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Eccentricity Modulation of Precessional Variation in the Earth’s Climate Response to Astronomical Forcing: A Solution to the 41-kyr Mystery

Rajarshi Roychowdhury

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ECCENTRICITY MODULATION OF PRECESSIONAL VARIATION IN THE EARTH’S CLIMATE RESPONSE TO ASTRONOMICAL FORCING: A SOLUTION TO THE 41-KYR MYSTERY

A Dissertation Presented

by

RAJARSHI ROYCHOWDHURY

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

DOCTOR OF PHILOSOPHY

February 2018

Department of Geosciences
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ABSTRACT

ECCENTRICITY MODULATION OF PRECESSIONAL VARIATION IN THE EARTH’S CLIMATE RESPONSE TO ASTRONOMICAL FORCING: A SOLUTION TO THE 41-KYR MYSTERY

FEBRUARY 2018

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The 41,000-year variability of Earth’s glacial cycles during the late Pliocene-early Pleistocene is usually attributed to variations in Earth’s obliquity (axial tilt). However, a satisfactory explanation for the lack of precessional variation in marine d$^{18}$O records, a proxy for ocean temperature and ice-volume, remains contested. Here, a physically based climate model is used to show that the climatic effect of precession is muted in global isotope records due to two different mechanisms, with each dominating as a function of eccentricity. At low eccentricities (e0.019), the time-integrated summer insolation and number of positive degree-days impacting ice sheets varies at precessional periods, but the variation is out-of-phase between the Northern and Southern Hemispheres. Each mechanism dominates at different times, leading to a net attenuation of precessional variability in globally integrated proxy records of ice volume.
Recently, several interglacials (MIS 9, 11, 31, 49, 55, 77, 87 and 91) have been identified as warmer than others and have been termed “Super-interglacials”. It has been shown that the warmest of these interglacials follow exceptionally low eccentricity periods, with a lag of ~50kyr. The explanation proposed for this low eccentricity preconditioning of the super interglacials is directly linked to the fact that the polar ice sheets respond differently to precessional changes at different eccentricities, as described above. Using a series of GCM and ice-sheet model simulations covering MIS 11 and 31, it is shown that Southern Hemisphere ice-sheets respond to Northern Hemisphere insolation at lower eccentricities, switching to local Southern Hemisphere insolation at higher eccentricities. This switch from northern forcing to southern insolation forcing leads to Antarctica missing a beat in its glacial-interglacial cycles, as northern and southern insolation intensities vary out-of-phase at 23 ka precessional periods. Thus, depending on the orbital conditions, Antarctica either has an unusually long glacial or interglacial period following a low eccentricity orbit. In the latter case, the prolonged warm conditions in the Southern Hemisphere preconditions the Polar Regions to produce a large response during the unusually warm interglacials like MIS 11 or 31.
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CHAPTER 1

INTRODUCTION

1.1 Motivation

The global scientific community has accepted the current warming of the climate system unequivocally, and the changes in our climate system have been summarized in the latest IPCC Fifth Assessment Report (IPCC AR5, 2014) based on the reports of the three working groups of the Intergovernmental Panel on Climate Change (IPCC). The IPCC AR5 provides an integrated view of climate change around the globe, and notes that the recent anthropogenic emissions of greenhouse gases are the highest in history, leading to unprecedented human influence on Earth’s climate. Present concentrations of carbon dioxide, methane and nitrous oxide are highest in at least the last 800,000 years. The effects of greenhouse gases along with other anthropological factors have been detected throughout the climate system and have been conclusively linked as the dominant cause of the observed warming since the middle of the 20th century (IPCC, 2014b). Continued emission of greenhouse gases will lead to further warming and long-term changes in all climate components, thus escalating the possibility of severe and irredeemable impacts upon ecosystems and societies at large.

Today, scientists and academia rely on the predictive capabilities of numerous climate models to assess the likely warming scenarios in the future. To effectively provide robust predictions of the future, the climate models need to be validated and benchmarked with geological records of past climate. Understanding the evolution of the present warming in the context of past warm periods (interglacials) is important in evaluating natural climate variability, in order to differentiate between natural and anthropogenic forcings. The properties of the climate system that determine the response to external forcings (solar, volcanic and orbital) have to be analyzed and quantified in order to provide robust predictions of Earth’s climate response to future global warming,
and eventually isolate the effect of anthropogenic climate change from the natural variability (Berger, 1995; DeConto & Pollard, 2016; etc).

Paleoclimate data and modeling provide a window into the Earth’s response to these external forcings, as well as internal forcings (greenhouse gases). Paleoclimate studies help scientists understand the Earth’s climate system better, and facilitates the study of Earth system response at various time-scales, beyond the scope of the short instrumental records available (limited to few hundred centuries). Thus paleoclimate studies, including this thesis, aim to improve the understanding of the Earth’s climate system and predictive capabilities of climate models (GCMs, RCMS and coupled ice-sheet models), which are critical for robust predictions of the Earth’s response to future global warming.

1.2 Methods

The research presented here utilizes a physically based model to study the response of Earth’s climate system to variations in orbital forcing. All original data contributions in this dissertation come from ensembles of GCM (General Circulation Model) and ice-sheet model experiments. I used the current version (v.3) of the Global ENvironmental and Ecological Simulation of Interactive Systems (GENESIS) GCM, originally developed by the Interdisciplinary Climate Systems Section of the Climate and Global Dynamics Division at NCAR (Pollard & Thompson, 1995; Thompson & Pollard, 1997). The GENESIS GCM has been validated against modern climate and used extensively for paleoclimatic simulations (Koenig, DeConto, & Pollard, 2011). The model is unique, because it can be coupled to a dynamical ice sheet model. For the ice-sheet simulations, I used Pollard and DeConto’s 3D ice-sheet/ice shelf model (ISM) capable of being driven by the GENESIS climate model. The model is designed for long-term continental scale applications, and has been used in numerous paleoclimate studies (DeConto et al., 2012a, 2012b, etc).

1.3 Astronomical Forcing of Insolation received by the Earth
Insolation is the prime and most well defined factor for forcing Earth’s climate system over long periods of time. Insolation is defined as the rate at which direct solar radiation is incident upon a unit horizontal surface at any point on or above the surface of the Earth. Total Solar irradiance (TSI) is the measure of the solar power over all wavelengths per unit area at the top of the atmosphere. It is a measure of the electromagnetic energy incident on a surface perpendicular to the incoming radiation at the top of the Earth’s atmosphere, and thus may be referred to as “flux”. In order to study the effects of solar radiation on the Earth’s climate system, it is necessary to determine the amount of energy reaching the Earth’s atmosphere and surface. Thus, for climate modeling, computation of radiative fluxes at the top of the atmosphere is an important component of understanding the Earth’s climate response to insolation forcing.

The energy available at the top of the atmosphere is the fundamental measure of insolation forcing affecting the Earth’s climate. For given latitude ($\Phi$), assuming a perfectly transparent atmosphere and constant solar constant $S_0$, the energy available at the top of the atmosphere depends on the Earth’s orbital and rotational parameters, which are a function of the gravitational effects of the sun, the moon and the planets (Berger, 1978). These are (i) the eccentricity, $e$; (ii) the obliquity, $\varepsilon$ (tilt of the Earth’s rotational axis relative to a perpendicular through the plane of the ecliptic); (iii) the semi-major axis ($a$) of the Earth’s orbit around the sun and the longitude of the perihelion ($\varpi$) measured from the moving vernal equinox. The latter two form the “precessional” component of the Earth’s orbital forcing.

Among the Earth’s orbital parameters, two of these have the strongest impact on the insolation forcing at the top of the atmosphere. The precession of the equinoxes alters the distance between the Earth and sun at any given time of the year, thus directly impacting the amount of incoming solar radiation. The eccentricity, which determines the shape of the Earth’s orbit around the sun, essentially determines the amplitude of this precession cycle. Apart from precession, obliquity plays a dominant role in calculating the insolation by affecting the seasonal contrast and the latitudinal gradient of insolation. Berger, in 1978, provided trigonometric formulae that allowed the direct spectral analysis
and computation of long-term variations of the Earth’s orbital elements described above. For the climatic precession parameter, the main astronomical frequencies are 23 and 19 kyr. For obliquity, the corresponding main astronomical frequencies are 41 and 54 kyr, and for eccentricity, 400, 125, 100 and 95 kyr (Berger, 1977; Berger and Loutre, 1991).

Milankovitch was among the first to study insolation quantitatively. Milankovitch introduced the concept of caloric summer, defined as the half of the tropical year during which daily mean insolation are greater than all days of the other half (Milankovitch, 1941). Milankovitch defined the half-year caloric seasons instead of using the variable length of the astronomical seasons. In Milankovitch’s half year caloric season, obliquity is in-phase in both hemispheres with a maximum effect at the polar latitudes. Precession is out-of-phase in both hemispheres, with a maximum effect at the equatorial latitudes. Vernekar recomputed Milankovitch’s results on radiation chronology with improved calculations of the variations in the Earth’s orbital elements and a more recent estimate of the Solar constant (Vernekar, 1972). Berger also calculated the annual cycle of daily irradiation for each 10-degree latitude, for both calendar and solar dates using a more accurate astronomical solution (A. L. Berger, 1979). Accuracy and spectral characteristics of the calculated daily irradiation were checked and analyzed by Pestiaux and Berger (Berger et al., 1984). The diurnal cycle was calculated by Ohmura (Ohmura, Blatter, & Funk, 1984) and by Tricot and Berger (Tricot & Berger, 1988), which computed the daily irradiation at the Earth’s surface for a given atmosphere of reference. Others analyzed numerically the insolation values, such as representing the insolation time series as Fourier-Legendre expansions, and rediscovered partly the rules used in generating them (Berger et al., 1984; North et al., 1979; Taylor, 1984). Fourier representations of orbitally induced perturbations in insolation were computed to improve the understanding of how each of the orbital parameters affects insolation.

Berger (1993) showed that the spectrum of instantaneous insolation, or irradiance, is dominated by climatic precession (e sin ϖ or e cos ϖ) displaying mainly 23 and 19 kyr periods. The instantaneous insolation or solar radiation striking the surface (W) is given by \[ W = S \left( \frac{a}{r} \right)^2 \cos z \]; where S is the solar constant, ‘a’ is the semi-major axis of the
Earth’s orbit around the sun, and ‘z’ is the zenith angle. The solar constant is the amount of energy received at the top of the Earth’s atmosphere on a surface oriented perpendicular to the Sun’s rays (at the mean distance of the Earth from the Sun). From various satellite and spacecraft observations, the value of the solar constant is generally accepted to be 1368 W/m², averaged over the year. Zenith Angle is the angle from the zenith (point directly overhead) to the Sun's position in the sky, and it is dependent upon the latitude, the solar declination angle, and time of day. The equation for \( W \) can be analytically simplified to show that the spectrum of \( (a/r)^2 \), or the distance factor, is dominated by climatic precession signal. Meanwhile, the term \( \cos z \), or the inclination factor, is dominated by the obliquity signal. Therefore, for a fixed distance of the Earth from the sun, there is only an obliquity signal in in the insolation spectra through geological time. For a fixed zenith angle, there is only precession signal in the insolation spectra through geological time. If neither is fixed, for a given hour of the day, the instantaneous insolation is a function of both precession and obliquity, with their individual spectral amplitudes depending upon the latitude being studied and the time of the year given by the longitude.

Another metric used for studying astronomical forcing is daily irradiation, calculated by integrating the daily instantaneous irradiation over 24 hours of true solar time, \( t_s \). However, true solar time is not regular because of the elliptical shape of the Earth’s orbit, and Kepler’s second law of orbital motion. One way to counter this is to use a regular evolving time, or the mean solar time, which is related to the true solar time through the equation of time, provided in the Astronomical Ephemeris for each day (A. Berger, Loutre, & Tricot, 1993). Even though the true solar time and regular evolving time is not exactly the same, and the difference being insignificant, both may be used interchangeably for calculation of daily irradiation (i.e. \( dt \sim dt_s \)). Total daily irradiation varies primarily at precessional frequencies for all months and latitudes, with obliquity being more dominant at higher latitudes as compared to lower latitudes (A. Berger & Pestiaux, 1984).
Diurnal irradiation is a time integrated insolation metric, with the insolation integrated over a time period defined by two different zenith angles (zenith distances, $z_1$ and $z_2$). The time period being integrated is not constant, and depends on the combination of zenith angles ($z_1$, $z_2$) chosen and obliquity. For given latitude, the zenith distance may correspond to different hours of the day depending upon obliquity. This metric varies at precession, eccentricity and obliquity frequencies, with the amplitude of each depending on the latitude being studied and the time of the year. It should be noted that this metric is different from daily irradiation defined above, as the time over which it is integrated is not constant, but depends on obliquity itself.

The spectrum of instantaneous insolation (irradiance) at the equinoxes is dominated by 23 and 19 kyr periods, corresponding to precessional variations. Low amplitude variations at half precessional periods (11.5 and 9.5 kyr) are also displayed by the irradiance spectrum. Similarly, the daily irradiation and the diurnal irradiation at the equinoxes are also only a function of precession, as shown by Berger et al. (1993)

The spectrum of the instantaneous insolation (irradiance) at the solstices shows strong precession and obliquity components. Precession dominates at all latitudes of the summer hemisphere, with the obliquity signal increasing from the equator to the pole. In the winter hemisphere, precession is dominant at the lower latitudes, while obliquity dominates precession at the higher latitudes. The analytic form of the spectrum also confirms the same. For daily irradiation at the summer hemisphere, precession is the dominant forcing factor at all latitudes, and obliquity has a stronger effect at higher latitudes than lower latitudes. For daily irradiation at the winter hemisphere, obliquity dominates precession at higher latitudes, while precession plays a stronger role in the lower latitudes.

1.4 Historical Background of Glacial-Interglacial cycles

The growth and decline of the polar ice sheets has been a subject of research in Earth science since the 18th century, when Scottish naturalist James Hutton (1726-1797)
observed erratics (boulders believed to have been transported by glacial action) in Switzerland and proposed that alpine glaciers were more extensive in the past. In 1837, Louis Agassiz (1807-1873) proposed that geological deposits in Europe and North America were remnants of vast ice sheets that spilled from the mountains. Based on his field findings, he proposed that the Earth had been subject to a past ice age. In 1842, the first attempt to explain the ice ages using an astronomical connection was made by French scientist Joseph Adhemar, who proposed that the ice ages were caused by the 22,000-year precession of the equinoxes. Adhemar proposed that glaciation occurs during anomalously long winters, which happens when winter coincides with aphelion. Kepler’s second law of planetary motion states that the speed of a planet increases as it nears the sun and decreases as it recedes from the sun. Thus, when winter coincides with Aphelion, the Earth experiences a longer than usual winter.

