DECIHERING 6 MY OF GREENLAND ICE SHEET HISTORY USING IN SITU ^{10}BE FROM MARINE SEDIMENT CORES

A Thesis Proposal Presented

By

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ABSTRACT

The Greenland Ice Sheet has been reduced in size, perhaps greatly, during past warm intervals. Ice Sheet dynamics during the Plio-Pleistocene are poorly constrained because knowledge of past ice extent is limited to inferences from sea-level indicators and ice rafted debris records - with the former only providing constraints on global ice volume and the latter indicating the timing but not extent of glaciation. My research will measure concentration of in situ^{10}Be in two ocean sediment cores as a different way to infer past ice extent. Cosmogenic ^{10}Be is produced in rocks at or near Earth's surface, which are not covered by ice, and so ^{10}Be accumulation is controlled by extent and duration of ice cover. Sediments eroded from the continent have been preserved in stratigraphically intact ocean sediment cores and thus may provide a record of Greenland Ice Sheet dynamics through time. Measurement of ^{10}Be in continental samples will help explain contemporary exposure, erosion, and transportation, which will serve as an analogy to the past and aid interpretation of the downcore records. I anticipate that the oldest sediments in the cores will contain the highest concentrations of ^{10}Be due to prolonged exposure prior to continental-scale glaciation. ^{10}Be concentration in exported rock and sediment will then presumably decrease through time as the continent was glaciated, but will increase abruptly during brief interglacial periods when the ice sheet was reduced. The records of ^{10}Be concentration will serve as a proxy for ice sheet extent over the past 6 million years, enabling me to better constrain dynamics of the Greenland Ice Sheet through past, and by extension future, climate states.

I. INTRODUCTION

Through my research, I hope to understand extent of ice cover on Greenland over the last 6 My (million years) and how growth and collapse of the ice sheet affects continental erosion and delivery of sediment to the deep ocean. Ice rafted debris (IRD) records show evidence of some glaciation on Greenland since 38 Ma (Eldrett et al., 2007) and a continental scale Greenland Ice Sheet (GIS) is thought to have existed since at least ~3 Ma (Rogozhina et al., 2011), but perhaps much longer (DeConto et al., 2008). Greenland ice cores are difficult to interpret prior to 100 ka due to basal ice deformation so little is known about response of the GIS to climate variations before the last glacial-interglacial cycle (Rogozhina et al., 2011).

This study will use the concentration of in situ ^{10}Be measured in quartz sand extracted from two ocean sediment cores as a proxy for ice sheet extent over the last 6 My. The cores, from the south and east coasts of Greenland, contain sediments eroded from the continent (Figure 1). I anticipate that the oldest core sediments will pre-date the onset of major continental glaciation when the ice sheet was rapidly eroding surficial material, which had been accumulating ^{10}Be prior to ice sheet inception. Thus, the record will likely show an initially high concentration of ^{10}Be. Deciphering changing ^{10}Be concentration through time will be aided by contemporary samples collected from rivers and outcrops on land, which will illuminate sediment exposure, erosion, and transportation during the current and, by analogy, prior interglacial periods.

^{10}Be is a cosmogenic isotope easily measured in quartz (Lal, 1988). In situ production of ^{10}Be occurs in rocks because of terrestrial exposure to cosmic radiation (Lal, 1988; Middleton et al., 1987). If bedrock or sediment is buried by ice, water, or more rock, then the material is shielded from cosmic rays and ^{10}Be is not produced. The cores contain sediment from the continent and thus the ^{10}Be concentration is controlled by a) duration and extent of landscape exposure and b) by the efficiency of erosion and transportation of sediment from the continent to the ocean.

If Greenland was ice-free at the beginning of the record, then concentrations of ^{10}Be will decline through time as the continent was repeatedly glaciated. Pre-glaciation ^{10}Be concentration would be high because the relatively stable landscape would accumulate ^{10}Be through prolonged cosmic-ray exposure. During glacial periods, ice shielding would reduce ^{10}Be accumulation and greater erosion rates would deliver deeper-sourced material containing lower amounts of ^{10}Be. During interglacial periods, ^{10}Be concentration will increase because of landscape exposure, and higher concentrations will represent greater and/or longer ice-sheet reduction (Figure 2).

In order to decipher the ^{10}Be signal in the core, I need to understand the processes through which sediments are eroded from the continent and transported to the ocean. While exposure will occur during periods of deglaciation, maximum erosion and deposition is likely to occur during periods of glacial growth, thus creating a lag between the timing of ^{10}Be production on the continent and the delivery of sediments to the core sites. For example, temporary sediment storage in basins such as lakes and fjords is likely during interglacials and delivery to the deep sea may not occur until the next period of ice advance to the continental shelf.

