.2 °C warmer than pre-industrial temperatures. Similarly to MIS 5e and 1, peak warmth coincides with perihelion during boreal summer however, a low eccentricity and obliquity attenuates the effects of precession relative to 5e and 1, making summer less intense, although longer in duration. This noticeable warmth is an obvious outcome of low to non-existent snow-ice albedo effect contributing to extreme warmth. Under the assumption sea level has risen due to ice sheet melt, increasing heat flux convergence under sea ice in the Arctic Ocean from 2 to 10 W m⁻² allowed us to test the hypothesis whether lessened sea ice and increased SSTs warmed the lake region by using a simplistic sensitivity test based on a modest calculation of increased, fourfold throughflow through the Bering Strait. Sea ice fractions during this experiment showed a 25 – 50 % decrease in summer Arctic sea-ice extent and increased summer ocean temperatures (fig 4.1). However, the effects of increased open water on Lake E temperatures were negligible, and only warmed the lake region and interior Beringia by +0.7 °C (fig 4.1 I, green dot).

Vegetation simulations do not pick up on possible forest biome in southern Greenland. However, the resolution of the vegetation component in our model may not be adequate enough to show extremely small-scale vegetation changes. Strong northern latitude interglacial forcing on terrestrial biome distribution is evident in our simulation by a poleward advance of evergreen needle-leaf forest during the interglacial around the lake which is in agreement with palynological analysis of tree species in the lake area
Analysis suggests forest-tundra and northern larch taiga environments with dark coniferous forests dominant of spruce, pine, birch, alder and larch controlled the lake region (Melles et al., 2012). Enhanced solar anomalies drove interior locales to warm allowing boreal forests to thrive. Surface warming as a result of increased low albedo needle-leaf forests accounts for some of the warming seen at the surface during this period. However, isolated forcing feedback of increased evergreen, terrestrial forest provides a net cooling effect during the summers and slight net warming effect during early fall (Sep. – Nov.; +0.3 °C).

A deglaciated Greenland has been shown to have regional effects on SSTs and sea-ice conditions, however warming of the circum-Arctic has been shown to be minimal (Koenig et al., 2012). This was demonstrated in our simulations by isolating the effects associated with the loss of the GIS leading to warming around the lake of only +0.3 °C. Analysis of 500 hPa geopotential height anomalies exhibit ridging (positive height anomalies of > 10 m) and troughing (negative height anomalies) to the west of the lake, indicating a slight change in the large-scale, synoptic planetary wave patterns over Beringia. Over the lake, positive height anomalies are also present indicating slightly warmer conditions and a slight shift eastward of an atmospheric ridge that may have been set up further west of the lake. The ridging evident in these simulations may also be related to a decrease in precipitation with the loss of the GIS. Extended high pressure over the lake associated with ridging would create somewhat drier conditions for the region. Generally, the exceptional warmth during MIS 11c would have melted the GIS affecting ocean overturning and creating a net cooling effect on the Northern Hemisphere. Unfortunately, the GCM spatial resolution is too coarse and cannot fully
reproduce all warming in the proxy records, making it difficult to explain the exceptional warmth during MIS 11c.