Later, James Croll suggested that the glacial-interglacial cycles were a result of variation in the severity of winter due to changes in the orbit of the earth. During periods of high eccentricity, glaciation occurred when winters coincided with aphelion, as weaker insolation led to colder winters (Croll, 1875; Muller, 1997). This implied that during periods of higher eccentricity, ice ages occur on 22,000-year cycles, alternating between Northern and Southern Hemispheres. Croll was the first to identify the important role of surface feedback processes necessary for major climatic changes to result from minor insolation changes. The insolation controlled glaciation theory was further advanced by Milankovitch who proposed that glaciation occurs during periods of low obliquity and summer coinciding with aphelion (Milankovitch, 1941). Milankovitch argued that when there is less insolation during summer, snow and ice persist throughout the year, leading to the formation of ice sheets.

Project CLIMAP was the first to empirically test Milankovitch’s theory of orbital cycles. In 1976, James Hays, John Imbrie and Nicholas Shackleton came to the conclusion that in the past 500,000 years, the variation in global climate corresponds to obliquity and precessional changes (Hays, Imbrie, & Shackleton, 1976). They found that the Oxygen isotope ratios in deep-sea sediment cores, which were calibrated to the
recently developed geomagnetic scale, varied at the same frequency of the changes in Obliquity and Precession of the Earth’s orbit.

1.5 100 kyr cycles

During the past 800,000 years, ice sheets followed a cycle of approximately 100,000 years (Ghil, 1994; Imbrie et al., 1992). The ice sheets took about 90,000 years to grow and only 10,000 years to collapse. Hays et al (reference) linked these 100,000-year cycles to the 100,000-year cycle of the earth’s eccentricity. However, the earth’s eccentricity has only a weak forcing on the insolation intensity reaching the top of the Earth’s atmosphere. The mid-Pleistocene cyclicity of the glacial cycles is complex, and a wide range of hypotheses have been proposed to explain it (Ashkenazy & Tziperman, 2004; W. H. Berger, Yasuda, Bickert, Wefer, & Takayama, 1994; Clark, 1999; Ghil, 1994; Imbrie, J., Hays, J. D., Martinson, D. G., McIntyre, A., Mix, A. C., Morley, J. J., Pisias, N. G., Prell, W. L., and Shackleton, 1984; Laepple & Lohmann, 2009; Maasch & Saltzman, 1990; Paillard, 1998; Saltzman Barry & Alfonso Sutera, 1987; Shackleton, Berger, & Peltier, 1990; Tziperman & Gildor, 2003). Most of these hypotheses attribute the 100kyr cycles to non-linear response of the climate system to the forcing or internal oscillations of the climate system. It has been observed that variation in ice volume at precession and obliquity frequencies do exist, and they appear to be directly forced and coherent with northern summer insolation.

1.6 41-kyr Cycles – The Achilles Heel of Milankovitch’s Theory of Climate Change

Today, geologists generally accept Milankovitch’s theory of glacial-interglacial climate change, and there is tendency to correlate insolation variations at specific latitudes directly with geological proxies recording changes in the earth’s climate (Sugden et al., 2014). However, there is one aspect of these glacial-interglacial cycles, which cannot be answered by Milankovitch’s theory. Before the Mid-Pleistocene Transition around 800,000 years ago, the glacial cycles during the late Pliocene to early Pleistocene (~1-3 myr) had dominant 40-kyr frequencies. The primary frequency
associated with the benthic $\delta^{18}$O records from this period corresponds to variation in the obliquity phase. This raises a major contradiction to Milankovitch’s theory of orbital forcing, which predicts precession should be the strongest frequency in glacial-interglacial cycles. High latitude summer insolation is primarily modulated by changes in the Earth’s precession, and summer insolation has been observed to drive glacial cycles in both hemispheres. However the paleoclimate records show strongest spectral power at 41kyr, which corresponds to the cycle of changes in the Earth’s axial tilt, or obliquity (Huybers & Curry, 2006). The absence of the precession signal and the presence of a strong obliquity signal are surprising and unaccounted for. Computer models predict a strong precessional signal in the modeled ice volume, but have been unable to recreate glacial cycles with spectral characteristics of the paleo ice volume records.

Several theories have been proposed to answer this anomaly. It has been proposed that the obliquity driven variations in the insolation gradient between high and low latitudes controlled polar climate during the late Pliocene-early Pleistocene (Maureen E. Raymo & Nisancioglu, 2003). Another suggestion is that high latitude snowfall variability, snowmelt variability over Antarctica and hemispheric changes in net snowfall are dominated by changes in the Earth’s axial tilt, which contribute towards the strong influence of obliquity forcing on global benthic $\delta^{18}$O records (Lee & Poulsen, 2009). It has also been suggested that the early Pleistocene glacial cycles are nonlinear oscillations with periodicity close to 40ka, and that these become phase-locked to obliquity cycles at the same frequency (Gildor & Tziperman, 2000; Tziperman, Raymo, Huybers, & Wunsch, 2006). More recently, it has been proposed that positive surface albedo feedbacks between high-latitude insolation, ocean heat flux and sea-ice coverage, and boreal forest/tundra exchange increase the strength of obliquity forcing on global ice-volume records (Tabor, Poulsen, & Pollard, 2015).

A recent hypothesis suggests that ice-sheets are sensitive to insolation integrated over the duration of summer, instead of intensity of summer insolation (Huybers, 2006; Huybers & Tziperman, 2008). Annual ablation is empirically related to an integrated summer insolation metric, which is a function of solar radiation intensity and duration of
the summer melting season. Kepler’s second law states that Earth’s distance from the sun is inversely proportional to its angular velocity. Thus, a summer with weak insolation intensity (related to Earth’s distance from the sun) would have a longer duration (related to Earth’s angular velocity), while a stronger summer would correspond to a shorter duration. Therefore, the integrated summer insolation metric is insensitive to precessional changes, due to the opposing effects of precession on intensity and duration.

On the other hand, it has been suggested that the amount of melting an ice sheet undergoes is controlled by local summer insolation which is dominated by the 23-ky precession period at nearly all latitudes (M E Raymo, Lisiecki, & Nisancioglu, 2006). However, the earth’s orbital precession is out-of-phase between hemispheres, i.e. when Northern Hemisphere has strong local insolation; the Southern Hemisphere has weak local insolation and vice-versa. Thus the ice sheet growth in one hemisphere is accompanied by melting of ice sheets in the other hemisphere. As a result, the ice volume changes in each hemisphere cancel out in globally integrated proxies such as ocean δ¹⁸O or sea level curves.

In this dissertation, I reconcile the “41-kyr problem” using physically based climate and ice sheet models, focusing on the late Pliocene – early Pleistocene period, when this 41-kyr periodicity in the glacial-interglacial cycles were most pronounced. Both Huyber’s and Raymo’s hypotheses are tested using complex Earth system models. Utilizing various statistical methods to verify the results. I use ensembles of GCMs and Ice-Sheet Models to better understand the mechanisms underlying the glacial-interglacial cycles during the late Pliocene–early Pleistocene.

1.7 Dissertation Outline

The following outline describes the main components of this dissertation. The dissertation consists of four chapters (Chapter 2-5) following this introductory chapter (Chapter 1). The four main chapters of this dissertation have been adapted from independent manuscripts for submission to peer-reviewed science journals. These
chapters were written as separate scientific manuscripts with independent literature reviews, methods, results and discussion sections. Consequently, there is some repetition of content.

Chapter 2 – Eccentricity modulation of obliquity-paced cyclicity in Plio-Pleistocene ice volume aims to address the “41-kyr world” anomaly, one of the longest standing questions in Earth sciences. Geological records demonstrate that during the Late Pliocene-Early Pleistocene (~1-3myr), there is an absence of strong 20-kyr precession signals in the proxy data of oxygen isotopes that record long-term variations in global ice volumes, contradictory to classical Milankovitch theory (Maureen E Raymo & Huybers, 2008). Making use of physically based model, I show that climate variations during intervals dominated by 40-kyr cyclicity are indirectly controlled by eccentricity, which modulates the phasing of precessional response between the Northern and Southern Hemispheres.

Chapter 3 – Interhemispheric Effect of Global Geography on Climate Response to Orbital Forcing focuses on improving our understanding of the bias in climate response of the Earth due to unequal distribution of land in the Northern versus Southern Hemispheres. Here, I investigate the asymmetric climate sensitivity to orbital forcing, with the aim to quantify the Land Asymmetry Effect (LAE) using a physically based model. The results of this research provide a baseline for interpreting contemporaneous proxy climate data spanning a broad range of latitudes and individual time-continuous records exhibiting orbital cyclicity.

Chapter 4 – Orbital Signature of “Super-Interglacials” from the Arctic Lake El’gygytgyn record. While it is generally accepted that glacial-interglacial variability is orbitally paced, the extent, duration and phasing of the climate cycles are complex, and difficult to constrain using simplified, theoretical methods (Tzedakis et al., 2009). The Lake El’gygytgyn record from northeastern Siberia has identified several instances of extreme warmth during the Plio-Pleistocene (“super-interglacials”). Each of the “super-interglacials” identified in paleoclimate archives, like the Arctic Lake-E record (Melles et
al., 2011) remains unique in terms of intensity, duration, orbital forcing and internal variability. This chapter attempts to place the orbital signature of the super-interglacials into a general mechanistic theory accounting for the pronounced and anomalous warming observed during these super-interglacials.
CHAPTER 2

ECCENTRICITY MODULATION OF OBLIQUITY-PACED CYCLICITY IN PLIO-PLEISTOCENE ICE VOLUME

2.1 Abstract

The 41,000-year variability of Earth’s glacial cycles during the late Pliocene-early Pleistocene is usually attributed to variations in Earth’s obliquity (axial tilt) (Imbrie, Berger, & Shackleton, 1993; Maureen E. Raymo & Nisancioglu, 2003). However, a satisfactory explanation for the lack of precessional variation in marine $\delta^{18}O$ records, a proxy for ocean temperature and ice-volume, remains contested (Maureen E Raymo & Huybers, 2008). Here, we use a physically based climate model to show that the climatic effect of precession is muted in global isotope records due to two different mechanisms, with each dominating as a function of eccentricity. At low eccentricities ($e<0.019$), the small response of summer temperatures to precessional variations in the intensity of summer insolation is balanced by changes in the duration of summer. At higher eccentricities ($e>0.019$), the time-integrated summer insolation and number of positive degree-days impacting ice sheets varies at precessional periods, but the variation is out-of-phase between the Northern and Southern Hemispheres. Each mechanism dominates at different times, leading to a net attenuation of precessional variability in globally integrated proxy records of ice volume.

2.2 Introduction

Alternating Northern Hemispheric glacial and interglacial cycles have dominated the Earth’s long-term climate variability for the past 3 million years (Emiliani & Geiss, 1959). Since the middle of the Nineteenth century, several theories have been proposed connecting these cycles with variations in the Earth’s orbital configuration. Milankovitch (Milankovitch, 1941) was among the first to provide a comprehensive theory associating...
the cyclic changes in Earth’s climate to variations in eccentricity, obliquity and precession; and his ideas were empirically demonstrated by variations in oxygen isotope ratios in deep-sea sediment cores, showing variations at the same frequency as changes in obliquity and precession of the Earth’s orbit (Hays, Imbrie, & Shackleton, 1976).

Before the Mid-Pleistocene Transition around 800,000 years ago, late Pliocene–early Pleistocene (~1-3Ma) glacial cycles recognized in benthic δ¹⁸O records were dominated by ~41kyr frequencies, corresponding to variations in orbital obliquity (axial tilt) (Huybers & Curry, 2006; Imbrie et al., 1993; Lisiecki & Raymo, 2005). This contradicts Milankovitch’s theory of orbital forcing, which states that glacial cycles are primarily forced by orbitally induced summer insolation changes over high northern latitudes. High latitude summer insolation is primarily modulated by changes in Earth’s precession, which determines the seasonal timing of perihelion and aphelion. An adequate explanation for the absence of a strong precession signal during this interval continues to be hotly debated (Gildor & Tziperman, 2000; Lee & Poulsen, 2009; Maureen E. Raymo & Nisancioglu, 2003; Maureen E Raymo & Huybers, 2008; Tabor, Poulsen, & Pollard, 2015; Tziperman, Raymo, Huybers, & Wunsch, 2006).

2.3 Two competing theories which explain the “41-kyr” anomaly

A recent hypothesis posed by Peter Huybers (Huybers, 2006; Huybers & Tziperman, 2008) suggests ice-sheets are sensitive to insolation integrated over the duration of summer, instead of summer insolation intensity. The integrated summer insolation is affected not only by intensity of summer insolation (which is controlled by precession), but also by duration of the summer. In this work, annual ablation is empirically related to an integrated summer energy metric, which is a function of insolation intensity and duration of the summer melting season. Summer insolation and summer duration are both primarily controlled by earth’s precession of its equinoxes. By Kepler’s law, a summer occurring at aphelion with weak insolation intensity (related to Earth-sun distance) would have a longer duration summer (related to Earth’s angular velocity), while a higher intensity summer insolation (with summer occurring at
perihelion) would correspond to a shorter duration. Therefore, the integrated summer energy metric is insensitive to precessional changes, due to the opposing effects of precession on intensity and duration. Consequently, the majority of the variation in glacial-interglacial cycles is observed to be in the obliquity periods.

On the other hand, it has also been suggested that ice-sheet melt is controlled by local summer insolation, which is dominated by the 23,000-yr precession period at nearly all latitudes (M E Raymo, Lisiecki, & Nisancioglu, 2006). However, Earth’s orbital precession is *out-of-phase* between hemispheres, i.e. when the Northern Hemisphere has intense summer insolation, the Southern Hemisphere has weak local insolation and vice-versa. Thus, melting of ice-sheets in one hemisphere could partially balance ice-sheet growth in the other hemisphere, muting precessional cyclicity in globally integrated ice volume proxies such as marine benthic $\delta^{18}$O as proposed by Maureen Raymo (M E Raymo et al., 2006).

In this study, I demonstrate that at low eccentricities, the total integrated summer insolation in insensitive to precession, and climate system responds primarily to obliquity forcing. At high eccentricities, the integrated summer insolation metric is sensitive to precession, but the effects are opposite in the two hemispheres. Consequently, the earth’s climate system responds to both precession and obliquity at higher eccentricities.