II. BACKGROUND

2.1. Climate Dynamics

Scientists have reconstructed Earth's climate over the last 65 My using deep-sea sedimentary records (Zachos et al., 2001). Compilation records of $\delta^{18}O$ and $\delta^{13}C$ values from sea-floor foraminifera serve as a proxy for ocean temperature and continental ice volume (Zachos et al., 2001). The Pliocene (~2–5 Ma) is characterized by relatively abrupt cooling as Earth transitioned from the warmer Miocene to the colder Pleistocene (Raymo et al., 2006) with cooling likely driven by plate motion and declining greenhouse gas concentrations (Figure 3). Throughout this cooling trend are periodic shifts between glacial and interglacial states called Milankovitch cycles, which are associated with changes in orbital precession, obliquity, and eccentricity.

2.2. Glacial History of Greenland

Evidence for ice sheet growth on Greenland comes from IRD data, which suggest intermittent glaciation since the Late Miocene (Alley et al., 2010). Ocean Drilling Program (ODP) site 918, which will be used in this study, has a pulse of IRD at ~7.3 Ma, and the next two fluxes occur at ~3.5 Ma and ~2.7 Ma (St John et al., 2002). The mid Pliocene pulse at ~3.5 Ma is in other Arctic records, which show major increases in IRD between 3.5 and 2.4 Ma (Kleiven et al., 2002; Moran et al., 2006; Shackleton et al., 1984). Major Northern Hemisphere glaciation began ~2.75 Ma (Cane et al., 2001; Lisiecki et al., 2007), and pulses of IRD at ~3.5 Ma suggest that expansion of the GIS preceded ice sheet growth over Northern Europe and North America (St John et al., 2002). Furthermore, the location of site 918 supports a proposal by Larsen et al (1994) that southeast Greenland, with high topography and precipitation rates, may have been the nucleation site for the continental-scale GIS (St John et al., 2002).

2.3. The Most Recent Million Years

Changes in the GIS over the last million years have been dominated by 100 ky glacial-interglacial cycles with amplitude of cycles linked to GIS mass balance (Alley et al., 2010). Cycles are numbered according to their Marine Isotope Stage (MIS), with even numbers corresponding to glacial periods and odd numbers to interglacials (Figure 4). Time periods of particular interest are interglacial stages 5, 9, and 11, when extent of the GIS was reduced, perhaps greatly, and glacial stage 6 and 2, when the GIS was at a maximum.

GIS extent during past interglacials can be inferred from changes in eustatic sea level and from the stable isotope composition of seawater (Figure 4), with the caveat that these quantities reflect the total volume of ice on the planet rather than any particular ice sheet. For reference, the modern GIS contains ~6 m sea-level equivalent of ice. During MIS 11 (~400 ka), sea levels were 6-13 m higher than today, indicating the absence or significant reduction of the GIS (Alley et al., 2010; Berger et al., 1991; Hearty et al., 1999; Kaufman et al., 1991; Poore et al., 2001; Raymo et al., 2011). Cosmogenic burial ages from beneath the GISP2 ice core site support complete loss of the GIS during MIS 11 (Nishiizumi et al., 1996). The GIS was also reduced during MIS 5e (120–130 ka) when sea levels were at least 6.6 m higher than today (Kopp et al., 2009), though the GIS may have contributed only ~2 m to this highstand (Colville et al., 2011).

During MIS 6 (~130–188 ka), GIS was at a maximum, with ice extending over areas of southeast Greenland not glaciated today (Funder et al., 1998). Iceberg plow marks and evidence of ice sheet grounding suggest thick ice shelves over the continental shelf, which would have rapidly delivered sediments to the coring site.

2.4. Sediment Budget

A sediment budget refers to “the sources and disposition of sediment as it travels from its point of origin to its eventual exit out of a defined landscape unit (Beylich, 2011).” To make a sediment budget, one must quantify and relate the processes that generate and transport sediment (Dietrich et al., 1978). Sediment budgets are calculated most reliably through long term monitoring (Beylich, 2011) because surface processes are highly variable in time and space and act over long time scales (Dietrich et al., 1978).

Since we will be considering ^{10}Be concentrations over the last 5–6 Ma, we are primarily concerned with understanding how the sediment budget for our study area might change over time. For both core sites, we can assume the lithology of sediments to be constant. The cores receive coarse sediment predominantly from the east Greenland margin and Iceland, and were relatively removed from the influence of the much larger Laurentide Ice Sheet over North America. Lithology is primarily dark gray continental and volcaniclastic silt (Jansen, 1996; Larsen, 1993), but since basaltic volcanic sediments from Iceland will not contain quartz, our analysis will mostly include sediment from Greenland.