Figure 4.1: Enhanced heat flux (+8 Wm$^{-2}$) simulations of summer sea surface temperature and sea ice anomalies relative to default heat flux simulations. A) Summer sea surface temperature change relative to default heat flux simulation (T °C) and B) Summer sea ice fraction anomalies relative to default heat flux simulation (%). With the +8 Wm$^{-2}$ of sub-surface heat flux convergence, Arctic Ocean temperatures rose >0.5 °C and sea ice fraction decreased 25-50 % in most areas.
Elevated GHG concentrations and a very warm orbit with a large precession can explain much of the exceptional warmth during MIS 31. Such large insolation anomalies at high latitudes allow thick boreal forests, such as needle-leaf and deciduous forests to grow. Average summer temperatures are about 12 °C or +2 °C warmer than modern summer temperatures around the lake. Biome model simulations derived from pollen analysis inside the lake core show maxima of trees and shrubs during peak northern hemisphere insolation of MIS 31 at 1072 kyr. Our BIOME4 model simulations also show similar results around the lake region with increased boreal forests and less tundra and small dwarf shrubs. The snow-albedo effect combined with thick low albedo, forest cover allows temperatures to increase in the Arctic during MIS 31. Peak precipitation rates derived from proxy analysis indicate about 600 mm year$^{-1}$, or about 350 mm year$^{-1}$ more precipitation than modern model simulations (Melles et al., 2012). GCM results indicate ~490 mm year$^{-1}$, the most annual precipitation out of all four interglacials modeled. Although modeling studies do not fully simulate the enhanced precipitation indicated in the proxy record, relative increase in precipitation is evident in the model and proxy records. Extraordinary warmth during MIS 31 correlates well with a diminished West Antarctic Ice Sheet (WAIS) (Pollard & DeConto, 2009) implying strong intrahemispheric coupling that can be related to reductions in Antarctic Bottom Water (AABW) formation during times of ice-shelf retreat and increased fresh water input into the Southern Ocean (Foldvik, 2004). Similar ice sheet collapses have been seen in other interglacials and could possibly be an explanation to the warmth in MIS 11c.
The described characteristics of facies C are most pronounced for MIS 11c, 31, 49, 55, 77, 87, 91, and 93 (red bars in Fig. 3), suggesting that these interglacials represent unusual “super interglacials” in the Arctic throughout the Quaternary. The exceptional character of these interglacial conditions becomes evident based on a comparison of MIS 1 and 5e (facies B) with MIS 11c and 31 (super interglacials of facies C), using additional biological proxies and pollen-based climate reconstructions (Fig. 3, I to L).

Sediments formed in Lake El’gygytgyn during MIS 1 and 5e have Si/Ti ratios only slightly higher than those formed during glacial and stadial conditions of MIS 2, 5d, and 6 (Fig. 3K). Pollen data show distinct increases in tree and shrub pollen percentages compared with spruce pollen content. Simulated July surface air temperatures (red and green dots) and precipitation (blue dots) at the location of the lake are shown for comparison. The location of the dots relative to the X-axis corresponds with the GHG and orbital forcing used in each interglacial simulation. The green dot indicates the results derived with a deglaciated Greenland and increased heat flux under Arctic Ocean sea ice by 8 Wm$^{-2}$. (Figure from (Melles et al., 2012)).
CHAPTER 5
ARCTIC SENSITIVITY TO GLACIAL BOUNDARY CONDITIONS

5.1 Introduction

An additional sensitivity test of Lake El’gygytgyn to changing boundary conditions associated with the buildup of major northern hemisphere ice sheets was also simulated and related to pollen analysis at ~2.7 Ma in the lake core. Such a substantial cooling in the Arctic has been demonstrated to coincide with a dramatic decrease in PANN values around the lake (Brigham-Grette et al., 2013). Climate model simulations were run with 300 ppm of $pCO_2$ and a cold, boreal summer orbit, like that of 116 kyr (Brigham-Grette et al., 2013). The simulations represent conditions similar to the late Pliocene, with an orbit favorable for the growth of major Northern Hemisphere ice sheets.

5.2 Method and Experiment Set-up

Two simulations were run using the GCM described earlier in this thesis (GENESIS GCM, v. 3.0) with (3HL116K) and without (3NG116K) Northern Hemisphere ice sheets. In both cases, the GCM was run to equilibrium with averages calculated from the last 10 years of the model’s history files. The first simulation used ice-free Northern Hemispheric climate conditions, while the second simulation adds the Greenland, Laurentide and Fennoscandian ice sheets, based on the LGM ice volume from ICE 4G (Peltier, 1994; Brigham-Grette et al., 2013) including a decreased sea level. This
simple sensitivity test is used to show the effect of large Northern Hemisphere ice sheets on Arctic climate.

<table>
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<th>CH$_4$ (ppbv)</th>
<th>N$_2$O (ppbv)</th>
<th>Eccentricity</th>
<th>Obliquity</th>
<th>Precession</th>
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<td>300</td>
<td>800</td>
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<td>3HL116K</td>
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<td>0.043988</td>
<td>22.52</td>
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**Table 5.1: List of glacial simulations.** Run 3NG116K is the simulation without Northern Hemispheric ice sheets and 3HL116K has all major Northern Hemispheric ice sheets. Obliquity and Precession are in degrees (°) and GHG concentrations are labeled.