2.4 Methods

Here, I use a General Circulation Model (GCM) to simulate the climatic response to orbital obliquity and precession between 2.0 and 1.0-Ma, when benthic $\delta^{18}$O is dominated by 40-kyr cyclicity.

2.4.1 GENESIS version 3 General Circulation Model

I use the current version of the Global ENvironmental and Ecological Simulation of Interactive Systems (GENESIS) 3.0 GCM with a slab ocean component (Thompson & Pollard, 1997) rather than a full-depth dynamical ocean (Alder, Hostetler, Pollard, &
Schmittner, 2011). The 50-m surface ocean model includes prognostic sea surface temperatures, diffusive heat transport, and thermodynamic sea ice, but still provides the computational efficiency required to run the 1000-simulation orbital sequence between 2 and 1Ma. The GCM has been used previously in many modern, future, and paleoclimate studies, including Pliocene-Pleistocene simulations (e.g. Coletti, DeConto, Brigham-Grette, & Melles, 2015; DeConto, Pollard, & Kowalewski, 2012; Koenig, DeConto, & Pollard, 2012). The 3-D atmospheric component of the GCM uses an adapted version of the NCAR CCM3 solar and thermal infrared radiation code (Kiehl et al., 1998). In the configuration used here, the model atmosphere has a spectral resolution of T31 (~3.75°) with 18 vertical layers, coupled to 2°x2° surface models including the slab ocean-sea ice model and soil, snow and vegetation components. For each experiment, the model is run for 50 model years, allowing quasi-equilibrium to be reached after about 20 years of integration. The results used to calculate PDDs are averaged over the last 10 years of each simulation.

2.4.2 Model Boundary Conditions, Forcing and Experiment Design

The GCM simulations are divided into 3 sets of experiments, with different transient orbital forcings (Table 1). The GCM uses a modern global geography, spatially interpolated to the model’s 2°x2° surface grid (Koenig et al., 2012). The geography provides the land-ice sheet-ocean mask and land-surface elevations used by the GCM. Greenhouse gas mixing ratios are identical in all experiments and set at preindustrial levels with CO₂ set at 280 ppmv, N₂O at 288 ppbv and CH₄ at 800 ppb. The default values for CFCl₃ and CF₂Cl₂ values are set at 0 ppm. The solar constant is maintained at 1367 Wm-2.

The orbital parameters used in the first two experiments are idealized and do not correspond to a specific time in Earth history. Rather, they are designed to provide a useful framework for isolating the effects of precession and obliquity on the climate response. The orbital parameters used in the third experiment are taken from the astronomical solutions in ref. (Berger & Loutre, 1991). Transient changes in the orbital
parameter configuration are applied at every 1-kyr intervals to simulate the whole range of orbital variation during the period 2.0 to 1.0-ma.

Both temperature and the length of the melt season (duration of summer) are important for net ice-sheet mass balance. One way to consider the melt season is by calculating the Summer Energy (J) as defined in Huybers, 2006. The Summer Energy is an integrated measure of changes in insolation intensity as well as duration of the melt season, and is defined as $J = \sum_i \beta_i (W_i \times 86,400)$, where $W_i$ is mean insolation measured in W/m$^2$ on day $i$, and $\beta$ equals one when $W_i \geq \tau$ and zero otherwise. $\tau = 275$ W/m$^2$ is taken as the threshold for melting to start at the surface of the earth. Rather than assuming a simple insolation-melt relationship (Huybers, 2006), we use the physically based climate model to calculate the sum of Positive Degree-Days (PDD) over the high latitudes of both hemispheres and its evolution from 2.0 to 1.0-Ma. In this case, Positive Degree-Days are calculated as $PDD = \sum_i \alpha_i T_i$, where $T_i$ is the mean daily temperature on day $i$, and $\alpha$ is one when $T_i \geq 0^\circ C$ and zero otherwise. The PDD captures the extremity as well as the duration of the melt season, and has been shown to be a good indicator of ice-sheet ablation potential (Braithwaite & Zhang, 2000).

2.5 Setup 1- Obliquity and Precession sensitivity experiments

In our first experimental setup, we aim to study the difference between how the climate responds to obliquity and precession by forcing the GCM with one parameter at a time, while keeping others constant. In our four GCM experiments, we vary either obliquity or precession at both high and low eccentricities. This provides a framework for investigating the inter-hemispheric climate responses to obliquity and precession.

2.5.1 Earth’s climate sensitivity to Obliquity

We first force the GCM with sequential changes in obliquity every 1-kyr (22.04° to 24.50°) over a complete 41-kyr cycle, while keeping precession constant with perihelion corresponding with the equinoxes, rather than the solstices. Increasing the obliquity from minimum (22.04°) to maximum (24.50°) results in an increase in the mean
summer insolation intensity of ~40 w/m² at 80°N. Two sets of simulations are run, with eccentricity set at the ~lowest (e=0.0001) and ~highest (e=0.05).

We calculate mean summer temperatures, integrated summer energy (J), and PDDs from our GCM simulations for Northern and Southern Hemispheres at 80°N and 80°S respectively (Fig 1 a–e). With an increase in obliquity, increases in summer insolation and summer energy (J) in both hemispheres are identical. Obliquity driven changes in summer insolation and summer energy are not impacted by eccentricity (Fig 1-b, d). Changes in mean summer temperature and PDD are also similar and in-phase between the hemispheres, and eccentricity has no effect on the high-latitude sensitivity to obliquity forcing.

2.5.2 Earth’s climate sensitivity to Precession

We next force the GCM with transient changes in precession over a 26-kyr cycle, while keeping obliquity constant (23.2735°). Precession is varied from NHSP (Northern Hemisphere summer at Perihelion) to SHSP (Southern Hemisphere summer at Perihelion). Two sets of simulations are run, again with eccentricity set at 0.0001 or 0.05. At high eccentricities, precessional changes aligning summer at perihelion versus aphelion results in an increase in summer insolation intensity at Top of Atmosphere (TOA) by ~120 W/m². However, at low eccentricities, the same precessional shift leads to negligible changes in summer insolation intensity.

We calculate mean summer temperatures, summer integrated energy, and PDD from our GCM simulations for Northern and Southern Hemispheres at 80°N and 80°S respectively (Fig 1 f–j). At low eccentricities, precessional changes in summer energy (J), mean summer temperature, and PDD are negligible. In contrast, at high eccentricities, precession affects summer insolation, summer energy (J) and the climatic responses, i.e. mean summer temperature and PDD. The precessional variation is clearly out-of-phase between Northern and Southern Hemispheres.
### Table 2.1: Experiment Design and Orbital Forcing

<table>
<thead>
<tr>
<th>Experiment</th>
<th>Eccentricity</th>
<th>Obliquity</th>
<th>Precession</th>
<th>Greenhouse Gases</th>
</tr>
</thead>
<tbody>
<tr>
<td>Experiment 1</td>
<td>0.0001</td>
<td>Transient</td>
<td>Perihelion at Equinoxes</td>
<td>Pre-Industrial</td>
</tr>
<tr>
<td></td>
<td></td>
<td>(22.04° to 24.5°)</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>0.05</td>
<td>Transient</td>
<td>Perihelion at Equinoxes</td>
<td>Pre-Industrial</td>
</tr>
<tr>
<td></td>
<td></td>
<td>(22.04° to 24.5°)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Experiment 2</td>
<td>0.0001</td>
<td>23.2735°</td>
<td>Transient (NHSP to SHSP)</td>
<td>Pre-Industrial</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>0.05</td>
<td>23.2735°</td>
<td>Transient (NHSP to SHSP)</td>
<td>Pre-Industrial</td>
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<tr>
<td></td>
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<td></td>
<td></td>
</tr>
<tr>
<td>Experiment 3</td>
<td>Transient</td>
<td>Transient</td>
<td>Transient</td>
<td>Pre-Industrial</td>
</tr>
</tbody>
</table>

**NHSP:** Northern Hemisphere Summer Solstice at Perihelion  
**SHSP:** Southern Hemisphere Summer Solstice at Perihelion
**Figure 2.1: Climate sensitivity to Obliquity and Precession forcing.** Solid lines correspond with simulations at HIGH eccentricity (0.05); dashed lines show simulations at LOW eccentricity (0.0001). Northern Hemisphere values are shown in blue; Southern Hemisphere values are red. **a-e. Climate response to obliquity.** a. Obliquity varies from 22.04° to 24.5° over a 41-kyr cycle. b. NH and SH insolation variation with changing obliquity. Note that both LOW and HIGH eccentricity simulations have similar variations in insolation. c. Summer temperatures are controlled primarily by local insolation. Northern and Southern Hemisphere mean summer temperatures vary in-phase at both low
and high eccentricities. **d.** Summer Energy (J), an integrated measure of insolation intensity and summer duration, shows similar variation in both hemispheres through an obliquity cycle and is not affected by eccentricity. **e.** Northern and Southern Hemisphere PDDs also show similar and in-phase variation, and are insensitive to eccentricity.

**f-j. Climate response to precession.** **f.** Precession varies from NHSP (Northern Hemisphere Summer at Perihelion) to SHSP (Southern Hemisphere Summer at Perihelion) over a 26-kyr cycle, **g.** NH and SH insolation variation for changes in precession, demonstrating out-of-phase insolation changes across Northern and Southern Hemispheres when eccentricity is high. **h.** Summer temperatures varying with local insolation. Northern and Southern Hemisphere mean summer temperatures vary out-of-phase at high eccentricities. **i.** Summer Energy (J) shows hemispheric out-of-phase variation at high eccentricities, and no variation at low eccentricities. **j.** Northern and Southern Hemisphere PDDs showing hemispheric in-phase variation at high eccentricities, and no variation at low eccentricities.
2.6 Orbital forcing of climate during the early Pleistocene

Next, we use realistically varying orbital parameters to investigate the effect of orbital forcing between 2.0 to 1.0 million years ago, when obliquity cyclicity dominates most proxy climate and ice volume records. Evolving orbital values (Berger & Loutre, 1991) are applied to the GCM in 1-kyr intervals (1000 GCM simulations). The high temporal resolution of our ensemble of simulations allows the possibility of direct comparisons with benthic δ¹⁸O records for this period.

High latitude summer insolation intensity is primarily controlled by the precession of the equinoxes, clearly evidenced by the spectrum of summer insolation intensity at 80°N and 80°S (Fig 2-c). Summer Temperature is largely controlled by local insolation, and consequently also varies at precessional frequencies (1/21 ± 1/100 kyr). Mean Northern Hemisphere and Southern Hemisphere summer temperatures have identical power spectral distributions, with >80% of their variation at precessional frequencies (Fig 2-d). When we consider both insolation intensity and summer duration, in context of the integrated summer energy (J), the strongest variation is observed at obliquity periods. This result agrees with the hypothesis (Huybers, 2006) that precessional changes in summer duration and intensity nearly balance each other, and obliquity is dominant in the variation of summer energy. The power spectrum of integrated summer energy (Fig 2-e) has 80% of its variation at frequencies corresponding to obliquity (1/41 kyr), with little variation at precessional frequencies.

While this may seem to solve the obliquity paradox, TOA calculations for summer Energy (J) used to infer orbital-climate-ice volume relationships don’t consider physical climatological processes, which play an important role in determining surface air temperature (and ablation potential) at any particular place. To account for such processes, we calculate PDDs from the surface air temperatures at 80° North and South simulated by the GCM, which accounts for both radiative and dynamical effects of changing orbits (Fig 2-f). While the PDDs still vary primarily at obliquity frequencies (~50% variation at 1/41 kyr), there is a strong variance at precessional frequencies in
both Northern and Southern Hemisphere PDDs (40% variation at 1/21±1/100 kyr). A wavelet transformation of the Northern and Southern Hemisphere PDD reveals that this precessional variation is present only during periods of high eccentricity. During periods of low eccentricity, variations at precessional frequencies are absent (Fig 3). This agrees with our previous result (Fig 1), which showed that out-of-phase precessional variation in PDD is significant only at high eccentricities. The wavelet transformations of both Northern and Southern PDD show excellent correspondence between periods of high eccentricity and strong variation at precessional frequencies.

Next, a windowed correlation was computed between the Northern Hemisphere PDD time series and Southern Hemisphere PDD time series. When the correlation is positive, the Northern and Southern Hemisphere PDD variation are positively correlated (i.e. in phase), and when the correlation is negative, the Northern and Southern Hemisphere PDD variation are negatively correlated (i.e. out-of-phase). The correlation between the Northern and Southern summer metrics show a strong variation at 100-kyr time periods (Fig 4), corresponding to eccentricity forcing. When eccentricity is higher, correlation is negative, i.e. Northern and Southern Hemispheres are out-of-phase. When eccentricity is low, correlation is positive, i.e. Northern and Southern Hemispheres are in-phase. This further reinforces our observation that eccentricity controls whether Northern and Southern Hemisphere have in-phase or out-of-phase climate responses to orbital forcing.
Figure 2.2: Orbital Forcing and climate variations during the early Pleistocene.

Northern Hemisphere values are blue; Southern Hemisphere values are red. **a.**

Eccentricity, **b.** Obliquity (degrees of axial tilt relative to the ecliptic) and Precession(Berger & Loutre, 1991). **c.** Insolation variation (Wm⁻²) during 2.0 – 1.0ma. The primary variation lies in precessional frequencies (purple), followed by the variations in obliquity band (pink). **d.** Mean summer temperatures for Northern Hemisphere (JJA; blue) and Southern Hemispheres (DJF; red). Summer temperature is largely controlled by local insolation; consequently the primary variation is in precession bands (purple). **e.**
Summer Energy (J) for Northern Hemisphere (blue) and Southern Hemispheres (red). Summer energy is a function of insolation intensity and summer duration; which varies primarily at obliquity periods (pink). PDDs, an indicator of ablation, show the influence of orbital forcing on ice-sheets. PDD for Northern Hemisphere (blue) and Southern Hemisphere (red) have ~50% of the variance at obliquity periods (pink) and ~40% of the variation at precessional periods.
Figure 2.3: Eccentricity control on interhemispheric phasing of PDDs. a. Eccentricity from astronomical calculations (Berger & Loutre, 1991). b. Evolutive spectrum of Northern Hemisphere and c. Southern Hemisphere PDD at 80°N and 80°S respectively, showing the evolution of precessional and obliquity frequencies. Strong power in precession and obliquity are seen during high eccentricities across the entire simulation period, while at low eccentricities, obliquity dominates. Variations at precession and obliquity bands at 95% or higher significance levels are indicated by black contour lines in b and c. Vertical dashed lines indicate eccentricity minima. d. Northern Hemisphere (blue) and Southern Hemisphere (red) PDD variation, with periods of in-phase variation
marked by gray shading, and periods of out-of-phase variation marked by yellow shading.