Sedimentation rates at site 918 differed during the Pleistocene–Holocene (~8 cm/ky), Pliocene (~20 cm/ky), and Miocene (~2 cm/ky) (Larsen, 1993). The high sedimentation rate during the Pliocene is due in part to abundance of IRD, suggesting that changes in climate influence sediment erosion and delivery rates (Larsen, 1993). Sedimentation rates at site 987 are less well constrained due to limited bio and magneto-stratigraphic markers, but range from 5–20 cm/ky over the last 3 My, with the highest sedimentation occurring during the Pliocene (Jansen, 1996).

Sedimentation rates show how sediment budgets change over million year timescales, but are not sufficiently resolved to show glacial-interglacial variability. Sediment budgets will differ during glacial-interglacial cycles due to changes in erosion rates, storage, and discharge. Glacial erosion is controlled by a number of factors including ice flow due to basal melting, basal water production, normal pressure, and rock strength (Hallet, 1979; Oerlemans, 1984). It is generally thought that erosion rates are higher in glaciated versus deglaciated terrains, however it is not fully understood how erosion rates for a given area might differ between glacial and interglacial states (Harbor et al., 1993). For instance, erosion is driven by basal sliding and erosion rates have been shown to increase during glacial advance (Shaw, 1994), but other studies have shown greater erosion rates during glacial retreat than advance, which is associated with accelerated ice flow due to climate warming (Koppes et al., 2010).

The timing of glacial erosion may not be concurrent with sediment transportation and marine deposition. Unless a glacier is calving directly into the ocean, sediments will be stored within the glacier, moraines, and in proglacial rivers, lakes, and fjords before they are discharged to the ocean (Harbor et al., 1993). In a study of Canadian rivers, contemporary sediment yields are dominated by remobilization of Pleistocene glacial sediments, suggesting a >10 ky lag between glacial erosion and fluvial discharge (Church and Slaymaker 1989). For my study, the lag time between erosion and deposition may be controlled by the duration of interglacial periods. Sediments, which are eroded during one glaciation, may be stored in fjords close to the source area and not evacuated until the next period of glacial growth (Koppes et al., 2010).

2.5. ^{10}Be

^{10}Be is produced in near-surface rocks and sediments by cosmic rays. Concentration of ^{10}Be is easily measured in quartz because of its simple chemistry, relative abundance, and resistance to weathering (Kohl et al., 1992). The ^{10}Be production rate for Greenland at sea level is 3.98 g a^{-1} (Briner et al., 2012), though production rates change over space with depth, altitude and latitude, and through time, due to modulation of cosmic-ray flux by the geomagnetic and heliomagnetic fields (Lal, 1988). Reactions that produce cosmogenic nuclides decrease exponentially with depth and thus production of ^{10}Be is negligible 2–3 m below the surface (Balco, 2011). These various spatiotemporal changes can be scaled so that the production rate for a specific time and place can be calculated using CRONUS-Earth (see http://hess.ess.washington.edu). The relatively long half-life of ^{10}Be, 1.39 My, (Chmeleff et al., 2009; Korschinek et al., 2009) allows ^{10}Be to be preserved in core sediments for millions of years.

Previous Greenland studies have used ^{10}Be to constrain GIS extent during the Last Glacial Maximum (LGM) (~20–25 ka) (Hakansson et al., 2007; Hakansson et al., 2011) as well as patterns and rates of ice retreat after the last glaciation (Corbett et al., 2011). Studies have also inferred erosion rates from ^{10}Be depth profiles (Goehring et al., 2010) and paired bedrock and boulder ^{10}Be concentrations (Corbett et al., 2011). ^{10}Be data suggest that the landscape has eroded differentially during the last glacial cycle with at least 2 m of erosion at some locations (Corbett et al., 2011; Goehring et al., 2010), but only limited erosion at others (Hakansson et al., 2011).

III. METHODS AND RESEARCH PLAN

3.1. Introduction

In order to interpret ^{10}Be concentrations in sediment cores, I need to understand how sediments accumulate ^{10}Be through exposure to cosmic rays and the processes through which the sediments are eroded, transported, and eventually deposited off shore. These processes will differ during glacial-interglacial cycles, and so the most recent cycle, which is well constrained, will be used to understand the time scale at which these processes occur. In order to understand better these processes, rock and sediment samples will be collected from deglaciated areas of southern Greenland (Figure 1). Samples from the ice margin will indicate the concentration of ^{10}Be in sediment coming out of the ice sheet today and ^{10}Be concentrations in river sediment will indicate ^{10}Be production since Holocene deglaciation. A detailed timeline of the sampling and laboratory work can be found in Table 1.