5.3 Results

5.3.1 Temperature

Mean Temperature of the Coldest Month (MTCM; Jan.) around Lake El’gygytgyn was simulated to be -40 °C with July temperatures about 3 °C (-5 °C relative to modern temperatures). These temperatures compare favorably with proxy reconstructions after 2.7 Ma and pollen reconstructions of the cool periods between interglacials (Brigham-Grette et al., 2013; Melles et al., 2012). Mean annual temperatures in the circum-Arctic decrease 5 to 25 °C in response to the increase of large ice sheets relative to the experimental run without Northern Hemispheric ice sheets.

5.3.2 Precipitation

Preliminary GCM analysis of mean annual precipitation (PANN) shows that most of the circum-Arctic becomes very arid with more than 150 mm year$^{-1}$ decrease in
precipitation, especially around the Arctic basin and parts of northern Beringia. This aridification is also consistent with drying seen in Melles, Brigham-Grette et al., 2012 during Pleistocene glacial periods. Aridification analysis, while not definitive, suggest that large Northern Hemispheric ice sheets initiate changes in the Arctic hydrologic cycle.

5.4 Discussion

Arctic aridification and temperature change can be linked to mechanical atmospheric forcing associated with large northern hemisphere ice sheets. Exceptionally large temperature decreases are thought to be associated with albedo-enhanced cooling from large ice sheets reflecting solar radiation back to the atmosphere. Likewise, enhanced cooling in the Arctic and expanded sea-ice cover contributed to circum-Arctic aridification (> 150 mm year$^{-1}$).

Comparable studies (Bromwich et al., 2004) using regional climate models to quantify mechanical forcing of large northern hemisphere ice sheets show important effects on mid-tropospheric westerly flow. The presence of a very large Laurentide ice sheets splits the jet stream into two branches: a northern most, polar jet and a southern branch (fig 5.1 A, B, C). Due to this split flow around the ice sheet, during January, surface cyclones tend to flow along the periphery of the Laurentide Ice sheet due to a very strong high-pressure system that forms over North America (fig 5.3, A). Due to a strong mid-level trough that forms on the south coast of western Beringia (fig 5.3, B), storms are frequent along the southern coast of Alaska and Beringia (Bromwich et al., 2004). During the summer (July), the jet stream is positioned directly over the ice sheet
allowing increased frequency of surface cyclones to migrate directly over the ice sheet dropping 42% of annual precipitation (Bromwich et al., 2004). This can be attributed to a large trough centered over southwestern North America (fig 5.3, C), allowing the storm track to push storms further south over this region and North America. Additionally, Beringia is encased in a very strong high-pressure system (fig 5.3, D), presumably limiting precipitation in Beringia and at the lake. This strong high-pressure system seems to be related to a considerable strengthening of the Siberian high. It is important to note that the strengthening of the Siberian high is seen when we have large Northern Hemisphere ice sheets. It can be concluded that mechanical forcing of northern hemisphere ice sheets led to aridification of the Arctic due to changes in the dominant storm track patterns. Even though these results are not definitive, the results suggest that the presence of large Northern Hemisphere ice sheets contributed to changes in synoptic weather patterns leading to aridification of Lake El’gygytgyn and the change of boreal/evergreen forest around the lake to shrub tundra, lichen and mosses.
Figure 5.1: Seasonal distribution of the 500hPa winds and geopotential heights over North America and the high latitudes with Laurentide Ice Sheet. A) Average annual positions of jet stream, B) Mean winter position of jet stream and C) Mean summer position of jet stream. Split flow is more evident in the annual and summer mean position of the jet. They correlate well.
Figure 5.2: Climate model simulations using 300 ppm CO$_2$ and a cold boreal summer orbit, like that at 116 kyr to isolation the effects of large Northern Hemispheric ice sheets on circum-Arctic climate and aridification. A) Simulated drying (difference) in mm year$^{-1}$ resulting from large northern hemisphere ice sheets, B) Mean Temperature of the Coldest Month (MTCM, Jan.) and C) Mean annual 2-m temperature difference between simulation with NH ice sheets and Control (no NH ice sheets).
Figure 5.3: Climate model simulations of mean sea level pressure during January and July. Both simulations have major northern hemisphere ice sheets. A) January mean sea level pressure (MSLP) over the continental United States (CONUS), B) MSLP over the Arctic Basin, C) July MSLP over CONUS and D) July MSLP over the Arctic Basin. Warm colors represent high MSLP and cool colors, low MSLP.
CHAPTER 6