Figure 2.4: Windowed Correlation of Northern and Southern Hemisphere PDDs

(a) Northern Hemisphere and Southern Hemisphere PDD variation (b) Windowed correlation between the Northern and Southern Hemisphere PDD, showing strong variation at 100-kyr band (c) The correlation coefficient (smoothed using a low pass filter) on the left axis, and eccentricity plotted on the right axis. It can be observed that negative correlation coefficients (out-of-phase PDD variation) correspond to high eccentricities, and positive correlation coefficients (in-phase PDD variation) correspond to low eccentricities.
2.7 Threshold Eccentricity for Precession control on PDD

After filtering the high frequency variations from each hemisphere, we simply use the first derivatives of the PDD time-series \( \frac{d\text{PDD}}{dt} \) to determine the phasing between Northern and Southern Hemispheres (Fig 3-d). If both Northern Hemisphere and Southern Hemisphere PDD increase or decrease simultaneously (i.e. derivatives have the same sign), the Hemispheres are in-phase. When PDD in one hemisphere increases while the other decreases, Northern and Southern Hemispheres are out-of-phase. As clearly seen in Fig. 3-d, in-phase PDD variation (gray shading) corresponds to periods of low eccentricity (i.e. no precession forcing), while out-of-phase variation (yellow shading) corresponds to high eccentricity (i.e. precession forcing is present). This is expected, because the effect of the precession of the equinoxes is opposite in the two hemispheres. Strong variation at eccentricity periods (100-kyr) in the correlation of Northern and Southern PDDs (supplementary) reinforces our observation of eccentricity control on the hemispheric phasing of climate responses to orbital forcing.

The first derivatives of the time series of Northern and Southern Hemisphere PDDs are calculated, and multiplied with each other to obtain ‘m’ as defined below:

\[
m = \frac{d}{dt} \text{PDD}_{\text{NH}} \times \frac{d}{dt} \text{PDD}_{\text{SH}}
\]

If Northern Hemisphere and Southern Hemispheres are in-phase, the derivatives of the PDD will have the same sign (i.e. PDDs are increasing or decreasing in both hemispheres), and therefore ‘m’ will be positive. If Northern Hemisphere and Southern Hemispheres are out-of-phase, the derivatives of the PDD will have opposite signs (i.e. PDDs are increasing or decreasing asynchronously in the two hemispheres), and therefore ‘m’ will be negative. By plotting ‘m’ as a function of eccentricity, we can calculate the threshold value of eccentricity at which the value of ‘m’ switches from positive to negative, thus going from in-phase to out-of-phase climate response to orbital forcing.
Figure 2.5: Threshold for eccentricity control of hemispheric phasing in climate response. ‘m’ is the product of the first derivatives of the PDD time series of Northern and Southern Hemispheres. For m>1, Northern and Southern Hemispheres have in-phase variation in PDDs; for m<1, Northern and Southern Hemispheres have out-of-phase variation in PDDs. The value of ‘m’ is plotted as a function of eccentricity, and the value of threshold eccentricity is equal to the eccentricity when m=0.
2.8 Empirical Mode Decomposition

Empirical Mode Decomposition (EMD) is a method of breaking down a natural signal into a set of Intrinsic Mode Functions (IMFs) which by themselves are sufficient to describe the original signal. The IMFs are all in the time domain and of same length as the original signal, and allows for varying frequency in time to be preserved. Each IMF represents a different part of the signal, thus providing a way to breakdown the different forcing factors of the climate signal. Using EMD, we decomposed the Summer Metric signal of each hemisphere into the constituting precessional (Fig 6-c) and obliquity components (Fig 6-d).

The precessional component is out-of-phase between the Northern and Southern Hemispheres, with a cross-correlation factor of -46.77. Consequently, with precessional forcing, the earth responds oppositely in the two hemispheres. However, the stronger obliquity component in the summer metric is in-phase between the Northern and Southern Hemisphere, with a cross-correlation factor of 59.29, resulting in a strong in-phase climate response to obliquity in both hemispheres.
Figure 2.6: Empirical Mode Decomposition (a) Obliquity and Precession variation during 2.0 – 1.0-myr (b) Northern and Southern Hemisphere PDD variation (c) 3rd IMF of PDD time series, corresponding to the precessional component. It can be observed that the precessional component of PDD variation is out-of-phase between the Northern and Southern Hemisphere (d) 4th and 5th IMF of PDD time series, corresponding to the obliquity component. The obliquity component is in-phase between both hemispheres.
2.9 Greenhouse Gas Feedback

In the GCM simulations used in the analysis explained above, greenhouse gas values were kept constant to isolate the effects of orbital forcing. Previous studies have shown that ice-driven responses to orbital forcing lead to lagged changes in atmospheric CO$_2$, which consequently provide positive feedback to the ice-sheets, thus strengthening the orbital-led variations (Ruddiman, 2003, 2006). The few time-continuous records of CO$_2$ variation spanning the time period of our simulations show a strong variation at obliquity frequencies, which should enhance the climatic response. To test the importance of greenhouse gas feedback, we repeated the 1000-year orbital sequence from 2.0 to 1.0-Ma, by adding CO$_2$ forcing with CO$_2$ concentrations varying with insolation, scaled to observations (Hönisch, Hemming, Archer, Siddall, & McManus, 2009). For upper and lower boundaries of our record, we use the glacial and interglacial pCO$_2$ extremes as estimated from boron isotopes in planktic foraminifers from Honisch et al. (2009). We forced the 1000 GCM simulations from 2.0 to 1.0-my with orbital forcing and CO$_2$ concentrations from the synthetic time series, which we created. As expected, we find the addition of CO$_2$ feedback enhances the obliquity response and attenuates the precessional response in Northern and Southern Hemisphere PDDs, but the effect on the eccentricity threshold determining obliquity versus precessional dominance remains unchanged.
Figure 2.7: Effect of time-varying CO2 concentrations on Northern and Southern Hemisphere PDDs (a) Eccentricity and (b) Obliquity and Precession variation during 2.0 – 1.0-myr (c) Northern and Southern Hemisphere summer insolation (d) Synthetic Carbon Dioxide record used as forcing (e) PDD variations from GCM simulations with constant CO2 concentrations (f) PDD variations from GCM simulations with time-varying CO2 concentrations.
2.10 Conclusion

In summary, our results demonstrate that the climate response to precession versus obliquity forcing is fundamentally controlled by eccentricity. Our model results show that the effect of high summer insolation intensity at the precession frequency is not cancelled out by the shorter summer season at all eccentricities as previously proposed. Except for those intervals when eccentricity is low and therefore precession forcing is weak, we find that surface temperatures, and presumably changes in polar ice volume, should record a strong precession signal as proposed by classic Milankovitch theory.

Only when eccentricity is below the threshold of ~0.019, do Northern and Southern Hemisphere polar climate response in-phase and with an obliquity beat. At these low eccentricities, the precession-forced changes in insolation intensity are small and are outweighed by changes in summer duration. Consequently, obliquity controlled changes in summer insolation dominate surface heating. At low eccentricities, precessional variation of PDD is weak in both hemispheres and obliquity is the dominant astronomical forcing impacting high-latitude climate and ice volume. Since obliquity affects both hemispheres similarly, Northern and Southern Hemisphere climates respond in-phase, and ice sheets in both hemispheres grow and melt synchronously.

When eccentricity is higher than ~0.019, Northern and Southern Hemisphere climate response to orbital forcing is asynchronous. At higher eccentricities, high-latitude summer temperatures become increasingly sensitive to precession, leading to interhemispheric out-of-phase variations in PDD, and potentially ice sheets. Consequently, when eccentricity is high, both precession and obliquity dominate the astronomical forcing and high-latitude climate response, supporting the (out-of-phase) hypothesis posed by Raymo (2006) that suggests that polar ice volume in each hemisphere does in fact record a precession signal but that it is missing in globally integrated records such as ocean δ¹⁸O since the Northern and Southern Hemispheric responses are out of phase at this frequency.
In summary, these results using a physically based model provides support for the Anti-phase hypothesis proposed to explain the problem of the “41kyr world”. We fail to find evidence to support the alternative view that the lack of precession observed in the “41kyr world” is caused by precessional forcing of summer duration counteracting insolation intensity. An important next step will be the inclusion of time evolving Antarctic and Northern Hemispheric ice-sheets in response to orbital forcing during this interval, to more directly compare the model results with marine isotope records.
CHAPTER 3

INTERHEMISPHERIC EFFECT OF GLOBAL GEOGRAPHY ON EARTH'S CLIMATE RESPONSE TO ORBITAL FORCING

3.1 Abstract

The climate response to orbital forcing shows a distinct hemispheric asymmetry due to the unequal distribution of land in the Northern versus Southern Hemispheres. This asymmetry is examined using a Global Climate Model (GCM) and a Land Asymmetry Effect is quantified for each hemisphere. The results show how changes in obliquity and precession translate into variations in the calculated interhemispheric effect. We find that the global climate response to specific past orbits is likely unique and modified by complex climate-ocean-cryosphere interactions that remain poorly known and difficult to model. Nonetheless, these results provide a baseline for interpreting contemporaneous proxy climate data spanning a broad range of latitudes, which maybe especially useful in paleoclimate data-model comparisons, and individual time-continuous records exhibiting orbital cyclicity.

3.2 Introduction

The arrangement of continents on the earth’s surface plays a fundamental role in the earth’s climate response to forcing. This global “geography” is primarily the result of the horizontal and vertical displacements associated with plate tectonics. While these processes are ongoing, the global continental configuration has been close to its present form since the mid-Cenozoic. Today, more continental land area is found in the northern hemisphere (68%) as compared to the Southern Hemisphere (32%). These different ratios of land vs. ocean in each hemisphere affect the balance of incoming and outgoing radiation, atmospheric circulation, ocean currents, and the availability of terrain suitable for growing glaciers and ice-sheets. As a result of this land-ocean asymmetry, the
climatic responses of the northern and southern hemisphere differ for an identical change in radiative forcing (Barron, Thompson, & Hay, 1984; Deconto et al., 2008; Kang, Seager, Frierson, & Liu, 2014).

A number of classic studies have shown interhemispheric asymmetry in climate response of Northern and Southern Hemispheres. Climate simulations made with coupled atmosphere-ocean GCMs typically show a strong asymmetric response to greenhouse-gas loading, with Northern Hemisphere high latitudes experiencing increased warming compared to Southern Hemisphere high latitudes (Flato & Boer, 2001; Stouffer, Manabe, & Bryan, 1989). GCMs also show that the Northern and Southern Hemispheres respond differently to changes in orbital forcing (e.g. Philander et al., 1996). While the magnitude of insolation changes through each orbital cycle is identical for both hemispheres, the difference in climatic response can be attributed to the fact that Northern Hemisphere is land-dominated while Southern Hemisphere is water dominated (Croll, 1870). This results in a stronger response to orbital forcing in the Northern Hemisphere relative to the Southern Hemisphere.

The changing continental configurations as a result of plate tectonics have been linked with climate change over a wide range of timescales (e.g. Crowley & North, 1996; DeConto, 2009; Fawcett & Barron, 1998; Hay, 1996). The distribution of continents and oceans has an important effect on the spatial heterogeneity of the Earth’s energy balance, primarily via the differences in albedos and thermal properties of land versus ocean (Trenberth, Fasullo, & Kiehl, 2009). The latitudinal distribution of land has a dominant effect on zonally averaged net radiation balance due to it’s influence on planetary albedo and ability to transfer energy to the atmosphere through long-wave radiation, and fluxes of sensible and latent heat. The latitudinal net radiation gradient controls the total poleward heat transport requirement, which is the ultimate driver of winds, and ocean circulation (Stone, 1978).

Even though modern estimates of atmospheric and oceanic heat transport remain poorly constrained, it is generally accepted that oceans contribute less than half of the
total heat transport required to maintain the earth’s meridional energy balance (Bryden & Imawaki, 2001; Trenberth, Caron, Trenberth, & Caron, 2001; Wunsch, 2005). Maximum poleward heat transport through the oceans take place in the low latitudes, however the ocean plays an important role in polar climate via its influence on atmospheric teleconnections to the tropics and its control on seasonal distributions of sea-ice. Oceans have a relatively slower response to seasonal changes in insolation due to the higher specific heat of water as compared to land, and mixing in the upper ~10-150 m of the ocean. As a result, in the ocean-dominated southern hemisphere, the surface waters suppress extreme temperature swings in the winter and provide the atmosphere with a source of moisture and diabatic heating. In the land dominated northern hemisphere, the lower heat capacity of the land combined with relatively high albedo results in greater seasonality, particularly in the interiors of large continents of Asia and North America.

The continentality of the Northern Hemisphere manifests itself in different hemispherically asymmetric climatic phenomenon, like the well-known Asian monsoonal circulation system. The intertropical convergence zone (ITCZ) is considered to be the region of low-level convergence and convective precipitation. The ITCZ moves further away from the equator during the northern summer than the southern one due to the continentality of the Northern Hemisphere (Kang, Held, Frierson, & Zhao, 2008; Philander et al., 1996). The land surface available in a particular hemisphere also affects the potential for widespread glaciation. The extreme cold winters associated with large continents provide the means of accumulation of winter snow, while the critical factor for formation of ice-sheets is annual ablation and can be estimated by the sum of Positive Degree Days (PDD) in a year (e.g. Huybers, 2006).

Continental geography has a strong impact on polar climates, as is evident from the very different climatic regimes of the Arctic and the Antarctic. Several early paleoclimate modeling studies using GCMs investigated continental distribution as a forcing factor of global climate (e.g. Barron et al., 1984; Hay, Barron, & Thompson, 1990). Theses studies demonstrated that an earth with its continents concentrated in the low latitudes is warmer and has lower equator-to-pole temperature gradients than an earth
with only polar continents. Although these early model simulations did not incorporate all the complexities of the climate system, the results provided valuable insights from comparative studies of polar versus equatorial continents in the earth and showed that changes in continental configuration has significant influence on climatic response to forcing.