3.2. Sampling

Samples include core sediments, which have already been collected, and rocks and sediments from the continent (Figure 1). In May 2011, 58 samples were collected from the west coast of Greenland from streams draining the GIS as well as non-glaciated drainages. I hypothesize that samples closest to the glacier terminus will contain the lowest amount of ^{10}Be because they will have limited cosmic-ray exposure. Samples from more distal locations, that represent a larger catchment area, will contain higher amounts of ^{10}Be because the landscape from which they eroded has been exposed to cosmic rays since early Holocene ice retreat. Exposure ages of the most distal samples will indicate ^{10}Be accumulation within the basin. Paired bedrock and boulder samples will allow me to infer the timing and pattern of deglaciation.

During the 2012 field season I will collect ~30 additional samples. Some 2011 sites will be re-sampled to determine temporal variability of ^{10}Be in stream sediments. Additional drainages on the east coast have been identified using Google Earth and will be sampled to provide drainage data more proximal to the coring sites.

3.3. Cores

The two cores that will be used in this study were collected by the Ocean Drilling Program (ODP) off the eastern coast of Greenland (Figure 1). Core 918A was collected in October 1993, is 332.7 m long, and spans the entire Plio-Pleistocene (Larsen, 1993). The core has been sampled at ~1.5 m intervals, from which I will create a low-resolution (30 amalgamated samples) record of ^{10}Be concentration over time. I also have smaller, high-resolution samples from the last ~500 ky of the core, from which collaborator Jeremy Shakun and I will create a planktic $\delta^{18}O$ record to develop a precise chronology for the late Pleistocene. After I have the $\delta^{18}O$ record, I will order an additional 60 samples (for ^{10}Be analysis) that span glacial-interglacial cycles.

ODP core 987 was recovered in August 1995, is 173.01 m long, and dates to the Pleistocene/Late Pliocene (Jansen, 1996). ODP staff will sample core 987 at a low resolution (30 samples) so that I can compare records from these two locations.

3.4. Lab Work

Processing has begun for core and field samples. Quartz grains are isolated by preferential dissolution of other minerals, which is done in sonicated acid baths, using methods adapted from Kohl and Nishiizumi (1992). Beryllium extraction is in progress in the UVM cosmogenic isotope laboratory (see www.uvm.edu/cosmolab). Measurements of the $^{9}Be/^{10}Be$ ratio for the first batch of samples have been made on the Accelerator Mass Spectrometer (AMS) at LLNL (Lawrence Livermore National Laboratory).

3.5. Data Analysis

From the cores, I will create decay-corrected records of ^{10}Be concentration using core age models (Jansen, 1996; Larsen, 1993), which will represent changes in continental ^{10}Be accumulation, and thus ice cover, over time. These ^{10}Be time series will be compared to climate records (Alley et al., 2010; Lisiecki et al., 2007) that span the last 5–6 My to better understand how Greenland responded to past climate changes. I expect ^{10}Be concentration to decrease as the land surface was shielded by ice and glacial erosion delivered deeper-sourced sediment to the ocean. Depending on how quickly the ice sheet expanded, I may see this as an abrupt or a more gradual change.

I also expect to detect abrupt increases in ^{10}Be concentration associated with some interglacial periods, such as MIS 5, 7, or 11 (Figure 4). My high-resolution sampling strategy for the most recent million years will allow me to document these brief but important changes in concentration. The amplitude of concentration increases should be controlled by the extent and duration of ice sheet loss. If I detect differing changes in concentration among the interglacial periods, this will indicate variable GIS responses to past warming. For example, an interglacial period with total ice loss will be represented by a greater change in ^{10}Be concentration over a longer time period than an interglacial during which the GIS was only partially reduced.

My analysis will only be feasible if the cores contain measurable concentrations of ^{10}Be. AMS results for the contemporary continental samples indicate that this is the case for sediments coming out of the GIS today. Additionally, a simple forward model of Greenland ^{10}Be concentration driven by the benthic $\delta^{18}O$ record indicates that ^{10}Be concentrations will be detectable in sediments that are several million years old (Figure 5). The model also shows the expected decrease in concentration over time with GIS expansion and suggests that potential ^{10}Be excursions during late Pleistocene interglacial periods will be relatively small but still detectable (Bierman, 2011).