CONCLUSION

Lake El’gygytgyn provides a high-resolution terrestrial proxy record of climate variability in the Arctic. Climate modeling studies described here show that the Arctic summers were significantly warmer during several Pleistocene interglacials by as much as $+2 \, ^\circ\text{C}$ in MIS 1 and 11c and by as much as $+4 \, ^\circ\text{C}$ during MIS 5e and 31 relative to pre-industrial. It can be inferred that the onset of this warming was caused by a combination of elevated GHGs and warm boreal summer orbits that lead the way for the super-interglacials in the Arctic. Other factors such as decreased productions of Antarctic Bottom Water (AABW) and long duration of open Arctic Ocean due to changes in sea-ice may have also contributed to exceptional warmth during this period. Thorough testing of these ideas will require additional simulations with atmosphere-ocean modeling, changes in glacial and post-glacial eustatic sea-levels, changes in continentality, changes in sea-ice distributions and the addition of melt-water inputs into northern hemisphere oceans.

General results of the simulations show that all the interglacials experienced similar warming. Looking at greenhouse concentrations, one would suggest that similar concentrations could lead to similar warming. Likewise, eccentricities for most of the interglacials (control experiments, MIS 1 and 11) were very similar resulting in similar warming results. Both exceptions, MIS 5 and MIS 31, also show similar warming, but have higher eccentricities. The net effect of an orbit, such as that of MIS 5e and 31, which created the greatest summer warmth out of all interglacials, produces high-
intensity summer insolation of >50 Wm$^{-2}$ at the surface and $\sim -0.98 - 1.89$ Wm$^{-2}$ of greenhouse gas forcing, relative to modern values (Melles et al., 2012). MIS 1 is an exception with lower CO$_2$ around the time of peak Holocene warmth producing -0.44 Wm$^{-2}$ less radiative forcing relative to pre-industrial levels (Melles et al., 2012).

Extreme warmth and changes in greenhouse gases shifted vegetation from mostly tundra with small shrubs as we see the Arctic today to thick, lush evergreen and boreal forest. Due to the extreme warmth, wetter conditions prevailed during the super-interglacials allowing biomes to thrive and increase their maximum extent poleward while making each interglacial unique based upon the different tree and shrubs species that dominant during each specific period. Ice sheets in the Arctic, such as the Greenland Ice Sheet, were significantly reduced during some interglacials, allowing summer temperatures to increase almost 2 to 5 °C warmer than present. The observed response of the region’s climate and terrestrial vegetation distribution to super-interglacial forcing is still not fully understood and creates a challenge for climate modeling and the study of Arctic amplification. Such examples are the extreme warmth at MIS 11c despite lower than modern GHG concentrations and the muted response in proxy records for MIS 5e, despite extreme summer insolation intensity. Additionally, modeling studies showed overall drier conditions in the earlier interglacials (11c and 31) relative to pollen analysis. The significant warming in the circum-Arctic can be linked to major deglaciation events in Antarctica, demonstrating possible intrahemispheric linkages between the Arctic and Antarctic climate on glacial-interglacial timescales.

Large northern hemisphere ice sheets during major glaciation events can be linked to Arctic aridification and extremely cold annual temperatures. The combination of
increased Arctic sea ice and increased surface albedo allows the Arctic to significantly cool and dry out during these events. This is demonstrated in the Lake El’gygytgyn core by multiproxy analyses and a transition to shrub vegetation due to the lack of precipitation. The climate modeling showed here suggests extreme Arctic aridification after 2.7 Ma was a consequence of the episodic expansion of ice sheets, which affected dominant atmospheric pressure patterns, the storm track and a general southward shift of precipitation in the Beringian sector of the Arctic.
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