3.3 Methods

3.3.1 Experimental Design

We use the latest (2012) version of the Global ENvironmental and Ecological Simulation of Interactive Systems (GENESIS) 3.0 GCM with a slab ocean component (Thompson & Pollard, 1997) rather than a full-depth dynamical ocean (Alder, Hostetler, Pollard, & Schmittner, 2011). The slab-ocean predicts sea surface temperatures and ocean heat transport as a function of the local temperature gradient and the zonal fraction of land versus sea at each latitude. While explicit changes in ocean currents and the deep ocean are not represented, the computational efficiency of the slab-ocean version of the GCM allows numerous simulations with idealized global geographies and greatly simplifies interpretations of the sensitivity tests by precluding complications associated with ocean model dependencies. In addition to the atmosphere and slab-ocean, the GCM includes model components representing vegetation, soil, snow, and thermo-dynamic sea ice. The 3-D atmospheric component of the GCM uses an adapted version of the NCAR CCM3 solar and thermal infrared radiation code (Kiehl et al., 1998) and is coupled to the surface components by a land-surface-transfer scheme (LSX). In the setup used here, the model atmosphere has a spectral resolution of T31 (~3.75°) with 18 vertical layers. Land-surface components are discretized on a higher resolution 2°x2° grid.

The GCM uses various geographical boundary conditions (described below) in 2°x2° and spectral T31 grids for surface and AGCM models, respectively. For each set of experiments, the model is run for 50 years. Spin-up is taken into account, and equilibrium is effectively reached after about 20 years of integration. The results used to
calculate interhemispheric effects are averaged over the last 20 years of each simulation. Greenhouse gas mixing ratios are identical in all experiments and set at preindustrial levels with CO$_2$ set at 280 ppmv, N$_2$O at 288 ppbv and CH$_4$ at 800 ppbv. The default values for CFCl$_3$ and CF$_2$Cl$_2$ values are set at 0 ppm. The solar constant is maintained at 1367 Wm$^{-2}$.

### 3.3.2 Asymmetric and Symmetric Earth Geographies

The GCM experiments are divided into three sets: 1) Preindustrial CONTROL 2) NORTH-SYMM and 3) SOUTH-SYMM. The Preindustrial CONTROL experiments use a modern global geography spatially interpolated to the model’s 2$^\circ$x2$^\circ$ surface grid (Koenig, DeConto, & Pollard, 2012). The geography provides the land–ice sheet–ocean mask and land–surface elevations used by the GCM.

To simulate the climate of an Earth with meriodionally symmetric geographies, we created two sets of land surface boundary conditions: NORTH-SYMM and SOUTH-SYMM. For the NORTH-SYMM experiments, the CONTROL experiment boundary conditions are used to generate a modified GCM surface mask, by reflecting the Northern hemisphere geography (land-sea-ice mask, topography, vegetation, soil texture) across the equator into the Southern hemisphere. Similarly, in the experiment SOUTH-SYMM, the land mask and geographic boundary conditions in the southern hemisphere are mirrored in the northern hemisphere. The NORTH-SYMM and SOUTH-SYMM boundary conditions are shown in figure 1(b) and 1(c), with the CONTROL (figure 1(a)) for comparison.
### Table 3.1 Experimental Setup of Model Boundary Conditions and Forcing

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<th>Run ID</th>
<th>LSX Configuration</th>
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<th>Obliquity</th>
<th>Precession $^a$</th>
<th>GHGs</th>
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<tr>
<td>CONTROL$_{NHSP}$</td>
<td>Modern</td>
<td>0.034</td>
<td>23.2735</td>
<td>270° (NHSP)</td>
<td>Preindustrial</td>
</tr>
<tr>
<td>CONTROL$_{SHSP}$</td>
<td>Modern</td>
<td>0.034</td>
<td>23.2735</td>
<td>90° (SHSP)</td>
<td>Preindustrial</td>
</tr>
<tr>
<td>CONTROL$_{HIGH}$</td>
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<td>0.034</td>
<td>24.5044</td>
<td>180°</td>
<td>Preindustrial</td>
</tr>
<tr>
<td>CONTROL$_{LOW}$</td>
<td>Modern</td>
<td>0.034</td>
<td>22.0425</td>
<td>180°</td>
<td>Preindustrial</td>
</tr>
<tr>
<td>NORTH-SYMM$_{NHSP}$</td>
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<td>23.2735</td>
<td>270° (NHSP)</td>
<td>Preindustrial</td>
</tr>
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</tr>
<tr>
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<td>0.034</td>
<td>23.2735</td>
<td>270° (NHSP)</td>
<td>Preindustrial</td>
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<td>0.034</td>
<td>23.2735</td>
<td>90° (SHSP)</td>
<td>Preindustrial</td>
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<td>24.5044</td>
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<td>SOUTH-SYMM$_{LOW}$</td>
<td>South-symmetric</td>
<td>0.034</td>
<td>22.0425</td>
<td>180°</td>
<td>Preindustrial</td>
</tr>
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</table>

**NHSP:** Northern Hemisphere Summer Solstice at Perihelion  
**SHSP:** Southern Hemisphere Summer Solstice at Perihelion

$^a$ Orbital precession in the GCM is defined here as the prograde angle from perihelion to the northern hemispheric vernal equinox.
Figure 3.1: Different geographies used in climate simulations (a) Modern continental geography (b) NORTH-SYMM geography and (c) SOUTH-SYMM geography
3.4 Asymmetry in the Earth’s Climate

We begin our study by investigating the asymmetry in the Earth’s climate. In our first experimental setup, we run the GCM with modern day orbital configuration, i.e. eccentricity is set at 0.0167, obliquity is set at 23.5° and precession such that perihelion coincides with Southern Hemisphere summer (The prograde angle from perihelion to the Northern Hemispheric vernal equinox is set at 101°). Figure 2(a) shows the present day summer insolation intensity and figure 2(b) shows present day Summer Energy for reference (The Summer Energy (J) is defined as defined as $J = \sum \beta_i (W_i \times 86,400)$, where $W_i$ is mean insolation measured in W/m$^2$ on day i, and $\beta$ equals 1 when $Wi \geq \tau$ and zero otherwise. $\tau = 275$ W/m$^2$ is taken as the assumed threshold for melting of ice at the earth’s surface).

Mean Summer Temperatures (ST) are calculated from the GCM as the mean of the average daily temperatures for the summer months in each hemisphere (JJA in Northern Hemisphere; DJF in Southern Hemisphere). Figure 2(c) shows the mean summer temperature for a simulation with modern orbit. The zonal averages (calculated for each latitude) demonstrate the inherent asymmetry in the Earth’s climate between Northern and Southern Hemispheres, especially evident in the higher latitudes. A better indicator of the Earth’s climate system, which quantifies both the intensity of summer as well as the duration of the melt season, is the sum of Positive Degree Days (PDD). Positive Degree-Days are calculated as $PDD = \sum T_i \alpha_i$ where $T_i$ is the mean daily temperature on day i, and $\alpha$ is one when $T_i \geq 0°C$ and zero otherwise. The PDD captures the intensity as well as the duration of the melt season, and has been shown to be indicative of the ice-sheet response to changes in external forcing. Fig 2(d) shows the PDD for modern orbit, and the zonal averages are plotted in the log scale. The extreme asymmetry between the Northern and Southern Hemispheres observed in the summer temperatures (Figure 2(c)) is also evident in the calculated PDDs (Figure 2(d)).

The observed asymmetry in the Northern and Southern Hemispheres can be attributed to three primary causes: (I) variation in insolation intensity across the Northern
and Southern Hemispheres caused by the precession of the equinoxes (today perihelion coincides with January 3, just after the December 21 solstice, leading to slightly stronger summer insolation in the Southern Hemisphere); (II) the effect of the continental geography on climate; and (III) the effect of interhemispheric continental geography on climate, i.e. the effect of Northern Hemisphere continental geography on Southern Hemisphere climate and vice-versa. Here, we attempt to isolate the effect of interhemispheric continental geography on climate (i.e. cause (III) above) by comparing results from GCM simulations using modern versus idealized (hemispherically symmetric) global geographies (Figure (1)).

Next, we maintain a modern orbit to test of effect of meriodionally symmetric continents (figure 2(e-h)). Figure 2(e) and 2(f) show the summer temperature and PDD from a simulation in which the Northern Hemisphere geography is reflected in the Southern Hemisphere (thus making the Earth geographically symmetric). Figure 2(g) and 2(h) show the summer temperature and PDD from a hypothetical simulation with symmetric Southern Hemisphere continents. Symmetric continents make the climates of Northern and Southern Hemispheres almost symmetric (>95%), with some small remaining asymmetry due to the current timing of perihelion with respect to the summer solstices.
Figure 3.2: **Simulations are forced by modern day orbit** (a) summer insolation; (b) summer energy*2 (See Notes); (c) Summer Temperature; (d) PDD; (e) and (f) Summer Temperature and PDD for NORTH-SYMM simulation; (g) and (h) Summer Temperature and PDD for SOUTH-SYMM simulation. The zonal averages are plotted on the right of each figure. Zonal averages of PDD are plotted on a log scale.

*Note: Higher resolution figures at end of chapter*
The simulations with modern and idealized (symmetric) geographies are used to quantify the different climate responses to a range of orbits. By comparing the climatic response from simulations with different geographies, we isolate and estimate the effect of interhemispheric continental geography and the influence of one hemisphere’s geography on the climate response of the opposite hemisphere.

### 3.4.1 Effect of Southern Hemisphere on Northern Hemisphere Climate

To estimate the effect of Southern Hemisphere continental geography on the Northern Hemisphere, we compare the Northern Hemisphere climate from the CONTROL simulation (asymmetric, modern orbit) and NORTH-SYMM (symmetric Southern continents, modern orbit). In these simulations, the only difference in setup is the Southern Hemisphere continental distribution. Thus the differences in Northern Hemisphere climate from the two simulations, if any, can be safely ascribed as the ‘effect of Southern Hemisphere continental geography on Northern Hemisphere climate’. We quantify this interhemispheric effect of Southern Hemisphere continental geography on Northern Hemisphere climate as:

\[
\hat{e}_{\text{Summer Temp}} = \frac{1}{n} \sum_{i} (T_{i}^{\text{control}} - T_{i}^{\text{north}})
\]

\[
\hat{e}_{\text{PDD}} = \text{PDD}^{\text{control}} - \text{PDD}^{\text{north}}
\]

Where \(T_{i}^{\text{control}}\) and \(\text{PDD}_{i}^{\text{control}}\) are the mean daily temperature on day \(i\) and PDD from the control simulation, and \(T_{i}^{\text{north}}\) and \(\text{PDD}_{i}^{\text{north}}\) are the mean daily temperature on day \(i\) and PDD from the simulation with the north-symmetric geography.

Figure 3(a) and 3(b) show the effect of Southern Hemisphere continental geography on Northern Hemisphere summer temperature and PDD respectively. For the Northern Hemisphere, the summer temperatures are calculated over the months of June, July, and August when the insolation intensity over the Northern Hemisphere is strongest. The asymmetry in the Southern Hemisphere landmasses leads to weakening of the
summer warming over North America and Eurasia (blue shaded regions correspond to cooling). Consequently, summer temperatures over Northern Hemisphere continents are lower by 3-6°C relative to a symmetric earth. There is a positive warming effect in the North-Atlantic Ocean, and in general the Northern Hemisphere oceans are slightly warmer relative to a symmetric earth. The general trends in the interhemispheric effect on PDD (Fig 3(b)) mimics those of the summer temperatures (Fig 3(a)).

3.4.2 Effect of Northern Hemisphere on Southern Hemisphere Climate

Similarly, we estimate the effect of Northern Hemisphere continental geography on the Southern Hemisphere by comparing the Southern Hemisphere climate to the CONTROL simulation (asymmetric, modern orbit) and the SOUTH-SYMM (symmetric Northern continents). In these simulations, the differences in Southern Hemisphere climate in the CONTROL and SOUTH-SYMM simulations, if any, can be ascribed as the ‘effect of Northern Hemisphere continental geography on Southern Hemisphere climate’. We quantify this interhemispheric effect of Northern Hemisphere continental geography on Southern Hemisphere climate as:

\[
e_{\text{Summer Temp}} = \frac{1}{n} \sum_{i} (T_{i}^{\text{control}} - T_{i}^{\text{south}})
\]

\[
e_{\text{PDD}} = PDD^{\text{control}} - PDD^{\text{south}}
\]

Where \(T_{i}^{\text{control}}\) and \(PDD_{i}^{\text{control}}\) are the mean daily temperature on day \(i\) and PDD from the control simulation, and \(T_{i}^{\text{south}}\) and \(PDD_{i}^{\text{south}}\) are the mean daily temperature on day \(i\) and PDD from the simulation with the south-symmetric geography.

Figure 3(c) and 3(d) show the effect of Northern Hemisphere continental geography on Southern Hemisphere summer temperature and PDD, respectively. For the SH, the summer temperatures are calculated over the months of December, January, and February when the insolation is most intense during the year. Southern Hemisphere landmasses, except Antarctica, generally show a cooling response during summer, due to
Northern Hemisphere geography. Summer temperatures are higher in the control simulations than in the symmetric simulations, leading to the inference that there is a warming (increase) in Antarctica summer temperatures due to interhemispheric effect of the Northern Hemisphere. Also, the Southern Ocean shows a strong positive temperature effect (warming) relative to a symmetric earth, although this Southern Ocean response might be different or modified if a full-depth dynamical ocean model were used.
Figure 3.3: Interhemispheric effect of Southern Hemisphere continental geography on (a) Northern Hemisphere Summer Temperature and (b) Positive Degree Days (PDD). Interhemispheric effect of Northern Hemisphere continental geography on (c) Southern Hemisphere Summer Temperature and (d) Positive Degree Days (PDD). Zonal averages are plotted on the right of each figure.

Note: Higher resolution figures at end of chapter
3.5 Interhemispheric effect on Earth’s Climate Response to Orbital Forcing

Next, we analyze the effect of the opposite hemisphere on the Earth’s climate response to changes in obliquity (axial tilt) and precession (positions of the solstices and equinoxes in relation to the eccentric orbit). The orbital parameters used in these experiments are idealized and do not correspond to a specific time in Earth’s history. Rather, they are chosen to provide a useful framework for studying the Earth’s climate response to precession and obliquity. HIGH and LOW orbits approximate the highest and lowest obliquity in the last three million years (Berger & Loutre, 1991). SP (Northern Hemisphere Summer at Perihelion) and SHSP (Southern Hemisphere Summer at Perihelion) orbits correspond to northern and austral summers coinciding with perihelion, respectively, and represent the two extreme configurations of precession, with obliquity set at its mean value averaged over the last 3 million years. Eccentricity is set at the same moderate value (mean eccentricity over the last 3 million years) for all simulations. Table 1 summarizes the orbits used in the ensemble of model simulations. Here, we focus only on the sum of the Positive Degree Days (PDD) calculated from our simulations. PDD is a better indicator of air temperature’s influence on annual ablation over ice-sheets than summer temperature, since this metric captures both the intensity and duration of the melt season.