My record will provide a continuous reconstruction of GIS variability over the past 6 My, adding to the limited knowledge of GIS extent through the Plio-Pleistocene. Depending on the record's length, I may provide a new way of dating the onset of continental-scale glaciation on Greenland. Furthermore, my detection of abrupt increases in ^{10}Be concentration over the past million years will allow me to infer the relative extent of the GIS during past interglacial periods. My record should thus provide valuable information about the dynamics of the GIS through past climate states and changes, helping to constrain climate-ice sheet models, shed light on its sensitivity to warming, and ultimately aid in projections of its response to ongoing and future climate change.

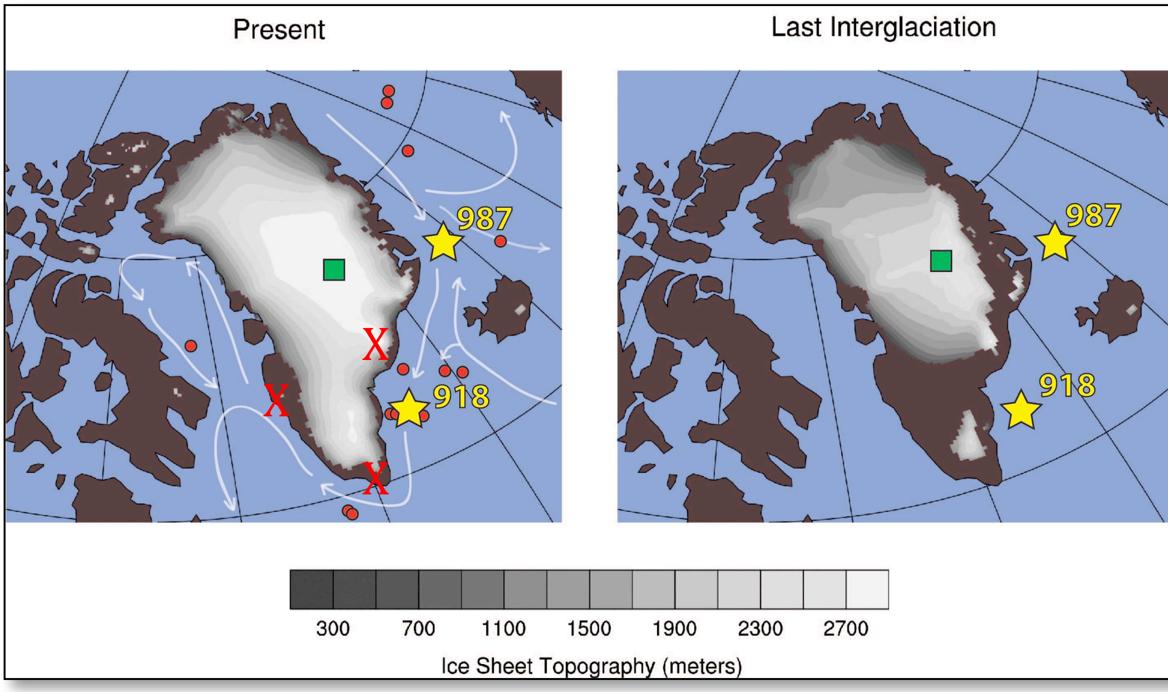

Figure 1: Height and extent of the GIS today (left) compared to the modeled height and extent during the last interglacial (130 ka) (Otto-Bliesner et al., 2006). Yellow stars indicate locations of the ODP coring sites, and red dots are other ODP and Deep Sea Drilling Program core sites. A red X shows the contemporary sampling locations and the green square is the site of the GISP2 ice core at Summit, Greenland. White arrows show modern currents, adapted from (Bierman, 2011).

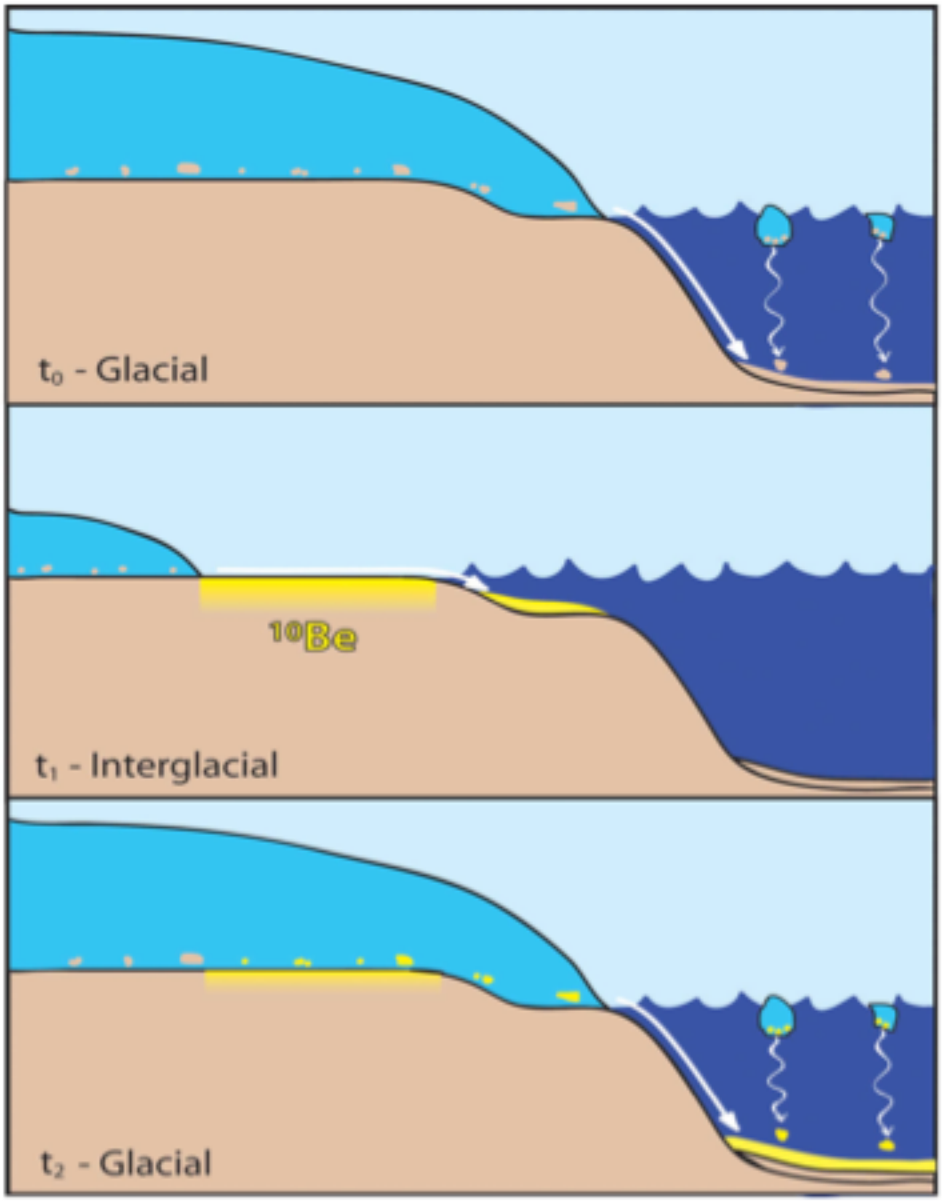

Figure 2: This schematic illustrates ^{10}Be accumulation and subsequent off shore deposition during different climate states (Bierman, 2011). At t_0, the continent is fully covered by a glacier and ice shielding prevents accumulation of ^{10}Be in continental sediments. Sediments eroding off the continent do not contain any ^{10}Be. At t_1, the glacier has receded and ^{10}Be accumulates at and near the surface of landscapes not covered by ice. During this time, eroded sediments containing ^{10}Be may be trapped in basins such as fjords. In t_2, the glacier advances, eroding sediments that have accumulated ^{10}Be. During initial stages of glacial growth, sediments, which were stored in basins during the previous interglacial, are evacuated to the coring site in the deep ocean. In glacial periods, sediments will also be delivered to the deep ocean via ice rafting.