3.5.1 Precessional Response of Earth’s Climate

Changes in precession primarily affect seasonal insolation intensity that is well known to be out-of-phase in both hemispheres (e.g. (Raymo, Lisiecki, & Nisancioglu, 2006)). The out-of-phase summer energy (J) variation is shown in figure 4(a) for reference. In one precessional cycle lasting ~23-kyr, the perihelion position of the earth’s orbit moves from the Northern Hemisphere summer solstice (NHSP) to the Southern Hemisphere summer solstice (SHSP), which are also the two extreme precessional configurations. We run the simulations at these two extreme precessions, keeping all other orbital parameters constant at their mean values. The difference in the calculated PDDs from the two simulations (represented as ΔPDD_{precession}) gives an estimate of the
earth’s climate response to the combined effect of the two precessional motions (wobbling of the axis of rotation and the slow turning of the orbital ellipse). Figure 4(b) shows the precessional response of the Earth in terms of PDD, and it is observed that the Northern and Southern Hemisphere responses are not symmetrical. Running the same simulations with a North-symmetric earth (Figure 4(c)) and a South-symmetric earth (Figure 4(d)) results in a nearly symmetrical climate responses to the precessional cycle.

3.5.2 Obliquity response of the Earth’s climate

In contrast to precession, obliquity alters the seasonality of insolation equally in both hemispheres (Figure 4(e)). A reduction in the tilt from 24.5° (HIGH) to 22° (LOW) reduces annual insolation by ~17 W/m² and summer insolation by ~45 W/m² in the high latitudes. In the tropics, summer insolation increases by up to ~5 W/m². Loutre et al (Loutre, Paillard, Vimeux, & Cortijo, 2004) among others predicted that global ice volume changes at the obliquity periods could be interpreted as a response to mean annual insolation and meridional insolation gradients. Similar to the experimental setup described above, we ran two simulations with the highest and lowest axial tilts, keeping all other orbital parameters constant at their mean values. The difference in the calculated PDDs (represented as ΔPDD_obliquity) provides an estimate of the Earth’s climate response to changes in tilt. Figure 4(f) shows ΔPDD_obliquity and the zonal averages reveal the asymmetry in the climate response to obliquity. Running the same simulations with a North-symmetric earth (Figure 4(g)) and a South-symmetric earth (Figure 4(h)) produces a nearly symmetrical climate response to the obliquity cycle.
Figure 3.4: Climate response in Modern Geography and Symmetric Geographies to Precession and Obliquity cycles (a) Summer Energy change for a transition from SHSP to NHSP orbit (Precession cycle) and the corresponding change in PDDs in CONTROL (b); NORTH-SYMM (c) and SOUTH-SYMM (d) simulations. (e) Summer Energy change for a transition from LOW to HIGH obliquity orbit and the corresponding change in PDD in CONTROL (b); NORTH-SYMM (c) and SOUTH-SYMM (d) simulations.

*Note: Higher resolution figures at end of chapter*
3.6 Effect of Southern Hemisphere on Northern Hemisphere climate

The effect of Southern Hemisphere continental geography on Northern Hemisphere at the two extreme precessional configurations is estimated using the same method described above, with Interhemispheric effect of Southern Hemisphere continental geography on Northern Hemisphere climate at 'NHSP' calculated as:

\[(e_{PDD})_{NHSP} = PDD_{NHSP}^{control} - PDD_{NHSP}^{north}\]

And Interhemispheric effect of Southern Hemisphere continental geography on Northern Hemisphere climate at 'SHSP' calculated as:

\[(e_{PDD})_{SHSP} = PDD_{SHSP}^{control} - PDD_{SHSP}^{north}\]

Figure 5(a) shows the spatial variation of \((e_{PDD})_{NHSP}\). The Northern Hemisphere landmasses show a strong negative response to PDD when perihelion coincides with Northern Hemisphere summer (NHSP). In this orbit, the Northern Hemisphere experiences elevated summer insolation, but the response is weakened due to the interhemispheric effect. This dampening effect is greatest in the interiors of the northern hemisphere continents (Fig. 5(a)). According to Milankovitch theory, the Northern Hemisphere should experience ‘interglacial’ conditions when perihelion coincides with boreal summer. However, because of the interhemispheric effect, interglacial (warm summer) conditions are muted relative to those on a symmetric earth. Figure 5(c) shows the spatial variation of \((e_{PDD})_{SHSP}\). When perihelion coincides with Southern Hemisphere summer (SHSP), the Northern Hemisphere continents have a weak positive effect, leading to slightly warmer conditions relative to a symmetric earth.

Next we try to observe the interhemispheric effect on \(\Delta PDD\) for a transition from NHSP to SHSP orbit whereby the Interhemispheric effect of Northern Hemisphere continental geography on Southern Hemisphere response to precession is:

\[e_{PDD} = \Delta PDD_{precession}^{control} - \Delta PDD_{precession}^{north}\]
The calculated effect is plotted spatially in figure 6(a), and shows a strong negative effect on Northern Hemisphere PDDs. For the Northern Hemisphere, the transition from NHSP to SHSP equates to a transition from warm to cold climate. The negative interhemispheric effect decreases the \( \Delta \text{PDD} \), thus weakening the effect of precession on the Northern Hemisphere.

The effect of Northern Hemisphere continental geography on Southern Hemisphere at the two extreme obliquity configurations are estimated as:

\[
(\hat{\varepsilon}_{\text{PDD}})_{\text{HIGH}} = PDD_{\text{HIGH}}^{\text{control}} - PDD_{\text{HIGH}}^{\text{north}}
\]

And

\[
(\hat{\varepsilon}_{\text{PDD}})_{\text{LOW}} = PDD_{\text{LOW}}^{\text{control}} - PDD_{\text{LOW}}^{\text{north}}
\]

At HIGH obliquity, there exists a negative effect on Northern Hemisphere continents (Figure 5(e)), which mutes the strong insolation intensity during summer months. In the Northern Hemisphere, as a result of continental asymmetry, a decrease in the equator to pole temperature gradient is observed as also seen by Lee & Poulsen (2008). A lowering of summer temperatures and temperature gradient due to the interhemispheric effect has a negative impact on the deglaciation trigger associated with HIGH obliquity orbits. Thus the interhemispheric effect would hinder the melting of ice during high-obliquity orbits. At LOW obliquity, the negative effect over Northern Hemisphere continents is generally less intense (Figure 5(g)). However, even the modest lowering of summer temperatures caused by the interhemispheric effect would support the growth of ice sheets during low obliquity orbits.

Further, we calculate the interhemispheric effect on \( \Delta \text{PDD} \) for a transition from LOW to HIGH orbit (obliquity cycle). This Interhemispheric effect of Northern Hemisphere continental geography on Southern Hemisphere response to obliquity is:
\[ e_{\text{PDD}} = \Delta \text{PDD}_{\text{obliquity}} \cdot \Delta \text{PDD}_{\text{obliquity}} \]

The calculated effect is spatially plotted in figure 6(c), and shows a small negative effect in the high latitudes, and a positive effect in the low latitudes. The transition from LOW to HIGH corresponds to a transition from cold to warm climate. The positive interhemispheric effect increases the \( \Delta \text{PDD} \), thus strengthening the climate response of obliquity cycle in the Northern Hemisphere.

### 3.7 Effect of Northern Hemisphere on Southern Hemisphere climate

The effect of Northern Hemisphere continental geography on Southern Hemisphere climate at two extreme precessional configurations is estimated as:

\[ (e_{\text{PDD}})_{\text{NHSP}} = \text{PDD}_{\text{NHSP}}^{\text{control}} - \text{PDD}_{\text{NHSP}}^{\text{south}} \]

and

\[ (e_{\text{PDD}})_{\text{NHSP}} = \text{PDD}_{\text{SHSP}}^{\text{control}} - \text{PDD}_{\text{SHSP}}^{\text{south}} \]

The spatial variation of \( (e_{\text{PDD}})_{\text{NHSP}} \) is shown in figure 5(b). During NHSP orbit, the Southern Hemisphere experiences ‘glacial’ (cold summer) conditions due to the weaker summer insolation. The positive effect in the southern hemisphere leads to weaker cooling relative to a symmetric earth. Thus, when perihelion coincides with Northern Hemisphere summer, the interhemispheric effect dampens the magnitude of ‘glacial’ versus ‘interglacial’ conditions in both hemispheres. When perihelion coincides with southern hemisphere summer (SHSP), the southern high latitudes experience intense summer insolation. The positive warming effect (Figure 5(d)) amplifies the ‘interglacial’ conditions in the Southern Hemisphere, predicted by Milankovitch theory.

The interhemispheric effect on \( \Delta \text{PDD} \) for a transition from NHSP to SHSP orbit, or the interhemispheric effect of Southern Hemisphere continental geography on Northern Hemisphere response to precession is:
The calculated effect is plotted spatially in Figure 6(c), and shows a strong negative effect on PDD over most of Southern Hemisphere. For the Southern Hemisphere, the transition from NHSP to SHSP equates to a transition from colder to warmer climate. The negative interhemispheric effect decreases the $\Delta$PDD, thus weakening the effect of precessional cycle in the Southern Hemisphere.

The interhemispheric effect of Northern Hemisphere continental geography on Southern Hemisphere climate at the two extreme obliquity configurations is calculated as.

$$(e_{PDD})_{HIGH} = PDD_{HIGH}^{control} - PDD_{HIGH}^{South}$$

And

$$(e_{PDD})_{LOW} = PDD_{LOW}^{control} - PDD_{LOW}^{South}$$

The spatial variations of $(e_{PDD})_{HIGH}$ and $(e_{PDD})_{LOW}$ are shown in figure 5(f) and 5(h), respectively. In the Southern Hemisphere, the positive interhemispheric effect on PDD over Antarctica and the Southern Ocean leads to overall higher temperatures in the high southern latitudes as compared to a symmetric earth. During high obliquity orbits, this positive effect contributes to deglaciation and during low obliquity orbits; the positive effect (warming) hinders the growth of ice sheets.

Lastly, we calculate the interhemispheric effect on $\Delta$PDD for a transition from LOW to HIGH orbit (obliquity cycle):

$$e_{PDD} = \Delta PDD_{obliquity}^{control} - \Delta PDD_{obliquity}^{South}$$

The calculated effect is plotted in figure 6(d), and shows largely a negative effect in the Southern Hemisphere, with a positive effect in the high latitudes. The transition
from LOW to HIGH corresponds to a transition from cold to warm climate. The positive interhemispheric effect increases the $\Delta$PDD, thus amplifying the effect of obliquity over Antarctica.

Figure 3.5: Interhemispheric effect of Southern Hemisphere continental geography on Northern Hemisphere PDD (left) and Interhemispheric effect of Southern Hemisphere continental geography on Northern Hemisphere PDD (right) at NHSP (a, b); SHSP (c, d); HIGH (e, f) and LOW (g, h) orbits.
Figure 3.6: Interhemispheric Effects on Precession cycles (left) and Obliquity cycles (right) (a) Interhemispheric effect of Southern Hemisphere continental geography on Northern Hemisphere PDD response to Precession cycle (NHSP to SHSP), (b) Interhemispheric effect of Northern Hemisphere continental geography on Southern Hemisphere PDD response to Precession cycle (NHSP to SHSP), (c) Interhemispheric effect of Southern Hemisphere continental geography on Northern Hemisphere PDD response to Obliquity cycle (LOW to HIGH), (d) Interhemispheric effect of Northern Hemisphere continental geography on Southern Hemisphere PDD response to Obliquity cycle (LOW to HIGH). Zonal averages are plotted on the right of each figure.

Note: Higher resolution figures at end of chapter
3.7 Conclusion

The unbalanced fraction of land in the Northern versus Southern Hemisphere has remained almost unchanged for tens of millions of years. However, the significance of this continental asymmetry on Earth’s climate response to forcing has not been previously quantified with a physically based climate model. We find that continental geography has an important control on the climate system’s response to insolation forcing, and this may help explain the non-linear response of the Earth’s climate to insolation forcing.

According to classical Milankovitch theory, the growth of polar ice sheets at the onset of glaciation requires cooler summers in the high latitudes, in order for snow to persist throughout the year. During warm summers at the high latitudes, the winter snowpack melts, inhibiting glaciation or leading to deglaciation if ice sheets already exist. Thus, the intensity of summer insolation at high latitudes, especially the northern polar latitudes, has been considered the key driver of the glacial-interglacial cycles and other long-term climatic variations (e.g. Milankovitch, 1941). At precessional periods, at which the high latitude summer intensity primarily varies, the land asymmetry effect plays an important role by amplifying (or weakening) the effect of summer insolation intensity.

In all the orbital configurations simulated here, we find that the geography of the Southern Hemisphere weakens the temperature response of the high Northern Hemisphere latitudes to orbital forcing. Consequently, this leads to a larger latitudinal gradient in summer temperatures in the Northern Hemisphere compared to a that of a symmetric Earth. In particular, the amplification (or weakening) of the response to insolation changes at precessional and obliquity periods might explain some of the important features of late Pliocene-early Pleistocene climate variability, when obliquity-paced cyclicity dominated precession in global benthic $\delta^{18}O$ records (Lisiecki and Raymo, 2005). In Figure 6, we have demonstrated that the interhemispheric effect causes a suppression of the effects of precessional cycle on the Earth’s surface. In other words, the real Earth has a smaller response to a precession cycle as compared to the
hypothetical symmetric earth. We have also showed that the interhemispheric effect causes an amplification of the effects of obliquity cycle on the Earth’s surface. In other words, the real Earth has a larger response to the obliquity cycle as compared to the hypothetical symmetric earth. Consequently, the interhemispheric effect of continental geography contributes to the muting of precessional signal and amplification of obliquity signal recorded in paleoclimate proxies such as benthic $\delta^{18}$O isotope records.

There are various ways in which the Earth’s continental asymmetry affects climate. Here, we have shown how these interhemispheric effects influence the Earth’s climate response to orbital forcing via the radiative and atmospheric dynamical processes represented in a slab-ocean GCM. While computationally challenging, future work should include complimentary simulations with AOGCMs, to explore the potential modifying role of ocean dynamics on the amplifying and weakening interhemispheric responses to orbital forcing demonstrated here.
3.8 High Resolution Figures

Figure 3.2(a): Summer Insolation in a GCM simulation forced with modern day orbit

Figure 3.2(b): Summer Energy in a GCM simulation forced with modern day orbit
Figure 3.2(c): Summer Temperature in a GCM simulation forced with modern day orbit

Figure 3.2(d): Positive Degree Days (PDD) in a GCM simulation forced with modern day orbit
Figure 3.2(e): Summer Temperature in a GCM simulation forced with modern day orbit and NORTH-SYMM geography

Figure 3(f): Positive Degree Days (PDD) in a GCM simulation forced with modern day orbit and NORTH-SYMM geography
Figure 3.2(g): Summer Temperature in a GCM simulation forced with modern day orbit and SOUTH-SYMM geography

Figure 3.2(h): Positive Degree Days (PDD) in a GCM simulation forced with modern day orbit and SOUTH-SYMM geography
Figure 3.3(a): Interhemispheric effect of Southern Hemisphere continental geography on Northern Hemisphere Summer Temperature (modern orbit)

Figure 3.3(b): Interhemispheric effect of Southern Hemisphere continental geography on Northern Hemisphere PDD (modern orbit)
Figure 3.3(c): Interhemispheric effect of Northern Hemisphere continental geography on Southern Hemisphere Summer Temperature (modern orbit)

Figure 3.3(d): Interhemispheric effect of Northern Hemisphere continental geography on Southern Hemisphere PDD (modern orbit)
Figure 3.4(a): Summer Energy change for a transition from SHSP to NHSP orbit (Precession cycle)

Figure 3.4(b): Corresponding PDD change for a transition from SHSP to NHSP orbit (Precession cycle)
Figure 3.4(c): Corresponding PDD change for a transition from SHSP to NHSP orbit (Precession cycle) in a simulation with NORTH-SYMM geography.

Figure 3.4(d): Corresponding PDD change for a transition from SHSP to NHSP orbit (Precession cycle) in a simulation with SOUTH-SYMM geography.
Figure 3.4(e): Summer Energy change for a transition from LOW to HIGH orbit (Obliquity cycle)

Figure 3.4(f): Corresponding PDD change for a transition from LOW to HIGH orbit (Obliquity cycle)
Figure 3.4(g): Corresponding PDD change for a transition from LOW to HIGH orbit (Obliquity cycle) in a simulation with NORTH-SYMM geography.

Figure 3.4(h): Corresponding PDD change for a transition from LOW to HIGH orbit (Obliquity cycle) in a simulation with SOUTH-SYMM geography.
Figure 3.5(a): Interhemispheric effect of Southern Hemisphere continental geography on Northern Hemisphere PDD (NHSP orbit - Northern Hemisphere Summer Solstice at Perihelion)

Figure 3.5(b): Interhemispheric effect of Northern Hemisphere continental geography on Southern Hemisphere PDD (NHSP orbit - Northern Hemisphere Summer Solstice at Perihelion)
Figure 3.5(c): Interhemispheric effect of Southern Hemisphere continental geography on Northern Hemisphere PDD (SHSP orbit - Southern Hemisphere Summer Solstice at Perihelion)

Figure 3.5(d): Interhemispheric effect of Northern Hemisphere continental geography on Southern Hemisphere PDD (SHSP orbit – Southern Hemisphere Summer Solstice at Perihelion)
Figure 3.5(e): Interhemispheric effect of Southern Hemisphere continental geography on Northern Hemisphere PDD (HIGH OBLIQUITY orbit)

Figure 3.5(f): Interhemispheric effect of Northern Hemisphere continental geography on Southern Hemisphere PDD (HIGH OBLIQUITY orbit)
Figure 3.5(g): Interhemispheric effect of Southern Hemisphere continental geography on Northern Hemisphere PDD (LOW OBLIQUITY orbit)

Figure 3.5(h): Interhemispheric effect of Northern Hemisphere continental geography on Southern Hemisphere PDD (LOW OBLIQUITY orbit)
Figure 3.6(a): Interhemispheric effect of Southern Hemisphere continental geography on Northern Hemisphere PDD response to Precession cycle (NHSP to SHSP)

Figure 3.6(b): Interhemispheric effect of Northern Hemisphere continental geography on Southern Hemisphere PDD response to Precession cycle (NHSP to SHSP)
Figure 3.6(c): Interhemispheric effect of Southern Hemisphere continental geography on Northern Hemisphere PDD response to Obliquity cycle (LOW to HIGH)

Figure 3.6(d): Interhemispheric effect of Northern Hemisphere continental geography on Southern Hemisphere PDD response to Obliquity cycle (LOW to HIGH)
CHAPTER 4

ECCENTRICITY FORCING AND PRECONDITIONING OF "SUPER-INTERGLACIALS"

4.1 Abstract

Important components of the long-term variability of the earth’s climate have been the alternating glacial and interglacial conditions, which have been present for at least the past 3 million years. Paleoclimate records, like the 3.58 million year long Lake El’gygytgyn record from the Siberian Arctic, identify some interglacial periods as particularly warm (Melles et al., 2012). These “Super Interglacials” include MIS 11c, 31, 49, 55, 77, 87, 91, and 93. Here, we use an ensemble of physically based climate models to show that during periods of extremely low eccentricity, favorable forcing conditions (high obliquity and moderate CO₂), lead to unusually long warm (interglacial) conditions in both Northern and Southern Hemispheres. The prolonged warming in the Southern Hemisphere preconditions the Polar Regions to produce an especially large response during the warm super-interglacials like MIS 31.

4.2 Introduction

An important aspect of the long-term variability of the earth’s climate has been the glacial-interglacial conditions that have been present for at least the past 5 million years (Emiliani & Geiss, 1959; Lisiecki & Raymo, 2005). Glacial-Interglacial cycles are divided into Marine Isotope Stages (MIS) based on benthic proxy records with odd numbers corresponding to interglacials and even numbers corresponding to glacial periods. In this paper, we focus on the orbital forcing leading to spatial and temporal variability in the intensity of the interglacials, characterized by low land ice extent and high sea levels (Berger et al., 2015). The interglacials have been shown to vary in intensity as well as duration. For example, in the past 800,000 years, Berger et al. (2015) showed that Marine Isotope Stages (MIS) 5e (last interglacial) and 11c were warmer than other interglacials during this period, while MIS 13a was cooler than other interglacials.
A sediment core from Arctic Lake El’gygytgyn (Northeastern Russia) revealed certain interglacials to be warmer than others based on sedimentological facies (Melles et al., 2012), with summer temperatures 3-5°C warmer than today. These “Super Interglacials” cannot be explained by a combination of greenhouse gas and astronomical forcing alone, which implies the presence of amplifying feedbacks leading to the onset of these super interglacials. These “Super Interglacials” are: MIS 9, 11, 31, 49, 55, 77, 87, 91 and 93.

4.3 Lake El’gygytgyn

Lake El’gygytgyn is a large, deep basin created by a meteorite impact 3.58 MA (Layer, 2000) on the continental divide of the Anadyr Mountains in central Chukotka, northeastern Arctic Russia (Figure 1). Located 100 km north of the Arctic Circle, the lake today measures 12 km wide and 171 meters deep. Deep drilling by the ICDP led consortium of countries in 2009 recovered a continuous sedimentary record dating to the time of impact (Melles et al., 2011). The complete record has been the focus of two summary papers (Brigham-Grette et al., 2013; Melles et al., 2012) plus a special volume in Climate of the Past. Because of geography and glacial aridity, the Lake El’gygytgyn basin escaped continental scale glaciation producing the longest, continuous terrestrial record of past climate in the entire Arctic (Brigham-Grette et al., 2013; Melles et al., 2012).

Super interglacials in the Lake El’gygytgyn were first identified by Melles et al. 2012 based on sedimentary facies and the co-occurrence of these sediments with other “warm” proxies. In particular, Facies C consists of reddish-brown fine muds with distinct laminations, unlike Facies A associated with cold, perennial ice cover on the lake and Facies B, more massive in character and associated with interglacial conditions. Facies C also commonly occurs with light values in the Lisiecki and Raymo (2005) marine isotope stack, and high regional July insolation. Many of the super interglacials are associated with a return to almost forested conditions around the lake based on palynological studies (Lozhkin et al. 2013, 2016) and high Si/Ti ratios suggesting much higher surface water primary productivity (Brigham-Grette et al., 2013; Melles et al., 2012). All of these
proxies suggest a longer ice free summer season. What remained puzzling about the
distinction between normal interglacial facies B and super interglacial facies C was the
observation that the pacing of superinterglacials appeared to be without a clear pattern
that could be easily associated with a distinct forcing mechanism such as CO₂ (where
data exist, aka Martínez-Botí et al., (2015); Parrenin et al., (2007), etc. or orbital forcing.
What could be the cause of superinterglacial pacing and forcing?
Figure 4.1: Location of Lake El'gygytgyn (star). The lake formed 3.58 Ma in NE Russia, when North America remained a contiguous landmass (and had a more continental climate), unbroken by tectonics and glacially eroded marine troughs.

(From Brigham-Grette et al., 2013)
4.4 Background Theory – Astronomical forcing of interglacials

Several theories have been proposed associating glacial-interglacial cycles with variations in the Earth’s orbital configuration. Milankovitch (Milankovitch, 1941) first provided a comprehensive theory associating the cyclic changes in Earth’s climate to variations in eccentricity, obliquity and precession. According to the Milankovitch’s theory of glaciation, interglacials in the Earth’s past correspond with high obliquity and perihelion coinciding with Northern Hemisphere summer. Milankovitch’s ideas were empirically demonstrated by variations in oxygen isotope ratios in deep-sea sediment cores, showing variations at the same frequency as changes in obliquity and precession of the Earth’s orbit (Emiliani & Geiss, 1959; Hays, Imbrie, & Shackleton, 1976).

During the late Pliocene and early Pleistocene, glacial-interglacial cycles recorded in benthic δ¹⁸O records were dominated by ~41 kyr frequencies, corresponding to variations in orbital obliquity (Huybers & Curry, 2006; Imbrie, Berger, & Shackleton, 1993; Lisiecki & Raymo, 2005). This is an apparent contradiction to Milankovitch’s theory of orbital forcing, which states that precession should be the strongest frequency in glacial-interglacial cycles. Precession determines the seasonal timing of perihelion or aphelion in the Northern (and Southern) Hemisphere, thus controlling the local summer insolation changes at high latitudes. Multiple theories have been given attempting to explain this anomaly (Gildor & Tziperman, 2000; Lee & Poulsen, 2009; Maureen E. Raymo & Nisancioglu, 2003; Maureen E Raymo & Huybers, 2008; Tabor, Poulsen, & Pollard, 2015; Tziperman, Raymo, Huybers, & Wunsch, 2006).

Two opposing theories, which try to explain the obliquity dominance during this time period, have gained prominence. The hypothesis posed by Peter Huybers (Huybers, 2006; Huybers & Tziperman, 2008) suggests that ice-sheets are sensitive to insolation integrated over the duration of summer, instead of summer insolation intensity. Annual ablation is empirically related to an integrated summer insolation metric, which is a function of solar radiation intensity and duration of the summer melting season. According to Kepler’s second law, Earth’s distance from the sun is inversely proportional
to its angular velocity. Consequently, a summer with weak insolation intensity (related to Earth’s distance from the sun) would have a longer duration (related to Earth’s angular velocity), while a stronger summer would correspond to a shorter duration. Therefore, an integrated insolation measure like the sum of Positive Degree Days is insensitive to precessional changes, due to the opposing effects of precession in each hemisphere on intensity and duration. On the other hand, the hypothesis posed by Maureen Raymo suggested that ice-sheet melt is controlled by local summer insolation which is dominated by the 23,000-yr precession period at nearly all latitudes (M E Raymo, Lisiecki, & Nisancioglu, 2006). However, Earth’s orbital precession is out-of-phase between hemispheres, i.e. when the Northern Hemisphere has intense summer insolation the Southern Hemisphere has weak local insolation and vice-versa. Consequently, melting of ice-sheets in one hemisphere could partially balance ice-sheet growth in the other hemisphere, muting precessional cyclicity in globally integrated ice volume proxies. In this paper, we build on these theories to arrive at an orbital hypothesis to explain the occurrence of the “super-interglacials”.

4.5 Low Eccentricity Preconditioning

Globally integrated proxies such as δ¹⁸O record the history of growth and collapse of ice sheets across the world. Global ice variability is best summarized in the “LR04” benthic δ¹⁸O stack (Lisiecki and Raymo, 2005), based on an analysis of 57 globally distributed marine sediment cores. We used a series of Generalized Additive Models (GAM) to study the response of astronomical forcing (eccentricity, obliquity and precession) on δ¹⁸O during late Pliocene - early Pleistocene. In our model, the δ¹⁸O values from the benthic LR04 stack, which is our response variable, is assumed to be some function of astronomical forcing, i.e. eccentricity, obliquity and precession. To model our response variable, we considered the Generalized Additive Model to incorporate the nonlinear forms of the predictors.

The fact that the relationship between astronomical forcing and the response of δ¹⁸O is highly complex is well known. Attempting to fit a standard GAM results in Figure 2(a),
which shows the effects of two predictors, viz. obliquity and eccentricity on δ\textsuperscript{18}O. The model predicts interglacials (lower δ\textsuperscript{18}O values) at higher obliquities and broadly at higher eccentricities, and conversely, low obliquity values are associated with glacial\textsuperscript{s} (higher δ\textsuperscript{18}O values). While our model has low explanatory power, these observations largely agree with Milankovitch’s theory of orbital forcing on glacial-interglacial cycles. However, when we add “lagged eccentricity” and “lagged obliquity” as predictors to our model, we see significant increase in the explanatory power of our model. Here, “lagged” refers to obliquity and eccentricity ~53ka before the observed δ\textsuperscript{18}O value. Different lags of 1-100 kyr were tested to get the highest correlation between the “lagged” orbital parameters and the δ\textsuperscript{18}O values, and a lag of ~53 kyr gave the best fit. It is observed that the deviance explained (which serves as a generalization of R-squared) increased from 13.2% to 35.1%, while the GCV decreased from 0.138 to 0.104 (Generalized Cross Validation score (GCV) is an estimate of the mean square prediction error, and is an efficient measure to compare the two models).