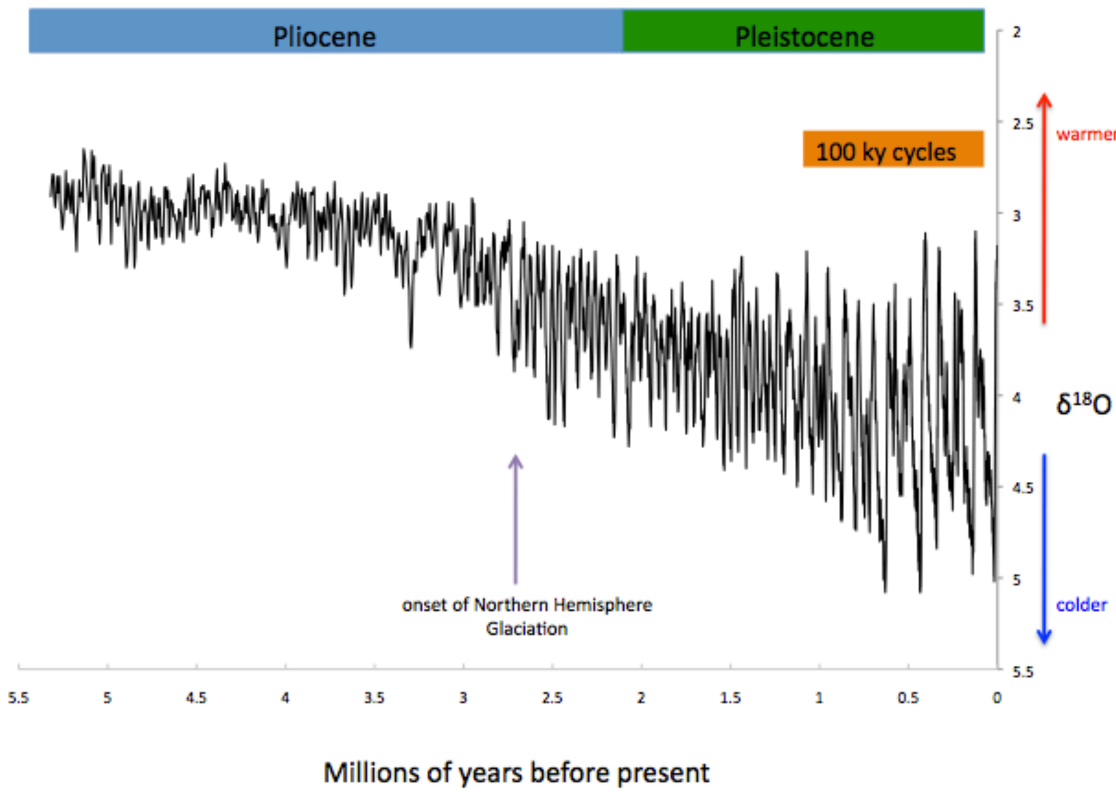

Figure 3: Oxygen isotope data for the last 5.3 Ma averaged from 57 globally distributed benthic $\delta^{18}O$ records (Lisiecki et al., 2005). Higher $\delta^{18}O$ values indicate greater continental ice volume and/or colder deep ocean temperatures. Northern Hemisphere glaciation began at ~2.75 Ma and climate over the last one million years has been dominated by one hundred thousand year glacial-interglacial cycles.

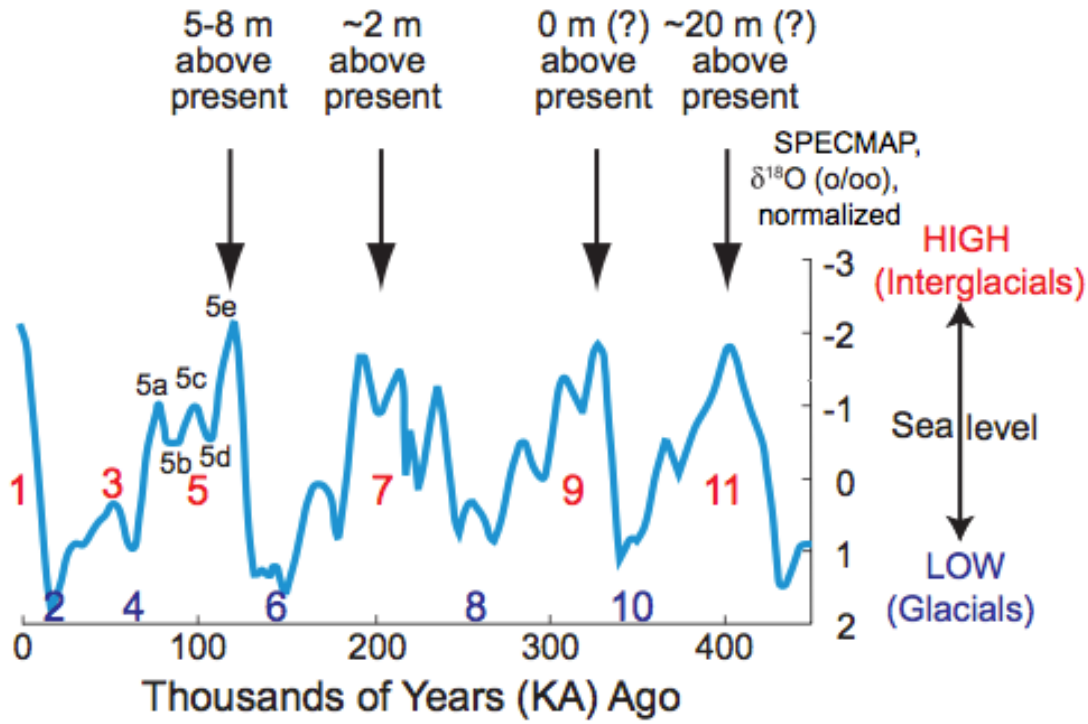

Figure 4: Oxygen-isotope data from the SPECMAP project, a standard chronology for oxygen isotope records (Imbrie et al., 1982). Numbers identify Marine Isotope Stages (1-11) and sea-level indications are given for MIS 5, 7, 9, and 11 (Alley et al., 2010)