Figure 2(b) captures the effects of the two new predictors viz. “lagged eccentricity” and “lagged obliquity” on δ\textsuperscript{18}O. Along with increased predictive power, this also provides some insight into the intensity of interglacials. For example, in Fig. 2(b), darker colors represent stronger interglacials; reds refer to the “super-interglacials”. We see a strong correlation between low eccentricity/high obliquity at a lag of ~53ka and the lowest δ\textsuperscript{18}O values, corresponding to the super interglacials. Thus, the combination of very low eccentricity (e < 0.01) and high obliquity (> 23.75) has a high probability of being followed by a super-interglacial, albeit after a lag of about 50ka. Indeed, we can observe from the “LR04” δ\textsuperscript{18}O stack and Berger’s orbital forcing that most of the super-interglacials are preceded by a period of low eccentricity (e <0.01) (Figure 3).
Figure 4.2: Response plots from GAM model predicting $\delta^{18}O$ values. (a) Obliquity (x-axis) versus eccentricity on (y-axis), with modeled $\delta^{18}O$ values (model response) given by the contours. (b) “Lagged Obliquity” (x-axis), “Lagged Eccentricity” (y-axis), with modeled $\delta^{18}O$ values (model response) given by the contours. “Lagged Obliquity” and “Lagged Eccentricity” refers to obliquity and eccentricity ~53 ka before the time period being modeled. Lighter colors indicate higher $\delta^{18}O$ values (glacials) and darker colors indicate lower $\delta^{18}O$ values (interglacials). The darkest reds indicate the lowest $\delta^{18}O$ values corresponding to the super interglacials. The figure shows that “Lagged Obliquity” and “Lagged Eccentricity” have greater explanatory power for “super-interglacials” as evidenced by closely spaced contour lines towards lower values of $\delta^{18}O$. 
Figure 4.3: Schematic diagram showing that most “super-interglacials” are preceded by periods of extreme low eccentricity. (a) Variations in the eccentricity component of orbital forcing over the last 2.4 million years together with $\delta^{18}$O values from the “LR04” benthic stack (Lisiecki & Raymo, 2005). (b) Variations in the eccentricity component shifted by 50 kyr together with $\delta^{18}$O values showing how the low eccentricity periods match up with the “super-interglacials”. Periods of extremely low eccentricity are shaded in green, and the “super-interglacials” identified from the Lake El’gygytgyn record are shaded in red.
4.6 An orbital hypothesis to explain low eccentricity preconditioning

During late Pliocene - early Pleistocene, the glacial-interglacial cycles in global proxy records such as the “LR04” stack are dominated by 41-kyr cyclicity, corresponding to the frequency at which Earth’s obliquity varies. Previously, it has been shown using a physically based climate model that the climatic effect of precession is muted in global isotope records due to two different mechanisms, with each dominating as a function of eccentricity (Chapter 1). At low eccentricities (e<0.019), the response of summer temperatures to precessional variations in the intensity of summer insolation is balanced by changes in the duration of summer. At higher eccentricities (e>0.019), the time-integrated summer insolation and number of positive degree-days impacting ice sheets varies at precessional periods, but the variation is out-of-phase between the Northern and Southern Hemispheres.

We extend the discussion from Chapter 1 to arrive at a hypothesis to explain the low eccentricity preconditioning of the super-interglacials. We used a Global Climate Model to simulate the climate response to astronomical forcing for each of the super-interglacial identified in the Lake El’gygytgyn record. In fact, we used a series of 1000 GCM simulations to model the climate and ice-sheet response from 2.0 to 1.0 ma, which covers three of the super-interglacials (MIS 31, 49 and 55) from the Lake El’gygytgyn record. We forced the GCMs with sequential changes in astronomical forcing parameters every 1-kyr using Berger & Loutre, 1991 (Chapter1 - Methods). Rather than using summer temperatures from the GCM simulations as a measure of ice-sheet mass balance, we use a time-integrated insolation metric, Positive Degree Days (PDD), which then takes into account both temperature and the length of the melt season (duration of summer). Using the simulations from the physically based climate model, we calculate the sum of Positive Degree Days as \[ \text{PDD} = \sum_{i} \alpha_i T_i, \] where \( T_i \) is the mean daily temperature on day \( i \), and \( \alpha \) is one when \( T_i \geq 0^\circ\text{C} \) and zero otherwise. The PDD captures...
the extremity as well as the duration of the melt season, and has been shown to be a good indicator of ice-sheet ablation potential (Braithwaite & Zhang, 2000).

It is observed that during periods of extremely low eccentricity, obliquity is the only dominant astronomical forcing impacting high-latitude climate and ice volume; while variations at precessional frequencies are absent. Consequently, the Northern and Southern Hemisphere polar climate response to orbital forcing is synchronous and in-phase with each other (hemispherically in-phase). During these intervals, Antarctic ice-sheets co-vary with Northern Hemisphere insolation. At higher eccentricities, high latitude climate responses become increasingly sensitive to precession; and precession is the dominant astronomical forcing impacting high-latitude climate and ice volume. Consequently, Northern and Southern Hemisphere polar climate response to orbital forcing is asynchronous, leading to interhemispheric out-of-phase variations in Positive Degree Days (hemispherically out-of-phase). During these intervals, Antarctic ice-sheets co-vary with Southern Hemisphere insolation.

For most of the Plio-Pleistocene, eccentricity is above the eccentricity threshold such that there is significant out-of-phase precessional variation in the Earth’s climate response. Only during periods when eccentricity falls below the threshold, i.e. the Earth’s orbit around the sun becomes essentially circular; does Antarctic ice-sheets start following Northern Hemispheric ice-sheets (corresponding to Northern Hemisphere insolation). During this time, local insolation in both Northern and Southern Hemisphere vary simultaneously, responding only to obliquity. As eccentricity increases, i.e. the Earth’s orbit becomes more and more elliptical; and out-of-phase precessional variation in local insolation starts. During this time, Antarctic ice-sheets start responding to local Southern Hemisphere insolation, which now varies asynchronously with the Northern Hemisphere local insolation. In the process of reverting back to Southern Hemisphere local insolation forcing, Southern Hemisphere misses a beat in the glacial-interglacial cycles at the 23-kyr precessional periods. Thus, depending on the prevailing obliquity conditions during the low eccentricity period, Antarctica either has a prolonged glacial or interglacial period, immediately following the low eccentricity orbit. For example, at
~1.13 Ma, eccentricity is below 0.01 and obliquity is high. In this case, Antarctica experiences prolonged warming (interglacial conditions) from about 1.14 – 1.1 Ma (Figure 4-d), which leads to a preconditioning effect for an unusually warm super-interglacial. In another example, at ~1.3 Ma, eccentricity is below 0.01, but obliquity is low. In this case, Antarctica experiences prolonged cooling (glacial conditions), which does not lead to the necessary preconditioning effect required for the super-interglacials. Indeed, it has been observed that low eccentricity orbits precede almost all super-interglacials, but the converse is not true, i.e. super-interglacials do not follow all low eccentricity orbits.

The explanation we propose for this low eccentricity preconditioning of the super interglacials is directly linked to the fact that at low eccentricities, there is a switch in the hemispheric phasing of the Earth’s climate response. This switch provides the necessary orbital forcing for prolonged interglacial climates in both hemispheres, much longer than expected from global oxygen isotope records. The prolonged warm conditions in the Southern Hemisphere preconditions the polar regions to produce a large response during the unusually warm interglacials like MIS 11 or 31.

4.7 Marine Isotope Stage 31 – a “Super-Interglacial”

We consider the case of MIS 31, which is one of the major interglacials of the early Pleistocene (~1.08 to 1.06 Ma). MIS-31 is marked by some of the lowest δ18O values in deep-sea benthic proxy records, which indicate both deep sea warming and reduced global ice volume (DeConto, Pollard, & Kowalewski, 2012). The peak warmth in MIS-31 corresponds to 1.07 Ma, and is preceded by a period of extremely low eccentricity (e = 0.001) at about 1.123 Ma. Prior to the period of low eccentricity (Figure 4-e; T1-T6), eccentricity is above the threshold of in-phase behavior. During this period (T1-T6), the PDD variation is out-of-phase between the Northern and Southern Hemispheres. During each half of a precession cycle (marked by boxes in Figure 4-e), the Northern and Southern Hemisphere climate responses are out-of-phase. When the Northern Hemisphere is undergoing glacial-type conditions (marked by yellow in Figure
4-d; lower than average summer temperatures), the Southern Hemisphere is undergoing interglacial-type conditions (marked by red in Figure 4-d; warmer than average summer temperatures). After each precession half cycle, the glacial-interglacial conditions flip across hemispheres. Therefore, after about 11ka for each precession half cycle, the Northern Hemisphere has interglacial conditions and the Southern Hemisphere has glacial conditions. However, when the eccentricity minimum is reached, both Northern and Southern Hemispheres start varying in-phase (T7-T9), becoming synchronized within about 10 kyr. Once synchronized, both hemispheres start responding in-phase to obliquity. As obliquity is high (T7, T8, T9), both hemispheres experience increased warming. As eccentricity again exceeds the threshold value for anti-phased precessional response, the two hemispheres start responding out-of-phase (T10-T18). So in effect, during T7-T10, both hemispheres experienced prolonged warming. This prolonged warming could have led to the collapse of WAIS and decrease in ice shelves and sea ice around Antarctica, as observed in the Ross Sea (Naish et al., 2009; Scherer et al., 2008). The prolonged warmth leading up to MIS31 ~1.08Ma, combined with extremely strong local insolation during 1.07 Ma could have provided the required one-two punch for the MIS-31 “super-interglacial” and loss of the West Antarctic Ice Sheet.
Figure 4.4: Orbital Forcing and climate Variation from 1.2 – 1.0 Ma (a) Eccentricity, (b) Obliquity (degrees of axial tilt relative to the ecliptic) and Precession (Berger &
Loutre, 1991) (c) Variations in $\delta^{18}$O values from the “LR04” benthic stack (Lisiecki & Raymo, 2005), and (d) Mean summer temperatures for Northern Hemisphere (JJA; upper) and Southern Hemispheres (DJF; lower). Red colors indicate summer temperatures higher than late Pliocene - early Pleistocene averages, while yellow colors indicate summer temperatures lower than the averages over the same period. (e) Positive Degree Days (Variation from Mean, %) for Northern Hemisphere (blue) and Southern Hemispheres (red). PDD is an indicator of ablation and shows the influence of orbital forcing on ice-sheets. Each shaded box refers to one half of a precession cycle, denoted by T1-T19. Each period has a duration of approximately 11 kyr.
CHAPTER 5

CONCLUSION AND FUTURE WORK

5.1 Key Findings

My dissertation details the first comprehensive application of a physically based climate model to a classic, unresolved problem in paleoclimatology: the apparent dominance of 41-kyr (obliquity) cyclicity in late Pliocene-early Pleistocene ice volume records. In my dissertation, I have shown that the 41,000-year variability of Earth’s glacial cycles during the late Pliocene-early Pleistocene may be attributed to two different mechanisms, with each dominating as a function of eccentricity. At low eccentricities (e<0.019), the response of summer temperatures to precessional variations in the intensity of summer insolation is balanced by changes in the duration of summer. At higher eccentricities (e>0.019), the time-integrated summer insolation and number of positive degree-days impacting ice sheets varies at precessional periods, but the variation is out-of-phase between the Northern and Southern Hemispheres. Each mechanism dominates at different times, leading to an attenuation of precessional variability in globally integrated proxy records of ice volume.

Further, I show that this mechanism might hold an insight into the astronomical forcing of “Super-interglacials”. Recently, several interglacials (MIS 9, 11, 31, 49, 55, 77, 87 and 91) have been identified as warmer than others and have been termed “Super-interglacials”. It has been shown that the warmest of these interglacials follow exceptionally low eccentricity periods, with a lag of ~50kyr. The explanation proposed for this low eccentricity preconditioning of the super interglacials is directly linked to the fact that the polar ice sheets respond differently to precessional changes at different eccentricities, as described above. Using a series of GCM and ice-sheet model simulations covering MIS 11 and 31, I show that Southern Hemisphere ice-sheets respond to Northern Hemisphere insolation at lower eccentricities, switching to local
Southern Hemisphere insolation at higher eccentricities. This switch from northern forcing to southern insolation forcing leads to Antarctica missing a beat in its glacial-interglacial cycles, as northern and southern insolation intensities vary out-of-phase at 23 ka precessional periods. Thus, depending on the orbital conditions, Antarctica either has an unusually long glacial or interglacial period following a low eccentricity orbit. In the latter case, the prolonged warm conditions in the Southern Hemisphere preconditions the Polar Regions to produce a large response during the unusually warm interglacials like MIS 11 or 31.

5.2 Future Work

My findings presented above are consistent with simulations from a numerical ice-sheet model, recreating the ice volume changes from both Northern and Southern Hemispheres. If a low eccentricity is prescribed, ice sheets in both hemispheres respond only to obliquity cycle, and grow and melt together. If the ice sheet variability is simulated with a high eccentricity, both hemispheres show dominant precessional variation.

One of the big challenges facing the precession-forced, interhemispheric antiphasing hypothesis is the assumption that precession time-scale ice-volume changes in the Southern Hemisphere can effectively cancel out changes in the Northern Hemisphere, where the ice-volume changes are much larger. To adequately answer this question, ocean isotopic variations from simulated ice-volume changes in each hemisphere should be compared to proxy isotope records from the deep sea. Applying similar modeling approaches as the transient GCM simulation, future work will involve modeling the Northern Hemisphere and Southern Hemisphere ice sheets, time continuously from 2.0 ma through 1.0 ma. While running a fully coupled ice-sheet/shelf model and GCM would be ideal, it is computationally time intensive. Hence, an asynchronous methodology, linking equilibrated GCM simulations with an ice sheet/shelf model every 1000 years between 2.0 and 1.0 Ma, will be used to to account for the simultaneous growth and collapse of the ice-sheets in both hemispheres. Using the simulated ice volumes from 2.0
to 1.0 ma, global benthic $\delta^{18}O$ variations can be empirically calculated based on the assumption that the relationship between changes in ice volume (above floatation) and sea level is linear and symmetric, and that the isotopic signature of the individual ice-sheets has not changed with time. These results allow direct comparison between modeled mean ocean $\delta^{18}O$ values and paleoclimate proxy records such as the LR04 stack. More importantly, will determine if the Antarctic ice sheet’s influence on the mean isotopic composition of the ocean is sufficient to cancel out the isotopic variation contributed by Northern Hemisphere ice sheets.
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