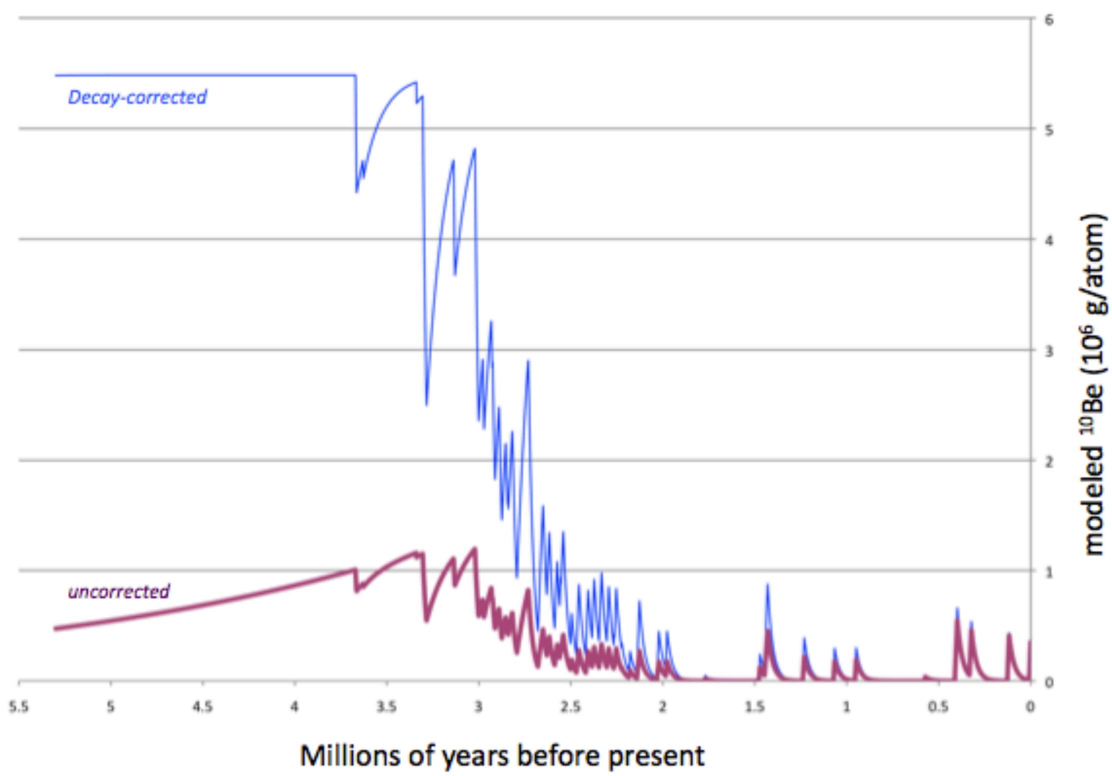

Figure 5: Modeled marine core in situ ^{10}Be record of GIS variability. The uncorrected line shows the expected ^{10}Be concentrations as measured by Accelerator Mass Spectrometry. The decay-corrected line represents the ^{10}Be concentration at the time of deposition. Modeling shows decreasing ^{10}Be values as the continent became glaciated and abrupt increases in ^{10}Be during interglacials. Data from (Bierman, 2011). The model was driven by the global benthic d18O record (see Figure 3). It has Greenland ice covered (ice free) whenever benthic d18O is above (below) early Holocene values, assumes 20 m/My (5 m/My) erosion rates during ice covered (ice free) intervals, and assumes instantaneous erosion and transport of Greenland surface sediment to the deep sea.

Timeline for sampling and laboratory work

• May 2011	• Samples collected from drainages and boulder/bedrock pairs along the east coast of Greenland
• Summer/Fall 2011	• Quartz extraction completed for the summer 2011 samples
• Fall 2011	• The first half of the summer 2011 samples were processed in the cosmogenic lab and the first round of measurements were made on the AMS at LLNL
• February 2012	• Testing of the quartz extraction process on three core test samples to ensure adequate yields
• March 2012	• Completed processing of the remaining summer 2011 samples and made AMS measurements at LLNL • Begin quartz extraction of core samples in the mineral separation lab
• June 2012	• Summer field work on the Southeastern and Southwestern coasts of Greenland to collect contemporary sediment samples
• July-August 2012	• Quartz extraction for summer 2012 field and core samples will be completed in the mineral separation lab by the end of summer
• September-November 2012	• Cosmogenic lab processing for all field and core samples will be completed by the end of the fall
• December 2012	• All remaining AMS measurements will be made by the end of the fall semester

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